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Relations between faulting and continuous deformation on the continents

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Abstract
The most reasonable way to describe the long wavelength deformation of the continental lithosphere is by a continuous velocity field, and yet the upper continental crust deforms discontinuously, by faulting. A kinematic description of the deformation in a region should thus include a knowledge of the long wavelength velocity field, and also an understanding of how faulting accommodates it in the crust. These concepts are illustrated with examples describing the active deformation of the Aegean Sea.

1. Introduction

Earthquakes are the most obvious manifestation of active deformation on the continents. Two features dominate the distribution of continental earthquakes: their epicentres are dispersed over broad regions that are usually several times the thickness of the lithosphere, and their hypocentres are usually restricted to the upper (10÷20) km of the continental crust, which is a small part of the total lithosphere thickness. The scale on which the large topographic features, such as mountain belts, plateaus and basins, occur suggests that at large length scales the deformation of the continents is dominated by the behaviour of the lithosphere as a whole, rather than by that of the thin seismogenic layer on top (*e.g.*, McKenzie, 1978; England and Houseman, 1986; Vilotte *et al.*, 1982; England and Jackson, 1989). It is probable that the lower (80÷100) km of the continental lithosphere deforms in a more distributed fashion than the discontinuous faulting that is important in the upper seismogenic layer. Thus at large length scales, by which we mean at dimensions comparable with or larger than the lithosphere thickness, it is reasonable to approximate the deformation of the continents to that of a continuum, which can then be described by a continuous velocity field (*e.g.*, England and McKenzie, 1982; McKenzie and Jackson, 1983). An important kinematic problem is then to obtain this velocity field, and to understand its relation to the motions of the rigid plates that bound the deforming region. Thus, for example, attempts have been made to understand how the velocity field in the Mediterranean region is related to the convergence between Africa-Arabia and Eurasia (*e.g.*, Jackson and McKenzie, 1984, 1988; Taymaz *et al.*, 1991a; Jackson *et al.*, 1992), and how that in central and eastern Asia relates to the India-Eurasia convergence (*e.g.*, Molnar and Deng, 1984; England and Houseman, 1986; Holt *et al.*, 1991a). However, continuous velocity fields that describe the long wavelength deformation of the lithosphere do not describe the detailed deformation of the seismogenic layer itself, which deforms discontinuously on faults. An additional problem is thus to understand how discontinuous faulting in the upper continental crust accomodates the velocity field that describes the deformation of the lithosphere as a whole.

The two questions outlined above, i) what is the continuous velocity field that describes the deformation of the lithosphere as a whole, and ii) how does discontinuous motion on faults in the upper crust approximate this velocity field, are kinematic: *i.e.* they are concerned with a description of the motions rather than with the forces or stresses that cause them. It is possible to separate these kinematic questions from those relating to the driving forces or dynamics in just the same way as proved so effective in the 1960s, when study of the geometrical aspects of deformation in the ocean basins led directly to the kinematic description of their behaviour known as Plate Tectonics. A complete understanding of the tectonics of a region would also include a knowledge of the origin, orientation and magnitude of the forces driving the deformation, and information about this may sometimes be inferred from characteristics of the deformation itself. However, a first step is to understand how the observed distributed deformation accommodates the motion of the bounding plates, and that is the subject of this paper. It will be illustrated by examples from the Aegean Sea region, which is one of the most rapidly deforming parts of the Alpine-Himalayan mountain belt.

2. Faulting and block motions in the Aegean

Figure 1 shows the seismicity of the Aegean region, and clearly demonstrates that the active deformation is distributed over a wide area. The part of interest to this paper is the northern Aegean region, between $(38\div42)^\circ$ N and $(22\div28)^\circ$ E, which is identified by shading in fig. 2. This is the most active area behind the Hellenic trench system, and the faulting within it must accommodate two types of motion: the westwards motion of the central Turkey relative to Eurasia, which occurs at about $(30\div40)$ mm/y, and the SSW motion of the almost aseismic central and southern Aegean (roughly $(35\div38)^\circ$ N, $(23\div27)^\circ$ E) relative to Eurasia, which occurs at about $(30\div60)$ mm/y. Taymaz *et al.* (1991a,b) summarize the evidence for these estimates.

A great deal is known about the active faulting in the northern Aegean, from geological studies (*e.g.*, Mercier *et al.*, 1991) and studies of the focal mechanisms of earthquakes (*e.g.*, Taymaz *et al.*, 1991a). East of ~31° E the North Anatolian Fault accommodates virtually all the motion between Turkey and Eurasia. West of 31° E the North Anatolian Fault splays into a series of sub-parallel right-lateral strike-slip faults that cross NW Turkey and the northern Aegean with a NE-ENE strike. These strike-slip faults do not cross Greece to meet the Hellenic Trench, but end abruptly in the western north Aegean Sea, where they meet the normal fault system with a NW-WNW strike that dominates the structure of central Greece. The way in which these two systems of faults move is illustrated schematically by a simple model proposed by Taymaz *et al.* (1991a), illustrated in fig. 3. The model consists of two sets of slats, which represent fault-bounded blocks, that are in relative motion. The slats are joined to the eastern and western margins of the model by pivots that cannot separate. Where the two sets of slats meet, they are joined by pivots that can move to change their separation. The eastern margin of the model rotates clockwise, representing the distributed right-lateral shear as the North Anatolian Fault splits into its several sub-parallel branches. The relative motion between the eastern set of slats is mostly right-lateral strike-slip, and they all rotate slowly anticlockwise relative to the top of the model (Eurasia). The western set of slats rotate faster and clockwise, and the motion between them is mostly normal (*i.e.*, extensional).

Taymaz *et al.* (1991a) show that the velocities of all points on these slats can be calculated analytically. The slip vectors between the slats (fault blocks) for a configuration of the model that best resembles the north Aegean are shown in fig. 4a), adapted from Taymaz *et al.* (1991a). Note that this configuration differs from the simple cartoon in fig. 3 in that the western margin of the model is also allowed to rotate clockwise, though at a slower rate than the eastern margin. This is in agreement with paleomagnetic declinations in Pliocene rocks, which show that the western seaboard of Greece has rotated clockwise relative to Europe at a rate of about 5°/My for the last 5 My (Kissel and Laj, 1988). Taymaz *et al.* (1991a) and Jackson *et al.* (1992) discuss how well this simple model describes the faulting, paleomagnetic, geodetic and geological observations in the northern Aegean. A limitation of the model is that, because the pivots on its eastern margin cannot separate, it takes no account of the N-S extension of western Turkey, which is manifest by the major E-W graben bounded by normal faults that dominate the geological structure there. With this proviso, the

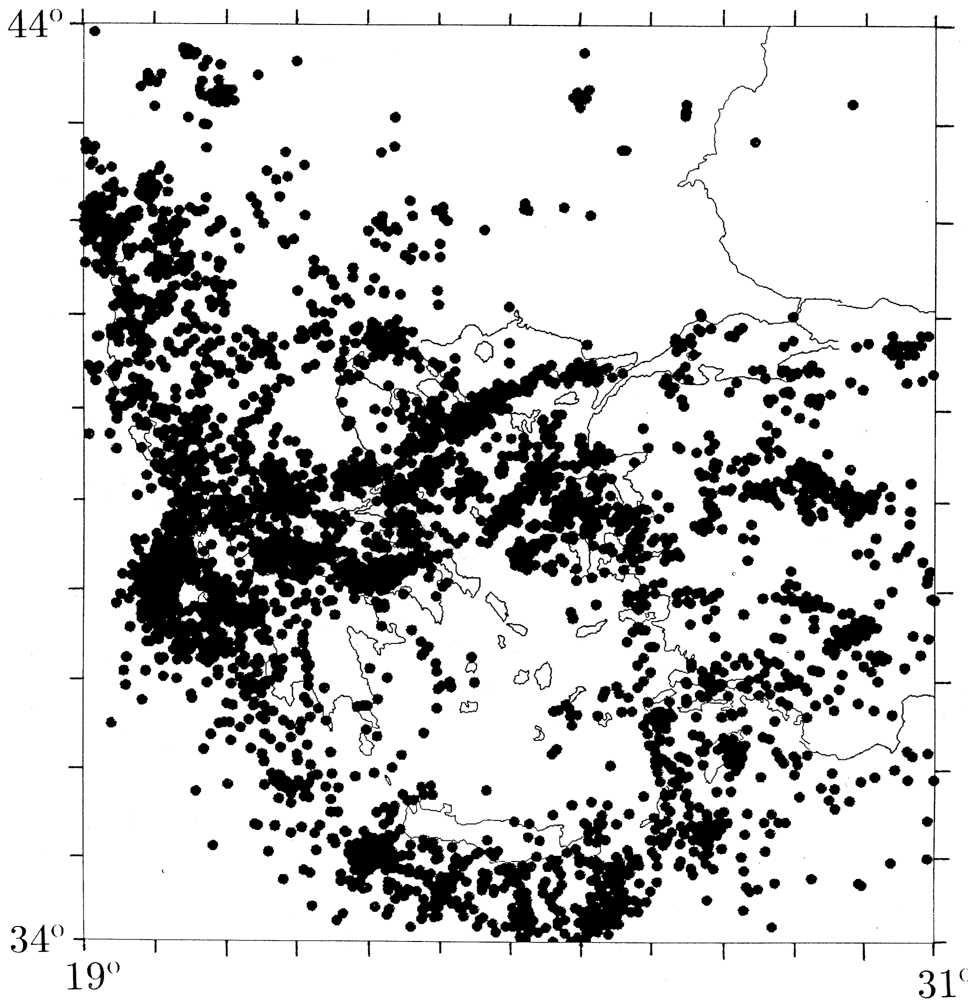

Fig. 1. Earthquakes with hypocentres shallower than 50 km, reported by the U.S. Geological Survey, 1961-1988.

model does an adequate job of describing how distributed strike-slip and normal faulting, combined with rotations about a vertical axis, take up the westward motion of Turkey and the N-S extension of the Aegean.

3. The velocity field

Whereas figure 4a) shows the slip vectors between the slats (*i.e.* the rates and directions of slip on the faults that separate them), figure 4b) shows the velocities of points on the slats, relative to the top pivot on the left-hand margin (*i.e.* conceptually relative to Eurasia). Figure 4b) is the *velocity field* accommodated by the faulting in the model. Note that the westerly component of velocity increases southwards (accommodating the E-W right-lateral shear between Turkey and Eurasia), and that the N-S component of

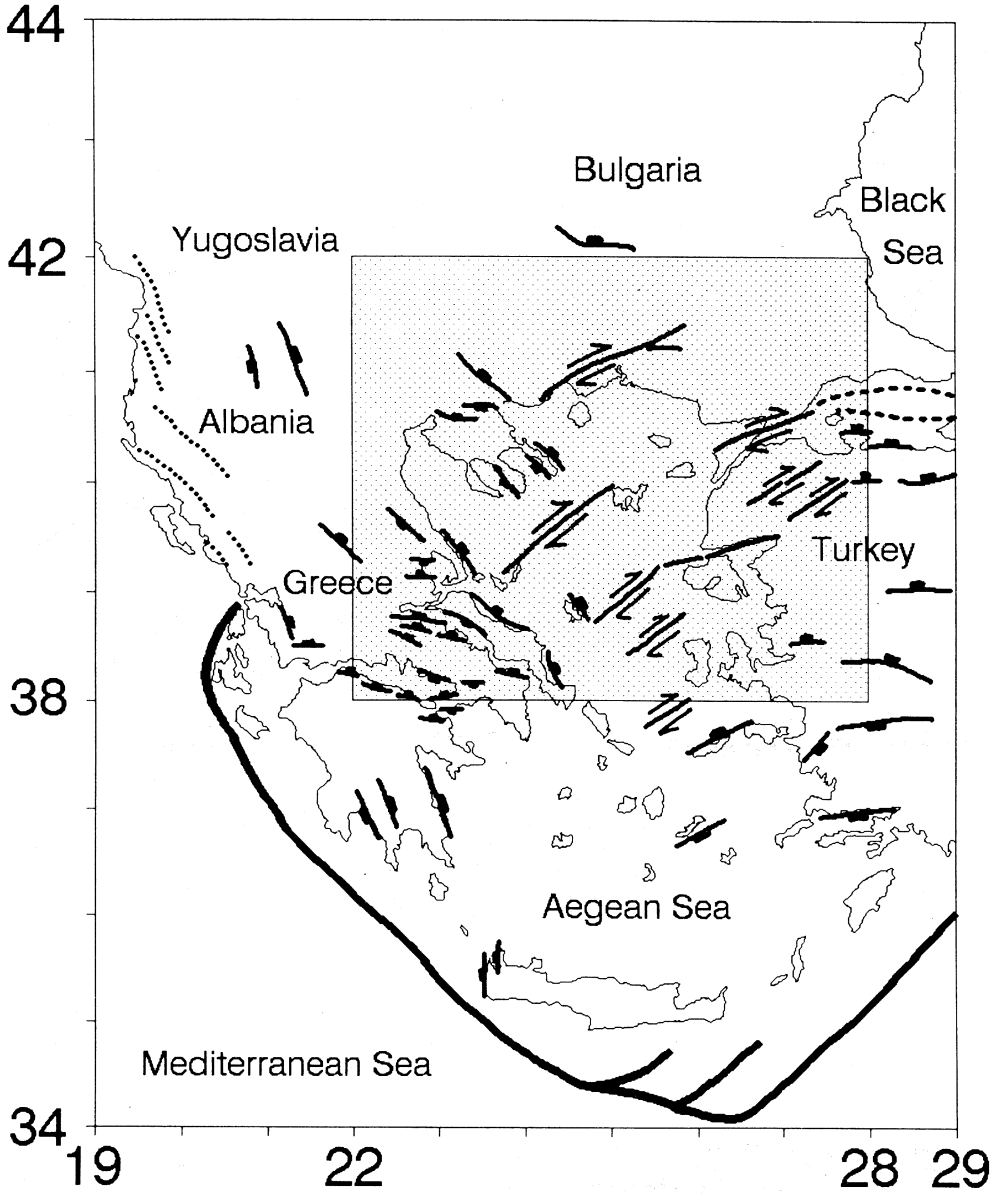

Fig. 2. Sketch map showing the orientations of the principal active structures in the Aegean region. The thick solid line in the south is the Hellenic Trench. The dotted lines in coastal Albania and NW Greece are anticline axes. Other thick lines are major active faults: those with a NE-ENE strike are mostly right-lateral strike-slip faults with some normal faulting component; those with a WNW-NW strike are predominantly normal faults, with slip vectors trending NNW-NNE (see Taymaz *et al.*, 1991a). Figures 3,4 and 6 refer to the stippled region in the central and northen Aegean.

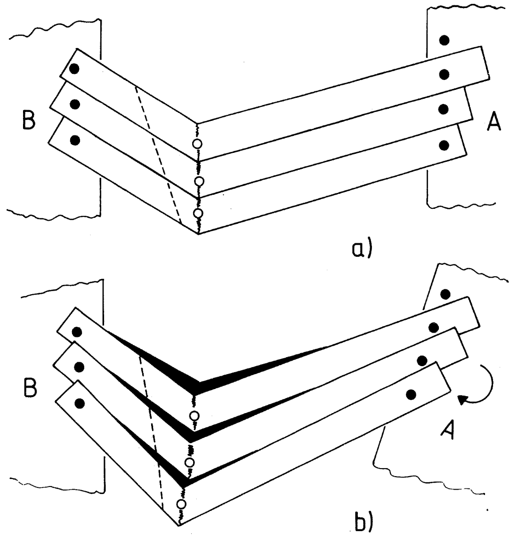

Fig. 3. Sketch of a simple model involving two sets of slats that are in relative motion. The model can easily be constructed from wood or card. Solid circles are pivots (screws) attached to plate A on the right or to plate B on the left. Open circles are screws joining two slats, but which are otherwise free to move. The configuration in a) moves to that in b) following a clockwise rotation of the right-hand margin. In this illustration the left-hand pivots, attached to B, do not rotate. In the analogy with the Aegean they rotate clockwise, but more slowly, than those on A (see fig.4). The black regions in b) represent a new surface area created by extension. The dashed line on the left-hand slats shows that the *relative* rotation between segments of an offset passive marker (such as an early Tertiary fold axis) is much less than the bulk rotation of the slats.

velocity also increases southwards (accommodating the N-S extension of the Aegean). With the rather few points shown in fig. 4b) the velocity field looks continuous, in that the length and direction of the velocity vectors change quite smoothly over the model. In fact, the velocity field is discontinuous across the faults (slat boundaries). The model in fig. 4 illustrates two different, but complimentary, ways of looking at the deforming lithosphere in the north Aegean. On the one hand, we can imagine that a smoothed version of the velocity field in fig. 4b) is a reasonable rapresentation of the pattern of continuous flow in the mantle part of the lithosphere that deforms by distributed creep. On the other hand, the slip (faulting) between the slats in fig. 4a) shows how discontinuous deformation in the seismogenic upper crust is able to achieve the same overall motion as the flow beneath it.

In fig. 4 we have considered a velocity field associated with a simple model designed to represent the faulting in the north Aegean Sea. Is it possible to obtain a velocity field that represents the long wavelength deformation of the lithosphere directly from observations? One of the difficulties we face here is that we can only make quantitative measurements of deformation rates at the Earth's surface, within the seismogenic

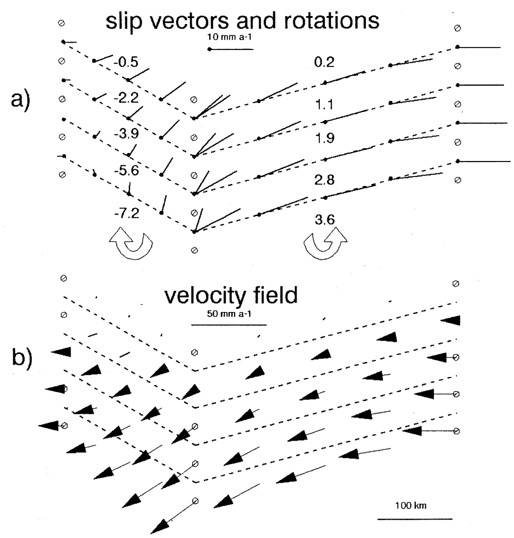

Fig. 4. The broken slat model of Taymaz *et al.* (1991), to illustrate the kinematics of fault motion in the central and northern Aegean Sea (roughly the area stippled in fig. 2). a) Block boundaries (faults) are dashed, slip vectors between blocks are shown by solid lines attached to points on the boundaries, and show the motion of the north sides relative to the south. Their magnitude is proportional to their velocity. Pivots (screws) are shown by open circles with a diagonal line. Rotation rates relative to the northern boundary are given in °/My, clockwise (negative) on the western slats, counterclockwise on the right. b) Arrows show the motions of points on the slats (the feet of the arrows) relative to the top left-hand screw, with a magnitude proportional to velocity.

upper crust that deforms discontinuously. Some progress can be made if we make the assumption that, provided we average over dimensions comparable with the lithosphere thickness, the deformation of the upper crust approximates the distributed flow in the lithosphere beneath it. This is the simplest assumption we can make, and there is some evidence that it is reasonable: the Moho is often elevated under regions of upper crustal extension and depressed beneath mountains ranges, and some of the thermal effects observed in basins and mountain ranges indicate that the deformation is not restricted to the crust, but involves the mantle as well (see, *e.g.*, McKenzie, 1978, Houseman *et al.*, 1981; England and Houseman, 1989). With this assumption, we can, in principle, use deformation rates observed at the surface to obtain an average velocity field. The most useful measurements come from geodesy, the slip rates on faults, rates of vertical motion (uplift and subsidence), and the seismic moment release in earthquakes.

Direct geodetic measurement of velocities is likely to become increasingly important over the coming decades, but at the moment there are relatively few places where geodetic measurements have been made over a sufficient time interval and with sufficient accuracy for the signal to be well above the noise. Examples include California (Lisowski *et al.*, 1991), New Zealand (Walcott, 1984) and part of central Greece (Billiris *et al.*, 1991). Attempts have been made to obtain the velocity field from the moment release of earthquakes in Asia (Holt *et al.*, 1991a; Holt and Haines, 1992) and the Aegean (Jackson *et al.*, 1992), and from Quaternary slip rates on faults combined with uplift rates in New Zealand (Holt *et al.*, 1991b). A problem with the use of all these data types is the choice of a reference frame. Most of the observations measure the symmetric part of the velocity field (*i.e.*, the strain rate tensor) and are unaffected by a rigid-body rotation (see Jackson and McKenzie, 1988). Yet abundant paleomagnetic data in deforming regions attest to the importance of rigid-body rotations in the deformation. This problem may potentially be overcome in two ways: either by extending the area of geodetic observations out of the deforming region into a rigid plate, which is now possible with space-based techniques; or by splitting the deforming region into small enough subregions that the *spatial variations* in the components of the strain rate tensor may be described. These spatial variations countain information about relative rotations, which can be used to reconstruct a velocity field relative to some given frame (see Haines, 1982; Walcott, 1984; Holt *et al.*, 1991a, b; Jackson *et al.*, 1992). This is illustrated by the cartoon in fig. 5. The question then is whether the data are of sufficient quality and density to justify this treatment. Figure 6 shows the velocity field in the Aegean obtained (by Jackson *et al.*, 1992) from the spatial variations in strain rates estimated from earthquake moment tensors over the time interval 1909-1983. Velocities are shown relative to the top of the picture (Eurasia). They agree favourably in orientation and magnitude with those in the simple model of Taymaz *et al.* (1991a) in fig. 4b), with the exception of those in western Turkey, which are directed SW (in fig. 6) rather than W (in fig. 4b), because fig. 6 takes account of the N-S extension in Western Turkey which is ignored in fig. 4b).

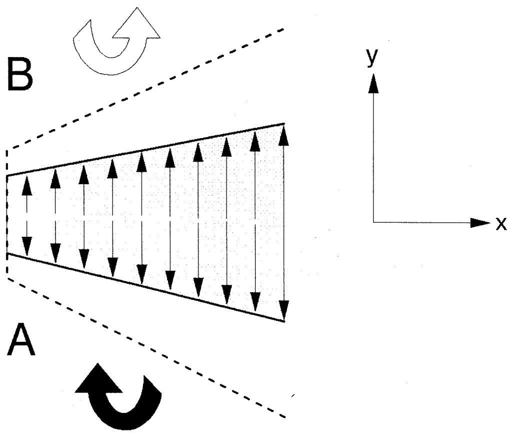

Fig. 5. Cartoon sketch of a plane view of a deforming zone (stippled) in which the extension in the y direction increases in the positive x direction. The boundaries of the deforming zone at a later time are shown dashed. Relative to plate A, plate B has rotated counterclockwise (open arrow), but relative to plate B, plate A has rotated clockwise (filled arrow). These relative rotations are clear *provided* we know that the extension increases in the x direction: if all we know is the *average* extension of the deforming region, we cannot determine any relative rotations between the blocks either side.

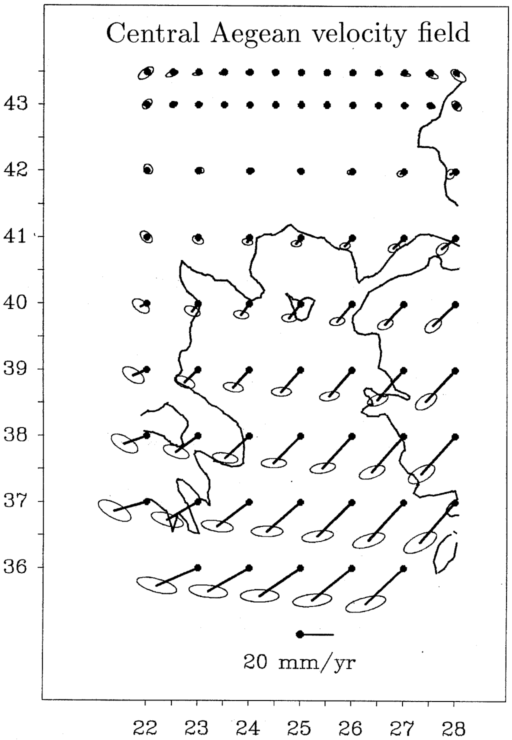

Fig. 6. The velocity field in the central Aegean Sea determined from the spatial distribution of seismic moment tensors during the period 1909-1983 (from Jackson *et al.*, 1992). Velocities are shown at points on a grid marked by filled circles, and are calculated relative to the top of the figure (essentially stable Eurasia). Error ellipses are shown at the ends of the velocity vectors. The lengths of the vectors are proportional to their magnitudes: a scale is shown at the bottom of the figure.

4. Discussion

What is the goal of such investigations? At one level, if a continuous velocity field is the most appropriate way to describe the long wavelength deformation of the continents, we want to know what that velocity field looks like. For example, we would like to know how fast, and in what direction, Crete is moving relative to Eurasia. If we know the velocity field, and also how faulting accommodates it in the upper crust, we can ask why the faulting does it this way rather than in any other. This is a significant geological question, as one consequence of the deformation being distributed is that there are usually several different possible ways in which the faulting can achieve the same overall motion. This is illustrated in fig. 7, which shows three different ways in which faulting can accommodate oblique extension across a deforming zone. In the case of the Aegean, an obvious question is why the right-lateral strike-slip faulting in NW Turkey does not simply continue across Greece to join the Hellenic Trench. The answer is probably that the pre-existing structure of Greece has a strong fabric aligned NW-SE inherited from the early Tertiary shortening deformation (*e.g.*, Aubouin, 1965), and that this fabric favours the motion of todays's normal faults that are in some cases

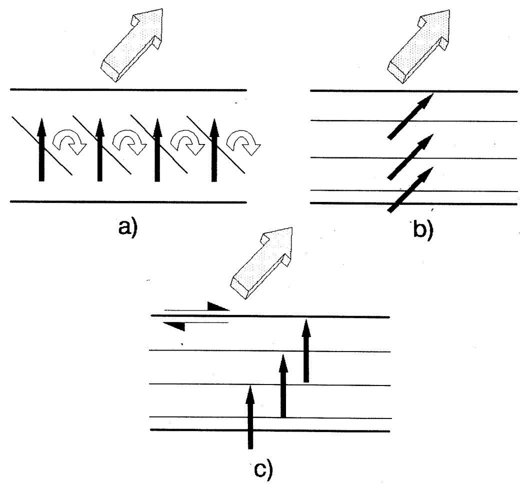

Fig. 7. Three plan views of deforming zones bounded by stable plates. The boundaries of the zones are shown by thick lines. The overall motion across each zone is the same, and is in the direction of the big shaded arrow: it involves extension and right-lateral shear. Thin lines are faults, and the black arrows show the horizontal projections of slip vectors on them. In a) the overall motion is achieved by left-lateral and normal slip on faults that are oblique to the strike of the zone and rotate clockwise about a vertical axis (open arrows). This scheme is that suggested by McKenzie and Jackson (1986). In b) the overall motion is achieved by right-lateral and normal slip on faults parallel to the deforming zone, with slip vectors parallel to the direction of overall motion. The faults do not rotate about a vertical axis. In c) the motion is partitioned into pure normal faulting and pure right-lateral strike-slip faulting.

almost sub-parallel to it, rather than the formation of new strike-slip faults that cross the fabric. The faults in central Greece then achieve the overall deformation of the velocity field by rotating clockwise about a vertical axis, relative to Eurasia (Kissel and Laj, 1988). Such rotations have attracted much attention, as they may contain information about the extent to which the upper crust deforms passively in response to the flow beneath (*e.g.*, McKenzie and Jackson, 1983; Lamb, 1987; Jackson and Molnar, 1990; Haines and Holt, 1992; Jackson *et al.*, 1992). Finally, the velocity field may contain clues to the dynamics of continental deformation: why do the high strain rates concentrate where they do, for example around the edges of plateaus and basins; why do some areas within the deforming region behave like almost rigid blocks (*e.g.*, the southern Aegean Sea); what controls the scale over which the velocity gradients are observed? A knowledge of the velocity field and how faulting accommodates it is thus a prerequisite for other questions of great current interest in continental tectonics.

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