STATE OF STRESS IN THE LITHOSPHERE

Mark D. Zoback
U.S. Geological Survey, zoback@stanford.edu

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STATE OF STRESS IN THE LITHOSPHERE

Mark D. Zoback
U.S. Geological Survey, Menlo Park, California 94025

Abstract. Recent estimates of the magnitude and orientation of lithospheric stresses are reviewed. Data come from a variety of sources including earthquake focal mechanisms, fault slip data, young volcanic dikes and feeders, in-situ stress measurements at depth, lithospheric flexure models, microstructure paleopiezometry, and consideration of constraints on maximum stress differences from laboratory-determined friction and flow laws.

Introduction

Studies related to understanding the magnitude and orientation of lithospheric stresses have been going on at an active pace during the past four years. In this brief report, I intend to highlight some of the recent work in this field, concentrating on investigations that have attempted to integrate geophysical observations and theory. Recent work on stress that is based solely on seismological observations (such as stress drop estimates), or laboratory studies, are not covered here [see other chapters of this report by D. Boore and S. Kirby]. An excellent compendium of recent work on the state of stress in the lithosphere can be found in a special issue of the Journal of Geophysical Research published in November 1980. The papers in this volume originated from a conference on the Magnitude of Deviatoric Stresses in the Earth's Crust and Uppermost Mantle convened by Thomas Hanks and C. Barry Raleigh in the summer of 1979 [Hanks and Raleigh, 1980]. Other notable recent publications that make repeated reference to the magnitude and orientation of lithospheric stresses are the compendiums Earthquake Prediction [edited by Simpson and Richards, 1981] and Mechanical Behavior of Crustal Rocks, The Handin Volume [edited by Carter et al., 1981] both published by the American Geophysical Union.

Magnitude of Lithospheric Stresses

Numerous techniques and methods of analysis have been advanced in the past four years to estimate the magnitude of lithospheric stresses. Unfortunately, different analysis methods often yield different types of stress estimates. For example, modeling of lithospheric flexure usually yields estimates of maximum bending stresses and differential stresses averaged over the upper 30-40 km of the lithosphere. Stress estimates based on the frictional or rheological properties of rocks...
limited by the frictional strength of faulted rock. For a wide variety of rock types Byerlee [1968] fit a simple relationship between laboratory measured frictional shear strength, and normal stress, $\sigma_n$. The simple relations

$$\tau = 0.85 \sigma_n \quad \text{for} \quad 30 < \sigma_n < 2000 \text{ bars}$$

$$\tau = 60 + 10 + 0.6 \sigma_n \quad \text{for} \quad \sigma_n > 2000 \text{ bars}$$

seem to be virtually independent of rock type (with the exception of some clays), displacement, and surface roughness. These relations have been referred to as Byerlee's law [Brace and Kohlstedt, 1980].

Assuming that there is at least one pre-existing fault plane that is well oriented with respect to the principal stresses, Brace and Kohlstedt [1980] show that in-situ stress data from southern Africa and Canada [compiled by McGarr and Gay, 1978] are limited by the frictional strength of faulted rock in the upper ~2 km of the crust (Fig. 1). Zoback [1982a] and Zoback and Hickman [1982] present several other examples, primarily in reverse-faulting environments, which support this contention by showing that seismically active areas have stress differences near the limiting value established by Byerlee's law. Byerlee's law also seems to hold for areas of incipient normal faulting [Hubbert and Willis, 1957; McGarr and Gay, 1978; Healy et al., 1982].

If Byerlee's law is nearly correct for most crustal materials, and pore pressures in the upper crust are approximately hydrostatic, it is straightforward to show that the approximate value of maximum shear stress in the crust increases with depth by about 200 bars/km in thrust-faulting environments, 80 bars/km in strike-slip faulting areas (if $n$ is equal to the average horizontal stress), and 50 bars/km in areas of normal faulting [see Sibson 1974, 1980]. These gradients correspond to maximum differential stresses ($p = 2$) of 400, 160, and 100 bars/km, respectively. Laboratory studies suggest that such stress gradients would be applicable until the thermally activated creep processes began to affect rock strength at depths of about 12 km in quartz-rich continental rocks, or 30 km in olivine-rich oceanic crust [Brace and Kohlstedt, 1980].

### Shear Stress on the San Andreas Fault

A crucial question regarding the mechanical behavior of the San Andreas fault is the magnitude of mean shear stress on the fault within the seismogenic layer (0-15 km). Although this subject has been widely discussed and debated, the issue basically remains unresolved.

As heat flow near the fault is proportional to the mean stress times the displacement rate, the lack of any observable heat-flow anomaly near the fault has led a variety of workers to suggest that the mean shear stress on the fault was at most a few hundred bars [Brune et al., 1969; Lachenbruch and Sass, 1973, 1980; Turcotte et al., 1980]. In the absence of the lack of an observable heat-flow anomaly, and other evidence, Raleigh and Evernden [1981] concur.
with this low stress estimate. The heat-flow constraint would seem to place an upper bound for the long-term mean shear stress of only 100-200 bars, but such low stress estimates are clearly incompatible with Byerlee's law [Hanks, 1977] unless the fault is composed of very low-strength clay-gouge [Norrer et al., 1982] or frictional strength is reduced during faulting [Lachenbruch, 1980; Raleigh and Evernden, 1980].

Because high pore pressure could significantly lower frictional strength [Hubbert and Rubey, 1959], Lachenbruch [1980] extensively investigates the manner in which transiently high fluid pressure could be generated during faulting. Raleigh [1977] discusses a thermally induced dehydration mechanism that could raise fluid pressure (and lower frictional strength), and Sibson [1980] points out that either transient high pore pressure or frictional melting [Jeffreys, 1943; Anderson, 1951; McKenzie and Brune, 1972; Richards, 1976] could be important mechanisms for lowering frictional strength.

Some workers have cited evidence of shear heating near major faults that they interpret as an implication of high shear stress [Sheppard et al., 1975; Scholz et al., 1979; Scholz, 1980; Sibson, 1980]. In the schists adjacent to the Alpine fault in New Zealand, Scholz et al. [1979] attribute argon depletion and isograds that are subparallel to the fault as evidence of shear heating. Scholz [1980] cites 16 other cases in which there is apparent evidence of shear heating. Sibson [1975] suggests that pseudotachlyte, found near some exhumed reverse faults in crystalline host rocks, is evidence for frictional melting on fault planes, but Sibson [1980] argues that there appear to be both high- and low-stress faulting on the basis of rock deformation textures from exhumed fault zones.

O'Neill and Hanks [1980] challenge the use of conductive heat-flow measurements as a valid constraint on the frictional strength of the San Andreas fault, and point to geochemical data which suggest that there may be a considerable amount of convective heat flow away from the fault plane. However, Lachenbruch and Sass [1980] point out that there is no evidence for appreciable convective heat flow (such as thermal springs) near the fault.

Unfortunately, in-situ stress measurements made to date near the San Andreas do not yet constrain estimates of shear stress on the fault at depth. In a well approximately 1 km deep near the San Andreas in the western Mojave desert, Zoback et al. [1980] report an increase in shear stress (resolved onto a plane parallel to the fault) with depth at a rate of about 95 bars/km. This rate is consistent with that expected from Byerlee's law, and extrapolation of this gradient to greater depths rapidly leads to stresses that exceed the 100-200-bar value allowed by the heat-flow data [see also McGarr et al., 1982]. Although these stress measurements seem to support a high-stress fault, the stresses measured in the 1-km-deep well increase with depth in a step-like manner, and simple linear extrapolation of the data to greater depth may not be reasonable. McGarr et al. [1982] point out, however, that the stress measurements from the western Mojave do not indicate particularly low values compared to other areas of the world. As the maximum magnitude of in-situ stresses from these other areas seem constrained by the frictional strengths of faults (as discussed above), they claim that this is indirect evidence that there is high stress on the San Andreas. Nevertheless, until stress measurements near the fault are available from greater depths, this issue will probably not be resolved.

Flexure and Gravitational Models

Estimates of the average magnitude of differential stresses and maximum bending stresses can be gotten from theoretical models of lithospheric flexure in response to various types of loads [Walcott, 1970a,b; Hanks, 1971; Watts and Taiwani, 1974; Caldwell et al., 1976; McAdoo et al., 1978] and the manner in which topography and gravity anomalies are compensated [Jeffreys, 1976; Artyushkov, 1973, 1974; McKenzie, 1969; McNutt, 1980; Lambeck, 1980]. Recent advances in these types of studies come from attempts to better constrain models by using more realistic rheological rock properties. That is, because of the importance of plastic deformation below depths of about 30 km, the use of lithospheric flexure models with only thick elastic plates can yield average regional differential stresses that exceed values based on laboratory rheological data [Kirby, 1977, 1980; Goetze and Evans, 1979] by as much as 5 kilobars.

In elastic-plastic models, the yield strength \( c \) is a function of both temperature \( T \) (or depth) and strain rate \( \dot{\varepsilon} \) through an equation of the type

\[
\sigma = A c \exp\left(-\frac{Q}{RT}\right) \tag{1}
\]

where \( A \) and \( n \) are constants to be derived experimentally, \( Q \) is the frictional energy, and \( R \) is the gas constant [after Carter, 1976]. In the case of subducting slabs, Forsyth [1980] suggests that models using an elastic-plastic rheology [Liu and Kosloff, 1978; Chapple and Forsyth, 1977] are not only most compatible with laboratory rheological data, they also best explain the location and concentration of earthquake activity near the trench, the depth to which normal faulting persists, and the cumulative seismic moment of the normal faulting earthquakes. On the basis of his analysis with elastic-plastic rheological models, Forsyth concludes that models which require large regional stress differences are generally unsatisfactory.

The basic problem with attempting to estimate the magnitude of differential stresses solely from flexure and gravitational models is that there is a fundamental trade-off between stress magnitude and elastic plate thickness - the thicker the plate, the lower the stresses. In fact, McNutt and Menard [1982] demonstrate that it is possible to constrain lithospheric plate models with eqn. (1) and fit flexure data by adjusting the plate thickness and activation energy. In other words, their estimate of
lithospheric stress is constrained as much by the laboratory-derived flow law than the flexure data. Similarly, from analysis of gravity anomalies and topography, Lambeck [1980] and Lambeck and Nakiboglu [1980] show that previous estimates of average differential stress as large as 5-10 kbar required to support crustal loads can be reduced to about 1 kbar by considering an elastic-plastic rheology.

Microstructure Paleopiezometry

Estimates of the maximum stress difference that a deformed rock has been subjected to can be made on the basis of recrystallized grain size, or subgrain size [Twiss, 1977, 1980; Mercier et al., 1977; Mercier, 1980] and dislocation density [Weathers et al., 1979]. Exhumed fault zone materials and peridotite xenoliths from the mantle have been used to estimate the maximum magnitude of paleostresses. Although an advantage of these methods is that they are relatively independent of temperature, an obvious disadvantage is that it is not known at what time during a rock's history a given microstructural texture developed. For example, Christie and Ord [1980] report that the quartz-bearing Coyote Mountain mylonite yields a maximum paleostress difference of 160-740 bars on the basis of grain size, but a maximum paleostress difference of 500-1300 bars on the basis of dislocation density. They attribute this difference to different deformation stages in the rocks' past. Presumably the grain size is the result of a low-stress annealing period, whereas the dislocation density results from a late-stage high-stress pulse.

Kohlstedt and Weathers [1980] summarize the results of microstructural analysis of rocks from various thrust and shear zones. These studies indicate maximum differential stresses of 1000-2000 bars for fault-zone materials exhumed from depths as great as 20-30 km. While recognizing problems such as that illustrated by Christie and Ord [1980], Kohlstedt and Weathers conclude that in most cases stress estimates made using different microstructural parameters agree well.

Mercier [1980] presents stress estimates from grain sizes in peridotite xenoliths. As these samples came from depths greater than 40 km, the results probably apply only to the lower crust and upper mantle, and indicate that differential stresses are in the 450-800 bars range.

Summary

The results of some of the studies discussed above are summarized in Table 1. In quartz-rich continental rock, a variety of evidence indicates that maximum differential stresses are in the 1-1.5 kbar range in the mid-crust (to depths of 10-20 km). Paleopiezometry indicates that maximum differential stresses decrease to about half that value at depths of 20-50 km. These results are generally consistent with average differential stresses in the upper 40 km of about 1-2 kbar yielded by studies of gravity and topography.

In oceanic areas, friction probably controls strength to a depth of about 30 kb. This yields a maximum strength of 3.0-6.0 kb. As average differential stresses in the upper 30-40 km indicated by topography and gravity modeling are only in the 1-1.5 kbar range, they are well within the limit of rock strength. More importantly, so are maximum bending stresses of about 5 kbar at 25 km depth yielded by flexure models with elastic-plastic rheologies.

Orientation of Horizontal Principal Stresses

There have been several attempts to compile data on the orientation of the in-situ stress field in the conterminous U.S. [Raleigh, 1974; Haimson, 1977; McGarr and Gay, 1978]. Due to the severe lack of data over very large areas, these studies were only marginally successful. Zoback and Zoback [1980] attempted to overcome the data shortage by synthesizing various types of-stress-field indicators (geologic, seismologic (focal plane), and in-situ stress measurements from depth). A rigid criterion was established for each type of data that was used. The different stress-field indicators gave results that were generally consistent with respect to each other as well as with the current tectonics within the various stress provinces (Figure 2).

The fact that stress measurements made at only a few hundred meters depth, or a few kilometers at most, yield horizontal stress orientations consistent with well-constrained focal-plane information from much deeper earthquakes indicates that the orientation and relative magnitude of horizontal principal stresses are generally consistent throughout much of the crust. Thus, tectonic forces seem to be reflected in the stress field at relatively shallow depths, and it seems reasonable to attempt to constrain tectonic models with both deep and relatively near-surface stress data. However, within a few tens of meters of the earth's surface the stress field often seems to be affected by topographic, or stress-relief, mechanisms [see Shor et al., 1979]. In general, such measurements cannot be expected to yield reliable tectonic stress orientations, although in some cases they do appear to agree with other stress-field indicators [e.g., Froidevaux et al., 1980].

Stress Provinces

The concept of stress provinces, or areas in which the relative magnitude and orientation of the in-situ stresses are uniform, has been evolving over the past few years as reliable data have accumulated. In addition to the discussion by Zoback and Zoback [1980] of stress provinces in the conterminous U.S., Nakamura and Uyeda [1980] discuss stress provinces associated with subduction zones and volcanic arcs. They find that (1) the area immediately landward of the trench is characterized by thrusting (with the direction of maximum compression parallel to the direction of convergence), (2) the volcanic arc is characterized by strike-slip faulting (with the direction of maximum horizontal compression still parallel to the convergence direction), and (3) the back-arc area is
<table>
<thead>
<tr>
<th>DEPTH RANGE</th>
<th>MAXIMUM DIFFERENTIAL STRESS</th>
<th>REMARKS</th>
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<tr>
<td>CONTINENTAL</td>
<td></td>
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<tr>
<td>0–12 km</td>
<td>1.2 kb (extensional)</td>
<td>Stress limited by frictional strength; hydrostatic pore pressure assumed</td>
<td>Brace and Kohlstedt [1980]; Sibson [1975, 1980]</td>
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<tr>
<td></td>
<td>1.9 kb (shear)</td>
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<td></td>
<td>2.5 kb (compressional)</td>
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<tr>
<td>4–10 km</td>
<td>0.5–2.0 kb</td>
<td>Paleopiezometry-Moine thrust</td>
<td>Twiss [1977]; Weathers et al. [1979]</td>
</tr>
<tr>
<td>7.5 km</td>
<td>0.3–1.5 kb</td>
<td>Paleopiezometry-Massif Central shear zone</td>
<td>Burg and Laurent (1978)</td>
</tr>
<tr>
<td>10–15 km</td>
<td>1.0 kb</td>
<td>Paleopiezometry-Shear zone in Colorado</td>
<td>Kohlstedt and Weathers [1980]</td>
</tr>
<tr>
<td>15–20 km</td>
<td>1.0–2.0 kb</td>
<td>Paleopiezometry-Shear zone in Wyoming</td>
<td>Weathers et al. [1979]</td>
</tr>
<tr>
<td>15–30 km</td>
<td>0.2–1.1 kb</td>
<td>Paleopiezometry-Shear zone in Greenland</td>
<td>Kohlstedt et al. [1979]</td>
</tr>
<tr>
<td>0–40 km</td>
<td>1.5–2.0 kb</td>
<td>Gravity and topography</td>
<td>McNutt [1980]; Lambeck [1980]</td>
</tr>
<tr>
<td>40–50 km</td>
<td>0.4–0.8 kb</td>
<td>Paleopiezometry-peridotite xenoliths</td>
<td>Mercier [1980]</td>
</tr>
<tr>
<td>OCEANIC</td>
<td></td>
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<tr>
<td>0–30 km</td>
<td>3.0 kb (extensional)</td>
<td>Stress limited by frictional strength; hydrostatic pore pressure assumed</td>
<td>Brace and Kohlstedt [1980]; Sibson [1975, 1980]</td>
</tr>
<tr>
<td></td>
<td>4.8 kb (shear)</td>
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<tr>
<td></td>
<td>6.0 kb (compressional)</td>
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<td></td>
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<tr>
<td>0–40 km</td>
<td>5 kb</td>
<td>Outer Rise</td>
<td>McNutt and Menard [1982]</td>
</tr>
<tr>
<td>0–35 km</td>
<td>1 kb</td>
<td>Seamount Load</td>
<td>McNutt and Menard [1982]</td>
</tr>
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</table>
Fig. 2. Compilation of stress data in the conterminous U.S. [from Zoback and Zoback, 1980]. Heavy shaded lines define the boundaries of the stress provinces discussed in the text; heavy dashed lines mark the approximate boundaries. Stress provinces are abbreviated as follows: SA, San Andreas; SN, Sierra Nevada; PNW, Pacific Northwest; NRM, northern Rocky Mountains; HL, Hebgen Lake-Centennial Valley; SRF, Snake River Plain-Yellowstone; CP, Colorado Plateau; and SOP, southern Great Plains. The Basin and Range-Rio Grande Rift stress province includes the northern (NBR) and southern (SBR) Basin and Range and the Rio Grande Rift (RGR). Physiographic provinces are shown by light solid lines for reference. The number by each data point refers to a description of the point.
characterized by normal faulting (where the least horizontal compressive stress direction is parallel to the convergence direction). Elements of this pattern can also be seen in Japan, Alaska, the Hellenic arc, and central Europe [see also Mercier, 1981 and Angelier et al., 1982].

The fact that some stress provinces can be geographically small in tectonically active areas appears reasonable if local sources of stress like shallow magmatism, variations in crustal thickness [Artushkov, 1973], and heat flow [McGarr, 1982] are considered. In the past, relatively small-scale stress provinces (those smaller than a few hundred kilometers) have generally been ignored in studies of global plate interaction aimed at resolving the forces responsible for plate motion [Richardson et al., 1976; Solomon et al., 1980].

Intraplate Stresses in the Conterminous U.S.

The origin of stresses in intraplate areas is of interest for a number of reasons. The uniform direction of the intraplate stress field in the central and eastern U.S. led Zoback and Zoback [1980] to suggest drag at the base of the lithospheric plate as a possible causative mechanism. Whether or not this hypothesis is correct will require further testing as more data become available and more sophisticated models are developed. However, one possible origin of intraplate stresses in the central and eastern U.S., asthenospheric counterflow [Chase, 1979; Hager and O'Connell, 1979], can apparently now be rejected; this mechanism predicts stress directions which are nearly at right angles to those observed in the region.

A better understanding of the consistency of the stress field within the central and eastern U.S. also helps constrain the causes and controls of the occurrence of intraplate seismicity. For example, Zoback and Zoback [1981] argue that because the stress field in intraplate seismic areas such as the New Madrid, Missouri area is consistent with the surrounding aseismic region, localized weak zones in the crust (such as repeatedly reactivated intracratonal rifts) may be controlling the locations of intraplate seismicity rather than areas of localized stress concentration [Kane, 1976; McKeown, 1978].

The delineation of stress provinces and determining how such areas correlate with physiography, crustal thickness, seismicity, and heat flow, is ultimately an interpretative problem. Sbar [1982], for example, presents an alternative interpretation to that shown in Figure 2 of stress provinces in the western U.S. The basic problems in comparing various interpretations of stress provinces are the extreme paucity of data in many critical areas and the accuracy of data needed for defining subprovinces. As more data become available, the stress provinces shown in Figure 2 will no doubt change. As they do, it will become more and more appropriate to constrain mechanisms of plate motion, crustal deformation, and intraplate volcanism and tectonism with stress data.

Measuring the Stress Field

The past four years have seen considerable progress in the development of methods to determine the magnitude and orientation of the crustal stress field. As mentioned above, comparison of various methods for determining the orientation of the principal stresses seems to yield consistent results. Moreover, the magnitudes of stresses determined with the hydraulic fracturing and overcoring techniques have recently been found to compare quite well in several control tests [Haimson, 1982]. In fact, it is just in the past few years that there have been sufficient stress measurements made with the hydraulic fracturing method (and by enough groups of investigators) to confirm the reliability of the method and data interpretation methods [see Zoback and Haimson, 1982].

Although there is considerable inherent uncertainty in determining the orientation of principal stresses from earthquake focal mechanisms [see McKenzie, 1969; Raleigh et al., 1972], recent progress has been made in analysis of such data. First, greater accuracy can be attained by considering focal mechanisms from earthquakes generated on multiple fault planes in a given area [Zoback and Zoback, 1980]. Second, in areas where body-wave data are sparse, surface-wave focal mechanisms [Herrmann and Canas, 1978; Herrmann, 1979] seem to work quite well. Finally, formal inversion of nodal plane and slip information from earthquakes on different fault planes in a given area [Ellsworth and Xa, 1980] is yielding information on both the orientation of the principal stresses, and the ratio of the differences of the principal stresses. This inversion method is based on methods for inverting fault slip data [see Angelier, 1979], which are yielding principal stress directions that are very similar to other stress-field indicators [Zoback, 1982b].

Finally, the use of stress-induced well-bore elongation [Cox, 1970; Babcock, 1978] seems to be an exciting new method for determining the orientation of the horizontal principal stresses because either oriented 4-arm caliper logs or borehole televiewer logs can be used. The results of using this method compare well with other stress data for various regions [Bell and Gough, 1979, 1982; Hickman et al., 1982]. As there are data now available which can be used to determine stress field orientations in numerous oil and gas wells, the use of well-bore elongation for determining the orientation of the in-situ stress field may have extremely far-reaching consequences.
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