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Stress Magnitude Estimates From Earthquakes in Oceanic Plate Interiors

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We propose a method to estimate stress magnitudes in oceanic plate interiors from focal depths and focal mechanisms. Using a depth-dependent rheology, we show it is possible to estimate the differential stress ($\sigma_1 - \sigma_3$), averaged over some reference lithospheric thickness. The resolving power of the method is investigated by evaluating the effect of uncertainties in parameters that are involved in the analysis. We apply the method to the Central Indian Ocean, where intraplate seismicity is high. From well-studied earthquakes we estimate differential stresses of the order of hundreds of megapascals. This result is consistent with the high level of stress that was found from numerical model calculations by Cloetingh and Wortel (1985, 1986). From the few intraplate events in the Pacific plate, we also estimate differential stresses in this area.

INTRODUCTION

Earthquakes relax stresses within the lithosphere, and therefore information on the stress distribution at the hypocenter is contained in seismograms. The relevance of intraplate stresses in the context of plate dynamics is an important motivation to study earthquakes in the interior parts of the plates. Directions of principal stresses are generally derived from focal mechanisms, either directly from P and T axes or via the criterion of *Raleigh et al.* [1972] if one of the nodal planes can be identified as the fault plane. The general consistency of stress trajectories obtained from focal mechanism data and their good agreement with stress orientations derived from various other indicators show that such directions have a regional significance [*Zoback et al.*, 1989]. However, a more complete specification of the stress field also requires magnitude information.

We propose a method to estimate differential stress magnitudes in oceanic plate interiors. By relating focal depth and focal mechanism information to a depth-dependent rheology, it will be shown that we can estimate the differential stress level in the lithosphere where an earthquake occurs. Hence, starting from high-quality seismological data and a model for the rheology of the lithosphere (which involves several assumptions and simplifications), we derive estimates for the stress level in the lithosphere. Composition and temperature structure within the lithosphere need to be known well enough to obtain a good approximation of the strength distribution with depth. Compared with continental lithosphere, oceanic lithosphere is therefore more suited for application of the method, although no principal objections exist to using it in continental areas. In this paper we focus on estimating differential stress magnitudes in oceanic plate interiors. Well-resolved earthquake depths are a prerequisite for

the method, so that intraplate earthquakes that have been the subject of depth relocation studies provide the best data for estimating stress magnitudes in oceanic plates.

The high seismicity in the Central Indian Ocean [e.g., *Wiens and Stein*, 1983, 1984; *Bergman and Solomon*, 1984, 1985] makes this area very well suited for studying the relation between seismicity and differential stress magnitudes. In a force modeling study of the Indian plate, *Cloetingh and Wortel* [1985, 1986] calculated stresses in the Central Indian Ocean that are very high (several hundreds of megapascals) in comparison with studies by *Richardson et al.* [1979] and *Richardson* [1987, 1989]. In agreement with Cloetingh and Wortel's results, *Zuber* [1987] and *McAdoo and Sandwell* [1985] found that hundreds of megapascals stresses are required to explain the gravity highs in the Central Indian Ocean that are attributed to basement undulations as a result of lithospheric buckling. We will test whether Cloetingh and Wortel's stress results are approximately an order of magnitude too high, as suggested by *Richardson* [1987, 1989]. Therefore, at stages in the analysis when assumptions need to be made, we will adopt those assumptions which (1) are considered to be realistic and (2) give low-end differential stress estimates. Finally, we will estimate differential stress magnitudes for the Pacific Ocean from its seismicity, which is significantly lower than in the Central Indian Ocean.

RHEOLOGY

In relating the depth of an oceanic intraplate earthquake to the stress field in the lithosphere, the rheology model adopted strongly affects the inferred differential stress values. In this section we briefly describe the rheology model adopted in this paper. Brittle deformation in oceanic lithosphere is expected to occur at shallow depth. We assume that the strength ($\sigma_1 - \sigma_3$) in the brittle regime is described by *Byerlee's* [1978] law. At deeper levels deformation occurs by temperature-activated ductile flow, controlled by creep of olivine [*Goetze and Evans*, 1979; *Kirby and Kronenberg*, 1987].

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Brittle Strength

The strength of brittle rock is to a first approximation insensitive to temperature and is mainly pressure-controlled. The stress required for initiation of slip on pre-cut rock faces was found by *Byerlee* [1978] to be insensitive to rock type. A linear frictional law provides a useful approximation of brittle strength in the lithosphere, particularly at effective pressures above approximately 100 MPa, where the initial surface roughness has little or no effect on friction [*Byerlee*, 1978].

An estimate of the pore fluid pressure at the fault surface is required to calculate the brittle strength with *Byerlee's* law. Fluid pressures may range from zero to superlithostatic and are not known in general. Lacking detailed information on pore fluid pressure distributions in the seismic source regions, we assume that the pore fluid pressure is hydrostatic. As the mantle source region for the growth of oceanic lithosphere is very likely to be extremely depleted in volatile elements [*Anderson*, 1989], very little water is available for building substantial fluid pressures in sub-crustal oceanic lithosphere [*Dixon et al.*, 1988]. Ophiolitic rocks show evidence of hot water percolation in the crust, where an open system of pores is likely to exist. Therefore, fluid pressures probably are close to hydrostatic in the upper 5–10 km and subhydrostatic in the rest of the oceanic lithosphere. Our assumption of hydrostatic pore fluid pressure therefore yields a lower limit to the brittle strength and is consistent with our approach of getting low-end stress estimates, within the limits of (what we consider to be) realistic assumptions.

The brittle strength ($\sigma_1 - \sigma_3$) depends on which of the principal stresses is vertical. We adopt the convention that compression is positive and that $\sigma_1 \geq \sigma_2 \geq \sigma_3$. If σ_3 is vertical, the resistance of preexisting faults to sliding is largest ("compression" curve in Figure 1). If σ_1 is vertical the brittle strength is smallest ("tension" branch in Figure 1). If both σ_1 and σ_3 are horizontal, the brittle strength may vary between these two extremes [*Turcotte and Schubert*, 1982; *Sibson*, 1974].

Ductile Rheology

Recent laboratory data have modified the depth-dependent strength model of *Goetze and Evans* [1979] only slightly, the main difference being the incorporation of "wet" deformed olivine, i.e., olivine containing trace amounts of water [*Chopra and Patterson*, 1981; *Tsenn and Carter*, 1987]. Steady state flow

laws for olivine are now well established under both anhydrous conditions and in the presence of water. The creep stress for olivine under conditions that favor dislocation processes may be represented by a powerlaw. Under conditions of low temperatures and high stress the powerlaw breaks down to exponential

Fig. 1. Relationship between focal depth of an intraplate earthquake and the stress distribution. (a) Three-dimensional image showing how Figures 1b and 1c are built up by stacking overlapping surfaces that have been calculated for various assumptions on brittle strength and strain rate. (b) Plan view of top surface of Figure 1a. Various strength envelopes are shown for oceanic lithosphere (age 53 Ma). Brittle strength branches labeled "compression," "intermediate," and "tension" are calculated for various principal stress orientations from *Byerlee's* [1978] law with hydrostatic pore pressure. The ductile strength (flow stress) is shown for strain rates of 10^{-15} s $^{-1}$ and 10^{-17} s $^{-1}$. The differential stress ($\sigma_1 - \sigma_3$) has reached the strength locally at the focal depth of 30 km (indicated by arrow), where ductile behavior is predicted. The depth-dependent stress distribution ($\sigma_1 - \sigma_3(z)$) is found by assuming that the stress in weaker parts of the lithosphere has reached the strength, whereas stronger parts elastically support a stress that is (at least) equal to the stress at the focal depth. Shaded surfaces are a measure of the (average) differential stress (equation (1)). The sensitivity of the average differential stress to assumptions on brittle strength and strain rate is shown by areas of different grey tones. (c) Earthquake with focal depth at 10 km, where brittle behavior is expected. The sensitivity to strain rate is seen to be relatively small for a brittle event. Both ductile and brittle earthquakes are very sensitive to the selection of one of the brittle branches.

behavior [Goetze and Evans, 1979; Tsenn and Carter, 1987]. In the present study we used a powerlaw flow law for "wet" olivine with an average grain size of 1 mm and an exponential law with data from Tsenn and Carter [1987]. As argued in our discussion on pore fluid pressures, we do not actually expect the oceanic lithosphere to be wet but use the present flow law in order to get a lower bound on the stress magnitude.

Oceanic geotherms are a function of lithospheric age. Temperature profiles in oceanic lithosphere are calculated using Crough's [1975, 1977] boundary layer model with a constant basal heat flux and a temperature at the base of the lithosphere of 1300°C. The basal heat flux is selected on the basis of a fit to bathymetry and heat flow data [Parsons and Sclater, 1977]. Lithospheric age is estimated from the magnetic anomaly map of Larson et al. [1985] and the geomagnetic time scale of Harland et al. [1982].

FOCAL MECHANISMS

Constraints on Brittle Strength From Vertical Slip Components

If a vertical cross section of brittle rock, deforming by plane strain, is subjected to a horizontal in-plane stress, faults exhibit normal slip in response to tensile stress and reverse slip due to compressive stress. As the brittle strength is weaker in tension than in compression, one could infer whether to use the "tension" or "compression" branch (Figure 1) from the vertical slip component. However, if stresses and strains are three-dimensional, the relation between vertical slip components and brittle strength is unclear.

The minimum shear stress required for initiation of slip on preexisting faults is described by Byerlee's [1978] law, whereas new faults will be created at higher shear stresses described by Coulomb's [1773] law. Byerlee's law predicts the most favorable fault plane orientation as well as the minimum shear stress that is required for initiation of slip [see Turcotte and Schubert, 1982]. However, if this particular fault is not present, the strength of the system is higher. Preexisting faults that are less favorably oriented might become activated at higher stress magnitudes.

We have used a three-dimensional numerical model, employing Byerlee's and Coulomb's laws, to determine bounds on the brittle strength from vertical slip components (Appendix A). We conclude that if the focal mechanism solution indicates a reverse slip component, the brittle strength is limited by "compression" and "intermediate" branches in Figure 1. In case of a normal slip component, the brittle strength ranges from "intermediate" to "tension". Strike-slip faulting does not provide any constraint on the brittle strength.

The sign of the vertical slip component is identical on both nodal planes (see Appendix B). Therefore, putting limits on the brittle strength by using vertical slip components does not depend on a proper selection of fault plane and auxiliary plane.

Earthquakes in Ductile Lithosphere

It will become clear in the next section where we explain how to infer differential stresses from earthquakes, that even if events occur in ductile rock, stress estimates would be more accurate if we could limit the range of brittle strengths in the same way we did for brittle earthquakes. Several models have been proposed to explain the occurrence of seismic slip at depths where crystal-plastic processes are thought to occur (we will refer to these earthquakes as "ductile events"). Sibson [1980] suggests that brittle frictional sliding within the ductile regime may occur due

to compositional differences. Variation of unstable to stable frictional slip on deeply penetrating faults is suggested by Tse and Rice [1986] to explain the occurrence of ductile earthquakes. Instability caused by injection of fluids into shear zones and fault gouges, for instance by dehydration reactions, has been put forward by Raleigh and Paterson [1965] and analyzed by Shimamoto [1985]. Finally, creep instability, i.e., catastrophic strain softening due to an increase in strain rate or temperature [Orowan, 1960; Griggs and Baker, 1968; Hobbs et al., 1986; Ogawa, 1987], has been put forward to explain ductile earthquakes. The seismic radiation pattern of most ductile earthquakes cannot be distinguished from that of a brittle shear fracture. Currently, it is impossible to select one of the proposed models on the basis of seismic or rheological data.

In the previous section we discussed how the vertical slip component of a brittle earthquake can be used to tighten the range of brittle strengths that are used to estimate upper and lower bounds to the differential stress. The well-established mechanism of brittle earthquakes was a principal ingredient in this discussion. For ductile earthquakes the mechanism is unknown, so that we consider it unwarranted to attribute significance to the vertical slip component inferred from a double couple representation of the source. Therefore, if a seismic event occurs at a depth where rocks are deforming ductily, we take the full range of brittle strengths into account to estimate the differential stress ("tension" to "compression" branches in Figure 1).

DIFFERENTIAL STRESS

The method is designed to yield a measure of stress magnitude that is directly comparable with results from modeling studies. Principal stress magnitudes inferred from modeling generally constitute averages over some (elastic) reference thickness L_{ref} (in the present study we used $L_{ref}=100$ km). This reference thickness has no physical meaning and is only used to facilitate comparison of stresses derived from force modeling studies with stresses estimated from earthquakes. The average differential stress is defined as

$$\overline{(\sigma_1 - \sigma_3)} = \frac{1}{L_{ref}} \int_0^L (\sigma_1 - \sigma_3) dz \quad (1)$$

L is the thermally defined, age dependent thickness of the lithosphere, i.e., the depth of the 1300°C isotherm in Crough's [1975; 1977] model. Bending stresses in the interior parts of the plate are assumed to be negligible.

Method Description

To calculate the average differential stress at an epicentral site, we note that the strength at the focal depth has been exceeded locally (see Figure 1). No major strength discontinuities occur within oceanic lithosphere, so it is a reasonable assumption that stresses are distributed evenly over the lithospheric thickness. Weak parts of the lithosphere reach their strength at low stresses, stronger parts deform elastically until stresses increase sufficiently to cause deformation either by flow or by fault slip. We conclude that the strength curve provides an upper limit for differential stresses at depths where the strength is less than that at the focal depth. The differential stress in the "elastic core", i.e., the strong part of the lithosphere where stress is supported elastically, is at least equal to the stress level at the focal depth [Wortel, 1986; Wortel and Vlaar, 1988]. After integration of the depth-dependent differential stress and scaling by L_{ref} we obtain

Fig. 2. Differential stress as a function of age and focal depth, for a strain rate of 10^{-16} s^{-1} : (a) for maximum brittle strength in case of reverse slip or strike slip mechanism, (b) for minimum brittle strength in case of normal slip or strike slip mechanism, (c) difference between maximum and minimum stress, a measure for method accuracy, if no constraints on the vertical slip components are available, (d) for maximum brittle strength in case of a normal slip event, and (e) difference between maximum and minimum stress (minimum in Figure 2b). Note that the the lower ductile part is not affected by the normal slip constraint, (f) for minimum brittle strength in case of reverse slip, and (g) difference between maximum and minimum stress for focal mechanism with a reverse slip component (maximum differential stress in Figure 2a).

$(\sigma_1 - \sigma_3)$. In the following we will refer to the average differential stress as "differential stress".

Sensitivity of Results to Input Parameters

Our results are sensitive to two different types of parameters: variables related to the method, that cause uncertainties in the rheology model, and "observational" parameters, like focal depth and lithospheric age. Relevant method-related parameters affecting the depth-dependent rheology model are the pore fluid pressure, the assumed strain rate, and the brittle strength.

The effect of decreasing the pore fluid pressure is that the brittle strength and, therefore, our stress estimates increase. As discussed before, we aim at getting a lower bound to differential stresses, within the limits of (what we consider to be) realistic assumptions. We therefore did not include pore pressures smaller than hydrostatic in our calculations.

Figure 1 displays the sensitivity of differential stress to brittle strength and strain rate for a ductile and a brittle event. In Figure 1, hatched areas are a measure of differential stress. The effect of varying the strain rate between 10^{-15} s^{-1} and 10^{-17} s^{-1} can be significant for ductile earthquakes (Figure 1b). The stress in the elastic core is strongly affected, so that the differential stress is sensitive to a change in strain rate. From Figure 1c it is clear that stresses calculated from events in the brittle part of the lithosphere are relatively insensitive to changes in strain rate.

Both for brittle and for ductile earthquakes the differential stress is very sensitive to the selected brittle strength. In the absence of constraints on the vertical slip component, upper and lower bounds to the differential stress are calculated from "compression" and "tension" strength branches, respectively. Figures 2a and 2b show upper and lower bounds as a function of focal depth and lithospheric age, for a strain rate of 10^{-16} s^{-1} . A measure of the resolution that can be obtained with our method is the difference between upper and lower bounds to the differential stress (Figure 2c). In older lithosphere we observe differences of several hundreds of megapascals between minimum and maximum differential stresses. Figure 2c can be used to estimate what can be gained by improving the focal depth accuracy after an approximate focal depth range has been determined: an estimated earthquake depth of $20 \pm 10 \text{ km}$ in 90 Ma lithosphere does not require additional waveform modeling to resolve differential stresses much better. However, if the observed focal depth is $40 \pm 10 \text{ km}$, it is worthwhile to try to improve the depth accuracy.

If the focal mechanism has a normal slip component the upper bound to the differential stress is not calculated from the "compression" brittle strength but from the "intermediate" strength branch (Figure 1). The lower bound to the differential stress again follows from using the "tension" brittle strength (Figure 2b). Comparing Figures 2d and 2a, we observe a significant reduction of the maximum differential stress for those focal depths that are in the brittle lithosphere. The exclusion of focal mechanism information for ductile events results in a discontinuity of calculated stresses at the brittle-ductile transition. The reduction of the maximum differential stress clearly improves the resolution of differential stress estimates for brittle events, as can be seen from comparing Figures 2e and 2c.

In case of reverse slip the "intermediate" instead of "tension" branch is used to calculate the minimum differential stress. The upper bound to the differential stress follows from the "compression" brittle strength (Figure 2a). Figures 2f and 2g show minimum stress and stress resolution as a function of age

and focal depth, assuming that the fault slip has a reverse component. Again, reduction of the range of brittle strengths improves the resolution significantly.

DIFFERENTIAL STRESS IN THE CENTRAL INDIAN OCEAN

The Central Indian Ocean, with its high intraplate seismicity, provides an excellent opportunity to study the relation between earthquakes and the level of intraplate stress. An important motivation to study intraplate stress fields is to improve our understanding of the dynamics of plate motion. With this goal in mind Richardson *et al.* [1979] performed numerical calculations of the stress field in all major plates. Subsequently, Wortel and

Fig. 3. Differential stresses in the Central Indian Ocean at epicentral locations of events listed in Table 1. Thin solid lines from the base of the columns end at epicentral locations. The height from the base of a column is a measure of the differential stress magnitude. Stress magnitudes constitute averages over a reference value (100 km) of the plate thickness. In each column, the darkest grey band corresponds to the inferred differential stress. The uncertainty increases for most events if the vertical slip component is neglected; in this case the light grey band plus the dark grey band correspond to the range of inferred differential stresses. The thick horizontal line in (or at the top of) the column corresponds with the differential stress calculated by Cloetingh and Wortel [1985, 1986].

Cloetingh [1981, 1983, 1985] performed a similar type of numerical modeling, with some important new aspects in their force modeling procedure. In contrast with the earlier models the ridge push was not represented as a line force but as an integrated pressure gradient [*Lister*, 1975] distributed over all contributing parts of the lithosphere. The slab pull was incorporated as an age-dependent force giving rise to significant lateral variations in the forces representing the subduction process. Whereas the force modeling for the Nazca plate [*Wortel and Cloetingh*, 1983, 1985] yielded stress values in agreement with *Richardson et al.*'s [1979] models, a later application of the same modeling to the Indo-Australian plate [*Cloetingh and Wortel*, 1985, 1986] resulted (Figure 3) in a significantly higher stress level than in the study by *Richardson et al.* [1979] (see also *Richardson* [1987, 1989]). This difference stems from a difference in modeling procedure. *Richardson et al.* [1979] made a variable parameter study of the various plate tectonic forces involved. *Cloetingh and Wortel* [1985, 1986] treated the driving forces (ridge push and slab pull) as known forces, which could be calculated from kinematic parameters (relative motion along convergent plate boundaries) and the age of the lithosphere involved, and derived the magnitudes of the resistive forces from the equation representing the balance of torques of all forces acting on the plate.

In view of the relevance of such discrepancies for ongoing and

future numerical modeling we address this issue by investigating the information which the unusually high seismic activity in the Central Indian Ocean [e.g., *Wiens and Stein*, 1983, 1984; *Bergman and Solomon*, 1984, 1985] provides concerning the level of intraplate stress. Previously, only the orientations of the principal stress directions have been used to test the results of stress modeling. For the Central Indian Ocean, good agreement was found between calculated orientations and focal mechanism data [*Cloetingh and Wortel*, 1986; *Bergman*, 1986]. Also the orientations of observed long wavelength basement undulations appear to be consistent with the calculated stress orientations [*Stein et al.*, 1989; *Petroy and Wiens*, 1989].

Large ($m_b \geq 4.9$) oceanic intraplate earthquakes in the Central Indian Ocean have been the subject of seismological waveform modeling studies by various authors. The earthquakes for which adequate information on depth and focal mechanism were available are listed in Table 1. The focal depth accuracy for shallow teleseismic earthquakes is discussed for various methods by *Stein and Wiens* [1986]. Depth relocation methods like long-period *P* waveform modeling [*Wiens and Stein*, 1983], Rayleigh waveform modeling with additional long-period *P* waveform modeling for shallow events [*Wiens and Stein*, 1984] and body waveform inversion [*Bergman and Solomon*, 1984, 1985] all are estimated to have a 2-km accuracy. Pre-WWSSN data, relocated with *P*, *SH*, Rayleigh and Love waveform modeling techniques

TABLE 1. Central Indian Ocean Intraplate Earthquake Data

Event ^a	Date	Latitude °N	Longitude °E	Depth ^b , km	Vertical Component ^c	Age, Ma	References ^d
7	May 25, 1964	-9.1	88.9	17 ± 2	SS	58 ± 2	BS85
8	October 31, 1965	-14.2	95.3	24 ± 2	SS	65.5 ± 0.5	WS83
9	September 14, 1968	-24.4	80.4	3 ± 2	SS	31 ± 3	BS85
				4 ± 2	T		BS85
10	October 10, 1970	-3.6	86.2	27 ± 2	SS	73 ± 9	BS85
				39 ± 2	T		BS85
11	June 26, 1971	-5.2	96.9	29 ± 2	T	52.5 ± 0.5	BS85
12	November 24, 1972	11.7	85.3	27 ± 2	T	116 ± 14	BS85
13	April 7, 1973	7.0	91.3	13 ± 2	SS	75 ± 7	BS85
				6 ± 2	SS		BS85
14	August 30, 1973	7.2	84.3	27 ± 2	T	109 ± 7	BS85
15	June 25, 1974	-26.0	84.3	23 ± 2	T	42.5 ± 3.5	BS85
				18 ± 2	T		BS85
				13 ± 2	T		BS85
16	August 3, 1978	-0.9	84.2	39 ± 2	T	80.5 ± 2.5	WS83
17	December 2, 1981	-15.7	88.4	23 ± 2	T	50 ± 5	BS85
18	April 23, 1967	1.6	80.2	32 ± 5	T	83 ± 6	SW90
19	June 3, 1978	-16.2	93.0	11 ± 5	T	65 ± 3	SW90
20	October 8, 1968	-39.9	87.7	11 ± 2	N	7.5 ± 2.5	WS84
21	November 2, 1976	-29.4	77.7	14 ± 2	N	9 ± 2	WS84
22	March 31, 1970	-3.8	69.7	12 ± 2	N	8.5 ± 2.5	WS84
23	September 19, 1975	-34.7	81.9	19 ± 2	N	20.5 ± 3.5	WS84
24	November 17, 1973	-1.6	69.8	19 ± 2	N	13 ± 2	WS84
25	December 12, 1981	5.0	70.2	15 ± 2	N	46 ± 7	BS84
26	October 2, 1957	-6.3	69.7	14 ± 5	SS	12 ± 3	W86
27	Februari 29, 1944	0.3	75.4	10 ± 5	T	76 ± 3	W86
45	December 19, 1965	-32.2	78.9	13 ± 2	N	5.5 ± 1.5	WS84
46	Februari 17, 1966	-32.2	78.9	11 ± 2	N	5.5 ± 1.5	WS84
47	April 25, 1970	-6.3	69.8	15 ± 2	N	12 ± 3	WS84
48	November 30, 1983	-6.8	72.0	16 ± 2	N	49 ± 7	WS84
49	September 12, 1965	-6.5	70.8	18 ± 2	N	30 ± 12	BS84
51	November 10, 1967	-6.0	71.3	18 ± 2	N	36 ± 11	BS84
52	November 11, 1967	-6.0	71.4	17 ± 2	N	36 ± 11	BS84
53	March 2, 1968	-6.1	71.4	13 ± 2	N	36 ± 11	BS84

^a Numbers 7-27 refer to *Cloetingh and Wortel* [1986], numbers 45-53 have been added to their list.

^b Depth from top of crust.

^c SS, strike slip; T, thrust; N, normal.

^d BS85, *Bergman and Solomon* [1985]; WS83, *Wiens and Stein* [1983]; SW90, *Stein and Weissel* [1990]; WS84, *Wiens and Stein* [1984]; W86, *Wiens* [1986]; BS84, *Bergman and Solomon* [1984].

[Wiens, 1986] have 5-km accuracies. Depth phase identifications from short-period *P/pwP/pP* arrivals [Stein and Weissel, 1990] are assigned a 5-km accuracy. Vertical slip components are considered significant if the slip direction makes an angle with the strike of more than 10°.

For all these events the differential stress is estimated, first without focal mechanism information taken into account and next with inclusion of such information (see Table 1). The results are displayed in Figure 3 in combination with the numerical model values obtained by *Cloetingh and Wortel* [1985, 1986]. Figure 3 clearly demonstrates what was already apparent from Figures 2c, 2e and 2g, namely, that hypocentral depth alone does not provide a firm constraint on the level of differential stress. If, however, reliable information on the vertical slip components of an event is available, the large contribution of the brittle slope variation to the uncertainty of the integrated differential stress can be reduced.

Gravity highs attributed to 200-km wavelength basement undulations as a result of lithospheric buckling have been observed in the region between 18°S and 10°N and 80°E and 100°E. Earthquakes 7, 10, 16, and 18 all occur near highs, so that the effect of flexural bending stresses in this compressive region would be to reduce stresses at the top and increase stress near the bottom of the lithosphere. Differential stress estimates in this region are very high, and the contribution of superimposed flexural bending stresses to the average differential stress generally is not large. More particularly, the stress found from event 7 probably is a lower estimate, and the stress from event 16 is estimated too high. Stresses from the other events are hardly affected by bending stresses, since their epicenters are in the flanks of the buckles. Bathymetric loading by the Chagos-Laccadive Ridge and Ninetyeast Ridge also might cause bending stresses. These would affect the stresses estimated from event 7, 17, 26, and 47-53.

Comparison of *Cloetingh and Wortel's* [1985, 1986] numerical model values (horizontal bars) with the differential stress ranges in Figure 3 indicates that for events 22, 24, 26, 45, 46, and 47, all near the ridge, the model values are somewhat too high. Only for events 11 and 13 the model values greatly exceed the inferred stress ranges. These events are near the Sumatra-Nicobar-Andaman trench system. All model values for the Central Indian Ocean (with the minor exceptions of one of the events of 10, and 19) fall clearly within the stress ranges. The model values for events 15, 18, and 27 are even near the minima of the seismicity derived stress ranges. We particularly note the good agreement for the high stress values of events 12 and 14.

Since our method yields low-end estimates of the differential stress, our results are more in agreement with those of *Cloetingh and Wortel* [1985, 1986] than with stress levels found by *Richardson* [1987, 1989], which are about an order of magnitude lower.

PACIFIC PLATE STRESSES

Seismicity in the Pacific plate is less abundant than in the Indian plate and focal depth and focal mechanism constraints are relatively poor compared to Indian Ocean data. Table 2 is a selection of $m_b \geq 4.8$ events we made from World Stress Map data. Figure 4 shows results of differential stress computations from focal depth and focal mechanism.

The highest stresses are estimated for the Samoa-Gilbert-Ralik area from event 6, 7, and 8. *Okal et al.* [1986] note teleseismically recorded swarms of intraplate seismicity ($4.0 \leq m_b \leq 6.0$) in this region and attribute them to large-scale deformation. However, nearby seamounts probably cause bending stresses so that differential stresses derived from these events do not reflect the intraplate stress field.

Stein [1979] studied event 3 in the northwest of the Pacific and concluded that the epicenter occurred on Emperor Trough, a dead

TABLE 2. Pacific Intraplate Earthquake Data

Event	Date	Latitude °N	Longitude °E	Depth ^a , km	Vertical Component ^b	Age, Ma	Reference ^c
1	March 6, 1965	-18.4	-132.8	5 ± 5	N	34 ± 1	O80
2	September 18, 1966	-18.4	-132.9	5 ± 5	N	34 ± 1	O80
3	April 28, 1968	44.8	174.6	10 ± 5	T	90 ± 10	S
4	July 29, 1968	-7.5	-148.3	5 ± 5	SS	89 ± 5	O80
5	August 6, 1969	-7.6	-148.1	5 ± 5	N	89 ± 5	O80
6	Januari 7, 1982	-3.4	177.6	15 ± 3	T	122 ± 3	LO, D83
7	Februari 15, 1982	-3.5	177.5	21 ± 3	SS	122 ± 3	LO, D83
8	March 16, 1982	-3.3	177.5	15 ± 3	T	122 ± 3	LO, D83
9	Januari 19, 1973	-7.6	-148.2	5 ± 5	N	89 ± 5	O80
10	Januari 19, 1973	-7.6	-148.1	5 ± 5	N	89 ± 5	O80
11	April 26, 1973	20.1	-155.2	45 ± 2	N	90 ± 5	B86, B, U
12	May 25, 1975	-18.4	-133.0	5 ± 5	SS	34 ± 1	O80
13	August 30, 1976	1.0	147.6	26 ± 2	SS	34 ± 3	B86, WS83
14	October 29, 1975	4.1	-103.5	9 ± 2	N	3 ± 1	WS84
15	July 31, 1983	-20.1	-126.9	17 ± 2	N	22 ± 2	WS84, D84, O84
16	June 30, 1945	16.6	-115.8	12 ± 2	N	16 ± 1	WO
17	August 20, 1968	5.4	147.1	6 ± 3	T	34 ± 3	B86, BS
18	October 5, 1984	20.1	-116.0	13 ± 2	N	18 ± 1	WO, D85
19	December 2, 1984	20.4	-115.8	15 ± 2	N	18 ± 1	WO, D85
20	May 28, 1986	20.0	-115.9	15 ± 2	N	18 ± 1	WO, D87
21	Januari 5, 1978	-20.9	-126.9	5 ± 5	T	23 ± 1	O80
22	July 25, 1978	-20.8	-126.9	5 ± 5	T	23 ± 1	O80

^a Depth from top of crust.

^b SS, strike slip; T, thrust; N, normal.

^c B86, *Bergman* [1986]; BS, *Bergman and Solomon* [1980]; B, *Butler* [1982]; D83, *Dziewonski et al.* [1983]; D84, *Dziewonski et al.* [1984]; D85, *Dziewonski et al.* [1985]; D87, *Dziewonski et al.* [1987]; LO, *Lay and Okal* [1983]; O84, *Okal* [1984]; O80, *Okal et al.* [1980]; S, *Stein* [1979]; U, *Unger and Ward* [1979]; WO, *Wiens and Okal* [1987]; WS83, *Wiens and Stein* [1983]; WS84, *Wiens and Stein* [1984].

Fig. 4. Differential stresses in the Pacific intraplate area at epicentral locations of events listed in Table 2. Columns have the same meaning as in Figure 3.

spreading center that has also been active as a transform fault between Kula and Pacific plates [Larson and Chase, 1972; Hilde *et al.*, 1976]. Due to the large uncertainties in lithospheric age and focal depth, the difference between maximum and minimum estimated stresses is large. The Caroline plate, north of New Guinea, has been proposed by Weissel and Anderson [1978] to be subducting under the Pacific plate at the Mussau Trench. The epicenter of event 17 lies on the proposed Caroline-Pacific plate boundary and might not be a true intraplate event.

The magnitude of intraplate stresses in the Pacific plate east of 170°W appears to be lower than in the west. The strongest constraint on intraplate stresses in the southeast Pacific comes from event 15. Other earthquakes in this area have less well resolved focal depths. Differential stresses southwest of Baja California also indicate that differential stresses in the eastern Pacific do not exceed the 100 MPa level. Earthquakes in the eastern Pacific that are related to bending stresses due to seamount loads are event 1, 2, and 12 [Okal *et al.*, 1980] and event 11 near Hawaii [e.g., Watts *et al.*, 1985].

DISCUSSION

Strength on Different Time Scales

Jeffreys [1959] observed that high, uncompensated mountain ranges can be supported by the lithosphere and inferred that the strength of the lithosphere must be hundreds of megapascals. Laboratory experiments on both brittle and ductile rocks showed that the lithosphere is indeed capable of bearing stresses of large

magnitude on long time scales [Byerlee, 1978; Goetze, 1978]. It was noted by Chinnery [1964] that the stress drop in an earthquake is 2 orders of magnitude lower. Thus, if the stress drop is taken to be a measure of strength, the strength on short time scales ("seismic strength") seems to be much lower than the strength on geological time scales ("tectonic strength"). The question arises whether we can use long-term strength envelopes to infer stress magnitudes from earthquakes? While further insight into this aspect has to be gained, we adopt the hypothesis concerning the relation between stress, strength (rheology), and earthquake generation proposed by Wortel [1986]. Recognizing that parts of the lithosphere in which the strength is low and where the stress is at or near the strength (comparable with the near-surface part of the oceanic lithosphere) earthquake generation is not observed, he postulated that seismic activity occurs when and where the width (or depth interval) of the anelastically deforming region increases at the expense of the "elastic core", in other words, when and where the stress reaches the strength for the first time. It was shown that this hypothesis adequately accounts for the distribution of seismic activity in subducting lithosphere [Wortel, 1986; Wortel and Vlaar, 1988]. In this hypothesis the seismic stress drop does not reflect the absolute level of stress but rather a stress adjustment to the equilibrium tectonic stress. Consequently, the seismic strength can be higher than the tectonic strength and earthquakes are envisaged to relax stresses that exceed the long-term strength. The stress drops involved are low, so the long-term strength envelope yields a good approximation of the stresses at the hypocenter.

Nonequilibrium Stress Distribution

In our analysis we assumed that ductile rocks in the "elastic core" are able to support stresses elastically on geologic time scales. If, however, ductile rock is subject to differential stresses it will show some elastic deformation, transient creep and eventually steady state creep. In steady state, the strain rate has equilibrated to the imposed stress. Therefore stress distributions like in Figure 1 cannot represent equilibrium, since ductile rock will eventually support a stress that is equal to the flow stress (the stress we previously named "strength") at some strain rate.

It is clear that the time required for establishing an equilibrium is a (highly nonlinear) function of boundary conditions and rheology. Probably, the stresses in the strongest layers will not change very rapidly with time, since their viscosity is very high. Therefore, the "elastic core" can be interpreted as a depth interval where creep in the ductile rocks very slowly alters the stress. We note that there is a great need for a dynamic approach to investigate the stress distribution with depth as it develops with time, if lithosphere is subjected to a stress boundary condition.

Brittle Strength

Our assumptions are aimed at getting a low-end estimate of the differential stress in oceanic lithosphere. However, the brittle strength inferred from Byerlee's [1978] relation is thought to be exceptionally high by a number of authors [Haimson and Doe, 1983; Zoback and Mastin, 1986]. In spite of the possible shortcomings we consider Byerlee's law to be the most suitable quantitative description available at present. Most likely, deviations from the assumed brittle strength are most pronounced near the brittle-ductile transition, and our method is not very sensitive to them.

Focal Depth

Many of the reported depths in Tables 1 and 2 are centroid depths. The centroid depth is the average depth of the fault plane and is therefore not necessarily equal to the initiation depth, the hypocenter. The difference in depth between hypocenter and centroid can be up to 5 km, depending on the wavelengths that were used in determining the centroid depth, the spatial source function asymmetry, the fault plane orientation, and the fault plane dimensions. In our analysis we assumed that hypocentral depth and centroid depth are equal.

CONCLUSIONS

The main objective of the current paper is to investigate the possibility of extending our knowledge of the intraplate stress field with magnitudes. Our present knowledge of lithospheric rheology required to relate intraplate deformation (earthquakes) to stress limits the accuracy of inferred stress magnitudes. Therefore, efforts to make the accuracy of focal depths better than 2 km are not expected to contribute to our knowledge of intraplate stress magnitudes. If the vertical slip component is not well constrained by the focal mechanism, if the faulting is pure strike-slip, or if an earthquake occurs in ductile lithosphere, the uncertainty in differential stresses is large. If the focal mechanism is known with confidence, tighter bounds can be placed on the stress magnitude.

Numerical calculations of the stress field in the Central Indian Ocean by Richardson *et al.* [1979] and Cloetingh and Wortel [1985, 1986] yield order of magnitude differences. From our

analysis we find that seismicity data require a stress level comparable with that calculated by Cloetingh and Wortel [1985, 1986]. Stress levels about an order of magnitude lower, as advocated by Richardson [1987, 1989] are less in agreement with differential stresses found in this study.

APPENDIX A

We investigate the relation between the vertical slip component of an earthquake and the brittle strength. Obviously, both are related by the stress field; slip occurs in the direction of maximum resolved shear stress if the strength is exceeded.

In general, the brittle strength according to a linear friction law can be written in terms of the difference between maximum and minimum principal stresses

$$(\sigma_1 - \sigma_3) = \frac{2S_0}{\sqrt{1+\mu^2}} + \frac{2\mu(q-\lambda)}{\sqrt{1+\mu^2}} \sigma_v$$

$$q = \frac{\sigma_1 + \sigma_3}{2\sigma_v}, \quad \frac{\sqrt{1+\mu^2} + \lambda\mu}{\sqrt{1+\mu^2} + \mu} \leq q \leq \frac{\sqrt{1+\mu^2} - \lambda\mu}{\sqrt{1+\mu^2} - \mu}$$

cohesion S_0 , coefficient of friction μ , overburden load σ_v , and pore fluid coefficient $\lambda = P_f/\sigma_v$. We assume that one of the principal stress directions is vertical and equal to the overburden pressure. Upper and lower limits to the brittle strength and q correspond to in-plane compression and in-plane tension, respectively, or more precisely, correspond to vertical σ_3 and σ_1 . Therefore, if we would know the principal stress directions at the epicenter, we could select a single brittle strength curve to estimate differential stress. Unfortunately, focal mechanism solutions only give principal deformation quadrants, not principal stress directions.

Fig. A1. Vertical fault slip component as a function of horizontal stresses σ_x and σ_y . The grey shaded areas indicate stresses at which slip along preexisting faults can take place. At lower stresses, in the internal part of the figure, the deformation is purely elastic. The failure criterion puts an upper bound to the stress level that can be reached. In the light grey shaded area the predominant vertical component is reverse slip. In the dark grey area normal slip is observed. The shape of the figure is invariant to depth; the figure simply shrinks or expands at smaller or greater depths (numbers on the axes correspond to a depth of 13 km).

To determine the relation between focal mechanism and applied stress, we performed a number of synthetic tests. For various three-dimensional stress fields we calculated the orientations of preexisting fault planes on which the conditions for slip are met; that is, the threshold value of shear stress has been reached or exceeded according to Byerlee's law. Horizontal stresses are limited by the failure strength of brittle rock, which was calculated by *Coulomb's* [1776] law.

Figure A1 shows a typical result of our modeling. Three basic response types are indicated; if horizontal tectonic stresses (σ_x and σ_y) are small deformation is elastic. If stresses increase slip on preexisting faults (if present) may occur. If stresses cannot be accommodated on preexisting faults, new ones may be created at even higher stresses. Faulting types typical for particular stresses may be recognized [Anderson, 1951; Sibson, 1974]; if both horizontal stresses are compressive mainly reverse faults become activated. Normal faulting typically occurs in response to horizontal tensile stresses. The switch of slip with normal component to slip with reverse component occurs approximately midway the transitional domain where σ_2 is vertical.

From our synthetic experiments we conclude that if a focal mechanism has a significant reverse component, the brittle strength is bounded by "compression" and "intermediate" curves in Figure 1. If a normal slip component has been observed, the brittle strength lies between "tension" and "intermediate" curves. A mechanism with a pure strike-slip character does not give any constraints on the brittle strength, so that the full strength range has to be taken into account.

APPENDIX B

Given the orientations of unit length **T**, **P**, and **B** axes we will show that the vertical slip component on both nodal planes has the same sign. Let **X** be north, **Y** east, and **Z** vertically downward. To specify the orientation of a particular vector, we use angles ϕ and δ . ϕ is the angle between the projection of the vector on the horizontal **X-Y** plane and the **X** axis, measured clockwise. δ is the clockwise angle between the horizontal projection of the vector and the vector itself. **P**, **T**, and **B** axes are typically given in lower hemisphere stereographic coordinates, so that $\delta \in [0, \pi/2]$ and $\phi \in [0, 2\pi]$.

We define a matrix **R** that rotates **X** to the **B** axis (ϕ_B, δ_B):

$$R = \begin{bmatrix} \cos\phi_B \cos\delta_B & -\sin\phi_B & -\cos\phi_B \sin\delta_B \\ \sin\phi_B \cos\delta_B & \cos\phi_B & -\sin\phi_B \sin\delta_B \\ \sin\delta_B & 0 & \cos\delta_B \end{bmatrix}$$

If we let R^{-1} work on **T** axis (ϕ_T, δ_T) and **P** axis (ϕ_P, δ_P), both $P' = R^{-1}P$ and $T' = R^{-1}T$ will lie in the **Y-Z** plane. We now define a third angle γ , to specify the angle in the **Y-Z** plane between the P' or T' axis and **Z**. We consider two cases: T' closest to the vertical, i.e., $\gamma_T \in [-\pi/4, \pi/4]$, and P' closest to the **Z** axis.

If T' is closest to the vertical in the system of north directing **B** axis, we define a second rotation matrix

$$V = \begin{bmatrix} 1 & 0 & 0 \\ 0 & \cos\gamma_T & -\sin\gamma_T \\ 0 & \sin\gamma_T & \cos\gamma_T \end{bmatrix}$$

so that $T'' = R V T'$. T'' is the vertical unit vector. Slip vectors in the system of vertical T'' axis and north directing B' axis are $S'_1 = (0, \frac{1}{2}\sqrt{2}, -\frac{1}{2}\sqrt{2})$ and $S'_2 = (0, -\frac{1}{2}\sqrt{2}, -\frac{1}{2}\sqrt{2})$. Slip vectors in the original system of **P**, **B**, and **T** axes can be calculated from

$$S_1 = R V S'_1$$

$$S_2 = R V S'_2$$

Evaluation of the vertical slip components in the original system of **P**, **B**, and **T** axes yields

$$S_1^z = -\frac{1}{2}\sqrt{2} \cos\delta_B (\cos\gamma_T - \sin\gamma_T)$$

$$S_2^z = -\frac{1}{2}\sqrt{2} \cos\delta_B (\cos\gamma_T + \sin\gamma_T)$$

for $\gamma_T \in [-\pi/4, \pi/4]$. For the given range of γ_T , both vertical slip components have the same sign.

By the same approach it can be shown that if the P' axis is closest to the vertical in the system of north directing **B** axis we have

$$S_1^z = \frac{1}{2}\sqrt{2} \cos\delta_B (\cos\gamma_P - \sin\gamma_P)$$

$$S_2^z = \frac{1}{2}\sqrt{2} \cos\delta_B (\cos\gamma_P + \sin\gamma_P)$$

for $\gamma_P \in [-\pi/4, \pi/4]$, so that both vertical slip components have either dip-slip or thrust components.

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