INTRODUCTION

In island arcs the oceanic lithosphere underthrusts episodically beneath the continent at the time of a great earthquake. Before an earthquake the continental lithosphere is dragged down with the descending slab of oceanic lithosphere. When the stress reaches some critical value, the continental lithosphere elastically rebounds, and hence a great earthquake results. In a simple elastic rebound theory the continental lithosphere returns to its original position after the earthquake, and postseismic deformation should be zero. It is well known, however, that an earthquake sequence leaves permanent uplift and landward tilting of coastal terraces and offshore islands. This mode of crustal deformation has accumulated at least through the late Quaternary [e.g., Yonekura, 1975]. A question arises as to what extent such landward tilting of the free surface can be traced back toward the trench. It is rather unlikely that it can be traced back to the trench axis. The vicinity of the trench axis is full of evidence for actively deforming features [e.g., Moore and Karig, 1976], and there is little evidence for permanent uplift and landward tilting to the degree expected from coastal observations. It may be noted in conjunction with the above question that the aftershock area of a great earthquake tends to be confined geographically to a deep-sea terrace and rarely extends further trenchward beyond the outer ridge of the deep-sea terrace [Den, 1968; Moge, 1969; Yonekura, 1975].

The outer ridge of the deep-sea terrace sometimes emerges as islands, an example being the Middleton Island in the Gulf of Alaska. An interesting geodetic observation on this island was reported after the Alaskan earthquake of March 27, 1964, which is one of the greatest earthquakes of this century. Figure 1 shows the aftershock area, the observed land deformation, and a presumed fault model of this earthquake. As is seen in Figure 1a, there is a considerable gap between the seaward boundary of the aftershock area and the trench axis. On the profile across the aftershock area, along the line PQ, the coseismic deformation was observed as shown in Figure 1b [Hastie and Savage, 1970]. Miyashita and Matsuura [1976] inverted these data to obtain a model of the fault plane (Figure 1c) and the slip dislocation across it (Figure 1d). Their model clearly shows that the thrust fault did not break the sea bottom at the coseismic stage. The upward and seaward end of the fault locates at a depth of about 10 km and about 100 km landward away from the trench axis. The aftershock distribution well delineates this seaward end of the fault plane (Figure 1a). Middleton Island in the Gulf of Alaska is situated farther trenchward beyond it (Figure 1a), where precise leveling has been repeated since 1964. The results show a continuous tilting of the island down to the northwest, normal to the trench axis, at a rate of 5 μrad/yr during 1964–1975 [Prescott and Lisowsky, 1977]. This tilt movement has been interpreted as a result of the seaward and upward extension of the rupture, which remained at depths at the time of the Alaskan earthquake [Prescott and Lisowsky, 1977]. Prescott and Lisowsky [1977] suggested two possible modes of seaward extension: main thrusting along the lithospheric boundary or subsidiary thrusting along imbricate faults that branch upward from the lithospheric boundary. The question of which is the more likely process may be directly related to the first question as to the episodic permanent uplift and landward tilting. Another related question is how a great earthquake process is relevant to a deep-sea terrace and a depressional feature of the trench. This consistency implies a causal relationship between great earthquake activities and geomorphological features near the trench.
Kurile Earthquake of October 20, 1963

A seaward extension of the rupture after a great earthquake appears to occur not only aseismically but also in a brittle way so as to generate a seismic shock. A typical example of this type of shock is the largest aftershock of the great Kurile earthquake of 1963 ($M_s = 8.2$), whose major sequence is given in Table 1. Figure 2 shows the aftershock areas of the mainshock, the largest foreshock, and the largest aftershock whose mechanisms are all typically low-angle thrust faulting [Stauder and Mualchin, 1976; Kanamori, 1970a; Ben-Menahem and Rosenman, 1972]. The aftershock area of the largest aftershock is located trenchward of the preceding two shocks. Redetermination of its hypocenter indicates a very shallow origin ($h \sim 5$ km [Takehara and Suzuki, 1977]). The spatio-temporal relation between the mainshock and the largest aftershock suggests that the latter event represents a seaward and upward extension of the rupture which remained at depths at the time of the mainshock.

This aftershock is unique in many aspects among thrust earthquakes in the related region. First of all it generated unexpectedly large tsunamis, although the tsunami source area of about $110 \times 45$ km$^2$ [Solomon, 1965a, b] is more or less the same as the aftershock area of about $100 \times 60$ km$^2$ (Figure 2). Figure 3 shows the tide gauge records of the mainshock and the largest aftershock at some stations in the Kurile Islands. As was pointed out by Solomon [1965a, b], maximum tsunami height of the aftershock is about 2.5 times as small as that of the mainshock. Such a tsunami height ratio is also indicated from the observations very far from the tsunami sources. For example, at the Pacific coasts of California the tsunami height is reported to be about 70 cm for the mainshock and 30 cm for the aftershock [Solomon, 1965b]. Near the tsunami sources, at the southern extremity of Urup Island, the inundation height of the aftershock is even greater than that of the mainshock [Solomon, 1965b]. Long-period seismic waves from the aftershock, on the other hand, are significantly weaker than those expected from the tsunami data. Figure 4 shows the Benioff strain seismograms (galvanometer period of 180 s) at Pasadena, California. The trace amplitude of the long-period G1 waves of the aftershock is less than one tenth of that of the mainshock. Ben-Menahem and Rosenman [1972] obtained an average amplitude ratio of about 10 for mantle Love and Rayleigh waves of the mainshock to the aftershock, on the other hand, are significantly weaker than those expected from the tsunami data. Figure 4 shows the Benioff strain seismograms (galvanometer period of 180 s) at Pasadena, California. The trace amplitude of the long-period G1 waves of the aftershock is less than one tenth of that of the mainshock. Ben-Menahem and Rosenman [1972] obtained an average amplitude ratio of about 10 for mantle Love and Rayleigh waves of the mainshock to the aftershock. Their result shows that the amplitude ratio has a slight tendency to decrease as the period decreases from 300 to 100 s [Ben-Menahem and Rosenman, 1972, Figures 12 and 13]. The ratio at the long-period end is about 11. This tendency for the spectral ratio to decrease with decreasing period suggests that the process time of the aftershock is smaller than that of the mainshock. Ben-Menahem and Rosenman [1972] took the spectral amplitude ratios for mantle Love and Rayleigh waves of the mainshock to the aftershock. Their result shows that the amplitude ratio has a slight tendency to decrease as the period decreases from 300 to 100 s [Ben-Menahem and Rosenman, 1972, Figures 12 and 13]. The ratio at the long-period end is about 11. This tendency for the spectral ratio to decrease with decreasing period suggests that the process time of the aftershock is smaller than that of the mainshock. A smaller process time was demonstrated more directly by Furumoto [1979], who analyzed the multiple Rayleigh waves for which great circle phase velocity and Q are accurately known [Nakanishi, 1978, 1979]. He obtained process times of 93 and 62 s from these phase spectra in the period range 150-300 s for the mainshock and the aftershock, re-

Table 1. Major Sequence of the Great Kurile Earthquake of 1963

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>$h$, km</th>
</tr>
</thead>
<tbody>
<tr>
<td>Largest foreshock Oct. 12</td>
<td>44.8°N, 149.0°E</td>
<td>40</td>
</tr>
<tr>
<td>Mainshock  Oct. 13</td>
<td>44.8°N, 149.3°E</td>
<td>60</td>
</tr>
<tr>
<td>Largest aftershock Oct. 20</td>
<td>44.7°N, 150.7°E</td>
<td>25</td>
</tr>
</tbody>
</table>
spectively. These values are qualitatively consistent with the spectral ratio curves of Ben-Menahem and Rosenman [1972].

The above process time of 62 s was determined from surface waves at periods below 300 s. A process with a time constant much longer than 300 s might not have been detected from the surface waves if it exists. A process with such a long time constant is, however, unable to generate effectively not only long-period surface waves but also tsunamis. Taking a characteristic length of the tsunami source as \( L \sim 100 \text{ km} \) and a characteristic water depth as \( H \sim 6 \text{ km} \) (Figure 2), we obtain a characteristic time for tsunami generation to be \( L / (gH)^{1/2} \sim 400 \text{ s} \), where \( g \) is the gravitational acceleration. Only a process with a time constant smaller than this characteristic time can produce substantial tsunamis, yet this characteristic time is not very different from the observed period range of surface waves. There is therefore little reason to believe that the tsunamis were mainly generated by a source whose process time is much longer than 60 s, a value determined from long-period surface waves. For the mainshock the free oscillation data with periods up to 600 s are consistent with a source time function determined from the surface wave data [Abe, 1970]. It may then be concluded that the tsunamis of the mainshock were generated by a source with a process time of about 100 s, which was determined from long-period surface waves. It appears almost certain that the process time of the aftershock is not longer than that of the mainshock and therefore that process time alone cannot explain why the efficiency of tsunami generation was so different between the mainshock and the aftershock.

**NEMURO-OKI EARTHQUAKE OF JUNE 10, 1975**

A very similar tsunami earthquake occurred off eastern Hokkaido, Japan, in 1975. Figure 5 shows tsunami source areas of large earthquakes off eastern Hokkaido [Hatori, 1974]. The tsunami source of the Nemuro-Oki earthquake of June 10, 1975 (43.0°N, 147.7°E, \( h = 15 \text{ km} \), \( M_s = 7.0 \)), is located trenchward of the preceding larger events, such as the eastern Hokkaido earthquake of August 11, 1969 (43.5°N, 147.4°E, \( h = 28 \text{ km} \), \( M_s = 7.9 \)), or the Nemuro-Oki earthquake of June 17, 1973 (43.2°N, 145.8°E, \( h = 48 \text{ km} \), \( M_s = 7.7 \)). The aftershock distribution also supports this trenchward location of the 1975 source, which, however, should be closer to the 1973 source than Figure 5 indicates [Takemura and Suzuki, 1977; K. Shimazaki, personal communication, 1977]. The 1975 event may be regarded as an aftershock of either the
1969 event or the 1973 event. Redetermination of its hypocenter indicates a very shallow origin \( h \sim 3 \text{ km} \) [Shimazaki and Geller, 1977] obtained a somewhat smaller aftershock area \( (\sim 50 \times 35 \text{ km}^2) \). The top illustration of Figure 6 shows the time of the preceding larger events.

The tsunami source area of the 1975 event is about \( 100 \times 60 \text{ km}^2 \) (Figure 5), which is roughly equal to the aftershock area of \( 75 \times 85 \text{ km}^2 \) [Suzuki and Sugimoto, 1975]. Shimazaki and Geller [1977] obtained a somewhat smaller aftershock area \( (50 \times 35 \text{ km}^2) \). The top illustration of Figure 6 shows the tsunami height versus distance plots [Hatori, 1975] for the 1973 and 1975 Nemuro-Oki earthquakes, whose mechanisms are both typically low-angle thrust fault [Shimazaki, 1975; Takemura et al., 1977; Shimazaki and Geller, 1977]. The tsunami height ratio of the 1973 event to the 1975 event is roughly 1.5, although the scatter is very large. The bottom illustration shows the tide gauge records of the two events at Miyagi-Enoshima, Japan [Hatori, 1975]. The high-frequency components are relatively dominant in the tsunamis from the 1975 event, whose trace amplitudes are about two thirds of those of the 1973 event. Excitation of long-period surface waves, on the other hand, is much smaller than expected from these tsunami data. Figure 7 shows the G2 waves at stations KTG, Greenland, and RIV and ADE, Australia. Figure 8 shows the R3 waves at COP, Sweden, and NUR, Finland, and the R2 waves at AFI, Samoa. For these waves we took the spectral amplitude ratios of the 1973 event to the 1975 event, and the results are given in Figure 9. The amplitude ratio at the long-period end \( (\sim 250 \text{ s}) \) is about 8.5 with a standard deviation of about 2. This amplitude ratio is significantly large compared with the observed tsunami height ratio even if various uncertainties of the measurements and scatter of the data are taken into account. The 1975 event is thus typically a tsunami earthquake. The spectral ratio curve is more or less flat in the long-period range, and there is no tendency for the ratio to decrease with increasing period. If the process time of the 1975 event were much longer than that of the 1973 event but still not much longer than the observed period range of surface waves, the spectral ratio should increase with decreasing period. We may conclude, following such discussion as in the last section, that the tsunamis and the long-period seismic waves of the 1975 event were generated by a source with a process time not longer than that of the 1973 event. Process time alone again cannot explain why the relative tsunami efficiency was so different between the 1973 and the 1975 event. Takemura et al. [1977] obtained a process time of 50 \( s \) for the 1975 event from the amplitudes of seismic waves having periods of 30–200 \( s \). It is apparent that this process time is not long enough to explain the above difference in tsunami efficiency.

### Seismic Wave Characteristics

A process time \( \tau \) is roughly expressed by

\[
\tau = \frac{(L + W/2)}{V}
\]  
(1)

where \( L \) is fault length, \( W \) fault width, and \( V \) rupture velocity [Aki, 1967; Savage, 1972]. Substitution of \( L \sim 250 \text{ km} \), \( W \sim 120 \text{ km} \) (Figure 2), and \( V \sim 3.5 \text{ km/s} \) [Kanamori, 1970a; Ben-Menahem and Rosenman, 1972] for the mainshock of the great Kurile earthquake of 1963 yields \( \tau \sim 89 \text{ s} \), which is not very different from the process time of 93 \( s \) actually obtained for the mainshock [Furumoto, 1979]. On the other hand, if we put \( L \sim 100 \text{ km} \) and \( W \sim 60 \text{ km} \) (Figure 2) and assume \( V \sim 3.5 \)
km/s for the largest aftershock (= tsunami earthquake), we obtain $r \sim 37$ s, which seems to be significantly smaller than the value of 62 s actually obtained [Furumoto, 1979]. This indicates a relatively slow process at the source of the tsunami earthquake. To be compatible with the observed value of $r$ a rupture velocity of about 2.0 km/s may be necessary. A slow rupture process is also indicated from the $P$ wave seismograms. Figure 10 compares the long-period $P$ wave seismogram of the aftershock to that of the foreshock at station GDH, Greenland. The waveform of the aftershock is very complicated, showing a multiple-event nature of the source (see also Figures 15 and 16). Although its maximum trace amplitude is more or less the same as that of the foreshock, the total duration time is vastly different. It is almost 100-120 s for the aftershock, while it is only 40-50 s for the foreshock. Substitution of $L \sim 60$ km, $W \sim 40$ km (Figure 2), and $V \sim 3.5$ km/s into (1) gives $r \sim 23$ s for the foreshock, which is not inconsistent with the observed $P$ wave duration. On the other hand, the $P$ wave duration of the aftershock indicates a source duration of at least 60 to at most 100 s, suggesting again a relatively slow rupture process.

For a slow rupture process we may expect a deficiency of high-frequency seismic waves. The low-frequency nature of the 1963 aftershock is in fact very apparent from a comparison of the seismograms of the foreshock to those of the aftershock. Figure 11 shows the seismograms recorded by three types of instruments having different passbands at the network of the California Institute of Technology (CIT). On the Press-Ewing seismograms (pendulum period of 30 s, galvanometer period of 90 s; abbreviated as 30-90) the trace amplitude of the long-period $R_2$ waves of the aftershock is several times as large as that of the foreshock. We have also examined several $G_2$ and $R_2$ seismograms of the WWSSN stations (Worldwide Standardized Seismograph Network). The average trace amplitude ratio of the aftershock to the foreshock is about 3 for the waves with the predominant periods of 100-200 s. On the other hand, the long-period Benioff seismograms (1-90) at the CIT network show that the 20-s Rayleigh wave amplitudes of the aftershock are only slightly larger than those of the foreshock (Figure 11). Excitation of short-period $P$ waves from the aftershock is even weaker than that from the foreshock as indicated by the short-period seismograms (1-0.2) of Figure 11. A weaker excitation of short-period $P$ waves is more apparent in Figure 12. This figure displays the $P$ wave seismograms recorded at station MTJ, Japan. The instruments are the wide band (1-20), the short-period (1-1), and the shortest-period (1-0.2) HES seismographs. It becomes more and more obvious on the shorter-period seismograms that the high-frequency components are deficient in the seismic waves from the aftershock. Of course, Figures 11 and 12 alone could be inter-

Fig. 8. Long-period $R_3$ seismograms at COP, Sweden, and NUR, Finland, and $R_2$ seismograms at AFI, Samoa, for the Nemuro-Oki earthquakes of 1973 and 1975.

Fig. 10. WWSSN long-period $P$ wave seismograms for the largest foreshock (October 12) and the largest aftershock (October 20) of the great Kurile earthquake of 1963 and for the Nemuro-Oki earthquake of June 10, 1975. Peak magnification, pendulum period, and galvanometer period are given for each station.
interpreted as an indication for a high-frequency nature of the foreshock rather than a low-frequency nature of the aftershock. Figure 13 shows, however, that the foreshock is a 'normal' earthquake. In this figure the seismic intensities measured at JMA (Japan Meteorological Agency) observatories are plotted for the foreshock, the aftershock, and the shock of January 21, 1976 (44.9°N, 149.1°E, h = 41 km). The first and third events have approximately the same hypocenter and the same aftershock area. The surface wave magnitude is also approximately the same, $M_s = 6.3/4$-7(PAS), 7(BKS), 63/4-7(PAL) for the first event, and $M_s = 6.4$-6.5(PAS), 6.7(BKS), 7.0(USSGS) for the third event. The magnitude of the second event is slightly larger, $M_s = 63/4-7$ (PAS), 71/4-71/2(PAL).

Seismic intensity $I$ gives a measure of the relative amplitudes of short-period waves at periods of around 1 s. The number of stations which report $I = 2$ are 3 for the third event, 1 for the first event, and none for the second event. The number of stations reporting $I = 1$ are 5 for the third event, 4 for the first event, and 2 for the second event. As far as the seismic intensities are concerned, there is thus no indication that the first event is a 'high-frequency' earthquake compared with the third event. The intensity map rather indicates a low-frequency nature of the second event. The low-frequency nature is thus demonstrated for the 1963 tsunami earthquake from the 20-s surface wave records, the short-period body wave seismograms, and the seismic intensity map.

The long-period $P$ wave seismograms of the 1975 tsunami earthquake are very complicated, and their durations are unusually long (~2 min), just as in the case of the 1963 tsunami earthquake (see Figure 10). The low-frequency nature of the seismic waves has already been discussed by many seismologists from 20-s Rayleigh waves, long- and short-period $P$ waves, and seismic intensities [see Kanamori, 1977]. By and large, the 1975 event seems to be a type of tsunami earthquake very similar to the 1963 event.

**Mechanism Change**

Another feature common to both the 1963 and the 1975 tsunami earthquake is a possible occurrence of focal mechanism change during the rupture process. The body wave radiation of the 1963 event has been studied by Stauder and Mualchin [1976]. A steeply dipping nodal plane can be determined unambiguously from the long-period $P$ wave seismograms. The trend and the plunge of the pole of this plane are 311° and 3°, respectively [Stauder and Mualchin, 1976]. Such an extremely shallow plunge has been confirmed by Takemura and Suzuki [1977, also personal communication, 1978], who discussed it in conjunction with a very shallow and seaward origin of the source. The second nodal plane cannot be determined accurately from the body wave data alone. An almost pure dip slip, however, is required from the surface wave radiation patterns for which one nodal line nearly coincides with the strike of the steeply dipping nodal plane [Ben-Menahem and Rosenman, 1972]; see also Figure 14. A reasonable mechanism is, then, a pure dip slip fault with a strike of N45°E. If we take the gently dipping nodal plane as the slip plane, the slip direction is almost exactly the same as that of the mainshock [Kanamori, 1970] and consistent with the slip direction expected from the plate tectonics [Le Pichon, 1968].

The mechanism represents a low-angle thrust whose dip can be constrained to 3° from the first motion data.

The dip angle of 3°, however, cannot be reconciled with the surface wave radiation pattern. Figure 14 shows the amplitude radiation pattern of Love waves at a period of 200 s [Ben-Menahem and Rosenman, 1972]. The pattern exhibits an almost complete node in the strike direction of the fault. On the other hand, the theoretical radiation pattern with a dip angle of 3° possesses in this direction an amplitude only 60% as small as that in the loop direction. To be consistent with the observation a dip angle of more than 10° is necessary (Figure 14). A similar conclusion can also be derived from the azimuthal pattern of the initial phase of long-period Rayleigh waves [Furumoto, 1979], yet the first motion data constrain the dip angle of the fault to less than 3°. We interpret this discrepancy in terms of a focal mechanism change during the rupture. What can be determined from the first motion data is a mechanism at the beginning of the rupture. The radiation patterns of long-period surface waves are, on the other hand, primarily controlled by an average mechanism during the rupture process.

A mechanism change is also suggested from the long-period $P$ wave seismograms. In Figure 15, station RIV, Australia, is fairly close to a $P$ wave nodal plane for both the mainshock and the aftershock. Station COP, Denmark, is far from the nodal plane. The $P$ wave amplitudes of the mainshock at RIV are, for this reason, considerably smaller than those at COP. In the case of the aftershock, on the other hand, relatively small amplitudes are seen at RIV only for the first 30 s. The amplitudes in the remaining portion of the $P$ wave seismogram are as large as those at COP. This feature was first discussed by Takemura and Suzuki [1977] and appears to be most simply explained by a focal mechanism change. Note that the direction of the first motion is opposite at RIV for the mainshock and the aftershock, manifesting a difference in the initial mechanism between the two events. The dip angle of the initial fault is 22° for the mainshock [Kanamori, 1970a] and 3° for the aftershock. If a possibility of a mechanism change from thrust to normal faulting is precluded, a change in dip angle of at least 30° may be necessary to explain the observation at RIV, where the maximum amplitude is comparable with that at the edge of the Plateau.
COP. A change of 20° makes the focal mechanism very similar to the mechanism of the mainshock, which still gives a large amplitude difference between COP and RIV. Figure 16 shows the $P$ wave seismograms belonging to the dilatational quadrant of the initial mechanism and to the compressional quadrant. For the first 20 s the polarity is opposite each other in the two quadrants, but after the first 30 s the polarity is apparently the same among all the stations. This indicates that all the stations shown in Figure 16 entered into the same quadrant presumably 20–30 s after the initial rupture as a consequence of mechanism change. A 45° thrust fault may explain the polarity of these later waves. A strikingly similar $P$ wave behavior can be seen for the 1975 tsunami earthquake in Figure 17 of Takemura et al. [1977], indicating again its considerable mechanism change.

Along the above line of evidence we suggest a substantial steepening of the thrust plane for both the 1963 and the 1975 tsunami earthquake, although the evidence may not be sufficient to warrant a quantitative discussion. The locations, the focal depths, and the initial mechanisms indicate that these tsunami earthquakes originated at the lithospheric plate boundary near the deep-sea trench. The subsequent steepening

![Image of seismograms](image)

Fig. 12. $P$ wave seismograms at MTJ, Japan, for the largest foreshock and the largest aftershock of the great Kurile earthquake of 1963. Letters U (up) and D (down) roughly mark the arrival of the $P$ waves.
Fig. 13. Seismic intensity map in JMA scale for the largest foreshock and the largest aftershock of the great Kurile earthquake of 1963 and for the shock of January 21, 1976. The aftershock areas of the first and third events overlap almost completely. Stations which report the zero intensity for all the events are not shown.

of the thrust plane therefore may be explained by either an upward branching of rupture from the lithospheric boundary toward the surface or a downward branching of rupture into the interior of the descending slab. The first alternative is apparently a more likely process. If this is the case, the upward branching took place through a wedgelike region at the leading edge of the continental lithosphere (see Figure 19).

Fig. 14. Amplitude radiation pattern of Love waves at a period of 200 s for the 1963 tsunami earthquake [Ben-Menahem and Rosenman, 1972]. Theoretical radiation patterns for pure reverse faults with dip angles of 3° and 10° are also shown.

Fig. 15. Comparison of P wave seismograms between the 1963 Kurile earthquake and its largest aftershock (= tsunami earthquake). Station COP is far from the nodal plane, and RIV is close to it. The NS components are shown for the mainshock because the vertical components are off scale. Note that the radial component rotated from the NS component has a different unit amplitude at COP and RIV.

Fig. 16. P wave seismograms belonging to the dilatational quadrants of the initial mechanism (GUA, RAB, and RIV) and to the compressional quadrants (COP and KEV). The polarity is the same after the first 30 s for all the stations.

Fig. 17 shows a structure of this wedgelike region across the central Aleutian arc [Grow, 1973]. It consists largely of thick deformed sediments [Grow, 1973]. This sedimentary wedge has a width of about 80 km in a direction perpendicular to the trench axis and an average thickness of about 8 km where an average P velocity is Vp ~ 3.2 km/s and an average density is ρ ~ 2.2 g/cm³. The presence of a similar sedimentary wedge has been indicated in many island arcs. For example, the sediment thickness beneath the inner wall of the deep-sea trench has been reported to be about 11 km (Vp ~ 3.7 km/s) in western Japan off Shikoku [Yoshii et al., 1970], 11 km (Vp ~ 3.7 km/s) in Java off Sumatra [Hamilton, 1977], and 20 km (Vp ~ 3.9 km/s, ρ ~ 2.5 g/cm³) in Lesser Antilles near Barbados [Westbrook et al., 1973], and 5 km (Vp ~ 2.8 km/s) in southern Kurile [Weizman, 1966]. The thickness of 5 km in southern Kurile is the one for the 2.8 km/s layer directly below which the 6.6 km/s layer appears in the related velocity profile [Weizman, 1966]. This direct appearance of the 6.6 km/s layer is perhaps a spurious result of inappropriate shot distances [Hotta, 1972], and possible intervening layers may be still sedimentary.

We thus conclude that the wedgelike region ruptured by a tsunami earthquake consists largely of thick sediments. These sediments with low elastic wave velocities and densities are indicative of a not only elastically but also inelastically very deformable nature of the wedge, which has been firmly established by deep penetration seismic reflection data and deep-sea drilling data [Seely et al., 1974; Karig and Sharman, 1975]. Inelastic deformation occurs in a form of complex folding and...
imbricate faulting. Imbricate thrusts are landward dipping and concave upward as if they follow stress trajectories due to downgoing motion of the underlying oceanic lithosphere [Seely, 1977]. Displacements along these thrusts may be so significant as to be a dominant mode of subduction near the trench [Seely et al., 1974; Karig and Sharman, 1975]. Substantial internal deformation of the wedge portion has also been suggested from geologic studies of the Franciscan mélangé, which has been interpreted as a fossilized Benioff zone [e.g., Hamilton, 1969]. Hsu [1971] postulated that the Cretaceous subduction of the Franciscan plate did not take place at shallow depths as thrusting along a single lithospheric interface but as penetrative shear over a 10-km-thick Franciscan package through which numerous shear surfaces developed. Because of this penetrative shear, brittle layers of the Franciscan graywackes were broken along extensional shear fractures associated with the ductile deformation of the shaly matrix. Ernst [1970] presented a similar model of the Franciscan subduction.

Seismic activities near deep-sea trenches seem to be consistent with the above mode of deformation. Figure 18 shows the cross sections of seismic activity on profiles across the Gulf of Alaska (section (a)), the eastern Aleutian arc (section (b)) [Jacob et al., 1977], the central Aleutian arc (section (c)) [Engdahl, 1977] and the Kamchatka-Kurile arc (sections (d) and (e)) [Isacks and Barazangi, 1977]. In cross section (a), Middleton Island is located about 60 km landward of the trench axis, where a very low seismic activity is apparent. On this island, as was described in a previous section, a substantial postseismic deformation has been observed after the great Alaskan earthquake. In cross section (b) the Aleutian tsunami earthquake of April 1, 1946, has been located about 40 km landward of the trench axis [Sykes, 1971], where the seismic activity is again low. This earthquake produced one of the greatest tsunamis of this century, although the tsunami-generating mechanism may not be the same as that for the 1963 and 1975 tsunami earthquakes [Kanamori, 1972]. In cross section (c), across the central Aleutian arc there is again an aseismic region at the leading edge of the continental lithosphere for which a structural cross section is shown in Figure 17. This aseismic region more or less corresponds to the sedimentary wedge. Presumably this sedimentary wedge is inelastically too deformable to sustain stress sufficient to produce stationary seismic activity. The presence of the aseismic wedge is, in this interpretation, an indication for active deformation rather than stableness. As seen in Figures 18d and 18e, the Kamchatka-Kurile arc also possesses an aseismic wedge at the leading edge of the continental lithosphere, in which the 1963 and 1975 tsunami earthquakes were generated.

**MECHANISM OF TSUNAMI EARTHQUAKES**

Fracturing through the thick sediments produces large displacement in the source region but relatively small displacement at far fields. For example, if we assume a P velocity of 4 km/s and a density of 2.5 g/cm³ in a focal region of a tsunami earthquake whose significant portion is sedimentary, we may expect an average rigidity of about $1.3 \times 10^{11}$ dyne/cm². This low rigidity is markedly contrasted with a rigidity of about $5 \times 10^{11}$ dyne/cm² ($V_p = 7$ km/s, $\rho = 3$ g/cm³) in a farther landward region where ordinary thrust earthquakes occur. Because of such a rigidity contrast the source displacement of a tsunami earthquake may be, in this case, about 4 times as large as that of an ordinary thrust earthquake when the far field displacements are the same. Moreover, the source displacement attains a substantial amount of vertical component at the sea bottom through the upward branching process of rupture so as to generate tsunamis more effectively. The upward branching process thus contributes to extensive tsu-
The above mechanism of a tsunami earthquake is quite different from a mechanism proposed by Kanamori [1972], who attributed its extensive tsunami to an unusually long process time. There is no doubt that a time constant is a controlling factor for some tsunami earthquakes. What we propose here, then, is the presence of another type of tsunami earthquake which is inherent to the subduction process near the trench. Our interpretation assumes a low rigidity in the source region which can be checked against a detailed study of the deep structure beneath the inner wall of a trench.

TECTONIC IMPLICATIONS

Near a deep-sea trench, active deformation takes place in a zone between the trench axis and the outer ridge of the deep-sea terrace, and this zone, corresponding to the lower continental slope, defines the surface expression of the subduction zone [Karig and Sharman, 1975]. In other words, the megathrust between the oceanic and the continental lithosphere does not come to the surface as a simple shear plane, but instead it merges as a complex shear zone across the lower continental slope [Hsü, 1971; Ernst, 1970]. The Franciscan mélangé has been interpreted as a fossil of this complex shear zone [Yonekura, 1975; Ernst, 1970]. In this zone, subsidiary reverse faults branch upward from the main thrust surface, and these imbricate faults become steeper toward the surface [Seely et al., 1974]. We suggest that rapid and intense deformation takes place in this zone either at the time of a great earthquake or shortly after it. If the deformation occurs at the coseismic stage, its effect on the seismic waves and the tsunamis would be indistinguishable from the effect due to a major slip along the lithospheric interface. Alternatively, the intense deformation may take place at the postseismic stage in either a brittle or a ductile way. The brittle mode of post-
seismic deformation causes a tsunami earthquake such as the 1963 and 1975 events. The ductile mode of postseismic deformation produces a tilt movement of an offshore island which has been observed at Middleton Island after the 1964 great Alaskan earthquake [Prescott and Lisowski, 1977].

The process just described corresponds to the stage shown in Figure 19c. The stages shown in Figures 19a and 19b correspond to the interseismic stage and the coseismic stage of a great earthquake sequence. At the interseismic stage (Figure 19a) the continental lithosphere is dragged down with the descending slab of lithosphere, and the stress gradually accumulates within it. The accumulated stress is partially relieved by ductile deformation in a wedge portion near the trench, where two dominant modes of stress relaxation are complex folding and imbricate faulting. The stress relaxation becomes progressively smaller landward away from the trench. The stress eventually reaches a critical value somewhere in a landward portion. Brittle fracture thus results as a great earthquake. If the rupture does not reach the free surface at this stage, the wedge portion at the leading edge of the continental lithosphere is loaded and strained. This corresponds to the coseismic stage shown in Figure 19b.

The above earthquake sequence leaves a permanent deformation on the free surface because thrusting associated with a great earthquake eventually branches upward from the lithospheric interface. The deformation is relative uplift landward of the zone of surface break and relative subsidence oceanward. This zone of surface break geomorphologically corresponds to the lower continental slope. An earthquake sequence thus produces relative uplift at the top and relative subsidence at the foot of the lower continental slope. Such crustal movement seems to be consistent with an actively rising feature of the outer ridge of the deep-sea terrace. It is also consistent with a depressional feature near the trench in the sense that if there were little sediment supply at the trench and if other tectonic factors were ignored, the foot of the lower continental slope would be progressively deepened by great earthquake activities. Anyway, the above consistency implies a causal relationship between great earthquake activities and geomorphological features near trenches. The presumed crustal movement is also consistent with a permanent uplift of coastal terraces that occurs episodically at the time of great earthquakes (K. Nakamura, personal communication, 1977). Southwest Japan, Alaska, and Chile are full of evidence for such an episodic coastal uplift (see for the detailed references Yonekura [1973]). We suggest that a coastal uplift at the time of a great earthquake can be traced back seaward to the outer ridge of the deep-sea terrace but not farther trenchward. Observation of sea bottom deformation before and after a great earthquake is an essential key for checking the above model of the shallowest subduction.

**CONCLUSION**

Many island arcs possess a thick sedimentary wedge at the leading edge of the continental lithosphere. This portion is characterized by low seismic activity which is presumably due to ductile deformation of sediments. The 1963 and 1975 tsunami earthquakes were generated in this wedge portion. We suggest from a study of these earthquakes that the sedimentary wedge plays an important role in great earthquake process and, in turn, that great earthquake activities are causally related to geomorphological features near trenches. We are currently investigating short-period records of more than 600 submarine shocks occurring along the Japan trench (Y. Fukao and K. Kanjo, unpublished manuscript, 1979). The results show that low-frequency earthquakes can be found almost exclusively in a belt of low seismic activity just landward of the Japan trench. This observation again indicates anomalous mechanical properties of a sedimentary wedge.

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**REFERENCES**


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