Stress in the Earth's Lithosphere

Mark D. Zoback
Stanford University

Mary Lou Zoback
U.S. Geological Survey

GLOSSARY

Critically stressed crust Stress magnitudes in the brittle crust are equal to the crust's frictional strength.

***In situ* stress** Forces in the lithosphere.

Sources of stress The mechanisms responsible for stress in the lithosphere.

Stress map Map showing the orientation and relative magnitude of horizontal principal stress orientations.

THE STATE OF STRESS in the lithosphere is the result of the forces acting upon and within it. Knowledge of the magnitude and distribution of these forces can be combined with mechanical, thermal, and rheological constraints to examine a broad range of lithospheric deformational processes. For example, such knowledge contributes to a better understanding of the processes that both drive and inhibit lithospheric plate motions, as well as the forces responsible for the occurrence of crustal earthquakes—both along plate boundaries and in intraplate regions.

While our topic is the state of stress in the earth's lithosphere, the comments below come primarily from the perspective of the state of stress in the brittle upper crust. As defined by the depth of shallow earthquakes, the brittle crust extends to a depth of ~15–20 km at most continental locations around the world. We adopt this perspective because nearly all the data available on lithospheric stress comes from the upper crust of continents. Furthermore, in the sections that follow, we argue that, to first order, the state of stress in the brittle crust results from relatively large-scale lithospheric processes, so that knowledge of crustal stress can be used to constrain the forces involved in these processes.

I. BASIC DEFINITIONS

Stress is a tensor which describes the density of forces acting on all surfaces passing through a point. In terms of continuum mechanics, the stresses acting on a homogeneous, isotropic body at depth are describable as a second rank tensor, with nine components (Fig. 1, left).

$$\tilde{S} = \begin{vmatrix} S_{11} & S_{12} & S_{13} \\ S_{21} & S_{22} & S_{23} \\ S_{31} & S_{32} & S_{33} \end{vmatrix} \tag{1}$$

The subscripts of the individual stress components refer to the direction that a given force is acting and the face of the unit cube upon which the stress component acts. Thus, in simplest terms, any given stress component represents a force acting in a specific direction on a unit area of given orientation. As illustrated in the left side of Fig. 1, a stress tensor can be defined in terms of any arbitrary reference frame. Because of equilibrium conditions,

$$\begin{aligned} S_{12} &= S_{21} \\ S_{13} &= S_{31} \\ S_{23} &= S_{32} \end{aligned} \tag{2}$$

so that the order of the subscripts is unimportant. In general, to fully describe the state of stress at depth, one must estimate six stress magnitudes or three stress magnitudes and the three angles that define the orientation of the stress coordinate system with respect to a reference coordinate system (such as geographic coordinates, for example).

We utilize the convention that compressive stress is positive because *in situ* stresses at depths greater than a few tens of meters in the earth are *always* compressive. Tensile stresses do not exist at depth in the earth for two fundamental reasons. First, because the tensile strength of rock is generally quite low, significant tensile stress cannot be supported in the earth. Second, because there is always a fluid phase saturating the pore space in rock at depth (except at depths shallower than the water table), the pore pressure resulting from this fluid phase would cause the rock to hydraulically fracture should the least compressive stress reach a value even as low as the pore pressure.

Once a stress tensor is known, it is possible to evaluate stresses in any coordinate system via tensor transformation. To accomplish this transformation, we need to specify the direction cosines (a_{ij}, as illustrated in Fig. 1, center)

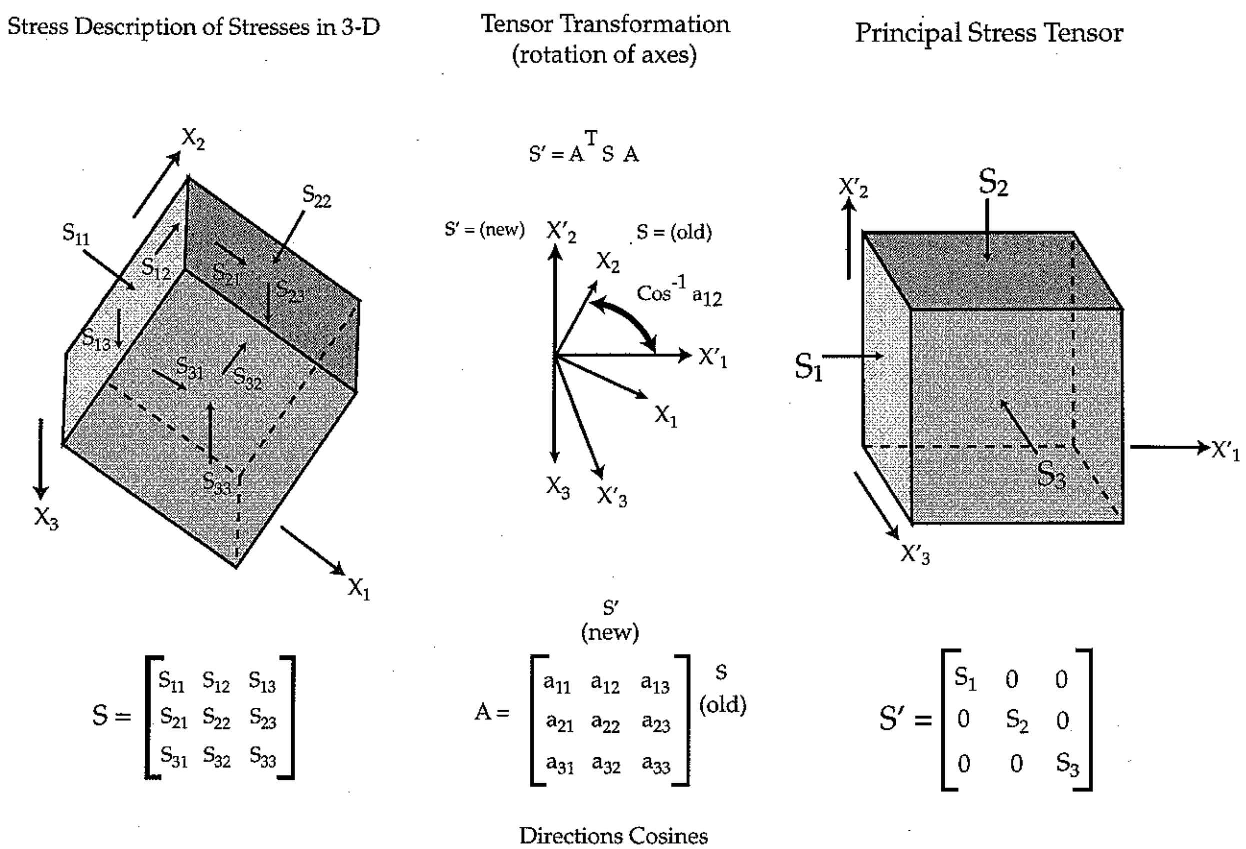

FIGURE 1 Definition of stress tensor in an arbitrary Cartesian coordinate system (left) and the rotation of stress coordinate systems through tensor transformation (middle) and principal stresses (right) as defined in a coordinate system in which shear stresses vanish.

that describe the rotation of the coordinate axes between the old and new coordinate systems. Mathematically, the equation which accomplishes this is

$$\bar{S}' = \bar{A}^T \bar{S} \bar{A}, \tag{3}$$

where

$$\bar{A} = \begin{vmatrix} a_{11} & a_{12} & a_{13} \\ a_{21} & a_{22} & a_{23} \\ a_{31} & a_{32} & a_{33} \end{vmatrix}.$$

The ability to transform coordinate systems is of interest here because we can choose to generally describe the state of stress in terms of the principal coordinate system. The principal coordinate system is the one in which shear stresses vanish and only three principal stresses, $S_1 \geq S_2 \geq S_3$, fully describe the stress field (as illustrated in right side of Fig. 1). Thus, we have diagonalized the stress tensor such that the principal stresses correspond to the eigenvalues of the stress tensor and the principal stress directions correspond to its eigenvectors.

$$\bar{S}' = \begin{vmatrix} S_1 & 0 & 0 \\ 0 & S_2 & 0 \\ 0 & 0 & S_3 \end{vmatrix} \tag{4}$$

The reason this concept is so important is that as the earth's surface is in contact with a fluid (either air or water) and cannot support shear tractions, it is a principal stress plane. Thus, one principal stress is generally expected to be normal to the earth's surface with the other two principal stresses acting in an approximately horizontal plane. While it is clear that this must be true very close to the earth's surface, compilation of earthquake focal mechanism data and other stress indicators (described below) suggest that it is also generally true to the depth of the brittle-ductile transition in the upper crust (Zoback and Zoback, 1989; Zoback, 1992). Assuming this is the case, we must define only four parameters to describe the state of stress at depth; one stress orientation (usually taken to be the azimuth of the maximum horizontal compression, S_{Hmax}) and three principal stress magnitudes: S_{v}, the vertical stress, corresponding the weight of the overburden; S_{Hmax}, the maximum principal horizontal stress; and S_{hmin}, the minimum principal horizontal stress. This obviously helps make stress determination in the crust a tractable problem.

In applying these concepts to the earth's crust, it is helpful to consider the magnitudes of the greatest, intermediate, and maximum principal stress at depth (S_1, S_2, and S_3) in terms of S_{v}, S_{Hmax}, and S_{hmin} in the manner originally proposed by E. M. Anderson. (1951). This is illustrated in Fig. 2. There are a number of simple but fundamental points about these seemingly straightforward relations.

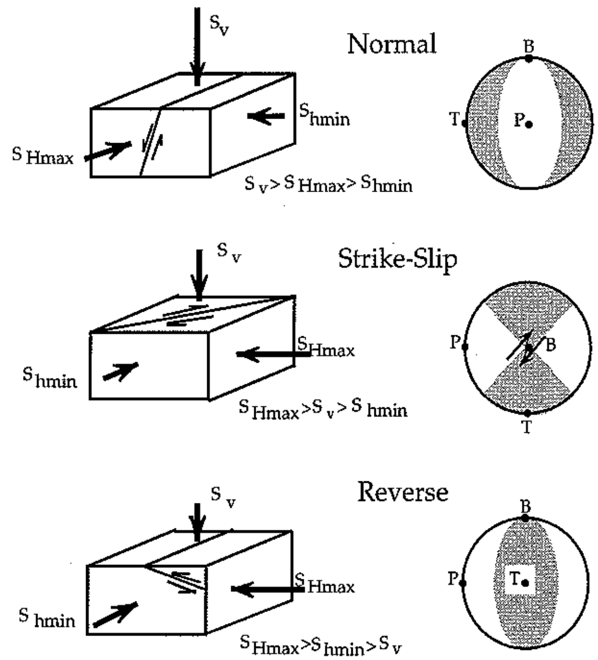

FIGURE 2 E. M. Anderson's classification scheme for relative stress magnitudes in normal, strike-slip, and reverse faulting regions. Corresponding focal plane mechanisms are shown to the right.

First, the two horizontal principal stresses in the earth, S_{Hmax} and S_{hmin}, can be described relative to the vertical principal stress, S_{v}, whose magnitude corresponds to the overburden. Mathematically, S_{v} is equivalent to integration of density from the surface to the depth of interest, z. In other words,

$$S_{\mathrm{v}} = \int_0^z \rho(z) g \, dz \approx \bar{\rho} g z, \tag{5}$$

where $\rho(z)$ is the density as a function of depth, g is the gravitational acceleration, and $\bar{\rho}$ is the mean overburden density. It is, of course, necessary to add atmospheric pressure and the pressure resulting from the weight of water at the earth's surface, as appropriate.

Second, the horizontal principal stresses are almost never equal and may be less than or greater than the vertical stress. In fact, the relative magnitudes of the principal stresses can be simply related to the faulting style currently active in a region. As illustrated in Fig. 2, (and following Anderson, 1951), characterizing a region by normal, strike-slip, or reverse faulting is equivalent to defining the horizontal principal stress magnitudes with respect to the vertical stress. When the vertical stress dominates in extensional deformational regions ($S_1 = S_{\mathrm{v}}$), gravity drives normal faulting. Conversely, when both horizontal stresses exceed the vertical stress ($S_3 = S_{\mathrm{v}}$), compressional deformation (shortening) is accomodated through reverse faulting. Strike-slip faulting represents an intermediate stress state ($S_2 = S_v$), where the maximum horizontal stress is greater than the vertical stress and the minimum horizontal stress is less than the vertical stress ($S_{Hmax} \geq S_v \geq S_{hmin}$).

The concept of effective stress is used to incorporate the influence of pore pressure at depth where a component of effective stress σ_{ij} is related to the total stress S_{ij} via

$$\sigma_{ij} = (S_{ij} - \delta_{ij} P_p), \tag{6}$$

where δ_{ij} is the Kronecker delta and P_p is the pore pressure.

Laboratory studies of the frictional strength of faulted rock carried out over the past several decades indicate that the Coulomb criterion describes the frictional strength of faults. That is, fault slippage will occur when

$$\tau = S_o + \mu\sigma_n, \tag{7}$$

where τ is the shear stress acting on the fault, S_o is the fault cohesion, μ is the coefficient of friction on the fault, and σ_n is the effective normal stress acting on the fault plane. The maximum shear stress is given by $1/2(S_1 - S_3)$.

Using the concept of effective stress at depth, we can extend Anderson's faulting theory to predict stress magnitudes at depth through utilization of simplified two-dimensional Mohr-Coulomb failure theory. Two-dimensional faulting theory assumes that failure is only a function of the difference between the last and greatest principal effective stresses σ_1 and σ_3 as given by Jaeger and Cook (1971)

$$\sigma_1/\sigma_3 = (S_1 - P_p)/(S_3 - P_p) = ((\mu^2 + 1)^{1/2} + \mu)^2. \tag{8}$$

Thus, a third point about crustal stress that can be derived from Andersonian faulting theory is that the magnitudes of the three principal stresses at any depth are limited by the strength of the earth's crust at depth. In the case of normal faulting, stress magnitudes are controlled by σ_v, and σ_{hmin}, which correspond to σ_1 and σ_3, respectively σ_{Hmax}, which corresponds to σ_2 is intermediate in value between σ_v and σ_{hmin}, does not influence faulting. Coulomb failure theory indicates that frictional sliding occurs when the ratio shear stress to effective normal stress on preexisting fault planes is equal to the coefficient of friction. As the coefficient of friction is relatively well defined for most rocks and ranges between ~0.6 and 1.0 (Byerlee, 1978), Eq. (8) demonstrates that frictional sliding will occur when $\sigma_1/\sigma_3 \sim 3.1$–5.8. For the case of hydrostatic pore pressure and commonly observed friction coefficients of 0.6 (e.g., Townend and Zoback, 2000), in extensional areas $S_{hmin} \sim 0.6\ S_v$, in reverse faulting areas $S_{Hmax} \sim 2.3\ S_v$, and in strike-slip faulting areas (when $S_v \sim 1/2(S_{Hmax} + S_{hmin})$)$S_{Hmax} \sim 2.2\ S_{hmin}$. As discussed below, these simple relationships have been confirmed by *in situ* stress measurements to depths of almost 8 km at a number of sites in intraplate areas.

II. INDICATORS OF CONTEMPORARY STRESS

Information on the state of stress in the lithosphere comes from a variety of sources—earthquake focal plane mechanisms, young geologic data on fault slip and volcanic alignments, *in situ* stress measurements, stress-induced wellbore breakouts, and drilling-induced tensile fractures. A stress measurement quality criterion for different types of stress indicators was developed by Zoback and Zoback (1989, 1991). This quality criterion was subsequently utilized in the International Lithosphere Program's World Stress Map Project, a large collaborative effort of data compilation and analyses by over 40 scientists from 30 different countries (Zoback, 1992). A special issue of the *Journal of Geophysical Research* (v. 97, pp. 11,703–12,014, 1992) summarized the overall results of this project and presented the individual contributions of many of these investigators in various regions of the world. Today, the World Stress Map (WSM) database has more than 9100 entries and is maintained at the Heidelberg Academy of Sciences and Humanities (Mueller *et al.*, 1997; http://www-wsm.physik.uni-karlsruhe.de/).

Zoback and Zoback (1991) discussed the rationale for the WSM quality criterion used in the WSM project in detail. The success of the WSM project demonstrates that with careful attention to data quality, coherent stress patterns over large regions of the earth can be mapped with reliability and interpreted with respect to large-scale lithospheric processes.

A. Earthquake Focal Mechanisms

While earthquake focal plane mechanisms are the most ubiquitous indicator of stress in the lithosphere, the determination of principal stress orientations and relative magnitudes from these mechanisms must be done with appreciable caution. The pattern of seismic radiation from the focus of an earthquake permits construction of earthquake focal mechanisms (right column of Fig. 2). Perhaps the most simple and straightforward information about *in situ* stress that is obtainable from focal mechanisms and *in situ* stress is that the type of earthquake (i.e., normal, strike-slip, or reverse faulting) defines the relative magnitudes of S_{Hmax}, S_{hmin}, and S_v. In addition, the orientation of the fault plane and the auxiliary plane (which bind the compressional and extensional quadrants of the focal plane mechanism) defines the orientation of the P (compressional), B (intermediate), and T (extensional) axes. These axes are sometimes incorrectly assumed to be the same as the orientation of S_1, S_2, and S_3.

For cases in which laboratory-measured coefficients of fault friction of ~0.6–1.0 are applicable to the crust, there is a nontrivial error of ~15–20° if one uses the P, B, and T

axes as approximations of average principal stress orientations, especially if the orientation of the fault plane upon which the earthquake occurred is known. If friction is negligible on the faults in question (but higher in surrounding rocks), there can be a considerable difference between the P, B, and T axes and principal stress directions. An earthquake focal plane mechanism always has the P and T axes at 45° to the fault plane and the B axis in the plane of the fault. With a frictionless fault the seismic radiation pattern is controlled by the orientation of the fault plane and not the *in situ* stress field (McKenzie, 1969). One result of this is that just knowing the orientation of the P axis of earthquakes along weak, plate-bounding, strike-slip faults (like the San Andreas) does not allow one to define principal stress orientations from the focal plane mechanisms of the strike-slip earthquakes occurring on the fault (Zoback *et al.*, 1987). For this reason, it is common practice to omit plate-boundary earthquakes from regional stress compilations (see below).

Principal stress directions can be determined directly from a group of earthquake focal mechanisms (or set of fault striae measurements) through use of inversion techniques that are based on the slip kinematics and the assumption that fault slip will always occur in the direction of maximum resolved shear stress on a fault plane. Such inversions yield four parameters, the orientation of the three principal stress and the relative magnitude of the intermediate principal stress with respect to the maximum and minimum principal stress.

The analysis of seismic waves radiating from an earthquake also can be used to estimate the magnitude of stress released in an earthquake (stress drop), although not absolute stress levels. In general, stress drops of crustal earthquakes are on the order of 1–10 MPa. Equation (8) can be used to show that such stress drops are only a small fraction of the shear stresses that actually causes fault slip if pore pressures are approximately hydrostatic at depth and Coulomb faulting theory (with laboratory-derived coefficients of friction) is applicable to faults *in situ*. This is discussed in more detail below.

B. Geologic Stress Indicators

There are two general types of *relatively young* geologic data that can be used for *in situ* stress determinations: (1) the orientation of igneous dikes or cinder cone alignments, both of which form in a plane normal to the least principal stress, and (2) fault slip data, particularly the inversion of sets of striae (i.e., slickensides) on faults as described above. Of course, the term "relatively young" is quite subjective, but it essentially means that the features in question are characteristic of the tectonic processes currently active in the region of question. In most cases, the WSM database utilizes data which are Quaternary in age, but in all areas represent the youngest episode of deformation in an area.

C. *In Situ* Stress Measurements

Numerous techniques have been developed for measuring stress at depth. Because we are principally interested here in regional tectonic stresses (and their implications) and because there are a variety of nontectonic processes that affect *in situ* stresses near the earth surface, we do not utilize near-surface stress measurements in the WSM or regional tectonic stress compilations (these measurements are given the lowest quality in the criteria used by WSM, as they are not believed to be reliably indicative of the regional stress). In general, we believe that only *in situ* stress measurements made at depths greater than 100 m are indicative of the tectonic stress field at midcrustal depths. This means that techniques utilized in wells and boreholes, which access the crust at appreciable depth, are especially useful for stress measurements.

When a well or borehole is drilled, the stresses that were previously supported by the exhumed material are transferred to the rock surrounding the hole. The resultant stress concentration is well understood from elastic theory. Because this stress concentration amplifies the stress difference between far-field principal stresses by a factor of four, there are several ways in which the stress concentration around boreholes can be exploited to help measure *in situ* stresses. The hydraulic fracturing technique takes advantage of this stress concentration and, under ideal circumstances, enables stress magnitude and orientation measurements to be made to about 3 km depth.

The most common method of determining stress orientation from observations in wells and boreholes is stress-induced wellbore breakouts. Breakouts are related to a natural compressive failure process that occurs when the maximum hoop stress around the hole is large enough to exceed the strength of the rock. This causes the rock around a portion of the wellbore to fail in compression. For the simple case of a vertical well drilled when S_V is a principal stress, this leads to the occurrence of stress-induced borehole breakouts that form at the azimuth of the minimum horizontal compressive stress. Breakouts are an important source of crustal stress information because they are ubiquitous in oil and gas wells drilled around the world and because they also permit stress orientations to be determined over a great range of depth in an individual well. Detailed studies have shown that these orientations are quite uniform with depth and are independent of lithology and age.

Another form of naturally occurring wellbore failure is drilling-induced tensile fractures. These fractures form in the wall of the borehole at the azimuth of the maximum horizontal compressive stress when the circumferential stress acting around the well locally goes into tension; they are not seen in core from the same depth.

III. DISTRIBUTION OF CRUSTAL STRESSES

Figure 3 shows maximum horizontal stress orientations for North America taken from the WSM database. The legend identifies the different types of stress indicators; because of the density of data, only highest quality data are plotted (A and B quality, shown by lines of different length). The tectonic regime (i.e., normal faulting, strike-slip faulting, or reverse faulting), where known, is given by color, as explained in the legend. The data principally come from wellbore breakouts, earthquake focal mechanisms, *in situ* stress measurements greater than 100 m depth, and young (<2 Ma old) geologic indicators. These data, originally presented and described by Zoback and Zoback (1989, 1991), demonstrate that large regions of the North American continent (most of the region east of the Rocky Mountains) are characterized by relatively uniform horizontal stress orientations. Furthermore, where different types of stress orientation data are available, see, for example, the eastern United States, the correlation between the different types of stress indicators is quite good.

Two straightforward observations about crustal stress can be made by comparison of these different types of stress indicators. First, no major changes in the orientation of the crustal stress field occur between the upper 2–5 km, where essentially all of the wellbore breakout and stress measurement data come from, and 5–20 km, where the majority of crustal earthquakes occur. Second, the criterion used to define reliable stress indicators appear to be approximately correct. In other words, data badly contaminated by nontectonic (near surface) sources of stress appear to have been effectively eliminated from the compilations.

IV. FIRST-ORDER GLOBAL STRESS PATTERNS

Figure 4 shows global maximum horizontal compressive stress orientations based on the 1997 WSM database. As with Fig. 3, only data qualities A and B are shown, and the symbols utilized in Fig. 4 are the same as those in Fig. 3. While global coverage is quite variable, the relative uniformity of stress orientation and the relative magnitudes in different parts of the world are striking and permit mapping of regionally coherent stress fields. Figure 5 presents a generalized version of the World Stress Map that is quite similar to that presented by Zoback (1992), showing, using large arrows, mean stress directions and stress regime based on averages of clusters of data shown in Fig. 4. Tectonic stress regimes are indicated in Fig. 5 by color and arrow type. Blue inward pointing arrows indicate S_{Hmax} orientations in areas of compressional (strike-slip and thrust) stress regimes. Red outward pointing arrows give S_{hmin} orientations (extensions direction) in areas of normal faulting regimes. Regions dominated by strike- slip tectonics are distinguished with green, thick inward pointing and orthogonal, thin outward pointing arrows. Overall, arrow sizes on Fig. 5 represent a subjective assessment of "quality" related to the degree of uniformity of stress orientation and also to the number of density of data.

A number of first-order patterns can be observed in Figs. 4 and 5.

1. In many regions a uniform stress field exists throughout the upper brittle crust, as indicated by consistent orientations from the different techniques that sample very different rock volumes and depth ranges.
2. Intraplate regions are dominated by compression (thrust and strike-slip stress regimes) in which the maximum principal stress is horizontal.
3. Active extensional tectonism (normal faulting stress regimes) in which the maximum principal stress is vertical generally occurs in topographically high areas in both the continents and the oceans.
4. Regional consistency of both stress orientations and relative magnitudes permits the definition of broad-scale regional stress provinces, many of which coincide with physiographic provinces, particularly in tectonically active regions. These provinces may have lateral dimensions on the order of 10^3–10^4 km, many times the typical lithosphere thickness of 100– 300 km. These broad regions of the earth's crust subjected to uniform stress orientation or a uniform pattern of stress orientation (such as the radial pattern of stress orientations in China) are referred to as "first-order" stress provinces (Zoback, 1992).

V. SOURCES OF CRUSTAL STRESS

As alluded to above, stresses in the lithosphere are of both tectonic and nontectonic, or local, origin. The regional uniformity of the stress fields observed in Figs. 3 and 4 argue for tectonic origins of stress at depth for most intraplate regions around the world. For many years, numerous workers suggested that residual stresses from

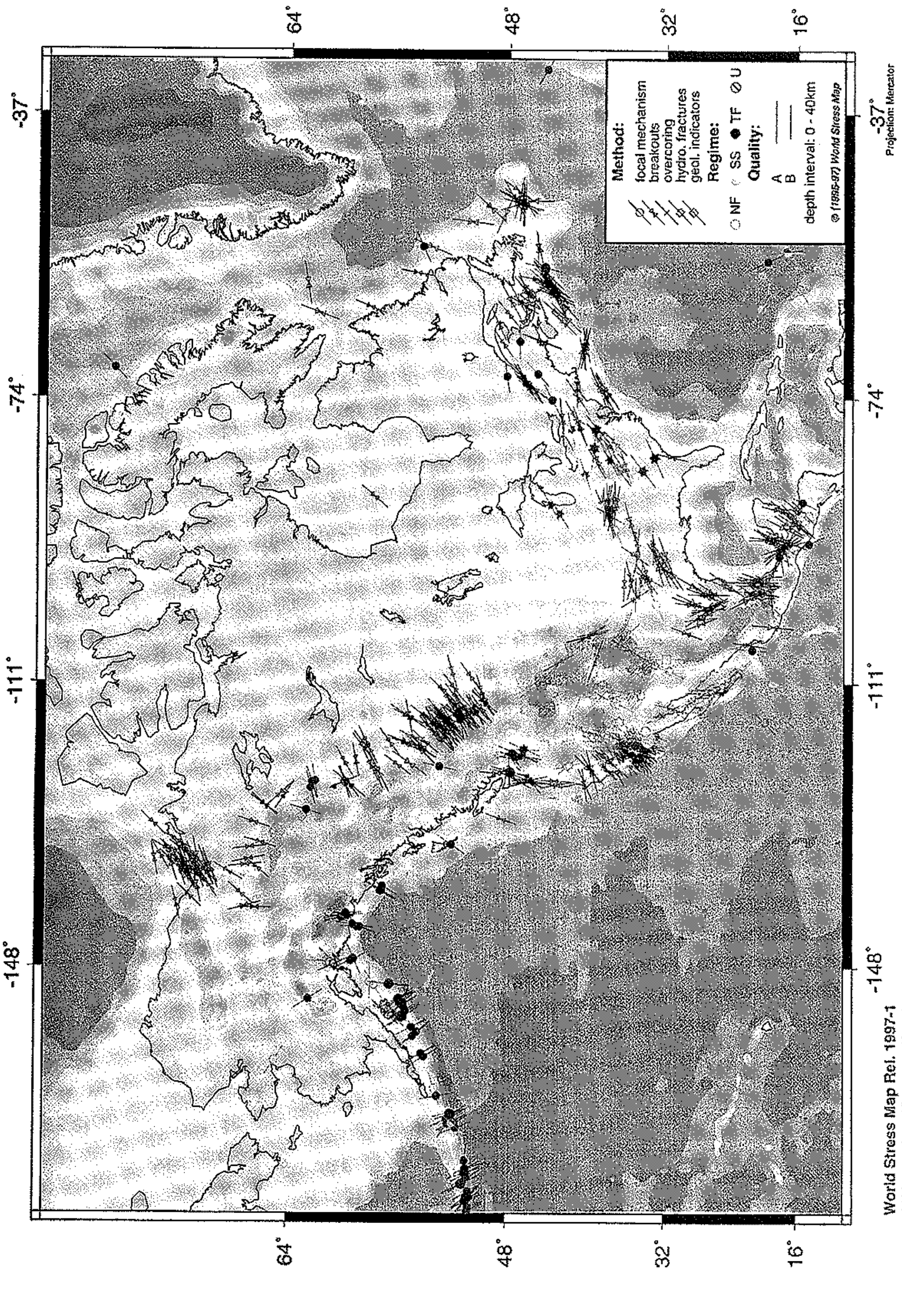

FIGURE 3 Directions of maximum horizontal stress from the WSM database superimposed on global topography and bathymetry. Only A and B quality data are shown. Data points characteristic of normal faulting are shown in red, strike-slip areas are shown in green, reverse faulting areas are shown in blue, and unknown stress indicators are shown in black.

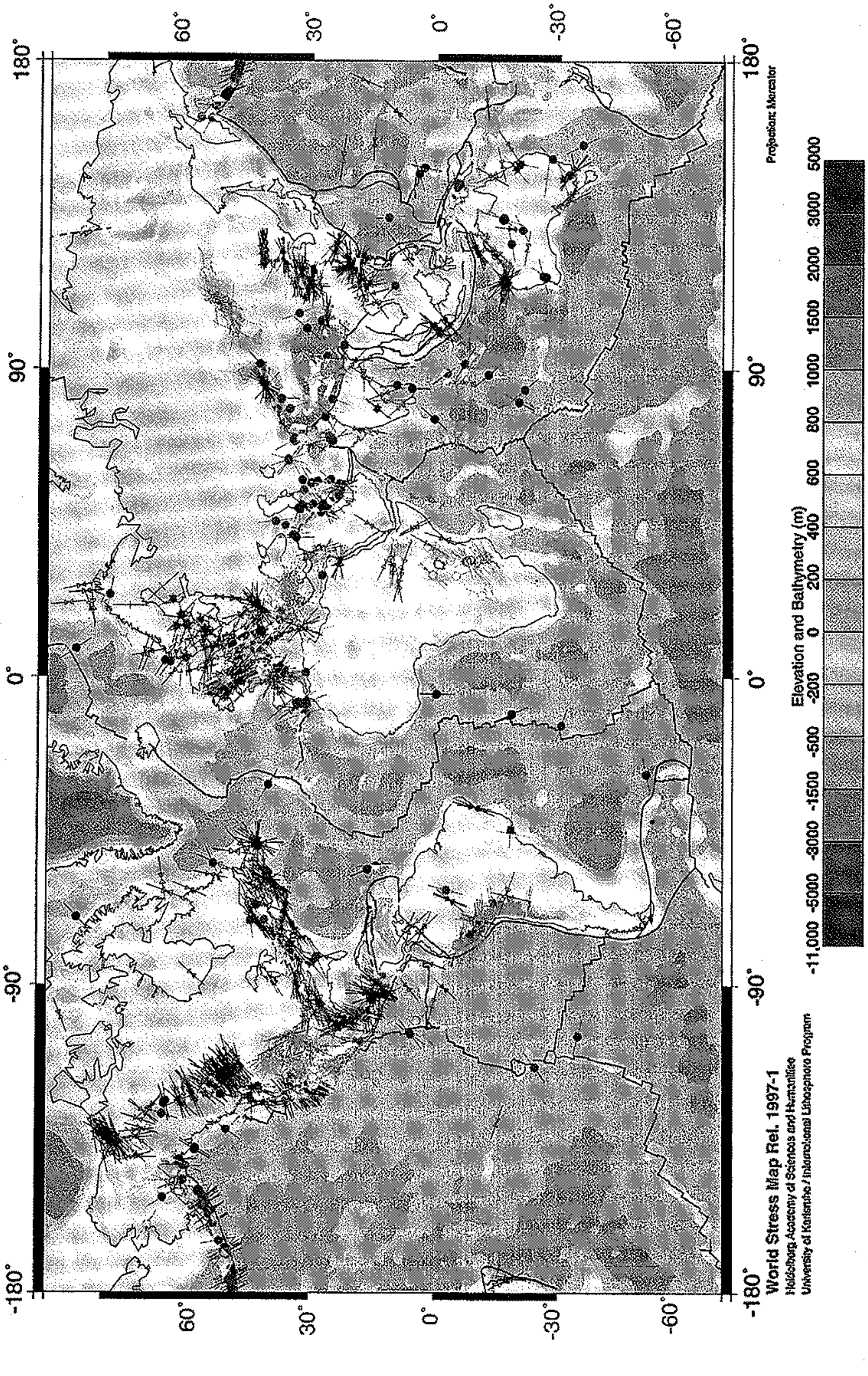

FIGURE 4 Directions of maximum horizontal stress from the WSM database. Colors are the same as in Figure 3. Only A and B quality data are shown.

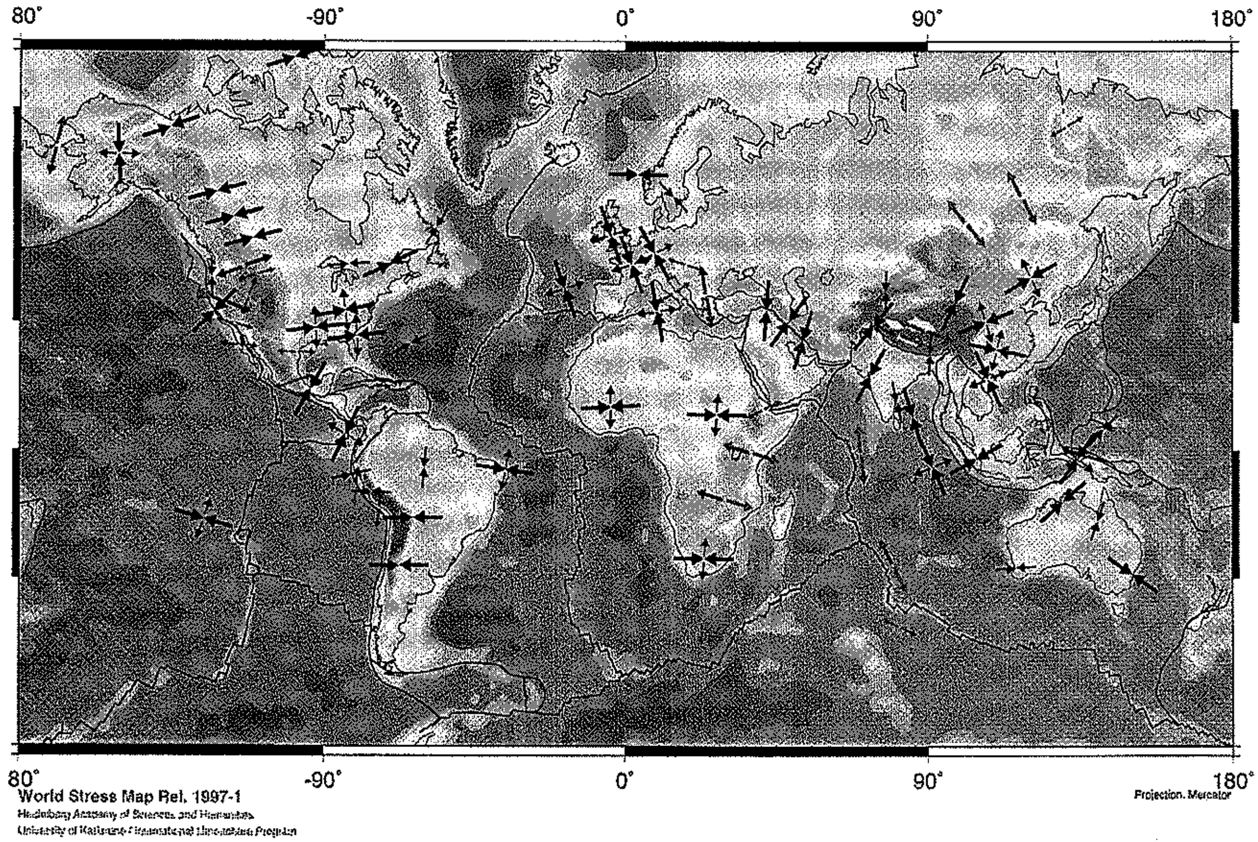

FIGURE 5 Generalized world stress map based on the data in Fig. 4 and similar to that of Zoback (1992). Inward pointed arrows indicate high compression as in reverse faulting regions. Paired inward and outward arrows indicate strike-slip faulting. Outward directed red arrows indicate areas of extension. Note that the plates are generally in compression and that areas of extension are limited to thermally uplifted areas.

past tectonic events may play an important role in defining the tectonic stress field. We have found no evidence for significant residual stresses at depth. If such stresses exist, they seem to be only important in the upper few meters or tens of meters of the crust where tectonic stresses are very small.

Similarly, no evidence has been found that indicates that horizontal principal stresses result simply from the weight of the overlying rock. This oversimplified theory is based on the "bilateral constraint," with the supposition that as a unit cube is stressed due to imposition of a vertical stress it cannot expand horizontally (because a neighboring unit cube would be attempting to expand in the opposite direction). If the bilateral constraint was applicable to the crust, the two horizontal principal stresses at depth would be equal and would always be less than the vertical stress. As demonstrated above, the dominance of compressional intraplate stress fields indicates that one or both horizontal stresses exceed the vertical stress. The broad regions of well-defined S_{hmax} orientations are clear evidence of horizontal stress anisotropy. Thus, the predicted bilateral stress state is generally not found in the crust. In fact, the assumptions leading to the prediction of such a stress state are unjustified, as the analysis assumes that an elastic crust exists in the absence of gravity (or any other forces) before gravity is instantaneously "switched" on.

In the sections below, the primary sources of tectonic stress are briefly discussed. Although it is possible to theoretically derive the significance of any particular source of stress, because the observed tectonic stress state is the result of superposition of a variety of forces acting on and within the lithosphere, it is difficult to define the relative importance of any one stress source. This can only be resolved by utilizing careful modeling and well-constrained observations.

A. Plate-Driving Stresses

Sources of the broad-scale regions of uniform crustal stress that immediately come to mind are the same forces that drive (and resist) plate motions (e.g., Forsyth and Uyeda, 1975). A ridge push compressional force is associated with the excess elevation of the mid-ocean ridges, whereas the slab pull force results from the negative buoyancy of downgoing slabs. Both of these sources contribute to plate motion and tend to act in the direction of plate motion. If there is flow in the upper asthenosphere, a positive drag force could be exerted on the lithosphere that would tend to drive plate motion, whereas cold, thick lithospheric roots (such as beneath cratons) may be subject to a resistive drag forces that would act to inhibit plate motion. In either case, the drag force would result in stresses being transferred up into the lithosphere from its base. There are also collisional resistance forces resulting either from the frictional resistance of a plate to subduction or from the collision of two continental plates. As oceanic plates subduct into the viscous lower mantle, additional slab resistive forces add to the collision resistance forces acting at shallow depth. Another force resisting plate motion is that due to transform faults, although, as discussed below, the amount of transform resistance may be negligible. Finally, it has been proposed that a suction force may act on the overriding lithosphere in a subduction zone. This force may tend to "suck" the overriding lithosphere toward the trench and result in back-arc spreading.

While it is possible to specify the various stresses associated with plate movement, their relative and absolute importance in plate movement is not understood. While many researchers believe that either the ridge push or slab pull force is most important in causing plate motion, it is not clear that these forces are easily separable or that plate motion can be ascribed to a single dominating force.

B. Topography and Buoyancy Forces

Numerous workers have demonstrated that topography and its compensation at depth can generate sizable stresses capable of influencing the tectonic stress state and style. Density anomalies within or just beneath the lithosphere constitute major sources of stress. The integral of anomalous density times depth characterizes the ability of density anomalies to influence the stress field and to induce deformation. In general, crustal thickening or lithospheric thinning (negative density anomalies) produces extensional stresses, while crustal thinning or lithospheric thickening (positive density anomalies) produces compressional stresses. In more complex cases, the resultant state of stress in a region depends on the density moment integrated over the entire lithosphere. In a collisional orogeny, for example, where both the crust and the mantle lid are thickened, the presence of the cold lithospheric root can overcome the extensional forces related to crustal thickening and maintain compression.

C. Lithospheric Flexure

Loads on or within an elastic lithosphere cause deflection and induce flexural stresses which can be quite large (several hundred megaPascals) and can perturb the regional stress field with wavelengths as much as 1000 km (depending on the lateral extent of the load). Some potential sources of flexural stress influencing the regional stress field include sediment loading, particularly along continental margins; glacial rebound; seamount loading; and the upwarping of oceanic lithosphere oceanward of the trench, the "outer arc bulge." Sediment loads as thick as 10 km represent a potential significant stress on continental lithosphere. Zoback (1992) suggested that a roughly 40° counterclockwise rotation of horizontal stresses on the continental shelf offshore of eastern Canada was due to superposition of a margin-normal extensional stress derived from sediment load flexure.

VI. THE CRITICALLY STRESSED CRUST

Three independent lines of evidence indicate that intraplate continental crust is generally in a state of incipient, but slow, frictional faulting: (1) the widespread occurrence of seismicity induced by either reservoir impoundment or fluid injection (Zoback and Harjes, 1997), (2) earthquakes triggered by small stress changes associated with other earthquakes, and (3) *in situ* stress measurements in deep wells and boreholes. The *in situ* stress measurements further demonstrate that the stress magnitudes derived from Coulomb failure theory utilizing laboratory-derived frictional coefficients of 0.6–1.0 predict stresses that are consistent with measured stress magnitudes (Townend and Zoback, 2000). This is well illustrated in Fig. 6 by the stress data collected in the KTB borehole in Germany to ~8 km depth. Measured stresses are quite high and consistent with the frictional faulting theory [Eq. (8)] with a frictional coefficient of ~0.7. Further demonstration of this "frictional failure" stress state was the fact that a series of earthquakes could be triggered at ~9 km depth in rock surrounding the KTB borehole by extremely low perturbations of the ambient, approximately hydrostatic, pore pressure (Zoback and Harjes, 1997).

That the state of stress in the crust is generally in a state of incipient frictional failure might seem surprising, especially for relatively stable intraplate areas. However, a reason for this can be easily visualized in terms of a simple cartoon as shown in Fig. 7. The lithosphere as a whole (shown simply in Fig. 7 as three distinct layers: the brittle upper crust, the ductile lower crust, and the ductile uppermost mantle) must support plate-driving forces. The figure indicates a power-law creep law typically used to

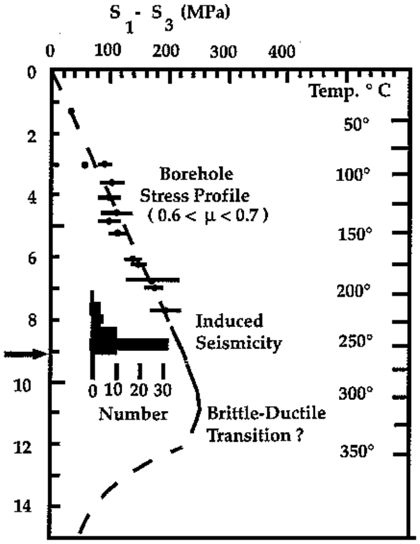

FIGURE 6 Stress measurements in the KTB scientific research well indicate a "strong" crust, in a state of failure equilibrium as predicted by Coulomb theory and laboratory-derived coefficients of friction of 0.6–0.7. [Modified after Zoback, M. D., and Harjes, H. P. (1997). *J. Geophys. Res.* **102,** 18,477–18,491.]

characterize the ductile deformation of the lower crust and upper mantle. Because the applied force to the lithosphere will result in steady-state creep in the lower crust and upper mantle, as long as the "three-layer" lithosphere is coupled, stress will build up in the upper brittle layer due to the creep deformation in the layers below. Stress in the upper crust builds over time, eventually to the point of failure. The fact that intraplate earthquakes are relatively infrequent simply means that the ductile strain rate is low in the lower crust and upper mantle. Zoback and Townend (2001) discuss the fact that at the relatively low strain rates characterizing intraplate regions, sufficient plate-driving force is available to maintain a "strong" brittle crust in a state of frictional failure equilibrium.

While appreciable evidence suggests that the Coulomb criterion and laboratory-derived coefficients of friction are applicable to plate interiors, major plate boundary faults such as the San Andreas fault (and other plate-bounding faults) appear to slip at very low levels of shear stress. Appreciable heat flow data collected in the vicinity of the San Andreas show no evidence of frictionally generated heat (Lachenbruch and Sass, 1980). This appears to limit average shear stresses acting on the fault to depths of ~15 km to about 20 MPa, approximately a factor of 5 below the stress levels predicted by the Coulomb criterion assuming that hydrostatic pore pressure and laboratory-derived friction coefficients are applicable to the fault at depth. Zoback *et al.* (1987) and Townend and Zoback (2001) present evidence that the direction of maximum horizontal stress in the crust adjacent to the San Andreas is at an extremely high angle to fault plane. Like the heat flow data, the stress orientation data imply low resolved shear stresses on the fault at depth. Thus, it appears that the frictional strength of plate boundary faults is distinctly lower than intraplate faults, which thus enables them to accommodate hundreds of kilometers of relative fault offset and correspondingly high strains.

VII. SUMMARY

Large portions of intraplate regions (lateral dimensions of 10^3–10^4 km) are characterized by relatively uniform stress fields, suggesting that the state of stress in the brittle upper crust is dominated by large-scale tectonic processes. In general, principal stresses in the crust are in vertical and horizontal planes. In intraplate areas, *in situ* stress measurements and inferences based on topography and flexure all suggest that shear stresses in the upper lithosphere is fairly large and seem to be controlled by the frictional strength of the faulted crust. Unlike the interiors of plates, plate boundary faults (like the San Andreas

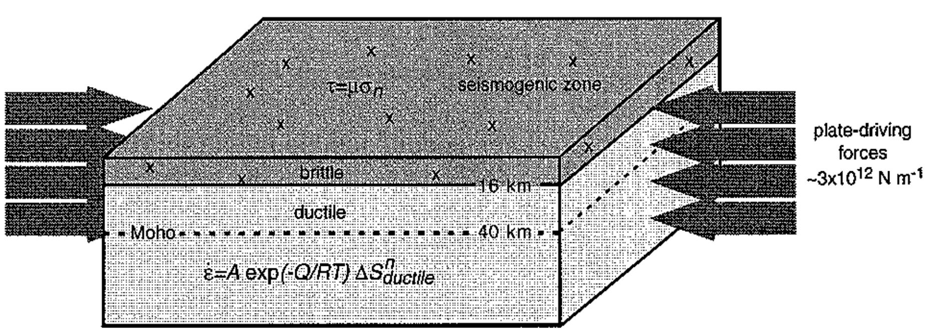

FIGURE 7 Schematic cartoon illustrating how the forces acting on the lithosphere keep the brittle crust in frictional equilibrium through creep in the lower crust and upper mantle. [Modified after Zoback, M. D., and Townend, J. (in press). *Tectonophysics.*]

fault and subduction zones) appear to be quite weak and slip at low levels of shear stress. At sufficient depth below the brittle upper lithosphere, stresses are likely controlled by the ductile flow properties of the constituent rocks and minerals. Plate-driving forces are sufficient in magnitude to maintain the intraplate lithosphere in a state of frictional failure. This frictional failure is manifest as slow, steady-state creep deformation in the lower crust and upper mantle and brittle deformation in the upper crust.

ACKNOWLEDGMENT

We thank Peter Bird for his comments on an early draft of this article.

SEE ALSO THE FOLLOWING ARTICLES

CONTINENTAL CRUST • EARTHQUAKE MECHANICS AND PLATE TECTONICS • EARTHQUAKE ENGINEERING • EARTHQUAKE PREDICTION • EARTH'S CORE • EARTH'S MANTLE • GEOLOGY, EARTHQUAKE • OCEANIC CRUST • PLATE TECTONICS • SEISMOLOGY, ENGINEERING • SEISMOLOGY, OBSERVATIONAL • SEISMOLOGY, THEORETICAL

BIBLIOGRAPHY

Anderson, E. M. (1951). "The Dynamics of Faulting and Dyke Formation with Applications to Britain," Oliver & Boyd, Edinburgh.

Byerlee, J. D. (1978). "Fusion of rock," *Pure and Applied Geophysics*, **116,** 615–626.

Forsyth, D., and Uyeda, S. (1975). "On the relative importance of the driving forces of plate motion," *Geophys. J. R. Astr. Soc.* **43,** 163–200.

Jaeger, J. C., and Cook, N. G. W. (1971). "Fundamentals of Rock Mechanics," 515 pp., Chapman and Hall, London.

Jones, C. H., Unruh, J. R., and Sonder, L. J. (1996). "The role of gravitational potential energy in active deformation in the southwestern United States," *Nature* **381,** 37–41.

Lachenbruch, A. H., and Sass, J. H. (1980). "Heat flow and energetics of the San Andreas fault zone," *J. Geophys. Res.* **85,** 6185–6223.

McKenzie, D. P. (1969). "The relationship between fault plane solutions for earthquakes and the principal stresses," *Seis. Soc. Amer. Bull.* **59,** 591–601.

Mueller, B., Wehrle, V., and Fuchs, K. (1997). "The 1997 Release of the World Stress Map," http://www-wsm.physik.uni-karlsruhe.de/.

Townend, J., and Zoback, M. D. (2000). "How faulting keeps the crust strong," *Geology* **28,** 399–402.

Zoback, M. D., and Townend, J. (2001). "Implications of hydrostatic pore pressures and high crustal strength for the deformation of interplate lithosphere," *Tectonphysics* **336,** 19–30.

Zoback, M. D., and Harjes, H. P. (1997). "Injection induced earthquakes and crustal stress at 9 km depth at the KTB deep drilling site, Germany," *J. Geophys. Res.* **102,** 18,477–18,491.

Zoback, M. D., and Zoback, M. L. (1991). Tectonic stress field of North America and relative plate motions. *In* "The Geology of North America: Neotectonics of North America" (D.B.Spemmons *et al*, ed.), pp. 339–366, Geol. Soc. Am., Boulder, CO.

Zoback, M. D., *et al.* (1987). "New evidence on the state of stress of the San Andreas fault system," *Science* **238,** 1105–1111.

Zoback, M. L., *et al.* (1989). "Global patterns of intraplate stresses; a status report on the world stress map project of the International Lithosphere Program," *Nature* **341,** 291–298.

Zoback, M. L. (1992). "First and second order patterns of tectonic stress: The World Stress Map Project," *J. of Geophys. Res.* **97,** 11,703–11,728.

Zoback, M. L., and Zoback, M. D. (1989). "Tectonic stress field of the conterminous United States," *Geol. Soc. Am. Memoir* **172,** 523–539.