THE MESOZOIC-CENOZOIC GEOLOGIC EVOLUTION OF IBERIA, A TECTONIC LINK BETWEEN AFRICA AND EUROPE

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ABSTRACT

The Mesozoic-Cenozoic geologic evolution of Iberia and adjacent land masses was controlled by Triassic-early Jurassic rifting, separation of North America and Africa in middle Jurassic (Bajocian), latest Jurassic-earliest Cretaceous rifting, opening of the Bay of Biscay and Rockall Trough in middle Aptian-early Albian, collision of Iberia, Africa, and Europe from latest Cretaceous to the Miocene, separation of Greenland and North America in the Maastrichtian, separation of Greenland and Rockall Plateau in early Eocene, and opening of the Balearic and Tyrrhenian basins in the Oligocene to the Quaternary. From mid Jurassic to early Cretaceous Iberia was part of the Eurasian Plate, but with the opening of the Bay of Biscay the peninsula became semi-independent. At the beginning of the Oligocene the plate boundary shifted to its present position in the Straits of Gibraltar and Iberia again became part of the Eurasian Plate. Sediment deposition within this complex arrangement of plates and ever changing styles of tectonism was controlled by subsidence of the seafloor, changes in the depth of the CCD, changes in sea level (glacially and tectonically induced), latitudinal migration of the plates, and the gradual expansion of the oceans.

Key words: Alpine System, Atlas Mountains, Betic System, Black shales, Calcium carbonate compensation Depth (CCD), Carbonate platforms, Continental collision, Glaciations (Würm; Wisconsin), Deep-sea erosion, Evaporites, Pyrenees, Rif System, Rifting, Sea-level changes, Seafloor spreading, Siliceous sediments, Subduction, Subsidence, Volcanism.

RESUMEN

La evolución geológica de Iberia y de las masas continentales adyacentes durante el Mesozoico y Cenozoico estuvo controlada por los siguiente acontecimientos: "rifting" Triásico-Jurásico inferior, separación de Norteamérica y África en el Jurásico medio (Bajociense), "rifting" Jurásico terminal-Cretácico basal, apertura del Golfo de Vizcaya y del Surco de Rockall en el Aptiense medio-Albiense inferior, colisión de Iberia, África y Europa entre el Cretácico terminal y el Mioceno, separación de Groenlandia y Norteamérica en el Maastrichtiense, separación de Groenlandia y la Plataforma de Rockall en el Eoceno inferior y, por último, la apertura de las cuencas Balear y Tirreniense entre el Oligoceno y el Cuaternario. Desde el Jurásico medio al Cretácico inferior, Iberia formó parte de la Placa Eurasiática, independizándose después parcialmente la península con la apertura del Golfo de Vizcaya. A comienzos del Oligoceno, el límite de la placa se trasladó hasta su posición actual en el Estrecho de Gibraltar, con lo que Iberia volvió a formar parte de la Placa Eurasiática. La sedimentación en esta compleja disposición de las placas, con contínuos cambios de estilo tectónico, estuvo controlada por la subsidencia de los fondos marinos, las variaciones de la batimetría de la CCD, los cambios en el nivel del mar (de origen glacioeustático y tectónico), la migración de las placas a diferentes latitudes y la expansión gradual de los océanos.

Palabras clave: Sistema alpino, Atlas, Béticas, pizarras negras, CCD, plataformas carbonatadas, colisión continental, Glaciaciones (Würm; Wisconsin), erosión marina profunda, evaporitas, Pirineos, Rif, *Rifting,* Cambios del nivel del mar, expansión oceánica, sedimentos silíceos, subducción, subsidencia, volcanismo.

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1. INTRODUCTION

The Iberian Peninsula forms a structural link between the Eurasian and African plates. From mid Jurassic to early Cretaceous it was part of the Eurasian Plate, from early Cretaceous to the Eocene it was a semi-independent plate, ands since the Oligocene it has been again part of the Eurasian Plate. The Permian to the present paleogeographic maps compiled from numerous published sources and described below illustrate the evolution of this interesting geologic terrain.

2. PALEOZOIC

The Paleozoic orogenic belts circling the North Atlantic Ocean are the result of the opening and closing of a latest Precambrian-Paleozoic Proto-Atlantic or Iapetus Ocean (Wilson, 1966). Tholeiitic basalts beneath transgressive lower Cambrian rocks in the Appalachians indicate that the opening began in latest Precambrian (Dewey and Kidd, 1974). Uppermost Precambrian Tayvallich Volcanics and lower Cambrian sediments indicate that the separation of the British Isles and Greenland also took place at this time. In Scotland there is evidente for three episodes of subduction extending from the Precambrian 750 m.y. ago (Moravian event) to the Ordovician Grampian event (Garson and Plant, 1973). In the Baltic region seafloor spreading marked by late Precambrian magmatic activiy followed by an early Cambrian transgression was preceded by a la te Precambrian metamorphism and tectonism resulting either from subduction or plate collision (Dewey, 1969: Baker, 1973).

The Proto-Atlantic attained its maximum width during late Cambrian to early Ordovician. It began to close soon after with the establishment of subduction zones along eastern North America, the northwest Baltic Shield, along the northern margin of the Avalon terrain, and within the oceanic basin itself. Subduction along eastern North America produced the Taconic/Grampian orogenies in the early Ordovician. Collision of the Baltic shield and North America-Greenland produced the late Ordovician to early Devonian Caledonian orogeny. When northwestern South America or northwestern Africa collided with North America it resulted in the middle Devonian Acadian orogeny. Northeast motion of North America-Baltic shield produce the late Devonian-early Carboniferous Fundian rift system in eastern Canada and the rift system cutting across the Caledonian orogen in the British Isles.

As the North America-Baltic shield block moved northeastward it rotated clockwise resulting in the collision of Europe with Africa north of the South Atlas fault in late Carboniferous (Hercynian orogeny), and the collision of Africa south of the fault with southeastern North America during early Permian (Alleghenian orogeny). Collision of North and South America during the Permo-Carboniferous is expressed by the Ouachita orogeny. Consolidation of the North American, South American, Eurasian, and African blocks led to the formation of a sinusoidal band of mountain chains wrapped around the older cratonic blocks and stable platforms (Hurley, 1974). This mega-continental mass known as Pangea was divided by an east-west trending ocean, the Tethys, into Eurasia to the north and Gondwanaland to the south. The present morphology of the North Atlantic is the product of the Mesozoic breakup of Pangea along rift valleys linked to plume-generated triple junctions (Burke and Dewey, 1973: Rankin, 1976).

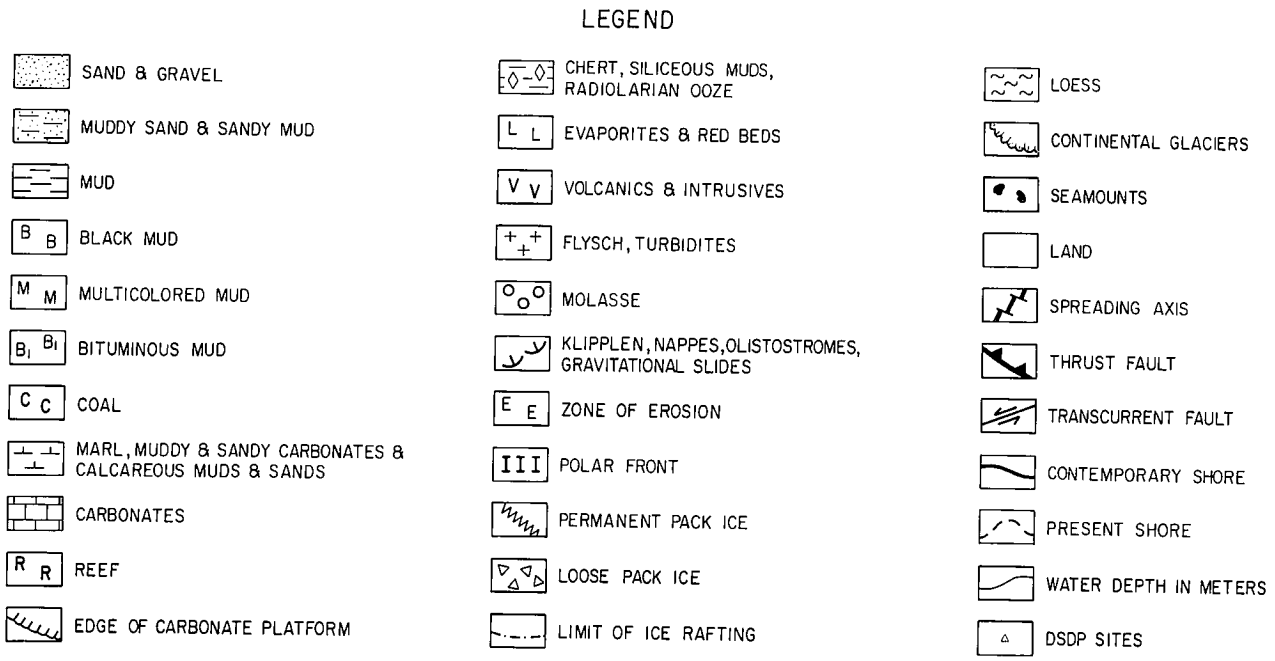

Fig. 1.—Legend of paleogeographic maps.
Fig. 1.—Leyenda de los mapas paleogeográficos.

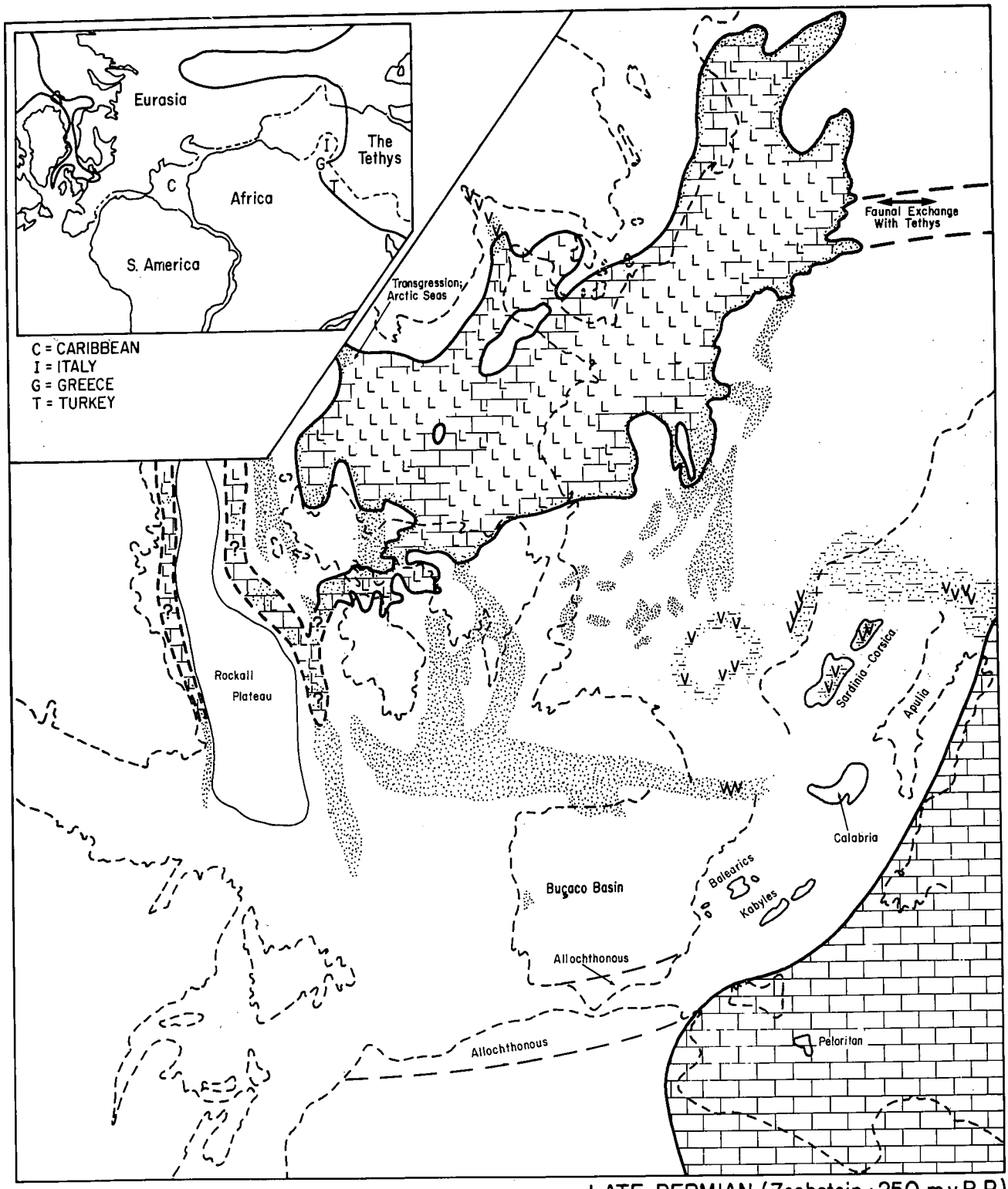

Fig. 2.—Paleogeography of the late Permian. (Insert map shows Permian reconstruction by Smith and Woodcock, 1982).
Fig. 2.—Paleogeografía del Pérmico Superior. (El encuadre muestra la reconstrucción pérmica de Smith y Woodcock, 1982).

3. PERMIAN (fig. 2)

Permian paleogeographic maps as the one illustrated in fig. 2 display a large wedge-shaped ocean between Eurasia and Gondwanaland. Geological evidence (Owen, 1976: Ahmad, 1978: Crawford, 1979), however, does not support the large oceanic basin implied by such a reconstruction. To overcome this objection Smith and Woodcock (1982) have proposed the Permian reconstruction shown in the insert in fig. 2 in which the southern and northern continents were much closer with South America located south of western Eurasia. Smith and Woodcock (1982) have proposed that this Permian Pangea evolved into the Late Triassic Pangea shown in fig. 3 by motion along a transform zone along the northern edge of Gondwanaland. As there is no evidence for such a transform zone I have used the Permian reconstruction where Pangea is assumed to be a rigid supercontinent which only underwent rotation throughout the Permo-Triassic.

Concurrent with and following the consolidation of the cratons Pangea underwent extensive erosion. The first manifestations of movements that led to the opening of the present North Atlantic began in the Permian with the collapse of the uplifted and eroded Variscan and older systems, the infra-rift phase of Falvey (1974). In eastern Greenland the late Permian sequence deposited in a gulf open to the northeast consists of reef carbonates, gypsum, black muds and red beds deposited in a subtropical to semi-arid climate at a paleolatitude of 20° to 10° (Surlyd *et al.,* 1981). In the Svalbard, the Barents Sea and the eastern Norwegian margin the Permian is represented by red beds and evaporites (Habicht, 1979). In western and central Europe which had drifted northward from an equatorial position into the trade wind belt early Permian sediments deposited in intramontane and peripheral collapse basins consit of gravels and sand along the edges of the lows, and red shales, anhydritic shales, and halites in the axial parts of the basins (Ziegler, 1982).

Along the western edge of the Tethys the Permian is developed in a carbonate sequence (Kamen-Kaye, 1972). Away from the North Sea and the Tethys the Permian is represented by red beds and volcanics that are similar to the overlying Triassic interval (Ager, 1980). In the Western Alps the Permian consists of coarse continental deposits with acidic intrusions and extrusions. Permian conglomerates, red sandstones and mudstones with land floras, and red rhyolites and quartz-porphyries which flowed from a land mass to the south, possibly Corsica, occur in the region of the Maures and Esterl massifs. In Corsica are ring dikes, rhyolitic intrusions and extrusions and ignimbrites, and in Sardinia the Permian consists of clastics with an Autunian flora intruded by a late tectonic phase quartz porphyry. In the Massif Central the Permian is represented by lacustrine-swampy, fluvial sediments, and "Saxonian" coarse red sediments with a distinct late Permian flora (Ager, 1980). Along the southern border of the central and western Pyrenees are Permian andesitic lavas capped by postorogenic red beds including thick conglomerates. Red sediments also have been reported from the Bucaço basin in Portugal (Alvarado, 1980). Permian quartzites and conglomerates have been described from the Betics and the Middle and High Atlas in northern Africa (Ager, 1980: Stets and Wurster, 1980).

In the late Permian (Tatarian) continued subsidence combined with a glacio-eustatic rise in sea level (end of glaciation in southern Gondwanaland) formed a pasageway connecting the basins in the Arctic, Barents Sea, northern and eastern Greenland, and the basins in western and central Europe (Ziegler, 1982). Possibly the passageway extended all the way to the Tethys by way of Poland. A combination of fluctuations in sea level and/or vertical tectonic oscillations causing instability in the Arctic-North Sea pasageway led to the deposition of the cyclical Zechstein sequence of carbonates (mounds and algal banks; transgression), and anhydrite banks, massive anhydrite and polyhalite beds (regression).

Like the Barents Sea and eastern Greenland where subsidence persisted throughout the Permian (Vischer, 1943: Haller, 1970: Ronnevik, 1981), the East Greenland, Rockall-Faeroe, Bay of Biscay, Porcupine, Celtic Sea, Bristol Channel, and Western Approaches troughs may have begun to subside in the Permian and were sites of continental and shallow, marine deposition. It is at this time that the evaporites that gave rise to the diapirs of eastern Greenland at 61°-62° N (Johnson *et al.,* 1975) may have been deposited.

4. TRIASSIC AND EARLY JURASSIC (fig. 3).

The Triassic was initiated by a regional regression of the Arctic Sea and the return of arid continental conditions to western and central Europe (Ziegler, 1982). As the breakup of Pangea accelerated a complex system of interlocking horsts and grabens expanded from eastern Greenland to the northern coast of South America and equatorial west America and equatorial west Africa (Burke and Dewey, 1973). Concurrent with this crustal extension massive shear faulting affected the northern and southern margins of Iberia and the northern margin of Africa (Manspeizer *et al.,* 1978). The Bay of Biscay rift and Aquitanian basin began to subside slowly during early and middle Triassic with the rate increasing during late Triassic. Wrench movements along the Bay of Biscay rift led to crustal extension in the French and Irish shelves, and on the Lusitania-Canadian Maritime provences-New England platform. Shearing in north Africa began in late Triassic (Norian) and in eastern North America somewhat earlier (Ladinian/Carnian). The onset of rifting in the Danish trough began in early Triassic and the North Sea structures which cut across the Caledonian and Permian basins structural grain also began to subside at this time (Ziegler, 1982). The Oslo graben, however, displays no evidente of Triassic subsidence. Rifting in eastern Greenland was initiated in latest Triassic-earliest Jurassic. The Rockall-Faeroe rift and East Greenland-West Norway

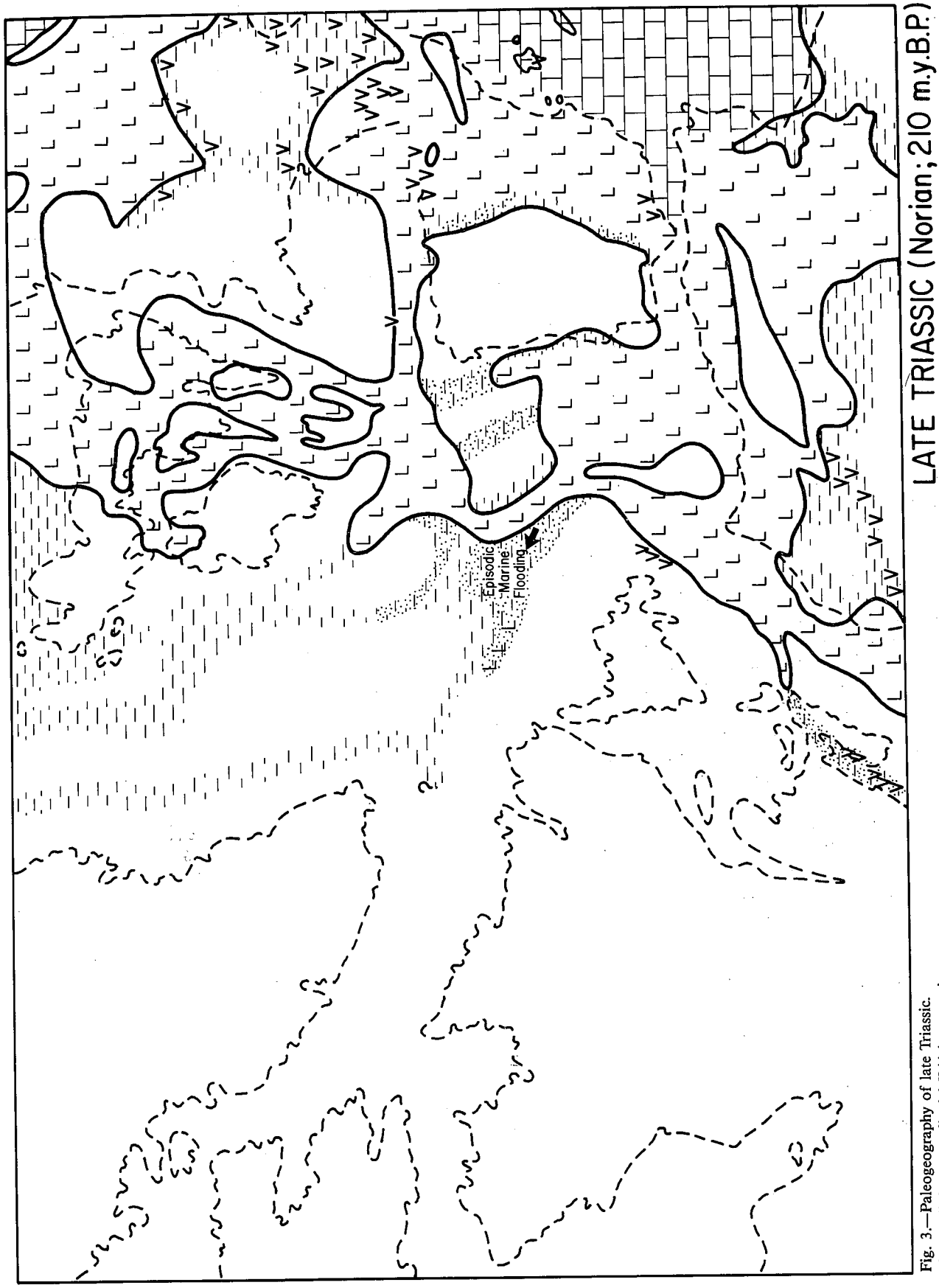

Fig. 3.—Paleogeography of late Triassic.
Fig. 3.—Paleogeografía del Triásico superior.

rifts also began to subside at this time. Suth of the North Sea Triassic rifting propagated into the Paris basin by way of the Hessian depression and Burgundy trough along zones of weakness established in late Carboniferous and early Permian destroying the Variscan fold belt.

Volcanism in the Triassic rifts of western and central Europe is at a low level, but the Aquitanian basin, Bay of Biscay rift, and the extensional structures in north Africa and eastern North America are characterized by extrusion and intrusion of lavas. The extrusion of basaltic "ophites" in the Bay of Biscay rift which peaked during the late Triassic and earliest Jurassic (Stévaux and Winnonck, 1974) was probably induced by contemporaneous wrench movements (Ziegler, 1982). The intrusions along the left-lateral Messejana fault in sothwest Iberia are of the same origin (Schott *et al.*, 1981). Initially andesitic lavas were emplaced along the shear zone in north African and granites (older White Mountain Magma Series) were intruded along the structures in eastern North America. During the final phases of rifting basalt lavas were extruded and intruded into the rift structures. Similar middle and late Triassic magmatic events also con be seen in the Alpine Mediterranean region (Bernoulli and Jenkyns, 1974).

Depositional patterns in the rift basins were influenced by syndepositional tensional tectonics, frequent sea level oscillations, a gradual rise in sea level which peaked in the early Norian, a mild regression during mid-Norian, and a transgression (Rhaetian) at the close of the Triassic (Haq *et al.*, 1987). Gravels, red sands and muds with evaporites occur in the High and Middle Atlas, and the Pre-Rif. Throughout most of western Europe the Triassic is developed in three facies (Germanic facies): lower Triassic (Buntsandstein) red bed deposited under continental to lacustrine, brackish marine to hypersaline conditions, middle Triassic (Muschelkalk) carbonates grading northward into evaporitic and dolomitic marls deposited during a relative rise in sea level, and late Triassic (Keuper) clastic evaporitic unit emplaced during a regression. During Keuper time when climatic conditions varied between humid and arid an appreciable thicknees of evaporites accumulated in lows (Ziegler, 1982). On the Betics are quartzites, gravel, gypsum, carbonates and muds with intercalations of gypsum. In this range also are Alpine type early and mid to late Triassic limestones. Red beds with evaporites, "Muschelkalk", and intrusive dolerites occur in the Pre-Betics, and continental red marls and well bedded carbonates with chert in the Sub-Betic. A similar sequence is found in the Celtiberian Chain, Catalonian Cordillera, Ebro basin, and in the Vizcaya basin (Western Pyrenees). In Portugal the Triassic is dominated by sandy deposits and evaporites.

In the Aquitanian basin north of the Pyrenees sedimentation began at the southern end of the basin and spread northward. Along the borders of the Massif Central are red sands and gravel. In the center of the basin are varigated muds with dolomite, anhydrite, and salt. On the southern end of the basin is a well developed Muschelkalk between Keuper anhydrite and salt above and Bunter with salt below (Ager, 1980). Another feature of interest in this part of the basin is the occurrence of dolerite sills. In northern Europe much of the Triassic strata are developed in continental to brackish marine red beds, shallow-marine carbonates, sulphates, and halites (Ziegler, 1982). In eastern Greenland are Triassic open marine shales, deltaic clastics, and subordinate evaporites. Marine influence was weak in eastern Greenland during the Rhaetian-Hettangian and sedimentarion was in fluvial and restricted marine environments (Surlyk *et al.*, 1981).

In Provence, southern France are gypsum and dolomites deposited in a broad trough around the Maures and Esterel massifs, and the Triassic in Corsica, Sardinia, and Sicily is developed mainly in carbonates. In the Alps, Apulia, and Dinarides are volcanics, continental red beds, orthoquartzites, evaporites, and tidal and sabkha platform carbonates. A Carnian rift episode ended platform deposition in the Western Alps and there was an influx of muds and sands, and the deposition of thick evaporites overlain by carbonates stromatolitic mats (Lemoine *et al.*, 1986). A second carbonate platform was initiated in the Norian. Its development was in two steps, the Hauptdolomit and Rhaetian cycles. The Rhaetian cycle consisting of coquinas, micritic limestones, calcareous blanc shales, and dolomitic mudstones is coeval with a transgresion. It bears evidente of some extensional tectonics that reached their peak in early Jurassic.

During the Rhaetian transgression a connection may have been established for the first time between the Arctic and the Tethys via the Irish Sea and the Rockall-Faeroe rift (Ziegler, 1982). At that time marine shales were deposited in the lows from the Celtic Sea to the Paris basin. Their margins underwent uplift and erosion during the Early Cimmerian distension phase that encompasses the latest Triassic to earliest Jurassic (Ziegler, 1982).

As Pangea continues to breakup and the rift system slowly subsided below sea level, it was flooded by marine waters terminating evaporite deposition and carbonates became dominant. In the High Atlas east of the Tichka massif a long narrow gulf extended across the Maghreb from the east in early Jurassic. Distal tubidites were deposited in the deeper parts of the trough. On the sides of the low and its western end are shallow-sub-tidal facies with reefs developed at several levels. Behind the reefs are lagoonal and supratidal carbonates (Ager, 1980). In the Middle Atlas are coral reefs with their more massive compound cores facing the open ocean to the northeast. Here also are shallow water carbonates, and an ammonite bearing muddier facies with a Tethyan fauna. On the Oran (High) Plateau in the *Pays des Horsts* are earliest Jurassic poorly fossiliferous dolomites. The middle Lias is represent by shallow facies including molluscan reefs, and the late Lias by a deeper water facier (Ager, 1980).

As the North Atlantic opened and Africa moved westward relative to Iberia a series of highs and lows arranged "en-echelon" were formed along southern Iberia, the North African Basin of Malod (in prepara-

tion). These highs and lows are comparable to those described from the translation margins of the Gulf of Guinea in equatorial West Africa, northern Brazil, and along the southwest margin of the Grand Banks of Newfoundland (Emery and Uchupi, 1984 and references therein: Keen and Haworth, 1985). As the lows (Sub-Betic and Pre-Rif) subsided rapidly from early to late Jurassic the Triassic marls and well bedded limestone with chert were mantled with radiolarian muds and red nodular carbonates. The highs continued to be sites of carbonate accumulation. The Pre-Rif zone or "Tellian Furrow" was uplifted above sea level at the end of early Jurassic, the first important movement of the Alpine orogeny (Ager, 1980).

Shallow carbonates are the dominant early Jurassic facies in the Pre-Betics, the Celtiberic Chain, Ebro basin, Catalonian Cordillera, and northern Spain with a fauna having strong Boreal affinities. In Portugal the early Jurassic is developed in thin beded carbonates with mud interbeds. Shallow water carbonates with a Boreal fauna also are present on either side of the Pyrenees. In early Jurassic basement was emergent along the present coast of France from the Bay of Arcachon to Biarritz, and a broad ridge north of Bordeaux divided the Aquitanian basin in two. At the beginning of the Jurassic a marine incursion flooded the basin from the southeast, and shallow water carbonates were deposited around the periphery of the basin and evaporites in its center. By the end of early Jurassic the basin attained its present configuration and shallow marine carbonate deposition prevailed throughout the region.

In the Paris basin thin transgressive Hettangian carbonates and bituminous muds are capped by a cyclical sequence of Sinemurian to Toarcian muds and carbonates. The bulk of the Liassic series in the North Sea, the English Channel, and the Irish and French shelves consist of open marine muds. In the northern North Sea are sand that persisted into early Sinemurian. Above them are late Sinemurian to Aalenian muds. By late Sinemurian the Arctic and Tethys were linked up via the North Sea.

In the Massif Central, Balearics, Provence, and Corsica the Early Jurassic is developed in shallow carbonates. In Sardinia are shallow water carbonates in the west and a deeper facies in the east. In the Alpine system the Triassic carbonate plaforms began to breaks up as block faulting became widespread. The first extensional phase took place during the Hettangian-earliest Sinemurian following by a quiet period during the Sinemurian-Pliensbachian. A second tensional phase took place during the Domerian to Toarcian with the distensional activity ending sometime between the Toarcian and Oxfordian. The highs continued to be sites of shallow water carbonate deposition, and the lows were sinks for reef detritus, turbidites, and gray to black limestones (Lemoine *et al.*, 1986: Bernoulli and Jenkyns, 1974). There ins litle volcanic activity associated with this extensional phase.

In North America a slow marine transgression (Rhaetian) to earliest Jurassic is reflected at the northeast edge of the Grand Banks by a hypersaline lagoonal and tidal flat dolomite-halite-anhydrite unit, a lagoonal-inner shelf dolomite-anhydite unit, and an outer shelf limestone unit. Above are calcareous and non-calcareous muds. To the southwest are shallow water outer shelf high energy environment carbonates. Off Nova Scotia the early Jurassic is represented by supratida muds and sands, dolomitic muds deposited in a sabkha environment, and peloid carbonates deposited on high energy outer shelf environment, and peloid carbonates deposited on high energy outer shelf environment. Off New England the early Jurassic is represented by siliclastics inshore, carbonate mounds resting on topographic higs farther offeshore, and carbonates on the outer shelf (Jansa and Wade, 1975).

Incursion of early Jurassic marine waters into the Atlantic rift system was mainly from the north. The transition of marine facies in the east to intertidal and continental facies to the west indicates that there was no passageway through the High Atlas for Liassic Tethys waters to reach the North Atlantic. The lower Lias of the Lusitanian basin in Portugal also has a Subboreal fauna indicating that basin was separated from the Tethys at that time by either a shallow sill or a subaerial ridge connecting the Iberian peninsula with the Grand Banks (Lancelot and Winterer, 1980). Similarities between the lower Liassic transgressive section in the Lusitania basin and in Grand Banks suggest that these two depocenters were connected at that time. The Pliensbachian-Toarcian transgressive section farther south off Nova Scotia and Georges Bank is comparable to the facies in western Morroco whose fauna displays affinities to those of the Lusitanian basin. Even if this transgression was from the Tethys it was limited (Lancelot and Winterer, 1980).

5. MIDDLE AND LATE JURASSIC (FIGS. 4 AND 5).

Sea-floor spreading in the North Atlantic probably began in early Bajocian about 183 m.y. ago (based on the scale by Palmer, 1983). I equate this event with the Mid-Cimmerian orogeny, and the early Bajocian disruption of the connection of the Tethys with the eas tern Pacific via northern South America (Westermann, 1975: Westermann and Riccardi, 1976). Others place the event earlier suggesting that the opening of the North Atlantic took place in early Toarcian when the first mixing of Subboreal and Tethys faunas took place (Lancelot and Winterer, 1980), or during the latest Toarcian regression when Aalenian/Bathonian sands prograded over the Lias carbonates on the Nova Scotia shelf (Jansa and Wade, 1975).

On the western High Atlas Middle Jurassic is developed in conglomerates, red sandstones, clays and dolomites. Offshore on the west African shelf the facies are predominantly carbonates with subordinate sands. The late Jurassic transgression cycle in the High Atlas is recorded by carbonates and marls and near Cape Rhir by thick-shelled faunas of a high energy environment. Three regressive phases are documented by the lower

Kimmeridgian and Tithonian red clays and evaporites. Platform carbonate deposition ended during the Berriasian regression when siliciclastic deposition was initiated (Jansa and Wiedmann, 1982). The last regression is associated with the uplift of the eastern High Atlas and Anti-Atlas (Schlager, 1980: Stets and Wurster, 1982). Offshore the late Jurassic is represented mainly by a masive carbonate platform. Seaward of the platform are pelagic to shallow water carbonates, a reflection of the diversity of depositional environments in the region.

In the Middle Atlas are middle Jurassic carbonates, and muddy carbonates and muds of Bajocian and Bathonian age. From mid Bajocian the marine deposits began to diminish, and plant debris together with fresh-water vertebrates (including a dinosaur deat asemblage) became abundant (Ager, 1980). Middle Jurassic dolomitic sediments are widespread on the Oran Plateau. By late Jurassic, however, the sea retreated northward depositing shallow water carbonates in *Pays des Horsts.* These carbonates grade eastward into sands in Algeria. The highs (Rif/Betic) between Africa and Iberia were sites of shallow water carbonate deposition and the lows (Sub-Betic, Pre-Rif) of radiolarian muds and red carbonates.

The middle and later Jurassic in the Pre-Betics, Celtiberic chain, Ebro basin, Catalonian Cordillera, and Vizcaya basin (western Pyrenees) are developed in shallow water carbonates. In Le Danois Bank are Bajocian to Tithonian capionellid micritic limestone and a shallow platform carbonate facies. The proximity of these two facies indicates that the bank must have been on an outer shelf/upper slope environment from late Jurassic to early Cretaceous. Near the end of the middle Jurassic the sea retreated and marine deposition ended in Santander in the Callovian and at the eastern end of the basin in the Oxfordian. Along the western side of the Iberian Peninsula are middle Jurassic argillaceous and oolitic limestones inshore and pelagic carbonates offshore. The late Jurassic sediments deposited after a short Oxfordian regression consit of neritic and reefal carbonates (Groupe Galice, 1979). On the low between the coast and Galicia Bank are Jurassic clastics or carbonates deposited above the Carbonate Compensation Depth (CCD) in a slowly subsiding basin. On Galicia Bank itself are shallow water carbonates and calpionellid pelagic micrites.

On the external and internal zones on both sides of the Pyrenees are middle and late Jurassic shallow-water carbonates. In the Aquitanian basin are shallow water carbonates to the east and argillaceous sediments with abundant ammonites to the west. In the Paris basin middle and late Jurassic are developed in platform carbonates. In earliest Callovian a combination of rising seas and crustal attenuation reoponed the seaway connecting the Tethys and the Arctic allowing the migration of boreal faunas into Paris basin and as far south as the Lusitania basin in Portugal.

During the opening of the North Atlantic the central North Sea was uplifted and volcanic center was initiated (Ziegler, 1982). Debris from the high were shed both northward and southward with Bajocian-Callovian deltas prograding southward from the Ringkobing-Fyn high. Shallow marine mud deposition predominated on the Celtic and French shelves. The only evidence of volcanism in this region are the Bathonian dolerite sills on the western extension of the Celtic Sea basin. Bajocian and Bathonian shallow water carbonates and shales overlie Lower Jurassic marine shales in the Channel area. In the northern Porcupine Trough are regressive continental clays and minor sands. In late Jurassic the southwest Norway-Faeroe rift was uplifted, an uplift accompanied by volcanic activity. Detritus from this high was deposited in the subsiding North Sea rift. As subsidence of the rift outpaced sedimentation and rising sea level it led to the deposition of deep water organic shales from the Callovian to the Kimmeridgian. At this time the outer shelf and upper slope west of the Western Approaches was a carbonate platform of Tethyan affinity that extended as far north as 48°N (Auffret *et al.,* 1979). In east Greenland rifting became pronounced in middle Jurassic with the low becoming a sink for Bajocian to earliest Oxfordian shallow water clastics, upper slope black shales and deep-water fans associated with a late Oxfordian distensional phase. Rifting culminated in the Tithonian.

In the Balearics and Corsica are middle to late Jurassic shallow carbonates, and in Provence are shallow water carbonates and Tithonian bathyal micrites. Westward of this region are Tithonian reefs. Easterly tilting in Sardinia become more pronounced at this time and the middle and late Jurassic section is more pelagic. In Sicily middle and late Jurassic carbonates are developed in both shallow and deeper water facies including volcanics and radiolarian muds. In late middle Jurassic or early late Jurassic Apulia broke away from Europe and seafloor spreading began in the Ligurian Tethys. Radiolarian cherts were deposited on top of the newly formed oceanic crust (Lemoine *et al.,* 1986). The Tethyan shelves including the Apulian platform Gagano zone in Italy, Venice basin, the Adriatic, the Dalmatian zone of the Dinarides, and the Gavrovo and Ionian zones of the Hellenides were sites of shallow carbonate deposition during middle and later Jurassic. The middle Jurassic facies on the tablemount-basin topography that characterized the northern margin of the Tethys range from red pelagic limestone with manganese nodules, subtidal stromatolites, crinoidal sands, and Lumachelle banks on the highs, and gray pelagic carbonates and marly pelagic carbonates on the lows. As the higs subsided the were mantled with red nodular limestone without manganese nodules and plegic pellet and "oolitic" facies in late Jurassic. The lows were sites of red marls, pelagic carbonates, and radiolarian oozes deposited below the CCD.

By late Bathonian-early Callovian the oceanic North Atlantic formed a continuous path from the Tethys to the eastern Pacific via northern South America (Westermann, 1975: Tucholke and McCoy, 1987). About 170 m.y. ago (latest Bathonian; Blake Spur magnetic anomaly time) the mid-ocean spreading axis jumped eastward toward the African margin stranding a segment of the African Plate with the North American Plate

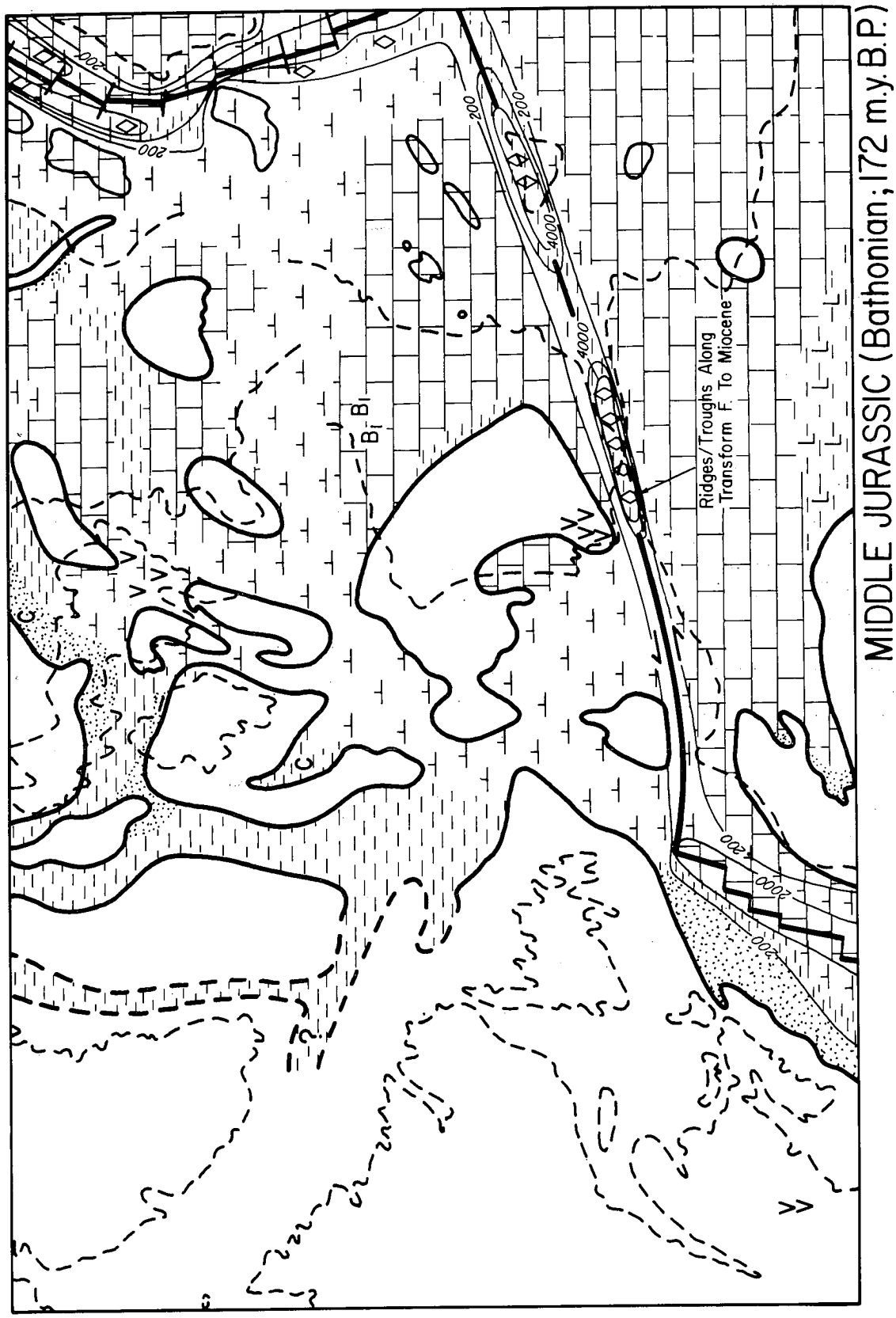

Fig. 4.—Paleogeografía del Jurásico medio. Contornos de profundidad en metros (como en los otros mapas) son de Tucholke y McCoy (1986). La topografía indicada por la región del Estrecho de Gibraltar es esquemática en las figuras 4 a 12. La morfología era muy irregular consistiendo en surcos y umbrales desde 2000 m a más de 4000 m de profundidad , con un estilo geométrico que cambiaba con el tiempo.

Fig. 4.—Paleogeography of middle Jurassic. Depth contours in meters on this and other maps are from Tucholke and McCoy (1986). The topography shown for the Straits of Gibraltar area in this and figs. 5 to 12 is generalized. The morphology of the region was very irregular consisting of linear highs and lows ranging in depth from less than 2000 to more than 4000 m, a geometric pattern that was in constant change with time.

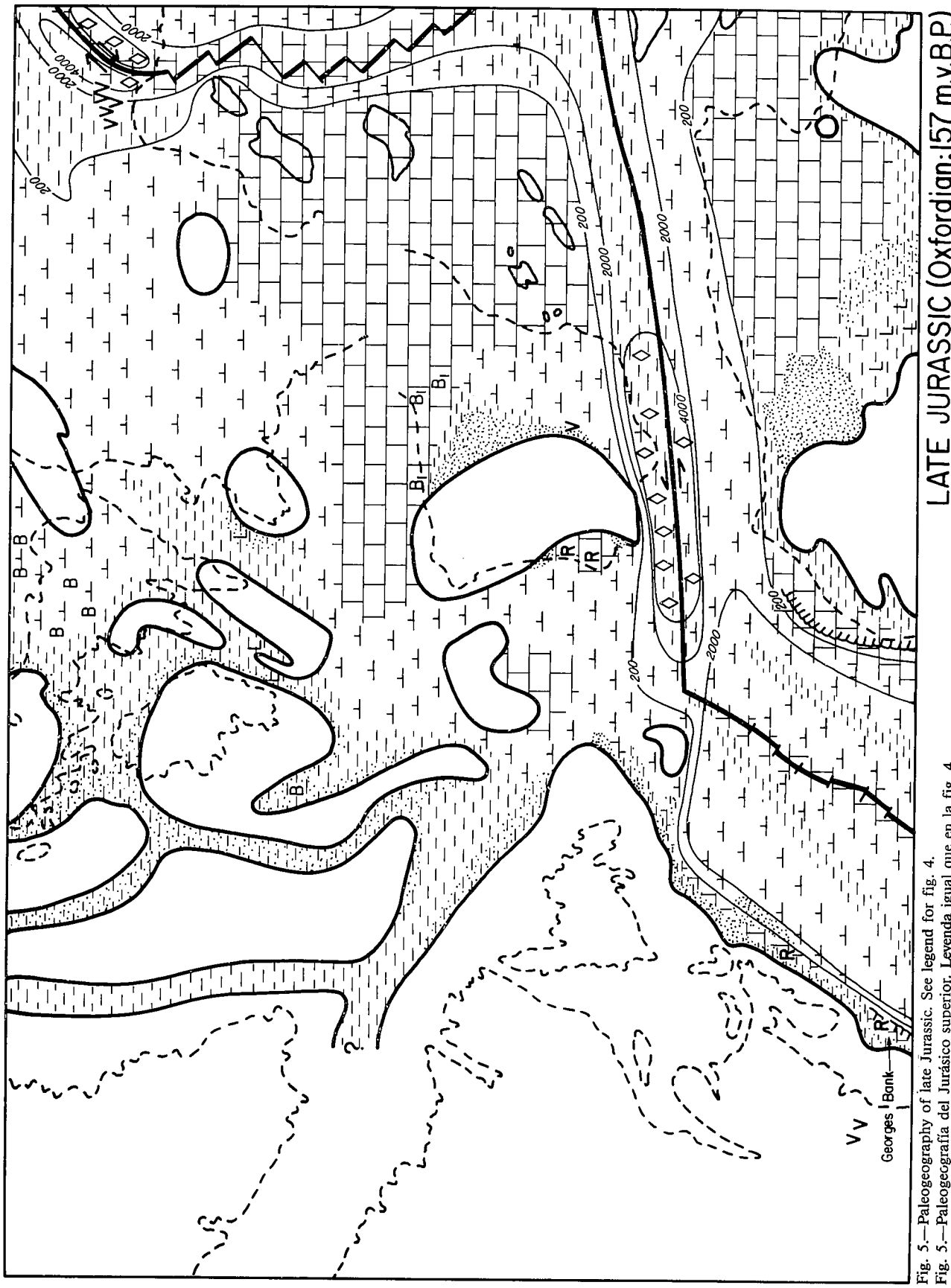

Fig. 5.—Paleogeography of late Jurassic. See legend for fig. 4.
Fig. 5.—Paleogeografía del Jurásico superior. Leyenda igual que en la fig. 4.

(Tucholke and McCoy, 1986). West of the narrow basins associated with the Betic/Rif transform fault system most of the oceanic North Atlantic was sediment starved at this time. The carbonate banks and reefs along the African and North American margins acted as dams blocking significant input from the adjacent continents. During late Jurassic the oceanic basin was a site of calcareous muds near the continents, and carbonate poor muds below the CCD at a 3 km depth (Tucholke and McCoy, 1986). Along the crest of the mid-ocean ridge were pelagic carbonates.

On the Grand Banks of Newfoundland, and the Newfoundland and Labrador shelves is a platform sequence of middle Jurassic calcareous shales and shallow water carbonates, and Bathonian to Kimmeridgian sandstones and shallow water carbonates. Orphan Knoll, a topographic high seaward of the Newfoundland shelf, was a site of calcareous carbonaceous sands and mud deposition in the Bajocian. During the opening of the North Atlantic, sediment sources in the Nova Scotia region were rejuvenated and clastics spread over the early Jurassic carbonates. Carbonate deposition was renewed in late Bathonian, a depositional cycle that was terminates with the deposition of muds during the Callovian transgression. A rapid shoaling during the Callovian/Oxfordian transition was followed by a deepening transgressive phase. This the decisive major invasion of the Atlantic Basin by the Tethys is recorded by the massive carbonate platforms on the African and North American margin. Off New England during the lower Bathonian regression sands prograded seaward and except for the outer shelf, terminated carbonate deposition. During the late Bathonian transgression carbonate deposition was initiated and continued until the end of the Jurassic.

6. LATEST JURASSIC AND EARLY CRETACEOUS (FIGS. 6 AND 7).

During a transgression in the Berriasian the Jurassic reefs on the north African continental shelf were drowned slowly (some were already drowned by early Valanginian) and carbonates deposition shifted landward of the present coast. As the sea advanced eastward sedimentation decreased in the oceanic basin, the Moroccan Meseta was flooded, and marine deposition extended to the Tichka massif along the eastern edge of the western High Atlas. During the transgression, which lasted until early Hauterivian, a coral reef developed parallel to the present coast. Landward of the reef are early Cretaceous marls, and patch reefs atop diapiric structures. Gaps in the reef served as passageways for turbidity currents that deposited a deep-sea fan at the base of the reef. Carbonate banks on the outer shelf were another source of the turbidites emplaced in the oceanic basin. During the late Hauterivian regression (or post-Valanginian according to some writers) a wedge of clastics, "Wealden facies", was deposited on both the African and North American margins of the Atlantic which at that time was only 1000 to 1500 km wide. This clastic wedge grades seaward into turbidites. The Aptian-Albian in the western High Atlas above the Wealden consists of a basal gravels, sands, marls and carbonates.

The early Cretaceous history of the High Atlas east of the Tichka massif is confusing with many local facies changes, uplift first in the north and then in the south (Ager, 1980). At the eastern end of the chain Cretaceous sediments prograded southward from southerly sources. The fauna in the region are northwest European and lack many of the "Alpine" elements present in the Rif (Ager, 1980). Latest Jurassic and early Cretaceous sediments are absent from the Middle Atlas and Oran Plateau and are scarce in the Moroccan Meseta. The early Cretaceous where present is developed in muds and carbonates.

From late Jurassic (Oxfordian) to earliest Cretaceous (Valanginian) calcareous flysch was deposited in the North Africa Basin, the low separating north Africa and Iberia (Malod, in preparation). In the Ultra-Betics are early Cretaceous carbonates with a deep water fauna of ammonite aptychi, belemnites, and aberrant pygopid brachiopds (Ager, 1980). From Barremian to Albian, large amounts of flysch were deposited in the basin, the Mauritanian unit to the north and the Massylian one to the south. Slight convergence of Africa and Iberia at this time produced a hiatus between early and late Cretaceous in the Sub-Betics (Alvarado, 1980). The early Cretaceous in the Pre-Betics consists of a shallow water limestone/marl sequence. In the Celtiberic chain the early Cretaceous consists of a Wealden-like shallow water or brackish facies followed by massive Urgonian (Barremian-Aptian) carbonates. In the Catalonian Cordillera the early Cretaceous succession of shallow water carbonates including an Urgonian facies of massive rudist bearing reef limestone (Ager, 1980).

Toward the end of the Jurassic Iberia began to break away from the North American continent. During this tectonic event which culminated with the initiation of seafloor spreading in the Bay of Biscay-Rockall Trough in middle Aptian-early Albian, source areas were rejuvenated, large areas of the shelves were uplifted and underwent erosion, and the mid-ocean ridge extended northward separating the southwest edge of the Iberian peninsula from North America. The rift axis that extended eastward from a triple junction at the Mid-Atlantic Ridge and was to lead to opening of the Bay of Biscay become the primary plate boundary. The boundary along the southern margin of Iberia became secondary (Schouten *et al.*, in preparation). Thus, from the end of the Jurassic to the end of the Eocene Iberia was attached on and off to the African Plate.

During the regression associated with this rifting phase (Late Cimmerian orogeny) Lusitania Basin was a site of red clay and sandstones deposition. The narrow graben between the coast and Galicia Bank was formed at this time or was reactivated. Forming the foundation of the low are Hauterivian platform carbonates and dark-colored muds (Sibuet and Ryan, 1979: Sibuet, Ryan *et al.*, 1980). This basin was either perio-

dically stagnated or there was a periodic migration of the oxygen minimum. Above the carbonates are Barremian distal tubidites and mud flows, and early Barremian to Cenomanian dark muds with abundant carbonaceous material derived from land plants. Galicia Bank west of the graben was a site of shallow-water carbonates on the topographic highs and calpionellid pelagic carbonates in the lows (Dupeuble *et al.,* 1976).

In North America detritus eroded from the uplifted Grand Banks of Newfoundland (Avalon upflit) were deposited along the periphery of the high as a series of alluvial, deltaic, and prodelta facies grading seaward into marine clays. Similar type of facies also occur on the Labrador shelf to the northeast. On Orphan Knoll, a topographic high at the eastern edge of the Newfoundland basin, the syn-rift/early drift episode is represented by an Albian sandy glauconitic carbonates resting unconformably on Bajocian calcareous carbonaceous sand and muds. On the Nova Scotian margin the the southwest sedimentation changed from predominantly carbonate-muds to a paralic siliciclastic one although carbonates continues to be deposited for some time along the fringes of the early Cretaceous delta (Given, 1977). It is also at this time that rifting began in the Labrador region. The syn-rift sequence here consists of fluviodeltaic clastics and 140 to 110 m.y. old volcanics.

The rift sequence along the north coast of Iberia is made up of a latest Jurassic to early Aptian Wealden Facies, a clastic lagoonal facies along the southern rim of the Vizcaya basin, and a thick fluvial unit at the western end. In the west a hiatus separates these sediments from the Jurasic deposits. This Dogger to middle Valanginian hiatus is well developed in Santander (Hines, 1985). The Ebro massif to the southeast broke into a serie of lows and highs during the rifting phase with the lows being sites of Wealden sedimentation followed by massive urgonian limestone deposition.

From mid-Aptian to early Albian the Mid-Atlantic Ridge extended northward separating Iberia and North America, and an incipient spreading axis separated the Rockall Plateau from Eurasia. The mid-ocean ridge southeast of the Grand Banks was quite shallow with some parts serving as foundations for carbonate banks and others rising above sea level to form islands (Tucholke and Ludwing, 1982). After this the shallow segment subsided to normal depths and was divided in two by seafloor spreading to from the J-Anomaly Ridge on the west and Madeira-Tore Rise on the east. As Iberia and North American drifted apart volcanic activity along leaky transform faults built the Southeast Newfoundland Ridge, and the Newfoundland and Fogo seamounts off the Grand Banks. Some of the seamounts became sites of carbonate deposition (Sullivan and Keen, 1977: Uchupi and Austin, 1979).

During the late early Cretaceous drift phase the Ebro massif was a site of Wealden deposition with brown coals and thin oyster banks (Ager, 1980). In Vizcaya Basin which underwent strong but differential movement during the Aptian-Albian, the drift sequence is made up of rudistid and coral reefs in the highs and muds in the lows. Around the southern rim of the basin are fluvial, paralic, and littoral siliciclastics. As Iberia continues to drift southwestward relative to Europe (Grimaud *et al.,* 1982) the "Black Flysch" trough was formed to the northeast and became a site of flysch and wildflysch deposition including black mud (see Crimes, 1976: Malod *et al.,* 1980: Hines, 1985: Pujalte, 1985: García-Mondejar, 1985: García-Mondéjar and Pujalte, 1985). Much of the sediment in the trough was derived from the north with the turbidity currents flowing down the north side of the basin and then turning westward parallel to the basin's axis. Intercalated with the flysch are pillow basalts. Complicating the sedimentary regime in the basin was the initiation of diapiric activity in late Aptian. Movement of the Keuper evaporites produced highs on the seafloor on top of which late Aptian sediments are thinner and were deposited under higher energy conditions (Hines, 1985). By late Albian the influx of this clastic unit buried the Urgonian carbonates.

Along the west side of the Iberian Peninsula the early Cretaceous drift unit is developed in shallow water shelly sands and Urgonian limestones in the area of Lisbon and Wealden facies farther north. In the trough between the shelf and Galicia Bank are late early Cretaceous black shales, and the bank itself continued to be a site of shallow and deep water carbonate deposition. On the eastern North American margin the Aptian-Albian drift sequence is made up of a transgressive unit of siliciclastics whose emplacement was influenced by the Aptian low sea level stand.

Along the internal and external zones on both sides of the Pyrenees central axis the early Cretaceous is developed in shallow water carbonates including massive Urgonian carbonates. These Urgonian deposits are locally bright red, such as the red limestone at Arteaga (Ager, 1980). As Iberia began to separate from the rest of Europe the North Pyrenean fault zone was reactivated, the Axial Zone was uplifted, and early Aptian coarse clastics were deposited on the eroded Jurassic surface (Ager, 1980). At the same time a series of narrow grabens separated by narrow horsts (sites of carbonate deposition) was formed along the southern margin (plane of continental separation) of the Aquitanian basin. The east-west trending Parentis basin formed by listric faulting was a site of fluvial deposition during the Valanginian-Hauterivian on the eastern part, and marine deposition with platform carbonates around the basin's western rim during Barremian-late Albian. Most of the Aquitanian basin was dry land by the end of the Jurassic. Wealden continental deposits occurred south of Arcachon, but farther south near the Pyrenees were Neocomian marine deposits with interbeds of anhydrite. With continued separation a marine gulf spread from the southeast and enlarged during the Aptian-Albian when carbonates, including Urgonian reef facies, and muds were deposited.

The latest Jurassic-earliest Cretaceous rifting event, the late Cimmerian orogeny, also affected the entire North-West European-Arctic rift system; it is expressed by a regression. It was followed by a transgression

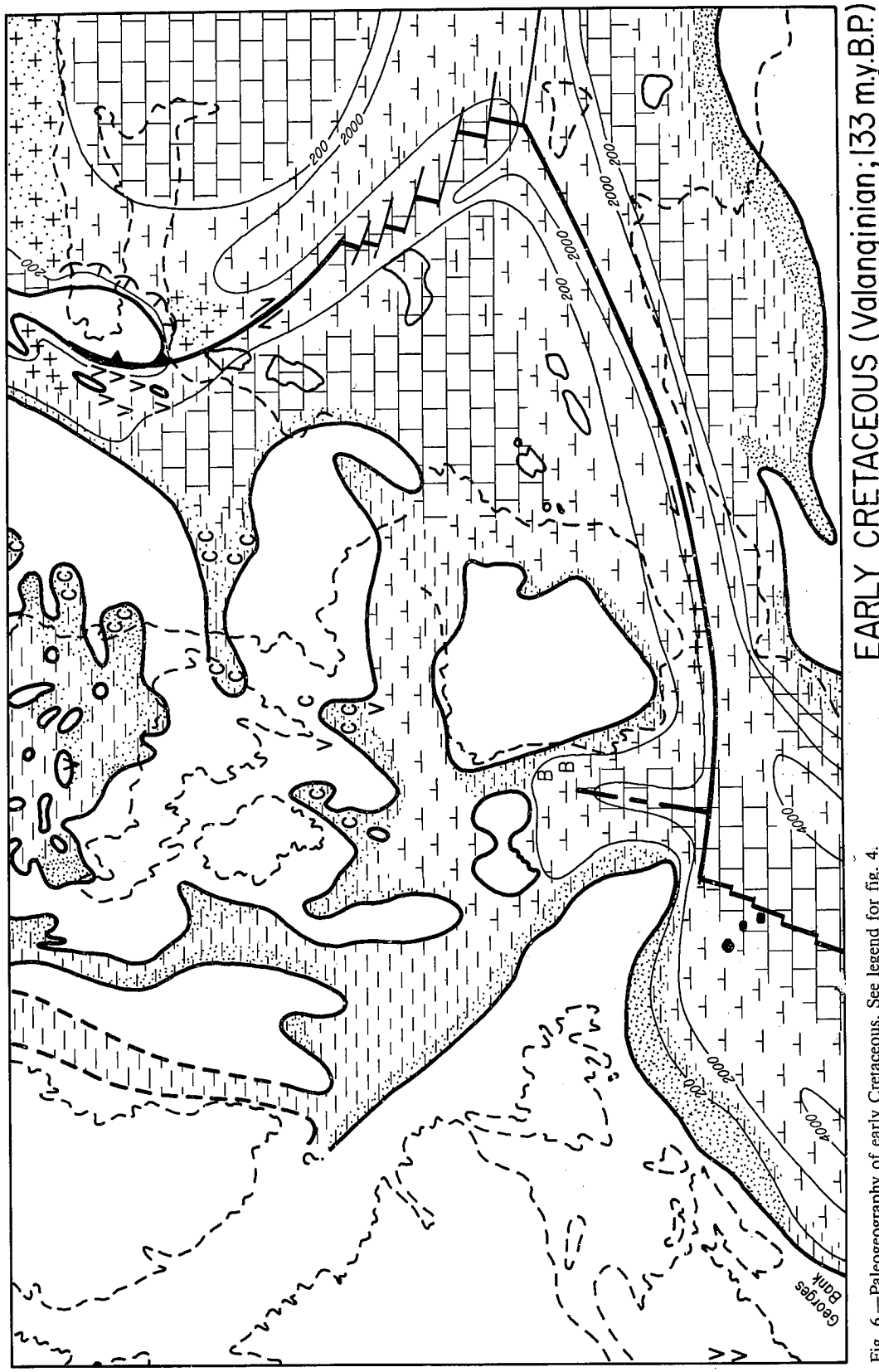

Fig. 6.—Paleogeography of early Cretaceous. See legend for fig. 4.
Fig. 6.—Paleogeografía del Cretácico inferior. Leyenda como en la fig. 4.

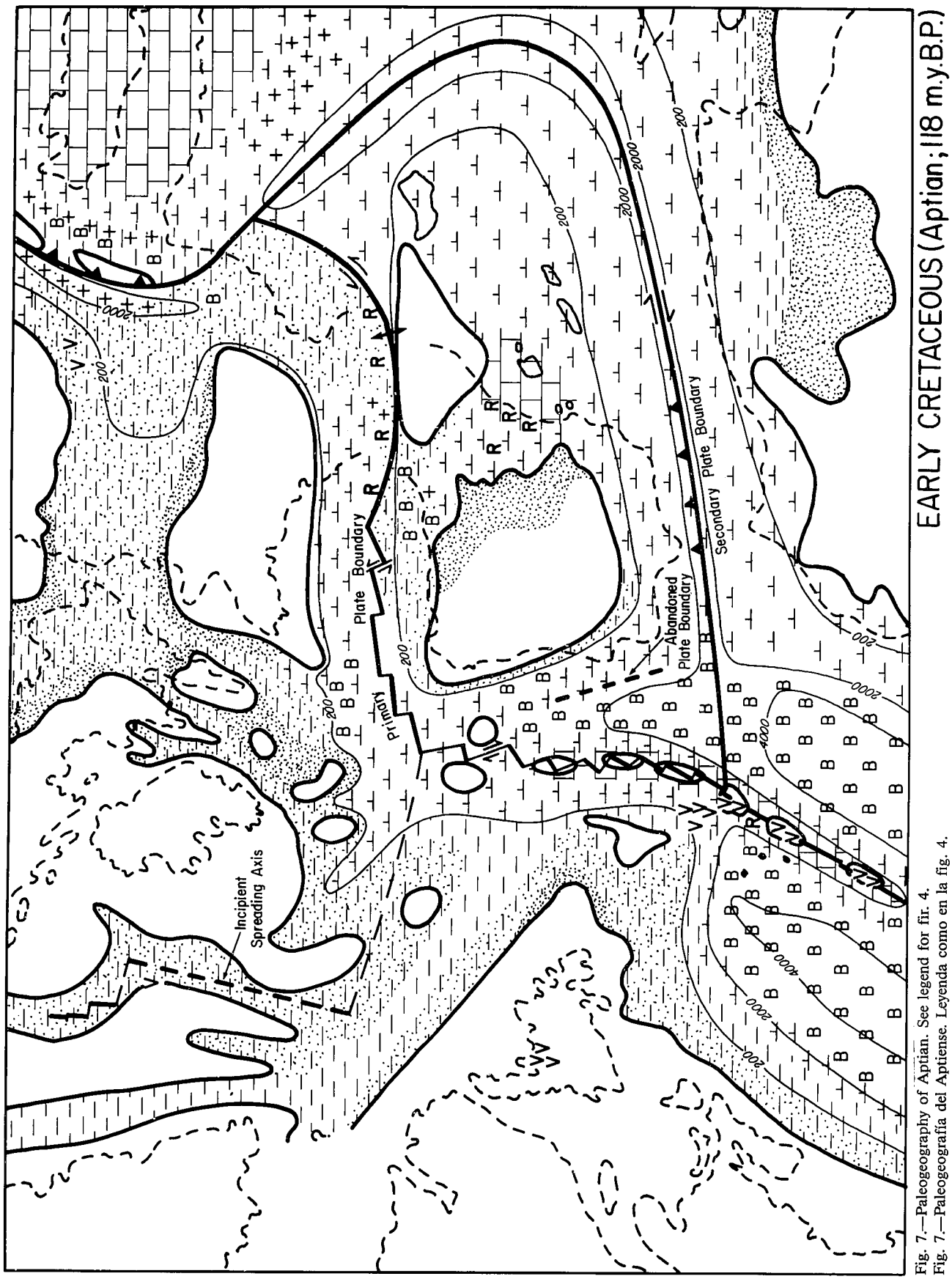

Fig. 7.—Paleogeography of Aptian. See legend for fir. 4.
Fig. 7.—Paleogeografía del Aptiense. Leyenda como en la fig. 4.

during the Valanginian to early Aptian, and a middle Aptian regression (Austrian tectonic pulse) during the onset of seafloor spreading in the Bay of Biscay (Ziegler, 1982). It is also at this time that the Aptian gabbroic intrusives were emplaced on Rockall Plateau (Harrison *et al.,* 1975). Crustal distension and listric faulting led to rapid subsidence and deposition of pelagic clays in the lows while adjacent highs were sites of non-deposition. Only minor volcanism is found in association with this phase of crustal attenuation.

In the Paris Basin the syn-rift unit consisting of siliciclastics derived from the adjacent Armorican and Brabant highs was deposited during the Neocomian transgression. As the basin was submerged slowly by Tethys waters a carbonate facies was deposited. This depositional cycle was disrupted by an early Aptian to early Albian hiatus. During a late Albian transgression the Paris basin became connected with the English Channel region. In the Channel area is a Wealden interval of lacustrine to brackish water facies. Large areas of this region were uplifted during the rifting tectonic pulse, a pulse which also affected the Celtic basin and Western Approaches. On the continental slope the Jurassic carbonate platform was disrupted by listric faulting. Sediments deposited on the slope during this rifting phase include early Cretaceous inner to outer shelf carbonates and muds. The drift unit is composed of early Albian carbonates above a late Barremian/early Albian hiatus, and middle Albian calcareous muds.

In the Balearics are shallow water carbonates, and in Provence are muds in the Vocontian trough to the north, and well bedded and massive Urgonian carbonates to the south. This episode of shallow water deposition was followed by uplift to the northwest, karst weathering, and formation of bauxite (Ager, 1980). In Corsica orogenic movements in mid-Cretaceous led to the development of flysch and klippen of granite and Jurassic limestone. Included in this allochthonous mass are spilitic pillow lavas, gabbros, and late Jurassic and early Cretaceous radiolarites. In Sardinia are pelagic carbonates on the eastern side of the island, and shallow water carbonates on the west including a lagoonal facies and urgonian carbonates with rudists. A pre-Aptian emergence led to the formation of bauxites, and an Aptian transgression brought a rich fauna from the east (Ager, 1980). As a result of the juxtaposition of allochthonous and autochthonous terrains the Mesozoic including the early Cretaceous facies in Sicily displays a complex pattern of shallow and deep water carbonates and radiolarian oozes. Carbonaceous Barremian to Albian sediments deposited on shallow pelagic plateaus, slopes, and oceanic crust are common in the Alps (Arthur and Silva, 1982). The Apulian block continued to be a site of shallow carbonate deposition. The Early Cretaceous oceanic anoxic event is represented in the Tethys by thin organic carbon rich marls or muds of Hauterivian and Barremian through early Aptian age (Arthur and Silva, 1982). As the organic carbon rich strate occur from shallow plateaus to oceanic crust the western Tethys must have been poorly oxygenated throughout the water column. Oxygenated conditions again prevailed in the late Albian and the site was carpeted by carbonates. During a limited closure between Apulia and Eurasia ophiolites were emplaced in the Hellenides (Biju-Duval *et al.,* 1976). This event was followed by extensive deposition of flysch in the Hellenides, Dinarides, and Alps.

In the oceanic North Atlantic the CCD was at a depth of 4.5 km during the pre-Barremian (Tucholke and McCoy, 1986) and pelagic carbonates blanketed most of the seafloor. The black marly interveds found within these sediments represent brief periods of anoxic or dysoxic bottom water conditions. Along the crest of the Mid-Atlantic Ridge are pelagic carbonates, and turbidites are common along the landward margins of the oceanic basin. From the Barremian to the end of early Cretaceous the CCD rose to a depth of 2.5-3.0 km. Barriers created by compressional tectonics barred bottom-water exchange with adjacent marginal sea creating anoxic and dysoxic bottom water conditions below a water depth of 2.5-3.0 km. This section of the oceanic basin was a sink for black and green muds deposited below the ambient CCD (Tucholke and McCoy, 1986). The Mid-Atlantic Ridge and lower continental margins rising above this general level continued to be sites of pelagic carbonate deposition.

7. LATE CRETACEOUS (FIGS. 8 AND 9)

Turonian facies near the western edge of the High Atlas are calcareous sediments with high organic content and chert deposited under reducing conditions (Stamm and Thiede, 1982). These deposits grade landward into calcareous micrites beyond which is an euxinic lagoonal facies with dolomitic laminites. As the High Atlas began to rise from the Coniacian to Santonian deposition shifted to the marginal basins formed during the uplift. Here are shell beds, marls, and deltaic sands. During the Campanian to Maastrichtian transgression these lows were sites of carbonate deposition with a phosphatic sand marking the worldwide regression at the Cretaceous/Tertiary boundary. Uplift of the High Atlas which lasted until the Quaternary was due to simple inversion as there is little evidence of crustal shortening (Stets and Wurster, 1982). As a result of this uplift the shelf also was elevated to form the Tafelney Plateau, and uplift of the plateau in turn initiated intensive salt diapirsm and large scale gravitational gliding seaward of the plateau (Price, 1980: Hinz *et al.,* 1982). East of the Tichka massif are Cretaceous marine and continental sediments, derived mainly from the south, which spread far south into what is today Africa (Ager, 1980). "Kasserine Island" at the eastern end of the trough was uplifted at the end of the Cretaceous.

During the late Cretaceous transgression large areas of the Moroccan Meseta were flooded and blanketed with carbonates and marls. On the Mazagan Plateau seaward of the meseta the late Cretaceous is developed in platform carbonates. The Cenomanian sea that flooded the meseta extended onto the Middle Atlas where shallow water sediments with abundant oysters were deposited. Red beds and evaporites reflecting oscillations

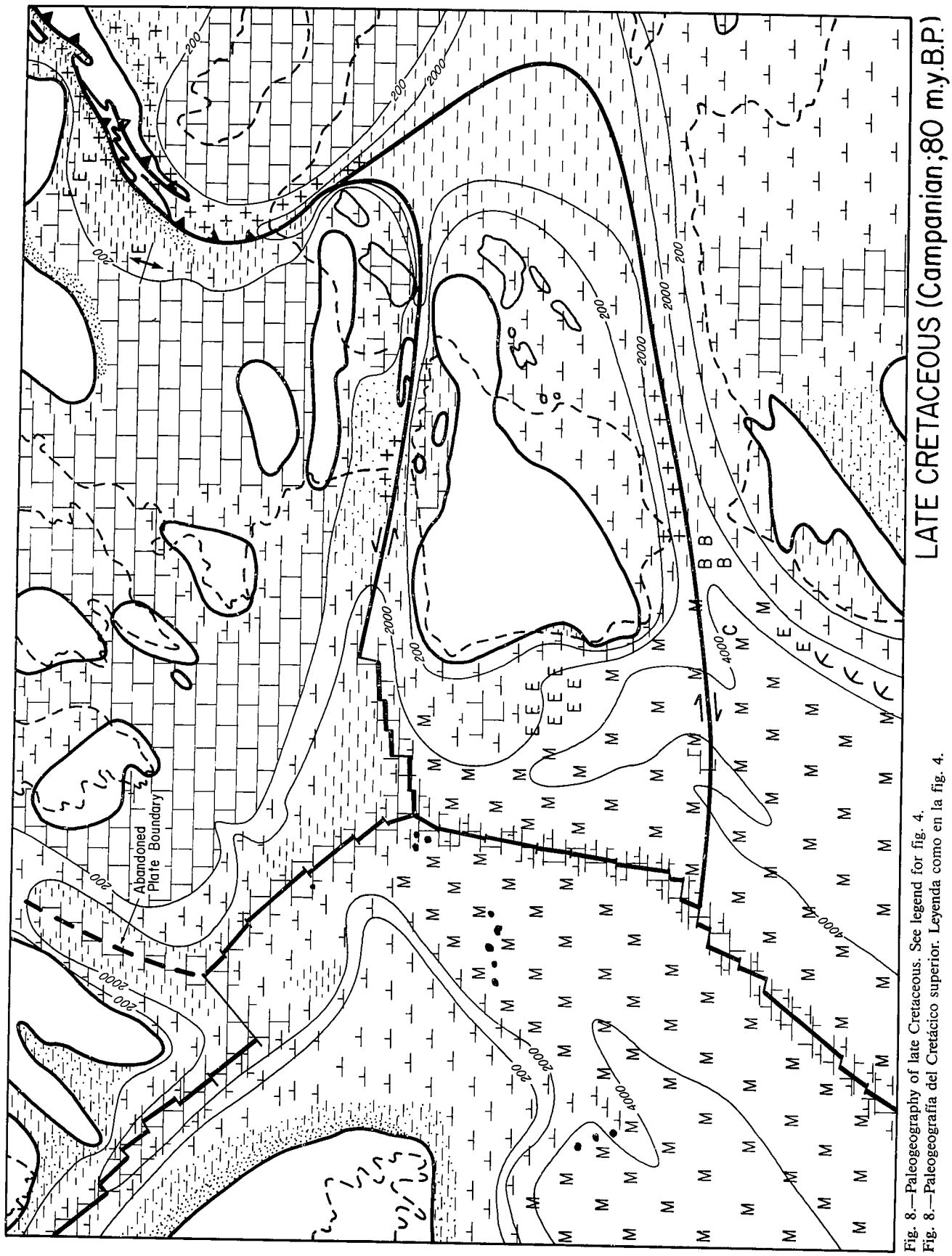

Fig. 8.—Paleogeography of late Cretaceous. See legend for fig. 4.
Fig. 8.—Paleogeografía del Cretácico superior. Leyenda como en la fig. 4.

in relative sea level aso occur as well as conglomerates marking the end of the Cretaceous (Ager, 1980). They are separated from similar deposits to the east by a belt of continental sediments. The sea also onlapped the southern edge of the Oran Plateau where shallow water carbonates were deposited and later covered by red sands and conglomerates. Carbonate flysch sedimentation increased within the North African Basin separating north Africa and Iberia. By latest Cretaceous the low begins to close as the motion between Africa became convergent along a north-south direction. In the Sub-Betics are late Cretaceous salmon colored marls similar to the "couches rouges" of the Alps (Alvarado, 1980).

Along the western edge of the Iberian Peninsula late Cretaceous carbonates interfinger eastward with fluvial sediments. In post Cenomanian time the southern part of the Lusitania Basin was uplifted and covered by volcaniclastic sediments (Groupe Galice, 1979). The northern part of the basin, however, continues to subside undergoing alternating transgressive-regressive phases during late Cretaceous. Mougenot (1981) ascribes the emplacement of the 76 and 72 m.y. old (Campanian-Maastrichtian) Sintra (at Libson), and Sines and Monchique (south of Lisbon) intrusives along faults to a tensional phase caused by the approachment of Africa and Iberia. The deformation that followed toward the end of the Campanian also resulted from the initial collision of Iberia and Africa as did the deformation in late Eocene. The crest of Galicia Bank also was uplifted at the end of the Cretaceous and the Maastrichtian carbonates were replaced by middle Eocene neritic carbonates. The Campanian/Maastrichtian and late Eocene uplift of Galicia Bank is due to the convergence of Iberia and Europe.

Black mud deposition in the narrow graben east of Galicia Bank ended in mid-Cenomanian with the transition to oxygenated conditions marked by a barren red clay. A hiatus extending from mid-Cenomanian to early Santonian or Campanian may reflect the onset of bottom circulation when the Rockall Plateau separated from Europe (Arthur, 1979). This separation allowed colder oxygenated waters from the Faeroe-Shetland shelf to flow southward by way of the trough between Galicia Bank and the shelf. On Galicia Bank are pelagic carbonates and shallow water bioclastics on the crest of the high.

The Pre-Betics was a site of shallow water carbonate deposition, and the Cenomanian in the Celtiberics is strongly transgressive (Ager, 1980). Along the northeast coast of Iberia are flysch sediments with the massive earliest late Cretaceous carbonates suggesting a slight shallowing of the basin at that time. Except for some easterly transport due to local elevation of the seafloor by the flow of Keuper evaporites (Hanish, 1978) the transport was westward. In Santander is a regressive Cenomanian clastic unit, a Cenomanian-Turonian transgressive carbonate facies, and a Coniacian to Maastrichtian deep water outer platform unit (Hines, 1985). Scarcity of late Cretaceous sediments on Le Danois Bank may be related to the hiatus noted in the trough east of Galicia Bank and on the Armorican continental slope.

During the Senonian a system of deep sea channels was eroded out of the Jurassic and Cretaceous strata parallel to the Pyrenean Axial Zone. These gullies flanked on the northern and southern flanks by carbonate platforms served as passageways for turbidity currents to the Bay of Biscay (Henry *et al.*, 1971). As Iberia and Europe began to converge in middle Senonian and their margins collided in the east in latest Cretaceous the continental realm migrated westward and the gully system was reduced to a narrow channel from Saint Gaudens to the Bay of Biscay where Maastrichtian bathyal marls accumulated.

The Parentis Basin was gradually filled during the Cenomanian. Vertical propagation of basement faults and/or salt tectonics has deformed the basin's sediment fill. The first halokinetic tectonic phase in the Landes high south of the Parentis Basin ended in the Cenomanian and was followed by a period of stability during the rest of late Cretaceous. Late Cretaceous sediments are quite extensive in the Aquitanian Basin being represented by a Cenomanian and Turonian transgressive facies including rudistid biostromes, and Coniacian to Maastrichtian sandy carbonates with broken shell debris (Ager, 1980). Sedimentation on the distal end of the basin (present shelf) was disrupted during late Cretaceous, possibly as a result of a tectonic event (Groupe CYMOR, 1981).

The Paris Basin was a site of shallow-water carbonate deposition, and in the North Sea carbonates with chert infilled the grabens and onlap the horsts. On the Shetland Platform marls and carbonates were deposited in sediment wedges filling the adjacent early Cretaceos rifts. In the West Shetland-Faeroe rift there is evidence of volcanic activity which will become pronunced during the Paleocene (Ziegler, 1982). The Celtic Sea and Western Approaches were masked with chalks, and the Porcupine Trough was a site of chalks and marls, sediments whose accumulation is disrupted by an intra-Turonian hiatus. On the outer shelf are pelagic carbonates, and on the slope a 30 m.y. hiatus separates the Campanian-Maastrichtian marl, carbonate, phosphatic interval from the Albian sediments. This hiatus appears to be oceanic in scale and is related to changes in plate geometry. During the Subhercynian orogeny from late Santonian to early Campanian the sediment fill in the grabens of northern Europe was folded, uplifted, and subjeted to erosion. During late Senonian to Danian the eroded Lower Chalk Series was submerged and the upper Chalk Series was deposited over them (Ziegler, 1982).

On the Balearics are pelagic carbonates (Alvarado, 1980), and in Provence the late Cretaceous was developed in sandy glauconites and carbonates with oysters and rudistids and muds. Toward the end of the Cretaceous (beginning of the Pyrenean orogeny) the sea withdrew from the region. The final Cretaceous unit is represente by land plant debris, dinosaur bones, and even dinosaur eggs (Ager, 1980). Uplift (beginning of the Pyrenean Orogeny) also produced torrential deposits.

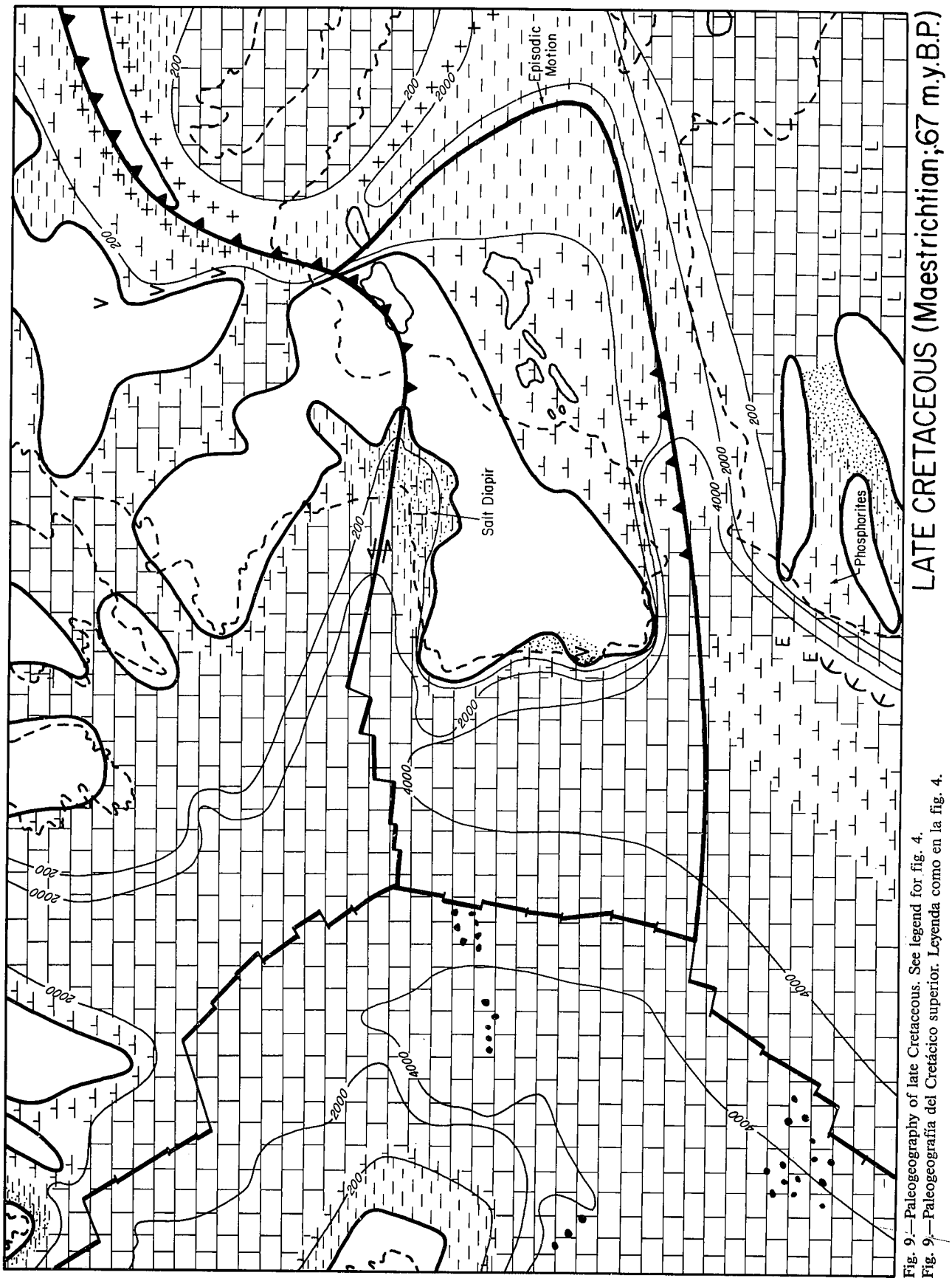

Fig. 9.—Paleogeography of late Cretaceous. See legend for fig. 4.
Fig. 9.—Paleogeografía del Cretácico superior. Leyenda como en la fig. 4.

Corsica was a site of sandy carbonate deposition, and in Sardinia the Cenomanian trangression was characterizated by Turonian peri-reef carbonates. In Sicily the late Cretaceous consisted of pelagic and shallow-water carbonates, and the Apulia block continued to be a site of shallow-water carbonate deposition. In early to middle Cenomanian the western Tethys was a setting of light-colored-pelagic carbonates verging on shades of pink and red (Arthur and Silva, 1982). Major upwelling in the late Cenomanian-Turonian resulted in the deposition of organic rich radiolarian bed in the Umbrian region. This event lasted for 0.25 m.y. and by late Turonian bottom waters became well oxygenated and red colored carbonates began to accumulate throughout much of the region. Oceanic crust to the north of the Apulian block had disappeared, but to the southwest the sea was opened broadly and connected with the Atlantic by way of the narrow troughs south of Iberia. The onset of these new paleogeographic settings in the western Tethys in late Cretaceous led to continental collision at the beginning of the Cenozoic. The earliest compressional events in the most internal domains of the Alps (Ligurian-Austro-Alpine) took place at this time (Lemoine *et al.,* 1986).

In late Cretaceous a ridge-ridge-ridge triple junction was formed in the North Atlantic and Greenland began to separate from North America along the northeast arm of the junction. With continued separation of Greenland and North America, western North Atlantic surface waters were able to flow as far north as Baffin Island (Gradstein and Srivastava, 1980). In latest Cenomanian to early Turonian as deep water connections were established between the North and South Atlantic and the North Atlantic and Tethys, black mud deposition ended and was preplaced by carbonate-free multicolored muds. The CCD was still shallow, about 2.5. km, and only the topographic highs above this level were mantled with pelagic carbonates. During the transition from anoxic to oxygenated conditions between late Cenomanian and Paleocene the outer edges of the northwest African margin were eroded by bottom currents. Lancelot and Winterer (1980) have proposed that the bottom water probably originated in the Southern Hemisphere. Toward the end of the late Cretaceous as the CCD deepened for a short time to a depth of more than 5 km pelagic carbonates and marls were deposited throughout the oceanic basin (Tucholke and McCoy, 1986).

Along the edges of the Labrador Sea and on the South Labrador margin are late Cretaceous clastics. On Orphan Knoll, a high along the eastern edge of Newfoundland basin, are Cenomanian glauconitic carbonates capped by phosphatic hardground and manganese oxide. Above this hiatus are Maastrichtian pelagic carbonates. On the Newfoundland shelf a late Cretaceous clastic unit divided in two by a carbonate layer represent a rapid transgression followed by a deepening (carbonate layer) and ending with a gradual regression (Jansa and Wade, 1975).

8. CENOZOIC

8.1. Paleocene and Eocene (figs. 10 and 11).

In the marginal trough between the High Atlas and the Moroccan Meseta are Paleocene and Eocene pelagic marls, phosphatic sands grading upward into dolomites, red marls and sands. A major regression associated with the late Eocene orogeny (Pyrenean phase) affected the basin west of the High Atlas. Above the hiatus are Oligocene conglomerates and sands. On the Moroccan Meseta are Paleocene and early Eocene carbonates rich in phosphate. The Oran Plateau and Middle Atlas were non-depositional areas at this time. Tertiary tectonics in the Middle Atlas are on a gentle scale although marked here and there by thrurst faulting and even nappes (Ager, 1980).

Continued closure of the North African basin Between Iberia and Africa led to the injection of flysch in the Eocene and subduction and obduction of oceanic crust (ophiolites in Ultra-Betics). Topographic highs that were to give form to the Betic/Rif internal zones, began to consolidate and the carbonate dorsal in the Rif was thrust southward. That uplift was taking place in the Gibraltar region is indicated by the sediment record at Deep Sea Drilling Project side 136 south of Horseshoe Abyssal Plain. Here late Cretaceous to Eocene abyssal plain turbidites were upliffted in early Eocene. From then to late Oligocene the site was non-depositional. Since then 235 m of pelagic carbonates have accumulated on the site. This partial closure led to limited subduction in the Gulf of Cadiz, rejuvenation of basement highs, and thrust faulting (Malod, la preparation). In the Pre-Betics the Paleocene-Eocene section consists of shallow-water carbonates. Along the western edge of the Iberian Peninsula are Eocene volcaniclastics and clastics on land, and early Eocene shallow water carbonates and late Eocene terrigenous limestones on the shelf. The terrigenous component in late Eocene probably reflects convergence of Africa and Iberia. On the graben east of Galicia Bank are Paleocene pelagic carbonates, and late Paleocene to Eocene calcareous turbidites. The crest of Galicia Bank is blanketed by shallow water Paleocene carbonates, but the younger sediments are predominantly pelagic carbonates.

Deformation of the Celtiberic Chain and Catalonian Cordillera occurred during mid to late Eocene. Along the southern rim of the Vizcaya basin are Paleocene-Eocene fluvial clastics. Farther north topographic highs are Paleocene calcareous turbidites, and flat bedded carbonates and muds. The Eocene in developed in a massive deep-water fan with the sediment source to the north (Kruit *et al.,* 1972). On the western end of the basin are Paleocene and lower to middle Eocene shallow water carbonates. Above are middle and late Eocene and Oligocenece deep-water carbonates. As a result of continental collision and subduction the north coast of Iberia underwent extensive deformation during the Pyrenean orogeny which in places lasted to post-Oligocene time. This compressional phase also af-

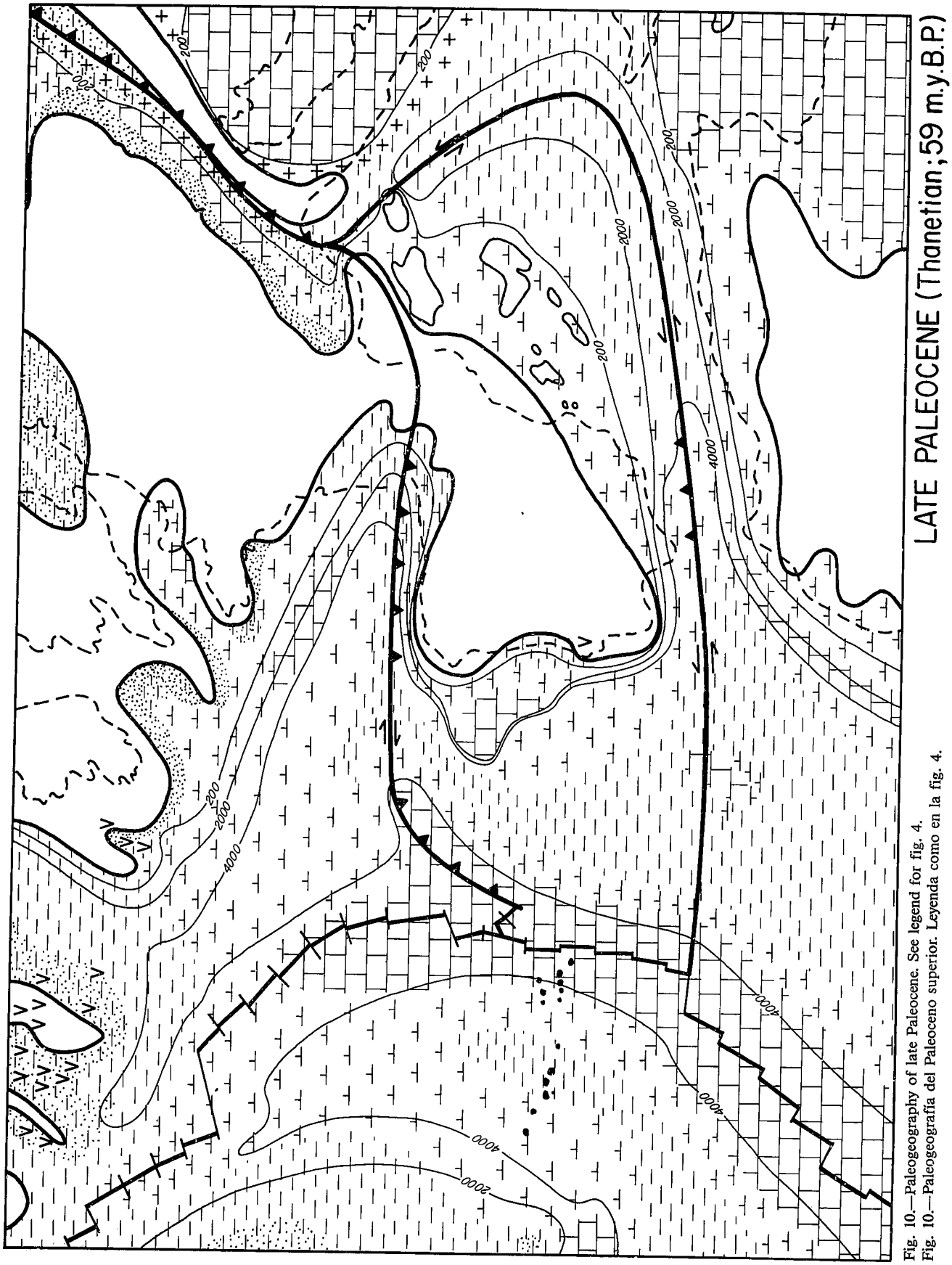

Fig. 10.—Paleogeography of late Paleocene. See legend for fig. 4.
Fig. 10.—Paleogeografía del Paleoceno superior. Leyenda como en la fig. 4.

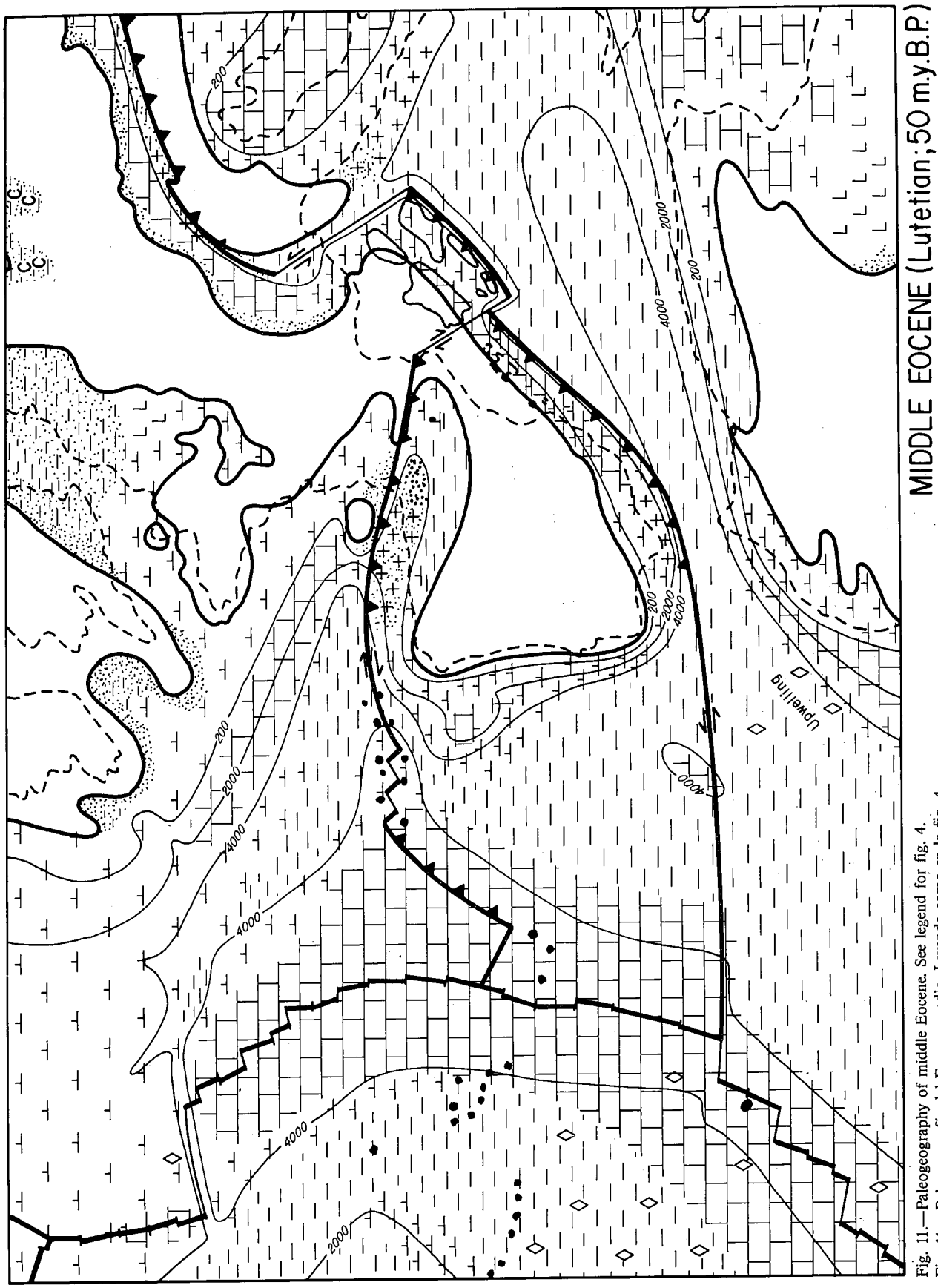

Fig. 11.—Paleogeography of middle Eocene. See legend for fig. 4.
Fig. 11.—Paleogeografía del Eoceno medio. Leyenda como en la fig. 4.

fected the margins off the English Channel and resulted in the uplift of the Cantabrian Mountains, Galicia Bank, and the Bay of Biscay seamounts exposing late Cretaceous-Paleocene turbidites, Paleocene red clays, and Paleocene middle Eocene calcareous turbidites.

At the end of the Senonian channel system in the Pyrenees are Danian limestones and flysch. As the collision between Iberia and Europe progressed during early and middle Eocene the channel system was reduced to Cap Breton Canyon. At the western and of the Pyrenean mountain chain are Eocene conglomeratic carbonates and Eocene carbonates and on the eastern end shallow water Eocene deposits. Following the collision flysch was replaced by molasse. The Parentis Basin to the north was being filled, slowly, and on the shelf west of the basin Eocene-Oligocene sediments were deposited on a Cretaceous erosional surface. The Landes high south of the basin underwent another phase of halokinesis during subsidence of the high. On the Landes Plateau west of the high is a Cenozoic prograding wedge atop a thin late Cretaceous unit. The lower part of the Cenozoic unit passes beneath Cap Breton Canyon, but the upper part is exposed along the northern flank of the canyon. The whole sequence south of the canyon is involved in the Pyrenean orogenic event. In the Aquitaine Basin are Danian micrites and Eocene sandy carbonates which extend up to earliest Oligocene. In the Paris Basin the Paleocene-Eocene is developed in shallow-water marine and lacustrine carbonates and siliciclastics whose deposition was controlled by cyclic transgressions and regression (Ager, 1980).

In late Paleocene (Laramide orogeny) large areas of north Europe and the North Sea were elevated and underwent erosion. Ziegler (1982) believes that both the Subhercynian and Laramide orogenies are due to closure of oceanic basins in the Alps and Carpathians. Accompanying this uplift was the Thulean volcanism in Scotland, Ireland, Rockall-Plateau, Greenland, and Baffin Island. This volcanism produced a land bridge from Great Britain through Greenland and across Davis Strait to Canada (McKenna, 1983). Uplift of the Yermak triple junction off northern Greenland formed another land bridge between Scandinavia and North America (Tucholke and McCoy, 1986). Volcanism ended in early Eocene as seafloor spreading began in the Norwegian-Greenland seas. Away from the region of thermal uplift the Paleocene-interval consists of pelagic muds, and on the outer shelf and slope are pelagic carbonates. Sediments on the slope are disrupted by hiatuses related to the Subhercynian and Laramide orogenies.

On the Rockall Plateau northwest of Great Britian the Paleocene consists of coarse siliclastics, volcanics and intrusives, with marginal troughs being blanketed by muds. The seamounts at the northern end of the Rockall trough east of the high were probably emplaced during the Thulean magmatic event. The early Eocene consists of muds and tuffs, and the middle Eocene is dominated by carbonates and siliceous deposits.

The hiatus separating the late Cretaceous from the Oligocene in Mallorca may reflect the Pyrenean movement. In Menorca, however, marine deposition appears to have been continous to early Miocene. At the end of the Cretaceous an extensive lake existed in Provence. This lake which lasted into the Eocene was filles with siliciclastics. Elsewhere are nummulitic carbonates comparable to those in the Pyrenees. Orogenesis in the region began in the Eocene. These east-west folds are obliterated to the west by the northwesterly trending Western Alps.

Alpine tectonism reached its maximun in Corsica at the end of the Eocene. In Sardinia are early Tertiary continental deposits, and in mid Eocene the northwest trending Campidana rift was formed. This low together with the Rhone-Saono trough, Rhine graben, Lower Rhine depression, the North Sea Central graben and the Viking graben, developed during the Alpine oro geny in response to a plate reorganization. In the Oligocene the rift system expanded into the Gulf of Lion and Balearic islands. Its southern limit was to envolve into a seaflor spreading phase in early Miocene producing the Balearic Basin. In late Eocene the Brese and Rhine grabens began to subside and by latest Eocene to earliest Oligocene a narrow strait extended from the Alpine deep to the southern end of the Rhine graben. The first peak of volcanism in the Rhine Graben was in late Eocene. In Sicily is a complex mixture of deep water pelagic and shallow water carbonates including radiolarian oozes. It was during the Eocene that the narrow trough between Iberia and north Africa was being consumed, Iberia was colliding with Europe, subduction was taking place in Corsica, and the main tectonogenesis was taking place in the Alps.

During Paleocene the CCD was 3.5 km to 4 km deep and the Atlantic basin was again blanketd by pelagic clays in its deeper parts with wast expanses of pellagic carbonates accumulating along the margins and the Mid-Atlantic Ridge. A triple junction was created south of Greenland with Greenland and Rockall Plateau separating along the northwest trending arm in early Eocene. By mid Eocene the Thulean land bridges between Great Britain and Greenland and in the Davis Strait had been severed, but volcanism continued in east Greenland and portions of the Greenland-Scotland Ridge. Although the bridge was breached it still formed an effective barrier isolating the Norwegian-Greenland seas from the regions to the south. The Artois bridge also subsided and circulation in the English Channel was renewed (Tucholke and McCoy, 1986). The CCD continued to be deep remaining at a depth of 3.5 to 4 km and extensive areas of the oceanic basin were carpeted with pelagic clays. As a result of high productivity in the surface waters associated with the circum-equatorial flow across the central Atlantic and upwelling zones off eastern North American and northwest Africa a large component of biogenic silica was deposited with these pelagic muds (Tucholke and McCoy, 1986). Additional silica was introduced into the basin by turbidity currents from the North American margin.

8.2. Oligocene, Miocene, and Pliocene (fig. 12, 13, 14 and 15).

In the trough between the western High Atlas and the Moroccan Meseta are late Oligocene conglomerates whose accumulation was followed by the late Oligocene-Miocene Alpine disturbance that produced the main folding in the Atlas (Dillon and Sougy, 1974). Although the main phase of this orogeny ended in latest Pliocene, raisded beaches, volcanism, and seismicity indicates that uplift still was taking place. Above the Oligocene are early Pliocene mars and sands capped by a hiatus which is carpeted by late Pliocene continental deposits. On the trough south of the western High Atlas are estuarine muds and shell sediments. During the Pliocene uplift of the High Atlas this low (Souss trough) was filled with siliciclastics. The Middle Atlas underwent gentle deformation and was a positive area lacking Neogene sediments. In the extreme north end of the Moroccan Meseta are some clays and sands with lignite and along the coast are scattered Plio-Pleistocene sediments including some raised beaches (Ager, 1980). There is limited Pliocene volcanism, but in the Quaternary there were extensive basaltic extrusions along the margins of the Middle Atlas. In the Oran Plateau are continental Miocene red sands and gravel, multicolored marls, and volcanics. Vast areas of Morocco and Algeria are blanket by Plio-Quaternary continental sands, lacustrine deposits, and ventifacts.

Although deformation along the north coast of Iberia continued until post-Oligocene time, by the beginning of the Oligocene the primary plate boundary had shifted southward to the Betic/Rif Zone and Iberia had been coupled to the Eurasian plate. Continued convergence of Iberia and Africa resulted in the closure of the North African Basin and the collision of Iberia and Africa. The Gibraltar klippe, the nappes in the Sub-Betics and the olistostromes in the Guadalquivir Depression and in the western approaches to the Straits of Gibraltar are the product of this orogenic event. Emplacement of the olistostrome consisting of a jumble of Mesozoic and Tertiary rocks in the Guadalquivir Depression must have taken place in mid Miocene as early Miocene sediments are involved in the slumping whereas the late Miocene is generally undisturbed (Ager, 1980). On the western approaches to the Straits of Gibraltar slumping must have taken place later as early Messinian sediments are involved in the nappes (Auzende *et al.,* 1981). In the Guadalquivir Depression the Neogene shore moved progressively northward with time until the Tortonian transgression when the shore retreatted to its present position. Miocene sediments in the low are a mixture of white siliceous marls and hard bands of carbonate and sand.

Deformation of the Pre-Betics and Balearics took place in the Miocene. Uplift of Gorringe Bank west of the Straits of Gibraltar occurred between the deposition of early Miocene pelagic carbonates and early Pliocene pelagic carbonates containing neritic bryozoans (Ryan, Hsü *et al.,* 1973). The neritic fauna probably was derived from southern edge of the bank which at that time must have been near sea level. The northward subduction of the African Plate also led to the deformation of the continental slope in the Gulf of Cadiz, and the extrusion of the 18.5 to 4 m.y. old calc-alkaline volcanics in southwest Spain. Convergence of Eurasia and Africa in the Paleogene produced the Carpathian and Alpine nappes. The foredeeps in front of the nappes were filled with deep water flysch, and the shelves north of the deeps were carpeted with thin shallow-water sequences (Ziegler, 1982). Downflexing of the shelves in early Oligocene caused the drowning of the carbonate platforms and the northward overstepping of the Molasse basin. As the Alpine nappes migrated northward the clastics deposited in front of the nappes became more and more involved in the deformation. Throughout the Oligocene and Miocene shallow marine conditions prevailed in the Swiss and Bavarian section of the Alpine foredeep, but deep water conditions presisted in the Austrian section of the Alpine foredeep and in the Carpathian foredeep. The Austrian part of the Alp ine foredeep became inactive at the onset of the Miocene and it became filled with upward shallowing water clastics. In the Carpathian foredeep northward tectonic transport continued until late Miocene.

As described by Ziegler (1982) Oligocene and Miocene orogenesis in the central and western Alps was accompanied by extensive crustal shortening. During latest Miocene and early Pliocene the southern parts of the Western Molasse basin were imbricated by thrust faults and partly overriden by the Helvetic nappes. It is a this time that the Jura Mountains were folded and the Mesozoic grabens of the Rhone Valley inverted. Volcanism in the Massif Central took place during the Oligocene, and in the Eger graben in late Oligocene. By mid-Oligocene the Rhine graben linked the Tethys with the North Sea, a link that was severed by early Miocene (Ziegler, 1980). The Neogene tectonic history of the west European rift system is one of volcanic activity, upflift of adjacent highs, subsidence of the lows, and strike slip faulting (see Ziegler, 1982).

Behind and within the Betic/Rif orogenic belt are several Neogene extension basins. The rift phase following domal uplift and subsequent erosion was succeeded by an Aquitanian transgression. Evidence of this Oligocene rifting phase can be found in the gulfs of Genoa, Lion, and Valencia. In the Gulf of Lion the graben system, a southern extension of the west European late Eocene rift, is filled with continental beds and/or evaporites, and the marginal faults show synsedimentary activity (Rehault *et al.,* 1985). In the Alboran Basin east of the Straits of Gibraltar are early Miocene evaporites or muds whose plastic flow has deformed the overlying sediments (Mulder and Parry, 1977). Opening of the Straits of Gibraltar in early Messinian led to strong erosion by bottom currents and the formation of a deep channel east of the straits. In they Sardinia rift late Oligocene volcanics are interbedded with the oldest sediments, Oligocene rift sediments in Sardinia and Menorca contain pebbles derived from the Pyrenees and Catalonian Cordillera, evidence for the proximity of these island and France before the opening of the

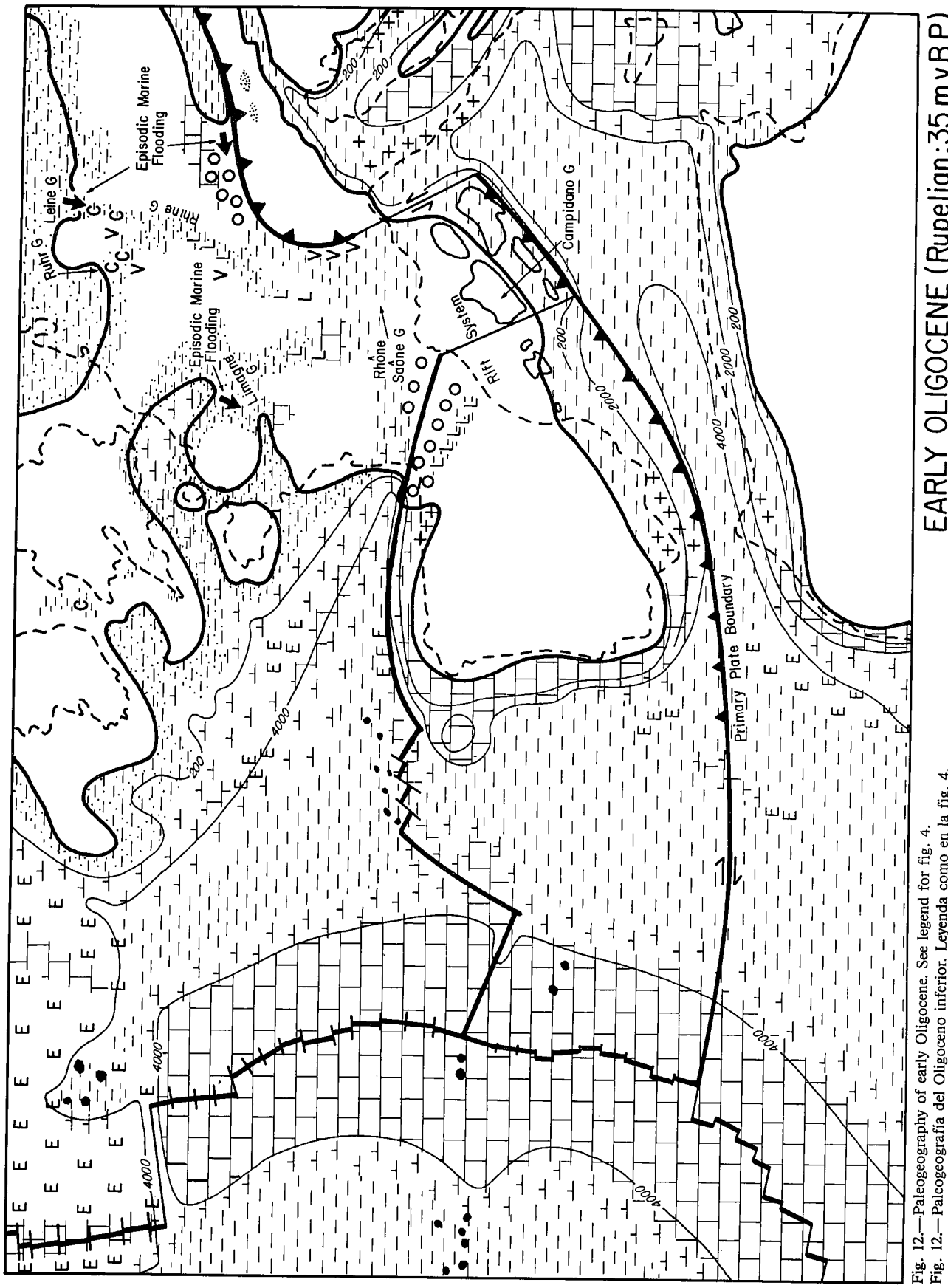

Fig. 12.—Paleogeography of early Oligocene. See legend for fig. 4.
Fig. 12.—Paleogeografía del Oligoceno inferior. Leyenda como en la fig. 4.

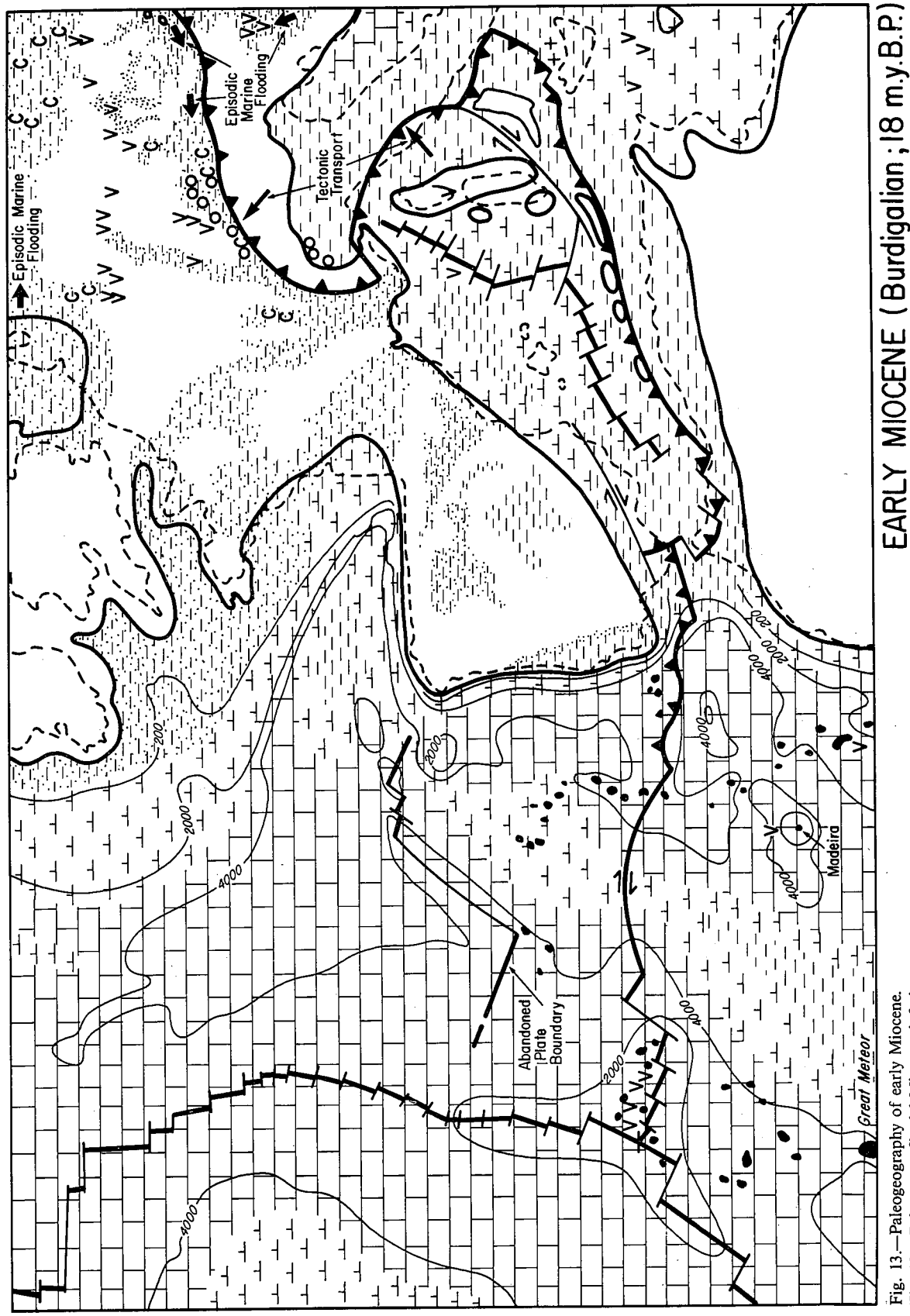

Fig. 13.—Paleogeography of early Miocene.
Fig. 13.—Paleogeografía del Mioceno inferior.

Balearic-Liguro-Provençal Basin (Rehault *et al.,* 1985 and references therein). DSDP Site 372 east of Menorca penetrated the flank of one of these northeast trending rifts filled with Oligocene clastics. The Aquitanian transgression following the rifting may be contemporaneous with seafloor spreading in the Balearic Basin which Rehault *et al.,* (1985) believe took place 21 to 18 m.y. ago.

During the opening of the Balearic-Liguro-Provençal Basin the Corsica-Sardinia block rotated counterclockwise and the Balearics drifted a small amount southeastward. The Kabyles and Calabria broke away from Iberia along the north African consuming plate boundary with the Kabyles colliding with Africa in early Miocene (Rehault et al., 1987). As the Balearic Basin slowly opened, the subduction belt migrated southward and eastward and the Apennines atop the Apulia block began to take shape. By 18 m.y. ago the rotation of the Corsica-Sardinia block ended and the site of extension shifted to the Tyrrehanian Basin. By then all the pre-existing western part of the Tethys between Africa and Europe had been subducted, the first units of the Apennines were entering the subduction zone, and calc-alkaline volcanism was taking place in Sardinia (Ryan, 1986). Continued sinking of the African-Adriatic lithosphere and eastward migration of the subduction zone led to the development of the Apennines, and extension of the Tyrrehanian basin. With migration of the subduction zone the Calabria block became attached to southern Italy and the Peloritan block to Sicily, and seafloor spreading was initiated at the southeast end of the Tyrrhenian Basin in the Plio-Quaternary. This tectonism accounts for the Plio-Quaternary volcanism in the Tyrrehanian basin, Sicily, Aeolian islands, and Italy (Selli, 1985; Wezel, 1985).

Continued motion along the primary plate boundary along the Beti/Rif system and the emplacement of nappes and olistostromes west of the Gibraltar arc led to the damming and isolation of the Mediterranean from the Atlantic during latest Miocene (Auzede *et al.,* 1981). As a result of this isolation the Mediterranean slowly dried up and the Messinian evaporites were deposited. This belt of evaporites extends from southern Spain across northwest Africa to Sicily, up the Italian Peninsula, and then eastward to Greece, Cyprus, Turkey, and Iran (Ryan, 1973). As the Messinian sea level fall, the early Messinian reefs in eastern Iberia, north Africa, southern France, Corsica-Sardinia, Sicily, and Italy died out (Cita, 1982). As a result of changes in base level the adjacent lands and margins underwent extensive erosion as exemplified by the subaerial erosion of the Stoechades and Saint Tropez canyons, entrenchment of the Var River, the subsurface channel in northern Lybia, fluvial erosion of the lows occupied by the Alpine lakes, the subsurface erosional surface in the Lombardian Plain, and the entrenchment of the Nile (Barr and Walker, 1973: Groupe ESTOCADE, 1978: Clauzon, 1978: Bini *et al.,* 1978: Finckh, 1978: Rizzini and Dondi, 1978: Rizzini *et al.,* 1978: Ryan, 1978). Erosion of this magnitude must have resulted in the emplacement of extensive siliciclastic wedges in the Mediterranean basins creating a bajada-playa complex comparable to that present in the Basin and Range province of western United States. Yet descriptions of the Medirranean Messinian landscape published to date are noticeable for their dearth of information on this unit; in some places the evaporites are even shown to be onlapping the fluvial erosional surface. A possible solution to this problem may be as follows. Initially as sea level was being lowered the adjancent highland underwent extensive erosion and the detritus was deposited in the lows. Along those marginal scarps facing away from major runoff erosion may have been negligible. During this phase the evaporites emplaced in the center of the depocenters had a considerable component of siliciclastics. With the onset of fluvial equilibrium episodic flooding of the Mediterranean by way of north Africa resulted in the deposition of the Messinian evaporites onlapping the clastic wedges. This massive extraction of salts from the world's oceans is reflected in the isotopic composition of benthic forams off Cape Bojador, north Africa (Cita and Ryan, 1979). The composition shows cyclically repetitive changes whose amplitude is one third to one half the amplitude of the Pleistocene isotopic composition at the same site. During the subsequent rise in sea level both the Messinian erosional surfaces and evaporites were blanketed by Pliocene sediments.

In the Catalonian Cordillera in eastern Iberia is a northeast trending graben filled with Oligocene, Miocene, and Pliocene sediments. Toward the southwest the Miocene is marine and to the northeast continental with a rich mammalian fauna. The Iberian massif is carpeted by continental deposits with evaporite interbeds. Present topographic features on the massif are the result of late Tertiary block-faulting and upwarping (Ager, 1980). In the Ebro basin the Oligocene is dominated by sands, marls, and gypsum. These sediments apparently were folded with the Mesozoic during a late phase of "Alpine orogenesis", a deformation aided by movement of the evaporites. The early Miocene in the Ebro Basin is made up of great fan deposits derived from the Pyrenees to the north and the late Miocene of conglomerates, sands, and marls. During the Oligocene and Neogene the north coast of Iberia underwent tension forming the rift structure on the shelf, rejuvenating the source areas and filling the inland basins. Along the west side of the Iberian Peninsula the Neogene interval is trangressive resting unformably on the older units and is found only on the outer shelf. On the continental slope off southern Portugal deposition on 10 to 140 km wide terraces ins controlled by Mediterranean overflow (Mougenot and Vanney, 1981). The early Oligocene to early Miocene in the graben east of Galicia Bank is characterized by biogenic silica and carbonate rock, sands and mud turbidites capped by an early Miocene hiatus (Alpine orogeny). Deposition of the middle Miocene to Pleistocene cyclic pelagic marls and carbonates and muds may have been influenced by climatic oscillations. On Galicia Bank to the west are pelagic carbonates..

Concurrent with and following deformation of the Pyrenees vast thickness of sediments were deposited

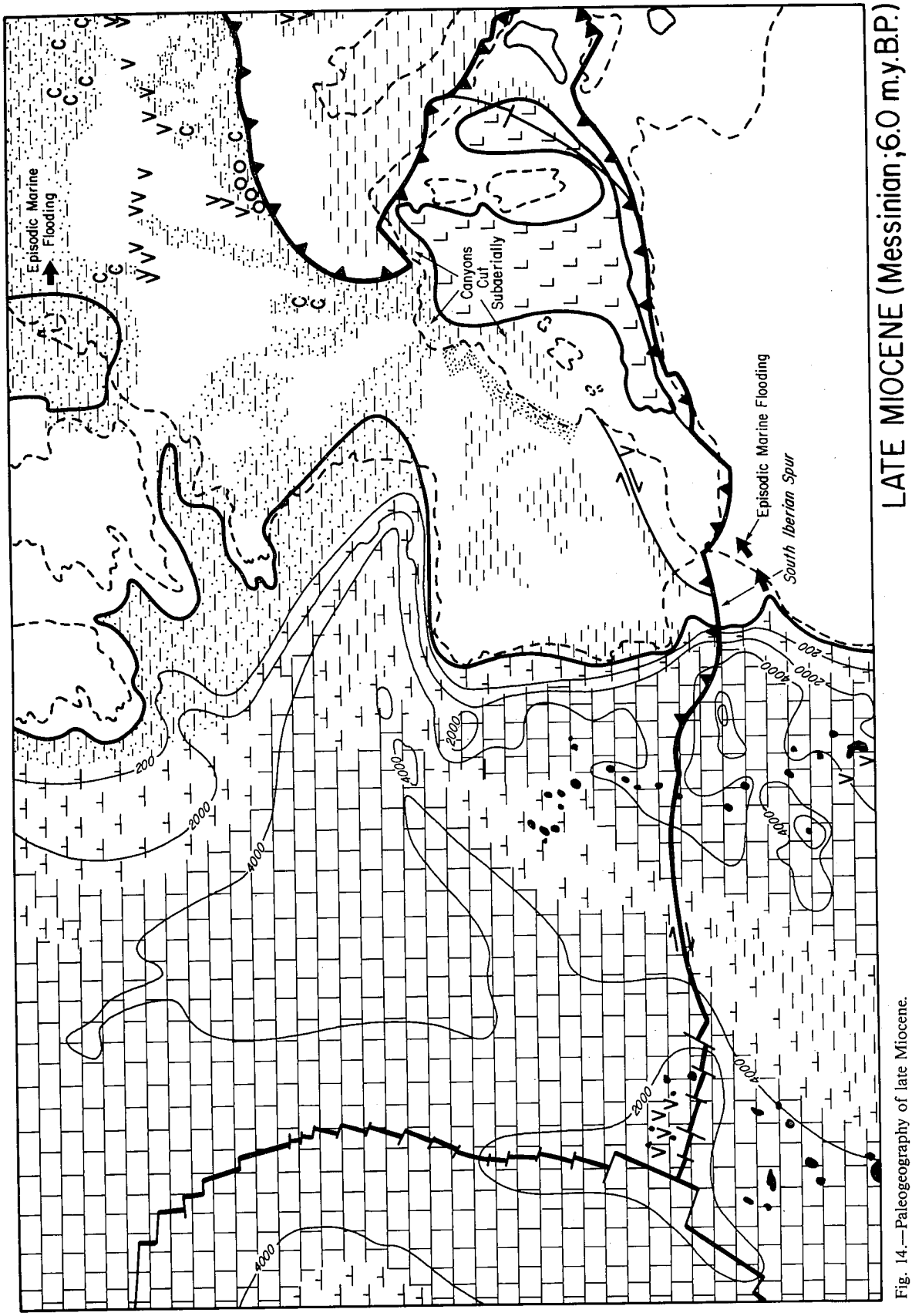

Fig. 14.—Paleogeography of late Miocene.
Fig. 14.—Paleogeografía del Mioceno superior.

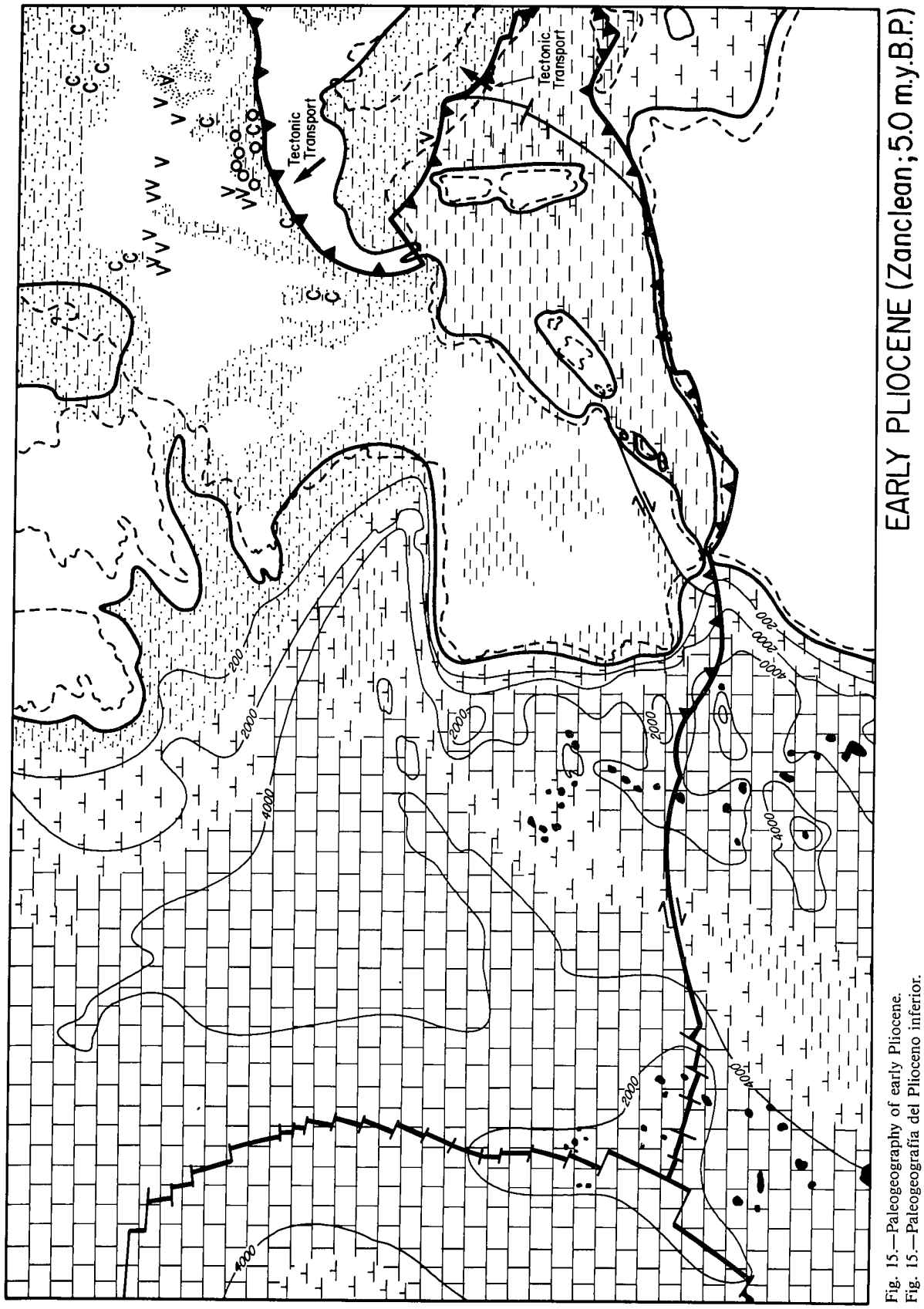

Fig. 15.—Paleogeography of early Pliocene.
Fig. 15.—Paleogeografía del Plioceno inferior.

north and south of the mountain range. Faulting in the Oligocene and Miocene renewed deposition. Associated with this vertical faulting is local late Tertiary volcanism (Ager. 1980). By late Tertiary the Parentis Basin was filled, and on the shelf west of the basin the Eocene--Oligocene wedge was deposited on a Cretaceous erosional surface. During the mid-Oligocene eustatic drop in sea level, the shelf was eroded and was followed by Miocene sediment progradational phase. During the Miocene an east-west fault cut across the shelf. Cap Ferret Canyon was eroded along this structural weakness. Oligocene sediments on the Aquitanian basin are to the north and east of the low. Its connections with the Paris Basin and Mediterranean came to an end at this time. Transgressions and regressions from the west continues into the Miocene with a marine carbonates to the west and brackish-water facies to the east. After the Tortonian transgression deposition was restricted to the present coast. Early Oligocene sediments in the Paris Basin are represented by semi-marine deposits and a late Oligocene more marine unit and fresh-water limestone (Ager, 1980). The Neogene is developed in freshwater carbonates, and continental clays and sands.

Oligocene sediments throughout much of the North Sea consist of pelagic clays. During the glacially induced low sea-level in mid-Oligocene sediments prograded eastward from the Shetland platform onto the central parts of the Viking graben. Miocene and Pliocene deposition of siliciclastic outpaced subsidence and there was a progressive shallowing of the region. On the outer shelf and slope are Oligocene and Neogene pelagic carbonates which extend onto the Bay of Biscay.

By earliest Oligocene a deep passageway developed between the Arctic and Norwegian-Greenland seas, and cool Arctic water began to flow into the North Atlantic over the Greenland-Scotland Ridge eastern end. This deep contour-following flow caused extensive erosion along the European, North American, and north African margins. The bottom water in the eastern Atlantic may have included water from southern latitudes that reached the region by way of the Equatorial fracture zones. In the Labrador Sea seaflor spreading ceased prior to earliest Oligocene and the basin segments peripheral to the spreading axis had subsided to a depth of 4 km (Tucholke and McCoy , 1986). With the CCD still at a depth between 4 and 4.5 km the seafloor seaward of the zone of erosion was blanketed by pelagic clays, but the Mid-Atlantic Ridge and the African margin were carpeted by pelagic carbonates. During the glacially induced sea-level drop in mid-Oligocene the shelf and slope underwent extensive erosion and clastic wedges were emplaced on the contiental rise.

When the primary plate boundary migrated to the Betic/Rif system in the Oligocene/Miocene, a volcanic plateau was emplaced at the Azores triple junction. Associated with the Alpine orogeny is an extensive deep-sea igneous activity that built the volcanic edifices in the Madeira-Tore Rise including the island of Madeira, constructed the Canary Islands, uplifted the seafloor to form the Cape Verde Plateau with the island being constructed by volcanic activity, and the volcanicity in Dakar, Senegal. As a result of this thermal/tectonic event the average depth of the eastern Atlantic was reduce significantly (Tucholke and McCoy, 1986). With the decrease in the intensity of bottom circulation during early Miocene large sediment drifts were formed along the western continental margin. These highs served as dams behing which turbidites slowly accumulated during the Mio-Quaternay. Uplift in the late Miocene terminated the flow from the eastern Tethys and the Pacific, and uplift of Panama in the Pliocene cut the Atlantic-Pacific connection. Containment of the Atlantic waters led to the intensification of the Gulf Stream and subsequent erosion of the Blake Plateau off southeastern United States (Emery and Uchupi, 1984 and references therein).

9. QUATERNARY

9.1. Pleistocene (fig. 16).

The Plio-Pleistocene glacial age was a time of drastic climatic changes and numerous transgressions and regressions resulting from the expansion and retreat of continental glaciers. Although there are indications of ice rafting at high latitudes as early as 5 m.y. ago, continental glaciation did not become well developed until 2.4. m.y. ago (Leg 81 Scientific Party, 1982). At the Würm maxima the west-east trending southern edge of the continental ice sheet was located at the southern end of the North Sea and north Germany. Mountain glaciers were extensive in the Alps and Pyrenees, and the higher peaks in Italy, southern France, Iberia, and north Africa had permanent ice cover. In the Pyrenees the glaciers appear to have extended farther down the colder northern side. Loess was deposited from eastern France to central Germany, in Algeria, and along the Atlantic coast of north Africa. Land vegetation was dominated by tundra.

During glacial peaks polar water moved southward to 42°N where an abrupt frontal system was formed separating ghe cyclonic Subpolar Gyre from the anticyclonic Subtropical Gyre (Ruddiman and McIntyre, 1976). As a result of ice cover productivity in the Norwegian-Greenland seas decreases and deposition of biosiliceous sediments was reduced drastically (Talwani and Udintsev, 1976). With the onset of the East Greenland Current with its low salinity Arctic Ocean outflow and the almost perenial ice cover the western segment of the Norwegian-Greenland seas became less productive than the eastern one. Reduced input of organic matter from the overlying waters also decreased the corrosiveness of bottom waters allowing the accumulation of biogenic calcareous sediments. In addition the Gulf Stream was forced southeastward by the Labrador Current leading to the development of a distinct subpolar bioprovince in the Norwegian-Greenland seas (Berggren and Schnitker, 1983). During the glacial maxima the warm North Atlantic Drift south of thermal front at 42°N flowed eastward impinging on the Iberian Peninsula. This thermal front also

marked the position of the Westerlies which during the winter must have been comparable in intensity to the "Howling Fifties" in present-day Antarctica (Sarnthein *et al.,* 1982). At its eastern end the thermal front was deflected southeastward off the Iberian Peninsula and Morocco. This caused the easterly flowing current to deflect to the southeast accounting for the ice rafted debris off southern Portugal and Morocco (Kudrass, 1973). During the glacial maxima the Canary Current waters were 8°C colder than today, and extended farther offshore and south, and the Sargasso Sea Water was displaced to the west. With the enhancement of the Trade-Winds coastal upwelling extended to the continental slope, and deep water circulation over the continental slope was intensified. As a result of the enhancement of the Trade-Winds and the strong upwelling offshore north Africa went through a phase of desertification. With the lack of plant cover, wind erosion increased and vast quantitites of eolian dust were transported to the ocean. During the summer the thermal front in the eastern Atlantic retreated to the center of the Bay of Biscay (Molina and Thiede, 1978) decreasing the intensity of the processes in north Africa.

Ruddiman (1977) has estimated that during the past 3 m.y. 200.000 km^3 of drift material was introduced into the deep North Atlantic. From 125,000 to 75,000 years ago the depocenter was located southeast of Greenland and northeast of Labrador. During Wisconsin (Würm), when icebergs reached farther south before reaching warm water (erratics have been found as far south as Great Meteor Seamount a 30°N; Pratt, 1963), the depocenter was along latitudes 46°N to 50°N. As described by Ruddiman (1977), drift input increased slightly 115,000 year ago during the next-to-last glacial, rose markedly 75,000 years during the inception of the Würm glaciation, and continued to rise toward the glacial maximum. As the glaciers expanded to their maximun runoff on the continents south of the glaciers was at a minimun ad contributed little to the marine evironment. According to Vita-Finzi (1971) very little sediment was supplied to the Mediterraneam from 20,000 to 10,000 years ago.

The high concentration fo erratics in the southern Greenland Sea northwest of the present North Polar Front indicates that summer melting of the sea ice in the Norwegian-Greenland seas began as early as 18,000 year ago (Grousset and Duplessy, 1983). This aided the southward drift of icebergs produced during the melting of the northern ice sheets. According to Ruddiman and McIntyre (1981) deglaciation form 40° to 65°N occurred in three steps: major warmings in the southeastern and central regions 13,000 years ago; warming in the central and northern sections 10,000 year ago; and a warming in the Labrador Sea 9,000 to 6,000 years ago. Within this framework there was a major cooling period from 11,000 to 10,000 years ago during the European "Younger Dryas" interval.

The Quaternary, however, is not just a time of climatic deterioration as described above, but it also is marked by volcanism in the Moroccan Meseta along the margin of the Middle Atlas, Sardinia, Sicily, and Italy. It is at this time by a combination of changes in sea level and tectonism that the terraces in Almeria and other parts of the Mediterranean coast, and along northern and western Iberia were eroded. Changes in base level also are reflected in the river terraces in the Ebro Basin, and the rias that dominate the north and nortwest coasts of Iberia. Changes in sea level also led to faunal isolation explaining, for example, the dwarf elephants on Pianosa Island 80 km east of Corsica (Ager, 1980). Uplift of the Pyrenees occurred at this time, and extensive gravels and calcretes were also deposited in the Pre-Betics.

9.2. Holocene (fig. 17).

The Holocene/Pleistocene boundary generally is placed at 11,000 years before the present. It is from 15,000 years ago when sea level was at 110 meters below its present level to 11,000 years when sea has risen to 60 meters, that extensive ablation of the northern ice sheets took place, a trend briefly interrupted by the cold Younger Dryas interval. In those areas where the margin had been depressed by the weight of the continental glaciers the sea extended well inland. By about 5,000 years ago the sea reached its present level. Isostatic rebound of formerly glaciated shelves has caused the sea to retreat locally, and in the peripheral bulge south of the glaciated area subsidence has produced local rises in relative sea level (Emery and Aubrey, 1985).

With the mass wasting of the continental ice sheets runoff increased and vast volumes of siliciclastics were transported and deposited on the outer shelf and slope. Rapid deposition led in turn to considerable instability on the slope producing slumps, debris flow, and finally turbidity currents. These currents, for example, transported their load as far as 700 km south of Iceland by way of the Maury Channel system. The currents also filled the Biscay Basin to its rim, and then were able to by-pass the Biscay Abyssal Plain by way of Theta Gap and deposited their load on the Iberia Abyssal Plain (Laughton, 1960). Off the Var, Rhone, Ebro, and off southeast Mallorca and at the eastern end of the Bay of Biscay massive deep-sea fans were emplaced by turbidity currents (Coumes *et al.,* 1982: Vanney and Gennesseaux, 1985). Plastic flow of the Messinian evaporites, in part as a result of sediment loading and in part due to basement tectonics, has deformed the Rhone Fan. Flooding of the Mediterranean by fresh water transformed the region into an estuary with deep Atlantic water inflow and the less dense Mediterranean outflow (Huang and Stanley, 1971). Diester-Hass (1973), however, has questioned the validity of this concept.

By 10,000 years ago when sea level was 40 meters its present level sediment input into the outer shelf decreased and turbidity currents diminished. According to Laughton (1960) this happened in the northern and western margins of the Iberian peninsula 10,000 years ago. With continued rise in sea level and the development of rias and estuaries much of the sediment from land was trapped in these regions creating the present "estuarine phase" of sedimentation. What little pelagic

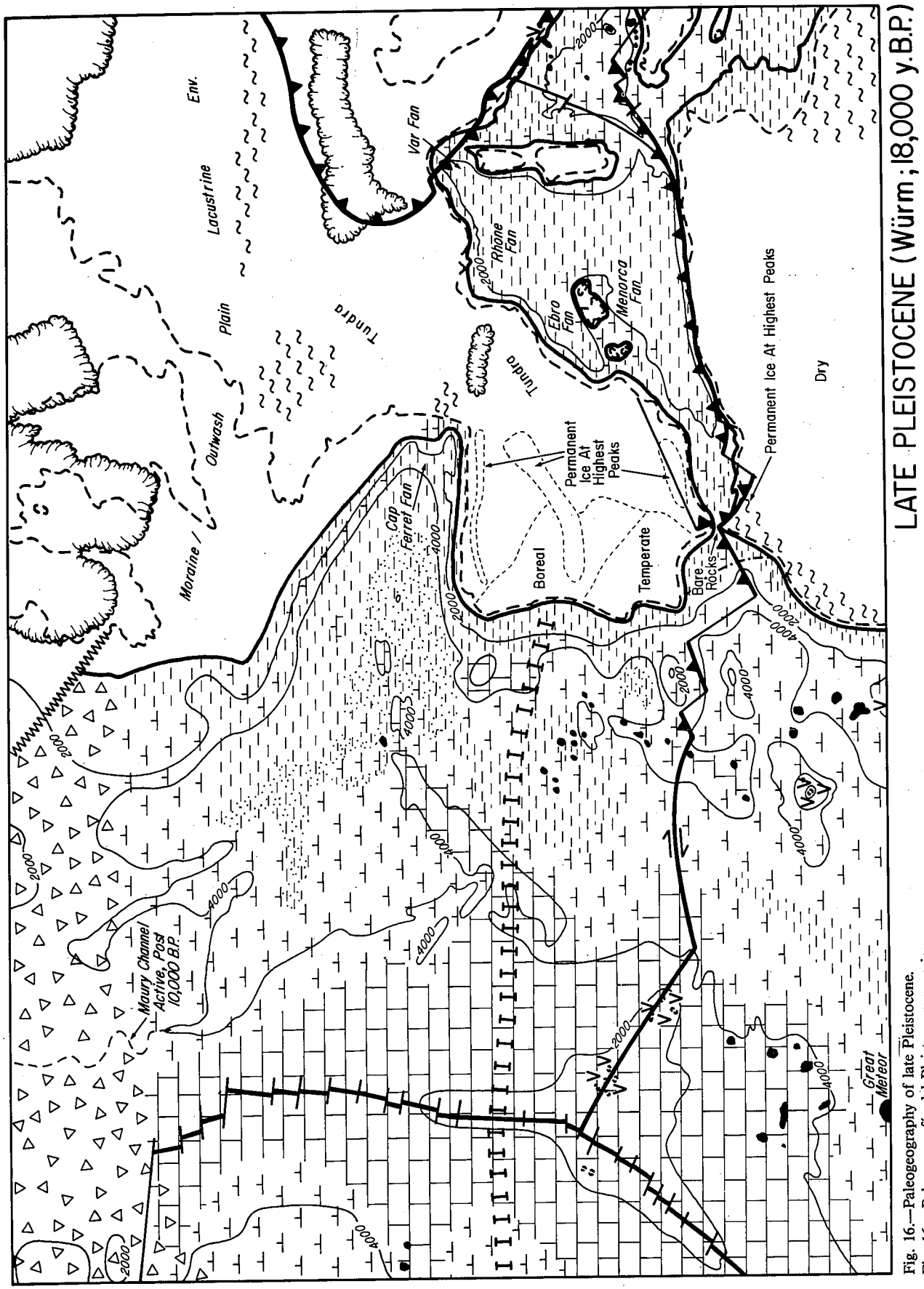

Fig. 16.—Paleogeography of late Pleistocene.
Fig. 16.—Paleogeografía del Pleistoceno superior.

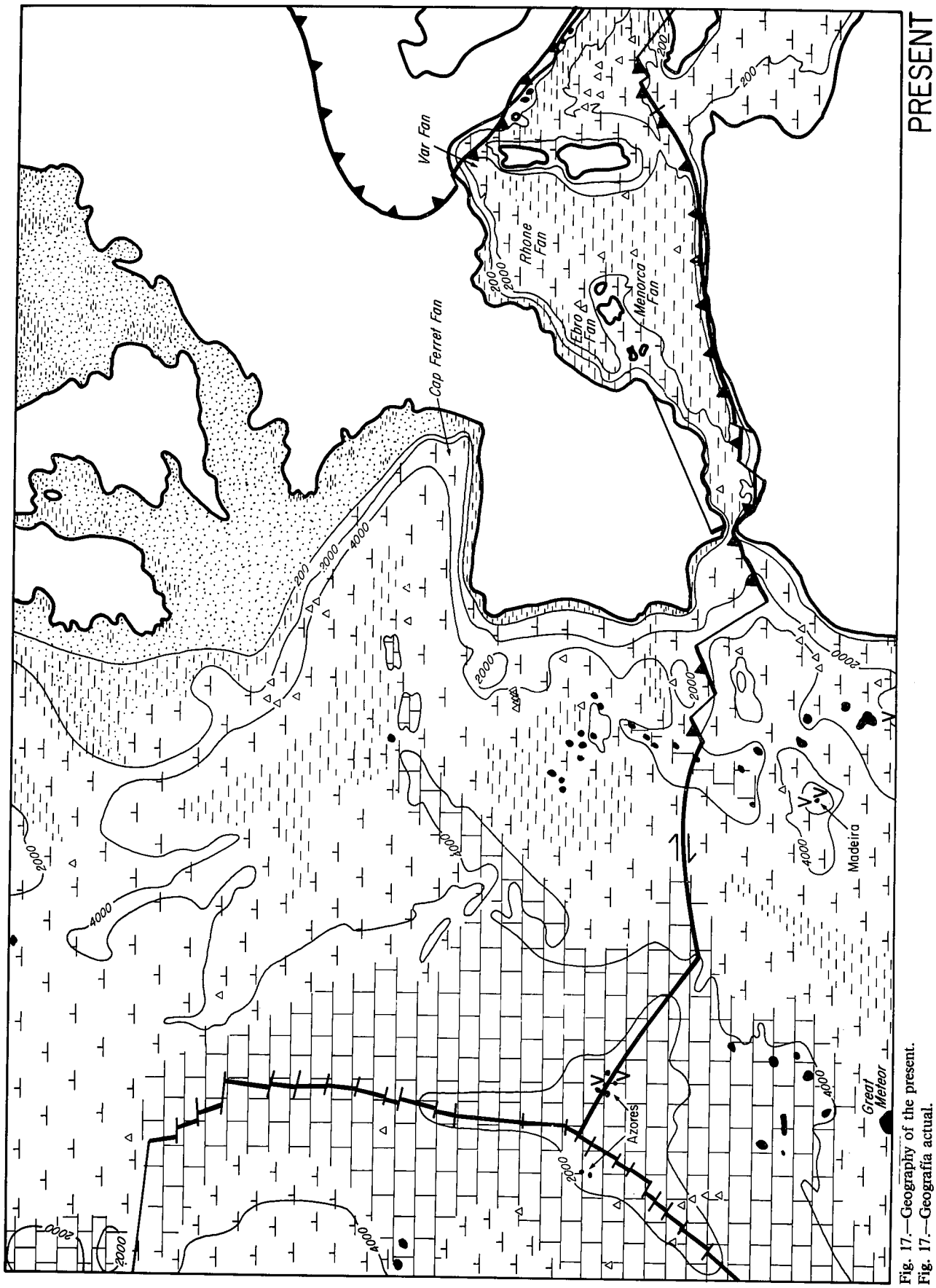

Fig. 17.—Geography of the present.
Fig. 17.—Geografía actual.

sediments escape the estuaries are by-passing the margins and coming to rest in the deep-ocean. Sedimentation in the margins consists predominantly of biogenic carbonates particularly in the topographic highs like those in the Alboran Sea (Milliman *et al.,* 1971). It is only at point sources like in the Ebro and Rhone deltas that there is any significant input into the margin. This is clearly demonstrated by Vita-Finzi (1971) who shows that eustatic records when combined with alluvial chronology of the Mediterranean catchment demonstrate that from 5,000 to 2,000 years ago an increasing proportion of the sediment was trapped in deltas. From 29 B.C. (2,000 years ago) to 1671 A.D. (300 years ago) little sediment was supplied to the region. The high sediment yield since then, much of which is stored in deltas, reflects the increase use of land for agriculture. Some of the sediment escaping the estuaries is carried laterally by littoral drift and then by way of submarine canyons into deeper basins. Such is the case in the Straits of Gibraltar region. After deposition in the Alboran Basin the sediment is reworked westward by the westerly flowing Mediterranean bottom water with the finer grades spilling eastward (Kelling and Stanley, 1971). As the Mediterranean water cascades over the Gibraltar sill it scours the bottom leaving essentially bare rock. West of the Straits of Gibraltar action by the Mediterranean overflow has produced a complex sediment wave topography along the north side of the Gulf of Cadiz (Heezen and Johnson, 1969). This type of morphology can be traced as far north as 38°N off Portugal marking the path of the Mediterranean overflow. At present tectonism is restricted to the Azores-Gibraltar Fracture Zone. Motion along this fault may be responsible for the deformation of the continental rise sediments off north Africa from 36° to 32°N and the sediments on the Madeira Abyssal Plain from 33° to 30°N (Duin *et al.,* 1984: Emery and Uchupi, 1984).

CONCLUSION (FIG. 18).

This brief summary presented above demostrates that Iberia with its roughly east-west trending mountains to the north and south, its easterly dip, and faulted western edge is the result of complex changes in plate geometry that began in the Triassic and are continuing to day. In my disccussion I have asumed that deformation along the south is principally due to north-south compression. This deformation, however, may be due to the lateral traslation of a small plate, the Alboran Plate, from the east that was trapped between Africa and Iberia, a concept suggested by some geologists familiar with the Betic/Rif system. Similarly orogenesis along the Pyrenees/north coast may have involved the interaction of small blocks in a subduction/translation regime comparable to that in northern Venezuela. The end product of these complicated tectonic convulsions is well illustrated by the photographs (fig. 18) taken by NASA from space.

SOURCES

They primary sources used to compile the paleogeographic maps include Emelyanov (1971), Kamen-Kaye (1972), Bernoulli and Jenkyns (1974), Jansa and Wade (1975), Johnson *et al.,* (1975), Jansen (1976), McIntyre *et al.,* (1976), Biju-Duval *et al., (1977),* Ríos (1978), Auffret *et al.,* (1979), Groupe Galice (1979), Ager (1980), Alvarado (1980), Denton and Hughes (1981), Prospero (1981), Surlyk *et al.,* (1981), Jansa and Wiedmann (1982), Ziegler (1982), Emery and Uchupi (1984), Rehault *et al.,* (1985), Leg 103 Scientific Drilling Party (1986), Leg 107 Scientific Drilling Party (1986), Lemoine *et al.,* (1986), Tucholke and McCoy (1986), and Rehault *et al.,* (1987). The base map used is from Schouten *et al.,* (in preparation). In all the maps Africa remains fixed.

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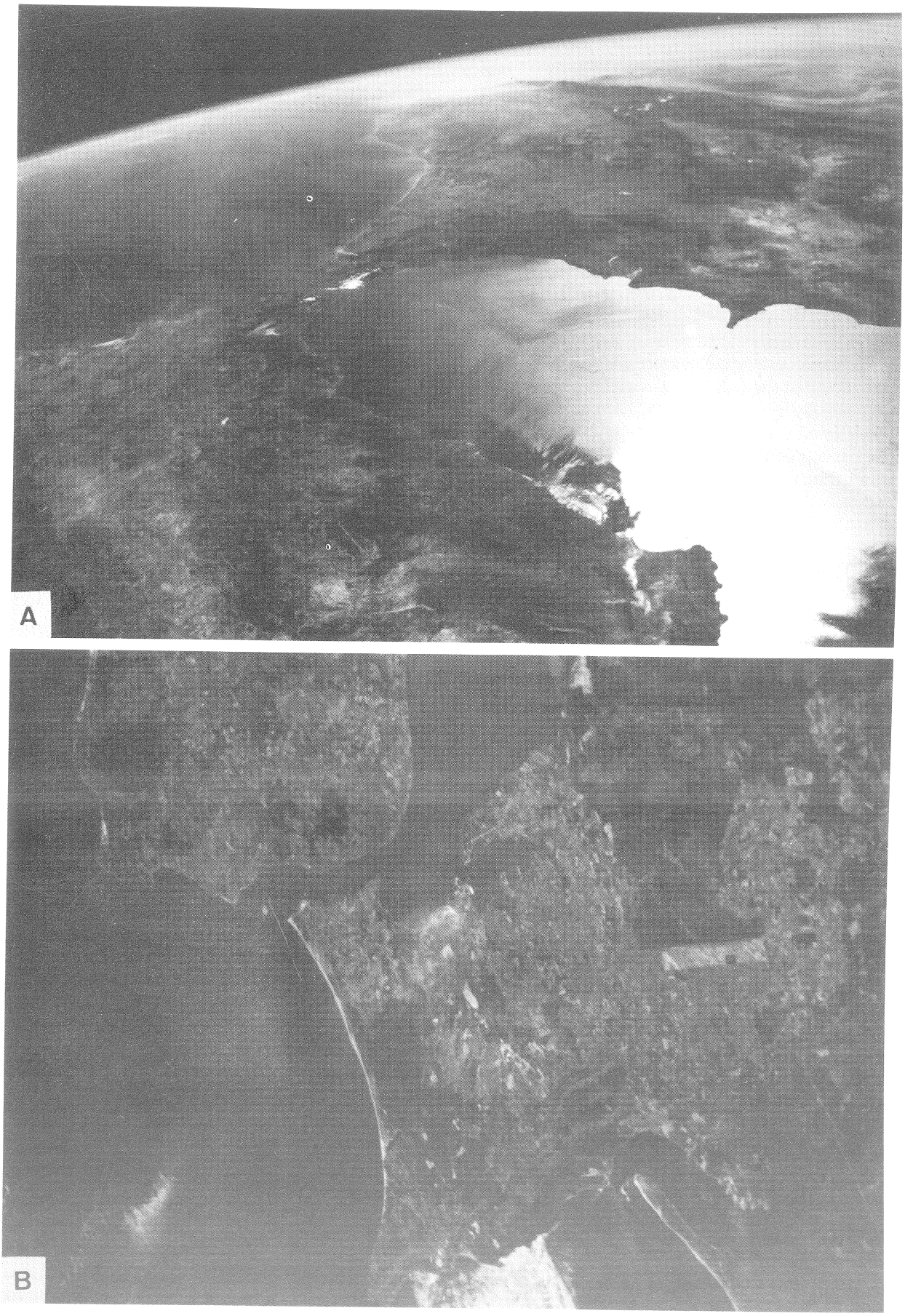

Fig. 18.—Photos courtesy of NASA. A. View from space of the Gribraltar arc. B.—The Lisbon estuary on the Lusitania basin.
Fig. 18.—Fotografías desde satélite suministradas por la NASA. A. Vista del arco de Gibraltar. B. El estuario de Lisboa en la cuenca lusitana.

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