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Proceedings of the VIIIth RCMNS Congress

# RCMNS RESEARCH ACTIVITIES

by

J. E. MEULENKAMP

Research goals of the RCMNS, set during the 7th International Congress on the Mediterranean Neogene, Athens, 1979, have become firmly established now. This may be concluded from the success of four interim-colloquia which were held in the time-span between the Athens' and Budapest' Congresses, as well as from the number and contents of contributions presented at the Budapest Congress. The interim-colloquia were meant to serve as preparatory meetings where specialists discussed problems pertinent to the three major topics the Executive Council and the Organizing Committee selected for the scientific programme of the Congress. These topics included Relative and Numerical Time scales, Geohistory of the Mediterranean and the Paratethys, and Paleoecology—Ecostratigraphy.

From the discussions during the interim-colloquia and the Congress we learned that further refinement of biostratigraphic and chronostratigraphic scales and the precise and accurate calibration of these scales with numerical time should remain the prime target of future RCMNS research activities. Irrespective of the progress made during the last decades through the joint efforts of the RCMNS and various IGCP projects, we must admit that we still face major problems as soon as we try to perform high-resolution correlations with a satisfactory degree of accuracy and precision between marine, as well as between marine and continental sequences for parts of the Neogene record. Other problems concern the definition of some stage-boundaries in connection with the designation of boundary-stratotypes.

The increased interest members of our committee show in problems pertinent to paleogeography and geodynamics is reflected by the high number of contributions to the topic Geohistory of the Mediterranean and the Paratethys and by their active participation in the IGCP Project 25 "Correlation Tethys—Paratethys Neogene", the IUGS Research Development Programme "Neogene Paleogeographic Map Series of Central and Eastern Europe" and the project "Neogene Sedimentological Map Series" of the Carpatho—Balkan Geological Association. The results of these multidisciplinary projects unambiguously demonstrate the crucial role pure stratigraphic research plays in any attempt to unravel the complex histories of the Neogene basins in the Mediterranean—Paratethyan realm. On the other hand, models accounting for the time and space-bound character of specific sediment successions as a function of the geodynamic evolution of the area may contribute to the understanding of the processes which control the distribution of time and environment-diagnostic faunal and floral elements. This, in turn, could lead to the further development of ecostratigraphic concepts.

Ecostratigraphy has become the central theme of RCMNS activities over the last years, which confirms the research trends we expected in response to the results of the Athens' Congress. Substantial progress has been achieved with respect to the effects of climatic changes on the distribution and the composition of fauna and flora in marine as well as in continental environments. The signals inferred from the biorecord, however, are not always consistent and interpretations from different sources cannot possibly be reconciled without a critical reevaluation of the methods applied. Improved methods to discriminate between the effects of local. Mediterranean-wide and global signals will no doubt enhance the reliability of correlations based on paleoclimatic criteria. This not only holds for the biorecord, it is equally true for stable isotopes and clay minerals. It is self evident that climatic changes alone do not account for the major and recurrent environmental changes we can infer from the Neogene record. Regional changes in basinal settings, which are as a rule tectonically-controlled, may provoke environmental changes, the visible effects of which on biomasses and on abiotic parameters can serve as a tool for time stratigraphic correlations on a Mediterranean-wide, or even a still larger scale. This is, for instance, also illustrated by the results of the SNS Working Group on Benthic Foraminifera, which results reveal parallel morphological trends in the development of Uvigerinids in the Mediterranean proper, the Paratethys and the North Sea Basin/Northern Atlantic.

The present state of knowledge on the Neogene seems an excellent starting point to initiate pioneer studies aiming at the establishment of models for the origin, accumulation and distribution of mineral resources as a function of the effects of geodynamic and environmental processes. With this in mind a special colloquium on European Late Cenozoic Mineral resources was organized. The large number of contributions presented during the plenary sessions and in workshops reflected the interest of both the scientific and the industrial world in this field of research. Tentative models on the origin and distribution of sedimentary resources, metallic ore deposits and hydrothermal energy were proposed, which may be considered highly promising for the planned continuation of this type of integrated research. Unfortunately, however, only a small number of papers on this topic could be included in the Congress Proceedings, since many authors felt that their ideas needed further elaboration before beeing published.

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Proceedings of the VIIIth RCMNS Congress

# THE ROLE OF NEOGENE DEPOSITS AMONG THE MINERAL RESOURCES IN HUNGARY

by

V. DANK

The successful exploration of Neogene mineral resources in Hungary required the knowledge of geodynamics, palaeogeography and basin development. This knowledge could be obtained by the combined use of the evidence provided by biostratigraphy, geophysics, geochemistry, magnetostratigraphy, radiometry, volcanology and other disciplines.

The present paper deals with the final results of this joint effort.

## Hydrocarbon geology

In the central part of the Pannonian basin Neogene deposits overlie Palaeo—Mesozoic and Paleogene sediments. They play an important role in hydrocarbon exploration and production in Hungary.

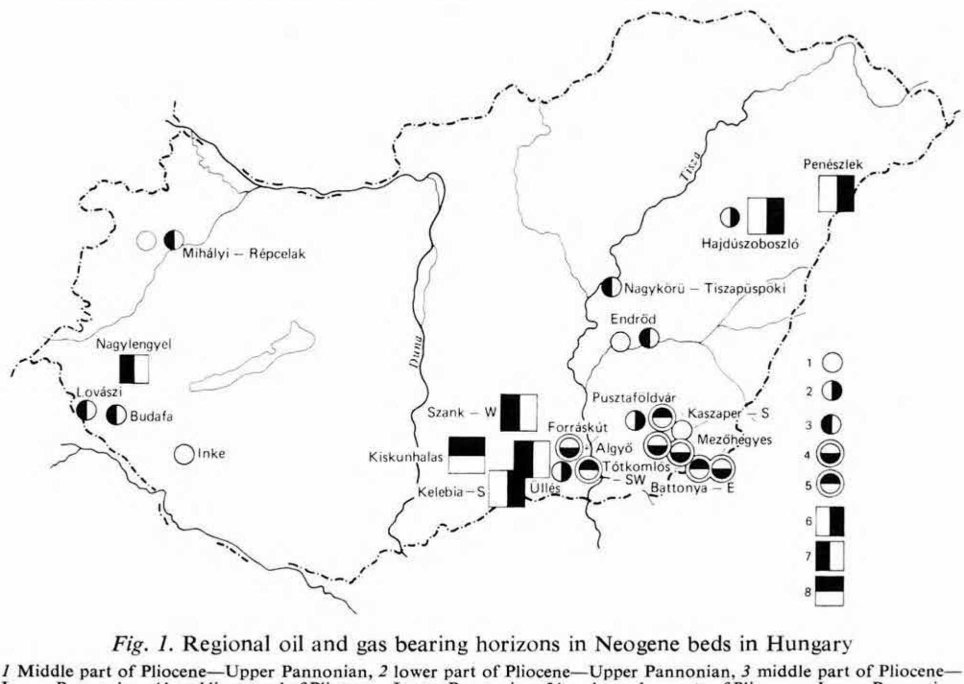


*Fig. 1.* Regional oil and gas bearing horizons in Neogene beds in Hungary

*1* Middle part of Pliocene—Upper Pannonian, *2* lower part of Pliocene—Upper Pannonian, *3* middle part of Pliocene—Lower Pannonian, *4* basal limy marl of Pliocene—Lower Pannonian, *5* basal conglomerate of Pliocene—Lower Pannonian. —*Miocene:* *6* Sarmatian beds, *7* Badenian beds, *8* Karpatian beds

Several oil and natural gas bearing horizons could be identified in the Neogene sequences (from top to bottom):

— in the middle third of the Pliocene—Upper Pannonian sequence the horizon of gas pools (Fig. 1, 1);

— in the lower third of the Pliocene—Upper Pannonian sequence and at the top of the Lower Pannonian the horizon of the oil and gas pools, mostly sandstone (Fig. 1, 2);

— in the middle third of the Pliocene—Lower Pannonian the horizon of oil and gas pools in sandstone layers (Fig. 1, 3);

— Pliocene fissured calcareous marl—oil bearing horizon (Fig. 1, 4);

— Pliocene basal conglomerate horizon of gas and oil pools (Fig. 1, 5);

— Sarmatian conglomerate—limestone—calcareous marl, horizon of oil (Fig. 1, 6);

— Badenian sandstone—conglomerate—breccia horizon of gas and oil pools (Fig. 1, 7);

— Karpatian clastic sequence horizon of gas pools (Fig. 1, 8).

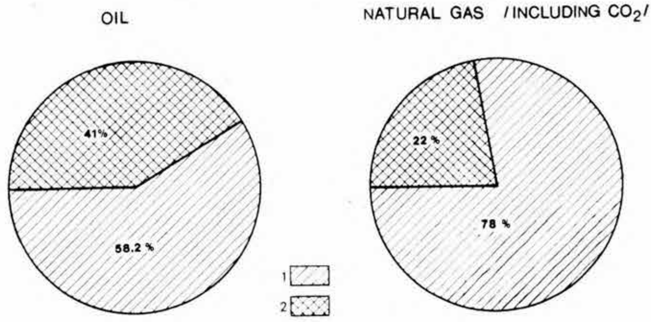


*Fig. 2.* Distribution of HC-reserves on the basis of their geological age
*1* Neogene, *2* pre-Neogene

The pelitic sediments of the Neogene sequence can be considered, in most cases, the source rocks of the hydrocarbon pools. The total initial hydrocarbon resources of the country, can be put to 538 million ton (Mt), 41.8 per cent of which is oil, while 58.2 per cent is natural gas.

In the basin regions 77% of the country's territory, altogether 12 Neogene and one Paleogene subbasins can be distinguished. In the decreasing order of the known initial hydrocarbon resources these are the Fig. 3. Further Hungarian possibilities are shown on the Table 1—3.

Finally, the hypotethic-speculative resources of the previous Neogene basins will be shown, with respect to the oil—natural gas ratio, in function of the age of the possible hydrocarbon bearing sequences (Table 4).

The distribution of the hypothetic-speculative hydrocarbon resources between the Neogene and Pre-Neogene reserves, without the Paleogene and undiscovered Neogene reserves, in the function of depth is the Table 5 (in per cent).

The basins, according to their hypothetic-speculative resources can be arranged according to the following order:

*1* Nagykunság 15.2%, *2* Kiskunság 13.2%, *3* Békés 11.6%, *4* Paleogene 10.7%, *5* Zala—S Balaton 8.8%, *6* Bihar 7.6%, *7* N Great Hungarian Plain 7.5%, *8* Hajdúság

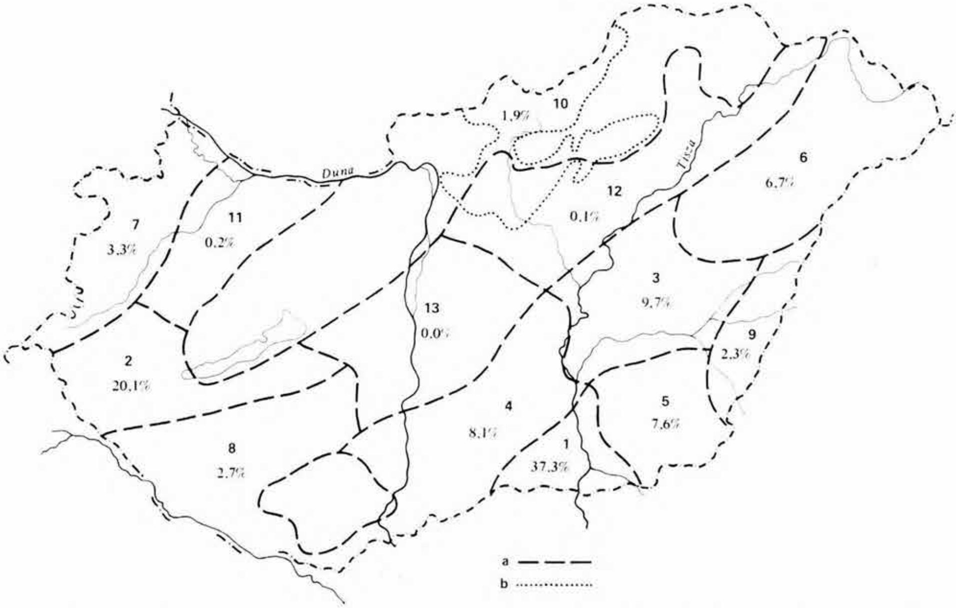


*Fig. 3.* Ranking of Hungarian subbasins on the basis of their original oil and gas in place reserves
*Subbasins: 1* Szeged, *2* Zala—S Balaton, *3* Nagykunság, *4* Kiskunság, *5* Békés, *6* Hajdúság, *7* W Little Hungarian Plain, *8* Somogy—Dráva valley, *9* Bihar, *10* Paleogene, *11* E Little Hungarian Plain, *12* N Great Hungarian Plain, *13* Duna.— *a* Boundary of Neogene subbasin, *b* boundary of Paleogene subbasin

6.3%, *9* E Little Hungarian Plain 4.6%, *10* Somogy—Dráva valley 4.4%, *11* Neogene (subtile) 4.4%, *12* Szeged 3.0%, *13* Danube 1.3%, *14* W Little Hungarian Plain 1.2%.

The majority of the oil and natural gas produced in Hungary comes from Neogene sequences.

In 1984 2.0 Mt of oil and 7.1 $Gm^3$ of natural gas was produced in Hungary. 79.74% of the oil and 80.33% of the natural gas came from Neogene hydrocarbon bearing sequences. Since the starting of the industrial scale production in 1937 altogether 60.1 Mt oil and 102.6 $Gm^3$ natural gas was exploited (1st January 1985) (Fig. 4).

**Distribution of the known and hypothetical speculative hydrocarbon resources according to stratigraphic units**

*Table 1*

| Stratigraphic unit | Initial potential resources (%) | Known reserves (%) | Hypothetical-speculative resources (%) |
|---|---|---|---|
| Neogene | 100 | 67.3 | 32.7 |
| Pre-Neogene (Mz + Pz) | 100 | 45.7 | 54.3 |
| Paleogene | 100 | 31.6 | 68.4 |
| Neogene (subtile traps) | 100 | — | 100.0 |
| Hungary's total | 100 | 66.5 | 33.5 |

**Distribution of the different types of hydrocarbon reserves according to geological ages**

*Table 2*

| Stratigraphic unit | Initial potential resources (%) | Known reserves (%) | Hypothetical-speculative resources (%) |
|---|---|---|---|
| Neogene | 58.6 | 69.8 | 44.1 |
| Pre-Neogene (Mz+Pz) | 32.6 | 26.4 | 40.7 |
| Paleogene | 6.8 | 3.8 | 10.8 |
| Neogene (subtile traps) | 2.0 | — | 4.4 |
| Hungary's total | 100.0 | 100.0 | 100.0 |

**Distribution of the quality of hydrocarbon reserves to be discovered**

*Table 3*

| Stratigraphic unit | Initial potential resources (%) | Known reserves (%) | Hypothetical-speculative resources (%) |
|---|---|---|---|
| Neogene | 32.1 | 54.0 | 43.3 |
| Pre-Neogene (Mz+Pz) | 3.3 | 5.5 | 1.9 |
| Paleogene | 23.6 | 1.5 | 3.1 |
| Neogene (subtile traps) | 41.0 | 39.0 | 51.7 |
| Hungary's total | 100.0 | 100.0 | 100.0 |

**Hypothetic-speculative hydrocarbon resources of Neogene basins according to stratigraphic units**

*Table 4*

| Name of the Neogene basin | OIL | | NATURAL GAS (incl. $CO_2$) | |
|---|---|---|---|---|
| | NG-series % | PRENG-series % | NG-series % | PRENG-series % |
| *1* W Little Hungarian Plain | 2.6 | 97.4 | 2.7 | 97.3 |
| *2* E Little Hungarian Plain | 7.4 | 92.6 | 23.7 | 76.3 |
| *3* Zala—S Balaton | 18.0 | 82.0 | 85.9 | 14.1 |
| *4* Somogy—Dráva valley | 52.5 | 47.5 | 43.6 | 56.4 |
| *5* Danube | 42.4 | 57.6 | 34.1 | 65.9 |
| *6* Kiskunság | 21.5 | 78.5 | 48.5 | 51.5 |
| *7* Szeged | 74.9 | 25.1 | 72.2 | 27.8 |
| *8* N Great Hungarian Plain | 64.3 | 35.7 | 61.0 | 39.0 |
| *9* Nagykunság | 72.4 | 27.6 | 69.5 | 30.5 |
| *10* Békés | 82.0 | 18.0 | 79.8 | 20.2 |
| *11* Hajdúság | 81.1 | 18.9 | 78.8 | 21.2 |
| *12* Bihar | 25.4 | 74.6 | 22.8 | 77 |

**Hypothetic—speculative hydrocarbon resources in function of the depth of occurence**

*Table 5*

| Depth range | Neogene % | Pre-Neogene % | Total % |
|---|---|---|---|
| 0—1500 m | 11.1 | 16.0 | 13.4 |
| 1500—3000 m | 52.0 | 52.8 | 52.4 |
| Deeper than 3000 m | 36.9 | 31.2 | 34.2 |
| | 100.0 | 100.0 | 100.0 |

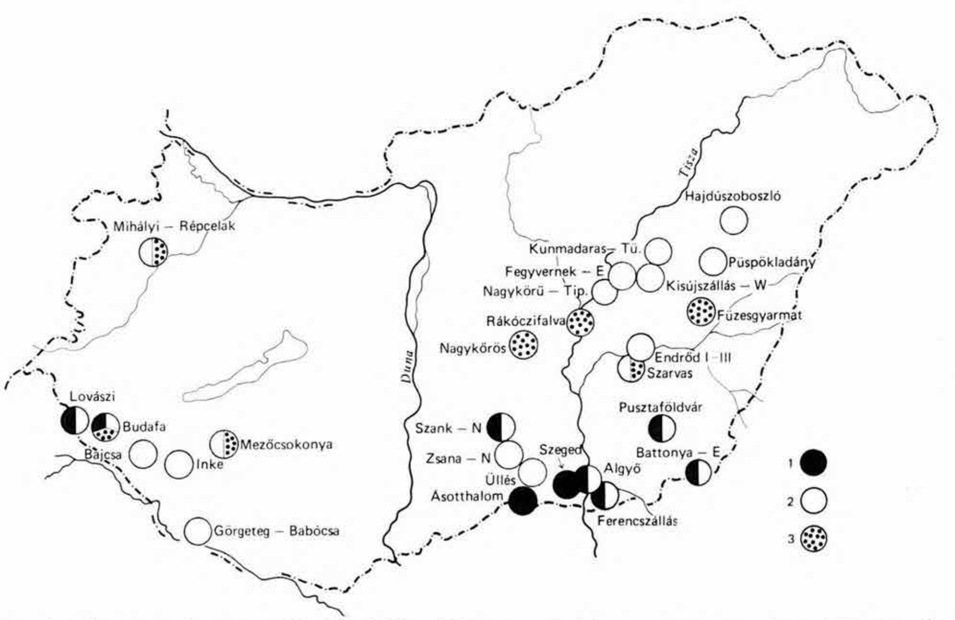


*Fig. 4.* Oil, gas and carbon dioxide fields of Hungary [with reserves more than 1 Mt ($Gm^3$)] in Neogene beds

*1* Oil, *2* natural gas, *3* carbon dioxide

During this period the oil production from Neogene sequences was 63.34%, while that of the natural gas reached the 95.40%.

In the next section the occurrences are enumerated where the Neogene hydrocarbon resources of the deposits exceed the 1 Mt ($Gm^3$) initial reserves (according to the data of 1st January 1984).

*Oil:* Szank (Szank-W), Lovászi, Budafa—Kiscsehi, Ferencszállás, Szeged, Ásotthalom, Pusztaföldvár, Battonya, Battonya-E, Algyő. — *Natural gas*: Inke, Mihály—Répcelak, Zsana-N, Görgeteg—Babócsa, Szank (Szank-W), Mezőcsokonya, Bajcsa, Lovászi, Budafa—Kiscsehi, Püspökladány, Szarvas, Kisújszállás-W, Fegyvernek—Fegyvernek-E, Nagykörü—Tiszapüspöki, Endrőd I—III, Ferencszállás, Hajdúszoboszló, Pusztaföldvár, Battonya, Üllés-basement, Tatárülés-E, Battonya-E. — $CO_2$: Mezőcsokonya, Nagykőrös, Budafa-basement, Mihályi—Répcelak, Füzesgyarmat, Rákóczifalva, Szarvas.

**Geology of brown coal and lignite**

The Hungarian Neogene brown coals and lignites occur in five basins (Fig. 5):
- the Borsod–Ózd (Ottnangian brown coal, ca 425 km²);
- the Nógrád (Ottnangian brown coal, ca 335 km²);
- the Várpalota–Herend (Badenian brown coal, ca 60 km²);
- the Cserhát–Mátra–Bükkalja (Upper Pannonian lignite, ca 485 km²);
- The Torony (Upper Pannonian lignite, ca 117 km²).

There are some more smaller Miocene brown coal occurrences that already have no registered coal reserves or the reserves of which are insignificant:
- the Brennberg deposits (Ottnangian);
- the Hidas deposits (Badenian).

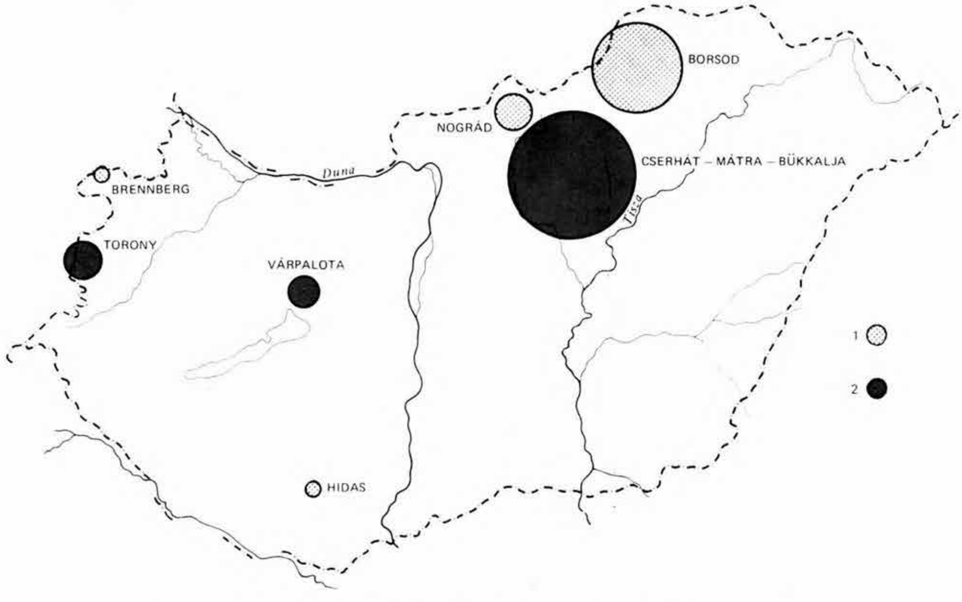


*Fig. 5.* Coal deposits of Neogene age in Hungary
*1* Brown coal, *2* lignite

The quantity of the Neogene coal reserves, and the extent of their exploitation, as compared to the total amount of the coal reserves in Hungary, is shown on Fig. 6.

The following conclusions can be drawn:
- almost three fourth (71–72%) of both the coal reserves and that of the industrial coal reserves can be found in the Neogene sequences;
- the prognostic possibilities of the Neogene coals are very significant (71.7%).

Within the annual 25 Mt coal production in Hungary the share of the Neogene coals is important (Fig. 7).

Accordingly,
- more than half of the coal produced in Hungary is of Neogene age (54.7%);
- 43.6% of the heat value is supplied by Neogene coals;
- the majority of the Neogene coals is exploited by open pit methods.

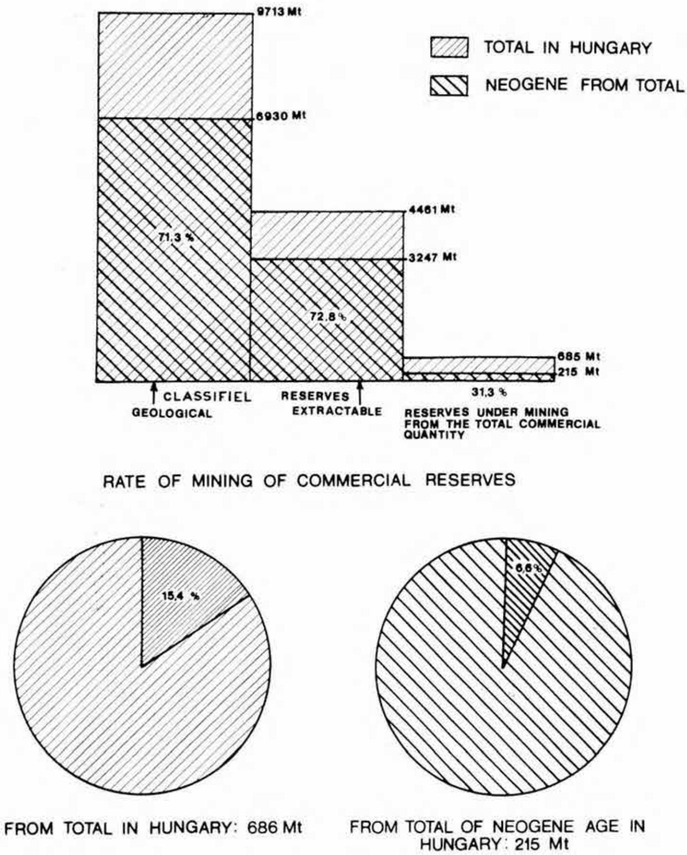


*Fig. 6.*
The quantity of Neogene coal deposits in Hungary and rate of their mining

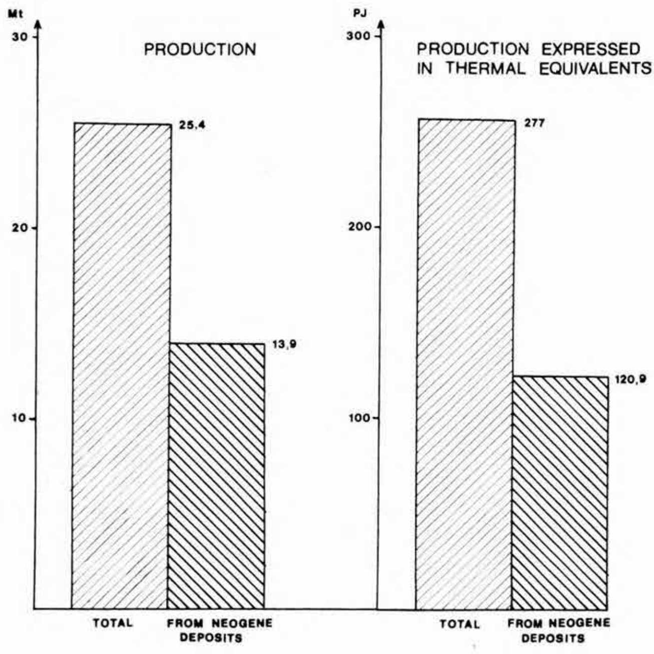


*Fig. 7.*
Coal production in 1984

**Non-metallic mineral resources**

Hungary is comparatively rich is non-metallic mineral resources (Fig. 8). The most valuable group of these resources occur in the *Tokaj* Mountains: noble clays (kaolinite, bentonite, illite, alunite), potassium tuff, perlites (perlite, pumicite), and tuffs containing zeolite, that are ever increasingly utilized in agriculture and food industry.

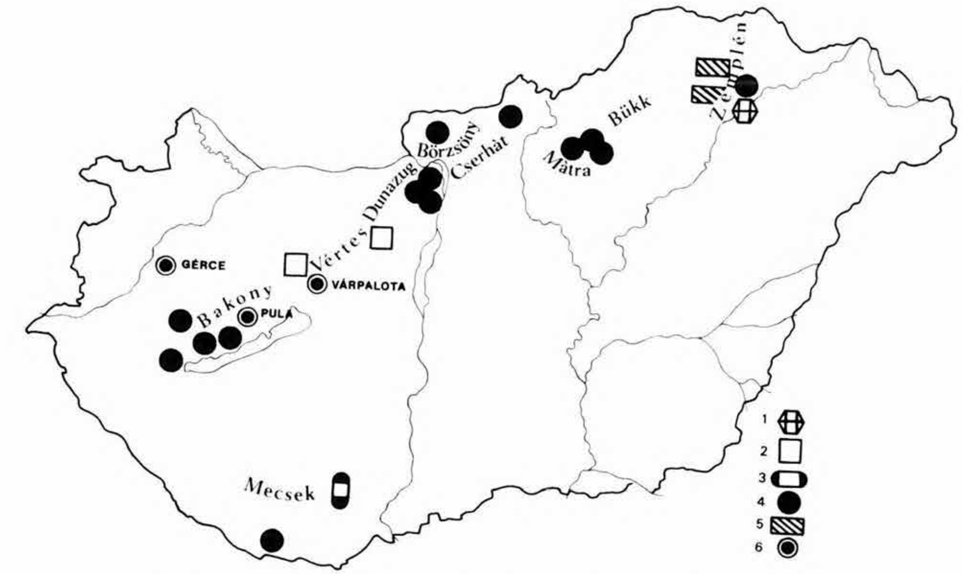


*Fig. 8.* Important mining sites of industrial minerals in beds of Neogene age in Hungary

*1* Rare clays, *2* siliceous sand, glass sand, foundry sand, *3* feldspar-bearing sand, *4* building and decorative stones, *5* ceramic clay, *6* amelioration material

In the *Bakony* and *Vértes* Mountains different siliceous sands (glass sands, soulding sands, and polishing sands (quartz sandstone, soft limestone for preparing coloured earth, etc. are exploited.

In the foreground of the *Mecsek* Mountains feldspar rich sand is stripped.

The Neogene ornamental and building stones occur mainly the volcanic mountain (Balatonfelvidék, Dunazug Mts, Börzsöny Mts, Cserhát Mts, Mátra Mts, Tokaj Mts). The reserves of andesite, dacite, basalt, and rhyolite rocks and their tuffs are at the disposal of the building industry in a quantity of milliard tons. A special ornamental-building stone is the Pannonian "card-type stratified" sandstone and the pearl gravel (Hévíz, Tapolca basin). Pannonian clay for coarse ceramic purposes (brick, construction ceramics, tiles) is available. Northern Hungary the Miocene (Eggenburgian and Karpatian) schlier is also suitable for brick making.

The bulk of the Neogene mineral raw materials for a melioration purposes are offered by the unconsolidated Leithakalk (ca 43%) and by the recently discovered Pannonian oil shale, the alginite. The occurrences of the latter at Gérce, Pula, Várpalota and in the Cserhát Mts provide an exploitable reserve of 100 Mt, and the prospective reserves are estimated to the order of magnitude of milliard tons.

The share of the Neogene resources among the explored and registered non-metallic mineral resources in Hungary amounts to a total of 30.4% (3.3 Mrd t). Of these, 47.4% can be exploited by open-cast mining.

### The economic significance of Neogene ore mineralization

Ore mineralization is bound to Neogene volcanites:

– in the Börzsöny Mts (Badenian andesites),

– in the Mátra Mts (Badenian andesites),

– in the Tokaj Mts (Sarmatian andesites).

Classified polymetallic ore reserves, can be found only at one site, in the Gyöngyösoroszi region. Here a smaller ore mine and ore dressing plant was operated. Unfortunately, however, for economic reasons the operation was suspended period in 1985.

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# NEOGENE CHRONOLOGY AND CHRONOSTRATIGRAPHY—NEW DATA

by

W. A. BERGGREN

*Introduction.* One of the main subjects on the program of the VIIIth RCMNS Congress in Budapest (15—22 September 1985) is Topic A: Relative and Numerical Time Scales. In addition to the symposium to be held in the plenary session, there will be a complementary Workshop A4: Neogene Chronology and Chronostratigraphy—New Data, organized by F. F. STEININGER and W. A. BERGGREN. This report has been prepared for presentation at the workshop. Its main purpose is to bring together data which has become available only recently dealing with Neogene isotopic and/or magnetochronology. The data are from both marine and terrestrial sequences and provide a valuable source of information in continuing efforts to refine Neogene geochronology. However, as is often the case in any area of scientific research, there are conflicting aspects to some of the data and prospective users would do well to heed the Roman interdiction: *caveat emptor.* Some of these conflicts are discussed in the papers cited here and the interested reader is urged to consult primary sources.

Original figures and tables (and their captions) are included in this report for ease of reference but new numbers have been assigned here for the sake of continuity in references to the text.

*Pliocene—Pleistocene.* Seven tuffs have been dated by $K/Ar$ and $^{40}Ar/^{39}Ar$ methods from the late Neogene (Pliocene—Pleistocene) of the Koobi Fora region, east of Lake Turkana, northern Kenya (McDOUGALL, 1985). Together they provide valuable data on the geochronology of the sedimentary sequence in the Koobi Fora region beginning about 4.3 Ma and in particular of hominid evolution between $\sim$2.0—1.4 Ma.

Of particular interest to this compilation are the virtually indistinguishable dates of $1.88 \pm 0.02$ Ma and $1.86 \pm 0.02$ Ma, respectively, on the KBS and Malbe tuffs of the Koobi Fora region (McDOUGALL, 1985., tables 5 and 6). The KBS date and its normal magnetic polarity indicate that it was erupted during the Olduvai normal subchron of the Matuyama Chron (latest Pliocene age).

The Chari Tuff, which lies at the top of the Koobi Fora Formation, is dated at $1.39 \pm 0.02$ Ma (McDOUGALL, 1985., table 7). Most of the hominid fossils from the Koobi Fora region were recovered from the interval between these two tuffs and extending a short stratigraphic interval below the KBS Tuff, thus spanning a time interval of about 2.0 to 1.4 Ma. Fossils assigned to the genus *Homo* and the genus *Australopithecus* have been retrieved from this interval and the Koobi Fora data are seen to provide the most definitive information regarding the contemporaneous presence of two late Pliocene—early Pleistocene hominid lineages in East Africa (Fig. 1).

The primary pyroclastic tuff, upper Bouroukie Tuff—2 (BKT—2u), which lies about 60 m above the fossil hominid *Australopithecus afarensis*, alias "Lucy" and associated hominids at Hadar, Ethiopia, has been subjected to an extensive series of

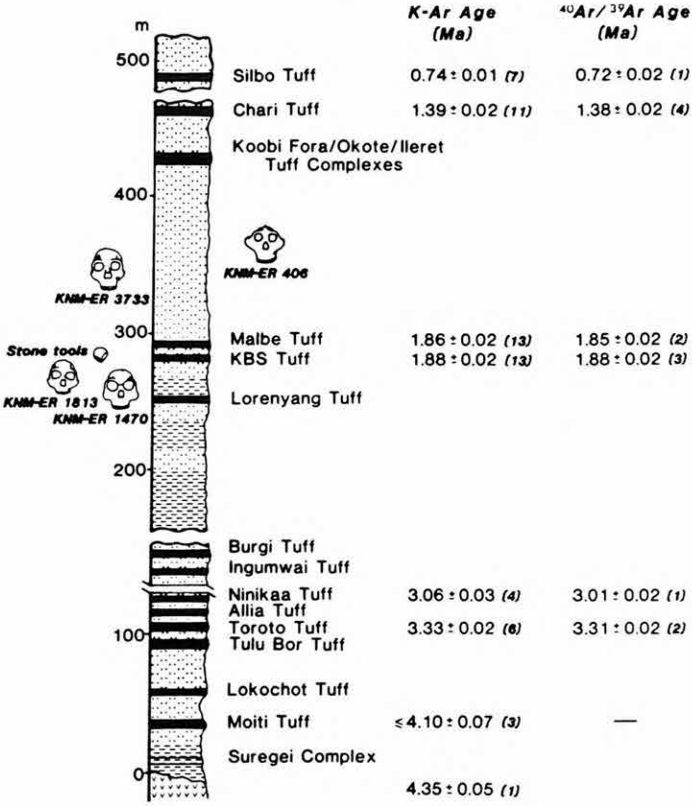



**Figure 1 Generalized tuff stratigraphy in the Koobi Fora region with mean K-Ar and $^{40}Ar/^{39}Ar$ ages shown. Number of samples used in calculating each mean is indicated in parentheses after each age.**

$^{40}Ar/^{39}Ar$ and conventional $K/Ar$ measurements on anorthoclase (HALL et al., 1985). Seven $^{40}Ar/^{39}Ar$ integrated ages yield a mean value of 3.19±0.4 Ma. Five of these ages are step-heating analyses whose plateau ages average to 3.14±0.2 Ma. Five conventional $K/Ar$ analyses yield an average age of 3.23±0.6 Ma. The results support an age for "Lucy" and associated hominids of the Hadar region of over 3 Ma.

"Lucy" is situated within a stratigraphic interval of normal magnetic polarity. These new results are inconsistent with previous studies which had bracketed "Lucy" between conventional $K/Ar$ dates of 2.6 Ma (on the BKT−2u tuff) and 3.0 Ma on a basalt layer (KMB) a short distance stratigraphically lower than "Lucy". These dates suggested that "Lucy" was situated between the Kaena and Mammoth subchron of the Gauss Chron. The ages were subsequently changed to 2.9 Ma and 3.3 Ma, respectively. The isotopic dates no longer agree with the paleomagnetic stratigraphy and one, or both, methods may be in error. A further possibility exists of error in stratigraphic correlation between various volcanic units.

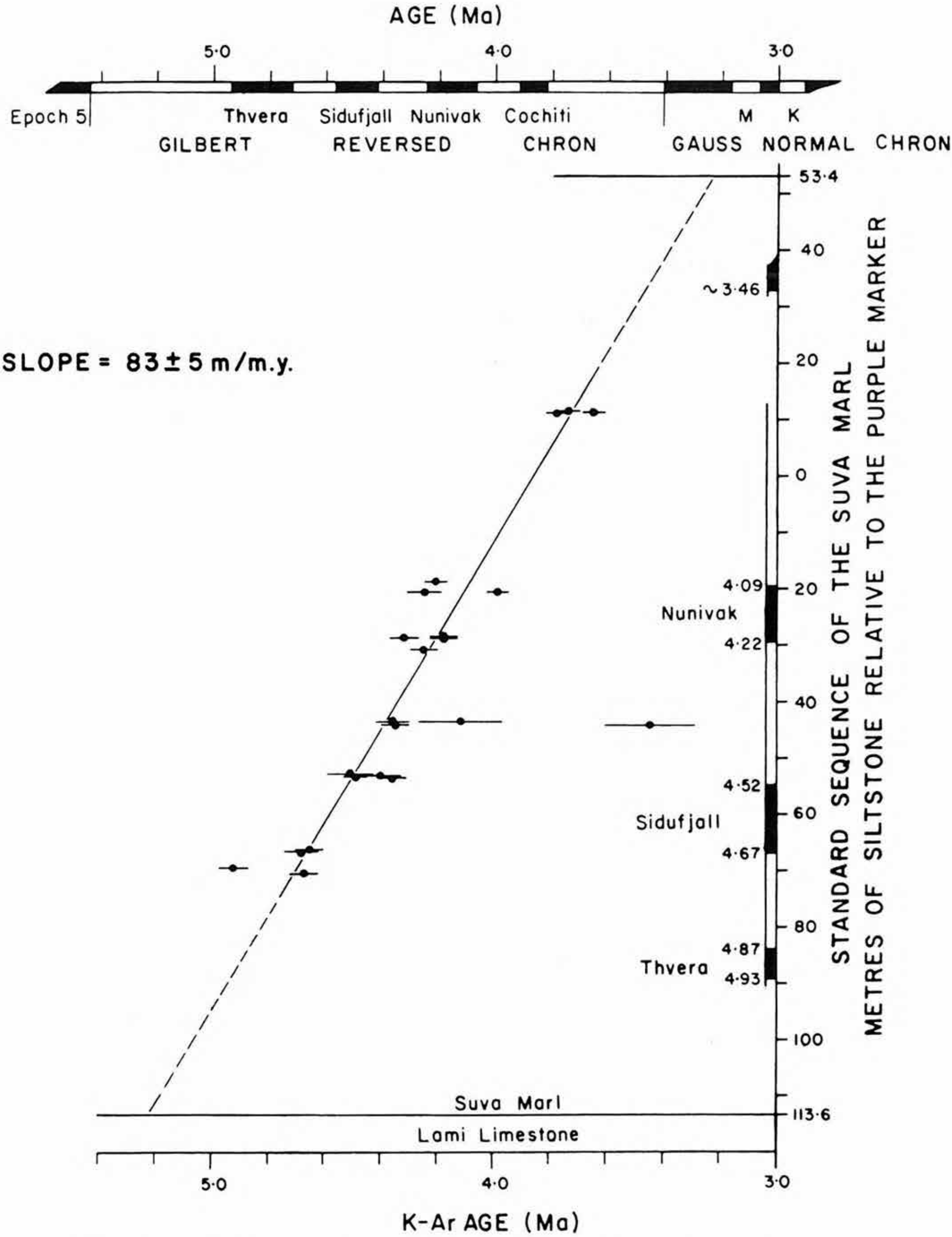



*Fig. 2.* Plot of *K/Ar* dates against stratigraphic position within the Suva Marl

At top is the paleomagnetic times scale after McDOUGALL (1979), and at right is the magnetic stratigraphy of the Suva Marl, based on studies by R. A. CASSIE (—90.2 m to +13.5 m) and YASIKAWA and others (+30.9 m to +37.5 m)

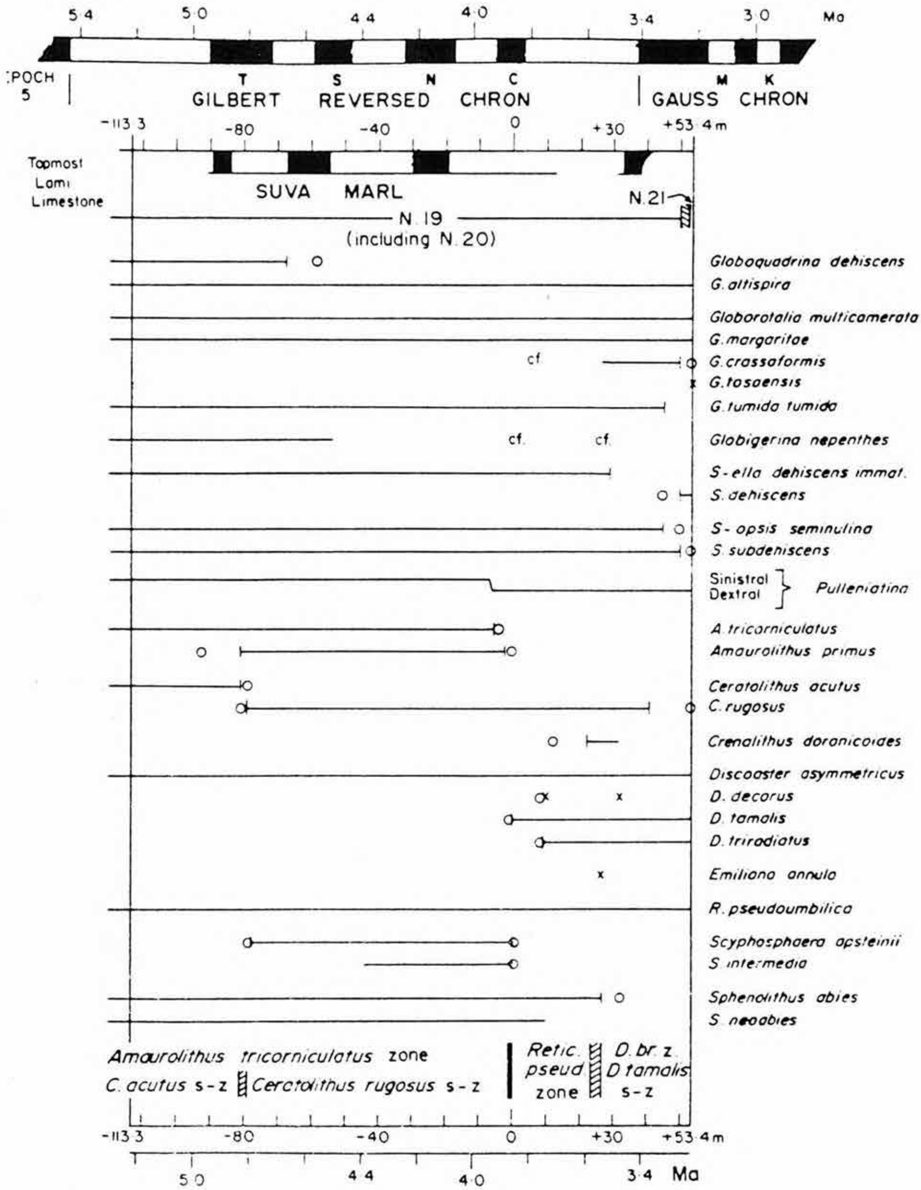



*Fig.* 3. Stratigraphic ranges of selected planktonic forminifera and calcareous nannoplankton within the topmost Lami Limestone and the Suva Marl, with paleomagnetic and $K/Ar$ data

*From top downward:* paleomagnetic time scale after McDougall (1979) positioned according to the $K/Ar$ dates from the Suva Marl; stratigraphic position within the Suva Marl with respect to the Purple Marker, and paleomagnetism of the Suva Marl; planktonic foraminiferal zones of the Suva Marl (the hatched bar indicates the range of uncertainty); ranges of planktonic foraminifera and calcareous nannoplankton (termination of range line = extreme sample containing the taxon, x = spot occurrence, o = first adjacent sample lacking the taxon); nannoplankton zone/subzone and stratigraphic position (hatched or solid bars representing the range of uncertainty); and age of the Suva Marl (partly extrapolated) from the $K/Ar$ dates

*Pliocene*. An integrated study on isotopic dates, magnetostratigraphy, and calcareous plankton biostratigraphy on the lower Pliocene Suva Marl, Fiji (RODDA et al., 1985) contains important data for early Pliocene magnetobiochronology. By combining data on nearly 40 *K/Ar* (biotite) dates (RODDA et al., 1985., table 1) on volcanoclastic tuffs (ranging sequentially in age from *ca* 4.8 Ma to 3.8 Ma) in an approximately 180 m thick sequence of predominantly reversed polarity (but with three normal events identified as the Thvera, Sidufjall, and Nunivak events of the Gilbert Chron; Fig. 2, this paper) and calcareous plankton biostratigraphy (Fig. 3), it has been possible to estimate ages for the normal polarity subchron boundaries as well as several biostratigraphic zonal boundaries and events. An age–stratigraphic height (relative to an 0–level marker bed) regression was used to estimate ages of the polarity boundaries which are seen to be in close agreement with the polarity scale of MCDOUGALL (1979) based predominantly on age and polarity data on subaerial volcanic rocks.

There is a close correspondence between the radiochronologic framework provided for the early Pliocene calcareous plankton biostratigraphy on Fiji (RODDA et al., 1985) and the magnetobiochronology derived by BERGGREN et al. (1985; see Fig. 11, this work).

*1.* 33 *K/Ar* dates (see RODDA et al., 1985., table 1) over the stratigraphic interval representing the combined zones NN 13 and NN 14 range from 4.9 to 4.0 Ma with estimated boundary ages of 4.82 Ma (NN 12/NN 13) and 3.86 Ma (NN 14/NN 15) based on a best fit regression through the dated values. By comparison the magnetobiochronologic estimates of BERGGREN et al. (1985) for these two boundaries are 4.6 Ma and 3.7 Ma, respectively.

*2.* 6 *K/Ar* dates (see RODDA et al., 1985., table 1) from the same stratigraphic level within Zone NN 15 yield average dates (on replicate samples) of 3.78 Ma, 3.74 Ma, and 3.65 Ma with an estimated age for the NN 15/NN 16 boundary of 3.52 Ma based on a best fit regression through the dated values. By comparison the magnetobiochronologic estimate of BERGGREN et al. (1985) for the NN 15/NN 16 boundary is 3.5 Ma.

*3.* Finally, the extinction of several planktonic foraminiferal taxa (i. al., LAD's of *Globigerina nepenthes*, *Sphaeroidinellopsis* spp., *Globorotalia multicamerata*, *Globoquadrina altispira*) as well as the sinistral to dextral coiling change in *Pulleniatina* and the upward increase in *Sphaeroidinella dehiscens* in the Suva Marl (RODDA et al., 1985; see Fig. 3) all seem to agree closely with values derived by BERGGREN et al., (1985).

*Miocene*. MCDOUGALL et al. (1984) have presented a paleomagnetic stratigraphy of two composite sections (Figs. 4 and 5) of tholeitic, low potassium flood basalts on the NW peninsula of Iceland. Potassium–argon dates (MCDOUGALL et al., 1984, tables 3–10) on over 70 lavas distributed throughout these composite section (on the eastern and western side of the peninsula) indicate that the lava sequences span a time interval of about 14–8 Ma (middle to late Miocene) (Figs. 6 and 7).

The authors presented a detailed discussion of the problems associated with the interpretation of the *K/Ar* ages, including both anomalously young and old ages. Age data were plotted versus cummulative stratigraphic thickness from the western (Fig. 6) and eastern (Fig. 7) sections, respectively. The general linear decrease in age upward through the two successions was interpreted as evidence that the ages provide a reliable estimate for the time of eruption and cooling of the lavas. Nevertheless, there is a spread in measured age which is uncorrelated with stratigraphy in some instances and some ages were eliminated from the final analysis, others combined to

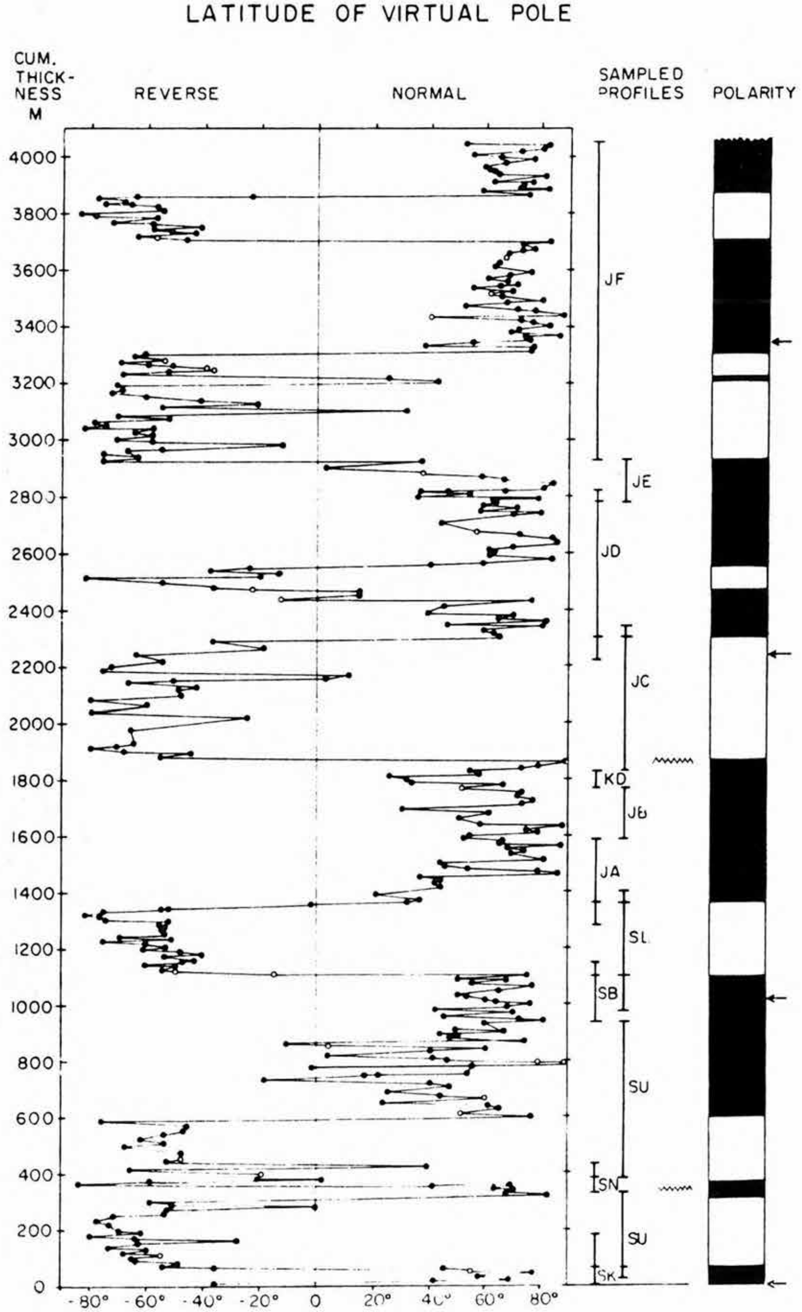



*Fig. 4.* Latitude of VGP derived from basalts of the western composite section plotted against cumulative stratigraphic thickness

Solid and open circles indicate flows where the $\alpha_{95}$ of the field direction mean is less or more than 23.5°, respectively. Short bars across vertical profile lines mark overlap with neighbouring profiles but only a single data set is plotted. Interpreted polarity log given on right, where black indicates normal polarity and white indicates reverse polarity. Arrows on the right indicate stratigraphic position of additional thin polarity zones, recorded in profiles overlapping with those of the main composite section provide a mean age. Regression analysis was then applied to the measured ages with an error of 2% assigned to the ages and uncertainty for the accumulative stratigraphic thickness parameter was arbitrarily set at 5%. In this way, ages were derived for the polarity reversal boundaries (Figs. 6 and 7).

One of the most significant results of this study is the revised estimate of 9.64 and 11.02 Ma for the younger and older limits of Chron C5N (= anomaly 5), respectively (Fig. 8). This is approximately 1 m. y. older than estimates used in several other magnetochronologic scales currently in use. McDougall et al. (1985) note that for the 15 polarity interval boundaries which they have recognized in the eastern Iceland sequence and can correlate with those in the Ness et al. (1980) time scale, there is a

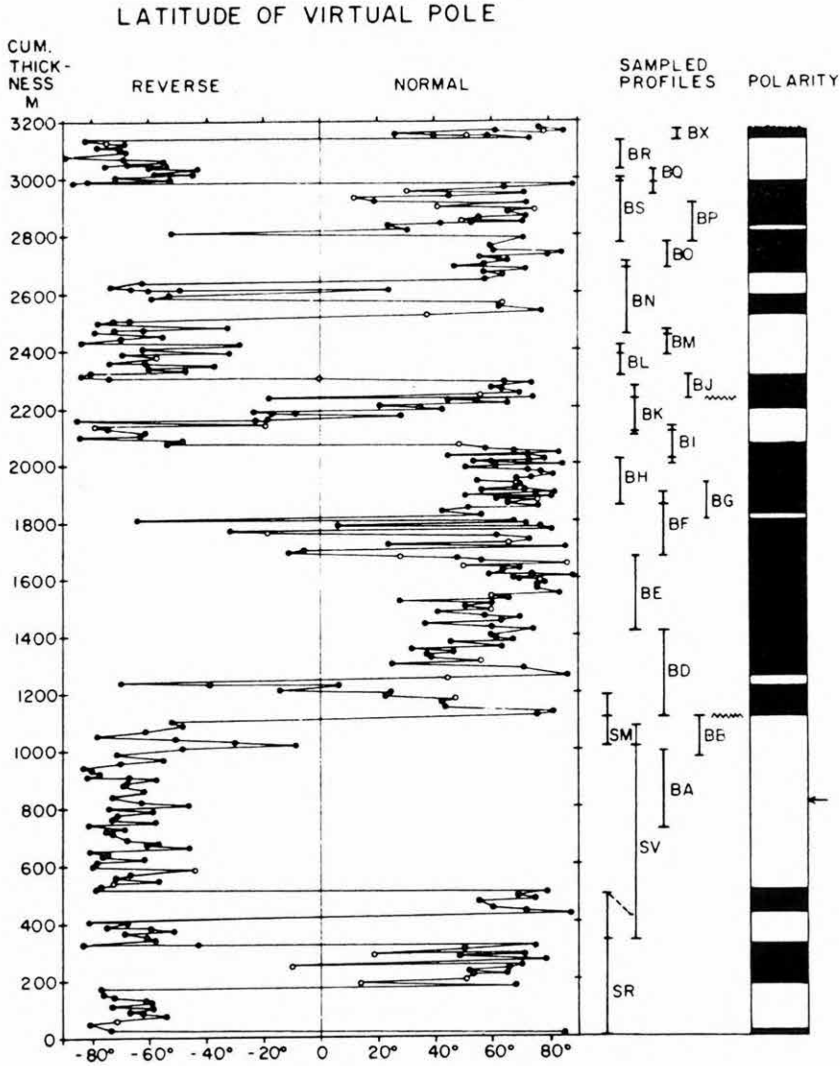



*Fig. 5.* Plot of latitude of VGP derived from basalt flows of the eastern composite section. Same conventions adopted as given in caption to Figure 4

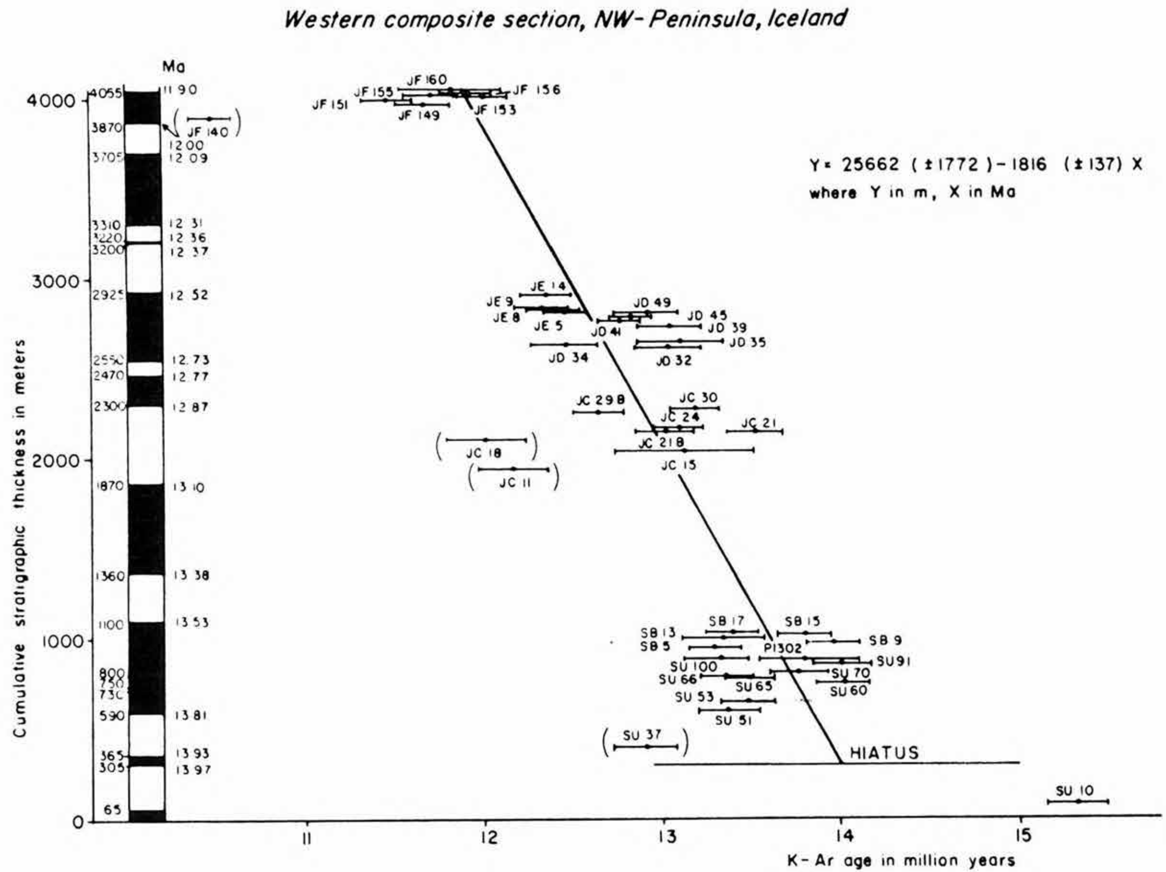



*Fig. 6.* Plot of measured *K/Ar* age against cumulative stratigraphic thickness for the western composite section

Error bars for *K/Ar* ages are one standard deviation. Best fit regression line through data is shown; age results in parentheses not used in regression analysis. Polarity log for the sequence is given adjacent to thickness axis, where black indicates normal polarity and white indicates reverse polarity. Ages for polarity interval boundaries derived from the regression

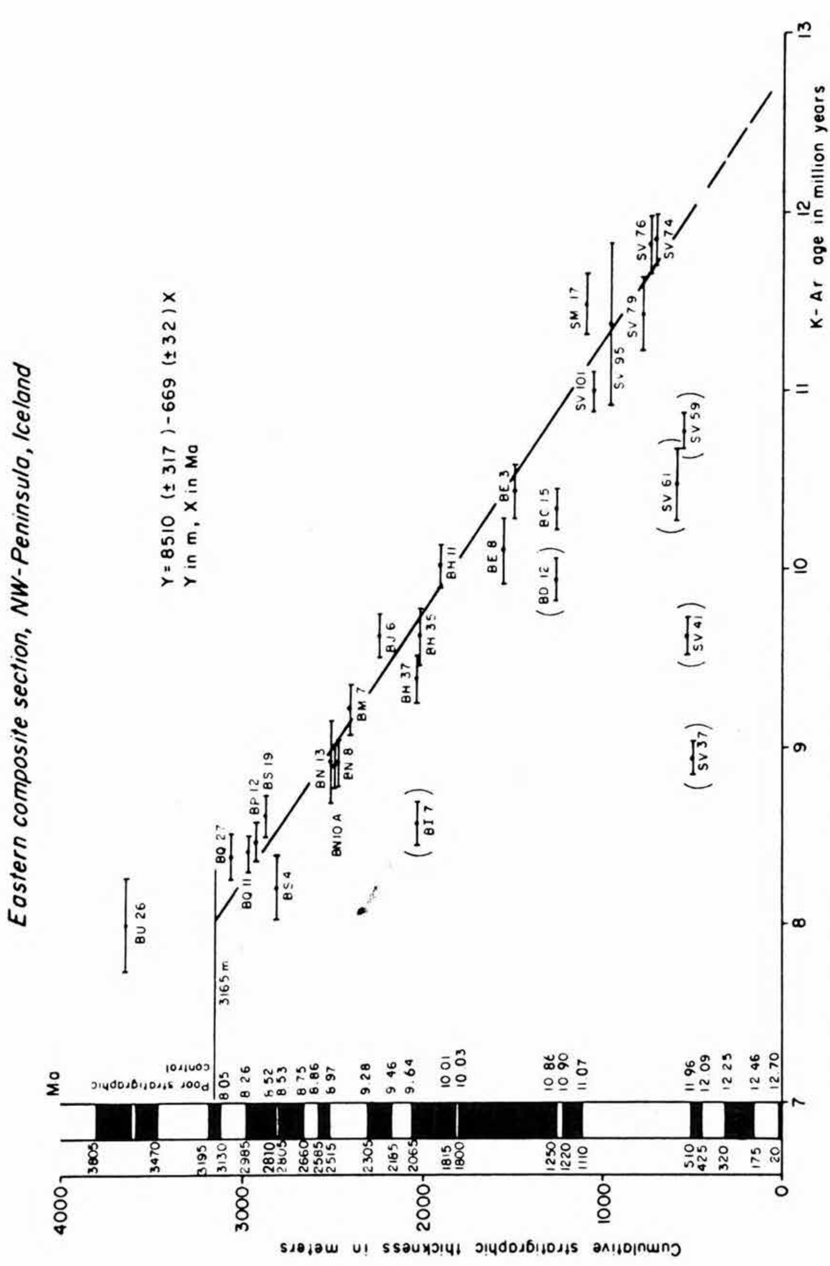



*Fig.* 7. Plot of measured *K/Ar* age against cumulative stratigraphic thickness for the eastern composite section

Best fit regression line through data is shown; age results in parentheses not used in regression analysis. Ages for polarity interval boundaries indicated on polarity log are derived from the regression

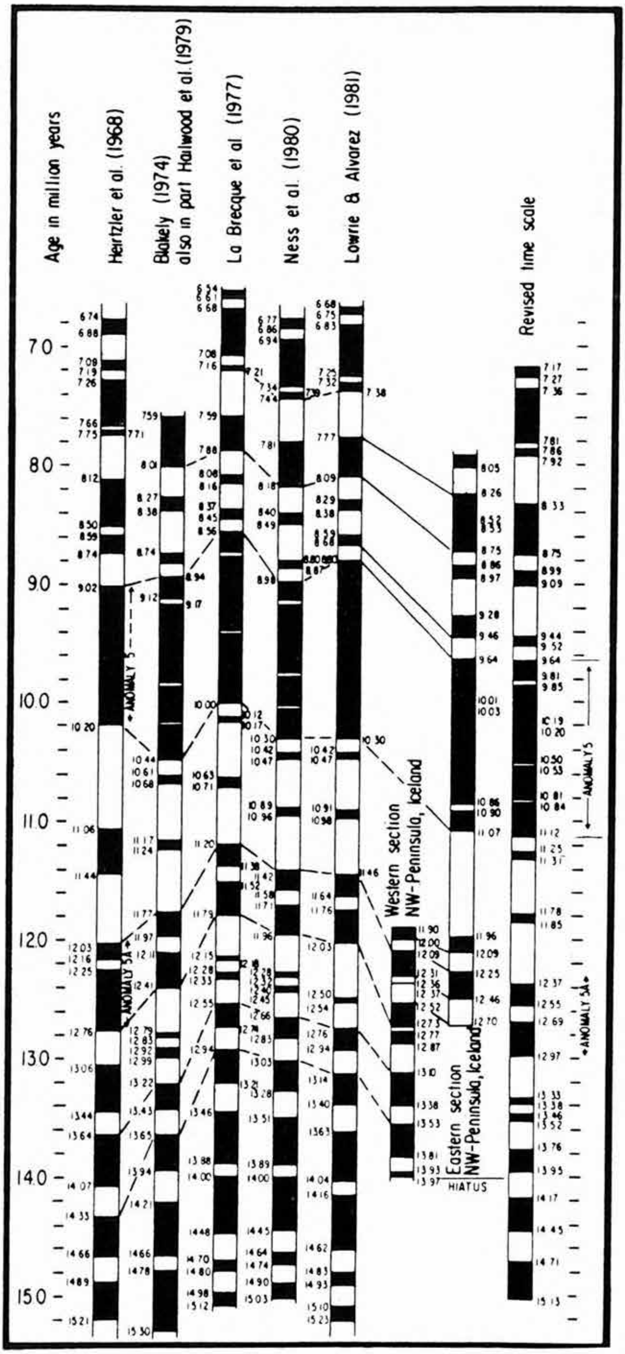



*Fig. 8.* Polarity time scales from various authors for the interval 7—15 Ma ago, adjusted where necessary to conform with current $^{40}K$ decay constants

Black represents normal polarity, white reverse polarity. Polarity logs for western and eastern composite sections of NW peninsula, Iceland, are shown to right. Suggested correlations indicated by tie lines. On extreme right is a derived polarity time scale based upon an age of 9.64 Ma for younger boundary of anomaly 5, an age of 3.40 Ma for the Gauss/Gilbert chron boundary and acceptance of the marine magnetic anomaly pattern consistently older age of from 0.45 and 0.83 Ma, and averaging 0.59±0.12 Ma which they take as indicative of the homogeneity and compatability of the two data sets. They admit the change in calibration of the polarity time scales to older ages by up to 8% is significant, but required if the age data are accepted as valid.

The older age derived for the base of anomaly 5 depends, in turn, on the interpretation of the base of anomaly 5 in the eastern composite section (Fig. 7). As McDougall et al. (1985) note, the base of anomaly 5 may be located either at the base of the long normal sequence at 1250 m which would yield an age estimate of 10.86 Ma for the base of anomaly 5, or at the base of a short normal event below the long normal at 1110 m, which is the position favoured by them.

They point out further that there are three independent estimates for the base of anomaly 5 in the literature: 10.3 Ma, 10.47 Ma, and possibly 11.07 Ma. An average age based on these three estimates is 10.61±0.40 Ma. Using the values 10.3 Ma, 10.47 Ma, and 10.86 Ma yields an average of about 10.5 Ma which is similar to published values (10.44 Ma) assigned by Blakely (1974), initially calculated by Ness et al, (1980) and derived (10.4 Ma) by Berggren et al. (1985) for the base of anomaly 5.

Tauxe et al. (1985) have presented an integrated study of the paleomagnetic stratigraphy and isotopic dating of the middle Miocene Ngorora Formation in the Barengo basin of the Kenya Rift valley. The magnetostratigraphy can be well correlated to the geomagnetic polarity time scale and spans the interval from Chron C5 to Chron C5ABR (Fig. 9). The isotope dates, 8 in all, fall into three discrete groups with average values of 12.5±0.22 Ma, 11.6±0.6 Ma, and 10.16±0.38 Ma (op. cit., see table 2).

A comparison of the Icelandic isotopic dates from McDougall et al. (1984) with the Kenyan dates of Tauxe et al. (1985) (Fig. 10) shows that the *K/Ar* dates on euhedral sanidines from Kenya overlap the scatter of the Icelandic *K/Ar* dates.

When comparison is made between a revised version of the LaBrecque et al. (1977) time scale (using the Kenyan data) and the revised time scale of McDougall et al. (1984), it can be seen that the two scales differ by more than 10%, which is considerably larger than the analytical error of the age determinations (~2%). For instance, whereas McDougall et al. (1984) suggest an age estimate for the older boundary of anomaly 5 of 11. 12 Ma, Tauxe et al. (1985) suggest a value of 10 Ma, a difference of over 1 m. y. Tauxe et al. (1985) believe that inasmuch as the correlation of the magnetostratigraphy to the magnetic anomaly pattern appears incontrovertible in both studies, particularly in the identification of magnetic anomaly 5, it would appear that the problem(s) lie with the reliability of the isotopic dates. I have pointed out above, however, that part of the problem may also lie in the interpretation of the lower (older) limit of anomaly 5 in the eastern composite section on Iceland. Tauxe et al. (1985), in reviewing the two discrepant sets of data, suggest that the Icelandic basalts may have been subjected to alteration (loss of potassium). They believe that the isotopic dates on euhedral sanidines from Kenya are more reliable than the whole rock dates from Iceland. The discrepancies may, in fact, be more complicated. There may be unresolved problems in the isotopic dating, the identification of the magnetic polarity stratigraphy or a combination of both.

*Lesser Antilles.* A number of recent isotopic dates in the Lesser Antilles have a direct bearing on Neogene geochronology (Andreieff et al., 1976; P. Andreieff, personal communication, February, 1985 and July, 1985). On the islands of Martinique and Guadeloupe a series of intercalated marine sediments and basalt flows dated by the potassium-argon method have yielded the following results (from older to younger; see Table 15).

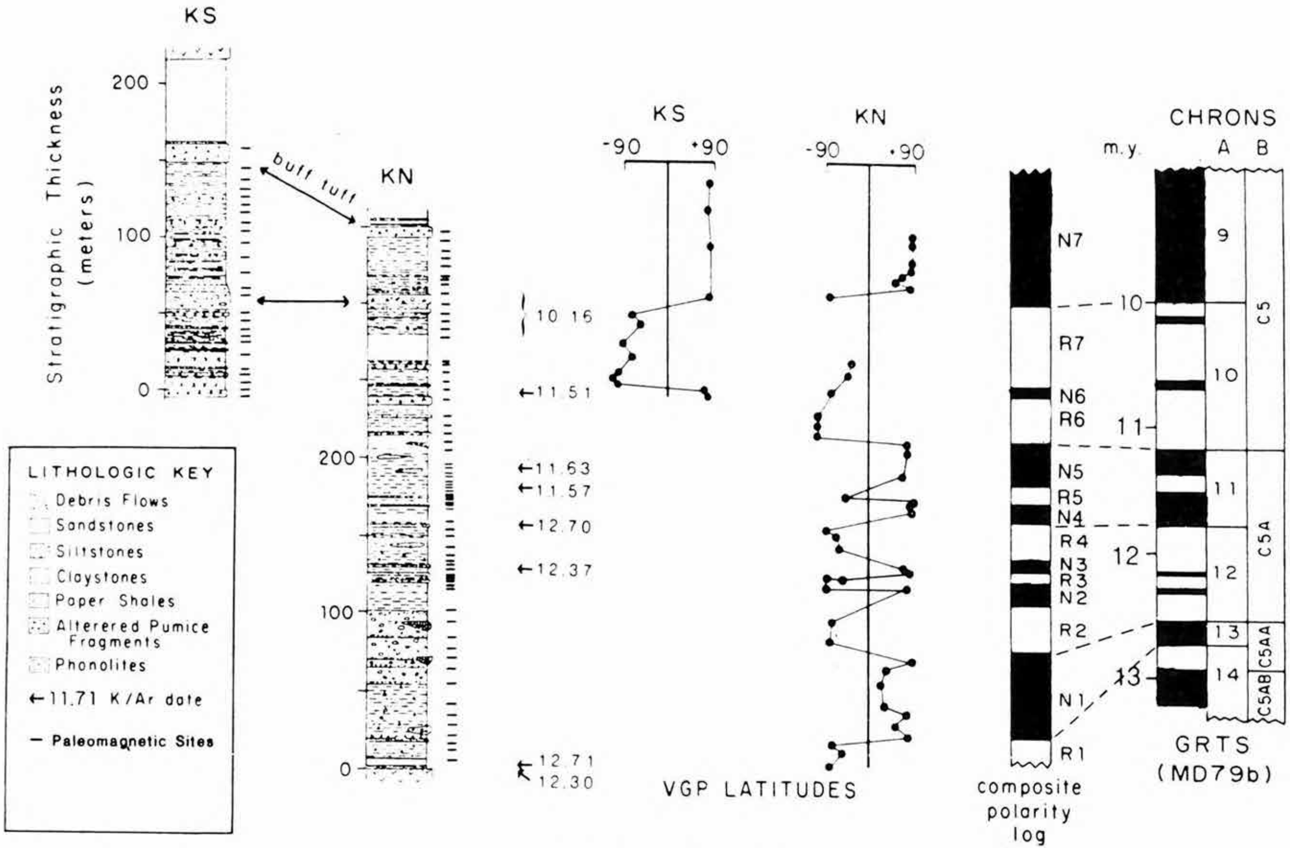



*Fig. 9.* Stratigraphic sections

Thickness in meters to left. *K/Ar* dates are those from Table 2. Columns labeled KN (Kabarsero North) and KS (Kabarsero South) are the stratigraphic sections as measured and described by us. Paleomagnetic sites are indicated by a tick mark to the right of lithology. Latitudes of the virtual geomagnetic pole (VGP) position are plotted to the right of the site location. These are interpreted in terms of polarity (normal/black, reversed/white). The magnetozones are labelled R1—N7 for reference in text. The magnetostratigraphy is correlated to the geomagnetic reversal time scale (GRTS) of Mankinen and Dalrymple (1979) (MD79b). Chron column B is the terminology of Berggren et al. (1985)

Attention is drawn to the close correspondence between the numerical values derived from studies in the Lesser Antilles and those proposed in a recent Neogene magnetobiochronologic scale (BERGGREN et al., 1985; see Fig. 11, this paper).

*Miscellaneous Data.* There are a number of data from different parts of the world which deal with Neogene isotopic chronology. Inasmuch as these data are scattered in the literature and do not form part of a more comprehensive study on a particular time interval, problem, or area, they are compiled in tabular form with appropriate explanatory comments and presented below (Tables 2, 3).

*Acknowledgments.* This report has been prepared at the request of Dr. F. STEININGER for Workshop A4 on Neogene Chronology and Chronostratigraphy—New Data, at the VIII Congress of the RCMNS, Budapest (September 1985). I should like to thank the Organizing Committee of the Congress, and in particular Dr. G. HÁMOR, President of the Congress, for inviting me to attend this meeting and to present this report and related results.

Drs. N. DE B. HORNIBROOK (Lower Hutt) provided literature and made critical comments on New Zealand data, I. MCDOUGALL (Melbourne) did the same with his recently published studies on Iceland, Fiji and East Africa B. ANDREIEFF (B.R.G.M., Orleans) provided me with a compilation of published and unpublished data from the Lesser Antilles, and D. V. KENT (New York) and M.-P. AUBRY (Woods Hole) reviewed with me various problems associated with late Neogene geochronology and read an early draft of this report. I am grateful to these colleagues for their help and stimulating critique. This work has been supported by the National Science Foundation Grant OCE8017040. This is Woods Hole Oceanographic Institution Contribution No. 6009.

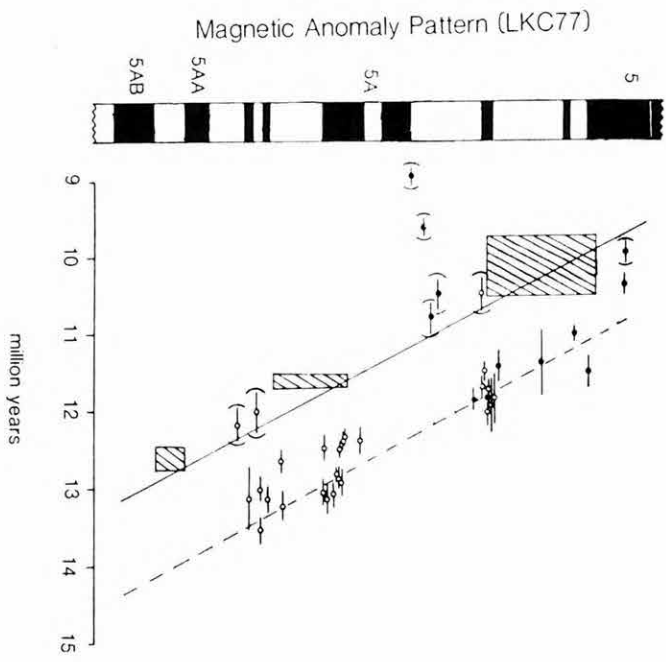
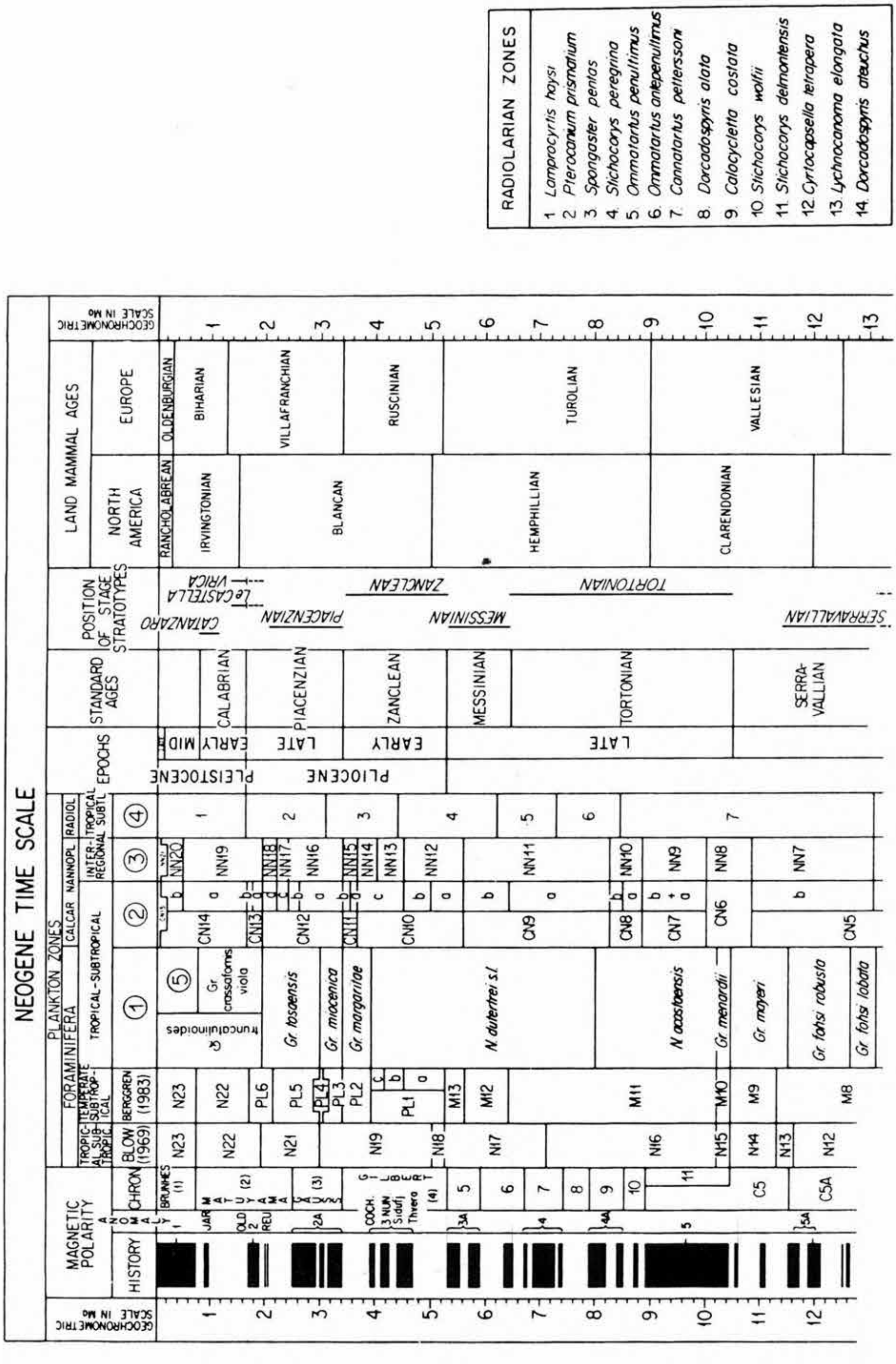



*Fig. 10.* Compilation of isotopic age dates from MCDOUGALL et al. (1984) and this paper with respect to the reversal pattern of LABRECQUE et al. (1977)

Solid (open) circles are from the eastern (western) composite section of northwest Iceland. Data in parentheses were excluded in the original study as anomalous Hatchured regions are the isotopic age data from Figure 4 as explained in text. The dashed line is the MKS84 time scale and the solid line is the LKC77* time scale


NEOGENE TIME SCALE
RADIOLARIAN ZONES
1 Lamprocyrtis haysi
2 Pterocanium prismatium
3 Spongaster pentas
4 Stichocorys peregrina
5 Ommatartus penultimus
6 Ommatartus antepenultimus
7 Cannatartus petterssoni
8 Dorcadospyris alata
9 Calocycletta costata
10 Stichocorys wolfii
11 Stichocorys delmontensis
12 Cyrtocapsella tetrapera
13 Lychnocanoma elongata
14 Dorcadospyris ateuchus
GEOCHRONOMETRIC SCALE IN Ma
MAGNETIC POLARITY
HISTORY
ANOMALY
CHRON
PLANKTON ZONES
FORAMINIFERA
CALCAR NANNOPL
RADIOL
EPOCHS
STANDARD AGES
POSITION OF STAGE STRATOTYPES
LAND MAMMAL AGES
NORTH AMERICA
EUROPE
PLEISTOCENE
PLIOCENE
CALABRIAN
PIACENZIAN
ZANCLEAN
MESSINIAN
TORTONIAN
SERRAVALLIAN
IRVINGTONIAN
BLANCAN
HEMPHILLIAN
CLARENDONIAN
BIHARIAN
VILLAFRANCHIAN
RUSCINIAN
TUROLIAN
VALLESIAN

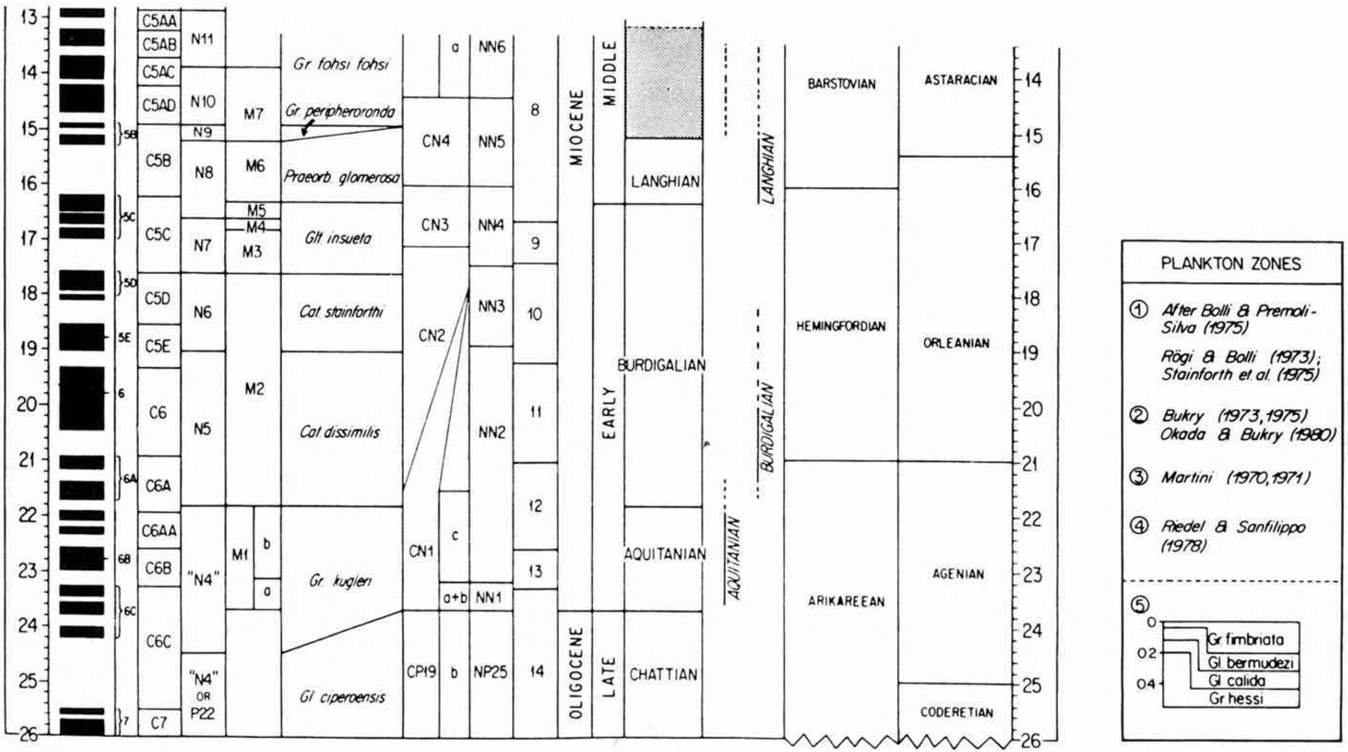



*Fig. 11.* Neogene geochronology

The geochronologic scale at the margins of the figure is derived from the magnetic polarity chronology which is in turn derived from paleontologically and/or paleomagnetically controlled radiometrically dated calibration points in the late Neogene, early Oligocene, middle Eocene and late Cretaceous (BERGGREN et al., 1985, for further explanation). The position of the calcareous planktonic zonal boundaries is based, for the most part, upon direct (first order) correlation between biostratigraphic datum levels and paleomagnetic polarity stratigraphy as determined in deep sea cores or continental marine sediments. In this way, a true 'magnetobiochronology' is possible. The extent (duration) of standard time-stratigraphic units and their boundaries and the position of stage stratotypes are estimated on the basis of their relationship to standard plankton biostratigraphic zones. (From BERGGREN et al., 1985, Fig. 2)

*Table 1*

| Location | Bio—magneto—chrono-stratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| *1* Tuffites de Fort de France, B.R.G.M. well MS3, Rivière Monsieur, eastern part of city of Fort-de-France, Martinique | Late *Neogloboquadrina humerosa* zone to "*Globigerinoides conglobatus* Zone" At 36 m: *N. humerosa*, *Nq. acostaensis* (sinistral), *Gr.* "*limbata*" (dom. dextral), *Gr. juanai*, *Gs. extremus*, *Gs. canimarensis*, *Gq. altispira*, *Gq. dehiscens*, *Ss. sphaeroides*, *Ss. seminulina*, *Gb. nepenthes*, *Cd. nitida;* late *Nq. humerosa* zone (~N17) | 6.5 ± 0.3 | K/Ar (whole rock) | ANDREIEFF and WESTERCAMP (in prep.); ANDREJEFF *et al.* 1976; ANDREIEFF (pers. commun., 1985) | Level at 36 m directly overlies andesite flow. |
| 2 Tuffites de Rivière—Pilote et l'Anse Dufour, Calcaire du Vauclin in Rivière—Pilote, Ste Marie and Vauclin areas, Martinique | *1* Rivière Pilote: *Gr. mayeri* Zone (lower part?) with *Gr. mayeri*, *Gb. nepenthes*, *Gb. druryi*, *Gr.* ex. *Gr. menardii*<br>*2* Anse Dufour (near Ste. Marie; very rich microfaunas from the *Gr. mayeri* zone<br>*3* Vauclin: *Gr. mayeri* and *Gr. menardii* zones | 10.6 ± 0.5, 10.6 ± 1.0, 10.2 ± 0.5, 11.1 ± 0.3, 10.3 ± 0.4 | K/Ar (whole rock) | same as above | Volcanics belonging to the "2d Effusive phase" are dated 10.6 ± 0.5 Ma at Pointe Bout, 10.6 ± 1.0 Ma at Quartier Medecin and 10.2 ± 0.5 Ma at Ste Luce. They are directly underlain by fossiliferous tuffs (*G. mayeri* zone) near Rivière—Pilote (Quartier Josseaud, Rocher Zombi, B.R.G.M. well SMA 3) and overlain by tuffs and limestones from *G. mayeri* zone or *G. menardii* zone in Vauclin area. At Pointe Vauclin, fossiliferous tuffs (*G. menardii* zone) are directly overlain by Montagne du Vauclin volcanics, and a lava flow from this locality is dated 11.1 ± 0.3 Ma. Another nearby lava flow belonging to the same volcanic complex is dated 10.3 ± 0.4 Ma at Pointe Faula |

| | | | | | |
|---|---|---|---|---|---|
| *3* Calcaires et tuffites du Marin s.s., Le Marin area, Martinique | *Globorotalia fohsi* zone to early *G. robusta* zone.<br>Samples from lower part of the formation at Quartier Bandelle yield *Gr. periphero-ronda*, *Gr. peripheroacuta*, *Gr. mayeri*, *O. suturalis*: early *G. fohsi* zone.<br>At Le Marin, a section of more than 150 m of marly tuffs and limestones has the following biostratigraphic succession:<br>— late *Gr. fohsi* zone with *Gr. peripheroacuta* and *Gr. praefohsi*<br>— *G. lobata* Zone with the nominate taxon<br>— early *Gr. robusta* zone with *Gr. fohsi* (*sensu* Benner and Blow), *Gr. ex. interc*, *G. lobata-robusta*, and *Ss. multiloba* | 12.6 ± 0.3 (Quartier Vieielle Terre) | K/Ar (whole rock) | same as above | This formation overlies the "First Effusive Phase" volcanics (dated 15.1 ± 0.3 Ma; cf. below) and is overlain by the Massif du Vauclin lava flows which is dated at 12.6 ± 0.3 Ma (see column 3) |
| *4* Calcaires du François and "Tufs" de Bassignac, Le François and Trinité areas, Martinique | *Globorotalia peripheroronda* zone (*Gr. peripheroronda*, *Gr. mayeri*, *Gr. praemenardii*, *Ss. disjuncta*, *Gs. obliquus*, *Globigerinopsis aquasayensis* Orbulinids) | 15.1 ± 0.3 | K/Ar (whole rock) | Same as above | The fauna (column 2) is directly overlain by the "First Effusive Phase" whose basalt flows are dated 15.1 ± 0.3 Ma (see column 3) at Quarry Pacquemar in ANDREIEFF *et al.* (1976). The "Tufs" de Bassignac were placed in the Miocene—Pliocene transition interval based on a paper presented by JULIUS and PONS at the same 7th Caribbean conference. Based on examination of a single sample from this formation, a footnote (p. 344) was entered in the galley proofs modifying the |

*Tabe 1. continued*

| Location | Bio—magneto—chrono-stratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| | | | | | biostratigraphic position to middle Miocene (zone N13) based on the presence of *G. aquasayensis*. Additional sampling since that time has shown that the Calcaires du François and "Tufs" de Bassignac are coeval and belong to the *Gr. peripheroronda* zone. |
| 5 Calcaires de Sainte Anne, Sainte Anne Peninsula and Macabou area, Martinique | *Globigerinatella insueta* zone (*Gs. bisphaericus*, *Gq. altispira*, *Gq. dehiscens*, *Gt. insueta*, *Gr. mayeri*, *Gr. peripheroronda*, and larger foraminifera: "*Operculinoides*" *cojimarensis*, *Miogypsina antillea*, *Lepidocyclina canellei* | $18.1 \pm 1.3$ (below) $15.7 \pm 1.0$ and $15.9 \pm 1.0$ (above) (see under remarks) | K/Ar (whole rock) | same as above | At Ste. Anne, these limestones and marly limestones are directly underlain by "Tufs de Fond Moustique" at Creve-Coeur which are associated with a basalt-flow dated $18.1 \pm 1.3$ Ma by K/Ar method (whole-rock). At Morne Carriere (Macabou area), they are overlain by François—Robert volcanics; lava flows belonging to this cycle are dated $15.9 \pm 1.0$ Ma (Quarry Bois-Soldat) and $15.7 \pm 1.0$ Ma (Ilet á Eau, Robert-Bay). In Andreieff *et. al.* (1976), the "Calcaires de Ste Anne" were thought to be coeval with Macabou complex (*Globorotalia kugleri* zone). After 10 years of detailed mapping (by D. Westercamp), intensive sampling and micropaleontological investigations (by P. Andreieff), ideas concerning the stratigraphy and the magmato-structural evolution of Martinique have changed considerably: calcaires de Ste. Anne are definitely younger than Macabou and belong to late early Miocene. |

| | | | | | |
|---|---|---|---|---|---|
| *6* Tuffites de l'Anse Pitou, Island of Marie-Galante, Guadeloupe | Late Miocene planktonic foraminiferal fauna (zone N17). Associated fauna: *N. acostaensis* (sinistral), *N. humerosa* (sinistral), *G. gr. menardii* (including *G. "limbata"*), *G. scitula*, *G. obliquus*, *G. extremus*, *G. trilobus*, *G. sacculifer*, *G. subdehiscens*, *G. altispira*, *G. nepenthes*, *G. venezuelana*, *H. siphonifera*, *Orbulina*. Assigned to lower part of *G. dutertrei* zone of Bolli (≃ Zone N17 Blow). | 6.81 ±0.31<br>6.65 ±0.28 | K/Ar (whole rock) | P. ANDREIEFF (personal communnication, June, 1985) | Two analyses on the same sample. |
| *7* Cariacou, Grenadines | *Catapsydrax stainforthi* zone | 18.1 ±0.5 | K/Ar (whole rock) | BRIDEN et al. (1979), WESTERCAMP et al. (1985) | Basalt from basal part of Belmont Formation within *C. stainforthi* zone. |

*Table 2*

| Location | Bio—magneto—chronostratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| *1* Mangapoike River section, North Island, New Zealand | Chron 6, Tongaporutuan Stage | 5.8 ±0.55 | fission track (zircon) | HORNIBROOK (1984) | Sample from thin tephra bed, just below unconformity in magnetically reversed interval in mid to upper part of Chron 6 (Tongaporutuan Stage; HORNIBROOK, 1984; 111, fig. 2). Fission track date from personal communication to N. de B. HORNIBROOK from D. SEWARD, 1983. Unconformity separates Tongaporutuan (*G. miotumida* zone of Jenkins, and above *Bolivinita pohana-pliozea* transition) (below) and upper Kapitean or basal Opoitian (*G. sphericomiozea* range zone) (above) |

*Table 2. continued*

| Location | Bio—magneto—chronostratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| *2* Momoe-a-toa Peninsula, Chatham Islands, Southwest Pacific | Early Pliocene planktonic foraminiferal fauna | 5.2±0.02 | K/Ar | HORNIBROOK (1984) | ADAMS in GRINDLEY et al. (1977) = 5.1±0.2 Ma; recalculated using new decay constant, C. J. D. ADAMS, pers. comm. to N. de B. HORNIBROOK, 1983; cited in HORNIBROOK 1984: 112. Associated with *Globorotalia mons*, (given originally as *G. conomiozea* by HORNIBROOK in GRINDLEY et al., 1977). *G. crassaformis*. *G. puncticulata*, *G. pliozea*. Magnetically normally polarized basalt lies *below* contact with the sediments (GRINDLEY et al., 1977; W. A. WATTERS, pers. comm. to N. de B. HORNIBROOK, 1983) and is stratigraphically *lower* than the level in a higher basalt flow from which the K/Ar dated sample was taken (J. T. LUMB, pers. comm. to N. de B. HORNIBROOK, 1983). Ref.: HORNIBROOK 1984: 113. |
| *3* Maunganui Bluff, 13 km southwest of Momoe-a-toa, Chatham Island | Early Pliocene planktonic foraminiferal fauna | 5.33±0.02 | K/Ar | HORNIBROOK (1984) | Thin calcareous sand underlying magnetically normally polarized basalt (originally dated 5.2±0.2 by ADAMS in GRINDLEY et al., 1977; recalculated using revised decay constant, C. J. D. ADAMS, pers. comm. to N. de B. HORNIBROOK, 1983). Contains *Globorotalia mons*, *G. crassaformis*, *G. puncticulata*, *G. pliozea*, providing confirmation of age relationships of basalt and fossiliferous sediments in item 2. Normal polarity of basalts on Chatham Island and K/Ar dates suggests correlation with part of Chron 5 or higher in Chron 4 (Thvera). Alternative explanation given by HORNIBROOK (1984:112) include: *a* biostratigraphic correlation too old; *b* basalt intrudes the fossiliferous sediments at Momoe-a-toa and is thus younger; *c* fossiliferous strata accumulated during reversed polarity interval between two normally |

| | | | | | |
|---|---|---|---|---|---|
| | | | | | polarized basalts. There may be some ambiguity between the field relationships of the basalts and the fossiliferous strata which requires further investigation (A. R. EDWARDS, pers. comm. to N. de B. HORNIBROOK, 1983). Ref.: HORNIBROOK, 1984: 113, 114. |
| *4* Northern Hawkes Bay, South Island, New Zealand (junction of Kohukohu Road and main highway, 4 km west of Nuhaka) | Waipipian Stage | 3.02±0.35 | fission track (zircon) | HORNIBROOK (1980) | Sample dated by D. SEWARD from a 2 m tuff bed immediately overlying the Whakapunake Limestone (Waipipian Stage) with *Chlamys (Phialopecten) triphooki marwicki* Beu. Correlated biostratigraphically to a level *above* the FAD of *G. inflata* and LAD of *G. pliozea*, within the range of *G. crassaconica* and locally sinistral *G. crassaformis* and *above* the LAD of *G. puncticulata* correlated with the Waipipian Stage and Gauss Chron. (N. de B. HORNIBROOK, written communication, 27 February, 1985) |
| *5* Muriwai Quarry, Muriwai Beach near Kumeu, Auckland, New Zealand (Sample N2 KA27) | "biostratigraphycally not far below the base of the *G. bisphericus* (revised Altonian Stage)" (HORNIBROOK, written communication, 19 February, 1985) | 16.0±0.2 | K/Ar (whole rock) | ADAMS (1975) | Volcanic flow within marine beds containing Awamoan—Hutchinsonian planktonic foraminiferal zonal fossils (within range of *Globorotalia zealandica* and *G. praescitula*). Originally dated at 16.1 Ma by STIPP and THOMPSON (1971), and 16.8 Ma by OBRADOVICS in BANDY et al. (1970) |
| *6* Drillhole, J. D. GEORGE No. 1 Sea field Petroleum drillhole near Ashburton, cuttings at 3900', Canterbury, New Zealand (Sample N2 KA87 of ADAMS, 1975) | Within range of *Globorotalia zealandica* (lower Altonian Stage) | 18.6±0.2 | K/Ar (glauconite) | ADAMS (1975) | |

*Table 2. continued*

| Location | Bio—magneto—chronostratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| *7* Drillhole at Awamoa Creek, near Oamaru, New Zealand (Sample N2 KA95 of ADAMS, 1975) | Near base of *Globorotalia praescitula* (above) Otaian/revised Altonian boundary) | 19.9±0.6 | K/Ar (glauconite) | ADAMS (1975) | Samples N2 KA87 and N2 KA95 should be approximately correlative although dates are about 1 my different (N. de B. HORNIBROOK, pers. comm. 27 February, 1985). |

*Table 3*

| Location | Bio—magneto—chrono-stratigraphic control | Date (Ma) | Method | Reference | Remarks |
|---|---|---|---|---|---|
| *1* DSDP 552A/36/3 135—138 cm (SW Rockall Plateau, North Atlantic) | NN5 | 15.8±0.8 | K/Ar (glauconite) | MARTAN et al. (1984) | Minerologically and chemically (low potassium) unsuitable for K/Ar dating. Nevertheless, stratigraphically concordant (with biochronologic age estimates) dates obtained |
| *2* DSDP 553A/8/3 124—126 cm (SW Rockall Plateau, North Atlantic) | NN5 | 16.2±0.9 | K/Ar (glauconite) | MARTAN et al. (1984) | Minerologically and chemically (low potassium) unsuitable for K/Ar dating. Nevertheless, stratigraphically concordant (with biochronologic age estimates) dates obtained |
| *3* DSDP 555/26/6 140—142 cm (SW Rockall Plateau, North Atlantic) | NN3 | 23.5±0.6 | K/Ar (glauconite) | MARTAN et al. (1984) | Minerologically and chemically (low potassium) unsuitable for K/Ar dating. Nevertheless, stratigraphically concordant (with biochronologic age estimates) dates obtained |

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Proceedings of the VIIIth RCMNS Congress

# PHYLOGENETIC AND BIOGEOGRAPHIC BASES FOR AN OLD WORLD HIPPARIONINE HORSE GEOCHRONOLOGY

by

R. L. BERNOR, ZH. QIU and H. TOBIEN

Hipparionine horses have been recognized for well over a century as ubiquitous in Old World later Neogene strata. Recent studies of this group of equids have sought to refine their phylogenetic systematics, identify lineage biogeographic ranges and employ these data for geochronologic correlations. We present here our current understanding of these aspects of this group's late Miocene and early Pliocene Eurasian and African record.

The first occurrence of Old World hipparionines has traditionally attracted the attention of mammalian geochronologists. MATTHEW (1929) actively promoted this "event" as a suitable marker for the base of the Pliocene. Despite strong disagreements with this view by a number of European stratigraphers, MATTHEW's (1929) ideas were not suitably rejected until BERGGREN and VAN COUVERING's (1974) popularization of this event as the "*Hipparion* Datum". Their tenets were that "*Hipparion*" first appeared in the Old World just prior to the basal late Miocene, ca. 12.5 Ma ago, that it underwent a rapid prochoresis throughout Eurasia and Africa, and that this constituted an instantaneous geochronologic event.

Corresponding with this heightened awareness about the applicability of "*Hipparion*" geochronologic correlations, efforts were focussed to erect a phylogeny of the group. SKINNER and MACFADDEN (1977) proposed direct species relationships between their North American genus *Cormohipparion* and primitive species of Old World "*Hipparion*". WOODBURNE and BERNOR (1980) presented evidence for the existence of several Old World hipparionine supraspecific groups. Later, BERNOR et al. (1980) proposed hypotheses of evolutionary polarity within these groups, and used these lineages for correlating several European, North American, and west Asian late Miocene localities. When compared to European regional biochronologic correlations and radiometric calibrations, their "*Hipparion*" species correlations were found to be highly corroborated. WOODBURNE et al. (1981) extended this work further and presented the morphologic, evolutionary and geochronologic bases for the relationships between North American Valentinian/early Clarendonian *Cormohipparion* and Old World Vallesian hipparionines. The "*Hipparion* Datum" and postdatum "*Hipparion*" geochronology had now acquired rudimentary systematic and phylogenetic bases.

Since these early studies, research efforts were focussed particularly on refining the systematics and phylogeny of Old World hipparionines, and detailing their biogeographic and geochronologic ranges. These studies endeavour to reconstruct the accurate evolutionary history of the group and further develop their use for intraprovincial to intercontinental scale correlations. In the series of figures presented here, we will summarize the major phylogenetic and biogeographic patterns of some supraspecific lineages, and their potential for future geochronologic applications.

The methods which we currently employ for "*Hipparion*" phylogenetic systematics include character state analyses of multiple morphologic complexes including the facial region of the skull, the cheek teeth, the mandible, and postcranial morphology and functional anatomy. We also employ statistical analyses of skeletal measurements to characterize species variability and scrutinize the homogeneity of our morpho-species groups. TOBIEN and BERNOR are engaged in research on the "*Hipparion*" species issue, and are studying population variability in the Howenegg quarry sample of complete skeletons. This sample is unquestionably a single species, "*Hipparion*" *primigenium* (s.s.), and is found with an extensive assemblage of *associated* vertebrate, invertebrate and plant fossil. We intend to estimate the expected range of variability in this hipparionine population as well as pursue phylogenetic and biogeographic studies.

QIU and BERNOR are in the process of revising the systematics of Chinese hipparionines, and documenting their biostratigraphic, geochronologic and biogeographic ranges. This assemblage is of paramount importance for establishing evolutionary relationships and biogeographic connections between North American and Eurasian hipparionine lineages. In order to define these relationships more precisely, BERNOR et al. (in progress) have undertaken a computerized Wagner 78 cladistic analysis of 25 Old World hipparionine species. Initial results of this study have strongly corroborated the phylogenetic patterns we cite here.

Fig. 1 illustrates the distribution of provincially first occurring Old World "*Hipparion*" species. Our investigation of this event has suggested that there is either a strong discordance in the dating of this "datum", or that the first appearance of "*Hipparion*" is diachronous throughout the Old World. Of the assemblage presented here, the most primitive species is the Central European form, "*Hipparion*" *primigenium*. As detailed by WOODBURNE et al. (1981) this taxon is morphologically very similar to the North American genus *Cormohipparion*. BERNOR and HUSSAIN (1985) argued that the species *Cormohipparion occidentale* is particularly similar to "*Hipparion*" *primigenium* in its development of a large, subtriangular-shaped, anteroventrally-oriented preorbital fossa, well defined buccinator fossa, shallowly retracted nasals and richly ornamented cheek teeth. "*Hipparion*" *primigenium* sensu strictu would appear to be restricted in its geographic distribution to the Central European Province (sensu BERNOR, 1983, 1984), and apparently underwent little provincial evolutionary change during the late Miocene. There is only one late Miocene locality in Central Europe, Dorn Durkheim (=Turolian age), where another species of "*Hipparion*" is known to occur.

The Chinese species "*Hipparion*" *weihoense* appears to be virtually identical to "*Hipparion*" *primigenium* in its skull and maxillary cheek tooth morphology, but has not yet been compared in its postcranial anatomy. The Spanish species "*Hipparion*" *catalaunicum* is relatively evolved in its more elongate facial fossa, but otherwise is closely similar to "*Hipparion*" *primigenium*. The North African Vallesian age species "*Hipparion*" *africanum* has a similar facial morphology to "*Hipparion*" *catalaunicum*, and in addition is known to have different limb proportions than "*Hipparion*" *primigenium* (BERNOR et al., in progress). The most primitive Siwalik species "*Hipparion*" *nagriensis* is poorly known, but appears to show a morphologic pattern very similar to "*Hipparion*" *primigenium*. However, its purported descendant "*Cormohipparion*" *(Sivalhippus) theobaldi* is known to retain the same facial morphology as "*Hipparion*" *primigenium*, but has distinctly higher crowned cheek teeth and strikingly more robust postcranials.

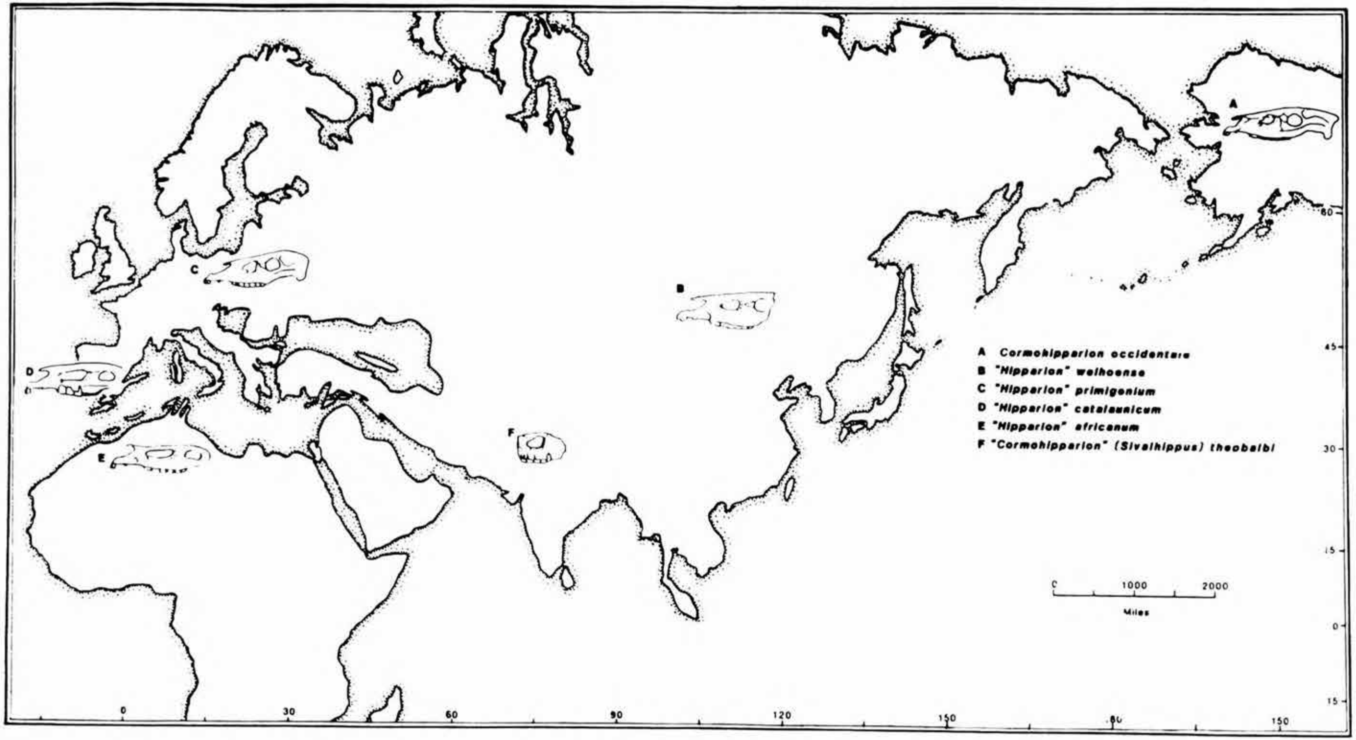


*Fig. 1.* Primitive Group 1 species of the "Hipparion Datum"

It would appear that the European first occurrence of "*Hipparion*" was 11 million years ago or more, while the occurrence in the Indian Subcontinent has been calibrated by OPDYKE et al. (1982) as having been no earlier than 10 million years. Recent investigations in East Africa and the Afar Triangle point to a 10 million year datum there also (BARRY, personal communication). Evolutionarily, Central Europe, China and the Siwaliks would appear to have the most primitive first occurring species. The peri-Mediterranean area, with an estimated first occurrence at 11 million years or more, has more evolutionarily advanced early Vallesian species. Our analysis suggests that there is a direct phylogenetic relationship between North American *Cormohipparion* and first appearing Old World hipparionines, that species evolutionary diversification was rapid in some Old World bioprovinces, while being virtually static in others, and that there probably was a significant diachroneity in the "*Hipparion*" Datum.

BERNOR and HUSSAIN (1985) have presented the most recent characterization of Old World hipparionine supraspecific groups. Fig. 2 illustrates what we have referred to as Group 1 derivatives, excluding taxa illustrated in the previous figure. China had an extensive evolutionary radiation of this group including "*Hipparion*" *dermatorhinum*, "*Hipparion*" *coelophys* and "*Hipparion*" *platyodus*. These taxa differ in details including size, snout proportions and slightly divergent facial morphologies. Their cheek are virtually identical in morphology. No postcranial morphological details have yet been discriminated for this group. In Western Europe there is evidence of an independent radiation of Group 1 hipparionines as is indicated by the species "*Hipparion*" *melendezi* which shows size reduction and an intermediate stage of facial fossa loss. In the eastern Mediterranean and southwest Asian region an advanced Group 1 horse with more elongate limbs, "*Hipparion*" *gettyi*, first occurs in early Turolian horizons. Another larger species, "*Hipparion*" *giganteum*, with an elongate snout, retracted nasals and reduced fossa, occurs in the western U.S.S.R. and the eastern Mediterranean.

BERNOR (1984; in press) has identified a number of lineages that would appear to be derived from advanced Group 1 species. This group has a known chronologic range of 8.5 to 4 m.y. and a geographic extension from Greece in the west, through southwest Asia and the western U.S.S.R., as far east as China (Fig. 3). Species of this lineage often develop multiple (3 or 4) well developed facial fossae and concomitantly retract the nasal bones, suggesting that they may have had a short, highly mobile proboscis and a specialized feeding apparatus. Two species of this lineage, "*Hipparion*" *matthewi* (southwest Asia) and "*Hipparion*" *richthofeni* (China) diverged from this pattern and instead underwent a marked reduction of the facial fossa, paralleling other lineages as we will discuss in a moment.

Group 3, or *Hipparion* s.s. (Fig. 4), had a geographic range extending from western Europe eastward through Greece, Iran and Pakistan. Species of this lineage show a striking reduction and eventual loss of the preorbital fossa. The cheek teeth undergo an evolutionary transformation from a primitive Group 1 complexity to a simplified occlusal ornamentation. The postcranials of these species are characteristically elongate and gracile in build and suggest an open country adaptation. MACFADDEN (1980, 1984) has argued that this lineage had its origin from Clarendonian age hipparionines based upon comparable facial morphology, but BERNOR and HUSSAIN (1985) and BERNOR (in press) have refuted this assertion, claiming evolutionary convergence in this character complex and citting differences between Eurasian and North American "*Hipparion*" s.s. in details of facial and cheek tooth morphology.

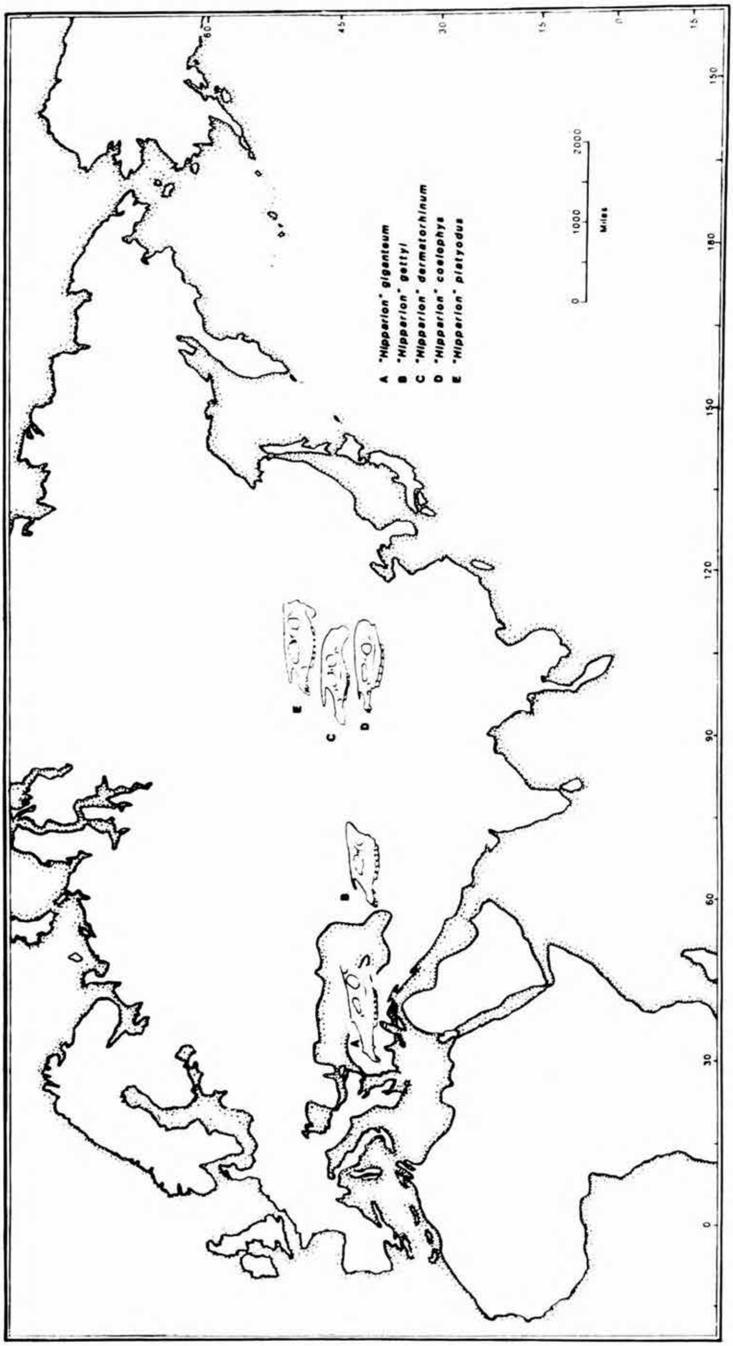


*Fig.* 2. Distribution of derived Group 1 hipparionine species

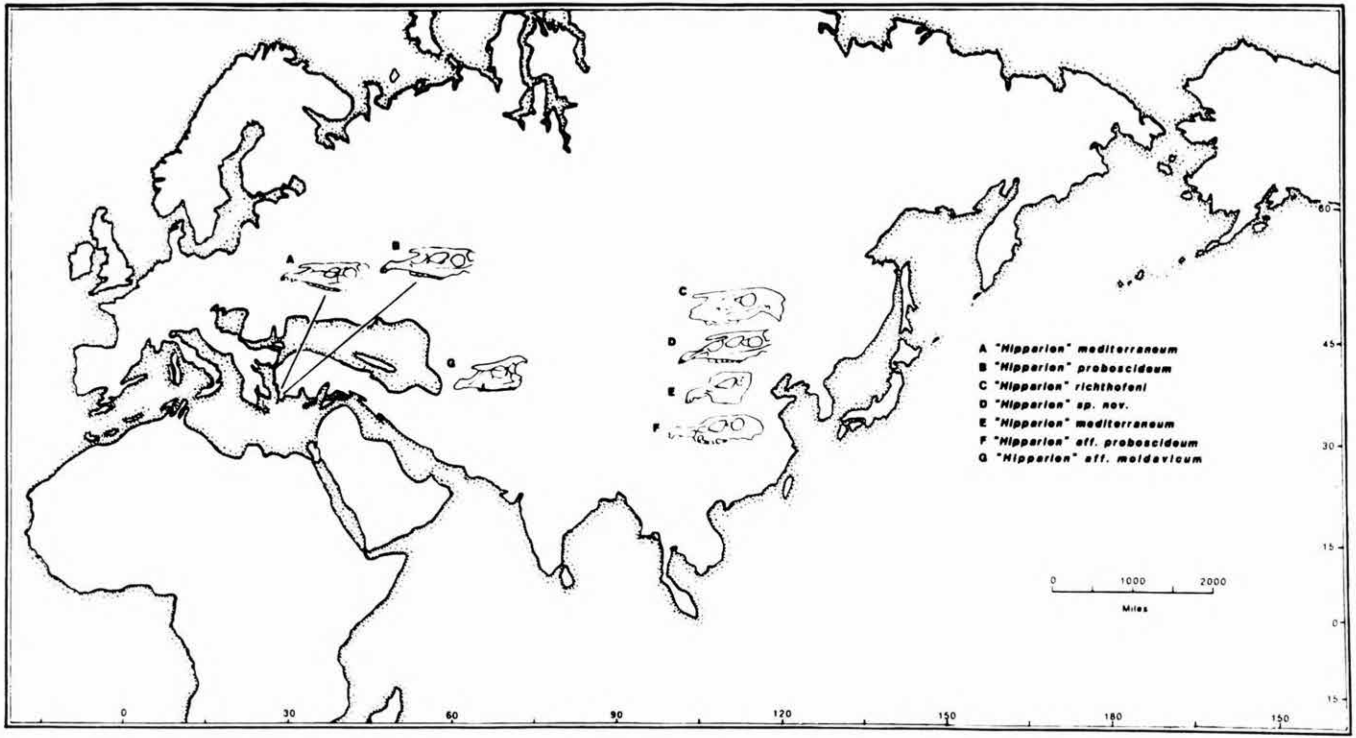


*Fig. 3.* Distribution of Group 2 hipparionine species

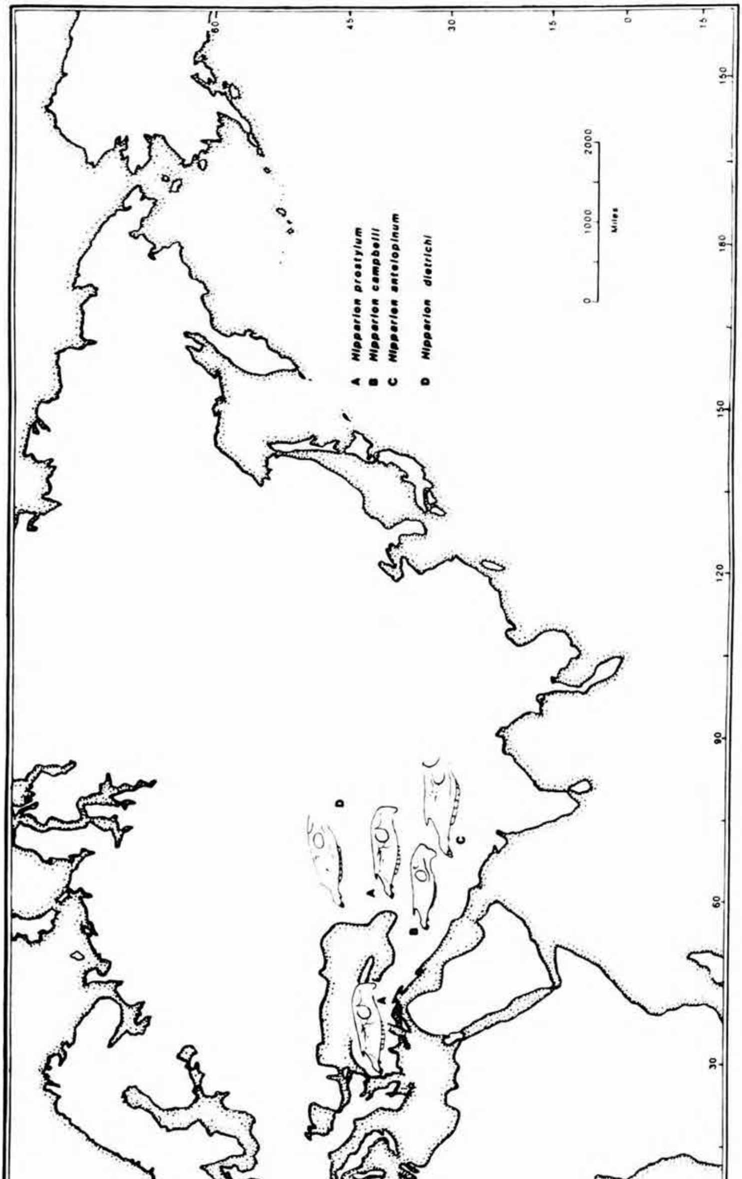


*Fig. 4.* Distribution of Group 3 hipparionine species

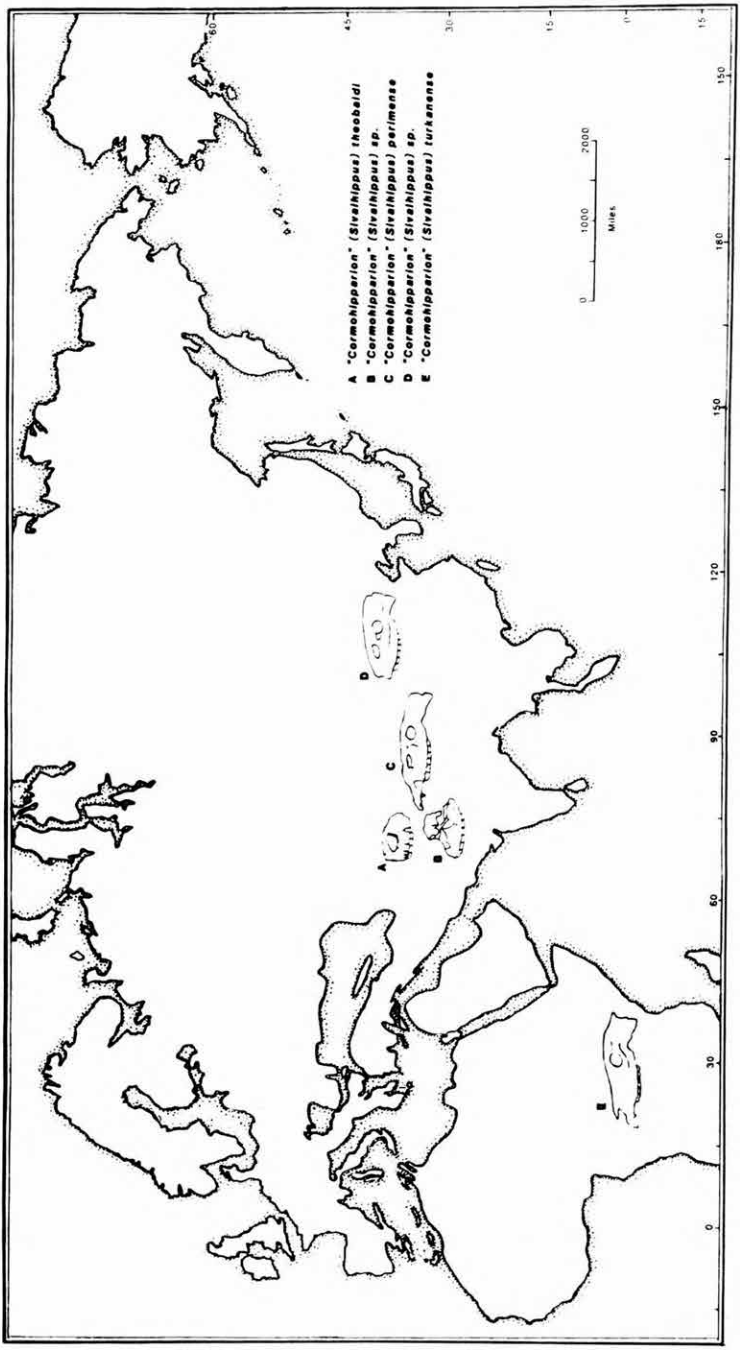


*Fig. 5.* Distribution of "Cormohipparion" (Sivalhippus) species

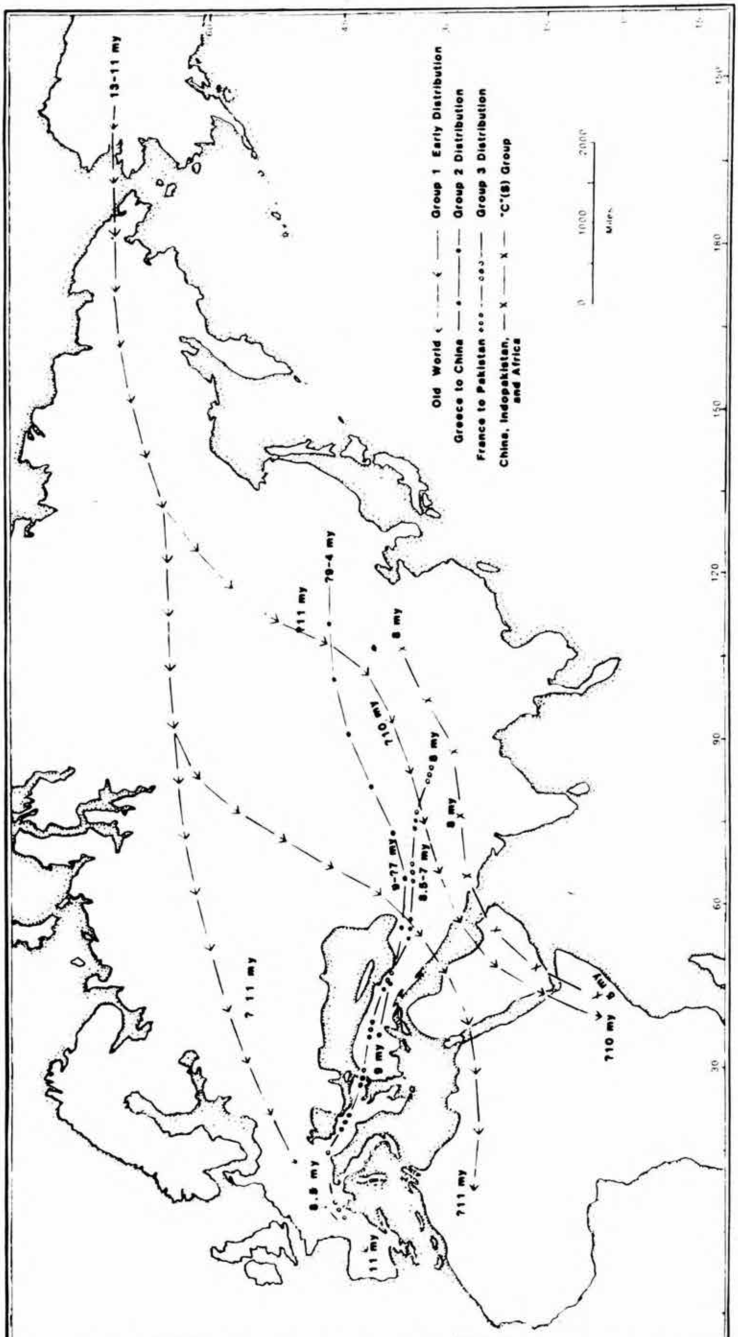


*Fig. 6.* Biographic tracks of Old World hipparionines

Yet another lineage, the "*Cormohipparion*" *(Sivalhippus)* group (Fig. 5), has been recognized by BERNOR and HUSSAIN (1985). This lineage includes "*Cormohipparion*" *(Sivalhippus)* sp. in China; "*Cormohipparion*" *(Sivalhippus) theobaldi*, "*Cormohipparion*" *(Sivalhippus)* sp. indet. and "*Cormohipparion*" *(Sivalhippus) perimense* in the Indian Subcontinent; and "*Cormohipparion*" *(Sivalhippus) turkanense* in East Africa. This lineage retains primitive characters of the cheek teeth, but shows an evolutionary transformation in preorbital fossa morphology which first places this structure far anteriorly on the face and eventually loses it altogether. In at least some species, the limb bones become very robust suggesting a possible woodland adaptation. The chronologic range of this lineage is estimated to be ca. 8 to 5 m.y. and potentially younger in East Africa.

Fig. 6 summarizes biogeographic tracks and chronologic ranges for the various supraspecific groups which we have discussed. Group 1 is known to range geographically from East Asia, through the U.S.S.R., Europe, southwest Asia and North Africa. The derived species of this group will, in many cases, probably be recognized as distinct lineages deserving supraspecific ranking. Chronologically, Group 1 species have a known range of 11 to 7 m.y. Group 2 ranges geographically from the eastern Mediterranean, through southwest Asia and the western U.S.S.R. to China. It has a known chronologic range of ca. 8.5 to 4 Ma. Group 3 extends geographically from western Europe, through Greece, Iran and Afghanistan as far as western Indopakistan. Its known chronologic range is ca. 8.5 to 6.5 Ma. The "*Cormohipparion*" *(Sivalhippus)* lineage ranges geographically from China, through Indopakistan and into East Africa. It has a known chronologic range of circa 8 to less than 5 Ma.

The phylogenetic and biogeographic results which we present today are not final; however, they do represent an ordered and congruent data set. When combined with recent work on Miocene Old World mammalian provinciality (BERNOR, 1983, 1984), this evidence suggests that regions which harbored the most active evolutionary diversification, such as the western U.S.S.R., western European, Subparatethyan and Northern Chinese Provinces, were ones where climatic change and biotic community responses were undergoing the greatest evolution. Areas of relative evolutionary stasis, such as the Central European Province, were apparently buffered from these climatic changes. These data reveal the great potential of this group, not only for. making intraprovincial to intercontinental scale correlations but also for pursuing studies of the relationships between climatic change, community evolution and biogeographic differentiation.

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# SEISMIC STRATIGRAPHY AS A TOOL FOR CHRONOSTRATIGRAPHY: PANNONIAN BASIN

by

GY. POGÁCSÁS

## Chronostratigraphic significance of seismic reflections

Seismic reflections are generated by physical surfaces in the rock, mainly acoustic velocity impedance contrasts along bedding, planes of stratal surfaces or unconformities. Therefore primary seismic reflections can be used as stratigraphic markers. The resolution of bedding units is limited by the resolution of the seismic waves.

The pysical surfaces that separate groups of strata within a sequence are essentially synchronous, so the seismic reflectors can be assumed to be time markers.

The unconformities can be identified according to the systematic terminations of reflections. The seismic reflections are chronostratigraphically significant because rocks above stratal or uniformity surface are younger than those below it (Fig. 1/A, B).

## Neogene seismic features in the Pannonian Basin

The Pannonian Basin is a Neogene—Quarternary system consisting of relatively narrow deep depressions 4 to 8 km depth that alternate with ridges of relatively high position 0.5 km to 2 km (KŐRÖSSY, 1980; POGÁCSÁS, 1980; 1985; KILÉNYI and RUMPLER, 1985). Fig. 2 reproduced from KILÉNYI and RUMPLER (1985) illustrates the depth of the pre-Tertiary basement.

Based on the seismic stratigraphic and tectonic features the following stages can be distinguished in the Neogene evolution and subsidence of the Pannonian Basin.

In the Early and Middle Miocene elongated depressions or troughs developed. Detrital sediments with nearby sources were deposited, and synsedimentary listric faults played a dominant role in controlling the structure of the troughs (HÁMOR, 1984; POGÁCSÁS, 1984).

It can be seen fairly well in the seismic profile intersecting the Kiskun depression (Fig. 3) that due to the backward moving of the listric faults the subsidence proceeded in several phases. The half grabens that are filled by folded and faulted Miocene sedimentary and volcanic rocks are characterized by different reflection pattern, truncated reflectors and diffractions. On the base of these reflectors it may be possible to distinguish between Miocene sedimentary units of different tectonic styles and thus to investigate the mechanisms responsible for the subsidence and sediment filling (see Fig. 3).

These formations (units 3 to 10 on Fig. 3) are overlain unconformably by Upper Miocene and Pannonian strata, sometimes with a large stratigraphic gap across the unconformity. This unconformity (Fig. 4) is very important from CH point of view (see MOLNÁR et al. in this volume).

The second phase of basin formation was characterized by rapid (Pannonian) subsidence of new depressions that were wider and less elongated than those of the

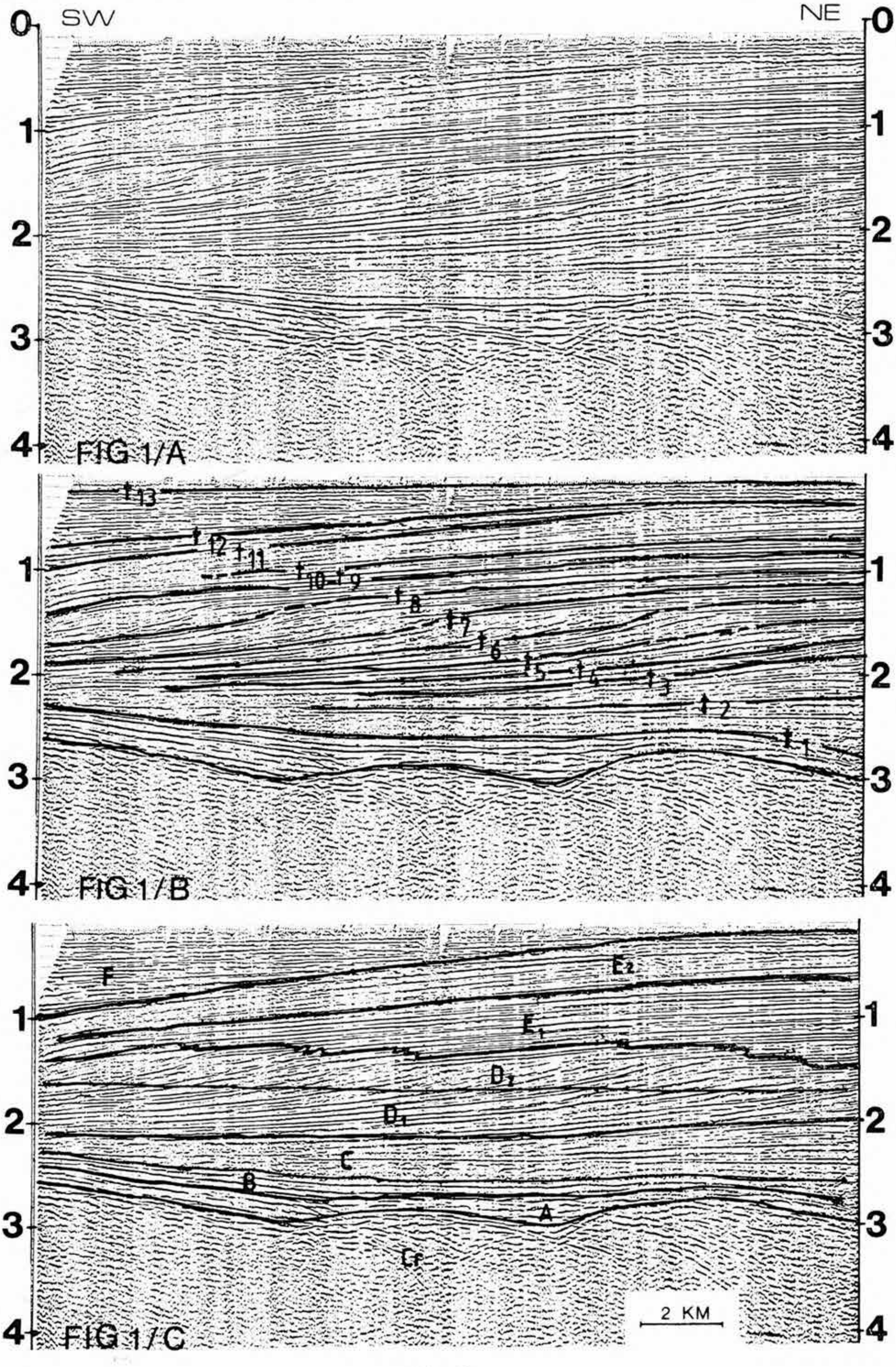
SW
NE
FIG 1/A
FIG 1/B
13
12
11
10
9
8
7
6
5
4
3
2
1
FIG 1/C
F
$E_2$
$E_1$
$D_2$
$D_1$
C
B
A
Cr
2 KM

*Fig. 1.*

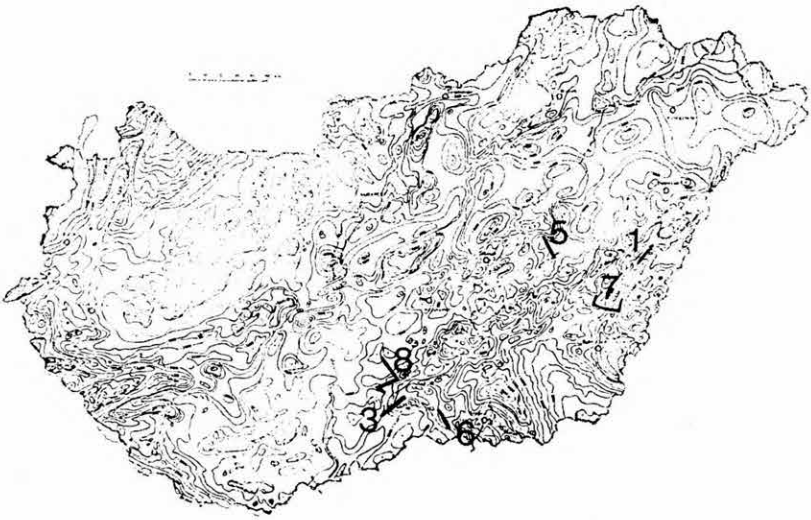


*Fig. 2.* Location map of the presented profiles

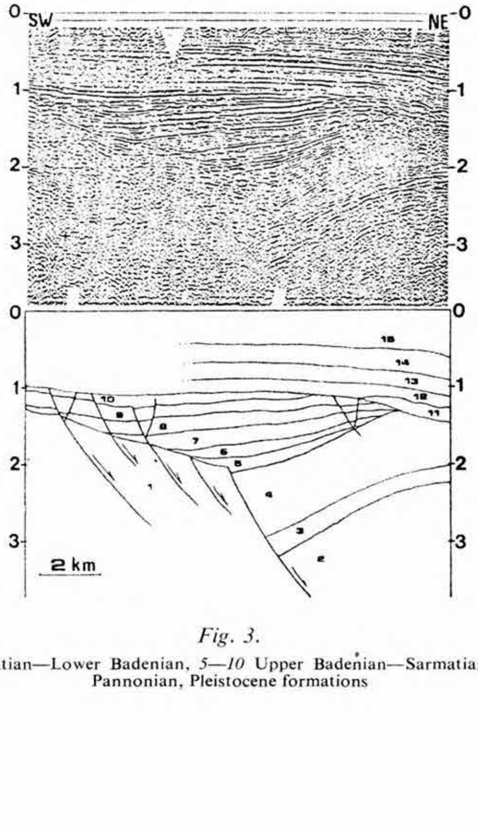


*Fig. 3.*
*1—2* Pre-Tertiary, *3—4* Karpatian—Lower Badenian, *5—10* Upper Badenian—Sarmatian (?), *11—15* Sarmatian (?), Pannonian, Pleistocene formations

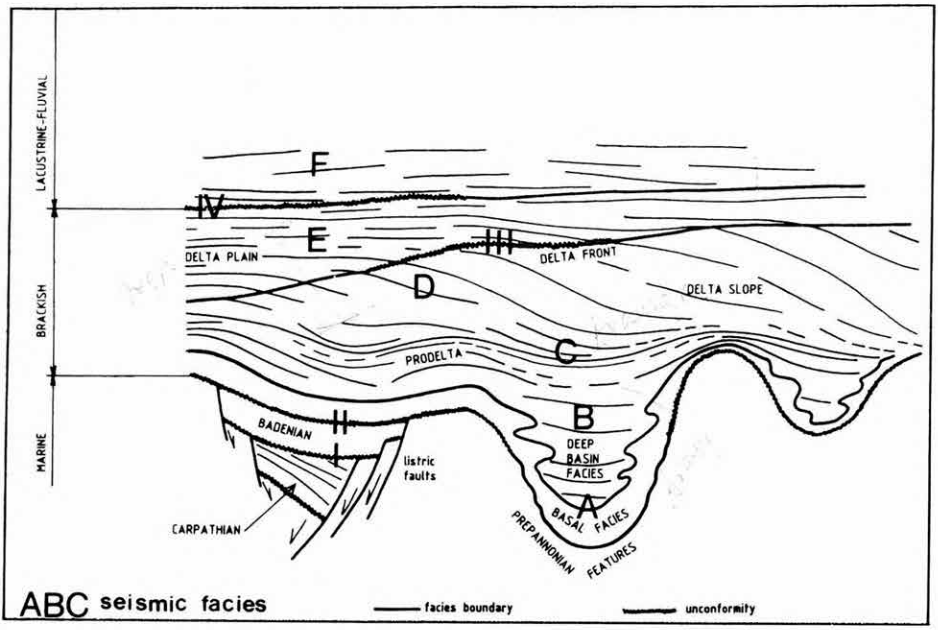


*Fig. 4.* General theoretical scheme of Neogene basin evolution and sediment filling of the Pannonian basin (modified after GAJDOS et al., 1983)

Middle Miocene phase. The trend of these new depressions bears little resemblance to that of the older depressions or to that of the pre-Neogene basement. At first the isolated depocenters subsided most rapidly and further on they became united. The several kilometers thick sedimentary sequence deposited during this phase. Different facies can be distinguished representing a great river dominated delta system (see POGÁCSÁS and RÉVÉSZ in this volume). In almost all Pannonian depocenter of the Pannonian Basin the seismic facies (representing different phases of the basin filling by prograding delta system) overlie each other in the same sequence. Nevertheless there are differences in the thickness, mode of occurrence and the number of seismic subfacies in the individual depressions.

Based on the lithostratigraphic units and subsequently overlying seismic facies the following main phase of palaeogeography and sediment filling can be stated.

In the area of deep depressions filling sedimentation took place in the Middle Miocene which continued also in the Early Pannonian (seismic facies A and B, Fig. 1/C). The Early Pannonian is characterized by gradual transgression (seismic facies C on Fig. 1/C). The transgression formations are overlain by regression sequences (seismic facies D and E on Fig. 1/C). The picture of seismic facies D characterized by prograding reflections reminds one of a delta front—delta slope and that, of facies E of a delta—plain and lagoon sediments.

This prograding sequence is present in most parts of the Pannonian Basin, but the unit is not continuous from one subbasin to the next. Outbuilding of sediments prograded from the marginal parts of the Pannonian Basin. Thus in the central part of the basin unit D, is younger that in more marginal areas.

Rapid filling of the Pannonian Basin terminated with the end of deposition of unit D. It is overlain by unit E, which may represent a partly contemporaneous delta plain facies, and indicates that a relatively even depositional surface had been established. This unit has characteristics typical of the third phase of basin subsidence (seismic facies F on Fig 1/C) that is fluvial and marshy sedimentation.

Within the Neogene sequence overlying the pre-Neogene basement by erosion gap generally four unconformities can be identified according to the systematic termination of seismic reflections (POGÁCSÁS, 1984) (Fig. 4).

i. The first unconformity separates the lower part of the Karpatian—Badenian sequence being tectonically deformed and constituting the starting member of the Neogene sequence, from the upper depositional unit represented by a seismic picture showing the original stratification of sedimentary origin.

ii. A marked unconformity surface is related to the bottom of the Pannonian formations except the deep depressions characterized by continuous deposition where this unconformity shows transition to correlative conformities.

iii. In the course of Pannonian delta filling the prograding delta front is accompanied by unconformity phenomena in the seismis profiles relating to local subaquatic redeposition and sediment removal.

iv. The fourth unconformity surface lies along the base of the Upper Pannonian—Pleistocene lacustrine—fluviatile formations.

Based on regional seismic profiles it can be stated that older depositional units of the Upper Miocene subside gradually into the depth form the marginal to the central parts of the basin, moreover, in many cases are downlapping on the dipping pre-Neogene basement surface.

The seismic features were compared with the published data of radiometric (BALOGH et al., 1983) and magnetostratigrapic measurements (RÓNAI and SZEMETHY; RÓNAI) in oder to investigate the chronostratigraphic position of the seismic units and the unconformities identified on seismic profiles.

The depth values of the radiometrically studied core samples were recalculated to seismic two way time data by using seismic velocity measurements.

The location of the wells providing the *K/Ar* and magnetostratigraphic data as well as location of the presented seismic profiles are seen in Fig. 2.

*1* The first *K/Ar* datum is the 18.25$\pm$0.3 Ma (BALOGH K. et al., 1983) determined in the depth of 1664 m in the boreholes Kisújszállás-NE-1 (Fig. 5).

The thin Miocene older than the Pannonian is overlain by the onlapping Pannonian sequence.

The 18 Ma age of the core samples at the base of the erosion hiatus gives a temporal limit but does not fix the start of the Pannonian transgression. Seismic profile fairly well shows a left lateral strike slip fault zone. This fault was still active during the deposition of unit F.

*2* The basalt from borehole Ruzsa-4 overlies the Badenian sequences with unconformity and is covered by Lower Pannonian sediments (Fig. 6). The depth of the studied basalt sample is 2657 m its age being 10.4$\pm$1.8 Ma (BALOGH K. et al., 1983) The thickness of the Badenian sequence is close to the seismic resolution in borehole Ruzsa-4.

The age datum of borehole Ruzsa-4 is valuable since it fixes the completion of the half-graben evolution developed along the listric fault southwest of the borehole. Nevertheless, the question remains open whether the listric faults bounded half grabens filled by Middle Miocene of the other parts of the country characterized by similar

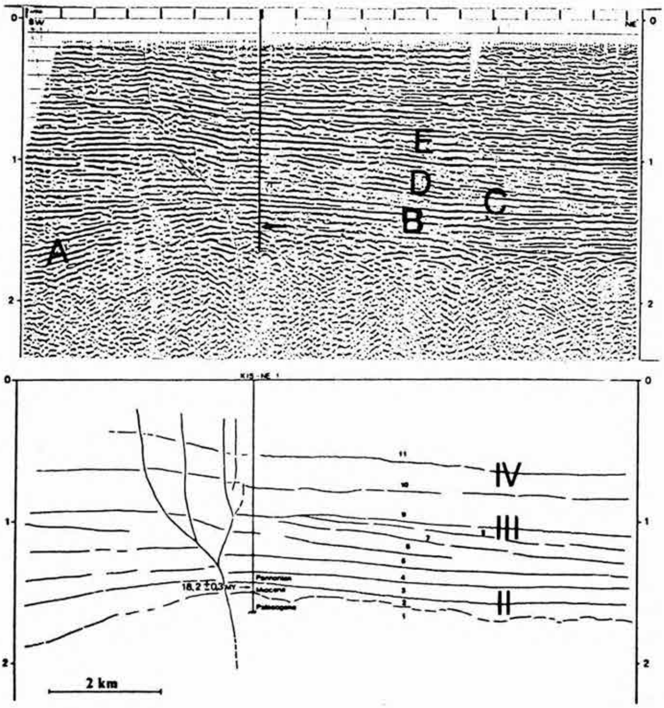
SW
NE
A
B
C
D
E
IV
III
II
18,2 ± 0,3 MY
Pannonian
Miocene
Paleogene
2 km

*Fig. 5.*

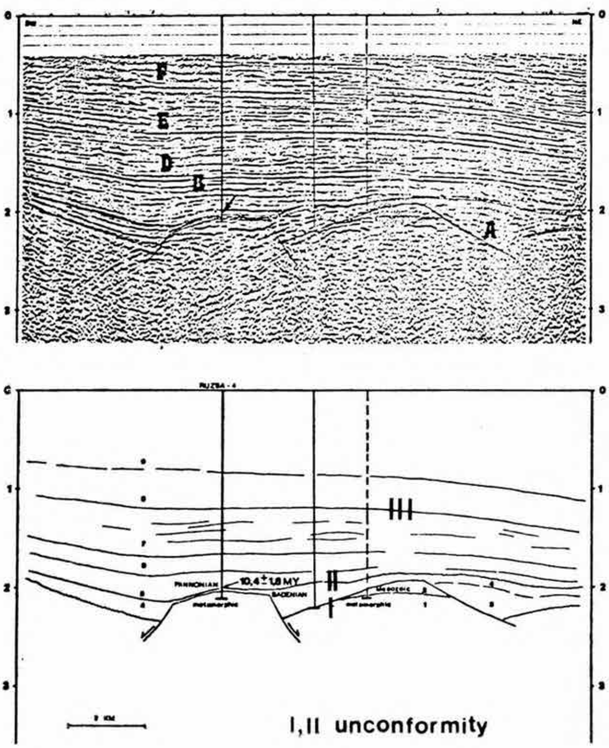
F
E
D
B
A
III
II
I
PANNONIAN
BADENIAN
10,4 ± 1,8 MY
I,II unconformity

*Fig. 6.*

tectonics are synchronous or not. The formations of the prograding delta sequence between 1300 and 1600 msec were deposited fairly well after the eruption of basalt lava of 10 million years.

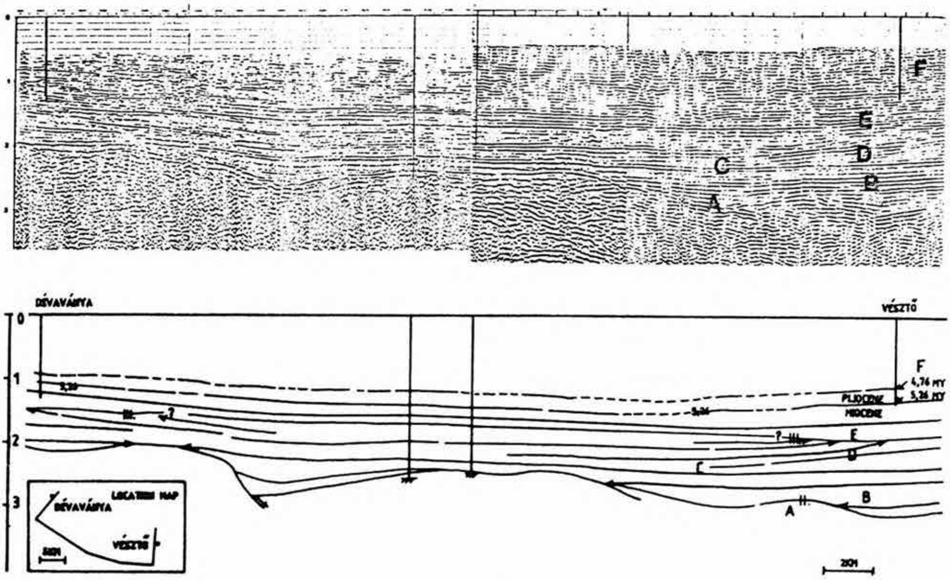


*Fig. 7.*

Seismic profile on Fig. 7 connects the Dévaványa and Vésztő boreholes drilled in eastern Hungary and studied by magnetostratigraphic method (Rónai and Szemethy, 1979; Rónai, 1981). In the Vésztő borehole the horizon of 5.26 Ma denoting the Miocene—Pliocene boundary is in a depth of 1250 m. Based on the seismic reflections the strata of the Vésztő boreholes can be fairly well correlated with the corresponding strata of the Dévaványa borehole. Based on the seismic correlation of the individual strata the Dévaványa borehole seems to explore the same formations in higher structural position. The seismic profile clearly demonstrates that the horizon of 5.2 Ma identified by magnetostratigraphic method lies by about one kilometre above the prograding delta sequence. The ages of 10.4 and 9.6 Ma fixed by the basalts in the boreholes Ruzsa-4 and Kiskunhalas-W-3 provide an upper limit to the age or the unconformity surface related to the base of the Pannonian. The prograding delta sequence showing characteristic seismic picture is obviously younger than 8.8 Ma in region of the profile.

*3* The next composite seismic profile (Fig. 8) gives possibility to compare the Kaskantyú-2 exploration well studied by magnetostratigraphic method (Hámor et al., 1985) with the *K/Ar* data of borehole Kiskunhalas-W-3.

The age of the basalt traversed in a depth of 1162 m in borehole Kiskunhalas-W-3, has been determined to be 9.61±38 Ma (Balogh et al., 1983). The coinciding reflexion horizonts with the first purely sedimentary sequence overlying the sequence consisting of the alternation of basalt lava, pyroclastics and sedimentary formations, can be fairly well followed in the depression lying in the centre of the profile, but it is wedged in its opposite emerging slope. Having correlated the in-

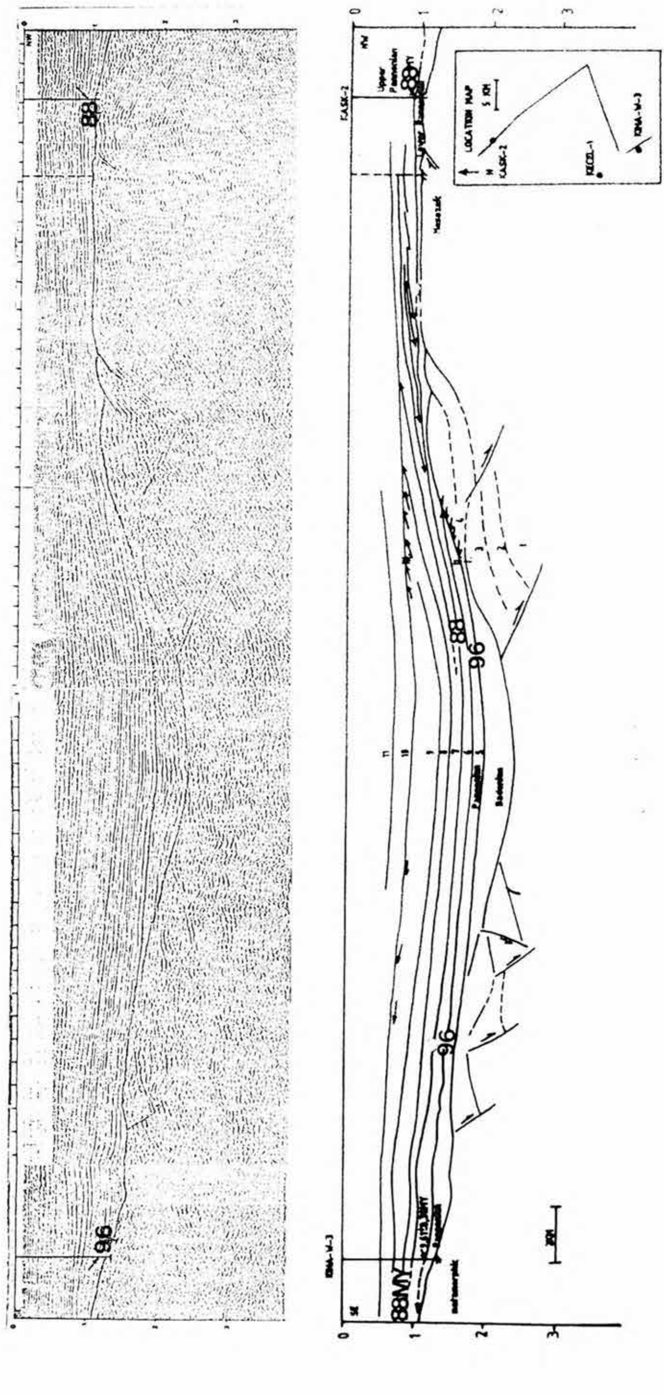

*Fig. 8.*
*1—3* Lower and Middle Miocene, *4* Upper Badenian, *5—6* Lower Pannonian, *7—11* Upper and Younger Pannonian →reflexion termination

complete reflections coinciding with the basalt of 9.61 Ma it can be seen that this horizon lies also above the unconformity surface coinciding with the base of the Pannonian.

As was stated by HÁMOR et al. (1985) a formation group of 8.8 Ma is in a depth of 865 m in borehole Kaskantyú-2 . The seismic reflection coinciding with this depth can be correlated in total length of the profile and can be traced up to borehole Kiskunhalas-W-3. The seismic profile clearly indicates that the horizon of 9.6 Ma is wedged in the surface of the trench-filling pre-Pannonian Miocene formations showing compressional structure.

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# PALYNOLOGY AS A STRATIGRAPHIC TOOL: THE WESTERN MEDITERRANEAN NEOGENE RECORD

by

J.-P. Suc

The use in the Mediterranean Neogene deposits of palynomorphs (pollen grains spores and microplankton) is steadily increasing for several reasons:

— their abundance in a very large number of localities,

— their permanence along thick sections,

— their presence in marine, coastal and lacustrine deposits.

Nevertheless, to be successful as a stratigraphic tool, Palynology is to be employed with the greatest caution. Neogene palynological studies consider the chronologic succession of floristic elements from a qualitative and/or a quantitative point of view. The increasing number of pollenanalysed localities reveals very important floristic differences between provinces (even adjacent) and it is very dangerous to intend to establish suitable palynostratigraphic correlations without elementary precautions:

— establishment of palynological reference series in limited geographic areas,

— detailed bio- and chronostratigraphic calibration of these reference sections by independent methods (foraminifers, nannoplankton, mammals, radiometric datings, paleomagnetic measurements),

— quest of lateral variations according to environmental or drifting changes.

L. Benda (Benda et al., 1977a and b; Becker-Platen et al., 1977; Benda and Meulenkamp, in press) established in accordance with the above-mentioned concepts a palynostratigraphic zonation from Late Oligocene to Earliest Pleistocene (six sporomorph associations) in the eastern Mediterranean region. The recognition of these sporomorph associations is chiefly based on the successive extinctions of "exotic elements" and the progressive increase of modern taxa (herbaceous mainly). Such a palynostratigraphy must be limited to a reduced area avoid geographic distortions. One must point out the difficulty to provide elaborated informations about flora, vegetation and climate, botanic attributions being only given for the generally well-known pollen-types.

## *The botanic conceptual approach*

The increasing knowledge of the modern world pollen flora allowed us to apply the Recent Quaternary pollen methodology which has been already successfully used in the Netherlands Plio—Pleistocene by W. H. Zagwijn (1960). According to the high diversity of the Mediterranean Neogene pollen flora, an unprecedented endeavour in pollen grains botanic determinations undertook sixteen years ago (Suc, 1976 and 1980; Diniz, 1984a; Bessedik, 1981 and 1985). Almost a hundred taxa (chiefly at the genus level, less often at the family one, sometimes at the species one) have been mentioned for the first time in the Mediterranean Neogene pollen flora, some of them were already known according to macrofloral remains. Nevertheless, such an approach is not devoid of problems. For instance, it is generally impossible to exceed the genus level because of the high variability in pollen morphology and the possible presence of extinct species. Moreover, if the determination at the species level is sometimes available, its exact ecological significance cannot be entirely guaranteed because of the peculiarity of the Neogene Mediterranean vegetational assemblages. A long period of investigations seems to-day necessary to overtop this obstacle and to reach a new step in the vegetational ecosystems reconstruction.

A consequence of this botanic way in determining is the establishment of detailed pollen diagrams just as those of the Late Quaternary. They constitute a paleobotanical document able to evidence the vegetational changes among which one must recognize those to be related to the climatic evolution (Cravatte and Suc, 1981; Suc and Cravatte, 1982; Diniz, 1984a and b; Bessedik, 1985; Zheng, 1986). The quantitative evaluation of the richness in pollen grains is measured for all the analysed samples and can be used to discriminate between different interpretations. Beyond the stratigraphic use of the successive extinctions* of climatically classified "exotic taxa", I consider that conclusive palynological progresses for the Neogene Stratigraphy in the whole Mediterranean region will proceed from the paleoclimatic understanding of pollen analyses.

### *The Miocene record*

M. Bessedik (1984 and 1985) applied for the first time this methodology to the Miocene deposits. He described the vegetational coastal ecosystems (*Avicennia* mangrove along shoreline, largely developped semi-arid open associations, forests with some tropical elements in humid places) of Lower and Middle Miocene and inferred a climatic (thermic as well as pluviometric) evolution from their respective representation. Nevertheless, this climatic curve which can be compared to the oceanic thermic continued curves (Müller, 1984; Vergnaud Grazzini, 1984) cannot be used from a climatostratigraphic point of view because of its discontinued aspect and the low duration of the investigated sections. The Upper Miocene data are now too fragmentary and partly contradictory. So, a climatostratigraphic pollen zonation of the western Mediterranean Miocene does not appear yet feasible.

### *The Plio—Pleistocene record*

Six main pollen zones (P I, P II, P III, P IV—Pl. I, Pl. II, Pl. III) have been evidenced for Pliocene and Lower Pleistocene from off-shore sections in Gulf of Lion and Adriatic Sea and from outcrops in southern France and northeastern Spain (Suc, 1982). The noted vegetational changes are referred to pluviometric (especially seasonal distributions) but also to thermic variations. Climatostratigraphic relationships have been proposed with northwestern Europe (Suc and Zagwijn, 1983; Zagwijn and Suc, 1984) then extended to southwestern Europe (Diniz, 1984a).

* Date and sequence vary from place to place.

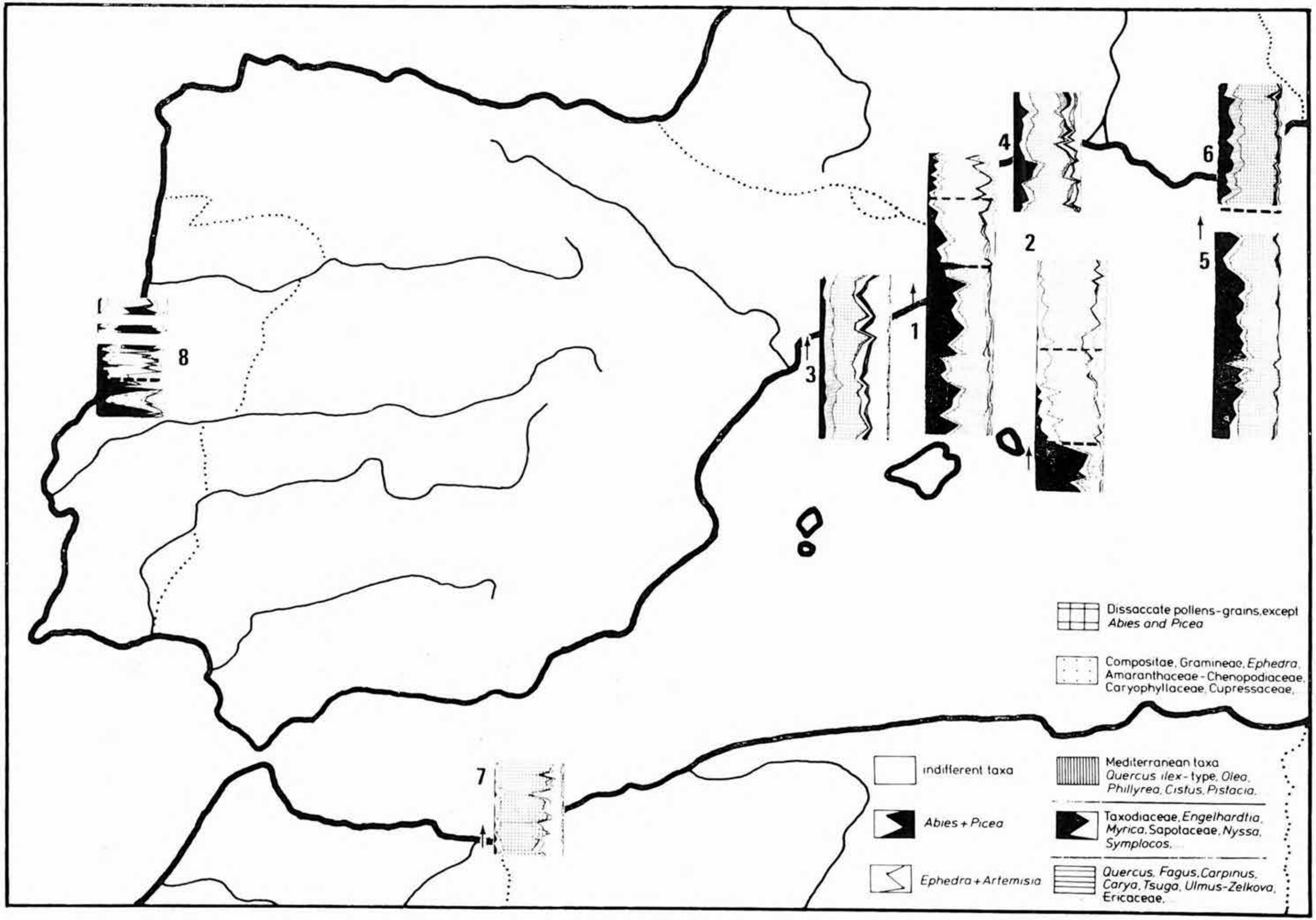


*Fig. 1.* Compared Pliocene synthetic pollen diagrams from the western Mediterranean region

*1* Garraf 1 (SUC and CRAVATTE, 1982), *2* Autan 1 (CRAVATTE and SUC, 1981), *3* Tarragona E 2 (BESSAIS, 1984), *4* Cap d'Agde 1, Languedoc-Zanclean (SUC, unpublished), *5* Saint-Martin-du-Var, Nice region-Early Zanclean (ZHENG, 1986), *6* Saint-Isidore, Nice region—Earliest Piacenzian (ZHENG, 1986), *7* Alboran Sea (SUC, unpublished), *8* Rio Maior F 16, Portugal (DINIZ, 1984). Legend: † Last occurrence datum of *Globorotalia margaritae*, - - - - - = P I—P II pollen zones boundary, - - - - - = P II—P III pollen zones boundary

From approximately −2.3 Ma (Late Pliocene), it is possible to date fields (inside the chronologic intervals independently defined) by reference to the glacial—interglacial northwestern Europe pollen curve (ZAGWIJN, 1982). So, several sections in southern France, northeastern Spain and southern Italy have been chronologically calibrated with a high level of precision (SUC, 1982; BRENAC, 1984); COMBOURIEU—NEBOUT, in preparation). The climatostratigraphic correlations are based on the correspondance both in northwestern Europe and northwestern Mediterranean region between forest phases (interglacial or interstadial) and open vegetation phases (glacial; respectively tundra-like and steppic associations).

Before this period characterized by Quaternary-type climatic cycles, climatostratigraphic correlations (local and between south and north Europe) are more difficult to establish according to the high diversity of the Mediterranean vegetation and to the low effects of the first Arctic coolings in the Mediterranean region (Messinian to earliest Piacenzian—Fig. 1). For instance, pollen diagrams from humid localities (from an edaphic point of view: Garraf 1, Autan 1, Rio Maior; from a climatic point of view: coastal Alps) reveal the passage between pollen zones P I and P II (which occurred just after the disappearance of *Globorotalia margaritae)* and even for some of them the secondary fluctuations inside P I pollen zone (Fig. 1). Pollen diagrams from less humid localities (Languedoc: Cap d'Agde 1) or from dry regions (Southern Catalonia, Algeria) only reveal minor changes even not at all (Fig. 1). In fact, the first kind of localities was rich in "exotic elements" which were sorely affected by the successive decreases in moisture and probably in temperature.

Moreover, regional correlations are also difficult to establish due to the vicinity of very different biotopes.

To conclude, I would like to emphasize some ideas:

— in general, palynology must be more considered as a stratigraphic correlations tool than a dating tool;

— whichever method, palynology has now to overcome an essential problem: how to enlarge its space validity (from local to regional inference before the whole Mediterranean basin), i.e. how to appreciate and supervise the peleogeographic differences due to latitude, marine and continental influences, mountains vicinity, . . . ;

— the climatostratigraphic method seems to be more utilizable in the recent periods (Pliocene and Quaternary, precisely from Late Pliocene) because of the relative vicinity (at the difference of Miocene) of the cold source;

— for this approach, marine deposits seems to be more useful for general climatic reconstructions because the relative pollen diversity does not directly appear in relation to the sedimentary paleoenvironments (sometimes, alas, to the preservation); moreover, such studies allow a direct confrontation with the other climatostratigraphic methods (e.g. the isotopic curves).

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# PLIOCENE AND PLEISTOCENE MARINE—CONTINENTAL CORRELATIONS

by

D. TORRE

*I* The main changes in climate shown by the continental biohistorical events can also be recognized in marine sequences, thus providing a first basis for correlations. The most reliable correlation is between the dispersal event that marks the beginning of the Montopoli faunal unit (MN 16B; G 3) and the climatic change that occured at about 2.5 Ma. At this time foraminiferal oxygen isotopic records from many DSDP sites show a significant shift which is interpreted as representing an ice volume growth in the northern Hemisphere (THUNELL and WILLIAMS, 1983). A sudden increas of ice rafted debris occurred in the North Atlantic and the North Pacific at 2.4—2.5 Ma (SHACKLETON et al., 1984; REA and SCHRADER, 1985). Relevant changes in the nannofossil assemblages are recorded round the world. In a 0.2—0.3 Ma interval all the *Discoaster* species faded out, except for *D. brouweri*, which remained as a rare species in the western Mediterranean (MUELLER, 1978) and in northern Italy (RIO et al., 1982) after 2.5 Ma. The benthic foraminifers were also affected by a reduction in diversity (SPROVIERI, in press; VAN DER ZWAAN, 1983). The pollen analyses show a strongly marked change in vegetation, calibrated at about 2.3 Ma, which produced tundra-like flora in northern Europe (Praetiglian unit) and a steppe environment in the Mediterranean region (P III—pollen zone) (SUC and ZAGWIJN, 1983; ZAGWIJN and SUC, 1983). As a matter of fact the Montopoli faunal unit corresponds to the most arid phase in the Pliocene mammal sequence but this is not the only factor in correlation. Many other lines of evidence support the correlation with the above mentioned biostratigraphic and paleoclimatic data. The Montopoli finds come from littoral deposits capping a marine sequence of the *Globorotalia crassaformis* zone (DE GIULI et al., 1983); LINDSAY et al. (1980) have placed the fossil site in a reverse magnetic polarity interval just above a section with normal polarity. Thus the transition from the Gauss to the Matuyama epoch is very likely to be represented. In the Jucàr valley (eastern Spain) the appearance of the Rincon 1 fauna, characterized by abundant *Gazella borbonica* and by a Montopoli-like largesize *Equus*, corresponds to a marked climatic deterioration evidenced by isotopic analysis in the lacustrine beds (ALBERDI et al., 1982; LEONE, 1985). In the Central Massive te Roca Neyra local fauna, which has provided the oldest record of *Equus* and the last record of *Hipparion* in France, has a *K/Ar* date of 2.5 Ma, according to SAVAGE and CURTIS (1970) or younger than 2.4 Ma date, according to MENG HOUR, CANTAGREL, DE GOER DE HERVE and VINCENT (1982). In the Dutch Praetiglian an *Equus* sp. was collected together with *Anancus arvernensis*, *Archidiskodon* cf. *gromovi* and a deer cf. *Eucladoceros falconeri*, the latter being the most primitive species of this genus. In southern Russia the *Archidiskodon—Equus* dispersal event marks the transition to the Khaprovian "complex", typified by the finds of Kapry, near the Azov Sea, and of Livenwovka, near Rostov on the Don. The Russian authors placed this event at about 2.5 Ma (SCHAN-

ZER, 1982). In the Indian subcontinent the arrival of *Equus*, along with *Rhinoceros*, *Elephas hysudicus* and *Stegodon insignis*, has been calibrated at the Gauss—Matuyama transition (LINDSAY et al., 1980; AZZAROLI and NAPOLEONE, 1982). In conclusion, the correlation of the Montopoli faunal unit with the upper part of the *G. crassaformis* zone can be considered sufficiently established.

*2* From the correlation of the Montopoli dispersal event with the shift in climate that occured at about the Gauss—Matuyama transition, it follows that the faunal change going on from the Perpignan to the Triversa unit (MN 15—16A;, G 1—2) is very probably connected with the climatic variation recorded at about 3.2 Ma. At this time an enrichment in 180 is observed at Site 132 (western Mediterranean) as well as in Atlantic and Pacific DSDP Sites. THUNELL and WILLIAMS (1983) have

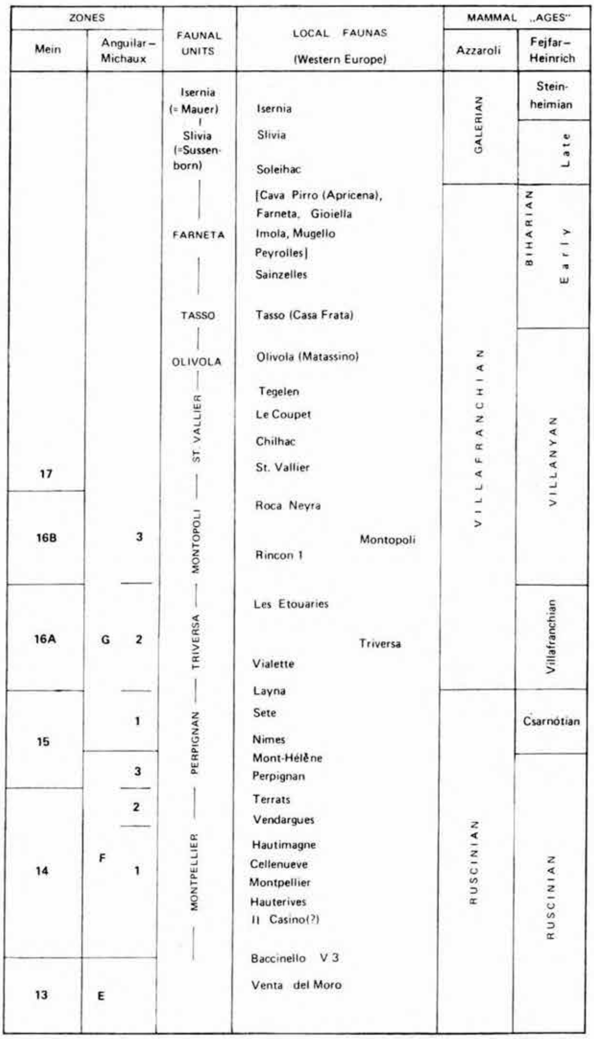


*Fig. 1.* Plio—Pleistocene mammal units of Europe

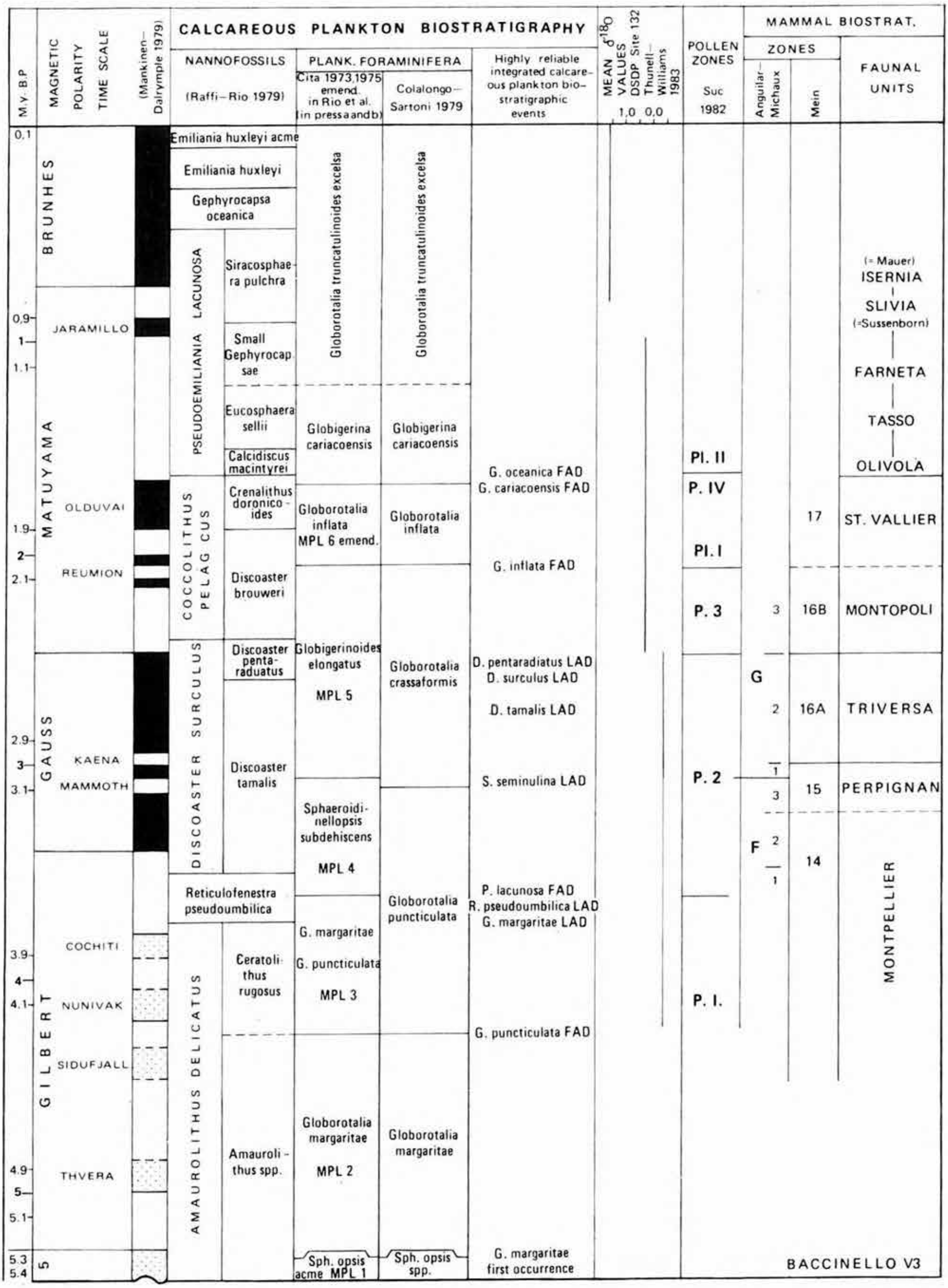


*Fig. 2.* Marine—continental stratigraphic framework of the Plio—Pleistocene of the Mediterranean area. Marine data from COLALONGO et al. (1984)

interpreted this datum to reflect a cooling event in surface waters. In Iceland, glacial deposits are dated approximately 3.1 Ma (McDOUGALL and WENSINK, 1966). In the Mediterranean many bio-events are observed around 3.2 Ma. About 10% of the benthic foraminiferal species died out (SPROVIERI, in press) and *Globorotalis crassaformis* immigrated together with *Globorotalia bononiensis*. These events were followed by the extinction of *Sphaeroidinellopsis* and *Globoquardrina altispira*. The correlation between the faunal change during the Perpignan unit resulting in the production of the Triversa fauna and the events observed in marine sequences at about 3.2 Ma is supported by several lines of evidence. The lacustrine beds around Villafranca d'Asti, where the Triversa finds were collected, represent the deposits of a shallow coastal lake. To the east, they change to a marine littoral facies capping a sandy clay section with *Uvigerina* cf. *rutila* nearby Valleandona (COLALONGO et al., 1972). LINDSAY et al., (1980) found a short normal polarity interval sandwiched between two reversed episodes at the Fornace RDB quarry, one of the typical Triversa fauna sites. They referred these episodes to the Kaena and Mammoth events. At Arcille (Grosseto, Italy), *Mimomys polonicus*, a typical rodent of the Triversa fauna, is present in a lignitic bed of a continental section overlying marine sediments referable to the *Globorotalia puncticulata* zone. The Vialette fossil site has been dated by BANDET et al. (1978) at a *K*/*Ar* age close to 3.3 Ma and probably not younger than 2.6 Ma. At Les Etouaires SAVAGE and CURTIS (1970) reported a radiometric age of 3.4–3.5 Ma for an ash bed underlying the fossil bed but more recently MENG HOUR et al. (1982) have pointed out that the fauna of this locality cannot be older than 2.6 Ma. All these elements make the correlation of the Triversa unit (MN 16A, G 2) with the lower part of the *Globorotalia crassaformis* zone very reliable. The local faunas of the Perpignan unit, which represent a period of climatic stress, may be correlated with an interval at about the transition between the *Globorotalia puncticulata* and *G. crassaformis* zones. The data on La Juliana (Murcia, Spain) are not in contrast with the present correlation. At this site lignitic marls with a rodent faunule, which can be ascribed to Perpignan unit, overlie marine beds with *G. crassaformis* and *G. puncticulata* (MONTENAT and DE BRUIJN, 1976). According to these marine stratigraphical data the age of the La Juliana rodents is somewhat younger than might be expected but the difference could stem from the uncertainty of correlation. Alternatively, the rodents could be somewhat retarded in this southernmost part of Spain. In Val di Pugna (Siena), *Alephis lyrix*, the bovid found at the Perpignan site, was collected in sediments of the *G. puncticulata* zone. SUC (1982a, b), SUC and CRAVATTE (1982), SUC and ZAGWIJN (1983, 1984) have correlated the Languedoc and Roussillon local faunas of the Montpellier and Perpignan units with the interval defined by the *Globorotalia margaritae* and *Sphaeroidinellopsis* last occurrence on the grounds of the palynostratigraphies obtained from continental sections and from the Autan 1 and Garraf 1 boreholes (western Mediterranean). Thus the faunas of the Montpellier unit should be compressed within the *G. puncticulata* zone. This conclusion is also supported by the location of the mammal fossil sites in the regressive phase of the Pliocene marine cycle of the Rhône basin and in the overlying continental deposits (BALLESIO, 1972). It is thus not possible to refer the Montpellier faunas of the Languedoc to the earliest Pliocene. Also the paleomagnetic data recorded in the Languedoc and Roussillon sites are not in conflict with this stratigraphic conclusion. The mammal fossils that come from the upper lignitic complex of Il Casino site (Siena) could be older than the Montpellier faunas but they have poor stratigraphic control: the only certainty is their position between gypsum beds, ascribed to the Messinian and Pliocene marine sediments of the *G. puncticulata* zone. The Baccinello V3 local fauna could represent the transition to Ruscinian fauna but its relationship to the marine stratigraphy is still undefined. The mammal fossils come from the upper part of continental deposits capped by marine sediments with *G. margaritae*. At present it is not possible to tell whether these mammal finds are correlated with the late Messinian or with the earliest Pliocene.

*3* As far as the mammal sequence following the Montopoli unit is concerned, the Pliocene local faunas where correlation may be possible are Tegelen, Le Coupes and Chilhac. At Tegelen most fossils were collected by quarry workers in previout years and their stratigraphic location was not exactly recorded, but it seems that there are very few doubts about their provenance from the Tc1—Tc6 interval of the Tiglian pollen unit (ZAGWIJNG, 1974). VAN MONTFRANS (1971) calibrated the Tiglian with the part of the Matuyama epoch that ranges from, about the oldest Reunion event to the Olduvai top. In the French Central Massif the bulk of the mammal finds of Le Coupet have an estimated age not older than 1.9 Ma, while the Chilhac fauna should not be younger than that age (BOEUF, 1983). The correlation of these two local faunas with the Tiglian pollen unit seems to be an obvious conclusion. Therefore it is indirectly inferred that the St. Vallier faunal unit is roughly correlated with the *Globorotalia inflata* zone. This statement is consistent with the fact that the appearance of *G. inflata* in the Mediterranean is correlated with the immigration of *Sphaeroidinella dehiscens* and *Globorotalia truncatulinoides*, showing the transition to a somewhat warmer climate (ZACHARIASSE and SPAAK, 1983). Evidence of a milder climate at this time have been also found in the Po basin (GASPERI et al., 1982). The local faunas representing the late Villafranchian faunal interval have to be located between the Olduvai and Jaramillo events in the Matuyama magnetic epoch. The dispersal event that marks the beginning of the Olivola faunal unit could be correlated with the transition from the Tiglian to the Eburonian pollen-phase; but no direct climatic evidence is presently available to support this inferred correlation. If the hypothesis were right and the beginning of the Olivola unit were so placed at the end of the Olduvai palaeomagnetic event, then it would be roughly correlatable with the beginning of the Pleistocene as defined at the Vrica section (Calabria, Italy). The Soleihac site, where the mammal assemblage already has a main Galerian character, is calibrated at the Jaramillo event (THOUVENY and BONIFAY, 1984). The Sainzelles finds, which are correlatable with the Tasso fauna or with a somewhat younger one, are located in a reverse magnetic interval underlying a basalt flow dated 1.3—1.4 Ma and resting on deposits with a normal polarity attributed to the Olduvai event (THOUVENY and BONIFAY, 1984). The Peyrolles, one of the local faunas, which mark the transition to the Galerian unit, has been very recently dated at about 1.2. Ma (CANTAGREL et al., 1985). The Imola (Po basin) and Pirro Nord (Apricena, Gargano) sites are surely younger than the *Hyalinea baltica* appearance in the Mediterranean.

*4* During the Jaramillo paleomagnetic event the mammal faunas have already attained a predominant Middle Pleistocene feature. This is recognized in western Europe (Soleihac fauna) and in Central Asia. The Lakhuti 2 fauna of Tadjikistan is a Middle Pleistocene mammal assemblage with few late Villafranchian hold-overs (AZZAROLI, 1983) and is situated by DODONOV (1980) at the top of the Jaramillo. The local faunas of Slivia, in Italy, Voigtsted, Süssenborn and Stranska Skala 2, in central Europe, are even more modern-looking. The dispersal of *Arvicola* marks the beginning of a new faunal unit characterized by the classic faunas of Mauer and "Graunes" Mosbach and by the finds of the new site or Isernia La Pineta (central Italy). The "Graues" Mosbach site has been referred to the Brunhes paleomagnetic epoch but the Isernia fauna has been calibrated with the late Matuyama reserved epoch and is immediately overlain by volcanics dated to .73±.04 Ma and .73±.07 Ma by two different laboratories (COLTORTI et al., 1982). Since the paleomagnetic and radiometric dates of the Isernia fauna seem reliable, and assuming that *Arvicola* is a valid marker, the normal polarities observed at Stranska Skala 2 and Voigtsted cannot be referred to the Brunhes, as these sites should be older than Isernia. Their normal magnetic fields should correspond to the Jaramillo or possibly to minor fluctuations within the late Matuyama. The stratigraphic problems of the early Middle Pleistocene remain to be worked out: many local faunas must still be adequately dated. It is clear that the renewal of the mammal fauna was practically completed during the time between the Jaramillo and the beginning of the Brunhes normal epoch. The biogeographic event that completed the faunal turnover must be correlated with the transition to the glacial Pleistocene, in turn, indicated by a significant shift in the mean values of the oxygen isotopic records from DSDP Sites.

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# PRELIMINARY CORRELATION OF OLIGOCENE TO PLEISTOCENE PHYTOSTRATIGRAPHIC UNITS OF THE MEDITERRANEAN AND THE PARATETHYS AREA

by

E. Velitzelos and H.—J. Gregor

An attempt is made to correlate the Paratethys floral zones (sensu Mai 1967: I—XIII and Gregor 1982 OMM to OSM-4) with floral complexes in the Mediterranean area, with the aid of the mammal units (sensu Mein) and other units. As many phytozones are still lacking in the regions under research we have to correlate typical floras of important localities.

In the Lower to Middle Oligocene we can compare Cereste (France) with Haselbach (GDR), the Upper Oligocene Spanish floras (for example Tarragona) with the (Molasse)-floras from W Germany (f.e. Peienberg, Rott) and Switzerland (Toggenburg) (Fig. 1a), or the Edirne flora (Turkey) with English ones (Bembridge). In the Lower Miocene it is possible to correlate the Mastixioidean floras from the Oberpfalz (Schwandorf, Phytozone OMM) and the Rhenish area (Eschweiler) or the Eastern countries (Turow, Hradek, Wiesa) with the one from Arjuzanx (France) and later (MN 4b) the Aliveri-flora (Greece) with those from Langenau and Niederpleis and Troisdorf (FGR), belonging to phytozone OSM-1 (Ottnangian Fig. 1b). In the Karpatian and the Badenian one can state the similarity of phytozone OSM-2 (incl. 3a) floras with Turkish ones (Saray) or Portuguese (Povoa, Fig. 1b). Middle Badenian to Sarmatian and Pannonian leaf floras from Germany (OSM-3b, -4), from Oehningen and Achldorf (FRG), Domanski Wierch (Poland), Vienna basin (Austria) etc. are comparable to those of Vegora (Greece), Soma (Turkey), Castellina mar. (Italy) etc. (Fig. 1c). Messinian and Pliocene Mediterranean floras from Italy (St. Barbara, Stirone), France (Cessenon, Pichegu), Greece (Pikermi) and others show a similar composition than the floras from Frankfurt or Frechen, the Molasse (FRG) and the Alsace (France) (Fig. 1c).

Climatic belt surely have not existed in the present-day sense. Only areas of comparable floral compositions, special forest systems can be distinguished. The authors made an attempt to work combined with index fossils like Ceratostratiotes (Ottnangian only bound to coastlines), Toddalia and Zanthoxylum different species, mesophytic elements (see Fig. 2) and to plot distribution maps for special species in space and time.

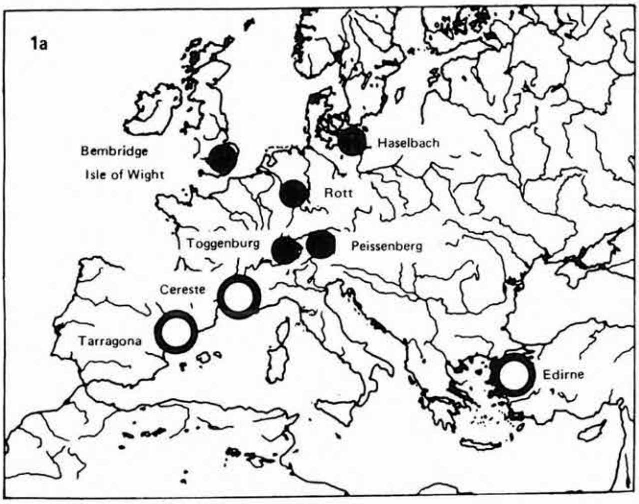
In the Uppermost Miocene we have a model in Europe concerning rich floras which we can compare. If the facies is equal, the age also is the same-proved by mammal units (see Fig. 3).

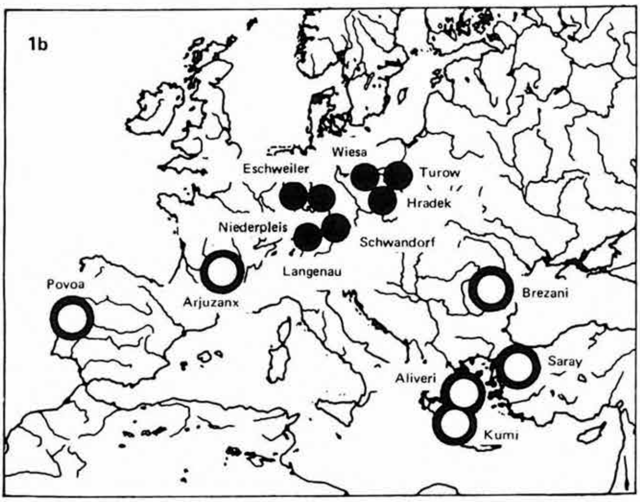
At least we try to give a chart with a number of fossil floras or localities and the floral zones as we have it in the Paratethys area [incl. mammal and pollen-zones (see fig. 4)].

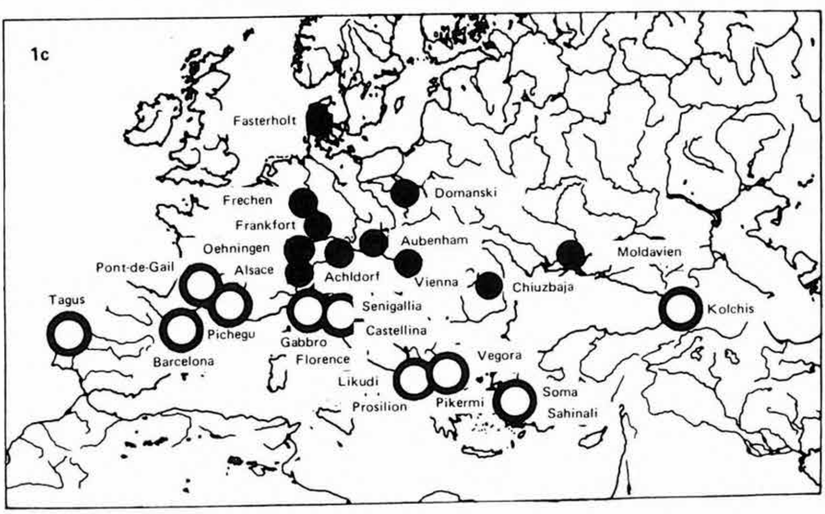
The future work will be, to define the floras typical of the Mediterranean and to compare them geographically and to establish new foral zones (Fig. 4). The relevant literature is very extensive. Some of the most important with more citated literature you find here.

*Fig. 1. a, b, c* Fossil floras from different localities in Europe

Paratethys area (black dots) and Mediterranean region (grey ring with white dot) from the Oligocene (1a) to the Lower Miocene (1b) and the Upper Miocene—Pliocene (1c)



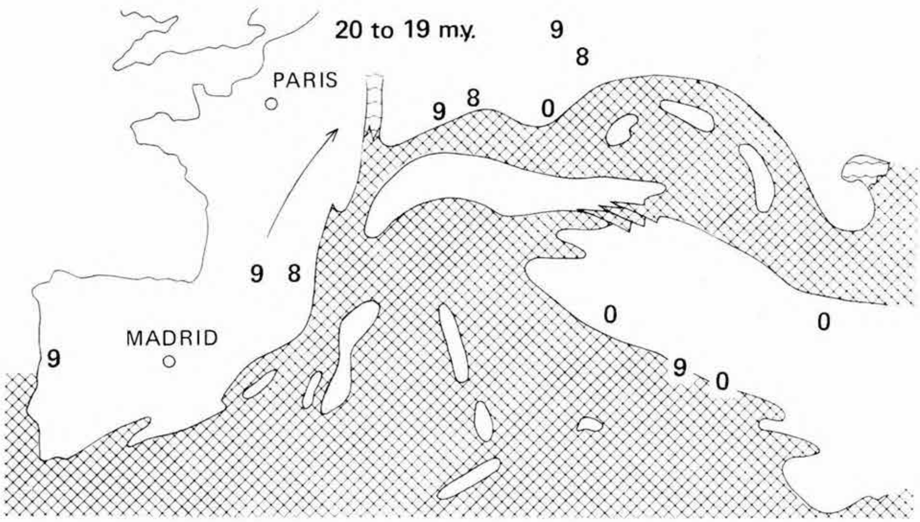



*Fig. 2.* Distribution of *Ceratostratiotes zapfei* (0) along the Otthangian coastline, of *Toddalia* div. spec. (9) and *Zanthoxylum* sp. (8) in Europe (first model)

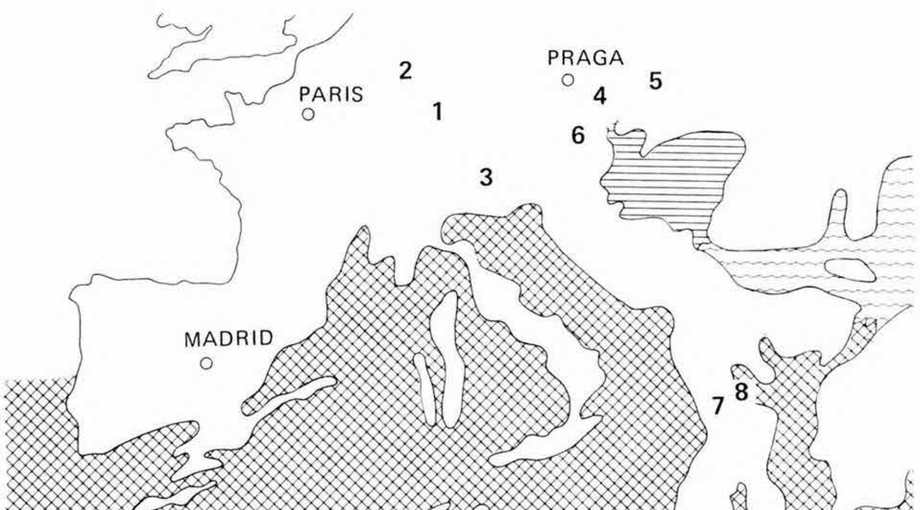



*Fig. 3.* Important leaf- and fruit-floras of the Uppermost Miocene—Pliocene (Pannonian s. l.) in Europe—comparable by their equivalent floral composition

*1* Mainz area, *2* Fischbach-clay, *3* Achldorf, *4* Moravska Nova Ves, *5* Ruszow, *6* Vienna basin, *7* Vegora, *8* Likudi

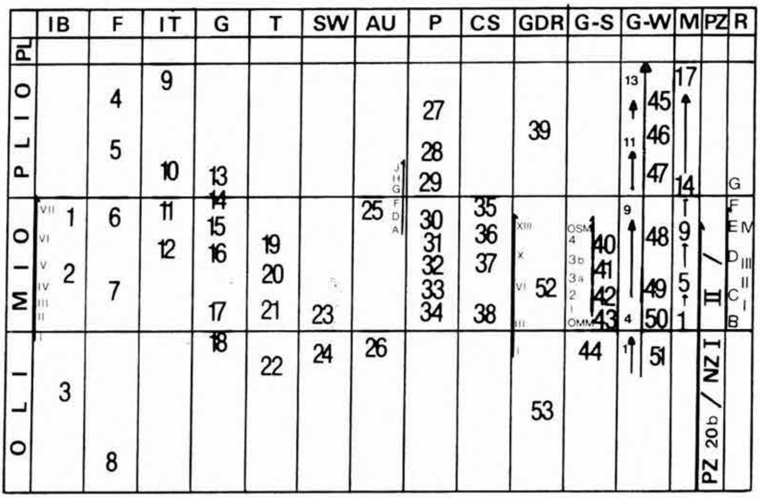



*Fig. 4.* Preliminary table for fossil floras from the Oligocene to the Pleistocene from Spain to Turkey (Mediterranean) and from Switzerland to Poland (Paratethys-area) and the floral zones as they exist today (micro- and macrofloras)

OLI = Oligocene, MIO = Miocene, PLIO = Pliocene, PL = Pleistocene, IB = Iberia, F = France, IT = Italy, G = Greece, T = Turkey, SW = Switzerland, AU = Austria, P = Poland, CS = Czechoslovakia, GDR = German Democratic Republic, G—S = West-Germany—southern part, G—W = West-Germany—western (Rhenish) area, M = Mammal units sensu Mein, PZ = European pollen zones (sensu Krutsch), R = Rhenish pollen zones (sensu v.d. Brelie et al.). — IB: Floral zones I—VII (sensu Pais), AU: Pannonian A—J (sensu Papp), GDR: floral zones I—XIII (sensu Mai), G—S floral zones OMM to OSM-4 (sensu Gregor), G—W: lithozones 1—13 (sensu Schneider et Thiele). — *1* Tagus, *2* Povoa, *3* Tarragona, *4* Pichegu, *5* Cessenon, *6* Pont-de-Gail, *7* Arjuzanx, *8* Cereste, *9* Stirone, *10* Florence, *11* Gabbro, *12* Castellina, *13* Skoura, *14* Pikermi, *15* Likudi, *16* Vegora, *17* Aliveri, *18* Kumi, *19* Sahinali, *20* Soma, *21* Saray, *22* Edirne, *23* lower freshwater molasse, *24* Toggenburg, *25* Vienna floras, *26* Eferding, *27* Mizerna, *28* Kroscienko, *29* Czorsztyn, *30* Domanski Wierch, *31* Stare Gliwice, *32* Wieliczka, *33* Nowy Sacz, *34* Turow, *35* Moravian floras of Vienna basin, *36* Moravska Nova Ves, *37* Cermniky, *38* Hradek, *39* Rippersroda, *40* Achldorf, *41* Oehningen, *42* Langenau, *43* Schwandorf, *44* Peissenberg, *45* Willershausen, *46* Frankfurt, *47* Fortuna, *48* Frechen, *49* Eschweiler, *50* Niederpleis, *51* Rott, *52* Wiesa, *53* Weisselster

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Proceedings of the VIIIth RCMNS Congress

# EVALUATION ON LOWER AND MIDDLE VILLAFRANCHIAN CHRONOSTRATIGRAPHY

by

M. T. ALBERDI and F. P. BONADONNA

In order to correlate the Villafranchian chronostratigraphy it is necessary to have many series for which it is possible to date the successive steps of Villafranchian stratigraphy.

First, we shall examine two deposits, Montopoli (Tuscany, Italy) and Casas del Rincon (Spain), Middle Villafranchian in age, to have some firm points for the top of Lower Villafranchian. The Montopoli section is constituted by marine clays with *Strombus coronatus* (Middle Pliocene) covered by yellow-red sands with *Ostrea* in which the vertebrate Montopoli fauna was recovered. The yellow-red sands are deeply eroded; resting on them, in a deposit few kilometres away, begins the cycle with northern guest *(Arctica islandica)*. For this reason we think it is possible to correlate the erosion on the top of Montopoli section to the Acquatraversan erosional Phase (3.4—2.5 Ma). This means that the Montopoli fauna was deposited in pre-Acquatraversan time. Furthermore a paleomagnetic study by LINDSAY et al. (1980) shows that the age of the Montopoli sands in very near to the Gauss—Matuyama paleomagnetic boundary (2.47 Ma) so the paleomagnetic age agrees with the stratigraphical one; AZZAROLI et al. (1982) regard the Montopoli fauna as Middle Villafranchian. The Middle Villafranchian age of the Montopoli fauna is further supported by the Spanish deposit of Casa del Rincon (level Rincon 1).

The Rincon 1 fauna is characterized by the same faunistic elements as Montopoli (for instance the same *Equus* and probably the same *Dicerorhinus*, ALBERDI and BONADONNA 1983, ALBERDI et al. 1982) and is equally considered Middle Villafranchian. The series (50 m thick), situated in the Southern Spanish Meseta, is formed by carbonate layers of a marshy and lacustrine environment and it is transgressive on other lacustrine series in which Ruscinian fauna was found. In the whole series close isotopic measurements were performed (LEONE, 1985) to determine the paleoclimatic trend of the deposition epoch. From the oxigen isotopic composition of carbonate layers and fresh water gastropods it is possible to build a paleoclimatic curve (Fig. 1); the comparison of this curve with THUNELL's curve (1979) for the Mediterranean paleoclimate shows that the sequence cold—warm (probably arid)—cold of Rincon is mirrored on THUNELL's sequence, in which the first cold has an age of 3.1—3.2 Ma the warm stage 2.7—2.6 Ma and the last cold 2.6—2.5 Ma. The Rincon 1 fauna is located in the second cold of the sequence while Rincon 2—3 just below the first one; the age of Middle Villafranchian fauna of Rincon 1 deposit is so fixed at 2.5—2.6 Ma.

*The problem of Lower Villafranchian.* The marine clays of Montopoli section are coeval to lacustrine clays of Upper Valdarno (Castelnuovo dei Sabbioni Group) in which only Lower Villafranchian fauna was found. According to LY et al. (1982) the age of the beginning of the Lower Villafranchian is the age of Les Etouaires de-

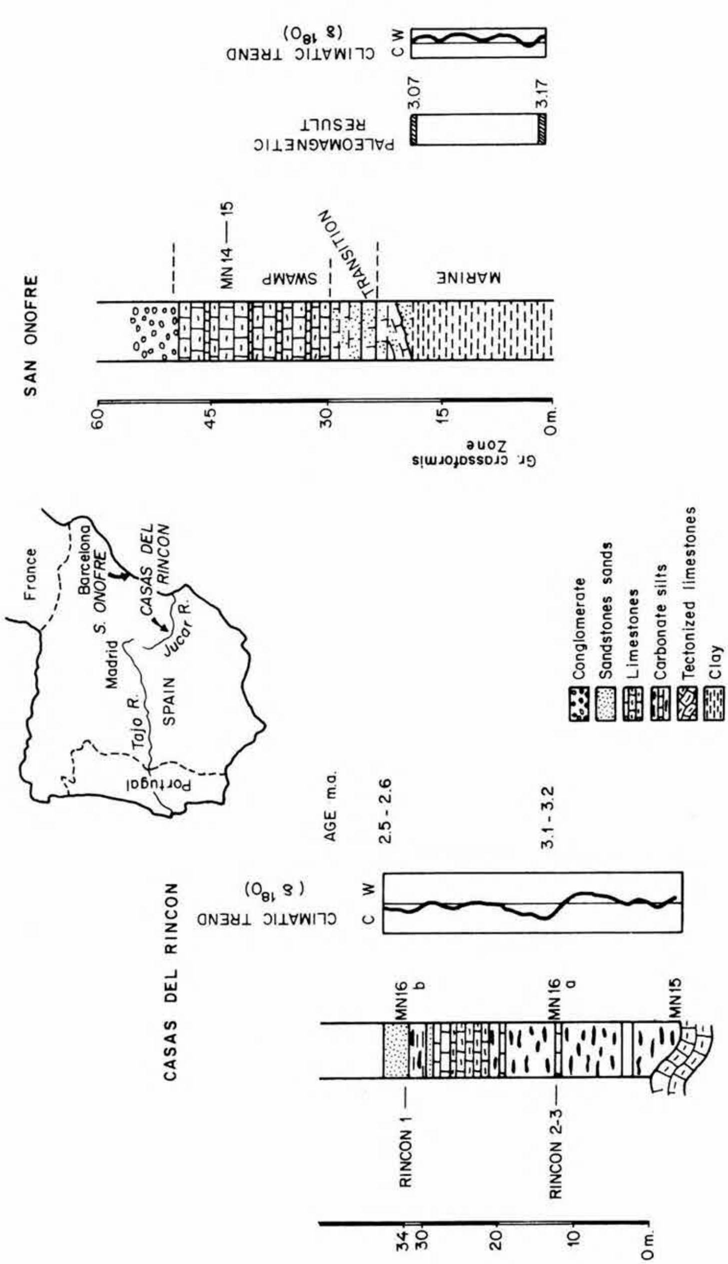
SAN ONOFRE
PALEOMAGNETIC RESULT
3.07
3.17
CLIMATIC TREND ($\delta$ $^{18}O$)
C W
MARINE
TRANSITION
SWAMP
MN 14 — 15
Gr. crassaformis Zone
60
45
30
15
0 m.
France
Barcelona
S. ONOFRE
CASAS DEL RINCON
Madrid
Jucar R.
Tajo R.
SPAIN
Portugal
CASAS DEL RINCON
CLIMATIC TREND ($\delta$ $^{18}O$)
C W
AGE m.a.
2.5 - 2.6
3.1 - 3.2
MN16 b
MN16 a
MN15
RINCON 1
RINCON 2-3
34
30
20
10
0 m.
Conglomerate
Sandstones sands
Limestones
Carbonate silts
Tectonized limestones
Clay

*Fig. 1.*

posit (between 2.6–2.4 Ma). The Les Etouaires deposit is strictly linked to a big erosion following the Grand Nappe volcanic event (the same erosion almost totally destroyed the fibrous pumices deposit linked to the Grand Nappe), so that the sediments of the deposit, resulting from a big erosion, are constituted by elements of different ages. This is indeed the criticism of LY et al. against the age (older than 3 Ma) obtained by fission tracks by CHAMBAUDET and COUTHURES (1981) and BIQUAND et al. (1981) on minerals from Les Etouaires deposit: according to LY et al. (1982) these minerals derive from some eruption older than the Grand Nappe event. But in such a geological situation the fauna also may be older than the deposit itself; it may derive from older nearby deposits, destroyed by the erosion. In our opinion the age of Grand Nappe volcanic event, obtained by LY et al. (1982) indirectly cannot date the faunistic assemblage of Les Etouaires; the 2.6–2.4 Ma age may be the age of the deposit as a sedimentological unit and not the age of the different elements, organic and inorganic, of the deposit.

Anyway the age of Les Etouaires deposit cannot be considered as the beginning of Lower Villafranchian both for the geological reason mentioned above and because of the existence of other deposits, at least four in all Mediterranean Europa of Lower Villafranchian age, but older than 2.6–2.4 Ma according to their radiometric and/or paleomagnetic ages.

The Vialette deposit (France) is concordingly considered Lower Villafranchian as it contains some typically Villafranchian animals as *Cervus pardinensis* and *Dicerorhinus jeanvireti;* it was reported by BANDET et al. (1979) to have a 3.3 Ma *K/Ar* apparent age (COUTHURES, 1979):, the dated samples show some alterations which caused almost certainly a loss of potassium. The age is thus questionable for two reasons: firstly the very existence of alteration at an unspecified time, secondly possible hydrocarbon. The contamination of hydrocarbons in Neogene–Quaternary volcanites has been demonstrated in Italy (I. M. VILLA, personal comunication) and Spain (BONADONNA and VILLA, 1984). The hydrocarbons have a mass spectrum that interferes with argon in all the masses between m/e=35 and m/e=45; for *K/Ar* measurements, the presence of isotope m/e=36 is very dangerous since the atmospheric correction and hence the age calculation is based on 36 *Ar* isotope. BONADONNA and VILLA (1984, Table 1) show what may happen if hydrocarbons are not under control. The first column lists the *K/Ar* results obtained with a routine gas purification. The ages are in rather good stratigraphical order: indeed VH3 representing the base of the series is the oldest but ages are geologically meaningless. The second column shows the *K/Ar* ages of the same series (Las Higueruelas deposit, Spain) obtained with the special procedure described in BONADONNA and VILLA (1984) to eliminate hydrocarbons: the results are clearly different. This problem is not mentioned for the Vialett eand Les Etouaires measurements: so we think that the published age of Vialette may possibly be only a minimum age for the deposit; furthermore pollen stratigraphy (LE GRIEL, 1983) states that the age of Vialette has to be older than 2.7 Ma, so that, the age 2.6–2.4 Ma for the Lower Villafranchian is contradicted again.

For the Vialette deposit we have another age (3.8 Ma , SAVAGE and CURTIS, 1970) which was contested by BANDET et al. (1979) because it contradicted the rodent biostratigraphy. The criticism of these authors is based on the age of La Juliana deposit (Spain) considered by MONTENAT and DE BRUIJN (1976) as Early Ruscinian and correlated to Late–Middle Pliocene (younger than *Globorotalia crassaformis* zone). After some researches in the field, in our opinion, the stratigraphy of La Juliana deposit is not so clear as MONTENAT and DE BRUIJN maintain. Moreover, the deposit was completely removed by agricultural activities and it is now impossible to subject it to a geological—paleontological revision. Furthermore, the definition of MN 14 and MN 15 (the rodent levels of La Juliana) in Spain is rather confused As it can be seen in Fig. 1, in the Tortosa area (San Onofre section) the level named "MN 14 or beginning of MN 15" by AGUIRRE et al. (1982) is found in a swamp sediment overlying the *Globorotalia crassaformis* zone, i.e. MN 16 b (also note that the top of the *Globorotalia crassaformis* zone was found to be 2.1 Ma old in Vrica section, Italy). On the other hand, in La Gineta deposit (Jucar valley, Spain), MN 15 underlies a conspicuous discordance (Ibero—Manchega I) which in turn underlies the Casas del Rincon series (Fig. 1).

In the latter series, rodents are only found in MN 16a and MN 16b (Early and Middle Villafranchian respectively, ALBERDI et al., 1982). The age of the entire series was determined by isotope paleoclimatology (LEONE, 1985) to lie between 3.2—3.3 and 2.6—2.5 Ma. Furthermore, the isotopic record of the marine clay sequence of San Onofre *(G. crassaformis* zone, also analyzed for paleomagnetism) indicates that the age of San Onofre is probably intermediate between the Ibero—Manchega, I discordance of La Gineta and the Rincon 2—3 level. Indeed, the paleomagnetic sequence of San Onofre shows a short normal episode at the beginning, a long reversal episode in the central part of the series and a new probably normal episode at the end. This palaeomagnetic sequence probably corresponds to the Mammoth reversal subzone (3.07—3.17, MCDOUGALL, 1979; 3.05—3.15, MANKINEN and DALRYMPLE, 1979) in good agreement with the inferred age of *G. crassaformis* zone and the palaeoclimatic sequence.

Given this status of uncertainty for the age of La Juliana deposit, we think that BANDET et al.'s criticism of the 3.8 Ma age of the Vialette deposit is totally groundless. In our opinion, the age of Vialette is equal to or older than 3.3 Ma.

To define the age of the basis of Lower Villafranchian there are other deposits for which there are chronological data: two in Italy (Triversa and Poggio Mirteto) and one in Spain (Las Higueruelas).

For Triversa (Fornace RDB near Villafranca d'Asti) paleomagnetic measurements are reported by LINDSAY et al. (1980), who argue that the fossiliferous deposit was formed between the Kaena and Mammoth reversal subzones of the normal Gauss zone, that is between 3.17 and 3 Ma.

We have the following data for Poggio Mirteto (Rome) (Fig. 2):

*1* the lignitiferous clays of vertebrate deposit are heteropical with Middle Pliocene marine clays *(Globorotalia crassaformis* zone);

*2* in the marine clays there is a volcanic ash level dated by *K/Ar* and fission tracks methods: the most consistent age is $3.32 \pm 0.30$ Ma (ARIAS et al., 1981);

*3* Paleomagnetic results on marine clays only show reversed magnetization, i.e. the series was sedimented during a reversal (ARIAS et al., 1976).

The combination of the above results shows that the Poggio Mirteto deposit is older than 3.0 Ma: it may have been deposited during the last part of the Gilbert reversal zone (i.e. older than 3.41 Ma) or during the Mammoth reversal subzone (3.17—3.0 Ma.)

In Las Higueruelas deposit, of Lower Villafranchian age (as shown by the presence of *Hipparion rocinantis*), *K/Ar* ages were obtained for the bed ($4.85 \pm 0.45$ Ma, for volcanic bombs ($3.82 \pm 0.34$ Ma; $3.52 \pm 0.45$ Ma) and for lava flows coeval to the deposit (BONADONNA and VILLA, 1984, Table 1); the volcanic bombs were indeed

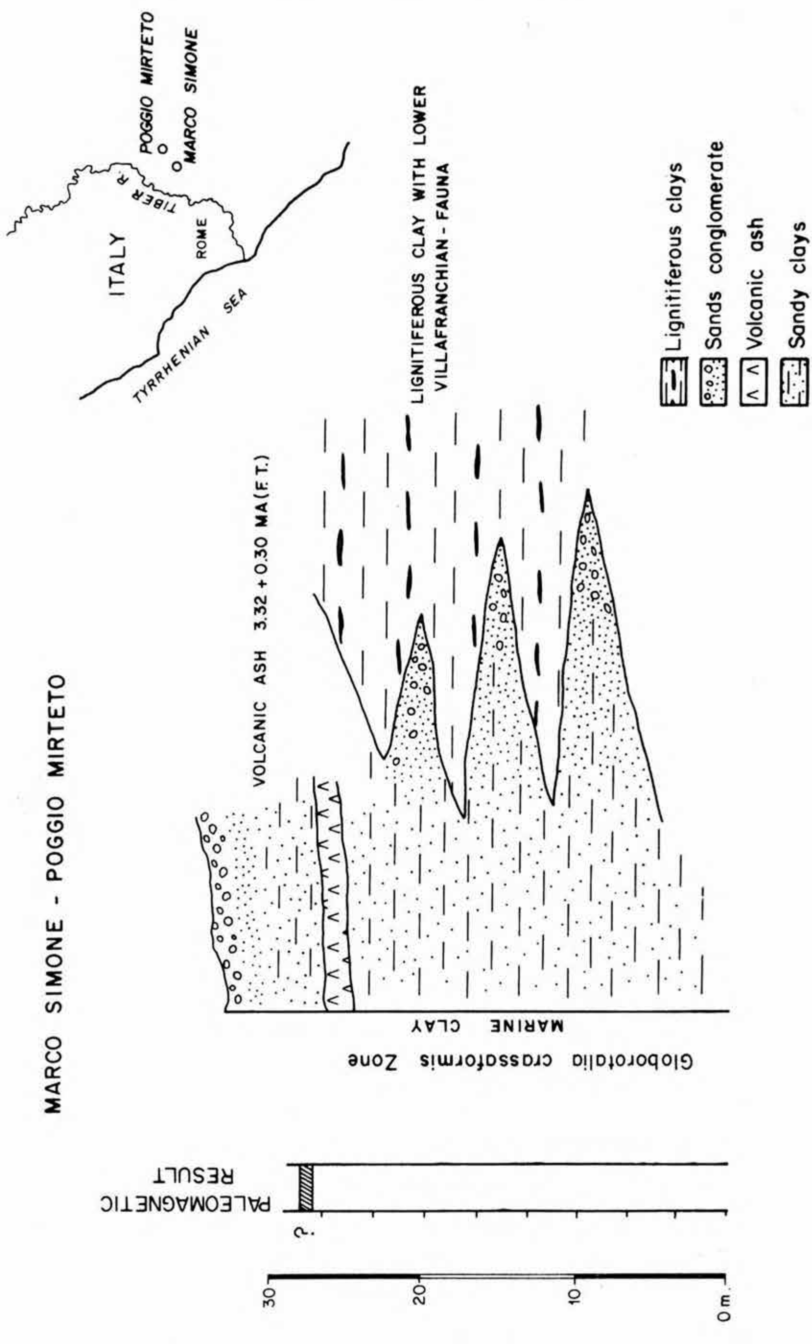
MARCO SIMONE - POGGIO MIRTETO
POGGIO MIRTETO
MARCO SIMONE
ITALY
TIBER R.
ROME
TYRRHENIAN SEA
VOLCANIC ASH 3.32 + 0.30 MA (F.T.)
LIGNITIFEROUS CLAY WITH LOWER VILLAFRANCHIAN - FAUNA
Lignitiferous clays
Sands conglomerate
Volcanic ash
Sandy clays
MARINE CLAY
Globorotalia crassaformis Zone
PALEOMAGNETIC RESULT
?
30
20
10
0 m.

*Fig. 2.*

found mixed to the fossils. These ages show that the age of Las Higueruelas deposit is surely older than 3 Ma; for the age of the bombs we may infer an age of 3.5 Ma.

The Las Higueruelas fauna is typified by the presence of *Hipparion rocinantis;* in Europe, this is the last form of this genus and it is followed by the appearance of *Equus*. In Italy, on the other hand, *Hipparion rocinantis* was never found; the latest italian *Hipparion* (probably *crassum*) was found in the Casino deposit (Tuscany) of pre-Villafranchian age. Almost the same happens in France: the reis a *Hipparion* (probably *rocinantis*?) in the Roca Neyra deposit but it is found together with *Equus stenonis* teeth (EISEMANN and BRUNET, 1973) and the geological setting of Roca Neyra deposit (it may be constituted by two different levels) is not so clear. Our opinion is that the two forms (*Hipparion* and *Equus*) cannot coexist for ecological reasons; so we think that either Roca Neyra faunistic assemblage results from mixing together different stratigraphical levels or that in Roca Neyra part of the present fauna is reworked.

As we have seen, for the age of the beginning of Villafranchian we have at least three dated deposits, and all the data are in good agreement, indicating an age older than 3 Ma. On the contrary the young age (2.6—2.4 Ma) of LY et al. (1982) for the beginning of Lower Villafranchian not only disagrees directly with the age of the other Lower Villafranchian deposists but also indirectly with the age of some dated deposits with Middle Villafranchian fauna such as Montopoli in Italy and Rincon 1 in Spain.

Considering all this evidence we believe that the age of 2.6—2.4 for the Lower Villafranchian is implausible. That young age is based only on one deposit, geologically ill-defined and only indirectly dated. We reaffirm that the beginning of Villafranchian should be older than 3 Ma (probably nearer to 3.5 Ma).

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# THE VALLESIAN IN THE TYPE AREA (VALLES-PENEDES, BARCELONA, SPAIN)

by

S. MOYÀ-SOLÀ and J. AGUSTÍ

*Introduction.* The Vallès-Penedès is an intramontainous basin situated to the northeast of the Iberian Peninsula. It is associated to the system of tectonic depressions which range from the North of Germany to Spain, across the southeast of France (Rhône basin, etc.), taking part of an incipient process of rifting developed during the Upper Oligocene and the Lower Miocene. This basin has a length close to 100 km and its widthranges between 7 and 10 km. Its orientation is ENE–WSW and it is limited by two important faults of different displacement: the Southern fault were less active than the Northern one and its activity ended during the Lower Miocene. On the contrary, the Northern fault is still in action. The Vallès-Penedès basin is filled by sediments which range from the Lower Miocene (MN3 zone) to the Lower Pliocene (MN14 zone). It has provided one of the most important assemblage of Mammal localities from western Europe (specially of *Hipparion*-faunas).

In this depression, it is possible to differentiate overall between five depositional units (AGUSTÍ et al., 1984):

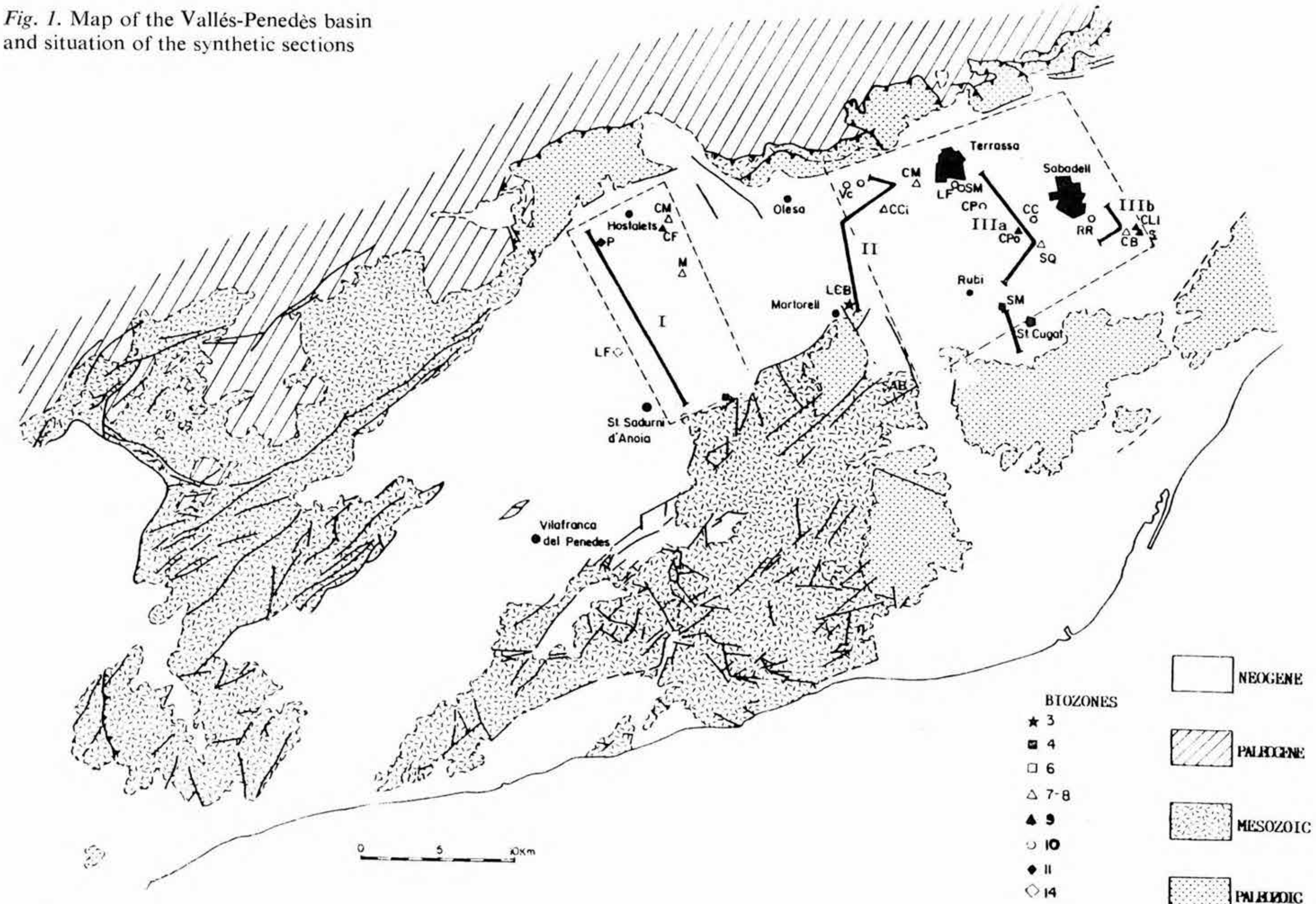
- A Basal Conglomerate Unit of indeterminate age (probably Aquitanian).
- A Lower Continental Complex (Orleanian, Lower Miocene).
- A Marine and Transitional Complex (Upper Burdigalian–Langhian).
- An Upper Continental Complex (Mid to Upper Miocene, Astaracian to Turolian).
- A Pliocene Continental Unit.

The sections in the Upper Continental Unit are the ones best represented in the basin outcropping in the more central and northern sectors (Fig. 1). The distribution of facies in this unit indicates that these are composed of alluvial fan deposits controlled by the northern fractures. This complex has yielded a high number of fossil localities ranging from the Astaracian (MN6 and specially MN7/8) to the Lower Turolian (MN11): in other words, the zones preceeding and following the Vallesian.

#### The Vallesian: outcrops and localities

The Vallesian is a Mammal stage established by CRUSAFONT (1950) to designate the faunas characterized by the first entry of the genus *Hipparion* into Europe. This author realized that most of the *Hipparion* localities in the Vallès-Penedès basin have a composition different of that from the classical localities of Pikermi and Samos in Greece. In his original definition CRUSAFONT defined the Vallesian as a transitional stage with a significant grouping of "Vindobonian" (now Astaracian) elements associated with *Hipparion* and few "Pikermian" (=Turolian) elements.

Fig. 1. Map of the Vallés-Penedès basin and situation of the synthetic sections



As regards outcropping conditions, continuity of the series and frequency of Vallesian localities, significant differencies exist between the two large sectors (Vallès and Penedès) into which the basin has been divided.

In the eastern sector, all the known localities of the Vallesian and, in general, of the Upper Continental complex have been found at essentially lutitic levels, developed in marginal zones and outer areas of alluvial fans. The relative position of these localities is well-established (Fig. 4, see AGUSTÍ et al., op. cit.). This sector has been divided into two areas. In the western one, only the Upper Vallesian outcrops (localities of Viladecavalls). There has been no record so far of the Lower Vallesian. The eastern area of the western sector is where the type-locality of the Vallesian, Can Llobateres, is located (CRUSAFONT, 1965). The poor quality of the outcrops in this area is due to the well developed quaternary overlay and the markedly fractured nature of the zone, which makes this sector a compartmentalized area of raised and sunken blocks. In spite of this, local partial successions have been established (although the stratigraphic columns are not totally continuous). The biostratigraphic record of the Vallesian in this sector is very complete, due to the frequency of the localities and the richness of their faunistic lists: Can Ponsic, Santiga, Can Llobateres, Polinya (MN9); Can Casablancas, Riu Ripoll (MN10a); Sta. Margarida, Can Perellada, Trinxera Autopista, Torrent Febulines (MN10b).

#### **Biozonation of the Vallesian in the Vallès—Penedès** (Fig. 2)

MEIN (1975) divided the Vallesian into two biozones, MN9 and MN10, which were in turn subdivided by AGUSTÍ (1981) and AGUSTÍ et al., (1984). Thus, the four phases established for the Vallès-Penedès are as follows:

*1) Hostalets de Pierola (Hipparion beds) phase (Fahlbuschia crusafonti partial range zone).*

This is a phase detected solely in the eastern sector (Fig. 4). The levels of this phase are in continuous transitional contact with the beds of the Upper Astaracian (MN8). The most important locality in this phase is Can Flaquer (Fig. 4). Biostratigraphically the onset of the Hostalets phase is marked by the appearance of *Hipparion* and the replacing of *Euprox furcatus* by *E. dicraniocerus*. In the rest, there are not changes in relation to the levels of the Upper Astaracian. The biotope of this phase was considered to be open and dry (AGUSTÍ *et al.*, 1984).

*2) Can Llobateres phase (Cricetulodon range zone).*

This phase is very well represented in the western sector and the type-locality is located here. Contact with the underlying deposits, which are well documented, is by fault (Fig. 4): St. Quirze/Can Ponsic, Castell de Barbera/Santiga, etc. This second phase of the Lower Vallesian is marked by the disappearance of the cricetids *Cricetodon, Fahlbuschia, Megacricetodon ibericus* and the rinocerotid species *Aceratherium simorrense*. Among the Rodents, *Cricetulodon* and the first murid, *Progonomys*, appear (the latter, at the top of the biozone). Quantitatively speaking, it represents a more important change than that which took place between the MN8 and the MN9. The biotope has been considered to be wet and wooded (AGUSTÍ et al., op. cit.).

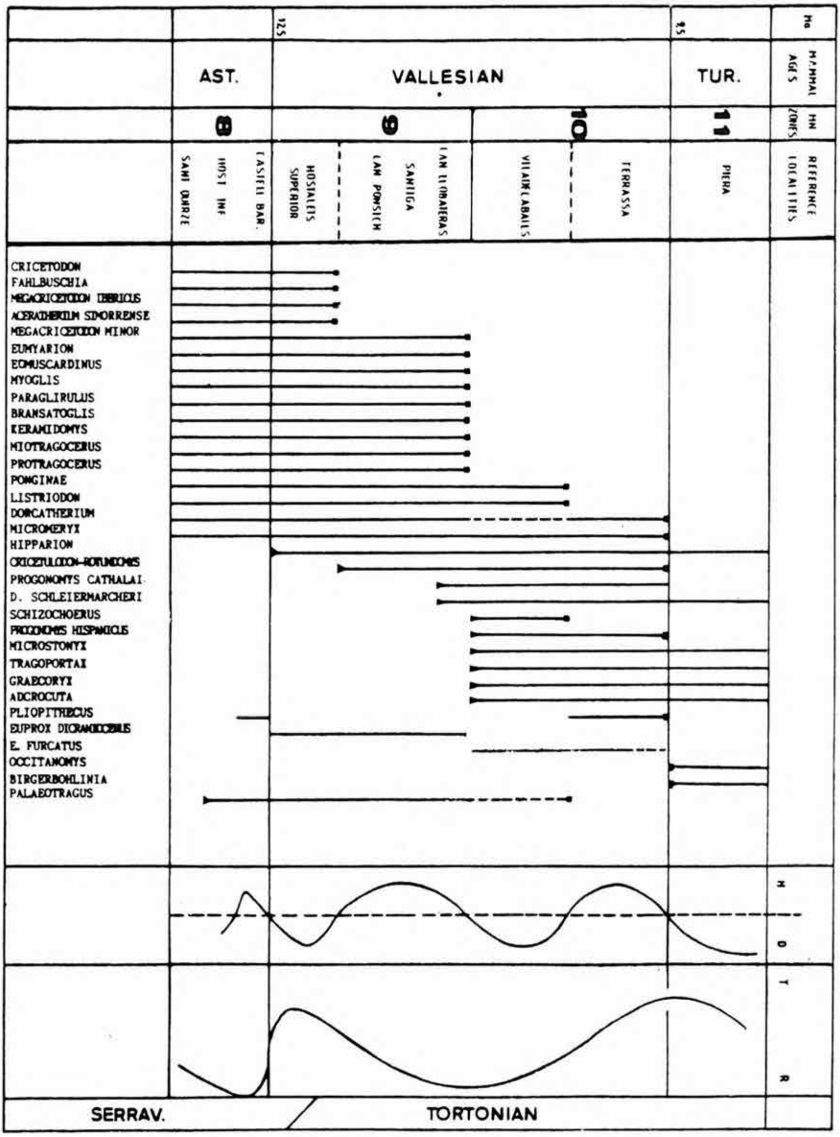



*Fig. 2.* Biozonation of the Vallesian in the type-area

*3) Viladecavalls phase (Schizochoerus range zone).*

This is well represented in the eastern sector (Viladecavalls, Riu Ripoll, Can Casablancas). The onset of this phase coincides with the boundary between the zones MN9 and MN10. This phase illustrates the most important faunistic rupture in the whole Vallesian, including its boundaries with the Astaracian and the Turolian (MN8/MN9 and MN10/MN11, respectively). It is characterized by the extinction of most of the species of Glirids (*Muscardinus* and *Myomimus* excepted), the species of

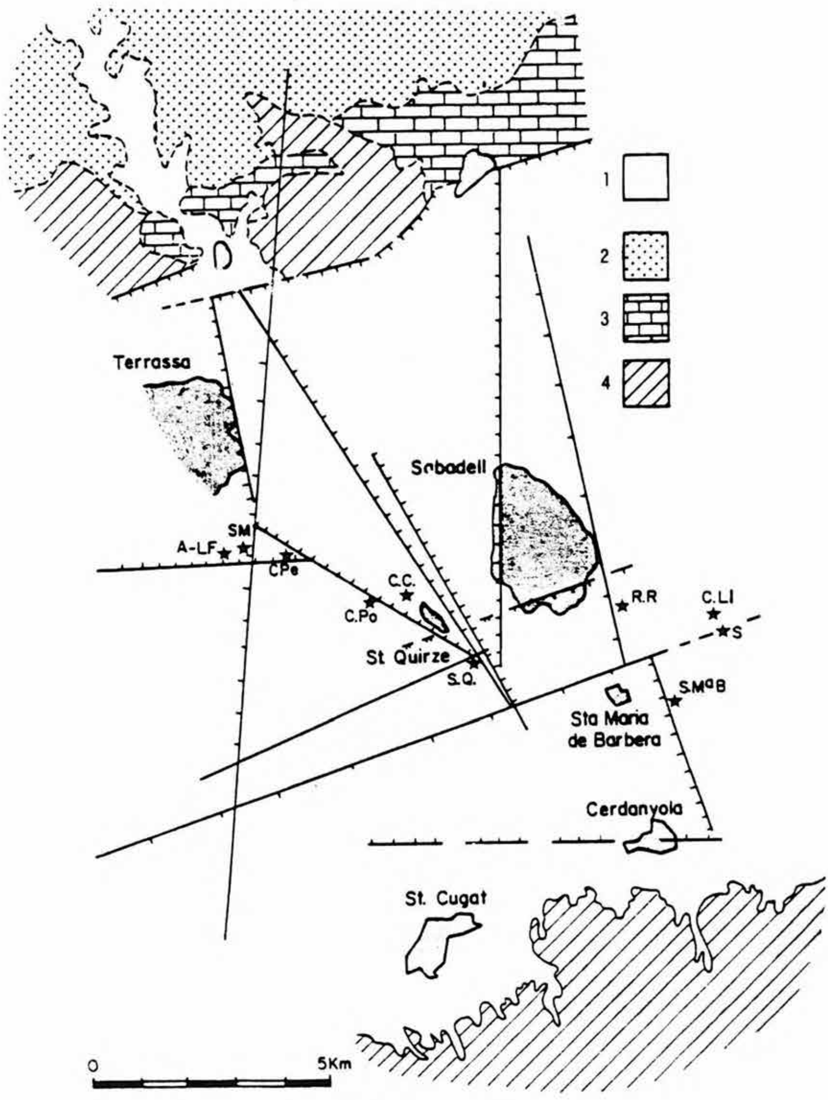



*Fig. 3.* Localities of the east Sector
*1* Neogene and Quaternary, *2* Paleogene of De Erbo basin, *3* Triassic, *4* Palaeozoic

Eomyids, two forms of Cricetids (*Megacricetodon* and *Eumyarion*), two species of Bovids and the Cervid *Euprox dicraniocerus*. It is marked by the appearance of a new species of Murid *(Progonomys hispanicus)*, the eastern Suids *Schizochoerus* and *Microstonyx*, the Bovids *Tragoportax* and *Graecoryx* and the hyaenid *Adcrocuta*. The predominant biotope in this phase has been considered to be dry and open.

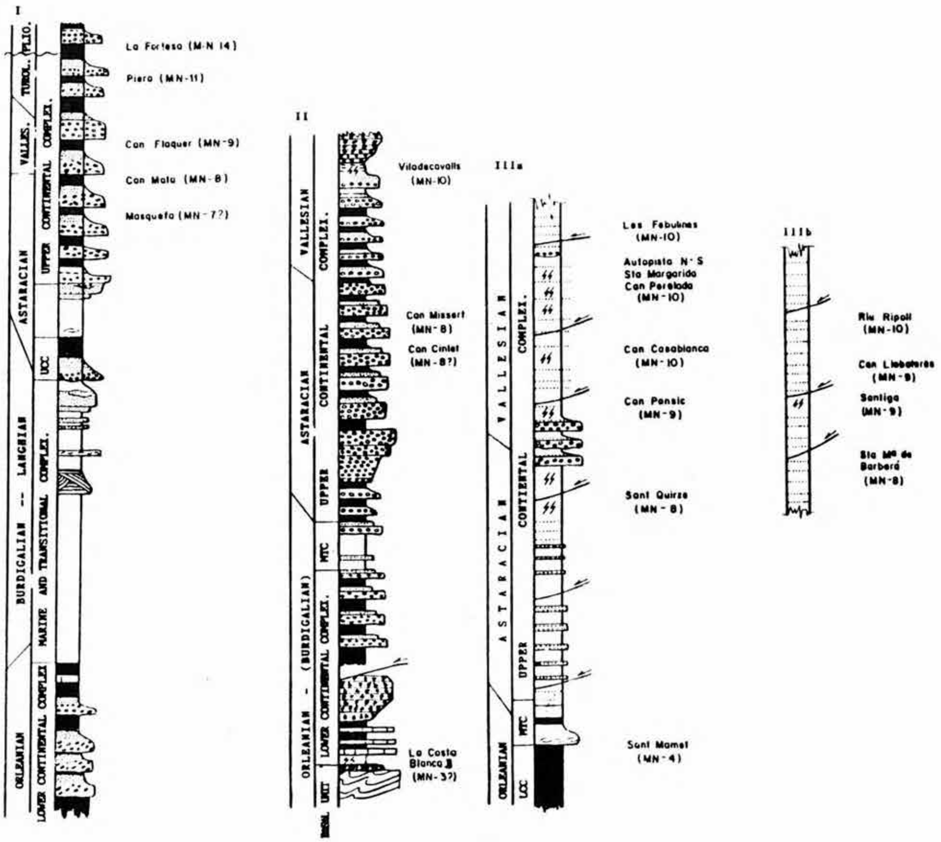



*Fig. 4.* Situation of the Mammal localities in the sections

*4) Terrassa phase (Rotundomys bressanus range zone).*

This phase is present in the eastern sector in various localities situated around the city of Terrassa (Fig. 1). It is characterized by the definitive disapparance of the Pongidae, *Listriodon* and the Astaracian—Vallesian *Palaeotragus* group. *Pliopithecus* reappear as well as a small sized form of *Euprox*. The biotope is considered to have been wet and wooded.

No complete sequence exists for localizing the transitional sedimentary passing from this phase to the Lower Turolian (MN11), which, however is represented in the Penedès sector by the locality of Piera.

*Conclusions.* AGUSTÍ et al. (op. cit.) proposed a model based on alternating dry and open episodes with others which were wet and wooded, for the Mammal associations of the Miocene of eastern Spain. During the Upper Astaracian and the Lower Vallesian this alternating of phases does not involve important faunistic ruptures. Generally speaking, the taxa existing in the previous wet phase recover in the following wet phase. However, this general pattern is broken on the boundary between the zones MN9 and MN10. In this limit we observe the final disappearance of *Protragocerus*, *Miotragocerus*, *Listriodon*, *Hyotherium*, *Parachleuastochoerus*, *Euprox* and, among the Rodents, the Eomyidae and most of the Cricetid and Glirid species. The immigratory movement on this boundary is, in addition, the most important in the whole Vallesian: *Tragoportax*, *Graecoryx*, *Adcrocuta*, *Paramachairodus*, *Schizochoerus*, *Microstonyx* and *Progonomys hispanicus* appear. It is also worthwhile nothing that these immigrants are, on the one hand, elements from drier biotopes than the preceeding ones in the previous wet phase (zone with *Cricetulodon*). On the other hand, they are typically Turolian elements (only *Schizochoerus* is not quoted for the Turolian in the Iberian Peninsula). The other group of elements forming the associations of the MN 10 are descendants of Astaracian elements (e.g., *Pliopithecus*, *Dorcatherium* or *Micromeryx*) or Vallesian ones *(Hipparion)*, the latter three forms lasting into the Lower Turolian of the Iberian Peninsula (Moyà-Solà, 1983).

Therefore, to sum up, the Lower Vallesian is a fauna essentially Astaracian in character although with *Hipparion*, whereas the Upper Vallesian fauna is largely Turolian in character with wet relicts from the Upper Astaracian—Lower Vallesian.

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# THE NEOGENE GEOLOGICAL EVENTS IN THE NORTH PACIFIC AND MEDITERRANEAN REGIONS

by

YU. B. GLADENKOV

During the last years our knowledge of the North Pacific Neogene has considerably advanced. Results of the investigations were mostly included in the Proceedings of the IGCP Project 114 (1984). Besides, by the 27th IGC held in Moscow, 1984 the soviet geologists had published a fundamental monograph generalizing their work on one of the Neogene key sections in the USSR Far East—west Kamchatka.

This area belongs mainly to the boreal belt (Fig. 1). There are wide-spread thick sedimentary and volcanic deposits of the young geosynclinal systems from the transitional zone between ocean and continent.

The study of the sedimentary deposits with organic remains from the northern Pacific Belt (Japan, Sakhalin, Kamchatka, Chukotka, Alaska, States of Washington, Oregon, California) and the adjacent oceanic regions revealed some synchronous geological events in the Neogene history: successive changes of faunistic and floristic assemblages, climatic fluctuations, change in the sedimentary regime, tectonic movements, etc.

A comparative study of these events and the geological phenomena recorded in Europe, particularly in the Mediterranean, makes their correlation possible.

Thus, in the North Pacific Neogene well pronounced were several climatic fluctuations (the optima of the end of the Early beginning of the Middle Miocene, end of the Middle Miocene, beginning of the Pliocene) (Fig. 2). They resulted in appreciable changes and migration of the planktonic and benthic assemblages and were reflected in changes of floristic associations. These fluctuations proved to be good markers to correlate the Neogene deposits. They have provided to a considerable extent the Neogene correlation both of northeastern Asia and North Asia and North America.

As indicated in the literature, some of these fluctuations were recorded in the North Atlantic and Mediterranean regions (warming in Burdigalian—Langian, cooling in the Late Miocene and Eopleistocene) (Fig. 3). This sallow us to reveal the common trends in the climatic changes during the Neogene all over the Northern Hemisphere.

At the beginning of the Middle Miocene in the most of the Pacific are the regressive regime turned to a transgressive one. Simultaneously abundant diatomites appeared.

In the North Pacific in the Late Miocene a notable regression took place, and intense tectonic movements produced, in fact, the modern structural plan (including the volcanic belts and the adjacent trough in eastern Kamchatka). These events coincided in time with the Messinian events in the Mediterranean. The Pliocene transgressions were also synchronous enough in many regions concerned.

An important geological event took place in the Pliocene. In the Middle Pliocene the Bering Strait was opened that caused new ways of migration of the marine fauna (via the Arctic region) and, in particular, appearance of considerable numbers of the Pacific migrants in North Europe (the Serripes zone in Iceland, Red Crags in England, etc).

During the Late Pliocene and Eopleistocene cooling the boreal assemblages migrated to the south of these areas. This explains the appearance of the assemblage with *Astarte borealis* in Japan and that with *Arctica islandica* in the Mediterranean.

It is worth of notice that in the Pleistocene (in Early Brühnes Epoch) the tectonic dislocations of both the Neogene and the Eopleistocene deposits took place. They were followed by formation of high terraces on the shores of open seas.

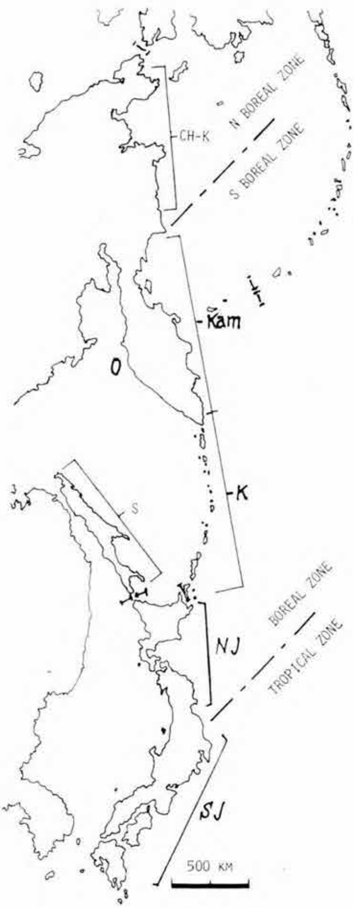


*Fig. 1.* Late Paleogene—Neogene biostratigraphic provinces in the marginal part of northeast Asia (The USSR Far East and East Asian provinces of the North Pacific boreal province)

*Biostratigraphic belts (zones):* boreal, tropical, subzones: h—b = high boreal, l—b = low boreal; *provinces:* Ch—K = Chukotka, Kam = Kamchatka, K = Kuril, S = Sakhalin, NJ = northern Japan, SJ = southern Japan; dotted lines = the USSR state boundary

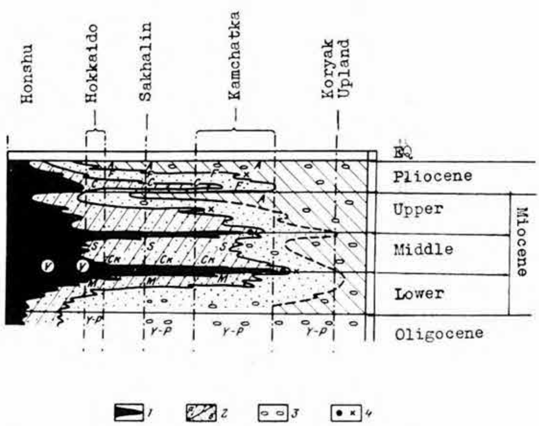


*Fig. 2.* Distribution of assemblages of the Neogene molluscs within the Northwestern Pacific area

*Assemblages: 1* tropical and subtropical; *2* boreal (*a* with the southern elements; *b* with the northern elements); *3* pebbles of marine ice rafting; *4* thermophilic assemblages of spores, pollen and diatoms. *Assemblages:* A = Astarte, F = Fortipecten takahashii, C = Chlamys cosibensis, heteroglipta, S = Securella ensifera chehalisensis, Ck = Chlamys kaneharai, V = Vicaria—Vicariella, M = Mytilus tichanovitchi, P = Papyridea harrimani, Y = Yoldia longissima

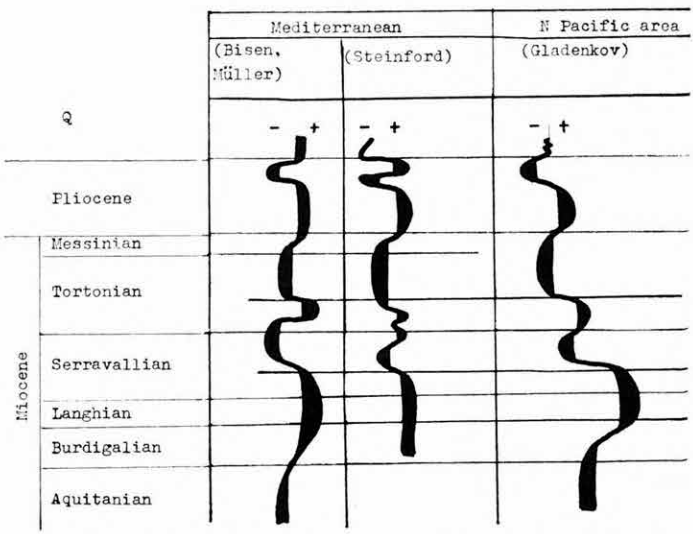


*Fig. 3.* Neogene climatic fluctuations (— relative coolings; + relative warmings)

A more detailed analysis can facilitate the correlation of the Neogene stratigraphical schemes of the North Atlantic and North Pacific regions, as well as the deciphering of global geological events.

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# NEOGENE—QUATERNARY BOUNDARY IN THE CONTINENTAL SEDIMENTS OF THE GAUDIX-BAZA BASIN (SOUTHEASTERN SPAIN)

by

J. AGUSTÍ, J. GIBERT, S. MOYÀ-SOLÀ, J. A. VERA

*Introduction.* The Guadix-Baza depression is an intramountainous basin situated in the Betic mountain range to the northeast of the province of Granada. The work under way in this basin permits the establishment of a continuous sequence from the Upper Miocene (Upper Turolian, MN13) to the Middle Pleistocene. The materials in this depression were studied by VERA (1970), who distinguished (disconformably over a basal deformed unit of Miocene age) three large formations with lateral changes:

A) *Guadix Formation.* This is formed of detritic materials belonging to a braided river system and fine meandering river deposits. They cover almost all the Western section and the margins of the Eastern section. The presence of *Hipparion gromovae granatensis* near Abla (CUEVAS et al., 1984) indicates an Upper Miocene age (Upper Turolian, MN13) for the bottom of the formation.

B) *Gorafe Huélago Formation.* This is formed of chemical deposits (micritic limestone, marls, dolomites and gypses) with lignitic intercalations, which are a lateral extension towards the centre of the basin of the fluvial deposits of the Guadix formation.

C) *Baza Formation.* This is richest formation in number of localities, especially within the Orce-Galera-Huéscar sector. In the centre of the formation, the dominant limestones and lutites change to gypses. Three members can be differentiated in this formation. The lowermost, mainly composed of limestone with intercalations of lignitic clays, has suffered some deformation.

The Middle member (Red detrital member) lies unconformably on the above mentioned one. Its materials were deposited in an alluvial plain with a slightly stabilized fluvial system. Finally, the levels of the Upper member lie conformably on the above mentioned one. They are composed of locally evaporitic, yelowish—white lacustrine beds deposited in the context of a very shallow lake which received some detrital contributions from time to time from the adjoining fluvial systems.

## *Trilophomys* cf. *castroi* zone

In the Guadix-Baza basin, the Lower Pliocene is characterized by micromammal assemblages where the dominant elements are the Muridae. In the Gorafe-Huélago formation the most characteristic locality belonging to this zone is Gorafe—4, which presents the following association: *Ruscinomys* sp., *Cricetus* cf. *barrierei*, *Trilophomys* cf. *castroi*, *Protatera* sp., *Stephanomys* cf. *margaritae*, *Paraethomys meini*, *Castillomys crusafonti*, *Apodemus dominans*, *Eliomys intermedius*, *Prolagus michauxi*, Leporidae indet., Suinae indet., *Gazella borbonica.*

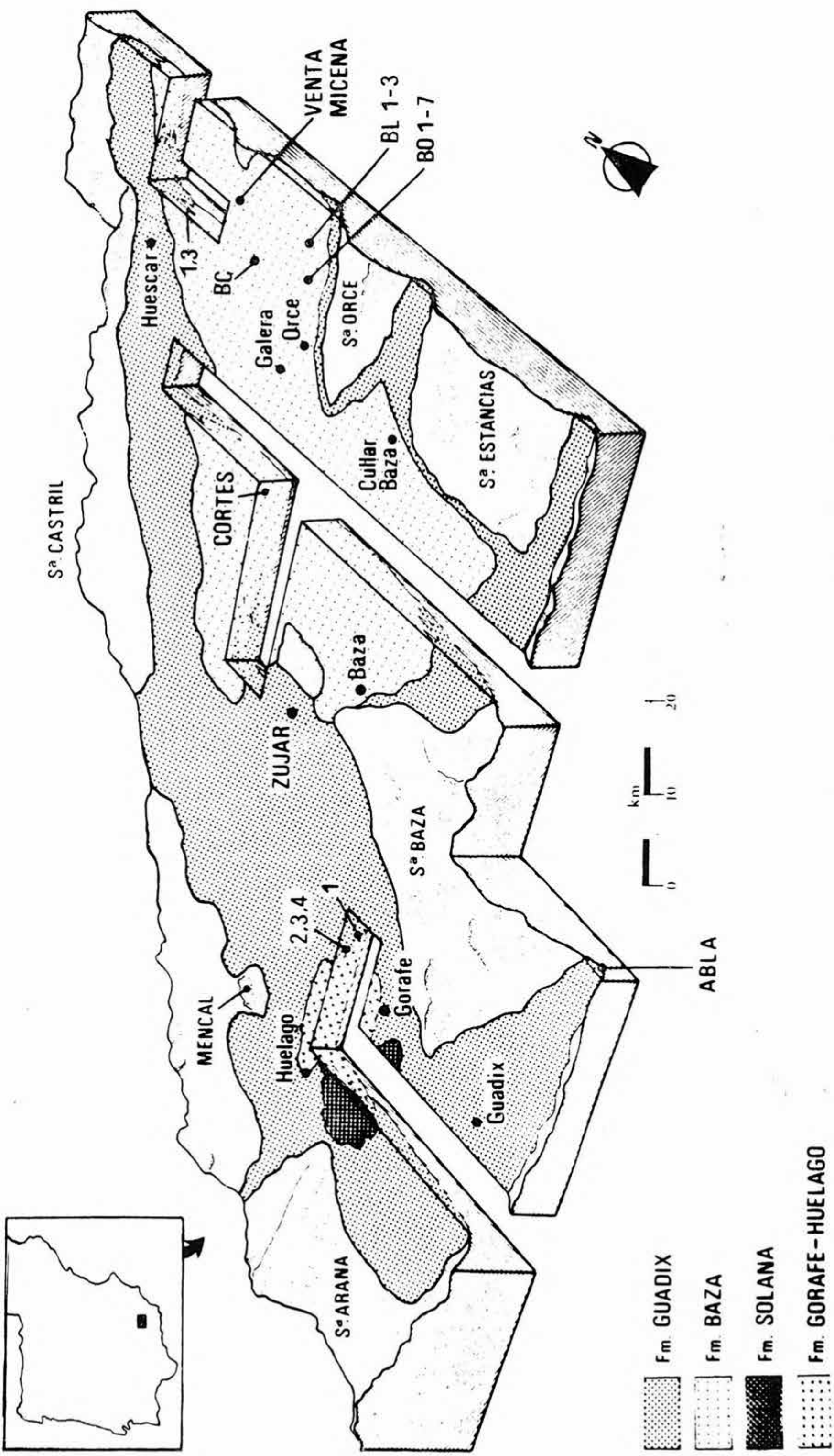


*Fig. 1.* Formations and paleontological sites of the Guadix-Baza basin (SE of Spain)

Some notable features of this fauna are:

— As stayed above, the Murids are the dominant element of the association, both from a quantitative and qualitative point of view.

— Among the Cricetids, two species (*Ruscinomys* sp. and *Cricetus* cf. *barrierei*) are of clear European origin, while *Trilophomys* cf. *castroi* is probably of Asiatic origin.

— A member of the Gerbillidae family is present in the association (*Protatera sp.*).

— No truly arvicolid species has been found in Gorafe—4.

*Protatera* sp. is present in other localities in the Guadix-Baza basin (Gorafe—1 in the Gorafe-Huélago formation; Botardo-C in the Baza formation). In Spain, this form appears at the top of the Turolian (MN13) and persists until the base of the Ruscinian. *Protatera* sp. is probably of African origin but an eastern one cannot be excluded.

Besides Gorafe—1, —3 and —4, in the Gorafe-Huélago formation, the localities of the Botardo section (Botardo B, C, 2 and 3, in the Baza formation) also belong to the *Trilophomys* cf. *castroi* zone. This zone may be correlated with the MN14.

*Mimomys occitanus* zone

The main difference between the levels of Gorafe—1, 3—4 and Botardo those belonging to this biozone is the appearance in the basin of the first true arvicolid, *Mimomys occitanus*. In the Gorafe-Huélago formation, this species is documented in the localities of Gorafe—2 and Gorafe—5. The latter has yielded the following microfauna:

*Ruscinomys* sp., "*Cricetus*" cf. *angustidens*, *Mimomys occitanus*, *Apodemus dominans*, *Occitanomys* sp., *Stephanomys* cf. *thaleri*, *Castillomys crusafonti crusafonti*, *Paraethomys meini*, *Eliomys intermedius*.

The absence of *Protatera* and *Trilophomys* should be noted. Nevertheless, the remainder of the microfauna strongly resembles that of the previous biozone (e.g. Gorafe—4).

In the Baza formation, a number of localities are also referable to this zone: Galera—1, Cañada del Castaño—1, Huéscar—1. This the microfauna from Huéscar—1 shows the following composition: *Mimomys occitanus*, *Apodemus dominans*, *Castillomys crusafonti crusafonti*, *Stephanomys* cf. *thaleri*. The level of Cañada del Castaño—1 has yielded somewhat different association with *Apodemus dominans*, *Castillomys crusafonti crusafonti*, *Paraethomys meini*, *Stephanomys* cf. *thaleri* and *Muscardinus pliocaenicus*. The presence of the latter taxon indicates wetter conditions than the previous levels. Curiously, all the localities in the Baza formation lack the cricetid "*Cricetus*" *angustidens*, an ever present species in the Gorafe section. Whether this absence is significant from a biostratigraphic point of view or is due to environmental factore is something to dilucidate in further fieldwork.

*Mimomys cappetai* zone

This zone is characterized by the presence the arvicolid *Mimomys (Kislangia) cappetai*. A number of taxa persist from the former zone: *Apodemus dominans*, *Castillomys crusafonti crusafonti*, *Stephanomys* cf. *thaleri* and *Eliomys intermedius*

**Biozones and Mammal associations in the Guadix-Baza basin**

*Table 1*

| Mammal stages | Biozones | Localities | Faunal associations |
|---|---|---|---|
| Upper Biharian | A. cantiana | Cúllar de Baza I, II | Arvicola cantiana, Microtus brecciensis, Allocricetus bursee duranciensis, Apodemus aff. sylvaticus, Eliomys quercinus |
| | P. arvalidens zone | Cúllar de Baza--A, -B, -C | Pitymys cf. arvalidens, Castillomys crusafonti, Eliomys quercinus, Oryctolagus cf. lacosti, Insectivora indet., Carnivora indet. |
| Middle Biharian | M. savini zone | Lome Quemada-1 Húescar-2,-3 Puerto Lobo-1, -4 | Mimomys savini, Allophaiomys cf. nutiensis, Apodemus aff. sylvaticus, Equus stenonis, Archidiskodon meridionalis, Dicerorchinus etruscus, Hippopotamus incognitus, Cervidae indet. |
| Lower Biharian | Allophaiomys pliocaenicus zone | Cañada de Murcia-1, -2 Venta Micena-1, -2 Fuentenueva-2 Orce-6, -7 Barranco Leon-1, -2, -3 Fuentenueva-C Orce-4, -5, P Cañada de Balmaez? Camino Yeseras? | Allophaiomys pliocaenicus, Castillomys crusafonti, Apodemus aff. sylvaticus, A. aff. mystacinus, Eliomys intermedius, Hystrix primigenia, Oryctolagus cf. lacosti, Desmana sp., Soricidae indet., Canis etruscus, Pachycrocuta brevirostris, Homotherium crenatidens, Ursus aff. etruscus, Archidiskodon meridionalis, Equus stenosis, Dicerorhinus etruscus, Hippopotamus incognitus, Megacerini indet., Cervidae indet., "Cervus" elaphoides, Praeovibos n. sp., Hemitragus n. sp., Ovis sp., Bison sp., Bovidae indet. |
| | Mimomys ostramosensis zone | Fuentecilla-5 Cementerio de Orce-B Cortijo D. Alfonso Cortijo D. Diego Barranco los Conejos Orce-1, -2 Orce-D | Mimomys ostramosensis, M. cf. pusillus, Apodemus aff. sylvaticus, Castillomys crusafonti, Eliomys aff. quercinus, Equus stenonis, Leptobos etruscus, Dicerorhinus etruscus, Archidiskodon meridionalis |
| Upper Villanyian | Mimomys cf. neidi zone MN 17 | Orce C Fuentenueva-1 Alqueria Fuentecilla-2? Cortes de Baza-3? | Mimomys cf. reidi, Apodemus aff. dominans, Castillomys crusafonti, Eliomys intermedius, Equus stenonis vireti, Gazella borbonica, Cervidae indet., Carnivora indet., Leporidae indet. |
| Lower Villanyian | M. cappetai zone MN 16 | Cañada de Murcia-3 Galera-2 Zujar | Kislangia cappetai, Stephanomys cf. tahleri, Apodemus aff. dominans, Castillomys crusafonti, Eliomys intermedius |

| Mammal stages | Biozones | Localities | Faunal associations |
|---|---|---|---|
| Upper Ruscinian | Occitanus zone MN 15 | Huéscar-1<br>Cañada del Castaño<br>Galera-1<br>Baza?<br>Gorafe-2, -3, -5 | Mimomys cf. occitanus, Paraethomys meini, Stephanomys cf. thaleri, Occitanomys brailloni, Castillomys crusafonti, Apodemus aff. dominans, Muscardinus pliocaenicus, Anancus arvernensis<br><br>Cricetus cf. angustidens, Trilophomys cf. castroi, Mimomys cf. occitanus, Paraethomys meini, Stephanomys cf. margaritae, Castillomys crusafonti, Apodemus aff. dominans, Eliomys intermedius, Prolagus michauxi, Leporidae indet., Soricidae indet. |
| Lower Ruscinian | Trilophomys castroi zone MN 14 | Gorafe-1, -4 | Ruscinomys sp., Cricetus cf. barrierei, Protatera sp., Trilophomys cf. castroi, Paraethomys meini, Stephanomys cf. medius, Occitanomys sp., Castillomys crusafonti, Apodemus cf. dominans, Eliomys intermedius, Atalantoxerus adroveri, Prolagus michauxi, Desmana sp., Microstonix sp., Trischizolagus maritsae, Gazella borbonica, Suinae indet. |

(Zújar, Galera—2, Cañada de Murcia—3, Cañada del Castaño—2). Most of the cricetid and murid species disappear: *Ruscinomys* sp., *Cricetus* cf. *angustidens*, *Occitanomys* cf. *brailloni*, *Paraethomys meini*. Thus, in contrast with the *M. occitanus* zone, the arvicolids (in this case, *M. cappetai*) become quantitatively dominant in the associations.

*Mimomys* cf. *reidi* zone

This zone marks an important change with respect to the previous one since, among the Rodents, only *Apodemus dominans*, *Castillomys crusafonti* and *Eliomys intermedius* persist. The *Mimomys (Kislangia)* and *Stephanomys* species disappear in younger levels. This is a surprising event, since both lineages persisted in other sites in the Iberian Peninsula and their remains are found in karst deposits.

The most characteristic element of the biozone is a small to medium sized hypsodont *Mimomys* with cement which maintains its mimomyan structures ("*Mimomys*"-ridge, enamel ring, etc). Nevertheless, this species (here provisionally called *M.* cf. *reidi*) is larger than the other European representatives of *Mimomys reidi*.

Besides *M.* cf. *reidi*, the locality of Fuentenueva—1 is illustrative of the associated fauna: *Apodemus dominans*, *Castillomys crusafonti*, Leporidae indet., Carnivora indet. Cervidae indet., *Gazella borbonica* and *Equus stenonis vireti*. Other sites in the Baza formation belonging to this zone are Alquería and Orce-C.

The *Mimomys* cf. *reidi* zone can be partially correlated with the MN17 zone (Upper Villanyian). All the localities in this part of the Baza formation are situated at the top of the deformed Lower Calcareous Member. The beds of the overlaying member (Red Detrital Member) were deposited unconformably over above mentioned. Nevertheless, the supposed gap seems not to be of great importance since the fossiliferous levels deposited immediately over this Red Detrital Member also belong to the Upper Villanyian.

### *Mimomys pliocaenicus ostramosensis* zone

All the fossiliferous beds after the *Mimomys* cf. *reidi* zone are situated in the Upper Calcareous Member. The youngest levels before those with *Allophaiomys pliocaenicus* are characterized by the presence of up to three different lineages of *Mimomys*.

In the oldest locality of the Upper Calcareous Member, Orce-D, a form close to *M. pliocaenicus* is associated with a small; sized *Mimomys* lacking the mimomyian structures (*Mimomys* cf. *pusillus*). In younger localities (Orce—1, Cementerio Orce A and B, Cortijo D. Diego, Cortijo D. Alfonso, Fuentecilla—5), the first species increases in size and hypsodonty (giving *M. pliocaenicus ostramosensis*) but the rest of the assotiation remains the same.

In the level of Orce—2, a new immigrant appears among the Rodents. Its generic attribution is doubtful, since this form lacks roots and mimomyian structures and has abundant cement. Nevertheless, its enanel is undifferentiated or of a Mimomys-kind. Its size is larger than any *Allophaiomys* species and it seems not to be related to *A. deucalion*. Besides the three arvicolid species, (*Mimomys p. ostramosensis*, *Mimomys* cf. *pusillus*, *Allophaiomys* sp.), the remainder of the Mammal fauna from Orce—2. is typical of the Upper Villanyian: *Apodemus dominans*, *Castillomys crusafonti* ssp., *Eliomys quercinus*, *Leptobos etruscus* and *Equus stenonis*. In the level of Barranco de los Conejos there still appear *Mimomys ostramosensis*, *Mimomys* cf. *pusillus* (as in Orce-D and Orce—2, but without roots) and *Allophaiomys* sp., but they were already associated with the cold—quaternary immigrant *Praeovibos* sp., In the upper levels of the Orce Section, the preceeding arvicolid species are substitued by *Mimomys* cf. *savini* and *Allophaiomys pliocaenicus*.

### *Allophaiomys pliocaenicus* zone

This is the best representationed biozone in the Guadix-Baza basin (Orce 3—7, Barranco León—1—3, Fuentenueva C and—2, Venta Micena—1 and —2, Cañada de Murcia—1, etc). *Allophaiomys pliocaenicus* appears in some levels associated with a *Mimomys* species close to *M. savini* but it usually constitutes the only arvicolid present. The most representative locality in this zone Venta Micena—2, which presents the following association (Moyà-Solà et al., 1981): *Desmana* sp., *Allophaiomys pliocaenicus*, *Castillomys crusafonti* ssp., *Apodemus* aff. *sylvaticus*, *Hystrix* sp., *Eliomys intermedius*, *Lepus* sp., *Ursus etruscus*, *Canis etruscus*, *Pachycrocuta brevirrostris*, *Panthera* sp., cf. *Panthera gomgaszogensis*, *Homotherium* cf. *latidens*, *Megantthereon* sp., *Vulpes praeglacialis*, *Lynx* cf. *spelaea*, Mustelidae indet., *Archidiskodon meridionalis*, *Equus stenonis*, *Dicerorhinus etruscus*, *Hippopotamus incognitus*, *Megaceros* aff. *verticornis*, *Cervus* (?) *elaphoides*, *Praeovibos* sp., *Capra* sp., Caprini indet. *Bison* sp., *Soergelia* sp., *Testudo* indet., *Lacerta* indet., *Amphibia* indet.

The *Allophaiomys* speciments from Venta Micena—2 show the typical morphotypes for *A. pliocaenicus*. Nevertheless, there is a certain tendency in some of them to develop an incipient BSA—4 in the lower M1. The remaing Rodent fauna (*Hystrix* excepted) is composed of elements from the previous biozones.

*Mimomys savini* zone

The most representative locality in this zone is Loma Quemada—1 with a microfauna which includes *Mimomys savini*, *Allophaiomys nutiensis*, *Apodemus* aff. *sylvaticus* and *Castillomys crusafonti* ssp. In the localities of Puerto Lobo—1 and Huéscar—2, *Mimomys savini* is associated with *Equus stenonis*, *Dicerorhinus etruscus* (large form), *Archidiskodon meridionalis*, *Hippopotamus incognitus*, Cervidae indet.

The main differences with this biozone in the *A. pliocaenicus* relation are the following:

— *Mimomys*, represented by *M. savini*, reappear, becoming the dominant arvicolid in the association.

— *Allophaiomys* is represented by two different species: *A. nutiensis* and *A.* aff. *burgondiae*. *A.* aff. *burgondiae* is the descendant in place of *A. pliocaenicus* from the previous biozone.

*Pitymys* cf. *arvalidens* zone

This zone is characterized by the appearance of the first arvicolae with *Pitymy* morphotypes (*Pitymys* cf. *arvalidens* in Cúllar de Baza—C). In the levels situated immediately below this locality (Cúllar de Baza—B) we still found *Castillomys crusafonti* (the last record in the basin).

Over these beds we found the site of Cúllar de Baza—I (RUIZ and MICHAUX, 1976) with a typical cromerian fauna including *Arvicola cantiana*, *Microtus brecciensis*, *Allocricetus bursae duranciensis*, *Apodemus* aff. *sylvaticus*, *Eliomys cuernicus*, etc.

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# MAGNETOBIOSTRATIGRAPHY OF THE MIDDLE—UPPER NEOGENE AND PLEISTOCENE DEPOSITS OF ROMANIA

by

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This paper attempts a review of some of rather significant magnetostratigraphic results acquired on the Sarmatian—Pleistocene deposits of Romania (ANDREESCU, 1981; 1982; 1983; ANDREESCU et al., 1981; ANDREESCU et al. in press; GHENEA et al., 1981; ALEXEEVA et al., 1983). In additicn, several pertinent observations regarding the present status of some Neogene biochronologic and chronostratigraphic units will be done.

In our palaeomagnetic scale the Sarmatian stage begins in lowermost Chron 14 ($\pm$14.0 Ma). The base of the Sarmatian stage seems to be not younger than 13.8—14.0 Ma, since the radiometric dating of the Upper Rhyolite Tuff (situated somewhere between the "Cerithium limestone formation" and "Ervilia claymarl formation", i.e. between Anomalinoides badenensis zone and Elphidium reginum zone) in Hungary revealed an average age of 14.0 Ma (HÁMOR et al. 1979; BALOGH, JÁMBOR, 1985 etc).

The Early Bessarabian falls within Chron 13 (Fig. 2), and an age of about 12.5 Ma may be assumed for the Volhynian—Bessarabian boundary. This age estimate is in a good agreement with Ganzei's (1984, vide AGADJANIAN, 1985) datings (12.5$\pm$ 0.57 Ma) of the Early—Middle Sarmatian boundary.

The palaeomagnetic polarity record in sections No 1—3 (Fig. 2) pointed out that the Khersonian—Bessarabian boundary could be situated below the reversed event of Chron 11, and an age of about 11.0 Ma may be assigned to this boundary, which is again in coincidence with Ganzei's datings (11.8$\pm$0.98 Ma).

The Meotian deposits encompass two chrons: the lower chron is prevailingly reversed, the upper one being revealed only by normal polarities. At least three quite different means of palaeomagnetic calibration and age assignation of the Meotian stage have so far been developed:

*a)* would correspond to Chrons 6 and 5 (6.8—5.4 Ma), including $MN_{12}$ (Early Meotian) and early $MN_{13}$ (Late Meotian) (CHEPALYGA et al., 1985);

*b)* in PEVZNER and VANGENGEIM's opinion (1984; 1985) it is correlated with Chrons 8 and 7 (9.0—6.9 Ma) and the Meotian mammals belong to the $MN_{11}$ zone;

*c)* the Meotian stage begins in early Chron 10 (ca 10.5—10.6 Ma) and ends in late Chron 9 (ca 9.0 Ma) (ANDREESCU, 1981; 1985 and this volume). In this interpretation the Meotian mammals are assigned to late $MN_{10}$ and early $MN_{11}$.

Since in dependence on the position of the Meotian in the Neogene chronostratigraphic scale the age assignation of the Sarmatian, Pannonian and Pontian stages is closely related and since the Meotian seems to represent a real turntable in attempts at correlation with both the western Paratethys and Mediterranean regions, some additional remarks need to be made:

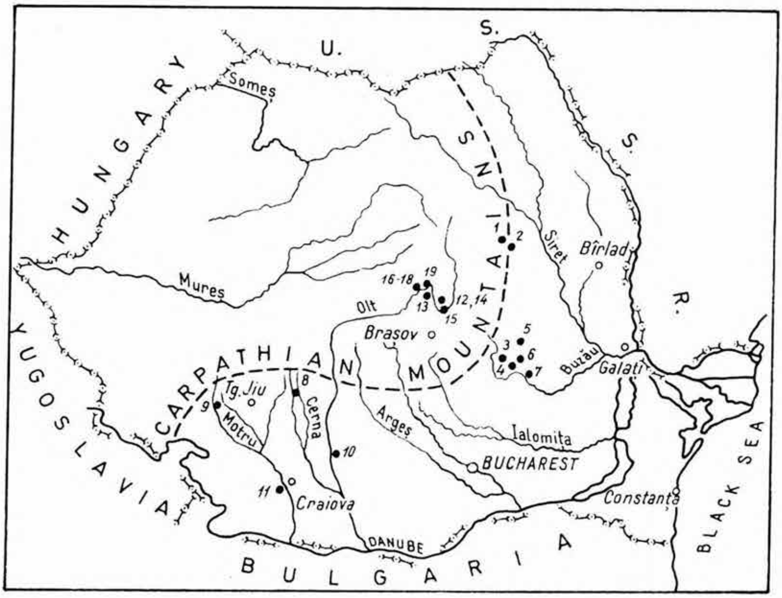


*Fig. 1.* Localities mentioned in the Table 1

*a)* nowhere in the Soviet Union have the mammal faunas of TARACLIA, EMETOVKA, NOVOELIZABETOVKA, TUDORA and other equivalent sites ever been found related to any undisputable Meotian molluscs;

*b)* unlike PEVZNER and VANGENGEIM, we consider that in the western Paratethys there is no gap corresponding to the Pannonian s. str. The weight of evidence indicates that the base of the Meotian, i.e. $NSM_{5a}$ subzone, is to be found in the Pannonian C zone and that the Upper Pontian corresponds roughly to the Upper Pannonian E zone (ANDREESCU, 1981). On the other hand, the Upper Pannonian deposits (zones D—E) bear Late Vallesian mammals (late $MN_{10}$), whereas the Early Pontian begins in the western Paratethys within the $MN_{11}$ zone (PAPP, 1981; RÖGL, STEININGER, 1983; RABEDER, 1985 etc). Since the base of the Pontian is unanimously considered synchronous in both the western and eastern Paratethys, one may assume that $MN_{11}$ straddles the Pannonian—Pontian and the Meotian—Pontian boundary respectively (see ANDREESCU, Fig. 1—3, this volume pp. 345—347). In this respect the mammal sites of Győrszentmárton, Rózsaszentmárton, Hatvan etc, closely associated with Upper Portaferrian molluscs (BARTHA, 1971), should rather be with $MN_{12}$ rather than with $MN_{11}$;

*c) radiometric data:* 10—10.6 Ma, Early Meotian with volcanics in continental deposits with *Hipparion garedzicum* in eastern Georgia (GABUNIA, RUBINSTEIN, 1977), $\pm$9.8 Ma, andesites interbedded mollusc-bearing Pannonian deposits (D—E zones) (EDELSTEIN et al., 1977; new dating, EDELSTEIN personal communication); about 9.5 Ma, Early $MN_{11}$ of Maragheh (CAMPBELL et al., 1980); $\pm$9.4—7.9 Ma, Kayadibi, Early Turolian (SICKENBERG et al., 1975).

*d) palaeomagnetic data:* Pannonian s. str.—Pontian boundary at top of Chron 9 (Anomaly 5), estimated at about 8.8 Ma (HÁMOR et al., 1985); Kastellios; Early

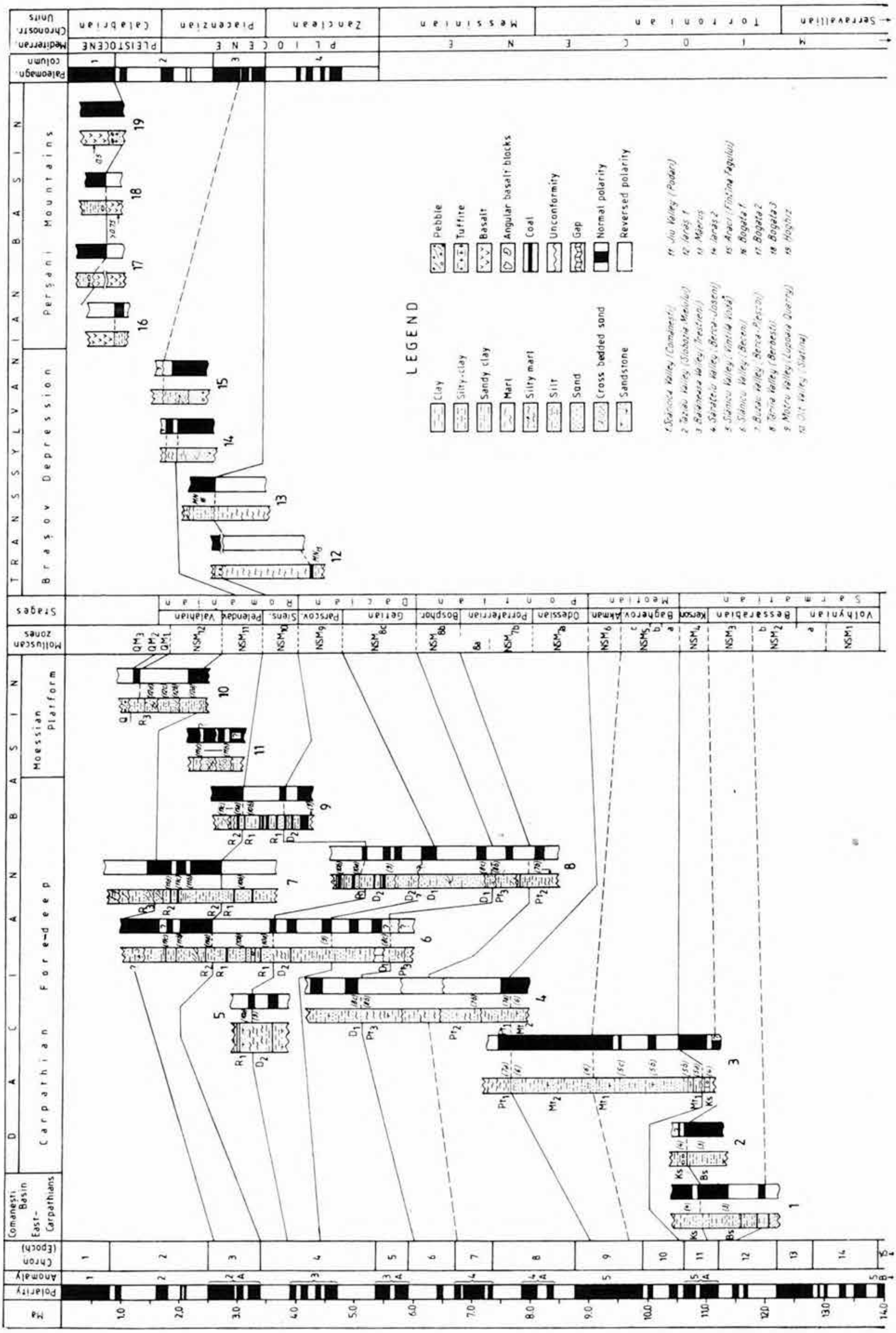


*Fig. 2.* Upper Neogene magnetostratigraphy

$MN_{10}$ related to early $N_{16}$ in a sequence with reversed polarities (MEIN, 1985; Roundtable, Budapest);

*e)* in Uppermost Meotian deposits SEMENENKO and LIULEVA (1982) pointed out a nannoplankton association of $NN_{10}$, while the Lower Meotian deposits revealed $NN_9$ *(Discoaster hammatus, Catinaster coalithus* etc*)*. In the Mediterranean area, early $N_{16}$ seems to be, more or less, linked with late $NN_9$ or, less probably, with early $NN_{10}$. Therefore if it Kastellios early $MN_{10}$ is in direct association with early $N_{16}$, then that interval should belong to Chron 10 (10.6— 9.9 Ma in our palaeomagnetic scale) because the base of the Turolian, i.e. that of $MN_{11}$, can not be younger than 9.0 Ma. Accordingly, the Early Meotian, with $NN_9$, should be assigned to the Chron 10, and the Early Meotian mammal fauna (GREBENIKI, REGHIU etc) pertains to $MN_{10}$ (Late Vallesian), where as TARACLIA, CIMISLIA, NOVOELIZABETOVKA etc, should be assigned to the Upper Meotian and or Lower Pontian (see: ANDREESCU, Fig. 3, this volume, p. 347).

*Correlations:* The Meotian stage correlates with the Pannonian s. str. (zones: C, D, E) and with the Middle and Upper Tortonian stage.

The Pontian stage spans, in our interpretation, uppermost Chron 9 and Chrons 8—6 (9.0—5.9 Ma). The Soviet authors pretend the Pontian deposits to be characterized only by reversed polarities, and they assign this stage either to Chron 6 (SEMENENKO, PEVZNER, 1979; PEVZNER, VANGENGEIM, 1982; 1984; 1985), or to the lowermost Gilbert (CHEPALYGA et al., 1985).

The lithologic and faunistic record suggests that during the Pontian especially in the Late Portaferian—Bosphorian interval (ANDREESCU, 1981) many non-depositional or erosional phases existed. Hence the frequency of sedimentary hiatuses in the Pontian sequences (Fig. 1, columns 4 and 8; see also Fig. 4, columns 3 and 4 in ANDREESCU, this volume, p. 348). Because of these gaps, in some instances, the palaeomagnetic record could exhibit, in spite of close-spaced sampling, only reversed polarities. However in section No 8 (Fig. 2) two sequences of normal polarity seem to occur (preliminary data). In section No 4, in addition to the absence of Early Portaferrian deposits due to a normal fault, several unconformities have been pointed out. In addition, the sampling density wass too poor: 18 samples for a pile of 650 m.

Therefore the palaeomagnetic record of the Pontian of the eastern Paratethys is far from being exhaustive. Given these constraints, our interpretation is based partly on the palaeomagnetic data provided by HÁMOR et al. (1985) from the Pannonian basin, and on the arguments put forward in discussion on the Meotian.

*Correlations:* Crevillente $_{4\text{-}5}$=$MN_{12}$=Messinian (G. conomiozea). Since Pikermi=Garkin (ca 8.6 Ma)=$Samos_{1\text{-}4}$ (8.5—8.0 Ma)=$Maragheh_2$ (8.5—8.0 Ma)= =late $MN_{11}$ and/or early $MN_{12}$, as suggested by various authors, this means that the base of the Messinian is situated at a level close to the Early Pontian. Consequently, a minimum age of 8.5—8.0 Ma is to be admitted for the base of the Messinian.

The base of the Dacian has been shown to be situated near the Chrons 6—5 boundary (5.8—5.9), which is in line with the Kimmerian—Pontian boundary (SEMENENKO, PEVZNER, 1979 etc).

The Romanian—Dacian boundary is practically coincident with the Cochiti event of the Upper Gilbert (3.9—4.0 Ma); the Early Romanian—Middle Romanian boundary correlates with the Gilbert—Gauss boundary, while the Middle—Upper Romanian boundary is located in the Upper Gauss (ca 2.7—2.6 Ma) ANDREESCU, 1981; 1982; ANDREESCU et al., 1981; ALEXEEVA et al., 1983).

The Romanian deposits contain in the Dacian basin, a very rich and diversified freshwater Mollusca fauna assigned to the $NSM_{10}$, $NSM_{11}$ and $NSM_{12}$ zones (Table 1, columns 5—11) (ANDREESCU, 1979 etc). Sometimes the Romanian deposits yielded mammal faunas of the $MN_{15}$, $MN_{16}$ and $MN_{17}$ zones (ANDREESCU, 1981; 1982; 1983; FERU et al., 1983; SAMSON, RĂDULESCU, 1985). In the Dacian basin, the best known of all the Romanian mammal sites are those of Berești, Mălușteni (Early Romanian = $MN_{15}$), Tulucesti, Cernătești, Covrig Podari (Middle Romanian = $MN_{16}$), Milcovu din Vale, $Slatina_{1-2}$, Cherleștii—Moșteni, Frătești etc (Upper Romanian = $MN_{17}$).

In the Transylvanian basin (Brașov depression) the so-called "coal-bearing complex" yielded a rich mammal fauna at Căpeni and Vîrghiș sites, which SAMSON and RĂDULESCU (1973, 1985) and GHENEA et al. (1981) correlate with Berești and Mălușteni ($MN_{15}$). The "coal-bearing complex" underlies the "marly complex" (Table 1, columns 12; 13) which, in turn, is overlain by the Iarăș formation ("sandy complex") with endemic molluscs and Early Villafranchian mammals ($MN_{16}$ at Iarăș (= middle—upper part of Iarăș—New Quarry, column 14) and Araci Fîntîna Fagului (column 15) sites. The "marly-complex" gave reversed polarities, whereas the Iarăș formation is characterized by normal polarities in the lower part of section $Iarăș_2$ (with Mammals $MN_{16}$), followed in the upper part of the section by a short reversed interval and then by a normal one. GHENEA et al. (1981) correlated the reversed polarities of the "marly complex" with the uppermost Gilbert and the normal ones of the Iarăș formation with the Lower Gauss. Consequently, an age of 4.0—3.6 Ma is inferred for the Căpeni and Vîrghîs Mammal sites and one of 3.4—3.2 Ma for $Iarăș_2$ (New Quarry) and Araci-Fîntîna Fagului may be accepted.

SAMSON and RĂDULESCU (1973, 1985) consider the Early Villafranchian sites of Brașov depression to be older than Cernătești and Tulucești sites of the Dacian basin. In agreement with this opinion, we think that Cernătești and Tulucești could be assigned to $MN_{16b}$, having an age of 3.2—2.9 Ma. This assertion is based on the occurrence of Mollusca (found together with mammal remains) at Cernătești which is an equivalent of the $NSM_{11b-c}$ subzones. The same mollusc assemblages occur in the Podar section (lower part) (= stratotype Middle Romanian), at Covrigi site (NW Oltenia) and in the stratotype sections of the Romanian stage (columns 6 and 7). As shown by magnetosrtatigraphic results these molluscs span the early—middle Gauss interval, i.e. prior to the upper part of Kaena event.

The correlations of Romanian biochronologic units and of Romanian substages with other equivalent units from Europe are discussed in detail in our above-cited papers and in those of SAMSON and RĂDULESCU (1973, 1985), FERU et al. (1983) as well.

The Early Pleistocene deposits have been investigated magnetostratigraphically in the Slatina section (Fig. 2, column 10) (ANDREESCU et al., 1981). In this section the occurrence of $QM_1$—*Unio apscheronicus* zone (= $Slatina_3$ level) has been shown to be coincident with the Olduvai event. Since the Pliocene—Pleistocene boundary is not considered to lie above the Olduvai event (INQUA Congress, Moscow, 1982), the $QM_1$ zone must be assigned to the uppermost Romanian. The same observation is valid for $Tetoiu_1$ and $Slatina_3$ Mammal sites, formerly included in the lowermost Pleistocene (ANDREESCU et al., 1981; FERU et al., 1983).

As regards other Early Pleistocene molluscan zones ($QM_2$ = *Bogatschevia sturi;* $QM_3$ = *B. scutum;* $QM_4$ = *B. caudata*), singled out by TSHEPALYGA (1972 and in NIKIFOROVA et al., 1976), in the Soviet Union, and subsequently identified in the Dacian basin as well (ANDREESCU, 1981 and unpublished data), no section suitable for palaeomagnetic sampling has hitherto been available.

At Izvoru, SAMSON and RĂDULESCU found *B. sturi* together with mammals which ANDREESCU et al. (1981), ALEXEEVA et al. (1983), FERU et al. (1983) and RĂDULESCU and SAMSON (1985) correlate with Tetoiu$_2$, Irimești, Drăgănești-Olt, Frătești$_2$ mammal sites, corresponding to the Upper Doaashkinian—Early Boshernitzian (= Upper Odessan) in the Soviet Union, and with the Early Eburonian in NW Europe.

Sections 16—19 in Transylvania are not related to any mammals or mollusc record. The palaeomagnetic polarities of these rocks have been calibrated by relying on some radiometric dates.

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# EGGENBURGIAN, OTTNANGIAN AND KARPATIAN (EARLY MIOCENE) ALONG THE BOHEMIAN MASSIF IN MORAVIA (CZECHOSLOVAKIA)

by

P. ČTYROKÝ

*Introduction.* The period of Eggenburgian to Karpatian, an interval from 23—16.5 Ma, was the period of the gradual underthrusting of the eastern margin of the Bohemian Massif under the Eastern Alps in Austria and the Outer West Carpathians in Moravia. This subduction provoked the neoid folding of the Outer West Carpathians and their displacement towards the Bohemian Massif.

The deposits of Eggenburgian to Karpatian on the southern and eastern slopes of the Bohemian Massif are not involved in nappes. For that reason they are sometimes called the Undisturbed Molasse. In this molasse region the stratotypes of Eggenburgian, Ottnangian and Karpatian stages were chosen earlier.

## Stratigraphy and palaeogeography

### *Eggenburgian*

The globally wide-spread transgression of this age (contemporaneous with the Burdigalian of Tethys) flooded the southern and eastern margin of the Bohemian Massif in Moravia during the highest sea-level culmination in Late Eggenburgian during which the deposition of so-called Eggenburg beds took place. The West Carpathians in Moravia and Slovakia were flooded by the sea and thicker sequence were deposited in gradually opened intramontane molasse basins and in the embryional West Carpathian fore deep. The elevation ridges of W Carpathians at that time made up a complicated archipelago amidst the Eggenburgian sea. It seems most probable, according to the deep-sea character of the deposists and their faunas, that the initial fore deep during the Eggenburgian period was located on the Paleogene—Ždánice flysch unit in Moravia and its prolongation to the south, into the Waschberg Zone in Lower Austria. The Eggenburgian strata of this region are called the Šakvice marls with rich planktonic foraminifers and calcareous nannoplankton of zones NN 2 and NN 3 (Stránik and Molčiková, 1980).

On the southern rim of the Bohemian Massif shallow marine biofacies with large Pectinids (genera Gigantopecten, Pecten, Macrochlamys, Chlamys) and other giant tropical molluscs is limited to the stratotype area in Lower Austria (Steininger and Seneš et al., 1971). In Moravia such biofacies is known from the north Moravia only where it was recorded from the Jaklovec in Ostrava (Ganss, 1936; Čtyroký 1958). We suppose that such rich Eggenburgian faunas invaded to this district directly from the East throught the West Carpathian archipelago in Slovakia.

Along the margin of the Bohemian Massif in South Moravia (along the Czechoslovak—Austrian boundary in approximate line Langau—Znojmo—Mikulov) the equivalents of the Eggenburg beds are littoral marine deposits with littoral molluscs

**Stratigraphy of the Eggenburgian to Karpatian in the Carpathian foredeep in Moravia**

*Table 1*

| | SOUTHERN and CENTRAL PART | | | NORTHERN PART | | |
|---|---|---|---|---|---|---|
| | Lithology | Fossils | Max. thick. | Lithology | Fossils | Max. thick. |
| Karpatian | calcareous silty clay and sand laminated calcareous silty clay (Schlier) | U. parkeri breviformis, G. tarchanensis, G. concinna, G. cipero-ensis ottnangensis | 1000 m | variegated beds | | 300 m |
| | | | | "grey beds" calcareous silty clay | G. concinna, G. opinata, U. parkeri brevif. | |
| | sand and calcareous sandy clay | molluscs of the stratotype in Slup | | "brown beds" bituminous clay coal clay | Vittoclithon, Melanopsis, Congeria div. sp. | |
| | | | | conglomerate, sand | Ch. macrotis, Ch. multi-striata, Discinisca, Hyotherium | |
| | diatomaceous clay | brackish and marine diatoms | | variegated beds | Vittoclithon, Unio | |
| Ottnangian | sand, sandstone and silty clay | Rzehakia socialis, Limnopagetia div. sp., Ctyrokya, Congeria | 500 m | UNKNOWN | | |
| | sand | mammals from Ořechov | | | | |
| | fish-bearing clay and sandy clay | fish-bones, teeth, scales | | | | |
| | gravel and sand | | | | | |
| Eggenburgian | rhyolite tuffite or bentonite sand and clay | Crassostrea gryphoides Pirenella moravica Hydrobia, Nematurella | 600 m | ? | ? | |
| | calcareous silty clay and sand | Cibicidoides budayi Thracia pubescens Laevicardium cingulatum | | calcareous silty clay and sand | G. bolli lentiana, Clobocas indulina subglobosa B. fastigia droogeri | |
| | glauconitic sandstone, sand and conglomerate | Glycymeris fichteli G. cor, Protoma cathedralis | | Bryozoa limestone basalt clastics | G. gigas, Ch. jaklowe-ciana, P. hornensis | |

*Glycymeris fichteli*, *G. cor*, *Crassostrea gryphoides*, *Thracia pubescens*, *Laevicardium cingulatum*, *Protoma cathedralis*, *Calyptrea chinensis* and others. Closer to the coast the estuarine and deltaic deposits with molluscs *Crassostrea gryphoides*, various cardiids, *Polymesoda convexa percostata*, congerias, *Pirenella moravica*, Vittoclithon, Hydrobia, Nematurella, Ctyrokya and others were deposited (ČTYROKÝ, 1982). From calcareous clays of the deeper littoral of the same period were reported foraminiferal assemblage with *Cibicidoides budayi*, *Lenticulina meznericsae*, *Cribrononion dollfusi*, *Elphidium ortenburgense* and others (Molčíková, 1976).

On the southeastern slope of the Bohemian Massif the final member of the Eggenburgian is a horizon of rhyolite to rhyodacitic tuffite or bentonite. This horizon is a good marker-horizon both in shallow estuarine deposits (with thickness of about 1.5 m) and as thin layer in calcareous clays of the deeper littoral (ČTYROKÝ, 1967; 1982). The rhyodacite tuffites at the top of the Eggenburgian in S Moravia are probably synchronous with a strong eruption of so-called Lower Acid Volcanism in the hinterland of W Carpathians in S Slovakia and Hungary which started on Ottnangian/Eggenburgian boundary (about 20–22 Ma; HÁMOR et al., 1979). The rhyodacite tuffite recorded in the faciostratotype well of the Eggenburgian stage ČČ-3 at Velká Čausa (Slovakia; STEININGER and SENEŠ et al., 1971) could be probable equivalents of the same tuffite in S Moravia.

The situation is quite different beyond the front-line of West Carpathian and Waschberg nappes where the Eggenburgian strata are known as autochthonous cover below nappes or tectonic "slices" incorporated within nappes. They are known from many deep wells in Lower Austria (BRIX, KRÖLL, WESSELY, 1977) and South, Central and Northern Moravia (MORKOVSKÝ, 1962; JURKOVÁ et al., 1983).

*Ottnangian*

The full-marine deposits of this stage, similar to these in the stratotype area at Ottnang in Austria (PAPP, RÖGL, SENEŠ et. al., 1973) are missing on the eastern slopes of the Bohemian Massif in Moravia, in spite of they occurrence in the Moravian part of the Vienna Basin.

The sequence of gravels, sands, fish-bearing clays, Rzehakia-bearing sands and silts are corresponding to this stage. They are deposited partly unconformably, over the estuarine tuffitic beds of the Upper Eggenburgian (ČTYROKÝ, 1982).

According to some views whole Rzehakia beds in W Carpathians should belong to the bottom of the Karpatian stage. This view is valid for the S Slovak–N Hungarian Rzehakia beds but it is not valid for the Rzehakia beds in the Alpine–Carpathian foredeep in Austria and Moravia. The different age of Austrian and Moravian Rzehakia beds (Late Ottnangian; ČTYROKÝ, 1967; 1972) and Slovak–Hungarian ones (Early Karpatian; KANTOROVÁ et al., 1967; HORVÁTH and NAGYMAROSSY, 1978) had probable roots in the gradual transgression of the Karpatian sea, spreading from the south through the Adriatic Sea. In the Alpine–Carpathian foredeep to the end of the Ottnangian the total degradation of the brackish sea basin took place and Rzehakia assemblages died out but in Hungary and Slovakia, the Karpatian sea invaded the still existing Rzehakia sea and did not interrupt its development suddenly.

Furthermore the Rzehakia beds with typical molluscan assemblage were recently found in S Moravia (SW from Brno) in some deep wells clearly over the estuarine beds of the Upper Eggenburgian (with *Crassostrea gryphoides* and *Pirenella moravica*) and below the marls rich in index–microfossils of the Karpatian.

In the deposits of the littoral margin of the Ottnangian brackish sea (most probably of Rzehakia beds) the mammal locality Ořechov at Brno is situated. Its mammal assemblage correlates with the zone MN4b (FEJFAR, 1974; FEJFAR and SCHMIDT-KITTLER, 1984).

The deposits of the Ottnangian stage are also known from the tectonic slices within the Ždánice-Flysch-nappe in south and central Moravia, but they are totally unknown from the Carpathian foredeep in northern Moravia.

*Karpatian*

During this age the next evolution of the W Carpathian foredeep in Moravia took place, it spreaded from the slopes of the Bohemian Massif in the W to the pre-Styrian Carpathian nappes in the E. The transgression penetrated into the Carpathian foredeep from the the southeast and it got connected with the Mediterranean—Adriatic basin. From the palaeogeographical point of view the W Carpathian foredeep in Moravia was the most distant and "blind" sea basin to the north in Europe during this period.

Both in the southern and the northern part of the Carpathian foredeep the initial members of the Karpatian are represented by brackwater and estuarine deposits with a mixture of brackish and marine diatoms, foraminifers and molluscs. But the higher strata are represented by up to 1000 m thick sequence of laminated calcareous silty clays (schlier), calcareous sandy clays and sands. They contain a rich assemblage of marine foraminifers, molluscs, echinoids and ostracods (CICHA, SENEŠ, TEJKAL et al., 1967).

The tectonized deposits of the Karpatian on the Ždánice-Flysch Unit were found in the same area as Šakvice marls (Eggenburgian) and Pavlovice beds (Ottnangian) at Velké Pavlovice (STRÁNIK and MOLČÍKOVÁ, 1980). It seems probable that these Karpatian deposits represent a marine connection between the Carpathian foredeep and the Vienna Basin over this flysch unit at that time.

During last 15 years new knowledge was collected in many deep wells for gas on the eastern marginal coast of the Karpatian sea in the northern Moravia. The existence of so-called "brown beds" was proved (JURKOVÁ and NOVOTNÁ, 1974). This sequence originated in estuarine and sublittoral marine environment with plenty of plant debris and estuarine molluscs of the genera Vittoclithon, Melanopsis, Hydrobia, Cerastoderma, Congeria and others (JURKOVÁ in MENŠÍK et al., 1983). These deposits interfinger with the so-called "grey beds" (schlier) with polyhaline micro- and mollusc-an faunas. Nearly whole width of the foredeep of this stage is covered by the Late Styrian nappes of the Outer Carpathians.

## Tectonics and orogeny

As was mentioned before, the deposists of Eggenburgian, Ottnagian and Karpatian on southern and eastern rim of the Bohemian Massif are not involved by strong tectonic and orogenic movements. Quite opposite is the situation on the eastern margin of the W Carpathian foredeep where the deposits of Eggenburgian and Ottnangian are covered by or incorporated into the Early Styrian nappes of the Outer Carpathians.

Strong Late Styrian movements are well documented in S Moravia too during which the Ždánice unit nappe was shifted over the Karpatian deposits at a distance of about 6 km. These movements were much stronger in northern Moravia, where the Outer Carpathian nappes are shifted over the Karpatian strata at a minimum distance of 30 km.

Considering this fact the Late Styrian movements (after the Karpatian and before the Early Badenian) were the strongest orogenic movements on the margin of the W Carpathians in Moravia.

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# BIOZONATION OF THE NEOGENE INVERTEBRATE MEGAFAUNA OF THE HELLENIC AREA

by

M. DERMITZAKIS and E. GEORGIADES-DIKEOULIA

*Introduction.* Studying for many years the invertebrate megafauna of the Neogene of the Hellenic area, the authors have the possibility to propose a tentative biozonation for this geological period. The studied Neogene outcrops are mainly located on Crete island, Attica, Peloponnese, north Greece, the Ionian islands and the islands of Aegina, Evia, Milos, Rhodos, Carpathos, Cos, Gavdos etc. The marine megafauna groups involved include bivalves, gastropods, echinoids, brachiopods and corals.

The intention of this study is to contribute to a better knowledge of the ranges and biostratigraphic distribution of megafaunal species, relative to the Mediterranean Neogene stages as well as to establish and propose a tentative biozonation for the Neogene of the Hellenic area based primarily on the molluscs.

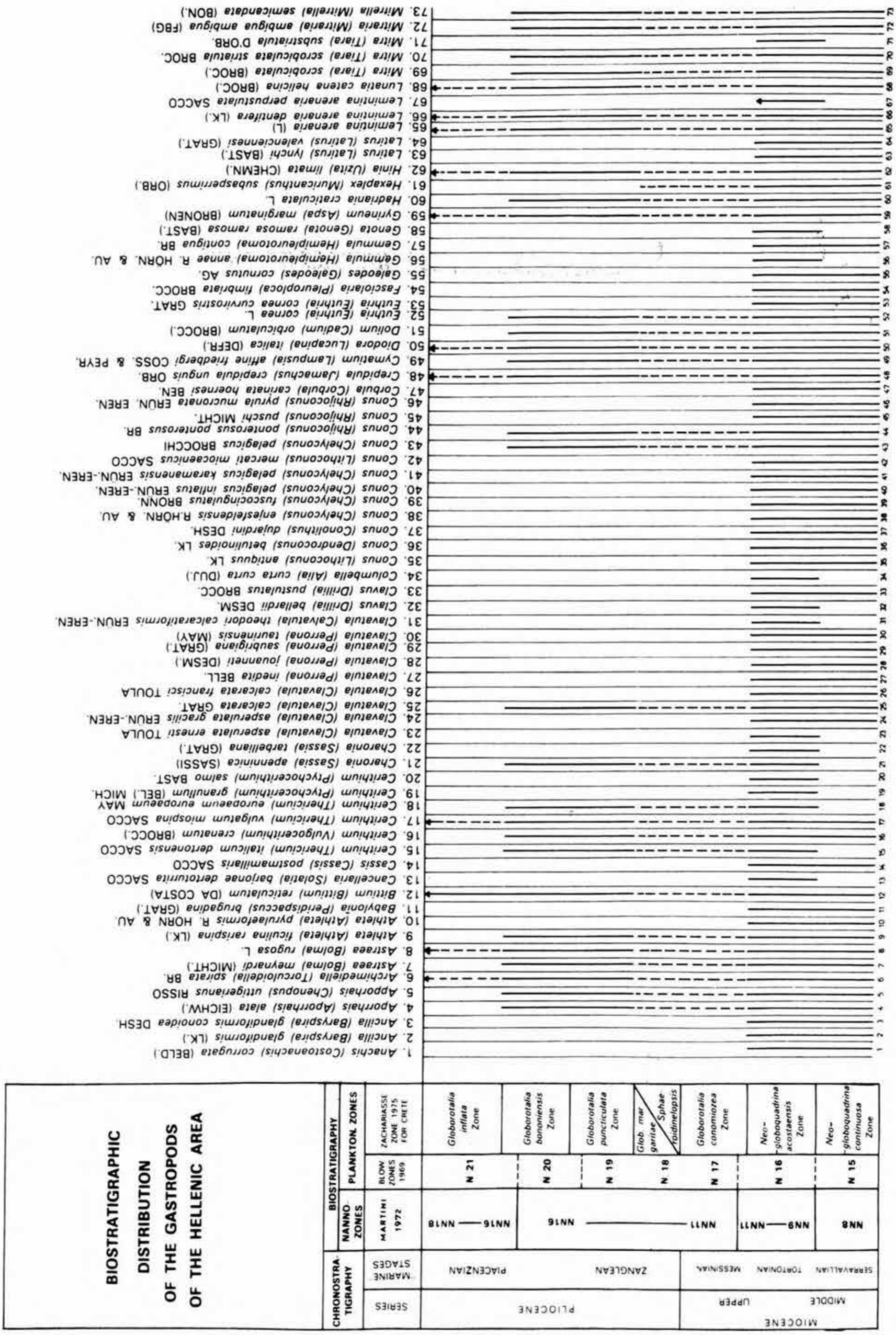
The palaeogeographic configuration and the way of formation of the basins during the Lower Miocene were in such conditions that megafauna was not very common. An exception to this fact, concerning the occurence of megafauna associations, are the domains of western Greece and the Mesohellenic trench in which faunas from the Lower Miocene are present but have not yet been well studied.

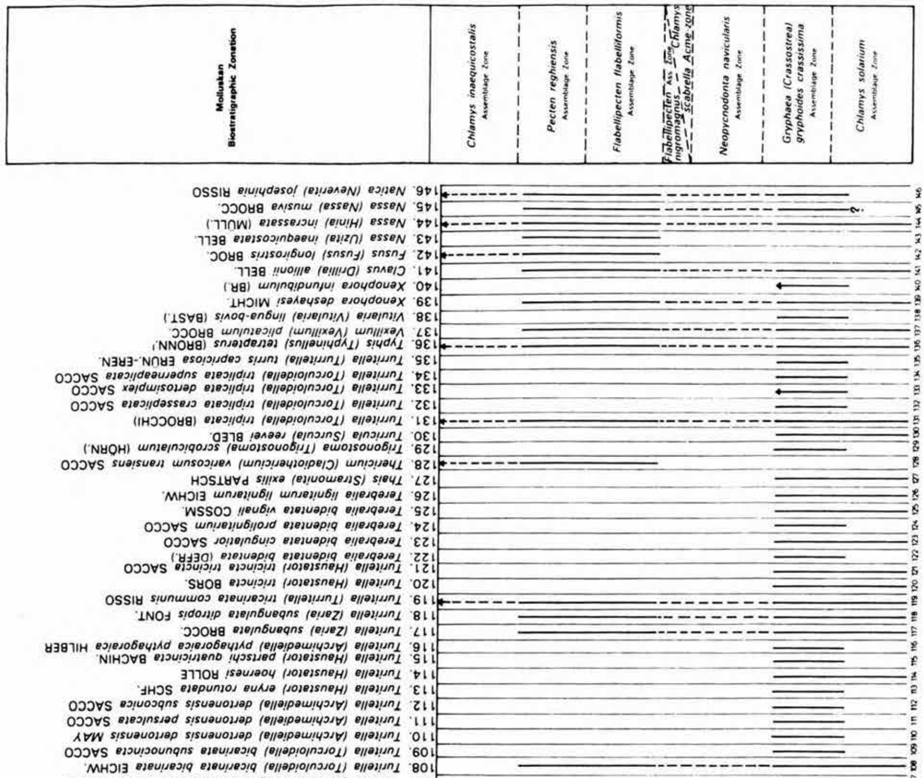
Most of the studied associations lived at depths between 10—300 meters in sandy environments or on carbonate platforms and as a rule these associations mirror the existence of fairly diverse bottom life during the Neogene. Their composition indicates that important changes occurred in the vertical water circulation and in the salinity of bottom waters. The effects of these changes cannot be easily separated from the changes caused by temperature. Although temperature played a role we think that food was far more important, with respect to the occurrence and distribution of megafauna taxa. The quality and quantity of food may have been related with the oxygen content of the bottom waters.

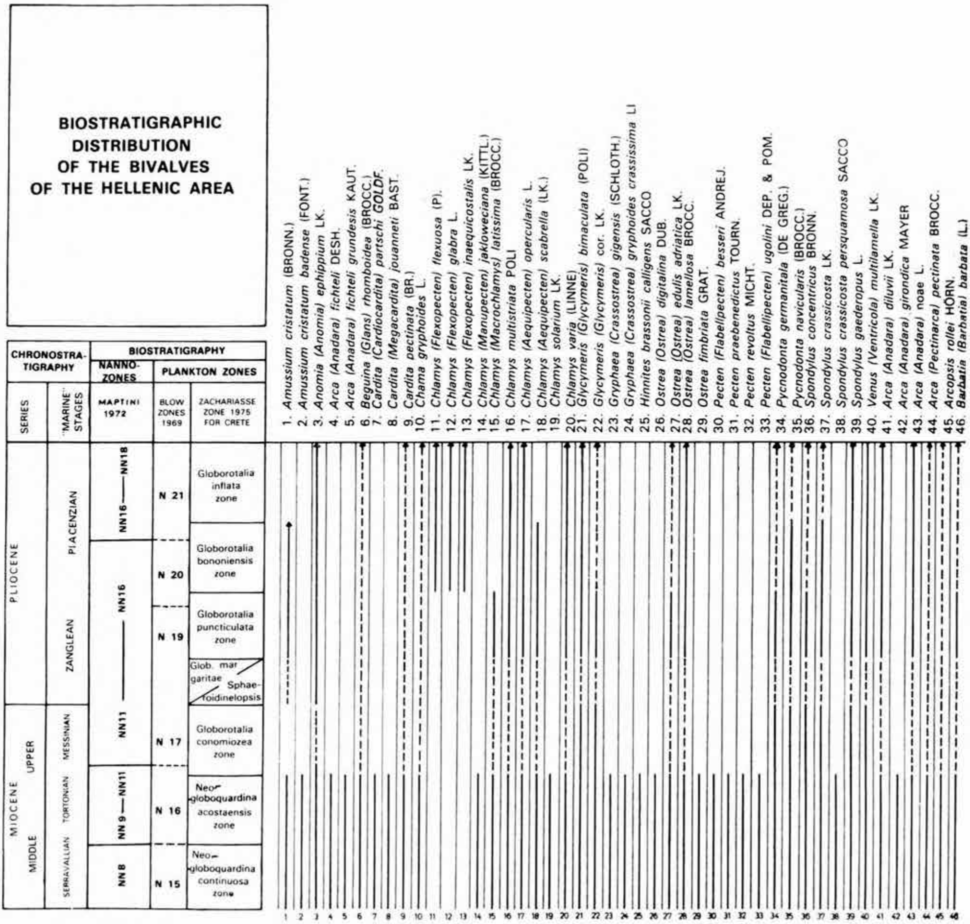
However, many problems were encountered as soon as we tried to get into detail and in particular when we tried to put a link with the ranges of taxa from noncoastal areas.

To achieve our goals we tried to correlate our marine megafaunal associations with marine microfossils collected from the same outcrops and particularly with planktonic Foraminifera, calcareous Nannoplankton, and, where possible, Ostracoda. The biozonation we propose is considered satisfactory for practical purpose. It concurs rather well with the planktonic foraminiferal and calcareous nannoplankton zonations of the Mediterranean bioprovinces.

Examination of the faunal succession shows a significant difference in faunal composition between the Miocene and Pliocene assemblages. The terminology for the zonation follows that of the International Subcommission on Stratigraphic Classification (1976). For the Greek Neogene, nine molluscan zones can be recognized.


BIOSTRATIGRAPHIC DISTRIBUTION OF THE GASTROPODS OF THE HELLENIC AREA
CHRONOSTRATIGRAPHY
SERIES
MARINE STAGES
BIOSTRATIGRAPHY
NANNO ZONES
MARTINI 1972
PLANKTON. ZONES
BLOW ZONES 1969
ZACHARIASSE ZONE 1975 FOR CRETE
PLIOCENE
MIOCENE
MIDDLE
UPPER
PIACENZIAN
ZANGLEAN
MESSINIAN
TORTONIAN
SERRAVALLIAN
NN16 —— NN18
NN16
NN11 —— NN16
NN9 —— NN11
NN8
N 21
N 20
N 19
N 18
N 17
N 16
N 15
Globorotalia inflata Zone
Globorotalia bononiensis Zone
Globorotalia puncticulata Zone
Glob. margaritae
Sphaeroidinelopsis
Globorotalia conomiozea Zone
Neo-globoquadrina acostaensis Zone
Neo-globoquadrina continuosa Zone
1. Anachis (Costoanachis) corrugata (BELD.)
2. Ancilla (Baryspira) glandiformis (LK.)
3. Ancilla (Baryspira) glandiformis conoidea DESH.
4. Aporrhais (Aporrhais) alata (EICHW.)
5. Apporhais (Chenopus) uttigerianus RISSO
6. Archimedielia (Torculoidella) spirata BR.
7. Astraea (Bolma) meynardi (MICHT.)
8. Astraea (Bolma) rugosa L.
9. Athleta (Athleta) ficulina rarispina (LK.)
10. Athleta (Athleta) pyrulaeformis R. HORN & AU.
11. Babylonia (Peridispaccus) brugadina (GRAT.)
12. Bittium (Bittium) reticulatum (DA COSTA)
13. Cancellaria (Solatia) barjonae dertoturrita SACCO
14. Cassis (Cassis) postmamillaris SACCO
15. Cerithium (Thericium) italicum dertonensis SACCO
16. Cerithium (Vulgocerithium) crenatum (BROCC.)
17. Cerithium (Thericium) vulgatum miospina SACCO
18. Cerithium (Thericium) europaeum europaeum MAY
19. Cerithium (Ptychocerithium) granulfum (BEL.) MICH.
20. Cerithium (Ptychocerithium) salmo BAST.
21. Charonia (Sassia) apenninica (SASSI)
22. Charonia (Sassia) tarbelliana (GRAT.)
23. Clavatula (Clavatula) asperulata ernesti TOULA
24. Clavatula (Clavatula) asperulata gracilis ERUN.-EREN.
25. Clavatula (Clavatula) calcarata GRAT.
26. Clavatula (Clavatula) calcarata francisci TOULA
27. Clavatula (Perrona) inedita BELL.
28. Clavatula (Perrona) jouanneti (DESM.)
29. Clavatula (Perrona) saubrigiana (GRAT.)
30. Clavatula (Perrona) taurinensis (MAY)
31. Clavatula (Calvatula) theodori calcaratiformis ERUN.-EREN.
32. Clavus (Drillia) bellardii DESM.
33. Clavus (Drillia) pustulatus BROCC.
34. Columbella (Alia) curta curta (DUJ.)
35. Conus (Lithoconus) antiquus LK.
36. Conus (Dendroconus) betulinoides LK.
37. Conus (Conolithus) dujardini DESH.
38. Conus (Chelyconus) enjesfeldensis R.HÖRN. & AU.
39. Conus (Chelyconus) fuscocingulatus BRONN.
40. Conus (Chelyconus) pelagicus inflatus ERÜN.-EREN.
41. Conus (Chelyconus) pelagicus karamanensis ERÜN.-EREN.
42. Conus (Lithoconus) mercati miocaenicus SACCO
43. Conus (Chelyconus) pelagicus BROCCHI
44. Conus (Rhijoconus) ponterosus ponterosus BR.
45. Conus (Rhijoconus) puschi MICHT.
46. Conus (Rhijoconus) pyrula mucronata ERÜN. EREN.
47. Corbula (Corbula) carinata hoernesi BEN.
48. Crepidula (Jamachus) crepidula unguis ORB.
49. Cymatium (Lampusia) affine friedbergi COSS. & PEYR.
50. Diodora (Lucapina) italica (DEFR.)
51. Dolium (Cadium) orbiculatum (BROCC.)
52. Euthria (Euthria) cornea L.
53. Euthria (Euthria) cornea curvirostris GRAT.
54. Fasciolaria (Pleuroploca) fimbriata BROCC.
55. Galeodes (Galeodes) cornutus AG.
56. Gemmula (Hemipleurotoma) annae R. HÖRN. & AU.
57. Gemmula (Hemipleurotoma) contigua BR.
58. Genota (Genota) ramosa ramosa (BAST.)
59. Gyrineum (Aspa) marginatum (BRONEN)
60. Hadriania craticulata L.
61. Hexaplex (Muricanthus) subasperrimus (ORB.)
62. Hinia (Uzita) limata (CHEMN.)
63. Latirus (Latirus) lynchi (BAST.)
64. Latirus (Latirus) valenciennesi (GRAT.)
65. Lemintina arenaria (L)
66. Lemintina arenaria dentifera (LK.)
67. Lemintina arenaria perpustulata SACCO
68. Lunatia catena helicina (BROC.)
69. Mitra (Tiara) scrobiculata (BROC.)
70. Mitra (Tiara) scrobiculata striatula BROC.
71. Mitra (Tiara) substriatula D'ORB.
72. Mitraria (Mitraria) ambigua ambigua (FBG)
73. Mitrella (Mitrella) semicandata (BON.)


Molluskan
Biostratigraphic Zonation
Chlamys inaequicostalis
Assemblage Zone
Pecten reghiensis
Assemblage Zone
Flabellipecten flabelliformis
Assemblage Zone
Flabellipecten Ass. Zone
nigromagnus
Chlamys
Scabrella Acme-zone
Neopycnodonta navicularis
Assemblage Zone
Gryphaea (Crassostrea)
gryphoides crassissima
Assemblage Zone
Chlamys solarium
Assemblage Zone
74. Narona (Sveltia) dertovaricosa SACCO
75. Nassa (Nassa) seyhanensis ERÜN. EREN.
76. Nassa (Nassa) subventricosa COSS. & PEYR.
77. Natica (Nacca) millepunctata LK.
78. Natica (Nacca) millepunctata tigrina DEFR.
79. Natica (Nacca) millepunctata magnopunctata SACCO
80. Natica (Nacca) millepunctata epiglottinoformis SACCO
81. Neverita olla (SERR.)
82. Ocinebrina sublavata sublavata (BAST.)
83. Olivancilaria (Aragonia) acuminata LK.
84. Patella (Patella) ferruginea GM.
85. Petaloconchus intortus LK.
86. Petaloconchus intortus turriteloides SACCO
87. Phos (Phos) connecteus hoernesi (SEMB.)
88. Polinices (Polinices) redemptus (MICHT.)
89. Polinices (Polinices) dertomamilla SACCO
90. Potamides tricinctus scalaritina SACCO
91. Potamides (Tympanotomus) trochlearispina subfenestrata SACCO
92. Purpura (Tritonalia) austriaca (R. HORN & AU.)
93. Purpura (Tritonalia) delbosiana friedbergi COSSM. & PEYR.
94. Pyrula lainaei BAST.
95. Ringicula (Ringiculella) auriculata paulucíae MORL.
96. Schilderia annullaria hoernesiana (SACCO)
97. Semicassis (Semicassis) miolaevigata SACCO
98. Sigaretus (Sigaretus) striatus SERR.
99. Strombus (Canarium) bonellii BRONGN.
100. Strombus (Strombus) coronatus DEFR.
101. Subula (Oxymeris) fuscata modesta (TRIST.)
102. Subula (Oxymeris) fuscata plicaria (BAST.)
103. Terebra (Terebra) acuminata acuminata BORS.
104. Terebra (Terebra) neglecta MICHT.
105. Turris (Turris) aquensis MICHT.
106. Turris (Turris) rotata DEFR.
107. Turitella (Archimediella) archimedis dertonatior SACCO
108. Turitella (Torculoidella) bicarinata bicarinata EICHW.
109. Turitella (Torculoidella) bicarinata subunocincta SACCO
110. Turitella (Archimediella) dertonensis dertonensis MAY
111. Turitella (Archimediella) dertonensis persulcata SACCO
112. Turitella (Archimediella) dertonensis subconica SACCO
113. Turitella (Haustator) eryna rotundata SCHF.
114. Turitella (Haustator) hoernesi ROLLE
115. Turitella (Haustator) partschi quatricincta BACHIN.
116. Turitella (Archimediella) pythagoraica pythagoraica HILBER
117. Turitella (Zaria) subangulata BROCC.
118. Turritella (Zaria) subangulata ditropis FONT.
119. Turritella (Turritella) tricarinata communis RISSO
120. Turitella (Haustator) tricincta BORS.
121. Turitella (Haustator) tricincta tricincta SACCO
122. Terebralia bidentata bidentata (DEFR.)
123. Terebralia bidentata cingulatior SACCO
124. Terebralia bidentata prolignitarium SACCO
125. Terebralia bidentata vignali COSSM.
126. Terebralia lignitarum lignitarum EICHW.
127. Thais (Stramonita) exilis PARTSCH
128. Thericium (Cladiothericium) varicosum transiens SACCO
129. Trigonostoma (Trigonostoma) scrobiculatum (HÖRN.)
130. Turricula (Surcula) reevei BLED.
131. Turritella (Torculoidella) triplicata (BROCCHI)
132. Turritella (Torculoidella) triplicata crasseplicata SACCO
133. Turritella (Torculoidella) triplicata dertosimplex SACCO
134. Turritella (Torculoidella) triplicata superneaplicata SACCO
135. Turritella (Turritella) turris capriciosa ERÜN.-EREN.
136. Typhis (Typhinellus) tetrapterus (BRONN.)
137. Vexillum (Vexillum) plicatulum BROCC.
138. Vitularia (Vitularia) lingua-bovis (BAST.)
139. Xenophora deshayesi MICHT.
140. Xenophora infundibulum (BR.)
141. Clavus (Drillia) allionii BELL.
142. Fusus (Fusus) longirostris BROC.
143. Nassa (Uzita) inaequicostata BELL.
144. Nassa (Hinia) incrassata (MÜLL.)
145. Nassa (Nassa) musiva BROCC.
146. Natica (Neverita) josephinia RISSO

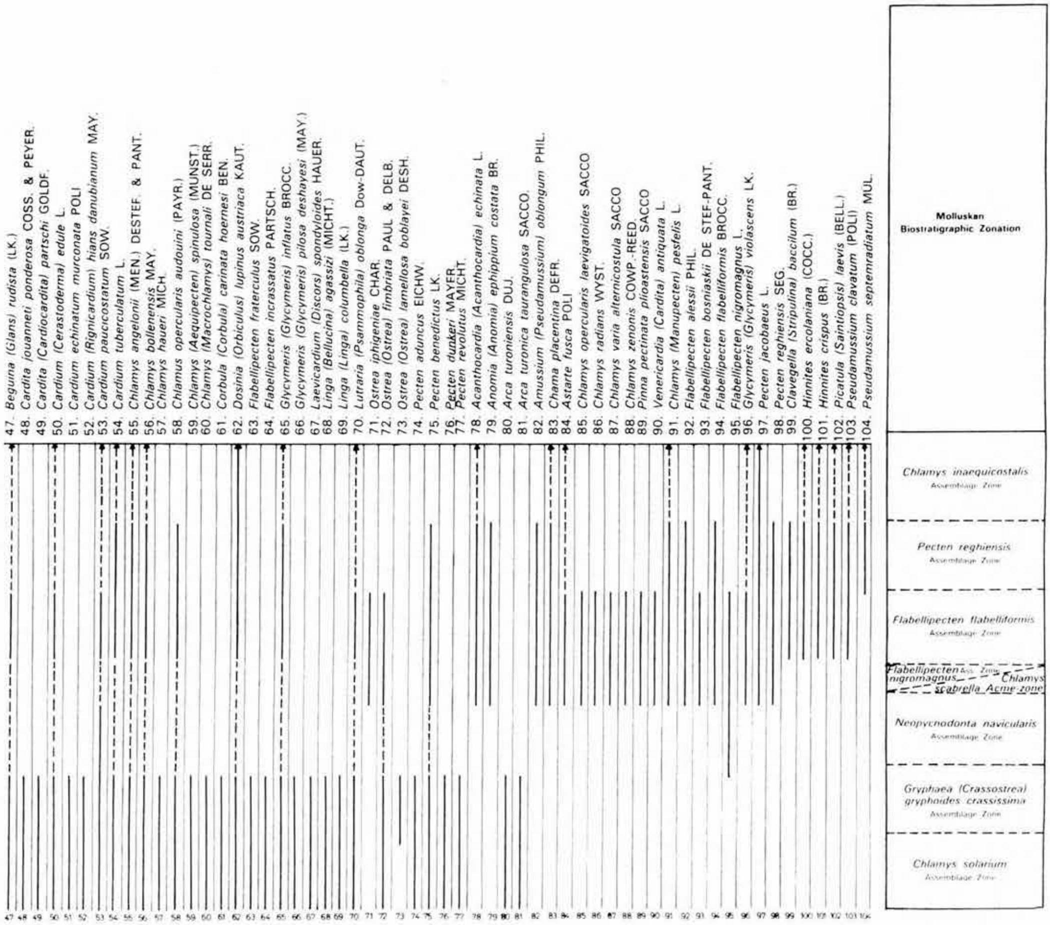



The proposed biozonation corresponds to assemblage biozones. A biostratigraphic assemblage zone is a body of strata whose content of fossils or of fossil of a certain kind, taken in its entirety, constitutes a natural assemblage or association that distinguishes it in biostratigraphic character from adjacent strata.

Only the *Chlamys scabrella* zone is an acme zone, characterized by the relative abundance (reflecting environmental conditions) of the nominate taxon.

The proposed assemblage zones are based mainly on molluscs and the name given represents the name of the most frequent and representative species.

A list of the most frequent co-cccurrent species is also given as well as the reference section and the distribution, for every proposed assemblage zone, which can be used as a standard of reference in the identification of the assemblage zone elsewhere.



## Mollusca zonation

### *Chlamys solarium* assemblage zone

Diagnosis: This zone is characterized by the presence of *Chlamys solarium* LK. It contains the range of *Chlamys solarium* LK. prior to the significant presence of *Chlamys latissima* (BR.) and *Gryphaea (Crassostrea) gryphoides crassissima* (LK.).

Analysis: This first zone is marked by the co-occurrence of *Pecten revolutus* MICHT., *Clypeaster tumescens* IMB., *Cardita partschi* GOLDF. etc.

It covers the biostratigraphic interval N15 of *Neogloboquadrina continuosa* biozone and the lowest part of *Neogloboquadrina acostaensis* zone that corresponds partly to the Serravalian. The top of the zone is marked by the presence of assemblages transitional between *N. continuosa* and *N. acostaensis*. The undefined lower boundary is beyond the interval dealt with in this paper.

Reference section: Section Ambelos of Gavdos Formation.

Distribution: This zone is found in the sections Ambelos and Panayia of the Gavdos Formation.

Both sections are caracterized by the presence of calcareous fine- or coarse-grained sandstones as also marly limestonesstratified or not and of a variable thickness (FREUDENTHAL, 1969, ZACHARIASSE, 1975, ANAPLIOTIS, 1975). To this zone correspond also the section: Kalamavka of Kalamavka Formation, and Vassiliki of Makrylia Formation (DERMITZAKIS, 1969, FORTUIN, 1977, DERMITZAKIS and THEODORIDIS, 1984), Chairethiana of Kissamou Formation (DERMITZAKIS and DIACANTONI, 1979).

*Gryphaea (Crassostrea) gryphoides crassissima* **assemblage zone**

Diagnosis: This zone is characterized by the presence of larger *Chlamys (Gigantopecten) latissima* (BROCC.) and *Flabellipecten besseri* ANDR. It contains the local range of *Gryphaea (Crassostrea) gryphoides crassissima* (LK.) prior to the significant presence of *Neopycnodonta navicularis* (BR.).

Analysis: The base of this zone coincides with the appearance of assemblages of *Neogloboquadrina acostaensis* types without transitional types between *N. continuosa* and *N. acostaensis*. From the megafauna the co-occurrence of *Pecten (Flabellipecten) ugolinii* DEP. et ROMAN, *Amussium cristatum badense* FONT., *Arca (Anadara) fichteli* DESH. *Arca (Anadara) turoniensis* DUR., *Cardita jouanneti portentosa* COSSMAN et PEYROT, *Ancilla (Baryspira) glandiformis* (LK.), *Clavus (Drillia) bellardii* DESM., *Conus (Conolithus) dujardini* SESH., *Genota (Genota) ramosa ramosa* BAST., *Polinices (Polinices) redemptus* (MICHT.), *Turritella (Archimediella) dertonensis dertonensis* MAY., *Terebralia lignitarum lignitarum* EICHWALD etc, is characteristic.

Apart from these molluscan assemblages the associations of *Clypeaster altus* LK. have been observed in several studied sections of this zone. The remaining fauna is not confined to this zone but it is a characteristic association including *Terebratula sinuosa* BR., *Clypeaster portentosus* SESM., *Helliastraea reussana* M. ELOW-HAME., *Ballanophyllia concina* REUSS etc.

Reference section: The Apostoli section of Apostoli Formation in Rethymnon Province of Grete.

Distribution: This zone is found in the Apostoli Formation in the section Apostoli (MEULENKAMP, TSAPRALIS (1969), FREUDENTHAL (1969), SISSINGH (1972), ZACHARIASSE (1975), GEORGIADES-DIKEOULIA (1979), as well as in the section Selli in Rethymnon district. Both sections mainly consist of marls and clays. The basal parts are commonly composed of conglomerates and sands. The middle and upper parts consist of organic limestones alternating with clays and marls (MEULENKAMP, 1969). Also this zone is present in the sections: Achladia in Sitia district (DERMITZAKIS et al., 1978), Vassiliki in lerapetra district (DERMITZAKIS and THEODORIDIS, 1984). Kalogeri and Sopates in Ierapetra district (DERMITZAKIS, 1980a; 1980b).

It covers the bistratigraphic interval $N_{16}$ *Neoglobaquadrina acostaensis* zone that corresponds to the Tortonian.

*Neopycnodonta navicularis* **assemblage zone**

Diagnosis: This zone is characterized by the co-occurrence of *Neopycnodonta navicularis* (BR.), *Chlamys fasciculata* MIL., *Chlamys multistriata* POLI, *Barbatia (Anadara) clathrata* (DEFR.), *Dorocidaris* sp. *Emarginula cancellata* PHIL., *Pecten aduncus* EICHW. etc.

Analysis: This zone is marked at its base by the significant presence of *Neopycnodonta navicularis* (BR.); this is shortly after the last occurrence of *Gryphaea (Crassostrea) crassissima* (LK.) and *Pecten (Flabellipecten) besseri* ANDR., which determine, together with other previously referred species, the upper part of the *Gryphaea (Crassostrea) gryphoides crassissima* zone.

In some sections like Aghios Charalambos and Myrtos in eastern Crete in Ierapetra district only *Neopycnodonta navicularis* (BR.) is present whereas the other taxa appear to be absent.

This zone covers the biostratigraphic interval of $N_{17}$ and part of $N_{18}$ zones of BLOW (1969). According to the Cretan planktonic zones (ZACHARIASSE, 1975), it corresponds partly to *Globorotalia conomiozea* zone. From chronostratigraphical point of view zone represents the interval of the Messinian and partly the lowermost Pliocene.

The impoverishment of the fauna, which is observed in the middle and upper parts of the zone, is reflected by the decreasing number of species. In several sections, these species-restricted fauna merely consist of a single species of *Chlamys fasciculata* MIL. or *Chlamys multistriata* P. or *Ostrea lamellosa* BR. or *Dorocidaris* sp. Samples in the interval may be barren.

R e f e r e n c e s e c t i o n : Section Akropotamos in Strymon basin (northern Greece).

D i s t r i b u t i o n : This zone includes: the section Akropotamos in Strymon basin (GEORGIADES-DIKEOULIA and VELITZELOS, 1983; STEFFENS et al., 1979); the upper part of Armenopetra section in Viannou district (DERMITZAKIS and GEORGIADES-DIKEOULIA, 1979); the lowermost part of Aghios Charalambos section in Ierapetra district (DERMITZAKIS, 1980): the lower and middle parts of Palaeokastron section in Sitia district (SONNENFELD, HUDEC and DERMITZAKIS, 1985); the Heraklion section (GEORGIADES-DIKEOULIA and MÜLLER, 1984).

At the reference section over the non-marine fluvial and lacustrine clastic sediments of the base, lense-shaped gypsum and marls alternate with small sapropelitic intercalations. Above these sediments marls thicker in size and sands alternate up to the upper part where there is travertine and sandstone.

### *Chlamys scabrella* acme zone

D i a g n o s i s : The zone is defined by relative abundance of representatives of the species *Chlamys scabrella* LK.

A n a l y s i s : The upper and lower boundaries of this zone coincide with the limits of the acme. In Kephallinia the zone has been observed in section Katelio, where the top of the zone is characterized by fairly abrupt decrease in frequency. In some samples the representatives of *Chlamys scabrella* LK. constitute up to 10% of the total fauna. Some samples contain two types of *Chlamys scabrella* LK. the type *Chl. scabrella* LK. and the type *Chl. bollenensis* MAY-EYM. Other samples contain mainly the type *scabrella* which by its relative abundance characterizes in fact this interval. But there are also some samples with only a few representatives of *Chl. scabrella* LK. (GEORGIADES-DIKEOULIA, 1965, DERMITZAKIS and GEORGIADES-DIKEOULIA in prepar.).

Although its lower boundary seems to coincide with that of the *Flabellipecten nigromagnus* zone, *Chlamys scabrella* LK. does not get extinct at the same level in the section Vroutsas of Carpathos. The *Chlamys scabrella* acme is succeeded by a fauna which contains *Aequipecten radians* NYST, *Anomia ephippium* L., *Ostrea lamellosa* BR., *Arca pectinata* BR., *Cardium papillosum* POL. (DERMITZAKIS and GEORGIADES-DIKEOULIA, in press). Such a fauna is characteristic of the upper part of *Flabellipecten nigromagnus* zone.

The *Chlamys scabrella* acme zone covers the biostratigraphic interval of the Lower Pliocene, approximately equivalent to *Globorotalia margaritae* zone and to *Sphaeroidinellopsis* acme zone (sensu ZACHARIASSE, 1975). Also this zone corresponds partly to $N_{18}$ of the BLOW zonation (1969) and also partly to $NN_{12}$, $NN_{13}$ of the nannofossil zonation of MARTINI (1972).

Reference section: Katelio section, Kephallinia island.

Distribution: The zone is recognized in the section Katelio of Kephallinia island (DERMITZAKIS and GEORGIADES-DIKEOULIA, in prep.).

Both sections mainly consist of sandy marls alternating with marly limestones and lightly cemented sandy calcarenites.

### *Flabellipecten nigromagnus* assemblage zone

Diagnosis: The zone is defined by the presence of the zonal marker.

Analysis: The base of *Flabellipecten nigromagnus* zone is placed at the first occurrence of the *Flabellipecten nigromagnus* SACCO type. This type species disappears in the upper boundary of the zone.

In this zone the co-occurrence of species like: *Chlamys bollenensis* MAY-EYMAR, *Aequipecten zenonis* C.—R., *Aequipecten radians* NYST, *Anomia ephippium* L., *Ostrea lamellosa* BR., *Arca pectinata* BR., *Cardium papillosum* POL., etc is also remarkable.

This zone represents the biostratigraphic interval of Lower Pliocene and seems to coincide with that of *Chlamys scabrella* zone.

Reference section: Section Alimos of Trachones formation in Attica district.

Distribution: The zone is recognized in the section Alimos of Trachones Formation (PAPP, STEININGER and GEORGIADES-DIKEOULIA, 1978). Also, in the lower part of the section Ag. Thomas in Aegina island (GEORGIADES-DIKEOULIA and DERMITZAKIS, 1983), in Strymon basin (STEFFENS et al., 1979) and in Serres basin (GEORGIADES-DIKEOULIA and KARYSTINEOS, in prep.). The base of the zone consists of coarse transgressive conglomerates, overlain by sands and sandy marls.

### *Flabellipecten flabelliforms* assemblage zone

Diagnosis: The zone is defined by the last occurrence of *Flabellipecten nigromagnus* SACCO (type) and the entry of *Pecten reghiensis* SEG. (type).

Analysis: The zone is well documented in Crete. It was identified in the section Profitis Ilias near Ploutis village. The section Armenopetra at Keratokampos village in Viannou area shows the transition between this zone and the next younger one. The transition is marked by the first occurrence of *Pecten reghiensis* SEG. The transitional part of this zone to the next one has also been recognized in Ammudhares section in Ierapetra district. Except the significant presence of *Flabellipecten flabelliformis* BR., the co-occurrence of species like: *Flabellipecten allessii* PHIL., *Amussium cristatum* BR., *Pecten jacobaeus* L., *Pecten benedictus* LK., *Spondylus crassicosta* LAM., *Ostrea iphigeniae* CHAR., *Lunatia catena helinica* BR., *Conus (Chelyconus) pelagicus* BR., is remarkable.

This zone covers the biostratigraphic interval of $N_{19}$ and partly of $N_{20}$ of the BLOW zonation (1969) as also well as the *Globorotalia puncticulata* zone (sensu ZACHARIASSE, 1975). Tentatively corresponds to the $NN_{14}$ and $NN_{15}$ nannofossil zones of MARTINI (1972). From the chronostratigraphical point of view this zone represent part of the Zanclean time-span. According to the calcareous nannoplankton biozonation of THEODORIDIS (1984) this interval belongs to the *Ceratolithus acutus* and *Ceratolithus rugolata* biozone. All the above data allow us to conclude that this biozone can be correlated with the late Early Pliocene Middle Pliocene time-span.

Reference section: Section Profitis Ilias at Ploumis village of Kourtes Formation is Heraklion district.

D i s t r i b u t i o n : The zone is recognized at the Kourtes Formation and at the section Profitis Ilias in Ploutis village of Heraklion district (DERMITZAKIS and GEORGIADES-DIKEOULIA, 1984), as well as in the upper part of Armenopetra section in Keratokampos village of Viannou district (DERMITZAKIS and GEORGIADES-DIKEOULIA, 1979, 1982; DERMITZAKIS and KOUROUNI, 1983), and at Paleopolis section, west of Avlemon, Kythera island (MEULENKAMP et al., 1977). The sections in general are characterised by sandy beds alternating with white greyish marls, sandy marls, breccias and sandstones.

**Pecten reghiensis assemblage zone**

D i a g n o s i s : The zone is defined by the first appearance of *Pecten reghiensis* SEG. (type) and the last occurrence of *Ostrea iphigeniae* CHARALAMBAKIS (type).

A n a l y s i s : The zone can be recognized at Aghia Marina section of Aegine island, but the biostratigraphic correlation is based only on nannofossil biozones $NN_{15}$ (upper part) and $NN_{16}$ (THEODORIDES pers. com.). The biostratigraphic interval seems to confirm the biostratigraphic evidence of Aghios Thomas section (BENDA et al., 1976). Since the impoverished planktonic fauna makes impossible the biostratigraphic correlation with planktonic biozones we can refer to the section Angistri (Angistri island) in which the zone is well documented with the *G. bononiensis* zone. The section contains the zonal marker *Pecten reghiensis* SEG. (type) and also the last appearance of *Ostrea iphigeniae* CHARALAMBAKIS, *Pecten benedictus* LK. and *Amussium oblongum* PHIL. This biozone can be recognized in Karpathos island in Pigadhia Formation we can recognize and also in Kythira island in section Paleopolis.

This biozone is characterized by the co-occurrence of *Chlamys angelonii* (MEN.), *Chlamys flexuosa* P., *Chlamys pesfelis* L., *Chlamys opercularis* L., *Pseudamussium clavatum* P., *Amussium oblongum* PHIL., *Hinnites ercolaniana* (COCC.) *Hinnites crispus* (BR.), *Pecten jacobaeůs* L., *Pecten benedictus* L., *Ballonophyllia* sp., *Dosinia exoleta* L., *Pinna tetragona* BR., *Amycyclina semistriata* BR., *Archimediella (Torculoidella) spirata* (BR.), *Tellina palmata* L., *Corbula gibba* OLIVI, *Terebratula* spp.

This zone covers the biostratigraphic interval of *Globorotalia bononiensis* biozone that corresponds to the upper Middle Pliocene. Good evidence of the zone of *Pecten reghiensis* was found in Prassas section (GEORGIADES-DIKEOULIA, 1979) and also in Kithyra island section Stenokambos.

R e f e r e n c e s e c t i o n : Section Aghia Marina in Aegina island.

D i s t r i b u t i o n : The zone is recognized in the following sections: Aghia Marina Aegina island (GEORGIADES-DIKEOULIA and DERMITZAKIS, 1983). Angistri island (SYMEONIDIS and DERMITZAKIS, 1973), Karpathos island Pighadia section (DERMITZAKIS and GEORGIADES-DIKEOULIA in press), Prassas section, Heraklion district (GEORGIADES-DIKEOULIA, 1979), Kithyra island, section Stenokambos (MEULENKAMP et al., 1977). Koufonisi island upper part of the section Papaloukos (DERMITZAKIS and THEODORITHIS, 1978).

The lithological character of this zone is exhibited by layers of laminated marls gradually becoming homogeneous with large distinct burrows overlain by a bioclastic limestone.

**Chlamys inaequicostalis assemblage zone**

D i a g n o s i s : The zone is defined at its base by the last appearance of *Ostrea iphigeniae* CH., *Pecten benedictus* LK. and *Amussium oblongum* PH.

A n a l y s i s : The base of the zone is coincided with the extinction of *Ostrea iphigeniae* CHAR., *Pecten benedictus* LK., *Amussium oblongum* PHIL.

The zone is well documented in the north part of Milos island at the section Appolonia in which the megafauna occurs in homogeneous marly deposits which are intercalated by sapropelic layers and dark laminated marls. Also in the upper part of Francocastello section in Chania province, samples contain *Chlamys inaequicostalis* LK., *Chlamys glabra* L., *Pecten jacobaeus* L., *Diodora italica* DEFRANCE etc in Rhodos island at Lindos section.

Reference section: Apollonia section, Milos island.

Distribution: The zone is recognized in Apollonia section, Milos island (DERMITZAKIS, GEORGIADES-DIKEOULIA, SYMEONIDIS, 1986, in press): Francocastello section. Chania Province (DERMITZAKIS, in press). Lindos section, Rhodos island (GEORGIADIS, 1978).

## Conclusions

In general it can be concluded that the ranges of the megafaunal taxa in the Hellenic area reflect the effects of eastwards migrations of bio(sub)provinces during some Neogene time spans.

If this conclusion is tenable, we have to accept that apart from differences between for example, such subbioprovinces as the Rhone basin and the Cretan basins there are comparable differences in the range of taxa for even smaller-scale areas within the Mediterranean. These criteria have to be kept in mind if we consider the diverging opinions on the biostratigraphic significance of several taxa.

In conclusion, we postulate that more refined local data sets are needed before a satisfactory biostratigraphic scheme that would be representative for the Neogene of the entire Mediterranean can be established.

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# EVOLUTION OF ENDEMIC MAMMAL FAUNAS IN THE GARGANO NEOGENE (ITALY): THE PROBLEM OF ENDEMIC VARIATION AS A CHRONOLOGICAL TOOL

by

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*Foreword.* The occurrence of rich faunal assemblages from the "terra rossa" fillings of karstified limestones near Apricena, in the Gargano peninsula, is known since 1971.

The age of the endemic faunas is unfortunately still uncertain but a Pliocene age is the most likely.

We made a tentative correlation with the marine stratigraphy by means of the reconstruction of the paleoenvironmental changes based on mammal data.

The features of the "terra rossa" fossils clearly show the occurrence of endemic evolutionary phenomena that took place in insular conditions. Both paleogeographic reconstructions and studies on the mammal assemblages have clearly shown that during the Neogene the Gargano area was a part of an archipelago that ranged in the southern adriatic region (DE GIULI et al., 1986, DE GIULI et al., in press a, b, and 1987). The occurrence of well documented fossil faunas from Gargano has suggested to analize the problems of the relative chronology of endemic fossil assemblages in islands.

In an island, time-sequence reconstructions are usually based on scant data and founded on:

*1* the occurrence of simple size-increase or size-decrease trends in any available taxon;

*2* the evaluation of the degree of morphological changes due to endemic evolution.

This approach, even if theoretically correct, may cause errors if the island is a part of an archipelago, since they can be contradicted by faunal exchanges between islands. In archipelagos, phenomena of evolutionary radiation may give rise to the occurrence of several sister species that live in different islands and differ only in size and evolutionary stage. Therefore, in a single island, large sized forms may disappear and be replaced by extremely similar immigrants of smaller size coming from the neighbouring islands. Likewise the immigration of a primitive species can replace an extinguished more evolved species.

We will try to demonstrate that a reliable chronological hypothesis is possible, in such cases, only if the available samples are so rich to allow a well-grounded reconstruction of the events (immigrations, extinctions and so on) that interested the studied area.

*Material and methods.* Eight samples collected in a very small area between Poggio Imperiale and Apricena were considered for our chronological analysis. The samples and the number of species per sample are indicated in Table 1. We made preliminary investigations on size and morphology changes in the glirid *Peridyromys*, in *Apodemus* and in a still unnamed small-sized galericine. Evolutionary changes in the endemic murid *Microtia* and in *Prolagus* have been object of a more detailed quantitative analysis as these taxa are the best represented ones in the samples and show marked evolutionary changes. The morphological changes and size variations have been analyzed in the first lower molar of *Microtia* and in the third lower premolar of *Prolagus*. Detailed data are reported in DE GIULI et al., in press, b.

The general criterion adopted for setting the samples in a chronological sequence is to assume as younger the sample which has at least one population with a significant frequency of a new apomorphic character, or, at a lower degree of confidence, the sample in which there is a population with a greater frequency of the more evolved morphotype. Size variations have been employed for chronological purposes when they could be recognized as a part of a phyletic trend.

*Table 1*

| TAXA \ SAMPLES | F15 | F21b | F21c | F1 | F8 | F9 | SG | F32 |
|---|---|---|---|---|---|---|---|---|
| Small Galericini | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 |
| Crocidurinae | 1 | 1 | 1 | / | / | / | / | / |
| Prolagus | 1 | 1 | 1 | 1 | 1 | 1 | 2 | 2 |
| Peridyromys | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 |
| Gliridae (giant-form) | 1 | 1 | 1 | 1 | 1 | 1 | 1 | / |
| Elyomys | 1 | / | / | / | / | / | / | / |
| "Kowalskia" | 1 | 1 | 1 | 1 | 1 | 1 | 1 | / |
| Apodemus | 1 | 1 | 1 | 1 | 1 | 1 | 1 | 1 |
| Microtia | 3 | 2 | 2 | 2 | 3 | 3 | 3 | 1 |
| Total number | 11 | 9 | 9 | 8 | 9 | 9 | 10 | 6 |

These criteria, applied to different taxa, may produce contradictory chronological sequences. The more parsimonious hypothesis should be chosen in such a case, since it is the one that implies the existence of fewer phylogenetic lineages, migrations and extinctions.

*The chronological reconstruction.* Data from *Peridyromys* and the galericine show the occurence of two distinct groups of samples. In the first group, including the samples F15, F21b and F21c, the glirid is represented by large sized individuals while the insectivore has primitive features in the upper p3. In the second group (including all the remaining samples) the glirid is smaller size. This group must be considered younger than the other one owing to the more evolved features of the upper p3 in the insectivore.

The data from *Microtia*, allow a more detailed chronological hypothesis that agrees with the data from the glirid and the insectivore. The general chronological sequence inferred from *Microtia* data is synthetized in Fig. 1A. This hypothesis assumes the occurrence at least of five phylogenetic lineages and four events of migration and extinction. The samples F8 and F9 do not show significant differences and must be considered of equivalent age. The sample F32 must be considered older than F8 and SG since the *Microtia* population from F32 belongs to the small lineage occuring in F8, F9 and SG and has a significantly lower frequency of evolved morphotypes.

The data from *Prolagus* agree with the previous hypothesized time sequence, except for the relative position of the samples F8, F9 and F32. The sequence of samples according to the data from *Prolagus* is shown in Fig. 1B. The sample F8 should be considered older than F9 owing to the low frequency of evolved morphotypes. The sample F32 is made up by two populations, as in SG. The largest population confidently belongs to the main evolutionary lineage. As a marked size increase trend can be detected along this lineage and the F32 population is significantly larger than the SG one; we should consider the F32 sample younger than SG. *Prolagus* data support, in conclusion, a chronological hypothesis alternative to the one desumed from *Microtia* data.

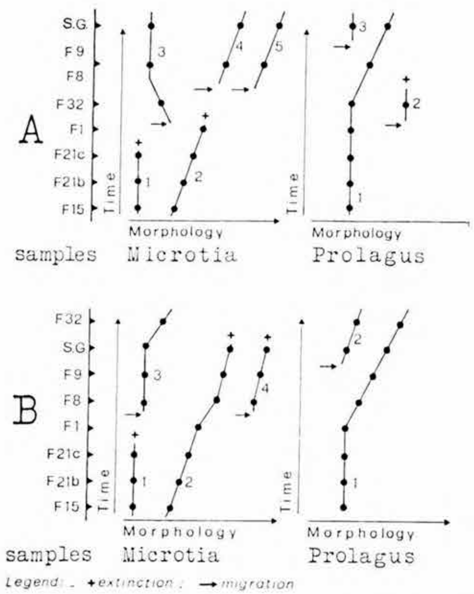


*Fig. 1.* Two different chronological sequences of the samples, involving alternative phylogenetic reconstructions for Microtia and Prolagus

A: Sequence according to the Microtia data, B: sequence according to the Prolagus data

The data from *Apodemus* agree with the chronological hypothesis desumed from *Prolagus*. The specimens from F32 are clearly more apomorphic than those of all the other samples, thus confirming the hypothesis of F32 being the youngest sample. Furthermore, the overall continuous trend towards a size-increase occurring along the whole sequence of samples strengthens the general chronological hypothesis.

In conclusion the hypothesis of chronological succession inferred from *Prolagus* and confirmed by *Apodemus* data is the most reliable since it is the one that assumes less phylogenetic lineages, less events of migration and extinction to have occurred. It also involves the minimum number of contradictions among the data of the analysed taxa. The discrepancy between the low morphological complexity degree in F32 *Microtia* and the higher complexity in the *Microtia* from the hypothesized older samples, may well be considered as a random floating of morphotypes in a phyletic lineage that does not show an overall significantly evolutionary change.

The previous discussion evidences the difficulties that are meet to order the endemic faunal assemblages from archipelagos in a chronologic sequence, particularly when the isolation of the area is so strong that no migrations from the mainland occur. In such cases a reliable chronological hypothesis is possible only by comparing each other, in a consistent number of taxa, the results of the analysis of evolutionary changes and the relevant phylogenetic reconstructions. At last, the chronological hypothesis requires the evaluation of the involved events of immigration and extinctions occured in the studied paleo-island.

*Palaeoenvironment*. The changes in relative abundance of taxa and the variations in diversity degree of micromammals along the hypotesized time sequence have evidenced the occurrence of two main environmental changes in the studied area:

*1* the evolution towards a dryer climate occurring from sample F15 to F1;

*2* the occurrence of the widening of the emerged area in the time before the F8 sample followed by a strong reduction that possibly caused the extinction of the largest micro-mammals before the F32 time.

Paleoenvironmental changes can be used to attempt a time correlation with the marine stratigraphy. The drying out of the climate and the sea regression may agree to place our samples in a latest Messinian—earliest Pliocene age as well as in the Middle Pliocene (*Globorotalia* gr. *crassaformis* zone). An emersion of the area near Apricena during the latter age is documented by local field geology data (VALLERI, 1985).

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# CORRELATION OF NEOGENE DIATOMACEOUS EARTH DEPOSITS IN HUNGARY

by

M. HAJÓS

A new and important endeavour of geological research in Hungary is to carry out the paleogeographic reconstruction and the most complete description of the lithostratigraphic units of Hungary. A basic prerequisite for achieving this is the proper understanding of the geological units.

The examination of fossil diatoms and other siliceous microfossils associated with them is a scientific tool which cannot be dispensed with in up-to-date and complex geological research and exploration.

Where the rock is uniform and homogeneous over tens or hundreds of metres in a vertical section, being apparently sterile or containing no calcareous fossils in a rock sequence deposited during continuous sedimentation, there the siliceous protists such as Archaeomonas, Silicoflagellata, Ebriida, Diatoma and Radiolaria etc are usually present.

The fossil assemblages of diatoms and other siliceous unicellular organisms provide clues to understanding variations of different kinds such as lateral and vertical changes in lithology and facies. They give information on the genetic history of the sedimentary basin involved and, if a larger chronological unit is concerned, on the date of sedimentation as well.

The author has since the fifties studied the Tertiary and Quaternary deposits of Hungary, from the Eocene up to the Recent. In this context she has examined a total of 678 fossil diatom taxa, in addition to the associated representatives of Archeomonas, Silicoflagellata, Ebriida, Phytolitharia, Silicospongia, etc which she has recovered from more than 5500 samples from a total of 72 Miocene localities taken in a wider sense.

The possibilities for sampling were rather limited. The examined rock samples came from surface exposures and boreholes in areas selected during geological mapping and given preference in national mineral exploration projects. In the first place, the diatomaceous deposits of the Mecsek Mountains and their surroundings, the North Hungarian Highland Range and the Tokaj Mountains were studied, and only a smaller fraction derived from similar deposits in the Transdanubian Central Range and the Miocene- and Pliocene-filled marginal basins of the Great Hungarian Plain (Fig. 1).

The deposits involved are connected with the Miocene—Pliocene volcanic and postvolcanic activities, being primarily areas of tuff ejecta and areas of tuffite deposition and geyserite accumulation subsequent to the former. The acidic volcanics were observed, almost as a strict rule, to be overlain by diatomaceous deposits that had been accumulated in a shallow-water sea, in nearshore, lagoonal and partly landlocked bay environments. Comparatively thicker diatomaceous earth, i.e. diatomite deposits were formed in dependence on the volcanic activity, at the optimum of pH and dissolved $SiO_2$ content of the seawater and, consequently, in a pure, oxygen-rich, well-oxygenated environment (e.g. Hidas, Szurdokpüspöki, Erdőbénye, etc) (Table 1). This accounts for the fact that, when examining deeper-water basin deposits layer by layer, regardless of whether Oligocene or Miocene or Pliocene deposits were intersected by the particular borehole, they are always found to be sterile in terms of diatomaceous earth accumulation.

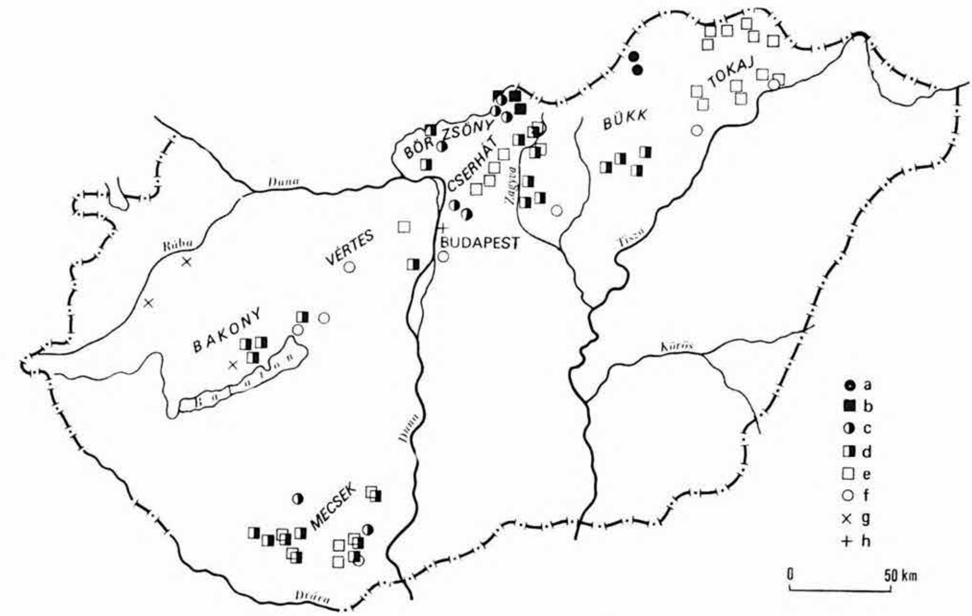


*Fig. 1.* Occurrences of diatomaceous deposits in Hungary
*a* Eggenburgian, *b* Ottnangian, *c* Karpatian, *d* Badenian, *e* Sarmatian, *f* Pannonian, *g* Pontian, *h* Holocene

Because of their biofacies the gravity points of diatomaceous earth deposition will always fall outside the areas of the carbonate sedimentation cycle, being lithologically associated preferentially with the fine-sandy, for the most part carbonate-free deposits. If the % $CaCO_3$ content and the percentage of Diatoma specimens in representative layer by layer samples be plotted as a function of sedimentation time, so the percentage maximum of Diatoma will almost appear at the minimum of the carbonate curve.

Hence the lack of regionally continuous diatomaceous deposits or beds in Hungary. They are restricted to local occurrences, forming lenses of varying size even within formations. We have had to study isolated rock sequences or lenses the vertical stratigraphic connections of which could not be traced from layer to layer even in case of no break in sedimentation. The stratigraphic succession and the even smaller interim changes in facies could be inferred from changes in the composition of the flora.

Accordingly, the diatomaceous deposits of Hungary do not fit directly in the local formation system established on a lithogenetic base, forming intraformational or possibly transformational "facies lenses".

*Table 1*

**Occurrences of diatomaceous deposits in Hungary**

● EGGENBURGIAN
Alsószuha
Sajókaza
■ OTTNANGIAN
Ipolytarnóc
Litke
Mihálygerge
◑ KARPATIAN
Nógrádszakál
Litke
Piliny
Diósjenő
Mogyoród
Fót
Magyaregregy
Zengővárkony
◨ BADENIAN
Szurdokpüspöki
Gyöngyöspata
Hasznos
Petőfibánya
Szokolya
Magyarkút
Bernecebaráti
Eger-Tihamér
Demjén
Novaj
Borsodgeszt
Mátraverebély
Mátraszőlős
Diósd
Herend
Szentgál
Bánd
Várpalota
Komló
Hidas
Pétervárad
Szilágy
Tekeres
Magyarhertelend
Magyarszék
Bodolyabér
□ SARMATIAN
Magyarszék
Tekeres
Szilágy
Hird
Hosszúhetény
Pécsvárad
Hidas
Budajenő
Kozárd
Ecseg
Buják
Bér
Hasznos
Mátraverebély
Sárospatak
Tállya
Erdőbénye
Hercegkút
Abaújvár
Cekeháza
Gönc
Pusztafalu
Erdőhorváti
Füzérkajata
Füzérradvány
Kéked
○ PANNONIAN
Szilágy
Vilonya
Csór
Csákvár
Budapest—Népliget
Gyöngyösvisonta
Bogács
Tállya
Sárospatak
× PONTIAN
Pula
Gérce
Várkesző
+ HOLOCENE
Újpest

Since the diatomaceous deposits do not constitute a continuous sedimentary chain, their correlation too can be performed on a bio- rather than lithostratigraphic basis. In attempting to determine their position in the stratigraphic scale we cannot help placing them in the afore-mentioned formations or in lenses in-between.

Ecological requirements of siliceous protists are: a considerable dissolved $SiO_2$ content of the water, its acidic or, at the most, neutral pH and its being suffiently penetrated by sunlight and well-oxygenated. It follows from the former that, regardless of the lithological order of the formations, the diatomaceous deposits constitute facies either dissecting, interrupting or intersecting them. In other words, the biozonal boundaries must not necessarily coincide with the formation boundaries.

The most important aim of this work has been to fix the stratigraphic position of the diatomaceous deposits. To achieve this goal, the author has sought to establish biozones and abundance zones by determining the stratigraphic ranges of a few short-lived taxa of typical morphology representing the most important members of a total of hundreds of identified species and present in greatest abundance (dominant taxa) (Table 2).

The author has listed for each particular stratigraphic unit just a few taxa considered to be the most important stratigraphic index taxa, because the biostratigraphic relationships, the phylogenetic evolution and the evolutionary lineages of some new taxa are more distinct and more easily traceable. The diatomaceous deposits of Hungary's lithostratigraphic units, formations, their order of sedimentation and biofacies are characterized by fossil assemblages (Table 2).

**Statigraphic position of Miocene diatom and**

| EPOCH | Paleomagnetic epoch | Radiometric ages M.Y. | Regional stages: Mediterranean | Regional stages: Central Paratethys | Regional stages: Eastern Paratethys | | Standard zones: Planktonic Foraminifers Blow, 1969 | Standard zones: Nannoplankton Martini, 1971 |
|---|---|---|---|---|---|---|---|---|
| UPPER | | | | Dacian | Kimmerian | | N18 | NN12 |
| | 5 | 54 | Messinian | | | | N17 | |
| | 6 | | | | | | | |
| | 7 | | Tortonian | Pontian | Pontian | | N16 | NN11 |
| | 8 | | | | | | | |
| | 9 | | | Pannonian | Meotian | | | NN10 |
| | | | | | Chersonian | Sarmatian | | |
| | 10 | 118 | | | | | N15 | |
| MIDDLE | 11 | | Serravollian | Sarmatian | Bessarabian | | N14 | NN9 |
| | 12 | | | | | | | NN8 |
| | | | | | Volhynian | | N13 | NN7 |
| | 14 | | | | | | | |
| | 15 | | | Badenian | Konkian | | N12 | NN6 |
| | | | | | Karaganian | | N11 | |
| | | | | | Tschokrakian | | N10 | NN5 |
| | 16 | 168 | Langhian | | Tarchanian | | N9<br>N8 | |
| LOWER | | | Burdigalian | Karpatian | Kozachurian | | N7 | NN4 |
| | 17 | | | Ottnangian | | | N6 | NN3 |
| | 18 | | | Eggenburgian | Sakaraulian | | N5 | NN2 |
| | 19 | | | | | | | |
| | 20 | | Aquitanian | Egerian | Caucasian | | N4 | NN1 |
| | 21 | 232 | | | | | | |
| | 22 | | | | | | | |

**silicoflagellata zones in the Central Paratethys** *Table 2*

| Biostratigraphy | | | | | |
|---|---|---|---|---|---|
| Zones | | | | | |
| Diatoma | | | Silicoflagellata | | |
| ŘEHÁKOVÁ, 1975 | ŘEHÁKOVÁ, 1977 | HAJÓS, 1985 | HAJÓS, 1985 | BACHMANN and MARTINI, 1972 | DUMITRICA, 1985 |
| | | *Anomoenois sphaerophora* | | | |
| | | *Actinoptychus trilobatus Coscinodiscus jambori* | | ? | |
| | | *Frag. bituminesa* | | *Dyctiocha rombica* | |
| *Coscinodiscus doljensis, Anaulus simplex* | *Coscinodiscus doljensis* | *Haynoldiella* | | | *Distephonus lemmermanni D. macilentus Distehanopsis longispinus* |
| ? | ? | *Anaulus simplex* | *Deflandriocha intercalaris* | | |
| | | *Navicula pinnata* | ? | | *Distephanopsis stauracanthus* |
| *Denticula lauta Actinocyclus ingens* | *Denticula punctata* | *Rhaponeis mediopunctata, Denticulopsis lauta* | *Distephanus crux v. longispina* | *Dictyocha triacantha* | *Paracannopilus picassoi* |
| | *Coscinodiscus lewisanus* | *Surirella costata C. pannonicus* | ? | | |
| *Diploneis microtatos Aulacodiscus grunowii* | *Raphidodiscus marylandicus* | *Rhaphoneis parilis* | *Mesocena elliptica* | | |
| *Actinoptychus truanii f. trivittata Coscinodiscus moronensis* | *Coscinodiscus moronensis* | *Rhaphoneis subtilissima* | *Corbisema triacantha v. flexuosa* | | |
| ? | *Actinoptychus amblyoceras* | *Melosira hispanica* | | *Naviculopsis navicula* | |
| *Cladogramma conicum* | *Cladogramma conicum v. campanutatum* | ? | ? | *Naviculopsis lata* | |

The diatomaceous areas are discussed on the basis of the type areas of the biozones by taking into consideration the degree of understanding of the areas concerned and, not in the last place, the sea currents pattern and plaeogeographic distribution of the diatomaceous sedimentary basins and by establishing their hierarchy of importance.

Having summarized the biostratigraphic and faciological data of three decades of research, I attempted at summing up the stratigraphic results concerning the Hungarian diatomaceous deposits. On the basis of the nomenclatural evaluation of Diatoma, by taking into consideration and critically evaluating the relevant literature available, I sought to determine their lateral distribution and vertical range.

In listing the fossil assemblages I have adopted the valid data of Van Landingham (1967—1979).

My stratigraphic statements and conclusions concern primarily the Hungarian deposits of the Central Paratethys. In comparisons with more remote areas, I have not restricted myself to my own data, having used data from J. Pantocsek (1886—1905) and other authors of a monographic coverage of the neighbouring countries (Krestel, Řeháková, Temiskova-Topalová, etc) as well.

The paleogeographic distribution of diatoms in the Central Paratethys (Fig. 2) has been reconstructed on the basis of data hitherto available. Finally, I made an attempt at a chrono-, litho- and biostratigraphic correlation of the deposits.

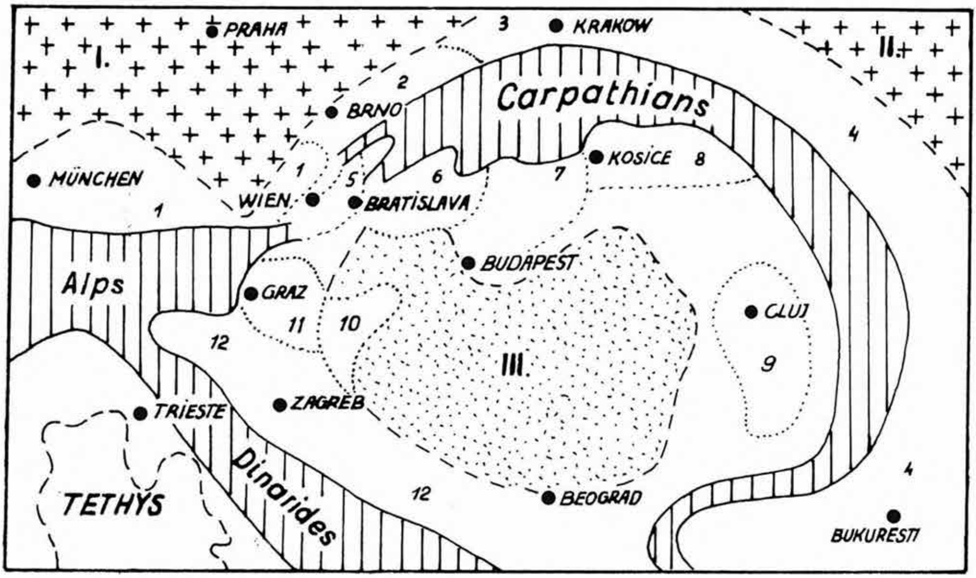


*Fig. 2.* Lower and Middle Miocene sedimentary basins of the Central Paratethys (Seneš et al., 1971)

*1* N foredeep of the Alps, *2—4* external foredeep of the Carpathians, *5* Vienna basin, *6—8* sedimentary basins in N Hungary and S Slovakia, *9* Transylvanian basin, *10* Transdanubian basin, *11* Graz basin, *12* Zagreb basin—Drava—Sava basin.—*I* Bohemian massif, *II* Podolian massif, *III* Pannonian basin

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# DATING OF MIOCENE ACID AND INTERMEDIATE VOLCANIC ACTIVITY IN HUNGARY

by

G. HÁMOR, L. RAVASZ-BARANYAI, J. HALMAI, K. BALOGH and E. ÁRVA-SÓS

Since 1974 a systematic *K/Ar* research has been carried out on Miocene volcanic rocks in Hungary as a joint project of the Hungarian Geological Institute and the Institute of Nuclear Research of the Hungarian Academy of Sciences.

The great number of dated rocks (from over 400 localities) and the critical evaluation of radiometric ages enabled us, in spite of the frequent disagreement of *K/Ar* and geologic ages, to establish the temporal evolution of Miocene volcanism in Hungary.

The first results of *K/Ar* dating were presented in 1977 at the XIth Congress of the CBGA, Kiev (KADOSA BALOGH et al., 1980). At the VIIth Congress of the RCMNS, Athens, the chronologic conclusions of *K/Ar* datings on rhyolitic pyroclastics have been summarized (G. HÁMOR et al., 1979). The first review of radiometric studies on Miocene volcanic rocks in Hungary was presented at the XIIth Congress of the CBGA, Bucharest (1981) by E. ÁRVA-SÓS et al. (1983) and an other paper dealt with the *K/Ar* results on Miocene volcanites from NE Hungary (K. BALOGH et al., 1983).

From among the radiometric ages presented here those of first order chronostratigraphic importance are used for the establishment of a revised radiometric time scale for the Central Paratethys Neogene by D. VASS et al. (this volume). The present work has a twofold aim.

*1* In view of the new age data an updated picture of the evolution of Miocene acidic and intermediate volcanic activity in Hungary is presented.

*2* The average ages of Miocene stage boundaries in the Central Paratethys deduced from radiometric ages and palaeomagnetic data on this territory (D. VASS et al., this volume) may slightly differ from those established for Hungary. As a step towards the assessment of regional variation of the ages assigned to stage boundaries, the differences will be emphasized which exist between the boundary ages defined for the Central Paratethys as a whole on one hand and for Hungary on the other. The sites of rocks referred to in the text are shown in Fig. 1.

In Hungary, the oldest Miocene volcanites occur in SE Transdanubia. In the Mecsek Mts the andesite at Komló overlies Eggenburgian sediments (according to palynology by M. SÜTŐ-SZENTAI, 1983). Rhyolite tuffs are intercalated in Eggenburgian strata and rhyolite tuffs and flood tuffs occur over the andesite too. According to palynological data (E. NAGY, 1969) the rhyolitic sequence is of Ottnangian—Eggenburgian age. Recent dating of the andesite resulted in $19.5 \pm 0.9$ Ma (borehole Komló 170, K. BALOGH et al., unpublished). The two oldest reliable ages on the rhyolite tuff are $19.6 \pm 1.9$ Ma (Szászvár, Szekernye valley) and $19.5 \pm 1.4$ Ma (Váraljaquarry) (G. HÁMOR et al., 1979). These dates are regarded as within the Eggenburgian, near the Eggenburgian—Ottnangian boundary.

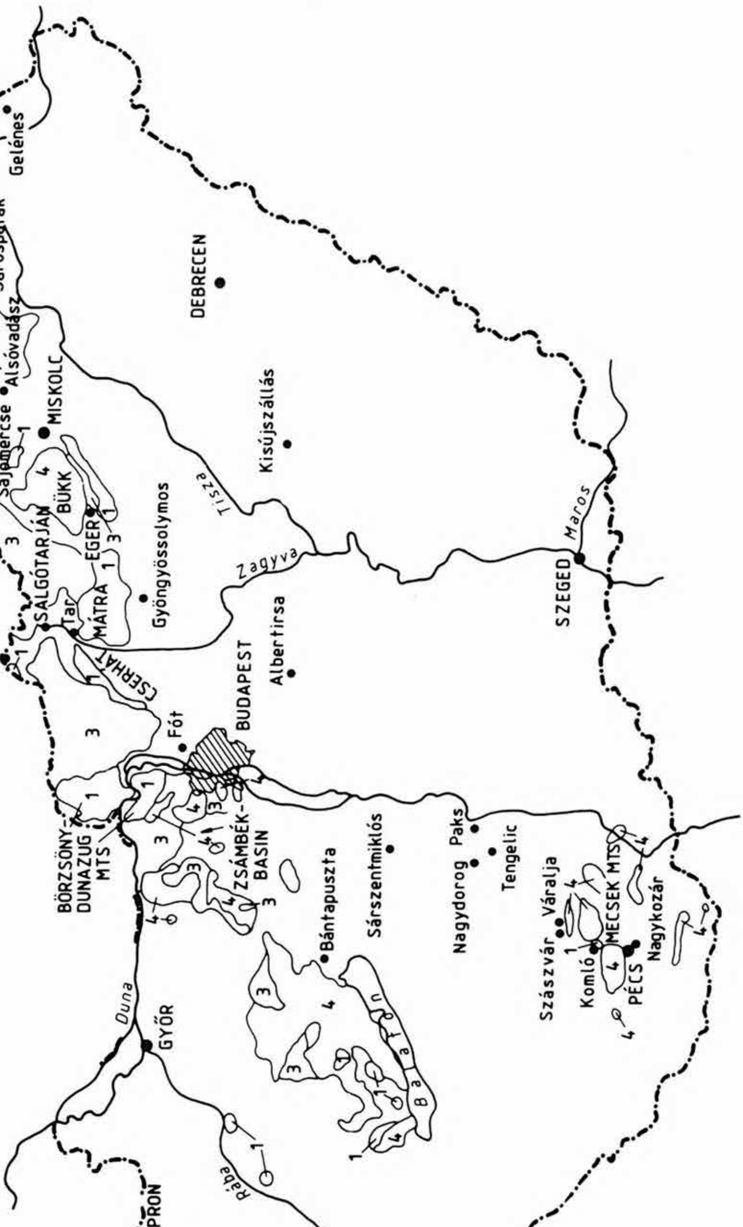


*Fig. 1.* Neogene volcanic areas in Hungary

*1* Neogene volcanic rocks, *2* Neogene sequence, *3* Palaeozoic sequence, *4* Mesozoic and Palaeozoic formations

Rhyodacitic tuffs belonging to the Middle Rhyolitic Tuffs are also present in the Mecsek Mts. Their mean age (16.4 Ma) is in a full accordance with the average age of the Middle Rhyolite Tuffs (16.4±0.8 Ma) calculated for the whole territory of Hungary (G. Hámor et al., 1979).

North of the Mecsek Mts borehole Tengelic 2 penetrated rhyodacitic volcanites covered by sediments belonging to zone NN5 (A. Nagymarosy, pers. comm.). The average age on 11 biotites from different depths is 15.6±0.7 Ma, but the mean age of 15.7±0.55 Ma measured for 9 biotites from lava rocks is more reliable (K. Balogh, 1984). According to stratigraphy and *K/Ar* age this sequence is a younger member of the Middle Rhyolite Tuffs.

Dacite was reached by borehole Nagydorog 1, N of the Mecsek Mts. Biotites from 4 cores from different depths gave resulted an average age of 19.29±0.37 Ma (K. Balogh, 1984).

A mean age of 18.5±1.7 Ma has been determined of 6 biotites from dacitic tuffs intersected by borehole Paks 2. (E. Árva-Sós et al., 1983).

Alkali rhyolite is exposed at the surface at Sárszentmiklós. According to its radiometric age of 17.0±0.7 Ma it is associated with the Middle Rhyolite Tuffs. This age is, however, somewhat doubtful since due to the absence of biotite the measurements were made on magnetically split fractions of the altered rocks and the presence of older contaminating material might be suspected. Similar alkali rhyolite from borehole Albertirsa 1 provided a younger age of 14.3±0.5 Ma (E. Árva-Sós et al., 1983).

At Bántapuszta the rhyolite tuff is situated in the lower part of the Upper Badenian (J. Kókay, Gy. Raincsák, 1983), which yielded an average biotite age of 15.0±0.4 Ma (K. Balogh, 1984).

In the Zsámbék basin (W of Budapest) boreholes Budajenő 3 and Perbál 6 intersected a rhyolite tuff layer in the lower third part of the Sarmatian (Á. Jámbor, 1976; Cs. Ravasz, 1978). Its average *K/Ar* age, measured on biotites, turned out to be 13.7±0.5 Ma (G. Hámor et al., 1979; K. Balogh, 1984).

In the Börzsöny–Dunazug Mts Miocene volcanic activity started with the ejection of andesite pyroclastics in the Early Badenian (T. Báldi, J. Kókay, 1970; G. Hámor, Á. Jámbor, 1971). According to radiometric dating (G. Hámor et al., 1979), the oldest volcanites are contemporaneous with the main eruption phase (16.4±0.8 Ma) of the Middle Rhyolitic Tuffs. On the basis of *K/Ar* dating on biotites and amfiboles (K. Balogh et al., unpublished) the production of volcanic material terminated about 15.0 Ma B.P., while according to stratigraphy it was restricted to the Early Badenian (T. Báldi, J. Kókay, 1970).

In N Hungary Miocene volcanism started with the eruption of the Lower Rhyolite Tuffs in Early Ottnangian time (G. Hámor, Á. Jámbor, 1971). Due to the strong alteration the radiometric dating of this level is difficult. This is illustrated by the $^{39}Ar/^{40}Ar$ release spectrum (Fig. 2), recorded by Y. Takigami (Tokyo University) on a biotite, from Ipolytarnóc which indicates strong recrystallization of the biotite. The total fusion age is 19.0±1.4 Ma, this deviates from the *K/Ar* ages of 16.3±1.6 Ma and 16.0±2.0 Ma measured in Debrecen (K. Balogh, 1984). The deviation may be attributed to the great analytical errors and the extreme sensitivity needed for $^{39}Ar/^{40}Ar$ measurement on young minerals. On the basis of the latest datings the most likely age of Lower Rhyolite Tuffs in N Hungary is about 19.0 Ma or somewhat younger.

In the Mátra Mts the Lower Rhyolite Tuffs were followed by Karpatian andesites (Gy. Varga et al., 1975). Due to their altered character these are still undated. The Karpatian andesites were followed by rhyodacite tuffs in the uppermost Karpatian (M. Hajós, 1968) dated at Tar as 16.4±1.1 Ma (G. Hámor et al., 1979). *K/Ar* age of

16.2±0.6 Ma has been measured (G. HÁMOR et al., 1979) on the rhyodacite tuff at Fót (NE of Budapest) which is in the same stratigraphic position (J. HALMAI, 1981). The rhyodacite tuffs were followed by andesites in the Early Badenian and these andesites are covered with rhyolite at Gyöngyössolymos (GY. VARGA et al., 1975). A highly reliable age of 15.9±0.5 Ma was measured on the rhyolite (K. BALOGH, 1984). Due to the frequent loss of radiogenic argon, it is difficult to date the end of the volcanism. The most reliable ages on the youngest andesite dikes fall in the range of 14.0—15.0 Ma (K. BALOGH et al., unpublished). This makes it likely that the volcanism continued in the Late Badenian, too. In the Zagyva trench (between the Cserhát and Mátra Mts) *K/Ar* ages as young as about 10 Ma have been measured (G. HÁMOR et al., 1978; K. BALOGH, 1984). This may be explained by postvolcanic tectonism or hydrotermal activity.

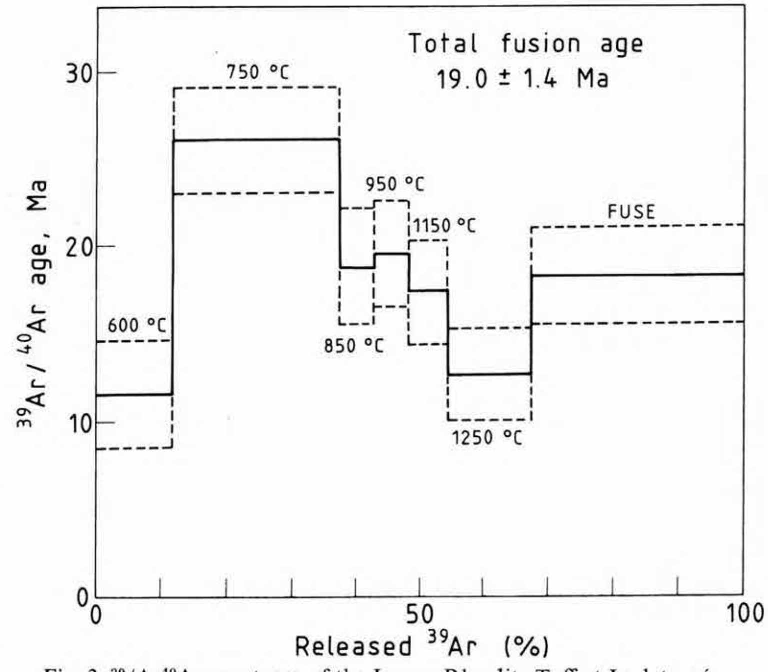


*Fig. 2.* $^{39}/Ar^{40}Ar$ spectrum of the Lower Rhyolite Tuff at Ipolytarnóc. Measured by Y. TAKIGAMI. Tokyo Univ 1984.

Andesitic and basaltic-andesitic volcanic activity is detected N of the Bükk Mts (Sajómercse, Dubicsány, Kelemér) in the Sarmatian and Pannonian. Radiometric ages (9.5—12.7 Ma, E. ÁRVA-SÓS et al., 1983; K. BALOGH, 1984) are in agreement with the stratigraphy.

In borehole Alsóvadász 1 (Cserehát) the rhyolite tuff is situated directly above the Karpatian—Badenian boundary. The biotite based average age (15.6±0.7 Ma) is in accordance with the stratigraphic position within the limits of experimental precision (E. ÁRVA-SÓS et al., 1983).

The chronology of Miocene volcanism in the Tokaj Mts and Trans-Tisza region has been treated in detail by K. BALOGH et al. (1983) and Z. PÉCSKAY (1983).

In the Tokaj Mts Miocene volcanic activity started with rhyolite tuffs in the Badenian and it is characterized by parallel rhyolitic, dacitic and andesitic material production till the end of Sarmatian. Alunite crystals from veins in the uppermost Sarmatian (T. ZELENKA, 1964; GY. RADÓCZ, 1969; Á. JÁMBOR, 1971) yielded *K/Ar* ages of 10.7—11.0 Ma (K. BALOGH et al., 1983; Z. PÉCSKAY, 1983) proving that the Sarmatian—Pannonian boundary can not be younger than 11.0 Ma in this area. In the Tokaj Mts volcanic activity continued with dacite and andesite production in the Pannonian and terminated with the basalt eruption at Sárospatak 9.4±0.5 Ma B.P. (V. SZÉKY-FUX et al., 1980).

In the Trans-Tisza region the oldest Miocene volcanites are rhyolites reached by borehole Kisújszállás-ÉK 1. The radiometric age of 18.25±0.3 Ma (E. ÁRVA-SÓS et al., 1983) may be accepted as the start of Miocene volcanism in this area. In the central part of the Trans-Tisza region Miocene volcanism was introduced by andesites in the Karpatian, it continued with alternating rhyolitic and andesitic products and terminated with rhyolite tuffs in the Sarmatian (K. BALOGH et al., 1983; Z. PÉCSKAY, 1983; V. SZÉKY-FUX, 1985). Along the boundary of Hungary and the USSR rhyolite tuff and rhyolite production finished near the end of Sarmatian (L. KULCSÁR, 1968). *K/Ar* ages measured on biotite and lava rock fall in the range of 11.0—11.3 Ma (K. BALOGH et al., 1983). Though a very small rejuvenation can not be excluded the radiometric dates support that the Sarmatian—Pannonian boundary is not older than 11.5 Ma. On the other hand Lower Pannonian dacite tuff from borehole Nagykozár 2 (S Hungary) yielded an older age (11.6±0.5 Ma; K. BALOGH, 1984). Accordingly, the Sarmatian—Pannonian boundary may be older in the southern part of Hungary.

In the Pannonian the acidic and intermediate volcanism terminated and alkaline basaltic volcanic activity started, which lasted till the end of Pliocene (K. BALOGH, et al., 1985).

## Chronostratigraphic conclusions

The age of Eggenburgian—Ottnangian boundary is 19.0—19.5 Ma. In S Hungary the older, in N Hungary the younger datum is more likely.

The age of Ottnangian—Karpatian boundary can not be directly established from the datings performed in Hungary.

The age of 16.4 Ma averaged for the Middle Rhyolitic Tuffs (G. HÁMOR et al., 1979) is a good approximation for the Karpatian—Badenian boundary. It is coherent with the age adopted by D. VASS et al. (1987), but it is slightly younger than suggested by F. RÖGL and F. STEININGER (1983). The Middle—Late Badenian boundary can not be younger than 15.0 Ma. The probable age of Badenian—Sarmatian boundary is about 14.0 Ma, somewhat older than suggested by D. VASS et al. (1987) and F. RÖGL and F. STEININGER (1983). 11.5±0.5 Ma is accepted for the Sarmatian—Pannonian boundary. This agrees well with the age given by F. RÖGL and F. F. STEININGER (1983). In S Hungary the older, in NE Hungary the younger datum is more likely.

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# GENERAL CHARACTERISTICS OF PANNONIAN s.l. DEPOSITS IN HUNGARY

by

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The dominantly fine-grained, brackish-water sediments with large thicknesses which are underlain by Sarmatian and overlain by Pleistocene sedimentary rocks are generally considered Pannonian formations. Their extension covers the Pannonian and Transylvanian basins which belonged to the Central Paratethys sea (Fig. 1). The Pannonian formations constitute the last but one major sedimentary cycle of filling-up of these basins, isolated from the world ocean and also from the eastern Paratethys. This inland sea lay on the unstable basement between the European and African plates. Its water progressively freshened and the depression was completely filled up in spite of the extensional subsidence of an average of 1300 m and a maximum of 5000 m.

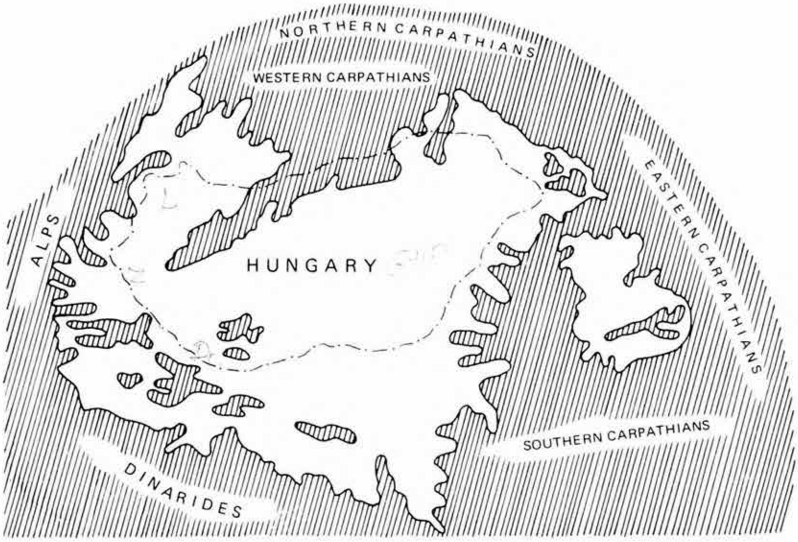



*Fig. 1.* Distribution of the Pannonian formations in the Carpathian basin

A detailed correlation between the Pannonian sedimentary complex and the marine sequences has not been solved, despite stratigraphic studies carried on for more than hundred years. The beginning of deposition of the Pannonian strata between 11 – 12 Ma seems to be well dated. The end of deposition is defined by the Pliocene/Pleistocene boundary. However, the age of this boundary is controversial and the determination of this boundary in the sedimentary sequence is often question-

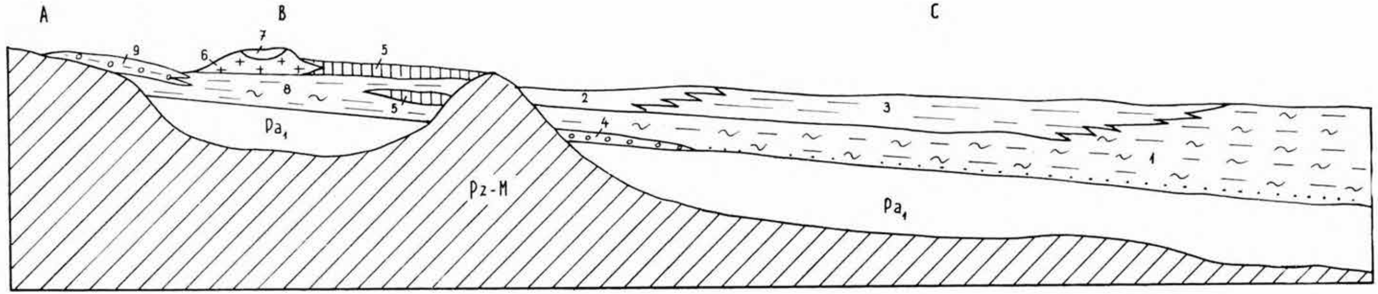



*Fig. 2.* Theoretical sketch of the position of facies units belonging to the Dunántúl Supergroup formations (compiled by Á. JÁMBOR, 1984)

*1* Basin sediments of argillaceous marl and sand, *2* mountain marginal variegated clay, sand, *3* lignite, huminitic mud, *4* abrasional pearl-pebbles, quartz-sand, *5* freshwater limestone, *6* basaltic tuffite, *7* alginite, 8 lagoonal argillaceous marl, sand, *9* fluviatile variegated clay and sand.—A, B, C, Pz—M: see Fig. 3, $Pa_1$ = Peremarton Supergroup beds

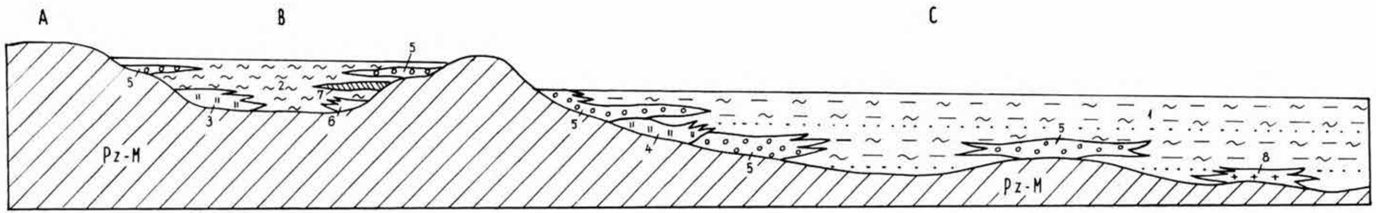



*Fig. 3.* Theoretical sketch of the position of facies units belonging to the Peremarton Supergroup formations (compiled by Á. JÁMBOR, 1984)

*1* Open water argillaceous marl, sand, *2* lagoonal argillaceous marl, *3* mountain marginal variegated clay, pebble, *4* basin-marginal variegated clay, *5* abrasional pearl-pebbles, *6* lignite, huminitic mud, *7* freshwater limestone, *8* volcano-sedimentary beds.—A = highlands-uplands, B = lagoon of the mountain margin, C = basin, Pz—M = formations predating the Peremarton Supergroup able. For the acceptance of the 1.8 Ma or 2.4 Ma boundary an international agreement is necessary. It is even more difficult to determine the boundary of the Lower and Upper Pannonian which is now called the Pannonian s. str. and Pontian boundary. The *K/Ar* radiometric age determination of Transdanubian basalts yielded 5—6 Ma; paleomagnetic data suggest 8 Ma, and 7 Ma was inferred from the Mollusca and Vertebrata faunal investigations. Moreover, the meaning of these data is not very clear because of the poor understanding of the contemporaneous facies conditions.

In Hungary the Pannonian (s.l.) formations can be found in more than three-quarters of the country (75,000 $km^2$) and are often covered by Pleistocene deposits. Only Paleo-Mesozoic inselbergs rise above the surface of the Late Cenozoic basin. The elevation of the surface of the basin varies from 79 m to 250 m above sea level, which is the consequence of an Early Pleistocene uplift accompanied by intensive erosion and followed by further subsidence. The outcrops of the Pannonian formations can be found at the margin of the basin and in the Transdanubian Central Range. In addition, the Pannonian formations can be studied on the basis of many thousand exploration wells for petroleum, water and other mineral resources, and by the help of about 30,000 kilometers of high resolution seismic profiles.

The Pannonian (s.l.) formations are monotonous petrologically. In the lower Peremarton Supergroup and in the inner parts of the basin a sequence of less variability can be found, while in the upper Dunántúl Supergroup and at the basin margins the gelogic column is more heterogeneous (Figs. 2 and 3).

The most important rockforming minerals are quartz, feldspars, micas (muscovite, chlorite, biotite), clay minerals (illite, smectite) and heavy minerals (mainly of metamorphic origin, dominantly granulites), authigenes: calcite, dolomite, bacteriogenic pyrite and limonite. Accordingly, the complex is considered to be of average molasse composition.

The thickness of the Pannonian basin-fill varies strongly from place to place. There are many smaller or larger deep troughs separated by elevated basement swells. The greatest thicknesses can be found in the southeastern Great Hungarian Plain (Fig. 4).

The Pannonian usually develops with gradual transition from the underlying Sarmatian. In elevated areas a paraconformable contact can be observed. In this case the dip of the two complexes is the same, and both are brackish and usually pelitic. There is no obvious evidence of a break in sedimentation or subareal erosion. The detailed palaeontologic investigations, however, prove in this case the lack of the lowermost biostratigraphic zones. This kind of contact is explained by simultaneous or subsequent subaquatic erosion. The Pannonian (s.l.) formations always overlie unconformably the pre-Sarmatian beds. Transgressive sequences can be observed in three different levels in the Pannonian complex: at the middle and upper levels of the Peremarton-, and at the lower level of the Dunántúl Supergroup. As a consequence of the transgressions, Pannonian sedimentation extended to larger and larger areas.

In the caprock of the Pannonian formation a Middle and Upper Pleistocene periglacial series can be found which usually consists of fluvial sand, gravel and loess. It overlies the Pannonian sediments with little angular unconformity over a great part of Hungary. This is due to a latest Pannonian uplift and subareal erosion in Transdanubia and northern Hungary. In deep basin areas, however, the Pleistocene sedimentary sequence is complete, the transition is gradual and no erosional hiatus can be supposed, regardless of whether 1.8 Ma or 2.4 Ma is taken to be the Pannonian/Pleistocene boundary.

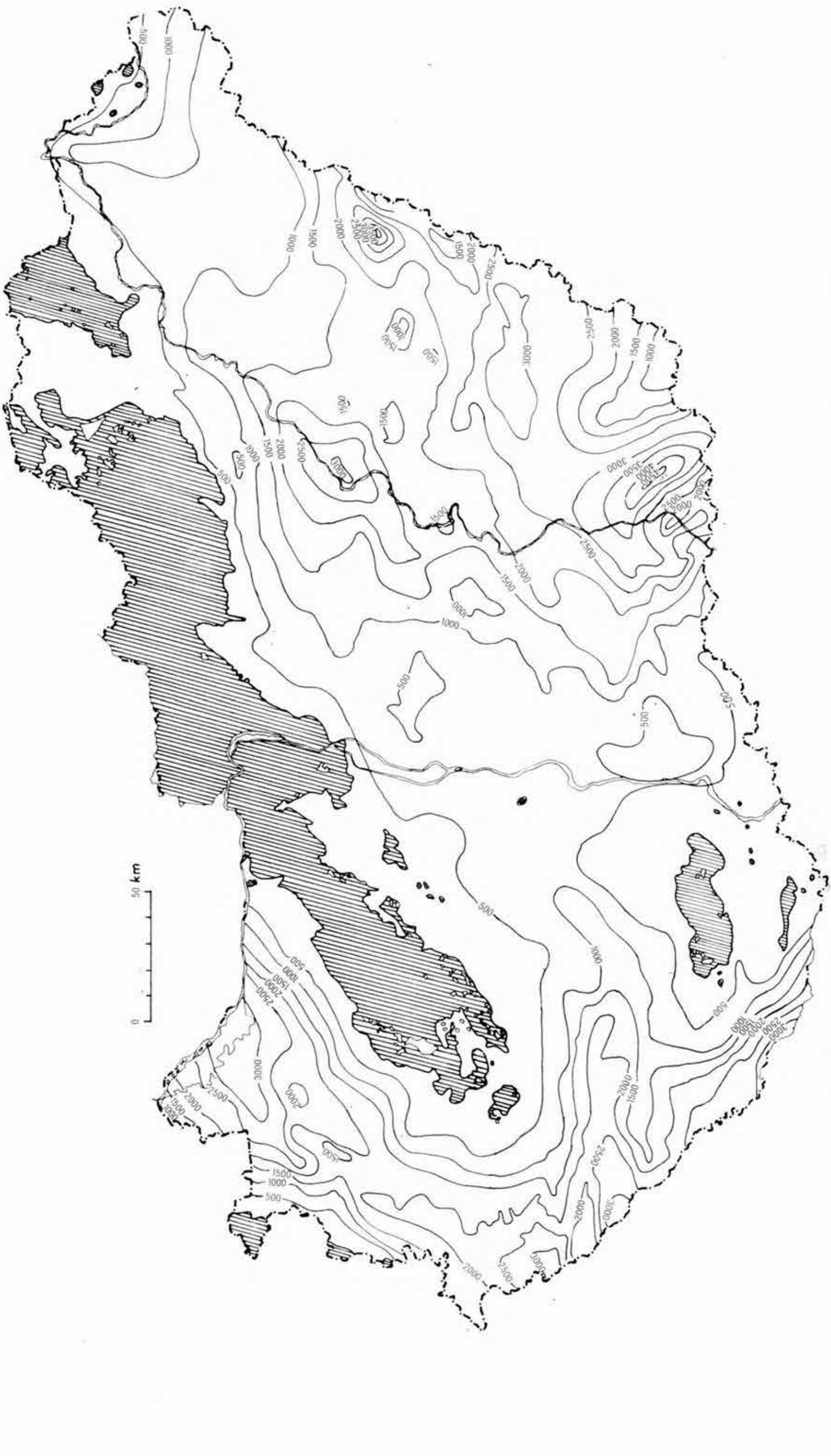



*Fig. 4.* Thickness of Pannonian s.l. deposits in Hungary (compiled by S. LENNER, 1984)

The process of filling-up of the Pannonian basin can be summarized as follows:

*1* A climatic change at the end of the Sarmatian significantly altered the former sedimentation regime which was characterized by pelitic deposition in deep basins and ooidic limestone formation on the bains margins. The humid climate further decreased the salinity of the inland sea and led to the deposition of calcareous pelites. In north-eastern Hungary the volcanism proceeded with a new phase of rhyolite-tuff eruption. Its marks can be found also in the southern and central parts of Transdanubia in forms of thin tuff interbeddings. Diatomite and bentonite deposited in lagoons developed in deeper parts of the volcanic Tokaj Mountains.

*2* The subsidence rate of the Pannonian basin increased due to important tectonic changes in the Early Pannonian. The subsidence was common, but varied in space. The source areas for clastic influx—the Alps and the Carpathians—uplifted.

Large amount of the erosional clastics was transported into the basin by rivers and thus a great deltaic sedimentation regime developed. In northern Hungary the delta-plain facies can be found. In the inner parts of the basin the maximum water depth could have reached 800 to 900 m, according to the seismic profiles (Fig. 5). In northern Transdanubia trachyte volcanism, in the southern Great Hungarian Plain basaltic eruptions took place which produced volcanic masses of a few cubic km.

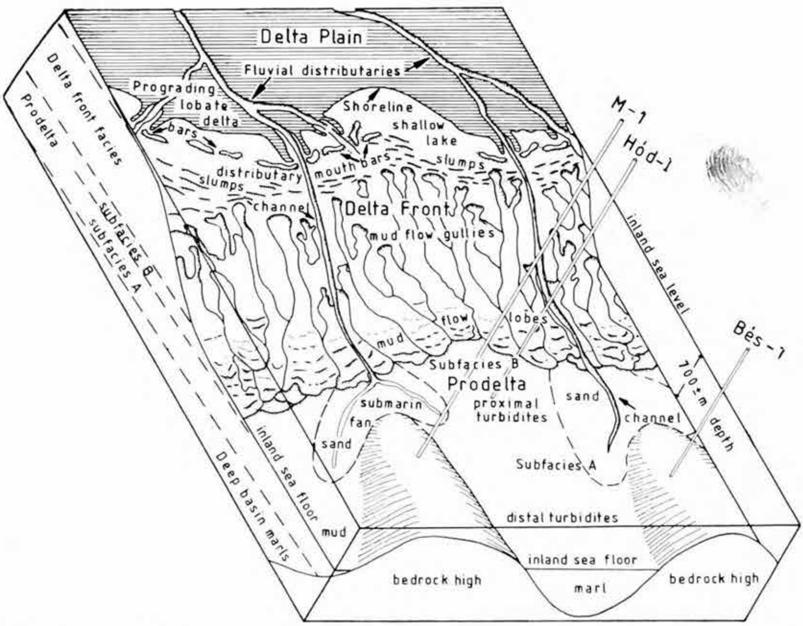



*Fig. 5.* General sedimentation system of the Pannonian s.l. sequence (compiled by I. BÉRCZI)

In the lagoons on the basin margins and in deep depressions pelites were deposited from suspension and mixed with biogenic sediments (diatomite, alginite). This sedimentation regime characterizes the upper two thirds of the Peremarton Group.

*3* The Dunántúl Group was deposited under similar conditions but significantly shallower water (200—400 m). The deposition took place in a delta system formed under different morphotectonic and climatic circumstances. The topography of the source area became more rough producing nerating more and coarser clastics, so the proportion of sand in the sediments increased. The basin got increasingly filled-up.

The delta-slope sediments, which were dominant at the beginning, were progressively replaced first by delta plain deposits and later by lacustrine ones and eventually the basins got completely filled-up. On the delta-plain in the southern and southeastern forelands of the mountains large marshes and Taxodium swamp forests were formed. Lignite seams characterizing the middle and upper parts of the Dunántúl Supergroup developed here.

On the basin margins freshwater limestones were formed in latest Pannonian times in the intramontane basins of the Transdanubian Central Range.

In the middle and upper part of the Dunántúl Supergroup Na-alkaline basaltic tuff and basaltic lava flows occur in the Bakony Mts, in the Little Plain and northern Hungary. In Transdanubia the volcanic activity generally took place in the basins, and the water of the Pannonian lake penetrated into these volcanic craters, and led to the formation of alginite beds. In northern Hungary the volcanism took place on emergent land and the activity continued during the Pleistocene, too.

*4* A marked break in sedimentation can be observed towards the end of Pannonian time. This is the consequence of renewed tectonic activity which led to uplift and erosion in the area of the present mountains and over much of Transdanubia. Over the rest of the country the subsidence continued and mostly terrestrial sediments were deposited.

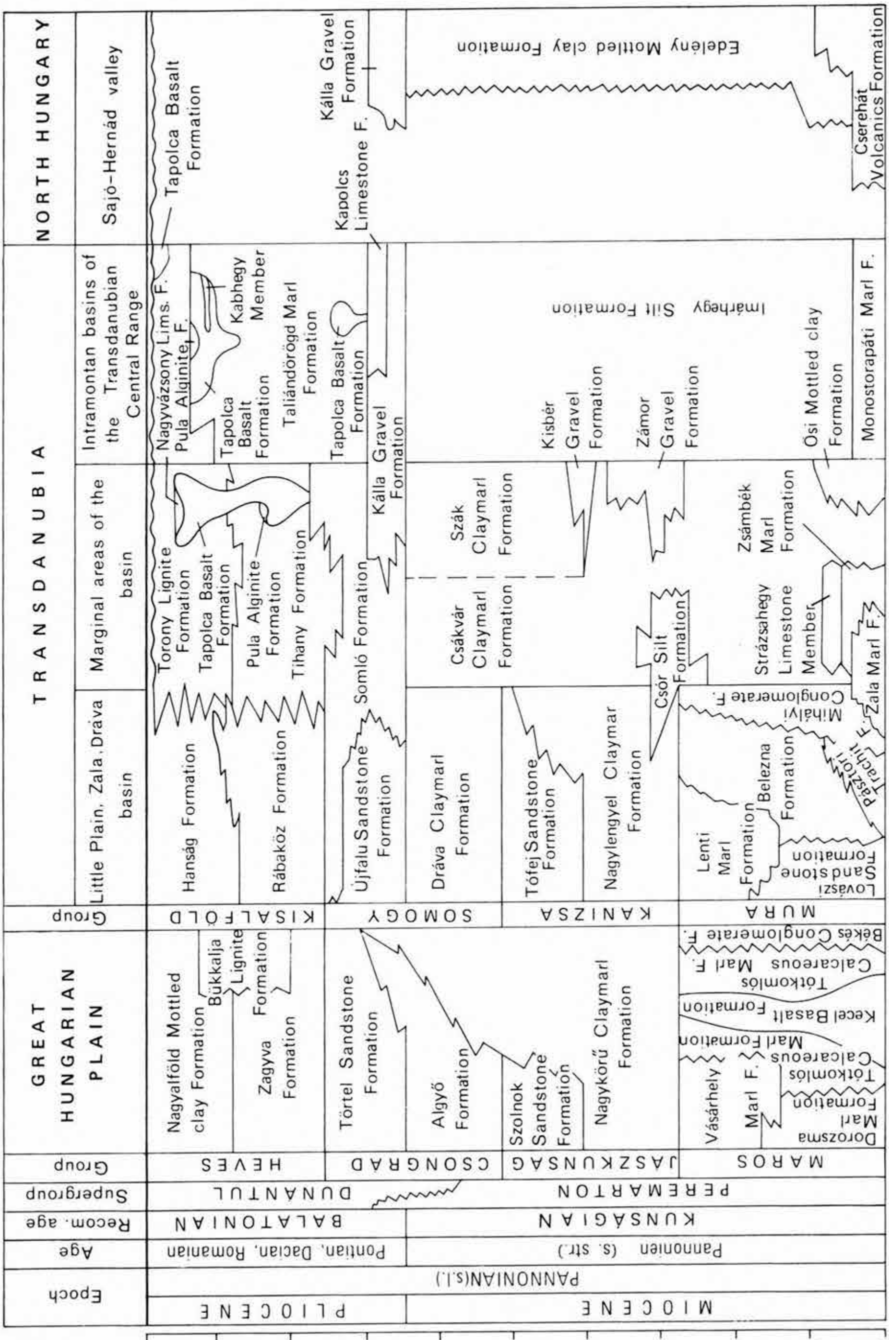
The two delta systems of the Pannonian basin mentioned above were similar from many points of view. Source areas for clastic inflow were in both cases the Alps and the western and eastern Carpathians. The clastic material was transported by rivers into the basins, where large delta systems were formed which prograded from the north and northwest towards the south. The younger delta system usually was superimposed with angular unconformity on the older one, as it can be seen on seismic profiles.

The bulk of the sediments was deposited mainly below the $O_2-H_2S$ boundary and therefore it contains more disseminated organic matter. The large-scale subsidence of the substratum of the basin was caused by the thinning of the crust and the lithosphere. As a result the heat flux in the Pannonian basin increased by about a factor of two compared with stable continental areas. Therefore the Pannonian (s.l.) formation could reach the oil generation window already below 1800 to 2500 m. This led to the generation of a number of minor oil and gas fields.

In Figures 6 and 7 the stratigraphic classifications of the Pannonian formations are shown. From the lithostratigraphic point of view the Pannonian strata can be divided into two main groups. The lower part containing dominantly pelites is called Peremarton Group, having previously been referred to as Lower Pannonian. The upper one is called Dunántúl Group. The lower part of the Peremarton Group consists of marls, calcareous marls, subordinately of conglomerates and sandstones. The upper part is made up of clays and marls with frequent sand and gravel interbeddings.

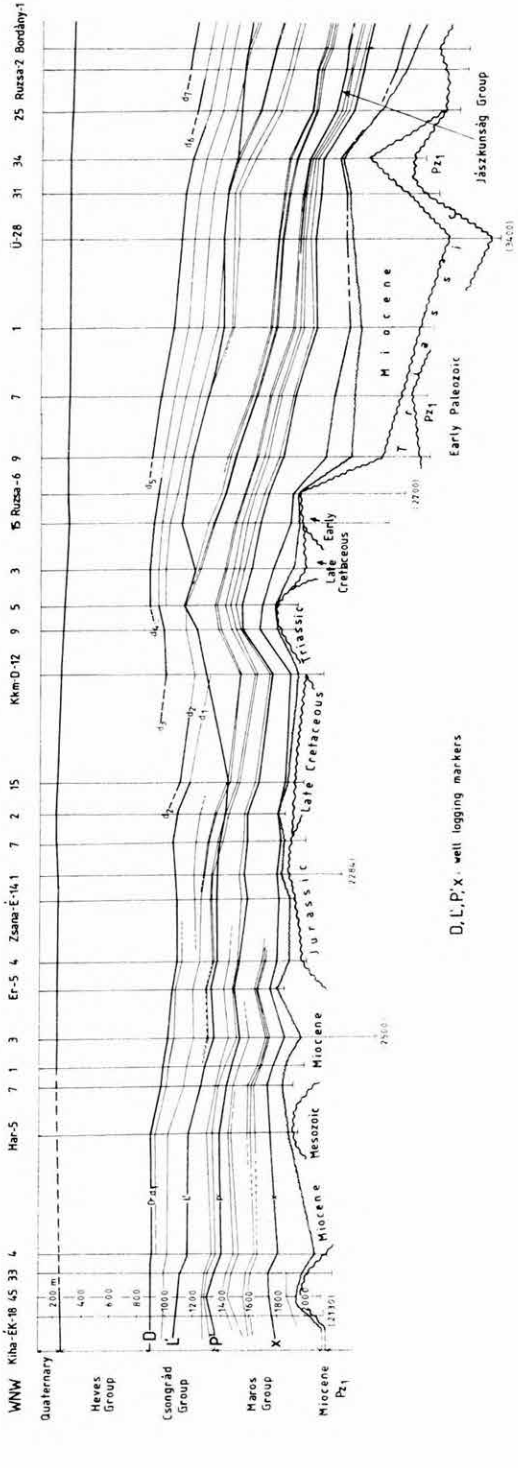
For the biostratigraphic subdivision of the Pannonian (s.l.) formation a rich and well-preserved mollusc fauna has been used since the end of the last century. The special Pannonian-brackish character is expressed in a fauna dominated by Congeria, Limnocardium and Melanopsis species and in the lack of all marine elements.

In the Recent decades the biostratigraphic importance of other fossil groups has been cleared up: vertebrates, ostracods, thecamoebans, diatoms, foraminifers, nannoplankton, spores and pollen grains, ichnofossils and sponge-spicules. All inland sea groups turned out to be of special Pannonian brackish character, i.e. low specific diversity and high density. Our recent biostratigraphic classification possibilities are shown in Fig. 8.




Compiled by Hungarian Stratigraphical Commission Pannonian Working Group 1984

*Fig. 6.* Lithostratigraphical classification of the Pannonian s.l. formations in Hungary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Har-5
7 1 3
Er-5 4
Zsana-É-14 1
7 2 15
Kkm-D-12 9 5 3
15 Ruzsa-6 9
7
1
Ü-28 31 34 25 Ruzsa-2 Bordány-1
Quaternary
Heves Group
Csongrád Group
Maros Group
Miocene
$Pz_1$
D
L'
P'
X
200 m
400
600
800
1000
1200
1400
1600
1800
Miocene
Mesozoic
Miocene
Miocene
Jurassic
Late Cretaceous
Triassic
Late Cretaceous
Early
Early Paleozoic
$Pz_1$
Triassic
Miocene
$Pz_1$
Jászkunság Group
D, L', P', X : well logging markers

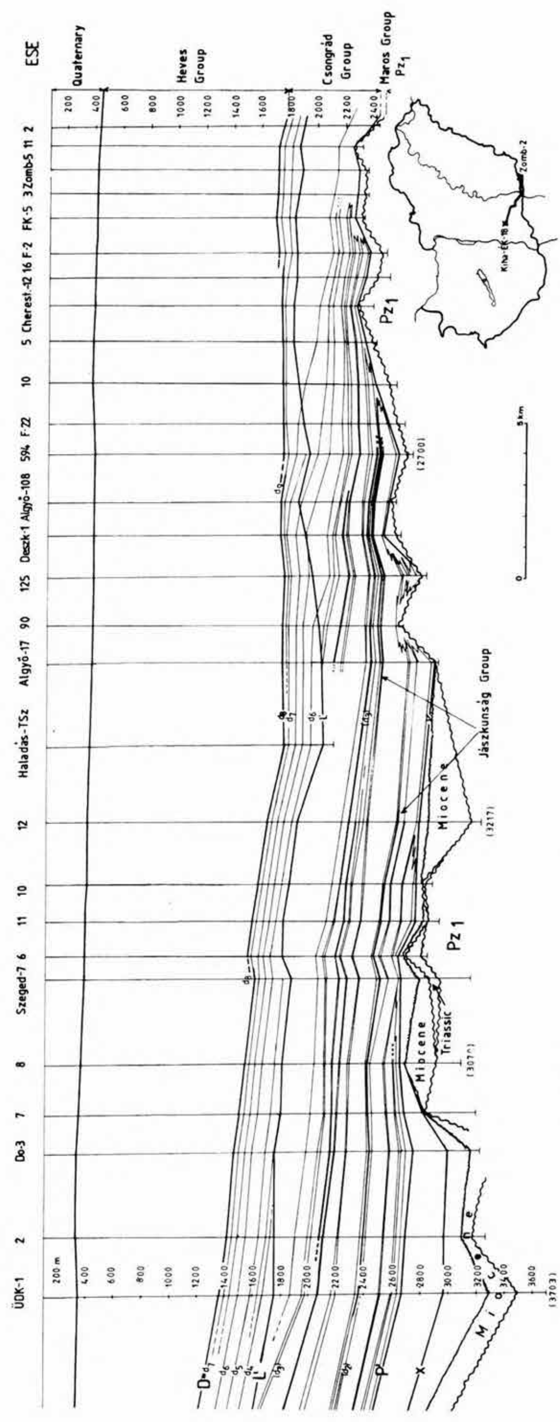


*Fig.* 7. Well-logging profile of the South Hungarian (Pannonian) (I. GAJDOS—S. PAP)

The molluscs, ostracods and the microplankton floras show a similar fourfold subdivision, which can be due to environmental constraints. At the end of the Sarmatian and at the beginning of the Pannonian the salinity of the inland sea varied within extremes of hypersaline and oligohaline due to a Mediterranean-type climate with hot summers and humid winters. It resulted in a strong selective pressure imposed on the biom. Hence the earliest Pannonian low specific diversity of fauna including the inland sea microfauna.

This special earliest Pannonian biome was succeeded by a very diversified Pannonian—brackish inland sea floral and faunal complex, of which ancestors partly had ingressed from the eastern Paratethys and then flourished in the Pannonian basin and partly survived the climatic changes at the end of the Sarmatian. They enriched the flora and fauna of the second level.

The next larger biotic change roughly coincides with the lithostratigraphic boundary between the Peremarton and Dunántúl Groups, when the previous predominantly pelitic sedimentation was succeeded by a balanced pelitic—psammitic depositions. The associated biotic changes unambiguously prove a significant and relatively rapid decrease in the salinity of the inland sea.

The still saline inland sea was essentially filled-up due to the increased rate of erosion. The brackish-water flora und fauna was followed by terrestrial flora and fauna. The distinction between these latter and the Pleistocene elements is therefore impossible. Furthermore, it is to be emphasized that the filling-up of the depression took place at different rates in different local basins.

For the stratigraphic subdivision of the marginal sequences mollusc, ostracods and planktonic microfossils are the best. In the inner parts of the basins the thick successions cannot be easily subdivided, because of the rarity of fossils due to greater water depths and unfavourable preservation conditions. The vertebrate, nannoplankton, foraminiferal, thecamoeban, diatom, spore and pollen remains and the ichnofossils often help the subdivision of some sequences, but general use is hampered by their limited occurrence. In areas of inner basins the Pannonian (s.l.) can be stratigraphically subdivided by means of well-logging markers, up-to-date seismic profiles and by lithological trend analysis.

Three important unconformities can be traced by seismic stratigraphic methods in the Pannonian formations.

*1* The oldest is at the base of the Pannonian and it appears above basement highs which may consist of either pre-Neogene or older Miocene rocks. In the deep depressions (Makó trough, Dráva basin, Little Plain, Derecske trough, Jászság depression), however, the Sarmatian to Pannonian transition was characterized by uninterrupted sedimentation and the unconformity continues here with correlative conformity.

*2* In association with the prograding delta system subaqueous redistribution of deposits occurred which led to local unconformities of rather different age.

*3* The younger unconformity can be observed between the Pannonian (s.l.) and Pleistocene sediments above basement highs. This boundary is, however, conformable in the deep basin areas.

Seismic stratigraphy units can be usually well correlated with other stratigraphic or facies units seismic facies units. The seismic facies unit A corresponds to the basal and deep basin facies, B and C correspond to the prodelta (turbiditic series), $D_1$ corresponds to the prograding delta-front, E corresponds to the deltaplain and lagoonal-facies, while F corresponds to the fluvial—lacustrine and terrestrial facies. This is demonstrated in Fig. 9 by a seismic profile from eastern Hungary.

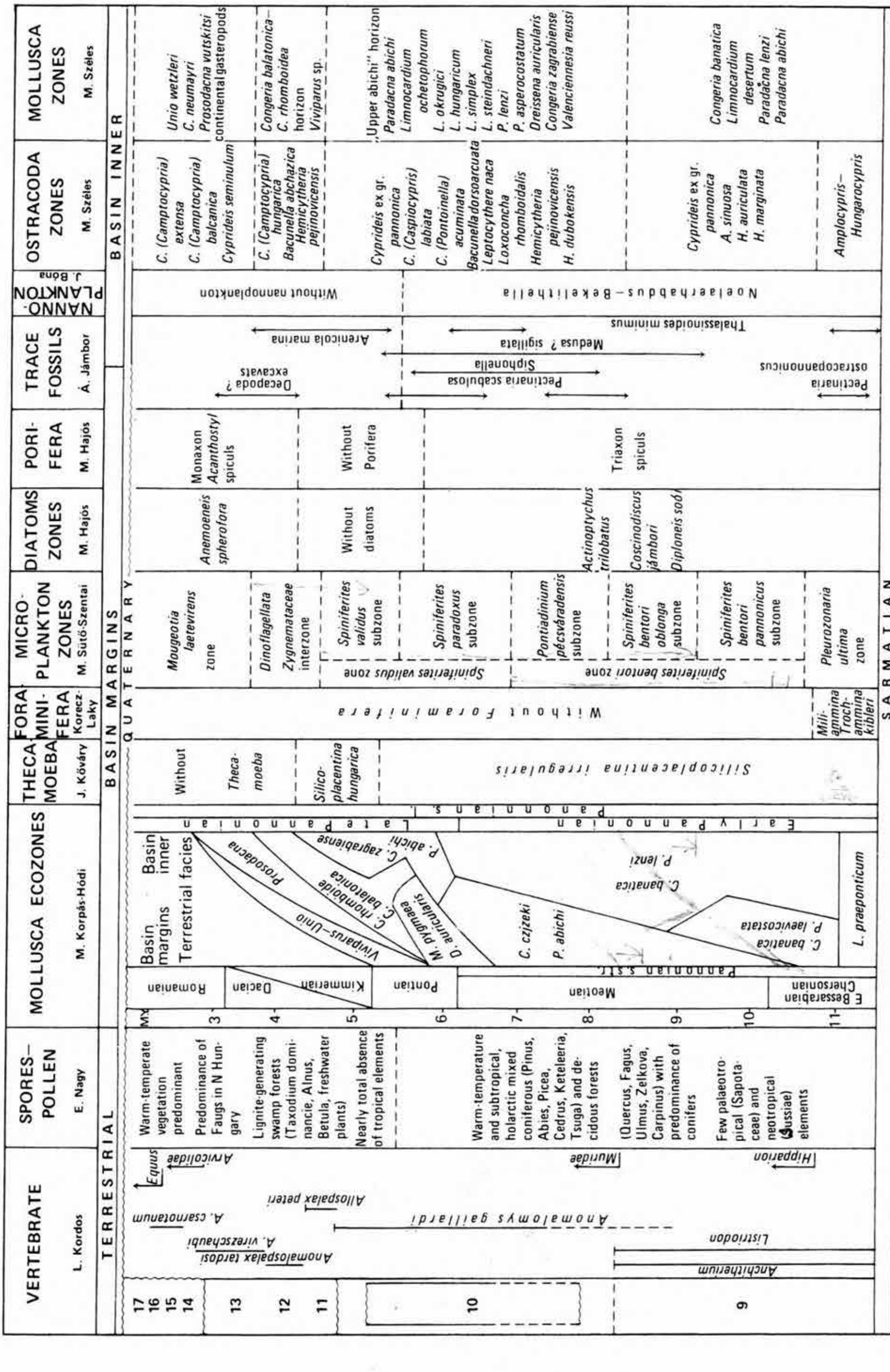


*Fig. 8.* Biostratigraphic subdivision of Pannonian sequences

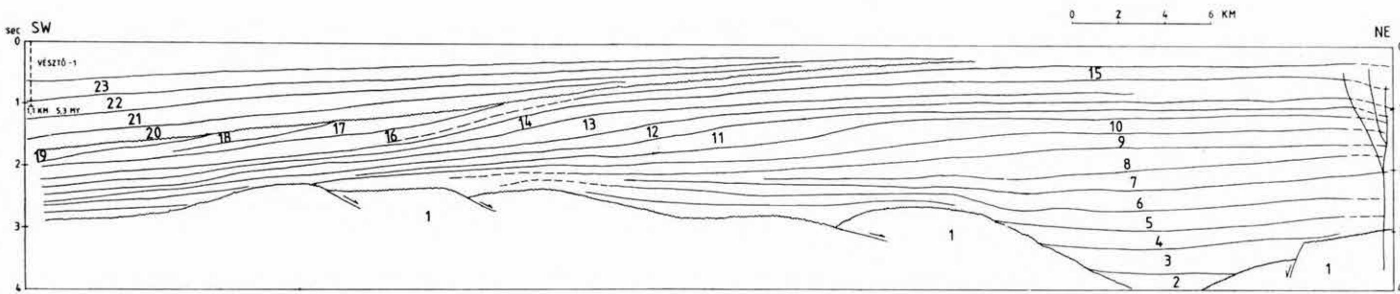


*Fig. 9.* Seismic profile parallel with the strike depression (after Z. BERKES, GY. POGÁCSÁS and B. SZANYI, 1983)
*1* Neogene basement formations, *2—23* Neogene depositional (sub)sequences

The Pannonian (s.l.) formations are tectonically quiet or slightly disturbed. The most characteristic structural forms are compaction anticlines. The length of these anticlines is about 5—10 km, with maximum of 35 km. Their width is about 1—2 km with a maximum of 10 km. In the marginal areas of the anticlines there are often normal faults with a throw of about 5—10 m, rarely 50—100 m. According to the seismic profiles typical structures are listric faults due to differential compaction.

In the southern and northern parts of Mecsek mountain overthrusts can be observed which involve the layers of the Dunántúl Supergroup. The most important fault zone crossing the Pannonian formations can be traced for several tens of kilometers in west—southwest direction in eastern Hungary. Dominantly strike-slip movement took place along this fault zone. Folded structures are probably in southwestern Hungary, where the "Sava folds" extend to Hungary from Yugoslavia.

The Pannonian formations are one of the most important complexes of the country from the point of view of mineral exploration. 60% of the gas and oil fields can be found in this formation. The lignite, drinking- and thermal water resources are also important. From the non-metallic minerals the following are the most important: quartzite sand, bentonite, kaoline, diatomite and building materials (clay, sand, gravel).

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# THE PONTIAN OF THE EASTERN PARATETHYS: ITS DURATION AND POSITION IN THE MAGNETOCHRONOLOGICAL SCALE

by

M. A. PEVZNER

For a long time Soviet geologists considered the Pontian duration to be about 2—2.5 Ma. In 1979 V. N. SEMENENKO and M. A. PEVZNER (1979) showed that the duration of the Pontian had been not more than 0.8 Ma. At present Soviet geologists do accept the "short" Pontian, but foreign scientists still keep to the "long" Pontian.

The duration of the Pontian has been estimated from 0.6 to 4.5 Ma (Fig. 1). The position of the Pontian in the magnetochronological and stratigraphic scales has also been discussed. The Pontian is situated either in the Upper Miocene or at the Miocene/Pliocene boundary, or in the Lower Pliocene. Different scientists place the Pontian within the time interval from 10 to 4.5 Ma in the chronological scale.

It should be noted that the conclusion concerning the "short" Pontian was drawn on the basis of paleomagnetic data. The Pontian deposits are characterized by reverse magnetization in all studied sections (Fig. 2).

The Pontian deposits were studied in eastern Georgia in the Atap section, on the Taman peninsula in the Zhelezny Rog section, on the Kerch peninsula in borehole 15 (PEVZNER, CHIKOVANI, 1978; SEMENENKO, PEVZNER, 1979) and in Romania in the Berka section (TRUBIKHIN et al., 1984).

The boundary between the normal and reverse zones coincides with the lower boundary of Pontian in the Atap section and in borehole 15. In two other sections the base of the Pontian lies slightly below the inversion. Thus, the Pontian may only correspond to one epoch of reverse polarity or to part of it.

In the considered interval of the magnetochronological scale (from 10 to 4.5 Ma) the Pontian can occupy one of three clearly fixed positions. It can be correlated either with epoch 8 or with epoch 6, or still with the beginning of the Gilbert epoch—between its lower boundary and the Thvera event.

In every correlation just mentioned the duration of the Pontian does not exceed 0.8 Ma, because epoch 8, epoch 6 and the lower part of Gilbert epoch have durations shorter than 0.8 Ma.

This lead us to the conclusion that stratigraphic scales in which the Pontian lasts over 0.8 Ma are not correct.

The nannoplankton finds are of great importance for the determination of the position of the Pontian within the magnetochronological scale. According to V. N. SEMENENKO and S. A. LULIEVA (1978) the nannoplankton was found in the Maeotian and Kimmerian deposits in normal stratigraphic succession. This fact excludes its redeposition.

The occurrence of *Discoaster quinqueramus* in the lower part of the Kimmerian in borehole 15 does not permit to place the Pontian in the Gilbert epoch, since the last occurrence of this form is within the upper part of epoch 5. Accordingly, that the Pontian must be placed into the Upper Miocene, but not in the Pliocene or at the Miocene/Pliocene boundary.

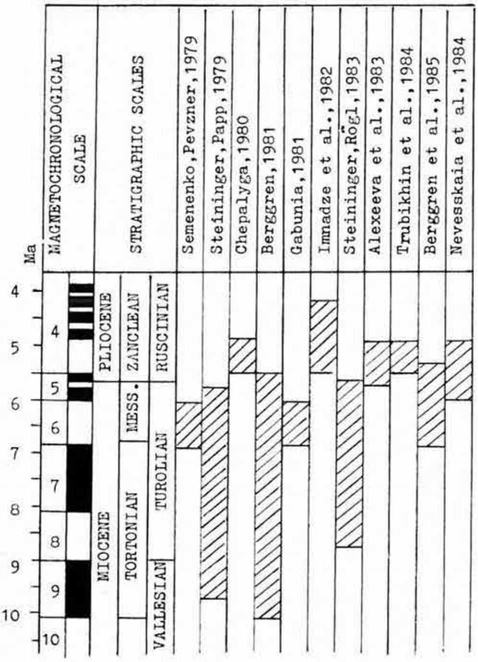


*Fig. 1.* The position of the Pontian in the magnetochronological and stratigraphic scales according to different authors

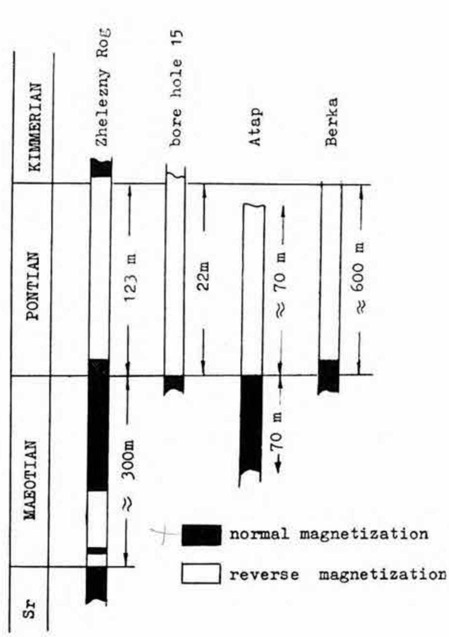


*Fig. 2.* Paleomagnetic characteristic of the sections

The data on mammals confirm this conclusion. Remains of mammals related with the Pontian deposits are known from a number of localities such as Aldamirovce, Gabor, Chrabarsko in Bulgaria (KOJUMDGIEVA et al., 1984), Eichkogel in Austria, Hatvan, Pestszentlőrinc and Rózsaszentmárton in Hungary (PEVZNER, VANGENGEIM, 1985), Mamai and Odessa in the Ukraina (GABUNIA, 1981). All these localities belong to the Turolian.

The Turolian corresponds to the Magnetic epochs 8, 7, 6 and to the part of epoch 5. The lower boundary of the Turolian coincides with the lower boundary of the epoch 8 and in the stratigraphic scale of the eastern Paratethys Neogene it coincides with the Sarmatian/Maeotian boundary. Its age is estimated at about 9 Ma. The Turolian upper boundary is drawn within the upper part of epoch 5 near the reverse event of this epoch and coincides with the Miocene/Pliocene boundary in the oceanic scale. Its age is estimated at about 5, 6, Ma (PEVZNER, VANGENGEIM, 1984).

Consequently, the Turolian mammal fauna of the Pontian does not permit to place the Pontian in the Gilbert epoch.

Both the Turolian mammal fauna and the reverse magnetization of the Pontian deposits allow to correlate the Pontian either with epoch 8 or epoch 6.

The evidence of mammal fauna makes impossible the comparison of the Pontian with epoch 8. If the Pontian is assumed to correspond to epoch 8, its lower boundary is to coincide with the lower boundary of Turolian. In this case there is no place left for the Maeotian the fauna of which also belongs to the Turolian.

Thus, both paleomagnetic and paleontological data make clear that the Pontian can be correlated only with Magnetic Polarity epoch 6 and with the lower half of the Messinian.

The age of the lower boundary of Pontian is close to that age of epoch 6 (6.8 Ma). The upper boundary of Pontian is slightly older than that of the epoch 6 (6 Ma) because the lowermost part of the Kimmerian belongs to the same epoch too.

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# A CASE STUDY OF THE STRATIGRAPHIC SUBDIVISION OF AR-RAJMAH FORMATION AND ITS IMPLICATION ON THE MIOCENE OF NORTHERN LIBYA

by

A. S. EL-HAWAT and M. J. SALEM

*Introduction.* Like most shallow neritic Neogene sequence the Ar-Rajmah Formation, Miocene, of Al Jabal al Akhdar (Fig. 1) covers an extensive area and exhibits rapid lateral and vertical facies changes, which may lack, in parts, time diagnostic fossils. Further complications are often caused by changes in syndepositional paleogeographies, sea level and syndepositional tectonism. When the area was divided and mapped by different geologic teams, lacking a unified regional approach the result was a series of controversial stratigraphic subdivisions, complicated nomenclature (Table 1) and mismatched geologic maps (KLEN, 1974; FRANCIS and ISSAWI, 1977; MAZHAR and ISSAWI, 1977; MEGERISI and MAMGAIN, 1979).

To resolve the stratigraphic controversy of the Miocene sequence a field sedimentological approach based on an actualistic method of interpretation as outlined in WALTHER's law of correlation of facies is used (WALTHER, 1893–94; see also MIDDLETON, 1973). Field measurements of stratigraphic sections, recognition of sedimentary facies and establishment of vertical facies profiles have led to the construction of a facies cross-section (Fig. 2) and the establishment of vertical and lateral facies relationships, geometries and palaeogeography. The nature of contacts between stratigraphic units or facies were recognized through careful field observations that led to the discovery of subtle but significant disconformities. These are used in the present study as a basis for correlation and stratigraphic subdivision. The same approach has been used successfully in settling a sedimentological controversy of the Miocene in Sirte basin, central Libya (EL-HAWAT, 1975; 1980a).

This paper aims to present a case study for the use of the sedimentological approach based on the recognition of depositional facies and events as a means towards the establishment of an actualistic stratigraphic order for similar depositional sequences.

## Sedimentary facies and cycles

Three facies associations are found to form a single shoaling-up depositional cycle of Ar-Rajmah Formation. These are, from the top: A. Open shelf; B. Shoal; and C. Restricted lagoonal.

*A. The open shelf facies association*

This association consists of several shallowing-up cycles, consisting of yellowish marl, algal-skeletal wackestone, and coarse skeletal whole shell packstone, that may change upwards into lenticular massive skeletal mudstone and oyster banks containing solitary corals. Both lithology and fauna indicate sedimentation under shallow marine shelf conditions. Similarly, oyster banks and lenticular coarse skeletal bodies are known to exist over the bathymetric highs of the modern shelf of the Arabian Gulf (PURSER, 1973).

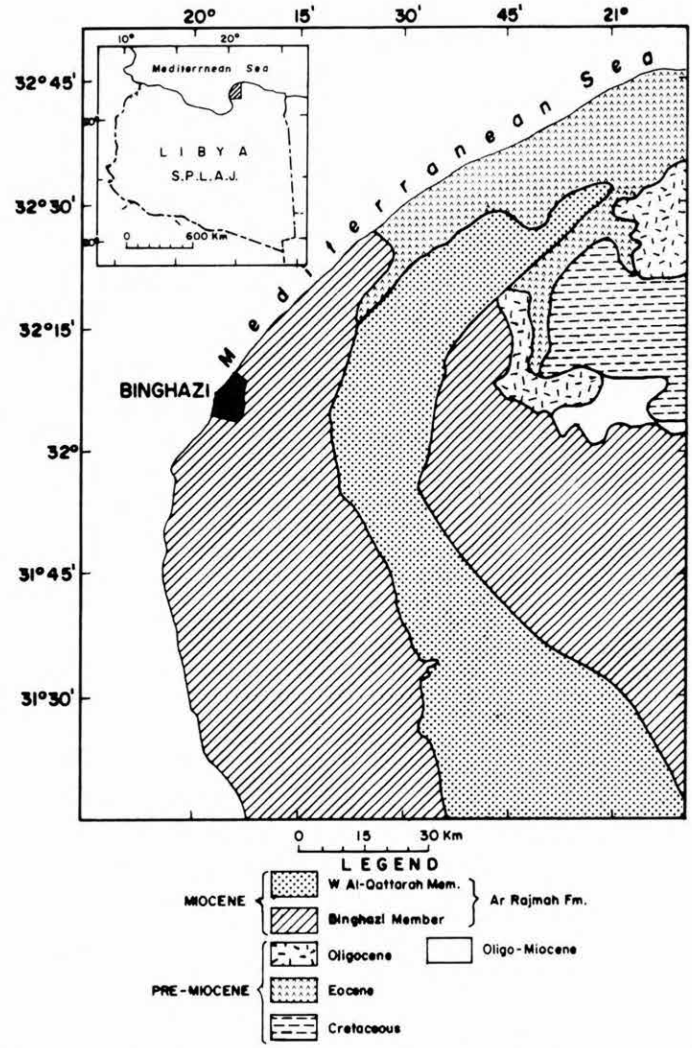
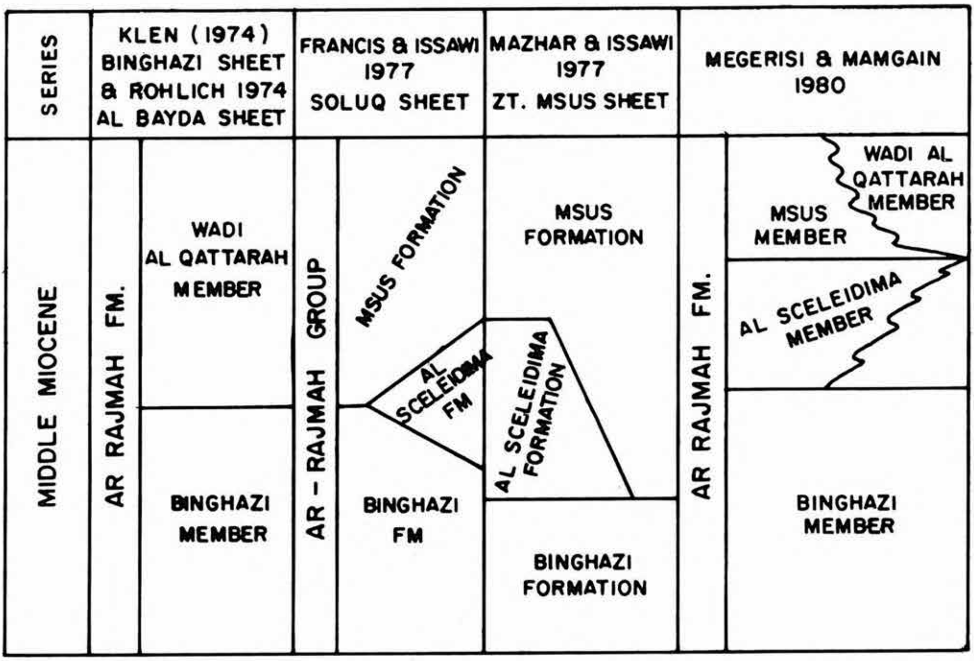



*Fig. 1.* Generalized geologic map of Al Jabal al Akhdar, NE Libya. The distribution of the Miocene is based on the stratigraphic subdivision proposed in this paper. Map compiled and modified after KLEN (1974), ROHLICH (1974), FRANCIS and ISSAWI (1977), MAZHAR and ISSAWI (1977)

East of the area and presently at a higher elevation this facies association changes into skeletal mudstone and baffelstone (EMBRY and KLOVAN, 1971) of branching *Porites* corals buried in foraminiferal grainstone (Fig. 3). *Porites* reefs are developed into massive patch reefal bodies embedded in skeletal packstone representing the back-reef facies. *Porites* reefal complexes were developed on the margin of Al Jabal al Akhdar high.

*Table 1*

**Stratigraphic correlation chart of the Miocene, Al Jabal al Akhdar, NE Libya. Note complications of stratigraphic subdivisions and nomenclature. Also, all earlier workers agree on the Middle Miocene age for the entire sequence**




*B. The shoal facies association*

This association occurs above the open marine facies and consists of cross-bedded skeletal, oolitic, pelletal and red algal-coated grainstone, packstone, and bioturbated skeletal wackestone. These facies are arranged in at least two successive shallowing-up cycles that form a linear geometry exposed in a trend parallel with the palaeostrike. This facies changes upwards and laterally towards the east into lagoonal facies associations. Palaeocurrent azimuthal distribution of the cross-bedded facies points to a net eastwards transport. These facies are interpreted as having been deposited as sand shoals, in channels, tidal deltas and spillover lobes advancing eastwards into a shelf lagoonal setting. The shoal facies complex sequence is shallowing upwards; their petrography and diagenesis suggest progressive emergence and subareal exposure (El-Hawat, 1986).

*C. The restricted lagoonal association*

The restricted lagoonal facies association forms the top part of Ar-Rajmah Formation and consists of several successive cycles. A type cycle consists of

i) cross-bedded lithoclastic, oncolitic, oolitic grainstone and packstone;

ii) algal stromatolites;

iii) lime mudstone and terminates in coarsely crystalline gypsum. These facies represent an evaporite salina complex, that was successively subjected to flooding by sea water and drying up (El-Hawat, 1980b). They are analogous to modern salinas described from Lake Marion, South Australia (Hardie and Eugster, 1975).

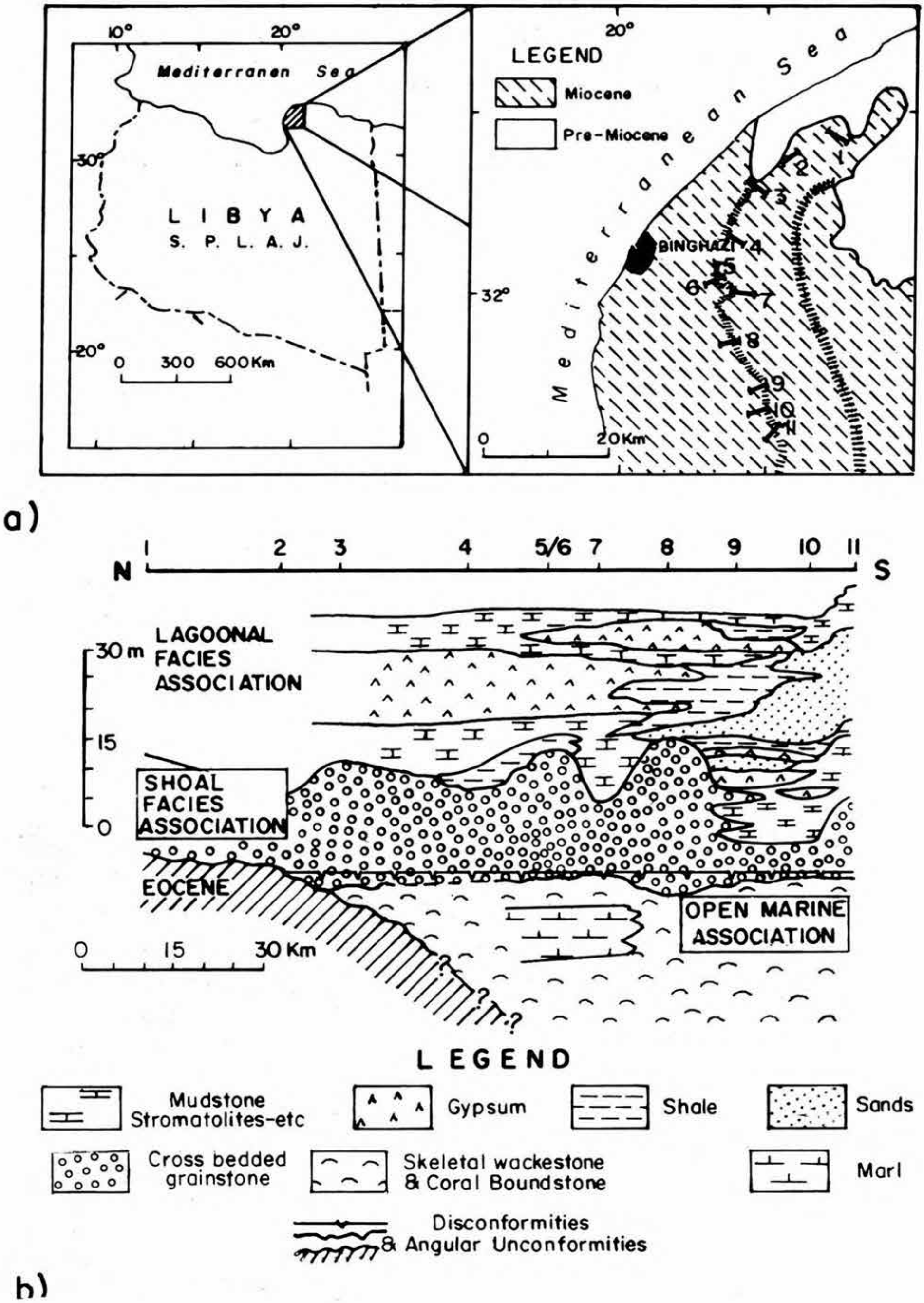



*Fig. 2.* (a) Measured sections location map. (b) N—S facies cross section of Ar-Rajmah Formation, Al Jabal al Akhdar, NE Libya

*Fig. 3.* Binghazi Member. Reefal facies with upwards growing, branching *Porites*, now dissolved and moldic cavities are filled by recent terra rossa. White areas consist of foraminiferal grainstone

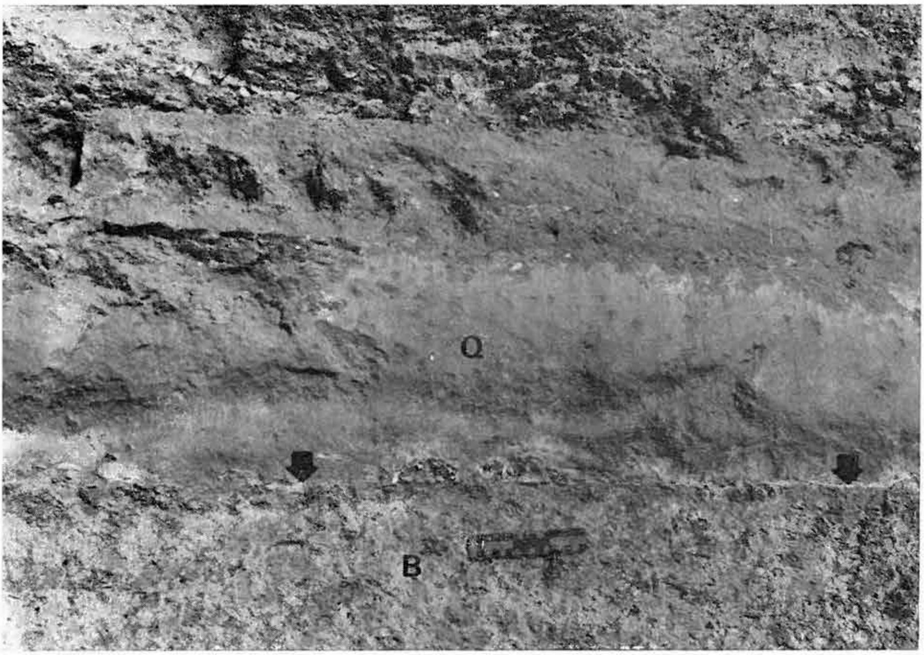



*Fig. 4.* Massive coarse skeletal wackestone facies of Binghazi Member (B) overlain by shoal facies association of Wadi al Qattarah Member (Q). The disconformity boundary (arrows) is occupied by *Echinolampas* sp. in living position. Scale = 15 cm

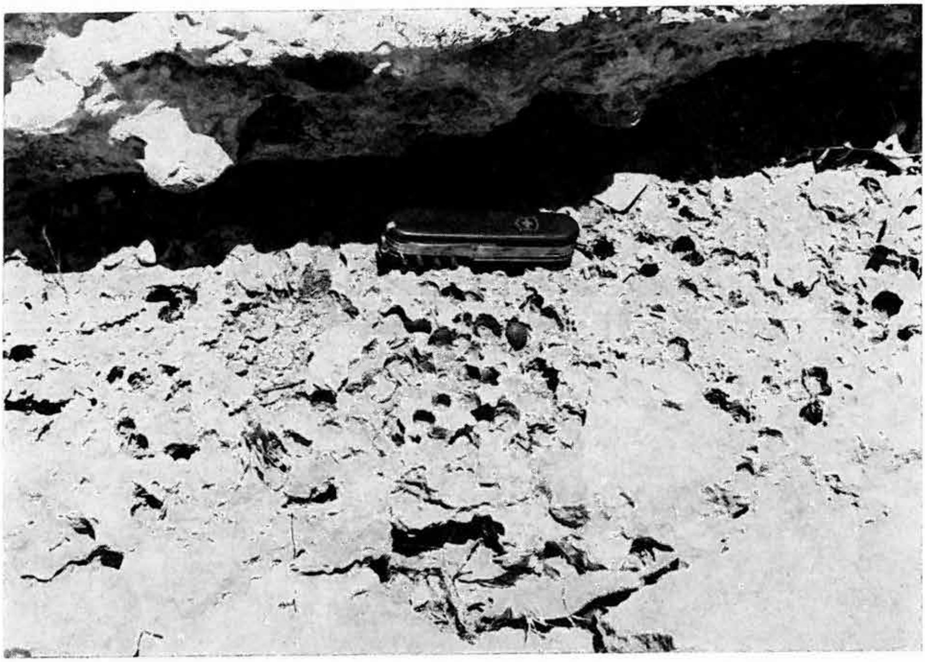


*Fig. 5.* Plane view of the second disconformity surface showing bivalve borings. Knife = 9 cm length


*Fig. 6.* Angular unconformity between the Eocene (E) and the shoal facies association of Wadi al Qattarah Member (Q). Note erosional boundary (arrows). Large divisions of the scale = 10 cm

The carbonate—evaporite facies cycles are found north of the area pass to the south into siliclastic-evaporite facies associations (Fig. 2). This is attributed to deltaic influence from the structurally low areas of Sirte basin.

### Boundaries, key horizons and events

The boundary between the open shelf and shoal facies associations is marked by two prominent disconformities that are traceable throughout the study area (Figs. 4, 5). The lower disconformity is distinctively sharp and erosional; its surface is bored by bivalves and sponges and encrusted with oysters, Bryozoa, serpulids and Foraminifera. In places open burrows, enlarged by solution, form a cavity system whose walls exhibit borings and are partly to completely filled by coarse skeletal sediments derived from the overlying shoal facies. Immediately above this surface large numbers of echinoids are to be found in living position.

A second disconformity is found above the base of the shoal facies, a few decimeters above the first disconformity. It is also extends throughout the area and has been used as a key horizon for the construction of the facies cross-section. This hard ground exhibits normal and inverse bivalve borings, as well as incrustations by various epifauna. Northwards, these hard grounds rest against the Eocene—Miocene unconformity surface. In this area the open marine shelf facies are missing due to non-deposition whereas the shoal facies rest over the Eocene. The unconformity is an irregular, weathered, karstified surface with solution pipes filled by terra rossa and Miocene material (Fig . 6.).

### Palaeogeographic reconstruction

From the field-established vertical and lateral facies relationships in relation to the regional disconformity surfaces it is possible to construct a series of palaeogeographic maps of Ar-Rajmah Formation (Figs. 7, 8, 9). The palaeogeographic map of the open shelf association (Fig. 7) exhibits the occurrence of a broad shallow marine shelf dominating the area, with possible local highs which coarser bioclastic sediments have accumulated. Eastwards these facies change into *Porites* reefal complexes marking the margin of Al Jabal al Akhdar high.

The second palaeogeographic map (Fig. 8) suggests the development of shoal-channel shelf lagoonal complex. Two palaeographic features are prominent in this map. First, the development of an elevated escarpment to the east, representing the *Porites* reefal wall of the earlier open marine shelf association. Second, the transgression of the shoal association further to the north. The first feature is associated with the lower disconformity, indicating a drop of the sea level, while the transgressive feature is associated with the development of the second hard ground.

Sedimentation of Ar-Rajmah Formation was terminated by a major drop in the sea level. It has led to the development of a beach-dune complex separating the salina complex from the open marine (Fig. 9). Flooding of the salina by sea water was achieved either by storms or through the porous barrier by ground water.

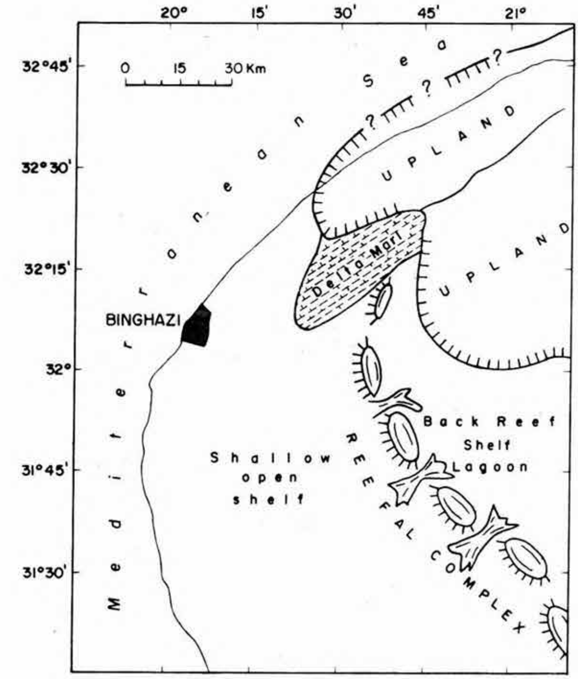



*Fig. 7.* Stage I—Palaeogeography during Langhian—Serravallian shallow shelf reefal complex development

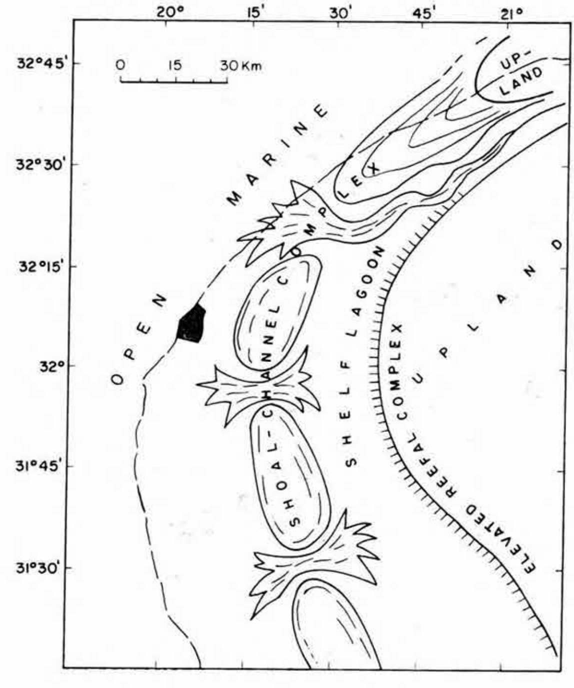



*Fig. 8.* Stage II—Palaeogeography during Tortonian time.
Development of the shoal complex

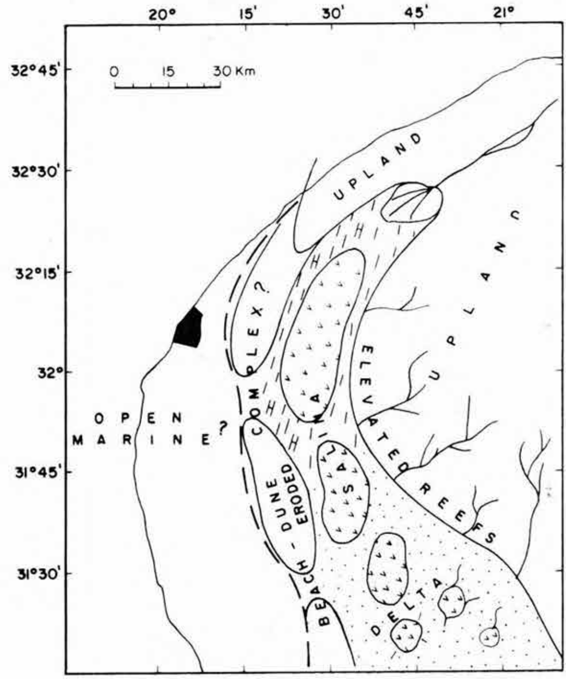



*Fig. 9.* Stage III—Palaeogeography during Messinian time. Development of salinas

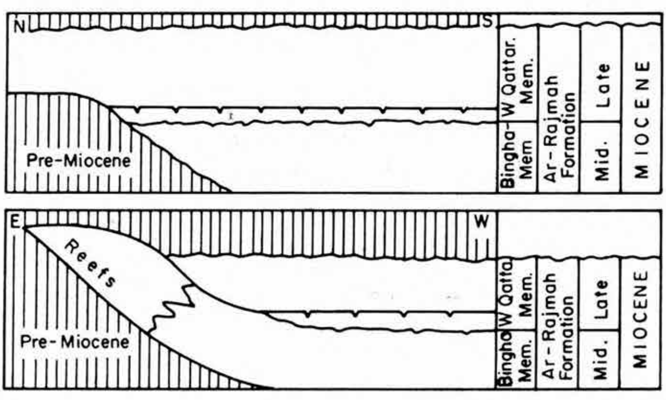
**Discussion**

## *Construction of a stratigraphic chart*

The sedimentologic study of Ar-Rajmah Formation has led to the establishment of depositional facies, cycles, events and construction of palaeogeographies at different relative time frames during the Miocene of Al Jabal al Akhdar. From these it is therefore possible to construct an actualistic stratigraphic subdivision of the Miocene (Table 2). Because of the methods used we do not regard the introduction of new names as an objective in itself, so the previously publised and simple nomenclature presented by KLEN (1977) is used thus avoiding unnecessary new names.

*Table 2*

**Stratigraphic chart of Ar-Rajmah Formation based on the field sedimentological approach**



The lower regressive disconformity is used here as a base for stratigraphic subdivision; it records the regressive event at the end of the Middle Miocene (VAIL et al., 1977). The second disconformity (hard ground) and the Eocene–Miocene unconformity, on the other hand, document the early Late Miocene transgression. It is reasonable to conclude, therefore, that the evaporite salina development is indicative of the Messinian drop in the Mediterranean sea level.

*Regional considerations*

Using the same approach it was possible to correlate the Miocene sequence throughout northern Libya (EL-HAWAT et al., 1985). To the south-west of Al Jabal al Akhdar, in Sirte basin, the top of the Middle Miocene sequence of Marada Formation is marked by a regional disconformity surface which was used as a base for correlation and construction of facies cross-sections (EL-HAWAT, 1980a; Figs. 10, 11). Northwards, in Sirte basin, as in Ar-Rajmah Formation, a regional disconformity surface is found to separate the Middle Miocene open shelf reefal facies associations (Langhian–Serravallian) from the Late Miocene (Tortonian–Messinian) shoals and evaporitic restricted lagoonal associations (DE HEINZELIN and EL-ARNAUTI, 1983; INNOCENTI and PERTUSATI, 1984; GIGLIA, 1984).

Finally, the case study of Ar-Rajmah Formation and the Miocene of northern Libya indicates that the use of an actualistic field approach based on sedimentology, as presented in this paper, is an effective method for the establishment of logical and simplified stratigraphic subdivisions of sedimentary sequences.

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# THE PLIOCENE/PLEISTOCENE BOUNDARY FROM THE POINT OF VIEW OF LATE CENOZOIC GEOHISTORICAL SCALE

by

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*Introduction.* The present-day state of art in Cenozoic stratigraphy is quite contradictory. On the one hand, there is an established system of traditional regional stages for different basins and countries of the world, which is a customary one and well suited for the purposes of mapping, however, enabling no minute interregional correlation. The General Scale of Cenozoic is, at present, in fact, limited to 6 series. On the other hand, in the past 10 to 15 years, an enormous amount of data was accumulated on nontraditional means of detailed stratigraphic subdivision and a high-resolution stratigraphic correlation of Cenozoic deposits by means of micropaleontological zones, dating levels, polarity zones, isotopic and climatic trends and chronometric data. Therefore, it is even now possible to outline the global high-precision and minute geochronological scale of the Cenozoic, differing from traditional Phanerozoic scales.

Practice calls for the quickest settling of this contradiction. However, there is as yet no single strategy two approaches and two basic philosophies exist.

The first approach, formulated by Hedberg and presented in the Stratigraphic Guide (1976), may be defined as a chronostratigraphic one. Hedberg places major emphasis on the choice of stratotypes of stratigraphic boundaries, which are eventually defined by agreement. All regional stratigraphic units should be correlated with the boundary stratotypes by "any means". Drawing of a new Pliocene/Pleistocene boundary with the stratotype in the Vrica section is suggested as an example, demonstrating a successful application of this pragmatic approach.

The second approach, proceeding from the experience of European stratigraphers and presented in the papers by LIBROVICH, MENNER, SCHINDEWOLF, ERBEN, WALLISER etc as well as the ISC documents, places major emphasis on searching and fixation of the natural geohistorical limits, which may be followed interregionally by traces of ecological reconstructions. This approach may be defined as signal or "event" stratigraphy. For drawing valid boundaries of global stratigraphic units within the framework of this approach, not only one "golden spike" is needed, which could be easily put in a wrong place, but a reliable synchronization of a possibly greater number of reference sections in different facies zones, different basins and at different latitudes. The authors of the present paper, who are in favour of the second approach, consider the drawing of the Pliocene/Pleistocene boundary with a stratotype in the Vrica section as a vivid example of an erroneous stratigraphic solution.

**Historical note**

The problem of the Pliocene/Pleistocene boundary and a closely related, but not identical problem of the Neogene/Quaternary boundary came into being at the very outset of stratigraphy. In 1829, DESNOYERS proposed to distinguish the Mastodon Formation in the Paris basin, corresponding to the contemporary Pliocene and a part of the Upper Miocene, as a Quanternary system. In 1830—33, LYELL distinguished Pliocene and post-Pliocene; and in 1839, he subdivided the post-Pliocene into "old" and "young" or Pleistocene. FORBES in 1946 used, without any good grounds, the term "Quaternary system" as a synonym of the new Pliocene and post-Pliocene, which gave rise to many misunderstandings and discussions. He also made a justified conclusion that the sense of distinguishing Pleistocene lies in the fact that its lower limit corresponds to the latest reconstruction of the latest reconstruction of the organic kingdom and that this subdivision is of climatostratigraphic nature. GIGNOUX (1910) made the basal boundary of Pleistocene coincident with the base of the Sicilian regional stage; and A. P. PAVLOV, with the base of the Baku and Chaudine regional stages of Paratethys.

On ZEINER's initiative, the IGC Session, held in London in 1948, passed a recommendation on lowering the Pliocene /Pleistocene boundary under the base of Calabrian and Villafranchian and on the choice of stratotype of this boundary in the marine section of southern Italy. However, in 1952, ZEINER himself acknowledged that this recommendation was erroneous, and SELLI showed that the base of Villfranchian is older than that of Calabrian. Later RUGGIERI and SPROVIERI (1977) proved that the Calabrian beds are a later synonym of the Sicilian beds. Thus, even the early history of this problem indicates that IGC recommendations (that of 1948 and a more precise one of 1952) were based on erroneous premises.

Nevertheless, an idea about drawing the Pliocene/Pleistocene boundary, and more exactly, the Neogene/Quaternary boundary at a lower level appeared to be highly vital owing to an apparent discrepancy between the Quaternary "system", reduced to the range of Pleistocene, and other Phanerozoic systems. These arguments were presented in a most distinct way by HAUG, and later by V. I. GROMOV (1952), who proposed to draw the boundary of the Quaternary system at a lower level, below Akchaghylian and even Pliocene. Overall, no less that 6 versions of lowering the basal boundary of Quaternary—Pleistocene have been proposed.

*The proposed versions of the lower boundaries of Quaternary system*
*(age in Ma according to the present-day estimates)*

*1* Under the traditional boundary of Pleistocene at the base of the Sicilian regional stage of the Mediterranean (GIGNOUX, 1910) and its equivalents in the Ponto-Caspian (= Chaudine—Baku) on continents, at the top of Villafranchian and at the base of the till of the first sheet glaciation in Europe (FORBES, PENK, PAVLOV, GORETSKY, MARKOV, MENNER, NEVESSKAYA etc), 1.0 to 1.15 Ma.

*2* At the base of Apsheronian (YAKOVLEV, 1956) and Santerno (NIKIFOROVA et al., 1982) —sapropelic bed "e" in the Vrica section (AGUIRA, PASINI, 1985)—base of Olduvai, 1.64 to 1.87 Ma.

*3* At the base of Amstelian and Pretiglian in Netherlands (ZAGWIJN, 1974)—Neogloboquadrina atlantica Zone in the Mediterranean (DRIEVER, 1984)—Wucheng Loess in China (LIU TUNG-SHENG et al., 1985), 2.3 to 2.5 Ma.

*4* At the base of Villafranchian (HAUG, 1930)—Akchaghylian (GROMOV et al., 1961)—the replacement of alluvial accumulation by lacustrine in the Pannonian Basin (KRETZOI, PÉCSI, 1982; RÓNAI, 1982), 3.2 to 3.5 Ma.

*5* At the base of Pliocene—Pontian—beginning of glaciation of the Northern Hemisphere (Nikolaev, 1950; Gromov, 1950; Yakhimovich, 1960; Zubakov, 1977), 6.6 to 7.4 Ma.

*6* By appearance of the Mastodon—Hipparion fauna (Desnoyers, 1829)—base of Sarmatian (Nalivkin)—Serravalian—beginning of the sheet glaciation in the Antarctic, 12 to 14 Ma.

The latest version (2), elaborated by the Working Group on IGCP Project 41, will be discussed in detail below. It should be noted that the 38-year—old history of this problem testifies, firstly, to the fact that it arose on the basis of a logically unfounded mixing of two problems, viz. the choice of the stratotype of the Pliocene/Pleistocene boundary, i.e. of units with fairly distinct stratigraphic range and content, and a classification and determination of the volume of deposits distinguished as the Quaternary system, which has always been and still remains debatable.

### What have we rejected

Earlier, stratigraphers, only by intuition, were convinced quite sure that the limit which was in different parts of the planet accepted as the Pliocene/Pleistocene boundary (under Sicilian—Chaudine—Baku and till of the first continental glaciation of Europe), was synchronous. In the USSR, on the basis of the first, in fact, reconnaissance paleomagnetic measurements, a theory was developed that the base of the Baku and Chaudine deposits is younger than the Brunhes/Matuyama boundary and it does not coincide with the Pleistocene boundary under the Sicilian—Menapian and Nebraskan, as accepted in Europe and America (Gromov et al., 1965; Nikiforova et al., 1958). This assumption, which, as will be shown below, is erroneous, was used as one of the reasons for drawing the Pliocene/Pleistocene boundary in the USSR at a lower level.

The investigations, carried out in the last years, have shown that the Chaudine regional stage in parastratotype sections of Georgia, where it is best represented, for 2/3 belongs to the Matuyama R-zone, and its base coincides with the second (from the top) event of normal polarity with a computed age of 1.1 Ma (Zubakov et al., 1975). In the Azov area, finds of the Taman mammal assemblage are correlated with the Lower Chaudine (Lebedeva, 1978), which in itself testifies to the Matuyama age of the Lower Chaudine. In the sections of Azerbaijan, the Tyurkyan subcontinental break in which a N/R inversion has been recorded, presumably of Brunhes—Matuyama age (0,73 Ma), was dated by the ash track method as 0.95 to 1.05 Ma (Ganzei 1984). And, finally, A. V. Mamedov and B. D. Aleskerov (1985) found *Didacna nalivkini*, which is an index fossil for the Baku regional stage, in drilling cores from boreholes, penetrating the Tyurkyan Formation in the Kura Lowland. Thus, the initial conclusion, drawn by G. I. Popov et al. (1947) about the synchroneity of the Chaudine and Baku deposits, has been confirmed in the light of the latest evidence. However, the age of the base of the Chaudine—Baku regional stage appeared to be 400,000 years older, than it had be assumed before, and reaches 1.0 to 1.1 Ma.

In the same years, it was established that the age of the base of the Sicilian regional stage is also estimated at 1.15 Ma (Rio, 1982; Colalongo et al., 1981). The Menapian Glaciation, according to the latest evidence, preceeded the Jaramillo, being dated between 1.2 and 1.1 Ma (Zagwijn, 1985). The age of the Nebraskan "B" till in North America also appeared to be about 1.2 to 1.0 Ma (Easterbrook, Boellstorff, 1981), as well as that of the maximum glaciation till in the Patagonian Andes (Mercer, 1978). The last major reconstruction in the organic kingdom, viz. the replacement of the Villafranchian fauna assemblage by mammals of the Tiraspolian—Galerian, i.e.

Pleistocene proper fauna, also took place in the interval from 1.3 to 0.9 Ma (Geochronology of the USSR, 1974; Azzaroli, 1983). The appearance of Homo genus is also recorced at this limit (Leinders et al., 1985). And, finally, the erosional phase in the Alps, accepted by Penk and Brückner (1909) for the beginning of Diluvium—Pleistocene, and similar phases in the mountains of Central Asia and Altai, according to paleomagnetic evidence, are dated to the interval from 1.1 to 0.9 Ma.

It is quite surprising that long before precise dating and correlation techniques came into being, our predecessors could find a synchronous stratigraphic level in different regions and define it unanimously as the Pliocene/Pleistocene boundary. Therefore, the history itself has confirmed the efficiency of the methodological principles of "event" stratigraphy. To the abovesaid, it should be added that this level is also easily distinguished in the deep-sea section by appearance of a small *Lephyrocapsa Kampteri* as 1.13 Ma (Rio, 1982); *Mesocena elliptica*, 1.3 to 1.0 Ma; *M. quadrangula*, 1.1 to 1.8 Ma (Bukry, 1984) etc.

## What have we come to

On agreement (and more exactly, by voting) of the Working Group 41 members* the sapropelic "e" horizon in the 300 m mighty Vrica section, south of the Crotone town in the Calabrian Peninsula, has been elected for the new stratotype of the Pliocene/Pleistocene boundary (Aguirre, Passini, 1985). This sapropelic bed lies 7 to 10 m above the top of the normal polarity zone N2 and 80 m below the base of zone N3. Thirty-five metres above sapropelic bed "e", there is an ash horizon, "m", with *K/Ar* dates of 1.99±0.8 Ma (Obradovich et al., 1982). At the level of the sapropelic bed "e" , *Neogloboquadrina atlantica* disappear and several other species of planktonic foraminifers appear, as well as ostracods, viz. *Cyteropteron testudo* etc.

Due to the fact that abundant datings of ash "m", the overlying tuff and the Pliocene ash, occurring 280 m below the sapropelic bed "e", give a considerable scatter in the interval of 1.99 to 3.6 Ma (Arrias et al., 1978; Obradovich et al., 1982), paleomagnetic data have been used for determining the computed age of sapropelic bed "e" of 1.64 Ma (Tauxe et al., 1983). Thus, the lower double zone of normal polarity "N1—N2" has been identified with the Olduvai (1.86 to 1.67). For instance, it was noted that it lies between LAD *Discoaster brouweri* and FAD *Gephyrocapsa oceanica* (Rio, 1982). In other words, the age of the so-called "golden spike", as well as that of N1—N2 polarity zone in the Vrica section, has been obtained by interpolating estimates of dating levels from the oceanic sections.

These interpolations are not unquestionable. Thus, in a more complete section of Santerno in the Appennines, Nakagawa et al. (1974) distinguish 7 intervals of normal polarity above the Globorotalia crassaformis Zone; the uppermost one should be assigned to Jaramillo on the basis of *Archidiskodon meridionalis* presence (Arrias et al., 1978), whereas the fourth one from the polarity zone N, dated by the appearance of *Arctica islandica*, should be assigned to the Reunion event. However, since Rio (1982) takes FAD *Arctica islandica* and *Cytoropteron testudo* as a synchronous level, in the Vrica section, the lower zones N1—N2 may be assigned to the Reunion. This is in full agreement with FAD *Gephyrocapsa oceanica*, recorded 25 m above the sapro-

* Including, which is highly significant, only those who adhere to drawing the boundary at a lower level, in different versions.

pelic bed "e", if it is dated as 1.77 Ma, according to BERGGREN et al. (1980). The last appearance of *N. atlantica* in the Vrica section, associated with "d—e" sapropels, is dated by DRIEVER (1984) as 2.27 Ma. This means that the lower polarity N1—N2 zones in Vrica could be older than Reunion, i.e. they could correspond to a normal polarity event with an age of 2.33 Ma. The doubts, presented herein, are in good agreement with the appearance of *Hyalinea baltica* in other parts of the world (e.g., near the coasts of Jawa or New Zealand) within the time interval of 2.5 to 2.3 Ma. As is seen, the age of the "golden spike" in the Vrica section is not uniform.

Let us assume that the estimate of sapropelic bed "e" as being 1.64 Ma old is correct, and let us discuss the problem of correlating continental sections with this stratotype. As it is known, the stratigraphy of the continental Pleistocene, as well as of the Upper Pliocene is also mostly based on paleoclimatic evidence. However, these data have been rejected by the Working Group on IGCP Project 41 as a criterion for drawing the Pliocene/Pleistocene boundary and a tool of its interregional tracing (AGUIRE, PASINI, 1985), and the Vrica section has not been studied at all from the viewpoint of climatostratigraphy. Therefore, the only means of correlating continental sections with the Vrica reference section are paleomagnetic and chronometric data.

However, radiological evaluations of continental formations in the interval between 2 and 1 Ma are quite scare and unreliable. As to the zone of normal polarity in the Matuyama orthomagneteme, the picture is also not clear. In different sections, a different number of N zones is distinguished, viz. from 3 to 7, and their correlation is, as yet, extremely difficult. It can be carried out only on the basis of a set of data by distinguishing bio-magnetostratigraphic steps or "seasons" (KOCHEGURA, ZUBAKOV, 1978). However, the position of the Olduvai event relative to biostratigraphic zones in the continental sections is quite unhappy, since it lies within the Upper Villafranchian and Lower Apsheronian. The newly proposed boundary with the age of 1.64 Ma can be "recognized" in continental sections only on the basis of rodent fauna, viz. FAD *Allophaiomys* and LAD *Mimomys pliocaenicus* (HORACEK, 1981). Such sections are still rare. That is why, the Siberian Group in its report on Project 41 makes quite a reasonable conclusion that "the stratotype in Vrica... would rather be called disappointing (similar to the earlier known ones of Le Castella and Santa Maria di Catanzaro), than giving hopes" (ARKHIPOV, 1984, p. 21).

Thus, the Pliocene/Pleistocene boundary at the Vrica "e" level, which, because of its correlation potentialities, is well-suited for deep-sea sediments, is less appropriate for continental formations. That is why, and not by chance, those, who are in favour of drawing the boundary of the Quaternary system on a lower level and are studying continental deposits, let it go down to (3) —2.5 Ma (ZAGWIJN, LIU THUNG-SHENG et al.) (4) —3.2—3.5 Ma (HAUG, GROMOV, KRETZOI, PÉCSI, RÓNAI), or (5) —5.2—7 Ma (YAKHIMOVICH). It is quite obvious, however, that when choosing a stratotype in a marine section, one should take into account the fact, how realistic this boundary is for continental sections.

An indispensable condition for the existence of the International Stratigraphic Scale is the stability of boundaries of valie stratigraphic units. If they are shifted, a new boundary should be better, more practical and convenient than the older one. This condition was not met, when the lower boundary of Pleistocene was transferred from the base of Sicilian (Calabrian) regional stage onto the Vrica "e" level. The new regional stage, viz. Selinuntian (RUGGIERI, SPROVIERI, 1977, etc) is an artificial unit, since the most significant stratigraphic limit is within this stage.

### *Characteristic features of modern practice of high-resolution stratigraphic correlation and its requirements to boundaries of general stratigraphic units of Late Cenozoic*

Below the authors will give a brief account of their ideas about the necessary qualities of boundaries of general stratigraphic units of the Late Cenozoic at a contemporary level. They should be clearly distinguished and easily followed in deep-sea, shelf, and continental sections. Consequently, they cannot be established by agreement of investigators about a particular section, but this should be done by a "natural selection" of signals, traced in practice in all three environments.

Such common signals for the ocean, shelf, and continent are polarity inversions and temperature trends. The second group is more important, for, firstly, they are not registered by one method, as the inversions are, but by a whole set of data and techniques, viz. lithological—facies changes, ecological—paleontological changes, variations of stable isotope composition, fluctuations of the ocean level, geochemical cycles etc. This allows to follow climatostratigraphic units and their boundaries from one section to another and mapping them. In this way, maps of Quaternary deposits are compiled, and this experience has been successfully applied to the Pliocene, too.

Secondly, climatic signals in the interval from the first tens of thousands years to 1.2 Ma are genetically related to variations in the Earth's orbital parameters, viz. precession, axis tilt to ecliptic, and eccentricity, i.e. to astronomic rhythms studied by MILANKOVICH. Modern stratigraphy empirically established climatic sedimentary cycles with duration of 13 to 20, 40, 80 to 100, 400 and 1,200 thousand years respectively, which may be regarded as "signals of precise geological time".

An enormous amount of data, obtained from the studies of deep-sea sediments, particularly isotopic curves (SHACKLETON, DUPLESSY, VERGNAUD—GRAZZINI, SAVIN et al.), temperature factor curves (KELLER, BARRON, THUNELL, BARASH etc), carbonate cycles (GARDNER etc), including information the Mediterranean Sea, to say nothing of a century-long experience of Quaternary geology, testify to the realistic character of climatostratigraphy. It has a long history and has been making rapid progress. Therefore, one cannot be surprised at the position of the authors of the new North American Stratigraphic Code, who have removed climatostratigraphic units from their classification. It is quite obvious that many geologists, among them members of the Working Group on Project 41, who discarded the use of a paleoclimatic criterion for following global stratigraphic boundaries, have got a wrong orientation due to the philosophic concept by Hedberg.

The modern stratigraphic scale with the ever increasing possibilities of dating stratigraphic limits and boundaries, should be very flexible ("spike" less). Valid limits in this scale will be determined more precisely, erroneous boundaries will disappear by themselves. An "event" (=signal) concept of stratigraphy provides for the best progress of the General Stratigraphic Scale, since it regards the latter simultaneously as a tool, and a synthesized draft of geohistorical periodization. The duration and significance of geological events are regarded as interrelated objective criteria for the taxonomic evaluation of certain stages in the geological history and the corresponding section intervals. The strategy, aimed at improving the General Stratigraphic Scale of the Cenozoic should, in our opinion, be based on three aspects:

*1* Following the stability of valid stratigraphic limits of the type of the traditional Pliocene/Pleistocene boundary drawn at the base of Sicilian (=Chaudine—Baku);

*2* Detailization of the Scale with the help of intergated climatostratigraphic, and micropalaeontological data;

*3* Revision of the relations between units, determining the geohistorical content of the Scale.

Let us dwell on point 3. It is clear, that the lowering of the Pliocene/Pleistocene boundary to the level of 1.6, 2.5, and 3.4 Ma does not make this interval a system, comparable with other systems. In this case, Pliocene looses its right to the rank of a series (MENNER, 1977). That is why, the problem of the number of systems, included into the Cenozoic Erathem is quite a special one, and should not be identified with the problem of the Pliocene/Pleistocene boundary. It is apparent, that there cannot, be more than 2 systems in the Cenozoic. These may be either Paleogene and Neogene or Tertiary and Quaternary. In the latter case, the boundary between systems could be made coincident with the base of Zanclian (5.2 to 5.3 Ma), or Serravalian—Sarmatian, 12 to 14 Ma*. The latter would be best founded. In any case, Pleistocene with a boundary of $1.1 \pm 0.1$ Ma remains the upper stage of the second Cenozoic system.

## Conclusions

*1* The traditional Pliocene/Pleistocene boundary drawn at the base of Sicilian (Chaudine—Baku), as its dating in different regions has shown, represents a distinct synchronous level, which is traceable on a global scale and equally well on continents and in the ocean. An objective value of this boundary has been confirmed by a long-term practice of the Geological Survey of the USSR.

*2* A resolution on making the Pliocene/Pleistocene boundary coincident with sapropelic bed "e" in the Vrica section, the age of which remains debatable, is a sad mistake, which continues a series of similar erroneous recommendations, passed earlier (in 1948 and 1972) concerning this problem. This boundary has got no reliable criteria, which enable fixing it in continental sections. Thus, practical works, which are sponsored by the Geological Survey of the USSR cannot be accomplished on the basis of such a boundary.

*3* The cause of erroneous recommendations on changing the position of the Pliocene/Pleistocene boundary lies in the mixing of two problems, viz. a more precise definition of the correlation datum, which is represented by the base of Pleistocene in its traditional sense and improving the geohistorical (-taxonomic) content of the General Scale of Cenozoic.

*4* Choice of a local limit, viz. the sapropelic bed "e" in the Vrica section, made on agreement by an initiative group of investigators, as a stratigraphic limit of two series, and, practically, systems, with regard for all the consequences of such a solution, demonstrates the ungrounded character of a new "chronostratigraphic" concept of stratigraphy, formulated by Hedberg and the Stratigraphic Guide (1976).

*Suggestions: 1* It is time to discuss urgently all problems, connected with the elaboration of a minute global geochronological scale of the Cenozoic and its geohistorical periodization. The data on the Mediterranean are of paramount interest

* Naturally, there can be no objections to solving such problems by agreement, since in this case, valid boundaries are retained, practice does not suffer, and geohistorical periodization is improved.

for this. Therefore, such a discussion should be undertaken at the next RCMNS Congress.

*2* Before this discussion, the solution concerning the choice of sapropelic bed "e" as a stratotype of the Pliocene/Pleistocene boundary should be frozen.

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# NEOGENE KINEMATICS OF THE CARPATHO—PANNONIAN REGION

by

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In the Carpatho—Pannonian region two units are usually delineated: the north Pannonian and the south Pannonian domains. Between them a poorly known strip of uncertain position exists being called middle Hungarian zone: either it can be a third independent unit or the boundary of the two domains lies somewhere within it.

Reconstruction in general means moving back of domains in a previous position. From both geological data and geometrical considerations it is clear that the south Pannonian domain has to be moved firstly. There are three main possibilities for its moving back (Fig. 1): (A) lateral displacement along the middle Hungarian zone, (B) lateral displacement along the south Carpathians and (C) rotation around the Moesian plate.

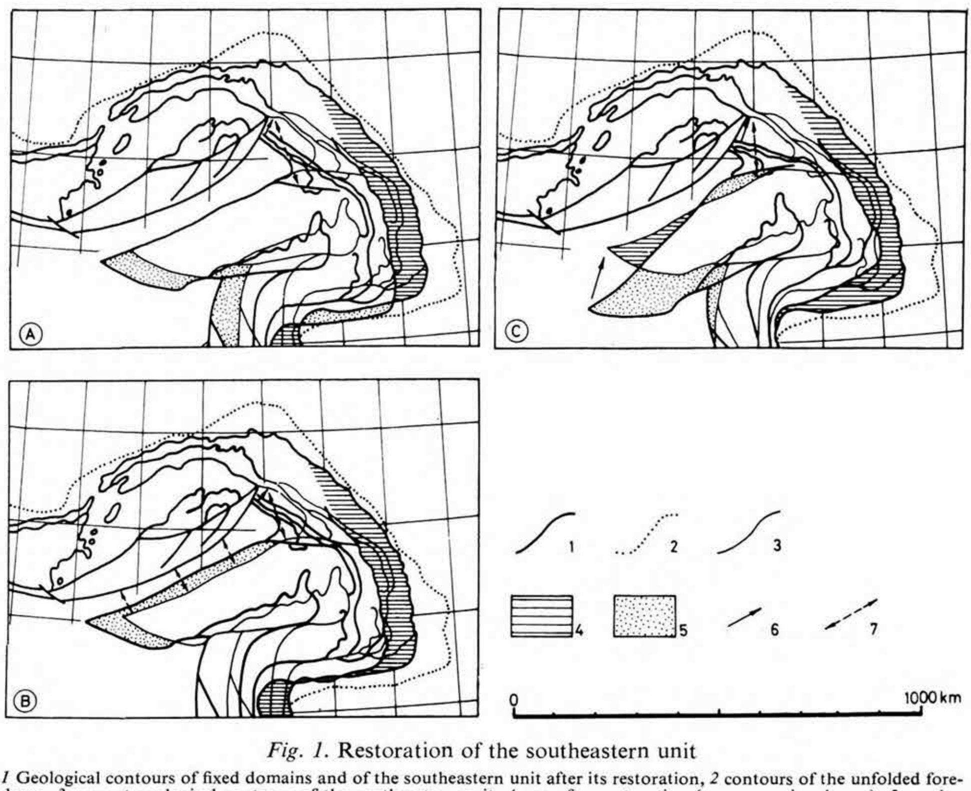



*Fig. 1.* Restoration of the southeastern unit

*1* Geological contours of fixed domains and of the southeastern unit after its restoration, *2* contours of the unfolded foredeeps, *3* present geological contours of the southeastern unit, *4* gap after restoration (=compressional area), *5* overlap after restoration (=extensional area), *6* direction of fitting, *7* direction of extension

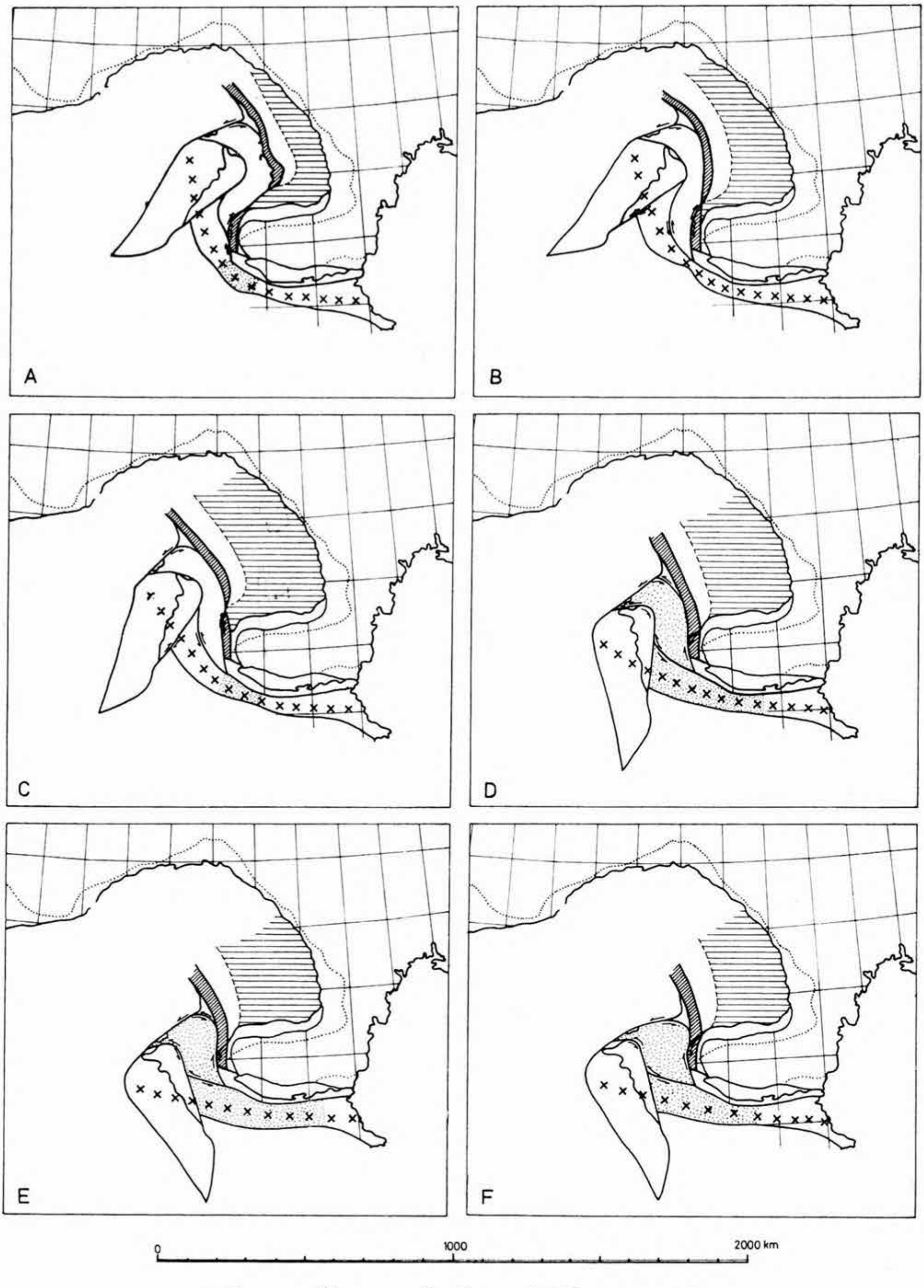



*Fig. 2.* Restoration of the southeastern unit

*1* Geological contours, *2* contours of the unfolded foredeeps, *3* trace of the present folding front after the restoration, *4* banatite belt, *5* Laramian flysch belt, *6* gap after the restoration (=compressional area), *7* area of deformations.— A Local stretching and local strike-slip displacements; maximum angle = 30°, B strike-slip displacements only; maximum angle = 30°, C strike-slip displacements with local stretching; maximum angle = 50°, D strike-slip displacements including the fault in the basement of the Transylvanian basin, with moderate stretching; rotational angle = 80°, E same with strong stretching; rotational angle = 100°, F same with more rotation in the basement of the Transylvanian basin; rotational angle = 100°

Lateral displacement along the middle Hungarian zone necessarily leads to an overlap in the south Carpathians. It would mean extension instead of convergence and is, therefore, inconsistent with geological data. Lateral displacement along the south Carpathians necessarily leads to a gap in east Serbia likewise in the previous version, also in contradiction with geological data for the Tertiary. Rotation around Moesia is the only kind of movements consistent with geology.

Main geological constraints for the rotation are as follows: (1) the Apuseni—Sredna Gora magmatic belt has to remain uninterrupted; (2) the south Carpathians in east Serbia have to remain welded with the Moesian plate; (3) the Laramian flysch belt of the east Carpathians has to remain on the outer side of the Magura zone of the west Carpathians and (4) the cross-section of the whole Balkan chain has not to suffer any interruptions.

The rotation of the south Pannonian domain around the Moesian plate means decrease of both the curvature of the Apuseni—Sredna Gora magmatic belt and the offset of the Laramian flysch belts of the east and south Carpathians. I suppose that the initial situation can be reached by means of total straightening of the magmatic belt which requires about 100° rotation. Rotation of the south Pannonian domain already by small angles requires lateral displacement of the Maramureș spur relative to the Bucovinian domain and deformation (Fig. 2A) or transcurrent displacements (Fig. 2B) within the south Carpathians or both of them (Fig. 2C). In this situation the Laramian flysch belt of the east Carpathians lies in the continuation of that of the south Carpathians. The rotation angle only can be increased by means of stronger deformation of the Bucovinian and Getic domains plus lateral displacements in the basement of the Transylvanian basin (Fig. 2D). Total straightening of the magmatic belt is possible by means of very strong deformations only (Fig. 2E—F). Since the magmatism ended in the Palaeocene the whole rotation must have taken place in the Tertiary.

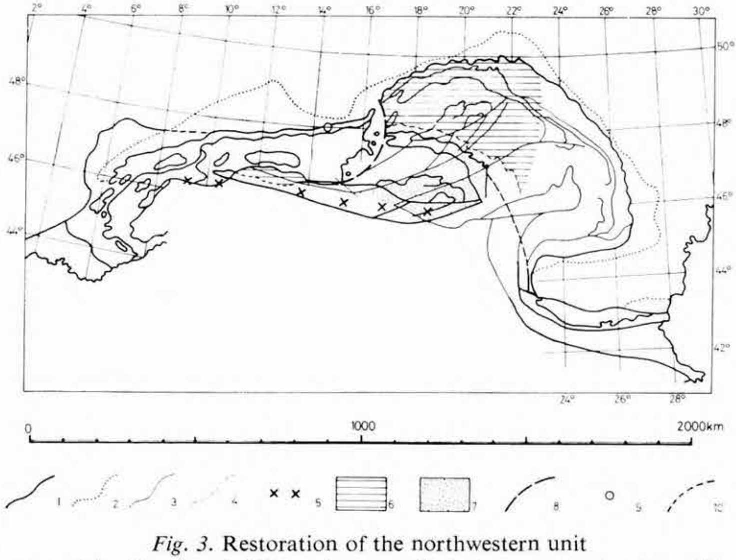



*Fig. 3.* Restoration of the northwestern unit

*1* Geological contours of fixed domains and of the northwestern unit after its restoration, *2* contours of the unfolded foredeeps, *3* present geological contours of units rotated, *4* trace of the present folding front after the restoration, *5* Late Eocene to Oligocene calc-alkali magmatic belt, *6* gap after the restoration (=compressional area), *7* overlap after the restoration (=extensional area), *8* rotational transform boundary of the Alps and Carpathians, *9* pole of rotation of the Carpathians relative to the Alps, *10* front of the Laramian flysch belt in the East and South Carpathians in the pre-rotational position

For the north Pannonian domain I suppose that in the initial situation the peri-Adriatic—north Hungarian magmatic belt is straight and the west Carpathians lie in the straight eastern continuation of the Alps after clockwise rotating back by 30° (Fig. 3). Overlaps clearly manifest that this rotation requires preliminary removal of the south Pannonian domain. Since the magmatism ended in the Oligocene the whole rotation must have taken place not earlier than in the second half of the Oligocene.

We can unite the initial positions for both domains (Fig. 4) and conclude that the clockwise rotation of the south Pannonian domain necessarily results in the anticlockwise rotation of the north Pannonian domain. Accordingly, the whole rotational history could not start prior to the Middle Oligocene. Additionally, we can suppose that the eastern half of the north Pannonian domain was deformed due to collision during the rotation.

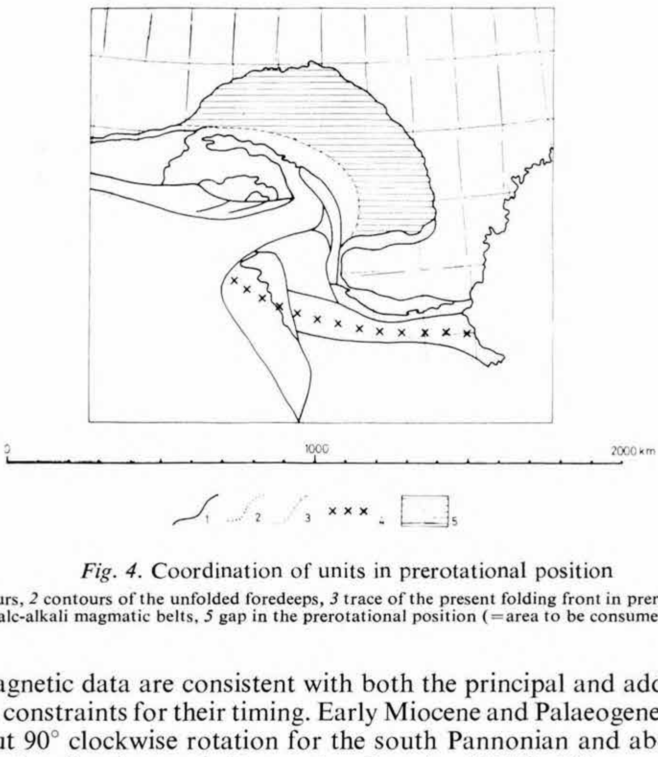


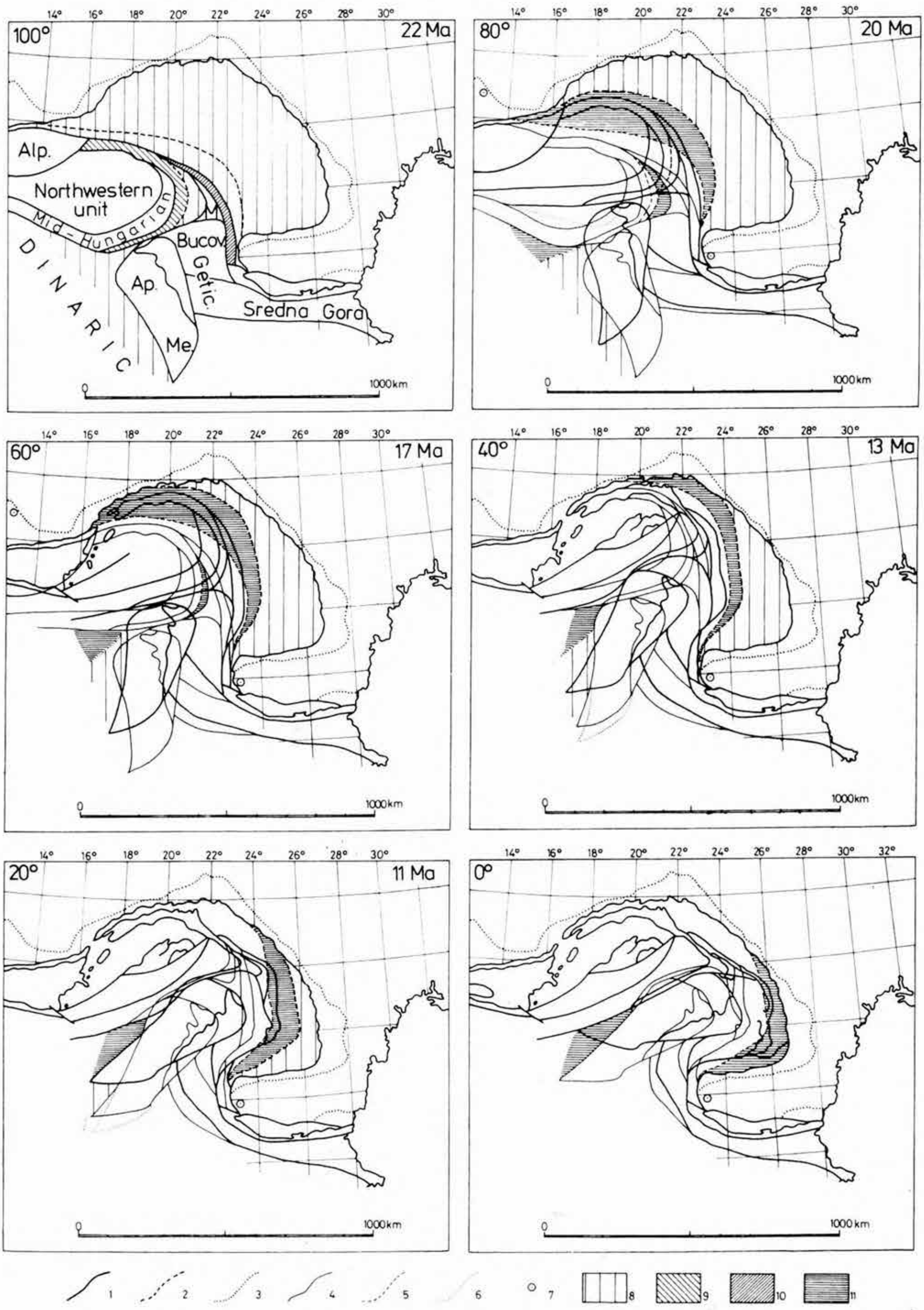

*Fig. 4.* Coordination of units in prerotational position

*1* Geological contours, *2* contours of the unfolded foredeeps, *3* trace of the present folding front in prerotational position, *4* calc-alkali magmatic belts, *5* gap in the prerotational position (=area to be consumed)

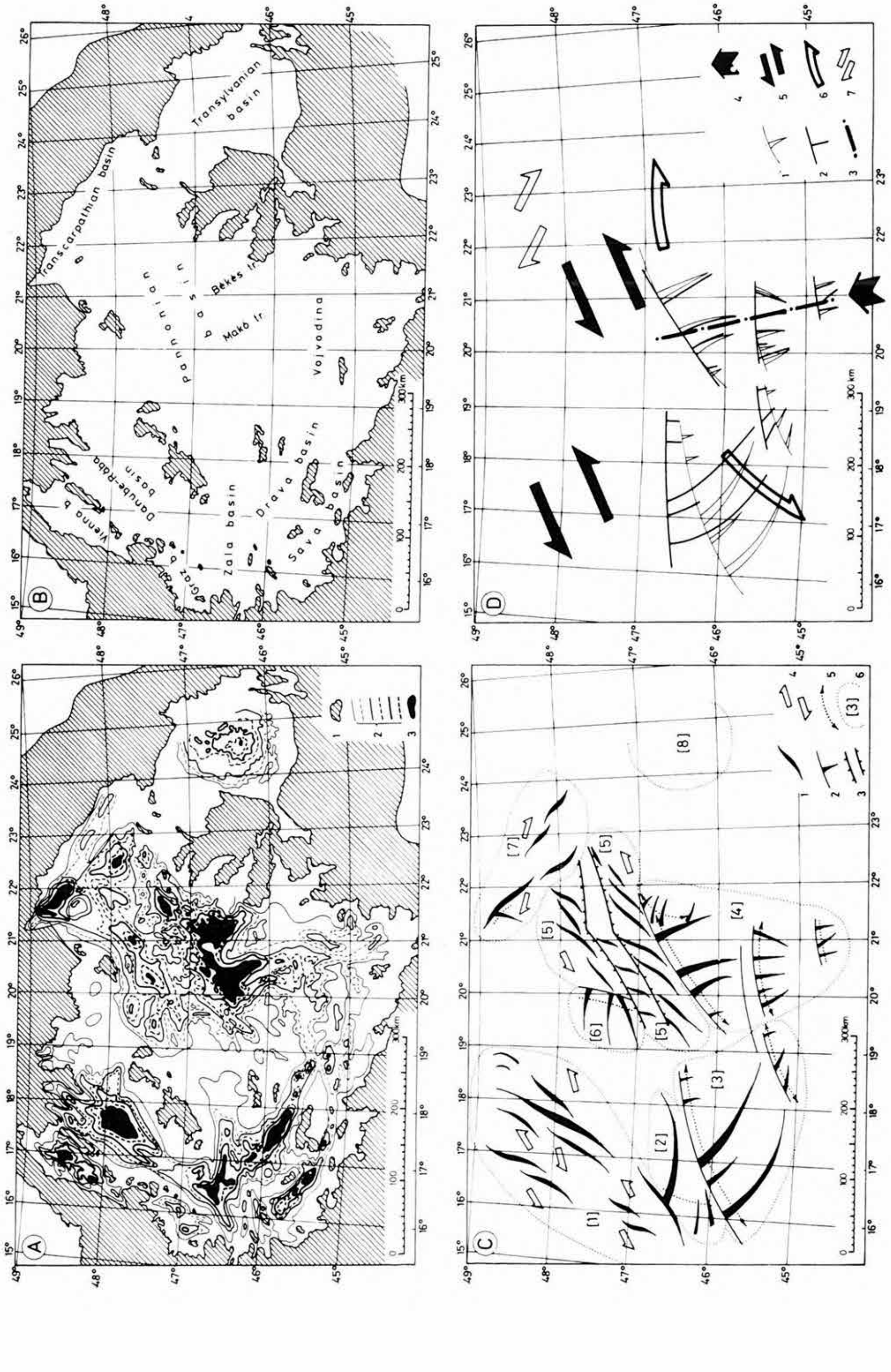
Palaeomagnetic data are consistent with both the principal and additional rotations and give constraints for their timing. Early Miocene and Palaeogene declinations manifest about 90° clockwise rotation for the south Pannonian and about 30° anticlockwise rotation for the north Pannonian domain. Middle Miocene declinations manifest no rotations in central Slovakia, about 10° anticlockwise rotation in north →

*Fig. 5.* Rotational history of the Carpatho—Pannonian realm

*1* Outlines in the given situation, *2* trace of the present folding front in the given situation, *3* trace of the present boundary of the foredeep, *4* outlines in the previous situation, *5* trace of the present folding front in the previous situation, *5* trace of the present folding front in the previous situation, *6* outlines of the south Transdanubian domain in the given situation relative to the Apuseni domain in the previous situation, *7* rotation pole, *8* area to be consumed, *9* folded Alpine and Magura—Maramureş—Szolnok flysch belt, *10* folded Ceahlău—Severin flysch belt, *11* area consumed prior to the given situation.—Bucov. = Bucovinian domain (=Centraleast Carpathians), Getic = Getic domain (together with the Danubian one), Ap. = Apuseni domain, Me. = Mecsek domain (=south Transdanubia of uncertain position because of the low probability of subduction in front of it during the whole Miocene), M = Maramureş domain ▷



100°
22 Ma
Alp.
Northwestern unit
Mid-Hungarian
DINARIC
Bucov
Getic.
Ap.
Me.
Sredna Gora
1000 km
80°
20 Ma
60°
17 Ma
40°
13 Ma
20°
11 Ma
0°
14° 16° 18° 20° 22° 24° 26° 28° 30° 32°
1 2 3 4 5 6 7 8 9 10 11


A
B
C
D
Transcarpathian basin
Transylvanian basin
Pannonian basin
Békés tr.
Makó tr.
Vojvodina
Vienna b.
Danube-Rába basin
Graz b.
Zala basin
Drava basin
Sava basin
[1]
[2]
[3]
[4]
[5]
[6]
[7]
[8]
300 km

Hungary and about 20—30° anticlockwise rotation in the east in agreement with kinematic conclusions. Late Miocene and Pliocene declinations manifest no rotation in the whole region.

The rotational history can be dated according to palaeomagnetic results and ages of the last Carpathian thrusting (Fig. 5). Accordingly, the early stage of the Miocene basin formation took place synchronously with the rotation. I suppose that the thickness arrangement (Fig. 6A) can be displayed in form of extensional zones (Fig. 6C) which can be related to lateral displacements in the north and to concentric extension in the south. In a generalized picture (Fig. 6D) we have sinistral displacements in the north due to the eastward-directed movement of the south Pannonian domain relative to the north Pannonian domain and concentric extension of the south Pannonian domain because of its pressing to the Moesian plate. Only the late basin formation in the Pannonian stage after the termination of all rotations can be related to something like the mantle diapirism.

I would like to emphasize that the rotational history outlined above (for details, see BALLA, 1984) is the only way to explain the Carpathian folding and overthrusts when considering a real geometric frame and the available palaeomagnetic data. Some problems can arise with the timing of the beginning of the process, the uncertainities, however, can cover the Late Oligocene—Early Miocene time span only.

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◁ *Fig. 6.* Kinematics of the Pannonian basin (with the present geographical network)

A Isopach map of the Neogene—Quaternary sedimentary complex: *1* Outcrops of pre-Tertiary complexes, *2* thickness isolines in km, *3* depths over 4.0 km. B Location of basins. C Interpretation: *1* Tensional zones connected with strikeslip movements, *2* tensional zones connected with concentric extension, and the concentring transform faults limiting them, *3* horst boundaries, *4* shearing forces, *5* extensional forces, *6* kinematically homogeneous areas: (1) Vienna, Danube—Rába and Graz basins—sinistral shearing, (2) Zala basin—north —south directed concentric extension, (3) Drava and Sava basins—west—east directed concentric extension, (4) southern Pannonian basin—west—east directed concentric extension, (5) northern Pannonian basin—sinistral shearing, (6) northwestern Pannonian basin—north—south directed extension, (7) northeastern Pannonian and Transcarpathian basins—dextral shearing, (8) Transylvanian basin. D Kinematics restored: *1* present position of tensional zones connected with concentric extension, *2* same in restored position, *3* fixed axis of the reconstruction, *4* active pressing, *5* active shearing derived from the relative movement of the northwestern and southeastern units, *6* concentric extension derived from the active pressing, *7* opposite shearing derived from the active one in accordance with the rotational kinematics (see Fig. 5)

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# THE NEOGENE TECTONIC PHASES OF THE NORTHERN APENNINES—SOUTH ALPINE SYSTEM: THEIR SIGNIFICANCE IN RELATION TO THE FOREDEEP SEDIMENTATION

by

M. Boccaletti, F. Calamita, E. Centamore, G. Deiana, L. Dondi, R. Gelati, F. Massari, G. Moratti and F. Ricci Lucchi

*Introduction.* The south Alpine and north Apennines chain make up a complex and sharply curved system with opposite and converging tectonic polarity. This system evolved during the Neogene in an ensialic post-collision phase after the complete closing, during Cretaceous—Eocene, of the oceanic ligurian piedmont basin (Boccaletti et al., 1982).

During the oceanic phase the two chains did not exist, because at that time the eoalpine chain was building up with the European polarity, while the north Apennines —south Alpine area represented the backland.

The development of the south Alpine—north Apennines chain had been taking place from the Neogene with inverted polarity for both sectors at the expenses of a sialic thinned crust, which formerly belonged to the eoalpine backland and later represented the common foreland of both sectors. They are built up by convergent migration—southwards for the south Alpine sector and northwards for the northern Apennines of a system of mobile foreland basin whose cratonward migration is closely related to that of the thrust.

## The tectonic setting

The tectonic setting of the study area is shown by three sections.

The first one (Fig. 1A) joins the westernmost sector of the northern Apennines with the south Alpine area. It can represent the common foreland where the external thrusts of the two chains face one another. We want to say in advance that, while in Apennines segment the most external thrusts cross also the Pliocene, in the south Alpine segment the Pliocene deposits seal the thrusts of the Miocene age. The highest levels of the Apennines chain consist of Ligurian units of oceanic realm which were at first involved in the eoalpine and later in the Apennines tectogenesis. They form an allochthonous cover overlying terrigenous foredeep sequence represented by:

*1* Macigno unit of the Late Oligocene—Early Miocene.

*2* Cervarola Unit of the Early Miocene.

*3* Marnoso/Arenacea of the Middle—Late Miocene.

At the south Alpine border the foreland sequence, where it outcrops, consists of the Gonfolite, Late Oligocene—Early Miocene resedimented formation, which has been found also below the Po Plain where its top is Middle Miocene.

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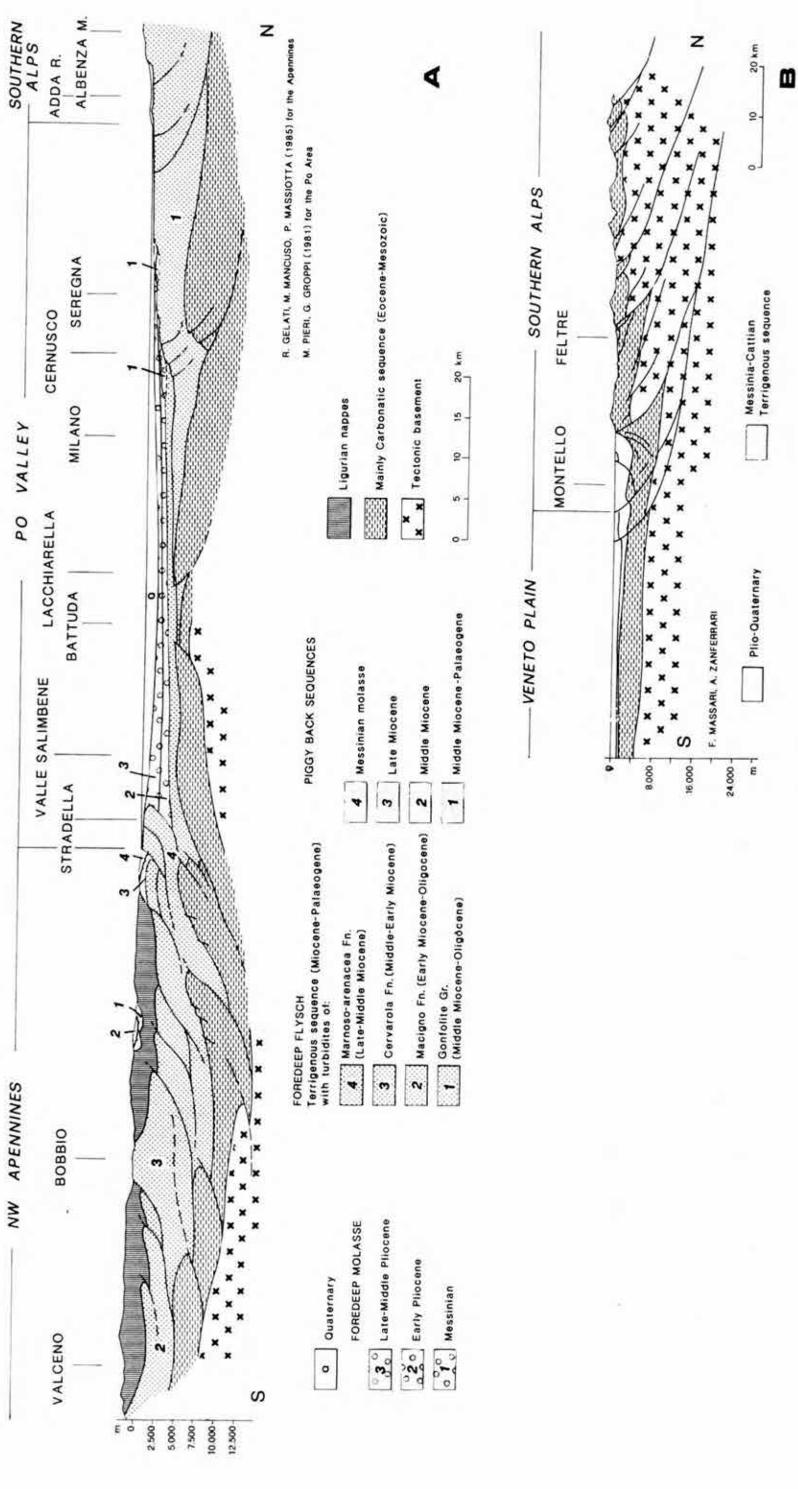

NW APENNINES
PO VALLEY
SOUTHERN ALPS
VALCENO
BOBBIO
STRADELLA
VALLE SALIMBENE
BATTUDA
LACCHIARELLA
MILANO
CERNUSCO
SEREGNA
ADDA R.
ALBENZA M.
S
N
A
Quaternary
FOREDEEP MOLASSE
Late-Middle Pliocene
Early Pliocene
Messinian
FOREDEEP FLYSCH
Terrigenous sequence (Miocene-Palaeogene) with turbidites of:
Marnoso-arenacea Fn. (Late-Middle Miocene)
Cervarola Fn. (Middle-Early Miocene)
Macigno Fn. (Early Miocene-Oligocene)
Gonfolite Gr. (Middle Miocene-Oligocene)
PIGGY BACK SEQUENCES
Messinian molasse
Late Miocene
Middle Miocene
Middle Miocene-Palaeogene
Ligurian nappes
Mainly Carbonatic sequence (Eocene-Mesozoic)
Tectonic basement
R. GELATI, M. MANCUSO, P. MASSIOTTA (1985) for the Apennines
M. PIERI, G. GROPPI (1981) for the Po Area
VENETO PLAIN
SOUTHERN ALPS
MONTELLO
FELTRE
B
F. MASSARI, A. ZANFERRARI
Plio-Quaternary
Messinia-Cattian Terrigenous sequence

*Fig. 1.*

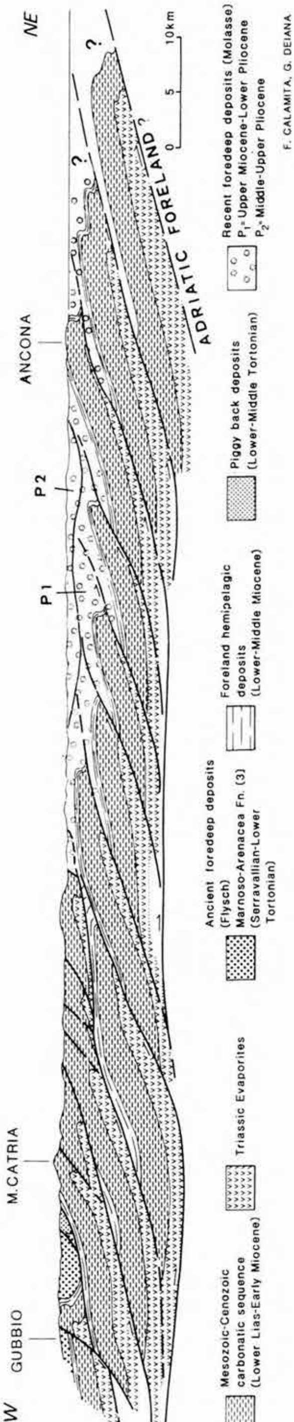

SW
GUBBIO
M.CATRIA
P 1
P 2
ANCONA
NE
ADRIATIC FORELAND
?
0 5 10km
Mesozoic-Cenozoic carbonatic sequence (Lower Lias-Early Miocene)
Triassic Evaporites
Ancient foredeep deposits (Flysch) Marnoso-Arenacea Fn. (3) (Serravallian-Lower Tortonian)
Foreland hemipelagic deposits (Lower-Middle Miocene)
Piggy back deposits (Lower-Middle Tortonian)
Recent foredeep deposits (Molasse) P1= Upper Miocene-Lower Pliocene P2= Middle-Upper Pliocene
F. CALAMITA, G. DEIANA

*Fig. 2.*

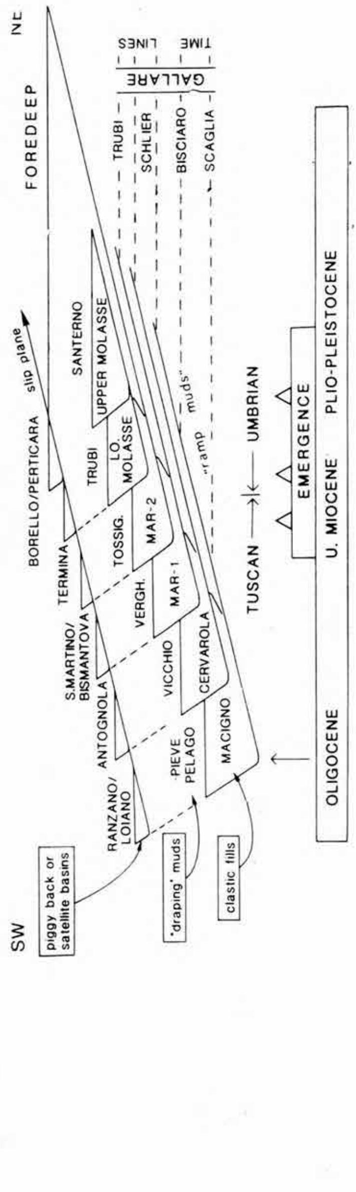

SW
NE
piggy back or satellite basins
'draping' muds
clastic fills
RANZANO/ LOIANO
ANTOGNOLA
S.MARTINO/ BISMANTOVA
TERMINA
BORELLO/PERTICARA
slip plane
SANTERNO
FOREDEEP
PIEVE PELAGO
VICCHIO
VERGH.
TOSSIG.
TRUBI
UPPER MOLASSE
LO. MOLASSE
MAR-2
MAR-1
MACIGNO
CERVAROLA
"ramp muds"
TRUBI
SCHLIER
BISCIARO
SCAGLIA
GALLARE
TIME LINES
TUSCAN
UMBRIAN
EMERGENCE
OLIGOCENE
U. MIOCENE
PLIO-PLEISTOCENE

*Fig. 3.*

The second section (Fig. 1B) crosses the eastern south Alpine sector where the external thrusts involve more recent sequences up to the Quaternary; here, according to CASTELLARIN (1978), the shortenings are up to 70%.

The third section (Fig. 2) crosses the external part of the Apennines chain from Umbria to the Adriatic sea, where it is possible to draw the values of the shortening which occurred within the sedimentary cover, exceeding 60%.

### The model of the tectonic and sedimentary evolution

The building of the south Alpine and north Apennines chains has been forming mainly during the Neogene with the migration of a complex foredeep—thrust belt system. An extensional wave followed this migration in the backland area starting from the Middle Miocene, as far as the Apennines area is concerned.

In the proposed model, already discussed in the article concerning the whole peri-Tyrrhenian area (BOCCALETTI et al., in this same volume), the different sedimentary basins are referred to diverse structural zone for each time interval. In particular, the terrigenous basins are subdivided into three types: foredeep basins, satellite or piggyback basins, back arc basins. The Fig. 3 (redcrafted from RICCI LUCCHI and ORI, 1985) shows a scheme of the space-time evolution of the basins of the system. The trust belt migration, here emphasized by the slip plane, was accompanied by the foredeep clastic wedge migration, starting from Oligocene draping muds which progressively sealed the clastic deposits and preceded (by a short time) the thrusting of the allochthonous covers. At the same time when the clastic fill occurred, muds heaped on the foreland ramp (Scaglia, Bisciaro, Schlier, etc). The main depocenters of the foredeep basin went through two stages of evolution (RICCI LUCCHI, 1985). The earlier stage, from Oligocene to Tortonian, is characterized by a flysch stage with the deposition of the Macigno, Cervarola and Marnoso—Arenacea formations. During this stage the foredeep margin was mostly submerged and the supplies came from sources located outside the thrust belt. The late stage, developed from the Late Miocene onward, was characterized by molasse deposits (molasse stage). The adjacent building was the main feeder to the foredeep, but subsidence rate was still higher than the sedimentation rate (1 m/1000 y), whereby these molasse deposits are mostly turbiditic.

Flysch and molasse wedges of the foredeep show similarities as well as differences (see also RICCI LUCCHI and ORI, 1985). Similarities are:

*a)* Outward wedging of sedimentary bodies with turbidite beds thinning and shaling towards the foreland.

*b)* Slide bodies in the basinal area, suggesting that the depocenters were close to the thrust front.

*c)* Sediment dispersal which occurs with longitudinal deflection after the lateral input.

Among the differences we find:

*a)* In the basins of the flysch stage the sedimentation stops suddenly in consequence of tectonic causes (for example, Ligurian Nappes interference) while the molasse basins have undergone a thorough filling.

*b)* The flysch sequences consist of deep water facies; those of the molasse deposits evolve towards facies of shallow water with closure conglomerate fans.

*c)* Horizontal displacements are predominant in the flysch stage, while vertical movements prevail in the molasse stage.

*d)* The supplies in the flysch stage were mainly longitudinal and of alpine origin. In the molasse stage, on the contrary, their provenance was mainly from the outcropping Apennines chain.

Also the depositional sequences of the satellite and foredeep basins show similarities and differences. The similarities are:

*a)* Common depositional trend: from a sedimentological point of view it is possible to distinguish, both in the satellite basins and in the foredeep basins, the flysch stage from the molasse stage.

*b)* Prevalence of turbiditic facies, same geometry of sand bodies and presence of slumps and olistostromes.

*c)* Multisource supply and longitudinal dispersal prevailing.

The differences characterizing the satellite basins from the foreland ones are as follows: presence of shelf facies in the Miocene, sudden and quick lateral variations, frequence of unconformities, more articulate basin topography, more immature and unorganized resedimented deposits, phenomena of cannibalism and reworking of sediments from the marginal areas.

## The north Apennines—south Alpine system: timing and features of a postcollisional evolution

In the Apennines system, during the flysch stage, the satellite basins were located on the Ligurian Nappes and often one single basin was closed and reactivated several times while in the molasse stage these basins were located also on the Tuscan–Umbrian units.

Regarding the space–time migration of the thrust belt–foredeep system, it is possible to affirm that the shifting of the compressional front and related subsidence axes of the foredeep were generally continuous, but distinct peaks of horizontal and/or vertical movements can be recognized by their characteristic sedimentary features: onset of clastic wedges, changes of sedimentation rate, gravity sliding, unconformities etc, well visible in the study area. Some of these peaks seem to represent true tectonic phases in as much as they can be traced throughout the whole western Mediterranean area.

We believe that Late Oligocene, Burdigalian, Serravallian, Tortonian and Early to Middle Pliocene phases are the most important. Particularly, during the Early Pliocene the actual molasse stage develops because of the uplift of all the internal chain. It's possible that this is connected with a significant outward jumping of the most external thrusts. At the same time a recrudescence of the tensional precesses took place in the internal zones with important collapses of the chain and consequent transgression.

The outward migration occurred in different ways related to some mountain chain segments limited by transversal lines. The main segments, from the north to the south, are separated by the following transversal lines: Ligurian Line, Livorno–Sillaro Line and Chienti–Bolsena Line. Similarly in the south Alpine area two important sectors can be distinguished, separated by a transversal line which is likely to be the Schio–Vicenza Line (Fig. 4).

We have drawn a scheme of the shifting through time for each one of these Apennines sectors (Fig. 5). In A the shifting marked by the external front of the thrust belt is evaluated about 60 km, in B the shifting is about 80 km, in C about 120 km. The shifting, therefore, increases from north to south and is accompanied by a corresponding increase of the shortening.

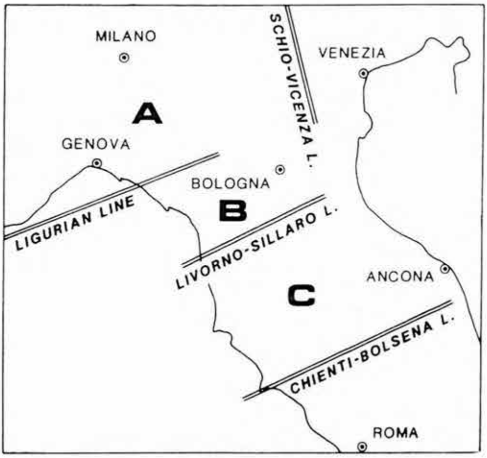



*Fig. 4.*

It is important to point out that the transversal lines of the northern Apennines are longitudinal to the south Alpine chain and vice versa suggesting an inter dependence, during the Neogene evolution, of the two domains. Therefore, probably it is not by chance that the orientation of the boundaries of the external Apennines basins are parallel to the south Alpine transversal lines. Furthermore, the space-time migration of the two sectors tightly corresponds to one another. In fact the well-known migration from SW to NE of the northern Apennines corresponds to the analogous northeastward displacement of the tectonic and sedimentary elements of the south Alpine chain. Similary the SE polarity of the south Alpine domain finds its corresponding occurrence in the shifting of some sedimentary elements of the northern Apennines in the same direction.

Full attention to the importance of the transversal features has been given only during the last years and therefore the structural data collected along them are still scarce and not sufficient to understand their significance. Anyway, the most recent geophysical researches carried out along the same, have shown how some of these lines have influenced the behaviour of the Moho isobaths thus revealing their deep significance (BOCCALETTI et al., 1985). Accordingly, our reconstruction emphasizes the importance of these transversal features, both for vertical and horizontal movements.

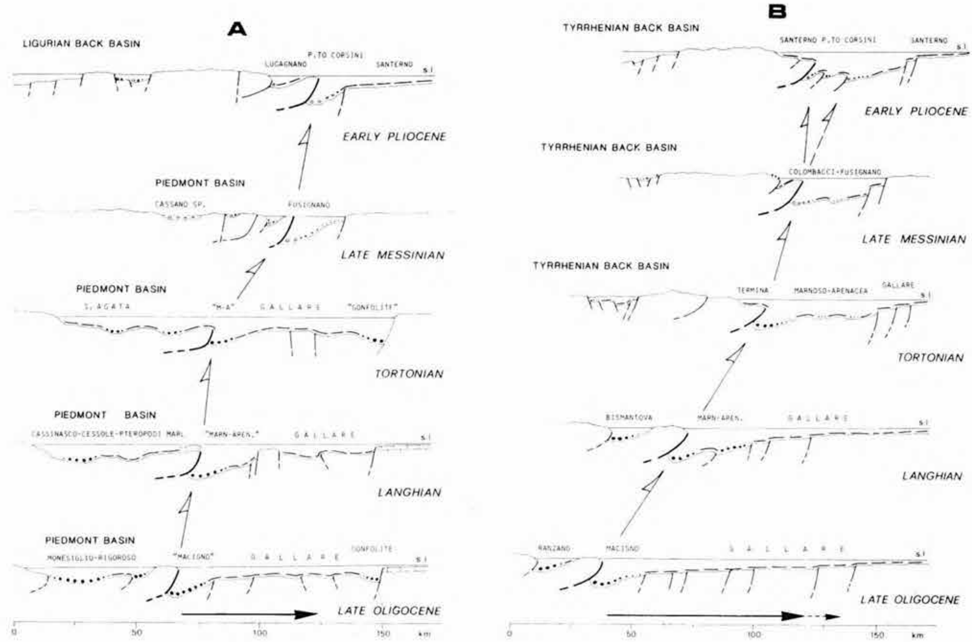
Among the most significant sedimentary elements connected with existence of these line we outline that:

*a)* They are important transport ways, which, in particular moments, break through the topographic barriers formed by the longitudinal structures.

*b)* In correspondence of these lines, in the foredeep, sudden longitudinal variation of thicknesses and facies are observed, such as to reveal that in certain moment these lines have played the role of barriers on steps.

*c)* The influence of the transversal lines can be seen also in the chain areas where they separate zones characterized by a different developing of both piggy-back and back basins.

*d)* Sometimes they give rise to topographical accidents within the sedimentary basins, creating minor basins in the chain as well as in the foredeep.



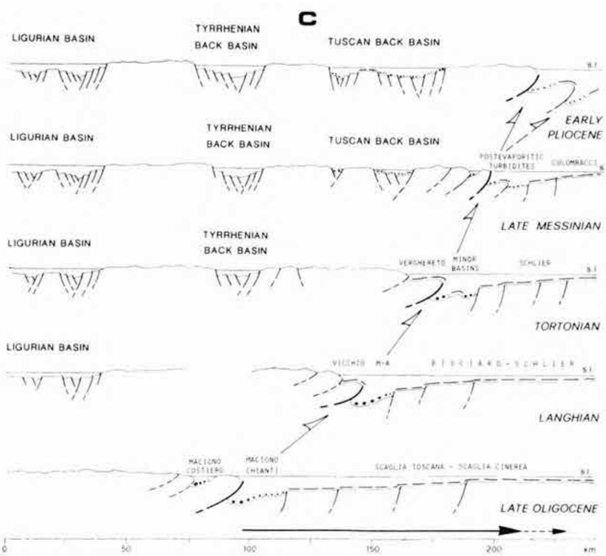




*Fig. 5.*

*e)* They represented boundaries perpendicular to the basins that at different time played reversed roles of subsidence and uplifting.

Their significance is emphasized also because the Plio–Quaternary tectonic activity differentiated in the diverse sectors.

*Conclusions*. During the Neogene evolution the south Alpine chain was linked with that of the northern Apennines. They formed a double thrust belt—foredeep system migrating toward a common foreland. The transverse lines of both chains belong to just one system that is connected to a principal stress acting NNW—SSE between the European and African blocks. In this frame the south Alpine thrusts would represent the main structures in accordance with the $\sigma_1$ oriented NNW—SSE. The northern Appennines thrusts represent the structures of a lateral chain coherent with a $\sigma_1$ NE—SW as well as the $\sigma_3$ of the main stress field system.

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# THE OPENING OF THE TYRRHENIAN SEA: TOWARDS A SEMIQUANTITATIVE APPROACH

by

M. BOCCALETTI, R. NICOLICH and L. TORTORICI

*Introduction.* The Tyrrhenian Sea is a basin characterized by an unevenly-distributed thin crust which developed within the peri-Mediterranean chains. While in the Maniaghi—Vavilov and Marsili sectors the Moho isobats show thicknesses of less than 10 km (CARROZZO et al., 1974; NICOLICH, 1981; FERRUCCI et al., 1982; RECQ et al., 1984), in the remaining part the crust varies between the 10 and 25 km marking the transitions with the Apenninic and Maghrebid chains to the east and south respectively (Fig. 1). Data from research in marine geology (SAVELLI and WEZEL, 1979; FABBRI et al., 1981; 1982; BARONE et al., 1982; NICOLICH et al., 1982; MOUSSAT, 1983; BOCCALETTI et al., 1984) show that the basin is affected by a brittle deformation. This is characterized by a N—S fault system which mainly develops along the eastern coast of the Sardinia—Corsica block but which also to be observed in the middle of the basin and in its eastern part along the Calabrian coast; a NE—SW trending fault system represented by a "central fault" (FABBRI and CURZI, 1979) and by the structures which delimit the C. San Vito basin, and an E—W system represented by the structures which separate the southern part of the basin from the Sicilian chain (BARONE et al., 1982; BOCCALETTI et al., 1984).

The structures on land are represented by normal faults bordering a Plio—Pleistocene basin which trends NNW—SSE in the southern Apennines and N—S and NE—SW respectively in the northern and southern parts of Calabria. In the northern sector of the external Calabrian Arc, the tectonic style is marked by deep structures along which the southern Apenninic chain is overthrusted on the Apulian foreland whereas the southern part is characterized by scattered thin-skinned thrusts and reverse faults which involve the sedimentary cover only (BARONE et al., 1982; BOCCALETTI et al., 1984).

The different models for interpreting the formation of this basin can be ordered into three main groups:

— the basin was primarily caused by vertical movements of the mantle; (VAN BEMMELEN, 1972; 1977);

— the present basin corresponds to a marginal basin caused by a subduction of the Ionian plate under the European plate (BOCCALETTI and GUAZZONE, 1972; BARBERI et al., 1973; DEWEY et al., 1973; GOERLER and GIESE, 1978; MOUSSAT, 1983);

— the basin is an asymmetrical megafissure which developed as a result of an African—European collision caused by a plastic-rigid deformation of the continental crust (BOCCALETTI et al., 1982; 1984).

According to this last and most recent hypothesis, the opening of the Tyrrhenian Sea has been influenced by an important E—W shear zone which developed in the southern part in time. It is obvious that in a model of this particular type a factor of primary importance is the shear line itself.

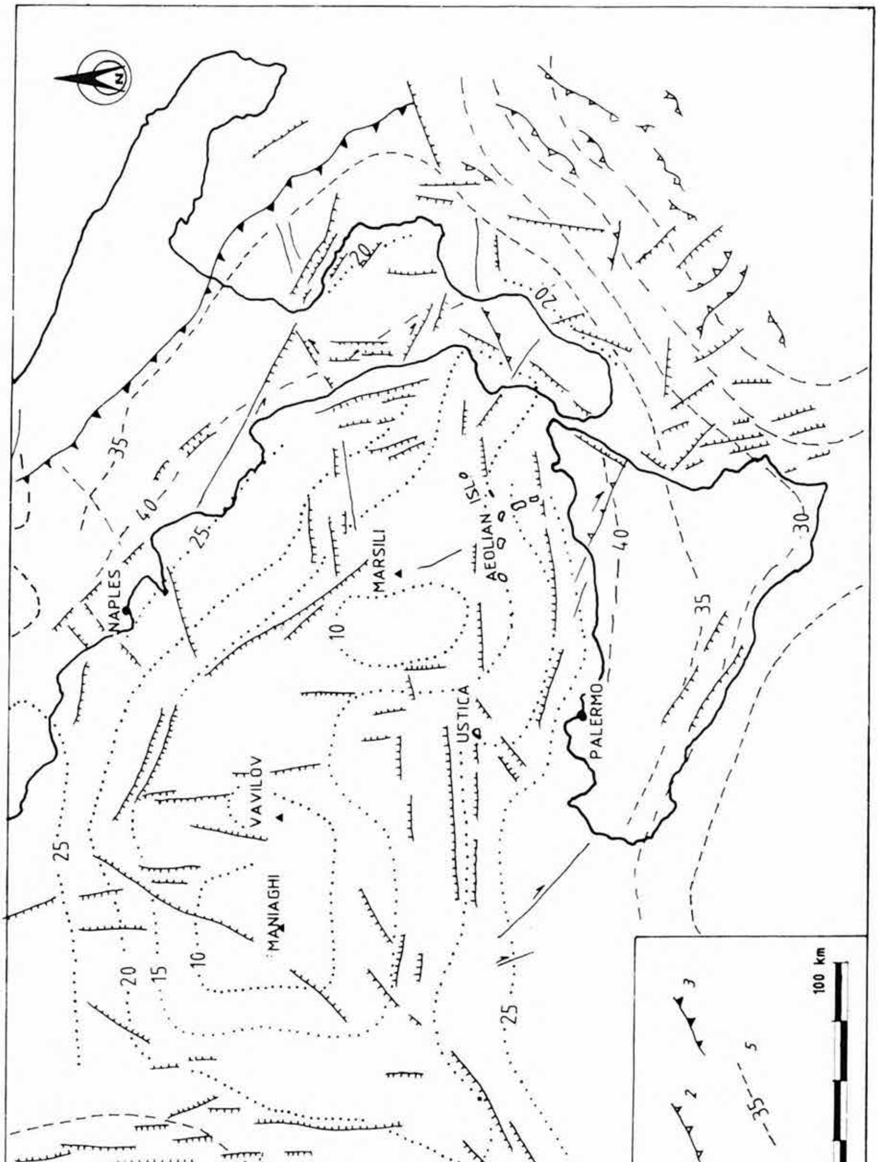


*Fig. 1.* Structural sketch map of the Tyrrhenian Sea

*1* Normal and/or strikeslip faults, *2* main thrusts, *3* external limit of the Apenninic chain, *4* Tyrrhenian Moho isobaths, *5* Moho isobaths of the external domains and of the Sardinia block

**Structural data**

The Tyrrhenian Sea is separated from the Sicilian chain in the south by a complex E—W trending belt of several structures which overlapped in Neogene. It is in fact still possible to detect the upthrust between the various tectonic units from which the chain has been constructed. It can be seen from data collected both off-shore and inland (NICOLICH et al., 1982; WEZEL et al., 1981; BARBIERI et al., 1984; CATALANO et al., 1985) that the inner part consists of several nappes of crystalline rocks which are similar in type to those found in the southern part of the Calabrian Arc (COMPAGNONI et al., 1985) to the east and in the Kabilia Massif to the west (Fig. 2). These units, which in the Straits of Sardinia are overlapped by the Sardinian basement, are overthrusted on the carbonate platform units of the Panormide complex. In its most external part the cain is overthrusted on the foreland terranes and involves Lower Pliocene and lower Pleistocene sediments in western and eastern Sicily respectively. It would appear that the NW—SE faults were partly responsible for the building of the chain. The trend of the fold axis of the Miocene—Pleistocene sediments outcropping in the Caltanisetta basin in Central Sicily suggest that these faults had a dextral movement. Along the east—west shear zone delimiting the Tyrrhenian Sea can be seen not only a NE—SW and NW—SE trending sedimentary basins but also volcanic centres. In particular this zone is composed of two principal lines (Fig. 3) one to the north and another to the south. The northern one would appear to cross the Calabrian peninsula at the level of the Catanzaro Strait where a series of E—W trending basins with a sedimentary thickness of more than 1000 m have developed (BARONE et al., 1982; C.N.R., 1985).

Between these two lines a number of different NW—SE trending normal faults delimit some very important deep basins which are characterized by 1200 m of lower Pleistocene sedimentary deposits. A volcanism whose most recent products are 1.8 million years old has been developed along these structures (BECCALUVA et al., 1982).

Another very well-developed system is represented by the NE—SW trending faults. These structures, which cut across those trending NW—SE, delimit the Pleistocene basins to be found in the C. San Vito basin and in the great Cefalù—C. Orlando basin which extends from Cefalù as far as the Aeolian Islands and the Gioia Tauro basin. That this system is superimposed on the older basins can be seen from the way in which the isopachs developed from less than 1000 m. along the eastern part of the Aeolian Islands where the Lower—Upper Pleistocene sediments are marked by a very important unconformity at the base (WEZEL et al., 1981; BARONE et al., 1982). The oldest example of volcanism (0.7—0.8 million years) in the Aeolian Islands (BECCALUVA et al., 1982) can be found within the Cefalù—C. Orlando basin. To the north of this EW shear zone tectonic activity from Middle Miocene was probably more homogeneous than elsewhere thus enabling the Paola, Crati and Crotone basins to be developed. It can therefore be seen that within the shear zone the trend of the basin changed from NW—SE to NE—SW and that this change probably began in Lower Pleistocene. This latter trend is in agreement with structural and seismological data for the neo-tectonic and present-day stress fields which are characterized by a $\sigma_{H}$min axis trending NE—SW (CELLO et al., 1982; GASPARINI et al., 1982).

It is clear that this shear zone divides the Calabrian arc into a northern and southern sector. To the east the Paola basin is separated from the Crati basin by the Calabrian coastal chain. The latter is characterized not only by upper Pleistocene to recent marine and continental terranes but also by the asymmetrical development of its principal drainage system which flows along the Sila Massif; this suggests that the

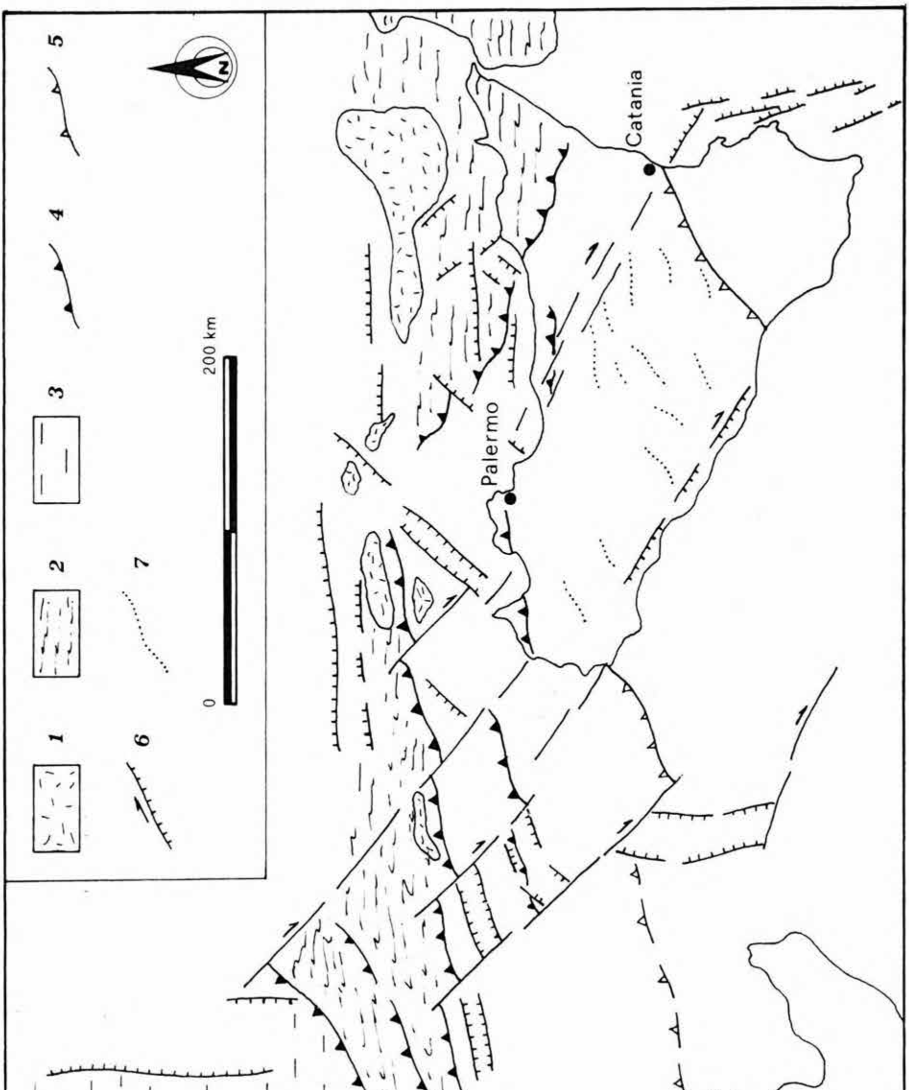


*Fig. 2.* Structural scketch map of the southern sector of the Tyrrhenian Sea
*1* Volcanics, *2* crystalline internal units, *3* Sardinia block, *4* internal thrusts, *5* external limit of the chain, *6* faults, *7* Plio—Pleistocene fold axes

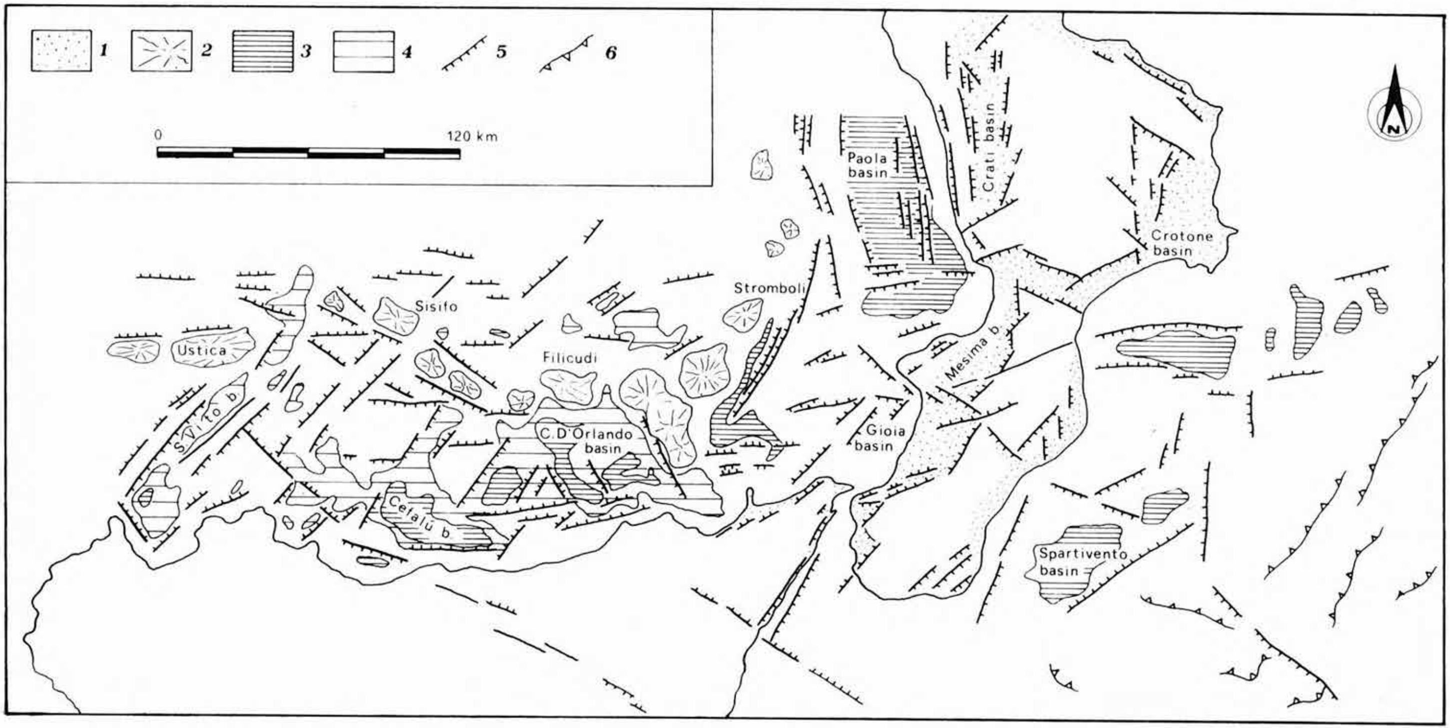


*Fig. 3.* Structural scketch map of the E—W trending shear zone

*1* Plio—Pleistocene basins, *2* volcanics, *3* Plio—Pleistocene sediment thickness (500—1000 m), *4* Plio—Pleistocene sediment thickness (0—500 m), *5* faults, *6* main thrusts entire sector is beginning to tilt along a growing fault. For a period of about 0.7—0.8 million years i.e. from Middle Pleistocene to the present day, the existence of this division is confirmed by the fact that the uplift characterizing the southern sector is much more accentuated than that in the north. In the northern sector the Paola basin is filled by Lower Pliocene sediments which are unconformably covered by Plio—Quaternary deposits (BARONE et al., 1982). Similar structures are to be seen to the west near the Marsili sea mounts where the sediments are different in age: they are Messinian and Lower Pliocene (WEZEL, 1982).

It can be also seen that along this sector, under Calabria massif, the Tyrrhenian Moho overlaps the Apulian Moho with a low-velocity layer interposed (GUERRA et al., 1981). Although in the southern sector the pattern is more or less identical with that of the north, it shows that the tilted blocks begin in Pleistocene time at the level of Aspromonte and continue to the east as far the Spartivento basin.

## Conclusions

The data provided can be seen as belonging to an evolutionary process which caused the Tyrrhenian Sea to open in time from west to east. Geophysical data based on heat flow indicate that its age varies from 7 to 10 million years in the west, 6 in the middle and 4 towards the east (DELLA VEDOVA and PELLIS, 1985). Such data are in agreement with the hypothesis that tilting phenomena affect the southern Tyrrhenian Sea which passes from Messinian in the Marsili area to Recent in the Crati valley.

The shear zone evolving to the north would confirm that the tilting phenomenon began in Messinian and continued through Early Pliocene to Recent. We are of the opinion that this phenomenon is connected with the fact that the Tyrrhenian crust is progressively thinning towards the east thus enabling a hot and softer mantle to be interposed between the Tyrrhenian and foreland Mohos. Whilst in the northern sector this phenomenon is continuous from Messinian to Recent in the south it only started, in Early Pleistocene.

This might suggest that the opening of the Tyrrhenian Sea is connected with movement along a horizontal plane within the crust. We are however of the opinion that the opening of the Tyrrhenian Sea is responsible for the shortening of the external zone. In fact if we compare the data concerning both thinning and shortening it can be seen that the quantities involved are in both cases very similar. Remembering that the normal thickness of the original crust is 35 km and that in the thinnest zones the extension is about 80% and in the others about 55%, the lateral extension from Middle Tortonian ist about 290 km with an average opening velocity of about 3 cm per year.

Since the shortening in the external zone is 180 km (BOCCALETTI et al., 1985) in 6.4 million years, it would appear that the phenomenon progressed at an average rate of about 3 cm per year. These velocity values are once more confirmed by calculations concerning the spread of the tilting phenomenon which is estimated at 2.9 cm per year along the cross-section of the northern part of the Calabrian arc.

In conclusion, it appears that the Tyrrhenian Sea is a basin which is developing as a megafissure. It is influenced in the south by an E—W shear zone which also cuts the Calabrian arc. Until Early Pleistocene this zone must have moved in a dextral direction. The fact that the northern part spread more rapidly than the southern part is suggested by the NW—SE direction of the basin.

In Early Pleistocene the continental crust blocked the northern part against the Apulian plate to such an extent that the southern zone spread onto the thinned crust of the Ionian basin. This movement was in fact inverted along the E—W shear zones where the system of structures developed (cf. the basin in a NE—SW direction) agrees with the hypothesis of a left strike slip.

It can therefore be affirmed that the opening of the Tyrrhenian Sea is only continuing along that part of the southern coast where the crust is thinnest. According to such a model the volcanism observed must be connected with crustal shears rather than with subduction phenomena.

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# NEOGENE PALAEOGEOGRAPHY IN THE CENTRAL AEGEAN REGION

by

H. BÖGER and M. DERMITZAKIS

Until recent times Neogene sedimentation pattern and paleogeography in the central part of the Aegean Region have remained enigmatic. This was primarily due to the ignorance of stratigraphic ages of some important sequences but also to the fragmentary preservation of the Neogene series in this particular area. We find them in a patchy distribution and on some islands only. Because of these two reasons any stratigraphic correlation seemed to be almost impossible. In addition, this situation was the more difficult in so far as most of the strata in question are of terrestrial or lacustrine facies and the fauna—consisting primarily of freshwater gastropods—was regarded to be without any stratigraphic value.

During the last ten years intensive work has been devoted to the palaeontology of those freshwater gastropods and it became possible to show that these faunas are quite useful as stratigraphic tools (BÖGER, 1981, 1983; BÖGER and WILLMANN, 1979a, b; WILLMANN, 1977, 1980, 1981, 1982). As a consequence of these exertions some of the most important Neogene sequences of terrestrial facies on the Aegean Islands could be put into a stratigraphic frame (BÖGER, 1983). Surprisingly, some of these turned out to be much older than hitherto suggested. This insight shed new light on the relationships between Neogene and "pre-Neogene" units in the Aegean Region also.

Almost all Neogene sequences within the Hellenic arc—being of marine or terrestrial facies—usually have been regarded as elements of a "postorogenic molasse stage". By this statement it was meant that they did not take part in the orogenic cycle proper. Instead it is believed that they originated merely as erosional products piled up within local troughs built by disruptive tectonics almost near the end of the orogenic cycle. This widely held opinion can no longer be maintained, at least not in the Aegean Region.

More than 30 years ago already MERLA (1952) was able to show convincingly that in the Northern Apennines molasse-series had been piled up on top of moving nappes and transported together with them over a distance of more than 100 km (see also SESTINI, 1970). At the same time equivalent parts of the molasse were overridden by the same nappe. After BÖGER (1983) had suggested that in SW Anatolia and on Kos island similar processes must have taken place, HAYWARD (1984) could prove that in the Göcek area (SW Turkey) Lycian Nappes overstep the Upper Miocene fill of the Kasaba basins. For processes like these, BÖGER (1983) introduced the term "Loiano-effect" and took the view that they had played an important role in the Central Aegean Region during Serravallian/Tortonian times.

After JANSEN (1977) had supposed the allochthonous position of Neogene strata on Naxos island, DERMITZAKIS and PAPANIKOLAOU (1980, 1981) put forward the hypothesis of a "Cycladic Nappe" part of which were those Neogene strata not only on Naxos but on other islands of the Cyclades as well. BÖGER (1983) took up this idea trying to combine observations on the Cyclades with others on the Dodecanese Islands and coined the term "Aegean Nappe".

Of particular interest is the Neogene sequence of Anaphi (SE Cyclades). Obviously it is completely allochthonous and appears to be overthrust on a metamorphic nappe pile with remnants of a Late Cretaceous high temperature belt (REINECKE et al., 1982) equivalents of which can be found in the uppermost nappe unit on Crete.

We are not quite sure yet as to the complete stratigraphic range of the Anaphi Neogene. What can be said is that it contains freshwater sediments of Upper Miocene and Pliocene age. The lowermost unit consists of conglomerates, sands and marls and is of terrestrial and fluviatile origin. The poorly preserved fauna consists of gastropods, pelecypods and ostracods, all belonging to a freshwater community. Occasionally a slight marine influence is traceable. Most of the conglomeratic components can not be derived from the underlying metamorphic rocks; their origin is unknown. Amongst them we find huge exotic blocks of dark or yellow limestone with ?Eocene *Nummulites*. The whole thickness of the Neogene on Anaphi can be calculated by at least 300 m.

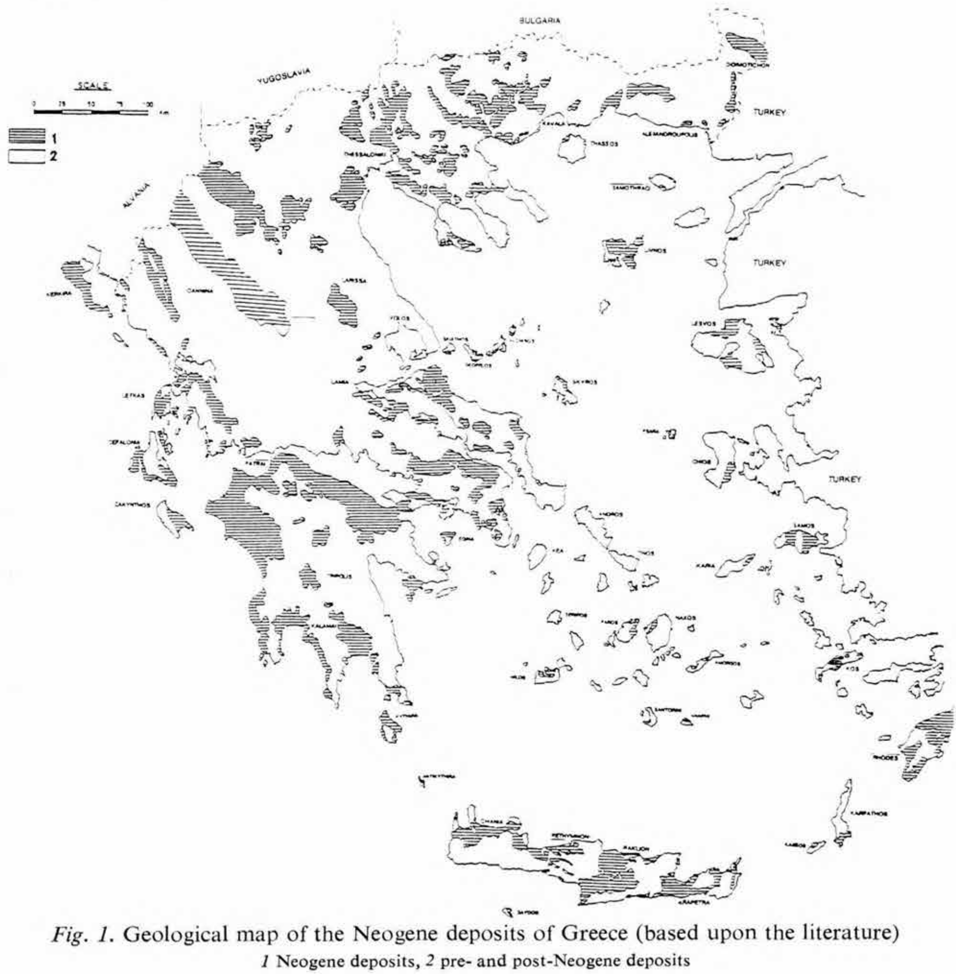


*Fig. 1.* Geological map of the Neogene deposits of Greece (based upon the literature)
*1* Neogene deposits, *2* pre- and post-Neogene deposits

The area of origin of this clastic material must have been a nearby and rather extended subaerial relief from which a sufficient freshwater supply must be supposed. To all appearance the only possibility of explanation is to assume that the position of the area of erosion and that of sedimentation was on top of still outward-moving "Central Aegean Nappes". During Late Miocene times these nappes sustained a paroxysmatic decoupling process by which different constituting parts moved off into different directions.

This hypothesis is supported by observations on some islands of the Central Cyclades. Anokoupho and parts of Katokoupho (two small islands south of Naxos) are built up of a thick sequence of fluviatile conglomerates, lacustrine travertines, marls and laminated freswater limestones. This Eremonisia formation (BÖGER, 1983) is of Upper Tortonian age. We find it also on the Makares islands east of Naxos.

Locally the Eremonisia formation unconformably rests on top of the Pesulia formation (BÖGER, 1983) which consists of dark marls, clays, sands and conglomerates. The marls contain a freshwater fauna of Upper Miocene age. On Naxos the Pesulia formation is tectonically connected with slices of marine Lower Miocene.

On Paros a marine Miocene rests on ophiolites and both units are unconformably overlain by lacustrine travertines of Tortonian age, the Damoulis formation (BÖGER, 1983). This whole complex is thrust over parts of the crystalline belt (DERMITZAKIS and PAPANIKOLAOU, 1980; 1981).

Whereas on the Cyclades slices of Miocene sediments of marine and terrestrial facies are piled up tectonically forming a complex medley, we draw the inevitable conclusion that their origin has been connected with a Late Miocene tectonic paroxysm by which parts of the Central Hellenic Nappes became decoupled. This obviously happened primarily to higher parts of the nappe pile which had been originated as a sort of "molasse" on top of the moving nappes ("Loiano-effect"). This paroxysm took place not only in the Central Aegean Region but also farther to the southeast where it is traceable especially on Kos island. It has been assumed by BÖGER, 1983, that similar events had happened in SW Anatolia but this remains to be proved.

On the other hand one can see that in surrounding areas a terrestrial Miocene rests unconformably indeed on top of the Pelagonian Nappes or their equivalents but is without any doubt autochthonous. This is true for all areas north of the Central Crystalline Belt (for instance Samos, Euboea) and on the Peloponnesos and on Kythera, too. It must be concluded nevertheless that these local "molasse basins" were connected with moving nappes, too.

Thus we see that the development of the Neogene palaeogeographic pattern in the Aegean Region is controlled by *(1)* the emplacement of the main Hellenic Nappe pile; *(2)* the emplacement of allochthonous masses during Serravallian/Tortonian times ("Aegean Nappe" or Nappes; "Tortonian Paroxism") and *(3)* fault tectonics dominating the Aegean Region since late Tortonian times.

One point remains to be stressed. There is still no clear evidence of Messinian sediments, whatsoever the facies in the Central Aegean Region may be. This seems to be due to at least three reasons: *(1)* Messinian sediments did exist but have been completely eroded; *(2)* there is a hiatus for this span of time; *(3)* the stratigraphic evidence has been overlooked so far. To enlarge on this particular point, special research remains to be done.

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# NEOGENE GEODYNAMICAL EVOLUTION OF A PYRENEO-MEDITERRANEAN GRABEN: THE ROUSSILLON EXAMPLE (SOUTHERN FRANCE)

by

G. CLAUZON

The Roussillon graben is located to the extremity of the Pyrenean "axial zone". It extends over the margin of the Gulf of Lion. As such, it is a structure that conjointly affects the marine and the continental domains.

Two major faults delimit this graben: to the north, the Prades' fault (N 60° E) and to the south, the Albéres' fault (N 85° E). Near the shoreline, in the heart of the basin, the well Canet 1 (GOTTIS, 1958) has rediscovered the Paleozoic basement underneath a sedimentary filling 1800 meters thick. That graben depth diminishes toward the West where this same Hercynian basement reappears.

The Neogene age of this tectonic unit is attested by the faunas delivered by this sedimentary filling. This one is organized in two superposed and inconformable detritic sequences:

– underneath, a Miocene series that outcrops along the edges of the basin and that was rediscovered in depth by the drills;
– over it, a Pliocene series, wildly outcropping on the surface.

These two stratigraphic units have also been recognized off-shore, in the Gulf of Lion: as well as by the sea drills (CRAVATTE and al., 1974) and by the sismic cross sections (GENNESSEAUX et LEFEVBRE, 1980).

### The Miocene rifting

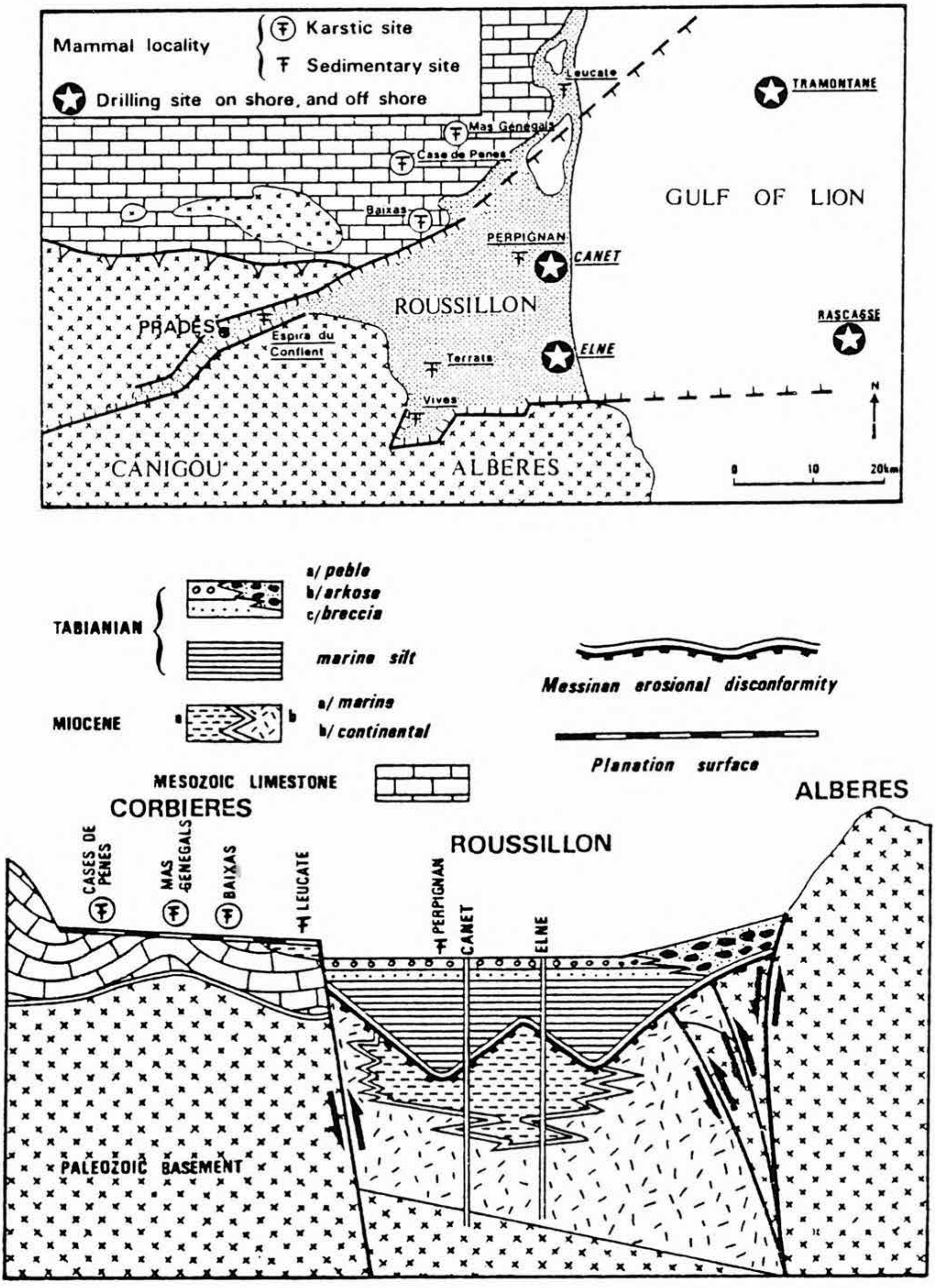
During this period, this rift showed two characteristics:
– an orientation at right angle to the shoreline,
– an amphibious disposition.

*The Miocene facies.* First, one observes a major opposition: to the East of the basin, the facies are exclusively marine (CRAVATTE et al., 1974) while, to the west, they are exclusively terrestrial (BANDET, 1975; CLAUZON et al., 1982, 1986).

Furthermore, within every one of these domains, one notices sedimentologic gradients. Within the marine domain, this gradient strikes west–east: it opposes sandy littoral formations with benthonic Foraminifera to the west, to the siltic deposits with pelagic faunas from the eastern open sea. Within the continental domain, it is an opposition between proximal (megabreccias) and distal (arkoses) facies which are disposed after a submeridian fashion, i.e. at right angle with the bordering accidents of the basin.

*Chronostratigraphy of the Miocene series.* On shore as well off-shore, this Miocene series is enframed by two unconformities:

– at the bottom of the graben, it rests in plane unconformity over the Palaeozoic basement;



Mammal locality
Karstic site
Sedimentary site
Drilling site on shore, and off shore
Leucate
TRAMONTANE
Mas Genegals
Case de Penes
GULF OF LION
Baixas
PERPIGNAN
CANET
ROUSSILLON
PRADES
Espira du Conflent
RASCASSE
Terrats
ELNE
Vives
N
CANIGOU
ALBERES
0 10 20km
TABIANIAN
a/ peble
b/ arkose
c/ breccia
marine silt
MIOCENE
a/ marine
b/ continental
Messinan erosional disconformity
Planation surface
MESOZOIC LIMESTONE
CORBIERES
ALBERES
ROUSSILLON
CASES DE PENES
MAS GENEGALS
BAIXAS
LEUCATE
PERPIGNAN
CANET
ELNE
PALEOZOIC BASEMENT


*Fig. 1.*

— at the top, it is truncated by the Messinian gullied unconformity which separates it from the overlaying Pliocene unit (BUROLLET et BYRAMJEE, 1974; BURRUS, 1984; CLAUZON et CRAVATTE, 1985).

Within the marine domain, the chronostratigraphic attributions rest upon Foraminifera and/or nannofossils (Canet, Rascasse, Tramontane and Autan drills) while, with respect to the littoral or continental domains, they rest upon rodents (BAUDELOT et CROUZEL, 1974; AGUILAR et MAGNÉ, 1978; CLAUZON et al., 1982, 1986; MICHAUX et AGUILAR, 1985).

In the off-shore drills, the Miocene series begins as early as the lower Aquitanian and ends—at the latest—within the Serravallian (CRAVATTE et al., 1974). On shore, within the satellite graben of Conflent, the rodent locality of Espira du Conflent (BAUDELOT et CROUZEL, 1974) provided a lower Burdigalian age, which allowed the datation to the Aquitanian of the 200 meters of arkosic deposits over which it rests (BANDET, 1975). To the upper limit of the series, included in a core sample cut off right under the Mio—Pliocene unconformity, the drill Canet delivered *Globigerinoides sicanus*, the Langhian planktonic marker (CLAUZON et CRAVATTE, 1985).

These results allow to say that—over the whole rift—the first levels of the filling belong to the lower Miocene while the upper part of this series is not older than the Serravallian. The chronological gap between the Miocene and Pliocene units corresponds to a duration of around 10 m.y. (CLAUZON et CRAVATTE, 1985).

*The planation surface of the northern basin border.* The Roussillon basin is overlooked to the north by a horst crowned by a Mesozoic carbonated cover: the Corbières unit. The folded structures which affect these Mesozoic carbonates are truncated by a planation surface which average elevation is around 200 meters (CORNET, 1975). This surface rises up, westward, very progressively. On the eastern side, it goes under marine deposits of a recognized Miocene age (DONCIEUX, 1903; MAGNÉ, 1978).

The geological testimony of this planation surface is twofold: palaeogeographical and geochronological. After the palaeogeographical point of view, this level—transgressed by the Miocene seas—makes up an excellent hypsometric landmark of the shorelines. Following the chronostratigraphic point of view—thanks to the numerous rodent karstic localities it delivered (MEIN et CORNET, 1973; CORNET et al., 1976; AGUILAR et MICHAUX, oral indication)—it is also a priceless landmark. Therefore, the datations provided by these karstic microfaunas are widely heterochronous, ranging from the Stampian to the Upper Pliocene.

Motionless in space (at least during the Miocene) and permanent in time (at least during the Neogene), that planation surface thus demonstrates that the horst —upon which it rests—has not been uplifted during Neogene times. Hence, the important throw (2000 m), which put it out of level with respect to the Roussillon graben, is to be exclusively imputed to the subsidence of this graben.

*Chronology of the rift subsidence.* Within the heart of the basin (Canet drill), the Hercynian basement has been reached at —1837 meters (GOTTIS, 1958). Out of that value, 200 meters are due to the Plio—Quaternary subsidence (CLAUZON et al., 1986). Thus, the Miocene subsidence is settled at about 1600 meters. An isochronous datum line—implanted on oneh and in the basin and, on the otherhand, upon the horst—permits to follow its chronological evolution during the Miocene. It is question, in the basin, of the already mentionned presence of the Langhian marker (Globigerinoides sicanus) from the Canet drill. Its faunistic association indicates a very shallow depth. On the other hand, on the Corbières horst, the locality of Leucate (AGUILAR et MAGNÉ, 1978) is concerned. This rodent microfauna (equally of Langhian age) has been delivered by a littoral deposit so its bathymetry have the same value than the preceding.

At the time of their setting, these two levels were localized at the same elevation. Today not including the Plio—Quaternary subsidence), a difference of 600 meters is recorded between them. It took place between the Langhian and the Messinian unconformity and gives the value of the subsidence in this apace of time. Accordingly, the subsidence belonging to the lower Miocene can be appreciated: its value is around 1000 meters (CLAUZON et al., 1986).

**The Upper Miocene compressive episode**

The southern edge of the basin (at leat, to the east of Maureillas) is marked out by a reverse fault. Following a simple or a repetitive superposition (Le Boulou thrust slices), the Palaeozoic basement is overlying the Tertiary material. Against the fault the layers are vertical and they have a megabrecciae facies. Their elements are issued from the southern Albères horst. They show a progressive unconformity.

These compressive structures are sealed by the Pliocene series that begins as early as the lower Tabianian (CLAUZON et CRAVATTE, 1985). Hence they are older. On the other hand, they are posterior to the Lower and Middle Miocene subsidence. Consequently this event occurred within the Upper Miocene.

**The Messinian gullying episode**

The geometry and the chronology of this gullying event are well-known (CLAUZON et al., 1982, 1985, 1986). During this episode—synchroneous of the Messinian salinity crisis—the three Roussillon rivers (Agly, Têt, Tech) have deeply cut through the Miocene filling series. Following a north—south cross section, established to the right of the present day shoreline, drops of several hundreds of meters are recorded between the palaeothal wegs of these rivers and the intermediate palaeocrest-lines (CLAUZON et al., 1986).

The dislevelments, meseared along the longitudinal profiles of these same rivers, are not less. Estimated along the whole lenght of the basin, they check around 600 meters, value corresponding to the average slopes plotted at 15 to 16‰, i.e. 4 to 5 times greater than the present-days.

The gullied topography—so obvious on the continental domain—extents offshore until the evaporites of the abyssal plains (MAUFFRET et al., 1973; BUROLLET et BYRAMJEE, 1974; RYAN, 1976; CITA et RYAN, 1978; MONTADERT et al., 1978; GENESSEAUX et LEFEBVRE, 1980; BURRUS, 1984).

Connecting the continental data and the submarine's, one succeeds restituting the continuity of the "Messinian erosional surface" (CITA et RYAN, 1978) from the Pyrenean mountain to the abssyal plains, over a distance of a hundred kilometers.

**The Pliocene ria filling**

Two major phenomenons occurred during the Pliocene:

— first one, the ingression of the Messinian canyon,

— second one, the following filling of these rias.

*Geometry and facies of the Pliocene filling up.* The top of this filling corresponds to the terminal level of the alluvial piedmont built by the Roussillon three rivers. This piedmont slopes eastward following the basin axis. It is exclusively made up of continental material: coarse (gravels) to the west, finer (arkoses and silts) to the east.

On the contrary, the bottom of this series is only made up of marine sediments. They are littoral to the west and epibathyal to the east.

The thicknesses of this filling series are, paradoxically, less important in western proximal position (300 meters) than in eastern distal situation (908 meters).

Finally, one notices a singularity in the stratigraphic relationship between the Miocene and Pliocene series: to the south—west of the basin, the latter is inset within that one (CLAUZON et CRAVATTE, 1985). One knows that such a setting cannot be explained by tectonics.

*The Pliocene tectonics.* Nevertheless, the Roussillon basin has not been spared by the Pliocene tectonic activity. In order to appreciate the undergone deformation, one disposes of a convenient reference—known as being stable during the studied time (VAIL and HARDENBOL, 1979)—the limit marine/continental. During the Tabianian, this limit has prograded over the whole length, today emerged, of the basin (CLAUZON et al., 1986). The curve of the Tertiary eustatic changes of sea level (VAIL and HARDENBOL, 1979) give a height of +80 meters to that reference level. Today, in the western part of the basin, the Tabianian coastal beds are laying between +130 and +200 meters while they are buried at −150 meters, right facing the shoreline (CLAUZON et CRAVATTE, 1985). So, we have an evidence that, during the Plio—Pleistocene, the basin has undergone a rocking motion: it went down around 250 meters to the east. The numerical ratio of the sedimentary accumulation (900 meters) and of the subsiding (250 meters) as well as the attenuation of these values westward, demonstrate that a loading subsidence is the prime cause of this movement.

*Age and geodynamics of the Pliocene filling up.* That loading subsidence is synchronous with the up lifting of the Albères horst, conjointly proved by the recurrent faulting of its northern accident and by the Villelongue dels Monts cyclopean breccia.

Still recently, this Roussillon Pliocene series was ascribed to the Plaisancian. Actually, it belongs only to the Tabianian stage (CLAUZON et CRAVATTE, 1985) of which it covers only the lower and middle parts since the Perpignan vertebrates beds, located at its roof, seem to be aged of 4 m.y. (MICHAUX et AGUILAR, 1985).

Thanks to the combined use of marine (planktonic Foraminifera) and terrestrial (rodents) pointers (CLAUZON et al., 1986), one has been able to demonstrate that:

— the marine transgressive levels, at the bottom of the Pliocene series, were isochronous and ascribed to the lowest Tabianian;

— the continental roof of the same series was also isochronous and dated of around 4 m.y.

Hence, between this two chronostratigraphic horizons, the building of the marine sedimentary prism, completed by the progradation of the alluvial piedmont that covers it (progradation carried out according to the west—east axis of the graben), have been realized within a lapse of about 1 to 1.5 m.y.

It follows that—from the geodynamical point of view—the Pliocene filling up of this basin has been determined: not by the volume generated by the subsidence of the graben (as such was the case during the Miocene) but by the volume eroded during the Messinian event and submerged at the outset of the Pliocene.

## Conclusion

The Neogene geodynamical evolution of the Roussillon basin records the three classical events recognized everywhere else in the French Mediterranean south:

— the compressive and orogenic phase of the Upper Miocene,

— the Messinian erosional surface,

— the filling up of the Pliocene rias.

The Roussillon originality does not lie in this ordinary sequence but in the previous rifting phase which, during the lower and middle Miocene, created this basin, stretched at right angle to the shoreline, allowing thus the further episodes to take advantage of an extraordinary recording.

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# LITHOSTRUCTURAL CHARACTERS OF THE ACOUSTIC BASEMENT OF THE SARDINIA CHANNEL (SOUTHWESTERN TYRRHENIAN SEA)

by

R. COMPAGNONI, E. MORLOTTI and L. TORELLI

The sea-floor of the Sardinia Channel (Fig. 1), located between the stable Sardinia block and the deformed belts of Sicily and Tunisia, is composed of several structural elements only partly affected by Neogene extensional tectonics (BATICCI et al., 1983; TORELLI et al., 1985). Therefore it can be considered an outstanding natural laboratory where to study and clarify the Tertiary Europe—Africa collisional history and the evolutionary stages of the adjacent Tyrrhenian basin.

The interaction of these geodynamic events is widely reflected by the depositional style of the sedimentary cover and by the complex lithostructural characters of the underlying acoustic basement considered here as a surface of maximum interpretable acoustic penetration. Both units have been extensively investigated since 1982 through the analysis of all the available seismic reflection profiles and interpretation of aeromagnetic data accompanied by a detailed sampling programme (Fig. 1) (BARBIERI et al., 1984; CATALANO et al., 1985).

We wish only to outline here the nature and characters of the sedimentary cover well described and contoured in a previous paper (BARBIERI et al., 1984). It generally unconformably overlies the acoustic basement and reaches a maximum thickness of about 2500—3500 m in the Upper Miocene subsiding basins of the Sicily continental slope as well as in the Cornaglia Terrace where a Middle Miocene rift axis was detected. When complete, the sedimentary cover includes a Plio—Quaternary unit, Messinian salts and evaporites and a pre-Messinian unit certainly not older than Early Miocene. Seismic reflection profiles across the Sardinia Channel (Fig. 1) show great variability in depth and characters of the acoustic basement, the features of which strictly depend on the Tertiary geodynamic evolution of the area. The seismic grid is not very dense and homogeneous and most information concerning the basement structures is provided by magnetic and sampling data which are used to interpolate between seismic lines and to produce a structural contour map (Fig. 2).

The seismic penetration is variable in many sectors depending on the quality of the processing as well as on the strong deformation of the subsurface. The acoustic and crystalline basement rarely coincide except in the southern margin of the Sardinia block where the acoustic basement surface reflects Palaeozoic metamorphic and igneous rocks (asterisks in Fig. 1). Furthermore along the axis of the Cornaglia Terrace, across a strongly block-faulted zone, a very thick Messinian basinal facies seems to overlie an acoustic basement characterized by a very high interval velocity and topped by a prominent reflector which may indicate an oceanic nature of the crust.

The acoustic basement between the Drepano—Aceste alignment and the Sicily—Sardinia Trough shows a diffraction pattern and a rugged topography typical of erosional processes in a subaerial environment. Sometimes it displays discontinuous, parallel and variable-amplitude reflectors which can be referable in the upper part to

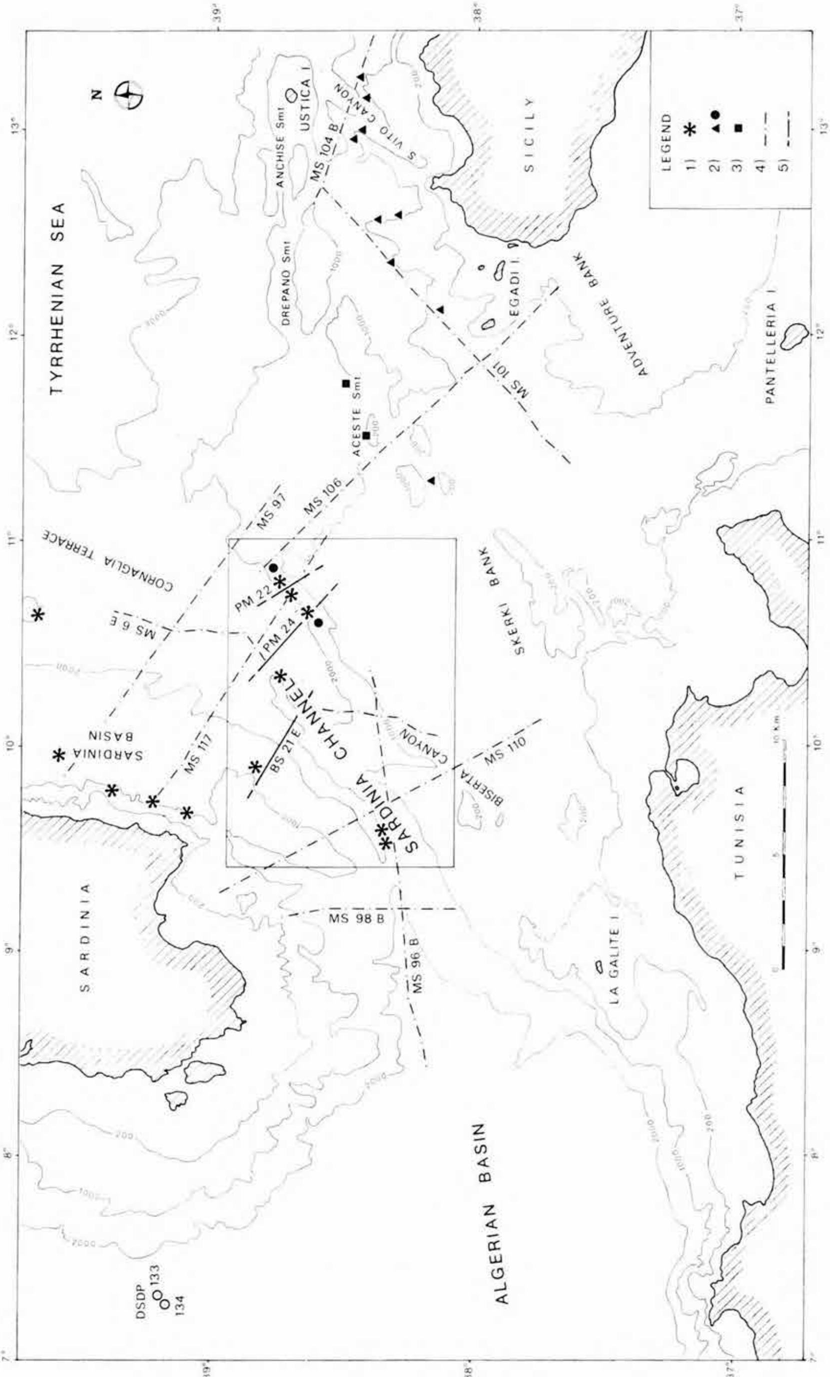


*Fig. 1.* Map of the investigated area with traces of main seismic profiles and location of dredging hauls
*1* Crystalline rocks, *2* carbonate rocks, *3* igneous rocks, *4* multichannel lines, *5* single-channel lines

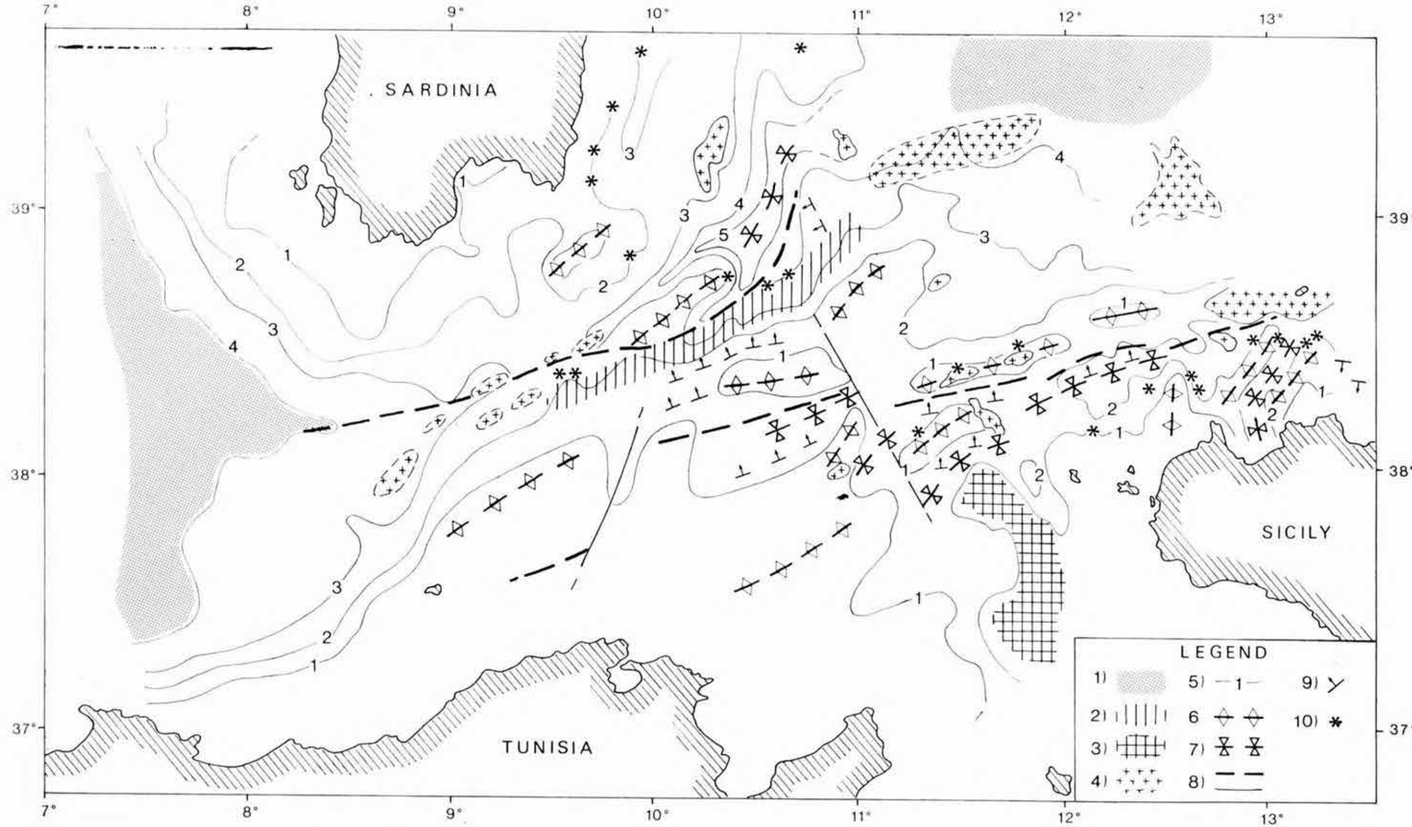


*Fig. 2.* Tectonic structural map of the acoustic basement of the Sardinia Channel

*1* Oceanic stretching zone, *2* low-grade Alpine metamorphic belt, *3* outcropping of the basement, *4* igneous rocks, *5* isobath of the acoustic basement in seconds, *6* axes of structural highs, *7* axes of structural lows, *8* major tectonic boundaries, *9* deepening of the basement inferred from magnetics, *10* dredging hauls a silicoclastic sedimentation. In fact dredge samplings of the basement in the western sectors of the central plateau led us to identify a contact between an Oligo—Miocene deep-sea clastic unit and crystalline—metamorphic rocks (Fig. 3) to form a stratigraphic superposition which can be traced to the Sardo—Tunisia district (Auzende et al., 1974).

Diffuse diffraction hyperbolas coupled with more continuous high-amplitude reflections characterize the seismic grain of the acoustic basement in the southern and southeastern sectors of the Sardinia Channel. Its direct sampling (Fig. 1) recovered also blackish shales and brown fine sandstones of the Numidian Flysch but mainly carbonate basinal lithologies, ranging in age from the Jurassic to the Palaeocene, well correlatable with the Panormide unit (part of the Maghrebian Africa-verging chain) widely outcropping on the mainland Sicily (Catalano et al., 1985).

The interpretations of the magnetic anomalies (Agip, 1981, 1982; Boccaletti et al., 1984) allow to provide depth estimates for the crystalline basement and to distinguish between its regional features and local shallow magmatic intrasedimentary bodies. Intense high-frequency anomaly patterns are localized along the Elimi Chain and are related to volcanic apparatus of the Ustica—Anchise complex and Aceste Seamount (Fig. 2). In this latter area the volcanic rocks are referable to the hawaiite-mugearite-trachyte transitional suite and are of Late Miocene—Early Pliocene age

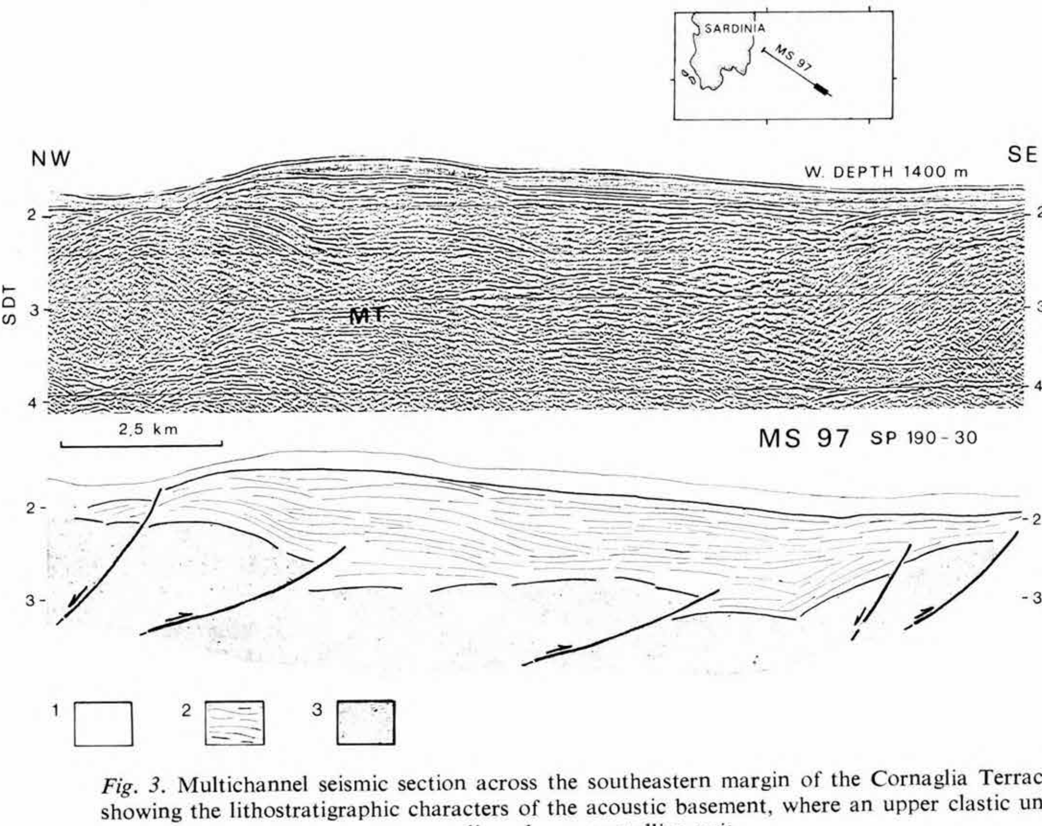


*Fig. 3.* Multichannel seismic section across the southeastern margin of the Cornaglia Terrace, showing the lithostratigraphic characters of the acoustic basement, where an upper clastic unit overlies a lower crystalline unit

*1* Plio—Quaternary, *2* Oligo—Miocene deep-sea clastic unit, *3* crystalline basement

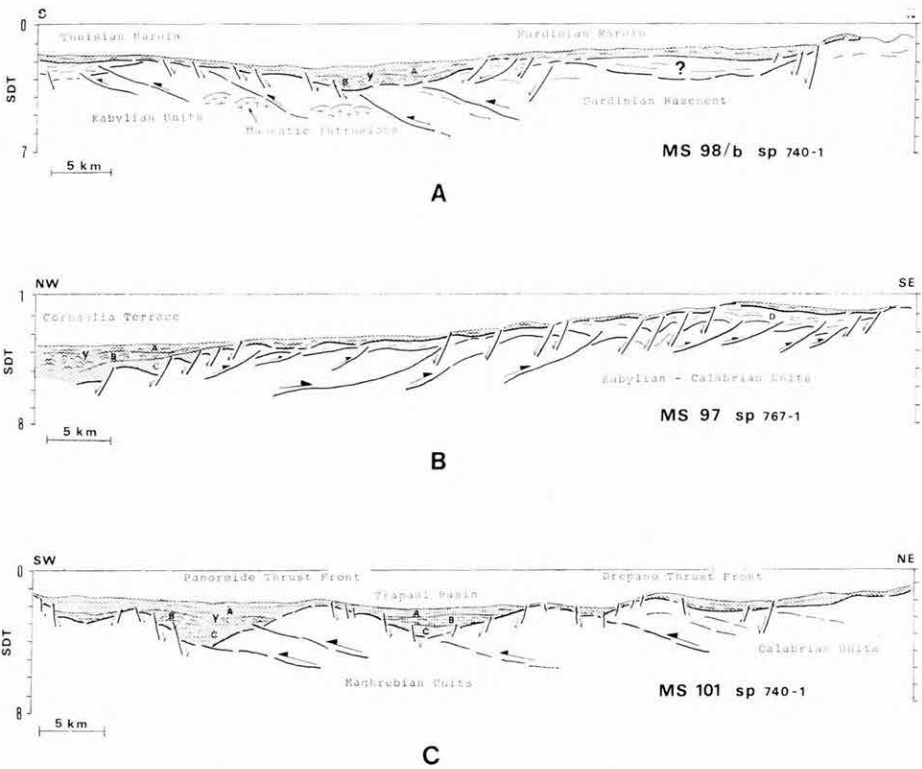


*Fig. 4.* Interpreted time-sections across the main thrust units of the Sardinia Channel. The stippled area represents post-orogenic cover

A Plio—Quaternary unit, B Messinian unit, C pre-Messinian unit. SDT: second double time. For location see Fig 1

(BECCALUVA et al., in press). Other local high-frequency anomalies are distributed to the south of the Sardinia block and may be caused by mafic intrusions along the trough connecting the oceanic sectors of the North Algerian basin and the Tyrrhenian bathyal plain. Nevertheless some igenous features of the Sardo–Tunisia district, which are interpreted on seismic reflection lines (Fig. 4), can be probably referred to Middle—Upper Miocene anatectic bodies, well-known and dated in La Galite Island (BELLON, 1981).

A medium deep basement (3–6 km), which becomes shallower in the Drepano Seamount and shows an acid–schistose crystalline character, appears to be responsible for the low-frequency anomalies recorded in the area. This magnetic pattern is well distinctive in the central sectors and the quantitative interpretations generally suggest a strong deepening of the basement towards the Sicily–Sardinia Trough (Fig. 2). It appears to be characterized by several block uplifts with ENE–WSW trending alignments to define tectonic slices of a complex thrust system mainly interpreted on the multichannel seismic sections.

As regards the composition the petrographic study of the crystalline basement revealed that a strong difference exists between the rocks coming from southern and northern scarps of the Sicily–Sardinia Trough (COMPAGNONI et al., in progress). The former ones mainly consist of medium-grade metamorphic and minor plutonic rocks with a peralouminous affinity, exhibiting a significant cataclastic to mylonitic deformation accompanied in most cases by the development of a rough foliation marked by low-grade metamorphic minerals. On the other hand, the Sardinia margin lithotypes are mainly granitoids and minor amphibolite-facies metamorphic rocks without any evidence of low-grade metamorphic overprinting (Carta Litologica e stratigrafica dei Mari Italiani, 1981; BORSETTI et al., 1979). Consequently a sharp difference may be detected between the Sardinia margin lithologies and the southeasternmost rocks here referred as belonging to the Kabilian—Calabrian palaeogeographic realm deformed in high-strain environment during the Upper Oligocene—Burdigalian eastward migration of the Corsica—Sardinia block (CHERCHI and MONTADERT, 1982). It is to point out that a similar low-grade metamorphic overprinting (Fig. 2) of probable Alpine age (younger than 30 m.y. ?) has been recently described in the Southern Calabria (Aspromonte units; BONARDI et al., 1984) and in Eastern Algeria (Little Kabilíe; BOUILLIN, 1982).

In summary the acoustic basement of the Sardinia Channel can be divided in three lithostructural domains (Fig. 2) on the basis of seismic grain, magnetic patterns and lithostratigraphic data. The boundaries between the above mentioned domains are interpreted on seismic profiles as low-angle major overthrusts, one of which is probably related to a Tertiary subduction zone (TORELLI et al., 1985). The southern margin of the Sardinia block superimposes on the intermediate Kabilian—Calabrian fold/thrust belt, which in turn overlies the Maghrebian units. The two thrust fronts are dissected by a set of strike-slip dextral faults which acted in connection with the development of the Upper Miocene—Lower Pliocene oceanic spreading of the Tyrrhenian Sea and appear to be responsible for the formation of the eastern pull apart basins and for the intense break up of the acoustic basement in the central sector. The tectonic framework so far defined can be clearly seen in the interpreted time sections of Fig. 4, where the SE-verging Kabilian—Calabrian and Maghrebian units are sandwiched between the Sardinia block and the Sicilian foreland as a result of crustal shortening phases which took place in Late Oligocene—Pliocene times.

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# SEDIMENTATION IN THE STRIKE-SLIP NORTH ANATOLIAN FAULT ZONE, THRACE, TURKEY

by

T. ERKAL

*Introduction.* The aim of the study was to describe and interpret the aspects of neotectonic history of European Turkey in terms of stratigraphical, structural and geomorphological data. The sedimentological observations relevant to the neotectonic history of the western end of the northern branch of the North Anatolian fault zone were made from the Neogene—Quaternary sediments.

The study area was chosen for investigation because it contains well exposed sequences deformed by structures and a range of geomorphological phenomena related to the fault zone. Mesoscopic scale fractures, tectonic landforms and sedimentary aspects of the Neogene—Quaternary succession were investigated in detail.

*Structure and tectonic setting.* The map shows the location of the study area which is situated in Thrace on the European side of the Sea of Marmara (Fig. 1). The ENE—WSW trending North Anatolian fault zone is boundary between the Anatolian and Black Sea plates as an intracontinental transform (HANCOCK and BARKA, 1980, 1981). The fault zône, which is represented by a single zone, is divided into three branches in the region (see Fig. 1). The northern one crosses southern Thrace and divides the region into two structural and morphological units. To the N of the main trace there is Ganos Mountain underlain by folded Eocene—Oligocene flysch (PLATT, 1959). High elevations to the N may be result of there having been overthrusting adjacent to the Sea of Marmara (ŞENGÖR, 1979, ŞENGÖR, et al., 1982). The North Anatolian fault is vertical beneath the sea (KAVLAKOÇLU and ÖZAKĞAY, 1973) and it forms a rift in the northern Aegean Sea (MALEY and JOHNSON, 1971).

*Stratigraphy.* Figure 2 shows the stratigraphy. The lowest part of the Miocene units uncomformably overlies pre-Neogene rocks. The Multicoloured (MC) member which is of early Miocene age (ÜNAY and DE BRUJIN, 1984) consists of clays of different colours at bottom and it passes upwards to siltstone and claystone. Miocene aged Iğdebağlar (I) member (ÜNAY and DE BRUJIN, 1984) is mainly composed of sandstone and interfingers with the overlying and underlying members of Kirazli and Multicoloured respectively (see Fig. 2). The Kirazli (K) member which is of middle Miocene age (ÜNAY and DE BRUJIN, 1984) overlies the I member and also interfingers with the underlying members of Iğdebağlar and Multicoloured. The uppermost part of the Miocene sequence (Şarköy, Ş member) consists of siltstone and claystone. Sediments deposited during late Pliocene—early Pleistocene time interval is mainly composed of oolithic limestone. Late Pleistocene to Holocene is represented by marine terrace gravels with marine fauna (NUTALL, 1982 pers. com.), alluvial fan, landslide and coastal plain deposits.

*Facies and environments of deposition.* Four main lithofacies have been recognised within the Miocene sediments (Fig. 3).

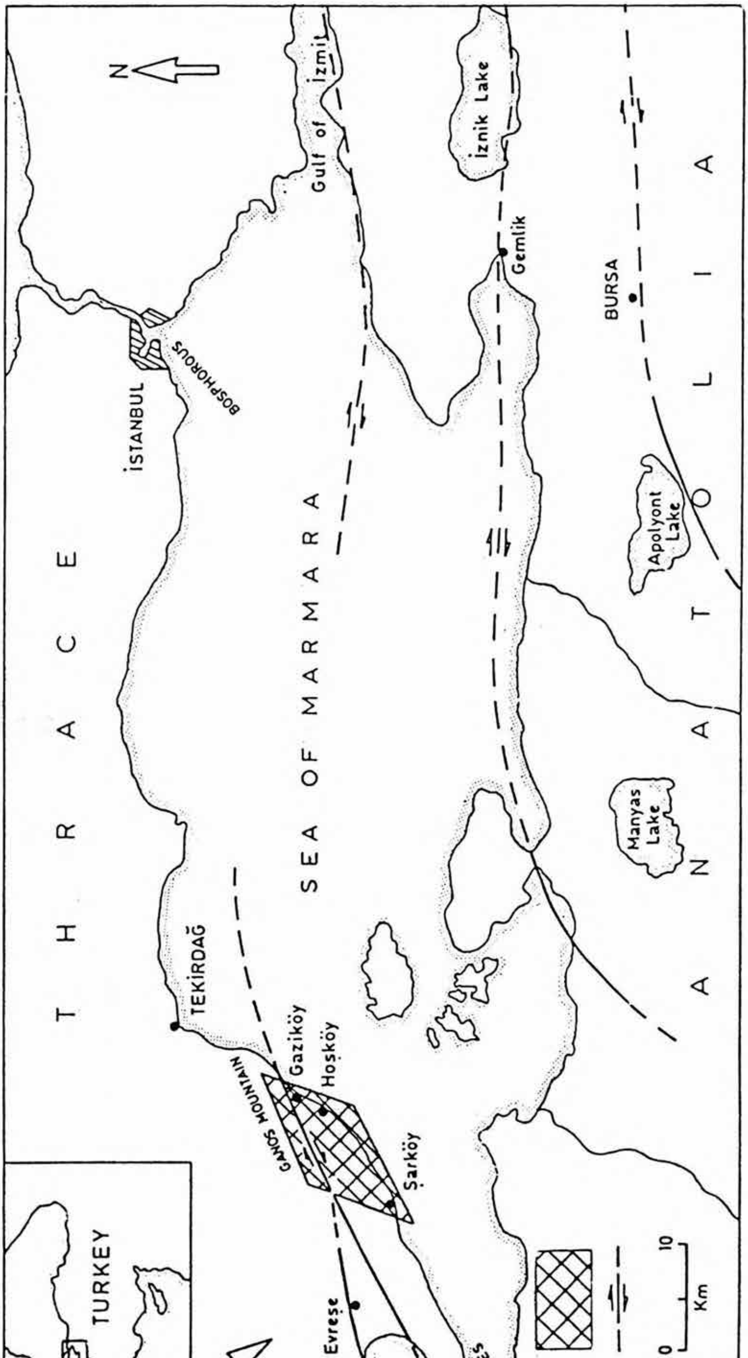


*Fig. 1.* Location map of the study area

*1* Study area, *2* principal branches of the N Anatolian fault, pecked where uncertain

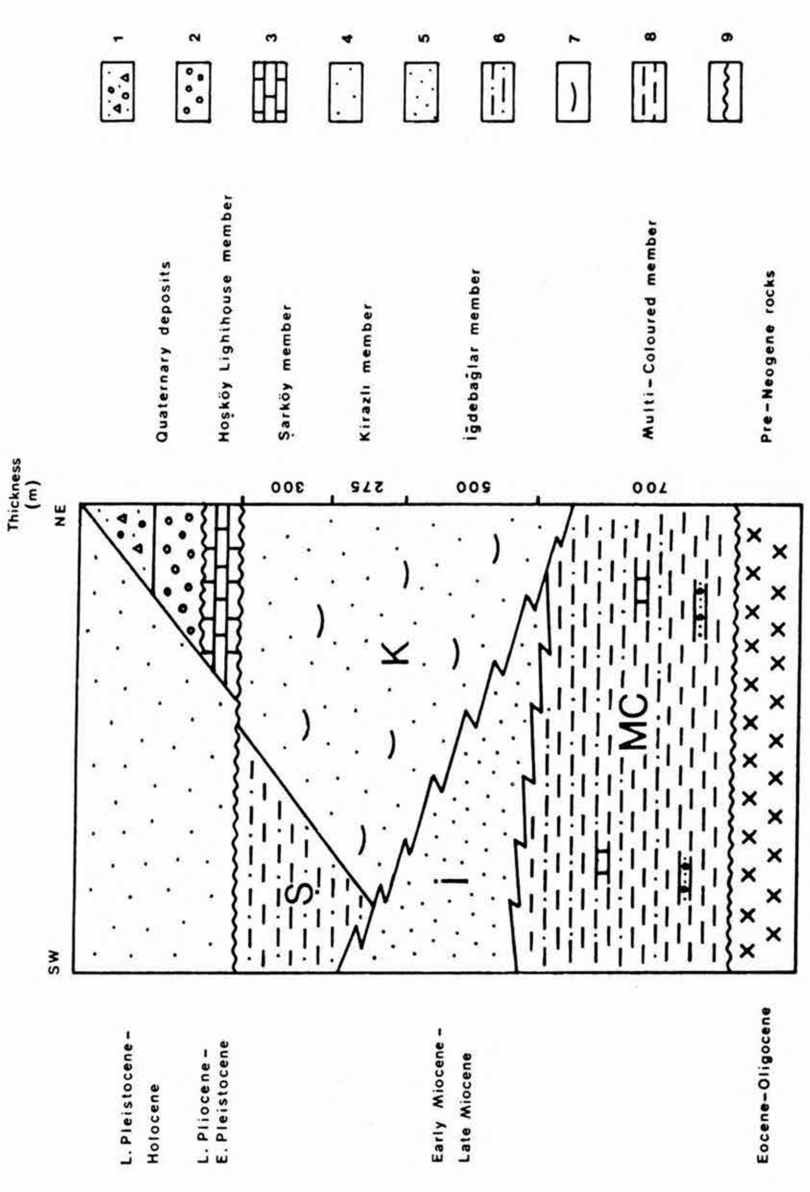


*Fig. 2.* Synoptic diagram showing relationship between Miocene—Quaternary facies units of the study area
*1* Colluvium, *2* gravel, *3* limestone, *4* sand, *5* sandstone, *6* siltstone, *7* cross-bedding, *8* clay/claystone, *9* unconformit

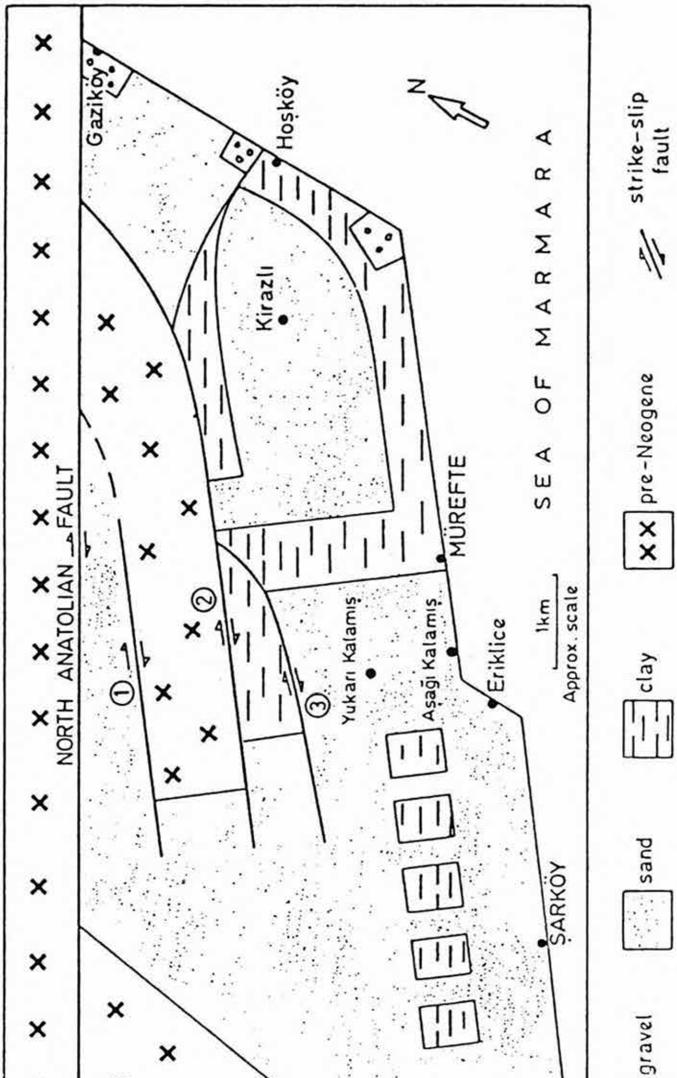


*Fig. 3.* Schematic map of major lithofacies in the study area (from ERKAL 1983. Fig. 3, 5)
*1* Çengelli fault, *2* N Tepeköy fault, *3* S Tepeköy fault

*1* Massive to stratified sands form the uppermost part of the sequence throughout the study area and are poorly cemented. They are medium-grained and white to yellow and greyish in colour. Sands can be divided into four subfacies:

*a)* Massive sands with some pebbles. They were probably deposited in a shallow marine environment with a strong bioturbation.

*b)* Parallel laminated muddy sands were probably laid down in a low energy environment such as a lagoon open to the sea (ERKAL, 1983).

*c)* Planar cross-bedded sands are interpreted as migrating straight-crested mega-ripples formed by unidirectional currents (JOPLING, 1965; ALLEN, 1966; COLEMAN and WRIGHT, 1975).

*d)* Trough cross-bedded sands were formed by unidirectional currents moving sinuous-crested mega-ripples (REINECK and SINGH, 1980; LEEDER, 1982).

*2* Clays are found next to the main trace and major faults. They are also wide-spread in the central part and are overlain by sands. The clays are thinly and parallel laminated. Two clay subfacies are recognisable:

*a)* Ostracod-bearing grey to green clays.

*b)* Multicoloured clays with ostracod, plant and fish remains. Ostracod-bearing grey to green coloured clays are indicative that the environment was a lagoon (DRUITT, 1959). Multicoloured clays are thought that the deposition occurred in a warm, lacustrine environment (DRUITT, 1959).

*3* Coal is rare and appears as lenses of variable thickness with a maximum thickness of 80 cm. Coal lenses suggest that they may have accumulated *in situ* (ERKAL, 1983).

*4* Limestones are in two different ages which occur in Miocene deposits enclosed by grey to green clays and are overlain by Quaternary gravels. Both of them are oolithic and micritic. Limestones are compact, well bedded and white coloured. Plio—Pleistocene aged limestones are of limited extend and no more than 2 m thick. It is interpreted as that limestones were probably deposited in a low energy environment, the scattered oolithes in limestones perhaps having been wind blown from neighbouring coastal dunes or having been transported by longshore currents (HAMILTON, 1982 pers. com.).

Marine terrace gravels are another facies that is of Pleistocene age (NUTALL, 1982 pers. com.). They comprise sandstone and limestone gravels which are subrounded to wellrounded. Gravels contain marine fauna.

*Interpretation.* Blockdiagrams show the palaeogeography and palaeotectonics in the study area (Figs. 4—7).The deposition commenced with multicoloured clays near the faults and at the bottom of the Miocene basin. It is suggested that they were laid down in a subaqueous environment such as a lake (Fig. 4). Grey to green clays above the multicoloured clays are indicative that deposition occurred in a lagoonal environment due to rapid deposition (Fig. 5); the lagoon resulted from a brief marine transgression over the former lake (Fig. 6). The deposition of both massive sands with pebbles and parallel laminated muddy sands around the margins of the lagoon resulted from progressive migration of the shoreline infilling the lagoon (see Fig. 6). Planar cross-bedded and trough cross-bedded sands were deposited in nearshore and alluvial environments of the study area. The deposition probably resulted from the activiation of the North Anatolian fault zone (see Fig. 6). During the late Pliocene—Pleistocene, limestones were laid down in a coastal zone similar to that of present topography. Marine terrace gravels were deposited by the interglacial transgression in Pleistocene (Fig. 7).

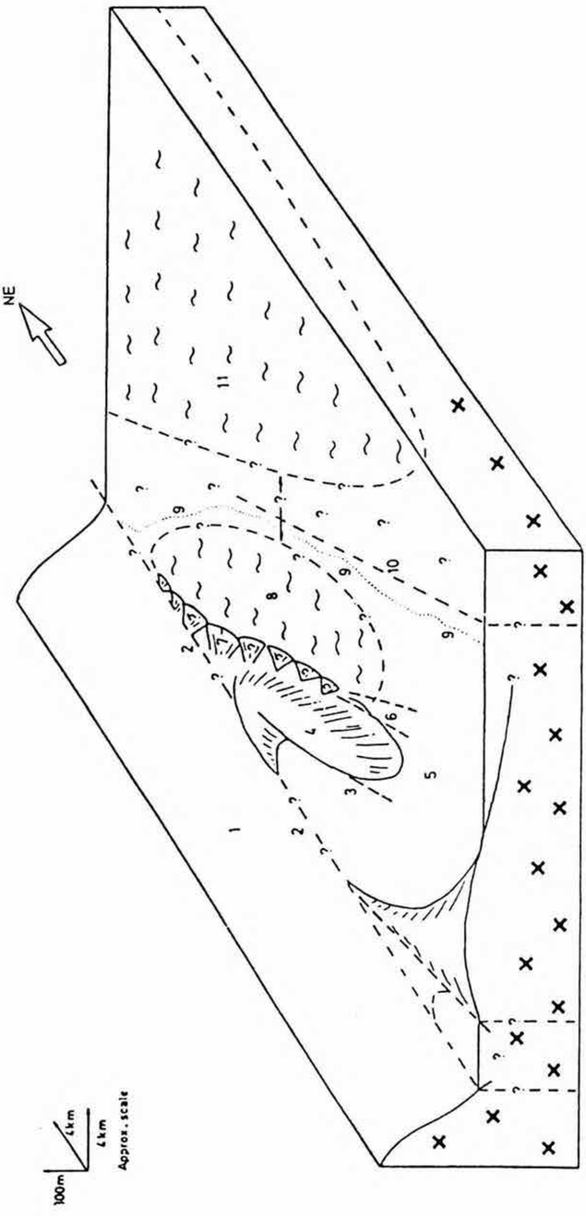


*Fig. 4.* Schematic reconstruction of Early—Middle Miocene palaeogeography and palaeotectonics in the Gaziköy—Sarköy area

*1* Ganos Mountain, *2* N Anatolian fault, *3* Çengelli fault, *4* pre-Neogene inlier, *5* Iğdebağlar member, *6* N and S Tepeköy faults, *7* Multicoloured member, *8* shallow lake, *9* present coastline, *10* Mürefte fault, *11* open sea

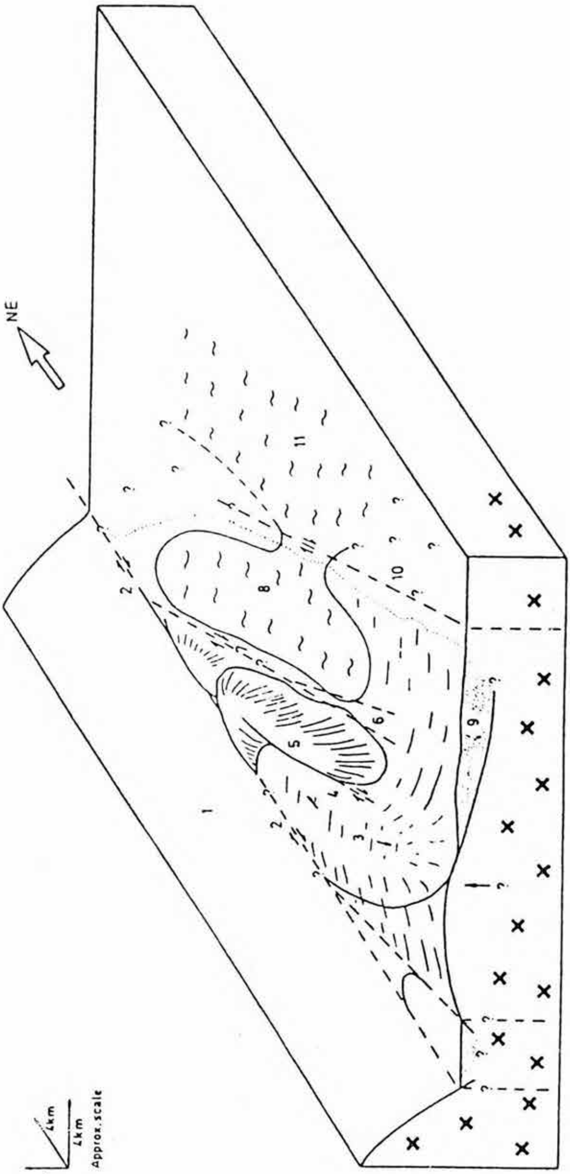


*Fig. 5.* Schematic reconstruction of Middle—Late Miocene palaeogeography and palaeotectonics in the Gaziköy—Şarköy area

*1* Ganos Mountain, *2* N Anatolian fault, *3* erosion surface overlain by Iğdebağlar member, *4* Çengelli fault, *5* pre-Neogene inlier, *6* N and Tepeköy faults, *7* Multicoloured member, *8* lagoon, *9* Iğdebağlar member, *10* Mürefte fault, *11* open sea

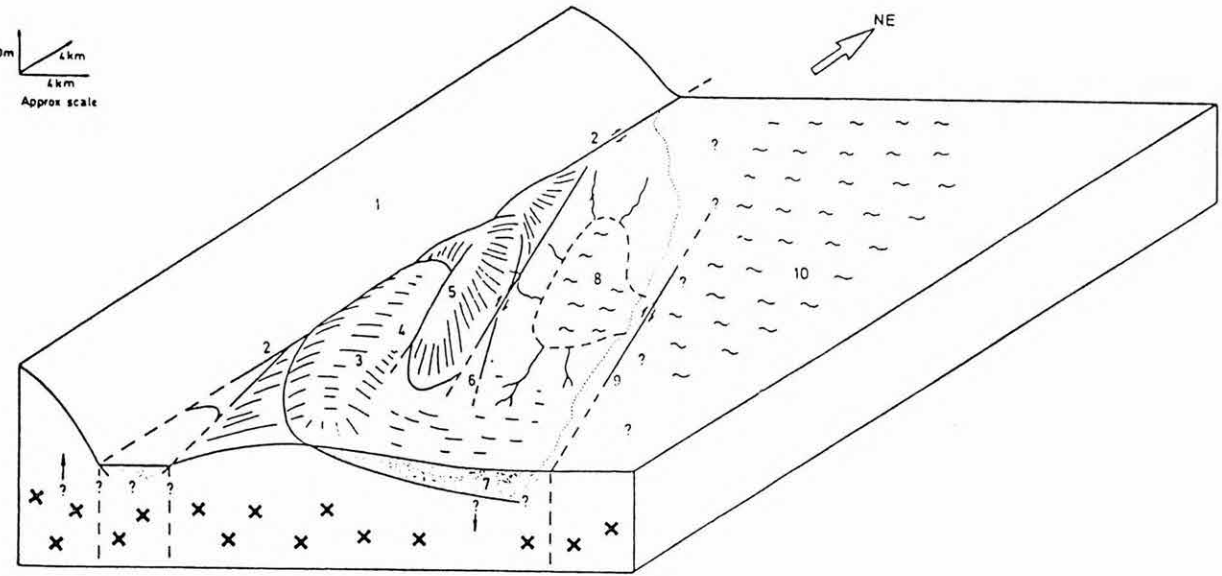


*Fig. 6.* Schematic reconstruction of Late Miocene palaeogeography and palaeotectonics in the Gaziköy—Şarköy area

*1* Ganos Mountain, *2* N Anatolian fault, *3* erosion surface overlain by Iğdebağlar member, *4* Cengelli fault, *5* pre-Neogene inlier, *6* N and S Tepeköy faults, *7* Iğdebağlar member, *8* lagoon during deposition of the Kirazli member, *9* Mürefte fault, *10* open sea

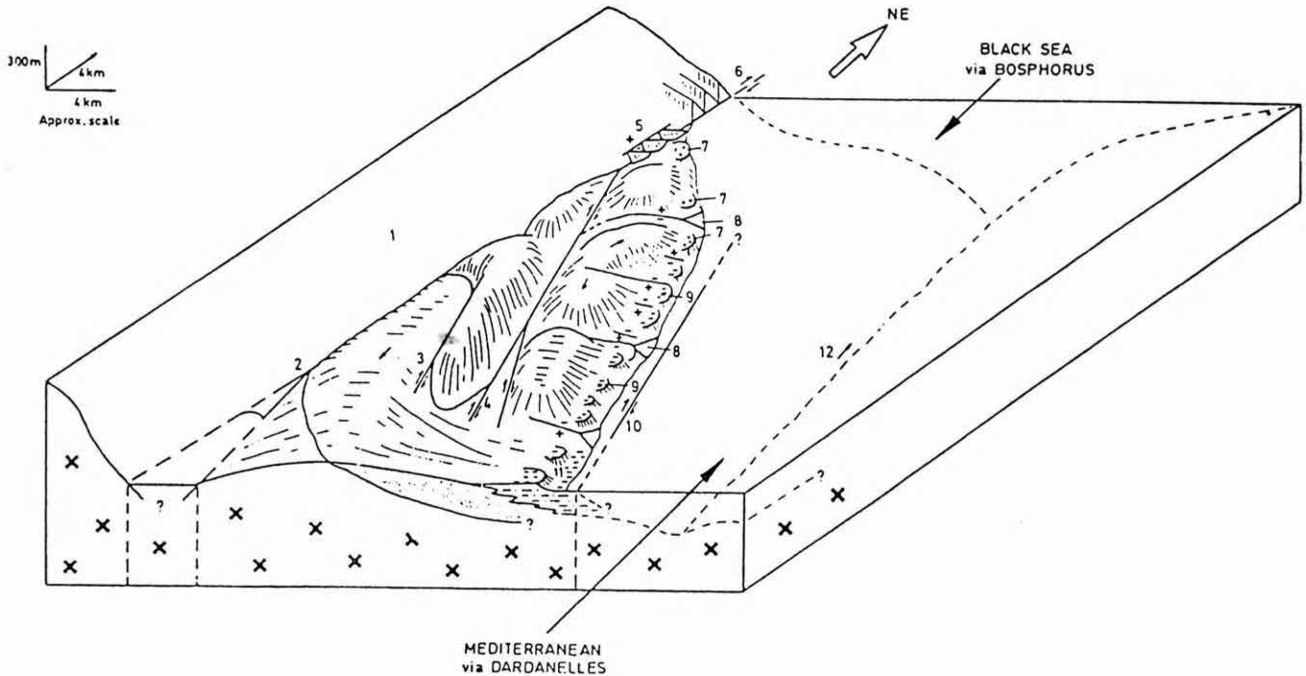


*Fig. 7.* Schematic reconstruction of Middle—Late Pleistocene palaeogeography and palaeotectonics in the Gaziköy—Şarköy area

*1* Ganos Mountain, *2* N Anatolian fault, *3* Çengelli fault, *4* N and S Tepeköy faults, *5* alluvial fans, *6* triangular facets related to strike-slip movement on the N Anatolian fault, *7* marine terraces, *8* deltas, *9* terrestrial (erosional) terraces, *10* Mürefte fault, *11* Kovalik coastal plain, *12* fluvial system during glacial episods

*Conclusion*. There is a little evidence such as the occurrence of synsedimentary mesofaults showing that the study area was tectonically active but the distribution of facies and thicknesses being unrelated to fault geometry and palaeocurrent directions indicate that there is no direct relationship to fault trend. Finally it can be said that sedimentation does not appear to have been controlled by strike-slip displacements on the North Anatolian fault zone.

*Acknowledgements*. I would like to express my thanks to Mr. M. KARABIYIKOĞLU of M.T.A. for his constructive comments, Dr. D. HAMILTON of the Department of Geology in the University of Bristol for the interpretation of limestone specimens and Mr. P. NUTALL of the Department of Palaeontology of the British Museum for his assistance of mollusc identification. This study forms a part of postgraduate study which was carried out under the supervision of Dr. P. L. HANCOCK of the University of Bristol.

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# GENETIC MODEL OF POST-SARMATIAN SEDIMENTATION IN THE GREAT HUNGARIAN PLAIN

by

J. GEIGER and I. RÉVÉSZ

The objects of this work are the Lower Pannonian clastic formations of the Great Hungarian Plain. Information has been obtained from 225 boreholes situated along the presented sections (Fig. 1).

The clastic rocks of the Great Hungarian Plain have been classified into eight formations on the basis of their lithological features and depositional characteristics (GAJDOS et al., 1983). These characteristics made general environmental interpretation possible years ago (Fig. 2). It has become known that the Early Pannonian transgression was succeeded by a delta accumulation arriving from different directions and the connecting environments of this delta. The turbidite environments of the deeper parts also produced a number of rocks bodies (RÉVÉSZ, 1984; BÉRCZI et al., 1984).

In order to reconstruct in detail the process of sediment accumulation in the basin, the revealing and mapping of the rock bodies developed in the same depositional environment. The areal distribution of the genetically equivalent rock bodies can be studied by using the palaeo-geomorphological means. As it is known "depositional

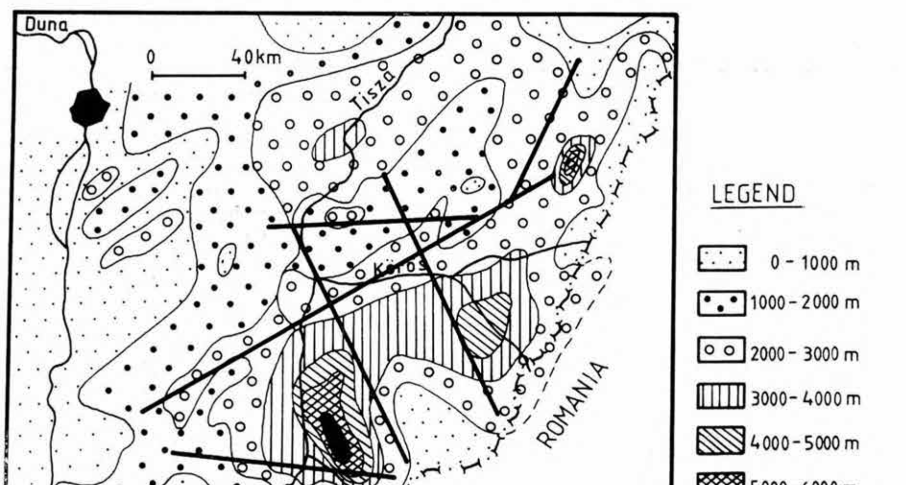


*Fig. 1.* Isopach map of the Pannonian (s.l.) deposits of the Great Hungarian Plain (after GEOS cooperative 1984)

**General information about the clastic Lower Pannonian (s. l.) formations (after RÉVÉSZ, 1984 and BÉRCZI et al. 1984)**

*Table 1*

| Formation | Rock types | Sedimentary features | Main genetic statement from core analyses |
|---|---|---|---|
| TÖRTEL | Alternating fine, medium and coarse sandstone, siltsone, argillaceous marl | Finning upward sequences. Ripple cross lamination, horizontal lamination, cross bedding etc. | DELTA PLAIN DEPOSITS |
| ALGYŐ | Sandstone, siltsone and argillaceous marl | Coarse upward seuqences. Ripple cros lamination, horizontal lamination, cross bedding | DELTA FRONT DEPOSITS |
| SZOLNOK | Various sandstone and argillaceous marl | Bouma divisions, bioturbation | DEPOSITS OF MASS GRAVITY FLOWS |
| NAGYKÖRŰ | Marl | Brownish gray, massive in some places pyritic concretion | |
| VÁSÁRHELY | Argillaceous marl, fine and coarse siltsone, fine sandstone | Bouma divisions | |
| DOROZSMA | Pebbly marl | Pyritic concretion, alternating pebble-poor and pebble-rich layers no grain orientation | TRANSPORTATION IN TURBIDITY CURRENTS |
| TÓTKOMLÓS | Calcareous marl | Pyritized root remains | SEVERAL REDUCING ENVIRONMENTS |
| BÉKÉS | From pebbly siltsone to conglomerates. Pebble. Lumachelle | Root remains, no grain orientation, fining upward sequences | COASTAL DEPOSITS |

environment" is a spatial term. So the spatial position of each formation is characterized by five independent parameters (Fig. 3). In this way, each borehole is represented by these during the examination of a given formation. Moreover, these parameters regarded as vectors arrange the boreholes of the tested formations in the sample space, too. Consequently the morphological systems of the formation can be revealed by a Q-mode cluster analysis of the sample space (Fig. 3). The sample classes defined in the dendogram will be exactly the morphogenetic units wanted. Each sample class contains the rock bodies formed by the same exogenous forces i.e. the rock bodies of the same depositional environment. In order to interpret the sedimentological content of these units the R-morde facto analysis of the parameters has been used. The factors are traced back to independent geomorphological processes. It has been proved a

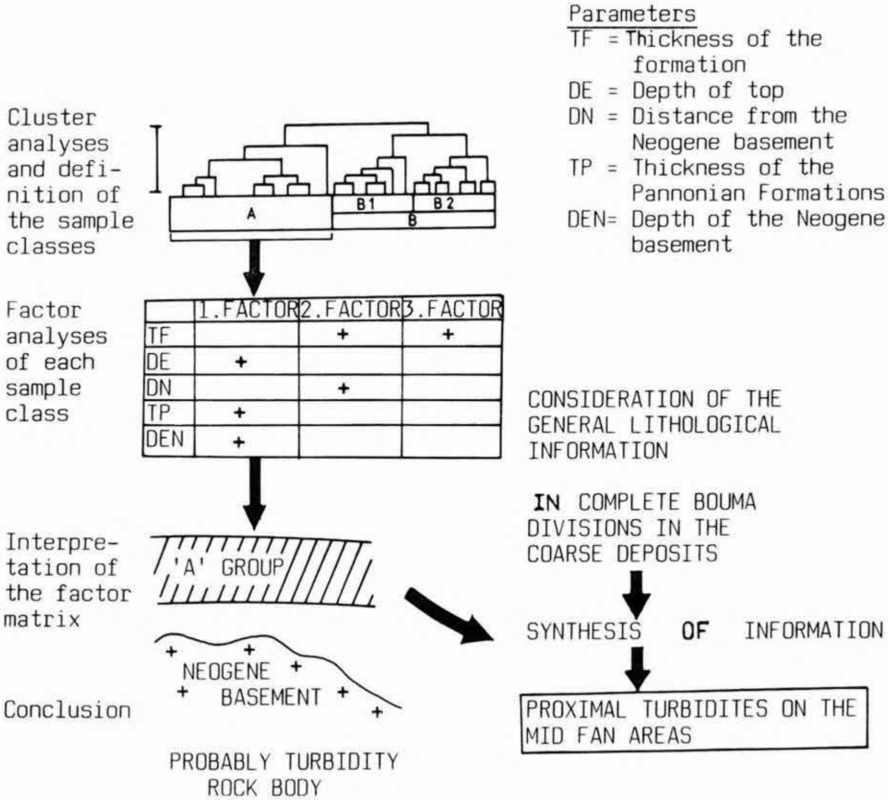


*Fig. 2.* The analysing system (after GEIGER, 1985)

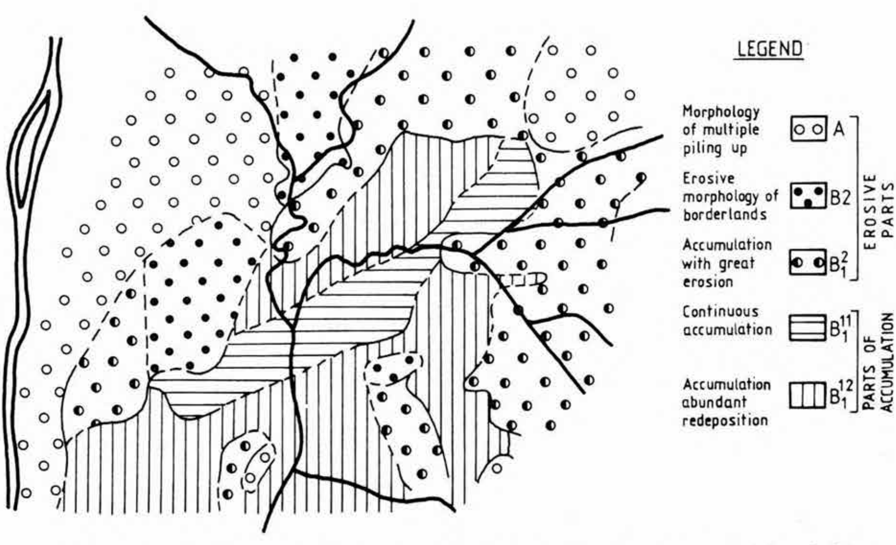


*Fig. 3.* Parts of accumulation and erosion of the Törtel Formation (delta plain)

theoretically that the effects of depositional environment, tectonism and compaction will be separated from each other. In the presented example on Fig. 3 the content of the factor matrix belonging to the defined sample class can be interpreted as the morphology of the subaqueous mass-gravity flows. If this is proved by the information provided by traditional sedimentological examination of the core samples, this concept of turbidity rock body can be extended to a large scale of rock bodies of this genetic type (GEIGER, 1985).

In this way, in the morphogenetic system of the Törtel Formation the rock bodies of the delta plain reworked by multiple progradation could be identified (Fig. 4). So one can obtain a theoretical succession of the areas of accumulation and erosion (Fig. 5). This pattern proves the existence of probably three progradational cycles during the basin accumulation (Fig .5).

The delta plain formation is characterized by the predominantly coarsening upward sequences of sandstones, various siltstones, argillaceous marls and lignite (Fig. 2). The percentage of sandstones is about 50%. The main sedimentary structures are root fragments in vertical position, quartz pebbles in the siltstones, cross bedding, ripple cross lamination in sandstones and coarse siltstones.

In the delta plain formation the boundary of a local base level can be clearly marked out (Fig. 4). This area belongs to the delta plain of the latest progradational cycle of course, every new progradation reworked the remains of the previous one. Consequently, in the rock bodies of the first cycle situated on the borderlands the effects of all three cycles can be recognized.

In the delta front formation different morphogenetic units belonging to different water depths could be identified (Fig. 6). In this way the direction of downward transport can be marked, too. The areal repetition of the same morphogenetic unit expresses the delta progradations. The textural characteristics are the following (Fig. 2). The predominant rock type is siltstone. The percentages of sandstone and argillaceous marl are very similar, about 15—15%. Convolutions, siltstone intraclasts, sliding marks are widespread in the formation. Some of the siltstone intraclasts are red coloured. In the sandy sequences the average grain size becomes finer and finer upwards, but the inverse process is also encountered.

The above mentioned delta plain and delta front formations are widespread in the Great Hungarian Plain.

The rock bodies of the prodelta area and the other parts of the foreland can be classified into several formations. The analysis has proved that these belong two main genetic units. The first one is controlled by the delta accumulation (Fig. 7). This group has developed fan type rock bodies displaying the upper, mid or lower fan morphology depending on the previous morphological evolution of the given area and the amount of the sediment transport. Farther from the prodelta area the deeper water fan type rock bodies are situated (Fig. 7). These have developed independently from the delta system. This group represent the second main genetic unit.

The prodelta fan type rock bodies are charakterized by decrease of the sandstone siltsone ratio. Marl and argillaceous marl are the predominant rock types. The Bouma divisions are characteristics. The facies of thin bedded turbidites suggested by these characteristics.

The deep water fan sedimentation has the same features, but the sandstone—siltstone ratio is smaller than in the previous case.

Beneath this fan type sedimentation marl, argillaceous marl deposited during the early Lower Pannonian transgression can be revealed. Some parts of these rock bodies have lower fan features.

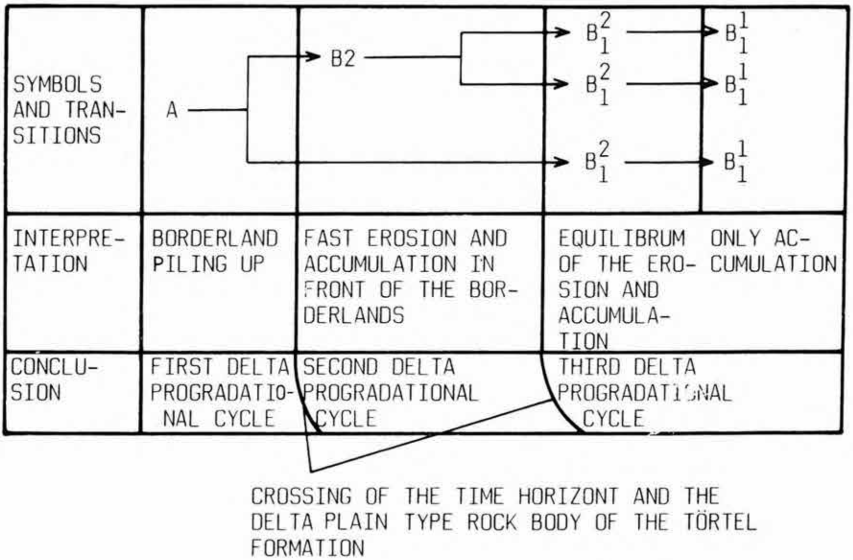


*Fig. 4.* Possible transport directions within the Törtel Formation

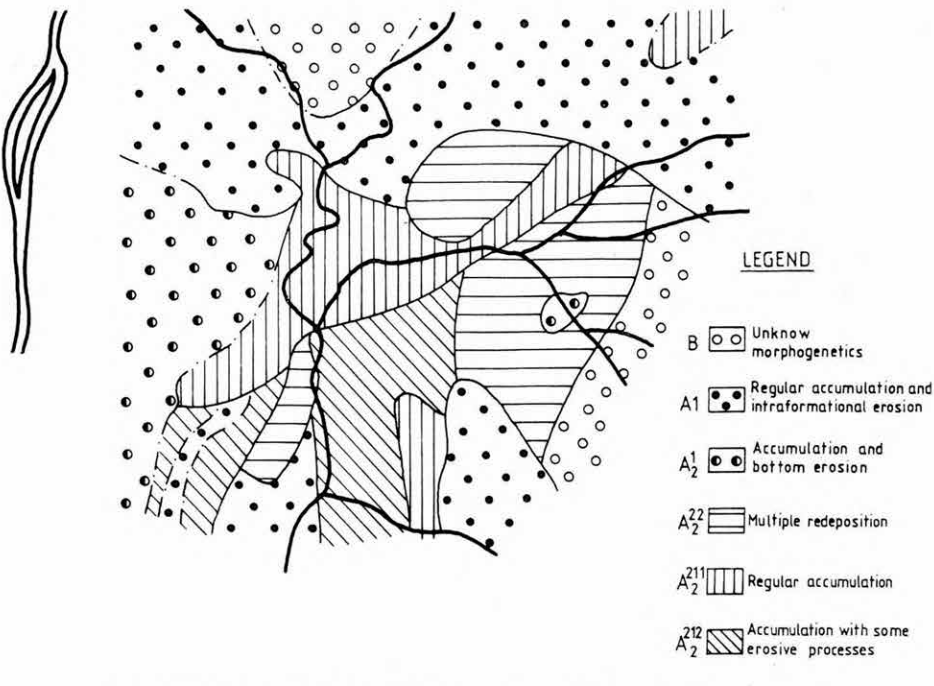


*Fig. 5.* Morphogenetic development of the Algyő Formation (delta front)

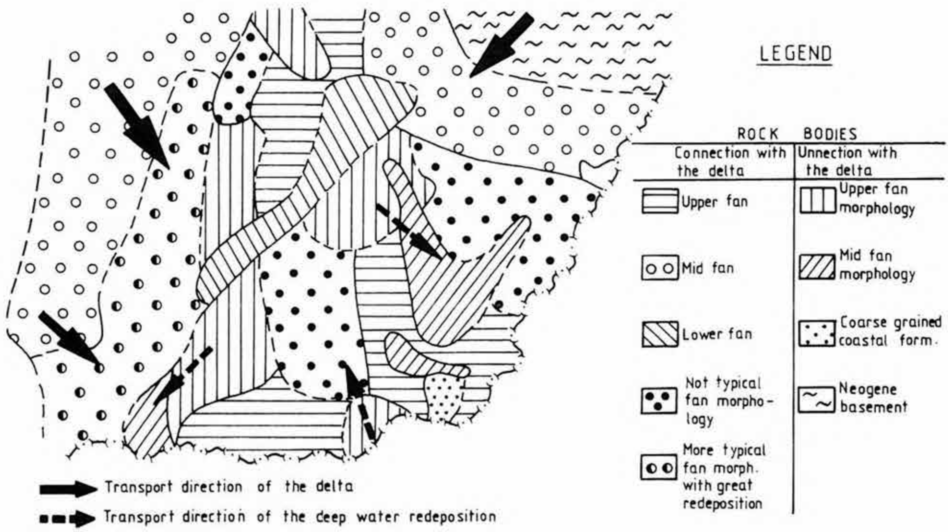


*Fig. 6.* Genetic system of the bottom formations of the delta front

**Summary**

*1* In the Great Hungarian Plain the Early Pannonian transgression is expressed by the deposition of marls and argillaceous marls. The deposition of the deeper water turbidity fans started at that time, too.

*2* During the Early Pannonian age the sedimentation of some deltas arriving from different directions can be identified.

*3* This delta accumulation consists of three progradational megacycles from the NNW, NW and W. During the time of the first progradation the delta plain, delta front and the fan type sedimentation of the prodelta areas and the deeper water fan deposits developed side by side. This evolution can be discussed as an analogy of Lake Superior (NORMARK and DICKINSON, 1976). The rock bodies of the delta front and the prodelta area developed during the time of the third progradation probably can be found in the north Yugoslavian Neogene basins.

*4* In a given progradational cycle the average grain size of the different rock types decreases from the delta front towards the deeper water fan of the prodelta area. The same tendency can be seen in the changing of the sandstone-siltstone ratio, too.

*5* The whole morphogenetic system of the Early Pannonian sedimentation can be revealed by using the presented method. This computerized system makes possible the automatical identification of the new data.

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# NEOTECTONIC AND PALAEOMAGNETIC RESULTS FROM NEOGENE BASINS OF MACEDONIA (N GREECE) AND THEIR GEODYNAMIC IMPLICATIONS

by

S. B. PAVLIDES and D. P. KONDOPOULOU

*Introduction.* Geological studies of both the alpine and neotectonic history of the Hellenides and the Aegean area, as well as geophysical and geodynamical ones, provide important information about the geological evolution of this area. The Tertiary and the Recent evolution in Greece is governed by the local microcontinental collisions and also influenced by distant events related to the active collision of African and European plates.

As far as the Neogene and Quaternary stress patterns are concerned, the detailed neotectonic studies of MERCIER et al., 1979, ANGELIER 1979, LYBERIS 1984 etc, have provided enough information about the Hellenic arc, the south Aegean and occasionally the north Aegean and the mainland of northern Greece.

An important point that may require further clarification is that the neotectonic and seismotectonic peculiarities of the north Aegean and the surrounding area can not easily be explained by the up to day suggested geodynamic models. Thus, for a deeper understanding of the geodynamic evolution of this area, it is necessary to determine the stress pattern of the Neogene—Quaternary times and to distinguish among different phases of rotational deformation. Palaeomagnetism is an important tool in the attempt to clarify the details of these events giving a quantitative technique for measuring relative displacement and rotations among different geological times, because palaeomagnetic measurements are generally made on small rigid block materials which have not been deformed and which are bounded by faults.

In this work we present new data concerning the stress pattern and palaeomagnetic directions of the Neogene basins in Macedonia (N Greece), alongside with similar data already published by others. A correlation is attempted between these sets of data in order to establish a possible geodynamic pattern during the "neotectonic stage".

## Methods

Methods used in the present tectonic analysis were the "right-dihedrons" and "the mean stress tensor" (ANGELIER, 1979). The chronology of fault events was established mainly by using the stratigraphical data of the sediments of the basins, as well as by taking into account criteria for successive fault motions.

Our detailed neotectonic analysis of Neogene and Quaternary fault mechanisms, as well as those published by MERCIER, 1981; LYBERIS, 1984; concerning the northern Greek mainland, enable us to characterize the direction of the regional stresses in the whole under investigation area (Fig. 1 and 2).

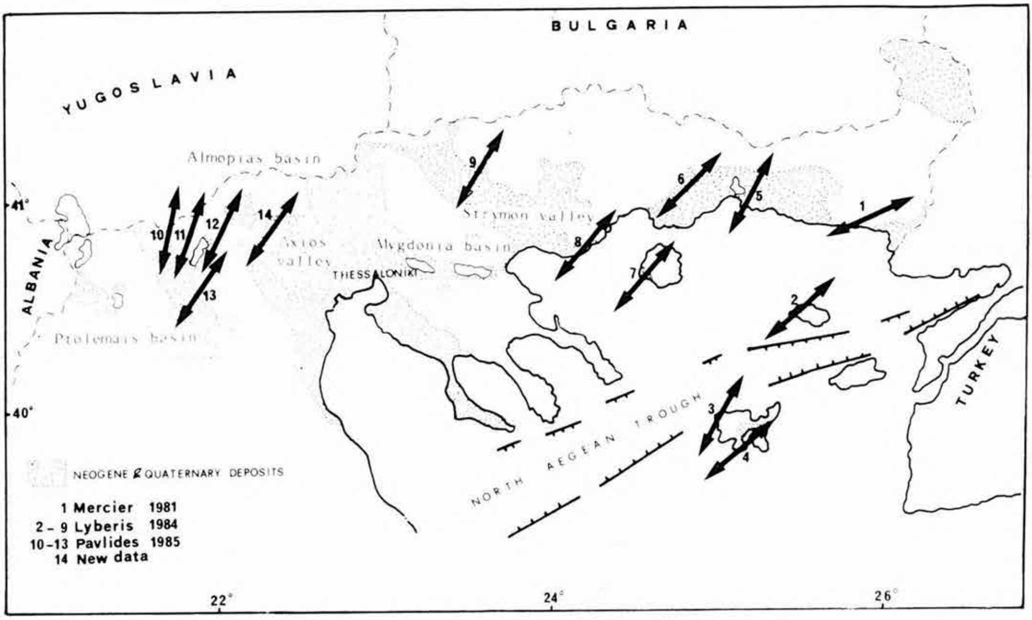


*Fig. 1.* A general picture of the investigated area (Macedonia and Thrace) in northern Greece with its Neogene and Quaternary deposits, which mainly fill up neotectonic basins. Arrows indicate the directions of tension for the Late Miocene—Pliocene period of extension

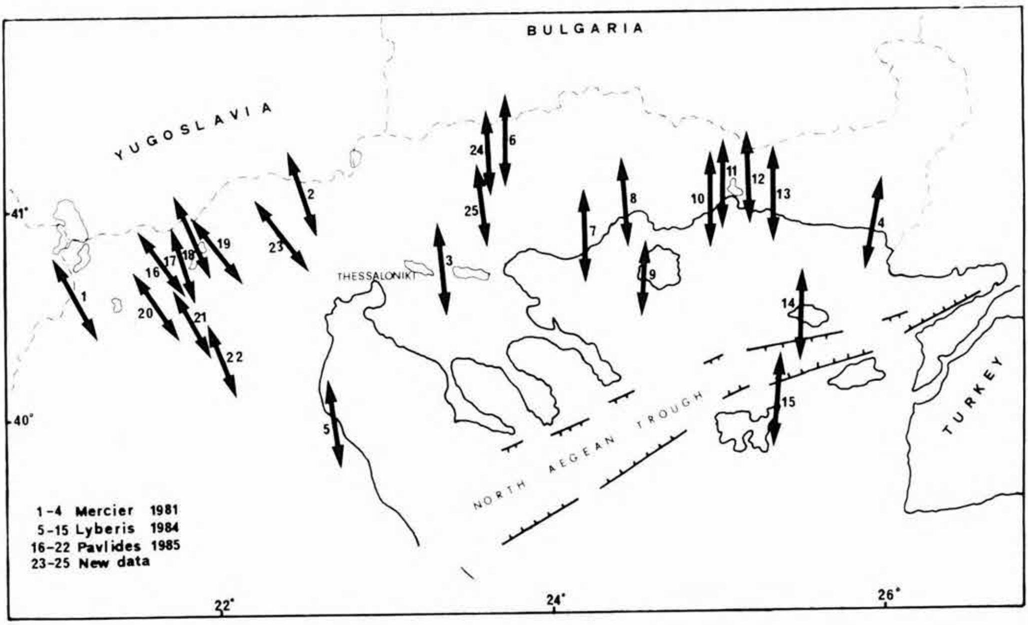


*Fig. 2.* Directions of tension for the period of Quaternary

As far as the palaeomagnetic sampling and measurements are concerned, we collected orientated handsamples from several sites, mainly volcanic, covering all the investigated area.

The samples were measured using standard laboratory techniques by a DIGICO spinner magnetometer and were demagnetized by A.C. field and thermally.

## Geological regime

*Ptolemais basin* (western Macedonia): It has a clear tectonic origin and is filled up by Neogene and Quaternary deposits. According to biostratigraphic determinations the lower formations are of Latest Miocene—Pliocene age (Ruscinian) and the upper ones of Pleistocene age (Villafranchian and Biharian). Taking into account the chronology of the earliest deposits of the basin and the published results for the beginning of the neotectonic activity in the whole Aegean region, a Post Middle Miocene age could be considered as the most acceptable age for the initial creation of the Ptolemais basin.

Two phases of extensional neotectonics are distinguished in the area by detailed field observations and quantitative tectonic analysis. The first one of latest Miocene—Pliocene, has been calculated to be NNE—SSW. That is, the extensional axes, $\sigma_3$, are almost horizontal trending to NNE—SSW direction, while the compressional ones, $\sigma_1$, are vertical and those of intermediate, $\sigma_2$, trend to NW—SE direction. The second extensional phase of Quaternary trends to NW—SE (Pavlides, 1985). *Almopias basin:* The origin of Almopias basin is also tectonic and the volcanic centers of the area, as well as those of Kozuf (S Yugoslavia) are connected with large faults and subsidences. The volcanics of Almopias, occupying an area of approximately 200 km$^2$, are mainly of trachy-andesitic composition and they have been extruded along the Axios (Vardar) zone. Their Pliocene age is determined by *K/Ar* methods as 2.5 to 4 Ma. On the other hand, the lacustrine sediments, which have been determined as Pliocene too by palynological analysis, are in a close relationship with the volcanism (Chorianopoulou et al., 1982).

Two neotectonic extensional phases have also been determined in the Almopias area, similar to that of Ptolemais basin. The *Serbomacedonian zone* and *Strymon valley:* In central and eastern Macedonia eleven exporuses of acidic volcanic rocks have been mapped. Nine of them lie in the Serbomacedonian geological zone and the rest in the western and eastern edges of the Strymon valley. Based on geological data, the age is considered to be Late Pliocene or Plio—Pleistocene.

There is no tectonic relation between this volcanism and any active subduction zone. This is inferred by the alcalic petrochemical character of the volcanic rocks and by the absence of any significant intermediate depth seismic activity in the area (Papadopoulos, 1982).

The Mygdonia basin, which extends in the central part of the Serbomacedonian massif, was created during Middle to Late Miocene. Two different extensional tectonic phases could be distinguished during its evolution; the first is of Miocene—Pliocene age and the second of Early Pleistocene one (Psilovicos, 1977).

## Palaeomagnetic results

The up-to-date published palaeomagnetic results of Greece, suggest that northwestern Greece has undergone two distinct clockwise rotations, of about 20°—25°.

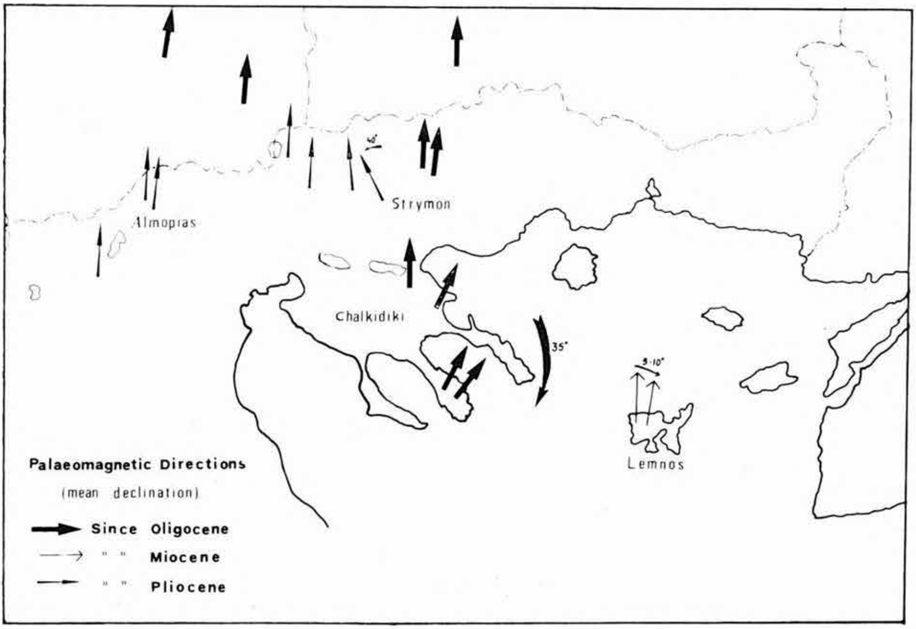


*Fig. 3.* Paleomagnetic directions (mean declinations) as arrows. Those of Bulgaria and Yugoslavia are from Nozharov and Petkov 1976, 1977, Nozharov et al., 1977a (see Kondopoulou and Westphal, 1985 for references) and Stefanović and Veljović 1972

The first one took place between Early to Middle Miocene and the second, also known from Peloponnesos and Ionian islands (west edge of the Hellenic arc), during the Pliocene and Quaternary. These two rotational phases are separated by a period of, at least, 7 Ma (Late Miocene) during which no major rotation occurred (Laj et al., 1982; Kissel et al., 1984).

Concerning Northern Greece and according to our palaeomagnetic results, as well as those refering to the Bulgarian Rhopode massif and south Yugoslavian Serbo-macedonian zone, no clear rotation occurred during Late Tertiary in this region (see also Kondopoulou and Westphal, 1985). Only one segment of Chalkidiki peninsula has undergone a clear clockwise rotation of about 35°, since the late Oligocene. So only this part of the studied area is close to the rotation of western Greece, but the limit between the unrotated part and the rotated one is still unknown (Fig. 3).

Furthermore, according to our latest palaeomagnetic results from Pliocene formations (Table 1), we can conclude that no significant rotation has been detected in the Almopias volcanics and those of Strymon area. The results from one site of the west edge of Strymon valley (Strymoniko, SRY) indicate a large scatter, the reason of which is not fully understood. Another site from the same area (WW) is very coherent, with a mean direction close to Almopias results, that is no clear rotation has been detected for this site. Similar evidence arises from some measured samples of Pliocene sediments (marls and lacustrine limestones) of Ptolemais basin.

The only case which looks significantly different is that of PW situated close to WW (central Macedonia). It is the first time that a strong counterclockwise rotation

*Table 1*

**Palaeomagnetic results of Neogene formations in Northern Greece**

ALMOPIAS (2.6—4 Ma)

| N° of sites | D° | I° | k | $a_{gs}$ | Ref. |
|---|---|---|---|---|---|
| 4 | 17 | 54 | 32 | 16 | Kondopoulou (1982) |
| 6 | 195 | —66.5 | 40 | 10 | Bobier (1968) |
| 3 | 356 | 56.4 | 36 | 14 | This paper |

STRYMON AREA (Plio—Quaternary?)

| *Site* | | | | | |
|---|---|---|---|---|---|
| SRY (6 samples) | 117 | —30 | 5.9 | 33.9 | This paper |
| WW (6 samples) | 349.3 | 58.8 | 43.7 | 10.1 | This paper |
| PW (6 samples) | 318.9 | 42.1 | 13.9 | 18.5 | This paper |

of about 40° is detected in this area. The samples of this site are very strongly magnetized and a possible explanation for the direction mentioned above could be that they have been thundered. In this case, the results should not be coherent. This hypothesis is rejected because results are coherent and normal as well as reverse directions are present which is a good indication of stability. Another problem concerning this site is the age which is known only from geological estimations. We are aware of the fact that no definitive conclusions could be drawn out of this only site but we draw the attention to the peculiarity of this counterclockwise rotation.

## Conclusion

From the overall neotectonic picture of Northern mainland Greece, it is inferred that the tectonics of the Late Tertiary times were uniform throughout its extent. The initial creation of the basins in the area was the result of the tectonic processes of Middle (?) to Late Miocene, while their evolution was completed in two principal extentional phases, the first one of the Late Miocene—Pliocene and the second of Quaternary.

According to the palaeomagnetic studies, no significant rotation has been detected in the area since Pliocene, except the clockwise one refering to southern Chalkidiki peninsula. Low values of mean declination (5°—10°) present in some sites (for instance Lemnos, Almopias) are not significant as they lie in the limit of the reliability of the method.

Only one site belonging to a rigid block of the western edge of Strymon valley (eastern Macedonia) indicates a strong counterclockwise rotation. More samples need in this area in order to establish this result more accurately.

A comparison of the neotectonic data with the palaeomagnetic ones from Macedonia leads us to investigate the following two possibilities.

*a)* If no significant rotation of the region can be seen, then the change of the direction of the stress field is not in relation with any rotation of the area. In this case the change of the stress field from NE—SW to NNW—SSE is rather the result of the mutual exchange of the principal stress axes $\delta_3$ and $\delta_2$.

*b)* If a real counterclokwise rotation exists are in some rigid blocks in the area, as it was observed in one case, then a hypothesis of the same rotation for the stress field with that of the region, could be examine. These two problems are open to future research.

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# THE PRESENT-DAY TARANTO GULF AND THE MIOCENE IRPINIAN BASIN FOREDEEPS OF THE SOUTHERN APENNINES (ITALY)

by

T. PESCATORE and M. R. SENATORE

*Introduction.* The Southern Apennine chain derives from the Cenozoic deformation of the continental margin of the Adria plate.

During the Burdigalian a sedimentary basin was formed (Irpinian basin, COCCO et al., 1972; PESCATORE, 1978) on the front of the thurst sheets.

The aim of this paper is to compare the Miocene foredeep with the Taranto gulf (Fig. 1), the present day foredeep of the Apennine chain (SELLI and ROSSI, 1975; PELFIORE et al., 1981).

The Irpinian basin and the Taranto gulf can be considered sensu lato foredeep areas. Different depocenters can be distinguished: piggyback (ORI and FRIEND, 1984) basins (on the thrust sheets), foredeep sensu stricto at the toe of the thrust sheets and minor foreland basins.

## Morphological framework of the Taranto gulf

Three sectors can be distinguished in the gulf of Taranto having specific morphological characters and different areas of sedimentation (Fig. 2).

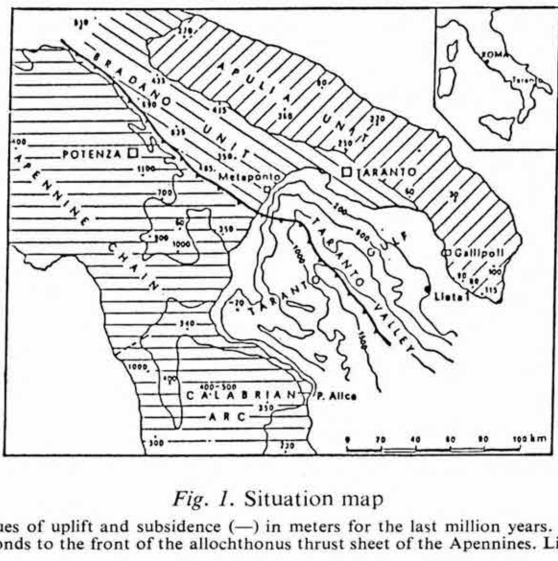


*Fig. 1.* Situation map

The map reports the values of uplift and subsidence (—) in meters for the last million years. Bathymetry is in meters. The toothed line corresponds to the front of the allochthonus thrust sheet of the Apennines. Lista 1 is well (Agip, 1977)

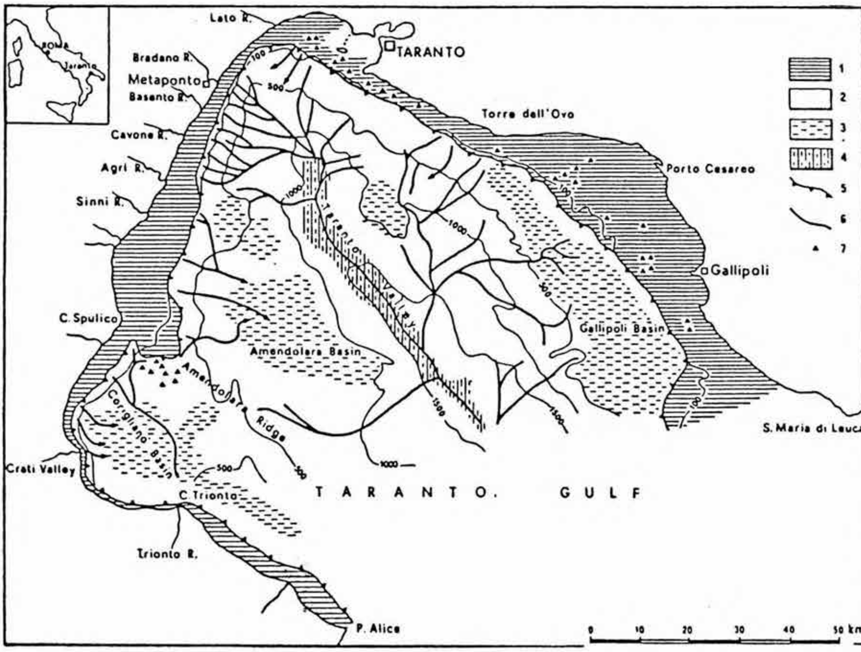


*Fig. 2.* Morphology map of the gulf of Taranto
*1* Continental shelf, *2* continental slope and ridges, *3* flat areas (basins), *4* Taranto valley (foredeep basin), *5* shelf break, *6* main channels, *7* corraline algal banks

*Western sector*. This sector extends between the margin of the Taranto valley and the land. The shelf is either prograding or regrading (Mougenot et al., 1983). Two basins are individualized, Corigliano and Amendolara basins, separated by a ridge (Amendolara ridge).

*Central sector*. The central sector is the Taranto valley, a NW—SE trough located between the allochthonous thrust sheets to the west and the Apulia foreland to the east. The terrigenous materials from the shelf accumulate in the valley; there are some channels coming from the Apulia sector which carry into the valley bioclastic calcareous deposits. The coarser materials are transported into the valley by turbidity currents (Senatore et al., 1982).

*Eastern sector*. The eastern sector is situated between the valley of Taranto and Apulia. The shelf is characterized by Holocene terraces; the slope by slump deposits which cover more than 30% of its surface. This sector comprises two Plio—Quaternary depocenters. The most extended one, the Gallipoli basin, has Plio—Quaternary deposits more than 500 m thick; they lie unconformably on the carbonate rocks of the Apulia foreland (Lieta 1 Well; AGIP, 1977).

#### Structural framework of the Taranto gulf

Our knowledge of the tectonic structures of the gulf of Taranto is based principally on the interpretation of seismic profiles (Finetti, 1976; Rossi et al., 1983; Tramutoli et al., 1984; Cello et al., 1981).

Using FINETTI's interpretation (1976) and the observations by TRAMUTOLI et al. (1984), it is possible to individualize the following principal elements in the gulf of Taranto:

— The western sector represents the continuation southward of the allochthonous Apennine units. It is characterized by an imbricate structure with two main thrusting fronts. Towards the west the fronts limit asymmetrical basins, where sedimentation occurred during deformation.

— The Taranto valley is the central sector of the gulf: this is a trough at the toe of the allochthonous thrust sheets, the foredeep sensu stricto of the Apennine chain.

— The eastern sector (Apulia margin) is the foreland, with Pliò—Quaternary basins.

**Irpinian basin**

The Irpinian units or Irpinids, at the present outcrop in Campania and Lucania, along two major belts, both trending northwest—southeast (PESCATORE, 1978; 1985). The southwestern belt includes terrigenous deposits resting unconformably on the older allochthonous thrust sheets (Castelvetere Formation—upper part—and Gorgoglione Flysch) while the northeastern belt is formed by calcareous and terrigenous turbidites conformably following the underlying deposits (Serra Palazzo Formation and Faeto Formation).

The palaeogeographic domains existing during the Mesozoic and Early Cenozoic in this sector of the Adria continental margin, from west to east, have been named as follows: Campania-Lucania platform, Lagonegro basin, Abruzzi-Campania platform, Molise basin and Apulia platform (IPPOLITO et al., 1975).

Upper Oligocene—Lower Miocene terrigenous successions consisting of an alternation of quartzose sandstone and pelites (Numidian Flysch, Auct., OGNIBEN, 1969) represent the beginning of the terrigenous synorogenetic sedimentation in the external domain of the Southern Apennines.

In the Burdigalian an important tectonic pulse significantly modified the preexisting domain. In particular, it deformed the terrains of the Campania-Lucania platform and in part the ones of the Lagonegro basin, which structured the Irpinian basin. Depocenters may be separated into three groups (Fig. 3):

*1* Upon the allochthonous thrust sheets: terrigenous successions are deposited at the base of the slope (Castelvetere Formation—upper part—) or in elongate deepsea fans parallel to the structural belts (Gorgoglione Flysch).

The terrigenous deposits (arkosic—lithic) were supplied by erosion of the thrust sheets.

*2* At the front of the thrust sheets: a calcareous and terrigenous succession was deposited at the toe of the thrust sheets (Serra Palazzo Formation).

*3* In the areas not yet deformed, i.e. the Molise basin, turbidite and hemipelagic calcareous successions were deposited (Faeto Formation). The clastic materials were fed from carbonate platforms.

Sedimentation occurred while the axis of the foredeep migrated eastward as suggested by the diachronism of the terrigenous facies, more recent towards the east, and the regressive trend of the successions (PESCATORE, 1978).

Sedimentation in the Irpinian basin terminated as a result of a new and important tectonic pulse in the Early Tortonian.

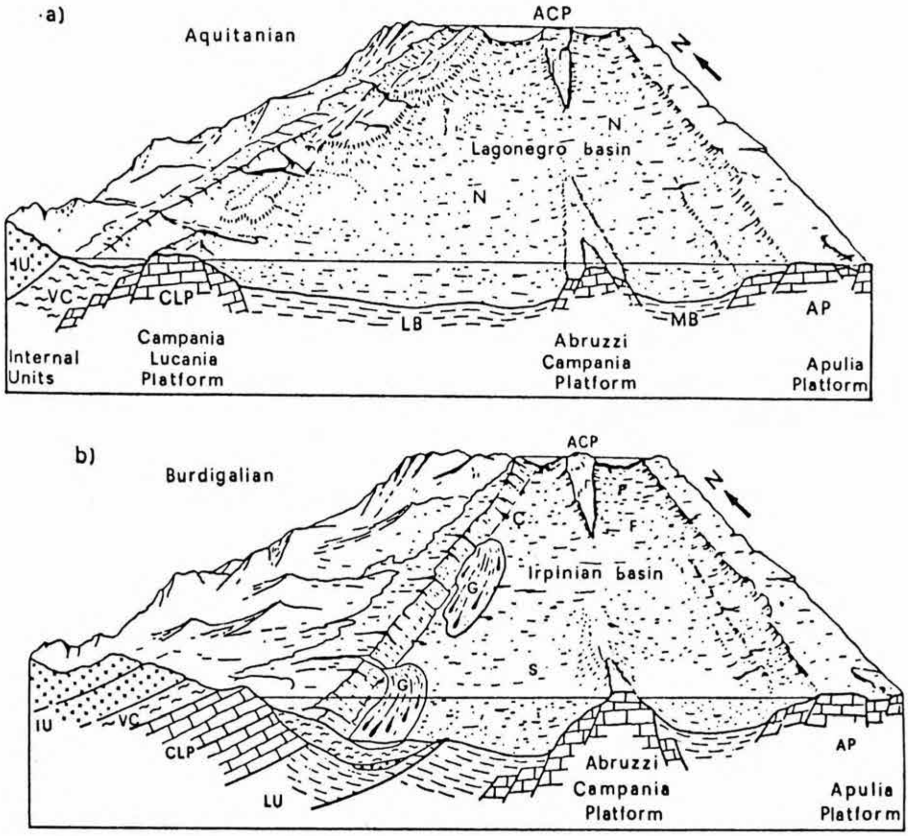


*Fig. 3.* Palaeogeographic scheme before (a) and after (b) the Burdigalian tectonic pulse

IU = Alpine and Apennine units, VC = variegated clays, CLP = Campanian Lucania platform, LB = Lagonegro basin, ACP = Abruzzi Campania platform, MB = Molise basin, AP = Apulia platform, N = Numidian flysch (quartzose sandstones), LU = Lagonegro unit, G = Gorgoglione flysch, C = Castelvetere Formation (upper part), S = Serra Palazzo Formation, F = Faeto Formation.—Not in scale

## Discussion

The distribution of the Irpinian Units and the Plio–Quaternary depocenters in the gulf of Taranto shows, from west to east, the same succession of basins (Fig. 4).

*Table 1*

| | Irpinian basin | Gulf of Taranto |
|---|---|---|
| PIGGYBACK BASINS | Castelvetere Formation —upper part— | Amendolara basin |
| | Gorgoglione Flysch | Corigliano basin |
| FOREDEEP BASIN | Serra Palazzo Formation | Taranto valley |
| FORELAND BASIN (minor) | Faeto Formation | Gallipoli basin |

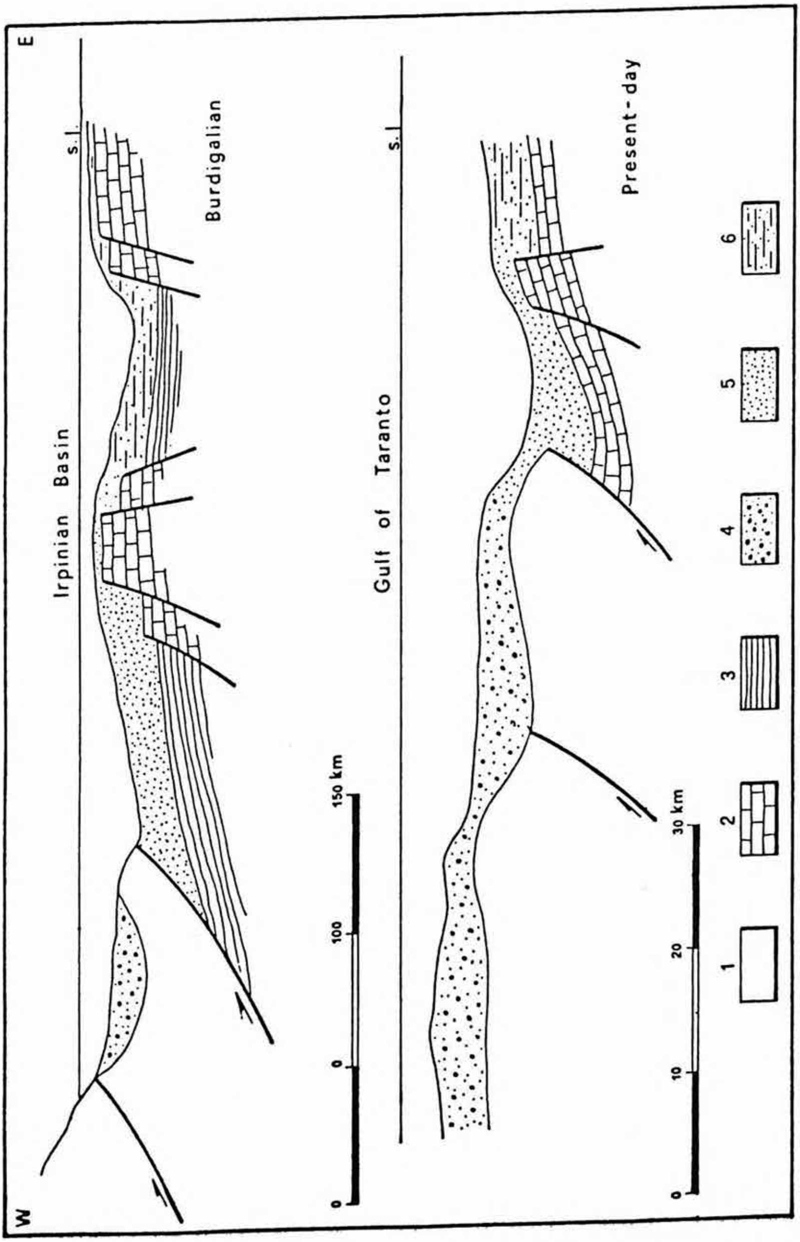


*Fig. 4.* Comparative sections across the gulf of Taranto (interpreted) and the Irpinian basin after the Burdigalian tectonic pulse (reconstructed)

*1* Allochthonous thrust sheets, *2* neritic platforms, *3* pelagic basins, *4* piggyback basins, *5* foredeep, *6* foreland basins. —The scale of length is greatly approximated for the Irpinian basin, hights are not in scale

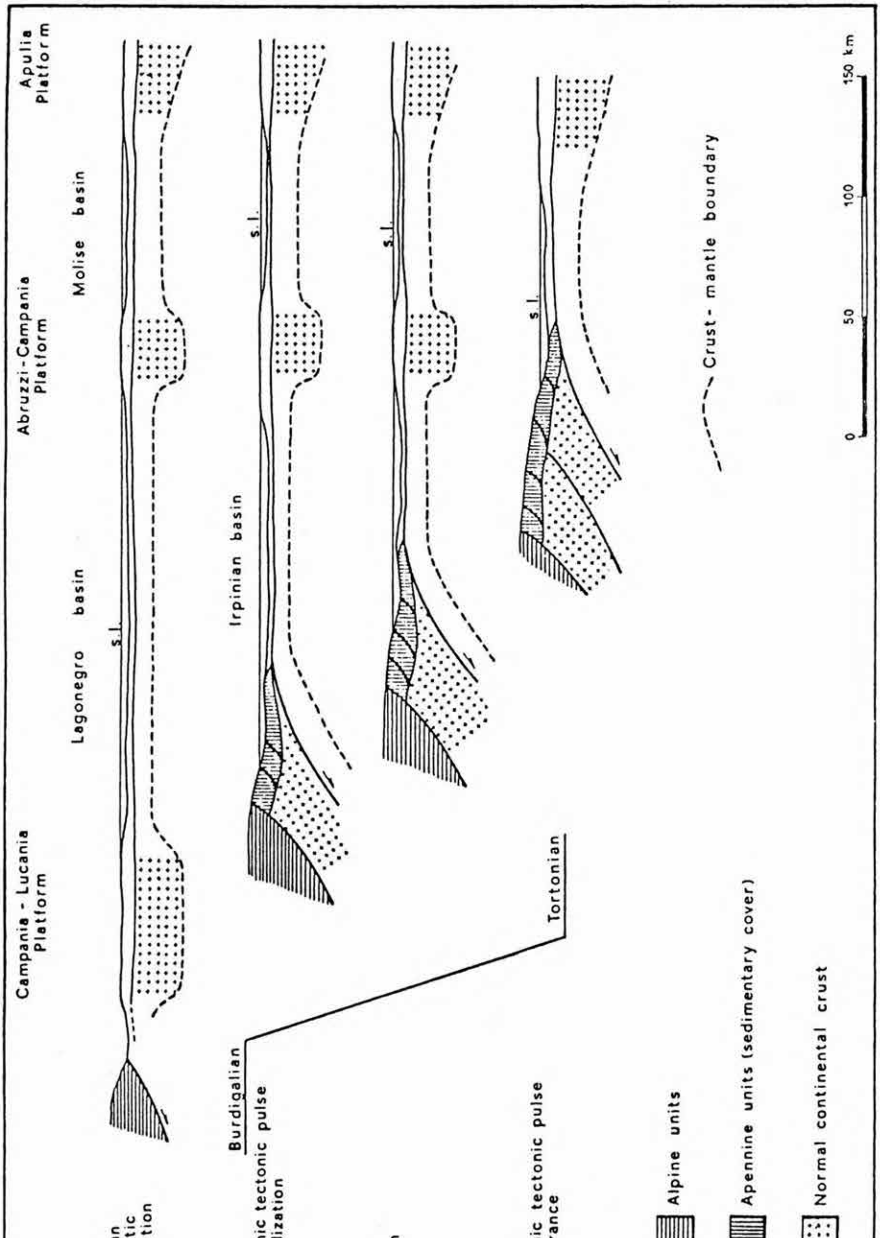


*Fig. 5.* The evolution of the Irpinian basin

The evolution of the Irpinian basin spanned a peiiod of ca. 10 m. y., from the Burdigalian to the Tortonian. Contrarily, for the Taranto gulf, only 2 m. y. are taken into consideration.

The width dimensions of the Irpinian basin can be estimated as being by a factor of two or three greater as compared to the Taranto gulf, while there are no references for a definition of the length of the basins.

During the compressive stage, the evolution of the Irpinian basin should have been influenced by the structures acquired in the previous crustal stretching (Fig. 5).

It would seem possible to conclude that:

I The discontinuities of the crustal thickness along the margin of the Adria plate, linked to the tensile regime of the Mesozoic, controlled the development of the flysch-like basins during the compressive regime. The flysch-like basins seem to be localized in areas with a thinned crust.

II During a continuous compressive regime, the margin of collision evolves in a discontinuous manner according to the crustal thickness of the colliding blocks: normal crust collision gives rise to distrophic pulses; normalcrust—thinned crust collision gives rise to flysch-like basins.

If an analogous model could be applied to the Taranto gulf, we ought to retain that the gulf is in its final stage, i.e. when the front of the thrust sheets has reached the margin of the Apulia platform with normal crust. The proceeding stage would have been characterized by A-type subduction of thinned crust existing below the Molise basin, to the west of the Apulia platform.

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# SEISMIC STRATIGRAPHIC AND SEDIMENTOLOGICAL ANALYSIS OF NEOGENE DELTA FEATURES IN THE PANNONIAN BASIN

by

GY. POGÁCSÁS and I. RÉVÉSZ

On the basis of seismic stratigraphy, seismic faciology and sedimentology the Upper Miocene—Pliocene delta formations have decisive role in the Neogene infilling of the Pannonian basin. Delta features can be identified in the Great Hungarian Plain in SW Hungary (Zala and Dráva basin) and in the Little Hungarian Plain Fig. 1).

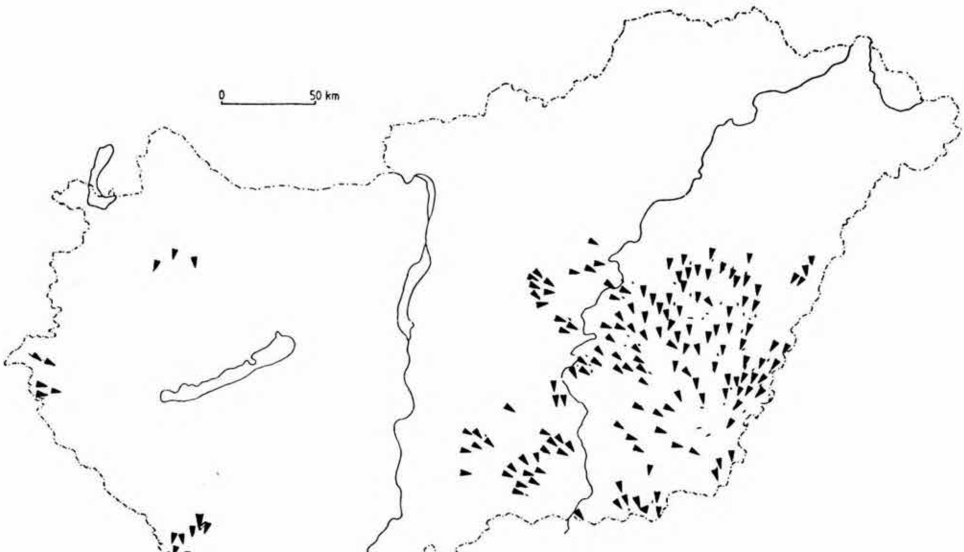



*Fig. 1.* Distribution of dip direction vectors of the prograding deltaic sequences

Fig. 6 presents the seismic, lithogenetic and lithostratigraphic units of the Great Hungarian Plain. Parts of the delta sequence are marked. On the Fig. 2 can be seen the idealized depositional feature of a delta body and the connected turbidite sequence after BERG (1983).

The delta sequence (illustrated by diporiented seismic profile on Fig. 3) represented by characteristic oblique progradational and sigmoid progradational seismic facies implies special depositional conditions, i.e. combinations of relatively high sediment supply, slow to no basin subsidence and a stillstand of the sea level which allows rapid

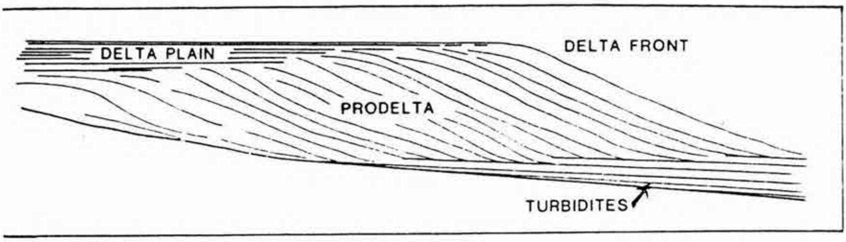



*Fig. 2.* Model of a prograded fluvial-dominated delta and associated turbidite sequence

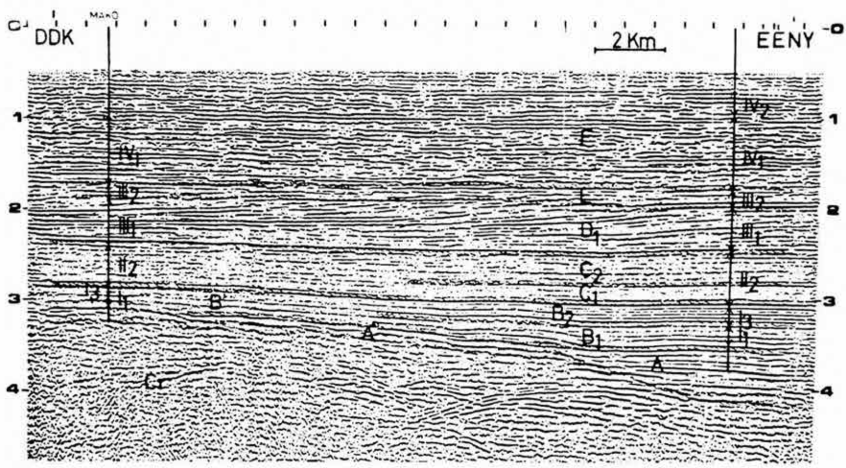



*Fig. 3.*

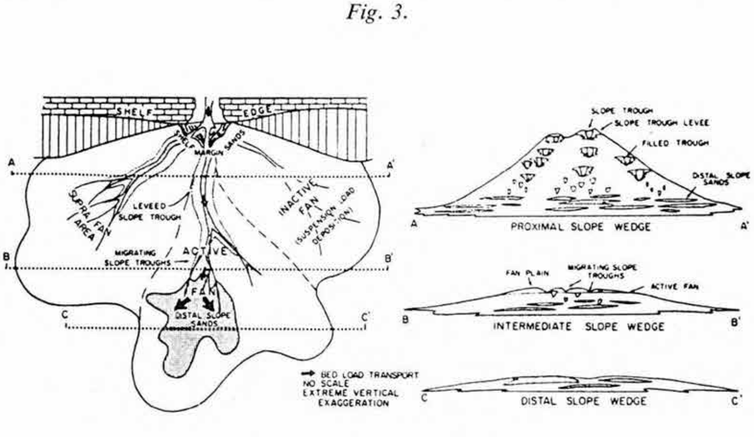



*Fig. 4.* Slope fan model

basin filling and sedimentary bypass of the upper depositional surface. The strike oriented sections of the delta fans can be seen of Fig. 4 (after BROWN and FISCHER, 1977).

Deposition of sediments prograded gradually from the marginal part of the basin one towards the central, thus the delta formations are the youngest in the central parts of the basin (SE Hungary).

The seismic reflections within the lowest Pannonian depositional sequence (Fig. 5, D—E) indicate that these sediments lie unconformably on the pre-Tertiary basement (showing onlap and downlap structures) and may lie either conformably or unconformably on the older Miocene formations. Reflectors within the sequence are parallel or slightly divergent due to differential subsidence of the basement. This sequence can often be divided into two seismic facies, B and C, characterized by different intensity and continuity of reflectors, and can further be subdivided into several vertically and laterally interfingering seismic subfacies.

The lowermost member of the delta sequence is the Vásárhely Formation (seismic unit $B_2$) characterized by great amplitude and strong continuity of seismic reflections (Fig. 5, B—C) and by pelitic, pelitic—carbonatic sediments with thin sandstone intercalations. The presence of Bouma sequence in the formation can be recognized. The microfaults and folded-layered structures referring to sediment slides are characteristic. The quiet aquatic sedimentation is dissected by the deposition of the thinning margins of distal turbidites caused by gravitational flows from the shore. This section of the sequence can be interpreted as a prodelta sequence belonging to the delta front. This formation is characteristic first of all in Southeast Hungary.

A sedimentary sequence with similar sedimentation mechanism is found in the central and northern parts of the Great Plain (Nagykörű Claymarl Formation, seismic unit $C_1$), the only difference being the higher carbonate and correspondingly the lower sandstone proportions.

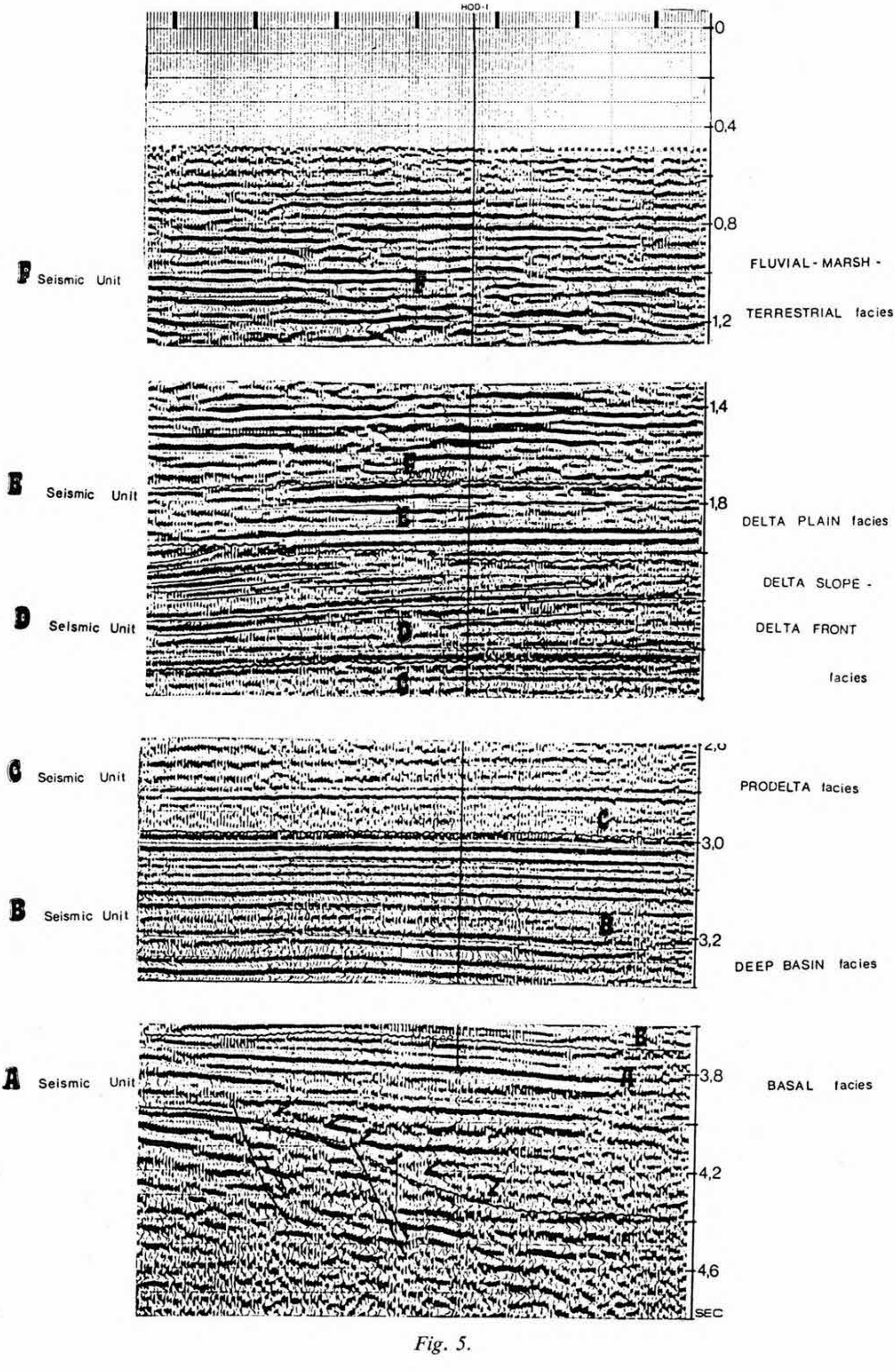
The Szolnok Sandstone Formation (seismic unit $C_2$) is found also in the deeper basin parts, that can be assigned also to the prodelta formations of the delta front and characterized by distal turbidites consisting mainly of argillaceous marl and siltstone and containing more sandstone that the formations above. Above the basement ridges this formation is wedged. Small bioturbations convolutions and siltstone intraclasts also occur here in addition to the Bouma sequences. On the jointing planes flute clasts abound; the plant remnants are also common. In sandstones plates of tenth of millimeter thickness consisting of mica and coalified plant remnants can be observed. In harmony with the electrofacies, these features refer to sandy, silty, gravitational redeposition, with thinner strata characteristic of distal turbidites and with normal gradation occasionally with bioturbated pelites separating the turbidite bodies.

The areas subsiding most rapidly in early Pannonian time can be identified by increased sediment thickness and by the greater difference in character of the seismic facies B and C.

The Algyő Formation (seismic unit D) consists of several seismic subfacies which grade vertically and horizontally into each other referrins to the deposition conditions of delta slope. The most characteristic of these can be correlated with the oblique progradational and sigmoid progradational seismic facies. These formations were observed first by KILÉNYI and RÁKÓCZY (1966) on reflection profiles.

The oblique and sigmoid progradational sequences are characterized by downlapping reflectors at their base, representing the development of sediments from relatively shallow water into deep water. Thinning of the outer toes of individual beds, if present, is beyond seismic resolution. The upper part of this sequence consists of sedimentary rocks deposited by a fluvio-marine system.

The sequence is characterized by thicker sandstone strata belonging to proximal turbidites, by quiet-aquatic argillaceous marl and siltstone strata. Based on rock


HOD-I
0
0,4
0,8
1,2
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3,2
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SEC
F Seismic Unit
FLUVIAL-MARSH -
TERRESTRIAL facies
E Seismic Unit
DELTA PLAIN facies
D Seismic Unit
DELTA SLOPE -
DELTA FRONT
facies
C Seismic Unit
PRODELTA facies
B Seismic Unit
DEEP BASIN facies
A Seismic Unit
BASAL facies

*Fig. 5.*

| SEISMIC FACIES | | LITHOGENETIC UNITS | LITHOSTRATIGRAPHIC UNITS: FORMATION GROUP | LITHOSTRATIGRAPHIC UNITS: FORMATION | |
|---|---|---|---|---|---|
| F | | $Pa_2$ | HEVES IV | Nagyalfold Clay Formation | |
| | | | | Zagyva Sand-Clay Formation | Fluviatile-lacustrine |
| E | $E_1$ $E_2$ | | CSONGRÁD III | Tortel Sandstone Formation | Delta plain |
| | | | | | Delta front |
| D | $D_1$ $D_2$ | $Pa_1^2$ | | Algyő Marl-Sandstone Formation | Delta slope Proximal turbidites |
| C | $C_2$ | | JÁSZKUNSÁG II | Szolnok Sandstone-Marl Formation | Prodelta distal turbidites |
| | $C_1$ | $Pa_1^{1b}$ | | Nagykörü Clay-Marl Formation | Prodelta distal turbidites |
| B | $B_2$ | $Pa_1^{1a}$ | MAROS I | Vásárhely Marl Formation; Tótkomlós Clay-Marl F. | Prodelta distal turbidites |
| | $B_1$ | | | Békés Conglomerate F.; Dorozsma Marl F. | |
| A | A | - | MIOCENE | | Basal facies |
| Cr | Mz | - | BASEMENT | | |

*Fig. 6.* Correlation table of seismic facies and lithostratigraphic units

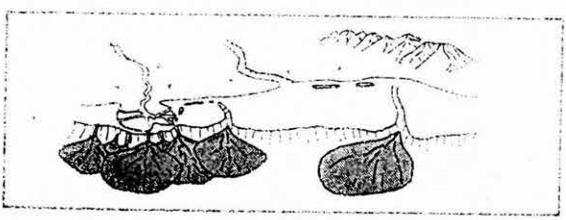

*Fig. 7.*

structural features all the types of gravitational sediment transport can be presumed. In well log profiles subaquatic beds, mouth bars can be distinguished. The upward coarsening sandstone microrhythms and red colour appearing in the upper section of the formation refer already to the margin of the delta front.

The vertical distance between the lowest and the highest points of the seismic reflections belonging to unit D may reach 700–900 m showing the maximum water depth during the deposition of this unit.

The up-to-date seismic reflection profiles of the Geophysical Exploration Company provided the possibility to study the characteristics of the seismic facies, the dipping conditions, and the spatial distributions of the prograding sequences.

At some hundred points on the basis of two intersecting seismic profiles, the true directions and dip angle of the reflecting surfaces belonging to unit D were determined. Fig. 1 shows the distribution map of the "dip vectors" obtained in this manner. At some points the dip directions of the reflecting surfaces lying over each other were also determined. These results may be seen in Fig. 8 (numbering of the dip vectors assigned to the reflecting surfaces starts form the bottom). The scattering of the dip directions determined at the same locations refers to the fact that the palaeo-slope conditions had changed rapidly as well as the transport direction in the course of the deposition of the unit. D. Fig. 7 reproduced from BERG (1983) illustrates the palaeogeographic environment during the deposition of unit D.

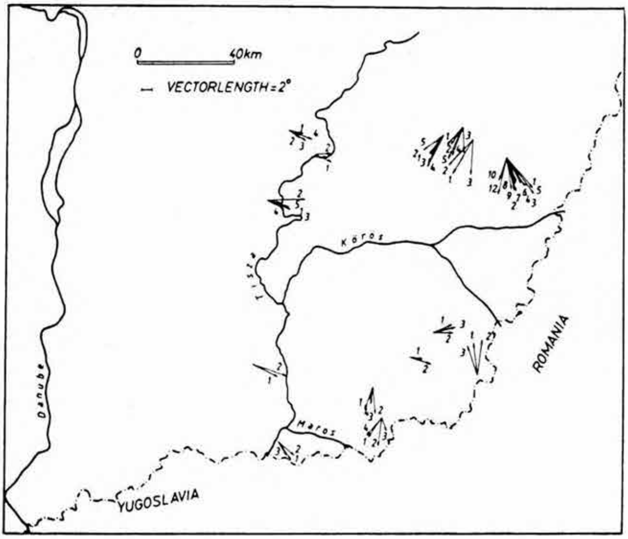


*Fig. 8.* Dip directions of the reflectors lying over each other in delta sequences (after LUKÁCS et al. 1983)

Unit D is overlain by Törtel Sandstone Formation (unit E) which, in some depressions, consists of two subunits, $E_1$ and $E_2$, and is characterized by parallel onlapping and downlapping reflectors. $E_1$ is characterized by high amplitude and strong continuity, the overlying $E_2$ unit by medium to low amplitude and medium to weak continuity. The upper boundary of the unit is conformable with internal reflections. Unit E represents the delta plain facies composed of alternating horizontally bedding sandstone—siltstone and argillaceous marl. Unit E may be interpreted as the partly contemporaneous facies (off-shore and lagoonal) of the prograding unit D. These shallow water formations consist of alternating sequences of lagoonal or terrestrial formations and fluvial sediments deposited in high and low depositional energy conditions.

In this formation the sand content of the sequence suddenly increases. The upward coarsening sedimentary rhythms become predominating. Woody lignite and coaly clay lenses also occur. Both the rock texture and structure, together with the marks cited above refer to the near-shore part of the delta front and the coastal zone of the delta plain as depositional environment. In the hydrocarbon exploration practice the lower boundary of this formation is used as the boundary between the Lower and Upper Pannonian. This, however, does not mean a temporal boundary, it rather follows the temporally long-lasting process of the progradation of the delta front-coastal environment, that is of an event of filling starting from the basin margins.

The third and shallowest depositional sequence the Zagyva Formation (unit F) makes up the youngest member of the Pannonian formations. In certain regions (e.g.

Danube—Rába lowland, the north Jász region, the foreground of the Mátra Mts, the southern part of the Derecske depression) an unconformity within the Upper Pannonian sequence could be identified seismically (see POGÁCSÁS in this volume). The Unit F is characterized by parallel and slightly divergent reflectors, with low to moderate continuity and moderate to high intensity (Fig. 5, F). The seismic and sedimentary characteristics of unit F (Zagyva Formation) suggest fluviatile, flood-plain, marshy, terrestrial and lacustrine sedimentation. The Nagyalföld Clay Formation containing variegated clays and fluviatile sand is the final member of the Pannonian sedimentation.

The sequence introduced above from the bottom towards the top indicates simultaneously the migration of the shore-line and it reflects fairly well both the temporal succession and the lateral contemporaneity of the facies, respectively.

Summarizing one can conclude that in the Neogene the Hungarian part of the Pannonian basin was filled up by sediments of prograding delta systems and connected environments from the margins of the basin (Fig. 1). The Great Hungarian Plain was filled from northwest and north east direction, the Zala basin from northwest and the Dráva basin from north.

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# NEOGENE VOLCANISM IN HUNGARY

by

Cs. Ravasz

The territory of Hungary is a part of the suture zone between the collision of the Afro-Indian and Eurasian continents along the Alpine—Himalayan orogenic belt. Collision zones of this kind are the most complicated ones due to the crack of the marginal region into a series of microplates, which are capable of independent motion with respect to each other. Along continent—continent collisions, subduction is relatively rare. The territory of Hungary, as a part of the Carpathian Basin, is the area of this rare phenomenon (Fig. 1—2). During an approximately 10 million years within the Neogene, the development of the Neogene basins and sedimentation were accompanied by heavy outbursts of volcanic activity of calcalkaline character.

On the base of the three main levels of tuff accumulations within the Neogene sedimentary sequence, Lower, Middle and Upper Rhyolitic Tuff accumulations were distinguished in tradition Hungarian stratigraphy.

Based on new results and parallel *K/Ar* datings, four main volcanic events can be detected, each of which produced pyroclastic rocks as well as lava flows but the proportion of pyroclastic rocks is varying significantly.

The earliest event is characterized mainly by rhyolitic ignimbrites and locally by minor rhyolitic domes appearing approximately at the boundary of Eggenburgian—Ottnangian, with an average thickness of 70 m, maximum 300 m, covering an area of 1006 km$^2$ and comprising by 112 km$^3$. While partly eroded, the volcanic products can be detected on the surface and in the basins in 500 to 1500 m depth range, covered by younger sedimentary rocks.

The eruption centres are found along the strike slip faults of SW—NE trend, indicating the sites of perpendicular cracks due to the rotation of the plate on one side and to the shearing movement on the other side, the latter being a result of the situation of the subduction zone in space and time. It is worth of mention that in a situation, where the direction of subduction is oblique to the surface trace of the subduction zone, strike slip faults occur in the marginal zone of the facing plate.

Regarding the volcanic rocks of each volcanic event and as a whole, younger and younger eruptions occur step by step towards the NE, as demonstrated also by *K/Ar* data. This phenomenon can be explained by the oblique situation of the descending lithosphere and the periodic rotational movement.

Within the Karpatian stage the repeated volcanic activity can be regarded as following the same model, producing dacitic as well as mixed, rhyolitic, andesitic ash fall tuffs as blankets, followed by ignimbrites and foam lavas of rhyolitic—dacitic composition. Minor andesitic domes also occur. The average thickness of the volcanic material is about 60 m, maximum 1000 m, covering an area as big as 10,000 km$^2$, in a volume of 1000 km$^3$.

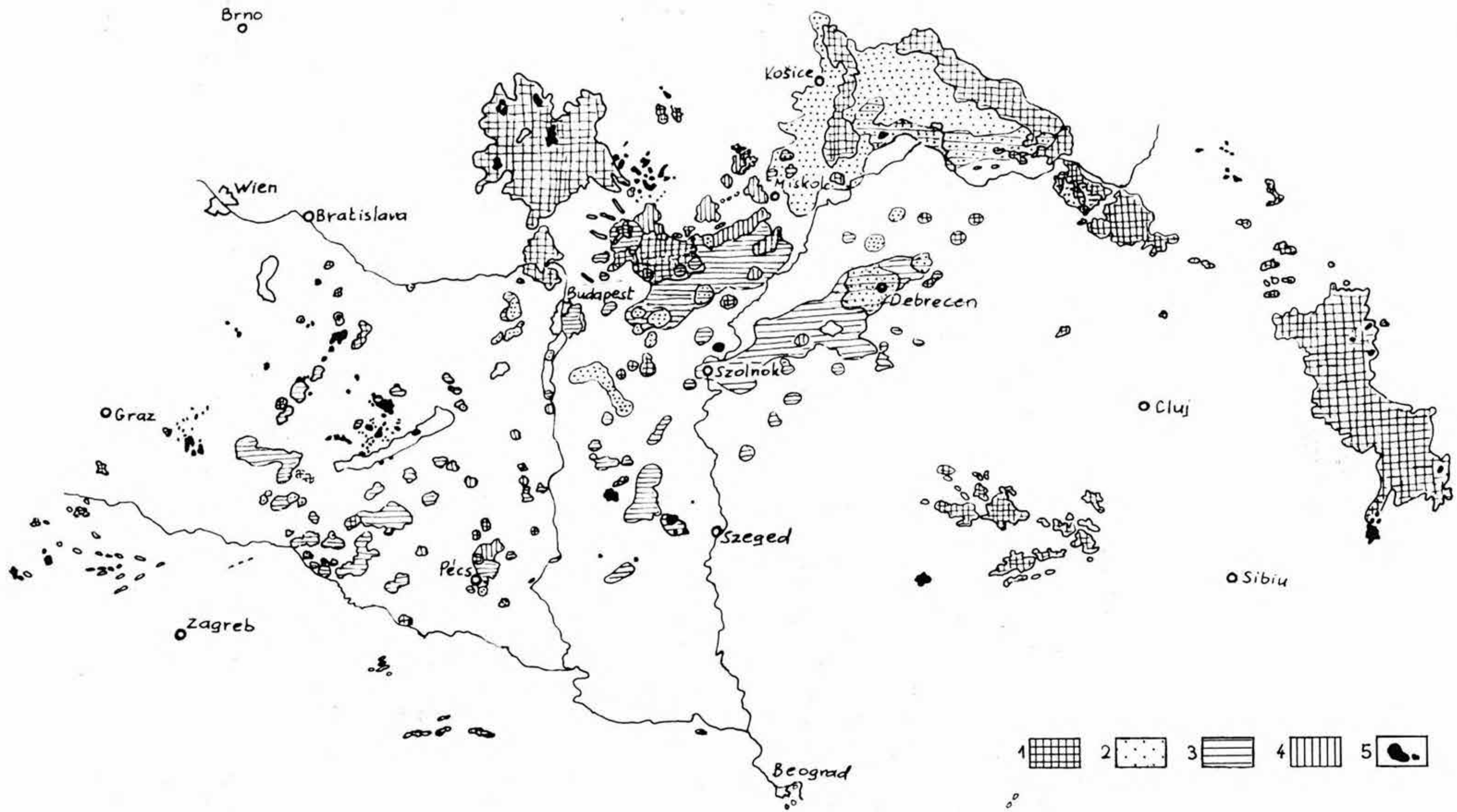


*Fig. 1.* Neogene volcanic rocks of the Carpathian basin

*1* Miocene andesite and associated rocks, *2* Sarmatian "Upper" rhyolitic tuff, *3* Karpatian "Middle" rhyolitic tuff, *4* Ottnangian "Lower" r hyolitic tuff, *5* Pliocene alkali basaltic rocks

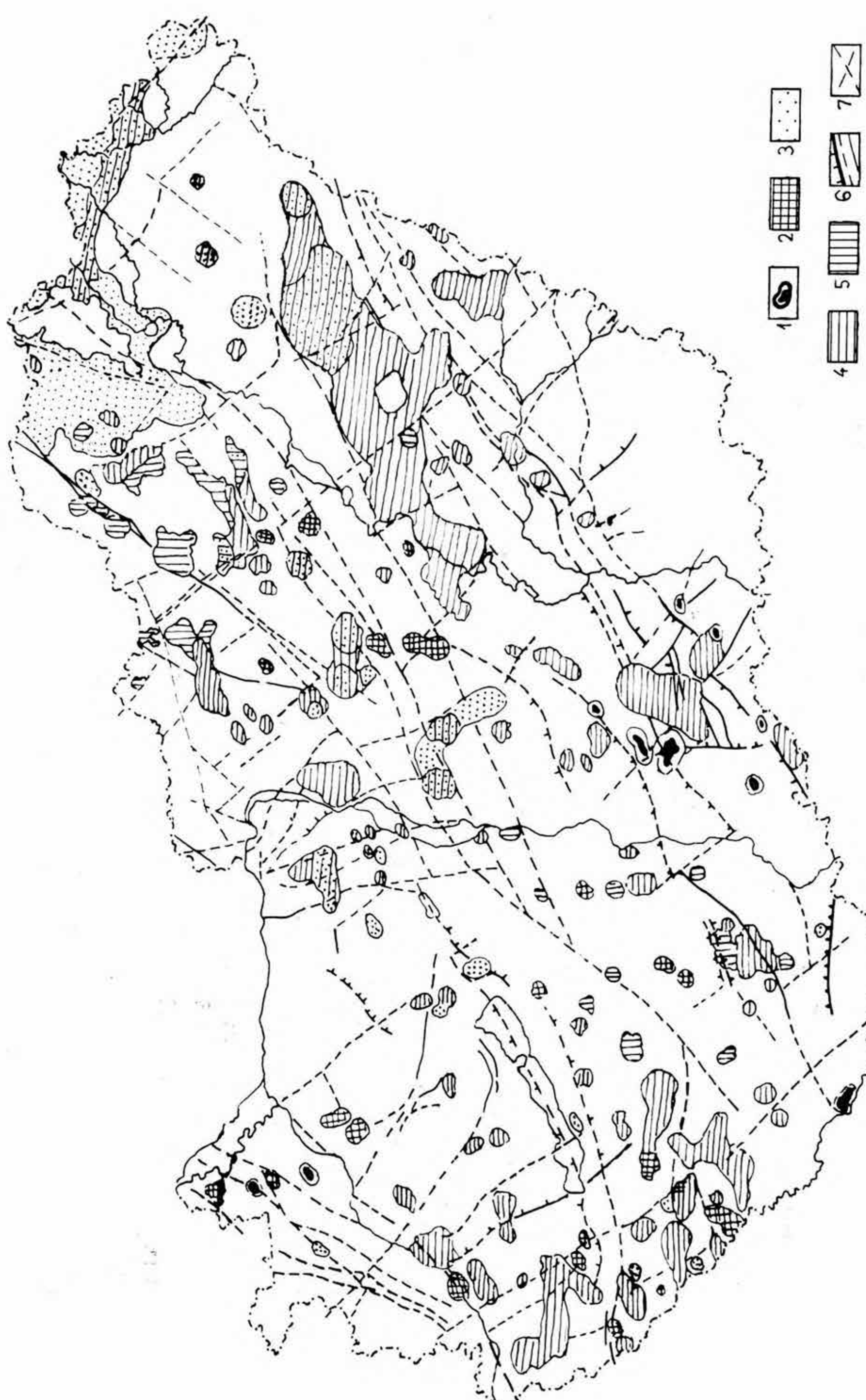


*Fig.* 2. Extension of Miocene volcanic rocks in the Hungarian basin
*1* Upper Miocene basaltic rocks, *2* andesite and associated rocks.—*Acidic pyroclastic rocks: 3* Sarmatian, *4* Karpatian, *5* Ottnangian.—*6* Dislocation lines, *7* fault

Depending on the rate of plate rotation and of the subduction process, volcanic material turned towards a more basic composition, resulting in large amounts of andesitic magma during the Badenian stage. The stratovolcanic complexes of Hungary have been formed during the Badenian, extending an area of 600 $km^2$ in a thickness up to 2500 m and in a volume of 500 $km^3$, in N Hungary.

On the contrary, the mean area of the fourth volcanic event of the calcalkalic volcanic activity during the Sarmatian is marked in the NE part of the country, in the Tokaj area respectively and the surroundings, characterized mainly by rhyolitic and dacitic tuffs, ignimbrites, glowing avalanches followed later by dacitic—andesitic lava flows and shields. Total volume of the volcanic products can be estimated about 900 $km^3$ here, while in the western part of the country, for example in Transdanubia, only thin tuffaceous accumulations are found.

Finally, as the rate of the rotation of the microplate prevailed over the subduction proceed, radial dyke swarm of basalto-andesitic composition closes the calcalcalic volcanic events.

Against the fact that Sr isotopes and rare earth determinations are sporadic, based on petrographic observations, field experiences and *K/Ar* data, we suppose that both or rhyolitic and andesitic magmas have been originated from a common source. This can be proved by exsolved pyroxene crystals being abundant in the rock types of rhyolitic composition and by the occurences of contemporaneous andesitic and rhyolitic products, beside each other. The volume and extension of the andesitic rocks are relating to the presence of a subducted plate with a composition of oceanic lithosphere (dense eclogite?) that time. Acidic members could be generated by plagioclase dominated fractional crystallization and some rate by contamination within the upper crust.

When dinamic process of the subducted slab has been ended while rotation of the plate continued, cracks along deep faults of NNW—SSE and N—S directions played a more important role along which alkali basaltic magmas of continental character have been upwelled during the Late Miocene and the Pliocene. Magmas of this type can be derived by partial melting of mantle material. These products of youngest period of volcanic activity indicates fundamental changes in tectonic environment and magma sources.

The Late Miocene and Pliocene volcanism can be characterized by a serie of alkali basaltic rocks included trachytes, basanites and peralkaline rock types, too. In the beginning, the ratio of pyroclastics and lava flows was rather high, while later decreased. From the Late Miocene towards the Pliocene, basaltic magmas became more undersaturated with a simultaneously growing significance of the continental character.

Based on the fact that in the northern part of the country basanites occur near the border of the Pliocene—Pleistocene while in the southern part of Hungary a variety of potash-rich mediterranean magma type is known with similar age, we can suppose a heterogeneity of the subcontinental crust or, probably of the mantle, within the territory of the country.

As a result of our researches, the Neogene volcanic products of great stratigraphic importance can be considered with some caution in the future due to their origin of a rather complicated mechanism and to the respect of the localized volcanic episodes.

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# GEODYNAMIC EVOLUTION OF THE TYRRHENIAN SEA NEW MULTICHANNEL SEISMIC REFLEXION DATA (ODP Leg 107 sites survey)

by

J. P. REHAULT, E. MOUSSAT, J. MASCLE and R. SARTORI

*Introduction.* The present-day Western Mediterranean Sea results from successive openings of back-arc basins associated with the consumption of the former Mesogean ocean areas.

The Tyrrhenian Sea is the youngest of these basins. Its western margin is edged by the Corsica and Sardinia islands, the drifting of which, related to Europe, stopped 18 m.y. ago (REHAULT J. P., 1981). Its northeastern and southern margins are edged by two compressive systems, the Apenninic and the Northern Sicilian Chains, the backbones of which have mainly been built between the Burdigalian and the Tortonian times (i.e. from 18 m.y. to 10 m.y. ago) though more recent compressive events have occured since then (AMODIO-MORELLI et al., 1976).

Both chains are linked in the east by the Tyrrhenian arc (so called Calabrian—Peloritan arc) which faces the Ionian sea, the only oceanic area of the foreland between the African continent and its Adriatic counterpart.

On the one hand, it is tempting to attribute the opening of the Tyrrhenian to the subduction of the oceanic lithosphere a relic of which the Ionian sea is. However, the occurrence of active margins without back-arc basins shows that the plunging of this slab into the asthenosphere is not sufficient for such a process. On the other hand, induction of a tensional state of stress in the Tyrrhenian area by a NNW/SSE shortening between Europe and Africa implies an anticlockwise rotation of the Adriatic foreland related to the African plate. However, the seismicity does not show any active plate boundary between them and no relevant observation support the hypothesis of such boundary in Neogene (HORVÁTH, 1984).

An attempt to conciliate both facts has already been published (MOUSSAT, 1983; MOUSSAT et al., 1985a, b). This model is discussed here in the light of the results of new multichannel seismic reflexion records carried out by IFREMER/IFP for the ODP (Leg 107) sites survey.

## Structure of the Tyrrhenian basin (Fig. 1)

The area of maximum stretching (Moho depth shallower than 10 km) and volcanic activity lies in the east of the so called "Faille Centrale". This area is divided into two depressions tightly delineated by the 3000 meter isobath, the Central and Eastern basins, in the middle of which elongated tholeitic volcanoes (Magnaghi, Vavilov and Marsili) form embryons of oceanic ridges (MOUSSAT, 1983; MOUSSAT et al., 1985b). According to a previous detailed fault mapping these deep basins (as well as the margins) are characterized by two sets of faults:

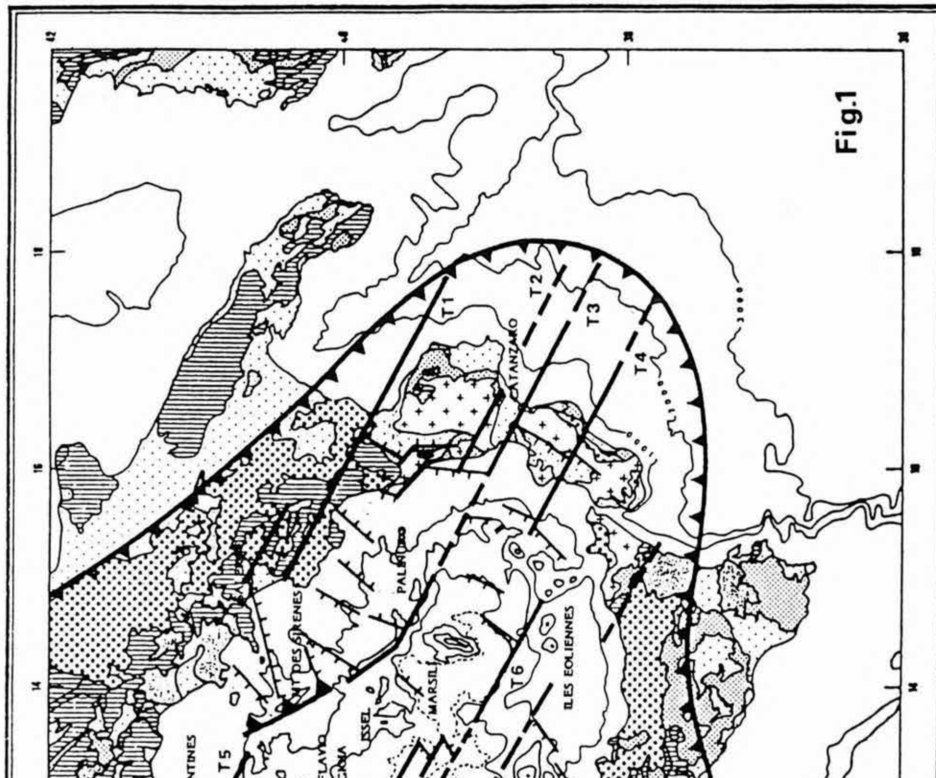


*Fig. 1.* Structural sketch map of the Tyrrhenian

ST8 location of seismic section of Fig. 2, *1* to *7* structural units, *8* isobaths, *9* main normal faults, *10* transform faults *11* thrust fault

— a set of listric normal faults whose strike is N 30 E to NS. They form the walls of both rifts which tilted blocks of continental crust have slided down.

— a set of faults cutting more or less perpendicularly the tilted blocks, the offsets of which have generated numerous small closed half grabens. These faults form facture zones, the direction of which lies in most cases approximately N 110 E/N 120 E, except the Siren seamount one, which is discussed later on (Fig. 1). Similar networks of faults have already been described on passive margins where the fractures oriented toward the ocean are considered as transcurrent faults whose strike is representative of the direction of opening—as transform faults are (LE PICHON and SIBUET, 1981).

The first contribution of the new seismic records lies in the fact that they corrobate the occurrence of such faults as well as the listric normal faults, although they do not enable to give us a better estimate of their mean direction. Nevertheless, the N 110—N 120 E direction of the opening of the Tyrrhenian is in agreement with the results of the tectonic analyses carried out in Calabria (MOUSSAT, 1983; MOUSSAT et al., 1985a).

### Tectonics and sedimentation

Except the N 140 E Siren seamount, no obvious compressive feature has been described so far whereas evidences of extensional movements can be observed everywhere.

The *post-rift sediments* are of Pliocene and Quaternary age in the deep basins, the northern part of the central one excepted. In spite of this apparent break of the extensional movements in these areas (numerous normal faults with throw of the order of 100 ms d.t.t. are still active during Pliocene and Quaternary times), the rifting is likely to be still progressing since the end of Messinian time elsewhere.

Detailed maps of the deposit centers of the successive sedimentary units observed on the seismic records as well as plots of the active faults at successive periods have shown that the main extensional movements have changed place through time: the more one goes further north of the rifts or the more one goes perpendicularly away from them, the younger they seem to be. Evidences of Pliocene, Quaternary and present active extensional faulting can be observed in the shallower areas of the Tyrrhenian margins. So, the *syn-rift sediments* are of early Pliocene age in the northern part of the central rift, whereas they are of Messinian and, at least, of late Tortonian age elsewhere in the deep basins. This last dating is based on field observations of the first extensional movements breaking up the backbone of the Apennines (AMODIO-MORELLI et al., 1976). However the opening of the Tyrrhenian Sea during the building of this chain is suggested by kinematic considerations (DEWEY and SENGÖR, 1979) —and a Middle to Upper Miocene age could be attributed to the sedimentary unit (B3) directly underlying the Messinian evaporites. However, no angular unconformity has been observed so far within this unit whereas effects of tectonic events weaker than the Middle Miocene one have been recorded by the sedimentation (unconformities of Messinian, Middle Pliocene and Middle Pleistocene (?) age). Thus this unit appears to be of upper Tortonian rather than Middle Miocene age. Few evidence of *prerifts sediments* underlying this unit have been reported (MOUSSAT, 1983; FABBRI et al., 1981). The sedimentary cover of the tilted blocks of the northern part of the central rift, on top of which the Pliocene deposits lie unconformably has been interpreted in such way (MOUSSAT, 1983). However, the upper Miocene sedimentary break observed in this area keeps us from determining its age.

Thus, the most important contribution of the new seismic records is backed up by the fact that they show the existence of thick sediments underlying the upper Tortonian

— Middle Miocene (?) ones (B3) (Fig. 2). They constitute the oldest sedimentary unit (B4) lying on the acoustic basement, observed so far in the bathyal plain.

This unit has been broken up by a drastic, probably distensive, tectonic event and then, recovered by prograding deposits of sediments coming from Sardinia. During this period of sedimentation, tilting of blocks down listric normal faults have occurred on the Sardinian margin. In spite of these facts, the variations of thickness of the lower sedimentary unit keep us from considering it as a pre-rift sedimentary sequence.

On the basis of kinematic considerations (8) as well as on the basis of field observations (REHAULT, 1981; AMODIO-MORELLI, 1976; CHERCHI et al., 1982), two hypotheses can be raised:

(a) The first rifting event is related to the opening of the Western Mediterranean, from late Oligocene (?) to Aquitanian—and is followed by the Tyrrhenian rifting s.s. starting (unconformity between B3 and B4) either in Middle Miocene time after the collision of Corsica and Sardinia against the Adriatic margin 18 m.y. ago or after the last of the main compressive events which give birth to the Apennines, i.e. in upper Tortonian time, 10 m.y. ago.

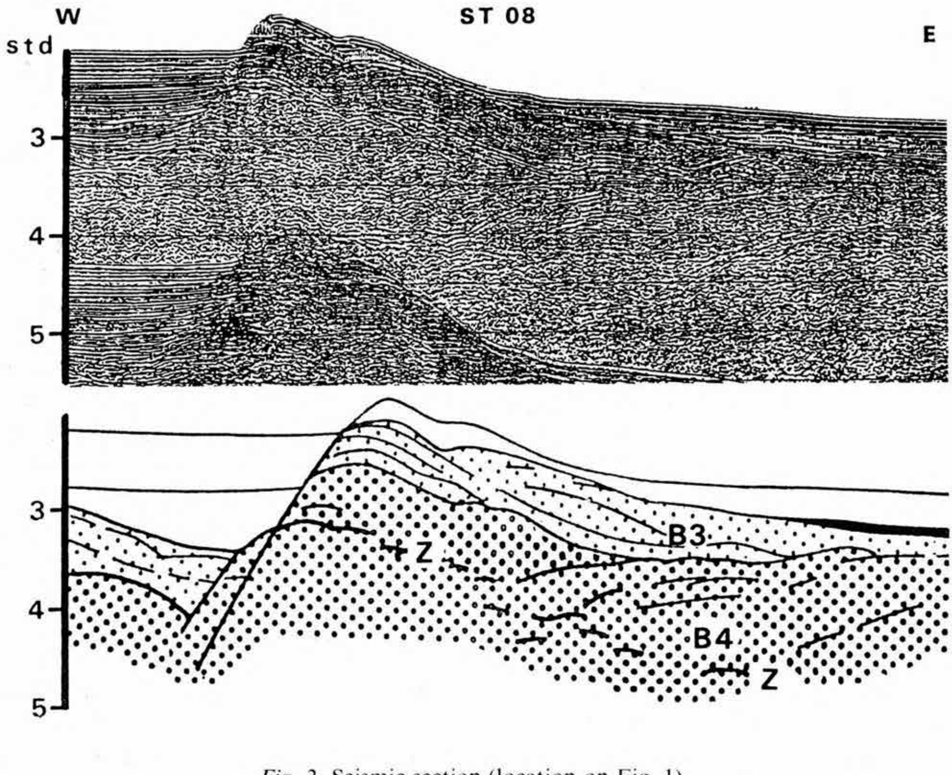


*Fig. 2.* Seismic section (location on Fig. 1)

Unit 1: Plio-Quaternary—in white, unit 2: Messinian—in black, unit B3: Middle (?) to Upper Tortonian—thin dots, unit B4: Late Oligocene to Lower Miocene (?)—thick dots. See text for discussion Z: unconformity of unit B4 on the acoustic basement—thick dots too

(b) The first rifting event started only after the continental collision of Corsica–Sardinia against the Adriatic margin 18 m.y. ago. Then, it has been followed by a second phasis of extension (unconformity between B3 and B4) which started 10 m.y. ago.

### Geodynamic

According to these observations, the Tyrrhenian sea has undergone a continuous stretching, at least from Tortonian to present time. However regional angular unconformities related to accelerations of the subsidence in Messinian, Middle Pliocene and Middle Pleistocene (?) (Fabbri et al., 1981) show that this stretching was not uniform through time. These accelerations appear correlated with compressive events in the Apennines and Sicilian chains as well as with accelerated motions of the part of the arc, free of continental foreland (Moussat, 1983; Moussat et al., 1985a). Thus the shortening between Europe and Africa is involved in the process of opening of the Tyrrhenian Sea.

However, the hypothesis of a stretching induced by the squeeze of the Tyrrhenian area is not supported by the direction of movements observed along the ESE–WSW fracture zones afore mentioned.

On the one hand, dextral movements are expected to exist along such faults on the Sicilian margin, from the southward thickening of the continental crust as well as from the EW "en échelon" directed folds of Middle Pliocene in Sicily. On the other hand, besides the northward thickening of the continental crust, the Campanian margin bears evidence of sinistral shear along such fracture zones:

(a) the opening of the pull-apart basin of Salerno (N 70 E) related to the thrust fault of the Siren seamount along the N 140 E segment of the fracture zone which runs from the northern end of the central rift to the southern ends of the tectonic troughs of the northern margin;

(b) the field observations of sinistral strike-slips along such faults in Calabria, e.g. along the SE extension of the fracture zone which bounds the northern end of the Salerno basin (Moussat, 1983; Moussat et al., 1985a, b). Such strike-slips of opposite directions, on both sides of the Tyrrhenian basin imply that the arc is draged by a force oriented toward the subduction zone rather than pushed by a force which squeezes its back arc area. The existence of such drag force is supported by the stresses analysis of the tectonic events affecting Calabria from Tortonian time to present (Moussat, 1983; Moussat et al., 1985a).

This force results from the continental collision which lock the African plate and its Adriatic promontory against Europe. Due to the low rate of convergence between these plates, the high density of the oceanic or thinned continental lithosphere makes the slab sink and shrink southeastward away from the overlying plate which is spreading out toward the subduction zone in a sort of gravity flowage. The successive riftings of the Tyrrhenian basin seem related to the same process, involving however continental collisions of second order (e.g. Corsica–Sardina/Adria collision) which lock the ends of the successive arcs, while transform faults allow the rest of them to move toward the subduction zone.

Besides, an additional drag force contributes episodically to the accelerations of the motion of the arc when the shortening seems to increase quickly between Europe and Africa. This is probably due to a force which acts on the subducted plate rather than on the overlying one, and which makes the slab shrink faster.

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# OLISTOSTROMES AND OVERTHRUSTING IN THE POLISH CARPATHIANS

by

A. ŚLĄCZKA and N. OSZCZYPKO

The flysch sediments of the Polish Outer Carpathians two types of olistolithes occur. The first type derived both from external and intra-basinal source areas contains exotic material. The second one derived from uplifted flysch sediments contains most flysch material.

The first type of olistolithes was deposited during the whole period of flysch sedimentation (uppermost Jurassic—Late Oligocene). The second one is connected with flysch sedimentation as well as with the subsequent period of folding. It is typical for the final stage of development of the Carpathian orogen (Oligocene—Miocene).

*1 Olistolithes with exotic material*

Material from the northern external source (southern prolongation of the epi-Variscian European Platform) is represented mainly by Precambrian crystalline rocks, Precambrian and Palaeozoic phyllites and carbonate rocks of Palaeozoic and Mesozoic age. The blocks are up to several metres in size. Of the sources situated within the flysch basins the most important one was the Silesian cordillera located between the Silesian and Magura basins (KSIĄZKIEVICZ, 1956). That source supplied granites, granitogneisses, psephitic and psamitic gneisses, garnetiferrous shales, effusive rocks (porphyries, dacite—andesites and felsites), pelitic and organodetritic Upper Jurassic and Berriasian limestones or sandstones with nummulites and lithothamniums (KSIĄZKIEWICZ, 1956; WIESER, 1958). An other important source area was situated along the southern margin of the Magura basin. From that source came alaskites, granodiorites, gneisses, quartzities and sedimentary rocks: radiolarites, Globochaete and tintinnid limestones (Tithonian—Berriasian), radiolarian limestones (Valanginian—Hauterivian), Urgonian limestones and Palaeocene lithothamnium—coral limestones (OSZCZYPKO, 1975). The last occurrence of the olistolithes of exotic character is known from the Lower Krosno Beds of Oligocene age. They were derived from sources situated between the Silesian and Dukla basins (phyllites, quartzites, marbles, amphibolites, zoogenic limestones (Upper Eocene), sandstones, green marls (ŚLĄCZKA, 1963) and between the Silesian—Subsilesian and Skole basins (granites, marbles, porphyries; GAWEŁ, 1932). It should be noted that in the olistostromes there are sometimes also fragments derived from the marginal part of the respective basin. During the flysch sedimentation both the external and the intrabasinal source areas were probably controlled by strikeslip faults of NWW—SEE direction.

*2 Olistolithes with predominantly flysch rocks*

These olistolithes which as a rule are bound to final stage of the Outer Carpathians development occur both within the flysch basin and in the foreland basins. The olistostrome bodies were connected with the successive emplacement of the Outer

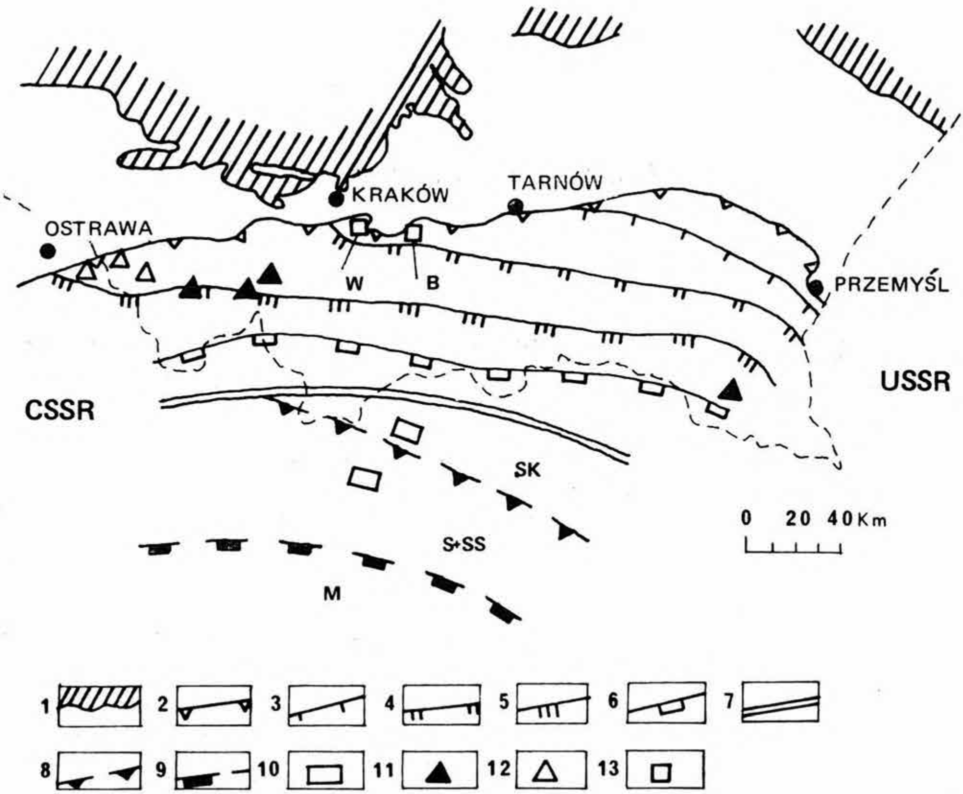



*Fig. 1.* Palinspastic sketch of the Polish Outer Carpathians and the distribution of olistostromes
*1* Northern extent of the marine Miocene in the Carpathian foredeep, *2* present-day front of the Carpathian overthrust *3* front in the Upper Badenian—Lower Sarmatian, *4* front in the middle part of Badenian, *5* front in the Lower Badenian *6* front in the Karpatian, *7* southern boundary of the platform basement, *8* front of the Silesian—Subsilesian overthrust in the Lower Miocene, *9* front of the Magura overthrust in the Lower Miocene, *10* Early Miocene olistolithes, *11* Karpatian olistolithes, *12* Lower Badenian olistolithes, *13* Middle Badenian olistolithes: SK = Residual Sloke basin, S + SS = Residual Silesian and Subsilesian basin, M = Magura nappe, W = Wieliczka, B = Bochnia

Carpathian nappes towards the north. They mainly represent marginal fragments of advancing nappes. It could be inferred that during the Oligocene all internal cordilleras which had supplied exotic rocks became subducted or covered by the flysch nappes advancing towards the north.

In the beginning of the Early Miocene the Magura nappe covered the internal ridge (Silesian cordillera) and reached the southern margin of the Silesian basin. The marginal part of this nappe became destroyed, slided to the still existing marine basin (Silesian basin) towards the north (Fig. 1, 2A) (SZYMAKOWSKA, 1976). During the next stage corresponding probably to the Eggenburgian the flysch deposits of the Silesian and Subsilesian units became folded and started to be overthrusted on the Skole basin. Consecutive olistolithes were detached from the margin of the Subsilesian nappe and deposited within the Upper Krosno beds in still existing Skole basin (Fig. 1, 2B) (JASIONOWICZ and SZYMAKOWSKA, 1963).

Subsequent olistostromes are already connected with the margin of the advanced Carpathian orogen and were deposited in the molasse foreland basin. In the Polish

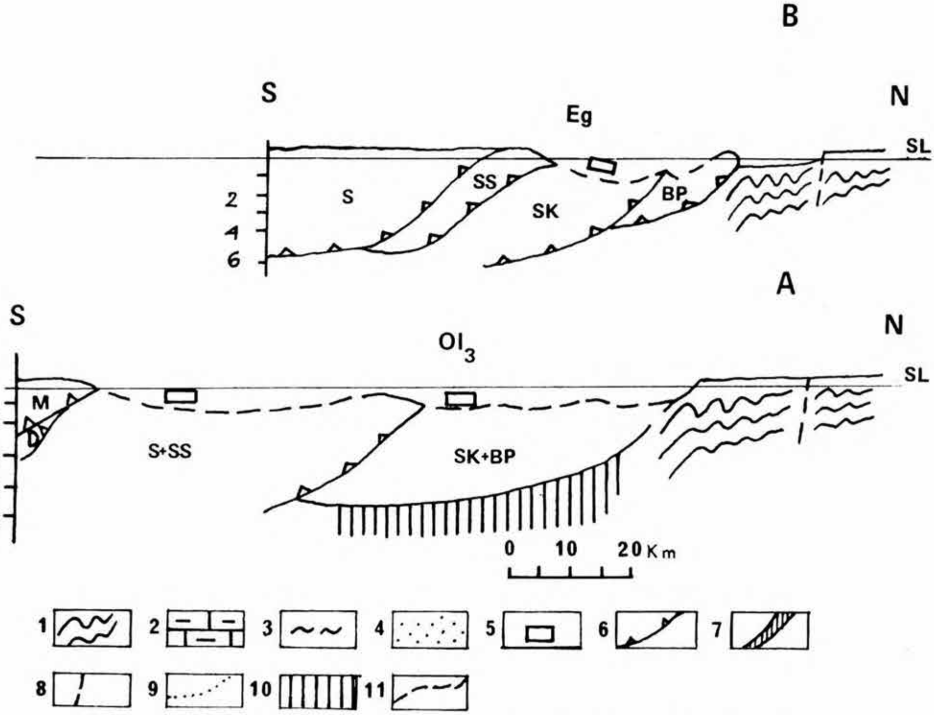



*Fig. 2.* Palinspastic cross section (eastern part of the Polish Carpathians)

*1* Precambrian and Lower Palaeozoic phyllites, *2* Upper Palaeozoic and Mesozoic, *3* Karpatian strata, *4* Lower Badenian strata, *5* olistolithes, *6* overthrusts, *7* Miocene scales, *8* faults, *9* area covered by the molasse deposits, *10* basement of the Carpathian flysch basins, *11* area covered by turbidite deposits: M = Magura nappe, D = Dukla nappe, S = Silesian nappe, SS = Subsilesian nappe, SK = Skole nappe, BP = Boryslav—Pokutie unit: A = Late Oligocene, B = Eggenburgian

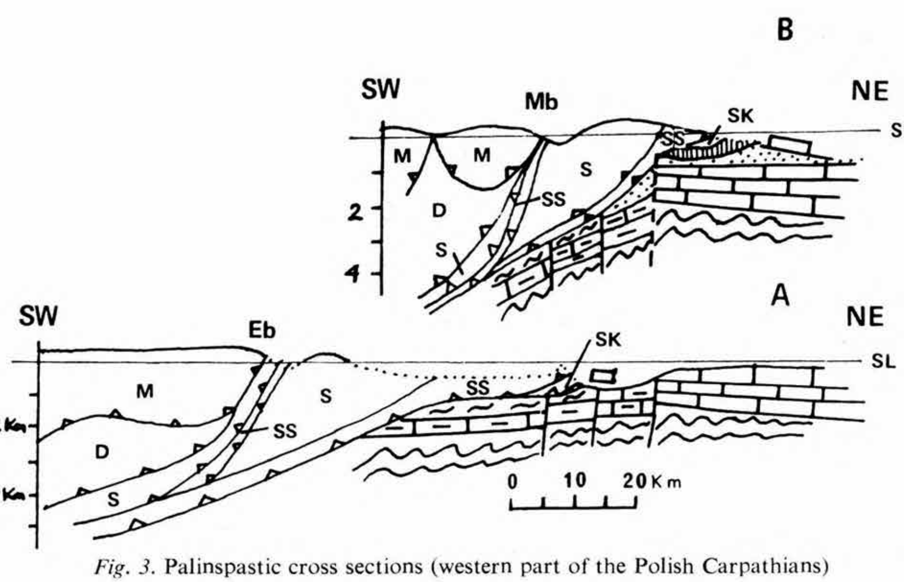



*Fig. 3.* Palinspastic cross sections (western part of the Polish Carpathians)

A = Lower Badenian, B = middle part of Badenian. Legend as in Fig. 2

Carpathians multistage movements are suggested by the occurrence of flysch olistostromes, known from the autochtonous and para-autochtonous Miocene molasse of different age. Palinspastic reconstructions (OSZCZYPKO and TOMAS, 1985) allow to establish the position of the margin of the Carpathians during the Karpatian and Sarmatian ages (Fig. 1).

The oldest olistostromes (Ottnangian—Karpatian) in the Polish part of the foreland basin have been recognized in its western part (borehole Sucha IG 1, ŚLĄCZKA, 1977) (Fig. 4, 7). The olistostromes, about 150 m mighty, are imbedded in red sandstones and shales. It suggests subareal condition of their origin. Clast composition is quite variable. In addition to fragments of the Subsilesian nappe (variegated shales and marls), clasts derived from the Silesian nappe (Upper Cretaceous sandstones and shales) and from older Miocene (anhydrite) also occur. The clasts range from few centimeters to more than ten metres in size.

In those time the margin of the Carpathians was situated from 45 to 80 km further to the south from its present position in the western and eastern parts, respectively (Fig. 1). It is worth to note that just after the deposition of the olistostromes a part of the foreland was uplifted and platform material was transported into the basin.

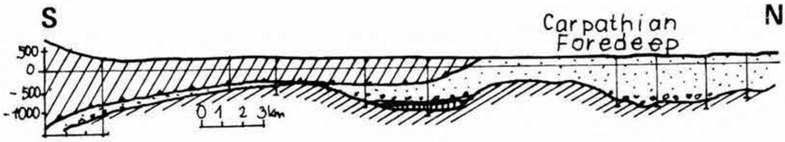
After the Karpatian, the margin of the Carpathian orogen shifted northwards covering the Sucha olistostromes, and SW from Ostrava (Moravian Gate) reached its present position (Fig. 1). Eastwards, the Carpathian margin was situated farther to the south as compared to its present position. Near Cieszyn, in the earliest Badenien, great slumps descended from the Carpathians margin towards the foredeep basin (BUŁA and JURA, 1983) (Fig. 6). In fact, on the Karpatian mudstones there is a unit of olistostromes up to 67 m thick, build up by blocks derived from the Subsilesian nappe. Similar deposits occur in the vicinity of Ostrava (comp. JURKOVA, 1976) and are interpreted here also as olistostromes (Fig. 5). The olistostromes pass gradually upwards, as in case of the Sucha ones, into massive conglomerates (Dębowiec conglomerate—Lower Badenian). The well rounded pebbles and clasts were derived generally from the platform and in a smaller amount also from the Carpathians. After the sedimentation of the conglomerates, still in the Lower Badenian, marine transgression affected the major part of the Foreland and the Carpathian orogen was being shifted towards the north, reaching in Kraków area its present position. East of Kraków the margin was situated more and more distant from its present position (Fig. 1). The area of slumping which descend from the Carpathian margin was also shifted towards the east during the middle part of Badenian it was situated SW from Kraków, between Wieliczka and Bochnia. These olistostromes are built up not only by material of the Carpathians but mainly by material derived from the southern margin of the foredeep basin (Badenian salt and marls; KOLASA and ŚLĄCZKA, 1985) (Fig. 8). The sequence is terminated with olistostromes consisting mainly of Karpatian rocks. It is a last occurrence of the olistostromes in the Carpathian foredeep, terminating the process of migration of slumping along the Carpathians. Farther to the east there are no more olistostromes. Only in vicinity of Rzeszów occur conglomeratic fan deposits (Upper Badenian) as probably equivalents of the olistostromes of the western Carpathian foreland.

The main reason for the lack of olistostromes in the Carpathian foreland east from Bochnia is that this part of the Carpathians was flat during the Badenian and Sarmatian. This interpretation can be supported by the following data:

*1* the margin of that part of the Carpathians was beeing overflooded in Badenian and Sarmatian times, and the margin of the Carpathians was covered by marine sediments with lithothamnian reef patches;

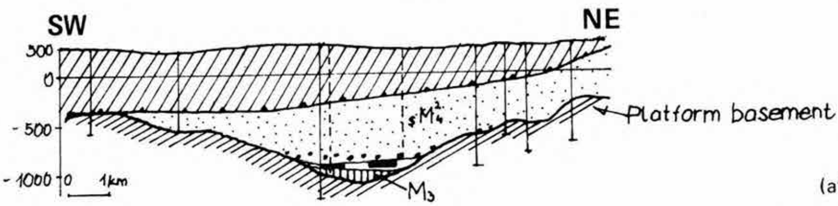
Fig. 4


Fig. 5



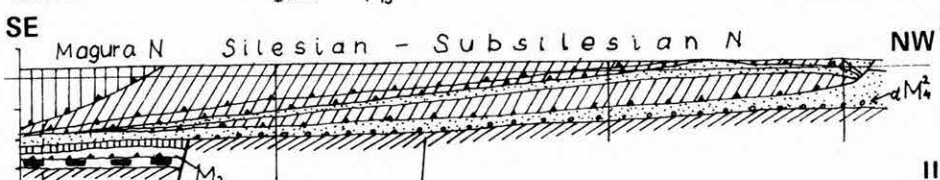

I—I
(after Jurkova, 1976)

Fig. 6



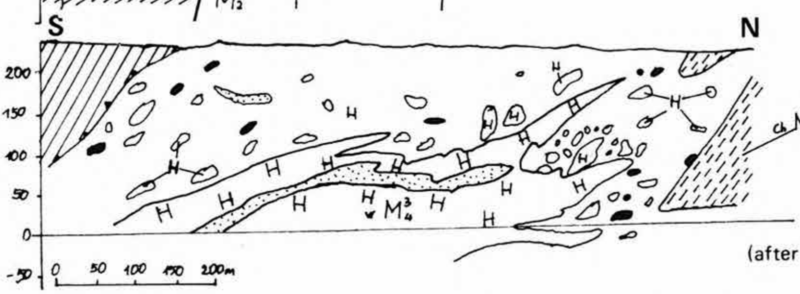

II—II
(after Bula, Jura, 1983)

Fig. 7




III—III

Fig. 8




IV—IV
(after Kolasa, Ślaczka, 1985)

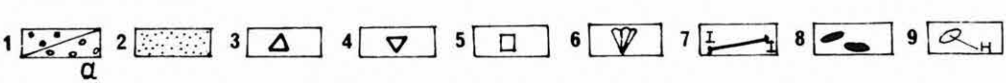



*Fig. 4—8*. Distribution map of the flysch olistostromes and geological cross sections

*1* Debowiec conglomerate, *a* = beneath of the Carpathians, *2* Miocene deposits on the Carpathians, *3* olistostromes within molasses ($M_2$—$M_3$), *4* olistostromes within flysch ($M_1$), *5* salt olistostromes ($M_4$), *6* fans, *7* cross sections, *8* flysch olistostromes (Fig. *5*—*7*), *9* salt olistrostromes (Fig. 8)

*2* in the internal part of the Carpathians a marine basin (the Skole basin) survived up to the Miocene, according to some ideas even up to Badenian (KOSZARSKI, 1985);

*3* east of Bochnia submarine conglomeratic fans connected with the Carpathians are less numerous than in the western part. In fact, only one fan is known, of Upper Badenian age (vicinity of Rzeszów, Fig. 4) which, moreover, is lying on the Carpathians (DOKTOR, 1983). The younger, Lower Sarmatian coarse grained sediments are unknown. After the Lower Sarmatian, the marginal part of the Carpathians between Tarnów and Przemyśl underwent the last episode of thrusting towards the northeast, reaching the present day position (Fig. 1).

The occurrence of the olistostromes and conglomerates derived from the Carpathian orogen is due to the complex interaction of tectonic and sedimentary processes and they record consecutive Miocene stages of the Carpathian overthrusting movements. Simultaneously with the overthrusting, the NE directed migration of the Miocene depositional centres took place. The most intense subsidence and accumulation occurred as a rule in front of the Carpathians with respect to their contemporary position (OSZCZYPKO and TOMAS, 1985). These processes were accompanied by the development of the olistolithes on the frontal part of the Carpathians. The olistostrome processes were being shifted along the advancing Carpathians. The older and more internal olistostromes occur in the western, while younger and more outer ones in the eastern part of the Carpathians.

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# BLOCKS OF WEST CARPATHIANS AND NEOGENE MOLASSE BASINS

by

D. VASS

According to the analysis of the regional and residual anomalies of the gravity field and DSS results in the West Carpathian region several expressive deep-seated physical boundaries can be determined. They probably reflect deep-seated faults (Fig. 1) associated with increased seismic activity, with distinct boundaries in the geothermal field and with recent vertical movements of the Earth's crust (FUSÁN et al., 1979).

Deep-seated faults disturb even the Moho discontinuity and divide the West Carpathian region into several crustal blocks. The crustal blocks are subdivided into partial blocks bounded by active, mostly seismically active faults penetrating the upper part of the crust.

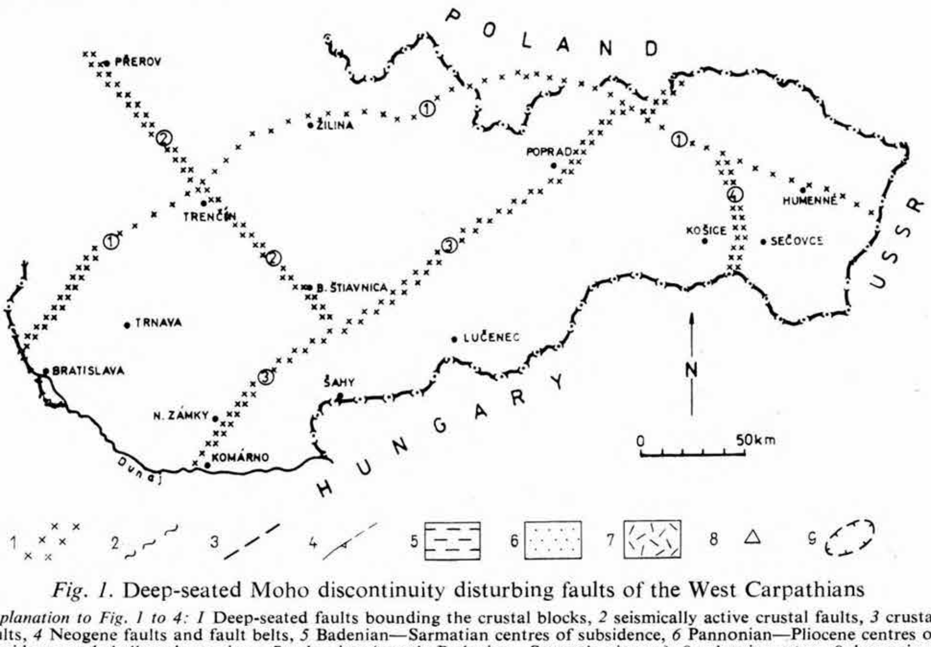


*Fig. 1.* Deep-seated Moho discontinuity disturbing faults of the West Carpathians

*Explanation to Fig. 1 to 4: 1* Deep-seated faults bounding the crustal blocks, *2* seismically active crustal faults, *3* crustal faults, *4* Neogene faults and fault belts, *5* Badenian—Sarmatian centres of subsidence, *6* Pannonian—Pliocene centres of subsidence and shallow depressions, *7* vulcanites (mostly Badenian—Sarmatian in age), *8* volcanic centres, *9* depressions identified by geophysics near the crossing and along the Záhorie—Humenné and Štiavnica—Přerov deep seated faults.— Numbers in circles are symbols for: 1 Záhorie—Humenné fault, 2 Štiavnica—Přerov fault, 3 Vepor fault, 4 Slanec fault, 5 Ludince fault, 6 faults bounding the graben of Žitny ostrov, 7 Hurbanovo fault, 8 Komárno fault belt, 9 Hron fault, 10 Turiec depression, 11 Žiar depression, 12 Horná Morava depression, 13 Močarany—Topla fault belt, 14 Trebisov fault belt, 15 Falkusovce fault belt, 16 Čičarovce fault belt, 17 Levice—Turovce horst, 18 Šahy—Lysec volcanotectonic zone, 19 Strháre—Trenč graben, 20 Kunesov—Tisovnik volcanotectonic zone, 21 Hornád fault, 22 Komárovce depression, 23 Čiž horst, 24 Kesov blocks

The breaking of the West Carpathians crust into crustal blocks may be geneticaly linked with the Middle—Late Miocene (16—11 m.y. B.P.) uplift and the creation of a mountain system (a morphogen). Recently the Middle Miocene uplift and emergence of West Carpathian crystalline rocks and intrusive bodies, originally deeply buried, are well proved by fission track method applied on apatites from the granitoid rocks of the West Carpathians (KRÁL', 1977).

It must be stressed that the Middle and Late Miocene was a period of fading out compression in the Outer West Carpathians (i.e. the Outer Flysch Zone).

The genesis of crustal blocks should be reflected in the development and structure of the molasse basins and in volcanic activity. The following general features of molasse basins and volcanic activity can be linked with the genesis of crustal blocks during the Middle and Late Miocene (VASS, 1980; 1981):

*1* culmination of the main molasse development,

*2* culmination of sedimentation rates in intramontane molasse basins and depressions (more than 20 cm/110 y.),

*3* radical change in the structural pattern of intramontane basins,

*4* culmination of fault activity including synsedimentary fault movements,

*5* culmination of andesite volcanism.

### Manifestations of crustal blocks structure—in particular molasse basins

The Danube crustal block (Fig. 2), bounded by a segment of the Záhorie—Humenné, Štiavnica—Přerov and Vepor deep faults was the home land of two molasse basins: the Galanta—Trnava intramontane Middle Miocene basin and the Gabčíkovo Upper Miocene—Pliocene basin which was a part of a large Pannonian—Pliocene basin in the back-deep area (predominantly on Hungarian territory). The Štiavnica—Přerov deep fault bounded the NE margin of both basins.

Near the crossing of the Záhorie—Humenné and Štiavnica—Přerov deep seated faults and along the both faults there are peculiar relatively deep depressions some of them mostly identified by geophysics. They are probably filled with Middle Miocene sediments (Beckov and Trenčín depressions), but the Bánovce depression is filled up partly with Lower Miocene sediments too. The Vepor deep fault between Komárno and Nové Zámky bounded on the E the Gabčíkovo basin principal depocentro.

The subsidence of the Danube block was differentiated and controlled by NW and NE crustal faults. The most important of them was the seismically active Ludince (Ölved) fault (K. TELEGDI-ROTH, 1929 fide T. BUDAY et al. 1967). The principal depocentres of the Trnava—Galanta basin filled with Badenian—Sarmatian deposits and volcanites 2000—3000 m thick are distributed around and to the NE of the fault. On the begining of the Pannonian an inversion in subsidence take place and gave rise to the Gabčíkovo basin. Its principal depocentre is situated to the SW of the Ludince fault.

Other seismically active NW fault passing from Komárno to Pezinok and another passing along the recent valley of Danube between Bratislava and Klížska Nemá limit the graben of Žitný ostrov (POSPÍSIL et al., 1978) and were active also during the Quaternary.

The NE trending crustal faults of the Danube block control the inner structure of the Galanta—Trnava basin dividing it into a system of horsts and grabens. One of the faults bounds the NW mangin of the basin.

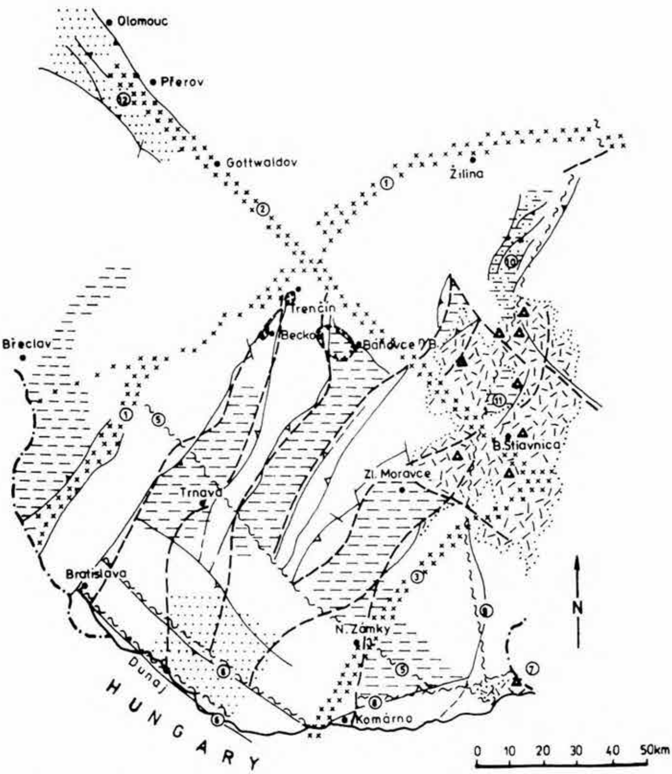


*Fig. 2.* The Danube crustal block and the structural features of alpine molasse basins reflecting the structure of the crust (for explanation see Fig. 1)

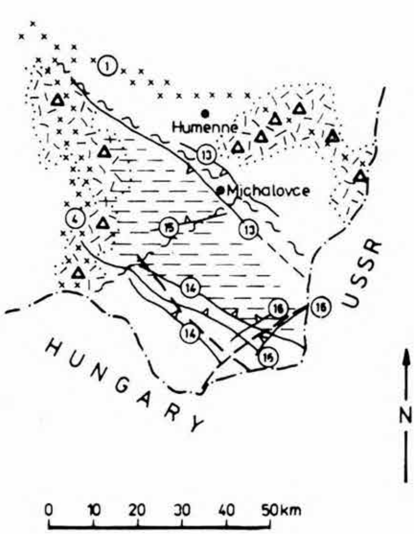


*Fig. 3.* The East Slovakian crustal block, the structural features of the East Slovakian alpine molasse basin reflecting the structure of the crust and spatial relation of Neogene volcanism to the deep-seated faults (for explanation see Fig. 1)

The east Slovakian crustal block is bounded by the Slanec and by a segment of the Záhorie—Humenné deep faults (Fig. 3). On the block there is the central part of the east Slovakian intramontane basin, east of the Slanec deep fault is located the most important depocentre of the basin filled mostly with Badenian and Sarmatian deposits and volcanoclasts, ca 4000 m thick. W of the deep fault the deposits of the same age are substantially thinner. There is also a Spatial relationship between the Slanec deep fault and the volcanic chain of the Prešov—Tokaj (or Slanec) Mts, Záhorie—Humenné fault and Vihorlat—Popričný volcanic Mts. The age of the volcanites is mostly Sarmatian. The east Slovakian block is dissected by crustal faults. One of the NW trending faults correponds to the Močarany—Topl'a fault belt and another to the Trebišov fault bolt as superficial structure (BUDAY et al., 1967). Both fault belts limit the central depression of the east Slovakian basin.

Seismically active fault trending NE wards corresponds to the Falkušovce fault belt, but the fault belt was active mostly in the pre-Badenian and Early Badenian times, respectively (ČVERČKO, 1977). Another crustal fault of the same direction corresponds to the Čičarovce fault belt (ČVERČKO, 1977) of superficial structure bounding SE margin of the principal depocentre of the east Slovakian molasse basin.

The Rudohorie—Pilis crustal block bounded by Vepor. Slanec and by a segment of the Záhorie—Humenné deep faults (Fig. 4) was mostly uplifted from the Middle Miocene till Pliocene.

The Vepor deep fault beside the role in the Gabčíkovo basin development had a deal for the origin of the Central Slovakian Volcanic Mts. The Badenian and Sarmatian volcanites are symmetrically distributed on both sides of the deep fault.

A crustal fault parallel to the Vepor deep fault corresponds to the Šahy—Lysec volcanotectonic zone of superficial structure with the volcanic manifestation from the Late Badenian till the Late Sarmatian (V. KONEČNÝ in VASS et al., 1979). The continuation of the crustal fault to the NE follows the northern margin of the Lučenec and Rimava molasse basins. The fault probably originated a slow and unextensive subsidence in the Uppermost Miocene resulting in sedimentation of the Poltár formation (Pontian age) with important ceramics raw material deposits. The seismically active Burbanovo fault had no direct on the thickness distribution of Badenian—Sarmatian molasse deposits, but had an important influence on the course of the Komárno fault system, causing a deviation from the SE trend to W—E. The Komárno fault system is an important structural element for Middle and Upper Miocene molasse sediments (B. GAŽA and M. BEINHAUEROVÁ, 1977).

The seismically active N—S crustal fault known as the Hornád fault (UHLIG, 1907) bounds the W margin of the east Slovakian molasse basin between Prešov and Košice.

Another N—S trending fault in the valley of the Slaná river tooks part in the structure of the Rimava molasse basin limiting the Číž horst and the Kesov blocks (D. VASS et al., 1985). The Číž horst is a rising structure till now. The fault controlled the distribution of Upper Miocene deposits in the E part of the Rimava basin but also the distribution of Quaternary river terraces.

The Slovak Karst and the Komárovce depression are limited by a crustal N—S fault. The Komárovce depression is a young one and is filled with Badenian—Pliocene sediments and volcanites.

In the frame of the Rudohorie—Pilis crustal block there are two crustal faults trending NW wards. Superficial manifestation of a crustal fault on the boundary between the Lučenec and Ipel' basins is the Strháre—Trenč graben (VASS et al., 1979) synsedimentary active during the Badenian. The NW continuation of the crustal fault

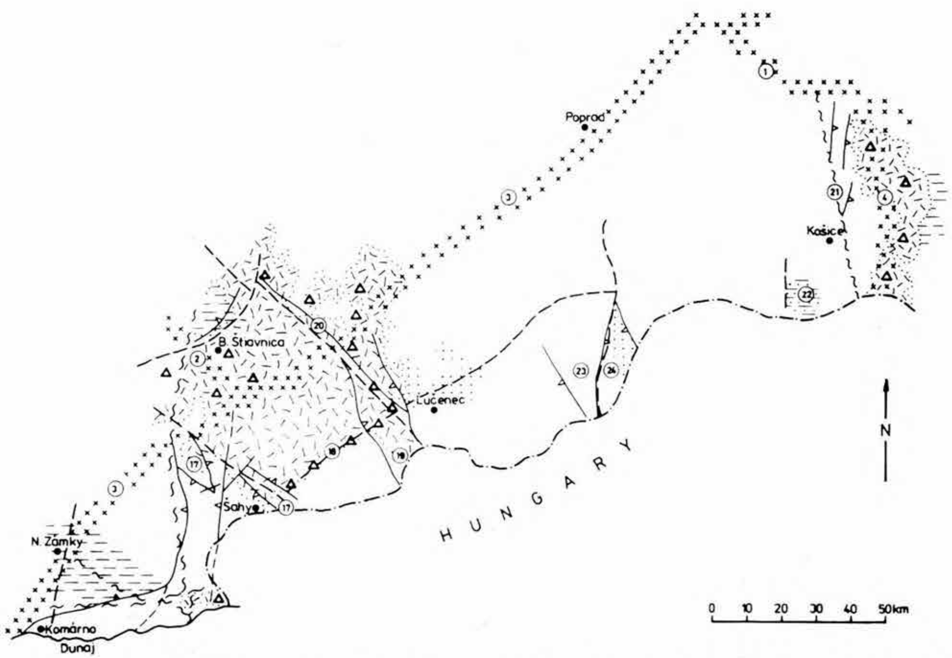


*Fig. 4.* The Pilis—Rudohorie (South Slovakian) crustal block, some crustal faults with the manifestation on the surface and relation of Middle Slovakian Neogene volcanites to the structure of the crust (for explanation see Fig. 1)

is represented on the surface by the Kunešov—Tisovník volcanotectonic zone (KONEČNÝ et al., 1978). Another NW crustal fault is overlapped by the Badenian Levice—Turovce horst (L. MELIORIS and D. VASS, 1982). The horst influenced essentially the evolution of Badenian sediments in the NE part of the Danube lowland.

The Slovak—Moravian crustal block bounded by the Záhorie—Humenné and by the Štiavnica—Přerov deep faults (Fig. 2) was an area of differentiated subsidence. Its southern part subsided during the Neogene (Vienna basin). The Záhorie—Humenné deep fault between Devínska Nová Ves and Jablonica limited SE border of the Vienna basin in the Middle and Late Miocene. The Štiavnica—Přerov deep fault gave rise to the Horná Morava depression which is a shallow graben filled with Pliocene sediments.

An important seismically active N—S crustal fault is running from Zázrivá to Štúrovo—Budapest and farther to the S follows the valley of Danube. Superficial manifestation of this crustal fault is the so-called Zázrivá—Budapest fault belt (KUBÍNY, 1962). Near Zázrivá it epigenetically disturbs the Klippen belt with a small strike-ship displacement. Small intramontane basins: the Turčianska kotlina and the Žiarska kotlina depressions filled mostly with Sarmatian—Panonian sediments generated on the mentioned fault belt. With the fault belt is connected the activity of middle Slovakian volcanism and metallogenesis (Badenian—Sarmatian in age, STOHL, 1976). Farther southwards the superficial manifestation of crustal fault is the northern branch of the Kravany—Hron fault (SENEŠ, 1963) bounding the E margin of the Panonian—Pliocene of the Gabčíkovo basin. The fault follows the lower valley of the river Hron predicting the distribution of the river Quaternary terraces.

Some deep-seated faults disturbing the Moho discontinuity and crustal faults apparently were active in the Early Miocene and/or in pre-Neogene time. Some surface faults coinciding with the Štiavnica-Přerov deep-seated fault were evidently active in pre-Neogene time (Skýcov fault, BIELY, 1962, Hrádok and Jastrabie fault, MAHEL, 1969), faults of the Horná Morava depression are pre-Devonian (RÓTH et al., 1962).

The Falkušovce fault felt in east Slovakian basin, as already mentioned, was active in the Early Miocene (ČVERČKO, 1977).

The crustal fault reflected on the surface by the Trebišov fault belt (east Slovakia basin) can be identified with Szamos line originating according to some authors in pre-Neogene time (T. BUDAY 1961, in T. BUDAY et al. 1967; P. GRECULA and I. VARGA 1979).

The Zázrivá—Budapest fault belt was active in the Eocene. The fault belt is responsible for a mobile zone by which the Buda and Inner Carpathian (Central Carpathian or Podhale) Paleogene were joined to each other (SAMUEL, 1973; VASS et al., 1979).

The crustal fault, manifested on the surface as the Šahy—Lysec volcanotectonics zone, is one of the NE fault system disturbing the pre-Neogene fundament of Krupinská planina Mts and Ipel'ská kotlina basin (S Slovakia). The fault belt come to existence in pre-Neogene time (VASS et al., 1979).

*Conclusions*. The blocks of the West Carpathians defined by geophysics reflected in the development of Neogene molasse basins and volcanism. A number of tectonic manifestations important for molasse forming epoch coincide in time with the supposed formation of crustal blocks (Middle—Late Miocene).

The supercrustal, deep—seated faults and crustal faults are manifested in the structure of molasse basins in the following manner:

- they control the principal depocentres and some of the molasse basin margins,
- they generated some small intramontane basins,
- young volcanic mountains are in close spatial relation with them,
- there is evident genetic relation between them and metallogenesis,
- surface manifestations of them are
- fault belts active in Middle—Late Neogene
- horst and/or grabens,
- volcanotectonic zones.

It should be noticed that some deep-seated, Moho discontinuity disturbing faults and crustal faults can be identified with the fault engaged in the tectonic or palaeogeographic evolution of the West Carpathians in pre-Badenian and/or pre-Neogene time.

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# THE MIDDLE SECTION OF THE ALPINE—MEDITERRANEAN BELT IN THE NEOGENE

by

Z. BALLA

For the Neogene, many reconstructions of the Tyrrhenian and Aegean regions exist. The link between these areas, i.e. the Adriatic region, however, is much less examined in this relation likewise the whole Alpine—Carpathian—Balkan realm. My goal is to outline the Neogene history of the whole middle section of the European Alpine—Mediterranean belt (Fig. 1) starting and ending with comparatively well known areas in the west and east (for details, see BALLA, 1986).

My reconstructions are based on the mutual position of Africa and Europe according to new French—Soviet results presented at the 27th International Geological Congress in Moscow. I have carried out reconstructions in four steps. In the first step, domains of constant shape are arranged. This is a rigid reconstruction containing a lot of defects which can be eliminated by means of correction of domain shapes i.e. of plastic deformations of the domains. This is a second-step or corrected reconstruction in which after the first correction some defects remain requiring additional corrections also by means of plastic deformations. This third-step or additionally corrected reconstruction is the base for the resulting picture, i.e. for the fourth-step or final reconstruction in which gaps are attributed to and overlaps are excluded from certain domains according to geological data. Final reconstructions, really, are palaeotectonic schemes.

In the first-step rigid reconstruction of the situation for 10 Ma Africa has been moved back relative to Europe, and a gap has been opened between them. The main problem of the reconstruction is where to open this gap. I start my analysis with the Alpine section. The simplest way is to keep Adria integral with Africa and to open the gap on the front of the southern Alps. It is known from palaeomagnetic data that the southern Alps have been rotated clockwise by about 10—15° relative to Adria. The most probable time for this rotation is within the time span under consideration, and the rotation means total elimination of the displacement along the Giudicaria line in the corrected reconstruction. By means of some deformations I pass the gap from the Periadriatic line on the northern boundary of the Alps in the additionally corrected reconstruction.

The Carpatho—Pannonian domains remain in their present position. In the first-step rigid reconstruction I keep the Lika—Dinara domain of the Dinarids integral with the southern Alps while the external zones are moved together with Adria and the internal zones are moved along the Zvornik line. Little deformations are enough to adjust internal zones with the Vardar—Pannonian domains in the corrected reconstruction. By means of stronger deformations the intra-Dinaric gap can be transferred on the Dinaric front in the additionally corrected reconstruction.

In the Hellenides the Ionian zone of Greece is rotated back according to palaeomagnetic data in the first-step rigid reconstruction and a gap is opened between the

Ionian zone and Africa in harmony with data on the recent subduction in the Hellenic trench. The alignment with the Dinarids requires no rotation of the Ionian zone of Albania. In the corrected reconstructions, the gap in the Hellenides can be widened by means of deformation of the Pelagonian and Vardar zones, of the Serbo—Macedonian and Rhodope massifs and by means of strike-slip movement along the Maritsa fault.

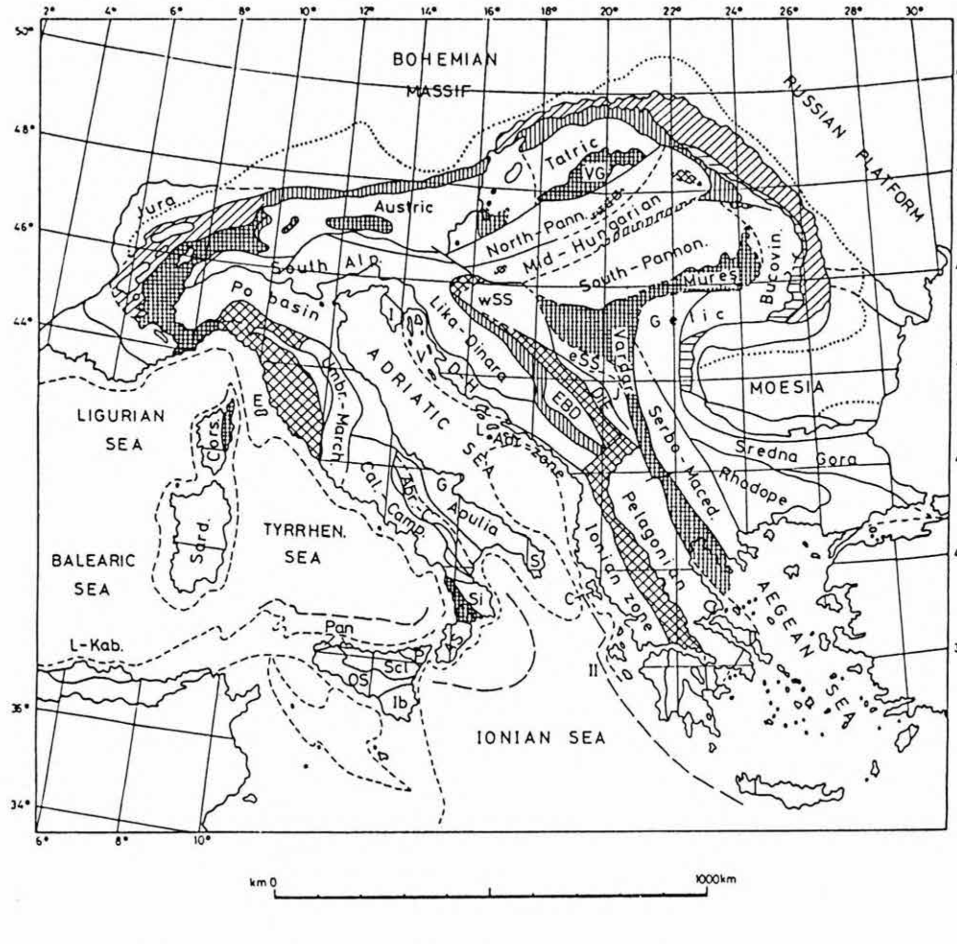


*Fig. 1.* Tectonic scheme of the middle section of the Alpine—Mediterranean belt (BALLA, 1986) —*4* Flysch, *1* Helvetic (Alps), Silesian (West Carpathians) and Moldavian (East Carpathians), folded, *2* same, unfolded (Fig. 3), *3* Rhenodanubian (Eastern Alps, Helvetic flysch included), Magura (West Carpathians), Maramureș (in the rear of the East Carpathians), Szolnok (Mid-Hungarian zone), Sarajevo (Central Dinarides), folded, *4* Ceahlău (East Carpathians) Severin (South Carpathians, neighbouring domains included), folded; *5—6* ophiolite-bearing belts, *5* in geosutures, Penninic (Alps), Vepor—Gemer (West Carpathians), Mureș (Transylvania), Vardar (Balkan peninsula), *6* possibly, in large nappes, Ligurian (northern Apennines), fragments in the Mid-Hungarian zone and its surroundings, ophiolite belt (central Dinarides), Subpelagonian (central Hellenides), *7* present geographical contours or their traces, geological boundaries, *8* on land: same, inferred, in seas: shelf contours, *9* contours of deep-see basins (selected), *10* foredeep contours, *11* hypothetical contours (Figs. 2 and 3). Letter symbols: Adr. = Adriatic, Alp. = Alpine, AS = Aspromonte—Serre, Bucov. = Bucovinian, C = Corfu, Cors. = Corsica, D—H. = Dalmat—Herzegovinian, DIE = Drina—Ivanjica element, E = Elba, EBD = = east Bosnian—Durmitor, eSS = eastern Slavonia—Srem, G = Gargano, Getic = Getic + Danubian, II = Ionian islands, = Jadar, March. = Marches, Pann. = Pannonian, S = Salento, Sard. = Sardinia, Serbo-Maced. = Serbo-Macedonian, Si = Sila, Umbr. = Umbria, VG = Vepor and Gemer, wSS = western Slavonia—Srem

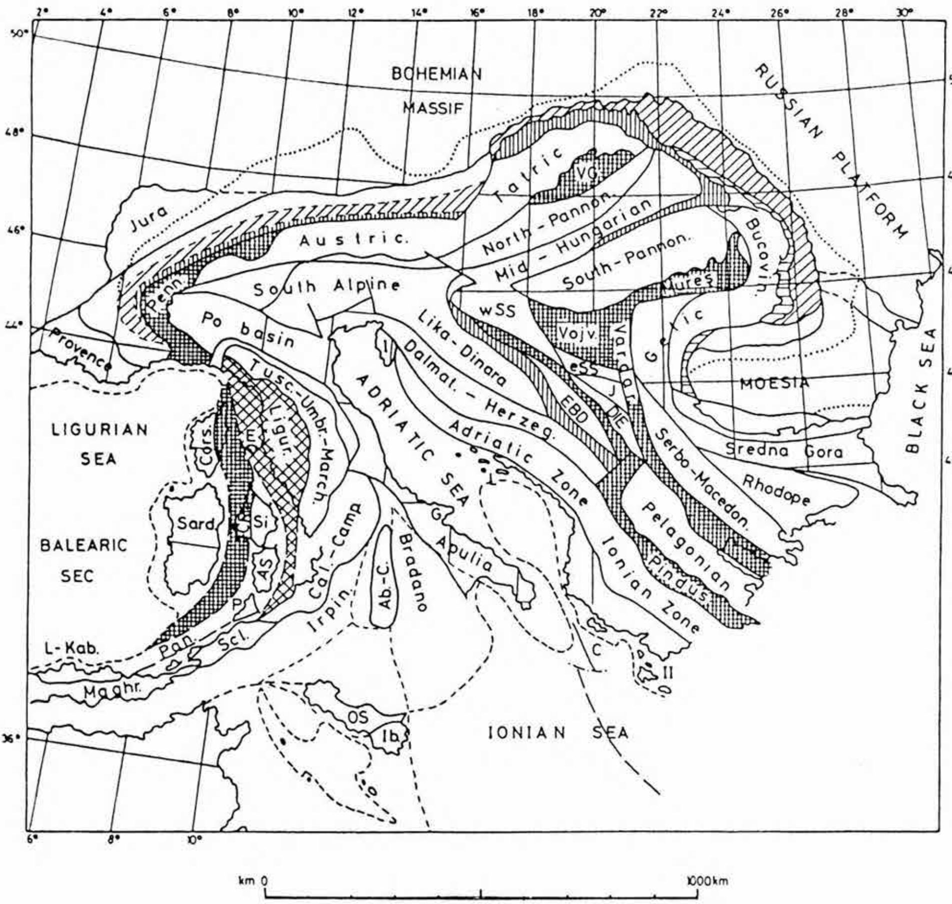


*Fig. 2.* Palaeotectonic scheme for 10 Ma. For legend, see Fig. 1

Letter symbols: Ab.—C = Abruzzi—Campania platform, Cal.—Camp. = Calabro—Campania platform, CR = Coastal range, Dalmat.—Herzeg. = Dalmatian—Herzegovinian zone, I = Istria, Ib = Iblean platform, Irpin. = Irpinian basin, Ligur. = Ligurian domain, L—Kab. = Lesser Kabylia, Maghr. = Maghrebides, Pan. = Panormides, Penn. = Penninics, OS = outer Sicilides, Scl. = Sclafani domain, Tusc. = Tuscany, Vojv. = Vojvodina

In the west the Tyrrhenian Sea is closed, and the surrounding domains are arranged according to well known models.

The resulting picture is the palaeotectonic scheme for 10 Ma (Fig.2). It is quite similar to the present one except for the Tyrrhenian and Aegean regions. I call attention to the main problem of the scheme: on the Dinaric—Hellenic front a gap has been opened which is inconsistent with geological data. It could be elimiated when supposing that Adria was not integral with Africa but moved towards the west or rotated anticlockwise around a pole in the north due to its pushing by the Turkish microplate. This possibility would yield numerous geometric solutions instead of the only one outlined above. Now I work on this problem and preliminarily have concluded that no important changes can exist in the general picture although many domains will have another shapes. That is why I have decided to present here further reconstructions which are based on the uncorrected situation (Fig. 2).

I start the reconstruction of the situation for 20 Ma with the Carpatho—Pannonian region. Here a large gap is opened due to the rotation back of Pannonian domains, the curvature of the Carpathians is much decreased, and a small gap is opened between the Maramureş spur and the Magura—Pieniny zone which is responsible for the Early Miocene thrusting in this area (for details, see BALLA, 1984.)

In the Alps the first-step rigid reconstruction results in a gap on both sides of the Drava range. By means of deformations I pass this gap on the outer boundary of the Alps in the corrected reconstructions.

According to palaeomagnetic data, the autochtonous domains of Adria, i.e. Istria, Umbria and Gargano have to be rotated back clockwise by about 25°. In the determination of the rotation pole I use following considerations: (1) the pole has to lie close to the Alpine—Apenninic whirl centre; (2) the rotating domain must have a transform boundary with the Alpine and Pannonian domains; (3) as it seen in the rigid reconstruction the rotating domain can not include the whole Dinaric system since in this case the Dinaric—Hellenic alignment would be interrupted, therefore, the southeastern transform boundary of the rotating domains has to lie north of the Shkodër—Peć line; I suppose it in the Gargano—Lastovo line.

The fit of the Dinarids with the Hellenides requires deviation of endings of both systems towards each other in the corrected reconstruction, i.e. no rotation for the southeastern third of the Dinarids and anticlockwise rotation of the Albanian section. The overlap of the Serbo-Macedonian massif is eliminated by means of its strong deformation. The Vardar domains have to be arranged in the gaps of the intra-Hellenic area. It is possible by means of their strong deformation in the additionally corrected reconstruction. The gap between the Inner Dinaric and South Pannonian domains in the rigid reconstruction can be passed on the front of the Dinarids by means of deformations in the corrected reconstructions.

In the west the Ligurian sea is closed, and the surrounding domains are arranged according to well known models, except for the southeastern areas. From the rigid reconstruction my main conclusion is that the palaeomagnetically detected rotation of Gargano, Umbria and Istria did not concern the southern part of the Adriatic domain. In the corrected reconstruction, only areas north of the Gargano—Lastovo line suffered this rotation and only these areas can be named "Adriatic microplate". In the additionally corrected reconstruction, between this unit and the African plate I delineate a new unit which moved towards the northeast relative to Africa with no rotation. It can be named "Apulian microplate" and consists of Apulia, Abruzzi—Campania and Calabro—Campania with basins between them. Its western boundary is unclear.

In the palaeotectonic scheme for 20 Ma (Fig. 3) the most imposing element is a mushroom-like body encircled by ophiolite-bearing and flysch zones. It is integral with Africa and penetrated into Europe and can be considered as the Adriatic promontory of numerous authors. The widening of the head of the promontory is clearly due to squeezing out during the continent-continent collision in the present Alpine area. Further moving back of Africa will allow further narrowing of the head of the promontory which necessarily means further shortening of the Alpine—West Carpathian range. This shortening is the main condition for its moving towards the south together with Africa to produce gap for the Palaeogene orogeny in the Alps. Besides this generality, we can note some unexpected and striking phenomena as follows:

*1* All Austric units of the Alps are assembled together and cover most of the Alpine domain.

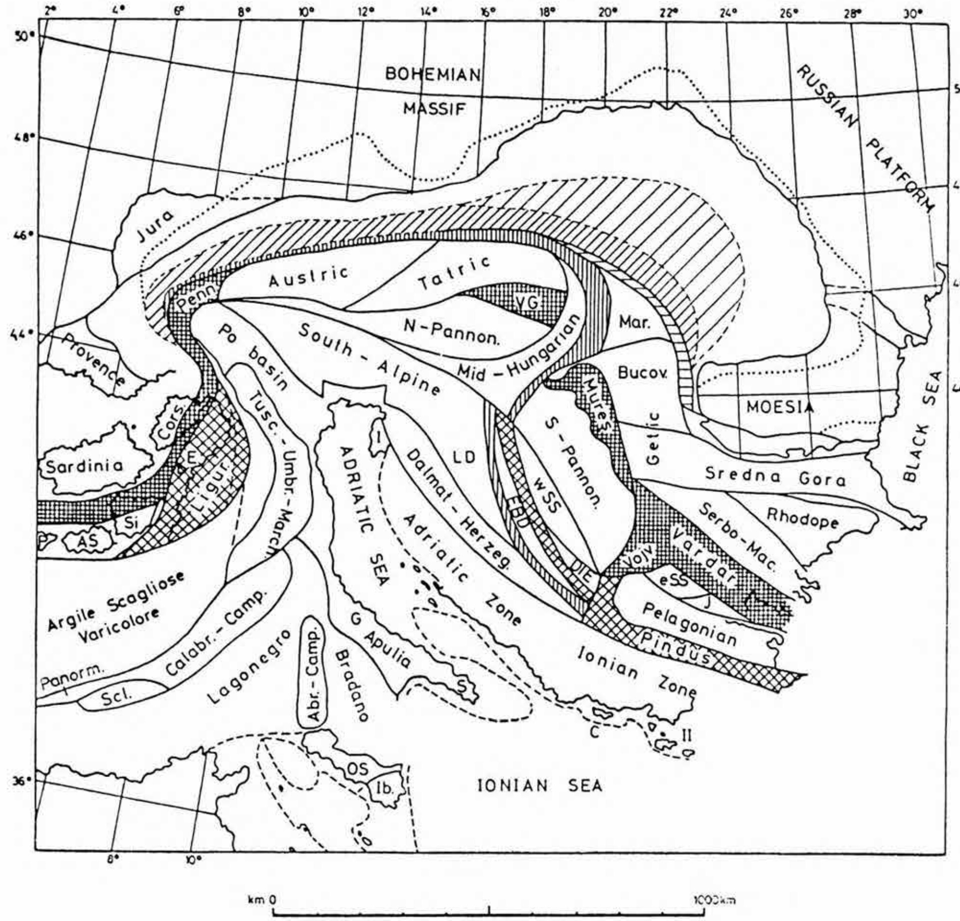


*Fig. 3.* Palaeotectonic scheme for 20 Ma. For legend, see Fig. 1
Letter symbols as in Fig. 2. — Mar. = Maramureş domain

*2* Around the northeastern corner of the Adriatic promontory a large bay is opened on the southern margin of Europe. Striking differences exist between the Alps and Carpathians in both the structure and the Oligocene—Miocene history and especially in the crustal structure. They all indicate that this bay was of thinned continental and, perhaps, oceanic crust.

*3* All domains around the Moesian plate are strongly shortened and widened in the reconstruction.

*4* The Dinaric—Hellenic zone is of slight curvature with uniform convexity in harmony with the direction of the subduction. Moreover, not only the Upper Cretaceous to Palaeocene but also the Eocene and the Oligocene—Miocene volcanic belts of the Balkan area are straight lying in east—west direction and wedging out towards the west. They clearly reflect a long-living convergent plate margin in the Hellenides and its passing into a transform boundaty within the Dinarids.

*5* The Sarajevo flysch belt lies in the continuation of the Rhenodanubian—Magura—Maramureș—Szolnok flysch belt marking together the northern and eastern boundaries of the Adriatic promontory.

*6* The situation with the sharp ending of the Betic—Calabrian range and with the location of its probable continuation in the Alps reminds a sinistral transcurrent boundary between the Alpine and Apenninic ophiolite-bearing belts. In a more general sense, the western transform boundary of the Adriatic promontory is visible here. Alignment of the Betic—Calabrian range with the Alpine one is expectable in Late Cretaceous reconstructions.

As seen, the Neogene history of the middle section of the European Alpides is the history of the break up of the Adriatic promontory of Africa and the history of interactions of its pieces with surrounding domains. Furthermore, the Neogene is the time when new phenomena appeared to make the Africa/Europe continent—continent collision in the Adriatic—Alpine region much more complicated than it was during the Late Cretaceous and Palaeogene. These new phenomena were (1) the appearance of an independent spreading centre in the Ligurian—Tyrrhenian area in the west and (2) the westward-directed movement of the Turkish microplate in the east induced by the Arabia/Europe continent—continent collision in the Caucasus region. These resulted in the movement of Adriatic—Apulian domains of the African plate and domains situated around Moesia on the European plate towards each other. Controversially directed movements were deviated towards the north and northeast, i.e. towards the Carpathian bay of Europe, and resulted in break up of the northeastern boundary of the Adriatic promontory and in closing of the Carpathian bay. That is why in this region we observe unusually large horizontal displacements in the Neogene, exceeding 300 km.

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# NEOGENE DYNAMICS OF THE PERI-TYRRHENIAN AREA IN AN ENSIALIC CONTEXT: PALAEOGEOGRAPHIC RECONSTRUCTION

by

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*Introduction.* A project on the palaeogeography of the peri-Mediterranean area has been developed in the frame of the Paleogeography and Geodynamics Working Group of RCMNS. Contributors to this project have been an Italian Group formed of researchers of the major Universities, of CNR and of the Italian Oil Company (AGIP).

The first results of the work done by the Group were presented at the Interim-Colloquium "Palaeogeography and Geodynamics of the peri-Tyrrhenian area" held in Florence in October 1984.

This paper reports on the synthesis of the research carried out so far, on the palaeogeography and geodynamics of the peri-Tyrrhenian segment, from the southern Alps up to Sicily.

This peri-Tyrrhenian segment is the most significant portion of the whole western Mediterranean, as it represents a connection between the Alpine—Dinaric and Maghrebid systems. Within this segment, the complex south Alpine—Apennines—Sicily orogenic system developed as a double orocline having the well-known S configuration.

The building of this system occurred in two major stages: a first, eo-Alpine, stage with a European polarity and a second, Apenninic Neogene stage, with an African polarity.

The latter discussed in this paper, is responsible for the formation of the S-like system, and it developed in a post-collisional ensialic context, starting from the Oligocene. This stage is characterized by the opening of the Ligurian basin up to the Burdigalian and, later on, by the opening of the Tyrrhenian basin (Boccaletti et al., 1982). All this process brought about important anti-clockwise rotations, with consequent outward migration of the compressional phases which are followed and overlapped by the tensional phases (progradation of the Tyrrhenian).

In this framework, the palaeogeographic reconstruction cannot be based solely on the present distribution of the sedimentary facies for the considered time intervals. It necessarily involves a palinspastic reconstruction which takes into account the contemporaneous processes of compression, tension and rotation. This entails the evaluation of various amounts of shortening and stretching (extension), both longitudinal and transversal to the chain axis. All this requires, besides a noticeable effort of synthesizing the knowledge so far achieved in various fields, to resort to a good measure of imagination.

### The back area—thrust belt—foredeep system

Fig. 1 shows a schematic cross-section of the chain, from the external to the internal areas, where the following subdivisions were made:

— the foreland, including an emerged zone, formed by platform carbonates, and a submerged zone, where platform and basin deposits are recognized, with terrigenous—calcareous sedimentation, mainly pelagic and subordinately neritic. The foreland zones are connected with the foredeep through a slope, generally controlled by normal faults, and a ramp zone, which is difficult to define, and are dominated by hemipelagic sedimentation, with subordinate gravitational sediment flows;

— the foredeep develops between the base of the slope and the front-thrust of the chain. It is an asymmetric depression, with a strong subsidence and a complex physiography related to the compressional activity of the thrust zone progressing towards the exterior: the trough is filled essentially by turbiditic sediments. The initial sediment contributions from both sides of the trough. The facies are very diversified and refer to base-of-slope deposits, submarine fans and basin-plain deposits;

— the chain starts to be built up, in the Apennine area, in Upper Oligocene. The chain front allows a discontinuous migration from the interior towards the exterior, until its present position in the Adriatic. In the tectogenetic events important stages of crisis (acmes?) are recognized. Some of these appear ascribable to mechanical response to, for instance, different crustal thickness of the foreland rather than to general variations of the thrust induced by the convergence of the two main, African and European blocks.

Discontinuous basins (piggybacks) develop within the thrust belt, which is their main sediment source. The sedimentation in these basins is clastic, generally coarse and even discontinuous.

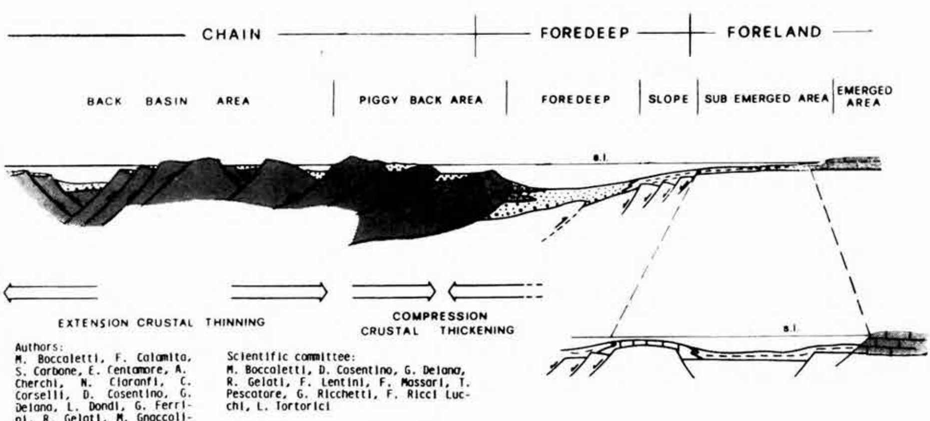


*Fig. 1.* Schematic cross-section of the chain from the internal to the external areas, with the attribution of the various basins to different structural zones

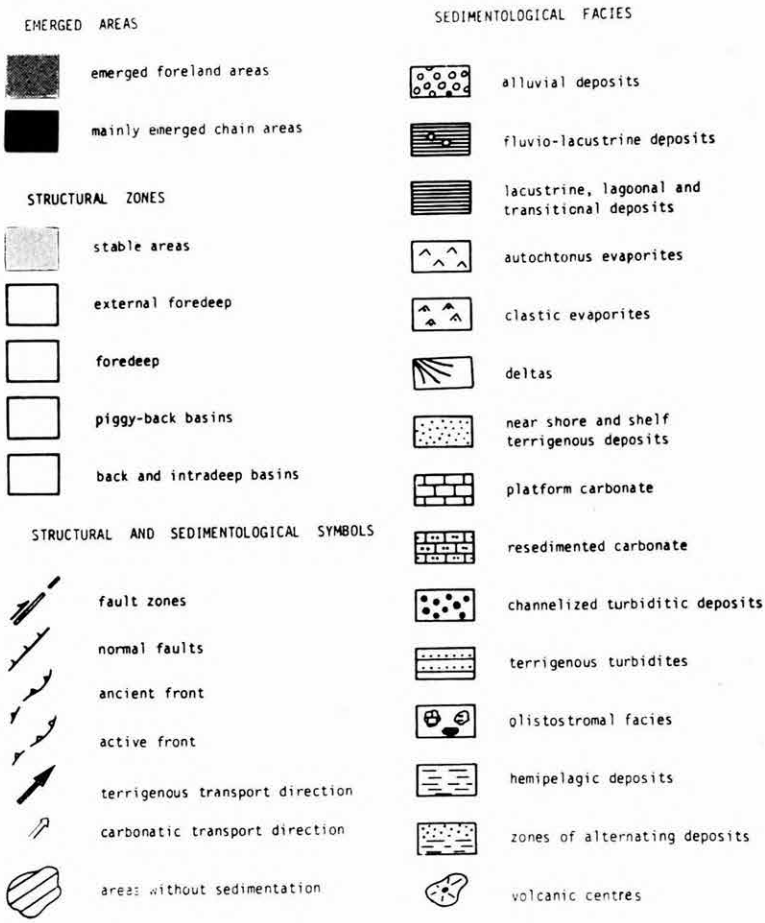


*Fig. 2.* See caption of Figs. 3—9

In the inner part of the chain, a tensional regime replaces the previous compressional regime producing the fragmentation of the thrust belt up to its destruction. This is caused by a process of stretching and crustal thinning. In this area, intrachain basins (back basin and intradeep) develop, grading from continental to marine.

Within the sedimentation areas the main facies are shown, as subdivided on the palaeogeographic maps (Fig. 2).

### Palinspastic and palaeogeographic reconstructions

The maps, prepared according to the previously mentioned criteria, represent the present-day distribution of the sedimentary facies referred to the following time intervals: Upper Oligocene (25 m.y.), Burdigalian (20–22 m.y.), Langhian (16—17 m.y.), Serravallian (13–15 m.y.), Tortonian (9–10 m.y.), Late Messinian (5.3–5.5 m.y.), Early Pliocene (4 m.y.).

These maps represent only a starting point for the palaeogeographic reconstructions. For this purpose, the amount of shortening of the external thrusts was evaluated through balanced cross-sections (see also BOCCALETTI, these volume). The values so obtained were employed as minimum orders of magnitude to assess the shortening in the internal zones. Furthermore, for each considered interval, an overall balance was attempted, between the total shortening and the total extension of the internal zones, assuming crustal values, before and after the extension, based on the present data.

The total displacement of the fronts, between Upper Oligocene and Lower Pliocene is approximately reckonable as 250 km in central northern Apennines, and as much as 600 km in the Calabrian arc.

*Upper Oligocene*

The Upper Oligocene marks the beginning of the Apenninic—Maghrebid tectogenesis, i.e. the birth of the chain sensu strictu (Fig. 3). During this period, Corsica and Sardinia are still in crustal continuity with the European continent. The

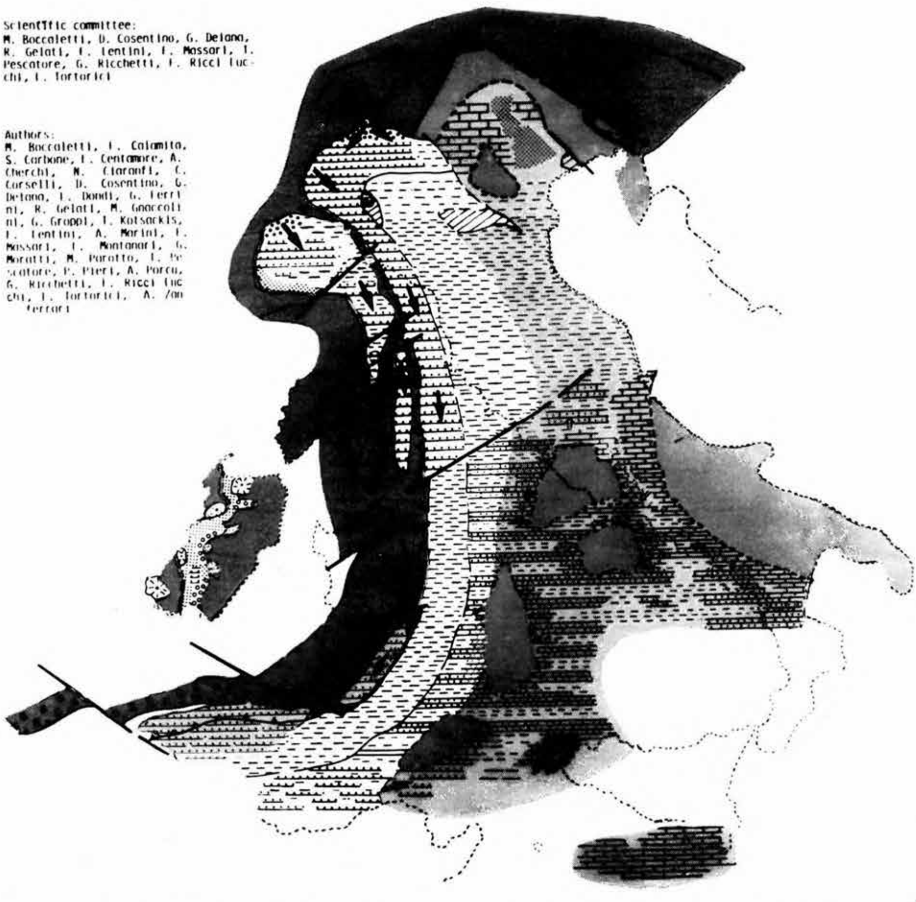


*Fig. 3.* Palaeogeographic reconstruction during Upper Oligocene (25 m.y.)

oceanization phase, which will set in at 23/24 Ma, has not yet begun, and the stage is still that of preoceanization rifting.

Corsica and Sardinia assume the role of hinterland with respect to the Apenninic system. Previously, in fact (Cretaceous—Eocene phases), they were part of the foredeep of the eo-Alpine chain.

The corrugated chain is included between the eo-Alpine fronts, already inactive, to the west, and the active Apenninic fronts, to the east.

On the active thrust belt some piggyback basins start developing, namely, from north to south: the Ligurian—Piedmont basin, the Ranzano basin, the basin of the Coastal Macigno, all with longitudinal sediment contributions. Towards the south, the Cilento basin and part of the Numidian flysch basin are forming. For what concerns the foredeep, the Macigno turbidites develop in the northern segment, with longitudinal sediment flow from the central western Alps. These turbidites stop abruptly in correspondence to the Grosseto—Chienti line (GCL). To the south, the deposits are pelitic and become arenaceous in the southernmost part (Numidian flysch).

The difference between the northern and the southern segments exists also on the slope. In fact, in the north, pelites prevail over calcarenites, whilst, in the south, turbiditic calcarenites are more frequently encountered.

More externally (foreland), north of the Grosseto—Chienti line, hemipelagic sedimentation continues. South of the mentioned line, basins with pelagic sedimentation develop. They are interposed amid neritic carbonate platforms, partially emerged at that time. The platforms feed clastic carbonate material to the basins and to the adjacent slope (white arrows).

Carbonate platforms develop also in the south Alpine segment. In the western south Alpine the occurrence of southward translations controls the entry-points of turbidites, while in the eastern part such movements are lacking during the time-span in question.

*Burdigalian (22—20 m.y.)* (Fig. 4)

In this interval, the translation of Corsica and Sardinia has already occurred, but the whole 30° rotation of Sardinia is not yet complete. Therefore, there is an increase, in the emerged areas, of the zones under stretching, with the formation of back-arc basins: Sardinia, Corsica and Finale Ligure basins.

An eastward migration of the external fronts follows, with also an increase of their curvature.

In the foredeep, siliciclastic, turbiditic sedimentation prevails. It is represented in the north by the Cervarola flysch, which is interrupted towards the south at the Grosseto—Chienti line. In the south the foredeep is filled by the quartz-arenitic Numidian flysch of African origin, which abruptly ends, northwards, at the Gaeta—Gargano line.

The Grosseto—Chienti line still separates the northern and southern segments; a mainly pelagic sedimentation continues in the north, both in the slope and in the foreland. To the south, where the Panormide and Campania platforms are already involved in the tectogenesis, the slope and the foreland areas are still influenced by the presence of neritic carbonate platforms, which continue feeding the adjacent basins with clastic material.

In the south Alpine area, the thrusts are still active in the western segment, but show a tendency to migrate towards the east, as indicated by the eastward migration of the entry-points of the turbidites, which feed the Cervarola formation in the foredeep.

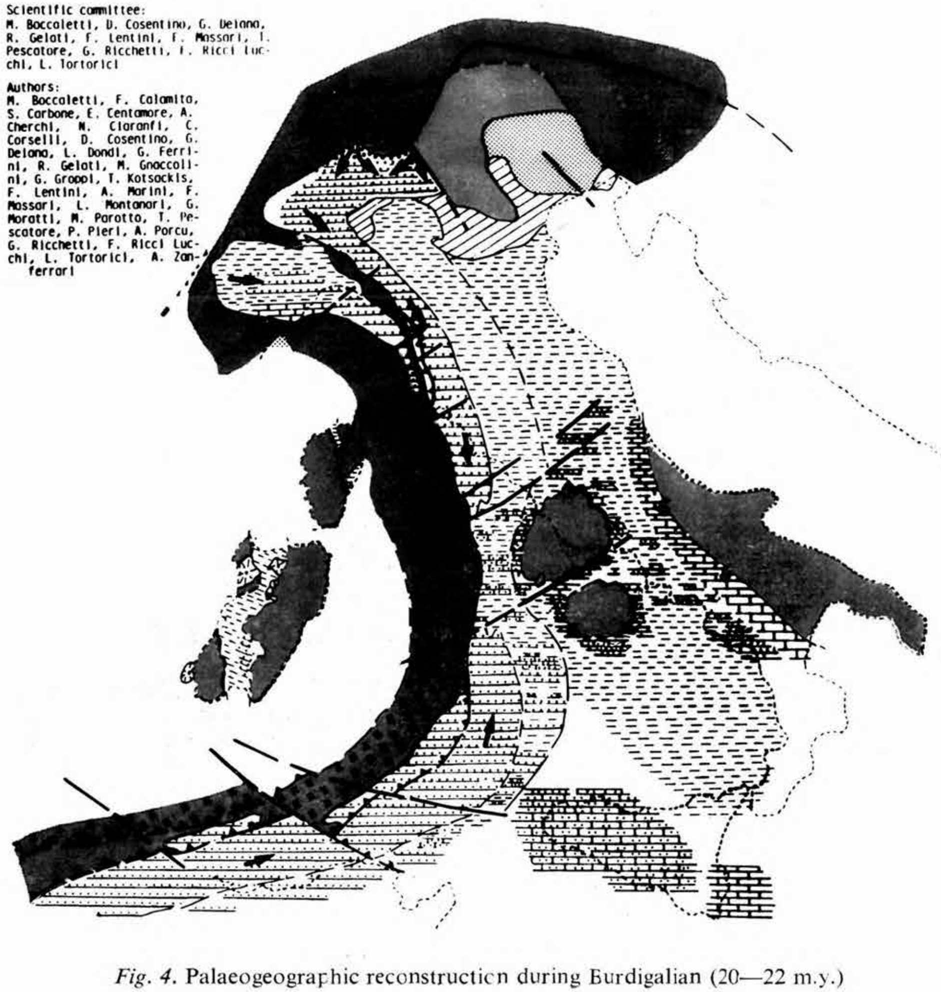


*Fig. 4.* Palaeogeographic reconstruction during Burdigalian (20—22 m.y.)

The eastern segment is quieter and is dominated by a broad platform area partly emerged and partly the site of glauconitic sandstone deposition.

*Langhian (16—17 m.y.)* (Fig. 5)

During the Langhian, the translation and rotation of Corsica and Sardinia are completed. This time represents an important palaeogeographic stage. It is characterized, in fact, by an extensive "transgressive" phase, revealed by a general tendency of the carbonate platforms to be drowned. An increase in frequency of carbonatic and hybrid mega-turbidites in the foredeep is noticed. This would suggest an underwater tectono-seismic (?) dismantlement of the platforms.

The Grosseto—Chienti transversal bundle still separates two major northern and southern segments. These, in turn, are further subdivided by other transversal lines. To the north, in the Apennines, the NE—SW transversal lines prevail, of which

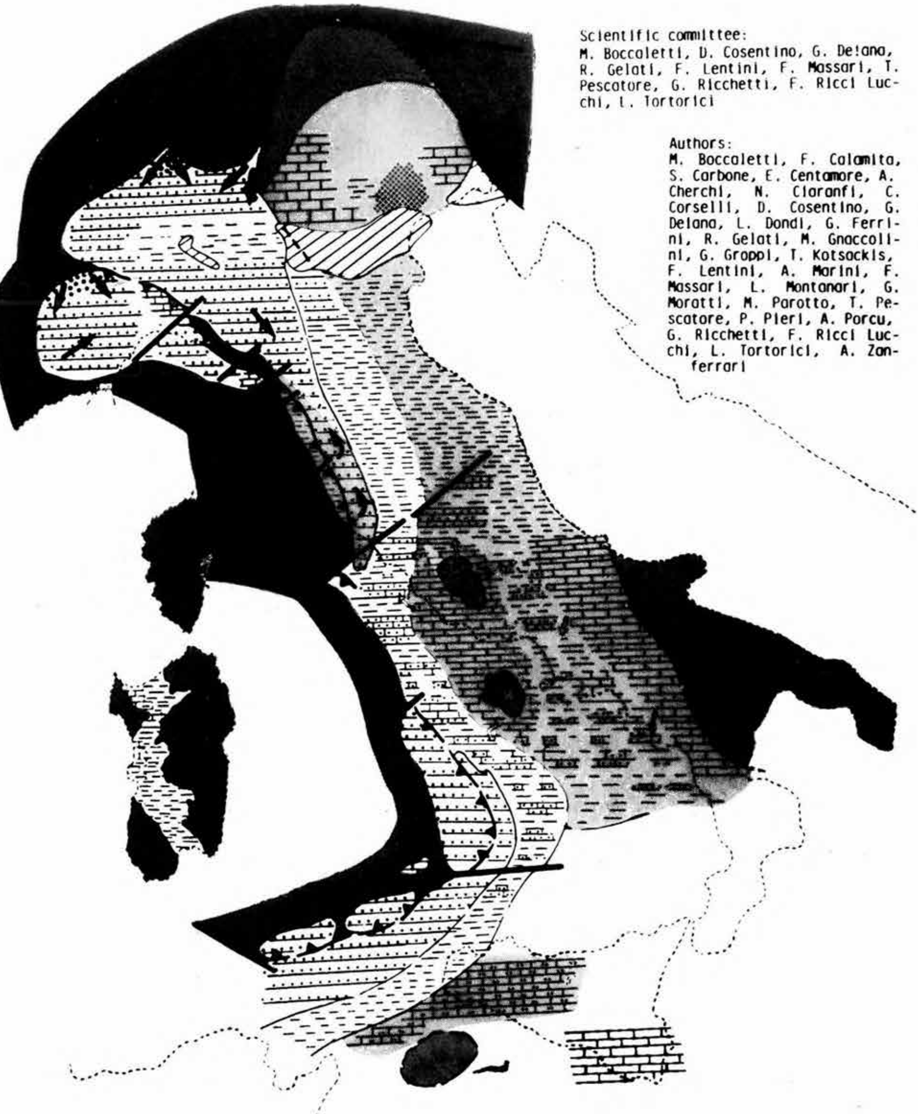


*Fig. 5.* Palaeogeographic reconstruction during Langhian (16—17 m.y.)

the Ligurian bundle, the Livorno—Sillaro and the Piombino—Faenza lines are the most conspicuous.

In the south Alpine unit, the NW—SE transversal lineaments are more frequent. Of these, the bundle approximately located on the Verona—Legnago alignment is the most important. This lineament acts as eastern limit of the thrusts and of the foredeep zones. To the east, in fact, the area is still quiet, with carbonate deposition, although the submerged area appears increased with respect to the Burdigalian interval.

To the south, the dominant transversal lineaments are oriented E—W (South Tyrrhenian system) or NW—SE (Ligurian—Balearic basin system).

Noticeable changes occur concerning the piggyback basins. The facies become more pelitic (Marne di Cessole, etc) and new basins are forming: Vicchio marls, Gorgoglione flysch, Capo d'Orlando flysch.

In the foredeep, turbiditic sedimentation continues: Marnoso—Arenacea in the north, Numidian flysch and Serra Palazzo flysch in the south. The clastic contributions related to the transversal accidents become important in the Marnoso-Arenacea. Still in the Marnoso-Arenacea, minimum deposition rates are observed, which confirm the transgressive tendency noticed at large. An important difference between the northern and the southern segments, is the further accentuation of the arching of the southern sector of the corrugated belt. This fact may be related to the 30° rotation of Sardinia, relative to Corsica, which ended 19 Ma BP, and possibly to the beginning of rifting in the Tyrrhenian.

*Serravallian (13—15 m.y.)* (Fig 6)

In this interval, a further transgressive tendency occurs. This is shown by the complete drowning of the southern platforms and by the deposition of some flysch, both in the foredeep and in piggy-back basins. In the north, in fact, both the Marnoso—Arenacea and the Bismantova show minimum rates of sedimentation, which are, however, compensated by regional tectonic events. In fact, in this period, the generalized transgression is accompanied by an important and strong deformational phase.

The Grosseto—Chienti line still separates two areas, to the north and to the south, with different evolution. The northern segment shows noticeable variations of shape and facies concerning the piggy-back basins. A higher segmentation and consequent isolation of the basins occur, in correspondence with the transversal lines. Along some of the latter, conditions of a "high" occur with platform deposits, which are brought into direct contact with the basinal turbidites. The influence of the transversal lines results also in important transversal sediment contributions to the foredeep.

Another major difference, with respect to the Langhian, is found in the south Alpine, where important thrusts affect the eastern segment, so far stable. This fact is accompanied by the fragmentation of the existing platform and the onset of piggy-back and foredeep areas with terrigenous sedimentation. The fragmentation is controlled by important transversal lines belonging to the Schio-Vicenza and Giudicarie systems. These lines show also transcurrent movements.

In the southern segment the most important event is the complete drowning of the foreland carbonate platforms, as previously mentioned. In the foredeep, the quartz-arenitic sedimentation of African origin of the Numidian flysch is definitely replaced by pelites (shales) and arcosic sandstones.

The arching of the southern thrust belt is still more marked and it already foreshadows the destination of the wouldbe Calabrian arc.

The accentuation of the arching is guided by an important E—W transversal accident (south Tyrrhenian bundle) and is, in our opinion, related to the incipient opening of the south Tyrrhenian (basin).

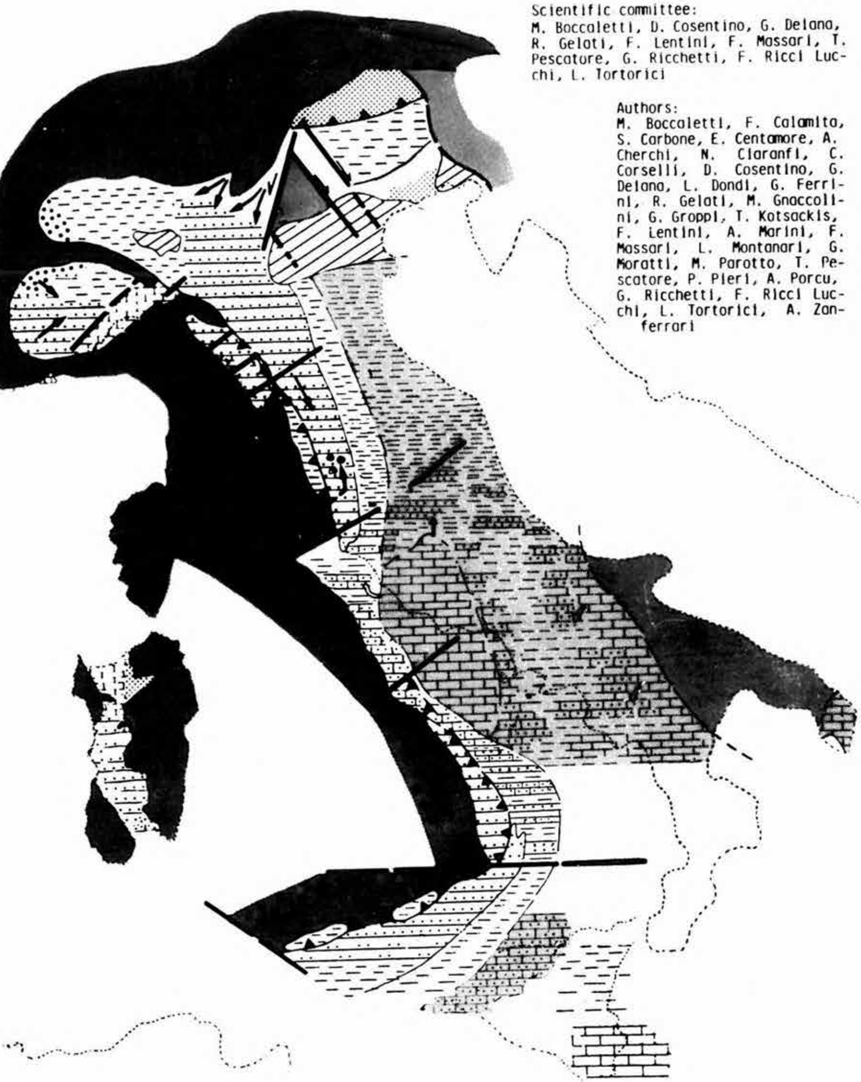


*Fig. 6.* Palaeogeographic reconstruction during Serravallian (13—15 m.y.)

*Tortonian (9–10 m.y.)* (Fig. 7)

This period is characterized by noticeable palaeogeographic changes related to a generalized major tectonic phase.

Large portions of the foreland, so far stable and characterized by platform deposits and intra-platform basins, are reached by the slope and foredeep zones, where sedimentation is mainly terrigenous, fine or coarse.

Even the Latium—Abruzzi carbonate platform is destroyed and replaced by a complex foredeep with turbiditic deposits confined to narrow longitudinal basins (Livi and Sacco basins).

The central southern segment shows the most striking physiographic changes with respect to the previous time-span. The first back-arc basins appear in the Calabrian arc zone, already reached by the extension following the futher opening of the Tyrrhenian (already realised more to the west, Cornaglia basin). Still in the Calabrian arc, portions of the eo-Alpine chain have proceeded eastward in respect to the slope and ramp-foreland areas.

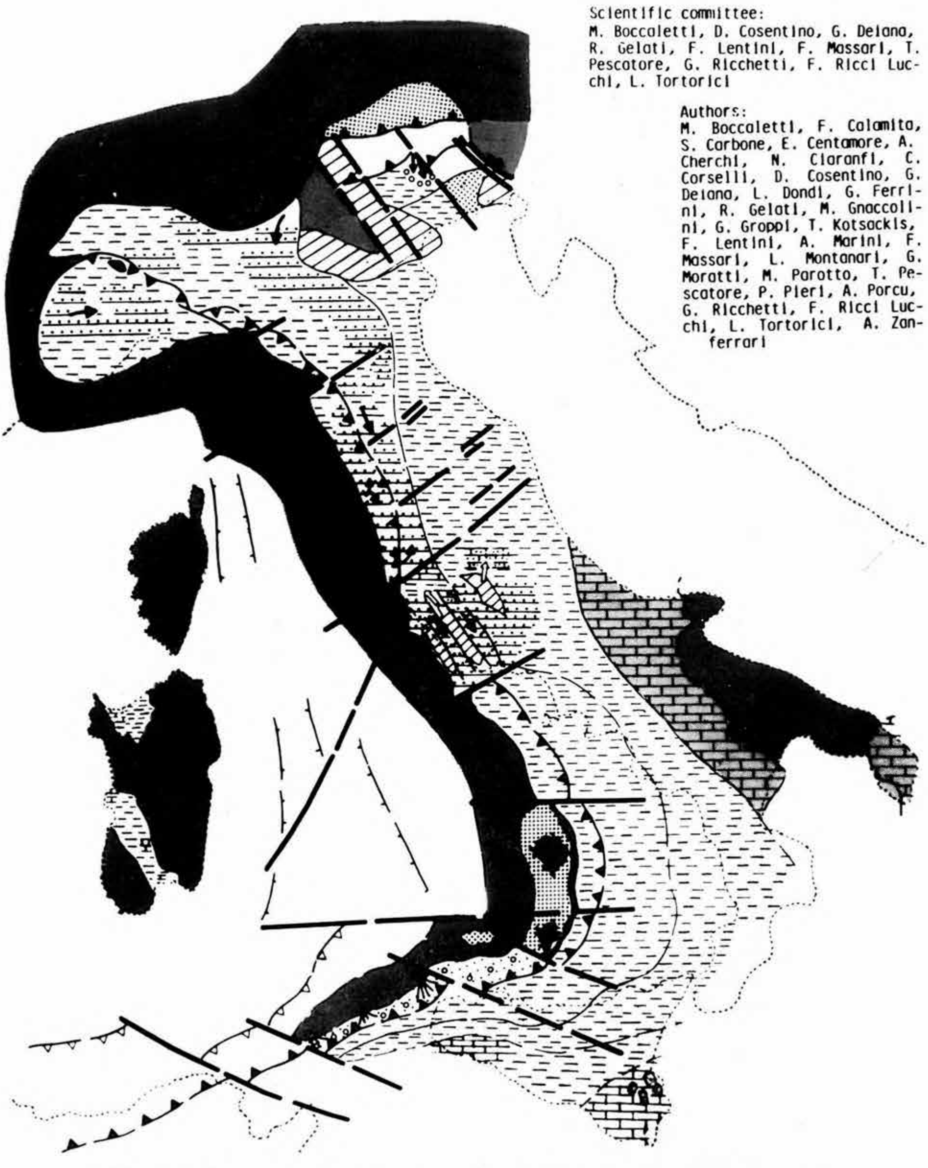


*Fig. 7.* Palaeogeographic reconstruction during Tortonian (9—10 m.y.)

In the south Alpine segment, a southward migration of the thrust fronts is noticed. These fronts, however dissected, appear by the Schio—Vicenza trend, which affects the area unto its easternmost part near the Dinarides. Especially in correspondence with the lineaments of this trend, the entry-points of terrigenous, generally coarse, deposits, are located.

*Late Messinian (5.3—5.5 m.y.)* (Fig. 8)

This interval follows immediately the salinity crisis. Two segments, separated this time by the transversal Gaeta—Gargano line, become well individuated.

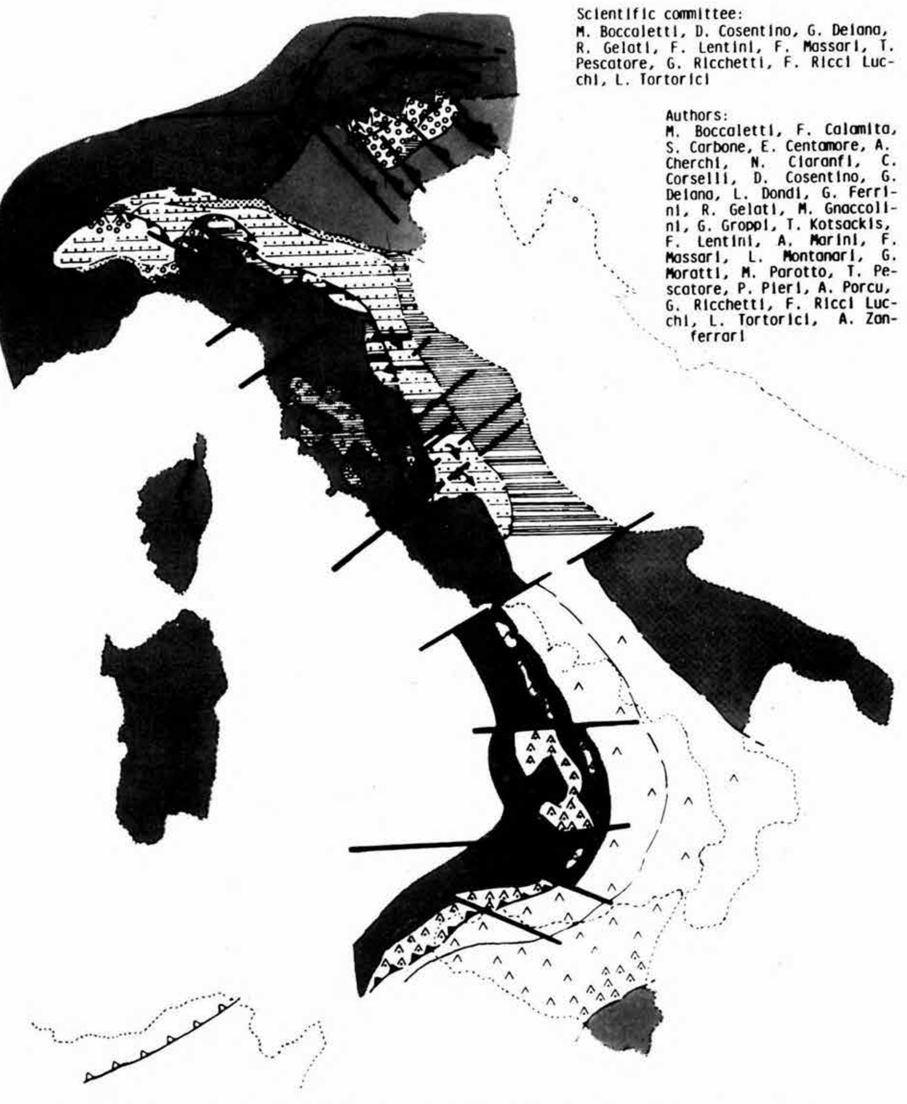


*Fig. 8.* Palaeogeographic reconstruction during Late Messinian (5.3—5.5 m.y.)

North of this line the evaporitic cycle is present only in the back-arc areas, to the north of the Piombino—Faenza line. In the foredeep and foreland it is replaced by prevalently pelitic (Colombacci Shale) or turbiditic (Laga and Fusignano Formations) successions.

South of the Gaeta—Gargano line, another evaporitic cycle can be recognized (Upper Evaporites). It lies above the previous evaporitic cycle in almost all the structural zones. The evaporitic deposits are primary or resedimented. In this area the distinction of the structural zones on the basis of the sedimentary facies remains therefore difficult.

The Gaeta—Gargano line, thus, results as an important barrier that separates, in the internal areas, two different worlds of problematic interpretation. Probably, among the principal causes of this are: the presence, to the north, of the Alpine chain, supplier of great quantities of water; and a greater supply, especially transversal, tied to an intra-Messinian tectonic activity, as it has been recently shown.

In the Po plain and Adriatic foredeep the molassic stage begins, sometimes unconformably on the underlying evaporites. The unconformity is widespread in the piggy-back areas, which are now reduced and mainly have fan-conglomerate sedimentation. In the back-basin basins of internal Tuscany, continental deposits overly the previous evaporitic and marine sediments.

In the south Alpine unit a foredeep, with pronounced subsidence, filled with fan-like coarse sediments, develops, in correspondence with the external thrust fronts, which are confined to the eastern part.

*Early Pliocene (4 m.y.)* (Fig. 9)

In the early Pliocene, a new, widespread and important tectonic event initiates. This produces extended palaeogeographic variations, as a consequence of the higher continuity of the structural zones, longitudinally, from north to south, even though a transversal segmentation of these zones is still evident, especially in the internal areas.

The chain approaches its definite shape, with the external thrusts fronts near their present position. The foredeep develops in continuity from the Po Plain southwards, up to the Adriatic area, the present-day Bradanic trough and the Hyblean foreland. In the Calabrian arc, the internal areas, together with fragments of the eo-Alpine chain, have proceeded towards E and SE along transversal tracks oriented NW—SE and E—W.

The Pliocene tectogenetic event is characterized, with respect to the Miocene events, by an evident and generalized jump of the frontal thrusts. Concurrently, the most important marine ingression occurs in the internal areas, with extended back-arc basin development. Although these back-arc basins seem longitudinally continuous, they are instead segmented by transversal lines, thus resulting in an articulated and complex physiography. This results from a strengthening of the tensional processes, tied to the opening of the Tyrrhenian sea, prograding eastward.

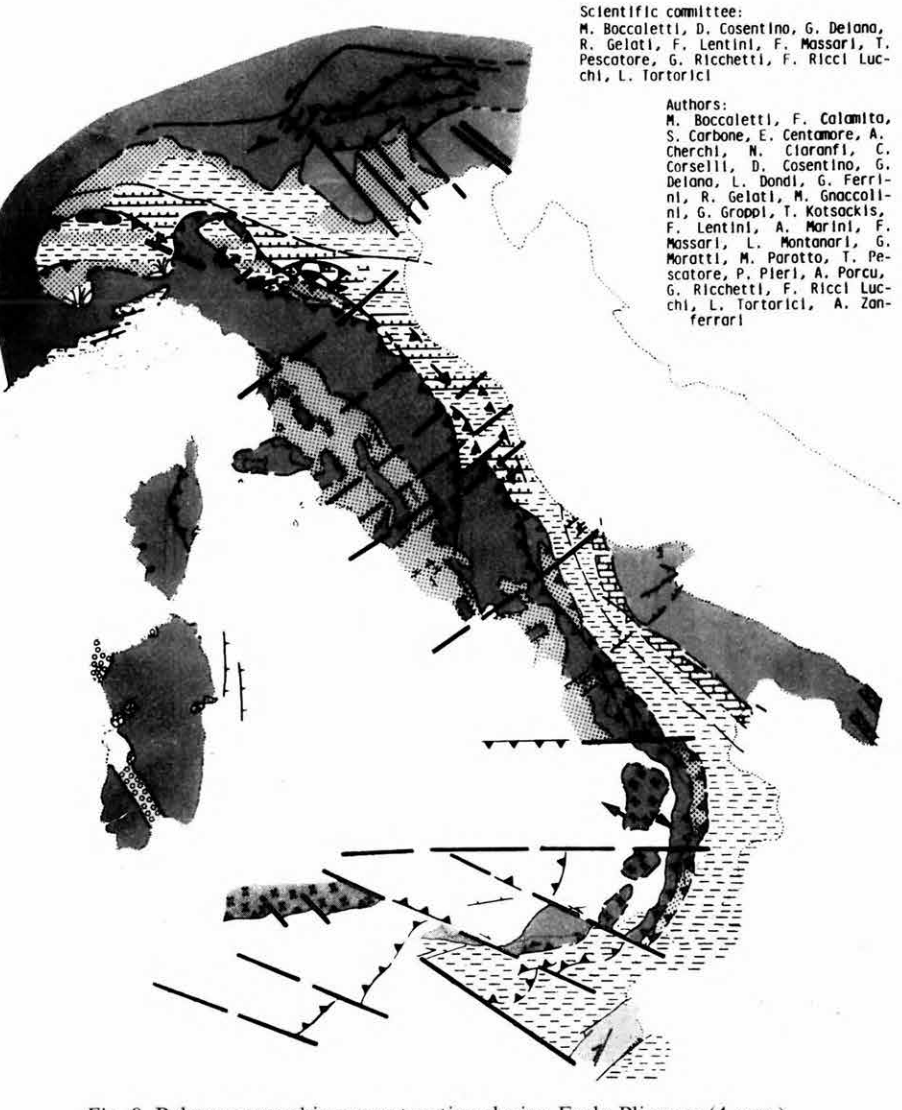


*Fig. 9.* Palaeogeographic reconstruction during Early Pliocene (4 m.y.)

**Concluding remarks**

The comparative analysis of the various time intervals considered allows to draw some general conclusions.

Four moments of primary importance can be clearly singled out, during which considerable geodynamic and palaeogeographic modifications occur in the whole peri-Tyrrhenian area:

*1* The late Oligocene is the first important stage, in which the actual tectogenesys of the Apennines and Maghrebids initiates. It occurs simultaneously with the beginning of rifting in the Ligurian—Provençal zones.

*2* The second stage delast from the late Burdigalian till the early Langhian. There is a significant migration of the external fronts, especially with the rotation of the Sardinia block. Possibly these events are already related to the beginning of rifting in the Tyrrhenian sea.

*3* The third stage occurs in the Tortonian, with the well-known tectonic phase developed throughout the Apennines. It results in a better identification of the southern arc, definitely more advanced than the northern arc. This fact is to be related to a greater extensional process affecting the southern Tyrrhenian, with respect to the northern part.

*4* Finally, the fourth stage develops between the end of the Lower Pliocene and the base of the Middle Pliocene. There is a generalized jump of all the fronts towards the exterior. Particularly, there is a better identification of the various segments of the Calabrian arc. The northern segment of the arc has proceeded towards the east, guided along a transversal bundle oriented E—W. This bundle bounds to the south the extension of the Tyrrhenian.

Two other events, generalized throughout the area, of palaeogeographic rather than tectonic importance, are referable to the Langhian and Serravallian. The tectonic events are, in these periods, associated with a widespread transgression, witnessed by an increase in marly—pelitic sedimentation, by the complete drowning of the carbonate platforms and by a minimum in the rate of sedimentation of turbidites in the foredeep.

In any case, the geodynamic evolution of the whole area seems to be clearly controlled by the activity of major bundles of tectonic lines, transversal to the chain.

The most important of these bundles have a different orientation according to the various segments of the chain: NW—SE (south Alpine), NE—SW (northern central Apennines), NW—SE and E—W (southern Apennines and Sicily). They determine a different dynamic behaviour from segment to segment, as well as they influence the facies distribution, representing, in some moments, preferential channelways for the turbiditic flows. The most important of these lines are:

— the Ligurian line that releases the fulcrum zone for the anticlockwise rotation of the northern Apennines;

— the Grosseto—Chienti line. It allows an independent evolution of the northern Apennine with respect to the southern Apennines, often impeding the basins;

— the Gaeta—Gargano, which allows an independent evolution of the Calabrian arc with respect to the central Apennines;

— lastly, the south Tyrrhenian line, of E—W orientation. It forms the southern limit of the Tyrrhenian opening, and also allows Calabria to migrate towards the east with respect to the contiguous Maghrebid area.

The role that each line played in time and space is nevertheless complex. During the same phase, in fact, they act in the outernmost part of the thrust belt as lines separating fronts with different entity of advancement, to which different values of shortening correspond. In other parts, the same lines separate totally different facies, showing vertical movement components. In the most internal areas, they even separate segments having a different amount of extension. In any case these tectonic lines certainly have a crustal, or even deeper, significance; they might have acted along pre-Neogene, or even pre-Alpine discontinuities.

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# COMMENTS ON THE EARLY HISTORY OF PARATETHYS

by

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In the last few years, the opinion has prevailed (above all BÁLDI, 1979; 1984) that, from the viewpoint of the evolution of the independent sedimentation realm of the Paratethys, the interval exhibiting the first signs of endemicity and isolation in the older section of the regional Kiscellian stage rather than at its base is of prime importance. This time span is characterized by the appearance of horizons with *Spiratellae* and by the deposition of beds with *Cardium lipoldi* and *Ergenica cimlanica.* The appearance of Spiratellae etc is a prominent anoxic event in the evolution of the older Oligocene sedimentation area of the sedimentation realm. For instance, BÁLDI (1979) supposes that the full separation of the Tethys and Paratethys took place at that time.

However, detailed studies on the history of sedimentation areas in the Pre-Caucasian region, the southern part of the USSR, the West Carpathians, and the Alpine foredeep have shown that the appearance of anoxic regime probably did not take place during a limited time interval. For instance, in the Pouzdřany sedimentation area this regime appears neither in zones NP21 and 22 respectively nor does it show up in the regions of the pre-Alpine molasses or the southern territories of the USSR. The first evolutional stage of Paratethys should also examined from the viewpoints of sedimentology and tectonics.

The broad and longitudinally rather dissected area of flysch sedimentation at the external side of the Alpie—Carpathian orogenic belt attained its maximum in Early Cretaceous time. From the area bearing the character of marginal seas (with oceanic depths and a crust more or less of oceanic type), it extended as far as the platform margin, gradually passing there to non-flysch epiplatform facial types.

As a result of subduction movements, the process of intensive reduction started at least since the Laramian movements, that means by the and of the Createaceous but probably as early as the movements during the Late Cretaceous. The area was gradually folded while nappe subunits were being formed and thrusted over one another toward the foreland during the reduction of space. Sedimentation went on more or less continuously in the area left in front of the unit immediately formed, resulting in sequences that, mostly, were stratigraphically more complete, provided that the lower members were not separated tectonically.

This "inheritance" of a part of the sedimentation area of the older (and broader) area of sedimentation is one of the universal evolutional features of the whole region that can be traced to the incipient molasse sedimentation. The culminating spatial reduction due to the folding of the internal units and gradual shallowing results in the onset of molasse sedimantation that—upon completed subduction of the remnants of the basement of the former flysch basins and incorporation of the oldest molasse sediments into the last nappes—advances to the platform margin into the new-forming molasse basins.

New molasse basins were formed where the mobility of the basement provided adequate conditions, not only in the platform foreland but also within the new-forming mountains and at their inner sides (intramontane and backdeep basins) as a result of tensile stresses beyond the zone of fading-out compression. In the depressions at the back part in some places sedimentation lithologically identical with the molasse one went on since the Upper Cretaceous, as at the back part existed conditions for forming of the basins with "molasse like" sedimentation. Dilatation took place here since the outset of shortening of the external parts of the Tethys area in which the flysch was deposited.

Vass (1980) calls the sediments originating in this period "early molasse", because they were formed subsequently to the main folding. From a more or less formal point of view, the sedimentation areas on the backpart of the forming mountains, in which molasse-type sedimentation occurred synchronously with flysch sedimentation in the Tethyan realm proper, should better be called premolasse areas, because they were situated on an other plate o plate fragment the displacement of which caused the compression of the northern (oceanic) branch of the Tethys. These basins were not formed by the disintegration of the Tethys, but were incorporated into the Paratethyan molasse region later, after the disintegration.

In the external Alpine—Carpathian zone, the stage of Tethys disintegration and Paratethys formation is that of the transition from flysch to molasse sedimentation. This process, occurring in close interrelation with the tectonic movements, can be traced, step by step, from the very similar types of lithological development. Interregional comparison makes possible to distinguish transverse facies zonality as well and to establish common features therein.

### Flysch-to-molasse transition in the Alpine—Carpathian region

Generally, it can be said that between the typical flysch sequences that developed over a long period of time, locally as early as the Lower Cretaceous up to the Paleogene, and the typical molasse, series comprising a bituminous clay formation have been deposited. It contains menilitic cherts and is called the Menilitic Formation that, in its typical development, exhibits neither flysch nor molasse character. It should be noted, however, that the upper section of the Menilitic Formation is frequently substituted by the Krosno lithofacies that, on the contrary, does show flysch character in its typical development.

However, the sedimentology of the Menilitic Formation has not yet been fully characterized. Its faunal or floral characteristic is full of contradictions. The majority of authors consider it a formation deposited on the lower part or at the base of the continental slope (Hanzlíková, 1981; Krhovský, 1981; Stráník, 1981). Even though the formation is characterized, by the occurrence of bituminous shales in the broadest sense of the word, it exhibits by far no uniform development throughout the external Carpathians. Locally, transverse facies transitions with varying amounts of clastics can be found, forming thicker or thinner horizons in the menilitic facies.

Mostly, transverse zonality is apparent in the facies development. For instance, in the East Carpathians of Roumania (according to Dumitrescu and Sandulescu, 1974; Micu in Sandulescu et al., 1980), the innermost part of the Tarcău unit still exhibits prevailing Fusaru flysch development during Oligocene time comprising a Slonc facies with olistoliths in its uppermost part, whereas the outer part of this unit already displays the typical development of the Menilitic Formation.

There, the latter comprises two horizons of Kliwa sandstones separated by the flysch sequence of the Podu Morii beds or by the Vinetisu sequence of strata that (DUMITRESCU and SANDULESCU, 1974) represent the latest occurrences of flysch facies. In the Tarcău unit, the sedimentation of bituminous mudstones with menilites seems to be the most high-reaching one throughout the Carpathians—as high as zone NN 2—3—apparently substituting there also the lowermost sections of the Lower Miocene Salt Formation. The equivalents of the Salt Formation—the lower Gypsum Horizon or the Lopătari Formation—overlie the Supramenilitic Horizon with a distinct disconformity (MICU in SANDULESCU et al., 1980).

The Supramenilitic Horizon is still developed as the highest (nonmolassic) member in the southern part of the more external unit of the Marginal folds. In the northern part and also further in the most external Subcarpathian unit we can see the beginning of the first appeareance of molassic developments at the end of Oligocene as the Gura Soimului beds, resp. as the Goru Misina beds (MICU, 1982) They represent a transition between the Menilitic Formation and the Lower Miocene Salt Formation and they are not lithologically uniform. They are generally characterized by longitudinal and transversal interfingering of the bituminous facies with the molassic sandyconglomeratic facies (MICU in SANDULESCU, 1980), but there differences exist from west to east. The calcareous nannoplancton of the Gura Soimului beds contains a Lower Miocene assemblage with species continuing their evolution from the Oligocene.

In the Ukraina, in Poland, and partially also in Czechoslovakia, the upper section of the Menilitic Formation is commonly substituted by the Krosno lithofacies that invariably shows flysch character. In a number of units such as the Premagura, Dukla, Čorna-Gora, Silesian, Subsilesian, and SkoleUnits, sedimentation terminates with this lithofacies, while it continues with molasse facies in other—the most external-units such as the Boryslaw, Pokuty, Stebnik, Sambor—Rozniatow or Ždánice and Pouzdřany units in Moravia.

The Krosno lithofacies seems to extend to a stratigraphically rather high position, at least to the Egerian or even Eggenburgian, as can be concluded from studies by WIESER (1979) and OLSZEWSKA (1982).

The Krosno lithofacies is extremely thick (up 4000 m) in the Silesian unit, particularly in its southeastern part in Poland, evidencing, in this area, an extraordinary subsidence during the late flysch sedimentation in the Carpathians.In accordance with its microfaunistic character, OLSZEWSKA (1980) correlates the upper section of the Krosno lithofacies to the Polanica beds that play an important role at the flysch/molasse transition in the so-called internal zone of the foredeep in the Boryslaw—Pokuty unit (Stebnik unit) and extend stratigraphically to the upper sections of the Egerian or the lower part of the Eggenburgian (OLSZEWSKA, 1980; JINORIDZE, 1979; DANYŠ et al., 1974; NAY et al., 1974). The Polanica beds are practically flyschoid in character or substituted by the conglomeratic Sloboda facies, or by the Vorotyšča Salt Formation in their upper section.

Thus, the facies zonality established in the East Carpathians of Roumania also appears in the Ukrainian and Polish regions, similarly as do the transitions from flysch to flyschoid or molasse peri-Carpathian saliferous development in comparable tectonic units in the outermost zone.

The Pouzdřany sedimentation area represents a certain evolutional transition between the pre-Alpine basins and the sedimentation areas adjoining the Carpathian belt. There, the Moutnice limestones and, particularly, the Pouzdřany marls the sedimentation of which terminates in zone NP22 were deposited at the Eocene/

Oligocene boundary. At present the Eocene/Oligocene boundary is unanimously placed in zone NP21 between the foraminiferal zones of *Turborotalia cerroazulensis* (Upper Eocene) and *Globigerina angiporoides*, *Pseudohastigerina naguwichiensis*. As a whole the Pouzdřany marls are bathypelagic sediments of the open sea. The deposition of the Pouzdřany marls was followed by the sedimentation of non-calcareous diatomites, dolomites and calcareous diatomites assigned to zone NP23 (or to the lower section of the Uherčice Formation). The period adverse to the evolution of marine fauna continued up to the brown or grey clays of the zones NP23 and mainly NP24 overlying the diatomite series. Clay sedimentation persists up to zone NP25, i.e. to the oldest Egerian time during which glauconitic sands were deposited, while pelagic sediments of the upper bathyal zone were deposited in the course of younger Egerian time. The latter sediments are called Boudky marls and terminate the Egerian sedimentation together with the freshened Křepice Formation in the original sedimentation area of Pouzdřany. The diatom series of strata (KRHOVSKÝ, 1981) is considered to be equivalent to the older section of the Menilitic Formation.

In general, the biostratigraphic evaluation of the principal younger Tertiary sedimentation areas in the Alpine—Carpathian region points to certain differences in the western, i.e. pre-Alpine section and in the region adjacent to the Carpathians. In the pre-Alpine regions of Bavaria, Vorarlberg and Austria, the boundary facies of the Eocene and Oligocene placed in zone NP21 are characterized mainly by Lithothamnium limestone facies in the basement of which Discocyclina marls are present. Upwards, to the overlying strata up to the older section of zone NP23, appear signs of a certain transition to brackish character, impoverishment in faunas, probably in connection with a certain isolation of the sedimentation space. But simultaneously at the Eocene/Oligocene boundary the flysch Deutenhausen beds "Altdorfer Flysch" etc were still depositing in the inner side of that time sedimentary area.

As follows from the general review, the typical Menilitic Formation did not develop in this region. Nevertheless, partly the so-called "Fischschiefer"correspond to it facially and stratigraphically.

The following Formations "Heller Mergelkalk, Bändermergel" (zone NP23), "Tonmergelschiefer" (NP24) are, according to FUCHS, of flyschoid character and to a certain extent they are analogous with the Krosno-Ždánice lithofacies in the Carpathians. In the Alpine external zone the conditions are similar to those in the Carpathians including the cross-facies zonality.

## Conclusions

*1* A large-area unification of flysch troughs that were not closed with the Old Pyrenean—Illirian phase (between the Lutetian and Priabonian) or later, during the Pyrenean phase (between Priabonian and the Oligocene) took place in the Carpathians and partly in the Alps by the end of the Eocene and in the beginning of the Oligocene. Lithofacies of the Globigerina marls and especially of the Menilitic Formation and its equivalent "Fischschiefer" extended in the outer external zone of the Carpathians (contingently of the Alps) prove this unification.

*2* In the course of the sedimentation of menilite—bituminous claystones (with or without menilite cherts) which has various facial differentiations and a maximum time extent beginning of Kiscellian—beginning of the Miocene, flysch and non-flysch facies of bituminous claystones (eventually with transitions into molasse developments) substitute each other. Transversal facial zonality is often obvious.

*3* In connection with the beginning of Paratethys formation (i.e. the outset of Tethys desintegration), it is important to mention that during Late Oligocene still the sedimentation of flysch and flyschoid sequences took place both in the Carpathians and in the pre-Alpian region. Some of these (e.g. the Krosno Formation in the Silesian unit of the Carpathians) are thousands of meters mighty. In this period one still cannot be speak about any signs of desintegration in the area of Tethys itself situated on the external side of the forming fold mountains.

*4* Shallow seas situated at the back part of the mountains have in the individual basins their own development essentially different from the flysch sedimentation areas of the Tethys s.s. northern branch. Numerous paleogeographical changes including the main anoxic period mentioned e.g. by BÁLDI et al. (1984) are more or less only of local importance. "Molasse like" sedimentation at the back-part of the fold mountains synchronous with the external flysch sedimentation should be differentiated from the molasse proper itself and designated as the premolasse.

*5* The molasse sedimentation begins no sooner than after termination of the collision, i.e. after the connection of the newly formed shallow molasse basins of foredeep type with intramontanes depresions and back-arc basins. At the same time during the Savian movements this area became isolated from the deep ocean spaces of Tethys rests. Since that time we can speak about Paratethys. This conception corresponds to the redefinition of the term Paratethys presented by SENEŠ (1959, 1961).

In case of the Carpathians it is undoubtedly after the Savian movements, i.e. since Eggenburgian. In the Alps the Purchkirchner Series has a character lithologically close to the molasse, however, according to FUCHS (in OBERHAUSER, 1980) it lacks flysch material that appears in Eggenburgian "Nagelfluh" with glauconitic sandstones with conglomerates. This most probably indicates the end of flysch nappes moving toward the front of the Alps and since this very moment the Alpian molasse zone gets a character of a genuine foredeep. Consequently, the Purchkirchner Formation can be also considered as a premolasse and the molasse sedimentation in the proper sense also begins no sooner than in Eggenburgian.

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# THE OCCURRENCE OF TUFFACEOUS HORIZONS IN THE TERTIARY OF THE POLISH LOWLAND AND THE CARPATHIAN FOREDEEP

by

K. Matl and M. Wagner

The well-surveyed Tertiary deposits of autochthonous character are distributed over two distinct areas in Poland (Fig. 1): one exceeding 150 000 km$^2$ in the Polish Lowland (central and northern part of the country); the other, considerably smaller, is situated in the south—in the Carpathian foredeep (Ney et al., 1974). The two sedimentary basin are separated by denudation zone connected with the metacarpathian ridge in places up to several dozen kilometres wide, that zone is almost completely devoid of Tertiary deposits. The two separate basins differred distinctly in tectonic regime, rate of basement subsidence, type of deposits, thickness of profile and stratigraphic range.

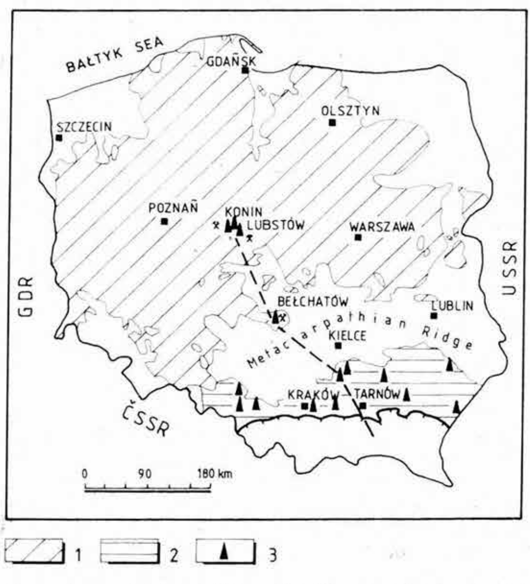


*Fig. 1.* Some important localities of tuffaceous horizons in Poland
*1* Boundary of the Polish Lowland Miocene, *2* boundary of the Carpathian fore-deep Miocene, *3* exposures of the tuffaceous horizons

The differences are substantial; in consequence, up to the present the correlation of the profiles of the two basins has remained unsolved.

The Carpathian foredeep deposits the main part of which has been assigned to the Middle Miocene (Upper Karpatian, Badenian, Lower Sarmatian), consist of marine sediments with rich micro- and macrofauna of biostratigraphic importance.

On the other hand, the Miocene of the Polish Lowland consists of deposits of limnic type with abundantly coal-bearing swampy sediments of practical importance. Its stratigraphy was based on differentiated spore—pollen assemblages with an auxiliary utilization of lithological characteristics (CIUK, 1980).

In recent years the search for correlable horizons led to the identification of pyroclastic rocks in the Lowland Miocene, in the Bełchatów and Konin areas (Fig. 1—3); these rocks can be connected with the extensive, effusive Carpathian volcanism of the Badenian—Sarmatian ages (MAHEL, 1983).

The discovery of pyroclastics in a few localities of the Polish Lowland region permits to attempt the correlation—on a general scale—of some basic links of the Polish Lowland Miocene with stratigraphically the well-documented Badenian and Sarmatian deposits of the Carpathian foredeep. This is possible on the assumption that the a common source of volcanogenic material was situated south of both those zones, i.e. in the Carpathian Mountains. It might have been a relatively common late-geosynclinal rhyolite—andesite neovolcanism of Badenian—Sarmatian age, from the region of Slovakia and northern Hungary (MAHEL, 1983). The tuffaceous intercalations of the Carpathian foredeep are of the same age and petrogenetic type. They have outcrops (Fig. 1) at many localities in the whole foredeep region. The majority of them are connected with the Skawina and Chodenice Beds (Lower and Middle Badenian). According to ALEXANDROWICZ and PAWLIKOWSKI (1980), tuffaceous intercalations in the Chodenice Beds, which are also found far beyond Poland, have the farthest distribution and the greatest accumulation of volcanogenic material.

The petrographic type (Table 1) of all the pyroclastic rocks in the foredeep is similar, indicating relationship with the rhyolite-andesite volcanism. The rocks are tuffites or, exceptionally, tuffs and, sporadically, also bentonites.

The tuffaceous intercalations $T_1$—$T_3$ in the Polish Lowland Miocene deposits—from the "Konin" brown coal mine and the Bełchatów tectonic trough, to have analogous genetic characteristics. They represent a rhyolitic and rhyolitic—dacitic volcanism. The composition of the pyroclastic material is very similar bat its percentage is considerably smaller, especially at Konin, due to a greater distance from the source of eruption. The grain size of the pyroclastics is finer; due to this suffered —to a greater extent—degradation and secondary transformation. The secondary clay components are of a different nature a fact connected with conditions of the sedimentation in a low-pH peat-bog basin. In the Konin area the basin was characterized by poor subsoil drainage, which resulted in the development of metabentonites and wetzsteins. At Bełchatów it had intensive drainage which, in turn, contributed to the formation of tonsteins (paratonsteins). Alumosilicate decomposition proceeded towards illitization and kaolinization.

From the petrographic point of view the tuffogenic $T_1$—$T_3$ (Tab. 1) markers are tonsteines (paratonsteins), meta-bentonites and honestones (wetzschiefers). The $T_1$ layer is a paratonstein in the Bełchatów mine but in the Konin one it changes into a meta-bentonite grading locally into a honestone. In paratonstein the pyrogenic components are represented by pyrogenic quartz, feldspars (sanidine), partly apatite and volcanic glass. Alumosilicates (biotite, feldspars) and volcanic glass were partly altered into kaolinite in the peat-bog environment of fairly high drainage. In meta-bentonite autigenic illite is very common whereas the pyroclastic components are similar to those identified in paratonsteins. Illite was formed by degradation of kaolinite.

Fine-grained volcanic glass forms the matrix of the rock. Its chemical composition could not be determined in details because of the basic troubles met in seperation of sufficient amounts for analysis. However, high content of pyrogenic quartz

Table 1

**Comparison and mineral characteristics of the main occurrences of Middle Miocene rocks containing pyroclastics**

| Area | | Age | Litostratigraphic unit | Mineral composition of the pyroclastic fraction | | | | | Petrografical characteristic |
|---|---|---|---|---|---|---|---|---|---|
| | | | | pyrogenic quartz | feldspars | volcanic glass | femic minerals | biotite | |
| foredeep | Chełm Wielki | Middle Badenian | Skawina Beds | + | + | + | — | — | Tuffite: rhyodacitic type |
| | Wieliczka | Middle Badenian | Skawina Beds | + | + | + | + | + | Tuffite: andesitic type |
| | Chmielnik | Lower and Middle Badenian | Baranów Beds, Spirialis clays | + | + | + | + | + | Bentonite |
| Carpathian | Wieliczka Bochnia | Middle Badenian | Wieliczka Beds | + | + | + | + | + | Tuffite: rhyo-andesitic type |
| | Chodenice | Upper Badenian | Chodenice Beds | + | + | + | — | + | ? |
| | Ochojec | Upper Badenian | Chodenice Beds | + | + | + | — | + | Tuffite: rhyolitic type |
| Polish Lowland | Bełchatów | Middle Miocene | Middle Polish, Adamów, Scinawa B. | + | + | + | — | + | Paratonstein: rhyodacitic type |
| | Konin Area | Middle Miocene | Middle Polish Beds | + | + | + | — | + | Metabentonite: rhyolitic type |

(5–9%) and domination of K-feldpars over plagioclases point to acidic, presumably rhyolitic or rhyo-dacitic type of volcanic dust.

Volacanogenic layers occurring in the Carpathian foredeep are tuffites, bentonites and, rarely, lithoclastic tuffs. Apart from volcanic glass they contain pyrogenic quartz, biotite, femic minerals (amphibolites and pyroxenes) and feldspars. Types of feldspars and the chemistry of volcanic glass point to acidic, rhyolitic, rhyo-dacitic and/or andesitic volcanism.

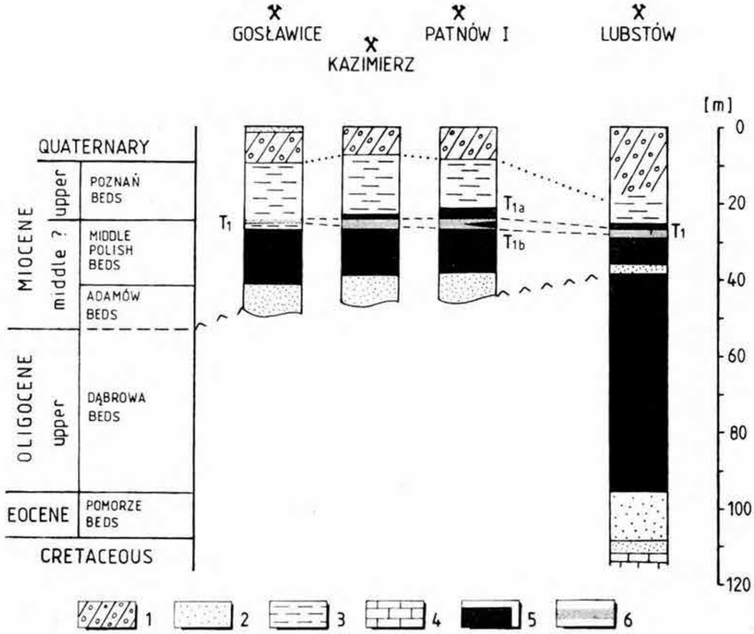


*Fig. 2.* Tuffaceous horizon $T_1$ in the Miocene profile of the Konin area brown coal mines

*1* Gravels, *2* sands, *3* clays, *4* cretaceous limestones, *5* soft brown coal, *6* tuffaceous intercalations

The uppermost tuffaceous horizon $T_1/T_{1a}+T_{1b}$) has the widest extent in the Polish Lowland. It is found in all brown coal exposures of the Konin area (Kazimierz, Pątnów, Lubstów outcrops) and at Bełchatów mine (Fig. 2, 3). Horizon $T_1$ is relatively thick (up to 40 cm), has partly a two-layered structure, and is characterized by graded bedding. Its distribution over a relatively large area and the stability of structure predispose it to a role of an important marker horizon. It played that role in the correlation of Miocene profiles of Konin and Bełchatów. It seems that horizon $T_1$ can be simultaneously correlated with an extensive tuffite horizon of the Chodenice Beds (the Bochnia Tuffite, according to Alexandrowicz and Pawlikowski, 1980) in the Carpathian foredeep (Fig. 4). The latter authors (1980) ascribe to it a role of a marker horizon of lithostratigraphic importance in the Paratethys region. Evidence from the Polish Lowland indicates that it can also be identified beyond the Carpathian region. Accordingly, it has a considerably increased extent and is important as a common marker horizon for Miocene profiles in separate basins: one in the Carpathian foredeep, the other in the epivariscan platform area of Central Poland. The latter finding

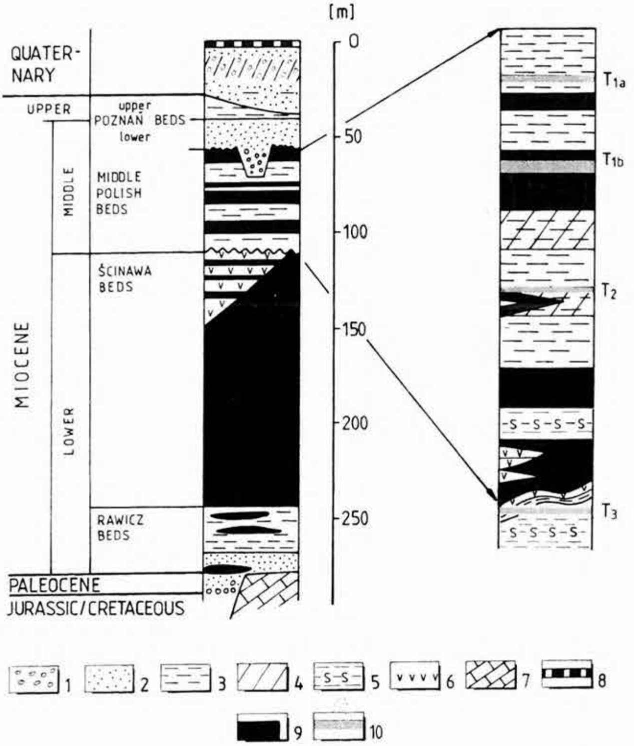


*Fig. 3.* Tuffaceous horizons in the Miocene profile of the "Bełchatów" brown coal outcrop
*1* Gravels, *2* sands, *3* clays, *4* coal clays, *5* sapropelites, *6* lacustrine limestones, *7* limestones, *8* peat, *9* brown coal, *10* tuffaceous intercalations

is important, especially for the Miocene of the Polish Lowland where there are no biostratigraphic horizons which might facilitate a direct correlation with the marine Miocene of the Carpathian foredeep.

As a results of the above accepted correlation, the "Middle Polish" brown coal seam, as well as the Middle Polish Beds and Adamów Beds, assinged up to now to the Upper Miocene (Fig. 2, 3) according to Ciuk (1980), should be shifted to the Middle Miocene. This would make them correspond to the Chodenice Beds (Upper Badenian) in the Miocene profile of the Carpathian foredeep. Similarly, it is postulated that the age of the clay-coal complex at Bełchatów should be lowered (Fig. 4).

Tuffaceous intercalations $T_2$ and $T_3$ from the Bełchatów mine (Fig. 3) presumably correspond to tuffite horizons of the Lower Badenian Skawina Beds in the Carpathian foredeep; according to Alexandrowicz and Pawlikowski (1980) these horizons are also distributed over a very large area. They were reported from numerous localities on the southern side of the Holy Cross Mts. It is likely that the fall of pyroclastic material extended farther northward, beyond the zone of the Carpathian foredeep.

On the basis of the occurrence of tuffaceous horizons it can be assumed that the Badenian and Lower Sarmatian (Middle Miocene, acc. to Rögl and Steininger, 1983) deposits of the Carpathian foredeep have their equivalent in the Bełchatów

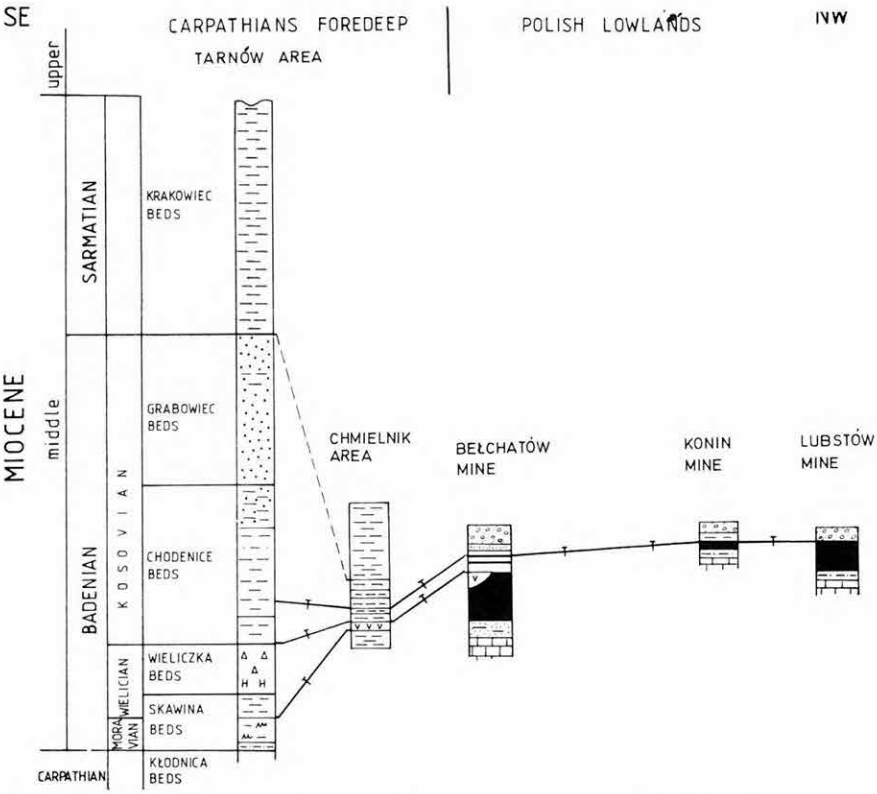


*Fig. 4.* Correlation of tuffaceous horizons of the Miocene of the Carpathian foredeep and the Polish Lowland

clay—coal sequence and corresponding to the Middle Polish Beds and Adamów Beds. On the other hand, the main seam of the Bełchatów area is an equivalent of the Karpatian coal-bearing deposits, and thus belongs already to the Lower Miocene. This finding may have further implications if it is accepted that the main Bełchatów deposit is an equivalent of the brown coal seam of group II (Ścinawa coal seam) in the profile of the Tertiary Lowland.

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# HISTORY OF PARATETHYS

by

L. A. NEVESSKAJA, I. A. GONCHAROVA, L. B. ILJINA, N. P. PARAMONOVA, S. V. POPOV, A. A. VORONINA, A. L. CHEPALYGA and E. V. BABAK

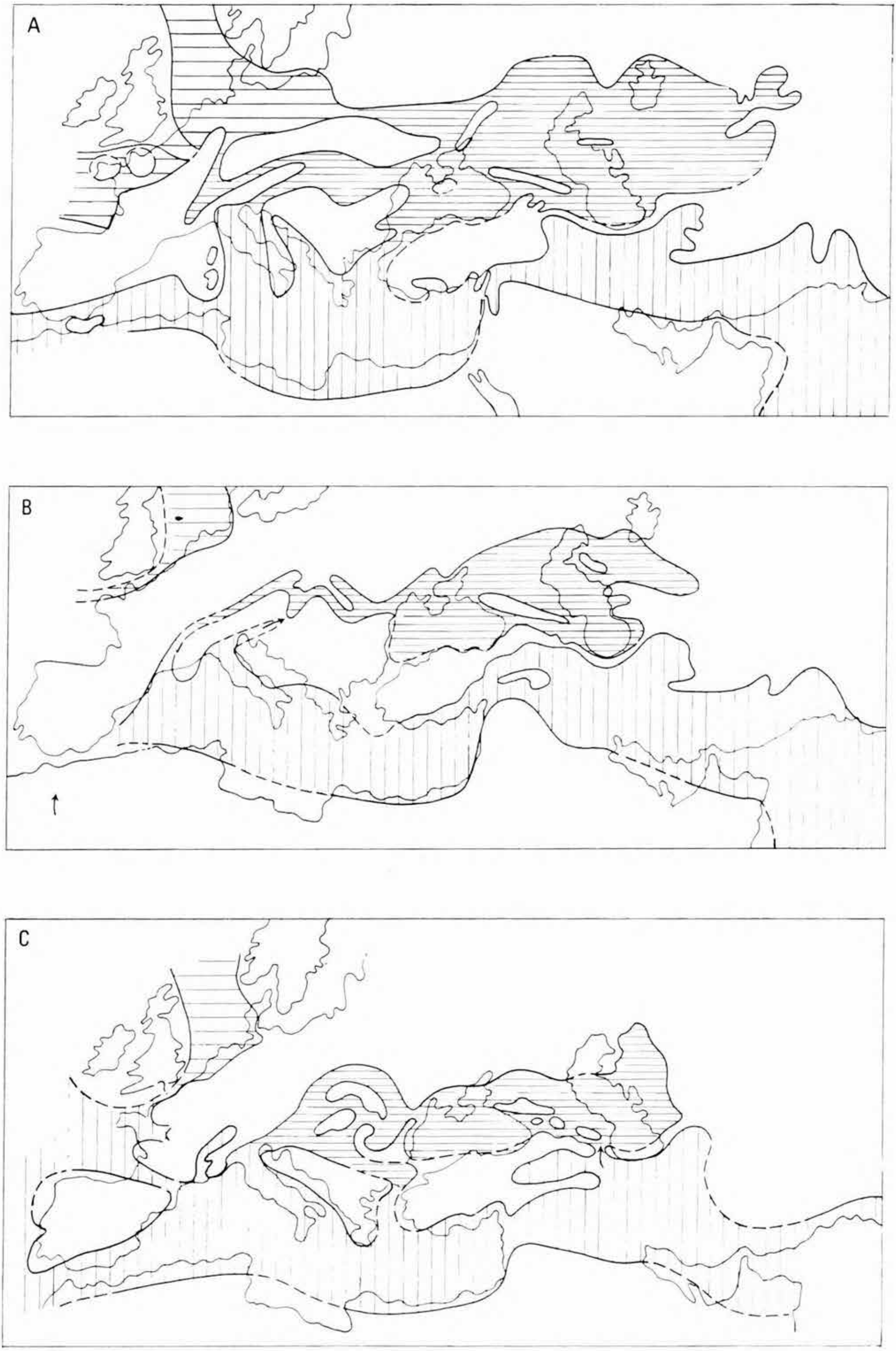
The largest innercontinental Paratethys sea appeared in Oligocene (Fig. 1, A) due to the isolation of the extreme northern part of Tethys as a result of orogenic movements of the Alpine belt, caused by the drawing of the continental plates of Africa, Arabia and Hindustan closer to Eurasia. The principal factors determining the history of Paratethys development were tectonic movements (alpine orogenesis), global changes of the ocean level and climatic variations.

From the end of Early Miocene Paratethys, for the first time since the beginning of its history, subdivides quite definitely into western and eastern parts (Fig. 1, C), its eastern (Euxino—Caspian) part was twice as large as its western (Pannonian) one. Further differentiation occurred later. At present exist two basins only—the Azov—Black sea and the Caspian sea, in the place of Paratethys.

Against the historical background of the process of Paratethys disintegration, connections both within Paratethys, in particular between its Western and Eastern basins, and between Paratethys as a whole and open marine basins appeared and disappeared repeatedly. This changed the character of the whole Paratethys and of its separate parts. On this basis N. I. ANDRUSOV singled out three cycles of development in the Eastern Paratethys (Tarchanian—Karaganian, Konkian—Sarmatian and Maeotian). Each of them started with the entering of saline waters of the world ocean into the basin and the penetration of marine organisms and finished with complete or partial closing of Paratethys and the formation of specific, usually brackish-water fauna. It caused the alternation of basins of various types. A number of cycles in the Western Paratethys have been introduced by J. SENEŠ.

The amount of data available at present makes it possible to single out in the Eastern Paratethys not less than seven cycles during Oligocene and Neogene (NEVESSKAJA et al., 1984). The *initial (I) cycle* (Early Oligocene—beginning of Late Oligocene) was characterized by a rather extensive connection with the Atlantic ocean (Fig. 1, A) and considerable hydrogen sulphide contamination. In the second half of the cycle (Solenovian time) the first closing of Paratethys took place. It was accompanied by the formation of endemic fauna of molluscs and ostracods.

*Cycle II* (Late Oligocene—Early Miocene) started with the immigration of marine fauna from the Atlantic area. Later (Sakaraulian, Eggenburgian) the joining with Tethys happened. In the first part of the cycle hydrogen sulphide contamination was the strongest, due to which fauna at the end of Oligocene—beginning of Miocene is practically unknown. It brings about special difficulties in the determination of the Oligocene—Miocene boundary and basin reconstruction (Fig. 2, IIa). The Sakaraulian basin contained diverse, very warm-water mollusc fauna with a great number of species and genera common with Eggenburgian basin of Western Paratethys. However, there are here many species known neither in Western Paratethys nor in


A
B
C

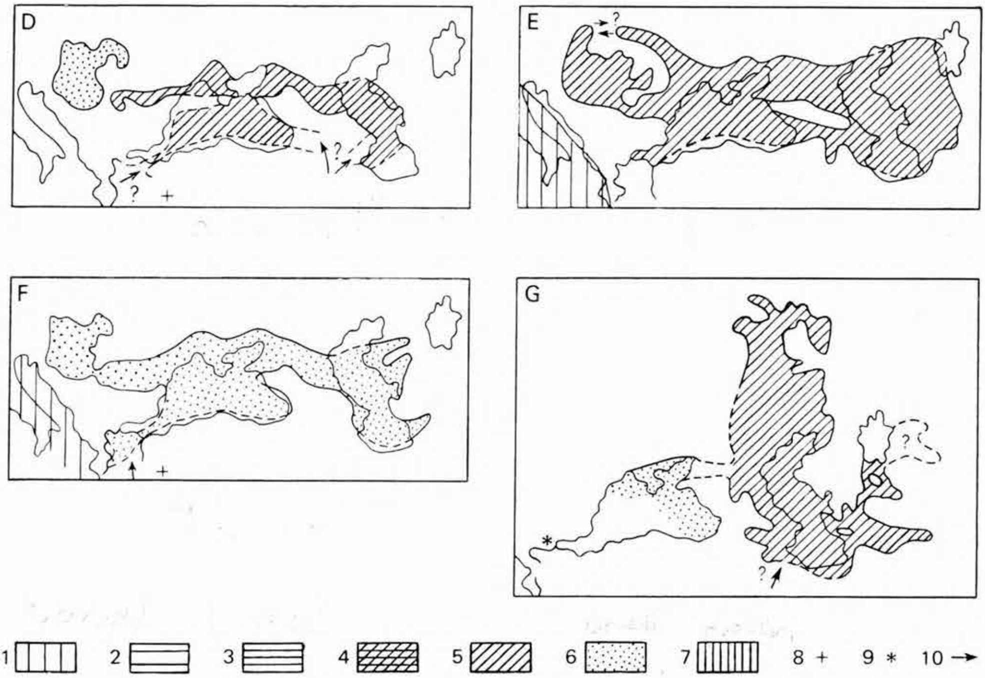



*Fig. 1.* (A—G) Palaeogeographic sketches of the Paratethys and adjacent basins

Tethys: *1*, *2* boreal basins; Paratethys: *3* marine, *4* marine with a somewhat abberrant salinity, *5* semimarine, *6* brackish-water, *7* freshwater.—*8* The location of molluscs of Pontian type (TANNER, 1974), *9* the location of molluscs of Akchagylian type (TANNER, 1982), *10* probable communication of Paratethys with open marine basins

the Mediterranean Tethys at that time. This permits us to assume the possibility of connection of the Eastern Paratethys not only with the Mediterranean Tethys but with also Indo-Pacific region of Tethys (Fig. 1, B). The end of this cycle (Kozachurian, Ottnangian) was marked by the closing of Paratethys and the formation of brackish-water fauna of molluscs and other groups. At that time there were episodical connections between the Western and the Eastern Paratethys owing to which some specific brakish-water species penetrated from the Western Paratethys into the Eastern and vice versa (POPOV, VORONINA, 1983). Cycle II corresponds to Eoparatethys according to SENEŠ. This cycle was characterized by still faint differentiation of the Eastern and the Western Paratethys and by presence of hydrogen sulphide contamination.

*Cycle III* (Mesoparatethys according to SENEŠ) was characterized by precise division into the Eastern and the Western Paratethys (Fig. 1, C). At the beginning of the cycle (Tarchanian) the Eastern Paratethys had communications both with the Eastern Paratethys (Carpathian basin) and the Indo-Pacific regions of Tethys. Early Tarchanian fauna included polyhaline molluscs and foraminifera, echinoids and others. In the second part of Tarchanian the salinity slightqlitly decreased, though the basin has been still marine.

The salinity of the Tschokrakian basin following the Tarchanian was probably high enough but deviated from the normal to a certain extent. This basin had close

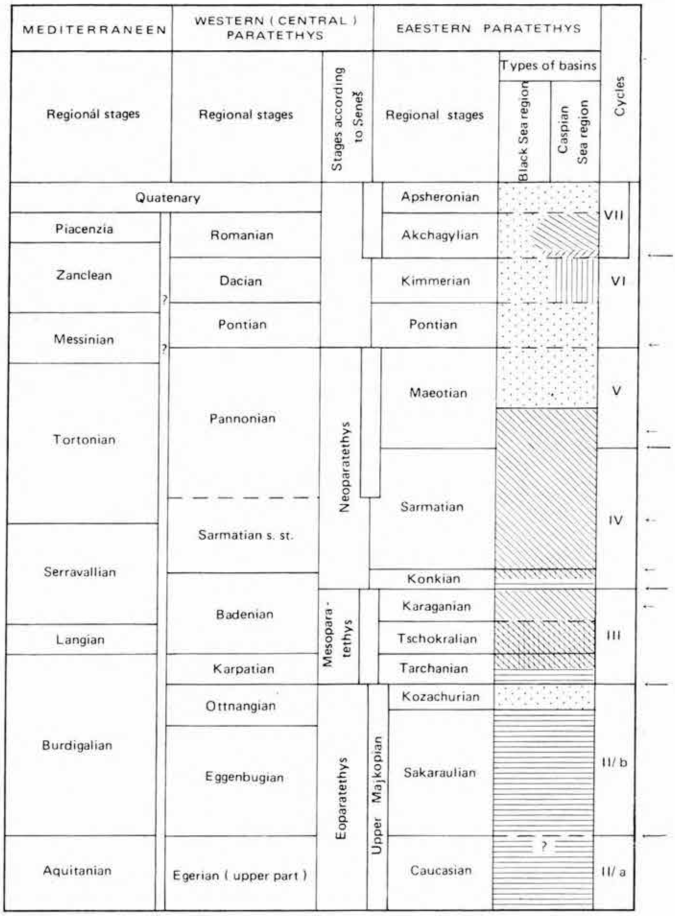



*Fig. 2.* Geological time table

connections with Tethys. This communication must have taken place in the south-eastern part of the Eastern Paratethys, as it is here that the most diverse fauna is observed. In Late Tschokrakian and Karaganian times the Eastern Paratethys began closing up and freshening. This caused extinction of the overwhelming majority of marine species and the evolution of endemic forms. Only in the second half of Karaganian time the episodical communication with Tethys waters was marked in the southeast of Paratethys. This caused a brief appearance of some marine species. Gypsum formation was characteristic of the Karaganian time. This stage may correspond to the stage of gypso- and salt accumulation of Middle Badenian. The Western Paratethys at the beginning and at the end of cycle III (Early and Late Badenian) had rather extensive communications with the Tethys (RÖGL et al., 1978).

*Cycle IV* of the Eastern Paratethys began with the Konkian time and lasted till the end of Sarmatian, i.e. it enveloped the second half of Middle and the beginning of the Late Miocene. This cycle corresponds to Neoparatethys according to SENEŠ. At the beginning of the cycle rather broad communication elisted with the Tethys.

The salinity became close to normal marine one, and the basin got inhabited by polyhaline species of molluscs, Bryozoa, Echinodermata, Foraminifera and others. The communication with the Tethys was probably in the southeast of Paratethys supported by the presence of most polyhaline assemblages in these regions (Transcaspian and Eastern Georgia). Some connection might also have existed between the Konkian basin and the Late Badenian Western Paratethys. During the first half of Sarmatian time Paratethys was again a unique basin (Fig. 1, D), and then complete isolation of the Western Paratethys from Eastern one occurred. In comparison with the previous time in Early Sarmatian there was impoverishment of all groups of organisms (ILJINA et al., 1976; PARAMONOVA et al., 1979) as a result of the extinction of a large number of polyhaline genera, caused by the reduction of salinity. In such situation few euryhaline species of molluscs, ostracods, foraminifera and other groups because widespread. In the course of time some of them gave birth to numerous endemic species. The character of the Middle and Late Sarmatian fauna in general points to the continuing decrease of salinity, though in the first half of Middle Sarmatian episodical communications with the Mediterranean probably still appeared. Absolute dating (ILJINA et al., 1976; VASS, 1979; CHUMAKOV et al., 1984) points to the possibility of correlation of Lower Sarmatian to Upper Serravallian and of Middle and Upper Sarmatian to Lower Tortonian. Beginning with the second half of Middle Sarmatian the Pannonian brackish-water basin was formed as a result of the isolation Western Paratethys from the Eastern one. Fully endemic mollusc fauna appeared there during Pannonian time.

In the post-Sarmatian time in the Eastern Paratethys *cycle V* — (Maeotic)-began whereas in the western part of Paratethys the previous cycle continued (Fig. 1, E; Fig. 2). The Early Maeotian transgression was accompanied by the coming of marine species from some basin which was communicating with the Mediterranean basin. This connection might probably take place in the southeast but might have passed through southwestern regions of the Eastern Paratethys (ILJINA et al., 1976; ILJINA, 1980; STEVANOVIĆ, ILJINA, 1982). It is possible that the maximum of the Maeotian transgression reflects the maximum of Tortonian transgression. By the end of Early Maeotian isolation of the Eastern Paratethys began, salinity decreased and a brackish-water fauna replaced marine one.

*Cycle VI* of the development of the whole of Paratethys began in post-Pannonian time in Western Paratethys and in post-Maeotian in the Eastern Paratethys. The vast Early Pontian basin was formed, Paratethys again becoming united (Fig. 1, F). In the Eastern Paratethys this cycle began with the coming of brackish-water fauna, alien to the previous late Maeotian one, as well as some marine elements. This factor makes it possible to single out this cycle as an independent one (Fig. 2). The majority of Early Pontian molluscs of the Eastern Paratethys is of Aegean origin. Here were only single common species which had genetic ties with the Pannonian fauna and inhabited the Western Paratethys. Consequently, the communication between the Western and Eastern Paratethys in Early Pontian was complicated (NEVESSKAJA, STEVANOVIĆ, 1985). Beginning with the second half of Pontian Paratethys started to skrink thus by the end of Pontian time the Pannonian basin disappeared and the Eastern Paratethys was devided into the Euxinian and Caspian basins.

In the euxinian region cycle VI lasted till the end of Pliocene (Tchaudian time), whereas in the Caspian region it continued to the end of Kimmerian.

In post-Kimmerian time the new *cycle VII* began in this area (Fig. 2). It lasted till the end of the Pliocene (Akchagylian—Apsheronian). Its was marked by the coming of the marine Akchagylian fauna which had been evidently caused by the communication of the Eastern Paratethys with the Mediterranean (Nevesskaja, Trubikhin, 1984) which had reappeared somewhere in the southeast (Fig. 1, G). The Akchagylian molluscs are definitely of mediterranean origin and have no genetic ties with the Sarmatian ones (Paramonova, 1977). The Akchagylian basin in the Caspian area of the Eastern Paratethys replaced by the brackish-water Apsheronian basin and later by the also brackish-water Baku, Khazarian, Khvalynian and Neocaspian basins so that cycle VII which started at the beginning of Akchagylian continued in the Pleistocene, too.

In the euxinian region the last Neogene Ponto-Tschaudian cycle (VI) was replaced by the Quaternary cycles reflecting interchanges of brackish-water (ancient euxinian and neo-euxinian) and marine (Uzunlarian, Karangatian and Black Sea proper) basins.

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# CONTROVERSIAL APPROACHES TO THE USE OF MIDDLE—UPPER NEOGENE CHRONOSTRATIGRAPHIC UNITS FROM THE TETHYS AND THE PARATETHYS

by

I. ANDREESCU

### General remarks on circum-Mediterranean Neogene stages

The Tethys or Mediterranean area s. str. represents the Neogene classical area, where Neogene stages were first established. Consequently, these stages should be considered as a standard for the global Neogene. For various reasons, more or less objective, the present status of the Mediterranean Neogene stages, however is rather unsatisfactory, especially when the Miocene stages are taken into account.

From among the most contradictory features, still hampering a clear meaning of these stages, the following can be mentioned:

— the lack of an unambiguous definition of the base of the Aquitanian stage and of the Miocene—Oligocene boundary respectively;

— the time-span of the Burdigalian stage is still obscure;

— the time-span of the Langhian stage is unknown. Early Langhian transgression took place after a pronounced continentalization time-interval and, as a result, there is a gap between the Burdigalian and the Langhian;

— other hiatuses seem to be located between the Langhian and the Serravallian and between the Serravallian and the Tortonian;

— the relationship between the Upper Tortonian and Lower Messinian is obscure and, on account of this, the lower boundary of the Messinian is often arbitrarily considered;

— the time-span of the Messinian stage is not yet precised and, closely related to this problem is the question of the position of the Miocene—Pliocene boundary.

As regards the Paratethys area, it is sufficient to mention that at least three, if not more, so-called "official" chronostratigraphic scales are currently used for the Neogene deposits. Beside these "official" scales many authors utilize a kind of scales of their own, contributing to make the puzzle of Neogene chronostratigraphy still more discouraging. As a result in the Paratethys the most controversial features regarding the Neogene stages include:

— the position of the Karpatian stage and its correlation with potential equivalent units from the E Paratethyan and Mediterranean area

— the meaning of the Badenian stage and its subdivisions;

— the time-span of the Sarmatian stage in the Western and Central Paratethys on the one hand and in the E Paratethys, on the other;

— the meaning of the Pannonian: Pannonian s.l. versus Pannonian s.str.;

— the time-span of the Meotian stage, its subdivision and correlation in various basins of the E Paratethys;

— the time-span of the Pontian stage in the W Paratethys by comparison with the same unit in the E Paratethys and the chronostratigraphic position of this stage in different basins.

A detailed analysis of all of these highly disputable questions is beyond the purpose of this contribution. Consequently, only some features considered to be crucial in Tethyan—Paratethyan Neogene stratigraphy and correlation will be pointed out.

### Main reasons leading to controversial use of Neogene stages

The controversial use of these stages is, above all, conditioned by the meaning and accuracy in which the biochronologic and/or biostratigraphic units have been individualized.

In the Mediterranean area neither planktonic foraminifera and calcareous nannoplankton are always suitable for any reliable zonation or long-distance correlation (BIZON, 1979; MÜLLER, 1979; ELLIS, 1979 etc). On the other hand, magnetostratigraphy is still of limited use in the Mediterranean Neogene record, since the available data are poor and conflicting (RIO et al., 1984). As a result, many datings regarding biotic events in the Mediterranean have been taken, as a rule, from events calibrated in extra-Mediterranean medium- and low latitude realms (RYAN et al., 1974; BERGGREN, VAN COUVERING, 1974; CITA, 1975). In addition, it seems that many planktonic taxa, recorded in various Mediterranean sectors, had a delayed or, on the contrary, an earlier occurrence in comparison with other areas of mid—low latitude. Thus the first occurrence of *Neogloboquadrina acostaensis* in the N Pacific and N Atlantic is dated at about 10.2 Ma, whereas in tropical regions its first occurrence is delayed by 1.6 Ma (BARRON, 1985). On the other hand, the first occurrence of *Globorotalia margaritae* is delayed in Mediterranean in comparison with the extra--Mediterranean regions (RIO et al., 1984). As regards the nannoplankton, MÜLLER (1979) and ELLIS (1979) noted that in the Mediterranean the paucity, or absence, of several index taxa makes difficult, or even impossible, the recognition of Miocene standard zones.

In the circum-Mediterranean region large areas are covered by Neogene sequences with brackish or freshwater fossils (Mollusca, Ostracoda, etc). Since the brackish-water organisms evolved faster than the marine or the freshwater ones, they frequently provide valuable markers for distinquishing minor stratigraphic units (substages or even "horizons"), but their usually restricted area of distribution does not allow, in many cases, any long-distance correlation. As a result, a stage, or substage, established in one basin, may not accurately be separated in other, neighbouring basins. Thus, in the Paratethyan area, the Sarmatian substages of the E Paratethys could not be identified in W Paratethys; the Meotian stage of the E Paratethys is roughly equivalent to the middle—upper part of the Pannonian s.str. (W Paratethys); the Dacian stage (E Paratethys), individualized in the Dacian basin on the basis of Lymnocardids and Congeria taxa, has been adopted in the W Paratethys as well in spite of the fact that, in this region, the diagnostic molluscs of the Dacian stage are absent.

In the circum-Mediterranean sections the marine sequences occasionally alternate, or intertongue, with brackish, freshwater or terrestrial deposits. When terrestrial deposits contain mammal remains, these represent tenable tie-points in age assignation and interregional correlation attempts. Unfortunately, in many instances, mammal fossils have been found unrelated to any other fossils. Moreover, beside the scarcity of mammal sites, no correlation between larger sites and smaller mammal sites has so far been available. As a result, the stratigraphic position of most mammal zones, and especially of $MN_9$—$MN_{13}$ zones, i.e. the biochronologic units defining the Vallesian and Turolian land stages, is still highly controversial. These controversies are closely dependent on the following features:

— insufficient knowledge of some mammal taxa ranges;

— possible diachronism of various taxa, or assemblages, in different regions;

— misinterpretation of palaeomagnetic polarity reversals (if available), leading to an inaccurate age assignation of the mammal sites.

In our opinion, most confusion is induced by miscalibration of sequences containing mammals. Hence the need for some pertinent considerations as to the palaeomagnetic method, increasingly felt in biostratigraphy in recent years as it is.

As pointed out by RYAN et al. (1974), BERGGREN and VAN COUVERING (1974) and BERGGREN et al. (1983) the palaeomagnetic scale is a relative time-scale. Its weakest aspect consists in a somewhat subjective polarity interval delimitation and nomenclature. Since in both the ocean floor lineation and sedimentary columns no unique palaeomagnetic polarity pattern is available at present, several palaeomagnetic scales for the Neogene are in use. Consequently, there is a wide field in optioning for one scale or another and, in close dependence on age assignation, the adopted scales often differ substantially even for one and the same site, when investigated by various authors. A spectacular example in this sense is given in Table 3, showing that the

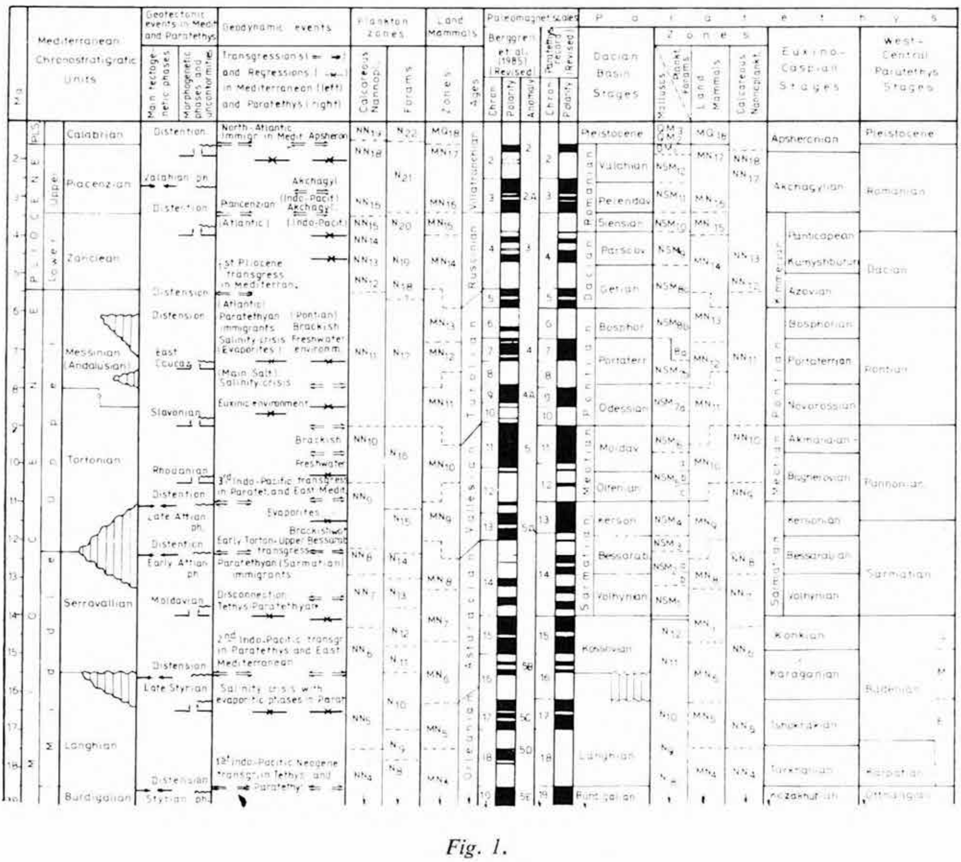

*Fig. 1.*

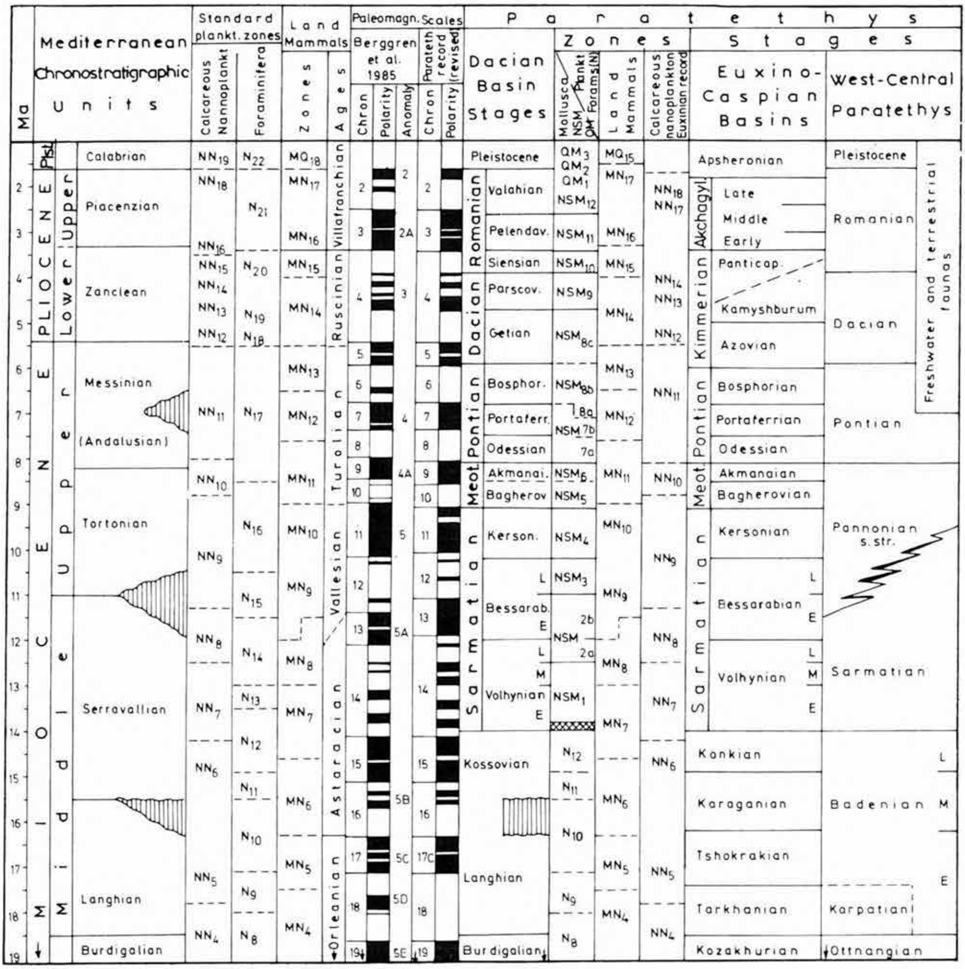


*Fig. 2.*

$MN_{11}$ zone has been assigned to the Upper Sarmatian (Kersonian), the Meotian, or the Upper Meotian–Lower Pontian interval, respectively.

Similar paradoxical chronostratigraphic conclusions can be reached for any group of fossils, when palaeomagnetic polarities are used without any proper tie-point correlation. Thus HSÜ (1978), relying on lithologic record and spore-polen and diatom assemblages and taking into account some palaeomagnetic data acquired at Site 380, Black Sea (DSDP, Leg. 42/B), concluded that the lower segment (210 m) of the column would pertain to the Messinian stage. In fact these deposits are Sarmatian in age, as KOJUMDGIEVA (1983) pointed out. The following sequences of the same site assigned by HSÜ (or his co-workers) to various chronostratigraphic units, are shown in Table 4. As a matter of fact this table needs no further comments because it is significant enough to express the controversial results which could sometimes be reached in circum-Mediterranean Neogene chronostratigraphy.

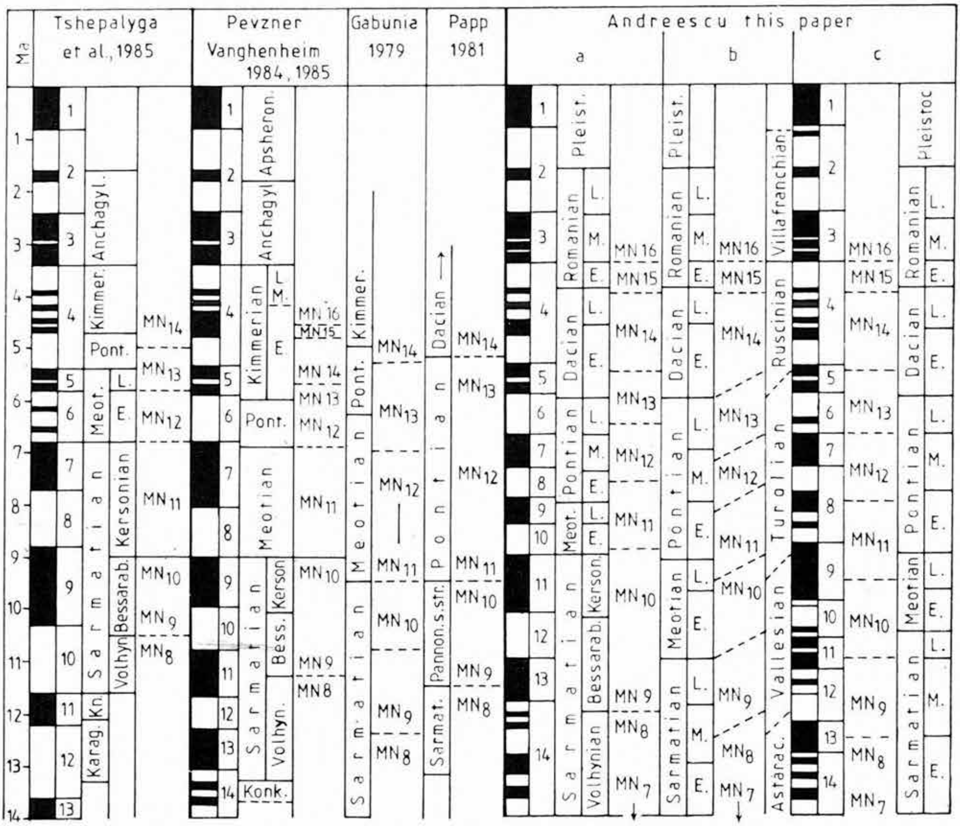


*Fig. 3.*

**Conclusions**

Some features leading to the controversial use of Neogene chronostratigraphic units from the Tethys and the Paratethys have deliberately been overlooked, or only suggested in this too short contribution. However, we think the reader will be able to detect them, and finally to discern himself the difficulties which may discourage those who are less familiar with this real welter of Neogene stages. Unfortunately, given the limited space available, we have been unable to argue for our own opinion. Nevertheless, some of them are rendered out in the accompanying tables (1—4) and in ANDREESCU et al., this volume, p. 113. In Tables 1 and 2 an integration of biochronologic units of the Langhian—Early Pleistocene interval with the main geotectonic and geodynamic events of Tethys and Paratethys has been attempted. The calibration and age assignation of these units have been made by using two somewhat different palaeomagnetic scales (both being quite plausible). Note that in Table 2 in spite of the fact that same palaeomagnetic scales, as in Table 1, were used, the biochronologic and chronostratigraphic units of the Sarmatian—Pontian time span have been calibrated in a sensibly different manner. This calibration seems to be rather plausible as far as the Sarmatian is concerned but it is less reliable for the Meotian and Pontian (see also: ANDREESCU et al., this volume).

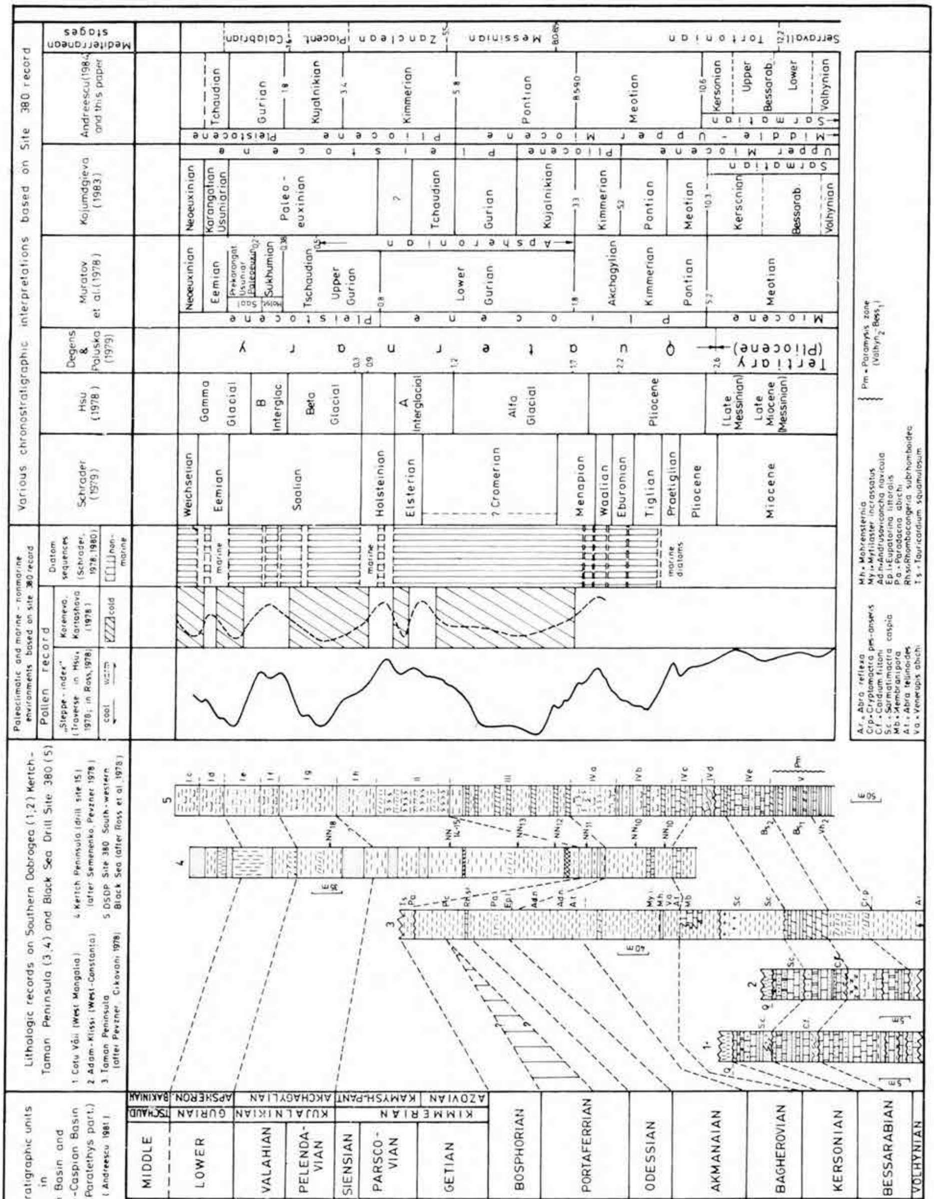

*Fig. 4.*

In Table 3 various ways of age assignation of Neogene Mammal units have been illustrated by using 4 different palaeomagnetic scales. In this table, our option goes towards alternative (c).

Concerning some references cited in the text, the reader is kindly requested to consult the exhaustive bibliography presented by F. RÖGL in the second part of the volume "Mediterranean and Paratethys Neogene, Report on Activity of the RCMNS Working Groups and Bibliography, 1979—1984.— Budapest, 1985."

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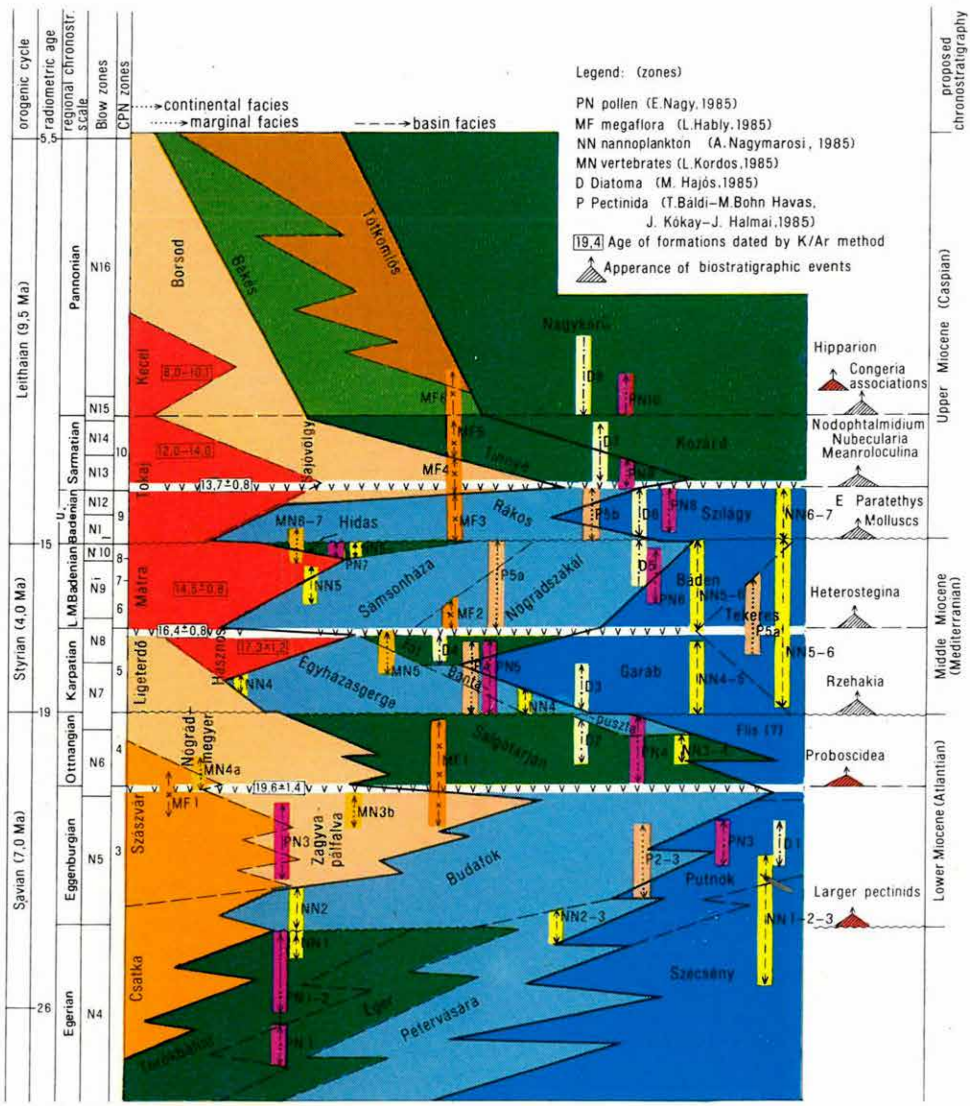


*Fig. 1.* Miocene bio-, litho- and chronostratigraphy of Hungary

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# THE BIO-, LITHO- AND CHRONOSTRATIGRAPHY OF THE HUNGARIAN MIOCENE

by

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The paper presents, from among the stratigraphic sequences spanning the interval between Hungary's Egerian and Pontian formations, the Eggenburgian—Sarmatian stratigraphy. The Pannonian is dealt with in a separate paper.

*1* The above mentioned part of the Miocene is made up of 37 formations forming a continuous vertical succession. Out of these, 26 formations studied in detail are shown in Fig. 1. These formations are very well exposed in surface or drilling key sections, their research history and nomenclature are very well-known. Their boundaries and lithological, sedimentological and facies characteristics are studied in fair detail. The paper synthesizes the results obtained by about 15 biostratigraphic methods, by geochronological methods, by the study of orogenic cycles and palaeogeographical reconstructions.

The most important of these are:

a) the Hungarian Miocene is displays a complete lateral succession of facies: continental deposits—continental and/or basin-margin volcanics—basin-margin facies—central deep-basin facies,

b) the isochronity of these isochronous facies counterparts (the so-called "heteropic" facies) is proved:

— by the isochronous markers represented by the three tuff ejections all over the country and their radiometric dates,

— the interbedding mode of occurrence of the continental and marine (in some cases, volcanic) facies,

— the direct and indirect methods of biostratigraphic zonation.

*2* Major motives for the assignation to the Parathethys regional chronostratigraphic system:

*2.1* The lower boundary of the *Eggenburgian* stage is characterized by a regional unconformity and the appearance of the "larger pectinids" (pectinid zone P2—3).

Zone PN3 is characterized by the appearance of Faveotriletes rueterbergensis, the extinction of Laevigatosporites pseudodiscordatus, the exclusive occurence of Faveotriletes pessinensis. The N5 Blow zone encompasses the whole marine interval (Budafok Sand Formation, Putnok Schlier Formation). In the marine vertebrate zonation the predominance of the representatives of the genere Hemipristis, Isurus and Lamna is characteristic. At the end of the Eggenburgian a subtropical megaflora (zone MF1) and the zone MN3b appear.

The transition to Ottnangian time and the kinship relation are proved by the continuous sedimentary cycle and the transient Platanus neptuni—Engelhardtia orsbergensis—Laurophyllum—Calamus noszkyi megafloral zone MF1.

*2.2* The lower boundary of the *Ottnangian* stage is drawn at the episodical rhyolite tuff ejection event 19.6±1.4 Ma (Gyulakeszi Rhyolite Tuff Formation). Its upper boundary is marked by the extinction of some of the palaeotropical elements (e.g. Platanus neptuni) and of Proboscidea (zone MN4a).

Zone PN4 is characterized by the predominance of Salixipollenites helveticus, S. densibaculatus, Myricipites species and plenty of ferns. The Rhaphoneis subtilissima D2 diatom zone proves the persistence of subtropical climate. The terrestrial vertebrate fauna belonging to zona MN4a is amplified, along with surviving Gomphotherium and Prodinotherium, by Zygolophon, Deinotherium, Palaeochoerus and Rhino I—II forms.

All these features readily correlable with zones NN3 and CPN4 identified in the upper part of the Salgótarján Browncoal Formation.

*2.3* The lower boundary of the *Karpatian* stage is marked by a regional unconformity, the basal formations of a new sedimentary cycle, the appearance abundant of Chlamys forms and marine vertebrates (23 taxa) as well as by the joint appearance of Rzehakia forms and Helicosphaera ampliaperta zone NN4.

Zone PN5 is marked by the exclusive occurence of Rudolphisporites species, Phaecocerosporites transversus and Ricciaesporites hungaricus. The megaflora is unsuitable for evaluation because of the marine environment. The Rhaphoneis parilis and Surirella costata—Coscinodiscus pannonicus D3—4 diatom zones are indicative already of a subtropical-mediterranean climate. Only the lower two-thirds of the cycle can be assessed stratigraphically in an exact way on the basis of the zones NN4, N7 and CPN5. The Flabellipecten pasinii—Pecten expansior—Amussium cristatum badense (P4) pectinid zone spans the whole interval of the cycle. At the Karpatian—Badenian boundary, the Tar Dacite Tuff Formation (16.4±0.8 Ma) is found.

*2.4* In *Early Badenian* time a break in sedimentation can be recognized over a part of the study area owing to a latest Karpatian emergence. In case of openwater, continuous sedimentation, the lower boundary can be drawn by the appearance of Heterostegina costata, H. simplex, Orbulina bilobata and O. suturalis.

The lower Badenian is marked by zone PN6 with the predominance and then extinction of Bifacialisporites grandis-Mecsekisporites miocaenicus—M. aequus—M. zengoevarkonyensis. In addition to its stratigraphic value, the Parrotia pristina (predominant) — Quercus pontica miocaenica (first appearance) MF2 megafloral zone proves that a riverian flora came into prominence, i.e. the landmass area widened. The age of the formations assigned to the Lower Badenian is determined, along with nanno NN5 zone and the foraminiferal zones N9, CPN7—8, by the pectinid zones Chlamys elegans and Ch. revolutus (P5a). The transient Actinocyclus ingens diatom zone supports primarily the peristence of the subtropical—mediterranean climate.

*2.5* The time interval of the *Middle Badenian* is spanned, practically in full, by the andesitic volcanism. The stratigraphic assignation was done on the basis of the intertonguing with the NN5 nanno zone, a number of radiometric *K/Ar* age determinations (14.5±0.4 Ma Mátra Andezite Formation) and of the underlying formations. Long-distance, interregional correlation is extremely difficult, because within the Carpathian realm, west of the zone of volcanic activities, a biostratigraphically based interregional correlation is handicapped by brackish-water—paralic—palustrial brown coal formations, and east of it by hypersaline—lagoonal evaporitic accumulations.

*2.6* The lower boundary of the *Upper Badenian* substage is marked by the appearance of sedimentary formations, a regional unconformity and basal deposits. In the bios, the extraordinary changes are marked by the appearance of Coniferae pollen grains, the P. leythaianus—P. aduncus D5b pectinid zonezone and, most of all, by that of a Caspian—brackish Mollusca fauna (Modiolus, Musculus, Ervilia, Cerithium taxa). The *Late Badenian* is determined by the predominance of deciduous arcto-tertiary elements (Populus populina, P. palsamoides—Salix—Ulmus, MF3 megafloral zone), the Navicula pinnata D6 diatom zone, NN6—7, N12, CPN9 zones, and the terrestrial vertebrate zone MN6—7 at the base furthermore by the contemporaneous predominance of marine vertebrates Charcharodon—Myliobatis.

*2.7* In the *Sarmatian* stage, zone PN9 is characterized by the first appearance of Tsugaepollenites helenensis, by the exclusiveness of Manikinipollis tetradoides and Echinotisporites cserhátensis and the extinction of E. longechinus. The megafloral zone Zelkova zelkovaefolia—Quercus kubinyii—Lauraceae MF4 and the D7 diatom zone Anaulus simplex testify to the coming into prominence of thermophilous species of Near East—Mediterranean affinitinity. Along with the zones NN6—7, N13—14 and CPN10 that can be identified with some difficulty, additional contributions to the stratigraphic assignation of the Sarmatian were provided by the radiometric age of the third rhyolite tuff ejection (Galgavölgy Rhyolite Tuff Formation, 13.7±0.8 Ma).

**Proposals**

a) It is timely and justified to revise and reformulate the notions Lower, Middle and Upper Miocene that are now being used according to different interpretations.

b) Spanning the interval between 24—5.5 Ma, the Miocene can be subdivided into 3 superstages: Lower Miocene, Middle Miocene and Upper Miocene.

c) On the basis of the regional stratigraphic nomenclature of the Central Paratethys the Lower Miocene spans the Eggenburgian—Ottnangian interval, the Middle Miocene the Karpatian—Lower Badenian—Middle Badenian interval, the Upper Miocene the Upper Badenian—Sarmatian—Pannonian interval. The authors propose to use the term Atlantian for the new Lower Miocene superstage, the Mediterranean for the new Middle Miocene superstage and the Caspian for the new Upper Miocene superstage. This proposal is motivated by the data presented in Fig. 1 and by quantitative biostratigraphic evidence.

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# NEW CRITERIA FOR THE CORRELATION OF THE ANDALUSIAN AND MESSINIAN STAGES

by

F. J. SIERRO, J. A. FLORES, J. CIVIS and J. A. GONZALES DELGADO

*Introduction.* In the history of geologic research, the Mediterranean was one of the first areas to be studied. Indeed, most of the Neogene stratotypes have been defined in this region and this has been of decisive importance in some of the great problems related to stratigraphic correlations.

Recent studies in biostratigraphy are now reaching a high degree of resolution which permits a very exact correlation, above all in the Neogene and Quaternary. Such works are based on the qualitative and quantitative analysis of different microfossil groups of wide geographical distribution. This permit the definition of events related with important environmental changes.

The characteristics of the Mediterranean are typical of those of a marginal sea, strongly influenced by local factors (tectonic activity), which may mask the overall events observable in the oceans.

This situation is particularly highlighted in the Messinian stratotype (Pasquasia—Capodarso series) proposed by SELLI (1960) to characterize an interval of time during which specially critical conditions prevailed throughout the Mediterranean and Paratethys. In this sense, the Messinian would not be a stage but rather a facies.

In the light of the peculiar characteristics of the sediments of the Upper Miocene throughout this domain, different authors have proposed the creation of new stratotypes outside the Mediterranean. Along such lines, PERCONIG (1964, publ. 1966) proposed the Andalusian, situated in the Guadalquivir basin, as a new stratotype to substitute the Messinian. According to this author, in the Guadalquivir basin there is a continuous marine sedimentation from the Tortonian to the Pliocene which converts this area in the most suitable one for the definition of a Unit equivalent to the Messinian in a marine facies.

Numerous studies carried out at the Dept. of Palaeontology at the University of Salamanca have provided new data which offer a better vision of the Messinian/Andalusian relationship and their boundaries.

## The stratotypes

*Messinian stratotype.* D'ONOFRIO et al. (1975) situate the Tortonian/Messinian boundary approximately 80 m below that defined by SELLI (1971) and suggest that it coincides with the first appearance of *Globorotalia conomiozea.* This view has received wide acceptance. However, its use outside the Mediterranean is somewhat questionable. Indeed, this taxon is common in the S Atlantic, the Pacific (some authors such as SCOTT, 1980 believe that it is not the same species) and in the northernmost part of the Atlantic, but is absent or only appears sporadically in the Atlantic adjacent to the Mediterranean. For other authors, the FAD of *G. conomiozea* in the Pacific and S Atlantic is evolutionary, whereas in the Mediterranean it is migratory (LANGEREIS, ZACHARIASSE and ZIGDERVELD, 1984) such that both events would not be synchronous. This poses certain problems in the correlation of this boundary, though its definition is valid.

A more difficult question is the Miocene/Pliocene boundary which is recognizable only in the Mediterranean at present and is impossible to establish in any other area since it coincides with an abrupt change of facies, restricted to this domain. Although it is possible that the event defining the boundary in the Mediterranean might be linked to an overall environmental change, it has still not been possible to confirm such an idea. However, the possibility should not be ruled out that this crisis might have been caused by local phenomena.

*The Andalusian Stratotype.* Since PERCONIG (op. cit.) proposed the new stratotype, considerable controversy has arisen concerning its validity. An idea of this can be gained from the large volume of publications dealing with the topic.

The base of the Andalusian was first situated in the Carmona section and then in the section of "Arroyo Galapagar". PERCONIG (1971, publ. 1974) suggested that the Tortonian/Andalusian boundary coincides with the first appearance of the "ancestral forms" of *Globorotalia margaritae*, but this event is not a good indicator. Accordingly, this boundary has always been very imprecise. A similar situation occurs with the Mio/Pliocene boundary, situated according to PERCONIG at the top of the "caliza tosca"; this author suggests that this material might be equivalent to the Mediterranean evaporitic deposits, whereas the green marls would constitute the base of the Pliocene transgression. Nevertheless, this has not been confirmed and this limit is also very doubtful. These are the main problems which have hindered the Andalusian from being widely accepted.

Recent works carried out by this team (SIERRO, 1984; 1985; SIERRO et al., submitted for publication) in the Guadalquivir basin and other transition areas between the Atlantic and Mediterranean have underlined the importance of the keeled Globorotaliids in the biostratigraphy of the Upper Miocene of the area. In this sense, three assemblages of keeled Globorotaliids have been described which are temporally successive. This phenomenon has been observed in a large number of stratigraphic sections: the Gibraleón, the Beas-Trigueros, the Guillena, the Cantillana and the Arroyo Galapagar sections, all of them in the Guadalquivir basin; the Oued Akrech section (in the South Riff basin, Morocco) and in the Arejos section of the Sorbas basin (Almeria, SE Spain) as well as at some DSDP sites in the NE Atlantic.

From the taxonomic point of view three groups of keeled Globorotaliids may be distinguished: the *G. miotumida* group and groups I and II of "*G. menardii*".

The *G. miotumida* group (sinistral) is composed almost exclusively of this species, though occasionally individuals of *G. conomiozea* may appear. The "*G. menardii*" group II (dextral) has previously been described by a member of this team and includes specimens with a similar morphology to those of *G. cultrata*, *G. merotumida* and *G. plesiotumida*. Group I of this taxon is easily distinguishable from the previous group both stratigraphically and with respect to its sinistral coiling; furthermore, in this latter group specimens with a tighter spire are predominant.

The sequencing of these assemblages defines three events:

*1* The massive disappearance of group I of "*G. menardii*".

*2* The massive appearance of group II of "*G. menardii*", after a short interval in which the keeled Globorotaliids are practically absent, with a predominance of the group *G. scitula*.

*3* The substitution of group II of "*G. menardii*" by that of *G. miotumida* (Fig. 1).

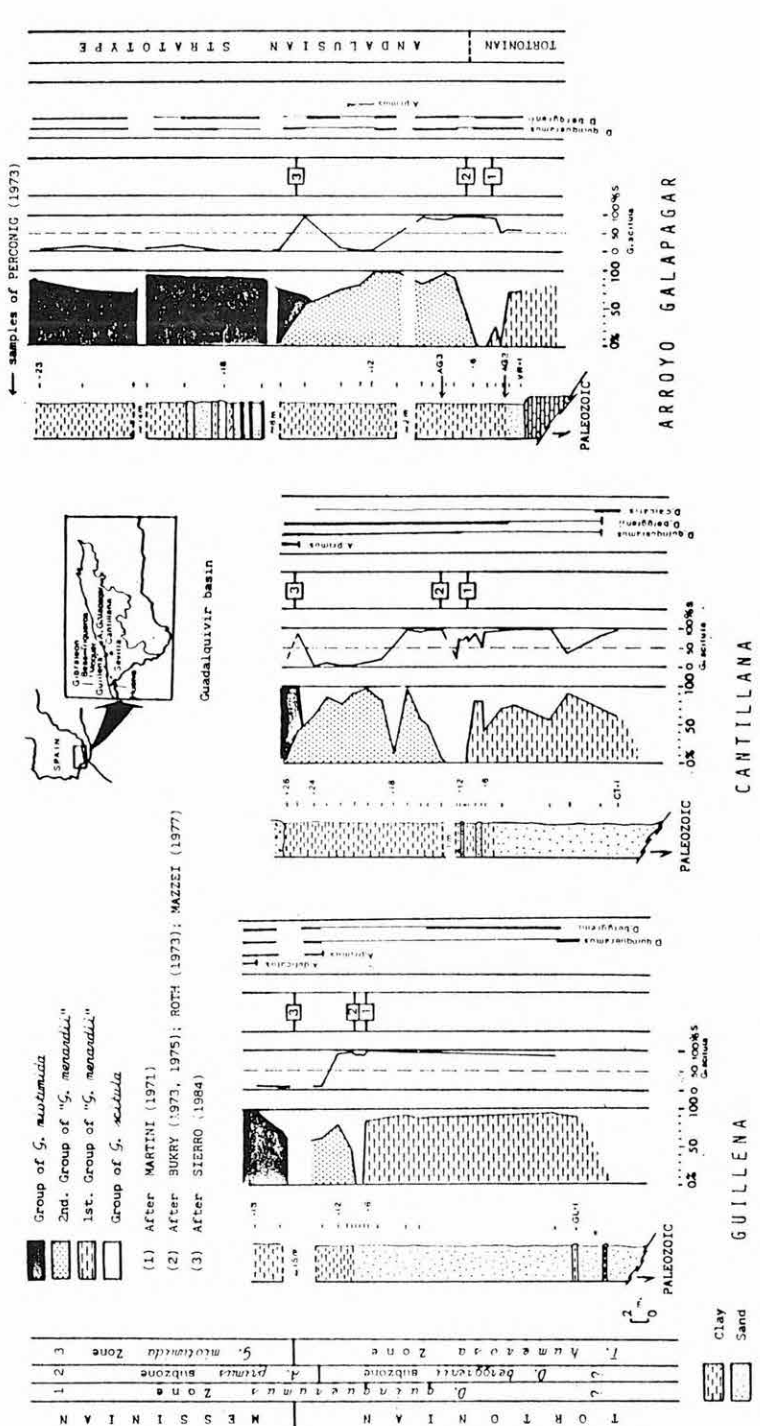


*Fig. 1.* Variation in the assemblages of keeled Globorotaliids in 3 sections of the Guadalquivir basin

At present the data are insufficient to support definitive conclusions regarding the variation in the coiling direction in the *G. scitula* group, though it has been observed that prior to Event 1 the group usually exhibits a preferential sinistral coiling which continues to a horizon slightly above event 2. Within this horizon a change may be observed from sinistral to dextral; this was observed both in the three sections presented and in other sections situated more to the west in the Basin.

The temporal sequence of the three assemblages reported here is similar to that described by ZACHARIASSE (1975, 1979), BOSSIO et al. (1976) and others, both in the Mediterranean and Atlantic realms.

In a study carried out by one member of this team (SIERRO, 1985) it is pointed out that the replacement horizon of the "*G. menardii*" group by that of *G. miotumida* (= *G. conomiozea* group of the authors working in the Mediterranean) coincides with the T/M boundary. To do so, we addressed the data of ZACHARIASSE (1975, 1979) in Crete and in the Messinian stratotype.

This very important change affects all the populations of *Globorotalia* in the NE Atlantic and the Mediterranean since in the Tortonian the "*G. menardii*" group inhabited the whole of this domain, whereas in the Messinian, the populations of *G. miotumida* displaced those of the "*G. menardii*" group which were restricted to the lower latitudes of the Atlantic. During this period, the whole of the NE Atlantic and the Mediterranean was populated by the *G. miotumida* group (SIERRO et al. submitted for publication).

This succession of assemblages of keeled Globorotaliids is very useful to correlate the Messinian and Andalusian stratotypes. In the Arroyo Galapagar section we have noted the succession of the three groups of keeled Globorotaliids mentioned above.

Event 3 is situated between samples VR15 and 16 such that we have reached the conclusion that the base of the Andalusian stratotype defined by PERCONIG is situated some 18 m below the T/M boundary (Fig. 1).

According to the studies of FLORES (1985) on the calcareous nannoplankton of this section, the previously described T/M boundary would be situated slightly above the first record of *Amaurolithus* and before the point at which a remission in the record of *Eudiscoaster berggrenii* and *Eudiscoaster quinqueramus* is observed. In a large number of sections its location can be pinpointed between the first record of *Amaurolithus primus* and *Amaurolithus delicatus*. Unfortunately, the conditioning of these ceratoliths by the fluctuating bathymetry has meant that their record is not as continuous as would be desirable, thus accounting for the imprecision.

These data permit, however, to establish an approximate equivalence of this interval with that defined in the zones or subzones of *Discoaster quinqueramus* (MARTINI, 1971; BUKRY, 1973, 1975) and *A. primus* (BUKRY, 1973; 1975; ROTH, 1973; 1974).

With respect to the upper part of the stratotype, defined in the Carmona section (Fig. 2), this must be situated above a fourth event which coincides with the sinistral to dextral coiling change in the group of *Neogloboquadrina acostaensis* observed in several sections of the Guadalquivir basin. It should be remembered that this event predates the Mediterranean salinity crisis and is recognizable in other Atlantic areas. *Globorotalia margaritae* s.s. appears (event 5) at some 30 m below the "caliza tosca". Accordingly, we believe that a large part of the Carmona marls and possibly the "caliza tosca" may be correlated with the Mediterranean salinity crisis.

In the upper green marls, *G. margaritae* but not *G. puncticulata* was found (the FAD of the latter constitutes event 6), such that it is possible to infer that these beds

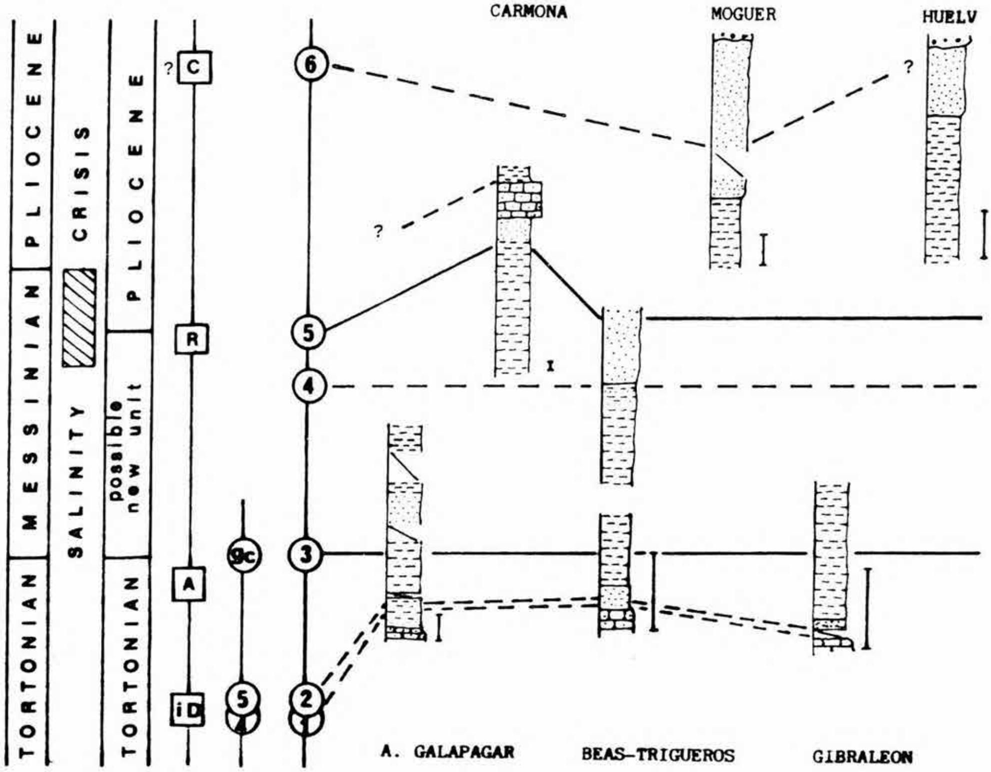


*Fig. 2.* Location of some stratigraphic sections of the Guadalquivir basin with respect to the events discussed in the text

The vertical line appearing next to each section represents 10 m; iD = dominance of the "Small placoliths" over the *R. haqii R. minutula* group; A = first record of *Amaurolithus;* R = reduction of *E. quinqueramus/E. berggrenii*; C = first record of *C. rugosus* (not observed in the Guadalquivir basin); 4, 5, gc = events of ZACHARIASSE in LANGEREIS et al. 1984

should be assigned to the latest Messinian or basal Pliocene. The absence of this latter taxon is very significant in view of the fact that in nearby sections it is relatively common.

In this section, the use of calcareous nannoplankton in biostratigraphy presents certain difficulties when litorality becomes more pronounced since this factor affects in a special way the marker taxons: Ceratoliths and Asteroliths. It is possible, however to detect certain events more or less correlatable with those defined using Foraminifera. In this sense, event 5 is more or less equivalent to the event defined by FLORES (1985) as "a reduction in the record of *E. berggrenii* and *E. quinqueramus*". It should be noted that such a "reduction" does not seem to be related to the beginning of carbonate sedimentation ("caliza tosca"), although the low number of Asteroliths may reflect similar results.

From this point onwards, the assemblage of calcareous nannoplankton becomes poorer, both in diversity and in abundance, with no evidence of the beginning of the record of Ceratoliths which the abovementioned author reports in nearby oceanic cores. Studies carried out by other authors (RIO et al., 1976; VISMARA SCHILLING and

STRADNER, 1977; RAFFI and RIO, 1979; MAZZEI et al. 1979) coincide, however, in that the appearance of *G. puncticulata* took place at the same time as that of *Ceratolithus rugosus* (shortly after the first record of *C. acutus*).

Thus, according to the data obtained with planktonic foraminifera and calcareous nannoplankton, the M/P boundary cannot be established with precision in the area according to its present definition.

### Importance of the Guadalquivir basin in studies on the Upper Miocene

The numerous studies conducted in this area in recent years highlight the importance of the region for the study of the Atlantic/Mediterranean relationship since it is situated between both.

This team has performed numerous studies in the area which have evidenced changes in the microfauna and microflora associations susceptible to being correlated on a large scale.

According to the above mentioned data it is clear that the Messinian stratotype cannot continue being used as a reference. There are two solutions to the problem:

*a)* To define a chronostratigraphically equivalent unit. In this sense, there is no better area than the Guadalquivir basin or the S. Riff basin, close to the Mediterranean, which were unaffected by the salinity crisis.

*b)* To define a new chronostratigraphic unit, maintaining the lower boundary and modifying the upper one to make it coincide with an important event which can be correlated on a global scale.

From such possibilities it may be deduced that the nexus of the problem is to define in a precise fashion a stratotype for the M/P boundary in an area of normal marine conditions. With respect to the first possibility, we feel that both in the Guadalquivir basin and in the S Riff basin this event could be recognized by conducting a more thorough mutidisciplinary study, although we are unaware of whether this event could be correlated on a large scale outside the Mediterranean.

With respect to the second possibility, the M/P boundary could be made to coincide with the first appearance of *G. margaritae* s.s. which took place in the latest Messinian. This event could be compared with that defined by the reduction in the representation of *E. berggrenii* and *E. quinqueramus* or with the first appearance of *Ceratolithus*. This boundary could be recognized over large areas with the exception of the Mediterranean.

According to this viewpoint, the definition of a new unit in the Guadalquivir basin is currently feasible, though no section is presently known which covers the whole interval owing to the low dip of the strata, such that this unit should composed of several partial sections.

The base and the top of the new unit could be located in the Arroyo Galapagar—Carmona series (Fig. 2); the latter of these would be the most suitable, according to the data currently available, for definition of the stratotype of the new M/P boundary. However, this series is composed of two sections, and between the two there is a distance of 30 km covered by Quaternary sediments which mask part of the series. Event 4 would be recorded in this obscured stretch.

Another suitable area would be between Sevilla and Huelva. In particular the Beas-Trigueros series, composed of two close sections, includes the whole of the interval under consideration. In these sections events 3, 4 and 5 are recorded (Fig. 2). The only unsuitable feature is the shallow nature of the silty—sandy sediments of the upper part.

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# DELIMITATION AND CORRELATION OF THE PONTIAN AND THE MESSINIAN STAGES ON THE BASIS OF MALACOFAUNA

by

P. M. STEVANOVIĆ

*Introduction.* In the present review of some of the latest attempts to correlate the Messinian in the Mediterranean province and the Pontian of the Paratethys, reference is made foremostly to the published papers from the 7th RCMNS Congress in Athens and the rich literature on the Pontian used in preparing the book "The Pontian" to be edited and published in Yugoslavia in the book series "Chronostratigraphie und Neostratotypen in der zentralen Paratethys".

My speculations are based on the results of general malacofauna considerations and my field investigations in various regions of the Paratethys in Greece and Italy.

In a paper recently published under the title "Die Äquivalente des Tortoniano und Messiniano in der zentralen Paratethys", it is claimed that after the Badenian, stages from the Mediterranean can be correlated with those from the Paratethys only "mit Hilfe radiometrischen Daten" or only of mammals (PAPP and STEININGER, 1979), thus excluding other possible correlations.

While certain authors, like J. SENEŠ, maintain the "equivalence de l'âge du Messinien et du Pontien supérieur" (SENEŠ, 1981, p. 54), another group of palaeomalacologists have a different approach to the Messinian/Pontian correlation, correcting their concept with the time (GILLET, 1958, 1963, 1969; SELLI, 1960, 1973; PAPP, STEININGER and GEORGIADES-DIKEOULIA, 1978; BALESIO, ARCHAMBAULT-GUÈZOU, 1980; STEVANOVIĆ, 1963, 1966; ARCHAMBAULT-GUÈZOU, ILINA et al., 1979; PAPP, 1980; ILINA, 1980; RÖGL and STEININGER, 1983; and others).

Approaches of other specialist to the Messinian/Pontian relationship such as of mammalogists, ostracodologists, micropalaeontologists, palynologists,geochemists, geophysicists, etc are not considered in the present paper.

Among the researchers various approaches are noted or rather variants in approaches and estimates of the possible Messinian/Pontian correlation. Some of the attempts are mentioned below.

Messinian synchronous with Pontian
Messinian synchronous with Upper Pontian
Messinian synchronous with Upper Pontian + Lower Kimmerian (= Lower Dacian)
Messinian synchronous with Maeotian + Lower Pontian
Messinian synchronous with Lower Pontian (possibly the whole Pontian)
Messinian synchronous with Chersonian + Maeotian + Lower Pontian.

There is an extreme opinion that the Messinian is synchronous with the Middle and Upper Sarmatian, Maeotian and Pontian (GILLET, 1959).

Supporters of the "Messinian synchronous with Upper Pontian" are vague or are not mentioning their understanding of the Upper Pontian. No reference is made to its biostratigraphy, malacofauna, climatic—palaeographic circumstances of the time, etc. The Upper Pontian in the western ("central") Paratethys is known, however, to be a more general term (STEVANOVIĆ, 1951) than in the Dacian or euxinian Paratethys where it is represented only by Bosporian stage of vaguely marked boundaries (ANDRUSOV, 1917; 1923).

By its principal properties (facies, fauna, sediment types) the Messinian is rather a geological formation than a stage. Its lithological—faunal aspect is highly variable: Lower Messinian consists of laminite deposits, the middle of evaporites with laminites and the upper of "Lago mare facies", while also coral and algal reefs occur on western and eastern Mediterranean margins. If it was not separated as a "stage" in the last century, it would be much easier to compare, correlate with relative sediments in the eastern Paratethys.

Similar difficulties in geological correlations of deposits from two or more separate basins are rarely faced in the stratigraphy. All this is a consequence of great diversity in Messinian sediments from the base upward to the top. Another difficulty in correlating with the Messinian is inadequate knowledge of Mediterranean geologists of Neogene deposits in the Paratethys or the knowledge of central and southeastern European geologists of formations in the Mediterranean.

Further consideration will cover the following:

*1* biostratigraphic boundaries of the Pontian stage;

*2* are Pontian and Pannonian in the western ("central", by SENEŠ) Paratethys synchronous or not?;

*3* relation of the Messinian from the Mediterranean (Italy, etc) to the Pontian in Greece and eastern Paratethys;

*4* what biocorrelation is possible between Messinian and Pontian?;

*5* relation of the Tortonian from the Mediterranean basin to Maeotian and Pontian stages.

When correlating the Messinian with the Upper Miocene or Lower Pliocene in the Paratethys first must be answered the principal question:

Has it been proved in any published work that evaporites in Mediterranean basin lie above argillaceous-marl sediments with Lago mare malacofauna, i.e. over an undoubted correlate with the Lower Pontian (Novorossian substage)? If evaporites are nowhere lying so they must be together with laminites as Lower and Middle Messinian, older than the Lower Pontian, i.e. be equivalent in geological age to the Chersonian and the Maeotian stages. This even more so because no discordance has been proved between the evaporite complex (Middle Messinian) and "Lago mare facies".

**1** *Delimitation of the Pontian*

*Lower boundary* in the Carpathian province or the Pannonian basin is considered by P. STEVANOVIĆ (1951, 1977) and in the Proceedings under the title "The Pannonian" (1985, published by the Hungarian Academy of Sciences) as a boundary between the Pannonian and Pontian stages. The boundary in the Vienna basin is between the "subglobosa zone" ("E" zone, in PAPP, 1953) and the "zahalkai zone" ("F" zone, also in PAPP). Related to malacofauna this boundary in the south of the Pannonian basin lies between "subglobosa zone" (Upper Pannonian) and "Praerhomboidea zone" (Lower Pontian) or "upper Abichi beds".

The Lower Pontian boundary in Dacian and Euxinian basins is marked by first occurrences of Limnocardids, Pseudoprosodacna and Pseudocatillus in the so-called Eupatoria-horizon.

*The Upper Pontian boundary* in Carpatho—Pannonian basin is characterized by the disappearance of last Congeria, Limnocardium and Valenciennius.

This boundary is not sufficiently marked in Dacian and Euxinian basins; the top of the Bosporian substage (ANDRUSOV, 1917) marks the end of the Pontian stage. This substage, according to some authors (SENEŠ, 1981), should be synchronous with the whole of the Messinian. Other authors (TAKTAKISHVILI, NEVESSKAIA, and STEVANOVIĆ, 1985) take that the Bosporian substage is virtually inseparable from the Portaferrian. At that time Transcarpathian connections between Pannonian and Dacian basins and between Black Sea (Euxinian basin) and Caspian basin were still existent; in other words, there was a single Paratethys from the Alps to the Aral sea in Asia.

### 2 *Were Pontian and Pannonian in the western ("central") Paratethys synchronous or not?*

An answer to the above question is partly given in the mentioned Proceedings "The Pannonian" (PAPP edits., 1985,Budapest). My discussion on the Pannonian, between the Sarmatian and the Pontian, with JEKELIUS (1943) and PEVZNER and WANGENGEIM (1982), will not be presently referred to (it will be published elsewhere); only an unsustainable passage will be quoted from the disputable questions of the volume and stratigraphic position of the Pannonian by PEVZNER and WAGENGEIM, which reads: "Thus, in our opinion the whole of the Pannonian (= Congeria beds) in Hungary can be correlated with the Pontian" (see p. 50).

The mentioned authors are of the opinion that a regression i.e. dry land existed in the Carpatho—Balkan basin between Pannonian zone with Congeria ornithopsis and three younger zones (C, D, E) of the same stage. This opinion fully agrees with that of R. HOERNES (1900) and his follower JEKELIUS (1943), though it was generally abandoned long ago after deep borings and instructive surface sections. The opinion about interrupted sedimentation and retreat of the Pannonian sea has been contradicted by the long established evolution of Mollusca (Dreissenids, Limneids, Limnocardids) through the Pannonian and the Pontian from the level with Congeria ornithopsis (the oldest Pannonian) to the end of the Pontian level with Congeria rhomboidea and Prosodacna vutskitsi.

### 3 *Relationship of the Messinian from the Mediterranean (Italy) to the Pontian in Greece, Turkey, and Eastern Paratethys*

The Messinian is divided, as mentioned above, into three levels: (1) praevaporites and laminites, (2) evaporites and (3) postevaporites or into Lower, Middle and Upper Messinian. Postevaporites or Lago mare facies are known (GILLET, 1962, 1969; SELLI, 1960, 1973; and others) to be of the greatest importance for defining the Messinian to Pontian relationship. Both in past and the present time, many palaeomalacologists have inferred that soon after evaporite deposition, in early Pontian, an association of molluscs, which had common properties with molluscs of Eupatoria beds of the Lower Pontian in the Paratethys, entered the Mediterranean basin and its periphery or partly as autochtone developed in it. It was discovered in a peripheral area of the Mediterranean from Spain in the west trough southern France and Italy to Greece and Near East. Similarly, Ostracods in this region were brackish (Caspibrackish fauna).

Mollusca from the Upper Messinian, i.e. Lago mare facies, have been studied by many authors in Italy, Spain, Greece, Turkey, southern France and certain common species were identified for the Lower Pontian of the Paratethys and Upper Messinian of the Mediterranean (GILLET, 1960, 1961, 1965, 1969; ARCHAMBAULT-GUÈZOU et al.,

1976, 1979; STEVANOVIĆ, 1963, 1964, 1966). Particularly notable among Limnocardids in both basins are small Pseudoprosodacna (P. litoralis), Paradacna (P. abichi), Euxinicardium (E. subodessae), etc., all forms known and first described from the Lower Pontian in the Paratethys. These originally euxine species are jointed with some endemic Limnocardid species such as Didacna and Pontalmyra (GILLET, 1961, 1965).

"Lago mare fauna" of this "Italian type" is also found in the Lower Pontian of the Aegean region, e.g. southeast of Thessaloniki, in Alatini suburb (STEVANOVIĆ, 1963) and near the village of Trilophos (GILLET et FAUGÈRE, 1970; GILLET et GEISSERT, 1971; PAPP, 1980). Moreover Lower Pontian with this species is known from Tracones Limestones south of Athens on Attican peninsula (GILLET, 1937, 1957; STEVANOVIĆ, 1963, 1964; PAPP, STEININGER and GEORGIADES-DIKEOULIA, 1978). Related to the present molluscan species typical of the Eupatoria horizon in the Euxine basin, all of the mentioned authors describe the three mentioned localities to the Lower Pontian. The Workshop for the Paratethys, assembled in Sofia (1978), consulted specialist in Pontian of the eastern Paratethys, at a proposal by A. PAPP, asking them to confirm Lower Pontian age of sediments near Thessaloniki, which they did by having agreed with the earlier determinations (see PAPP, 1980).

Simple comparison of Upper Messinian malacofauna from Italy, e.g. vicinity of Ancona (GILLET, 1969)*, and CAPELLINI's collections from Toscana (GILLET, 1962) with fauna from Lower Pontian of the Aegean basin, Greece, readily shows the common caspi-brackish character of the association in addition to certain local, autochthonous elements in both regions. This type of Upper Messinian malacofauna closely related to the Lower Pontian euxine type in the Paratethys, can also be identified in regions of the Mediterranean Sea other than Italy and Greece as mentioned above.

In the light of these facts, the opinions: "Therefore, beds in Trilophos and Trakones were older than the Messinian", or: "...Geological age of these deposits should be determined as the Upper Tortonian" (PAPP, 1980, p. 243), about the geological age of the mentioned localities in Greece can hardly be accepted because they are not based on required malacological data but only on time determination by physical or similar methods.

### 4 *What correlation between the Messinian and the Pontian is possible?*

As there is not any discordance between evaporites and postevaporites of Lago mare facies, the deposition of evaporites must be geologically older than the Lower Pontian, equivalent to Maeotian and Chersonian (or Pannonian). Both Maeotian and Chersonian are characterized by great regression of the Paratethys, same as the Lower and Middle Messinian in the Mediterranean. The marine connection between western and eastern Paratethys provinces ceased to exist at that time; the connection between the eastern province and the Mediterranean simultaneously broke in the Chersonian and existed at intervals in the Maeotian; during the deposition of laminites and marginal Messinian reefs, occasional communication with the Paratethys may have existed. Significant reduction in molluscan species in deep sea of eastern Paratethys during the Chersonian in particular, corroborates its complete isolation both from the

* A close examination of Lago mare fauna from Ancona area (GILLET, 1969) may reveal occasional Pannonian (s. str.) or prae-Pontian forms of Limnocardids, e.g. ex gr. Limnocardium cekuši, L. veselinovići, L. viquesneli.

Mediterranean and Pannonian provinces of the Paratethys, as has been documented by the absence of common species in these three basins.

Correlation of the Lower and Middle Messinian or evaporite formation and Lower and Upper Pontian is not possible from a climatic aspect as it can be done using malacofauna. The Lower Pontian is characterized by relatively lower temperature than that of the immediately older Maeotian; there are opinions that Ukrainian rivers of the time froze in the winter*. The Upper Pontian (Portaferrian and Bosporian substages) as directly younger, on the other hand, is characterized by clearly warm and humide climate especially in Dacian and Pannonian basins of the Paratethys. In other words, in the Pontian stage, both the Lower (Novorossian substage) and the Upper (Portaferrian and Bosporian substages), the conditions were not suitable for sedimentation of evaporites: the older Pontian because of low temperature and humidity, and the younger, with increased temperature but even more humid climate.

Dry and warm epoch in the younger Neogene, suitable for homogene sediments, is represented in the western Paratethys by the younger Sarmatian, upper Bessarabian and Pannonian. At that time enormous masses of semilagoonal marl sediments, known as "white marlstone formation", were deposited in the south of the Pannonian basin. They are accompanied by flaglike limestones poor in malacofauna but rich in fish remains and continental flora. This formation in Yugoslavia is including "white marls" from Zagreb area (Radoboj locality with sulphur occurrence, "praepontic" marls in the same area, GORJANOVIĆ-KRAMBERGER, 1900). In various areas of Yugoslavia it is known as: "Radix beds", "Provalenciennesia beds" (MOOS, 1944; OŽEGOVIĆ, 1944; JENKO, 1944) in Croatia; "Banatica beds" in Slavonia, "Beočin marls" in Srem, Belgrade area, Roumanian Banat and Transylvania, etc.

Middle and Upper Pontian (Portaferrian) is marked by warm and very humid climate along the southern Pannonian margin, which, according to the above authors, should be synchronous with evaporites in the Mediterranean. Thus on the territory of Yugoslavia, from Krško in the west (Slovenija) to the Carpathians in the east (eastern Serbia), the Upper Pontian (Portaferrian) typically has thick deposits of brown coal (lignite)—an evidence of humid climate. This is used by national palaeobotanists (PANTIĆ, 1956, 1957) to describe warm and very humid climate in the Upper Pontian. The fact is that there are also different opinions (HSÜ et al., 1977; PAPP and STEININGER, 1980) about the climate in northern areas of the Pannonian basin.

Pontian coal basin at Krško (Globoko) in Slovenia and the nearest Messinian evaporites in Marche province, Italy, are approximately situated at the same parallel, some 300 kilometres apart in air line. When coal was deposited in Krško (Globoko), the Dinarides, according to palaeobotanic data, were by about 800 to 1000 metres lower than at the present time; similarly was with the Alps. The question arises: are two extreme types, one very arid, suitable for evaporite deposition, and the other, very humid, providing conditions for coal deposition, conceivable in one geological time at one geographic parallel of the north latitude over a relatively short distance?

Similar examples of allegedly simultaneous deposition of climatologically extreme types of sediments at a short distance can be given for various parts of Balkan peninsula, e.g. southern Serbia, Macedonia, Greece, Turkey. One example is the Pontian coal deposits 50 metres thick in Kosovo coal basin and Messinian gypsum, anhydrite, only 150 kilometres away, near Tirana. These could not be simultaneous from the climatological aspect. Given examples therefore contradict the mentioned understanding of correlation.

* Recent palaeofloral examinations in the Lower Pontian of southern Ukraina contradict this opinion (see SULADZE, 1984).

Above considerations of the correlation based on malacofauna and continental flora (including palaeoclimatic factor) oppose the published opinions that the Messinian was synchronous with the Upper Pontian and Lower Kimmerian (SENEŠ, 1981; SEMENENKO, 1979). On the contrary, a different correlation is possible that would suit both the nature of malacofauna in the Mediterranean and the Paratethys and the nature of fossil continental vegetation, including in consideration also palaeogeographic—climatic circumstances in the two regions during the Messinian, viz.:

| Tethys | | Paratethys |
|---|---|---|
| U. Messinian (Postevaporites) | | — L. Pontian (Novorossian and Odessian) |
| M. Messinian (Evaporites)<br>L. Messinian (Preevaporites) | gessoso-<br>-solfiferra<br>formazzione | — Maeotian<br>— Chersonian |

**5** *Relations of the Tortonian from the Mediterranean to the Chersonian, Maeotian and Pontian from the Paratethys*

If correlation, as proposed in the preceding chapter using malacofauna, climate, etc could be accepted, i.e. that Messinian of the Mediterranean is synchronous with the Chersonian, Maeotian and Lower Pontian then the table of stratigraphic correlation in RÖGL and STEININGER (1983) must be revised. Three stages mentioned in the above subtitle are evidently younger than the Tortonian in Italy, underlying the Messinian. Tortonian as one of Mediterranean Neogene stages cannot be correlated with the Lower Pontian of the Paratethys only the Upper Messinian can.

*Conclusion.* Considering the question of correlation of Neogene stages at Miocene/Pliocene boundary, the proposal in the 6th RCMNS Congress resolution may be closer to the right answer from the malacofauna and continental vegetation points of view. It related the Messinian to the Maeotian and the Lower Pontian. If left at that, we would not discuss today the correlation of the Tortonian to the Lower Pontian, which is absurd from the viewpoint on which the present work is based. Two stages, which are certainly in superposition in the Mediterranean, are placed in one segment of time!

Synchronisation of Messinian—Pontian will be further considered because there are various formulations of the proposals based on radiometric data*, nanoplanktons and the like. It is not a final, generally acceptable solution that is negotitated for; nothing the weakness of the arguments presented by certain authors and some advantages of the methods used, the attempt is made in this review to eliminate the doubts in the validity of using malacofauna in younger Neogene stratigraphy. Comparative stratigraphic tables published recently are known to completely neglect the division into stages, horizons and biozones derived using malacofauna. When introducing new correlation methods, however, the tested methods should not be neglected which were successful in providing data when the first "classical" division of the Neogene was made for the Tethys and the Paratethys.

* In a handwritten paper for the Proceedings "The Pontian", which is to be edited by me and published, TRUBIKHIN is concluding as follows: "As Messinian deposits cover the sixth and fifth palaeomagnetic epochs and the lowermost Gilbert epoch, the whole of the Pontian or at least most of the Lower Pontian is equivalent to the Upper Messinian" (1985, manuscript). If so, the older Messinian (evaporites and laminites) must certainly be Maeotian or Praepontian. Evidently, neither results acquired by physical methods are identical with all authors.

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# PALAEOTHERMIC EVOLUTION DURING THE NEOGENE IN MEDITERRANEA THROUGH THE MARINE MEGAFAUNA

by

G. DEMARCQ

The data collected from continental floras and faunas are not alone in illustrating climatic variations of a period. Using significant groups of marine megafauna, it is possible to estimate the relative evolution of the temperature in marine waters. The benthic forms of the littoral margins (basins, gulfs, shelfs, archipelagos) live in shallow water and give good information on thermal variation. Furthermore, the Neogene period relatively recent, allows comparative analysis with present taxa (family, genus) living in the warm bioprovinces.

From this data, it is possible to design some major palaeoecologic events in Mediterranea during the Neogene. The implications of these results concern also the paleogeographic reconstructions and the geostructural evolution of the margins, into the Tethys area, but more generally with atlantic and indopacific influences. At once, it is necessary to precise that the medium temperature of the sea-waters from Aquitanian to Middle Pliocene is more hot that at present. There is no comparison; the ecological methods of the Actuology must be considered on the bases of this fact: the actual period is abnormal. There is mega- or mesotherm climax in the megafaunal assemblages in the Tertiary. But in the detail, it is possible to give a very clear paleothermic evolution with some etaps of variation.

*Where are the significant groups?*

In the Madrepora, with hermatypic genera or species, it is possible to give relative and quantitative indications ($t^\circ > 23\ ^\circ C$); the quality of the associations allows slight differences. For example, the proportion of hermatypic species all along the time is more significant. Important during aquitanian and burdigalian ages—recifal constructions with Poritidae and Faviidae in the Lower Miocene of the south of Rhône valley—their number fall mostly after those times.

The Gasteropoda contain many families with megatherm elements (Apporrhaiidae, Strombidae, Eratoidae, Cypraeidae, Cymatiidae, Pirulidae, Muricidae, Pyrenidae, Galeodidae, Fasciolariidae, Olividae, Mitridae, Cancellariidae, Turridae and Conidae). In the Bivalva, the Pectinidae, among others, give precious informations on certain species of southeastern bioprovinces: some megatherm Pecten (e.g. *P. pharaoni*, *P. cristato-cristatus*, *P. ziziniae*, *P. pseudobeudanti*, *P. concavus*, *P. fraasi*). The genus Flabellipecten in its ensemble is "warm" (e.g.: *Fl. schweinfurthi*, *Fl. expansus*). Others subgenera are mega or polytherm; in the group of "Macrochlamys" "*Chl.* gr. *albina*, *Chl. gigas*, *Chl. latissima*, *Chl.* gr. *solarium)* and some species of Chlamys *(Chl. zitteli*, *Chl. sub-malvinae*, *Chl. zenonis*, *Chl. scissa*, *Chl. lilli)*. But generally in Mollusca some nordic "cold" migrants indicate the end of the Pliocene. The Bryozoa have some genus (e.g.: Steginoporella) and species characteristics of an intertropical environment, like the Echinodermata (e.g.: Clypeaster, Histocidaris, Scutella). Balanidae and Terebratulinae are also megatherm markers with large species as *Megabalanus* gr. *tintinabulum*, *Terebratula grandis*... In some groups, giant forms of differents taxa are common, especially during the Lower Miocene and the Messinian. It is the same for big calciferous tests in typical species or morpha (e.g.: Crassostrea, Spondylus, Clypeaster), associated with alga Melobesiae (e.g.: Lithothamium), Serpulidae, Vermetidae, etc...

*What is the method?* (DEMARCQ, 1984a)

Some taxa (families, genera, species) characterize marine bioprovinces: subtropicals, tropicals, intertropical. For every period they are found in homogeneous latitudinal zones; they advance rather far to the north with time, indicating relatively hot periods. For every area in the Mediterranea their number and frequency vary, according to the temperature of the littoral waters. Their actuel refuge in the equatorial seas confirms this paleoecological data. Four examples among many: Steginoporella (Bryozoa), Cymbium, Strombus (Gasteropoda), Scutellidae (Echinodermata). Migrations during the Neogene are sometimes fallowed between adjacent domains: Central Atlantic, Indo-Pacific, or North Sea (at the time of preglacial cooling). In particular to the indopacific direction, more communications are established through the channel of Taurus and Central Iran (Adana—Tabriz-Qom) and through the Syrian archipelago and the mesopotamian basin: straits considerated as "seuils de diffusion" (DEMARCQ, 1984b) with two channels Lattaquié—Aboukemal and Tripoli—Hama. These migrants forms prove the biogeographical communications but in the condition of an optimal paleothermic possibility... sine qua non! (DEMARCQ, 1984b).

*First result: Upper Oligocene hiatus*

In Mediterranea a significant hiatus, quantitative and qualitative, marks the transition between Oligocene and Lower Miocene. There are very important modifications in the populations of marine megafaunas and, for some groups, in the evolutive lines. A disruption seems to occur during the Upper Oligocene, in any case before the beginning of the Miocene. The principal reason must be founded in tectonic, paleogeographic and paleoecologic events. In this causality-context the climatic events are very determining. The mediterranean Oligocene megafaunas are rare and not various: little number of taxa, poor communities, low biomass. Examples: stations of the Egerian in Paratethys; "couches d'Escornebeou" in the terminal Oligocene of south Aquitaine; poor megafauna with Pecten arcuatus in the Upper Oligocene of Cherichira basin in Tunisia, of Syria, etc... These data are according to the results of the continental characteristics (not warm climat, hydric instability) and also to the results of the stable isotopes ($^{13}C$, $^{18}O$) in the oceanic waters (VERGNIAUD-GRAZZINI, 1983).

This hiatus does not coincide with the Oligocene/Neogene boundary but it is anterior to it: into the Upper Oligocene. The evolutiv continuity of the phyla and lineages is good between terminal Oligocene and lower Aquitanian megafaunas but not between middle Oligocene (Stampian, Rupelian) and terminal Oligocene. Many events have taken place on this time, quick, various and repeated.

In Mediterranea, the megafaunas are reconstitued from local "preserved islands" into the Tethys (e.g.: N Italy, Malta, Syria and Lebanon) but especially with the arrival of atlantic and indopacific migrants. The thermic ascent of the sea-waters in the beginning of Aquitanian (e.g.: basis of the Carry formations, in south France) gives a favorable impulsion to the neritic biocenoses (e.g.: large benthic Foraminifera).

*Upper Burdigalian maximum*

The condition of equilibrium became progressively established at the Aquitanian but the blooming of the benthic faunas really takes place at the Burdigalian. Generous exchanges are realized then among the stable margins from East to West and from South to North. In the Lower Aquitanian of Carry, near Marseille, a little but typical recifal bioherm with Poritidae is known (43°15' lat. N). At this time a very important thermic maximum occurs during the Upper Burdigalian and the Langhian. In the all Mediterranea, from Egypt to Rhône valley, megathermic associations are extended. Tropical and particularly intertropical taxa are common in all the groups. The terminal Burdigalian show the maximum migration towards the nordic latitudes of megatherm species, genera and families. Among hundred examples the most remarquable are the presence of Clypeaster in the median Rhône valley (Montsegur near Saint-Paul-Trois-Châteaux, 44°16' lat. N), and large Pectinids near Pontarlier in East-Jura (46°53' lat. N, RANGHEARD et al., 1985). In Austria a famous tropical association persist in the Leithakalk formation in the Badenian (47°54' lat. N). In the littoral of Languedoc exist beautiful recifal constructions (43°15' lat. N, CHEVALIER, 1961) according to the proximity of paleomangrove with Avicennia (SUC, 1980) and Crassostrea. The figure below gives the curve of this thermic variations during the Lower Miocene.

*Serravallian crisis*

A rather general inbalance in the faunas occurs during the Serravallian: quantitative and qualitative impoverishment, triteness of associations and biotopes. All the groups prove this marine megafaunal crisis. The reasons of this phenomene are triple: strong resumption of the tectonic activity (causing detritism and infavourable conditions of environment) palaeobiogeographical disruptions (causing more difficulties in the migrations) but also a very important climatic degradation. The intertropical taxa disappear; the tropical taxa are more rarous in all the groups; the thermic ambiance of the biocenoses—not rich and not various is more or less subtropical. This thermic minimum (see the curve) correspond principally to the second part of the Serravallian (along 13 m.y.). These data are in conformity with other documents: mesotherm neritic faunas in Portugal; stabilization and inertia of the deep water-mass in Mediterranea (VERGNIAUD-GRAZZINI, 1983); formations of first permanent cold water-mass in Atlantic (ibidem); establishment of the first permanent antarctical ices. The Mediterranean megafaunas are withness of a very general first thermic degradation during the Tertiary.

*Tortonian intervall*

An irregular ascent of the temperature of littoral marine waters during the Tortonian, with one or two relative maxima, is attested to new megafaunas associations. From Turkey to Spain, more rich biocenoses are cited with ecological equilibrium in all the groups principaly in the communities of Gasteropoda. New seaways regional transgressions and local structural inaugurations favorise this thermic climax between subtropical and tropical ambiance in the littoral biocenoses.

*Messinian ascent*

A second but short maximum occurs at the Messinian. The salinity crisis remove all the euhalin faunas in Tethys, except in the occidental part of W Mediterranea: it is the "andalusian sanctuary" in the Alboran basin. Here, in south of Spain, north of

Morocco and Oranie, some rich intertropical communities are known. Megatherm taxa are cited in Madrepora, Bryozoa, Brachiopoda, Gastercpoda, Bivalva, Balanidae, Echinodermata, Pisces. They are also observed: intensity of the calcareous metabolism; big and large forms or varieties; constructions with Serpulidae and Vermetidae; typical recifal biohermes (e.g.: recif of Santa-Pola, near Murcia in SE of Spain, at 38° lat. N). This little thermic optimum is the last episode of the normal temperature before the actual periods.

*Pliocene: difficulties and definitiv degradation*

The different groups of neritic Invertebrate show not any megatherm taxa and not much polytherm taxa. At two moments (part of Tabianian, beginning of Plaisancian) are still observed some subtropical elements: big Terebratula, warm-atlantic Mollusca and Echinodermata, a few Bryozoa but only in the Oriental Tethys (e.g.: Hellenic areas) and in the south of occidental Tethys (S Spain, Maghreb). In Roussillon, Rhône valley and N Italy, the megafaunas have already banal characters; their communities are rich and various but more similary than the actual. At the end of Plaisancian, the ecozone with Chlamys inaequicostalis and *Chl. glabra* marks the beginning of the actual mediterranean ambience. These thermic conditions in the littoral waters are established near 2.8—3 m.y., in the upper part of the *Gl. crassaformis* zone. At the lower Quaternary the penetration in Mediterranea of the north atlantic migrants *(Panopea norvegica, Arctita (= Cyprina) islandica, Buccinum undatum)* is a classic and significant prove of the cooling of the seawaters: banal

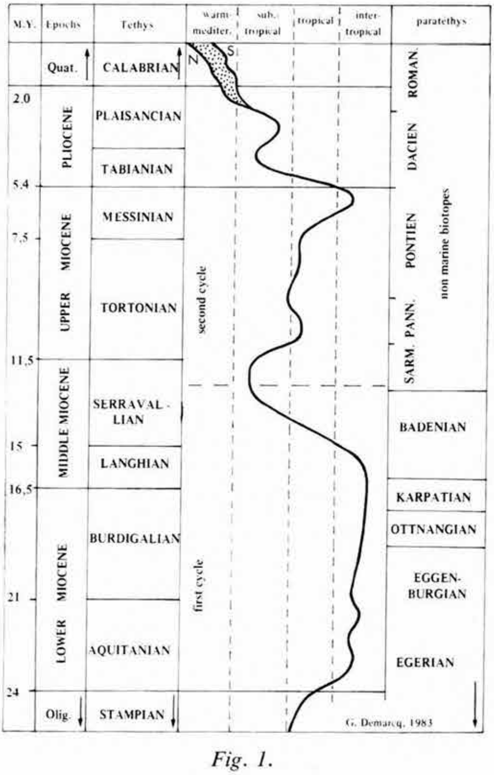


*Fig. 1.*

actual taxa, triteness of associations. But during the Pliocene this irregular and continuous thermic degradation lead to two phenomens: a relative compartimentalization of the biocenoses and a biothermic cleaving between north and south mediterranean areas; consequently, from the second part of Plaisancian it is necessary to cleave definitively the thermic curve (see the figure, with letters N and S).

## Conclusion

The benthic megafaunas are good indicators of the medium temperature of the not deep littoral waters. The results, completed from seven years (DEMARCQ, 1979), are reliable and give a clear view of the thermic evolution from Upper Oligocene to Quaternary. These results are according to the other methods: continental faunas (DEMARCQ et al., 1983), palynology (SUC, 1980), nannoplankton (MÜLLER, 1983) and stables isotopes (VERGNAUD-GRAZZINI, 1983). It is possible to propose a quantitative calibration of the thermic curve through the Madrepora: intertropical between 29°—26° (recifal optimum), tropical 26°—23° (minimum recifal tolerance), subtropical 23°—20° (proportion of the hermatipic taxa), warm-mediterranean lower (20°—17°), the whole in medium compensated temperatures (DEMARCQ, 1984). The establishment of the curve (see figure) indicats four thermic events. Two warm: Upper Burdigalian (long and strong), Messinian (short and little strong). Two cold: Serravallian (serravallian crisis, near 13 m.y.), Pliocene (progressiv and irregular). The whole appears to form two cycles, the first more intense than the second, each of them lasting 11 m.y. It is very interesting to note that the boundary between these two cycles (relative cooling at Upper Serravallian) coincids practically with the separating of the Paratethys.

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# ENVIRONMENTAL CHANGE AND ECOSTRATIGRAPHY IN THE CARPATHIAN BASIN

by

L. KORDOS, M. HAJÓS, P. MÜLLER and E. NAGY

The evolution of the Neogene environment of the Carpathian basin was basically controlled by two impact factors, i.e. the proportion of the flooded and terrestrial areas and the local influence of the general climatic conditions.

By analysing the fossil remains indicative of different environmental factors, that are used in the course of investigations in Hungary, the following sequence of ecological events can be traced. The Pannonian basin, i.e. the depression located between the Alpine ranges and the Carpathian arcs had already existed as a geographical background. The geographical position of the area in terms of latitude and its connection with the other major European units had stabilized.

Given the above position, the crucial factors were the following:

- the horizontal extension of the Tethys and Paratethys and their connection with other seas,
- the changes in the depth and the regime of the marine basins,
- the transformation of the terrestrial mesorelief.

For the determination of the extension of the diversified saline-water facies filling up the Carpathian basin, several palaeogeographical maps prepared according to different principles are available. Among them the manuscript map shown in the course of the present Congress was prepared within the frame of international cooperation. Although these versions differ in some details, they are suitable for providing sketches about the proportion of the terrestrial and marine regions in certain characteristic periods.

The size of the drowned areas in Eggenburgian and Ottnangian time in the course of the Neogene epoch was insignificant and it was not until the end of the Pannonian that similar floods occurred. The Karpatian transgression affected only a small area. The largest area was flooded during the Badenian, and the maximum (70–80 per cent) was reached in the Late Badenian. In the Sarmatian the flooded area somewhat diminished, but during the Early Pannonian and at the beginning of the Late Pannonian the marine–terrestrial ratio reached that the Badenian. The drastic increase in the proportion of the terrestrial areas and then the development of the present-day topography and fluvial system set in from the end of the Late Pannonian.

Along with the palaeogeographical maps prepared first of all on the basis of sedimentology, the vertebrates living both in water and on land are also suitable for the reconstruction of the two factors.

Because of the sporadical occurrence of the fossils and the different way of accumulation of the bones, the marine–terrestrial ratio differs from the results achieved by the former method. First of all the front of the transgression is indicated rather sharply by the shark-tooth horizons both in the Lower Eggenburgian and in the Karpatian. The vast biotope that resulted from the new connections is fundamentally different from the preceding ones, as clearly marked by the great number of Badenian fish and marine mammals. The Sarmatian regression, beside the zoogeographical spreading, also increased the number of factors that could lead to their fossilization. Thus the marine—terrestrial ratio indicates more abrupt change than may have been the reality. Why the sediments of the Pannonian lake are so poor in fish remains is a peculiar and so far unexplained phenomenon. From the Miocene—Pliocene boundary onwards no vertebrate fossil suggestive of considerable water coverage can be found in the Carpathian basin.

The size of the area covered with water, beside being in itself the determination factor, also exerted a climatic impact. No continental climate due to any of the cyclon systems could develop till the Carpathian basin was covered with water. At the same time, the balance of the Earth-atmosphere radiation coming from the albedo of the water surface present in the subtropical climatic zone is positive. This "warming-up factor", considering the extension of the whole Paratethys up to the period of its remarkable decrease, i.e. till the Pontian, must have been significant. After this decrease and with the preservation of the subtropical climate the balance of radiation became negative, and thus the atmospheric conditions became changeable, and this resulted in a more marked continental climate and its biogeographical events. Corals, coral reefs and diatoms are important indicators of the regions of the Carpathian basin covered with water during the Neogene. Within the basin in a more strict sense only the Badenian was rich in corals during the Neogene. In the Early Miocene the spreading of the terrestrial impacts that restrict the Oligo—Miocene marine environment was characteristic. The coral reefs appeared simultaneously with the Karpatian transgression and became dominant with the expansion of the transgression in the Badenian. The corals disappeared from the basin due to a change in salinity in the Sarmatian, and perhaps because of a decrease in the temperature of the water.This means that the corals mark the balanced optimum stage of the salinity and temperature in the Early Badenian.

The number of diatom species in the time-span between the Ottnangian and the Upper Pannonian shows three peaks in the Karpatian, the Middle Badenian and the Sarmatian, respectively. An increase of 300 per cent can be perhaps attributed to volcanism. The Karpatian peak can be tied to the eruption of the Eggenburgian—Ottnangian boundary (Gyulakeszi Rhyolite Tuff Formation), the Upper Badenian peak is the result of the andesite volcanism (Mátra Volcanics Formation) while the Sarmatian peak is due to the Early Sarmatian volcanic event (Galgavölgy Rhyolite Tuff Formation). If, however, only the characteristic diatom species are discussed, only the Karpatian and Sarmatian peaks are the strongest rhyolitic phase, can be mentioned. Besides, the number of Sarmatian species was increased by those coming from the Mediterranean region of the Middle East. The corals together with the diatoms represent a biological signal system that cast light upon the zonation and stability of the marine biotope. During the maximum of the water cover in the Badenian, especially during the Lower Badenian, large areas underlay shallow, lagoon-like seas of uniform salinity and temperature. All these factors caused a period of stability of the terrestrial environment. Its duration marks an ecozone and its end is an ecozonal boundary.

By means of fossils coming from terrestrial areas the palaeoclimatological conditions can be reconstructed. The fossils available were analysed for palaeovertebrates, fossil flora and pollen. The rodent finds from the available vertebrate localities are suitable for drawing up their succession from the end of the Oligocene till the beginning of the Pleistocene. The same significant evolutionary stages can be found in the Carpathian basin as in Spain and Greece:

– the dominance of Eomyidae and Gliridae till the first part of the Middle Miocene,
– the dominance of Cricetidae from the second half of the Middle Miocene till the Early Pannonian,
– the appearence and spread of the Muridae till the Middle Pliocene, and
– the great frequency and density of the Arvicolidae.

This succession indicates an environment, where similar vertebrate faunas could develop in the Mediterranean region and in the Carpathian basin. This, at the same time, refers also to similar palaeoclimatological conditions, i.e. the first remarkable climatic change took place in the Middle Badenian, together with the desiccation. The climate became humide in the Early Pannonian. This was followed by another dry and warm period at the end of the Miocene, where the Cricetidae and Muridae assemblage became widespread. In the second half of the Pliocene great humidity and high temperature became characteristic for a short time, followed by the dry period of the Villanyium.

The most comprehensive investigation of the paleoclimate of the Carpathian basin was carried out by G. ANDREÁNSZKY with help of the rich fossil flora recovered. The present paper gives a revision of the chronology of the Badenian and Sarmatian flora established by G. ANDREÁNSZKY, but the climatic data have been borrowed from his studies. Accordingly, the initial great humidity and high temperature decreased significantly only in the Middle Badenian. In later periods no significant changes in the temperature could be experienced. The humidity began to increase at the end of the Sarmatian and this increase continued in the Late Pannonian. Winter frosts can be detected already in the Sarmatian.

On the basis of the frequency of tropical, subtropical and temperate elements the palynological studies suggest tropical and subtropical climates up to the Middle Badenian. At the end of the Badenian, following the subtropical desiccation, greater quantities of pollen from tropical plants appear in the Sarmatian. Pollen grains indicative of a temperate climate with a cooling-down trend become dominant in the Lower Pannonian. In the Late Pannonian the Carpathian basin is characterized by extreme ecotypes. In the shallower sections of the basin marshes with thermophilous subtropical plants evolved, while on the adjacent hill-sides much drier forests developed.

According to the ecological data available, three cardinal events can be clearly identified in the palaeoenvironmental history of the Carpathian basin:

– Desiccation in the Middle Badenian. In spite of the maximum water surface the physico-chemical equilibrium of the balance was thrown off. A significant change in the terrestrial flora and fauna took place as late as in Miocene time. Some sedimentological data also refer to a short period of desiccation.
– Dry period at the end of the Miocene. With the filling-up of the Pannonian lake and the setting-in of normal fluvial regime and a possible increase in wind erosion the steppe-dwelling vertebrate fauna greatly increased. A greater variability of several genera of animals and plants could be experienced. This event can be related to the Messinian salinity crisis.
– Humidity maximum of the Late Pliocene. At present it is a high temperature wet period, followed by a drier period, that can be inferred from the vertebrate fauna only.

These three environmental events are suitable for ecostratigraphical correlation. For the establishment of the ecozones of the Neogene in the Carpathian basin not only the regional climatic changes, but also the local changes in the environment can be used. On this basis seven ecozones, marked provisionally with letters from A to G, could be distinguished.

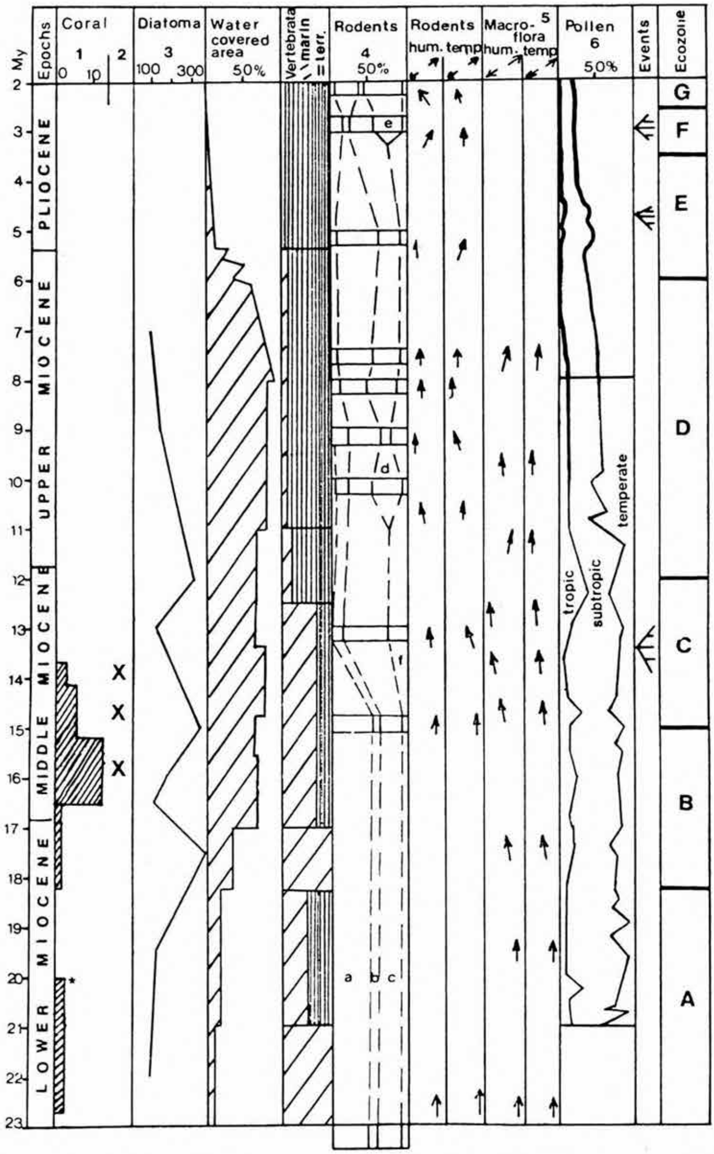


*Fig. 1.* Ecostratigraphical elements and ecozones of the Neogene in the Carpathian basin

*1* Number of hermatypic coral genera, *2* presence of coral reefs (P. MÜLLER), *3* number of diatoma (M. HAJÓS), *4* a = Eomyidae, b = Gliridae, c = Cricetidae, d = Muridae, e = Arvicolidae, f = others (L. KORDOS), *5* after G. ANDREÁNSZKY, 1954, *6* after E. NAGY *outside the basin (Eggenburg)

A ecozone — Relatively small water surface with a balanced tropical—subtropical climate (Eggenburgian—Ottnangian).
B ecozone — Large-scale inundation by seawater and its stabilization with tropical and subtropical climate (Middle Miocene, Karpatian and older Badenian).
C ecozone — Desiccation in the Middle and Late Badenian.
D ecozone — From the unambiguous and constant process of cooling down till the adult stage of the filling-up of the Pannonian lake. Stabilized and cooler climate with changes in humidity.
E ecozone — Desiccation at the end of the Miocene.
F ecozone — Humidity maximum of the Late Pliocene.
G ecozone — Appearance of continental climate at the end of the Pliocene.

These ecostratigraphical stages developed for the Carpathian basin for the first time can later be used for correlation, especially those regions we do not have direct sedimentary contact with. Ecozones A, B, C and D completely agree with the complex major lithostratigraphical and biostratigraphical units that have already been described the Neogene Congress by G. HÁMOR. He also proposed the introduction of new superstages, i.e. Atlantian that is equivalent to ecozone A, Mediterranean corresponding to B and Caspian to C and D.

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# APPROACH TO THE SPANISH CONTINENTAL NEOGENE SYNTHESIS AND PALAEOCLIMATIC INTERPRETATION

by

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*Introduction.* Integrated studies on Neogene geology have been scarce in Spain, but attemps to stratigraphic and sedimentological analysis of continental Tertiary basins have increased considerably lately. The large extent of Neogene basins in Spain, the good quality of the outcrops and the abundance of fossil provide an excellent basis for this kind of studies.

Advances in basis analysis, supported by the achievement of 10 national meetings of sedimentology, continental biostratigraphy programs and regional mapping projects sponsored by public funding, enable us to present this approach assuming that it adequately reflects the present day knowledge about the continental Neogene in Spain.

Previous synthesis on this topic have been outlined by Vertebrate palaeontologists specially (AGUIRRE, 1974; ALBERDI et al., 1975; AGUIRRE, 1975; CRUSAFONT et al., 1975) but the stratigraphic background has barely been taken into account. Numerous regional works have provided extensive information which has made possible the present synthesis. It is supported by a considerable amount of data, and proposed to be tested and compared with other synthetic attempts on Mediterranean Neogene geology.

*Basin analysis.* Constructing a sketch summarizing the information on different basins (Fig. 1) needs previous discussion and data selection. Stratigraphic units have been distinguished in the simplified logs representing to the basin sedimentary successions. The method used in this paper approaches the infill unit of continental basins to the tectosedimentary unit (TSU) defined by MEGIAS (1973, 1982). The depositional sequence defined by MITCHUM et al. (1977) for marine sediments are geometrically similar to the TSU, but they propose an essentially eustatic control in their model. This makes difficult its conceptual use in continental basins with their own base level, and where the control of the depositional units is considered mainly tectonic.

The limits between the stratigraphic units (Fig. 1) are sedimentary discontinuities or ruptures of basinal range. Their origin and identification criteria were discussed by MEGIAS (1973, 1982). In this paper the ruptures limiting stratigraphical units have been recognized in the field as hiatuses (with or without karstification), erosions, unconformities (and their relative conformities), horizontal and vertical changes of polarity during the depositional processes and boundaries between megasequences. Not all co-authors of this paper have identical view on the work method; data have been selected according to the accepted criteria given above.

*Biostratigraphy and marine—continental correlations.* Regional biozonation based on Micromammal assemblage zones have been used for biostratigraphic correlations. Overlied assemblage zones have been defined in the Teruel (WEERD, 1976; MEIN et al., 1983), Daroca (DAAMS et al., 1981), Vallés-Penedés (AGUSTÍ, 1981;

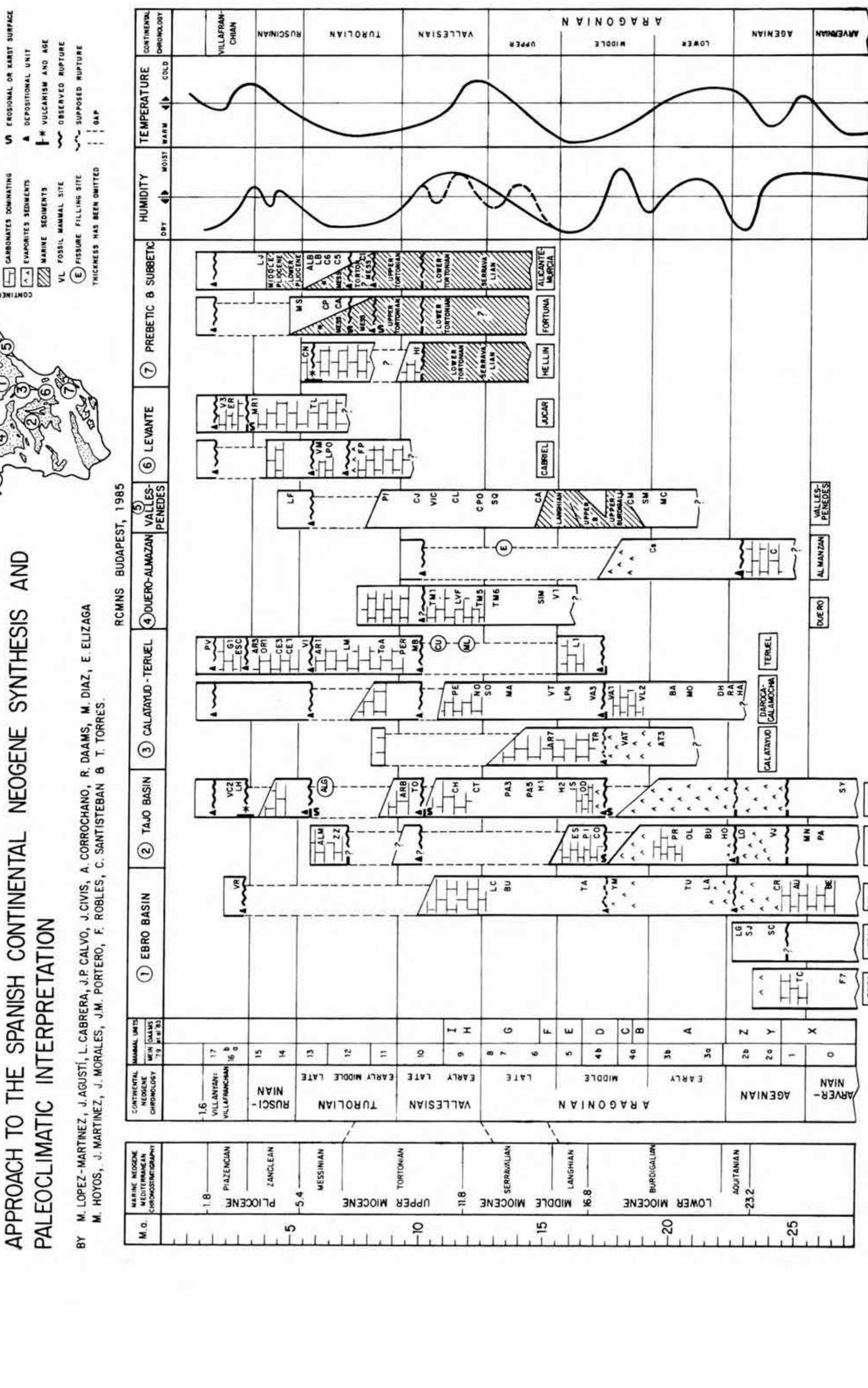


*Fig. 1.* Correlations of the Spanish continental Neogene sedimentary sequences. Sedimentary units and ruptures have been dated on the basis of Micromammals fossil sites; several of them are coded on the logs

ALB — Alberca
ALG — Algora
ALM — Almendros
AR 1, 3 — Arquillo 1, 3
AR 7 — Armantes 7
ARB — Arbancón
AT 3 — Ateca 3
AU — Autol
BA — Bañón
BE — Bergasa
BS — El Buste
BU — Buciegas
C — Cetina
C 1, 5, 6 — Crevillente 1, 5, 6
CAL — Can Almirall
CC — Casa del Acero
CE 1, 3 — Celadas 1, 3
CH — Chiloeches
CJ — Can Jofressa
CL — Can Llobateras
CM — Can Marti Vell
CN — Cenajo
CO — Córcoles
CP — Casa de las Palomas
CPO — Can Ponsic
CR — Carretil
CS — Ca Soria
CT — Cendejas de la Torre
CU — Cucalón
DH — La Dehesa
E — Escobosa
ES — Escamilla
ESC — Escorihuela
ER — El Rincón
F 7 — Fraga 7
FP — Fuentepodrida
G 1 — Gea 1
H 1, 2 — Henares 1, 2
HI — Hijar
HO — Huerta Obispalía
IS — San Isidro
LA — Los Agudos
LB — Librilla
LC — La Ciesma
LF — La Fortesa
LG — La Galocha
LH — Las Higueruelas
LI — Libros
LJ — La Juliana
LM — Los Mansuetos
LD — Loranca
LP 4 — Las Planas 4
LPO — La Portera
LVF — Los Valles de Fuentidueña
MA — Manchones
MB — Masía del Barbo
MC — Molí Calopa
ML — Molina de Aragón
MN — Moncalvillo
MD — Moratilla
MR 1 — Marmota 1
MS — Molina Segura
NA — Navarrete
ND — Nombrevilla
OD — O'Donnell
OL — Olmeda
OR — Orrios 1
P — Pineda
PA — Parrales
PA 3, 5 — Paracuellos 3, 5
PE — Pedregueras
PER — Peralejos
PI — Piera
PR — Priego
PV — Puebla de Valverde
RA — Ramblar
SC — Santa Cilia
SIM — Simancas
SJ — San Juan
SM — San Mamet
SO — Solera
SQ — San Quirce
SY — Sayatón
TA — Tarazona
TC — Torrent de Cinca
TL — Tolosa
TM 1, 5, 6 — Torremormojón 1, 5, 6
TO — Torija
TOA — Tortajada
TR — Torralba de Ribota
TU — Tudela
V 1 — Valladolid 1
V 3 — Valdeganga 3
VA 1, 3 — Valdemoros 1, 3
VAT — Valtorres
VC 2 — Valverde Calatrava 2
VI — Villastar
VIC — Viladecabals
VJ — Vallejo
VL 2 — Villafeliche
VM — Venta del Moro
VR — Villarroya
VT — Valalto
YM — Yesos de Monteagudo
ZZ — Zafra de Záncara

Agustí et al., 1984 a) and Duero (Alvarez et al., 1985) basins. The biochronological scale of Mein (1975) has been used together with that of Daams et al. (1984), more adequate for the Spanish fauna. The migration events of Large Mammals, traditionally used to define Mammal Ages, occur within the Micromammal zones. Changes in the composition of Mammal fauna seems controlled by environmental and climatic changes, because each assemblage has a different ecological meaning. The entry of immigrants is related also to palaeogeographic changes of intercontinental range.

Sedimentary ruptures have been dated according to the age of the oldest overlying bed or the youngest underlying bed. Ruptures do not coincide with faunal changes. All the eight dated generalized ruptures occur within biozones. No wonder, since sedimentary ruptures are mainly related to tectonics, while faunal changes to climate.

The controversed correlation of marine and continental scales has been summarized by Rögl et al. (1983), though the equivalence is not definitely stablished. We have used their results but we propose to change it at three points (see Fig. 1):

*a* Late Aragonian—Langhian; this is a first order correlation in Vallés-Penedés basin (Agustí et al., 1984 a).

*b* Early Vallesian—Serravallian; the oldest *Hipparion* was calibrated in Europe at—12.5 Ma., and a first order correlation has been indicated by Mein (1985).

*c* Middle Turolian—Messinian; in the Levante basins a second order correlation is pointed out in Crevillente 5 and Casa del Acero (Bruijn et al., 1975; Agustí et al., 1975; Agustí et al., 1981, 1984). Many of the marine-continental correlation problems derive from the criteria of marine stage boundaries, as arbitrary historical boundaries cannot be translated to continental basins only biostratigraphic correlations are possible (see Fig. 3).

*Summary of the Spanish continental Neogene.* Spanish continental Neogene basins cover more than 100 000 km$^2$. Eastern basins show marine interbedded sediments, but in northern and central basins only Neogene continental sediments have been indicated. Thickness can reach more than 1000 m. Lithologies are varied, according to the origin of the materials but the amount of evaporites is striking.

Simplified logs of each area (Fig. 1) are the results of careful lithostratigraphic and sedimentologic analysis, sampling and correlation of fossil sites, biostratigraphic and basin analysis and the integration of both kind studies. The discontinuities, both stratigraphical and palaeontological are the rule; but when a sedimentary rupture has been well-dated, the underlying and overlying fauna showed the same composition. An intra-biozone position has been verified for most of the major ruptures indicated in Fig. 1.

*1* An *Early Agenian* rupture is revealed in Ebro and Tajo basins as a change from well-developed fluvial system sedimentation to a mud-flat, playa-lake and lacustrine deposits. *Rhodanomys* faunas are present both below and above the rupture. This event corresponds probably to the Upper Oligocene.

*2* A sedimentary rupture in *Late Agenian* or *Early Aragonian* has been recognized in the Ebro, Tajo and Duero (Almazán) basins. This rupture appears as an angular unconformity on the margins of the basin and a change of sedimentary polarity in its centre of it. *Ligerimys* (one species), *Anchitherium* and *Gomphotherium* are recorded both below and above this rupture. A possible correlation may be suggested with the Aquitanian based on marine—continental interbedded deposits in Lisboa R—1 (Antunes et al., 1973).

*3* A *Middle Aragonian* sedimentary rupture seems to affect each of the studied basins. Some of them display angular unconformities; others have "cannibalistic" autophagic detritical sediments, and finally others seem to have formed at this moment. *Hispanotherium*, recorded above this rupture, and other faunistic criteria indicate increased aridity, coinciding with sepiolite deposits in the Tajo basin. The marine-continental interbedding in Vallés-Penedés basin allows to correlate this rupture with the Upper Burdigalian (AGUSTÍ et al., 1984).

A minor sedimentary rupture within the Late Aragonian deposits has been detected in Madrid basin. It has not been marked in the sketch because it is not clearly observed in other basins.

*4* A *Late Vallesian* rupture is commonly located as a palaeokarst on the top of a structural folded surface of a carbonate depositional unit in most of the basins. The overlying unit is a complex deposit of terrigenous sediments and peculiar palustrine carbonates. *Hipparion* and *Progonomys* are present both below and above this rupture. A correlation with the Lower Tortonian has been verified in the Prebetic basins (CALVO et al., 1978).

*5* A *Middle Turolian* rupture is well-characterized in most of the basins. *Parapodemus* are recorded both below and above this rupture. A correlation with marine deposits can be established in the Fortuna and Alicante basins. The Middle Turolian site of Casa del Acero overlies shallow open marine deposits of Messinian age (with *Globigerinoides elongatus* and *G. extermus*) and an evaporite diatomitic unit with *Globorotalia mediterranea.* Between this evaporite unit and the overlying terminal complex (evaporites, reefs and associated continental deposits), a major rupture can be correlated with the intra-Messinian regional rupture (SANTISTEBAN, 1981; MEGIAS et al., 1983; AGUSTÍ et al., 1984 b).

*6* A *Late Turolian* sedimentary rupture appearing as a major progradation of terrigenous over chemical sediments is recorded in many basins. This rupture has a probable absolute age of about $-5.7 \pm 0.3$ Ma dated at the Monagrillo volcanic event (BELLON et al., 1981). *Apodemus* faunas are recorded both below and above this rupture.

*7* A *Late Ruscinian* or *Early Villafranchian* rupture is well-known in the Ebro, Tajo, Júcar and Teruel basins. A new depositional unit with terrigenous and carbonated sediments overlie a folded and karstified surface. *Mimomys* aff. *cappettai* (sensu WEERD, 1976) has been recorded both below and above this rupture (MARTINEZ and ESTEBAN, 1985). This absolute age is near $-3.5 \pm 0.3$ Ma according to the volcanic event of Campo de Calatrava (Las Higueruelas, ALBERDI et al., 1984).

*8* A *Late Villafranchian* rupture can be distinguished prior to the recent fluvial system incision. A large conglomeratic unit overlies an erosional incrusted surface. *Equus* and *Mammuthus* faunas occur below and above this rupture.

Other basins which have not yet been studied can be used to test this model. For example, the Guadix-Baza basin studied by AGUSTÍ et al., (1985) corroborates the Late Villafranchian rupture in a different geographic and geologic framework.

*Palaeoclimatology*. The palaeoclimatic interpretation of this synthesis is mainly based on faunal data, because the palaeobotanical data so far are scarce. The interpretation is based on the criteria proposed by WEERD et al. (1978) and DAAMS et al., (1984), who use percentages of Micromammal climatic indicators (glirids, castorids, eomyids etc). Invertebrates are also good temperature indicators, and in some cases oxigen isotopes in shells have been measured (ALBERDI et al., 1982). Organisms are more sensible to the variations in climate than sediments, which reveal to be more controlled by other factors. Several examples of the Spanish Neogene sediments show that sedimentological criteria are ambiguous as climatic indicators;

*a* the contemporaneous deposition of both types, arid alluvial fan and wet fluvial fan in the same basin;

*b* alluvial fan deposits where sedimentary facies are mainly mass transport deposits, having a fan radius longitude similar to those of the wet fluvial fan models (SCHUMM, 1977); these examples would indicate that all drainage system variables (SCHUMM, 1981) and not only climate, have to be considered;

*c* evaporitic deposits do not always correlate with warmer and drier intervals but can be associated to moister and colder times in Tajo basin (Fig. 1) where, on the basis of field criteria, a control of the source area lithology can be demonstrated.

Deduced climatic changes should be attributed either to humidity oscillations or temperature variations. Temperature changes were more difficult to detect. Four relative cooling phase have been recognized in Late Oligocene (X-zone), Early Miocene (Z–A zones). Aragonian/Vallesian limit and Ruscinian/Villafranchian limit. The last three of them may be roughly related to those outlined by KEIGWIN et al. (1979) and MULLER (1983) at –23 Ma, –12 Ma and –3 Ma in the marine Neogene.

Relative humidity changes can be detected easier than temperature variations. The Neogene faunas in Spain seem to indicate a more arid climate than that of the areas located north of it. Relative increases in humidity are recognized during the Middle Agenian, Early Aragonian, Early Vallesian, Early Ruscinian and Late Ruscinian. Other humidity oscillations in the Late Aragonian and Vallesian of the Tajo and Vallés-Penedés basins have been detected.

*Chronostratigraphy and global events.* Global ruptures have been recognized in marine sedimentary successions by VAIL et al., 1979 (changes of sea-level based on coastal onlap, and unconformities), KELLER et al., 1983 (hiatuses) and SOLER et al., 1983 (regional tecto-sedimentary ruptures). Based on marine-continental biostratigraphic correlations, we have verified that most of the ruptures in both marine and continental series coincide chronologically (Fig. 2). The major ruptures are at –22,5 Ma (intra-Aquitanian ≃ Late Agenian), al –10 Ma (intra-Tortorian ≃ Late Vallesian), al –7 Ma (intra-Messinian–Middle Turolian), and at –3 Ma (intra-Piazencian ≃ Early Villafranchian). Other ruptures appear also well-correlated, such as at –81 Ma (intra-Burdigalian ≃ Middle Aragonian), at about –5,5 Ma (Mio/Pliocene limit ≃ Upper Turolian) and at –1.8 Ma (Plio/Pleistocene limit ≃ Upper Villafranchian).

Sedimentary ruptures divide the Neogene into tecto-sedimentary units with a chronostratigraphic significance (as indicated by MEGIAS, 1973, 1982; MITCHUM et al., 1977; VAIL et al., 1982). Most of the global ruptures do not coincide with classic chronostratigraphic limits. Boundaries between stages, defined from stratotypes, are located within the TSU (Fig. 2) because of the methodology of the classic stages definition. They are based on stratigraphic record (stratotypes) of transgresive events that usually correspond to the terminal episodes of coastal onlap (Fig. 3). For this reason, stages boundaries only can be correlated on the basis of biostratigraphic successions. On the contrary, sedimentary ruptures can be correlated by means of basin analysis and biostratigraphy, even between marine and continental successions; therefore they are a better reference for chronostratigraphic limits.

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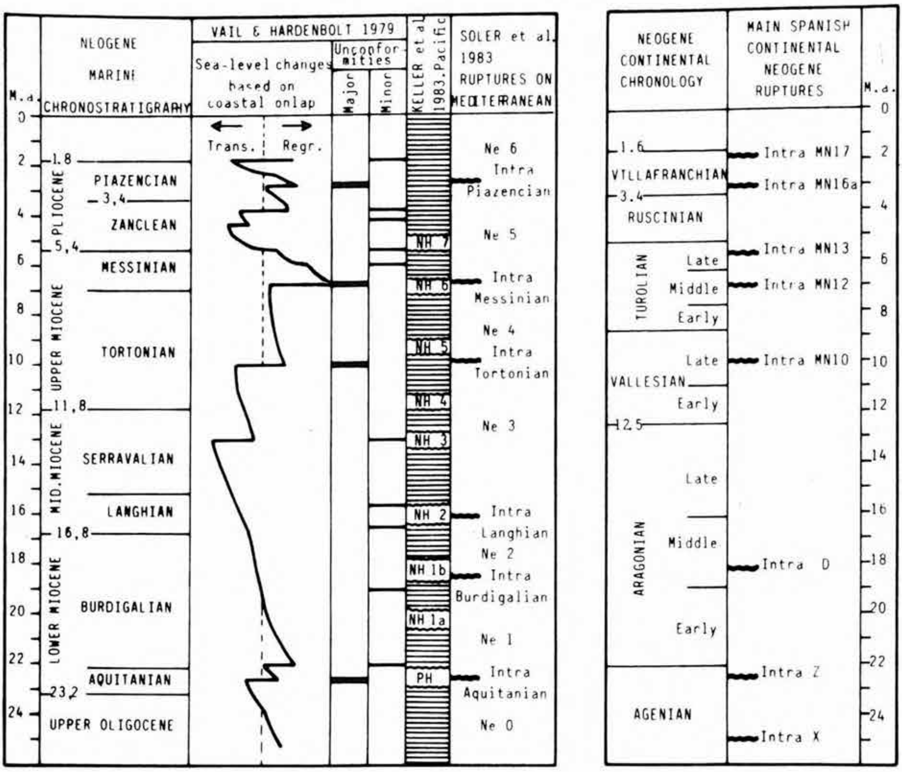


*Fig. 2.* Correlation of global changes in marine and continental Neogene. Note the position of the classic chronostratigraphic limits within the transgressive events, and the sedimentary ruptures within the stages

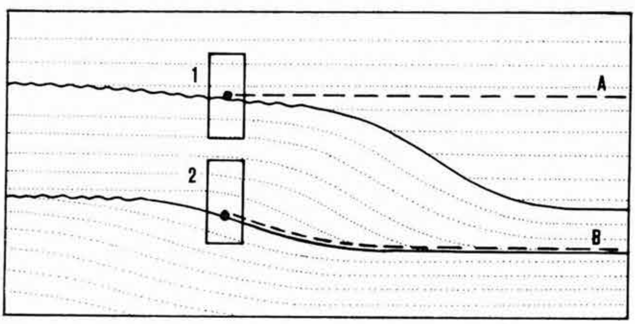


*Fig. 3.* Sketch representing several transgression—regression cycles with two ruptures dividing the record into three tecto-sedimentary units

*1* Situation of the classic stratotypes on the coastal onlap episodes. The transgression event is diachronous, so the chronostratigraphic limit is defined on the basis of the transgression in the stratotype (isocron A). It can only be correlated biostratigraphically. *2* situation of a posible stratotype, being the sedimentary rupture the chronostratigraphic limit (isocron B). It can be correlated by basin analysis and/or biostratigraphy

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# NEOGENE VERTEBRATE BIOSTRATIGRAPHY IN HUNGARY

by

L. KORDOS

The main aspects of the biostratigraphical evaluation of the Neogene vertebrates discussed in the course of the VIIIth RCMNS Congress (Budapest, 1985) included:

- collection of all possible finds from Hungary, classification into lithostratigraphic formations, their mechanical correlation with the MN Zones;
- content distinction of well-correlated horizons within the Neogene Mammalian Zonation (Szentendre, Hasznos, Tardosbánya, Polgárdi 4, Tihany, Nyárád) by excavating new sites and by determining their fauna;
- the elaboration of an independent, regionaly valid biostratigraphy of vertebrates (Anomalomyidae, Cricetidae) by vertical systematic and evolutionary reevaluation of the major rodents.

At the present stage no vertebrate biostratigraphy elaborated in detail can be mentioned, due first of all to the lack of finds from certain stages or to the incompleteness of modern revisions, respectively.

In the course of the evolution of the Neogene vertebrates in Hungary the marine and terrestrial cycles based on different formations can be sharply differentiated. From the Eggenburgian up to the Pannonian a total of three marine and two or three terrestrial cycles can be differentiated. In the Pannonian, in spite of the fact that the largest area was covered by water in this period, hardly any aquatic fauna can be found. Rather the rich terrestrial fauna, mostly from local karstic sediments is characteristic. The transgression of the Eggenburgian, beside the shark-teeth sandstone at Ipolytarnóc, is marked by reptiles and marine mammals, being followed by the terrestrial—marshy facies of the Ottnangian. In the browncoal layers of the Ottnangian, rich in proboscidians and Ungulata, shark, dolphin or crocodile remains scarcely occur. The Karpatian transgression is quite distinctly introduced by the shark-teeth horizon at Kozárd. At the same time, exclusively allochtonuos, poor mammalian finds are also known. On the varied landscape of the Badenian, a new wave appears in both the marine and terrestrial faunas. In the former the appearance of Carcharodon—Myliobatis associations, in the latter that of modern Rhinocerotidae is the characteristic. On the basis of the vertebrate fauna the division within the Badenian is not yet possible. No marine mammals are practically known from the Sarmatian. There are some difficulties concerning the determination of the Sarmatian—Pannonian boundary, in spite of the fact that in terms of the *F*irst *A*ppearance *D*ata (FAD) of Hipparion there should not be any problem.

The attached table indicates that there are some differences in opinion concerning the various stages and MN Zones in the chronologies applied in Hungary and abroad. The reason for these differences is partly the lack of fauna in the Lower and Middle Miocene, and partly the arbitrary use of guiding horizons (rhyolite tuff) as applied

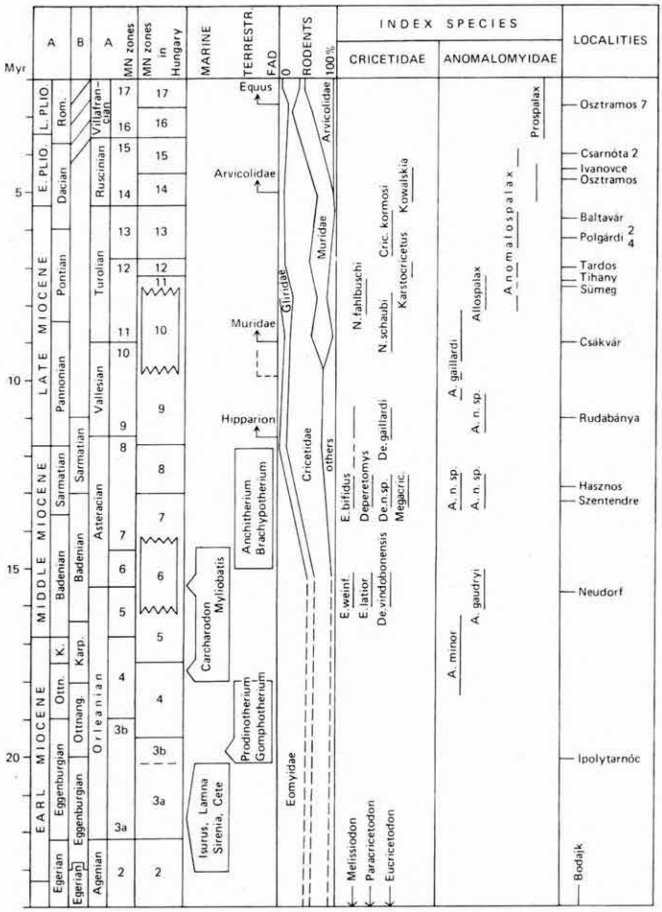


*Fig. 1.* Stratigraphical correlation of the Neogene vertebrate fauna in Hungary
A = after STEININGER and RÖGL, 1983; B = after HÁMOR et al., 1985

for practical purposes in Hungary. In case of the Pannonian, Pontian and Pliocene formations the correlation of the horizons based on aquatic molluscs and terrestrial layers, due to the presence of proper complex localities, has so far been rather dubious. For the moment the Vertebrate fauna found at Tihany is of utmost importance, that can serve as a right basis for proper chronological correlation.

By and large the evolution of the rodents was the same in the Carpathian basin as in the southern regions of Europe. For the stratigraphic identification of larger units the FAD and LAD of different families and genera are suitable. Only Muridae appeared later in the Carpathian basin than in southern Europe. The majority of certain species and sometimes genera mainly in the southern and western foreland of the Alps and the Carpathians, offers relationships providing a possibility for direct correlation. The obvious eastern connections of the Neogene vertebrates are not still sufficiently known.

The revision of two rodent families, Anomalomyidae and Cricetidae has been carried out by the present author to extent enabling a more precise and slightly different chronological classification. *Anomalomys minor* FEJFAR found in the limnic sequence in the Mecsek Mts (Váralja 21 boring, range 264.7—265.4 m, Szászvár Formation) is Karpatian at the type locality (Franzensbad, Czechoslovakia), while the new finds coming from the Alpine region (Niederbayerns) can be classified as Badenian. The same species at Aliveri (Greece) appears already in the MN 3 zone. On the above basis the *Anomalomys minor* in the Mecsek Mts enables us to extend the range of the local limnic sequence towards the Karpatian.

In the Carpathian basin, *Anomalomys gaudryi* GAILLARD is known only from the locality at Neudorf-Spalte (Devinská Nová Ves) from the MN6 Zone (Badenian). At recently discovered important localities at Szentendre and Hasznos (Hungary), that consist of marshy terrestrial, sometimes diatomaceous sediments covering the Upper Badenian eruptive rocks, an Anomalomys species can be identified. The species is related to the Opole finds (Poland, MN7 Zone) and to *Miospalax monacenis* STROMER. Also at the Hasznos locality there is another *Anomalomys* n.sp. that morphologically completely agrees with the *Anomalomys gaudryi* finds (inv. no. UUG. 167) from Neudorf 2.

From the Prehominidae locality at Rudabánya only one and new hypselodont Anomalomys species is known. At Csákvár (Hungary), however, typical *Anomalomys gaillardi* VIRET et SCHAUB was found. Systematically it is important that KRETZOI's *Allospalax plenus* described from Sümeg (Hungary) agrees with *Prospalax petteri* BACHMAYER et WILSON from Kohfidisch (Austria) that, with regard to the rule of priority, should be referred to as *Allospalax petteri* in the future. This species is characteristic of the MN11 zone, being known from the karst locality at Sümeg and from Tihany and Nyárád (Hungary) are open-air localities correlated with aquatic molluscs.

From the rich vertebrate fauna of fossil karst sediments in the quarry at Tardosbánya (Gerecse Mts), a new Anomalomyidae genus, *Anomalospalax tardosi* KORDOS (1985) was described by the present author, from the MN12 Zone. It originates from "*Prospalax*" *gernoti* DAXNER-HÖCK, and this line is continued by the *Anomalospalax viretschaubi* (KRETZOI) at Polgárdi (MN13 Zone) and by a new Anomalospalax species considered as *Prospalax priscus* (NEHRING) from borehole Csarnóta 2 zone MN15. This latter probably totally or partially agrees, with the finds formerly considered *Prospalax priscus* as coming from Weze (Poland) and Ivanovče (Czechoslovakia), also zone MN15. The tooth-morphology of the *Prospalax priscus* classified from Hajnačka (Czechoslovakia) to zone MN16 indicates the further evolution of the species found at Csarnóta. Systematically, it is significant that Late Pliocene—Early Pleistocene *Prospalax priscus* (NEHRING) belongs to Anomalomyidae via Anomalospalax rather than to Spalacidae. At present only the type species is known from this genus. The species, that were formerly assigned also to this genus *(Prospalax gernoti, petteri* and *kretzoii)*, belong to other genera. The other revised family is Cricetidae. From the earlier part of the Neogene (Late and Middle Miocene) no hamster-like rodents are known in Hungary. The period seems to be characterized by further evolution of the fauna that contains rich Melissiodon, Eucricetodon and Paracricetodon remains from the Oligocene locality at Bodajk (Hungary). From the Neudorf—Spalte locality (Czechoslovakia), after FEJFAR, *Eumyarion weinfurteri*, *E. latior* and *Democricetodon vindobonensis* species are known. The exact biostratigraphic situation of the Szentendre and Hasznos localities (boundary of MN9—8 zones, Badenian—Sarmatian) is marked by *Deperetomys hagni* ssp., *Democricetodon hasznosensis* KORDOS, *Eumyarion* aff. *bifidus* and *Megacricetodon minor-group*. In Rudabánya locality MN9 zone, *Eumyarion bifidus* is still present, *Democricetodon gaillardi freisingensis* frequently occurs and Megacricetodon is already missing.

Cricetidae from the Pannonian, first of all the Kowalskia genus has undergone an important systematic revision, in which mainly finds from Hungary have been involved. *Neocricetodon schaubi* KRETZOI from Csákvár (MN10) is valid and can be identified at Sümeg (MN11) as well. At this latter locality it is coexisting already with "*Neocricetodon transdanubicum* KRETZOI" that is used as synonym for "*Kowalskia*" *fahlbuschi* (Kohfidisch locality, MN11) that is transferred to the Neocricetodon genus.

A new Cricetidae genus appears in the Tardosbánya fauna (Karstocricetus) that is followed by the finds in the freshwater limestone at Budapest—Széchenyi Hill (MN 12).

Zone MN13 is marked by *Cricetus kormosi* SCHAUB and its predecessor, *Cricetus* n.sp., that were all found at the Polgárdi localities. *Cricetus kormosi* makes possible correlations of finds discovered at larger distances, i.e. it can be found at Crevillente 6, d'Arguillo, Venta del Moro, Librilla, La Alberca and d'Alcoy (Spain), Gargano (Italy).

*Kowalskia polonica* and *K. magna* FAHLBUSCH, corresponding in age their counterparts at the type locality Podlesice, Poland (MN14) can be found in borehole Osztramos 1 in Hungary.

At present, in the Carpathian basin, mainly on the basis of localities in Hungary, the following biostratigraphical units can be singled out and correlated:

- Deperetomys—Megacricetodon zone (boundary of the MN8—9 zone Badenian—Sarmatian)
- Democricetodon gaillardi—Hipparion FAD zone (MN9 zone Lowermost Pannonian)
- Neocricetodon schaubi—Anomalomys gaillardi zone (MN10 zone Pannonian = "Lower Pannonian")
- Neocricetodon fahlbuschi—Allospalax petteri zone (MN11 zone boundary of the Pannonian and Pontian)
- Karstocricetus—Anomalospalax tardosi zone (MN12 zone Pontian)
- Cricetus kormosi—Anomalospalax viretschaubi zone (MN13 zone end of the Pontian, Messinian)
- Kowalskia zone (MN14—15 zone, Early Pliocene)

The number of vertebrates families included in the revision and thus involved in stratigraphic correlation will increase, and that will make possible the elaboration of a more detailed stratigraphy in the long run.

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# REPORT ON THE ROUND TABLE DISCUSSION: "MEDITERRANEAN AND PARATETHYS CORRELATIONS"

by

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with contributions and remaks by

AGUILAR J. P.: letter dated December 3, 1985.
AGUSTÍ J.: Upper Miocene correlations in Eastern Spain.
ALBERDI M. T. and F. B. BONADONNA: Results on Pliocene marine—continental correlations in Spain and Italy.
ANDREESCU I.: letter dated December 12, 1985.
ČTYROKÝ P.: The stratigraphic position of the first Hipparion occurrence in Southern Moravia, Vienna basin, Czechoslovakia.
KOJUMDGIEVA E.: Tarkhanian—Tshokrakian correlation problems.
MARINESCU F.: letter dated February 27, 1986.
MENNER V. V., PEVZNER, M. A. and VANGENGEIM E. A.: letter dated December 16, 1985.
NOSOVSKY M.: letter dated January 28, 1986.
PEVZNER M. A. and VANGENGEIM E. A.: letter to F. F. STEININGER, dated October 9, 1985.
STEININGER F. F.: Remarks to the letter of PEVZNER and VANGENGEIM (October 9, 1985) and the letter of MENNER et al. (December 16, 1985).
VASS D.: letter dated December 1, 1985.

*Introduction.* Since 15 years one of the main goals of IGCP Project No 25, "Stratigraphic Correlation Tethys—Paratethys Neogene", has been to provide a more accurate correlation between the different main regions of the Mediterranean, the Central Paratethys and the Eastern Paratethys. As it turned out lately by the work done of our colleagues, significant differences still exist with regard to the final results of this Project.

Similar problems refer to different versions of the correlation of planktonic biostratigraphies of the world oceans and the paleomagnetic time scale (e.g. W. A. BERGGREN et al., 1985; J. A. BARRON, 1985 and C. MÜLLER, 1984) (Fig. A, B, C). Lately another version of planktonic biostratigraphic correlations appeared: H. M. BOLLI et al., 1985: Plankton Stratigraphy (Cambridge Univ. Press).

At the round table meeting of the "RCMNS Working Group on Chronostratigraphy and Geochronology" during the Budapest Congress the discussion turned up several correlation key points which were then summarized by us for the Mediterranean (Table 1), the Paratethys (Table 2) and additional correlation points and problems (Table 3). Based on this material we tried to come up with a revised tentative correlation which follows (1) the proposed Mediterranean correlation by MÜLLER (1984)—see Fig. D—and (2) the standard correlation for the Pacific by BARRON (1985)—see Fig. E. The summarized results of the Budapest round table and figures A to E were sent to all colleagues interested to raise more discussion on additional correlation points. The following questionaire was added to gain more precise information:

QUESTIONAIRE

*A) Biostratigraphic data:*

A. 1. Correlation points of aquatic (e.g. marine, euxinic etc) and continental environments;
correlation points between Central and Eastern Paratethys;
correlation points between Mediterranean and Paratethys.

A. 2. Those correlation points should be documented by:
a lithological section with biostratigraphic data points;
a list of species for each data point (please do not list families or genera only).

A. 3. References.

*B) Radiometric data:*

Radiometric data points should be documented by:

B. 1. Lithological section with position of radiometric samples and biostratigraphic data points (see A).

B. 2. Radiometric results and constants used.

B. 3. References.

*C) Palaeomagnetic data:*

Palaeomagnetic data should be documented by:

C. 1. Lithological section with palaeomagnetic sample points and polarity measurements.

C. 2. Biostratigraphic and radiometric data points (for indications see A and B).

C. 3. Palaeomagnetic interpretations.

C. 4. References.

An updated Neogene correlation chart was expected to be produced by this discussion and to be included in this volume of the Proceedings of the Budapest RCMNS Congress. However, since the response was rather meagre (see below), we prefer to include the answers received only without compiling an updated correlation chart.

F. RÖGL and F. F. STEININGER

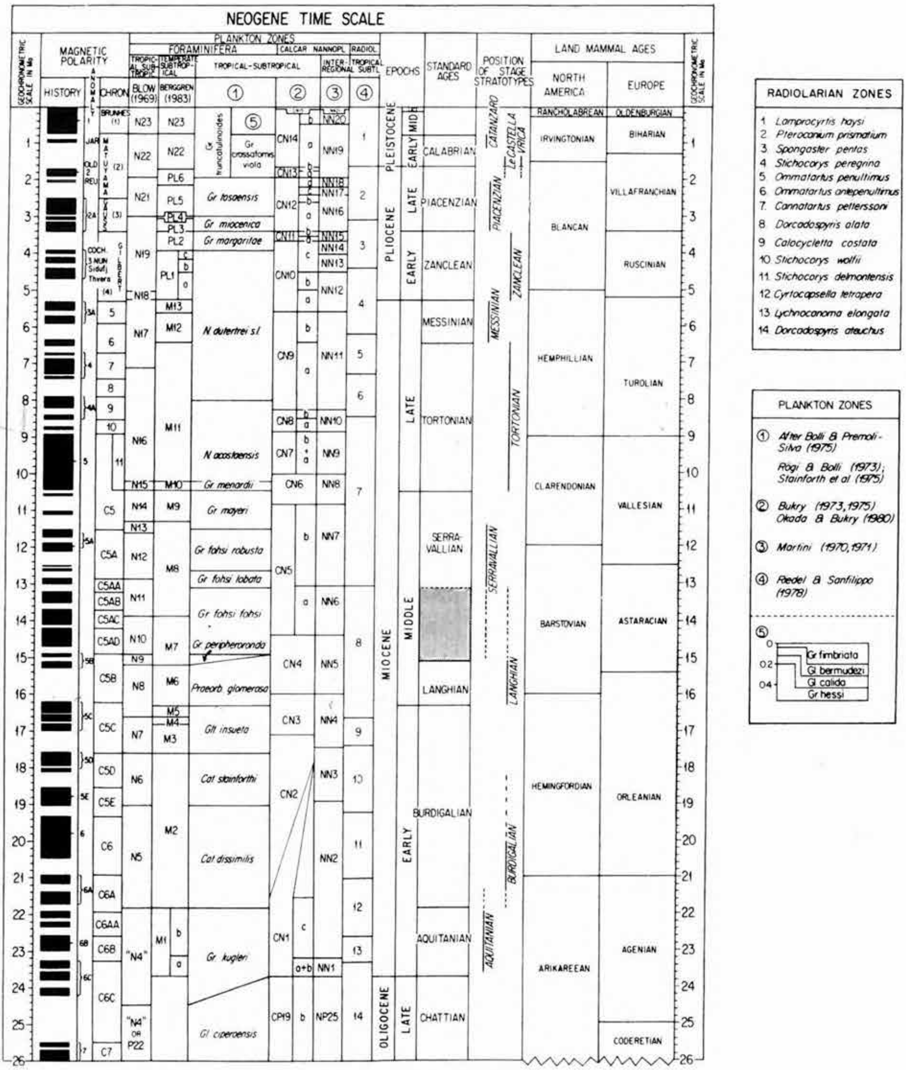


*Fig. A.* Neogene Geochronology and Chronostratigraphy by W. A. BERGGREN, D. V. KENT et J. A. VAN COUVERING (1985)

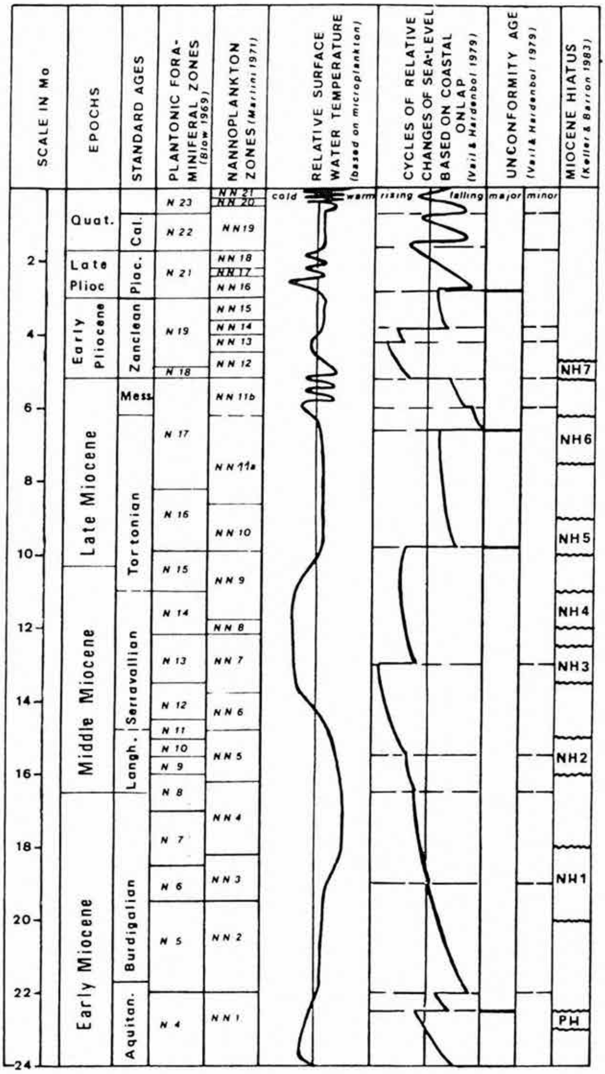


*Fig. B.* Neogene time-scale, surface water temperatures and changes of coastal onlap and Miocene hiatuses by MÜLLER (1985)

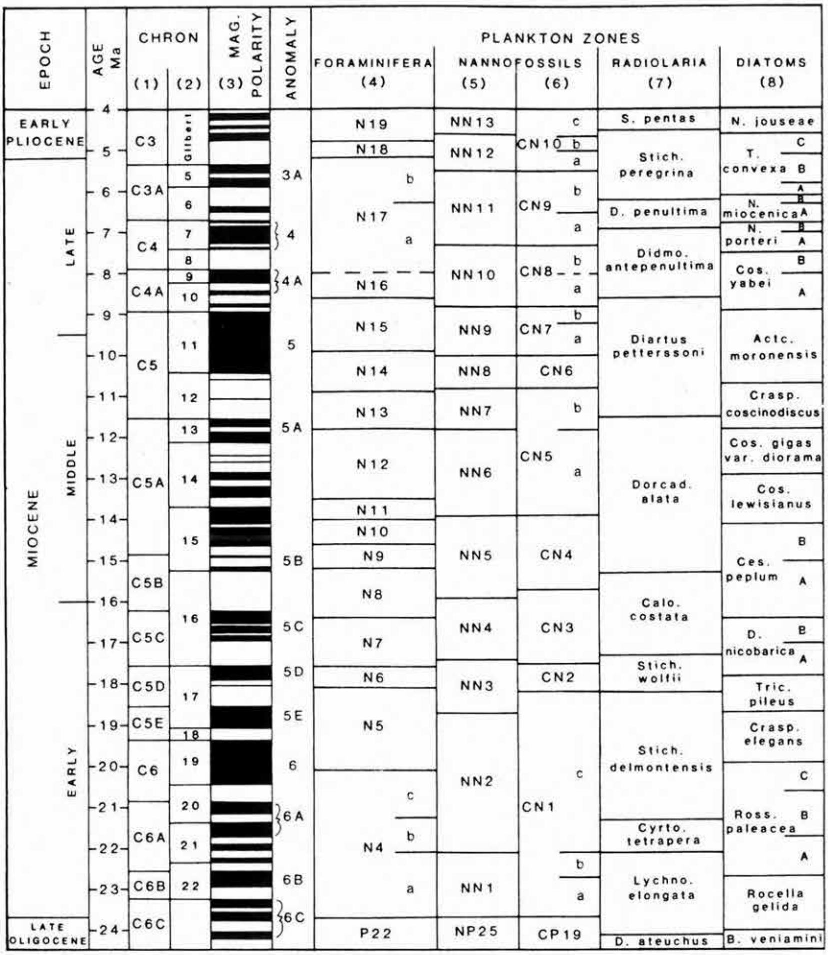


*Fig. C.* Miocene time scale by BARRON (1985)

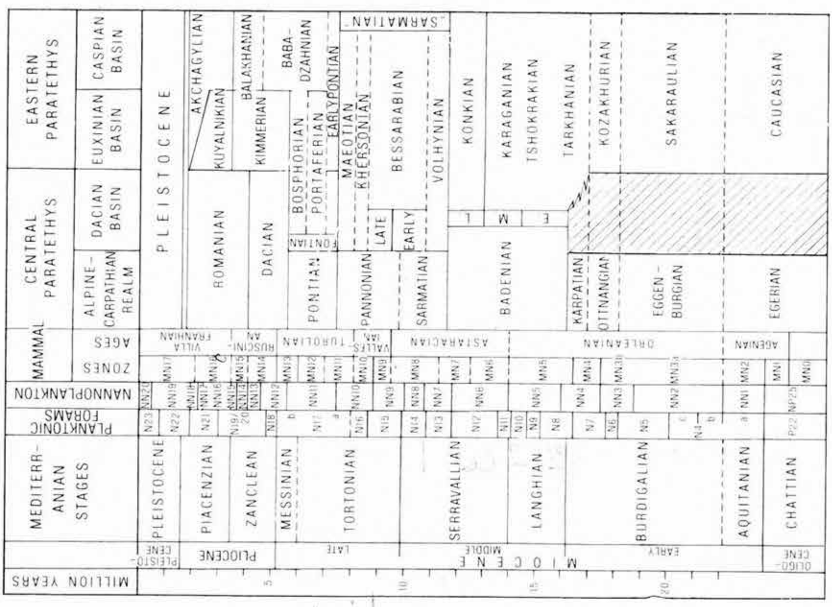


*Fig. D.* Tentative Neogene Mediterranean—Paratethys Correlation table

*1* Correlation of calcareous nannoplankton zonation—planktonic foraminifera zonation and radiometric ages according to C. MÜLLER (1984—see also Fig. B in this article), *2* radiometric ages of Central Paratethys stage boundaries according to D. VASS et al. (1985—Abstract volume of VIII. RCMNS Congress)

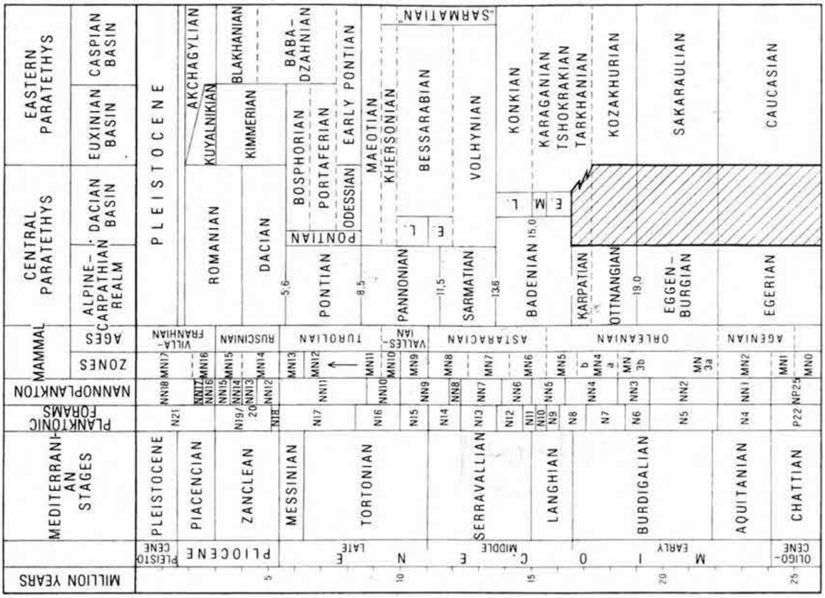


*Fig. E.* Tentative Neogene Mediterranean—Paratethys Correlation table

*1* Miocene correlation of calcareous nannoplankton zonation—planktonic foraminifera zonation and radiometric ages according to J. A. BARRON (1985—see also Fig. C in this article), *2* Pliocene correlation according to W. A. BERGGREN et al. (1985—see also Fig. A. in this article)

*Table 1*

**Mediterranean correlation key points**

| |
|---|
| Marine—continental correlations: |
| MN15 La Juliana = *Globorotalia crassaformis?* |
| late MN 13 La Alberca = N18? |
| MN13 Crevillente 6 = Messinian |
| MN12 Crevillente 4 = *Globorotalia conomiozea* |
| Crevillente 1—3 = N16—MN11 |
| early MN10 Kastellios = early N16 (reversed) |
| MN5 Chelas 2 = N9 |
| MN4 Chelas 1 and Quinto de Pombeja = N8 |
| MN3b Navigao = N7 |
| MN3a Lisboa, Univ. Catholica = NN2 |
| MN3a Maigen (Austria) = NN2/3 |
| MNO Cap Janet = below P/N-boundary |

*Table 2*

**Paratethys correlation key points**

| | |
|---|---|
| Apsheronian = NN17—18 upper | 0,95 ± 0,11 my |
| | 1.05 ± 0.12 my |
| Akchagylian = NN17—18 | 2.19 and 2.55 my |
| MN16 = Bevesti, Malusteni, Jaros—with palaeomagnetics | |
| early Kimmerian = NN11—12 | |
| late Maeotian = NN10 | |
| middle Maeotian = NN9, *Discoaster hamatus* | |
| Maeotian, with *Unio flabellatus*, no mollusc correlation to mammals | |
| Khersonian: Kalfa with molluscs | |
| Hipparion appearance = middle Bessarabian | |
| "Sarmatian" (sensu lato), abs. age 13.5—9.5 my | |
| early Maeotian with *Catinaster coalithus* | |
| early Sarmatian with *Catinaster coalithus* also | |
| Konkian = NN6/7 | |
| Tarkhanian = NN4 and Karpatian molluscs | |
| Caucasian = NP25—NN1 | |

*Table 3*

**Additional correlation points and problems**

*1* Pliocene—Pleistocene boundary according to the new boundary stratotype (AGUIRRE & PASINI, Episodes vol. 8, no. 2, 1985) at 1,6 m.y.

*2* Eastern Paratethys stage system according to NEVESSKAYA et al. (1984, Int. Geol. Congr.)

*3* Correlation of Eastern Paratethys partly by nannoplankton (SEMENENKO and LULIEVA, 1985, Abstracts; NOSOVSKY, 1985, Abstracts).

*4* Subdivision of Pontian in absolute time not prooved, only estimated. Correlation of Pontian to Late Tortonian (partly) according to VEKUA by means of ostracods (1985, Abstracts).

*5* Dacian/Romanian boundary according to ANDREESCU (Athens, 1981).

*6* Base of Pontian in MN11 by correlation of molluscs of the Eichkogel and Tihany localities (ČTYROKÝ, pers. comm. at the RCMNS meeting).

*7* Correlation of mammal zonation according to MEIN at the discussion meeting and MEIN (Bratislava and Athens), lower part compare to RÖGL and STEININGER (1983); additional correlations:

MN1 = Paulhiac = in Oligocene (RINGEADE, 1978)

MN6 = Devinska Nova Ves, sand hill (Neudorf) = Middle Badenian

MN7 = La Grive, La Grenatière = late NN6 (AGUILAR, 1982) = about Sarmatian according also to Steinheim—Nexing correlation

MN8 = Late Sarmatian s. str./Volhynian with molluscs at Comanești 1 (FERU et al., 1980)

MN9 = first *Hipparion* = Pannon C/D at Comanești 2 (FERU et al., 1982); = Gaiselberg, Pannon C (ZAPFE, 1949) = uppermost Pannon 8 at Hovorany (comm. of ČTYROKÝ); late Bessarabian at Nessebar (KOJUMDGIEVA, 1971)

MN10 = Pannonian E of Vösendorf

MN16/beginning of Villafranchian in *Gr. crassaformis* zone (ALBERDI and BONADONNA, Abstr. 1985).

*8* Position of the Langhian according to CITA and RYAN, N8 pp—N10 lower part (beginning of *Praeorbulina* to FAD *G. druryi*); see also "Activity reports and Proceedings of Bratislava RCMNS" nannoplankton zone NN4 to NN6, top according to RYAN et al. (1974) coinciding with top of NN5.

The uncorrect stratigraphic position of the Langhian within NN4 (HAQ, 1983) is without any explanation or reason, compare also the discussion of BANDET et al. (C. R. Acad. Sc. Paris, t. 299, sèr. II, no. 10, 1984, p. 651).

*9* Tortonian according to MARTINI (Bratislava, 1975) NN 9—11, boundary Serravallian/Tortonian at base of N15 (RYAN et al., 1974).

*10* Boundary Egerian/Eggenburgian at the $NN^1/_2$ boundary, no NN1 in the Eggenburgian. Caucasian must reach below the P/N—boundary, corresponding at least to late Egerian also by the occurrence of NP25.

*11* The lower part of Fig. E: N4—5 is according to KELLER (Micropaleont., vol. 26. p. 372, 1980). The correlation to the nannozonation does not correspond to any other correlation. N4 as total range zone of *Gr. kugleri* does not cover much of NN2 but correlates with NN1 and the upper part of NP25.

**J. P. AGUILAR** (letter dated December 3, 1985)

*I Mediterranean key point* (voir Table 1) corrections proposées:

*Chelas 2* (N9—N10) zone C2 (AGUILAR, 1982) probablement MN5. Pour le gisement de Chelas 2, un âge plus récent devrait être envisage, âge voisin de celui de Sansan (zone C3 AGUILAR, 1982) ou MN6.
*Chelas 1* entre N8 et N9 niveau à *Hispanotherium* zone C2 (AGUILAR, 1982) = MN5.
*Quinto de Pombeiro* N8 zone C1 (AGUILAR, 1982) = MN4b.
*Lisboa R2*. Quinto de Navigao entre N7 et N8 MN4a (MEIN, 1975, ANTUNES in 1982; REBEIRO et al. 1979). Pourquoi MN3b? Est-ce sur la corrélation donnée par AGUILAR (1982)?
*Lisboa Univ. Catolica* N4—N5 zone A5 (AGUILAR, 1982) = MN3a
*Cap Janet:* pas de relation entre P et N. Ce niveau est de l'Aquitanien parastratotypique (ALVINERIE, ANGLADA, CARALP et CATZGRAS, 1977).' Si la zone N4 est reconnue bien au-dessus, rien n'indique clairement que c'est la base de N4 puisque N3 ou P22 n'y sont pas reconnues dans ces formations très littorales.

*Eléments de corrélations supplémentaires:*

— Les Cévennes N4 zone A2 (AGUILAR, 1982), ce qui implique pour la zone MN1 un âge aquitanien.
— Beaulieu âge radiométrique 17.8 ± 0.5 M.A. (BAUBRON et al., 1975) nouvelle datation à paraître 17.5 ± 0.5 (AGUILAR, BAUDET, CLAUZON)
Ceci implique pour la partie supérieure de la zone B (AGUILAR, 1982) un âge de 17.5 M.A. environ.
— Veyran NN6 Partie supérieure de la zone C2 (AGUILAR, 1982) zone MN5 ou MN6??

*II Figure de corrélation*

Fig. D (corrélation MÜLLER):
— La limite Plaisancien—Zancléen est admise à —3.3 M.A., il faudrait donc déplacer la limite.
— La zone MN13 commence dans le Tortonien supérieur (zonation de MEIN, 1984).
— La zone MN5 est caractérisée par le gisement de Langenmoosen. Ce gisement est attribué au Karpatien d'après CICHA, FAHLBUSCH et FEJFAR, 1972, ainsi que dans le tableau de corrélation du volume "Corrélation du Néogène de la Paratéthys centrale" (Geological Survey Prague, 1975).
Dans le tableau d'après l'extension de la zone MN5, Langenmoosen serait du Badénien.
— L'Ottnangien est caractérisé par le gisement d'Orechov que l'on doit corréler avec la base de la zone C1 (AGUILAR, 1982) ou de la zone MN4, mais cette limite Ottnangien/Eggenburgien ne peut être fixée à 19 M.A., puisque le gisement de Beaulieu qui appartient à la partie supérieure de la zone B ou à la zone MN3b est un gisement plus ancien que celui d'Orechov, ayant un âge radiométrique de —17,5 ± 0,5 M.A.
— Pour la limite Aquitanien—Burdigalien, GOMINARD et al. (1985) proposent un âge de —20,6 M.A.

**J. AGUSTÍ:** Upper Miocene correlations in Eastern Spain

Two kinds of data are presented in this report. The first ones occur in four sections which permit the correlation between the continental and the marine scale. The sections are as follows (Fig. 1):
A) Casa del Acero. Fortuna basin (Murcia, SE of Spain).
B) La Hornera section. Fortuna basin. (Murcia, SE of Spain).
C) Molina de Segura section. Fortuna basin (Murcia, SE of Spain).
D) San Onofre section. Tortosa area (Tarragona, NE of Spain).
Palaeomagnetic data are also available from the Lower part of the last section.
The other data occur in a continental section with radiometric samples and biostratigraphic points (La Celia section, Murcia, SE of Spain)

A) *Casa del Acero section* (Figs. 1, 2, 3)

The Fortuna basin is an intramontainous basin filled by marine sediments ranging from the Tortonian I and II up to the Messinian (MONTENAT, 1977). A detailed study of the geology of this basin was done by SANTISTEBAN (1981). In the borders of the basin, the terminal Miocene passes into reef and evaporitic deposits. These ones are covered by deltaic and continental beds which sometimes include fossiliferous horizons with mammals. Three evaporitic groups were distinguished by SANTISTEBAN (op. cit).. The lower one overlies clays and marls of marine origin which contain *Globigerinoides elongatus* and *G. extremus*. *Globorotaria mediterranea* is also present in this lower evaporitic group. The continental beds of Casa del Acero, as well as those from la Hornera and Molina de Segura, were deposited over the Upper evaporitic group (Fig. 2). Casa del Acero has yielded a mammal association belonging to the MN12 zone (AGUSTÍ et al., 1981):

*Petenyella repenningi*, *Schizogalerix* sp. I, *Schizogalerix* sp. II, Echinosoricinae indet., *Hispanomys adroveri*, *Kowalskia meini*, *Parapodemus barbarae*, *Occitanomys adroveri*, *Valerimys turoliensis*, *Eliomys truci*, *Atlantoxerus* sp., *Hipparion concudense* ssp., *Metailurus* n. sp., Cervidae indet., *Tragoportax gaudryi gaudryi*, "*Mastodon*" sp.

Thus, levels belonging to the Middle Turolian (MN12) overlie marine beds from the Messinian.

B) *La Hornera section* (Figs. 1, 2, 3)

Like the Casa del Acero section, this sequence belongs to the third evaporitic group (above the horizons with *G. elongatus* and *G. extremus*). It shows a transition from green marls of marine origin to lacustrine beds, some of them with mammal remains. The fauna found at La Hornera is composed of the following elements:

*Apodemus gudrunae*, *Stephanomys ramblensis*, *Eliomys truci*, *Muscardinus vireti*, *Hispanomys* sp., *Prolagus* cf. *michauxi*. This association is typical of zone MN13 (Upper Turolian).

C) *Molina de Segura section* (Figs. 1, 2, 4)

The lower part of the sequence is formed by green and bluish marls with intercalations of gypsum and sand. These beds overlie the beds of the third evaporitic group with *G. elongatus* and *G. extremus*. Samples MSA 1 and 3 have yielded many valves of the ostracod species *Cyprideis pannonica*, possibly indicating a late Messinian age (ZACHARIASSE, oral com.). In the upper part of the section up to 11 fossiliferous sites with mammals have been recovered (from base to top):

*Molina de Segura-D: Occitanomys adroveri*, *Valerimys turoliensis*, *Stephanomys* cf. *ramblensis*, Suinae indet.

*Molina de Segura -1: Valerimys turoliensis*, *Stephanomys ramblensis*, *Hipparion* aff. *concudense*.

*Molina de Segura-E: Hipparion gromovae*, Cervidae indet., ?Antilopini indet.

*Molina de Segura-2: Prolagus* cf. *michauxi*, *Occitanomys* sp.

*Molina de Segura-3: Stephanomys ramblensis*, *Paraethomys miocaenicus*, *Hispanomys* sp.

*Molina de Segura-4:* aff. *Propotamochoerus* sp.

*Molina de Segura-6: Stephanomys ramblensis*, *Prolagus* cf. *michauxi*.

*Molina de Segura-7: Occitanomys* sp., *Paraethomys miocaenicus*.

*Molina de Segura-8: Stephanomys ramblensis*, *Paraethomys miocaenicus*, *Prolagus* cf. *michauxi*, Leporidae indet., *Hipparion* sp.

*Molina de Segura-9: Apodemus gudrunae*, *Paraethomys miocaenicus*, *Stephanomys ramblensis*, *Occitanomys* sp., *Hispanomys* sp., *Critcetus kormosi*, Cervidae indet.

*Molina de Segura-10: Stephanomys* cf. *margaritae*, *Paraethomys miocaenicus*.

The Molina de Segura horizons D and 1 could be situated at the base of the zone MN13 or in the top of the zone MN12 (because of the presence of *Valerimys turoliensis*). The horizons from Molina de Segura-E to M. de Segura-10 probably belong to the base of the Ruscinian, MN14, because of the presence of *Stephanomys* cf. *margaritae*.

D) *San Onofre section* (Figs. 1, 5)

It consists, from base to top, of

1) Grey marine marls.

2) Marls with thin sandstone beds.

3) Transitional sequence of yellow to brown clays.
4) Continental sequence of dark clays with hydromorphic horizons.
5) Limestones, occasionally travertinic.

Disconformably over the latter, the section continues with cemented breccias coming from indeterminate alluvial fans.

At the San Onofre-1 site of the continental sequence, a discrete mammal microfauna was recovered *(Apodemus jeanetti, Occitanomys* sp., *Muscardinus* aff. *vireti* and *Prolagus* cf. *michauxi)* indicating a MN14 zone or base of the MN15. In the lower part of the section a foraminifera fauna is composed of *Globigerinoides extremus, Globigerina decoraperta, Florilus bouenus, Elphidium crispum, Ammonia beccarii tepida* and *Bulimina* cf. *Fussiformis baccata* (see AGUIRRE et al.). This association indicates an age older than the *Globorotalia inflata* zone. The ostracod association corresponds to SISSING zone 9 (AGUIRRE et al., op. cit.) which partially overlaps the *G. crassaformis* zone. The palaeomagnetic imprints show a sequence which goes from normal to reverse and again to normal at the top of the marine sequence.

E) *La Celia section* (Figs. 1, 6, 7)

The main interest of this section consists in the existence of lava beds intercalated within fossiliferous deposits. One of these extrusive beds immediately above a lacustrine deposit has yielded the following mammal association (Los Gargantones):

*Parapodemus lugdunensis, Occitanomys sondaari, Prolagus crusafonti, Alilepus* cf. *turoliensis,* Cervidae indet., *Hipparion* aff. *concudense, Microstonyx* sp., *Tragoportax gaudryi crusafonti.*

This association is typical of MN11 zone (Lower Turolian). The volcanic episode from La Celia is also documented in the area by many pipes and intrusions which transsect the fluviatile and lacustrine beds. The age of the whole episode was established by NOBEL et al. (1981) by different methods, giving the following values:

| | |
|---|---|
| sanidine | $7.2 \pm 0.3$ |
| K-richterite | $7.6 \pm 0.2$ |
| apatite fission tr. | $7.2 \pm 1.4$ |

These values also coincide with other samples from Jumilla area (Las Cabras, Las Minas, Calasparra), the dates of which range from 7.2 to 7.6. Because of chemical particularities all the volcanic episode belongs to a single source (NOBEL, oral com.).

## Conclusions

*Conclusions:* According to the data presented, the following correlations are proposed:

*1)* The Lower Turolian (MN11) must be correlated with the Upper Tortonian (7.2 Ma.) according to the data from La Celia section.

*2)* The Middle Turolian (MN12) is correlated with the Messinian, since, at Casa del Acero, horizons of that MN zone overlie marine evaporitic deposits of this stage. Until now, the Middle Turolian was considered to be time-equivalent with the Upper Tortonian.

*3)* The Upper Turolian (MN13) must be considered equivalent to the extreme top of the Messinian, according to data from La Hornera and Molina de Segura sections.

*4)* The zone MN14 (Lower Ruscinian) is partially correlated with the *G. crassaformis* zone (Middle Pliocene).

In Fig. 4 by MÜLLER (1984) and Fig. 5 by BARRON (1985), both authors correlate the base of the Vallesian with the base of the Tortonian. But we know (after MEIN) that *Hipparion* (the Vallesian indicator) is present in Serravallian beds of the Rhone basin. So, the lower part (or, at least, part of the lower part) of the Vallesian must be correlated with the Upper Serravallian. On the other hand, it seems clear (section of Kastellios Hill, Greece, DE BRUIJN et al., 1971) that the Upper Vallesian may be correlated with the Lower Tortonian. So, the scheme of correlations between the marine and continental scales is as follows:

Lower Vallesian—Upper Serravallian
Upper Vallesian—Lower Tortonian
Lower Turolian—Upper Tortonian
Middle and Upper Turolian—Messinian
Lower Ruscinian—Middle Pliocene

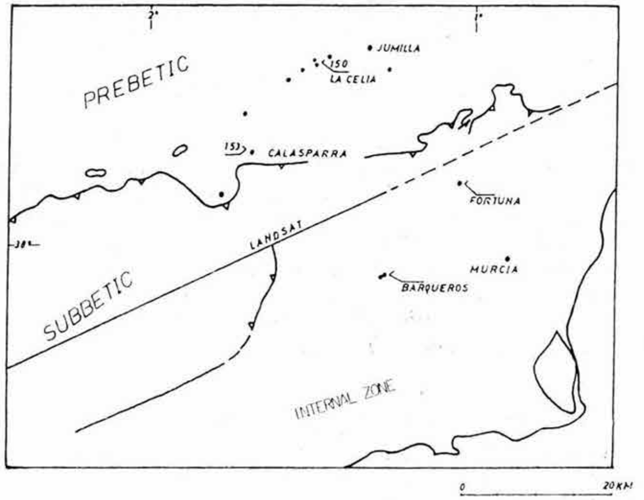


*Fig. 1.* Map of situation of the localities of SE Spain cited in the text (after AGUSTÍ et al., 1985)

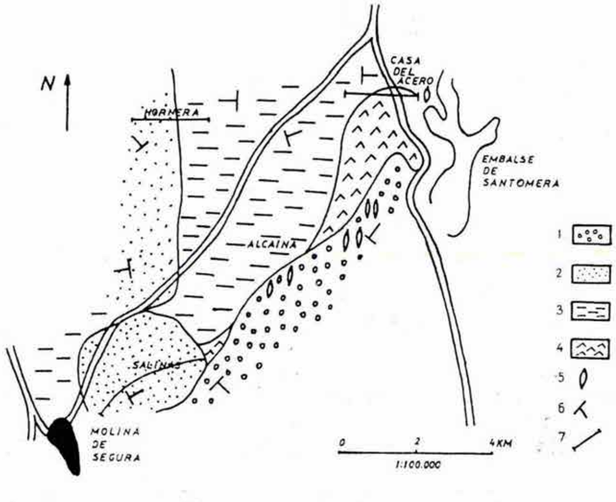


*Fig. 2.* Location of the sections in the Fortuna basin (after AGUSTÍ et al., 1985)
*1* Alluvial system, *2* deltaic facies, *3* marine clays and muds, *4* evaporitic facies, *5* reefs, *6* dip, *7* situation of the sections

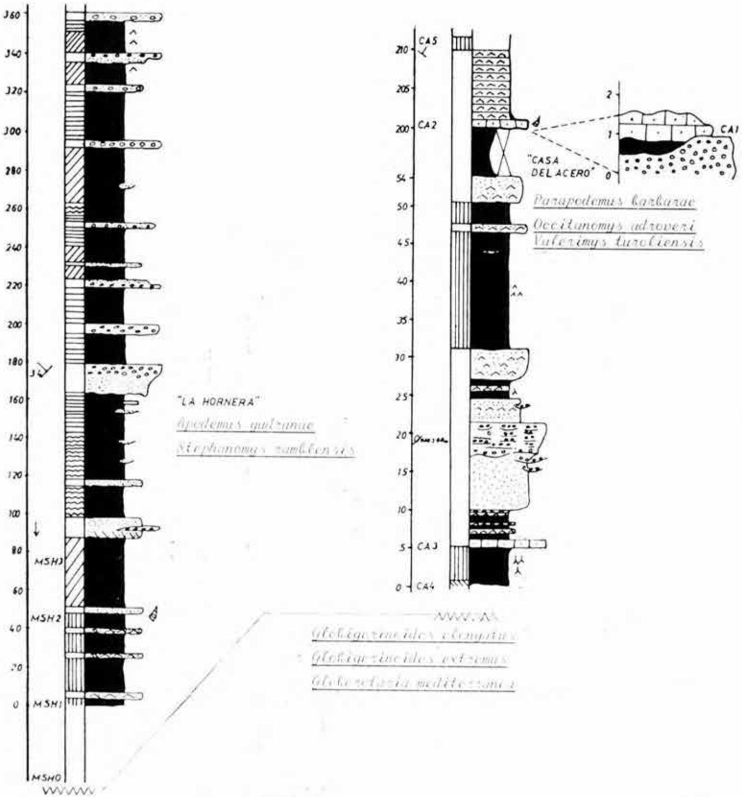


*Fig. 3.* Sections of Casa del Acero and La Hornera

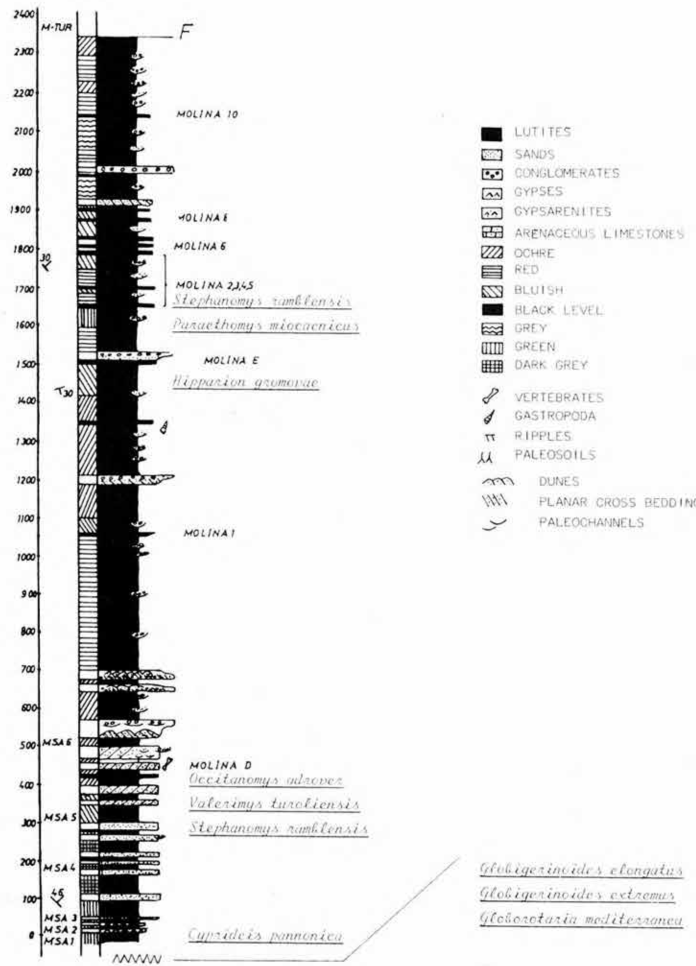


*Fig. 4.* Section of Molina de Segura (after AGUSTÍ et al., 1985)

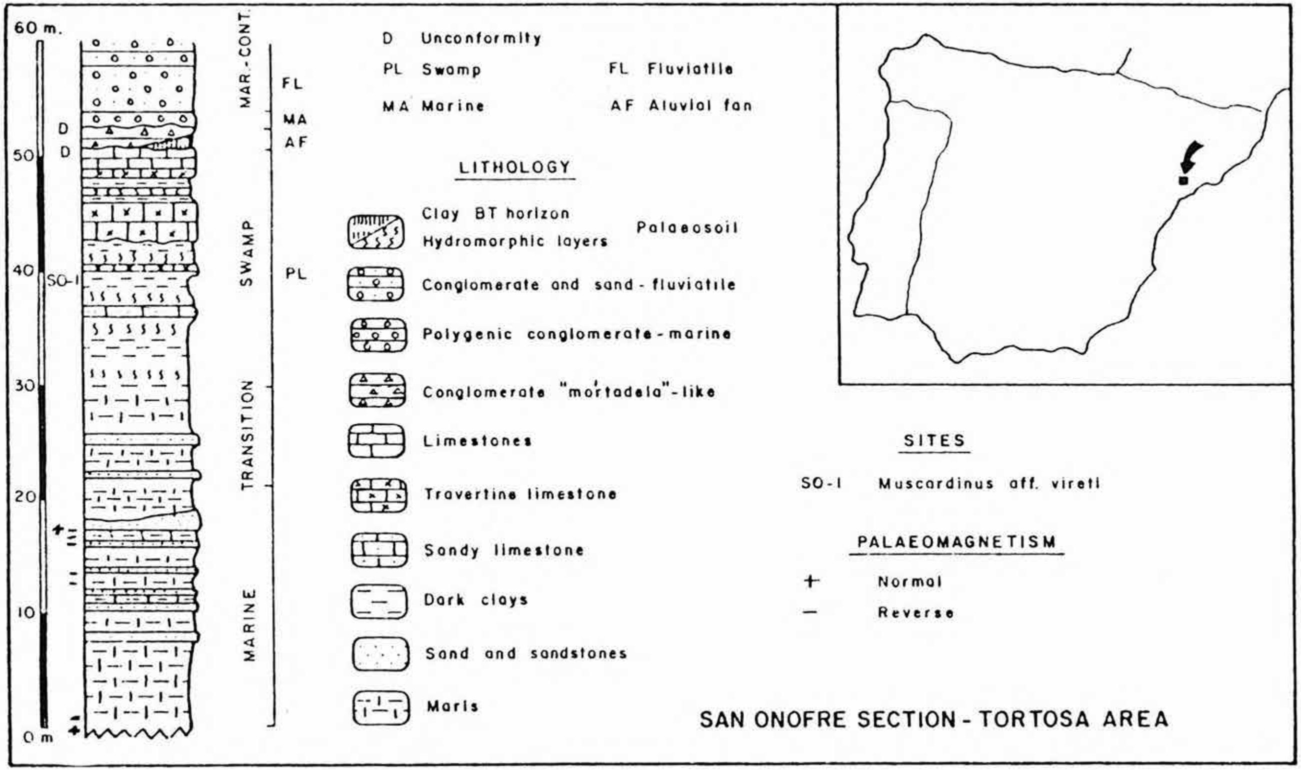


*Fig. 5.* Section of San Onofre (after AGUIRRE et al., op. cit., see also AGUSTÍ et al., 1983)

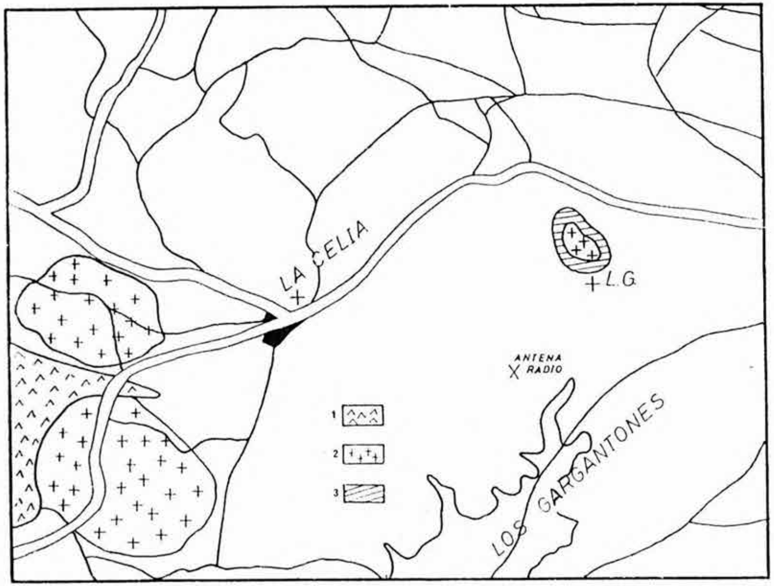


*Fig. 6.* Location of the sections from La Celia and Los Gargantones (after AGUSTÍ et al., 1985)

*1* Diapiric gypsum (Trias), *2* intrusion, *3* extrusive level

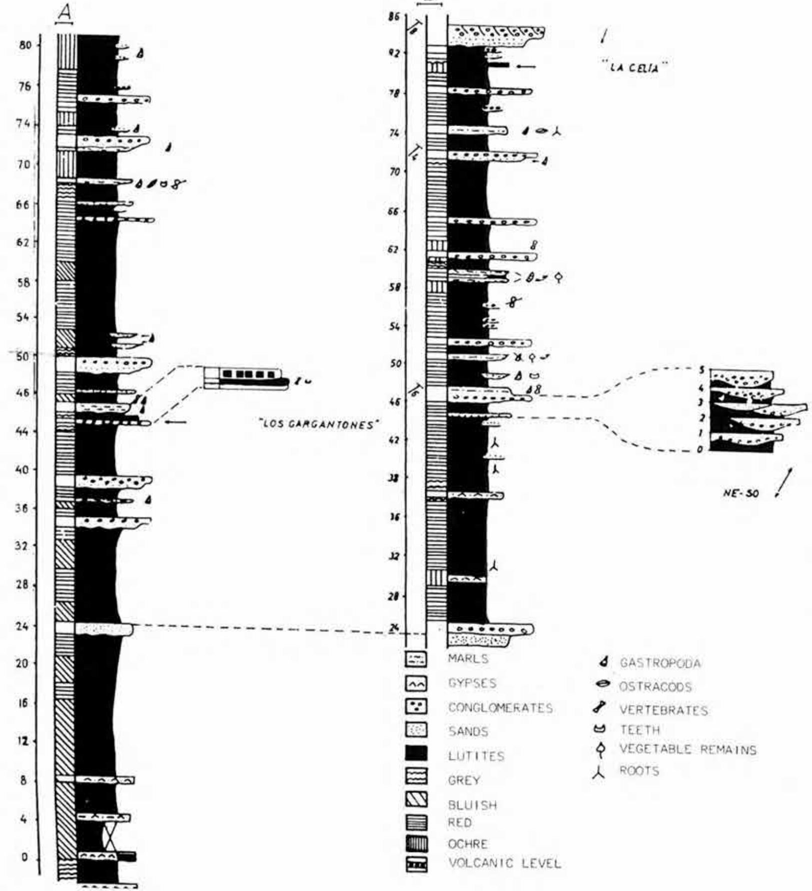


*Fig. 7.* Sections of Los Gargantones and La Cella (after AGUSTÍ et al., 1985)

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**M. T. ALBERDI** and **F. P. BONADONNA:** Results on Pliocene marine—continental correlations in Spain and Italy

The age of La Juliana deposit (Spain) is considered by MONTENAT and DE BRUIJN (1976) as Early Ruscinian and correlated with the Late—Middle Pliocene (younger than *Globorotalia crassaformis* zone). ("The Early Ruscinian thus seems to be coeval with the Late Pliocene, but a correlation with the middle Pliocene cannot be excluded" MONTENAT et DE BRUIJN, 1976, p. 255). After some research in the field, in our opinion, the stratigraphy of La Juliana deposits is not so clear as MONTENAT and DE BRUIJN maintain. Moreover, the deposit was completely removed by agricultural activities and it is now impossible to subject it to a geological—palaeontological revision. Furthermore, the definition of MN14 and MN15 (the rodent horizons of La Juliana) in Spain is rather confused. As can be seen in Fig. 1, in the Tortosa area (San Onofre section) the horizon called "MN14 or beginning of MN15" by AGUIRRE et al. (1982) is found in a swamp sediment overlying the *Globorotalia crassaformis* zone, i.e.MN16b (also note that the top of the *Globorotalia crassaformis* zone was found to be 2.1 MA old in Vrica section (Italy). On the other hand, in the La Gineta deposit (Jucar valley, Spain), MN15 underlies a conspicuous discordance (Ibero-Manchega I) which in turn underlies the Casas del Rincon series (Fig. 1) (ALBERDI and BONADONNA, 1985).

The series (50 m thick), situated in the Southern Spanish Meseta, is formed by carbonate layers of a marshy and lacustrine environment, and it is transgressive on other lacustrine series in which Ruscinian fauna was found. In the whole series close isotopic measurements were performed (LEONE, 1985) to determine the palaeoclimatic trend of the deposition epoch. From the oxygen isotopic composition of carbonate layers and fresh water gastropods it is possible to build a palaeoclimatic curve (Fig. 1); the comparison of this curve with THUNELL's curve (1979) for the Mediterranean palaeoclimate shows that the cold—warm (probably arid)—cold sequence of Rincon is mirrored on THUNELL's sequence, in which the first cold has an age of 3.1—3.2 Ma, the warm stage 2.7—2.6 Ma and the last cold 2.6—2.5 Ma. The Rincon 1 fauna is located in the second cold of the sequence while Rincon 2—3 just below the first one; the age of Middle Villafranchian fauna of Rincon 1 deposit is so fixed at 2.5—2.6 Ma.

In the Rincon series, rodents are only found in MN16a and MN16b (Early and Middle Villafranchian respectively, ALBERDI et al., 1982). Furthermore, the isotopic record of the marine clay sequence of San Onofre (*G. crassaformis* zone, also analyzed for palaeomagnetism) indicates that the age of San Onofre is probably intermediate between the Ibero—Manchega I discordance of La Gineta and the Rincon 2—3 horizon. Indeed, the palaeomagnetic sequence of San Onofre shows a short normal episode at the beginning, a long reversal episode in the central part of the

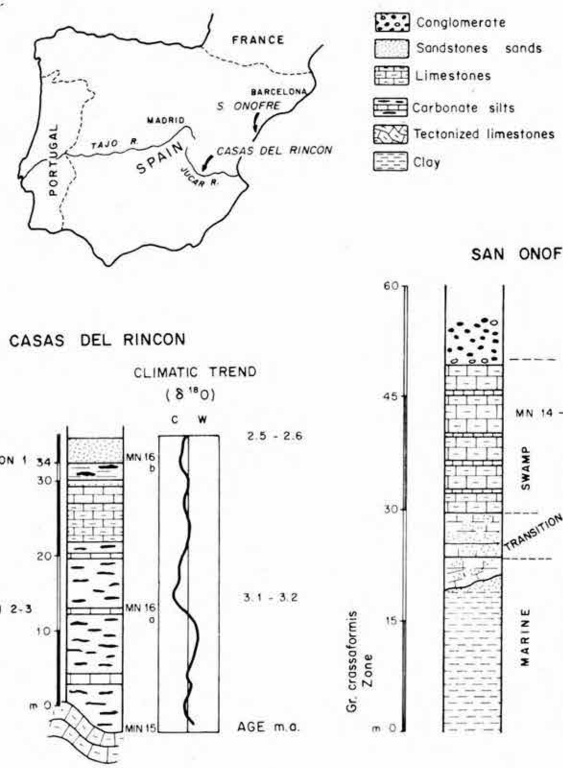
FRANCE
BARCELONA
S. ONOFRE
MADRID
PORTUGAL
TAJO R.
SPAIN
CASAS DEL RINCON
JUCAR R.
Conglomerate
Sandstones sands
Limestones
Carbonate silts
Tectonized limestones
Clay
SAN ONOFRE
CASAS DEL RINCON
CLIMATIC TREND
(δ 18O)
C W
RINCON 1
RINCON 2-3
MN 16 b
MN 16 a
MN 15
2.5 - 2.6
3.1 - 3.2
AGE m.a.
MN 14 — 15
SWAMP
TRANSITION
MARINE
PALEOMAGNETIC RESULT
3.07
3.17
Gr. crassaformis Zone

*Fig. 1.*

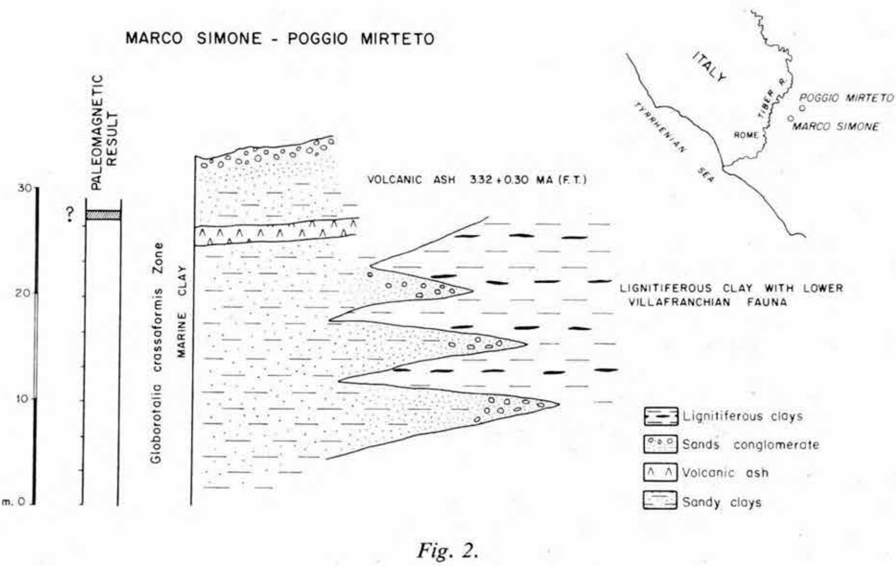
MARCO SIMONE - POGGIO MIRTETO
PALEOMAGNETIC RESULT
?
Globorotalia crassaformis Zone
MARINE CLAY
VOLCANIC ASH 3.32 + 0.30 MA (F.T.)
LIGNITIFEROUS CLAY WITH LOWER VILLAFRANCHIAN FAUNA
ITALY
TIBER R.
POGGIO MIRTETO
MARCO SIMONE
ROME
TYRRHENIAN SEA
Lignitiferous clays
Sands conglomerate
Volcanic ash
Sandy clays

*Fig. 2.*

series and a new probably normal episode at the end. This palaeomagnetic sequence probably corresponds to the Mammoth reversal subzone (3.07—3.17, McDOUGALL, 1979; 3.05—3.15 MANKINEN and DALRYMPLE, 1979) in good agreement with the inferred age of *G. crassaformis* zone and with the palaeoclimatic sequence.

For Poggio Mirteto (Rome), we have the following data (Fig. 2):

1) The lignitiferous clay of vertebrate deposits *(Tapirus arvernensis, Mastodon arvernensis)* are heteropical with Middle Pliocene marin clays (*Globorotalia crassaformis* zone);

2) in the marine clays there is a volcanic ash horizon dated by *K/Ar* and fission tracks methods: the most consistent age is $3.32 \pm 0.03$ Ma (ARIAS et al., 1981);

3) palaeomagnetic results on marine clays show only reversed magnetization, i.e. the series was sedimented during a reversal (ARIAS et al., 1976).

The combination of the above results shows that the Poggio Mirteto deposit is older than 3.0 Ma it may have been deposited during the last part of the Gilbert reversal zone (i.e. older than 3.41 Ma) or during the Mammoth reversal subzone (3.17—3. 0 Ma).

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**I. ANDREESCU** (letter dated December 12, 1985)

In my opinion I consider that the figures D and E are fitting pretty well most of the data (biochronologic, magnetostratigraphic, radiometric, paleogeographic etc) available till now. However, as concerns the Paratethys, I feel that these figures have a "weakness" that, between ourselves, could be essential in Western-Eastern Paratethys correlation. Namely, I have adopted the opinion according to which in the Western Paratethys the Pontian stage starts inside of MN11 zone. As you know, in the Eastern Paratethys both MN11 and MN12 zones are considered by many authors (see GABUNIA, 1979 etc) as belonging to the Maeotian stage. This assumption is maintained in spite of the fact that neither in Soviet territory nor in Romania the mammal remains of the MN12 zone have ever been found related directly to undisputable Maeotian mollusc faunas. On the other hand I am not convinced yet, that the Western Paratethys Vallesian and Turolian Mammal sites are being correctly assigned to an age or another. In this respect compare, for example:

a) Csákvár = Pannon B = Bessarabian; Kohfidish = Pannon C = Kersonian; Eichkogel, Podlesice = Pannon E1 = Maeotian (PAPP, 1975);

b) Vösendorf = Pannon E1; Eichkogel = Early Pontian; Polgárdi—Baltavár = Late Pontian (PAPP, 1978; Sofia);

c) Csákvár, Gaiselberg = Pannon C = Berislav, Eldar = Kersonian; Kohfidisch, Vösendorf = Pannon D = Taraclia, N Elizabetovka, Grebeniki, Bazaleti = Early Maeotian (GABUNIA, 1975); Vösendorf = Grebeniki = MN11 (GABUNIA, 1979);

d) Csákvár, Gaiselberg = Pannon B = Vösendorf = Pannon E; Polgárdi = Uppermost Pannon—Lowermost Pontian; Eichkogel, Kohfidisch = Early Pontian (STEININGER, 1975);

e) Csákvár, Vösendorf = MN10 = Sevastopol, Varnitza; Kohfidisch, Eichkogel = MN11 = = Grasulovo, Eldar, Berislav (MEIN, 1975);

f) Gaiselberg, Rudabánya = MN9 = Pannon C—D; Vösendorf = MN10 = Pannon E; Kohfidisch Eichkogel = MN11 = Early Pontian; Csákvár = MN12 etc (RABEDER, 1985).

It's amazing, isn't it? However, in my interpretation, I consider Vösendorf in the upper part of MN10 or lowermost part of MN11 = Late Pannonian = Late Maeotian; Eichkogel, Kohfidish if = Tihany (= Portaferrian), then they have to be assigned rather to the MN12 zone (lower part). This does not contradict the statement according to which the Lowermost Pontian is located inside of MN11 zone.

**P. ČTYROKÝ:** The stratigraphic position of the first *Hipparion* occurrence in Southern Moravia, Vienna Basin, Czechoslovakia.

According to JIŘIČEK (1985) and RABEDER (1985) the teeth of *Hipparion primigenium* were found in the Kyjov-coal seam in the lignite coal pit in Hovorany and Sardice. These two localities have surely the same stratigraphic position because they represent only two historic stages of a coal pit, changing from one to the another of the districts of the two villages—Hovorany and Sardice.

The mammal remains collected during the last 30 years from these localities are, besides *Hipparion*, unidentified, well-preserved mastodont teeth and other bone remains. They were found at the bottom of the Kyjov-coal seam which is the final coal-bearing member of the zone B of the Pannonian (according to PAPP's molluscan zonation). It was proved in the area of the Sardice coal mine in many cores that the top of the coal seam is overlain by the basal sandy member of zone C with a typical mollusc fauna (see ČTYROKÝ, 1975). From the Hovorany lignite is cited also *Dinotherium giganteum* KAUP, deposited in collection of Natural History Museum in Vienna (PIA-SICKENBERG, 1934).

The *Hipparion* remains from Hovorany and Sardice were deposited in the Moravian Oil Mines at Hodonín and were checked by A. PAPP personally during his visit at Hodonín in 1967. Unfortunately, this *Hipparion* material seems to be lost now. The mastodont remains are under study by O. FEJFAR (Praha).

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**E. KOJUMDGIEVA:** Tarkhanian—Tshokrakian—correlation problems

The subdivision and correlation of Tarkhanian and Tshokrakian are controversial. The two principal correlations are:

| | | | | |
|---|---|---|---|---|
| Tshokrakian | Up. | Kwalitian beds with Lutetia intermedia | Up. | Tshokrakian |
| | Low. | Zukian beds with typ. Tshokrakian fauna | Mid. | |
| Tarkhanian | Up. | Argunian beds with Spirialis | Low. | |
| | Mid. | Terian beds with Amussium denudatum | Up. | Tarkhanian |
| | Low. | Kuvinian (Kamyshlakian) beds with Foraminifers | Low. | |

This subdivision is mainly valid for argillaceous sediments. The correlation with limestones and sands is also controversial, e.g. the Gorian beds (Ostrea beds) are attributed by GONTCHAROVA and KVALIASHVILI to the Kuvinian beds, and by ANANIASHVILI to the Terian beds. The correlation with other realms requires complementary studies of planktonic foraminifers and calcareous nannoplankton.

*Karpatian.* The presence of NN5 zone in the Karpatian (also below the FAD Praeorbulina) must be verified.

*Khersonian—Maeotian.* The Khersonian/Maeotian boundary is at 9.8 Ma. (CHUMAKOV et al., 1984).

**F. MARINESCU** (letter dated February 27, 1986)

Commençons par les informations: par le magnétisme le Bessarabien supérieur et Khersonien semblent être vers 11.8 MA (s'accordant avec les 9–10 MA de la zone C du Pannonien et du Méotien inférieur). En plus n'oublie pas que les zones CD du Pannonien sont approximatives. Méotien inférieur et zone E sont approximativement Méotien supérieur.

Des données préliminaires du mangétisme donnent le Bosphorien (Pontien supérieur) vers les époches 5+6 et le Portaferien le 7.

La limite basale du Dacien semble passer vers 5.5–5.6 tandis que la base du Romanien correspond à l'épisode Cochiti (donc vers 3.9 MA).

Maintenant voilà le problème: il concerne la base du Badénien qui, conformément au tableau, correspond à 16.5 MA. Presque la même valeur a été obtenue aussi chez nous par le magnétisme, mais les mensurations ont été faites à la base des "marnes à globigérines" et du " tuf de Slănic " aux environs de Slănic Prahova, où dans la région la base du Badenien (d'après *Praeorbulina*) est à +300 m plus en bas. De même on ne peut pas oublier les 18.5 MA obtenus par BERGGREN pour le tuf de Dej, qui sont une réalité. D'ailleurs le problème est encore plus difficile (et maintenant voilà la question) en ce qui concerne cette limite dans les Carpates Orientales : elle (la limite basale du Badénien) passe à l'apparition, ou à l'acmée de *Praeorbulina =(Candorbulina?)*. S'il s'agit de l'apparition, alors les choses sont compliquées parce que aux plusieurs coupes cette apparition (par des rares exemplaires) est synchrone au maximum de *sicanus* (et les informations ne proviennent pas seulement de POPESCU , mais de plusieurs micropaléontologistes de chez nous). Donc ainsi même le Karpatien est mis en cause. Si l'on commence le Badénien par l'acmée de *Praeorbulina*, alors tout est O.K., cela correspond à environ la base des tufs (Dej, Slănic etc) et donc vers 16.5

MA, mais la limite du Badénien a été définie "mit dem Auftreten" donc... voilà le problème important pour la Paratéthys en générale et surtout pour les Carpates Orientales, qui ne peuvent pas être passée comme rien, tout comme le problème ne peut pas être escamoté.

**V. V. MENNER, M. A. PEVZNER** and **E. A. VANGENGEIM** (letter dated December 16, 1985)

I At present there are three stratigraphic levels to correlate the Central and Eastern Paratethys Neogene deposits:

1) the base of the Dacian = the base of the Kimmerian;

2) the base of the Sarmatian (s. SUESS) = the base of the Sarmatian s.l.;

3) the base of the Tarkhanian corresponds to the base of the Karpatian due to common malaco faunas and nannoplankton (GONCHAROVA, 1985; ANANIASHVILI, SAKHELASHVILI, 1984; MUSYLEV, PEVZNER, 1983; NOSOVSKY, BOGDANOVICH, 1984; NOSOVSKY et al., 1975; MINASHVILI, 1981, 1983).

According to palaeomagnetic data the base of the Tarkhanian is: 17.1 Ma. (PEVZNER and VANGENGEIM, 1985, Budapest Congress, Abstracts p. 461 ff). The radioisotopic dating of the Karpatian base is 17.5±0.5 Ma. (VASS, 1985, Budapest Congress, Report on activity of the RCMNS, p. 20 ff.).

II In The Eastern Paratethys the Pontian corresponds to Magnetic Polarity Epoch 6 and the end of Epoch 7. In the thickest Pontian sequence of the Taman peninsula the lower 9 m of 123 m only are normally magnetized, while the rest of the Pontian strata and the Kimmerian base (3 m) have reversed magnetization. According to the palaeomagnetic data the Pontian top is aged 6.1 Ma and its base — 6.8—6.9 Ma. So, the Pontian lasted not more than 0.8 Ma.

In case the Pontian shows reverse magnetization the deposits are to be correlated with Epoch 8 or with the lower part of Gilbert Epoch, its duration cannot exceed 0.6 Ma. For this, all our stratigraphic constructions must proceed from a short Pontian.

III The zonal scales demonstrated at the Congress differed from each other both in boundary datings for N and NN zones and in relations between N and NN zones themselves as well as in radiometric and paleomagnetic scales. Therefore on a general correlation scheme the boundaries between N and NN zones as well as MN zones must be omitted.

IV On Figures D and E the Babadzhanian horizon should correspond to the Portaferian and Bosphorian ones.

*Remarks to: Mediterranean key points* (see Table 1)

MN12 Crevillente 4 — *Globorotalia conomiozea*, this is the Messinian s.str. Therefore the Messinian should correspond not only to MN13 zone (Fig. 4 and 5) but partly to MN12 zone, as a minimum.

*Remarks to: Paratethys correlation points* (see Table 2)

The Aspheronian = parts of NN18 and NN19 zones;

The Akchagylian = NN16, NN17 and partly NN18 zones (the figure is correct);

The locality Kalfa (MN19 zone) coincides with the deposits with Bessarabian molluscs. *Catinaster coalithus* has been found in Upper Bessarabian (MINASHVILI, 1983).

The Konkian is corresponding to the uppermost part of NN5 and not younger than NN6. *Coronocyclus nitescens* (does not occur higher than NN6) has been found at the Konkian top and the Volhynian base (MINASHVILI, 1983).

The Caucasian equals the Aquitanian only but does not contain any specific zonal nannoplankton species (NOSOVSKY, 1984, and this article).

*Remarks to: Additional correlation points* (see Table 3)

The Tihany and the Eichkogel localities cover the beds with *Congeria rhomboidea* (BARTHA et al., 1971, pp. 37, 90—93), i.e. they should be assigned to the Pontian top but not its base.

There are no distinct criteria to draw boundaries between MN zones for the Turolian in the Paratethys. These zones should be better connected with narrow MN11 →MN13.

Attention should be paid to the fact that in Vösendorf dating beds on molluscs do not coincide with those on mammals: the latter are actually related to MN10 zone (i.e. the Sarmatian) but the presence of *Congeria balatonica* speaks that these deposits belong to the F—G—H—zones of PAPP (the Pontian), i.e. the mammal remains have been redeposited (see details in the letter of DRS. M. A. PEVZNER and E. A. VANGENGEIM to Prof. STEININGER from 9th October 1985; see this letter in this article and the reply by F. STEININGER).

**M. NOSOVSKY** (letter dated January 28, 1986)

Die Begründung der Korrelation des Neogen der Zentralen und Östlichen Paratethys ist in meinem Vortrag beim Kongress dargelegt, dessen englischen Text Sie besitzen (Budapest Congress, Abstracts p. 424 ff).

Was die Kaukasische Regionalstufe betrifft, so sind ihre Lage und ihr Alter in dem publizierten Vortrag des Kongresses von Athen (1979) und in dem oben erwähnten Vortrag von Budapest genau beleuchtet.

Deshalb sollte in den mir übersandten Figuren D und E die Kaukasische Regionalstufe nur dem oberen miozänen Teil der Eger-Regionalstufe im Sinne des Aquitan entsprechen.

Im gegebenen Fall bringe ich nicht nur meine persönliche Meinung zum Ausdruck, sondern den offiziell akzeptierten Beschluss von 1981 des Zwischen-behördlichen Stratigraphischen Kommitees der UdSSR über das untermiozäne Alter der Kaukasischen Regionalstufe.

**M. A. PEVZNER** and **E. A. VANGENGEIM** (letter to **F. F. STEININGER** dated October 9, 1985)

One of us has promised you in Budapest to write you in more detail about our considerations concerning the Vösendorf locality.

*1)* The presence of *Congeria balatonica*, *C. partschi* and *C. zsigmondyi* at Vösendorf are mentioned in the publications by PAPP (1951.: 113—117) and PAPP and THENIUS (1953.: 5—10). According to BARTHA (1971, see also Table 2 in PEVZNER, VANGENHEIM, 1985), *C. balatonica* is a form characteristic of the middle horizon of the Upper Pannonian (s. l.); *C. partschi* and *C. zsigmondyi* are only characteristic of the Lower Pannonian. Co-presence of these forms may be, apparently, explained by redeposition of the last two.

*2) C. balatonica* is of the same stratigraphic range as *Melanopsis fuchsi* which is characteristic of the Eichkogel locality. Thus, on the basis of the mollusc fauna, Vösendorf (with the Vallesian fauna of mammals) is of the same age as Eichkogel (with the Turolian mammal fauna).

*3)* Vösendorf mammalian fauna is quite correctly attributed to the MN10 zone of the Vallesian. The E zone of the Pannonian is correlated both by you and PAPP with the Upper Maeotian of the Eastern Paratethys. In the Maeotian of the Eastern Paratethys the mammalian fauna is of a Pikermy type (Chimishlia, Taraklia and other localities) and should be attributed to the Turolian. The boundary between the Vallesian and Turolian in the Eastern Paratethys corresponds to the boundary between the Sarmatian and Maeotian. On the basis of the mammalian fauna the E zone of the Pannonian should be attributed to the Sarmatian. You hardly agree with it. There is nothing for us but suggest that remains of mammals in Vösendorf are redeposited.

*4)* Data on Vösendorf indicate just redeposition:

a) presence of pebbles and concretions from Pannonian conglomerates;

b) rolled mollusc shells;

c) rolled bones of mammals (PAPP, 1951.: 113—117; PAPP and THENIUS, 1953.: 5—10; Miozän $M_6$, Pannonien, 1985.: 187, 190).

The rich faunal complex of Vösendorf suggests, probably, a near transportation during redeposition.

Thus we have no doubt that the Vösendorf mammalian fauna belongs to the MN10 zone of the Vallesian, but its belonging to the E zone of the Pannonian is rather doubtful (i.e. we believe that in zone E this fauna is not in situ).

**F. F. STEININGER:** Remarks to the letter of PEVZNER and VANGENGEIM (October 9, 1986) and the letter of MENNER et al. (December 16, 1986).

At the Budapest Congress PEVZNER and VANGENGEIM (see: Budapest Congress Abstracts, 1985, pp. 461—462) presented a new correlation chart for the Eastern and Central Paratethys and the Mediterranean stages. In this contribution the correlation of the Late Miocene Eastern and Central Paratethys stages differs extremely from the results obtained so far. Therefore this contribution in the letters of MENNER et al. (letter dated December 16, 1985) and PEVZNER and VANGENGEIM (dated October 9, 1985) is published above.

PEVZNER and VANGENGEIM (Budapest Abstracts, p. 462) correlate the Eastern Paratethys Pontian with the entire Pannonian of the Central Paratethys respectively and they state, by their correlation table, that a gap exists in the Central Paratethys between the Sarmatian (sensu SUESS) and the Pannonian resp. the Pontian. This gap would span about 4 million years in their correlation. That would mean there are no equivalents of upper Bessarabian, Chersonian and Maeotian known in the Central Paratethys and there are no mammal faunas of zone MN10 and MN11 in situ. There is no need to discuss this miscorrelation at length—since it seems to be based on palaeomagnetic and radiometric data only, disregarding all other stratigraphic evidence. The evidence for the equivalents of late Bessarabian, Chersonian and Maeotian in the Central Paratethys have been published recently in: PAPP et al., 1985: $M_6$—Pannonien (Slavonien and Serbien). — Chronostrat. and Neostrat., 7. Budapest (Akad. Kiadó). However, since most of this miscorrelation by PEVZNER and VANGENGEIM seems to be caused by an erroneous interpretation of the biostratigraphic principles of PAPP's Pannonian zonation and the Vösendorf locality, I will try to summarize these points very briefly:

(1) PAPP's Pannonian biozonation (1951; 1953; PAPP et al. eds. 1985), as far as one can jugde by his published work, is based on the evolution of different phylogenetic lineages (Melanopsids, Congeriids and partly Limnocardiids) and the biozones (A—to—E) represent the acmes in the evolution of these different taxa. In each lineage, there seems to be evidence for a gradual transition between the different taxa within one given phylogenetic line and therefore scarce appearances are explained and evident already before the acme of the taxon and also of course still after the acme of the taxon. The Pannonian biozones A to E are defined therefore only by the acme of the evolution of the typical assemblage of the zone. PAPP's biozonation which is based on molluscs was duplicated and refined by ostracods, this biozonation based on ostracods was treated lately extensively by JIŘIČEK (1985 in PAPP et al. (eds. 1985).

(2) *Vösendorf:* PEVZNER and VANGENGEIM state directly and indirectly in their letter (see above) that because of the scarce record of *Congeria balatonica* (which has its acme in the Pontian) in Vösendorf, the Vösendorf locality belongs to the Pontian and is therefore of the same age as the Eichkogel locality (which is according to mammals MN11 Turolian in age). The MN10 (Vallesian) mammal fauna from Vösendorf locality is, in their opinion, redeposited. However, according to PEVZNER's and VANGENGEIM's correlation table, also the exceptionally rich and typical MN11 mammal faunas of Eichkogel (a lacustrine lake deposit) and Kohfidisch (a fissure filling) have to be redeposited, since the gap shown spans also the MN11 zone in the Central Paratethys! This also means that the MN10 and MN11 mammal faunas known from Czechoslovakia, Hungary, Roumania and elsewhere in the Central Paratethys are, according to this correlation chart, redeposited! In this relation it will be interesting to see how the radiometric ages between 11.8 Ma and 8.4 Ma coming from biostratigraphically well-dated Pannonian horizons are explained (see

VASS, 1985 in PAPP et al. 1985 and VASS, 1985 p. 23.—Report of RCMNS Working-Groups, Budapest Congress) and the palaeomagnetic dates coming from the Hungarian Pannonian (see Budapest Abstract volume).

The Vösendorf section shows in general (for details see PAPP, 1951: 113 ff; PAPP and THENIUS, 1954: 3 ff and PAPP, 1985: p. 187 ff in PAPP et al. 1985): greenish marls at the base

— followed by 0.15 up to 1.5 m sand, rich in fossils, "transgressively" overlying the greenish marls

— followed, respectively passing into greyish sandy marls up to 12 meters thick, rich in fossils.

The greenish marls at the base, the sandy horizon and the overlying greyish marls contain a typical and rich ostracod fauna indicating PAPP's Zone E and, according to JIŘIČEK's ostracod zonation, they belong to the upper part of zone E ($E_2$ to $E_3$) (see PAPP and THENIUS, 1954 p. 25; JIŘIČEK, 1985. 378 ff in PAPP et al. 1985).

No macrofossils were recovered from the greenish marls. The sandy horizon is rich in plants (fruit and wood remains), ostracods, molluscs, fish-, reptile- and mammal-remains (see PAPP et THENIUS, 1954; revised faunal lists in PAPP et al. 1984). *Congeria balatonica* is recorded as a rare element, respectively, as a very rare element from the base of this horizon only (see PAPP and THENIUS, 1954, p. 13). PAPP (1951, p. 115) considers that *Congeria balatonica* could have originated already in Zone D.

The greyish sandy marls on top of the section contain a rich leaf flora, ostracods and molluscs. It is essential to note that the bivalves—*Congeria, Limnocardium*—are frequently double-valved and in "living position". The sandy horizon, with its "transgressive" character is explained by PAPP as a sort of near-shore lumachelle, cutting channels and moulds into the underlying greenish marls where fossils were trapped. No wonder that lots of fossils are worn in this horizon, e.g. the larger melanopsids, congeriids and mammal remains. However, there is no indication at all that the mammal remains should have been redeposited through time.

The geological situation of Vösendorf is fixed by several drill hole sections in the near-surroundings. There is also a complete section between Zone E at the base of the Eichkogel hill and the top of this hill which is made up of lake deposits (limestones)—the type locality for the MN11 Eichkogel mammal fauna.

For the literature, see PAPP et al., 1985 cited above.

**D. VASS** (letter dated December 1, 1985)

May I turn your attention to the most recent modifications of the Paratethys Neogene radiometric time scale: For other details, see the article of D. VASS, I. REPČOK, K. BALOGH et J. HALMAI in this volume:

1) *The numeric age of the Pannonian/Pontian respectively the Maeotioan/Pontian boundary:*

My present opinion is that a more realistic age of the base of Pontian stratotype in the Euxino—Caspian region is about 7.0 Ma. The numeric age suggested for the above-mentioned boundary by ANDREESCU (1981) is too old. It seems that the Pontian of the Dacic basin (ANDREESCU) and of the Euxinic—Caspian region—where the Pontian was originally described in the Odessa region—do not have the same volume. Because of these differences there are discrepancies in palaeomagnetic records of the Pontian in Dacic basin and the Pontian of the Kertch and Taman Peninsula (SEMENENKO and PEVZNER, 1981; PEVZNER, 1985; see Abstracts volume). I am more inclined to accept the opinion of SEMENENKO and PEVZNER and the still existing radiometric dates of the Pontian show that the Pannonian/Pontian boundary should be about $7.0 \pm 0.2$ Ma. Now we plan a borehole for the magnetostratigraphic investigation of a "Pontian" section in the Danube lowland (Bratislava—Komárno) to verify what really represents the Pontian in the West Carpathians. But this is a song for the future.

2) The numeric age of Sarmatian (s. SUESS)/Pannonian boundary is shifted back to 11.0 Ma. The majority of radiometric ages of volcanic rocks closely related to this boundary support this this numeric age of 11.0 Ma.

3) The numeric age of Eggenburgian estimated in our radiometric time scale is supported by a F.T. age of 20.5 ±0.9 Ma from the Upper Krosno Beds in the Silesian tectonic unit of outer Carpathians. The dated tuff is found in a sequence of Eggenburgian age (NOWAK et al., 1985—Abstracts), but some other dates mentioned by NOWAK et al., especially the Badenian dates, are unrealistic.

4) The Badenian/Sarmatian boundary is near the boundary NN6/NN7 zones, or within the zone NN7 (R. LEHOTAYOVÁ, 1982; N. MÉSZÁROS, 1985: Abstracts, top of Konkian). According to BARRON (1985) the NN6/NN7 boundary is in chron 13—app. 11.9 Ma; according to BERGGREN et al. (1985) in chron CS 5 AA—13 Ma. The radiometric age of the Badenian/Sarmatian boundary 13.6 ±0.2 Ma, is not in accordance with either BARRON's or with BERGGREN et al.'s correlations, but the numeric age of 13.6 Ma is supported by 6 dates concerning the Upper Badenian and by 26 dates concerning the Lower Sarmatian.

5) Palaeomagnetic investigations on the Hajnačka section with a mammal fauna of MN16 (OPDYKE, paper given at the Budapest Congress) shows a normal polarity and is considered as chron 3 or Gauss (2.48—3.40 Ma). The mammal fauna of Hajnačka was recovered from the strata considered as a maar-fill. Another maar or diatreme, that of Hajnačka Castle Hill, near the fauna locality is cut by a basalt dike, the radiometric age of which was dated at 2.75—0.44 Ma (isochron age, BALOGH, MIHALIKOVA, VASS, 1981).

6) Some additional comments on the Pannonian/Pontian boundary: If the Late Maeotian really corresponds to NN10 zone, then BARRON's correlation of the top of NN10 zone with lower part of the chron 7—ca 7.3 Ma is close to the base of Pontian numerically calibrated at 7.0 ±0.2 Ma but BERGGREN et al.'s correlation of NN10 zone with the lower half of the chron 9 and chron 10 is not in accordance with these results.

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# REVISED RADIOMETRIC TIME-SCALE FOR THE CENTRAL PARATETHYAN NEOGENE

by

D. VASS, I. REPČOK, K. BALOGH and J. HALMAI

At the 6th Congress of RCMNS (Bratislava, 1975) and, later, at the 25th International Geological Congress (Sydney, 1976) a radiometric time scale for the Central Paratethyan Neogene was presented (D. VASS et al., 1975; D. VASS and G. P. BAGDASARYAN, 1978). During the past 10 years many new radiometric dates of volcanic rocks from the Paratethys realm appeared. On the basis of new datings, the authors have tried to complement the radiometric time-scale of the Central Paratethyan Neogene and to add more precision to it.

A reliable radiometric age of the Egerian is still missing. New and reliable age data of the Eggenburgian, Karpatian, Upper Miocene (the Pannonian and the Pontian) and Pliocene were obtained. The new data concerning the Badenian and the Sarmatian are more or less conformable to the older ones and they only add precision and detail to the existing numeric scale.

*Eggenburgian*

Up to the present only two radiometric datings of a volcanic rock from biostratigraphically identified Eggenburgian deposits have been done. The rhyodacite tuff near the village Lipovany (S Slovakia, ČSSR) was dated by the F. T. method (I. REPČOK) as 20.6±0.5 Ma. The Eggenburgian age of the formation from which the dated tuff comes (Filaková Fm., Lipovany Member) is proved by molluscs (A. ONDREJIČKOVÁ, 1972), by Foraminifera (*Globigerinoides primordius*, *Uvigerina bononiensis primiformis*, *U. parkeri breviformis*, V. KANTOROVA's unpublished data) and by calcareous nannoplankton of the zone NN2 (including the index form *Discoaster druggi*, R. LEHOTAYOVÁ, 1984 and unpublished data).

The tuff layer inside the Upper Krosno formation of the Silesian succession (tectonic unit of the outer West Carpathians in Poland) was dated by F. T. method as 20.5±0.9 Ma. According to W. NOWAK et al. (1985), the dated tuff is Lower Burdigalian, i.e. Eggenburgian, in age.

*Eggenburgian—Ottnangian*

In S Slovakia and Hungary a rhyodacite—rhyolite tuff horizon (Gyulakeszi Rhyolite Tuff Formation or Lower Rhyolite Tuff in Hungary) is at the Eggenburgian/Ottnangian boundary. It is biostratigraphically well defined, representing the lowermost part of the Ottnangian stage. In the past the tuffs were dated by the *K/Ar* method applied to whole rock. The radiometric ages showed a great variation (19.9—21.9 Ma) and a great error. New and more precise dating were carried out on minerals separated from the rock. The ages obtained are younger and we consider them as closer to the actual age of the tuffs.

The age of a tuff from the vicinity of the village Lipovany (the tuff lies above the biostratigraphically and radiometrically dated Eggenburgian) is 20.1±0.4 Ma (F. T. method, I. REPČOK). Rhyolite tuffs, floodtuffs and andesites in the Mecsek Mts (Hungary), according to the biostratigraphy proposed by E. NAGY (1969) and M. SÜTŐ-SZENTAI (1983), are closed to the Eggenburgian—Ottnangian boundary. The radiometric age of tuffs from the Váralja-quarry and Szászvár, Szekernye valley (Hungary) is 19.5±1.4 Ma and 19.6±1.9 Ma (G. HÁMOR et al., 1981). The age of the andesite coming from the borehole Komló-170 is 19.5±0.9 Ma (K. BALOGH, unpubl.). The tuff coming from the Upper Krosno of Skole succession (tectonic unit of the Outer West Carpathians in Poland) is probably Ottnangian in age. The radiometric age of the tuff is 18.4±1.0 Ma and the tuff should be, according to W. NOWAK et al. (1985), Upper Burdigalian in age (i.e. Ottnangian?). Because of the uncertainty of the biostratigraphic age record we do not consider the age as a key datum.

*Karpatian*

Three radiometric dates concerning the Karpatian are mentioned in a D. VASS—G. P. BAGDASARYAN (1978) paper (19.4±20.7 Ma). They seem to be unreliable. Rhyodacite tuffs from Fót and Tar (N Hungary, uppermost Karpatian) (M. HAJÓS, 1968; J. Halmai, 1981) were dated as 16.2±0.6 Ma and 16.4±1.1 Ma, respectively (G. HÁMOR et al., 1981), and the average age of the Tar Dacite Tuff Formation (Middle Rhyolite Tuff) corresponds to the uppermost Karpatian, being quite close to the Karpatian—Badenian boundary (G. HÁMOR and Á. JÁMBOR, 1971), is 16.4±0.8 Ma (G. HÁMOR et al., 1981).

F. RÖGL and F. STEININGER (1983) suggested a numeric age of 17.5 Ma for the Ottnangian/Karpatian boundary. Unfortunately, this numeric age could not be proved by radiometric datings because reliable radiometric ages from the Lower Karpatian as well as the Ottnangian are still missing.

*Lower Badenian (Moravian)*

The radiometric age of the Lower Badenian was documented by 9 dates (15.6*—16.9* Ma: D. VASS—G. P. BAGDASARYAN, 1978). New datings: Andesite volcanoclastics at the village Kamenica at Hron (S Slovakia, ČSSR) intercalated with marine sediments and containing Lower Badenian fauna (fauna and calcareous nannoplankton described by R. LEHOTAYOVÁ, E. BRESTENSKÁ, D. VASS, A. ONDREJIČKOVÁ in borehole K—5 Salka in A. PAPP et al., 1978, p. 175, 181—184) were dated by the K/Ar method as 15.6±1.2 Ma (G. P. BAGDASARYAN, unpubl.) and by the F. T. method as 15.7±1.4 Ma and 16.1±0.4 Ma. From the borehole GK—3 near village Horné Rykyncice (S Slovakia, ČSSR) andesite volcanoclastics were dated by the F. T. method as 16.3±0.2 Ma (I. REPČOK, 1981). The Lower Badenian age is proved by Foraminifera (R. LEHOTAYOVÁ, E. BRESTENSKÁ in V. KONEČNY et al., 1983).

On the southern periphery of the Krupiná plateau (S Slovakia, ČSSR) andesite volcanoclastics of the Vinica formation, Lower Badenian in age (D. VASS et al., 1979), were dated by the F. T. method. The age is 16.5±0.6 Ma and 16.8±0.8 Ma (I. REPČOK).

In the Mátra Mts, at Gyöngyössolymos (N Hungary), the rhyolite is younger than lowermost Badenian (M. HAJÓS, 1968, GY. VARGA et al., 1975, J. HALMAI, 1981).

* K/Ar ages recalculated by constants: $\lambda_K = 0.581 \times 10^{-10} y^{-1}$; $\lambda_\beta = 4.96 \times 10^{-10} y^{-1}$; $^{40}K = 0.01167\%$.

The average *K/Ar* age measured on biotite and whole rock samples is 15.9±0.5 Ma (K. BALOGH, 1984).

All new datings agree with the older ones and support the numeric calibration of Karpatian/Badenian boundary at 16.5±0.6 Ma as suggested by D. VASS and G. P. BAGDASARYAN (1978).

Reliable radiometric dates from the Middle Badenian (Wielician) are still missing. A preliminary F. T. datum—the age of rhyolite tuffite from Poziom I—WT, from Wieliczka (Poland)—is of 15.0±3.5 Ma. The tuffite was sampled by E. LUCZKOWSKÁ and dated by J. KRÁL (Bratislava, pers. comm.).

*Upper Badenian (Kosovian)*

The radiometric age of the Upper Badenian is based on 5 datations (from 14.3* to 15.4* Ma: D. VASS—G. P. BAGDASARYAN, 1978). A new dating comes from Hungary. The rhyolite tuff near Bántapuszta (Transdanubia) has been dated. The tuff is interbedded in the lower part of the Upper Badenian (J. KÓKAI, GY. RAINCSÁK, 1983). The *K/Ar* age of the tuff is 15.0±0.4 Ma (K. BALOGH, 1984).

*Sarmatian*

The Sarmatian (in the sense of E. SUESS, 1866) was dated by 24 radiometric age data (11.5*—13.9* Ma) and three dates came from rocks considered as Upper Sarmatian—Lower Pannonian (11.0*—11.44* Ma: D. VASS, G. P. BAGDASARYAN, 1978).

Two datings carried out in 1968 seem to have basic data for the Lower Sarmatian: The andesites at Horša and Brhlovce (near the town of Levice, Southern Slovakia, ČSSR) were dated both as 13.2±0.5* Ma (G. P. BAGDASARYAN et al., 1968). Both dated andesites are covered with Lower Sarmatian sediments, and their lateral equivalents rest on the Upper Badenian.

In the Štiavnické vrchy Mts and the Phronský Inovec Mts (Central Slovakia, ČSSR) andesite ignimbrites of the Drastvica Formation have been dated. In the upper part of the Drastvica Formation there are pelitic intercalations containing Lower Sarmatian fauna (M. VAŇOVÁ, E. BRESTENSKÁ in V. KONEČNÝ et al., 1983. p. 101—102). The average of 15 datings by the F. T. method is 13.3±0.2 Ma (I. REPČOK, 1981 and unpublished data). West of the town Zvolen, andesites of the Breznica complex interbedded with Lower Sarmatian sediments (V. Konečný et al., 1983, p. 105) were dated by the F. T. method. The four obtained age data vary between 12.2±0.4 Ma and 12.7±0.5 Ma (I. REPČOK).

In the Zsámbék basin (Hungary, W of Budapest) boreholes Budajenő 3 and Perbál 6 penetrated a rhyolite tuff layer in the lower third part of the Sarmatian (Á. JÁMBOR, 1976; CS. RAVASZ, 1978). The average *K/ar* age on biotites is 13.7±0.5 Ma (G. HÁMOR et al., 1981; K. BALOGH, 1984).

The Lower Sarmatian of Moldavia, USSR, was dated by the F. T. method. Obtained dates: 13.6±0.89, 13.54±0.89, 13.75±1.08 Ma (S. S. GANZEJ, 1984, J. S. TCHUMAKOV et al., 1984). S. S. GANZEJ dated the Upper Volhynian (Middle Sarmatian in sens of SUESS, 1866) of Kerch peninsula, USSR: 12.24±0.97 Ma.

Along the Soviet—Hungarian border rhyolite tuff and rhyolite production ended near the end of the Sarmatian (L. KULCSÁR, 1968). *K/Ar* ages fall in the range of 11.0—11.3 Ma (K. BALOGH et al., 1983b). Since very little rejuvenation cannot be excluded, these data may be used for the approximation of age of the Sarmatian—Pannonian boundary.

In the Štiavnica Mts and the Kremnica Mts (Central Slovakia, ČSSR) rhyolites and tuffs of the Jastrabie Formation were dated by the F. T. method. According to palynological data the Jastrabie Formation is Upper Sarmatian—Lower Pannonian (V. KONEČNÝ et al., 1983). The resulting 11 age data vary between 10.9±0.5 Ma and 12.3±0.1 Ma (I. REPČOK, 1981 and unpublished data).

*Pannonian*

Several new results have been added to the seven radiometric ages concerning the Pannonian (7.8*—10.9* Ma D. VASS and G. P. BAGDASARYAN, 1978) and the biostratigraphic control of the newly dated rocks is better than that of the previous ones.

In NE Hungary rhyolite tuff production ended near the Sarmatian—Pannonian boundary (T. ZELENKA, 1964; GY. RADÓCZ, 1969; Á. JÁMBOR, 1971). Therefore the *K/Ar* dates of 10.7—11.0 Ma measured on alunite crystals from veins in the uppermost Sarmatian tuff correspond to the Pannonian (K. BALOGH et al., 1983b). On the other hand the Lower Pannonian dacite tuff from borehole Nagykozár 2 (S Hungary) yielded an older age of 11.6±0.5 Ma.

Basalts from boreholes Kecel 1, 2 and Kiskunhalas—Ny 3 (S Hungary) are accompanied by marls with a Middle and Upper Pannonian fauna (B. CSEREPES MESZÉNA, 1978). The radiometric age of basalts is 8.47±0.77 Ma, 8.13±0.71 Ma and 9.61±0.38 Ma respectively (K. BALOGH et al., 1983a).

An andesite agglomerate from Valea Poiana, Ignis Mts, Roumania, is in a formation containing Middle — Upper Malvensian fossils. The radiometric age of an andesite fragment is 11.08* Ma (O. EDELSTEIN et al., 1977).

From the equivalent of the Pannonian (Khersonian) from the Kerch and Taman peninsulas, USSR, the following F. T. were obtained: 11.18±0.74 (lower ash level), 10.6±0.75 (middle ash level), and 5 data fell in the range of 9.45—10.20 Ma (upper ash level). The age of 7.14±0.58 Ma comes from the Meotian (Upper Pannonian), (S. S. GANZEJ, 1984).

*Pontian*

In the previous radiometric time scale the Pontian was not calibrated. In the past few years several Pontian volcanic rocks were radiometrically dated.

A basalt lava flow at Podrečany (near the town of Lučenec, S Slovakia) is in the Poltár Formation which is Pontian in age according to palynological data (E. PLANDEROVÁ in lit.). The basalt was dated by two laboratories. The isochronous *K/Ar* age is 6.19±0.43 Ma (K. BALOGH et al., 1981), the analytical *K/Ar* age being 7.15±0.23 Ma (J. KANTOR and V. WIEGEROVÁ, 1981).

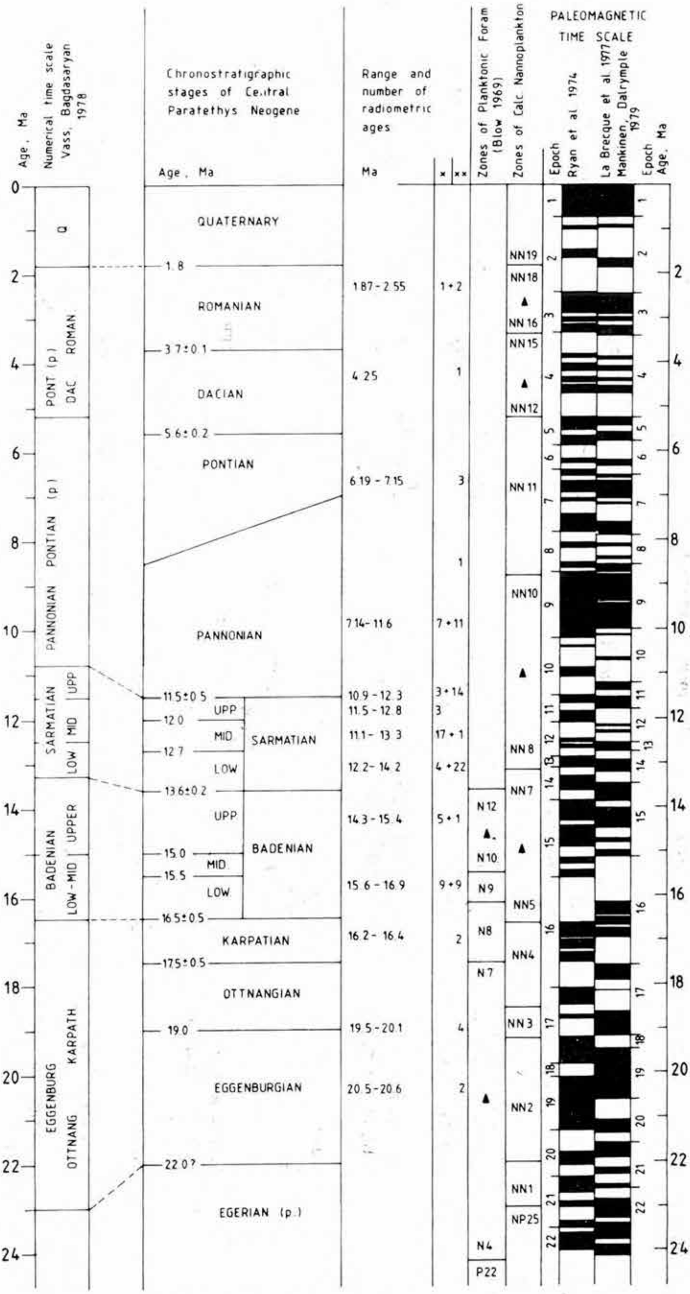
The Pontian of the USSR was dated by the F. T. method as 7.07±0.6 Ma (S. S. GANZEJ, 1984).

The Meotian—Pontian basalt from the valley of the river Liachva (Geogia, USSR) was radiometrically dated as 8.0 Ma (A. T. ASLANYAN et al., 1982).

*Dacian*

In the previous radiometric time scale the Dacian was not calibrated.

The basalt from the vicinity of Pula (Hungary), according to the biostratigraphic control (Á. JÁMBOR et al., 1981), poured out at the end of the Congeria balatonica zone. Its *K/Ar* age is 4.25±0.17 Ma (K. BALOGH et al., 1983a).




× Ages taken from Vass, Bagdasaryan, 1978
×× Ages in this paper

*Fig. 1.* Radiometric time-scale for the Central Paratethyan Neogene

**Review of the basic radiometric dates of Paratethyan Neogene time-scale**
(measured between 1976 and 1984)

*Table 1*

| Stratigraphic unit | Site | Rock | Method (Mineral) | Radiometric age (Ma) | Analyst | Published |
|---|---|---|---|---|---|---|
| Eggenburgian | Lipovany, ČSSR | rhyodacite tuff | F.T. (biotite) | 20.6 ± 0.5 | REPČOK | |
| | S Poland | tuff | F.T. | 20.5 ± 0.9 | | NOWAK et al, 1985 |
| Upper Eggenburgian | Lipovany, ČSSR | rhyodacite tuff | F.T. (biotite) | 20.1 ± 0.4 | REPČOK | |
| Lower Ottnangian | Komló 170, Hungary | andesite | K/Ar (w.r.) | 19.5 ± 0.9 | BALOGH et al. | unpublished |
| | Váralja, Hungary | rhyolite tuff | K/Ar (biotite) | 19.5 ± 1.4 | BALOGH et al. | HÁMOR et al., 1979 |
| | Szászvár, Hungary | rhyolite tuff | K/Ar (biotite) | 19.6 ± 1.9 | BALOGH et al. | HÁMOR et al., 1979 |
| Karpatian | Fót, Hungary | rhyodacite tuff | K/Ar (biotite) | 16.2 ± 0.6 | BALOGH et al. | HÁMOR et al., 1979 |
| | Tar, Hungary | rhyodacite tuff | K/Ar (biotite) | 16.4 ± 1.1 | BALOGH et al. | HÁMOR et al., 1979 |
| Lower Badenian | Kamenica at Hron, ČSSR | andesite volcanoclasts | K/Ar (w.r.) | 15.6 ± 1.2 | BAGDASARYAN | unpublished |
| | | canoclasts | F.T. (hornblende) | 15.7 ± 1.4 | REPČOK | REPČOK, 1981 |
| | | | (biotite) | 16.1 ± 0.4 | REPČOK | REPČOK, 1981 |
| | GK 3 900 m | andesite | F.T. (hornblende) | 16.6 ± 0.3 | REPČOK | unpublished |
| | Horné Rykyncice, ČSSR | volcanoclasts | (biotite) | 16.3 ± 0.2 | REPČOK | REPČOK, 1981 |
| | Sirákov, ČSSR | andesite volcanoclasts | F.T. (hornblende) | 16.8 ± 0.8 | REPČOK | REPČOK, 1981 |
| | Hrusov, ČSSR | andesite volcanoclasts | F.T. (hornblende) | 16.5 ± 0.6 | REPČOK | REPČOK, 1981 |
| | Gyöngyössolymos, Hungary | rhyolite | K/Ar (w.r. biotite) | 15.9 ± 0.5 | BALOGH et al. | BALOGH, 1984 |
| Upper Badenian | Bántapuszta, Hungary | rhyolite tuff | K/Ar (biotite) | 15.0 ± 0.4 | BALOGH et al. | BALOGH, 1984 |
| Lower Sarmatian | W Zvolen, ČSSR | andesite | F.T. (biotite) | 12.2 ± 12.7 (4 data) | REPČOK | |

*Table 1 continued*

| | | | | | | |
|---|---|---|---|---|---|---|
| | Štiavnica and Pohron Inovec Mts, ČSSR | andesite igminbrites | F.T. (biotite) (hornblende) | 13.2±0.2 | REPČOK | REPČOK, 1981 and unpublished |
| | Moldavia, USSR | volc. tuff | F.T. | 13.6±0.89 | GANZEJ | TCHUMAKOV et al., 1984 |
| | Moldavia, USSR | volc. tuff | F.T. | 13.54±0.89 | GANZEJ | TCHUMAKOV et al., 1984 |
| | Moldavia. USSR | volc. ash | F.T. | 13.75±1.08 | GANZEJ | TCHUMAKOV et al., 1984 |
| | Horsa, ČSSR | andesite | K/Ar (w.r.) | 13.2±0.5 | BAGDASARYAN | BAGDASARYAN et al., 1968 |
| | Brhlovce, ČSSR | andesite | K/Ar (w.r.) | 13.2±0.5 | BAGDASARYAN | BAGDASARYAN et al. |
| | Budajenő 3, Hungary | rhyolite tuff | K/Ar (biotite) | 14.2±1.3 | BALOGH et al. | BALOGH, 1984 |
| Middle Sarmatian | Perbál 6, Hungary<br>Kerch, USSR | rhyolite tuff<br>? | K/Ar (biotite)<br>F.T. | 14.2±1.3<br>12.2±0.97 | BALOGH et al.<br>GANZEJ | BALOGH, 1984 |
| Upper Sarmatian | Gelénes 1, Hungary | rhyolite tuff | K/Ar (biotite) | 11.0±0.6 | BALOGH et al. | BALOGH et al. 1983b |
| | Barabás 1, Hungary | rhyolite | K/Ar (w.r.) | 11.3±0.5 | BALOGH et al. | BALOGH et al., 1983b |
| | Barabás 1, Hungary | ignispumice | K/Ar (w.r.) | 11.2±0.5 | BALOGH et al. | BALOGH et al., 1983b |
| Upper Sarmatian—Lower Pannonian | Štiavnica, ČSSR<br>Kremnica Mts, ČSSR | rhyolite | F.T. (biotite, glass) | 10.9—12.3 (11 data) | REPČOK<br>REPČOK | REPČOK, 1981 unpublished |
| Lower Pannonian | Szerencs Hills, Hungary | alunite veins | K/Ar | 10.7—11.0 (3 data) | BALOGH et al. | BALOGH et al., 1983b |
| | Nagykozár 2, Hungary | dacitic tuff | K/Ar (biotite) | 11.6±0.6 | BALOGH et al. | BALOGH et al., 1984 |
| Middle Pannonian | Valea Poiana Ignis Mts, Roumania | andesite agglomerate | K/Ar | 11.08±0.5 | SOROIU | EDELSTEIN et al. |

*Table 1 continued*

| Stratigraphic unit | Site | Rock | Method (Mineral) | Radiometric age (Ma) | Analyst | Published |
|---|---|---|---|---|---|---|
| Khersonian (= Middle Pannonian) | Kerch, USSR | volc. ash | F.T. | 9.45 ± 0.88 | Ganzej | Tschumakov et al., 1984 |
| | Kerch, USSR | volc. ash. | F.T. | 9.59 ± 0.59 | Ganzej | Tschumakov et al. |
| | Taman, USSR | volc. ash | F.T. | 10.6 ± 0.75 | Ganzej | Tschumakov et al. |
| | Taman, USSR | volc. ash | F.T. | 11.18 ± 0.74 | Ganzej | Tschumakov et al. |
| Middle—Upper Pannonian | Kecel 1, Hungary | basalt | K/Ar (w.r.) | 8.47 ± 0.77 | Balogh et al. | Balogh et al., 1983a |
| | Kecel 2, Hungary | basalt | K/Ar (w.r.) | 8.13 ± 0.71 | Balogh et al., | Balogh et al., 1983a |
| | Kiskunhalas-Ny 3, Hungary | basalt | K/Ar (w.r. and fractions) | 9.61 ± 0.38 | Balogh et al. | Balogh et al., 1983a |
| Upper Meotian (= Upper Pannonian) | Kesamen, USSR | volc. ash | F.T. | 7.14 ± 0.58 | Ganzej | Ganzej, 1984 |
| Meotian—Pontian | Liachva Riv. | basalt | K/Ar (w.r.) | 8.0 | Rubinstein | Aslanjan et al., 1982 |
| | Kobystan, USSR | volc. ash | F.T. | 7.07 ± 0.6 | Ganzej | Ganzej, 1984 |
| | Podrečany, ČSSR | basalt | K/Ar (w.r. isochron) | 6.19 ± 0.43 | Balogh et al. | Balogh et al, 1981 |
| | Podrečany, ČSSR | basalt | K/Ar (w.r.) | 7.15 ± 0.6 | Kántor | Kántor, Wiegerová, 1981 |
| Upper Pannonian (uppermost Congeria balatonica z.) | Pula, Hungary | basalt | K/Ar (w.r. isochron) | 4.25 ± 0.17 | Balogh et al. | Balogh et al., 1983a |
| Romanian | | | | | | |
| Lower Akschagylian | Kvarab, USSR | volc. ash | F.T. | 2.55 ± 0.20 | Ganzej | Ganzej, 1984 |
| Upper Akschagylian | Kvarab, USSR | volc. ash | F.T. | 1.87 ± 0.15 | Ganzej | |

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# MIGRATION OF MIOCENE ECHINOIDS BETWEEN THE WEST INDO-PACIFIC AND THE MEDITERRANEAN REGIONS

by

M. S. M. ALI and O. H. CHERIF

*Introduction.* Since the earliest phases of the development of modern geologic sciences, the study of the migration of fossil fauna has been considered as one of the most important tools to decipher palaeogeographic conditions.

This paper deals with the study of the distribution of Early and Middle Miocene echinoids in the Eastern Mediterranean and the West Indo-Pacific regions as an attempt to understand the connection between those two main provinces and the mechanism of migration of the considered fauna.

In order to reach the proposed aim, the following points must be taken into consideration:

*1* The palaeogeographic situation (distribution and extention of lands and seas) of the considered region should be determined. It is done by reviewing the available literature on plate movements during the Miocene in the area comprising Africa, the eastern Mediterranean lands and India.

*2* The migration of marine organisms is influenced by many factors, which can be summarized as factors depending on the nature of the organism itself (its life habit, its life cycle, its ecological valency, etc...) and on the environmental conditions under which the organism lives and their suitability to help migration (distribution of lands and seas in which the organism is prone to migrate, existence of currents, existence of barrier to migration, etc...).

*3* By reviewing previous literature care must be taken to use a unified scheme for defining the species, which are often differently evaluated and named by different authors. In general, it has been observed that many ancient authors tend to split too much taxonomic units. For example more than 139 species of *Clypeaster* have been recorded in the Middle Miocene of the Mediterranean region (ALI, 1983). This shows that a great effort should be undertaken to revise the taxonomy of echinoids in this part of the world. In the present work only the forms showing striking similarities in the different localities investigated have been taxonomically revised. Only publications with illustrations have been taken into consideration.

In the following paragraphs the first two mentioned points regarding the factors influencing the migration of marine organisms are dealt with briefly before presenting the discussion on the distribution of the echinoid faunas in the various Mediterranean and West Indo-Pacific occurrences and its implications regarding the migration mechanism of these faunas.

**Palaeogeography of the Eastern Mediterranean and the West Indo-Pacific regions in Middle Miocene time**

The maps of figure (1), show roughly the relative position of the various continental masses, marine bodies and sea ways during the Middle Miocene in the Eastern Mediterranean.

FELL (1967), explained the direction of the main marine currents acting in the Indo-Pacific. It is reasonable to suppose that the same causes which produce the movements of major marine currents in the seas on the northern and southern sides of the equator were also acting in the same way during the Miocene times. These forces creat currents rotating in a clockwise direction and sweeping the equator from east to west in the northern hemisphere. Thus, during the Miocene, the main warm southern currents were comming from India towards Africa. The reverse movement being probably lost in the continental mass of the Arabian Peninsula and Asia.

**Distribution of Early and Middle Miocene echinoids in the Mediterranean and in the coasts of the Indian Ocean**

It is interesting before our discussion to give some informations about the stratigraphy of the Miocene successions which have yielded the echinoids in the localities chosen in the present study of the West Indo-Pacific region.

The most notable works of the Miocene echinoids in Somalia are the works of STEFANINI (1932) and SOCIN (1942, 1956). The faunas of this country have been collected from Hafun Peninsula, the most eastern point of Africa, coasts of N–E Somalia on the Indian Ocean. The Hafun Series which has yielded the echinoid faunas should therefore now be regarded as being of Middle Miocene age (EAMES and SAVAGE, 1975) and not Burdigalian as mentioned by SOCIN (1942 and 1956).

In Zanzibar, the Miocene echinoid faunas have been collected and described by STOCKLEY (1927), from Chake Chake Beds of the Pemba Series in Pemba Island. He assigned a Lower Miocene age to this series. EAMES and SAVAGE (1975), considered it to be Middle Miocene on the basis of Foraminifera.

According to KING (1953), the echinoid faunas of Zululand, northeast of South Africa, is contemporaneous with that from the Chake Chake Beds of the Pemba Island. Its age therefore should be Middle Miocene not Burdigalian as it was cleared by KING (1953).

The Dam Formation crops out near the coast of the Arabian Gulf in northeastern Saudi Arabia. This formation is Middle Miocene according to POWER et al. (1966). But KIER (1972), and ADAMS et al. (1983), considered it as Burdigalian.

Table (1), shows the distribution of the echinoid species found in both the Mediterranean and West Indo-Pacific regions during the Miocene in the investigated localities. These species and their distribution have been recognized from published data on echinoids.

The Middle Miocene echinoid forms of East Africa and the Arabian Gulf region were compared to these from the Early and Middle Miocene in the Mediterranean. As it is known that the connection between the Indian Ocean and the Mediterranean was severed in the Early Miocene (Burdigalian) time (ALI, 1983 and ADAMS et al., 1983), it is not likely to find in sediments of this age species that would be common to both areas (EAMES and SAVAGE, 1975). This means that the Early Miocene echinoids of the Mediterranean show a great deal of endemism.

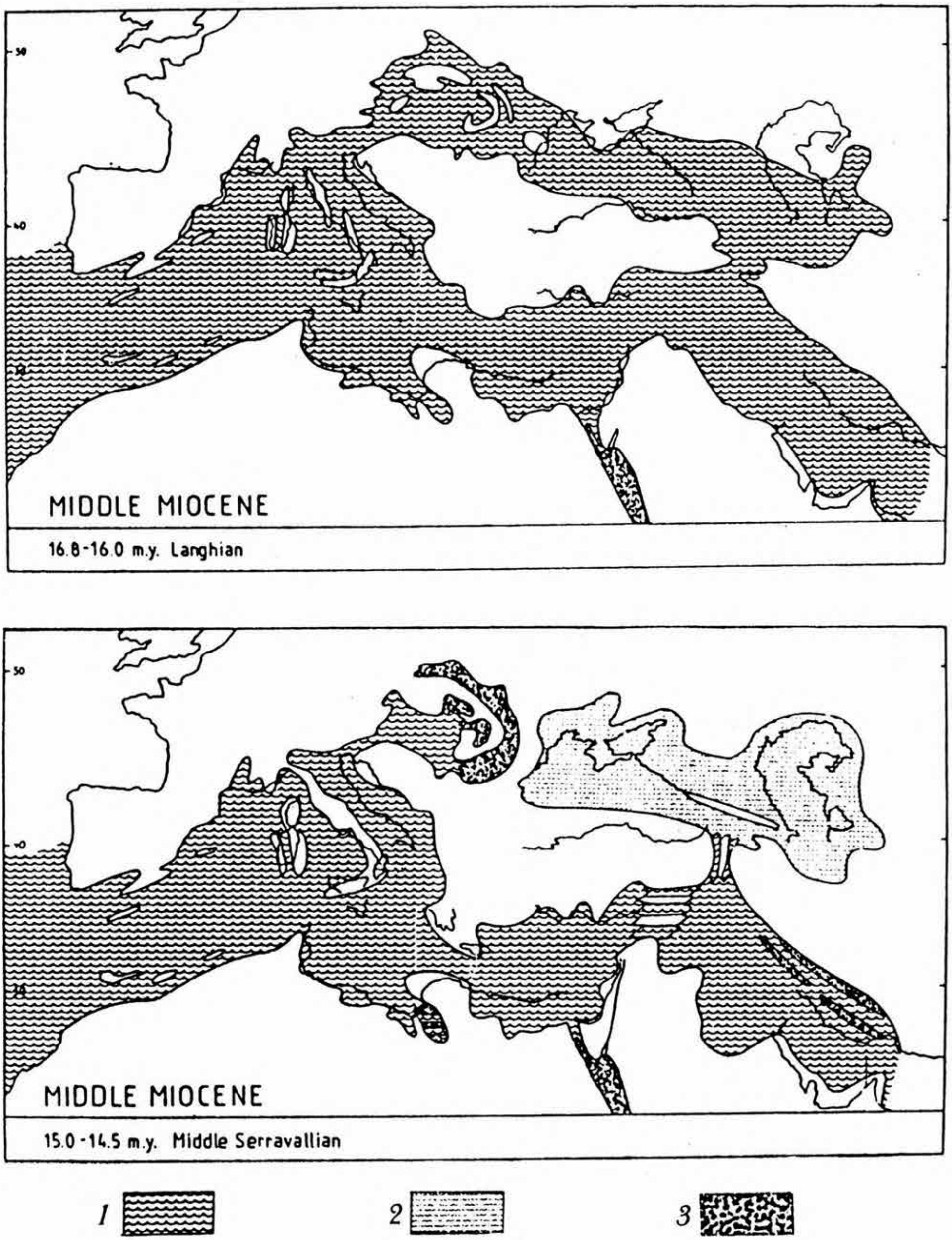


*Fig. 1.* Explanation of the relative position of the various continental masses, marine bodies and sea ways during the Middle Miocene in the Eastern Mediterranean
(after STEININGER and RÖGL, 1984)

*1* Marine realms, *2* reduced salinity realms, *3* evaporitic realms

**The distribution of some Miocene echinoid species found in the Mediterranean and West Indo-Pacific regions**

*Table 1*

| Species | Mediterranean region | West Indo-Pacific region |
|---|---|---|
| *Clypeaster martini* | Aqui. and Burd.: Italy and France<br>M. Miocene: Libya, Egypt and Sardinia | M. Miocene: Somalia, Zanzibar and Zululand |
| *C. latirostris laganoides* | Langh.: Libya and Italy<br>M. Miocene: (Hel.): Italy, Lybia, Portugal and Sardinia | M. Miocene: Somalia |
| *C. marginatus* | M. Miocene: Malta, Italy, Portugal, France, Algeria, Spain and Egypt | M. Miocene: Zanzibar |
| *Lagamm depressum* | Tor.: Malta | Burd.: Saudi Arabia<br>M. Miocene: Zululand |
| *Echinolampas complanata* | M. Miocene: Syria | Burd.: Iran |
| *Schizaster eurynotus* | Burd.: Malta<br>Langh.: Malta, Italy, France and Libya<br>M. Miocene: Italy, France, Greece, Egypt, Libya and Portugal | M. Miocene: Somalia |
| *S. parckisnoni* | Aqui.: Malta<br>Langh.: France, Spain and Italy<br>M. Miocene: Malta and Italy | M. Miocene: Somalia |
| *S. desori* | Helv.: Corsica | M. Miocene: Somalia |
| *S. baylei* | Burd.: Sardinia and Corsica | M. Miocene: Somalia |
| *S. valabrequei* | Aqui.: France | M. Miocene: Somalia |
| *Pericosmus latus* | Burd.: Libya and Egypt<br>Langh.: Malta<br>M. Miocene: Egypt, Malta and Libya | M. Miocene: Somalia |
| *Macropneustes saheliensis* | Helv.: Italy<br>Tor.: Algeria | M. Miocene: Somalia |

Aqui. = Aquitanian, Burd. = Burdigalian, Langh. = Langhian, Helv. = Helvetian, M. = Middle, Tor. = Tortonian.

Here, also it can be found that some species of the Early and Middle Miocene of the Mediterranean region are found in the Middle Miocene of Somalia, Zanzibar and Zululand. Thus attesting for a migration of some Early and Middle Miocene echinoids from the Mediterranean region towards the West Indo-Pacific during the Middle Miocene.

Common Middle Miocene echinoid species are also found between East Africa, India and Pakistan. However, no common species whatever have been found between India or Pakistan and the Mediterranean region. These peculiarities were controlled by current circulation as shown here under.

**Migration of echinoid species in the Mediterranean Sea and the Indian Ocean**

Ten Early and Middle Miocene species recorded from the Mediterranean region (France, Portugal, Spain, Italy, Malta, Greece, Libya and Egypt), are also found in the Middle Miocene rocks of the Indian Ocean (Somalia, Zanzibar and Zululand) (Table 1) These are: *Clypeaster martini*, *C. latirostris laganoidea*, *C. marginatus*, *Schizaster eurynotus*, *S. parckinsoni*, *S. desori*, *S. baylei*, *S. valabrequei*, *Pericosmus latus* and *Macropneustes sahelensis*. These similarities indicate west—east migration from the Mediterranean to the West Indo-Pacific region. At the same time, two species *Echinolampas complanata* and *Laganum depressum*, which are Early Miocene (Burdigalian) West Indo-Pacific forms (recorded from Iran and Saudi Arabia respectively), have been recorded from the Middle Miocene of Syria and Malta. This indicates east—west migration from Iran and the Arabian Gulf to the Eastern Mediterranean during the Middle Miocene time.

The general absence of common Miocene species between India and the Mediterranean region suggests that no migration path between these two areas existed at that time. However, common forms between East Africa and India are frequent (*Coelopleurus forbesi*, *Opechinus rosseaui*, *Clypeaster depressus*, *C. profundus*, *C. pulvinatus*, *Echinolampas spheroidalis* and *Breynia carinata*). These phenomena may be explained as a result of the distribution and the currents direction shown by FELL (1967), which indicate that migration of larval planktonic echinoids between the two areas (India and East Africa), is more likely to have proceeded in an east—west direction, from India towards East Africa.

It is also interesting to notice here that *Laganum depressum*, which is an Early Miocene of the Eastern Saudi Arabia, is to be found in the Middle Miocene of Zululand and in the Tortonian of Malta (ZAMMIT-MAEMPEL, 1978). This suggests migration in two directions from the Arabian Gulf to East African coast on one hand, and towards the Mediterranean Sea on the other hand.

**Conclusion**

The analysis and discussion of published data on the distribution of echinoids in the Mediterranean and the West Indo-Pacific regions show that these two basins were connected by a passage way during the Middle Miocene time as represented by the occurrence of identical echinoids in both regions. This passage way permitted to these species to migrate from the Mediterranean towards the Indian Ocean and vice versa although the main trend was from west to east.

The Middle Miocene echinoid faunas of the Gulf of Suez in Egypt show definite Mediterranean affinities at the generic and specific level (ALI, 1984). On the other hand, the marine Miocene faunas of Iran have much in common with both the West Indo-Pacific and the Mediterranean (EAMES and SAVAGE, 1975), suggesting that the passage way between the two regions during the Middle Miocene existed through the region of the Arabian Gulf and Iran and not through the Gulf of Suez and the Red Sea.

Here, it is interesting to notice that ADAMS et al. (1983), consider that a land bridge existed in Langhian time between the Indian Ocean and the Eastern Mediterranean. They reached their conclusions through the study of the foraminifera and macrofossils in Iran, Iraq and Arabia and around the Arabian Gulf. These conclusions not necessarily contradict our findings. The chronostratigraphical subdivisions referred to in the literature we used have not the degree of precision to decide if what is called Langhian in Arabian Gulf by ADAMS et al. (1983) corresponds to what is generally referred to as "Middle Miocene" in the present paper. It is well-known that the Early Langhian is considered by most Mediterranean Neogene geologists as related to the "Early Miocene" (HARLAND et al., 1982). In this case the Langhian of ADAMS et al. (1983), may be Early Miocene and a continuation of the period of separation of the West Indo-Pacific from the Mediterranean. The Middle Miocene of our discussion refers probably to the Late Langhian and Serravallian ages. However, if we accept the conclusions of ADAMS et al. (1983), about the Middle Miocene land bridge between the two regions, we suggest that there were several intervals of relatively high sea level during the Middle Miocene in which the land bridge was probably of low relief or in the form of island chains, so allowing for partial migration of some echinoid species between the Indian Ocean and the Mediterranean Sea by temporary connections between them.

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# PECTINID ASSEMBLAGE ZONES OF THE MIOCENE IN HUNGARY

by

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The Miocene formations in Hungary (from the Eggenburgian and the top of the Egerian respectively, up to the top of the Badenian inclusive) can be readily characterized by the frequent occurrence of pectinids in different facies.

A particularly rich fauna with "larger pectinids" and Chlamys forms of littoral, nearshore facies has become known from the Bretka, Budafok, Bántapuszta, Egyházasgerge, Fót, Pécsszabolcs and Rákos formations, whilst the representatives of the genus Amussium (Lentipecten, Propeamussium, Palliolum, etc) that are generally indicative of a deeperwater environment are characteristic of the Putnok, Garáb and Szilágy formations.

All these results have enabled the authors to make an attempt at developing a pectinid biozonation of the 23 to 13.8 Ma time span extending from the Eggenburgian to the end of the Badenian.

In the 1970's several proposals for subdividing the European Oligocene and Neogene on the basis of molluscan biozones were submitted (FRENEIX, 1974; CATZIGRAS, 1974 etc). The most important biozonation was that based on pectinids (BÁLDI, 1973; 1975; DEMARCQ, 1974; 1979). Although these data have been used by the present writers, because of the dissimilarities of the Mediterranean and Paratethyan Miocene deposits they could not be simply adapted to Hungary.

For the 23 to 13.8 Ma part of the Hungarian Miocene, the introduction of 5 pectinid assemblage zones and the distinction of 4 subzones are proposed (Fig. 1 and 2). The zones are to be documented with the following data:

*1* motivation for the choice of the zonal index fossil, derivation of the name;

*2* general characterization and boundaries of the zone (most important or alternative species, dates of entry and exit);

*3* possibilities for correlation (nannoplankton, planktonic foraminifers);

*4* relation to the corresponding lithostratigraphic units;

*5* major Hungarian occurrences;

*6* collection in which the respective zonal index species are to be found.

## Chlamys rotundata assemblage zone

*1* Demarcq's zonal index fossil, *Flabellipecten carryensis*, does not occur in Hungary, but *Chlamys rotundata*, a species having the same date of entry, is quite common.

*2* Entry of *Chlamys rotundata* and *Chl. varia*, exit of *Chl. biarritzensis*, *Chl. deleta* and *Pecten arcuatus*. Alternative species (replacing the former in basin facies): *Lentipecten denudatum* of typical size.

*3* *Miogypsina gunteri* chron (base of Miocene: topmost Egerian? or base of Eggenburgian)

*4* Bretka Limestone Formation.

*5* Rudabánya, margin of Szendrő Mountains (Imola, Trizs).

*6* Department of Geology, ELTE (Loránd Eötvös University, Budapest).

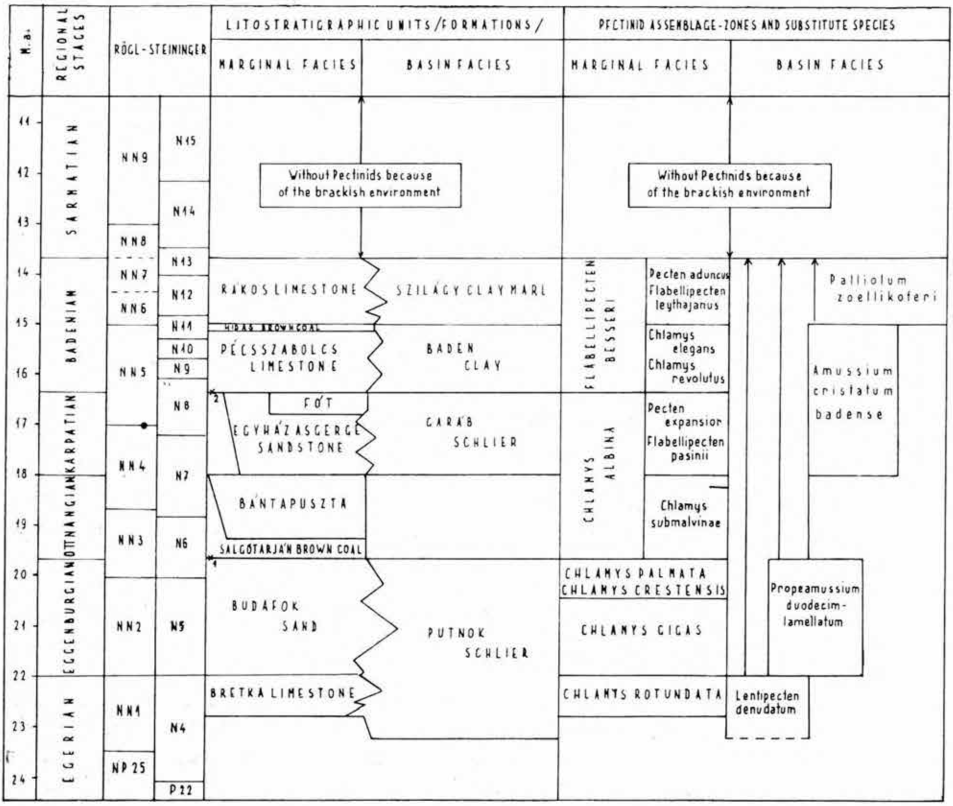



*Fig. 1.* Relationship between the lithostratigraphic units and the Pectinid assemblage zone
*1* Gyulakeszi Rhyolite Tuff Formation, $19.6 \pm 1.4$ Ma; *2* Tar Dacite Tuff Formation, $16.5 \pm 0.6$ Ma; † range of the substitute species in basin facies, ● NN4/NN5 boundary modificated by BÁLDI-BEKE—NAGYMAROSI (1979)

## Chlamys gigas assemblage zone

*1* Characteristic and common species.

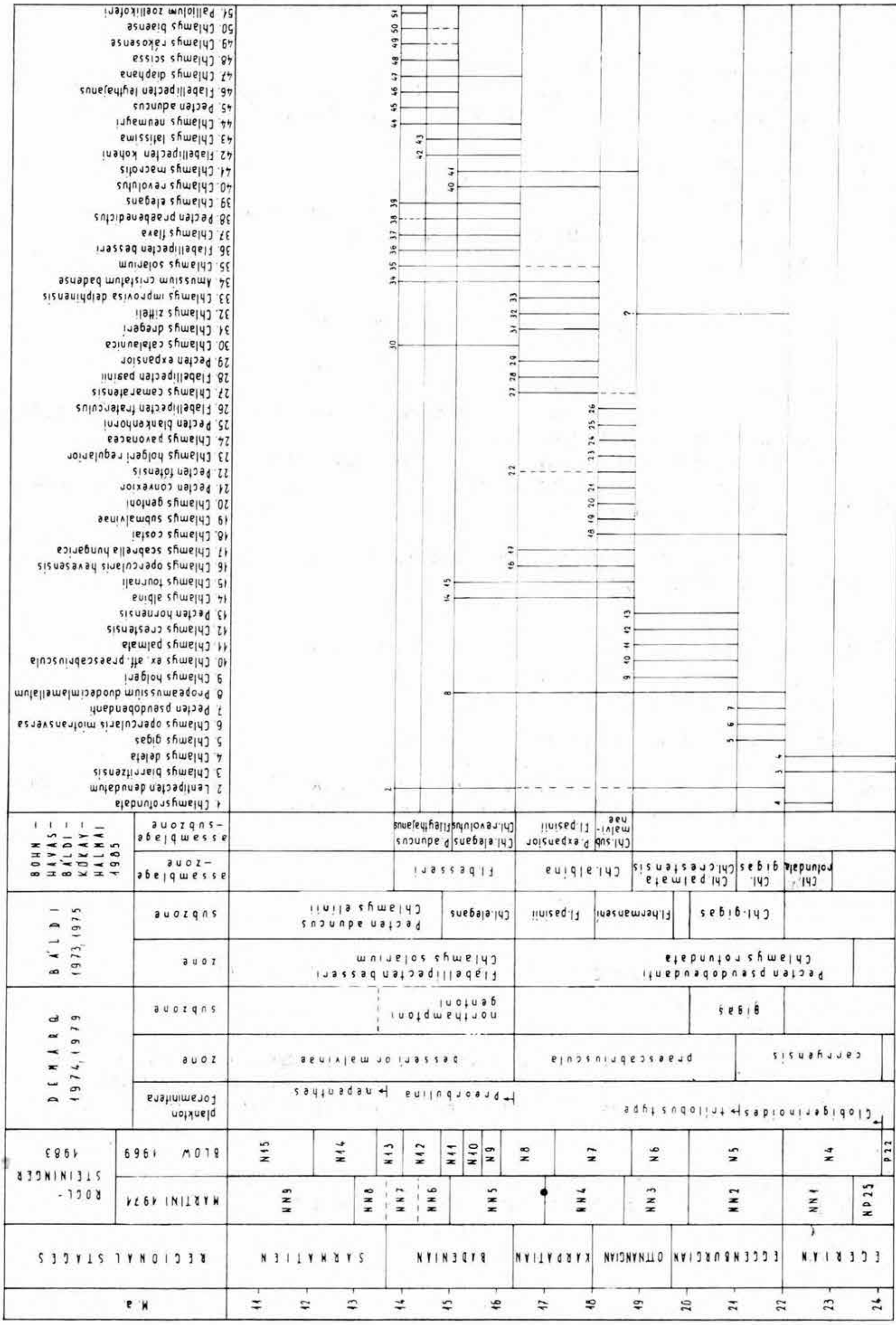
*2* Entry of *Chl. gigas*, *Pecten pseudobeudanti*, *Chl. opercularis miotransversa* and *Chl. costai.* Alternative species (replacing the former in basin facies): *Propeamussium duodecimlamellatum* accompanied by *Lentipecten denudatum.*

*3* At the base of NN2 chron, appearance of typical *Globigerinoides trilobus.*

*4* Budafok Sand Formation, Putnok Schlier Formation.

*5* Borehole Budafok 2 (50.5—108.2 m), Törökbálint, Dunabogdány, Salgótarján (Karancsalja), Iliny, boreholes in the Borsod basin: Alsószuha-1 (62.0—805.0 m), Sajókaza 292 (188.0—214.0 m), etc.

*6* Hungarian Geological Institute, Museum; Department of Geology, Loránd Eötvös University, Hungarian Natural History Museum.



• NN4/NN5 boundary modificated by Báldi-Beke-Nagymarosi /1979/

*Fig. 2.* Correlation with the standard plankton and Pectinid zonations

## Chlamys palmata—Chlamys crestensis assemblage zone

*1* Characteristic and common species.

*2* Entry of *Chl. palmata, Chl. crestensis, Chl. gentoni, Chl.* ex gr. *praescabriusculus, P. hornensis, Chl. holgeri.* Alternative species (replacing the former in basin facies): *Pp. duodecimlamellatum, L. denudatum.*

*3* Upper part of zone NN2.

*4* Budafok Sand Formation, Putnok Schlier Formation.

*5* Borehole Budafok 2 (30—50.5 m), Borsodnádasd, Bekölce, Budafok, boreholes in the Borsod Basin (see in the description of the Chlamys gigas zone).

*6* Hungarian Natural History Museum.

## Chlamys albina assemblage zone

*1* Characteristic and common species.

*2* Entry of *Chl. albina, Chl. tournali, Chl. gentoni, Chl. opercularis hevesensis, Chlamys scabrella hungarica, P. fótensis, P. blankenhorni, Chl. holgeri regularior, Chl. submalvinae.* Alternative species (replacing the former in basin facies): *L. denudatum, Amussium cristatum badense.*

*3* Date initiating the zone is provided by a short-interval overlapping of the zones NN3 and CN3 of Martini's and Okada-Bukry's nannozonation, the Praeorbulina date being the upper boundary.

*4* Bántapuszta Formation, Egyházasgerge Sandstone Formation, Fót Formation.

*5—6* For details, see the subzones.

Within the zone, two subzones could be distinguished:

## Chlamys submalvinae subzone

*1* Characteristic and common species.

*2* Its lower boundary is the date of entry of the eponymous species of typical size (the form appearing deeper is one of anomalous size, smaller than the type). The upper boundary of the subzone is provided by the exit of *Chl. holgeri regularior, Chl. pavonacea,* and *Chl. submalvniae.*

*3* —

*4* Bántapuszta Formation.

*5* Bántapuszta.

*6* Private collection of J. KÓKAY (Budapest).

## Pecten expansior—Flabellipecten pasinii subzone

*1* Characteristic and common species.

*2* The entry of *P. expansior, Flabellipecten pasinii, Chl. dregeri, Chl. catalaunica, Chl. improvisa delphinensis;* presence of associates *Chl. opercularis hevesensis* and *Chl. scabrella hungarica;* exit of *Chl. zitteli.* Alternative species (replacing the former basin fcaies): *A. cristatum badense.*

*3* —

*4* Egyházasgerge sandstone Formation, Fót Formation, Garáb Schlier Formation.

*5* Egyházasgerge, Várpalota, Borsod basin (boreholes Balaton 27; 28; Mikófalva 3; Nekézseny, Fót, Csomád, Garáb 1; Sirok 1; Alcsútdoboz 3).

*6* Private collection of J. KÓKAY (Budapest).

## Flabellipecten besseri assemblage zone

*1* Demarcq's eponymous zonal index fossil is typical of the Badenian of Hungary, too.

*2* Entry of *Flabellipecten besseri, Chl. flava, Pecten praebenedictus, Chl. neumayri, Chl. elegans*. Alternative species (replacing the former in the basin facies): *A. cristatum badense, L. denudatum, Palliolum zoellikoferi, Pp. duodecimlamellatum.*

*3* Initiation of the zone is provided by the Praeorbulina date; middle of the zones NN5, NN6, up to the middle of NN7?

*4* Pécsszabolcs Limestone Formation, Rákos Limestone Formation, Sámsonháza Formation, Baden Clay Formation.

*5—6* For details, see the subzones.

Within the zone, two subzones could be distinguished:

### Chlamys elegans—Pecten revolutus subzone

*1* Both the species are common in the Lower Badenian of Hungary.

*2* Entry of F. besseri, Chl. flava, P. praebenedictus, frequency of F. koheni, Chl. latissima, F. solarium, exit of P. revolutus and Chl. macrotis. Alternative species (replacing the former in the basin facies): A. cristatum badense and Pp. duodecimlamellatum.

*3* Its lower boundary is the Praeorbulina date, from the middle of zone NN5 up to the NN5/NN6 boundary.

*4* Pécsszabolcs Limestone Formation, Nógrádszakál Marl Formation, Tekeres Schlier Formation, Sámsonháza Formation.

*5* Borsod basin (boreholes Putnok 5; Tardona 78; Sajógalgóc 10 etc), boreholes Pécsszabolcs 1; Nógrádszakál 2; Tekeres 1; key-section of Sámsonháza etc.

*6* Hungarian Geological Institute, Museum.

### Flabellipecten leythajanus—Pecten aduncus subzone

*1* Characteristic and common species.

*2* Entry of *F. leythajanus, Chl. diaphana* and *Chl. scissa*, frequency of *Chl. biaense, Chl. rákosense* and *Chl. neumayri* in the upper segment of the subzone. Alternative species (replacing the former in the basin facies): *A. cristatum badense, L. denudatum* and *Palliolum zoellikoferi.*

*3* The NN5—NN6 boundary is the probable starting point of the subzone, its upper boundary is marked by the appearance of Sarmatian brackish-water forms.

*4* Rákos Limestone Formation, Szilágy Claymarl Formation.

*5* Budapest—Rákos; boreholes Tengelic 2; Tekeres 1, etc.

*6* Hungarian Geological Institute, Museum.

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# BRYOZOAN FAUNAS AND THE MESSINIAN SALINITY CRISIS

by

P. MOISSETTE and S. POUYET

*Introduction.* In the Mediterranean, the end of the Miocene is notably characterized by widespread deposition of evaporites (ROUCHY, 1982). The so-called "Messinian salinity crisis" has given rise to numerous controversies concerning the causes and amplitude of this sedimentological phenomenon and its biodynamical effects on the Mediterranean floras and faunas at the Miocene/Pliocene boundary. As a matter of fact, rich marine faunas are found in Tortonian to Lower Messinian (preevaporitic) formations whereas a gap is inferred during the deposition of evaporites, followed by reappearance of marine faunas in the lower Pliocene.

Based on these facts, two main hypotheses have been advanced. Some believe that marine floras and faunas were decimated because of the increase in salinity and subsequent drying up of the Mediterranean in the late Messinian. Reimmigration from the Atlantic came afterwards during the Pliocene transgression (HSÜ et al., 1973; CITA, 1976; RUGGIERI and SPROVIERI, 1976). Others are of the opinion that most of the taxa persisted, at least in some basins, due to an uninterrupted connection between the Mediterranean sea and the Atlantic ocean (GAUDANT, 1978; DAVID & POUYET, 1985).

From this point of view some groups have already been studied more or less in detail (RUGGIERI et al., 1969; BENSON, 1976a; CITA, 1976; GAUDANT, 1978). Conclusions drawn from these works supported one or the other of the hypotheses mentioned before. From an analysis of Messinian fossil deposits, we present here observations on how, and if, bryozoan faunas were affected by the massive deposition of salt and gypsum in the Mediterranean at the end of the Miocene.

## Messinian bryozoan fauna

Although bryozoans have been found in various Messinian levels, the only deposits studied as yet (Fig. 1) are those of the western parts of the Mesogean realm: Spain, Morocco (EL HAJJAJI, 1982) and Algeria (MOISSETTE, 1984) where more than 200 species of cyclostome and cheilostome bryozoans were identified. From a stratigraphic point of view, most of the deposits studied may be attributed to the Early or Middle Messinian, although some of them may possibly be coeval with evaporite deposition (Late Messinian).

In this paper, we shall confine ourselves to the study solely of the cheilostome bryozoans whose species are better known and more significant. However, a complete understanding of the Messinian bryozoan faunas will have to include, finally, also the Cyclostomata.

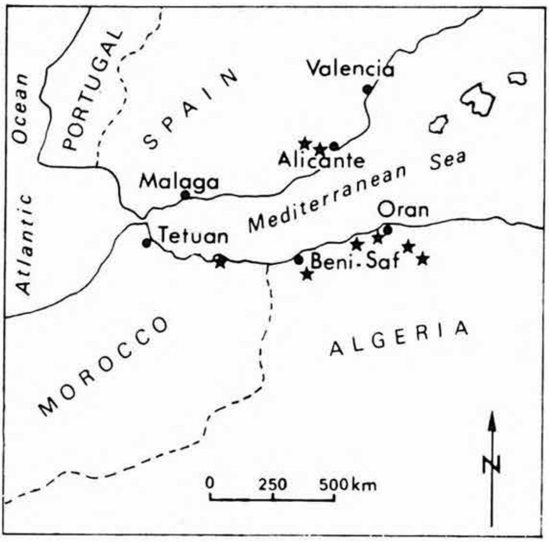


*Fig. 1.* Geographic distribution of Messinian bryozoan deposits in the Alboran basin (western Mediterranean)

**Analysis of the cheilostome faunas**

Among the 154 species of cheilostome bryozoans found in the three countries concerned in this study, some are new to science, others were left in open nomenclature. For statistical purposes, only those species already described or new species discovered in at least two countries will be taken into further account. Thus, we only use 122 taxa for our comparisons (Fig. 2).

To establish whether Messinian species ranged into Pliocene, existing literature on Mediterranean Pliocene bryozoans faunas from Italy (CIPOLLA, 1921; POLUZZI, 1971 and 1975), Spain (POUYET, 1976) and Tunisia (BUGE, 1956) was used but no account was taken in Fig. 2 of the species appearing in the Pliocene.

A relatively high number of species (83=68%) are common to the Messinian and the Pliocene (groups 2, 3, 5 and 6). Some of them (26=21%) appeared in the Messinian (or earlier?) and were also found in the lower or upper Pliocene and even in the Pleistocene (group 5) or the Recent (group 6). Among the Messinian (or lower) to Recent species (groups 2, 3, 5 and 6), Mediterranean endemics represent almost 17%. These species are therefore restricted to the Mediterranean basin and could not have been reintroduced from the Atlantic in the Early Pliocene, after the salinity crisis. This percentage could be much higher if we had included taxa that are known in the fossil stage only in the Mediterranean sea but that are found nowadays in the "near Atlantic", beyond the Straits of Gibraltar (Morocco, Portugal, Madeira ...). Their centre of distribution is in the Mediterranean and they are assumed to have migrated to the Atlantic after the Miocene or the Pliocene. Besides, these endemic species in the broader sense are often not distinguished by biologists (GAUTIER, 1962; HARMELIN, 1969; HAYWARD, 1974) from the "real" endemics.

In the Recent cheilostome bryozoan faunas of the Mediterranean, the percentage of endemic species ranges between 32 (western Mediterranean: GAUTIER, 1962) and 44 (eastern Mediterranean: HARMELIN, 1969), whereas in our Messinian faunas, this percentage amounts to 35. This fairly pronounced endemism bears testimony to a prolonged isolation, especially from the Atlantic ocean. However, only 1 *sensu stricto* and 7 *sensu lato* endemic species are known from the Messinian to the Recent. The others disappeared in the interval and were replaced by remnants of Atlantic immigrants and species evolving in situ from indegenous stocks.

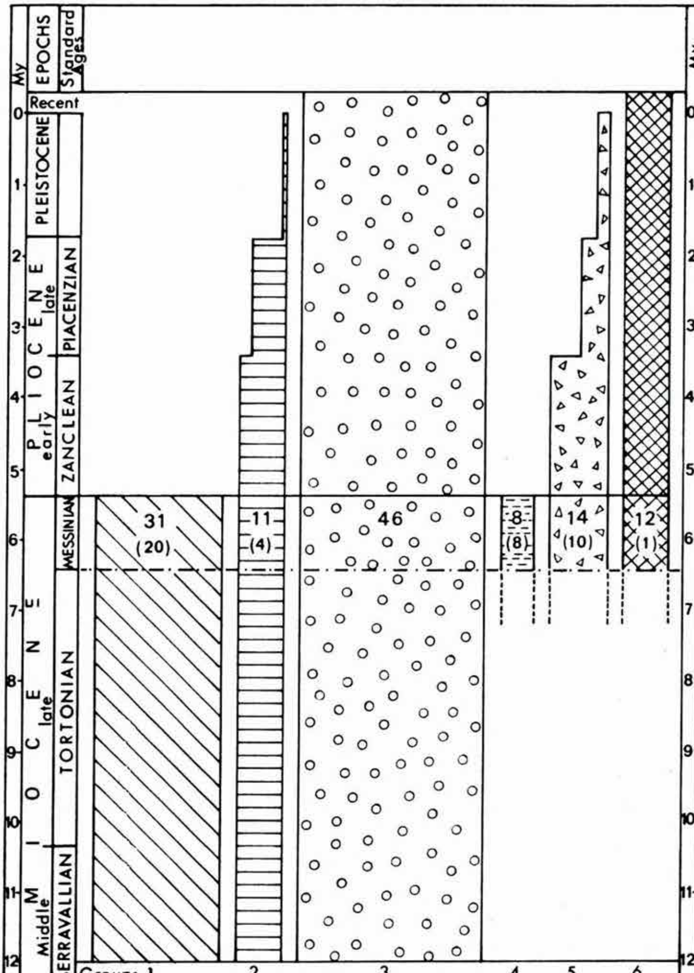


*Fig. 2.* Diagram showing the stratigraphic distribution of Messinian cheilostome bryozoan species For each group indication is given of the total number of species and, in parenthesis, of the number of endemic species to the Mediterranean basin

**Zoogeographical affinities of the Messinian bryozoa**

Apart from endemic, temperate-boreal Atlantic and cosmopolitan species, types with tropical affinity—species still living or extinct but belonging to tropical genera (VAVRA, 1980; DAVID & POUYET, 1985)—are found among Messinian to Recent bryozoans. Among the 154 taxa of western Mediterranean Messinian cheilostome bryozoans, the following belong to this group:

*Biflustra savarti*
**Antropora lecointrei*
*Antropora* sp.
*Chaperia annulus*
*Tremogasterina catenularia*
**Tremogasterina* nov. sp. 1
*Tremogasterina* nov. sp. 2
*Tremogasterina* sp.
*Tremopora radicifera*
*Arachnopusia* sp.
**Steginoporella cucullata*
**Steginoporella iberica iber.*
**Steginoporella manzonii*
**Steginoporella montenati*
*Thalamoporella neogenica*
*Gephyrophora rostrigera*
**Emballotheca* nov. sp. 1
*Escharina dutertrei*
*Stenopsis*
*Margaretta cereoides*
*Metrarabdotos canui*
*Metrarabdotos elegans*
*Metrarabdotos helveticum*
*Metrarabdotos lecointrei*
*Metrarabdotos moniliferum*
**Anoteropora* aff. *persimplex*
**Kionidella* aff. *excelsa*
*Lacrimula* sp.
*Conescharellina* sp.

Noteworthy is the fact that the number of tropical types decreases steadily during the Pliocene and the Pleistocene. We have thus 21 species of this group in the Messinian, 10 in the Pliocene, 5 in the Pleistocene and only 2 in the Recent.

Another distinctive feature of the faunas under study is the presence among the tropical taxa of species or genera whose present-day distribution is limited to the Indo—Pacific region:

*Tremopora radicifera*
*Arachnopusia* sp.
*Gephyrophora rostrigera*
**Emballotheca* nov. sp. 1
*Escharina dutertrei*
*Margaretta cereoides*
**Anoteropora* aff. *simplex*
**Kionidella* aff. *excelsa*
*Lacrimula* sp.
*Conescharellina sp.*

*endemic species

Their presence here must be related to the persistence of a stock of taxa in the Mediterranean since the end of the Early Miocene, when connection with the Indian ocean was disrupted. These forms remained in the Mediterranean and could not have been reintroduced from the Atlantic in the Pliocene; they belong to relics of the palaeomediterranean fauna. Just as for the afore-mentioned group, their number declines progressively: 11 species (7%) in the Messinian and only 3 (2%) in the Recent western Mediterranean.

**Conclusions on Messinian cheilostome bryozoan faunas**

The data reviewed above indicate that part of the Miocene bryozoan faunas remained in the Mediterranean sea till the Pliocene or further. This could only happen as a result of the persistence of normal marine conditions throughout the Messinian, at least in some basins (DAVID & POUYET, 1985).

Rarefaction and then disappearance of certain species have been progressive, without any major break, and is due essentially to the cooling of the climate (BUGE, 1982).

Nevertheless and in the future, a better knowledge of Neogene bryozoan faunas will still be necessary. This seems particularly true for the Atlantic side (Morocco, Portugal, Spain...), for the western and especially eastern (poorly known) Mediterranean and also for the Indian ocean. We shall then be able to better assess the extent of the faunistic modifications connected with the various events occurring near the Miocene/Pliocene boundary.

## Other evidence

Comparable phenomena are noted in some other groups of marine organisms: molluscs (RUGGIERI et al., 1969, GRECCHI, 1977), fishes (GAUDANT, 1978)... although foraminifera (ADAMS, 1976; CITA, 1976) or ostracods (BENSON, 1976b; SISSINGH, 1976) seem to exhibit different patterns and be more sensitive to the salinity crisis.

The complete and definitive disappearance of the hermatypic corals at the end of the Messinian, often invoked in support of "catastrophic" models is only the consequence of modifications—essentially of climatic origin—already initiated during the Tortonian or before: decrease in the number of species and predominance of the genera *Porites* and *Tarbellastraea* (ESTEBAN, 1979; SAINT-MARTIN, 1984).

It should be pointed out, however, that at the end of the Messinian marine faunas hitherto known are principally represented by planktonic forms (often dwarfed) and that, consequently, a continuity of marine conditions is hard to presume for benthic forms such as bryozoans. A set of concordant field observations shows nevertheless the persistence of normal marine conditions in some places of the Mediterranean, even at the end of the Messinian, during deposition of evaporites. We may mention for example the intercalation of normal marine deposits inside gypsiferous sequences —with planktonic but also benthic faunas (MONTENAT et al., 1980)—and the presence of marine deposits above evaporites (ROUCHY, 1982). On the other hand, the evaporitic phenomenon is not so general, so complete than one may think at first: some Messinian basins do not have evaporites at all, Morocco, western Oranie (ROUCHY, 1982) and a marine continuity of sedimentation between Miocene and Pliocene has been suggested by several authors, notably in the western basins (MONTENAT et al., 1976). Lastly, gypsum depositions may occur before the late Messinian. We have thus gypsiferous intercalations inside diatomites (Spain: Lorca; ROUCHY, 1982) and gypsum may be found in some sequences at the end of the Tortonian (Spain: Fortuna, Albanilla; MONTENAT, 1977). There thus exists during the Messinian a diachronism (ROUCHY, 1982) in evaporite deposition: normal marine conditions continued to predominate in some basins, whereas in others growing confinement led to the deposition of evaporites.

## Conclusion

On the basis of evidence presented here, a complete disappearance of marine faunas during the salinity crisis is not favoured. Planktonic and benthic organisms remained in situ in some basins during all the Messinian. Atlantic water inlet probably never ceased totally through the Betic-Rif zone. Climatic changes—cooling (BIZON &

MÜLLER, 1976; DEMARCQ, 1979; HEIMANN, 1979)—are then the main factors that induced modifications in the bryozoan and other marine (PÉRÈS & PICARD, 1964) faunas of the Mediterranean and other oceans (BOURROUILH-LE JAN, 1979).

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# STRATIGRAPHICAL AND BIOGEOGRAPHICAL SIGNIFICANCE OF BRYOZOAN FAUNAS FROM MIOCENE TO RECENT IN TETHYS AND PARATETHYS

by

S. Pouyet and L. David

*Introduction.* For twenty years, the studies about the bryozoan faunas of the tethyan realm have been sufficiently numerous to give a good idea of stratigraphical scale and palaeoecological significance. Several palaeontological studies have been done on the Burdigalian of the Rhone basin, the Badenian of the Vienna basin, the Messinian of Morocco, Algeria and Spain, Italy and Tunisia (Fig. 1).

We collected data about Neogene or recent deposits or sites (Fig. 2). Finally, a list of 452 taxa of the Cheilostome bryozoans has been available for our study.

Among these data sets, we have chosen three examples:

– six recent sites in the western and eastern Mediterranean;

– four Burdigalian deposits from the south Rhone basin, four Badenian localities from the Vienna basin and one in the Serravallian from the north Rhone basin;

– three localities from the western mediterranean Messinian and three from the Pliocene of Spain, Tunisia and Portugal.

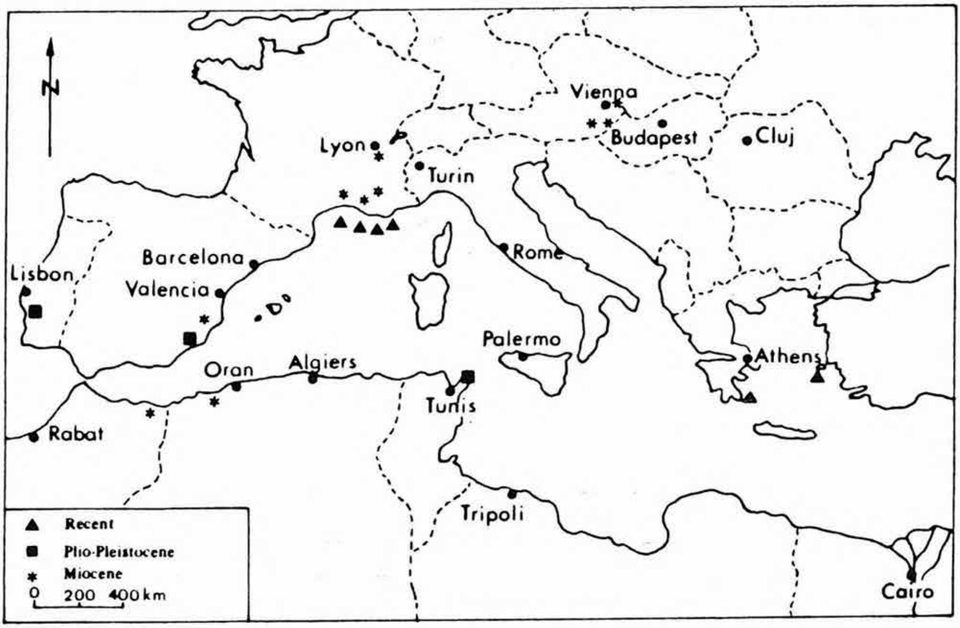


*Fig. 1.* Geographical distribution of main Neogene and Recent bryozoan deposits from the Mesogean realm

These three examples are enough to show the interest of comparison tests between the bryozoan faunas of Tethys and Paratethys.

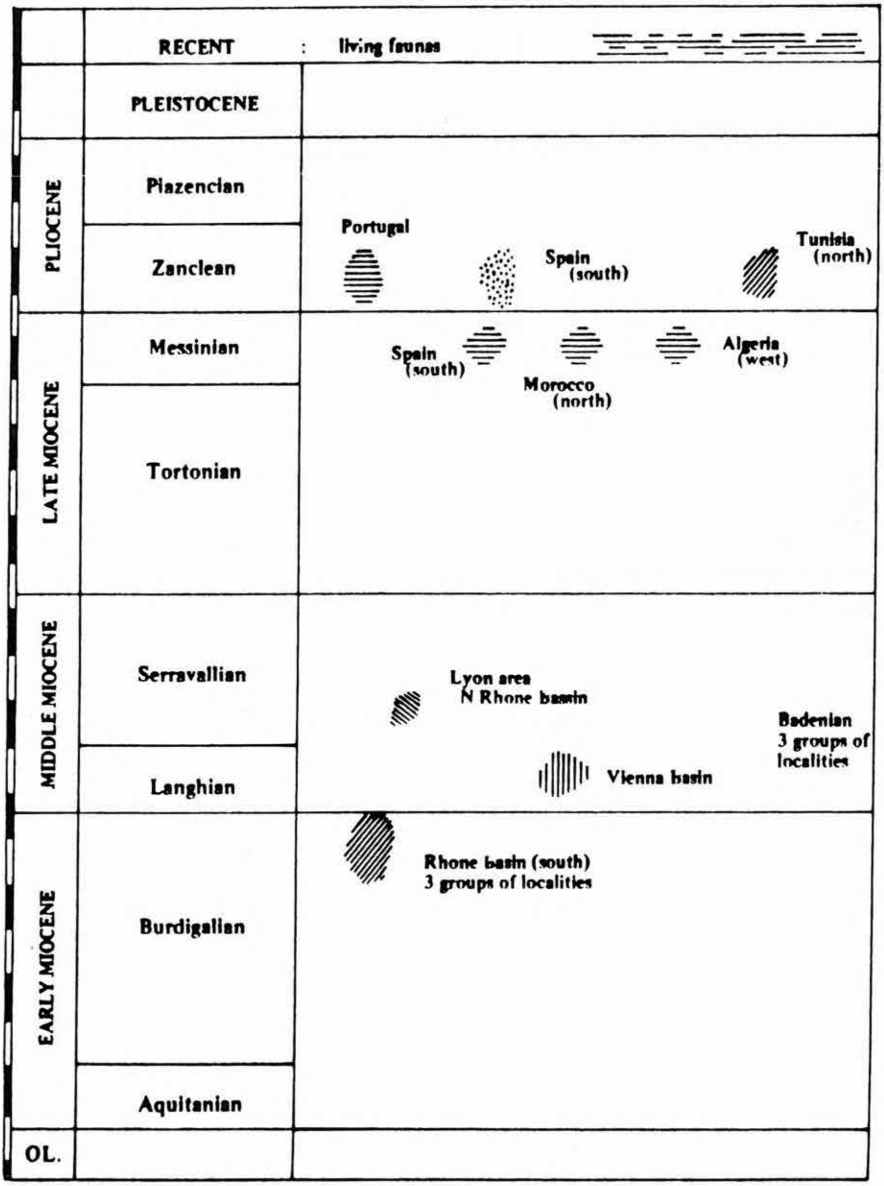


*Fig. 2.* Stratigraphical distribution of main bryozoan localities

## Methods

For each recent site or fossil locality, we established a list of species: we confine ourselves to the study solely of the cheilostome species. A complete study of the bryozoan faunas will have to include ultimately also the Cyclostomata.

We kept all the specific taxa, included doubtful species (cf, aff, ?). We excluded generic taxa and, for the living species, the ones which cannot be fossilized. Subspecific species are considered only as one taxa.

We used a similarity test to compare our lists the Kojumdgieva coefficient (1976), it allows a better comparison between localities in which the number of species is very unequal:

$$K = \frac{Ca\% + Cb\%}{2}$$

$$Ca\% = \frac{C}{A} \times 100 \qquad Cb\% = \frac{C}{B} \times 100$$

A = number of species from *a* locality
B = number of species from *b* locality
C = number of species both in *a* and *b* localities

Then, we used cluster analysis to calculate a coefficient of distance between the percentages of similarity coefficients and to produce dendrograms. The program was written by DR. GILLES CARBONNEL of the Claude-Bernard University (Lyon); we thank him for his assistance.

## Results

FIRST EXAMPLE: comparison between recent sites of the Mediterranean (Figs. 3, 4 and Table 1).

We chose four sites from the Tyrrhenian Sea studied by HARMELIN (1976) and two sites from the Aegean Sea: Chios Island (HAYWARD, 1974) and the Charcot cruise (HARMELIN, 1969).

The 4 sites of the Tyrrhenian Sea correspond to four different biocoenoses: rocky ground (97 species), Posidonia beds (70 species), detritic sea bottom (87 species) and minute substrate (70 species). The coefficient of similarity K and the dendrograms (Fig. 3 and Table 1) show a very strong correlation between the two hard substrata then a good correlation with the Posidonia beds and finally a slight one with the sandy substratum (= detritic sea bottom).

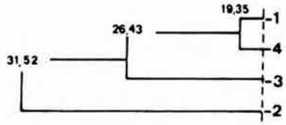


*Fig. 3.* Dendogram showing the distance coefficient between four sites of living faunas in the Tyrrhenian Sea (see Table 1)

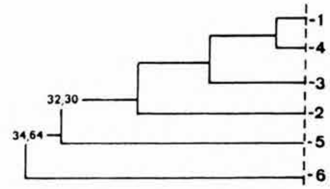


*Fig. 4.* Dendogram showing the distance coefficient between four biocenoses of the Tyrrhenian Sea and two sites of the Aegean Sea (see Table 1)

*Table 1*

| | | rocky ground<br>1 | posi- donies<br>2 | detritic sea bottom<br>3 | minute substrate<br>4 | Chios Island<br>5 | Charcot cruise<br>6 |
|---|---|---|---|---|---|---|---|
| W Mediterranean Tyrrhenian Sea | 1 | 100 | 63,95 | 68,68 | 73,17 | 49,56 | 53,33 |
| | 2 | 63.95 | 100 | 48.99 | 64.76 | 51.04 | 38.85 |
| | 3 | 68.68 | 48.99 | 100 | 61.07 | 38.64 | 51.98 |
| | 4 | 73.17 | 64.76 | 61.07 | 100 | 51.46 | 53.62 |
| E Mediterranean Aegean Sea | 5 | 49.56 | 51.04 | 38.64 | 51.46 | 100 | 38.60 |
| | 6 | 53.33 | 38.85 | 51.98 | 53.62 | 38.60 | 100 |

The sites of the Aegean Sea are weakly correlated with the ones of the Tyrrhenian Sea. The biocoenoses are similar but the biogeographic distance is great. The geographic distance is the main factor which marks the differences between the sites.

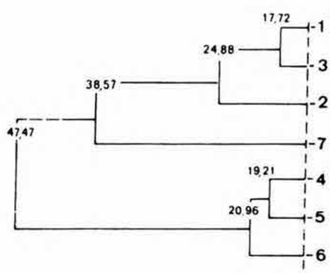


*Fig. 5.* Dendogram showing the distance coefficient between the seven localities of Table 2

SECOND EXAMPLE: comparison between several localities from Tethys and Paratethys in Rhone basin and Vienna basin (Fig. 5 and Table 2).

In the south part of the Rhone basin, we studied numerous outcrops from three areas:

— four outcrops from the Mus area (Gard, Languedoc) (David, Mongereau & Pouyet, 1972; Pouyet, 1973). The bryozoan faunas are very similar and we compiled a single list of 83 species for the whole region;

— three outcrops from the Valreas area (Drome) (David, Mongereau & Pouyet, 1970) for which we established a list of 59 significant species;

— thirteen outcrops from the Faucon-Mollans-Malaucene area (Vaucluse) (Pouyet & David, 1984). We obtained a list of 76 species.

*Table 2*

| | | Rhone basin | | | Vienna basin | | | Rh. b. |
|---|---|---|---|---|---|---|---|---|
| | | Mus | Valréas | Faucon | North | South | West | Lyon |
| | | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
| Burdigalian | 1 | 100 | 56.55 | 68.06 | 22.55 | 28.99 | 30.91 | 29.10 |
| | 2 | 56.55 | 100 | 54.20 | 23.16 | 24.09 | 23.65 | 34.78 |
| | 3 | 68.06 | 54.20 | 100 | 30.75 | 33.53 | 34.84 | 36.18 |
| Badenian | 4 | 22.55 | 23.16 | 30.75 | 100 | 64.43 | 60.75 | 17.18 |
| | 5 | 28.99 | 24.09 | 33.53 | 64.43 | 100 | 62.47 | 16.44 |
| | 6 | 30.91 | 23.65 | 34.84 | 60.75 | 62.47 | 100 | 25.39 |
| Serravallian | 7 | 29.10 | 34.78 | 36.18 | 17.18 | 16.44 | 25.39 | 100 |

These three groups of localities have the same age: Upper Burdigalian. Their comparison shows a very strong similarity (K = 54 to 68) and very short distance (17 and 24).

In the paratethyan Vienna basin, a great deal of deposits were studied in three areas (David & Pouyet, 1974; Vavra, 1977):

— nine localities from the northern part of the intraalpin basin at the north of Vienna. A single list of 62 species was established;

— twelve localities from the southern part of the intraalpin basin at the south of Vienna. The complete list includes 76 species;

— ten localities from the eastern Pannonian basin with a list of 92 species.

These three groups of deposits have the same age: Badenian i.e. Langhian—Serravallian. This age is slightly younger than the one of the south Rhone basin. The comparison of these three groups shows also a very strong similarity (K = 60 to 64) and a very short distance (19 and 20).

In conclusion, the similarity is very strong when the deposits have an identical age and when they are in the same palaeogeographical area.

Then, we compared a Serravallian deposit from the north of Rhone basin (Lyon area) (David, 1965) with the south Rhone basin and the Vienna basin. The age of this locality is nearer of that of the Vienna basin, however the similarity is greater with the south of the Rhone basin. The palaeogeographical unity is more significant than the stratigraphical position.

THIRD EXAMPLE: comparison between fossil deposits from Upper Miocene and Lower Pliocene of the western Mediterranean sea (Fig. 6 and Table 3).

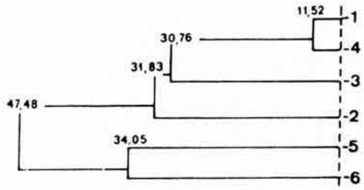


*Fig. 6.* Dendogram showing the distance coefficient between the six localities of Table 3

At first, we compared three groups of Messinian deposits from western Algeria, south Spain (Moissette & Pouyet in press) and north Morocco (El Hajjaji, 1982). The similarity co-efficient is 46 to 64: the deposits belong to the same palaeogeographical area: the Alboran basin.

Then, we compared three groups of Lower Pliocene localities from south Spain (Pouyet, 1976), Tunisia (Buge, 1956) and Portugal (Galopim de Carvalho, 1971). The similarity co-efficient is weak: from 30 to 41. This result confirms that the palaeogeographical unit is the most important fact: indeed the localities belong to three different palaeogeographic basins: Atlantic, Alboran and central Tethys.

At last, the comparison between Messinian deposits from south Spain and Pliocene deposits from south Spain leads to a value of 89 for the similarity co-efficient and a value of 11 for the distance. We can suppose a very high palaeogeographical endemism in a part of the Alboran basin only.

*Table 3*

| | | ALGERIA W 1 | SPAIN S 2 | MOROCCO N 3 | SPAIN S 4 | TUNISIA 5 | PORTUGAL 6 |
|---|---|---|---|---|---|---|---|
| | 1 | 100 | 64.90 | 59.32 | 89.54 | 33.08 | 31.86 |
| Messinian | 2 | 64.90 | 100 | 46.85 | 43.85 | 21.07 | 18.40 |
| | 3 | 59.32 | 46.85 | 100 | 49.22 | 33.52 | 30.42 |
| | 4 | 89.54 | 43.85 | 49.22 | 100 | 38.71 | 30.90 |
| Pliocene | 5 | 33 08 | 21.07 | 33.52 | 38.71 | 100 | 41.36 |
| | 6 | 31.86 | 18.40 | 30.42 | 30.90 | 41.36 | 100 |

## Conclusion

Comparing the results of both recent and fossil faunas as illustrated in the three examples, the following conclusion may be drawn: the fossil asscciations are dependent of three main factors, time, ecology and palaeogeography.

TIME is the fundamental factor which is always present. The standard systems of comparison of faunas (percentage of living faunas) are found upon the stratigraphical distribution i.e. time only.

ECOLOGY is a very important factor. The comparison between living faunas (1st example) shows the influence of ecology very well. This factor is less important in the fossil assemblages which are less dependent of the environment because there are several various biotopes, not only one. The similarity co-efficient is enough to estimate the effect of environment.

PALAEOGEOGRAPHY: in appearance, it is the main factor because it takes evolution (migrations, endemism), environment and geodynamic into account. Palaeogeography has a synthetic significance and the best mean to appreciate it, is cluster analysis.

This study represents the initial step in the comparative analysis of Neogene and Recent mediterranean bryozoan faunas. Further works will be necessary to confirm our results.

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# MORPHOLOGICAL VARIATION IN THE CENOZOIC ECHINOID CLYPEASTER AND ITS ECOLOGICAL AND STRATIGRAPHICAL SIGNIFICANCE

by

E. P. F. ROSE and R. H. PODDUBIUK

## The importance of Clypeaster

Six characteristics give *Clypeaster* particular importance among Neogene echinoid genera:

*1 Clear morphological distinction.* Even in poor fossil material, echinoids of the order Clypeasteroida can easily be recognized by their accessory tube-foot pores and internal skeletal pillars. Clypeaster is among the most distinctive and best-known of clypeasteroids, with a fossil record that is especially well documented.

*2 High preservation potential.* Rigid interplate suturing plus support and stress distribution by its internal pillar network, make the Clypeaster test particularly robust (SEILACHER, 1979; SMITH, 1984). Preservation potential is often further enhanced by an infaunal mode of life.

*3 Wide geographic distribution.* Clypeaster is one of the most widespread clypeasteroid genera in Recent seas (GHIOLD, 1985) and appears to have been tropicopolitan throughout the Neogene. Individual species have been reported to range widely (ALI, 1983), but tend to be confined to single faunal provinces.

*4 High species diversity.* With more than 400 named species (DURHAM, 1966), Clypeaster has the highest recorded internal diversity of any echinoid genus.

*5 Characteristic Cenozoic range.* The order Clypeasteroida is exclusively Cenozoic, originating in the Palaeocene (KIER, 1982). The earliest species of Clypeaster presently known in Middle Eocene (ROMAN, 1968). By early Oligocene times the genus was common and widespread in the Tethyan region, but did not reach its maximum geographic range or species diversity until the Miocene.

*6 Local abundance.* Ecological success and high preservation potential have ensured that Clypeaster is numerically, as well as taxonomically, among the best represented of all Neogene echinoid genera. Beds rich in Clypeaster specimens occur widely within Oligo-Miocene shallow water carbonate marine facies (BOGGILD and ROSE, 1985).

## Factors limiting biostratigraphic use

Despite high recorded species diversity, rapid evolutionary change and common, widespread occurrence, Clypeaster has been little used in biostratigraphy. There seem to be five main reasons for this omission:

*1 Species concepts differ widely between authors.* For example, ROSE (1985) has demonstrated how differences in species concept, even between recent contemporaneous authors, lead to differences in name and number of Clypeaster species recorded from Malta, and to strikingly different faunal lists from the late Miocene of Malta compared to that of Crete. In general, species recognition and comparison from the literature are difficult.

*2 Substantial infrapopulation variation occurs in many test characters commonly regarded as species diagnostic*. In consequence, the number of species described from any region tends to be grossly exaggerated (Challis, 1980; Poddubiuk, 1985).

*3 Few structural innovations punctuate the evolutionary history of the genus*. There are still no generally accepted criteria for recognition of clades, and hence for subgeneric grouping of species.

*4 Gradual phyletic change is seldom discernable*. Most species of Clypeaster are defined on the basis of quantitative differences in a small number of shared test characters. In consequence, phyletic relationships between species are usually obscure, and serial changes that might reliably serve for correlation are not recognizable.

*5 Adaptive strategies are commonly repeated*. Similar morphologies have repeatedly evolved quite independently. For example, the major changes leading to the extant Caribbean species *C. rosaceus* were paralleled during the evolution of some other large and inflated Clypeasters, such as the Mediterranean Neogene *C. altus*. Individual species morphology may thus be of more obvious palaeoenvironmental than taxonomic or stratigraphic significance.

## Main features of interspecific variation

In Clypeaster, interspecific (as constrated with infraspecific) test variation occurs mainly in seven features:

*1 Test size*. Adult size in echinoids is subject to environmental modification, notably by food supply and temperature but may still be an important taxonomic character. While it is not the best indicator of overall size because if shape variation, test length is the measure traditionally used in taxonomy. Within the genus it varies from around 20 mm up to nearly 250 mm.

*2 Ambital outline*. Two characteristics almost invariably covered during description of a Clypeaster specimen are pentagonality and elongation of the test outline in the ambital plane. Degree of pentagonality is very striking in individual specimens, but shows such great infraspecific variation that it is rarely of taxonomic significance. In contrast, degree of elongation is infraspecifically stable, and has both taxonomic and autoecological significance. Width/length ratios in the genus range from 70% to 105%, but in individual species they normally vary by less than 15%. Ratios under 85% are typically associated with an active deep burrowing mode of life; 85–95% with superficial burrowers and also with epibenthic forms; over 95% with low superficial burrowers and ploughers.

*3 Test profile*. The most important aspects of longitudinal test profile are: height relative to length, ambital tumidity and its variation, shape of the ambital plane in cross-section, size and distinctness of the apical mound, nature of any adoral convexity and characteristics of its adambital margins. Test height varies from $<15\%$ to $>65\%$ of test length, but wide infraspecific variation is normal, so it is important as a taxonomic character only at the extreme ends of its range. Marginal tumidity, the thickness of the strongly curved edge zone of the test relative to total test height, may show considerable variation round the test perimeter. It is normally greatest at the anterior end, decreasing to a minimum at a posterior trailing edge. Its maximum value is often of considerable taxonomic use, being generally stable within individual species but varying from 5% to 100% within the genus as a whole. High marginal tumidities and large posterior decrease in tumidity are typical of deep burrowing and ploughing species. Ambital elevation, relative to marginal tumidity or to test height, can also be stable within species and of use in species differentiation and palaeoecological interpretation. In burrowing forms it is always about 50% of margin thickness, and typically 20–50% of test height, while in epibenthic forms it is frequently less than 40% of margin thickness and may be less than 5% of test height. Ambital plane curvature is a much less widely recognised and understood morphological characteristic in Clypeaster. Normally it is negligible but in a few species (e.g. *C. cotteaui*) a distinctive anterior-posterior upward concavity is developed, producing a highly distinctive arching of the lower surface of the test by over 5% of test height. Adoral concavity in Clypeaster ranges from negligible to about 50% of test height. It may occupy only the central part of the oral surface and form a distinct buccal cavity around the peristome or extend more or less uniformly almost to the ambitus, although wide infraspecific variation is common. The adambital margins may be well rounded or severely flattened adorally, the former typical of relatively deep burrowers and ploughers, the latter of epifaunal and superficially burrowing forms where the spines borne on these marginal zones are almost entirely responsible for locomotion.

*4 Test construction.* Whether they act as skeletalized tethers (SEILACHER, 1979) or compressive struts (TELFORD, 1985), the distribution of internal pillars within a clypeasteroid test is generally related to test morphology. In low sharp-edged species pillars are normally small, numerous, and distributed evenly from mouth to ambitus. Forms with higher tests and a less distinct margin have fewer pillars, concentrated midradially. Some large epibenthic Clypeasters have a double rather than single test wall. This can increase test thickness to more than 7 mm (6% of test length) in Recent *C. rosaceus.*

*5 Petal character.* Traditionally the most significant characters in Clypeaster species differentiation have been associated with their petal systems, functionally very important as gas exchange organs. The five petals often differ considerably and should be analysed individually. Infraspecifically the most stable and important elements of petal morphology are length relative to the perradial distance from ocular to ambitus, maximum width relative to length, position of this width maximum, degree of closure relative to it, width of poriferous zones and details of the respiratory pores, most obviously their number. Petal length ranges from <40% to >85% of perradial length. Although subject to some allometric growth and ecological controls its infraspecific variation for any particular petal rarely exceeds 10%. Maximum petal width and its distance from the ocular are both infraspecifically stable but respectively range within the genus from 35–65% and 40–100% of the petal's length. The width of the petal at its distal end is also important taxonomically and ranges from 0 to 100% of maximum petal width. Width of poriferous zones is a measure of respiratory tube-foot width, a direct control on the area of the gas exchange surface. Within the genus it ranges from <15% to >25% of maximum petal width. Number of respiratory pores per petal normally increases throughout life but once adulthood has been reached the addition rate is low. In sexually mature specimens of any one species it rarely varies by more than 10 but within the genus as a whole it ranges from under 20 to over 60. Degree of petal inflation, although highly distinctive in individual specimens, is subject to very considerable infraspecific variation and is of little taxonomic or morphotypic value.

*6 Periproct position.* With few exceptions, periproct morphology and size seem to be of little taxonomic importance in Clypeaster. This is not true of periproct position, which ranges from inframarginal to submarginal, the separation of the rear of the periproct from the posterior margin of the test varying from less than 1% to over 8% of test length. For use in species discrimination, the distance is stablest when described in terms of periproct lengths, ranging from 20—400% on this scale.

*7 Tuberculation.* Clypeasteroid tubercles are partially or completely sunken beneath the test surface and often very well preserved. It is normally possible to deduce from them absolute measures and patterns of spine size, packing density, articulation angle and power stroke direction as well as direct tubercle characteristics such as packing style and area utilization efficiency. Patterns tend to be stable within a species, but may vary significantly between species with differing modes of life.

### Common morphotypes and their distribution

The main features of gross test morphology do not vary independently. Many features almost invariably occur together, thus limiting the number of commonly occurring overall test morphologies. Distinctive and recurring combinations of test characters characterize "morphotypes".

Many morphotypes were regarded as subgenera (LAMBERT and THIÉRY, 1925; MORTENSEN, 1948) until proven polyphyletic origins and the existence of intermediate forms led to their rejection (DURHAM, 1966). The character combinations they possess do, however, have considerable functional and ecological significance. Each main morphotype represents a particular adaptive strategy, rather than a monophyletic taxonomic group. That similar morphologies evolved independently many times is evidence that adaptive strategies were repeated, presumably to exploit the commonly available environmental niches.

The principle can be illustrated by six particularly widespread and distinctive morphotypes. Major features of their morphology are summarized in the accompanying table, time of first occurrence and inferred depth of burrowing in a separate figure. For convenience, each morphotype has been designated here with the name of an early, well-known and widely occurring representative species.

1 '*biarritzensis*' Named after the *C. biarritzensis* species group (which includes *C. trotteri*), widespread in the early Oligocene of S Europe, N Africa and the Middle East (ROSE, 1966; DURHAM and MOJAB, 1983). Many species once included in the subgenera *Biarritzella*, *Laginidea*, and *Guebhardanthus* are closely comparable, including Eocene *C. marbellensis*.

2 '*oxybaphon*' After *C. oxybaphon*, widespread in the mid-Cenozoic of southern USA and the Caribbean islands. Although relatively rare in *Clypeaster* itself, this morphotype is well seen in other clypeasteroids, particularly some Laganids.

3 '*cotteaui*' After the *C. cotteaui* species group (which includes *C. batheri*, sensu PODDUBIUK and ROSE, 1985; PODDUBIUK in prep.) widespread in the late Oligocene of the Caribbean islands; species closely similar in morphology and stratigraphic age also occur in the Mediterranean region (ROSE, 1966). Comparable species have been referred by MORTENSEN (1948) to the subgenera *Pavaya*, *Paleanthus*, *Laubeanthus*, and sometimes *Rhaphidoclypus*.

4 '*concavus*' After *C. concavus*, widespread in the Early/Middle Miocene of the Caribbean region (PODDUBIUK, 1985; in prep.). In the past, species with this morphotype were generally referred to *Bunactis* or *Stolonoclypus*.

5 '*marginatus*' After *C. margitanus*, well-known from the late Middle Miocene of the central Mediterranean region (CHALLIS, 1980). Similar species have generally

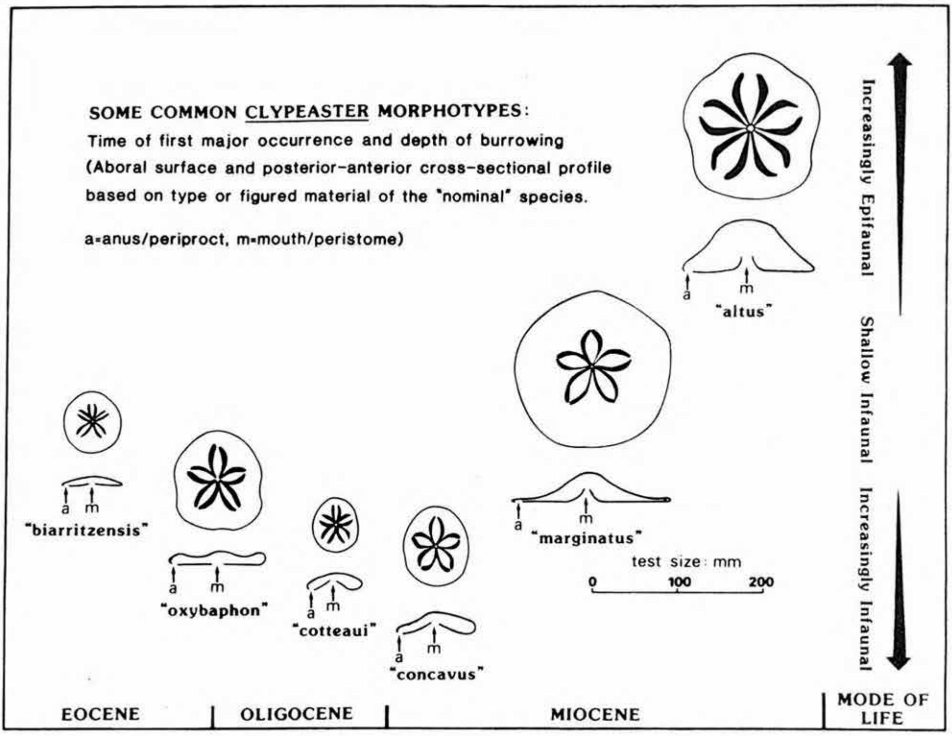


*Fig. 1.*

been ascribed to *Paratinanthus*, *Platyclypeina*, sometimes *Stolonoclypus*, and occasionally *Coronanthus*.

6 '*altus*' After *C. altus*, widespread in the later Miocene of the Mediterranean region (CHALLIS, 1980). This morphotype includes *Clypeaster* (sensu stricto), *Tholeopelta*, *Oxyclypeina*, *Pliophyma*, plus some species formerly referred to *Bunactis* and *Rhaphidoclypus*.

## Conclusion

Time of first appearance and subsequent quantitative importance vary from one morphotype to another with a general pattern of increasing diversity through time. Clypeaster morphotypes, which can easily be recognized in the field, are thus significant indices of maximum stratigraphic age. The '*biarritzensis*', '*cotteaui*', '*concavus*', and '*altus*' morphotypes are broadly characteristic of U. Eocene/L. Oligocene, L/M. Miocene, and U. Miocene/Pliocene strata respectively, at least in the West Tethyan/Caribbean area, although examples of all are still extant.

By analogy with Recent species and analysis of functional morphology it can be inferred that early Clypeasters belonging to the '*biarritzensis*' morphotype were adapted to very shallow burrowing modes of life. Subsequent morphotype development seems to reflect an expansion of the ecological range of the genus, as it adapted

Table 1

**Major features of some common *Clypeaster* morphotypes**

| Morphotype | Test size | Ambital outline | Marginal tumidity | Test profile | Test construction | Petal character | Periproct position | Aboral tuber-culation |
|---|---|---|---|---|---|---|---|---|
| '*biarritzesnis*' | small; typical adult length 70 mm | subpentagonal, subradial | thin; little anterior —posterior differentiation | very depressed; petaloid area little inflated; flat adorally | pillars many, evenly spaced; wall single | widely open, subradially arranged; long—short | well separated from margin | fine, sparse-dense |
| '*oxybaphon*' | medium; typical adult length 110 mm | subpentagonal, bilateral | thick; slight-moderate anterior—posterior differentiation | depressed; low, dished submarginally, petaloid area slightly inflated/concave adorally | pillars few, concentrated ambitally; wall single | slightly open bilaterally arranged; long | separated from margin | very variable |
| '*cotteaui*' | small; typical adult length 70 mm | suboval, bilateral | thick; slight anterior —posterior differentiation | depressed; petaloid area not inflated; concave adorally | pillars few, evenly spaced; wall single | open, bilaterally arranged; mod.—long | well separated from margin | coarse, sparse |
| '*concavus*' | medium; typical adult length 90 mm | subpentagonal, bilateral | thick; strong anterior —posterior differentiation | mod. inflated; distinct wedged petal mound; concave adorally | pillars few, concentrated ambitally; wall single | slightly open bilaterally arranged; long | adjacent to or slightly separated from margin | fine, dense |
| '*marginatus*' | large; typical adult length 150 mm | subcircular, subradial | thin; very slight anterior —posterior differentiation | mod. inflated; distinct but unwedged petal mound; flat adorally | pillars many, evenly spaced; wall single | closed, subradially arranged; short | well separated from margin | fine, dense |
| '*altus*' | large; typical adult length 120 mm | subpentagonal to subcircular, subradial | very thick; little anterior —posterior differentiation | highly inflated; petal mound often indistinct; flat adorally | pillars v. few, distant from ambitus; wall may be double | open, sub-radially ar-ranged; long —v. long | adjacent to margin | coarse, sparse |

to more varied burrowing and surface-dwelling strategies. Initially ('*cotteaui*', '*concavus*'), morphotypes represent increasingly infaunal adaptations. Efficient epifaunal adaptations (e.g. '*altus*' morphotype) occur relatively late in the stratigraphic record.

Variation in the qualitative importance of morphotypes generally may in part reflect changing shallow water carbonate sedimentation patterns. Increasingly epifaunal trends in eastern Caribbean Clypeasters seem to correlate with the spread of seagrass communities there (PODDUBIUK, 1985). Seagrass arrival reduced the amount of ecospace available for shallow burrowers whilst providing a large new food source for epibenthos.

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Proceedings of the VIIIth RCMNS Congress

# PALAEOGEOGRAPHY AND MAMMAL FAUNAS IN THE APULO-DALMATIC AREA

by

C. DE GIULI, F. MASINI, D. TORRE and G. VALLERI

Three fossil mammal complexes have been preserved in the karst structures of the Gargano Peninsula (southern Italy). From the youngest one it can be recognized:

a) *Allocricetus* fauna, made up by horses, bovids, cervids and, among micromammals, *Allocricetus bursae* and *Microtus*. In one case the fauna has been recovered together with pre-Musterian artifacts. The age is Middle Pleistocene.

b) *Allophaiomys* fauna, an assemblage with typical villafranchian elements and some peculiar new species. *Allophaiomys pliocaenicus* is present. The age is late Early Pleistocene (DE GIULI et al., 1985a).

c) *Microtia* fauna, abundant endemic assemblages whose change can be followed over a geologically significant period. The age referred to is the Early Pliocene, but some genera show relationships with Late Miocene immigrants while others can be connected even with Early Miocene forms (DE GIULI et al., 1985b).

This work deals with the palaeogeographic evolution of the South Adriatic area during the Neogene in order to define the correlation between the geographic history of the area and the possible migration routes of the mammal faunas.

In the South Adriatic region it is possible to recognize different structural sectors from east to west:

*1* the Dalmatic platform,
*2* the meso-Adriatic basin,
*3* the Apulian foreland,
*4* the foredeep of the Apennines Range,
*5* the Apenninic system.

These sectors formed a part of an emerged unit during the Neogene which existed in the South Adriatic area; present- day remnants are in the Balkanic and Italian Peninsulas. This unit has been defined as Apulo-Dalmatic Realm (DE GIULI et al., 1986). The main elements of this Realm were the whole Apulian region (Murge highlands, Salento Peninsula, the Tavoliere and Gargano Peninsula), with a northern extension in the Fortore—Sangro area and the shallow sea between the Gargano and the Split—Dubrovnik region (Mid-Adriatic Ridge). To the west and the northwest, no nearby continental area can be envisaged up to the Late Messinian. An emerged area was in existence to the north, made up by the Istria region and the facing part of the Adriatic Sea. Some authors envisage a NW—SE narrow ridge connecting this area with the Apulo-Dalmatic Realm. We are not yet able to define the eastern extension of the Realm in the Balkans and its relationship with other emerged units.

The endemic character of the *Microtia* fauna, namely the low number of supposed immigrants, suggests that the connection with the neighbouring continental areas has never been easy. The low number of immigrants is in antithesis with the strong specific diversity in most assemblages from several fissures. As sister species are the rule in the *Microtia*, fauna an hypothesis of evolution in an archipelagus is highly probable. Thus the Apulo-Dalmatic Realm can be envisaged as formed by structural high blocks often emerged and discontinuously connected.

Palaeographic maps have been drawn concerning different geologic periods:
*1* Early Miocene,
*2* Middle Miocene—Tortonian,
*3* Messinian,
*4* Pliocene (*Sphaeroidinellopsis* and *Globorotalia margaritae* zones),
*5* Pliocene (*Globorotalia puncticulata* zone),
*6* Pliocene (*Globorotalia gr. crassaformis* zone),
*7* Pliocene (*Globorotalia inflata* zone),
*8* Early Pleistocene.

The Pliocene biostratigraphic zonation is according to COLALONGO and SARTONI, 1979. These subdivisions are those allowed by the available borehole stratigraphic data, published by Agip and Montedison oil companies.

On the base of the palaeogeographic sketches it is possible to recognize that an ingressive phase, flooding the Oligocene largely emerged lands, started in the Early Miocene (Fig. 1). Later in the Middle Miocene but mainly in the Tortonian (Fig. 2) the sea spread largely and caused a fragmentation of the emerged lands. Also in the Pannonian basin a distensive phase, correlating to the Tortonian, occurred during the Leithiaian orogenetic cycle (HÁMOR, 1984).

During the Messinian (Fig. 3) the domain of the sea reduced and many formerly isolated areas were again connected. To the north the emerging Apennines, already uplifting since the Tortonian, approached the western border of the Fortore—Sangro area. During the earliest Pliocene (*Sphaeroidinellopsis* and *Globorotalia margaritae* zones) emerged land conditions still persisted (Fig. 4). A new ingressive phase started during *Globorotalia puncticulata* time and progressively advanced southwards even during later periods. In the infra-Pliocene basins of the foredeep the sediments, often turbiditic, were continuously deposited on the Messinian ones only to the north. While

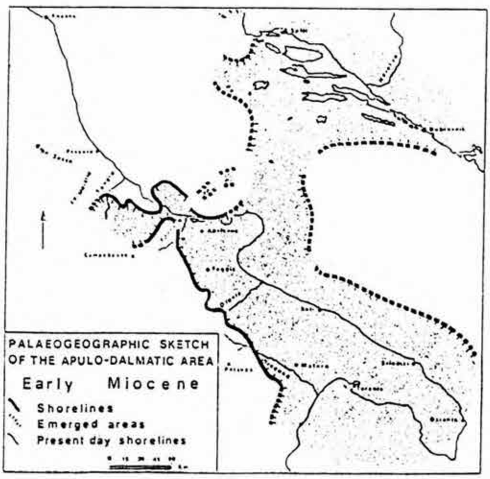


*Fig. 1.*

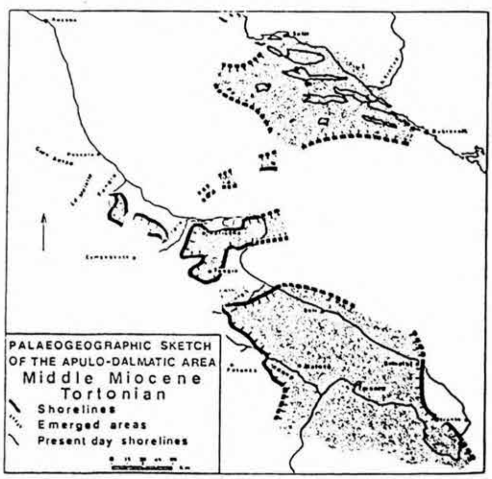


*Fig. 2.*

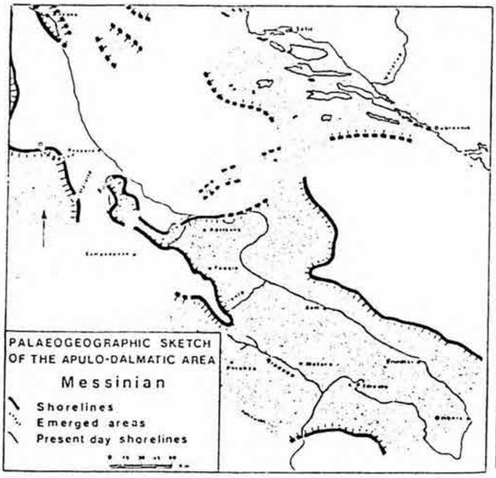


*Fig. 3.*

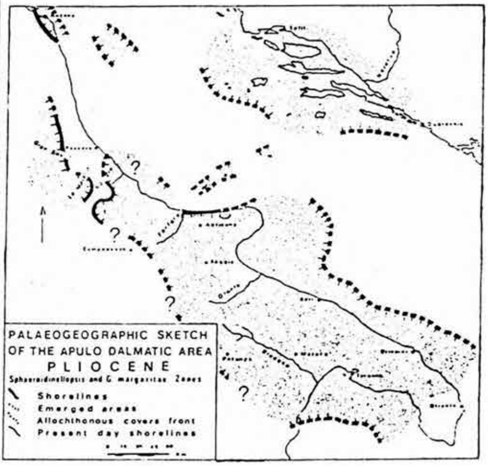


*Fig. 4.*


*Fig. 5.*


*Fig. 6.*

prograding to the south, the sea transgressed a progressively aging substratum, from the Middle Miocene to the Cretaceous. In the palaeogeographic sketch concerning *Globorotalia puncticulata* zone (Fig. 5) one may notice the northern Apennines moving eastward. To the south it is possible to appreciate the Apennines, emerging from the time of the Early Pliocene, approaching the Gargano area and the allochthonous covers to prograde from the west. During the later period, *Globorotalia* gr. *crassaformis* zone (Fig. 6), the general ingressive phase progressed and the basins of the foredeep subsided and spread eastwards. However local tectonics caused a reduction of the sea on the western border of the foredeep, due to the compression generating from the Apennine movement. Meanwhile, to the eastern border, a distensive tectonics renewed the horst-graben structures with fragmentation of the foreland: extremely variable local conditions occurred. Sedimentary gaps or reduced series took place on structurally high blocks which were isolated by sea branches. This situation is sketched in the *Globorotalia* gr. *crassaformis* zone map in which the maximal narrowing of the foredeep is in evidence.

In the Val Marecchia, a locality to the north out of the Apulo-Dalmatic Realm, fishe with some endemic characteristics have been found in Pliocene sapropelitic sediments whose age is referred to *Globorotalia* gr. *crassaformis* zone (SORBINI, 1982). Sapropels imply anoxic conditions in the northern Adriatic sea. The limited water circulation to the west has already been demonstrated from the emerging Apennines, thus we have to believe that to the east the limited circulation in the northern Adriatic basin was produced by the emerging Mid Adriatic Ridge.

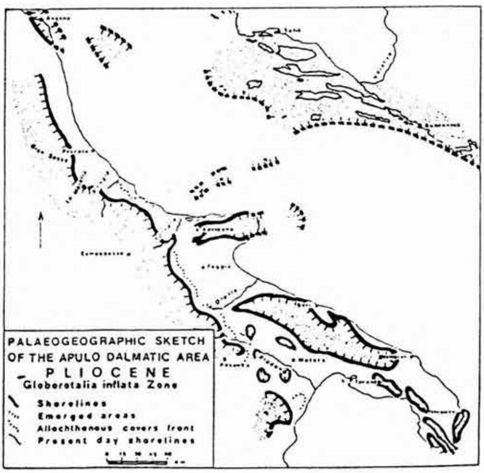


*Fig. 7.*

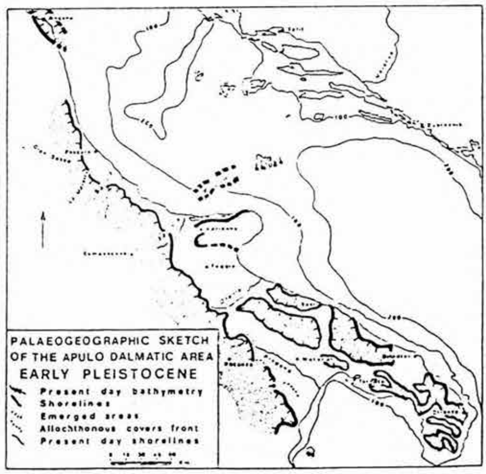


*Fig. 8.*

During the time of *Globorotalia inflata* zone (Fig. 7) and the Early Pleistocene (Fig. 8), the foredeep basins were still subsiding and sea domain was spreading. A contemporary high rate of sedimentation maintained neritic conditions. At the end of the Early Pleistocene a regressive phase started and the coastlines progressively approached the present day ones. No sufficient data are available at the moment to sketch palaeogeographic evolution during the whole Pleistocene.

The reconstruction of palaeographic maps thus confirms the existence of two regressive phases during the Miocene which could allow immigration: an Early/Middle Miocene one and a Messinian/earliest Pliocene one. In the *Microtia* fauna two groups can be recognized whose migration could be correlated to these phases.

The first group, including the erinaceids, glirids, ochotonids and cervoids, was derived from colonizers who immigrated during the first part of the Miocene. The Mid-Adriatic Ridge continuously connected the Apulian platform to the Dalmatic one and could allow mammal arrivals from the east (Fig. 1).

The features of cricetids, *Eliomys*, *Apodemus* and *Microtia* require a younger colonization phase corresponding to the Messinian regression and to the following earliest Pliocene emerged land condition (Figs. 3, 4). Also these migrations were possible only with an eastern provenance.

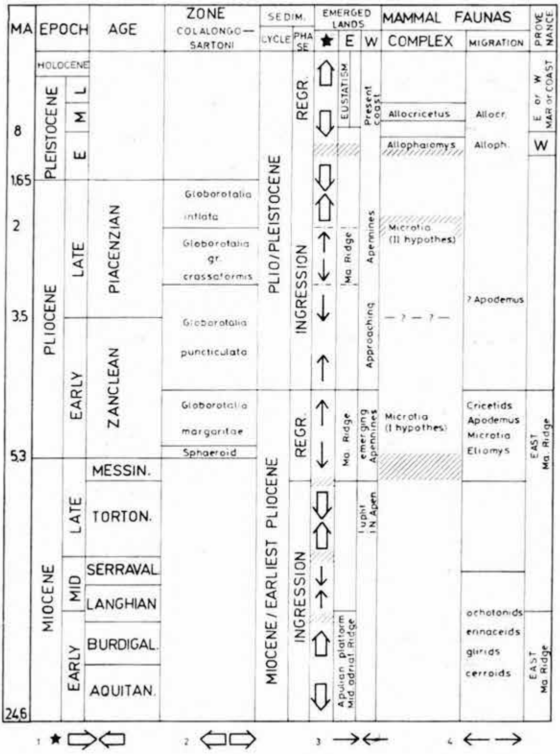


*Fig. 9.* Geologic and biologic events in the Apulo-Dalmatic Realm
*1* Very limited emerged lands, *2* very largely emerged lands, *3* limited emerged lands, *4* largely emerged lands

Two hypotheses may have derived about the *Microtia* fauna age. The first, based on the assumption of the age (*Globorotalia puncticulata* zone) of the oldest sediments capping the "terra rossa" fillings, claims a Pliocene age, older than the last occurrence (about 3.1. Ma) of *G. puncticulata* (Valleri, 1984). The second hypothesis suggests an age limited by the later sea ingression during the *G. inflata* zone (first occurrence about 2 Ma).

The second mammal complex with *Allophaiomys pliocaenicus* is referred to a late Early Pleistocene. Evolutive considerations and correlations with other european faunas suggest an age of about 1 Ma. The migration of this fauna corresponds to sufficient emerged land conditions and dates the regressive phase which cannot be dated by marine biostratigraphic data. In fact this fauna most probably arrived at the Gargano through the Apennines and marks the definitive closing of the foredeep to the North.

The pattern of the *Allocricetus* complex is like the one of other Middle Pleistocene fauna of Italy. Only the occurrence of *Allocricetus bursae* claims an eastern provenance, as this species is known only in the northeastern part of Italy. Once more, geological information does not allow any conclusion as a migration along the Adriatic sea shore could have been possible, although not proven.

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# NEOGENE BIOGEOGRAPHY OF HYRACOIDEA (MAMMALIA)

by

T. KOTSAKIS

Members of the Order Hyracoidea (family Pliohyracidae) are very common in Palaeogene deposits of Africa but unknown from layers of other continents. But in the Neogene both Europe and Asia received immigrants.

During Palaeogene time the subfamily Sagatheriinae is the dominant one. During Rusingan (African equivalent of Agenian and Orleanian) members of the genus *Pachyhyrax*, belonging to this subfamily, are found in East Africa (at Bukwa/Uganda, Rusinga/Kenya etc) (MEYER, 1978). Members of the same genus were discovered at Ad Dabtiyah and at Al-Sarrar, both belonging to the Dam Formation, in East Saudi Arabia, near the shores of Persian Gulf (HAMILTON, WHYBROW and MCCLURE, 1978; THOMAS, 1982). Beside the fact that many rodents of these sites are of Asian origin (SEN, 1982a) the whole fauna indicates African connections. This is not surprising as free migration was apparently possible between Arabia and both North and East Africa during this time. Perhaps a Sagatheriine is present also in Gebel Zelten (Libya) (SAVAGE and HAMILTON, 1973).

But the subfamily who undertook a large expansion during Neogene was, by far, the subfamily Pliohyracinae. Following MEYER (1978) the most ancient findings of this subfamily come from the Rusingan of East Africa, both Lower Rusingan (a, Bukwa) and Late Rusingan (at Rusinga). This material is referred to the genus *Meröehyrax*. A more recent finding in Africa is that of Beni Mellal (Morocco—Middle Astaracian). The specimens discovered in this layer are classified as *Parapliohyrax* (GINSBURG, 1977).

Out of Africa the most ancient findings of Pliohyracine hyracoids, belonging to the genus *Pliohyrax*, were these of the Aegean area, Lower Vallesian in age. They are collected at Melambes in the Island of Crete (Greece) and in Turkey at Esme-Akçakoy (KUSS, 1976; BECKER PLATEN, SICKENBERG and TOBIEN, 1975a, 1975b). A migration of hyracoids from Africa to Aegean area at the moment (or moments) of the arrival (or arrivals) of *Progonomys* and *Hipparion* in Europe and N. Africa is possible but the rodents of Aegeid testify extinctions rather than migrations of African forms at the time of the boundary Astaracian—Vallesian (SEN, 1982b). Thus, a previous migration from Africa, cannot be excluded. In fact, by some palaeontologists (LEINDERS and MEULENKAMP, 1978), the layer of Melambes is considered of the same age with the Cretan layer of Plakias, of Middle or Upper Astaracian age. The genus *Pliohyrax* is unknown from the very few layers of Upper Vallesian age of the Aegeid but is well represented or rather, less rare, in the Lower and Middle Turolian deposits of Greece and Turkey. At Balçikli Dere and perhaps at Kayadibi (Lower Turolian) and at Garkin, Pikermi, Halmyropotamos, Samos (Middle Turolian), remains of one or two species of the genus *Pliohyrax* were collected (MELENTIS, 1966; BECKER PLATEN, SICKENBERG and TOBIEN, 1975a, 1975b; SOLOUNIAS, 1981a, 1981b).

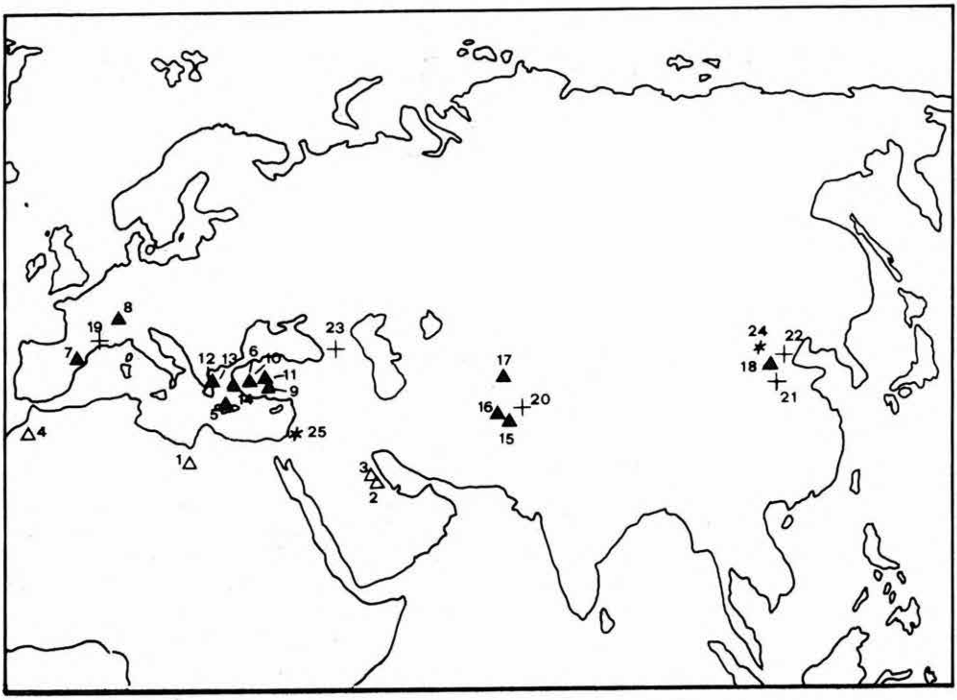


*Fig. 1.* Geographical distribution of the Hyracoidea during Neogene and Quaternary (excluding Sub-Saharan Africa)

*Lower and Middle Miocene: 1* Gebel Zelten; *2* Ad Dabtiyah; *3* Al-Sarrar; *4* Beni Mellal. *Upper Miocene: 5* Melambes (?); *6* Esme Akçakoy; *7* Can Llobateres; *8* Soblay; *9* Kayadibi; *10* Balçikli Dere; *11* Garkin; *12* Pikermi; *13* Halmyropotamos; *13* Samos; *15* Taghar; *16* Molayan; *17* Sor; *18* Baode. *Pliocene: 19* Montpellier; *20* Jalalabad (?); *21* Yüshe basin, zone II; *22* Cap Travertin; *23* Kvabebi. *Pleistocene: 24* Nihewan; *25* Iraq e Zigan

In the other side of Europe, at Can Llobateres in Spain, a well-known site of Lower Vallesian age, few fossils of a hyracoid assigned with doubt to ?*Pliohyrax* are reported (GOLPE POSSE and CRUSAFONT PAIRO, 1981). Remains of a local species of *Pliohyrax* are discovered also in southern France, at Soblay, a layer of Upper Vallesian age (VIRET, 1949). The Upper Astaracian—Lower Vallesian marco- and micromammals discovered in the Iberian Peninsula are very different from those of Maghreb and we must conclude that a Betico-Rifain way of dispersion is highly improbable (BRUIJN and MEULEN, 1981; KOTSAKIS, 1986). Thus, a migration of hyracoids from Southeast Europe is the most probable. Anyway, the scanty record of hyraxes in West Europe and their absence from the deposits of Central Europe indicate that these animals were very rare.

Two Pliocene records of big hyracoids, perhaps representing the final stages of this Upper Miocene expansion, are also known. In the Upper Ruscinian or Lower Villafranchian (sensu AZZAROLI, 1977=Villanyan) layer of Kvabebi, in the Georgian SSR, a great number of hyracoid remains were collected. The specimens are attributed to an endemic Pliohyracine, *Kvabebihyrax* (GABUNIA and VEKUA, 1966, 1974). Very probably this form is the last survivor of the expansion of Pliohyracines in the Near and Middle East. Also in France, a Pliohyracine of big size was found in the Lower Ruscinian of Montpellier (VIRET and THENIUS, 1952). The relationships of this species, attributed either to *Pliohyrax* or to *Postschizotherium*, with the Vallesian ones from the same area or with oriental elements of the Order Hyracoidea is a matter of question.

During Upper Miocene Pliohyracines expanded their area also to Central and Eastern Asia. Remains of members of the genus *Pliohyrax* are collected in the Turolian deposits of Afghanistan and more precisely in the Lower Turolian site of Taghar and in the Middle Turolian site of Molayan (HEINTZ and BRUNET, 1982; BRUNET, HEINTZ and BATTAIL, 1984). It is interesting to note that hyracoids never reached the Siwaliks in spite of the proximity of the Afghan layers (BRUNET, HEINTZ and BATTAIL, 1984).

The layer of Sor, in the SSR of Tadzhikistan, might be also of Turolian age. In this locality remains of a Pliohyracine, named *Sogdohyrax* by DUBROVO (1978), are discovered.

During the last phases of Miocene, Pliohyracine hyracoids undertook an expansion further to east reaching China. The most ancient findings came probably from Baode (Shanxi) from sediments of Baodean age (Chinese equivalent of Turolian) and are assigned to the genus *Pliohyrax* (TUNG YOUNG-SHENG and HUANG WAN-PO, 1974) but this attribution is criticized (DUBROVO, 1978). The other remains of Chinese hyraxes are classified within the genus *Postschizotherium*. Three species are described, but some fossils were bought in drugstores and therefore their geographic and stratigraphic position is uncertain (KOENIGSWALD, 1966). Anyway remains belonging to the genus *Postschizotherium* are found in the second zone of Yüshe basin, of Jinglean age (Chinese equivalent of Ruscinian) and in the Cap Travertin (Hebei-Henan), of Youhean age (Chinese equivalent of Lower Villafranchian or Villanyan). The genus survives during the Lower Pleistocene at Nihewan and other localities (LI, WU and QIU, 1984).

A radius discovered in the (probably) Upper Ruscinian site of Jalalabad in Afghanistan is assigned also to *Postschizotherium* (HEINTZ, GINSBURG and HARTENBERGER, 1978). At the moment of the discovery of this fossil the presence of *Pliohyrax* in the Turolian layers of the same country (Afghanistan) was unknown. At present four hypotheses are possible: *1)* the hyracoid from Jalalabad represents the last Central Asiatic survivor of the expansion of *Pliohyrax* during the Turolian; it may be a relict population like *Kvabebihyrax* in Caucasus and perhaps the hyracoid of Montpellier; *2)* it belongs to the genus *Kvabebihyrax; 3)* it belongs to the genus *Sogdohyrax; 4)* it is really a *Postschizotherium* coming back from China through Central Asia. Perhaps it may be also that the age of the layer of Jalalabad is older than Upper Ruscinian (interv. of DR. E. HEINTZ during the Congress). More material should be available for a solution of this problem.

The history of the recent family Procaviidae begins with the genus *Prohyrax* in the Lower Miocene of Africa. This family includes also a few fossil species of the genera *Gigantohyrax* and *Procavia* from the Plio—Pleistocene deposits of the same continent (MEYER, 1978). A specimen referred to cfr. *Prohyrax* is reported from the Libyan layer of Gebel Zelten (MEYER, 1978).

The recent species *Procavia capensis* PALLAS, the only one of the Order living not only in Africa but also in Syria, Lebanon, Israel, Arabian Peninsula, is known from the Upper Pleistocene deposits of Near East (HELLER, 1978). The mention of the genus *Dendrohyrax* as fossil in my Abstract (KOTSAKIS, 1985) is due to a mere error of typewriting. (For the Mammalian Stages—Ages of Europe, Africa and Asia see MEIN, 1975; SAVAGE and RUSSELL, 1983; CH. LI, W. WU, and ZH. QIU, 1984.)

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# CRICETID RODENTS OF LOWER SIWALIK DEPOSITS, POTWAR PLATEAU, PAKISTAN AND MIOCENE MAMMAL DISPERSAL EVENTS

by

E. H. LINDSAY

Siwalik deposits in Pakistan record a 17 Ma interval of vertebrate history in southern Asia. This record includes the evolutionary appearance and radiation of murid rodents, the most successful group of terrestrial mammals; the radiation of bovids in Asia; the Miocene appearance of the three-toed horse *Hipparion* and the Pliocene appearance of the one-toed horse *Equus* from North America; plus the radiation, and extinction of many other groups of mammals. Siwalik deposits have provided one of the best known records of mammalian evolution. Knowledge of the Siwalik vertebrate record has increased markedly during the last ten years. It was fueled by the quest for knowledge about hominoids that are part of the Siwalik record; it has resulted in a precise chronologic framework based on biostratigraphy and magnetostratigraphy (BARRY et al., 1982; JOHNSON et al., 1982).

The Siwalik fossil record is important to European Neogene stratigraphy because many of the mammals recorded from the Siwaliks are also recorded from deposits in Europe. The Siwalik record, with fossils placed in superposed biostratigraphic sequence and calibrated by magnetostratigraphy, can serve as a test for the European fossil record where superposition and magnetostratigraphy are less frequently available. This paper focuses on the oldest Siwalik deposits to illustrate the radiation and dispersal history of cricetid rodents in those deposits. The Siwalik record of cricetid rodents and their derivatives, support the interpretation that small mammals dispersed between continents during at least two separate intervals during the Miocene. These dispersal intervals are broadly defined at the present time; however, their precise chronologic resolution can very likely be attained with a better fossil record. The intervals I wish to emphasize are at 8—10 Ma and 15—17 Ma. These intervals are represented in the Siwalik stratigraphic sequence in the Nagri and Kamlial formations, respectively, illustrated in Fig. 1 (from BARRY et al., 1985).

Two discrete lineages of Miocene cricetid rodents, the Democricetodonini and the Megacricetodontini, were identified by FAHLBUSCH (1964) and were further exemplified by MEIN and FREUDENTHAL (1971). The Megacricetodont lineage is represented in Siwalik deposits by Megacricetodontini and its presumed derivatives, the Dendromurini and Myocricetodontini. Megacricetodont cricetids can be characterized by an $M_1$ with a strongly bilobed anterocone, upper molars that tend to have a single loph joining the paracone and protocone, and $M_1$ that is usually long with a relatively narrow anteroconid. The Dendromurini differ from the Megacricetodontini primarily in development of lingual cingula and cusps on the upper molars. The Myocricetodontini differ from the Megacricetodontini primarily in reducing the longitudinal connection between cusps (the mure) in both upper and lower molars. Murid rodents are believed derived from Megacricetodonts by developing a lingual cingulum and accessory cusps on the upper molars and reducing the longitudinal connection of cusps, in order to derive the distinctive chevron folds of upper murid dentitions.

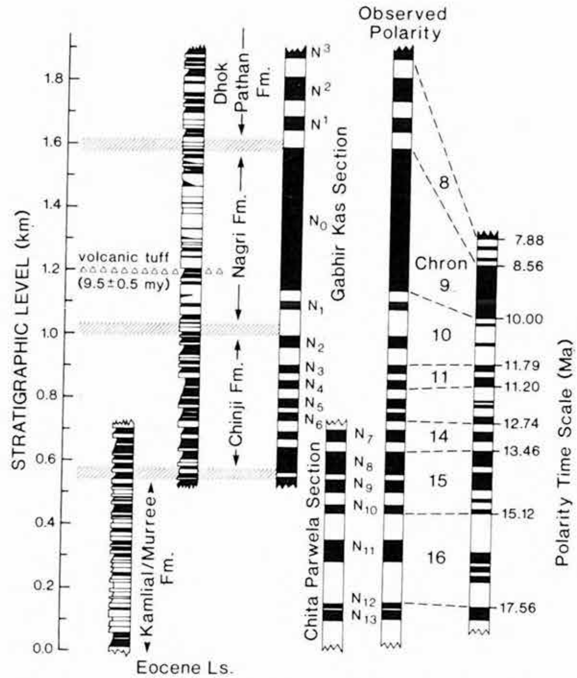


*Fig. 1.* Lower Siwalik stratigraphic and magnetostratigraphic sequence

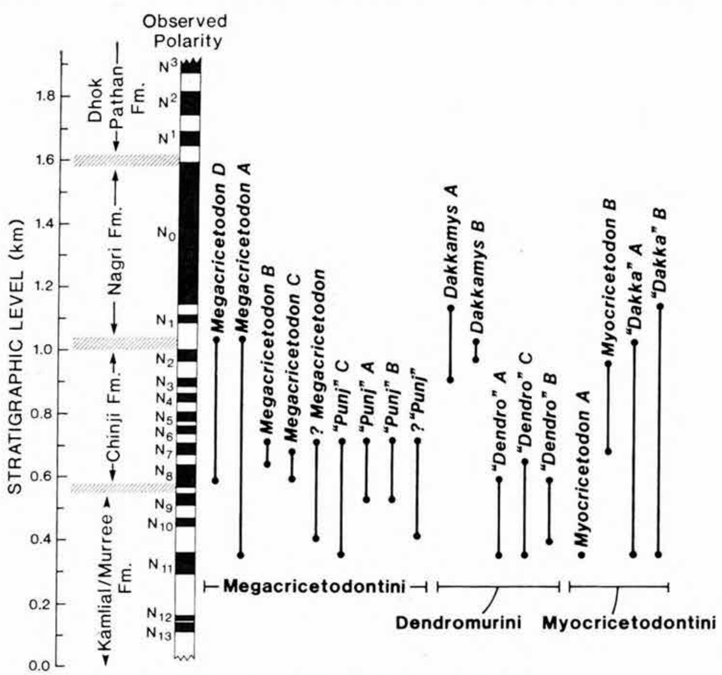


*Fig. 2.* Stratigraphic ranges of Megacricetodontinae

The Siwalik record of Megacricetodont rodents in the Potwar plateau is illustrated in Fig. 2. It should be pointed out that several of these taxa have never been published, although a relatively large sample has been studied. Also, the sample is being expanded, so some of the identifications are likely to be changed. In spite of this, the sample is presently adequate to characterize most of the species and further taxonomic changes are expected to be minor. Note that Megacricetodontini, Dendromurini and Myocricetodontini are all known from the oldest Siwalik faunas with cricetids considered about 16 Ma. This suggest that all of these rodents were thriving in southern Asia just prior to 16 Ma or they underwent an explosive radiation in southern Asia about that time. JAEGER et al. (1985) described a similar dendromurid *(Antemus thailandicus)* and an Asian rodent *Diatomys*, but none of the other cricetid groups, from middle Miocene deposits in Tailand.

The highest diversity, as well as maximum abundance, of these megacricetodont rodents occurs in the Lower Chinji Formation, about 13—14 Ma. Surprisingly, the Megacricetodonts had virtually disappeared from the Potwar plateau by the time the Nagri Formation was deposited about 10—11 Ma.

The second lineage of cricetids in these deposits is the Democricetodontini. Democricetodonts are characterized by $M_1$ having a wide and weakly bilobed anterocone (medial lobe much smaller than labial lobe); upper molars commonly have more than one loph directed toward and frequently joining the opposing cusp; posterior molars are slightly reduced in size especially in the lower dentition. In addition, the longitudinal crest (mure) is very persistent and the mesoloph (mesolophid) is commonly present but short in all molars. Democricetodonts probably gave rise to the Rhizomyidae in southern Asia, although an ancestral species has not been identified.

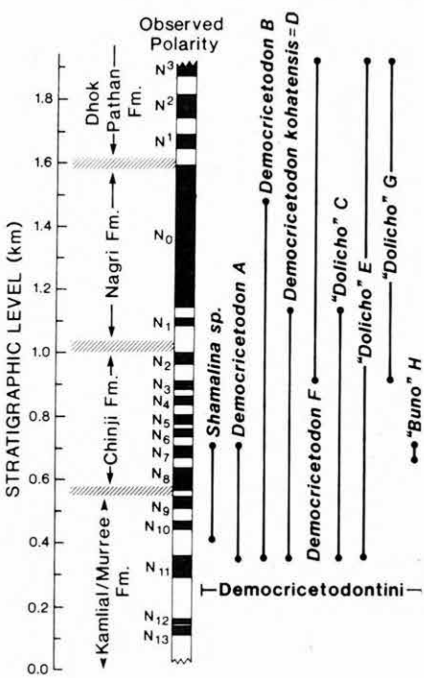


*Fig. 3.* Stratigraphic ranges of Democricetodontinae

*Shamalina*, a small cricetid from early Miocene deposits of Saudi Arabia is also recorded from the Potwar Plateau; *Shamalina* appears closely related to both cricetid lineages.

The record of democricetodont rodents in the Potwar Plateau is shown in Figure 3. Democricetodonts are less diverse but have a longer record than the megacricetodonts; however their maximum abundance is in the Lower Chinji Formation, about the same time as the "blooming" of Megacricetodonts. Probably the decline of cricetid rodents in Middle Siwalik deposits results from competition with the murids, their descendants. It is only the larger cricetids that continue into the Dhok Pathan Formation. By about 8—10 Ma all of the smaller cricetids had been replaced by small murids.

With this stratigraphic—taxonomic framework in mind, we can postulate the intercontinental dispersal history of these rodents. The oldest dipsersal interval, 15—17 Ma, is illustrated in Fig. 4. This interval precedes the appearance of murids; it also records the presence of both Megacricetodontini and Democricetodontini in Europe (e.g., La Grive and Sansan), and in China (the Xining Basin (*Megacricetodon* and *Plesiodipus*)). In the Siwalik deposits, Megacricetodontini, Democricetodontini, Dendromurini, and Myocricetodontini all occur together in the same site. Southern Asia probably records a higher diversity of cricetid rodents during this interval than anywhere else in the world. In Africa both Dendromurini (e.g. *Dakkamys*) and Myocricetodontini (e.g. *Myocricetodon*) are recorded during this interval (e.g., at Beni Mellal, and Ngorora).

The second interval of intercontinental dispersal that I emphasize is the 8 to 10 Ma interval, illustrated in Fig. 5. This interval occurs later than the evolutionary appearance of murids in southern Asia (about 14 Ma), and the appearance (by dispersal) of *Hipparion* into Eurasia. It also approximates the youngest known record of hominoid primates in southern Asia as well as the dispersal appearance of murid rodentsin Europe (Can Llobateres) and Africa (Bou Hanifia). Note that the Megacricetodontini cricetids had become extinct in SE Asia by 10 Ma although some Democricetodontini cricetids persisted with Muridae in Europe (Can Llobateres), in China (Lufeng), and in the Siwaliks as noted above.

AGUILAR and others (1984) have reported the Myocricetodontini *Myocricetodon* and the Dendromyini *Dendromus* along with murids (*Apodemus*, *Stephanomys*, *Occitanomys*, *Castillomys* and *Paraethomys*), Democricetodontini (*Cricetus* and *Calomyscus*) and the gerbilline *Protatera* from deposits near Salobrena in southern Spain. These authors not that the assemblage from Salobrena correlates best with latest Miocene or early Pliocene; that is near 5 Ma rather than 8—10 Ma. These are the only known records of *Myocricetodon* and *Dendromus* from Europe. These genera are well-known from the southern shores of the Mediterranean and perhaps the Salobrena record indicates a shift of southern Mediterranean mammals toward the north at the end of the Miocene. At the least, the Salobrena Spain record indicates that populations of Myocricetodontini and Dendromurini in the Mediterranean long after their extinction in SE Asia.

Myocricetodontini (and Dendromurini?) cricetids occur with murids in North African faunas associated with the 8—10 Ma dispersal interval (e.g. at Oued Zra and Bou Hanifia). Thus this interval of dispersal is best described as the radiation and dispersal of murid rodents associated with "relict" Democricetodontini in Eurasia and with "relict"; Myocricetodontini and Dendromurini in Africa.

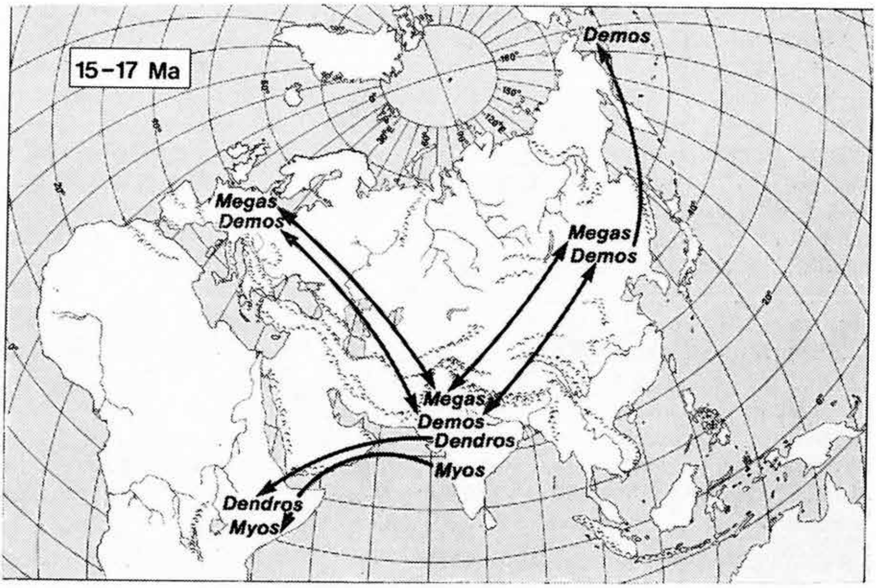


*Fig. 4.* The 15—17 Ma dispersal interval

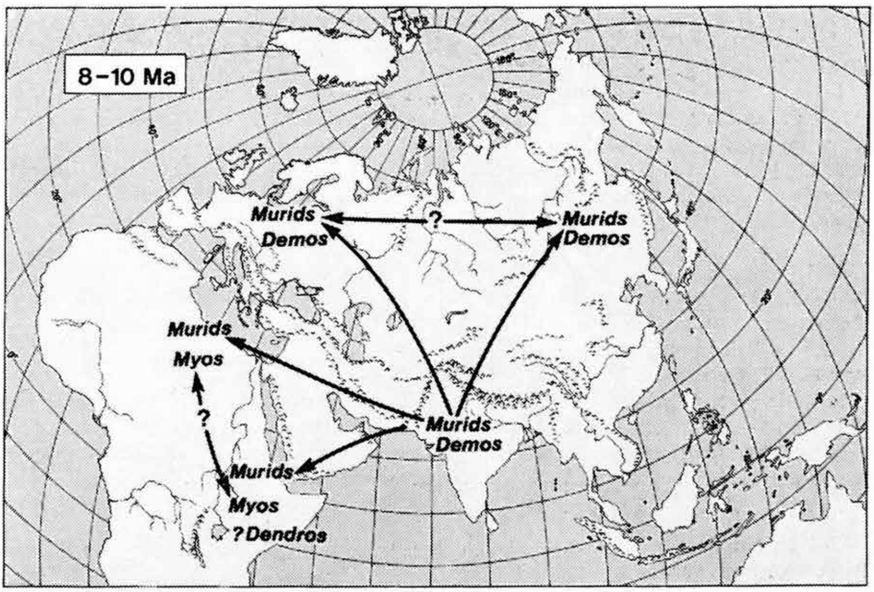


*Fig. 5.* The 8—10 Ma dispersal interval

To summarize, muroid rodents are small, rapidly evolving, locally abundant and widely distributed. They are good travelers and have provided a good fossil record. They are proving extremely useful for biostratigraphic and biochronologic subdivisions. They also promise to clarify the timing and direction of intercontinental mammal dispersal events during the Miocene. These muroid rodents suggest intercontinental dispersal between Asia, Europe and Africa during two broadly defined intervals 15—17 Ma and 8—10 Ma. The first interval (15—17 Ma) probably brought muroids from Europe and Africa to Asia. However, the records of muroid rodents in Asia prior to 17 Ma is poor, so the direction of dispersal cannot be determined with confidence. The later dispersal interval (8—10 Ma) is more clear. This interval appears dominated by dispersal of murids from Asia into Europe and Africa.

In drawing the conclusions present here, I have borrowed from the work of a large consortium of colleagues associated with the Geological Survey of Pakistan and Peshawar University in Pakistan plus Dartmouth College, Harvard University, Yale University, Columbia University, the University of Arizona, Southern Methodist University and the Smithsonian Institution in the United States. I particularly wish to acknowledge my colleagues Larry FLYNN (American Museum of Natural History) plus Louis JACOBS and Will DOWNS (Southern Methodist University) for sharing their knowledge of Siwalik small mammals with me.

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# EVOLUTION OF NEOGENE MEDITERRANEAN VEGETATION AND THE QUESTION OF A DRY UPPER MIOCENE PERIOD (SALINITY CRISIS)

by

H.—J. GREGOR and E. VELITZELOS

Not only in Northern European countries are known rich fossil floras (FRG, GDR, Poland, ČSSR etc) but also in the Mediterranean area, for example from Spain (Tarragona, Izarra, Barcelona, Siurana) and Portugal (Povoa de Santarem, Tagus) to France (Cerestè, Pont-de-Gail, Cessenon, Pichegu, Cerdagne), Italy (Senigallia, Gabbro, Castellina maritima, Florence, Stirone), Greece (Kumi, Aliveri, Vegora, Likudi, Prosilion) and Turkey (Çanakkale, Soma, Sahinali, Saray, Edirne) (see Fig. 1). Most of these are of leaf character, but also fructifications and pollen-grains occur in these floras, belonging to a period which reaches from the Oligocene through the Miocene to the Pliocene, including one part of the Pleistocene, too (see Documenta naturae 25, 1985; GREGOR, 1983).

Through the ages there are equivalent facies like open water-, reed-, swamp-, bottomland forest—and mixed mesophytic forest (see MAI, 1981).

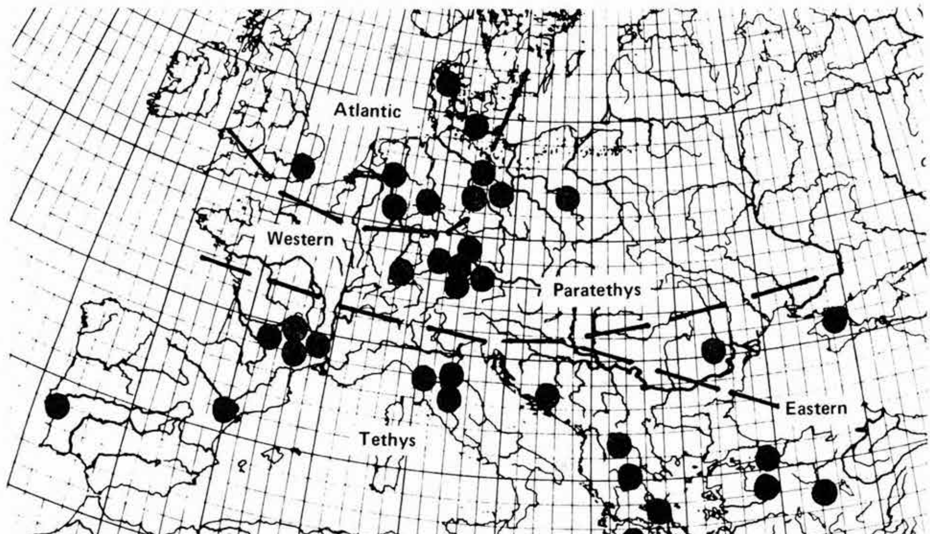



*Fig. 1.* Localities of fossil floras in Europe in the different areas: Tethys, Paratethys (eastern and western), Atlantic—Boreal

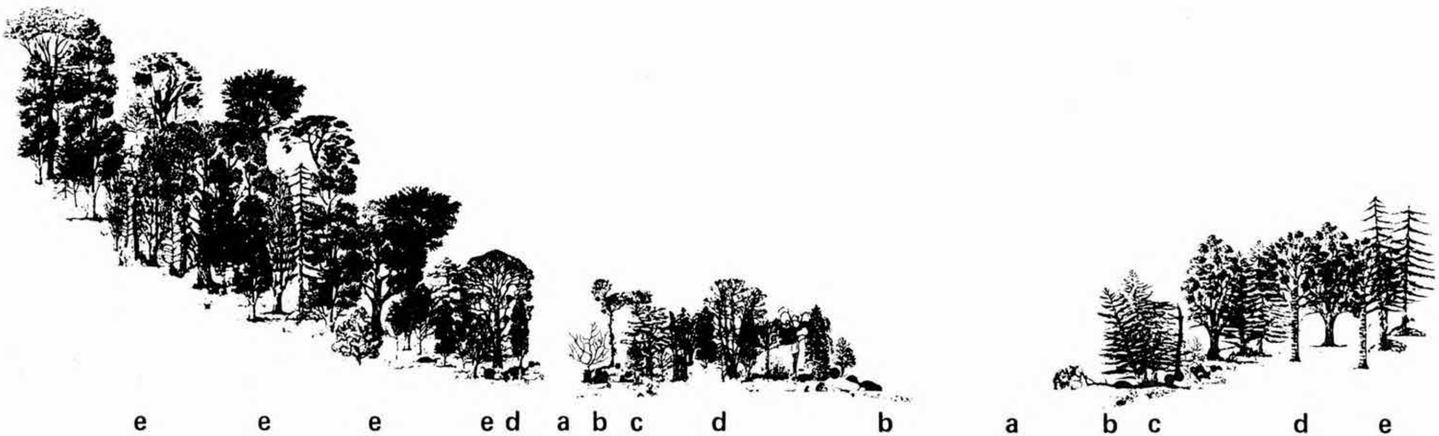



*Fig. 2.* Different facies-types of fossil floras, equivalent in Paratethys and Mediterranean floras
a = open water, b = reed, c = swamp, d = bottomland forest, e = mixed forests

Important plant elements for these types were for example (see Fig. 2):

*a)* open water: Nymphaeaceae, Brasenia, Stratiotes, Ceratostratiotes;

*b)* reed: Bolboschoenus, Scirpus, Schoenoplectus, Carex;

*c)* swamp: Glyptostrobus europaea, Taxodium dubium, Myrica, Spinophyllum;

*d)* bottomland forest: Populus, Salix, Gleditsia, Cinnamomum (vel Daphnogene), Alnus, Betula, Fraxinus;

*e)* mixed-mesophytic forest: Quercus, Laurus, Cinnamomum (vel Daphnogene) Sapidus, Acer, Ulmus, Fagus, Paliurus, Pinus, Carpinus, Cephalotaxus, Liquidambar, Mastixia.

If we have a look on the composition of the different facies in the run of the time we see a slow change from a highly evergreen to less exotic and more native vegetation, running to the Recent one in later Pleistocene times (Fig. 3).

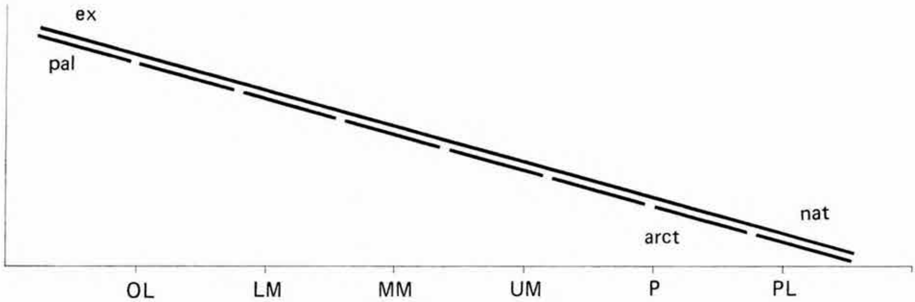



*Fig. 3.* Decline of exotic (ex) and palaeotropical (pal) elements in Paratethys floras (—) and Mediterranean ones (----) to native (nat) and arctotertiary (arct) composition (a model)

We have same shift from palaeotropical elements to arctotertiary elements for example in Germany (GREGOR, 1982), Poland etc. If we compare the elements of the Mediterranean and the Paratethys-floras we discover a high degree of similarity, not comparable to the Recent floras as these are connected with vegetation zones and different climate types. Some differences in the floral record may be mentioned here. In contrast to Paratethys-areas, the common Mastixias we know from there are nearly all missing in the Tethys area, including many subtropical elements like Symplocaceae, Theaceae, Magnoliaceae etc (see GREGOR, 1978; 1980) (Fig. 4).

All floras from the Paratethys area and the Tethys area indicate wet or humid subtropical (warm-temperate) climate of Cfa-type (Virginia-climate, sensu KÖPPEN, see Fig. 5). We only have to distinguish between the palaeotemperature curve over wide areas from an Eocene Af-Cfa-climate to a Plio—Pleistocene Cfa (Cw)-Cs-climate (see Fig. 5). The special climate-diagramms of certain smaller areas (sensu WALTHER, see Fig. 6) give an idea about local factors like humidity, wind, frost, desiccation etc.

To give an impression about the real concordance of fossil floras in the Paratethys and Tethys areas, some examples are brought here (as extracts only), see Documenta naturae 25, 1985; 29, 1986; GREGOR, 1978, 1980, 1982;

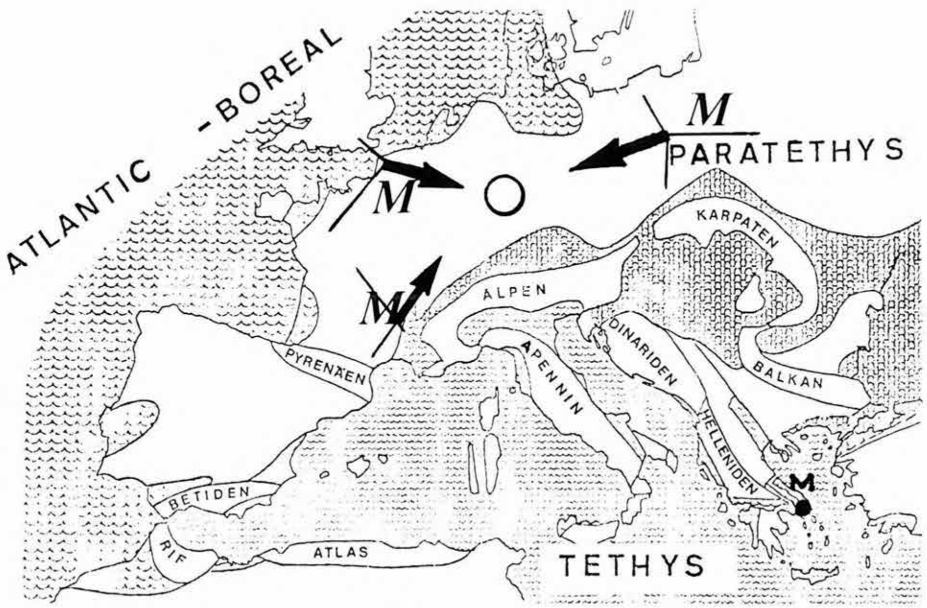



*Fig. 4.* Europe in the Lower Miocene with Mastixioidean floras (M) a possible centre of this typical flora and the only Greek locality yielding such types

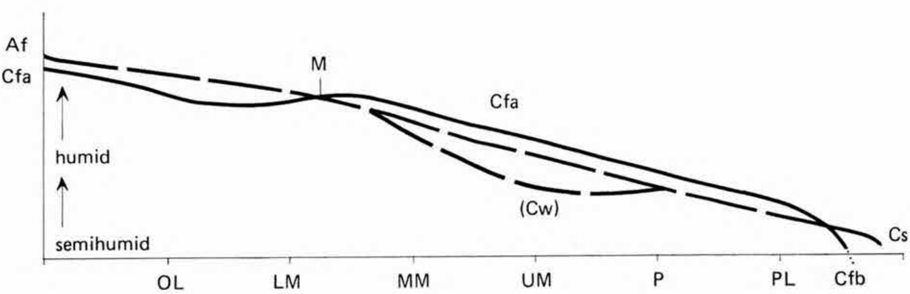



*Fig. 5.* Principal model of the decline of the palaeotemperature curve from Oligocene (OL) to Pleistocene (PL) in Paratethys-area (—, with Mastixioidean maximum M) and the Mediterraean (----, with elongated line in Pleistocene and missing Lower Miocene temperature maximum, lack of Mastixias)

*Lower Miocene*

Floral list from Arjuzanx (France) and Schwandorf (FRG):

Carya ventricosa
Liquidambar magniloculata
Magnolia sp.
Mastixia sp.
Myrica stoppii
Nyssa ornithobroma
Ocotea rhenana
Parabaena europaea
Retinomastixia oerteli
Sapindoidea globosa
Spinophyllum daemonorops
Symplocos div. sp.
Toddalia rhenana
Trigonobalanus sp.
Zanthoxylum sp.

*Lower Miocene*

Floral list from Aliveri (Greece), Langenau and Schwandorf (FRG):

Glyptostrobus europaea
Myrica div. sp.
Rubus laticostatus
Toddalia div. sp
Zanthoxylum ailanthiforme
Paliurus sibiricus
Sambucus pusilla
Sparganium camenzianum
Carex plicata

*Middle Miocene*

Floral list from Soma (Turkey) and Oehningen (FRG):

Cinnamomum div. sp.
Myrica lignitum
Laurophyllum princeps
Quercus div. sp.
Acer trilobatum
Salix varians
Populus latior
Gleditsia sp.
Glyptostrobus europaea
Magnolia ludwigii
Tilia sp.
Cercis sp.

*Upper Miocene*

Floral list from Likudi and Vegora (Greece), Achldorf and Massenhausen (FRG):

Paliurus thurmanni
Pinus sp.
Alnus sp.
Carpinus grandis
Corylus sp.
Populus sp.
Ceratophyllum vösendorfense
Sassafras sp.
Platanus leucophylla
Zelkova zelkovaefolia
Fagus sp.
Quercus div. sp.
Ostrya sp.
Alnus ducalis
Acer div. sp.

*Plio—Pleistocene*

Floral list of Stirone river and St. Barbara (Italy), Frankfurt and Willershausen (FRG):

Carya angulata
Juglans bergomensis
Liquidambar europaea
Cephalotaxus sp.
Corylus avellana
Picea sp.
Pinus sp.
Platanus aceroides
Populus latior
Salix varians
Quercus div. sp.
Acer integrilobatum
Carpinus grandis
Juglans acuminata

The question arises whether there was a "Messinian crisis", a desiccation of the Mediterranean sea—or not. Now as we have seen—all floras are of wet character and show no hint of any arid climate or a typical arid phase neither in the Messinian nor anywhere else. In contrast—the floras in gypsum or salt are of the same type as before or after the deposits of Messinian age (6 m.y. ago, see Fig. 4, 6).

To state this floral behaviour we will show here some typical Messinian floras (and from the time shortly before and afterwards) in their composition of elements.

The floral remains all show a humid (semihumid!) hinterland area with dense forests. As we have so many similar floras the question of a local "oasis"-type vege-

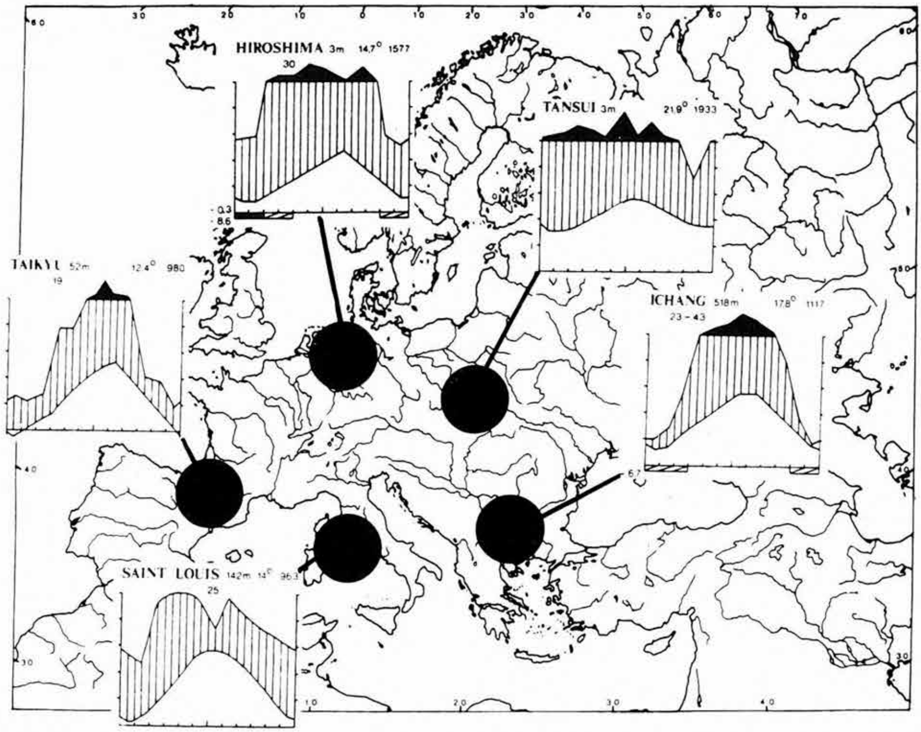



*Fig. 6.* Theoretical model for climate diagrams in Europe for Upper Miocene times (salinity crisis) to show the differences in a Cfa-climate. In the Mediterranean we have less rain per year and a certain "winterdry" season

| *Taxa* | | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| Platanus | + | + | | + | + | + | + | + |
| Populus | + | + | + | + | + | + | + | + |
| Acer | + | + | | + | + | + | + | + |
| Quercus | + | + | + | + | + | + | + | + |
| Ostrya | | + | | + | | + | | + |
| Carpinus | + | | + | + | | + | + | + |
| Betula/Alnus | | | + | + | + | + | | + |
| Zelkova | + | | | + | + | + | + | + |
| Pterocarya | | + | | + | | + | + | + |
| Cinnamomum | + | + | + | | | + | + | + |
| Glyptostrobus | | | + | + | | + | | + |
| Taxodium | | | | | | + | + | + |
| Ulmus | | + | | + | | | + | + |
| Leguminosae | + | | | | | + | + | + |
| Salix | + | + | | | | + | + | + |
| Fagus | + | | + | + | | + | | + |
| | *1* | *2* | *3* | *4* | *5* | *6* | *7* | *8* |

Fossil floras from *1* Chios: Nenita-Strata, *2* Korfu: Paghi-Strata, *3* N Evia, *4* NW Macedonia, *5* Strymon Basin, *6* Turkey, *7* Italy, *8* SW Germany.

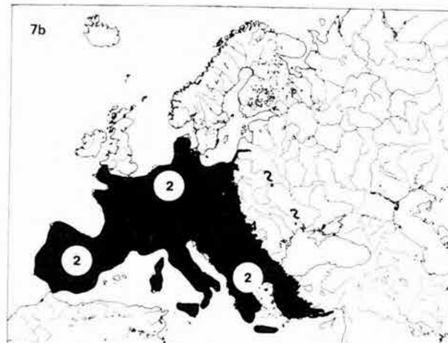
tation is *not* given. The reason for the deposition of huge gypsum-horizons at that time is (as we think) not due to on arid climate but to semihumid conditions or a Cw--climate-type (see Fig. 5). This means that we have no desert climate with summer desiccation and partly winter draught (like in a Cs- or Bs-climate), but a drier (semi-humid means for example 700—1000 mm rain/year instead of 1500—2000) one or a winter dry Cw-climate with one to three months less rainfall rather than desiccation (Fig. 6).

Accordingly the gypsum deposits in the Messinian time-span and other ones perhaps need another explanation, which is not our problem (but see SONNENFELD, 1984).

To prove the given information we bring here some additional data from faunal and other records at this time. In Pikermi for example 80% of the animals are of wetland or swamp-character, only 20% grazing animals in more open countries.

Also the huge gravel beds with well rounded pebbles from Pikermi gorge (Megalo-rhema) cannot be explained by less rainfall but by heavy seasonal rainfall (as in Cw or Cfa-climate); the flora (Doc. nat. 25. p. 3, 1985) is also of typical wetland facies. On the contrary many special question remain, including monsoon-climate, in which way the rain was distributed per year (one or two peaks e.g.) or the endurance of frost or snow or the question of local swamp or wetland areas in an relatively open country forest.

The following picture tries give a model of a mediterranean vegetation in Early Miocene, Late Miocene (Messinian) and Plio—Pleistocene times. The forest systems cover the existing areas at the mentioned times and are connected with those of the Paratethys (Fig. 7, climate and vegetation models).

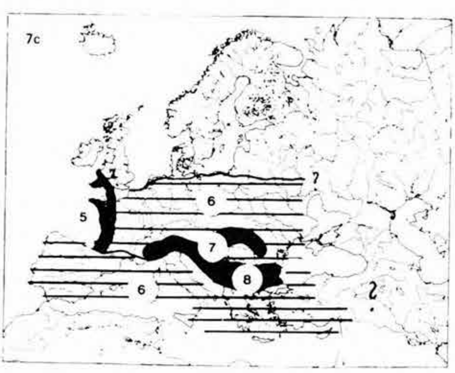




*Fig. 7a, b, c* Model of a Neogene vegetation of the Tethys- and Paratethys-area in connection with the climate (Cfa)

*a* = Recent climate-types in Europe, *b* = Neogene climate-type (Oligocene—Pliocene) in Europe, *c* = vegetation types at the end of the Neogene.—*1* Cfb-climate (Fagus-climate), *2* Cfa-climate (Virginia-climate), *3* Cs-climate (Mediterranean-climate), *4* alpine-climate (elevation over 1000 m), *5* oceanic Pinus—Sequoia—Ericacean vegetation, *6* mixed mesophytic forest with Taxodium, Glyptostrobus, Nyssa, Fagus, Quercus, Sciadopitys, Mastixia, *7* montane Tsuga—Chamaecyparis forest, *8* Kolchis forest

The next important question concerns the Plio—Pleistocene boundary, the Villafranchian. The flora from the Stirone river for example, of Calabrian age (Pleistocene!) is of *typically* Pliocene character (see Doc. nat. 25. p. 31) comparable with those of Frankfurt, Czorsztyn, Kroscienko. So the first cold phase had, as it seems, no effect in the Mediterranean, and we have to postulate a divergence of fossil floras in the run of the Pleistocene—between the Tethys and Paratethys floras. Here we have another topic—the question of the glacial and fluvial episodes and the floras of these times.

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# A COMPREHENSIVE STUDY OF THE HUNGARIAN NEOGENE FLORAS

by

L. HABLY

This paper reports on the complex computer-assisted study of the 31 most significant Hungarian Neogene flora assemblages based partly on literature as well as the scientific investigations of the author. The evaluation includes the change of the Arctotertiary/Palaeotropical elements' ratio, the evolution of the lamina margin types, the relation (descent) of the floras, as well as their distribution according to biotopes. The dendrogram obtained by cluster analysis presents the similarity relations of the floras.

Fig. 1 demonstrates the ratio of Palaeotropical and Arctotertiary elements in the 31 floras studied. It is conspicuous that the number of *P* elements in the Lower Miocene floras are higher than or equal to that of the *A* elements. This is especially prominent in case of the Ipolytarnóc flora. Subsequently, *A* elements are always more important than the *P* ones.

Fig. 2 shows the evolution of the leaf lamina margin. Three categories were established that is the entire leaf toothed leaf and needle-leaf types. The line corresponding to the entire leaf and toothed leaf categories run roughly parallel to the *P* and *A* lines, respectively, as one could suppose. This means that the majority of the palaeotropical elements are entire-leaf types, with the laurels dominating, corresponding to the laurels forests characteristic of the subtropical climate.

In case of the grouping according to biotopes we can see, that the number of lacustrine and palustrine species can be high in case of some Lower Miocene and Sarmatian floras, becoming generally high, however, in the Pannonian period. Some Sarmatian floras might contain species characteristic of the flood plain areas, while the species of laurophyllous forests dominate in the Lower Miocene, floras as well as some in Lower Sarmatian ones. Alpine species are generally rare, while dry biotopes are common in case of the Lower and Upper Sarmatian floras.

In the genetical grouping of the species those geographical regions are named which are the present areas of the modern equivalents of the fossil species. As it is apparent from Fig. 3, species with SE Asian (SEA), North American (NAM), and Caucasian (CAU) affinity were the most important in the Neogene. These were followed by the Mediterranean species and those of Asia Minor. The Central European, South European and East Mediterranean elements were subordinate. The SEA elements are present in almost any Neogene flora, and dominate the Lower Miocene and Lower Sarmatian assemblages. Caucasian elements are scarce in the Lower Miocene floras but they are represented in considerable numbers in the Lower Sarmatian and younger assemblages. The North American affiliation species reach the highest values in some Sarmatian floras as well as on a Pannonian locality. Mediterranean elements are present in course of the whole Neogene, especially increasing, in the Lower Sarmatian. Eastern Mediterranean elements are met in the

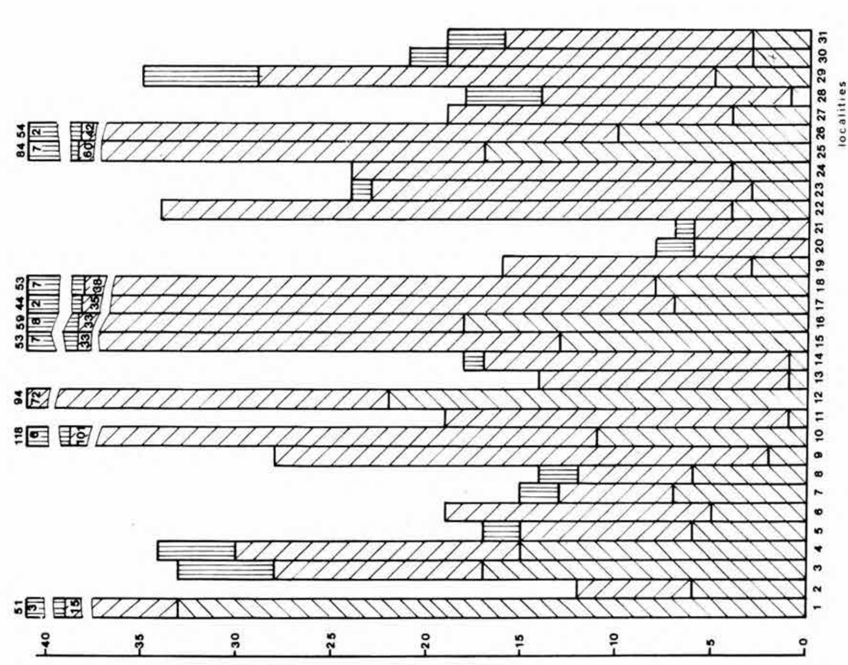


*Fig. 1.* The ratio of arctotertiary and palaeotropical elements in the Neogene floras of Hungary

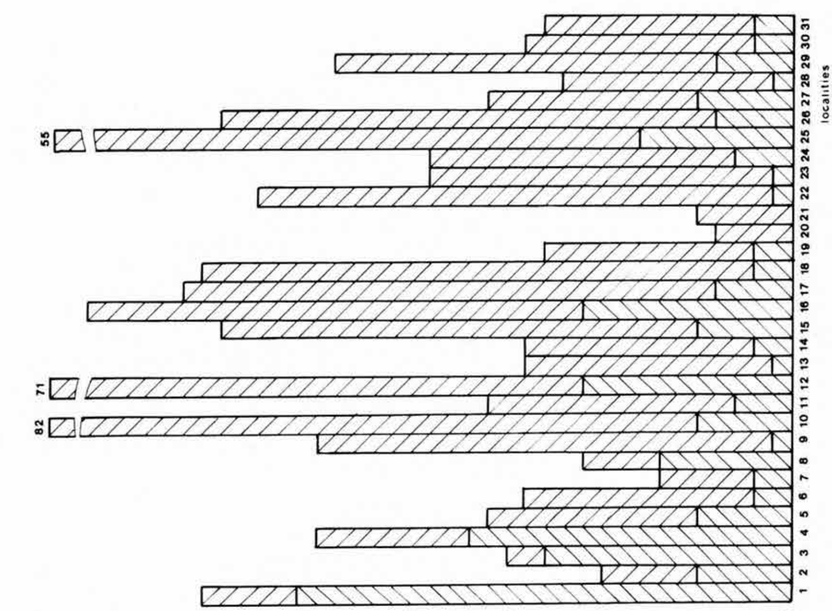


*Fig. 2.* Types of lamina margins in the Neogene floras of Hungary

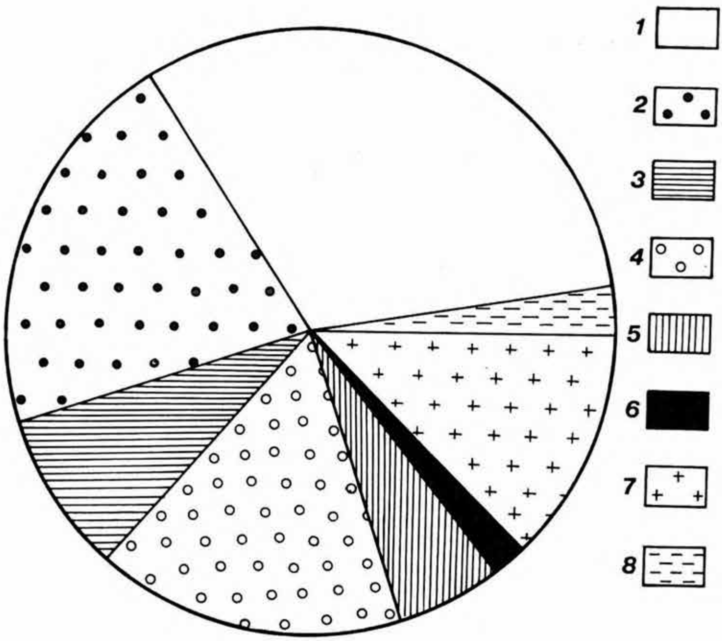


*Fig. 3.* The geographical distribution of the species in the Neogene floras of Hungary
*1* Southeast Asia, *2* North America, *3* Asia Minor, *4* Caucasus, *5* Central America, *6* East Mediterranean, *7* Meditteranean, *8* South Europe

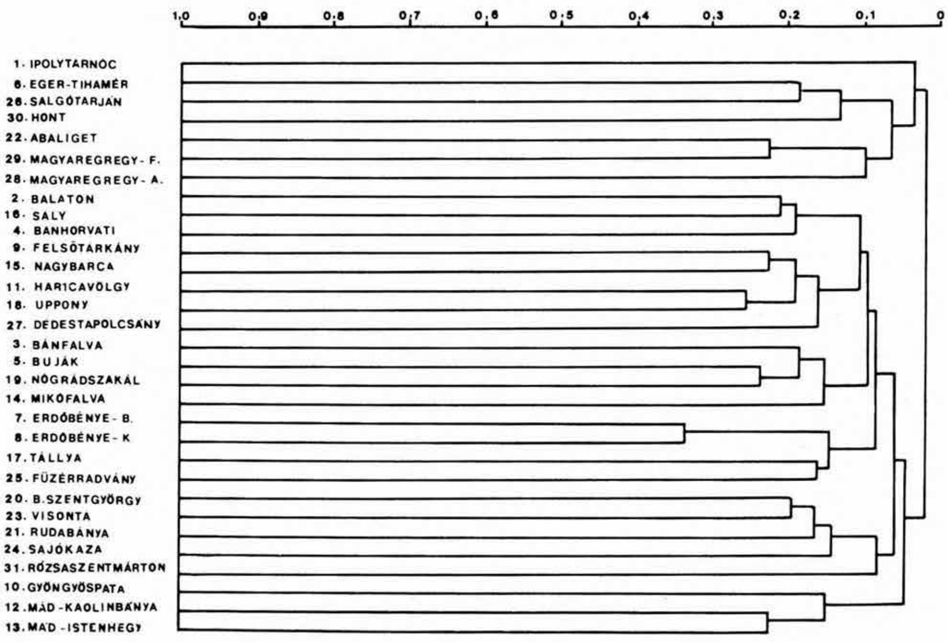


*Fig. 4.* The Hungarian Neogene leaf flora dendogram of Czekanowski index

Sarmatian only. The Central European elements are constantly present in minor numbers. The South European ones appear in some floras only. Species of Asia Minor occur only sporadically in assemblages preceding the Sarmatian, while they are present, to some extent, in almost all Sarmatian floras. On the dendrogram (Fig. 4) there are four distinct units that can be fairly well separated from each other: I Ipolytarnóc—Magyaregregy, II Balaton—Füzérradvány, III Balatonszentgyörgy—Rózsaszentmárton, IV Gyöngyöspata—Mád (Isten-hegy). In the first group, floras of Ottnangian, Karpatian and Badenian age are assigned only. The second group comprises Sarmatian flora with the exception of one locality, the third group consists of Pannonian assemblages with one Sarmatian locality. The fourth group consists of Badenian and Sarmatian floras.

*I* The first group comprises seven localities. Their age is Ottnangian, Karpatian and Badenian. The Eger (Tihamér fields = Lower Badenian) and Salgótarján (Ottnangian) floras show greater similarity, to each other than with their respective contemporary floras. The Eger "Tihamér" flora is rather far from the Nógrádszakál flora, belonging also to the Badenian. Its connection with the Salgótarján flora is due to the fact that both of them are dominated by palustrine species. This reflects the effect of factors other than geological age in the similarity values namely climate and ecology. The characteristic plants that are common in these assemblages are Myrica, Phragmites, Typha, Pronephrium stiriacum. The Lower Badenian of Hont flora contains, apart from the palustrine species shared with the above localities considerable amount of laurels like Daphnogene and Laurophyllum. The Lower and Middle Miocene localities of the Mecsek Mountains appear separately from the above north Hungarian sites on the dendrogram. Abaliget (Ottnangian—Karpatian) is nearest to Magyaregregy (Farkasordító ditch) (Karpatian). These floras are composed of swamp vegetation as well as of laurels. The Ottnangian flora of Magyaregregy (Almás fields) is connected to the above localities on account of the laurophylles. As opposed to the above floras, this site contains no palustrine vegetation beside the palaeotropical elements, but that of the flood plain area (Liquidambar, Populus, Parrotia). Floras of northern Hungary and the Mecsek Mts are connected only at this level. The Ipolytarnóc Ottnangian flora can be connected to the above floras at a rather low mean similarity value. It can be attached to the above floras on account of the Lauraceae (Daphnogene, Laurophyllum).

In the group discussed above the Nógrádszakál Badenian assemblage as well as the Gyöngyöspata Lower Badenian flora are missing. These differ not only from the other Lower and Middle Miocene floras, but from each other as well. In Nógrádszakál, the dominating species are typical of the flood plain area environment, especially the higher river flats. Palustrine species, as well as the laurels ones are missing or they are present in very low numbers. As it is apparent from the dendrogram, the flora stands nearest to the Sarmatian floras of group II. The Gyöngyöspata flora assemblage is the nearest to the floras of Mád—Kaolinbánya and Mád (Isten-hegy), both of them preserved in kaolinite (Group IV); however, the mean similarity values are rather low. These floras comprise, on one hand, the swamp vegetation, on the other hand, some xerophilic elements. Their similarity can be explained by the latter. The flora of these three sites are, due to their special biotopes, clearly distinct from all other assemblages.

*II* Floras of the greatest similarity belong to this group. Except for the Nógrádszakál locality, all assemblages are the Sarmatian. Harica valley with Uppony and Felsőtárkány with Nagybarca: similarity here is apparent. Environment must have been a decisive factor. Most of the species are of the food plain area, they are dominating the assemblages. Beside the species characteristic of the river flats, the xerophilic *Quercus pontica-miocenica* is also significant in the flora. The Dédestapolcsány assemblage is only slightly different from these. Balaton—Dellő with Sály and with Bánhorváti form a slightly separate unit, differing from the other assemblages on account of the presence of *Zelkova zelkovaefolia*, *Fagus* div. sp., *Carpinus grandis*. Their similarity arises from the presence of the flood plain area elements like Parrotia, Alnus, Betula, Ulmus. Buják with Nógrádszakál: The Nógrádszakál flora of Badenian age shows the greatest similarity with that of the Sarmatian Buják assemblage. This is due partly to the presence of the species *Parrotia pristina*, *Quercus pontica-miocenica*, both of them belonging to the east Mediterranean food plain area types, as well as the high river flats' Ulmus species. Essential differences between the two floras are, for example the absence of the Liquidambar at Nógrádszakál, which is dominating in the Buják flora. The Buják flora is connected to the Sarmatian ones by the presence of *Zelkova zelkovaefolia*, *Quercus kubinyii*, that are not present in the Nógrádszakál assemblage. The flora of Bánfalva is connected to the above assemblages by the presence of the east Mediterranean elements. The Mikófalva flora is fairly different from these. Their connection can be demonstrated in the species characteristic of the flood plain area biotope, such as *Populus* div. sp. Within group II, the following 4 localities are very close to each other. Erdőbénye (Barnamáj) with Erdőbénye (Kővágó-tető) are, naturally enough, the most similar among the Neogene floras. They are very near to each other in time and space. Xerophillic elements are dominating here. Tállya with Füzérradvány: these two essentially different floras are linked together with the Erdőbénye flora by *Gleditsia knorrii*. The flora of Tállya is fairly similar to that of Erdőbénye. From the presence of the laurophyles (*Daphnogene* div. sp.), however, it can be inferred that a subtropical vegetation of higher humidity requirements existed contemporary to these assemblages at Tállya. In the Sarmatian flora of Füzérradvány, supposed to be younger, swamp vegetation had also some role. The presence of the *Quercus* div. sp. and the *Gleditsia knorrii*, however, connect this assemblage to the above mentioned floras.

*III* Five assemblages belong to this group. Apart from the Sarmatian Sajókaza, they represent Pannonian sites. All of them contain typical swamp vegetation. In the Sarmatian flora of Sajókaza as well as the Lower Pannonian of Rudabánya, the species Pterocarya can still be found, but it is already absent from the younger floras. Typical Pannonian swamp vegetation is formed by the species *Glyptostrobus europaeus*, *Byttneriophyllum tiliaefolium*, *Phragmites* sp., *Typha* sp., *Salix* div. sp. (Balatonszentgyörgy with Visonta). We find same young flora assemblage in the Upper Pannonian of Rózsaszentmárton as well. From more distant areas, a Daphnogene leaf had been drifted to the sedimentary basin. This is the youngest occurrence known for the genus Daphnogene in Hungary.

The floras belonging to Group IV have been already analysed in connection with the Gyöngyöspata Badenian assemblage.

As we can see from the above discussion cluster analysis cannot be used for a chronological ordering of the localities, i.e. for biostratigraphical purposes. Its biostratigraphical relevance is only approximate, ecological and climatological factors are also taken into consideration.

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# PALYNOLOGICAL CORRELATION OF CONTINENTAL AND MARINE NEOGENE SEQUENCES IN ISRAEL

by

A. HOROWITZ and B. DERIN

*Introduction.* The continental Neogene is well developed in Israel, comprising fluviatile, lacustrine and lagoonal sediments deposited in successive drainage systems, which were controlled by the changing sea levels and the various phases of tectonic activity along the Jordan—Dead Sea rift system (GARFUNKEL and HOROWITZ, 1966; HOROWITZ, 1979). The dating of these formations was quite problematic since hardly any fossils were ever recovered from the predominantly clastic rocks and views differed as to the stratigraphic and palaeogeographic frameworks.

Several deep boreholes have been drilled in the last years through the rift valley fill in search for oil, penetrating almost continuously the Neogene formations which had been accumulated in great thicknesses in the Jordan—Dead Sea region. These boreholes have been drilled in the deeper parts of the basins thus the sections penetrated were rather continuous, consisting of finer grained sediments and much less oxidized when compared with outcrops. Consequently, palynomorphs in significant numbers could be extracted from the continental deposits. This resulted in obtaining a detailed palynostratigraphic framework (HOROWITZ and HOROWITZ, in press).

The marine Neogene formations are quite well developed in the western part of Israel and the off-shore (GVIRTZMAN, 1970) and their foraminiferal assemblages are well-known (DERIN and REISS, 1973; and many others).

The present study is an attempt to correlate the two sets of stratigraphic data, the palynozonation of the continental sequences and the planktonic foraminifera zonation of the marine sediments, so that the Neogene palaeogeography of Israel could be reconstructed.

Bravo No. 1 borehole, drilled off-shore Israel was chosen as a preliminary target. It proved to contain both palynomorphs and planktonic foraminifera, thus enabling the correlation. It should be noted, however, that this is only the beginning of the study and the presented results (Fig. 1) only refer to a single borehole in which the two sets of data have been compared.

The palynostratigraphy and planktonic foraminifera have been studied in more than 10 boreholes each so these seem to be quite well established (DERIN and REISS, 1973; HOROWITZ, 1984).

*Results.* The results of both pollen and foraminifera analyses of Bravo No. 1 borehole are shown in Fig. 1. Only the most diagnostic foraminifera are listed, together with the suggested environments of deposition at the drilling location, off the coast of Ashqelon in the southern coastal plain of Israel.

The pollen diagram is divided into three columns: the left one shows relations of various tree groups—the Mediterranean *Quercetalia*, conifers, which require a rather dry climate, *Picea orientalis* which although a conifer seems to prefer humid climates, and Betulaceae—Juglandaceae of subtropical affinity; calculations are made on the basis of total AP in the sample, taken as 100%.

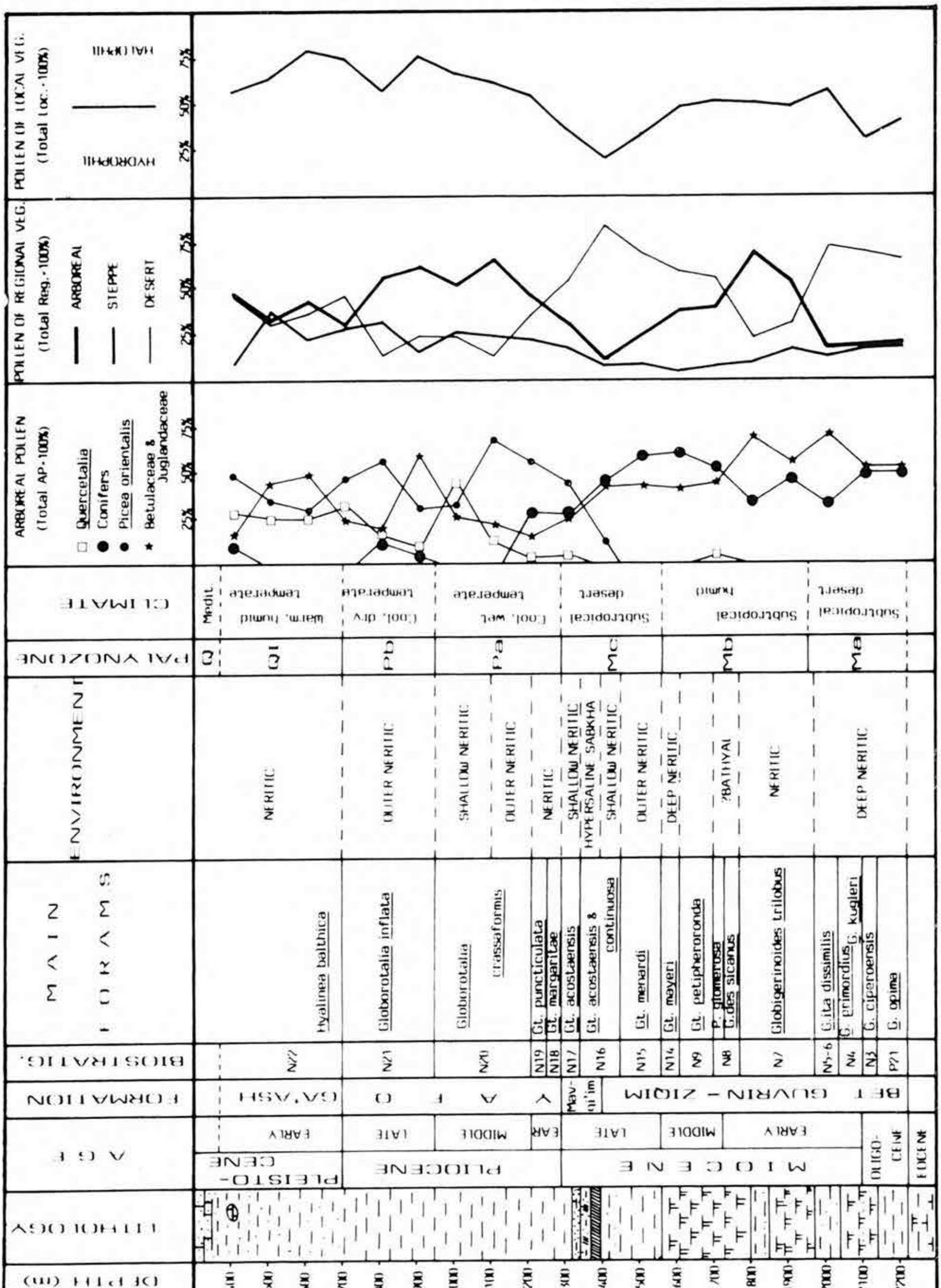


*Fig. 1.* Bravo No 1 borehole

The middle column displays the regional vegetation by its environment. The pollen spectra have been divided into two groups: those of a regional affinity, represent by plants growing in desert, steppe or arboreal domains; and those of a local affinity, such as hydrophil and halophil plants which characterize the salinity of water bodies close to the place of deposition. It should be noted that correlations between borehole sequences are only based on the composition of the regional pollen spectra, while the local elements differ from one locality to the other. Thus the regional elements are calculated on their total number taken as 100%, regardless of the number of local elements.

The column to the right shows the local, coastal environment in terms of salinity or freshwater affinity and is only based on the total number of pollen derived from the hydrophil and halophil plants, taken as 100%.

*Stratigraphy*. The lowermost Neogene palynozone is Ma, which comprises pollen spectra rich in non-arboreal components, of which the most abundant are those derived from Compositae, indicating desert conditions. The arboreal pollen comprise Betulaceae, Juglandaceae and conifers, always in subordinate quantities of the regional vegetation spectra. Steppe elements, mainly *Artemisia* and *Ephedra* have values quite similar to the arboreal pollen. The spectra indicate a warm, dry subtropical environment. Typical foraminifera of the Ma palynozone in Bravo No. 1 are, from bottom to top, *G. opima*, *G. ciperoensis*, *G. kugleri*, *G. primordius* and *G. ita dissimilis*, indicating that the sequence was deposited in P21 through N6 stages, in a deep neritic environment. N5–6 are extremely reduced.

The overlying, palynozone Mb is much richer in arboreal pollen than the preceeding one, basically of the same types. However Betulaceae and Juglandaceae are more common than conifers. The steppe elements maintain their previous values but a considerable drop is apparent in the pollen produced by desert plants. These pollen spectra indicate a warm, rather wet subtropical environment for Mb times. Foraminifera recovered for the Mb sequence in Bravo 1 are *Globigerinoides trilobus*, *G. sicanus*, *P. glomerosa* and *Gt. peripheroronda* indicating N7 through N9 for its lower part, and *Gt. mayeri* for the top of the palynozone indicating N14 zone. N10 through N13 zones seem to be altogether missing in this borehole, which is most probably a regional phenomenon. N7 was deposited in a neritic environment at Bravo No. 1, N8 probably in the bathyal and N14 in the deep neritic zone. However it is not entirely clear at which environment N9 was deposited and what is the nature of the overlying missing section.

Palynozone Mc is again characterized by very low arboreal pollen percentages, dropping to nil in many samples from the Jordan Valley. At the base conifers show an increase, while towards the top *Picea orientalis* makes its first appearance. Within the regional vegetation the desert elements prevail, while within the local flora the halophils show a considerable peak. The region was probably a warm, dry desert with very little trees and steppe plants, dominated by the desert vegetation. The picture at the coastal plain of Israel is somewhat less harsh than in the hinterland, where no tree pollen are recorded for much of the palynozone Mc. The main foraminifera in Bravo No. 1 include *Gt. menardii*, *Gt. continuosa*, and *Gt. acostaensis*, indicating N15 through N17 zones and the environment indicates an oscillating regression from outer neritic at N15 times to a hypersaline sabkha at N16, depositing evaporites which somewhat predate the Messinian crisis becoming marine again at the base of N17. The Messinian evaporites *per se* are almost totally missing in the Israeli coastal plain, being eroded or not deposited at all.

The overlying, palynozone Pa is again typified by a considerable increase of the arboreal pollen. This time the increase is mainly caused by the spreading of *Picea orientalis*, together with an increase of the *Quercetalia*. Betulaceae and Juglandaceae slowly decrease, together with the conifers typical for the underlying Miocene deposits. Steppe elements show some increase and the desert plants pollen are much less abundant. Within the local vegetation the hydrophil plants again take the upper hand. These all seem to indicate that during Pa times the environment was of cool, wet temperate character. The foraminifera typical for this interval at Bravo 1 are *Gt. margaritae*, *Gt. puncticulata* and *Gt. crassaformis*, in Early—Middle Pliocene N18 through N20 zones. The environment of deposition is neritic at the base, outer neritic at the middle and shallowing towards the top of the palynozone.

During palynozone Pb the arboreal pollen show some decrease, while an increase is apparent at the pollen produced by the steppe vegetation. Within the arboreal pollen a peak is seen in the conifers. The environment seems to have been not much different than during the preceeding Pa times, only somewhat drier, of a cool, dry temperate climate. The foraminifera are dominated by the late Pliocene *Gt. inflata*, indicating N21 zone, deposited in the outer neritic zone.

The Neogene sediments in Bravo No. 1 borehole, as in other boreholes in Israel, are overlain by a suite of Quaternary formations typified palynologically by the abundance of pollen derived from trees of the Mediterranean *Quercetalia* with an abundance of *Picea orientalis* at the base and the occurrence of *Hyalinea balthica* at the bottom of N22 zone.

*Conclusion.* Palynozone Ma is characterized by pollen spectra indicating a dry, subtropical environment, in Late Oligocene—Earliest Miocene age.

Palynozone Mb, typified by pollen spectra indicating a wet subtropical environment is in early through Middle Miocene age.

Palynozone Mc, in which the desert plants pollen prevail, indicating a rather extreme desert conditions, is in a Late Miocene age.

Palynozone Pa, dominated by arboreal pollen flora signifying a cool, wet temperate climate is in an Early—Middle Pliocene age.

Palynozone Pb, of essentially similar character to the underlying Pa, only somewhat drier, is in a Late Pliocene age.

The Neogene sequence is overlain in most parts of Israel by a suite of Quaternary sediments.

The Miocene sequence at Bravo No. 1 is not continuous, a hiatus is apparent where zones N10 through N13 are missing, probably due to a regional regression.

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# THE PROBLEM OF RED-BEDS IN NORTHERN GREECE

by

A. PSILOVIKOS, G. KOUFOS and G. SYRIDES

*Introduction.* The area of Northern Greece is a complex of grabens, basins (plains) and horsts (mountains) arranged in a primary NW—SE and secondary NE—SW trend. Intermountaneous high level grabens (Ptolemais, Volax) or low level grabens (Mygdonia, Drama), large coastal basins and plains (Axios—Thermaikos, Anthemous, Chalkidiki, Serres—Strymon) were filled with Neogene and Quaternary sedimentary deposits (Fig. 1). The most characteristic of them are the red-beds. They are continental, terrestrial—fluvial clastic sediments of red-brown colour, that appear along graben sides on the surface, or can be traced in the central parts of the grabens, in the subsurface by drilling. The widespread of red-beds and the associated rich mammalian fauna offer a good basis for stratigraphic and palaeogeographic correlation with other sedimentary deposits in the wider Mediterranean realm.

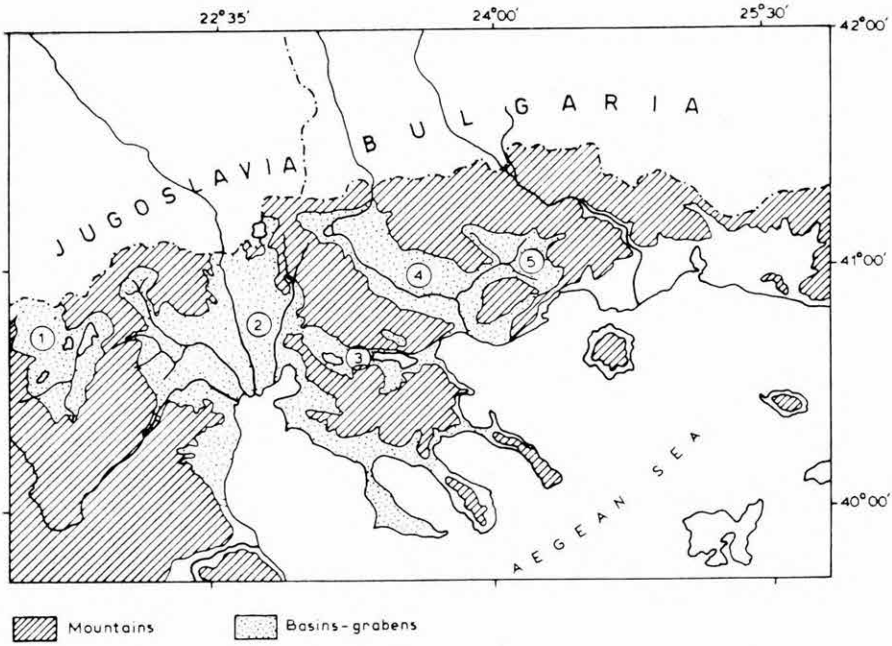


*Fig. 1.* Sketch map indicating the major basins and grabens, containing red-beds in northern Greece

*1* Kozani—Ptolemais grabens, *2* Axios—Thermaikos basin, *3* Mygdonia—N Chalkidiki graben, *4* Serres—Strymon basin, *5* Drama basin

**Lithology—texture**

The red-beds are characterized by great lithological variety, particularly the coarser members such as conglomerates and fanglomerates. Their lithology is related to the source rocks, found along the mountain sides that border the grabens.

A similar variety in texture characterize the red-bed. Textural analysis was applied to several samples that contain coarse and fine sediments (Table 1, and Fig. 2). The size parameters (M, δ, sk, ku) indicate mixed sand/silt grains very badly sorted, positively skewed and leptokyrtic. The shape parameters (Sphericity, Roundness) of the quartz grains show low-middle sphericity (0.6 to 0.7) and very low roundness (0.15—0.25).

The size-shape parameters indicate textural immaturity which is further related to the mineralogical immaturity of the gravel and sand.

**Size and shape parameters of representative red bed-samples**

*Table 1*

| *Neogene* | Size | | | | Shape | |
|---|---|---|---|---|---|---|
| | M | q | sk | ku | sph | rd |
| Anthemus graben | 5.25 | 2.58 | 0.10 | 0.96 | 0.66 | 0.22 |
| Petralona (Chalkidiki) | 3.85 | 2.30 | 0.13 | 1.91 | 0.62 | 0.18 |
| N. Fokea (Kassandra Peninsula) | 4.77 | 2.45 | 0.07 | 1.39 | 0.63 | 0.22 |
| Ag. Nikolaos (Sithonia Peninsula) | 4.57 | 2.25 | 0.46 | 1.44 | 0.60 | 0.15 |
| *Pliocene—Pleistocene* | | | | | | |
| Mygdonia basin: | | | | | | |
| Gerakarou | 4.80 | 2.02 | 0.23 | 0.88 | 0.69 | 0.25 |
| Krimni | 3.64 | 1.52 | 0.23 | 0.99 | 0.64 | 0.21 |
| Chalkidiki: | | | | | | |
| Gerakini | 5.30 | 2.30 | 0.12 | 1.51 | 0.61 | 0.17 |
| Moudania | 3.19 | 2.46 | 0.13 | 1.18 | 0.59 | 0.19 |

**Structure**

The structure of red beds varies from place to place, even in the same graben. But the basic structural unit of the red-beds is a rythm, having three individual beds of varied dimensions:

— a lower bed of coarse clastics mainly gravel (pebble with sand matrix);
— a middle bed of sand, associated with gravel and silt;
— an upper bed of silt-clay, associated with sand.

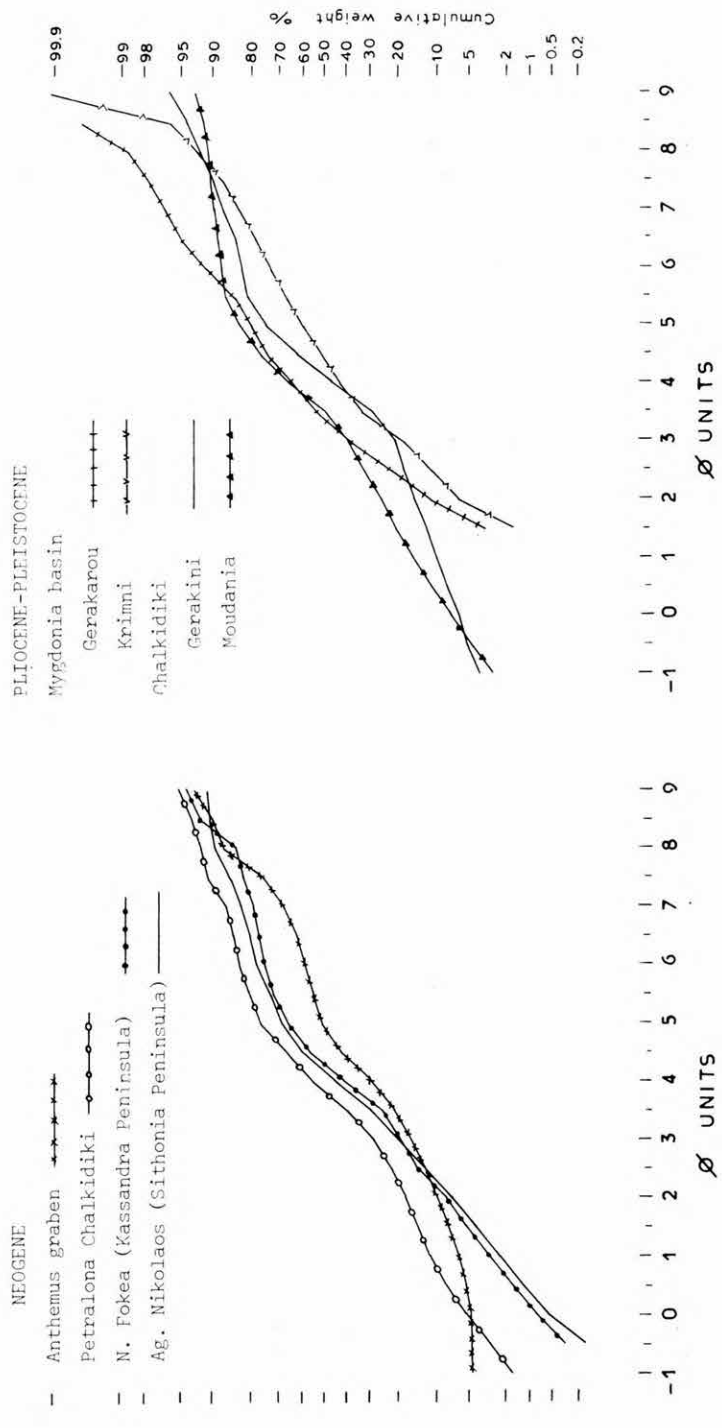


*Fig. 2.* Size curves of representative red-bed samples

The above described rythm has a varied thickness and bed arrangement. Three main types of bedding were distinguished within the rythm (Fig. 3).

- *Type A. Cross-bedding.* The three beds of the rythm are well developed, have clear border surfaces and include thin layers of material, arranged in a "fining upwards" way.
- *Type B. Graded bedding.* The material is arranged in a fining upward way within each rythm. There is a transitional zone from the coarse to fine grains, with no clear distinction of the beds.
- *Type C. Lenticular bedding.* In a mixed, unsorted coarse and fine material, having a rather massive character, several lenses of gravel and coarse sand are developed.

The structures of the red-beds indicate a rythm in transport and deposition of material, probably due to a similar rythmic character of the climate during the time of their formation.

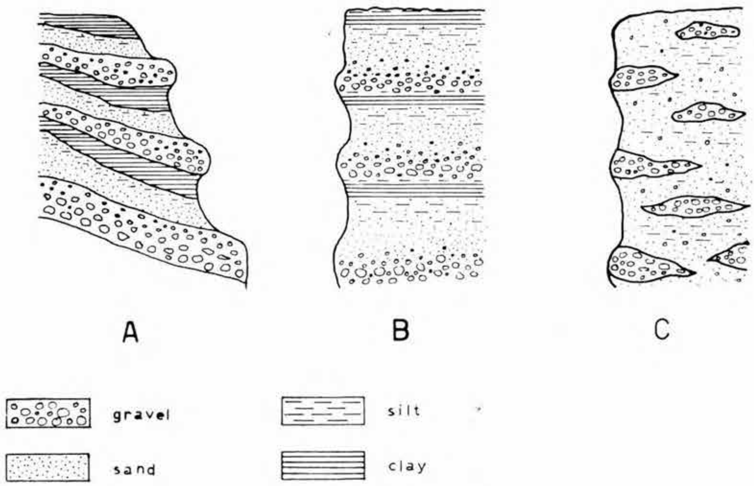


*Fig. 3.* Main types of red-bed structure

**Stratigraphy**

A rich mammalian fauna was recovered from the red-beds of Northern Greece. The fauna identified so far allowed the following stratigraphic grouping (Table 2).

*Upper Miocene.* It begins with the Nea Mesimvria Formation (Axios basin) that ncludes fossils of zone MN10.

It continues up with the Vathylakkos Formation (Axios basin) and Thermi-Trilophos one (Anthemous graben—western Chalkidiki). The first contains Lower Turolian fauna while the second contains fossils of Turolian age.

It ends up with the Ano Metochi/Palios Mylos Formation (Serres basin), with fossils indicating an Upper Turolian age, that can be correlated with zone MN13 of Mediterranean Neogene.

The same age was also suggested for the red-beds of the base formation (Ptolemais basin) Agios Nicolaos (Chalkidiki, Sithonia) and the Nea Fokea Formation (Chalkidiki—Kassandra), where no fossils were found. The overlying Pliocene lacustrine—brackish sedimentary deposits justify the above made suggestion.

*Upper Pliocene/Lower Pleistocene.* The greatest fossil assemblages in red-beds were recovered from the Mygdonia basin—northern Chalkidiki area. It includes two major localities close to the villages of Gerakarou (Langada graben) and of Krimni (Marathousa graben). The recovered fauna reveals an Upper Villafranchian (Villányian) age. Of the same age is also the fauna recovered by SICKENBERG (1968) in the small graben of Volax North of the Drama basin.

*Upper Pleistocene.* There is a good distribution of Upper Pleistocene red-beds that contain mammalian fauna. In Ptolemais basin the Perdikas Formation contains fossils of Late Pleistocene age (PAVLIDES, 1985). The top red-beds of western Chalkidiki (Moudania—Gerakini) can also be correlated with the Perdikas Formation, although this need further study and consideration.

In Drama basin the Aggitis Formation is an alluvial fan that contains a Late Pleistocene fauna (KOUFOS, 1981). The above indicate a period of red-beds formation during the glacial/interglacial epochs of the Late Pleistocene. Glaciation in Greece was limited only on the highest mountaineous zones during this period (VAVLIAKIS, 1981; PSILOVIKOS, 1981).

## Palaeoclimate

For the continental red-beds of N Greece the palaeoclimatic conditions during the time of their formation were the following.

— For the red-beds of the Upper Miocene the climatic conditions were characterized as warm and progressively dry, from Vallesian (savannah with small trees and shrubs) to Turolian (savannah). It should be considered as a subtropical zone that became drier during the Messinian (KOUFOS, 1980).

— For the red-beds of the Upper Pliocene/Lower Pleistocene the climate was warm semiarid, characterized by wet and dry seasons.

— For the red-beds of the Upper Pleistocene there is a controversial opinion about the prevailed climatic conditions. It seems that both the glacial (KOUFOS, 1981, VAVLIAKIS, 1981) and the interglacial epochs (MARINOS, 1964) favoured the formation of red-beds in this period. The role of the climate needs further consideration for the Upper Pleistocene.

**Stratigraphic and climatic characteristics of the red-beds in northern Greece**

*Table 2*

| M.Y. | Epochs | Mammal zones | Continental stages | Climate | Localities | Mammalian fauna |
|---|---|---|---|---|---|---|
| 1 | Pleistocene | | | | Aggitis (Drama) | *Ursus spelaeus*, *Mammontheus* cf. *primigenius*, *Equus caballus*, *Coelodonta antiquitatis*, *Megaloceros giganteus*, *Cervus* sp. |
| | | MNQ20 | BIHARIAN | glacial interglacial | Perdikas F. (Ptolemais) Gerakini—Moudania (Chalkidiki)? | *Coelodonta antiquitatis*, *Bos* cf. *primigenius*, *Cervus* sp. |
| 2 | Pliocene | MNQ19 | VILLANYIAN | warm semiarid | Volax (Drama) Gerakarou-Krimni (Mygdonia B.) | *Megantereon*, *Nyctereutes*, *Vulpes*, *Euctenoceros*, *Eucladoceros*, *Macedonitherium*, *Leptobos*, *Gazella*, *Nemorhedus*, *Gazellospira*, *Didermoceros*, *Equus*, *Alloippus* *Canis etruscus*, *Equus stenonis senezensis*, *Sus* sp., *Croizetoceros ramosus* cf. *minor*, *Cervus* cf. *philisi*, *Gazella* sp., *Dicerorhinus etruscus*, *Equus stenonis* |
| 3 | | MN17 MN16 | VILLAFRANCHIAN | | | |
| 4 | | MN15 MN14 | RUSCINIAN | warm humid | | |
| 5 | Miocene | MN13 | | warm arid | Ano Metochi (Serres) Base Formation (Ptolemais) Nea Fokea (Chalkidiki)? | *Helladotherium* cf. *duvernoyi*, *Hipparion* cf. *mediterraneum*, *Gazella* sp., *Prostrepsiceros woodwardi*, *Gazella* cf. *gaudryi* |
| 6 | | | | | Agios Nikolaos (Chalkidiki)? | *Hipparion* sp. |
| 7 | | MN12 | TUROLIAN | savannah warm | Termi-Trilophos (Axios—Chalkidiki) | *Hipparion mediterraneum*, *Gazella deperdita* |

| | | | | | | |
|---|---|---|---|---|---|---|
| 8 | Miocene | MN11 | TUROLIAN | semiarid | Vathylakos (Axios) | *Adcrocuta eximia, Ictitherium robustum, I. hipparionum, Plioviverrops orbignyi, Plesiogulo crassa, Hipparion dietrichi, H. mediterraneum, Dorcatherium puyhauberti, Samotherium boissieri, Miotragocerus* sp., *Prostrepsiceros (Helicotragus) zitteli* |
| 9 | | | | | | |
| 10 | | MN10 | VALLESIAN | savannah with grasslands | Nea Mesimvria (Axios) | *Progonomys cathalai, Adcrocuta eximia leptoryncha, Hipparion primigenium, Mesembriacerus melentisi, Decenatherium pachechoi, Ouranopithecus macedoniensis* |
| 11 | | MN9 | | warm semiarid | | |
| 12 | | | | | | |

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# BIOSTRATIGRAPHY AND PALAEOECOLOGICAL INTERPRETATION OF MICROMAMMAL FAUNAL SUCCESSIONS IN THE UPPER ARAGONIAN AND VALLESIAN (MIDDLE—UPPER MIOCENE) OF THE DUERO BASIN (N SPAIN)

by

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and R. Daams

*Introduction.* The Duero basin is located in the northwestern part of central Spain (Fig. 1). The fluvio—lacustrine sediments of the continental Miocene are well-exposed in the central part of the basin, where their thickness exceeds 200 m. The exceptional conditions of the exposures, and the absence of tectonical deformation, favour the establishment of biostratigraphic sequences. Two sections (Torremormojón and Ampudia, see Fig. 1) have been studied in detail, yielding 11 micromammal-bearing localities covering the Middle Miocene and part of the Upper Miocene.

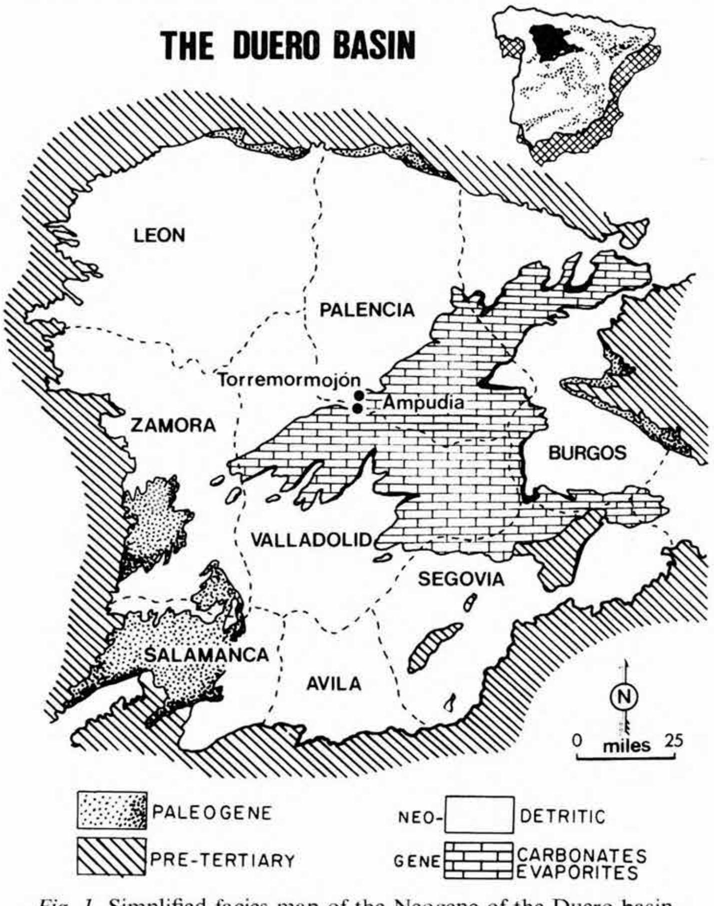


*Fig. 1.* Simplified facies map of the Neogene of the Duero basin

The aim of this study is the establishment of a local biozonation, which may help to test the European continental chronology, not defined over stratigraphic sequences. Hitherto, only a few sections have been described covering more than one biozone (VAN DE WEERD, 1976; DAAMS and FREUDENTHAL, 1981; MEIN et al., 1983). In the Duero basin the faunas may be arranged in four formal biozones covering the Upper Aragonian and part of the Vallesian. This is the only basin known in Spain in which a continuous documentation of this interval is present. A further purpose of this study is the palaeoecological interpretation on the basis of the small mammals.

Previous studies on the biostratigraphy of this basin are by LÓPEZ and SANCHIZ (1982), ALVAREZ SIERRA (1983), GARCÍA MORENO (1983), ALVAREZ et al., (1985) and LÓPEZ MARTÍNEZ et al. (in press).

## Lithostratigraphy

Five lithologic units have been recognized. These are:

*1* The Villalba de Adaja unit, defined by CORRALES et al. (1978). Consisting mainly of reddish arkosic silts it is present in the southeastern part of the basin. The thickness measures some 40 m and it is of Middle Aragonian age. It is furthermore characterized by grey-greenish arkosic silts and intercalations of small, sand-filled channels. The sands and silts form fining upward sequences that may be overlain by limestones.

*2* The Dueñas unit, defined by OLMO et al. (1982), consists mainly of marls and limestones. Exposed in the southern part of the basin, its thickness may measure up to 40 m and it is of middle to late Aragonian age. It is characterized by white and light-grey marls and calcareous clays. At the basis some slightly bioclastic limestone intercalations are present. This unit wedges out towards the south and the west. The sedimentation mechanism is attributed to more or less saline playa-lake deposits.

*3* The Tierra de Campos unit, defined by HERNANDEZ PACHECO (1915), mainly consists of clays and channeled sands. It is present in the northern and western sector of the basin. Its thickness measures between 30—45 m, and it is of late Aragonian age. It consists of yellow silt- and claystones with a carbonate content of up to 15%, and at some places diagenetic gypsum at the top. The sands may be channeled or they may form regular bedding planes. In the clays and silts, palaeosols may be present. The unit is thought to have been formed in a distal fluvial plane with meandering rivers and flood plains.

*4* The Cuestas unit, defined by HERNANDEZ PACHECO (1915), mainly consists of marls, limestones and gypsum. Exposed in the central sector, its thickness measures between 50—70 m, and it is of Late Aragonian—Early Vallesian age. Towards the southwest the gypsum wedges out. The unit is thought to have beend eposited in a shallow lacustrine environment with variable salinity.

*5* The Paramo limestone, defined by HERNANDEZ PACHECO (1915), is exposed at various places in the central part, and its thickness is 20 m or more. It would be of Turolian age.

Our mammal-bearing localities are found in the Tierra de Campos (3) and Cuestas (4) units.

## Palaeontology and biostratigraphy

The faunas are subdivided in 4 assemblage zones (Fig. 2), these are:

*a* The *Megacricetodon lopezae* zone, characterized by *Megacricetodon lopezae*, *M. minor*, *Microdyromys* cf. *koenigswaldi*, *Peridyromys rex* and *Prolagus major*.

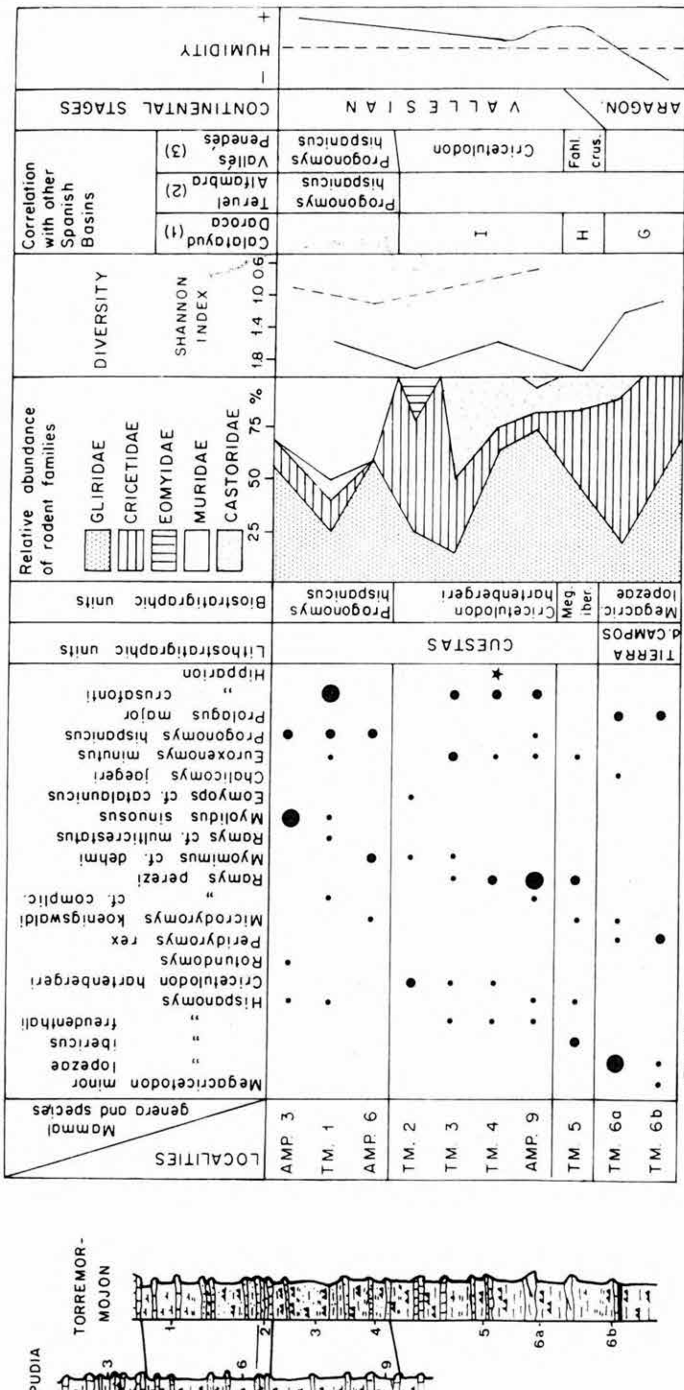


*Fig.* 2. Range chart of mammal genera and species from the Ampudia and Torremormojón sections in the Middle—Upper Miocene of the Duero basin

In the correlation column the (1) refers to the biozonation of DAAMS and FREUDENTHAL, 1981, (2) to the biozonation of VAN DE WEERD, 1976 and (3) is that of AGUSTÍ, 1981 *b* The *Megacricetodon ibericus* zone, characterized by *Megacricetodon ibericus*, *Hispanomys*, *Ramys perezi* and *Euroxenomys minutus*. *Microdyromys* cf. *koenigswaldi* is also present in this zone.

*c* The *Cricetulodon hartenbergeri* zone, characterized by *C. hartenbergeri*, *Megacricetodon freudenthali*, *Microdyromys* cf. *complicatus*, *Myomimus* cf. *dehmi*, *Eomyops catalaunicus* and *Prolagus crusafonti*. *Ramys perezi* and *Euroxenomys minutus* are also present. In AMP 9 *Progonomys hispanicus* is present in low numbers. The only remains of *Hipparion*, found in the two sections, are from TM 4.

*d* The *Progonomys hispanicus* zone, characterized by the abundance of this species, and by the presence of *Myolidus sinuosus*, *Ramys* cf. *multicrestatus*, *Euroxenomys minutus*, *Prolagus crusafonti*, *Myomimus* cf. *dehmi*, *Microdyromys* cf. *complicatus*. *Rotundomys montisrotundi* is present in low numbers.

This biostratigraphic sequence is correlated to the biozonations of other Spanish basins (Fig. 2). The differences in the faunal compositions and the presence of various endemic taxa in the Duero and in the other basins make it difficult to give a precise correlation. In spite of these differences, a rough correlation is possible.

The *M. lopezae* zone has the following taxa in common with zone G from Daroca: *Megacricetodon crusafonti*, *M. minor*, *Prolagus major* and *Microdyromys*.

The *M. ibericus* zone shares the following taxa with zone H from Daroca: *M. ibericus*, *Hispanomys* and *Hipparion*. With the *Fahlbuschia crusafonti* zone from the Vallès Penedès it has *M. ibericus* and *Hispanomys* in common.

The *Cricetulodon hartenbergeri* zone shares with zone I from Daroca: *Cricetulodon* en *Progonomys hispanicus*. With the *Cricetulodon* zone from the Vallès Penedès it shares *Cricetulodon*, *P. hispanicus* and *Eomyops catalaunicus*.

The *Progonomys hispanicus* zone shares with the zone of the same name from Teruel *P. hispanicus* and *Myomimus* cf. *dehmi*, and with the *P. hispanicus* zone from the Vallès Penedès *P. hispanicus* and *Rotundomys*.

In our sections the only locality with *Hipparion* is that of TM 4 (*C. hartenbergeri* zone). LÓPEZ et al. (in press) mention that the first appearance known of *Hipparion* in the three best documented Spanish basins on the basis of micromammals during this time (Duero, Daroca, Vallès Penedès) is diachronic. The immigration of *Hipparion* does not coincide with any major change in the micromammal fauna (VAN DE WEERD and DAAMS, 1978; AGUSTÍ, 1981; ALVAREZ SIERRA et al., 1985). In the type area of the Vallesian, the limit between the Aragonian and the Vallesian is situated within the *Fahlbuschia crusafonti* zone (AGUSTÍ, 1981). In our opinion it should be avoided to use large mammals to define stage boundaries, whereas the stages are furthermore subdivided on the basis of small mammals. Because micromammal faunas are much more frequently found than large-mammal faunas, we propose to use micromammal events to define the stage limits.

## Palaeoecological interpretation

The Gliridae are relatively abundant in the entire composite section. Through time the representatives of this family display more complicated dental patterns (*Ramys perezi*, *R.* cf. *multicrestatus* and *Myolidus sinuous*). The Cricetidae are abundant in the lower zones and less common in the upper zone; its decline coincides with the increase of Muridae. VAN DE WEERD and DAAMS (1978) observed a similar substitution in the Calatayud-Teruel basin, and interpreted it as the result of a possible ecological competition between the two families. In the same time a strong increase of Muridae is observed in the Vallès Penedès, but the Critidae remain abundant. The Castoridae are common in the second and third biozone, and rare in the last one.

The quantitative composition of the faunas at family level from the Duero basin is significantly different from that of the other two basins. While the Gliridae remain abundant in our composite section, a strong decrease of this family is observed in Teruel and the Vallès Penedès. In the Upper Vallesian of the Duero and Teruel the Cricetidae are poorly represented, whereas in the Vallès Penedès they remain abundant. In the Duero basin the Castoridae are far more abundant than in the other two basins. The Eomyidae appear to be relatively common in the Vallesian from the Vallès Penedès, whereas in the Duero basin only one locality has Eomyidae, and in Aragon this family is absent. The differences in the qualitative and quantitative compositions between the Duero basin on the one hand, and the other two basins on the other, show that during the late Aragonian and Vallesian the Duero basin was a separate bioprovince. In our composite section small mammals that favour moist and forested biotopes are abundant: Gliridae with complicated dental pattern (VAN DER MEULEN and DE BRUIJN, 1982; DAAMS and VAN DER MEULEN, 1984); Eomyidae, which are supposed to be representatives of humid biotopes (VAN DE WEERD and DAAMS, 1978); and Castoridae are also indicators of humid biotopes. The decrease of this latter family in the uppermost zone may be explained by sedimentary causes. In the upper portion of our sequences the fluvial influence decreases and it is substituted by lacustrine environments, which is not the preferred habitat of beavers.

We also calculated the Diversity Index of Shannon. Fig. 2 shows that there are no sudden changes in the diversity values. In conclusion we may say that on the basis of small mammals the Duero Basin had a relatively stable, humid climate during the middle and late Miocene. During the Vallesian there is a trend towards more humidity.

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# NEOGENE ORE MINERALIZATIONS OF HUNGARY

by

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The majority of the hilly range along the northeastern borders of Hungary is built up by volcanic rocks of Miocene age. This volcanic belt is a member of the so called Inner Carpathian Volcanic Chain. Volcanic activity had been centered here around several eruption centers more or less reflected also by present-day morphology. The volcanic landscape, beginning with the Dunazug Mts on the west, includes the Börzsöny, the Cserhát, the Mátra and finally the Tokaj Mts, the latter occupying the easternmost corner of the country. The chain does not terminate in Hungary: it continues on Czechoslovakian and Romanian territories with the Zemplín, Vihorlat, and Gutin Ranges respectively. The age of the volcanism is mainly Badenian, although older and younger volcanics can also be recognized within the chain. Radiometric data suggest that the duration of the volcanic activity might have been about 15 million years and it has begun 25 million years ago. The volcanics are younger at the eastern end of the chain that it is on the west: volcanic rocks of the Tokaj Mts are mostly of Late Sarmatian age.

Formation of the volcanic chain is the direct result of plate tectonic movements responsible for the present shape of the Carpathian region. Due to the mighty Neogene basinfilling of the Great Plain and the complex geological setting of the Northern Mountain Range, the exact nature of these plate tectonic movements is rather difficult to deciper. Although the question is yet far from settled, latest research suggests that the present shape of the geotectonic units of north Hungary began to emerge some time during the Miocene; and that the Neogene structural boundaries became superimposed onto the older structures. It is the masking effect of these younger structures which makes it almost impossible to reconstruct the details of the pre-Miocene structural setting. According to this interpretation the most characteristic lineaments such as the Zagyva, the Darnó and the Hernád lines are most probably not much older than Early Neogene. As to the Darnó line, its Neogene age has been proved lately on the basis of geological considerations.

### Volcanism and its beginnings are obviously closely related to the formation of these young Neogene structures

The palaeovolcanic reconstruction of the Börzsöny, Mátra and Tokaj Mts is not yet complete and—what is more— opinions widely differ concerning the genetic interpretation of field and laboratory evidence. All these mountains are built up by stratovolcanic series mostly of andesitic composition. It is only the Tokaj Mts where chemical assays show a more acidic assemblage of volcanics. Exploratory drilling and geochemical survey have disclosed several new details of the geology of the Mátra and Börzsöny Mts during the past decade, whereas much less is known of the deeper horizons of the Tokaj Mts. In the Börzsöny and Mátra Mts the afore mentioned exploration activity disclosed the subvolcanic source of the known ore mineralizations while in the Tokaj Mts the presence of hypothetic intrusive bodies can be inferred on the basis of analogies and general geological considerations only. The fact that exploration has proved the presence of several previously surmised intrusive facies is of utmost importance from the point of view of further prospecting. We have good reason to suppose that a vertically zoned mineralization similar to these known from several occurrences of the Carpathe—Balkanian region, is present also in the Börzsöny, Mátra and Tokaj Mts. Near-surface polymetallic and gold—silver mineralizations of the stockwerk or the breccia-pipe type usually indicate deep-seated subvolcanic bodies the discovery of which is the very task of modern exploration. Succesful prespecting for the mineralized facies of those subvolcanic rocks, which were discovered by drilling activity in the Mátra and Börzsöny Mts, may prove the pertinence of hypotheses concerning an analogy of the north Hungarian region with the great Neogene ore deposits of Slovakia and Transylvania. In the Tokaj Mts the supposed vertical continuation of the known near-surface gold—silver mineralization is still to be verified. This requires largescale drilling and the understanding of several still obscure details of the genesis of the deposit.

**Genetic types of the ore mineralizations**

*Table 1*

| | |
|---|---|
| 1st BÖRZSÖNY RANGE | — porphyry copper and molybdenum shows<br>— polymetallic Fe—Pb—Zn (Bi—Te—As—Au—Ag) hydrothermal veins and mineralized breccias |
| 2nd MÁTRA RANGE | — polymetallic Pb—Zn—Cu (Au—Ag) hydrothermal veins centered around Gyöngyösoroszi and Parádsasvár<br>— Hg + Ba + Sb shows at Asztagkő east of Gyöngyösoroszi |
| 3rd TOKAJ RANGE | — hydrothermal polymetallic and piritic veins (Fe—Pb—Zn) with Au + Ag at Telkibánya<br>— epithermal Hg shows at Sárospatak |

Mining activity in the *Börzsöny Mts* had been centered around the village of Nagybörzsöny in the vicinity of which the two historic ore occurrences of Fagyosasszony-bánya and Rózsa-bánya are situated. The only large-scale prospecting campaign, covering the whole area of the Börzsöny Mts was the one launched in the early seventies. As a result of this campaign, both the volcanotectonic pattern and the outlines of the deep structural make-up of the area were disclosed. Also several new details of the petrology and geochemistry of the subvolcanic andesite suite became known and a new porphyry copper—molybdenum indication was discovered. As to the paragenesis and the morphological characteristics of the ore bodies it is the near-surface mineralization which is fairly well-known. There are two main types to be distinguished. The older one is a mesothermal impregnation-like mineralization bound to brecciated zones and breccia-pipes and is called the Rózsa-bánya type. (The name comes from the old mining site where this type of ore was mined from.) The

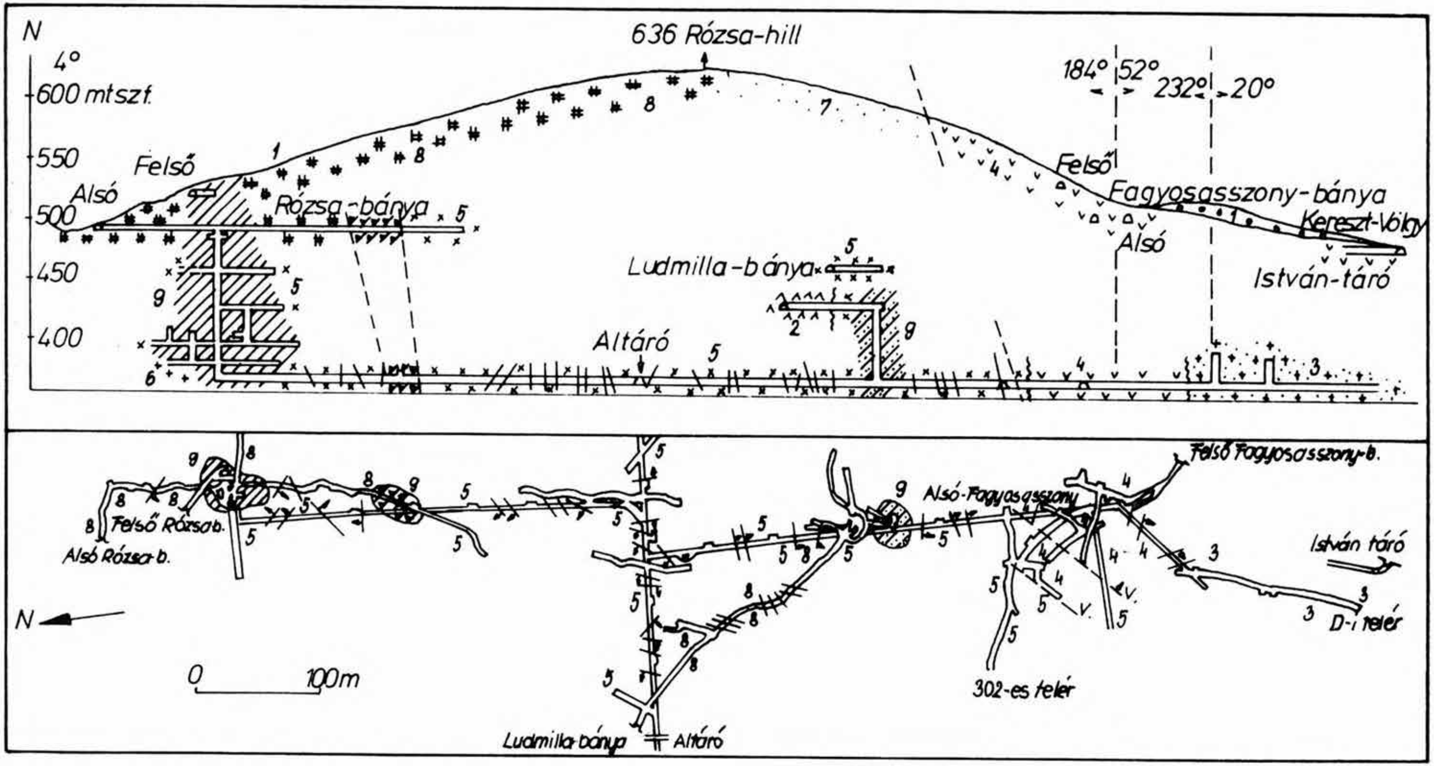


*Fig. 1.* Vertical and horizontal sections of the Nagybörzsöny ore district (after L. Mikó)

*1* Surface, *2* Neogene amphybole-augite andesite, *3* Neogene biotite-amphybole andesite, *4* Neogene amphybole-augite andesite, *5* Neogene amphybole-hyperstene andesite, *6* altered andesite, *7* agglomerate, *8* dacite, *9* mineralized bodies ore consists mainly of pyrrhotite accompanied by melnikovite, pyrite, marcasite, arsenopyrite, sphalerite and chalcopyrite. The presence of bismuth is characteristic of Rózsa-bánya with abundant *Bi*-minerals like bismuthin, cosalite, glaucodote, jamesonite, sartorite, proustite, stefanite, sternbergite etc, quite frequently with gold and silver as accessories. Non-metallic gangue minerals are calcite, quartz and siderite.

On the other side of the Rózsa-hill right opposite to Rózsa-bánya, the other historic mining site, Fagyosasszony-bánya is situated. Mineralization is of the vein type here and relatively younger than the impregnation type ore bodies of Rózsa-bánya. Formation temperatures are epithermal: beside pyrrhotite, which is obviously the earliest phase, galena and sphalerite are predominant usually replacing pyrrhotite. Less often arsenopyrite, chalcopyrite and pyrite also occur, alongwith a number of accessory ore minerals. The ore has a low but characteristic gold and silver content. Gangue minerals are, calcite and quartz.

Although the history of mining activity dates back as far as medieval times at the Nagybörzsöny area and there were several periods of prosperity, even the results of the last drilling campaign could not justify the reopening of the abandoned mines. Thus the area is inactive at the moment.

In the *Mátra Mts* Neogene ore mineralizations occur in the form of hydrothermal sulphide veins centered around Gyöngyösoroszi and Parádsasvár in the western and central Mátra respectively. The age of the mineralized pyroxene andesites and pyroclastics is Badenian. The most common wallrocks are mentioned as "variegated" andesites by Hungarian literature. The name comes from the manyfold textural variations observed in connection with this kind of rocks. The fresh-looking almost black pyroxene andesite, formed during the second stage of the eruptive activity is called the "cover"-andesite. The solidification of this youngest andesite is thought to have taken place after the main stage of ore mineralization. In accordance with this, up to now no traces of any kind of ore mineralization is known from the "cover"-andesite. Its radiometric age being Badenian it serves as an excellent marker regarding the age of the mineralization bound to the underlying—likewise Badenian—"variegated" andesite. Premineralization tectonism having affected the host rock prior to deposition of the ore minerals controlled the pattern of the sulphide veins. According to strike there are three main groups of veins to be distinguished. Related to the central zone they may be either radial or oblique. The radial set of joints and fissures seems to be much more of the "open" type and their filling is richer and more diverse mineralogically than that of the others. The reason for this lies most probably in the peculiarities of volcanotectonic movements preceeding the ascendence of the mineralizing fluids. Twenty veins are known at the Gyöngyösoroszi occurrence, about two thirds of which is of "central" position. The most famous of them is the "Károly" vein, the known length of which is nine-hundred meters along strike with a thickness of one to five meters. Its known vertical extension reaches fourhundred meters. As to the paragenesis of the epi to mesothermal vein filling, there are seven separate stages of mineral deposition recognized up to now. Main ore minerals are sphalerite, galena, pyrite and chalcopyrite with a number of accessories. The percentages of chalcopyrite show a definite downward increase. The higher parts of the vein are relatively enriched in gold an silver. Gangue minerals are quartz, amethyst, calcite and chalcedony. Although at a reduced rate production is still underway at Gyöngyösoroszi.

As to the paragenesis of the ore minerals, the epithermal mineralization of the *central Mátra* is basically similar to that of Gyöngyösoroszi with the exception of sphalerite which is much more prevalent here than at Gyöngyösoroszi. Differences regarding the gangue minerals are, however, much more substantial. Instead of quartz

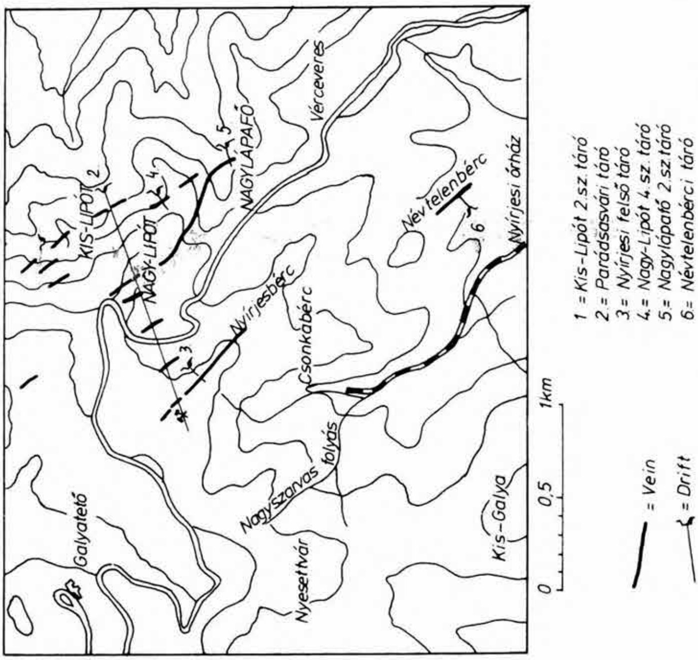


*Fig. 3.* Locality map of the Middle Mátra ore district (after A. VIDACS)
*1—6* Different drifts

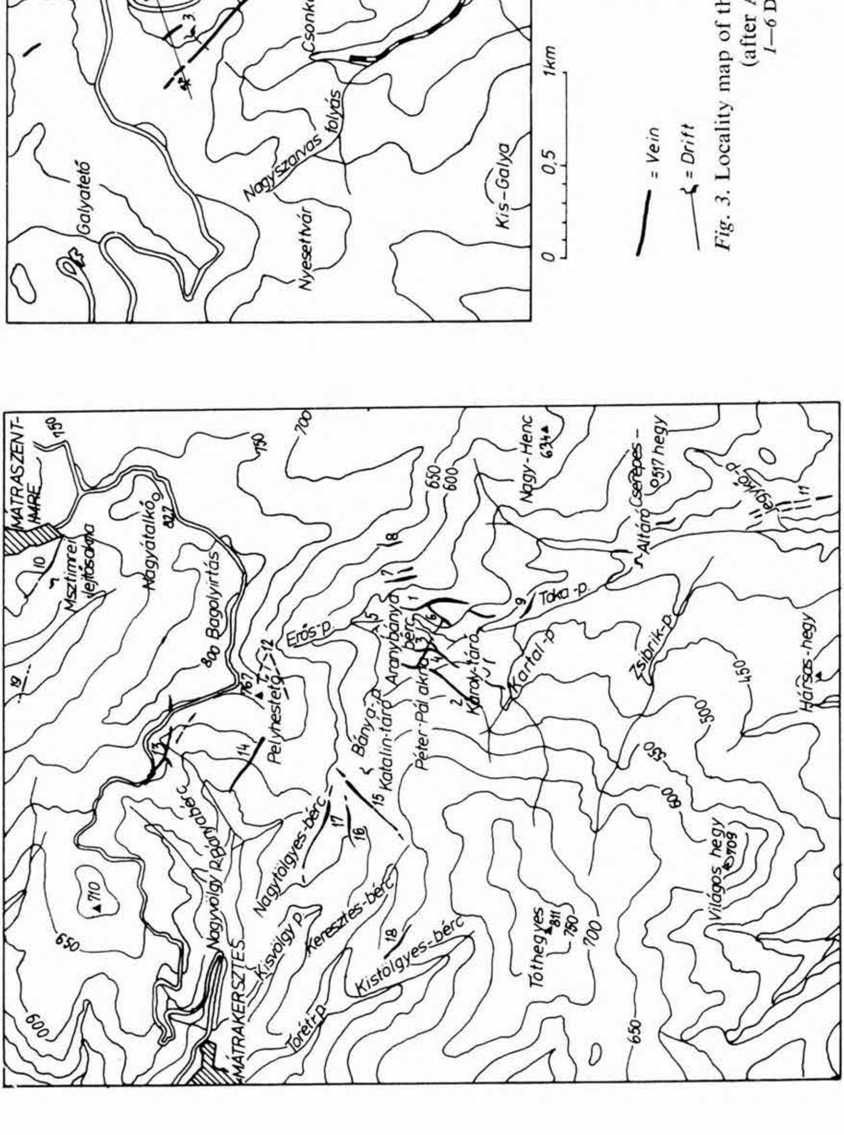


*Fig. 2.* Locality map of the Gyöngyösoroszi ore district in the W Mátra Mts (after A. VIDACS)
*1—19* Mineralized veins

and amethyst, calcite is the sole gangue mineral in the veins of the central Mátra. The sphalerites are invariably high in cadmium which sometimes occurs even in the form of greenockite. Latest research suggests that a direct connection has to be supposed between the two mineralized areas—namely between Parádsasvár and Gyöngyösoroszi.

The *Telkibánya* ore deposit is hosted by the intensely altered Sarmatian volcanic complex of the Tokaj Mts. The strike of the mineralized veins is north—south; the wallrocks are potash—rich trachytes formed by potassium—metasomatism. Silicification of the wallrocks is also a characteristic phenomenon at Telkibánya. Most veins are found in brecciated, cataclastic zones. In all fourteen veins are known from the area. The paragenesis is rather monotonous. Temperatures of formation are meso to

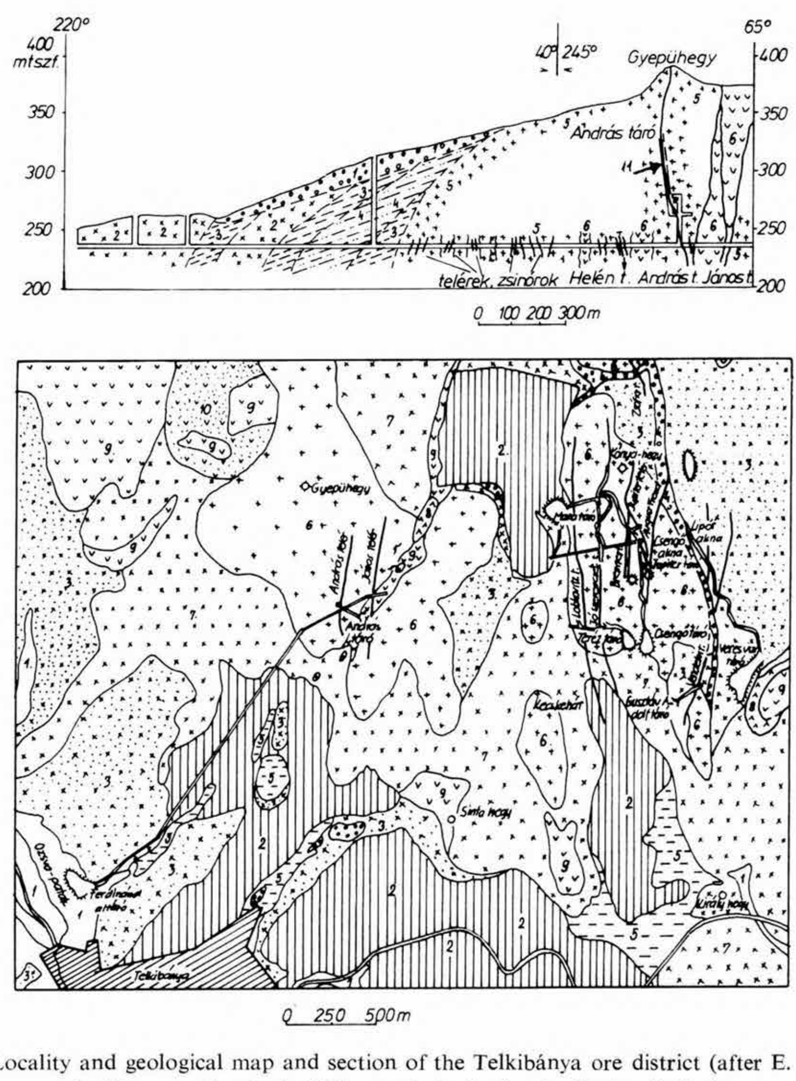


*Fig. 4.* Locality and geological map and section of the Telkibánya ore district (after E. SCHERF)
*1—4* Neogene sediments, *5—10* Neogene stratovolcanic series, *11* mineralized vein

epithermal. The prevailing sulphide mineral is auriferous pyrite the gold content of which may reach even the ton to twenty grams pro ton. Additional ore minerals are sphalerite, galena and chalcopyrite—the latter becoming more and more abundant towards the depth. The gold to silver ratio of the uppermost oxidized part of the vein filling varies from one to ten to one to thirty, whereas towards the depth the ratio decreases to one to eighty or even to one to a hundred. All available data suggest that the mineralization of the upper levels is the near-surface facies of a yet unknown polymetallic deposit situated somewhere at the intrusive subvolcanic level.

Much like in the case of the Börzsöny Mts also here the history of mining dates back to medieval times. Not only mining activity but also prospecting had been stopped however more than thirty years ago at Telkibánya. Revision of the results of previous exploration activities and reevaluation of the viability of another project in terms of present technical-economic conditions is underway.

Taking also the current state of metal prices into consideration, we may conclude that to revive the prosperity of the submarginal, partly exhausted polymetallic mines of north Hungary would require systematic exploration covering not only the near-surface indications but rather concentrating on the subvolcanic facies.

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# PALAEOGEOGRAPHY OF THE FORMATION OF INDUSTRIAL SAND DEPOSITS IN HUNGARY

by

GY. BIHARI

Several millions of tons of industrial quartz sand is mined and used all over the world every year.

A large portion of the territory of Hungary is covered by unconsolidated clastic sediments, mostly sands. Most of these sands are airborn or fluviatile but at some places, mainly in Transdanubia, off-shore sands and beach deposits can also be found.

There are hundreds of smaller or larger occurrences in Hungary where quartz-based sand is mined. The production of the majority of these mines goes to the state building industry or serves local building purposes.

From our point of view industrial quality sands or high grade sands are those quartz sands which meet the requirements of the glass or the foundry industry:

*1 Grain size distribution:* the glass quality sand should be well sorted in a narrow range; the molding sand is classified according to its AFS and ACS numbers.

*2 Grain morphology:* in case of molding sand the morphology determines the specific surface which in turn is directly proportional to the amount of binding material needed in molding mixtures.

*3 Chemical composition:* the most important parameter is the $SiO_2$ content. This should be between 85 and 96% for molding sands and over 99% for glass quality sands. A most characteristic impurity is $Fe_2O_3$. It must not exceed 0.1% in glass quality sands, its standard amount is between 0.015 and 0.1%. Other impurities important for the qualification of sands are carbonate and $Al_2O_3$.

*4 Mineral composition:* quartz or quartzite make up about 99,8% of the sands. Some undesired impurities are chromite, sillimanite and corundum.

*5 Heat resistance:* molding sands should be refractory.

Molding and glass quality sands are purified by different technological processes and separated according to their grain size. The complexity of the necessary processes greatly depends upon the quality of the raw material. So it is a must that the purest parts of sandstone and sand deposits be located and used for mining. The only successful way to do so is to study the genetics of high grade sand deposits.

There are only a few places in Hungary where high grade sand is mined. The most important glass quality sand deposit is the Fehérvárcsurgó one which meets 95% of the industrial demand. As far as molding sand is concerned 65% of the total national consumption is supplied by the the Kisörspuszta occurrence.

According to the latest reserve forecasts these two mines can meet the industrial needs for only twenty more years. Therefore prospecting and exploration of further high grade sand deposits is an inevitable necessity. Large scale prospecting has to be based on a suitable genetic and palaeogeographic concept.

High grade sands should be well sorted with no or only limited clay and $Al_2O_3$ content i.e. it should be deposited from a transporting medium with high current velocity. Silica enrichment may occur under hot and humid climatic conditions while a slightly acid transporing fluid rich in dissolved organic componds can easily strip the sand of its limonite content. The necessary low carbonate content is known to be brought about by $CO_2$-rich solutions.

Now the most important question is where and when these optimum conditions existed in the palaeohistory of Hungary.

The answer for the "where" question is fairly straightforward since these places must have been smaller or larger embayments where the following conditions were met:

*1* marshy back-lagoon environment rich in humic acid (to keep iron in solution)

*2* a current water flow within the bay directed towards the open sea (removing iron and resulting in a well sorted deposition of the sand);

*3* an area of denudation in the background producing large amounts of clastics, enriching the sand in quartz and removing the impurities such as placer minerals;

*4* presence of carbonatic bedrock and a carbonatic shoreline.

In Hungary the above geomorphological situation was existed in the course of the Late Pannonian. The Late Pannonian in Hungary can be characterized by a highly dissected shoreline with a great number of lagoons and bays and by a mountainous background morphology with the mountains emerging like islands.

Oligo—Miocene sands covering the surfaces provided an excellent denudation area for the quartz grains.

In time the Carpathian Basin became more and more detached from the Mediterranean and its salinity sharply decreased firtst in the bays than throughout the whole Basin. Most of the high grade sand deposits were formed at the beginning of the Late Pannonian along margins of the mountains. During most of the Late Pannonian quartz sand deposition took place in and around off-shore bars separating marshy back lagoons from the open sea. In the central part off these off-shore bars there is always a so called white-sand zone which contains only very limited amount of small pebbles and kaolinitic clay and is practically free of carbonates and iron oxides. Both seaward and shoreward from this central zone the grade of the sand gradually deteriorates. Shorewards from the white-sand zone grain size gradually increases, pebbles occur in form of thicker and thicker interbeddings and they may be mixed with the sand too. Approaching the shoreline three kinds of quality deterioration can be recognized:

*1* In the mouth of smaller creeks and rivulets, along with an increase of gravel also the ammount of clay, bauxitic clay and mottled clay admixtures is increased in the sand. Horizontal lamination may also occur.

*2* In marshy back lagoons, separated by off-shore bars from the open sea, a characteristic bluish-green, greyish-black clay appears while sand and pebbles become minor constituents.

*3* The alluvial fans of incoming streams are characterized by an increase of gravel and enrichment in limonite. These formations are obliquely laminated.

The seaward deterioration of the quality of the sand can be characterized by the lack of gravel accumulation of limonite (as a result of a more alkaline environment) and by the appearance of $CaCO_3$ in the sediment in the form of concretions and limestone banks. Further away from the shores the mica content of the limonitic sand increases and interbeddings of grey, greenish-grey micaceous clay occur.

The vertical division of the white-sand zone shows a definite transgressive character.

As a result of diagenetic changes postaccumulation up grading plays an important role in the establishment of the final grade of the sands. In this respect the drainage system of the carbonatic basement is also of crucial importance.

The above outlined genetic concept has some rather important practical implications. Prospecting activities have to be focussed on lagoons and bays along the Pannonian shoreline. As soon as in the course of exploration the succession of the layers is established, both shoreward and seaward quality deteriorations can be recognized within the white-sand zones. Based on these data the most promising directions where further exploration have to be concentrated can be established.

Based on this genetic concept the whole area of Hungary was evaluated and prognostic geological map showing the most promising areas was compiled.

In one of the most promising territories, the so called "sand belt" of the E Bakony Mts and of the S—SE marginal region of the Vértes Mts a systematic drilling campaign was undertaken. As a result, several new white sand deposits were discovered demonstrating that the above described genetic concept was correct. (Fig. 1)

Finally, a short is given account on two specific sand deposits: first the glass sand occurrence of Fehérvárcsurgó and then the molding sand deposit of Kisörspuszta.

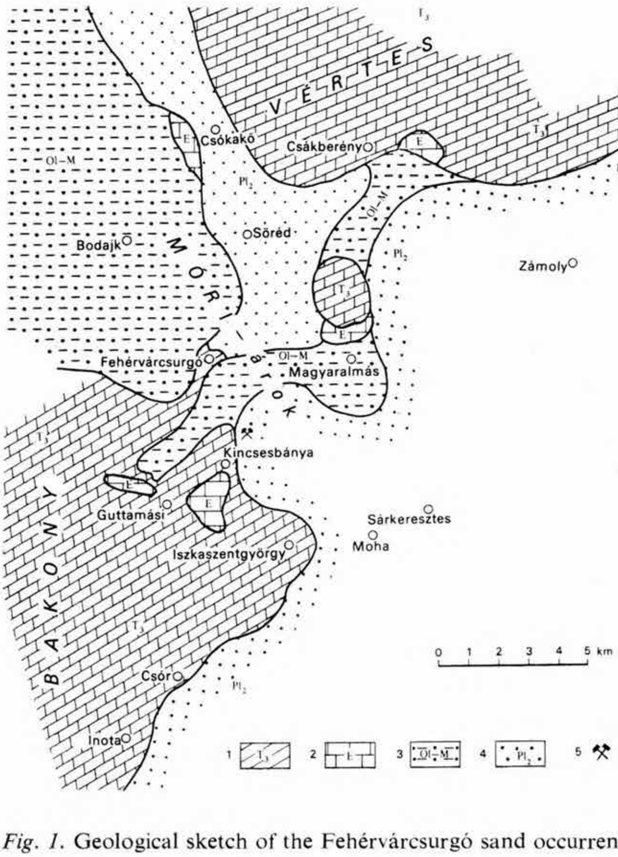


*Fig. 1.* Geological sketch of the Fehérvárcsurgó sand occurrence

*1* Upper Triassic dolomite, *2* Eocene limestone, *3* undistinguished Oligocene—Miocene sediments, *4* Upper Pannonian sand, *5* Fehérvárcsurgó sand mine

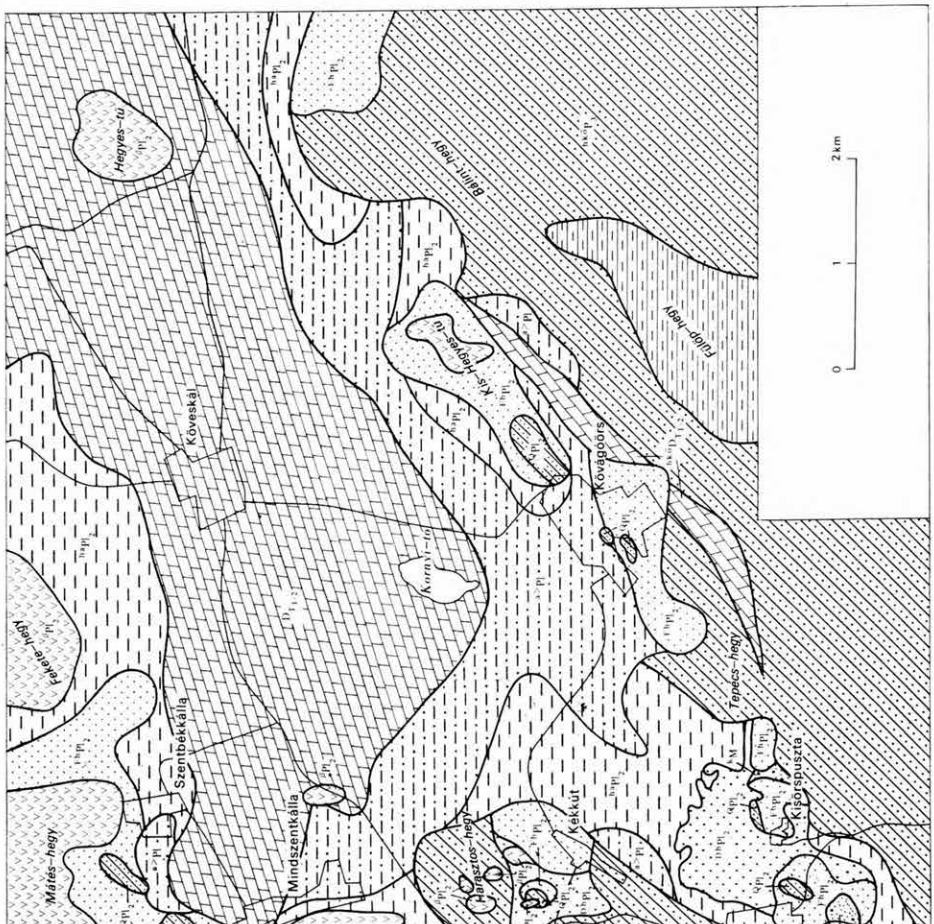


*Fig.* 2. Uncovered geological map of the Káli basin (Gy. BIHARI, 1984)

The sand deposit at Fehérvárcsurgó is located in one of the Pannonian embayments along the margins of the Bakony and Vértes Mts. This particular basin is underlain and also surrounded by Upper Triassic dolomites. The Upper Pannonian sand deposit strikes the NNW—SSE. The thickness of the sand complex increases from 3—4 m on the West up to 40—45 m as maximum on the East. In the ENE zone of the deposit, where the thickness is the greatest, a 4 to 5 m thick lignitiferous clay interbedding can be observed. In the SE part of the area the complex begins with white sand immediately overlaying the Upper Triassic dolomite. The glass sand of this deposit is a well-sorted white sand 99,5% of which is quartz. The processed product contains 0,025—0,050% $Fe_2O_3$ beside the 99.3% $SiO_2$.

The molding sand occurrence at Kisörspuszta (Fig. 2) is situated in a finger-like bay of the Kál-basin, one of the smaller basins of the Bakony Mts. Both the surroundings and the basement of this basin are built up of Permian sandstones. The light-grey Upper Pannonian sand is coarse-grained. Its $SiO_2$ content (98%) and the very low percentage of the iron oxide, carbonatic and feldspar impurities make it heat-resistant over 1400 °C.

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# DIATOMITE DEPOSITS IN SOUTHEASTERN SPAIN: GEOLOGIC AND ECONOMIC ASPECTS

by

J. P. CALVO and E. ELIZAGA

*Introduction.* Commercial diatomite deposits are widely extended in circum-Mediterranean areas, and elsewhere within Neogene formations in Europe. Spanish contribution of diatomites to the world production is limited in comparison with other main producer countries, but it is significant for regional markets in Western Europe and North Africa. Total production of diatomites in Spain exceeded 60,000 tons in 1982–1983 (official statistics, Ministery of Industry). It means a considerable increase, bearing in mind that spanish diatomite production reached up only 20,000 tons in the last seventies.

This noteworthy increase has been reached due to extensive explotation of diatomite deposits in some areas of SE Spain (prov. of Albacete), which have replaced other previously main producer areas.

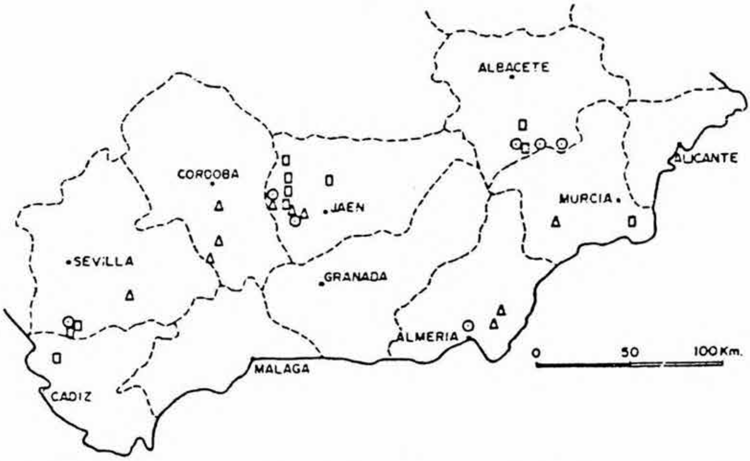


*Fig. 1.* Main productive areas in South Spain (after CALVO, 1978). Circles: active mining, squares: previous productive areas, triangles: non economic diatomite deposits

Diatomite deposits in Spain are restricted to the southern regions (Fig. 1). Diatomites occur in Miocene and Pliocene formations, both continental and marine. The highest concentration of active mining is sited, at present-day, in south Albacete. First references on diatomites in this area belong to AREITIO (1873), who published some brief reports on diatoms found in a few levels near Hellín. These former data were gathered up by AZPEITIA (1911) in his synthetical study on spanish diatoms. No more references on these deposits can be found until the paper by MARGALEF (1953), who updated the previously used palaeontological nomenclature and made an extremely detailed study on palaeoecology inferred from the diatom associations. The stratigraphy of the Neogene formations in which diatomite deposits are included was studied by JEREZ MIR (1973), CALVO et al. (1978), and BELLON et al. (1981). Further descriptions of diatomites in the region belong to CALVO et al. (1978), CALVO (1980), CALVO and ELIZAGA (1985).

## Geologic setting

Diatomite productive areas are located in the most external zones (Prebetic Zone) of the Betic Ranges, to the southeast of the Iberian Peninsula (Fig. 2). A major structural feature, the Subbetic Front, is extended in the vicinities of the study area. Miocene continental basins were the result of generalised extensional movements, subsequent to a main compressive phase during the Early Tortonian (CALVO et al., 1978). Afterwards, sedimentary filling of those basins occurred during the Vallesian and Turolian, as indicated by mammals (CALVO et al. o.c.) and radiometric data.

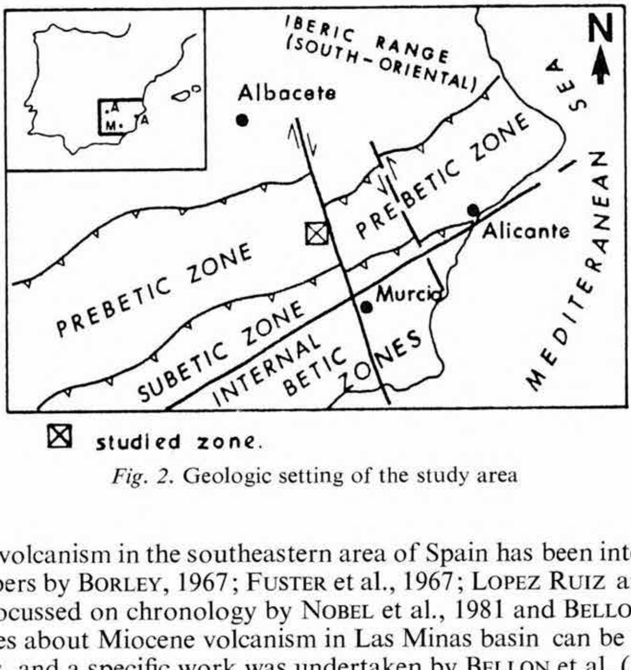


*Fig. 2.* Geologic setting of the study area

Neogene volcanism in the southeastern area of Spain has been intensively studie (petrologic papers by BORLEY, 1967; FUSTER et al., 1967; LOPEZ RUIZ and RODRIGUEZ 1980; papers focussed on chronology by NOBEL et al., 1981 and BELLON et al., 1983). Some references about Miocene volcanism in Las Minas basin can be found in some of these papers, and a specific work was undertaken by BELLON et al. (1981) to determine the radiometric age of volcanic rocks in the Cerro del Monagrillo. A $K/Ar$ age of $5.7 \pm 0.3$ Ma. was reported for these rocks. However, some uncertainties remained on the lithostratigraphic relationships between the volcanism and the Neogene sequence.

**Neogene lithostratigraphy**

Continental bearing diatomite formations disconformably overlied previous marine sequences of Middle Miocene age in the study area. Middle Miocene marine deposits consist of monotonous sequences of marly platform facies and biocalcarenitic, more littoral, deposits (CALVO, 1978) that discontinuously covered the area before retreat of the sea towards the south. The youngest marine deposits in the study area have been dated as lower Tortonian (USERA et al., 1979; BELLON et al., 1981).

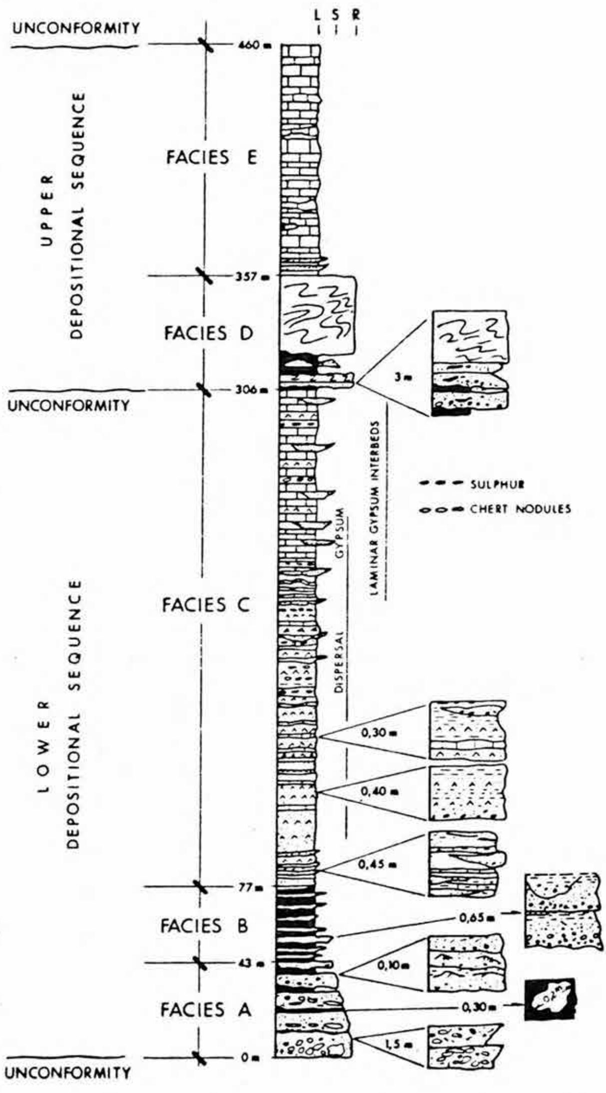


*Fig. 3.* Lithostratigraphic log of Miocene continental deposits in the western side of the Cenajo basin (after CALVO and ELIZAGA, 1985)

Investigations on diatomites have been focussed on three distinctive, although partially interconnected, basins. Two of them (Las Minas and Cenajo basins) show the most complete sequences of continental Miocene formations. Greatest thickness observed reaches up about 460 m in the western side of the Cenajo basin. Probably, this thickness value is greater in the central part of Las Minas Basin, but no drilling or geophysic data are available.

Surrounding reliefs (Mesozoic: conglomerates and sandstones, limestones and dolostones; middle Miocene: bioclastic carbonates and marls) supplied detritus to marginal areas in the basins. Outcropping terrigenous marginal deposits are specially well exposed where Triassic protruded the Neogene formations.

Two major Depositional Sequences have been recognized within the continental Miocene deposits. The Lower Depositional Sequence is up to 306 m thick in the western part of the Cenajo basin. It is made up by a fining upward sequence of conglomerates, interbedded sandstones and shales, and shaly gypsiferous beds passing upwards to predominant carbonate sediments. Three distinctive facies have been consequently defined within the Lower Depositional Sequence (Fig. 3).

Upper Depositional Sequence reaches up some more than 150 m in the described section. Two main facies can be easely distinguished. Clastic and strongly deformed-to-brecciated carbonate and diatomaceous sediments form the lower facies (facies D), overlying deposits of the previous sequence. A continuous succession of laminated chalks and diatomites (facies E) tops the section at this point.

Both main Depositional Sequences are largely extended along the Cenajo basin and also they can be detected in the larger Las Minas basin. The latter one displays a more definite gypsiferous sequence (tabular continuous gypsum beds with formerly economic deposits). An outstanding feature in this basin is the occurrence of the above mentioned volcanic rocks of the Cerro del Monagrillo. Besides the radiometric data obtained, physical evidences for an intra-Miocene age of this volcanism also may be shown. They will be discussed below.

## Sedimentary evolution

Miocene continental basins in the study area behaved as tectonically controlled, close lake systems. At least during some periods, partial connection among the different basins must be suspected, but no definitive criteria have been yet found to prove it. Sedimentary fill of the basin was accomplished by means of both detrital supplies from surrounding reliefs and autochtonous chemical and/or biogenic lacustrine deposits. Vertical sequences of the major facies, as previously described, closely reflect a progressive enlargement of the lake facies and, probably, a relative deepening of the water bodies.

Results from the sedimentological analysis, the Lower and Upper Depositional Sequence (CALVO and ELIZAGA, 1985) may be summarized as follows:

### Lower Depositional Sequence

Facies A (thinning-upwards conglomerate—sandstone sequence): it represent a first stage of lake fill by adjacent small alluvial fans. Coarse detrital sheets entered the lake leading to a slow fill and subsequent transgression of the lacustrine facies.

Facies B (channeled and sheet sandstones interlayered with lutites): they are interpreted as even-exposed, marginal lacustrine deposits with a progressive minor clastic inflow.

Facies C (mudstones with thin sandstone sheets and gypsum): it represents a steady transgressive trend in lake deposition. Lower part of this facies may be ascribed to sedimentation in more or less shallow, slightly saline, lacustrine areas, with gypsum precipitation and further sulphur nodules development. Upper part of the facies is carbonate predominant, with minor clastic inflow (thin lenticular sandy sheets) by micro-density currents.

**Upper Depositional Sequence**

A sharp sedimentary discontinuity indicates the lower boundary of this sequence. Basinal conditions were reached at the end of the underlaying depositional unit. An initial event of large clastic inflow is represented by conglomerate—sandstone turbidite beds, that was continued by a 30 m thick single intraformational sedimentary slump structure (Fig. 4). Major descriptive features in this slump are translational and rotational slides, as well as slump-breccias. As pointed out above, facies D includes frequent diatomite beds, altogether with laminated chalks and thin sandstones.

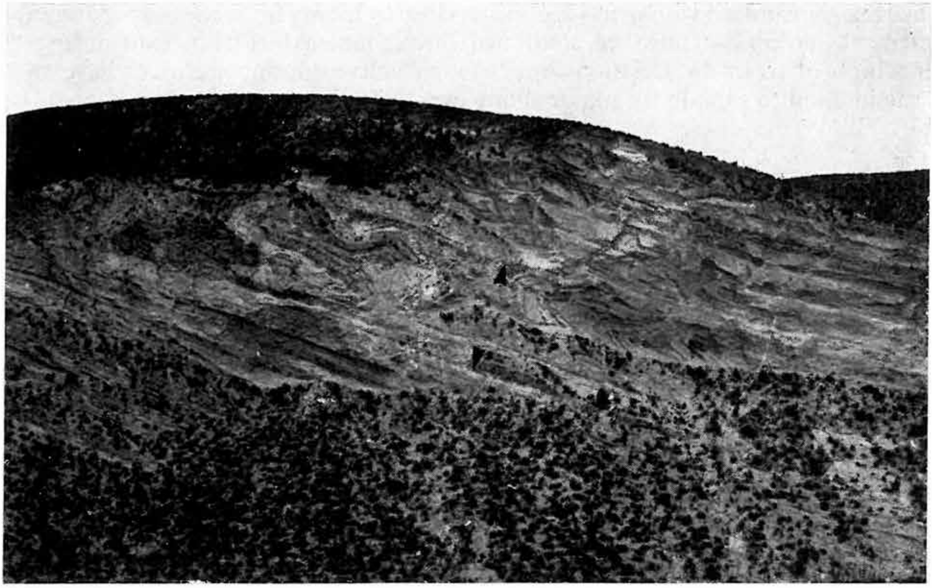

*Fig. 4.* General view of strongly deformed lacustrine deposits (Facies D). Arrows mark the base and the top of the slumping facies

Upper limit of the slump deposits is also marked by a sharp discontinuity. A thick, up to a hundred meters, sequence of very regular laminated chalks and diatomites tops the Miocene sequence in the study area.

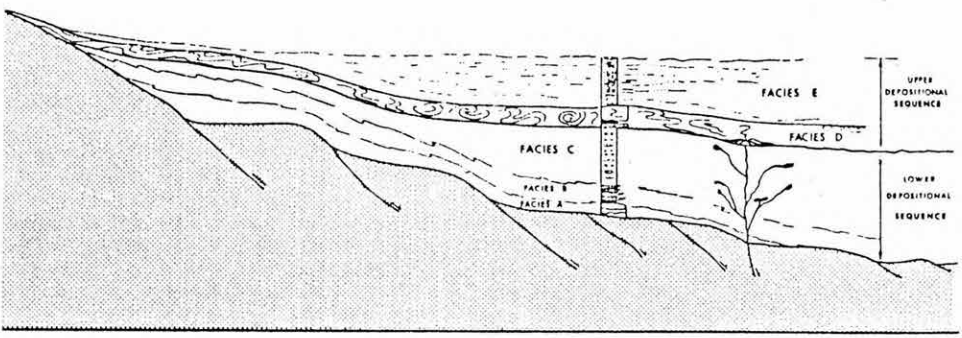


*Fig. 5.* Sketch of successive stages of sedimentation in the studied basins. Note tectonic readjustment of the lake floor, as well as the proposed localization of vulcanism in the Neogene sequence

Sedimentary evolution of the basins has been summarized in Fig. 5. Progressive deepening of the basins was accomplished through relative fault-controlled movements of the floor lake. The emplacement of volcanic rocks took place in the area after sedimentation of the Lower Depositional Sequence. Physical evidences of volcanic rock fragments mixed with marly sediments close to the slump facies corroborate that statement. Volcanism must be continued throughout a further period during the deposition of facies D. Relationships between volcanism and tectonics have to be borne in mind to explain the major slump event affecting lacustrine facies.

## Discussion

The lithostratigraphic and sedimentological analysis of continental Miocene deposits in two of the main diatomite producer basins in SE Spain allow us to point out some main conclusions concerning the origin of these diatomite deposits. The continental character of the deposits is revealed by associated faunas (fishes, ophidia, amphibians, micromammals (Calvo et al., 1978) and the diatoms themselves, although some periodical brackish conditions can be deduced from the analysis of this flora (Servant-Vildary, 1986). Also, the geological evolution of the area asserts this idea.

A close relationship between post-Alpine lamproitic volcanism (Nobel et al., 1981) and the flourishing of diatoms in the basins may be proved, bearing in mind the occurrence of volcanic traces within the Upper Miocene sequence. So, volcanic traces have been found after the sedimentation of the Lower Depositional Sequence. Volcanism was indeed coincident with a tectonic readjustement of the basins, that triggered largescale delapsional events. Volcanic rocks have been dated at $5.7 \pm 0.3$ Ma, i.e. Turolian. It is also in agreement with the chronostratigraphic data deduced from mammal faunas found near the top of the sequence (Upper Turolian) (Calvo et al., 1978).

Diatomite beds are restricted to the Upper Depositional Sequence. So, they have been recognized more or less discontinuously along 154 m. Silica content was analysed thoroughout this interval. Palaeoecological data from fauna and flora associations have not yet been entirely evaluated but new studies are going on now. Anyway, present day evidence deduced from preliminary analysis of diatoms and associated faunes, seem to indicate a rather shallow, this is not very deep, lake during deposition of diatomites and chalks. Palaeobathimetric interpretation in lake basins are usually risky due to the variability of conditions in particular cases. As pointed out by GIBLING et al. (1985), "the depth of water in a perennial lake need not to have been great, and good preservation of laminae directly depends of the upper boundary of anaerobic bottom, that at times may be close to the sediment—water interface". The same idea has been expressed by BUSSON and NOEL (1972), who pointed out the relative independence of bathimetry and euxinic conditions at the lake bottom. So, specific thermal and/or chemical stratification of water in the lake seems to be sufficient for laminite preservation.

Relatively shallow conditions, perhaps ten meters or less depth, characterized the deposition of diatomites in the studied basins. It may be supported by interfingering and alternation of laminites with more littoral deposits (limestones and marls made up by fixed benthic algae and gastropoda) in some areas. Finally, seasonal control of sedimentation may be suspected by common occurrence of laminite couplets. An average thickness of 0.3—0.5 mm has been usually observed for these couplets.

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# QUANTITATIVE ATTEMPT TO RECONSTRUCT PALAEO PORE-PRESSURES BASED ON NEOGENE SEDIMENTATION HISTORY IN THE PANNONIAN BASIN

by

M. HARGITAY and L. DUSZA

*Introduction.* Young sedimentary basins are often characterized by the occurrence of anomalously high pressure both in the basin fill and in the basement. A number of mechanisms for the formation of high fluid pressures have been suggested and compaction of sediment is believed to be the main cause of overpressuring (DICKINSON, 1953; RUBEY and HUBBERT, 1959; BREDEHOEFT and HANSAW, 1968; SMITH, 1971; MAGARA, 1978; KEITH and RIMSTIDT, 1958). In an evolving sedimentary basin, burial is often so rapid that the pore fluid cannot escape quickly enough for fluid pressures to attain hydrostatic values, and for sediments to compact normally. This process may be augmented by aquathermal pressuring defined as pore-pressure created by thermal expansion of the pore fluid during the burial (BARKER, 1972; DAINES, 1982).

Some authors suggested that clay-mineral transformation also contributes to the formation of overpressure (HANSAW and BREDEHOEFT, 1968; BURST, 1969). MAGARA (1975) and CHAPMAN (1980) argue that this factor is of secondary importance and the numerical analysis of KEITH and RIMSTIDT (1985) confirms their opinion. Therefore, the compaction of sedimentary rocks may be taken to be the prime factor in overpressure generation.

## The method

We have considered the one-dimensional problem and used that version of TERZAGHI's (1943) equation which is not restricted by the infinitesimal strain assumption. This is the assumption that the dimensions of a volume element do not change with time.

The infinitesimal strain equation is the following:

$$\partial/\partial x[K/\mu(\partial u/\partial x)] = a\partial u/\partial t,$$

where u = the excess pore-pressure, a = the coefficient of compressibility, K = the permeability, and $\mu$ = the viscosity. We basically use this equation, but, with a new vertical coordinate which is fixed to the material (rock matrix). The material coordinate is:

$$z = \int_{o}^{x} [i - \varphi(x)]\, dx.$$

The equation is solved numerically, and parameters are calculated step by step by means of a finite-difference method (Crank—Nicholson scheme). Moreover, the constant parameter a in the Terzaghi's equation is replaced by $\sigma(\varphi)$ which describes the relationship between effective stress and porosity for sedimentary rocks. The definition of effective stress is given by the equation (HUBBERT and RUBBEY, 1959):

$$\sigma = S - p,$$

where S and p are the total and pore-fluid pressure, respectively. The effective stress gives the stress which is actually present in the rock matrix, therefore it is often called as rock-frame pressure. The porosity vs effective stress function is different for different rock types and can be obtained by using empirical porosity-depth curves (Fig. 1).

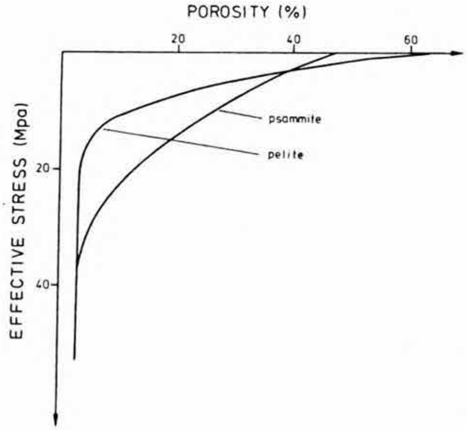


*Fig. 1.* Effective stress vs porosity relationships derived from SZALAY'S (1982) porosity vs depth relationships for normally compacted sediments in the Pannonian basin

The porosity vs depth function, measured at hydrostatic pressure, is called the normal porosity trend. In zones of overpressure the porosity is higher than the normal value depending on the rock-frame pressure.

A relationship to describe the dependence of the permeability on the porosity is also required. Unfortunately there is no general mathematical relationship expressing permeability in terms of porosity, that can be applied to all cases. It is because, there can be very high porosity with very low permeability if the pores are not connected, or the connecting channels are so fine that the surface tension is strong enough to prevent fluid movement between the pores. The controlling factor is, therefore, not the porosity itself, but the geometry of pores and the way they are connected.

In our modelling we used the following equation:

$$\log K = a\varphi + b,$$

where K is the permeability in milidarcy and $\varphi$ is the porosity in percent (KEITH and RIMSTIDT 1985). The value of a and b parameters were taken from KEITH and RIMSTIDT (1985) for clays. For sandstone we assumed a = 4 and b = 3.5 and found that these parameters gave good results.

The sediments of the Pannonian basin can be divided into two broad rock types: pelites (clay, shale) and psammites (gravel, sandstone, conglomerate). These rocks are intermixed on a finer scale than the practical sizes of the blocks used in the finite differences method. In our case each block is divided into two volume fractions occupied by a pelite and a psammite component. This division allows a simple calculation of the bulk properties.

For permability we used the following formula:

$$K = K_{sh}(K_s/K_{sh})^{\langle 1-R \rangle},$$

where $K_{sh}$ is the shale permeability, K- is the sand permeability, and R is the pelite to psammite ratio.

In making the calculation, the time-span of sedimentation is divided into small $\Delta t$ intervals and the sedimentation is modelled by addition of a thin layer in each $\Delta t$ time interval. The load generated by a new thin layer is carried at the beginning totally by the pore-fluid and it is transferred progressively to the rock matrix. On the surface the pressure is always hydrostatic, the bottom can be permeable (hydrostatic pressure) or impermeable ($dp/dz = 0$).

## Results

The results of the calculations are shown in Fig. 2 and 3 for boreholes Hód—I and Békés—1 respectively. In the case of borehole Hód—I it can be seen that a large overpressure has developed in the deeper strata and that it drops down to the hydrostatic pressure close to the basement. This pressure vs depth relationship is very important in controlling the migration of fluids. Namely, the peak of the overpressure constitutes a kind of "seal". Migration can occur upward in sedimentary rocks above the peak and downward in sedimentary rocks below the peak. The boundary conditions were defined at the surface and at the top of the basement. The excess of pressure

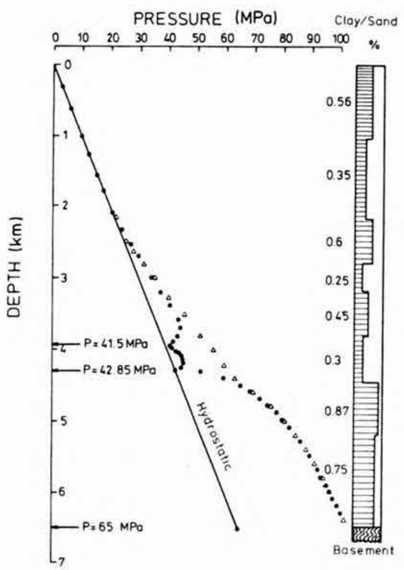


*Fig. 2.* The calculated pressure vs depth curve for the borehole Hód-I by supposing permeable basement (△), and, in addition, two nearly hydrostatic boundary conditions close to a depth of 4 km (●)

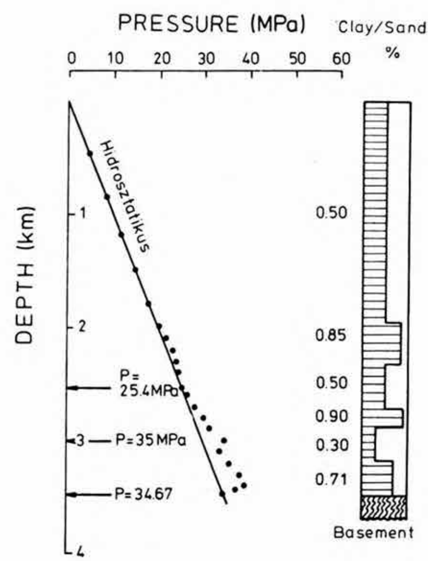


*Fig. 3.* The calculated pressure vs depth curve for borehole Békés-1 with permeable basement and two other boundary conditions

steadily increases with depth and the rate of increase changes according to the pelite—psammite ratio. The observed pressure seal can be modelled satisfactorily only by defining other boundary conditions. We assumed that the fluid pressure is close to hydrostatic at 3950 m and hydrostatic at 4285 m. Fig. 2 shows that the definition of these boundary conditions leads to good results. The pressure seals are not symmetric to the peak point of the pressure curve because of the downward decrease of the porosity. The depth position of the peak is not constant with time but moves downward.

In case of borehole Békés—1 the calculated pressure vs depth curveshows overpressures smaller by one order of magnitude than in well Hód—I. This is mainly the consequence of the boundary conditions. At a depth of 2540 m hydrostatic pressure was measured, the basement was assumed to be permeable and therefore hydrostatic pressure was taken here, too. At a depth of 3020 m, 60 MPa pressure was measured in the pore fluid which is approximately equal to the lithostatic pressure. This high overpressure was unexpected, because:

i the layer consists of sand of relatively great permeability,
ii 30 MPa excess of pressure between two nearby hydrostatically pressured points, seems to be unrealistic.

We think therefore that either this was a wrong measurement or the result of a strong lateral effect which cannot be taken into consideration in a one-dimensional model.

*Conclusions.* According to our model calculation it can be understood that overpressures occur usually below 1200—1500 m only in the Pannonian basin. During sedimentation alow overpressure in the near-surface layers can occur only for a short time because, as a result of the high porosity and permeability, the fluid could migrate away very rapidly.

To explain the evolution of pressure seals, lateral fluid migration has also to be taken into consideration. In one-dimensional calculation it can be modelled only by choosing suitable boundary conditions. Such boundary conditions can be defined if the position of highly permeable layers in a sedimentary basin are known. This can be obtained by interpretation of seismic sections and well-logs.

Two-dimensional models are more suitable to describe lateral migration. Their success, however, is also strongly dependent on the choice of adequate boundary conditions.

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# ORIGIN, DISTRIBUTION AND ECONOMIC IMPORTANCE OF THE MESSINIAN GYPSUM IN THE MEDITERRANEAN

by

A. HERRMANN

Gypsum ($CaSo_4 \cdot 2H_2O$) was probably the first material to be used as mortar: its use was first reported in connection with the erection of the towers of Jericho about 7000 BC. It can be assumed that the art of calcining gypsum to plaster has been known to man for about 10,000 years. This is certainly due to the low temperature of 120 to 180° C required for driving off the cristallisation water to produce the semi-hydrate $CaSO_4 \cdot {}^1/_2H_2O$.

It was evidently *not* Messinian gypsum that was used for the world premier of gypsum calcination. Most probably, however, it *was* Neogene gypsum from the Mediterranean region. We know that beginning in the Eocene, through the Oligocene and especially during the early and middle Miocene, gypsum deposits were formed in an area which was considerably larger than the Messinian evaporite basin which itself was not much more extended than the present Mediterranean. As shown in probably the best current general representation by ROUCHY (1982, Fig. 1), the major sedimentation areas of Late Miocene gypsum are beneath the present Mediterranean. Only at the margins do they "rise" to the mainland, where some of these deposits can be profitably worked.

The structure of sulfate rocks differs depending on the depth of the sea in which they were formed. That has been known for decades and made clear by G. RICHTER-BERNBURG (1955), using the example of the Central European Zechstein: the lithofacies in the deeper parts of the basin is thinbedded and of small thickness, that on the flanks of the swells or near the margins of a basin is thick and massive. These findings, however, cannot necessarily be applied to the conditions in the Mediterranean area of the late Miocene because partly completely different types of sulfate sediments occur in the Messinian, particularly the cristalline gypsum, i.e. the selenite. In contrast to numerous older evaporite formations, special conditions prevailed in the Messinian. This is clear from the "dynamic model for a shallow basin setting at world ocean sea level" by FABRICIUS, HEIMANN and BRAUNE (1978). ROUCHY (1981, 1982) modifies this model by imagening that the main basin was subdivided into several secondary basins which are situated one behind the other. Shallow water deposits and even distinct sebkha formations were of relatively great importance in the Messinian at least in the areas accessible to "land-locked people", as DROOGER (1973) put it. But even there, non-selenitic, normal, i.e. compact to fine-crystalline gypsum or anhydrite rocks are found locally, e.g. in the core of the rapidly subsiding Chelid Trough in Algeria or of the Caltanisetta Trough in Sicily (RICHTER-BERNBURG, 1973; GERSONDE, 1980). Their occurrence indicates a bit deeper water environment.

There is no doubt, however, that there are large areas—mainly below sea level—with sulfate deposits for which the "deep-basin model" of the minority of authors whom DROOGER (1973) calls "sea-borne people" does not necessarily apply. Typical

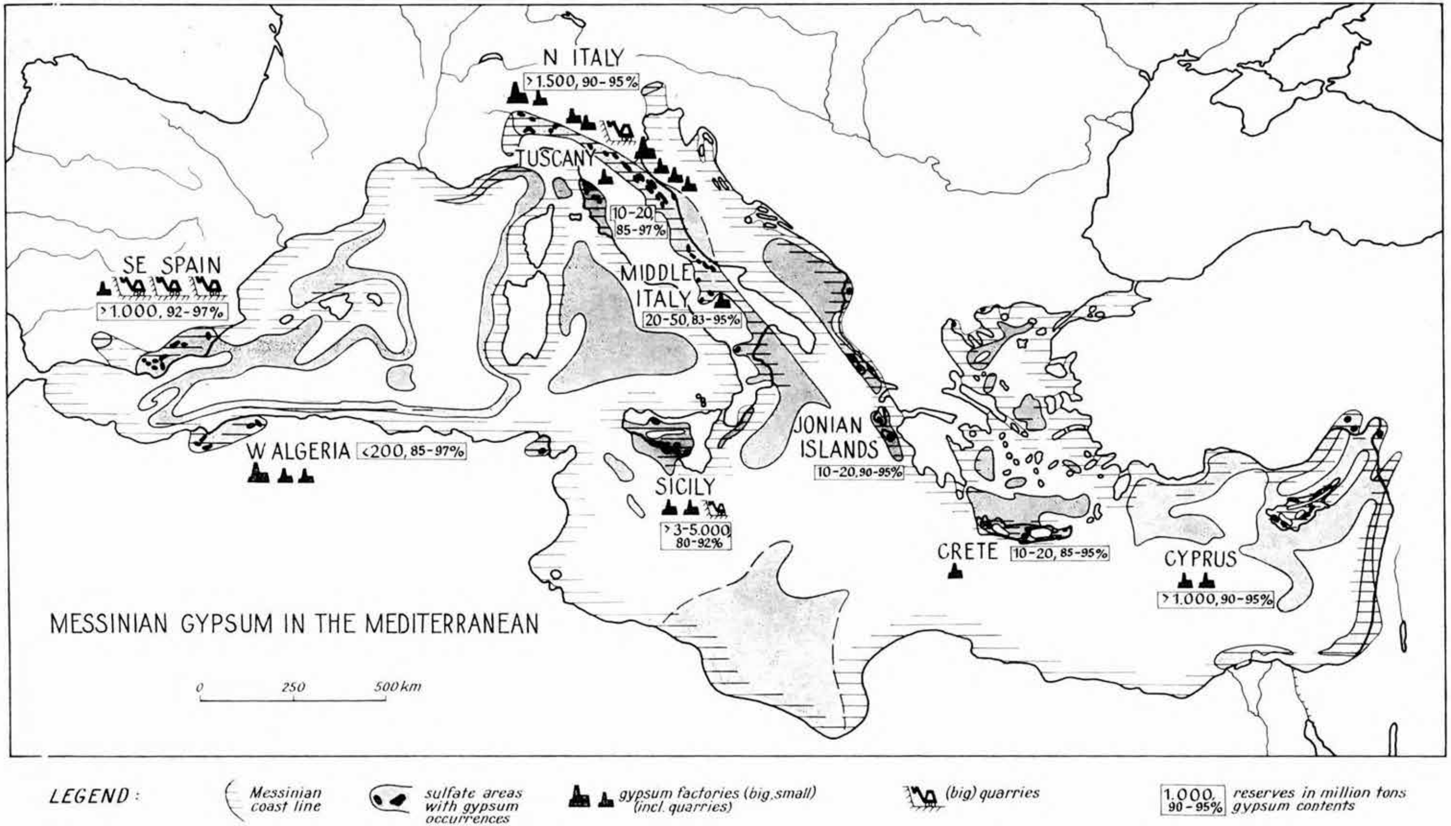


*Fig. 1.* The Messinian basin in the Mediterranean (adapted from ROUCHY, 1981/82) showing distribution, quality and quantity of gypsum as well as important exploitation areas and enterprises

sediment structures are found in the sulfate rocks in the few drilling samples collected during the Glomar Challenger campaigns: in addition to layered structures, there are especially so-called "chickenwire" gypsum or anhydrite (in German "Hühnerdraht-Anhydrit", in French "l'anhydrite grillage à poulet") with a cloudy structure. There are also various types of redeposited rocks, e.g. gypsum turbidites. Most probably they are the result of submarine landslides of varying size from the margin to the basin (HERMANN and RICHTER-BERNBURG, 1955).

These processes indicate considerable differences in the seafloor level in the whole area of deposition (SCHREIBER, 1985): At the same time as the sulfate was deposited on the slopes the deeper parts of the basin were more or less rapidly filled with rock salt once the necessary salt concentration had been reached; later on only rock salt was formed in the whole basin.

It can be seen in the only area of Messinian deposition of thick rock salt, or even potash, that has become land (in Sicily) that also during the Messinian the thickest sulfate layers were deposited in more shallow water. Both towards the depth and in the direction of the basin margins, the thickness and purity of the gypsum decrease (RICHTER-BERNBURG, 1973). Additionally, the gypsum content, which is so important for its processing, is reduced even more due to the occurrence of anhydrite, i.e. calcium sulfate without crystal water, in both the deep and extremely shallow environments. The gypsum is purest in areas with thick sulfate beds, which in the Messinian are characterized by the "selenite" facies in its various forms: from large crystals measuring several meters across to microscopic size. In the selenite facies, even in areas where it attains a thickness of more than 150 m, *no* anhydrite has been found up to now, whereas its occurrence is characteristic of the core of sulfate layers in older formations.

The stratigraphic position of the crystalline gypsum occurrences in the late Tertiary is responsible for further positive characteristics: the mostly rather unlithified cover, i.e. overburden, is easily erodible and the diagenetically only slightly altered crystalline rock is very resistant to leaching, i.e. karstification.

These stratigraphic conditions result, at least in the crystalline gypsum facies, in sufficiently thick and pure gypsum occurrences without an anhydrite core. These occurrences have only a thin cover over wide areas. Illuviation of foreign rock material due to solution is a minimum. All in all, the crystal facies of the Messinian gypsum is an especially high quality mono-mineralic rock that is available in large quantities in various regions. Because preference is given, for a wide variety of reasons, to other priorities, perhaps only half of the indicated reserves of 8–10 thousands of millions of tons can be exploited, but even these quantities will probably be sufficient to meet demand for more than a thousand years. As far as I know, such immense reserves of gypsum are available nowhere else in the world.

It is interesting that the economic importance of Messinian gypsum varies so widely in the different parts of the Mediterranean region. Let's begin with *Spain*, which is among the countries with the highest gypsum reserves in the world, distributed throughout the country in extensive deposits in several geological stages. The Messinian gypsum deposits are situated on the southeast coast. This gypsum, in crushed form, has been shipped to the American east coast and to Scandinavia for many years. After a power plant fired only with coal imported from the US was erected on the coast near the deposits of gypsum, sales abruptly increased due to the more favourable freight costs to and from the US.

The situation in *Algeria* is quite different: Algeria is an oil and gas-producing country with a rapidly growing population. This country is forced to develop a building materials industry, especially for housing. Several modern gypsum plants have been erected in rapid succession at sites with Messinian gypsum.

In *Italy*, especially in the north, there are numerous plaster plants of varying capacity, but no plasterboard factory. The coarse-spathic crystalline gypsum is also used for wall paneling and table tops; the latter are sold as "lava del mare". The largest accumulations of Messinian gypsum are found in Sicily. Their extent can hardly be imagined, but they are exploited only on a comparatively small scale.

In contrast to the western Mediterranean, the exploitation of Messinian gypsum in the eastern part, chiefly on the *Greek islands*, is rather underdeveloped, although this type of gypsum occurs quite frequently there. The only plasterboard factory in the entire Mediterranean region was in Cyprus until the 1973 Arab–Israeli war closed its markets in the Near East.

Some final remarks on the controversy between the adherents of the "deep-basin model" and the supporters of a "shallow-basin model" that has been discussed, sometimes vehemently, at numerous congresses held exclusively on this subject over the 15 years since the first cruise of "Glomar Challenger" in 1970. This discussion can be regarded as more or less settled on the basis of a modified shallow-basin model of ROUCHY (1981). It can be traced, however, in a textbook, edited by STANLEY and WEZEL. As to the formation of evaporites during the Messinian, Mrs. SCHREIBER (p. 31) comes to the conclusion that various sedimentation environments, due mainly to differing water depths, can exist *within* any one basin. The palaeontological findings have been generally interpreted to determine that the water depth in the Messinian basins did *not* exceed 500 m (MEULENKAMP, 1977, verbal commentary from L. BENDA). And ROUCHY (1981, 1982) has developed the reasonable concept that the deposition of evaporite sequences did not necessarily take place at the same time throughout the whole Mediterranean region.

Evidence for this that can be recognized by everyone is provided by an example of the Paratethys in the Balkans: A "salinity crisis" occurred in the early to middle Miocene similar to that of the Messinian. ROUCHY (1982, Fig. 9) shows that corresponding conditions apply also to the individual subbasins of the Messinian, disregarding the fact that only part of the sedimentation of the carbonate, sulfate, and chloride evaporites took place contemporaneously. The primary formation of anhydrite can also be easily included in this concept: on the one hand, because of the high temperatures prevailing in supratidal areas and, on the other hand, because of the hypersalinity of the descending, highly concentrated brines in the deeper parts of the basins.

Finally, using a "shallow-basin model" extended to a certain depth as presented by FABRICIUS, HEIMANN and BRAUNE (1978), it is much easier to explain the numerous, sometimes more than twenty sequences of evaporite deposition, in some places containing repeated illuviations from the surrounding land. In any case, it is considerably more difficult to explain using the "deep-basin model" (HSÜ, 1973; HSÜ et al., 1973; 1978; GVIRTZMAN and BUCHBINDER, 1977; RYAN and CITA, 1974), which assumes repeated, more or less complete desiccation of the entire basin of the Messinian. Eustatic fluctuations of the sea level of the ocean, hypothesized on the basis of other evidences, are more likely, but only to explain the *cycles* during the Messinian.

As quite correctly remarked by SCHLAGER and BOLZ (1977, p. 601: "We are aware that the extrapolations are made from very limited data.") the details of the saline sequences in the present-day deep parts of the Mediterranean are still almost unknown. If total desiccation repeatedly took place, it should be possible to furnish evidence for a quantitative relationship between the evaporable water column and the thickness resp. thickness ratios of the various evaporites, especially of gypsum and rock salt.

Van Couvering (1976) also supports this view, with his opinion that a complete evaporation of the Mediterranean according to Hsü's theory (et al., 1973; 1978) cannot have taken place because the evaporites in the deep areas are too thick. In his conception the evaporites could have attained such a thickness only through the evaporation of brines following the first desiccation.

The large, more or less uninterrupted thickness of gypsum, e.g. up to 300 m in the Gessoso Solfifera Formation in Sicily (Gersonde 1980, p. 23), cannot, in any case, plausibly the explained using the "deep-basin model". Rather, a gradual increase in the salt content in a continuous lateral brine current in the sense of Richter-Bernburg (1955) or Rouchy (1981) must be hypothesized.

Neither should the fact be ignored that the boundary between the African and the European plates runs through the Mediterranean area: sediments of the same lithofacies can be observed in altitudes which differ in level by up to 2000 m. And then large areas are supposed to have existed unaltered as a structural low since the Middle Miocene? (see also Leenhardt, 1973).

Summarizing, I come to the conclusion that the "Messinian event", the often cited "salinity crisis", was not so unique as made out to be by a number of authors. Instead, it took place in a saline basin under the usual conditions that prevailde several times before in the history of earth, although with typically Mediterranean, i.e. "Messinian" aspects.

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# GEOCHEMICAL AFFINITY OF THE MAIN IRON—MANGANESE ORES OF EGYPT WITH LATE TERTIARY BASALTIC ACTIVITY

by

O. A. KAMEL, E. A. NIAZY and A. Y. ABDEL AAL

*Introduction.* The Late Tertiary volcanism is represented by different basaltic occurrences and the activity of thermal springs (Fig. 1). The *K/Ar* ages indicate that volcanicity started towards the Late Oligocene: 24 Ma, and have continued through the Lower Miocene: 16 Ma (MENEISY and ABDEL AAL, 1984). This is similar to the onset of volcanism around the Red Sea margins.

Ascending mineralized hydrothermal solutions or thermal waters have developed through replacement and cavity filling some important Fe—Nn deposits in Egypt. The common relation between the sites of these deposits and those of Tertiary basalts, and the marked silisification of the country rocks present additional criteria that most of these mineralized solutions are related to the Tertiary volcanism. The epigenetic Fe—Mn ores (Bahariya Oasis), Mn-veins (Elba) and modified sedimentary Mn—Fe deposits (Um Bogma) are chosen to carry out extensive studies on them.

## Iron—manganese ores of Bahariya Oasis

The Bahariya Oasis lies in the central plateau of the Western Desert (Fig. 1). It represents a large oval-shaped depression characterized by a bounding escarpment and a large number of separated conical hills within the depression. The exposed sediments forming the floor of the depression and the scarpment walls are ranging in age between Lower Cenomanian and Quaternary.

The Fe—Mn ores occur in the northern and north-eastern parts of the oasis in the form of bed-like or lenticular bodies. They consist mainly of goethite, hematite, hydrogoethite, quartz, halite, chalcedony, psilomelane—cryptomelane and barite. Siderite, pyrolusite, manganite, chamosite, calcite and aragonite are less commonly encountered. KAMEL (1971), mentioned that the ores could be genetically classified to

*a)* massive pisolitic and oolitic hydrogoethite of Lower—Middle Eocene age and

*b)* hydrothermal—metasomatic hematite, goethite and goethite—hematite.

*Geochemistry.* The average chemical analyses and trace elements of the Fe—Mn ores of the Bahariya Oasis after MAHGOUB and AMER (1964), SHERIF (1969), KAMEL (1971) and NIAZY (1975) are given in Table 1. It is clear that the Fe—Mn ratio varies widely in the ores and no clear relationship between Fe and Mn can be observed. A positive correlation has been noticed between Ba and Mn (Fig. 2). which is similar to that of BONATTI et al. (1972a) for the hydrothermal Mn-deposits of the Afar Rift, although the line is shifted lower, due to the Mn-content encountered. Furthermore BERGER (1968) indicated that psilomelane is more abundant in hypogene deposits than cryptomelane and that these deposits are enriched in Ge and Ba than in supergene deposits.

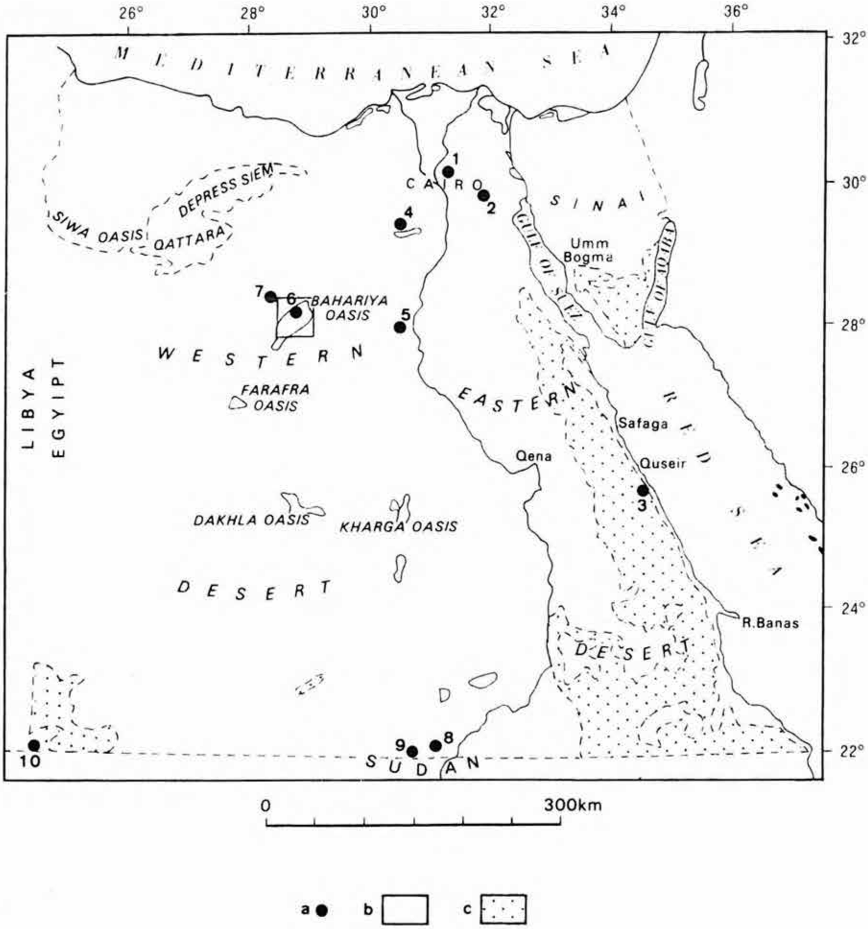


*Fig. 1.* Location map

*a* Basaltic occurrences, *b* Fe—Mn mineralization, *c* basement complex.—*1* Abu Zaabal, *2* Cairo—Suez district, *3* South El-Quseir, *4* Quatrani, *5* Bahnasa, *6* Bahariya oasis, *7* Naqb Siwa, *8* Gebel Siri, *9* Gebel El-Fantas

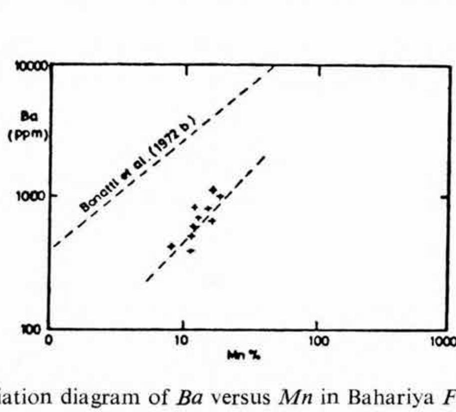


*Fig. 2.* Variation diagram of *Ba* versus *Mn* in Bahariya *Fe—Mn* ores

**Chemical composition of Bahariya iron manganese-ores**

*Table 1*

| Element | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| Major elements in weight percent | | | | | | | |
| $SiO_2$ | 1.42—6.28 | 23.40 | 3.01—10.15 | 5.76 | 10.00—13.40 | 8.70 | 35.00—44.57 |
| $Al_2O_3$ | 0.10—0.60* | 2.15 | 2.16—3.03 | 1.94 | 4.32—4.88 | 1.10 | 6.99—10.95 |
| Fe | 50.00—61.27 | 49.12 | 45.60—57.35 | 43.24 | 47.50—49.06 | 44.89 | 22.00—29.00 |
| Mn | 0.98—2.25 | 3.03 | 1.93—4.28 | 11.58 | 0.50—0.52 | 0.79 | 0.15—0.30 |
| Ca | 0.40 | 2.23 | 0.49—2.76 | 2.96 | 2.50—2.79 | 2.43 | — |
| Mg | trace | 0.29 | trace —0,05 | 0.58 | 0.39—0.046 | 0.94 | — |
| $Cl^{-1}$ | 0.48—0.83 | 0.82 | 0.76—1.41 | 0.83 | — | — | — |
| S | 0.44—1.35 | — | 0.43—0.77 | — | 0.07—0.17 | 0.60 | — |
| Fe/Mn | 27.23—51.02 | 16.21 | 10.65—22.72 | 3.90 | 94.35—95.00 | 56.82 | 70.0—200.0 |
| Trace elements, in ppm. | | | | | | | |
| Ba | 8330** | 400—600 | 400—1000* | 712 | — | 300 | 135—140 |
| P | 1600—2100 | 568 | 700—2050 | 830 | 8100—12500 | 4000—5000 | 10 |
| Ti | n.d.—300 | 100 | 975—1870 | 300 | 120—1200 | 100—1600 | — |
| Sr | n.d.—200 | n.d. —60 | 10 | 68.4 | — | 100—270 | — |
| V | 6—20 | 100—300 | 50 | 19 | 100 | — | — |
| Co | n.d. | n.d. | 20 | n.d. —18 | 10—30 | — | 17—22 |
| Cu | 60—100 | 20—40 | 20 | 20 | 1000 | 2396 | 11—60 |
| Ni | n.d. | 10 | 50 | 18.6 | 8—30 | — | 5—22 |
| Zn | 600—8000 | 200—500 | 50 | 22.1 | — | 5624 | n.d |
| Mo | n.d. | n.d. | 2 | 5.9 | — | — | — |
| Pb | 60 | n.d. | 20 | 20.9 | 0—100 | 80 | — |
| Cr | n.d.—200 | 20—100 | 80 | n.d. —8 | 0—10 | — | — |
| As | 200—400 | n.d. | — | — | 30 | — | — |
| Ge | 10—20 | n.d. | 8 | — | — | — | — |

*1* Hematite-goethite ores, El-Gedida (MAHGOUB and AMER, 1964; KAMEL, 1971), *2* pisolitic limonite ore, Gebel Ghorabi (NAKHLA and SHEHATA, 1967; KAMEL, 1971), *3*** average of different Fe-deposits (SHERIF, 1969), *4* Mn-rich ores, average of 15 analyses (NIAZY, 1975), *5* average of Aswan sedimentary Fe-ores (GHEITH, 1955), *6* amorphous goethite facies, Red Sea brine deposist (BISCHOFF, 1969), *7* Fe-rich deposits, Afar Rift (BONATTI et al., 1972a). *KAMEL (1971)

Therefore, the present authors believe that Mn was introduced into the Bahariya iron ores by sources related to volcanism, most probably by solution of low-temperature. The determined values of Mn content are distinctly higher than the values given for the oolitic limonite ironstone, hematite rich sedimentary rocks and even the chamositic ironstones (JAMES, 1966). Similarly, the values of $Al_2O_3$ Mg and P are quite lower than the typical sedimentary Fe-ores. In addition, the hematite—goethite ores are marked by higher values of Ba, $Cl^{-1}$ and $SO_4^{-2}$. It seems quite possible to correlate the association of these elements to certain waters of volcanic activity (WHITE et al., 1963).

The contents of trace elements are generally low, much lower than the Aswan Fe-ores of shallow marine environment, which are characterized by the presence of Cu, V and As. The Bahariya pisolitic—oolitic ores are characterized by the presence of V and possibly Ni similar to the minette sedimentary ores of Lorraine (JAMES, 1966). The hematite—goethite ores, however, are similar to the goethite facies of Red Sea brine deposits which are characterized by noticeable concentration of Ba, Zn, As, Ge and possibly Cu, Pb and Sr. GERMANN (1973) mentioned that a Mn/Co ratio greater than 300 is indicative of volcanogenic origin which is the case with the "Mn rich" ores. Therefore, it is suggested that these elements have been concentrated during the development of the hematite—goethite ores by the action of epigenetic solutions similar in composition to the Red Sea brines, after the deposition of the pisolitic—oolitic hydrogoethite bed. In addition, As is a characteristic trace element for volcanogenic sedimentary Fe-oxides of Mexico (ZANTOP, 1981).

**Manganese ores of Elba**

The Mn-ores are located in the southeastern Desert (Fig. 1), in the form of veins. More than twenty-four occurrences are known, distributed in a narrow belt trending for more than 70 km. They occur within the Red Sea coastal plain which consists mainly of Quaternary (sands, gravels and coral limestone) and Miocene (lime-grits and evaporites) sediments. These rocks unconformably overlie the Precambrian Basement Complex which are well exposed in the western portion of the area. The Mn-veins, in most localities, cut through the Middle Miocene rocks. They are variable in length, ranging from 300 to 1500 m with a maximum observed downward extension of 10 m.

NIAZY (1975) mentioned that the ores consists essentially of pyrolusite ($\beta$-$MnO_2$), psilomelane—cryptomelane ($\alpha$-$MnO_2$), ramsdellite ($\gamma$-$MnO_2$), manganite (disordered form of $\alpha$-$MnO_2 \cdot H_2O$), goethite ($\alpha$-$Fe_2O_3 \cdot H_2O$) and hematite ($\alpha$-$Fe_2O_3$). The main gangue minerals are black calcite (Mn-bearing), quartz, barite and celestite.

*Geochemistry*. The average chemical analyses and trace elements of Elba Mn-veins after BASTA and SALEEB (1971), NIAZY (1975) together with some data for other Mn-deposits are given in Table 2. The Fe/Mn ratio varies from 0.008 to 0.019, a fact which makes a sedimentary origin of such veins a far possibility. Normal sedimentary Fe/Mn mineralization have Fe/Mn ratio close to 1 (BONATTI et al., 1972). Extreme separation of Fe from Mn with Fe/Mn ratio between 0.002—0.14 characterize encrustations of hydrothermal origin from Fe—Mn modules of hydrogenous origin (RONA, 1978).

The abundance of transitional elements Ni, Co, Cu, Cr, Zn etc are low. Similar concentration of these elements and such Fe/Mn ratio were also recorded from the hydrothermal Mn rich deposits of the Afar Rift (BONATTI et al., 1972a). Such low values are in contrast to the deep marine sedimentary Mn nodules where the concentration of these elements is generally high (MANHEIM, 1965). The manganite facies deposited from the Red Sea brines (BISCHOFF, 1969) are also characterized by the presence of Zn, Pb, Cu, Ba and Sr. These elements are present in the Mn-veins of Elba but with lower concentration except for Ba. Many authors (BONATTI et al., 1972a; RONA , 1978) have pointed out the role of rapid deposition in hydrothermal solutions, which most probably prevents any extensive "scavening". Therefore, low concentrations of trace metals are encountered in the hydrothermal deposits.

**Chemical composition of manganese veins of Elba**

*Table 2*

| Element | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| Major elements in weight percent | | | | | | | |
| $SiO_2$ | 1.99 | 2.55 | 6.61 | 16.20 | 7.50 | 0.430—8.79 | 15.40 |
| $Al_2O_3$ | 0.27 | 0.61 | 0.41 | 4.60 | 0.70 | 0.0—2.27 | 1.00 |
| Fe | 0.38 | 0.93 | 2.45 | 11.68 | 21.33 | 0.30—1.64 | 0.30 |
| Mn | 49.11 | 48.96 | 6.39 | 15.98 | 25.58 | 31—54 | 38.10 |
| Ca | 4.09 | 5.88 | 22.79 | 1.57 | 2.07 | — | 5.80 |
| Mg | 0.09 | 0.21 | 0.66 | 1.39 | 0.93 | — | 0.70 |
| Ba | 3.45 | 2.49 | 5.82 | 0.15 | 0.036 | 0.013—6.25 | 0.81 |
| S | — | — | — | — | 0.60 | — | — |
| Fe/Mn | 0.008 | 0.019 | 0.38 | 0.62 | 0.834 | 0.001—0.03 | 0.008 |
| Trace elements, in ppm. | | | | | | | |
| P | n.d.—301 | 323 | n.d. | 1921 | 1746 | — | 10 |
| Ti | n.d.—112 | 180 | n.d. | 5640 | 300 | — | 500 |
| Co | n.d. | 6.4 | 2 | 2800 | — | 5—17 | 10 |
| Cu | 300 | 24 | 3 | 4000 | 799 | 5—25 | 1900 |
| Ni | trace—100 | 10.6 | 2 | 5800 | — | 5—20 | 5 |
| Zn | n.d.—300 | 43 | 5 | 400—4000 | 11248 | — | 50 |
| Mo | n.d.—100 | 5.6 | 2 | 385 | — | — | 1700 |
| Pb | 1000 | 65 | 10 | 1000 | 928 | — | 500 |
| Sr | n.d.—1070 | 98 | 1100 | 600 | 211 | — | 0.50—1500 |
| V | n.d.—2000 | 69 | 5 | 440 | — | — | 800 |
| Cr | n.d. | 5.4 | 2 | 10 | — | — | 10 |
| As | n.d. | — | — | — | — | — | 2500 |

*1* Mn-veins (Basta and Saleeb, 1971), *2* Mn-veins (Niazy, 1975), *3* black calcite (ibid.), *4* deep ocean Mn-concretions (Manheim, 1965), *5* Mn-facies, Red Sea brine deposits (Bischoff, 1969), *6* Mn-rich deposits, Afar Rift, (Bonatti et al., 1972a), *7* volcanogenic Mn-oxides, Mexico (Zantop, 1981)

The concentration of Ba in sea-waters is such that they are undersaturated with respect to $BaSO_4$ (Hanor, 1966). Ba is found to be concentrated in Fe—Mn deposits along oceanic ridges, where it is probably supplied by hydrothermal activity. Ba tends to concentrate with Mn rather than with Fe. The relation between Ba and Mn in the Elba deposits, Afar region and in Mn-nodules is illustrated on Fig. 3. It shows the difference between the fields of hydrothermal and sedimentary Mn-deposits. Bonatti et al. (1972a), concluded that the barite mineralization associated with Fe—Mn deposits of Afar indicates an association with active ocean rift. In other words, the

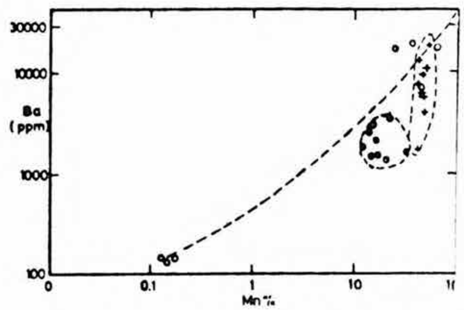


*Fig. 3.* Variation diagram of *Ba* versus *Mn* in Elba *Mn*-veins
+ = Elba: after BASTA and SALEEB (1971), and present work, ● = Mn nodules: after CRONAN (1972), ○ = Afar: after BONATTI et al. (1972b)

source of hydrothermal solutions that led to the formation of Elba Mn—Fe—Ba deposits is most probably linked with the Red Sea Rift activity.

The circulating hot waters are able to perform high-temperature leacing and mass transfer of transitional elements (including Fe, Mn) from sediments and basalts which leads to the formation of hydrothermal solutions. RONA (1978) also mentioned the possible addition of Ba, transported in volatile phases from the mantle and injected through the hydrothermal system.

### Manganese—iron ores of Um Bogma

The Mn—Fe ores of Um Bogma area, lie in west-central Sinai (Fig. 1). The geologic section of the area consists mainly of Carboniferous rocks (sandstones, siltstones, shales and dolomite), overlying igneous and metamorphic basement of Precambrian age. The ores form lenses in the dolomite member which vary from 1 to 220 m in diameter, and from 0.2 to 6 m in thickness.

The stratigraphic sequence of west-central Sinai includes two volcanic episodes

*1* Permo—Triassic episode comparising basaltic sills at the top of the Palaeozoic section,

*2* Post-Miocene episode which produced dolerite and basalt dykes.

The ores consist of the following minerals in a decreasing order of abundance; manganite, psilomelane, goethite, quartz, pyrolusite, cryptomelane, hematite, calcite and barite (NIAZY, 1975).

Geochemistry: The average chemical analyses and trace elements of the Um Bogma Mn—Fe ores after GILL and FORD (1956), NAKHLA and SHEHATA (1963), NIAZY (1975), together with the data of other deposits of different geological environments are given in Table 3. The analyses show that the Fe/Mn ratio varies markedly in the ores (1.68—2.33), it is highest in the Fe rich samples which occur at the peripheries of the lenses. The silica is mainly attributed to free quartz and/or clay particles Fe rich varieties contain a higher amount of quartz than the Mn rich varieties which are admixed with clay. It is clear, that 1% $P_2O_5$ can be considered as common, especially when shallow marine concretions are associated with semi-stagnent or stagnent sediments.

**Chemical composition of Um Bogma manganese deposits**

*Table 3*

| Element | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| | | | Major elements, in weight percent | | | | |
| $SiO_2$ | 5.80 | 4.71 | 3.61 | 6.36 | 16.40 | 7.50 | 16.20 |
| $Al_2O_3$ | 2.38 | 2.05 | 1.72 | 2.21 | 2.90 | 0.70 | 4.60 |
| Fe | 36.03 | 41.38 | 13.55 | 29.73 | 22.45 | 21.33 | 11.68 |
| Mn | 21.46 | 20.15 | 44.04 | 22.77 | 14.02 | 25.58 | 18.98 |
| Ca | 0.56 | 0.63 | 1.07 | 2.61 | 1.21 | 2.07 | 1.57 |
| Mg | 0.06 | 0.22 | 0.43 | 0.58 | 0.57 | 0.93 | 1.39 |
| Ba | 0.89 | minor | 0.33 | 0.19 | 0.25 | 0.36 | 0.15 |
| S | 0.15 | — | — | — | 0.08 | 0.60 | — |
| Fe/Mn | 1.68 | 2.33 | 0.31 | 1.31 | 1.60 | 0.834 | 0.62 |
| | | | Trace elements, in ppm. | | | | |
| P | 1048 | 573 | — | — | 6986 | 11746 | 1921 |
| Ti | — | — | — | — | 1320 | 300 | 5640 |
| Co | trace | — | 140 | 305 | 160 | — | 2800 |
| Cu | 399 | trace-minor | 550 | 410 | 48 | 799 | 4000 |
| Ni | trace | faint trace | 390 | 85 | 750 | — | 5800 |
| Zn | trace | Probable trace | 155 | 85 | 80 | 11248 | 400—4000 |
| Mo | — | — | 115 | 85 | 130 | — | 380 |
| Pb | 650 | trace | 395 | 270 | 38 | 928 | 1000 |
| Sr | — | trace | 96 | 115 | — | 211 | 600 |
| V | — | faint trace | 115 | 200 | 150 | — | 400 |
| Cr | — | faint trace | 7 | 7 | 10 | — | 10 |
| As | 100 | faint trace | — | — | — | — | — |

*1* Average shipping ore (GILL and FORD, 1956), *2* average of different deposits (NAKHLA and SHEHATA, 1963), *3* average of Mn-rich deposits (NIAZY, 1975), *4* average of Fe-rich deposits (ibid.), *5* Mn—Fe concretinos (MANHEIM, 1965), *6* manganite facies, Red Sea brine deposits (BISCHOFF, 1969), *7* deep ocean Mn-concretinos (MANHEIM, 1965)

The relative large amount of Ba may come partly from the stagnent sediments around which the concretions form, as in the case of Gotland region, Baltic Sea (MANHEIM, 1965). However, the large part of the sulphate had persisted reduction. The amount of trace elements encountered is rather comparable to the average given for shallow-water Fe—Mn concentrations, but is much lower than that in Mn nodules of the deep-ocean. Tha amount of transitional elements is somewhat higher than that normally encountered in hydrothermal deposits (N $\sim$ 10, Co $\sim$ 15, Cu $\sim$ 100 ppm, BONATTI et al., 1972b). Most probably, this indicates that the trace elements found in Um Bogma ores were not only syngenetic with the ore but other amounts particularly of Cu, Pb and Sr were added later during an epigenetic or diagenetic process.

Ni, Cu, Zn and Pb are more enriched in the Mn rich samples, while Co and V are more enriched in Fe-bearing samples. BURNS and FUERSTENAN (1966). mentioned that there is a distinct correlation between Fe, Co and Ca in the Mn nodules (Fig. 4).

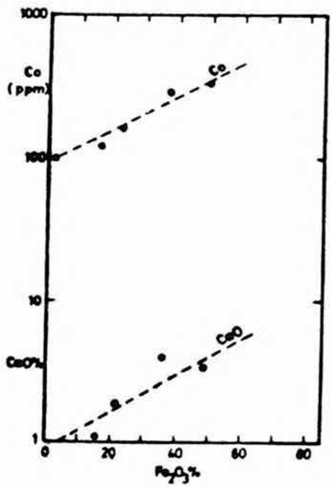


*Fig. 4.*
Variation diagram of *Co* and *CaO* versus $Fe_2O_3$% in Um Bogna Mn-Fe ores

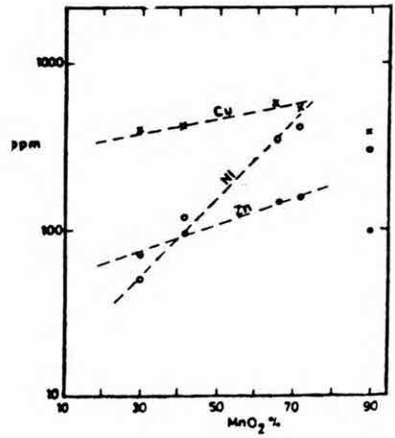


*Fig. 5.*
Variation diagram of *Zn, Cu* and *Ni* versus *Mn,O*% in Um Bogma *Mn—Fe* ores

Similarely, there is a pronounced element coherence between Ni, Cu, Zn and Mn (Fig. 5, notice that one sample is excluded as it is "pure" pyrolusite). These interelement relationships suggest that isomorphic substitution may play an important role in the distribution of these elements. $Ni^{2+}$, $Cu^{2+}$, $Zn^{2+}$ can substitute for $Mn^{2+}$ in the manganite phase and $Co^{3+}$ for $Fe^{3+}$ in the hydrated Fe-oxide phase.

The Um Bogma Mn—Fe ores are characterized by sharp differentiation of Mn and Fe that has been started mostly during the early processes of chemical weathering and sedimentation. It is suggested that the authigenic minerals, ore textures and advanced concentration of Mn and other trace elements as Ba, Ni, Cu and Zn in the Mn lenses were formed during diagenesis. In later stage of epigenesis, some of the trace elements as Cu, Pb, Sr and partly Ba were introduced to the ores by the effect of thermal solutions related to the Tertiary volcanism.

## Late Tertiary volcanism

The Late Tertiary volcanism in Egypt is represented mainly by basaltic occurrences in the form of sheets, dykes, plugs, mounds and small ridges. The hydrothermal activities that accompaned the volcanicity seem to have long survived it.

The average chemical composition for major and trace elements of forty basaltic samples collected from ten occurrences (Fig. 1) are given in Table 4. The world averages of basalts as quoted from different authors are cited in the same table for the sake of comparison.

The distribution of transitional and some other elements in the different basaltic rocks of Egypt show relative concentration above their abundance figures of the world averages. The elemental composition of the different three basaltic varieties (ABDEL AAL, 1981) can be briefly outlined as follows:

**Chemical composition of tertiary basaltic rocks of Egypt**

*Table 4*

| Element \ Rock type | Tholeiitic basalt | Sodic transitional basalt | Alkali basalts | | | World averages | |
|---|---|---|---|---|---|---|---|
| | | | Hawaiite | Sodic basalt | Potassic basalt | Tholeiitic[1] basalt | Alkali[2] basalt |
| **Major oxides, in weight percent** | | | | | | | |
| $SiO_2$ | 48.27 | 46.86 | 46.28 | 45.70 | 45.70 | 50.83 | 45.73 |
| $TiO_2$ | 2.60 | 2.55 | 2.62 | 2.45 | 3.52 | 2.03 | 2.63 |
| $Al_2O_3$ | 14.91 | 15.17 | 15.07 | 13.79 | 16.22 | 14.07 | 14.64 |
| $Fe_2O_3$ | 4.41 | 5.29 | 4.19 | 4.63 | 4.87 | 2.88 | 3.16 |
| FeO | 8.61 | 8.57 | 8.49 | 8.07 | 9.51 | 9.00 | 8.73 |
| Total Fe | 9.77 | 10.36 | 9.53 | 9.51 | 10.80 | 9.00 | 8.99 |
| MnO | 0.24 | 0.22 | 0.19 | 0.19 | 0.18 | 0.18 | 0.20 |
| MgO | 5.17 | 6.10 | 6.76 | 8.79 | 5.58 | 6.34 | 9.39 |
| CaO | 9.90 | 9.89 | 9.07 | 9.28 | 7.77 | 10.42 | 10.74 |
| $Na_2O$ | 2.56 | 2.83 | 4.29 | 3.26 | 3.25 | 2.23 | 2.63 |
| $K_2O$ | 0.79 | 0.71 | 1.60 | 1.26 | 1.93 | 0.82 | 0.95 |
| $P_2O_5$ | 0.35 | 0.33 | 0.42 | 0.46 | 0.41 | 0.23 | 0.39 |
| $H_2O^+$ | 1.07 | 1.08 | 1.02 | 1.47 | 1.32 | 0.91 | 0.76 |
| $H_2O^-$ | 0.67 | 0.57 | 0.39 | 0.64 | 0.87 | — | — |
| **Trace elements, in ppm.** | | | | | | | |
| Mn | 1865 | 1700 | 1481 | 1439 | 1437 | 1163 | 1550 |
| P | 1573 | 1536 | 1903 | 2088 | 1895 | 1100 | 1400 |
| Cr | 118 | 137 | 271 | 354 | 297 | 170 | 200 |
| Ti | 15600 | 15300 | 15720 | 14700 | 21120 | 13800 | 9000 |
| V | 273 | 277 | 201 | 206 | 250 | 250 | 200 |
| Co | 39 | 41 | 70 | 74 | 50 | 48 | 45 |
| Ni | 70 | 77 | 187 | 200 | 150 | 130 | 160 |
| Cu | 86 | 79 | 47 | 53 | 41 | 87 | 100 |
| Zn | 121 | 125 | 88 | 82 | 95 | 105 | 130 |
| Pb | 0.7 | 3 | 6 | 1 | — | 6 | 8 |
| Mo | 1.8 | 1.9 | 4 | 3.4 | 4.5 | 1.5 | 1.4 |
| Ba | 323 | 331 | 437 | 410 | 613 | 330 | 300 |
| Sr | 368 | 410 | 550 | 500 | 713 | 465 | 440 |
| S | — | 1000 | — | — | — | 300 | 300 |

*1* NOCKHOLDS (1954), and TUREKIAN and WEDEPOHL (1961) *2* NOCKHOLDS (1954), and VINOGRADOV (1962)

*1* Tholeiitic basalts show relative high concentration of Fe, Mn, Ti and V. This variety mainly occur along Cairo—Suez district, and in Gebel Qatrani.

*2* Sodic transitional basalts with relatively high concentrations of Fe, Mn, Ti, V, possibly S. These rocks are exposed along Cairo—Suez district, south El-Quseir and Bahnasa.

*3* Alkali basalts, having relatively high concentrations of P, Cr, Ti, Co, Ni, Mo, Ba and Sr and are differentiated into:

*a)* Hawaiites, represented by Mandisha, Maesera and Basalt Hill in the Bahariya Oasis and north Uweinate.

*b)* Sodic basalt, represented by El-Hefuf in the Bahariya Oasis and Naqb Siwa.

*c)* Potassic basalts, represented by Gebel Siri and Gebel El-Fantas.

The effect of the Tertiary volcanic thermal waters have been recorded in several regions. The occurrences of ferruginous tubes in the Oligocene sandstones and quartzites of Gebel Ahmer near Cairo, are attributed to the uprising fluids carrying Fe and Mn (SHUKRI, 1953). KOLBE (1957), reported that the thermal springs of Bahariya Oasis contain up to 69% Fe and Al-oxides. SOLIMAN (1961) described the occurrence of a dolerite dyke in Wadi Feiran, southern Sinai, where the Lower Eocene limestones are silicified and coloured black due to introduction of Mn-oxides.

**Partial chemical analysis of altered tertiary basaltic rocks of Egypt**

ATTIA, 1979

*Table 5*

| | Bahariya Oasis | | | | Cairo—Suez District | | | | |
|---|---|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 3 | 1 | 2 | 3 | 3 | 3 |
| $Fe_2O_3$ | 3.13 | 3.85 | 5.90 | 6.58 | 4.45 | 5.30 | 6.32 | 7.90 | 7.77 |
| FeO | 7.39 | 6.36 | 3.41 | 1.95 | 8.97 | 7.80 | 6.48 | 4.87 | 4.87 |
| MnO | 0.16 | 0.14 | 0.13 | 0.08 | 0.19 | 0.18 | 0.18 | 0.17 | 0.17 |
| Fe | 7.93 | 7.63 | 6.78 | 6.12 | 10.08 | 9.77 | 9.60 | 9.30 | 9.21 |
| Mn | 0.124 | 0.108 | 0.101 | 0.062 | 0.147 | 0.139 | 0.139 | 0.132 | 0.132 |
| Fe/Mn | 63.95 | 70.65 | 67.13 | 98.71 | 68.57 | 70.29 | 69.06 | 70.45 | 69.77 |
| I. L. | 1.41 | 1.87 | 3.43 | 4.39 | 0.91 | 1.10 | 2.10 | 2.68 | 2.83 |
| | South Quseir | | | | | Abu Zaabal | | | |
| | 1 | 2 | 3 | 3 | | 1 | 2 | 3 | 3 |
| $Fe_2O_3$ | 3.47 | 3.59 | 5.68 | 5.90 | | 5.12 | 6.54 | 7.00 | 7.08 |
| FeO | 7.68 | 7.39 | 5.19 | 4.64 | | 7.69 | 5.33 | 4.59 | 4.40 |
| MnO | 0.17 | 0.19 | 0.18 | 0.16 | | 0.16 | 0.14 | 0.12 | 0.14 |
| Fe | 8.40 | 8.25 | 8.00 | 7.74 | | 9.54 | 8.71 | 8.46 | 8.37 |
| Mn | 0.132 | 0.147 | 0.139 | 0.124 | | 0.124 | 0.108 | 0.093 | 0.108 |
| Fe/Mn | 63.64 | 56.12 | 57.55 | 62.42 | | 76.94 | 80.65 | 9.907 | 77.90 |
| I. L. | 1.44 | 1.55 | 2.13 | 3.70 | | 1.69 | 2.41 | 2.64 | 5.71 |
| | Bahnasa | | | | | Qatrani | | | |
| | 1 | 2 | 3 | 3 | | 1 | 2 | 3 | 3 |
| $Fe_2O_3$ | 4.09 | 5.59 | 7.44 | 7.46 | | 4.44 | 5.56 | 5.92 | 6.34 |
| FeO | 9.39 | 7.71 | 6.33 | 4.88 | | 8.22 | 7.08 | 6.36 | 5.10 |
| MnO | 0.16 | 0.19 | 0.19 | 0.17 | | 0.14 | 0.13 | 0.13 | 0.12 |
| Fe | 10.16 | 9.90 | 10.12 | 9.01 | | 9.49 | 9.39 | 9.08 | 8.39 |
| Mn | 0.124 | 0.147 | 0.147 | 0.132 | | 0.108 | 0.101 | 0.101 | 0.093 |
| Fe/Mn | 81.94 | 67.35 | 68.84 | 68.26 | | 93.98 | 92.97 | 89.90 | 90.21 |
| I. L. | 0.56 | 0.92 | 2.03 | 2.75 | | 1.59 | 1.71 | 2.41 | 3.26 |

*1* Fresh basalt, *2* slightly altered basalt, *3* aftered basalt

*Alteration of Tertiary Basalts.* Some basaltic occurrences of Late Tertiary had suffered alteration. In most cases, the alteration is essentially active on their upper parts. However, in Gebel Mandisha, Bahariya Oasis, a reverse situation is observed and the alteration is particularly active on lower parts. The chemical alteration of the basalts is primarily related to their hydrolysis, oxidation and leaching, with development of montmorillonite, chlorite, illite and goethite (ABDEL AAL, 1975).

Chemical analysis for twenty-five samples representing fresh, slightly altered and altered basalts (ATTIA, 1979), are given in Table 5. From the data it is quite evident that Fe and Mn are depleted from the different fresh basaltic occurrences during the process of alteration. A fact which may represent a possible source for these elements in the waters affecting the alteration of basalts. RONA (1978), has also pointed out the role of depleted Fe, Mn and other transitional elements during the formation of hydrothermal mineral deposits in oceanic crusts.

## Conclusions

*1* The tholeiitic and transitional basalts are characterized by relative concentration of some trasitional and other elements above their abundance figures: Fe, Mn, Ti, V possibly and S. The alkali basalts are relatively enriched with P, Cr, Ti, Co, Ni, Mo, Ba and Sr except for potassic basalts which show normal distribution of Co and Ni.

*2* The chemical alteration of the basaltic rocks, caused their marked depletion in Fe, Mn and some other elements. The alteration resulted in lateral disolution of depleted elements in the active thermal solutions.

*3* The epigenetic Fe—Mn deposits of Bahariya Oasis are characterized by high contents of Fe, Mn and relative concentration of Ba, Zn, As, Ge, low Al-content, and high Fe/Mn ratios (27.23—51.02), similar to metalliferous sediments of predominant hydrothermal origin. The concentration of Mn, Ba, Cl, $SO_4^{-2}$ might be correlated with thermal waters of the volcanic activity.

*4* The Mn-veins of Elba display extreme chemical fractionation of Fe from Mn, while the Fe/Mn ratios are between 0.008 and 0.019. They are also characterized by their low trace-metal content, and distinct correlation between Ba and Mn, and between Ba and Fe/Mn ratio, indicating their similarity with the hydrothermal Mn-deposits of Afar, Red Sea region.

*5* The Mn—Fe ores of Um Bogma, exhibit Fe/Mn ratios of the order of 1.68—2.33, with relative concentration of the trace-metal content (Ni, Co, Cu, Zn), characteristic for hydrogenous sediments. However, the deposits display diagenetic fractionation of Fe from Mn. It is suggested that the replacement phenomena above the Mn-horizon, as well as the leaching effects and recrystallization have took place as a result of the epigenetic modification of the ores. Thermal waters along faults, have probably introduced additional amounts of Cu, Pb, Sr, parlty and Ba.

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# INFLUENCE OF NEOGENE GEOLOGICAL EVENTS ON THE ORIGIN, ACCUMULATION AND DISTRIBUTION OF OIL AND GAS IN THE PO BASIN

by

L. MATTAVELLI

*Introduction.* The Po Basin was characterized by a complex geological evolution, due to its particular location in the regional plate tectonic context of the Mediterranean realm (northwestern edge of the African plate or an African microplate). The sedimentary sequence in this basin of thick, terrigenous Tertiary and Quaternary deposits overlying the mainly carbonate Mesozoic sediments (Fig. 2). Two main sedimentary cycles characterize the Neogene deposits: the first cycle, from the Oligocene to the Messinian, and the second cycle from the Upper Messinian to the Quaternary. The present geological setting of the Po Basin was brought about by the tectonic and sedimentologic events of the Neogene. In short, the Po Basin served as a foreland to both the northern Apennines and the southern Alps characterized by opposite vergences towards the north and the south, respectively.

Though hydrocarbons exploration began in the Emilian Apennines during the early 19th Century, only after the World War II the intense use of modern seismic techniques led to the discovery of 85 gas fields and 10 oil and condensate gas fields in the Po Plain and in northern Adriatic (Fig. 1). Most of the reservoir rocks of the gas

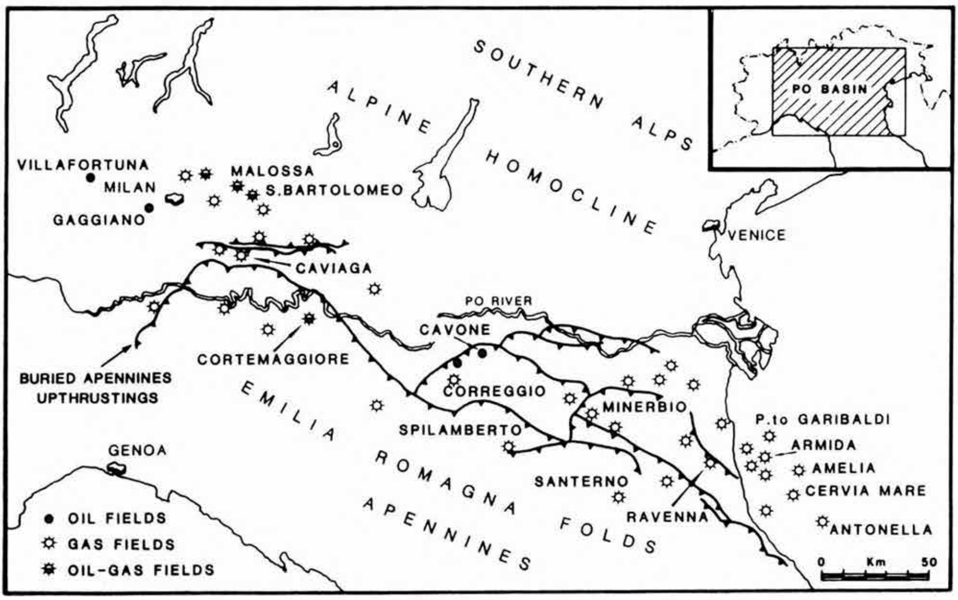



*Fig. 1.* Location of the main oil and gas fields in the Po Basin

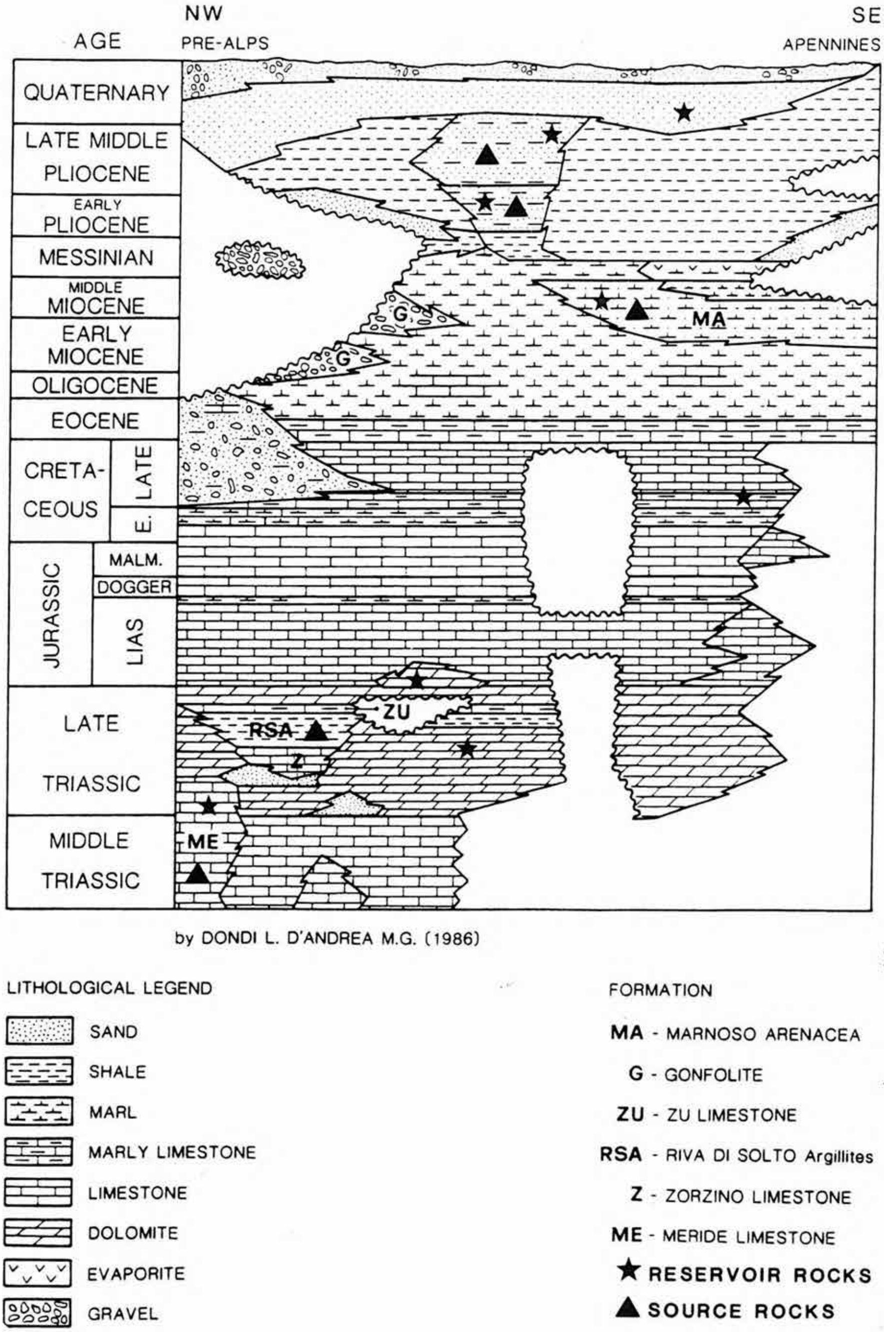



*Fig. 2.* Simplified schematic lithostratigraphic model of the central Po Basin

fields terrigenous Plio—Pleistocene turbidites. Only a few oil and condensate-gas fields have been found in the Mesozoic carbonate rocks (Fig. 2). In order to determine the origin of the hydrocarbons, detailed geochemical analyses (stable isotope, GC, GC—MS, R-, kerogen composition etc) were carried out in the AGIP laboratories on the natural gases, oils and on the organic matter dispersed in the sediments. Moreover, all analytical data were considered in the light of tectonic and sedimentologic events of the Po Basin.

### Gas fields

The Po Basin a classical subalpine gas province natural gas (computed in equivalent tons of oil) represents 95% of the total discovered hydrocarbons. Moreover the data listed in Table 1 indicate that 98% of the natural gases are contained in the Neogene sediments, whereas only a small amount of gaseous hydrocarbons are located in Mesozoic carbonate rocks.

**Percentage of gaseous and liquid hydrocarbons computed in equivalent tons of oil, and their distribution in sequences**

*Table 1*

| TYPE OF HYDROCARBONS | | HYDROCARBONS IN SEDIMENTARY SEQUENCES | | |
|---|---|---|---|---|
| GAS | OIL | AGE | GAS | OIL |
| 95% | 5% | NEOGENE | 98% | 8% |
| | | MESOZOIC | 2% | 92% |

*Analytical results.* Chemical and isotopic analyses have been carried out on the gaseous hydrocarbons from 50 gas fields which represent 90% of the original, "in situ" gas of the basin.

All these analytical data are reported in a previous study (Mattavelli et al., 1983). This paper summarize the main conclusions of the above-mentioned-study, and, in addition, will provide some new considerations which concern the comparison of the Po Basin data with those relevant to the central and southern basins of Italy.

In short, on the basis of the previously mentioned chemical and isotopic analyses, three types of natural gases can be distinguished in the Po Basin (Tab. 2).

*1* Biogenic gases are characterized by almost pure and isotopic, light methane ($C_2+<-2\%$, $\delta 13C_1$ between $-75$ and $-60$ ppt). Their chemical and isotopic compositions are similar to those measured on the interstial gases of bacterial origin in recent marine sediments. Therefore, this fact is considered to be a strong evidence for an indigenous formation of these gases in the Plio—Pleistocene sediments by bacterial or diagenetic processes.

*2* Thermogenic gases are formed by 13C from their contained methane which is heavier than $-50$ ppt, and with their $C_2+$ contents ranging from 3% to 10%. The more positive methane values of 13C ($-31$ to $-36$ ppt) are observed in condensate-

**Types of natural gases in the Po Basin, distinguished on the basis of chemical and isotopic analyses**

*Table 2*

| Genetic type of natural gases | Stable izotopes | | Heavier homologues |
|---|---|---|---|
| | $\delta^{13}$ C | $\delta$ D | $C_{2+}$ |
| 1) BIOGENIC GASES ('B') | —75—60‰ | —180—210‰ | <0.2% |
| 2) MIXED GASES ('M') | —60—50‰ | —180—210‰ | 0.1—5% |
| 3) THERMOGENIC GASES ('T') | —50—30‰ | —150—180‰ | 3—10% |

gas fields found in the Mesozoic carbonate rocks at depths ranging from 4200 to 6 000 m (central western part of the Po Basin).

*3* Mixed gases are characterized by 13C methane values ranging between $-50$ and $-60$, and hy $C_2+$ concentrations ranging from 0,1% to 3%. Accumulations of this types of natural gases originated where thermogenic gases migrated from deeper strata mixed in the same trap with indigenous bacterial gas. Moreover, an estimate of the percentage of the above mentioned three types of gases, with respect to the original gas in place, has given the following surprising results: 80% biogenic or diagenetic gases; 10% thermogenic gases and 10% gases of mixed origin.

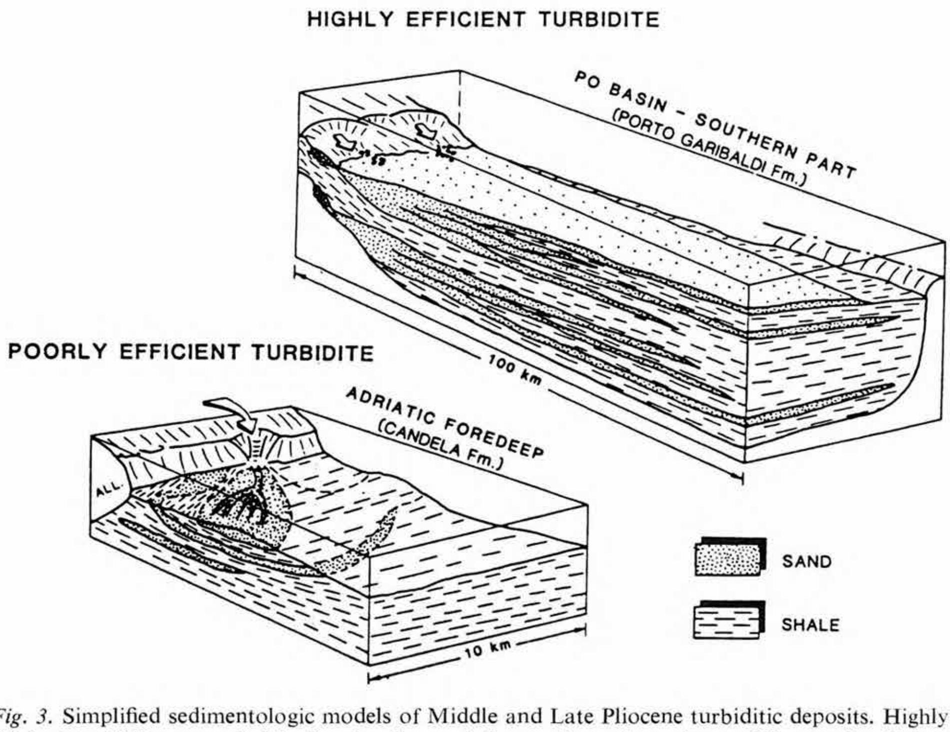



*Fig. 3.* Simplified sedimentologic models of Middle and Late Pliocene turbiditic deposits. Highly efficient turbidites prevailed in the foredeep of the northern Apennines, while poorly efficient turbidites occurred mainly in the foredeep of the central and southern Apennines

*Relationship between the geological setting and genetic types of natural gas.* The sedimentological and tectonic factors during the Neogene clearly exercided controlled the occurrence of genetic types of natural gases. In fact, most of the gas fields are located along the Emilia Romagna and Adriatic fold belt, which represent the subsurface continuation of the Apennines in the Po Basin. These folds were generated during the Miocene, were active until Plio—Pleistocene times and modified the sedimentation processes, thus producing several kinds of structural and stratigraphic traps. In this foredeep domain of the northern Apennines the following three factors favoured the accumulation of the gases that were generated very early in the sediments by bacterial or diagenetic processes:

*1* synsedimentary tectonism;
*2* high rate of subsidence;
*3* turbiditic sedimentation.

In fact, 70% of the bacterial gases produced in the whole basin are found in a small area around Ravenna. They are all bacterially formed gases and mostly occur in multipay zone gas fields. This great concentration of bacterial gases, which is typical of the Ravenna area, has not been found in other foredeep domains of Central and Southern Italy even though the three above-mentioned factors were present. In my opinion, a critical factor was represented by the characteristics of the turbiditic deposits. In fact, highly efficient turbidites prevailed in the Ravenna area, while poorly-efficient turbidites (Mutti, 1984) predominated in the Tertiary basins of central and southern Italy (Fig. 3). In practice, as regards the accumulation of bacterial gas,

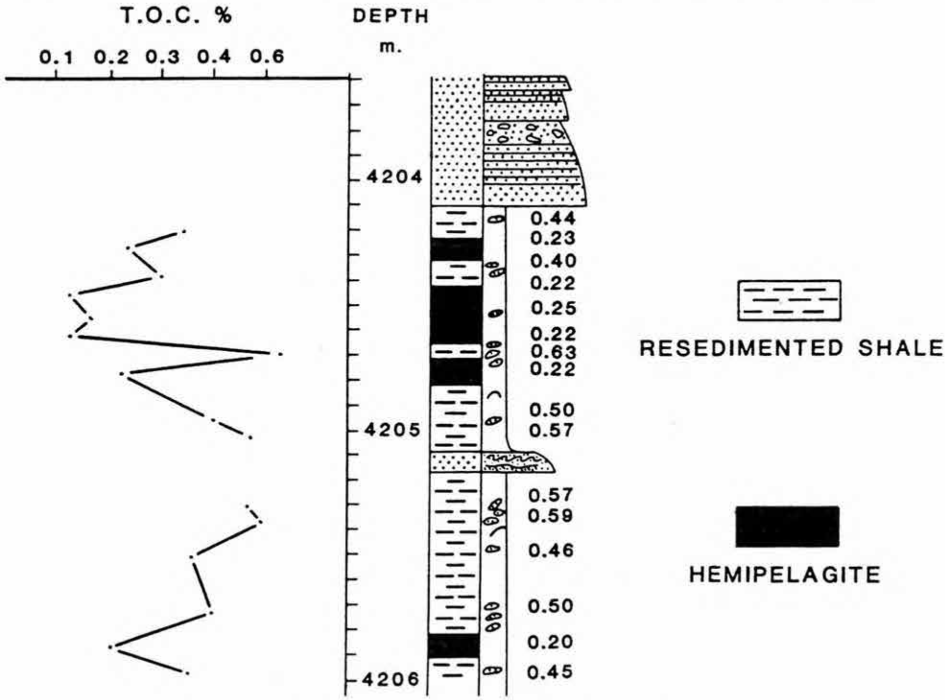



*Fig. 4.* Total organic carbon values in the Late Pliocene turbiditic sediments. Note that the organic matter content of resedimented shales is two to three times greater than that of the hemipelagites

these highly efficient turbidites must be considered ideal types of sediments, because reservoirs and cap rocks are arranged in sandwich mode to form multipay zone gas reservoirs. In addition in the resedimented clay the organic matter was preserved by complete oxidation during rapid burial. This seems to be confirmed by the fact that the amount of organic carbon in the turbiditic shales is 2 to 3 times greater than it is in the pelagic shales (Fig. 4).

## Oil fields

Oil fields were discovered in the western and central part of the Po Basin; that is in the Lombardy area and in the external part of the Emilia folds (Fig. 1). As previously mentioned (Tab. 1), although oil represents, only 5% of the total hydrocarbons found to date in the Po Basin, it is still interesting from both the economical and geological points of view. In order to divide the liquid hydrocarbons into distinct groups and identify their source rocks modern geochemical techniques were used (for more analytical details, see RIVA et al., 1985; in press).

*Analytical results.* On the basis of the stable carbon isotopes two main groups of oil can be identified in the Po Basin: *1* Cortemaggiore area oils, characterized by saturated and aromatic compounds having δ13C values around −23%; *2* Malossa area oil having saturated and aromatic compounds δ13C values of about −30% (Fig. 5). Biological markers also support this subdivision mainly because of the occurrence of oleanane in the Cortemaggiore group (Tab. 3). Furthermore, taking into account the same analytical data relevant to the extracts of the candidate source rocks, it has been proved that the Cortemaggiore oils originated from Miocene flysch (Marnoso Arenacea formation); while the Malossa and Gaggiano oils were generated by the Riva di Solto shales (Late Triassic) and the Meride limestone (Middle Triassic), respectively.

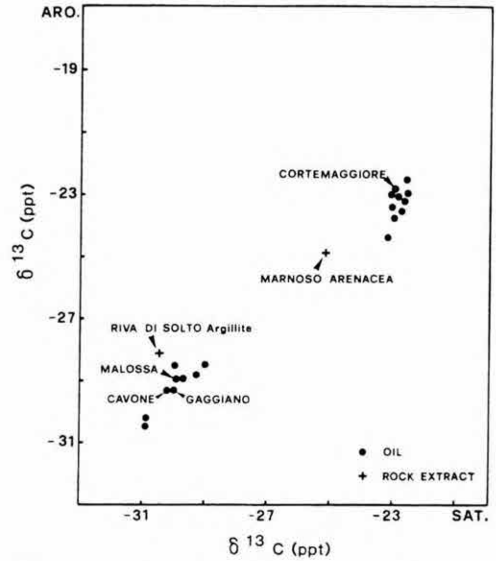



*Fig. 5.* The carbon isotopic values of the saturated and aromatic compounds indicate that the Malossa and Cortemaggiore area oils have a different origin

**The different origin of the Malossa and Cortemaggiore oils is pointed out by several physical, chemical and isotopic parameters**

*Table 3*

| Field | depth | AGE | API | Pr/Ph | $\delta^{13}$ C HCS | Olea-nane / C 30 Hop | 5 α C 28 R / 5 α C 29 R | C 34 SHop / C 33 SHop | Ts / Tm |
|---|---|---|---|---|---|---|---|---|---|
| MALOSSA | 5947 | TRIASSIC | 52 | 1.00 | 29.8 | — | 0.28 | 0.77 | 1.5 |
| CORTE-MAGGIORE | 1934 | MIOCENE | 34 | 2.30 | 22.8 | 0.28 | 1.6 | 0.46 | 0.95 |

In practice the occurrence of the oleanane in Cortemaggiore oil clearly indicates its Tertiary source origin (Marnoso Arenacea Formation)

*Influence of geologic setting on occurrence of oil fields.* The Cortemaggiore and Malossa area oils originated in geological settings characterized by different thermal and burial histories relating to the Neogene. In particular, in the Cortemaggiore field, a structural trap pertaining to the external part of the Apennines overthrusts, oil was found in the sandy layers of the Tortonian and the Upper Messinian. In this field in order to reconstruct the thermal history, the methods developed by ANGEVINE and TURCOTTE (1983) were used; and for the generation of the hydrocarbons, the technologies realized by TISSOT and ESPITALIE (1975) were used. The results of these calculations are illustrated in a depth-converted seismic line going from the Cortemaggiore field towards the Apennines (Fig. 6). The oil generation probably occurred (3 m. y. ago) at a depth of 5500 and 7500 m either in the thrust sheets or in the autochthonous sediments of the Marnoso—Arenacea formation (Miocene). Moreover, the rapid temperature increase caused by the emplacement of the nappes (thermal blanket effect), and the composition of the kerogen (mainly terrestrial), could have favoured the generation of gaseous hydrocarbons. This is confirmed by the fact that the quantity of gas in the Cortemaggiore field (calculated in equivalent tons of oil) is twelve times greater than the original amount of oil. In the western part of the Po Basin (Lombardy area), oil fields and condensate gas fields were discovered in the Mesozoic carbonate sequences at depth of 4200 and 6200 m. In the Gaggiano and Villafortuna fields overpressured hydrocarbons are represented exclusively by oils (34° and 42° API respectively), while in the Malossa, Seregna and San Bartolomeo fields only a small amount of light oil (43 to 53 API) is associated with the gas. In practice the reservoir fluids in these latter fields are monophases with pressure values ranging from 880 to 1067 kg/cm$^2$ (4200 m and 6200 m respectively). Furthermore, on the basis of different maturity parameters (Ro, TAI, Tmax), the organic matter looks immature in many wells down 6000 m, despite the fact that the present temperature are generally higher than 150 °C. This anomalous maturation trend could be caused, in my opinion, by the two following factors:

*1* rapid burial during the Neogene;
*2* abnormal pressures.

The role of abnormal pressure with respect to organic matter maturation is still a subject of debate. Recent laboratory experiments (SAJGÓ et al., 1985) and data relevant to deep wells of the Lombardy area (CHIARAMONTE and NOVELLI, 1985; in

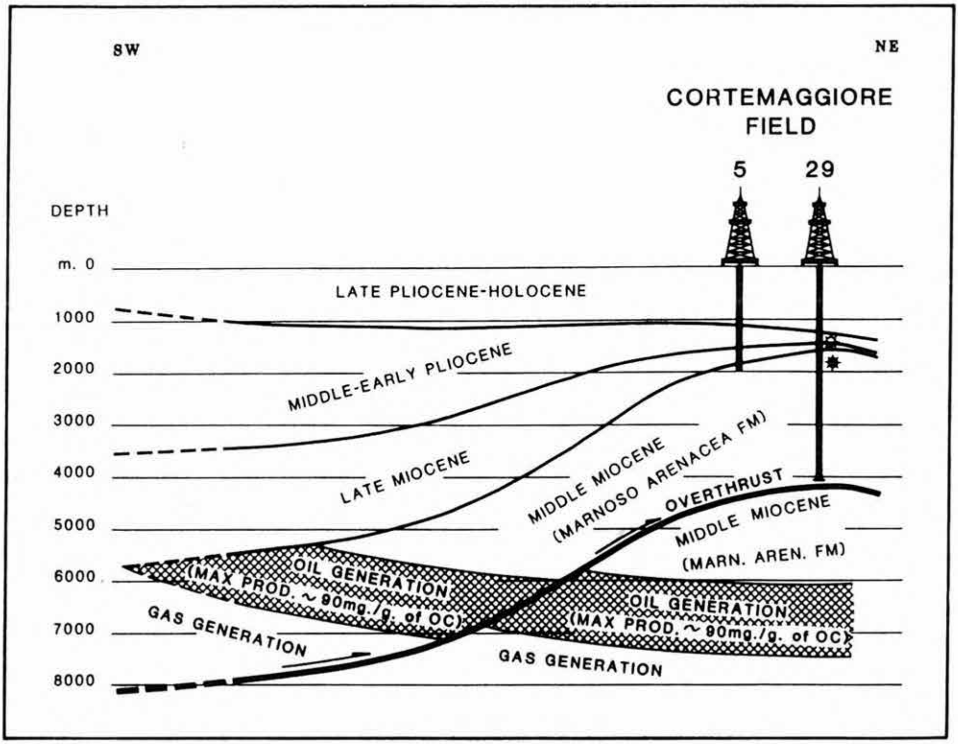
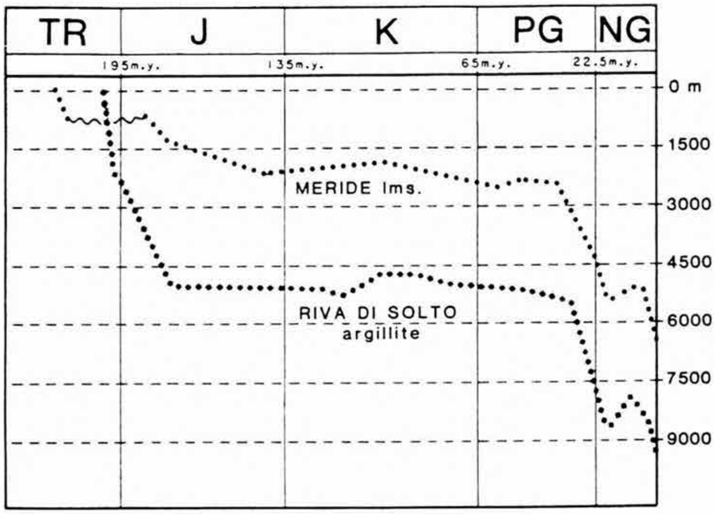



*Fig. 6.* Depth converted seismic line going from the Cortemaggiore field towards the Apennines. Mathematical models were used for determining the thermal history and the amounts of hydrocarbon generated. The hydrocarbons either originated in the thrust sheets or in the autochtonous sediments of the Marnoso Arenacea formation

press) seem to indicate, however, that pressures above 800 atm could have a retarding effect on the maturity of organic matter. Moreover, in order to determine the timing of the hydrocarbon generation, the burial histories of the two main source rocks responsible for the origin of the Malossa condensate gas field and Villafortuna oil field have been reconstructed (Fig. 7). In particular, the burial diagram of the Riva di Solto Argillites could indicate that oil originated during Jurassic, Cretaceous and Palaeogene times. Nevertheless, only the hydrocarbons generated at great depths (over 7000 m) during the Neogene were accumulated, because the traps were formed during the thrusting of the southern Alps; in other words, during the Upper Miocene. Moreover, this assumption seems to be confirmed by isotopic values of the Malossa gas ($\delta 13C_1 - 36$ ppt; $\delta D - 153$ ppt), which suggest highly mature source rocks close to an anthracite coalification level. On the other hand, the Meride limestone remained at a shallow depth until the Palaeocene and the organic matter entered the oil generative window only during the Neogene. The Villafortuna oil field therefore is a significant example of hydrocarbon generation that resulted from recent, rapid burial. In fact, only oil (42 API) was found down to 6200 m in spite of the present temperature of 180°C. In practice, the Villafortuna field represents one of the deepest oil fields in the world.




TR TRIASSIC J JURASSIC K CRETACEOUS PG PALEOGENE NG NEOGENE

*Fig. 7.* Simplified burial histories of two main source rocks responsible for the origin of the Malossa condensate gas (Riva di Solto Argillites) and Villafortuna oil (Meride Limestone). Both these accumulations were generated during the Neogene. Oil had also been produced by Riva di Solto Argillites before the Neogene, but the lack of traps probably prevented its accumulation

## Conclusions

In the Po Basin, the accumulation of gas and oil was greatly influenced by Neogene sedimentological and tectonic events related to the formation of Apennines and the thrusting of the southern Alps. In brief, the main results of this study can be summarized as follow:

*1* In Plio—Pleistocene sequences, owing to the immaturity of the organic matter, only the indigenous formation of biogenic or diagenetic methane took place. Great amounts of this gas, which represent 80% of the original "in situ" gas of the basin, is found in the foredeep of the Apennines; actually in a small area around Ravenna. In this area, three main factors determined the optimum conditions for gas accumulation:

*a* synsedimentary tectonics;
*b* high rate of subsidence (1000 m/m.y.);
*c* highly efficient turbidites.

*2* Miocene sediments reached the oil and gas generative windows only in the external part of the Apennines Nappes (Cortemaggiore area), where a strong increase in temperature was caused by the emplacement of the thrust sheets (thermal blanket effect).

*3* Furthermore, rapid burial and the tectonic movement of the southern Alps which occurred during the Neogene (Malossa area) caused the origin of the deep, condensate gases and the oil fields found in the Mesozoic sequences.

*4* In practice, the Po Basin could serve as a genetic model for other Mediterranean, subalpine, Tertiary basins the gas and oil of which came about as a result of the previously—described processes.

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# NON-METALLIC MINERAL DEPOSITS OF THE TOKAJ MOUNTAINS NEOGENE VOLCANIC AREA

by

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*Introduction:* The Tokaj Mountains are of outstanding importance in the extraction and beneficiation of non-metallic mineral raw materials. This role is similar to that of the Transdanubian Central Range in the bauxite mining (Fig. 1).

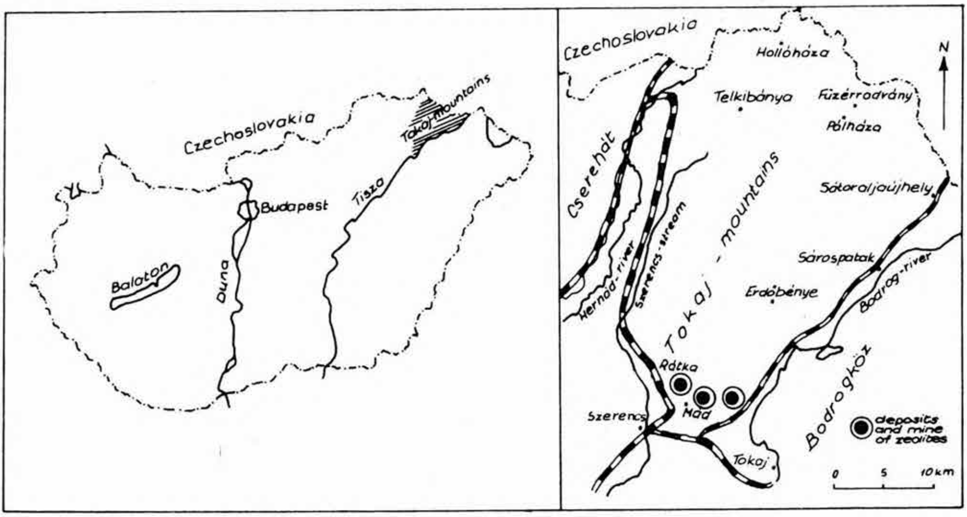



*Fig. 1.* Geographic situation. Map of the region

The 70×35 km area of the mountains, genetically belonging to the inner Carpathian volcanic zone, accomodates 13 mineral types deserving a special rank among the other volcanic aluminium-silicate rocks not for their metal or energy content but their advantageous features due to their mineral composition (Fig. 2).

The exploration, mining and utilization of mineral deposits in the area of the mountains started in the historical past. At present, the raw materials from the mountains are exploited in 13 mines and processed and beneficiated in four dressing plants. The annual production is over 300 thousand tonns. The semi-products or products manufactured from the raw material are delivered to about 1200 users in 45 main directions, to be utilized in remote areas of the country or to cross the border as exported goods (Fig. 3).

Among the mined minerals there are raw materials, like quartzite, used by prehistoric man for his tools which find their application in the industry of today as

| Raw material | | | | Useful mineral componens | Main cation of the useful mineral | Relativ degree of change of the parents rocks | Begining of using or mining | | in Tokaj | |
|---|---|---|---|---|---|---|---|---|---|---|
| Serial number | Denomination | Sign | Useful quality | | | | in the world | in Tokaj | Number of type of raw material | Number of deposits |
| 1. | Quartzite | | hardness, solidity | quartz, opal | Si | | prehistoric age | prehistoric age | 2 | 7 |
| 2. | Kaoline | | whiteness, fire-resisting plasticity | kaolinite | Si Al | | antiquiti | 1569. Sárospatak | 5 | 15 |
| 3. | Potash tuff | | low melting point, whiteness | kaolinite, adulare sanidine | Si Al K | | XX. century in Europe | 1960. Szerencs | 2 | 2 |
| 4. | Colouring earth | | colouring material | hydrohematite limonite | Si Al Fe | | prehistoric age | 1924. Mád | 2 | 2 |
| 5. | Bentonite | | plasticity, ion exchange swelling | montmorillonite | Si Al Fe Ca Mg | | 1925. Fort Benton USA | 1940. Komlóska | 5 | 9 |
| 6. | Illite fine-clay | | whiteness, plasticity melting | illite | Si Al K | | XIX. century in Europe | 1820. Füzérradvány | 2 | 2 |
| 7. | Zeolithe | | whiteness, ion exchange absorption | clinoptilolite | Rock-forming main elements | | 1880. Japan | 1960. Mád | 1 | 2 |
| 8. | Trass | | building material | volcanic glass | | | Antiquity in Roma | 1955. Rátka | 2 | 4 |
| 9. | Pumicit | | building material | pumice | | | 1960. Italy | 1840. Szegi | 1 | 1 |
| 10. | Perlite | | therm, swelling, low density porosity | volcanic glass | | | 1941. USA | 1854. Pálháza | 1 | 2 |
| 11. | Kalitrachite | | k-fertilizer | adulare | | | - | 1880. Hollóháza | 1 | 1 |
| 12. | Diatomite | | absorption, low density porosity | fragments of sil. algae | Si | | Antiquity in Roma | 1937. Erdőbénye | 2 | 4 |
| 13. | Loess | | building material | quartz, koolinite illite | Si Al K | | Antiquity in Babylon | 1569. Sárospatak | 1 | 1 |
| sum total 13 non-metallic raw material | | | quality of acidic volcanic rock or minerals theirs alteration | volcanic or clay minerals | Rock-forming main elements and theirs mobilization | | prehistorical or industrial minerals | | 27 | (52) 47 |

*Fig. 2.* Non-metallic raw materials in the Sarmatian volcano-sedimentary formation on the Tokaj Mountains

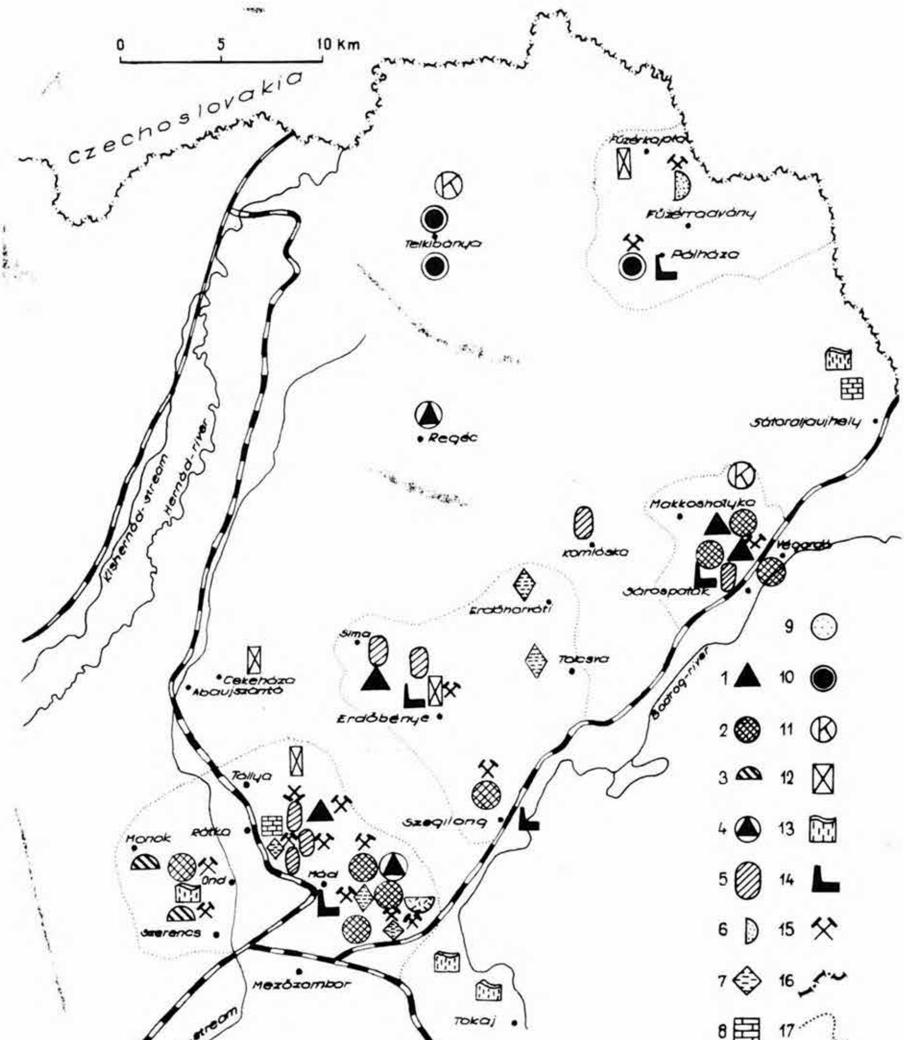



*Fig. 3.* Areal distribution of the mining works, explored mineral deposits and mineral-dressing plants in the Tokaj Mountains

*1* Quartzite, *2* kaolite, *3* potash tuff, *4* bole, coloring earth, *5* bentonite, *6* illitic fine clay, *7* zeolitic rhyolite tuff, *8* trass, *9* pumicite, *10* perlite, *11* kalitrachite, *12* siliceous earth, *13* loess, *14* mineral-dressing plant, *15* operating mine, *16* national boundary, *17* boundary of the area of mineral resources for mining

drum mill lining and millstone in mills eliminating *Fe* contamination. There are also entirely new minerals owing their ranking as raw materials to the technological industrial needs of today. The rhyolitic tuffs with natural zeolite belong to this latter category.

1920 illitic fine clay
1937 siliceous earth
1940 bentonite
1955 perlite
1965 trass
1972 potash tuff
1979 zeolitic rhyolite tuff

The pumiceous rhyolite tuff with relatively high content of natural zeolite (above 35%), referred to as zeolite in the industry, is one of the youngest mineral raw materials in the study area. Attracting the interest of potential users, the explored mineral reserves exceed 100 million t in this area. Further reserves of the same amount and similar quality are registered as prospective. The mineral base for the efforts aimed at the utilization of natural zeolitic minerals, the realization of their value and creation of natural zeolite based products was available as early as 1979.

The utilization of zeolitic minerals is based upon the large-scale distribution of the pure mineral and its useful qualities. The accumulation of relative knowledge and the development of the local zeolite industry based on this knowledge is the result of the research of the past 10 to 15 years.

### Research of natural zeolites

In the 1960's Hungarian specialists, T. MÁNDY, E. NEMECZ, GY. VARJÚ and G. KLOPP, found that the cemented rhyolitic tuffs contained two characteristic zeolitic minerals, *mordenite* and clinoptilolite. Geological investigations indicated that the thickness of the rhyolitic tuffs containing at least 30% of *mordenite* or clinoptilolite amounted to 80 to 100 m. They are locally traced over several square km.

These large homogeneous masses are genetically related to subaquatic volcanic eruptions. They are distributed in flat lense-like accumulations of tuffs around subaquatic eruption centres. The commercial value of the rhyolitic tuffs is determined, besides their amount, by their advantageous characteristics.

These are determined by the structure of the chemical components rather than the composition. Thus the determination of the advantageous features was performed simultaneously with the development of investigations aimed at a determination of the structure of *silicates*, *aluminium silicates*. It was proved that other mineral components of natural zeolite-bearing rocks such as clay minerals and volcanic glass, did not influence negatively the main features of zeolite-bearing rocks such as biogenic gas absorbtion, ionic exchange and activable trace- and rare element content. On the contrary, these qualities did improve.

By the end of the 70s the "new mechanism" had been completely acquired in the industry as well. The agriculture switched over to industrial methods both in plant cultivation and stock-breeding. Town planning increased and the first threatening signs of pollution were registered. The resolution of the actual problems in the industrial plant growing, livestock breeding and environment protection could be the use of rhyolitic tuffs bearing the advantageous qualities of the natural zeolites after a suitable dressing procedure.

**The zeolitic minerals of the Tokaj mountains and their occurrence**

Only two of the 15 zeolitic minerals occurring in Hungary (Fig. 4) are found in the Tokaj Mts. These are clinoptilolite and mordenite. The other zeolite varieties occurring in the pores of basalts or andesites or in the fissures of these rocks and forming decorative crystals are insignificant in quantity. The concentration of zeolite as referred to the total rock mass does not exceed 1% even in the most enriched parts.

In the large-pored pumiceous pyroclasts, rhyolite tuffs of the Tokaj Mts, zeolite crystals are of rather small, microscopic or submicroscopic size. The phenocrysts, visible under microscope or even with the unaided eye formed, in this case too, in the micropores or fissures, but *zeolitization* in the tuff or tuffite material itself is also rather important. As opposed to the crystals of the pores and fissures here, xenomorphic crystals resembling zeolite in their internal lattice structure are characteristic. Zeolite formation in the Tokaj Mts followed the first subaquatic eruption periods of Pacific sacidic volcanism with three intensive phases during the Neogene. The formation of zeolite crystals, as seen from laboratory and factory experience, requires the coincidence of the following parameters in time and space:

*1* Parent rock forming the lattice of the natural zeolitic minerals with the main rockforming elements.

*2* Energy neccessary for the decomposition of the parent rock and zeolite formation.

*3* External pressure required for the formation of the zeolitic lattice.

*4* High humidity and the relative tension, due to water-vapour pressure.

The prerequisites for zeolite formation were provided in the subaquatic acidic volcanic environment of the Tokaj Mountains as follows:

*1 Parent rock:* Pacific or slightly alkaline, acidic volcanic glass.

*2 Energy:* Heat energy from the volcanic eruption and hydration in the sludge belt.

*3 Water-vapour steam tension:* pressure of the water head above the volcanic channel and the internal pressure of the heated system.

These factors proceeded paralelly in time and space three times during the three intensive volcanic phases in the Neogene evolution of the Tokaj Mountains. Therefore, there are three volcanosedimentary horizons, especially rich in natural zeolites in the study area (Fig. 5). The mineable horizons on the surface and near it were formed in the Sarmatian as well as at the Sarmatian—Tortonian boundary. The uppermost, oldest zeolite horizon crops out only in a local spot and only in the NE. Commercial deposits are found mainly in the SW, the Mád—Rátka—Bodrogkeresztúr area, formed during the explosions of the Sarmatian volcanic megarhythm. In this area, detailed exploration discovered five explosive horizons. The 1st of these horizons is at several hundred metres below the surface. It is mainly due to argillization being, tuffitic character.

*The several hundred metre thick pyroclastic rocks of the II^nd^ and III^rd^ horizons in the Rátka—Mád—Bodrogkeresztúr area bear the largest complex of zeolitic rhyolite tuff ever found in Central Europe.* The Rátka area (Fig. 6) is characterized by cemented pyroclastic rocks with 35 to 55% clinoptilolite. The identified, categorized reserves are over 50 million ton.

The III^rd^ explosive horizon in the Bodrogkeresztúr area is characterized by subaquatic explosions. The blanket-like eruption products locally exceed 500 m.

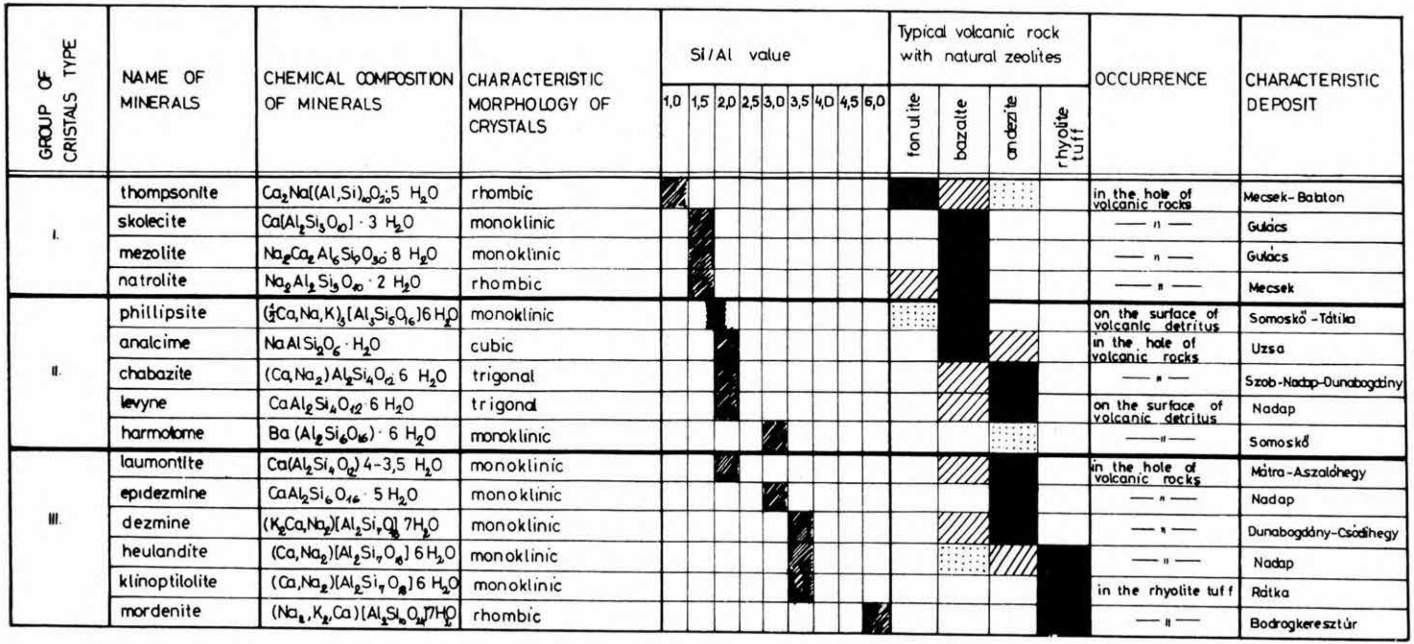


Fig. 4. The natural zeolites of Hungary

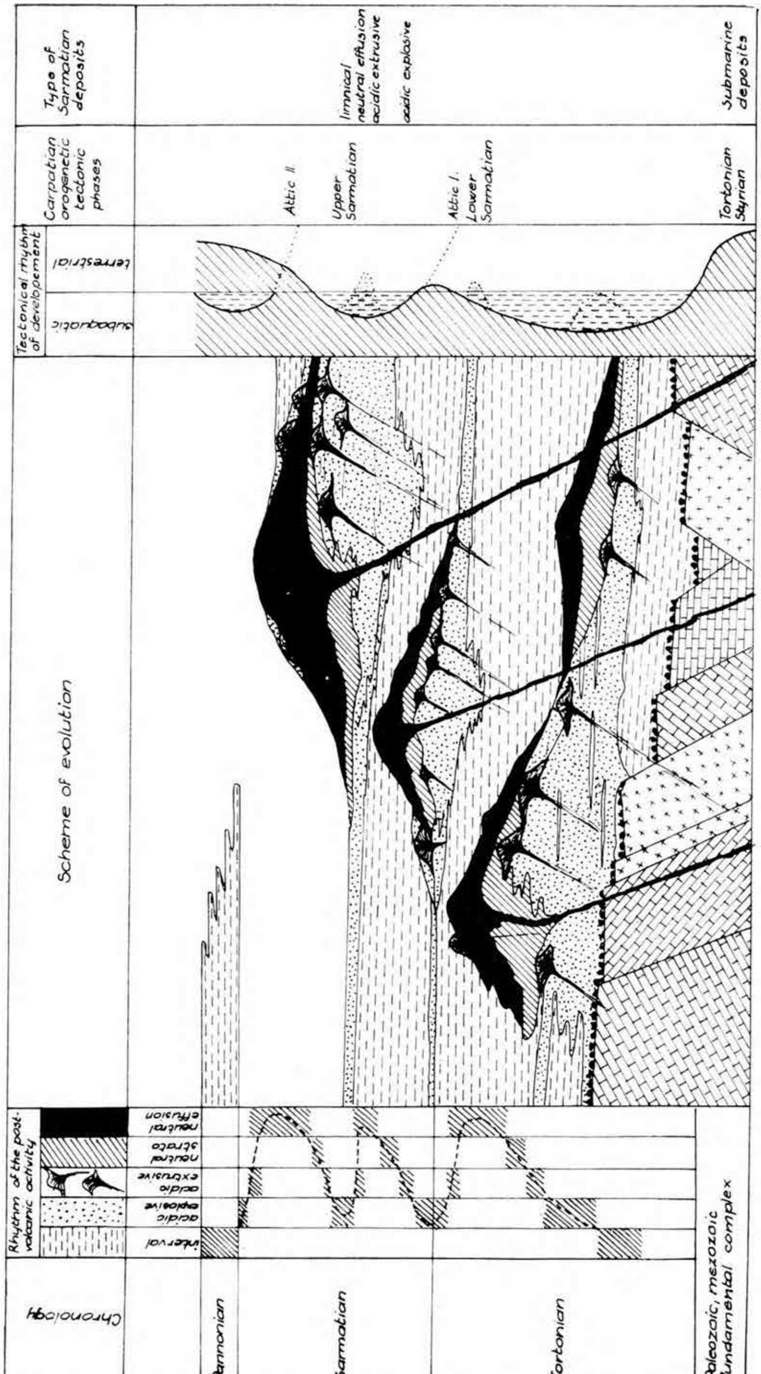



*Fig. 5.* Theoretical evolution scheme of the Neogene volcanic activity in the Tokaj Mountains

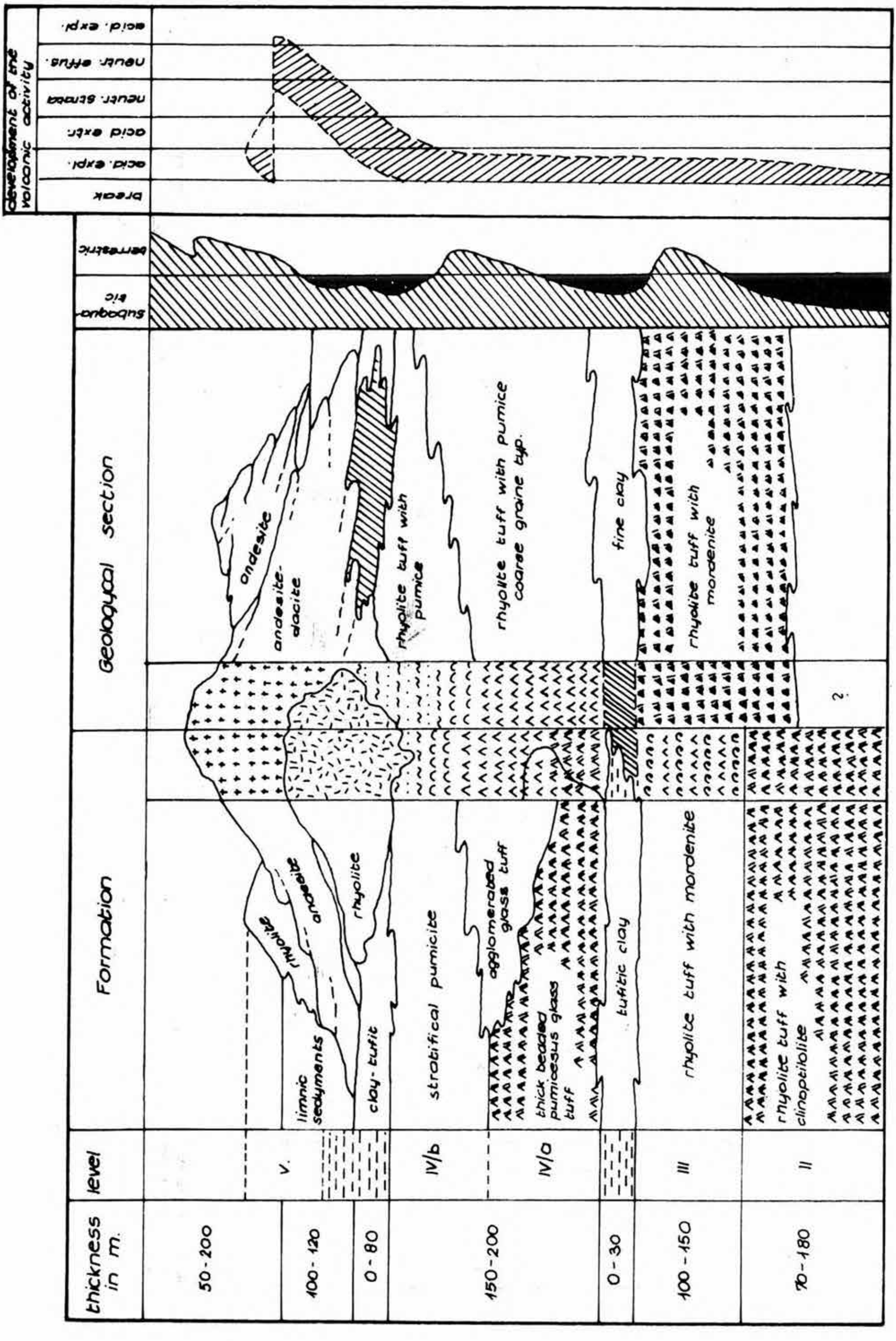



*Fig. 6.* Local geological development of formation

The mordenite content varies between 25 to 70 and 75%. The pyroclastic material with a mordenite content of 60 to 70% is found in separate, 1 to 2 m thick lenses within the tuff complex. *These deposits of pyroclastic rocks have the highest mordenite content in Hungary.*

Stratigraphically, the zeolite-bearing pyroclastic rocks are in the lower part of the Sarmatian volcanosedimentary complex and crop out of the tectonically tilted "erosion windows" in the mountain margin area, providing a possibility for open-pit mining. The valley of the *Szerencs* creek in the Rátka area is such a region of morphological inversion and weathering. The largest mordenite-bearing rhyolite tuff deposit in the Tokaj Mts is 5 km long, having been uplifted and thrust over the mountain margin between Bodrogkeresztúr and Mezőzombor. Therefore the exploration of natural zeolites is concentrated in this part of the mountains. As to the dressing of the zeolites it is favourable that the extraction of other non-metallic minerals is also concentrated in the Mád area. Therefore the special infrastructure of the mining and dressing industry has been available (Fig. 6).

#### Favourable parameters of the natural zeolites from the Tokaj Mountains

One of the good qualities of the zeolite cemented rhyolite tuffs, their relatively high porosity and insulation characteristics, was known to man in the Middle Ages already. Other qualities have been discovered in the last 10 to 15 years. As to the practical use the most important parameters are as follows (Fig. 7):

*1* Biogenic gas adsorption.
*2* Ionic exchange capability.
*3* Trace and rare element content.
*4* Antiparasitic effect.
*5* Coherence with organisms of higher order.
*6* Heavy-metal-trapping effect.

The discovery of the good qualities of zeolites and the emergence of social demand led to the launching of a programme for the utilization of natural zeolites in Hungary. Immediately at the beginning it became evident that the rocks from the deposits with zeolite content are not suitable for odour control, feeding, improvement of soil quality and environment protection without dressing, modification, regardless of their outstanding qualities.

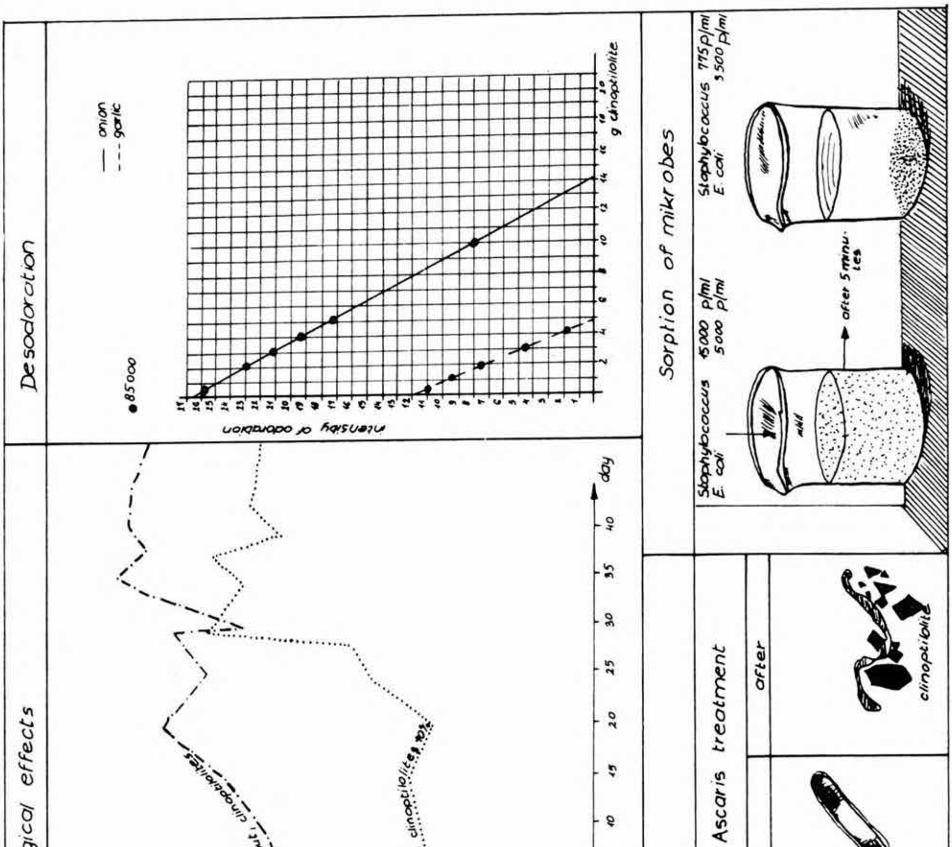
Suitably prepared zeolitic rocks as well as rocks with associated minerals mixed with other non-metallic materials such as siliceous earth, kaoline, bentonite, illite, perlite, etc and dressed will represent products of natural zeolite content, suitable for the elimination of the above-mentioned problems.

It is especially advantageous in the Tokaj Mts that these minerals are mined in zeolite areas (Fig. 3).

These composite mineral products bear the advantageous qualities of both the natural zeolitic minerals and other non-metallic minerals.

*1* A special composite-mineral product, displaying ionic exchange and gas absorbtion qualities with an appropriate trace rare element composition, was developed so as to be used for enhancing the growth of the crops and improvement of soil quality.

*2* When used for feeding and making up for trace- and rare element deficiency, products of different grain size, consistency and mineralogical and chemical composition are required for different animals. Considering these requirements a special



Absorption of NH3
Biological effects
Desodoration
Antiparasitic effects
Sorption of mikrobes
Oocysta (coccidiosis) treatment
Ascaris treatment
befor
during
after
clinoptilolite
pTorr
day
g clinoptilolite
intensity of odoration
concentration of „N" in excrement
without clinoptilolites
with clinoptilolites
onion
garlic
85 000
Staphylococcus 15000 p/ml
E. coli 5000 p/ml
after 5 minutes
Staphylococcus 775 p/ml
E. coli 3500 p/ml


Fig. 7.

product was developed under the name **ZEOVIT—RCL—O**. Given its trace element content with qualities improving the climate of poultry-houses controlling odours and sering as a natural sorbent, it is readily utilized in poultry farming.

*3* As to environment protection, **ZEOIRIX—8—M**, a multi-purpose animal litter is designed to solve the problems concerning animals, kept in blocks of flats (odour control, desinfection, feeding). This product has all the advantageous qualities of the natural zeolites and besides, it has an increased moisture capacity.

**ZEOIRIX—8—O** is a similar product with a moisture capacity of 40 to 50%, suitable for the absorption of hydrocarbons, quick elimination of oil puddles and other contaminations from workshops.

*4* **AKVAROSORB** is another zeolitic product deserving attention. This product has a good effect on the environment of large-scale fish ponds by filtering out ammonia and ammonium ions and absorbing the toxic materials discharged by the fish.

*5* Bacterio-sorbency, desinfection and antiparasitic effect are important with products used for the desinfection of playgrounds, parks and sand-boxes. These products are harmless for higher organisms (children).

*6* We have developed special filter-lining for the treatment of liquid manure. These filters will trap the harmful components so effectively that it will be possible to canal the filtered material right into live waters. Other zeolite-based products help solving minor household problems like odour-control of refrigerators and ash-trays. Other products, such as *Zeolin* paste, *Surolit-ZHS* scrubber are useful in household scrubbing and cleaning. These products are harmless for the environment and facilitate sewage cleaning.

*The utilization of special-quality products made from zeolites is advantageous not only for their immediate effect in the phase of application but their secondary effect as well.*

With the approximately 20 varieties of natural zeolite-based special-quality products developed so far do not deplete our possibilities are not exhausted. Zeolites may play an important role in the future resolution of practical industrial problems owing to their good qualities. Specialists are experimenting with mordenite-rich rocks to produce filters for cleaning gases discharged from nitrosamine factories. Natural zeolite-based products will find application in cleaning industrial gases and sewaging slurry from electroplating factories.

### Production of natural zeolite-based products in Hungary

The development project for the utilization of natural zeolites was launched in 1979 when resources and mineral qualities were studied. Since laboratory and semi-industrial experiments had given positive results, a pilot plant was put in operation. The rapid increase in the rate of production volume is seen in Fig. 8. Sales figures of natural zeolite-based products for the past production period (Table 1):

*Table 1*

| Period of production | Sold product in tons | Equivalents in hundred HUF |
|---|---|---|
| 1978 | 40 | 120 |
| 1979 | 491 | 1,532 |
| 1980 | 601 | 4,425 |
| 1981 | 5,779 | 19,516 |
| 1982 | 8,828 | 32,743 |
| 1983 | 8,902 | 32,168 |
| 1984 | 15,000 | 48,100 |

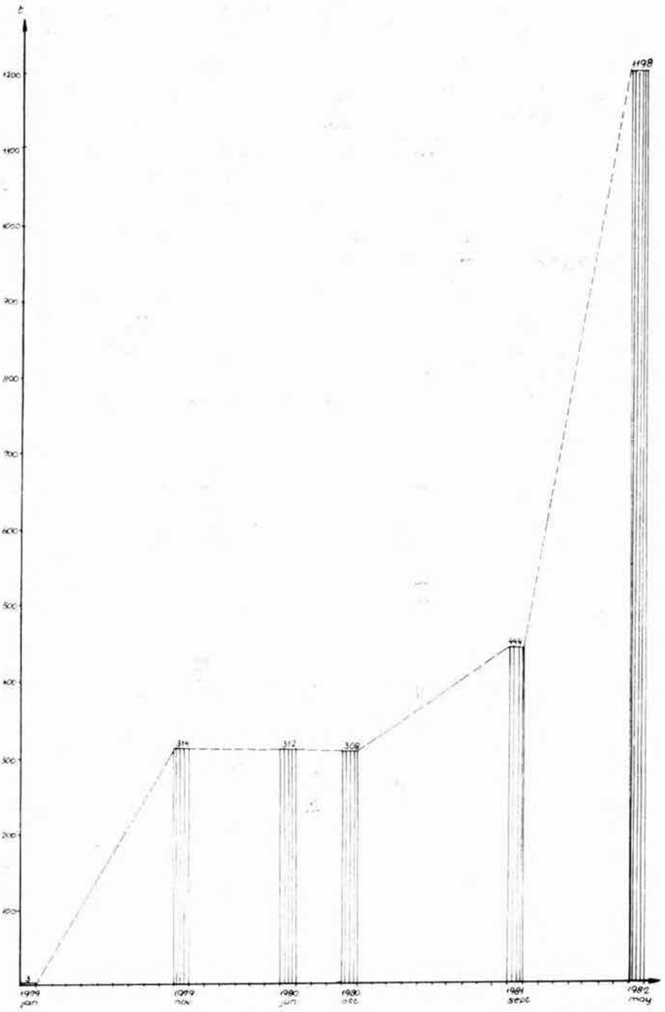



*Fig. 8.* Development of the utilisation of natural zeolites for animal feeding in Hungary

The special quality of the zeolite reserves and natural zeolites of the Tokaj Mts ni Hungary provides a guarantee for their playing an important role in the solution of problems concerning the improvement of soil quality, animal feed production and environment protection both in Hungary, elsewhere in Europe.

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# SEISMIC REFLECTION INVESTIGATIONS IN THE HUNGARIAN PART OF THE PANNONIAN BASIN: APPLICATION TO HYDROCARBON EXPLORATION

by

K. MOLNÁR, GY. POGÁCSÁS and J. RUMPLER

In the hydrocarbon exploration surface geophysics has been of primary role and its success or unsuccess considerably affect the chances of the subsequent exploration phases.

The first seismic surveys in Hungary were carried out in the late thirties. The Geophysical Exploration Company measured nearly hundred thousand km seismic profiles between 1952 and 1985 in Hungary. The early seismic surveys supplied basic information about the regional structures but, in most cases, only the upper part of the Neogene sediments could be explored in detail. Still, significant hydrocarbon bearing structures of the Neogene beds were explored. In the mid sixties with analog magnetic recording instruments shallow and medium depth ranges could be surveyed but the deeper zones could not be mapped.

Since the early seventies by means of digital data recording and multifold coverage it is possible to get good penetration and to trace the base of the Neogene all over Hungary. Moreover, in many cases it has been possible to get structural and tectonic information about features below the Neogene sequence (MOLNÁR, 1976; 1982).

Based on some thousand boreholes drilled for examination of structural forms indicated by many ten thousand kilometers seismic profiles and seismic measurements in the Pannonian basin, first of all the thickness of the prospective Neogene sequence (Fig. 1) is roughly known (KŐRÖSSY, 1980; POGÁCSÁS, 1980: KILÉNYI and RUMPLER, 1985).

The maximal thickness of the Sarmatian and older Miocene sediments accumulated in the Neogen depressions amounts to 3 km in certain zones. That of the Pannonian and younger strata overlying unconformly or by correlative conformities these sediments, is as much as 5 to 5.5 km.

As we know so far, the maturated Neogene sediments filling the deep depressions constitute the most significant hydrocarbon generating sequence. Hydrocarbons deriving from Neogene sediments were trapped in Mesozoic and Palaeozoic formations partly by lateral, partly by downward migration.

The Geophysical Exploration Company's modern seismic profiles bear geometric, tectonic, stratigraphic as well as facial information. By means of their seismostratigraphic analyses the individual depositional units, their spatial and temporal position and the unconformity surfaces dissecting the sedimentary sequence can be identified. Depending on the density and accuracy of the available borehole information as well as on the basis of the seismic information and of the resolution of seismic data the events of basin evolution, the subsidence and filling of the individual basin parts can be reconstructed and the geothermic history of the sediments can be determined.

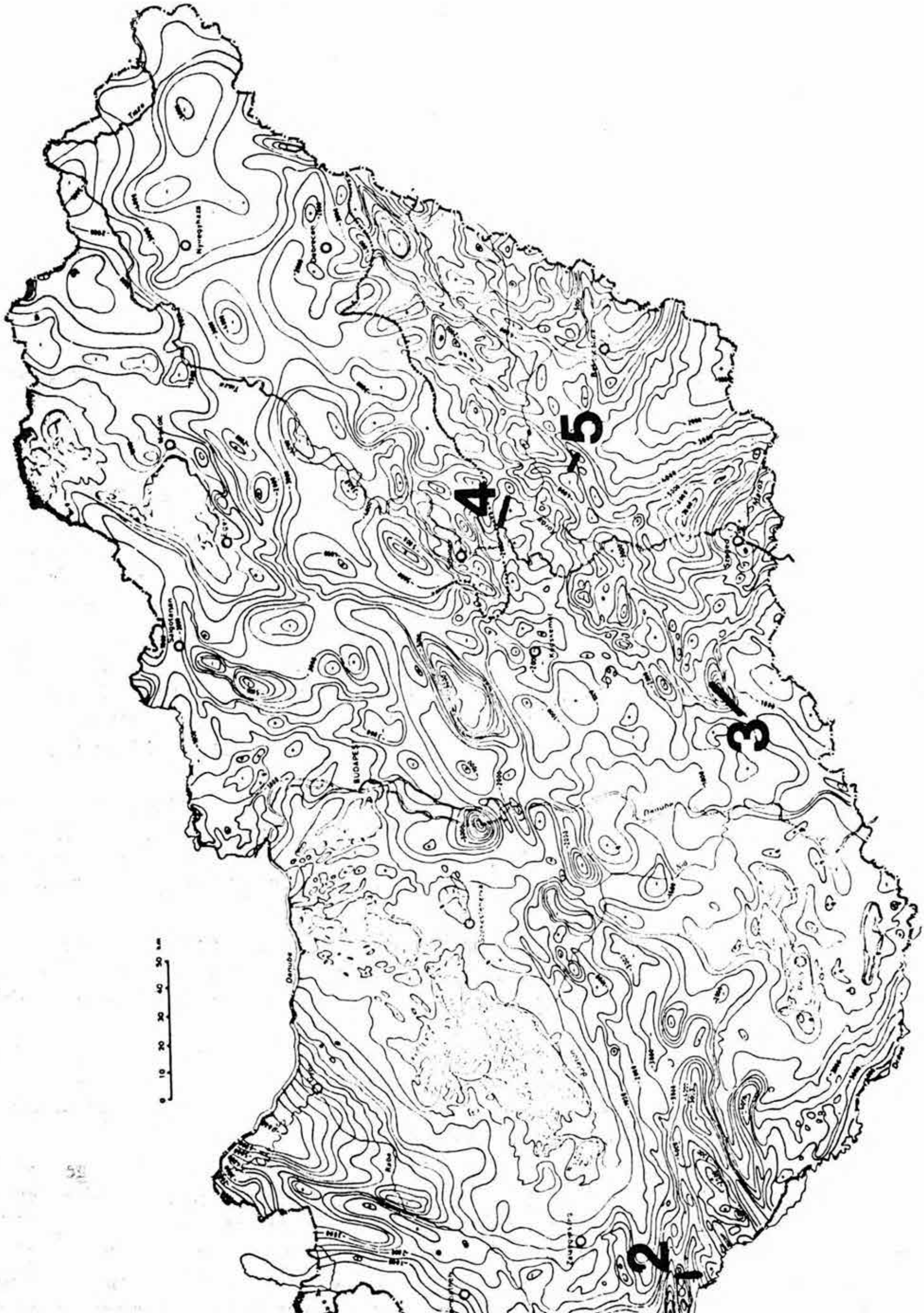


Fig. 1.

Comparing and supplementing the data from the seismic facial analyses with the sedimentological results the reconstruction of basin evolution can be carried out also in regions where no suitable borehole data are available (e.g. deep basins). The determination of the deposition environment according to seismic data provides a possibility to assess the content and type of the organic matter as well as the expected position of hydrocarbon accumulation zones related to subtle traps. Case studies on different types of hydrocarbon deposits (compressional anticlines, structural and unconformity traps, growth faults related traps palaeogeomorphic traps) explored in the Pannonian basin as well as their appearance in the seismic profiles will be presented. The location of seismic profiles are shown on Fig. 1.

*i Compressional anticlines*

On the seismic profile (Fig. 2) measured in the Budafa deep zone (west Hungary) hydrocarbon fields relating to compression anticline are seen. Above the strongly dipping Karpatian, Badenian, Sarmatian and Pannonian sediments filling the depression, the surface terrain shows also elevated position. In the exploration of the hydrocarbon fields in the Pannonian strata the surficial dip measurements carried out by geological compass for half a century ago played important role.

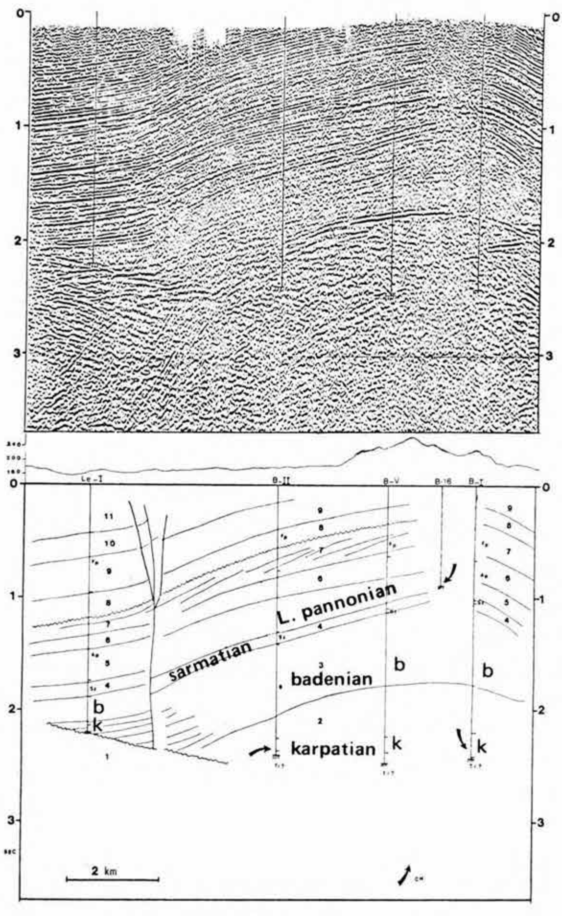


*Fig. 2.*

In harmony with the seismic picture, the Pannonian strata do not form deposition archs with sedimentary origin but are shown active folds (zone of the Savian folds).

In this region the recent seismic measurements aim to investigate the Neogene basement. For the sake of the subsequent exploration of the gas and gas-condensate field explored in the last years the stratigraphic, structural and tectonic conditions of the deep-lying Neogene formations should be revealed.

*ii Structural and unconformity traps*

On the seismic profile (Fig. 3) measured in the region of the Kiskun depression a hydrocarbon accumulation zone is seen related to compression folding. In this region in relation with the Miocene horizontal plate motions along wrench faults a pull-apart-type basin developed. Parallel with progress of the strike-slip movement the southeastern part of the depression subsided ever deeper along listric faults (POGÁCSÁS, 1985). Faults are ever younger when moving toward the margin of the depression. Parallel to or shortly after the extension subsidence of the southwestern part, due to the compression regime developed in the northeastern part the Miocene

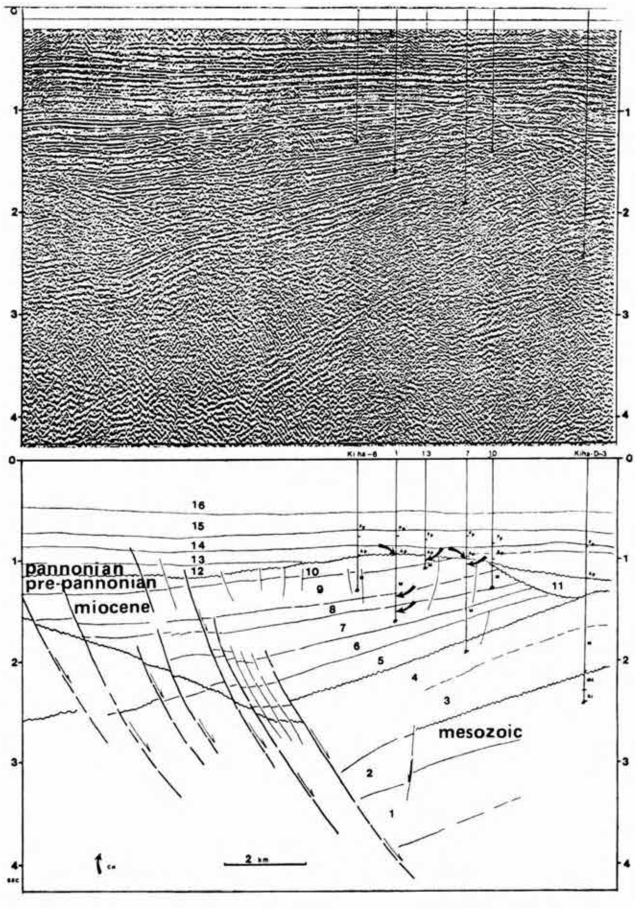


*Fig. 3.*

sediments of that part were arched and the top of the arch was uplifted. This uplift led to erosion. In the profile it can be fairly well seen that the formations of the Pannonian cycle overlie older Miocene by erosion unconformity.

The hydrocarbon traps were developed partly below this unconformity, partly in the deeperlying Middle Miocene sandstone sequences.

In the northeastward rising Mesozoic basement also gas and oil reservoirs are found in higher tectonic position.

*iii Palaeogeomorphologic traps*

On the next seismic profile (Fig. 4) the Martfű CH field can be seen. In different regions of the area hydrocarbons and carbon dioxide are found in formation of different geological age.

The oldest reservoir rocks are Lower Cretaceous diabase and limestone. The Miocene sedimentary rocks as well as the Lower Pannonian marls overlying these rocks are gas-bearing. Gas fields are known in the Lower Pannonian pinching out sandstones as well.

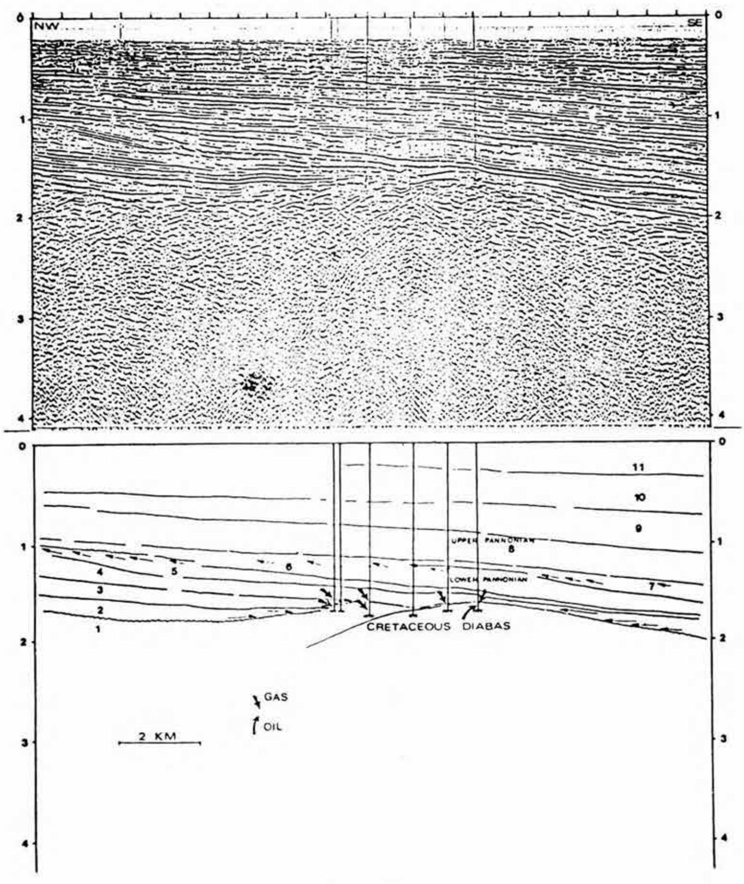


*Fig. 4.*

The individual CH pools are bound to stratigraphic–palaeogeomorphological traps.

The profiles intersect the progradational delta series nearly in dip-direction. The progradation filling progressed from NW to SE.

*iv Traps related to growth faults*

On the seismic profile intersecting the Szarvas hydrocarbon field (Fig. 5) several reservoir horizons are seen. In the Upper Pannonian six, in the Lower Pannonian fourteen traps were discovered so far. The movements along the rejuvenated tectonic lines of the basement might affect also the Pannonian formations, thus the Pannonian reservoir sequence was also dissected. This is why in the Lower Pannonian traps of the individual areas gas of different composition was accumulated.

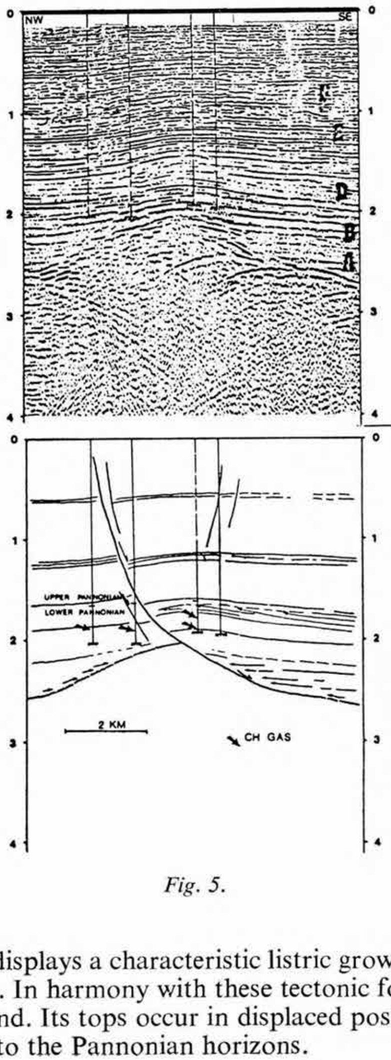


*Fig. 5.*

The Szarvas region displays a characteristic listric growth fault that affected also the Quaternary sequence. In harmony with these tectonic forms, in the downthrown roll over structure is found. Its tops occur in displaced position, i.e. displaced along the fault plane similarly to the Pannonian horizons.

The fault plane is flattened parallel with increasing depth and is probably continued in a low angle normal fault.

*CH perspectives*

The geochemical investigations (KONCZ, SZALAY, VETŐ, VÖLGYI, VÁNDORFI) proved that the Palaeogene and Mesozoic sedimentary formations are also hopeful from the point of view of hydrocarbon generation.

Their maximal thickness can be assumed in the central part of the Great Hungarian Plain (Nagykunság). Total thickness may reach as much as 8000 m.

The major part of the Mesozoic formations that may be potential hydrocarbon source rocks reached the temperature window needed to hydrocarbon generation in the Neogene.

The lower boundary of the depth range to be investigated by hydrocarbon exploration seismic measurements can be put to the lower boundary of the thickest sedimentary sequence.

This depth boundary coincides with the average focal depth of the earthquakes in Hungary (ZSÍROS, 1984) that indicates stresses accumulated during tectonic movements and which generate deformation first of all in this or in smaller depths. The lower boundary of the tectonically active upper zone probably coincides with the high-conductivity zone indicated in the Neogene basement by the geoelectric measurements (ÁDÁM, NAGY).

In some areas the hydrocarbons emigrating from the sedimentary formations accumulated in the traps relating to zones of the metamorphic basement of secondary porosity. For this reason, within this depth interval it is reasonable to explore the internal structure of the Metamorphic Basement, too.

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# NEOGENE BROWN COAL DEPOSITS IN HUNGARY

by

Gy. Radócz, M. Bohn-Havas and Gy. Szokolai

This paper is aimed at a short characterization of the stratigraphy of Neogene brown coal deposits in Hungary.

Brown coal deposits are known in Hungary from all Miocene stages (Fig. 1). They are found in several areas and at different stratigraphic levels according to their palaeogeographic and tectonic position.

Brown coal seams generally occur in the marginal areas of the deep basins and in intramontane molasses (Fig. 2). These coal seams lie in some areas on older as a basal seam. In other areas they are observed as intermediate members of a cyclic sequence locally several hundred-meter-thick (freshwater, brackish and marine).

Two, relatively small brown coal seams of Eggenburgian age have been known for 25 years under the "Lower Rhyolitic Tuff" Felsőnyárád, Borsod basin, N Hungary. Ottnangian coal seams, mined for a long time in the N Hungarian region (Borsod, Nógrád) as well as in W Hungary at Brennbergbánya. In Borsod, the brown coal-bearing sequence of cyclical structure is transitional, to the Karpatian. The presence of Ottnangian and Badenian brown coals is indicated in several localities in the Mecsek—Bakony—Vértes—Dunazug and Börzsöny Mts areas. The Middle Badenian brown coal deposits at Hidas are the most important in the Mecsek Mts. These deposits have been mined earlier. There are brown coal indications in the Ottnangian as well. Both the Várpalota and Herend occurrences in the Bakony Mts are Middle Badenian. The Várpalota brown coal deposits are now being mined. Mining at Herend was carried out only temporarily, just like at Hidas. The first coal mine in Hungary was established at Brennbergbánya in 1759; it was closed down in the 50s.

The map also shows several small Middle Miocene (Ottnangian — Badenian) coal deposits in the Hidas and Mecsek areas. There are small occurrences near Nagygörbő and Devecser (on the basis of one borehole per locality), and also at Balatonföldvár, Pusztamiske and Kapolcs (in the Bakony region), NE of there at Fehérvárcsurgó — Gánt, near Tököl and Dömös and in the Börzsöny Mountains (Nógrádverőce, etc).

Upper Sarmatian brown coals (or lignites) are known in some localities of Transdanubia, at Inota, Csákvár, Mány and in N Hungary (at Felsőnyárád and Edelény in the Borsod basin, as well as in the Cserehát region). The seams are better developed in the surroundings of Edelény where lignite had been mined before.

Lower Pannonian formations, recently placed into the Miocene, bear considerable brown coal and lignite seams in N Hungary, in the surroundings of Szendrő, Rudabánya and Komjáti. In the Szendrő, Rudabánya, Szuhogy, Galvács and Abod areas these deposits had been mined for some time. Sites are also known in the Rakaca, Debréte, Teresznye and Szuhafő areas, along the margin of the Szendrő — Rudabánya and Gömör Karstic Mountains. A Lower Pannonian deposit is known in Transdanubia, in the Várpalota area near Öskü.

*Fig. 1.*
Neogene coal occurrences in Hungary

| Age of seams | Name of occurrences (or basins) | Brown coal of medium quality | Lignite (brown coal of poor quality) | Underground mining | Open pit mine | Explored reserves | Prospective resources | Depth of prospective seams | Water hazard | Gas and explosion hazard | Tectonics | Note |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| UPPER PANNONIAN (PLIOCENE) | Alföld (indications) | | ǂ | — | — | — | o | ⊘ | ⦶ | — | — | Locally depth as well |
| | Taktaharkány | | ǂ | — | — | o | o | ⦶ | ⦶ | — | — | |
| | Bükkalja | | ǂ | ⚒ | ⊗ | ⊘ | ⊘ | ⦶ | o | — | — | Planned open pit mine: Bükkábrány |
| | Mátraalja | | ǂ | ⚒ | ⊗ | ⊘ | ⊘ | ⦶ | o | — | — | Open pit mine Visonta: (from 1967) |
| | Cserhátalja | | ǂ | — | — | ⦶ | ⦶ | ⦶ | ⦶ | — | — | Open pit mining is possible |
| | Kisalföld (indications) | | ǂ | — | — | — | — | ⦶ | ⦶ | — | — | On the basis of boreholes |
| | Nyugat - Magyarország | | ǂ | ⚒ | ⛏ | ⊘ | ⊘ | ⦶ | o | — | — | Planned open pit mine: Torony |
| | Dunántúli - Középhegység (indications) | | ǂ | — | — | — | — | ⦶ | o | — | — | Small ⚒: Vállus |
| | Dél - Dunántúl (indications) | | ǂ | — | — | — | — | ⦶ | o | — | — | Small ⚒: Kiscsehi |
| LOWER PANNONIAN | Komjáti | | ǂ | — | ⛏ | o | o | ⦶ | ⊘ | — | — | Water danger only the alluvium of Bódva. Mines established after 1910 |
| | Szendrő - Rudabánya (indications) | | ǂ | ⚒ | — | o | o | o | ⊘ | — | — | |
| | Szendrő - Szuhogy - Galvács | | ǂ | ⚒ | — | o | o | o | ⊘ | — | — | |
| | Öskü (indications) | | + | — | — | — | — | o | o | — | — | |
| SARMATIAN | Borsod - Cserhát (indications) | | + | — | — | o | o | ⦶ | o | — | o | On the basis of boreholes |
| | Edelény (Császta - puszta) | | + | ⚒ | — | o | o | o | o | — | o | Operated as early as 1855 |
| | Inota — Csákvár - Mány (indications) | | + | — | — | — | — | ⦶ | o | — | o | |
| BADENIAN | Dunazug — Börzsöny (indications) | | + | ⚒ | — | — | — | ⦶ | o | — | o | Nógrádverőce mine opened in 1767 |
| | Iszkaszentgyörgy — Gánt — Tököl (indications) | | + | — | — | — | — | ⦶ | o | — | o | |
| | Várpalota | | + | ⊗ | ⚒ | o | o | ⦶ | ⦶ | — | o | Mine opened in 1876 |
| | Herend — Szentgál | | + | ⚒⛏ | ⚒ | o | o | ⦶ | o | — | o | Mine opened after World War I. |
| | Balatonföldvár (indications) | | + | — | — | — | — | ⦶ | — | — | o | On the basis of one borehole |
| | Hidas | | + | ⚒ | — | o | o | ⦶ | o | — | o | Mine opened in 1860 |
| OTTN. and BAD. | Nagygörbő and Devecser | | + | — | — | — | — | ⦶ | | — | o | On the basis of one borehole per locality |
| | Mecsek (indications) | + | + | — | — | — | — | ⦶ | | — | | |
| OTTN.-KARP. | Kelet - Borsod (Sajóvölgy) | + | | ⊗ | ⊗⚒ | ⦶ | ⦶ | ⦶ | o | — | ⦶ | In Cserehát depth as well. Extraction near Miskolc since 1830 |
| OTTNANGIAN | Nyugat - Borsod (Egercsehi - Ózd) | + | | ⊗ | — | ⦶ | ⦶ | ⦶ | o | — | ⦶ | Area mined continuously since 1840 |
| | Nógrád | + | | ⊗ | ⊗⛏ | o | o | ⦶ | o | — | o | In the Southern part depth as well. Mine opened in 1848 Coal known since 1767 |
| | Brennbergbánya | + | | ⚒ | ⚒ | o | o | ⦶ | o | ȯ | o | Mine opened in 1759 (first industrial coal mine in the country) |
| EGGENBURG. | Felsőnyárád | + | | — | — | o | — | ⦶ | o | — | ⦶ | |
| OLIGO - MIOCENE | Kocs and Leányfalu (indications) | + | | — | — | — | — | ⦶ | | — | o | |
| | Naszlop | + | | ⚒ | — | — | — | o | ⦶ | — | ⦶ | Know for a long time. Mine reopened after World War I. |

Legend

- \+ Occurrence of few seams type
- ǂ Occurrence of several thin seams type
- ⚒ Closed mine
- ⊗ Working mine
- ⛏ Planned mine
- — Does not exits or unimportant
- ⊘ Important } relative characters
- ⦶ Medium } relative characters
- o Small } relative characters

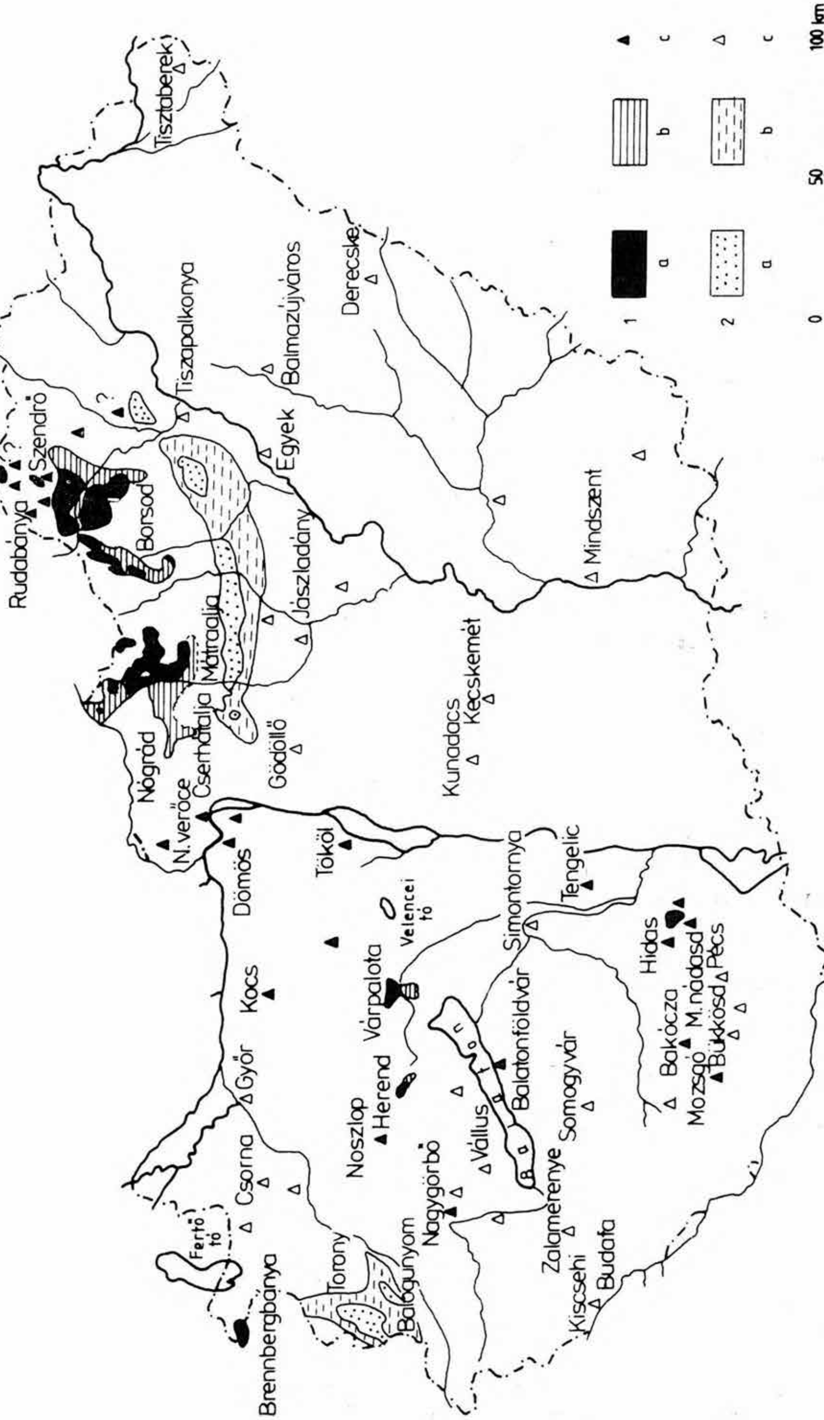


*Fig.* 2. Outline map of Neogene coal occurrences in Hungary

*1* Miocene brown coal deposits, *2* Pliocene coal deposits; *a* explored areas, *b* prospective areas, *c* indications

In the Upper Pannonian beds there are considerable brown coal seams mixed with palaeo soil and detrital wood (lignite). Very large swamp areas were formed in the marginal areas of the Pannonian sedimentary basin.

The Cserhát—Mátra—Bükkalja lignite seams are dipping down to 300 m and growing thinner from the margin of the North Hungarian Range towards the plain. The relationship of these with the generally thin seams in known several parts of the plain as well as with the very small occurrences at different depths in Transdanubia, is not exactly known. It is certain that largest swamps in the area of the present Hungary existed in the Pliocene.

The most important brown coal and lignite occurrences are as follows:

1 *Eggenburgian brown coal bearing sequence*

The 300 m thick Eggenburgian Felsőnyárád Formation under Ottnangian rhyolite tuffs contains two coal seams. Reconnaisance of the seams is completed and further exploration is suspended. The limnic seams bear no fauna, although they often contain beds with siderite nodules, indicating a palustrine facies. The cover is well studied. A brackish fauna appears approximately 15 m above the seams. Mytilus- and Pirenella-bearing assemblages alternate according to the oscillation of the shoreline. Mytilus-bearing assemblages can be found in the agitated, sandy part of the shoreline whilst Tympanotonus—Pirenella associations prefer sludgy coastal plains, overgrown by vegetation.

With the progress of transgression, assemblages appear that are characterized by the predominance of Pitaria and Turritella as well as Flabellipecten belonging to several types. These communities indicate a normal-salinity shallow sublittoral environment. The Felsőnyárád Formation is of Eggenburgian—age a fact proved by its position as well as by micro- and macrofaunal studies.

2 *Ottnangian—Karpatian brown coal bearing sequences*

2.1. *Brennbergbánya*

The average thickness of the coal bearing sequence at Brennbergbánya is 50 to 60 m, with a 4 to 10, seldom 15 m thick basal seam. The fauna is rather poor, only coalified faunal remains (Brotia sp., Unio sp. and Bythinia) were found both in the footwall and the hanging wall of the coal seam. The seam is lenticular derived from a lacustrine marsh, as the upper 50 m thick sequence. Its age is Ottnangian, determined on the basis of its stratigraphy. Its fauna and flora is insufficient for a biostratigraphic age determination.

The brown coal seam continues into Austria. Before the mine closed down mining was going on under Austrian territory by agreement between the two governments. Since the most significant parts of the seam have been extracted, it is estimated that there is very little Brennberg coal left.

2.2. *Nógrád (Salgótarján) brown coal basin*

The brown coal bearing sequence in the Nógrád basin (Salgótarján Brown Coal Formation) is 40 to 180 m thick and generally accommodates 3 seams. Exploitation of these seams and the study of their fossil flora and fauna started in the past century. Industrial coal reserves are low today. The formations are Ottnangian as evidenced by foraminifers, pollen, nannoplankton and macrofauna.

The sequence can be divided into 3 members and contains a characteristic fauna:

The lowermost (Nógrádmegyer) member, is a 20 to 40 m thick continental, fluviatile and swamp formation with fossil vertebrates. Mastodon, Prodinotherium, Testudo, etc were found.

The middle (Kisterenye) member contains 3 coal seams. The lower seam is limnic, with coalified plant remains and represents a freshwater—marsh facies. Rocks, bearing a characteristic Congeria fauna follow and contain the second seam representing, an oligohaline—mesohaline estuary. The overburden is represented by nearshore brackish lagoon facies (Vizslás Sand and Mátranovák Claymarl Beds) with a Cardium fauna beds rich in trace fossils, ichnofossils and fish remains.

The hanging wall of the Salgótarján Brown Coal Formation is composed of beds bearing a purely marine Chlamys fauna (Egyházasgerge Sandstone Formation).

### 2.3. *Borsod basin*

In the eastern part of the Borsod basin the thickness of the brown coal bearing sequence exceeds 300 m. Seams in the eastern part of the basin are paralic, as evidenced by their fauna. The complex is Ottnangian possibly extending into the Karpatian. Most of the complex belongs to zone NN3. It was formed by a oscillatory transgression during which swamp intervals were interrupted by nearshore brackish lagoonal shallow water marine facies. Eleven macrofaunal assemblages have been distinguished which, due to their facial sensitivity, were suitable for the reconstruction of the palaeoenvironment and provided important data on the evolution of the basin.

Among the 11 associations 3 are brackish (with Brotia, Congeria, Theodoxus) 3 are transitional with both brackish and marine environments (Mytilus, Ostrea, Psammobia) and five are marine (with Pitaria, Anadara, Corbula, Zostera—Spirorbis). It can be clearly seen from the enclosed profile that the first significant ingression occurs after the IVth seam. Marine episodes are often observed between seams although these are not so significant as the first ingression.

In the so-called western basin area (between Egercsehi and Ózd) few fossils have been found so far. The facies analysis of the sequence is still going on.

It can be concluded, that the three most important Ottnangian brown coal areas in some respects differ from each other. The Brennberg area is a local lacustrine—palustrine formation without detectable connections with the sea. The Salgótarján Brown Coal Formations are due to the Ottnangian transgression, forming a continuous transition between the freshwater—marsh to the estuary and lagoonal environment. No marine faunas or formations are know in the area from this period. The sea over flooded the site only in the Karpatian. The brown coal and fauna of the Borsod basin, in contrast, evidence frequent ingressions as early as the Ottnangian. The area covered by five main seams in Borsod and three seams in Nógrád as shown on the map is relatively large, although their productivity is expected to be low.

## 3. *Badenian brown coal bearing sequences*

### 3.1. *Herend*

The thickness of the brown coal bearing sequence in the Herend region is 20 to 25 m. Extent and reserves of the seams are small, the quality of the brown coal (lignite) is also poor. Mining is at present suspended.

Three seams are known in the sequence. The lowermost, IIIrd lacustrine seam is of autochtonous type. The footwall is lacustrine clay with $CaCO_3$ concretions, bearing *Brotia esheri*, *Planorbis* sp. Seams II and I are allochtonous and bear a brackish fauna.

Clayey beds between the seams bear palustrine assemblages with Planorbis, Helix, Brotia and Unio predominant. They also contain partly brackish Theodoxus and Hydrobia-bearing assemblages in the lower part. Assemblages characterized by the appearance of the predominant *Pirenella picta mitralis* in the clay interbeddings with carbonate mud occur in the upper intervals. The overburden of the brown coal-bearing sequence contains fossil assemblages representing an upward transition to the marine Corbula-bearing beds with an abundant and varied Mollusca fauna.

3.2. *Várpalota*

The Várpalota brown coal bearing sequence usually comprises one coal seam only. The footwall is continental clay without fauna. Planorbis and Lymnea were found in the brown coal seam, evidencing freshwater environment. The cover, however, bears a brackish fauna. The immediate hanging wall contains a Congeria, Theodoxus and Bulimus-bearings assemblage. Then follow beds with leaf—im prits, and fossil fish. The Bulimus—Theodoxus-bearing assemblage appears again higher up in the sequence. The sedimentary cycle ends with a 1 to 2 m thick clayey coal bed.

The faunal study revealed that the coal had been formed in the lagoonal part of the Badenian sea, where brackish conditions became predominant after a swamp-forming cycle.

Most of the coal is of poor quality in the generally 5 m thick seam. It is used in electric power plants.

3.3. *Hidas*

Hidas is the type area of the Middle Badenian Hidas Brown Coal Formation. The brown coal-bearing sequence is 30 to 110 m thick and comprises 7 seams. The low-quality Hidas brown coal (lignite) was mined since 1860 with long interruptions. No coal has been brought to surface the 60s. Faunistically, the sequence is divided into three parts. The lower member bears assemblages with Unio, Planorbis, Brotia. The middle part is characterized by as Hydrobia, Cerithium and Rissoa assemblages. Ostrea, Anomia, Cardita, and Corbula assemblages occur in the upper part. Thus the lower, middle and upper parts bear freshwater, brackish and marine faunas, respectively.

There are also local interbeds of different facies. The Ostrea, Anomia and Pecten-bearing intercalations appear in the hanging wall of seam VI, as a record of the first marine ingression.

These brown coal seams are most probably of paralic origin, as evidenced by the marine ingressions. Coal formation took place in a lagoonal part of the Badenian sea, where swamp conditions prevailed periodically due to sea level oscillations. Both the footwall and the hanging wall of the Hidas Brown Coal Formation are shallow water marine sediments. The study of their fauna showed that the footwall is Lower Badenian and the hanging wall is Upper Badenian.

The foraminifers from the ingressive bed sequence are Middle Badenian.

4 *Sarmatian brown coal bearing sequences*

Upper Sarmatian brown coal and lignite seams are relatively thin and of low quality. Therefore, their mining is not economical in either of the localities. However, the Borsod coal (Edelény) had been mined before. The age of the seams has been defined on the basis of their position as well as by palaeontological evidence.

5 *Pannonian brown coal bearing sequences*

Pannonian formations of characteristic biostratigraphy and lithology are represented by sequences 100 to 600 m thick in the marginal areas of the basin and between inselbergs, as well as 600 to 4500 m thick in the central parts of the basin. Lignite beds are known at different levels throughout the basin. Important lignite deposits, however, were formed only around the mountain margins.

Potentially productive areas are:

— Deep basin areas (thickness of Upper Pannonian sequence usually over 5 to 8 thousand m), where important seams are in considerable depths (e.g. Great Hungarian Plain)

— Marginal areas of inselbergs without productive lignite occurrence (e.g. foregrounds of the Transdanubian Central Range: Bakony, Vértes, etc.).

The W Hungarian lignite basin (in the Torony area) in the foreland of the extensions of the eastern Alps.

— North Hungarian lignite range in the southern foreland of the Cserhát—Mátra and Bükk Mountains.

— In small depressions in the intramontane bays (e.g. Rudabánya, Szendrő area). For exploitation high capacity seams, suitable for open-pit mining are great importance. The Cserhát—Mátra—Bükkalja lignite region is the largest continuous coal area in Hungary. All the former underground mines (Petőfibánya, Gyöngyös) are now closed down. A large open-pit mine is still in operation at Visonta. A much smaller deposits in W Hungary (Torony) extends over the state border to Austria but it lies far from the industrial axis of the country. Exploration aimed at the investigation of possible open-pit mining are in an advanced stage for example Bükkábrány. There are several more adjacent prospective localities.

The Mátra—Bükkalja sequence in the exposed Upper Pannonian interval is composed in 50 to 80 per cent of fine to medium-grain sands, less silt, clay and 2 to 15 m thick lignite seams consisting of several beds. The main complexes are traceable over long distances, but interfinger towards the centre of the basin. Interfingering of the seams is characteristic in the N areas as is their decreasing coal content, although this is obscured in several localities, due to weathering. The productive lignite region

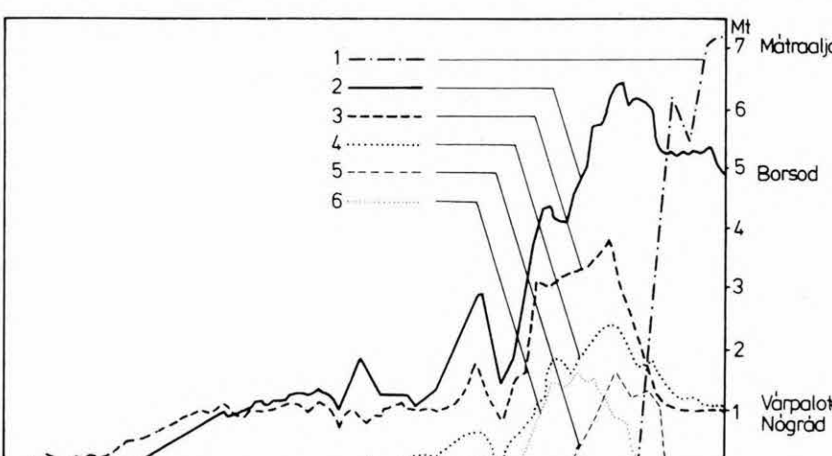


*Fig. 3.* Production data of the Borsod, Nógrád, Várpalota and Mátraalja mines (based on different sources)

*1* Mátraalja open-pit mine Visonta, *2* Borsod, *3* Nógrád, *4* Várpalota, *5* Mátraalja the former open-pit mine et Ecséd, *6* abandoned subsurface mines at Mátraalja contains important lignite reserves. Their extraction, however, is economical only if strip-mining is used. In the vicinity of Visonta, the Thorez open pit mine produces 7 million tons of lignite annually. This lignite is burned in electric power plants.

This production rises from exploitation of Hungarian coal basins (Fig. 3 and Table 1).

**Data for 1980 of Neogene coal mines and explored areas** (after J. FÜLÖP, 1981)

*Table 1*

| Collieries, mining works, and coal basins | Production in 1980 (Mt) | Mines in operation or under development | | | Explored areas | | Coal consumption in 1980 (%) | | |
|---|---|---|---|---|---|---|---|---|---|
| | | No | Coal resources | | | | Power plants | Industry | Population |
| | | | Mt | kJ/kg | Mt | kJ/kg | | | |
| Nógrád coal mines | 0.926 | 6 | 18.1 | 13109 | — | — | 80 | 7 | 13 |
| Borsod coal mines | 5.317 | 13 | 83.1 | 12690 | 1758 | 12430 | 48 | 8 | 44 |
| Várpalota collieries | 1.231 | 3 | 57.5 | 10065 | 230 | 8750 | 63 | 10 | 27 |
| Thorez mine (Visonta) | 7.253 | 1 | 98.3 | 6294 | — | — | 100 | — | — |
| Karácsond | — | — | — | — | 260 | 7910 | — | — | — |
| Kápolna, Füzesabony | — | — | — | — | 1230 | 6455 | — | — | — |
| Bükkábrány | — | — | — | — | 594 | 6920 | — | — | — |
| Torony | — | — | — | — | 431 | 7365 | — | — | — |

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# GENETIC TYPES OF OIL SHALES IN HUNGARY

by

Cs. Ravasz and G. Solti

As a result of the research work during the reambulation of geological mapping started in 1973, four occurrences of oil shale deposits were discovered in Hungary (Fig. 1). Except for two indications, the oil shale deposits were formed during the Late Neogene. Regarding the environments of sedimentation, two main types can be distinguished:

*1* oil shales deposited in maars,
*2* oil shales deposited in intramontane lagoons.

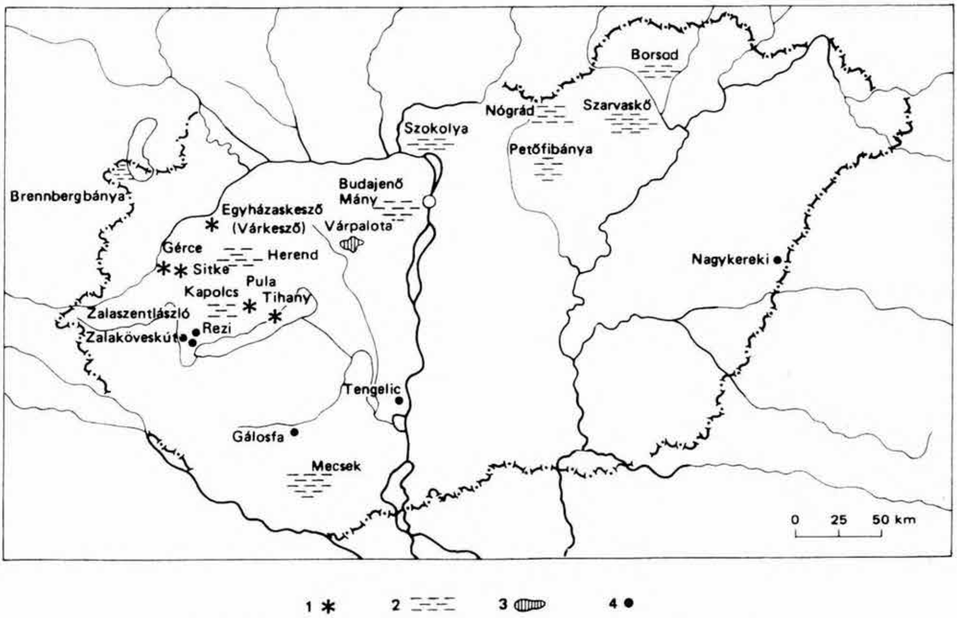



*Fig. 1.* Oil shale occurrences in Hungary (after G. Solti, 1984)
*1* Maar volcano, *2* lagoon type oil shale indication, *3* oil shale deposit, *4* oil shale indication in borehole

In further detail an attempt has been made at showing the major characteristics, of the two different genetic types of oil shales (Fig. 2, 3), in order to facilitate the discovery of similar deposits beyond the borders of Hungary.

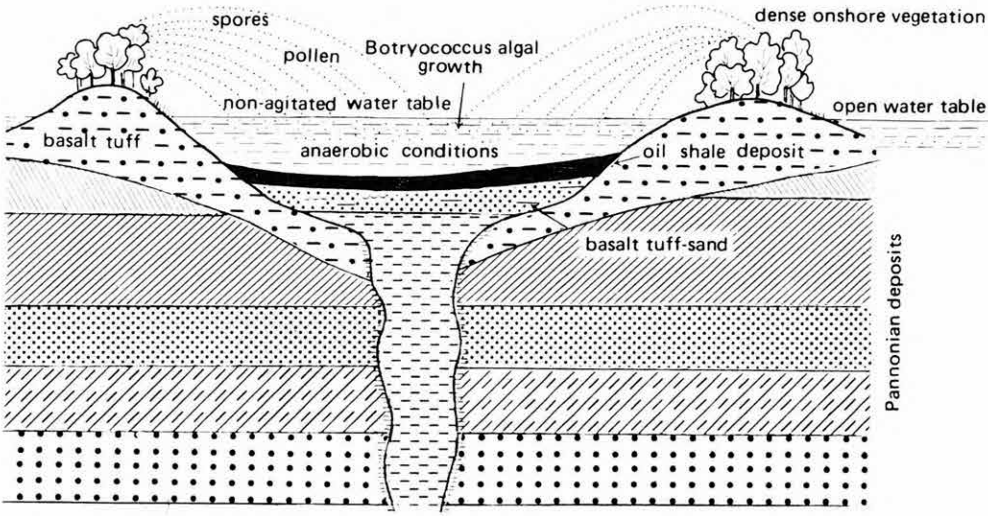



*Fig. 2.* Oil shale formation in a volcanic crater (after G. Solti, 1984)

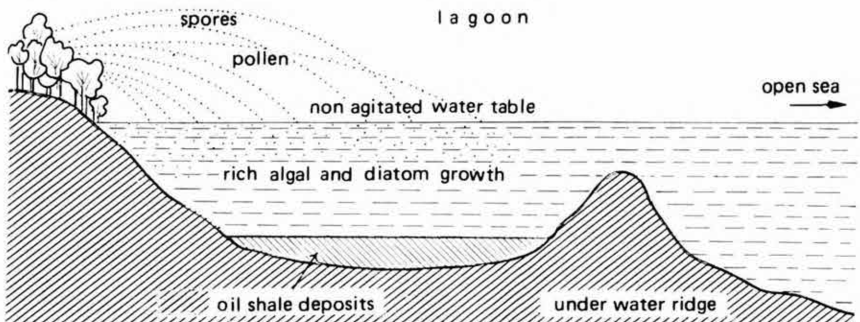



*Fig. 3.* Oil shale formation in a lagoon (after G. Solti, 1984)

**Maar-type deposits**

In the Little Plain and the Balaton Highland, the process of sedimentation in a series of small basins characterized by shallow, brackish water during the Late Pannonian was interrupted by a volcanic activity of alkali basalt type. As a result of repeated eruptions, circular accumulations of pyroclastic material were built up, like rings, around the volcanic craters. Where the top of these tuff rings emerged above the sea level, a small, closed sedimentary basin with special characteristic features could develop. In the region of the extended lake system that time, this type of crater lakes was also represented.

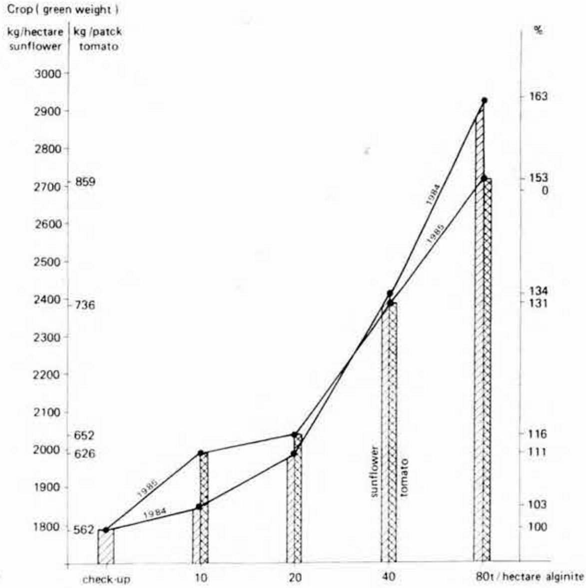
Represented by deciduous forest, as proved by remnants of Ulmus, Carpinus, Fagus, Quercus and also the undergrowth, the vegetation worked against large-scale erosional processes and against the transportation of coarse detritus into the lake.

The water of these crater lakes was oligohaline type as a result of rainfall and periodical brackish water inflow with a salinity of 3% a pH of 7.6 and an average temperature of 20 °C. Because of the slow erosion of the basaltic tuff along the slopes and the rapid weathering of the hydrated, basic pyroclastic material and the contemporaneous streams of postvolcanic activity, the water of the lake got enriched in silica and essential rare elements.

This environment was favoured by particular organisms, algae and diatoms, respectively. Among others, the *Botryococcus braunii*, an oil-generating algal species, occurred in unusual abundance. During the warm season a biomass was formed by the flourishing algal culture in the lake. Anorganic components, siliceous, calcareous and argillaceous material played a major role in sedimentation, while organic matter was deposited in greater amount during the cold season. Laminae 0.1 to 5 mm thick and alternately rich and poor in organic matter resulted in a varved sedimentary sequence of annual rhytmicity. During lithification, laminae rich in organic matter were altered into alginite, calcareous alginite or diatomaceous alginite by the process of diagenesis. A series of rock types of different composition was defined by petrographic examinations but, here and now, we confine the petrographic definition of the alginite is understood in a more strict sense. The rock type called alginite consists of figured algal remains and fragments of algal colonies, both being infiltrated and cemented by opal enclosing also tests of diatoms and containing bituminites. It forms thin layers, patches and dispersed drops of microscopic size. Megascopically, it can be characterized by a brownish-green, grayish-green colour, laminated structure. Its bulk density is approximately 0.9, the bituminite content being over 10%. When lit by striking a match, the rock can be easily burnt.

Following the uplift of the area the crater lake ran dry and under favourable circumstances it was buried by Pleistocene, Holocene sedimentary rocks and thus prevented from erosional processes.


*Fig. 4.* Alginit accumulation can be supposed in Graz basin, Slovakia and Transylvania




| | pH (KCl) | Arany-type bond | $CaCO_3$ | Humus | $P_2O_5$ | $K_2O$ | N | Mg | Mn | Na | Zn | Cu |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | | Soluble in ammonium lactate total | | | | | | | |
| | | | % | % | ppm | | | | | | | |
| Alginite | 7,68 | >90 | 14,5 | 14,3 | 410 | 1020 | 1086 | 295 | 102 | 62 | 4,6 | >10 |
| Sandy soil | 7,33 | 24 | 17 | 0,97 | 187 | 164 | | 55 | 75,3 | 51 | 22 | 10,9 |

*Fig. 5.* Increase of sunflower and tomato yield due to the influence of amelioration with alginite at Izsák

Seven crater lakes of this type covered by younger sediments were investigated in detail. Three of these have a lacustrine sedimentary sequence with alginite deposits. The thickness of these sedimentary rocks varies from 40 to 90 m, the diameter of the tuff rings is between 330 and 2000 m. The total volume of the accumulated raw material is 130 million tons.

Alginite and oil shale deposits in crater lakes bordered by tuff rings were first recognized in Hungary. This characteristic sequence, now called the Pula Alginite Formation after the type locality at Pula, is explored by boreholes.

In terms of palaeogeography and geological environments, the Early Pliocene Carpathian basin during is supposed to have offered favourable conditions for alginite accumulation. And this holds true for territories outside Hungary as well (Graz basin, Slovakia, Transylvania). Some of these further occurrences could be detected along the fault line of the river Rába as a continuation of the alginite deposits of the Kemeneshát, up to the Alpine range, where a few rings of maar-type basaltic volcanoes, probably filled with alginite deposits, are found.

### Lagoon-type deposits

Detailed information about the deposition of alginite in crater lakes induced geologists to detect similar material in sedimentary basins of similar environment.

Sedimentary basins, closed periodically by bars, with nonagitated brackish waters of shallow depth, filled by tuffaceous, diatomaceous, pelitic layers of claymarl and marl and containing also organic material, were taken under searchlight. Such facies represented by intramontane basins and lagoons of the Neogene sea are well known within the country. This kind of environment seems to be similar to places of evaporite deposition. Using the bar theory, a similar model can be adopted. The basin, a bay or lagoon, is partly closed against the open sea. In the relatively warm brackish water, vegetation-generating hydrocarbons could start flourishing. Accumulations of organic material were supplied simultaneously by large amount of spores and pollen grains of a lush vegetation along the coast. The remains of flora, rich in hdyrocarbons deposited on the basin floor in an anaerobic environment. In case of swampy shores no coarse-grained anorganic detritus was transported into the basin.

The oil shale deposited in this type of basins is megascopically similar to the alginite of maars. The rock is greenish-brownish gray in colour, thin lamellae rich in organic component alternate with lamellae or thin layers of silty clay-marls, marls, diatomaceous, tuffaceous layers commonly occur, too. Due to the reducing, anaerobic conditions the benthonic fauna is completely missing.

The shale oil content of the layers rich in organic material varies between 5 and 55% mainly as a derivative of spores and pollen grains.

Significant deposition of lagoon-type oil shale can be found in the Várpalota basin in Transdanubia, at the foot of the Bakony Mountains. Covering an area of approximately 50 km$^2$ it was deposited during the Badenian and is overlain by regressional lignite layers.

With a view to using the oil shales, i.e. alginite, for practical purposes, numerous inventions were developed as documented by sixteen patents and licences, by the Hungarian scientists. Most significant of these is, as the authors believe, the utilization of natural alginite and oil shale in agriculture for soil melioration. Alginite, as a natural fertilizer, can stimulate plant growth considerably and this kind of fertilizer is valuable from the point of view of environment protection, too.

As proved by agricultural experiments run continuously since 1977, the alginite deposited in crater lakes is quite promising for agricultural utilization, being considered a fossil biomass. The intense growth of Algae, among others, can be attributed to the weathered volcanic material, the microelement content, incorporated in the organism of Algae. It is assumed that, as a result of a reverse process, the slightly diagenized alginite, when fed into the soils, may act again as a nourishing and microelement-supplying agent. This material contains humus needed for crop production, together with the necessary nutritive macro- and microelements in a form that can be directly taken up by the plants. The 10–15% humus contained in the alginite exceeds several times the humus content of the best chernozem. The high clay content (60–80%) enables the fixation of loose sandy soils and the improvement of their composition. Simultaneously, the water storage capacity of the soil increases, the water regime improves and a more balanced absorption of water by the plants becomes possible for a longer period.

As a result of the alginite's organic matter supply, nutritive capacity, composition improvement and water-regime control of the ameliorated soils depending on the dosage and on the soil type (1–5 kg/m$^2$), the crop yields can be increased by 15–50%. As an example an experiment will be described that was carried out on sandy soil, short of humus. In 1984 on soils ameliorated with alginite from Gérce, sunflower was grown, while in 1985 on soils ameliorated with alginite from Pula (Vázsony), tomato was grown. On experimental plots of 100 m$^2$, and 500 m$^2$, respectively, by applying doses of 10, 20, 40 and 80 t/ha (kg/m$^2$) the yield in case of sunflower was 3—63% higher, while in case of tomato it increased by 11—53% as compared to the untreated control plots (Fig. 5).

The ameliorating effect lasts for 4—6 years. The alginite is practically applicable to all soil types and any crop.

A lime content of about 30% makes possible the neutralization of soils that got acidized as a result of the excessive use of fertilizers.

Perhaps the environment-protecting impact of the alginite is even more important than the amelioration and regeneration of the soils got exhausted due to forced increase of crop yields. Once fed to the soil, the clay minerals of the alginite (illite, smectite) bind the fertilizers, which then, enhanced by rainwater, will gradually and evenly get to the roots of the plants. By this the removal of potassium and phosphorus is reduced, the live water is less polluted, and the nitrate content of the subsurface waters (drinking water) is diminished. Also, on account of the clay mineral content, the alginite's absorptive ability to prevent the disintegration of organic matter will improve the microclimate of animal-breeding sites and stables, the odour control and the blotting-up and compaction due to liquid manure.

For the utilization of the oil shales and alginites in their natural state not only the Hungarian alginites of the volcanic crater lakes are suitable. This statement is proved by pedological experiments carried out by using Yugoslavian and Maroccan oil shales in plant-growing plots.

At present, alginite is extracted and sold from two pits in Hungary and the material is used for agricultural purposes, such as amelioration, preparation of various soil mixtures, and for solving problems of environment protection. Its application is first of all preferred in minor cash-producing gardens, where the crop- and flower production is performed without using chemicals or fertilizers.

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# DEPOSITION ENVIRONMENTS OF THE WIELICZKA SALT DEPOSIT

by

A. ŚLĄCZKA

The Wieliczka salt deposit is part of a vast area of evaporites which developed during middle Badenian time on the Carpathian foreland. The stratigraphic position of the Wieliczka salt is based on foraminifers (LUCZKOWSKA, 1978). The salt deposits are underlain and made up by the zones with: *Candorbulina suturalis* (Moravian), *Uvigerina costai* (Lower Wielician) and are covered by the *Neobulimina longa* zone. The sedimentation of salt was preceded by marine fine-grained deposits with sporadic marine conglomeratic fans along the southern margin of the basin. The material of these fans derived from the Carpathians (ALEXANDROWICZ, 1965; DOKTOR, 1983).

The development of the evaporites was rendered possible by:

*1* partial cut off the western part of Paratethys from the eastern part—an open sea,

*2* constant influx of salty water from the East joined with intensive evaporation (GARLICZKI, 1979).

The salt accumulated in a narrow furrow situated in the southern part, the above mentioned evaporitic basin (Fig. 1). The depth of this furrow just before salt sedimen-

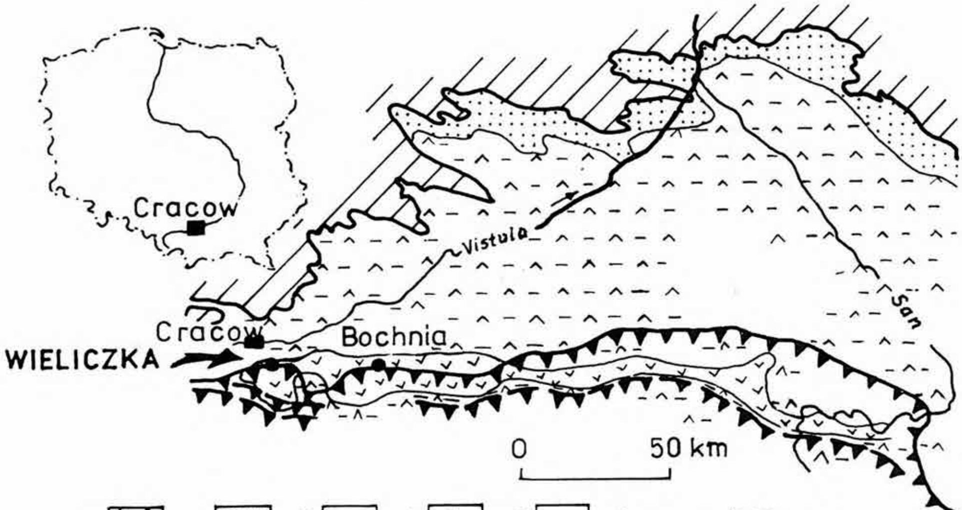


*Fig. 1.* Facies map of the Middle Miocene (Badenian) evaporites developed in the fore Carpathian depression (after GARLICKI, 1979; modified)

*1* Mesozoic and Palaeozoic substrate, *2* littoral facies, *3* sulphates, *4* chlorides, *5* region without evaporites, *6* present-day boundary of the Carpathian overthrust, *7* supposed boundary of the Carpathians at the Badenian time

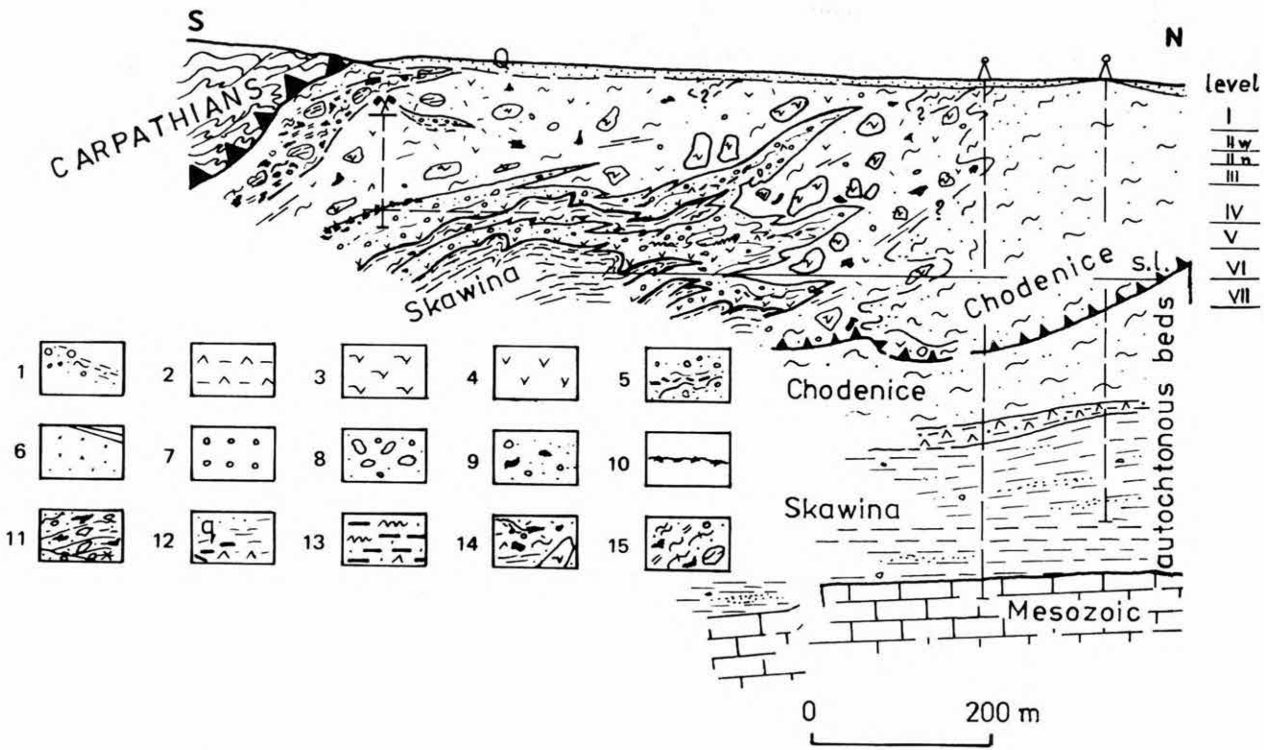


*Fig. 2.* Transverse section of the Wieliczka Salt Mine (after GAWEL, 1962; GARLICKI, 1968; modified)

1 *Stratified Member: 1* Littoral facies, *2* sulphates, *3* southern salt, *4* lower salt, *5* upper salt (Spisa); types of Spisa salt *6* pelitic laminated, *7* white (pure), *8* conglomerate, *9* gravelly saltstone, *10* breccia, *11* breccia and conglomerate with corals, *12* sand, clay, anhydrite, *13* mudy anhydrite intercalation. 11 *Chaotic Member* (olistostromes): *14* Zuber with huge salt boulders and lenses of redeposited salt, *15* upper breccia of Miocene and Carpathian rocks but without salt

tation was established on several hundred metres on the basis of foraminiferal assemblages (Luczkowska, 1978). There is no evidence of shallowing of the basin, the salt was deposited in a deep basin according to the model, proposed by Schmalz (1969). Moreover, in marly layers within the upper part of the salt sequence there were found pelagic foraminifers similar to those occurring beneath the salt. It shows also that in the upper part of the basin, salinity was, at least periodically, normal. On both sides of the chloride facies sulphate and littoral (fossiliferous) facies developed. These facies belts are more narrow on the southern margin then on the northern one. Along the uplifted part of the Carpathians there were probably subaerial cones of coarse material redeposited from the Carpathians. A problem which is still unsolved is the relation ship between the Wieliczka salt and the folded Carpathians. The main reason for it is a lack of preserved contact of the Wieliczka salt and the underlaying marls with their primary substratum. The whole sequence was striped off (Fig. 2). In our paper in the first place is presented idea that the Wieliczka salt basin was developed only on the Carpathian foreland (Fig. 3.). According to another idea the Wieliczka salt basin embraced also the marginal Carpathians part and Lower Badenian together with salt sediments was deposited not only on the platform but also on the Carpathians. This possibility is shown in Fig. 3a.

The Wieliczka salt sequence was deposited the central part of the furrow. It is composed of two units: the Stratified Salt Member, and the overlying Salt Breccia Member. The profile of the Stratified Salt Member begins with layers of coarse to finegrained salt originated from precipitation (Fig. 3, I). There are secondary intercalations of marls with anhydrite and layers of gravely mudstones with fragments of the Carpathian rocks (Gawel, 1962; Garlicki, 1979). The latter sediments can be an evidence of a tectonic movement within the area of the Carpathians. The next stage corresponds to a quiet period when several metres of pure coarse grained salt was precipitated. Higher in the profile together with salt also quartz grains and detritus of fauna have been deposited (lower part of Spisa salt). The salt displays often bands of dark and light colour and thin intercalations of mudstone with anhydrite. The layers are often strongly folded. These structures may be due to synsedimentary slumping as well as to tectonism. During this stage the first layers of redeposited salt with distinct gradation appear. The redeposited salt and slumping can be proofs of an another period of tectonic disturbances which embraced at that time not only the Carpathians but also the southern part of the salt basin. In this southern part during the fore-mentioned stages, mainly thick (up to 30 m) fairly pure laminated salt was deposited. As subordinate to this one there were e.g. stained-glass type salt, dolomitic salt, and coarse grained salt intercalated by marls. These salt deposits are known from redeposited boulders only and their primary pattern is unknown. This stage (lower Spisa salt) is terminated by a complex of mudstones, finegrained sandstones with anhydrite-halite cement and layers of anhydrite. There are fragments of carbonized flora. The clastic sediments contain a wide variety of primary structures (cross-bedding and ripplemarks) as a result of current action. Often they are disturbed by convolution. Current direction was approximately from the West to the East along the axis of the basin.

After the deposition of the clastic sediments once more salt precipitated but with a considerable amount of intercalations of redeposited salt lenses or layers (upper Spisa salt). Exclusively redeposited salt terminate the sequence. During that time the southern part of the salt basin was submitted to uplift, partial desintegration and salt clasts with barren rocks were deposited by density flows to the central part of the basin (Fig. 3, II). As an effect in the Wieliczka salt mine deposits similar to those of

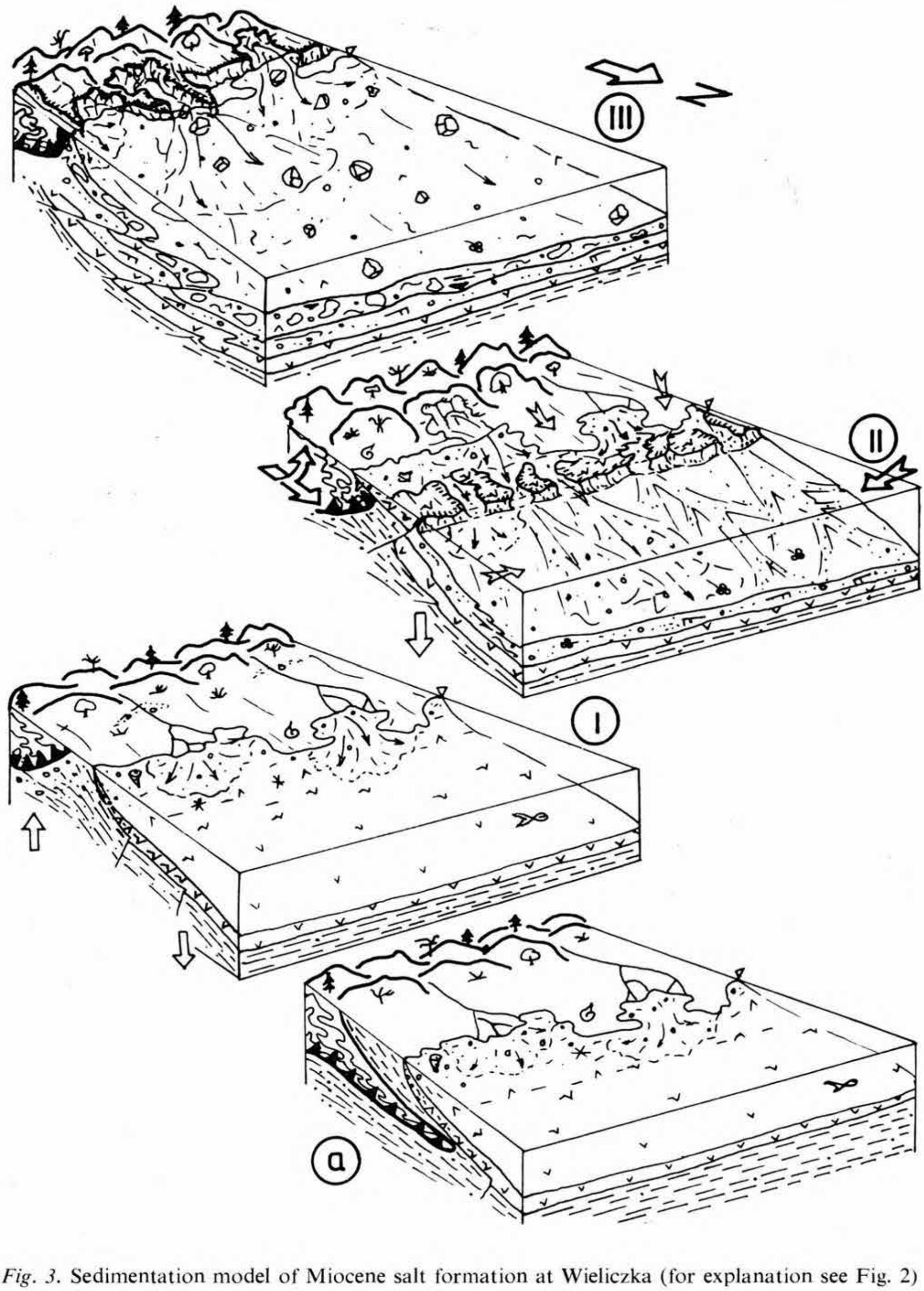


*Fig. 3.* Sedimentation model of Miocene salt formation at Wieliczka (for explanation see Fig. 2)

submarine fans can be distinguished with internal, middle and outer fan associations. The interfingering of these associations due to the lack of continous exposures are not always clear. The internal part of the fan situated in southernmost area of the mine consist of alternated gravelly saltstones, laminated sandstones and conglomerates deposited from various gravity flows (Fig. 4a). The sequence shows an overall coarsening upward tendency. These sediments are built up of redeposited salt fragments with subordinate amounts of pebbles of Miocene mudstones and marls, and also of Carpathian rocks (sandstones, shales, variegated marls, exotic rocks). Characteristic is the disappearance of the Carpathian rocks along current direction. Locally the whole sequence is terminated by cones of sedimentary breccia (Fig. 4a). This breccia is dominated by clast supported conglomerates which at places pass laterally into matrix supported ones. Those beds lack pronounced bedding plane separation and display overlapping contacts. The breccia is polimictic, composed of unsorted pebbles of rock salt and in smaller amount of the Miocene marls, anhydrite, Carpathian sandstones and shales. Very characteristic is the sporadic occurrence of coral fragments (Coenocyathus). The breccia might have be on deposited by avalanche grading into high density turbidity currents. The conglomeratic cone very rapidly wedges out as well in current direction as perpendicularly to it.

Sediments exposed in the central part of the Wieliczka Salt Mine are interpreted here as representing deposits of middle fan (Fig. 4b). The main differences are lack of rocks derived from the Carpathians, lack of sedimentary breccia and greater thickness of conglomerates. They are matrix-supported. Both matrix and clasts are made up by salt. The conglomerates usually grade upwards into laminated salt sandstones. These bodies are interpreted as deposits of high and low density turbidity currents. The conglomerates overlie crude laminated salt layers which do not show clear evidence of redeposition and a part of them may represent salt deposited by precipitation. The central part of the fan is remarkbaly aboundant in carbonized plant remnants although smaller amounts of these occurs in the whole sequence of the salt deposits. These plants belong to the so-called younger mastixioid floras known from the European Miocene, which comprise a number of plants of Mediterranean ecology (Łańcucka—Środoniowa, 1984). Only one genus—Tetraclinis—associated with clearly arid biotope was found. The flora of Wieliczka consists of 136 taxons, with predominance of arboreal plants: dicotyledous and coniferous trees or shrubs, distinctly few herbs (Ceratonia, Myrtus, Olea, Nerium, Paliurus, Pinus, Pistacia, Vitis, Juniperus etc). A distinct predomination of plants from the higher elevations could be stated allochtonous fauna rich of phytophagous small land- and fresh-water gastropods (Helix, Pupa, Planorbis) extracted from the Spisa salt (Kowalewski, 1933) indicate the conditions of the surrounding land area.

The distal part of the fan display generally decrease in clast size and in thickness of layers but increase in quantity of non-salt grains (quartz, cherts, fragments of fauna and glauconite). The salt fan sediments described above are covered by huge debris flow deposits (Salt Breccia Member) of olisthostrome type derived mainly from the southern part of the salt basin (Kolasa and Ślączka, 1984; 1985), Fig . 3. III. The boundary is sharp and at places there are traces of synsedimentary erosion. Sometimes, however, on the top of the fan deposits there is preserved a thin layer of marl with poor planktonic foraminiferal assemblage. The debris flow deposits contain a poorly sorted collection of pebbles and boulders (salt, Badenian barren rocks, Karpatian rocks) embedded within finegrained matrix. The largest (up to thousend of cub. m.) and most common are salt boulders. The matrix is composed of marly of clayey mudstone with scattered small fragments of salt crystals (Zuber). In some places

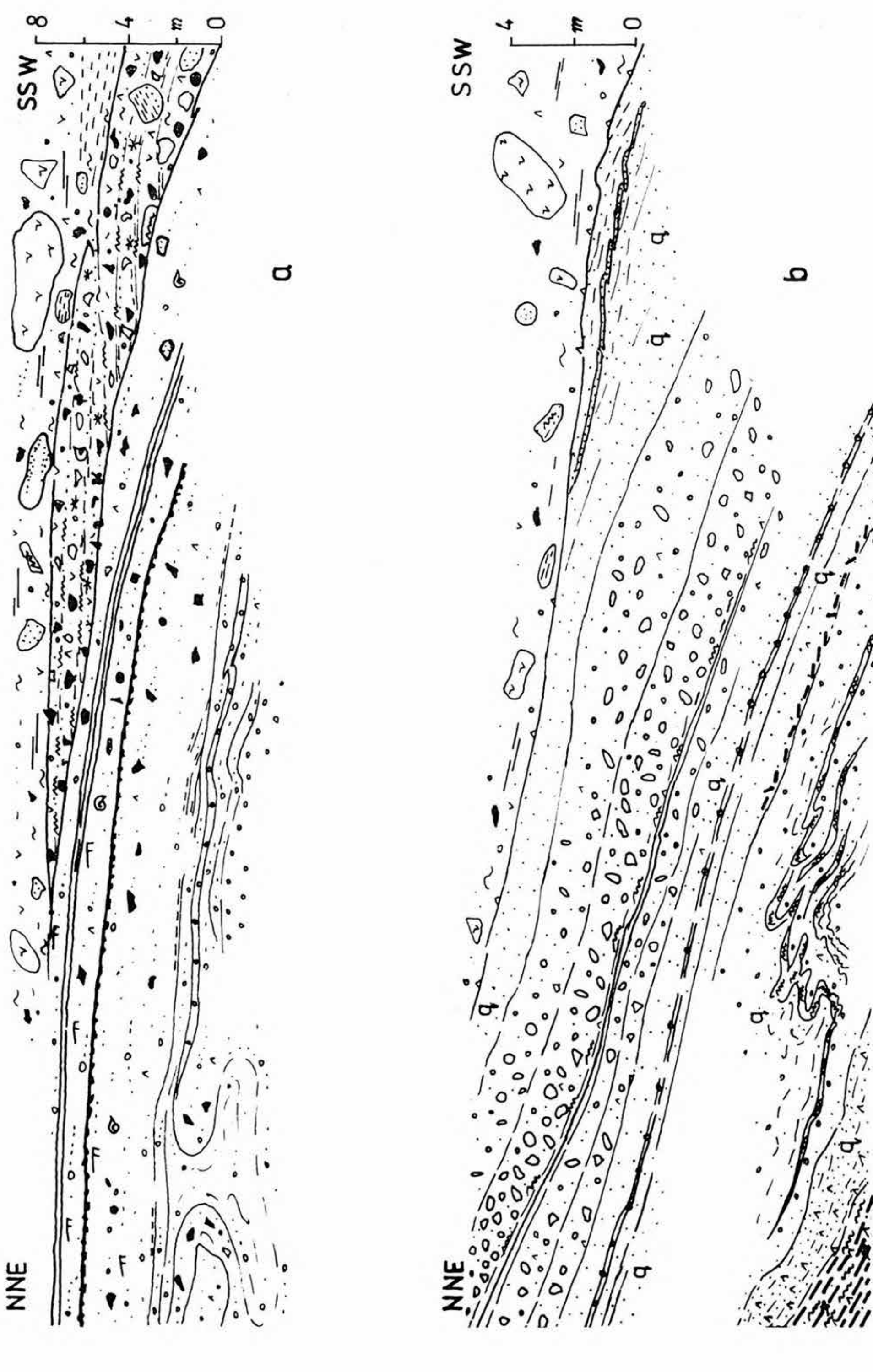


*Fig. 4.* Upper part of the Spisa salt: **a** the southern area and **b** northern one (for explanation see Fig. 2)

debris flow deposit pass upwards into clast-supported conglomerates and finally into laminated saltstones.

The Chaotic Member is terminated by another debris flow deposit with lack of rock salt but abounding in rocks derived from the Carpathians. The debris flow deposits are covered by marls shales, and sandstones of open basin with planktonic foraminifers (Łuczkowska, 1978).

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# HYDROCARBON PROSPECTS AROUND NEOGENE EVAPORITES

by

P. SONNENFELD

*Introduction.* A relationship between gas and oil pools and evaporites has been recognized for a long time (WOOLNOUGH, 1937; WEEKS, 1958; MOODY, 1959; HEDBERG, 1964; GONCHAREV and KULIBAKINA, 1971). It goes back to HARBORT (1913), was again revided by SZATMÁRI (1980) and SONNENFELD (1984, 1985). Worldwide, the vertical and lateral environs of ancient evaporite basins, the surroundings of salt stocks and some of the seams in underground mines have been the sites of commercial and non-commercial hydrocarbon occurrences. This relationship is not accidental.

Oil field palaeo-latitudes cover a belt almost identical to that of evaporite deposits (IRVING and BRIDEN, 1962; IRVING and GASKELL, 1962; GORDON, 1975); over three-quarters of all hydrocarbons known in Europe, the Middle East and the USSR are stored under evaporites or occur in lateral contiguity to them (SOLOWJOW, 1978; KOZLOV, 1978). Hydrocarbon reserves are associated in North America with evaporites in the Michigan, Paradox, Delaware, Midland, and Elk Point basins, to name but a few.

### The facies distribution in an evaporite basin

Brine saturation can occur in a marine embayment whenever run off and inflow cannot match evaporation and seepage losses. The surface inflow is charged with nutrients that allow reef tops to flourish; the expansion of reefs further decreases the cross sectional area of the inflow, hinders the outflow, and thus accelerates the progressive concentration of the resident brine.

To concentrate a brine to saturation for halite and other chlorides requires a water surface several times the size of the halite precipitation area; the shoals and shelves become the sites of gypsum precipitation, act as saturation shelves (Fig. 1). Rapidly subsiding parts of the basin collect brines that contain a variety of chloride complexes in a gradually decreasing amount of solvent. Without synsedimentary subsidence, the basin would fill up promptly, once halite precipitation commences. Halite preservation rates are at least 0.1–4.0 mm/year (BARNETT and STRAW, 1983). Dissolved potassium and magnesium chlorides become both less dense and less soluble upon cooling. They do not mark the deepest part of the basin because they eventually precipitate on basin slopes and outer shelves.

### Organic matter in evaporite basins

Well oxygenated surface waters in any evaporite basin are rich in biota. The swept-in surface waters carry a very prolific fauna and flora, the generation of organic

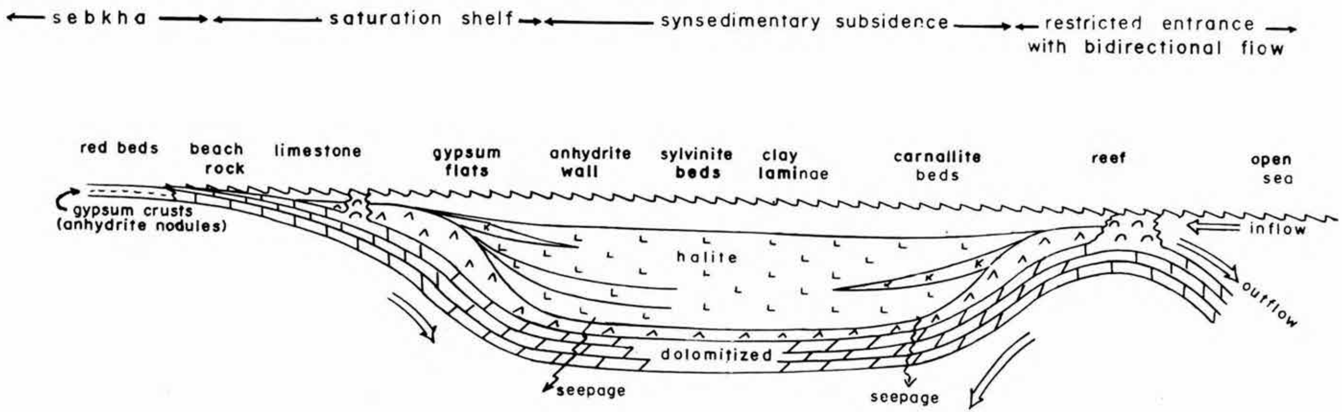


*Fig. 1.* Model of an evaporite basin with synsedimentary subsidence (after SONNENFELD, 1984)

matter per square meter of bay surface is 2—3 orders larger than that of the open ocean. An elaborate food chain is developed from planktonic algae to schools of fish. A Messinian eel (STURANI, 1973) that had ventured far from shore in search of food was preserved in the evaporites because upon sinking through the density interface it died but did not decompose before burial.

The bottom brines become anoxic as soon as they are separated from the atmosphere. The increasing osmotic pressure of bottom waters eliminates the bottom dwellers and burrowers, and organic matter dropping through the pycnocline is not consumed. Only bluegreen algae persist within the photic zone and anaerobic bacteria in the remaining brine column, as bottom dwelling faunas, grazers and scavengers are eliminated and bioturbation of sediments ceases.

The disappearance of this fauna reduces recycling of organic matter, which accumulates in hypersaline brines at an extremely fast rate. Sediments beneath a density stratified brine have a negative redox potential; low solubility of oxygen in hypersaline brines prevents decomposition, leaving it up to anaerobic bacteria to macerate the organic matter. Organic matter continuously raining down from surface waters is decomposed by bacteria, goes into solution or settles on the bottom. Additional hydrocarbons are produced by algae, diatoms and bacteria. Anaerobic fermentation process in the Cariaco Trench, Venezuela, produce both alkanes and molecular nitrogen (LINNENBOM and SWINNERTON, 1971). In an anaerobic environment, more organic matter enters the bottom sediments when rates of accumulation accelerate, not when decomposition is slowed down (RICHARDS, 1970).

Halite and gypsum appear to be at first sight devoid of organic matter. However, the evidence for its presence is manifold. Evaporites contain many direct fossil forms, e.g., embedded fish, plankton, insects, spores, pollen, leaves, tree trunks, or stromatolites replaced by gypsum or halite. Fluid inclusions in evaporite minerals invariably contain both liquid and gaseous hydrocarbon fractions and some organic nitrogen. Organic matter controls the rate of precipitation of evaporite minerals, the crystal size and habit of both sulphates and chlorides. Indirect evidence for organic matter and its maceration products is offered by discoid gypsum crystals, favoured over prismatic ones in the presence of phenols, by an octahedral habit of NaCl and KCl due to adsorbed amines. Amino acids raise gypsum solubility and have been found in ancient anhydrites; unflushed organic matter causes slower setting of gypsum from salinas than from quarries. Sylvite nucleates much more slowly than carnallite; its precipitation at ambient temperature is due to organic nitrogen compounds, as alcohols and other organic surfactants reduce KCl-solubility but urea complexes Mg-chloride preventing its coprecipitation with sylvite. Oxidation of nitrogen compounds leads to carnallite precipitation; dissolved bivalent iron complexes are concurrently also oxidized producing very fine hematite needles embedded in primary carnallite. The presence of aliphatic acid anions can contribute a major portion of the measured alkalinity in a brine (CAROTHERS and and KHARAKA, 1978).

### Bitumina in evaporites

Many bitumina, notably members of the alkane series, are produced in the evaporite environment (Fig. 2). They form by metabolism of anaerobic bacteria, some algae and diatoms. Dissolved organic matter alters the viscosity of the brine, its ability to hydrate ions, and its proficiency to transport organic and inorganic substances.

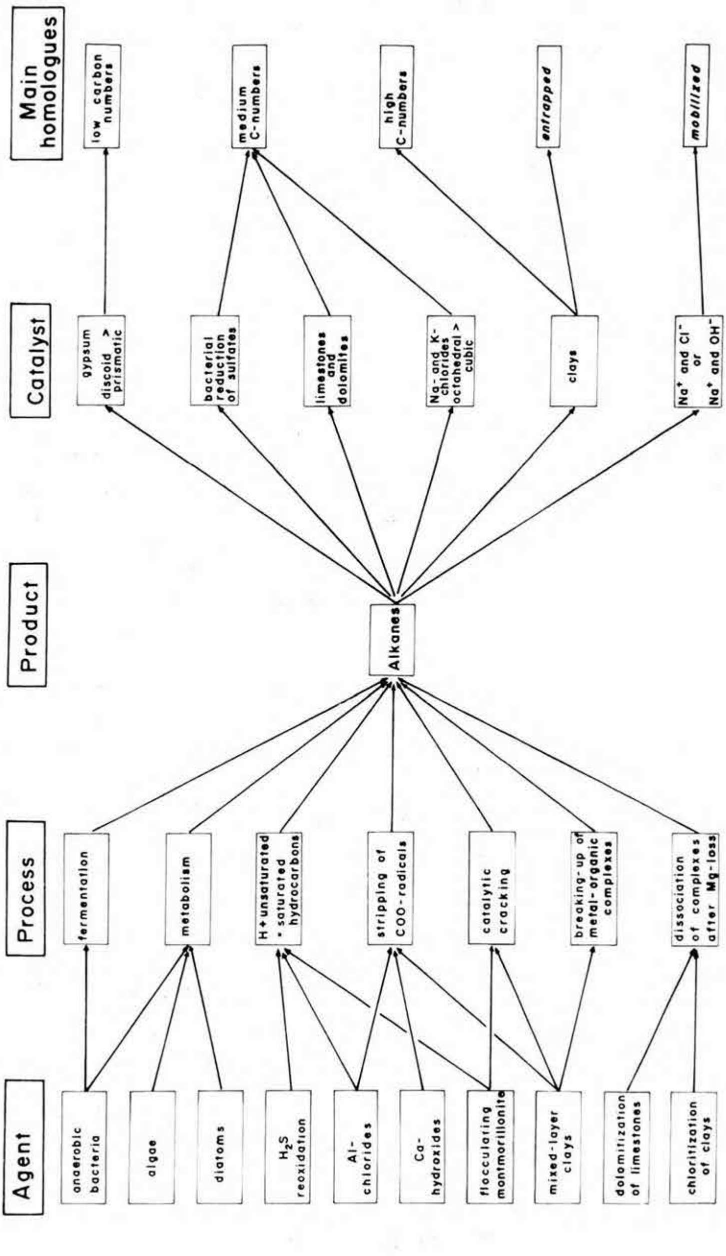


*Fig.* 2. Flow chart of alkane production and hydrocarbon development

Sulphate ions are digested by bacteria within the anaerobic water column but hydrogen sulphide does not become available in the bottom sediment for pyritization of available iron complexes. It either escapes or is reoxydized by bluegreen algae in the photic zone. Hydrogen liberated in the reoxidation of hydrogen sulphide causes unsaturated compounds to change to saturated ones. Adsorption of swept-in organic compounds onto Ca-sites in a gypsum or calcite lattice or even the stripping COO-radicals by Ca-hydroxides leads to the production of alkanes. Mixed-layer clays perform the same function.

Another path to alkane produktion is a bacterial degradation of algal matter that accepts hydrogen from flocculating montmorillonites. Montmorillonites change in the anaerobic brine to mixed-layer clays with brucite pillars and induce catalytic cracking (illitization inactivates these clays). The brucite incorporation leads to the break-up of any available hygroscopic alcohol-Mg-chloride complexes. Alkanes also dissociate from brines stripped of magnesium by dolomitization or by ion exchange with clays. Especially the red bed facies, which is closely related to evaporites, contains clays that have a pronounced catalytic effect on sedimentary organic matter and hydrocarbons (VEBER and GORBUNOVA, 1969). Conversion of blown-in clays into Mg-chlorites liberates both silica and alumina. The presence of Al-chlorides in the brine is indicated by the formation of koenenite or loewigite. These aluminium chlorides turn animal fats and naphthalenes into alkanes.

Limestone and dolomite, reacting with fatty acids, generate liquid bitumina. Calcium sites in gypsum fosters the formation of gaseous homologues, but permeable dolomite intercalations in evaporites frequently yield commercial oil and gas. A bacterial reduction of sulphate to hydrogen sulphide prevents the formation of methana and stimulates the formation of liquid hydrocarbons (SONNENFELD, 1984). If bacterial sulphate reduction is a prerequisite to the production of liquid petroleum, the absence of such activity may explain why ancient gypsum shelves, such as the Miocene Po Valley of Italy (SELLI, 1973; RIZZINI and DONDI, 1978) and the northern Adriatic Sea, located a thousand km of several interconnected gypsum-containing subbasins away from the edge of a salt basin, have yielded natural gas, but little liquid petroleum. Gas has been reported in Sicilian gypsum for a century (SJOEGREN, 1893).

Hydrocarbons in salt and potash beds must be formed in situ: Crude oil from evaporite beds tends to be immature, a high-gravity gasoline-range petroleum. Seeps of crude oil are common in undisturbed halite and potash beds (PETERSON and HITE, 1969), blow-outs are well-known from Miocene potash mines in Poland, occasionally leaving behind a white, waxy "mountain tar." Fluid inclusions of light oil are common in recrystallized halite (KITYK and PETRICHENKO, 1978) and extractable oil in crushed sylvite. The apparent immaturity of evaporite-related crude oils is a consequence of their mode of genesis. One has to bear in mind, that vitrinite reflectivity, a maturation index, sets oil genesis in clastics at 0.5–0.7, but at less than 0.4 in evaporites, i. e. 20–40% lower (HOLLERBACH, 1980).

In short, the organic compounds are altered catalytically into hydrocarbons by Ca-montmorillonites, calcites, gypsum and even by potash deposits. As Ca-montmorillonite and other clays change to mixed-layer Mg-varieties, dispersed organic matter transforms into syngenetic hydrocarbon fluids. Hydrocarbons generated in anaerobic bottom brines contain saturated homologues, whereas oxygenated surface waters generate primarily undersaturated olefins, a rare component of unrefined crude oils (LINNENBOM and SWINNERTON, 1970).

The amount of hydrocarbons generated in evaporite basins is substantial: For each gram of organic matter contained in 1 $m^3$ of incoming seawater, at least 0.003

barrels of crude oil, 57° API-gravity, accrue per $m^3$ of salt precipitated. Organic carbon may reach 150 ppt in uncompacted Holocene evaporites but most ancient salt sequences still contain 2–5 ppt (WILSON, 1984). Translated into liquid petroleum that represents about 300 bbls of oil generated per 100 $m^3$ of salt, of which 4–10 bbls of oil are still retained in ancient sequences. Ancient evaporite deposits measuring several hundred km were capable of generating many billions of barrels of oil; much of this was again consumed by anaerobic bacteria but a significant quantity was able to migrate into surrounding reservoirs.

### Metallic ores

Seawater contains substantial quantities of base metals, which are concentrated in the residual brine when precipitation of evaporite minerals commences. The prevalent whitish color of salt proves that iron remained in solution as bivalent chloride — or organochloride complexes. Some metals are selectively removed by algae and are later released from decomposing organic materials, again as organo–metallic complexes. Organic complexes travelling alongside metal chloride complexes increase solubilities with rising salinities of the brines. Organometallic complexes are quite capable of transporting quantities of base metals adequate to form a deposit. Wherever the pH of the percolating brine is lowered and hydrogen sulphur bearing formation waters are encountered, the metals are then precipitated, the organics cleaved from their chlorides. An association of organic matter and ore deposition has been noted in many localities. However, when mobilized, the heavy metal-bearing brines then travel downward, and the hydrocarbons tend to move up dip, to be trapped in geographically and stratigraphically separated reservoirs. Messinian evaporite deposition occurred on such a large scale that exploration for commercial deposits of lead–zinc or copper ores, or for light-gravity petroleum could be rewarding especially on the periphery of Messinian salt basins.

### Migration of hydrocarbons

The settling out of organic matter occurs indiscriminately over the whole floor of the brine basin and organic materials are trapped only when they settle out over impervious parts. Precipitating gypsum or halite slush is very porous; the accruing organic matter is flushed into the subsurface and delivered to reservoirs. The brine seeping carries out with it entrained organic matter as small molecules and dissolved compounds. High rates of salt precipitation increase both the geostatic pressure on underlying beds and their pore pressure; they increase the pressure gradient between hydrocarbon-bearing strata and reservoirs and promote flushing and migration. The diffusion co-efficients for potassium chloride or sodium chloride and amino acids are entirely related to their volume and shape of diffusing molecules and not to their molecular weight (MEHL and SCHMIDT, 1937) and are temperature-dependent. This is true also of many other organic compounds.

Where the substrate is porous, but not permeable, such as in micritic limestones and marine clay beds, the organic matter is trapped. Clays offer adsorption surfaces more than two orders larger than those of the finest-grained carbonates. Large hydrocarbon molecules are easily trapped when they settle out over impervious clays or dens calcilutites; by echanging Al-hydroxide cations clay can attain a pore volume of over 40 percent and a surface area of 600–2200 $m^2/cm^3$ ; its highly active microspace is easily filled to 70% with alkanes (SUESS, 1970; SERRATOSA, 1979; OCCELLI et al., 1981). Clays are thus able to adsorb large quantities of bitumine before polymerizing them into solids.

In that manner the clays trap the organic matter in muddy bottoms of a bay, later compacted into bituminous shales. The shales act as retaining sponges. However, there is no indication how much of the original organic matter has been flushed into aquifers through adjacent previous substrates. Indeed, impermeable beds adjacent to evaporites (shales or micritic limestones) also are often bituminous: organic matter forms pyrite-rich sapropels and is buried in stinkstones, finely grained, bituminous nearshore limestones, dolomites and shales that originally lined the evaporite basin.

Once entrapped in the claystones the organic matter is unable to move until a substantial pressure is applied. Thermal oil generation from bituminous shales requires a burial depth at which compaction would have completely eliminated the effective permeability; moreover, the clays then act as semi-permeable membranes capable of filtering out all but the smallest molecules (SZATMÁRI, 1980). Mere burial pressure is insufficient to expel the crude oil: Clays containing about equal amounts of crude oil and water still keep almost 50% of the crude oil at a pressure equivalent to more than a 5 km burial (SNARSKY, 1962). POTONIÉ argued in 1928 that if crude oil migrated out of compacting shales, compounds ought to be segregated geographically according to individual threshold pressures and temperatures of fractionation.

The farther the ancient shores of a barred evaporite basin the higher is the hydrogen content of hydrocarbons and thus the greater the oil yield (BREGER and BROWN, 1963). As the brine concentrates and becomes more dense, more and more of it seeps through an initially porous substrate, displaces lighter formation waters underneath the basin floor and carries with it into aquifers dissolved or complexed base metals, silica and organic maceration products. The hydrocarbons move into reservoirs underneath at the flanks and later even into overlying reservoirs (SONNENFELD, 1984). Because oil tends to migrate up dip, the shelf edges and structural traps within a basin, rather than nearshore sites become important exploration targets. Favourable entrapment areas are reef chains, basin-margin fault traps and post-Miocene erosional unconformities. Hydrocarbons, because of their buoyancy tend to separate from the direction of brine movement, wherever an antiform structure such as a reef is passed. Tectonically squeezed evaporites yield their hydrocarbons to aquifers; these hydrocarbons fill reservoirs abutting salt domes. BURK et al. (1969) found highly aromatic and asphaltic high-gravity oils in the caprock of Jurassic Louann salt in the Sigsby Knoll, Gulf of Mexico.

Rock salt is permeable to non-aqueous fluids and crude oil moving through it is undergoing changes (SONNENFELD, 1985): a few ppt are adsorbed, alkane structures increase their chain lengths and are favoured over aromatics, resins and asphaltenes increase. Heavy oil filtered through potash beds becomes alkane-rich light oil (GRAEFE, 1910). Halite veinlets in anhydrite often smell of gasoline; carnallite and at times halite decrepitate when dissolved as gases escape: Miocene "crackle salt" in Poland has a weakly bituminous taste (ROSE, 1839). Gasoline-range hydrocarbons and their gaseous homologues are seeping into overlying sediments from Upper Miocene evaporites beneath the Mediterranean Sea (MCIVER, 1973).

Rising brine salinity lowers the solubility of oil (FAINGERSH, 1977); excess sodium or potassium ions in the percolating brines convert oil-wet to water-wet rocks and thus mobilize any hydrocarbons adsorbed onto pore walls (LEACH et al., 1962). It is no coincidence that Miocene oil flanks Fars evaporites in Iran, Mediterranean salts in Libya, Transylvanian salts at Ploiesti, occurs off-shore Thrace close to salt-filled grabens; similar grabens in the Tyrrhenian Sea have not been investigated.

Both Neogene and Permian salt precipitation are unique in that they were shortly followed by a glacial period with large-scale entrapment of ocean water in ice caps. A drop in sea level bared gypsum shelves, meteoric waters became quickly saturated with calcium sulfate and were able to penetrate into the young, as yet undercompacted evaporite sequence, sulphatizing all K—Mg-salts and most of the time increasing thereby the tenor of the potash ores. However, where such well oxidized brines penetrated into surrounding aquifers, they changed the redox potential into a positive one, the pH into an acidic one and much of the organic matter in all likelihood was destroyed.

## Conclusion

If a genetic relationship exists between gas field and gypsum or anhydrite shelves, and between liquid hydrocarbons and the flanks and substrates of salt basins then this would provide a focus to any exploration strategy; a heavy emphasis would be placed on exploring the environs of that portion of the evaporite basin where primary chlorides were precipitated. Penecontemporaneous migration patterns then determine the early maturation history of hydrocarbons because of additional catalytic influences of host rocks. Hydrocarbons that migrate with brines through the initially porous substrate, move into reservoirs underneath, at the flanks and later even into younger reservoirs. Crude oil can travel up dip and this favours within a basin as exploration targets shelf edges and structural traps, such as reef chains and basin-margin faults, rather than nearshore sites. A regional evaluation of palaeomigration patterns of percolating brines and bitterns from evaporite basins would prove to be a powerful exploration tool for stratigraphic traps around Middle and Upper Miocene salt basins.

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# COMPUTER MODELLING OF THE THERMAL AND MATURATION HISTORY OF THE GREAT HUNGARIAN PLAIN

by

Á. Szalay, F. Horváth and P. Dövényi

*Basin analysis* is a new complex tool for prospect appraisal during petroleum exploration. This is a purely deterministic and predictive approach, which has as the main objective the effective reduction of exploration risk. The basic idea is that the process of generation, migration and accumulation of hydrocarbons is attempted to be reconstructed in space and time during the formation and evolution of sedimentary basins. It can be performed by computer simulation of these processes determining characteristic parameters at any time in different places of the three-dimensional sedimentary complex.

*Basin analysis* can be used successfully at very different levels of knowledge. During the reconnaissance phase of exploration *Basin analysis* may be used to describe the general pattern of basin evolution and to outline the possible maturation of source rocks. In a fairly well-known basin amounts of hydrocarbons generated in the source rocks, migration paths, possible location of traps and their ranking, quantity and quality of reserves can also be predicted.

The flow diagram of *Basin analysis* can be seen in Fig. 1. This diagram shows the kind of input data required to carry out model calculations and the output results. In Fig. 2 to 7 principles of the applied models are explained and some of the results are illustrated.

The first step of basin analysis is the determination of the *Lithogenetic model* of the basin (Fig. 2). It relies primarily on geophysical well logs which are usually available continuously in exploratory drill holes. The sedimentary sequence is divided into pelite and psammite layers, and—with a simple plotting technique—a lithological trend diagram is constructed. Depth intervals with consistent lithological development define lithogenetic units, which characterize the different phases of basin evolution. Some features of the lithogenetic trend diagram (e.g. heterogeneity, thickness of individual beds) make possible to apply them for facies analysis. Lithogenetic units defined in separate boreholes can be correlated and extended into a regional system by the use of reflection seismic profiles. Depth and thickness maps showing also facies changes can then be prepared. Seismic correlation combined with available age determinations and palaeontological data determines the chronostratigraphic meaning of the lithogenetic units.

The second step of basin analysis is a precise determination of the thickness of sedimentary columns through time. This can be done by the help of the *Compaction model* (Fig. 3), which assumes that the decrease of porosity with depth is the consequence of rock compaction during progressive burial. Actual porosity versus depth relationships can be derived from well logs combined with direct porosity determinations on core samples. Thickness reconstruction at any time unit is performed by step backstripping of the present sedimentary succession. This leads to an understanding

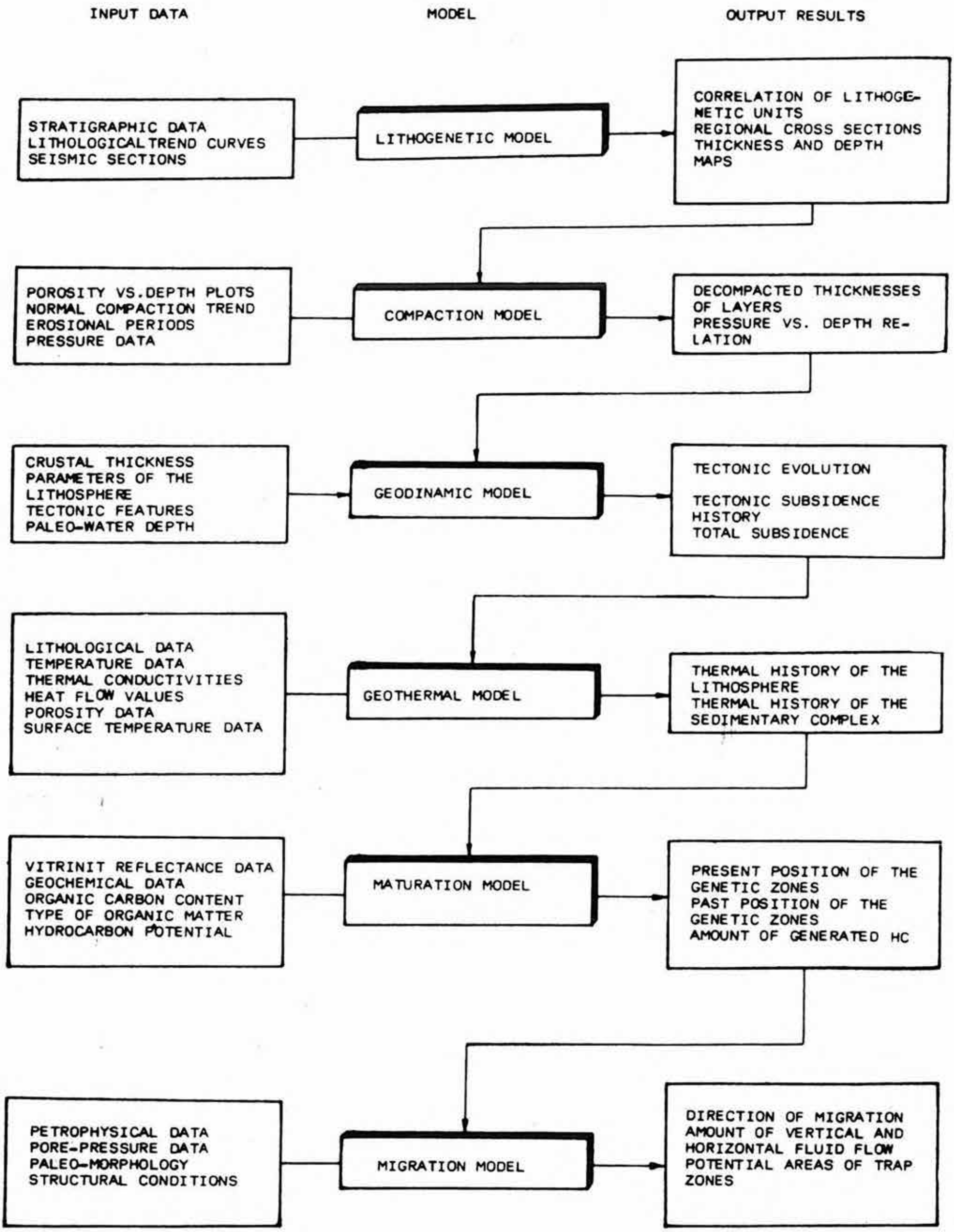


*Fig. 1.* Basin analysis

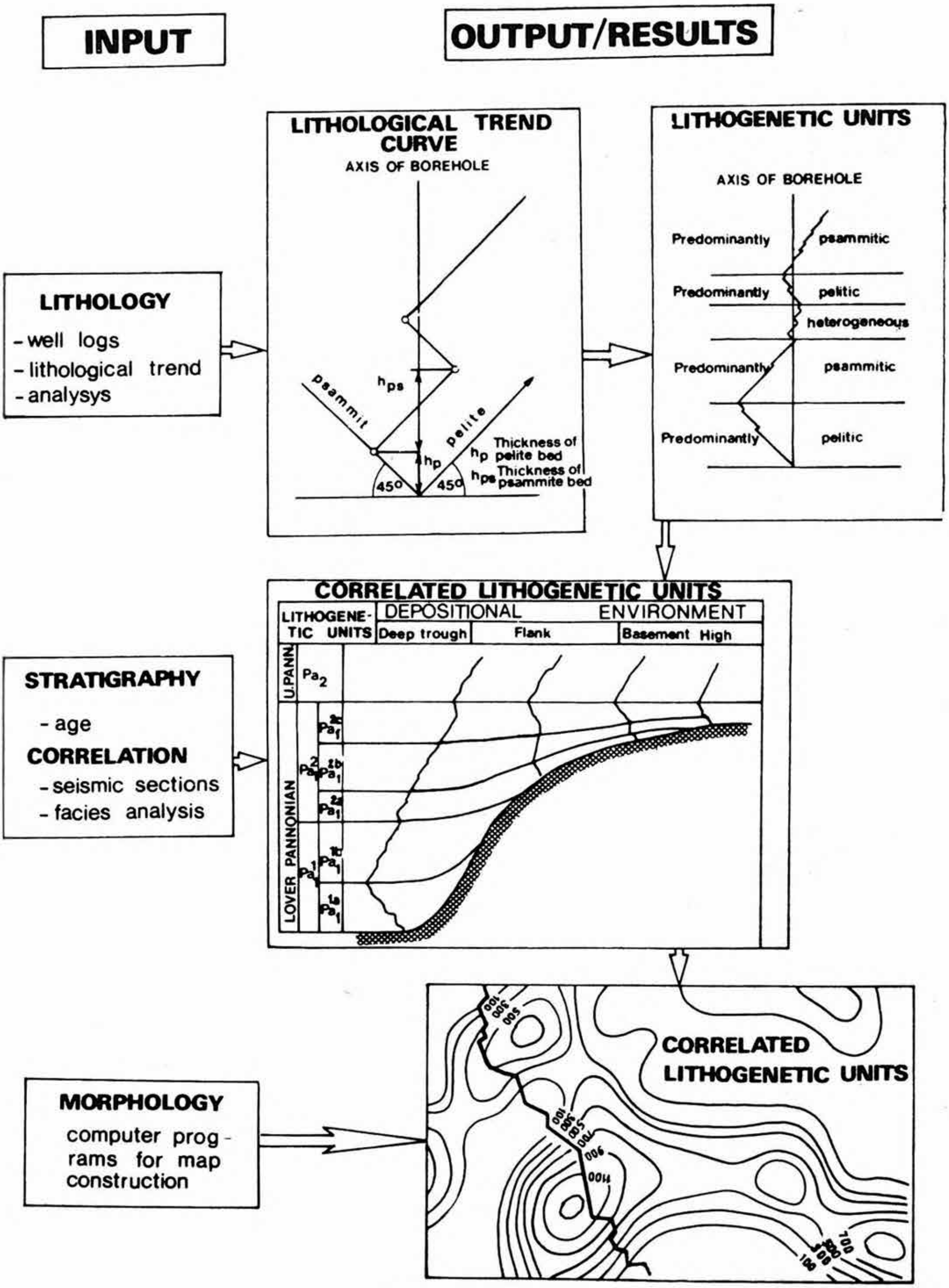


*Fig. 2.* Lithogenetic model

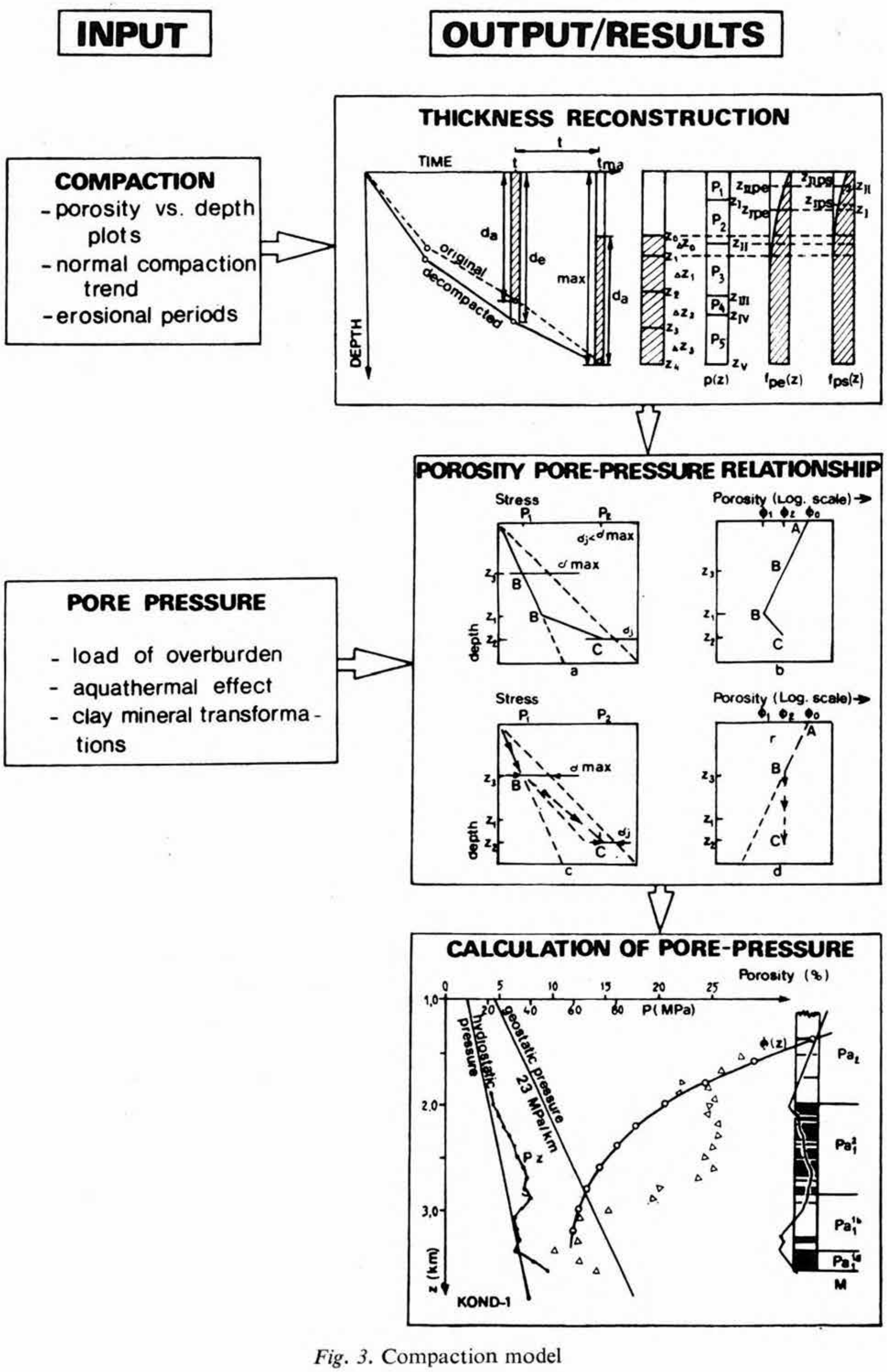


*Fig. 3.* Compaction model

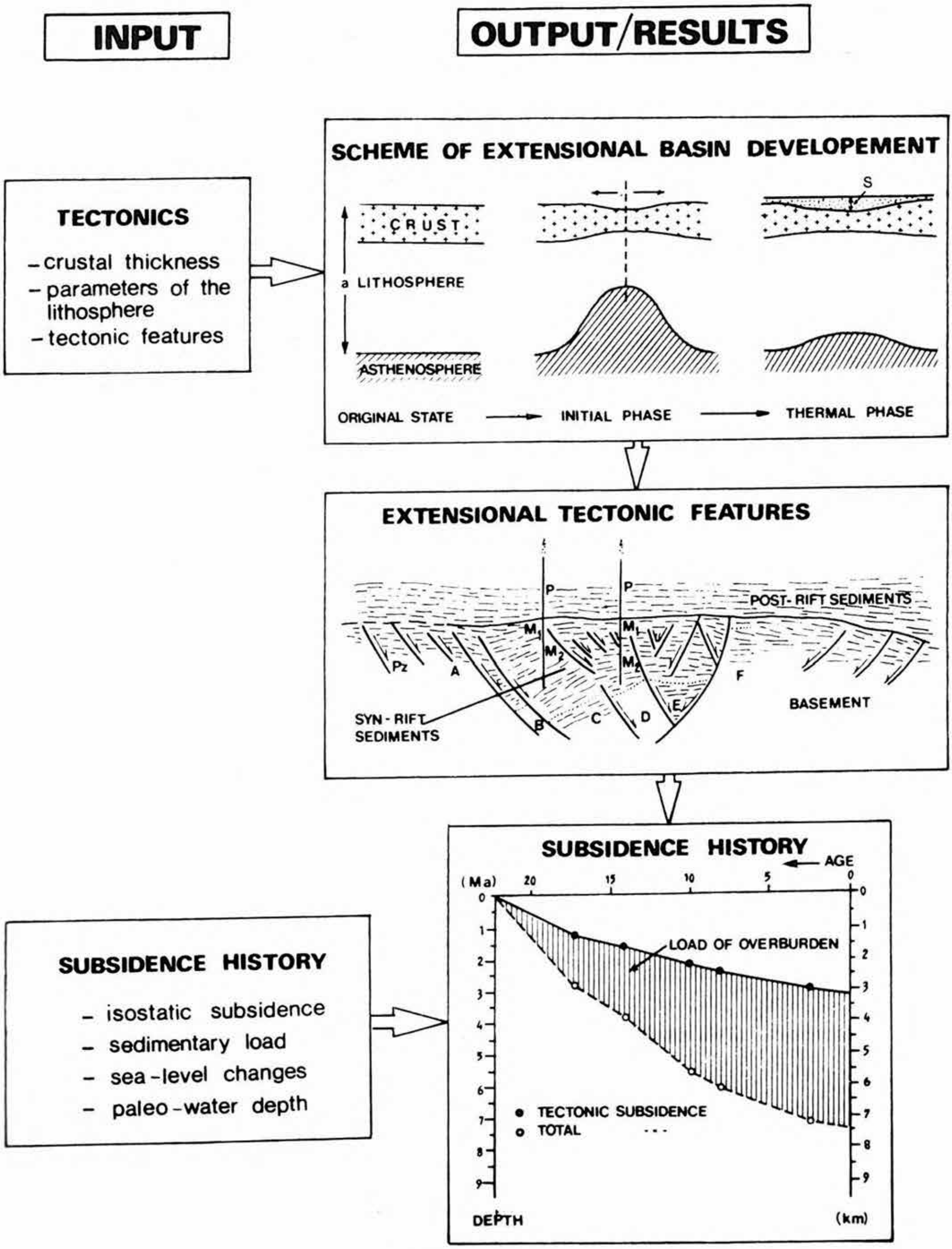


*Fig. 4.* Geodynamic model

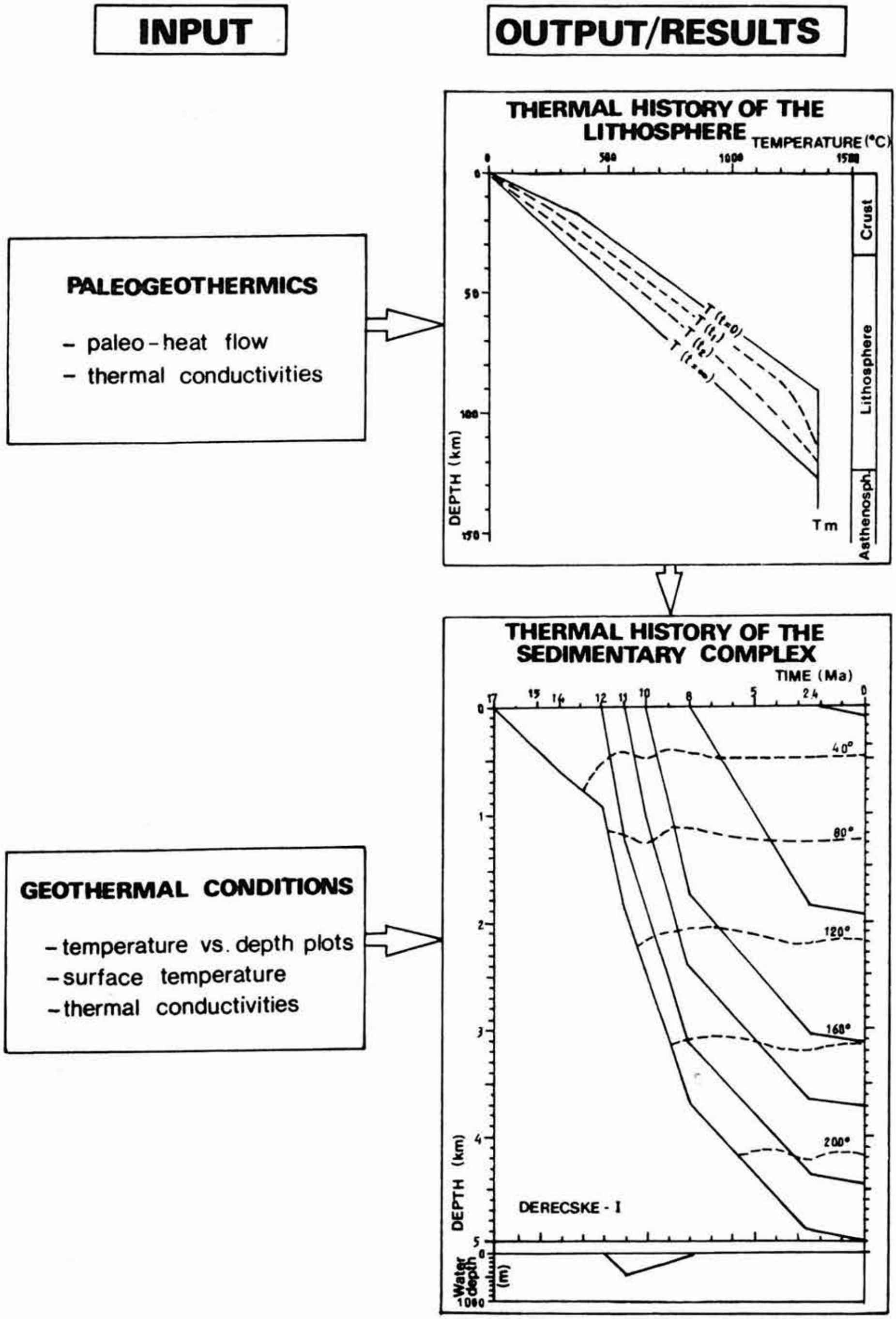


*Fig. 5.* Geothermal model

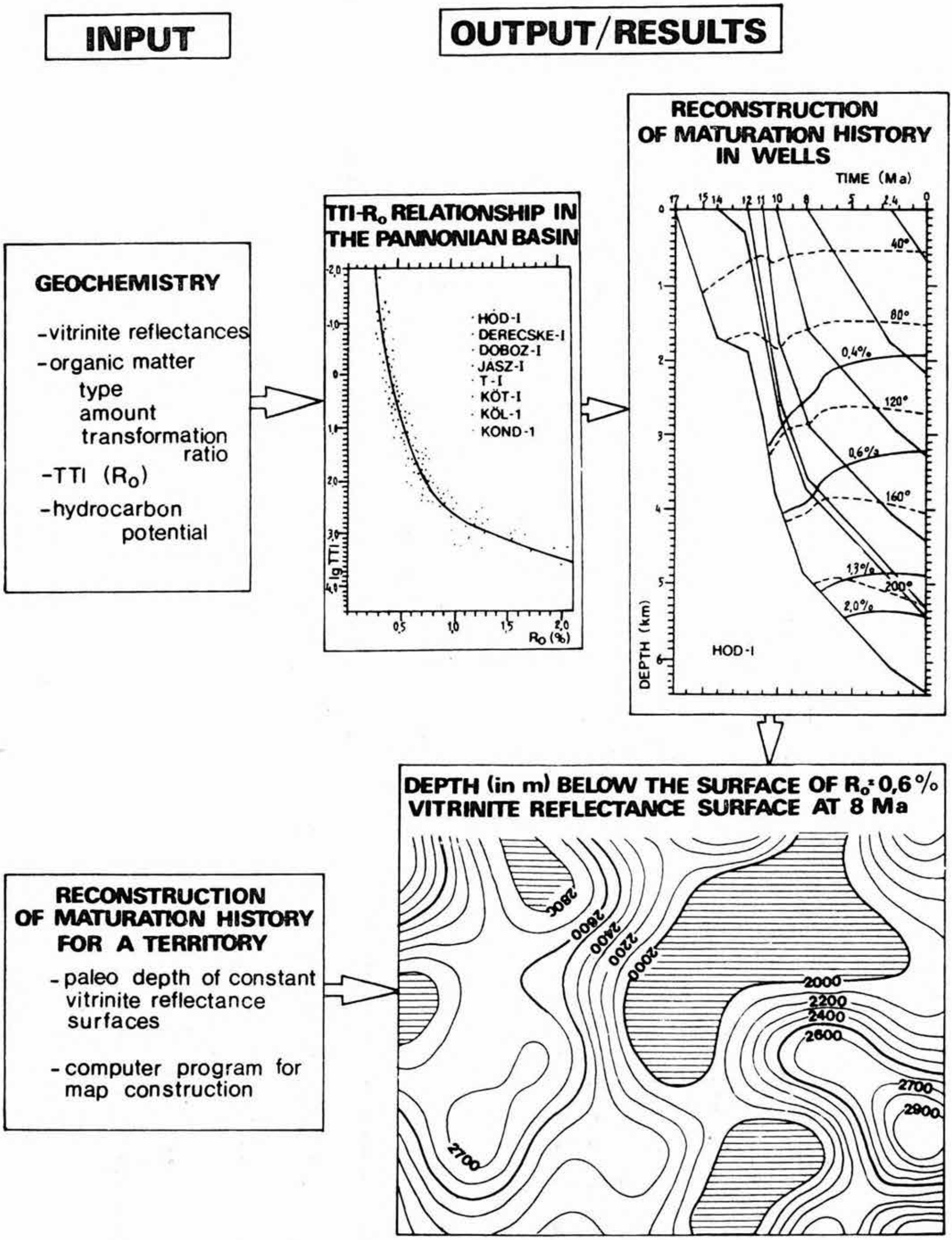


*Fig. 6.* Maturation model

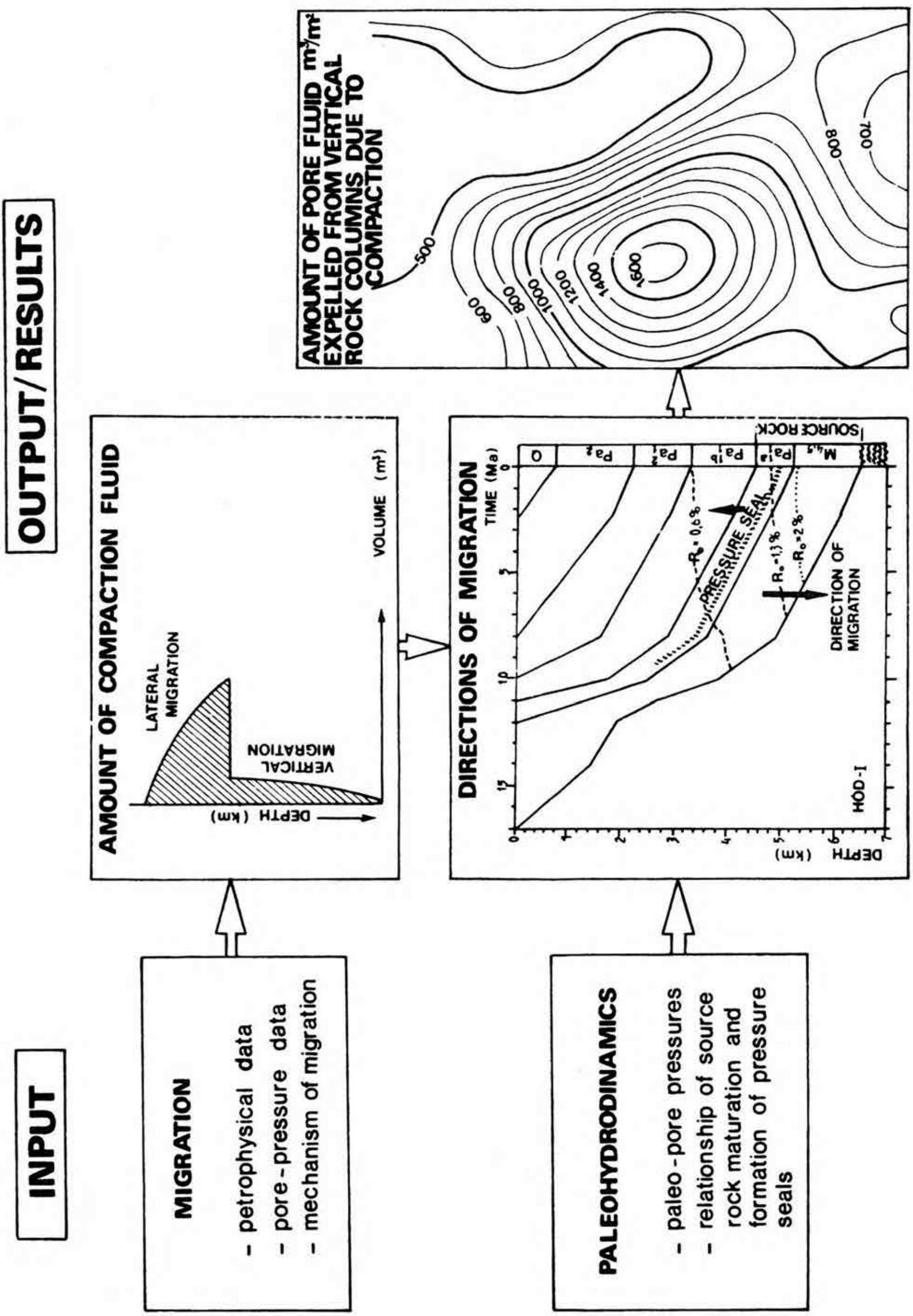


*Fig. 7.* Migration model

of the depositional history and can result in maps showing the palaeo-thicknesses of the different lithogenetic units. Furthermore, pore pressures and their change with time can also be predicted. It relies on the assumption, if that compaction is restricted the load of the overburden is partially supported by the fluid in the pore space. Accordingly undercompacted zones are characterized by overpressure and low effective stress. Palaeo-pressure maps can also be constructed, and it is often very useful to combine with paleo-thickness maps.

The next step of basin analysis includes the elaboration of the *Geodynamic model* of the basin (Fig. 4). Continental shelves and many inland sedimentary basins of the world are of extensional origin. Analysis of seismic sections and drillhole data usually enables the division of the basin fill into three major stratigraphic units: prerift, synrift and postrift sediments. Pre-to synrift sediments are often separated from the postrift sediments by a marked unconformity (often referred to as the "break-up" unconformity). If the time period of active rifting is known, a lithospheric extensional model can be worked out which predicts the subsidence history. Observed subsidence can be obtained by combining the depositional history with estimates about the palaeo-water depths. Unknown parameters of the lithospheric extension can be fixed by comparing the predicted and observed subsidence histories for a number of wells in a basin. Subsidence due to the load of sediments can be calculated by using local or regional isostatic compensation models.

A very important part of basin analysis is the *Geothermal model* (Fig. 5). Once the parameters of lithospheric extension are fixed temperatures and heat flow in the lithosphere can be calculated by numerical integration of the appropriate differential equation. If there is a good idea about the thermal conductivity of sedimentary rocks and its change with porosity and temperature, then the temperature and heat flow history of the basin fill can also be calculated. Present steady-state temperature profiles and heat flow determinations are good constraints on these model calculations.

The next step of basin analysis is the determination of transformation of organci matter by the *Maturation model* (Fig. 6). This is again a simulation in four dimensions by using the timetemperature history of the basin. The Lopatin-Waples relationship was accepted to calculate the maturity level. However, the concersion between Time—Temperature Index and vitrinite reflectance has been recalibrated. Geochemical data measured on drill cores combined with well log interpretation can identify the potential source rocks in a basin. Their maturation history can be followed during basin evolution and the amount of hydrocarbons generated can be estimated. It is often very useful to map the depth of onset of the oil generation window ($0.6\% \leq R_0 \leq 1.3\%$), or to show the position of source rocks relative to this window as a function of time.

The last step of basin analysis is the *Migration model* (Fig. 7) which attempts to trace the path of generated hydrocarbons from the source rocks up to the traps. Expulsion of hydrocarbons from the source rock and primary migration are thought to be a diphasic migration described by the Darcy's law through the relative permeability concept. Therefore determination of the pressure conditions for the time when source rocks are situated within the oilgeneration window are particularly important. Secondary migration takes place in carrier beds, and a certain amount of hydrocarbons can be trapped if adequate seals are available. Delineation of possible migration paths and analysis of the contemporaneous facies and structural conditions can help to predict the most probable location of traps which may contain producible hydrocarbons.

In summary, *Basin analysis* is a deterministic and predictive model which can significantly contribute to the reduction of exploration risk by offering the following informations:

i) better understanding of basin evolution,

ii) general evaluation of the hydrocarbon potential of a basin,

iii) forecasting of zones favourable for oil and gas accumulation,

iv) ranking of promising traps and an estimate of the volume of producible hydrocarbons,

v) predicting geologz and rock physics relevant to the planning of deep drillings.

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# MIOCENE HYDROCARBON RESERVOIRS AND POOLS IN EASTERN HUNGARY

by

M. SZENTGYÖRGYI

The location map shows the hydrocarbon fields where oil and gas accumulated in Miocene reservoirs (Fig. 1) of the fields 12 are located in Transdanubia and the rest in the Great Hungarian Plain. These fields represent nearly 10% of the proved hydrocarbon reserves. One fourth of Hungary's oil and gas resources can be found in multiple reservoirs hydrodynamically connected with Miocene rocks. This paper deals only with that part of the country which is east of the river Tisza (Fig. 2).

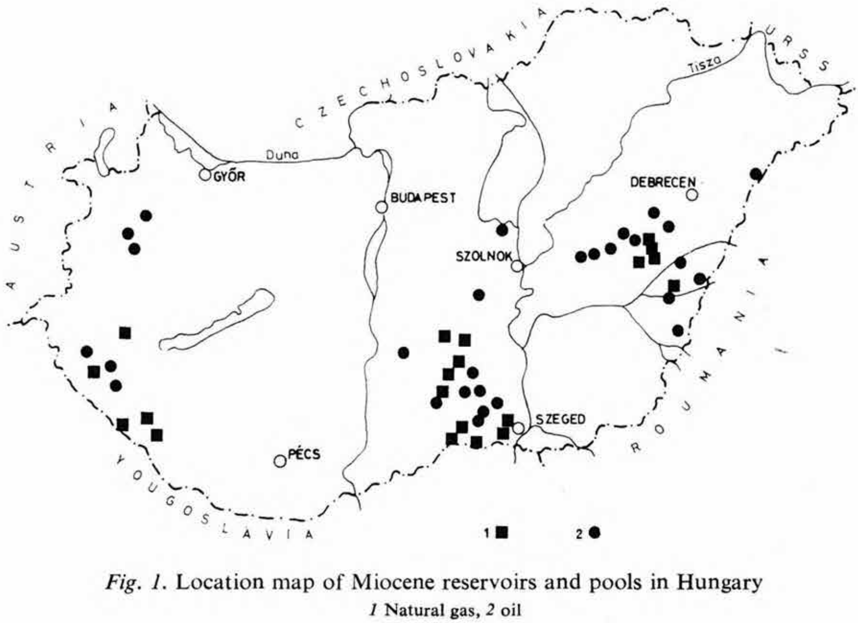


*Fig. 1.* Location map of Miocene reservoirs and pools in Hungary
*1* Natural gas, *2* oil

On this territory there are five Neogene basins, the borders of which where drawn along the negative watersheds. In these basins out of 68 hydrocarbon occurrences 25 reservoirs consist of Miocene rocks or are hydrodynamically connected with Miocene formations. 22% of the discovered hydrocarbon resources can be found in them. The most significant occurrences are the pools at Hajdúszoboszló and Szeghalom.

The Miocene formations rest unconformably on the surface of the Paleozoic, Mesozoic and Paleogene basement, respectively. Hydrocarbon accumulated in multiple reservoirs and rarely bed reservoirs. The impermeable seal rocks are the Lower

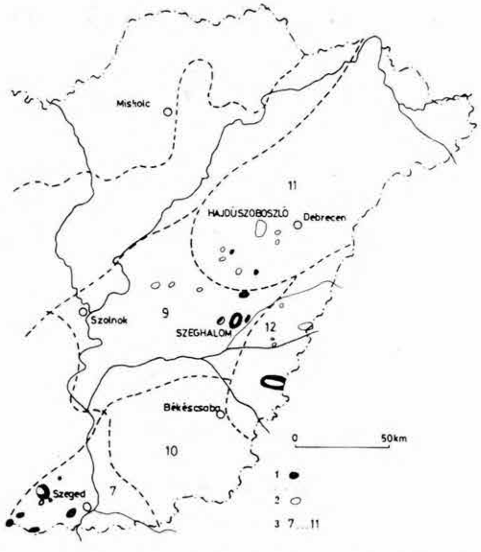


*Fig. 2.* Location map of the pools dealt with

*1* Oil, *2* natural gas, *3* 7—11 area of Neogene partial basins

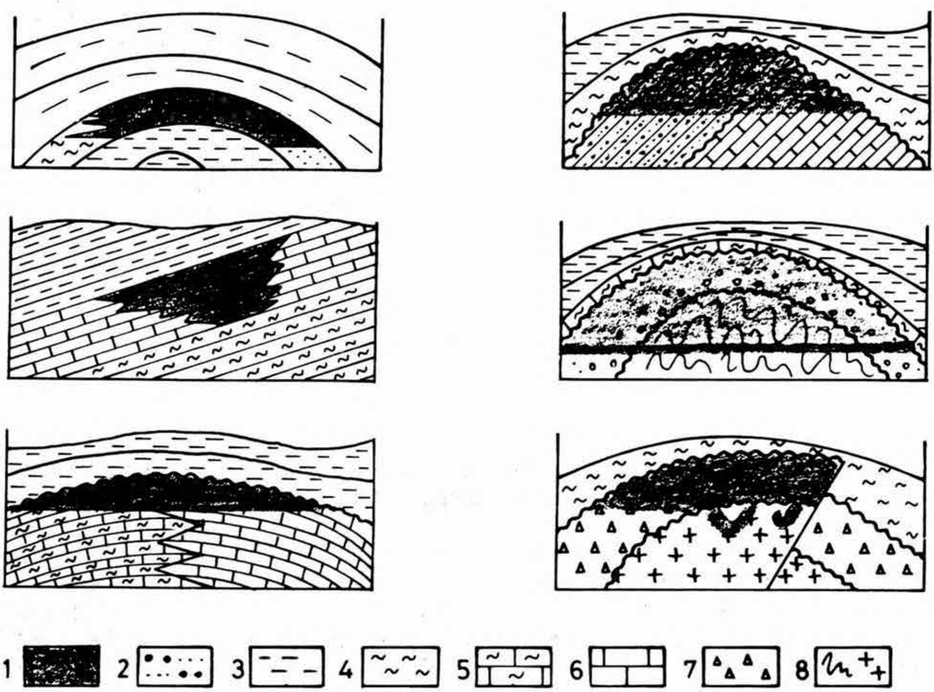


*Fig. 3.* Different types of traps

*1* Fluid content (hydrocarbon), *2* sandstone, *3* clayey marl, *4* marl, *5* calcareous marl, *6* limestone, *7* breccia, *8* metamorphite

Pannonian calcareous marls and marls. The traps are due to either lithological or permeability changes or by truncation and sealing. However the combination of structural and stratigraphic features constitutes the most important type (Fig. 3).

Even the bed reservoirs are lithologically and petrophysically varied. This type of reservoirs can be found in 14 fields of the 39 beds of this pools 27 stone free gas (Fig. 4). Coarse-grained reservoirs (Miocene breccias, conglomerates, sandstones) are the most common. Another important reservoir type consists of heterogeneous sediments: alternation of fine to medium grained detrital sediments, with slightly marl-shale and limestone interbedding. The carbonate reservoir rocks are mostly lithothamnian or oolitic limestones and calcareous marls. The Neogene volcanic complex (agglomerates, tuffs) is not important as reservoir.

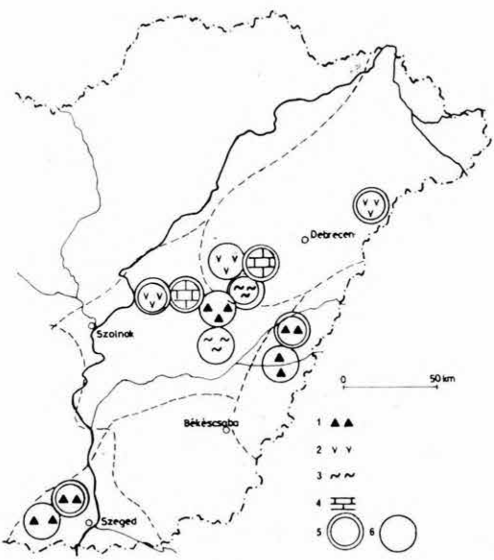


*Fig. 4.* Reservoir rocks of bed reservoirs
*1* Coarse-grained sediments, *2* volcanic rocks, *3* heterogeneous sediments, *4* carbonate rocks, *5* natural gas, *6* oil

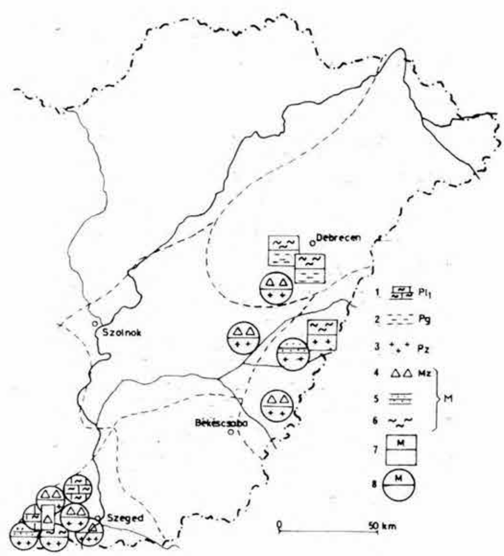


*Fig. 5.* Rocks of multiple reservoirs (with the age of rocks)
*1* Calcareous marl, *2* flysch, *3* metamorphite, *4* breccia, *5* sandstone, *6* lithologically heterogeneous sequence, *7* natural gas, *8* oil

Oil and gas accumulated in 17 multiple reservoirs of 14 fields in the Great Hungarian Plain. In some cases the hydrocarbons come partly from fractured basement rocks (two-porosity system). Fig. 5 shows the dominant fluid content and the age of the storing rocks. Lower Pannonian and Miocene calcareous marls contain some oil only in one pool in the southern Great Hungarian Plain. The most significant free gas resources are stored in the sandstone matrix and narrow fractures of the Palaeogene flysch sequence and they are connected with the overlying Miocene carbonate formations.

At one place there is hydrodynamic connection between the Triassic fractured dolomite and the Miocene breccia. The most important is the combination of Miocene reservoir rocks with Precambrian metamorphites (gneis, amphibolite, granite). The basement rock stores in the fractures (Fig. 6).

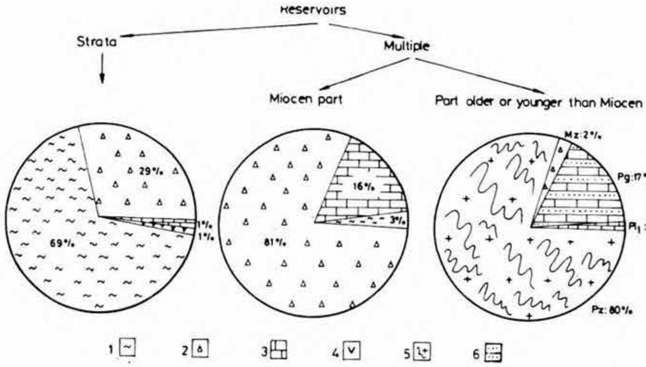


*Fig. 6.* Distribution of reserves in the study area

*1* In heterogeneous rocks, *2* in coarse-grained rocks, *3* in carbonate rocks, *4* in volcanic rocks, *5* in metamorphites, *6* in sandstones

The petrophysical properties of Miocene reservoirs are rather varied. The average matrix porosity of the rocks is between 5.1–18.6%. The extreme porosity values measured on core samples are the following:

– coarse detrital sequence of reservoirs caracterized by 2.5–15.9%. The maximum is 29.6%.

– porosity of limestones is 5.0–19.3% (max. 29.6%),

– average porosity of volcanics 6.0% (max. 16.0%),

– shale porosity is 1.1–8.2% (average is 5.1%).

The Szeghalom pool developed mostly in Miocene reservoir rock is one of the most significant gas-cupped oil pools. The Miocene reservoir consists of sandstone and conglomerate (Fig. 7). Measured average porosity is 12.9% in reservoir part.

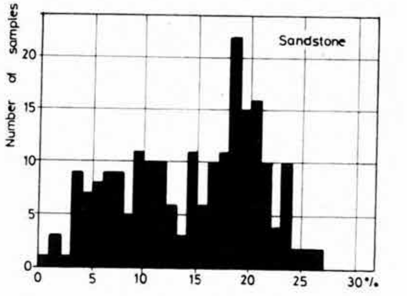


*Fig. 7.* Distribution of porosity in Miocene reservoir rocks pool of Szeghalom (data provided by the testing of core samples)

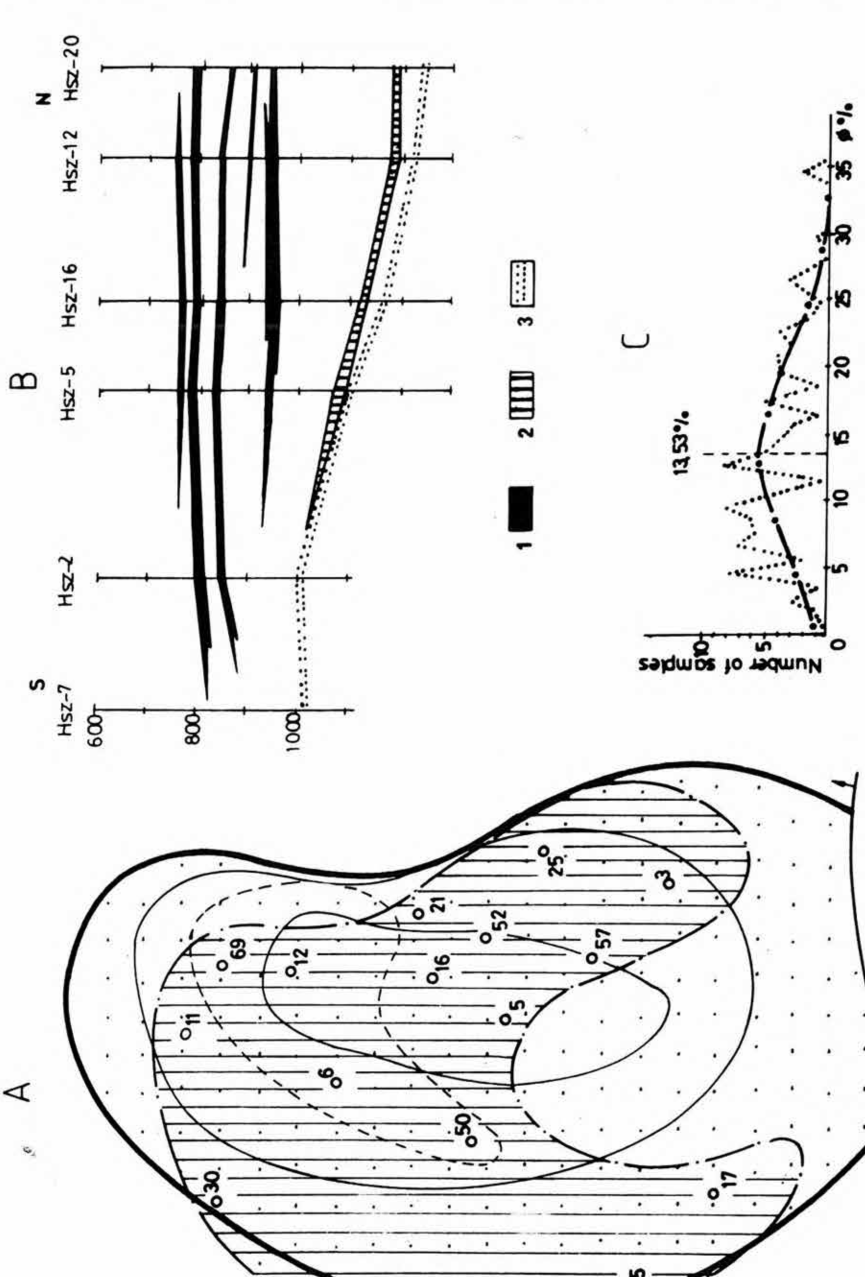


*Fig. 8.* Geological sketch of pool Hajdúszoboszló

A: Extension and position of Miocene carbonate reservoir rocks in the pool.—B: Position of natural gas reservoirs.—*1* Pannonian sandstone reservoirs, *2* Miocene oolitic limestone reservoir, *3* Paleogene flysch reservoir.—C: Distribution of porosities in Miocene oolitic limestones

An important free-gas pool can be found at Hajdúszoboszló in the Miocene limestone and the Paleogene flysch series (Fig. 8). The flysch complex is made up by alternating layers of sandstones, shales and siltstones; porosity is due to fracturing caused by faulting. The Miocene reservoir part is oolitic limestone; its porosity is due to oolitic texture with partially cemented interstices. Average porosity of the limestone is 13.5%. Calculated average permeability for the total reservoir range between: 10 and $843 \cdot 10^{-3}$ $\mu m^2$. Estimated values of connate water saturation are between 20–75%.

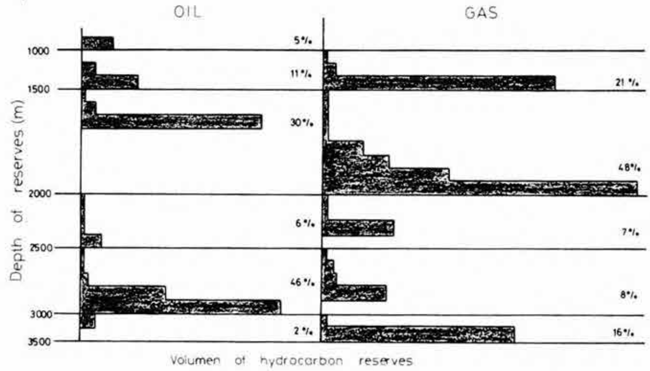


*Fig. 9.* Distribution of oil and gas reserves versus depth

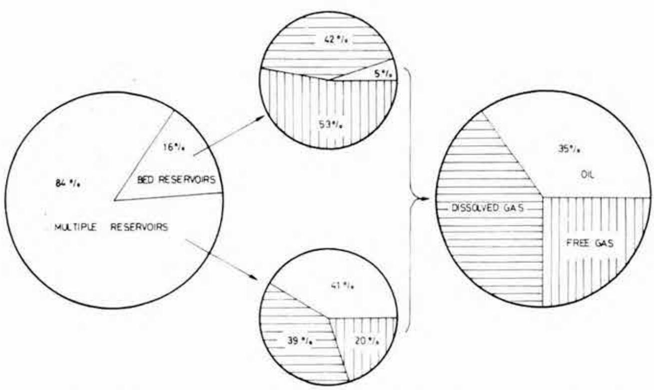


*Fig. 10.* Distribution of hydrocarbons in the two types of reservoirs

Approximately the half of oil resources in the Miocene reservoirs of the Great Plain area is to be found in the interval between 2500–3000 m and 30% between 1500–2000 m (Fig. 9). Half of the natural gas resources are also bound to the same intervals. Only 16% of resources of partly or totally Miocene reservoirs is contained in bed reservoirs. However, only 5% of their total hydrocarbon volume is oil. 41% of the resources stored in multiple reservoirs is oil combined. 35% of the hydrocarbons to be found in these two types of reservoirs is oil (Fig. 10). On the basis of hydrocarbon prediction prospectivity of the combined Neogene and pre-Neogene structures has been increased. Accordingly, the importance of Miocene reservoirs and their contribution to the hydrocarbon reserves of Hungary has increased as well.

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# NEOGENE EFFECTS ON THE HUNGARIAN BAUXITE DEPOSITS

by

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On the southern flank of the "Central Range Bauxite Belt" in Hungary, bauxite deposits are known at the contacts of Upper Triassic and Upper Cretaceous, Upper Triassic and Eocene and between Upper Cretaceous and Eocene rocks. These deposits occur on the southern margins of the one-time Late Cretaceous and Eocene sedimentary basins. Out of these, first of all the Eocene-covered ones are associated with basins, in which the primary overburden near the margin was partially or totally removed by infra-Oligocene denudation. Beside these, there may have been deposits which were formed beyond the farthest transgression shoreline, now unidentifiable, and which even originally were not covered by sediments. These deposits and the ones that had been exposed by the infra-Oligocene denudation could be preserved or reworked in varying measure owing to the preserving tendency of the lithoclases and to the intensity of denudation. After the Oligocene, owing to the Savian and then to the Styrian orogenic phases a gradual uplift of the one-time sedimentary basin of the Transdanubian Central Range and a subsidence of the surrounding areas set in. This movement on the southern rim took place along the so-called "Balaton Line". At the beginning of the Miocene, the foreland of the Transdanubian Central Range was already a landmass. This time, the overlying Eocene rocks in the marginal zone and where present their Oligocene counterparts and subsequently the bauxite itself were repeatedly attached by erosion. At some of the Nyirád bauxite deposits this process can be traced in time, up to the appearance of the bauxite.

The Miocene sea could reach the Central Range around Várpalota only in the Ottnangian. Transgression, over a larger area took place at the beginning of Badenian. Parallel to marine sedimentation in the coastal foredeep (at Fehérvárcsurgó) terrestrial—lacustrine sedimentation, while in other places, terrestrial accumulation took place. The reworking of the bauxites, which had partly, or later totally, been deprived of its overburden, and the accumulation of Miocene- or Pliocene-covered bauxite in the southeastern foreland of the Central Range, must have taken place during this period, prior to the Badenian. The polymictic bauxite-scree in a pelitomorphic matrix, the local abundance of quartz and carbonate clastics in the deposits, the frequency of lamination and the low grade of the bauxite prove the reworking of bauxite formed before. The overlying beds of these deposits show a diversified lithology, varying from Badenian to Upper Pannonian in age. Strata of littoral facies, like the Badenian "Fertőrákos Limestone Formation" at Ódörögd, or the Upper Pannonian "Kálla Gravel Formation" at Öcs, which are immediately superimposed on the local bauxite bodies, show an unconformity between bauxite and overlying beds and sand or pebble mixtures with bauxitic material at the top of the bauxite sequence. The reduction of bauxite grade was caused by mechanical contaminations (quartz and sand) from the abrasion of Badenian layers also to some of the bauxite primarily Eocene-covered deposits now outcropping in denudational windows around Nyirád–Izamajor. Certain alteration of bauxite could be observed under basalts, too. Neogene effects on bauxite deposits will be discussed later in detail in the context of the Nyirád–Nagytárkány, Öcs and Iszkaszentgyörgy deposits (Fig. 1).

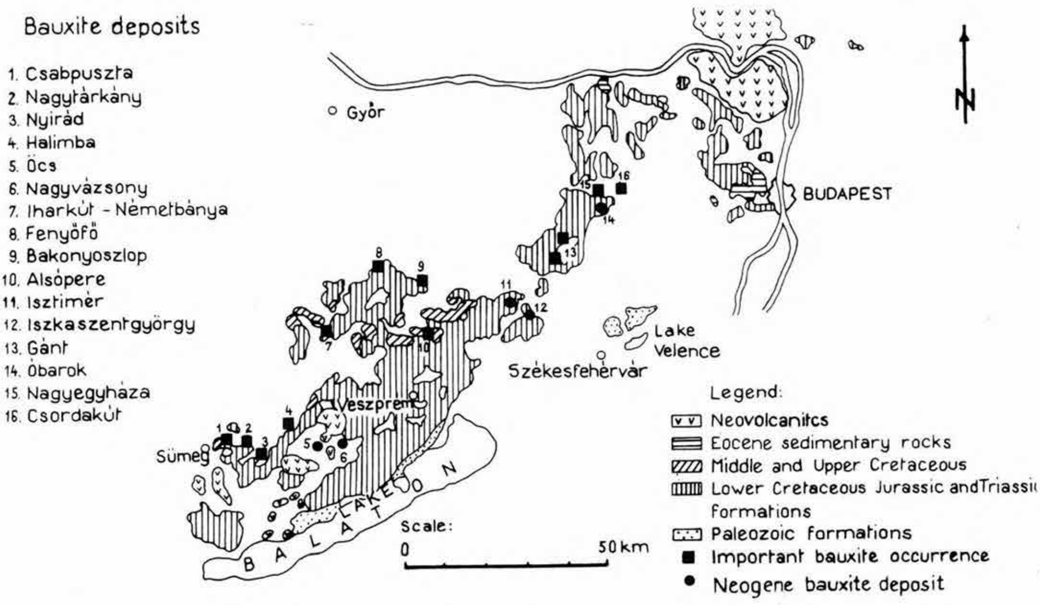


*Fig. 1.* Bauxite deposits in the Transdanubian Central Range

Most of the bauxite bodies at the Nyirád–Nagytárkány deposit are filling karstic dolinas in the Carnian–Norian Hauptdolomit overlain by Middle Eocene or, in places, by Lower Eocene layers.

The Triassic basement has a gentle dip to the N, so that a thicker and thicker overburden covers the bauxite in this direction, and the profile ends with the Csatka Formation of Oligocene age. In the southern part of the deposit, the Eocene is pinching out owing to denudation. Merely the oldest members of the Eocene complex have remained intact in tens of metres in thickness, to pinch out farther on. A few hundred m N of the pinching-out zone, denudational windows occur, in which denudation has reached down to the Triassic basement. This was partly due to the rough pre-Eocene basement surface and partly to tectonic movement.

Above the Eocene and older rocks , Lower Badenian gravel, sand and limestone layers can be found, or variegated clay and coaly clay are locally developed at their base. Sedimentation in the Miocene begins as result of the Late Styrian orogenic phase. This time there was certainly a rejuvenation of the older tectonical lines, but these movements could not be definitely identified owing to the lack of the Oligocene and the Lower Miocene layers. The distribution and partly the form of these denudational windows without an Eocene cover probably seem to have been delineated by earlier tectonical movements of the uplifted blocks, prior to their denudation.

At the beginning of the Badenian sedimentation, bauxite deposits of removed cover began to be reworked. Only the upper part of the bauxite body was reworked in the case of bauxite lenses of deeper position, while the bauxites of higher position were reworked totally. Reworking is proved by the presence of sand and gravel intercalations of overburden provenience even in bauxite bodies of commercial grade, as is obvious from our geological profile (Fig. 2). (Only the upper part of the ore is reworked in Darvastó lens N° XVIII., while the material of lens N° XIX is totally reworked.) The lower quantity of sandy-micaceous material has resulted in less reduction in grade. In case of total denudation, sandy clay of bauxitic origin and/or variegated clay were formed in the place of the entire commercial ore body or its upper part. The commercial ore body and the degraded clay of bauxitic origin are often intertongued.

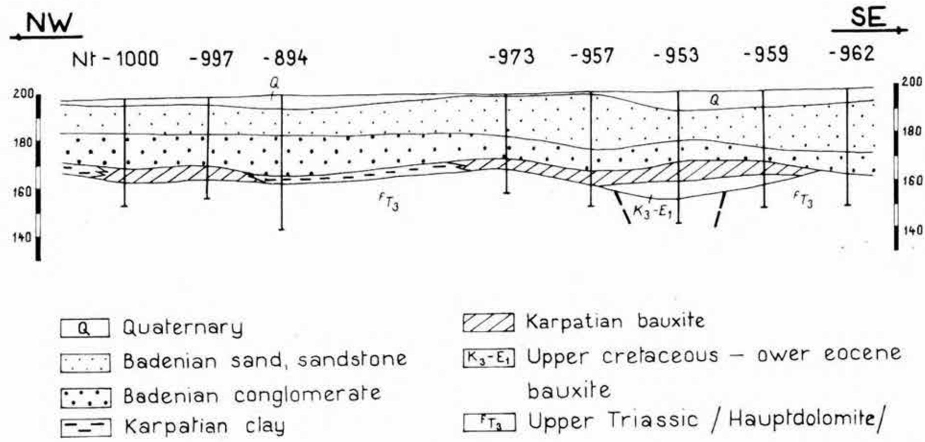


*Fig. 2.*

Within the extension of Eocene formations in the so-called "denudational windows", the bauxite is usually characterized by a relatively high grade, a low degree of degradation and generally a slight reworking being restricted only to the upper part of the sequence. The ore grade here is the same or only slightly lower than in the case of the Eocene-covered ore, in spite of its mixed material combined with macroscopic sand and gravel grains of quartz. In parts of the deposit the ore grade, with a modulus 12–13, exceeds the average.

South of the limit of extension of the Eocene, the bauxite deposits are characterized by a stronger denudation and reworking of the whole complex. Consequently, an alteration of different-grade bauxites is very common, resulting in lower average grade of the complex (an average of 2–5 in terms of modulus).

Farther S in what was probably an emergent land in the Miocene, the Upper Pannonian so-called Kállai Gravel Formation covers the bauxitic rocks, containing often quartz-sand and pebble and/or, sometimes, dolomite-scree reaccumulations. These layers are most often of red clay or of bauxitic clay quality, with 24–40% $Al_2O_3$, 30–50% $SiO_2$, and with a modulus of 0.4–1.4. Their repeated reaccumulations in the Pannonian and their resulting greater degradation are marked by the afore-mentioned low grade.

High-grade bauxite deposits do occur of course even just beneath the overlying Pannonian beds, like they occur under the Miocene layers too, preserved by morphological depressions of their original setting and thus affected by only partial removal, though their grade reduction trend is generally felt.

Bauxitic rocks resulting from secondary or multiple redeposition can be locally identified by their textures. Contrary to the Eocene-covered bauxites of pelitomorphic or often pizoidal texture, and of conchoidal, granular or earthy-like fracture, they are of small-grained, fractured type, being densely intersected by sliding paths and planes. Their basic material is pelitomorphic or pelitomorphic—micrograined, including iron-poor and iron-rich zones and showing chaotically folded or fluidal structures. These bauxites contain coarse-grained bauxitic scree material in various quantities, which are either iron-rich or kaolinitized to some extent. They contain fine sand grains in various amount in places, or small carbonaceous grains in several cases.

Degradation usually takes place in form of decomposition of allitic minerals. Originally Eocene-covered bauxites are predominantly composed of boehmite with a small quantity of gibbsite. These two allitic minerals are characteristic of partially reworked Miocene-covered bauxites as well, but they contain, in addition, up to 10—30% kaolinite. Totally reworked bauxites are predominantly kaolinitic, though minor amounts of boehmite and gibbsite may also be present in them. The aluminium content of the Pannonian red clays is invariably bound to kaolinite.

Several kilometres farther S, SE—E of the Halimba—Szőc bauxite deposit, in the Öcs, Nagyvázsony and Kab Mts areas, a red clay sequence, discontinuous, but of significant extension, is developed. This overlies the Hauptdolomit, or in places in the Kab Mts area, it rests on the surface of Middle Eocene limestones. Its oldest overburden is the Öcs Limestone Formation of Badenian—Sarmatian age, consisting of mottled clays, limestones and calcareous marls of lacustrine facies, intertonguing with the upper part of the red clay. On the basis of the afore-mentioned stratigraphical evidence the age of accumulation of the red clay sequence can be identified as Karpatian. The bauxitic origin of the complex is out of question, evidence as proved by the presence of pebbles, coarse bauxite grains and bauxite lenses within the clay. The complex is characterized by 25—44% $Al_2O_3$, and equal amounts of $SiO_2$, and with an alumina to silica ratio of 0.5—1.5 on the average, but bauxitic parts with a modulus of $>2$, or exceptionally even of commercial grade (with a modulus $>5$) do also occur in places. Mineralogically, the higher-grade bauxites contain 20—50% gibbsite, 0—13.3% boehmite, and up to 26—60% kaolinite, while the lower-grade parts predominantly consist of kaolinite, with illite, but without any allite minerals. Over the most part of the study area the red clays are overlain by Upper Panonnian sediments.

The "Kálla Gravel Formation" may also rest on the red clays in some places. According to field-observations, here the contact between overburden and red clay underneath is abrasional, with a reworking of quartz pebbles and a substantial grade-reduction. Such circumstances cannot be observed at the contact of the sequence of the "Taliándörögd Marl Formation" (facies of an ephemeral swampy lagoon). At one site, in the red clay overlain by the Kab Mt basalt and basaltic tuff the clay is observed to be much darker red in color than elsewhere, because of the effect of the superimposed basalt-lava flows. In other places there were basalt lapillis or debris in the upper part of the sequence.

The footwall of the Iszkaszentgyörgy bauxite deposit consists of Upper Triassic rocks, predominantly of dolomite. Primary bauxite can be found more often beneath Middle Eocene neritic sedimentary deposits (Fig. 3). There the bauxites are of the highest grade and the deposits have mostly a northeastearly dip. The post-Eocene Pyrenean orogeny mostly resulted in fractures of NE—SW direction, thus determining the preservation and, consequently the distribution of the bauxites formed before.

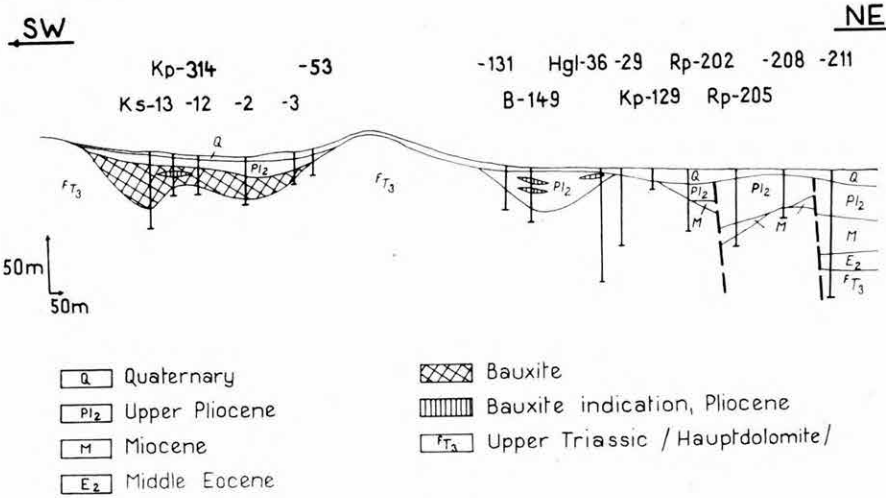


*Fig. 3.*

In post-Eocene time, the Eocene formations the Eocene-covered dolomite and the marginal parts of the bauxite deposit were removed by erosion. This denudation, was followed by Karpatian—Sarmatian terrestrial—lacustrine sedimentation and terrestrial transport and accumulation of clastic material.

At present, the higher-grade ore deposits are situated near the surface, being covered by clay with dolomite scree. Deposits of medium depth are more common, however, being overlain by mottled clay. In general the Miocene-covered bauxite horizon is thinner than those of Eocene-covered one, but its $Al_2O_3/SiO_2$ content indicates that it still belongs to the bauxite complex: its average modulus is 1.0—2.

Higher-grade bauxite occurs under an overburden of dolomite debris while the clay-covered bauxites are usually characterized by a somewhat lower grade. This bauxitic material contains more bauxitomorphic debris and also carbonate scree.

Stryrian movements led to the rejuvenation of older faults. They are of little importance, however, because they resulted in small-scale morphological differences only. This led to the disintegration of the bauxite deposits of Eocene age.

During the Pannonian a slow ingression of the sea took place, redepositing and mixing up the clastics of the bedrock and of the formerly accumulated bauxite.

This complex has transgressed onto the "Csatka Formation" of Oligocene age, the Miocene mottled clay and, sometimes also, onto Eocene and Triassic sediments. However, the bauxite horizon rests almost exclusively on Triassic rocks. As a result of repeated reworking, the bauxite bodies were emplaced to their present stratigraphic position just before the Late Pannonian sedimentation. These bauxitic rocks are distributed over larger areas and are thicker than their Miocene-covered counterparts. Their grade shows a more uniform distribution, pay-ore bodies, however, are rather rare, being much thinner, than the Miocene-covered bauxite. Their average modulus is: 0.78. Therefore they can be considered only just as traces or indications. Mainly quartz debris, and also muscovite or more seldom, biotite, chlorite, zircon and other submicroscopical silicate debris can be found in them.

The ore-grade tends in the majority of the cases, to decline owing to the higher silica content. The ore contains more silica than alumina. Consequently, the effect of Neogene tectonic movements cannot be identified directly in Hungarian bauxites. The distribution of the reworked bauxitic rocks is usually controlled by the karstic morphology. Denudation and reworking of the original bauxite bodies, however, are dependent on the morphological conditions which are the results of tectonic deformation. Reworking resulted in degredation, petrological and mineralogical alterations and in deterioration of the grade as a consequence of the latter.

The Neogene palaeoclimates were unfavourable for further bauxite deposition. As a general phenomenon, the grade was reduced by repeated reworking of the bauxite overlain by younger and younger beds. The reason for that was, obviously, mechanical contamination.

Additional mechanical contamination resulted in a more efficient grade reduction when the presence of overlying littoral deposits is an indication of abrasion. Denudation certainly resulted in a significant decrease of the tonnage of the original bauxite reserves. Unfortunately, no numerical data that would corraborate this suggestion are available.

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|---|---|---|---|---|---|---|
| *1* Algeria | 1 | — | — | — | — | 1 |
| *2* Austria | 6 | — | 2 | — | — | 8 |
| *3* Belgium | 1 | — | — | — | — | 1 |
| *4* Bulgaria | 2 | — | — | — | — | 2 |
| *5* Canada | 2 | — | — | — | 1 | 3 |
| *6* China | 2 | — | — | — | — | 2 |
| *7* Czechoslovakia | 16 | — | — | — | — | 16 |
| *8* Egypt | 5 | 1 | — | — | — | 6 |
| *9* Finland | 1 | — | — | — | — | 1 |
| *10* France | 14 | — | — | — | — | 14 |
| *11* F. R. Germany | 13 | — | — | — | 3 | 16 |
| *12* Greece | 11 | — | 11 | — | — | 22 |
| *13* Hungary | 139 | 4 | — | — | — | 143 |
| *14* India | 1 | — | — | — | — | 1 |
| *15* Indonesia | — | 1 | — | — | — | 1 |
| *16* Iran | 1 | — | — | — | — | 1 |
| *17* Ireland | 1 | — | — | — | — | 1 |
| *18* Israel | 4 | — | — | — | 1 | 5 |
| *19* Italy | 27 | 1 | — | — | 1 | 29 |
| *20* Japan | 2 | 3 | — | — | — | 5 |
| *21* Libya | 2 | 1 | — | — | — | 3 |
| *22* Maroc | 1 | — | — | — | — | 1 |
| 23 the Netherlands | 6 | — | 8 | — | — | 14 |
| *24* Poland | 13 | — | — | — | — | 13 |
| *25* Portugal | 1 | — | — | — | — | 1 |
| *26* Roumania | 5 | 2 | — | — | — | 7 |
| *27* Spain | 16 | 1 | 2 | 3 | — | 22 |
| *28* Switzerland | 5 | 1 | — | — | 1 | 7 |
| *29* Turkey | 5 | — | — | 1 | — | 6 |
| *30* United Kingdom | 7 | 1 | — | — | 2 | 10 |
| *31* USA | 15 | 4 | — | — | 2 | 21 |
| *32* USSR | 21 | — | — | — | 2 | 23 |
| *33* Yugoslavia | 24 | 1 | — | — | — | 25 |
| Non-member participants and guests | | | | | | 48 |
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195 DE GREIFF STR.
4150 KREFELD

HERRMANN AXEL
GEBR. KNAUF WESTD. GIPSWERKE
BAHNHOF STR 10
8715 IPHOFEN

KUSTER HANS
NIEDERS. LANDESAMT F. BODENF.
STILLEWEG 2
3000 HANNOVER 51

KUSTER-WENDENBURG ELISABETH
ÜBERSEE MUSEUM IN BREMEN
BAHNHOFSPLATZ 13
2800 BREMEN

MOORKENS THIERRY
DEMINEX
GRAF BERNADOTTE STR. 70.
4300 ESSEN—1

SCHMIDT-KITTLER N.
J. GUTTENBERG UNIVERSITÄT
INST. FÜR GEOWISSENSCHAFTEN
SAAR STR. 21
POSTFACH 3980
6500 MAINZ

TOBIEN HEINZ
UNIVERSITY OF MAINZ
SAAR STR. 21
P.O. BOX 3980
6500 MAINZ

## GREECE

DERMITZAKIS MICHAEL
ATHENS UNIVERSITY
PANEPISTIMIOPOLIS-ZOGRAFOU
15784 ATHENS

FERMELI GEORGIA
NATIONAL TECHNICAL HIGHSCHOOL
42 PATISION
10682 ATHENS

FEORGIADES-DIKEOULIA E.
ATHENS UNIVERSITY
PANEPISTIMIOPOLIS-ZOGRAFOU
15784 ATHENS

IOAKIM CHRYSSANTHI
IGME
70 MESSOGHION STR.
11527 ATHENS

KOUFOS GEORGE
UNIVERSITY OF THESSALONIKI
54006 THESSALONIKI

KOUTSOUVELI ANASTASIE
IGME
70 MESSOGHION
11527 ATHENS

MARCOPOULOU-DIACANTONI
ANASTASIA
DEPT. OF GEOLOGY-PALEONTOLOGY
ATHENS UNIVERSITY
MARATHONO DROMOU 92
15125 ATHENS-MAROUSSI

MIRKOU MARIA-RHEA
UNIVERSITY OF ATHENS
DINOCRATOUS 109 KOLONAKI
ATHENS

PAVLIDES SPYROS
UNIVERSITY OF THESSALONIKI
54006 THESSALONIKI

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ATHENS UNIVERSITY
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15784 ATHENS

SYMEONIDIS NIKOLAUS
ATHENS UNIVERSITY
PANEPISTIMIOPOLIS-ZOGRAFOU
15784 ATHENS

VAKOPOULOU VASSILIKI

## HUNGARY

ALBU ISTVÁN
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

ÁDÁM OSZKÁR
KFH
1251 BUDAPEST PF. 22
ISKOLA U. 19—27

ÁRVA-SÓS ERZSÉBET
ATOMKI
4001 DEBRECEN PF. 51
BEM TÉR 18/C

BAKSA CSABA
OÉÁ
1406 BUDAPEST PF. 34
NÉPKÖZTÁRSASÁG ÚTJA 126

BALLA KÁLMÁN
KŐOLAJKUTATÓ VÁLLALAT
5100 SZOLNOK
MÉSZÁROS LAJOS U. 28

BALLA ZOLTÁN
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

BALOGH KADOSA
ATOMKI
4001 DEBRECEN PF. 51
BEM TÉR 18/C

BALÁZS ENDRE
SZKFI
2443 SZÁZHALOMBATTA PF. 32

BALÁZS-VARGA MÁRIA
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

BARDÓCZ BÉLA
KŐOLAJ ÉS FÖLDGÁZBÁNYÁSZATI V.
8801 NAGYKANIZSA PF. 126

BARTKÓ LAJOS
1203 BUDAPEST
RÁKÓCZI ÚT 51

BÁLDI TAMÁS
ELTE
1088 BUDAPEST
MÚZEUM KRT. 4/A

BÁLDI-BEKE MÁRIA
MÁFI
1442 BUDAPEST PF. 106

BENCE GÉZA
MÁFI
1442 BUDAPEST PF. 106

BÉRCZI ISTVÁN
SZKFI
2443 SZÁZHALOMBATTA PF. 32

BÉRCZI-MAKK ANIKÓ
SZKFI
2443 SZÁZHALOMBATTA PF. 32

BODOR ELVIRA
MÁFI
1442 BUDAPEST PF. 106

BIHARI GYÖRGY
OÉÁ
2085 PILISVÖRÖSVÁR

BOHN PÉTER
MÁFI
1442 BUDAPEST PF. 106

BOHN-HAVAS MARGIT
MÁFI
1442 BUDAPEST PF. 106

BREZSNYÁNSZKY KÁROLY
KFH
1251 BUDAPEST PF. 22
ISKOLA U. 19—27

CHIKÁN GÉZA
MÁFI
1442 BUDAPEST PF. 106

CSEH-NÉMET JÓZSEF
OÉÁ
1406 BUDAPEST PF. 34
NÉPKÖZTÁRSASÁG ÚTJA 126

CSERNY TIBOR
MÁFI
1442 BUDAPEST PF. 106

CSILLAG JÁNOS
OÉÁ KUTATÓ ÉS TERMELŐ MŰVEI
3301 EGER PF. 39
KERTÉSZ U. 128

CSILLAG-TEPLÁNSZKY ERIKA
MÁFI
1442 BUDAPEST PF. 106
NÉPSTADION ÚT 14

DANK VIKTOR
KFH
1251 BUDAPEST PF. 22
ISKOLA U. 19—27

DEÁK JÓZSEF
VITUKI
1095 BUDAPEST
KVASSAY J. U. 1

DÖVÉNYI PÉTER
ELTE-GT
1083 BUDAPEST
KUN BÉLA TÉR 2

DUDICH ENDRE
MÁFI
1442 BUDAPEST PF. 106

FODOR-NAGY PIROSKA
KFH
1251 BUDAPEST PF. 22
ISKOLA U. 19—27

FRANYÓ FRIGYES
MÁFI
1442 BUDAPEST PF. 106

FÜLÖP JÓZSEF
ELTE REKTORI HIVATAL
1364 BUDAPEST PF. 109
EGYETEM TÉR 3

GEIGER JÁNOS
SZKFI
6701 SZEGED PF. 30

GERZSENY ISTVÁN
KŐOLAJ ÉS FÖLDGÁZBÁNYÁSZATI
VÁLLALAT
8801 NAGYKANIZSA PF. 126
SZABADSÁG TÉR 22

GRASSELY GYULA
JATE
6722 SZEGED
EGYETEM U. 2

GRÓNAY ISTVÁNNÉ
SZKFI
2443 SZÁZHALOMBATTA PF. 32

GYARMATI JÁNOS
KŐOLAJ ÉS FÖLDGÁZBÁNYÁSZATI V.
8801 NAGYKANIZSA
SZABADSÁG TÉR 22

GYARMATI PÁL
MÁFI
1442 BUDAPEST PF. 106

HABLY LILLA
TERMÉSZETTUDOMÁNYI MÚZEUM
NÖVÉNYTÁRA
1476 BUDAPEST PF. 222

HAHN GYÖRGY
MTA FÖLDRAJZTUD. KUT. INTÉZET
1388 BUDAPEST PF. 64
NÉPKÖZTÁRSASÁG ÚTJA 62

HAJDÚ DÉNES
KŐOLAJKUTATÓ VÁLLALAT
5100 SZOLNOK
MUNKÁSŐR U. 43

HAJÓS MÁRTA
MÁFI
1442 BUDAPEST PF. 106

HALMAI JÁNOS
MÁFI
1442 BUDAPEST PF. 106

HÁMOR GÉZA
MÁFI
1442 BUDAPEST PF. 106

HERMESZ MIKLÓS
NÓGRÁDI SZÉNBÁNYÁK
3101 SALGÓTARJÁN PF. 124
FELSZABADULÁS U. 44

HORVÁTH FERENC
ELTE-GT
1083 BUDAPEST
KUN BÉLA TÉR 2

HORVÁTH RÓBERT
OKGT
1502 BUDAPEST PF. 22
SCHÖNHERZ Z. U. 18

JANKOVICH ISTVÁN
MÁFI
1442 BUDAPEST PF. 106

JASKÓ SÁNDOR
1122 BUDAPEST
PETHÉNYI KÖZ 4

JÁMBOR ÁRON
MÁFI
1442 BUDAPEST PF. 106

KAISER MIKLÓS
MÁFI
1442 BUDAPEST PF. 106

KASZAI-SELMECZI ILDIKÓ
MÁFI
1442 BUDAPEST PF. 106

KÁDÁR-JUHÁSZ GYÖRGYI
SZKFI
2443 SZÁZHALOMBATTA PF. 32

KEDVES MIKLÓS
JATE NÖVÉNYTANI TANSZÉK
6722 SZEGED PF. 657

KONCZ ISTVÁN
SZKFI
8801 NAGYKANIZSA PF. 126
SZABADSÁG TÉR 22

KOMJÁTI JÁNOS
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
GORKIJ FASOR 42

KOMLÓSI ZSOLTNÉ
SZKFI
2443 SZÁZHALOMBATTA PF. 32

KORDOS LÁSZLÓ
MÁFI
1442 BUDAPEST PF. 106

KORECZ ANDREA
MÁFI
1442 BUDAPEST PF. 106

KORECZ-LAKY ILONA
MÁFI
1442 BUDAPEST PF. 106

KORPÁS LÁSZLÓ
MÁFI
1442 BUDAPEST PF. 106

KORPÁS-HÓDI MARGIT
MÁFI
1442 BUDAPEST PF. 106

KÓKAI JÁNOS
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

KÓKAY JÓZSEF
MÁFI
1442 BUDAPEST PF. 106

KOVÁCS ANDRÁS
KŐOLAJKUTATÓ VÁLLALAT
5001 SZOLNOK PF. 85
MUNKÁSŐR U. 43

KOVÁCS ZSOLT
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

KÖRÖSSY LÁSZLÓ
1124 BUDAPEST
VAS GEREBEN U. 1

KŐVÁRY JÓZSEF
SZKFI
2443 SZÁZHALOMBATTA PF. 32

KRETZOI MIKLÓS
1024 BUDAPEST
LÖVŐHÁZ U. 24

KROLOPP ENDRE
MÁFI
1442 BUDAPEST PF. 106

KURUCZ BÉLA
KŐOLAJKUTATÓ VÁLLALAT
5900 OROSHÁZA
CSORVÁSI ÚT

KUTI LÁSZLÓ
MÁFI
1442 BUDAPEST PF. 106

LANTOS MIKLÓS
MÁFI
1442 BUDAPEST PF. 106

LELKES ÁKOS
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

LELKES GYÖRGY
MÁFI
1442 BUDAPEST PF. 106

LUKÁCS ANDREA
SZKFI
2443 SZÁZHALOMBATTA PF. 32

MAGYAR IMRE
ELTE ŐSLÉNYTANI TANSZÉK
1083 BUDAPEST
KUN BÉLA TÉR 2

MÁRTON GYULÁNÉ
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

MÁRTON PÉTER
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

MÁRTON-SZALAY EMŐKE
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

MÁTYÁS ERNŐ
OÉÁ HEGYALJAI MŰVEI
3909 MÁD

MÉSZÁROS LÁSZLÓ
KŐOLAJ- ÉS FÖLDGÁZBÁNYÁSZATI V.
8801 NAGYKANIZSA PF. 126
SZABADSÁG TÉR 22

MIHÁLY SÁNDOR
MÁFI
1442 BUDAPEST PF. 106

MINDSZENTHY ANDREA
ELTE ÁSVÁNYTANI TANSZÉK
1088 BUDAPEST
MÚZEUM KRT. 4/A

MOLNÁR KÁROLY
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
GORKIJ FASOR 42

MONOSTORI MIKLÓS
ELTE ŐSLÉNYTANI TANSZÉK
1083 BUDAPEST
KUN BÉLA TÉR 2

MÜLLER PÁL
MÁFI
1442 BUDAPEST PF. 106

NAGY BÉLA
MTA X. OSZTÁLY
1051 BUDAPEST
MÜNNICH F. U. 7

NAGY LÁSZLÓNÉ
MÁFI
1442 BUDAPEST PF. 106

NAGYMAROSI ANDRÁS
ELTE FÖLDTANI TANSZÉK
1088 BUDAPEST
MÚZEUM KRT. 4/A

PANTÓ GYÖRGY
MTA GEOKÉMIAI KUTATÓ INTÉZET
1112 BUDAPEST
BUDAÖRSI ÚT 45

PAPP SÁNDOR
KŐOLAJKUTATÓ VÁLLALAT
5001 SZOLNOK PF. 85
MUNKÁSŐR U. 43

PÉCSKAY ZOLTÁN
ATOMKI
4001 DEBRECEN PF. 51
BEM TÉR 18/C

POGÁCSÁS GYÖRGY
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
NÉPKÖZTÁRSASÁG ÚTJA 59

POLCZ IVÁN
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

PÓKA TERÉZ
MTA GEOKÉMIAI KUTATÓ INTÉZET
1112 BUDAPEST
BUDAÖRSI ÚT 45

RADÓCZ GYULA
MÁFI
1442 BUDAPEST PF. 106

RAVASZ CSABA
MÁFI
1442 BUDAPEST PF. 106

RAVASZ-BARANYAI LÍVIA
MÁFI
1442 BUDAPEST PF. 106

RÁDLER BÉLA
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
GORKIJ FASOR 42

RÁNER GÉZA
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

RÉVÉSZ ISTVÁN
SZKFI
6701 SZEGED PF. 30

RÓNAI ANDRÁS
MÁFI
1442 BUDAPEST PF. 106

RUMPLER JÁNOS
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
NÉPKÖZTÁRSASÁG ÚTJA 59

SÁG LÁSZLÓ
MTA FÖLDRAJZTUD. KUT. INTÉZET
1388 BUDAPEST PF. 64
NÉPKÖZTÁRSASÁG ÚTJA 62

SCHWEITZER FERENC
MTA FÖLDRAJZTUD. KUT. INTÉZET
1388 BUDAPEST PF. 64
NÉPKÖZTÁRSASÁG ÚTJA 62

SOLTI GÁBOR
MÁFI
1442 BUDAPEST PF. 106

SÜTŐ-SZENTAI MÁRIA
OFKFV LABORATÓRIUM
7300 KOMLÓ
KOSSUTH LAJOS U. 1

STRAUSZ LÁSZLÓ
8600 SIÓFOK
SZENT LÁSZLÓ U. 83

SZABADVÁRY LÁSZLÓ
ELGI
1440 BUDAPEST PF. 35

SZANYI BÉLA
GEOFIZIKAI KUTATÓ VÁLLALAT
1391 BUDAPEST PF. 213
GORKIJ FASOR 42

SZALAY ÁRPÁD
KŐOLAJKUTATÓ VÁLLALAT
5001 SZOLNOK PF. 85
MUNKÁSŐR U. 43

SZANTNER FERENC
BAUXITKUTATÓ VÁLLALAT
8221 BALATONALMÁDI PF. 31

SZARVAS IMRE
NÓGRÁDI SZÉNBÁNYÁK
3101 SALGÓTARJÁN PF. 124
FELSZABADULÁS ÚT 44

SZEIDOVITZ GYŐZŐNÉ
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

SZENTGYÖRGYI KÁROLY
SZKFI
5100 SZOLNOK
MUNKÁSŐR U. 43

SZERECZ FERENC
OKGT
1502 BUDAPEST PF. 22
SCHÖNHERZ Z. U. 18

SZÉKY-FUX VILMA
KOSSUTH LAJOS TUDOMÁNYEGYETEM
4010 DEBRECEN PF. 4
EGYETEM TÉR 1

SZÉLES LAJOS
BÁNYÁSZATI EGYESÜLÉS
2800 TATABÁNYA
TÓTH BUCSOKI ÚT 12

SZÉLES MARGIT
SZKFI
2443 SZÁZHALOMBATTA PF. 32

SZÖŐR GYULA
KOSSUTH LAJOS TUDOMÁNYEGYETEM
4010 DEBRECEN PF. 4
EGYETEM TÉR 1

SZÖRÉNYI ZOLTÁN
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

SZŐTS ANDRÁS
BAUXITKUTATÓ VÁLLALAT
8221 BALATONALMÁDI PF. 31

SZUROVY GÉZA
1022 BUDAPEST
BIMBÓ U. 41

SZTRÓKAY KÁLMÁN
1124 BUDAPEST
PAGONY U. 2

T. KOVÁCS GÁBOR
KŐOLAJKUTATÓ VÁLLALAT
6701 SZEGED PF. 30

TANÁCS JÁNOS
MÁFI
1442 BUDAPEST PF. 106

TENKEI SÁNDOR
KŐOLAJKUTATÓ VÁLLALAT
5100 SZOLNOK
MUNKÁSŐR U. 43

THAMO-BOZSÓ EDIT
MÁFI
1142 BUDAPEST PF. 106

TIMÁR ZOLTÁN
ELGI
1440 BUDAPEST PF. 35
COLUMBUS U. 17—23

TÓTH GYÖRGY
MÁFI
1142 BUDAPEST PF. 106

TÓTH KÁLMÁN
BAUXITKUTATÓ VÁLLALAT
8221 BALATONALMÁDI PF. 31

VÁNDORFI RÓBERT
OKGT
1502 BUDAPEST PF. 22
SCHÖNHERZ Z. U. 18

VEREBÉLYI SÁNDOR
BAUXITKUTATÓ VÁLLALAT
8221 BALATONALMÁDI PF. 31

VÖLGYI LÁSZLÓ
OKGT
1502 BUDAPEST PF. 22
SCHÖNHERZ Z. U. 18

ZARÁND CSABA
KŐOLAJ- ÉS FÖLDGÁZBÁNYÁSZATI V.
8801 NAGYKANIZSA PF. 126

ZELENKA TIBOR
OÉÁ
1406 BUDAPEST PF. 34
NÉPKÖZTÁRSASÁG ÚTJA 126

ZSITVAI SZILÁRD
OKGT
1111 BUDAPEST
SCHÖNHERZ Z. U. 18

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GOVINDAN ABHIRAMAN
OIL AND NATURALGAS COMISSION
MAKARPURA ROAD
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T. T. UNIVERSITY
116 AULEY BARZEGAR ST.
19559 CHEMIRAN

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LEFLEF DOGAN
PETROCONSULTANTS LTD
POTTERY ROAD KILL D'GRANGE
DUN LAOGHAIRE
CO DUBLIN

## ISRAEL

AHARONI EFRAIM
PAZ OIL COMPANY
JAFFA ROAD 3
TEL AVIV

GERRY EPHRAIM
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INSTITUT OF ARCHEOLOGY
TEL AVIV UNIVERSITY
RAMAT AVIV
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SERVIZIO GEOLOGICO D'ITAL.
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50121 FIRENZE

BONADONNA FRANCESCO PAOLO
DIPARTIMENTO DI SCIENZE TERRA
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56100 PISA

BORSETTI-CATI A.M.
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VIA ZAMBONI 65
40127 BOLOGNA

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CHIOCCHINI UGO
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VIALE BETTI 1/A
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DEIANA GIOVANNI
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AGIP STIG-NUOVI LABORATORI
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ET D'EXPL. PÉTROLIERE
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RABAT

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RIJKSMUSEUM. GEOL. MIN.
HOOGLANDSE
KREK GRACHT 17
LEIDEN

MEULENKAMP JOHAN E.
STATE UNIVERSITY UTRECHT
BUDAPESTLAAN 4
3508 TA UTRECHT

VAN DER ZWAAN BERT
INST. OF EARTH SCIENCES
BUDAPESTLAAN 4
3508 TA UTRECHT

ZAGWIJN WALDO H.
GEOLOGICAL SURVEY
NETHERLANDS
SPAARNE 17 P. B. 157
2000 AD HAARLEM

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30—059 KRAKÓW

MACZYNSKA STEFANIA
POLISH ACADEMY OF SCIENCES
20—26 AL. NA SKARPIE
00488 WARSAWA

MATL KAZIMIERZ
UNIVERSITY OF MINING
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30-059 KRAKÓW

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00975 WARSAWA

OSZCZYPKO NESTOR
JAGELLONIAN UNIVERSITY
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