Conditions for a crustal block to be sheared off from the subducted continental lithosphere: What is an essential factor to cause features associated with collision?

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1. Introduction

Crustal blocks are detached from the subducting lithosphere at various depths. I call such phenomena shearing-off and investigate the mechanical conditions for shearing-off to occur in the continental lithosphere. The buoyancy of the crustal block does not work selectively as a driving force until it has been sheared off and I suppose that shearing-off occurs when the shear traction at the thrust, whose maximum value is calculated from the strength along the thrust, overcomes the strength of the boundary faults of the block, and determine the depth of the leading edge of the subducted crustal block when shearing-off occurs. The temperatures of the slab and at its surface are calculated by the steady state formula of Molnar and England (1990) as a function of the surface heat flow, dip angle, and convergence velocity of the subducted continental plate, as the shear stress and temperature at the ductile part of the thrust zone are consistent with the flow law. The shear stress at the brittle part is a strong function of the pore fluid pressure ratio $\lambda$. I show that shearing-off of an upper crustal block is possible only for a small value of $\lambda$ such as $\lambda \leq 0.4$ when the surface heat flow value is $\leq 70 \text{ mW/m}^2$. I suggest that the essential factor to cause shallow offscraping of crustal blocks in collision zones, is not a buoyancy of the continental crust, but a low value of $\lambda$, which would result from a little amount of dehydration from the subducted continental plate.

Table 1. Phenomena Related to Shearing-Off From the Subducting Lithosphere

<table>
<thead>
<tr>
<th>Depth of Shearing-Off</th>
<th>Oceanic Lithosphere</th>
<th>Continental Lithosphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shallowest (d &lt; 10 km)</td>
<td>offscraping of sediments (rarely seamounts)</td>
<td>(</td>
</tr>
<tr>
<td>Shallow (d = 10~30 km)</td>
<td>ophiolite obduction, offscraping of aseismic ridges and plateaus</td>
<td></td>
</tr>
<tr>
<td>Intermediate (d = 30~60 km)</td>
<td>exhumation of high-pressure (HP) metamorphic rocks</td>
<td></td>
</tr>
<tr>
<td>Deep (d &gt; 60 km)</td>
<td>exhumation of ultrahigh-pressure (UHP) metamorphic rocks</td>
<td></td>
</tr>
</tbody>
</table>

Banno and Sakai, 1989; Marsuyama et al., 1996. Recent detailed studies of HP metamorphic rocks of the Sanbagawa belt, Japan, showed that part of the subducted oceanic plate was exhumed from the depths of ~100 km, close to those of the UHP rocks in collision zones [Aoya et al., 2003; Ota et al., 2004].

[5] The fate of subducted crustal and mantle rocks, therefore, seems to be diverse in both space and time. Many researchers have tried to solve the mechanism of such variation by relating the buoyancy of continents, continental fragments, plateaus, and seamounts to their subductability [Molnar and Gray, 1979; Nur and Ben-Avraham, 1982; Cloos, 1993]. They, however, do not seem successful enough so far in explaining, for example, why a thin sheet of continental crust is accreted in some places, but not in others, why UHP rocks are exhumed often in the initial stage of collision, or why ophiolites are obducted in rare cases. This would imply that subductability of such features would not be a simple matter of buoyancy.

[6] In this study, I call separation of the upper coherent block (sheet) of the subducting continental or oceanic lithosphere from the rest of it (due to the shear traction at the thrust, as will be shown later) shearing-off. I use here the terminology such as subduction and slab also for continental plates, because it is now known that continental plates can in some cases penetrate deep into the mantle. The purpose of this study is to address a question what is a controlling factor that determines the depths of shearing-off for a plate capped by a continental crust with enough thickness. Answering this question seems important to understand the geological processes associated with collision, where plates with thick continental crust are impinging beneath the overriding plate, causing phenomena related to shearing-off at various depths and times as stated above. It would also have implications for shearing-off in subducting oceanic plates as will be shown in the discussion section.

[7] van den Beukel [1992] first investigated the conditions of shearing-off (breakup of the subducting lithosphere in his term) for the continental plate based on his thermo-mechanical models of the subducting lithosphere. He assumed that shearing-off occurs when the driving forces overcome the resistive forces to break the boundary of the detached block (Figure 1), and I follow his scheme in this paper. He, however, included the buoyancy force operating to the crustal block in the driving forces. I show that the buoyancy force of the subducting crust does not work as a driving force for shearing-off, and do not include it in the driving forces.

[8] He also assumed that the pore fluid pressure ratio \( \lambda \), defined by \( P_w = \lambda \sigma_n \) where \( P_w \) is the pore pressure and \( \sigma_n \) is the normal stress at the thrust [Hubbert and Rubey, 1959], as high as 0.96, which was inferred to be typical for oceanic plate subduction [e.g., van den Beukel and Wortel, 1988; Lamb, 2006]. He used hydrostatic \( P_w \) in the block boundary inside the subducting slab. He obtained the results that shearing-off occurs when the leading edge of the detached block reaches at a depth of 30~50 km for the continental lithosphere with a high to moderate surface heat flow (>~60 mW/m\(^2\)) prior to subduction, and it does not occur with a low surface heat flow (<50 mW/m\(^2\)). With these results, however, it seems difficult to explain the diversity in depth of shearing-off as stated above in a straightforward way.

[9] In this study, I take \( \lambda \) ranging from 0 to 0.95. This wider range is because there is no a priori reason to assume values of \( \lambda \) similar to those at the thrust of the subducting oceanic plate and there has been evidence for little dehydration from the subducting continental or island arc crust [Seno and Yamasaki, 2003; Seno, 2007]. I show that, without the buoyancy force, the effect of \( \lambda \) is more important and shearing-off can occur only when \( \lambda \) is small, if the surface heat flow prior to subduction is moderate or low. I try to demonstrate that a controlling factor to cause features characteristic of collision, such as shallow offscraping of a continental crust and intense deformation in the hinterland, is not the buoyancy of the crust but small \( \lambda \). I also try to explain the observed diversity in shearing-off among collision zones, e.g., the difference seen between the Himalaya and the W. Alps, by the different evolution of \( \lambda \) through their histories of collision.

2. Conditions for Shearing-Off to Occur

[10] van den Beukel [1992] asserted that the buoyancy force acts on the subducting crust and serves as a driving force for shearing-off. However, on the basis of the Archimedes’ principle, the buoyancy force arises from the pressure of the fluid integrated over the surface of the object. If the crustal and mantle part of the lithosphere is not separated, the asthenospheric pressure does not act at the base of the crust as far as the lithosphere behaves as an elastic plate. The buoyancy force acts to the lithosphere as a whole, if it is embedded within the asthenosphere, by the difference in asthenospheric pressure at the top and the bottom, not selectively to the crustal block to be sheared off. Therefore prior to accomplishment of shearing-off, it is evident that the buoyancy force does not help to peel the crustal block off from the rest of the lithosphere. This shows that the shear stress, or its integrated value, on the top of the block (Figure 1) is the only driving force for shearing-off.

[11] I assume that shearing-off occurs when the shear traction at the thrust overcomes the forces necessary to break the faults (\( F_1, F_2, \) and \( F_3 \)) surrounding the block. Without the shear traction, all the continental lithosphere
may be subducted into a greater depth, if an enough amount of the slab pull is available [Molnar and Gray, 1979]. I calculate the depth of the leading edge of the block, when shearing-off occurs. Although the traction provides a driving force for shearing-off, it is often difficult to estimate an exact amount of this force, because it might vary temporarily due to occurrence of earthquakes at the thrust zone. In this paper, I therefore use the maximum value of the shear stress, i.e., strength $\tau_c$, and its integrated value, traction $T_c$, over thrust zone $F_c$.

$T_1$ is the shear traction to cause shear failure along $F_2$. On the other hand, fault planes $F_1$ and $F_3$ are broken by reverse faulting and normal faulting, respectively. Within the slab near the fault edge, $\sigma_1$ and $\sigma_3$, the principal stresses, are aligned in the dip and perpendicular directions of the slab, respectively [Matsumura, 1997, Figure 1]. Let $T_1$ and $T_3$ be the integrated values of $\sigma_1$--$\sigma_3$ needed to fracture $F_1$ and $F_3$, respectively, from the upper to the lower ends of these faults. Because the stress perpendicular to the slab is close to the lithostatic (Appendix A), these forces to fracture $F_1$ and $F_3$ are able to be compared with $T_c$ or $T_2$. Then, I assume that shearing-off occurs when

$$T_c > T_1 + T_2 + T_3.$$  \(1\)

### 3. Driving Force

For the brittle part of $F_c$, $\tau_c$ is given by the Byerlee’s friction law, as

$$\tau_c = \mu \sigma_u^* + \tau_0 = \mu (\sigma_n - P_w) + \tau_0 = \mu (1 - \lambda)\sigma_n + \tau_0.$$  \(2\)

where $\mu$ is the coefficient of static friction, $\sigma_u^* (= \sigma_n - P_w)$ is the effective stress, and $\tau_0$ is the cohesive stress. Compressive stresses are taken to be positive throughout this study. $P_w$ is normalized to $\sigma_n$ by $P_w = \lambda \sigma_n$. As shown in Appendix A, $\sigma_n$ is approximated by the lithostatic pressure. The thickness of the upper crust $h_u$ and that of the lower crust $h_c$ are both assumed to be 18 km, which are typical for the continental crust [Christensen and Mooney, 1995]. The densities of the crust and the mantle ($\rho_c$ and $\rho_m$) are assumed to be 2800 and 3300 Kg/m$^3$, respectively.

I try to see the effect of $P_w$ on the occurrence of shearing-off by changing the value of $\lambda$ from 0 to 0.95. The higher end of this range is the one favored for subduction zones [van den Beukel and Wortel, 1988; Tichelaar and Ruff, 1993; Peacock, 1996; Cattin et al., 1997; Lamb, 2006]. The lower end implies that no fluid is supplied to the pore space. The high value of $\lambda$ in subduction zones is probably caused by dehydration of subducting altered crust and serpentinized mantle. Therefore there is no reason to assume this high value in collision zones, where a small extent of dehydration of the subducted continental crust is expected [see Seno and Yamasaki, 2003; Seno, 2007]. $\mu$ and $\tau_0$ are 0.85 and 0 MPa, respectively, for $\sigma_n < 200$ MPa, and 0.6 and 50 MPa, respectively, for $\sigma_n > 200$ MPa [Byerlee, 1978].

For the ductile part of the thrust zone, the shear stress cannot be given in an a priori manner because the shear stress depends on temperature $T_t$ at the slab surface through the flow law, and on the other hand, the temperature depends on the shear stress through shear heating, as noted by Lamb [2006]. The ductile shear stress at the fault plane is represented using the power law rheology by

$$\tau_c = \left[\varepsilon / A'\right]^{1/n} \exp[\{E^* + v^*P\}/nRT],$$  \(3\)

where $\varepsilon$ is the shear strain rate, $A'$ is the constant converted from the constant $A$ of the triaxial compression experiment to that of the 2-d simple shear problem, $n$ is the number close to 3, $E^*$ is the activation energy, $v^*$ is the activation volume, $P$ is the lithostatic pressure, and $R$ is the gas constant. $A'$ is the same constant in the generalized constitutive equation and converted from $A$ as

$$A' = 3^{(n+1)/2} A/2$$  \(4\)

(see Appendix B).

To calculate the temperature at the surface and inside of the slab, I ignore convection in the mantle wedge, and
Table 2. Parameters Used in the Modeling

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Definition</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_c$</td>
<td>shear strength at the thrust</td>
<td></td>
</tr>
<tr>
<td>$T_u$</td>
<td>maximum shear traction integrated over the thrust</td>
<td></td>
</tr>
<tr>
<td>$T_1$</td>
<td>traction needed to fracture $F_1$</td>
<td></td>
</tr>
<tr>
<td>$T_2$</td>
<td>traction needed to slip along $F_2$</td>
<td></td>
</tr>
<tr>
<td>$T_3$</td>
<td>traction needed to fracture $F_3$</td>
<td></td>
</tr>
<tr>
<td>$T_{\text{all}}$</td>
<td>$T_1 + T_2 + T_3$</td>
<td></td>
</tr>
<tr>
<td>$P_w$</td>
<td>pore fluid pressure</td>
<td></td>
</tr>
<tr>
<td>$\lambda$</td>
<td>pore fluid pressure ratio</td>
<td></td>
</tr>
<tr>
<td>$\mu$</td>
<td>coefficient of friction</td>
<td></td>
</tr>
<tr>
<td>$\sigma_0$</td>
<td>cohesive stress</td>
<td></td>
</tr>
<tr>
<td>$\theta^*$</td>
<td>fracture angle that satisfies $\mu = -\cot 2\theta^*$</td>
<td></td>
</tr>
<tr>
<td>$T_s$</td>
<td>surface temperature of the slab</td>
<td></td>
</tr>
<tr>
<td>$T$</td>
<td>temperature within the slab</td>
<td></td>
</tr>
<tr>
<td>$Q_s$</td>
<td>surface heat flow of the continental plate</td>
<td></td>
</tr>
<tr>
<td>$a$</td>
<td>plate thickness</td>
<td>65–117 km</td>
</tr>
<tr>
<td>$V$</td>
<td>convergence velocity</td>
<td>1.5–6 cm/a</td>
</tr>
<tr>
<td>$\delta$</td>
<td>dip angle of the underthrusting slab</td>
<td>15–45°</td>
</tr>
<tr>
<td>$A_0$</td>
<td>radiogenic heat production at the surface</td>
<td></td>
</tr>
<tr>
<td>$D$</td>
<td>scaling depth for exponential decay of radiogenic heat production</td>
<td></td>
</tr>
<tr>
<td>$k$</td>
<td>thermal conductivity 2.7 W/mk</td>
<td></td>
</tr>
<tr>
<td>$\kappa$</td>
<td>thermal diffusivity $10^{-6}$ m$^2$/s</td>
<td></td>
</tr>
<tr>
<td>$h_{uc}$</td>
<td>thickness of the upper crust 18 km</td>
<td></td>
</tr>
<tr>
<td>$h_{lc}$</td>
<td>thickness of the lower crust 18 km</td>
<td></td>
</tr>
<tr>
<td>$\rho_c$</td>
<td>density of the crust 2800 Kg/m$^3$</td>
<td></td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>density of the mantle 3300 Kg/m$^3$</td>
<td></td>
</tr>
<tr>
<td>$\varepsilon$</td>
<td>strain rate $10^{-36}$s in the thrust</td>
<td></td>
</tr>
<tr>
<td>$\varepsilon_s$</td>
<td>strain rate $10^{-36}$s in the slab</td>
<td></td>
</tr>
</tbody>
</table>

use the expressions derived for the subducting continental plate by Molnar and England [1990]. Then, temperature $T_s$ at the slab surface is a function written as

$$T_s = T_s(Q_s, V, A_0, D, \delta, \tau_c),$$

where $Q_s$ is the surface heat flow of the plate prior to subduction, $V$ is the convergence velocity, $A_0$ is radiogenic heat production at the surface, $D$ is the scaling depth for the exponential decay of radiogenic heat production, and $\delta$ is the dip angle of the slab. The formula of (5) is a sum of the three terms due to advection, radioactive heating, and shear heating [Molnar and England, 1990] and their expressions are given in Appendix C. I solve (3) and (5) simultaneously to determine $T_0$ and $\tau_c$ at the slab surface. This is conducted as follows. Starting from small initial trial values of $T_0$ and $\tau_c$ satisfying (5), substituting $\tau_c$ into (3), $T_0$ is obtained. If $T_0$ is larger than $T_0$, the next values of $\tau_c = \tau_c + \Delta \tau$ and $T_0$ satisfying (5), where $\Delta \tau$ is a small fraction of $\tau_c$, are tried. This process is repeated until $T_0$ becomes smaller than $T_0$, where $n$ indicates the n-th iteration. The obtained value of $T_0$ has an accuracy smaller than a few to several degrees, which is enough for the present purpose.

[17] The temperature within the plate at the trench prior to subduction is determined by the surface heat flow, $Q_s$, that is varied in the range of 50–90 mW/m$^2$. $Q_s$ is composed of the mantle heat flow, $Q_{sm}$, and the heat flow due to radiogenic heat production in the plate, $Q_{sr}$. The former and the latter are assumed to be 60% and 40% of $Q_s$, respectively [Pollack and Chapman, 1977]. The temperature of the asthenosphere, $T_m$, is taken to be 1300°C. The plate thickness, $a$, is then determined from the heat conduction equation as

$$a = (kT_m - Q_{sr}D)/Q_{sm},$$

where $k$ is the thermal conductivity, which is assumed to be constant for the whole lithosphere. I use $k = 2.7$ W/mk, which is the average of the crust and the upper mantle values [Schatz and Simmons, 1972], and $D = 10$ km. The thickness of the plate then varies between 58 and 110 km. $V$ and $\delta$ are varied between 1.5 and 6 cm/a and between 15 and 45°, respectively. The symbols and values of the parameters used in this study are summarized in Table 2.

[18] Because the thrust zone is the plate interface, and the upper plate material in contact with the slab changes from the upper crust, through the lower crust, to the mantle. The material of the lower plate in contact with the upper plate is always the upper crust of the subducting continental lithosphere. Because the weaker material deforms mainly at the shear zone [Yuen et al., 1978], I regard that the ductile shear zone at the plate interface forms in the surface layer of the lower plate, whose rheology is represented by that of granite [Carter et al., 1981]. The rheological parameters used for the crust and mantle materials are listed in Table 3. The activation volume of $17 \times 10^{-6}$ m$^3$/mol of olivine [Kirby, 1983] is used for both the crust and the mantle, because that of the crust is not well known. The strain rate at the ductile shear zone of the thrust is taken to be $10^{-14}$s, which corresponds to a differential velocity of 5 cm/a across the shear zone having a width of 2 km. This width is a representative one observed for outcrops of shear zones [Nicolas et al., 1977; Ponce de Leon and Choukroune, 1979].

[19] Figure 2 shows examples of calculation of $\tau_c$ and $T_s$ along the thrust with $V = 3$ cm/a, $\delta = 30°$, and $Q_s = 70$ mW/m$^2$ for $\lambda = 0.95$ and 0. For $\lambda = 0.95$, $\tau_c$ is small and the brittle-ductile transition occurs at a downdip distance of ~83 km from the trench (~42 km depth), due to low shear heating. For $\lambda = 0$, $\tau_c$ is larger, and the brittle-ductile transition occurs at a downdip distance of ~16 km from the trench (~8 km depth) due to large shear heating. In this case, $T_s$ in the ductile shear zone does not rise quickly along with the depth, because the combination of (3) and (5) forms a feedback system that stabilizes $T_s$. For example, if $\tau_c$ becomes larger, $T_s$ becomes larger through (5) due to shear heating. This makes in turn $\tau_c$ smaller through (3), and then making $T_s$ smaller through (5). Therefore $\tau_c$ at the thrust zone falls gradually as shown in Figure 2b. This is important to provide an enough amount of $T_c$ to cause shear-ing off for small $\lambda$.

4. Resistive Forces

[20] From the Coulomb-Mohr fracture criterion, the strength for $F_1$ and $F_3$ in the brittle part is,

$$\sigma_1 - \sigma_3 = 2 \tan \theta_\theta + (\tan^2 \theta - 1) \sigma_3,$$

Table 3. Rheological Parameters for Crust and Mantle Materials

<table>
<thead>
<tr>
<th>Section</th>
<th>Material</th>
<th>A</th>
<th>E*</th>
<th>n</th>
<th>Ref</th>
</tr>
</thead>
<tbody>
<tr>
<td>U. crust</td>
<td>granite</td>
<td>1.26E-9</td>
<td>105.8E3</td>
<td>2.9</td>
<td>Carter et al. [1981]</td>
</tr>
<tr>
<td>L. crust</td>
<td>diabase</td>
<td>5.2E2</td>
<td>356E3</td>
<td>3.0</td>
<td>Caristan [1982]</td>
</tr>
<tr>
<td>Mantle</td>
<td>dunite</td>
<td>3.4E4</td>
<td>544E3</td>
<td>3.5</td>
<td>Zech [1983]</td>
</tr>
</tbody>
</table>

*A and E* are in the unit of (MPa)$^{-1}$s and J/mol, respectively. Activation volume of $17*10^{-6}$ m$^3$/mol [Kirby, 1983] is used for all materials.
where $\sigma_1$ and $\sigma_3$ are the principal stresses aligned in the downdip and perpendicular direction of the slab (see section 2), $\theta$ is the fracture angle that satisfies $\mu = -\cot \theta$, where $\mu$ is the internal friction. Equation (6) is obtained from the geometry of the Mohr circle and the friction line tangential to it as

$$(\sigma_1 - \sigma_3)/2 = [(\sigma_1 + \sigma_3)/2 + \tau_0 \tan(180^\circ - 2\theta)] \cos(180^\circ - 2\theta).$$

The values of $\mu$ and $\tau_0$ in section 3 are used here. Although these values are for static friction [Byerlee, 1978], they seem appropriate for fractures within the crust. $\theta$ determines the fault geometry, as it is the angle between the fault normal and the direction of the larger principal stress. Because different values of $\mu$ give different values of $\theta$, as $\theta = 60.5^\circ$ for $\mu = 0.6$ and $\theta = 65.2$ for $\mu = 0.85$, I use $\theta = 60.5^\circ$ for the fault geometry. For the calculation of the strength, appropriate values of $\theta$ and $\mu$ are used depending on $\sigma_x$. [21] For $F_1$ which is the reverse fault, using the result that $\sigma_3 = \sigma_x$ in Appendix A, equation (6) gives

$$\sigma_1 - \sigma_3 = 2 \tan \theta \tau_0 + (\tan^2 \theta - 1) \sigma_x$$

Similarly, for $F_3$ which is the normal fault, using the result that $\sigma_1 = \sigma_x$ in Appendix A, equation (6) gives

$$\sigma_2 - \sigma_3 = 2 \tau_0 / \tan \theta + (1 - 1/ \tan^2 \theta) \sigma_x.$$  

For faults $F_1$ and $F_3$, I assume $P_w = 0 (\lambda = 0)$. This is because dehydration by compaction of the cracks in the granitic crust would be finished at shallow depths and the breakdown of white mica and biotite occurs at depths >100 km [Ernst et al., 1998]. The small extent of dehydration from the subducting continental lithosphere will be discussed later. However, because van den Beukel [1992] assumed a hydrostatic $P_w$ within the slab, I also test such a case, i.e., $\lambda = 1/3$. [22] The differential stress for the ductile part of these faults can be estimated from the power laws of the crustal and mantle material, strain rate, and temperature $T$ and pressure $P$ in the subducting plate, using

$$\sigma_1 - \sigma_3 = [\varepsilon / A']^{1/\alpha} \exp[(E^* + \nu^* P)/nRT],$$

where $A''$ is the constant converted from the constant $A$ of the triaxial compression experiment to that of the plain strain problem (Appendix B), as

$$A'' = (3/4)^{(n+1)/2} A$$

I calculate temperature $T$ within the slab using the expressions of Molnar and England [1990], on the basis of temperature $T_s$ at the slab surface obtained in section 3 (see Appendix C). I use the rheological parameters listed in Table 3. The strain rate within the plate is lowered by a factor of 100 from that at the thrust zone.
Examples of the calculation of strength profiles along \( F_1 \) and \( F_3 \) are shown in Figure 3. The strength minima appear at the bases of the upper crust and the lower crust. It is required that \( F_2 \) is along these minima. I assume that \( F_2 \) is located at the base of the upper crust, i.e., that only the upper crust is sheared off, because shearing-off of the lower crust requires a few times larger driving force. Consistently, subduction of the continental plate leads generally to the detachment of the upper crust only [van den Beukel, 1992, p. 316]. Shear traction \( T_2 \) along \( F_2 \) is calculated by using equation (3), with the same strain rate for \( F_1 \) and \( F_3 \). Traction \( T_1 \), \( T_2 \) and \( T_3 \) required to fracture \( F_1 \), \( F_2 \) and \( F_3 \), respectively, are obtained by the integration of the strength over these faults.

5. Depths Where Shearing-Off Occurs

As the length of \( F_c \) becomes larger, \( T_c \) becomes larger, and if equation (1) is satisfied, shearing-off occurs. If not, the continental crust will subduct deep into the mantle. In this study, the updip end of \( F_1 \) is fixed at a distance of 10 km from the trench on the thrust zone, taking into account that the shallowest portion of the subduction boundary has a low stable friction [Byrne et al., 1988; Scholz, 1998]. The values of \( T_1 \), \( T_2 \), and \( T_3 \), and the sum of them, \( T_{all} \), are calculated as a function of the distance of the updip end of \( F_3 \) along the thrust. The values of \( \lambda \) are set to be zero for these intraslab faults and it is varied only for \( F_c \).

Figure 4 shows an example of a plot of these tractions, with \( T_c \) versus the distance from the trench, for \( \lambda = 0, V = 3 \) cm/a, \( \delta = 30^\circ \), and \( Q_s = 70 \) mW/m\(^2\). The value of \( T_{all} \) increases with the distance because \( T_2 \) increases with it. \( T_2 \) also increases due to the activation volume effect. In contrast, \( T_1 \) does not increase significantly because the effect of the activation volume and that of the temperature counteract each other. The activation volume effect on \( T_2 \) makes the curve of \( T_{all} \) concave upward, and \( T_2 \) does not cross \( T_{all} \) at all if it does not before \( T_{all} \) starts to deflect upward.

Figure 5a shows the depths of the updip end of \( F_3 \) at which shearing-off occurs versus \( Q_s \) for various \( \lambda \) ranging from 0 to 0.95 (\( V = 3 \) cm/a and \( \delta = 30^\circ \)); no symbol for \( Q_s < 70 \) mW/m\(^2\) means that there was no solution. As \( Q_s \) becomes higher, shearing-off occurs at shallower depths, as noted by van den Beukel [1992]. For \( Q_s \) as high as \( 90 \) mW/m\(^2\), \( \lambda \) as large as 0.95 makes shearing-off occur if the leading edge subducts to a depth of \( \sim 100 \) km. However, for \( Q_s \sim 70 \) mW/m\(^2\), which is typical for continents [Pollack and Chapman, 1977], shearing-off does not occur unless \( \lambda \) is smaller than \( \sim 0.4 \). In other words, in the environment with high \( \lambda \), all the continental crust would subduct into the mantle without being sheared off, unless \( Q_s \) is very high.

These results significantly differ from those of van den Beukel [1992] in which \( \lambda \) of 0.88–0.97 are used, and shearing-off still occurs for low to moderate heat flow values of 65–80 mW/m\(^2\). This distinction may arise from different assumptions used in his models; they are inclusion of the buoyancy of the crust as a driving force, hydrostatic \( P_w \) within the crust of the subducting lithosphere, and usage of the wet quartzite rheology for the upper crust. I make sure that the latter two are not the factors that bring the shallow depths of shearing-off in the work of van den Beukel [1992]. I test the hydrostatic \( P_w \) (\( \lambda = 1/3 \)) in the subducting crust and obtain the depths of shearing-off...
shallower only by a few km. The usage of the rheological parameters of wet quartzite [Koch et al., 1980] for the upper crust in the way I use in the present study makes the depths of shearing-off rather deeper by 20–30 km. This may be caused by the reduction of $T_c$ in the ductile part of the thrust more significantly than at $F_1$ and $F_2$. Because details of the calculation of $T_c$ by van den Beukel [1992] are not known, further investigation on this point is difficult. I further test the cases including the buoyancy of the crust as a driving force, with hydrostatic $P_w$ and the rheological parameters as used in this study, and obtain depths of shearing-off similar to those of van den Beukel [1992] for $\lambda \sim 0.93$. The magnitude of the buoyancy force is $\sim \Delta \rho gh_m L \sin \delta$, where $\Delta \rho$ is the density difference between the upper crust and the mantle ($\sim 500$ Kg/m$^3$), $g$ is the gravity of acceleration, and $L$ is the length of the thrust zone. For the case shown in Figure 4 where $L \sim 60$ km, this amounts to be $2.7 \times 10^{12}$ N/m, about a half of $T_c$. This assures that even for a large $\lambda$, inclusion of the buoyancy makes shearing-off at shallow depths possible. I therefore believe that the difference of the results by van den Beukel [1992] from the present study mainly come from the inclusion of the buoyancy force of the crustal block.

[draft: Figure 5b shows the depths of shearing-off versus $Q_s$ when $V$ is varied between 1.5 and 6 cm/a. Dip angle $\delta$ is fixed at $30^\circ$. The smaller the convergence rate, the shallower shearing-off occurs. This is because the plate is warmed by the slower rate of convergence, being hotter, resulting in reducing $T_1$ to $T_3$ and making shearing-off to occur earlier. Conversely, larger $V$ makes the slab colder, and the shearing-off occurs at greater depths. Because of this effect, at the slower convergence rate such as 1.5 cm/a, shearing-off occurs even at relatively higher values of $\lambda$ such as 0.8–0.6 at depths of 74–87 km for $Q_s = 65–60$ mW/m$^2$, yet having no solution for $Q_s < 55$ mW/m$^2$.]

29 Figure 5c shows the depths of shearing-off versus $Q_s$ when $\delta$ is varied between 15 and $45^\circ$. $V$ is fixed at 3 cm/a. The smaller the dip angle, the shallower shearing-off occurs. The smaller dip angle makes the upper edge of $F_3$ shallower for the same thrust zone length and, thus, $T_3$ smaller. Because $T_c$ is roughly proportional to the length of $F_c$ (Figure 4), the crossing point of $T_c$ and $T_{\text{all}}$ becomes shallower. This explains why the smaller $\delta$ the shallower shearing-off occurs. Conversely, the larger the dip angle, the deeper shearing-off occurs. Because of this effect, when $\delta$ is $15^\circ$, shearing-off occurs even at relatively higher values of $\lambda$ such as 0.8–0.6 at depths of 32–34 km for $Q_s = 70–60$ mW/m$^2$, yet having no solution for $Q_s < 50$ mW/m$^2$.

6. Discussion

In the previous section, I have shown that $Q_s$ and $\lambda$ are the main controlling factors to determine the depth where shearing-off occurs. For typical continents with $Q_s < 70$ mW/m$^2$ [Pollack and Chapman, 1977], shearing-off occurs for only $\lambda$ smaller than $\sim 0.4$. Although this value shifts to a larger one, when the dip angle or the convergence
velocity is small, it is still much smaller than seen in subduction zones \( (\lambda \sim 0.95) \) [e.g., \textit{van den Beukel and Wortel}, 1988; \textit{Tichelaar and Ruff}, 1993; \textit{Peacock}, 1996; \textit{Cattin et al.}, 1997; \textit{Lamb}, 2006]. This dehydration is probably manifested by the intermediate-depth intraslab seismic activities [\textit{Kirby et al.}, 1996; \textit{Seno and Yamanaka}, 1996; \textit{Peacock and Wang}, 1999; \textit{Hacker et al.}, 2003; \textit{Yamasaki and Seno}, 2003]. For example, dehydration from the altered basaltic crust starts around \( \sim 30 \text{ km} \) depth for young slabs and \( \sim 70 \text{ km} \) depth for old slabs [\textit{Hacker et al.}, 2003; \textit{Yamasaki and Seno}, 2003]. The fluids dehydrated from the slab would migrate through the fractured conduit in the thrust zone or escape upward into the wedge. This would be a unique source for the high pore fluid pressure at the thrust where an oceanic plate is subducting, because no other metamorphic reaction is expected at such depths in the subduction zone.

Compared with the subduction of an oceanic plate, there are pieces of evidence indicating that the extent of dehydration is small for the subducting continental plate. This was first noticed by \textit{Seno and Yamasaki} [2003] in the Nankai Trough, SW. Japan, on the basis of the fact that, where island arc type crust is subducting, no low-frequency tremor is observed [\textit{Obara}, 2002]. They also showed that no intraslab earthquakes are occurring in the subducting crust in such places. These provide evidence for a small extent of dehydration from the subducting island arc crust.

intraslab seismicities are generally absent in collision zones, such as the Himalayas and the Zagros. This suggests no or scarce dehydration from the subducting slab in collision zones. The crustal deformation associated with underthrusting in these places is described by the detachment model, in which the shallow portion of the thrust is adhered. In contrast, in subduction zones, the crustal deformation is described by the back slip model [Savage, 1983]. This implies that the adherence at the thrust is stronger in collision zones than in subduction zones [Seno, 2007]. On the basis of these facts, Seno [2007] proposed that the continental or island arc crust, when it is subducted to depths of tens of km, does not scarcely dehydrates. This is probably due to the absence of dehydration reactions of major hydrous minerals in the continental crust at these depths. If white mica and biotite are contained, they dehydrate at greater depths (>100 km [Ernst et al., 1998]). [35] The small extent of dehydration from the slab with continental crust would lead to small P_w and λ at the thrust zone, because there is no other effective source of fluids filling the thrust zone at depth, as mentioned above. In contrast, the metamorphic reactions in the oceanic slab provide enough amount of fluids at the thrust, although details of the mechanism of the fluid transport from the source to the shallower portion of the thrust (10–60 km) would be a subject of investigation in future studies. No matter the transport mechanism is, however, the nearly absence of dehydration will lead to small λ at the thrust zone, leading to shearing-off of the upper crust at shallow depths from the underthrusting continental plate. [36] The offscraping of the continental crust at shallow depths has been believed to be one of the most typical features of collision zones, as seen in the Himalayas [Molnar, 1984; Butler, 1986; Mattauer, 1986] and the Zagros [Synder and Barazangi, 1986; Bird et al., 1975]. The intense deformation in the hinterland of the collision zone, like those in Tibet and C. Asia [Molnar and Tapponnier, 1975] has been believed to be due to indentation of the buoyant colliding continent. It should be noted, however, shearing-off may not occur, with large λ, even if the continental plate is underthrusting. In fact, it is known that continental crust, by happen, can subduct into the mantle of a depth >100 km in many collision zones with UHP exhumation [Liou et al., 1994; Coleman and Wang, 1995; Ernst and Liou, 1999; Chopin, 2003]. Molnar and Gray [1979] also showed that the continental plate, from the viewpoint of buoyancy, might subduct into such a large depth, if the slab pull force is available. If λ is large, the shear stress at the thrust becomes small, resulting in modest deformation in the hinterland. These imply that the buoyancy of the subducted crust is not an essential factor to cause features associated with collision, such as shallow offscraping of a crustal sheet and intense deformation in the hinterland, but rather the absence of dehydration and small λ might be the essential factor.

6.2. Two Types of Collision Zones

[35] Although the Himalayas and the W. Alps are both type localities of collision zones, the features associated with them are quite different. Offscraping of the Indian continental crust has occurred in the foreland fold and thrust belt of the Himalayan collision zone [e.g., Molnar, 1984; Butler, 1986; Mattauer, 1986]. The overriding Asian continent has been extensively deformed by indentation of the Indian continent as shown in the crustal thickening [Allegre et al., 1984; England and Houseman, 1988] and in the extrusion [Molnar and Tapponnier, 1975]. Although offscraping at shallow depths has been dominant in the Himalayas, exhumation of the UHP terranes is known to have occurred in the early stage of the collision (~50 Ma) in the Hindu Kush [Kaneko et al., 2003]. In contrast, in the W. Alps, exhumation of HP/UHP terranes, forming numerous nappes, has been dominant, and the offscraping of the colliding continent is minor [e.g., Platt, 1986; Hsu, 1991; Escher and Beaumont, 1997]. The deformation of the overriding Adriatic plate has been also minor.

[36] I show that these differences can be explained by the evolution of P_w and λ in the thrust zone through the history of collision. Several collision zones that accompany intermediate-depth seismic activities, such as Vlancia, Hindu Kush, Timor, and Taiwan, give us a hint for investigating this problem. In these zones, dehydration from the leading oceanic slabs is manifested by the intermediate-depth seismic activities [Seno, 2007]. These can be regarded as transient or intermediate cases between subduction of an oceanic plate and underthrusting of the continental plate. It is expected that P_w and λ at the thrust zone of these places would be moderately large by the fluids released from the attached oceanic slab, making λ higher than typical collision zones. In such a case, the continental crust could subduct deep into the mantle due to the higher λ, as observed in the collision zones cited above [Seno, 2007].

[37] Prior to the collision of the Indian continent, the Tethys Sea had been subducting beneath the Asian continental fragments [e.g., Sengor, 1992]. In the early stage of the collision, the dehydration of the leading oceanic plate would have made the subduction of the continental crust to large depth possible, having made it suffer the UHP metamorphism. As the collision proceeded, the ocean-continent boundary would have subducted deep into the mantle, having, in turn, made the thrust zone starved of fluids. \( T_c \) would have gradually become large as λ decreased. When it satisfied the condition \( T_c > T_{di} \), part of the subducted continental plate would have been sheared off, having provided a source of the exhumed UHP terranes at the initial stage of the collision. Afterward, the continental plate continued to underthrust, and the upper crust would have been offscraped at shallow depths because the thrust zone has been devoid of fluids. This scenario explains why only the offscraping at shallow depths, along with the exhumation of the low-grade metamorphic rocks [Le Fort, 1986], occurred in the Himalayas during the succeeding stages. The level of \( T_c \) of the thrust would have been kept high during these stages, on the order of 100 MPa (Figure 2b), having resulted in the intense deformation of Asia by the indentation of India.

[38] In the case of the W. Alps, in contrast, a series of continents and oceans have been subducting beneath the Adriatic plate since Late Cretaceous [Platt, 1986; Escher and Beaumont, 1997]. When the continental plate was first subducted, similarly to the initial stage of the Himalayas, dehydration from the leading oceanic plate would have made λ large along the thrust, and the continental crust would have subducted into a great depth. After the oceanic
slab went deeper, part of the subducted continental crust was sheared off and was exhumed as the UHP rocks, such as the Dora Maira Massif [Chopin, 1987]. In contrast to the Himalayas, however, subduction of another oceanic plate succeeded [e.g., Platt, 1986], and the cycle of oceanic plate subduction - continental plate subduction - exhumation - oceanic plate subduction has been repeated. Offscraping at shallow depths, therefore, did not occur, except for the recent one of the European continental crust, for which a succeeding oceanic plate does not exist. \( P_w \) and \( \lambda \) are likely to have been kept at a relatively high level throughout the collision history in the W. Alps, leading to small \( \tau_e \) at the thrust zone. This would have made a small extent of indentation and deformation of the hinterland in the upper Adriatic plate.

6.3. Ophiolite Obduction

[39] Some types of ophiolite obduction [Coleman, 1971; Dewey and Bird, 1971; Dewey, 1976] and exhumation of HP/UHP rocks in subduction zones [Aoya et al., 2003; Ota et al., 2004] indicate that sheet-like bodies are possibly sheared off also from the subducted oceanic plate, not only from the continental one. To discuss these phenomena quantitatively, we need to model the temperature of the subducting oceanic slab, and calculate driving and resistive forces for shearing-off, in a similar manner to the present study. Leaving such work in the future, I discuss the problem only qualitatively using the results obtained in the previous sections.

[40] I note here that shearing-off them from the oceanic plate is generally difficult. First, in the scheme of Figure 1, \( T_2 \) is generally very large because the oceanic crust is as thin as \(~6\) km, which prevents the development of the ductile shear zone at the base of the crust. Secondly, \( P_w \) and \( \lambda \) are usually very large in subduction zones [e.g., van den Beukel and Wortel, 1988; Tichelaar and Ruff, 1993; Peacock, 1996; Cattin et al., 1997; Lamb, 2006], which reduces the level of \( T_e \).

[41] There might be two possible cases that overcome the first difficulty. If the oceanic plate is extremely young, the oceanic crust could still be hot enough to develop a ductile shear zone at its base, or within it. The temperature at a depth of 6 km from the surface of an oceanic plate becomes greater than 300°C for ages younger than 5 Ma, if the temperature is calculated by the plate model with the asthenosphere temperature of 1300 °C [Parsons and Sclater, 1977]. Next, if the mantle part of the subducting plate is serpentinized, dehydration in the slab mantle occur as it subducts deeper and warmed by conduction [Nishiyama, 1992; Seno and Yamanaka, 1996; Peacock, 2001; Hacker et al., 2003; Yamasaki and Seno, 2003]. The dehydration front of serpentinite in the subducting slab is expected to be located a few tens km below the slab surface based on the phase diagram [Yamasaki and Seno, 2003]. Since the strength along this dehydration front in the slab would be small [Fyfe, 1985], this could be used as a lower boundary of a block to be sheared off.

[42] However, even if the 1st difficulty is remedied by these situations, the 2nd one remains; \( T_e \) in subduction zones is generally small because \( P_w \) and \( \lambda \) are large due to dehydration from the altered basalt. I note here that there are some exceptional cases, however, in which \( \lambda \) might be very small. For example, where the Kinan Seamounts, the extinct spreading centers affected by the volcanism during 10 m.y. after the cessation of the spreading of the Shikoku Basin [Sato et al., 2002], are subducting beneath SW. Japan, intraslab seismicity does not occur in the subducted crust [Kodaira et al., 2002; Kurashimo et al., 2002], or no low-frequency tremor is observed [Obara, 2002]. Seno and Yamasaki [2003] suggested the absence of dehydration in the crust in this place, which might lead to small \( \lambda \). Another case is the subduction of the Izu-Bonin forearc beneath Kanto, central Honshu. The absence of intraslab seismicity in the crust [Hori, 2006] and low-frequency tremor [Obara, 2002] here also suggests that dehydration might be absent in the crust [Seno et al., 2001; Seno and Yamasaki, 2003]. These are possible places where shearing-off of part of the crust or the mantle from the oceanic lithosphere could occur, due to small \( \lambda \), if a weak zone is developed in the slab. The exhumation of the Sanbagawa HP/UHP terrane occurred at the end of Cretaceous in the SW. Japan forearc. The subducting slab was very young in this case [Isozaki, 1996; Aoya et al., 2003], and Terabayashi et al. [2005] reconstructed seamount sequences from the geological sections of the exhumed UHP terrane. Therefore the exhumation of this terrane might have occurred in a favorable situation, satisfying both the conditions depicted above. Ben-Avraham et al. [1982] once proposed that ophiolites are emplaced in association with collision of buoyant features such as the continental fragments and plateaus/seamounts. This might be, in the context of this study, interpreted as small \( \lambda \) and large \( T_e \) due to insufficient dehydration from these features upon subduction facilitate shallow offscraping of such features, when a ductile shear zone is developed inside the slab.

6.4. Future Scopes

[43] There are several directions that might be explored in the future. The model in this study can be extended to the cases of shearing-off of the oceanic plate, as mentioned above, once the temperature of subducting oceanic lithosphere is calculated. The obduction of ophiolites and exhumation of HP/UHP terranes would be able to be discussed in a more quantitative manner. There are also many buoyant features having thicker crust with low densities on the oceanic plate, such as remnant ridges, plateaus, and seamounts [Ben-Avraham et al., 1981]. The fate of these buoyant features could be discussed in terms of \( \lambda \), their spatial dimensions, and crustal thickness. Discussion of their fate in terms of buoyancy only [e.g., Cloos, 1993] would not be adequate.

[44] I infer that a low value of \( \lambda \) is caused by a small extent of dehydration of the subducting continental plate. The connection between the dehydration process from the crust and \( \lambda \) is still obscured, and this should be more substantiated through high-pressure experimental and theoretical work. Simultaneously, the fate of fluids derived from metamorphic reactions of altered basalt/serpentinites should be investigated more through theoretical work. This is not only applicable to the subducting oceanic plate generally, but also important for collision zones where subduction of the oceanic plate leading the trailing continent may gradually shifts to subduction of the continental plate.
The result that \( \lambda \) controls the depth of shearing-off may lead to the basic idea that the tectonic style of the Earth is controlled by \( \lambda \). Although this idea was explored by the earlier paper of the same author [Seno, 2007] and the effects of \( \lambda \) in collision zones are discussed in a quantitative manner in this study, effects of \( \lambda \) and dehydration should be pursued in various tectonic, volcanic, and seismic activities on and within the Earth, and a general model might be explored to explain the different tectonic styles of the Earth on the basis of \( \lambda \) in a unified manner.

7. Conclusions

I study the conditions for a crustal block to be sheared off from a subducting continental lithosphere. I show that the buoyancy does not operate selectively to the crustal block. I suppose that shearing-off occurs when the shear traction at the thrust, whose maximum value is calculated from the strength along the thrust, overcomes the strength of the boundary faults of the detached block, and determine the depths of the leading edge of the block when shearing-off occurs. The temperatures of the slab and of its surface are calculated by the steady state formula of Molnar and England [1990] as a function of the surface heat flow, the dip angle, and the convergence velocity. The shear stress and temperature at the ductile part of the thrust zone are determined to satisfy both the ductile flow law and the temperature due to shear heating. The shear stress at the brittle part is represented by the Byerlee’s friction law using the pore fluid pressure ratio \( \lambda \). I show that shearing-off of an upper crustal block is possible only for small \( \lambda \) if the surface heat flow is below 70 mW/m\(^2\) unless the convergence rate is very small.

These results imply that temporal and/or regional variation of \( \lambda \) might help us to understand various features associated with shearing-off, such as exhumation of HP/UHP terranes, offscraping of continental crust at shallow depths, and obduction of part of the oceanic lithosphere. Taking into account the possibility that the continental or island arc crust does not dehydrate when it is subducted [Seno, 2007; Seno and Yamasaki, 2003], the above condition for shearing-off implies that the essential factor to cause features characteristic of collision might not be the buoyancy of the subducting crust, but is the absence of dehydration from it. In the early stage of collision, dehydration from the subducted oceanic plate leading the continent would lubricate the thrust zone, making \( \lambda \) to be large. The continental plate is then able to subduct to great depths. As the collision proceeds, the thrust zone becomes less lubricated. \( \lambda \) becomes smaller, resulting in shearing-off and providing a source of exhumation of the UHP terranes in the early stage of collision. In later stages, absence of dehydration from the slab causes offscraping of the continental crust at shallow depths, producing a stack of sliced sheets often seen in the fold and thrust belts. The shear stress at the thrust zone becomes as high as \( \sim 100 \) MPa, leading to the intense deformation of the hinterland. This could be the case of the Himalayas.

If the subduction of an oceanic plate succeeds in later stages, the dehydration of the oceanic plate makes \( \lambda \) larger again, subduction of another trailed continental plate becomes possible, and the shear stress level of the thrust zone remains low on the average. The W. Alps, where exhumation of HP/UHP terranes have occurred extensively and repeatedly, with a small amount of both offscraping and deformation in the hinterland, could be this case.

For a typical oceanic plate, it is generally difficult to cause shearing-off, i.e., obduction of ophiolite or exhumation of HP/UHP terranes, because of the absence of conditions of developing ductile shear zones within the subducting crust, except for the case of the very young plate. Shearing-off is also difficult because of large \( \lambda \) due to dehydration from the subducted altered crust. For shearing-off to occur, two conditions seem to be satisfied. (1) Weakening along a dehydration front bordering a region of serpentinized slab mantle, or at the base of the crust in the plate younger than 5 Ma, and (2) small \( \lambda \) at the thrust zone caused by the absence of dehydration from the subducted crust. Some places of the subduction zones near Japan show that these conditions are partly satisfied. Exhumation of HP/UHP terranes or ophiolitic obduction might have occurred in the situation where these conditions were met.

In the future, shearing-off from the subducting oceanic plates should be treated in a more quantitative manner, by calculating the temperature of the slab. Further studies are needed to elucidate the connection between the extent of dehydration of the slab and the value of \( \lambda \). The results in this study lead to a general view that the tectonic style on the Earth, like subduction and collision, is controlled by \( \lambda \). Pursuing this view through various tectonic, seismic and volcanic activities may provide a new paradigm beyond plate tectonics.

Appendix A: Normal Stress Operating at the Thrust

In this appendix, I show that normal stress \( \sigma_n \) operating at thrust zone \( F_c \) is given by the lithostatic pressure based on the force balances in the shallow portion of the subducting plate and the lithospheric wedge of the upper plate. A schematic cross-section of the shallow portion of the subduction zone is depicted in Figure A1. The downgoing plate is in contact with the upper plate at the thrust zone \( AC \) with average shear stress \( \tau^* \), length \( L \), and dip angle \( \delta \). BC is the location of the aseismic front, where interplate coupling is terminated [Yoshii, 1979].

I first try to obtain relations between the slab pull \( T_{sp} \), ridge push \( T_r \), and \( \tau^* \). I ignore in the following discussion all the forces operating at the interfaces with the asthenosphere, regarding them small compared with \( T_{sp} \) and other tectonic forces. \( T_{sp} \) is the negative buoyancy multiplied by \( \sin \delta \) integrated over the slab deeper than point \( C \) [McKenzie, 1969]. It could be a force on the order of \( 10^{13} \) N/m, if the slab has a length of \( \sim 300 \) km [Molnar and Gray, 1979].

I first consider the force balance in rectangle ACEF. I assume that \( T_{sp} \) is directing downdip because forces perpendicular to the slab surface are likely to be balanced with the fluid dynamic forces associated with subduction [Stevenson and Turner, 1977]. \( T_{sp} \) should then be balanced with the downdip component of \( T_r \) and the shear traction at the thrust, and I obtain

\[
T_{sp} + T_r \cos \delta = \tau^* L. \tag{A1}
\]
Next, I consider the force balance in trapezoid GBDF. The horizontal component of $T_{sp}$ is balanced with $T_r$, and collision force $T_c$ operating at the asismatic front of the upper plate, then I obtain

$$T_{sp} \cos \delta + T_r = T_c.$$  \hspace{1cm} (A2)

The vertical forces at the asismatic front $T_v$ and at the trench $T_{vo}$ are balanced with the vertical component of $T_{sp}$, then I obtain

$$T_{sp} \sin \delta = T_v + T_{vo}.$$  \hspace{1cm} (A3)

Normal traction at the thrust zone $T_n$, in excess of the lithostatic pressure, is

$$T_n = T_c \sin \delta - T_r \cos \delta.$$  \hspace{1cm} (A4)

Substituting $T_c$ and $T_v$ using (A2) and (A3) in (A4), I obtain

$$T_n = T_r \sin \delta + T_{vo} \cos \delta.$$  \hspace{1cm} (A5)

Because the ridge push force is on the order of $10^{12}$ N/m [Parsons and Richter, 1980] and gives a stress of a few MPa, it is negligibly small compared with the lithostatic pressure, which amounts to $\sim 1$ GPa at a depth of 30 km. Similarly, because the shear force $T_{vo}$ at the trench is on the order of $10^{12}$ N/m [Caldwell et al., 1976], it can also be neglected. Therefore we can regard that normal traction $T_n$ acting at the thrust is independent on $T_{sp}$ and negligible compared to the lithostatic pressure. This is used in the estimation of the brittle strength at AC based on the Byerlee's friction law in section 3 and in the force balance between $T_c$ and $T_{all}$ in section 4. The smallness of $T_r$ also implies that the average shear stress at the thrust zone is balanced by $T_{sp}$ through equation (A1).

**Appendix B**

Constants $A'$ and $A''$ in equations (3) and (9) in the text are converted from $A$ in the triaxial compression test as follows. Let us write the generalized constitutive equation as

$$\varepsilon_{ij} = A' \sigma^{n-1}_{ij} \sigma_{ij} \exp[-(E^* + v^*P)/RT].$$  \hspace{1cm} (B1)

where $A'$, $n$, $E^*$ and $v^*$ are material constants, $\varepsilon_{ij}$ is the strain rate tensor, $\sigma_{ij}$ is the deviatoric stress tensor, and $\sigma$ is the 2nd invariant of the stress tensor, as $\sigma^2 = \sigma_{ij}^2/2$. In the triaxial compression test, let $\sigma_1^*$ be the axial principal deviatoric stress, and $\sigma_2^*$ and $\sigma_3^*$ are the other principal deviatoric stresses in the orthogonal directions. Then, $\sigma_2^* = \sigma_3^* = -\sigma_1^*$, and we obtain $\sigma^2 = 3\sigma_1^2/4$. The constitutive equation for the triaxial compression test in the form of (B1) is

$$\varepsilon_1 = A' (3\sigma_1^2/4)^{(n-1)/2} \sigma_1^* \exp[-(E^* + v^*P)/RT]$$

$$= 2A' 3^{-(n+1)/2} (\sigma_1^* - \sigma_3^*)^n \exp[-(E^* + v^*P)/RT].$$  \hspace{1cm} (B2)

We thus obtain

$$A = 2A' 3^{-(n+1)/2}$$

or

$$A' = 3^{(n+1)/2} A/2.$$  \hspace{1cm} (B3)

In the 2-dimensional simple shear strain problem, which is suitable for $F_c$ and $F_2$, taking $x_1$ along the thrust plane, and $x_3$ orthogonal to it, other components than $\sigma_{13}^* = \sigma_{31}^*$ and $\varepsilon_{13} = \varepsilon_{13}$ are zero. We obtain $\sigma^2 = \sigma_{13}^2$. Inserting these into (B1),

$$\varepsilon_{13} = A' \sigma_{13}^* \exp[-(E^* + v^*P)/RT]$$

or

$$\sigma_{13}^* = \tau_c = [\varepsilon/|A'|]^{1/n} \exp[(E^* + v^*P)/nRT].$$  \hspace{1cm} (B4)

This indicates that the constant for the simple shear problem is $A'$. Rewriting equation (B1) in the form using $E^2 = \varepsilon_{ij} \varepsilon_{ij}/2$,

$$\varepsilon_{ij} = A' \sigma^{n-1}_{ij} \sigma_{ij} \exp[(E^* + v^*P)/nRT].$$  \hspace{1cm} (B5)
In the 2-dimensional plain strain problem, applied to the strength of \( F_1 \) and \( F_3 \), \( \varepsilon_1 = -\varepsilon_3 \) and \( \varepsilon_2 = 0 \). We obtain \( E^2 = \varepsilon_1^2 \) and \( \varepsilon_1 - \varepsilon_3 = 2\varepsilon_1 \). Inserting these into (B5),

\[
\sigma_1 - \sigma_3 = A^{(1/2)}\varepsilon_1^{(n-1)/2}e_1\exp[(E^* + v*\varepsilon_1)/nRT] = 2A^{(1/2)}\varepsilon_1^{n/2}\exp[(E^* + v*\varepsilon_1)/nRT].
\]  

(B6)

Rewriting this, we obtain

\[
\varepsilon_1 = 2^{-\theta}A'(\sigma_1 - \sigma_3)^{\theta}\exp(-\varepsilon_3).\]  

(B7)

Therefore the constant for the 2-d plain strain problem is

\[
A'' = 2^{-\theta}A' = (3/4)^{(n+1)/2}A.
\]  

(B8)

**Appendix C**

[55] Temperature at the thrust zone \( T_s \) is composed of the three terms due to (1) advection \( T_{s1} \), (2) internal radiogenic heating \( T_{s2} \), and (3) frictional heating \( T_{s3} \). From *Molnar and England* [1990, pp. 4838–4839], the expression for advection is

\[
T_{s1} = Q_s z_t/k/S,
\]  

(C1)

where \( Q_s \) is the surface heat flow, \( z_t \) is the depth of the point of the thrust plane, \( k \) is the thermal conductivity. \( S \) is the divisor, which is common to all the terms, and written as

\[
S = 1 + b \sin \delta u_T V/k \beta, 
\]  

(C2)

where \( b \) is the constant close to 1, which depends on the exponent of the temperature increase with time [Molnar and England, 1990, p. 4837], \( \delta \) is the dip angle of the thrust, \( u_T \) is the distance from the trench axis along the thrust in the downdip direction, \( V \) is the convergence rate, and \( \beta \) is the thermal diffusivity.

The expression for radiogenic heating is

\[
T_{s2} = (A_0 D^s (1 - \exp(-z_T/D))(1 + z_T/D) - A_0 D^s \exp(-a/D))/(ks),
\]  

(C3)

where \( A_0 \) is the radiogenic heat generation at the surface, \( D \) is the scaling depth of the radiogenic heat generation assuming the exponential decay, \( a \) is the plate thickness.

The expression for shear heating is

\[
T_{s3} = (\tau_s V/k)/S,
\]  

(C4)

where \( \tau_s \) is the shear stress at the thrust. The temperature within the subducting plate is determined by the temperature gradient given by

\[
\partial T/\partial y = -bT_s(u_T)/\sqrt{su_T}/V,
\]  

(C5)

where \( y \) is the direction perpendicular to the slab surface [Molnar and England, 1990, equation (13), p. 4837].

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