Reprinted from Nature, Vol. 335, No. 6186, pp. 131-137, 8 September 1988

Continental tectonics in the aftermath of plate tectonics

Peter Molnar

Department of Earth, Atmospheric and Planetary Sciences, MIT, Cambridge, Massachusetts 02139, USA

The success of plate tectonics required an acceptance of continental drift, and thus a reinterpretation of the large-scale geological history of most of the Earth. But the basic tenet of plate tectonics, rigid-body movements of large plates of lithosphere, fails to apply to continental interiors, where buoyant continental crust can detach from the underlying mantle to form mountain ranges and broad zones of diffuse tectonic activity.

TWENTY-FIVE years ago, to explain some subtle, almost insignificant spatial variations in the Earth's magnetic field, Vine and Matthews[1] integrated two wholly unproven and otherwise unrelated hypotheses—the creation of new sea floor at mid-ocean ridges and recurrent reversals of the Earth's magnetic field. In doing so they helped launch a revolution in the Earth sciences. Reduced to its essentials, plate tectonics is the quantitative description of the kinematics of the lithosphere, the strong outer layer of the Earth including the uppermost mantle and its overlying crust. Vine and Matthews presented a method for measuring how rapidly plates diverge from one another at oceanic ridges, or spreading centres. Because most plate boundaries are very narrow, often consisting of only one fault or fault zone, accurate mapping of the bathymetric expression of these zones and the study of earthquakes at these boundaries allowed the directions of relative motions between neighbouring plates to be determined. By the late 1960s, geophysicists had discovered an internal consistency of the rates and directions of relative motions of these vast aseismic plates, at least where the boundaries between them are indeed narrow, and therefore had demonstrated that the plates move as effectively rigid bodies over the surface of the Earth. This is plate tectonics.

Plate boundaries in oceanic regions are narrow. In contrast, seismicity and active faulting in continental regions are commonly dispersed over thousands of kilometres. Today, one of the central controversies among students of continental tectonics concerns how best to describe this widespread deformation: by the relative motion of rigid blocks ('platelets'), or by the deformation of a continuous medium that undergoes internal strain and rotation (or vorticity). This is not to say that plate tectonics has failed to help us understand how continents deform or evolve, but that the fundamental assumption of plate tectonics, rigid plate motion, provides an incomplete description of the broad zones of deformation within continents.

Growing appreciation for this limitation has defined the study of continental deformation for the last fifteen years, but not until after attempts had been made to apply plate tectonics to continental regions. As marine geologists established plate tectonics as the paradigm for the tectonics of oceanic terrains, land-based geologists were quick to realize that the geological record contained evidence for plate motions earlier in the Earth's history. For example, they recognized that the mélanges of oceanic sedimentary rock, basalt, and ultrabasic rock, typical of oceanic crust and mantle (such as those found along the coast of northern and central California), and the parallel belts of granite (cropping out in the Sierra Nevada in eastern California), constitute geological evidence for the underthrusting and off-scraping of oceanic floor beneath the granitic belt[2,3] (eastward beneath western North America). A present day analogue is the subduction of ocean floor beneath the volcanically active Andean margin of South America. Such 'geological corollaries' for processes clearly occurring elsewhere today not only allowed but required a reinterpretation of the geological histories of vast areas, such as the western United States. Regional synthesis became a respectable domain of serious scientists, a status that had been lost since the very perceptive insights of geology's grand thinkers, Wegener, Seuss and Argand, 50 to 100 years ago.

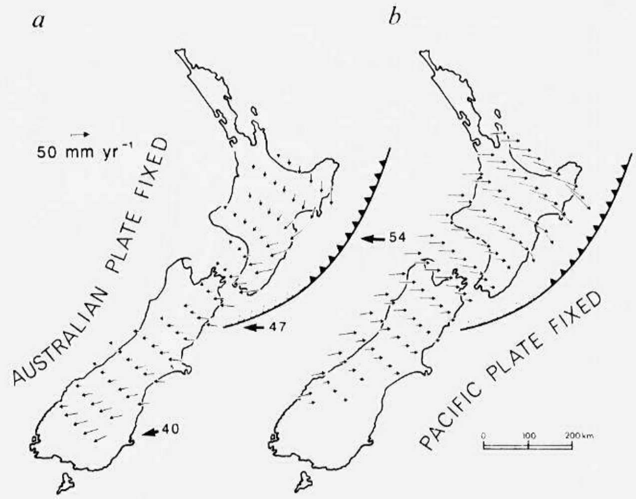

Fig. 1 Maps of New Zealand (after ref. 6) showing velocity fields over the islands relative to the Australian (*a*) or Pacific (*b*) plate. Note diffuse simple shear across both islands, as well as subduction of Pacific Ocean floor beneath the North Island perpendicular to the Hikurangi Trough (barbed line). The three large arrows indicate the relative velocity between the Australian and Pacific plates, in mm yr^{-1}.

Plate motions and continental deformation

The acceptance of continental drift, seafloor spreading and subduction of oceanic crust—ancestral concepts linked together by plate tectonics—required a qualitative reinterpretation of continental geology, but the application of the quantitative aspects of plate tectonics revealed both the full power of plate tectonics applied to continental regions, and its limitations. The Vine-Matthews hypothesis applied to the sea floor of the Pacific west of North America enabled Atwater[4] to determine its age

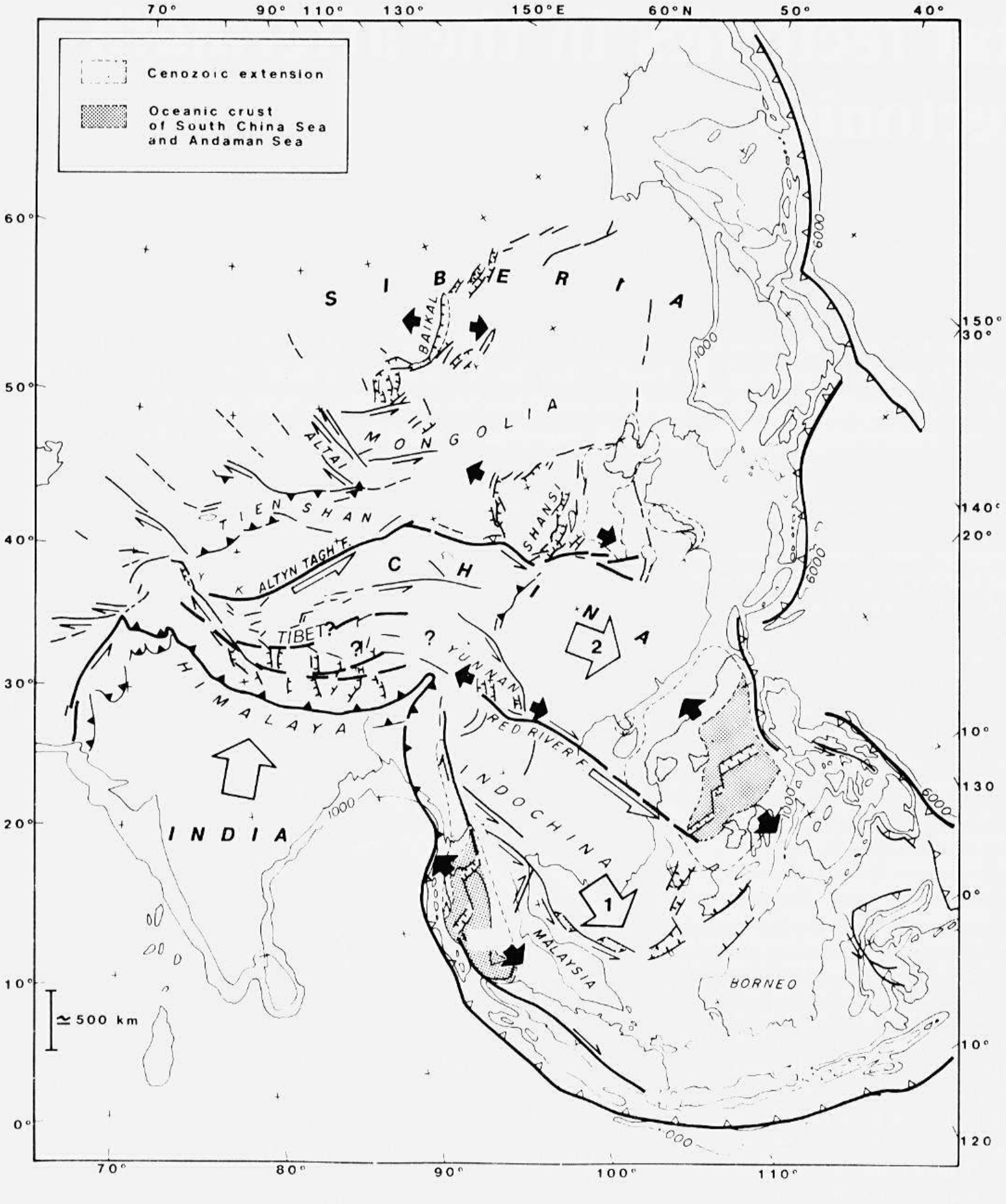

Fig. 2 Map of active faults and tectonics of Asia (after refs 8, 18). Northward penetration of India towards Siberia causes thrust faulting (lines with barbs) at the Himalaya and Tien Shan, and strike-slip faulting, primarily left-lateral, on faults emanating from Tibet and across Mongolia, as far north as Baikal. Extrusion of material sideways from in front of India's path may have displaced Sundaland and south-east China to the south-east[7,8,17,18]. In a thrust fault, one block of crust is pushed over another; on a strike-slip fault, blocks of crust slide past each other, moving horizontally.

and evolution and then the implications of that evolution for the geological history of western North America. Not only had oceanic crust been thrust beneath California, but that process had continued until much more recently than could be inferred from ages of rocks either along the coast or in the Sierra Nevada. Atwater also showed that slip on the San Andreas fault in southern California, initially considered to be the plate boundary between North America and the Pacific plate, has been much smaller than the displacement of those two plates since they came in contact. Most significantly, her work emphasized that an understanding of tectonic processes required much more than the analysis of rock exposed at the Earth's surface. Although Atwater's application of the tenets of plate tectonics to the oceanic realm in order to study the timing and scale of processes within continental North America forced a reassessment of the geological evidence for the tectonic evolution of California, it provided little insight into how the interior of western North America had deformed. She simply wrote this off as deformation of a "wide, soft zone"[4].

The interrelationship between rigid motion of oceanic lithosphere and more diffuse deformation in continental regions is particularly clear in New Zealand, the islands of which straddle a part of the boundary between the converging Pacific and Australian plates. The Pacific plate, which includes the southeastern part of New Zealand, rotates anticlockwise with respect to Australia, the Tasman Sea and the rest of New Zealand about a point ~1,000 km south of New Zealand. Pacific ocean floor is underthrust beneath the North Island of New Zealand, but deformation occurs throughout the crust of both islands (Fig. 1). Re-surveys of bench marks installed over much of the country in the last century revealed relative displacements of the bench marks that could be described only as diffuse, inhomogeneous strain. Nevertheless, an integration of that strain across the South Island yields a rate and direction of convergence of its margins that agrees remarkably well with the velocity calculated using the Vine–Matthews hypothesis and data from spreading centres in the South Pacific and the Indian Ocean[5,6]. The agreement is a tribute both to plate tectonics and to surveying techniques developed in the last century, but perhaps the most important result is the need to describe the deformation in New Zealand in terms of strain and vorticity instead of rigid-body motion. Moreover, the internal rotation (or vorticity) deduced from the re-surveys is corroborated by rotations, measured using palaeomagnetism, of small areas within New Zealand[6]. Thus, although one might conceivably be able to describe the deformation of New Zealand as relative displacements and rotations of numerous small blocks of crust, the number and the dimensions of such blocks would probably make such a description impractical.

The region where intracontinental deformation occurs on its grandest scale is Central Asia (Fig. 2). The penetration of the Indian subcontinent into the rest of Asia seems to be responsible for the world's highest mountains (the Himalaya, a consequence of the overthrusting of slivers of India's northern margin onto the rest of the subcontinent) and the largest and highest plateau (Tibet, a consequence of apparently distributed, crustal thickening), as well as the world's deepest lake (Baikal, filling a rift system caused by a splitting apart of the Asian landmass)[7,8]. Within the area of active deformation, ~3,000 km across, are large regions, such as the Tarim basin, that do not seem to be undergoing any deformation, and regions of comparable dimensions, such as the Tibetan Plateau, that are sliced by numerous faults, which isolate many tens of separate small blocks. The deformation that we see at the Earth's surface is neither homogeneous nor accurately described by the relative motion of a few rigid blocks.

Thus, whereas the success of plate tectonics on a global scale encouraged Earth scientists to analyse areas as large as Central Asia in terms of general processes, and whereas many of the same techniques that contributed data fundamental for establishing plate tectonics were also vital for obtaining a qualitative image of how deformation occurs in continents, we still lack a quantitative description of the kinematics of this deformation comparable with that used to describe plate motions in oceanic regions. The broad, diffuse deformation of the western United States, of New Zealand, or of Central Asia is much more complex than the rigid-body displacements of a small number of large plates, and finding a simple and accurate way to represent the deformation of continents remains a major task.

In short, plate tectonics is a poor approximation for the tectonics of many continental regions because vast portions of continental crust and mantle do not seem to move together as parts of the same, coherent rigid plates. Why should this be?

Differences between continents and oceans

Oceanic lithosphere has only a few kilometres of crust overlying its more dense mantle portion, but continental lithosphere commonly includes 35 km of relatively light crust. The difference is crucial: whereas the thin oceanic crust appears to be dense enough to plunge back into the mantle at subduction zones, the buoyancy of thick continental crust keeps it afloat. If continental lithosphere were strong enough to maintain its integrity at a subduction zone, the buoyant continental crust would not only resist being subducted, but the subducting plate would abruptly grind to a halt when the continental 'passenger' reached the trench.

The strength of continental lithosphere also contrasts with that of oceanic lithosphere. The strongest part of oceanic lithosphere seems to lie in the mantle, between 20 and 60 km depth, beneath a brittle upper part and above its increasingly ductile lower part, which grades downward into the asthenosphere (Fig. 3). In the same depth range where oceanic lithosphere is strongest, however, continental lithosphere consists of crust, not mantle. At temperatures typical of the lower crust (400–700 °C), the minerals comprising the crust appear to be much weaker than olivine, the strong mineral that comprises most of the upper mantle. Consequently, continental lithosphere could be much weaker than oceanic lithosphere. Oceanic lithosphere behaves as a virtually rigid plate because of its strong core, but, as the late C. Goetze noted in the mid-1970s, continental lithosphere

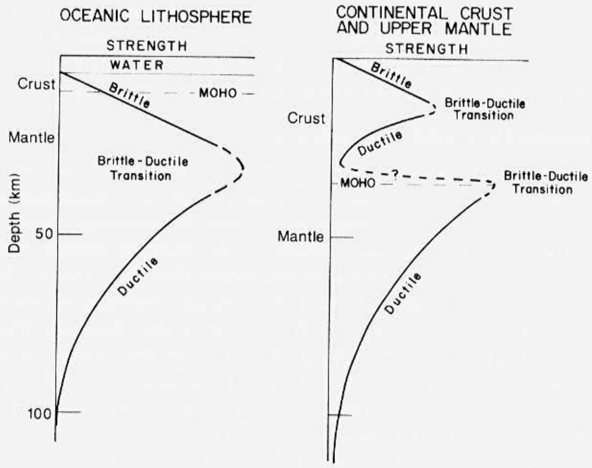

Fig. 3 Profiles of strength across oceanic and continental lithosphere, based on frictional strengths at shallow depths and ductile flow laws of minerals at greater depths[9]. Note the strong core of oceanic lithosphere at depths of 20–60 km and the probable weak, ductile region in the lower continental crust. Obviously, other phenomena, such as fluids, impurities in minerals or melting, can modify these curves, and given the uncertainties in the flow laws, the strengths cannot be usefully quantified. The existence of a low-strength zone in continents, however, seems very likely. The boundary between crust and mantle, known as the Moho, is defined chemically; mantle rocks are much richer in magnesium and iron than either continental or oceanic crust. By contrast, the boundary between lithosphere and asthenosphere is defined by a change in rheology; the lithosphere comprises the crust and uppermost upper mantle, which together make up a relatively strong shell overlying the weaker asthenosphere.

might consist of three layers: a brittle upper-crustal layer, a weak lower crust and a stronger uppermost mantle, which, nevertheless, would not be as strong as the strongest part of the oceanic lithosphere[9]. This jam-sandwich-like rheological profile (Fig. 3) is also suggested by the frequent occurrence of earthquakes (brittle deformation) in the upper crust, their nearly complete absence in the (presumably weak, ductile) lower crust, and their occasional presence in the underlying upper mantle[10].

These differences in rheology and buoyancy are probably responsible for mountain ranges, the most spectacular features of the continents, and for their virtual absence in oceans. (Mid-ocean ridges and seamounts owe their relief primarily to volcanic constructional processes and elevated temperatures in the underlying mantle, rather than to 'mountain building' as described below.) Most mountain ranges are built by crustal thickening. When masses of crust are pushed together, their buoyancy inhibits subduction into the mantle, and slices of upper crust, detached from the region below, are thrust atop one another. Crustal thickening does not seem to occur in oceanic regions (where, instead, subduction of the entire lithosphere allows sustained convergence of two lithospheric plates), but it dramatically affects the style and amount of deformation in continents.

For crustal thickening to occur, and for mountain ranges to be built, work must be done against gravity. Mountain ranges and high plateaux store gravitational potential energy, and this potential energy varies with the square of the mean height[11]. A rapidly increasing amount of work must be done to increase the mean height of an already elevated region, and therefore high plateaux created by crustal shortening and thickening should reach limiting elevations[11,12]. Continued convergence will then manifest itself by a growth in area of the range or plateau, as thrust faulting leads to crustal thickening on its margins. Hence, a high plateau can serve as a 'pressure gauge' for the average pressure applied to its margins, and it can transmit this pressure, much as a contained fluid transmits pressure laterally across itself[11,12]. Note the important difference between the sustained, virtually steady-state subduction of oceanic lithosphere into the asthenosphere and the transient growth of continental plateaux, by a process that has no analogue in plate tectonics.

Kinematics and dynamics

A key to the simplicity of plate tectonics is that the strength of lithospheric plates enables the analysis of their kinematics to be isolated and treated separately from the dynamic processes controlling plate motions; relative velocities of plates can be analysed without reference to the forces that give rise to them. Conversely, the much different rheological structure of the continents, which allows rapid deformation of the crust and mantle and facilitates the decoupling of one from the other, leads to an inseparable dependence of the kinematics of continental deformation on the dynamic processes controlling that deformation. In addition, because of crustal thickening, deformation varies in all three dimensions, and horizontal variations in mean elevation and in crustal thickness require horizontal gradients in some components of stress. Thus, not only do we lack a well-defined procedure for describing the kinematics of continental deformation, but we must also recognize the ultimate necessity of considering the dynamics in three dimensions.

We cannot do this yet, and even progress in understanding the dynamics of continental deformation in two dimensions is recent. In an important study, England and McKenzie[13] analysed the deformation of a thin viscous sheet as an analogue for the continental crust and upper mantle. By considering vertically averaged stresses and strains, they could study two-dimensional variations in the deformation of such a sheet in a gravity field. Calculations of stress, strain and vorticity fields and of distributions of equivalent elevations and crustal thicknesses as a function of time depend on only two non-dimensional numbers; one, the Argand number, measures the relative importances of gravity and of material strength in inhibiting deformation, and the other is a measure of the nonlinear dependence of strain rate on stress. With only two non-dimensional parameters, a full series of numerical experiments could examine the dependences of the calculated fields on them and on the boundary conditions. The most significant result of these numerical experiments is the weak dependence of the deformation field on the geometry of the boundary conditions; deformation of a sheet indented by a rigid die decays rapidly at distances from the die comparable to its width[14,15]. By analogy with the Earth, England and Houseman[16] inferred that the penetration of India (a rigid die) into Eurasia (a thin viscous sheet) during the last 40–50 Myr caused crustal thickening in the area ~1,000 km north and north-east of the Himalaya, but that south-east China has not been extruded very far to the east, out of India's northward path (see Fig. 2).

In another approach, laboratory studies by Peltzer and Tapponnier[17,18] of the indentation of plasticine, unconfined on its top and bottom and on two sides but unconfined on one lateral edge, addressed the rapid eastward extrusion of parts of China and Indochina, which had been inferred from the prevalence of strike-slip faulting in Asia[7,8]. They found a large lateral extrusion of coherent, if deformed, pieces of plasticine[8,17,18]. The small expulsion calculated for the thin viscous sheet and the large expulsion of plasticine are largely a consequence of thickening of the indented viscous sheet but not of the plasticine, but the difference between the rheological properties (or constitutive laws) of the two media also contributes to the differing observed styles of deformation.

Role of crustal blocks

This difference in assumed constitutive laws underscores one of the principal questions asked about the dynamics of continental deformation. The thin viscous sheet deforms homogeneously, but plasticine undergoes strain softening, with the localization of shear in narrow zones, analogous to faults and fault zones in the Earth. The concentration of deformation in shear zones allows intervening regions to be displaced coherently, as blocks (or pieces of plasticine), but homogeneous deformation does not include blocks that translate with respect to one another. Clearly, the crust that we see and walk on is not flowing or deforming with large strains, and the issue, therefore, is not whether or not there are blocks of crust at the surface, but what role they play.

In one extreme, the upper crust is imagined to consist of decoupled blocks of rock whose composite relative displacements serve primarily as passive markers that reflect the average deformation of ductile material below[6,13–16,19,20]. Hence, the velocity or displacement fields of points on the surface of the Earth would yield a rough image of ductile deformation of (possibly stronger) material at depth. In the other extreme, the outer part of the Earth deforms by localization of strain into strong, if deformable, blocks, whose relative motions are controlled by stresses on their edges[8,17,18,21]. Effectively, such blocks float on a weak substratum, the asthenosphere, in a manner similar to ice fragments floating on water. In the former case, slow motion of the underlying fluid carries the blocks, and in the latter, boundary conditions on the lateral margins of the deforming continent form blocks that rub on one another and move above a passive fluid. Because studies of the viscous sheet ignore localized weaknesses, and because the strongest part of continental crust and upper mantle is thought to lie just below the Moho[9], the thin viscous sheet approximates the former more than the latter analogue. A medium, like plasticine, cut by prominent shear zones is more analogous with the latter.

Clearly, these two oversimplified images of crustal blocks driven by flow beneath them or by stresses on their margins represent end-members which embrace the likely relationship between the observable kinematics of continental deformation and the dynamic processes that we try to infer. Consideration of them as extremes, however, allows us to pose very simply a number of specific problems that should reduce our ignorance of how the dynamics and kinematics are related.

For example, suppose the thin viscous sheet were an appropriate analogue for continental tectonics, and specifically suppose that England and Houseman[16] were correct in suggesting that most of the extrusion of Tibet due to India's penetration into Asia has been absorbed by crustal thickening on the eastern edge of Tibet and not by southeastward expulsion of south-east China. Then, slip on the major strike-slip faults on the margins of Tibet and south-east of it should be small, less than 100 km. Alternatively, suppose that Peltzer and Tapponnier[8,17] were correct that the deformation is controlled by the localization of slip on major strike-slip faults and shear zones, and, in particular, that expulsion of blocks comprising Indo-China and South China has been large. Then, the cumulative slip on several faults should also be large, more than 100–300 km on each fault, summing to at least 1,000 km of cumulative displacement. Burchfiel's[22] and Zhang Peizhen's[23] measurement of only 10–15 km of Cenozoic slip on the Haiyuan fault, the only significant strike-slip fault strand in China with a well constrained total offset, does not imply a large explulsion, but clearly the displacement on only one fault is insufficient to prove this. In any case, classical geological techniques for mapping offsets on major faults hold the key to resolving this question, and quantifying the amounts of deformation or displacement looms as one of geology's great challenges for the future.

Although geological methods are honed for mapping both displacements of coherent rock units at faults and large internal strain of individual terrains, rotations of small regions or blocks about vertical axes can easily go undetected; yet such rotations may hold a key for understanding the relation between surficial and deeper deformation. Beginning largely from the work of Beck[24] and Luyendyk[25], a wealth of palaeomagnetic data now show large, late Cenozoic rotations of isolated crustal blocks[24–28], and such rotations comprise an important part of

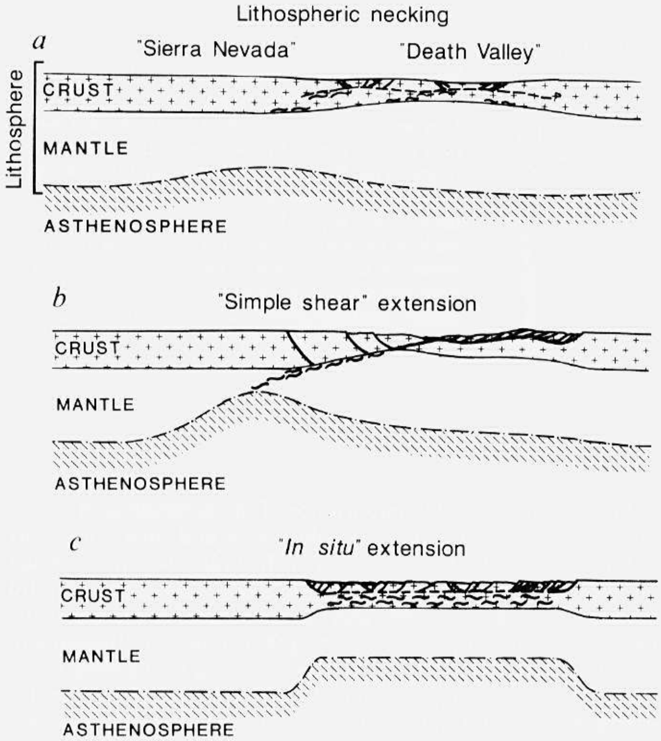

Fig. 4 Idealized cross-sections across the Basin and Range Province and the Sierra Nevada in California (after ref. 38), illustrating the likely possibility (*a*, *b*) that crustal extension in Death Valley may be detached from the mantle directly underlying it, and the extension of the mantle lithosphere may underlie the Sierra Nevada. The youthful uplift of the Sierra Nevada would be a consequence of the intrusion of hotter material where the underlying mantle lithosphere had thinned. Cross-section *c* shows the older, apparently incorrect idea that the crust and its directly underlying mantle would be stretched and thinned by the same amount.

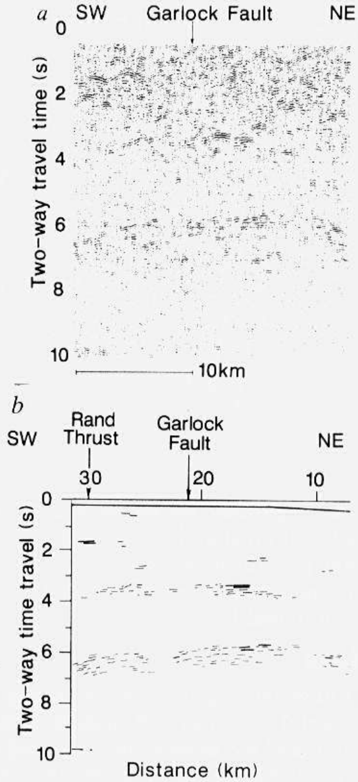

Fig. 5 Seismic reflection profile across the Garlock Fault in California (*a*) and simple interpretation of it (*b*). The Garlock Fault is a major strike-slip fault, with many tens of kilometres of displacement[39], but the fault does not seem to penetrate below ~10–12 km depth (two-way travel time of 3.5 s). Instead, nearly flat reflections pass beneath the surface trace of the fault and could mark detachment of the upper from the lower crust. (From refs 40 and 41.)

the deformation in broad shear zones[6,29,30]. If such blocks were dragged along (like ball bearings) by horizontal displacements at the edges of the shear zone, the rate of rotation would be simply the rate of slip of the margins of the shear zone divided by the radius of the block. If, instead, the block floated on a viscous medium undergoing distributed homogeneous shear, the rate of rotation would be half as large (because only the vorticity, and not the pure shear, of the viscous material would contribute to the rotation[19,20]). Thus, rates and amounts of rotations of blocks driven by stresses on their edges or by flow beneath them differ from one another. To distinguish between these two extremes requires accurate measurements both of displacement of opposite sides of the shear zone and the rotation of the intervening material, as well as a belief that the areas that rotate are equidimensional[31]. What measurements exist (see, for example, refs 6, 20, 28, 32) favour blocks floating on a viscous substratum undergoing shear, but whether this process is common or not is not known.

Detachment of crustal fragments

If blocks were driven by stresses on their margins, then boundaries between them would probably pass directly from the Earth's surface through the stronger mantle beneath them. If they were driven by flow beneath them, then the boundaries between such blocks might be truncated by flat shear zones in the crust. One of the exciting discoveries of the 1980s is the widespread detachment of small blocks of the upper crust from the lower crust and uppermost mantle. Following the work of Anderson[33] and Armstrong[34], field geologists working in the western United States have mapped abundant evidence for large horizontal translations of highly fractured blocks of upper crust over ductilely sheared, strongly metamorphosed (obviously hotter) lower-crustal rock. Whether the contacts between these brittly deformed upper-crustal rocks and ductilely deformed lower-crustal rocks mark single faults[35] or a more complicated juxtaposition of two entirely different styles of deformation[36,37] is unresolved, but it seems likely that the stretching of the crust in one area might continue into the mantle elsewhere[38] (Fig. 4).

A lack of coherence between the upper continental crust and its uppermost mantle is also virtually certain along some strike-slip faults. Where strike-slip faults mark the boundaries between oceanic plates (where they are known as transform faults), the rigidity of the plates requires that the rate and amount of displacement be the same at all points along the fault. In continents, however, the deformation of material on both sides of the fault can be so large that the amount of strike-slip displacement varies markedly along the fault[39]. If this deformation involved the entire crust, there would be marked differences in crustal thickness across the fault. Seismic imaging of some intracontinental strike-slip faults, however, suggests that these faults do not penetrate deeper than 8–10 km[40,41] (Fig. 5). Even portions of the San Andreas fault in California may not continue through the upper mantle directly beneath it[42]. Although the surface trace is so overwhelmingly clear that pilots flying between Los Angeles and San Francisco can navigate by it, part of the mantle lithosphere beneath the Transverse Ranges seems to detach and plunge into the asthenosphere instead of being displaced coherently with the crustal blocks on either side of the fault[43]. Elsewhere, the San Andreas fault seems to pass directly into the mantle as a narrower vertical zone[41]. Thus,

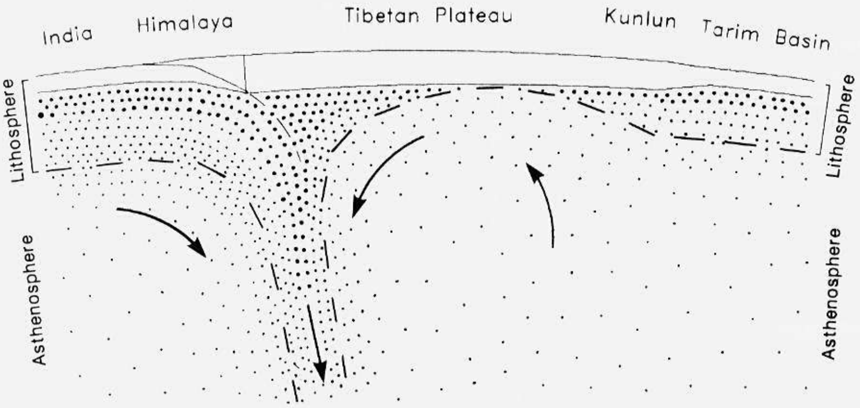

Fig. 6 Inferred dynamics of the mantle beneath Tibet and the Himalaya (from ref. 47). Low seismic-wave velocities in northern Tibet may mark a zone of upwelling in the asthenosphere. The flexure of the Indian lithosphere beneath the Himalaya seems to require a bending moment applied to it, and convective downwelling beneath southern Tibet could be the source of that moment[46]. Large, closely spaced dots indicate colder temperatures, grading into hotter asthenosphere with smaller, widely spaced dots.

small crustal blocks separated by minor faults probably are detached from the lower crust and upper mantle beneath them, but major faults may pass through the strong uppermost mantle as localized shear zones which separate plates, or small fragments, of intact continental lithosphere.

Future directions

Ultimately, a full understanding of both the kinematics and the dynamics of continental tectonics will include aspects of both a continuous medium and the movements of strong blocks. The question is how small blocks must be before their relative motions are more accurately described as parts of a continuum. At one extreme, a crustal block the size of Siberia is a plate, as is the Tarim Basin, which at $500 \times 1,000$ km behaves as a large rigid block surrounded by zones of diffuse deformation. The existence of only a few strike-slip faults in Asia along which rates of slip exceed 10 mm yr^{-1} also implies a concentration of deformation in a few zones separating extensive, less rapidly deforming regions, 200–500 km across. Near the other extreme, each range in the Basin and Range Province of the western United States, ~10–20 km across, seems to move with respect to the others. A description of the velocity field for the Basin and Range Province will surely require many more parameters than the four elements of a two-dimensional velocity gradient tensor. Moreover, if these small blocks are indeed detached from the underlying lower crust and mantle, the deformation of this deeper region could be continuous, and perhaps relatively homogeneous. Hence some aspects of continental deformation can be described well by relative motions of relatively rigid blocks, which behave as small lithospheric plates, but others seem to require the consideration of a continuous medium. Like all complex questions, this becomes one not only of kind, but also of degree.

Thus, one focus of future studies of continental tectonics will surely be on measuring the kinematics of deformation. Traditional geological field mapping applied to quantifying deformation will continue to flourish, and its burgeoning offspring, the study of Quaternary faulting, will surely yield the most precise average velocities of slip on faults. The task of determining rotations, however, will fall largely on the palaeomagnetists. Because deformation in the outer part of the Earth occurs largely by slip during earthquakes, seismology will always hold a seat in the presidium of continental tectonics. But the most rapid advances in our understanding of the kinematics of continental deformation may come with the imminent renaissance in surveying, brought about by the implementation of Global Positioning System satellites, which will make repeated measurements of relative positions of points tens to hundreds, and even thousands of kilometres apart easy, and eventually inexpensive. One can foresee a time when the deformation of continents will be monitored continuously.

I have focused on the importance of the kinematics of deformation as crucial for understanding why and how continents deform as they do, but obviously such studies will not provide all the answers. Two examples illustrate this. First, careful laboratory work led Goetze and his colleagues[9] to recognize the likely difference in profiles of strength through continental and oceanic lithosphere (Fig. 3). Second, imaging of the deep structure of continental regions, largely with seismic waves (see, for example, refs. 40–43), is vital for understanding how and where the upper crust detaches from the underlying material—for example, on narrow flat faults or a broad shear zone.

Back to dynamics

The tectonics of continents has found plate tectonics an inadequate paradigm. Nonetheless, it shares one unsatisfactorily answered and profoundly important question: What is the driving mechanism?

Plate motions are the kinematic manifestations of large-scale convection in the Earth; oceanic lithosphere is the cold boundary layer of that convection. It is a simple matter to ascribe the driving force to gravity causing plates to slide downhill from mid-ocean ridges and pulling them into the asthenosphere at subduction zones, but it is a rare fluid dynamicist who would contend that these processes are understood or that such a description constitutes a physically rigorous, quantitative description of the driving mechanism. Similarly it seems probable that mountain belts, at which crustal shortening is a major process, overlie zones of convergence and downwelling convective flow in the mantle[43–47] (Fig. 6). What is not known about the structure of the upper mantle beneath major mountain belts, however, dwarfs what is thought about such regions. A growth in our knowledge of lateral variations in the structure of the upper mantle is vital for an understanding of convection in the mantle and the forces driving plate motions, which will, in turn, bring a refined understanding of the forces that drive mountain building and continental tectonics.

The contribution of plate tectonics to continental tectonics has been enormous. The demonstration of large horizontal displacements of terrains over the surface of the Earth required not only the consideration of large-scale processes on the continents but also a very different understanding of the geological record than had previously been widely held. The most profound effect of plate tectonics on the Earth sciences, however, was to expedite its transition from a largely qualitative, data-cataloguing natural science to a quantitative physical science. One obvious indication of this has been the enormous growth of geophysical methods for solving geological problems—those same methods that had demonstrated plate tectonics. A second is the growing emphasis on problem solving. Most geologists now map regions, not to put colour on previously blank maps, but rather to solve specific problems that require better geological maps than currently exist. Similarly, marine geologists no longer ply the seas in a spiderweb of ship tracks, but now choose specific areas for detailed study of particular phenomena. Perhaps the most rewarding and the most useful result of plate tectonics was the recognition that processes that had seemed too difficult to study except qualitatively could, in fact, be analysed quantitatively with simple physical concepts replete with mathematics and meaningful uncertainties. Although continental tectonics has not had a revolution like plate tectonics, it has nevertheless progressed enormously in the past 20 years.

I am grateful to B. C. Burchfiel, P. C. England, J. A. Jackson, R. McCaffrey, D. P. McKenzie, S. Neustadtl and P. Tapponnier for critically reading earlier drafts of this manuscript and D. L. Frank for numerous retypings of it. R. Allmendinger, C. H. Jones, S. Kaufman, P. Tapponnier and R. I. Walcott kindly supplied original figures. I thank NASA and NSF for continued support.

1. Vine, F. J. & Matthews, D. H. *Nature* **199**, 947–949 (1963).
2. Hamilton, W. *Geol. Soc. Am. Bull.* **80**, 2409–2430 (1969).
3. Dewey, J. F. & Bird, J. M. *J. geophys. Res.* **75**, 2625–2647 (1970).
4. Atwater, T. *Geol. Soc. Am. Bull.* **81**, 3513–3536 (1970).
5. Bibby, H. M. *Geophys. J. R. astr. Soc.* **66**, 513–533 (1981).
6. Walcott, R. I. *Geophys. J. R. astr. Soc.* **79**, 613–633 (1984).
7. Molnar, P. & Tapponnier, P. *Science* **189**, 419–426 (1975).
8. Tapponnier, P., Peltzer, G. & Armijo, R. in *Collision Tectonics* (eds Coward, M. P. & Ries, A. C.) 115–157 (Geological Society, London, 1986).
9. Brace, W. F. & Kohlstedt, D. L. *J. geophys. Res.* **85**, 6248–6252 (1980).
10. Chen, W.-P. & Molnar, P. *J. geophys. Res.* **88**, 4183–4214 (1983).
11. Molnar, P. & Lyon-Caen, H. *Spec. Pap. geol. Soc. Am.* **218**, 179–207 (1988).
12. Molnar, P. & Tapponnier, P. *J. geophys. Res.* **83**, 5361–5375 (1978).
13. England, P. C. & McKenzie, D. P. *Geophys. J. R. astr. Soc.* **70**, 295–321 (1982).
14. Houseman, G. A. & England, P. C. *J. geophys. Res.* **91**, 3651–3663 (1986).
15. Vilotte, J. P., Madariaga, R., Dagnières, M. & Zienkiewicz, O. *Geophys. J. R. astr. Soc.* **84**, 279–310 (1986).
16. England, P. C. & Houseman, G. A. *J. geophys. Res.* **91**, 3664–3674 (1986).
17. Peltzer, G. & Tapponnier, P. *J. geophys. Res.* (in the press).
18. Tapponnier, P., Peltzer, G., Le Dain, A. Y., Armijo, R. & Cobbold, P. *Geology* **10**, 611–616 (1982).
19. McKenzie, D. & Jackson, J. *Earth planet. Sci. Lett.* **65**, 182–202 (1983).
20. McKenzie, D. & Jackson, J. *J. geol. Soc. Lond.* **143**, 349–353 (1986).
21. Nur, A., Ron, H. & Scotti, O. *Geology* **14**, 746–749 (1986).
22. Burchfiel, B. C. *et al. Geol. Soc. Am. Bull.* (submitted).
23. Zhang Peizhen *et al. Geol. Soc. Am. Bull.* (submitted).
24. Beck, M. E. *Am. J. Sci.* **276**, 694–712 (1976).
25. Luyendyk, B. P., Kamerling, M. J., Terres, R. R. & Hornafius, J. S. *J. geophys. Res.* **90**, 12454–12466 (1985).
26. Ron, H., Freund, R., Garfunkel, Z. & Nur, A. *J. geophys. Res.* **89**, 6256–6270 (1984).
27. Kissel, C. & Laj, C. *Tectonophysics* **146**, 183–201 (1988).
28. Nelson, M. R. & Jones, C. H. *Tectonics* **6**, 13–33 (1987).
29. Nicholson, C., Seeber, L., Williams, P. & Sykes, L. R. *Tectonics* **5**, 629–648 (1986).
30. Sonder, L. J., England, P. C. & Houseman, G. A. *J. geophys. Res.* **91**, 4797–4810 (1986).
31. Lamb, S. H. *Earth planet. Sci. Lett.* **84**, 75–86 (1987).
32. Jackson, J. *Nature* **334**, 195–196 (1988).
33. Anderson, R. E. *Geol. Soc. Am. Bull.* **82**, 43–58 (1971).
34. Armstrong, R. L. *Geol. Soc. Am. Bull.* **83**, 1729–1754 (1972).
35. Wernicke, B. *Nature* **291**, 645–648 (1981).
36. Miller, E. L., Gans, P. B. & Garing, J. *Tectonics* **2**, 239–264 (1983).
37. Davis, G. A. & Lister, G. S. *Spec. Pap. geol. Soc. Am.* **218**, 133–159 (1988).
38. Jones, C. H. *Tectonics* **6**, 449–473 (1987).
39. Davis, G. A. & Burchfiel, B. C. *Geol. Soc. Am. Bull.* **84**, 1407–1422 (1973).
40. Cheadle, M. J. *et al. Tectonics* **5**, 293–320 (1986).
41. Lemiszki, P. J. & Brown, L. D. *Geol. Soc. am. Bull.* **100**, 665–676 (1988).
42. Hadley, D. & Kanamori, H. *Geol. Soc. Am. Bull.* **88**, 1469–1478 (1977).
43. Humphreys, E., Clayton, R. W. & Hager, B. H. *Geophys. Res. Lett.* **11**, 625–627 (1984).
44. Panza, G. F. & Müller, St. *Mem. Sci. Geol.* **33**, 43–50 (1979.
45. Houseman, G. A., McKenzie, D. P. & Molnar, P. *J. geophys. Res.* **86**, 6115–6132 (1981).
46. Lyon-Caen, H. & Molnar, P. *J. geophys. Res.* **88**, 8171–8191 (1983).
47. Molnar, P. *Phil. Trans. R. Soc.* **A326**, 33–88 (1988).