A GRAPHIC DISPLAY OF PLATE TECTONICS IN THE TETHYS SEA

by

Anthony B. Williams
S.B. California Institute of Technology
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Signature of Author Signature redacted
Department of Earth and Planetary Science

Certified by Signature redacted 30 Aug 1971
Thesis Supervisor

Accepted by Signature redacted
Chairman, Departmental Committee on Graduate Students

ABSTRACT

The hypothesis of plate tectonics is applied to the Mediterranean Sea region in order to explain its Tertiary mountain chains and to derive previous positions for the component plate fragments. Consuming plate boundaries are selected on geophysical and geological grounds, and are assumed connected through modern oceanic regions by shearing and accreting boundaries. The resultant plate fragments are rotated to fit geological and morphological similarities, and the timing of tectonic events, without violating available constraints. The cumulative rotations are plotted to illustrate behind-arc spreading in the western Mediterranean basins and closure upon the Adriatic and Ionian seas from east and west following northward motion of the central Alpine plate fragment.

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INTRODUCTION

This paper is an attempt to interpret the present geological structure of the circum-Mediterranean Alpine mountain chains according to the hypothesis of plate tectonics. In the next two sections, we shall combine observations made of several plate boundaries in order to make a simple model of one way in which mountain chains may evolve, and in the following sections force sparse geological and geophysical data from the study area to fit into this model. We then present an approximation to the history of the Tertiary Mediterranean Sea region, as a sequence of reconstructions of the probable positions of plate fragments determined by the applied model. Discrepancies between this approximation and other evidence from the Mediterranean will either refine the approximation, indicate our misinterpretation of the data employed, or lead to an improved hypothesis. In the remainder of this section, we introduce the postulates of plate tectonics, the general geological background of the study area, and briefly preview the evolution of the Mediterranean Sea which we believe has occurred.

The history of the development of the hypothesis of plate tectonics has been reviewed by Vine (1970). Plates are taken to be rigid segments of a spherical shell, the lithosphere, which comprises the outermost portion of the Earth and is perhaps 50 - 100 km thick. These plates extend laterally for distances as great as several thousand kilometers, and move across the Earth at speeds of a few centimeters per year relative to each other. At present, the active plate boundaries are the loci of most faulting and tectonism, and may have sustained activity for as much as 200 x 10^6 years. These boundaries may have dimensions of up to many thousands of kilometers along strike, but tectonically affect no more than several hundreds of kilometers transversely. At boundaries, new lithosphere is added on to plates (mid-ocean ridges, basins behind island arcs, continental rifts), old lithosphere is returned to depth (island or continental margin volcanic arcs, with an associated trench, possibly sediment filled), or lithosphere is sheared horizontally (active segments of transform faults between triple point junctions). The relative motion of any pair of plates may be specified by the instantaneous location and magnitude of a pole of rotation fixed to both plates or their imaginary extensions.

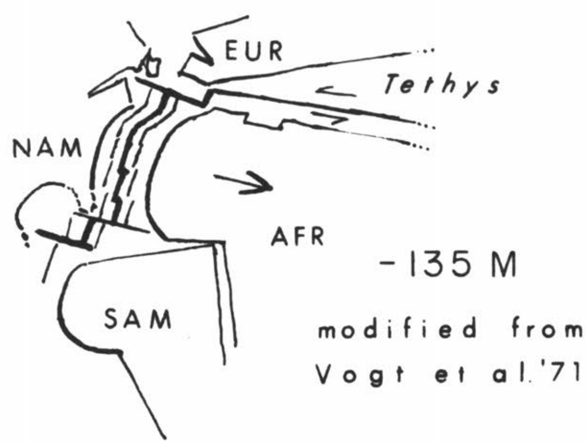
For plates coupled to an opening ocean, the amount of opening may be obtained from the study of linear magnetic anomalies fixed in the newly created lithosphere under appropriate conditions, and the direction of opening may be indicated by fault traces. The presence of consuming boundaries around an ocean not only obliterates the effects of earlier opening, but decouples the abutting plates, so that their previous positions are not as easily inferred.

Attempts to sum poles known from the opening side of the plates involved may lead to large accumulated error (Morgan, 1968), so that indirect evidence is required to constrain possible plate motions.

From the immense accumulations of sediments bearing many indications of deep water deposition (Rutten, 1969), from paleomagnetic orientations of the surrounding continents (Francheteau, 1970) and the fit of their margins (Bullard, Everett & Smith, 1965), it is apparent that the present Mediterranean Sea is a remaining expression of the Tethys Sea which once separated India, Arabia and Africa from Asia and Europe. Since the breakup of Pangea (Dietz & Holden, 1970), deformation of the Tethys is required by the opening of the Atlantic Ocean (Funnell & Smith, 1968), although only part of the sequence of motion of the abutting plates can be specified from this source (Vogt, Higgins & Johnson, 1971). We here instead take the approach of assuming continuity among several geological structures described around the Mediterranean Sea, and assuming that the plate tectonic hypothesis holds within the Tethys Sea on a scale of several hundred kilometers.

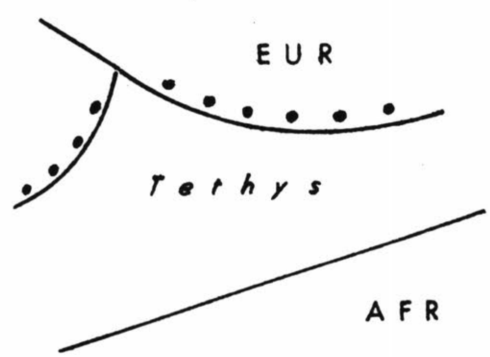
In order to represent the motion of rigid bodies on the surface of a sphere, maps of previous plate fragment positions were generated by a computer program (HYPERMAP, R. L. Parker) which transforms coordinates in the required manner. This program has been modified to permit live display and keyboard control of the strings of geodesic segments which represent the plate fragments under discussion, in order to try various positions before selecting the ones presented here. No mechanical optimization of these positions was performed, as no sufficiently simple and uniform criteria are yet available within the present hypothesis of plate tectonics, nor within the present knowledge of Mediterranean geology.

The descriptions of the Tethys Sea referred to above indicate that near the beginning of the Triassic, its shape was roughly that of a V lying on its side. The point of the V was located between southern Spain and northern Africa, and the sides opened eastward along the southern shores of Europe and Asia, and along the northern shores of Africa, Arabia, and India. As Africa and South America moved away from North America and Europe, left lateral shear must have occurred somewhere within this V, which with the opening of the central North Atlantic, made Tethys into a long seaway, open to at least the Caribbean Sea by the Jurassic (Hallam, 1971). The later opening of the North Atlantic between North America and Europe in turn spread Africa away from Europe (Funnell & Smith, 1968).



We know that India was moving north toward a subduction zone dipping north under Asia during the Mesozoic and early Cenozoic (Dietz & Holden, 1970), and distortions of magnetic anomaly patterns suggest that Siberia sheared right laterally with respect to Russia during the same period of time (Hamilton, 1970), so that the eastern end of Tethys was closing at least during the Cretaceous. The counterclockwise rotation of Spain relative to internal Europe determined by paleomagnetic evidence (Van der Voo & Zijderveld, 1971) and other material discussed in the later section concerning the Iberian Peninsula support the idea that the eastern end of the Tethys was closing during the Cretaceous also.

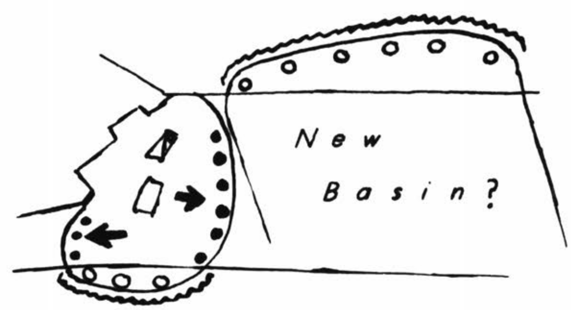
We now summarize the conclusions drawn in the body of this paper, and sketch the history of the western Mediterranean basins which will be displayed in more detail in figures 3 through 9. During the last part of the Cretaceous through the Paleocene (figures 9 and 8), an island arc to the south of Europe with subduction zone dipping north was consuming the Tethys Sea in what is now the central Mediterranean Sea between the Gulf of Sirte and the Adriatic Sea. This arc collided with the northern



Late Cretaceous

edge of the African continent and was thrust back over Europe from southern France east to the Balkans, raising the Helvetic Alps and the Carpathian Mountains in the Eocene. The rotation of the Iberian Peninsula requires that another arc, convex to the southeast, bordered southern Spain with a subduction zone dipping to the northwest during the Cretaceous. The rotation of the Iberian Peninsula was complete at least by the Late Eocene, when the associated arc struck northern Africa, raising the central part of the Atlas Mountains in Algeria, figure 7.

The two portions of the arc which were not yet in collision with Africa, in the west and the east, then continued to consume the western Tethys Sea, while spreading behind the eastern portion



Late Eocene

of the original arc broke Corsica and Sardinia away from Spain and France.

The western portion of the original arc moved to the west, and thrust the Riff Mountains onto Africa to the south and the Betic Cordillera onto Spain to the north in the Oligocene and Miocene (figures 5 and 6). During this time, the eastern branch of the original arc bordering the Iberian Peninsula moved east along the northern coast of Africa, extending the Atlas Mountains through Tunisia. By the end of the Miocene, Corsica and Sardinia were caught between Europe and Africa, so that the locus of spreading behind the arc jumped to the Tyrrhenian Sea. The change in locus may be related to the fact that the eastern branch of the earlier arc around the Iberian Peninsula must have been consuming the new basin behind the plate fragment which had raised the Helvetic Alps, rather than the older Tethys seafloor. The northern part of this eastern arc, recognized as the Apennines, swung northeast into the Adriatic Sea, while the southern portion swung southeast as the Calabro-Sicilian arc during the Pliocene (figure 4). The schematic representation of the plate fragments in figure 3, which is the starting point for the motions and subsequent positions in figures 4 through 9, should be compared with the conventional illustration and place names depicted in figures 1a and 1b. The schematic representations are Mercator projections, from the computer program mentioned above, with plate boundaries added by hand. Double lines limit newly created sea floor, and active consuming margins are marked with triangles on the overriding side. Arrows indicate motion of plate fragments since the earlier projection.

The early Tertiary positions of Northern Sicily and the later Tertiary positions of the Calabrian fragments are not well specified by the available data. The size of these fragments is perhaps about the size of the expected error in the positions of the larger fragments, and about the smallest size of the plates which are considered within the discussion of plate theory which follows. It is strongly emphasized that much of the constraints which are discussed in the interpretation of this theory are obtained from spot data rather than complete coverage, that only a small fraction of the available literature has been examined, and that the available data has been forced into the model. We are attempting to develop a first approximation to a complex structure and its evolution, rather than some sort of photographic record of great accuracy.

SPREADING BOUNDARIES

We here attempt to obtain a general idea of the structure of one opening ocean presumably formed by rifting initiated within a continent, not later modified by consuming boundaries, so that we may later compare structures found elsewhere. A mature spreading boundary such as the Mid Atlantic Ridge forms new basement, which remains connected to the plates on both sides of the boundary until returned to depth by a subduction zone. Mid Atlantic Ridge dredge hauls contain olivine tholeiitic basalt to gabbro in pillow lava structure (Myashiro, Shido & Ewing, 1969, 1970a, 1970b) with greenschist to greenstone (Van Andel & Bowin, 1968; Melson & Van Andel, 1966). A seafloor model consistent with these samples, with geophysical data, and with present geochemical theory has been proposed by Cann (1970); away from the ridge, layer 2 may be a built up pile of pillow lava basalts with local clays and greenschist, and layer 3 may be green hornblendes replacing basalt ferromanganese among dike swarms more coarsely grained downward to granitic-dioritic layered gabbro intruded by serpentinitic diapirs from the (peridotitic) mantle. This model independently approximates the Troodos massif of Cyprus, mostly oversaturated tholeiites (Gass, 1968), and the Vourinos and Pindos complexes in northern Greece, chert over pillow lavas over gabbro over pyroxenites (Moores, 1969) and is taken to represent the hot emplacement stage of basic and ultrabasic igneous rock found as ophiolites (Dewey & Bird, 1971).

Typical sediments found on the Atlantic floor from the ridge crest to the continental rise are local volcanic sands (Fox & Heezen, 1965), followed by carbonate changing to silicate ooze at depth, mixed with a fine aeolian contribution before terrigeneous sediments dominate, chiefly turbidite sands (Hersey, 1965; Van der Lingen, 1969) on the abyssal plains. On the continental rise, contourites (Schneider et al., 1967) interbed the turbidites, often in extensive slumps (Emery et al., 1970), providing the material for a miogeosynclinal prism (Dietz, 1963).

From the continental rise to the shore, basement structure, sedimentary deposition, and igneous petrology are affected by the initial rifting stage, and are quite distinct from the products of mature spreading. The continental slope of eastern North America is underlain by deep, linear, subparallel troughs (Drake, Ewing & Stockard, 1969), usually parallel to the coast but possibly extensions of half-grabens observed on shore (Sanders, 1963; Grim, Drake & Heirtzler, 1970). The troughs are separated by a crystalline ridge, and acted as sediment traps during the Mesozoic, so that continental rise deposits did not prograde extensively until the Cenozoic (Emery et al., 1970). Around the North Atlantic coast, Triassic and Jurassic diabase dikes appear parallel to what may be tensional components of some pre-drift central stress field (May, 1971), which seems to show NW-SE compression in the southeastern United States and NE-SW sinistral shear in the northeastern United

States (De Boer, 1967). If the African Rift system represents an early stage of continental drift, we may gain some insight into the igneous petrology of rocks now eroded from and buried under the Atlantic continental margin. The Kenya Rift Valley is reported to have a sequence of epeirogenic uplift, episodes of basic volcanism, regional uplift with intrusive silicic volcanism, and central collapse through the eruption of trachytic ignimbrites, with less than 10 km of lateral extension over 20 myr (Baker & Wohlenberg, 1971). In the Ethiopian Rift, flood basalts have erupted continuously for about 40 myr (Rex, Gibson & Dakin, 1971). Le Bas (1971) has summarized the behavior of some 20 regions in Africa and along the coasts surrounding the South Atlantic, to show that the locus of continental separation connects irregularly distributed regions which are uplifted as much as 1 km to form domes 500 to 1000 km across, generally broken by central grabens, and which contain a peralkaline suite of basalts, so that the rifting stage is distinct from the separation stage of continental drift. We conclude that the origin of the passive boundary between ocean and continent may be complex and diffuse in time and space, compared to the mature spreading boundary between plates.

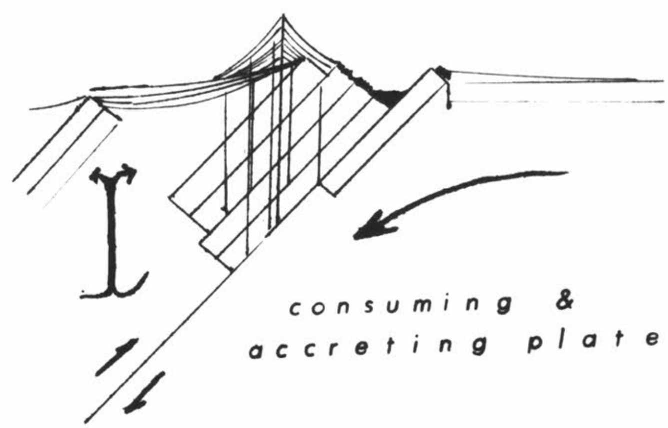
As a result of the uplift of the rifted site, the accumulation of sediments does not occur until there is sufficient separation for the development of traps. Shortly after rifting, the restricted circulation and isolation of the new seaway favor accumulation of bedded salts, possibly later modified by diapirism. The salts of Mexico (Viniegra, 1971) and of the South Central United States may correlate with rifting in the Gulf of Mexico (Freeland & Dietz, 1971), and the Aptian Gabon evaporites may mark the later opening of the South Atlantic (Rona, 1969; Wright, 1971). The Mississipian evaporites of the Canadian Maritime Provinces (Evans, 1970) have been taken to support rifting in the Carboniferous to Permian (Belt, 1968; Dewey & Bird, 1970).

Our model for an ocean which results from initiation of a spreading center within a continent has two distinct stages. The rifting stage involves uplift with peralkaline volcanics, and may continue without further development. Salts may be deposited as continental separation begins, and the separating stage develops as two independent sequences; erosion of the continental margin leads to the filling of tectonic sedimentary traps near shore, followed by mature development of the continental rise, while sea floor is generated at the ocean ridge and accumulates a characteristic sedimentary cover. This mature phase remains until affected by a consuming boundary, which we describe next.

CONSUMING BOUNDARIES

A consuming boundary is the site at which the consumed plate disappears from the Earth's surface, marked by an oceanic trench, possibly sediment filled. If the consuming plate is oceanic, it will carry an island arc such as the Kurile-Kamchatka arc; if it is continental, it will carry a volcanic mountain range such as the Andes. Oceanic arc structure is typically double, with a young volcanic ridge inside an older quiescent ridge inside the trench (Gorshkov, 1967; Sykes et al., 1969; Karig, 1970). The trench may be 100 ± 50 km wide, separated by 175 ± 100 km from an arc of volcanic islands 75 ± 25 km wide (Dickinson, 1971). A detailed survey of earthquake foci below Japan found that 75% of recorded shocks could be plotted on two surfaces, separated by 45 ± 15 km vertically, extending along the arc in segments about 190 km long, with the active volcanoes located where the upper surface was 90 to 180 km deep (Carr, Stoiber & Drake, 1971), over folds plunging downdip in the seismic surfaces (Stoiber & Carr, 1971). Displacement associated with motion on the consuming plate boundary suggests reverse fault uplift immediately behind the Japan trench with subsidence past a hinge line further inside the arc (Fitch & Scholz, 1971). Reflection profiles across the Antilles arc suggest underthrust faults on the consumed plate, dipping toward the trench, with crumpling at the elevated end of the thrust slices, which in turn may become covered with sediments slumping down the trench hanging wall (Chase & Bunce, 1969).

Dickinson (1971) has concluded from a study of circum-Pacific arcs that the descending slab carries marine sediments toward the trench, under crumpled turbidites along the hanging wall, down the Benioff zone through high pressure, low temperature metamorphism of blueschist grade, to release andesite eruptives and granitic intrusives. The ratio of K / Si in the volcanics increases inward from the trench, and the volcanics weather to copious graywackes, which accumulate in structural traps toward the trench. The high temperature, low pressure region about 150 km behind the trench is an andalusite, sillimanite, kyanite isograd associated with high surficial heat flow (Oxburgh & Turcotte, 1971).



It seems plausible that the sediment fraction of the descending slab cannot be carried to depth, under the same buoyancy rule that prevents consumption of continents (McKenzie, 1969). Successive upper slices of the descending slab may accumulate, separated by reverse faults dipping under the volcanic pile (Hsü, 1971). The point of maximum descent of these slices is at the trench, which represents the inward limit of active thrust faulting. Uplift and thrusting may occur on the consumed plate outside of the trench, e.g., the Mediterranean Ridge to the south of the Hellenic arc (Rabinowitz & Ryan, 1970).

If the bottom portion of these slices include the oceanic layers 2 and 3, cold thrust emplacement of the ophiolite suite will occur (Dewey & Bird, 1971). In classical terminology, these slices are nappes in an eugeosyncline, covered at their front with crumpled trench turbidites and broken fragments of their own composition, and cut to the rear by igneous diapirs. If the descending slab contains continental material, the diapirs may be anorthositic, with layered gneiss and quartz monzonite (Martignale & Schriver, 1970), dated close to the time of napping (Naylor, 1970).

As the assemblage of upper lithosphere slices, volcanic pile, and local sediment moves with the consuming plate, it slowly grows in volume, analogous to the process of chip formation with a built up edge in materials processing. As the consuming plate approaches a continent, this assemblage will be crumpled against and over the continental margin, thrusting and crumpling continental rise and shelf deposits shoreward, completing the tectonic phase of the eu- and mio-geosynclinal couple, before a change in either site or sense of plate consumption must occur.

This model is able to account for most of the features associated with mountain chains that have been described as true geosynclines (Aubouin, 1965).

Karig (1970) suggested that small ocean basins such as the Lau Basin behind the Tonga-Kermadec arc result from ocean floor spreading behind the consuming margin, resulting from the rise of light, surficial material from the downgoing slab. Sleep (personal communication, 1970) has modeled this situation and prefers spreading as a result of secondary circuits in the asthenosphere forced by the drag of the downgoing slab, and rising behind the arc. This behind-arc spreading and plate accretion may explain the apparent separation of the Japanese island arc and the Kurile-Kamchatka arc from their respective mainlands (Hess, 1962), and has been modeled by Dewey & Bird (1971).

We suppose that the age of the oldest sediment in the basin dates the initiation of structural traps as behind-arc spreading occurs, and that sialic fragments of the accumulated eugeosynclinal pile remain in the wake of the modern arc, separated from it by basement of mixed new oceanic and old continental character, bounded by transform faults of opposite sense to either side of the plate fragments, as in the marginal sketch above.

The minimum overall width of such a small consuming and accreting plate is the distance from the trench to the spreading locus behind the arc, 235 $\pm$ 125 km in Karig's hypothesis, over 600 km in Sleep's hypothesis. Both of these hypotheses may apply, in that rifting could be initiated by diapiric action and later accelerated by the induced secondary flow, so that the plate fragment width may be about 250 km in the beginning, increase to about 600 km, and then remain at that size. The smallest observed length along the arc of a plate fragment may be the 190 km average segment length found for the Japan arc by Stoiber, Carr & Drake (1971). A reasonable minimum for the size of an active plate fragment is therefore about two degrees of latitude. This is the smallest scale of continuous geologic structure which we expect to be caused by a single plate fragment in the discussion of the geology of the circum-Mediterranean Alpine mountain chains which follows.

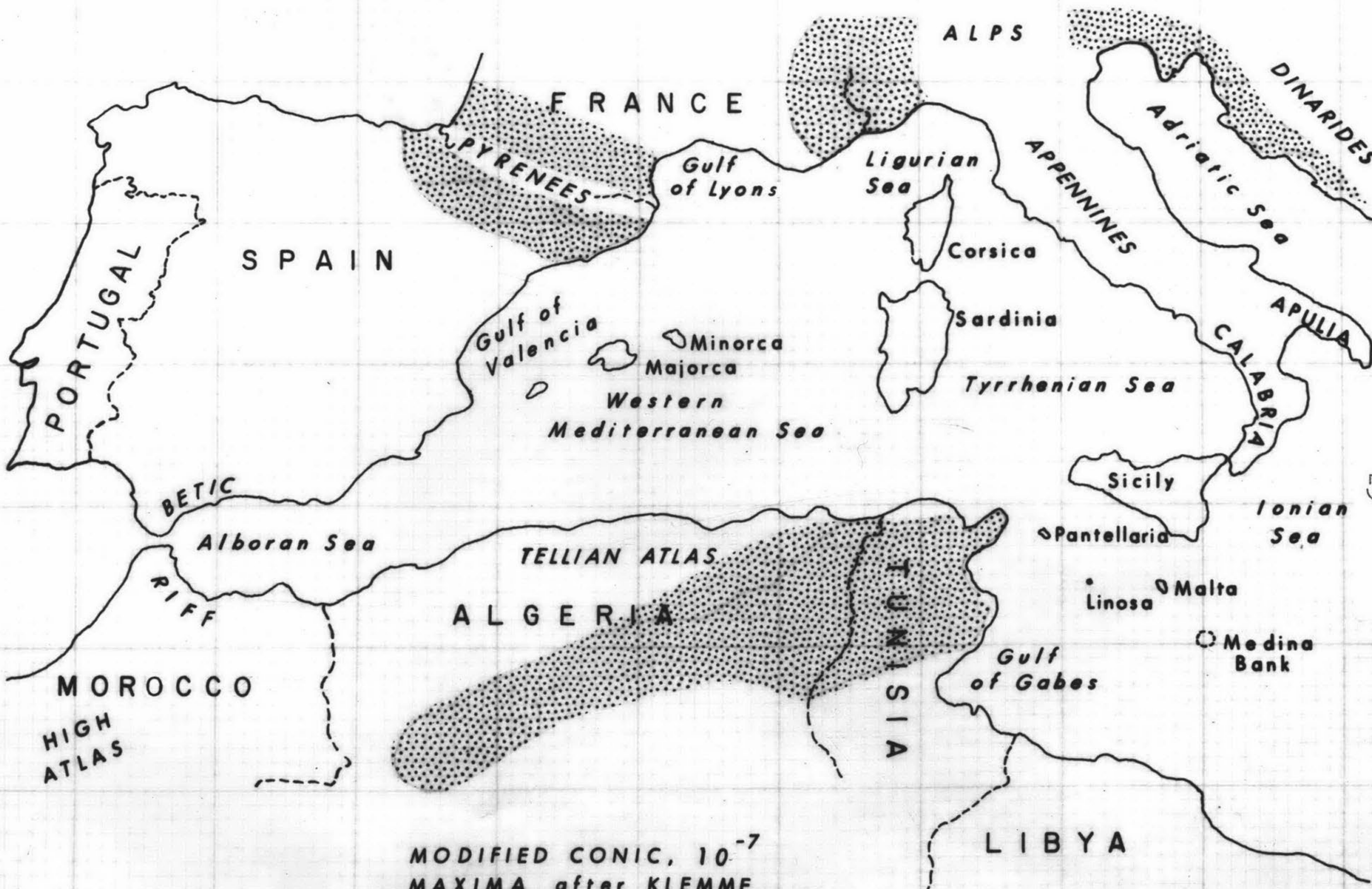
MEDITERRANEAN PLATE BOUNDARIES

Rutten (1969) has comprehensively summarized European geological literature concerning the Alpine mountain chains around the Mediterranean Sea (figures 1a and 1b); of this division of European landforms, he states "Geographically, it comprises the Betic Cordillera in southeastern Spain and the Pyrenees between Spain and France and their westerly continuation; further parts of the Provence in southern France and all of the Alps; French, Swiss, Austrian and Italian. Included also are the Jura Mountains in eastern France and northwestern Switzerland, the Carpathians, several Mediterranean islands, the Apennines of Italy, and most of the Balkan." Outside of western Europe, we also include the Riff and Atlas mountains of northwestern Africa (De Sitter, 1956), the Zagros Mountains of Iran (Wells, 1969), and the Tauride and Pontide mountain chains of Turkey (Bailey & McCallien, 1950). Although the Alpine orogeny extends through the Himalayas to southeastern Asia, the motion along the Urals (Hamilton, 1970), the diffuse fracturing of continental regions in collision (McKenzie, 1970) and the dearth of reported geological control limit the present exercise to the region west of the Caspian Sea; and particularly to the Western Mediterranean Sea region.

After the topographic expression of uplift and folding, we notice the variation in the amount of sedimentary material included in these mountain chains. Klemme's (1958) isopachous-lithofacies maps for the Triassic through Eocene display three long sets of narrow sedimentary maxima, extending along the Atlas, along the Western and Southern Alps, the Dinarides and Hellenides, and northeast through Turkey and Syria before striking southeast through Kurdistan. These geosynclinal deposits are indicated schematically in figures 1a and 1b, and are interpreted to have been formed off the shores of the Tethys Sea (Wezel, 1970).

Late Jurassic ophiolites, carbonate breccias, and radiolarian cherts found in these geosynclines require that deep sea conditions existed in the Tethys by the Jurassic (Hallam, 1971). Jurassic sea floor spreading in the Atlantic with an average left lateral shear in the Tethys is required by the Atlantic magnetic anomalies (Vogt, Higgins & Johnson, 1971), by the distribution of Mediterranean foraminifera and circum-Atlantic paleomagnetic data (Berggren & Phillips, 1969), and by the geometric constraints upon the Permian-Triassic continental fit (Bullard, Everett & Smith, 1965; Funnell & Smith, 1968).

Plate spreading and shear alone are unable to account for the present geography of the Mediterranean, as the great continental plates of Europe and Africa are also separated by the pre-Triassic Iberian Peninsula, Corsica, Sardinia, Sicily, and Italy. The thrust sheets or nappes, folding and uplift which comprise the tectonic elements of the Alpine chains mark consuming plate boundaries, evidence that these older plate fragments consumed the Tethys to form the present Mediterranean. We consider these Alpine chains in the



MODIFIED CONIC, 10^{-7}
MAXIMA after KLEMME

figure 1a


PONTIDES
TURKEY
TAURIDES
Aegean Sea
HELLENIDES
Ionian Sea
Antalya Bay
SYRIA
Eastern Mediterranean Sea
Gulf of Sirte
LIBYA
EGYPT
SAUDI ARABIA
Red Sea
MODIFIED CONIC, 10^-7
MAXIMA after KLEMME

figure 1b

Western Mediterranean after a discussion and presentation of the figures which illustrate most of our argument.

REPRESENTATION OF PLATE MOTIONS

Figures 3 through 14 illustrate our discussion of the motions of the plate fragments which we believe to have been significant in the evolution of the present Mediterranean Sea. Two kinds of boundaries are shown; the fine dotted lines generated by digital computer depict modern shorelines and mountain chains, while the inked lines drawn by hand represent plate boundaries.

A FORTRAN computer program (HYPERMAP, R. L. Parker) for the generation of plotter output of geographical maps in various projections and rotations was modified to run in an interactive mode on the dual PDP-7 system with a digital CRT and high speed precision plotter located at the Lincoln Laboratory Group 28 Seismic Discrimination Center. A card deck containing the coastlines of the world digitized at $.1^{\circ}$ intervals included with the original program was divided into segments associated with the plates fragments discussed in this exercise. These segments were joined into closed figures along highly simplified linear approximations to the mountain chains which we believe to have been formed by the interaction of the plate fragments. The Mercator projection in figure 3 is thus a straight line segment approximation to the shoreline of the present Western Mediterranean Sea. The remaining figures show where this modern shoreline would have been located in the past, not the location of the earlier shoreline. Using these outlines as a guide, finite poles of rotation were chosen in the coordinates which we apply to modern Europe, taken fixed for convenience in all the illustrations. It should be stressed that the location and magnitude of all but two of the poles employed were derived from weak constraints, such as continuity of structure, paleomagnetic pole rotations, and morphological similarity, rather than from magnetic lineation patterns and fault traces such as used by Morgan (1968) or from a least squares fit to some isobath as used by Bullard et al. (1965). Since these poles are essentially subjectively picked, and since their location and magnitude was seldom critical, the exact values used are not quoted here, in order to avoid the implication of accuracy from mere precision.

After a complete set of trial rigid rotations on a sphere was generated without violating these weak constraints and others discussed later, plausible plate boundaries were drawn by hand. Transform or shear boundaries were taken as small circles about the relevant pole of rotation connecting consuming boundaries shown with triangles on the consuming margin and identified by application of the model discussed above to the available geological literature. The process by which spreading behind island arcs takes place is poorly understood, so that the question of the location of spreading boundaries may be meaningless for this case. Without spreading behind island arcs, however, we cannot explain the recent subsidence inferred from the thin sedimentary cover, nor the presence of continental rise material thrust above the shoreline. In order to represent this spreading, a suitable line reference was chosen in the wake of each arc before each motion illustrated, and this line was then traced on the diagram representing the completed motion, first in the position which it would have had if attached to the plate left behind, then in the position it would have had if attached to the moving plate fragment. In this way, the area of newly spread ocean floor was obtained, represented by the region between the double lines in the figures. The reference line was also rotated half way and traced as a double dashed line to indicate the approximate position of the spreading boundary if it obeyed the rules of symmetrical spreading associated with mature ocean ridges. Where possible, the reference line was chosen to lie along some freely interpreted trend in the magnetic anomaly field reported by Vogt et al. (1971). Otherwise the reference line was chosen to be a radius from the pole of rotation where possible, and arbitrarily elsewhere. Since the plate boundaries, especially spreading boundaries, are not well defined by the available evidence, and since the stability of triple junctions was not required, no vector sums (McKenzie & Morgan, 1969) were computed where three plate fragments join.

Certain critical ambiguities relevant to this exercise occur in the data which has been published about the Mediterranean Sea, roughly divisible into two categories: the motion of Africa relative to Europe, and the date of the rotation of the Iberian Peninsula. Our main stream of discussion is directed toward a complex evolution which is most strongly constrained: Africa is assumed fixed relative to Europe, and the Iberian Peninsula is still moving in the Tertiary, figures 3 through 9. Figures 10 through 12 illustrate a simpler model in which Africa is allowed to close toward Europe, in contrast to figures 4 through 6, in order to examine the effect of the first ambiguity. Figures 13 and 14 illustrate another simpler model in which the Iberian Peninsula is not moved relative to Europe after the Mid Cretaceous, in order to examine the effect of the second ambiguity. Each alternate model is discussed in the appropriate section.

Our discussion of the evolution of the Mediterranean Sea falls naturally into a spatial and temporal arrangement, since the Calabro-Sicilian arc is critical in the Pliocene (figures 3 and 4; alternate in figure 10), the Apennine arc in the Miocene (figure 5; alternate in figure 11), the Riff-Betic arc in the Oligocene (figure 6; alternate in figure 12), the Alpine arc in the Eocene (figures 7 and 8), and the Iberian Peninsula at the opening of the Tertiary (figures 8 and 9; alternate in figures 13 and 14). With this key in mind, all of these diagrams are displayed in the next few pages for quick comparison. Although the main sequence of motions appears complex, it is the simplest model which fits the tightest set of constraints and relaxation of these constraints simplifies the model, as we shall see.

FIGURE 3 WESTERN MEDITERRANEAN SEA (-0M)

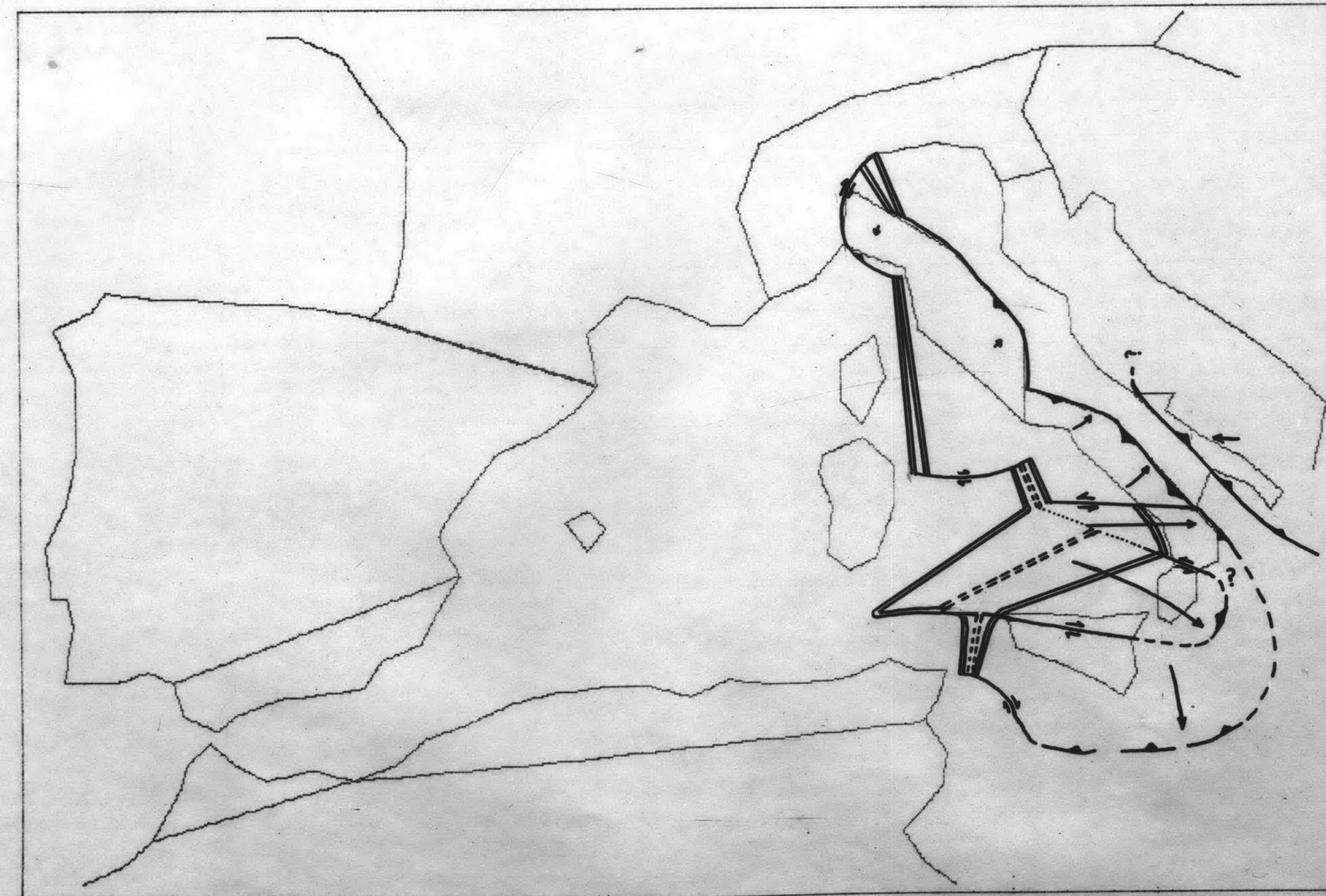


FIGURE 4 MID PLIOCENE WESTERN TETHYS (~5M)

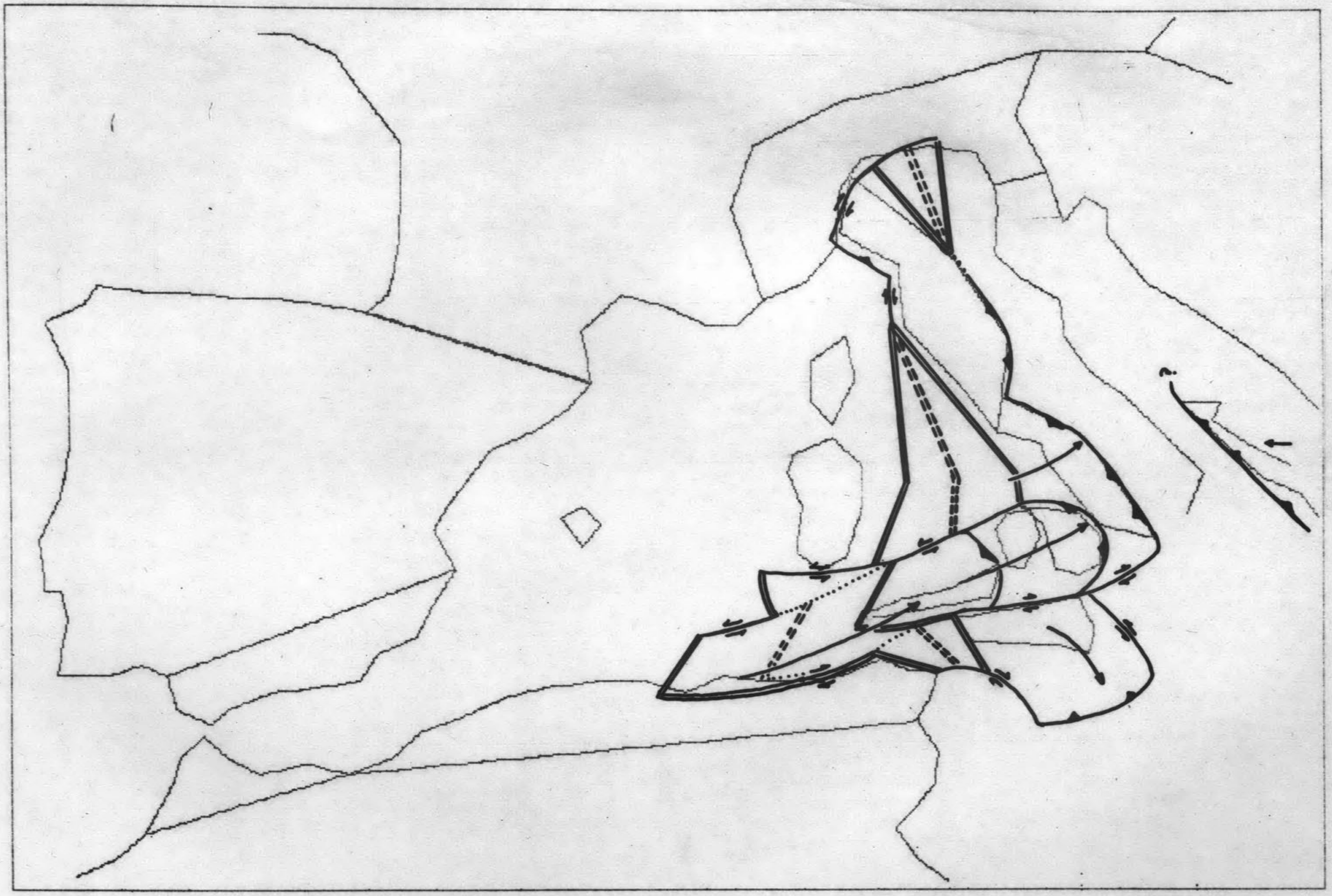


FIGURE 5   LATE MIOCENE WESTERN TETHYS (-15M)

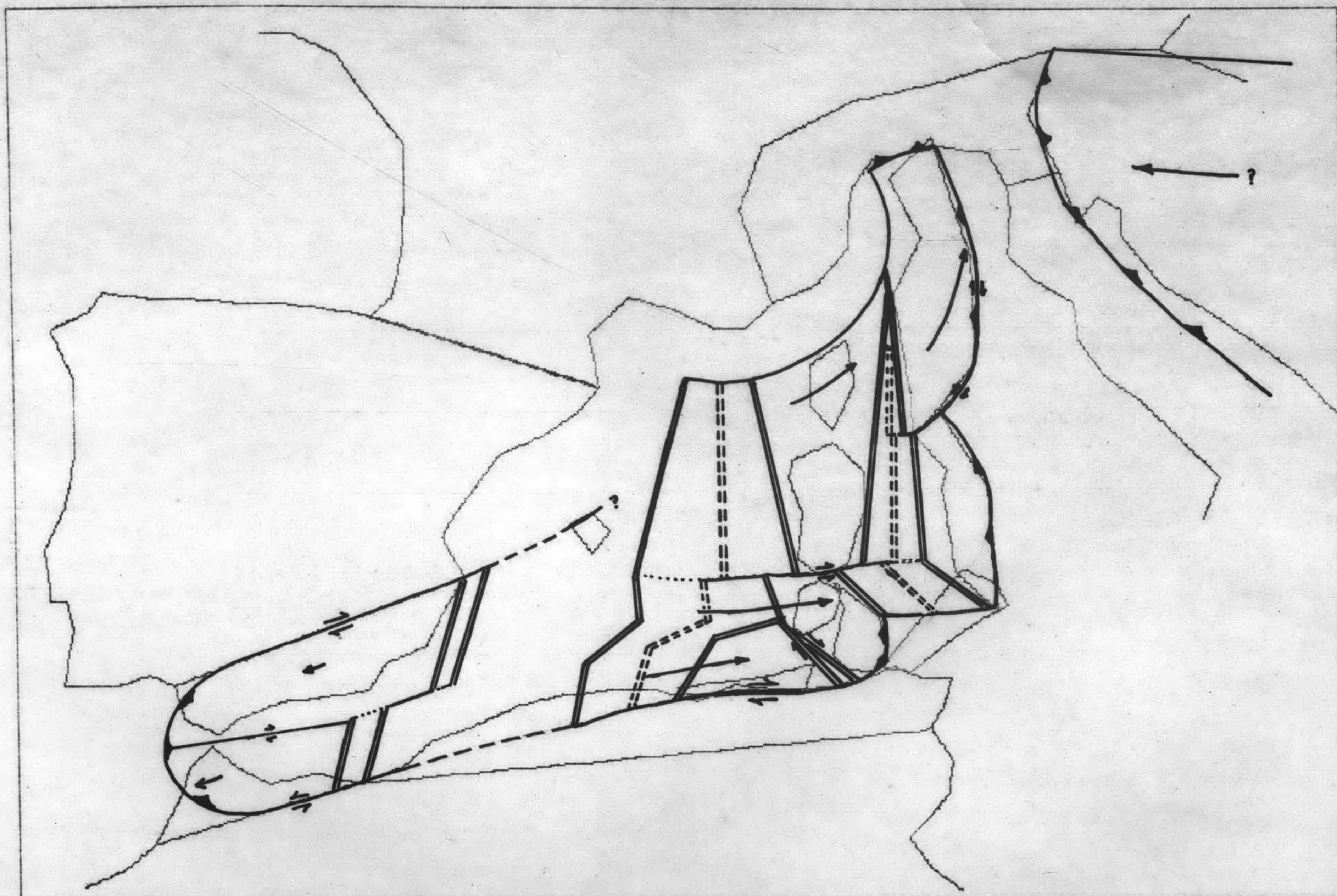


FIGURE 6   LATE OLIGOCENE WESTERN TETHYS (-30M)

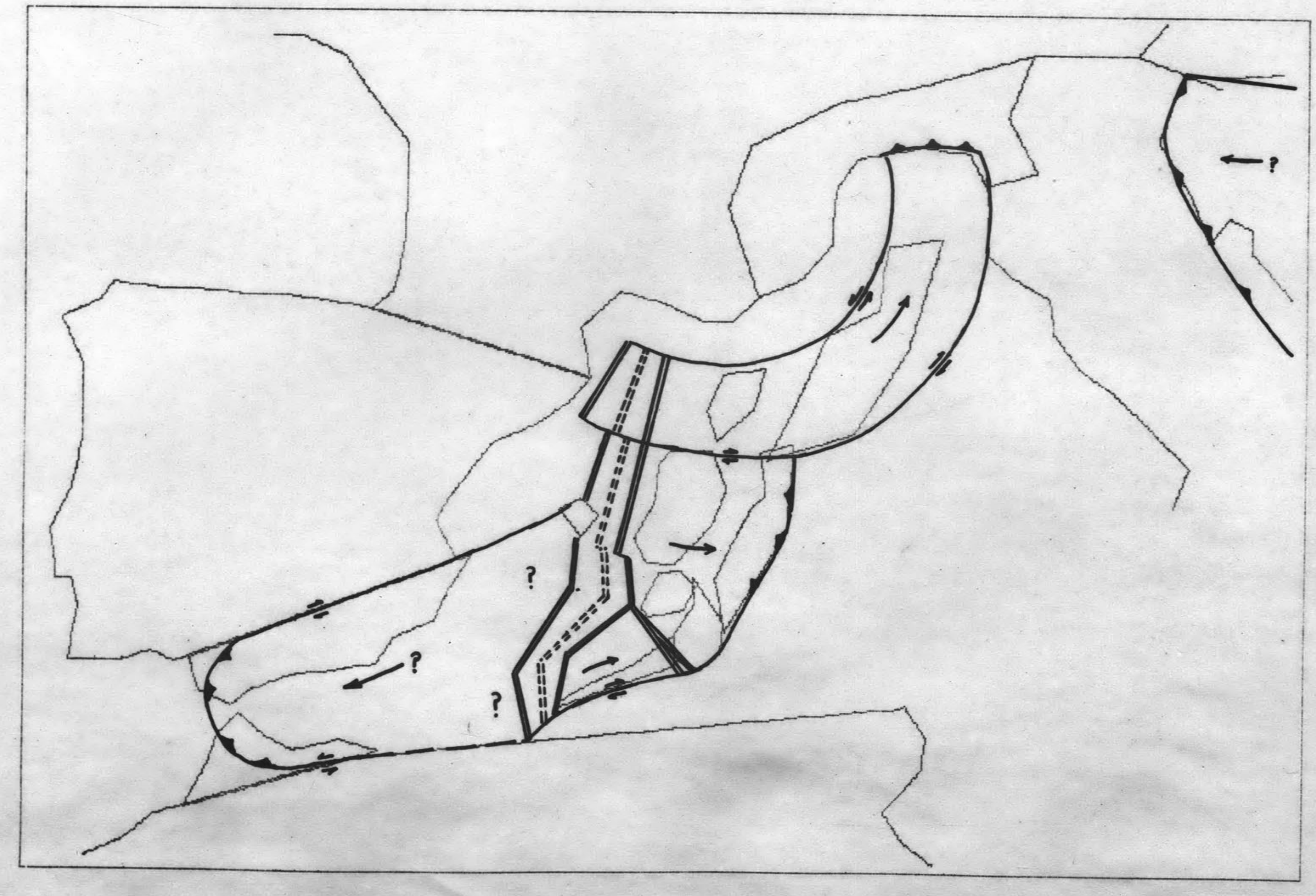


FIGURE 7 LATE EOCENE WESTERN TETHYS (-45M)

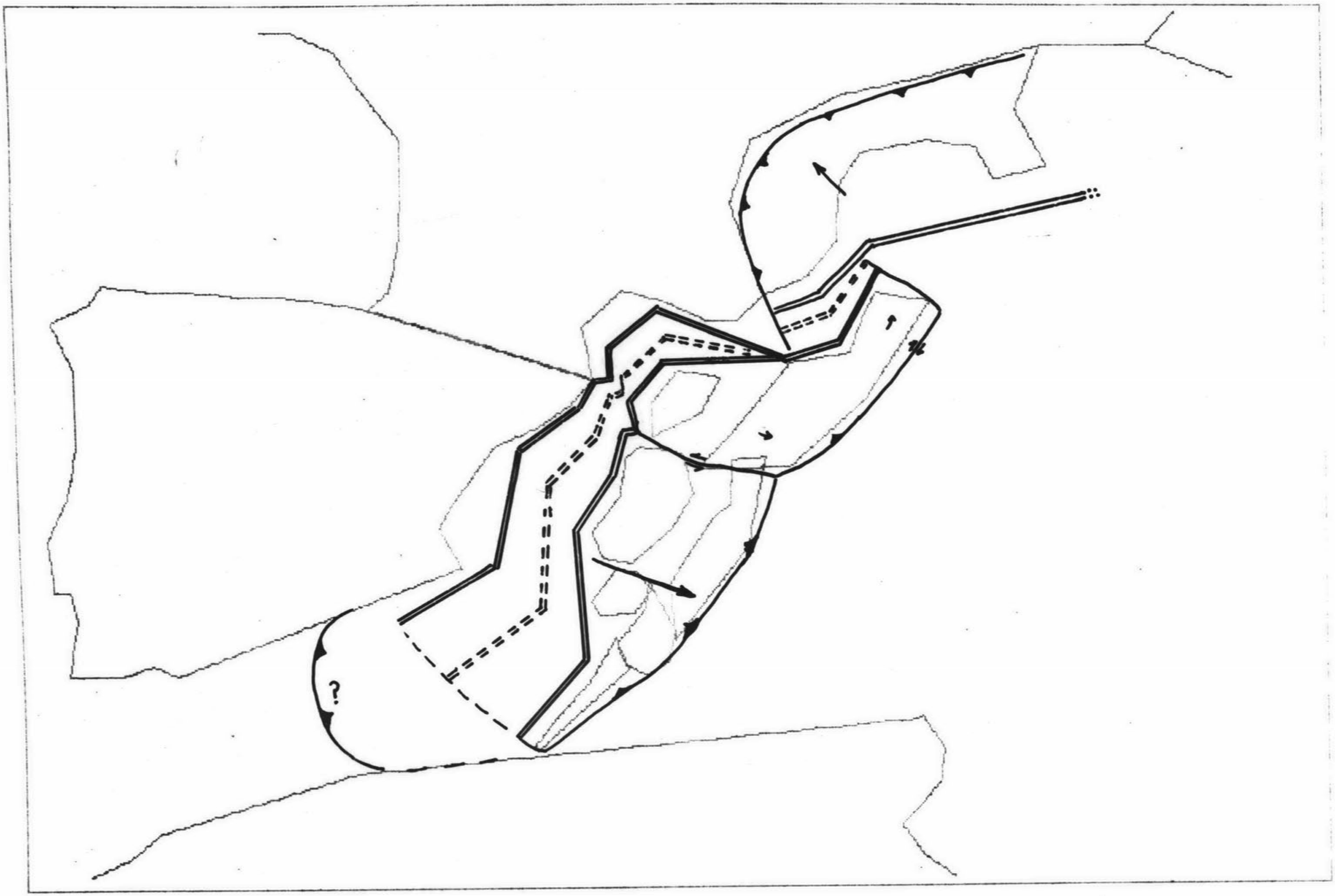


FIGURE 8   EOCENE-PALEOCENE WESTERN TETHYS (~60M)

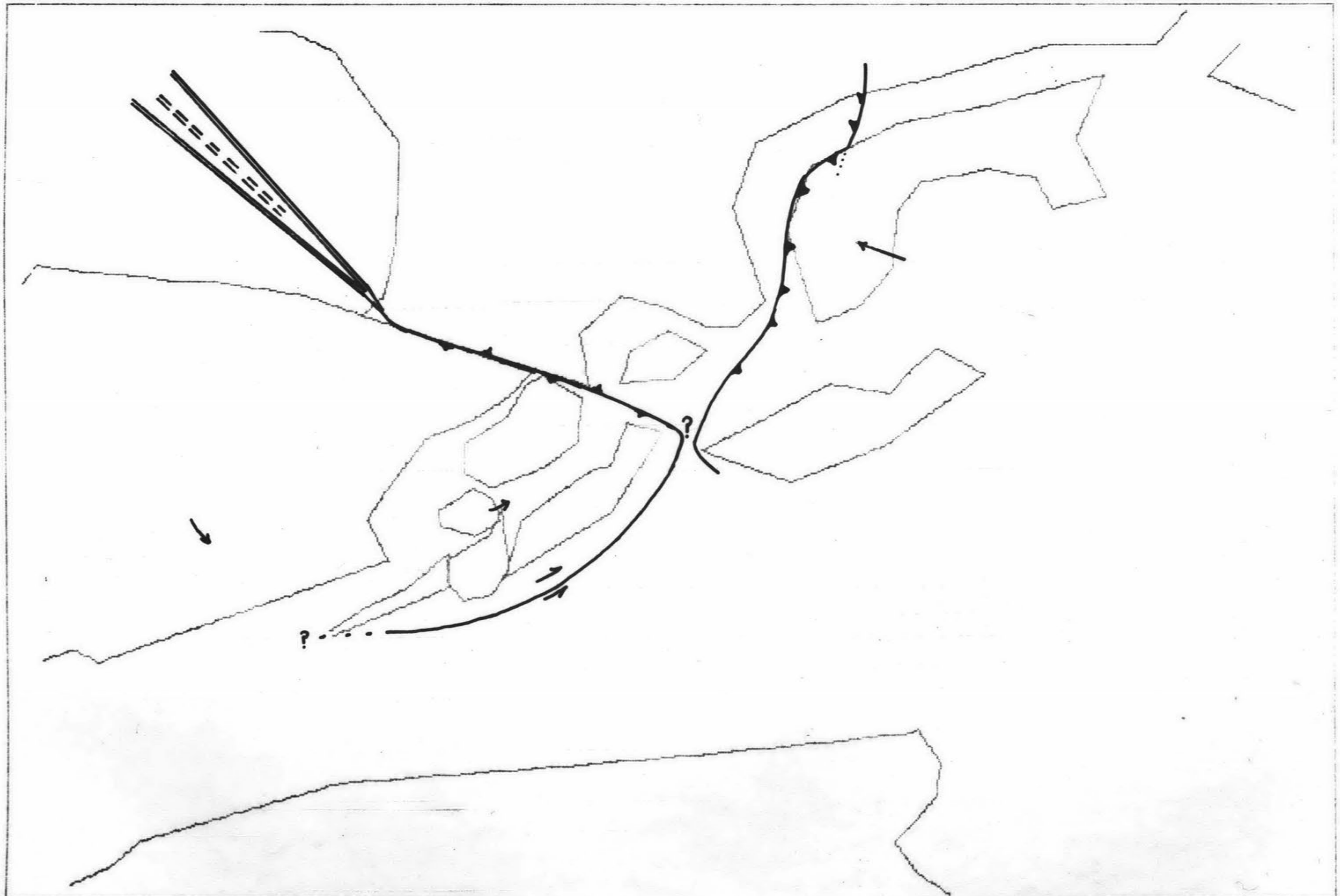


FIGURE 9   LATEST CRETACEOUS WESTERN TETHYS (-75M)

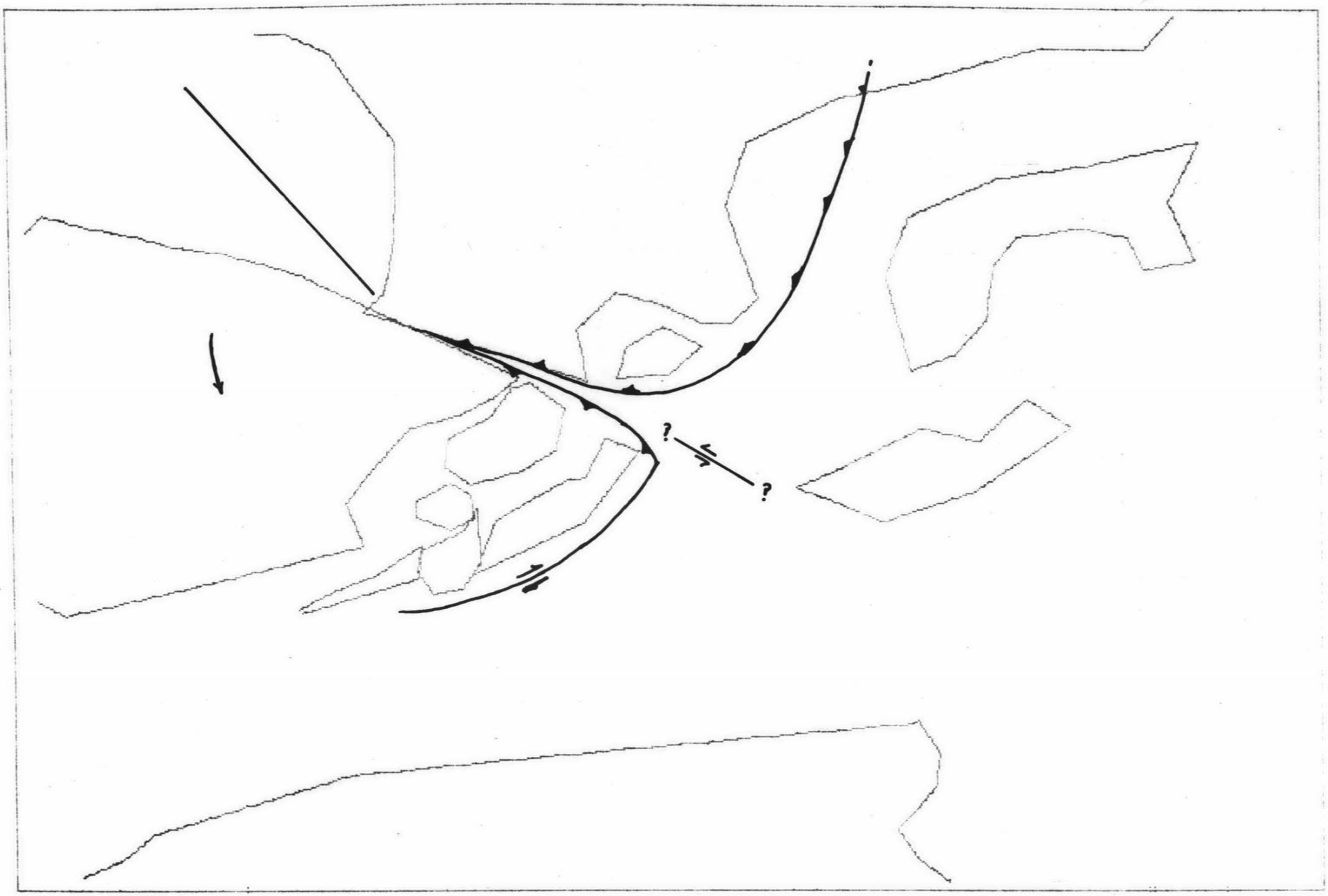


FIGURE 10 MID PLIOCENE WESTERN TETHYS (-5M) ALT 1

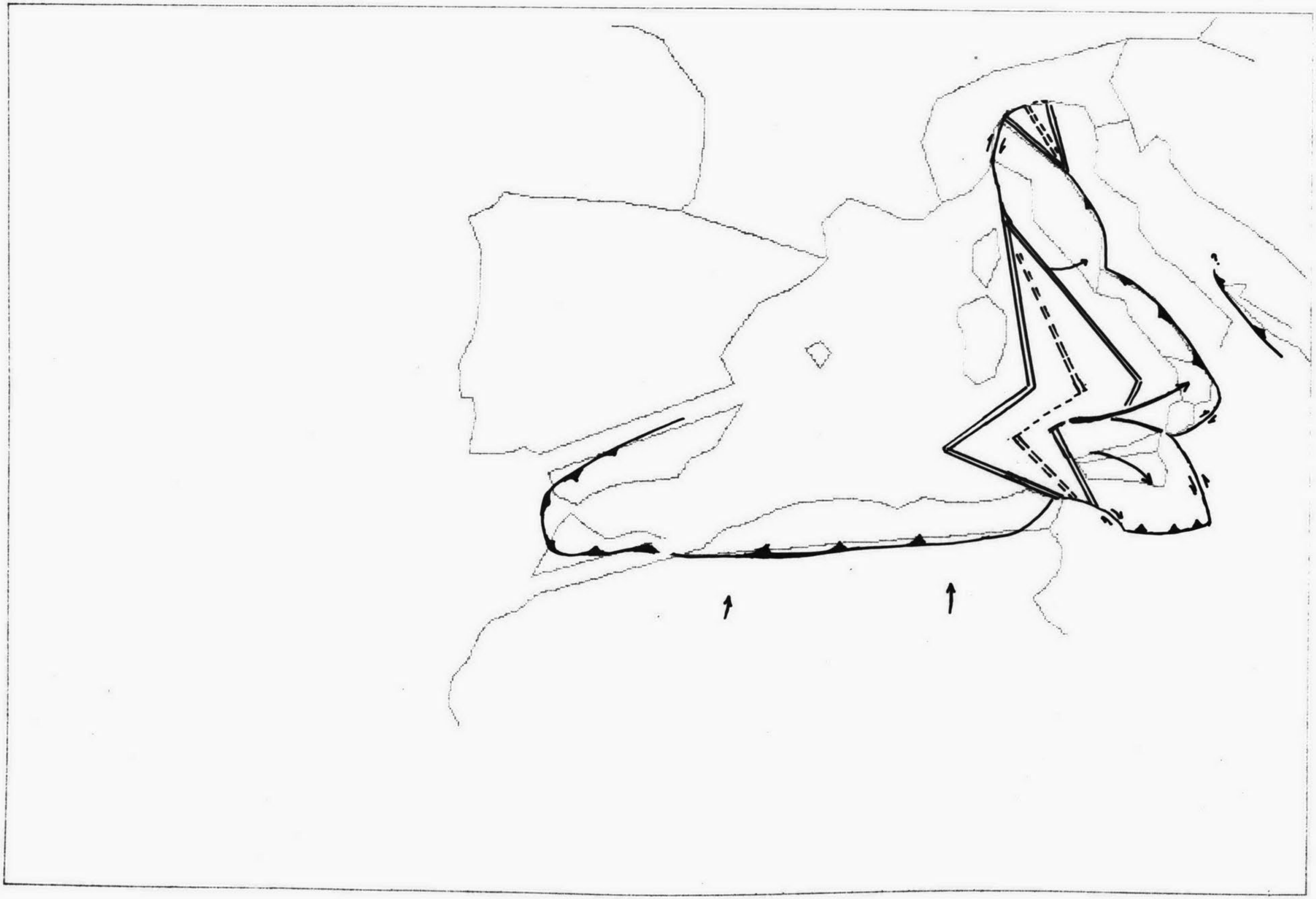


FIGURE 11 LATE MIOCENE WESTERN TETHYS (-15M) ALT 1

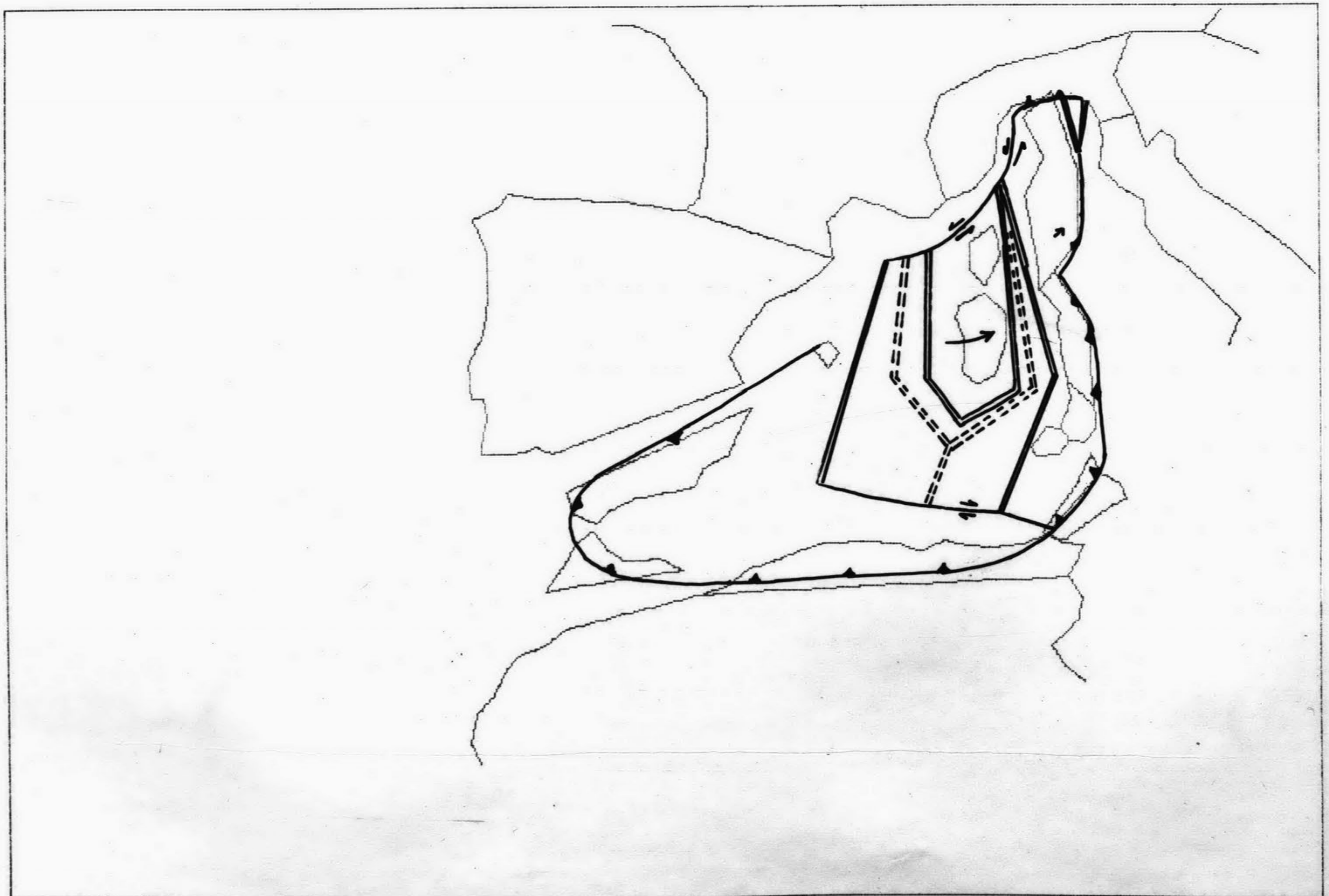


FIGURE 12 LATE OLIGOCENE WESTERN TETHYS (-30M) ALT 1

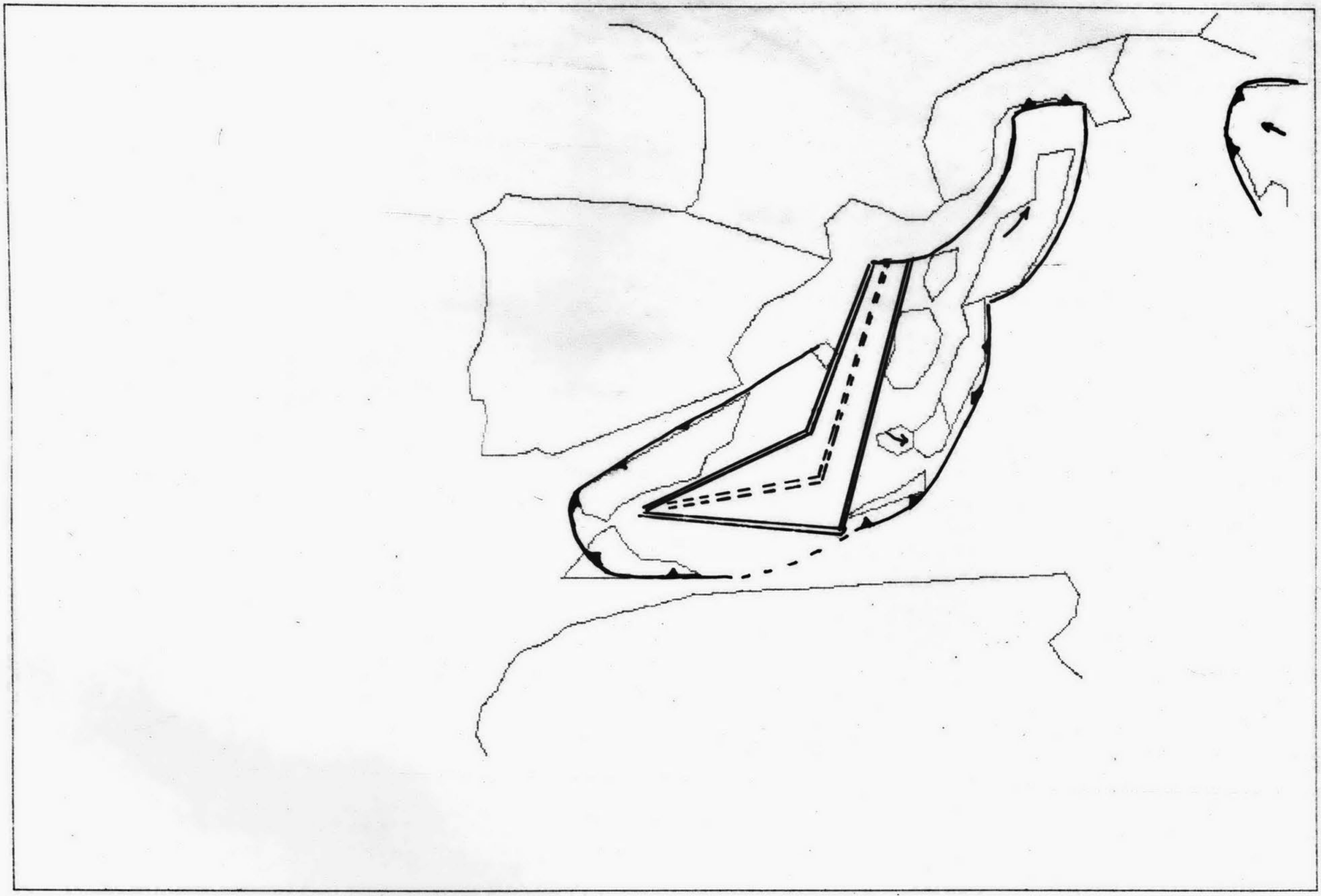


FIGURE 13   LATE EOCENE WESTERN TETHYS (-45M) ALT 2

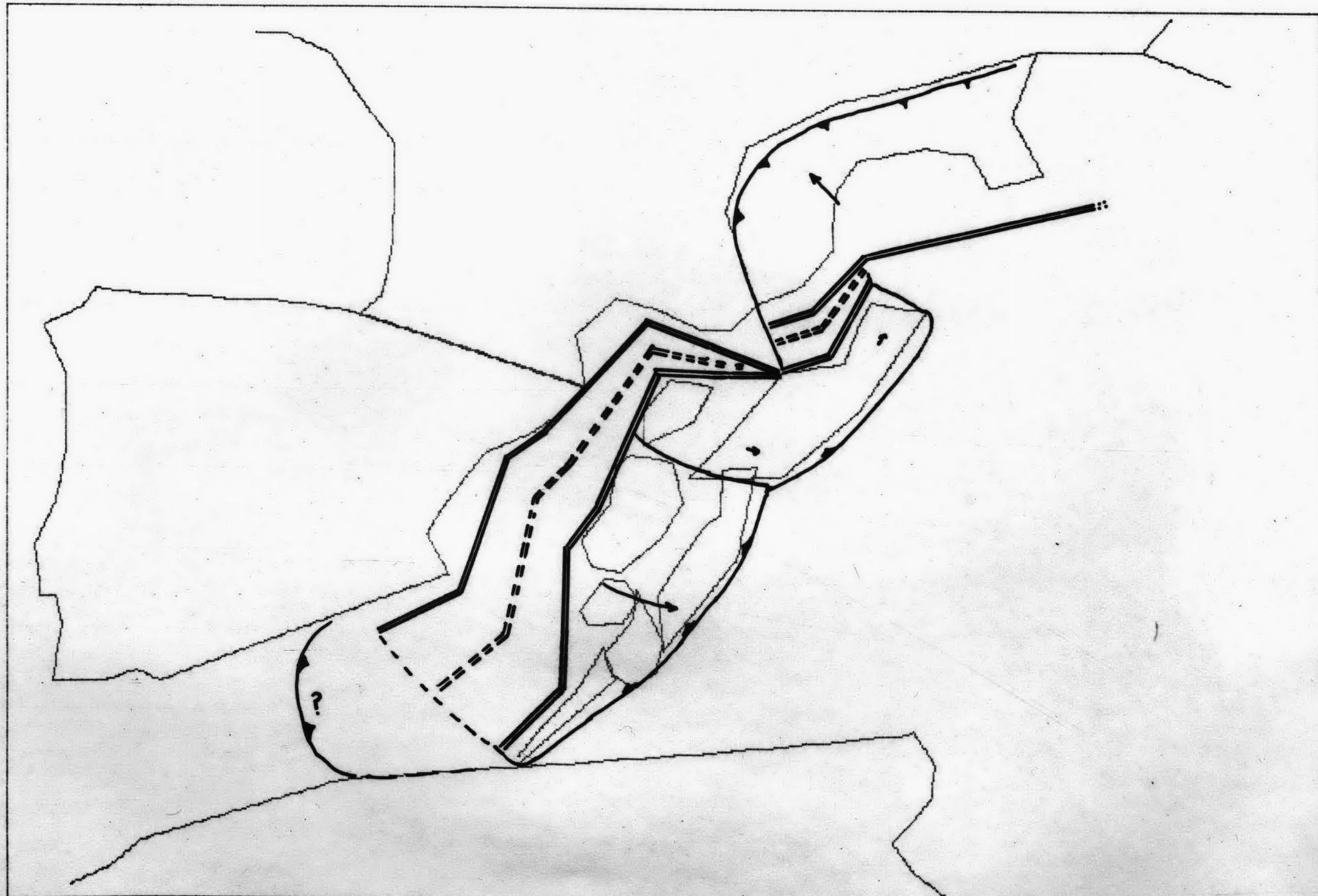
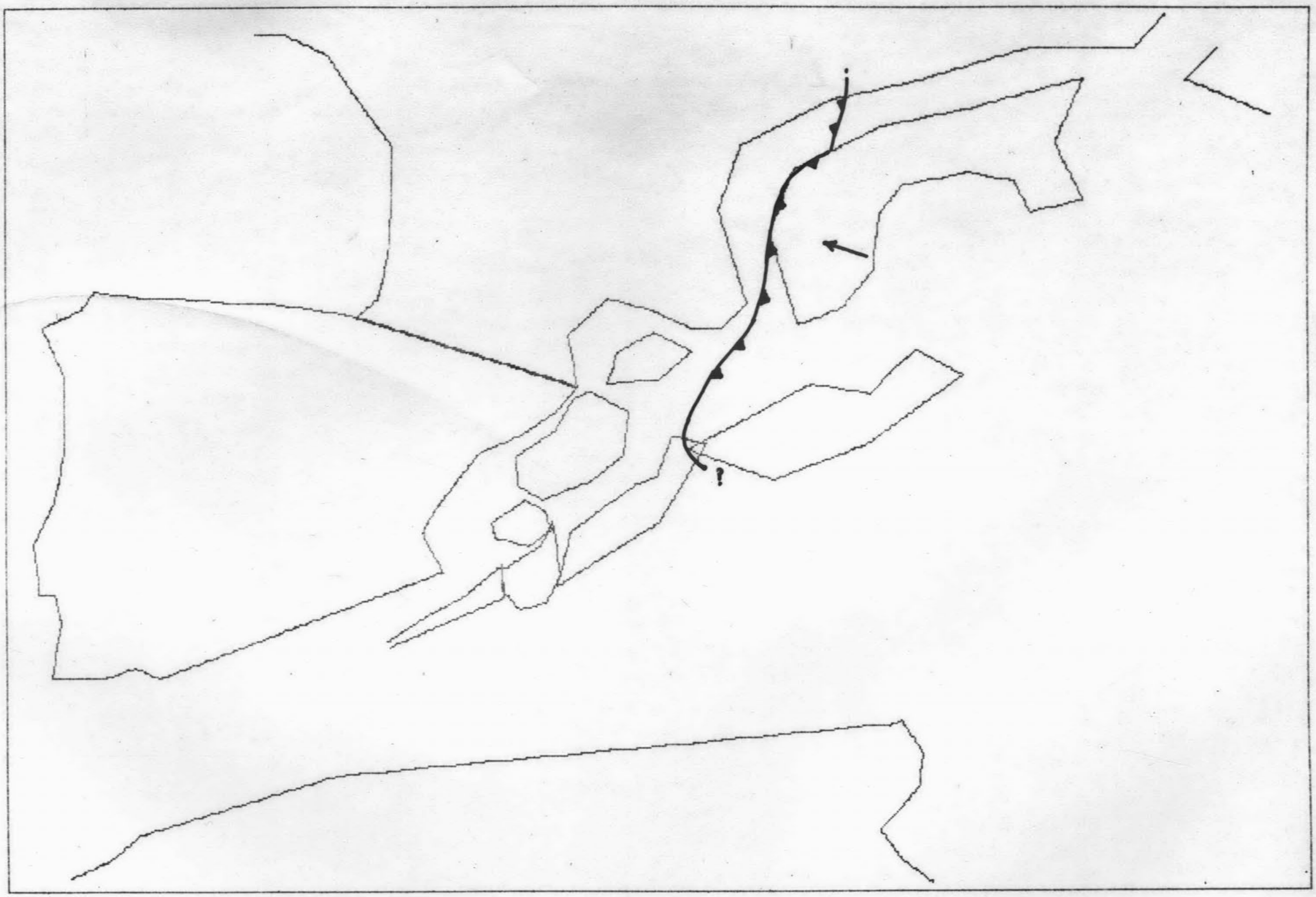


FIGURE 14 EOCENE-PALEOCENE WESTERN TETHYS (-60M) ALT 2



THE CALABRO-SICILIAN ARC

The Western Mediterranean Sea is surrounded by a continuous chain of mountains, from the Balearic Islands, along southern Spain as the Betic Cordillera, turning across the Straits of Gibraltar to become the Riff Mountains, along northern Africa as the Atlas Mountains, across northern Sicily, and up the Italian peninsula as the Apennines, to end near Genoa. The chain is characterized by thrusting away from the Western Mediterranean basins, and its outer thrust sheets in many places are composed of the Numidian Flysch, deposited in the Tethyan continental rise during the Oligocene - early Miocene (Wezel, 1970). The southeastern corner of this chain, in Sicily and Calabria, is significant in our reconstruction of the Miocene Mediterranean.

The tip of Tunisia, northeastern Sicily and Calabria have been described as a single mountain arc (Caire, 1962, 1970; Auzende, 1971). Three nappes in this chain, with the oldest on top, overthrust the Apennines to the north northwest before the Middle Oligocene, resulting in glaucophane or blueschist metamorphism (Dubois, 1970). The Numidian Flysch was in turn overthrust by the Reitano Flysch during the Upper Tortonian orogeny (15 myrbp) in northern Sicily and southern Calabria, and both series were deformed about the Middle Pliocene (Wezel, 1970). The nappe structure and the blueschist metamorphism involved in thrusts of the continental rise material indicate consuming plate activity, at least on three occasions, about 35, 15 and 5 myrbp.

The Calabro-Sicilian Arc has been identified as a modern, active consuming plate margin (Caputo et al., 1970), and the Tyrrhenian Sea taken to be the associated subsiding behind-arc basin (Erickson, 1970). The evidence for the existence of this plate boundary includes seismicity, heat flow measurements, seismic refraction and reflection surveys, airborne and marine magnetic surveys, volcanic petrology, dredge hauls and other studies. We shall here briefly examine the available data in order to determine the motion of the associated plates.

Peterschmitt (1956) noted that a Benioff Zone dips west under Sicily. Caputo et al. (1970) found that the epicenters of seismic shocks lie on a curved surface which dips 58° NNW and 60° WNW from Mt. Etna to Naples, and chose the location of the plate boundary so as to pass through Mt. Vesuvius, Lipari Island, Mt. Etna, Linosa Island, Pantellaria Island and on to Tunisia. As these volcanic islands must lie above the descending slab, the surface intercept of the plate boundary is actually outside of their line. The depths of 18 epicenters are plotted in figure 15, and hand contoured at the 250, 260, 300 and 310 km isobaths, although errors in the depths given may be as large as the extremes of the contours. The locations of active volcanoes are marked by asterisks, the dash and dot line is the trace of a narrow submarine canyon extending WNW from the Gulf of Euphemia on the map of Ryan et al. (1965), and the strike-slip fault trace through Catanzaro marks the northern boundary of the

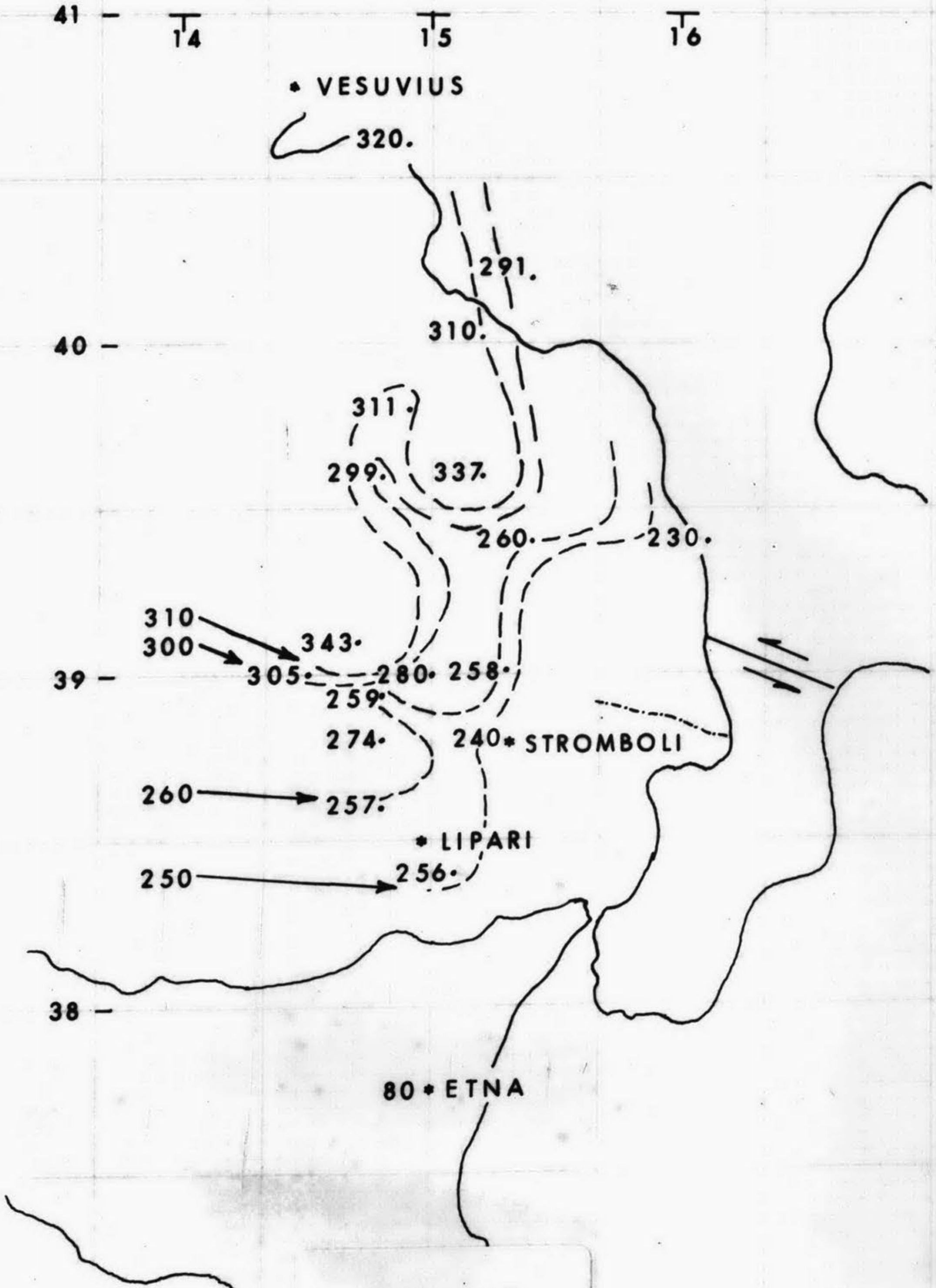


figure 15

1962 - 1969 DATA FROM CAPUTO et al

* VOLCANO · DEPTH in KM

SEISMOBATH PLOT - TYRRHENIAN SEA

mountain chain described by Caire (1962).

Lacking evidence of other structural trends, the WNW trend of the lineaments shown for Calabria may be a surface expression of the complexities of the seismic surface shown by the recurvature of the contours, which seem congruent with the offsets of the Calabrian coast. If the modern motion of Sicily is east southeast, any eastern trench is buried under the sediments derived from the arc and the nearby continental sources. To the south and southeast spot oceanic gravity measurements and land geological evidence suggest that a buried trough extends from the island of Pantellaria to southern Malta, where Oligocene-Miocene rocks are downfaulted to the northwest (Harrison, 1955). If this trough is taken as the trench of the arc, disturbance of some sort may have taken place further south. Marine refraction shots indicate more than 1 km depth of sediments from Tunisia to Medina Bank versus 1/3 km elsewhere (Gaskell & Swallow, 1953). Eastward from Medina Bank, aeromagnetic surveys detected a positive 500 gamma anomaly spur along 35° N (Vogt & Higgs, 1968) which may mark the leading edge of deformation related to subduction of a new slice of the African Plate, through uplift of the leading edge of the slice as it tilts into the presumed trench, thus bringing the magnetic sea floor closer to the surface.

Behind the Calabro Sicilian arc, four measurements of heat flow in the southeastern Tyrrhenian Sea average $3.38 \pm .16$ heat flow units (HFU). The Western Mediterranean averages 2.83 $\pm$ 1.0 HFU over 12 points, while the Eastern Meditarranean averages only 0.74 $\pm$ .3 HFU at 33 points (Erickson, 1970). This may be explained as successive stages of spreading behind the Calabro-Sicilian arc, with the most recent episode occurring in the southeast Tyrrhenian Sea. The 300 - 400 m depth of sediment found by reflection profiles in the Tyrrhenian (Ludwig et al., 1965) indicates Neogene formation of sediment traps which have since filled with turbidites, volcanic ashfall, and pelagic sedimentation (Ryan et al., 1965; Zarudski et al., 1971). Heezen et al. (1971) dredged schist, phyllite and marble from a linear ridge trending N 15 E near 40° 15' N, 12° 15' E against which turbidite sequences thin, indicating that foundering of continental material has taken place and that subsidence continues. We take this ridge to be a fragment of the Calabro-Sicilian Arc, possibly separated along an old subduction surface from the accumulation of volcanicized oceanic slices described in our consuming boundary model, and take its strike to be roughly perpendicular to arc motion during separation.

Although aeromagnetic surveys over the Mediterranean show no parallel linear anomalies typically associated with spreading mid-ocean ridges (Vogt & Higgs, 1968), marine magnetic surveys found a 700 gamma north-south linear trend along approximately 10° 20' E, to the east of Corsica, similar to the 1000 gamma anomaly over the spreading margin northwest of the Aegean Plate (Taylor & Rankin, 1971). One exception to the generally smooth aeromagnetic pattern is a group of short wavelength anomalies roughly following the orientation of the coasts of Sardinia, Corsica, Calabria, and the islands of Vulcano and Stromboli (Vogt & Higgs, 1968), taken to indicate spreading behind the arc (Vogt et al., 1971). This spreading must have been responsible for the subsidence of the Early Pliocene (7 myrbp) erosion surface which was penetrated by JOIDES hole 132, located east of Sardinia (Ryan et al., 1970).

The outer boundary of the Numidian Flysch thrust sheet described by Wezel (1970) is discontinuous from Mt. Etna to the Gulf of Taranto, and has a strike in Sicily about 58° CW from its strike in Tunisia. We take the discontinuity and the change in strike to be a result of the motion of the Calabro-Sicilian arc, rather than the relative motion of Africa and Europe, as the same boundary is not significantly offset near the Straits of Gibraltar. In order to connect this feature around the Late Miocene Tyrrhenian Sea, we rotate Sicily 58° CCW backwards from its present location (figure 3) to obtain its Late Miocene location (figure 5). We assume that the motion is steady for its duration, so that one third of this backward rotation is shown for the Middle Pliocene (figure 4). The pole of rotation in the coordinates of Europe, taken to be stable, is chosen to be 35.7 N, 12.0 E, consistent with the spreading and consuming boundaries discussed above.

Southern Calabria is included with Sicily in the Pliocene rotation, shearing along the boundary given by Caire (1962, 1970), and central Calabria is moved inside northern Calabria in the Early Pliocene to the Late Miocene, so that the metamorphic and granitic rocks of the Calabrian Mountains (Heezen et al., 1971) are located inside the Numidian Flysch boundary before the Late Miocene orogeny in Calabria. Pliocene and earlier motions have also occurred along the Italian Peninsula and within the Western Mediterranean Sea, and will be described next; a counterclockwise rotation about a pole near Genoa moves the peninsula away from the Calabro-Sicilian arc allowing the motion described in this section.

In effect, our model shows the Pliocene Calabro-Sicilian arc overtaking, distorting, and renewing activity on a portion of the Miocene island arc which was responsible for the main Alpine chains around the western Mediterranean Sea. The Calabro-Sicilian arc rotated Sicily away from the Tyrrhenian Sea, leading to a modern volcanic arc with subsidence and sedimentation in its wake, separating the previously continuous structures on either side of it. In particular, we conclude that the arc length of the Miocene consuming boundary has been increased by the collision with the Pliocene arc, so we may proceed with rotation of the Miocene arc shortened by subtraction of the Pliocene segment. An alternative to this complicated sequence of events is also presented in the next section.

THE APPENNINE ARC

The Apennine mountain chain extends along the Italian Peninsula from Calabria to the north of Genoa. The southern part of the chain contains a continental rise deposit, the Numidian Flysch, thrust to the northeast (Wezel, 1970) over the Bradano Trough, which is an Eocene graben with mostly Lower Pliocene fill, located between Calabria and Apulia (mongelli & Richetti, 1970). The northern Apennine Mountains were formed by a sequence of flysch deposition, nappes thrust northeastward, normal faulting, and volcanism, which progressed from northwest to southeast. The eugeosynclinal components were thrust over the miogeosynclinal components of the chain near the time of the Oligocene-Miocene boundary (Abbatte et al., 1970; table 2, appendix 1). Boccaletti & Guazzone (1970) demonstrate that the Tuscany Basins of the northernmost Apennines could have been thrust northeastward from a continental shelf and rotated counterclockwise to their present position by an island arc moving to the east and the north in the Tyrrhenian Sea, with a subduction zone dipping to the west. From the continuity of the Numidian Flysch thrust boundary, we take the rest of the Apennines to have been formed by the same mechanism.

Geological and geophysical studies suggest that subsidence of the Tyrrhenian Sea has occurred in the Late Miocene (Ryan et al., 1965) and the Middle Pliocene (Zarudski et al., 1971), which we take to indicate spreading behind the Apennine arc, so that one plate fragment in the Western Mediterranean Sea extends from the easternmost thrust boundary of the Apennines to some spreading line in the Tyrrhenian Sea, figures 3 through 5. The total area of new ocean floor, shown bounded by double lines, is implied by the motion of the plate fragments, but the location of the spreading boundary is ambiguous, since the process by which new ocean floor is created behind a consuming boundary is poorly understood. The north-south magnetic anomaly in the Tyrrhenian Sea (Taylor & Rankin, 1970) was taken as a persistent spreading center, figures 3 and 4, but the remaining spreading boundaries were simply placed where convenient. For the Pliocene, spreading is indicated to the north of the northernmost Apennine Mountains by the post-orogenic subsidence of the Po Basin (Rutten, 1969, p. 301).

In our representation of the Late Miocene, figure 5, we have arranged the western Italian peninsula and southern Sicily so that the Numidian Flysch thrust boundary is one continuous feature across the Mediterranean, and as can be seen from the several small overlaps, the fit is too tight. The shorelines which are displayed take up less space than, say, the 500 fm isobath as used by Bullard et al. (1965), so that the fit is worse; if the weakly constrained motions of the Calabrian fragments are in error, it is likely that even more room is required. We conclude that the configuration portrayed in figure 5 represents a minimum distance between Africa and Europe, and we would prefer Africa moved south instead of held fixed. However, measurements of paleomagnetic pole locations from the Turkana lavas of northwest Kenya indicate that no change has occurred from between the Late Oligocene and the Middle Miocene until the present (Raja, Reilly & Mussett, 1968), so the African plate cannot be moved much from its position as indicated, at least in latitude or orientation. Since paleomagnetic measurements do not fix longitude, some change in the fit might be sought in strike-slip motion along the Azores-Gibraltar ridge; however, the Riff and Betic mountain chains are not offset across the Straits of Gibraltar by more than 50 km. We conclude that if the positions of Africa and Europe shown are correct, some post-Miocene increase in arc-length has not been accounted for.

Since paleomagnetic pole positions provide weak constraints, we shall attempt to extrapolate evidence for the modern motion of Africa relative to Europe. McKenzie (1970) has examined the seismicity of the Mediterranean region, and describes the inferred motion along the Azores-Gibraltar boundary as strike-slip with a small amount of extension near the Azores, changing to overthrust to the south of Spain and in North Africa, consistent with Le Pinchon's pole at 9 N, 46 W. In figures 10, 11, and 12, Africa is moved 1.5^{o} counterclockwise with respect to Europe between the present and the Mid Pliocene, and the same again between the Mid Pliocene and the Late Miocene. The Riff-Betic arc, described in the following section, is rotated by half that amount, and opened slightly to indicate some change in shape resulting from compression between Europe and Africa. In order to illustrate the looseness of this fit, the Calabrian fragments are not moved behind the Tyrrhenian arc, although the Numidian Flysch boundary is assumed continuous from Tunisia through the Apennines as before, and the remaining fragments are moved as in figures 3 through 6. It is clear that once the fragments composing the Numidian Flysch thrust boundary are moved past a north-south axis in the Late Miocene, figure 11, there is neither requirement for nor constraint upon further separation between Europe and Africa; 3^{o} is sufficient opening, and would be hard to detect by the usual paleomagnetic methods.

In the basins to the west of Corsica and Sardinia, a 1 km thick layer of Upper Miocene salt has been identified from explosive reflection profiling ("Flexotir"), and is thought to have subsided in the Pliocene (Auzende et al., 1971). The presence and age of the salt has been confirmed in three JOIDES bore holes on leg 13 of the Glomar Challenger (Ryan et al., 1970). Since we relate subsidence to the spreading behind the Apennine arc, the islands of Corsica and Sardinia are moved slightly eastward during the Pliocene, figure 3 and 4, although no spreading margins are indicated in their wake, as the amount of spreading was slight and the location of the spreading margin is unknown and probably diffuse. A much more significant Miocene spreading west of Corsica and Sardinia is indicated in figure 5, in order to create the isolated basin in which the salt was deposited.

Off Esterel, southern France, the narrow continental shelf and a continental slope dipping as steeply as 10 to 20 degrees seaward, cut by submarine canyons following fault zones (Pautot, 1970) suggest a rapid subsidence of the Ligurian Sea. In the French and Italian Maritime Alps, the Annot sandstone, a flysch of Late Eocene to Early Oligocene age, contains consistent evidence of paleocurrents from the south (Kuenen et al., 1959). Seismic refraction profiles shot by Fahlquist (1953) recorded a compressional wave velocity of only 7.7 km below the Moho sec at two of four places where mantle arrivals were thought to have occurred in the northern part of the Western Mediterranean Sea. The morphological similarity among Corsica, Sardinia and Italy, the low mantle velocity under the Ligurian Sea, paleomagnetic evidence for the rotation of Corsica, and the continuing eastward dislocation of the source area for Eocene and Miocene Alpine sediments led Stanley and Mutti (1968) to support the idea of an emerged land mass in the Ligurian Sea during that time. The Eocene through Miocene representations, figures 5, 6, and 7, show how the plate fragments behind the Apennine arc may have satisfied these conditions, as Corsica, Sardinia, and western Italy moved to the northeast around Esterel. In figures 5 and 6, the transform fault boundary between these plate fragments and Europe is shown close to the northwest margin of Corsica, but this shear boundary may have been active over a broad region, or closer to Esterel.

The Miocene and Pliocene motion of Corsica and Sardinia is taken to be roughly due east, and rotation about a nearby pole is allowed before the Oligocene, in order to agree with the paleomagnetic pole locations found for these two islands. In Corsica, west of the north-south trending westward thrust boundary, Stephanian andesite-rhyolite volcanics over granite near the town of Osani have a paleomagnetic pole which is consistent with a rotation of $53^{\circ} \pm 14^{\circ}$ CCW about a pole located near Genoa (Ashworth & Nairn, 1965). Possible support for this rotation was given by investigation of the Ota gabbro-diorite, but the age of the unit is not well known (Nairn & Westphal, 1967). Oligocene to Early Miocene phenotrachyandesite ignimbrites near Alghero on Sardinia yielded a paleomagnetic pole at 54 N, 95 W (De Jong, Manzoni & Zijderveld, 1969), and the magnetization of Permian rhyolitic ignimbrites and red sandstones had an inclination comparable to that found for Africa, while the declination required a rotation of 50° CCW with respect to Europe (Zijderveld et al., 1970). Lower limits on the amount of rotation and the youngest limits on the date at which rotation was complete have been given by Van der Voo and Zijderveld (1969); Corsica must have turned at least 21° CCW and Sardinia 40° CCW, ending around the Miocene-Oligocene. These limits are consistent with the rotations shown in figures 3 through 9, which total about 65° for Corsica and about 58° for Sardinia; no rotation of the magnetic pole occurs after the Late Miocene, although motion due east does occur.

In the earliest Tertiary, figure 8, Corsica is fitted into the Gulf of Lyons, while Sardinia is located behind the Balearic Islands in the Gulf of Valencia, with the southern fragments of the Apennine arc and Calabro-Sicilian arc packed around it. The location of these two islands is dictated by the similar pre-Tertiary sediment patterns on the islands and their respective mainlands, shown on the pre-Tertiary isopachous-lithofacies maps of Klemme (1958). The location of Sardinia in the Gulf of Valencia overlaps the Balearic Islands of Majorca and Minorca. These islands are part of another arc, which will be described next.

THE RIFF-BETIC ARC

The westernmost arm of the Mediterranean Sea is the Alboran Sea, bordered to the north from Gibraltar to the island of Majorca by the Betic Cordillera of southern Spain. This mountain range contains individual sierras with Oligocene-Miocene nappes thrust north and west, separated by discontinuous Neogene basins (Rutten, 1969, p. 392 et seq.). To the south of the Alboran Sea, the Riff or Rif Mountains of Morocco are composed of nappes which were thrust south and west toward the Hercynian Atlas Mountains, which have been newly folded along an ENE-WSW trend after the Eocene and along an E-W trend after the Miocene (de Sitter, 1956), p. 258). Both the Betic and Riff mountain chains contain the Numidian Flysch, thrust away from the Alboran Sea after the Early Miocene (Wezel, 1970). Although the question of the continuity of the Riff and Betic chains across the Straits of Gibraltar was raised in the earlier literature, Rutten (1969, p. 407) has stated that "Today, all authors seem to consider the Rif to be the continuation of the Betic Cordillera."

To the west of the Riff-Betic arc, Bouguer gravity anomalies reach -50 mg (F. Lamy, personal communication, 1970) and extensive accumulations of modern sediments in the Rharb and Quadalquivir river plains indicate subsidence of the Moroccan and Iberian mesetas in front of the arc. In Portugal, a reversed refraction profile was shot along a line 250 km in length from Sines in the northwest to Fuzeta in the southeast. Mantle velocity (8.15 km/sec) was reached at a depth of 30 km under Sines, 36 km under Fuzeta (Mueller et al., 1971), so that the Moho dips toward the Alboran at least to the north of the Straits of Gibraltar consistent with depression in the front of the Riff-Betic arc. Within the arc, JOIDES hole 121 penetrated oceanic basalt under Late Miocene sediments (Ryan et al., 1970), so that the Alboran Sea must have been formed before that time.

Although the width of the Riff-Betic arc is less than 350 km, the presence of continental rise deposits in nappes thrust away from the Alboran Sea, and the evidence for subsidence in front of the arc suggest that this structure is the result of a Miocene island arc moving to the west with a subduction zone dipping east and colliding with the continental margins of Africa and the Iberian Peninsula, figures 5 and 6. Spreading behind this arc resulted in the formation of the Alboran Sea. The Riff-Betic arc was probably continuous with the Apennine arc along some southern boundary now represented by the Tellian Atlas, since the Numidian Flysch thrust boundary retains the same sense of thrusting away from the Mediterranean Sea along the north coast of Africa, although we have no detailed description of the orogeny of the Tellian Atlas itself.

The fragment of northern Sicily which we assume was connected to the Calabro-Sicilian arc cannot have been moved backward beyond the Miocene without overlapping some part of the boundary of the Riff-Betic-Apennine arc system unless it is rotated counterclockwise relative to the Apennine plate fragment. Since northern Sicily contains crystalline Paleozoic rocks, we place it to the south of Spain in figures 7, 8, and 9, inside of any plate boundary around the Iberian Peninsula. The position which would require the least movement not accountable to some known boundary is parallel to southern Spain and very close to the Calabrian fragments to the southwest of the Gulf of Valencia, figures 8 and 9. However, northern Sicily must then have moved away from Spain before the end of the Oligocene, so that the Riff-Betic arc had room to form Majorca in its wake; some support for such an idea is given by the pre-Middle Oligocene date for overthrusting in Sicily mentioned in our initial description of the Calabro-Sicilian arc. If this is the case, the correct position for the northern Sicily fragment was along the thrust boundary approaching Algeria, figure 6. Earlier positions would then lie along extensions of the consuming plate boundary bordering the Apennine arc in figure 7, and along the southern shore of Spain in figures 8 and 9. The Riff-Betic arc was then continuous with Majorca in figure 6, as required by geological evidence, and moved southwest from the gap between Spain and the Calabrian fragments in figure 7. The movement of the Riff-Betic arc has obliterated evidence of any earlier arc which bordered southern Spain.

THE ALPINE ARC

The isopachous-lithofacies maps of Klemme (1958) clearly indicate that the most significant regions of geosynclinal sedimentation around the Mediterranean Sea are located to the north of Italy in the Alps. As described by Rutten (1969), the Western Alps (table 4, appendix 1) turn from a NW-SE trend north of the Ligurian Sea, to strike north in southern France, northeast as they pass through Switzerland, and finally east in Austria. To the south, the Dinaride mountain chain strikes NW-SE along the east coast of the Adriatic Sea, and is cut off by the Eastern Alps, rather than joining them in a smooth curve. The Carpathian mountains extend to the east from north of the Eastern Alps and form a broad reversed S curve to the west of the Black Sea, with the Transylvanian Alps marking the center section of the S. Within the Swiss Alps a subparallel sequence strikes WSW-ENE and consists from north to south of the miogeosynclincal Helvetic Nappes, the uplifted Central Massifs, the eugeosynclinal Penninic Nappes, the vertically tectonized Root Zone, and the Southern Alps.

Hsü & Schlanger (1971) have interpreted the sedimentary and igneous geology of the region in terms of plate tectonics and speculate that a Late Cretaceous and Paleocene arc bordered Europe with a subduction zone dipping north; during the Eocene, this Habkern arc attempted to consume a Penninic plate approaching from the south, but was overridden instead, so that the subduction zone changed sense and dipped to the south, thrusting the shelf and basin deposits which lay behind the Paleocene arc back onto the European shore until the Oligocene. It is accepted that a consuming boundary must have existed within the Tethys near the beginning of the Tertiary, in order to permit the rotation of Africa so that its northern boundary moved toward Europe (Dietz & Holden, 1970), but the consumption of the Habkern arc by the Penninic plate as proposed above requires a complex history for the subduction zone. A simpler solution would consist of an island arc in the Tethys with subduction zone always dipping south until it collided with Europe, so that the Helvetic Nappes once comprised the European continental rise, and the Penninic Nappes were the island arc, following the model employed earlier in this exercise. Hsü & Schlanger depend upon the reconstructed southeastward convexity of the European shore as indicative of the initial northwestward dip of the subduction zone, and upon the modern absence of the Habkern arc to indicate its consumption by the Penninic plate. The difference between their model and the simpler alternative suggested is not great, since the result in either case is to close the Tethys along southern Europe and form the Alps from a subduction zone which dipped south after the Eocene. Since their model explains the detailed local geology, it is accepted within the more general framework illustrated here.

Figure 9 shows the Habkern arc of Hsü & Schlanger as a consuming boundary along the margin of southwestern Europe, with the Penninic plate approaching from the southeast. Since we lack any constraint upon the Cretaceous position of Africa, the Penninic plate might be part of the African plate, in which case the shear boundary indicated for the rotation of the Iberian Peninsula would also have a consuming component, most likely with a subduction zone dipping to the northwest. Alternatively, the Penninic plate may have broken away from the African plate, so that a spreading boundary would be required somewhere to the southeast. A third possibility, employing the simpler alternative to the Hsü & Schlanger model proposed above, would make the northern boundary of the Penninic plate a consuming margin with subduction zone dipping southeast, spreading away from Africa behind the arc.

Figure 8 indicates that the final third of the rotation of the Iberian Peninsula about a nearby pole had been completed. The Penninic plate had begun to override the Habkern arc near France, and overthrust the present northeastern corner of Corsica. In figure 7, the Penninic plate had completed its collision with Europe, and Corsica, Sardinia and the Calabrian fragments had broken away from Spain, which moved slightly into the Mediterranean by left lateral strike-slip motion along the Pyrenees.

In figures 5 and 6, we illustrate how the plate fragment which consisted of the northwestern part of the Italian peninsula and Corsica was consumed along its northern boundary by a subduction zone near the Southern Alps, in the vicinity of the Po basin. Evidence for the existence of this subduction zone is weak, as this area is presently subsiding and hence covered, but thrust boundaries around the Southern Alps do face south toward the Adriatic Sea (Wezel, 1970), and left lateral strike-slip motion along the NE-SW trending Judicaria Line has offset the Root Zone of the Alps, consistent with the indicated motion.

The Alpine arc is thus a continued locus of plate consumption, with as many as three distinct phases of thrusting, and one of subsidence, tectonizing both its interior and exterior. As a result, it contains the thickest pile of sediment derived from the Tethys Sea of any of the structures discussed. The eastward encroachment of the Apennine and Calabro-Sicilian arcs upon its past connection with Africa have obliterated the most recent traces of its motion. As we shall see next, motion in the eastern Mediterranean has been to the west, so that traces of the Alpine arc motion have been disturbed on that side also. The remaining Penninic plate is not yet consumed in the Ionian Sea, to the south of the Mediterranean Ridge, so investigation of that area may yield information relevant to the history of the Alps.

THE IBERIAN PENINSULA

Primary support for the idea that Spain has been moved relative to stable Europe comes from measurement of paleomagnetic field directions frozen into rock samples of both regions. Van Dongen (1967) found that the declinations of Permo-Triassic andesites and clays near Seo de Urgal were 30° counterclockwise (CCW) from the field directions found in Europe, a result supported by the study of Triassic sandstone and Silurian andesites and basalts from the Spanish Meseta (Van der Voo, 1967). The indicated rotation must have been complete before the Eocene, since subsequent virtual pole positions from the Lisbon Volcanics were in agreement with those from Europe (Watkins, Richardson, & Van der Voo, 1968). Van der Voo (1969) concluded that the rotation amounted to 40° CCW since pre-Hercynian times, 35° between the Late Triassic and the Late Cretaceous, with a relative pole of rotation between Spain and Europe in the Western Pyrenees. De Boer (1965) obtained a 65° CCW rotation beginning in the Late Triassic, and Zijderveld et al. (1970b) found a minimum of 50° CCW after the Permian, 35° CCW after the Triassic, confirming earlier work (Van der Voo & Zijderveld, 1969). Completion of the rotation by the Eocene has also been confirmed through further studies of the Lisbon Volcanics (Van der Voo & Zijderveld, 1971).

Evidence for the counterclockwise rotation of the Iberian Peninsula has been found in the Bay of Biscay. Three non-magnetic seamounts north of the Iberian Peninsula may be a marginal plateau which has subsided 3 km since the Late Cretaceous (Black, Hill & Laughton, 1964). A fan shaped pattern of linear magnetic anomalies which trend WNW to W is consistent with opening about a pole of location near the Pyrenees (Matthews & Williams, 1968). Reflection profiling and bottom samples taken in the Bay of Biscay are consistent with a pre-Eocene opening of more than 22° (Jones & Ewing, 1969). Bacon and Gray (1970) extended earlier gravity surveys into the eastern end of the bay and found a positive free air anomaly on strike with the North Pyrenean Fault, which they suggest may be a result of lherzolite intrusion associated with pre-Cretaceous sinsitral motion of the Iberian Peninsula. Along the Brittany shelf, low refraction velocities, low magnetic anomalies, and estimated continental densities suggest significant marginal downfaulting (Bacon & Gray, 1971). The JOIDES drilling project did not reach basement in the Bay of Biscay, hole 119, but bottomed in Paleocene (60 myr old) pelagic sediments continuous from Mid-Oligocene. Hole 118, also in the Bay of Biscay, reached spilitic basalt in a high temperature intrusive sill below Early Eocene (52 myr old) turbidites (W. A. Berggren, WHOI Sack Lunch Meeting, 1970), so that the bay is at least Paleocene in age, with volcanic activity at least in the Eocene.

Further off shore, Day (1959) reported a 10° slope off the northwest trending continental margin between Brittany and Ireland, which he suggested was a tension scarp. The continental slope deposits north and west of the Iberian Peninsula were steeper and less prograded than those west of the British Isles and France, which were a Cretaceous product of massive shelf erosion preceding the Tertiary downwarping and outbuilding (Stride et al., 1969). General agreement has been reached (Symposium, 1970) that the Cantabria, Biscay and Gascony seamounts mentioned above resulted from a linear east-west dislocation and uplift in the Eocene. Near 43N, 20W, the linear Palmer Ridge between Peake and Freen Deeps trends SE away from King's Trough, and has been an igneous and metamorphic basement formed about Early Eocene (60 myrbp), draped with Early Eocene to Mid-Miocene pelagic sediments (Cann & Funnell, 1967); Ramsay, 1970). This structure may have resulted from faulting and spreading required for the rotation of the Iberian Peninsula, and hence may be taken as a segment of the western boundary of the plate carrying Spain and Portugal.

The Pyrenees (table 3, appendix 1) are strongly eroded in the axial zone, but Late Cretaceous flysch on both flanks culminating in Eocene folding and emergence suggest that collision of opposed consuming boundaries between the Iberian Peninsula and the European mainland occurred in the Eocene. The absence of an eastward increase in the tectonism along strike suggest a great distance to the pole of rotation in the Eocene, consistent with the little or no change observed in the Eocene paleomagnetic pole location. In our main model, we therefore take the rotation of the Iberian Peninsula to be about a nearby pole (44N, OE, 7CCW) in the Late Cretaceous (60-75 myrbp) followed by rotation about a distant pole in the Eocene (70N, 25E, 2CCW; 45-60 myrbp) after which time the plate boundaries in figures 7 - 9 are no longer active.

Williams & McKenzie (1971) have interpreted shipboard magnetometer profiles recently gathered in the eastern North Atlantic Ocean and have correlated the fan shaped pattern of magnetic anomalies in the Bay of Biscay mentioned above with another fan shaped pattern opening southward and located to the southwest of Spain, and conclude that the Iberian Peninsula was rotated relative to Europe during the Upper Cretaceous and Lower Jurassic. If this dating is correct, we conclude that the Eocene folding of the Pyrenees Mountains is not a result of this rotation about a nearby pole, and must be assigned either to transform fault motion among Europe, North America and the Iberian Peninsula, or to plate consumption among the Iberian Peninsula, Africa, and Europe. The former alternative is difficult to accept, as the axis of the Pyrenees is roughly perpendicular to the Mid Atlantic Ridge, so little compression would be expected. The latter alternative requires a consuming boundary in the Bay of Biscay, say in the region suspected of marginal downfaulting by Bacon & Gray, mentioned above, and connected to the Azores-Gibraltar Ridge via some shear boundary to the west of Portugal, for which we have no evidence. The effects of using the above dating for the rotation of the Iberian

Peninsula are shown in figures 13 and 14, where the motions of the plate fragments are relative to a single rigid plate containing Europe and the Iberian Peninsula, in contrast to figures 7, 8, and 9. These plate fragments have evolved from another agency than the rotation of the Iberian Peninsula since consuming boundary activity has occurred on both sides of the southeastern termination of the Pyrenees mountains, which were taken to be the division between the fixed and rotating plates in either the Cretaceous or the Early Tertiary. This consuming boundary activity is illustrated in the Eocene and later motion of the Calabrian, southwestern Italian, northern Sicilian and Sardinian fragments which move away from the Iberian Peninsula to the southwest of the Pyrenees boundary, and of the Corsican and northwestern Italian fragments which move away from France to the northeast of the Pyrenees Boundary. The difference between figure 8 and figure 14 is that the assumption of no Early Tertiary motion of the Iberian Peninsula implies no necessity for an active plate boundary between Europe and Africa to the west of the position shown for Sardinia in the Early Tertiary. Evidence to the west of the Gulf of Lyons naturally proves difficult to unravel, as a result of the Oligocene-Miocene deformation associated with the motion of the Riff-Betic arc, so that no information from southern Spain is available to decide upon any single extrapolation past the end of the Cretaceous. We therefore close our examination of the Western Mediterranean Sea and turn to a brief discussion of the region to the east.

EASTERN MOTIONS

Albania and Greece from the Peloponnesus through Crete to Rhodes are the site of a discontinuous mountain chain, the Hellenides (table 1, appendix 1). The general sequence of overthrusting toward the southwest progressing southwest and culminating near Early Miocene time has been described in detail by Aubouin (1965). Although the geology of Yugoslavia is not well reported, it is apparent that the Dinaride mountain chain is a northwestward continuation of the Hellenides which ends near Trieste, to the south of the Eastern Alps. Modern plate activity at the southern end of the Hellenides and east to Turkey has been derived from seismicity and fault plane solutions by McKenzie (1970), who has shown that two small, rapidly moving plates are present in the Eastern Mediterranean Sea, the Aegean plate moving to the southwest, and the Turkish plate moving west. To the south and west of the Aegean plate, the Mediterranean Ridge extends in an arc from the Gulf of Taranto, southern Italy, to end southeast of the Island of Rhodes. Shipboard gravity measurements and Bouguer corrections based on various assumed densities have led Rabinowitz & Ryan (1970) and Woodside & Bowin (1970) to conclude that this ridge results from downwarping of the African plate and its consumption by the Aegean plate, not from uplift at a locus of plate spreading. East of Greece, continuous seismic reflection profiles indicate that the deformation of sediments on this ridge is greatest toward the north and west (Wong et al., 1971), consistent with the plate activity described above.

Applying the consuming boundary model developed in an earlier section, the outer edge of the Mediterranean ridge is taken to be the presently active consuming plate boundary between Africa and Europe in the Eastern Mediterranean Sea, and this consuming boundary must extend northward through the Bradano Trough mentioned in the section on the Apennine arc. We suppose that it further extends through the Northern Adriatic Sea, and turns to the east as a former right lateral shear boundary south of the Eastern Apennines, in order to maintain continuity of the Dinaride - Hellenide mountain chains.

The Carpathian Mountains to the north and east of the region discussed contain the Kliwa Sandstones, a continental rise deposit thrust toward the Russian platform (Wezel, 1970). The roughly east-west trend of the Carpathians is bent into a reversed S, with the Transylvanian Alps forming the center of the S and the Balkan Mountains continuing to the east as the bottom of the S. If the Carpathian Mountains were formed by an island arc system moving toward the Russian platform, as indicated by our model for a consuming boundary, this S curvature must have resulted from right lateral deformation after the collision of the arc and the continental margin, else no motion could have occurred. It seems reasonable to conclude that the former shear boundary south of the Eastern Alps extends eastward to the Transylvanian Alps, and that the modern shear boundary on the northern sides of the Turkish and Aegean plates was initiated quite recently.

We speculate that the region north of the Eastern Mediterranean Sea might have evolved as follows. Plate collision thrust the Carpathians over the Russian platform, and the locus of plate consumption shifted to the Dinaride-Hellenide arc. The region north of an eastwest trending shear boundary between the Eastern Alps and the Transylvanian Alps remained fixed to the Eurasian plate, while the southern region between this shear boundary and Greece moved west until it collided with the Apennine arc. The southernmost part of this southern region became the new locus of shear, at the northern boundaries of McKenzie's small plates, and the locus of plate consumption shifted south to its modern position southwest of Greece. This speculative evolution is conceived only for use as a working hypothesis to examine possible motions to the east of the area discussed in more detail in the previous sections.

We require some such evolution to explain the presence of the Dinaride Mountains in the wake of the Penninic plate, which was responsible for the Eocene formation of the south central European Alps, as discussed in the section on the Alpine arc. If the Dinarides had maintained their present position to the south of the Eastern Alps throughout the

Tertiary, then the Eastern Alps, and necessarily the Western Alps, could not have been formed by northward motion of the Penninic plate, no matter what was the direction in which the Penninic subduction zone dipped.

The presumed motion of the Dinaride-Hellenide arc is illustrated in figure 4, where the Apulian peninsula is withdrawn from the Italian Peninsula, and in figure 6, where the Dinaride mountain coast of the Adriatic Sea is also withdrawn from the Italian peninsula. The eastern coast of the Italian peninsula is assumed uplifted as a result of the approach of two facing consuming boundaries, so is not included in any rotation; it is left as a visual guide to the motion of the Apennine fragments, and is not otherwise significant except in figure 3, where it represents the modern shoreline.

CONCLUSION

The complex geology of the Mediterranean Sea region has been the source for many theories of mountain building, and its mountain chains have been intensively studied. If the hypothesis of plate tectonics is to be fully accepted by land geologists as a useful description of the regional background for their work, then the hypothesis must eventually be tested in an application to this classical area. We have synthesized a simple model for mountain building from the interaction of plates, and have squeezed that model into a long loop around the Western Mediterranean Sea.

The greatest theoretical defect in this exercise has been our dependence upon the idea of spreading behind island arcs. We were moved toward its use by the young, subsided nature of the Western Mediterranean basins and the generally outward sense of thrusting of the mountain chains around the basins, particularly the Riff-Betic arc. Our lack of understanding of this spreading process has prevented rigorous application of the tenets of plate tectonics, except for the use of rigid rotations upon a sphere.

The greatest experimental defect in this exercise has been our use of spot data in a complicated area, and the assumption of continuity across many hundreds of kilometers. In particular, we lack descriptions of the Tellian Atlas

Mountains, the southern Apennine Mountains, and the islands of Corsica and Sardinia. Paleomagnetic data from the Italian Peninsula and from the island of Sicily would have been greatly preferred to the use of the orientation of thrust boundaries. The question of the amount, direction, and timing of the motions of Africa and the Iberian Peninsula has been mentioned above.

We conclude that the hypothesis of plate tectonics can be applied to the Western Mediterranean Sea region with limited success, at a size scale somewhat smaller than has been involved in the development of the hypothesis; that this application can successfully include many of the geological and geophysical constraints found in the literature; and that the evolution of the Western Mediterranean may have taken place either along the path illustrated in figures 3 through 9 or its alternates in figures 10 through 14, or some variation of those paths toward a looser fit.

APPENDIX I

Geological Outlines

1) Hellenides; southern Italy, Albania, Greece

2) Northern Appenines; northern Italy, Genoa to Rome

3) Pyrenees; Spain, Andorra, France common boundaries

4) Western Alps; France and Italy, common boundary

TABLE 1

HELLENIDES

Upper Triassic	shallow shelf sea, algal limestone subsidence pelagic limestone and jasper in Pindus
Upper Liassic	block normal faulting in Ionian, sub-sidence, terrigenous clastics on Apulian side of Ionian, radiolarites in Pindus
Upper Jurassic	pillow lavas, spilites in Gavrovo, Pindus, ophiolites at greenstone metamorphic grade in sub-Pelagonian, Pindus
Lower Cretaceous	
Barremian	flysch to Pindus
Aptian-Albian	radiolarite microbreccia to Pindus
Cenomanian	radiolarite microbreccia to Pelagonian
Maestrichtian	flysch to Pindus and sub-Pelagonian
Eocene	
Middle-Upper	flysch fills Pindus, sandstone and con-glomerate on Pelagonian side. fines overflow Pindus to Gavrovo and Ionian. Vardar receives Molasse
Late Eocene-Early Miocene	Pindus overthrust Gavrovo to SW, conglomerates to Meso-Hellenic, Gavrovo uplift
Burdigalian	Ionian above sea level
Helvetian	Meso-Hellenic above sea level Ionian overthrust Apulian, Molasse to Milise Trachyandesite volcanics and gradiorite plutons in Pelagonian, Vardar and Rhodope
Pliocene	basaltic volcanics in sub-Pelagonian subsidence of Aegean, Thessaly, Bradano

SW						NE
Apulian (ridge)	Ionian (furrow)	Gavrovo (ridge)	Pindus (furrow)	sub-Pelagonian (margin)	Vardar (ridge)	Rhodope (Massif)

TABLE 2

NORTHERN APENNINES

Upper Triassic	Verrucano sandstone, last terrigenous section carbondates and evaporites, then neritic limestone
Lower Jurassic	begin slow subsidence, carbonate-siliceous limestone, marl, chert, radiolarites in Tuscan-Umbrian; shallow water calcareous dolomites in Latium-Abruzzi
Upper Jurassic	ophiolites (serpentinites, lherzolites, gabbro, pillow lava) in eugeosyncline until Neocomian. calcareous dolomites until Tithonian. Clays until Turonian
Lower Cretaceous	pre-flysch clays, marl and cälcareous trubidites in Tuscan
Cretaceous	slumping in eugeosyncline, Olistothromes with ophiolites and Corsica-Sardinia granites
Upper Cretaceous	arenaceous Helminthoides flysch from Corsica-Sardinia, first napping in southern part of eugeosyncline, to E and NE. pre-flysch limestone and marl in Umbrian, facies migrate E. Latium-Abruzzi emerges
Paleocene	end of Helminthoides flysch
Eocene	first napping in northern part of eugeosyncline, with calc-arenaceous flysch
Early Middle Oligocene	flysch, thick turbidites from NW in Tuscan and from Corsica-Sardinia to the SW
Oligocene-Miocene boundary	first napping of eugeosyncline over miogeosyncline, moving east
Early Miocene	end Tuscan flysch, begin Umbrian flysch andesite-dacite volcanic ash, Latium-Abruzzi sedimentation resumes
Late Miocene	end Umbrian flysch, block faulting; acid volcanics, magmatism migrate east
Upper Lower Pliocene	end napping, end faulting, begin conglomerate in Po Valley

TABLE 2 (cont.)

Upper Pliocene	Eugeosynclinal uplift, normal faulting
Quaternary	granodiorite and quartz monzonite to west, mixed extrusive and intrusive volcanic elsewhere

West				East
"Eugeosyncline"	Tuscan	Umbrian	Latium-Abruzzi	Po (North)

after Abbate et al., 1970

TABLE 3

PYRENEES

Permian- Triassic	generalized continental red beds evaporites Volcanics to south, intrusive and extrusive, andesite, much basic greenstone, "ophites"
Jurassic- Lower Cretaceous	sporadic sediments on north flanks west and east on south flanks
Upper Cretaceous Cenomanian- Campanian	begin carbonates; marl, limestone, dolomite to S
Campanian	conglomerates, clastics
Maestrichtian	flysch to north
Eocene Landenian- Bartonian	flysch to 4 km on south flanks, derived from S
Lutetian	folding, vergenz away from axis, emergence stronger metamorphism to North
Miocene Girondian	flysch on north flanks moves north
Pliocene post Pontian	uplift of axial zone

after Rutten, 1969

TABLE 4

WESTERN ALPS

Carboniferous-Permian	Brianconnais subsides; shallow sediments, pyroclastic and effusive volcanics to 3 km
Lower Triassic	peneplained, dune and aeolian derivatives
Middle Triassic	Muschelkalk sea transgresses, unites w Tethys
Upper Triassic	and regresses
Lower Jurassic	Dauphinois and Piemont subside, shale and marl
Middle Jurassic	breccias on Brianconnaise and Tarantaise
Upper Jurassic-Lower Cretaceous	Dauphinois and Subbrianconnais receive crystalline debris, basic submarine volcanics, spilites and serpentinites intrude (Ophiolites)
Upper Cretaceous-Eocene	spasms of flysch, uniform pelagic sediments
Upper Eocene	Helminthoids Flysch napping W
Oligocene	schistes lustres napping
Middle Miocene	maximum instability and thrusting NW most intense to NE; metamorphism increases eastward: lawsonite-pumpellyte in outer Pennine nappes, glaucophane-chloritoid inner. kyanite-sillimanite in Lepontine, later

E W

Piemont Brianconnais Sub-Brianconnais Dauphinois

after Ramsay, 1963
and Debelmas and Lemoine, 1964

APPENDIX II

Geological Time Scales

A) 0-80 myr

B) 80-200 myr

TABLE 5a

TIME SCALE A

0 - 80 myr

0	Villafranchian		Pleistocene
	Levantinian		
	Dacian	Astian	
	Plaisancian		Pliocene
	Pontian	Pannonian	
-10	Meotian		
	Sarmatian =Messinian		
	Toarcian	Tortonian	
		Vindobonian	
-20	Helvetian		Miocene
	Burdigalian		
		Girondian	
	Aquitanian		
-30	Chattian		
	Stampian		Oligocene
	Sannoisian		
-40			
	Ludian		
		Priabonian	
	Bartonian		
-50			Eocene
	Lutetian		
	Ypresian		
-60			
	Landenian		
	Montian		Paleocene
	Danian		
-70			
	Maestrichtian		
	Campanian		
-80			
	Santonian		
		Senonian	
			Upper Cretaceous

after Van Eysand
2 myr/line

TABLE 5b

TIME SCALE B

80 - 200 myr

				UPPER
- 80	•	Santonian	Senonian	CRETACEOUS
	•			
	•	Coniacian		
- 92				
		Turonian		
	•			
	•	Cenomanian		
-104	•	Albian		
	•	Aptian		
	•	Barremian		
-116				LOWER
	•	Hauterivian		CRETACEOUS
	•	Valangian		
-128		Berriasian		
	•			
	•	Tithonian		
-140				
	•	Kimmeridgian	Malm	
		Oxfordian		
-152	•			
	•	Callovian		
				JURASSIC
	•	Bathonian	Dogger	
-164				
	•	Bajocian		
		(Aalenian/Toarcian)		
		Pliensbachian	Lias	
-176	•			
	•	(Sinemurian/Hettangian)		
	•	Rhaetian		
-188				
	•	Norian	Keuper	
	•	Carnian		
-200				
	•	Ladinian		
			Muschelkalk	
	•	Anisian = Virglorian		TRIASSIC

after Van Eysand
3 myr/line

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