Three-dimensional numerical modeling of detachment of subducted lithosphere

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Abstract. Recent seismological studies suggest that slab detachment has occurred in the Mediterranean and the New Hebrides subduction zones. Subducted slabs in these regions are recognized to be torn at depths ranging from 100 to 300 km, presumably caused by the lateral migration of the tear along the strike of the slab. To investigate the physical mechanism of the slab detachment and in particular its migration, we constructed a viscoelastic three-dimensional finite element model and introduced a small initial tear from one side of the slab. We investigated spatio-temporal variations in the state of stress within the slab, as a function of tear length, rheology, and a variety of force distribution. Our results show that an area of high shear stress concentration of the order of several hundred megapascals forms near the tip of the tear inside the slab, which is probably sufficient to cause further lateral migration of the tear. The stress concentration increases with the length of the tear and lower viscosity values of the surrounding mantle and increases with downdip tension. From our modeling, we conclude that favorable conditions for slab detachment are characterized by a high interplate frictional force at a subduction zone and a low convergence rate, forming in-plate tensile stress at intermediate depths. Such a condition is indeed observed in the Dinarides/Hellenic and the New Hebrides subduction zone.

Introduction

Soon after the hypocentral depth distributions of intermediate depth and deep earthquakes started to be analyzed in terms of subduction zone geometry, it appeared that several subduction zones exhibit a clear gap in seismicity in the depth range of about 100 to 300 km. Most notable among these is the New Hebrides zone [see Isacks and Molnar, 1969, 1971]. Explanations for such seismicity gaps can be summarized as follows: either the conditions within the subducting slab, in particular stress level and/or rheology, preclude the generation of earthquakes [see Wortel and Vlaar, 1988] or the seismicity gap is an expression of a real gap in the slab structure.

Clearly, high-quality information on the structure of subducted lithosphere would allow an assessment of the validity of the latter explanation. Such information recently became available for some interesting regions of plate convergence, in the form of a detailed three-dimensional (3-D) P wave velocity model for the lithosphere and upper mantle in the Mediterranean region determined by seismic tomography [Spakman, 1988, 1990, 1991] and P wave velocity and Q structure of the upper mantle for the New Hebrides region [Chatelain et al., 1992]. For both regions the structural information shows strong evidence for discontinuities in the descending slab, which were proposed to be the result of detachment (separation and sinking) of the lower parts of the slab [Wortel and Spakman, 1992; Chatelain et al., 1992].

For the Hellenic (Aegean) subduction zone, the studies by Spakman et al. [1988, 1993] [see Spakman, 1990] provide strong evidence for the existence of an interruption of the high-velocity slab in parts of the subduction zone, indicating that slab detachment has occurred in the depth range of 100-250 km. The geometry of the gap led Wortel and Spakman [1992] to interpret the structure in terms of a slab detachment process migrating in the strike direction of the convergence zone. This postulated process is referred to as “lateral migration of slab detachment.” In addition, they drew attention to the geodynamic consequences of slab detachment, as did Chatelain et al. [1992] in their work on the New Hebrides region. In particular, lateral migration of the tear in the Mediterranean subduction zone would introduce along-strike (along-arc) variations in vertical motions and the state of stress. Those changes in the stress field and vertical motions have important implications for development of fault patterns, formation and evolution of sedimentary basins, tectonic transport of nappes, sedimentation patterns (including large-scale
depocenter shifts), and volcanism. Such variations in state of stress and vertical motions and the associated processes are observed in some areas, and slab detachment may be a promising mechanism to account for them [Wortel et al., 1993].

To evaluate the detachment process and, in particular, the hypothesized aspect of its lateral migration more quantitatively, an approach using an efficient numerical method is required. We constructed a 3-D finite element model, incorporating a realistic rheology of the upper mantle and several pertinent force distributions acting on the subducted slab. This enabled us to study spatio-temporal variations in the state of stress within a slab arising from the slab detachment process. The purpose of this paper is to numerically evaluate the stress tensor in the descending slab and to investigate the physical conditions which allow lateral migration of slab detachment to occur.

Evidence for Slab Detachment

The evidence for slab detachment in the Hellenic (or rather Dinarides/Hellenic) and the New Hebrides subduction zones is briefly outlined in Plate I and Figure 1, respectively. Plate 1 shows P wave velocity anomaly patterns beneath the Mediterranean region obtained from travel-time tomography by Spakman et al. [1993]. Slab detachment has been proposed for the Tyrrhenian subduction zone by several authors, but evidence for a continuous slab down to approximately 500 km was presented by Anderson and Jackson [1987]. In view of the higher resolution obtained in the tomographic results for the Hellenic zone [Spakman et al., 1993], we focus in the present study on the Hellenic subduction zone in the eastern Mediterranean.

In the Hellenic zone, at depths of 95 km and 195 km the tomography results show a continuous belt of high-velocity anomalies extending from the northwest (from the northern end of the Adriatic Sea) along the Dinarides to the southeast where a prominent anomaly is found below and just north of Crete. These high-velocity anomalies are considered to be the image of subducted lithosphere [Spakman et al., 1988; Spakman, 1990]. This interpretation is warranted in view of the correspondence between high velocity anomalies and intermediate-depth and deep earthquake zones in other parts of the Mediterranean (Tyrrhenian Sea, southern Spain) and by extensive numerical modeling of geological reconstructions of the region [de Jonge et al., 1993]. Around a depth of 145 km, however, the high-velocity anomaly is only found in the Aegean region. Thus below Crete the subducting slab appears to be continuous (down to at least 600 km, as shown by Spakman et al. [1993]), whereas to the northwest (underneath the Dinarides), the tomography results point to the presence of a gap in the slab. Wortel and Spakman [1992] hypothesize that this geometry came into existence by a tear, originally present in the northwestern part of the plate convergence zone, which expanded laterally toward the southeast. The slab detected by seismic tomography at depths greater than about 200 km is aseismic. As shown by Wortel et al. [1990], the combined absence of seismic activity and presence of a detectable high-velocity anomaly (at these depths) can be explained if the subduction zone with pertinent convergence rate and thermal structure of the lithosphere is taken into account.

Figures 1a and 1b show horizontal and vertical projections of seismic activity along the strike of the sub-
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(a) New Hebrides


(b) o

..c: 200

21s 18s 15s

168E 165E

500 km
distance

171E

Figure 1. Map of the New Hebrides region. (a) Horizontal projection of seismicity down to a depth of 400 km along the island arc. The three lines A-A', B-B', and C-C' are cross sections of seismicity perpendicular to the island arc shown in Figures 10a, 10b, and 10c, respectively. (b) Vertical projection of seismicity (the area between the two dashed lines in Figure 1a) along the island arc, from the Aneytum island to the Torres island.

ducted slab in the New Hebrides region, respectively. The minimum magnitude plotted on the Figure 1 is 3.0, and the data are taken from the International Seismological Centre (ISC) catalogue from 1964 to 1987. In this region, the oceanic plate is subducting along the New Hebrides trench, with a very steep dip angle, in an ENE direction underneath the Fiji plateau. Beneath the islands of Malekula and Efate, seismic activity within the subducted slab in the depth range of 50-200 km is very low. Below the Malekula island, the width of the seismic gap in depth is several tens of kilometers, and the width appears to increase along the strike of the slab towards the SSE direction. Beneath the Santo island, however, there is no seismic gap, and the slab appears to be continuous in the depth direction. Based on seismic data from a local network, a sharper delineation of the wedge-shaped seismic gap was obtained by Chatelain et al. [1992].

Seismic waves passing through the seismic gap are characterized by severe attenuation and low velocities [Grasso et al., 1983; Marthelot et al., 1985; Prevot et al., 1991]. From this evidence, Chatelain et al. [1992] suggested that the seismic gap corresponds to the absence of the subducted slab. Furthermore, we interpret the seismic gap as a tear within the slab which has migrated along the strike of the slab in the NNW direction, as was similarly suggested by Wortel and Spakman [1992] for the Hellenic subduction zone. In this case, the tip of the tear may have reached the region below the Santo island.
The 3-D Finite Element Model of the Slab Detachment

Geometry and Rheology of the Model

Figure 2 shows our 3-D finite element model from the Earth’s surface down to a depth of 700 km. We adopted isoparametric hexahedron elements with eight nodes [Zienkiewicz and Cheung, 1967]. The total number of the finite elements (thick lines) and unknown nodal displacements are 5616 and 19,950, respectively. For the solver, we used the wave front direct solution method [Irons, 1970] to reduce core memory usage on the computer.

Figure 3a represents the geometry of the subducted slab in the modeling space shown in Figure 2. Here, we consider a general slab to investigate the state of stress within the slab associated with the slab detachment. We assumed that a planar slab with a width of 230 km in the strike direction, a dip of 65°, and a uniform thickness of 100 km is subducting to a depth of 600 km. The thickness of the overriding continental lithosphere is assumed to be 90 km. To investigate the state of stress associated with slab detachment, we introduced a small initial tear with a uniform width of 10 km from one side of the slab at a depth of about 125 km. It should be noted, however, that throughout this paper we deal with neither the formation process of the initial tear nor realistic propagation of the tear along the strike of the slab.

We constructed a finer finite element mesh in the vicinity of the tip of the tear, where large stress concentration is expected to occur. The overriding continental lithosphere, the slab, and the surrounding upper mantle (all of which are assumed to be isotropic) are represented as an elastic body, a standard linear solid, and a Maxwell body, respectively. The assigned material constants, which are relevant to the upper mantle, are given in Table 1. These values are taken from Hager et al. [1983], Hashimoto [1984], and Ghose et al. [1990]. Since we consider a force balance pertaining to the descending slab only (which is presented below), we do not take into account the density contrast of the continental lithosphere and the horizontal part of the oceanic lithosphere (see Figure 3b) relative to the underlying mantle. Incidentally, if the density contrast is taken into account, downward displacements of an amount comparable to those of the downgoing slab are obtained for the continental lithosphere and the horizontal part of the oceanic lithosphere. This is not consistent with observations and an unreal situation. However, even if this is the case, the resultant state of stress is not so much different from that for the model in which the density contrast is not incorporated.

To solve the time-dependent viscoelastic problem, we used the Laplace transform, which enables us to solve it as an equivalent elastic problem for several values of the Laplace variables [Peltier, 1974]. In this paper, we follow the method by Hashimoto [1984], in which Smith’s [1974] method is extended. We also employed the approximate methods of transform inversion [Schapery, 1962] to obtain the solution in the time domain.

Force Distributions on Subducted Lithosphere

As the probable forces acting on the subducting lithosphere, gravitational negative buoyancy (slab pull), ridge push force, negative and positive buoyancy accompanying phase changes from olivine to spinel at 400 km depth and spinel to perovskite and magnesiowüstite at 670 km depth, interplate friction at the subduction zone, viscous drag from the surrounding mantle, and viscous resistance forces or hydrodynamic pressure acting on the leading edge of the slab have been proposed [e.g., Forsyth and Uyeda, 1975; Schubert et al., 1975; Chapple and Tullis, 1977; Davies, 1980; Sekiguchi, 1985; Harper, 1986; Goto et al., 1987; Wortel et al., 1991]. In these previous studies, the relative importance of the slab pull force ($F_{SP}$), the slab resistance force acting in the deeper portion of the slab ($F_{S}$), and interplate friction at the subduction zone ($F_{F}$) has been pointed out.

In the present study, therefore, we consider static force equilibrium in the downdip direction of the slab:

$$ F_{SP} = F_{R} + F_{F}, $$

namely,

$$ g \cdot \sin \varphi \int_{V} \Delta \rho dV = \int_{S_1} \tau_{Rd} dS_1 + \int_{S_2} \tau_{Pd} dS_2, $$

where $g$ is the acceleration of gravity, $\varphi$ is the dip angle of the slab measured down from the horizontal, $\Delta \rho$ is the density contrast between the slab and the surrounding mantle, $\tau_{R}$ is the resistive shear stress acting on the upper, lower, and side plate boundaries or the resistive stress in the updip direction (hereafter re-
Figure 3. (a) The configuration of the planar subducting lithosphere in the modeling space. A small initial tear is introduced from one side of the slab at a depth of about 125 km. Finer mesh is constructed near the tip of the tear. (b) Schematic cross section of force balance acting on the subducted lithosphere. The three thick solid lines and the thick dashed line represent cross section of areas where surface integral $\int_S dS_1$ and $\int_S dS_2$ are carried out, respectively. Density contrast relative to the underlying mantle is not considered for the shaded portions. (c) A projection of the slab with a 30-km horizontal length of the tear on to the vertical plane perpendicular to the $y$ direction. Dashed lines denote the finite element mesh. (d) A projection of the slab with a 80-km horizontal length of the tear on to the vertical plane perpendicular to the $y$ direction. (e) A projection of the slab with a 130-km horizontal length of the tear on to the vertical plane perpendicular to the $y$ direction.
ferred to as updip stress $\tau_R$) on the bottom surface of the slab, and $\tau_F$ is shear frictional stress on the plate boundary at the subduction zone (see Figure 3b). $\tau_F$ may also include the effect of compositional buoyancy of the oceanic crust [Oxburgh and Parmentier, 1977]. The magnitude of $\tau_F$ has been estimated to be of the order of ten to a hundred megapascals [e.g., Hanks, 1977; Davies, 1980; Honda, 1985; Sekiguchi, 1985; Vanden-Beukel, 1992]. Once $\tau_F$ is determined, we can calculate $\tau_R$ from equation (2). In this study, we assigned the same magnitude for the resistive shear and updip stresses $\tau_R$. The stresses $\tau_R$ and $\tau_F$ are applied directly to the slab-mantle and the slab-continental plate interfaces, respectively. It should be noted that unlike previous research based on two-dimensional models [e.g., Davies, 1980; Wortel and Vlaar, 1988], the volume integral $\int_V dV$ and the surface integral $\int_{S_1} dS_1$ are carried out over the entire (3-D) volume of the slab except the portion of the initial tear ($\int_{S_1} dS_1$ is taken below 90 km in depth). The surface integral $\int_{S_2} dS_2$ is calculated over the boundary between the overriding continental plate and the subducting plate at depths shallower than 90 km. In the model, the full forces $F_{SP}$, $F_R$, and $F_F$ are applied instantaneously at the time $t=0$ year and are kept constant subsequently in time.

**Boundary Conditions**

For boundary conditions, we assumed the Earth's surface to be a free surface. We attached infinite domain elements (thin lines in Figure 2) [Beer and Meek, 1981] to the rest of the five boundaries to reduce the

**Table 1. Properties for Material**

<table>
<thead>
<tr>
<th>Material</th>
<th>Rigidity, $\sigma$</th>
<th>Poisson's Ratio, $\nu$</th>
<th>Density, $\rho$</th>
<th>Viscosity, $\eta$</th>
<th>Relax Ratio, $\mu_2/\mu_1$</th>
<th>Material Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Continental lithosphere</td>
<td>$3.2 \times 10^{10}$</td>
<td>0.23</td>
<td>2.3</td>
<td>0</td>
<td>0.0</td>
<td>elastic</td>
</tr>
<tr>
<td>Descending slab</td>
<td>$7.1 \times 10^{10}$</td>
<td>0.26</td>
<td>66.7</td>
<td>$5.5 \times 10^{22}$</td>
<td>0.6</td>
<td>standard linear solid</td>
</tr>
<tr>
<td>Horizontal slab</td>
<td>$7.1 \times 10^{10}$</td>
<td>0.26</td>
<td>0</td>
<td>$5.5 \times 10^{22}$</td>
<td>0.6</td>
<td>standard linear solid</td>
</tr>
<tr>
<td>Surrounding mantle</td>
<td>$6.7 \times 10^{10}$</td>
<td>0.30</td>
<td>0</td>
<td>$1.0 \times 10^{20}$</td>
<td>0.0</td>
<td>Maxwell body</td>
</tr>
<tr>
<td>Infinite domain element</td>
<td>$6.9 \times 10^{10}$</td>
<td>0.28</td>
<td>0</td>
<td>0.0</td>
<td>0.0</td>
<td>elastic</td>
</tr>
</tbody>
</table>

Relax ratio is defined as $\mu_2/(\mu_1+\mu_2)$, where $\mu_1$ and $\mu_2$ are rigidity of springs 1 and 2, respectively, connected in parallel in the standard linear solid model. The spring 1 is directly connected with a dashpot in series. For the elastic and Maxwell rheology, spring 2 is not used.
Table 2. Stress and Force Distribution Acting on the Descending Slab

<table>
<thead>
<tr>
<th>Force</th>
<th>(\tau_R), (\tau_F), (F_R), (F_F)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>(\text{Pa} )</td>
</tr>
<tr>
<td>Force 1</td>
<td>(1.93 \times 10^7)</td>
</tr>
<tr>
<td>Force 2</td>
<td>(1.30 \times 10^7)</td>
</tr>
<tr>
<td>Force 3</td>
<td>(6.39 \times 10^6)</td>
</tr>
<tr>
<td>Force 4</td>
<td>0</td>
</tr>
</tbody>
</table>

\(\tau_R\), shear and up-dip resistance stress acting in the deeper part of the slab; \(\tau_F\), shear frictional stress on the plate boundary at the subduction zone; \(F_R\), slab resistance force acting in the deeper part of the slab; and \(F_F\), shear frictional force on the plate boundary at the subduction zone.

The grid size. We assumed a perfect elastic body for the material of infinite domain elements (Table 1). Since the material of the upper mantle behaves as a fluid after a long loading time has elapsed, the elastic material plays a role to hamper flow of the upper mantle material through the finite element-infinite domain element boundaries.

Figures 3c, 3d, and 3e represent the projection of the slab with three different horizontal lengths of the tear, namely, 30, 80, and 130 km along the strike of the slab on to the vertical plane perpendicular to the y direction. With these models, we mimic different stages in the horizontal propagation of the tear. The tear has a uniform thickness of 10 km in the z direction and cuts the slab entirely in the y direction at the time \(t=0\) year (Figure 3a). We assumed that the tear is filled with a Maxwell body with the same rheology as the surrounding upper mantle. The material of the upper mantle can flow into and out of the tear freely from +x and ±y directions after a long time in which the material behaves as a fluid. Therefore the upper and the lower horizontal and a vertical interface between the slab and the tear are simply defined as the material boundary and no forces are applied on these surfaces through the time concerned.

Model Specification

Several combinations of stresses and forces considered here are listed in Table 2; the force distributions are referred to as force 1 to force 4. Although the values of \(\tau_R\) and \(F_R\) vary with the change in \(F_{SP}\) in relation to the length, and hence to volume, of the initial tear, its effect is negligible because of the small volume and surface changes of the tear relative to the entire slab. Hence, for the determination of their values, we took the average of the cases differing in the length of the tear. In force 1, the resistive shear and updip forces \(F_R\) acting on the subducted plate are dominant (see Figure 3b). On the other hand, interplate frictional force \(F_F\) at the subduction zone is dominant in force 4. Therefore force 1 corresponds with fast plate convergence, whereas force 4 with extremely slow or no convergence. Forces 2 and 3 are intermediate between them in terms of the force distribution. Incidentally, if the values of \(F_R\) for forces 1 to 3 are divided by the width of the slab in the strike direction (230 km), we obtain the force per unit length.

Table 3. Model Description

<table>
<thead>
<tr>
<th>Model</th>
<th>Force</th>
<th>Length of the Tear, (\text{km})</th>
<th>loading Time, (\text{years})</th>
<th>Viscosity of the Mantle, (\text{Pa s})</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-1</td>
<td>1</td>
<td>30</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>1-2</td>
<td>1</td>
<td>80</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>1-3</td>
<td>1</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-1</td>
<td>2</td>
<td>30</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-2</td>
<td>2</td>
<td>80</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-3</td>
<td>3</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>3-1</td>
<td>3</td>
<td>30</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>3-2</td>
<td>3</td>
<td>80</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>3-3</td>
<td>3</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>4-1</td>
<td>4</td>
<td>30</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>4-2</td>
<td>4</td>
<td>80</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>4-3</td>
<td>4</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-4</td>
<td>2</td>
<td>130</td>
<td>(0)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-5</td>
<td>2</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-6</td>
<td>2</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-7</td>
<td>2</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{20})</td>
</tr>
<tr>
<td>2-8</td>
<td>2</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(10^{21})</td>
</tr>
<tr>
<td>2-9</td>
<td>2</td>
<td>130</td>
<td>(1 \times 10^6)</td>
<td>(2 \times 10^{20} \rightarrow 10^{24})</td>
</tr>
</tbody>
</table>
of the order of $10^{13}$ N/m, which is consistent with the
values estimated in previous two-dimensional research
[e.g., Davies, 1980; Wortel et al., 1991].

In the following sections, each model is specified by
the combination of two numerals, e.g., model 1-2, as is
shown in the first column of Table 3. The first numeral
(1 to 4) refers to the force distribution and the second
one (1 to 3) to the length of the tear. The values 4
to 9 for the second numeral denote different models in
loading time or viscosity of the mantle.

**Results**

**Time and Viscosity Dependence of Stress Concentration**

Before describing detailed results, we first examine
the effects of time and viscosity dependence on the state
of stress within the slab. The state of stress is speci-
fied by six independent components of the stress ten-
sor. Solving for the eigenvalues of the stress tensor,
the minimum ($\sigma_{\text{min}}$) and maximum ($\sigma_{\text{max}}$) principal
stresses, and the maximum shear stress ($\sigma_{\text{mss}}$) are ob-
tained. The maximum shear stress $\sigma_{\text{mss}}$ is defined as

\[
\sigma_{\text{mss}} = (\sigma_{\text{max}} - \sigma_{\text{min}})/2
\]

Although we do not incorporate any specific mech-
nism of fracture formulation and propagation, this pa-
rameter is considered to be of primary importance in
any possibly operating mechanisms in fracture.

Figures 4a and 4b represent the loading time and vis-
cosity dependence, respectively, of the maximum shear
stress in a vertical cross section parallel to the $y$-$z$ plane
(see Figure 3a) passing through the vicinity of the tip
of the initial tear (3 km away from the tip on the un-
detached slab side; hereafter referred to as the "cross-
section VIC TIP"). The length of the initial tear is 130
km for all of these models. The horizontal axis in Fig-
ures 4a and 4b is taken as the distance measured from
the Earth’s surface in the downdip direction along the
upper surface of the slab. The relatively high stresses
at distances shorter than 150 km are due to the applied
interplate frictional force $F_F$.

Figure 4a shows the loading time dependence for
models 2-4, 2-5, 2-7, and 2-6 of Table 3. For all these

![MSS (tear length 130 km)](image)

**Figure 4.** Time and viscosity dependence of the maximum shear stress in the vertical cross
section passing through the vicinity of the tip of the tear (3 km away from the tip on the side
of the undetached slab; the cross-section VIC TIP) for models 2-3 and 2-4 to 2-9 of Table 3. The
abscissa is a distance measured from the Earth’s surface in the downdip direction along the upper
surface of the slab. (a) Loading time dependence of the magnitude of the maximum shear stress
(MSS). The viscosity is assumed to be $10^{21}$ Pa s. (b) Viscosity dependence of the magnitude of
the maximum shear stress (MSS). The loading time is assumed to be 1 m.y. For model 2-9, a
depth-dependent viscosity for dislocation creep under cold (an old suboceanic upper mantle or
continental shield) and wet conditions with an activation volume of 15 cm$^3$/mol by Kato and
Wu [1993] is taken.
models, the viscosity of the material of the tear and the surrounding upper mantle is assumed to be $10^{20}$ Pa s uniformly. There is no stress concentration at the time $t = 0$ year. An area with high stress concentration is gradually formed over time near the tip of the tear, reaching 200 MPa in the maximum shear stress for model 2-6. This time dependence of the stress concentration is due to the process to reach the steady state of the model. Although the tear is filled with a Maxwell body, the dashpot of the Maxwell body does not move just after the application of the full forces $F_S$, $F_R$, and $F_F$ at the time $t = 0$ year. Therefore the slab recognizes the tear as an elastic body with values of rigidity and Poisson ratio which are a little different from those of the slab. This results in no stress concentration at the tip of the tear initially. Because of the stress relaxation of the Maxwell material of the tear and the surrounding mantle, the material behaves as a fluid after a long loading time has elapsed. Hence, over time, the slab will recognize the tear as a tear in which a fluid is filled, forming the gradual stress concentration at the tip of the tear.

In general, the Deborah number $D$, which indicates a degree of fluidity of the system, is defined as

$$D = \frac{t_{rel}}{t_{load}}$$

with

$$t_{rel} = \frac{\eta_m}{\mu_m}$$

where $t_{rel}$ is the relaxation time, $t_{load}$ is the loading time, and $\eta_m$ and $\mu_m$ are the viscosity and the rigidity of the mantle material, respectively. If $D$ is nearly 0, the system is regarded as a fluid. The larger the value of $D$, the more elastic the system. $D$ is close to 1 when the elastic and viscous effects are comparable. The Deborah number of the model 2-3 is approximately $5 \times 10^{-5}$; thus the mantle material surrounding the slab behaves like a fluid after 1 m.y. has elapsed. This model also appears to almost reach the steady state at that time (Figure 4a).

To examine the effect of viscosity, we constructed models with variable viscosities for the material in the tear and for the mantle material surrounding the subducted lithosphere (models 2-7, 2-3, 2-8, and 2-9 of Table 3). We considered the case with a loading time of 1 m.y. for all these models. Models 2-7, 2-3, and 2-8 show that higher viscosity plays a role to suppress the stress concentration (Figure 4b).

Recently, Karato and Wu [1993] calculated the depth variation of the effective viscosity for suboceanic upper mantle on the basis of dislocation and diffusion creep. The viscosity for the dislocation creep steeply decreases with depth and has a minimum of $2 \times 10^{15}$ Pa s at a depth of 150 to 200 km and gradually increases to $10^{21}$ Pa s at a depth of 400 km. As the stress concentration tends to become larger with lower viscosity of the surrounding mantle, the minimum viscosity may be an important factor to induce slab detachment at that depth range. To investigate the effect of depth-dependent viscosity on the stress concentration, we also calculated the state of stress within the slab based on their viscosity for dislocation creep under cold (an old suboceanic upper mantle or continental shield) and wet conditions with an activation volume of 15 cm$^3$/mol. Although the viscosity is obtained assuming nonlinear rheology, we simply use the viscosity values in a frame of a linear Maxwell body. For the depths deeper than 400 km, we extrapolated their result, reaching $10^{24}$ Pa s at a depth of 675 km. The model (model 2-9) also shows the stress concentration at the tip of the tear. From a comparison with the isoviscous models, the degree of stress concentration appears to depend on the viscosity near the tear. The viscosity for the diffusion creep, which is more favorable for cold and wet conditions, has slight depth dependence [Karato and Wu, 1993]. This is approximately comparable to the model 2-3. In summary, Figure 4b indicates that the stress concentration surely occurs for the range of the probable viscosity values of the upper mantle.

Therefore, throughout the following sections, we discuss the state of stress after 1 m.y. as a loading time, being independent of the length of the tear, assuming the homogeneous viscosity of $10^{20}$ Pa s as the tear and the mantle material. Also we examine the three different lengths of the tear, namely, 30, 80, and 130 km along the strike of the slab, mimicking different stages in the horizontal propagation of the tear. Hence it should be noted that the loading time has nothing to do with the real stress concentration or lateral propagation of the tear but represents the process in which each model with different length of the tear reaches the steady state in the viscoelastic system.

**Directions of Stresses Within the Slab**

**Directions of the maximum and minimum principal stress axes within the slab due to the slab detachment.** Before describing the calculated directions of stresses, we briefly describe how we express and display them. In this paper, we define the minimum and maximum principal stresses depending on the arithmetic magnitude, not absolute value, and take compressive and tensile stresses as negative and positive values, respectively. We project the two principal stress axes on a lower hemisphere of the Wulff net, together with two planes bisecting the angle between these two stress axes. This representation is similar to the fault plane solution of an earthquake with the $P$ (minimum principal) and $T$ (maximum principal) axes and the two nodal planes.

First, we consider the effects of the length of the initial tear on the directions of the principal stress axes. For the purpose of illustration, we choose examples of force 2 because the resulting state of stress in the slab closely corresponds with the depth dependence of fault plane solutions observed in several deep subduction zones: downdip tension dominant in a shallower portion, downdip compression dominant in the deeper portion [e.g., Isacks and Molnar, 1969, 1971; Zhou, 1990]. The directions of stresses within the subducting slab for models 2-1 and 2-3 of Table 3 are shown in Figures 5a
Figure 5. Directions of the principal stress axes within the slab in the vertical cross-section VIC TIP. Dotted area denotes the depth range of the initial tear. Small solid squares in each circle is the maximum principal stress axis plotted on the lower hemisphere of the Wulff net (projected on the vertical cross section). Two nodal planes calculated from the minimum and maximum principal stress axes are also shown. For the coordinate, refer to Figure 3a. (a) For model 2-1 of Table 3 and (b) For model 2-3 of Table 3.
lower surfaces of the slab toward the inside of it. If the applied shear stress \( \tau_R \) is large enough, one of the two nodal planes becomes nearly parallel to the downdip direction along the upper and the lower surfaces of the slab. However, it should be noted that their sense of shear is opposite in direction as a result of the applied shear stress \( \tau_R \) on both sides. Although the applied shear stress \( \tau_R \) is not so large for model 2-1, we can recognize such tendency in Figure 5a. The second effect is the body force due to negative buoyancy in the dipping direction of the slab. This would be dominant near the central part of the slab along the dipping direction, where the effects of applied stress \( \tau_R \) becomes weaker. The third effect is the magnitude of the resultant stress concentration near the tip of the tear. This stress decreases with increasing distance from the tear and also has an asymmetrical distribution at a particular depth.

In contrast to model 2-1, in model 2-3 (Figure 5b) the directions of the maximum principal stress axes are not parallel to the downdip direction except around the depth of the initial tear. In the upper portion of the slab below the depth of the tear, the maximum principal stress axes deviate from the downdip direction and have a component in the strike direction. In the lower portion, the minimum and maximum principal stress axes tend to orient parallel to the strike of the slab. As we shall see in a later section, compression of the minimum principal stresses is dominant within the deeper portion of the slab, because the magnitude is discernibly larger than the magnitude of the maximum principal stresses. This situation is schematically illustrated in Figure 6. It should be noted, however, that the deformation of the slab is of the order of several hundred meters, which means that reconstruction of the finite element mesh with the loading time is not required. Since the variation of the deformation along the strike of the slab is one order smaller than the deformation itself, we do not consider the lateral variation of the resistive force \( F_R \) in our model.

Figures 5a and 5b indicate that the stresses in the
3-D subducted slab with the long initial tear are redistributed more strongly than those in the slab with the short tear. Therefore the difference in directions of the principal stress axes between the two models is interpreted as the difference of the intraplate deformation of the detached slab in the downdip direction relative to that of the undetached slab, as illustrated in Figure 6. In a vertical cross section passing through the detached slab in both models 2-1 and 2-3, patterns similar to those in Figures 5a and 5b can be identified.

**Directions of down-dip stresses within the slab.**

For a more quantitative discussion, focusing on the state of stress in the downdip direction, we introduce the following parameter $\kappa$:

$$\kappa = (| \cos \theta_{\text{min}} | - | \cos \theta_{\text{max}} |) \cdot \sin \theta_{\text{int}}$$

which is modified from Zhou [1990]. Here, verticals indicate absolute value, and $\theta_{\text{min}}, \theta_{\text{max}}, \text{and } \theta_{\text{int}}$ are defined as angles between the downdip direction and the directions of the minimum, maximum, and intermediate principal stress axes, respectively. $\kappa$ is dimensionless, with a value in the interval of -1.0 to 1.0. $\kappa$ equals -1.0 and 1.0 when the maximum and minimum principal stress axes, respectively, coincide with the downdip direction. $\kappa$ equals 0 when the intermediate principal stress axis coincides with the downdip direction or when one of the two nodal planes is parallel to this direction. In summary, a larger positive value (up to 1.0) indicates increasing downdip compression within a downgoing slab, whereas a smaller negative value (up to -1.0) shows increasing downdip tension.

Figure 7 shows the values of $\kappa$ as a function of distance measured downdip from the Earth's surface along the central part of the downdip slab. These represent the results for models 1-1 to 4-3 of Table 3 in the cross-section VIC TIP. In each model, the values of $\kappa$ are approximately 0 at distances less than 100 km. This is because one of the two nodal planes is parallel to the downdip direction due to the applied interplate frictional force $F_F$.

Models of force 1, in which the slab resistance force $F_R$ is dominant, show downdip compression within the entire slab, independently of the length of the tear. However, $\kappa$ shows small positive or negative values for the distances from 200 to 500 km. This is because one of the two nodal planes is nearly parallel to the downdip direction due to the strong shear stress $\tau_R$ applied on the upper and the lower sides of the slab even along the central part of the downdip slab. Another nodal plane is nearly perpendicular to the downdip direction. These indicate that the intermediate axis is parallel to the direction of the strike of the slab. At distances greater than 500 km, because of the strong updip stress $\tau_R$ applied on the bottom of the slab, downdip compression becomes dominant, resulting in the increase of the values of $\kappa$.

In contrast to models of force 1, force 4, in which the interplate frictional force $F_F$ is dominant, shows downdip tension within the entire slab. The state of stress in the deepest portion of the slab gradually changes from downdip compression to downdip tension, corresponding to models of forces 1 to 4, with the changes being proportional to the magnitude of the applied interplate frictional force $F_F$. Considering mechanics of subducted lithosphere, Davies [1980] also showed quantitatively that downdip compression becomes dominant within the entire slab with increasing deep mantle resistance force, which is consistent with our results.

Observed state of stress along the downdip direction of the slab in the northwestern Pacific region by Zhou [1990] shows that a distance of transition from downdip tension to downdip compression is short (less than 150 km) and downdip compression with $\kappa \approx 1$ is dominant in the deeper portion of the slab for more than 300 km. However, such dominance of downdip compression in the deeper part of the slab is not recognized in our models, in which we assumed that the amount of the shear stress $\tau_R$ acting on the upper, lower, and side boundaries of the slab is the same as that of the updip stress $\tau_R$ acting on the bottom of the slab. Therefore the updip stress would be considerably larger than the shear stress for real subducting slabs. There is also a possibility that the body force due to the phase change at a depth of 400 km, which is not considered in our model, might contribute to dominance of downdip compression in the deeper part of the slab.

In our models, except for force 1, there exists a depth range in which downdip tension is dominant locally at
Figure 7. The values of $\kappa$ in the vertical cross-section VIC TIP (see the text) for models 1-1 to 4-3 of Table 3 as a function of distance measured from the Earth's surface in the down-dip direction along the central part of the slab.

the distance of around 140 km, which is the depth of the tip of the initial tear. The longer the tear, the weaker the down-dip tension is for forces 2 to 4. This is attributable to the stress redistribution within the slab accompanying the larger intraplate deformation of the detached slab. A short initial tear would not influence the 3-D state of stress within the slab seriously; hence a two-dimensional approximation in the vertical cross-section would hold. For the longer tear, however, the component of down-dip tension is weakened by the tensile and compressive stresses in the strike direction of the slab (Figure 6).

Magnitude of the Stresses Within the Slab

The state of stresses near the tip of the initial tear. To reveal more detailed mechanical aspects of slab detachment, a better understanding of the magnitude of stresses is essential, in addition to the determination of their directions. This is very important to evaluate the possibility of lateral migration of the slab detachment and understand its formation process. We show three-dimensional contours of the magnitude of the minimum ($\sigma_{\text{min}}$) and maximum ($\sigma_{\text{max}}$) principal stresses, the maximum shear stress, and $\kappa$ within the subducted lithosphere in Plates 2 and 3. The values of $\kappa$ are plotted only within the descending part of the slab because of its definition. These display the state of stress in the vertical cross-section VIC TIP for model 2-3 of Table 3. We took force 2 as the example because the resulting state of stress in the slab closely corresponds with the depth dependence of fault plane solutions observed in several deep subduction zones. The horizontal and oblique cross sections within the slab are taken along the center of the elements which form the top and upper surfaces of the slab, respectively.

In Plates 2a and 2b, an area with large tensile stresses
of the order of hundred megapascals is formed horizon-
tally both in the minimum and maximum principal
stresses in the vicinity of the tip of the tear, especially
near the upper surface of the subducted plate. The po-
tive value in the minimum principal stress indicates that
tensile stress is predominant there. This means that
owing to the tensile volumetric changes in the vicinity
of the tip of the tear, the trace of the stress tensor is
not zero there. Weak compressive stresses in the mini-
imum principal stresses (but larger than tensile stresses
in the maximum principal stresses) are found within the
deeper portion of the slab. This area of compression
extends in the updip direction with increasing length of
the tear. Taking into account the systematic directions
of the minimum principal stress axes within the slab
in Figure 5b, this is indicative of lateral compressive
stresses arising from the larger deformation of the de-
tached slab, as illustrated in Figure 6. The maximum
shear stresses also show a similar high stress concentra-
tion except for their magnitude (Plate 3a). The high
stress values extending upward along the upper surface
of the descending slab at depths shallower than 90 km
are mostly due to the frictional force $F_F$ applied on
the plate boundary. The values of $\kappa$ indicate that downdip
tension becomes weak with the increase of depth. Plate
3b also clearly shows that downdip tension is dominant
locally in the area of stress concentration. These are
comparable with the results in Figures 5b and 7.

In our model, we assumed that the rheology of the
slab is represented by standard linear solid. We can
also obtain nearly the same results for the case of the
elastic or the Maxwell body. However, one to two orders
higher viscosity must be given for the Maxwell body, in
which the stress tends to dissipate easily.

We assumed a perfect elastic body for the material of
infinite domain elements (Table 1). Since the material
of the upper mantle behaves as a fluid after a long load-
ing time has elapsed, the elastic material plays a role
to hamper flow of the upper mantle material through
the finite element-infinite domain element boundaries.
As a result, return flow associated with the motion of
the plate in the downdip direction occurs for the upper
mantle material. Incidentally, if we assign Maxwell ma-
terial for infinite domain elements, the flow of the upper
mantle material through the finite element-infinite do-
main element boundaries are easily done after the long
loading time has elapsed. This results in no return flow
for the upper mantle material. Even in this case, how-
ever, the pattern of the state of stress within the slab
is similar to that for the case which is assigned elastic
body for infinite domain elements, except for a little
greater magnitude of stress values. This is presumably
because tensile stress within the slab becomes larger
easily for the case which has Maxwell material for in-
finite domain elements.

We also checked the effects of the depth of the initial
tear. To investigate this, we introduced an initial tear
at a depth of 200 km instead of 125 km. Except for a
little smaller values of the stress concentration, nearly
the same pattern is formed in the vicinity of the tip of
the tear for this case. The smaller values are probably
due to the smaller gravity force $F_S$ due to the decrease
of the entire volume of the detached slab below the tear.

Lateral variation of the state of stresses within
the slab. Plates 4 and 5 represents the state of stresses
in a cross section passing through the central part of the
slab, which is taken parallel to the oblique upper surface
of the slab. The horizontal and vertical cross sections
within the slab are taken along the center of the ele-
ments which form the top and side surfaces of the slab,
respectively. The respective figures correspond to those
in Plates 2 and 3. In Plates 4a, 4b, and 5a, the areas
with high stress concentration are formed in the vicin-
ity of the tip of the tear. These patterns appear to
agree well with the values of $\kappa$ (Plate 5b); the area with
high stress concentration corresponds to downdip ten-
sion, and vice versa. Downdip compression is dominant
in the deeper part of the detached slab (Plate 5b). This
is probably due to the larger intraplate deformation of
the detached slab in the downdip direction (Figure 6).
Lateral variation of $\kappa$ at depths around the tear along
the strike of the slab is also remarkable; the state of
stress varies from the downdip tension to the downdip
compression (Plate 5b).

The maximum shear stresses near the tip of
the initial tear. In Figure 8 we show the maximum
shear stresses as a function of the horizontal distance $x$
(see Figures 3a, 3c, 3d, and 3e) along the upper surface
of the subducting slab. These are the stresses passing
through a depth of the central part of the tear for mod-
els 1-1 to 4-3 of Table 3. The locations of the tips are at
distances $x = 200, 150,$ and $100$ km for the tear of
the length of 30, 80, and 130 km, respectively. As the dom-
inant stress within the slab becomes downdip tension
from forces 1 to 4, and as the tear becomes longer, the
asymptotic curve just ahead of the tip inside the unde-
tached slab becomes steeper. In contrast to this, the
maximum shear stress drops approximately to 0 MPa
inside the tear, where the material is filled with the
Maxwell material.

The above results bear a close resemblance to the
stress concentration and release associated with propa-
gation of a dynamic growing crack, especially for mode
I crack (the tensile or opening mode in which the crack
wall displacements are normal to the crack), where the
stress has a square root singularity near the crack tip
and becomes 0 inside the crack [e.g., Aki and Richards,
1980; Scholz, 1990]. However, detailed comparison be-
tween the two models would be difficult because of the
difference in rheology, the discrete computational val-
ues, and the asymmetry of the stress distribution nor-
tal to the horizontal tear plane (see Plates 2, 3, 4, and
5).

In Figure 9 we represent the values of the maximum
shear stresses as a function of the distance measured
from the Earth's surface in the downdip direction along
the upper surface of the subducted plate. These are
Plate 2. The state of stress within the slab for model 2-3 of Table 3. The vertical cross section coincides with the VIC TIP section. The horizontal and oblique cross sections within the slab are taken along the center of the elements which form the top and upper surfaces of the slab, respectively. It should be noted that the stress value is calculated at the center of each element. (a) The magnitude of the minimum principal stress (the positive and negative values denote tensile and compressive stresses, respectively). (b) The magnitude of the maximum principal stress (the positive and negative values denote tensile and compressive stresses, respectively).

Plate 3. The state of stress within the slab for model 2-3 of Table 3. The vertical cross section coincides with the VIC TIP section. The horizontal and oblique cross sections within the slab are taken along the center of the elements which form the top and upper surfaces of the slab, respectively. (a) The magnitude of the maximum shear stress. (b) The values of \( \kappa \).
Plate 4. The state of stress within the slab for model 2-3 of Table 3. The oblique cross section is taken so as to pass through the central part of the slab along the strike direction. The horizontal and vertical cross sections within the slab are taken along the center of the elements which form the top and side surfaces of the slab, respectively. It should be noted that the stress value is calculated at the center of each element, resulting in the apparent thicker tear. (a) The magnitude of the minimum principal stress (the positive and negative values denote tensile and compressive stresses, respectively). (b) The magnitude of the maximum principal stress (the positive and negative values denote tensile and compressive stresses, respectively).

Plate 5. The state of stress within the slab for model 2-3 of Table 3. The oblique cross section is taken so as to pass through the central part of the slab along the strike direction. The horizontal and vertical cross sections within the slab are taken along the center of the elements which form the top and side surfaces of the slab, respectively. (a) The magnitude of the maximum shear stress. (b) The values of $\kappa$. 
Figure 8. The magnitude of the maximum shear stress (MSS) in the horizontal cross section passing through a depth of the central part of the tear for models 1-1 to 4-3 of Table 3. The abscissa is a distance in the horizontal direction \( x \) (see Figure 3a) along the upper surface of the subducting slab.

Discussion

Evidence on the State of Stress From Fault Plane Solutions

In a global study of subduction zones, Isacks and Molnar [1969, 1971] investigated downdip stresses in descending slabs, on the basis of fault plane solutions. From their analyses, they concluded that at intermediate depths downdip tensional stresses are predominant in zones characterized either by gaps in the seismicity as a function of depth or by an absence of deep earthquakes. Downdip compressional stresses are prevalent in every subduction zone which is seismically active at depth greater than about 300 km. They also showed that especially at depth ranging from 100 to 200 km, there are many earthquakes whose fault plane solutions are neither downdip tension nor downdip compression. Such earthquakes are often observed below the New Hebrides region. They suggested that these events might be attributable to unresolvable contortions of the downgoing slab or other sources of stress such as thermal gradients within the plates or local density variations.

However, a more detailed global study of fault plane solutions of intermediate depth earthquakes by Fujita and Kanamori [1981] clearly shows predominance of in-
Figure 9. The magnitude of the maximum shear stress (MSS) in the vertical cross-section VIC TIP for models 1-1 to 4-3 of Table 3. The abscissa is a distance measured from the Earth's surface in the downdip direction along the upper surface of the slab.

Plate tensile stress beneath the New Hebrides region, where the possibility of the slab detachment is suggested. Since the configuration of the slab in the observed region is smooth and straight along the strike of the slab, the possibility of contortion would be ruled out as the explanation of the in-plate tensile stress. Also Chung and Kanamori [1978] found a couple of unusual intermediate depth earthquakes whose fault plane solutions are strike-slip type below the east of the Santo island. They suggested that these are presumably related to the subduction of the D'Entrecasteaux fracture zone - aseismic ridge system. However, the region is too limited to explain general characteristics of the in-plate tensile stress along the arc. Although in-plate tension is not unusual for the state of stress of a lower plane of a double seismic zone [Fujita and Kanamori, 1981], accurate hypocentral determinations using local network by Pascal et al. [1978] show no such double zone within the subducted slab beneath the New Hebrides region.

Disregarding the effect of the angle of internal friction of observed fault plane solutions, we can compare our calculations with observations. The calculated state of stress within the slab arising from the slab detachment also shows the predominance of in-plate tension at intermediate depths (Figures 5a and 5b). In Figure 10, we show fault plane solutions in some cross sections beneath the New Hebrides region (Figure 1a). We can also recognize dominance of in-plate tension from the fault plane solutions of intermediate depth earthquakes. In contrast, for the shallower portion, thrust faulting presumably due to the interplate frictional force between the slab and the overriding plate appears to be dominant, which is also consistent with our calculations. These agreements between observations and calculations strongly support the hypothesis of slab detachment occurring beneath the New Hebrides region.
A Possible Lateral Migration of the Tear and Its Migration Velocity

Results from high-pressure experiments of rock deformation during extension [Kirby, 1980] limit the strength of nonsubducted oceanic lithosphere to 100-1000 MPa in terms of differential stresses $|\sigma_{\text{min}} - \sigma_{\text{max}}|$. These values were obtained assuming a temperature distribution according to the plate cooling model [Parsons and Sclater, 1977] and anhydrous olivine rheology with the strain rate of $10^{-14}$ s$^{-1}$. Based on the depth dependent rheological and thermomechanical subduction zone model by Wortel and Vlaar [1988], we can estimate the strength in tension (in terms of the differential stress) of the subducted slab at a depth of 100 to 200 km as 600-900 MPa. Noticing the relation of equation (3), Figure 9 also shows that the differential stresses for forces 2 to 4 are of the same order as those estimated in these previous studies. This suggests that the stresses obtained here are sufficient to reach the yield stress and to allow the tear to migrate laterally.

This implication is very important to know the formation process of slab detachment. In the New Hebrides region, Chatelain et al. [1992] considered two possibilities for the cause of seismic gap. First, they considered that the lack of continuity of the downgoing slab suggests that the gap formed within the downgoing slab after it was subducted. They also suggested that there had been a gap within the oceanic plate before it was subducted. Our results and the clearly defined wedge-shaped seismic gap along the strike of the slab suggest the possibility that the slab detached partially and the tear migrated laterally along the strike of the slab after the subduction of the oceanic plate. If this is the case, how fast could the lateral migration of the tear occur?

We can roughly estimate the average migration velocity $V_M$ (centimeters per year) by assuming

$$V_M \sim L_M/((D_B - D_A)/V_B)$$

(7)

where $L_M$ (kilometers) is the horizontal length of the tear along the strike of the slab, $D_A$ and $D_B$ (kilometers) are the depth of the slab just above and below the tear, respectively, at its end (opposite side of the tip of the tear), and $V_B$ (centimeters per year) is a falling velocity of the detached slab. $L_M$, $D_A$, and $D_B$ can be

Figure 10. Cross sections of seismic activity perpendicular to the island arc in the New Hebrides region, together with fault plane solutions, which are projected on the back hemisphere of the focal sphere (modified from Pascal et al. [1978]). The hypocenters are relocated from the data of the International Seismological Summary (ISS), the International Seismological Centre (ISC), and the Earthquake Data Reports (EDR) used for the preliminary determination of epicenters (PDE) for the period 1962-1973. These data are reread from the seismograms of the World-Wide Standard Seismograph Network (WWSSN). (a) A-A' in Figure 1a. (b) B-B' in Figure 1a. (c) C-C' in Figure 1a.
estimated from a shape of a seismic gap or the results of seismic tomography. We estimate the possible range for $V_B$ as follows. In order to make a tear within the downgoing slab, $V_B$ must be faster than a convergence rate of the region concerned. On the other hand, $V_B$ should be slower than the velocity obtained by solving Stokes's problem because the detached slab is still connected with the undetached slab along the strike of the slab.

Here, following the procedure by Chatelain et al. [1992], we estimate the velocity of a falling sphere through the viscous upper mantle. As is well known, velocity $V_S$ of a sinking rigid sphere through a viscous medium in Stokes's problem is given by

$$V_S = 2\Delta \rho r^2 g / 9\eta,$$

where $\Delta \rho$ is density contrast between the sphere and surrounding viscous material, $g$ is the acceleration of gravity, and $\eta$ is viscosity of the viscous material. Assuming that $\Delta \rho = 66.7$ kg/m$^3$, $r = 50 - 100$ km, $\eta = 10^{20}$ Pa s, we obtain $V_S = 11.5 - 45.8$ cm/yr. However, as demonstrated by Morgan [1965], if we take account of the effect that the Earth's surface should alter the flow, the sinking velocity $V_T$ is approximately expressed by

$$V_T = (1 - 3r/4D)V_S,$$

where $D$ is the depth of the center of the sphere below the surface. Substituting $D = 250 - 300$ km into equation (8), we obtain $V_T = 9.8 - 34.4$ cm/yr beneath the New Hebrides and the Dinarides part of the Hellenic subduction zone.

In the New Hebrides region, if the Fiji plateau is assumed to be part of the Pacific plate, the NUVEL-1 model of DeMets et al. [1990] predicts a convergence rate of about 8.5 cm/yr. However, if the Fiji plateau is considered to be a separate microplate, Ruff and Kanamori [1980] estimates a rate of about 2 cm/yr, assuming that the Fiji plateau is stationary in the absolute frame. However, tensile stress is dominant for the intermediate depth earthquakes and has a very steep dip. Based on these discussion, Fujita and Kanamori [1981] suggested that a slower convergence rate (2 cm/yr) is preferable in the New Hebrides region. Substituting $L_M = 300 - 400$ km, $D_A = 100$ km, $D_T = 250$ km (see Figure 1), and $V_B = 2 - 34.4$ cm/yr into equation (7), we obtain the average migration velocity $V_M$ as 4 - 91.7 cm/yr.

On the other hand, in the Mediterranean region, de Jonge and Wortel [1990] obtained average convergence velocity at 10 Ma based on paleogeographic reconstructions proposed by Dercourt et al. [1986]. Their results show a convergence rate in the Dinarides part of the Dinarides/Hellenic subduction zone of about 0.2 cm/yr which drastically increases southeastward toward the Hellenic arc. Using $L_M = 800$ km, $D_A = 130$ km, $D_T = 190$ km (see Plate 1 [Spakman et al., 1993]), and $V_B = 0.2 - 34.4$ cm/yr, the average migration velocity is obtained as 2.7 - 459 cm/yr.

Our results show the viscosity of the tear and the surrounding mantle plays an important role to control the degree of stress concentration (Figure 4b); lower viscosity corresponds with a higher stress concentration. Therefore we can conjecture that the higher viscosity causes lower migration velocity of the tear. The estimated slow migration velocities of the tears below the New Hebrides and the Dinarides in the Mediterranean region are probably controlled by the high-viscosity values of material of the tear and the upper mantle near the tip of the tear in the respective regions.

From a comparison between Figures 7 and 9, we find that the downdip tension plays an important role to form the stress concentration. Also the dominant stress within the slab gradually changes from downdip compression to downdip tension, corresponding to forces 1 to 4, with the decrease of the slab resistance force $F_R$ (Figure 7). This would imply that an extremely low convergence rate (force 4) plays an important role in causing a more effective stress concentration than a fast convergence rate (force 1). Fujita and Kanamori [1981] suggested that old and slow slabs are under in-plate tension for the intermediate-depth earthquakes since these slabs would sink at a rate faster than the rate of surface convergence. Therefore the slow convergence rates in the Dinarides part of the Dinarides/Hellenic subduction zone and presumably the New Hebrides regions would cause the tensile stress within the subducted slab, resulting in the formation of the slab detachment.

Conclusions

We have investigated the spatial and temporal state of stress induced by slab detachment using a three-dimensional finite element method. Significant results obtained from our modeling are as follows:

The state of stress in the subducted lithosphere depends on the distribution of forces exerted on the slab. In general, downdip tension is promoted by a concentration of resistive forces in the shallow parts of the subduction zone, e.g., shear stress along the plate contact between subducting plate and overriding plate $F_R$. For models with such resistive forces (forces 2, 3, and 4), stress concentrations, with a magnitude of several hundred megapascals in the maximum shear stress, are found near the tip of the tear in the subducted slab. The magnitude of the stress concentration becomes larger, with weaker slab resistance force $F_R$ and with longer initial tear length. The area with the high stress concentration is formed over time in the vicinity of the tip of the tear. The stress concentration also has a tendency to be suppressed by the higher viscosity of the mantle material. The level of stress concentration at the tip of the tear is probably sufficient to cause further propagation of the detachment process in the horizontal (along strike) direction. Our model shows that the state of stress associated with the slab detachment produces in-plate (not everywhere downdip) tensile stress at intermediate depths. This is indeed observed in the New Hebrides subduction zone, where slab detachment is suggested.
The modeling results indicate that lateral migration of slab detachment may be expected in subduction zones with a low convergence rate, where the slab resistance shear and up dip force \( F_R \) acting in the deeper part of the slab is weak. The Dinarides/Hellenic subduction zone and presumably the New Hebrides regions are indeed characterized by such convergence conditions. Roughly estimated velocities of the lateral migration of the tear are of the order of centimeters to meters per year.

For better understanding of the physical mechanisms of the slab detachment, further investigations considering the effects of geometrical and compositional parameters and other forces acting on the slab are necessary. Incorporation of temperature-dependent viscosity of the mantle material, and the nonlinear rheology such as power law creep, taking into account a high water content in the upper mantle beneath island arc regions [e.g., Ito et al., 1983] should also be considered. After introducing subcritical rupture criteria by physico-chemical mechanisms and constructing a more realistic time-dependent dynamic model of lateral migration of the tear, the velocity of the migration and its direction, and the dominant parameter to control them should be evaluated. However, the present study should give a direction and constraints for further investigation of the slab detachment process.

Acknowledgments. We are indebted to M. Hashimoto for use of an original computer program of FEM. W. Spakman is gratefully acknowledged for providing a graphical software and for permission to use a figure of his results. We are also grateful to K. Furlong, M. Wallace, and G. Nolet for their critical reviews. Discussions with G.F. Davies and J. Braun were helpful in writing our manuscript. We also thank to H.C. Natal, R. Govers, P.E. van Keken, M.J.N. Remkes, M.R. de Jonge, and K. Hirahara for their kind help with the calculation. This research was supported by the Netherlands Organization for Scientific Research (NWO) through the Pioneer project “Detailed structure and dynamics of the upper mantle” (PGS 76-144).

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(Received July 7, 1993; revised May 6, 1994; accepted May 13, 1994.)