

SOURCE CHARACTERISTICS OF HYPOTHETICAL SUBDUCTION EARTHQUAKES IN THE NORTHWESTERN UNITED STATES

BY THOMAS H. HEATON AND STEPHEN H. HARTZELL

ABSTRACT

Historic earthquake sequences on subduction zones that are similar to the Cascadia subduction zone are used to hypothesize the nature of shallow subduction earthquakes that might occur in the northwestern United States. Based on systematic comparisons of several physical characteristics, including physiography and seismicity, subduction zones that are deemed most similar to the Cascadia subduction zone are those in southern Chile, southwestern Japan, and Colombia. These zones have all experienced very large earthquake sequences, and if the Cascadia subduction zone is also capable of storing elastic strain energy along its greater than 1000 km length, then earthquakes of very large size ($M_w > 8\frac{1}{2}$) must be considered. Circumstantial evidence is presented that suggests (but does not prove) that large subduction earthquakes along the Cascadia subduction zone may have an average repeat time of 400 to 500 yr.

INTRODUCTION

This is the third in a series of four papers that lead to an estimation of the seismic hazard associated with the subduction of the Juan de Fuca and Gorda plates beneath North America. In the first paper, Heaton and Kanamori (1984) compared physical characteristics of the Cascadia subduction zone (also referred to as the Juan de Fuca subduction zone) with those of other subduction zones and concluded that the Cascadia subduction zone is similar to other subduction zones with strong seismic coupling, and thus may be capable of producing great shallow subduction earthquakes. In the second paper, Hartzell and Heaton (1985) compared the nature of the time history of seismic energy release for 60 of the largest subduction earthquakes to occur in the past 50 yr. In this paper, we extend the results of the previous studies to estimate the nature of shallow subduction earthquakes that could be postulated if the Cascadia subduction zone is assumed to have strong seismic coupling. In the fourth paper, Heaton and Hartzell (1986) estimate the nature of strong ground motions that may result from earthquakes hypothesized in this study.

Key issues addressed here are: (1) the dimensions and geometry of hypothetical Cascadia subduction earthquakes; (2) plausible repeat time; (3) identification of analogous historic earthquakes; and (4) estimation of possible tsunami amplitudes.

Subduction of the Juan de Fuca and Gorda plates has presented earth scientists with a dilemma. Despite compelling evidence of active plate convergence (Riddihough, 1977; Snively *et al.*, 1980; Hyndman and Wiechert, 1983; Adams, 1984a; Nishimura *et al.*, 1985), subduction on the Cascadia zone has often been viewed as a relatively benign tectonic process (Riddihough, 1978; Ando and Balazs, 1979; Acharya, 1981, 1985). There is no deep oceanic trench off the coast; there is no extensive Benioff-Wadati seismicity zone; and most puzzling of all, there have not been any historic low-angle thrust earthquakes between the continental and subducted plates. The two simplest interpretations of these observations are: (1) the Cascadia subduction zone is completely decoupled and subduction is occurring aseismically, or (2) the Cascadia subduction zone is uniformly locked and storing elastic energy to be released in future great earthquakes. Full resolution of this issue

may prove elusive. Although it is somewhat surprising that no shallow subduction earthquakes have been documented in this region, the duration of written history is relatively short. It seems certain that great shallow subduction earthquakes have not occurred in this region since the 1850's and highly probable that they have not occurred since the 1790's. If large shallow subduction earthquakes do occur on the Cascadia subduction zone, we can infer their characteristics only by studying the nature of subduction earthquakes that have occurred on other subduction zones that can be considered as analogous. Unfortunately, no subduction zone is exactly the same as the Cascadia subduction zone, and thus, the search for analogs will always lead to fundamental ambiguities.

Systematic variations in the nature of seismic energy release with physical characteristics of subduction zones have been discussed by Kelleher *et al.* (1974), Kanamori (1977), Uyeda and Kanamori (1979), Ruff and Kanamori (1980), Lay *et al.* (1982), Peterson and Seno (1984), and Uyeda (1984). Many types of correlations have been suggested, but since quantification of the physical characteristics of subduction zones and their earthquakes is often rather subjective, not all of these studies seem to reach compatible conclusions. Perhaps the most consistent observation is that subduction of young lithosphere is associated with strong seismic coupling whereas subduction of old lithosphere is associated with weak seismic coupling. Ruff and Kanamori (1980) demonstrate that the seismic coupling is well-correlated with a simple linear function of subducted plate age and convergence velocity in which decreasing age and increasing convergence velocity imply stronger seismic coupling. Heaton and Kanamori (1984) show that the Cascadia subduction zone is clearly different from the class of aseismic subduction zones that is characterized by the subduction of old lithosphere. They further show that it shares many characteristics with subduction zones that have experienced very large shallow subduction earthquakes. In this report, we do not restate all of the arguments given by Heaton and Kanamori (1984), but instead we extend that study by systematically comparing trench bathymetry and shallow seismicity for a world-wide sampling of subduction zones.

TRENCH BATHYMETRY AND GRAVITY

Uyeda and Kanamori (1979) suggest that strongly coupled subduction zones are accompanied by relatively shallow trenches, whereas weakly coupled subduction zones are accompanied by deep oceanic trenches. Similarly, they conclude that free-air gravity anomalies tend to be larger for those trenches with weak seismic coupling. The Cascadia subduction zone is somewhat unusual in that it has virtually no bathymetric trench. In order to assess just how anomalous this is, we have constructed profiles of bathymetry and free-air gravity for many circum-Pacific convergent boundaries. Figure 1 shows shiptrack segments that were used to construct the profiles shown in Figure 2. The shiptrack data were obtained in digital form from NOAA (1981). All data are plotted on a common scale, and the subducted plate is always on the right-hand side of the profile. Bathymetry and free-air gravity are plotted together. Unfortunately, gravity data were not available for all the shiptracks selected, and thus, only bathymetry is plotted for some of the profiles.

Comparing the profiles from differing subduction zones, we see that the Cascadia subduction zone (profiles 38–43) is indeed remarkable for its lack of a bathymetric trench, very shallow ocean floor, and very small gravity anomalies. However, there are several factors that may help to explain these characteristics. The shallow 3 km depth of the Juan de Fuca plate before subduction is near the average depth of very

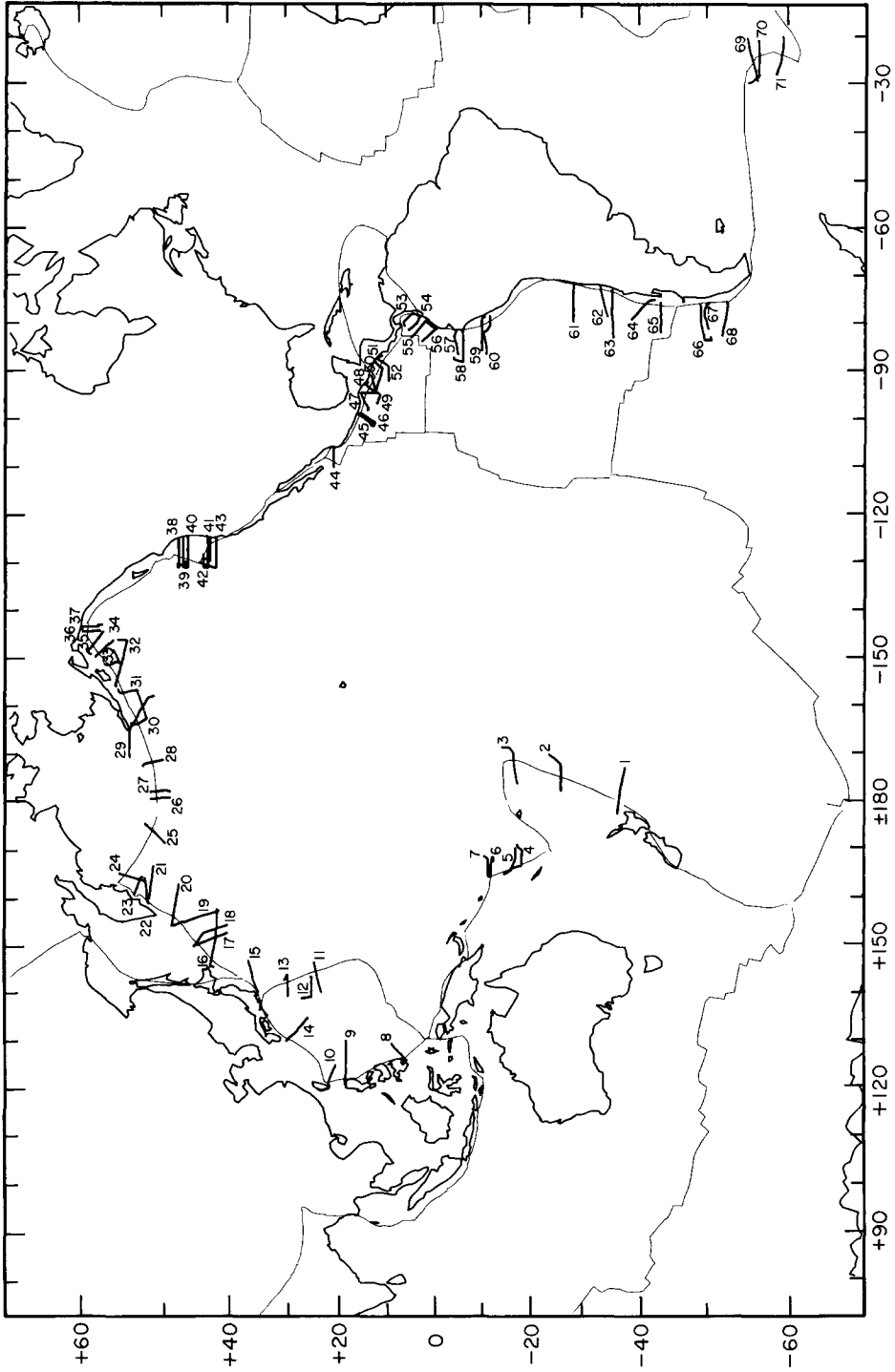


FIG. 1. Shiptracks used to construct profiles of bathymetry and free-air gravity for world-wide subduction zones shown in Figure 2.

young oceanic lithosphere (Parsons and Sclater, 1977). Furthermore, reflection profiles (Snively *et al.*, 1980; Kulm *et al.*, 1984) reveal the presence of a shallow trench that has been completely inundated by approximately 2 km of sediments. Grellet and Dubois (1982) show that there is a good correlation between the age of subducted plate and both the absolute trench depth and the trench depth relative

a

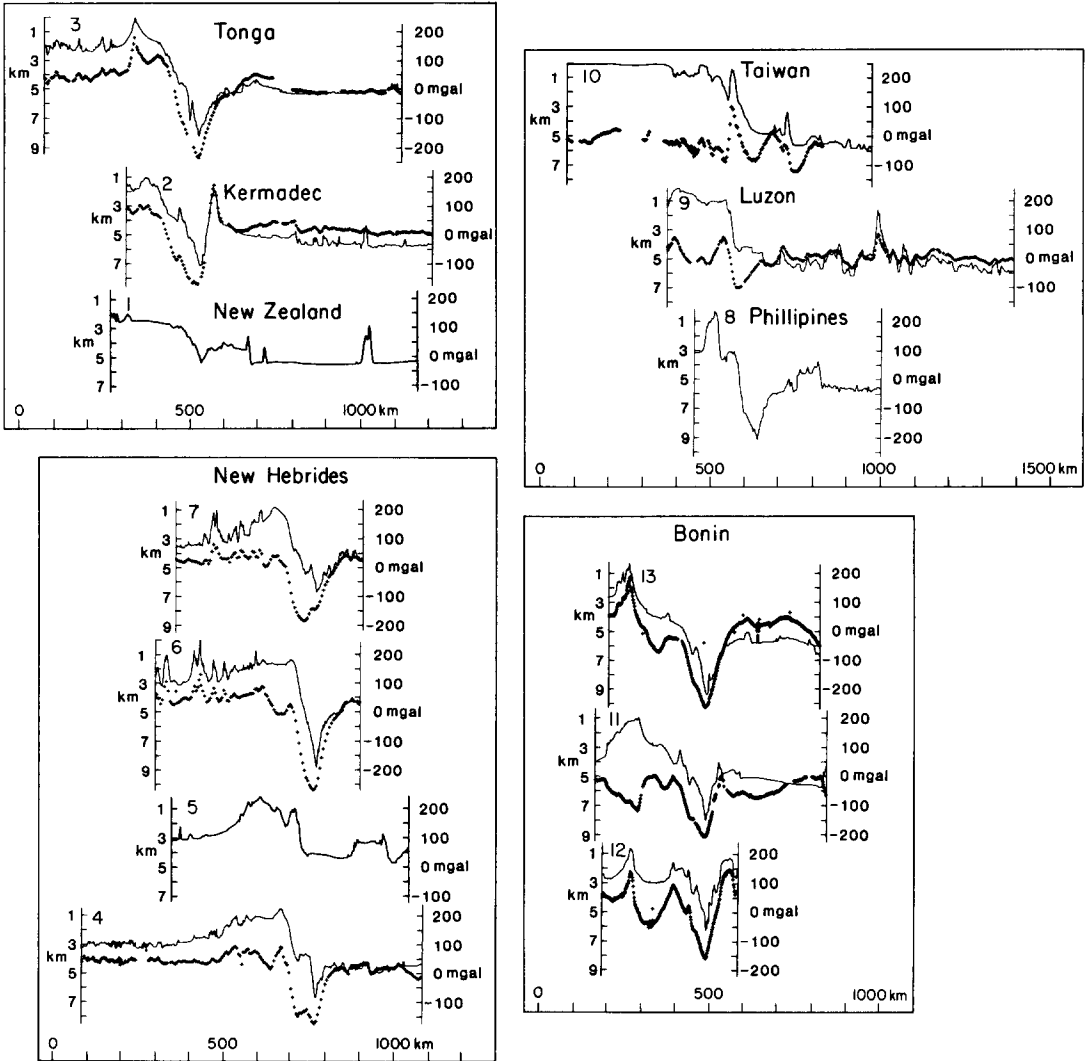


FIG. 2. (a) to (g) Profiles of bathymetry (light solid line) and free-air gravity (dotted heavy line) for shiptracks shown in Figure 1. All profiles are plotted on a common scale and are aligned approximately perpendicular to the trench axes. The subducting oceanic plate is always on the *right-hand* side of each profile. The number in the *upper left* corner of each profile corresponds to the shiptrack numbers in Figure 1.

to the adjacent abyssal plain. Once again, the very shallow depth of the Cascadia trench is consistent with the subduction of very young oceanic lithosphere.

Searching through the profiles in Figure 2, we see that there are other subduction zones that have features similar to those observed for the Cascadia subduction zone. New Zealand, Nankai Trough, Alaska, Colombia, and southern Chile are all notable

for the absence of a well-developed bathymetric trench. Hilde (1984) reports extensive trench sediments for all of these areas, and with the possible exception of New Zealand, these trenches have experienced very large shallow subduction earthquakes. We particularly note striking similarities in both bathymetry and gravity profiles between the Cascadia subduction zone, Colombia, and southern Chile.

SEISMICITY

One of the most striking features of the Cascadia subduction zone is the remarkable paucity of shallow earthquake activity between the trench axis and the coastal

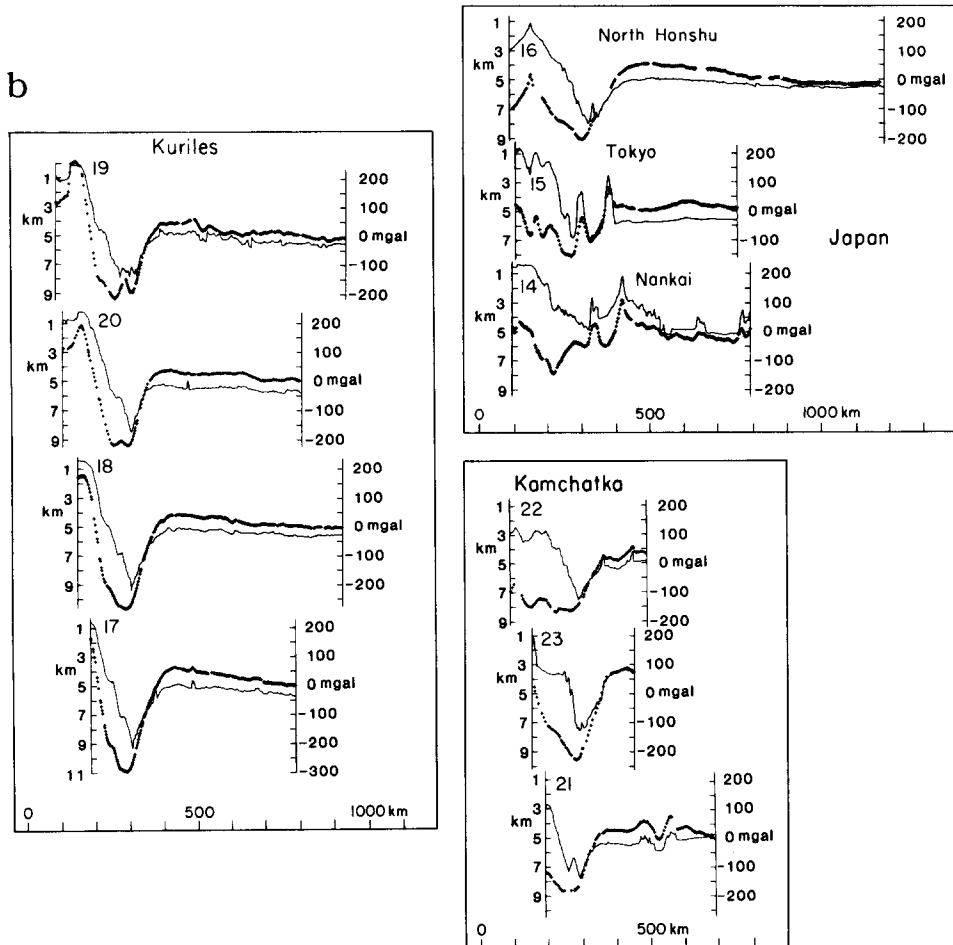
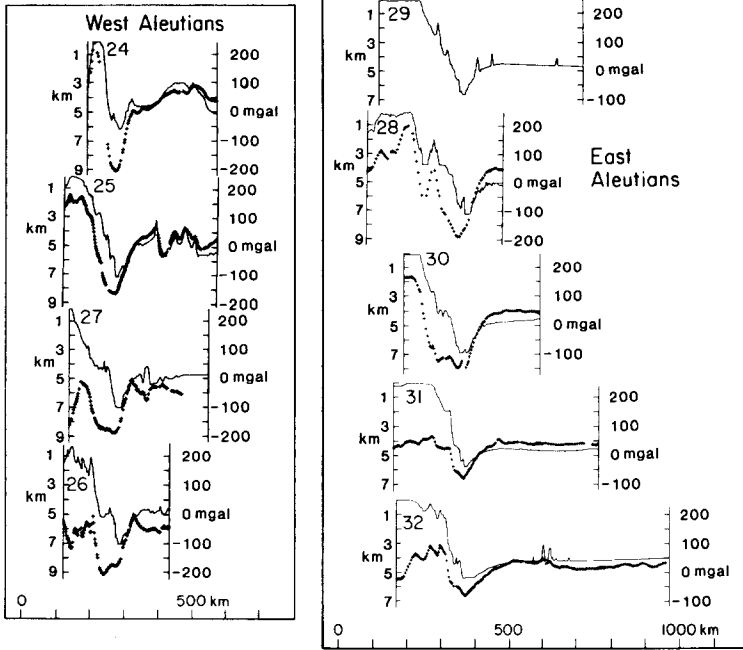


FIG. 2. Continued.

mountain ranges (Taber and Smith, 1985). Despite the continuous monitoring of this region with a local seismometer network since 1970, very few earthquakes of any size have been observed in the coastal regions of Washington and Oregon. This seems somewhat puzzling for a major plate boundary that is undergoing 3 to 4 cm/yr of convergence. Heaton and Kanamori (1984) suggest that this behavior may be analogous to that observed on the San Andreas fault in California. That is, while the central section of the San Andreas fault, which appears to slip principally through aseismic creep, experiences numerous small earthquakes, the sections of

C



d

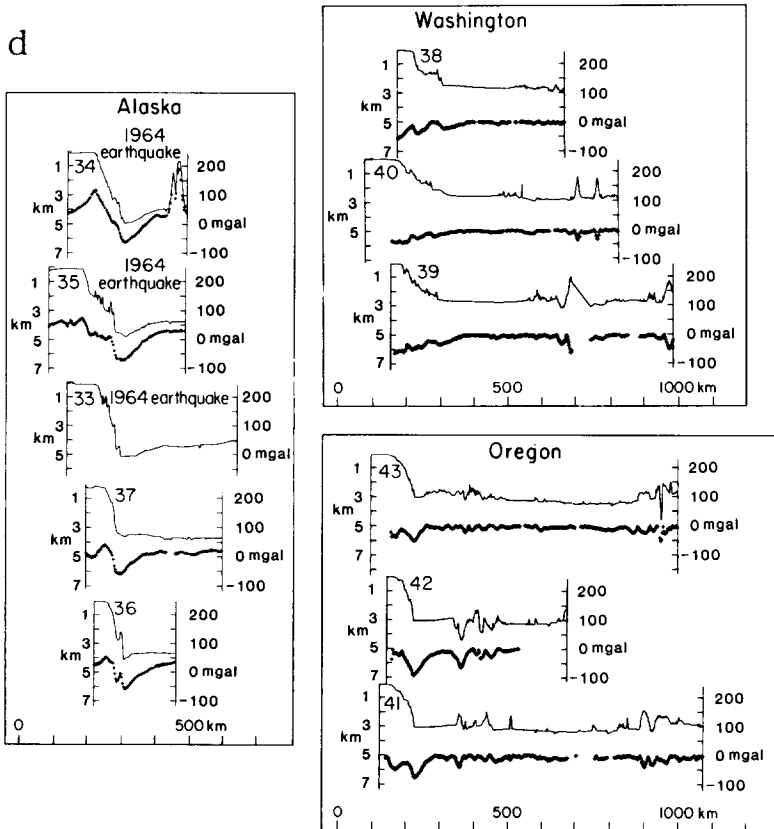
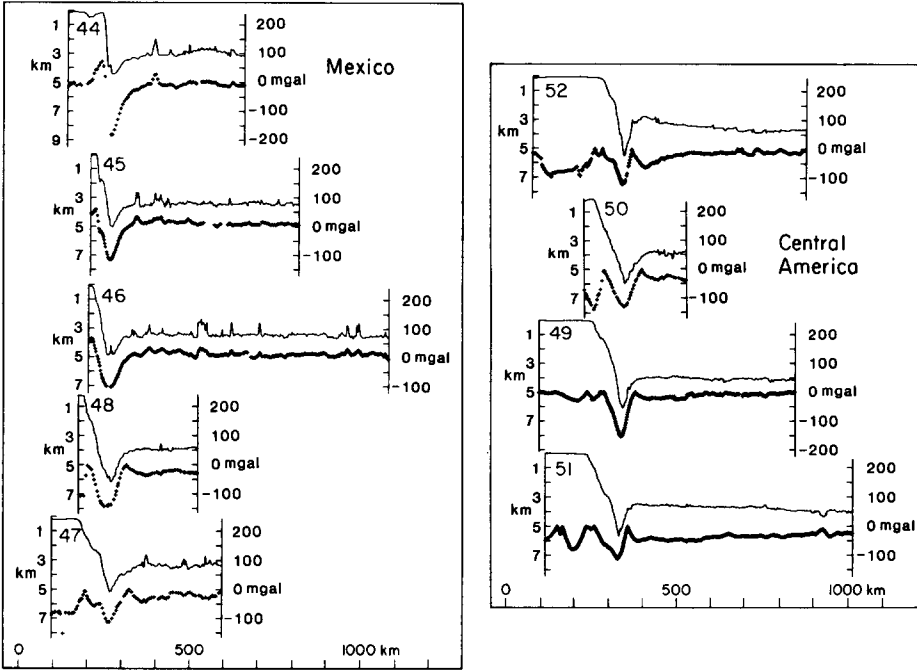


FIG. 2. Continued.

e



f

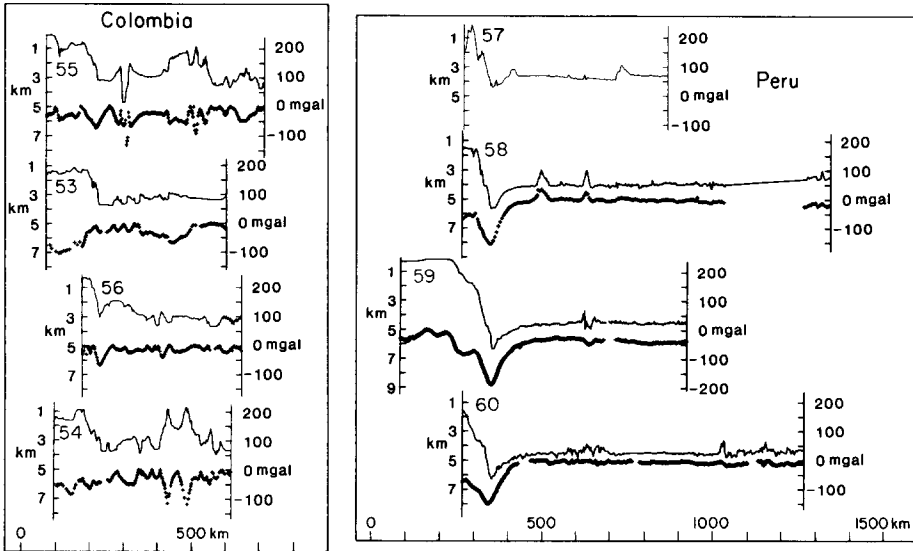


FIG. 2. Continued.

the fault that experienced great earthquakes in 1857 and 1906 are almost totally devoid of present-day seismicity. One hypothesis is that stress increases smoothly and uniformly on fault zones that are strongly coupled over large areas, whereas numerous local stress concentrations occur on faults having large areas that undergo aseismic slip. Although this model may help to explain the paucity of shallow

activity along the Cascadia subduction zone, it has not yet been demonstrated that it generally applies to subduction zones.

Unfortunately, catalogs of local seismicity are not available for many subduction zones that are thought to be strongly coupled. However, there is remarkably little shallow seismicity in the Nankai Trough (Takagi, 1982), a well-instrumented subduction zone having many similarities to the Cascadia subduction zone. In particular, seismicity is very low in the region offshore of the Tokai district (Japan Meteorological Agency, 1984), an area that is often considered as the potential source of a large, shallow subduction earthquake.

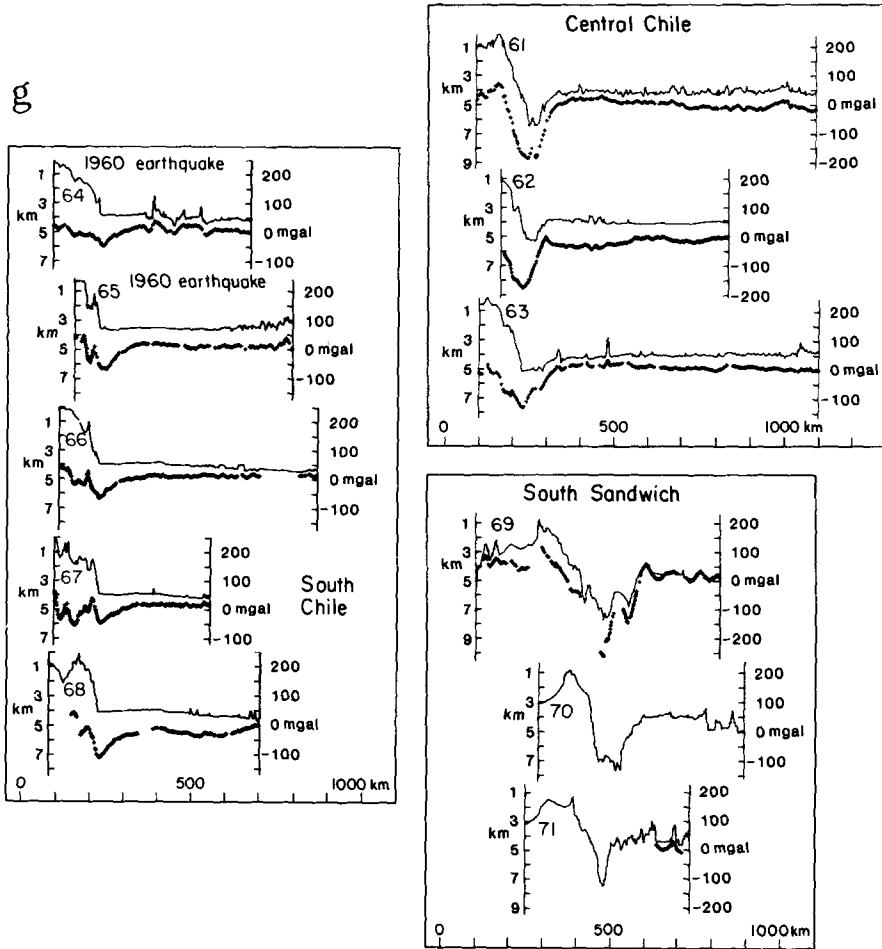


FIG. 2. Continued.

Lacking local seismicity catalogs for most subduction zones, we have systematically compared shallow seismicity from different subduction zones as reported by the NOAA catalog. We wish to identify those zones that have experienced significant time periods of very low seismicity. To this end, we have selected earthquakes with hypocentral depths of less than 50 km from the regions shown in Figure 3. We then plotted the cumulative number of earthquakes in each region as a function of time for the period from 1930 to 1979 (Figure 4). The number of earthquakes in each

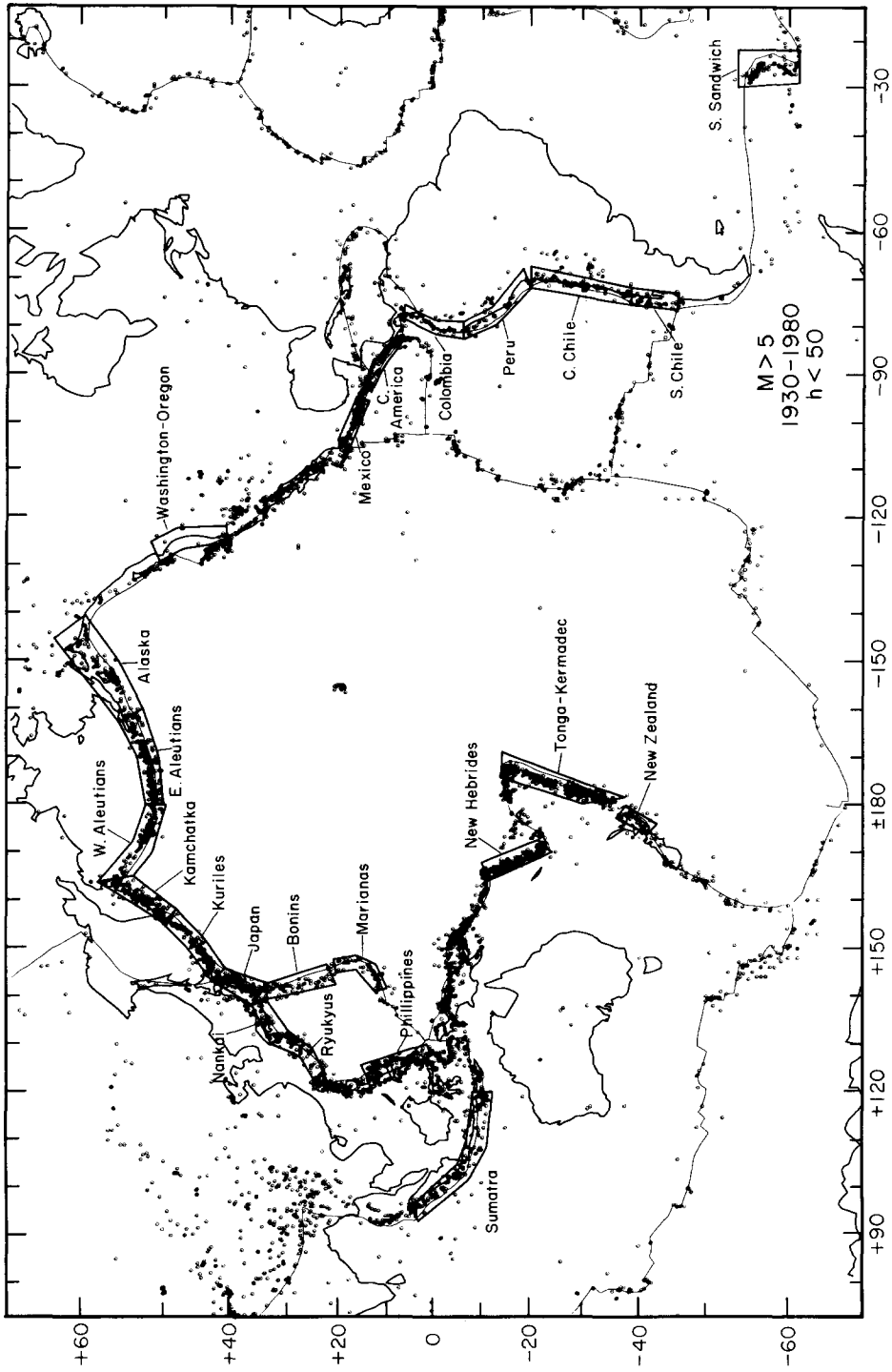


Fig. 3. World-wide map of shallow (depth < 50 km) earthquakes of $M > 5$ taken from the NOAA catalog for the period from 1930 to 1980. Boxes outline the regions for which cumulative seismicity plots (shown in Figure 4) were constructed.

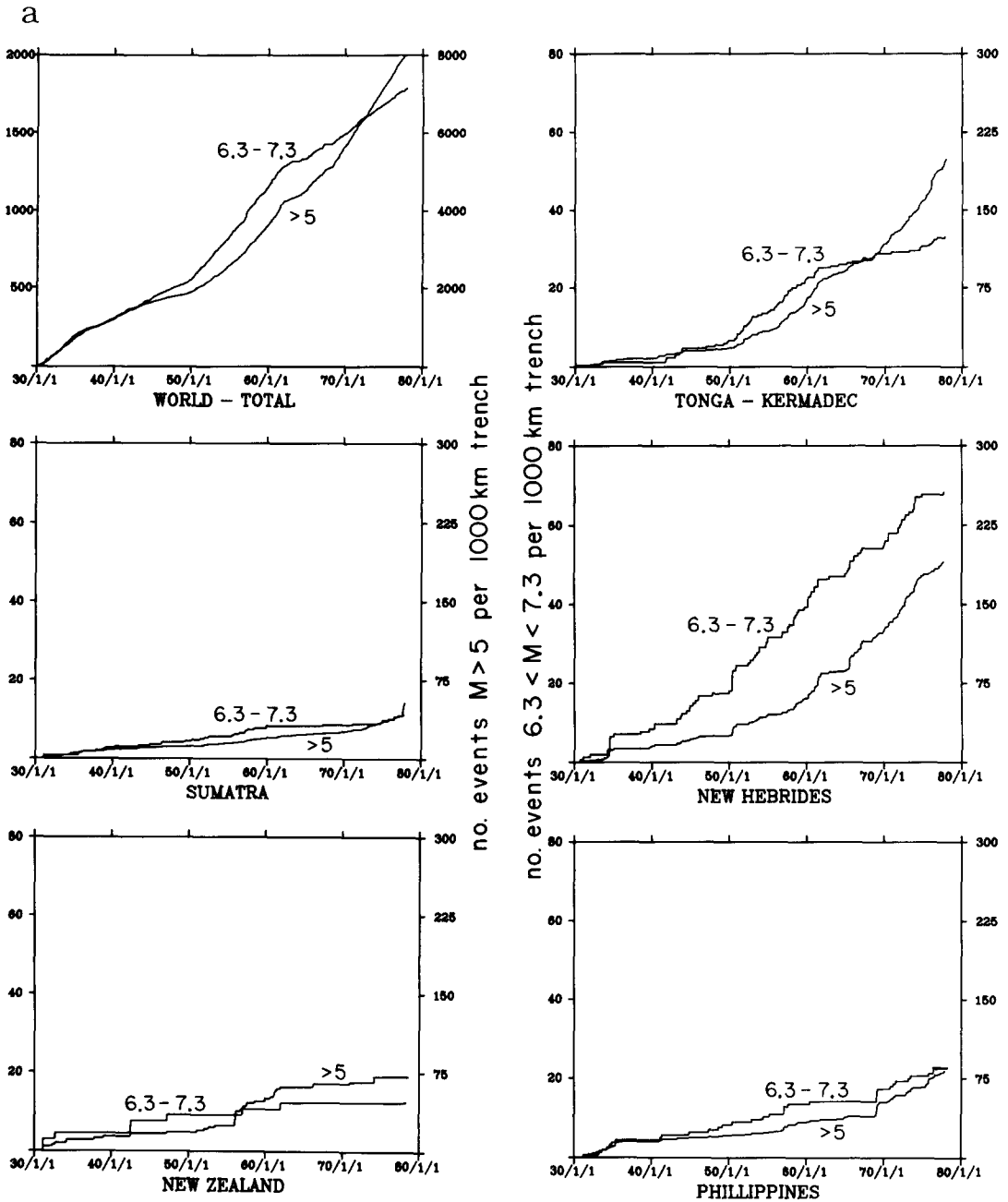


FIG. 4. (a) to (d) Cumulative number of earthquakes shown as a function of time for the regions defined in Figure 3. Both the cumulative number of earthquakes above $M 5$ (right-hand scale) and the cumulative number of earthquakes between $M 6.3$ and $M 7.3$ (left-hand scale) are plotted for the period from 1930 to 1980. The NOAA catalog is probably incomplete for $M > 5$ earthquakes, but should be complete for $M 6.3$ to $M 7.3$ earthquakes. Plots have been scaled for the length of the trench section used, but have not been scaled for convergence rates.

region has been normalized by the trench length, but has not been normalized by convergence rate. Unfortunately, there are almost certainly important spatial and temporal variations in the completeness of the NOAA catalog. Thus, we have selected two magnitude windows: (1) all events between $M 6.3$ and $M 7.3$ (amplitude

scale on the left axis of each plot), and (2) all events larger than M 5 (amplitude scale on the right axis of each plot). We believe that the catalog detection is approximately homogeneous with respect to differing regions for earthquakes larger

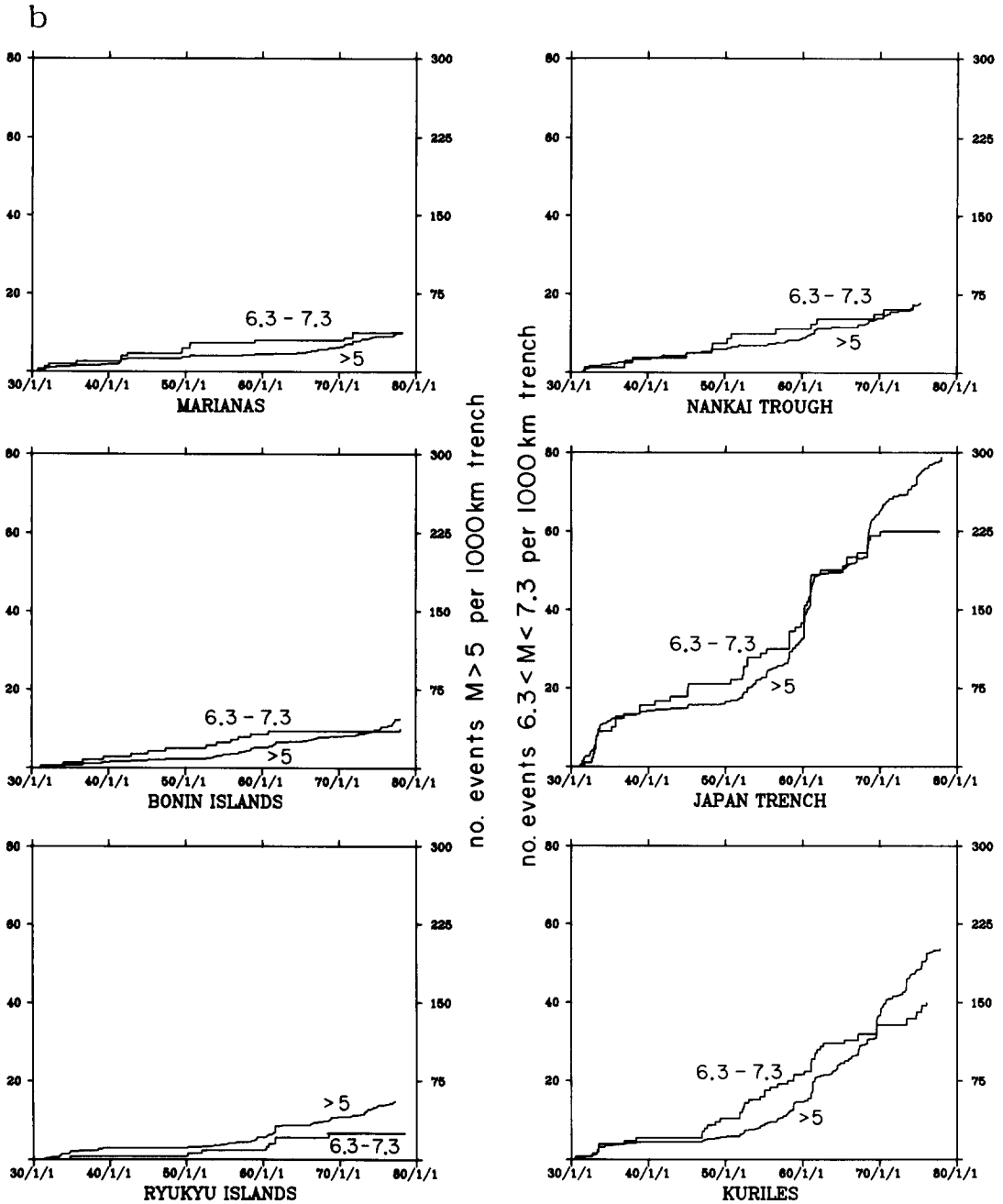


FIG. 4. Continued.

than M 6.3. However, the fact that there are significant temporal variations in the world-wide rate of occurrence of M 6.3 to M 7.3 earthquakes suggests that there may be systematic temporal variations in the calculation of magnitude. Although

the completeness of the NOAA catalog for M 5 earthquakes almost certainly varies with time and region, it is useful to recognize that these earthquakes have occurred in some regions, even if completeness problems will not allow us to prove that they

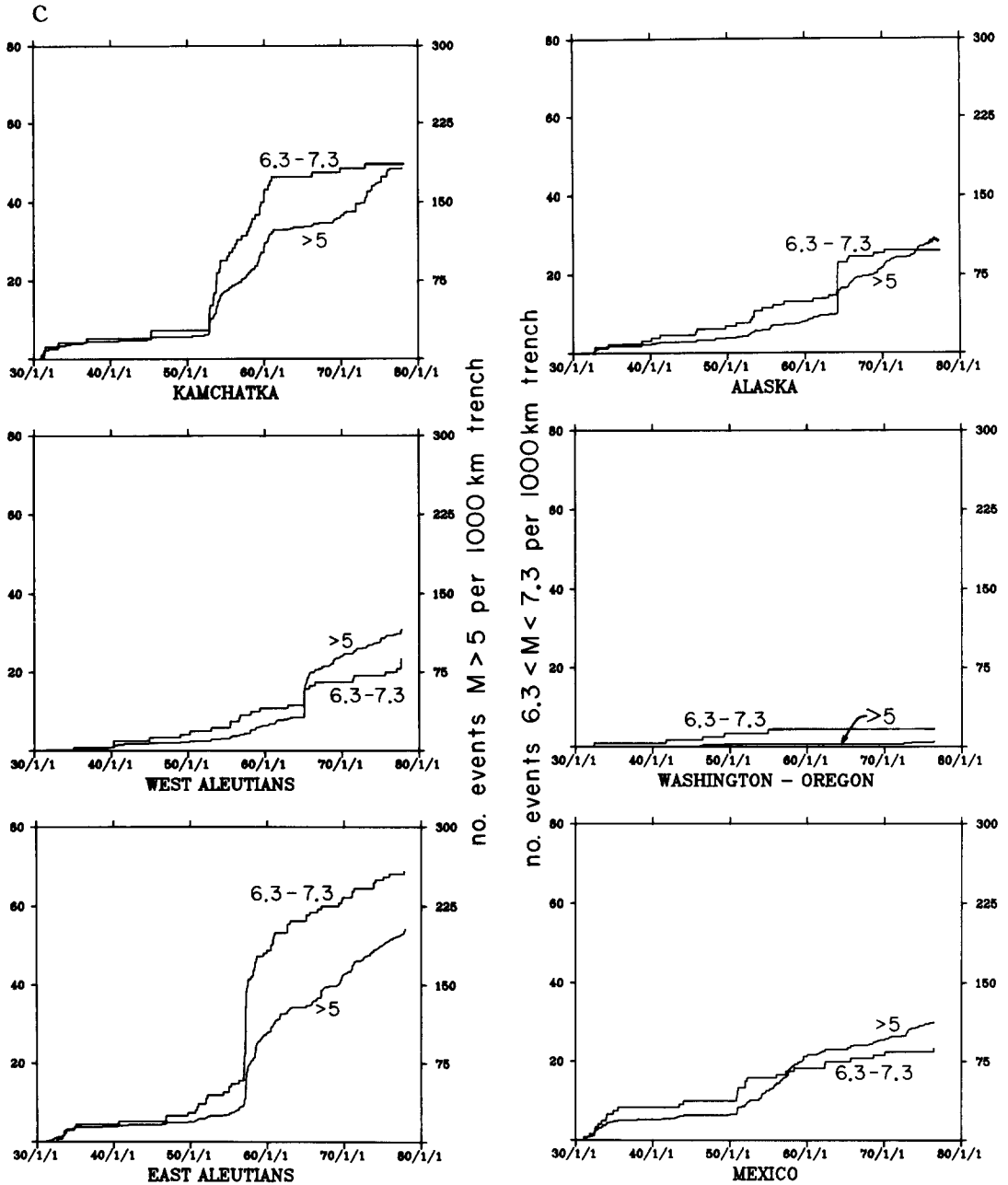


FIG. 4. Continued.

have not occurred in other regions. That is, regions for which there are numerous M 5 earthquakes reported must certainly have a higher level of seismicity than the Cascadia subduction zone, whereas those regions for which few M 5 earthquakes

are reported may or may not have seismicity levels comparable to the Cascadia subduction zone.

A careful study of Figure 4 reveals that coastal Washington and Oregon are

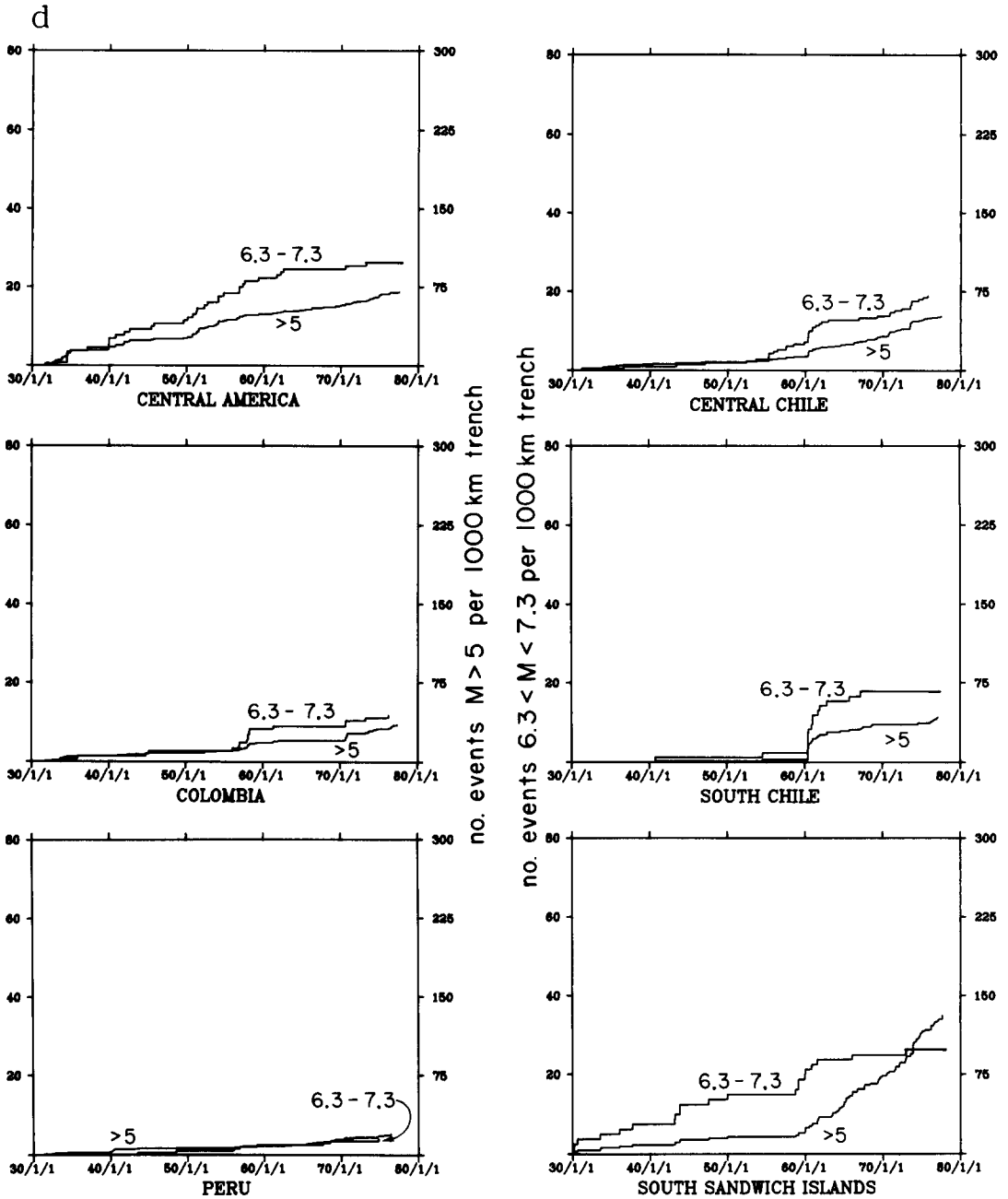


FIG. 4. Continued.

indeed remarkable for their very low levels of earthquake activity. In particular, the Marianas and Bonin Island regions, which are representative of the aseismic subduction of very old lithosphere, have experienced low, but continuous, seismicity.

Regions experiencing high levels of seismicity (Tonga-Kermadec, New Hebrides, Japan Trench, Kurils, and Aleutians) can all be classified as moderately to strongly coupled. Despite high convergence rates and the occurrence of very large earthquakes, South American subduction zones have a relatively low seismicity level. The seismicity in southern Chile prior to the 1960 M_w 9.5 earthquake is particularly noteworthy. Most of the earthquakes shown for this region are aftershocks of the 1960 earthquake, and Duda (1963) reports that earthquake activity was very slight for the region south of 37°S latitude for a period of at least 17 yr before the earthquake. Although questions about the completeness of seismicity catalogs prevents us from concluding that the area of the 1960 Chilean earthquake experienced a seismic quiescence as profound as that observed in Washington and Oregon, it does seem clear that seismicity in the southern Chilean region was very low relative to most other subduction zones.

SPECIFIC EXAMPLES

In Figures 5, 7, and 9, we show selected subduction zones where relatively young oceanic crust is subducted. These regions are all plotted on the same scale and show the relative geometry of these subduction zones. Approximate locations of seafloor magnetic lineations, Quaternary volcanoes, and the rupture extent of the largest shallow subduction earthquakes are also shown.

Southern Chile. Figure 5 (a and b) shows southern Chile and the northwestern United States, respectively, and represent subduction zones where very young oceanic crust is presently subducting. The coseismic vertical deformation from the 1960 Chilean earthquake (Plafker and Savage, 1970) is contoured and illustrates several key points. The epicenter of the 1960 earthquake is at the northern end of the zone, which extends nearly 1000 km southward to the point at which the South Chile rise is subducted beneath South America. The rupture propagated across several major fracture zones, with the age of subducted oceanic lithosphere varying between 30 and 5 m.y. The magnetic anomaly time scale is shown in Figure 6. Seismicity prior to the 1960 earthquake appears to have been very low in the region south of 41° where the youngest oceanic lithosphere is subducting (Duda, 1963). The predominant onland coseismic vertical deformation was subsidence; uplift was only observed along the northern coastal region and along southern islands located relatively near to the trench axis. Significant subsidence and uplift of up to 6 m were observed in the south where the youngest oceanic crust is presently subducting. Thus, it appears that large earthquakes in southern Chile involve the subduction of very young oceanic crust.

Several features of subduction in southern Chile are also observed for the Cascadia subduction zone. We have already noted similarities in trench bathymetry as well as the suggestion that both zones may experience significant periods of seismic quiescence. The lengths of the two zones are roughly comparable and the distance between the trench axis and the volcanic arc is slightly greater in the northwestern United States. One to two kilometers of clastic sediments overlie the oceanic plates in both trenches (Schweller *et al.*, 1981; Herron *et al.*, 1981; Kulm *et al.*, 1984), and relatively high heat flows of 1 to 2 HFU have been reported in both areas (Herron *et al.*, 1981; Kulm *et al.*, 1984). Although much of the South American subducting margin is characterized by the juxtaposition of the great heights of the massive Andes Mountains and the great depths of the Peru-Chile Trench, the region of the 1960 earthquake is better characterized (from west to east) by a broad continental margin, low coastal ranges, a central valley that grades southward into an inland

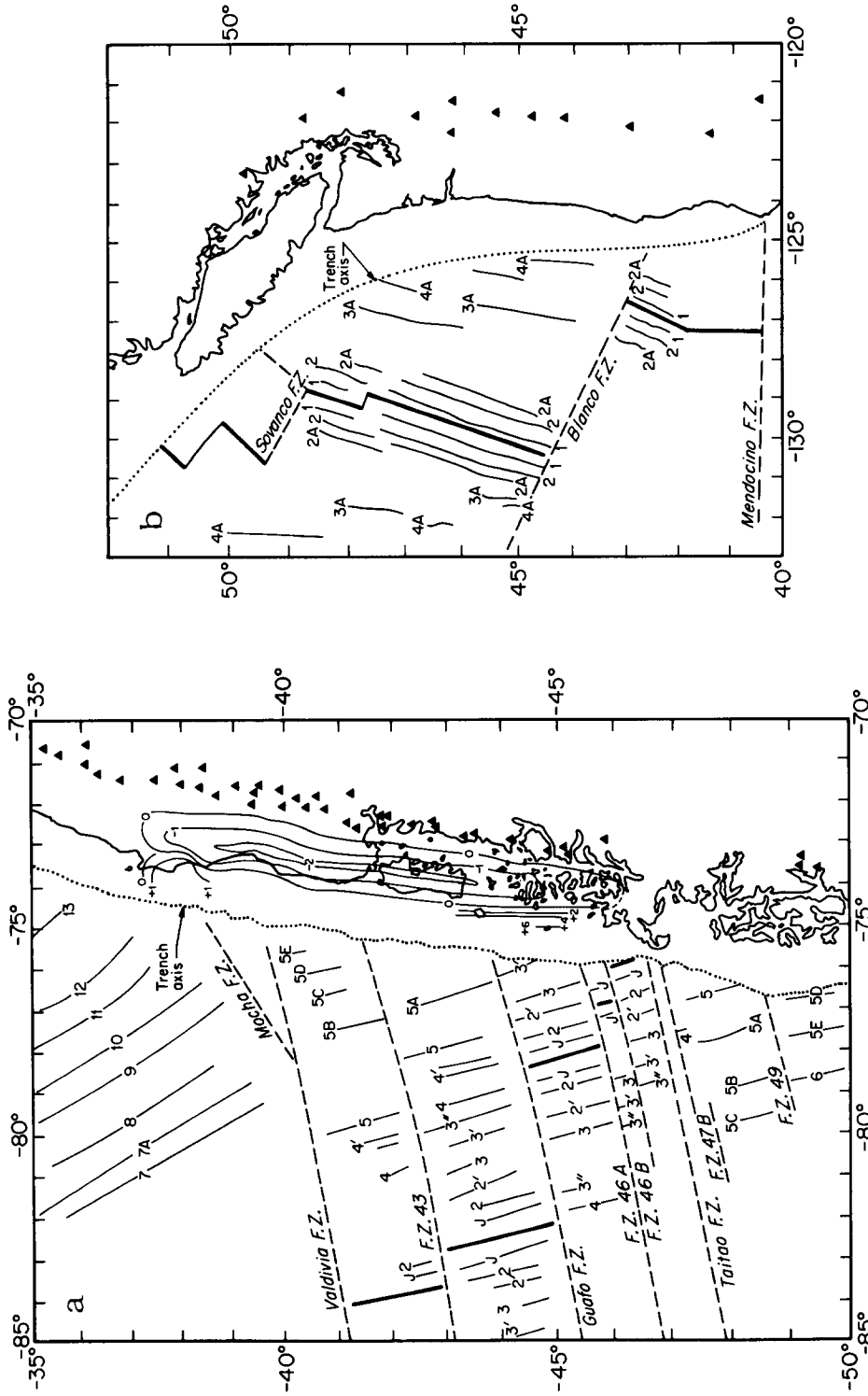


FIG. 5. Comparison of subduction in southern Chile (a) and the northwestern United States (b). All maps shown in Figures 5, 7, and 9 are plotted on approximately the same scale. Active spreading centers (heavy solid lines), seafloor magnetic lineations, and Quaternary volcanoes (solid triangles) are shown. Contours (meters) of the coseismic vertical deformation associated with the 23 May 1960 Chilean earthquake (M_w 9.5) are also shown (Pflafer and Savage, 1970).

sea, and finally a chain of strato-volcanoes (Lowrie and Hey, 1981). These landforms are very similar to those encountered in the northwestern United States.

Lomnitz (1970) reports that earthquakes similar to that in 1960 may have occurred in 1575, 1737, and 1837, thus giving an average repeat time of 128 yr. However, Plafker and Savage (1970) estimate the average dislocation of the 1960 earthquake to be in excess of 20 m, and Minster and Jordan (1978) estimate that the convergence rate is about 9.0 cm/yr. Thus, we might expect the recurrence interval for 1960-sized events to exceed 200 yr. Nishenko (1985) suggests that the events in 1737 and 1837 were somewhat smaller than the one in 1960. In particular, the teleseismic tsunami from the 1837 Chilean earthquake was 57 per cent of the size of the 1960

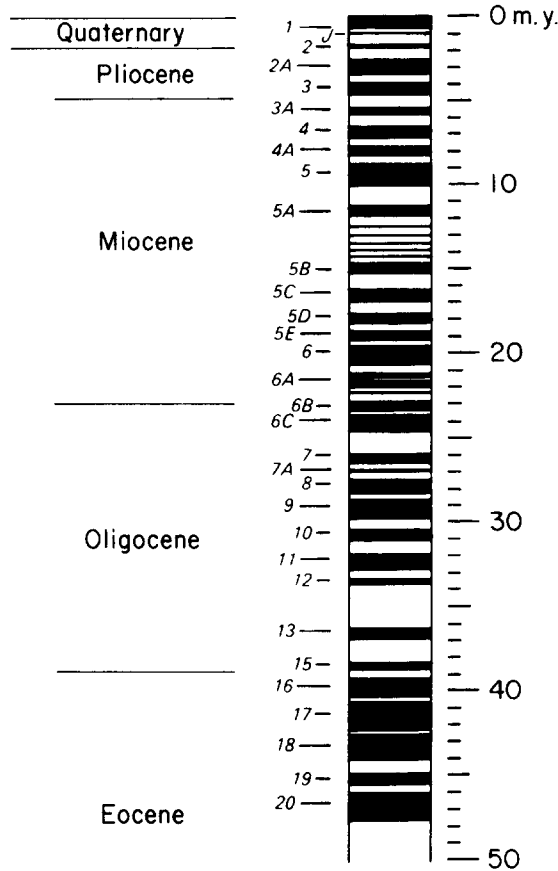


FIG. 6. The geomagnetic time scale.

tsunami as measured at Hilo, Hawaii (Nishenko, 1985). Furthermore, the tsunami from the 1960 earthquake is considered to be the largest teleseismic tsunami in Japanese history (Yoshikawa *et al.*, 1981). Plafker (written communication, 1985) also suggests that the 1960 earthquake was significantly larger than previous events since the type of inundation damage from widespread coastal subsidence that was reported in 1960 was not reported for earlier events. Kanamori and Cipar (1974) report that the 1960 sequence was accompanied by anomalous long-period seismic radiation. They show long-period strain records from Pasadena, California, on which very large 300- to 600-sec waves arrive shortly after the times expected for major seismic phases from the M_S 6.8 foreshock that occurred about 15 min prior

to the main shock. At periods of greater than 300 sec, the foreshock may have been larger than the main shock, and the moment of the entire sequence may have exceeded 6×10^{30} dyne-cm (M_w 9.8). Since it is difficult to explain the observed coseismic coastal deformation with such a large moment, Kanamori and Cipar (1974) speculate that a large-scale, deep-seated deformation was associated with this earthquake. In any case, it appears as though most of the interplate motion in southern Chile occurs during great earthquake sequences.

If the Cascadia subduction zone is analogous to that in southern Chile, then what might we infer about the recurrence time of Cascadia subduction zone earthquakes? Nishimura *et al.* (1984) report convergence rates along the Cascadia subduction zone that range from 3.3 ± 0.9 cm/yr in the south to 4.3 ± 0.9 cm/yr in the north. A simple scaling of convergence rates between the northwestern United States and southern Chile would lead us to postulate great earthquakes along the Cascadia subduction zone with average dislocations in excess of 10 m and recurrence times of 250 to 500 yr. Furthermore, if an earthquake similar to the 1960 Chilean earthquake were to occur, then we might expect the hinge line for coseismic vertical deformation to lie near the coastline, with much of the coast experiencing coseismic subsidence.

Nankai Trough. As can be seen in Figure 7a, there are many geometric similarities between the Nankai Trough and the Cascadia subduction zone. As is the case in southern Chile and the northwestern United States, there are 1 to 2 km of clastic sediments filling a shallow structural trench (Hilde, 1984; Shephard and Bryant, 1983), a broad continental margin, and coast ranges that bound an inland sea or central valley. Unlike southern Chile or the northwestern United States, southwestern Japan does not have a zone of Quaternary strato-volcanoes. Although the oceanic lithosphere subducted beneath southwestern Japan is very young (about 20 m.y.; Kobayashi and Nokada, 1978), it is about twice as old as the lithosphere subducting beneath the northwestern United States. This young age is consistent with the relatively high heat flows (1 to 2 HFU) that are observed in the Philippine seaplate off southwestern Japan (Uyeda, 1984). One unusual feature of this subduction zone is the fact that its seafloor magnetic lineations are oriented perpendicular to the trench axis.

Southwestern Japan is of particular significance because the convergence rate of 3.3 to 4.3 cm/yr (Seno, 1977) is very close to that in the northwestern United States and also because there is an extensive historical earthquake record for this region. This historical record was used by Imamura to anticipate the great earthquakes of 1944 and 1946 (Usami, 1982) despite the fact that, in the words of Richter (1958), they occurred "in a part of the Pacific seismic belt which had been almost completely quiet during the rise of international seismology." More recently, Ando (1975) proposed that coseismic geodetic data from the 1944 and 1946 earthquakes, historical earthquake data, and coastal geomorphic features can be explained by a model that segments the subduction zone into the four semi-independent patches shown in Figure 8. According to Ando's model, these four patches have each ruptured eight times in the last 1300 yr. The patches rupture either simultaneously or individually, but in most instances, failure of the four segments occurred within a time period of less than 3 yr. This pattern has apparently been broken by the absence of rupture in the Tokai region during the sequence of 1944 through 1946.

Although the Nankai Trough is remarkable for its pattern of clusters of great earthquakes that extend the length of the interface with an average repeat of 180 yr, there is apparently still considerable variation in the character and timing of

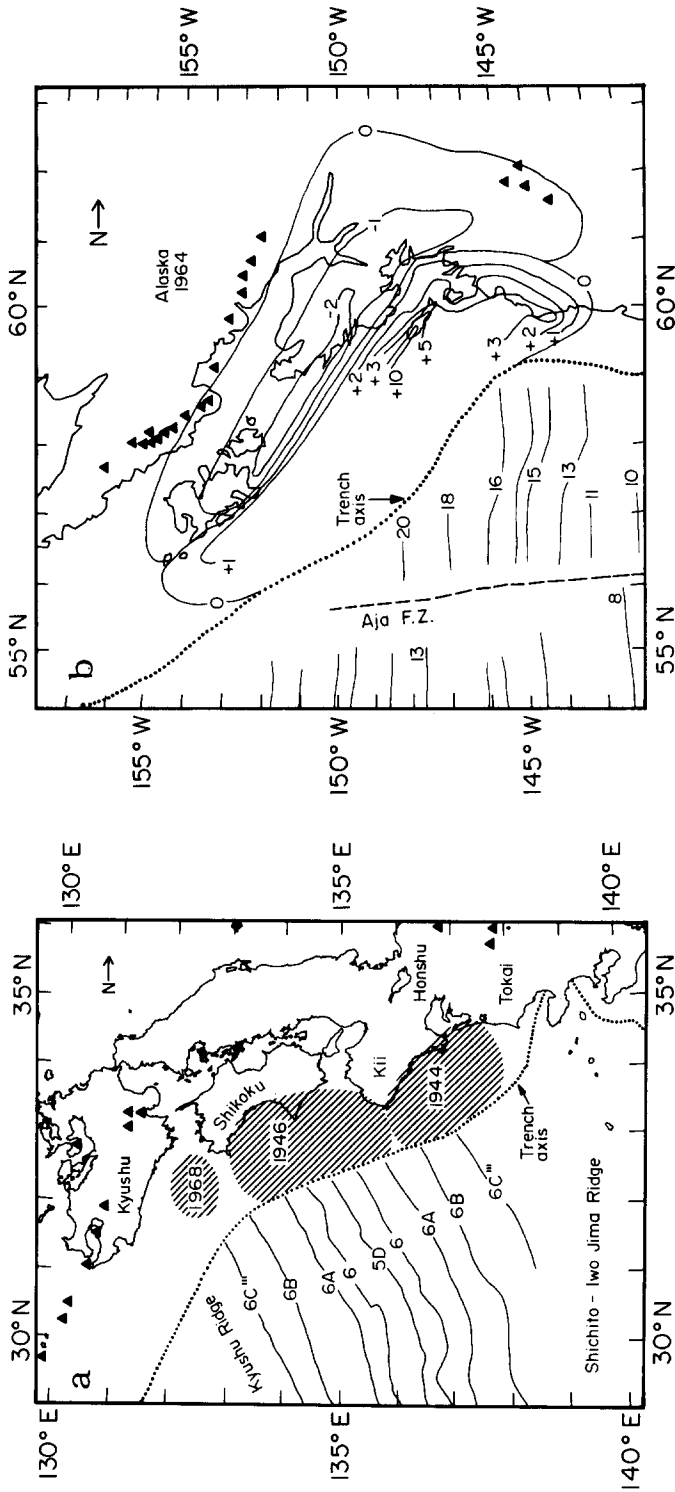


FIG. 7. Similar to Figure 5, but for the Nankai Trough of southwestern Japan (a) and Alaska (b). The approximate rupture areas are indicated for the largest earthquakes in the Nankai Trough in this century (1944 Tonankai earthquake, M_s 8.1; 1968 Hyuganada earthquake, M_s 7.5). Contours (meters) of the coseismic deformation associated with the 28 March 1964 Alaska earthquake (M_w 9.2) are also shown (Pflafer, 1972).

these sequences. Although there is some uncertainty about the completeness of this record prior to 1707, the apparent time interval between sequences has been as little as 90 yr and as much as 260 yr. Shimazaki and Nakata (1980) have studied uplifts on the Muroto peninsula (large peninsula on Shikoku near the middle of the 1946 rupture) from the earthquakes in 1707, 1854, and 1946, and they report uplifts of 1.8, 1.2, and 1.15 m, respectively. From this, they infer that a larger dislocation was associated with the 1707 earthquake than for events in 1854 and 1946. They interpret this data as evidence for the “time-predictable” recurrence model in which

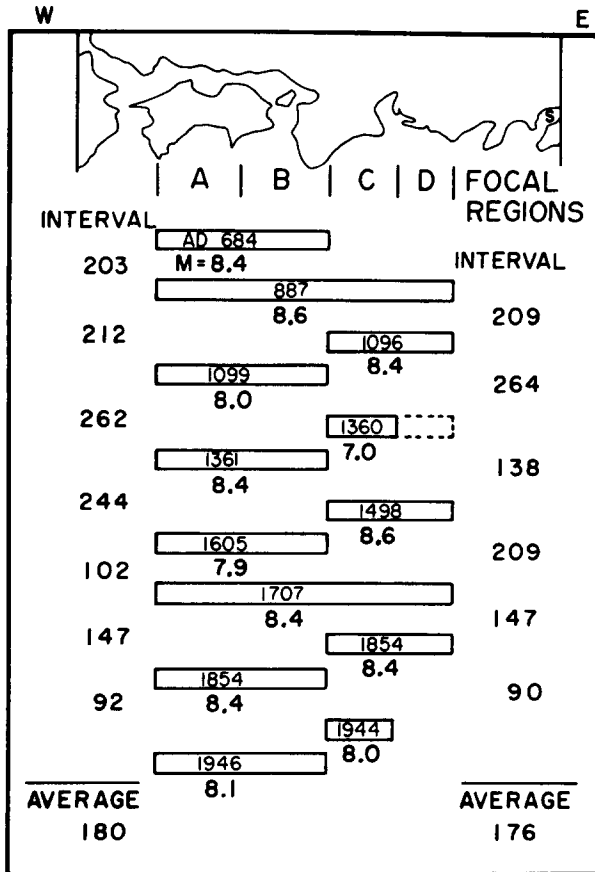


FIG. 8. Ando's (1975) model of historic ruptures along the Nankai Trough in southwestern Japan (from Sieh, 1981). The Tokai seismic gap is indicated by region D.

the time between two earthquakes is determined by the dislocation from the first. However, if the “time-predictable” model is appropriate here and if the historic record is complete, then one must wonder about the nature of earlier events with repeat times in excess of 200 yr.

Estimating the size of the largest Nankai Trough events is somewhat problematic. Kanamori (1972) studied long-period surface waves, and Hartzell and Heaton (1985) studied teleseismic body waves and they both report that the 1944 and 1946 earthquakes have comparable moments of about 1.5×10^{28} dyne-cm (M_w 8.05). Kanamori (1972) estimates a seismic slip of about 3 m for each event. However, Ando (1975) modeled geodetic data from these earthquakes and infers a moment of

6.6×10^{28} dyne-cm (M_w 8.5) and 1.8×10^{28} dyne-cm (M_w 8.1) for the 1946 and 1944 earthquakes, respectively. The 1944 and 1946 earthquakes are clearly not the largest historic events in southwestern Japan. It appears that the 1707 Hoi earthquake ruptured most of the 700 km length of the Nankai Trough (Ando, 1975) and was one of the largest earthquakes in Japanese history. Historic uplift data is scarce, but uplifts in central Shikoku were at least 50 per cent larger in 1707 than in later earthquakes, and Ando's estimates of rupture parameters for this event would indicate a moment of about 1.5×10^{29} dyne-cm (M_w 8.7). Although the scarcity of data makes it difficult to confirm this large moment estimate, it does appear that the estimated rupture length alone, suggests an energy magnitude of at least M_w 8.5 for the 1707 Hoi earthquake.

The subduction zones in southwestern Japan and the northwestern United States are similar in their dimensions, convergence rates, and overall physical characteristics. Although this in no way proves that they experience similar earthquakes, it does suggest that we should consider the possibility that the Cascadia subduction zone may experience sequences of several great earthquakes ($M_w > 8.0$) with repeat times of the sequences from 100 to 250 yr.

Alaska. Although the oceanic crust that is subducting at a rate of 5.7 cm/yr beneath Alaska is relatively young (40 to 50 m.y., Figure 7b), it is considerably older than the oceanic lithosphere in the Cascadia or any other subduction zone considered in this study. One of the most remarkable features of the Alaskan subduction zone is its extensive accretionary wedge; the width of the Alaskan continental margin is about twice that of other subduction zones considered in this study. Although there is a bathymetric trench, it is a relatively subtle feature, and it is filled with 2 km of clastic sediments (Von Huene *et al.*, 1978) having properties similar to those found in the Nankai Trough and the Cascadia subduction zone (Shephard and Bryant, 1983).

Contours of the coseismic vertical uplift associated with the M_w 9.2 1964 Alaskan earthquake are shown in Figure 7b (Plafker, 1972a). Localized uplifts of up to 12 m appear to be associated with activation of high-angle imbricate faults on Montague Island. Coseismic horizontal motions were even more impressive than the vertical deformations; much of the coastline moved oceanward by more than 10 m, and the maximum horizontal displacement was 20 m (Plafker, 1972a). Alewine (1974) modeled these vertical and horizontal deformations and concluded that the fault dislocation exceeded 10 m over a rupture width of greater than 200 km with dislocations of nearly 30 m over the central 70 km of the rupture zone.

Unfortunately, the period of recorded history is very short in Alaska, and the timing and sizes of previous earthquakes must be inferred from geologic evidence. Middleton Island lies far out on the continental margin near the trench axis and displays a sequence of six marine terraces, the oldest of which has a maximum elevation of 46 m and a radiocarbon age of 4500 y.b.p. and the youngest of which was formed during the 1964 earthquake and has an elevation of $3\frac{1}{2}$ m (Plafker, 1972b). Detailed analyses of these terraces indicate time separations of as little as 500 yr for the older terraces and as much as 1350 yr for the most recent terrace, which is also the smallest terrace observed (Plafker and Rubin, 1978). If each terrace can be associated with a great subduction earthquake, then the Middleton Island record implies an average recurrence time of 800 yr, an exceedingly long time when compared with the recurrence time of great historic earthquakes in southern Chile. However, permanent uplifts of Middleton Island are most likely due to slippage along imbricate faults in the accretionary wedge, and there may not be a one-to-

one correspondence between slippage on the main thrust plane and slippage on individual imbricate faults. Although the terrace sequence on Middleton Island may yield an incomplete record of great earthquakes, Plafker (1972b) notes that there is evidence for gradual widespread subsidence of many shorelines for a period of at least 930 yr prior to the 1964 earthquakes. This may be corroborating evidence for a very long repeat time for great Alaskan subduction earthquakes. It is interesting to note, though, that the recurrence times for great earthquakes in the other sections of the Aleutian Trench lying to the west are generally less than 100 yr (Sykes *et al.*, 1981). Although there is evidence suggesting that repeat times of great Alaskan earthquakes may be very long, further corroborating evidence must be found before this result can be considered to be compelling.

We have noted differences in the age of the subducted lithosphere and in the dimensions of the accretionary wedge between the Alaskan and the Cascadia subduction zones. On the other hand, there are similarities in the convergence rate, trench bathymetry, and sedimentation processes. Furthermore, the length of the Cascadia subduction zone is comparable to the rupture length of the 1964 earthquake. If similar earthquakes occur on the Cascadia subduction zone, then it may be that repeat times exceed 1000 yr.

Colombia. The oceanic lithosphere that is subducted at a rate of 8 cm/yr (Minster and Jordan, 1978) beneath southwestern Colombia and northern Ecuador is very young, ranging in age from 10 to 15 m.y. (Figure 9a). As is the case in other regions of young lithosphere subduction, thick sediments fill a shallow structural trench in this region (Hilde, 1984). Colombia has an extensive western continental margin that is somewhat wider, on average, than that encountered in the northwestern United States. A low coastal plain, ranging in width from 30 to 100 km and consisting of several large river deltas, is bounded on the east by three separate mountain chains separated by narrow linear valleys. The central mountain range has peak elevations from 4000 to 6000 m and contains a chain of Quaternary volcanoes. Although some of these landforms are found in other subduction zones, the presence of large river deltas is somewhat unusual and may indicate relative stability or slow subsidence of the coastal region.

Although large earthquakes along Colombia's western coast have been relatively infrequent this century, there have been large subduction earthquakes in 1906, 1942, 1958, and 1979. The area of the rupture zone of the largest of these events, the 1906 earthquake, has been estimated by Kelleher (1972) and is shown in Figure 9a. Seismic radiation from all of these events has been studied by Kanamori and McNally (1982), and they report that despite the fact that the combined rupture areas of the 1942, 1958, and 1979 earthquakes seem to cover the rupture surface of the 1906 earthquakes, the combined moment of these later events is only about one-fifth of that of the 1906 event. They estimate that the 1906 earthquake had a seismic moment of 2×10^{29} dyne-cm (M_w 8.8) with an average dislocation of 5 m over a rupture area of 1.1×10^5 km². The 1979 earthquake is the second largest earthquake having a seismic moment of about 2.9×10^{28} dyne-cm (M_w 8.2) with an average dislocation of 2.7 m over a rupture area of about 2.8×10^4 km² (Kanamori and McNally, 1982).

Unfortunately, historical earthquake records in this region are very sketchy. Ramirez (1933) compiled records of earthquakes during the period from 1575 to 1915, but there is not enough information to clearly identify other great coastal earthquakes. Ramirez does report a great earthquake in 1797 that caused extensive damage in a region extending more than 400 km from southwestern Colombia to

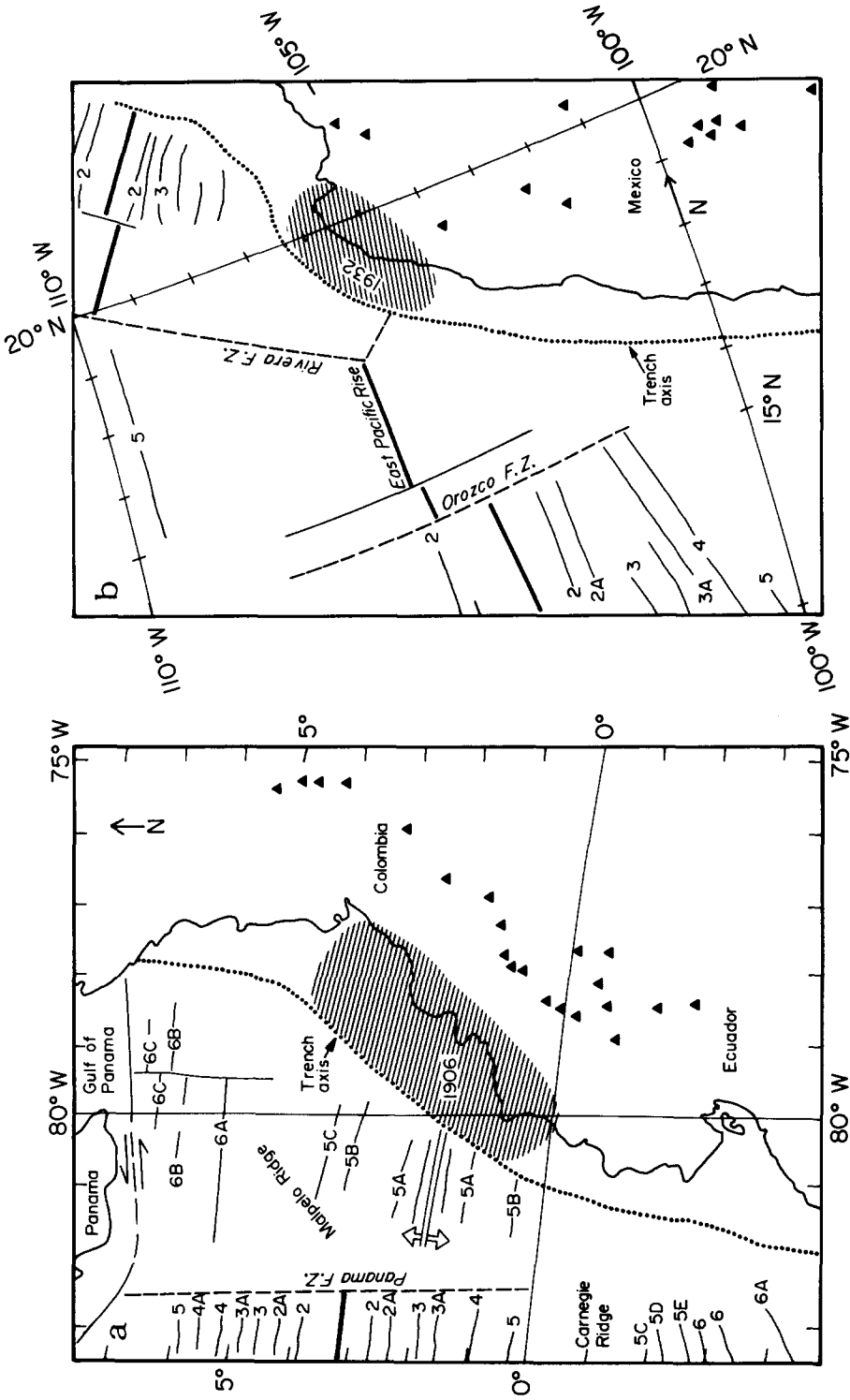


FIG. 9. Similar to Figure 5, but for Colombia (a) and western Mexico (b). The approximate rupture areas of the largest events of this century for each regions are also shown.

central Ecuador. There are no known tsunami reports from this shock, and the source region of this earthquake is unknown. However, the region of high damage seems to extend several hundred kilometers south of that of the 1906 earthquake.

As is apparently the case in the Nankai Trough, it appears that the Colombian subduction zone may rupture either during single very large earthquakes or during a sequence of smaller ones.

Mexico. Subduction of the Rivera plate beneath the North American plate in the region of Jalisco, Mexico, is of considerable interest to us since it represents the slow subduction of a very young and small plate (Figure 9b). Klitgord and Mamerickx (1982) suggest that the subducting Rivera lithosphere is younger than 10 m.y., while the northernmost part of the Cocos plate subducting beneath Colima, Mexico, may be younger than 5 m.y. Eissler and McNally (1984) estimate that in the region of the Rivera fracture zone, the Rivera and Cocos plates are subducting at rates of about 2.3 and 5.5 cm/yr, respectively. Unlike the other areas considered in this report, the continental margin of central Mexico is relatively narrow. Furthermore, there is a shallow bathymetric trench with relatively little clastic sediment (0.6 km; Shepard and Bryant, 1983).

The historic rate of seismicity is relatively high along most of the Middle American Trench with many events in the magnitude 7 to 8 range. Singh *et al.* (1981) and Astiz and Kanamori (1984) suggest that the average repeat times for individual segments of the zone vary between 30 and 60 yr. The largest earthquake sequence documented along the Middle American Trench occurred in 1932 in the Jalisco region of Mexico and the approximate aftershock area as estimated by Singh *et al.* (1985) is shown in Figure 9b. The 1932 earthquake sequence contained two large shocks, a M_S 8.2 (Abe, 1981) event on 3 June that has an estimated moment of 10^{28} dyne-cm (Singh *et al.*, 1985) and a M_S 7.8 (Abe, 1981) event on 18 June with an estimated moment of 7.3×10^{27} dyne-cm (Singh *et al.*, 1985). The average dislocation for the 1932 sequence is estimated to be 1.9 m over an area of 230 km by 80 km. Both events were accompanied by local and teleseismic tsunamis (Abe, 1979), and there is little doubt that these events represent thrusting of the Rivera plate beneath the North American continental margin.

Singh *et al.* (1981) identify other large earthquakes in the Jalisco region in 1837 (M_S 7.7), 1875 (M_S 7.5), 1900 (M_S 7.9), and 1911 (M_S 7.7). However, none of these events appears to be as large as the 1932 sequence. Furthermore, seismicity in the Jalisco region has been remarkably low since the 1932 earthquakes (Astiz and Kanamori, 1984; LeFevre and McNally, 1985).

Although there are clear differences in the physiographic features and the frequency of moderately large coastal earthquakes between subduction of the Rivera and Juan de Fuca plates, the earthquake history of the Jalisco region clearly demonstrates that the slow subduction of a small, very young plate does not necessarily imply aseismic slip.

CASCADIA SUBDUCTION ZONE: LOCKED OR UNLOCKED?

We have noted that there is a large variation in the size and recurrence interval of shallow subduction earthquakes in subduction zones that share physical characteristics with the Cascadia subduction zone. However, it does seem clear that the subduction of very young oceanic lithosphere is often, if not usually, associated with the occurrence of very large shallow earthquakes. Although this does not prove that great earthquakes will occur on the Cascadia subduction zone, it does suggest that

it is inappropriate to assume that great earthquakes will not occur based on observations of bathymetry, lithospheric age, trench sediments, heat flow, convergence rate, physiography, overall size of the subducted plate, Quaternary volcanism, or the rate of background seismicity. A summary of the comparison of physical characteristics of the subduction zones just discussed is found in Table 1.

Geodetic strain. The observation and interpretation of geodetic strain may eventually lead to a fairly clear picture of the subduction process along the Cascadia subduction zone. Savage *et al.* (1981) and later Lisowski and Savage (1986) discuss Geodolite surveys in the Seattle region from 1972 to 1984 and triangulation surveys in the Strait of Juan de Fuca from 1892 to 1954. Their analyses indicate that both regions show maximum contraction in a direction nearly parallel to the proposed east-northeast plate convergence direction. Lisowski and Savage (1986) report shortening at a rate of $0.05 \pm 0.02 \mu\text{strain/yr}$ in the Seattle region and shortening at a rate of $0.22 \pm 0.07 \mu\text{strain/yr}$ in the Strait of Juan de Fuca. Repeated leveling surveys along much of the coastline adjacent to the Juan de Fuca plate (Ando and Balazs, 1979; Reilinger and Adams, 1982; Riddihough, 1982) show steady uplift of the coastal region at a rate of up to 3 mm/yr and subsidence of the inner coastal areas at a rate of about 1 mm/yr. Although Ando and Balazs (1979) interpret the uplift data as evidence of aseismic slip, Lisowski and Savage (1986) show that the combined geodolite, triangulation, and uplift data may be better explained by a model in which the shallow thrust zone is locked between the trench axis and the coastal region. It is interesting to note that the coseismic extension in the central valley of southern Chile ranged from 25 to 50 μstrain for the 1960 earthquake (Plafker and Savage, 1970). Furthermore, coseismic extensions from the 1964 Alaskan earthquake ranged from about 15 μstrain in the inland coastal areas to over 50 μstrain in the outer coastal areas (Plafker, 1972). At the current strain rate, it would take several hundred to 1000 yr for comparable horizontal strains to accumulate along the Cascadia subduction zone. Weaver and Smith (1983) interpret focal mechanisms in southwestern Washington as evidence for crustal compression parallel to the plate convergence direction, and they suggest that these observations are best explained by a model in which the Cascadia subduction zone is locked.

Geomorphology. Holocene geomorphic and depositional features often record the occurrence of great subduction earthquakes. These features vary from region to region and are usually recognized only after a modern earthquake creates features that are repeated in the geologic record. Sequences of uplifted Holocene marine terraces are some of the more easily recognizable features. However, such features are relatively rare since their preservation requires the right combination of high, long-term uplift rates, large coseismic uplifts, and moderate to low coastal erosion rates. Adams (1984a, b) discusses long-term coastal uplift rates along the Cascadia subduction zone and reports relatively slow emergence and probably submergence for most of Washington and northern Oregon. He further reports uplift rates of as high as 1.5 mm/yr in southern Oregon. However, coastal erosion is very strong along this coastline and few, if any, Holocene strand lines have been identified in this region. Curiously, large tree stumps in apparent growth position are found in the intertidal zone at Sunset Bay in southern Oregon. This indicates either short-term or short-wavelength subsidence in a region that has risen at a rate of about 0.6 mm/yr for the last several hundred thousand years (Adams, 1984a, b).

The highest geologic uplift rates documented anywhere along the Cascadia subduction zone occur near Cape Mendocino, California, in the region of the Gorda-

TABLE 1
SUMMARY OF SUBDUCTION ZONE CHARACTERISTICS

	Cascadia	Southern Chile	Nankai Trough	Alaska	Colombia	Mexico (Riviera Plate)
Age (m.y.) of subducted plate	10	5 to 30	18 to 24	40 to 50	10 to 15	10?
Convergence rate (cm/yr)	3.3 to 4.3	9	3.3 to 4.3	5.7	8	2.3
Character of trench	Shallow, sediment-filled	Shallow, sediment-filled	Shallow, sediment-filled	Shallow, sediment-filled	Shallow, sediment-filled	Shallow, few sediments
Free-air gravity anomaly	Small (50 mgal)	Small (50 mgal)	Moderate (100 mgal)	Moderate (100 mgal)	Small (50 mgal)	Moderate (100 mgal)
Background seismicity	Very low	Very low	Low	Moderate	Low	?Moderate?
Heat flow	High (1-2 HFU)	High (1-2 HFU)	High (1-2 HFU)			
Quaternary volcanism	Yes	Yes	No	Varies along the trench	Yes	Yes
Inland sea or valley	Yes	Yes	Yes	Yes	?No?	No
Width of continental margin	30-120 km	100 km	110 km	180 km	130 km	80 km
Largest historic earthquakes (M_w)	None in 150 yr	9.5 (1960)	8.1 (1944, 1946) ?8.5? (1707)	9.2 (1964)	8.8 (1906)	8.2 (1932)
Average repeat time	?400 yr?	128 yr	180 yr	?800 yr?	Unknown	?50 yr?

Pacific-North America triple junction. Lajoie *et al.* (1983) report Holocene terraces along a 40 km stretch that runs southward from Cape Mendocino. A flight of nine emergent terraces and beachridges, uplifted as much as 16.8 m, have been reported in the Cape Mendocino region and ^{14}C dates indicate a maximum uplift rate of 3.6 mm/yr (Lajoie *et al.*, 1983). This would indicate an average repeat time of about 500 yr. However, the relationship between these beach terraces that lie near the structurally complex triple junction and large subduction earthquakes along the length of the subduction zone is unclear.

Turbidites. Adams (1984a) has suggested that extensive Holocene turbidites studied by Griggs and Kulm (1970) may have been triggered by large earthquakes along the continental shelf. Since the deposition of Mazama Ash 6,600 yr ago, there have been approximately 16 major turbidites in the Cascadia channel, indicating an average repeat time of 410 yr. Furthermore, there are hemipelagic layers of relatively uniform thickness between the turbidites, indicating a relatively uniform repeat time. Griggs and Kulm (1970) suggest that this time interval is controlled by the sedimentation rate, with turbidity currents occurring spontaneously due to gravitational instability. However, Adams (1984b) points out that several channels that feed the main Cascadia channel have turbidity sequences comparable to that in the main channel. He suggests this indicates the turbidites were synchronously triggered in several independent channels and may be evidence for the recurrence of great earthquakes every 400 to 500 yr. However, these channels lie within 50 km of each other, and simultaneous turbidity currents may have been triggered by moderate-size earthquakes in the continental margin.

Other evidence. Snavelly (1986) reviews evidence of Cenozoic folding and faulting in the continental margin of Washington and Oregon and concludes that several major and numerous minor faults have been active in the Late Pleistocene and Holocene. Upper Pleistocene abyssal sediments have been uplifted by as much as 1100 m along anticlinal ridges that are bounded by thrust faults in the outer continental margin (Snavelly *et al.*, 1980). Snavelly (1986) interprets the combination of geologically active faults in the accretionary wedge together with a lack of historic earthquakes as evidence for episodic activity of the main thrust zone. However, Adams (1984a) suggests that these deformations may also be occurring through steady aseismic creep.

Snavelly (1986) also suggests that major coastal earthquakes may have triggered major landslides. He cites archaeological evidence (Mauger and Daughert, 1979; Samuels, 1983) that mudflows buried Indian structures at Ozette on the northwest Olympic Coast about 800 y.b.p. and again about 350 y.b.p. Snavelly (1986) also suggests that multiple landslides at Lake Crescent adjacent to the Strait of Juan de Fuca in northern Washington may have been triggered by subduction earthquakes. A submerged tree that was probably transported by the youngest slide has a ^{14}C date of 350 y.b.p. Unfortunately, the significance of these observations is unknown. While it is highly probable that the occurrence of great subduction earthquakes would trigger major landslides, it is also likely that major landslides would occur even if major earthquakes never occur in this region.

Heaton and Snavelly (1985) discuss legends of Washington coastal Indians that were recorded in the 1860's and that suggest the occurrence of a large tsunami along the northwestern Washington coast. They suggest that a large prehistoric subduction earthquake could be responsible for these legends.

Much of the evidence that we have presented is compatible with a simple model in which the Cascadia subduction zone fails during great earthquakes with a repeat

time of 400 to 500 yr. However, it is important to recognize that the logical structure behind this hypothesis is quite weak, consisting of many poorly constrained inferences and conjectures. Does this evidence clearly demonstrate that the Cascadia subduction zone is locked? We don't think so. Furthermore, this issue may remain ambiguous for quite some time. Proving that the subduction process is aseismic seems even more difficult since it requires that one prove that great earthquakes have not occurred. While the 150-yr paucity of historic earthquakes supports this alternative explanation, it may be very difficult to prove that prehistoric earthquakes have not occurred.

HYPOTHETICAL CASCADIAN EARTHQUAKES

If the Cascadia subduction zone is locked, then what is the nature of the earthquakes that may occur there? This question must be answered if we are to provide estimates of the shaking and tsunami hazard from Cascadian subduction earthquakes. However, we address this question under duress; the question is of central importance, but our answers are clearly speculative. We present several hypothetical earthquake sequences that may be plausible for the Cascadia subduction zone.

Completely aseismic zone with no interplate earthquakes. In this case, we assume that slip along the entire boundary between the accretionary wedge and the oceanic lithosphere occurs as aseismic creep. Although we need not worry about great low-angle thrust earthquakes if this condition exists, we must still ask whether the geologically recent faulting and folding observed within the accretionary wedge (Snavely, 1985) has been associated with major earthquakes. Deformation of the accretionary wedge often occurs simultaneously with great interplate earthquakes, and there are not many clear examples where independent large earthquakes are due to subsidiary faults in the accretionary wedge. The M 7.1 Mikawa earthquake (1945) of southwestern Japan may be an example. This earthquake was locally very damaging and occurred along a high-angle thrust fault in the near-coastal region adjacent to the 1944 Tonankai earthquake (Richter, 1958).

Large-scale submarine landsliding or slumping along the continental slope may also present a significant tsunami hazard even if large interplate earthquakes are absent. Kanamori (1985) has suggested that the 1929 Grand Banks, Canada, earthquake (M_S 7.2) and the 1946 Aleutian Islands earthquake (M_S 7.4) may be best described as large landslides in the continental shelf. Despite its small surface wave magnitude, the 1946 Aleutian Islands event generated one of the largest local and teleseismic tsunamis in recent history (Abe, 1979). However, such events are rare and poorly understood.

Mainly aseismic slip with isolated earthquakes of $M < 8$. This is thought to be the situation for what is traditionally called "aseismic" subduction zones such as the Bonins or Marianas. However, there are clearly many physical differences between the Cascadia subduction zone and those of the Marianas' type. If several earthquakes of magnitudes less than 8 were the ultimate culmination of at least 150 yr of plate convergence, then the Cascadia subduction zone would still be considered to be a weakly coupled zone. If isolated moderate-size interplate events do occur here, then one must wonder why isolated earthquakes of M 4, 5, or 6 have not already been observed. Although there is no compelling reason to believe that this is the failure mode of this subduction zone, our present understanding is too limited to allow us to rule out a moderate-size ($8 > M > 7$) interplate earthquake anywhere beneath the continental margin of the northwestern United States.

Moderately coupled with earthquakes of $M < 8\frac{1}{4}$. This is the situation that seems to exist for many world-wide subduction zones; northern Japan and central America may be examples. However, the Cascadia subduction zone does not seem to fit into this category. That is, a moderately high level of seismicity with earthquakes up to $M 8\frac{1}{4}$ is the distinguishing feature of these zones. Thus, the Cascadia subduction zone's historic quiescence seems to eliminate this category.

Strongly coupled with earthquakes of $M > 8$. We have seen that the Cascadia subduction zone shares many features with other zones that are strongly coupled and have experienced very large earthquakes. In this hypothetical situation, we assume that interplate slip occurs only during great earthquakes. Based on our observations of the Nankai Trough, Colombia, and southern Chile, we hypothesize that the Cascadia subduction zone may experience either a sequence of several great earthquakes, or alternatively, a single giant earthquake that ruptures the entire zone. As proposed by Adams (1984a), the average recurrence time of 410 yr for Cascadia basin turbidity currents suggests that the average repeat time for great earthquakes is at least this long.

We can somewhat arbitrarily subdivide the Cascadia subduction zone into three segments: the southern 250 km segment along which the Gorda plate is subducting; the central 800 km segment from the Blanco fracture zone to the middle of Vancouver Island along which the Juan de Fuca plate is subducting; and the northern segment (250 km?) along which the Explorer plate may be subducting. There is little dispute that the central segment is subducting, probably at about 4 cm/yr, but the kinematics of the northern and southern segments are less certain. In the model of Nishimura *et al.* (1984), the Gorda plate is subducted at a rate of 3.3 cm/yr. However, models by Riddihough (1980) and Knapp (1982) suggest that there may be significant complications introduced by internal deformation of the Gorda plate. Riddihough (1984) suggests that the Explorer plate may have a hot-spot rotation pole that is located within the plate itself. Nevertheless, according to Riddihough's model, the North American plate is overriding the Explorer plate at a rate of 2 to 3 cm/yr.

The width of the hypothetical locked zone is another important parameter about which there is considerable uncertainty. The models of Lisowski *et al.* (1985) suggest that, in northern Washington, the plate boundary may be locked from the coastal region to the trench axis, a distance of about 120 km. This may be comparable to the width of the locked zone encountered in southwestern Japan, southern Chile, and Colombia. However, it is probably considerably less than that encountered in Alaska and considerably greater than that encountered in Mexico.

The width of the locked zone may not correspond with the width of the rupture zone of great subduction earthquakes. That is, coseismic rupture may extend beyond the locked zone into the regions that experience aseismic slip (Kanamori and McNally, 1982). The width of the rupture zone may be a function of the length of the rupture zone as well as the width of the locked zone. For the present, we will sidestep the issue of just how wide the rupture zone may be for Cascadia earthquakes. Instead, we will consider the types of sequences that would cover the length of the Cascadia subduction zone.

We begin by proposing that earthquakes on the Cascadia subduction zone are analogous to those in the Nankai Trough. We have seen that the Nankai Trough has repeatedly experienced sequences of great earthquakes that cover the 700 km length of this subduction zone. Four or five earthquakes similar to the 1944 or 1946

Nankai earthquakes (M_w 8.1) would probably cover the length of the Cascadia subduction zone. Such earthquakes may be closely spaced in time as has been the case for many sequences in southwestern Japan. Larger earthquakes, such as the 1707 Hiei earthquake that apparently ruptured over a 700 km length of the Nankai Trough, should also be considered as a possibility. As we have already discussed, it is difficult to estimate the true size of the 1707 earthquake, but it seems likely that it was at least in the M_w 8.5 range. Convergence rates in southwestern Japan are similar to those along the Cascadia subduction zone, and thus we might expect the average repeat time of such sequences to be between 100 and 250 yr in the Cascadia subduction zone.

Several earthquakes of the size of the 1906 Colombian earthquake (M_w 8.8) could also cover the length of the Cascadia subduction zone. Such earthquakes may have rupture lengths of about 500 km.

Finally, the overall dimensions of the Cascadia subduction zone are similar to the zone that ruptured during the 1960 Chilean earthquake (M_w 9.5). This is the largest event recorded in this century, and thus, we consider it to be a reasonable upper bound for hypothetical Cascadia subduction earthquake. Dislocations in an event of this nature may exceed 20 m, and the repeat time would probably be at least 400 yr.

TSUNAMI POTENTIAL

If a great subduction earthquake were to occur along the Cascadia subduction zone, then it is very likely that it would generate a large local tsunami. It is beyond the scope of this study to attempt to predict tsunami run-up heights from specific models of the source and coastal physiography. However, we can make several general observations of tsunami run-up heights for great world-wide subduction earthquakes. Figure 10, modified from a figure prepared by Abe (1979), shows the maximum observed local run-up heights of various world-wide earthquakes for which Abe calculated a tsunami magnitude, M_t . Abe (1979) defines the M_t scale to be the logarithm of tsunami amplitude as measured at calibrated sites at great distances. Abe (1979) shows that there is a close correspondence between M_t and energy magnitude M_w . Thus, Figure 10 is similar to a plot of maximum local run-up height as a function of energy magnitude, M_w . As is obvious from this plot, maximum run-up height is not simply a function of earthquake size. Local tsunami run-up heights often vary considerably along the shoreline adjacent to major earthquakes, and the maximum run-up height may be several times the average run-up height.

All of the regions that we considered to be possible analogs for the Cascadia subduction zone have experienced large local tsunamis. The 1944 and 1946 Nankai Trough earthquakes generated tsunamis that had maximum local run-up heights of 7.5 and 6.0 m, respectively, and the 1707 Hiei probably generated an even larger tsunami (Lockridge and Smith, 1984). The 1906 Colombian earthquake also generated a large local tsunami that extensively damaged much of the coastal regions of southwestern Colombia and northern Ecuador (Lockridge and Smith, 1984). The 1960 Chilean earthquake generated one of the largest well-documented tsunamis in recent times with heavy damage occurring both locally (Sievers *et al.*, 1963) and also in Hawaii (Cox and Mink, 1963) and Japan (Yoshikawa *et al.*, 1981). Although the maximum local run-up may have been over 20 m (Lockridge and Smith, 1984),

it seems likely that most coastal regions adjacent to the earthquake experienced local run-up heights of less than 10 m (Sievers *et al.*, 1963).

It is very difficult at this point to provide reliable estimates of the tsunami run-up heights that may occur following any large subduction earthquakes along the Cascadia subduction zone. However, run-up heights ranging from several meters to several tens of meters have been observed along other subducting boundaries after events of the type that we consider to be plausible for the Cascadia subduction zone.

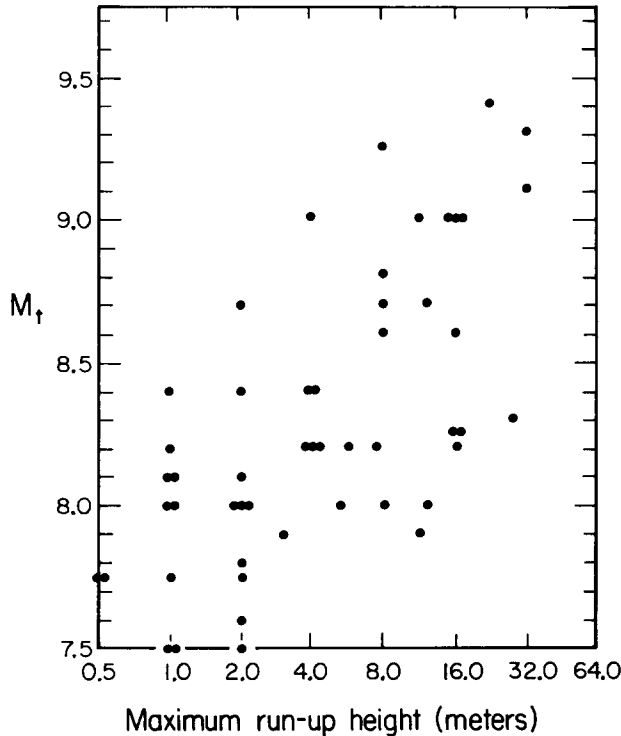


FIG. 10. Maximum local tsunami run-up heights as a function of Abe's (1979) tsunami magnitude which corresponds closely to energy magnitude M_w (modified from Abe, 1979).

CONCLUSIONS

Is slip along the Cascadia subduction zone a benign process, occurring slowly as aseismic creep? Or alternatively, is elastic strain energy accumulating along this zone, and if it is, what is the nature of earthquakes that may result? Unfortunately, this study has not satisfactorily resolved these important issues. However, of other world-wide subduction zones, the Cascadia subduction zone seems most similar to southern Chile, the Nankai Trough, and Colombia, each of which have experienced very large historic earthquake sequences. The length of the subducting margin along the Cascadia subduction zone probably exceeds 1000 km, and if the zone is locked, then earthquakes of very large size must be considered. Several observations seem compatible with a model in which great earthquake sequences occur with an average repeat time of 400 to 500 yr, but because of the nature of this data, we consider this hypothesis to be highly speculative.

ACKNOWLEDGMENTS

This research was supported by the U.S. Nuclear Regulatory Commission. The authors benefited from discussions with Hiroo Kanamori, Greg Davis, Parke Snively, Jr., John Adams, Larry Ruff, Craig Weaver, Yoko Ota, Atsumasa Okada, Kerry Sieh, George Plafker, and Mike Lisowski. We thank Jim Savage and Peter Ward for reviewing the manuscript.

REFERENCES

- Abe, K. (1979). Size of great earthquakes of 1837–1974 inferred from tsunami data, *J. Geophys. Res.* **84**, 1561–1568.
- Abe, K. (1981). Magnitude of large shallow earthquakes from 1904 to 1980, *Phys. Earth Planet. Interiors* **27**, 72–92.
- Acharya, H. (1981). Juan de Fuca Plate—Aseismic subduction at 1.8 cm/yr, *Geophys. Res. Letters* **8**, 1123–1125.
- Acharya, H. (1985). Comment on seismic potential associated with subduction in the northwestern United States by T. H. Heaton and H. Kanamori, *Bull. Seism. Soc. Am.* **75**, 889–890.
- Adams, J. (1984a). Active deformation of the Pacific Northwest continental margin, *Tectonics* **3**, 449–472.
- Adams, J. (1984b). Vertical crustal deformation of the Pacific Northwest continental margin: some enigmas with bearing on great-earthquake risk, *Proceedings of Chapman Conference on Vertical Crustal Motion*.
- Alewine, R. W., III (1974). Application of linear inversion theory toward the estimation of seismic source parameters, *Ph.D. Thesis*, California Institute of Technology, Pasadena, California, 303 pp.
- Ando, M. (1975). Source mechanisms and tectonic significance of historical earthquakes along the Nankai Trough, Japan, *Tectonophysics* **27**, 119–140.
- Ando, M. and E. I. Balazs (1979). Geodetic evidence for aseismic subduction of the Juan de Fuca plate, *J. Geophys. Res.* **84**, 3023–3028.
- Astiz, L. and H. Kanamori (1984). An earthquake doublet in Ometepec, Guerrero, Mexico, *Phys. Earth Planet. Interiors* **34**, 24–45.
- Cox, D. C. and J. F. Mink (1963). The tsunami of 23 May 1960 in the Hawaiian Islands, *Bull. Seism. Soc. Am.* **53**, 1191–1210.
- Duda, S. J. (1963). Strain release in the Circum-Pacific belt, Chile: 1960, *J. Geophys. Res.* **68**, 5531–5544.
- Eissler, H. K. and K. C. McNally (1984). Seismicity and tectonics of the Rivera plate and implications for the 1932 Jalisco, Mexico, earthquake, *J. Geophys. Res.* **89**, 4520–4530.
- Grellet, C. and J. Dubois (1982). The depth of trenches as a function of the subduction rate and age of the lithosphere, *Tectonophysics* **82**, 45–56.
- Griggs, G. B. and L. D. Kulm (1970). Sedimentation in Cascadia deep-sea channel, *Geol. Soc. Am. Bull.* **81**, 1361–1384.
- Hartzell, S. H. and T. H. Heaton (1985). Teleseismic time functions for large shallow subduction zone earthquakes, *Bull. Seism. Soc. Am.* **75**, 965–1004.
- Heaton, T. H. and H. Kanamori (1984). Seismic potential associated with subduction in the northwestern United States, *Bull. Seism. Soc. Am.* **74**, 933–941.
- Heaton, T. H. and P. D. Snively, Jr. (1985). Possible tsunami along the northwestern coast of the United States inferred from Indian traditions, *Bull. Seism. Soc. Am.* **75**, 1455–1460.
- Heaton, T. H. and S. H. Hartzell (1986). Estimation of strong ground motions from hypothetical earthquakes on the Cascadia subduction zone (manuscript in preparation).
- Herron, E. M., S. C. Camde, and B. R. Hall (1981). An active spreading center collides with a subduction zone: a geophysical survey of the Chile margin triple junction, *Geol. Soc. Am. Memoir* **154**, 683–701.
- Hilde, T. W. C. (1984). Sediment subduction versus accretion around the Pacific, *Tectonophysics* **99**, 381–397.
- Hyndman, R. D. and D. H. Wiechert (1983). Seismicity and rates of relative motion on the plate boundaries of western North America, *Geophys. J. R. Astr. Soc.* **72**, 59–82.
- Japan Meteorological Agency (1984). Hypocentral distribution in the Tokai and southern Kanto districts, 1963–1983, *Report of the Coordinating Committee for Earthquake Prediction* **32**, 212–215.
- Kanamori, H. (1972). Tectonic implications of the 1944 Tonankai and 1946 Nankaido earthquakes, *Phys. Earth Planet. Interiors* **5**, 129–139.
- Kanamori, H. (1977). Seismic and aseismic slip along subduction zones and their tectonic implications,

- in *Island Arcs, Deep Sea Trenches and Back-Arc Basins*, Maurice Ewing Series, M. Talwani and W. C. Pitman, Editors, American Geophysical Union, Washington, D.C., 163-174.
- Kanamori, H. (1985). Non-double couple seismic source, 23rd General Assembly of IASPEI, Tokyo, Japan.
- Kanamori, H. and J. J. Cipar (1974). Focal process of the great Chilean earthquake of May 22, 1960, *Phys. Earth Planet. Interiors* **9**, 128-136.
- Kanamori, H. and K. C. McNally (1982). Variable rupture mode of the subduction zone along the Ecuador-Colombia Coast, *Bull. Seism. Soc. Am.* **72**, 1241-1253.
- Kelleher, J. A. (1972). Rupture zones of large South American earthquakes and some predictions, *J. Geophys. Res.* **77**, 2087-2103.
- Kelleher, J., J. Savino, H. Rowlett, and W. McCann (1974). Why and where great thrust earthquakes occur along island arcs, *J. Geophys. Res.* **79**, 4889-4899.
- Klitgord, K. D. and J. Mammerickx (1982). Northeast Pacific Rise: magnetic anomaly and bathymetric framework, *J. Geophys. Res.* **87**, 6725-2750.
- Knapp, J. S., Jr. (1982). Seismicity, crustal structure, and tectonics near the northern termination of the San Andreas fault, *Ph.D. Thesis*, University of Washington, Seattle, Washington.
- Kobayashi, K. and M. Nakada (1978). Magnetic anomalies and tectonic evolution of the Shikoku Inter-arc Basin, *J. Phys. Earth* **26** (Supplement), 391-402.
- Kulm, L. D., et al. (1984). Western North American continental margin and adjacent ocean floor off Oregon and Washington, Atlas 1 Ocean Margin Drilling Program, Regional Atlas Series: Marine Science International, Woods Hole, Massachusetts, 32 sheets.
- Lajoie, J. R., G. L. Kennedy, S. A. Mathieson, A. M. Sarna-Wojcicki, S. A. Morrison, and M. K. Tobish (1983). Emergent Holocene marine terraces at Cape Mendocino and Ventura, California, U.S.A., *Proc. International Symposium on Development of Holocene Shorelines*, Tokyo, Japan.
- Lay, T., H. Kanamori, and L. Ruff (1982). The asperity model and the nature of large subduction zone earthquakes, in *Earthquake Prediction Research*, Terra Scientific Publishing Co., Tokyo, Japan, 3-71.
- LeFevre, L. V. and K. C. McNally (1985). Stress distribution and subduction of aseismic ridges in the Middle America subduction zone, *J. Geophys. Res.* **90**, 4495-4510.
- Lisowski, M. and J. C. Savage (1986). Geodetic strain in northwestern Washington (manuscript in preparation).
- Lockridge, P. A. and R. H. Smith (1984). *Tsunamis in the Pacific Basin*, map published by the National Geophysical Data Center, NOAA, Boulder, Colorado.
- Lomnitz, C. (1970). Major earthquakes and tsunamis in Chile during the period 1535-1955, *Geol. Rundsh.* **59**, 938-960.
- Lowrie, A. and R. Hey (1981). Geological and geophysical variations along the western margin of Chile near lat 33° to 36°S and their relation to Nazca plate subduction, *Geol. Soc. Am. Memoir* **154**, 741-754.
- Mauger, J. E. and R. D. Daugherty (1979). Ozette Archaeological Project, Interim Final Report, Phase XI, Project Report No. 68, Washington Archaeological Research Center, Washington State University, Pullman, Washington.
- Minster, J. B. and T. H. Jordan (1978). Present-day plate motions, *J. Geophys. Res.* **83**, 5331-5354.
- Nishenko, S. P. (1985). Seismic potential for large and great interplate earthquakes along the Chilean and southern Peruvian Margins of South America: a quantitative reappraisal, *J. Geophys. Res.* **90**, 3589-3615.
- Nishimura, C., D. S. Wilson, and R. N. Hey (1984). Pole of rotation analysis of present-day Juan de Fuca plate motion, *J. Geophys. Res.* **89**, 10283-10290.
- NOAA (1981). The marine geophysical data exchange format 'MGD77', Key to Geophysical Records Documentation No. 10, National Geophysical and Solar-Terrestrial Data Center, Boulder, Colorado, 17 pp.
- Parsons, B. and J. G. Sclater (1977). An analysis of the variation of the ocean floor bathymetry and heat flow with age, *J. Geophys. Res.* **82**, 803-827.
- Peterson, E. T. and T. Seno (1984). Factors affecting seismic moment release rates in subduction zones, *J. Geophys. Res.* **89**, 10,233-10,248.
- Plafker, G. (1972a). Tectonics, in *The Great Alaska Earthquake of 1964, Seismology and Geodesy*, National Academy of Sciences, Washington, D.C.
- Plafker, G. (1972b). Alaskan earthquake of 1964 and Chilean earthquake of 1960, implications for arc tectonics, *J. Geophys. Res.* **77**, 901-925.
- Plafker, G. and J. Savage (1970). Mechanism of the Chilean earthquakes of May 21 and 22, 1960, *Geol. Soc. Am. Bull.* **81**, 1001-1030.

- Plafker, G. and M. Rubin (1978). Uplift history and earthquake recurrence as deduced from marine terraces on Middleton Island, Alaska, *Proc. Conf. VI: Methodology for Identifying Seismic Gaps and Soon-to-Break Gaps, U.S. Geol. Surv., Open-File Rept. 78-943*, 687-722.
- Ramirez, J. E. (1933). Earthquake history of Colombia, *Bull. Seism. Soc. Am.* **23**, 13-22.
- Reillinger, R. E. and J. Adams (1982). Geodetic evidence for active landward tilting of the Oregon and Washington coastal ranges, *Geophys. Res. Letters* **9**, 401-403.
- Richter, C. F. (1958). *Elementary Seismology*, W. H. Freeman and Company, San Francisco, California, 768 pp.
- Riddihough, R. P. (1977). A model for recent interactions off Canada's west coast, *Can. J. Earth Sci.* **14**, 384-396.
- Riddihough, R. P. (1978). The Juan de Fuca plate, *EOS* **59**, 836-842.
- Riddihough, R. P. (1980). Gorda plate motions from magnetic anomaly analysis, *Earth Planet. Sci. Letters* **51**, 163-170.
- Riddihough, R. P. (1982). Contemporary vertical movements and tectonics on Canada's west coast: a discussion, *Tectonophysics* **86**, 319-341.
- Riddihough, R. P. (1984). Recent movements of the Juan de Fuca plate systems, *J. Geophys. Res.* **89**, 6980-6984.
- Ruff, L. and H. Kanamori (1980). Seismicity and the subduction process, *Phys. Earth Planet. Interiors* **23**, 240-252.
- Samuels, S. M. (1983). Spatial patterns and cultural processes in three Northwest Coast longhouse floor middens from Ozette, *Ph.D. Thesis*, Washington State University, Pullman, Washington.
- Savage, J. C., M. Lisowski, and W. H. Prescott (1981). Geodetic strain measurements in Washington, *J. Geophys. Res.* **86**, 4929-4940.
- Schweller, W. D., L. D. Kulm, and R. A. Prince (1981). Tectonics structure, and sedimentary framework of the Peru-Chile Trench, *Geol. Soc. Am. Memoir* **154**, 323-349.
- Seno, T. (1977). The instantaneous rotation vectors of the Philippine Sea plate relative to the Eurasian plate, *Tectonophysics* **42**, 209-226.
- Shephard, L. E. and W. R. Bryant (1983). Geotechnical properties of lower trench inner-slope sediments, *Tectonophysics* **99**, 279-312.
- Shimazaki, K. and T. Nakata (1980). Time-predictable recurrence model for large earthquakes, *Geophys. Res. Letters* **1**, 279-282.
- Sieh, K. E. (1981). A review of geological evidence for recurrence times of large earthquakes, in *Earthquake Prediction an International Review*, Maurice Ewing Series 4, American Geophysical Union, 181-207.
- Sievers, H. A., G. Villegas, and G. Barros (1963). The seismic sea wave of 22 May 1960 along the Chilean Coast, *Bull. Seism. Soc. Am.* **53**, 1125-1190.
- Singh, S. K., L. Astiz, and J. Havskov (1981). Seismic gaps and recurrence periods of large earthquakes along the Mexican subduction zone: a reexamination, *Bull. Seism. Soc. Am.* **71**, 827-843.
- Singh, S. K., L. Ponce, and S. P. Nishenko (1985). The great Jalisco Mexico earthquakes of 1932 and the Rivera subduction zone, *Bull. Seism. Soc. Am.* **75**, 1301-1313.
- Snively, P. D., Jr. (1986). Tertiary geology and petroleum potential of the western Oregon and Washington continental margin, in *Geology and Resource Potential of the Continental Margin of Western North America and Adjacent Ocean Basins—Beaufort Sea to Baja California*, D. W. Scholl, A. Grantz, and J. G. Vedder, Editors, American Association of Petroleum Geologists Memoir Series (in press).
- Snively, P. D., Jr., H. C. Wagner, and D. L. Lander (1980). Interpretation of the Cenozoic geologic history, central Oregon continental margin: cross-section summary, *Geol. Soc. Am. Bull.* (Part I), **91**, 143-146.
- Sykes, L. R., J. B. Kisslinger, L. S. House, J. N. Davies, and K. H. Jacob (1981). Rupture zones and repeat times of great earthquakes along the Alaska-Aleutian Arc, in *Earthquake Prediction*, D. W. Simpson and P. G. Richards, Editors, American Geophysical Union, Washington, D.C., 73-80.
- Taber, J. J. and S. W. Smith (1985). Seismicity and focal mechanisms associated with the subduction of the Juan de Fuca plate beneath the Olympic Peninsula, Washington, *Bull. Seism. Soc. Am.* **75**, 237-249.
- Takagi, A. (1982). How small earthquakes occur, in *Earthquake Prediction Techniques*, T. Asada, Editor, University of Tokyo press, Tokyo, Japan, 63-88.
- Usami, T. (1982). Great earthquakes of the past, in *Earthquake Prediction Techniques*, T. Asada, Editor, University of Tokyo Press, Tokyo, Japan, 11-29.
- Uyeda, S. (1984). Subduction zones: their diversity, mechanism, and human impacts, *Geol. J.* **8.4**, 381-406.

- Uyeda, S. and H. Kanamori (1979). Back-arc opening and the mode of subduction, *J. Geophys. Res.* **84**, 1049–1061.
- Von Huene, R., J. C. Moore, and G. W. Moore (1978). Cross section of Alaska Peninsula–Kodiak Island–Aleutian Trench, Geological Society of America, MC-28A.
- Weaver, C. S. and S. W. Smith (1983). Regional tectonic hazard implications of a crustal fault zone in southwestern Washington, *J. Geophys. Res.* **88**, 10371–10383.
- Yoshikawa, T., S. Kaizuka, and Y. Ota (1981). *The Landforms of Japan*, University of Tokyo Press, Tokyo, Japan, 222 pp.

U.S. GEOLOGICAL SURVEY
CALIFORNIA INSTITUTE OF TECHNOLOGY
PASADENA, CALIFORNIA 91125

Manuscript received 22 September 1985