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Active Tectonics of the Mediterranean Region

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Summary

Examination of more than 100 fault plane solutions for earthquakes within the Alpide belt between the Mid-Atlantic ridge and Eastern Iran shows that the deformation at present occurring is the result of small continental plates moving away from Eastern Turkey and Western Iran. This pattern of movement avoids thickening the continental crust over much of Turkey by consuming the Eastern Mediterranean sea floor instead. The rates of relative motion of two of the small plates involved, the Aegean and the Turkish plates, are estimated, but are only within perhaps 50 per cent of the true values. These estimates are then used to reconstruct the geometry of the Mediterranean 10 million years ago. The principal difference from the present geometry is the smooth curved coast which then formed the southern coast of Yugoslavia, Greece and Turkey. This coast has since been distorted by the motion of the two small plates. Similar complications have probably been common in older mountain belts, and therefore local geological features may not have been formed by the motion between major plates.

A curious feature of several of the large shocks for which fault plane solutions could be obtained for the main shock and one major aftershock was that the two often had different mechanisms.

1. Introduction

The concepts of plate tectonics have principally been established by the study of oceanic plate boundaries. The seismicity of such boundaries is confined to a zone no wider than a few tens of kilometres, and therefore ridges, trenches and transform faults can be considered as lines on the Earth's surface along which plates are produced, destroyed, or slide past each other. Though the processes which operate along these features are of considerable interest, the very success of plate tectonics shows that it is not necessary to understand them thoroughly in order to describe the tectonics of the oceans.

Continental tectonics is more complicated. Many authors have remarked that the distribution of epicentres on continents is quite different from that of those beneath the oceans. This difference is clearly shown in Figs 1 and 2, where the epicentres along the Mid-Atlantic ridge follow a thin strip whereas those in the Mediterranean and in Iran in particular are distributed over a broad belt. This scatter is not due to errors in the location of epicentres, which are usually less than 30 km, or about twice the

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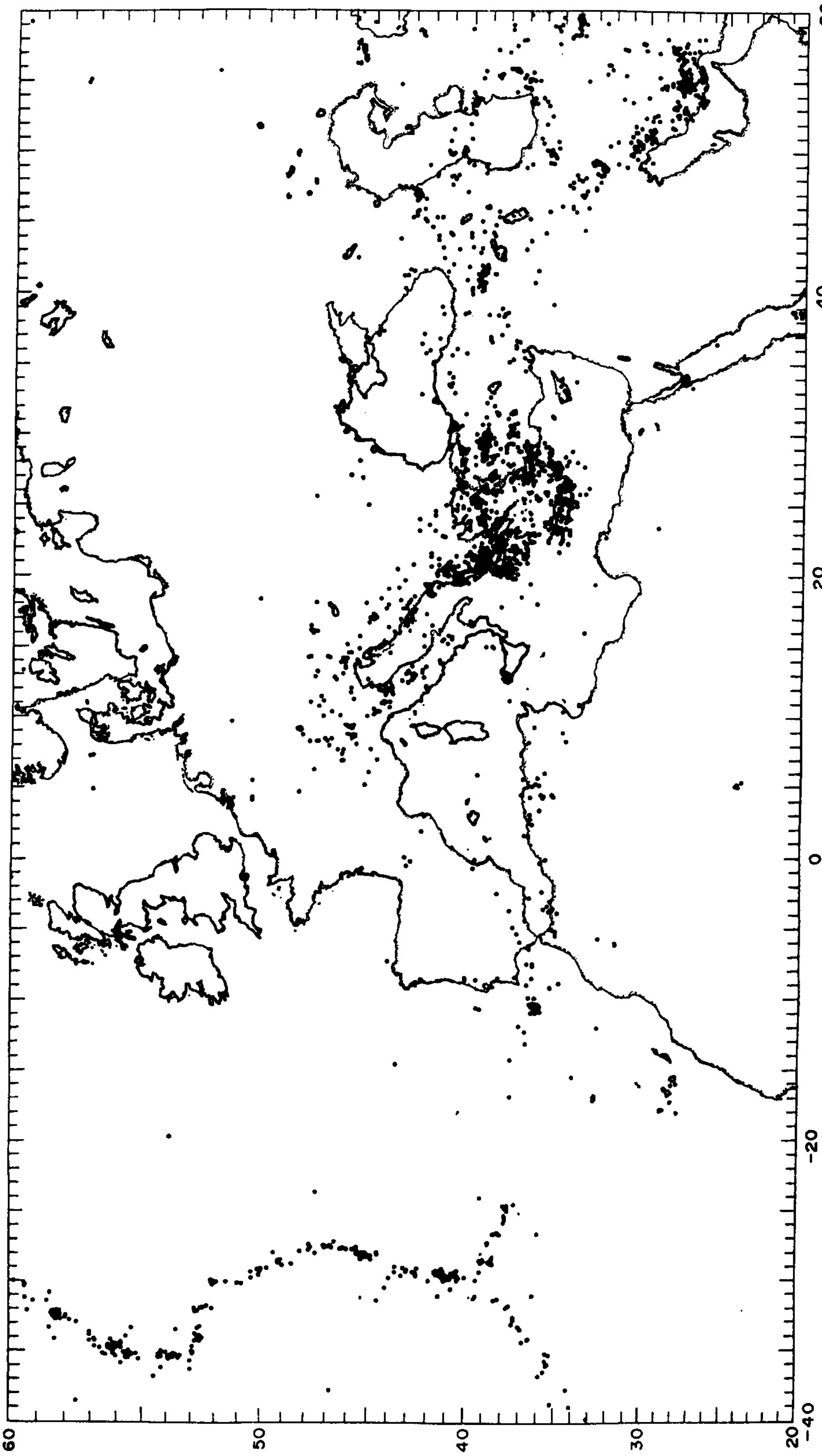


Fig. 1. Positions of all epicentres of earthquakes between 40° W and 60° E, 20° N to 60° N located by the U.S. Coast and Geodetic Survey between 1961 January 1, and 1970 July 30.

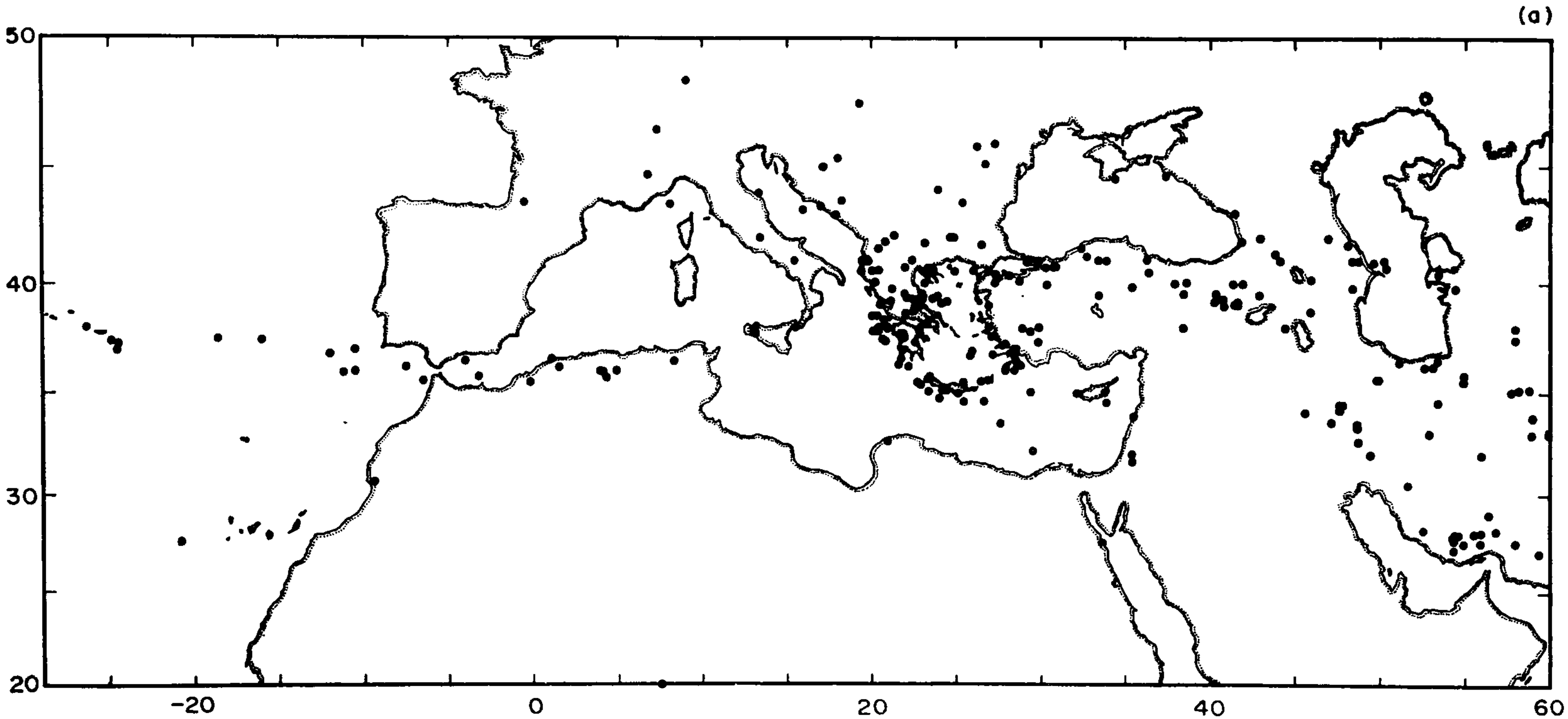


FIG. 2(a). Major shocks on the Azores Gibraltar Ridge, the Mediterranean and Iran between 1922 and 1970. The magnitude of most earthquakes shown is 6 or greater, but in several cases is not accurately known. These epicentres are taken from Gutenberg & Richter (1954), the ISS bulletins and the USCGS listings, and have not been relocated. Notice that the activity on the major plate boundaries is more striking than in Fig. 1.

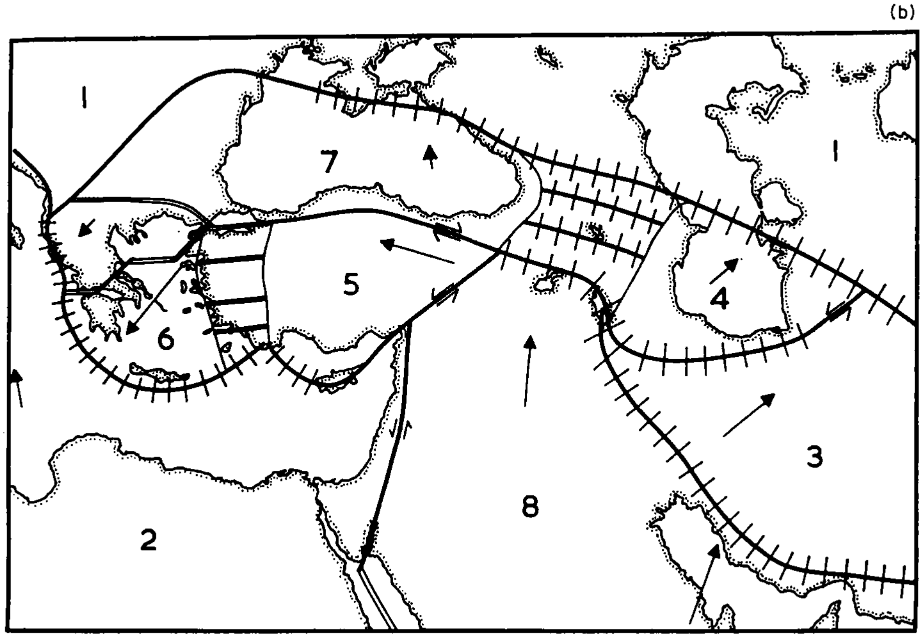


FIG. 2. (b) Sketch of plate boundaries and motions. The arrows show the directions of motion relative to Eurasia and their lengths are approximately proportional to the magnitude of the relative velocity. Plate boundaries across which extension is occurring are shown by a double line, transform faults by a single heavy line and boundaries across which shortening is occurring by a solid line crossed by short lines at right angles. The activity in western Turkey and in the Caucasus is due to normal faulting and thrusting respectively, but the plate boundaries shown in the figure in these regions do not correspond to individual plates and their boundaries, but only represent the general nature of the deformation. The plates are assigned the following numbers: 1 Eurasian, 2 African, 3 Iranian, 4 South Caspian, 5 Turkish, 6 Aegean, 7 Black Sea and 8 Arabian.

diameter of the dots in Fig. 1. Nor is it due to the use of different methods of locating earthquakes, since one global network of stations is used for all earthquakes. This difference is therefore real, and shows that it is not in general possible to describe the tectonics of continents in terms of the motion of a few large aseismic plates. This result follows directly from Fig. 1 if active platé margins are defined to be faults on which earthquakes occur (Wilson 1965).

Two possible explanations for the difference between the deformation of continents and oceans have been suggested. The first depends on the existence of many older faults which act as lines of weakness. Such faults do not exist on ridge axes because they are the sites of sea-floor production. The fracturing of existing ocean floor is, however, often controlled by the positions of old transform faults and ridge axes. The other possible explanation of the difference between oceanic and continental seismicity is that the rocks which form continental plates behave in a different way from oceanic plates when subjected to shearing stresses. Such a difference would not be unexpected because continents contain rocks with more silica than those of the oceans. Silica-rich rocks like granite deform at low temperatures more easily than

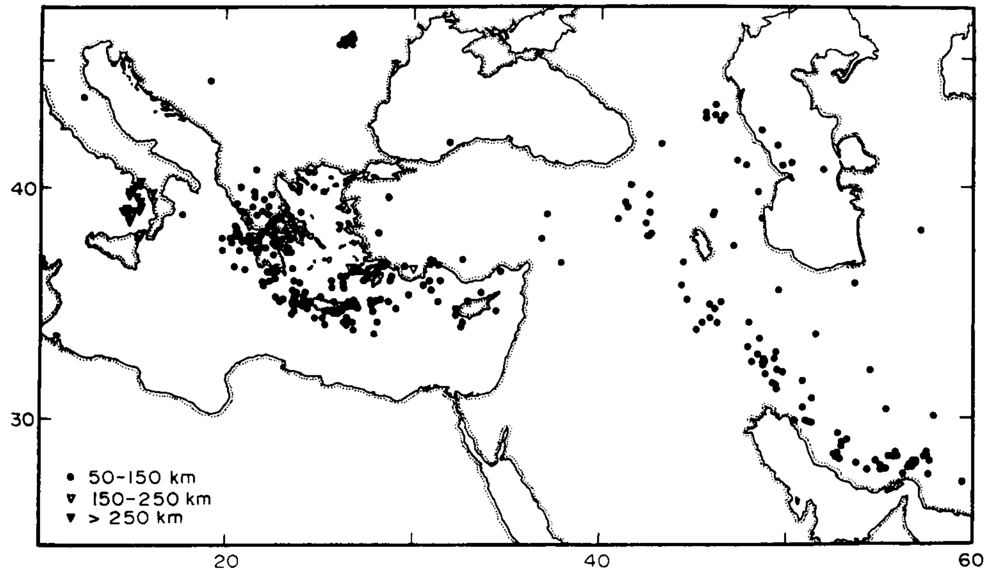


FIG. 3. Epicentres of foci at and below 50 km between 1961 and 1970. The depth of the foci given by the USCGS were used to obtain this figure, and may well be in error by 50 km. Since focal depths of greater than 50 km are believed to occur only where plates are being destroyed, the activity in this figure suggests that crustal shortening is now taking place over a wide region in the Eastern Mediterranean and South-west Asia.

basalts, and therefore may lead to differences in the behaviour of continental and oceanic regions. Though this problem is important it is more likely to be solved by laboratory experiments than by field studies, and will not be further discussed in this paper.

The principal problem with which this study is concerned is whether the simple concepts of plate tectonics can usefully be applied to the continents, and in particular to the Mediterranean region. This was also the subject of a brief earlier report (McKenzie 1970), hereafter called (I). In the period since (I) was written several more fault plane solutions have been obtained which now permit a more detailed discussion of the tectonics of Italy, Yugoslavia and Iran than was previously possible.

The main conclusions of this present study are shown in Fig. 2(b), which also gives the names of the major and minor plates and their directions of motion relative to Eurasia. None of the conclusions reached in (I) have been altered by the more recent data, though various new plate boundaries are suggested below to account for the new data.

The principal plates in Fig. 2(b) are Africa, Eurasia and Arabia. The motion of Africa with respect to Eurasia may be obtained from the spreading rates and strikes of fracture zones in the Atlantic to the north and south of the Azores triple junction, together with slip vectors from shocks on the Azores–Gibraltar Ridge. The pole and rate so obtained are shown in Fig. 4 and Table 1, and show that Africa is moving northward towards Eurasia from Gibraltar eastward. This movement is taken up on the plate boundary which crosses North Africa and continues through or south of Sicily. The boundary probably continues around the Adriatic through Italy and Yugoslavia rather than crossing the entrance to the Adriatic directly to Greece. This suggestion that the Adriatic is part of the African plate is, however, still based on few reliable fault plane solutions and may be modified by later work.

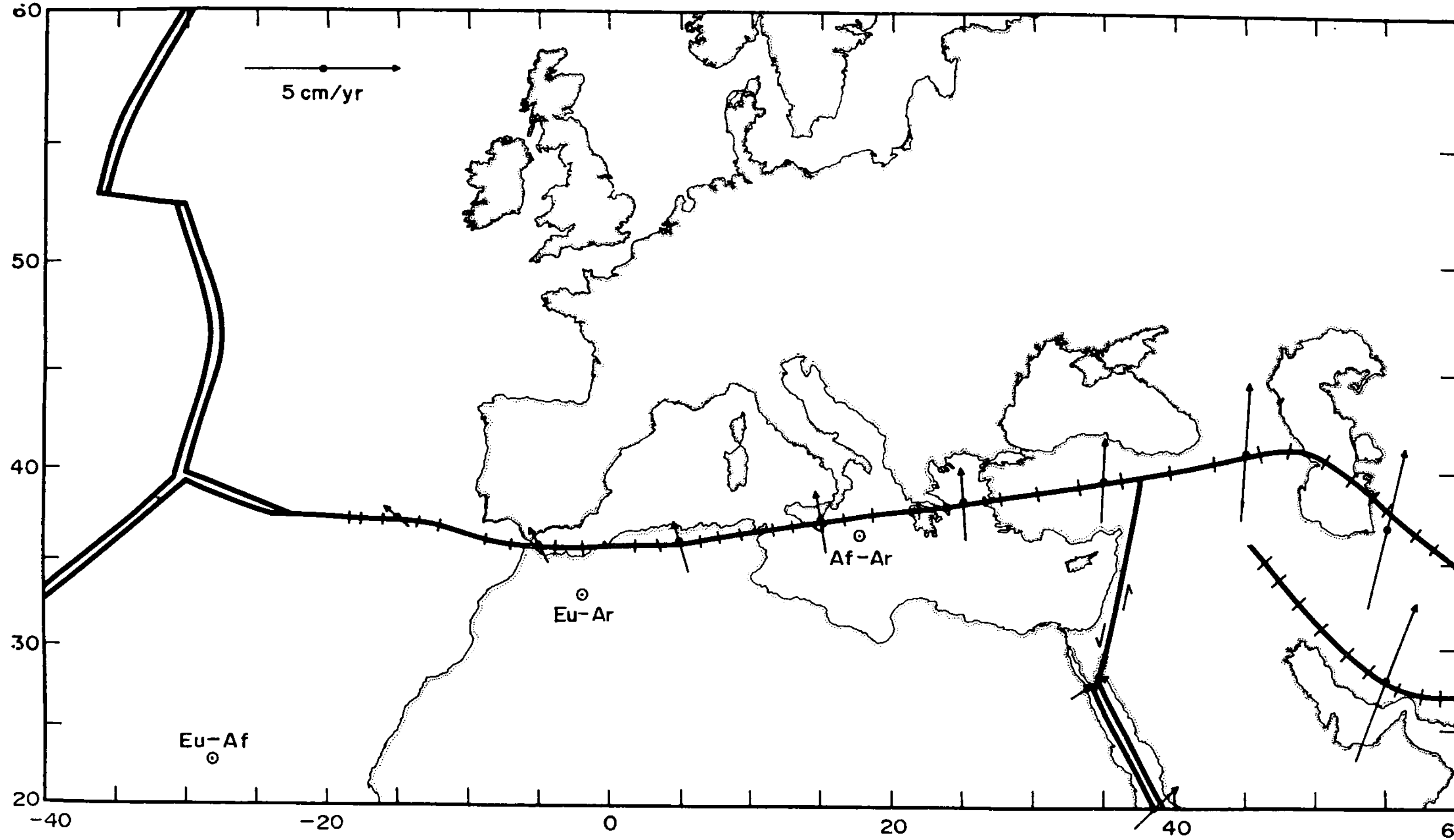


FIG. 4. The relative motion between Africa, Arabia and Eurasia obtained from observations in the surrounding oceans. The plate boundaries are simplified to a single line in order to show the direction and magnitude of the relative motion which must be taken up by all the various motions taking place across the active belt. Extensional plate boundaries (ridges) are shown as a double line, transform faults as a single heavy line and boundaries where shortening is occurring as a solid line with short cross lines. Two boundaries are shown in Iran, but the slip vectors on each show all the relative motion between Arabia and Eurasia. The interpretation of the active processes in South-west Asia and the Mediterranean would have been extremely difficult if the motion between the three major plates in this figure had not already been obtained from oceanic magnetic lineations. The vector in the North-east Atlantic shows the scale of all the other vectors in the figure, and does not represent the motion between America and Eurasia.

Table 1

Instantaneous rotation poles

		Latitude	Longitude	Rate units of 10^{-7}°/yr
America	Eurasia(1)	58·6	152·5	2·44
America	Africa (1)	68·3	−28·9	3·38
Eurasia	Africa (1)	22·7	−28·2	2·73
Africa	Arabia (1)	36·5	18·0	3·10
Eurasia	Arabia (2)	32·2	−5·2	5·44
Eurasia	Turkey (1)	18·8	35·0	6·20
Africa	Turkey (2)	10·2	60·4	5·30

The rate is positive if the first plate listed is taken to be fixed. Poles labelled (1) were obtained directly from observations, poles labelled (2) calculated from the vector triangles. The Eurasia–Turkey pole was obtained from the data in Section 8.

East of 20° E the motion between Africa and Eurasia is not taken up on one plate boundary, but is confused by the rapid motion of two small plates, the Aegean and Turkish plates. This suggestion was originally made in (I) to explain the large number of earthquakes on the Aegean boundaries and the focal mechanisms of shocks in northern Turkey and the northern Aegean. Many more focal mechanisms have since been obtained especially along the Cretan Arc, and they have strongly supported the original suggestion. There is also now some evidence that the Black Sea and surrounding regions form another plate moving towards Eurasia, though with a slower rate of motion than that of the Aegean and Turkish plates. The boundary between the Aegean and Turkish plates forms a north–south belt of seismicity across western Turkey and the eastern Aegean. It is quite different from any oceanic plate boundary and focal mechanisms in this area show in some detail how the motion is taken up.

At the eastern end of the Turkish plate the motion is taken up by thrust faults associated with the Caucasus. No single fault forms the boundary, but the motion on all faults is approximately in the same direction as that between Arabia and Eurasia. The result of this geometry is to thicken the continent throughout the active region, and hence to continue to elevate the Caucasus. This behaviour, like that of western Turkey, is not similar to that on oceanic plate boundaries.

East of the Caucasus the present deformation is concentrated in three belts. The northernmost crosses the Caspian from the Baku peninsula to the Kopet Dag range on the border between the USSR and Iran. The other two start as one belt near the Caucasus, then a branch occurs and the northern belt follows the Elburz mountains south of the Caspian to join the northernmost belt in the Kopet Dag. The southernmost belt follows the Zagros range close to the northern boundary of the Arabian Shield. Mechanisms on all three boundaries show thrusting. Though it is not yet possible to determine the strike slip component for most shocks because most of the foci are in the crust, it is probably less than the thrust component. Also the seismic activity of Iran is different from that of Turkey because large earthquakes occur in regions far from the three belts discussed above. Even when the shocks are within the belts they are often not produced by slip on major fault systems like the North Anatolian Fault in Turkey (Ganser 1969).

The principal reason why the deformation of the Mediterranean region is of interest is that it is an accessible and reasonably well-studied area where the motion between the major plates involved is well known. The poles and the angular velocities between Africa, Eurasia, America, Arabia and Turkey are given in Table 1. The motion between Africa, America and Eurasia was determined by Chase (private communication) using the Atlantic spreading velocities and fracture zones as well as four

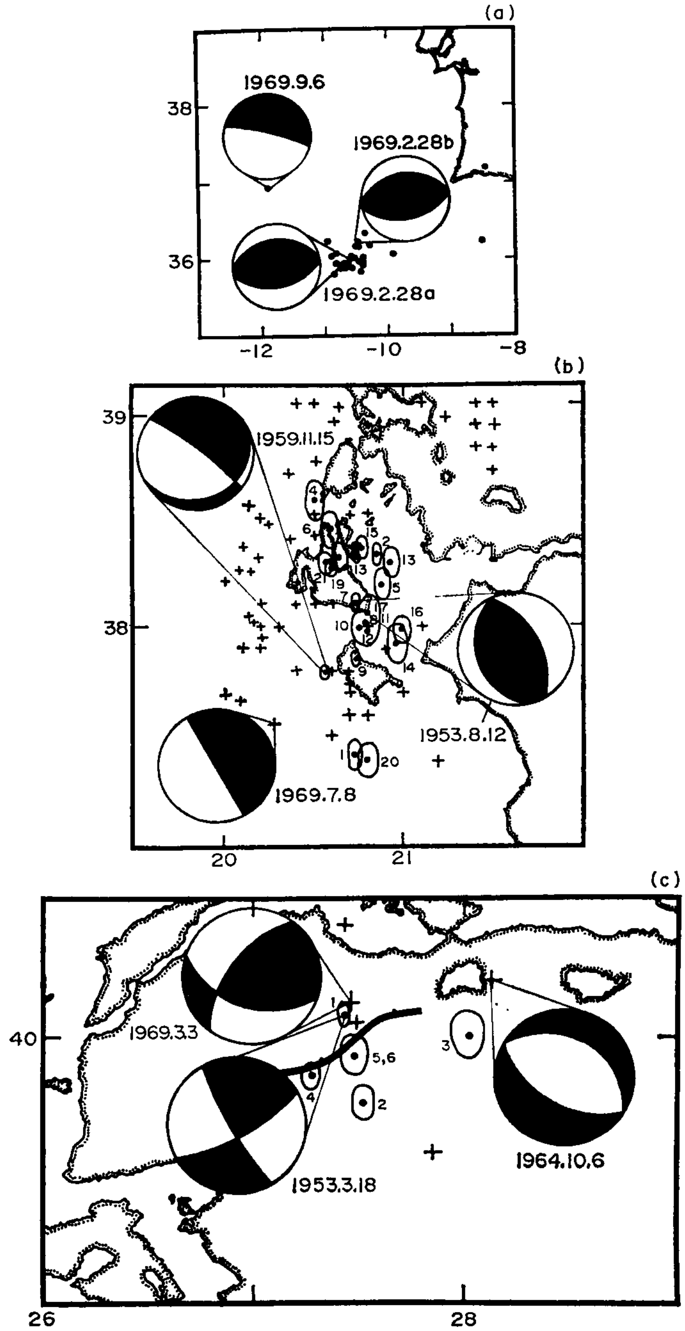


FIG. 5(a)–(c)

slip vectors obtained from fault plane solutions in this paper. The motion between Africa (Nubia) and Arabia was obtained from McKenzie, Molnar & Davies (1970), assuming that the Red Sea started to open 15 My ago, and has done so at a constant rate thereafter. Since the movement between the Arabian and African plates is principally taken up on the Dead Sea Rift, the Africa–Eurasia pole describes the total deformation across the Mediterranean west of about 37° E, the Arabia–Eurasia pole governs that across the active belt between 37° E and 60° E, where this study stops.

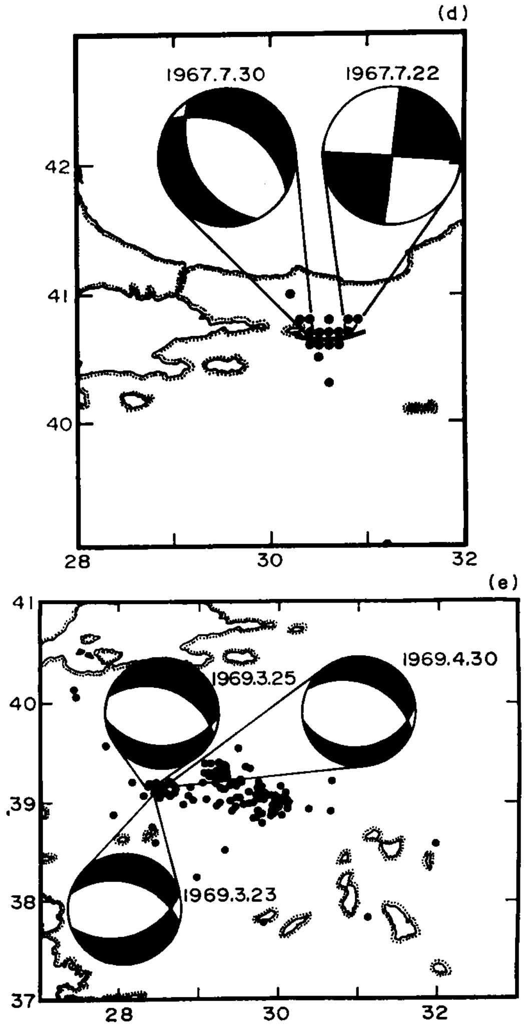


FIG. 5(d) and (e)

The motions derived from these poles are shown in Fig. 4. The success of plate tectonics elsewhere is so striking that these poles and rotation rates can be used with considerable confidence to describe the total deformation which is presently occurring across the Alpide belt. It is fortunate that this is the case because the complications of the active belt east of Sicily are such as to permit only the crudest checks of the motion between the major plates. The deformation of the Alpide belt east of 15° E is of interest not because it provides an accurate test of the assumptions of plate tectonics, but because it tests the usefulness of the concepts of plate tectonics as a description of complicated continental deformation. In a pedantic sense plate tectonics can describe any two-dimensional deformation which can occur, but if the number of independent plates required by such a description becomes large, then plate tectonics ceases to be useful. The equations of fluid dynamics are a good analogy to this problem. Newton's second law must apply to each molecule of any fluid, but does not provide a useful description of the motion. Instead the equations governing the movement of large numbers of molecules, or the continuum approximation, form the starting point of most studies of fluids. The problem of continental deformation is similar, since it is always possible to describe every thrust as a trench and every normal

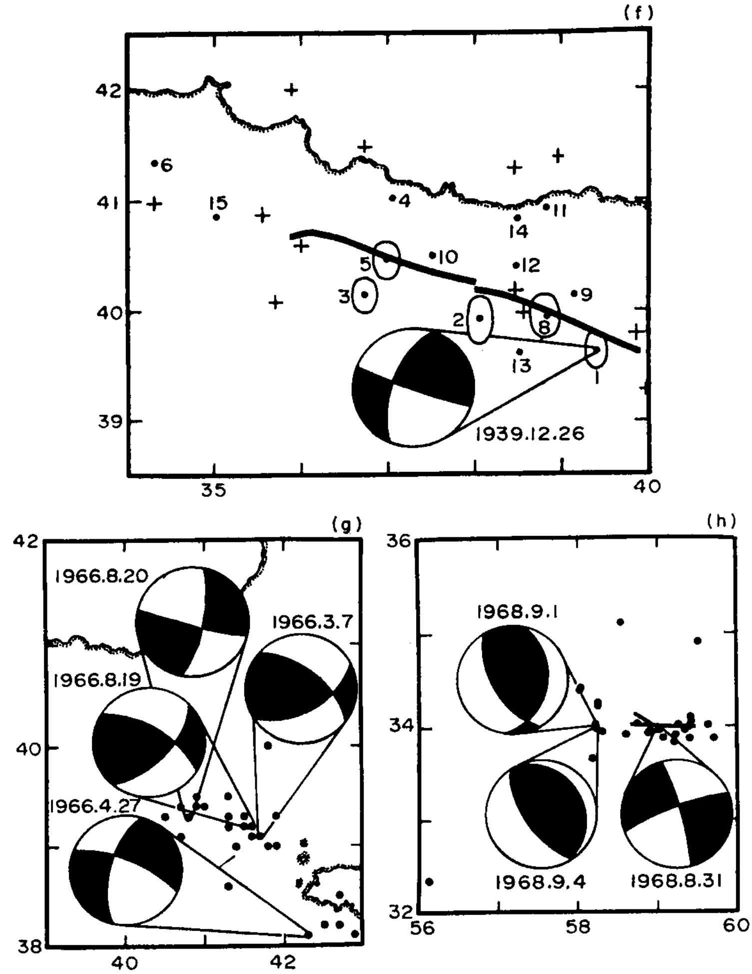


FIG. 5(f)–(h)

FIG. 5. Various large earthquakes and their aftershock sequences, to show in three cases, (d), (g) and (h) the difference in the mechanism of the main shock and the principal aftershock. From W to E. (a) On the Azores–Gibraltar Ridge, on 1969.2.28. USCGS locations. (b) Ionian Islands, starting on 1953.6.21 with epicentre labelled 1. All epicentres relocated and positions given in Table 4, p.154 Sequence was from 1 to 21, with 7 being the largest shock. The ellipses surrounding each epicentre show the probable error in location calculated by the relocation program, but unless the distribution of recording stations is uniform over the focal sphere this error does not include systematic errors in location. USCGS epicentres between 1961 and 1970 are shown as crosses. The mechanisms of three earthquakes are shown as bubbles, with the compressional quadrant black (see Table 3 and the individual mechanisms for details). The distribution of epicentres and the fault plane solutions both suggest that the motion is taken up on an almost horizontal thrust with a lateral extent of at least 50 km. Such a feature would closely resemble the base of a large alpine nappe. (c) The aftershock sequence of the Yenice–Gonen earthquake of 1953.3.18 (point 1), showing the continuity of the faulting of this earthquake and that of 1964.10.6. The surface break is shown as a continuous line. Other symbols as in (b). Notice the remarkable difference in mechanism between 1953.3.18 and 1969.3.3. (d) Mudurnu earthquake of 1967.7.22, and aftershock sequence located by the USCGS. Surface break from Ambraseys & Zatopek (1969). The triple junction between the Eurasian, Turkish and Aegean plates must be between the epicentres of 1967.7.22 and 1967.7.30. (e) Sequence of shocks starting on 1969.3.23 and continuing eastward with the Gediz earthquake of 1970.3.28, associated with the formation of an east–west graben by extension (Ambraseys & Tchalenko 1970). (f) Surface break and aftershock sequence of

the Great Turkish Earthquake of 1939.12.26. Numbered epicentres without ellipses are all poorly located because of poor azimuthal distribution of observations. Conventions otherwise as for (b). This is the largest shock to occur since 1920 within the region in Fig. 1, and was produced by up to 3·7 m of strike slip motion on a long straight section of the fault. (g) Various shocks at the eastern end of the North Anatolian Fault, USCGS locations. Notice the difference in mechanism between the Varto earthquake of 1966.8.19 and its aftershock on 1966.8.20. The triple junction between the Turkish, Eurasian and Arabian plates lies between these epicentres. (h) The Dasht-e-Bayaz earthquake of 1968.8.31 and its aftershocks. USCGS locations. The main shock and the surface break show strike slip motion, yet two of the aftershocks were produced by thrusting. Surface break from Ambraseys & Tchalenko (1969).

fault as a ridge. If, however, the number of such boundaries required is large it may be more useful to formulate a continuum theory rather than attempt to use plate tectonics in its present form. Such a continuum theory would treat the region between Africa and Eurasia as a mobile belt without strain rate discontinuities, and would in some ways resemble the earlier concepts of orogenic belts. The results below show that plate tectonics can describe most of the deformation at present taking place, but in parts of the Northern Aegean and in Greece there are some earthquake mechanisms which appear to reflect stresses within the plates rather than the motion between them. If further work supports this result then a continuum theory may be required.

Plate boundaries are by definition the seismically active zones of the Earth. Therefore the epicentres in Figs 1 and 2(a) must be used to mark the plate edges in the Mediterranean.

Since the only known occurrences of intermediate and deep focus earthquakes are where oceanic lithosphere has been consumed in the last 10 My, the position of such foci in Fig. 3 should mark places where such consumption has recently taken place, and where plates are probably still approaching each other. Of the various techniques at present available for studying the motion between plates only fault plane solutions of earthquakes can be used to investigate the details of the motions which cause the activity in Figs 1 and 2(a). There is, however, one important disadvantage of the technique. The mechanism of earthquakes determine the direction, but not the rate of motion between plates. Brune (1968) has overcome this problem by using the cumulative moment of earthquakes on a given plate boundary to estimate the total displacement which occurred over several decades, but estimates of the rate of slip obtained by this method are unlikely to be as accurate as those derived from oceanic magnetic lineations. Fault plane solutions of earthquakes have now been determined in large numbers, and form the principal observational evidence for the conclusions of this paper. Though several authors (Honda 1962; Stauder 1962; Sykes 1967) have discussed the theory and the principles involved, the use of the wave form and of the first motion of the *S* waves has not been sufficiently emphasized. All known fault plane solutions require a double-couple source mechanism. It is therefore impossible to distinguish between the fault plane and the auxiliary plane by using first motion observations of *P* or *S* waves. Such a distinction can be made by using the distribution of aftershocks, especially if the mechanism was strike slip. If the shock produced a surface break then the fault plane must have the same dip and strike as the observed break. McKenzie & Parker (1967) pointed out that plate tectonics can also be used in the case of strike slip earthquakes, since the motion between two plates must be consistent with rotation of one with respect to the other about some axis. However the ambiguity cannot be resolved in the case of a thrust or normal fault mechanism beneath the sea. Since it is the horizontal projection of the slip vector which is of importance in plate tectonics it is only necessary to determine which is the fault plane in the case of faulting with a strike slip component. Therefore it is usually possible to resolve the ambiguity in cases where it affects the direction of motion between the plates.

One observation applies to all studies which use seismicity and fault plane solutions alone to study regional tectonics. Since good determinations of epicentres can only be made for earthquakes which occurred during the last 20 yr, and fault plane solutions have only become reliable since the installation of the World Wide Network 10 yr ago, the plate motions and boundaries are strictly only a valid description of the tectonic deformation which has occurred during the last 20 yr. The seismic observations alone do not apply to motions before 1950, and must be extrapolated to earlier times only on the basis of other evidence. Therefore the principal concern of this paper is with instantaneous motions and not with the processes by which the geological features have been produced. This problem is discussed in detail in Section 8.

This study is principally concerned with the deformation of continents and not with the mechanism by which the seismic energy is generated from the stress field. However, one rather surprising observation was that earthquakes of sufficient magnitude for a fault plane solution to be obtained for both the main shock and the principal aftershock often had different mechanisms for the two events. Three such shocks occurred in the area after 1963, and plots of their aftershock distribution surface breaks were mapped and mechanisms are shown in Fig. 5(d), (g) and (h). In two cases, Fig. 5(d) and 5(g) the earthquake occurred on one branch of a triple junction and the aftershock on another (see Section 6). In the other case, Fig. 5(h) it is not the direction of motion between the plates which was different in the main, shock and the aftershock, but the strike and the dip of the fault plane. The other plots are of focal mechanisms and aftershock distributions discussed below. In all cases except the three already mentioned it was impossible to discover if the main shock had a different mechanism from the largest aftershock.

The main shock in Fig. 5(d) was at the eastern end of the fault break which probably propagated westward. The aftershock on 1967.7.30 had a completely different mechanism. The sequence of shocks in eastern Turkey in Fig. 5(g) shows a similar though less striking change. The foreshock of 1966.3.7 was followed by the main shock, the Varto earthquake, of 1966.8.19 with a considerable thrust component. On 1966.8.20 the main aftershock occurred on the North Anatolian fault, and the mechanism was pure strike slip. This sequence shows how the thrusting in eastern Turkey and the Caucasus changes to strike slip motion between the Turkish and Eurasian plates. The triple junction between these two and the Arabian plate lies between 1966.8.19 and 1966.8.20 (see below).

The other shock to show this behaviour is the Dasht-e-Bayaz earthquake, Fig. 5(h), on 1968.8.31. The main shock was strike slip but the major aftershock on 1968.9.1 was a thrust. In this case the slip vectors of all shocks were parallel but the fault changed direction. These observations suggest that fractures easily propagate only if the fault is a plane. If the fault has any sharp changes in direction then the propagating break stops and the resulting stress concentration produces aftershocks on the extension of the fault. Stauder (1968) suggested a similar explanation for the Rat Island earthquake and aftershocks. This idea is also supported by the limited occurrence of earthquakes above magnitude 8 in the Mediterranean region.

The two largest shocks recorded were the Lisbon earthquake of 1755 and the Great Turkish Earthquake of 1939. The first of these probably occurred on the Azores–Gibraltar ridge (Section 3), the second on a straight section of the North Anatolian Fault. The surface breaks and relocated aftershocks of the second of these earthquakes are shown in Fig. 5(f). The section which broke is considerably less complicated than either the eastward or westward extension of this fault (Section 6). In other parts of the Mediterranean region where shocks have occurred which have been mapped in detail the faulting is extensive but the faults are rarely continuous for large distances. This pattern is reflected in the seismicity, which consists of many earthquakes of magnitude 6·5–7·5 and no recorded larger events. If fault breaks cannot propagate round bends large earthquakes should be restricted to long straight faults, and this does indeed appear to be the case. Unfortunately formal statistics cannot be used to support this suggestion because very large earthquakes are rare everywhere, and reliable instrumental magnitudes are only available for the last 60 yr.

2. Earthquake mechanisms

All fault plane solutions in this paper for earthquakes which occurred during or after 1963 were determined from the polarity of the long period vertical components of the World Wide Standardised Seismograph Network (WWSSN), and all records of this type were read by the author. A few solutions could only be determined from the short period records, either because the earthquake did not generate long period waves of sufficient amplitude, or because it occurred when the long period instruments were disturbed by the surface waves from an earlier large earthquake. Solutions of this type are noted in the figure captions. The station distribution around the Mediterranean is sufficiently good that *S* wave polarization (Stauder 1962) need rarely be used to define fault planes. In a few cases, however, in Iran and in Yugoslavia the *S* wave polarization did have to be used to distinguish between strike slip and thrust solutions. Earthquakes during 1962 were not recorded by the WWSSN instruments at sufficient stations to permit solutions to be obtained using these observations alone. What observations there were were combined with short period observations from older non-standard instruments to obtain a solution. Before 1962 short period polarity measurements alone are available. These measurements were mostly obtained by the operator of each station, and are less reliable than they would be if they were read by one observer using all the available records. Furthermore short period polarities are less consistent than long period readings, partly because of interference due to crustal structure and partly because of the narrow frequency response of short period instruments. Therefore observations before 1962 are less consistent and less useful than those obtained from WWSSN records.

Two principal collections of short period observations used in this study are those of Wickens & Hodgson (1967) and of Canitez & Ucer (1967a, b). Wickens & Hodgson (1967) used a computer to determine a fault plane solution from the available observations, and this technique often produced apparently reliable mechanisms which were not in agreement with later WWSSN solutions in the same area. The reason for this disagreement is clear from Fig. 6, which shows a solution for a large earthquake in the Straits of Gibraltar carried out by the International Seismological Centre in Edinburgh using the Wickens–Hodgson method, but plotted on the standard equal area projection rather than the extended distance projection used by Wickens and by the ISC. This plot should be compared to the one in Fig. 10 obtained by the author from WWSSN seismograms, and with similar solutions carried out independently (Section 3). All the long period solutions are in substantial agreement that the mechanism is a thrust with a small component of strike slip. The ISC solution, however, is dominantly strike slip, principally because of a few inconsistent polarity observations and an absence of data in the NE quadrant. If the mechanism had been determined from the plot in Fig. 6 using an equal area net it is doubtful if the N–S plane would have been drawn. The solution would then have agreed rather well with the long period solutions. This example illustrates a general problem with the Wickens–Hodgson compilation. The number of discrepant observations in a fault plane solution is only a measure of its accuracy if the observations are equally distributed on the focal sphere. This requirement is rarely satisfied. In particular if distant stations alone are available and are unreliable, some dilatations and some compressions will be reported, and all will be close to the centre of the equal area plot.

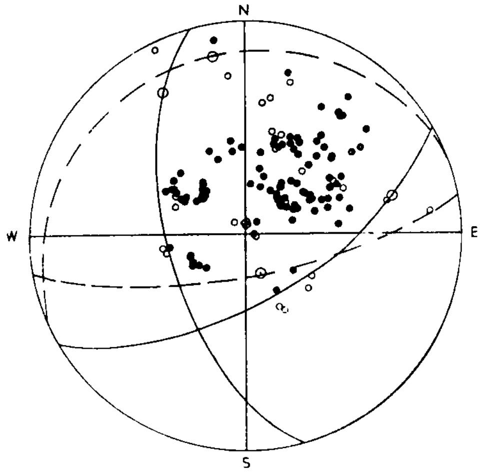


FIG. 6. Short period observations of polarity from the shock on 1964.3.15, published by the International Seismological Commission. The continuous curved lines are the nodal planes from the same source. The dotted nodal lines are from Fig. 10(c) and are based on long period observations. The disagreement between the two solutions is principally caused by the lack of short period observations in the NW and SW quadrants, combined with two inconsistent observations. If nodal planes had been drawn by hand it is unlikely that the N–S plane of the short period solution would have been drawn, then the agreement between the two solutions would have been good.

The program will attempt to separate compressions from dilations by using the two nodal planes, and in doing so will produce a strike slip solution regardless of whether this was in fact the earthquake mechanism. This phenomenon appears to be the explanation of the early belief that all the seismic zones around the Pacific were strike slip faults (Hodgson 1957). Fig. 6 shows that this difficulty may be avoided if the observations collected by Wickens & Hodgson are replotted on an equal area projection and nodal planes are drawn by hand. All mechanisms for earthquakes before 1962 were determined in this way, the polarity observations coming principally from the two sources already mentioned and from various other less extensive compilations mentioned in the captions. Where these early mechanisms can be compared with solutions for earthquakes recorded by the WWSSN the agreement is generally good. The explanation of the discrepancies discussed by Stevens & Hodgson (1968) is therefore that the mechanisms, not the observations, are incorrect.

A different problem makes the interpretation of even WWSSN fault plane solutions ambiguous for earthquakes in the eastern part of the region in Fig. 2. The depth of many of the shocks in Turkey and Iran determined from teleseismic observations lies within the continental crust. Since the locations of earthquakes in this region depend principally on distant stations the depth control is poor, the depths determined by the standard procedure are unlikely to be reliable to better than 50 km. It is, however, essential to know the P wave velocity at the focus if the position of a seismic station is to be plotted in the correct position on the focal sphere. All the equal area plots were first run on the assumption that the focus was in the Upper Mantle beneath the Moho with a P wave velocity of 8 km s^{-1}. In a considerable number of cases it was impossible to draw two mutually orthogonal planes to separate the compressional from the dilatational readings. Two such cases, one for a normal fault and one for a thrust are shown in Fig. 7. Though neither of these shocks produced an observed surface break, several earthquakes did so whose focal spheres resembled those in Fig. 7. The dominance of dilatations in one case and

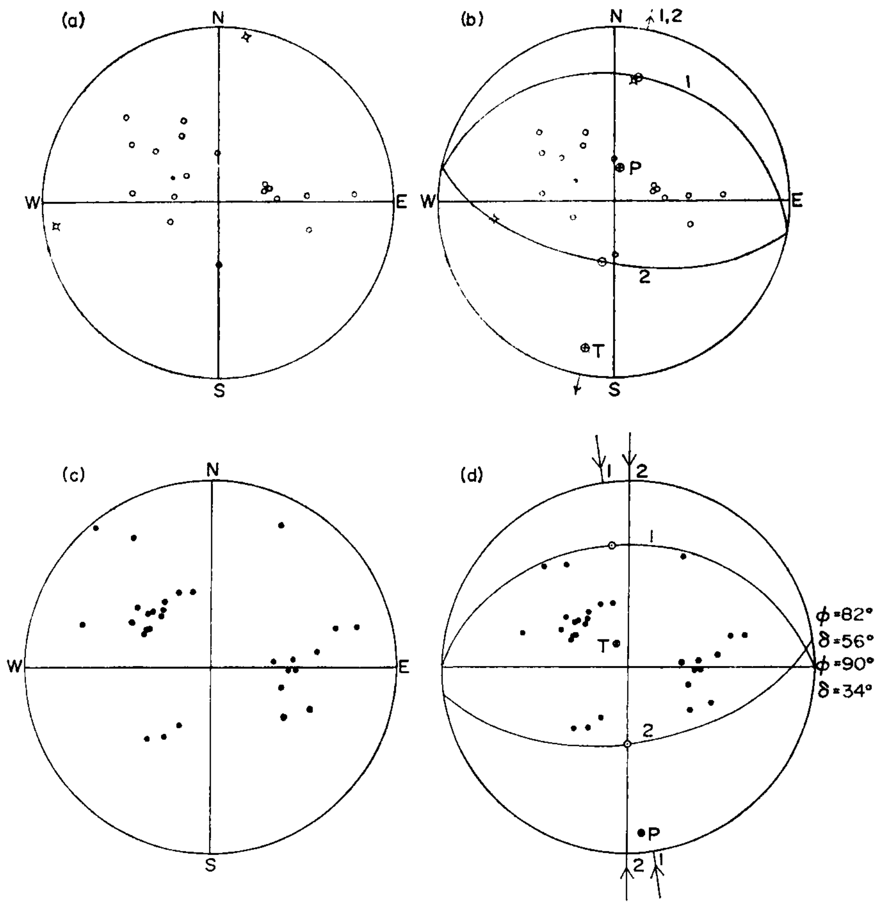


FIG. 7. Focal mechanisms of two crustal shocks. All mechanisms in this paper show an equal area projection of the lower hemisphere of the focal sphere (Sykes 1967). Long period polarity observations obtained from the vertical long period record at each station are shown as dots, open if the polarity was dilational (first motion downwards), solid if it was compressional (first motion upwards). If the polarity was not completely clear, or if it had to be determined from short period records it is shown as a small dot or circle in the figures. It is sometimes possible to recognize that a station is close to one of the nodal lines of the solution from the long period waveform itself, and such observations are shown as dot or a circle with a cross through it if the polarity could be read, or as a simple cross if this was not possible. All mechanisms in this study were consistent with a double couple as a source function, and therefore the dilatations could be separated from the crosses by two planes at right angles. These planes are shown as curved lines in the mechanism plots because of the projection of the focal sphere into a plane. They are generally labelled 1 and 2 arbitrarily. One plane is the fault plane and the other is the auxiliary plane, but it is not possible to determine which is which from the mechanism alone. The normals to the two planes are shown as circles with dots in the centre, and a necessary and sufficient condition that the two planes are at right angles is that each pass through the other's normal. The lines outside the circle with arrows and numbers on them show the horizontal projections of the slip vectors. If plane 1 is the fault plane, then the projected slip vector is in the direction labelled 1, with arrows pointing inward if the faulting is thrusting, outward if normal faulting occurred. Slip vectors are always at right angles to the auxiliary plane. The so-called Pressure and Tension axes bisect the angles between the normals to the two planes, and the Pressure axis lies in the dilatational quadrant, the Tension in the compressional. These axes are not usually related to the regional stress field in any simple manner. The two mechanisms (a) and (c) show the

projected focal spheres for two shocks on 1969.3.28 and 1966.9.18, obtained for foci in the upper mantle. It is clear that orthogonal nodal planes cannot be drawn for either solution. If, however, the foci are taken to be in the crust with a P wave velocity of $6 \cdot 8 \text{ km s}^{-1}$, then the focal spheres change to (b) and (d), when it is easy to draw orthogonal nodal planes. Many of the shocks in the eastern part of the region studied here resembled those in (a) and (c), but in all cases planes could be drawn if a P wave velocity of $6 \cdot 8 \text{ km s}^{-1}$ at the focus was used. Such solutions are noted as occurring in the crust in the figure captions.

compressions in the other need not therefore be due to implosive or explosive mechanisms, but to the use of an incorrect P wave velocity at the focus (Ritsema 1967). In both cases it became possible to draw two planes at right angles to each other if a velocity of $6 \cdot 8 \text{ km s}^{-1}$, rather than 8 km s^{-1}, was used (Fig. 7). The effect of this change in velocity is shown in Fig. 8. All stations plot nearer the centre of the projection because of refraction at the Moho. This effect shows clearly in the two plots of the world in the focal sphere in Fig. 8(b) and (c), the first using a source velocity of 8 km s^{-1}, the second using $6 \cdot 8 \text{ km s}^{-1}$. All mechanisms were first plotted using 8 km s^{-1}, then, if two mutually orthogonal planes could not be drawn, rerun using $6 \cdot 8 \text{ km s}^{-1}$. In all cases two planes could then be drawn. This success does not, however, mean that $6 \cdot 8 \text{ km s}^{-1}$ is the correct P velocity at the focus, since in most cases a lower P wave velocity would still permit two planes to be drawn, but with a larger component of strike slip than is shown. This problem is especially important in Iran, where faults may have a larger component of strike slip than that obtained from the fault plane solutions. This difficulty can only be overcome if the crustal velocity structure of Iran and Turkey can be accurately determined, and if the depth of earthquake foci can be obtained from observations of P wave arrival times at close stations. Until such observations can be carried out the amount of strike slip motion involved cannot be determined with confidence from teleseismic observations. These remarks do not apply to solutions which are dominantly strike slip, which are not affected by the crustal velocity.

3. Azores to Sicily

Gutenberg & Richter (1954) observed that an active seismic zone extends from the Mid-Atlantic ridge close to the Azores eastwards towards the Straits of Gibraltar and across North Africa to Sicily. This belt can be seen in Fig. 1, though it is not sufficiently active to show clearly using only ten years' data. Fig. 2(a), however shows all large shocks since 1925 and clearly outlines this active belt. Some very large earthquakes have occurred on this boundary, especially west of Gibraltar. Several studies have been made of the seismicity and fault plane solutions of this active belt. Di Filippo (1950) carried out a fault plane solution for the shock on 1941.11.25, and Udias (1967), Shirokova (1967), Udias & Arroyo (1970), Ritsema (1969) and Banghar & Sykes (1969) have also examined the mechanisms of earthquakes in this region. The seismicity of the belt was discussed by Gutenberg & Richter (1954), and more recently by Munuera (1965), and by Barazangi & Dorman (1969). Two surface breaks from earthquakes in this area have been observed. The first occurred in Algeria as a result of the Orleansville earthquake (1954.9.9) and was mapped by Rothé (1955), the second was observed by Le Pichon *et al.* (1971) on the sea floor south of Spain and was presumably the result of the 1969.2.28(a) shock in this region. The active belt defines the plate margin between Eurasia and Africa and Le Pichon (1968) proposed that the activity was the result of rotation of Africa relative to Europe about a pole at $9 \cdot 3°$ N, $-46°$ E. This pole produced right lateral strike slip in the Azores and overthrusting in North Africa, in general agreement with the long period mechanisms published by Banghar & Sykes (1969), but not with Ritsema's (1969) short period solutions in North Africa. Ritsema proposed that the

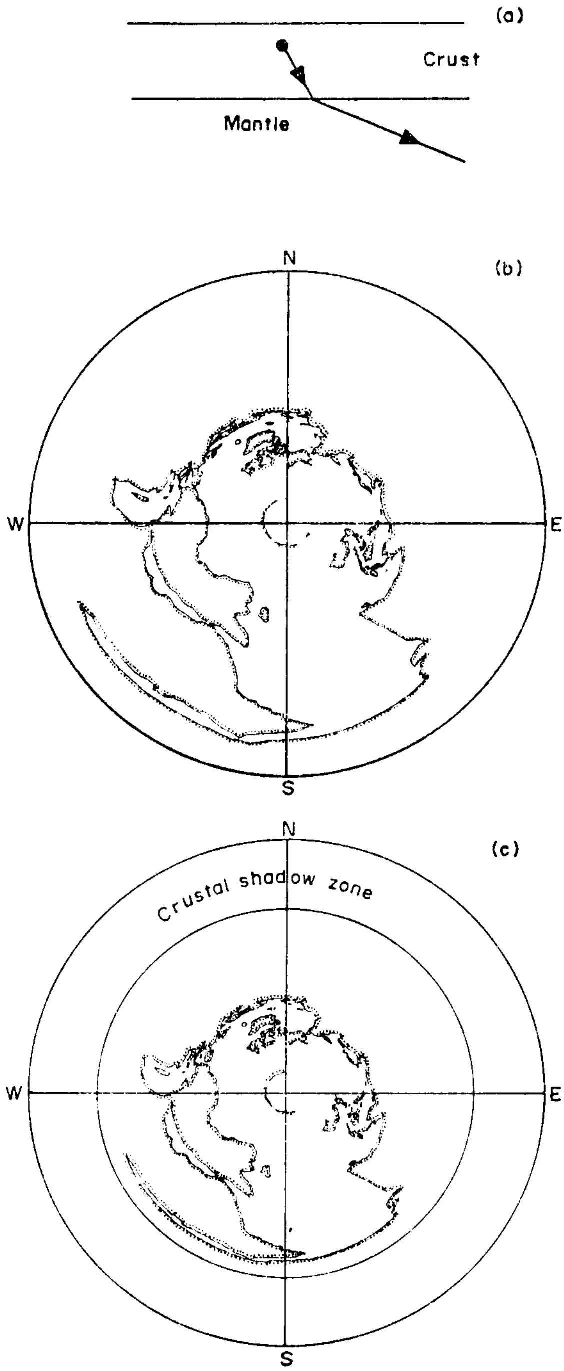


FIG. 8(a). Ray diagram to illustrate the consequence of a shock occurring in the crust. Because the ray is bent away from the normal on passing into the mantle only part of the focal sphere can be examined using teleseismic records. (b) and (c) Show the coast lines of the World plotted in the projection of the focal sphere for a shock in NE Iran (Davies & McKenzie 1969). In (b) the focus is in the mantle, and shows the shadow zone caused by the core near the centre of the plot. (c) shows the same plot but with a P velocity of $6 \cdot 8$ km s^{-1} at the focus. There is now a shadow zone due to the crust at the edge of the plot (see also Ritsema 1967.)

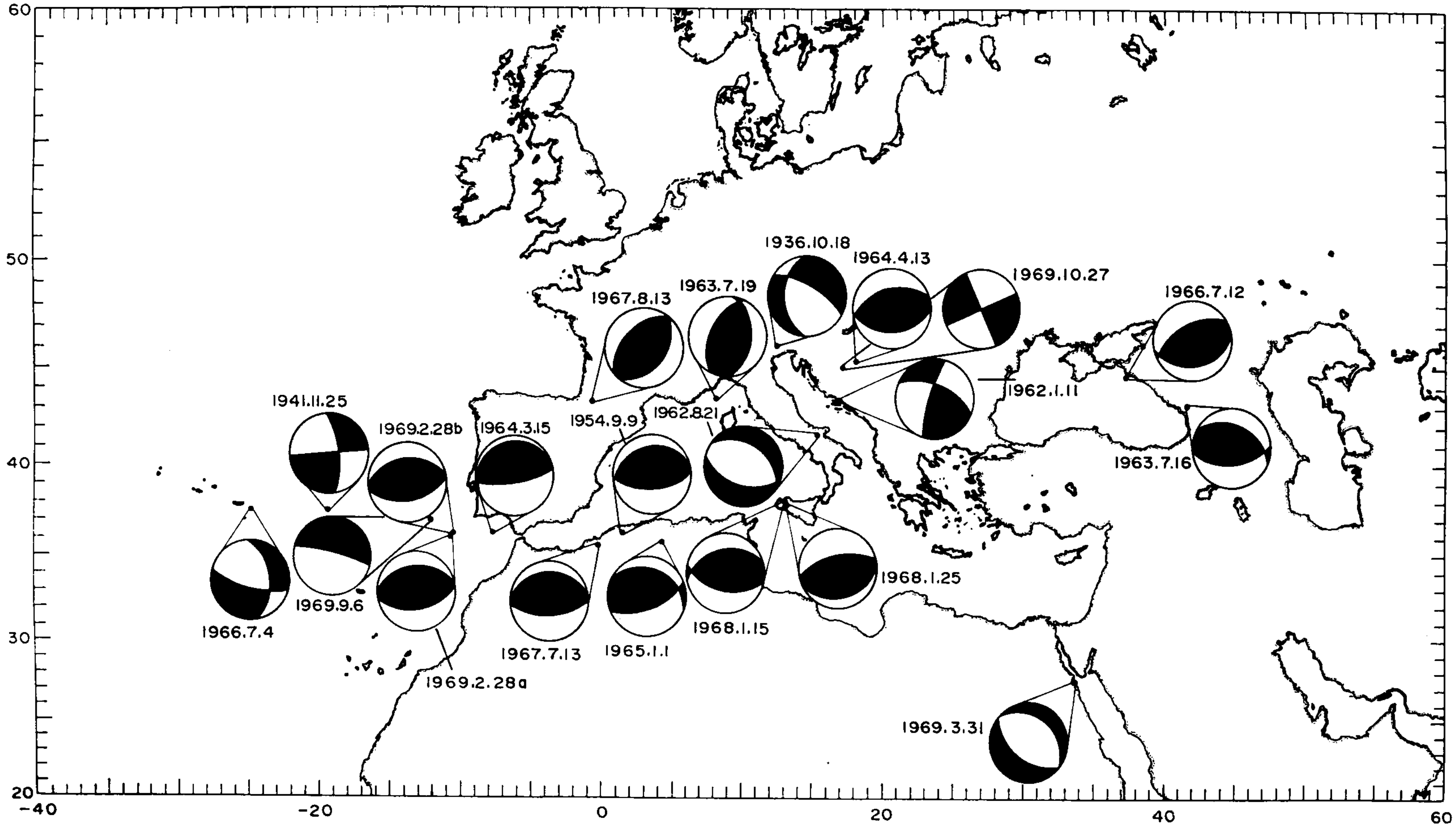


FIG. 9. Earthquake mechanisms in the Western Mediterranean, Eastern Atlantic and Black Sea regions. Conventions as for Fig. 5. See Table 3 for locations and mechanism figure numbers.

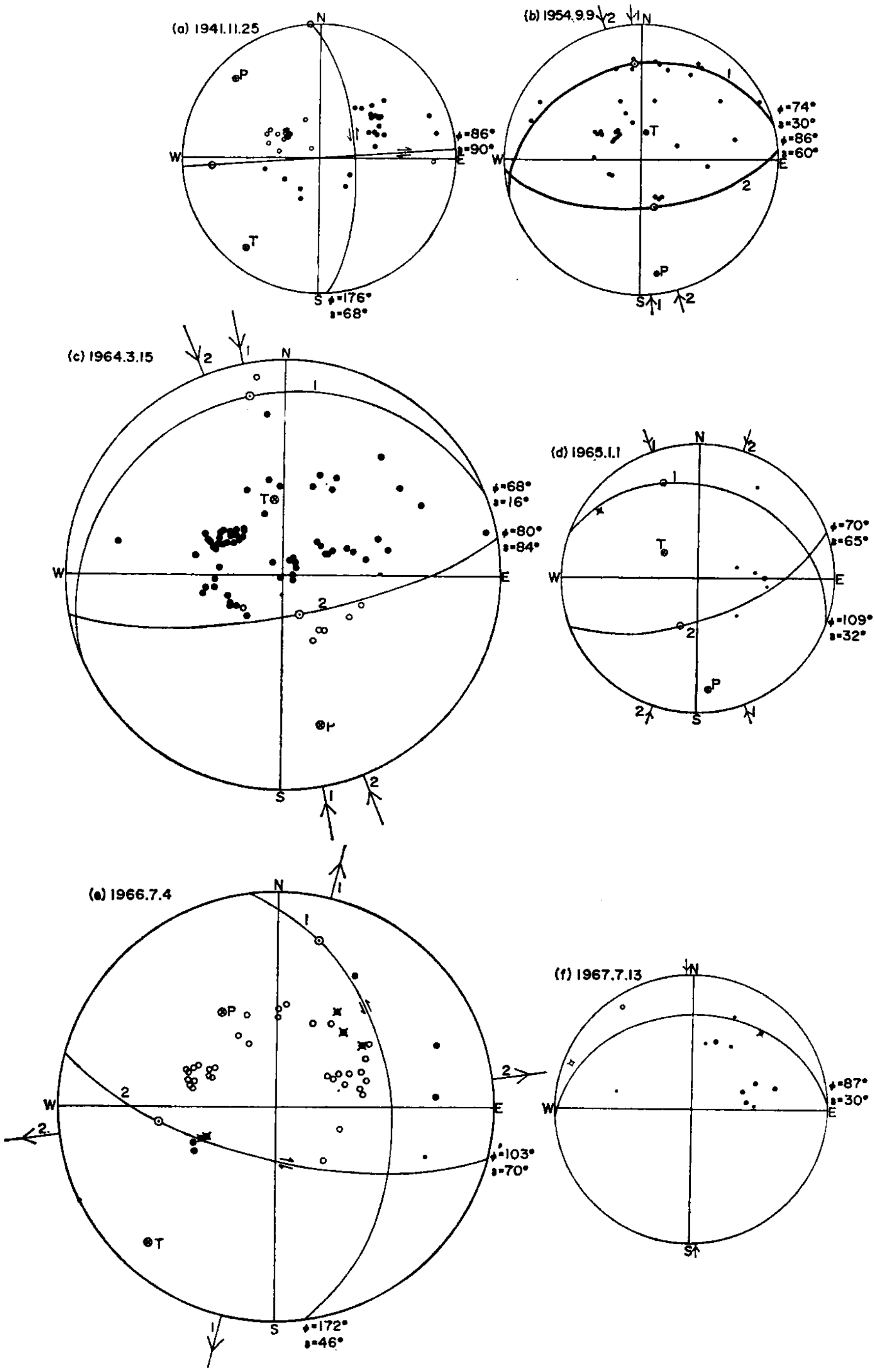

(a) 1941.11.25
N
W
E
S
P
T
ϕ=86°
δ=90°
ϕ=176°
δ=68°
(b) 1954.9.9
ϕ=74°
δ=30°
ϕ=86°
δ=60°
(c) 1964.3.15
ϕ=68°
δ=16°
ϕ=80°
δ=84°
(d) 1965.1.1
ϕ=70°
δ=65°
ϕ=109°
δ=32°
(e) 1966.7.4
ϕ=103°
δ=70°
ϕ=172°
δ=46°
(f) 1967.7.13
ϕ=87°
δ=30°

FIG. 10(a)–(f)

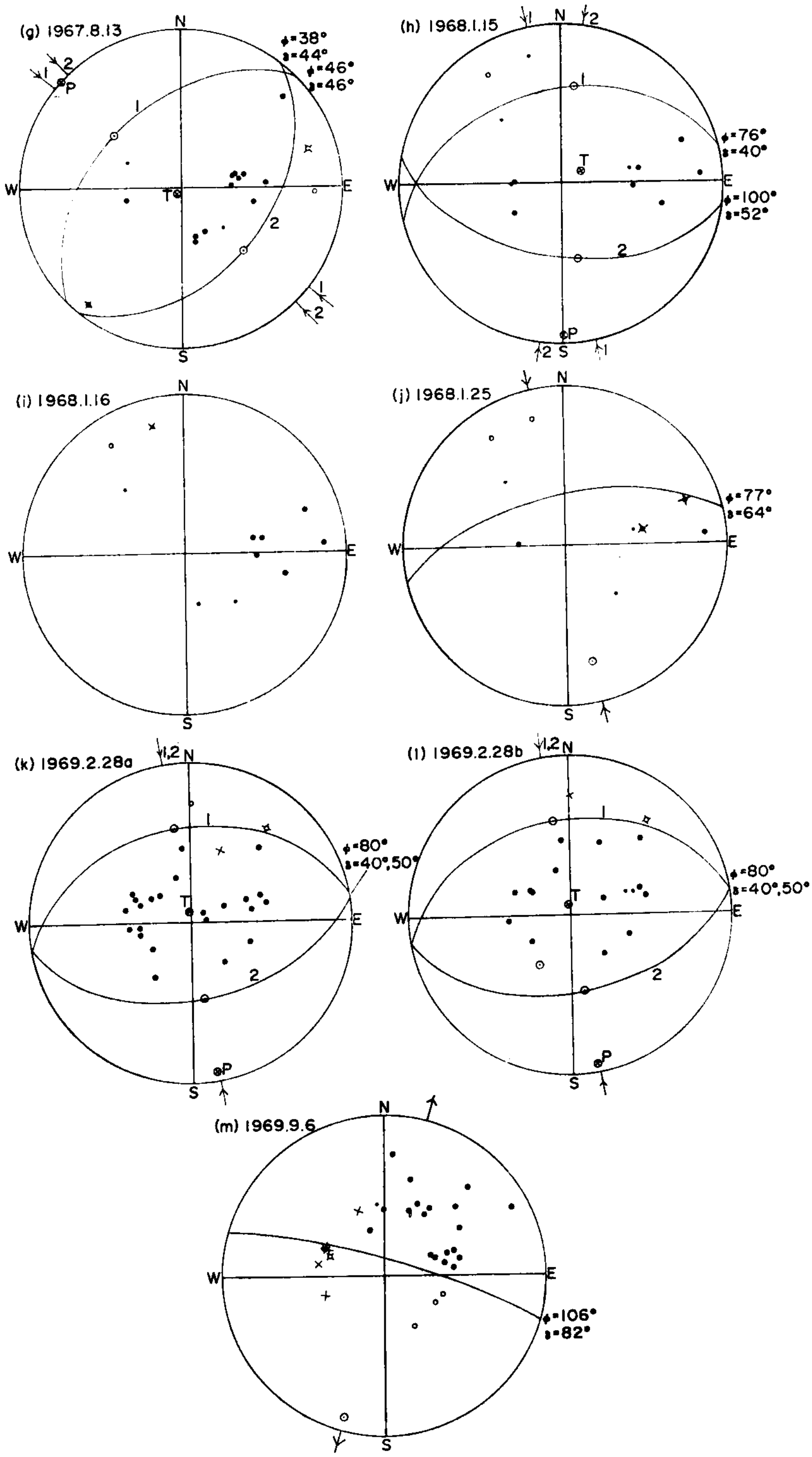


FIG. 10. Solutions for earthquakes in Fig. 9 in and west of Sicily. All mechanisms in the upper mantle, and all except (a), (b), (g) and (l) based on long period WWSSN observations. Di Filippo's (1950) polarity observations were used for (a), those of the ISS for (b). Short period WWSSN seismograms were read by the author to obtain (g) and (l). For an account of the conventions used in the mechanisms see the caption of Fig. 7.

entire boundary between the Azores and Sicily is a right lateral transform fault, a conclusion that principally depends on the short period focal mechanism of the Orleansville shock which is discussed in detail below.

The mechanisms obtained in this study are shown in Figs 9 and 10. The arrivals were clear and easy to read on 1966.7.4, 1969.2.28(a), 1964.3.15 and less clear for the other seven shocks after 1963. The ambiguity between the fault plane and the auxiliary plane cannot be resolved for any of these shocks by using the surface break or the aftershock distribution. In the case of 1966.7.4 the seismic zone in its vicinity is approximately parallel to plane 2, and if this is chosen to be the fault plane the slip vector is in good agreement with that calculated from the pole position of Le Pichon (1968) and that in Table 1. Udias (1971) has, however, applied Ben Menahem's (1961) technique to the surface wave radiation pattern of this earthquake and shown that the choice of 1 as the fault plane produces a better fit to the observations than does 2. This choice is not compatible with any simple interpretation of the tectonics of the region and is not in agreement with more recent work on this shock by the same author (Udias, private communication). Whichever plane is the fault plane the mechanism has an appreciable normal faulting component. Di Filippo's (1950) short period observations have been used for the mechanism 1941.11.25. The east–west plane 2 is parallel to the seismic zone and is probably the fault plane. The slip vector is then also east–west and right lateral. The shock of 1969.2.28(a) (Fig. 5(a)) which was the largest event which occurred in the region covered by Fig. 1 in the last 10 yr, produced clear observations at all stations, but because of its mechanism the strike of neither of the planes is well defined by the *P* wave solution. Furthermore all seismographs were off scale before the *S* waves arrived, so their polarization cannot be used to restrain the solution. Fukao (private communication) has studied the surface wave radiation from 1969.2.28(a) and has been able to determine the strike and dip of the nodal planes with more accuracy than is possible using the *P* waves alone. He has shown that the northward dipping plane strikes about 55° E, whereas the southward dipping plane strikes 81° E. Furthermore he has used the locations of aftershocks to show that the northward dipping plane is the fault plane, in agreement with seismic reflection records (Le Pichon, private communication). The slip vector is therefore north–south and the shock was caused by Gorringe Bank thrusting over the sediments to the south. The first aftershock 1969.2.28(b) has a similar mechanism, but what may be the second is quite different. Its mechanism is drawn as normal faulting, but is not sufficiently well defined to rule out thrusting; a more probable mechanism in this region. The large shock which occurred on 1964.3.15 close to the Straits of Gibraltar has been studied by Udias (1967), Banghar & Sykes (1969), Udias & Arroyo (1970), and Udias (1971). All solutions closely resemble each other, but even so there is probably an uncertainty of 15° in the strike of the steeply dipping plane, which, by analogy with the mechanisms in island arcs, is most likely to be the auxiliary plane. The strike of the other plane, labelled 1 in Fig. 10, is poorly defined.

The shocks in Algeria since 1963 have all been small, and little can be said about 1967.7.13 and 1965.1.1 except that they are both compatible with overthrusting and have been drawn as such on Fig. 9. Though these two solutions are poor the observations are not compatible with right-handed strike slip on an east–west plane. One large shock, the Orleansville earthquake on 1954.9.9 occurred before the world-wide network was installed. The earthquake produced a number of surface breaks which were described by Rothé (1955) in some detail. The major fracture was east–west and about 6 km long, and was approximately an arc of a circle of similar radius. In places there was considerable vertical displacement of up to 1 m across this fault, but in other parts the fault broke up into many small fissures. It is not clear from Rothé's description whether the sense of motion of the southern block was consistently up or down relative to the northern part, but his maps and photographs suggest that the displacement occurred on a fault dipping to the south, and that the fault was a thrust. Other fault breaks associated with this shock occurred further west, and formed a series of north–south lines. Rothé did not find any evidence of consistent vertical motion across this system of faults, though he describes a graben formed by two parallel breaks. The important question is whether these surface breaks are of tectonic origin or merely mark the boundaries of a large landslide. Rothé studied the breaks while they were still fresh and believes that most are the result of an incipient landslide (1955 and private communication), whereas Ambraseys (private communication) examined the fractures some years after they were formed and believes that they could mostly be of tectonic origin. The discussion below assumed that Ambraseys' interpretation is correct; if this assumption is wrong then the surface breaks are superficial and are irrelevant to the region tectonics.

The surface breaks of this shock do not resemble those of strike slip earthquakes such as the Mudurnu shock (1967.7.22) in Western Turkey (Ambraseys & Zatopek 1969) or the Dasht-e-Bayaz earthquake (1968.8.31) in eastern Iran (Ambraseys & Tchalenko 1969). Both these earthquakes produced long and relatively straight surface breaks and their fault plane solutions show that the displacement was strike slip with no extension or shortening. The surface break of the Orleansville shock closely resembles that of the San Fernando earthquake, which occurred on the north side of the Los Angeles basin and whose surface break was mapped in considerable detail by the Division of Geological Sciences, California Institute of Technology (1971). The San Fernando earthquake involved about equal thrust and left-handed strike slip movements, and the surface displacements did not follow a straight fault but were offset by a complicated series of north–south faults. The fault plane solution of the San Fernando shock agreed excellently with the surface motions. That for the Orleansville event is shown in Fig. 10(h). The first motions are taken from the ISS bulletin, and are based on short period records. The mechanism based on these points must contain a large thrust component in agreement with the solution published by Shirokova (1967). Ritsema (1969), however, believes that the right-handed strike slip motion which occurs at the western end of the Azores–Gibraltar ridge at its junction with the Mid-Atlantic ridge continues across North Africa. The direction of motion obtained in this paper differs from that of Ritsema (1969) because he assumed that the Orleansville shock occurred in the crust where the velocity was $6 \cdot 25\ \mathrm{km\,s^{-1}}$, and hence was able to obtain a solution with a considerably larger component of strike slip motion (Ritsema private communication). Ritsema's other solutions between Gibraltar and Sicily are poor, and therefore there are no observations in conflict with the direction of motion obtained from the Africa–Eurasia pole (Table 1, Fig. 4).

Three small earthquakes 1968.1.15, 1968.1.16 and 1968.1.25, occurred in Sicily at the eastern end of this active belt. The mechanisms of the first and the last shocks are clearly thrusts, and all observations are consistent with all three shocks being the result of thrusting on east–west planes. The ambiguity between the two planes cannot be resolved for any of these events. The mechanisms of shocks between the Azores and Sicily has been discussed in considerable detail because this is the only part of the seismic belt between Africa and Europe where the mechanisms directly reflect the motion between these two plates. Since any interpretation of the complications east of Sicily must depend on a knowledge of the motion between Africa and Eurasia it is essential to test the pole obtained from magnetic lineations, focal mechanisms and fracture zones in the Atlantic against the seismic observations of the active belt. It is clear from the discussion above that all reliable mechanisms at present available in this area are consistent with the pole of motion between Africa and Eurasia.

As Fig. 1 clearly shows the seismicity is not restricted to the major active zone across North Africa. Though some of the events not on this major line have done

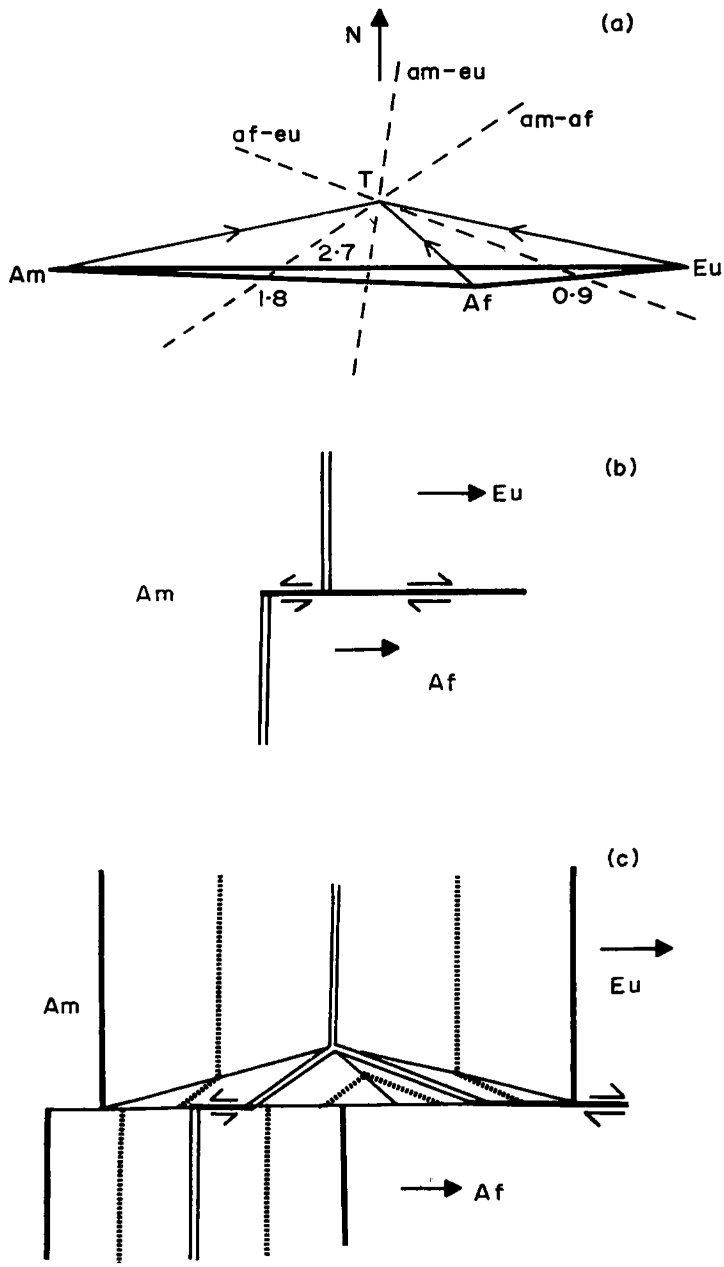


FIG. 11(a). The Azores triple junction in velocity space obtained from the poles in Table 1. The notation follows that of McKenzie & Morgan (1969) and the strikes of the ridges were obtained from Fig. 1. Notice that the ridges are not at right angles to the directions of relative motion, and therefore the junction can be stable for many different ridge orientations. If the junction was once a Fault-Fault-Ridge (FFR) junction like that in (b) it could have changed to an RRR junction in (c) without any change in the motion directions between the plates provided oblique spreading can occur.

considerable damage, such as the Agadir earthquake of 1960.2.29, they have all been of small magnitude. The only solution available is from the Pyrenees on 1967.8.13, and suggests that the thrusting which formed these mountains still continues in a minor way. There is, however, no evidence for major motions except in North Africa and on the Azores–Gibraltar Ridge.

No mention has yet been made of two famous earthquakes which occurred south of Spain. The first of these is the Lisbon earthquake of 1755 November 1, which has been discussed by many authors with little knowledge of seismology. The seismological observations have, however, been analysed with some thoroughness by Reid (1914), Davison (1936) and Richter (1958). It is the only earthquake in the Atlantic known to have produced a destructive tsunami, and was therefore most probably the result of movement on a thrust fault (Isacks, Oliver & Sykes 1968). Richter estimates the magnitude of this shock as between 8·5 and 8·75, principally because of the area over which seiches were excited. This value is supported by the field effects of the shock which occurred on 1969.2.28, which had an instrumental magnitude of about 8 yet produced no large tsunami or seiches. In the absence of instrumental records the epicentre of 1755 earthquake cannot be accurately located. Reid believes that the observations are consistent with a location of 38° N, $-10°$ E, but Davison argues that a large event must have occurred much closer to Lisbon, even though this may not have been the location of the main shock. In the absence of good observations of the time which elapsed between the arrival of the earthquake waves and the tsunami at different points on the North African and Portuguese coasts, an absence which is indeed scarcely surprising in view of the damage done by the shock in these regions, the most likely location of the event was on the Azores–Gibraltar ridge.

The other important earthquake near Spain occurred on 1954.3.29 at a depth of 630 km. Its mechanism is well determined (Fig. 23) and was studied by Hodgson & Cock (1956), and by Di Filippo & Peronaci (1959). The depth of the focus is very well determined, since many of the European stations lay in the upper hemisphere of the focal sphere. The magnitude of the shock was about 7, yet no aftershocks were recorded. No other deep earthquakes have been recorded from this region. Isacks & Molnar (1971) discussed this shock and its mechanism and suggested it could have been caused by an isolated slab of lithosphere produced by underthrusting near the Straits of Gibraltar. The present motion could not produce a slab dipping east as required by the mechanism and therefore if such a slab exists it must have been produced by the motion of small plates in the western Mediterranean which has now ceased. Any proposed history of the evolution of this part of the Alpide belt during the Tertiary must account for the existence of this shock.

The mechanisms in this section are all consistent with the poles in Table 1, and therefore these poles may be used to discuss the evolution of the Azores triple junction. An attempt to do so has previously been made by Krause & Watkins (1970). They did not use the methods proposed by McKenzie & Morgan (1969), and produced a remarkably complicated evolutionary history to explain their observations, which was incorrect in one important respect. The problem is of considerable importance, since the evolution of this junction can be used to study the history of the relative motion between Africa and Eurasia, which in turn is essential to the tectonic history of the Mediterranean. For this reason the triple junction is discussed here, even though its evolution is not relevant to the instantaneous motions. The present motions of the plates at the Azores triple junction are best obtained from the poles in Table 1. They are listed in Table 2, as also are the strikes of the plate boundaries measured from the strike of the seismic zones in Fig. 1. The plate velocities and boundaries are shown in Fig. 11(a) using the conventions of McKenzie & Morgan (1969). The plate boundaries pass rather accurately through T. Therefore the triple junction is stable and T marks its motion in velocity space relative to

Table 2

Motions at the Azores triple junction

Fixed		Moving	Rate (cm/yr)	Direction
Eurasia	—	Africa	0·9	264°
Eurasia	—	America	2·7	269°
Africa	—	America	1·8	271°

Strikes of ridges at the Azores triple junction

Eurasia	—	America	7°
Eurasia	—	Africa	111°
Africa	—	America	55°

the three plates, and the lines Am–T, Af–T and Eu–T show the relative motion between each of the three plates and the triple junction. There are two surprising features about Fig. 11(a). The relative motion between the three plates is in all cases taken up by a ridge, but only in the case of America–Europe is the plate boundary even approximately at right angles to the motion direction. The earthquake mechanism for 1966.7.4 shows that the Africa–Europe boundary is truly oblique where the earthquake occurred and does not there consist of an echelon arrangement of ridges and transform faults. The other feature of Fig. 11(a) is that the evolution of the triple junction shortens the ridge between America and Europe as it evolves. This unexpected behaviour is shown more clearly in Fig. 11(c), which shows the arrangement of the magnetic lineations which would be produced by the evolution of the junction. Such shortening is always present if T lies outside the velocity triangle in Fig. 11(a) and is not a consequence of the oblique spreading.

The evolution of this triple junction is dominated by the oblique spreading of all the ridges involved. Oblique spreading was not possible for the RRR (ridge–ridge–ridge) junction of McKenzie & Morgan (1969) because they only discussed ridges which lay at right angles to the motion direction. Once this restriction is relaxed T may lie anywhere in the plane Am, Eu, Af, and in particular its position may change with time. For instance since the motions between the three plates are all in the same direction within experimental error, the Azores junction could change to the FFR junction (fault–fault–ridge) junction in Fig. 11(b) without any change in the plate motions. Krause & Watkins (1970) believed that this junction was originally an FFR junction like that in Fig. 11(b) and has recently become an RRR junction. Though their evidence could be differently interpreted, even if they are correct this change with time does not, as they supposed, require a change in plate motions.

This discussion shows that the evolution of this part of the Atlantic is dominated by the position of the point T in Fig. 11(a), yet there are at present no criteria for preferring one position to another. Without such rules it is likely to be difficult to determine the history of the relative motion of the three plates involved from the shape of the magnetic lineations produced by this junction.

4. Italy, North and Central Yugoslavia

The seismicity of this region is moderate compared to that of Greece and Turkey further east, but large earthquakes in Italy (Davison 1936) have in the past done a considerable amount of damage, principally due to the poor construction of the houses. A large number of papers have been published on the focal mechanisms of Italian shocks in the past (Di Filippo & Marcelli 1951, 1954; Di Filippo & Peronaci 1959, 1961, 1962a, b, 1963, 1964), but unfortunately most of the observations are from close short period stations and therefore few of the mechanisms are reliable. Two previous attempts have been made to discuss the regional tectonics of this part

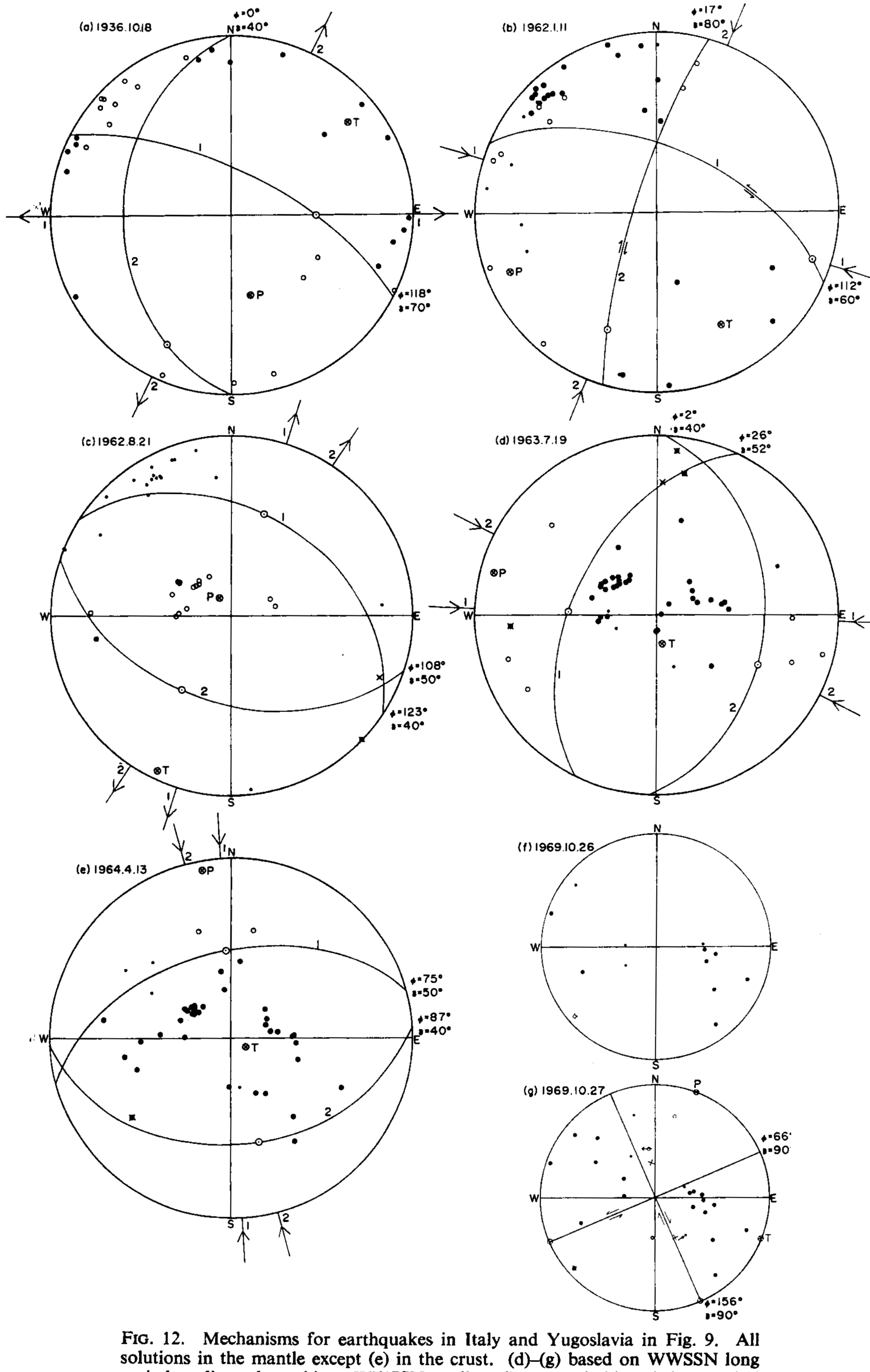


FIG. 12. Mechanisms for earthquakes in Italy and Yugoslavia in Fig. 9. All solutions in the mantle except (e) in the crust. (d)–(g) based on WWSSN long period readings alone, (c) on WWSSN readings (large symbols) and short period readings from Di Filippo & Peronaci (1963). (b) on short period observations only from the same authors, and (a) from Di Filippo & Peronaci (1962a). For account of the conventions used in the mechanisms see the captions of Fig. 7.

of the Mediterranean. Ritsema (1969) used the fault plane solutions which have been published and believed that normal faulting predominates in Italy, and strike slip and overthrusting in Yugoslavia. This faulting pattern is consistent with that in Fig. 9. Lort (1971) argued that the Adriatic forms a small plate moving north-west relative to Europe, and believed that this motion could account for the seismicity and the faulting. The fault plane solutions in Fig. 9 and Fig. 12 contain three reliable solutions, 1963.7.19, 1964.4.13 and 1969.10.27. Of the earlier shocks only 1962.8.21 is partly based on long period observations, all of which are dilatations. This mechanism must therefore be dominantly normal faulting, though neither of the nodal planes is well controlled. Little confidence can be placed in the other two solutions.

The information at present available is not sufficient to propose any plate tectonic interpretation with any confidence. The seismicity in Figs 1 and 2(a), and the fault plane solutions in Fig. 9 are consistent with Ritsema's and Lort's interpretation, and suggest that the motion between Africa and Eurasia is not taken up by thrusting on an east–west feature at the southern end of the Adriatic but by normal faulting in Italy and thrusting in Yugoslavia. A sketch of a plate tectonic scheme consistent with the observations is shown in Fig. 13. This region is probably much more complicated than is shown by the available observations, and has a relatively low level of seismic activity. Fault plane solutions of future shocks will be of great interest, and may well change the interpretation in Fig. 13.

Though the observations in this section are of relatively little use from the point of view of plate tectonics, they are sufficient to show that present motions cannot account for the important features of the Calabrian Arc: the intermediate depth shocks beneath the Tyrrhenian Sea and the associated vulcanism in Italy and Sicily (Fig. 14). Gutenberg & Richter (1954) noticed that small shocks occurred fairly frequently at depths up to 350 km, and Peterschmitt (1956) observed that the foci lay on a cone which dipped westward beneath Italy and Sicily, and intersected the Earth's surface beneath the Adriatic to the east. These earthquakes have since been relocated by Ritsema (1971) and both the improved locations and the USCGS

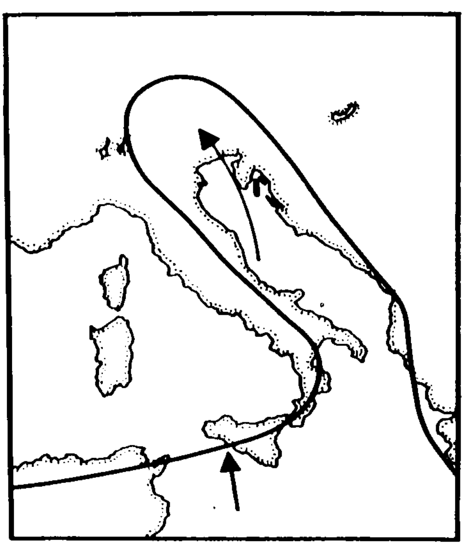


FIG. 13. Sketch of the possible motion of the Adriatic. The seismicity is too weak to define the plate boundaries, and the fault plane solutions are too scattered to show the motion clearly. Hence this arrangement of plate boundaries is compatible with the data, but may well be incorrect.

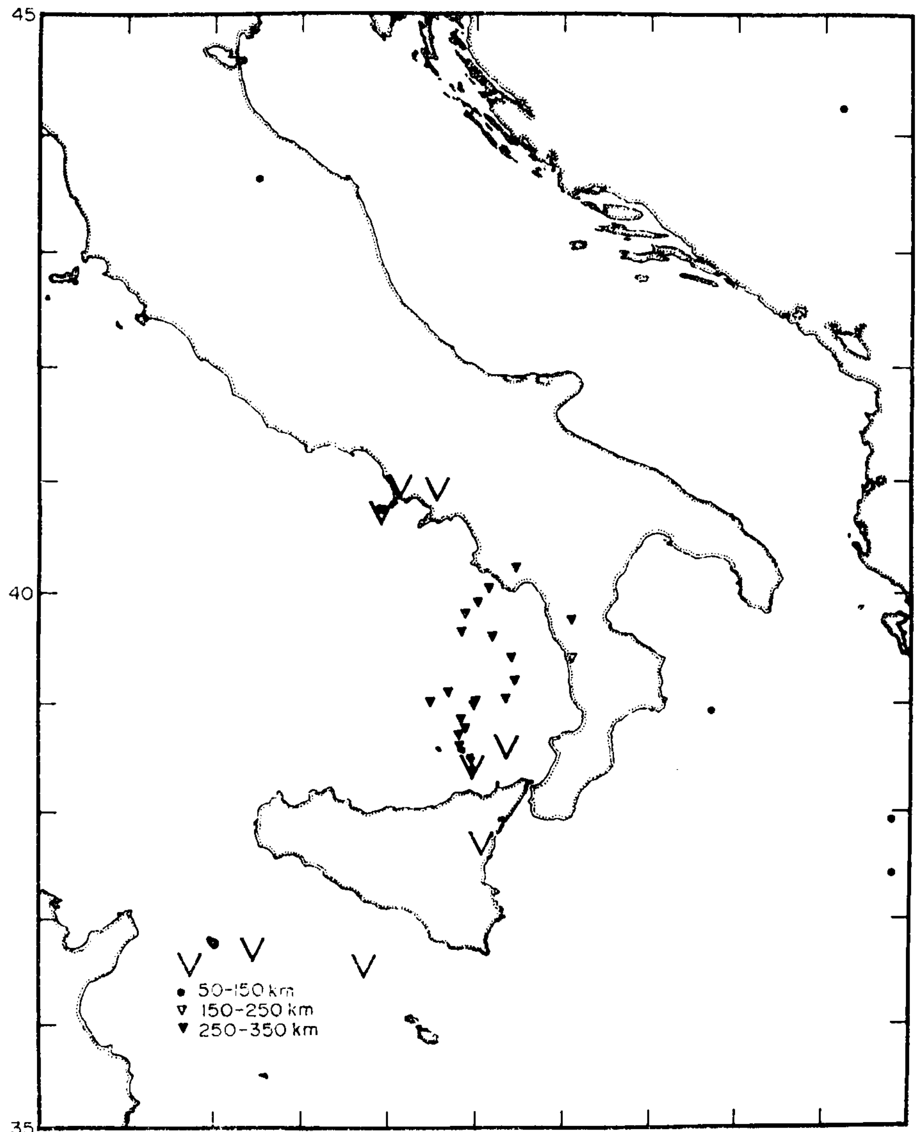


FIG. 14. Epicentres of shocks deeper than 50 km beneath the Tyrrhenian Sea, with different symbols to mark the depth ranges. Volcanoes are shown as V and are from Gutenberg & Richter (1954).

hypocentres support Peterschmitt's interpretation. There is, however, an absence of shocks between 50 and 200 km. This gap could either be caused by the absence of the dipping slab in this depth range, or by an absence of sufficient stresses within the slab to produce earthquakes. Until detailed studies of travel times have been carried out it is not possible to choose between these two possibilities. Though most of the shocks occur within a well-defined slab, one shock on 1955.2.17 was located by Ritsema (1971) further west and deeper (450 km) than any others in the region. The location is accurate and the depth control good, therefore this earthquake resembles the deep shocks beneath the North Island of New Zealand (Hamilton & Gale 1968) and does not obviously form part of the shallower structures.

Since 1963 no large intermediate shocks have occurred beneath the Tyrrhenian Sea. Ritsema (1969) has, however, collected short period observations on earthquakes before 1963, and Isacks & Molnar (1971) have shown that these observations imply that the sinking slab is in compression along its dip at the depth where the shocks occur. In this respect the dipping seismic zone closely resembles similar zones on the Pacific margins.

There is, however, one major difference between the intermediate activity beneath Italy and that of the margins of the Pacific. Figs 4 and 13 show the directions of plate motions relative to the Tyrrhenian Sea, a part of the Eurasian plate. These plate motions cannot account for the observed distribution of activity. If the slab were the result of the motion of Africa northwards it would not be restricted to beneath the Tyrrhenian Sea, but would extend along the whole plate margin as does the activity from Tonga to the South Island of New Zealand (Isacks *et al.* 1968; McKenzie 1969a). Also if the thermal time constant of the slab is about 10 My, at the present rate of consumption the maximum depth of shocks should be 150 km rather than the observed depth of about 350 km. Though it is possible to argue that a piece of the slab has broken off and sunk to a greater depth than it would otherwise have reached, this explanation cannot account for the localization of the activity and the westerly dip of the slab. The northward motion of Africa can produce a slab which strikes east–west but not one which strikes north–south. The difficulty is not removed by the complications suggested in Fig. 13. These arguments therefore suggest that the slab is a relic, and was produced by the motion of Italy and part at least of Sicily in an eastward direction with respect to Eurasia with a velocity considerably greater than that between Africa and Europe. The restricted region in which intermediate shocks now occur can then be explained if the rapidly moving plate was relatively small. If this argument is correct then the existence of a slab composed of what was formerly oceanic lithosphere may be used to study the evolution of this part of the Mediterranean (see Section 8 below). A general rotation of Italy is also implied by the palaeomagnetic observations of Van der Voo & Zijderveld (1969).

The other curious feature of the tectonics of Italy is the composition of the lavas erupted by the volcanoes, especially those near Vesuvius. The igneous rocks in this region often contain high concentrations of potassium and sodium (Washington 1906), and are quite unlike the volcanic rocks of the Pacific arcs. In terms of the tectonic evolution of the area outlined above there appear to be two differences between the Calabrian arc and the larger Pacific arcs. The thickness and composition of the sediment beneath the Mediterranean sea floor is much thicker and of different composition (Ryan *et al.* 1970) from that beneath the Pacific. Since many geologists now believe that the sediments are carried down with the sinking slab (Dickinson 1970), the composition of the volcanic rocks should be affected by that of the sediments carried down. This explanation is unlikely to account for the composition of the alkali-rich Italian rocks because the Aegean arc is consuming the same Mediterranean sea floor, but is not erupting alkali-rich volcanic rocks from the volcanoes in the Aegean and in Turkey (Washington 1926). Therefore the alkali-rich character of the Calabrian arc vulcanism is probably related to the change in the motion directions of the plates involved in this area, which has in consequence changed the thermal structure of the sinking slab and the rate of magma generation.

5. The Aegean and surrounding regions

There have been a large number of moderate sized earthquakes since 1961 in the active belts surrounding the Aegean, and therefore more is now known about the earthquake mechanisms in this region than in any other region in Figs 1 and 2. The seismicity of this area was studied by Gutenberg & Richter (1954), but at that time

Table 3

Year/month/day	Time h m s	Lat °N	Long °E	Depth (km)	Magnitude	Fig.
1969.10.27	8 10 58·3	44·92	17·23	33	5·3	12(g)
1969.10.26	15 36 51·8	44·87	17·29	33	5·3	12(f)
1969.10.13	1 2 28·5	39·86	20·64	8	5·6	20(α)
1969.9.6	14 30 39·5	36·94	−11·89	33	5·7	10(m)
1969.7.8	8 9 17·5	37·56	20·28	33	5·4	21(t)
1969.6.12	15 13 31·1	34·40	25·06	25	5·8	21(s)
1969.4.30	20 20 31·8	39·16	28·58	9	5·1	20(z)
1969.4.16	23 21 4·9	35·34	27·77	45	5·2	21(r)
1969.4.6	3 49 33·5	38·50	26·42	14	5·5	20(y)
1969.3.31	7 15 54·4	27·67	33·99	33	6·0	*
1969.3.28	1 48 30·4	38·59	28·45	9	6·0	20(x)
1969.3.25	13 21 32·4	39·18	28·37	23	5·6	20(w)
1969.3.23	21 8 42·6	39·16	28·48	12	5·6	20(v)
1969.3.3	0 59 10·5	40·12	27·43	4	5·6	20(u)
1969.2.28b	4 25 36·9	36·23	−10·48	33	5·7	10(l)
1969.2.28a	2 40 32·5	36·01	−10·57	22	7·3	10(k)
1969.1.14	23 12 7·9	36·18	29·20	33	5·5	21(q)
1969.1.3	3 16 38·1	37·13	57·90	11	5·6	32(p)
1968.12.5	7 52 11·0	36·58	26·97	35	5·5	21(p)
1968.9.14	13 48 31·2	28·44	53·11	33	5·8	32(o)
1968.9.4	23 24 47·2	33·99	58·24	15	5·4	32(n)
1968.9.3	8 19 52·2	41·79	32·31	5	5·7	26(n)
1968.9.1	7 27 30·2	34·04	58·22	15	5·9	32(m)
1968.8.31	10 47 37·4	33·97	59·02	13	6·0	32(l)
1968.5.30	17 40 24·4	35·49	27·96	21	5·3	20(t)
1968.4.29	17 1 57·6	39·20	44·30	34	5·3	32(k)
1968.2.19	22 45 41·2	39·40	25·00	7	—	20(s)
1968.1.25	9 56 48·7	37·80	13·20	33	5·1	10(j)
1968.1.16	16 42 44·3	37·90	13·10	14	5·1	10(i)
1968.1.15	2 1 8·5	37·90	13·10	33	5·4	10(h)
1967.11.30	7 23 51·5	41·50	20·50	29	6·0	20(r)
1967.8.13	22 7 47·5	43·20	−0·50	15	5·3	10(g)
1967.7.30	1 31 1·7	40·70	30·40	16	5·6	20(q)
1967.7.26	18 53 1·3	39·50	40·40	33	5·6	26(m)
1967.7.22	16 56 53·3	40·70	30·80	4	6·0	26(l)
1967.7.13	2 10 20·0	35·50	−0·10	13	5·0	10(f)
1967.5.1	7 9 0·5	39·70	21·30	15	5·6	20(p)
1967.4.7	18 33 31·3	37·40	36·20	39	5·0	26(k)
1967.3.4	17 58 6·4	39·20	24·60	33	—	20(o)
1967.1.11	11 20 45·7	34·10	45·70	34	5·6	32(j)
1966.12.10	17 8 32·2	41·00	33·50	13	4·9	26(j)
1966.10.29	2 39 29·4	39·20	21·20	20	5·7	20(n)
1966.9.18	20 43 53·8	27·90	54·30	18	5·9	32(i)
1966.8.20	11 59 8·8	39·40	40·90	12	5·3	26(i)
1966.8.19	12 22 10·7	39·20	41·60	33	—	26(h)
1966.7.27	14 49 2·1	32·60	48·80	33	5·3	32(h)
1966.7.12	18 53 10·4	44·60	37·30	46	5·7	27(b)
1966.7.4	12 15 26·5	37·50	−24·70	20	5·4	10(e)
1966.5.9	0 42 54·2	34·50	26·40	20	5·5	21(o)
1966.4.27	19 48 52·4	38·20	42·50	40	5·0	26(g)
1966.4.20	16 42 6·0	41·70	48·20	36	5·4	32(g)
1966.3.7	1 16 9·7	39·10	41·60	38	5·3	26(f)
1966.2.5	2 1 45·5	39·10	21·70	22	5·6	20(m)
1965.12.20	0 8 15·3	40·20	24·80	33	5·3	20(l)
1965.11.28	5 26 7·4	36·30	27·50	89	5·8	23(f)
1965.7.6	3 18 42·7	38·40	22·40	20	5·9	20(k)
1965.6.21	0 21 16·1	28·10	55·90	40	6·0	32(f)
1965.6.13	20 1 48·0	37·80	29·30	16	5·3	21(n)
1965.4.27	14 9 7·1	35·70	23·50	50	5·5	21(m)
1965.4.9	23 57 3·2	35·10	24·30	51	6·0	21(l)
1965.4.5	3 12 54·2	37·70	21·80	34	5·7	21(k)
1965.3.31	9 47 30·7	38·60	22·40	78	6·3	23(e)

Table 3 (*continued*)

Year/month/day	Time h m s	Lat °N	Long °E	Depth (km)	Magnitude	Fig.
1965.3.9	17 57 53·7	39·40	24·00	18	5·7	20(j)
1965.1.1	21 38 29·2	35·70	4·40	10	5·2	10(d)
1964.12.22	4 36 34·7	28·20	57·00	42	5·5	32(e)
1964.10.6	14 31 19·2	40·30	28·20	10	6·0	20(i)
1964.7.17	2 34 26·9	38·20	23·70	150	5·4	23(d)
1964.6.14	12 15 31·3	38·00	38·50	8	—	26(e)
1964.4.29	4 21 6·7	39·30	23·70	33	5·1	20(h)
1964.4.13	8 30 3·6	45·30	18·10	33	5·4	12(e)
1964.4.11	16 0 42·8	40·50	25·00	33	5·1	20(g)
1964.3.15	22 30 26·0	36·20	−7·60	27	6·2	10(c)
1963.12.16	13 47 56·4	37·10	20·90	15	5·6	21(j)
1963.9.18	16 58 12·5	40·90	29·20	33	5·2	20(f)
1963.7.26	4 17 12·5	42·10	21·40	5	5·5	20(e)
1963.7.19	5 45 28·0	43·40	8·20	33	5·5	12(d)
1963.7.16	18 27 18·4	43·10	41·50	33	5·8	27(a)
1963.2.21	17 14 30·7	32·60	21·00	5	5·5	21(i)
1962.9.1	19 20 39·9	35·63	49·87	27	7·2×	32(d)
1962.8.28	10 59 56·3	37·80	22·90	100	6·3×	23(c)
1962.8.21	18 19 33·3	41·40	15·50	34	6·0×	12(c)
1962.1.11	5 5 4·1	43·30	17·10	33	5·7×	12(b)
1961.5.23	2 45 16·0	36·60	28·30	72	6·3×	23(b)
1959.11.15	17 8 40·5	37·80	20·56	0	6·6×	21(h)
1959.4.25	0 26 19·5	37·05	28·55	43	6·1×	21(g)
1957.12.13	1 45 4·6	34·35	47·65	40	7·2×	32(c)
1957.7.2	0 42 23·0	36·14	52·70	10	7·3×	32(b)
1957.5.26	6 33 31·6	40·66	30·89	0	7·0×	26(d)
1957.4.25	2 25 41·9	36·45	28·59	0	7·0×	21(f)
1957.4.24	19 10 12·8	36·37	28·61	50	6·9×	21(e)
1956.7.9	3 11 38·6	36·69	25·81	0	7·5×	21(d)
1956.2.20	20 31 38·1	39·86	30·49	9	6·0×	20(d)
1955.9.12	6 9 19·3	32·28	29·68	1	6·7×	21(c)
1955.7.16	7 7 11·2	37·66	27·19	6	6·8×	21(b)
1954.9.9	1 4 37·7	36·25	1·53	0	6·7×	10(b)
1954.4.30	13 2 35·1	39·24	22·17	0	7·0×	20(c)
1954.3.29	6 17 6·8	37·02	−3·56	630	7·2×	23(a)
1953.8.12	9 23 51·2	38·11	20·72	0	7·2×	21(a)
1953.3.18	19 6 13·3	40·07	27·39	0	7·2×	20(b)
1951.8.13	18 33 30·0	40·95	32·57	0	6·7×	26(c)
1949.7.23	15 3 30·1	38·66	26·29	0	6·7×	20(a)
1948.10.5	20 12 4·3	37·78	58·41	5	7·3×	32(a)
1943.6.20	15 32 50·6	40·70	30·38	0	6·3×	26(b)
1941.11.25	18 3 54·6	37·47	−19·10	0	8·3×	10(a)
1939.12.26	23 57 16·9	39·70	39·41	0	8·0×	26(a)
1936.10.18	3 10 3·7	46·05	12·50	18	5·6×	12(a)

Date, time of origin and position of the focus of all earthquakes whose focal mechanisms are used in this paper. The parameters for shocks after 1960 are those given by the USCGS. All earthquakes before 1960 have been relocated. The body wave magnitudes of earthquakes after February 1963 are those given by the USCGS and are systematically low (Bune *et al.* 1970). Those for earlier shocks, marked with a cross, are from a number of different sources. All shocks are followed by the figure number of their mechanism except 1969.3.31, which will be found in McKenzie *et al.* (1970).

there were insufficient seismic stations to provide good epicentral determinations. Accurate epicentres have been determined since 1961 by the USCGS, and Barazangi & Dorman (1969) showed that there was an aseismic region in the centre of the Aegean surrounded by active belts on all sides. This plate was called the Aegean plate in (I), and its boundaries are now reasonably well defined by the seismic activity (Figs 15 and 16).

The first authors to recognize the aseismic nature of the southern part of the Aegean were Galanopoulos (1967) and Ergin (1966). Galanopoulos also used

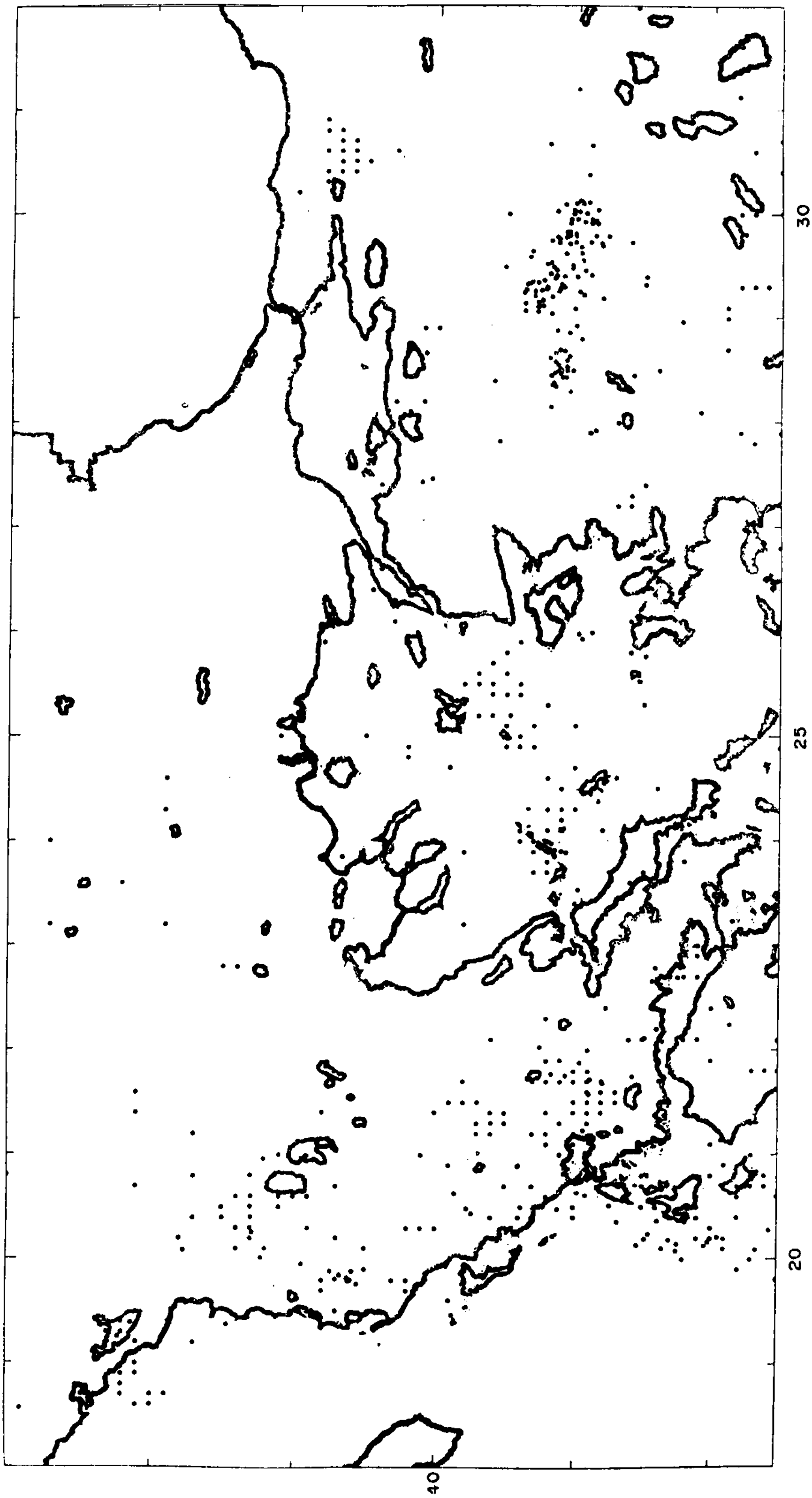


FIG. 15. USCGS epicentres between 1961 and 1970 in the Northern Aegean. Only shocks with foci of 50 km or less depth are plotted.

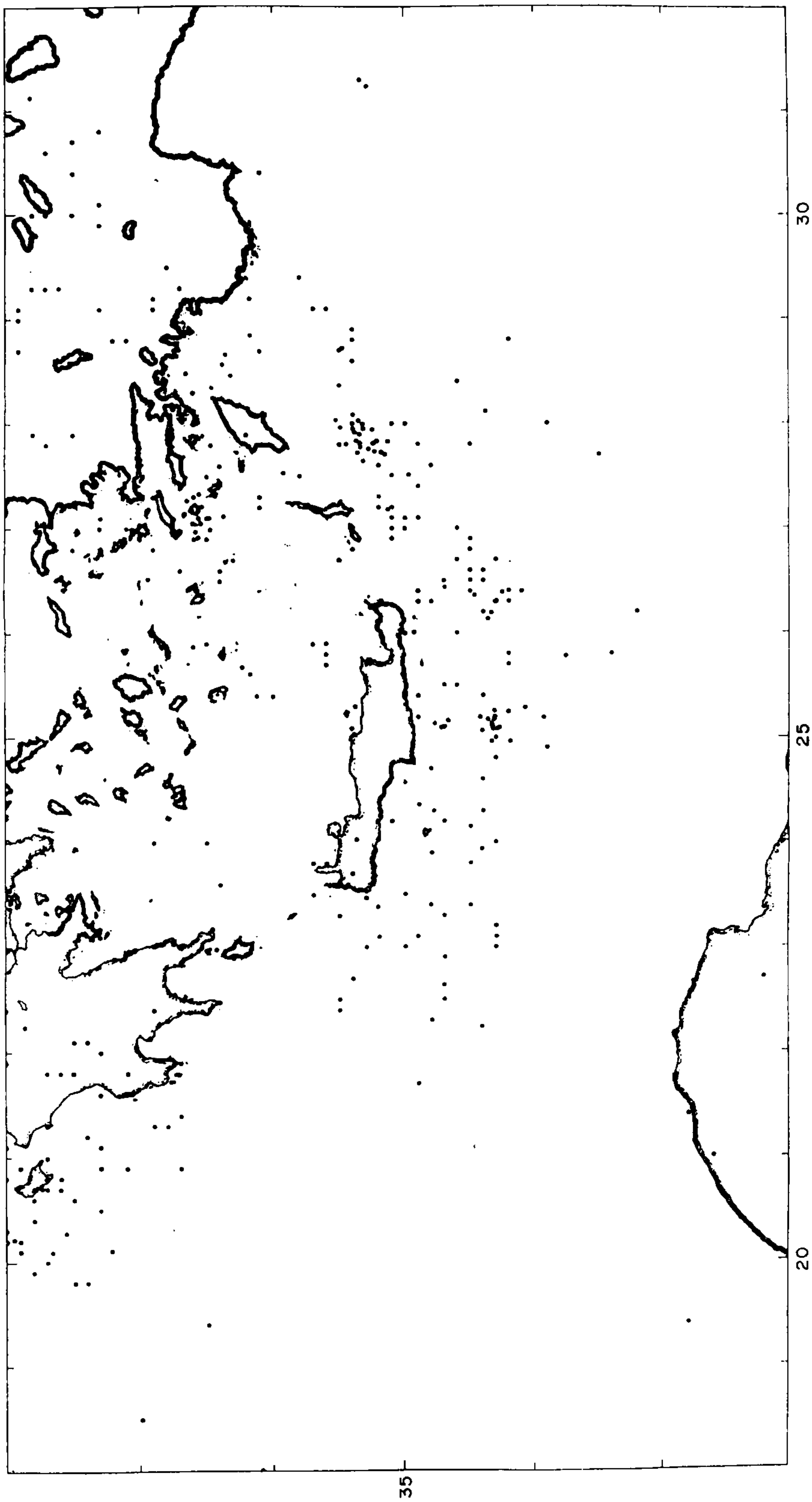


Fig. 16. As for Fig. 15 but for the Southern Aegean.

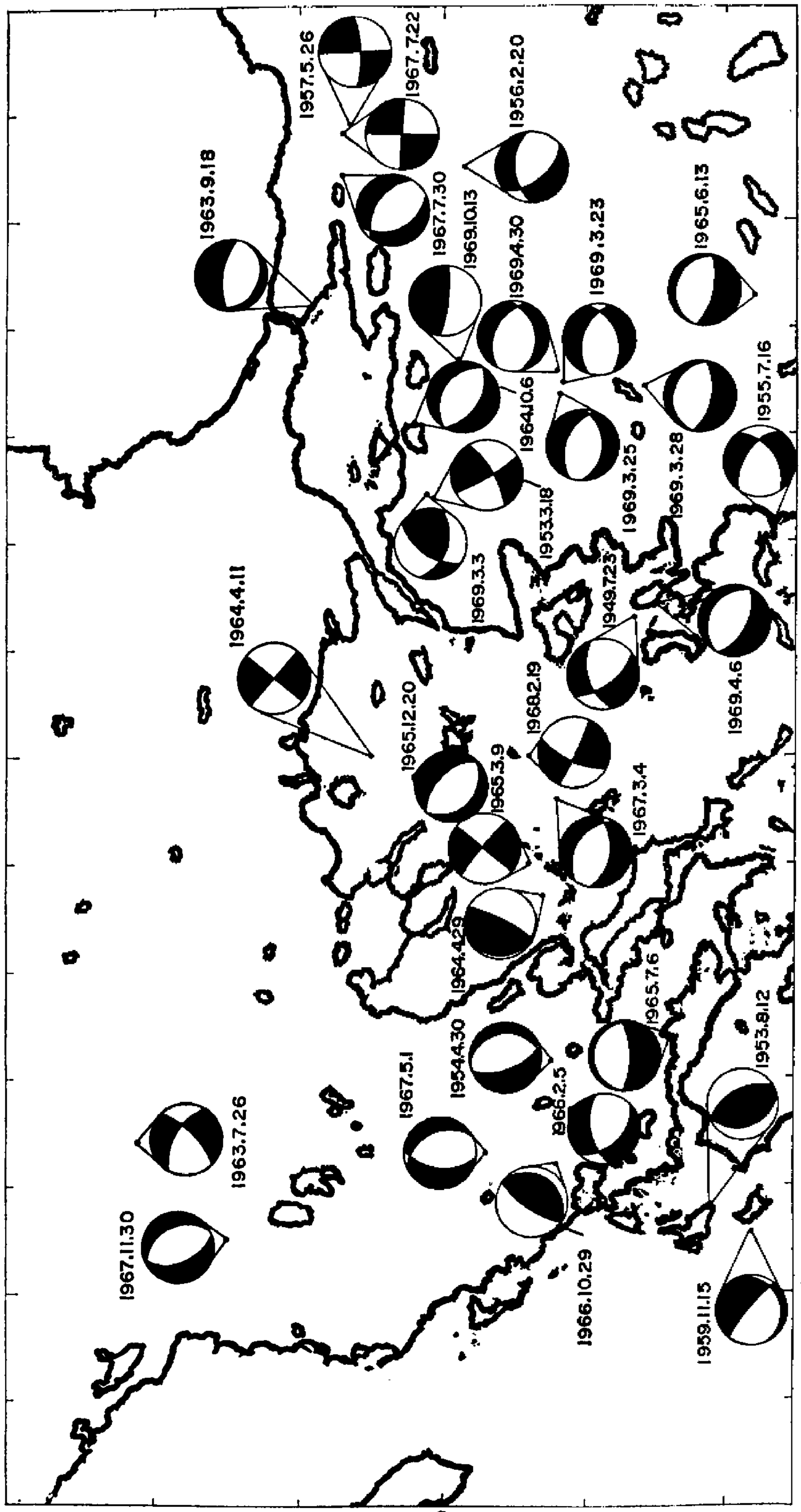


FIG. 17. Mechanisms of earthquakes in the N. Aegean. Conventions as for Fig. 5, and the mechanisms are shown in Fig. 20.

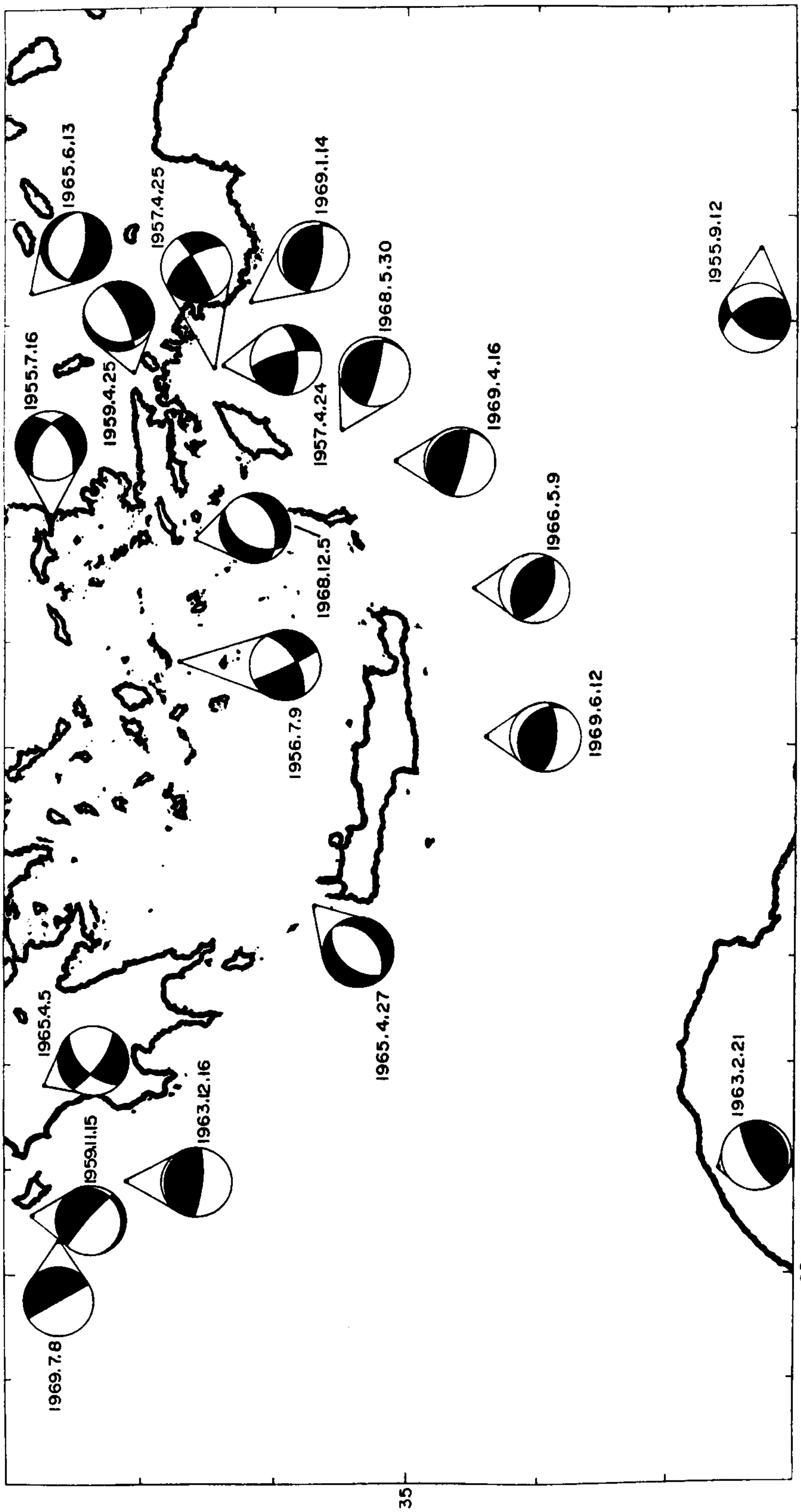


FIG. 18. As for Fig. 17. Mechanisms are shown in Fig. 21.

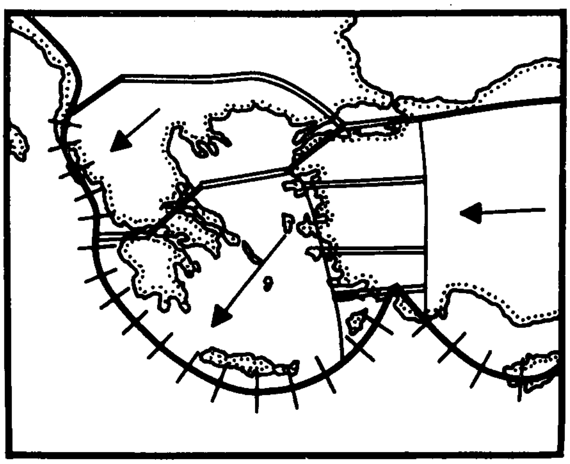


FIG. 19. Sketch of plate boundaries and motions surrounding the Aegean obtained from the seismicity and fault plane solutions.

Hodgson & Wickens' (1965) fault plane solutions in an attempt to account for the earthquake mechanisms by means of a regional stress field. He believed that the active belt in the northern Aegean and the Cretan Arc formed a conjugate fault system in the sense of Anderson (1951) and Hill & Dibblee (1953). This interpretation is not followed here because the success of plate tectonics has shown that it is the slip vectors, and not the orientation of the *P* axes, which are of importance for regional tectonics (McKenzie 1969b). Galanopoulos also pointed out that the right-handed strike slip motion of the North Anatolian Fault continued across the northern part of the Aegean. This result is strongly supported by the fault plane solutions in Fig. 17, though the motions cannot be as simple as Galanopoulos believed. A comparable synthesis to that of Galanopoulos has not been previously attempted in Turkey, though the seismicity (Ergin 1966) and fault plane solutions (Canitez & Ucer 1967a, b) have been studied in some detail. These authors recognized that the strike slip character of the North Anatolian Fault changed west of about 30° E to predominantly normal faulting, but they did not account for this change.

The fault plane solutions in Figs 17 and 18 show that every type of faulting occurs in the active boundaries of the Aegean, and it is difficult at first to make any sense of the observations. This difficulty is partly caused by the lack of continuity of many of the active tectonic structures in the region. Most of the mechanisms can, however, be accounted for by the south-westward motion of the Aegean plate relative to both Eurasia and Africa. A sketch of motion in Fig. 19 shows the general outline of the tectonics, though many of the details have been omitted. The arguments in favour of this interpretation are contained in the rest of this section.

One unexpected observation was episodic nature of the region activity. Between 1961 and 1968 only two shocks occurred in western Turkey and on the Cretan Arc which were sufficiently large for fault plane solutions to be obtained. From 1968 May to 1969 June, however, eight shocks of sufficient magnitude occurred in various parts of this area, and the high activity has been maintained since then. Though less striking than the sequential activity on the North Anatolian fault (Section 5), this behaviour does suggest that stresses are propagating across the Aegean, as well as Turkey, in the manner described by Elsasser (1967) and Savage (1971).

The least complicated part of the Aegean is its southern margin (Figs 16, 18 and 21). Several well-determined mechanisms show thrusting with little or no strike slip component, and these solutions show that the motion between the Aegean and the

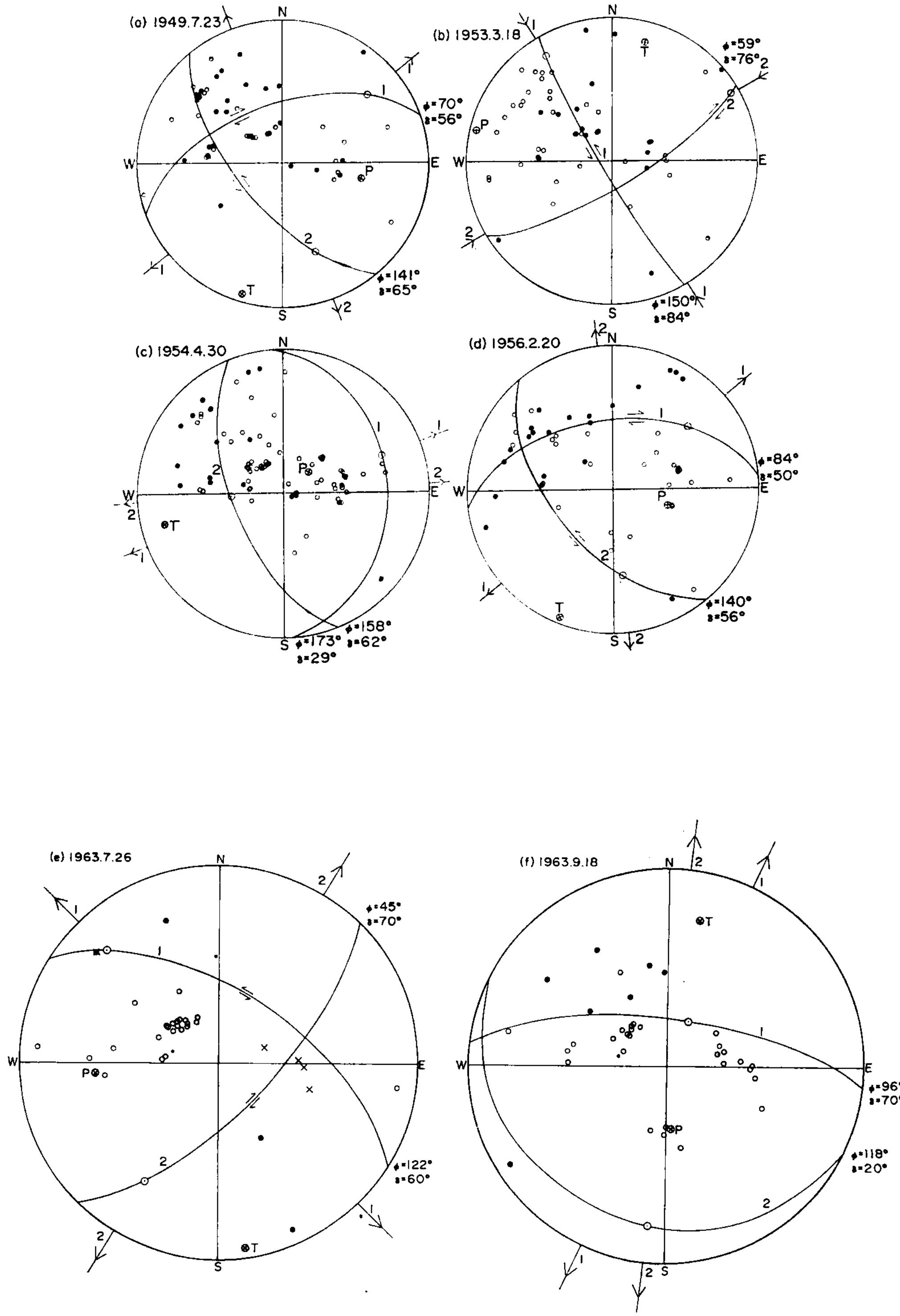
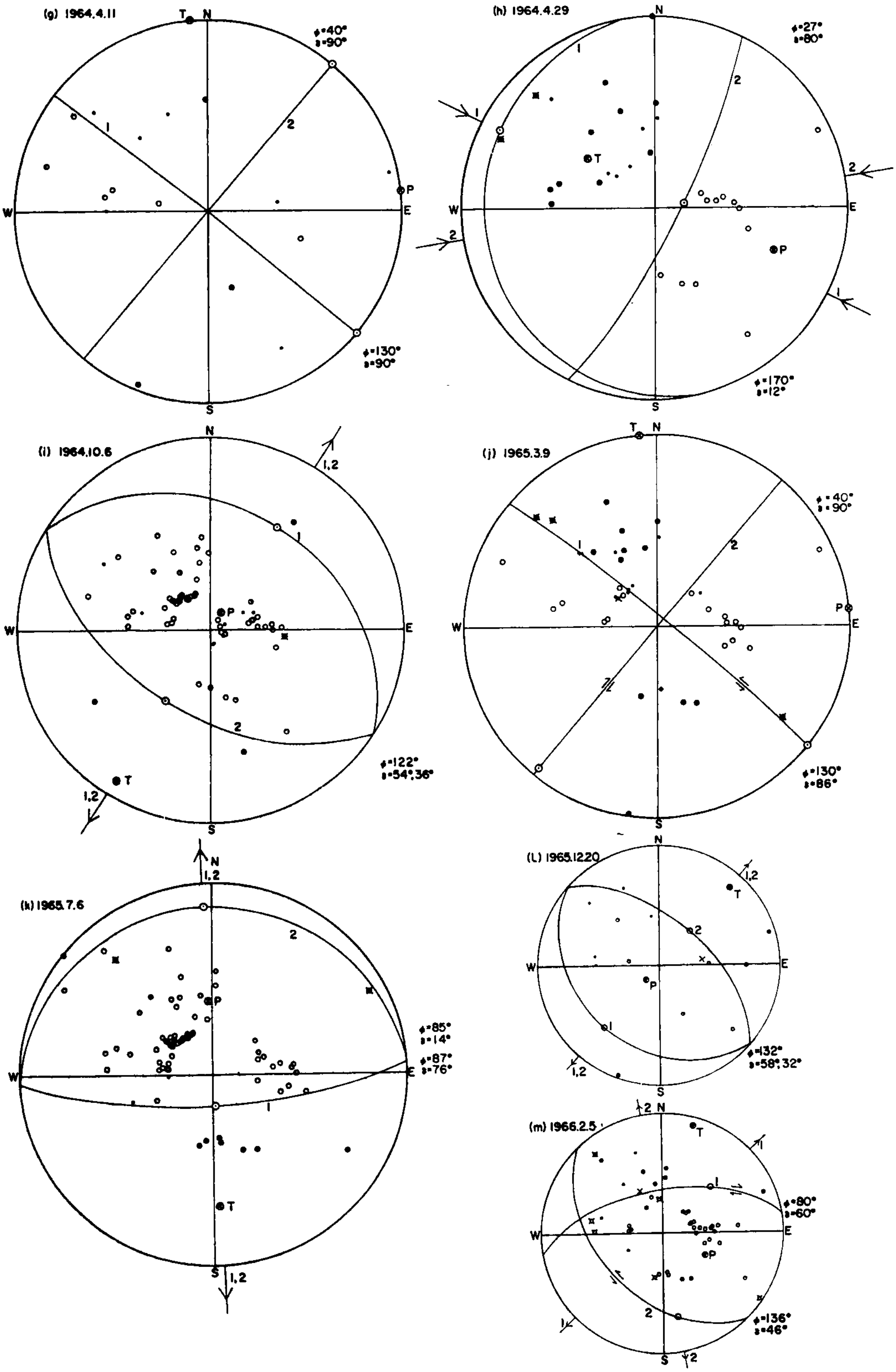

(a) 1949.7.23
N
W
E
S
P
T
1
2
ϕ=70°
δ=56°
ϕ=141°
δ=65°
(b) 1953.3.18
N
W
E
S
P
T
1
2
ϕ=59°
δ=76°
ϕ=150°
δ=84°
(c) 1954.4.30
N
W
E
S
P
T
1
2
ϕ=158°
δ=62°
ϕ=173°
δ=29°
(d) 1956.2.20
N
W
E
S
P
T
1
2
ϕ=84°
δ=50°
ϕ=140°
δ=56°
(e) 1963.7.26
N
W
E
S
P
T
1
2
ϕ=45°
δ=70°
ϕ=122°
δ=60°
(f) 1963.9.18
N
W
E
S
P
T
1
2
ϕ=96°
δ=70°
ϕ=118°
δ=20°

FIG. 20(a)–(f)



(g) 1964.4.11
φ=40°
δ=90°
φ=130°
δ=90°
(h) 1964.4.29
φ=27°
δ=80°
φ=170°
δ=12°
(i) 1964.10.6
φ=122°
δ=54°, 36°
(j) 1965.3.9
φ=40°
δ=90°
φ=130°
δ=86°
(k) 1965.7.6
φ=85°
δ=14°
φ=87°
δ=76°
(L) 1965.12.20
φ=132°
δ=58°, 32°
(m) 1966.2.5
φ=80°
δ=60°
φ=136°
δ=46°

FIG. 20(g)–(m)

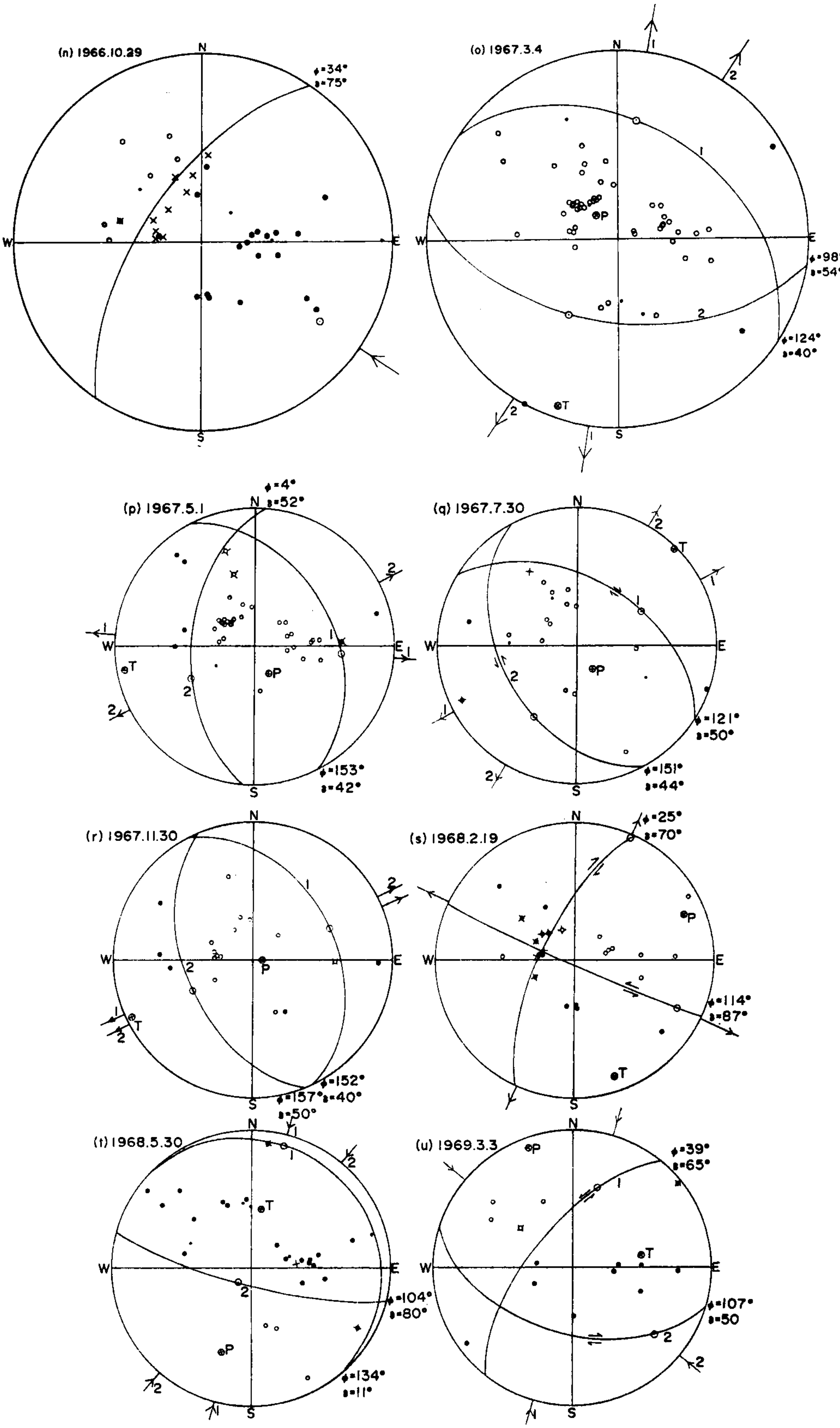

(n) 1966.10.29
φ=34°
δ=75°
(o) 1967.3.4
φ=98°
δ=54°
φ=124°
δ=40°
(p) 1967.5.1
φ=4°
δ=52°
φ=153°
δ=42°
(q) 1967.7.30
φ=121°
δ=50°
φ=151°
δ=44°
(r) 1967.11.30
φ=152°
δ=40°
φ=157°
δ=50°
(s) 1968.2.19
φ=25°
δ=70°
φ=114°
δ=87°
(t) 1968.5.30
φ=104°
δ=80°
φ=134°
δ=11°
(u) 1969.3.3
φ=39°
δ=65°
φ=107°
δ=50

FIG. 20(n)–(u)

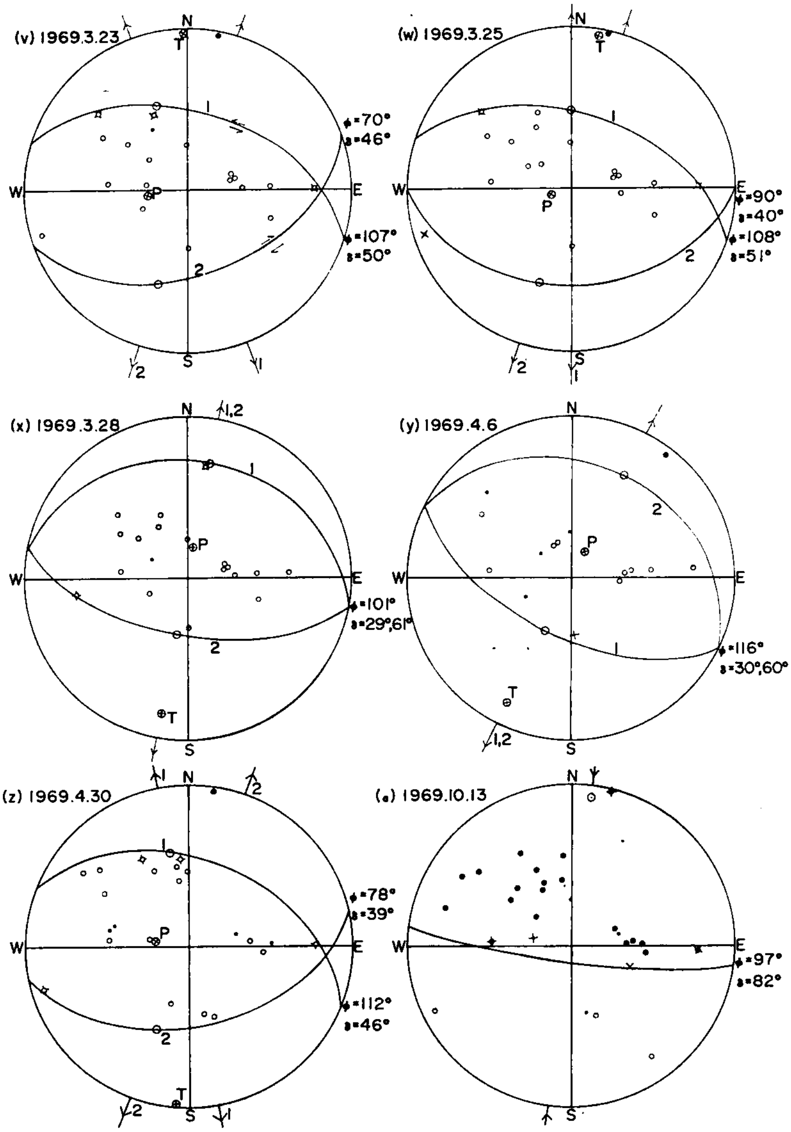


Fig. 20. Mechanisms for earthquakes in Fig. 17. Solutions (e)–(α) based on long period WWSSN observations, and (a)–(d) on short period observations from Wickens & Hodgson (1967) supplied by Wickens. All mechanisms except (i), (n) and (x) in the mantle, these three in the crust. For an account of the conventions used in the mechanisms see the caption of Fig. 7.

African plate is in the north–south direction. It is not possible to resolve the ambiguity between the fault plane and the auxiliary plane for any of these earthquakes since none of them produced known surface breaks. However the intermediate shocks and volcanoes lie to the north of the shallow activity (Fig. 22). Thus the Mediterranean sea floor is probably underthrusting the Cretan Arc and the northward dipping plane is the fault plane. If this choice is correct then the average slip vector direction for the seven shocks since 1963 in Fig. 16 is 17° E. The four mechanisms for shocks before 1963 are all very poor, with many inconsistent observations. The

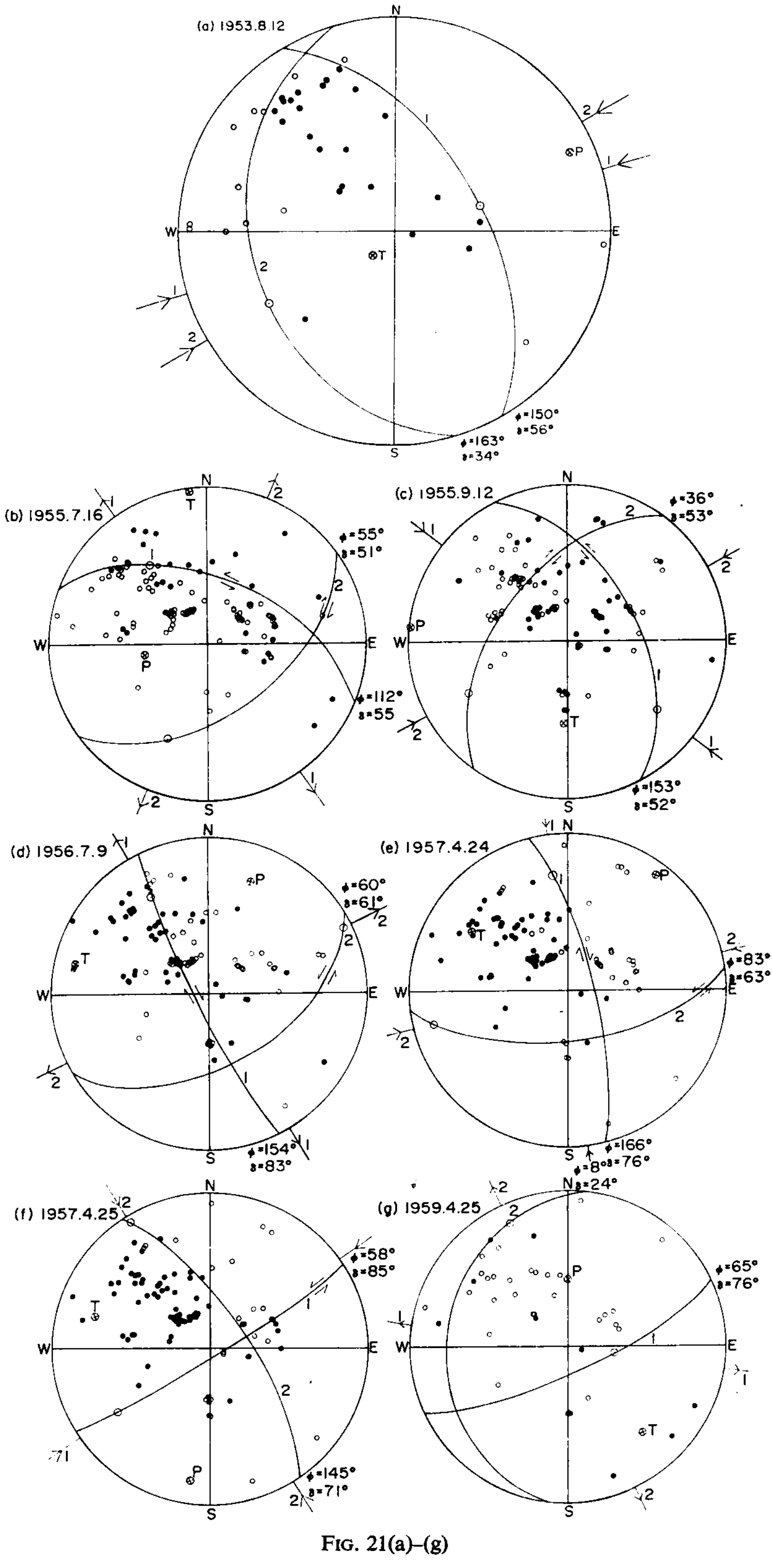

(a) 1953.8.12
(b) 1955.7.16
(c) 1955.9.12
(d) 1956.7.9
(e) 1957.4.24
(f) 1957.4.25
(g) 1959.4.25

FIG. 21(a)–(g)

two large shocks near Rhodes, 1957.4.24 and 1957.4.25, both appear to be produced by strike slip motion, but if this between Africa and the Aegean then the motion must be left lateral, or on planes 2 and 1 respectively in Fig. 21. The resulting slip vectors do not agree well with those from later shocks, but little confidence can be placed in either of these two mechanisms. At the western end of the arc there are three mechanisms, 1953.8.12 (Figs 8, 15 and 21), 1959.11.15 and 1969.7.8, all of which suggest that the relative motion direction is more north-east than north. Two of the mechanisms, 1959.11.15 and 1969.7.8, are shown as normal faulting solutions, though they could well be thrusts. If the shallow angle plane is taken to be the fault plane then the mechanisms are those that would be expected during the formation of nappes and gravity slides, such as those of the Alps (Trümpy 1964; Hsü & Schlanger 1971) and the Appenines (Maxwell 1959). This suggestion could easily be tested by a microearthquake survey, and the resulting structures could probably be detected by marine seismic reflection observations.

There is at present no evidence that the Cretan Arc extends north of 39° N along the coasts of Albania and Yugoslavia. Few shocks north of this latitude occur at sea (Fig. 15), and only one mechanism for a shock beneath Greece, 1966.10.29, shows thrusting. This latitude also marks the northern limit of the intermediate shocks and vulcanism (Fig. 22), and therefore the motion on the Cretan Arc must be taken up on a boundary across Greece which is marked by the line of shallow focus earthquakes between the west coast of Greece and the western end of the North Anatolian Fault marked by the epicentre of 1967.7.22. This activity must be produced by the motion between the Aegean and Eurasian plates, and the slip vectors should not therefore be parallel to those on the Cretan Arc which takes up the slip between the Aegean and the African plates.

The scattered activity in the northern Aegean suggests that the plate boundary in this region is not simple, and the remarkable variety of the fault plane solutions shows that every type of faulting is indeed present in a small region. Three shocks produced surface breaks, and therefore can be used to determine the strike of the slip vectors. The most important of these earthquakes was on 1953.3.18, and its fault plane solution is shown in Fig. 20. The surface break was mapped by Ketin & Roesli (1953) and by Dilgan & Hagiwara (1955) and showed up to 4·3 m of right-handed strike slip motion parallel to plane 2 in Fig. 20(b). The slip vector is therefore at right angles to plane 1. The surface break for 1964.10.6 was much less clear, and was mapped by Ketin (1966) who believed the surface features were the result of right-handed strike slip motion on an east–west fault. The fault plane solution shows that the dominant type of faulting was normal faulting, but the strike of plane 1 is not well controlled, and if it was the fault plane, an east–west strike is not excluded by the observations. There then would indeed be a considerable component of right lateral strike slip motion, and the slip vector would strike 32° E, compared with 60° E for 1953.3.18.

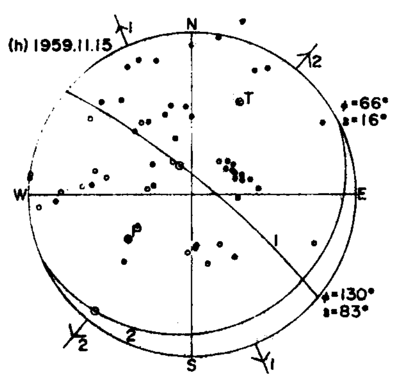


FIG. 21(h)

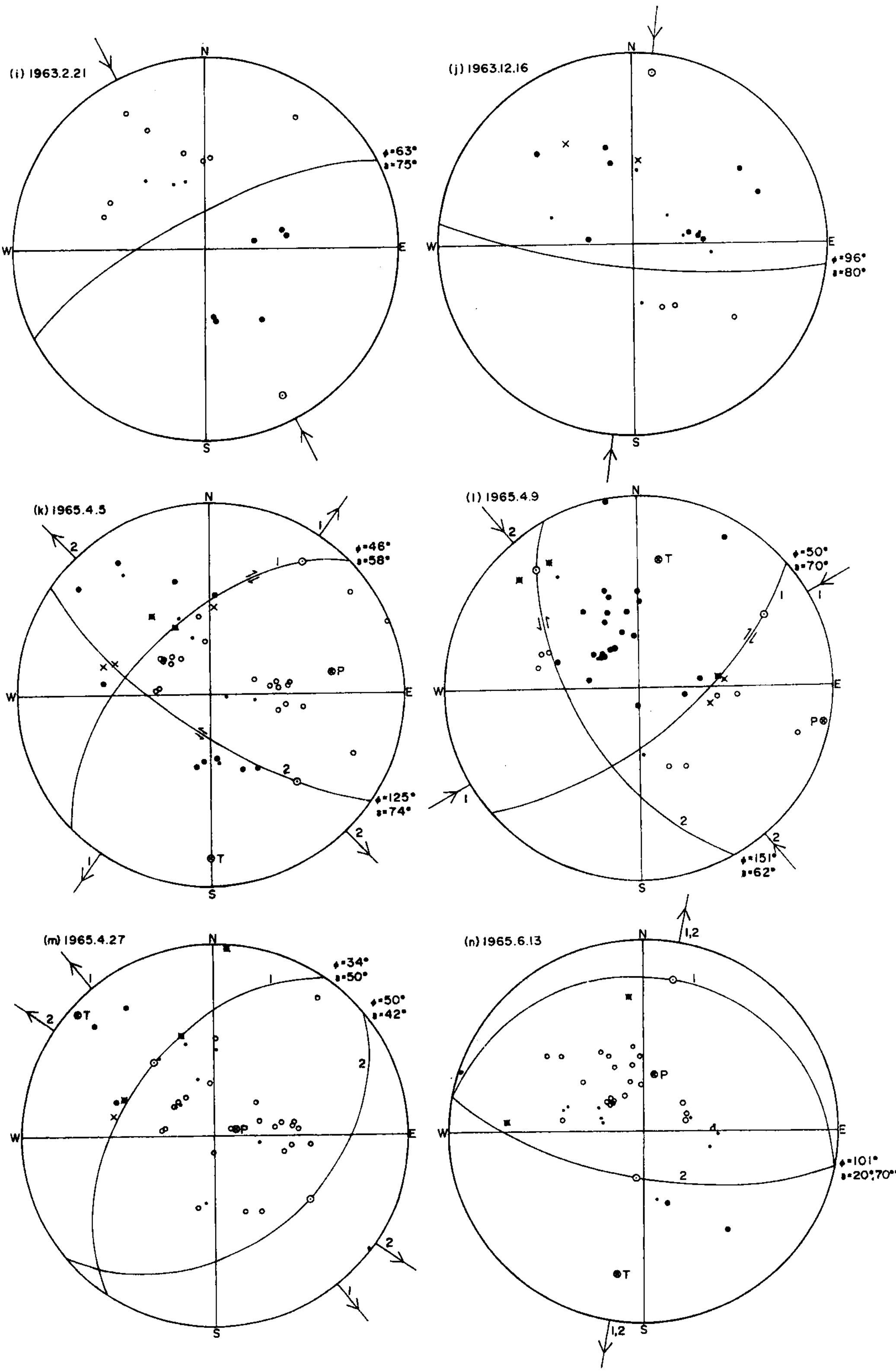

(i) 1963.2.21
N
W
E
S
φ=63°
δ=75°
(j) 1963.12.16
φ=96°
δ=80°
(k) 1965.4.5
φ=46°
δ=58°
φ=125°
δ=74°
P
T
(l) 1965.4.9
φ=50°
δ=70°
φ=151°
δ=62°
(m) 1965.4.27
φ=34°
δ=50°
φ=50°
δ=42°
(n) 1965.6.13
φ=101°
δ=20°,70°
1,2

FIG. 21(i)–(n)

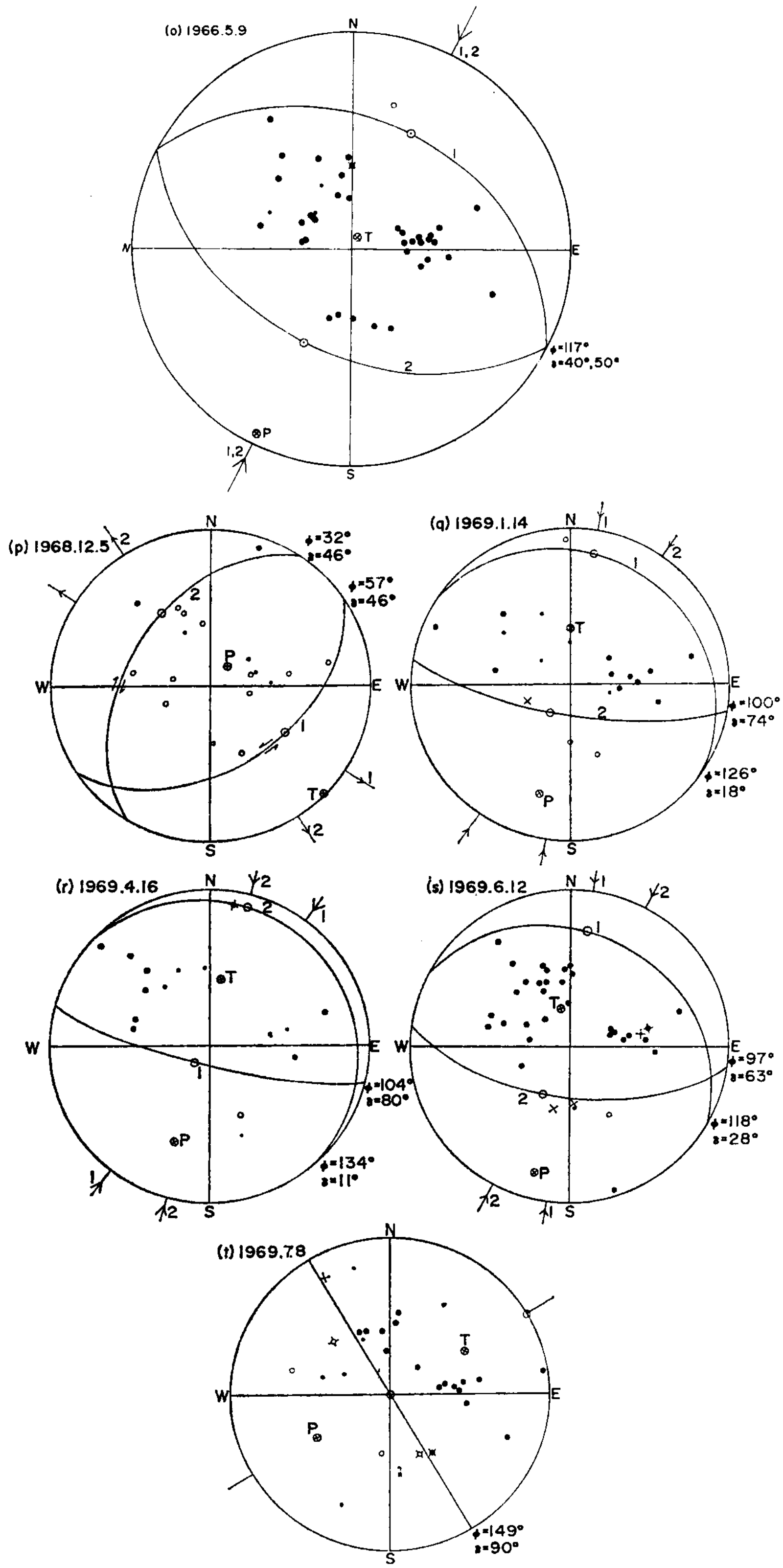


FIG. 21. Mechanisms for earthquakes in Fig. 18. Solutions (i)–(t) based on long period WWSSN observations, (b)–(h) on short period observations from Wickens and Hodgson supplied by Wickens, and (a) on short period observations from Di Filippo & Marcelli (1954). All mechanisms except (j), (n) and (o) in the mantle, these three in the crust. For an account of the conventions used in the mechanisms see the caption of Fig. 7.

The last shock which may have produced an observed surface break was an aftershock of 1967.7.22 on 1967.7.30. This aftershock occurred at the western end of the surface break mapped by Ambraseys & Zatopek (1969), but it is not clear whether the break originated during the main shock or the aftershock. However, plane 1 in Fig. 20(q) agrees well with the strike, sense of motion and surface displacement mapped by these authors, and if this choice is made the strike of the slip vector is then 61° E. Furthermore since the polarity of the *P* wave at Istanbul 300 km from the epicentre was dilatational from the main shock on 1967.7.22 but compressional for 1967.7.30, the difference in mechanism explains Ambraseys & Zatopek's observation that the polarity of many aftershocks recorded at Istanbul was different from that of the main shock. These three breaks all require the Aegean plate to be moving south-west relative to Eurasia, and it is easy to test whether this relative motion continues across the remainder of the boundary further west. The simplest test is to examine the mechanisms in Fig. 17 to discover if each solution has a plane striking north-west, and if so this plane must be the auxiliary plane and is normal to the slip vector. This test shows that the majority of the solutions for shocks beneath the Aegean satisfy this condition, and that there are two types of faulting involved. Three shocks, 1964.4.11, 1965.3.9 and 1968.2.19 have strike slip mechanisms which closely resemble that of 1953.3.18, and therefore involved right-handed strike slip faulting on planes striking north-east. The other type of faulting which is related to the motion between the Aegean and Eurasian plates is normal faulting on east–west faults, combined with some right-handed strike slip. If the northwest striking plane is taken to be the auxiliary plane then the fault plane of 1967.3.4 dips south, that of 1963.9.18 and 1966.2.5 dip north. However, the control on the strike of the nodal planes of some of these solutions is not always good and therefore the wrong plane may have been taken to be the fault plane.

The eight mechanisms discussed above are all consistent with the sketch in Fig. 19. This shows obliquely spreading ridge segments joined by transform faults extending across the northern Aegean. It is not yet clear whether there is one continuous plate boundary in this area, or whether the motion is taken up on a series of parallel grabens connected by strike slip faults. It is, however, clear that there is no feature present which resembles a normal oceanic ridge. Magnetic surveys have failed to show any linear magnetic anomalies associated with the shallow seismicity (Vogt & Higgs 1969), nor is there any large trough of the type that would be expected if sea floor was being produced. Magnetic and bathymetric surveys have, however, shown a remarkable deep linear trough about 50 km north of the major seismic activity (Vogt & Higgs 1969; Lort 1971). This feature is associated with large magnetic anomalies, and in some ways resembles features such as the Red Sea and the Bay of Biscay, both of which are underlain by oceanic crust and neither of which possess typical linear magnetic anomalies. Therefore this trough could also be floored by oceanic crust and have been produced by the present motion of the Aegean plate. Since it is not now associated with the larger shocks it is not now the boundary on which the principal motion is taken up.

Several mechanisms in Fig. 17 can quite clearly not be related to the motion between the Eurasian and Aegean plates if the sketch in Fig. 19 is correct. Four of these are thrust faults 1964.4.29, 1969.3.3, 1969.10.13 and 1968.9.3 in Fig. 25. The last shock occurred further east than the others, but appears to be of similar type. Though two of these shocks, 1968.9.3 and 1969.3.3 have a considerable strike slip as well as a thrust component, and also have one plane striking northwest, they cannot be simply interpreted in terms of the Eurasia–Aegean motion. The difficulty becomes clear if the north-west plane is chosen as the auxiliary plane, since the resulting strike slip motion is then left handed. A similar difficulty arises with 1969.10.13, which is a thrust on a north-westerly striking plane. Though it is not impossible to relate such motions to the Eurasia–Aegean motion, the contortions

Table 4

Year/month/day	Time	Lat °N	Long °E	Depth (km)	Fig.	Number of shock in sequence
1939.12.26	23 57 16·9	39·70	39·41	0	5(f)	1
1939.12.27	2 48 26·1	39·93	38·12	0	5(f)	2
1939.12.27	22 34 9·4	40·80	36·78	0	5(f)	3
1939.12.28	2 23 24·5	41·05	37·09	0	5(f)	4
1939.12.28	3 25 21·2	40·48	37·00	0	5(f)	5
1939.12.28	17 40 43·0	41·37	34·30	0	5(f)	6
1939.12.28	19 9 54·6	39·78	39·88	0	5(f)	7
1939.12.29	11 33 34·5	39·94	38·83	0	5(f)	8
1939.12.29	13 18 19·9	40·23	39·17	0	5(f)	9
1939.12.29	16 4 27·9	40·56	37·56	0	5(f)	10
1940.1.4	20 44 59·6	40·91	38·84	0	5(f)	11
1940.1.26	20 56 1·3	40·40	38·50	0	5(f)	12
1940.2.1	5 12 58·7	41·69	33·04	0	5(f)	13
1940.2.3	19 35 3·1	40·85	38·51	0	5(f)	14
1940.3.7	5 4 36·0	40·84	35·05	0	5(f)	15
1953.3.18	19 6 13·3	40·07	27·39	0	5(c)	1
1953.3.18	21 18 3·8	39·75	27·47	0	5(c)	2
1953.3.19	12 53 42·7	39·98	28·00	28	5(c)	3
1953.3.19	21 13 54·8	39·84	27·24	1	5(c)	4
1953.3.26	15 10 26·6	39·93	27·45	0	5(c)	5
1953.4.1	1 47 34·6	39·92	27·43	0	5(c)	6
1953.6.21	8 11 31·1	37·44	20·73	49	5(b)	1
1953.8.9	7 41 10·0	38·34	20·85	19	5(b)	2
1953.8.11	3 32 22·5	38·35	20·72	0	5(b)	3
1953.8.11	12 3 30·3	38·57	20·50	45	5(b)	4
1953.8.11	13 11 3·7	38·19	20·88	0	5(b)	5
1953.8.12	6 8 6·7	38·44	20·58	26	5(b)	6
1953.8.12	9 23 51·2	38·11	20·72	0	5(b)	7
1953.8.12	11 33 46·9	38·02	20·77	21	5(b)	8
1953.8.12	12 5 20·1	37·86	20·73	0	5(b)	9
1953.8.12	13 39 21·3	38·00	20·76	4	5(b)	10
1953.8.12	14 8 38·1	38·02	20·79	0	5(b)	11
1953.8.12	16 8 32·5	37·99	20·78	15	5(b)	12
1953.8.13	3 22 5·8	38·29	20·92	0	5(b)	13
1953.8.17	2 12 22·3	37·93	20·96	17	5(b)	14
1953.9.14	14 56 13·6	38·34	20·73	0	5(b)	15
1953.10.10	21 29 12·0	37·99	20·98	4	5(b)	16
1953.10.16	21 44 40·9	38·06	20·79	0	5(b)	17
1953.10.21	11 31 5·3	38·31	20·62	0	5(b)	18
1953.10.21	18 39 52·9	38·27	20·57	0	5(b)	19
1953.11.28	20 17 27·7	37·31	20·79	0	5(b)	20
1953.12.28	2 38 43·6	38·28	20·56	0	5(b)	21

Foreshock and aftershock sequences of some of the major shocks before 1960 in the Mediterranean

required in the shape of the plate boundary do not make this explanation particularly convincing. Until the required geometry of the plate boundary can be established using the seismicity or reflection seismic surveys a simpler explanation discussed below in Section 8 seems preferable.

Most of the fault plane solutions for shocks on the mainland of Greece, Albania and Southern Yugoslavia show a component of normal faulting. The only exception took place on 1966.10.29. The solutions are mostly well controlled, but the ambiguity cannot be resolved for any of these shocks because surface faulting following an earthquake has not yet been observed anywhere in this region. Since the strike of normal faults is not easily determined accurately from fault plane solutions it would probably be useful to carry out a special survey to attempt to find surface displacement associated with one of the larger shocks in Fig. 17, for instance 1967.5.1. Perhaps the most striking feature of Figs 15 and 17 is the absence of any obvious structure crossing Greece from the western end of the northern Aegean plate boundary near the epicentre of 1967.3.4 to the northern end of the Cretan Arc north of 1953.8.12. The general features of the activity on both sides of Greece suggest that such a connection does exist, but the only shock which could have occurred on it during the interval was on 1966.2.5. Since the displacement on this structure has probably been more than 100 km (Section 8), it should easily be detected by careful geological mapping. The other solutions in Greece, Albania and Southern Yugoslavia appear to be related to a region of moderate seismic activity which extends westward from the western end of the North Anatolian Fault across Bulgaria to Yugoslavia and Albania. This zone has been the site of some large shocks in the past, and two in Bulgaria on 1928.4.14 and 1928.4.18 produced clear E–W fault traces showing normal faulting (see Richter 1958, p. 619–21). The same motion direction could probably account for 1963.7.26, but not for the normal faulting on N–S faults in Greece. This direction follows that of the major tectonic structures of Greece, and much of the topography of the country is the result of normal faulting during and after the Miocene (Aubouin *et al.* 1963). This pattern of faulting could be the result of the south-west motion of the Aegean plate, which cannot easily be taken up on north–south structures and therefore produces general disruption and seismic activity over a large region of northern Greece and Albania.

Reactivation of old trends has also taken place much more extensively in western Turkey on the eastern boundary of the Aegean. Here the seismic activity extends over a broad north–south band whose width is about 400 km, and all fault plane solutions show normal faulting. The north-eastern limit of this band is between the epicentre of the 1967.7.22 shock and its major aftershock 1967.7.30 (Fig. 17). The first of these shows clear strike slip motion and occurred on the North Anatolian fault, the second shock was the result of normal faulting. Hence the triple junction between the Aegean, Turkish and Eurasian plates must lie between these epicentres. This point also marks the beginning of the horst and graben structures which dominate the physiography of western Turkey, and whose development causes most of the earthquakes of the region. A particularly striking example is the sequence of earthquakes starting on 1969.3.23, whose epicentres are shown in Fig. 5(e). The largest of these shocks was the Gediz earthquake on 1970.3.28 and the resulting surface faulting after this shock has been shown by Ambraseys & Tchlenko (1970) to have occurred on east–west and to a lesser extent north–south faults. A fault plane solution has not yet been determined, but the distribution of aftershocks and the three mechanisms from the earlier shocks in Fig. 5(e) strongly suggest that the Gediz earthquake was the result of the same type of faulting further east. Since surface faulting associated with earthquakes has not previously been observed in western Turkey, the ambiguity between the two planes cannot be resolved in any of the cases. However, since few of the mechanisms show any major component of strike

Table 5

Intermediate and deep focus shocks

	P		T		Null Axis		
Date	strike	dip	strike	dip	strike	dip	Fig.
1965.11.28	246°	21°	349°	30°	128°	52°	23(f)
1965.3.31	235°	27°	30°	60°	139°	11°	23(e)
1964.7.17	219°	24°	97°	50°	324°	30°	23(d)
1962.8.28	353°	2°	85°	65°	263°	25°	23(c)
1961.5.23	179°	17°	359°	73°	89°	0°	23(b)
1954.3.29	86°	47°	271°	42°	178°	3°	23(a)

slip faulting the motion between the Aegean and Turkish plates must be in a north–south direction whichever plane is the fault plane. This boundary is therefore taking up a motion which is approximately parallel to the strike of the band. In the oceans such a feature would be a transform fault with a north–south strike. In western Turkey, however, the motion is taken up on a series of east–west normal faults, presumably because this is the direction of the tectonic structures in the area. Therefore this faulting, like that of Greece, appears to be the consequence of the interference of a new direction of motion with old structural trends. This has reactivated pre-existing faults in a complicated manner to accommodate the new motion. Such reactivation is a common feature of continental but not of oceanic tectonics. It is, however, not easy to understand how a number of east–west graben structures can take up the southward motion of the Aegean relative to Turkey.

Three mechanisms in the Aegean have not yet been mentioned. These are all shown on Fig. 18 and occurred on 1956.7.9, 1965.4.5 and 1965.4.27. They do not

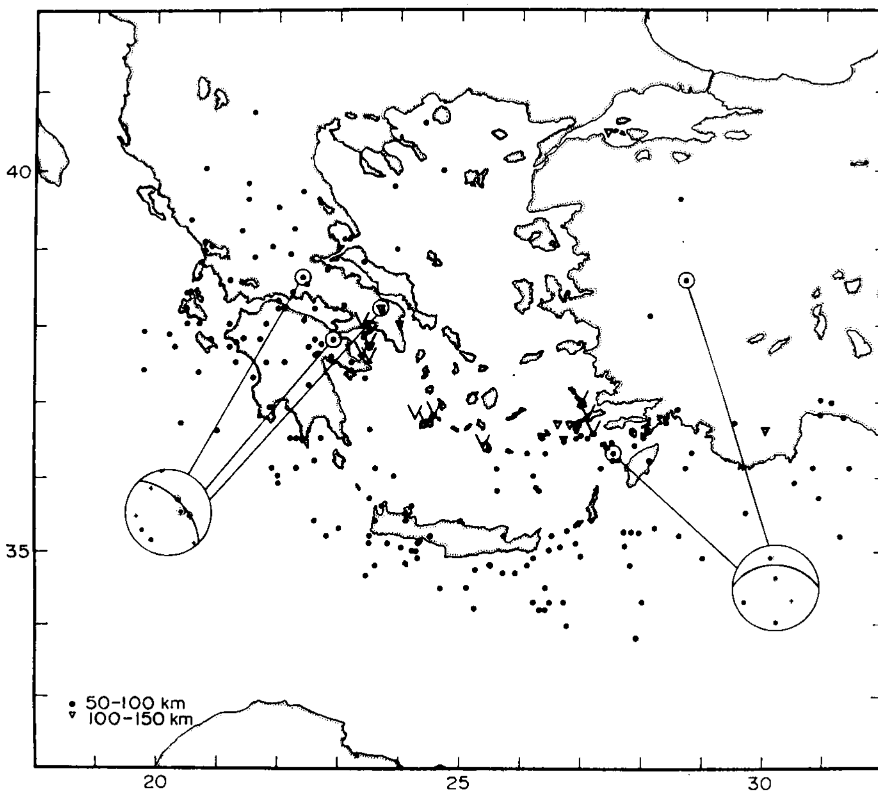


FIG. 22. Epicentres of earthquakes deeper than 50 km, and sites of quaternary volcanoes marked by a heavy V and taken from Washington (1926). The mechanisms of five intermediate shocks, three in the western Aegean and two in the eastern Aegean, are shown in the two circles. The intermediate stress axis, or null vector is shown as a cross, the T axis as a solid dot and the P axis as an open circle with a dot in the centre. The convention is that of Isacks & Molnar (1971), and each circle represents an equal area projection of the lower hemisphere of the sphere on which the stress directions are plotted. The curved line in each circle represents the approximate plane in which the intermediate earthquakes occur, and strikes NW–SE and dips at 60° beneath the western Aegean, changing to E–W with a 45° dip beneath the eastern Aegean. The stress directions agree with the general observations of Isacks & Molnar (1971), who found that the T axis for intermediate earthquakes was parallel to the dip of the sinking slab.

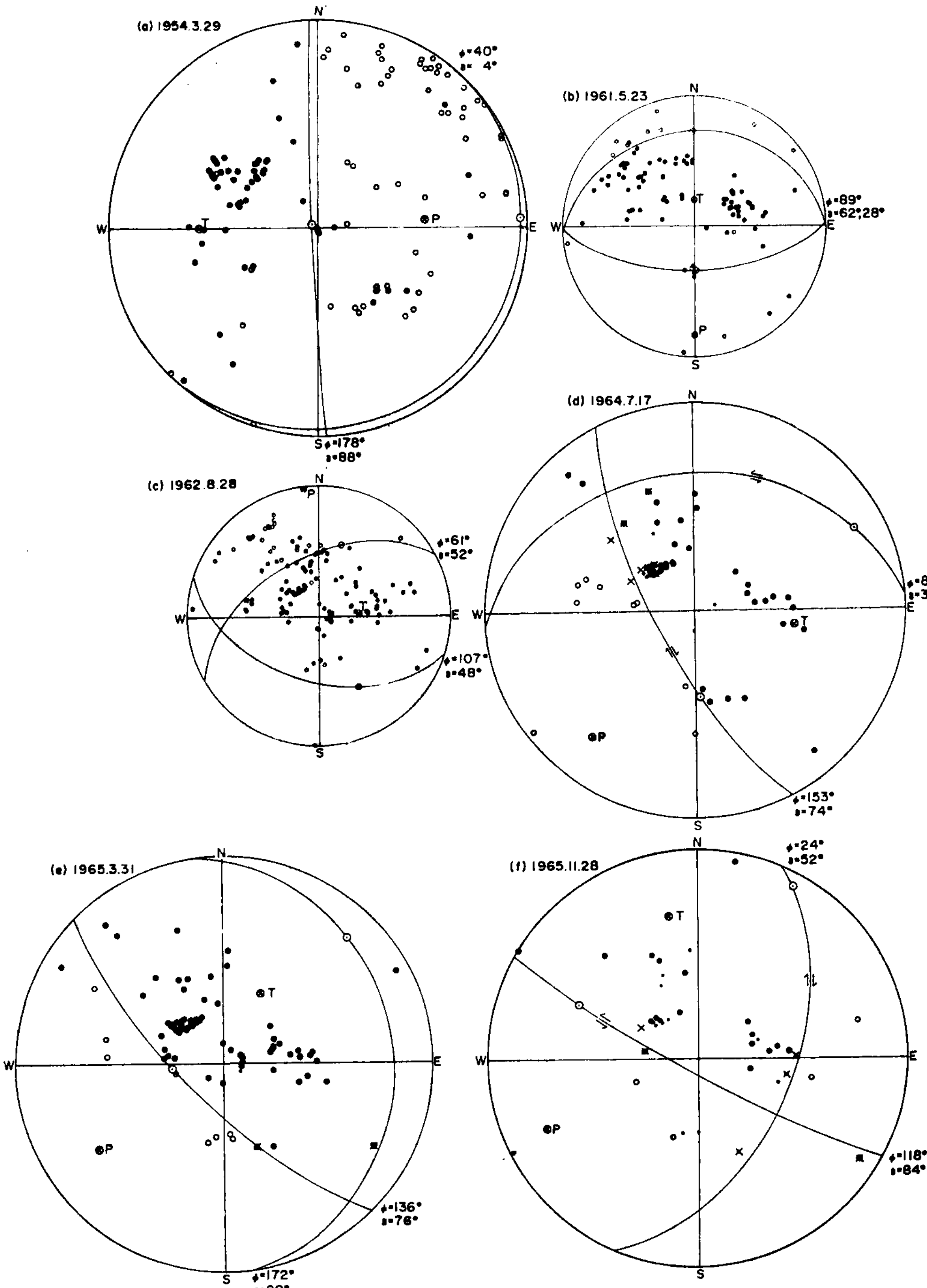


FIG. 23. Solutions of all shocks with focal depths below 70 km. (d)–(f) based on long period WWSSN observations, (c) on some long period WWSSN reading and some short period readings of Wickens & Hodgson (1967) supplied by Wickens. (b) is entirely based on short period observations of the same authors, and (a) on the short period observations of Hodgson & Cock (1956). The strikes and dips of the P and T axes for these shocks are given in Table 3, and an account of the conventions used in the mechanisms is given in the caption to Fig. 7. The shock on 1961.5.23 could be of shallow, rather than of intermediate depth, and would then be of similar mechanism to that of 1969.1.14 (Fig. 21(q)).

appear to be connected in any simple way with the motion of the Aegean plate, but perhaps their importance will become clearer when more good locations and mechanisms are available. Two other mechanisms in Fig. 18, one in north Africa on 1963.2.21 and one north of Egypt on 1955.9.12 are both very poor solutions, the first because the shock was small, the second because of the number of inconsistent observations (Fig. 21).

Little mention has been made of the intermediate earthquakes and their mechanisms, principally because they reflect stresses and deformation within the lithosphere, rather than motions between plates. If present theories are correct intermediate and deep focus shocks are restricted to those parts of the mantle which have been part of an oceanic plate within the last 10 My. Figs 22 and 23 show the locations of the shocks and demonstrates that these earthquakes occur only beneath the southern part of the Aegean to the north of the shallow focus activity. Isacks & Molnar (1971) have used some of the mechanisms in Fig. 23 to demonstrate that the T axes for these shocks are parallel to the dip of the slab. This conclusion is supported by the additional mechanisms in Fig. 23, and therefore the Cretan Arc closely resembles other arcs where the deep focus shocks do not occur. The P, T and null vectors are plotted in Fig. 22 using the same convention as Isacks & Molnar (1971) to show their relationship to the dipping slab.

The mechanisms discussed in this section show the usefulness of fault plane solutions as a tool for understanding regional deformation. Much of the present activity around the Aegean is either beneath the sea or in poorly mapped regions whose geology is little understood. Earthquake mechanisms can be obtained with little expense, and permit the general outline of the tectonics to be studied. The main shortcoming of the method is that it cannot disentangle the tectonics on a scale of less than about 20 km, since this is about the error in epicentre location. This limitation may well explain why the deformation of Greece, Albania and the neighbouring parts of the Aegean sea floor remain at present confused because the blocks are too small to be resolved using teleseismic observations.

6. Turkey, Cyprus and the Black Sea

The main feature of this area is the North Anatolian Fault, whose surface breaks since 1939 are shown in Fig. 24. The North Anatolian Fault is a right-handed strike slip fault between the Turkish and Eurasian plates, and extends from the epicentre of 1967.7.22 in the west to that of the 1966.8.20 in the east. A triple junction occurs at each end of the fault. At the western end the motion is taken up by the boundaries crossing the northern Aegean and western Turkey discussed in the last section, whereas at the eastern end the nature of the faulting changes from strike slip to strike slip and thrust. The term North Anatolian Fault will only be used to refer to the fault between these two triple junctions. The most remarkable feature of this fault is the progression of activity which has occurred from the eastern end in 1939 westward, first remarked by Ketin (1948) and recently discussed in some detail by Ambraseys & Zatopek (1969) in connection with the most recent fault break which was associated with the Mudurnu earthquake on 1967.7.22. Savage (1971) has attempted to explain this progression in terms of propagation of dislocations along a plate boundary. The seismic activity associated with movements on this fault shows clearly on the seismicity maps published by Gutenberg & Richter (1954), Ergin (1966) and in Fig. 2. Since the central part of the fault has not been the site of major earthquakes since 1961 it does not show clearly on the maps of Barazangi & Dorman (1969) or in Figs 1 and 24. Clearly ten years of data is not always sufficient to demonstrate the existence of even active structures. A considerable number of fault plane solutions have already been published by Turkish authors for shocks on this fault. Collections of polarity observations by Ocal (1960, 1966), and by Canitez & Ucer (1967a, b)

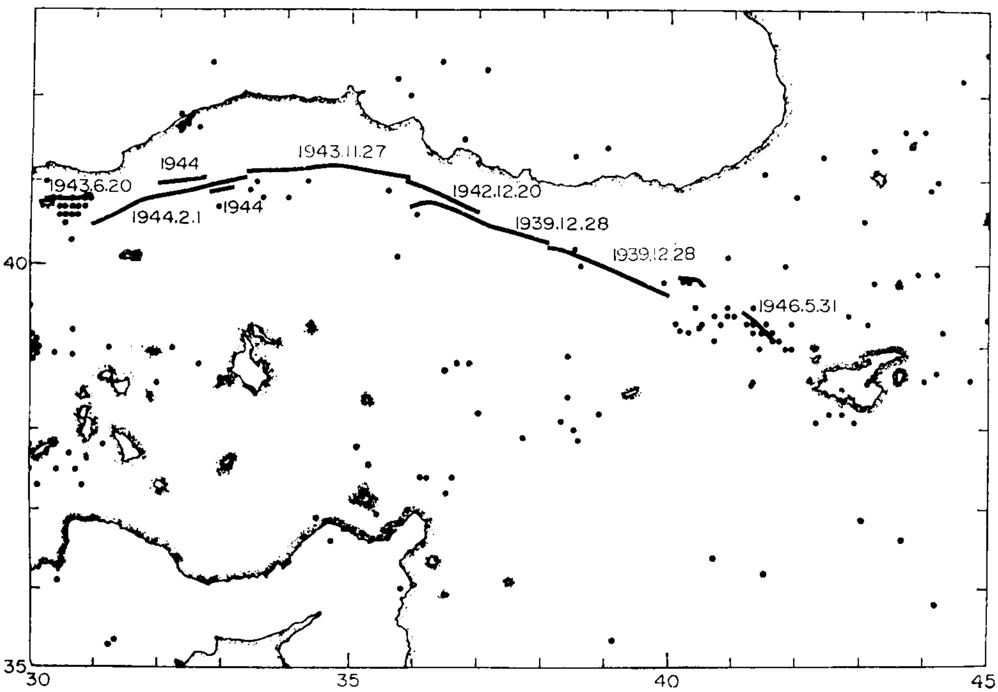


FIG. 24. USCGS epicentres of earthquakes shallower than 50 km and fault breaks of the North Anatolian Fault taken from Ketin & Roesli (1953).

were found especially useful. Several of the surface fractures have also been mapped; that of 1953.3.18 by Ketin & Roesli (1953), who also published a map showing earlier breaks, those of 1966.8.19 by Wallace (1968) and by Ambraseys & Zatopek (1968), and that of the most recent break of 1967.7.22 by Ambraseys & Zatopek (1969). Allen (1965, 1969) has also described this fault and has mapped some of the breaks associated with shocks before 1953.

Throughout the northern part of Turkey the fault plane solutions are in excellent agreement with field observations of surface faulting. Since no large shocks have occurred since 1939 without producing surface displacement there is no ambiguity between the fault plane and the auxiliary plane for any of these events, all of which are the result of right-handed strike slip. The direction of slip given by the normal to the auxiliary plane changes from 105° E at the eastern end to 90° E at the western end. Though most fault plane solutions cannot be used to obtain the slip direction to an accuracy of better than 10° or 20°, the nature of these mechanisms, their self consistency and the agreement of the strike of the fault plane obtained from teleseismic observations and from detailed mapping all suggest that this difference of 15° may in fact be real. If this is indeed the case then the pole position for the motion of Turkey relative to Europe can be determined to be at 18·8° N, 35° E using the method discussed by McKenzie & Sclater (1971). Unfortunately no accurate measurements of the rate of movement exist. The best is probably that of Brune (1968), who estimated that the slip rate was 11 cm/yr on the basis of the seismic moments of earthquakes on the fault. This estimate gives an angular rotation rate of $25{\cdot}4 \times 10^{-7}$ deg y^{-1}, a value which is probably too large (see Section 8 below).

At the eastern end of the North Anatolian Fault the fault plane solutions change from strike slip solutions to ones which involve approximately equal components of strike slip and overthrust, such as 1966.3.7, 1966.8.19 and 1968.4.29. This type of

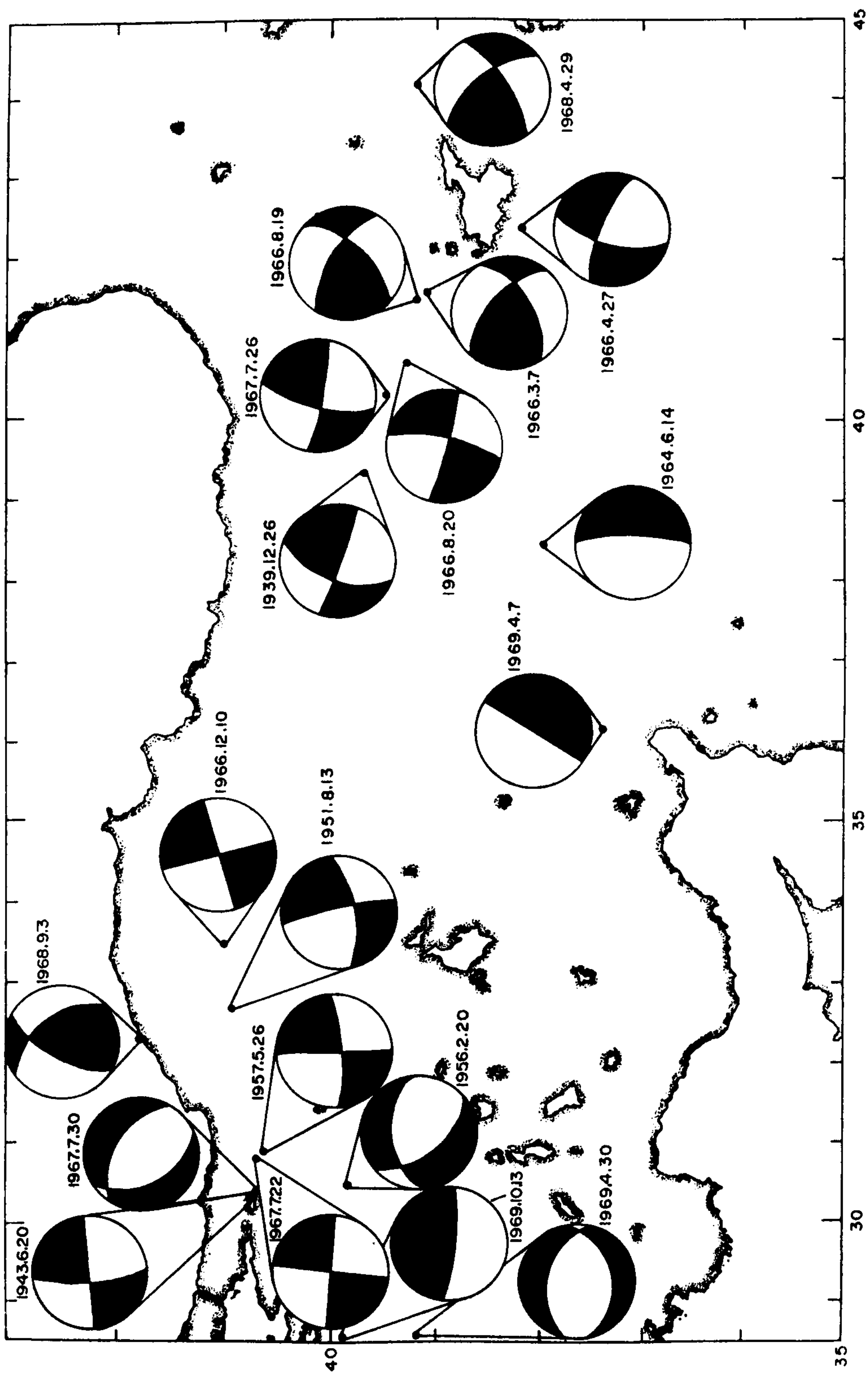


FIG. 25. Fault plane solutions of earthquakes in Turkey.

mechanism does not produce a straight surface break, and the shatter zone does not erode in the characteristic manner of the rest of the strike slip fault further west (Allen 1969, also see the *Times Atlas* 1968). As Allen has pointed out the change in the type of faulting occurs very abruptly, between the Varto earthquake of 1969.8.19 and its major aftershock on 1966.8.20 (Figs 5(g) and 25). The position of this triple junction based on Allen's field mapping agrees very well with that based on seismic evidence, and as he points out is the point at which a large fault striking south-west joins the North Anatolian Fault. This fault is part of what Ambraseys (1971) has called the Border Zone, which is marked by a zone of weak seismicity in Fig. 24. The zone runs from between the epicentres of 1966.8.19 and 1966.8.20 south-west to the Gulf of Iskenderum, and its activity between 1961 and 1971 (though not before) has been comparable with that of the North Anatolian fault. Ambraseys believes that there have been periods in the past 2000 years when large shocks frequently occurred in the Border Zone, perhaps even as frequently as they have on the North Anatolian Fault since 1939. In contrast to the work which has been carried out on the North Anatolian Fault, very little is known about the nature of the faulting in this region. The two poor fault plane solutions, 1964.6.14 and 1969.4.7, are similar and are not compatible with strike slip motion on a steep NE–SW plane. They could, however, be the result of strike slip motion on a plane with a shallow dip to the north since only one nodal plane is defined by either of these solutions. If this one plane is the auxiliary plane in both cases, then the solutions are consistent with left-handed strike slip in the Border Zone, but there is at present no means of deciding whether the single plane in these solutions is the fault or the auxiliary plane. Though nothing can be obtained at present from studies of the Border Zone itself, the change of mechanism at the eastern end of the North Anatolian Fault requires left-handed strike slip motion on the fault systems to account for the change in the strike of the slip vector (Section 8). The continuation of the Border Zone beyond the Gulf of

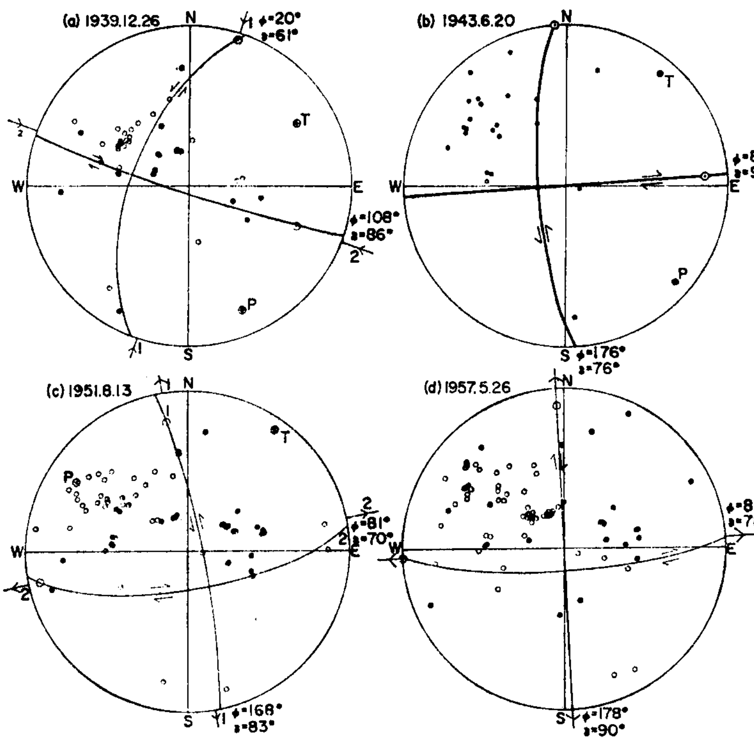


FIG. 26(a)–(d)

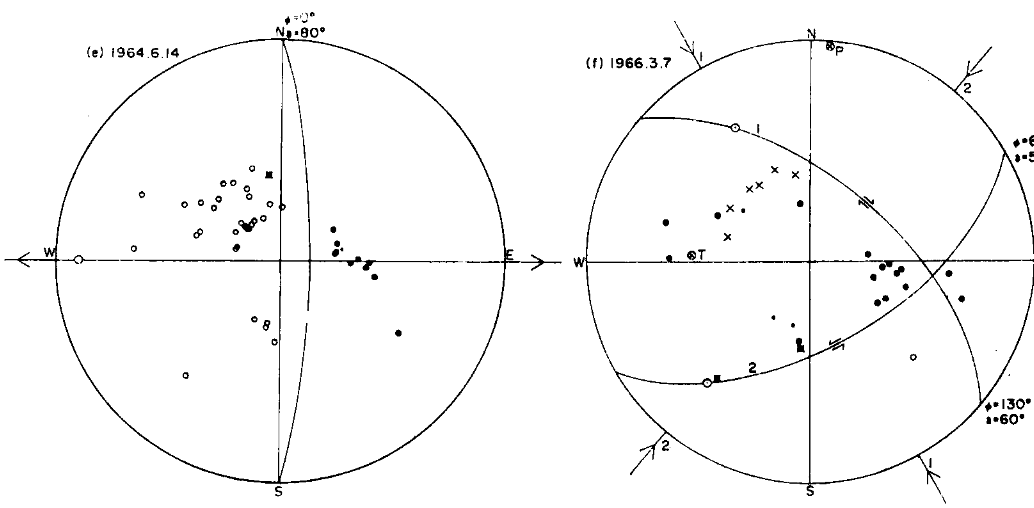


FIG. 26(e) and (f)

Iskenderun could either be the Dead Sea Rift and the faulting in Syria, or could extend in a loop through Cyprus to join the Cretan Arc near Rhodes. Allen (1969) favours the first alternative, whereas McKenzie (I) believed that the slip rate on the Border Zone was too rapid to be taken up by the Dead Sea system. He believed that the activity was continuous with that of Cyprus (Galanopoulos & Delibasis 1965; Ergin 1966), and that the southern boundary of the Turkish plate runs in a large loop from the Gulf of Iskenderun south of Cyprus to Rhodes. Though this interpretation is retained here there is no additional seismic evidence to favour this model rather than that of Allen (1969).

Though the North Anatolian Fault is the most active feature near the coast of the Black Sea, some shallow focus activity extends north-west from the Caucasus along the shores of the Black Sea through Crimea to Roumania and Bulgaria. Two solutions were obtained for earthquakes in this region, shown in Figs 27 and 9. If the auxiliary plane is taken to be the southward dipping plane which has approximately E–W strike for both solutions, then the fault plane dips beneath the continent in both cases. This choice therefore seems reasonable, especially for 1963.7.16 which lies on the southern edge of the Caucasus, but without surface breaks or special surveys of the aftershock distribution it remains merely a plausible guess. Whichever plane is the fault plane makes little difference to the relative motion between the Black Sea and the Eurasian plate, which must be in a north–south direction with the Black Sea moving towards Eurasia. This motion, must, however, be rather slow, since the seismicity of this boundary has been weak for most of this century (Gutenberg & Richter 1954, Figs 1 and 2). For this reason the Black Sea has been considered as part of the Eurasian plate in discussions of the tectonics of Greece, the Aegean and Turkey.

One remarkable feature of the seismicity near the margins of the Black Sea is the active source of intermediate earthquakes beneath a bend in the Carpathians in Roumania. Roman (1970) has relocated these shocks and has shown that they lie in a vertical slab striking NE–SW. This structure is on a smaller scale but otherwise very similar to the source of intermediate shocks studied by Nowroozi (1971) beneath the Hindu Kush. The difficulty in generating such slabs has already been discussed above in connection with the intermediate and deep focus shocks beneath the Tyrrhenian Sea, and in (I). In Roumania it is clear that the present motion of the Black Sea relative to Eurasia is unable to produce such a localized source of intermediate earthquakes. Such a source could, however, be produced by the rapid motion of a

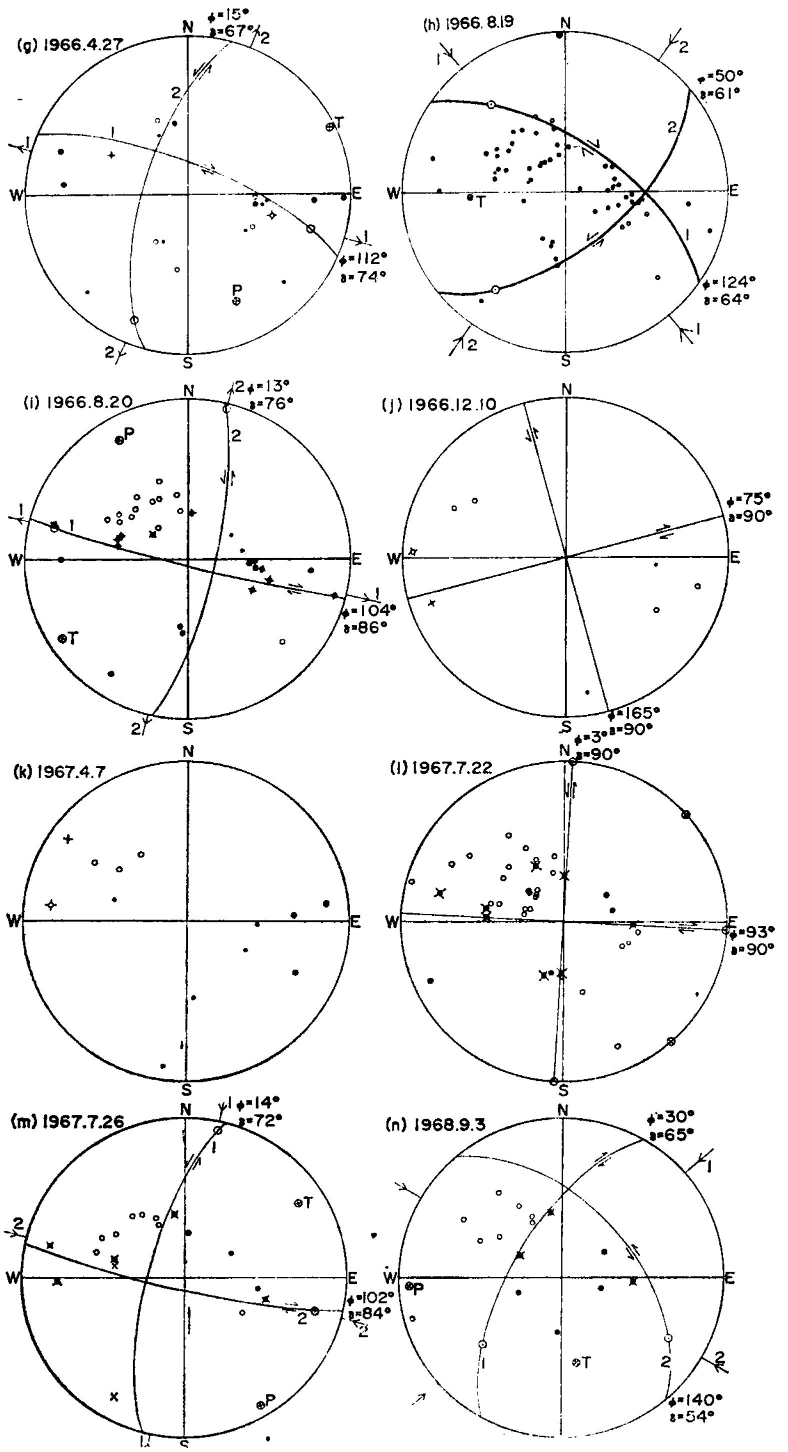


FIG. 26. Mechanisms for earthquakes in Fig. 25. (e)–(n) based on long period WWSSN observations, (c) and (d) on short period observations of Wickens and Hodgson (1967), (b) on short period observations of Ocal (1966) and (a) on those of Canitez and Ucer (1967b). All solutions in the mantle except (e) and (f) in the crust. The caption of Fig. 7 describes the conventions used in drawing the mechanisms.

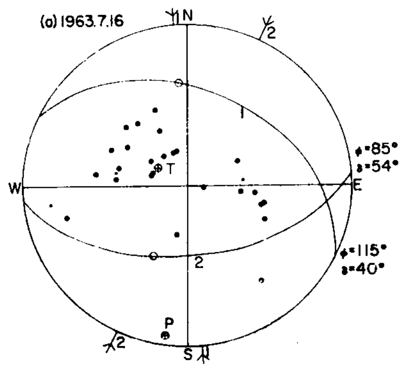


FIG. 27(a)

small plate containing the Carpathian mountains and surrounding regions south-eastward relative to the Black Sea, in the same way as the Aegean plate is now moving southward across the Eastern Mediterranean. Roman (1970) has discussed this possibility in some detail, and has shown that it is consistent with the known surface and subsurface geology of this part of the Carpathians. In particular the crust to the south-west of the shocks appears to be oceanic crust overlain by a large thickness of Miocene and more recent sediments. Consumption of such crust in the past could probably have produced the slab in which the seismic activity occurs. A sketch of this mechanism is shown in Fig. 28.

7. The Caucasus and Iran

The general seismicity of this area has been discussed by Gutenberg & Richter (1954) who pointed out that the active zone is much wider in Iran than in most other parts of the Alpide belt. They also recognized that intermediate shocks occur beneath southern Iran. These results have been confirmed by Barazangi & Dorman (1969) who plotted the USCGS epicentres, and by Nowroozi (1971) who relocated a large number of earthquakes in the area as part of a detailed study of its seismicity. The location and energy release of earthquakes has also been studied by Niazi & Basford (1968) and by Peronaci (1958). Ambraseys (1968) has discussed the seismicity of North-Central Iran from 850 to 1900 AD, and has shown that with one exception the areas active during this period have also been seismically active since 1900. The period for which good locations are available only extends from 1950 to 1970, and could be too short to mark even the most active boundaries.

The early locations in Gutenberg & Richter were not sufficiently accurate to resolve the wide belt of activity between 40° E and 60° E into separate zones. However the relocated epicentres obtained by Nowroozi (1971) and the USCGS epicentres in Fig. 29 show that most of the activity is concentrated in three bands. The northernmost of these runs from the Caucasus south-east across the Caspian Sea to join the middle band in north-eastern Iran. The middle and southern belts appear to originate as a single active zone near the eastern end of the North Anatolian Fault, and then divide into two bands, one following the Elburz mountains along the southern edge of the Caspian Sea, the other more southerly one continuing along the Zagros mountains to the Straits of Hormuz at the entrance to the Persian Gulf, then changing strike and continuing along the south coast of Iran and Pakistan. Nowroozi in his Fig. 3 also shows a band of epicentres in east central Iran with a strike of NNW–SSE. Some of these epicentres are also plotted in Fig. 29, but the existence of a distinct active zone is less clear. Such differences are not unexpected, since Nowroozi's data covers the years 1950–1965, whereas that in Fig. 29 is from 1961–1970.

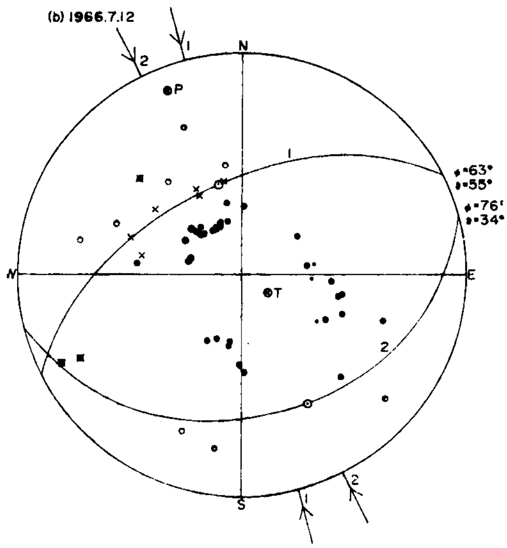


FIG. 27. Both shocks occurred on the northern margin of the Black Sea. Both mechanisms are based on long period WWSSN observations and are in the mantle.

A previous attempt to understand the tectonics of this region by the use of fault plane solutions has been made by Shirokova (1962, 1967), who used the records of stations in southern Russia as well as published polarity observations. Her solutions for shocks here and within the frontiers of the Soviet Union are likely to be more reliable than those obtained using the short period instruments commonly installed outside Russia because the broad-band low magnification instruments most frequently used in the Soviet Union give a more accurate record of ground motion than do the western short period instruments. Thus certain of Shirokova's fault plane solutions

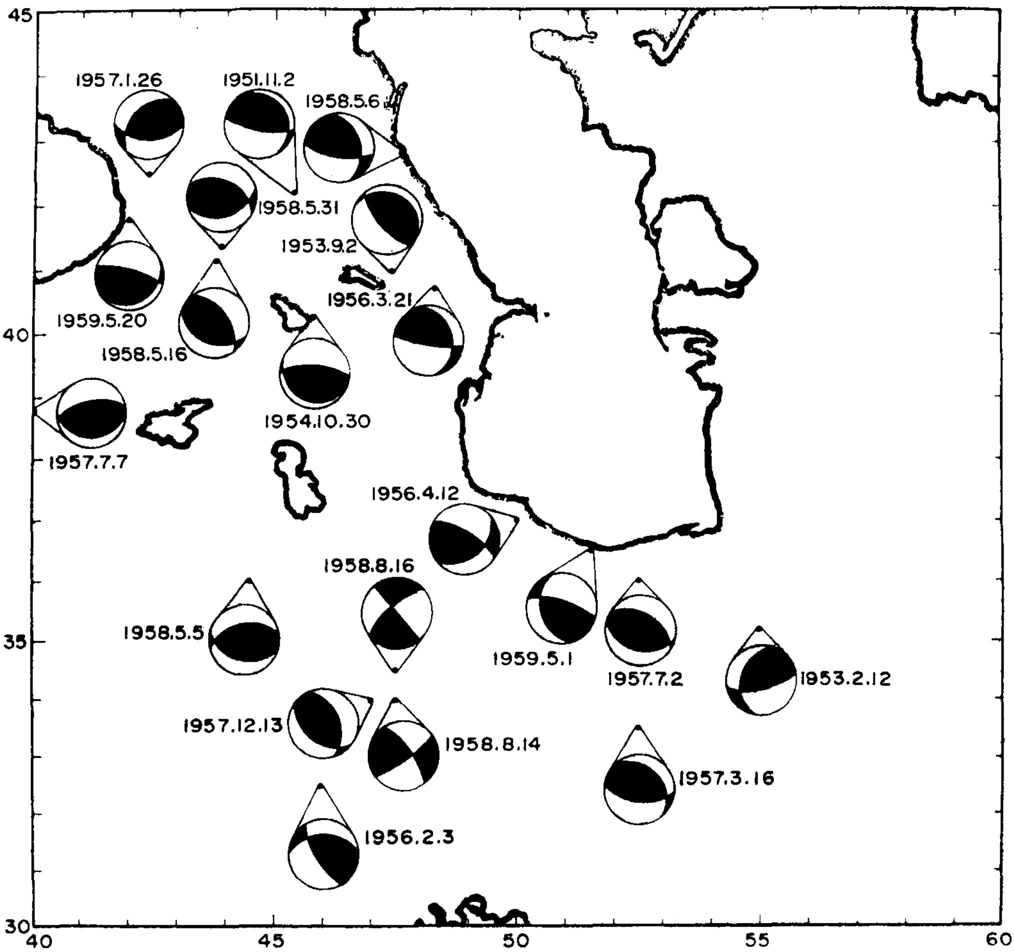


FIG. 30. Focal mechanisms taken from Shirokova (1962). Though she does not list the polarity observations at the seismograph stations used, these mechanisms are probably more reliable than those obtained with western short period instruments at this time.

the mechanism of the main shock of the Dasht-e-Bayaz earthquake on 1968.8.31 and of its major aftershock on 1968.9.1. His solution for the main shock agrees reasonably well with that in Fig. 32, but that for the aftershock is quite different. The onsets for the main shock were difficult to read in many cases because the shock was preceded by a small foreshock. The difference between the mechanism for the aftershock in Fig. 32 and that of Niazi (1969) appears to be principally due to the difference between the way the nodal planes are drawn rather than in the polarity observations themselves.

The principal problems concerning the present tectonics of Iran and southern Russia are the nature of the connection between the activity of Iran with that of Turkey, and the relative motions between the Arabian, Iranian, South Caspian and Eurasian plates. The boundaries of these plates in Fig. 2(b) are marked by fault systems rather than single faults, and the concept of plates in this region should be used with some caution since deformation is occurring at present over most of Iran. Despite these qualifications the plates in Fig. 2(b) are useful for understanding the general motions now taking place.

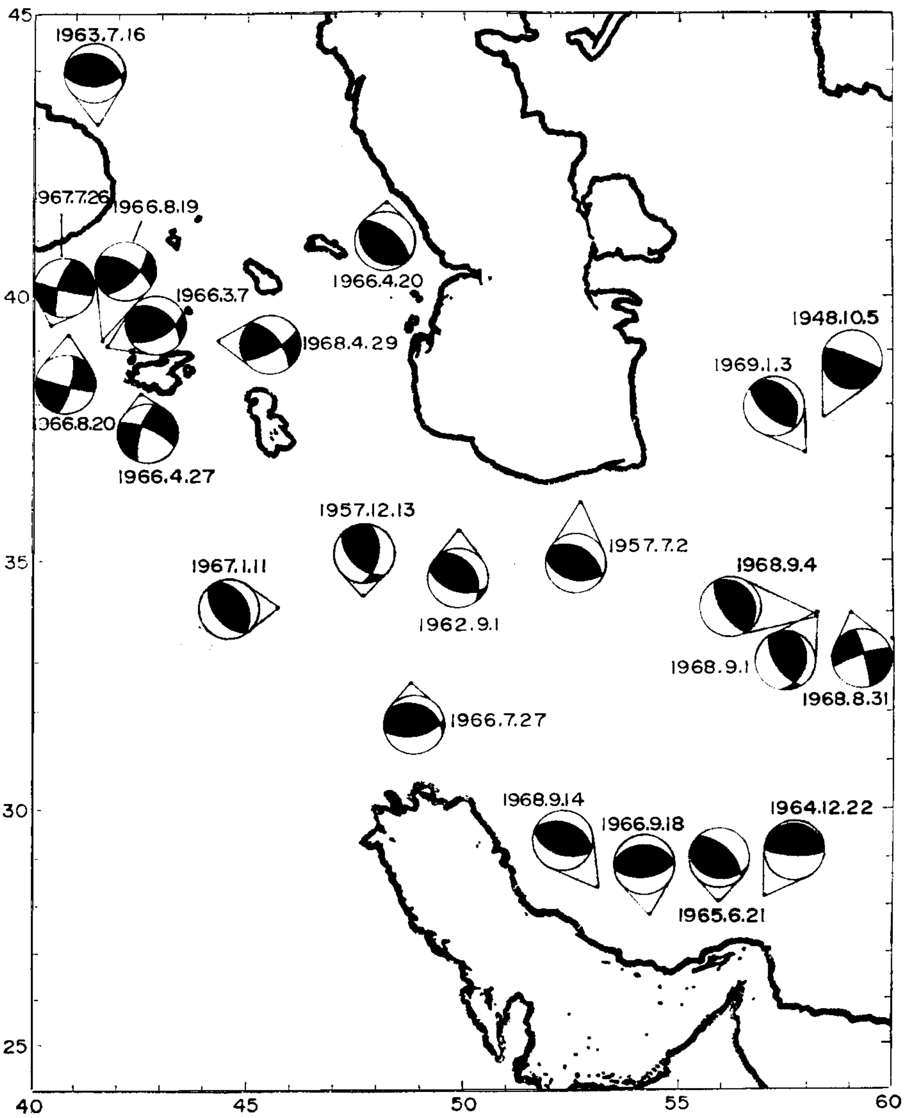


FIG. 31. Fault plane solutions for earthquakes in eastern Turkey, southern USSR and Iran. The individual solutions are shown in Fig. 32.

The connection between the eastern end of the North Anatolian Fault and the Border Zone, and the activity in the Caucasus further north is not yet clear from the focal mechanisms at present available. The most easterly shock which has occurred since 1963 was on 1968.4.29. Though no surface break has been mapped, Mortazavi (private communication) has mapped the faults in the epicentral region and shown that they strike NW–SE. If the plane of 1968.4.29 parallel to this direction is taken as the fault plane, then the mechanism and slip direction is remarkably similar to that of 1966.3.7 and 1966.8.19 further west on what is probably the south-eastward extension of the activity associated with the North Anatolian Fault. The shock of 1968.4.29 did not occur on this fault zone, however, but on one offset to the north-east. Therefore if the above choice of fault plane is correct this earthquake cannot be on

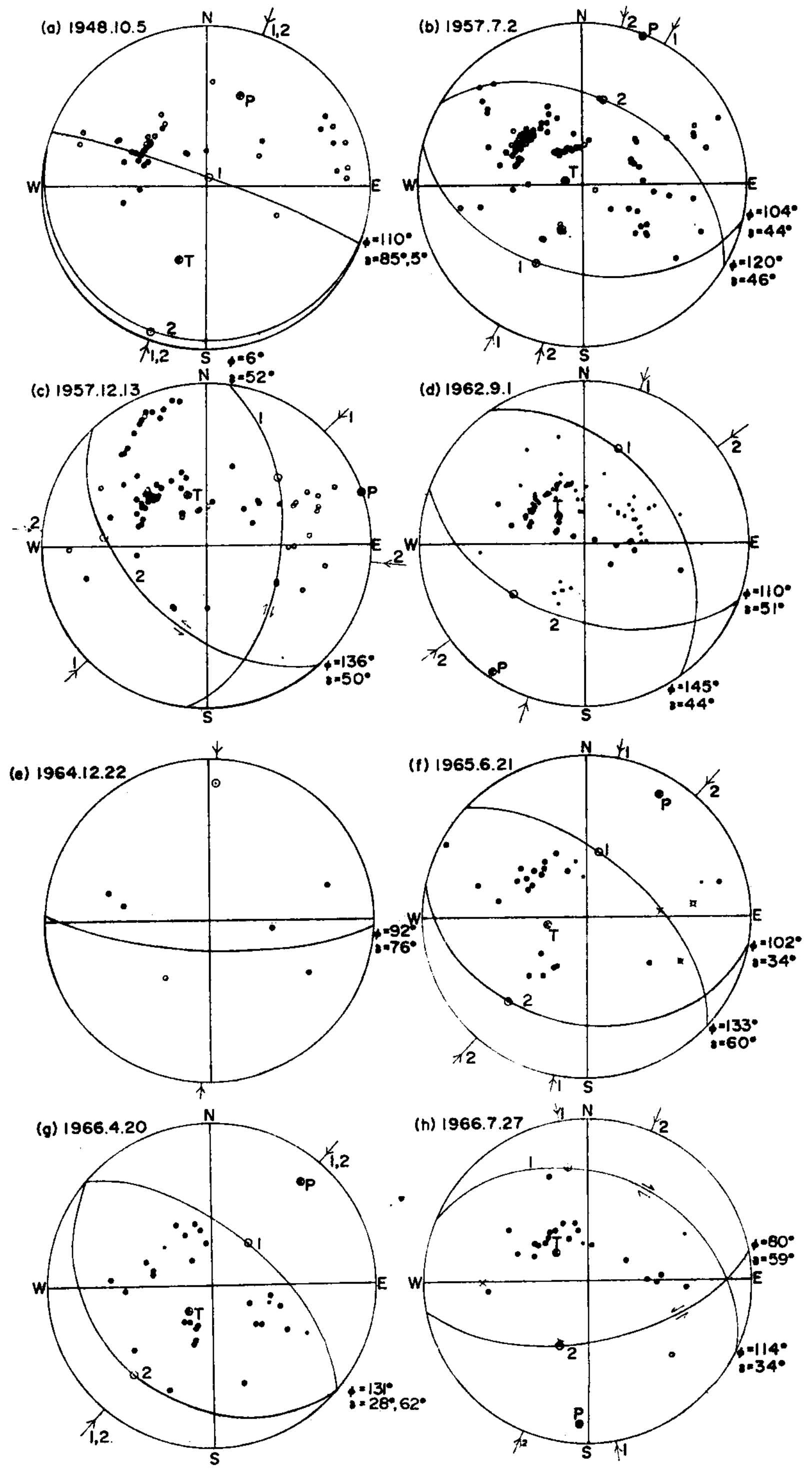

(a) 1948.10.5
ϕ=110°
δ=85°.5°
(b) 1957.7.2
ϕ=104°
δ=44°
ϕ=120°
δ=46°
(c) 1957.12.13
ϕ=6°
δ=52°
ϕ=136°
δ=50°
(d) 1962.9.1
ϕ=110°
δ=51°
ϕ=145°
δ=44°
(e) 1964.12.22
ϕ=92°
δ=76°
(f) 1965.6.21
ϕ=102°
δ=34°
ϕ=133°
δ=60°
(g) 1966.4.20
ϕ=131°
δ=28°, 62°
(h) 1966.7.27
ϕ=80°
δ=59°
ϕ=114°
δ=34°

FIG. 32(a)–(h)

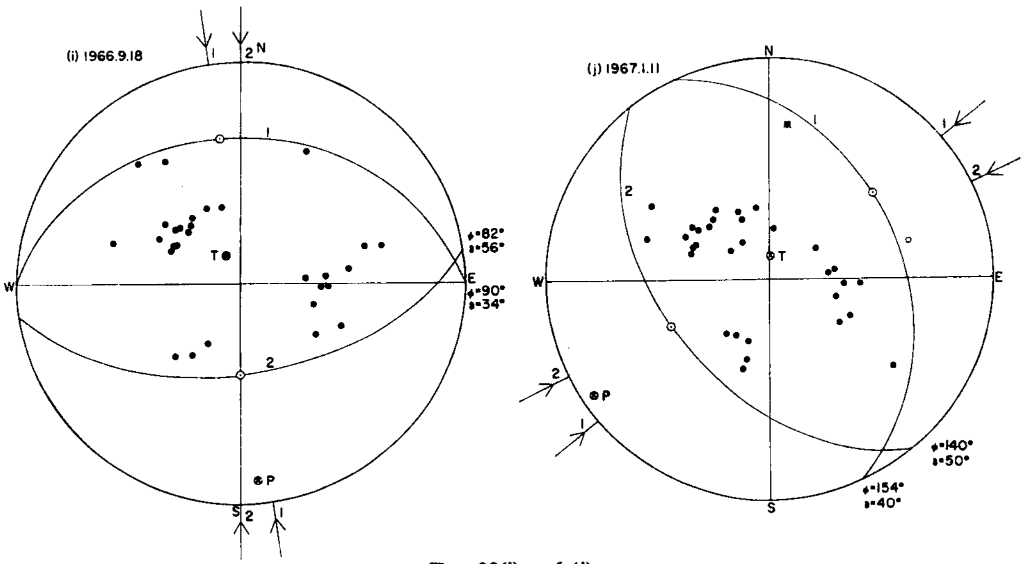


FIG. 32(i) and (j)

the active boundary joining the North Anatolian fault to the Caucasus further north, and indeed no shocks for which mechanisms are available appear to lie on this structure. If, however, the other plane of 1968.4.29 is taken to be the fault plane then the motion across the fault is in a north-east direction, with a large component of left-handed strike slip motion. This choice is less likely to be correct than the one mentioned above, since it requires a remarkable change in faulting mechanism in a small distance. If it is, however, then this shock could lie on the structure joining the North Anatolian fault to the Caucasus. Another fault which could serve this purpose is shown on Nowroozi's (1971) map and strikes NE–SW south-east of the Black Sea. This problem will remain unsettled until detailed maps can be combined with fault plane solutions to disentangle the possible forms of the connection.

In the Caucasus two WWSSN solutions are available, one shown in Fig. 27 on 1963.7.16 on the southern edge of the western part, and the other on 1966.4.20 in Figs 31 and 32 on the northern edge of the eastern mountains. If the fault plane is the northward dipping plane for 1963.7.16 and southward dipping plane for 1966.4.20, which is a reasonable choice if the position of the epicentres relative to the mountains is taken into account, then these two solutions show that the strike of the slip vector changes by about 40° between the two extremities of the mountain chain. This observation also shows that some active zone just join the Caucasus between these two shocks. Shirokova's solutions (1962, 1967) in Fig. 30 show the same feature and also demonstrate that thrusting still continues over most of the Caucasus range. The activity associated with the Caucasus extends eastward across the Caspian, then joins with the active belt south of the Caspian to form a belt running SE within which the shocks of 1948.10.5 and 1969.1.3 occurred. The mechanism of the first of these was published by Rustanovich & Shirokova (1964), who combined this solution with the results of levelling lines to show that the plane dipping at a shallow angle to the SW was the fault plane. The ambiguity cannot be resolved for 1969.1.3, but since both planes have similar strikes the projection of the slip vector is in a NE direction whichever is chosen. This direction and that for 1966.4.20 is not parallel to the motion direction obtained from the Arabia–Eurasia pole (Fig. 4). The Iranian and South Caspian plates are therefore not moving in the same direction relative to Europe as the Arabian plate. Two of the possible arrangements of plate motions are shown in Fig. 33. Fig. 33(a) is the simplest scheme which can explain

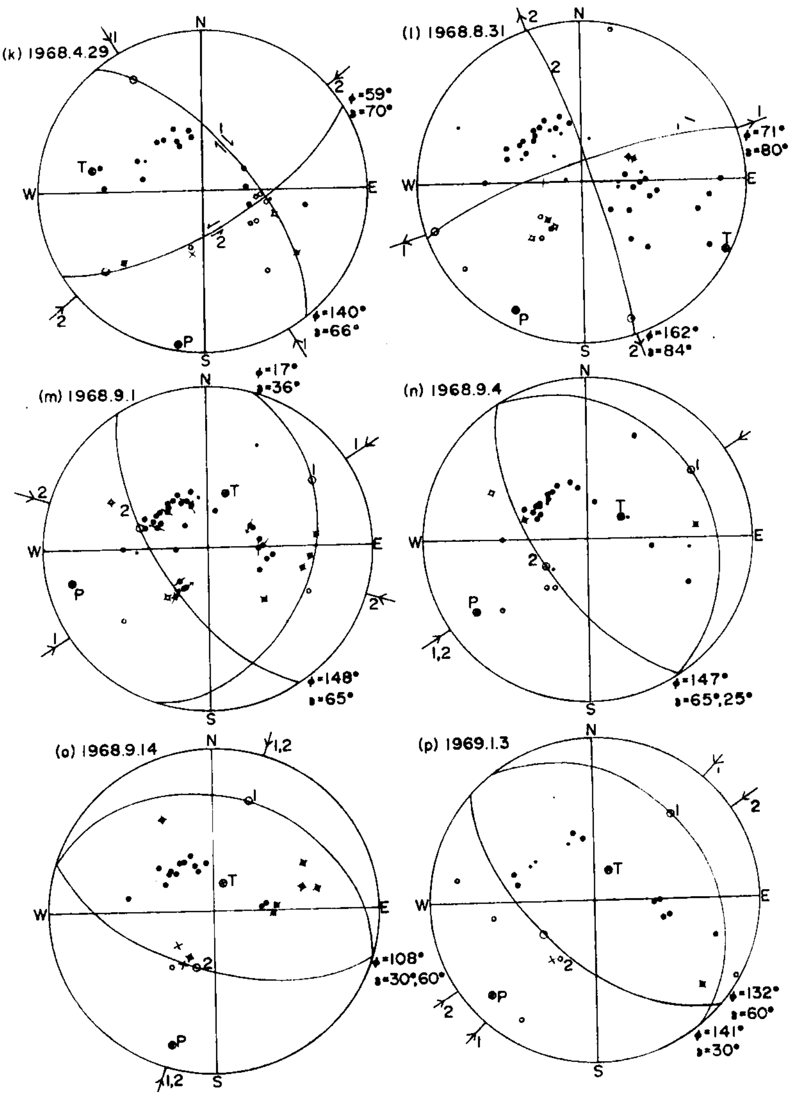


FIG. 32. Mechanisms of earthquakes in Fig. 31. Solutions (d)–(p) based on long period WWSSN reading, (b) and (c) on short period observations of Wickens & Hodgson (1967), (a) on short period observations of Rustanovitch & Shirokova (1964). Solutions (d), (g)–(k) and (m)–(p) in the crust; (a)–(c), (e), (f) and (l) in the mantle. For an account of the conventions used in the mechanisms see the caption of Fig. 7.

the width of the seismic zone. Fig. 33(b) is more complex, since the relative motions between A and C, and B and C, are quite different from that between A and B. The motions in Fig. 33(b) appear to be more common than Fig. 33(a) in active seismic belts generated by shortening continental crust.

Two shocks have occurred on the boundary between the Iranian and South Caspian plates. The best studied of these is 1962.9.1, the Buyin-Zara earthquake, whose surface break was mapped by Ambraseys (1963). He showed that the faulting had a large thrust component, with also some left lateral strike slip, and occurred on a fault striking 100° E. This is in excellent agreement with the fault plane solution in

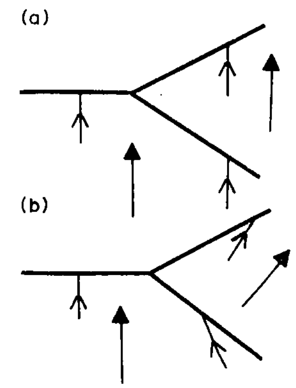


FIG. 33. The plate on the right of both (a) and (b) is a small plate such as the Turkish or Iranian plate. If it is moving in the same direction relative to the northern plate as the major plate to the south, case (a), then the slip vectors on both its northern and southern boundaries will be parallel. If, however, it is moving sideways and so moving continental crust away from the approaching major plates this movement will affect the slip vectors on both its northern and southern boundaries as shown in (b). The Turkish plate is clearly behaving in the manner shown in (b), and there is some evidence that the Iranian plate is also (see text).

Fig. 32 if plane 2 is chosen as the fault plane. The slip vector then strikes 55° E. The other shock for which a solution was obtained was 1957.7.2. All observations of polarity were obtained from short period records and therefore this solution is less reliable than the previous one. Neither nodal plane is well controlled, but the motion is consistent with that obtained from 1962.9.1.

The only other evidence of the nature of faulting on this boundary comes from Wellman's (1966) study of the aerial photographs of Iran and neighbouring countries. He believed that the motion between the Iranian and South Caspian plates was marked by a large left lateral fault (see also Nowroozi 1971). Though his interpretation of the observations has been questioned (Falcon 1967; Canitez 1969), the horizontal displacements which have occurred after earthquakes have in general supported Wellman's suggestion, though in some cases such as 1962.9.1, there has also been considerable vertical movement as well as strike slip. Thus what observations are at present available for this zone suggest the motion between Iran and the Southern Caspian is in a north-east direction, or the same as that between the southern Caspian and Eurasia. The strike of the plate boundary is such that this motion produces some left lateral strike slip movement as well as thrusting.

In north-eastern Iran there are further complications. Wellman (1966) and Nowroozi (1971) both show active left lateral faults south of the main activity. Slip on one of these faults resulted in the Dasht-e-Bayaz shock, whose fault plane solution is shown in Fig. 32. This solution together with the field reports of Ambraseys & Tchalenko (1969), and Bayer, Heuckroth & Karim (1969) shows that the faulting was left-handed strike slip on an east–west vertical plane, with no appreciable dip slip motion. The mechanism of two of the aftershocks, was, however, very different. The largest of these was on 1968.9.1 and was dominated by thrusting. The south-west dipping plane has a strike similar to that of the auxiliary plane of the main shock, and is therefore probably also the auxiliary plane. If this choice is correct the fault plane dips eastward, but its strike is poorly controlled. A smaller after-shock on 1968.9.4 has a similar mechanism to that of 1968.9.1, though its solution is less well controlled. These mechanisms could explain the branched nature of the surface break, since the east–west left lateral motion on the main break could be taken up by thrusting on a fault with a more northerly strike. Such an arrangement could permit a connection with the main active zone further north through a thrust fault striking north-west.

The third major seismic zone is the Zagros zone. Wellman (1966) believed that a large right-handed strike slip fault follows the coast of Iran from the Straits of Hormuz to 33° N, 46° E, and is about 200 km south west of the Zagros thrust. Nowroozi (1971) does not show this fault on his maps, and the seismicity does not appear to be restricted to a single line as it is in northern Turkey. It is therefore unlikely that the motion between Arabia and Iran is taken up on a single fault.

The interpretation of the fault plane solutions in this area is hindered by the absence of observed surface breaks, and by the velocity structure near the foci. The polarity observations for only one shock 1965.6.21, of the seven solutions obtained for this active belt could be divided into four quadrants by two orthogonal planes when the P wave velocity at the focus was assumed to be 8 km s^{-1}. One of the other solutions 1964.12.22, was too poorly defined to determine two planes, and one short period solution for 1957.7.2 contained sufficient inconsistencies near the nodal planes to permit a solution with a P-wave velocity of 8 km s^{-1} at the focus. The polarity observations from other shocks could only be fitted by two orthogonal planes if the P-wave velocity at the focus was 6·8 km s^{-1} or less. 6·8 km s^{-1} was used for all these shocks, and none of the resulting solutions had an appreciable component of strike slip motion. However, since the velocity structure in the crust is not known beneath this part of Iran, and since the depth of the foci in this region could be in error by at least 30 km, a large component of strike slip cannot be ruled out if the P-wave velocity at the focus was appreciably less than 6·8 km s^{-1}. Such motion is unlikely to occur on the east–west active zone in south-eastern Iran, since the one solution for a shock in the mantle on 1965.6.21 shows little strike slip motion. Furthermore intermediate focus earthquakes (Gutenberg & Richter 1954; Nowroozi 1971, Fig. 34) and andesitic volcanism has been active in the recent past (National

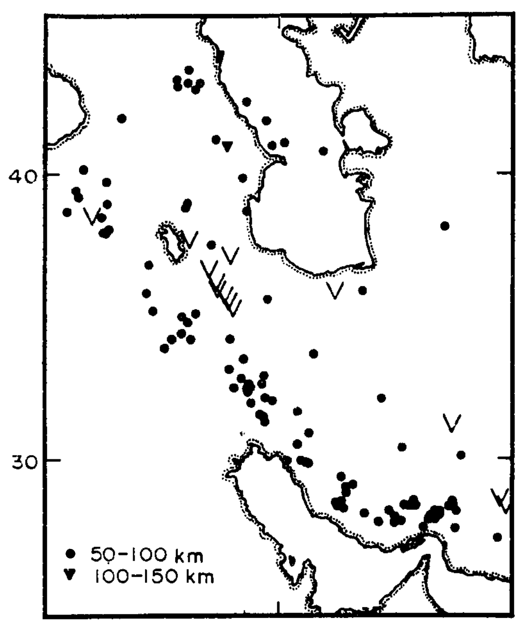


FIG. 34. Earthquakes below 50 km in the Eastern part of the Mediterranean region. Very few shocks occur below 100 km, and the few that do may be the result of destruction of oceanic lithosphere. Quaternary and active volcanoes are shown as V's, and their positions are taken from the National Iranian Oil Company Geological Map (1959) and from Gutenberg & Richter (1954).

Iranian Oil Company 1959). Therefore all observations are consistent with overthrusting on east–west faults, with perhaps a small component of strike slip. The possible existence of intermediate shocks also suggests that Iran is overthrusting the oceanic plate since continental crust cannot sink to produce a slab in the mantle. Most of the relative motion must therefore be taken up on a northward dipping thrust. The width of the active belt does, however, suggest that other faults are active in this region.

It is not possible to use any of these arguments to rule out a large strike slip component of motion on the active zone south-west of the Zagros thrust in western Iran. Such motion is not excluded by the lack of strike slip on the east–west part of this active zone further south, since the pole of relative motion between Arabia and Iran could be close to the boundary. A similar case is the motion between Africa and Eurasia discussed in Section 3, where thrusting near Gibraltar changes to strike slip near the Azores. It is clear from these arguments that there are insufficient observations of the nature of the deformation between Iran and Arabia to determine the direction of movement of these two plates in Western Iran. However, the motions between the Iranian, South Caspian and Eurasian plates are closely related, and since the observations in northern Iran favour the motions shown in Fig. 33(b) rather than Fig. 33(a), a component of right-handed strike slip is probably present across the Zagros active belt. The activity of Iran continues eastwards to join the seismic belts crossing northern India and Tibet. The activity in these regions has been studied by Fitch (1970) east of 75° E, but it is not clear from his solutions or from the seismicity maps of Barazangi & Dorman (1969) and Nowroozi (1971) how the belts crossing North and Southern Iran join to the activity further east. Complications similar to those in the Aegean, Turkey and Iran must be present in Central Asia to account for the extension now taking place near and beneath Lake Baikal and the large strike slip component of the motion observed after the 1957 Mongolian earthquake (Florensov & Solonenko 1963). The solutions obtained by Fitch (1970) show that the tectonics of other parts of Central Asia is not related in any simple manner to the motion of India relative to Eurasia. It is therefore unlikely that the remarkable motions in Turkey and Iran are peculiar to the western end of the Alpide–Himalayan belt, even though the observations at present available in Central Asia are too sparse to understand the motions in detail. If this is indeed the case, then the motions in ancient mountain belts such as the Caledonides were likely to have been equally complicated.

8. The pattern of tectonic deformation and its history

The plate boundaries and their motions relative to Europe are shown in Fig. 2(b). The plate margins in the Caucasus and in Iran are rarely well defined, and are still poorly known. The lines marking plate boundaries in these areas are therefore schematic only, and cannot yet be drawn with any accuracy. The arrows on the plates show the direction of motion relative to Europe, and their length is approximately proportional to their velocity also relative to Europe.

The most striking feature of Fig. 2(b) is the tendency of all small plates to move away from a north–south line through the Caucasus. The motion of plates to the west of this line has already been discussed in (I); that of those to the east is less clear and involves more thrusting, but appears to follow a similar pattern. The symmetry of the motions strongly suggests that the pattern of deformation is the result of the collision between northern Syria and the continental part of the Eurasian plate. The motions of the small plates to the east and west of the Caucasus are such as to move continental crust away from Asia Minor towards either the Mediterranean or the Arabian Sea. In the west the consequences of this motion are relatively simple, since there is a considerable area of what is presumed to be oceanic crust in the Eastern

Mediterranean. Hence the motion between Africa and Arabia, and Eurasia can be taken up by consumption of oceanic plate rather than by thickening of continental plate near the Caucasus.

To the east the deformation is less simple because it is less easy for the continental material which has been wedged aside by the northward movement of Arabia to find any oceanic plate to consume. The fragment of possible oceanic plate that remains in the southern Caspian (Neprochnov 1968) appears to be insufficient to produce a strong effect, and the Arabian Sea is a considerable distance from the Caucasus and also is difficult to reach because of obstruction by the rigid continental mass of the Arabian Shield. Therefore the eastward movement of the two small plates to the east of the Caucasus is probably principally taken up by thrusting within the continental plates, in North and South Iran, in Iraq, in southern Russia and in Afghanistan; and only to a small extent by consumption of oceanic plate along the southern coasts of Iran and Pakistan. If this explanation is correct large strike slip faults may well have existed in Iran before it became trapped between the continental masses of Arabia and Asia.

The symmetry about an axis through the Caucasus is also striking on a topographic map of Asia Minor. The high fold mountains in Turkey and Iran lie on either side of the shield of Iraq and Syria, and the topography has been produced by the shield pushing aside and folding these weaker parts of the Eurasian plate.

This pattern of deformation was first recognized by Hill & Dibblee (1953) on the San Andreas fault in California. They mapped two large left-handed strike slip faults in southern California which intersect the San Andreas where it bends (Fig. 35) through the transverse ranges. They interpreted their observations in terms of a regional stress field, and believed that the system of conjugate faults in Fig. 35 was produced by compression in a north–south direction. The observations are, however, equally well explained by ideas discussed above, since one consequence of motion on the faults described by Hill & Dibblee is to move continental material away from the bend in the San Andreas, where overthrusting would otherwise have to occur on a large scale. This pattern of deformation is therefore rather similar, though on a much smaller scale, to that at present taking place in the eastern Mediterranean and south-west Asia. The existence of such features is a simple consequence of the inability of trenches to consume continental crust. There is no

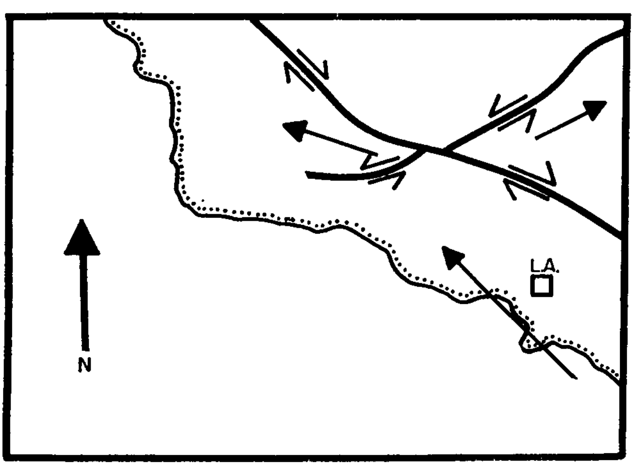


FIG. 35. A copy of a figure due to Hill & Dibblee (1953) redrawn to resemble Fig. 2(b); to show that the type of faulting they described closely resembles that in Fig. 2(b) though it is on a smaller scale. L.A. is Los Angeles. The arrows show the relative motion between the blocks and the American plate.

evidence to support a regional stress field of the type proposed by Hill & Dibblee in the Middle East, though it is not in conflict with the observations. Such stresses would, however, have to be maintained by the plate motions, since otherwise they would quickly be relaxed by the stress released by earthquakes. It is therefore simpler at present to neglect the stress field, which must be exceedingly complicated, and account for the features produced by major deformation directly in terms of plate motions.

If the scheme in Fig. 2(b) is correct it is simple to calculate the motions of the Turkish and Aegean plates relative to Africa and Eurasia, using the rate of motion between Africa, Arabia and Eurasia and the directions of motion obtained from the fault plane solutions. At the eastern end of the North Anatolian Fault there is a stable triple junction between Turkey, Arabia and Eurasia. If the active zone between Turkey and Arabia is assumed to be a strike slip fault, as Allen (1969) has suggested, then the rates of motion on both the Border Zone and the North Anatolian Fault are easily obtained from the velocity triangle (Fig. 36(b)). Some of the motion is taken up further north by thrusting in the Caucasus, and therefore the rates given in Fig. 36(b) are the maximum rates. There is therefore a considerable difference between Brune's (1968) estimate of 11 cm/yr on the North Anatolian fault and 4 cm/yr obtained from the velocity triangle. For the Border Zone the contrast is even more striking, since few large shocks have occurred on it since instrumental recording began, yet the rate of movement is 5 cm/yr if Fig. 36(b) is correct. These differences are probably caused by the sporadic nature of strain release by earthquakes on these boundaries. Allen (1969) has remarked that the high activity on the North Anatolian Fault appears to have started in 1938, and is not typical of its behaviour before this date. Ambraseys (1971) has also shown that there are as many historical records of disastrous shocks in the Border Zone as there are on the North Anatolian Fault. It is therefore probably not possible to obtain accurate estimates of the slip rates in this area from the seismicity, and for this reason the present

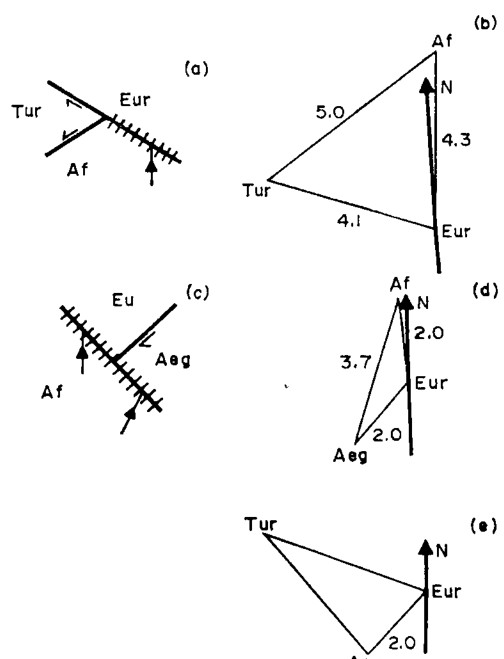


FIG. 36. Vector velocity triangles for motions between Africa, Eurasia, the Aegean and Turkey (see text). Numbers give rates in cm/yr.

difference in seismicity between the north and south boundaries of the Turkish plate is not in conflict with Fig. 36(b). The other place where a similar estimate can be carried out is at the junction between the northern boundary of the Aegean plate and the Cretan Arc. The mean direction of motion between the Aegean plate and the Eurasian plate to the north is 220° E, whereas that between the Aegean and African plates is 197° E. If this difference is a result of motion between Africa and Eurasia, then the vector velocity triangle Fig. 36(d) can be drawn. Unfortunately the difference between the Africa–Aegean and Eurasia–Aegean directions is only 23°, therefore the distance of the vertex labelled 'Aegean ' in Fig. 36(d) from Eurasia is uncertain. The motion between Turkey and the Aegean can now be obtained by using Fig. 36(b) and (d), and is shown in Fig. 36(e). Since the fault plane solutions show that the motion is in a north–south direction, this triangle is a check on the accuracy of the other two. The calculated direction is about 45° different from that observed, and therefore none of the velocities are particularly accurate. If the velocities in Fig. 36(b) are decreased by a factor of 1·5 and those in Fig. 36(d) are increased by the same amount the Turkey–Aegean direction agrees with the observations. This calculation therefore provides some test of the accuracy of these triangles. If their accuracy could be improved then the calculations should be carried out with angular velocity vectors, not velocity tirangles, since the plates have finite extent. Such subtleties do not at present seem worth including.

One other crude check on these calculations can be carried out. The depth of the deepest shocks beneath the Aegean is about 250 km, and occur about 250 km behind the Cretan trench. The length of the slab is therefore 350 km and if this corresponds to the distance the Aegean has moved towards Africa in the last 10 My, then the rate of consumption must be 3·5 cm/yr. This figure agrees rather well with that of 3·7 cm/yr in Fig. 37(d), but should not be trusted on this account. In particular the motion of a small plate such as the Aegean is unlikely to remain constant for long periods. Since the velocity may either increase or decrease with time, 3·5 cm/yr is not either the upper or the lower limit of the range of possible velocities. Similar estimates are not possible at present east of the Caucasus because the motion directions are too poorly known.

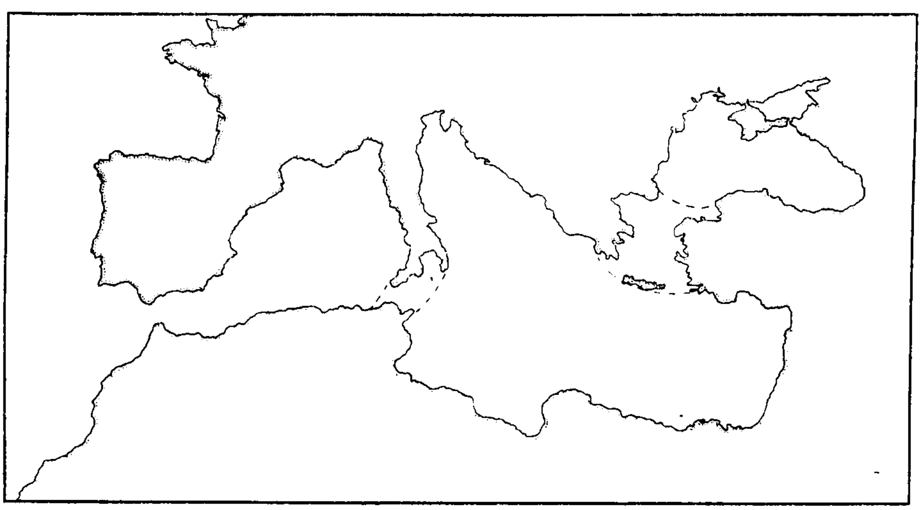


FIG. 37. An attempt to reconstruct the arrangement of the plates around the Mediterranean 10 My before present. The positions of Greece and Turkey are likely to be more reliable than that of Italy. All islands except Crete have been omitted because of lack of evidence. Many of them contain old rocks and must therefore have existed at this period.

Though it does not seem likely that the existence of the boundaries in Fig. 2(b) is the result of a regional stress field, the motions of the plates must be maintained by forces which act on them. There is no reason to believe that they are sliding downhill, and therefore the forces maintaining the motion must act either on their bases or on the plate boundaries themselves. Though there are now strong arguments in favour of the major plate motions being maintained by convection in the upper mantle (McKenzie 1969a), it is difficult to believe that this is the mechanism by which the motions of small plates such as the Aegean and Turkish plates are moved. One difficulty is that the viscous resistance to small-scale convective motions is very great. Another is that there are a number of small plates in the Eastern Mediterranean which have been produced by the collision of two continents, but it is difficult to understand how their existence could affect the motions below. If therefore the motions of small plates are due to convection in the Upper Mantle there should be other parts of the Earth's surface in the middle of shield and oceanic regions where the plate geometry and motions are as complicated as those in the Eastern Mediterranean. There are no obvious cases, though the East Indies is still a possible candidate, and therefore the complicated motions are most easily maintained by forces acting on the edges of the plate. This mechanism also permits complications of the type which now exist in the Eastern Mediterranean to be produced when parts of two continents collide. If this argument is correct then large forces must act near the plate boundaries, and since the complications are the result of a continental collision, and not of fragmentation, the axis of greatest principal stress will be close to the horizontal. Such a stress field is consistent with the mechanisms in Turkey, the Caucasus and Iran, though the direction cannot be obtained from the mechanisms themselves (McKenzie 1969b). It can also account for the four thrust solutions, two in Turkey, one in the Aegean and one in Greece discussed in Section 5.

It is at present the fashion to attempt reconstructions of the Mediterranean region at periods as far back as the Upper Cretaceous (Smith 1971; Hsü & Schlanger 1971). This author does not believe there is sufficient evidence at present for such attempts to be useful. There are essentially two difficulties which must be overcome before any such arrangements can be based on sound principles. The positions of plate boundaries as a function of time must be disentangled from the geological record and very extensive palaeomagnetic measurements must be undertaken. There is no reason to believe that the present plate boundaries in the Mediterranean have remained unchanged for long periods, and any changes to them profoundly affect the reconstructions. It is perhaps helpful to illustrate some of the difficulties by using the rates and directions in Fig. 36 to produce a reconstruction on the assumption that the present rates remained unchanged for the last 10 My. For this purpose the rates in Fig. 36(b) have been reduced by a factor of $1 \cdot 5$ and those in Fig. 36(d) increased by the same factor. The positions of the African, Aegean and Turkish plates can then be obtained relative to Europe, and are shown in Fig. 37. Cyprus, Corsica, Sardinia and Sicily have been omitted because there is no evidence for their relative positions. That of Italy has been obtained by moving the southern part 500 km westward relative to Africa. Such a displacement is required because the length of the dipping slab beneath the Tyrrhenian Sea is 500 km, and probably has been generated in the last 10 My. The northern part of Italy was taken to be attached to Europe. The evidence on which this reconstruction is based is very poor compared with those based on magnetic lineations (Pitman, Talwani & Heirtzler 1971; McKenzie & Sclater 1971), and therefore the plate positions in Fig. 37 may well be incorrect. Despite these cautions the Mediterranean reconstruction has some interesting features. The arc formed by Yugoslavia, Greece and Turkey is relatively simple and shows that the present shape of the west coast of Greece and the Cretan Arc are the result of the motions of the Aegean plate. Furthermore the motion has stretched much of what was once a land mass connecting Greece and Turkey. The thinning of continental crust by normal faulting in the Aegean region has resulted in large-scale subsidence to form the present shallow sea. Italy and perhaps Sicily form a barrier between the two basins, and the Straits of Gibraltar were probably also closed at this period. Hence the evaporites of the western Mediterranean found during the Deep Sea Drilling Project are not in conflict with this reconstruction. A good test of the arrangements in Fig. 37 should be provided by the palaeomagnetism of Italian lavas, since these should show a considerable anticlockwise rotation.

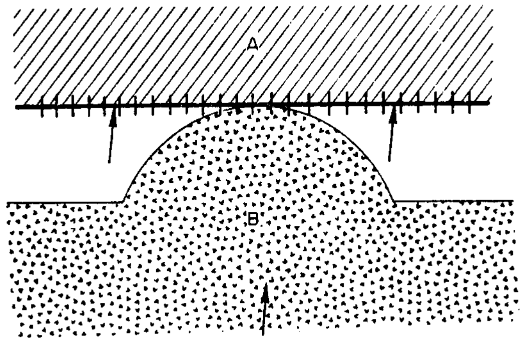
It appears unlikely that present motions have continued for much more than 10 My. The normal faulting in Greece associated with the present movements began in the Miocene (Aubouin *et al.* 1963), and the simple shape of the Cretan Arc in Fig. 37 suggests that it formed the southern boundary of a single plate before this period. The break up of this plate to the north was produced by the movement of Arabia northwards, and it is in Asia Minor that the geological record of this collision should be best recorded. It is perhaps worth observing that the formation of a new small plate by fracturing a previously continuous larger plate could thrust oceanic crust on top of a thick pile of sediments provided the boundary of the new plate lies within oceanic crust. Thus the formation of a new plate boundary provides an opportunity to form ophiolites which is normally lacking. Most mechanisms which have been suggested for thrusting oceanic crust on top of sediments involve two-dimensional models (Dewey & Bird 1971) and some sudden change in plate geometry or direction of underthrusting. The mechanism suggested above essentially depends on three-dimensional motions, and can therefore produce more complicated arrangements of fragments of oceanic crust.

The final problem of some considerable importance concerns the usefulness of this study as a guide to the type of deformation now occurring in other continental areas, and that which has occurred in older mountain belts. Though the area covered by this study is considerably greater than that of any geological field mapping project, it still only covers a minor part of the Alpide–Himalayan belt. It is therefore of interest to enquire if a similar pattern of deformation is at present occurring in other parts of this active belt. Unfortunately little work has yet been carried out on either the earthquake mechanisms or the tectonic geology of most of Central Asia. Fitch (1970), Balakina *et al.* (1967) and Shirokova (1967) have published fault plane solutions for this region, and apart from the overthrust solutions on the Himalayan front, have been unable to find any simple pattern of deformation. The presence of extensional features such as Lake Baikal associated with a fold mountain belt produced by shortening does not suggest that the deformation is simple. There is a similar difficulty in accounting for the existence of a large strike slip component on faults such as the fault in Mongolia which slipped in 1957 (Florensov & Solonenko 1963) unless a small plate is moving in a direction different from that of India towards Asia. Therefore the tectonics of the Mediterranean region do not appear to be specially complicated. Nor does the tectonics of older mountain belts appear to be simpler. The geometry of the main alpide chains of Italy, France and Switzerland is difficult to relate to the motion of Africa and Eurasia, and the transverse structures such as the Uinta mountains in the North American Rockies suggest that complications exist here also. Therefore there do not at present appear to be strong reasons for believing that the present tectonics of the Mediterranean are atypically complicated. In the absence of such reasons it is clearly dangerous to use any features of local deformation as a guide to the relative motion of major plates.

9. Conclusion

The reason for this study was to discover what happens when continents collide. This process is believed to have produced most wide fold mountain belts such as the Himalayas. Though mountains such as the Andes are unlikely to have a similar

(a)



(b)

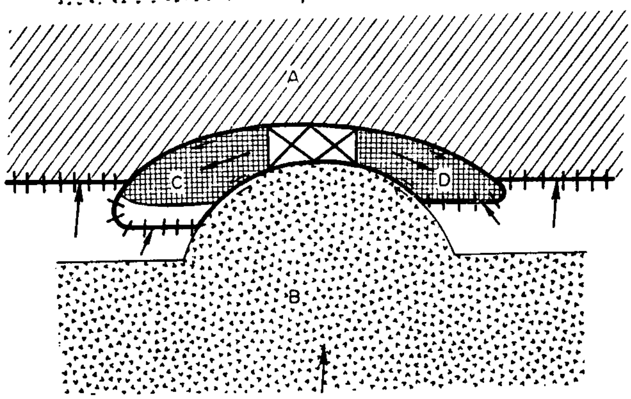


FIG. 38. Sketches to show the sequence of events when a protruberance on a continent B collides with A. (a) a trench lies on the continental margin of A and consumes the sea floor of plate B. One point on the continent on plate B reaches the trench, consumption can no longer continue because continental material will not sink through the mantle, (b) two new small plates form to move the continental material away from the point where the two continents have met. Sea floor on plate B is consumed to the left and right of the point where the collision has occurred. The formation of the new plates C and D required new plate boundaries to form.

origin, many ancient mountain belts appear to have been formed by the approach of two continental masses and the deformation of the sediments in the ocean in between. The work discussed in this paper suggests what might happen during such a collision, and is illustrated in the cartoons in Fig. 38. The evolution is taken to begin with a trench forming the plate margin between plates A and B, and consuming the oceanic part of B. At this stage there is only one plate boundary and two plates. In general approaching continental edges will not have the same shape, and therefore when the first point of the continent on B meets A pieces of oceanic crust will remain between them (Fig. 38(a)). Further consumption along the plate margin between A and B is impossible because continental material will not sink into the mantle (McKenzie 1969a). Therefore either the continental crust is thickened, or small plates are formed to consume the remaining ocean or both processes take place. Two small plates are shown in Fig. 38(b); when they meet the continental edge of B more small plates will be produced until all the sea floor between the two continents is consumed. If the motion between A and B still continues the mobile belt between them will consist of the smashed fragments of many small plates, and is unlikely to behave in a simple manner when further shortened. The Mediterranean region shows examples of all three stages. In the west the African continent has not yet collided with the Eurasian (except at Gibraltar). In the Eastern Mediterranean the Aegean plate is consuming the oceanic part of the African plate. In Iran there is no sea floor remaining between

Arabia and Iran, and large scale thrusting is therefore taking place. Though this description is a simplified outline of the events, it is probably sufficient for many purposes. The most likely time for sea floor to be emplaced by thrusting is that at which the small plates C and D form by fracturing A, since there is no reason why the new plate boundaries should not cross oceanic lithosphere.

The whole process resembles the impaction of some hard die on a soft piece of metal. The metal is extruded on all sides of the die, since it is relatively incompressible. Flow of this type is driven by the boundary forces, and is remarkably similar to that discussed in Fig. 38. It is, however, important to notice that the boundary force produced by continental crust is due not to its strength but to its buoyancy. The surface expression of this buoyancy is a mountain belt, and it is therefore possible to regard the gravitational forces caused by surface elevations as the driving forces for the motion of small plates such as the Aegean.*

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* Dielines of Figs 1 and 9 at a scale of 1:12 233 000, Figs 15–18 at 1: 1 million and Figs 24, 25, 29 and 31 at 1:2·5 million can be provided at cost.

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