JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 91, NO. B5, PAGES 4811-4825, APRIL 10, 1986

DOUBLE SEISMIC ZONES: KINEMATICS

Hitoshi Kawakatsu[1]

Department of Geophysics, Stanford University, California

Abstract. The kinematics of a double seismic zone of subducting lithosphere at an intermediate depth is studied. A simple geometrical argument shows that the strain rate due to unbending of subducting lithosphere can be as large as $10^{-15}\ s^{-1}$ and large enough to account for the seismic strain rate in this depth range. The geometrical unbending of subducting lithosphere in the mantle and the presence of a double seismic zone there can be, therefore, considered essentially equivalent phenomenon. We model a double seismic zone required from this unbending in order to investigate what constraint the presence of a double seismic zone can place on the properties of the descending lithosphere. The thickness of a double seismic zone is controlled by the thermal structure and the rheological properties of the subducting slab. A preliminary result for the double seismic zone beneath Tohoku, Japan, indicates a possibility that the currently available flow law of "dry" olivine does not properly describe the double seismic zone or at least presents some inconsistency in our knowledge of the mechanical properties (e.g., thermal structure, rheology) of oceanic lithospheres.

Introduction

In the theory of plate tectonics, the zone of intermediate depth and deep focus earthquakes (Wadati-Benioff zone) is considered the trace of subducting oceanic lithosphere. Above a depth of ~50 km, oceanic lithosphere bends when it subducts into the mantle. From seismicity, we know that the subducting lithosphere is almost straight in the mantle below a depth of ~200 km. This indicates that the lithosphere must unbend in a depth range from ~50 km to ~200 km to become straight [Isacks and Barazangi, 1977; Engdahl and Scholz, 1977; Tsukahara, 1977]. The kinematics of the theory of plate tectonics, therefore, requires unbending of the subducting oceanic lithosphere at an intermediate depth range, and it is likely that the strain due to this unbending is partly released by earthquakes. In this paper, we demonstrate essential equivalence of the so-called double seismic zone of intermediate earthquakes to this geometrical unbending of the subducting lithosphere and suggest that the presence of a double seismic zone is a natural consequence of the theory of plate tectonics.

A double seismic zone is a double-planed distribution of intermediate depth earthquakes. It was first discovered beneath Tohoku, Japan, based on the data from local seismograph networks [Umino and Hasegawa, 1975; Hasegawa et al., 1978a,b]. The two planes of this double seismic zone are separated by about 40 km at the shallow end (a depth of ~60 km) [Stefani et al., 1982], and gradually merge at depth (~180 km). The focal mechanisms of the upper and lower planes are characterized by downdip compression and downdip tension, respectively. This stress (or strain) pattern is exactly the opposite of what is expected for bending of the plate, and thus it is what is expected when the plate is unbending.

This discovery of a double seismic zone beneath Tohoku, Japan, combined with the earlier suggestions of the presence of a two-layered seismic zone beneath Kanto, Japan [Tsumura, 1973], and in the Kuriles [Sykes, 1966; Veith, 1974], led to speculation that double seismic zones are a general worldwide feature of Wadati-Benioff zones [Isacks and Barazangi, 1977; Engdahl and Scholz, 1977]. Later, after compiling a large number of focal mechanism solutions for intermediate depth earthquakes in various subduction zones, Fujita and Kanamori [1981] suggested that double seismic zones were not a feature common to all subduction zones.

Double seismic zones have been reported beneath Tohoku, Japan [e.g., Umino and Hasegawa, 1975; Hasegawa et al., 1978a, b; Yoshii, 1979], Tohoku-Hokkaido junction, Japan [Umino et al., 1984], Hokkaido, Japan [Suzuki et al., 1983], Kurile-Kamchatka [Veith, 1974; Stauder and Mualchin, 1976], eastern Aleutians [Reyners and Coles, 1982; House and Jacob, 1983], and Peru [Isacks and Barazangi, 1977]. Suggestions of double seismic zones beneath the central Aleutians [Engdahl and Scholz, 1977] and Marianas [Samowitz and Forsyth, 1981] are rather questionable [Topper, 1978; Engdahl and Fujita, 1981]. Kawakatsu [1985a, b, 1986] recently suggested the presence of a double seismic zone in Tonga, where it has been known that downdip compressional stress dominates the stress state throughout the slab [Isacks and Molnar, 1969, 1971; Billington, 1980].

The cause of the double seismic zones has also been discussed by many researchers, but no consensus has been reached as yet. Proposed mechanisms are as follows: stresses associated with phase changes [Veith, 1974], unbending of the slabs [Engdahl and Scholz, 1977; Isacks and Barazangi, 1977; Tsukahara, 1977, 1980; Samowitz and Forsyth, 1981], sagging of the plate [Yoshii, 1977; Sleep, 1979], and thermo-elastic stresses [Goto and Hamaguchi, 1978, 1980; Goto et al., 1985; House and Jacob, 1982].

We suggest that it is appropriate and instructive to separate the problem into kinematical and dynamical parts. In this paper, we will focus on the kinematical part of the problem. Separating kinematics from dynamics, we are able to avoid difficult problems, such as estimation of forces acting on a plate. By estimating the seismic strain rate from intermediate depth earthquakes and the geometrical unbending strain rate of the subducting slabs, we first demonstrate that geometrical unbending of subducting slabs and the presence of a double seismic zone can be considered essentially equivalent. Following Tsukahara [1980], we then model a double seismic zone required from unbending in order to investigate what constraint the presence of a double seismic zone can place on the properties of the descending lithosphere. It will be shown that the thickness of a double seismic zone is controlled by the thermal structure and the rheological properties of the subducting slab. A preliminary result for the double seismic zone beneath Tohoku, Japan, indicates a possibility that the currently available flow law of "dry" olivine which is used for the rheology of subducting lithosphere does not properly describe the observed double seismic zone there, or at least that there exist some inconsistencies in our understanding of the mechanics of the subduction process.

[1]Now at Seismological Laboratory, California Institute of Technology, Pasadena.

Paper number 5B5689.
0148-0227/86/005B-5689$05.00

Strain Rates Within the Subducting Slabs

In regions such as subduction zones, where large-scale deformation is taking place rapidly, what properly describes the state of the deformation is the strain rate, rather than the stress. For example, in Figure 1 the principal stress (and strain) axes of the earthquakes which make up the Tonga double seismic zone [Kawakatsu, 1985a, b, 1986] are plotted. For most events, either the compression or the tension axis of each earthquake is located near the dip direction of the slab. Furthermore, the second axes (tension or compression) are clustered around a direction perpendicular to the slab. This suggests that a plane strain type deformation is occurring within the vertical plane perpendicular to the strike of the trench. A similar conclusion can be drawn from Appendix B of Fujita and Kanamori [1981], which summarizes observations from other subduction zones where double seismic zones are known to exist. In this section, assuming that the deformation is occurring in a two-dimensional manner, we make order of magnitude estimates of the seismic strain rates and the geometrical unbending strain rates of the intermediate depth part of the slabs.

Seismic Strain Rate

Following Richter [1979], the rate of the seismic moment release of intermediate depth (70 km$<$depth$<$150 km) earthquakes is estimated from events with m_b between 5.0 and 5.5, for which the seismic catalog (Preliminary Determination of Epicenters, PDE) is assumed to be relatively complete (see Figure 2). The depth range is chosen to avoid possible bias due to errors in depth estimates (particularly for shallow thrusting events). For events smaller than 5.0, there is a detection problem, and for the events larger than 5.5, there is a sampling problem (due to the short time window) and a problem due to the saturation of the body wave magnitude scale [Geller, 1976; Richter, 1979].

As the empirical relation between seismic moment (M_o) and body wave magnitude (m_b) for intermediate depth earthquakes (50-200 km), we use the following equation:

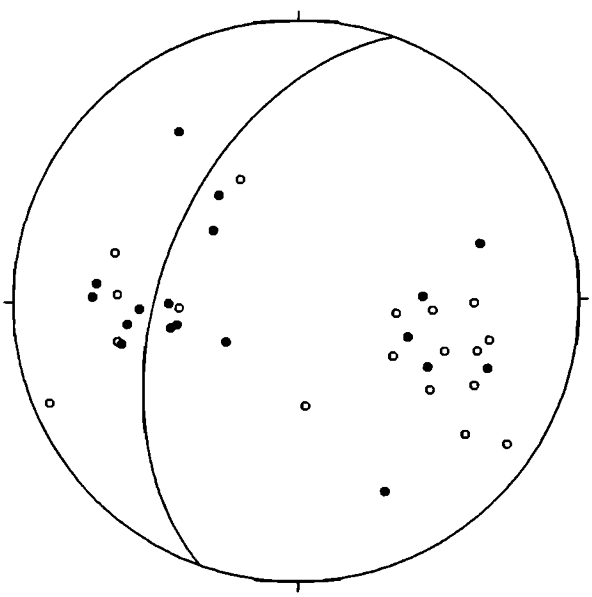

Fig. 1. The principal stress axes of the events which make up the Tonga double seismic zone are plotted on a lower hemisphere projection. Solid and open circles denote compression and tension axes, respectively. The great circle corresponds to the dip of the slab.

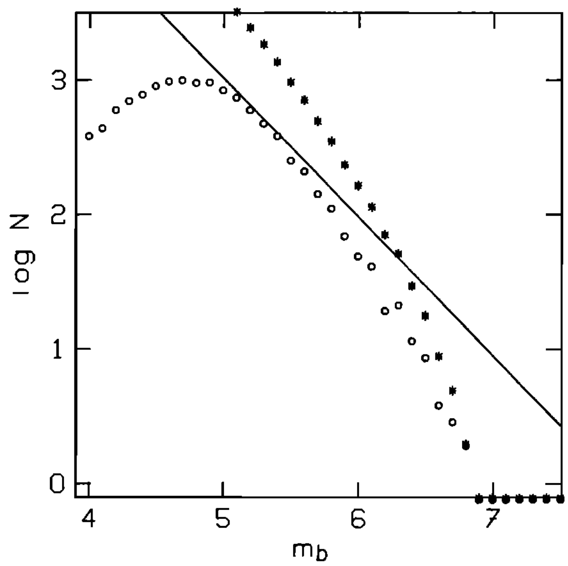

Fig. 2. Frequency-magnitude relation for all the earthquakes between depths 70 km and 150 km. Data are from PDE from 1964 to 1983. The circles are the number of events with m_b in 0.1 unit range($N_{(m_b)}$). Asterisks represent total number of earthquakes with m_b greater than the indicated value. The line has a slope of 1. Note the departure of the circles from this line above $m_b \sim 5.5$ and below 5.0, suggesting the saturation of the m_b scale and the incompleteness of data, respectively.

$$\log M_{o(m_b)} = 11+2.4m_b \tag{1}$$

where M_o is given in units of dyne centimeters. This is obtained rather arbitrarily (to be easily compared with Abe's [1982] formula) within the error bar of the least squares fit to moment-magnitude relation of the centroid moment tensor (CMT) solutions, and it slightly underestimates moments from magnitudes (Figure 3). Abe [1982] obtained $\log M_o = 10+2.4m_B$, where m_B is body wave magnitude measured from amplitudes in the period range 3-15 s. On the average, m_B is larger than m_b by ~0.7. Therefore (1) underestimates the moment by a factor of 5 compared to Abe's formula, which seems to overestimate the moment slightly (Figure 3). Giardini [1984] obtained a slightly different formula using a similar data set. Using (1), the total seismic moment released by earthquakes with m_b between 5.0 and 5.5 is first estimated by using

$$M_{o(5.0-5.5)} = \sum_{m_b=5.0}^{5.5} M_{o(m_b)} N_{(m_b)} \tag{2}$$

where $N_{(m_b)}$ is the number of events with magnitude m_b. Assuming the empirical relation between magnitude and frequency,

$$\log N_{(m_b)} = a - bm_b \tag{3}$$

where $b = 1$, the total seismic moment can be expressed as

$$M_{o(5.0-5.5)} = 10^{11+a} \sum_{5.0}^{5.5} 10^{1.4m_b} \tag{4}$$

It is then possible to extrapolate the total moment release estimate up to a certain maximum size of earthquakes [Smith, 1976],

$$M_{o(5.0-m_{bmax})} = \sum_{5.0}^{m_{bmax}} M_{o(m_b)} N_{(m_b)} = 10^{11+a} \sum_{5.0}^{m_{bmax}} 10^{1.4m_b}$$

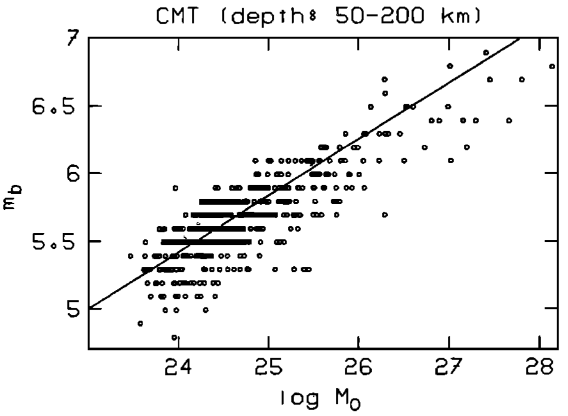

Fig. 3. Moment versus magnitude relation of CMT solution (depth 50-200 km). The line corresponds to equation (1), which is used to convert PDE magnitudes to seismic moments. Unit of seismic moment is dyne centimeters.

$$= M_{o(5.0-5.5)} \sum_{5.0}^{m_{bmax}} 10^{1.4m_b} / \sum_{5.0}^{5.5} 10^{1.4m_b} \quad (5)$$

$$\equiv \alpha_{m_{bmax}} M_{o(5.0-5.5)}$$

$\alpha_{m_{bmax}}$ is 5.6, 29, 146, and 740 for $m_{bmax} = 6$, 6.5, 7, and 7.5, respectively. Therefore under all these assumptions, once we know (or assume) a maximum size of earthquake that can occur in a certain area, it is possible to calculate the total seismic moment release for events with $m_b > 5.0$ from the seismic moment release of events with $5.0 \leq m_b \leq 5.5$. The effect of excluding events with $m_b < 5.0$ is negligible, since magnitude enters into (4) as a power of 10.

Figure 4 show the magnitude distribution of earthquakes in the depth range 70-150 km for each subduction zone. Data are from (PDE) in the time period 1964-1983. Using the method described above, total seismic moment release within slabs between 70 km and 150 km depth is estimated for each subduction zone. Average seismic moment release rates for the intermediate part of subducting slabs per unit surface area are then estimated by

$$\dot{M}_{o(m_{bmax})}(\text{dyn cm/km}^2\ \text{yr}) = M_{o(5.0-m_{bmax})}/(LWT) \quad (6)$$

where L, W, and T are the length along the dip between 70 and 150 km depth, the width of trench, and the total time period (20 years in this case), respectively. Corresponding seismic strain rates ($\dot{\epsilon}_s$) are obtained from

$$\dot{\epsilon}_{s(m_{bmax})} = \dot{M}_{o(m_{bmax})}/(2H_s\mu) \quad (7)$$

where H_s and μ are the thickness of the seismic zones (~40 km) and rigidity (~7×10^{11}dyn/cm^2), respectively. Here strain rates are uniaxial along the dip of the slab and earthquakes are assumed to occur on planes dipping 45° to the dip of the slab. Chapple and Forsyth [1979] used a similar approach in their study of bending earthquakes in the outer rise region. Table 1 summarizes the parameters used for various subduction zones. The poles of the arcs are estimated from the distribution of the earthquakes between depth 125 km and 175 km by least squares fit, and the lengths of the trenches are estimated using these poles for each subduction zones.

Table 2 summarizes the estimated strain rates. The quantities in the parentheses are the estimated $M_{o(5.0-5.5)}$ based on the observed number of earthquakes between m_b 5.0 and 5.5,

$$N_{(5.0-5.5)} = \sum_{5.0}^{5.5} N_{(m_b)}$$

instead of (2). The differences between the two numbers (before and in parentheses) are the simplest measures of the validity of assumption (3) in the magnitude range 5.0-5.5.

Quantification of earthquakes is a difficult problem, particularly for small earthquakes. Furthermore, the statistical assumptions which have been made may not hold. The values obtained in Table 2 should thus be considered no better than order of magnitude estimates. It is also safe to interpret the variation of estimated values among the different subduction zones as indicative of the error of this kind of analysis.

For the subduction zones where double seismic zones are observed, average seismic strain rates range from 1.0 to 8.1×10^{-17}s^{-1} and 0.3 to 2.1×10^{-15}s^{-1} for $m_{bmax} = 6.0$ and 7.0, respectively; $m_{bmax} = 6.0$ is a fairly small estimate of the maximum size of intermediate depth earthquakes. Therefore any mechanism which is proposed to explain the presence of double seismic zones must be able to produce strain rates at least of the order of ~10^{-17} s^{-1}.

The maximum thickness of the double seismic zone beneath Tohoku is ~40 km. Assuming that both halves of the double seismic zone are equally active, the maximum thickness of each is ~20 km. Considering that fault planes dip 45° with respect to the dip of the slab (i.e., one of the principal axes is downdip, Figure 1), the maximum length of the earthquake fault is ~30 km. The fault area of 30×30 km^2 corresponds to a seismic moment of 10^{26}~10^{27} dyn cm for a stress drop of 1 ~ 10 MPa [Kanamori and Anderson, 1975] and to a magnitude of 6.25 ~6.67 from (1). It is reasonable to think that the maximum magnitude m_{bmax} is ~6.5 and that seismic strain rate is ~10^{-16} s^{-1}.

Geometrical Unbending Strain Rate

Knowing the geometry of the slab, its thickness, and convergence velocity and assuming a steady state of subduction, it is possible to estimate the unbending strain rate. The unbending strain rate is proportional to the rate of change of the curvature and thus a function of the third derivatives of the surface geometry of a subducting plate. Thus it is almost impossible to measure the unbending strain rate accurately from the geometry of the slab observed from seismicity. It is, however, still possible to estimate the order of magnitude of an average unbending strain rate.

Figure 5 shows a bending-unbending model of subducting lithosphere. The lithosphere starts bending downward at the trench (T) and increases its curvature until it reaches the point of maximum curvature (S_0). Then it starts unbending and the curvature decreases until the point (S_1) beyond which it is straight. Assuming constant unbending, the rate of geometrical unbending (γ), which is the change of curvature, is given by

$$\gamma \equiv \frac{d\kappa}{ds} = 2(\theta_1-\theta_0)/(s_1-s_0)^2 \quad (8)$$

where κ is the curvature, s measures distance along the surface of the plate and θ is the dip angle. Since it is impossible to locate points S_0 and S_1 precisely from seismicity, we assume that unbending starts at the depth of 60 km and ends at the depth of 150 km. From a variety of published cross sections of seismicity (mostly from Isacks

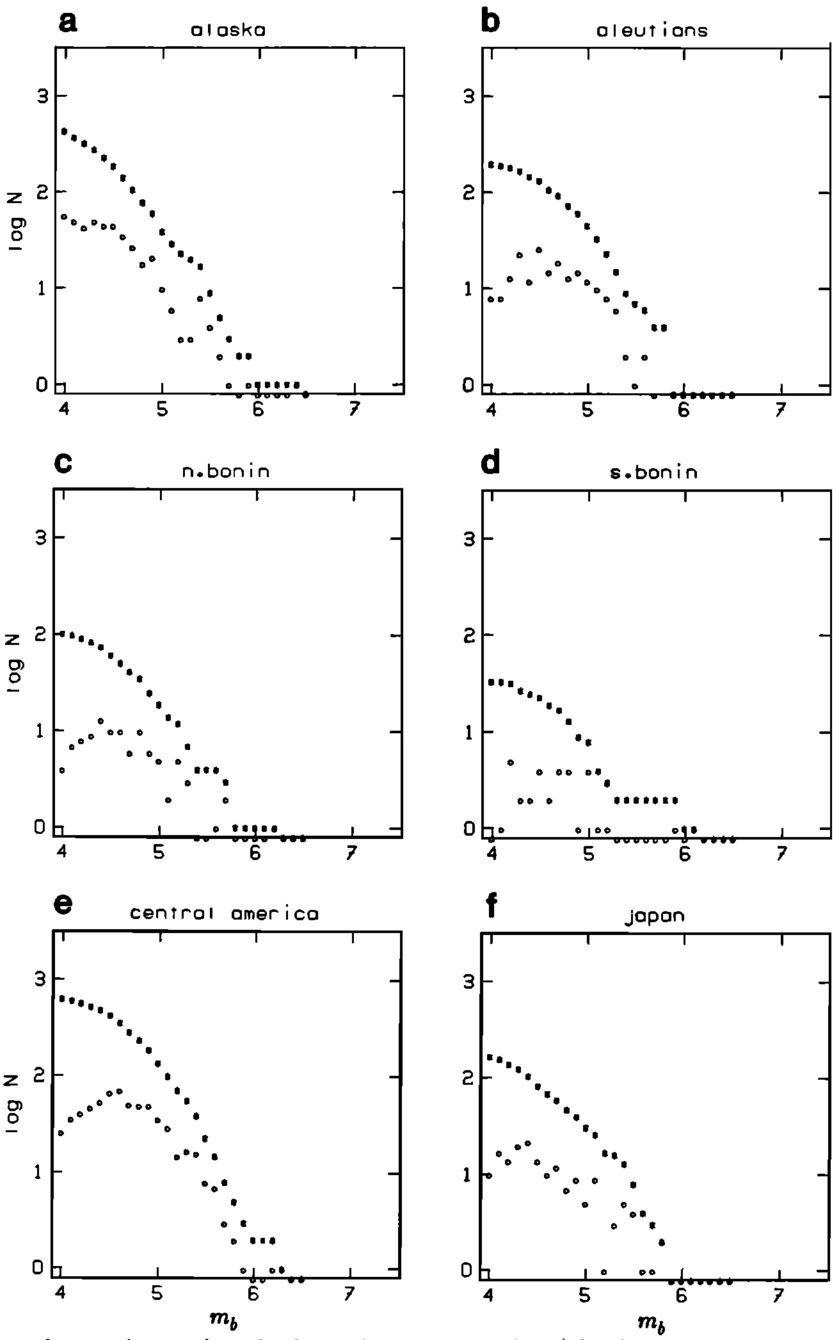

Fig. 4. Data used to estimate the seismic strain rate for each subduction zone. Arc parameters are in Table 1. Symbols as in Figure 2. (a) Alaska, (b) Aleutians, (c) northern Bonin, (d) southern Bonin, (e) central America, (f) Japan, (g) Mariana, (h) New Hebrides, (i) northern Ryukyu, (j) southern Ryukyu, (k) Scotia, (l) Sumatra, (m) Java, (n) Kamchatka, (o) northern Kuriles, (p) southern Kuriles, (q) Tonga, and (r) Kermedec.

and Barazangi [1977]), geometrical unbending rates are thus calculated (Table 3).

Assuming that the subduction process is steady state, it is possible to convert geometrical unbending rates into strain rates. Following Tsukahara [1980], a rectangular coordinate system is used as shown in Figure 5. When the lithosphere is deformed by pure unbending, the distribution of the uniaxial strain rate in the dip direction (i.e., direction of y axis) is given by

$$\dot{\epsilon}_y(x) = \dot{\Delta\epsilon}(x_n - x)/H \quad (9)$$

where H is the thickness of the lithosphere and $x = x_n$ represents the neutral plane. $\Delta\epsilon$ is the difference of the uniaxial strain rate at the upper surface of the plate and that at the lower surface and is related to the unbending rate as follows.

In bending models of a thin plate, the difference between the uniaxial strain at the top surface of the plate and that at the bottom one is

$$\Delta\epsilon = H/R = \kappa H \quad (10)$$

where R and κ are radius of curvature and curvature. It follows that

$$\dot{\Delta\epsilon} = \frac{d(\Delta\epsilon)}{dt} = H\frac{d\kappa}{dt} = \frac{ds}{dt}\frac{d\kappa}{ds}H = V\gamma H, \quad (11)$$

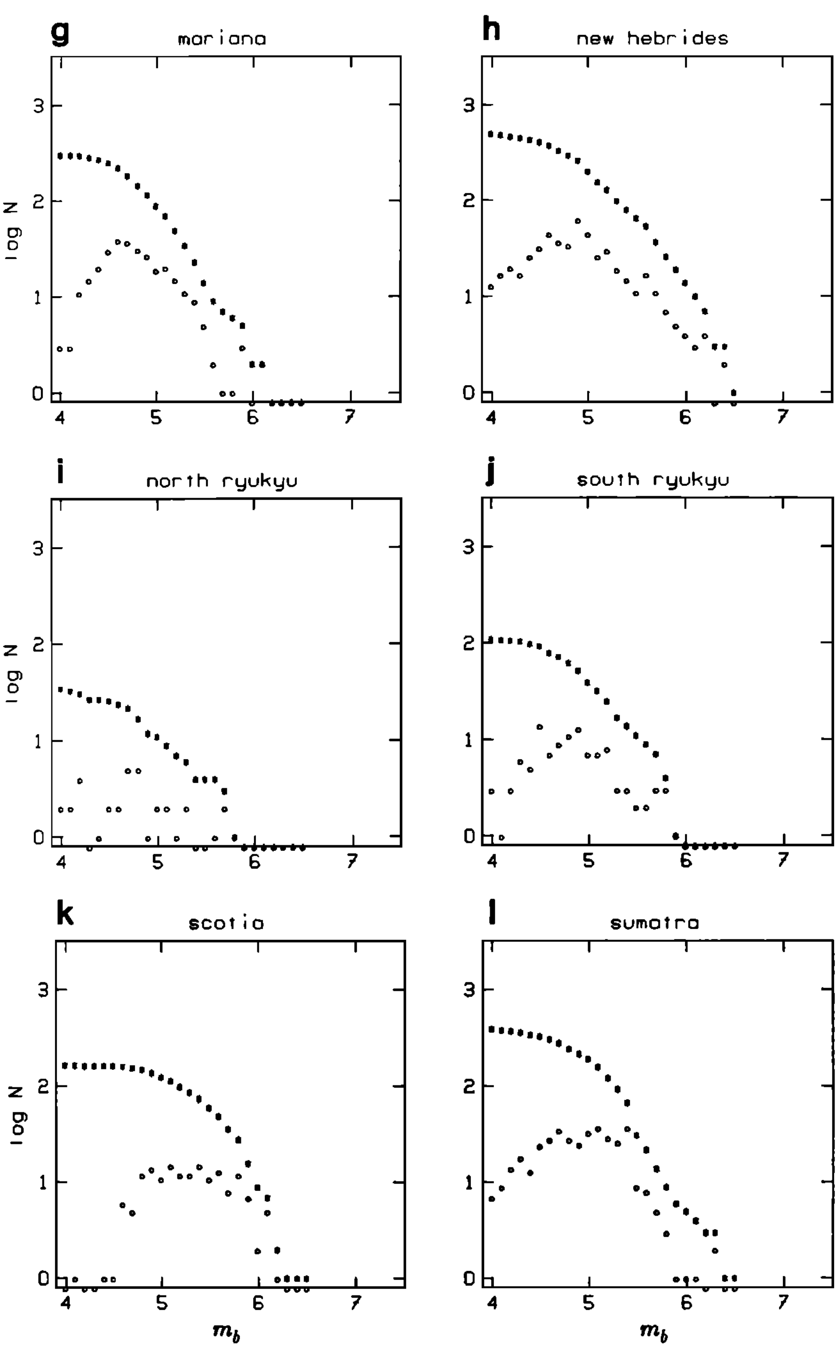

Fig. 4. (continued)

where V is the convergence velocity and γ is the unbending rate. Taking the thickness of the plate (H) as 100 km and the thickness of the seismic portion of the plate as 40 km, the average unbending strain rate over x within the seismic portion of the lithosphere is

$$\dot{\epsilon}_b = \frac{1}{4}\dot{\Delta\epsilon}\frac{40}{100} \tag{12}$$

where we assume that the neutral point, $x=x_n$ is located at the middle of the seismic portion. Within the seismic portion of the lithosphere, some of this unbending strain is released by earthquakes and the rest is released aseismically. In the lower portion of the plate, the temperature is so high that all the bending strain is released aseismically.

The average strain rates for different subduction zones are listed in the last column of Table 3, where the convergence velocity is set equal to 10 cm/yr. They are of the order of $10^{-16}\sim10^{-15}$ s^{-1}. These values exceed the seismic strain rate with $m_{bmax} = 6.0$ ($\epsilon_{s(6.0)}$) estimated in the previous section by a order of magnitude and are comparable to $\epsilon_{s(7.0)}$. Thus the strain rate expected from a simple geometrical unbending is comparable to that of intermediate depth earthquakes and can explain the double seismic zone.

It is often claimed that the geometry of the deeper portion (say, between 100 km and 200 km) of double seismic zone is apparently straight and that the unbending strain rate should be thus small [e.g., House and Jacob, 1982]. It is, however, easy to show that even an apparently straight part of a slab can have an unbending strain rate of $10^{-17}\sim10^{-16}$ s^{-1} as follows.

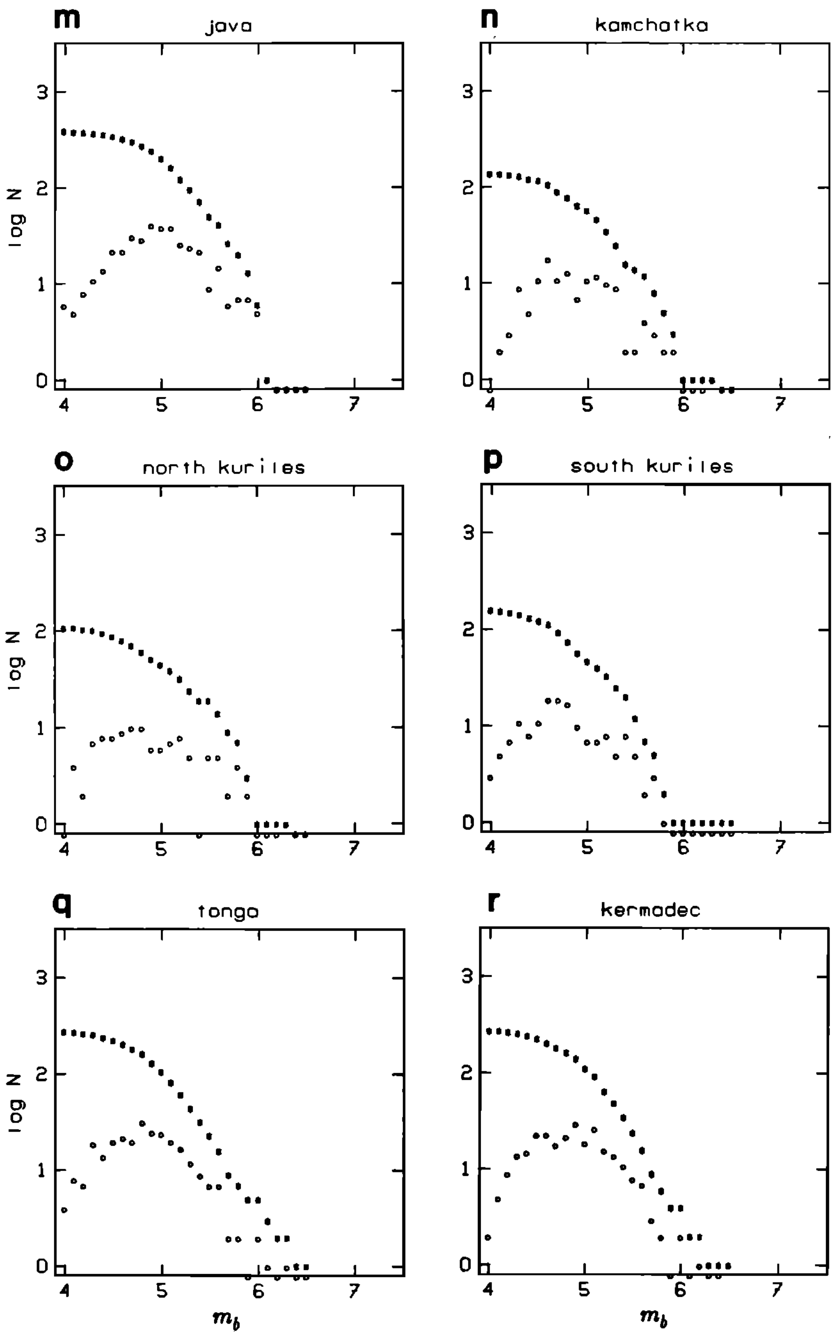

Fig. 4. (continued)

Let us take an average unbending strain rate ($\dot{\epsilon}_b$) equal to 10^{-17} s^{-1}. From (11) and (12), the corresponding unbending rate is

$$\gamma = 4\dot{\epsilon}_b/(VH_s) \sim 3.2\times10^{-7}\ km^{-2}$$

where H_s=40 km is the thickness of the seismic portion of the slab and V = 10 cm/yr. From (8), this unbending rate gives a change of the dip to be 0.09° between two points 100 km apart on the slab surface along the dip direction. We cannot detect such a small rate of change of the dip angle (0.09°/100 km) from seismicity. It is even impossible to detect a change of 0.9°/100 km, which corresponds to an unbending strain rate of 10^{-16} s^{-1}. In other words, even an apparently straight portion of the slab can easily have an unbending strain rate of $10^{-17}\sim10^{-16}s^{-1}$.

The stress distribution expected from this unbending strain is the same as that of double seismic zones (i.e., downdip compressional in the upper layer and downdip tensional in the lower layer). It is shown above that the strain rate due to this unbending is of the order of $10^{-16}\sim10^{-15}$ s^{-1} and is large enough to account for all the strain energy released by earthquakes, while the strain rate due to thermal expansion ($10^{-18}\sim10^{-17}$ s^{-1}) [House and Jacob, 1982] is too small. Unbending of the subducting plate is required as a natural consequence of the theory of plate tectonics. Therefore it is quite natural to think that the strain from this unbending, which must be released

TABLE 1. Arc Parameters

Arc	Pole (Latitude, Longitude)		Radius, deg	Left End Point		Right End Point		Dip, deg	Length, km
Alaska	65.0,	180.4	12.7	63.3,	209.5	57.0,	201.0	45	855
Aleutians	65.0,	180.4	12.7	57.0,	201.0	52.9,	173.0	55	1990
Northern Bonin	6.4,	75.1	65.4	33.5,	140.0	30.0,	140.5	50	391
Southern Bonin	6.4,	75.1	65.4	30.0,	140.5	25.0,	141.0	60	557
Central America	29.4,	282.3	19.3	11.3,	274.5	17.4,	264.7	50	1238
Japan	37.9,	113.1	21.1	40.5,	141.3	36.2,	140.0	30	476
Java	39.2,	120.8	47.7	-8.0,	124.0	-6.0,	106.0	60	2020
Kamchatka	57.8,	128.9	17.1	57.0,	160.0	48.5,	157.5	50	879
Northern Kurils	57.8,	128.9	17.1	48.5,	157.5	45.0,	152.5	45	485
Southern Kurils	57.8,	128.9	17.1	45.0,	152.5	44.0,	147.0	40	385
Tonga	22.4,	80.1	110.2	-16.0,	185.6	-24.0,	183.6	55	912
Kermadec	-22.4,	159.9	21.1	-25.0,	183.2	-36.0,	178.0	62	1333
Lesser Antilles	13.9,	285.1	13.2	18.0,	298.0	12.0,	299.0	45	669
Marianas	17.3,	139.1	6.3	22.5,	143.0	12.5,	143.0	70	1303
New Hebrides	-1.7,	-157.9	36.8	-21.0,	170.0	-11.0,	166.0	70	1191
Northern Ryukyu	35.5,	116.3	12.3	33.5,	133.5	28.2,	131.0	70	549
Southern Ryukyu	35.5,	116.3	12.3	28.2,	131.0	23.3,	125.0	45	702
Scotia	-58.0,	-31.8	3.0	-55.6,	331.7	-60.6,	332.3	50	598
Sumatra	33.2,	159.9	65.0	-6.0,	106.0	8.0,	94.0	70	2047

somehow, is partly released by earthquakes, producing double seismic zones.

Modeling a Double Seismic Zone

Considering the plastic-brittle property of the material composing the lithosphere, Tsukahara [1980] modeled the essential features of double seismic zones by a simple geometrical unbending of the subducting plate. In this section, using Tsukahara's formulation with more realistic parameters, we investigate what constraint the presence of a double seismic zone can put on the subduction process.

Brittle–Plastic Rheology

Figure 6 (after Figure 3 of Tsukahara [1980]) schematically shows the idealized model of brittle-plastic rheology. It should be noted that the term "brittle" fracture here represents any kind of instantaneous failure that could be detected as an earthquake. When the strain rate is small and the expected steady-state stress (σ_p) is smaller than the strength of the material (σ_s), deformation occurs aseismically (line A). (In the rest of this paper, stress σ always refers to the uniaxial differential stress $\sigma_1 - \sigma_3$, where σ_1 and σ_3 are the maximum and minimum principal stresses, respectively.) When the strain rate is so large that the expected steady state stress exceeds the strength of the material, fracture occurs (line B). In this case, deformation occurs both seismically and aseismically. The average stress in this case is fixed at the strength of the material.

For the plastic flow, we use the power law creep formula of "dry" olivine summarized by Kirby [1983]:

$$\dot{\epsilon} = A\, \sigma_p^n \exp[-(E^* + PV^*)/(RT)] \qquad (13)$$

TABLE 2. Seismic Strain Rates

Arc	$\dot{M}_{o(5.0-5.5)}$, 10^{19} dyn cm/km^2 yr	$\dot{\epsilon}_{s6.0}$, 10^{-17} s^{-1}	$\dot{\epsilon}_{s7.0}$, 10^{-15} s^{-1}
Alaska	0.9 (0.7)	2.9	0.7
Aleutians	0.3 (0.3)	1.0	0.3
Northern Bonin	0.5 (0.7)	1.6	0.4
Southern Bonin	0.1 (0.2)	0.3	0.1
Central America	1.7 (1.8)	5.5	1.4
Japan	0.9 (0.7)	2.9	0.8
Java	1.6 (1.6)	5.2	1.4
Kamchatka	0.8 (1.0)	2.6	0.7
Northern Kurils	1.2 (1.1)	3.9	1.0
Southern Kurils	2.1 (1.7)	6.8	1.8
Tonga	2.5 (2.5)	8.1	2.1
Kermadec	1.6 (1.6)	5.2	1.4
Lesser Antilles	0.1 (0.0)	0.3	0.1
Marianas	1.3 (1.4)	4.2	1.1
New Hebrides	2.7 (2.8)	8.8	2.3
Northern Ryukyu	0.2 (0.3)	0.6	0.2
Southern Ryukyu	0.7 (0.7)	2.3	0.6
Scotia	3.3 (2.4)	10.	2.8
Sumatra	2.1 (1.9)	6.8	1.8

Quantities in parentheses are the estimated $M_{o(5.0-5.5)}$ based on the observed number of earthquakes between m_b 5.0 and 5.5.

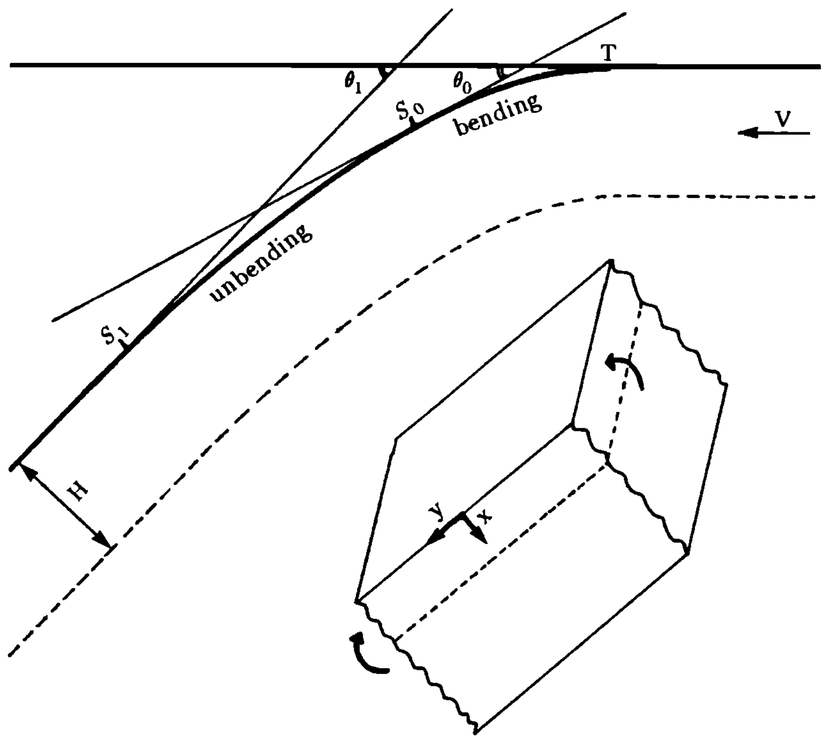

Fig. 5. Bending and unbending model of a subducting plate.

where P, R, and T are pressure, universal gas constant, and absolute temperature, respectively. A, n, E^*, and V^* are material constants and are given by

$$\log_{10} A = 4.8 \pm 1.2 \quad \text{A in } s^{-1}\,MPa^{-n} \tag{14a}$$

$$n = 3.5 \pm 0.6 \tag{14b}$$

$$E^* = 533 \pm 60 \ \text{kJ/mol} \tag{14c}$$

$$V^* = (17 \pm 4) \times 10^{-6} \ m^3/\text{mol} \tag{14d}$$

The range of differential stresses over which these power-law constants apply is 10-500 MPa [Kirby, 1983].

The average strength of the lithosphere (i.e., apparent stress in the lithosphere) has always been a point of controversy. Tsukahara [1980] chose it to be 5~20 MPa and Samowitz and Forsyth [1981] 500 MPa. Here, it is treated as a variable.

Distribution of Strain Rate

Only the strain rates due to pure unbending of a thin plate will be considered. We assume that the effect of the preexisting stress and strain before initiation of unbending is negligible. Considering the high unbending strain rate ($\sim 10^{-15}\,s^{-1}$), this assumption is reasonable. For discussion of the effect of an uniform, uniaxial, additional strain (e.g, strain due to so-called ridge push or slab pull forces), the reader should refer to the discussion by Tsukahara [1980]. The strain rate field is given by (9). The neutral plane,

TABLE 3. Geometrical Unbending Strain Rates (60-150 km Depth)

Arc	Slab Dip, deg		s_1–s_0	Unbending Rate,	Strain Rate*,
	60 km	150 km	km	10^{-5} km^{-2}	10^{-15} s^{-1}
Aleutians	29	53	134	4.6	1.5
Northern Bonin	39	55	120	4.0	1.3
Southern Bonin	42	65	109	6.6	2.1
Central America	44	74	102	10.0	3.3
Japan	28	32	173	0.41	0.13
Java	30	58	130	5.7	1.9
Kamchatka	37	48	130	2.1	0.68
Northern Kurils	37	46	132	1.8	0.59
Southern Kurils	37	46	143	1.6	0.52
Tonga	37	52	126	3.4	1.1
Kermadec	44	64	110	5.8	1.9
Marianas	40	67	110	7.6	2.5
New Hebrides	58	70	99	4.1	1.3
Northern Ryukyu	44	66	109	6.6	2.1
Southern Ryukyu	38	63	114	6.8	2.2

* Convergence rate of 10 cm/yr and plate thickness of 100 km are assumed.

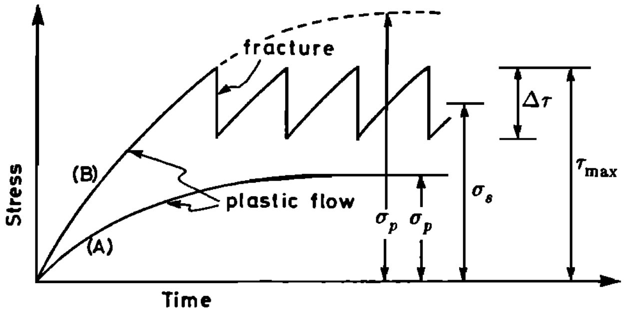

Fig. 6. Schematic illustration showing the idealized deformation with fracture and plastic flow. σ_s, σ_p, $\Delta\tau$, and τ_{max} denote average strength, stress due to steady state plastic flow, stress drop of earthquakes, and maximum strength, respectively. Line A, $\sigma_p \leq \tau_{max}$; line B, $\sigma_p > \tau_{max}$. After Tsukahara [1980].

$x = x_n$ can be obtained from the balance equation

$$\int_0^H \sigma(x)\,dx = 0 \tag{15}$$

where $\sigma(x)$ is the uniaxial unbending stress and can be calculated from (13) as a function of the strain rate. When the value exceeds the strength (σ_s) of the material, it is set equal to σ_s. Thus if $\Delta\epsilon$ in (12) (or the average unbending strain rate, $\dot{\epsilon}_b$) is specified, and if T(x) and P(x) are known, it is possible to calculate the strain field, $\dot{\epsilon}_y(x)$ and the stress field, $\sigma(x)$ to satisfy (9), (13), and (15).

Model

Figure 7 shows the temperature profile of the model of the subducting lithosphere. It is calculated by the finite difference scheme discussed by Minear and Toksoz [1970] and implemented by N. H. Sleep. Sleep's program incorporates an induced corner flow in the mantle wedge above the slab and entrained flow in the vicinity of the slab and thus predicts a lower temperature field within the slab than his previous models [e.g., Sleep, 1973; Fujita et al., 1981]. The thermal constants are chosen to be comparable to those of the seafloor spreading model of Parsons and Sclater [1977] (e.g., the thermal diffusivity is 0.88×10^{-6} m^2/s). The oceanic lithosphere subducts at a thermal age of 100 Ma (calculated by error function with a mantle temperature of 1370°C) with a velocity of 105 km/Ma and a dip angle of 0.5 rad. The computation is terminated at ~7 Ma after the initiation of subduction. These values are chosen such that the model mimics the subduction zone beneath Tohoku, Japan, where the nature of the double seismic zone is best known.

The stress and strain fields were modeled along the three profiles (depths at the slab surface are 50 km, 100 km, and 150 km) perpendicular to the slab dip (lines A-A′, B-B′, and C-C′ in Figure 7). Figure 8 shows temperature, hydrostatic pressure, and strain rate profiles. The difference of strain rate at the top and the bottom of the slab, $\Delta\epsilon$ is prescribed to be 10^{-15} s^{-1}, 10^{-16} s^{-1}, 10^{-17} s^{-1} for A-A′, B-B′, and C-C′, respectively. These numbers are chosen because the number of earthquakes beneath Tohoku apparently decreases by one order of magnitude between depth of ~60 km and ~110 km [Anderson et al., 1980, Figure 6].

The result is shown in Figure 9. Positive stress corresponds to downdip compression, and negative stress corresponds to downdip tension. Results of two different sets of parameters for (13) are shown. The thinner line in each box represents the stress distribution obtained from parameters in (14). The thicker lines are the results from parameters for which the viscosity takes the lowest value within the error range given in (14). The strength of the slab is set equal to 50 MPa. Parameters which give the highest viscosity within the error in (14) apparently give a viscosity so high that the whole slab (~100 km thick) becomes seismic.

Although the basic stress distribution of a double seismic zone (i.e, downdip compression in the upper part and downdip tension in the lower part) is consistently reproduced in the models (as expected from the unbending model), the thickness changes from one model to another. The temperature distributions in Figure 8 do not vary much among the different profiles. The thicknesses of the brittle failure zone (i.e., the thickness of the double seismic zone) in Figure 9 does not change in each column because the effect of pressure on viscosity cancels out the effect due to the change of strain rates.

The seismicity beneath Tohoku [e.g., Hasegawa et al., 1978a] shows that the thickness of the double seismic zone stays almost constant down to a surface depth of ~100 km, as predicted by the model. The lower plane is almost straight throughout the double seismic zone, which is also consistent with the model (i.e., the temperature profile does not change much along the dip in the lower portion of the slab; Figures 7 and 8a). The upper plane of the observed double seismic zone, on the other hand, starts bending (although it does not necessarily mean the plate itself is

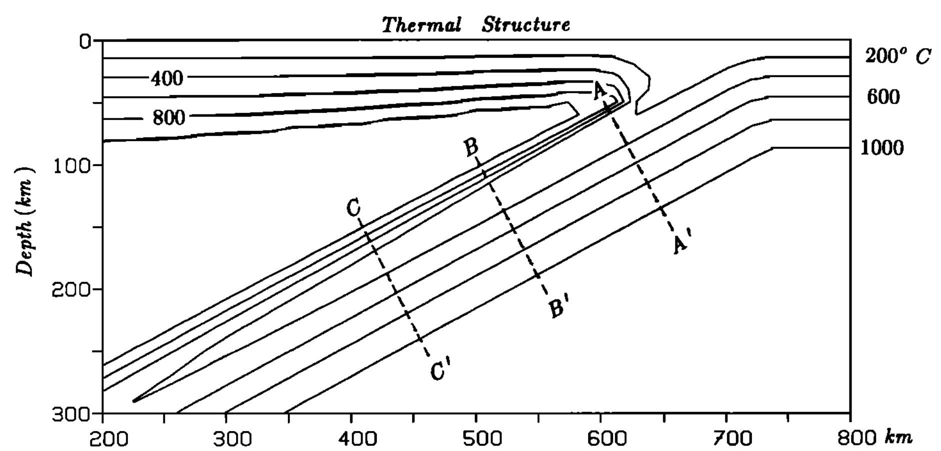

Fig. 7. Thermal model for the subduction zone beneath Tohoku, Japan. Line A-A′, slab surface depth 50 km; line B-B′, 100 km; line C-C′, 150 km.

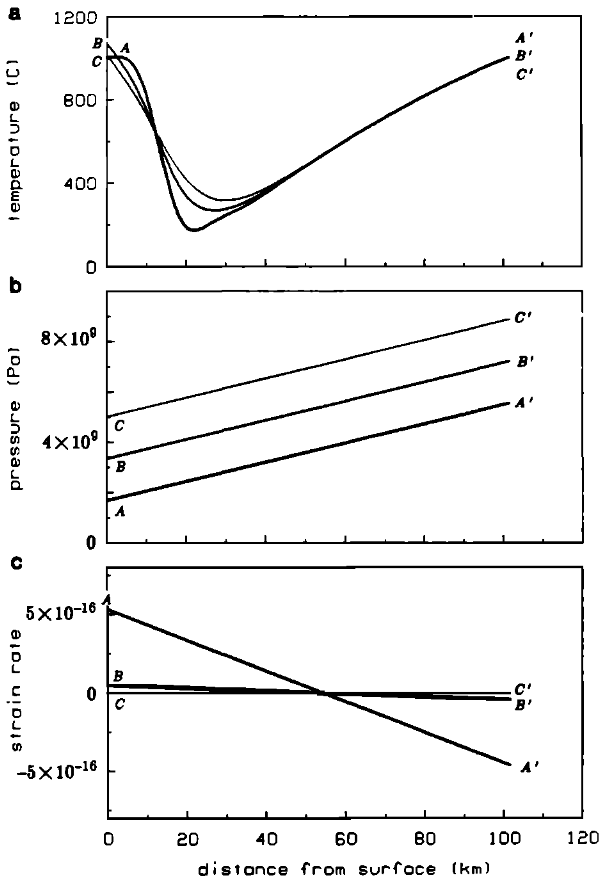

Fig. 8. Distributions of (a) temperature, (b) hydrostatic pressure, and (c) uniaxial unbending strain rate perpendicular to the dip of the slab. Thick, medium, and thin lines correspond to lines A-A′, B-B′, and C-C′ in Figure 7.

bending) toward the lower plane at a depth of ~100 km and they converge around ~180 km. This feature is not represented in the model, and it may mean that the upper portion of the slab tends to get heated faster than the temperature model predicts. The mantle temperature is held constant in the numerical modeling. The observed convergence of the two seismic zones may suggest that the temperature of the mantle beneath Japan is in fact higher than that of the mantle beneath the Pacific ocean [e.g., Honda, 1985]. In the second and the third column of Figure 9, in order to see the effect of temperature change, profiles are uniformly increased by 100°C and 200°C, respectively. The thickness of the double seismic zone decreases significantly. This suggests that the temperature is one of the essential factors controlling the double seismic zone.

The stress field model in Figure 9 does not show the spatial gap between the upper and the lower plane of the double seismic zone observed in seismicity, but the separation may be qualitatively understood. Since the unbending strain rate is always largest at the outer edges of both the compressional and the tensional seismic zones in the model, more earthquakes are expected to occur in the outer regions than inside. Rupture should also tend to initiate at the edges where the strain rate is highest. Since hypocentral locations are usually those of the initial point of rupture (as opposed to the centroid), they are expected to be located in the outermost region of both seismic zones.

In the present rheological model (Figure 6), the temperature of the outer edges of the double seismic zone (critical temperature, T_c, above which σ_p with a given strain rate is too small to cause earthquakes) is related to the strength of the slab through (13), where $T = T_c$ and $\sigma_p = \sigma_s$. Figure 10 shows this critical temperature as a function of depth with given strain rate and the strength. It shows that at the current level of knowledge of the rheology of the lithosphere, from observations of the location of the lower edge of the double seismic zone, it is only possible to predict the temperature within an error of $\sim\pm$ 100°C, even if the strain rate and the strength are known precisely. It also shows the significant effect of the strength on the critical temperature and the lesser significance of strain rates. On the other hand, this means that if the temperature structure can be constrained by other means (e.g., petrologically), it is possible to constrain the strength of the lithosphere.

Figure 11 shows the relation between the strength and the critical temperature at a depth of 80 km, where the lower plane of the double seismic zone in Tohoku appears to start. Strain rates are the highest around this depth and probably of the order of 10^{-15} s^{-1}. In Figure 12, seismicity beneath Tohoku [after Hasegawa et al., 1978a] is plotted with the thermal model. The thick line denotes 600°C isotherm, and most events of the double seismic zone are located inside of this isotherm. A critical temperature of 600°C at a depth of 80 km predicts an unrealistically high value for the strength of the lithosphere (off scale in Figure 11). At a depth of 80 km, the effect of heat conduction from the surrounding mantle is small, and the thermal structure of the inside of the slab is almost static and similar to that of the oceanic lithosphere at the trench. The thermal structure of the oceanic lithosphere is fairly well known [Parsons and Sclater, 1977]. Although their estimates of the standard errors for each parameter are large, they are primarily due to the coupling among the parameters and do not necessarily mean the thermal structure is uncertain. The essential parameter which determines the thermal structure is the temperature at the bottom of the lithosphere. If we assume that the bottom temperature is correct within ±100°C, then the thermal model in this depth range (~80 km) should not be wrong by more than 100°C. From bending earthquakes in the outer rise region in Tonga, Wortel [1982] estimated this critical temperature to be 580±70°C at a depth of ~45 km. Data on earthquakes occurring near oceanic trenches, compiled by Chen and Molnar [1983], also appear to suggest 600°C for the critical temperature. Even taking the highest estimate of the critical temperature to be 700°C and the lowest estimate of viscosity, the strength still has to be bigger than 500 MPa. This implies either that the lithosphere can sustain a differential stress higher than 500 MPa without having any earthquakes, that the olivine flow law of (13) and (14) is not appropriate for the lithosphere, or that the thermal model is significantly incorrect (i.e., by much more than 100°C).

Since the olivine flow law of (13) and (14) is that of "dry" olivine, it is possible that neglecting the effect of water causes the unreasonably high computed strength for the lithosphere. Neither the flow law for "wet" olivine nor the water content in the lithosphere at depths is well known [Kirby, 1983]. Thus it is not possible at this stage to determine whether or not the existence of a double seismic zone requires that high stresses exist in the lithosphere.

It is also possible that the flow law obtained of the laboratory scale (both in time and space) is not applicable to geologic scale phenomena [e.g, McNutt and Menard, 1982]. Figure 13 shows the relation between the yield strength and activation energy, $Q = E^* + PV^*$ for two different critical temperatures. If the strength of the lithosphere is on the same order as the stress drop of earthquakes (~10 MPa), Figure 13 implies that the activation

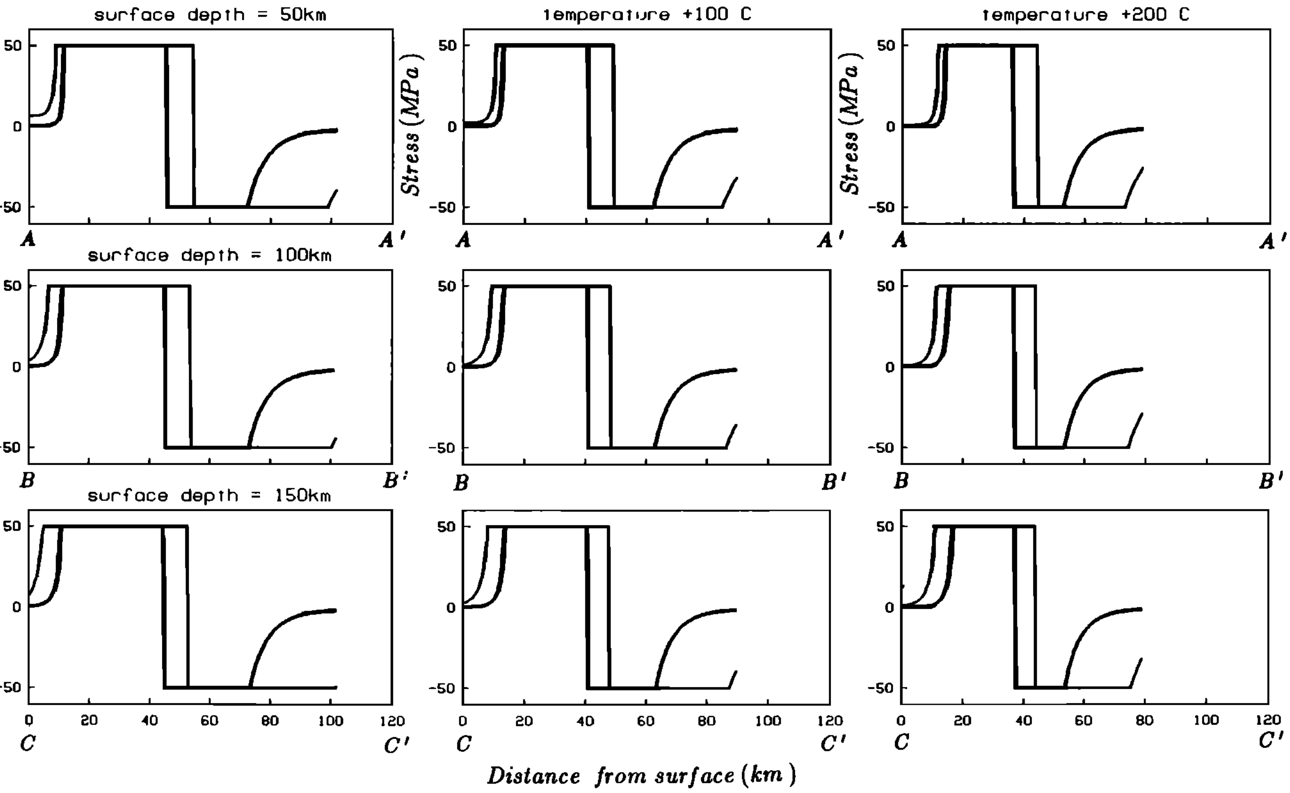

Fig. 9. Stress field perpendicular to the dip of the slab. Thin lines are the results using the average values in (14) and the thicker lines are using the values in (14), with which the viscosity is the smallest. First, second, and third row are for profiles A-A′ , B-B′ , and C-C′ , respectively. For the second and the third column, temperatures are uniformly increased by 100°C and 200°C, respectively.

energy has to be less than ~450 kJ/mol. Since the activation energy Q is expected to be more than 600 kJ/mol in (14) at a depth of 80 km, this suggests that the flow law obtained in laboratories may be seriously in error when applied to geologic scales. Brittle behavior of rock (or lithosphere) may be scale dependent (due to inhomogeneities), and this may change the flow law for the semibrittle regime (or brittle-ductile transition [Kirby, 1980]). Thus both strength and activation energy could be much smaller than the values obtained from laboratory experiments for the semibrittle regime. Estimating the bending moment of oceanic lithosphere following Goetze and Evans' [1979] formulation, McNutt and Menard [1982] suggested that the activation energy was significantly less than 420 kJ/mol [Kirby, 1983]. This value also predicts a relatively low yield strength (<50 MPa) at the lower edge of the double seismic zone.

The thermal model itself could be significantly incorrect, possibly due to our inability to determine the temperature at the bottom of the oceanic lithosphere. Possible effects of shear heating at the interplate thrust zone are also not included [e.g, Honda, 1985]. At this moment, we do not have access to the earthquake location data; therefore errors involved with seismicity can not be estimated. Thus definitive conclusions cannot be drawn from the preceding arguments which should be considered as a preliminary analysis. We, nevertheless, showed such potential that studying double seismic zones can constrain the thermal or mechanical properties of the oceanic lithosphere and demonstrated a possible inconsistency in our understanding of those properties.

The stress field calculated here is an instantaneous stress field, and neither the variation along the dip of the slab nor that in time is considered. The subducting slab with a strain rate of 10^{-16} s^{-1} and a velocity of 100 km/m.y. moves 3 km in the dip direction to obtain a strain of 10^{-4}. If the average critical strain of 10^{-5}~10^{-4}, obtained from shallow earthquakes [Tsuboi, 1933; Kanamori and Anderson, 1975; Rikitake, 1976], is applicable for intermediate earthquakes, it is, therefore, not necessary to consider the variation in time and in space.

What Unbends the Slab?: Dynamics

We have so far concentrated just on the geometrical (and kinematical) part of the problem and made no dynamical considerations. A question still remains unsolved: what unbends the slabs in the mantle? The answer probably varies from one subduction zone to another, and we only list the possible mechanisms. (1) Pressure due to the mantle flow [e.g., Hager and O'Connell, 1978; Yokokura, 1981]: Large-scale mantle flow exerts forces normal to the slab. Dynamical sagging of Sleep [1979] is essentially the same as this, because the normal force due to the mantle flow is, on the first order, proportional to the viscosity and the viscosity increases in the mesosphere. (2) Gravitational forces: Even without the flow pressure, the gravitational force itself unbends the slab when the slab dips too steeply (e.g., Marianas). (3) Overriding plate push: When an overriding plate is moving toward the trench, the slab tends to maintain a shallower dip angle, resulting in unbending. The strain rate expected from the thermal expansion of the slabs is 10^{-18}~10^{-17} s^{-1} [House and Jacob, 1982] and a few orders of magnitude smaller than the unbending strain rate. Therefore the thermo-elastic stress cannot unbend the slab and cannot be the main cause of the double seismic zone.

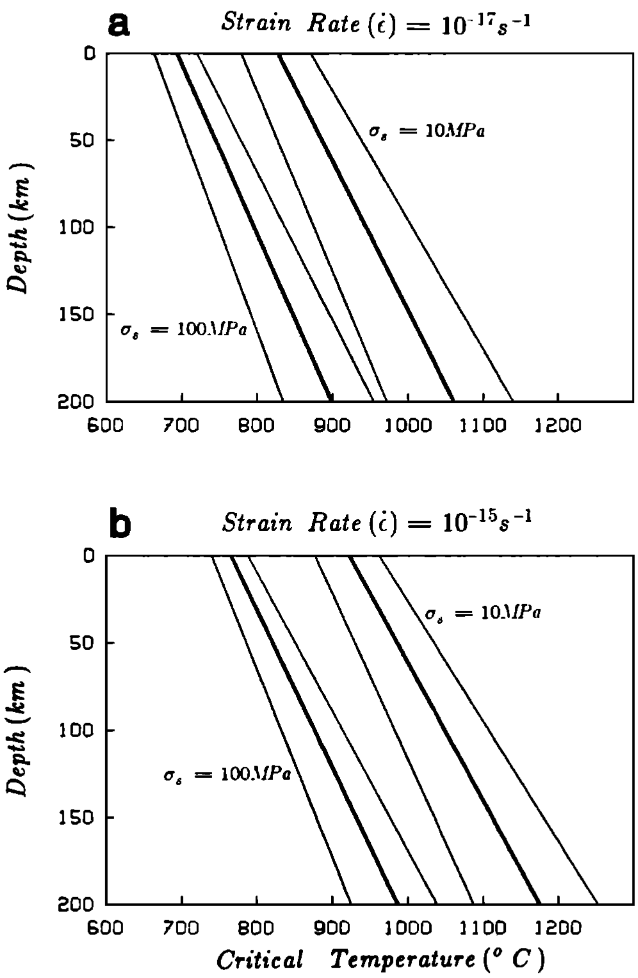

Fig. 10. Critical temperature as a function of depth with given strain rates and strength. Thick lines are for the average values in (14), and the thin ones surrounding them show the effects of uncertainties in these parameters.

Discussions

Fujita and Kanamori [1981] suggested that the age of the slab and the arc normal convergence rate are the two major factors which control the stress state within the slab and the presence of a double seismic zone. Although the general effects of these two factors on the intermediate depth stress are as they describe, force balances and the dynamics of the subduction process are probably much more complicated, and it is questionable whether their line of argument can be applied to all subduction zones in the world. It seems more appropriate to apply it to local problems [e.g., Forsyth, 1975; Shiono et al., 1980].

If the stress state of the slab is dominated by downdip compression (or downdip tension) in the intermediate depth range where unbending is taking place, it means that there is an additional uniaxial compressional (or tensional) strain whose magnitude is equal to or more than that of the maximum unbending strain. The average unbending strain rates between depths of 60 km and 150 km are estimated to be of the order of 10^{-16}~10^{-15} s^{-1}. Above a depth of 100 km where the unbending takes place relatively rapidly, the actual strain rate should be higher than the average value. Therefore a uniaxial strain rate of ~10^{-15} s^{-1} due to some force other than unbending is required for the slab to be totally compressional (or tensional) and not to have a double seismic zone.

A uniform uniaxial compressional (tensional) strain rate of ~10^{-15} s^{-1} corresponds to a shortening (elongation) of ~5%, when the slab subducts 160 km along dip with a velocity of 10 cm/yr. This rate of deformation seems to be large and may require a highly localized deformation, such as necking or (brittle) buckling of the slab. Therefore neither compressional nor tensile stresses may be completely dominating the stress state of the slab in the intermediate depth range where unbending is taking place, and thus double seismic zones may exist in most subduction zones where deformation is mainly taking place in a two-dimensional manner. In other words, unbending strain rates dominate the deformation of slabs, and the state of stress in the intermediate depth range and additional uniaxial strains simply shift the neutral stress plane of a double seismic zone higher or lower depending on their sign.

Based on the above discussion and considering the suggestion of a double seismic zone in Tonga [Kawakatsu, 1985a, b, 1986], we suggest that a double seismic zone is a more common feature of subducting lithosphere at a intermediate depth and may be found in many subduction zones where the deformation is taking place mainly in two-dimensional manner. The observation of double seismic zones may have been hampered by the lack of high-quality local networks in most subduction zones and the fact that the time span of reliable observations of earthquake focal mechanisms is too short.

Fujita and Kanamori [1981] pointed out that "young and slow" slabs (Ryukyu, Aleutians, Scotia) had thin seismic zones and that it became easier for tensile or compressive stresses to dominate short segments of the slab

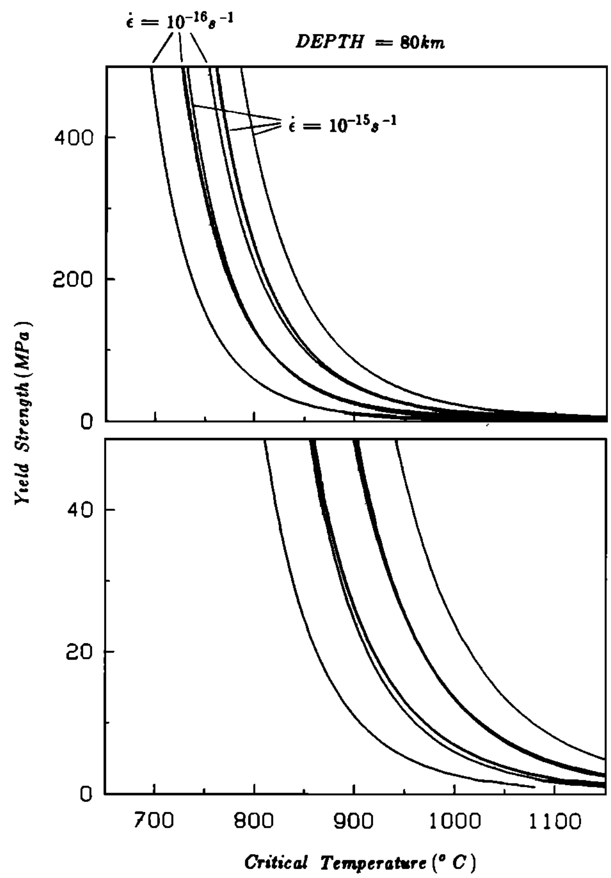

Fig. 11. Critical temperature as a function of yield strength at a depth of 80 km, where the double seismic zone appears to start. Lines are the same as in Figure 10. Strain rates are 10^{-16} s^{-1} and 10^{-15} s^{-1}.

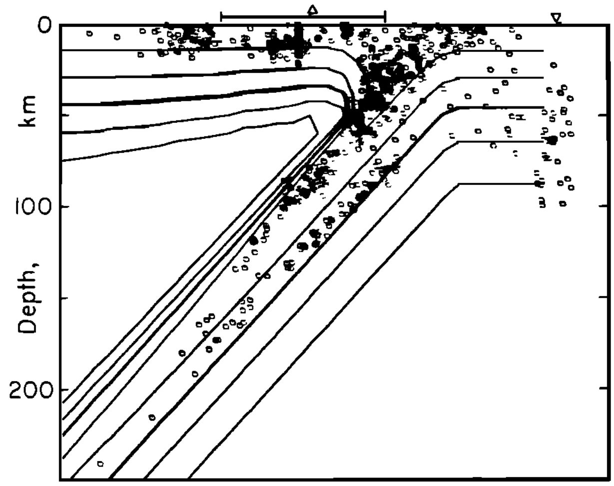

Fig. 12. Double seismic zone beneath Tohoku [after Hasegawa et al., 1978a] is compared with the thermal model of Figure 7. Contours are drawn for every 200°C. The thick contour corresponds to 600°C isotherm. The figure is vertically exaggerated by a 2:1 ratio.

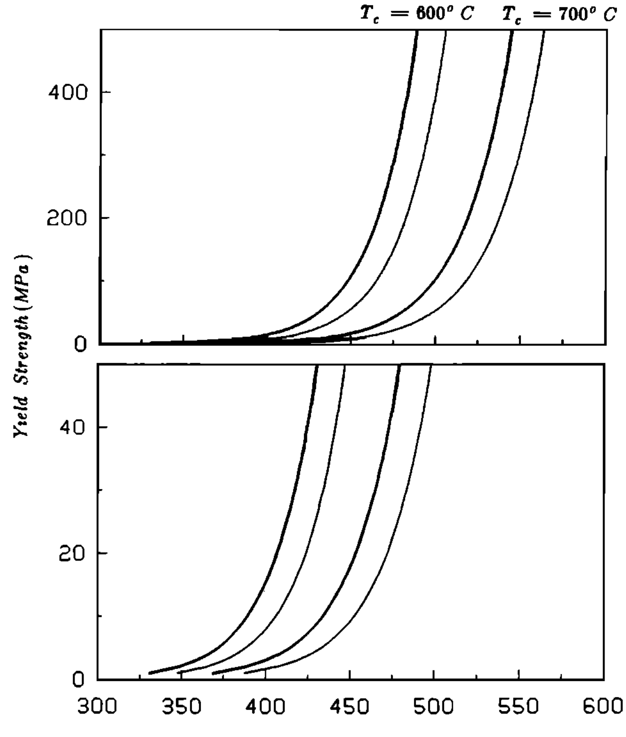

Fig. 13. Yield strength as a function of activation energy with $T_c = 600°C$ and 700°C. Thick and thin lines are results with strain rates 10^{-15} s^{-1} and 10^{-16} s^{-1}, respectively.

with no single segment exhibiting both. They called these slabs "stress segmented" because short segments in close proximity show different stress states. The explanation of Fujita and Kanamori [1981] is certainly attractive and probably true to some extent, but uniaxial strain rates of the order of 10^{-15} s^{-1} are still required for either tensile or compressive stresses to dominate completely the stress fields there. Therefore, we suggest that tensional and compressional earthquakes may be actually located in the lower and upper edges of seismic zones, respectively, and that if reliable data for smaller earthquakes become available, double seismic zones may be found (see addendum).

The activity of deep (>300 km) earthquakes in Tonga is the highest of the all subduction zones, and the compressional stress (or strain) is expected to be transmitted upward more than anywhere else in the world. The presence of a double seismic zone in Tonga therefore seems to indicate that the compressional stresses would not be likely to dominate the stress field of the intermediate part of the slab anywhere in the world at present. On this basis, we suggest that a double seismic zone may exist beneath the Izu-Bonin arc where the general seismicity is similar to that of Tonga [Fujita and Kanamori, 1981]. Further speculations concerning where double seismic zones may exist can be found in the work by Kawakatsu [1985a].

Although the unbending strain rate is large and is expected to exist within slabs over a long time average, it may be possible that temporally the stress pattern of a double seismic zone due to unbending is significantly disturbed by local stress perturbations (e.g., before and after large earthquakes) and one of the two zones is masked. Kawakatsu and Seno [1983] reported the temporal and spatial variation of activity of the double seismic zone beneath the northern Honshu arc. They attributed the behavior to the variation of seismic coupling at the interplate thrust zones; i.e., there are more downdip tensional events and fewer downdip compressional events in double seismic zones prior to the occurrence of large thrust earthquakes. A similar feature for bending earthquakes is also reported by Christensen and Ruff [1983]. Even in this case, it is expected that downdip compressional and downdip tensional events tend to occur near the upper and lower edges of the seismic portion of the downgoing slab and over a long time average there may be still a double seismic zone.

Concluding Remarks

The strain rate due to geometrical unbending of the subducting slab, which is suggested equivalent to the presence of a double seismic zone, is large, and it is expected to exist within all subducting slabs if they are straight at some depth in the mantle. This strain due to unbending has to be released by some mechanism, and if it is released by earthquakes, a double seismic zone results. A double seismic zone of intermediate depth earthquakes, therefore, is a natural consequence of the theory of plate tectonics, and its presence within a subducting lithosphere can be considered as another line of supporting evidence for the theory, which states not only that the Wadati-Benioff zone is the trace of subducted oceanic lithosphere but also that the oceanic lithosphere is moving as a rigid plate.

Modeling the double seismic zone beneath Tohoku, we showed that the properties of the oceanic lithosphere could be constrained. Studying double seismic zones has an advantage over studying other areas (e.g., outer rise [e.g., Chapple and Forsyth, 1979], intraplate earthquakes [e.g., Wiens and Stein, 1983]) for the purpose of constraining the mechanical properties of oceanic lithosphere because seismicity of microearthquakes of the double seismic zone in some area (e.g., Tohoku, Japan) is well studied and the strain rate can be reasonably well estimated. We believe that studying double seismic zones in many subduction zones will give a rare opportunity to constrain the mechanical properties of the oceanic lithosphere.

Addendum

During the process of the revision of this paper, Ishikawa [1985] reported a possible existence of a double seismic zone beneath Kyushu, Japan (a region classified as northern Ryukyu by Fujita and Kanamori [1981]), in the letter section of Zisin. Ishikawa relocated earthquakes using data of the Japan Meteorological Agency and showed the possibility. Although the number of events studied is small and the focal mechanism for none of the events in the upper plane is determined, his suggestion seems reasonable. This suggests that Shiono et al. [1980], who used only teleseismic data, might have been observing only the earthquakes in the lower plane of the possible double seismic zone. If that is the case, Ishikawa's result is consistent with this work (see also Kawakatsu [1985a]).

Acknowledgments. I thank Norm Sleep for his computer program of thermal modeling and general discussions. I am grateful to Eric Peterson, Bob Geller, Heidi Houston, and Rob Haar for their comments on the manuscript. I thank Kazu Fujita for sending me a copy of Ishikawa's paper. Discussions with Satoru Honda helped to clarify the contents of this paper. I thank J. Dewey for his comments on the manuscript as the refree and D. Giardini of Harvard University for sending me a tape of CMT solutions. This research was supported by National Science Foundation grants EAR82-18961 and EAR83-06555.

References

Abe, K., Magnitude, seismic moment and apparent stress for major deep earthquakes, J. Phys. Earth, 30, 321-330, 1982.

Anderson, R. N., A. Hasegawa, N. Umino, and A. Takagi, Phase changes and the frequency-magnitude distribution in the upper plane of the deep seismic zone beneath Tohoku, Japan, J. Geophys. Res., 85, 1389-1398, 1980.

Billington, S., The morphology and tectonics of the subducted lithosphere in the Tonga-Fiji-Kermadec region from seismicity and focal mechanism solutions, Ph.D. thesis, 220 pp., Cornell Univ., Ithaca, N. Y., 1980.

Chapple, W. M., and D. W. Forsyth, Earthquakes and bending of plates at trenches, J. Geophys. Res., 84, 6729-6749, 1979.

Chen, W.-P., and P. Molnar, Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res., 88, 4183-4214, 1983.

Christensen, D. H., and L. J. Ruff, Outer-rise earthquakes and seismic coupling, Geophys. Res. Lett., 10, 697-700, 1983.

Engdahl, E. R., and K. Fujita, Comment on "Double seismic zone beneath the Mariana Island arc" by Ilene R. Samowitz and Donald W. Forsyth, J. Geophys. Res., 86, 7023-7024, 1981.

Engdahl, E. R., and C. H. Scholz, A double Benioff zone beneath the central Aleutians: An unbending of the lithosphere, Geophys. Res. Lett., 4, 473-476, 1977.

Forsyth, D. W., Fault plane solutions and tectonics of the South Atlantic and Scotia Sea, J. Geophys. Res., 80, 1429-1443, 1975.

Fujita, K., and H. Kanamori, Double seismic zones and stresses of intermediate depth earthquakes, Geophys. J. R. Astron. Soc., 66, 131-156, 1981.

Fujita, K., E. R. Engdahl, and N. H. Sleep, Subduction zone calibration and teleseismic relocation of thrust zone events in the Central Aleutian islands, Bull. Seismol. Soc. Am., 71, 1805-1828, 1981.

Geller, R. J., Scaling relations for earthquake source parameters and magnitude, Bull. Seismol. Soc. Am., 66, 1501-1523, 1976.

Giardini, D., Systematic analysis of deep seismicity: 200 centroid-moment tensor solutions for earthquakes between 1977 and 1980, Geophys. J. R. Astron. Soc., 77, 883-914, 1984.

Giardini, D., and J. H. Woodhouse, Deep seismicity and modes of deformation in Tonga subduction zone, Nature, 307, 505-509, 1984.

Goetze, C., and B. Evans, Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics, Geophys. J. R. Astron. Soc., 59, 463-478, 1979.

Goto, K., and H. Hamaguchi, A double-planed structure of the intermediate seismic zone-Thermal stress within the descending lithospheric slab (in Japanese), Annu. Meet. Seismol. Soc. Jpn. Abstr., 2, 36, 1978.

Goto, K., and H. Hamaguchi, Distribution of thermal stress within the descending lithospheric slab (in Japanese), Annu. Meet. Seismol. Soc. Jpn. Abstr., 2, 48, 1980.

Goto, K., H. Hamaguchi, and Z. Suzuki, Earthquake generating stresses in a descending slab, Tectonophysics, 112, 111-128, 1985.

Hager, B. H., and R. J. O'Connell, Subduction zone dip angles and flow driven by plates, Tectonophysics, 50, 111-133, 1978.

Hasegawa, A., N. Umino, and A. Takagi, Double-planed structure of the deep seismic zone in the northeastern Japan arc, Tectonophysics, 47, 43-58, 1978a.

Hasegawa, A., N. Umino, and A. Takagi, Double-planed deep seismic zone and upper-mantle structure in the northeastern Japan arc, Geophys. J. R. Astron. Soc., 54, 281-296, 1978b.

Honda, S., Thermal structure beneath Tohoku, northeast Japan-A case study for understanding the detailed thermal structure of the subduction zone, Tectonophysics, 112, 69-102, 1985.

House, L. S., and K. H. Jacob, Thermal stresses in subducting lithosphere can explain double seismic zones, Nature, 295, 587-589, 1982.

House, L. S., and K. H. Jacob, Earthquakes, plate subduction, and stress reversals in the eastern Aleutian arc, J. Geophys. Res., 88, 9347-9373, 1983.

Isacks, B. L., and M. Barazangi, Geometry of Benioff zones: Lateral segmentation and downwards bending of the subducted lithosphere, in Island Arcs, Deep Sea Trenchs and Back-Arc Basins, Maurice Ewing Ser., vol. 1, edited by M. Talwani and W. C. Pitman III, pp. 99-114, AGU, Washington, D. C., 1977.

Isacks, B. L., and P. Molnar, Mantle earthquake mechanisms and the sinking of the lithosphere, Nature, 223, 1121-1124, 1969.

Isacks, B. L., and P. Molnar, Distribution of stresses in the descending lithosphere from a global survey of focal-mechanism solutions of mantle earthquakes, Rev. Geophys., 9, 103-174, 1971.

Ishikawa, Y., Double seismic zone beneath Kyushu, Japan (in Japanese), Zisin, 38, 265-269, 1985.

Kanamori, H., and D. L. Anderson, Theoretical basis of some empirical relations in seismology, Bull. Seismol. Soc. Am., 65, 1073-1095, 1975.

Kawakatsu, H., Double seismic zones: A first order feature of plate tectonics, Ph.D. thesis, Stanford Univ., Stanford, Calif., 1985a.

Kawakatsu, H., Double seismic zone in Tonga, Nature, 316, 53-55, 1985b.

Kawakatsu, H., Downdip tensional earthquakes beneath the Tonga arc-A double seismic zone?, J. Geophys. Res., in press, 1986.

Kawakatsu, H., and T. Seno, Triple seismic zone and the regional variation of seismicity along the northern Honshu arc, J. Geophys. Res., 88, 4215-4230, 1983.

Kirby, S. H., Tectonic stresses in the lithosphere: Constraints provided by the experimental deformation of rocks, J. Geophys. Res., 85, 6353-6363, 1980.

Kirby, S. H., Rheology of the lithosphere, Rev. Geophys., 21, 1458-1487, 1983.

McNutt, M. K., and H. W. Menard, Constraints on yield strength in the oceanic lithosphere derived from observations of flexure, Geophys. J. R. Astron. Soc., 71, 363-394, 1982.

Minear, J. W., and M. N. Toksöz, Thermal regime of a downgoing slab and new global tectonics, J. Geophys. Res., 75, 1397-1419, 1970.

Parsons, B., and J. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res., 82, 803-827, 1977.

Reyners, M., and K. S. Coles, Fine structure of the dipping seismic zone and subduction mechanics in the Shumagin islands, Alaska, J. Geophys. Res., 87, 356-366, 1982.

Richter, F. M., Focal mechanisms and seismic energy release of deep and intermediate earthquakes in the Tonga-Kermadec region and their bearing on the depth extent of mantle flow, J. Geophys. Res., 84, 6783-6795, 1979.

Rikitake, T., Earthquake Prediction, 357 pp., Elsevier, Amsterdam, 1976.

Samowitz, I. R., and D. W. Forsyth, Double seismic zone beneath the Mariana Island arc, J. Geophys. Res., 86, 7013-7021, 1981.

Shiono, K., T. Mikumo, and Y. Ishikawa, Tectonics of the Kyushu-Ryukyu arc as evidence from seismicity and focal mechanism of shallow to intermediate-depth earthquakes, J. Phys. Earth, 28, 17-43, 1980.

Sleep, N. H., Teleseismic P-wave transmission through slabs, Bull. Seismol. Soc. Am., 63, 1349-1373, 1973.

Sleep, N. H., The double seismic zone in downgoing slabs and the viscosity of the mesosphere, J. Geophys. Res., 84, 4565-4571, 1979.

Smith, S. W., Determination of maximum earthquake magnitude, Geophys. Res. Lett., 3, 351-354, 1976.

Stauder, W., and L. Mualchin, Fault motion in the larger earthquakes of the Kurile-Kamchatka arc and of the Kurile-Hokkaido corner, J. Geophys. Res., 81, 297-308, 1976.

Stefani, J. P., R. J. Geller, and G. C. Kroeger, A direct measurement of the distance between hypocenter in a Benioff-Wadati Zone and the slab-asthenosphere contact, J. Geophys. Res., 87, 323-328, 1982.

Suzuki, S., T. Sasatani, and Y. Motoya, Double seismic zone beneath the middle of Hokkaido, Japan, in the southwestern side of the Kurile arc, Tectonophysics, 96, 59-76, 1983.

Sykes, L. R., Seismicity and deep structure of island arcs, J. Geophys. Res., 71, 2981-3006, 1966.

Topper, R. E., Fine structure of the Benioff zone beneath the central Aleutian arc, M.S. thesis, 148 pp., Univ. of Colo., Boulder, 1978.

Tsuboi, C., Investigation on the deformation of earth's crust found by precise geodetic means, Jpn. J. Astron. Geophys., 10, 93-248, 1933.

Tsukahara, H., Double-planed structure of the deep seismic zone and plasticity for the descending lithosphere (in Japanese), Annu. Meet. Seismol. Soc. Jpn. Abstr., 2, 149, 1977.

Tsukahara, H., Physical conditions for double seismic planes of the deep seismic zone, J. Phys. Earth, 28, 1-15, 1980.

Tsumura, K., Microearthquake activity in the Kanto district, in Special Publication for the 50th Anniversary of the Great Kanto Earthquake, pp. 67-87, Earthquake Research Institute, Tokyo, 1973.

Umino, N., and A. Hasegawa, On the two-layered structure of deep seismic plane in northeastern Japan arc (in Japanese), J. Seismol. Soc. Jpn., 28, 125-139, 1975.

Umino, N., A. Hasegawa, A. Takagi, S. Suzuki, Y. Motoya, S. Kameya, K. Tanaka, and Y. Sawada, Focal mechanisms of intermediate-depth earthquakes beneath Hokkaido and northern Honshu, Japan, Zisin, 37, 523-538, 1984.

Veith, K., The relationship of island arc seismicity to plate tectonics, Ph.D. thesis, South. Methodist Univ., Dallas, Tex., 1974.

Wiens, D. A., and S. Stein, Age dependence of oceanic intraplate seismicity and implications for lithospheric evolution, J. Geophys. Res., 88, 6455-6468, 1983.

Wortel, R., Seismicity and rheology of subducting slabs, Nature, 296, 553-556, 1982.

Yokokura, T. Viscosity of the earth's mantle: Inference from dynamic support by flow stress, Tectonophysics, 77, 35-62, 1981.

Yoshii, T., Structure of the earth's crust and mantle in northeast Japan, Kagaku, 47, 170-176, 1977.

Yoshii, T., A detailed cross-section of the deep seismic zone beneath Japan, Tectonophysics, 55, 349-360, 1979.

H. Kawakatsu, Seismological Laboratory 252-21, California Institute of Technology, Pasadena, CA 91125.

(Received June 24, 1985;
revised November 26, 1985;
accepted December 9, 1985.)