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Extensional–compressional tectonics associated with the Aegean Arc: comparison with the Andean Cordillera of south Peru – north Bolivia

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Extensional–compressional tectonics are analysed along the convergent boundaries of the Aegean Arc and of the Andean Cordillera of south Peru and north Bolivia. This paper deals only with deformations of Miocene, Pliocene and Pleistocene ages.

The prevailing régime in the Aegean basin is extensional. Yet two compressional tectonic phases of U. Miocene–L. Pliocene and of L. Pleistocene age have been recorded. They are also known in many places all around the Mediterranean. Thus forces that induce compressional tectonics in the Aegean seem to have their origin outside the region. They seem to be linked with changes in the conditions of convergence (change in rate?) between the Arabo-African and Eurasian plates.

Beneath the Andean Cordillera of south Peru – north Bolivia, the subduction zone plunges at an angle of 25–30° as it does under the Aegean Arc. There, as in the Aegean Sea, tectonics are predominantly extensional and several short-lived compressional phases break up this extensional régime. The latter may also be induced by changes of the conditions of convergence, here between the Nazca and South American plates.

Beneath central Peru, north of the Nazca Ridge, the subduction zone plunges at a small angle, about 10°. There, neotectonic deformations are different. In the Recent they are essentially compressional or expressed by strike-slip faulting.

Geological data from the Aegean and Andes seem to demonstrate that the compressional deformation in the crust above subduction zones is strongly controlled by the stress conditions along convergent boundaries.

Extensional tectonics are associated with subduction zones. This is manifest in the marginal sea basins of the W Pacific arcs as well as in the continental basins (Basin and Range, Altiplano) close to the E Pacific Cordilleras. Basins are also associated with the Mesogean Arcs: the Tyrrhenian Sea with the Calabrian Arc, the Aegean Sea with the Aegean Arc, the Jaz Murian basin with the Makran Arc.

Within continental collision zones, deformation is more complicated. Nevertheless, extensional tectonics are associated with the Himalayan collision in regions north of Tibet.

With the two chosen examples, the Aegean and Andes, I show that continental crust situated above subduction zones is, in space and time, predominantly stretched. Yet the extensional régime is interrupted by short-lived periods of compressional tectonics.

I. Extensional–compressional tectonics associated with the Aegean Arc

A neotectonic analysis of the Aegean Arc has been undertaken by our team (Pegoraro 1972; Philip 1974; Sorel 1976; Carey 1976; Lemeille 1977; Sébrier 1977; Jarrige 1978; Gauthier 1979). The analytical methods applied and the principal results have already been published (Mercier 1979; Keraudren 1979; Dufaure *et al.* 1979; Bellon *et al.* 1979; Carey 1979; Mercier

1976; Mercier *et al.* 1979*a*). The last two publications are referred to respectively as A and B in this paper, especially for detailed illustrations of tectonic structures. Readers should consult these publications for more precise data and detailed references, and also two other recent papers (Le Pichon & Angelier 1979; Angelier 1979).

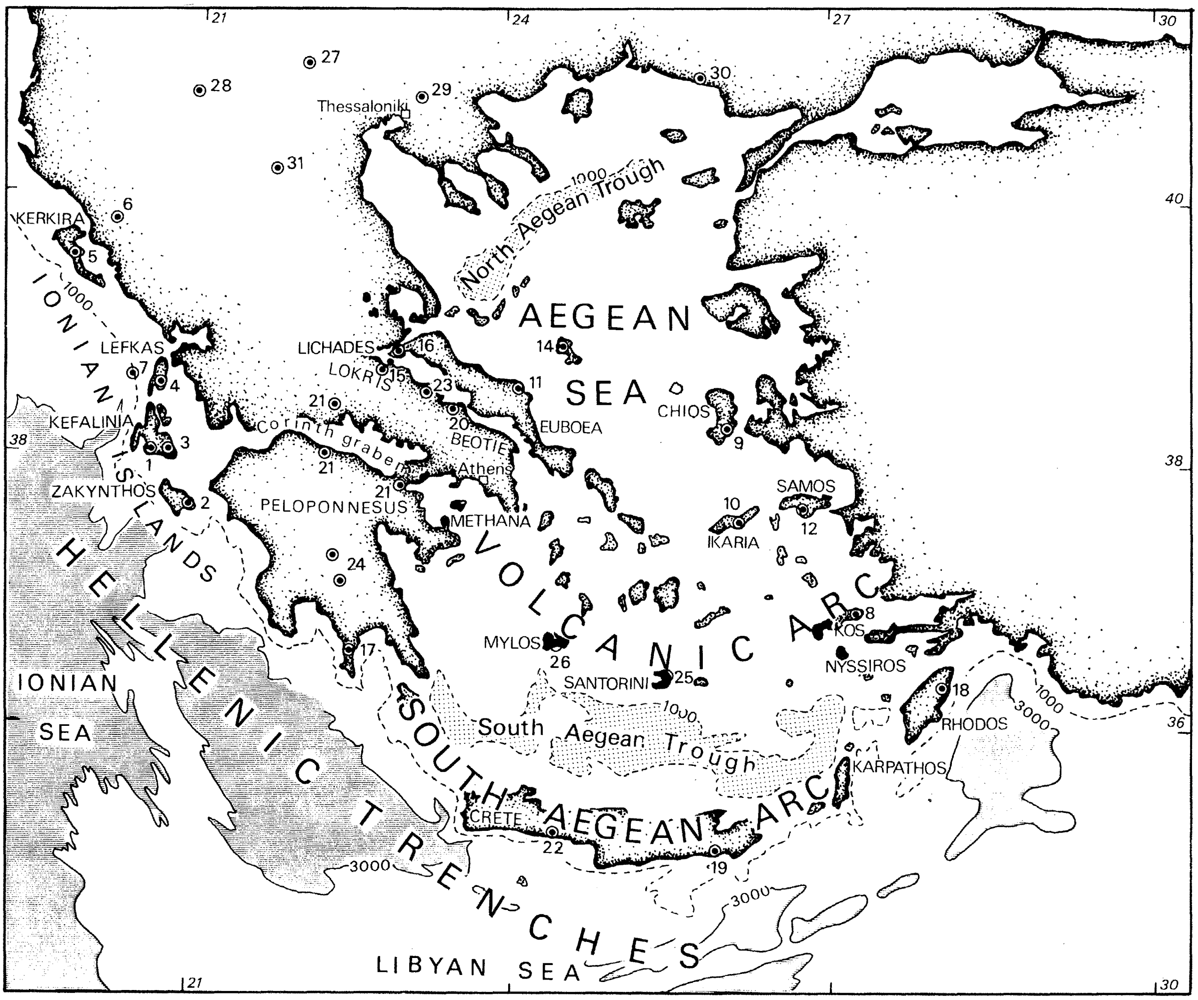

Figure 1. Map showing essential geographic and bathymetric features of the Aegean, Ionian and Libyan seas. Black patches represent U. Pliocene and Quaternary volcanoes constituting the Cycladic Volcanic Arc. The Aegean Arc edge outcrops on the Ionian Islands of Lefkas, Kefalinia and Zakynthos and in NW Peloponnesus (see figure 3); towards the S it seems to bound the Hellenic Trenches. Numbers refer to sites mentioned in the text.

1. *The Aegean Arc*

In the eastern Mediterranean, the Aegean Arc lies between an Adriatic continental collision in the NW and Arabic continental collision to the E.

On its seaward side, the Aegean Arc is flanked by a submarine trench. Landwards extends a back-arc basin, the Aegean Sea (Boccaletti *et al.* 1974), with its active volcanoes (figure 1) This pattern resembles, on a small scale, that of the W Pacific arcs. The seismic zone dips at an angle of 30° towards the centre of the Aegean Sea (Caputo *et al.* 1970; Papazachos & Comminakis 1971). Yet the crust of the back-arc basin is essentially continental. It is thinner than normal, typically 20 and 30 km in the south and north Aegean troughs (Makris 1977).

Seismicity within the eastern Mediterranean is diffuse, with the result that several small rigid plates located between the major Arabo-African and Eurasian plates have been proposed (McKenzie 1972). One of these is the Aegean plate. However, tectonic and neotectonic analyses have shown that the Aegean crust is cut by faults of Mesozoic and Cainozoic age and that this inherited fracture pattern has been rejuvenated during Pliocene–Pleistocene time. Continuum deformation models appear to explain the intraplate deformation of the Aegean better than rigid microplate models. Furthermore, the absence of clearly defined plate limits to the N and E indicate that the Aegean constitutes a deformed part of the Eurasian plate (Mercier 1977).

2. *Timing of the neotectonic deformations*

(*a*) *The pre-Aegean Arc stage*

At about 10 Ma, most of the present Aegean Sea area was land. This Aegean land had been strongly folded, metamorphosed and faulted during the Mesozoic and Cainozoic. An epicontinental sea transgressed periodically from the NE. From the S, the Mediterranean Sea also invaded some parts of the Aegean during the Tortonian and Messinian (see fig. 1A of Keraudren 1979).

The Mediterranean bordered this land approximately at the limit of the S Aegean and the W margin of Greece. The western marine domain of the Preapulian platform (Ionian Islands) had not yet been folded (Aubouin 1973). Until the L. Pliocene, it was affected by extensional tectonics expressed by normal, syn-sedimentary faults at Kefalinia (point 1 on figure 1) (B, fig. 3A). Between the unfolded Preapulian platform to the W and the folded Ionian zone in the E, there was a convergent boundary, the L.–mid-Miocene 'Ionian Arc' (figure 3*a*).

During the Lower Pliocene, the convergent limit jumped to the W of the Ionian Islands. The compressed edge of the plate was thus displaced towards the W. Neotectonic analysis in the Aegean has also shown an important intraplate deformation (Mercier 1977). This paper deals with the latter after the Miocene–Pliocene boundary (5 Ma) when a new 'Aegean Arc' was formed. Two distinct domains will now be analysed: the arc edge and the back-arc domain.

(*b*) *Neotectonics of the Aegean Arc Edge*

The arc edge domain outcrops on the Ionian Islands and in NW Peloponnesus. Further to the S, it constitutes the inner wall of the Hellenic trench.

(i) During the Lower Pliocene (5 Ma), the first compressional events affect both the external Ionian and the Preapulian zones (Sorel 1976).

The frontal part of the external Ionian zone is thrust onto the Preapulian platform (figure 3*a*) of Zakynthos, Kefalinia and Lefkas (points 2, 3 and 4 on figure 1). Compression probably also reactivates older thrusts in the Ionian zone, e.g. at Kerkira (point 5) and Filiates (point 6) (I.G.R.S.–I.F.P. 1966). The Preapulian zone is under compression. On Kefalinia (point 1), major reverse faulting occurs on the pre-existing discontinuities represented by NNE and SE orientated normal faults (B, fig. 3B). On Zakynthos (point 2) the main anticline is formed. The NNE and SE trending faults of the Ionian Islands affect the continental plateau and seismic reflexion near Lefkas (point 7) shows compressional features (Sorel *et al.* 1976). So at the NW edge of the Arc, the active compressional continental margin is formed as early as the Lower Pliocene (after *G. margaritae* zone).

(ii) As early as the L. Pliocene (*G. puncticulata* zone), following compression and emergence, the sea transgresses (B, fig. 3C). Marine sedimentation is sustained until Calabrian time. A certain instability is identified: on W Kefalinia (point 1), kilometre-sized olistoliths are deposited in the Pliocene sea. It is not clear whether this event reflects compressional or extensional activity.

(iii) During the L. Pleistocene (about 1 Ma), after the Calabrian and before the Milazzian, a new important compressional phase affects the arc edge. In the external Ionian zone, structures formed during the L. Pliocene are reactivated. On Zakynthos (point 2), marked shortening is expressed by complex folds and by reverse faults (Sorel 1976). In the Preapulian zone, e.g. on Kefalinia (point 1), reverse faults and reverse strike-slip faults are rejuvenated (A, fig. 2). Marked folding of the Plio-Calabrian formation also occurs (B, fig. 3C). The importance of this compressive phase in terms of structures (vertical throws of about 1 km) is comparable to that of the L. Pliocene.

(iv) Throughout the mid-Pleistocene–Recent, the arc edge remains under compression. Structures formed (B, fig. 3D) in Kefalinia (point 1) include reverse or reverse strike-slip faults of post-Milazzian, Riss and post-Neotyrrhenian ages (A, figs 3 and 4). In NW Peloponnesus, they include rejuvenated anticlinal structures (Sorel 1976). These structures, however, are relatively modest: visible displacement along reverse faults attains a maximum of 50 m. Today this compression along the arc edge is reflected by intense seismic activity (McKenzie 1972, 1978).

(c) *Neotectonics of the back-arc domain* (*south Aegean*)

This domain includes (figure 1) the S Aegean Arc, the S Aegean Sea (interarc basin), the volcanic Arc, the back arc basin s.s. and their prolongations in continental Greece. Boundaries between this domain and the north Aegean are not clearly defined.

(i) During the uppermost Miocene–L. Pliocene (about 7–5 Ma), two compressional phases occur in this region. On Kos (point 8) (Jarrige 1979), the first phase is expressed by folds (axial direction b_1 = N 00° to 40° E) and by reverse faults (B, fig. 4A). These affect sedimentary formations of uppermost Miocene–L. Pliocene age but do not affect the overlying 'Levantine' series of L. Pliocene age. The second phase is expressed by folds (b_2 = N 115° to 145° E), by reverse faults and by small thrusts (B, fig. 4A). These deform the 'Levantine' sediments but do not deform the Pliocene series containing *Paludina* fossils, immediately above. By contrast this last horizon is affected by normal, syn-sedimentary faults (B, fig. 4B). Other folds affecting U. Miocene–L. Pliocene series have been recorded within the Aegean but have not been accurately dated. Possibly they have the same age on Chios (point 9), on Ikaria (point 10) islands, and in Euboea at Kimy (point 11). On Samos (point 12), a first phase could be older (before 9 Ma) (Angelier 1979). These compressional structures are barely seen within the S Aegean Arc (Crete, Rhodos). They are moderate (hectometric folds and faults) N of the Cycladic Arc, e.g. on Chios. They are strong (kilometric folds and thrusts) within the Cycladic Arc, e.g. Kos (point 8), Samos (point 12) and on Skyathos (point 14) and Euboea (point 11).

During the uppermost Miocene–L. Pliocene the back-arc domain suffers compression. Thus, the formation of the new Aegean Arc may have begun during the uppermost Miocene within the S sector. This is earlier than in the sector of the Ionian Islands.

(ii) During the Pliocene, after these last compressional tectonics, the back-arc domain enters an extensional period. This is well demonstrated (Philip 1974) in Lokris (point 15) and NW Euboea (point 16). There, a family of normal faults with N 110–20° E strike (B, fig. 5A) are post-U. Pliocene and pre-date the 'Older Quaternary' compression (B, fig. 5B, C). During this period, the Corinth graben takes shape with an accumulation of deltaic Pliocene deposits, about 1000 m thick (Dufaure *et al.* 1979). The S Aegean Arc, briefly uplifted during the U. Miocene–L. Pliocene, is subsequently submerged. This submergence is expressed in the S Peloponnesus (point 17) (Dufaure *et al.* 1979) and at Rhodos (point 18) (Keraudren 1979). Syn-sedimentary extensional Pliocene deformation can be observed on Rhodos (B, fig. 7D) and on E Crete (point 19) (A, fig. 6). At that time, the back-arc domain collapses and an epicontinental sea transgresses over the whole Aegean region.

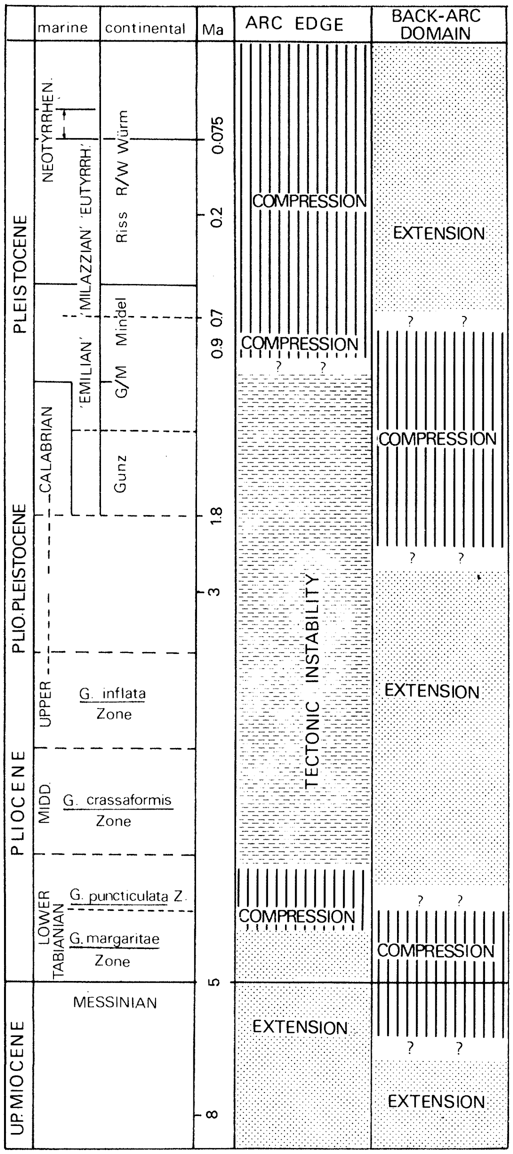

Figure 2. Table showing schematic chronology of the phases of deformation from U. Miocene to Pleistocene within the arc edge (Ionian Is) and in the back-arc domain (synthesis of data from the Lokris–Euboea (points 15, 16 and 23), the Gulf of Corinth (points 21) and the islands of Kos (point 8) and Rhodos (point 18)). A strong compressional deformation occurs in the arc edge during the L. Pliocene. From the L. Pliocene to the Calabrian a tectonic instability is evident; it has not yet been possible to determine whether this is compressional or extensional. A strong compressional deformation occurs again within the arc edge; this is of post-Calabrian–pre-Milazzian age. Within the back-arc domain, compressional phases alternate with longer periods of extension. These compressional phases seem to be formed by several compressive events within an extensional period.

(iii) During 'Older Quaternary', the back-arc domain suffered compression again. In Lokris (point 15) and NW Euboea (point 16) compression is expressed by reverse faults that cut into pre-existing normal, Pliocene fault-planes (B, fig. 5C; A, fig. 11), by reverse displacements along older normal faults (A, figure 16) and by small-scale thrusting (B, fig. 5A, B, D). In NW Euboea, these thrusts supply frontal breccias (B, fig. 5B, E) which are interbedded within lacustrine formations containing a fauna of transitional Pliocene–Pleistocene age. In Lokris (B, point 4 on fig. 5A), they are associated with syn-sedimentary folds affecting these same

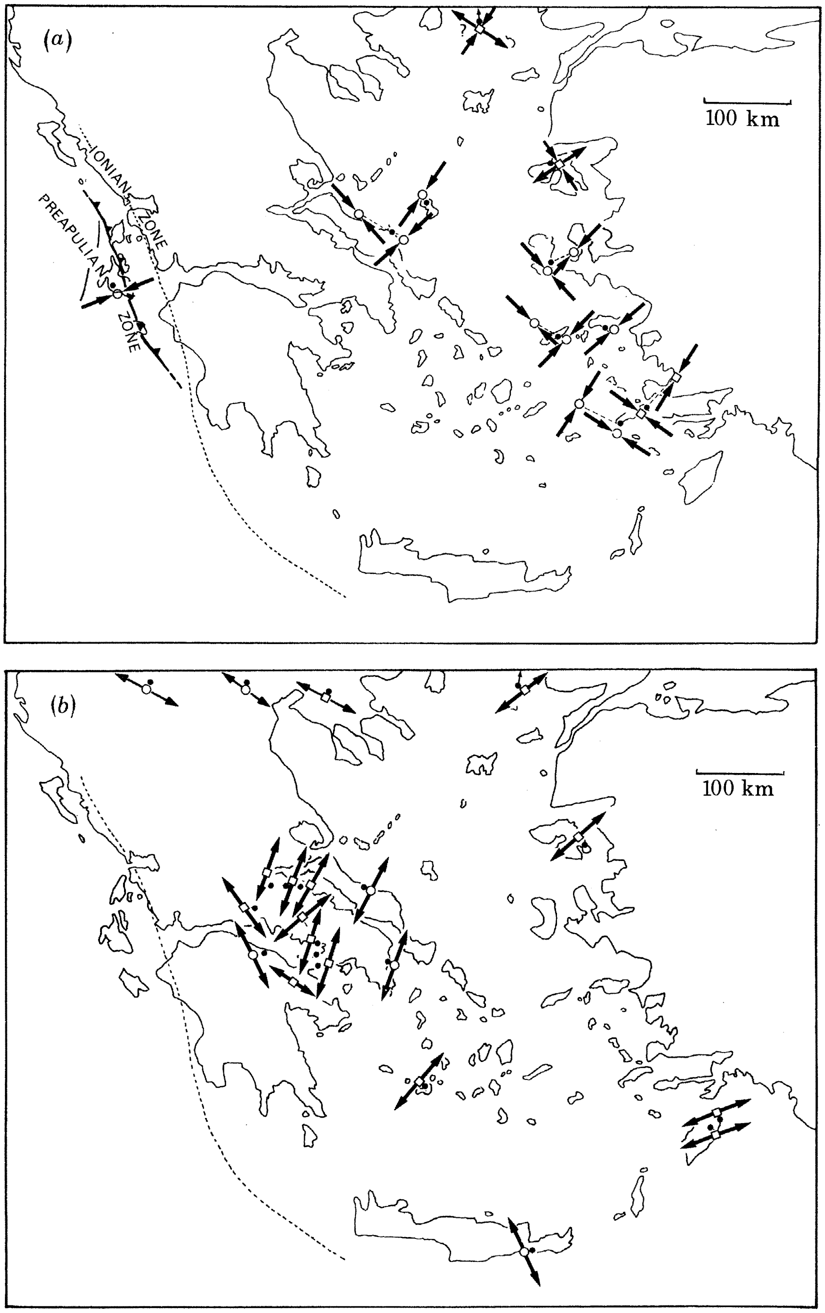

FIGURE 3*a* and *b*. For description see opposite.

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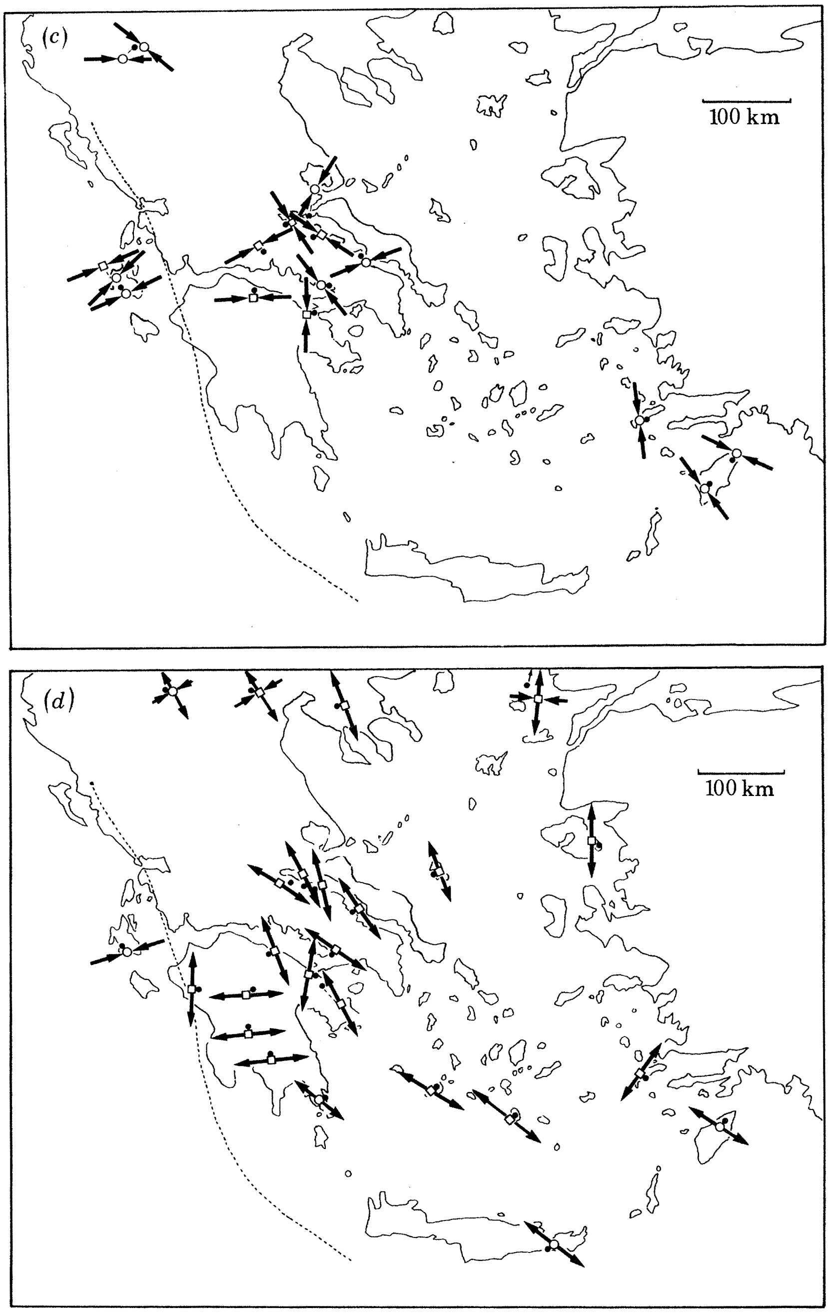

FIGURE 3. Directions of compression (convergent black arrows) and of tension (divergent black arrows) for the four principal tectonic phases of Upper Miocene to Recent age within the Aegean domain. At open squares, arrows represent the σ_3 and σ_1 principal stress directions calculated from a family of faults by the numerical method of Carey (1979). At open circles, arrows represent the directions of shortening and of lengthening determined by graphical methods. Directions of shortening obtained from folds are presumed to be orthogonal to the axial direction of folds. In the north Aegean (*a* and *d*), convergent and divergent arrows at the same site represent strike-slip faulting.

(*a*) Directions of compression during the uppermost Miocene–Lower Pliocene. Within the arc edge, the compression is of L. Pliocene age. In the back-arc domain at a given site, two successive compressional phases occur during the U. Miocene–L. Pliocene. The two directions of compression are approximately orthogonal. However, all structures that have resulted from the same compressional direction do not seem to have the same age.

(*b*) Directions of tension for the Lower to Upper Pliocene period of extension.

(*c*) Directions of compression for the 'Older Quaternary' to mid-Pleistocene period of compression.

(*d*) Directions of tension for the period mid-Pleistocene–Recent.

lacustrine sediments. This compressional phase appears to be very widespread around the Aegean. It is also known in Beotia (point 20) (B, fig. 6A), around the Gulf of Corinth (points 21), on Kos (point 8) (Jarrige 1978), on Rhodos (point 18) (B, figure 7C) and in W Crete (point 22) (Angelier 1977). Thus compressional structures seem to have begun earlier ('Older Quaternary') within the back-arc domain than along the arc edge where they formed after the Calabrian.

(iv) During the mid-Pleistocene–Recent, deformation in the back-arc domain is once more extensional. Once again collapse leads to transgression of an epicontinental sea across the Aegean. On NW Euboea (point 16) and Lokris (point 15), this extension is renewed with certainty after the base of the mid-Pleistocene. It is expressed by normal and normal-strike slip faulting along fault planes formed during the Pliocene or earlier, and by the development of new faults striking N 70° E. These faults are (see fig. 2) of post-transition Plio-Pleistocene, ante-Mindel–Riss (B, fig. 5B), post-Mindel, post-Mindel–Riss (B, fig. 5B), post-Riss, intra-Würm (A, fig. 10) ages. This extension has persisted until today, yielding seismic activity. An example is the reactivation of the Martienon–Atalanti fault (point 23) during the 1894 earthquake (B, fig. 6B). This extension is illustrated in numerous locations within the Aegean: Kos (point 8) (B, fig. 4D), Gulf of Corinth (point 21) (A, fig. 8; Dufaure *et al.* 1979), central Peloponnesus (point 24) (Sébrier 1977), E Crete (point 19) (Angelier & Gigout 1974; A, fig. 7). Karpathos (Angelier 1977), Rhodos (B, fig. 7B), Santorini (point 25) and Mylos (point 26) (Jarrige 1978).

(*d*) *Neotectonics of the north Aegean domain*

The neotectonic deformation of the north Aegean is similar to that of the back-arc domain, yet compressional structures are very rare.

During the Pliocene, central Macedonia is under extension. The Almopias graben (point 27) takes shape and a high-K volcanism persists until the U. Pliocene (4–2.5 Ma) (Bellon *et al.* 1979). Normal faults striking N 30° E near the Greek, Albanian and Yugoslav border (point 28) are probably of the same age. Similar normal faults affect older Pleistocene formations north of Chalkidiki (point 29). During the U. Pleistocene–Recent, extensional deformation is widespread in Macedonia. Near Thessaloniki (point 29), it is expressed by neotectonic and active faulting (Mercier *et al.* 1979*b*). From there towards Turkey (point 30) to the E and towards Albania (point 28) to the W, strike-slip faulting becomes more and more important.

The first compressional phase is probably present on Chios (point 9) where two families of folds affect U. Miocene–L. Pliocene formations as in the back-arc domain. Near the Turkish–Greek border, compressive strike-slip faults cut the U. Eocene, Oligocene and (?) Miocene series. They may be a result of this compressional phase. 'Old Quaternary' compressional deformations are only known in W Macedonia (point 31), where reverse faults cut conglomerates of L. Pleistocene age (Faugères & Vergely 1974).

(*e*) *Summary of the neotectonic evolution of the Aegean*

The four principal stages are shown in figure 2.

(i) During L. Pliocene time (about 5 Ma), the overall Aegean domain suffers compression (figure 3*a*). Important thrusts develop on the Preapulian edge, which becomes the active convergent margin of the new Aegean Arc. Within the back-arc domain, compressional structures seem to have formed earlier, in U. Miocene. Thus in the S sector of the arc the convergent margin may have been active since the uppermost Miocene (about 6–7 Ma). No data have been

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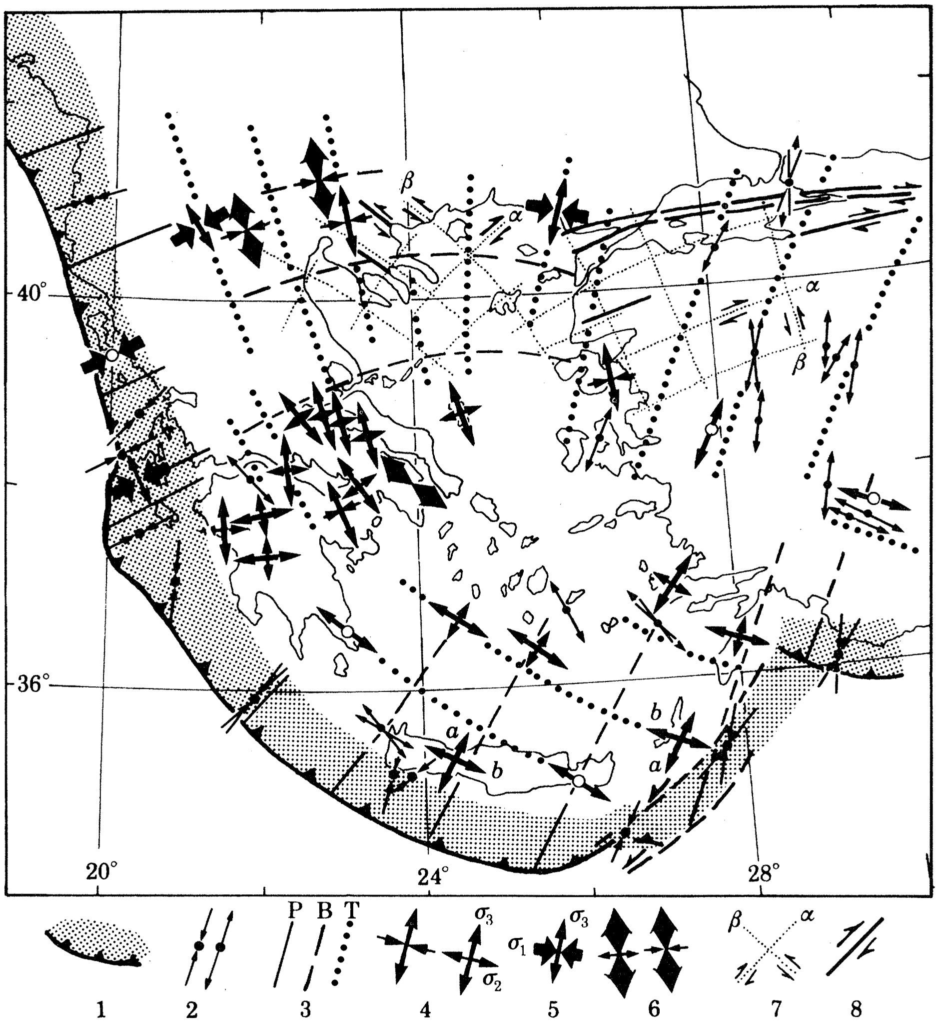

FIGURE 4. Stress trajectories in the Aegean domain from (*a*) focal mechanisms of shallow earthquakes; (*b*) structural analysis of neotectonic and active faults; (*c*) *in situ* stress measurements.

(1) Aegean subduction: the dotted area approximately represents the compressed convergent margin of the Aegean arc and of the Adriatic collision; (2) Horizontal projections of the slip-vectors determined from focal mechanisms of compressional (convergent arrows) and extensional (divergent arrows) shallow earthquakes (McKenzie 1972, 1978) (3) (P, B, T). Regional principal directions of compressional, intermediate and tensional stresses determined from clusters of focal mechanisms (from A. R. Ritsema (1974), modified along the convergent margin, in SE Aegean and W Turkey with the help of McKenzie's data (1972, 1978). (4, 5). Deviatoric compressional (σ_1), intermediate (σ_2) and extensional (σ_3) principal stress directions determined from structural analysis of neotectonic and active faults. The two principal directions shown are those approximately situated in the horizontal plane. The intermediate deviatoric principal stresses are compressive (small convergent arrows) or extensional (small divergent arrows). In W Crete and on Karpathos (from J. Angelier), the two arrows marked *a* and *b* represent two σ_3 directions belonging to distinct Quaternary extensional phases. Arrows diverging from open circles represent extensional directions determined from graphical methods. (6) Principal extensional stress directions ($\delta_{\mathrm{h,\,min}}$, negative) from *in situ* stress measurements (Froidevaux *et al.* 1979); the other principal stress directions, having the smallest absolute values ($\sigma_{\mathrm{h,\,max}}$), are compressive (small convergent arrows) or extensional (small divergent arrows). Measurements are made at a small depth (less than 4 m). (7) In the north Aegean, slip-lines have been drawn bisecting at a 45° angle the P and T trajectories. (8) Strike-slip faults (see Mattauer & Mercier 1980).

Data from: Ionian Islands, Sorel (1976) and Mercier *et al.* (1979*a*); Peloponnesus and Corinth Gulf, Sébrier (1977); Lokris, Euboea and Beotie, Pegoraro (1972), Philip (1974), Mercier *et al.* (1979*a*) and Lemeille (1977); E Crete, Mercier *et al.* (1974); W Crete and Karpathos, Angelier (1977, 1979); Islands of Kos, Santorini and Milos, Jarrige *et al.* (1976) and Jarrige (1978); Thessaloniki, Mercier *et al.* (1979*b*); Thrace & Lesbos, Mercier, unpublished; W Turkey, Dumont *et al.* (1979). See references in Mercier *et al.* (1979*a*). Stress tensors have been computed by Mlle E. Carey.

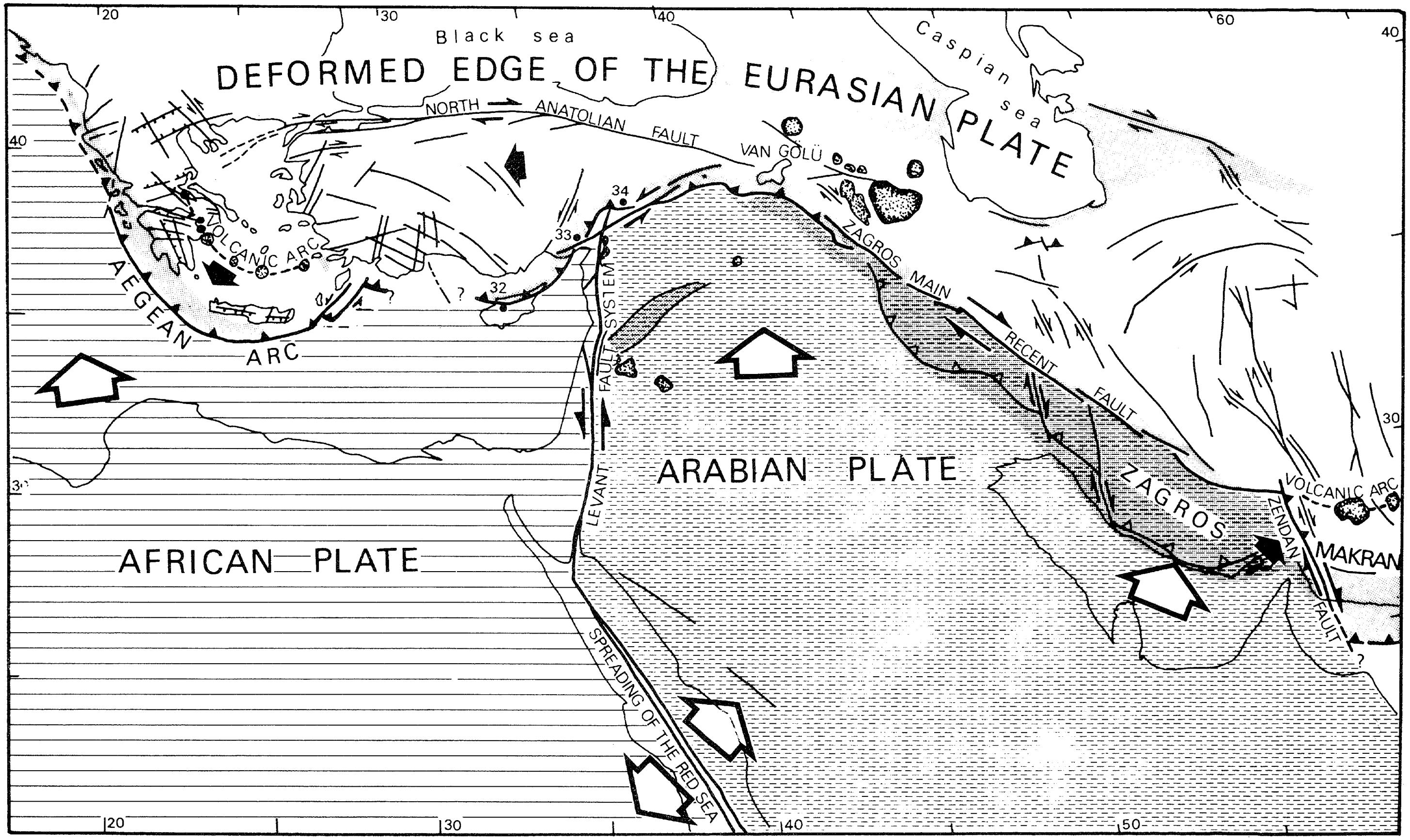

FIGURE 5. Location of the Aegean Arc with respect to the major deformational units of the Middle East and the E Mediterranean. Zones of uppermost Miocene–Quaternary compressional deformation are represented in grey. Dotted patches represent Quaternary volcanism. The Aegean, Turkish and Iranian domains are considered as belonging to the edge of the Eurasian plate. The Makran Arc, situated to the E of the Arabo-Eurasian collision, exhibits characteristics comparable to those of the Aegean Arc. Large white arrows show approximate movements of the Arabian and African plates with respect to the Eurasian plate. Small black arrows indicate the 'flow direction' of Turkish and Aegean material with respect to the Eurasian plate (after McKenzie 1972). Even where strike-slip movements are the major mechanism of continental deformation, deformation is not purely planar so that dip-slip faults also occur. Numbers refer to sites quoted in the text.

put forward for an earlier formation of the new Aegean convergent margin. However, from early extensional tectonics in the Aegean, it has been argued that this goes back to 13 Ma (Le Pichon & Angelier 1979). In fact, the Miocene calc-alkaline volcanism and plutonism lasts from 25 Ma to 8 Ma. Between 8 and 4 Ma, alkaline volcanism arises. Around 2.5 Ma, calc-alkaline volcanism starts again and forms the present volcanic arc (Bellon *et al.* 1979).

(ii) During mid–Upper Pliocene time, the back-arc domain suffers extension (figure 3*b*). Along the Preapulian edge, tectonic activity is weak. It has not been possible to establish whether it was compressional or extensional.

(iii) During 'Old Quaternary' time, the Aegean domain is once more under compression (figure 3*c*). After the Calabrian, important compressional structures form again along the Preapulian edge. As far as stratigraphic correlations are trustworthy, it appears that compression acts earlier in the back-arc domain (figure 2). Moreover, this compressional phase seems to consist of several compressive events within an extensional period.

(iv) From mid-Pleistocene to Recent, the back-arc domain was again extensional. Compression persists in a narrow zone along the Preapulian edge but the structures are moderate (figure 3*d*).

Thus, in the convergence process between the Arabo-African and Eurasian plates, it appears that under 'normal' conditions the Aegean includes a convergent margin under compression that is narrow and weakly deformed plus an extensional back-arc domain. Short compressional phases perturb this normal régime. During these phases the convergent margin is strongly compressed and deformed and compression invades the back-arc domain.

Further east, one notices an important convergent boundary in the northern part of Cyprus (point 32, figure 5). This same boundary is found in E Turkey in the Mysis (point 33) and Bitlis (point 34) mountains (Biju-Duval *et al.* 1976). In Cyprus, compression takes place between the U. Messinian and L. Pliocene (Barroz 1979).

Around the W Mediterranean, compressional activity of U. Messinian and/or L. Pliocene age is reported for Morocco (Frizon de la Motte 1979), Spain (Armijo 1977) and Tunisia (Rouvier 1977). The second compressional phase of L. Pleistocene age is also present in S Spain (Armijo *et al.* 1977), Morocco (Rampnoux *et al.* 1977), Algeria (Thomas 1977), Tunisia (Jauzein 1967) and Sicily (J.-C. Bousquet, personal communication).

Thus, compression in the Aegean seems to be part of a wider tectonic phenomenon occurring at the contact between Arabo-Africa and Eurasia. Changes in convergence velocity could account for this widespread behaviour.

Along the Preapulian edge the duration of the compressional phases are estimated to be no more than some hundreds of thousands of years (Sorel 1976).

3. *Stress patterns and spreading in the Aegean*

(*a*) *Stress analysis*

These have been determined for each of the four principal neotectonic phases (figure 3*a*–*d*). The present discussion is restricted to deformation of U. Pleistocene–Recent age.

The regional stress pattern is drawn (figure 4) from data provided by focal mechanisms of shallow earthquakes (McKenzie 1972, 1978; Ritsema 1974), by *in situ* stress measurements (Froidevaux *et al.* 1979) and by structural analysis of recent and active faults (see A, B and Carey 1979). In a highly fractured medium, deformation takes place by small displacements of rigid blocks along pre-existing fault-planes. One can assume that the slip marked by striations $\boldsymbol{S}$ is parallel to the tangential stress $\boldsymbol{\tau}$ applied to the fault. The orientation of $\boldsymbol{\tau}$ is a function of the orientation of the principal stress directions and of the ratio $R = \sigma_1 - \sigma_2/\sigma_3 - \sigma_2$, where the σ's are deviatoric stresses. Conversely, the measured striations on a family of faults of the same generation can be used to compute the above parameters of the stress deviator.

This simple model implies an isotropic material, so that the principal strain and stress directions are parallel. Coherence tests must be carried out. First, the discrepancy in orientation between the computed $\boldsymbol{\tau}$ and the observed $\boldsymbol{S}$ must be small (less than 15°) for each fault. Secondly, the differences between computed principal directions for each site in a given region must be small (less than 25°). These tests are necessary to ensure the validity of the model. They also help one to appreciate the regional significance of the computed solutions.

It should be emphasized that data obtained by focal mechanisms, *in situ* stress measurements and neotectonic analysis are similar (figure 4). Thus, the inferred regional crustal stress pattern must provide a rather good approximation to reality.

(*b*) *Aegean spreading*

The stress pattern (figure 4) is markedly different in the north and south Aegean.

(i) In the north Aegean the principal extensional direction σ_3 changes from NNE in W Anatolia to NNW in W Macedonia, near Albania. Slip-lines have been drawn bisecting the angle formed by the horizontal principal stress directions. These slip-lines are parallel to well known neotectonic and active strike slip-faults in W Anatolia and Macedonia. Slip-line theory states that compressional deviatoric stresses must decrease in value along slip-lines running from W Anatolia (or Albania) to the central Aegean. Similarly, extensional values must increase. This is based on a model of plane deformation in a rigid-plastic medium. Neotectonic analysis and *in situ* stress measurements both agree with the above theory (Mercier *et al.* 1981). Indeed, principal stresses striking E–W (figure 4) are compressional (σ_1) in Thrace (and probably near Albania), compressional intermediate (σ_2) in central Macedonia and extensional intermediate (σ_2) in the central Aegean. Spreading in the north Aegean seems to be associated with southwards flow (extrusion) of the European lithosphere (Tapponnier 1977; Mercier 1977). The real deformation is, however, not purely planar (Mercier *et al.* 1979*b*).

(ii) In the south Aegean, the principal extensional direction σ_3 is approximately parallel to the convergent boundary of the arc. Stress values change along trajectories normal to principal extensional directions. Three principal stress directions fall into line in the NNE direction: σ_1 at the arc edge (Crete), extensional σ_2 in the interarc basin and finally σ_3 in W Turkey.

Thus, it seems that the southward flow of the Aegean lithosphere leads to a lengthening nearly parallel to the arc edge in the south Aegean.

The above pattern prevails during the U. Pleistocene–Recent but is not in a steady state. For instance, along the south Aegean Arc (central Peloponnesus, Crete, Karpathos), σ_3 directions change and become nearly perpendicular to the arc boundary. This may be associated with occasional buckling of the crust near the convergent boundary. On Kos, on the other hand, the present σ_3 is WNW–ESE (McKenzie 1978) but it was NNE–SSW during the U. Pleistocene (Jarrige 1978) as it is now in W Turkey, 100 km N of Kos.

Possible mechanisms will be discussed later. For now, note that in the extended south Aegean crust, stress trajectories seem to be controlled by the African–Aegean convergence. A similar pattern will now be described in the Andes.

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II. Extensional–compressional tectonics associated with the Andean Cordillera of south Peru and north Bolivia

The chronology of Miocene and Pliocene deformations of south Peru and north Bolivia is rather well known (see Wilson & Garcia 1962; Noble *et al.* 1974; Audebaud *et al.* 1973; Aubouin *et al.* 1973; Mégard 1978; Dalmayrac *et al.* 1977; Martinez 1978). However, neotectonic studies describing the Neogene strain patterns in this region are only now in progress (Soulas 1978; Lavenu 1978; Sébrier *et al.* 1981; Lavenu & Mercier 1981). A short summary follows.

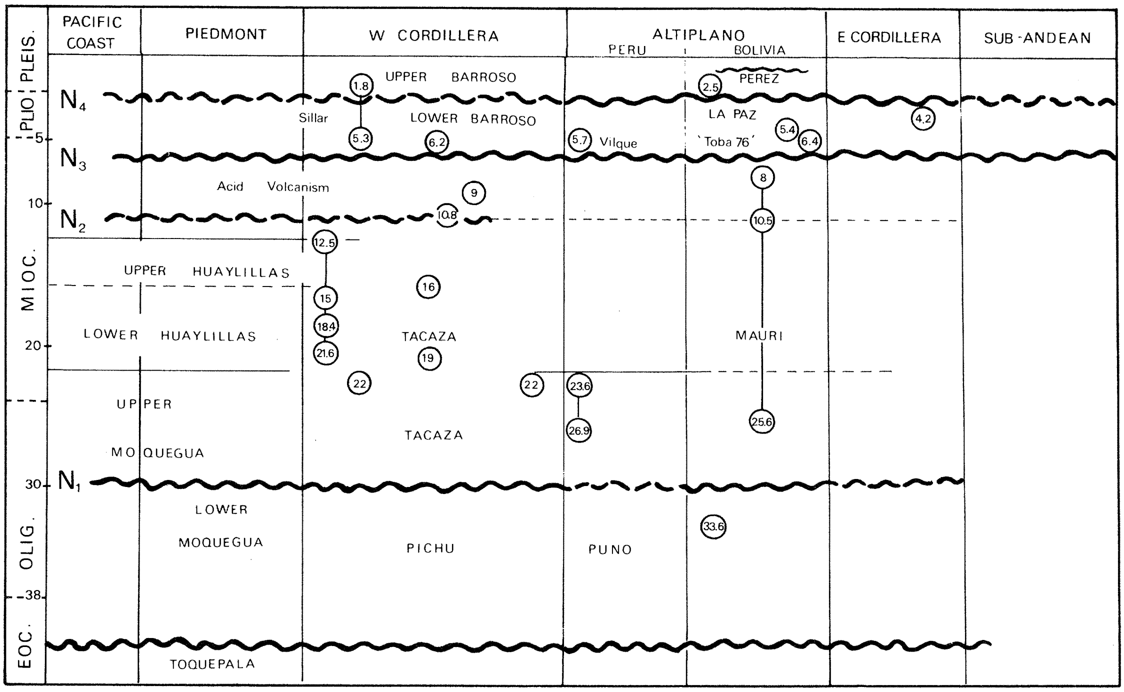

Figure 6. Timing of the Neogene compressive deformations of the Andean Cordillera of south Peru–north Bolivia (after Sébrier *et al.* 1981). From the Pacific Coast to the eastern Cordillera (see figure 8), the Andes are affected by four major compressive phases referred as N_1, N_2, N_3 and N_4 (N_1 seems to be of U. Oligocene age rather than L. Miocene as originally suggested). Local names of the principal volcanic and sedimentary formations of the Andean zones are given here. Numbers refer to radiometric ages (megayears) of volcanic formations. They have been used to establish the chronology of the compressive deformations.

1. *The Andean Cordillera of Peru and Bolivia*

Along the south Peru–north Chile segment (lat. 15–27° S), the Nazca plate plunges at an angle of 25–30° under South America. There a wedge of asthenospheric material appears to separate the sinking slab from the continental lithosphere (Barazangi & Isacks 1976). Quaternary and active volcanoes and a subsiding high plateau (the Altiplano) are present on the overriding plate. In spite of the strong uplift of the Andes, this structure shows some similarities with the Aegean Arc: an inclined seismic zone with similar dip, active calc-alkaline volcanism and subsidence behind the arc.

By contrast, north of the Nazca Ridge, along the central and north Peru segment (lat. 2–15° S), the slab has a dip of about 10° and asthenospheric material is absent between the oceanic slab and the continental lithosphere (Barazangi & Isacks 1976). The Altiplano and active volcanoes do not extend into this segment (figure 8).

2. *Timing of the Neogene deformations of the Andean Cordillera of south Peru – north Bolivia*

(*a*) *Compressional tectonics*

Figure 6 summarizes the chronology and the extent of the Neogene compressive phases (Sébrier *et al.* 1981). It shows two important points. First, from the Pacific coast to the eastern Cordillera, the Andes are affected by several major compressive phases. Only the youngest phases are known in the easternmost region (sub-Andean zone). Secondly, these compressional events are short-lived, probably less than a few megayears.

Both the short duration and the broad extent of these compressive events are incompatible with the idea (James 1971) that their basic mechanism is the dilatation of the W Cordillera due to magmatic intrusions.

(*b*) *Extensional–compressional tectonics within the Altiplano*

Thick accumulations of clastic material in the Altiplano and of volcanic material in the W Cordillera suggest that extensional tectonics have taken place between the compressive phases. However, normal faults of Miocene and Pliocene age are exceptional in the Andes. Most of them must have been reactivated during compressive phases so that reverse and strike-slip faulting prevails in the geological record. Yet compressional–extensional tectonics have been demonstrated on the Bolivian Altiplano (Lavenu 1978; Lavenu & Mercier 1981). This is shown in figure 7.

(i) During the U. Miocene, compression affects the Altiplano. In the W Altiplano, folds are recorded in a clastic formation containing interbedded welded tuffs 10.5 and 8 Ma old. These folds have been eroded and then covered by a Pliocene formation containing a welded tuff dated 6.4 Ma. They have an axial direction near N 115° E and seem to form an en-echelon pattern along the NNW–SSE San Andres fault zone. Thus the regional shortening direction is roughly N–S.

(ii) During the Pliocene, extension takes place. On W Altiplano (W of Curahuara), a graben of a few kilometres width cuts the Pliocene formation. This graben was formed before the outpouring of the 2.5 Ma old welded tuff. Slip on fault planes shows that the lengthening direction is approximately E–W.

(iii) During the Upper Pliocene, strike-slip faulting prevails. On the E Altiplano (near La Paz), strike-slip faults and folds affect the Pliocene formation. They are older than the first Pleistocene glacial deposits. Structural analysis shows that they result from a nearly N–S extension σ_3 and a nearly E–W compression σ_1, both horizontal. On the W Altiplano (E of Curahuara), the Pliocene formation is folded and unconformably covered by 2.5 Ma old welded tuff. Folds have an axial direction N 135° to 160° E and appear to form an en-echelon pattern along the San Andres fault.

(iv) During the L. Pleistocene, the Altiplano is once more compressed. In the W (Curahuara, Charaña), the 2.5 Ma old welded tuff is cut by numerous flat reverse faults. These are older than the lacustrine formation related to early Pleistocene climate changes. Compressional σ_1 direction is near N 25° E.

(v) During the Pleistocene and Recent, extension occurs once again. Normal faults are frequent in the Quaternary and Recent deposits. They are linked with a σ_3 having a N 10° to 20° E direction. Strike-slip faulting may have taken place during the Pleistocene in this region.

(*c*) *Compression throughout the Andean Cordillera*

It has been suggested that gravitational forces acting on a plunging slab may be overcome by hydrodynamic forces so that the slab dip becomes flat (Jischke 1975). Friction between the oceanic and continental plates gets stronger and compressive stresses arise in the overriding plate (Barazangi & Isacks 1976). Indeed, above the flat slab of central and north Peru, compression is expressed by seismicity (Stauder 1975) and by Quaternary and active strike-slip faulting (Soulas 1978; Mégard & Philip 1976), an exception being the White Cordillera. The above mechanism can explain compressional deformations in the Andean crust.

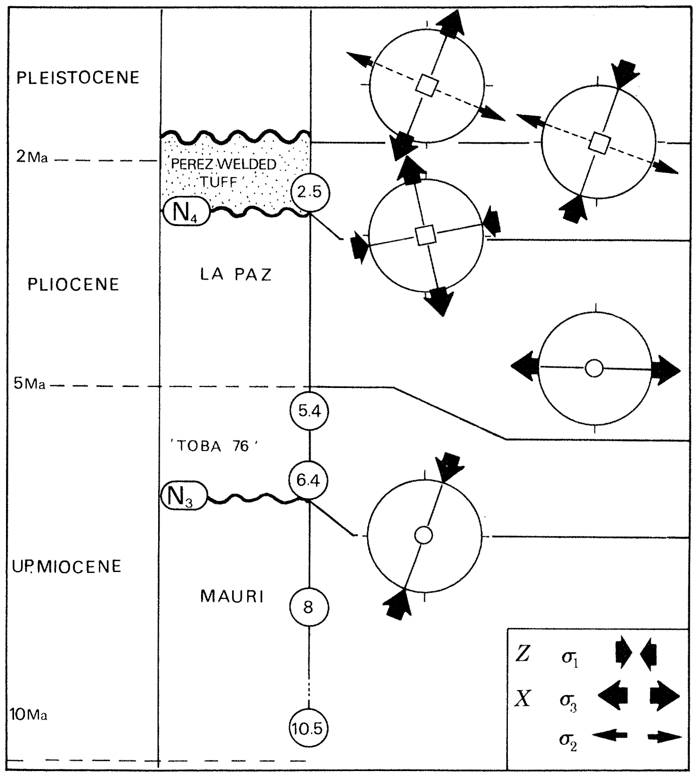

FIGURE 7. Extensional–compressional tectonics within the Bolivian Altiplano during the U. Miocene–Pleistocene (after Lavenu 1978; Lavenu & Mercier 1981). During the U. Miocene, the regional shortening (N_3) is roughly N–S; during the Pliocene, extension is E–W; during the U. Pliocene, strike-slip faulting (N_4) prevails (nearly E–W compression and N–S extension); during the L. Pleistocene, compression is nearly N–S; during the Pleistocene and Recent, extension is roughly N–S. Open circles in the centre of the nets indicate that shortening (Z) or lengthening (X) have been determined by graphical methods; open squares indicate that the principal deviatoric stress directions ($\sigma_1, \sigma_2, \sigma_3$) have been determined by numerical methods. Radiometric ages, local names of volcanic formations and tectonic phases are shown in figure 6 (Altiplano, Bolivia).

However, major compressive phases have the same age (U. Cretaceous, U. Eocene, U. Miocene, ? L. Pleistocene) for both segments of the Andes considered here. My feeling is that, as for the Mediterranean, this must result from modifications of the conditions of convergence.

3. *Stress pattern and extension in south Peru and north Bolivia*

(*a*) *Pleistocene–Recent stress pattern* (figure 8)

Off the Pacific coast, focal mechanisms show compression in E–W to NE–SW directions (Stauder 1975). On land, on the coast and piedmonts, Quaternary tectonics are predominantly extensional. Normal faults have a predominant E–W strike; thus the direction of lengthening is approximately N–S. In the W Cordillera (near Huambo), normal faulting is associated with

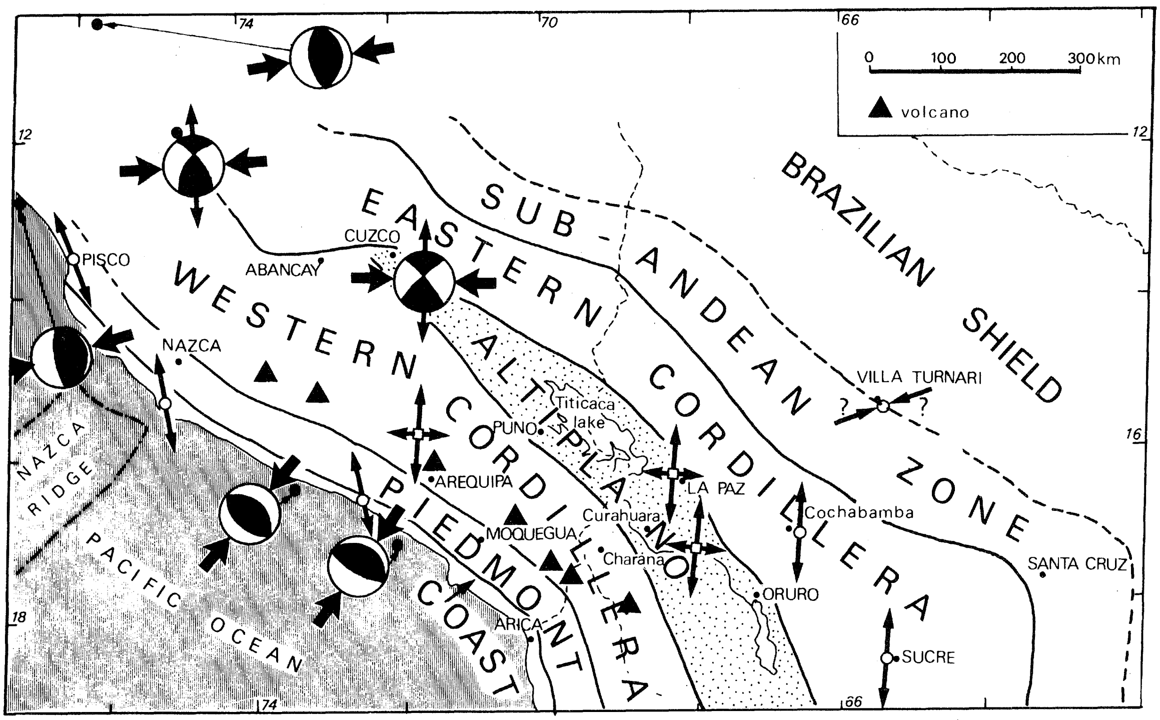

FIGURE 8. Pleistocene–Recent stress pattern in south Peru and north Bolivia from (*a*) focal mechanisms of shallow earthquakes (Stauder 1975) and (*b*) structural analysis of Recent faults; Pacific coast, Soulas (1978); W Cordillera, M. Sébrier (personal communication); Altiplano and E Cordillera, Lavenu (1978) and Lavenu & Mercier (1980). Off the Pacific coast, focal mechanisms show E–W to NE–SW compression. On land, in south Peru–north Bolivia, normal faulting prevails with a nearly N–S extension. In central Peru, strike-slip faulting seems to be predominant with an E–W compressional direction. Active deformation within the sub-Andean zone appears to be compressive.

Recent volcanic flows. The extensional direction σ_3 has N 10° E strike (M. Sébrier, personal communication). On the Altiplano we have seen that Recent tectonics are extensional in a N 10° to 20° E direction. In the E Cordillera, the Cochabamba basin contains Pliocene–Pleistocene deposits. These are affected by several phases of deformation apparently similar to those of the Altiplano. The scarcity of stratigraphic data makes it difficult to prove that normal faults yielding N–S extension are of Quaternary age. In the sub-Andean zone, vegetation does not permit good tectonic observations. Along its E border (near Villa Turnari), Quaternary deformation seems to be compressional.

In central and north Peru, i.e. in the northern segment, seismicity in the sub-Andean zone clearly indicates an E–W compression direction (Stauder 1975).

(*b*) *Extension*

In south Peru–north Bolivia, as in the south Aegean, a narrow compressed edge borders the trench. Further inland, the crust is in extension in a direction roughly normal to the velocity of convergence of the plates. This extensional domain is probably bounded by the E–W compressed crust of the sub-Andean zone and Brazilian shield.

Now a question arises: is the superficial extensional tectonics representative of the deviatoric stress field throughout the crust? Or is it a superficial feature caused by topography? Only one focal mechanism is known beneath the Altiplano (Stauder 1975). It is located in its NW part, between the north Chile–south Peru and the central–north Peru segments. This mechanism is strike-slip and is in agreement with a N–S extension and an E–W compression. This earthquake could be situated within the transition zone between a N–S extended crust in south Peru and a E–W compressed crust in central Peru.

If the crust in south Peru is in extension, one has to look for a mechanism. The N–S direction of extension means that gravitational sliding towards the trench is not predominant. On the other hand, a wedge of asthenospheric material associated with volcanism, back-arc subsidence and extensional tectonics is a common feature of south Peru and the Aegean. The extensional mechanism must be related to the presence of asthenospheric material between the dipping slab and the continental plate.

III. Concluding remarks

(*a*) Extensional tectonics associated with convergent boundaries have been explained in different ways: sublithospheric convection (see McKenzie 1978), gravitational sliding towards the trench (Makris 1977; Le Pichon & Angelier 1979) and/or secondary features related to large scale intraplate deformation (Tapponnier & Molnar 1976).

Above a subduction zone a correlation has been found in the Andes and the Aegean between the presence of an asthenospheric wedge and the occurrence of extensional tectonics. Gravitational sliding towards the trench certainly exists but is not the major phenomenon. In central Peru, for instance, where the asthenospheric wedge is absent and topography is important, the prevailing deformation is by reverse and strike-slip faulting. On the other hand, in south Peru and the Aegean, extension does occur but in a direction roughly at right angles to the direction of convergence.

In intracontinental conditions like the north Aegean, extensional neotectonics is compatible with a continuous crustal deformation described by means of slip-line theory. Yet the real strain is not planar but three dimensional.

(*b*) Temporal variations of the compressional–extensional neotectonics, seem to be too rapid to be explained by mechanisms characterized by a large time constant. These variations seem to be controlled by stress conditions at the boundaries, particularly at the convergent boundary.

In an 'unlocked' situation, the compressive deformation is spatially restricted to a narrow zone next to the trench. Elsewhere, extensional deformation is widespread. This is the normal régime for situations with a steeply plunging slab.

In a 'locked' situation, compressive stresses are transmitted throughout the crust of the overriding lithosphere. Locking may be caused by a stronger friction between the plates related to a flatter slab dip. Major compressive phases affecting the overriding plate could be caused by changes in the direction or rate of convergence.

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