47.5 TIME SCALE AND GENERAL SYNTHESIS

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BACKGROUND

Deep earth sampling beneath the floor of the Mediterranean has acquired continuous sequences of marine pelagic ooze which were deposited during an interval of time ranging from the end of the Late Miocene "crisis of salinity" to the present. A recognizable evolutionary trend in assemblages of planktonic foraminifera and calcareous nannofossils has permitted the Leg 13 scientific party to document in this volume a sequence of biostratigraphic zones covering this period of time—that is from the earliest Pliocene onward.

We are indeed fortunate that a majority of the criteria used to establish the boundaries of these zones (see Chapter 47.1) can also be identified in deep-sea sediment cores from other parts of the world's ocean. Many of the biostratigraphic datums used (for example, first appearances, evolutionary changes, and extinctions) have been demonstrated prior to the Mediterranean drilling venture to be, for all purposes, time-synchronous on a worldwide scale (Glass *et al,* 1967; Hays *et al,* 1969; Hays, 1971; Kobayashi *et al,* 1971).

Our confidence in this synchroneity is supported by paleomagnetic measurements carried out on the same sediment samples that contain the fossil record. Back at least into the Gilbert epoch the pattern of reverse and normal magnetization in the deep-sea cores is identical to the sequence of sea floor magnetic anomalies near the crests of spreading mid-oceanic ridges, as well as the reversal history of the geomagnetic field recorded in radiometrically data lava flows (see Chapter 47.2).

We have combined in this chapter the results of the biostratigraphic, chronostratigraphic, and paleomagnetic investigations on the Mediterranean cores in order to arrive at an applicable absolute time scale which we believe is reliable and useful. The purpose of our paper is to document this time scale, evaluate it in terms of previous attempts at formulating a time scale for the same period, and to then use it to give a narrative of changes in the sedimentary environment of the Mediterranean Region from the Late Miocene to the present.

TIME SCALE

The absolute chronology for the Mediterranean sequences is adopted directly from the carefully refined geomagnetic time scale of Talwani *et al.* (1971). This time scale, discussed in Chapter 47.2, is based on a statistical study of the width and spacing of marine magnetic anomalies in the Reykjanes Ridge of the North Atlantic (*op cit*, p. 483 Figure 10). Their premise is "that the age of anomaly 5 in the time scale of Heirtzler *et al.* [1968] is correct and that the rate of spreading in the Reykjanes Ridge has been constant since anomaly 5 time." Back to 3.35 my, this time scale is in close agreement with that of Cox (1969), established principally in the dating of lava flows, with only some minor modifications suggested by Foster and Opdyke (1970). An age of 5.18 my has been adopted for the Gilbert/epoch 5 boundary.

Age of the Miocene/Pliocene Boundary

The Miocene/Pliocene boundary, as recognized at Site 132 in the Tyrrhenian Basin (see Chapter 13) by a sedimentary and paleontological break in Section 2 of Core 21 lies within the upper part of geomagnetic epoch 5 at an extrapolated age of ≈5.4 my. This age assignment is in good agreement with the various time scales given by Berggren (1969, 1971, 1972a), particularly the last one, which takes into consideration the latest deep-sea drilling results in the Atlantic and Pacific oceans. The age is also not inconsistent with the age of 6.5 my proposed by Tongiorgi and Tongiorgi (1964), based on the age of the Isle of Elba, pebbles of which have been found in the basal conglomerate of the Pliocene succession at Roccastrada (Toscana Marittima). This value, as clearly indicated by the authors, has to be considered as a maximum limit; in other words, the Miocene/Pliocene boundary cannot be older than 6.5 my.

Age of the Pliocene/Pleistocene Boundary

The Pliocene/Pleistocene boundary occurs at ≈1.85 my in the lower part of the Olduvai event of the Matuyama epoch in accordance with biostratigraphic criteria summarized in Chapters 47.1 and 47.2. The Lower Pliocene/Upper Pliocene boundary has been put at 3.32 my, coincident with the Gauss/Gilbert epoch boundary.

New Radiometric Dates

The resulting time span of the Pliocene from 5.4 to 1.85 my is consistent with several recently published and still unpublished dates (K/Ar method) of volcanic rocks in contact with marine Pliocene sediments in Italy. These absolute ages, summarized in Ambrosetti *et al.* (1972) and Selli (1970), are shown by large solid dots in Figure 1 and are indicated below.

1) 2.3 ± 0.2 my—Roccostroda, Tuscany. This is a lava flow overlying Upper Pliocene marine sediments reported by Borsi *et al.* (1965).
2) 3.7 ± 0.15 my—Cerveteri (Cerite volcanic group). This is another lava flow in contact with probably Lower Pliocene marine clays reported by Bonadonna and Bigazzi (1970).
3) 4.2 ± 0.2 my—Lava Simoni, Via Palambaresse, Rome, These are dates on tuffs (plagioclase and hornblende) interbedded in Pliocene clays heteropic with the lacustrine basin of Poggio Mirtelo with Lower Villafranchian fauna (Ambrosetti *et al,* 1968).

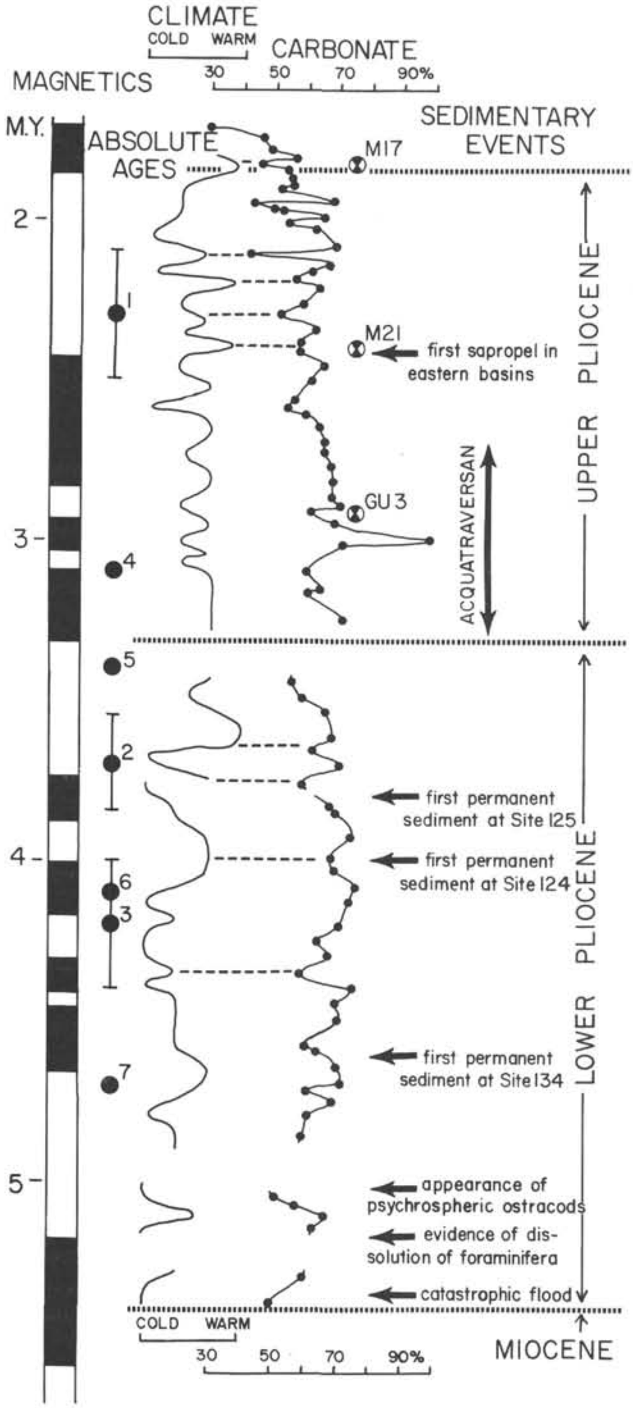

Figure 1 *Variations in the inferred climates and calcium carbonate during the Pliocene, based primarily on data from Site 132 in the Tyrrhenian Basin. The geomagnetic reversal sequence is from Talwani* et al *(1971) with large solid dots (numbers identified in text) showing absolute age determinations on igneous rocks and volcanic ashes in contact with marine Pliocene sediments in Italy.* M 17, M 21 *and* GU 3 *stand for carbonate minima identified in Pacific Ocean cores. During the Upper Pliocene, carbonate minima generally coincide with warm climates, whereas in the lowermost parts of the lower Pliocene the opposite relationship is indicated.*

4) 3.1 my–S. Leonardo di Cutro, Crotone Basin, Calabria. Dates have been obtained by Dymond (unpublished) on glass shards from a very deep, fresh volcanic ash interbedded in clays at the base of the Papanice formation (Emiliani *et al.*, 1961). An age of Middle Pliocene (lower part of the *Globorotalia crassaformis* Zone of Cati *et al.*, 1968) is given by Selli (1970) to the fauna from this formation.
5) 3.4 my–Capo Colonne, Crotone Basin, Calabria. Ages here are from the same volcanic ash interbedded at the base of the above mentioned Papanice formation (Dymond and Selli, unpublished) and this discordance is believed to be the results of argon leakage.
6) 4.1 ± 0.13 my–Montecatini Val di Cecina, Tuscany. This is an age from selagite (biotite) metamorphosing Miocene-Pliocene clays reported by Borsi *et al* (1967).
7) 4.7 my–Orciatico, Pisa, Tuscany. This date reported in Selli (1970) is on a mafic trachyte (another selagite) that metamorphoses clays considered to be of Lower Pliocene age.

Pre-Pliocene Chronology

It is interesting to note that no absolute ages have yet been reported for volcanic rocks associated with marine sediments which lie outside the inferred range of the Pliocene as delineated by the paleomagnetic stratigraphy on cores from Site 132. The age of the Miocene/Pliocene boundary reported here, though significantly younger than that interpolated by Kulp (1959 and 1961) and Holmes (1965), seems reasonable in light of recent investigations on the chronology of Upper Miocene strata. The *Hipparion* fauna, long thought to belong to the Pliocene, has been found in pumice-rich strata on the island of Samos in Greece, where it has been dated at ≈9.3 my (Van Couvering and Miller, 1971). The same continental fauna has since been found in Crete in association with a marine foraminiferal fauna correlative with the Tortonian Stage of the lower part of the Upper Miocene (De Brujin *et al*, 1971). This correlation to the type Tortonian is based on the occurrence of *Globorotalia acostaensis* and on uniserial uvigerinids.

As to the exact age of the top of the Tortonian, only estimates can be made at the present time. Perhaps the best inferences come from an examination of continuously cored marine sections at DSDP Sites 62 and 77 from the western and eastern equatorial Pacific, respectively. Upon correlating the base of the *Ceratolithus tricorniculatus* Zone in these holes with its occurrence in the Mediterranean close to the Miocene/Pliocene boundaryry at ≈5.4 my, one is able to establish that the succeeding strata were deposited at a remarkably constant sedimentation rate. If one then assumes that this same rate applies to slightly older strata in the next lower *Discoaster quinqueramus* Zone, the top of the Tortonian (i.e., lower part of Zone N.17 according to Cita and Blow, 1969) occurs at about 7.5 my. This extrapolation suggests that the entire Messinian stage, during which the Mediterranean evaporites were deposited, was less than two million years in duration.

The fact that the Tortonian extends upwards in time to strata certainly as young as 9 my (Van Couvering and Miller, 1971) and perhaps even 7.5 my in age, leads us to discount a single date reported by Selli (1970) on Messinian strata at S. Ginesio, Macerata, Italy. One of us (W.B.F.R.) wrote of the apparent discrepancy to J. R. Dymond who made this K-Ar determination on isolated glassy shards from a volcanic ash layer a few meters above the *gessoso-solfifera* formation; he replied that there were indeed problems with this particular measurement and that as far as he was concerned it could be in error.[1]

CHRONOSTRATIGRAPHY

The range distributions of key species of planktonic foraminifera and calcareous nannoplankton from Late Miocene times to the present are shown in Figure 2. The data have been correlated to the geomagnetic time scale of Talwani *et al*, (1971) by means of the paleomagnetic measurements on sediment samples from Sites 125 and 132, discussed in Chapter 47.2.

Key biostratigraphy levels based on the foraminiferal faunas include:

1) The first appearance of *Globorotalia puncticulata* at 4.25 my between events B and C_1 of the Gilbert epoch.
2) The extinction horizon of *Globorotalia margaritae* at 3.32 my coincident with the Gauss/Gilbert epoch boundary.
3) Extinction horizon of *Sphaeroidinellopsis* spp. at 2.85 my in the Kaena event of the Gauss epoch.
4) Extinction horizon of *Globigerinoides obliquus extremus* at 2.22 my in the Matuyama epoch.
5) First evolutionary appearance of *Globorotalia truncatulinoides* from *Globorotalia tosaensis* at ≈1.85 my in the Olduvai event of the Matuyama epoch (documented only at Site 125).

The first evolutionary appearance of *Globorotalia truncatulinoides* and the extinction horizons of *Sphaeroidinelopsis* spp. and *Globorotalia margaritae* (group) correspond to Datums IV, V, and VI of Hays *et al.* (1969), first recognized in deep-sea cores from the Equatorial Pacific.

Key biostratigraphic levels based on the calcareous nannofossils include:

1) Last occurrence of *Discoaster quinqueramus* at ≈5.4 my, at or near the top of the Miocene in event A of geomagnetic epoch 5.
2) First appearance of *Ceratolithus rugosus* at 4.06 my in event B of the Gilbert epoch.
3) First appearance of *Discoaster asymmetricus* at 3.46 my, between the base of the Gauss epoch and event A of the Gilbert epoch.
4) Last occurrence of *Ceratolithus tricorniculatus* at 3.28 my near the base of the Gauss epoch.
5) Last regular occurrence (at Sitew 125 and 132) of *Reticulofenestra pseudoumbilica* at 2.88 my in the Kaena event of the Gauss epoch.
6) Last occurrence of *Discoaster surculus* at 2.39 my near the base of the Matuyama epoch.
7) Last occurrence of *Discoaster penetaradiatus* at 2.32 my, also near the base of the Matuyama epoch.
8) Extinction horizon of *Discoaster brouweri* at 1.85 my within the Olduvai event of the Matuyama epoch.
9) Last occurrence of *Pseudoemiliania lacunosa* at 1.03 my slightly below the Jaramillo event of the Matuyama epoch.
10) First appearance of *Emiliania huxleyi* at ≈0.58 my in the Brunhes epoch. The extinction horizon of *Discoaster brouweri* corresponds to Datum III of Hays *et al.*, and is practically coincident with the evolutionary appearance of *Globorotalia truncatulinoides*, observed in Core 4 of Hole 125.

Variations in Climate and Carbonate Productivity

The changes in the Pliocene climate of the Mediterranean as inferred from assemblages of planktonic foraminifera, has been presented in Chapter 47.3. The generalized climate curve established for the continuously cored section of Site 132 in the Tyrrhenian Sea is shown in Figure 1 plotted against the geomagnetic time scale along with the measured percentages of calcium carbonate from the same core sections.

We point out that generally from about 4.5 my onwards, periods of inferred warm climates appear to correlate with carbonate minima, much like the relationship first recognized by Arrhenius (1952) to be the case in the Pacific. He reasoned that during the glacial intervals the trade wind circulation would be greatly accentuated, leading to a more vigorous equatorial current system which in turn would stimulate greater upwelling and hence greater productivity at the equatorial divergence. The higher productivity would be reflected in a proliferationa of calcium carbonate in the surface water and a higher flux of

[1]Since Selli (1970) apparently placed great faith in this date, his entire time scale is in significant discrepancy with ours. Unfortunately, his paper represents an offician report of the Working Group in Absolute Ages of the Committee for Mediterranean Neogene Stratigraphy, and with this gains great attention, if its inconsistenciecies are not brought to attention. We offer the following points of criticism:

1) By extending the base of the Messinian back to 11.8 my, Selli infers that foraminiferal Zone N.18 of Banner and Blow (1965) is the longest in duration of the Neogene, when, in fact, in most other sections and in all DSDP cores (see Berggren, 1969, 1971, and 1972a) this zone is the shortest and belongs, in part, to the Lower Pliocene.

2) The correlation presented by Selli of Bolli's (1966) and Banner and Blow's (1965) zonations places the top of the *Globorotalia margaritae* Zone in Zone N.20, where as the range charts published by Blow (1969) never show it further up than about three quarters of Zone N.19.

3) The Miocene/Pliocene boundary is put at the N.18/N.19 boundary instead of within zone N.18.

4) The Tortonian Stage is indicated by Selli as correlative to Zone N.14 (pars) and to Zone N.16 (pars). But it is well known from the literature that *Globorotalia acostaensis* first appears about 35 meters above the base of the stratotype Tortonian. The main body of the type Tortonian corresponds to Zone N.16, and an extension to the lower part of Zone N.17 has been documented by Cita and Blow (1969).

The Tortonian Stage is indicated by Selli as extending from 14.1 to 11.8 my, with a duration (2.3 my) less than one-half that of the Messinian Stage (4.8 my). The anomalous interpretation is based on two radiometric ages, one of which (by Dymond and Selli) has been discussed in our text. This age of 22.3 my is indicated as the base of Middle Messinian, a few meters above the *gessoso-solfifera* formation (Selli, op. cit., p. 55). This statement strongly disagrees with the definition of the neostratotype Messinian by the same investigator (Selli, 1960). In the type section, the *gessoso solfifera* formation represents the main body of the stratotype, with the exception of a few meters of marls overlying the *gessi superiori*. It seems strange to us indeed to have these few meters represent more than 3 my of deposition.

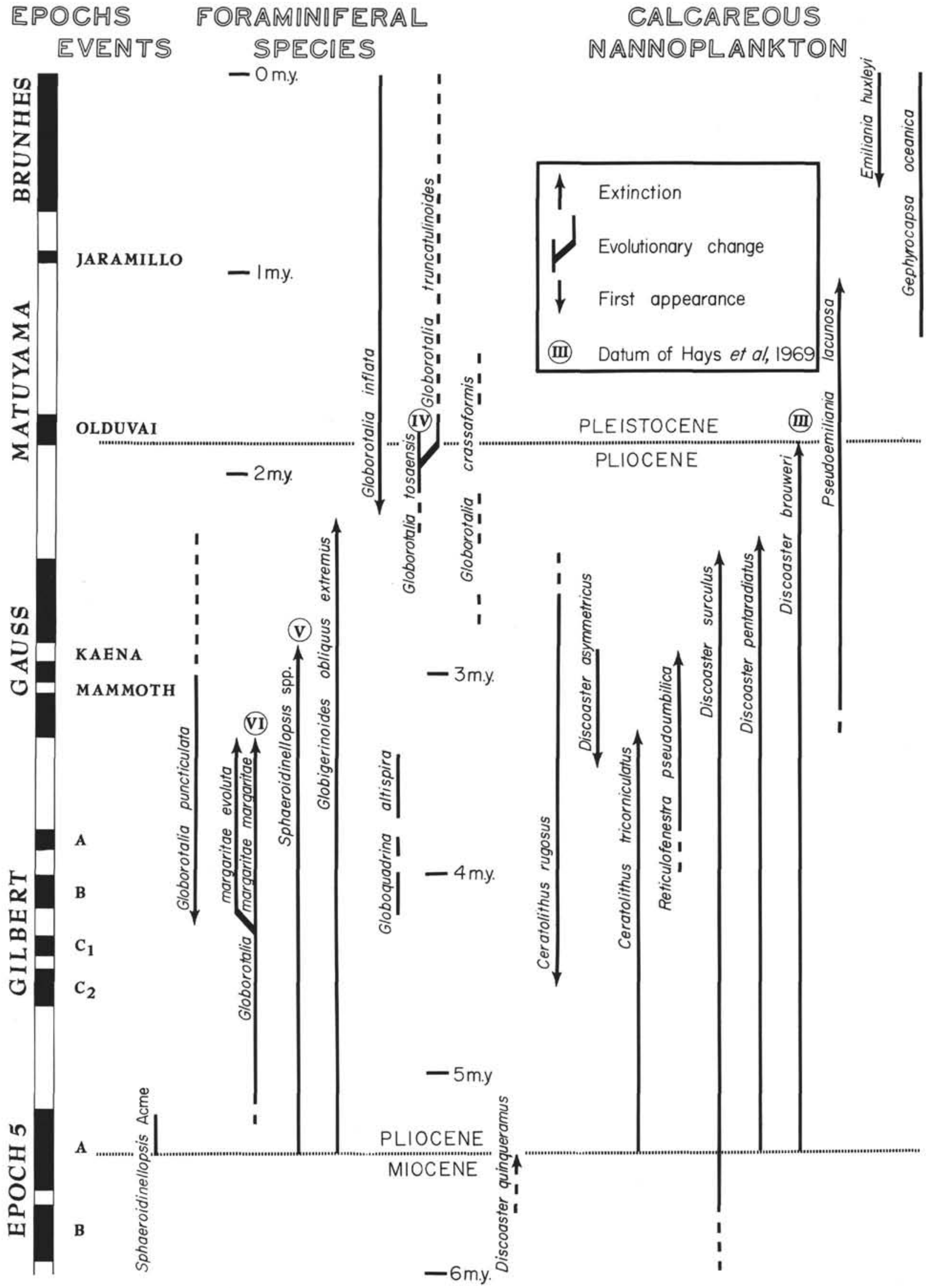

Figure 2. *Range distribution of key planktonic foraminifera and calcareous nannofossils in the Mediterranean deep-sea sequences from the Late Miocene to present. The oldest Pliocene sediment belong to the* Sphaeroidinellopsis *Acme-zone extending from ≈5.2 to 5.4 my. The ranges of* Globoquadrina altispira *and* Globorotalia crassaformis *employed in the zonation schemes of Bolli (1966) and Cati* et al *(1968) are also shown, though in the Mediterranean their occurrences seem greatly influenced by local environmental conditions.*

carbonate skeletal debris to the bottom, inhibiting dissolution.

The carbonate cycles first systematically analysed from Quaternary cores of the equatorial Pacific have been extended back into the Gauss epoch by Hays *et al.* (1969). Three of their minima in the Upper Pliocene (i.e., M17, M21, and GU 3) are shown in Figure 1 at levels established by paleomagnetic reversal boundaries in both the Pacific and Mediterranean sequences.

The highest value of $CaCO_3$ in the Mediterranean occurs during climatic episode "brown" (see Chapter 47.3).

The carbonate maximum here, at 3.0 my, is perhaps correlative to a cold period referred to in Italy as the Acquatraversan erosional phase (see discussion in Ambrosetti and Bonadonna, 1967) not to mention the pronounced cooling recognized at that same time in Iceland (Rutten and Wensink, 1960), which is considered to reflect the onset of high latitude glaciation in the North Atlantic region (McDougall and Winsink, 1966; W. A. Berggren, personal communication).

The occurrence of warm peaks with carbonate minima, particularly well-expressed in the Upper Pliocene, is not typical of the Late Quaternary sequence recovered in piston cores, where jsut the opposite trend is observed, nor is it apparently characteristic of the lower parts of the Pliocene either. In fact, from about 4.7 my back to the Late Miocene there is a direct correlation of warm peaks with carbonate maxima. Perhaps this sympathetic relationship ship is somehow related to special geomorphologic conditions in the Mediterranean set up by the terminal flooding following the Late Miocene "salinity crisis." We might speculate that at the time when sea level rose dramatically, fine-grained terrigenous sediments were trapped to some degree and for some while well back in deep river-estuaries, and that high stands of the eustatic sea level were accompanied by less effective clastic input from the continents to the deep basins, (i.e., less dilution of the carbonate skeletal productivity). Cold periods synchronous with the growth of high latitude glaciation in Antarctica (Denton and Armstrong, 1969; Rutford *et al*, 1968; Denton *et al*, 1971) and Alaska would be expected to have at least some modest sealevel lowering that in turn might then aid in flushing the detrital sediment out of the estuaries and deltas across the shelf break to the vicinity of the drill sites on the basin floor.

Dilution of biogenic production during cold periods by excess terrigenous input is the most preferred mechanism to explain the Late Quaternary carbonate fluctuations in pelagic sediment of the Mediterranean Sea. At this more recent time, however, exchange with the Atlantic Ocean was drastically reduced at Gibraltar so that vigorous circulation there would not necessarily penetrate the Meditarranean to produce greater upwelling.

A BRIEF HISTORY OF SEDIMENTATION IN THE MEDITERRANEAN

As a starting point, we assume that in Late Miocene surface water masses of the open Atlantic and Pacific oceans, marine faunas were evolving and flourishing in response to normal evolutionary changes and in adaptation to different environments. This assumption is certainly not contradicted by the preliminary DSDP drilling in these oceans.

Late Miocene Setting

Commencing somewhere around 7.5 my ago,[2] the Mediterranean embayment of the Atlantic Ocean began to be restricted by what we surmise to have been a narrowing of the passageway between the Iberian peninsula and North Africa.[3] This restriction is seen in the euxinic and pyritic marls of the Tripoli formation of the lower Messinian Stage (Sahelian time of Catalano and Sprovieri, 1971). This level has been correlated with the *Globorotalia plesiotumida* Zone in Sicily (Base of Blow's Zone N.17) and is time equivalent to the upper part of geomagnetic epoch 7. Findings at Site 62 in the equatorial Pacific indicate that the base of Zone N.17 corresponds to the nannofossil *Discoaster quinqueramus* Zone (see Martini, 1971). The Miocene restriction eventually led to complete isolation of the Mediterranean from the Atlantic, initiating the "crisis of salinity" (Ruggieri, 1967). The subsequent desiccation of the Mediterranean basins is treated in Chapter 43, to which reference is made. It suffices here to say that the severe periods of desiccation resulted in the deposition of evaporites and the subsequent sterilization of the Mediterranean, except for local marginal lakes with Sarmatian faunas.

Therefore, as brief marine incursions of open-ocean waters spilled into the Mediterranean basins (as found, for example, in Cores 9 and 10 of Site 124 and Cores 22 and 23 of Site 132) they carried in marine planktonic organisms of the northeastern Atlantic. However, since the time span of the "salinity crisis" occupies but a single foraminiferal zone and a single nannofossil zone, fossiliferous records of these marine incursions could appear to give a false impression of a continuity in marine sedimentation.

The Beginning of the Pliocene

Our story begins with the terminal flooding of the Mediterranean (Figure 3). This event, like the previous incursions, also carried in foraminifera which have been found in the pyritic marls of Section 2, Core 21 in the Tyrrhenian Basin (Site 132). The admixture of fully developed planktonic tests with dwarfed benthonic forms characteristics of littoral environments, apparently records the transgression of the Atlantic flood across the Tyrrhenian Rise. The difference between this event and the previous ones is that this one has remained permanent to the present day.

The filling of the Mediterranean was not a short-term event. The association of dwarfed faunas with shallow-water benthonic foraminifera is strong evidence that the laminated pyritic deposits of Core 21 at Site 132 were laid down when the Mediterranean sea level in the Tyrrhenian

[2]This age is extrapolated from the location of the base of zone N.17 at Site 62 in the equatorial Pacific as previously discussed.

[3]The actual passageways have been identified as The South Betic and North Rifian straits.

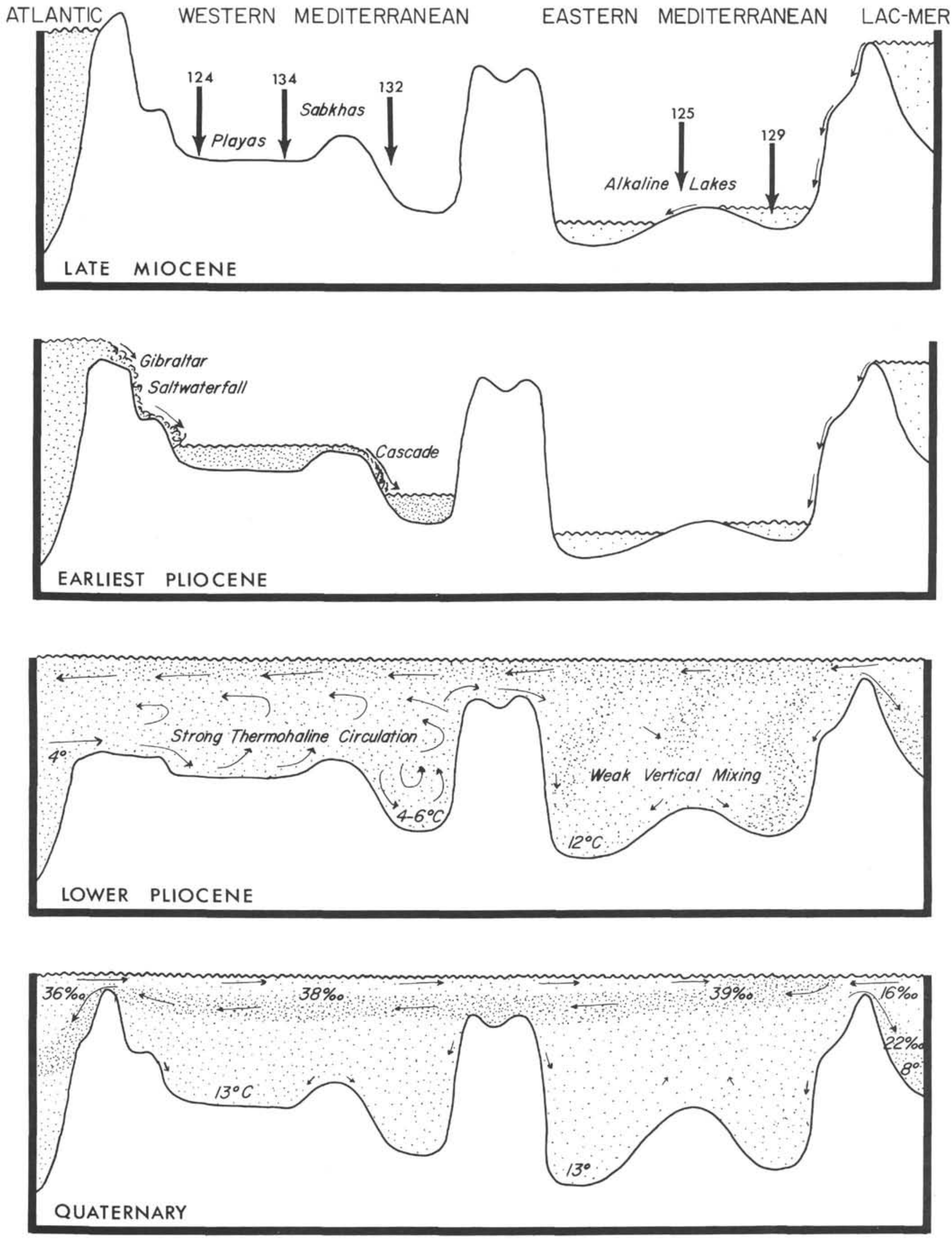

Figure 3. *Stages in the desiccation, flooding, and oceanic circulation of the Mediterranean Sea, as discussed in the text.*

Basin was still some thousand meters or more below the Atlantic surface.

The evidence obtained in Core 21 and previously noted in many outcrop areas around the Mediterranean is that the uppermost strata of the desiccated basin are immediately overlain by open marine pelagic sediments. Furthermore, there are no results from the analyses of the accompanying fauna, mineral composition, or lithogenesis to indicate that this sediment was deposited in a basin shallower than the present Mediterranean.

All we can deduce is that the transition from a period of gradual brine dilution to a complete open exchange with the Atlantic ocean was geologically instantaneous.

Evidence of Initially Cold Water

The one particularly unusual aspect of the first pelagic sediments is that they contain an assemblage of planktonic foraminifera dominated by two species–*Sphaeroidinellopsis seminulina* and *S. subdehiscens*. Neither of these species is living today, but their descendants with massive tests are known to adapt best to a deep habitat in the open-ocean thermocline (Be and Tolderlund, 1971).

We conjecture that if the Mediterranean filled to the brim by a shallow overflow of the Straits of Gibraltar, the mean water temperature of the new sea would be that of the Atlantic Ocean surface water mass. In this sea we should expect that species of planktonic foraminifera living in the uppermost water layers would be dominant. If a shallow sill or cataract remained at Gibraltar, this sea would eventually be stocked by a limited assemblage as the Black Sea is today.

One remarkable piece of evidence contradicting the above conjecture was provided by Benson (in press) who discovered, in the early Trubi marls, benthonic ostracods of the same type as are found today only on the deep floors of the open ocean where the water temperature is less than 6°C. These creatures could not have merely been swept into the basin, having been siphoned over the Gibraltar sill, since they have subsequently in the Mediterranean reproduced throughout more than three million years of the Pliocene.

The alternate conjecture is that the dam at Gibraltar broke down, some time after the brines transgressed the Tyrrhenian Rise, and some time before the basin filled to the depth level of the six degree isotherm of the Atlantic Ocean (i.e. with the Mediterranean sea level depressed below 1500 meters). If the collapse was sudden, providing a deep erosional opening to the Atlantic, we should expect that the inrushing waters would consist of both Atlantic Ocean surface water and cold Atlantic Deep Water. As a consequence, the resulting fill would be well mixed and cool. How cold the water was would depend on criteria we have no knowledge of at the present time. All we can say is that the major dominance of a deep-water species of foraminifer is consistent with the supposition that deep water from the Atlantic accompanied the flood, and that enough cold, deep water came into the basin to support the psychrospheric ostracod population. The evidence of an initial cold environment is also indicated by oxygen isotope measurements on the tests of the planktonic and benthonic foraminifera, as presented in Chapter 30.4. What strikes us as perplexing is that unless the opening at Gibraltar was deep, the bottom water of the Mediterranean would soon warm up and exterminate the ostracods. The only actualistic model of such an event is seen in the emptying of glacial lakes (see Malde, 1968).

Consequences of a Catastrophic Flooding

The model of a catastrophic flooding is illustrated in Figure 3. We would predict a strong vertical circulation in the thermally homogeneous water mass of the Mediterranean basins and would expect vigorous mixing with the Atlantic Ocean until the basin reached a thermohaline equilibrium. It is interesting to note that at the level of the Miocene/Pliocene boundary in the Andalusian section of western Spain facing the Atlantic, there is an abrupt change in facies from gray clays to calcareous sandstones synchronous with the first occurrence of *Globoratalia margaritae* and *Ceratolithus tricorniculatus*, a biostratigraphic event that occurred at the end of the Messinian "Crisis of salinity" (See Chapter 47.1). Perhaps the basal conglomerate found beneath the Tabianian type section and at the base of the Trubi in other outcrops represents the violence of the final flooding.

One further puzzling deduction from the simple model of catastrophic flooding is that the Mediterranean basins would be expected to reach thermal equilibrium with the Atlantic in a few tens of thousands of years. Why does the abundant population of *Sphaeroidinellopsis* persist for five to ten meters of the sedimentary record?[4] To suggest that the inferred thickness actually represents a short time interval due to increased productivity and sediment dispersal is possible; however, we have no clear evidence to prove it. We do know that the flooded basins would be initially grossly out of isostatic equilibrium. The studies of crustal rebound of the Laurentian and Scandinavian shields following the melting of the last ice sheets indicate that flowage in the upper mantle would start immediately in order to compensate the basins (Crittenden, 1963; Bloom, 1967). The sinking of the Mediterranean sea floor would not only flex down the continental margin, but the accompanying increase in volume of the basin would require the continued input of Atlantic Deep Water over the Gibraltar Sill. This we can be assured of because the density stratification of the Mediterranean basins would require the newly added water to come in across the sea bed in order to maintain eustatic equilibrium with the Atlantic water column. The logical consequence of the subsidence of the Mediterranean sea floor would be to bring in water from the Atlantic threshold depth. This inflow would not only keep the Mediterranean basins well ventilated, but as more and more of the Mediterranean basins became stratified, the *in situ* bottom water temperature would approach that of the Atlantic threshold temperature.

[4]The top of the *Sphaeroidinellopsis* Acme-zone is found in Section 3 of Core 20, Site 132. Neither Core 20, nor Core 21, with the evaporite-Trubi contact, was completely full. Thus an uncertainty exists for the true thickness of the acme-zone depending on exactly where in the nine meters interval cut for each core the recovered sediments came from.

We noted in Chapter 40 evidence of partial solution (i.e., oliogolytic facies) of foraminifera in Sections 3 and 4, Core 19, Site 132, some 9 meters above the top of the *Sphaeroidinellopsis* Acme-zone. Perhaps, although we cannot be sure of this point, the selective corrosion observed at this level in the sediment column of early Pliocene age, (see Figure 2) indicates the presence of Atlantic Ocean Deep Water unsaturated with $CaCO_3$ which temporarily formed as a discrete water mass within the western basins of the Mediterranean Sea. The period of time represented by the oligolytic facies is on the order of a hundred thousand years, and this facies was never observed at any other levels at other sites, including even he sediments deposited on the floor of the Hellenic Trench at depths close to 5000 meters.

The *Sphaeroidinellopsis* Acme-zone occurs only in the lowermost part of the Trubi formation and is characterized at Site 132 by reddish colored marls containing traces of iron hydroxides. Perhaps this color comes from the contamination of the biogenous components with sediments washed from lateritic soils of the former desiccated basin.[5]

Communications Between the Eastern and Western Basins

Since the psychrospheric ostracod faunas apparently also penetrated into the eastern Mediterranean (e.g., found in eastern Sicily, southern Calabria, and Crete, see Benson and Sylvester-Bradley, 1971), it is possible that in earliest Pliocene times the barrier between the eastern and western basins was not as high as it is today. This supposition is further supported by the necessity to get marine waters upon occasion into the eastern basins during the Messinian to produce the great thicknesses of halite found in offshore exploration south of Turkey and west of the Levant coast (personal communication from Gerald Friedman). The connection was probably north of Sicily since there is no compelling geologic evidence to suggest that the Straits of Sicily were deep in the Upper Miocene.

Tectonic activity in Sicily and Calabria has been capable of lifting Pliocene and Pleistocene marine sequences over one thousand meters above their original site of deposition (Ogniben, 1970). Exactly when the restriction between the two basins began is difficult to pinpoint. In Chapter 47.4 it was pointed out that variations in the δO^{18} composition of pelagic foraminifer (*Orbulina universa*) in Cores 18 to 13 at Site 132 in the Tyrrhenian Basin suggest that the western Mediterranean basins during Lower Pliocene time were sharing the isotopic composition of the Atlantic Ocean. The paleontological investigation of *Globigerinoides* in Chapter 47.3 showed that the surface waters of the Ionian Basin were warmer than those of the Tyrrhenian Basin at identical time intervals. The contradictory finding, however, of higher δO^{18} values in the fossil tests from Sites 125 and 125A than those from Site 132 perhaps indicates that the Ionian Basin shared waters of a higher δO^{18} composition as the result of a partial isolation of this basin, since excess in evaporation over precipitation would enhance the δO^{18} ratio in marine waters (Dansgaard, 1961). It was noted further, that the δO^{18} values in the Ionian Basin increase from an average of +0.38 in the Lower Pliocene to an average of + 1.17 in the Upper Pliocene. Although the increase could be used to suggest that the Upper Pliocene water temperatures were colder, the paleontological evidence contradicts this inference. The resolution of the apparent incompatability of the two approaches is found in hypothesizing a progressive isolation of the eastern Mediterranean from the influx of Atlantic waters by a gradual constriction of the portals either through the Straits of Sicily or the Strait of Messina.

In Chapters 6, 13, and 14, there are lengthy discussions concerning the evidence for vigorous, deep thermohaline circulation in the Balearic Basin and Tyrrhenian basins in Pliocene times (i.e., the construction of "lower continental rise hills," the presence of erosional hiatuses, and the formation of hardgrounds, etc.). No comparable evidence of current reworking in the sediment has yet been observed in the drill cores from the Ionian and Levantine basins.

Climatic Deterioration in the Upper Pliocene

Commencing around 3 my ago, climatic deterioration led to the deposition of ice-rafted debris directly overlying tropical faunas in the Labrador Sea (Berggren, 1971b) and a major change to diatomaceous sediments in the northernmost Pacific Ocean and Bering Sea (DSDP Leg 19, Scholl *et al*, 1971). This cooling is recognizable in our climatic evaluations at the "brown" episode referred to previously. An increase in sedimentation rate in the Tyrrhenian Basin from 2.85 to 4.0 cm/10^3 y has been observed coincident with this interval. Eustatic drop in worldwide sea level associated with the high latitude glaciations (Evernden *et al*, 1964, Denton and Armstrong, 1969) was probably responsible for the major Pliocene erosional phase (Acquatraversan) observed in Italy.

The paleontological evidence of cooling is not only supported by the occurrence of a restricted number of the taxa used to build up the *Globigerinoides* curve," but is also apparent in the almost complete absence of keeled *Globorotalia*, the abundance of *Globorotalia puncticulata*, which apparently behaves as *G. inflata* in the present seas, and the recorded absence of *Globoquadrina altispira*—a subtropical form whose range in other seas extends upward to the interval under discussion.

Communication with the Black Sea

The cooling episodes in the Upper Pliocene shown in Figure 2, if reflecting high latitude glaciations, should be accompanied by some eustatic lowering of sea level. With a sufficient drop, the eastern Mediterranean could be cut off from direct communication with the Black Sea, allowing the latter to gradually freshen, as according to Late Quaternary analogues (Ross *et al*. 1970).

The climatic deterioration in the eastern Mediterranean would be expected to decrease the temperature difference between the surface water and bottom water mass so as to facilitate convective overturn and enhance the oxygen ventilation of the deep water. Our knowledge of the Pliocene history of the Black Sea does not permit us to speculate on eustatic changes in that basin, except to note that no crisis of high salinity (i.e. period of evaporite

[5]See Chapter 13 for for further discussion of the soild cored at Site 132.

deposition) has yet been reported from this time period.

Stimulated ventilation of the Mediterranean is shown by the highly oxidized light-colored oozes of Lower Pliocene age. The warming of climates, on the other hand, would tend to develop a greater density stratification leading to oxygen depletion. In fact, the first sapropel is recognized in the eastern Mediterranean at ≈2.4 my, just above the Gauss/Matuyama epoch boundaries (see Figure 1). This time period (peak of the "green" episode) is represented in Core 6 of Hole 125, Core 2 of Core 125A, and Core 10 of Site 132, all of them, occurring in the *Globigerinoides obliquus extremus* Interval-zone and in the lower part of the *Discoaster pentaradiatus* Zone. A marked decrease in carbonate content of equatorial Pacific Cores V 24-62 and V 24-59 has previously been discussed. Hays *et al.*, 1969, have interpreted this minimum (M 21) to represent an episode of decreased productivity in the equatorial divergence associated with an interglacial climate. Simultaneous melting of high latitude ice sheets is demonstrated by rafting of pebbles and sands into the Labrador Sea (Berggren, 1972b).

Development of Sapropels

Sapropels recovered in the DSDP cores and in piston cores seem to occur at levels in the sediment column coincident with the warming trends on the infered climate curves (see further discussion of this point in Chapter 46). Sapropels indicate short periods of euxinic conditions (stagnations) in the Mediterranean basins. Figure 4 illustrates a model first proposed by Olausson (1961), where the rise in the world sea level following a glacial event establishes a communication between the salt waters of the eastern Mediterranean and the fresh waters of the Black Sea. The new communication produces a major density disequilibrium such that the heavier Mediterranean waters rush through the Bosporus Straits and underflow into the Black Sea. The invading salty water displaces the Black Sea fresh water into the Mediterranean, as occurs today, but initially with much greater volume. The warming of the Mediterranean surface waters that accompanies the rise in sea level, produces the aforementioned density stratification that inhibits the efficient sinking of the new fresher film of surface water and prevents reoxygenation of the deep water body. The cooccurrence of a sudden lens of low density fresh water across the stratified basin thus apparently completely halts vertical mixing and initiates a brief period of basin-wide stagnation and the concomittant deposition of sapropels. This process repeated itself numerous times in the last 2.4 my, the youngest such event being post-glacial, about 7 to 9 thousand years ago.

The Quaternary Circulation Pattern

The present circulation scheme of the Mediterranean is shown in the lowermost profile of Figure 1. The Gibraltar Portal diminished in depth somewhere near the end of the Pliocene, as evidenced by the disappearance of phychrospheric faunas (Benson, in press). Also, commencing about 0.5 my ago the episodes of oxygen depletion were felt in the Tyrrhenian Basin as well as in the Ionian and Levantine basins, indicating that perhaps by this time the inflow of North Atlantic Deep Water had been terminated altogether.

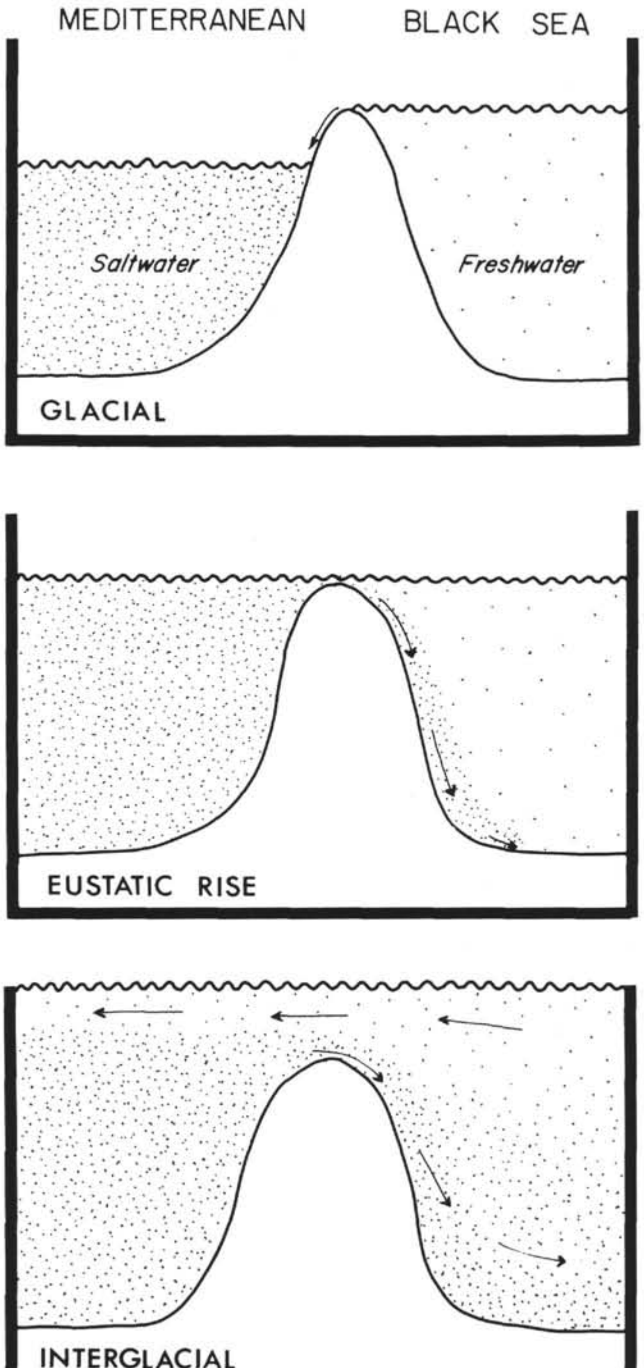

Figure 4. *Stages of communication between the Mediterranean and Black seas. Low eustatic sea stands in the Mediterranean lead to isolation of the latter sea and its freshening. Subsequent rise of sea level allows denser marine waters to invade across the Bosphorus sill as an underflow, flushing out the former fresher water of the Black Sea to form a brackish lens on the surface of the eastern Mediterranean. The resulting density stratification inhibits convective overturn and ventilation of the deep water there, thus initiating basin-wide stagnation and deposition of the sapropels.*

Today the Mediterranean balances its salinity budget by discharging the high salinity Levantine Intermediate Water into the Atlantic as an underflow across the Gibraltar sill (Miller, 1963). The outflow creates a marked impoverishment of the nutrient content of the Mediterranean (McGill, 1961). Perhaps the change in sedimentation rate from 4 to 3.5 cm/103 y observed coincident with the Jaramillo (despite high volcanic contributions at that time) reflects the initiation of exchange of Intermediate Water across the Gibraltar sill and the blockage of the former deep water communications. For discussion of the productivity in planktonic foraminifera in the Ionian Basin during periods of stagnation, reference is made to Chapter 40.

CONCLUSIONS

The interpretation here proposed reflects considerations of the roles played by ocean circulations and communications between the Atlantic, Mediterranean, and Black seas. Veritical changes in sea level have been interpreted to have had a great influence on the cyclic sedimentaion, more so than unidirectional tectonic movements. However, the eventual constriction of the Gibraltar Portal has been significant. The aforementioned models depicting stages in the Pliocene history of the Mediterranean are provisional and subject to thorough, open criticism. They are given here to stimulate thought on complex geologic phonomena not at all fully comprehended by the authors. It is our hope that subsequent detailed studies of the continuously cored Pliocene and Pleistocene sections at Sites 132 and 125 will shed light on the complex history here only briefly and incompletely outlined.

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