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THE NATURE OF DEEP-FOCUS EARTHQUAKES

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INTRODUCTION

Wadati (1928) first proved convincingly that some earthquakes occur at depths well beneath the Earth's crust. He essentially proposed the nomenclature used today by the International Seismological Centre (ISC), i.e. earthquakes with focal depths exceeding 300 km are deep earthquakes, and those with depths between 70 and 300 km are intermediate earthquakes. However, when there is little reason to distinguish between the two groups, both types are called simply deep earthquakes. To avoid confusion, in this paper we refer to those with focal depths exceeding 300 km as *very deep earthquakes*, and those with focal depths between 70 and 300 km as *intermediate earthquakes*.

Deep earthquakes are of interest for at least four reasons. First, they are exceedingly common. Between 1964 and 1986 they constituted no less than 22% of all earthquakes having m_b greater than 5.0 in the ISC catalog. Second, they most often occur in association with deep ocean trenches and volcanic island arcs in subduction zones. One of the great achievements of twentieth century geophysics was the recognition that their occurrence in inclined planar groups, or *Wadati-Benioff zones*, apparently delineates the cold downgoing cores of convection cells in the uppermost mantle. Third, seismologists use body waves of deep earthquakes disproportionately more often than those of shallow earthquakes to investigate core, mantle, and crustal structure. This is because (*a*) deep earthquakes often possess relatively more impulsive sources, (*b*) their body phases traverse the heterogeneous uppermost mantle only once on the ray path from hypocenter to station, and (*c*) surface waves do not contaminate these body phase arrivals on seismograms. Finally, deep earthquakes are

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important in their own right because they are mechanically different from shallow earthquakes. The essential mechanical process responsible for their existence is still poorly understood and remains one of the outstanding unsolved problems of geophysics and rock mechanics.

Surprisingly, the present paper is the first overall review of the properties of deep earthquakes since those of Gutenberg & Richter (1938), Jeffreys (1939), and the truly remarkable paper of Leith & Sharpe (1936). Recently, Frohlich (1987b, 1989) described the events leading to the discovery of deep earthquakes and reviewed the research that occurred prior to World War II. Gutenberg & Richter (1954), Rothe (1969), and Abe (1981, 1982, 1984) have provided lists of large historical deep earthquakes. The publication most representative of recent research is a special section of the *Journal of Geophysical Research* (December 1987) devoted to deep earthquakes. The present paper does not review the geographic distribution of deep earthquakes and the geometry of particular Wadati-Benioff zones; for this, see Yamaoka et al (1986) and Burbach & Frohlich (1986). Instead, I focus on the observed properties of deep earthquakes, which ultimately might provide some insight into their essential mechanical nature.

To this end, I first review the observed characteristics of deep earthquakes: their depth distribution, size distribution, stress drops, moment tensors, aftershocks, focal mechanisms, and tendency to cluster. This is followed by a discussion of some of the mechanical theories that have been proposed to explain the failure process occurring within the source region of deep earthquakes. Finally, I mention some unanswered questions that deserve future research attention.

OBSERVED CHARACTERISTICS OF DEEP EARTHQUAKES

Depth Distribution

The global occurrence of earthquakes (Figure 1) decreases approximately exponentially with depth down to about 400 km, then increases between about 450 km and 600 km before decreasing quite abruptly between 600 km and about 680 km. Between 70 and 300 km, earthquakes are numerous in several different geographic areas (Figure 2). In most areas the activity exhibits the distinct exponential decrease with depth that is evident in the global distribution (e.g. see Isacks et al 1968, Vassiliou 1983). Beneath 300 km depth, fully two thirds of the world's earthquakes occur in the Tonga-Kermadec region. For those regions having earthquakes at depths exceeding 500 km, generally there are fewer earthquakes between about 300 and 450 km (near the depth of the olivine-spinel phase transition) than at shallower or greater depths.

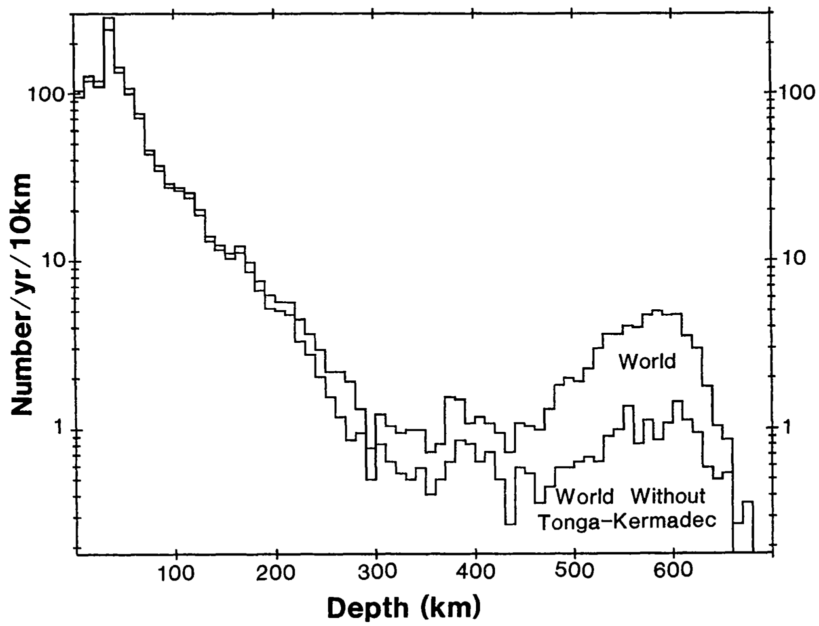

Figure 1 Number of earthquakes per year in 10-km depth intervals, as reported by the ISC as occurring between January 1964 and February 1986 and having m_b of 5.0 or greater. The upper line shows rates for the entire world, while the lower line excludes earthquakes in the Tonga-Kermadec region (Flinn & Engdahl 1965, regions 12 and 13). Between about 70 km and 330 km, the global rate is approximately proportional to $10^{-h/145\ \mathrm{km}}$, or $e^{-h/63\ \mathrm{km}}$, where h is depth.

Beneath about 600 km the rate of activity decreases markedly. Indeed, Stark & Frohlich (1985) noted that if the standard deviation in determining focal depths were as large as 30 km, then it would be possible to explain all the earthquakes beneath 630 km simply as a statistical "tail" caused by location errors. However, their study and that of Rees & Okal (1987)

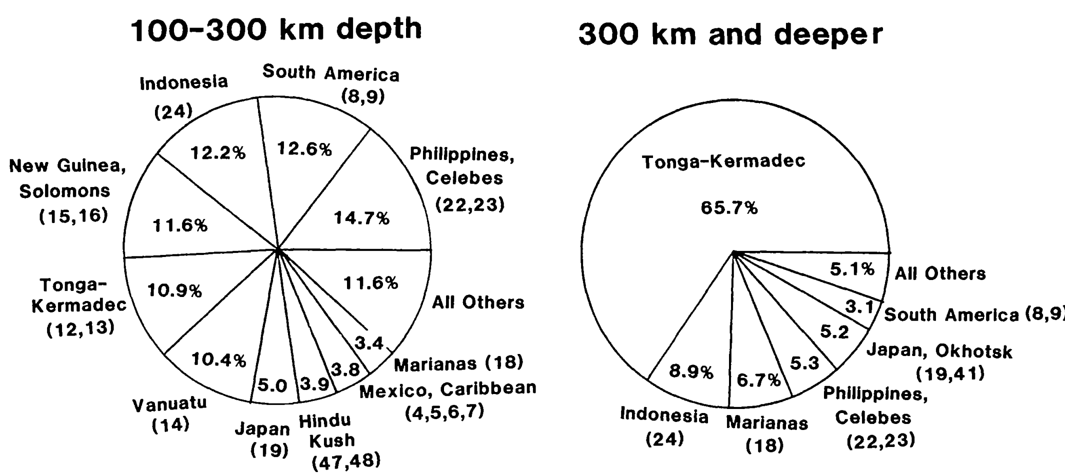

Figure 2 Pie graphs of the geographic distribution of earthquakes with depths 100–300 km (*left*) and >300 km (*right*). Numbers in parentheses are the Flinn & Engdahl (1965) region numbers included in each group.

showed quite clearly that earthquakes do occur between 630 km and about 680–690 km, and thus that the tail is real and not a statistical artifact. When earthquakes have depths determined independently by more than one method, Stark & Frohlich (1985) found that the differences between depths had standard deviations of only about 10–15 km.

Furthermore, some earthquakes beneath 630 km are exceptionally large and/or well recorded. The largest occurring in the last 25 years (Abe 1982) was the Colombia earthquake of 31 July 1970 (moment $= 2.2 \times 10^{28}$ dyn-cm, $m_w = 7.6$). The ISC determined depths from P residuals and $pP-P$ time differences as 653 km and 645 km, respectively. Perhaps the deepest well-recorded earthquake to date occurred in Tonga on 17 June 1977 (moment $= 2.35 \times 10^{25}$ dyn-cm, $m_b = 5.6$). The ISC determined depths of 673 km from P residuals and 684 km from $pP-P$ time differences, whereas Giardini (1984) found a depth of 690 km from waveform inversion. Gutenberg & Richter (1954) report an earthquake in the Flores Sea occurring on 29 June 1934 ($m_B = 7.0$) with a depth of 720 km. While this earthquake is often referenced as the "deepest known earthquake," more careful analyses by Jeffreys (1940) and by Rees & Okal (1987) favor a depth of 630–650 km. Stark & Frohlich (1985) and Rees & Okal (1987) question the reliability of all reported focal depths exceeding about 680–690 km. However, they do not entirely rule out the possibility that small, poorly recorded earthquakes occasionally occur at greater depths.

Size Distribution

While the largest deep earthquakes are somewhat smaller than the largest shallow earthquakes, between about 200 and 650 km there is no obvious indication that the size of the largest earthquakes varies with depth (Abe & Kanamori 1979; Figure 3). The maximum spatial dimension of the rupture zone can be as large as 500–1000 km for shallow earthquakes, causing moments of 10^{30} dyn-cm and more. For deep earthquakes, Willemann & Frohlich (1987) and Fukao & Kikuchi (1987) found that the largest earthquakes had moments of about 10^{28} dyn-cm and rupture or aftershock zone dimensions of about 80–150 km. The largest values occurred both for intermediate earthquakes (Banda Sea, 4 November 1963, depth = 100 km, moment $= 3.1 \times 10^{28}$ dyn-cm, $m_w = 7.7$; Kuriles, 6 December 1978, depth = 110 km, moment $= 6.40 \times 10^{27}$ dyn-cm, $m_w = 7.4$) and very deep earthquakes (Colombia, 31 July 1970, depth = 653 km, moment $= 2.2 \times 10^{28}$ dyn-cm, $m_w = 7.6$). In general, however, for both intermediate and very deep earthquakes, rupture or aftershock dimensions exceeding 80 km and moments exceeding 10^{28} dyn-cm, are quite rare.

The most common statistic used to characterize the proportion of large

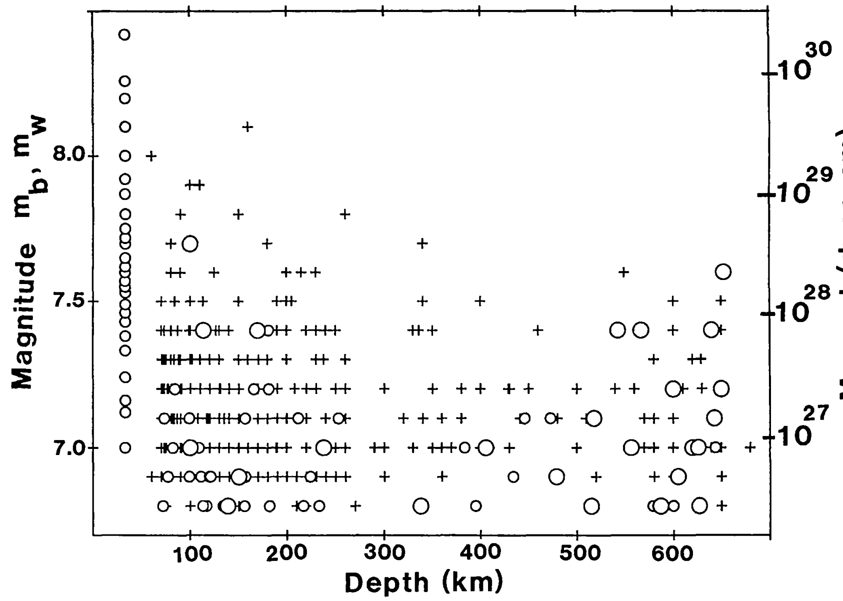

Figure 3 Beneath about 200 km there is no systematic change with depth in the size of the largest observed earthquakes, which have magnitudes of about 7.6 and moments of 10^{28}–10^{29} dyn-cm. Here crosses are magnitudes m_B from Abe (1981, 1984), and large circles are moment magnitudes m_w from Abe (1982). Small circles are moments M_0 from large shallow earthquakes listed by Kanamori (1977) and from deep earthquakes reported by the Harvard group (Domenico Giardini, personal communication, 1986). The relationship between M_0 and m_w is $m_w = [\log_{10}(M_0 - 10.1)]/2.4$ (Kanamori 1983).

to small earthquakes is the *b*-value. However, meaningful *b*-values are difficult to determine because (*a*) detection of small earthquakes is seldom uniform, (*b*) particular magnitude scales are generally reliable only over a limited range of earthquake sizes (e.g. Bath 1981), and (*c*) there exist physical constraints on the size of the largest possible earthquakes.

We evaluate here the relative proportions of moderate and large earthquakes (Table 1) by comparing Abe's (1981, 1984) compilation of large earthquakes (m_B of 7.0 or greater) occurring between 1904 and 1980 with the ISC catalog between 1964 and 1986 for moderate-sized earthquakes (m_b between 5.0 and 5.5). Abe's lists are the most complete and uniform available for large earthquakes (see Perez & Scholz 1984), and the ISC data should be uniform for moderate-sized earthquakes (Habermann 1982). For comparison with other studies, we compute a "*b*-value" by

$$\text{``}b\text{''} = \frac{\log_{10}(r_{\mathrm{ISC}}/r_{\mathrm{Abe}})}{\Delta M},$$

where r_{ISC} and r_{Abe} are the annual rates of moderate and large earthquakes, respectively, and ΔM is 2.0, the approximate difference in magnitude between the two groups.

This analysis (Table 1) and that of Giardini (1988) show clearly that

Table 1 Number of earthquakes per year occurring in two magnitude ranges, and associated "b-value"

	Moderate earthquakes[a] ($5.0 \leq m_b \leq 5.5$)	Large earthquakes[b] ($7.0 \leq m_B$ or M_S)	"b-value"[c]
All shallow ($0 < h < 70$ km)	821	12.7	0.91
All intermediate and deep ($h \geq 70$ km)	314	4.7	0.91
Tonga-Kermadec ($h \geq 300$ km)	44.7	0.34	1.06
Rest of world ($h \geq 300$ km)	19.4	0.74	0.71
South America ($h \geq 300$ km)	1.40	0.21	0.41

[a] Rates for moderate-sized earthquakes are determined from the ISC tape for events having magnitudes between 5.0 and 5.5 and occurring between January 1964 and February 1986.

[b] Rates for large earthquakes are determined from lists of Abe (1981, 1984) and are events having m_B or M_S of 7.0 or greater.

[c] If R_{ISC} and R_{Abe} are rates for moderate and large earthquakes, respectively, then the "b-value" is $\frac{1}{2}\log_{10}(R_{ISC}/R_{Abe})$.

there exist distinct geographic variations in the size distributions of very deep earthquakes. Taken as a group, shallow (0–70 km), intermediate (70–300 km), and very deep (>300 km) earthquakes have nearly identical "b"-values of about 0.9. However, the very deep earthquakes in the Tonga-Kermadec regions ("b" = 1.06) differ significantly from those in the rest of the world ("b" = 0.71). Clearly, Tonga-Kermadec, with two thirds of the world's very deep earthquakes, is an anomalous region in that it has so many moderate-sized very deep earthquakes and so few large ones. In contrast, in South America, where "b" = 0.41, large very deep earthquakes are relatively common and smaller very deep earthquakes seem to be rare (Suyehiro 1967). For the Colombia deep earthquake of 31 July 1970 (depth = 653 km, moment = 2.2×10^{28} dyn-cm, $m_w = 7.6$), the ISC catalog reports no earthquakes of any size whatsoever within 500 km of the hypocenter. Historically, the only known earthquakes that were possibly nearby occurred in 1911, 1921, and 1922 (Strelitz 1980); these are poorly located (Gutenberg & Richter 1954) and very large (m_B = 6.9, 7.5, and 7.4, respectively). Spain is even more anomalous than South America, as only two very deep earthquakes are known, one occurring on 29 March 1954 (depth = 630 km, moment = 7.0×10^{27} dyn-cm, $m_w = 7.0$; see Chung & Kanamori 1976) and the other occurring at nearly the same place on 30 January 1973 ($m_b = 4.0$; see Udias et al 1976).

Stress Drop and Apparent Stress

If an earthquake occurs having slip S over a fault with smallest linear dimension W and total area A, and if μ is the rigidity, then the moment M_0 and stress drop $\Delta\sigma$ are (e.g. Madariaga 1977):

$$M_0 = \mu AS,$$

$$\Delta\sigma = c\mu S/W,$$

where c is a constant that depends on the specific fault geometry assumed (e.g. circular, rectangular bi- or unilateral, etc.). Eliminating μS, we obtain

$$\Delta\sigma = cM_0/WA.$$

It is seldom possible to determine W and A directly, and so most studies determine stress drop as

$$\Delta\sigma = cM_0/A^{3/2} = cM_0/L^3 = cM_0/(V_R\tau)^3,$$

where L is some characteristic linear dimension of the earthquake, and τ is a source time or corner period over which rupture occurs with nominal rupture velocity V_R. While finding the moment M_0 is fairly straightforward, the determination of A, W, L, τ, or V_R is highly model dependent and may produce serious sytematic errors in the calculation of $\Delta\sigma$. Furthermore, investigators may either determine an average stress drop over the entire fault area, or instead determine stress drop for individual subevents occurring during a complex rupture process. All these factors severely complicate any comparison of stress drops from one study to another.

The source for large deep earthquakes usually consists of a series of two or more subevents (e.g. Sasatani 1980, Brustle & Muller 1987, Fukao & Kikuchi 1987). The rupture velocity is less than the shear velocity as implied by the distance and time difference between initial and subsequent subevents (Willemann & Frohlich 1987).

Several different investigations find that stress drops of intermediate and deep earthquakes are somewhat higher than those of shallow earthquakes. Wyss & Molnar (1972) and Molnar & Wyss (1972) determined corner periods and stress drops for 34 shallow and 19 deep earthquakes of moderate size (m_b of about 6.0) occurring in Tonga. They reported stress drops of 1–25 bars for the shallow earthquakes and 4–70 bars for the very deep earthquakes. Kikuchi & Fukao (1987) and Fukao & Kikuchi (1987) determined stress drops from the moments and pulse widths of subevents of 9 shallow and 18 deep, exceptionally large earthquakes (M_w of 7 or greater), none of which occurred in the Tonga-Kermadec region. They found stress drops of 170–460 bars for the shallow subevents, 60–1870 bars for the intermediate subevents, and 450–920 bars for the very deep subevents. While most investigations find that shallow earthquakes have lower stress drops than deep earthquakes, there is no agreement otherwise as to how stress drops vary with depth. Wyss & Molnar (1972) found no change with depth, Sasatani (1980) and Mikumo (1971) found an increase with depth, Fukao & Kikuchi (1987) found that they are highly variable at intermediate depths but nearly constant at greater depths, and Chung

& Kanamori (1980) found that they are highest near the depths of the 400-km and 650-km mantle boundaries.

As with shallow earthquakes (Scholz 1982), there is evidence that the stress drop of deep earthquakes depends on earthquake size. Pennington & Isacks (1979) found that pulse widths and pulse shapes of deep earthquakes in Tonga were nearly the same for magnitudes from 3.5 to about 5.0. This suggests that the earthquakes were simple and that stress drop varied from less than 0.1 bar to more than 1 bar. A careful analysis by Choy & Boatwright (1981) of digital seismograms of subevents of two moderate-sized earthquakes (Bali Sea, 10 June 1978, depth = 520 km, moment = 8.33×10^{24} dyn-cm, $m_b = 5.5$; Kuriles, 21 June 1978, depth = 371 km, moment = 6.49×10^{25} dyn-cm, $m_b = 5.9$) found stress drops of about 50 bars. Mori (1983) obtained similar values for five intermediate Aleutian events with magnitudes 5.3–5.8. Again, Fukao & Kikuchi (1987) obtained stress drops of 60–1870 bars for deep earthquakes having magnitudes of 7 and larger.

An earthquake's apparent stress σ_{app} is determined from the radiated seismic energy E_{rad} and the moment M_0 by

$$\sigma_{app} = \mu E_{rad}/M_0 = \mu\eta E_{tot}/M_0 = \eta(\sigma_{bef}+\sigma_{aft})/2.$$

Here E_{tot} is the total strain energy released by the earthquake, σ_{bef} and σ_{aft} are the shear stress levels at the source before and after the earthquake, respectively, and η is the seismic efficiency, i.e. the fraction E_{rad}/E_{tot} (e.g. Aki & Richards 1980).

Abe (1982) determined apparent stresses for 45 deep earthquakes having magnitudes of 6.0 and larger and found values from a few bars to more than 100 bars. The average apparent stress was about 60 bars, very close to that determined by Choy & Boatwright (1981). Because it seems unlikely that the stress drop exceeds σ_{bef}, the observation that σ_{app} is often much smaller than the stress drop suggests that the seismic efficiency must be small, say 0.1 or less. This indicates that a considerable fraction of the strain energy E_{tot} released by deep earthquakes is available to produce frictional heating, melting, etc, in the source region.

Moment Tensors

If we model earthquake waves as emanating from a point source, then the moment tensor M is a 3×3 symmetric matrix describing the radiation pattern (Aki & Richards 1980). Any source describable by a moment tensor M can be decomposed (e.g. Knopoff & Randall 1970, Riedesel & Jordan 1988) as the sum of three tensors representing an isotropic component (explosion or implosion), a double-couple component (slip along a fault plane), and a compensated linear vector dipole (CLVD) component (sourceward motion along one axis, accompanied by simultaneous motion away from the source along the two perpendicular axes). In particular, if e_1, e_2, and e_3 are the eigenvalues of M in order of decreasing size, the isotropic component is proportional to $e_1+e_2+e_3$. If M_D is the deviatoric (i.e. nonisotropic) part of M, then the usual measure of the proportion of CLVD component is $e_{\text{middle}}/e_{\text{largest}}$, where e_{largest} is the eigenvalue of M_D with largest absolute value, and e_{middle} is the eigenvalue of intermediate size (i.e. smallest absolute value). For a pure CLVD source, $e_{\text{middle}}/e_{\text{largest}}$ is 0.5, whereas it is zero for a pure double-couple source.

Almost as soon as deep earthquakes were discovered, scientists such as Stechschulte (1932) and Leith & Sharpe (1936) considered the possibility that they might have isotropic sources caused by phase transitions accompanied by a sudden change in volume. However, they rejected this hypothesis because careful observations such as those of Honda (1932) showed clearly that the radiation pattern possessed both dilational and compressional quadrants, closely resembling a double couple.

Subsequently the hypothesis of isotropic sources for deep earthquakes has reappeared more than once, especially as it has become plausible that metastable phase transitions do occur in the upper mantle (Sung 1974, Sung & Burns 1976a,b). For example, Benioff (1963) saw evidence for sourceward compression on a strainmeter for two very deep earthquakes in Peru (19 August 1961, $m_B = 7.2$; 31 August 1961, $m_B = 7.3$). Dziewonski & Gilbert (1974) and Gilbert & Dziewonski (1975) analyzed hand-digitized World-Wide Standardized Seismograph Network (WWSSN) records for 65 stations that recorded the 31 July 1970 Colombian deep earthquake, and they concluded that there was an isotropic component beginning about 70 s prior to the origin time determined from high-frequency data. However, Okal & Geller (1979) have argued that it is not possible to discriminate such a source using the data of Gilbert & Dziewonski.

More recently, Silver & Jordan (1982) developed methods to distinguish whether the total moment tensor had a statistically significant isotropic component. They concluded that a deep earthquake occurring beneath Honshu on 7 March 1978 (depth = 434 km, moment = 5.38×10^{26} dyn-cm) had an isotropic component, as measured in some (but not all) frequency bands. However, using a different approach, Riedesel (1985) was unable to detect a significant isotropic component for this earthquake, or for any of 20 other intermediate and deep earthquakes that he studied.

In summary, these and other published studies (e.g. Rogers & Pearce 1987, Stimpson & Pearce 1987) demonstrate rather conclusively that if isotropic components do occur in deep earthquakes, the relative amount must be quite small, e.g. 10% or less of the total moment. Furthermore, to date there has not been a single deep earthquake that all investigators agree possesses an isotropic component. Finally, considering the systematic uncertainties caused by Earth structure, attenuation, and station calibration, it is entirely possible that deep-earthquake sources have no isotropic component whatsoever. Any successful model of the deep-earthquake source must explain why deep earthquakes are predominantly deviatoric.

In contrast, it is becoming ever clearer that some deep earthquakes are not simple double couples, as many do possess a CLVD component. Knopoff & Randall (1970) were the first to propose the CLVD source for deep earthquakes. Randall & Knopoff (1970) found CLVD components in all of the five deep earthquakes they studied, although they questioned their statistical method because they also found nonzero implosive components for three earthquakes and explosive components for two earthquakes.

More recently, the Harvard group (Dziewonski et al 1981, Dziewonski & Woodhouse 1983, Giardini 1984) has routinely determined deviatoric moment tensors for several hundred earthquakes each year, including 956 intermediate and deep earthquakes occurring between 1977 and 1986. The proportion of CLVD component for these earthquakes ranges up to 90% ($e_{\mathrm{middle}}/e_{\mathrm{largest}} = 0.45$) with a mean of 22% ($e_{\mathrm{middle}}/e_{\mathrm{largest}} = 0.11$). There is little indication that the proportion depends either on the size or the depth of the earthquakes (Figure 4), although the proportion may be slightly higher for earthquakes between 70- and 150-km depth than for shallower and deeper earthquakes.

Some large and extremely well-recorded earthquakes are clearly different from a double couple. For the Tonga earthquake of 26 January 1983 (depth = 224 km, moment = 3.66×10^{26} dyn-cm), Dziewonski et al (1983) found $e_{\mathrm{middle}}/e_{\mathrm{largest}}$ to be 0.39. A more statistically rigorous analysis of this earthquake by Riedesel (1985) and Riedesel & Jordan (1988) found $e_{\mathrm{middle}}/e_{\mathrm{largest}}$ to be 0.41. They concluded that this event was significantly different from a pure double-couple source.

Physically, what allows sources with significant CLVD components to occur? Knopoff & Randall (1970) proposed that rapidly running, volume-preserving phase transitions might produce CLVD sources. If so, why couldn't phase transitions with volume changes occur at the same rate and produce sources with significant isotropic components? More recently, Julian & Sipkin (1985) interpreted a shallow CLVD source near Mammoth Lakes, California, as the opening of a tensile crack by high-pressure pore fluids; however, it is difficult to imagine such a process occurring in the mantle.

Finally, because the sum of two suitably oriented double-couple sources will produce a pure CLVD source (Figure 5), several studies (e.g. Sipkin

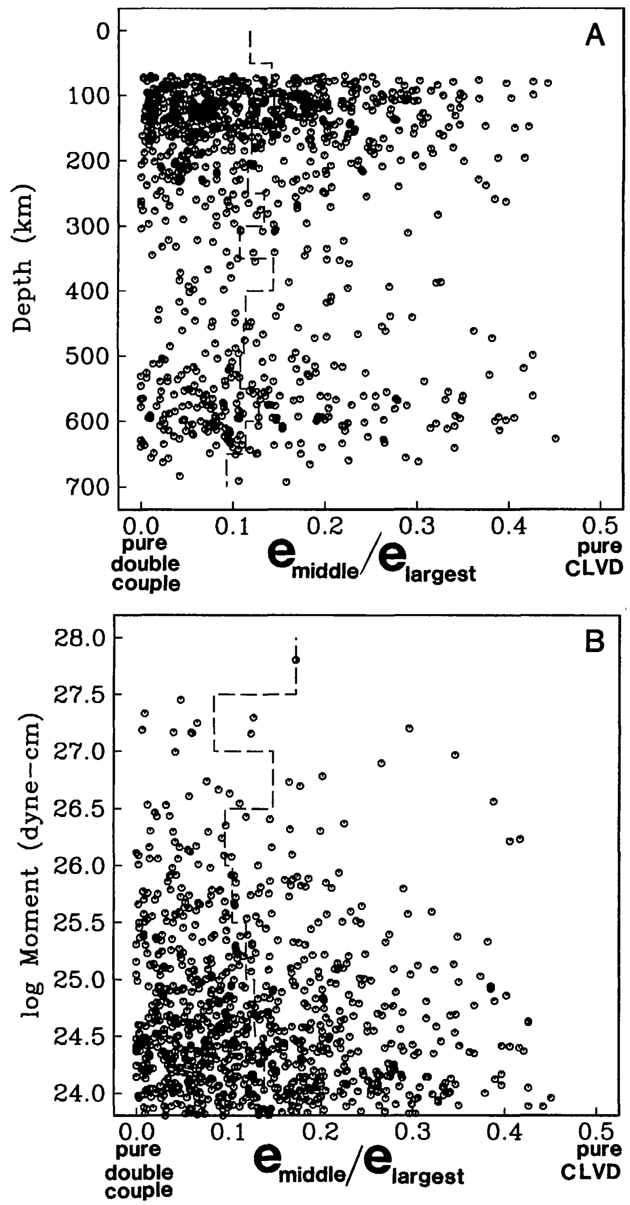

Figure 4 Depth (*A*) and moment (*B*) distributions of the ratio $e_{middle}/e_{largest}$ of the compensated linear vector dipole (CLVD) component to the double-couple component for 956 earthquakes occurring between 1977 and 1986, as determined by the Harvard group (e.g. Dziewonski et al 1981, Dziewonski & Woodhouse 1983, Giardini 1984). Dashed line is the mean value of the ratio for earthquakes in intervals of 50 km for depth and half an order of magnitude for moment.

1986) have suggested that CLVD sources come about in regions of "extreme structural complexity" when an earthquake consists of two or more subevents having different focal mechanisms, or when rupture occurs on a nonplanar fault surface. This explanation seems the most plausible, even though the Tonga earthquake described above does not occur in a

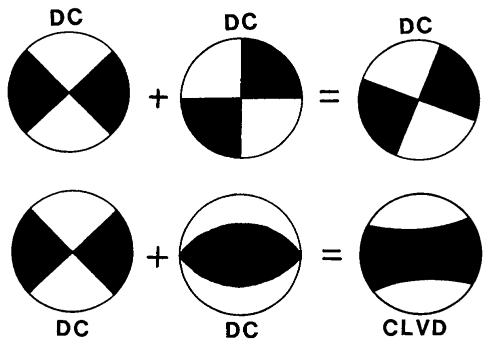

Figure 5 A combination of two suitably oriented double-couple sources of equal strength will produce a pure CLVD source (*bottom*). The two double couples must have perpendicular *B* axes and either the same *P* or the same *T* axes. Combining two double couples having the same *B* axis produces a pure double couple with no CLVD source component (*top*), even when the two double couples differ in strength.

region where the Wadati-Benioff zone geometry is complex (Burbach & Frohlich 1986).

Aftershocks

The spatial and temporal relationship of aftershocks with one another and with the initial event contrains the scale of the source process, while the relationship to focal mechanism parameters and to Wadati-Benioff zone geometry constrains how the failure process depends on factors such as stress. There is ambiguity in the literature concerning aftershocks because not all investigations define aftershocks in the same way. For example, Page (1968) required aftershocks to be numerous, to be smaller in magnitude than the main event, and to die off according to Omori's law (i.e. the number of aftershocks per unit time is proportional to $1/t^p$, where t is time and p is a constant). In some studies the term aftershocks includes what we call "rupture subevents," which occur during the rupture process of the main event. In the following an earthquake is a "true" aftershock only if a statistical test suggests that it is not independent of the initial event. Thus an initial event may have none, one, or any number of aftershocks, which may be either smaller or larger than the initial event.

Recent observations demonstrate clearly that aftershocks are much less common for deep earthquakes than for shallow earthquakes, but that a significant fraction of deep earthquakes possess at least one aftershock. However, as noted by Page (1968), sequences possessing numerous aftershocks dying off according to Omori's law occur exclusively at shallow depths. Kagan & Knopoff (1980) and Prozorov & Dziewonski (1982) investigated the stochastic properties of global earthquake catalogs and found that the branching rate was 10 times lower for intermediate than for shallow earthquakes. Unlike Kagan & Knopoff (1980), Prozorov & Dziewonski (1982) found detectable aftershocks at depths of 450–650 km, while neither study detected aftershocks between depths of 250 and 450 km.

Using a somewhat more sensitive method, Frohlich (1987a) detected aftershocks in the ISC catalog in all depth ranges but found them to be less common between about 100 and 450 km than at greater or lesser depths. For initial events having magnitudes of 5.0 and larger and depths exceeding 100 km as constrained by $pP-P$ intervals, only about 5% had aftershocks. Of these, only about one fourth had more than one aftershock. While aftershocks were more common for large-magnitude earthquakes, some extraordinarily large earthquakes such as the Colombian deep earthquake of 31 July 1970 had no aftershocks whatsoever. Willemann & Frohlich (1987) found a few earthquakes with magnitudes of 5.5 or less that had aftershocks 20–50 km distant from the initial hypocenter, and thus clearly occurred outside the source region.

Frohlich & Willemann (1987a) and Willemann & Frohlich (1987) analyzed 59 deep sequences to determine whether aftershocks occur on focal mechanism nodal planes, as might be expected if deep earthquakes were progressive shear failures along a planar surface. Surprisingly, they found no preferential clustering near initial-event nodal planes for either true aftershocks or rupture subevents (Figure 6), a result in contrast with most previous studies (e.g. Oike 1971). Most of the other studies either failed to recognize that half the focal sphere lies within 16° of a nodal plane or simply assumed that rupture subsevents occurred on a nodal plane (e.g. Fukao & Kikuchi 1987). Giardini & Woodhouse (1984) relocated numerous very deep earthquakes (not aftershocks) and found that they

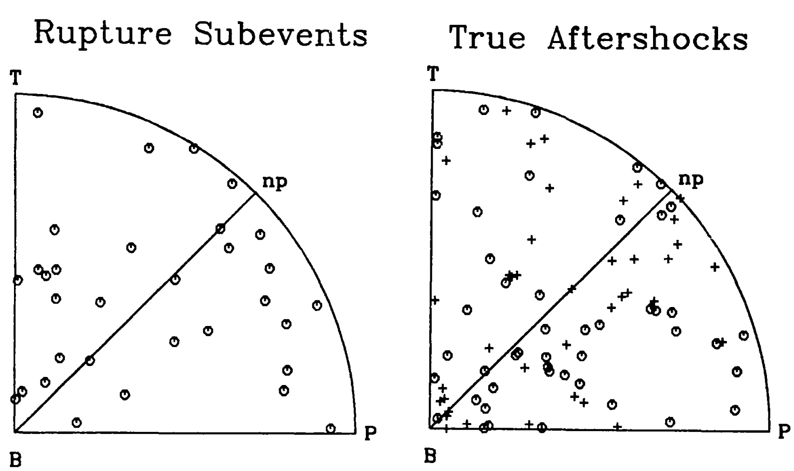

Figure 6 Comparison of aftershock direction and orientation of initial-event focal mechanism for deep earthquakes (reproduced from Frohlich & Willemann 1987a). In this equal-area projection, an aftershock plots as P if it occurs along a line having the orientation of the P axis and passing through the initial-event hypocenter, and similarly for B and T. Aftershocks occurring on the extension of the nodal plane plot along the line joining B and "np" [see Frohlich & Willemann (1987b) for a more complete explanation]. Circles are data of Willemann & Frohlich (1987), and pluses are data of Oike (197). Note that neither true aftershocks nor rupture aftershocks appear clustered with respect to the nodal planes.

sometimes formed elongated groups or planar clusters lying roughly parallel to the focal planes of large earthquakes.

Michael (1988) noted that the larger aftershocks of shallow earthquakes often do not occur on nodal planes and suggested that Willemann & Frohlich (1987) had insufficient data to conclude that shallow and deep earthquakes differed in this respect. Nevertheless, a conservative interpretation of Willemann & Frohlich's (1987) results is that the data available at present do not indicate that deep aftershocks cluster near nodal planes. Furthermore, if more data subsequently show that clustering does occur, it cannot be very intense, as a substantial fraction of all deep aftershocks are clearly not near nodal planes.

Focal Mechanisms and Wadati-Benioff Zone Geometry

Earthquake focal mechanisms provide the most direct information available concerning the orientation of the stresses that produce deep earthquakes. Isacks & Molnar (1971) reviewed 204 focal mechanisms determined primarily from P-wave first motions. For earthquakes deeper than 300 km they found that compressional axes (P axes) were generally aligned along the downdip direction of the Wadati-Benioff zone. For intermediate-depth earthquakes, the P-axis orientation was downdip in a few regions such as Tonga and Japan, while the tensional axes (T axes) were downdip in most other areas, including South and Central America, the Philippines, and the New Hebrides. More recently, Vassiliou (1984) and Apperson & Frohlich (1987) found similar results from mechanisms determined from moment tensors derived from digital seismograms of earthquakes occurring since 1977 (e.g. see Figure 7).

Generally, the interpretation of these investigations is that gravitational

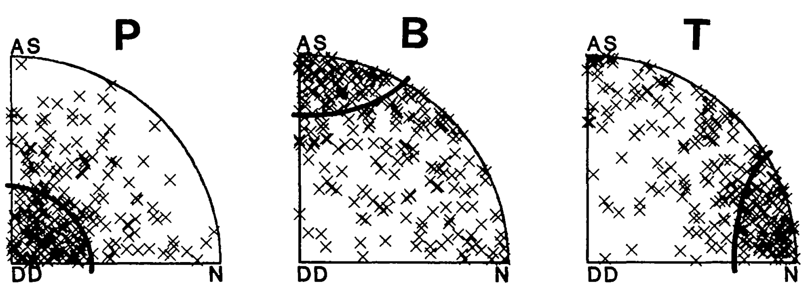

Figure 7 Equal-area projections showing the relationship between Wadati-Benioff zone geometry and the P, B, and T axes of 249 very deep earthquakes evaluated by Apperson & Frohlich (1987). DD, AS, and N are locations of axes lying along the downdip, alongstrike, and normal axes of the Wadati-Benioff zone. Solid curved lines show the portion of the focal sphere within 30° of the downdip (*left*), alongstrike (*middle*), and normal (*right*) axes. Note that a substantial fraction of all events possess P, B, or T axes lying more than 30° from the DD, AS, or N axes. Focal mechanism axes are as determined by the Harvard group (e.g. Giardini 1984), as in Figure 4.

forces acting on the sinking lithospheric plate provide the downdip control of P and T axes (e.g. Richter 1979, Vassiliou et al 1984). In regions with downdip T the denser sinking plate pulls itself into the mantle, whereas in regions of downdip P it meets resistance because the viscosity of the mantle increases significantly beneath about 650 km.

However, recent investigations show that within individual Wadati-Benioff zones there is considerable variability that cannot be attributed to statistical or systematic errors in the determination of focal mechanisms. For example, at intermediate depths in Japan and Tonga, Hasegawa et al (1978), Fujita & Kanamori (1981), Kawakatsu (1986a), and Hasegawa & Takagi (1987) found both downdip P and downdip T occurring together, which they attributed to bending of the subducting plate. The P and T axes of Giardini's (1984) 200 deep earthquakes possess considerable scatter, which he attributed to complexity in the regional stress field. Apperson & Frohlich (1987) studied 865 earthquakes and found that for "typical" deep earthquakes, P or T axes lay downdip and B axes lay alongstrike of Wadati-Benioff zones. However, more than 70% of all earthquakes were not "typical," as they either had a P or T axis more than 30° from downdip or a B axis more than 30° from alongstrike. The axial orientation of earthquakes with large CLVD components does not differ in any obvious way from that of earthquakes that are mostly double couples (Figure 8).

Clustering and Isolation of Hypocenters

In several regions between about 100-km and 200-km depth there exist intense spatial clusters, or "nests," of earthquakes. The best known of these is the Bucaramanga nest in Colombia, which was investigated with a temporary local network by Schnieder et al (1987). They recorded about 15 earthquakes day^{-1} emanating from a zone having dimensions 8 km ×

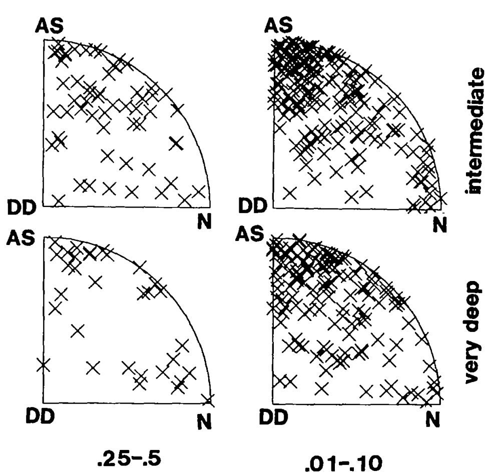

Figure 8 The orientation of the B axis with respect to the Wadati-Benioff zone does not differ for earthquakes with a large CLVD component (*left*: $e_{\mathrm{middle}}/e_{\mathrm{largest}} > 0.25$) and those with a small CLVD component (*right*: $0.01 < e_{\mathrm{middle}}/e_{\mathrm{largest}} < 0.10$). This figure shows B axes on equal-area projections (see Figure 7) for intermediate earthquakes (*top*) and very deep earthquakes (*bottom*) studied by Apperson & Frohlich (1987).

4 km × 4 km, surrounded by a zone more than 100 km in extent having a considerably lower rate of activity.

Several other intermediate-depth clusters have often been compared to the Bucaramanga nest, although all appear to have a larger volume, a lower activity rate, and less isolation from other nearby activity. These include nests in the Hindu Kush (Chatelain et al 1980), activity near Vrancea, Romania (Oncescu 1984, Oncescu & Trifu 1987), the Socampa nest in Peru (Sacks et al 1967), and the Iliamma cluster beneath Cook Inlet, Alaska (Pulpan & Frohlich 1985). Similar concentrations of activity occur near tears, edges, or bends in subducted lithospheric segments in the Caribbean arc, the Scotia arc, and elsewhere. However, except for the Bucaramanga nest, it is not clear whether "nests" represent a distinct phenomenon or merely normal fluctuations in the spatial rate of earthquake activity.

At depths exceeding 400 km there exist several regions where rare, isolated earthquakes occur, often 500 km or more from the nearest Wadati-Benioff zone. Examples include the two Spanish deep earthquakes at about 630-km depth (Chung & Kanamori 1976, Udias et al 1976), four earthquakes at about 580–600 km depth beneath North Island, New Zealand (Adams 1963, Adams & Ferris 1976), and the exceptionally large Colombian earthquake of 31 July 1970 at 650-km depth (Furumoto 1977, Strelitz 1980). The existence of these earthquakes suggests that in the future, large isolated earthquakes may well occur in areas where no activity has occurred previously.

MECHANICAL PROCESS

When scientists such as Leith & Sharpe (1936) and Jeffreys (1939) first realized that deep earthquakes occurred, they asked three questions. First, how could mantle rock have sufficient strength to permit earthquakes to occur, when observations of isostasy suggested that it would flow to reduce any shear stress (Figure 9)? Second, what was the source of stress causing deep earthquakes, particularly inasmuch as laboratory observations showed that for ordinary fracture, rock strength increased with increasing hydrostatic pressure? Finally, what physical process took place in the source region as deep earthquakes occurred?

Concerning mantle strength, scientists now agree that convection (i.e. plate tectonics) causes the mantle to be very heterogeneous. Beneath subduction zones there are regions of lower temperature, different composition, and higher strength, making earthquakes possible, even as most of the mantle undergoes ductile flow. Furthermore, rock failure is highly

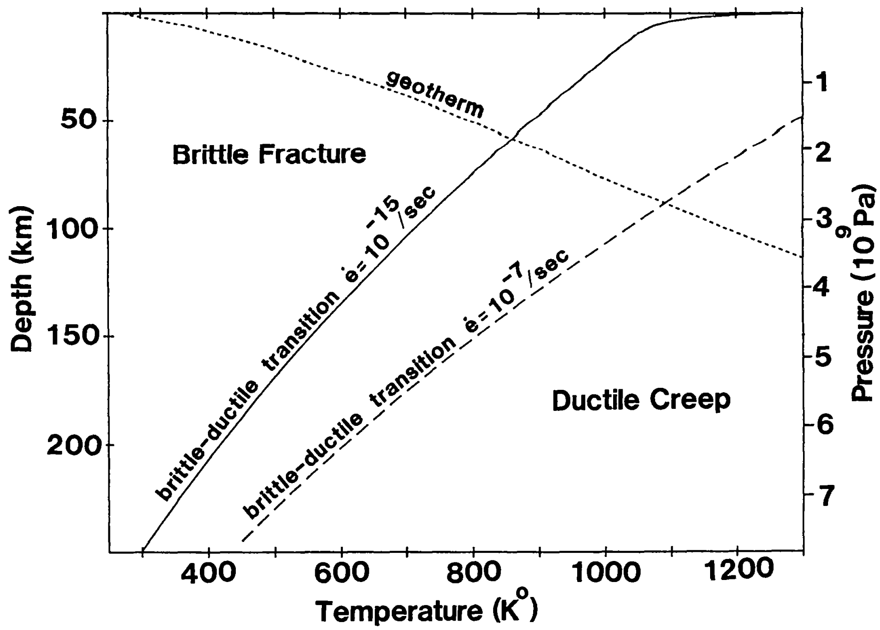

Figure 9 Relationship between pressure (depth), temperature, strain rate, and the mechanism of rock failure. At low temperatures, low pressures, and high strain rates rocks tend to fracture, whereas at high temperatures, high pressures, and low strain rates they fail by ductile creep. Solid line indicates transition for a strain rate $\dot{\varepsilon}$ of 10^{-15} s^{-1}; dashed line is for $\dot{\varepsilon}$ of 10^{-7} s^{-1}. The dotted line is an approximate mantle geotherm. For this figure we calculate the pressure and temperature where the transition occurs by using the expressions of Houseman & England (1986), equating the fracture strength (Byerlee's law) and the shear stress necessary to maintain the given strain rate. At temperature/pressure conditions corresponding to depths exceeding about 100 km, brittle fracture does not occur, which suggests that some other failure mechanism must be responsible for deep earthquakes.

strain rate dependent, so that rock, like pitch or Silly Putty, may flow at low strain rates and fracture at higher strain rates.

Concerning sources of stress, there are numerous processes that singly or in combination could produce stresses large enough to generate deep earthquakes. These include the following:

1. Tensional stresses caused by gravity as a plate attached to the surface sinks into the mantle (e.g. Spence 1987).
2. Compressional stresses caused by gravity as the edge of a sinking plate impinges upon the more viscous lower mantle (e.g. Richter 1979, Vassiliou et al 1984).
3. Bending stresses produced as the plate buckles or changes shape as it subducts (Kawakatsu. 1986b, Fukao et al 1987).
4. Thermal stresses occurring because the outside of the subducting plate warms faster than the interior (e.g. Wortel 1986).
5. Phase transitions causing volume reduction in subducting material,

producing tensional stresses in neighboring regions (Ringwood 1972, McGarr 1977).

Concerning the mechanical process, deep earthquakes clearly are unlike shallow earthquakes, even though they generally appear to involve a shear failure. Theories of rock fracture rely on the properties and coalescing of cracks (e.g. Dmowska & Rice 1986), which could not stay open at mantle pressures. Furthermore, at high pressures and temperatures, Byerlee's law breaks down (e.g. Kirby 1983), and rocks begin to yield by creep or flow rather than by fracture (Figure 9). Apparently, the failure process involves some shearing mechanism other than ordinary brittle fracture.

Melting or Runaway Creep Along Shear Zones

Most rock creep processes are strongly temperature dependent (e.g. Turcotte & Schubert 1982). Thus, if shear stresses induce rock creep, the creep produces heat that will influence the rate of creep if conduction cannot dissipate the heat quickly enough. As creep accelerates it may cause melting along a planar surface, effectively reducing the rock strength to zero. Several investigators have suggesed that this phenomenon might explain deep earthquakes (Griggs & Handin 1960, Griggs & Baker 1969, Ogawa 1987). This mechanism is attractive because it is generally independent of the particular physical mechanism responsible for the creep. Ogawa (1987) calculates that the critical strain rate necessary to allow runaway creep is $10^{-14}\ s^{-1}$, which is comparable to or lower than the rates expected within portions of the subducting slab (Bevis 1988).

However, an important uncertainty is whether the creep instability can proceed quickly enough or over a large enough region to produce earthquakes. The mechanism is favored if there exist suitably oriented planar regions of lower strength, a condition that may not occur within subducted lithosphere. It seems plausible to me that accelerating creep would commonly produce foreshocks, spatial chains of aftershocks, long and very complex source time functions, and a general swarmlike behavior as neighboring regions within the slab readjusted to the stresses produced by a particular earthquake. These phenomena are not generally characteristic of deep earthquakes. However, these objections are admittedly intuitive and nonquantitative, and the runaway creep hypothesis remains a reasonable candidate to explain deep-earthquake activity.

Plastic Instabilities

Resistance to any creep process may either decrease or increase as strain rate increases. If the resistance increases, the process is stable and called strain hardening. If resistance decreases and strain softening occurs, the process is unstable and creep accelerates. In plastic materials this may occur even if no significant temperature change occurs during the creep process. However, a temperature increase often occurs and contributes to strain softening, and so this mechanism is not always distinct from melting or runaway creep. Several investigators (Tse & Rice 1986, Strehlau 1986) have suggested that the temperature and pressure dependence of creep resistance may explain a number of features of faulting and seismic activity for shallow earthquakes.

Orowan (1960) and Hobbs & Ord (1988) have suggested that plastic instabilities may be the fundamental process of deep earthquakes. Laboratory investigations of some metals, polymers, and metallic glasses find plastic instabilities occurring fast enough to cause audible pings of clicks. The major uncertainty is whether this process occurs in actual mantle materials, as there has been virtually no experimental work on olivine-rich rocks at appropriate temperatures and pressures.

Solid-Solid Phase Transitions

The idea that deep earthquakes might be implosions caused by sudden phase transitions within the subducting lithosphere is so intellectually appealing that it is reborn regularly (e.g. Bridgman 1945, Evison 1963, 1967, Liu 1983), in spite of the absence of any compelling seismological observations to support it. Knopoff & Randall (1970) suggested that phase transitions without volume changes might produce compensated linear vector dipole sources. While many deep earthquakes do possess linear vector dipole components (Figure 4), the double-couple components are generally larger. An important question for any phase transition hypothesis is whether over large volumes the transition can occur quickly enough to radiate seismic energy (Ringwood 1972).

Recently, Kirby (1987) proposed a new phase transition mechanism, which depends on the observation that in some materials solid-solid phase transitions occur at lower confining pressures if the material is under substantial shear stress. Thus the transition occurs preferentially along the planes of maximum shear stress, and if this transition occurs suddenly, it is likely to radiate energy as a double-couple source. In the laboratory these shear-stress-dependent transitions occur quickly enough in materials such as ice and tremolite to produce acoustic emissions, suggesting that this process might be capable of radiating seismic energy in larger samples. Meade & Jeanloz (1988) observed acoustic emissions associated with phase transitions in samples of pure silicon and pure germanium at pressures corresponding to depths of 500 km and 1750 km, respectively. However, in samples of pure olivine and pyroxene they observed no acoustic emissions. They did record acoustic emissions in olivine-serpentinite mixtures and in pyroxene-serpentinite mixtures at a temperature of 900 K and pressures corresponding to depths of about 600 km.

While these experimental results are preliminary, they suggest that phase transitions permit fracture at depths exceeding those indicated by Figure 9. Phase transitions provide an attractive mechanism for deep earthquakes because we know that solid-solid phase transitions do occur in the upper 700 km of the mantle. Furthermore, since the transformation of olivine to the spinel and then to the perovskite structure is complete at pressures corresponding to about 700 km, this mechanism might explain why no earthquakes occur at greater depths.

Pore Fluids Reducing Effective Stress

A final possibility is that trapped pore fluids in the mantle reduce the effective hydrostatic stress, bringing the rock into a regime where shear stresses cause it to fail by ordinary brittle fracture or frictional sliding (Griggs & Handin 1960). This mechanism has helped to explain how injected fluids or reservoir loading can induce earthquakes at shallow depths (Gupta 1985, Simpson 1986, Davis & Frohlich 1989). In the mantle, the pore fluids might arise from water trapped in subducting sedimentary material (Shreve & Cloos 1986) or from the thermochemical dehydration of subducting rocks such as serpentinites (Raleigh & Paterson 1965, Raleigh 1967). Alternatively, the fluid might be a substance other than water produced by partial melting along grain boundaries (Raleigh 1967). This provides an explanation for Meade & Jeanloz' (1988) laboratory observation that acoustic emissions occur in olivine-serpentinite mixtures at pressures corresponding to depths of 600 km, but that acoustic emissions are absent in pure olivine samples.

However, it is questionable whether the porosity in subducted rocks is sufficient to communicate pore pressures well enough to support this mechanism. It is also not clear if ordinary fracture or frictional sliding will work at all under mantle temperatures and pressures, regardless of pore pressure. There exists no obvious concentration of earthquakes at a particular depth in Wadati-Benioff zones, as might be expected if they were due to a chemical dehydration reaction taking place at specific pressure-temperature conditions. Nevertheless, it is at present impossible to rule this out as a candidate mechanism for deep earthquakes. Furthermore, because it is now possible to observe fracture in the laboratory under temperature/pressure conditions appropriate for deep earthquakes, further experimentation in the near future should allow us to choose among the various candidate mechanisms.

SUMMARY AND CONCLUSIONS

Both theory and observation suggest that deep earthquakes are fundamentally different from shallow earthquakes. Laboratory investigations of brittle fracture and stable sliding, the mechanical processes responsible for shallow earthquakes, show that they cannot explain the presence of earthquakes under the pressure-temperature conditions within most of the upper mantle. Theoretical models of rock failure that rely on the opening, propagation, and coalescing of cracks cannot explain seismic activity in the mantle, where the opening of cracks is not possible. As a group, deep earthquakes have significantly fewer aftershocks and higher stress drops than shallow earthquakes of comparable size.

However, in some ways deep and shallow earthquakes are similar. For example, large earthquakes at both shallow and deep depths often have complex source time functions, which suggests that rupture in one location affects stress in neighboring regions, generating subevents. Much of the reputed simplicity of the seismograms of deep earthquakes comes about because of the more distinct separation between direct phases and surface or crustal reflections, as well as the lower amplitude of surface waves. Another similarity between deep and shallow earthquakes is that both seldom or never possess sources having isotropic components, as might be expected if they were generated by a catastrophic volume change or implosion.

Nevertheless, there are reasons to conclude that phase transitions either directly or indirectly affect the mechanism of deep earthquakes. There is an abrupt cessation of all activity at about the depth of the spinel-oxide transition, and a pronounced minimum in the activity at approximately the depth of the olivine-spinel phase transition. The rate of activity decreases exponentially with depth from the surface to about the depth of the olivine-spinel transition. Aftershocks are somewhat more common for very deep earthquakes than for intermediate-depth earthquakes. In any case, these phase transitions clearly do occur in the upper mantle, the associated volume changes must affect the regional pattern of stress, and heat liberated or absorbed must affect the thermal and mechanical structure of the subducted lithosphere. If these transitions occur catastrophically at lower confining pressures when there is shear stress present, it is possible that the transitions are the primary process responsible for deep earthquakes.

There exist distinct geographic variations in the size distribution of very deep earthquakes, with smaller very deep earthquakes being relatively common in the Tonga-Kermadec region and relatively rare (relative to the number of large earthquakes) in South America and Spain. Because about two thirds of all reported very deep activity occurs in Tonga-Kermadec, these earthquakes are the "typical" very deep earthquakes. However, if we sample by geographic region rather than by event, regions such as South America or Okhotsk with relatively few small earthquakes and occasional very large earthquakes are more typical. This size-related geographic bias is seldom recognized in the very deep earthquake literature, but it is always present, since studies of the properties of moderate-sized earthquakes usually focus on the Tonga-Kermadec region. Similarly, studies of the properties of large very deep earthquakes (e.g. Fukao & Kikuchi 1987) are biased against earthquakes in Tonga-Kermadec. For intermediate-depth earthquakes, both large and small earthquakes occur in abundance in several different geographic regions.

Unanswered Questions—Future Research

In the 50 years since the appearance of the reviews by Leith & Sharpe (1936), Gutenberg & Richter (1938), and Jeffreys (1939), about a thousand articles concerning deep earthquakes have been published, with about half appearing since 1977. This research has clarified and in some cases contradicted the conclusions of the earlier reviewers. Unlike a half century ago, today we are quite certain about the presence of phase transitions and/or compositional boundaries within the upper mantle. We are also fully aware of the presence of both convective and conductive regimes in the lithosphere and upper mantle. This lateral heterogeneity makes seismic activity possible in some geographic regions but absent in others. However, the earlier reviewers reported periodicities of deep activity at tidal and other frequencies, a conclusion not confirmed by more recent studies (McMurray 1941, Curchin & Pennington 1987). Furthermore, Leith & Sharpe (1936) and Jeffreys (1939) disagreed with Gutenberg & Richter (1938) concerning whether aftershock activity was different for deep and shallow earthquakes and whether there had to be different fundamental mechanical processes causing deep and shallow earthquakes. Today we agree that the mechanical processes must be different, as are some of the observed characteristics, such as the pattern of aftershocks.

What is the mechanical process of deep earthquakes? When a deep earthquake happens, what physical process actually take place in the source region? This is still one of the outstanding unsolved problems in geophysics, even though it has been regularly noted as such (e.g. Adams 1947) ever since the discovery of deep earthquakes. Until we make more progress on this question, it will be difficult to explain why some properties of deep earthquakes vary with depth or geographic region, or to find answers to any of the questions listed below.

Why do earthquakes not occur beneath about 680 km? Also, what allows such remarkably large ($m_B > 7.0$), very deep (depth > 630 km) earthquakes to occur in so many different geographic regions, namely South America, Tonga-Kermadec, Okhotsk, the Philippines, Indonesia, and Spain? Does upper-mantle convection proceed beneath the distinct velocity increase occurring at about 650–670 km depth, and is this boundary a phase boundary or a compositional boundary? While the most recent seismological evidence tends to favor convection that proceeds into the lower mantle (e.g. Silver & Chan 1986, Silver et al 1988, Fischer et al 1988), the observational results are subject to multiple interpretations, and these questions must still be considered unanswered (e.g. see Anderson 1987).

Are all deep earthquakes related to subduction? Since the acceptance of plate tectonics, scientists have often presumed that all deep earthquakes occur in subduction zones. However, Hatzfeld & Frogneux (1981) and Chen & Molnar (1983) argue that some deep earthquakes occur in areas such as northern Africa where there is no obvious ongoing subduction. While it is clear that the vast majority of deep earthquakes occur in subduction environments, can we at present dismiss the possibility that a small fraction (say 2%) of all deep earthquakes might be "intraplate deep earthquakes," unrelated to subduction?

How similar or different are intermediate and very deep earthquakes? Most investigations consider intermediate and very deep earthquakes together because they often occur together geographically, and because both are distinctly different from shallow earthquakes. However, many regions possess intermediate but no very deep earthquakes. Where both occur, they are spatially separated by a quiescent or aseismic zone and by the olivine-spinel phase boundary. In several regions, intermediate and very deep earthquakes differ with respect to their *b*-values and in their liklihood of possessing aftershocks. What is the significance of the remarkable exponential decrease with depth in the activity at intermediate depths? Do intermediate and very deep earthquakes represent distinctly different mechanical phenomena, or are they essentially similar in spite of the differences in the pressure, temperature, and strain rate conditions in their source regions?

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