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Microplate versus continuum descriptions of active tectonic deformation

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Abstract. Whether deformation of continents is more accurately described by the motions of a few small rigid plates or by quasi-continuous flow has important implications for lithospheric dynamics, fault mechanics, and earthquake hazard assessment. Actively deforming regions of the western United States, central Asia, Japan, and New Zealand show features that argue for both styles of movement, but new observations are necessary to determine which is most appropriate and at what scale the description applies. Geologic, geodetic, seismic, and paleomagnetic measurements tend to sample complementary aspects of the deformation field, so an integrated observation program can utilize the strengths of each method and overcome their separate spatial or temporal biases. Provided the total relative motion across each region is known and the distribution of active faults is well mapped, determination of fault slip rates can provide potentially decisive constraints. Reconnaissance geological studies supply useful slip rate estimates, but precise values depend upon detailed intensive investigation of individual sites. Geodetic survey measurements can determine the spatial pattern of contemporary movements and extract slip rate information, but the sometimes elusive effects of cyclic elastic strain buildup and relief must be accounted for in relating current movements to the long-term deformation pattern. Earthquake catalogs can be applied to determine seismic strain rates and relative velocities but must be averaged over large regions and are usually limited by the inadequate duration of historical or instrumental seismicity catalogs. Paleomagnetic determinations of vertical axis rotations provide estimates of block rotation rates but are often locally variable and averaged over many millions of years. Which of the two descriptions of continental tectonics is more nearly correct depends on the local rheological stratification of the lithosphere, especially the strength and thickness of the elastic crust relative to the ductile lithosphere, and dynamical models can provide contrasting forecasts of observable features with testable consequences.

1. Introduction

From the earliest years of the development of the plate tectonic model it was recognized that continental and oceanic lithosphere deform differently, with movements broadly distributed in plate boundary regions on continents and narrowly confined in ocean basins [e.g., *Isacks et al.*, 1968; *Dewey and Bird*, 1970]. Why this should be so, how best to describe the continental deformation, and what dynamical models are most appropriate in explaining the observed movements have been widely discussed. The purpose of this paper is to review the main developments in this research, outline unsolved problems, and suggest new work. Related reviews published recently include *Molnar* [1988], *England and Jackson* [1990], and *Gordon and Stein* [1992].

For what follows, it is useful to consider two contrasting kinematic models proposed to describe the distributed deformation seen on continents (Figure 1). In the first, analogous to plate tectonics, relative motion is taken up on a few major faults with displacements and strains thus concentrated near these features and with little deformation occurring in the intervening crustal blocks (Figure 1a). In the second, movements occur on numerous faults of comparable importance and straining is rather evenly distributed throughout the deforming region (Figure 1b). The two models clearly represent end-member idealizations. Gradational behavior is thus likely, and features of both models may well be embedded in the deformation pattern of a particular study area.

A variety of methods have been applied to delineating the patterns of movements and distinguishing between the two contrasting models. In much of the Mediterranean, Middle East, and central Asia, methods analogous to those applied to studies of oceanic plate boundaries have been used to define broadscale patterns of deformation. Reconnaissance-scale geologic mapping and interpretation of satellite photographs have outlined the tectonic framework of each active region, while earthquake locations, fault plane solutions, and historical and instrumental seismicity catalogs have been applied to determine the distribution and style of activity and estimate rates of deformation. In many active regions, crustal blocks lying within deforming zones are rotated about vertical axes, and paleomagnetic measurements can detect these motions, map their spatial distribution, and provide estimates of rotation rate. In a few intensively studied regions (e.g., western United States, New Zealand, Japan) a considerable refinement in the estimation of defor-

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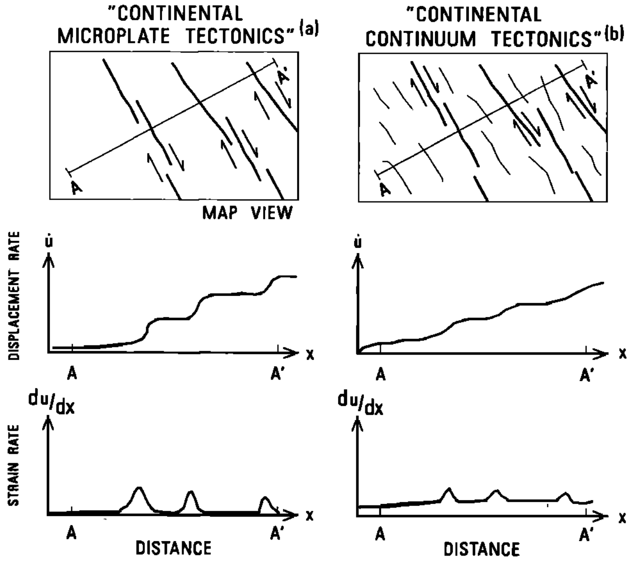

Figure 1. Two kinematic models for describing continental deformation: (a) Continental microplate tectonics and (b) continental continuum tectonics. Shown in each case are (top) a map view of the deforming region (major faults heavy lines, smaller faults fainter), (middle) the relative displacement rate across line AA′, and (bottom) the strain rate across the same profile. For simplicity, fault motion is assumed to be of the same type and sense across all faults.

mation rates is provided by geological determinations of fault slip rates and geodetic measurements of contemporary movement patterns.

Which of the two kinematic models shown in Figure 1 is more nearly correct has important implications for the material properties of continental lithosphere and its spatial variability. Observed movement patterns can thus be applied to obtain bounds on the rheological behavior of continental lithosphere, and recent work has begun to exploit this capability.

In this paper I review progress in mapping continental deformation in active regions and in constructing dynamical models to account for observed movements. Section 2 shows inferred deformation patterns from two different regions to demonstrate the range of observed movements and to illustrate the advantages and limitations of different data sources in delineating current motions. Section 3 reviews ideas on the origin of differences between oceanic and continental deformation and discusses several approaches to modeling observed deformation on continents. Finally, section 4 discusses trends in current research and suggests how new work might be focused to best address outstanding unresolved issues.

2. Observed Deformation

Middle East

The collision of the African, Arabian, and Indian plates with Eurasia has produced dramatic and widespread deformation throughout much of southern Europe, the Middle East, and central Asia [*McKenzie*, 1972; *Molnar and Tapponier*, 1975]. These movements nicely illustrate several characteristic features of continental deformation and highlight some major unsolved problems in understanding the patterns of observed deformation.

Figure 2 shows the recent seismicity of the Middle East region alongside *Jackson and McKenzie*'s [1984] interpretation of the main features of the current deformation derived from seismicity, earthquake focal mechanisms, surface faulting data, and interpretation of Landsat (satellite) photographs. In this view of the deformation the Middle East is roughly equally divided between active belts and intervening inactive regions. The deforming belts are 200–500 km wide and up to 2000 km long. The inactive blocks in central Turkey and Iran are 500–800 km long and roughly equidimensional. If these blocks are assumed to be perfectly rigid, the known convergence rate between Arabia and Eurasia and the slip vectors of earthquakes along the block boundaries can be used to determine internally consistent esti-

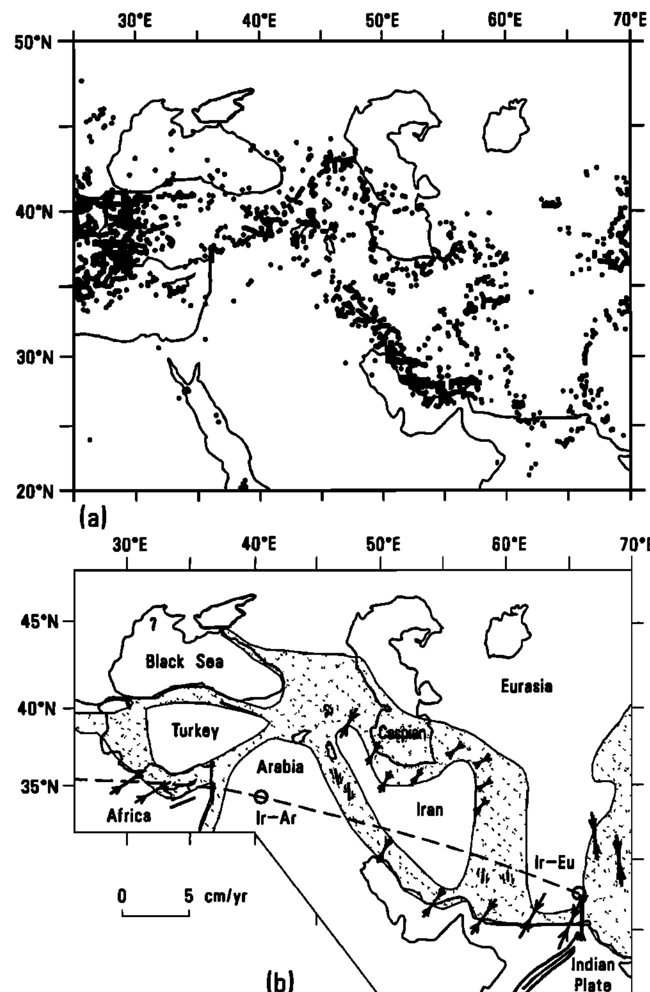

Figure 2. Earthquake epicenters and inferred broadscale tectonics of the Middle East region [*Jackson and McKenzie*, 1984]. (a) Seismicity shallower than 50 km determined by NOAA during 1961–1980. (b) Contemporary deformation pattern derived from seismicity, faulting patterns, and earthquake fault plane solutions. Regions of most intense deformation are shown stippled. Slip vectors within these regions are determined from relative motions between Iran-Arabia (Ir-Ar) and Iran-Eurasia (Ir-Eu), whose rotation poles are shown by circles. Dashed line marks great circle connecting these poles.

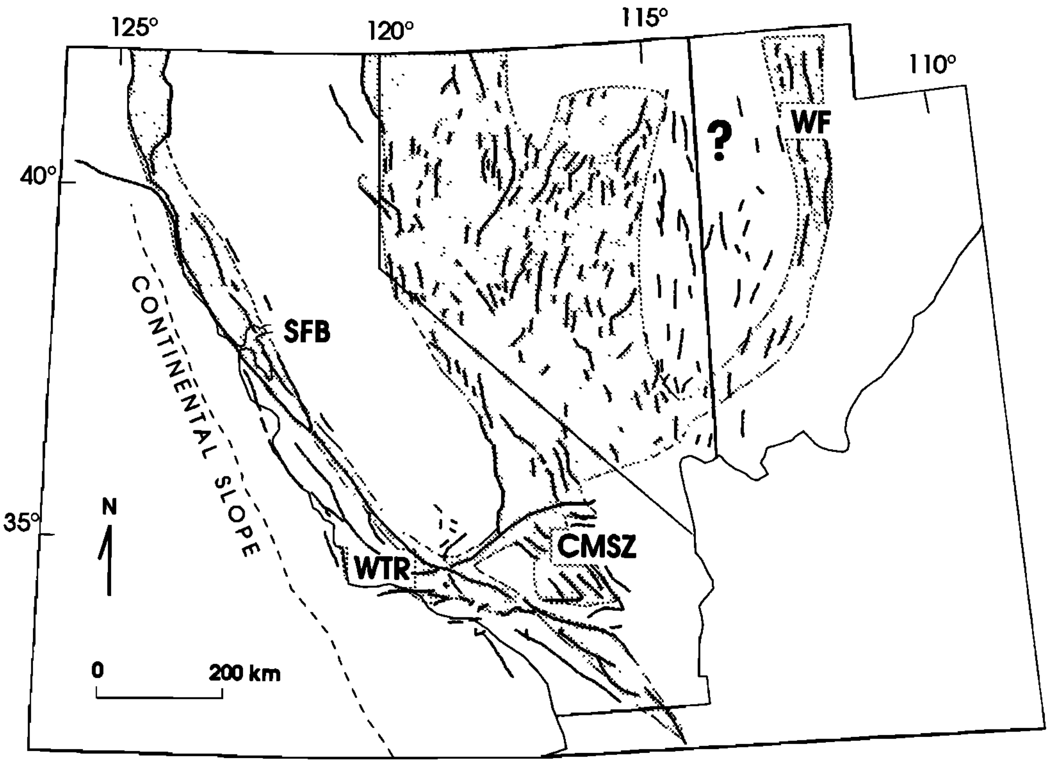

Figure 3. Active faults and inferred deforming zones (stippled) in the western United States. Stippled regions were defined based on deformation and slip rate data obtained from detailed geologic and geodetic investigations. WTR western Transverse Ranges; CMSZ, central Mojave Shear zone; and WF, Wasatch fault; SFB, San Francisco Bay.

mates of the relative motions of central Turkey and Iran [*Jackson and McKenzie*, 1984] (see Figure 2). The gross picture that emerges from these estimates is that the convergence of the Arabian shield with Eurasia is causing the lateral expulsion of Turkey and Iran from the central deforming zone lying between the Black and Caspian Seas.

Obtaining reliable estimates of rates of movement within deforming zones like the stippled regions in Figure 2 is difficult. A complete specification of the deformation field requires knowing the sense and rate of slip across all major faults within the region of interest. Style and sense of faulting can frequently be determined from reconnaissance-scale geological mapping and earthquake fault plane solutions, but only arduous geologic studies or detailed geodetic surveys can accurately determine fault slip rates and local rates of deformation. Such studies have only recently begun in the Middle East, and thus it is not yet possible to say whether deformation is quasi-continuous within the stippled regions of Figure 2 (Figure 1b) or is concentrated in much narrower zones on one or a few major faults (Figure 1a).

Western United States

In the western United States, regional fault mapping and earthquake distributions define the broadscale movements and delineate active and inactive regions (Figure 3). In addition, detailed geologic, geodetic, and seismological data provide the local measurements necessary to characterize the movement patterns within several of the main deforming regions.

Geologic and geodetic data from throughout California [*Brown*, 1990; *Thatcher*, 1990] have been used to define the zones of most intense deformation in Figure 3. Although motions are relatively narrowly confined near the San Andreas fault system throughout its 1300-km length, the deformation zone varies in width from <1 km in central California to >100 km in its northern and southern sections. In the western Transverse Ranges of southern California (Figure 3), paleomagnetic measurements complement the available geologic and geodetic data, showing evidence of block rotations of as much as 35° during Neogene time [*Hornafius et al.*, 1986]. Movements on the 50 to 100-km-wide central Mohave shear zone (near 35°N, 118°W in Figure 3) [*Dokka and Travis*, 1990; *Savage et al.*, 1990] appear to transfer motions from the southernmost San Andreas system northward into the Basin and Range Province. There, in contrast to California, both crustal extension and strike-slip motions occur, and Quaternary faults are spread over a 800-km-wide zone. Whether motions are uniformly distributed across this region is not known. However, geological evidence [*Wallace*, 1981] suggests that late Quaternary and Holocene fault slip is concentrated in the central Nevada seismic belt (a north-south zone of active faults near 118°W in Figure 3), where four large earthquakes with extensive surface faulting have ruptured much of this 500-km-long zone during the past century. In addition, detailed geological studies of the Wasatch fault in central Utah (near 112°W in Figure 3) indicate active dip-slip faulting at a rate of 1–2 mm/yr [*Schwartz and Coppersmith*, 1984].

In western California, seismicity is regionally rather dif-

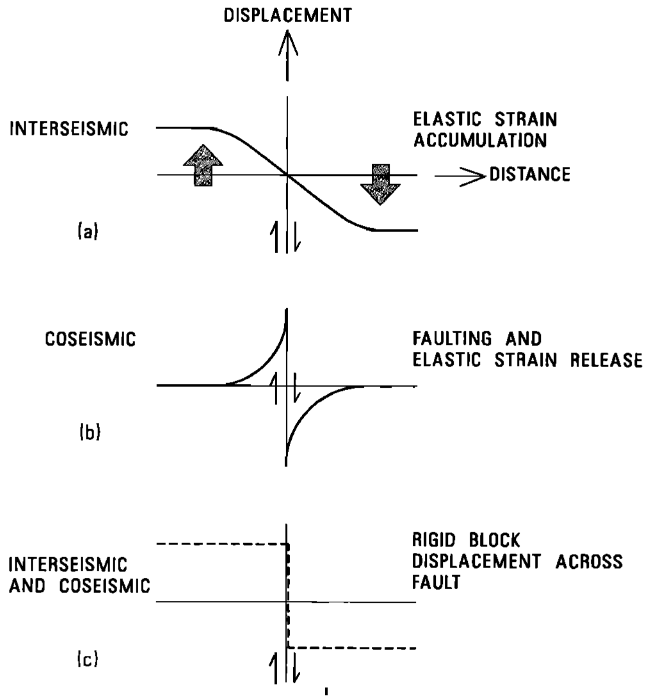

Figure 4. Idealized earthquake deformation cycle for vertical strike-slip fault. Horizontal displacement versus distance from fault for (a) time interval between successive large earthquakes (interseismic period), (b) earthquake (coseismic) strain release, and (c) sum of interseismic and coseismic displacements.

fuse [see *Engdahl and Rinehart*, 1988], but local microearthquake network data show clear concentrations on the major strike-slip faults of the San Andreas system [see *Hill et al.*, 1990, Figure 5.6]. However, since the microearthquakes account for only a small proportion of the total relative motion across the San Andreas and its major strands, they cannot be used to provide a quantitative picture of the distribution of movements. Although a seismicity catalog complete back to about 1800 for events larger than about $M6.5$ has been compiled by *Ellsworth* [1990], it is unlikely that this record is long enough to be useful in quantitatively characterizing the movement pattern.

Results from geologic and geodetic data on the one hand and global plate tectonic reconstructions on the other provide consistent estimates of the total right-lateral slip rate across the entire San Andreas fault system. In central California, where the San Andreas itself takes up all of the right-lateral slippage of the fault system, geologic [*Sieh and Jahns*, 1984] and geodetic [*Thatcher*, 1979] data give values of 34 ± 3 and 33 ± 1 mm/yr, respectively. Plate tectonic data suggest Pacific-North American plate relative motion rates of 49 ± 3 mm/yr [*Demets et al.*, 1990]. However, correcting for the contribution of movements in the Basin and Range Province and the intervening Sierra Nevada block to these motions suggests relative movements of about 35 mm/yr across western California [*Argus and Gordon*, 1991], in accord with the geologic and geodetic estimates of total displacement rate across the San Andreas system.

With this constraint on total motion across the San Andreas system, geodetic survey measurements spanning the broader, more complex portions of the San Andreas system and geologic estimates of slip rates on individual faults can be applied to infer long-term motions within the deforming zone. Such data can thus be used to distinguish, on this more local scale, between microplate and continuum descriptions of the deformation.

However, in seismogenic regions near active faults, crustal movements are often dominated by the cyclic buildup and release of elastic strains, and in order to relate geodetic measurements to long-term movement patterns these effects must be accounted for. This is illustrated in Figure 4, which shows a simplified view of the earthquake deformation cycle first suggested by *Reid* [1910]. In this view, crustal movements between successive earthquakes represent elastic strain energy stored in the seismogenic upper crust. This built-up strain is then released by the cycle-ending event, with the cumulative interseismic strain accumulation exactly balanced by coseismic strain release. Although geodetic measurements made during the interseismic period would show movements distributed quasi-continuously across the fault, the long-term average pattern is that shown in Figure 4c, a rigid block displacement across the fault.

The effect of interseismic straining on geodetic measurements made across a complex deforming zone is schematically illustrated in Figure 5. The detailed pattern depends on the number and spacing of faults traversed and their type, dip, and slip rate. However, the spatial extent of the zone of cyclic straining is determined by the downdip seismogenic fault width, and this zone will be wider for low-angle faults than for vertical ones. If the spacing of faults is much greater than the zone of cyclic straining, the net offset across

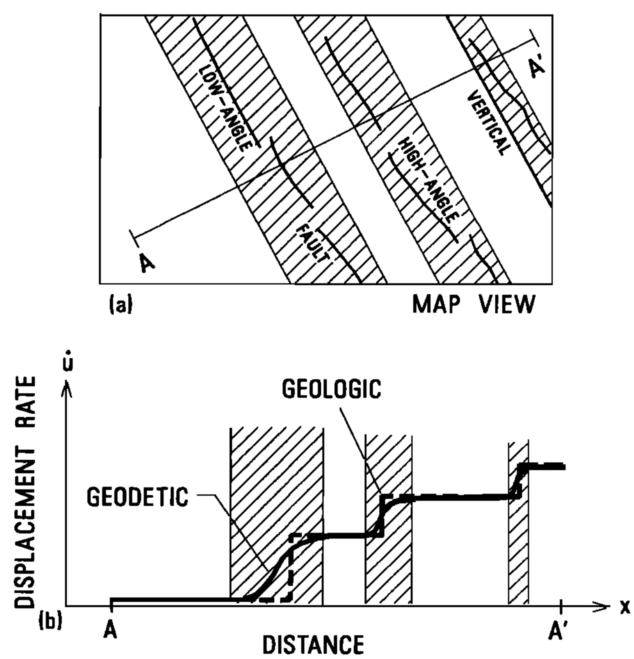

Figure 5. Effect of interseismic strain accumulation on geodetic survey measurements of displacement rate across profile AA′. Hatched areas, (a) map view and (b) diagram, are regions within which measurements are affected by cyclic strain buildup and release. Solid line in Figure 5b denotes geodetically measured velocities, and dashed line shows long-term velocities assuming all deformation is due to rigid block motions across the faults indicated.

individual faults can be readily determined (see Figure 5). However, if the faults are more closely spaced, individual offsets cannot be distinguished, and the displacement rate profile may be roughly continuous.

Figure 6 shows geodetic data from the San Francisco Bay area (see Figure 3 for location) that demonstrate some of these features. In this region the major strike-slip faults are vertical and are separated by only 10–30 km, comparable to the 10–15 km thickness of the seismogenic upper crust. The cumulative right-lateral movement rate across the network is about 34 mm/yr, close to the geologic and geodetic estimates for the entire San Andreas system, suggesting this net has completely spanned the deforming zone. The geodetic data show that straining is relatively continuous across the network, suggesting the continuum tectonic model may be appropriate here. However, historical large earthquakes have occurred on both the San Andreas and Hayward faults [*Ellsworth*, 1990], where geological estimates of slip rate are 10–20 mm/yr and 10 mm/yr, respectively [*Brown*, 1990]. Thus despite the character of the geodetically measured deformation it seems likely that a kinematic description close to the block model applies for the long-term movements across this complex deforming zone.

3. Mechanics of Continental Deformation

The patterns of lithospheric deformation are determined by conditions applied at the boundaries of the deforming region, by forces acting within it, and by its rheological properties (i.e., how it deforms in response to applied stresses). The boundary conditions are often constrained by the rates and senses of movement of the major bounding tectonic plates, and the effects of internal forces like gravity are usually known. In contrast, relatively little is certain about the rheological properties of continental lithosphere, in particular, how rheology varies laterally and as a function of depth. Observed movement patterns can be applied to obtain bounds on the rheological behavior of continental lithosphere, and recent work has begun to exploit this capability.

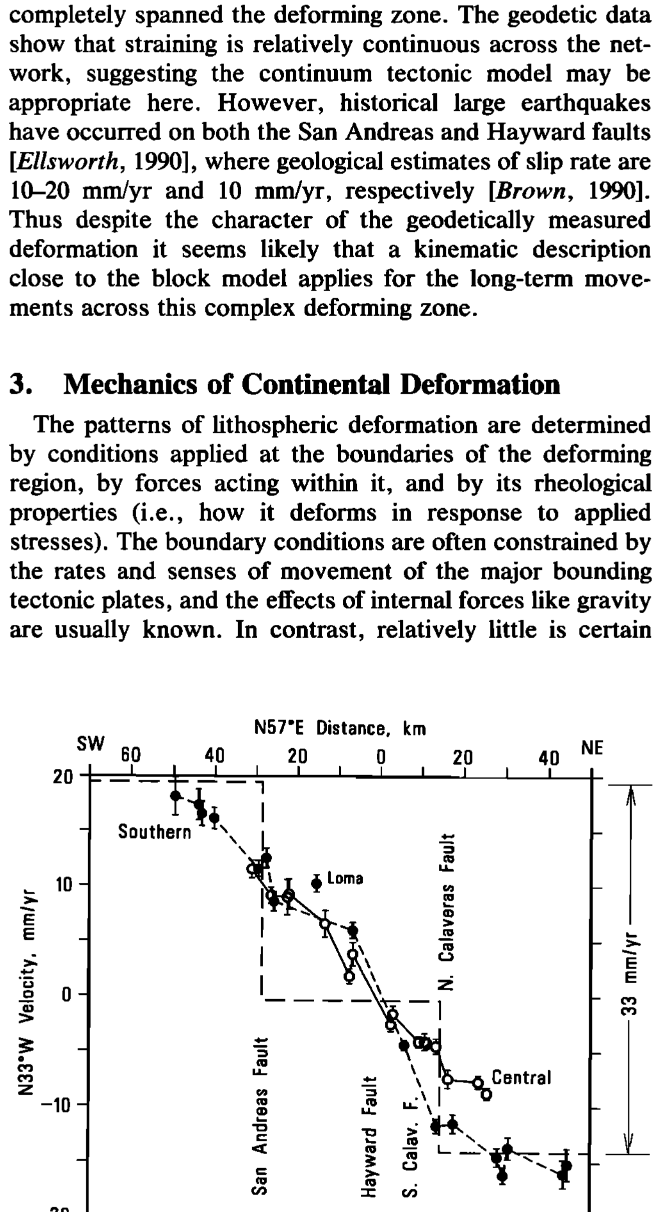

Figure 6. Geodetic measurements of velocities parallel to the major right-lateral strike-slip faults of the San Francisco Bay area [*Lisowski et al.*, 1991]. Results for two profiles, one across the central bay and the other across the southern bay are shown, and one standard deviation error bars are shown for each point. Heavy dashed line shows velocity profile for the case of rigid block motions of 20 mm/yr across the San Andreas fault and combined slip rate of 13 mm/yr across the Hayward and Calaveras faults.

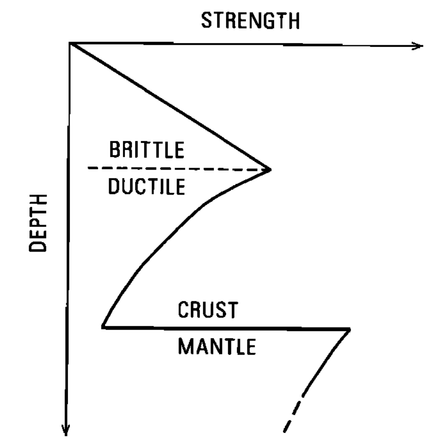

Figure 7. Maximum stress as a function of depth for the continental lithosphere inferred from laboratory rock mechanics measurements.

The variation of lithospheric strength with depth inferred from laboratory rock mechanics data (see review by D. L. Kohlstedt, Rheology of the lithosphere, submitted to *Journal of Geophysical Research* 1994) is sketched in Figure 7. In the upper crust the limiting shear stress τ is determined by the frictional resistance to slippage on preexisting faults

$$\tau = \tau_0 + \mu \sigma_n \tag{1}$$

where τ_0 is cohesive strength, μ is the coefficient of friction, and σ_n is the effective normal stress across the fault. The normal increase of lithostatic load implies that τ increases linearly with depth. Except for slippage on faults, upper crustal deformation is predominantly elastic and reversible. At greater depths the predominant deformation mechanism is thought to be ductile flow controlled by dislocation creep, which is insensitive to normal stress but very temperature- and strain-rate-dependent. For this deformation mechanism the maximum shear stress τ is of the form

$$\tau = B\varepsilon^{(1/n-1)} \exp\,[Q/RT] \tag{2}$$

where B, n, and Q are constants that depend on rock type, R is the Boltzman constant, ε is strain rate, and T is absolute temperature. Since temperature normally increases with depth, τ is greatest at the shallowest depths for which the flow law applies and thereafter decreases exponentially with increasing depth. The strength envelope for the crust is determined by the lowest stresses at a given depth in (1) and (2), and their point of intersection is called the brittle-ductile transition. Since quartzo-feldspathic rocks deform at lower stresses than more basic rocks, there may be a further increase in strength at the crust-mantle boundary.

The degree to which the laboratory-derived strength profiles in Figure 7 apply to the continental lithosphere is uncertain. Their applicability within major fault zones is doubtful. The absence of a local heat flow anomaly across the San Andreas fault in California has long been cited as

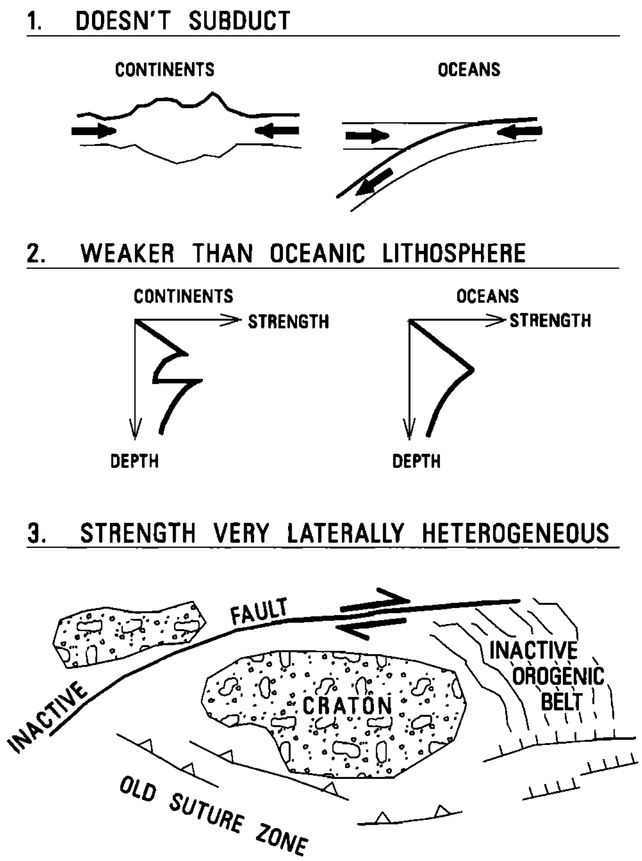

Figure 8. Three reasons cited for differences in the patterns of deformation of continental and oceanic lithosphere.

evidence for fault stresses in the brittle upper crust much lower than those predicted by (1) [*Brune et al.*, 1969; *Henyey and Wasserberg*, 1971; *Lachenbruch and Sass*, 1973, 1980]. More recently, evidence for a fault-normal compressional stress field across the San Andreas and other major transform faults has independently reinforced these arguments [*Zoback et al.*, 1987; *Mount and Suppe*, 1987]. These data have also been used to argue that upper crustal stresses increase rapidly with distance from major faults, possibly reaching magnitudes consistent with (1).

The applicability of Figure 7 to the ductile portions of the continental lithosphere is even more uncertain. Laboratory-derived frictional properties of rocks are largely independent of rock type [*Byerlee*, 1978] and do not depend upon temperature or deformation rate. However, the exponential dependence of temperature and activation energy on stress in (2) shows that ductile deformation is very sensitive to changes in rock type and thermal regime, and both of these parameters are expected to vary considerably in active regions with a complex thermal-tectonic history. Furthermore, because the ductile lithosphere is even less observationally accessible than the brittle upper crust, no useful constraints have yet been proposed for the maximum stresses that can be supported during ductile flow of the lower crust or upper mantle.

Differences in the patterns of deformation of oceanic and continental lithospheric have been attributed to three main factors (Figure 8). Because continental lithosphere is capped by crust averaging 35 km thick composed largely of quartz- and feldspar-rich rocks, it is less dense and hence more buoyant than oceanic lithosphere (crustal thickness about 8 km). It thus resists subduction, and plate convergence on continents is largely accommodated by thickening of the crust and lithospheric mantle and the formation of elevated plateaus and mountain belts. Furthermore, the stresses required to deform lower crustal rocks by ductile flow are thought to be less than those needed to deform rocks of the oceanic upper mantle, suggesting the integrated strength of the lithosphere beneath continents is less than that beneath oceans. Finally, because continents do not subduct, they are composed of fragments of varying age, tectonic history, and thermal state. Both the depth of the brittle-ductile transition and ductile strength of rocks are strongly temperature- and lithology-dependent, suggesting that variations in thermal state and rock type will lead to large spatial variations in lithospheric strength. Preexisting faults may offer additional zones of decreased strength along which new slippage may be accommodated.

Modeling active deformation and comparing predicted and observed movements offers one means of quantitatively testing these inferences. Since so little is known about lithospheric rheology, a useful starting point has been to consider very simple models with uniform or nearly uniform material properties. Models that idealize the lithosphere as a thin viscous sheet have been widely applied to a range of tectonic environments [e.g., *Bird and Piper*, 1980; *England and McKenzie*, 1982; *Vilotte et al.*, 1982]. In these models, rheology (and hence stress) does not vary as a function of depth, and lateral changes in lithospheric strength are usually ignored. Subsequent support for the first of these idealizations has come from the analysis of *Sonder and England* [1986]. They showed that applying the laboratory data embodied in (1) and (2) and using reasonable values of thermal and material parameters, the thickness-averaged rheology of the entire lithosphere could be approximated well by a single power law relation of the form

$$\tau = BE^{(1/n'-1)}\varepsilon \qquad (3)$$

where B and n' are constants, ε is the strain rate, and $E = \varepsilon^2$. The exponent n' ranges from about 3–10, with lower values corresponding to rheologies where ductile deformation dominates and higher values indicating a more important role for the brittle upper crust.

Under these approximations some general statements can be made about the length scales of thin sheet deformation for strike-slip, convergent, and divergent boundary displacements [*England et al.*, 1985]. Figure 9 illustrates these effects, showing that displacements perpendicular to the boundary result in deformation that, for the same boundary length D and power law exponent n', extends much farther into the sheet than do motions caused by boundary-parallel displacements. This inference generally accords with observed large-scale deformation on continents. Regions of plate convergence and extension are usually much broader than zones of strike-slip movements.

The deformation of western North America (Figure 3) provides a particular example of these features. There, extension is widely distributed across the 800-km extent of the Basin and Range Province, and strike-slip motions are confined within a zone of a few hundred kilometers or less in western California. However, as noted previously, the width of strike-slip deformation varies considerably along the

1300-km length of the San Andreas fault system, indicating the influence of factors other than boundary displacement on deformation zone width.

In regions of continental convergence, deformation is influenced by the internal stresses resulting from lateral variations in the thickness of the buoyant continental crust. *England and McKenzie* [1982] have shown that this effect can be economically incorporated into thin viscous sheet models by defining an additional dimensionless parameter, the Argand number, which is the ratio between stresses due to crustal thickness variations and the viscous stresses that resist lithospheric deformation. With this formulation, once boundary displacements are specified, the motions within the deforming sheet of lithosphere are determined by only two parameters. The rheological exponent n' determines the length scale of the deformation, and the Argand number controls the height of the "mountains" that can be supported by the strength of the viscous, fluid lithosphere.

This model of mountain building and evolution has been applied to the tectonics of the Himalayas and central Asia with increasing elaboration over the past decade [*England and McKenzie*, 1982; *Vilotte et al.*, 1982, 1984, 1986; *England and Houseman*, 1985, 1986, 1989; *Houseman and England*, 1986, 1993]. Despite successive inclusion of additional influences, the models remain conceptually simple, ignoring variations of rheology with depth and including at most a single rigid microplate within a continuously deforming fluid. Nonetheless, such models can account for several of the large-scale features of central Asian tectonics. As Figure 10 shows, the predicted topography mimics the actual smoothed elevation contours rather well. Although the Tibetan plateau is deforming about 4 times slower than the model predicts [*England and Houseman*, 1986, Table 3], the interior of Tibet is significantly less active than its boundaries. The Tarim Basin, north of the Tibetan plateau, is nearly completely inactive, a feature that can be accommodated within the thin sheet formulation by ad hoc inclusion of a 1000-km-scale rigid microplate in the fluid [*Vilotte et al.*, 1984, 1986; *England and Houseman*, 1985].

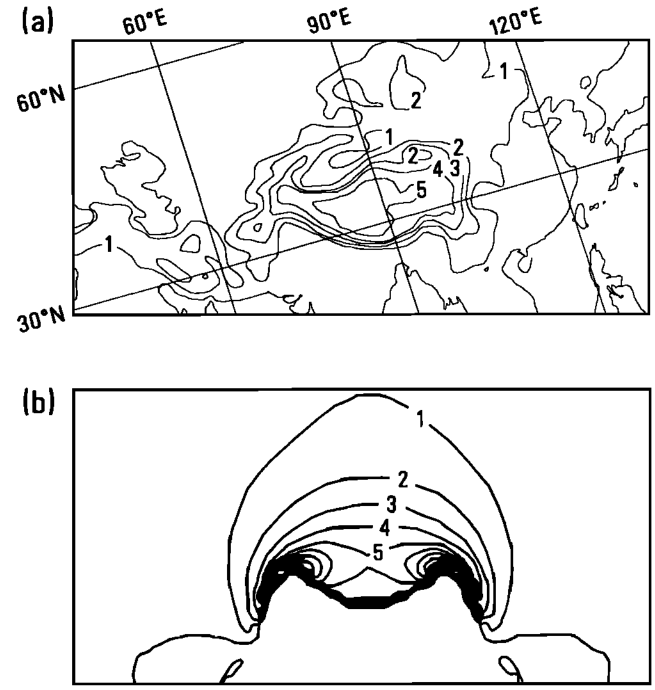

Figure 10. (a) Observed elevation of central Asia (in kilometers) compared with (b) synthetic elevation contours obtained using thin viscous sheet model of Indian-Eurasian collision (modified from *England and Houseman* [1986]).

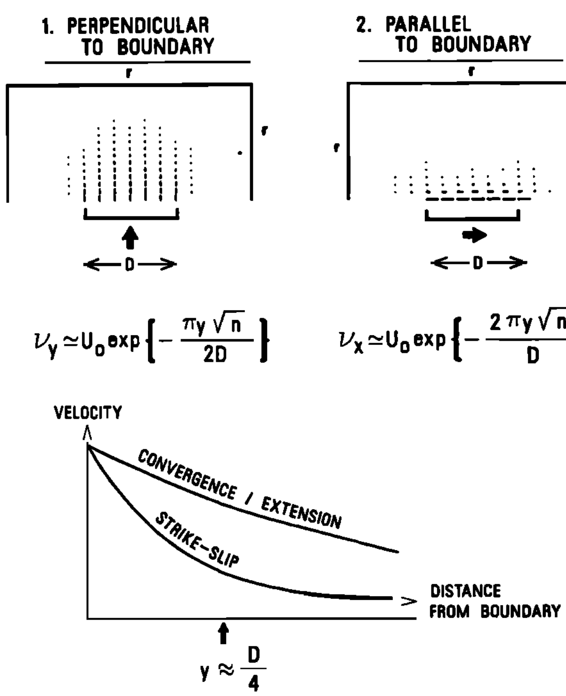

Figure 9. Scale lengths of deformation for differing boundary conditions in thin sheet model of continental deformation. (top) Velocities in interior of sheet for imposed (1) normal and (2) tangential velocities over length D of bottom boundary. Solid bounding lines labeled r denote boundaries where velocities are fixed to be zero. In each diagram the x axis is parallel to the boundary and the y axis is perpendicular to it. (middle) Approximate equations for velocity as a function of distance (y) from the boundary are shown for the two cases. (bottom) Velocities versus distance from boundary for $n = 3$. (Modified from *England and Jackson* [1990].)

Whether such continuum models are considered better or worse than rigid microplate models of the same region depends on what features are to be accounted for as well as how well they are matched. For example, *Avouac and Tapponier* [1993] proposed a kinematic model for central Asian tectonics that includes only two rigid plates, Tibet and the Tarim Basin, and obtained acceptable matches between model predictions and available slip rate data for the major plate-bounding faults. The kinematic models can thus incorporate specific tectonic features such as major faults and their slip rates, and *Lamb* [1994b] has shown how postulated microplate configurations can be tested for internal consistency and agreement with external constraints like convergence rates of the major tectonic plates. However, even if all of these requirements are met within data uncertainties, such models cannot address the dynamics of the deformation or directly account for features such as mountain belts. Conversely, simple dynamical models, however successful they might ultimately be in reproducing the broadscale deformation patterns with mechanical consistency, may not match details on length scales less than the lithospheric thickness.

4. The Future

The rapid development and widespread use of the Global Positioning System (GPS) surveying method for measuring crustal movements ensure that an increasing number of actively deforming regions will be mapped in the detail that now exists in only a few areas. Because the GPS measurements are referred to an external reference frame, rigid body rotations as well as strains can be determined, permitting significant refinements in estimates of contemporary rotation rates over paleomagnetically determined values, which are usually averaged over many millions of years. At the same time, Quaternary geologic studies in many of these same areas can elucidate the active tectonic framework and provide information on deformation styles and slip rates that complement the geodetic results. Together, these two approaches, supplemented by seismological and other geophysical methods, will produce significantly refined images of the styles and patterns of continental deformation and reveal how it is expressed under a range of driving conditions and a variety of preexisting structural states.

Such detailed deformation mappings will contribute importantly toward resolving whether deformation is predominantly confined to narrow active zones or is broadly distributed quasi-continuously (Figure 1), but results from other disciplines will be needed to determine why this is so. Although the geological and geophysical observations may themselves point to a mechanical explanation, modeling experiments will be necessary to provide rigorous tests and to suggest further investigations. With the introduction of sophisticated numerical computation codes and increasingly more powerful computers, the scope and complexity of such modeling studies are potentially very great. However, a principal limitation to our understanding of deformation processes may be neither the power of modern computers nor the ingenuity of modelers but our ignorance of the rheology of the lithosphere and its variability laterally and as a function of depth.

The problem of constraining the true variation of strength as a function of depth (Figure 7) is a particularly significant unresolved matter with important implications for tectonics and earthquake mechanics. Extrapolation of laboratory results to the continental lithosphere suggests the seismogenic upper crust is relatively strong, the lower crust is weak, and the uppermost mantle is strong again. However, there is as yet no consensus on whether this is actually the case in the Earth, what the magnitudes of the stresses are, and how much these strength profiles vary laterally.

The relative strength of the brittle upper crust and ductile lithosphere determines which of these elements controls the patterns of large-scale tectonic deformation. Thus the thin sheet models discussed in section 3 will be most applicable if the ductile lithosphere is the stronger and lateral variations in strength are modest. If the upper crust is stronger, it not only determines the patterns of surface deformation but also exerts an important influence on ductile flow in the subseismogenic crust, at least locally [*Lachenbruch and Sass*, 1992; *Lamb*, 1994a].

Whether the lower crust is as weak as laboratory data suggest and is actually pervasively ductile influences the depth to which discontinuous fault motions extend and determines the degree to which motions at depth are coupled to near-surface movements (and vice versa). Strain-softening mechanisms [see *Kirby and Kronenberg*, 1987, pp. 1227–1228] can also weaken the ductile roots of faults, narrowing the zone of deformation and promoting downward continuity of upper crustal faults.

Whether deformation below the brittle upper crust is narrowly focused on the downdip extensions of seismogenic faults or broadly distributed in bulk flow has important implications for both long-term and cyclic earthquake-related deformation. Models of the energetics and long-term deformation of the continents that consider the rheological layering of the lithosphere [e.g., *Lachenbruch and Sass*, 1980; *Bird and Piper*, 1980] commonly assume the correctness of the laboratory data and the pervasive ductility of the lower crust that follows from it. Some models of cyclic deformation also explicitly allow for ductile flow of the subseismogenic lithosphere [*Savage and Prescott*, 1978; *Thatcher and Rundle*, 1984; *Li and Rice*, 1987]. On the other hand, many, if not most, models of earthquake mechanics explicitly or implicitly assume that upper crustal faults remain narrow zones of concentrated shear well into the lower crust or uppermost mantle [e.g., *Savage and Burford*, 1973; *Thatcher*, 1979; *Savage*, 1983; *Tse and Rice*, 1986]. While the geodetic data upon which the cyclic deformation models rely are generally insensitive to the distinction between focused slip or distributed shear [*Savage and Prescott*, 1978; *Thatcher*, 1983], the mechanical implications of the two modes of lower crustal straining are very different, and their impact on modeling long-term deformation can be very great.

The rheological layering of the lithosphere also controls whether the upper crust can be mechanically decoupled from the underlying lithosphere in regions of extension or compression, either by low-angle listric faults located near the brittle-ductile transition [e.g., *Wernicke*, 1981; *Dahlen*, 1990] or by "delamination," actual detachment and sinking, of the denser mantle lithosphere [*Bird*, 1978; *Humphreys et al.*, 1984; *England and Houseman*, 1989].

Field, laboratory, and computational studies are needed to address these issues by beginning to narrow the uncertainties in rheology that limit our understanding of how the lithosphere deforms. For the brittle upper crust the state and physical properties may be sampled directly by drilling, and investigation of active tectonic processes is one of the principal justifications for the establishment of an international program of scientific drilling [*Zoback and Emmermann*, 1994]. However, much of the active lithosphere is not directly accessible, and other methods must suffice. Studies of exhumed sections of now inactive lower crust and upper mantle shear zones have led to progress in understanding the detailed mechanisms of ductile deformation [*Carreras et al.*, 1980], but a focus on the larger-scale context and physiochemical state of these regions would contribute importantly to the understanding of the active roots of major faults and their surroundings. Perhaps most critical, however, is a thorough interdisciplinary research program that delineates the range of candidate rheological profiles of the continental lithosphere and defines observable differences that permit them to be distinguished.

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