DOUBLE-PLANED STRUCTURE OF INTERMEDIATE-DEPTH SEISMIC ZONE AND THERMAL STRESS IN THE DESCENDING PLATE

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(Received December 20, 1983)

Recent seismic studies using a high-gain seismograph network have demonstrated the existence of a double-planed seismic zone in the descending plate beneath island arcs such as northeastern Japan, Kurile, and Central Aleutian. Several hypotheses in terms of plate unbending, phase changes, mechanical models have been proposed to explain the characteristic features of the double-planed structure. This paper presents a new hypothesis that thermal stress due to non-uniform temperature distribution in the descending plate is the main causative force for genesis of earthquakes in the double-planed seismic zone.

Based on the estimated temperature distribution in the plate with a dip angle \( \theta \) and a convergence velocity \( V_c \), the thermal stress is calculated analytically under several assumptions. According to the results of these calculations, the upper and lower parts of the plate are characterized by compressional stress, and the central part by tensional stress. This stress pattern is well consistent with the focal mechanism solutions of earthquakes in the two planes of seismic zone. To verify out hypothesis quantitatively, a new parameter \( R \), defined as the ratio of deviatoric stress to the mean normal stress at a depth, is introduced as an index of the possibility of earthquake occurrence. In the case of the descending plate, for example, beneath northeastern Japan (\( \theta = 30^\circ \), \( V_c = 8 \) cm/yr), two regions with \( R \geq 0.04 \) exist at the uppermost and central parts of the plate. These regions are parallel to each other with a distance of about 30 km. The upper and central regions are characterized by compressional and tensional deviatoric stress, respectively. These regions terminate at a depth of about 250 km. The above features explain the observed seismic activity under the northeastern Japan arc. This value of \( R = 0.04 \) is not too different from the data of rock fracture experiments at high temperature and pressure. The value of \( R \) at the center of the plate is the largest in the case of \( \theta = 20-30^\circ \) and \( V_c \geq 3 \) cm/yr and decreases with increasing dip angle.

1. Introduction

Owing to the dense distribution of local seismographic stations and developments in techniques of observation and analysis, the high quality of data provides us with a detailed understanding of the nature of intermediate depth earthquakes.
as well as shallow ones. UMINO and HASEGAWA (1975), TAKAGI et al. (1977), and HASEGAWA et al. (1978a, b) showed that the so-called Wadati-Benioff zone beneath the northeastern Japan arc is composed of double seismic planes. The double-planed zones in the intermediate depth are characterized by the following features: (1) The double-planed seismic zone exists in the depth range from 50 to 200 km. (2) The two planes are almost parallel to each other and the separation between them is 30 to 40 km. (3) Composite focal mechanism solutions for the earthquakes in the upper plane are of down-dip compression, whereas those in the lower plane are of down-dip extension. (4) The upper plane is located very close to the upper surface of the descending slab. (5) Seismicity in the upper zone is higher than that in the lower zone.

Recently similar double planed structure of the intermediate-depth seismic zone has been recognized in the Hokkaido and Kanto districts of Japan (TSUMURA, 1973; SUZUKI et al., 1983), Kurile (VEITH, 1974), and Aleutian (ENGDAHL and SCHOLZ, 1977), although they are not as definitively clear as in northeastern Japan. Therefore, the double-planed structure is taken to be a general feature in some subduction zones.

Several ideas to explain the double-planed seismic zone, as well as the characteristics of its source mechanism, have been proposed so far. VEITH (1974) proposed a qualitative interpretation by phase change from olivine to spinel. ENGDAHL and SCHOLZ (1977), ISACKS and BARAZANGI (1977), LINDE and SACKS (1978, 1979), and TSUKAHARA (1980) stated that the double-planed zone reflects the stress due to bending and unbending effects in the descending slab. The stress due to the plate sagging into surrounding asthenosphere was discussed by SLEEP (1979) as a cause. The present writers proposed thermal stress by non-uniform temperature distribution within the plate as the main cause of observed double-planed seismicity (HAMAGUCHI and GOTO, 1978; SEGAWA et al., 1982; GOTO and HAMAGUCHI, 1983). Although the comparison and combination of these proposed ideas will be made in detail in another paper (GOTO et al., 1983), some cause among them produces only negligibly small stress compared to others, and the validity of some other causes is essentially dependent on assumed models and values of parameters. Therefore, in the opinion of the present authors, the thermal stress is of more importance than others in many aspects at least for intermediate-depth earthquakes.

The aim of this paper is to elucidate how the thermally induced stress can explain the observed features of double-planed seismic zone taking the intermediate depth events in northeastern Japan as a typical case. Only the analytical solution of the problem based on simple assumptions will be described here for the sake of simplicity, and the examination on effects of these simple assumptions will be made in another paper (GOTO et al., 1983) based on calculations in a more realistic case.

A new parameter representing the ratio of deviatomic stress to confining pressure is introduced to connect the stress state with seismic activity, because the inter-
mediate-depth earthquakes occur under high pressure and temperature, where the criterion of brittle fracture should be controlled mostly by deviatoric stress. The distribution of the parameter is compared with observed seismological features mainly for a typical case of northeastern Japan.

Finally, the validity of the idea of thermal stress as the cause of brittle fracture is discussed based on icequake observations and acoustic emission experiments.

2. Temperature Distribution in the Descending Plate

Since the cold oceanic lithosphere thrusting under the continental plate is surrounded by hotter mantle, the plate is heated at both upper and lower boundaries and a non-uniform temperature distribution is generated within the plate and it results in thermal stress.

Figure 1 illustrates the orthogonal coordinates $0$-xyz adopted in the calculation. The $x$ axis is taken parallel to the dip direction of the descending plate and the $y$ axis coincides with the strike of the oceanic trench which is indicated by a reverse triangle in the figure. The $z$ axis is in the direction perpendicular to the surface of plate. The position of the origin, $0$, is fixed on the upper surface of a plate. For simplicity, the following description of analytical solution is limited to the portion where the plate is considered to be a rectangular prism, and the result is not suitable for interpreting the observation at a shallower depth than about 50 km.

The general equation of thermal conduction for a moving material is expressed as

\[ \frac{\partial T}{\partial t} = \alpha \nabla^2 T - \mathbf{v} \cdot \nabla T + S. \]
Fig. 2. Temperature distributions in the plate with (θ = 30°, V_c = 3 cm/yr), (θ = 30°, V_c = 8 cm/yr), and (θ = 60°, V_c = 8 cm/yr).

θ = 30°
V_c = 3 cm/yr

θ = 30°
V_c = 8 cm/yr

θ = 60°
V_c = 8 cm/yr
Fig. 3. The distribution of thermal stress $\sigma_x$ in the plate with ($\theta=30^\circ$, $V_\theta=3$ cm/yr), ($\theta=30^\circ$, $V_\theta=8$ cm/yr), and ($\theta=60^\circ$, $V_\theta=8$ cm/yr). Equal contours of thick line indicate the compressional stress field and those of broken line indicate the tensional stress field. The unit of stress is in kilobars. The parameters used in the calculation are given in the text.
where \( T \) is temperature, \( t \) time, \( \rho \) density, \( C_p \) specific heat at constant pressure, \( \kappa \) thermal conductivity, \( V \) velocity of material, and \( H \) internal heat production per unit volume (McKenzie, 1969). In solving Eq. (1), we introduce the following assumptions: (1) Temperature is independent of \( y \), i.e., \( \partial T/\partial y = 0 \) and \( \partial^2 T/\partial y^2 = 0 \). (2) The velocity \( V \) has only \( x \)-component of a constant value of \( V_c \). (3) The thickness \( a \) and the dip angle \( \beta \) of the plate are unchanged with respect to depth. (4) The coefficients \( \rho \), \( C_p \), and \( \kappa \) are constants throughout the plate. (5) Internal heat production including shear heating at plate surfaces is neglected. (6) The plate has enough length in \( x \) direction. (7) The calculation of thermal regime is reduced to a steady state. These simple assumptions are convenient to examine the dependency of the solution on adopted parameters. Then, Eq. (1) reduces to

\[
\rho C_p \rho \frac{\partial T}{\partial x} = \kappa \left( \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right). \tag{2}
\]

Now we introduce two functions \( F(x) \) and \( G(x) \), which express the surface temperatures along the upper and lower rims of a descending plate, respectively, and are determined by the adopted geotherm. Under the following boundary conditions,

\[
\begin{align*}
T &= F(0) \quad \text{at } x \leq 0, \ z = 0 \\
T &= F(x) \quad \text{at } x > 0, \ z = 0 \\
T &= G(0) \quad \text{at } x \leq 0, \ z = a \\
T &= G(x) \quad \text{at } x > 0, \ z = a \\
T &= F(0) + \frac{(G(0) - F(0))}{a} z \quad \text{at } x = 0, \forall z,
\end{align*}
\tag{3}
\]

the solution of Eq. (2) at \( x = x_j \) is expressed by the superposition of the basic solution given by McKenzie (1969, Eq. 14) as

\[
T = F(0) + \frac{(G(0) - F(0))}{a} \frac{z}{a} + \sum_{n=1}^{\infty} \frac{1}{n \pi} \left[ A - (A^2 + n^2 \pi^2)^{1/2} \right] \frac{x - x_i}{a} \sin \left( \frac{n \pi}{a} z \right) \] 

\[
+ \sum_{i=1}^{\infty} \left[ f(x_{i+}) - f(x_{i-}) \right] \frac{a - z}{2a} + \frac{1}{n \pi} \left[ A - (A^2 + n^2 \pi^2)^{1/2} \right] \frac{x - x_i}{a} \sin \left( \frac{n \pi}{a} (a - z) \right) \right]
\tag{4}
\]

where \( A \) is the thermal Reynolds number and is given by
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\[ A = \frac{\rho C_p V_c a}{2x} , \quad (5) \]

and

\[ f(x) = F(x) - F(0) \]
\[ g(x) = G(x) - G(0) . \quad (6) \]

This gives the analytical solution for a temperature distribution within a plate.

The following constants are adopted throughout the calculation; \( \rho = 3.00 \) g/cm, \( C_p = 0.25 \) cal/g·deg, \( \kappa = 0.01 \) cal/cm·deg·sec, and \( a = 80 \) km. The temperatures at the upper and lower boundaries of the plate, denoted as \( F(x) \) and \( G(x) \) in Eq. (3), are given from the geothermal profile proposed by von Herzzen (1967). The summation in Eq. (4) is performed up to \( n = 20 \), which is sufficient to diminish the fluctuation due to small number of terms. Dip angle \( \delta \) and convergence velocity \( V_c \) essentially control the temperature field in a plate. If the plate descends slowly or with a low dip angle, it is reasonable to expect that the temperature distribution is more uniform compared to the case of faster descending or higher dip angle.

The temperature fields in the descending plate are estimated for all combinations of dip angles of 10, 20, 30, 45, 60, and 80°, and convergence velocities of 1, 2, 3, 4, 6, 8, and 10 cm/yr. The temperature distribution for \( \delta = 30° \) and \( V_c = 8 \) cm/yr, as well as \( \delta = 30° \), \( V_c = 3 \) cm/yr and \( \delta = 60° \), \( V_c = 8 \) cm/yr for comparison, is exclusively shown here in Fig. 2, because the comparison of computed thermal stress and observation, which will be described later, is made mainly in the case of northeastern Japan, where \( \delta = 30° \) and \( V_c = 8 \) cm/yr.

3. Thermal Stress Distribution in the Descending Plate

The thermal stress field in a plate can be estimated from the temperature distribution discussed above. The length (x direction) and thickness (z direction) of a plate are denoted by \( L \) and \( a \), respectively. The width in y direction and length \( L \) are taken to be large enough compared with \( a \). It is also assumed that the plate tip (\( x = L \)) sinks into the asthenosphere without any resistance, in other words, free boundary condition at the bottom tip of the plate.

The thermally induced stress is then expressed by the summation of stresses due to thermal expansion and that due to bending by a non-symmetric temperature distribution in z direction (Timoshenko and Goodier, 1951). Using the solution by Boley (1956), the calculated normal component of thermal stress in the direction of x axis, \( \sigma'_x \), is written as

\[
\sigma'_x = -\frac{\alpha ET}{1-\nu} + \frac{1}{1-\nu^2} \left\{ \frac{1}{a} \right\}^{a/2} \int_{-a/2}^{a/2} \alpha ET dz + \frac{\nu}{La} \int_{-a/2}^{a/2} dz \int_{-L/2}^{L/2} \alpha ET dx + \frac{1}{1-\nu^2} \left\{ \frac{12z-6a}{a^3} \right\}
\]
where \( \alpha \) is the linear thermal expansion coefficient, \( E \) the Young’s modulus, and \( \nu \) the Poisson’s ratio. \( H(x, z) \) is a function including second and higher order derivatives of temperature with respect to \( x \) as shown in the paper by Boley (1956).

In the same manner, the normal component of thermal stress in the direction of \( y \), \( \sigma_{y}^\prime \), is also obtained.

Numerical integration in Eq. (7) is performed on lattice points with an equal space interval of 5 km. The following values are adopted in the computation: \( \alpha = 1.1 \times 10^{-5} \, ^\circ\text{C}^{-1} \), \( E = 1.7 \times 10^{12} \, \text{dyn/cm}^2 \), \( \nu = 0.272 \), \( a = 80 \, \text{km} \), and \( L = 1,000 \, \text{km} \). Figure 3 shows the distribution of \( \sigma_{x}^\prime \) in the plate for the same values of parameters as in Fig. 2. The thermal stress field in Fig. 3 is characterized by tensional stress in the central part and compressional stress at upper and lower parts of the plate. In the case of northeastern Japan (\( \theta = 30^\circ \), \( V_c = 8 \, \text{cm/yr} \)), the maximum value of tensional stress is about 8 kbars at a depth of 350 km beneath the earth’s surface and that of the compressional stress is about 22 kbars at about 160 km depth. In other cases, for example, \( \theta = 30^\circ \), \( V_c = 3 \, \text{cm/yr} \) or \( \theta = 60^\circ \), \( V_c = 8 \, \text{cm/yr} \), the maximum stress is less than that in this particular case. It is also found that the compressional stress at the upper surface of the plate is always larger than that at the lower surface in all cases.

4. A Parameter \( R \) for the Criterion of Earthquake Occurrence

The thermal stress estimated in the previous section is not directly connected with the occurrence of earthquakes because of ambiguity concerning the critical strength of material in the intermediate depth where high pressure and temperature exist. Although the rock strength obtained for small size samples in the laboratory cannot directly be extrapolated to large bodies such as rocks in the plate, the experimental results give us a possible idea on fracture criterion.

A number of fracture experiments suggest the existence of limiting surface of fracture expressed by \( \sigma_1 = f(\sigma_2, \sigma_3) \) for a given material, where \( \sigma_1 \), \( \sigma_2 \), and \( \sigma_3 \) are the three principal stresses (e.g., Mogi, 1973). The conventional compression experiments \( \sigma_1 > \sigma_2 = \sigma_3 \) show that the compressive strength \( \sigma_1 - \sigma_3 \) of brittle rocks increases markedly with confining pressure \( \sigma_2 = \sigma_3 \). Based on various experiments, Robertson (1972) demonstrated the relation of \( \sigma_1 - \sigma_3 = 6\sigma_3 \) which provides a rough approximation for the strength of massive rocks in the crust. Mogi (1966) pointed out that the transition from brittle fracture to ductile flow for silicate rocks is expressed by a simple relation \( \sigma_1 - \sigma_3 = 3.4\sigma_3 \).
If the criterion of brittle fracture under the stress $\sigma_1 > \sigma_2 = \sigma_3$ can be formulated as

$$\sigma_1 - \sigma_3 = \beta \sigma_3,$$  \hspace{1cm} (8)

where $\beta$ is constant, the relation in triaxial stress state $\sigma_1 > \sigma_2 > \sigma_3$ may be written as

$$\sigma_1 - \bar{\sigma} = \gamma \bar{\sigma},$$  \hspace{1cm} (9)

where $\bar{\sigma} = (\sigma_1 + \sigma_2 + \sigma_3)/3$ and $\gamma = 2\beta/(3 + \beta)$. We may reasonably assume that Eq. (9), which is originally introduced in experiments at room temperature and low stress state, is generalized to the fracture criterion of rocks in high pressure and temperature, and this fracture criterion (Eq. (9)) is adopted as an indicator of the possibility of earthquake occurrence within a descending plate. The normal component in $z$ direction, $\sigma'_z$, may be negligibly small because of the assumptions that $a \ll L$ and there is no resistance at the bottom tip of the plate. Then, denoting the hydrostatic confining pressure as $P_h$, the resultant normal stresses $\sigma_x$, $\sigma_y$, and $\sigma_z$ inside the plate are given as

$$\sigma_x = \sigma'_x + P_h$$
$$\sigma_y = \sigma'_y + P_h$$
$$\sigma_z = P_h.$$  \hspace{1cm} (10)

The $x$-component of deviatic stress (down-dip component) is expressed by

$$\sigma_{xz} = \sigma_z - \bar{\sigma} = \frac{2\sigma'_x - \sigma'_y}{3}.$$  \hspace{1cm} (11)

Considering the criterion of rock fracture by Eq. (9), we introduce a new parameter,

$$R = \left| \frac{\sigma_{xz}}{\bar{\sigma}} \right|,$$  \hspace{1cm} (12)

as a measure of possibility of earthquake occurrence or shear faulting. In other words, earthquakes occur if the value of $R$ is greater than a threshold level.

The numerical value of the threshold can be roughly estimated from laboratory experiments, though results of various experiments are scattered to considerable extent. The experiments on compressional strength of limestones, gabbro, dunite, and serpentinite by Raleigh and Paterson (1965) under confining pressure less than 5 kbars and at room temperature indicate that the condition of occurrence of brittle fracture is roughly expressed by $R \geq 1.1$ at $P_h = 2$ kbars and $R \geq 1.4$ at $P_h = 5$ kbars. Mogi's compressional experiments (Mogi, 1966) for silicate and carbonate rocks at room temperature also showed the condition of $R \geq 1.1$ for $P_h \leq 5$ kbars. It has been demonstrated in these and other experiments (e.g., Raleigh and Paterson, 1965; Mogi, 1966; Byerlee, 1968; Robertson, 1972) that there exists a linear relation between differential stress and confining pressure larger than 2 kbars. When the experimental results are extrapolated to 30–40 kbars, which corresponds to the situation at the depth of intermediate-depth earthquakes, the threshold values of $R$ may be assumed to be 1–1.5 for room temperature.
The distribution of $R$ in the plate with $(\theta=30^\circ, V_c=3\text{ cm/yr})$, $(\theta=30^\circ, V_c=8\text{ cm/yr})$, and $(\theta=60^\circ, V_c=8\text{ cm/yr})$. Equal contours of thick line represent the compressional deviatoric stress and those of broken line represent the tensional deviatoric stress. The parameter $R$ is defined in the text.

Raleigh and Paterson (1965) examined the effect of temperature on ultimate strength of serpentinite and revealed that the strength dramatically decreases at
a critical temperature around 600°C. The strength beyond the critical value is less than one tenth of that for lower temperature. Since the temperature at the depth in concern is certainly above this critical value, the threshold of $R$ may be considerably less than 0.1, taking the effect of temperature into account.

GIARDINI (1974) showed the torsional shear strength of dunite, granodiorite, and other rocks for temperature range up to 900°C and pressure up to 50 kbars. His results indicated that the threshold of $R$ is almost constant within this range of temperature and that the values are approximately 0.13 and 0.10 for temperatures of 500 and 900°C, respectively. This gives the threshold value of $R$ around 0.1 for the confining pressure and temperature at 100–200 km in depth as a rough extrapolation.

Based on these considerations, it is not far from experimental results that the threshold of $R$ for intermediate-depth earthquakes is taken to be 0.1 or so.

**5. Distribution of $R$ in the Descending Plate**

The examples of distribution of $R$ within the plate are shown in Fig. 4, where the values of $\theta$ and $V_c$ are the same as those in Figs. 2 and 3. The solid and broken lines in the figure stand for the contours of $R$ for compressional and tensional deviatoric stresses, respectively. It can be clearly seen that both the upper and
lower surfaces of the plate are characterized by compressional field, and the central part by tensional stress. The variation of $R$ on the cross section perpendicular to the surface at 150 km depth is shown in Fig. 5 for combinations of $V_c=1, 4,$ and $8 \text{ cm/yr}$ and $\theta=10, 30,$ and $60^\circ$. The shaded and open areas indicate tensional and compressional fields, respectively. This figure demonstrates that the value of $R$ decreases with increasing dip angle and decreasing convergence velocity. It is also seen that the value of $R$ on the upper surface of plate is larger than that

Fig. 6. (a) Depth distribution of the iso-$R$ at the upper rim of the plate with the compressional deviatoric stress as functions of $V_c$ and $\theta$. The line with thick bar indicates $R=0.06$ and that with thin one $R=0.04$. (b) Depth distribution of the iso-$R$ at the center of the plate with the tensional deviatoric stress as functions of $V_c$ and $\theta$. The line with thick bar indicates $R=0.04$ and the thin one indicates $R=0.02$. Note that no iso-$R$ equal to 0.06 appears at the tensional stress field.
on the lower surface in most cases except for the case of $V_c=1$ cm/yr and $\theta=10^\circ$, where they are almost the same.

Figure 6 shows the variation of depth of the position corresponding to a fixed value of $R$ against $V_c$ and $\theta$, cases of $R=0.02-0.06$ being shown as examples. Figures 6(a) and 6(b) represent the variation of depth on the upper surface (compressional field) and in the central part (tensional field), respectively. As seen in Fig. 6, the variation with respect to dip angle $\theta$ is represented by an upward concave curve and it implies that there is a maximum depth at a contain value of $\theta$. Figure 6(a) indicates that the dip angle for this maximum depth decreases with increasing $V_c$. It is also evident that the absolute value of maximum depth increases with increasing $V_c$. For examples, it is about 200 and 440 km for $V_c=1$ and 10 cm/yr, respectively. Figure 6(b) shows a similar variation of the maximum depth in the tensional field in the central part of plate. It is noted that the values of $R$ in this figure are 0.02 and 0.04 instead of 0.04 and 0.06 in Fig. 6(a).

The comparison of Figs. 6(a) and 6(b) reveals some differences between compressional and tensional fields. First, there is no part corresponding to $R=0.06$ or larger in the tensional field for any combination of $V_c$ and $\theta$. Second, the region where $R$ exceeds a certain value appears for a larger range of $\theta$ in the compressional field than in the tensional one. For example, the region for $R \geq 0.04$ in the tensional field exists only when $\theta \leq 40^\circ$, whereas that in the compressional

Fig. 7. The maximum value of $R$ at the center of the plate with the tensional deviatoric stress as functions of $\theta$ and $V_c$. 

\begin{center}
\begin{tikzpicture}
\begin{axis}[
    title={R vs Dip Angle and Convergence Velocity},
    xlabel={Dip Angle (degree)},
    ylabel={R},
    xmin=0, xmax=90,
    ymin=0, ymax=0.1,
    xtick={0,20,40,60,80},
    ytick={0.01,0.02,0.03,0.04,0.05,0.06},
    xticklabels={0,20,40,60,80},
    yticklabels={0.01,0.02,0.03,0.04,0.05,0.06},
    grid=both,
    legend style={at={(0.5,0.95)},anchor=north},
]
\addplot3[surf,shader=interp,draw=black] coordinates {
    (0,0,0.01)
    (10,0,0.02)
    (20,0,0.03)
    (30,0,0.04)
    (40,0,0.05)
    (50,0,0.06)
    (60,0,0.07)
    (70,0,0.08)
    (80,0,0.09)
    (90,0,0.1)
};
\end{axis}
\end{tikzpicture}
\end{center}
field exists even for larger $\theta$ up to 90°. Third, the dependency of the shallowest
and the deepest depths of the region on $V_c$ is different in compressional and
tensional stresses. The shallowest depth for $R=0.04$ is kept almost constant, say
about 70 km, for $V_c$ from 1 to 10 cm/yr in the tensional field, and the deepest one
gradually increases with $V_c$ but the increment is much smaller than in the compres-
sional case. Actually it is about 170 and 270 km, respectively, for $V_c=1$ and
10 cm/yr, and is much smaller than that in the compressional mentioned previously.

Figure 7 gives a more intuitive representation about the change of maximum
value of $R$ in the tensional field with respect to $V_c$ and $\theta$ in a three dimensional
expression. The dependency of maximum $R$ in the compressional field on dip
angle and velocity has a nature similar to Fig. 7, although it is not presented here.
This figure reveals that the maximum value of $R$ in the central part of the plate
has a peak when the dip angle is about 20°, and the peak value does not vary much
for convergence velocity larger than 2 cm/yr. It implies that the possibility of
thermally induced occurrence of earthquake in the central part of a descending
plate strongly depends on dip angle but not so much on convergence velocity larger
than 2 cm/yr.

6. Correlation between the Distribution of $R$ and Seismicity

The northeastern Japan arc is one of the typical island arc systems and the
seismic activity there has been precisely investigated (HASEGAWA et al., 1978a, b;
TAKAGI et al., 1977). The characteristics of double-planed seismicity of inter-
mediate depth earthquakes, which were described in the first section of this paper,
are recognized most clearly among various subduction zones. The relation
between $R$ and earthquake occurrence, therefore, is discussed mainly for this region
as a typical case. In northeastern Japan, the plate subducts with a dip of 30°
and a convergence velocity of 8 cm/yr. The dipping of 30° has been established,
for example, by ray tracing of converted waves (HASEGAWA et al., 1978b). The
Pacific plate moves towards the Eurasian plate with a relative spreading velocity
of 8 cm/yr near the Japan trench according to LE PICHON et al. (1973), and this
value is taken as $V_c$.

The distribution of $R$ inside the descending plate with $\theta=30^\circ$ and $V_c=8$ cm/yr
is illustrated in Fig. 8. The distribution of hypocenters of intermediate-depth
earthquakes during the period from 1975 to 1980 determined by the Observation
Center for Earthquake Prediction, Tohoku University, is also seen in this figure
for reference. The comparison of seismicity and contour line of $R$ demonstrates
that the contour of $R=0.04$ generally coincides with the seismic zone of inter-
mediate-depth earthquakes. The value of $R=0.04$ is in accordance with the
consideration of criterion of brittle fracturing discussed previously.

Two regions of possible earthquake occurrence, where the value of $R$ exceeds
0.04, are indicated by hatched area in Fig. 8. Several features of the thermally
induced stress field are seen in this figure. The two hatched regions lie parallel
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Fig. 8. The distribution of $R$ in the plate with $\theta = 30^\circ$ and $V_c = 8$ cm/yr and the focal depth distribution (EW section) of intermediate-depth earthquakes beneath northeastern Japan (latitude between 39°N and 40°N) observed by the high gain seismograph network of Tohoku University in the period from 1975 to 1980. Equal contours of thick line represent the compressional deviatoric stress and those of broken line represent the tensional ones. The region with $R$ larger than 0.04 and temperature lower than 1,300°C are marked by vertical lines.

to each other, their dipping angle being almost the same as that of the plate. The seismic zone near the upper surface of the plate is characterized by compressional stress, while that in the central part has a tensional character. The directions of the principal axes in both compressional and tensional fields are parallel to the dipping direction of the plate. Judging from the value of $R$ and area size for $R \geq 0.04$, higher seismicity is expected near the upper surface than in the central part of the plate. All these qualities of thermal stress are in good agreement with observed characteristics of earthquakes.

There are, however, two discrepancies between seismicity and iso-$R$ contour in Fig. 8. One is that there is a region of $R = 0.04$ near the lower surface of the plate, though the area size is very small. But earthquakes have not been observed in this region. This difference may be due to partial melting, or transition from an elastic to a plastic state, which is not taken into consideration in the present study. The experiments for natural peridotite at a high pressure (KUSHIRO et al., 1968) show that melting begins at a temperature of about 1,300°C under hydrous conditions. If this is applied to the materials of the descending plate, the lower surface of the plate begins to melt at a depth of about 110 km under the earth’s surface. Then earthquakes should not occur below this depth even when the value of $R$ exceeds the threshold. On the upper surface, the calculation indicates that the depth of melting temperature descends down to about 250 km and can thus explain the observed seismicity.

Another discrepancy is that the region for $R \geq 0.04$ extends down to $x = 625$ km (390 km depth) at the upper part, whereas the lowest end of intermediate-depth seismicity is located at a depth of about 250 km. This may be also attributed to the
effect of melting, and the occurrence of deeper events, which are observed in some other subduction zones, may be, to considerable extent, due to other causes such as phase change from olivine to spinel. The above explanation of differences can be proved valid by more detailed calculation of stresses by various causes, which is discussed in another paper (GOTO et al., 1983). It is concluded, therefore, that thermal stress originating from a temperature distribution in a descending plate can explain the characteristics of intermediate-depth seismicity in northeastern Japan.

Thermal stress model can interpret other situations in earthquake occurrence. Recently, FUJITA and KANAMORI (1981) surveyed intermediate-depth earthquake mechanisms from arc to trench regions and found that the double-planed seismic zones are not a feature common to all subduction zones but exist only in old and fast slabs with a dip angle between 30° and 45°. The maximum value of \( R \) in the tensional field is given in Fig. 7 as a function of \( V_c \) and \( \theta \). This figure indicates that the maximum \( R \) beyond 0.04 appears either when \( \theta < 30° \) and \( V_c \geq 3 \text{ cm/yr} \) or \( \theta = 45° \) and \( V_c = 3 \) to 4 cm/yr. This result is in accordance with the conclusion of FUJITA and KANAMORI (1981). Figure 7 also explains why the double-planed seismic zone is most clearly found in northeastern Japan. Dip angle of 30° and convergence velocity of 8 cm/yr corresponding to the plate in northeastern Japan are the most likely condition for appearance of large maximum \( R \) in the central part as well as near the upper surface. The plate beneath Kurile arc has a velocity of 7.5 cm/yr (LE PICHON et al., 1973) and dip angle of 38–45° judging from the seismicity plot by ISACKS and BARAZANGI (1977). It gives a favorite condition for double-planed seismic zone, though the possibility is lower than in northeastern Japan. The other plates with a dip angle larger than 40°, for examples the Aleutian, Izu-Bonin, Mariana, Tonga, Kermadec, and New Zealand arcs, have less possibility of double-planed zone than does the plate beneath northeastern Japan. Exceptions are South American regions, where dip angles and velocities are not much different from northeastern Japan. Though the reason for this is not quite clear at the moment, the shape of plates in these regions have been reported to be complicated (HASEGAWA and SACKS, 1981), and the stress due to a cause like bending-unbending may play a considerable role together with thermal stress.

7. Discussion

The idea that the thermal stress is the main cause of earthquake generation in a descending plate came from the study on icequakes in a floating ice sheet at Lake Suwa, central Japan (HAMAGUCHI et al., 1977; HAMAGUCHI and GOTO, 1978; GOTO et al., 1980). The icequake activity was observed together with temperatures at various depths in the ice sheet. High icequake activity was found from 8 to 10 o'clock in the morning, when the ice plate was heated at the upper surface while the temperature at its bottom remained almost constant. Calcula-
tion based on the measured temperature distribution indicates that compressional stress larger than 8 bars and tensional stress exceeding 4 bars should appear at the top and center of the ice plate, respectively. The pair of compressional and tensional stresses in the ice plate is proved to have clear correlation with the occurrence of icequakes.

This idea was also confirmed by a model experiment using a sample of a pine-resin (SEGAWA et al., 1982). One surface of pine-resin plate is heated so that the acoustic emissions occur within the plate. The thermal stress distribution can be estimated by temperatures given at surfaces. Compressional and tensional fields exist near the heated surface and middle part of plate, respectively. The distribution of hypocentral depths of acoustic emissions was observed to be in good correlation with thermal stress.

In the cases of icequakes and acoustic emissions mentioned above, the magnitude of stress was adopted directly as the measure of occurrence of brittle fracture different from the parameter $R$ in the case of earthquakes in a descending plate. However, since the confining pressure is very low in these cases, the magnitude of stress itself is almost the same as that of $R$, which implies the ratio of deviatoric to confining stress.

8. Conclusion

We introduced a new hypothesis that the thermal stress due to a non-uniform temperature distribution in the descending plate is the causative stress for the occurrence of earthquakes in the double-planed seismic zone. The thermal stress in the descending plate is estimated for models with various convergence velocities and dip angles and is examined in connection with the seismicity in the descending plate with due regard to the changes of hydrostatic pressure with depth.

The most important results obtained in this study are that the thermal stress model can account not only for the upper, down-dip compressional and the lower, down-dip extensional stresses in the descending plate, but also for the spatial distribution of the upper and lower seismic planes. This model can also answer for the greater seismic activity at the upper seismic plane than that at the lower one and explains well the observed double-planed seismic plane.

Our model suggests that the tensional stress at the lower seismic plane is evidently dependent on a dip angle and a convergence velocity. It follows from Fig. 7 that the double-planed seismic zone is not always found in a descending plate beneath some island arcs. Descending plates with a dip angle larger than 40°, for examples the Izu-Bonin, Mariana, Tonga, Kermadec, and New Zealand arcs, are supposed to have less possibility of a double-planed seismic zone than that beneath the northeastern Japan arc with a dip angle of 30°.

The authors acknowledge Profs. A. Takagi and T. Hirasawa, and Dr. T. Masuda for their enlightening discussions, and Observation Center for Earthquake Prediction, Tohoku University for providing the earthquake data.
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