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# Seismic Gaps and Plate Tectonics: Seismic Potential for Major Boundaries[1]

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*Abstract* – The theory of plate tectonics provides a basic framework for evaluating the potential for future great earthquakes to occur along major plate boundaries. Along most of the transform and convergent plate boundaries considered in this paper, the majority of seismic slip occurs during large earthquakes, i.e., those of magnitude 7 or greater. The concepts that rupture zones, as delineated by aftershocks, tend to abut rather than overlap, and large events occur in regions with histories of both long- and short-term seismic quiescence are used in this paper to delineate major seismic gaps.

In detail, however, the distribution of large shallow earthquakes along convergent plate margins is not always consistent with a simple model derived from plate tectonics. Certain plate boundaries, for example, appear in the long term to be nearly aseismic with respect to large earthquakes. The identification of specific tectonic regimes, as defined by dip of the inclined seismic zone, the presence or absence of aseismic ridges and seamounts on the downgoing lithospheric plate, the age contrast between the overthrust and underthrust plates, and the presence or absence of back-arc spreading, have led to a refinement in the application of plate tectonic theory to the evaluation of seismic potential.

The term seismic gap is taken to refer to any region along an active plate boundary that has not experienced a large thrust or strike-slip earthquake for more than 30 years. A region of high seismic potential is a seismic gap that, for historic or tectonic reasons, is considered likely to produce a large shock during the next few decades. The seismic gap technique provides estimates of the location, size of future events and origin time to within a few tens of years at best.

The accompanying map summarizes six categories of seismic potential for major plate boundaries in and around the margins of the Pacific Ocean and the Caribbean, South Sandwich and Sunda (Indonesia) regions for the next few decades. These categories range from what we consider high to low potential for being the site of large earthquakes during that period of time. Categories 1, 2 and 6 define a time-dependent potential based on the amount of time elapsed since the last large earthquake. The remaining categories, 3, 4, and 5, are used for areas that have ambiguous histories for large earthquakes; their seismic potential is inferred from various tectonic criteria. These six categories are meant to be interpreted as forecasts of the location and size of future large shocks and should not be considered to be predictions in which a precise estimate of the time of occurrence is specified.

Several of the segments of major plate boundaries that are assigned the highest potential, i.e., category 1, are located along continental margins, adjacent to centers of population. Some of them are hundreds of kilometers long. High priority should be given to instrumenting and studying several of these major seismic gaps since many are now poorly instrumented. The categories of potential assigned here provide a rationale for assigning priorities for instrumentation, for future studies aimed at predicting large earthquakes and for making estimates of tsunami potential.

**Key words:** Seismic gaps; Earthquake prediction; Plate tectonics.

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## *1. Introduction*

This paper attempts to review and synthesize information about the potential of certain parts of plate boundaries to rupture in future large earthquakes. Over the last ten years it has become clear that some segments of major plate boundaries that have not been the sites of large earthquakes for tens to hundreds of years, i.e., have been seismic gaps for large shocks, are more likely to be sites of future large shocks than segments that have experienced major rupture during, say, the last 30 years. FEDOTOV (1965), MOGI (1968), TOBIN and SYKES (1968), SYKES (1971), KELLEHER (1972), KELLEHER *et al.* (1973) and others delineate major seismic gaps for several of the major plate boundaries of the circum-Pacific and Caribbean regions. Several of these gaps have since been 'filled in' by rupture during large earthquakes.

The seismic gap technique provides estimates of the location and maximum likely size of shocks, but it does not provide an estimate to better than tens of years of the time of occurrence of future large shocks. In this paper the word 'forecast' is used for estimates of location and size (magnitude or seismic moment) of large shocks by the method of seismic gaps. The term 'prediction' is reserved for estimates that involve a more precise calculation of the time and probability of occurrence as well as the size and location.

Although the concept of seismic gaps was applied by FEDOTOV (1965) to the active Kuril-Kamchatka seismic zone prior to the advent of plate tectonics, the plate tectonic model provides a foundation for understanding the build up and release of strain energy in large shocks and for understanding why the gap hypothesis works as well as it appears to. Most of the world's earthquakes occur along rather well defined belts that coincide with plate boundaries. Relatively little deformation occurs in most of the interiors of plates.

Most of the world's great (magnitude, $M \geq 7\frac{3}{4}$) earthquakes (category *a* of GUTENBERG and RICHTER, 1954) occur at shallow depths (depths less than about 40 km) along plate boundaries of the thrust (convergent) type where subduction of one lithospheric plate is presently occurring. Most of the Earth's other great shocks occur along long transform faults where relative plate motion involves strike-slip faulting. Along these two types of seismic zones, the slip during earthquakes reflects the relative motion of two interacting lithospheric plates. Active ridge crests where sea-floor spreading is occurring, however, do not appear to be sites of shocks larger than about magnitude 7. Along ridge crests, short transform faults, and along those parts of transform faults near spreading centers the lithosphere is young, hot, weak, and thin, and the region capable of releasing seismic energy is apparently insufficient for earthquakes larger than about magnitude 7.

This study is restricted to large shallow shocks along simple plate boundaries of the thrust or transform type. It attempts to include many (but not all) of the zones of this type in and around the margins of the Pacific Ocean as well as the major plate boundaries of the Caribbean, South Sandwich, and Sunda (Indonesian) regions. The zone of complex, multibranched deformation extending from the Mediterranean across the Middle East, Central Asia, and China is not covered here.

In this paper the term 'large' is used in a general sense for shocks of magnitude 7 or greater. The term 'major' is employed for $7 \leq M \leq 7\frac{3}{4}$ (class *b* of GUTENBERG and RICHTER, 1954). Great shocks are those of $M \geq 7\frac{3}{4}$.

Most of the cumulative seismic energy release, cumulative seismic moment or cumulative seismic slip along major plate boundaries occurs in large or great earthquakes (BRUNE, 1968). Shocks of magnitude smaller than 7 may be regarded as noise or as accessory to the major plate movement, which occurs mainly in conjunction with large earthquakes or which may occur in part by a seismic deformation. The strain energy that is released in large shallow shocks is believed to be built up slowly along simple plate boundaries for tens to hundreds of years. This strain comes from the movement of plates, which varies from about 2 to 12 cm/year for the major plate boundaries discussed here. Friction along plate boundaries prevents many major seismic zones from moving continuously on a scale shorter than tens to hundreds of years. Once stresses build up to a critical level, the plate interface moves suddenly about 1 to 20 m during the rupture associated with a large earthquake.

Plate tectonic theory indicates that plate boundaries are continuous and that they do not end suddenly unless the motion is transformed into some other type of tectonic feature such as a spreading ridge, subduction zone or transform fault. Hence, parts of a plate boundary that have not experienced large earthquakes for tens to hundreds of years are likely to be either: (1) sites of future large shocks, or (2) regions where plate movement is accommodated either aseismically or by the occurrence of only small to moderate-size shocks. In the long term, which is here taken to be thousands of years, plate movement is thought to be fairly even along the entire length of a simple plate boundary. For more complex, multibranched zones of deformation as in Central Asia, western China or Central Alaska, strain build up may occur less regularly and in a more complex manner which is less suitable for forecasts based on the concepts of seismic gaps.

While the concept of a seismic gap probably is applicable to shocks smaller than magnitude 7, we do not consider moderate shocks in this paper. The resolution in defining rupture zones from either the distribution of teleseismically-located aftershocks, felt reports or tsunami data is no better than a few tens of kilometers. Hence, the size of the smallest rupture zones that can be defined with confidence from data of those types is equivalent to an event of magnitude near 7. In areas where surface breakage is common or a local network has existed for a long time (as along parts of the San Jacinto fault zone in southern California, THATCHER *et al.*, 1975), seismic gaps can be defined for moderate-size earthquakes.

In subduction zones great earthquakes and at least some major shocks appear to rupture that portion of the interface between the two plates from the surface to a depth of about 40 km (DAVIES and HOUSE, 1979). In great events like the 1964 Alaskan and 1960 Chilean earthquakes, aftershocks extend from the surface at a point near the trench to a depth of about 40 km beneath the arc. Below this depth fault plane solutions of the thrusting type are less common; many of the solutions exhibit either down-dip tension or compression parallel to the downgoing lithospheric slab. They are inferred to represent stresses in the interior of the downgoing plate (ISACKS and MOLNAR, 1971). This transition from intraplate to interplate motion may not be sharp and the thickness of the lithosphere varies among the methods used to define it. One common assumption that is often made in analyses of seismic gaps is that large shocks rupture the entire plate boundary in depth. There are, nonetheless, a few examples where the deeper part of the plate boundary ruptures during a shock (usually a major and not a great event) and the shallow portion ruptures in a second later event. Rupture in great shocks is commonly initiated at a depth of about 40 km and usually near one of the ends of a seismic gap. Rupture then propagates both up dip and along the strike (KELLEHER *et al.*, 1973) of the interface.

As pointed out later in the text, the forecasts made here are for a specific set of assumptions, types of plate boundaries and range of magnitudes. The paper does not attempt to forecast events smaller than magnitude 7, large shocks of intermediate depth, earthquakes involving normal faulting in the bottom or along the outer (seaward) walls of deep-sea trenches, events adjacent to a nearby plate boundary as in Central Alaska, shocks along a multiple plate boundary (for example, along the Hayward fault in the San Francisco Bay area adjacent to the San Andreas fault), or seismic events along the set of grabens that are often found along the chain of volcanoes near subduction zones. Large shocks are less common in those areas than they are along the main plate boundaries. Nevertheless, earthquakes of those types, including ones of only moderate size can be destructive when they occur near centers of population.

If an area is identified as a seismic gap in this paper, this means that the region in question has not been ruptured by a large shock in the last 30 years or more. This identification should not be interpreted to mean, however, that the site necessarily has a high potential for being the location of a large shock during the next few decades. Other additional information, such as the seismic history or tectonic regime, are needed before the area can be designated as one of high seismic potential. In this paper major plate boundaries are classified into six categories of seismic potential. Areas along plate boundaries of lowest assigned potential for the next few tens of years (category 6) are those that ruptured in large shocks during the last 30 years. Of course, after many decades regions now assigned to category 6 will have a higher potential. Gaps assigned the highest potential (category 1) are those in which a great event is known to have occurred in the area but not within the last 100 years. Gaps that were the sites of large shocks more than 30 but less than 100 years ago (category 2) also appear to have a high potential for the occurrence of future large shocks. Many of the latter gaps, however, probably will not be sites of large shocks for the next few tens of years since a long time is needed to build up high tectonic stresses. In some places where the repeat time appears to be about 40 years as in parts of

Middle America, however, large shocks may occur in areas designated as category 2 within the next few tens of years. Hence, for categories 1, 2 and 6 the potential for large shocks within the next few decades is fairly clear.

The potential for large shocks is not as clear for the regions designated in categories 3, 4 and 5. These areas have either an ambiguous history of great earthquakes or no record of large shocks. For categories 3, 4 and 5 we resort to assessing the potential by comparing them with other regions that appear to have a similar tectonic setting and, we presume, a similar seismic potential. It should be remembered, however, that designations in these three categories are more hypothetical and that much work remains to be done in checking the validity of the bases for those hypotheses. It does appear nonetheless, that certain parts of plate boundaries are not typified by the occurrence of great shocks and that they probably will remain areas of low or negligible potential for thousands of years.

In the future it may be possible to parameterize the seismic potential according to more physically understandable quantities such as the long-term rate of plate motion, repeat time of large shocks or the configuration of the interface between the two interacting plates. At the present time the uncertainty in these quantities, especially in repeat time, is large enough that we have simply used cutoffs of 30 and 100 years in assigning regions to specific categories.

We find a relatively small number of regions that are assigned the highest seismic potential, i.e., category 1. In many of those areas, however, local seismic networks and strong-motion instruments are either non-existent or nearly so. Many of these gaps are located largely in submarine areas adjacent to land masses. These gaps clearly deserve high priority for instrumentation and intensive study both on land and at sea.

## 2. *Map of seismic potential for major plate boundaries*

### (*a*) *Six categories of potential*

Figure 1 displays our conclusions about the relative seismic potential for the major plate boundaries of the Pacific, Caribbean, Indonesian and South Sandwich areas. Before discussing these categories, it is important to remember that the forecasts are made subject to several assumptions and limitations which are as follows:

1. only shallow earthquakes of magnitude 7 or greater are considered or forecast;
2. some future events of magnitude near 7 may fail to be forecast because the error in mapping aftershock zones and rupture zones is comparable to the size of the rupture zone of an event near magnitude 7;
3. earthquakes of intermediate and deep focus are not considered;
4. only simple plate boundaries of the thrust (convergent) or transform type are considered;

Figure 1

Seismic potential for events with $M \geq 7$ for the next few decades along certain major plate boundaries. The six categories presented are based on historic and tectonic criteria. Shaded areas are those portions of plate boundaries about which we have the most data, historic and/or tectonic, and hence the most confidence in our evaluation. Dark areas (category 1) have not ruptured in a great earthquake in over 100 years, and are considered likely candidates for major or great shocks within the next decade or few decades. Similarly, lightly shaded areas (category 6) have ruptured in a great earthquake within the past 30 years, and are considered to have the lowest seismic potential for the present time. Some regions that are not shaded (categories 3, 5) are areas where more information is needed in order to more accurately assess their seismic potential. Areas that are cross-hatched (category 4) are characterized by plate motion subparallel to the arc, and appear to be tectonically similar. The seismic potentials presented on this map are meant as general forecasts, not specific predictions of the time of occurrence.

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5. events near a major plate boundary as in central Alaska or along the grabens and volcanic zone of Central America are excluded;
6. shocks involving normal faulting of the down-bent lithosphere along the outer walls of trenches or in the bottoms of deep-sea trenches are excluded;
7. zones of multibranched deformation (i.e., 2 or more major subparallel faults) are not considered;
8. large shocks are assumed to rupture the plate boundary from about 0 to 40 km in depth;
9. a second large event cannot occur along the same plate interface for many decades after a large earthquake until the stress is slowly built up again by plate movements;
10. the rupture zone is accurately reflected by the extent of the aftershock zone, by the extent of the area of intense shaking or damage, or by the area of tsunami generation. One case is known (Nankaido, Japan, 1946) in which the rupture zone is larger than the area inferred from aftershocks;
11. the forecasts are valid only for the next few tens of years;
12. the designation of high seismic potential is not a prediction in the sense that a precise origin time of a future shock is estimated. Each segment of these plate boundaries is assigned to one of six categories.

The categories defined below are designed to reflect our assessment of: (1) the relative seismic potential of the region; (2) the completeness of its seismic history; (3) our understanding of the tectonic regime. The areas thought to be of the highest potential are assigned to category 1; those of successively lower potential are assigned higher numbers up to six. The categories of seismic potential are as follows:

1. The region (portion of a plate boundary) has experienced at least one large shock in the historic past with the most recent event occurring prior to 1879, i.e., more than 100 years ago. This category represents the highest seismic potential.
2. The region has experienced at least one large shock in the past with the most recent event occurring between 1879 and 1949, i.e., more than 30 years ago, but less than 100 years ago.

3. The region has an incomplete history of large earthquakes. No historic event is clearly documented as having ruptured the plate boundary. There is no evidence, however that would indicate that the region may not be the site of a future large earthquake. A comparison of the tectonic framework with that of other areas known to be sites of historic large shocks may also suggest that the region is capable of being the site of a future large shock.
4. Motion between the plates is parallel or nearly parallel to the local strike of the subduction zone (trench). This category applies to the Puerto Rico-Virgin Island region, the Commander Islands in the westernmost Aleutians, and the Andaman-Nicobar region in the Indian Ocean. All appear to have a similar tectonic setting. A resolution of the question of seismic potential for one area may be useful in assessing the potential for the other two.
5. The region does not have a history of great earthquakes. Several tectonic hypotheses, which are proposed by various investigators, suggest that these regions will not be the sites of great shocks in the future.
6. The region has been ruptured by a large earthquake during the last 30 years (since 1 January 1949). This category is considered to represent the lowest seismic potential for the next few decades.

Regions assigned to category 1 or 6 are from our most clearly defined data set. In each case the regions have clearly ruptured during a large quake. Those in category 1 have been accumulating strain energy for at least 100 years. They are believed to be highly stressed and should be considered likely sites for great earthquakes in the next few to few tens of years.

We have used 100 years as an arbitrary cutoff for category 1. Repeat times of large shocks, even in areas with a long history, vary enough that they do not appear to be useful for forecasts of time of occurrence to better than a factor of about 1.5 to 2. The repeat time might be expected to be a function of the dip of the interface between plates, the age of the interacting lithosphere and the rate of relative plate motion. As yet these factors have not been taken into consideration in a quantitative sense. Most of the plate boundaries considered here have relative movements between 5 and 12 cm/year. Repeat times might be expected to be longer for boundaries with slow relative motion, such as the Caribbean plate where the relative motion is about 2 cm/year. The motion of the Philippine Sea plate relative to surrounding plates is also poorly known. Nevertheless, since the repeat time is poorly determined for most areas, some regions in category 1 may still not be sites of large shocks for many years to come.

None of the regions (Puerto Rico-Virgin Islands, Commander Islands, Andaman-Nicobar Islands) in category 4 has experienced a great shallow earthquake, during this century, that has ruptured an extensive portion of the plate boundary. The seismic potential for great earthquakes in these regions should be considered poorly constrained; major events are more likely. The physical mode of plate motion is also not well understood for these three areas. Focal mechanism solutions indicate thrust faulting along nearly horizontal planes with the slip vector nearly parallel to the trench (ISACKS, OLIVER and SYKES, 1968; MOLNAR and SYKES, 1969; CORMIER, 1975).

Some of the regions in category 2 may be the sites of large shocks in the next few years to tens of years as their recurrence times may be as short as 30 to 50 years (i.e., Central America, Solomon Islands). The record of large earthquakes for the regions in category 3 is such that historic reports are absent, too short in length, or yet to be researched. They may have a seismic potential in categories 1, 2 or 5 when more analysis is performed.

Category 5 also includes regions that have not been the site of a great earthquake in recorded history. In each case there are reasons for inferring that the area may never be the site of a great shock. These inferences stem from either the presence of a tectonic regime along the plate boundary that is thought to be the site of only occasional large shocks or the region is thought to be experiencing some modification of the subduction process such that large shallow shocks are rare or absent. A reduction in the size and number of large shocks is commonly observed in many regions where aseismic ridges and other bathymetric highs on the seafloor encounter subduction zones (KELLEHER and MCCANN, 1976). Category 5 includes subduction zones where the density contrast between the two interacting plates is thought to be unusually large, and the plates may be moving aseismically (MCCANN and NISHENKO, 1978; FRANKEL and MCCANN, 1979). These regions correspond, in general, with the 'Mariana type' of subduction of UYEDA and KANAMORI (1978). The density contrast hypothesis, the decoupling idea of KANAMORI (1971, 1977a) and the inference that the presence or absence of back-arc spreading plays an important role in the state of stress and mode of occurrence of large earthquakes need further investigation. Therefore, assignment of regions to category 5 should be considered temporary pending more detailed research on these hypotheses and on their possible role in governing the occurrence of large earthquakes. In several areas, like the Marianas, each of these hypotheses predicts that few if any great shocks are to be expected in the future.

### (b) *Other phenomena shown on map of seismic potential*

*Tsunami.* In Fig. 1 wavy lines are used to denote a possible tsunami risk in those regions in category 1 that have experienced an historic, destructive tsunami. This is not to indicate that these areas will have a destructive tsunami associated with the next great shock but only that a potential exists. Conversely, those regions on convergent plate boundaries with no tsunami potential indicated, nevertheless, may produce tsunamis during future large earthquakes.

*Focal Mechanism.* Typical focal mechanisms are presented (lower hemisphere projection, where shaded areas denote compressions) for shallow earthquakes along some of the major plate boundaries in Fig. 1. These solutions indicate the probable sense of motion during the next large earthquake on a particular plate boundary.

*Volcanoes.* Triangles are shown in Fig. 1 where active volcanoes occur in conjunction with the subduction process. These volcanoes may show fluctuations in activity before the next nearby great earthquake as discussed by NAKAMURA (1975), CARR (1977), and KIMURA (1978).

*Successful Forecasts.* Stars are placed in Fig. 1 on those segments of plate boundaries that experienced large shocks after the region was cited in the literature as a major seismic gap. Some of the successful forecasts are indicated in Table 1.

Table 1

| Forecast | | Earthquake | | |
|---|---|---|---|---|
| Investigator | Year published | Location | Year | Magnitude |
| Fedotov | 1965 | near Hokkaido, Japan (Tokachi-oki) | 1968 | 8.2 $M_\omega$ |
| Fedotov | 1965 | near southern Kuril Islands | 1969 | 8.2 $M_\omega$ |
| Fedotov | 1965 | near central Kamchatka | 1971 | 7.8 $M_s$ |
| Fedotov<br>Mogi | 1965<br>1968 | near Hokkaido, Japan (Nemuro-oki) | 1973 | 7.7 $M_s$ |
| Sykes | 1971 | near Sitka southeastern Alaska | 1972 | 7.6 $M_s$ |
| Kelleher | 1972 | near Lima, Peru | 1974 | 8.1 $M_\omega$ |
| Kelleher and others | 1973 | near Colima, Mexico | 1973 | 7.5 $M_s$ |

## 3. *Historic review and basic concepts*

### (*a*) *Initial research on seismic gaps*

The first work on seismic gaps appeared in 1965 by S. A. Fedotov (Fig. 2). His study covered a portion of the seismically active margin of the northwest Pacific extending from central Japan to the Kamchatka Peninsula near the Commander Islands. His examination of the occurrence of great ($M_s \geq 7\frac{3}{4}$), shallow earthquakes led him to conclude that great earthquakes tend to occur in regions that were not the sites of great earthquakes for at least several decades. He proposed that the rupture zones of large shocks were delineated by aftershock activity and observed that the aftershock zones of nearby large earthquakes tend to abut without overlapping. FEDOTOV (1965) also specifically mentioned several areas that are probable locations for future large earthquakes. Since 1965 several of the seismic gaps he delineated near the southern Kuril Islands and central Kamchatka were 'filled' by the occurrence of recent large earthquakes (see Table 1).

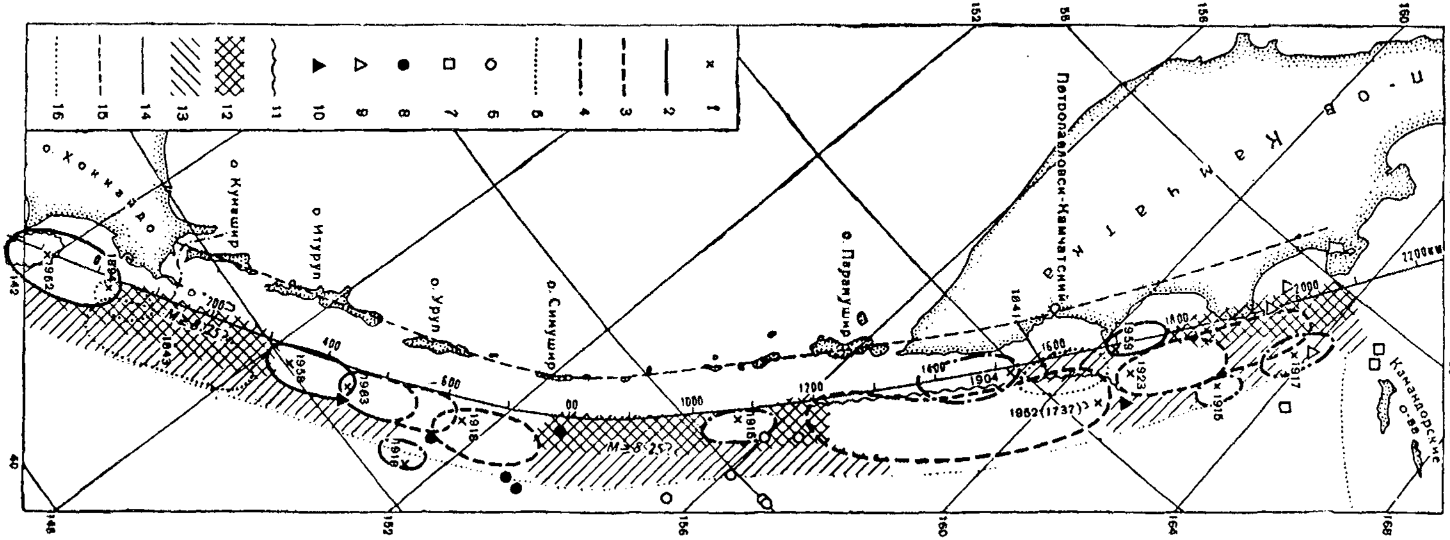



Figure 2

Focal regions of great (M ≥ 7¾) earthquakes in the Kuril-Kamchatka region from 1904 to 1963 and the probable areas of future earthquakes of that strength (from FEDOTOV, 1965). (1) instrumental epicenters of great earthquakes; (2) borders of the focal regions of great earthquakes; (3) uncertain part of the boundary or possible variations in the outline of the focal region; (4) possible focal region; (5) proposed (probable) focal regions of the strongest earthquakes of the past century, aftershocks of earthquakes; (6) 1 May 1915; (7) 30 January 1917; (8) 7 September 1918; (9) 3 February 1923; (10) strong foreshocks of the 1923 and 1963 earthquakes; (11) outline of the region of tsunami origin; (12) the most probable locations of future earthquakes with M ≥ 7¾; (13) less likely locations of future earthquakes with M ≥ 7¾; (14) distance along the Kuril-Kamchatka trench; (15) axes of the trench; (16) axes of the Kuril-Kamchatka volcanic belt.

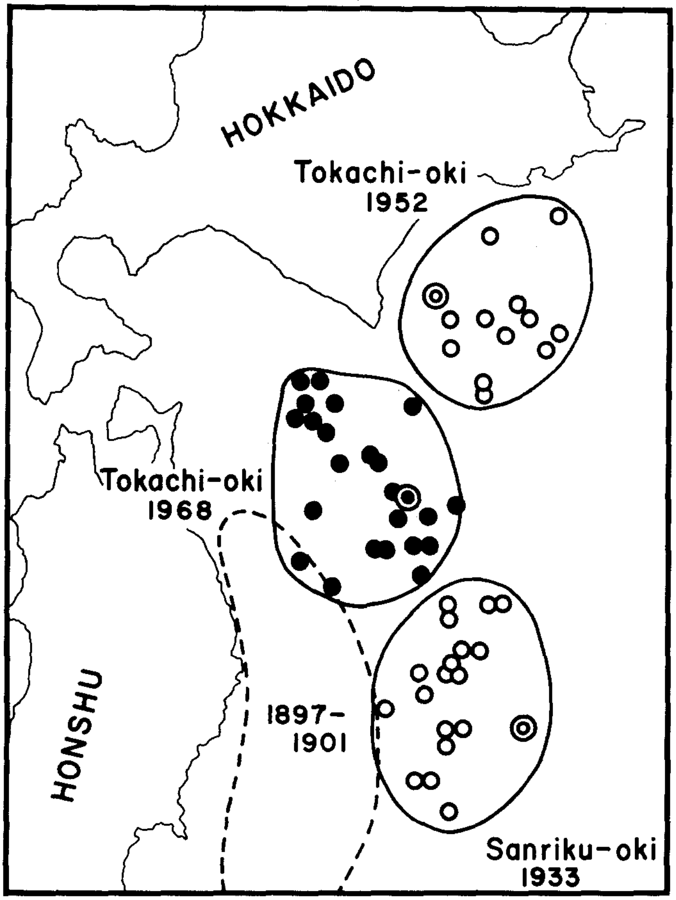



Figure 3

Large earthquakes near northern Japan (after MOGI, 1968 and SYKES, 1971). Double circles are epicenters of main shocks, single circles are aftershocks which define the rupture area (solid line). The dashed line marks the zone affected by a series of shocks occurring between 1897–1901. No reliable instrumental data are available for these events. Note how the rupture areas of the 1968 and 1952 zones abut as do the 1968 zone and that inferred for the 1897–1901 events. The 1933 Sanriku event is not related to the underthrusting of seafloor beneath the Japanese arc.

Figure 3 (from MOGI, 1968 and SYKES, 1971) shows an example of how aftershock zones near Japan tend to abut rather than overlap. Double circles are locations of main shocks; single circles represent aftershocks. The 1968 and 1952 zones abut, as do the 1968 and 1897 zones. The 1968 and 1952 zones were seismic gaps since the turn of the century. The 1968 zone was specified as a gap by FEDOTOV (1965). The 1933 Sanriku event was not associated with underthrusting beneath the Japanese arc but involved normal faulting in the deepest part of the Japan trench (KANAMORI, 1971).

ALLEN *et al.* (1965), TOBIN and SYKES (1968) and MOGI (1968, 1969) also discuss ideas similar to those proposed by Fedotov. In these papers, however, smaller magnitude events are considered. They find that small-scale earthquake activity is generally very low for several years, perhaps decades, before a large event occurs in a seismic gap. Thus, seismic gaps often appear to be regions of reduced activity for moderate activity and small shocks as well as regions of relative quiescence for large shocks. MOGI (1969) notes that activity often increases in a ring-shaped region immediately surrounding a quiescent zone, that coincides with the rupture zone of its coming large shock. Together these form a 'donut' pattern of small and moderate earthquakes. The level of seismic activity before great earthquakes is examined in detail by KELLEHER and SAVINO (1975); their conclusions generally confirm the observations of low activity in areas that later rupture in large earthquakes. They do find, however, forerunning activity near the epicenter of the coming shock and in the region surrounding its rupture zone as found by MOGI (1969).

KELLEHER (1970) examined the spatio-temporal relationship of large earthquakes in the Alaska-Aleutian area, among others. Several areas exhibited migrations of activity along their seismic zones. In some cases the main shock of a large earthquake occurred at one end of the aftershock zone and fault rupture progressed unilaterally along the seismic zone. In many cases the main shock occurred near the aftershock zone of a former large event and rupture propagated away from that aftershock area.

SYKES (1971) examined the seismic history of the Alaska-Aleutian seismic zone using plate tectonic theory. Aftershocks of large events were relocated by computer to increase the location accuracy. He also finds that aftershock zones of large earthquakes do not overlap but tend to abut and that large events tend to occur in regions that were quiescent for several decades prior to the main shock. In these nearly quiescent zones strain energy, which appears to result from the large-scale motion of rigid lithospheric plates, has been accumulating for tens to hundreds of years prior to the occurrence of a large shock. If the plates are to maintain their rigidity on a time scale of thousands of years, large events must occur in these gaps and release the strain energy that is stored.

*(b) Criteria for determining seismic gaps*

Expanding on the work of SYKES (1971) and KELLEHER (1972) for individual subduction zones, KELLEHER *et al.* (1973) analyzed seismic zones along several regions of plate convergence (subduction) and transform faulting. Their study included the northern, eastern and western margins of the Pacific and most of the border of the Caribbean plate. It was directed solely at large, shallow, interplate events, i.e., shocks in which rupture involves the relative motion of two lithospheric plates.

The initial criteria used by KELLEHER *et al.* (1973) to determine a seismic gap were:

1. 'The segment is part of a major seismic belt characterized predominantly by strike-slip or thrust faulting.'
2. 'The segment has not ruptured for at least 30 years.'

These criteria appear to be very simple, but large shocks occurring since 1973 show that they are very effective tools for determining possible sites of future large events. Several segments of the plate boundaries studied by KELLEHER *et al.* (1973), however, have no historic record of ever being affected by a great thrust earthquake. Hence, either the historic quiescence is representative of a long-term stationary process or it is a temporary phenomenon in which the repeat times of large shocks exceed the length of the historic record. It is necessary, then, to distinguish among different types of seismic gaps; supplementary criteria must be developed to indicate whether a given gap either has a record of previous large shocks or some other evidence indicates that large events could be expected to occur there. These supplementary criteria include:

1. an historic record of one or more large earthquakes for the region in question;
2. the recurrence interval for large events in a given segment is nearly equal to the time interval since the last large event and hence a large event could be expected;
3. the site appears to be the next event in a series of earthquakes in a regular space–time progression of large events along a plate boundary; and
4. the region is tectonically similar to other areas in which large shocks either have or have not occurred in the historic record.

As will be discussed later, supplementary criterion 1 is not trivial, as several regions not meeting this criterion appear to be similar tectonically. Some of these regions probably will be the source regions of future great earthquakes, while others may be permanently aseismic for large shocks. Some possible tectonic criteria are developed later in an attempt to distinguish these two types of gaps.

### (*c*) *Sizes of future shocks and the geometry of the subduction zone*

KELLEHER *et al.* (1974) find that for many of the subduction zones of the Pacific the source dimensions of the large shallow earthquakes have a characteristic maximum length. This characteristic maximum appears to be strongly influenced by the geometry of the interface zone between the two interacting plates. It appears to be particularly influenced by the down-dip width of the interface between the overthrust and underthrust slabs of lithosphere. KELLEHER *et al.* (1974) find that variations in the inferred width, $w$, of the interface, as defined by the zone of shallow ($<70$ km) hypocenters, correlate with source size (magnitude and dimensions) of many large shallow earthquakes of this century. Regions with wide interfaces ($w \geq 100$ km) are the source areas for great thrust earthquakes in which the maximum lengths of ruptures along a subduction zone may exceed 400 km. Along transform fault zones this interface is nearly vertical and interplate seismic slip extends to a depth of about 15 km. Along subduction zones the interface has a shallow dip and appears to extend to a depth of about 40 km (rather than 70 km as assumed by KELLEHER *et al.*, 1974). This depth is ascertained from the depths of aftershocks of great earthquakes and the transition from interplate thrust faulting found at shallower depths to deformation within the downgoing plate at greater depths (DAVIES and HOUSE, 1979).

Figure 4 presents a diagrammatic model of a subduction zone. Oceanic lithosphere (lower plate) approaches the trench from the upper left-hand corner of the diagram. Upon entering the trench region the oceanic plate is deflected or bent downward beneath the overriding plate. Normal faults that develop on the oceanic plate result from this deformation. The zone of contact (i.e., the interface) between the overthrust and underthrust plates accumulates energy in the form of elastic strain during the long interval between great earthquakes. When the shear stress across the boundary exceeds the static friction on the fault, rupture occurs. Rupture initiates at the hypocenter of the main shock (largest star pattern), its epicenter (largest ellipse) is the position of the hypocenter projected onto the surface of the overthrust plate. Motion initiated at the hypocenter of the main shock then generally spreads both up dip and along the strike of the zone of contact between the two plates. Aftershocks (smaller

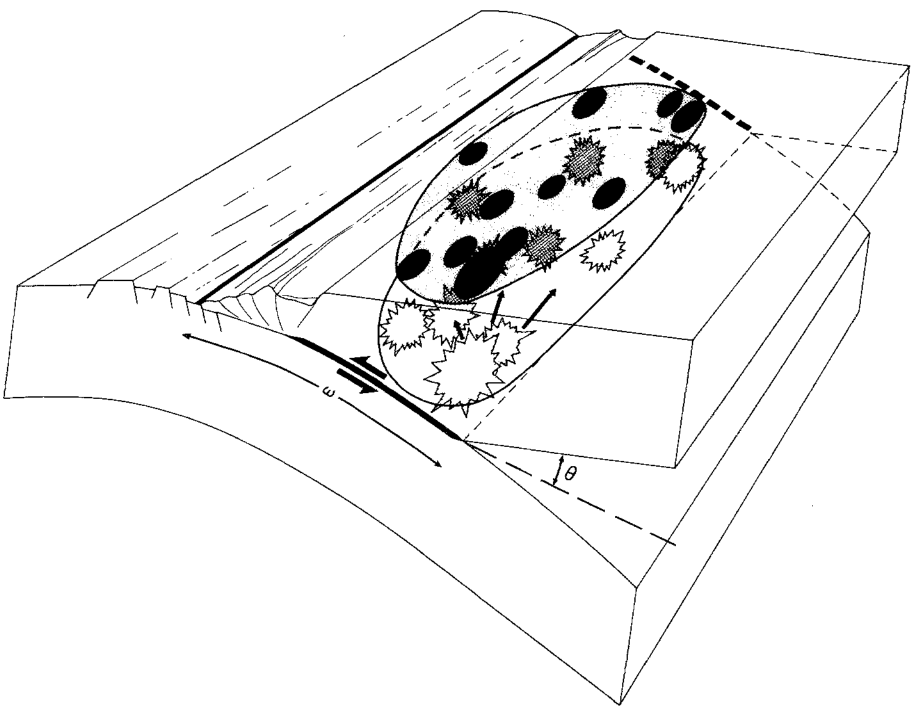



Figure 4

Typical rupture propagation during a large thrust earthquake along a convergent plate boundary. Hypocenter of the main shock (largest hatched star) is landward and usually deeper than most aftershocks (smaller hatched stars). Rupture propagates upward and outward (solid arrows) along the plate interface (thick black line). For some great earthquakes the rupture propagates hundreds of kilometers laterally along the plate interface (dashed line). Epicentral locations of aftershocks (solid ellipses) define the surface projection of the rupture zone (shaded area). The width of the interface ($\omega$) coincides with the zone of shallow earthquakes and is strongly influenced by the dip of the underthrust slab ($\theta$) and the direction of convergence.

star patterns and ellipses) occur along the rupture zone during the first few weeks or months after the main shock. As these aftershocks represent readjustments occurring on the ruptured fault surface, the spatial extent of the aftershock activity can be used to infer the configuration of the rupture zone of the main shock (shaded area). The relationships among the width of interface ($\omega$) the width of the zone of shallow earthquakes and the angle of dip ($\theta$) of the shallow portion of the underthrust lithosphere are also illustrated. It can be seen that a wide zone of shallow earthquakes results from either a low angle of dip or an increased thickness of the overthrust plate of lithosphere.

Differences in the width of the interface are inferred using the epicenters of shallow earthquakes. The width of the zone of shallow earthquakes is taken to be the distance between the trench axis and the contour defined by hypocenters 70 km deep. The 70 km contour is determined using the locations of hypocenters of both shallow ($<70$ km) and intermediate ($>70$ km) depth. It is employed here as it was by KELLEHER *et al.* (1974) since calculated depths are subject to greater errors for foci near 40 km. The upper limit of the zone of plate contact occurs either near the axis of the trench or along the inner wall of the trench. In detail, regional variations in the development of the arc, such as the presence of an accretionary sedimentary wedge, may influence the dimensions of rupture zones. In general, however, the aftershock zones of great earthquakes extend to near the trench axis. This information is an important aid in understanding the generation of large events at thrust boundaries, for if the width of interface is reasonably well known, then estimates of the largest possible shock in any particular gap can be made.

### (*d*) *Permanent gaps*

Returning now to supplementary criterion 1, KELLEHER and MCCANN (1976, 1977) find that the distribution of large shallow earthquakes along subduction boundaries does not always agree with the distribution pattern that might be predicted from a simple model derived from plate tectonics. For extensive sections of some island arcs large shocks have occurred infrequently or not at all during recorded history. Most of these zones of long-term quiescence (category 5) are nearly coterminous with segments of plate boundaries where groups of seamounts, aseismic ridges or other topographic features on the underthrust lithospheric slab appear to be interacting with and modifying the subduction process. Besides a decrease in the level of large shocks, other manifestations of a modified subduction process include gaps and offsets in the line of active volcanoes, wide diffuse zones of deformation or abrupt jumps in the plate margin, infrequent shallow earthquakes with thrust-type mechanisms and the diminution or disappearance of hypocenters at intermediate depths. In contrast, nearly all segments of subduction zones landward of smooth, low-lying ocean floor have a history of at least one large shock.

Differences in the seismic and tectonic regimes among various island arcs suggest that the lack or near lack of large shocks in some regions probably results from specific differences in the tectonic framework and not from the brevity of the seismic record. KELLEHER and MCCANN (1976) hypothesize that these variations are related to the density contrast between the overthrust and underthrust lithospheric plates. Major topographic rises and seamount provinces, for example, represent regions with lithospheric densities intermediate between those of typical oceanic and continental lithosphere. Buoyant material on the downgoing plate may lead to marked changes (and in the most extreme case possibly even to a cessation) in the subduction process. Variations in geological and geophysical parameters suggest that this buoyant interaction is also capable of producing modifications in the subduction process at deeper levels.

These complexities in the subduction process, which appear to affect the occurrence of large shocks, make it necessary to use the terms gap and seismic potential with care. We use the term seismic gap to refer to any region along an active plate boundary that has not experienced a large thrust or strike-slip earthquake for at least a minimum period of time, which here is taken to be 30 years. In most regions 30 years is the minimum time between the recurrence of large events; in many active zones the repeat time is 100 to 300 years. The term seismic gap does not imply the potential for a large event to occur in that region. A region of seismic potential, however, is here taken to be a seismic gap that, for tectonic or historic reasons, is thought to be capable of producing large shocks in the future.

### (e) *Aseismic slip*

Quiescence for great earthquakes and near quiescence for moderate-size shocks is observed along several arc segments that are not presently subducting bathymetric features. Each of these regions subducts relatively old ($>60$ m.y.) seafloor beneath a young island arc reared by an active marginal basin and has been assigned to category 5.

UYEDA and KANAMORI (1978) propose that the presence or absence of an active marginal basin on the overthrust plate may be responsible for the observed variation in the number of large shocks occurring along various convergent plate boundaries. They delineate two modes of subduction using criteria based on the spreading activity in the region behind the arc. They associate Chilean-type subduction with the occurrence of great, shallow earthquakes, the absence of spreading centers behind the arcs, and little or no aseismic slip on the thrust boundary. Mariana-type subduction is characterized by a lack of great earthquakes along the thrust boundary at shallow depths, and active back-arc spreading, with a substantial portion of the relative motion between the plates possibly occurring aseismically. UYEDA and KANAMORI (1978) propose two possible mechanisms for the presence of marginal basins and the absence of great earthquakes on the thrust boundary.

1. the contact between the converging plates changes, through an evolutionary process, from tight coupling to decoupling;
2. if the subducted slab is anchored at its bottom to the mantle, a retreating overthrust plate would create marginal basins near the trench; an actively converging overthrust plate would not lead to the development of marginal basins.

At present it is not possible to choose among these models of aseismic and seismic behavior. Aseismic slip appears to play a larger role in plate motion than was generally assumed a few years ago (KANAMORI, 1977a). Hence, these models deserve more attention. The occurrence or non-occurrence of large earthquakes along convergent plate boundaries may be strongly influenced not only by the density contrast between the plates but also by the absolute plate motions and by the mode of interplate coupling. Until we understand why large shocks do not occur along some parts of plate boundaries, it is not possible to accurately forecast whether some areas that have been aseismic for the last 100 to 300 years will remain so in the future.

### (*f*) *Basins on the upper slopes of trenches, deep-sea terraces, and large earthquakes*

Variations in the structures on the inner wall of trenches appear to reflect changes in both the lengths of rupture zones and in the source areas of tsunamis that are associated with large shallow earthquakes (NISHENKO and MCCANN, 1979). Crustal deformation that occurs during large earthquakes and related tsunamic activity play important roles in the development and maintenance of topographic features on the inner walls of deep-sea trenches. A regional comparison of the average length of upper slope basins and terraces with the maximum length of earthquake rupture zones shows that longer basins and terraces characteristically occur in regions with larger rupture zones. Examples in Japan, Alaska and the Aleutians clearly show how these topographic features reflect the size and spatial distribution of seismic-tsunamic source areas. In many cases, this relationship may be explained by the coseismic reactivation of structural units on the inner wall of the trench. The two principal areas of coseismic reactivation are the trench slope break and the frontal arc region. Thus, upper slope basins, deep-sea terraces and other topographic features may serve as indicators of the tectonic regime and seismic-tsunamic risk along convergent plate margins. Also, gross features of the seismic regime (i.e., sizes of rupture zones) for at least the last tens of thousands of years appear to be recorded on the inner wall of deep-sea trenches. This gives us a valuable tool for further studying the seismic interactions between plates.

Variations of the dimensions of rupture zones also appear to be influenced by the block-like behavior of the overthrust plate (CARR *et al.*, 1974; ANDO, 1975; SPENCE, 1977). In many instances, the dimensions of upper slope basins and terraces are nearly equivalent to those of the crustal blocks. Regions with basin or terrace lengths greater than 100 km have histories of large ruptures involving adjacent segments of the arc (i.e., rupture lengths of several hundred kilometers). Regions with basin or terrace dimensions less than 100 km tend to rupture independently of adjacent segments (NISHENKO and MCCANN, 1977).

### (g) *Temporal-spatial relationships between volcanism and large thrust earthquakes*

NAKAMURA (1975) studies the short-term relationship between volcanic eruptions and the occurrence of nearby large earthquakes. He finds that volcanic activity may increase during the years immediately surrounding the time of large shocks. The study of individual volcanoes, however, has not yet been shown to be a useful tool for forecasting the location of large shocks. CARR (1977) examines regional volcanic activity before several great earthquakes. He observes a marked reduction in volcanic activity about ten years before a main shock. KIMURA (1978) relates long-term volcanic activity and its temporal-spatial relationship to great earthquakes of the thrust type. Volcanoes standing landward of converging plates show increased eruptive activity up to 30 years before several main shocks. The volcanic activity, which is often characterized by lava flows or effusion of great amounts of volcanic material, either ceases or markedly decreases after the occurrence of large shocks. In some cases volcanic activity decreases before the earthquake occurs. After the occurrence of the earthquake, volcanic activity in neighboring seismic gaps tends to increase. This relationship between volcanic activity and the occurrence of large shocks is most clearly seen along the arcs of the northern Pacific (e.g., Aleutians, Kamchatka, Kuriles) and suggests that the timing of regional volcanic activity is at least in part controlled by regional tectonic movements associated with the generation of large earthquakes of the thrust type. Hence, changes in volcanic activity should be carefully monitored and studied along some of the plate boundaries we identify as having the greatest seismic potential for the occurrence of future large shocks.

## 4. *Discussion of seismic regions*

### (a) *Alaska-Aleutians*

The Alaska-Aleutian seismic belt forms part of the northern and eastern boundary of the Pacific plate. Right-lateral strike-slip motion occurs along the Fairweather and Queen Charlotte transform faults as the Pacific plate moves northward relative to the North American Plate (TOBIN and SYKES, 1968). In southern Alaska, the Pacific sea-floor underthrusts the Alaska peninsula and Aleutian Islands, but the direction of convergence becomes more oblique to the arc and trench in the central and western Aleutians. In the Commander Islands, at the far western end of the Aleutian chain, the motion between the Pacific and North American plates is nearly parallel to the axis of the trench (CORMIER, 1975).

Figure 5 shows the rupture areas of great earthquakes along the Alaska-Aleutian seismic zone since about 1920. The seismic history of this region is only complete to about the turn of the century. Nevertheless, the instrumental record indicates that the release of seismic energy in the region is dominated by infrequent, great earthquakes that rupture extensive (> 500 km) portions of the plate boundary (KELLEHER, 1970; SYKES, 1971). Three great earthquakes (1957, 1964 and 1965) have recently ruptured most of the zone of underthrusting along the arc. One seismic gap lies at the western end of the Aleutian arc.

No large earthquakes have been recorded on the thrust boundary of the western end of the Aleutian arc during this century. A large earthquake did occur in this region in 1849 (MEDVEDEV, 1968) and was accompanied by a locally destructive seawave. An independent source (ELLA, 1890) indicates that this event probably occurred on the southern side of the island chain as a seawave was observed in the Samoan and New Hebrides Islands in the southwest Pacific about ten hours after the main shock. A second line of evidence also points to a source to the south of the islands. Events occurring on the northern flank of the island arc typically have strike-slip mechanisms with little or no vertical motion (CORMIER, 1975). It is, therefore, unlikely that a large event on the northern flank of the island chain would produce a large seawave capable of being observed in the southwest Pacific. Although we cannot be sure that this event was not one involving normal faulting in the trench, the sketchy historical record does indicate that this region might be capable of storing enough strain energy to generate large destructive earthquakes with recurrence times of 130 (1979–1849) years or more. This region is assigned to category 4. This category has been assigned to three regions that have slip vectors nearly parallel to the trench and that show evidence of a downgoing seismic zone.

The second gap lies between the 1957 and 1964 rupture zones. Important activity near this gap includes a great earthquake in 1938 ($M = 8.2$). The western limit of the aftershock zone of the 1938 event is uncertain, as only a few poorly located events lie to the west of the main shock. The existing aftershock data suggest that only about half of the plate boundary between the 1964 and 1957 shocks ruptured in 1938. Hence, that region appears to be a major gap. This area has been noted by KELLEHER (1970) as the possible location for the next large event in a space–time progression of large events moving north and then west along the Alaskan coast (1949, 1958, 1964).

Historically only one large shallow event is reported for this region. It occurred in 1788 and was associated with a locally destructive tsunami on Unga Island in the Shumagin Islands and on Sanak Island about 200 km to the southwest (COX and PARARAS-CARYANNIS, 1976). The intensities reported for this event, however, only reached Modified Mercalli intensity V (MEYERS *et al.*, 1976). The intensity and tsunami damage associated with this event are similar to the macroseismic effects of the 1946 earthquake. Although the 1946 event was only moderate in size ($M_s = 7.4$), it was associated with an extremely destructive seawave (EPPLEY, 1965). More thorough

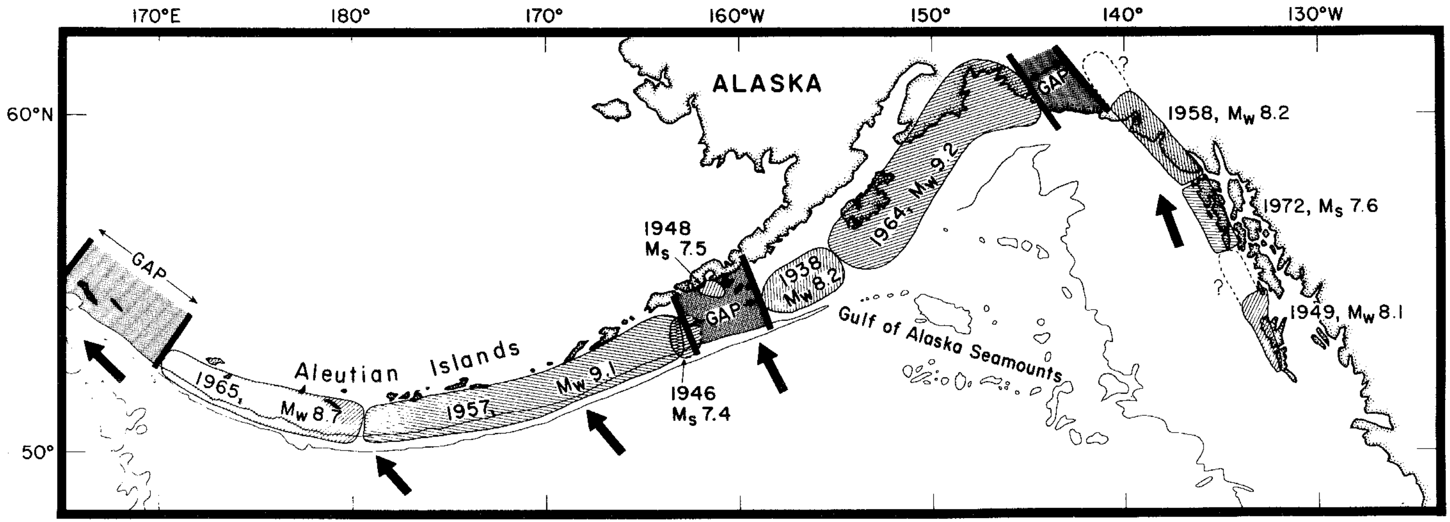



Figure 5

Recent large earthquakes of the Alaska-Aleutian seismic zone. Note that the rupture zones (hatched regions) tend to extend for several hundred kilometers. Year and magnitude of the events are shown. Arrows give direction of motion of the Pacific plate relative to North America (MINSTER *et al.*, 1974). Three large gaps are stippled regions. Lines in Gulf of Alaska are 2000 fathom contours.

documentation of the 1788 event is necessary. Until those data are collected, however, the region between the 1957 and 1938 rupture zones should be considered one of high risk and possibly the site of the next large earthquake in a space–time progression.

Recent work in this region has focused on two earthquakes that occurred in 1974. ARCHAMBEAU (1978) calculated a stress drop of several hundred bars for one of these events. Both earthquakes ($m_b$ $5\frac{1}{2}$ – 6) triggered a strong-motion instrument in the Shumagin Islands. Detailed analysis of the first motions of $P$-waves and the strong-motion records indicate that the earthquakes were of the thrust type and each event is associated with a stress drop of about 500 bars (HOUSE and BOATWRIGHT, 1979). These high values of stress drop may be indicative of a broad region that is under considerable tectonic stress. ARCHAMBEAU (1978) also reports high stresses for other nearby events in this area. Because the seismic history is unclear, this region has been assigned a potential of category 3.

A third gap lies between the rupture zones of the 1964 and 1958 earthquakes in southern Alaska. This region is near the site of several large events that occurred around the turn of the century (RICHTER, 1958). TARR and MARTIN (1912) found that uplift in those shocks was restricted to Yakutat Bay. THATCHER and PLAFKER (1977) conclude that the region affected by those shocks was limited in extent, with motion possibly occurring on a small portion of an east-trending thrust fault and a small segment of the Fairweather fault. As no tsunami was recorded outside the immediate area of the events, it is unlikely that they ruptured a great distance along the coast of Alaska. The region affected by these events lies in or near the area ruptured by the 1958 event. The remaining gap, therefore, lies in a region of transition from transform to thrusting motion and has no clear history of being associated with a large earthquake. We have no evidence, however, that would indicate that this region could not be the site of a future large earthquake. Therefore, it is assigned to category 3.

There are other faults near the Alaska-Aleutian plate boundary, such as those of central Alaska and the northwestern extension of the Fairweather fault. These faults have been the sites of large earthquakes in the past and can be expected to be the sites of large shocks in the future. Because these regions do not lie on clearly defined plate boundaries, however, the categories used here may not apply and we do not consider their seismic potential in this paper. Also, to the south of the 1949 Queen Charlotte event the maximum size of instrumentally recorded earthquakes is considerably smaller than that to the north. KELLEHER and SAVINO (1975) relate this change in rupture size to the thinner lithosphere that lies to the south, adjacent to the spreading centers that lie off the Washington-Oregon coast. The region to the north of these spreading centers of thin lithosphere has been the source of large earthquakes (KELLEHER and SAVINO, 1975; GÜTENBERG and RICHTER, 1954), however, because of the locally thin lithosphere most of the events in this region will fall below the threshold of $M_s \geq 7.0$. Thus, we will not consider the area in detail.

*(b) California*

The San Andreas fault system extends along much of the coast of California (Fig. 6). Right-lateral strike-slip motion has been well documented along much of the fault system. Great earthquakes with extensive rupture zones have occurred along two sections of the fault. The San Francisco earthquake of 1906 broke the northern segment of the fault from near Shelter Cove to San Juan Bautista (REID, 1910). The Fort Tejon earthquake in 1857 ruptured from north of Cholame Valley to near San Bernardino in the south (ALLEN, 1968). The fault segment that lies between

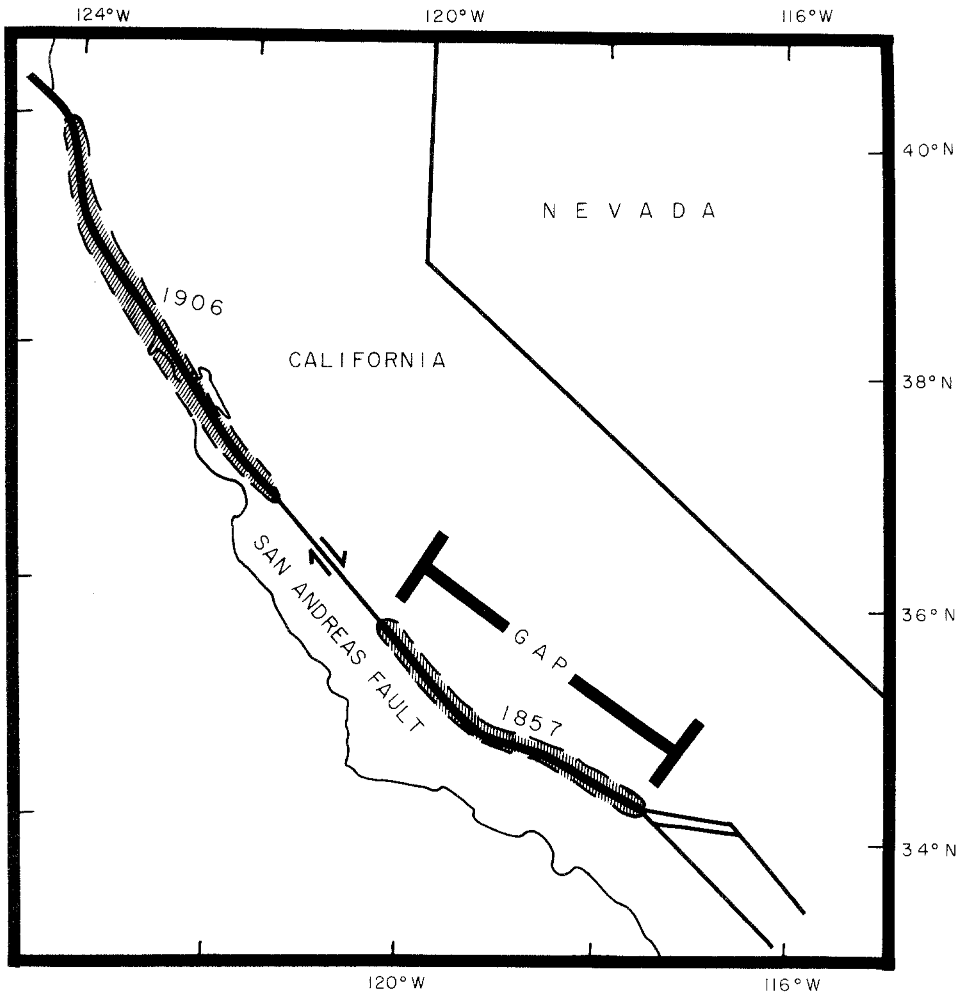



Figure 6

Location and rupture zones for the great earthquakes of 1906 and 1857 that broke the San Andreas fault. Fault segment between these rupture zones is creeping and thus may not be the site of future great earthquakes. That portion of the fault labeled as a gap is the most likely site of the next great earthquake along the San Andreas fault.

these rupture zones does not appear to have the potential for being the site of future large shocks as the creep rate along the fault in much of that segment is nearly equal to the rate of plate movement determined geodetically (SAVAGE and BURFORD, 1973).

Historic records for California are not very complete for more than about 100 years. There is no clear evidence of a great earthquake prior to the 1857 event. ALLEN (1968) indicates that events in 1838 near San Francisco are not associated with ground breakage as extensive as that of the 1906 event. SIEH (1978) and JAHNS (1977) recently completed studies of Holocene displacement along the southern portion of the San Andreas fault. Average recurrence times from the two sites studied are rather different: 225 years for one and 160 or 323 years for the other, depending on the interpretation of the smaller offsets at that site. There is much scatter in the observed return intervals themselves, 57 to 255 years. Because the reported recurrence times scatter a great deal, we do not believe that they can be used at present to provide much control on the times of future large earthquakes. That portion of southern California by the 1857 event has not experienced a great earthquake during the last 100 years. Thus, it has been assigned to category 1.

The area of the 1906 rupture, which appears to have extended to the Mendocino fracture zone, is assigned to category 2. The Gorda ridge and the Mendocino fracture zone represent the continuation of the Pacific-North American plate boundary to the northwest of the San Andreas fault system. They have experienced large shocks in the past (GUTENBERG and RICHTER, 1954). Although none of these shocks are as large as the great earthquakes to the south, future major shocks along the eastern parts of these features could be very damaging in northern California. Those features are also assigned to category 2.

The region of southern California to the south of the zone that ruptured in 1857 is characterized by a series of faults trending parallel or subparallel to the trend of the San Andreas fault. Some of these fault segments have experienced large shocks, although most sizeable events in the region are in the $6 \leq M \leq 7$ range (THATCHER *et al.*, 1975). The largest size of events may be related to the distance from spreading centers in the Imperial Valley and the northern Gulf of California. This area is not discussed further in this paper.

### (*c*) *Middle America*

The Middle America seismic zone extends along western Mexico and Central America between 82.5° and 105°W, defining the northeastern boundary of the Cocos plate. Focal mechanism solutions for shallow earthquakes along this zone indicate thrusting of the Cocos plate beneath the Caribbean and North American plates (MOLNAR and SYKES, 1969; DEAN and DRAKE, 1978) with about a N30°E direction of convergence (CARR, 1974).

Both the seismic gaps and rupture zones of earthquakes of this century are relatively small (100–200 km) compared to the dimensions of rupture zones along the Alaska-

Aleutian or Kurile-Kamchatka seismic zones. Northwest of 95°W, the Middle America trench is subducting somewhat younger sea-floor than the segment to the southeast. Consequently, the inferred dip of the seismic zone northwest of 95°W is generally shallower than in the southeast (KELLEHER and MCCANN, 1976; DEAN and DRAKE, 1978). The Tehuantepec ridge divides these two regions of sea-floor and intersects the trench near 95°W. Rupture zones of large earthquakes occurring along the segment of the arc that is subducting young sea-floor tend to be somewhat larger than those occurring southeast of 95°W. CARR and STOIBER (1977) suggest that variations in the geometry of the inclined seismic zone influence the spatial location of large earthquakes and other subduction-related phenomena. Along the inner wall of the trench northwest of 95°W, the lengths of fore-arc basins correlate well with the rupture lengths of the largest known shocks. Several basins are bounded by submarine canyons, which may be fault controlled (FISHER, 1961). The alignment of these canyons with other transverse breaks mapped by CARR *et al.* (1974) further supports Fisher's interpretation.

The Middle America seismic zone has seven major areas that have not experienced shallow earthquakes larger than magnitude 7 for 45 years or more; these seismic gaps are presented in Fig. 7. The northwestern end of the Middle America seismic zone last ruptured during 1932–34. The short recurrence times for this arc, 40–50 years (KELLEHER *et al.*, 1973), indicate that some of these segments have a high

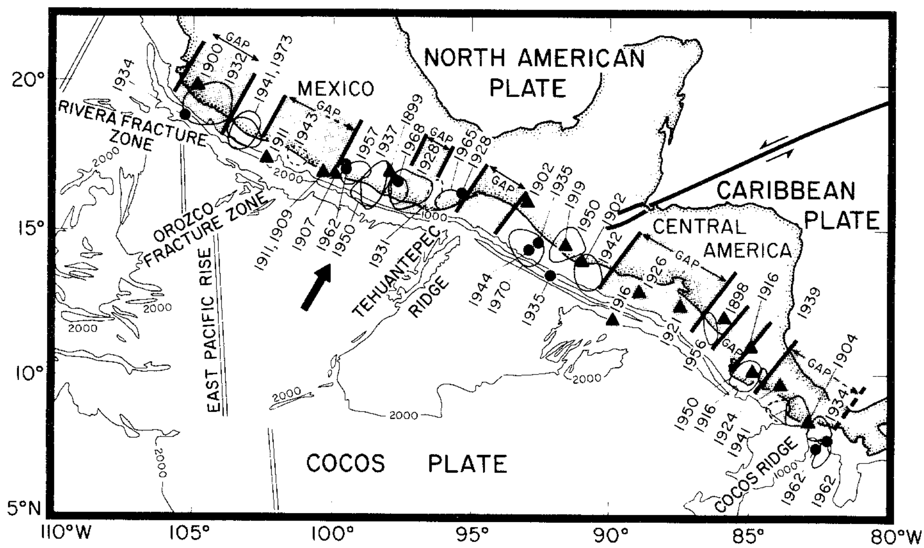



Figure 7

Rupture zones (ellipses) and epicenters (triangles and circles) of large shallow earthquakes (after KELLEHER *et al.*, 1973) and bathymetry (CHASE *et al.*, 1970) along the Middle America arc. Note that six gaps which have earthquake histories have not ruptured for 40 years or more. In contrast, the gap near the intersection of the Tehuantepec ridge has no known history of large shocks. Contours are in fathoms.

potential for being the sites of future large earthquakes. The northwestern end of the arc appears to have that potential and is placed in category 2.

A major gap is located where the Orozco fracture zone and other associated features intersect the trench. These features may be interfering with the normal subduction process; hence, this region may not have rupture zones or recurrence intervals similar to the rest of the arc. A series of major earthquakes occurred in this gap in 1907–09, 1911 and 1943; however, there are insufficient data to accurately determine the size and precise location of the rupture zones. This area is assigned to category 2. Note that the 1973 Colima earthquake ($M = 7.5$) apparently ruptured the same area as the 1941 event and filled the adjacent long-term gap to the southeast (REYES *et al.*, 1978).

A third seismic gap, which has recently attracted much attention, is located near the coast of Oaxaca, Mexico. It is assigned to category 2. The last large earthquake in this region occurred in 1928 ($M = 7.9$). A decrease in local seismic activity in mid-1973 is noted by OHTAKE *et al.* (1977). They think the decrease is similar to patterns of seismic quiescence preceding previous large earthquakes. Observations of variations in seismic activity are crucial in efforts to predict earthquakes; however, as GARZA and LOMNITZ (1978) note, these variations need to be evaluated with respect to the long-term seismic history. That is, do changes of this type occur without being followed by large shocks or are they only associated with the occurrence of large earthquakes?

A major long-term seismic gap is located near 95°W, where the Tehuantepec ridge intersects the Middle America trench. This segment has no clear association with large historic earthquakes (KELLEHER *et al.*, 1973). Hence, we conclude that it has a relatively low potential for being the site of a future large earthquake and assign it to category 5.

Three seismic gaps are located southeast of 95°W. The large gap near the coast of El Salvador experienced large events in 1921 ($M = 7.3$) and 1926 ($M = 7.1$). It is not clear that these shocks ruptured the entire gap, however, their magnitudes would suggest that they did not. Prior to those dates, the historic record indicates that the region offshore from El Salvador may have ruptured in two space–time progressions in 1847–50 and 1898–1902 (CARR and STOIBER, 1977). This gap has been seismically quiescent for shallow events of $M \geq 4$ since at least 1950. Like the Oaxaca gap, this region deserves more attention as a potential site for a future large earthquake. This gap has the potential (category 2) for being the site of one to a few large shocks or a great earthquake of magnitude exceeding 8. A smaller gap (category 2) is also located between the 1950 and 1956 rupture zones in southern Nicaragua and Costa Rica.

The southeastern end of the Middle America trench, near the west coast of Costa Rica, is located in a rather complicated tectonic regime. It is in this region that the Cocos ridge intersects the trench. The collision of this topographic feature with the subduction zone undoubtedly complicates and thus reduces the reliability of any analysis of this area using the techniques of seismic gaps. Large events occurring in

1934 (M = 7.7) and 1941 (M = 7.5) may have only partially ruptured the plate boundary. Prior to that, the last great earthquake occurred in 1851 (CARR and STOIBER, 1977). Thus, this area has the potential for a future great earthquake and is assigned to category 2.

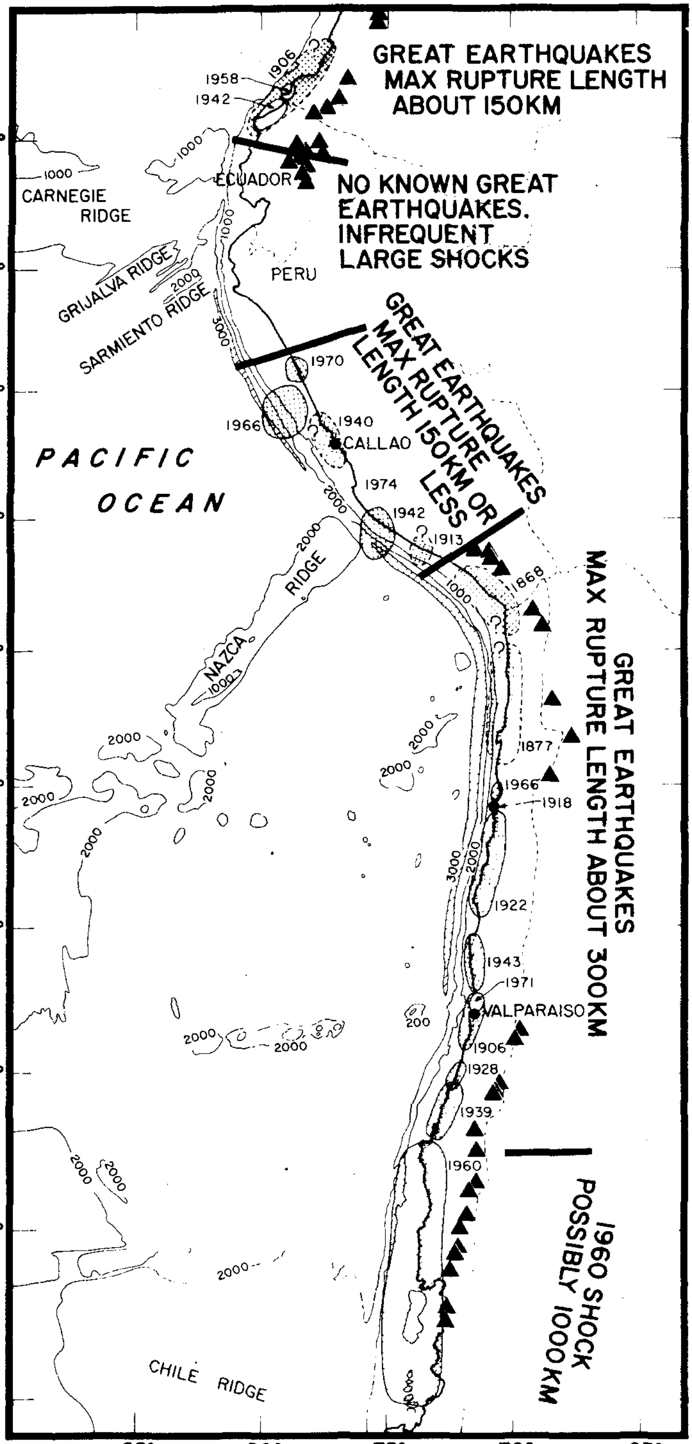
Several moderate-size earthquakes that have occurred along the grabens and volcanic zone of Central America during the past 400 years have been quite damaging to nearby centers of population. These shocks, however, are not located along the main plate boundary and hence are not included in the forecasts made here. It can be expected that similar shocks will pose a hazard in the future to cities like Guatemala City and Managua.

### (*d*) *South America*

The South American seismic zone extends along western South America from 5°N to 46°S, forming the eastern boundary of the Nazca plate. Typical focal mechanism solutions for shallow earthquakes are of the thrust type with the direction of relative plate motion about N80°E (STAUDER, 1973). The maximum dimensions of rupture zones of large South American earthquakes vary considerably along the strike of the active seismic zone. This large variation correlates strongly with the width of the zone of contact between the overthrust and underthrust plates (KELLEHER *et al.*, 1974), which may be influenced by the segmentation of the underthrust plate (CARR *et al.*, 1974; BARAZANGI and ISACKS, 1976). Variations of the morphology of the inner wall of the trench (KULM *et al.*, 1977) also appear to correlate with the segmentation of the seismic zone as well.

Five regions have been designated as seismic gaps. Many other areas have not ruptured in the last 30 years. Historic records, however, indicate that recurrence intervals along the South American seismic zone tend to be larger than about 60 years (KELLEHER, 1972). The northernmost gap, in northern Ecuador-southern Colombia (Fig. 8), includes the northern portion of the rupture area of the 1906 event; the southern portion of this zone ruptured during events in 1942 (M = 8.3) and 1958 (M = 7.8). The northern portion of the zone is assigned to category 2. If the recurrence interval is as short in the northern part as it is in the southern region (40 to 50 years), then this region is a prime location for a future large earthquake.

An extensive gap exists from 0° to 9°S. Historic records give no indication of a great earthquake occurring in this region during the last 400 years (KELLEHER, 1972); hence, this gap is assigned to category 5. Unless the recurrence times for great events in this region are unusually long (more than 400 years), one would not expect any change from the past pattern of relative seismic quiescence for large shocks. This zone corresponds to that portion of the South American trench that is consuming the Carnegie ridge, a large aseismic feature that is thought to have been formed at the Galapagos hot spot. The interaction of both the Carnegie and Nazca ridges with the subduction zone may explain the extremely shallow dip of the seismic zone from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40°
85°
80°
75°
70°
65°

0° to 15°S, as well as the absence of appreciable Neogene volcanism and historic great earthquakes. Major structural changes along the Andes are noted where these ridges intersect the coast (GANSSER, 1973). The Amotape zone, which forms the boundary between the northwest-trending central Andes and the northeast-trending northern Andes, occurs near the intersection of the Carnegie ridge with the coast. In addition, the Amotape zone may continue eastward into the Amazon depression, which has been suggested to follow a reactivated lineament in the shield (DE LOCZY, 1968).

The region from 9° to 15°S has a history of great earthquakes; however, the extent of rupture is usually limited to 150 km or less (KELLEHER and MCCANN, 1976). A previously noted gap near 12°S was the site of a large event in 1974 ($M = 8.1$). The aftershocks of that event nearly fill the seismic gap. The southern boundary of this region, near 15°S, occurs near a major tear in the descending Nazca plate (BARAZANGI and ISACKS, 1976), which may be associated with the collision of the Nazca ridge with the trench. On land, the coastal cordillera of the central Andes and Neogene volcanics terminate abruptly near the intersection of the Nazca ridge and the coast (GANSSER, 1973).

Another extensive seismic gap is located along the coast of southern Peru and northern Chile. This region, along the big bend in the South American coastline, has a history of highly destructive earthquakes and tsunamis. This region is delimited by a high concentration of Neogene volcanism and a relatively steeply dipping seismic zone. Along the northern portion of this gap, great earthquakes in 1604 and 1868 completely destroyed Arica and other coastal towns and generated damaging Pacific-wide tsunamis. The only great event known in the southern portion of this gap occurred in 1877 and was also accompanied by a destructive tsunami (LOMNITZ, 1970). The earthquake history for the northern region suggests a recurrence time for great tsunamic earthquakes of about 260 years. Hence, this area may continue to accumulate tectonic strain for many years before a great shock occurs. Nevertheless, the town of Arica has been destroyed five times in the historic record (LOMNITZ, 1970), indicating the destructive potential of large earthquakes in this region as well. This region is assigned to category 1 based on the long time interval since the last great events in 1868 and 1877.

Variations in population density suggest a rather incomplete historic record along segments of western South America. The regular occurrence of earthquakes near populated areas is suspect; however, they do provide an estimate of recurrence

Figure 8

Bathymetry (CHASE *et al.*, 1970), rupture zones of large earthquakes (cross hatching) and active volcanoes (solid triangles) along western South America (from KELLEHER and MCCANN, 1976). Note the absence of great shocks near the intersection of the Carnegie ridge complex with Peru-Ecuador. Discontinuities in the distribution of active volcanoes reflect the intersection of ridges with the continent and discontinuities in the already-subducted Nazca plate. Contours are in fathoms.

times. Earthquakes occurring in regions between large cities are not well documented and may provide a misleading interpretation of the seismic history, i.e., the inferred repeat times are larger than the actual recurrence intervals.

Historic earthquake activity near Copiapo, Chile (25°S) suggests a rather short recurrence interval of about 40 years. Since the last large earthquake occurred in 1918 (M = 7.7), the area is assigned to category 2. The region between Copiapo and Valparaíso (Transverse Valley region) is located along a segment of the arc that is characterized by a shallow dipping seismic zone and by a lack of Neogene volcanism. Both the 1922 (M = 8.4) and the 1943 (M = 8.3) events indicate that the region has a seismic potential; either low population or general lack of large shocks may explain the 'aseismic' history of this region prior to the 1922 event. The apparent lack of historic earthquake activity necessitates placing this region in category 3.

The region near Valparaíso, Chile is marked by the intersection of the Juan Fernandez ridge with the trench, an increase in the dip of the Chilean seismic (Benioff) zone and the abrupt appearance of Neogene volcanism to the south. A moderate-size gap exists south of Valparaíso and is placed in category 2. A recurrence time of approximately 85 years is suggested from the historic data. During this century the region first ruptured in 1906 (M = 8.2), and the northern segment ruptured again in 1971 (M = 7.9). As shown in Fig. 8, large events tend to migrate south along this section of the seismic zone. If this pattern continues, then the southern portion of the 1906 zone should be the site of the next large earthquake in central Chile.

South of the triple junction between the Peru-Chile trench and the Chile rise at 46°S the oceanic portion of the Antarctic plate is being subducted beneath the South American plate at a rate of about 2 cm/year (FORSYTH, 1975). Magnetic anomalies in the southeast Pacific indicate that segments of the Chile rise collided with the southern portion of the South American continent 26 m.y. ago (HERRON and TUCHOLKE, 1976). Since that time, there has been a great diminution of volcanic activity on the continent and a cessation of folding in sedimentary basins (WINSLOW, 1978). The seismicity of the region between 46° and 52°S is very low, with only a few events reported in the magnitude 4.5 to 6 range; hence, this region is assigned to category 5.

Two convergent left-lateral shear zones transect the southernmost South American continental margin near 53°S (Fig. 9). The main branch of the more easterly-trending shear zone includes the Magellan fault of KATZ (1964) which can be followed east into the North Scotia ridge. Both the Shackleton fracture zone and the North Scotia ridge correspond to the western and northern transform boundaries of the Scotia plate (FORSYTH, 1975).

Since the settlement of this region has been very recent, the earliest record of earthquake activity is the Magellan Strait earthquake of 1879 (M = 7–8 ?) (LOMNITZ, 1970). Two large earthquakes occurred within a 10 hour period in 1949 (M = 7.5, GUTENBERG and RICHTER, 1954). Field mapping of ground displacements associated with the 1949 events indicates a rupture length of about 300 km (WINSLOW, 1976,

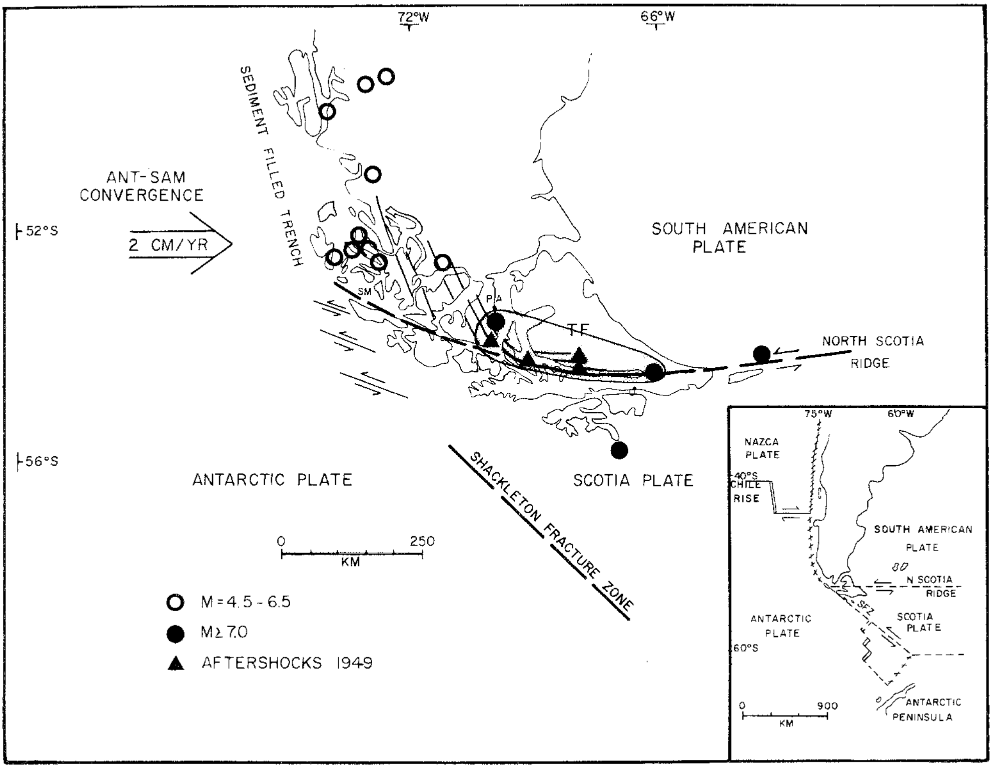



Figure 9
Seismic activity for southern South America since 1948. Modified after WINSLOW (1978). Events are from GUTENBERG and RICHTER (1954), ISS (1950), and FORSYTH (1975). Shaded area is inferred rupture zone of 1949–50 events, based on historical data at both extremities and by field mapping in the region in between. Aftershocks, as recorded by local sources, define the rupture zone. PA = Punta Arenas, Chile; U = Ushuaia, Argentina; TF = Tierra del Fuego; SM = Strait of Magellan.

1979). Damage to structures occurred in Punta Arenas, Chile as well as north of Ushuaia, Argentina, which is 300 km to the southeast (INTERNATIONAL SEISMOLOGICAL SUMMARY, 1949, 1950). The inferred rupture zone is shown in Fig. 9. This region has the potential for large earthquakes and is assigned to category 6, having ruptured 30 years ago or less.

(*e*) *Caribbean*

Large segments of the seismic zone surrounding the Caribbean plate have not ruptured in the last 100 years (Fig. 10). The rates of motion between the Caribbean and the North and South American plates is estimated to be approximately 2 cm/year (JORDAN, 1975); therefore recurrence intervals of at least 100 years, and probably more typically hundreds of years, may be more commonplace for this region. Since

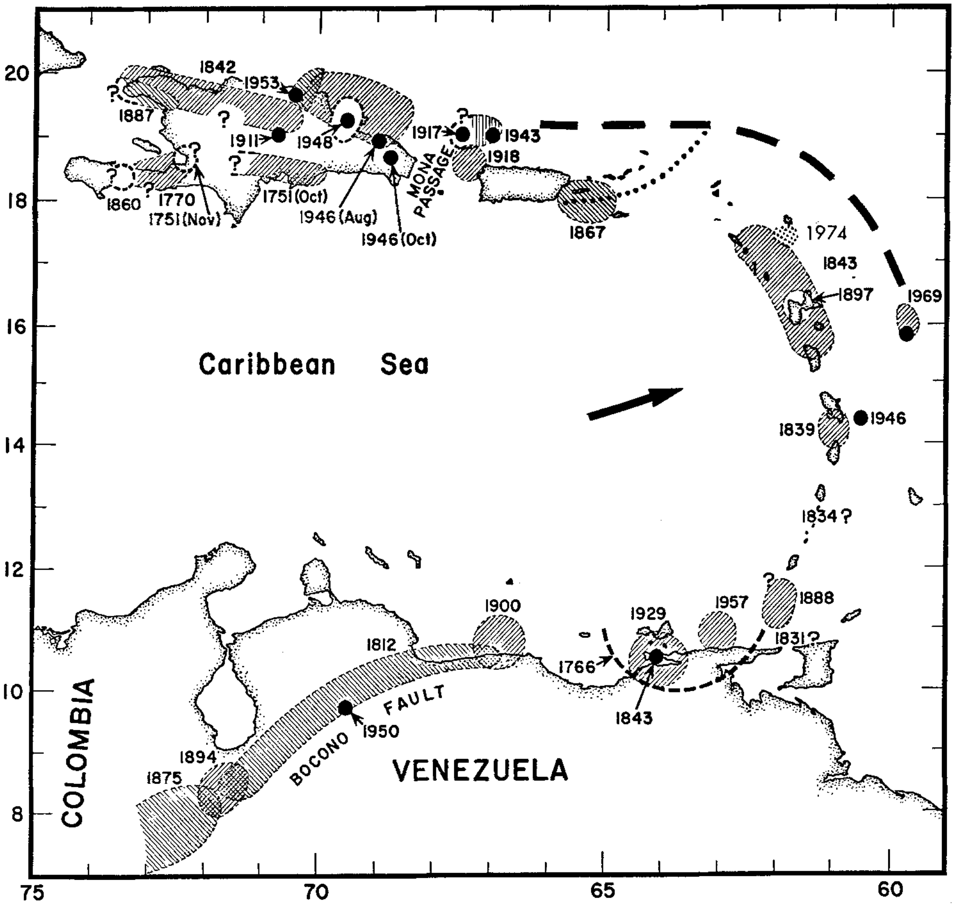



Figure 10

Estimated rupture zones of Caribbean shocks since 1800 (after KELLEHER *et al.*, 1973, with additions). Three great shocks of the eighteenth century are also included. Estimated rupture zones for the earthquakes of 1943, 1946, 1948, 1953, 1969 and 1974 are based on aftershock data. Other estimates are based on isoseismal areas where intensity IX or greater was experienced. Rupture zones of events during the nineteenth century along the Lesser Antilles were probably seaward of the areas indicated in the figure. Arrow shows the relative motion of the Caribbean plate with respect to the North and South American plates. Dashed line is Puerto Rico trench; dotted line is the Anegada trough.

the rate of plate motion is relatively low, it is very difficult to estimate recurrence times for the Caribbean. In many areas only one large shock is reported during the last 200 years. In other areas no shocks are known historically. The quiescence observed at many sites along the plate boundary, therefore, may not be related to an incomplete historical record but rather to the long intervals between large shocks.

The northern Lesser Antilles are, perhaps, the only classical subduction zone in the region in that they contain historically active volcanoes and subduction is occurring nearly perpendicular to the arc, i.e., in a WSW direction. This arc segment ruptured in great shocks in 1690 and 1843; the zones affected by each event appear to be very similar (KELLEHER *et al.*, 1973). There is, then, a crude indication of a repeat time of about 150 years for this segment of the arc. It has been assigned to category 1.

To the south of this region the earthquake history does not indicate that extensive great earthquakes have recurred along any portion of the arc. In fact, it appears that some segments of this arc have not ruptured in historic times. A large sedimentary wedge, a part of the delta of the Orinoco River, lies seaward of the central and southern Lesser Antilles. Large amounts of sediment may interfere with the subduction process (JACOB *et al.*, 1977). Therefore, the relatively quiescent past for the central and southern Lesser Antilles may continue for some time in the future.

The coast of Venezuela has felt the effects of several major and great earthquakes, the most extensive event being the great earthquake of 1812 (KELLEHER *et al.*, 1973). Its destruction was spread inland along the Bocono fault zone for several hundred kilometers. Undoubtedly much strain has accumulated along this fault zone and other areas that have experienced large earthquakes, but the complicated tectonics and relatively low rate (2 cm/yr) of plate motion in this region make it difficult to further identify regions of high seismic potential, with the possible exception of a zone between Caracas and Cumana (64°–67°W, 10°N, Fig. 10). That zone may not have ruptured during the 1900 event, which RICHTER (1958) estimated to have a magnitude of 8.4.

The Greater Antilles are also a region of complex tectonics. Fault plane solutions near these islands indicate plate motion trending WSW. This motion occurs on fault planes that dip gently to the south beneath the inner wall of the Puerto Rico trench (MOLNAR and SYKES, 1969). The motion in this region is therefore characterized by oblique thrusting. Great earthquakes have not occurred north of much of Puerto Rico and the Virgin Islands for at least the last 400 years (KELLEHER *et al.*, 1973). Large destructive shocks of magnitude near $7\frac{1}{4}$, such as that in 1844, apparently originated to the north of Puerto Rico, but great events from this region are unknown. The presence of recently-deformed sediments in the accretionary wedge north of these islands indicates that the possibility of future great earthquakes cannot be ruled out. A study of the relationship between the observed seismicity and these deformed sediments and associated sediment ponds (fore-arc basins?) is presently underway (MCCANN, MURPHY and FRANKEL, 1979). This region of oblique slip is similar to the Commander Islands in the western Aleutian Islands. Although the westernmost Aleutians have also been a seismic gap for this century, they may have been the site of a large interplate event in 1849 (MEDVEDEV, 1968). If so, by analogy the Puerto Rico-Virgin Islands region may be one of considerable tsunamic-seismic risk. These regions and the Andaman arc, which appear to have a similar tectonic setting, are assigned to an individual category, 4.

The Anegada passage, an extensional trough in the Virgin Islands was the source

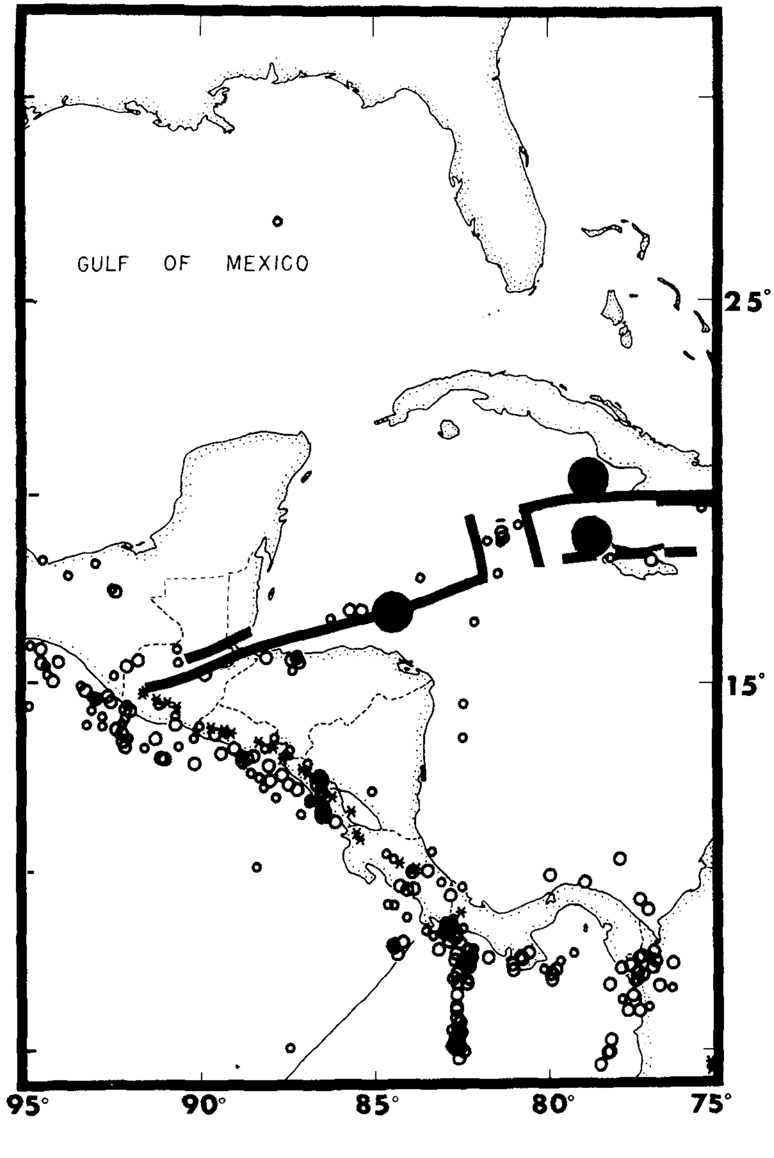



Figure 11

Relocated epicenters (circles) of shallow earthquakes (depths $\leq$ 70 km) for the western boundary of the Caribbean plate from 1954 to 1962 (after MOLNAR and SYKES, 1969, with additions). Larger symbols are more precise locations. Note the general lack of activity south of Cuba and thence west to Guatemala. Solid circles are sites of major earthquakes of this century. Solid arcs show regions affected by large earthquakes or tsunamis. Solid line shows inferred position of boundary of Caribbean plate.

of a large tsunamic earthquake in 1867 (REID and TABER, 1920). The felt area would suggest a magnitude of about $7\frac{3}{4}$ to 8. Although the tectonic history and present framework of the passage are yet to be unraveled, the trough has the potential for being a source of destructive earthquakes. A local network has recorded several microearthquakes along the north wall of the passage indicating that it is presently an active tectonic feature (MURPHY and MCCANN, 1979).

Large earthquakes have occurred along the western end of the Puerto Rico trench in 1911 (M = 7.0), 1917 (M = 7.0), 1918 (M = 7.5), 1943 (M = $7\frac{3}{4}$), 1946 (M = 8.1) and 1948 (M = 7.3) (GUTENBERG and RICHTER, 1954). These events were centered in either the Mona passage off the northwest coast of Puerto Rico or along the northeast coast of Hispañola, where subduction may be occurring more nearly perpendicular to the plate boundary that it is farther to the east or west.

The history of large shocks in Hispañola stands in marked contrast to the quiescence of much of the Puerto Rico-Virgin Islands region. Great earthquakes have occurred on both the north and south coasts of the island (KELLEHER *et al.*, 1973). Slip between the plates in this region may be accommodated by two or more fault zones striking easterly. The north and south coasts of the central and western portions of the island have not experienced large shocks for several scores of years and in some places for over two centuries. Therefore, any of these regions may have stored sufficient energy to generate a large shock.

Jamaica and southeastern Cuba have been the sites of several major earthquakes since the 17th century (TABER, 1920, 1922). Destructive shocks occurred in Jamaica in 1692 and 1907 with another strong shock in 1812. Seawaves from each of these events indicate sources to the north of the island (TABER, 1920). It is interesting to note that these shocks that are very destructive on Jamaica are not regarded as strong shocks on the southeastern coast of Cuba less than 200 km away. In fact, there are no reports of damage on Cuba from the 'great' Jamaican earthquake of 1692 (TABER, 1922), nor is the 'great' Cuban earthquake of 1678 reported on Jamaica. As each of these areas has experienced at least locally destructive shocks, they have the potential to be the future sites of major earthquakes. Preliminary analysis of the historic record indicates major Cuban shocks occur about every 100 years. The last occurred in 1852. Southeastern Cuba is assigned to category 1. Major shocks have occurred near Jamaica in 1692, 1812 and 1907 with the 1812 event being of lesser intensity. The events of 1692 and 1907 are associated with widespread destruction, and are probably great (M > 7.75) in size. The event in 1812 appears not to have been that large. The area near Jamaica is assigned to category 2.

Very little information exists about large shocks to the west of Jamaica (Fig. 11). The area is largely submarine and the islands are sparsely populated; therefore, only instrumental locations are available. The area of contact along this transform boundary is probably of sufficient strength to be the source of future large shocks involving strike-slip motion. The thin, weak crust near the Cayman spreading center near 82°W, however, will probably not be the site of large shocks in the future.

### (*f*) *South Sandwich arc*

In the central and southern parts of the South Sandwich arc, the subduction of young, oceanic lithosphere generated at an adjacent spreading center results in a sizeable region that lacks known shallow earthquakes of magnitude 6 or larger (Fig. 12; FRANKEL and MCCANN, 1979). Subduction of young buoyant lithosphere is also accompanied by uplift of the platform on which the southern islands are located and by shoaling of the trench. Bathymetric features on this incoming plate may further modify the subduction process in the far southern portion of the arc, as it is a region of intraplate seismicity. In the northern part of the arc where older (70 m.y.) lithosphere is being subducted, more events of moderate size ($6 \leq M \leq 7$) occur in the

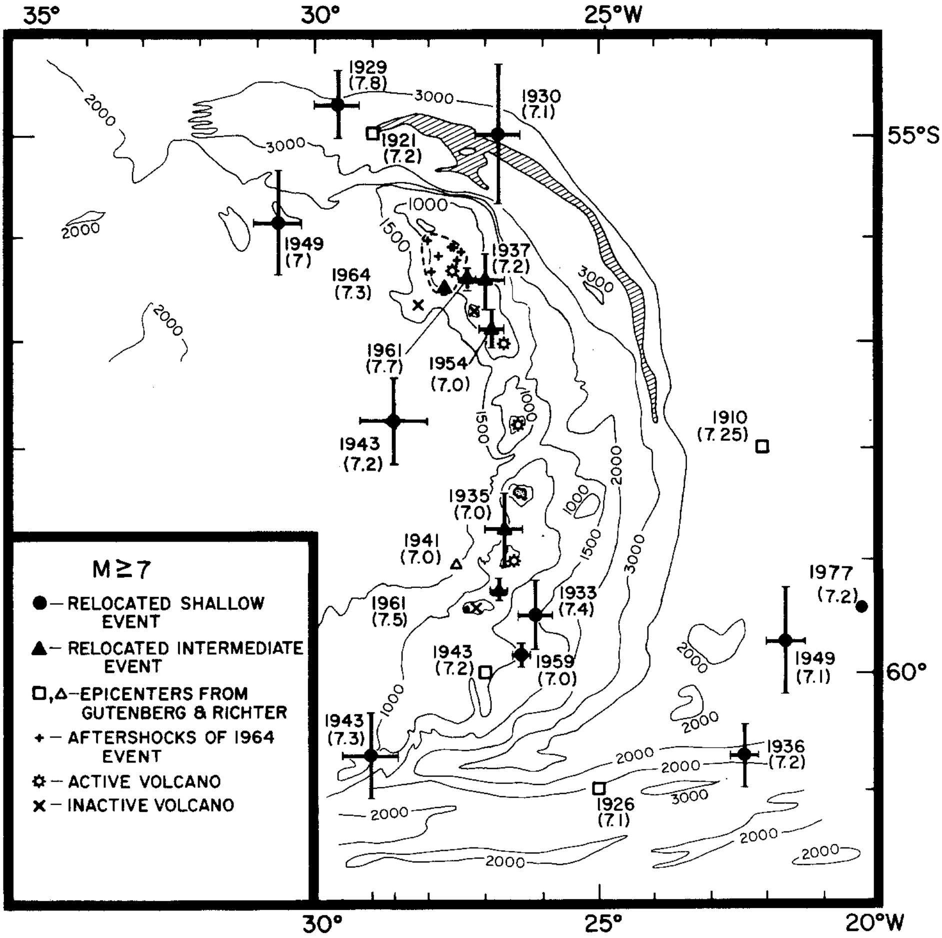



Figure 12

Large ($M \geq 7$) earthquakes in the South Sandwich region, 1910-August 1977 (from FRANKEL and MCCANN, 1979). Error bars indicate the 95% confidence limit on relocated epicenters. Bathymetry is in fathoms (1 fathom ~ 2 meters), shaded areas being deeper than 4000 fathoms.

region of underthrusting along the inner wall of the South Sandwich trench. Conspicuously absent, however, are large or great events involving thrust faulting. Perhaps the density contrast between the young overthrust plate (age approximately 7 m.y.) and the old underthrust plate is so large that only weak coupling exists between the plates. This may result in motion that is mostly aseismic. Another possibility is that back-arc spreading, which is occurring at a fast rate behind the arc, may lead to decoupling or low stress along the entire plate interface at the subduction zone.

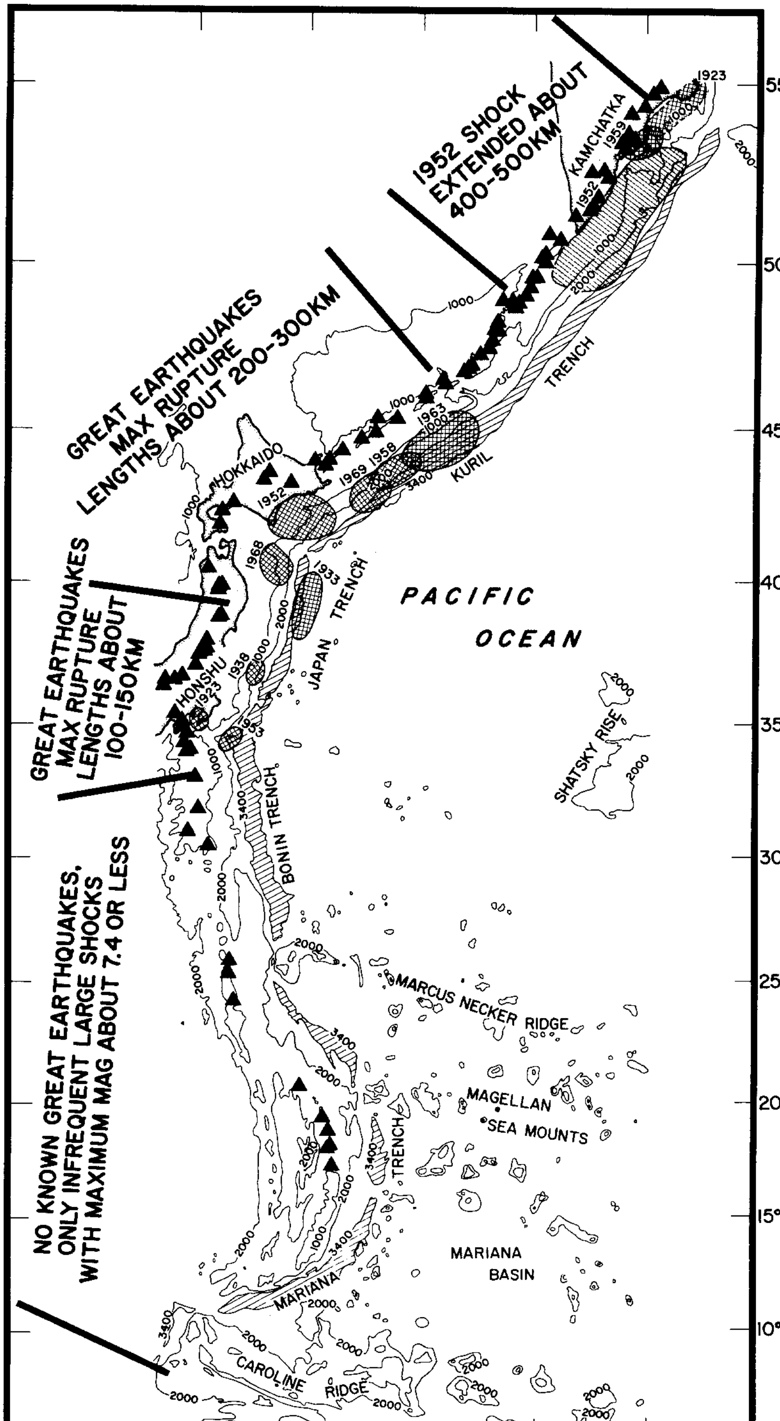
(*g*) *Western Pacific*

There are remarkable variations in the size and number of large thrust earthquakes that occur along the western margin of the Pacific plate (Fig. 13). Much of this variation can be explained by changes in the geometry and geophysical parameters that determine the tectonic regime of a particular segment of an arc; these include:

1. changes in the interface or zone of contact between the overthrust and underthrust plates;
2. modification of the subduction process by the interaction of bathymetric features on the incoming sea-floor with the overthrust plate [this condition is believed to lower the density contrast between the converging plates and hence to inhibit the subduction process (KELLEHER and MCCANN, 1976)];
3. large density contrasts between the overthrust and underthrust plates which may lead to poor seismic coupling and a reduction in the ratio of seismic to aseismic slip (KANAMORI, 1977a; MCCANN and NISHENKO, 1978).

(*1*) *Izu-Bonin-Mariana arcs.* The Izu-Bonin-Mariana arcs form the eastern edge of the Philippine Sea plate. In those areas the subducted Pacific plate appears to be reflected in a downgoing seismic zone which extends to depths of almost 700 km (KATSUMATA and SYKES, 1969). Neither GUTENBERG and RICHTER (1954) nor ROTHÉ (1969) report any shallow events of magnitude greater than 7.3 along these arcs between latitudes 10° and 35°N. The remarkable absence of great earthquakes and the small number of reported large shocks along this plate boundary may be attributed to one of the following factors:

1. There is little or no relative motion between the Philippine Sea plate and the Pacific plate.
2. Motion along the entire plate boundary occurs aseismically as a result of a decoupling of the plates (KANAMORI, 1977a). This could result from a large density contrast between the overthrust and underthrust plates (MCCANN and NISHENKO, 1978; FRANKEL and MCCANN, 1979) or from a weakening of the zone of interplate motion by frictional (or other) heating (KANAMORI, 1971).
3. Segments of the Pacific sea-floor are too light to be subducted beneath the overthrust plate. Deformation of the plates then occurs; intraplate activity may increase, but large thrust events become rare as subduction is interrupted.


1952 SHOCK EXTENDED ABOUT 400-500KM
GREAT EARTHQUAKES MAX RUPTURE LENGTHS ABOUT 200-300KM
GREAT EARTHQUAKES MAX RUPTURE LENGTHS ABOUT 100-150KM
NO KNOWN GREAT EARTHQUAKES, ONLY INFREQUENT LARGE SHOCKS WITH MAXIMUM MAG ABOUT 7.4 OR LESS
KAMCHATKA
HOKKAIDO
HONSHU
KURIL
TRENCH
JAPAN TRENCH
BONIN TRENCH
PACIFIC OCEAN
SHATSKY RISE
MARCUS NECKER RIDGE
MAGELLAN SEA MOUNTS
MARIANA BASIN
MARIANA
CAROLINE RIDGE
140°
145°
150°
155°
160°
55°
50°
45°
40°
35°
30°
25°
20°
15°
10°
5°

4. The recurrence time is extremely long (i.e., larger than the historic record which is complete only for about the last 75 years for most of the region).

At present the first hypothesis cannot be ruled out. The paucity of shallow activity and the lack of focal mechanism solutions of the thrust type can be used to argue for this possibility as well as for the second and third hypotheses. Until more information is gathered on this region we cannot resolve this matter. The deformation of the Mariana arc where the Marcus-Necker and Caroline ridges intersect the trench may indicate that the third possibility may apply there. KELLEHER and MCCANN (1976) and VOGT *et al.* (1976) favor that interpretation. KANAMORI (1977a) proposes that the absence of great earthquakes along this margin may be the result of poor coupling between the plates. He shows that a significant portion of the motion between the Pacific and Eurasian plates along that segment of the plate boundary off Honshu and the southern Kuril Islands may occur aseismically. He proposes that the decoupling is greater farther to the south, which results in nearly total aseismic motion on the zone of shallow thrusting in the Marianas, Bonin and Izu arcs.

Large shocks do occur occasionally along this boundary. The magnitude of one event in 1902, which is listed by RICHTER (1958) as a great shock ($M = 8.1$) is probably over-estimated (KANAMORI and ABE, 1978). The macroseismic effect of that shock did not include any indication of a seawave, although two islands 250 km apart are claimed to have exhibited effects similar to intensity IX on the Modified Mercalli scale (ALGUÉ, 1909). Guam has been the site of historic destructive earthquakes (ALGUÉ, 1909) but it has not experienced destructive seawaves. It is possible that these events are intermediate in depth and are therefore not directly related to the underthrusting process. The instrumental record and the short historic record of the Marianas give no indication of the occurrence of large events of the thrusting type that are typical of so many other subduction zones.

It is possible that recurrence times in this region are longer than the historic record. If this is so, then this region is very unusual. Convergent zones with relatively narrow interfaces, i.e., thin zones of plate contact, are found in Central America and along the Solomon arc, where short recurrence times (30–50 years) are typical. The Mariana arc also has a narrow interface. Given the present instrumental record, recurrence times would have to be at least 75 years. While this could be the case, it is difficult to explain why the whole of the Izu-Bonin-Mariana arc system from 10° to 35°N is presently in a quiescent stage. This type of synchronicity is unparalleled in

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Figure 13

A comparison of bathymetry (CHASE *et al.*, 1970), rupture zones of large earthquakes (cross hatching) and active volcanoes (solid triangles) along the western Pacific (from KELLEHER and MCCANN, 1976). Note the smooth ocean floor, large rupture zones and nearly continuous chain of volcanoes toward the north (Kamchatka-Kuril-Japan). Contrast these features with the irregular bathymetry, absence of great earthquakes and gaps and irregularities in the line of active volcanoes along the Bonin and Marianas arcs.

any other arc with the possible exception of Java, the Lesser Sunda Islands (which will be discussed later), and the South Sandwich arc.

Certain tectonic similarities between these arcs and others with no history of great shocks suggest to us that the 75 years of instrumental data are typical of the longer pattern of seismic energy release. Hence, we conclude that these arcs have a low potential (category 5) for being the sites of large or great shallow earthquakes of the thrusting type.

*(2) Northern Honshu and Hokkaido.* Smaller tsunamis and earthquakes have occurred in the region south of 38°N near the section of the Japan trench that is interacting with a group of seamounts on the Pacific plate (KELLEHER and MCCANN, 1977). To the north, smooth sea-floor is being subducted into a region that has been the site of several great earthquakes and destructive tsunamis. The observed spatial pattern

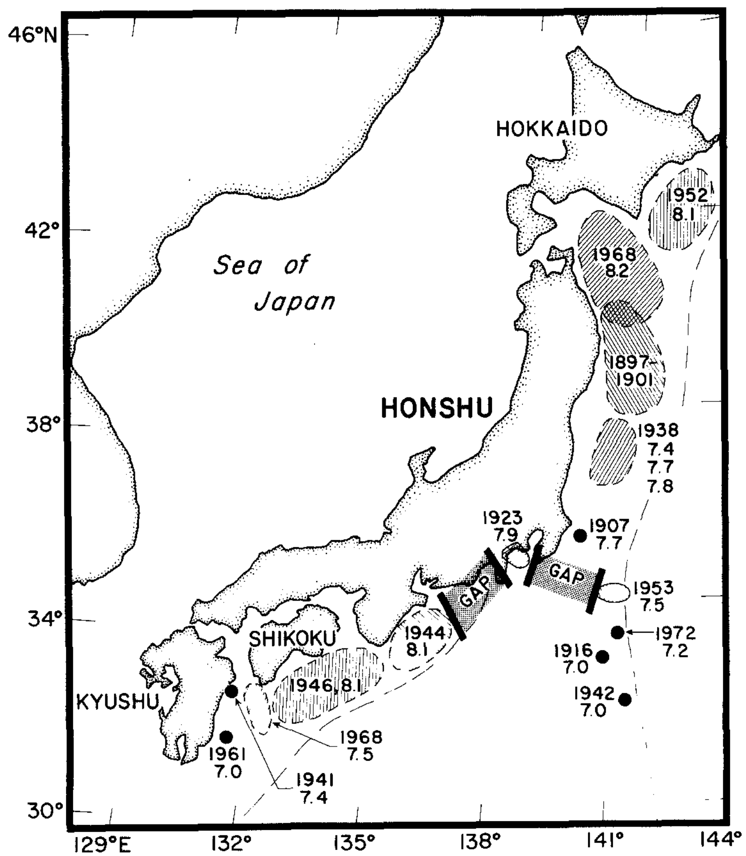



Figure 14

Seismic gaps and recent large earthquakes near the eastern and southern coasts of Japan. Hatched regions are rupture areas determined from aftershocks. Year and magnitude of the events are indicated next to the affected region.

of occurrence of large shocks is consistent with the historic record. Hence, we expect this pattern to continue in the future.

Figure 14 summarizes the known seismic gaps near the southern and eastern coasts of Japan. These gaps are nearly identical to those identified by MOGI (1968) and ANDO (1975).

(*3*) *Kuril-Kamchatka.* Much of this portion of the Pacific plate boundary has ruptured during the last 80 years. Figure 15 from KIMURA (1978) shows the rupture zones of great earthquakes and active volcanoes of this region. Two prominent seismic gaps

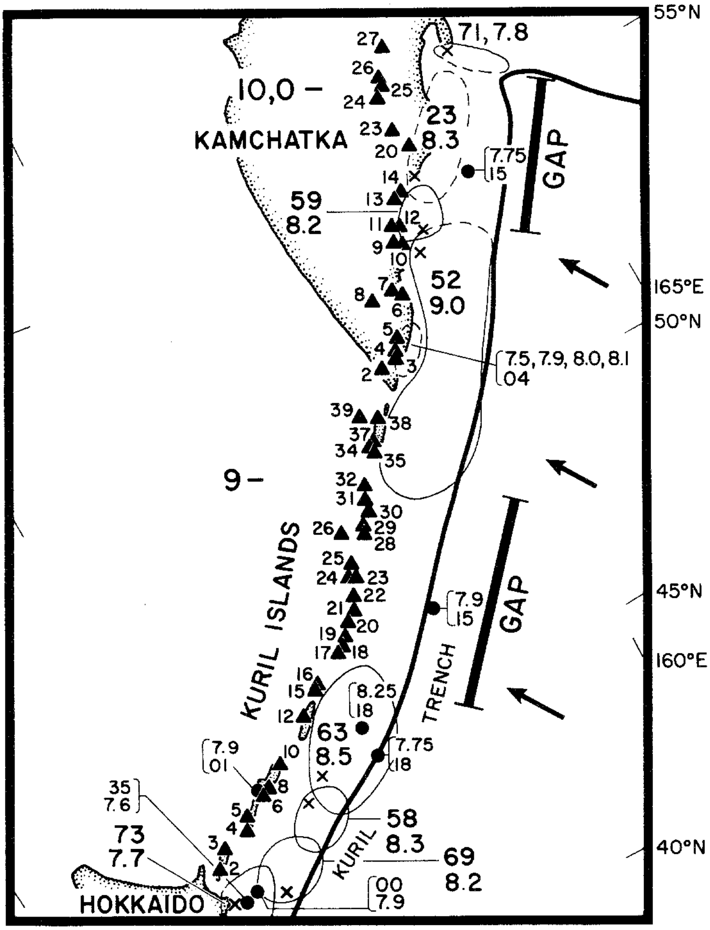
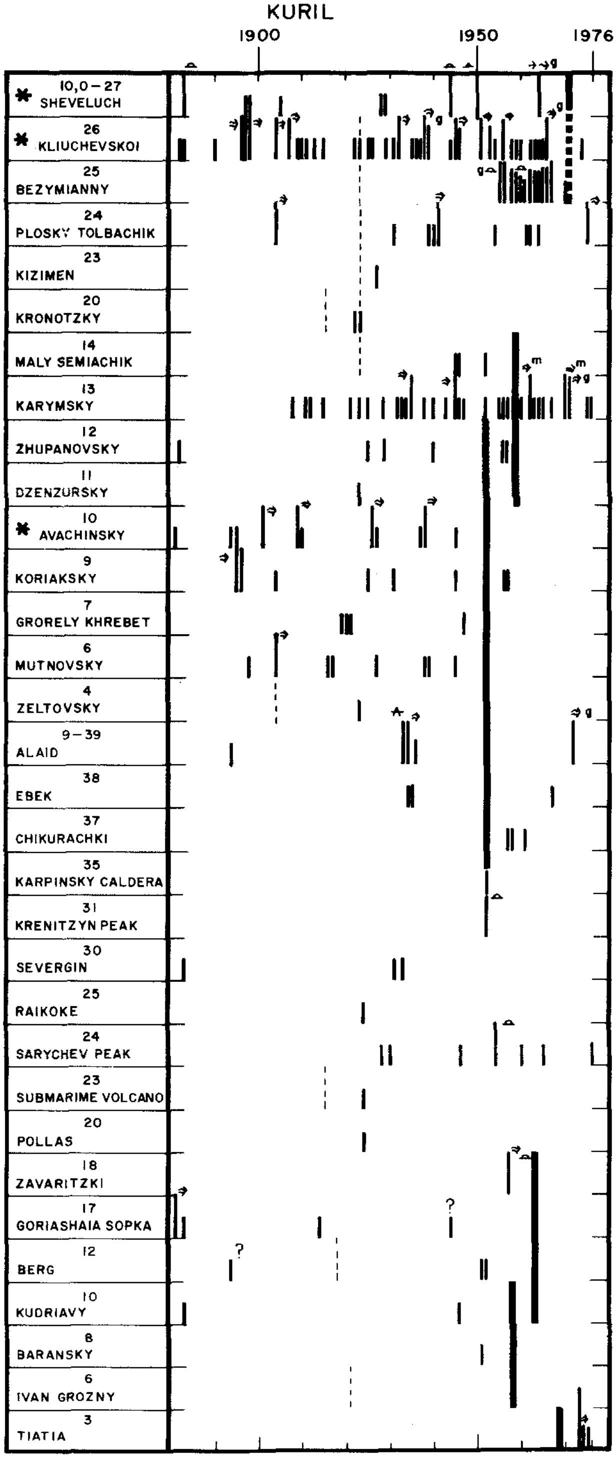



Figure 15

Great earthquakes along the Kuril-Kamchatka arc (from KIMURA, 1978). Solid lines represent rupture areas of recent great events. Dashed lines represent rupture zones inferred from felt reports. Dots and Xs represent poor and good epicentral locations, respectively, for main shocks. Dates and magnitudes of main shocks are indicated. Numbers next to volcanoes (solid triangles) are for reference to Fig. 19. Arrows denote directions of plate motion taken from MINSTER *et al.* (1974).


KURIL
1900
1950
1976
10,0 – 27 SHEVELUCH
26 KLIUCHEVSKOI
25 BEZYMIANNY
24 PLOSKY TOLBACHIK
23 KIZIMEN
20 KRONOTZKY
14 MALY SEMIACHIK
13 KARYMSKY
12 ZHUPANOVSKY
11 DZENZURSKY
10 AVACHINSKY
9 KORIAKSKY
7 GRORELY KHREBET
6 MUTNOVSKY
4 ZELTOVSKY
9 – 39 ALAID
38 EBEK
37 CHIKURACHKI
35 KARPINSKY CALDERA
31 KRENITZYN PEAK
30 SEVERGIN
25 RAIKOKE
24 SARYCHEV PEAK
23 SUBMARIME VOLCANO
20 POLLAS
18 ZAVARITZKI
17 GORIASHAIA SOPKA
12 BERG
10 KUDRIAVY
8 BARANSKY
6 IVAN GROZNY
3 TIATIA

are also indicated. The northernmost gap last ruptured during a great earthquake in 1923 (FEDOTOV, 1965). It has been assigned to category 2. The southern gap, near the central Kuril Islands, may have ruptured, in part, during an event in 1915. Since very little instrumental information is available for that event, the extent of rupture is uncertain. Earthquake catalogs (MEDVEDEV, 1968; KONDORSKYA and SHEBALIN, 1977) indicate that this region was less active for major shocks than the adjoining regions as far back as the early 1800s. More analysis of the geophysical data for this region is necessary to properly assess its potential as a source of future shocks. It is tentatively assigned to category 3.

Figure 16, also from KIMURA (1978), displays the history of eruptions for various volcanoes along the Kuril-Kamchatka arc. Heavy bars indicate the locations in time and space of all great earthquakes since 1950. For each volcano, bars are shown indicating the time of a major (full bar) or moderate (one-half bar) eruption. Volcanic activity adjacent to the rupture zones of the 1971 and 1952 earthquakes was clearly higher for several decades before each of those events compared to the years after those events. Therefore, observations of volcanic activity may provide a valuable tool for the long-term forecasting of future large earthquakes.

### (*h*) *Western margin of the Philippine Sea*

The western margin of the Philippine Sea plate can be divided into five zones, based on the tectonic character; the Nankai trough, the Ryukyu arc system, Taiwan, the Luzon Strait between Taiwan and the northern Philippines, and the Philippines. The Nankai trough is a young trench that has a historic record of great earthquakes, with extensive rupture zones and tsunamis, dating back to A.D. 684 (ANDO, 1975). The Ryukyu arc is a well-developed trench, with a well-defined dipping seismic zone. Taiwan, the Philippines, and the region between are complex island arc systems, which have undergone (and may be presently undergoing) polarity reversals. Parts of the latter experience frequent larger shocks, but lack well-developed subduction zones.

ANDO (1975) examined historic great earthquakes in the Nankai trough which meets the Ryukyu arc at the Palau-Kyushu ridge (Fig. 17). He divided the region into four blocks, west to east, A through D. He finds that great earthquakes tend to occur in cycles, which migrate from west to east, with large shocks occurring at a given place every 100 to 200 years. The latest cycle seems to have begun with the 1944–46 Tonankai and Nankaido earthquakes, which ruptured sections A through C.

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Figure 16

Space-time plot of the seismic and volcanic history of the Kuril-Kamchatka region (from KIMURA, 1978). Thick, long bars represent time and location of great earthquakes indicated on the previous figure. Small and full bars represent the time of small or moderate (m) to great (g) eruptions. Lava flows are indicated by wavy arrows.

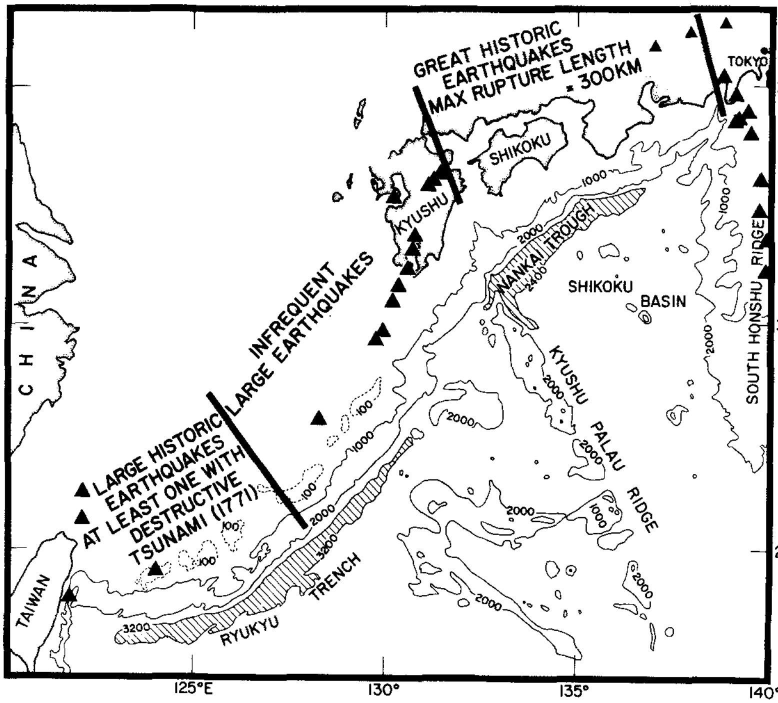


Figure 17
Bathymetry and active tectonics along the Ryuku arc and Southwest Japan (after KELLEHER and MCCANN, 1976). Seismic record near Southwest Japan is based on ANDO (1975). Triangles represent epicenters located by GUTENBERG and RICHTER (1954). Depth contours are in fathoms (1 fathom = 1.83 m). Solid triangles are Quaternary volcanoes.

The cycle has not yet reached section D, which last ruptured in the Ansei I earthquake of 1854 (ANDO, 1975). Hence, a great earthquake may be expected east of the Tokai district. That region has been assigned to category 1, i.e., highest seismic potential. The great historic earthquakes of the Nankai trough studied by Ando all have rupture zones that extend for several hundreds of kilometers, and terminate at the intersection of the Palau-Kyushu ridge with the Nankai trough and northern end of the Ryukyu arc (KELLEHER and MCCANN, 1976).

The Ryukyu arc is typical of a subducting margin: earthquakes are concentrated in a well-developed, planar zone that dips 35° to 45° to the northwest, and reaches depths of 250 to 300 km (KATSUMATA and SYKES, 1969); ROWLETT and KELLEHER, 1976). Behind the frontal arc and volcanic chain, the Okinawa trough represents a slowly-extending marginal basin (KARIG, 1973).

Historical data indicate that the southwestern portion of the Ryukyu arc has had a greater occurrence of large earthquakes than other portions of the arc. ROWLETT and KELLEHER (1976) note at least one severely destructive shock in 1771, which was accompanied by a destructive tsunami (Fig. 17). The northeastern portion, from Okinawa to Kyushu (Fig. 18), by contrast, has experienced only infrequent large

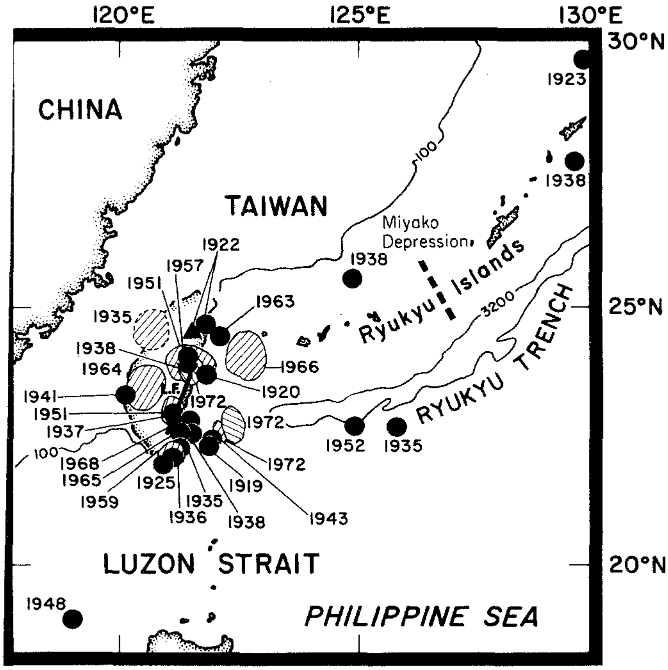


Figure 18

Epicenters of large shallow earthquakes and aftershock zones for the Luzon Strait-Taiwan-Ryukyu Islands region since 1918 (after ROWLETT and KELLEHER, 1976). Relocated epicenters (solid circles), epicenters located by GUTENBERG and RICHTER (1954) (solid triangles), aftershock data determined from relocated data (hachured regions), and Longitudinal fault of Taiwan (L.F., solid line) are shown. Note the intense activity for large earthquakes near Taiwan as compared to that near either the Ryukyu Islands or the Luzon Strait. Much of this activity is located on the 'oceanic' side of the Longitudinal fault.

shocks, with no reports of destructive tsunamis (ROWLETT and KELLEHER, 1976). Seaward of the southwestern portion of the arc, the bathymetry is relatively flat, whereas seaward of the portion from Okinawa to Kyushu a series of aseismic ridges intersect the trench and may be interfering with the subduction process (KELLEHER and MCCANN, 1976). KATSUMATA and SYKES (1969) cite a region from 24° to 25°N, 126° to 128°E as being nearly aseismic for small to moderate size earthquakes of shallow depth. In this region the Miyako Depression (Fig. 18) intersects the Ryukyu trench. To the northeast and southwest of this 'gap' KATSUMATA and SYKES (1969) and ROWLETT and KELLEHER (1976) noticed a clustering of activity since 1961. Similar patterns of gaps in short-term seismicity, with clustering of events at the edges, were observed by MOGI (1968) and KELLEHER and SAVINO (1975) along subduction zones prior to the occurrence of large earthquakes. Such areas may be likely to experience large shocks in the future. The northeast portion of the Ryukyu arc, which has no historic record of great earthquakes with extensive rupture zones or destructive tsunamis, is assigned to category 5. Because prominent bathymetric features similar to those seaward of the trench may be hindering the subduction process in this area, we do not expect great earthquakes for this region. The southeast portion, which may be exhibiting activity observed along other subduction zones prior to large earthquakes, and has not had a great shock since 1771, has been assigned to category 1, the highest seismic potential.

During this century Taiwan (Fig. 18) has experienced frequent large ($7 \leq M \leq 7.5$) shocks, with ruptures extending tens (as opposed to hundreds) of kilometers (ROWLETT and KELLEHER, 1976). Much of this activity tends to cluster along the Longitudinal Valley fault zone near eastern Taiwan. This feature probably represents the collisional boundary between the Asian continental plate and the northern part of the Luzon arc system (KARIG, 1973; ROWLETT and KELLEHER, 1976).

East of the Longitudinal Valley, Miocene to Pliocene melange and andesitic volcanic terrane comprise the Coast Range (KARIG, 1973). Karig notes the continuation of 'arc structural units' from the Batanes-Babuyan Islands to the south within the Coast Range, as well as a continuation of the east-dipping seismic zone of shallow (mostly less than 100 km) foci (KATSUMATA and SYKES, 1969) as evidence that the eastern section of Taiwan is part of the Luzon arc system. Taiwan is thought to have collided with the west-facing Luzon arc in the Late Pliocene (KARIG, 1973), and may be undergoing a reversal in polarity at the present time. Left-lateral offset along the Longitudinal Valley and along faults to the west are related by KARIG (1973) to intraplate shortening, which results from continued collision.

The present seismic regime of Taiwan is characterized by: (1) large, frequent shocks with small rupture zones; (2) focal mechanisms with scattered orientation; and (3) a poorly-defined seismic zone with an easterly-to-vertical dip beneath the east coast of Taiwan and maximum depths less than 200 km, and a tendency for shocks to be associated with the Longitudinal Valley fault (KATSUMATA and SYKES, 1969; ROWLETT and KELLEHER, 1976). The rather widespread distribution of seismic activity in Taiwan and as far south as the Philippines may be related to internal plate deformation associated with the collision of the Luzon arc system with continental material and to a possible reversal of polarity of the arc (KARIG, 1973; ROWLETT and KELLEHER, 1976). Taiwan is assigned to categories 1 and 2. Even though it has experienced large events less than 100 years ago, the repeat time of large shocks appears to be less than 100 years. Large shocks, however, have been a relatively rare phenomenon on Taiwan since about 1957. That relative quiescence for large shocks appears to be a temporary phenomenon.

The Luzon Strait (Figs. 18, 19), between the northern Philippines and Taiwan, consists of a series of ridges and troughs that are broken into en echelon segments, which suggest that left-lateral shear is operating in this region (KARIG, 1973). KARIG (1973) interprets these features as a continuation of the west-facing Luzon arc system, and traces them into eastern Taiwan. No large earthquakes are known to have occurred in this region during this century, but moderate ($M = 6$–$6.9$) shallow events are observed across a 200 km-wide zone between Luzon and Taiwan (ROWLETT

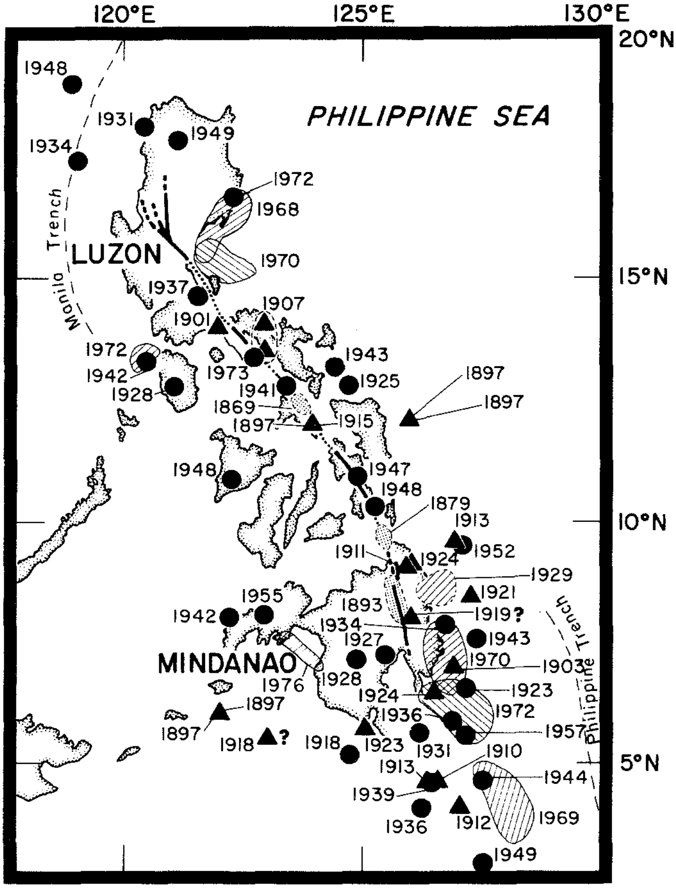


Figure 19

Large earthquakes near the Philippine Islands (after ROWLETT and KELLEHER, 1976). Relocated epicenters (solid circles), epicenters located by GUTENBERG (1956) or GUTENBERG and RICHTER (1954) (solid triangles), aftershock zones determined from relocated data (hachured regions), estimated rupture zones of several earthquakes prior to 1900 (dotted regions), and trace of the Philippine fault (solid and dotted lines) as mapped by ALLEN (1962) are shown. Dashed boundaries and queries indicate uncertain boundaries of aftershock zones or less reliable epicentral locations. Data prior to 1918 are not complete. Note that several large earthquakes have occurred during the past 105 years near the trace of the Philippine fault.

and KELLEHER, 1976). West of Luzon, the Manila trench (Fig. 19) is continuous as a depression to 20.5°N, where it merges with the series of north-northeast trending ridges and troughs of the Luzon Strait. KARIG (1973) and ROWLETT and KELLEHER (1976) suggest that plate motion in the region between southern Taiwan and the northern Philippines is accommodated by regional release of tectonic strain and that sharp plate boundaries have not yet developed. If this is the case, the absence of large earthquakes may be a permanent feature of this region (ROWLETT and KELLEHER, 1976). Hence, it has been assigned to category 5.

The Philippines are characterized by frequent large, shallow earthquakes that occur throughout the islands. The active features with which the seismicity is associated are the Manila trench, Philippine fault system and the Philippine trench (Fig. 19).

The present tectonic pattern of northwestern Luzon is that of a west-facing island arc system (KARIG, 1973). The Manila trench is associated with a poorly developed seismic zone that dips east. The deepest foci lie between 150 and 200 km. Seamounts and other bathymetric features on the underthrusting plate to the west of Luzon may be modifying the subduction process along the northern portion of the Manila trench (ROWLETT and KELLEHER, 1976). KARIG (1973) and ROWLETT and KELLEHER (1976) suggest that subduction may be in the process of terminating along the Manila trench. Focal mechanism solutions of shallow-focus events along the east coast of northern Luzon, which do not appear to be associated with the Philippine fault system, indicate westerly-directed underthrusting. These mechanisms are cited by KARIG (1973) as evidence that subduction may be developing in northeastern Luzon.

The Philippine fault system extends from Mindanao to central Luzon, where it branches into several subparallel faults which extend into northwestern Luzon. FITCH (1972) proposes that oblique convergence of the Philippine and Asian plates may result in thrust motion at the Philippine trench and left-lateral and strike-slip motion along the Philippine fault. Epicenters of large earthquakes for this century appear to lie close to the fault trace, and sections of the fault 75 km or more in length (Fig. 19) which have not ruptured for several decades appear to have the potential for being sites of future large shocks (ROWLETT and KELLEHER, 1976).

The Philippine trench has typical features of an active subduction zone from about 3°N to 1.5°N (ROWLETT and KELLEHER, 1976). Near 15°N it swings west as a shallower trough, which is almost sediment free, is characterized by shallow seismicity, and is presumed to be a zone of underthrusting and left-lateral shear (KARIG, 1973). Large earthquakes have occurred along most of the trench during the past century. South of Mindanao, the seismic zone is characterized by fault plane solutions involving thrust faulting (ROWLETT and KELLEHER, 1976). KARIG (1973) suggests that subduction may be beginning along this seismically active section. Scattered activity throughout the lesser Philippine Islands may indicate the boundaries of a small plate in the western Philippines (KATSUMATA and SYKES, 1969) or internal deformation (FITCH, 1972).

Most of the Philippines are assigned to category 2; large, frequent shocks are characteristic. Many large earthquakes have occurred there within the past 30 to 100 years.

### (*i*) *Sunda arc*

The seismotectonic nature of the Sunda arc changes throughout its length as the direction of relative plate motion changes from normal near Java to oblique near the Andaman Islands. East of the Sunda Strait, in Java and the Lesser Sunda Islands, there is a lack of historic great earthquakes and associated tsunamis, whereas to the west, in Sumatra, the historic record indicates great earthquakes and extensive damage by tsunamis. From the historic record there appears to be a higher potential for great earthquakes on and near Sumatra than near Java and the islands to the east (Fig. 20). At present the tectonic reasons for this are poorly understood and more work needs to be done in this area.

VISSER (1922), reporting on historic Sunda earthquakes from the eighteenth, nineteenth and early twentieth centuries states, 'Most earthquakes felt on the Isle of Java are weak'. KELLEHER and MCCANN (1976) note a prominent gap in activity for large shocks during the twentieth century between 105° and 122°E. This lack of earthquakes may be related to the interaction of the Christmas Islands ridge, a prominent bathymetric feature, with the subduction zone off central Java and to the interference of the Australian continent with the subduction process near the Lesser Sunda Islands. CARDWELL and ISACKS (1978) note a reduced number of shallow earthquakes as well as a paucity of focal mechanism solutions involving underthrusting for the eastern Sunda arc near Timor. They relate these factors to subduction of the leading edge of the Australian continental shelf. On the basis of 41 new and old focal mechanism solutions for the eastern Sunda arc, CARDWELL and ISACKS (1978) propose a model of a slab that is contorted and bent at depth as a result of curvature of the eastern end of the arc. North-northwesterly convergence would, according to their model, translate the bend of the slab at the eastern end from the surface down to depths of 600 km. The north-northwesterly direction of convergence between the Australian plate and the Sunda arc inferred by Cardwell and Isacks contradicts the northeasterly direction computed by LEPICHON (1968), MORGAN (1972), and MINSTER *et al.* (1974). Their poles of rotation were based on magnetic anomalies and hot spot traces (MINSTER *et al.*, 1974), with no instantaneous slip vectors derived from focal mechanism solutions. There is a dearth of shallow underthrusting mechanisms for the entire Sunda arc: only four have been reported, all of which lie near Sumatra (FITCH, 1972; CARDWELL and ISACKS, 1978). CARDWELL and ISACKS (1978) cite seven volcanoes at the eastern end of the arc near the Weber basin as evidence of north-northwesterly convergence. They propose that Southeast Asia is not part of the Eurasian plate, but that there is a separate Southeast Asian plate with a southeasterly direction of motion relative to the Australian plate. Their model is consistent with the direction of oblique convergence to the north-northwest between the Australian plate and Sunda arc near the Andaman Islands. Other evidence that may support the model of Cardwell and Isacks is the observed dextral slip on the Semangko fault in Sumatra and south-southeasterly extension in the Andaman basin (RODOLFO, 1969).

The tectonics of the eastern Sunda arc are complex and still poorly understood. The lack of great thrust earthquakes at shallow depths may be related to the contortions of the dipping slab as it bends around the eastern Banda Sea and to the interference of the Australian continental shelf, Christmas Island ridge, and other

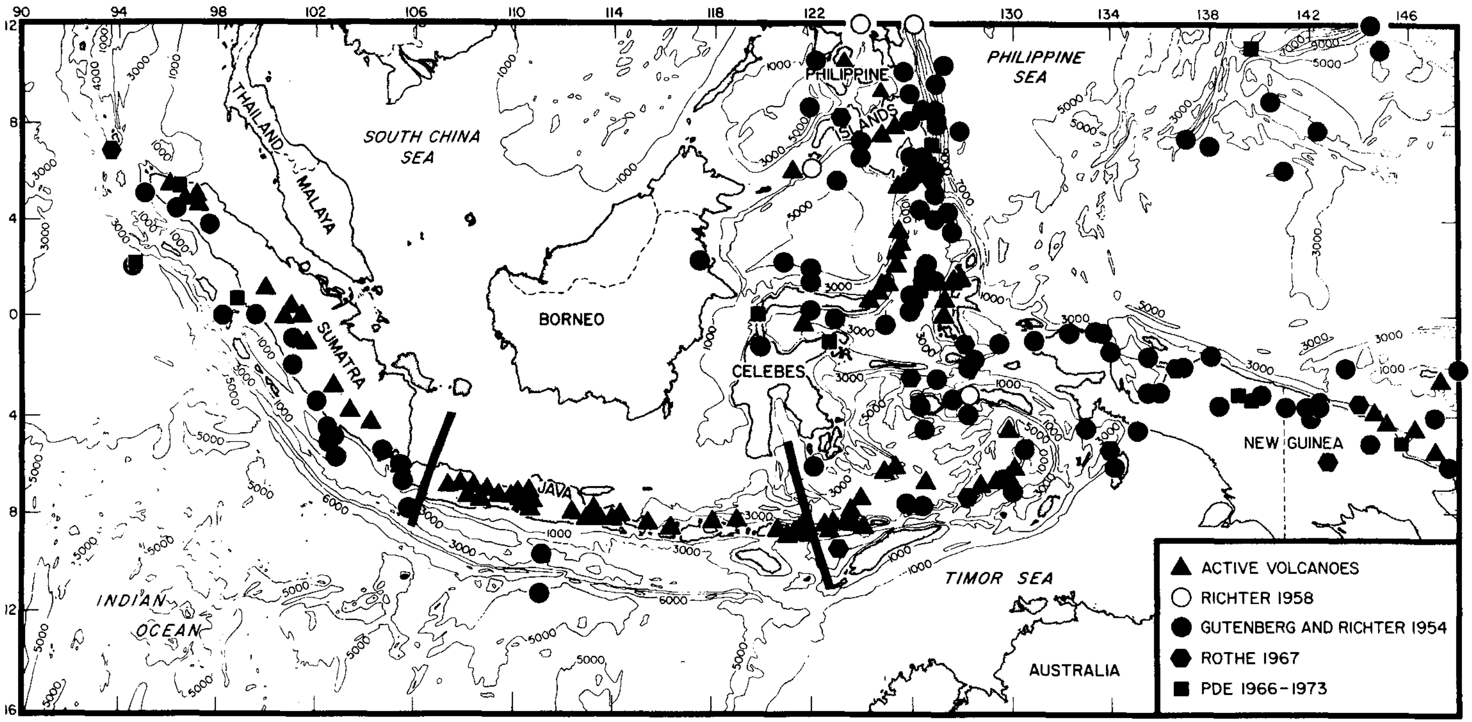


Figure 20

Large earthquakes (M $\geq$ 7.5), of shallow focus, of the twentieth century along the Sunda arc (from KELLEHER and McCANN, 1976). Note the gap in activity for large shocks from 106° to 122°E. This region has no history of large destructive earthquakes. The region offshore from Sumatra, however, has been the site of great tsunamic earthquakes of extensive length and should be considered as a potential site for the occurrence of future large and great events.

bathymetric features with the subduction process. Although more work is needed to understand the tectonic regime, there is clearly a lack of historic great thrust earthquakes and associated tsunamis for the eastern Sunda arc. Hence, on the map of seismic potential this region is shown as category 5. Under the present tectonic regime the lack of great earthquakes recorded for the past 200 years may be expected to prevail during the forseeable future. Nevertheless, the reasons we give for believing this region differs tectonically from subduction zones with a record of great shocks should be regarded as a hypothesis and not proven fact. The basins north of Java and the Lesser Sunda Islands do not appear to be characterized by high heat flow and recent extension as they do behind other arcs such as the Marianas and South Sandwich, which also do not have a history of great shocks. If current spreading along marginal basins is the important factor in governing the occurrence of great shocks, then the eastern end of the Sunda arc may, in fact, have the potential for great shocks to occur.

At the Sunda Strait the arc undergoes a sharp bend from an easterly to northwesterly strike, and the direction of relative plate motion becomes increasingly oblique. The maximum depth of earthquakes changes abruptly from 600 km east of the Sunda Strait to about 200 km on the west side. FITCH (1972) proposes that the Sunda Strait may be undergoing extension on the seaward (convex) side and compression on the landward side, which may be related to recent shallow activity in the area. The Semangko fault of Sumatra shows evidence of historic and recent right-lateral movement (RODOLFO, 1969), while the Lambang fault east of the strait, in Java, which strikes parallel to the arc, may indicate left-lateral offset (KATILI, 1970).

FITCH (1972) proposes a model for island arcs with oblique convergence in which there is decoupling along a pre-existing zone of weakness. According to his model, strike-slip movement occurs along a fault between the trench and volcanic front, while underthrusting takes place at the trench. In the western Sunda arc underthrusting is occurring along the inner wall of the Java trench, while horizontal shear is simultaneously occurring along the Semangko fault.

Great shallow events that are thought to involve thrust faulting are recorded historically, as well as large inland shocks associated with the Semangko fault. Rupture lengths for great historic earthquakes for the nineteenth and twentieth centuries, as estimated by reports of tsunami destruction by VISSER (1922) tend to be several hundred kilometers long (Fig. 21). Also shown in Fig. 24 are locations of events with reported intensities of VIII or IX from SOEDATI (1962) for the period 1900 to 1960, which appear to be associated with the Semangko fault, as well as events of magnitude $M \geq 7.5$ that occurred during the twentieth century. [Intensities reported by Soedati are based on the Rossi-Forel scale; equivalent Modified Mercalli intensities for R.-F. VIII to IX are VII to IX (RICHTER, 1958)]. It is evident that Sumatra, unlike Java and the eastern Sunda Islands, has had a history of large to great events, both shallow thrusting (associated with the trench), and strike-slip (associated with the Semangko fault).

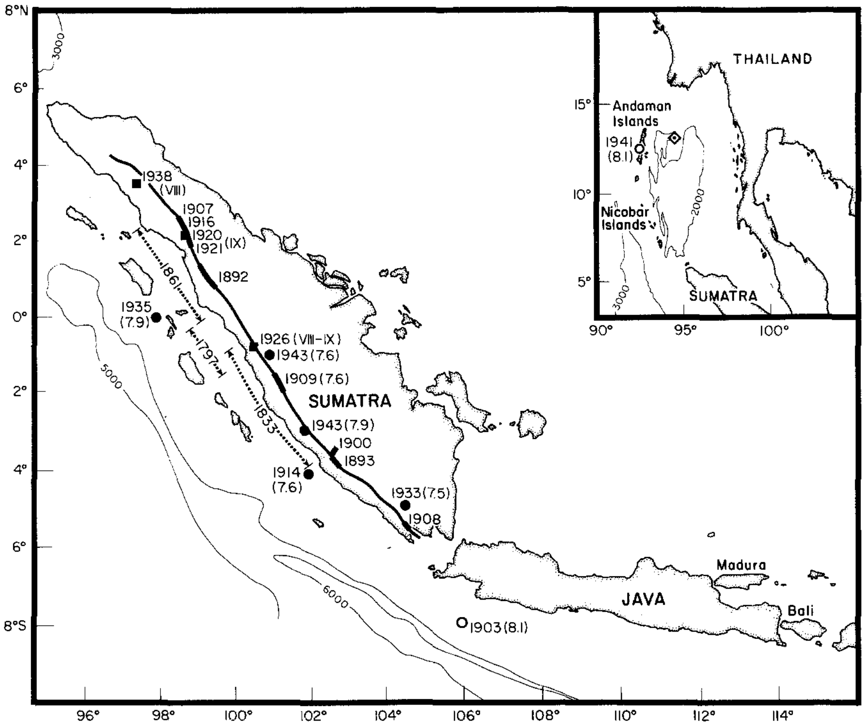


Figure 21

Large historic earthquakes in the Sunda region. Dashed regions are areas of reported damage by tsunamis associated with historic great earthquakes (VISSER, 1922). The Semangko fault in Sumatra is shown as a sinuous line. Heavily lined portions along the Semangko fault are areas reported by VISSER (1922) to have ruptured in the years shown. Squares are reported intensities VIII and IX (Rossi-Forel) by SOEDATI (1962). Solid circles are relocated epicenters originally reported by GUTENBERG and RICHTER (1954); open circles are Gutenberg and Richter epicenters which have not been relocated. Diamond with dot in insert is the only historically active volcano in the Andaman-Nicobar Islands region, reported by RODOLFO (1969).

The extent of the rupture zone of the 1861 event, as inferred from reports of tsunami inundation by VISSER (1922) may be as great as 800 km (Fig. 21). If this estimate is credible, then that event is the most recent great event that can be said to have ruptured a major portion of the plate boundary and relieved significant tectonic strain. Hence, the western portion of the Sunda arc off Sumatra has not experienced a great event for over 100 years, and is assigned a seismic potential of category 1.

The portions of the Semangko fault with historically recorded activity all appear to have ruptured between 30 and 100 years ago, and hence are assigned to category 2. [The oldest reported large event along the fault occurred in 1892 (VISSER, 1922), which resulted in 2 meters of offset (RODOLFO, 1969).]

The 1941 earthquake in the Andaman Islands, although given a magnitude of 8.7 by RICHTER (1958), was probably not of comparable size as is discussed later. Its rupture zone does not appear to be several hundred kilometers long.

The portion of the Sunda arc along the Andaman and Nicobar Islands is characterized by oblique convergence. There are no great earthquakes or extensive tsunamis reported historically. The earthquake of 26 June 1941, which occurred west of the Andaman Islands, is believed to have been the largest event of this century recorded along the western Sunda arc; however, it is probably not a great event. SINDVAHL *et al.* (1978), using RICHTER's (1958) magnitude of 8.7, infer a rupture zone of 800 km, but these estimates of magnitude and rupture length appear to be far too high.

GELLER and KANAMORI (1977) re-examined the methods by which RICHTER (1958) and GUTENBERG and RICHTER (1954) computed magnitudes. An appendix to the paper of Geller and Kanamori contains corrected values of $M_s$. The 1941 event has a corrected $M_s$ of 7.7. KANAMORI (1977b) calculated revised magnitudes, $M_\omega$, based on seismic moment for great earthquakes with extensive rupture zones. His new magnitude scale, $M_\omega$, extends the $M_s$ scale, which saturates at magnitudes near 8.0, so as to accurately measure the moment and energy release of large shocks. Events with rupture lengths of about 800 km have $M_\omega$ values near 9.0 on his scale. The $M_s$ and $M_\omega$ scales coincide up to approximately magnitude 8.0. Thus, an event such as the 1941 shock of $M_s$ near 7.7, is unlikely to have a rupture zone of hundreds of kilometers, as its magnitude would have been recalculated, by Kanamori's method, near 9.0. Also, there are no reports of tsunamis associated with the 1941 event, and few reported aftershocks. The lower re-calculated magnitude, general lack of aftershock activity and macroseismic effects for the 1941 event lead us to believe that SINDVAHL *et al.* (1978) over-estimated the extent of the rupture of this event. Hence, there may not appear to be great earthquakes associated with the Sunda arc in the Andaman-Nicobar region.

The northwestern part of the Sunda arc near the Andaman and Nicobar Islands is assigned to category 4 because plate convergence occurs subparallel to the subduction zone in that region. A dipping seismic zone extends to depths of 100 to 200 km beneath the Andaman-Nicobar ridge, but the dominant sense of motion as derived from the few available focal mechanism solutions for this region is dextral strike-slip, with nearly vertical nodal planes striking northwest (FITCH, 1972). RODOLFO (1969) notes only one historically active volcano in the area near the island of Narcondam (Fig. 21). He cites evidence that extension has been occurring in the Andaman basin since the Late Miocene. High heat flow measured in the Nicobar rift, subparallel north-northeast trending magnetic anomalies in the Andaman basin, and relatively thin (1.5 km) sediment cover in the southern Andaman basin are presented by Rodolfo as evidence of extensional tectonics, trending NW. FITCH (1972) compared the Andaman basin to the model proposed by SCHOLZ *et al.* (1971) for the western United States. As compressive stress was relieved off the coast of California in the late Tertiary, and the dominant plate motion changed to strike-slip along the San Andreas fault zone, extension was triggered in the Basin and Range province. FITCH (1972) suggests that the submarine continuation of the Semangko fault might be comparable to the San Andreas. The evidence for southeasternly extension in the Andaman basin cited by Rodolfo (1969) appears to support Cardwell and Isack's (1978) northwesterly motion of the Australian plate with respect to the Sunda arc and the existence of separate Eurasian and Southeast Asian plates.

The Andaman portion of the Sunda arc appears to be tectonically similar to the westernmost Aleutians, where plate convergence is subparallel to the arc and high heat flow is present behind the arc (Cooper *et al.*, 1977).

### (*j*) *Southwest Pacific*

During this century large shallow earthquakes have not been evenly distributed along the subduction zones of the southwest Pacific. Figures 22, 23 and 24 are from a study of large earthquakes occurring along the shallow thrust boundaries of the southwest Pacific arcs (McCann and Kelleher, 1976). The details of the tectonics in this complicated region will not be presented here. These figures show either the clustering or the absence of known large events along various convergent boundaries. Since these variations appear to be spatially related to the regional tectonic setting, they may represent long-term features of the seismicity.

A triple junction between the South Bismark, Solomon Sea and Pacific plates is located near 152°E, 6°S (Johnson and Molnar, 1972) (Fig. 22). The intense clustering of large events in this region appears to result from the rapid subduction of the Solomon

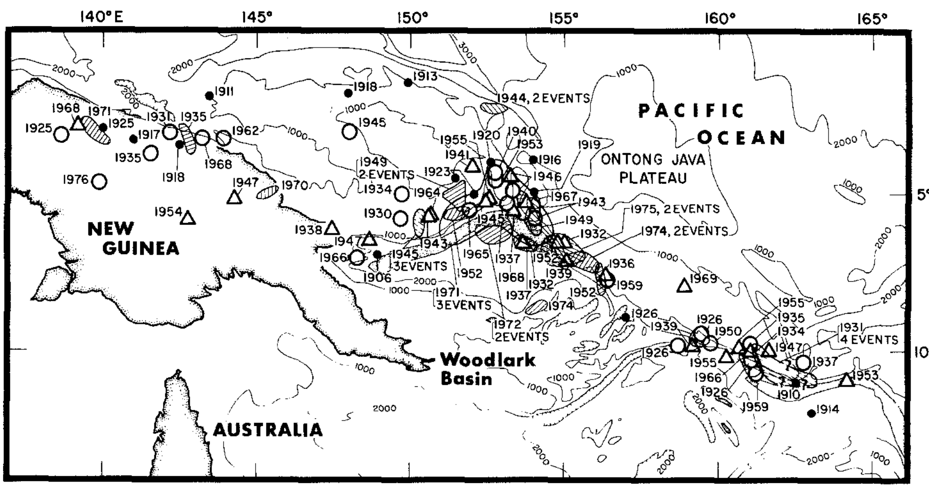


Figure 22

Large (M $\geq$ 7.0) earthquakes of the New Guinea-Solomon area for the twentieth century (from McCann and Kelleher, 1976). The large gap in activity near 157°E coincides with the segment of the arc subducting the Woodlark rise, a spreading center. This gap may therefore persist indefinitely. Contours are in fathoms. Solid circles are locations taken from Gutenberg and Richter (1958). Open circles are relocated epicenters; hatched regions are rupture zones; triangles are events with uncertain depth determinations. Small circle pattern in the region of the trench delineates water depths greater than 3000 fathoms.

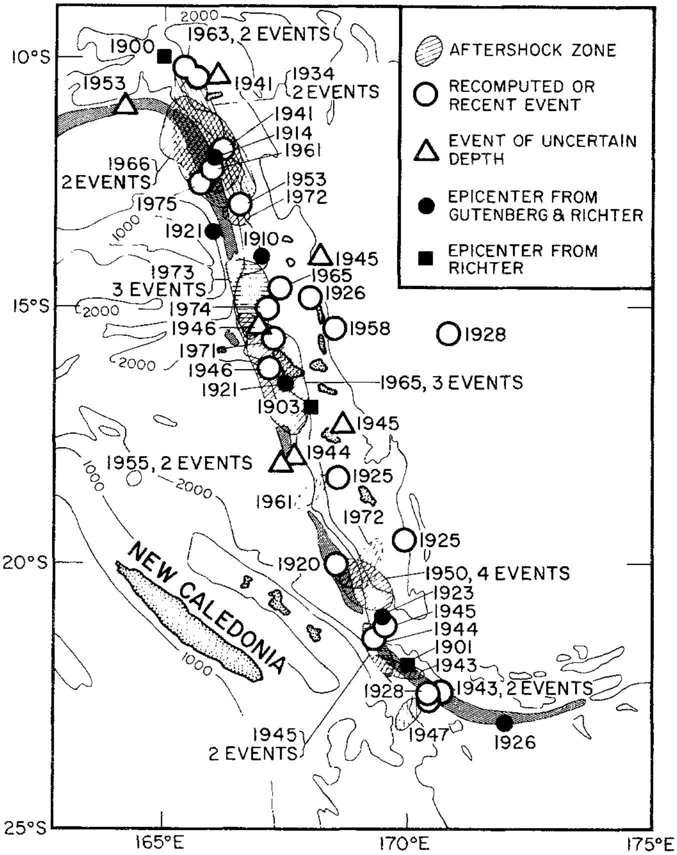


Figure 23

Large (M ≥ 7.0) earthquakes of the twentieth century and their associated aftershock zones for the New Hebrides arc (after McCann and Kelleher, 1976). Note the lack of recent large shocks to the south of 20°S. Heavily hatched regions is that portion of the New Hebrides trench with water depths greater than 300 fathoms.

Sea plate near this triple junction. This intense activity, therefore, may continue indefinitely as it appears to stem from a tectonic regime characterized by high strain rates. The large gap in the central portion of the Solomon arc (near 157°E) occupies that segment of the arc that intersects the Woodlark rise, a series of spreading centers that forms the boundary between the Solomon Sea and Australian plates. Lithosphere in that area appears to be too weak to store the energy sufficient to generate large earthquakes. The region between 159° and 161°E recently ruptured during a series of large events. The remaining segment of the Solomon arc from 161° to 163°E, which has not ruptured in over 30 years, is probably the most likely site for future large shocks along this arc.

The rupture lengths of large shallow earthquakes are only moderately long (maximum dimensions less than 150 km) along all of the subduction zones of this

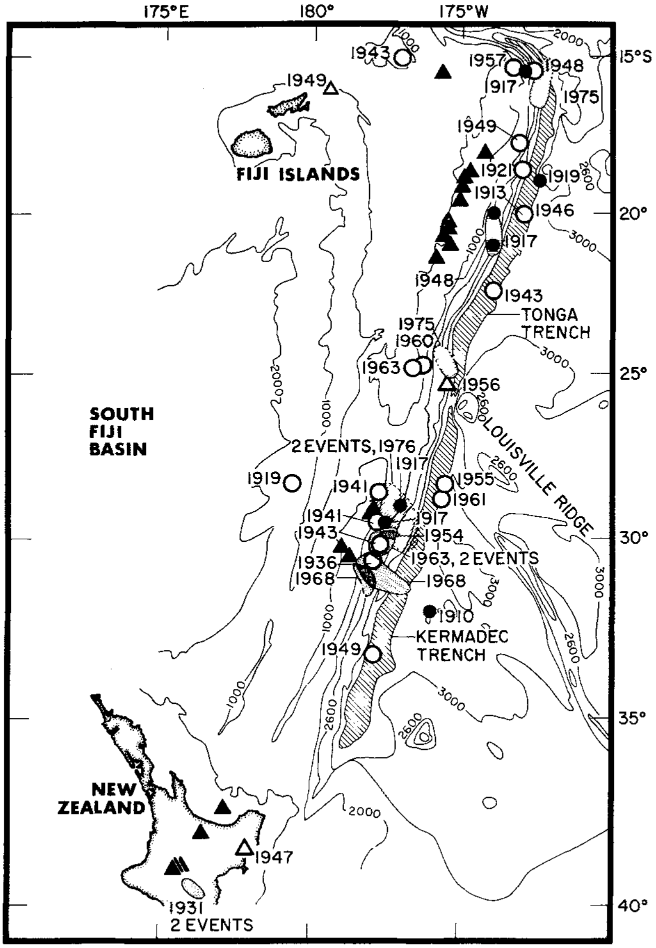


Figure 24

Large (M ≥ 7.0) earthquakes of the twentieth century along the Tonga and Kermadec arcs (from McCANN and KELLEHER, 1976). Note the sparse activity for large earthquakes in the Tonga arc region and the smaller number of large shocks from 32° to 38°S and from 23° to 28°S. These gaps appear to be of tectonic origin and may persist indefinitely. Symbols same as in Fig. 26.

region. These limited dimensions of rupture zones may result from the narrow width of interface, the zone of contact between the underthrust and overthrust plates, along the convergent boundaries. There is an observable variation in the maximum dimension of rupture zones from north to south along the New Hebrides arc. Rupture zones with large maximum dimensions occur in the northern portion of the arc. Heat flow values in this region are lower on both the incoming sea-floor and on the overthrust plate compared to values observed to the south. The very southern portion of the arc has a poorly-developed island arc platform and may only be Quaternary in age (KARIG and MAMMERICKX, 1972). In this region rupture zones are smallest when compared with the rest of the arc. The change in the rupture dimensions may result from a narrower interface between the younger plates of the southern portion of the arc.

The distribution of large shocks along the Tonga and Kermadec arcs is very uneven. In the Kermadec arc, which is about 1000 km long, large shocks of this century are clustered along a 300 km segment. In the Tonga arc earthquakes are distributed more evenly but are less frequent than in the Kermadec arc. The last large shock in the central portion of the Tonga arc occurred in 1949. This uneven distribution may not be random as the gaps in activity lie landward of regions subducting relatively shallow sea-floor. As discussed earlier, the subduction of bathymetric highs may inhibit the subduction process; regions interacting with bathymetric highs typically lack large shocks, and in some cases gaps appear in the volcanic chain. Note the good correlation among the presence of volcanoes, large shocks and low lying sea-floor ($>3000$ fathoms) for the Tonga and Kermadec arcs.

Category assignments in the southwest Pacific are numerous and they apply to only short segments of each arc. There is a lack of large shocks for much of the Tonga-Kermadec arc system. Thus much of it lies in category 3 or 5. The region near New Britain (and Taiwan north of the Philippines) has been coded with categories 1 and 2 to indicate that the region is extremely active for large shocks which repeat at intervals generally less than 30 years.

## 5. *Discussion*

### (*a*) *Successes and problems*

Seismic gap analysis has led to the successful forecasting of the sites of a number of large earthquakes (Table 1). In each case the area was discussed in the literature prior to occurrence of the shock. No great earthquakes have occurred in at least the last 5 years on segments of plate boundaries that were thought to be of low seismic potential. The general success of this method can probably be attributed to the regular manner in which large shocks occur along the shallow portions of simple plate boundaries.

Two unusual events, however, occurred in the Kuril Islands in 1975 and 1976. Their rupture zones are located near a segment of the arc that ruptured during a somewhat deeper-than-normal (about 50 km) event in 1958. Although each event was of magnitude $M = 7.0$, they were associated with a large number of aftershocks that were spread over about a 400 km segment of the plate boundary. Events of this size cannot be forecast for this region because most of the seismic energy release occurs during great earthquakes ($M \geq 7\frac{3}{4}$). Thus, despite the large aftershock areas, these events probably did not contribute significantly to the cumulative seismic motion between the plates.

KANAMORI (1977a) examined seismograms of the 1975 event and found an excess of long-period energy (20 sec waves) compared to that for shorter periods. Such an excess of long-period energy appears to be characteristic of a long rise time for motion at a given point on the rupture zone, which appears to be associated with larger than normal tsunamis. It may reflect a reduced coupling between the oceanic and continental lithospheric plates in that area (KANAMORI, 1977a). Kanamori also reports that a significant amount of plate motion in the southern Kuril Islands occurs aseismically.

Historic records of the southern Kuril Islands indicate that a portion of the plate boundary may not have ruptured during the sequence of great earthquakes in the nineteenth century. The one segment that may not have ruptured was the site of the large shock in 1958. This event is anomalous since the hypocenter of the main shock and the depths of aftershock activity (40–60 km) were deeper than is commonly observed. Slippage during the 1958 event was greater than that during each of the other events of the twentieth century that ruptured segments affected in the nineteenth century (KANAMORI, 1978). The excess slippage experienced during that event, therefore, may have released strain energy stored during the eighteenth and nineteenth ceuturies but not released when the region was skipped during the great earthquakes of the nineteenth century. This region may then be capable of accumulating varying amounts of strain energy. This may reduce the reliability of the seismic gap techniques used here. In that area very old oceanic lithosphere is being subducted. That may account for the greater depth at which earthquakes are reflecting relative motion between plates. Hence, in that region the assumption that great shocks break the entire range of the plate interface may not always hold.

Problems are presented in applying gap theory to regions in which the relative plate motion occurs along a series of faults rather than a single boundary. Strain accumulation on these faults may not be a steady state process, and motion along these faults may be episodic, jumping from one fault to another in time. Although primary faults can be assigned to various categories of seismic potential, their splay faults cannot be accurately classified, and the presence of splay faults reduces our confidence in forecasts for these regions. Thus, some of the regions assigned to categories 2 or 6 may have a potential for large shocks to occur along splay faults that are subparallel to the main plate boundary.

### (*b*) *Future research using existing data*

Much work still needs to be done to improve the historic record of many regions along plate boundaries. Those areas shown in category 3 in Fig. 1 could probably benefit most from a more complete historical record, as they could be placed into a more clearly defined category of seismic potential. Much of the historic record needs to be re-examined in the light of our present understanding of plate tectonics, and the hypotheses of aseismic plate movement and modified subduction. We also need a clearer understanding of the parameters involved in the generation of tsunamis. In some cases the damage from a tsunami is more extensive than that from the earthquake itself. If tsunamic earthquakes (KANAMORI, 1977a) could be regionally and tectonically categorized, tsunami risk could be more definitely forecast. An examination of historic records appears to be one of the most effective methods for achieving a better understanding of tsunamis.

The larger and generally more complete historical record of earthquakes and tsunamis in Japan yield important information about the recurrence times of large shocks for particular segments of the plate boundary near Japan. Some information about repeat times of historic events or those deduced from geologic movements along particular faults during the past few thousand years is, of course, very much needed for most other plate boundaries. More precise repeat times would be preferable, in determining the seismic potential of a region, to the use of an arbitrary cutoff of 100 years. At present there is much scatter in many of the reported recurrence times even for regions with a long historical record. We must gain confidence in the completeness of the historical records; only then can we be certain that the apparent scatter in observed repeat times is not caused by the missing of events.

Variations in intraplate activity near active seismic zones need to be studied more thoroughly. CARR (1977) examined large shocks in Central America, differentiating between inter- and intraplate earthquakes. However, no conclusions were drawn from the variation in activity. The occurrence of intraplate events in either the frontal arc or volcanic arc regions may reflect the state of stress on the thrust zone. A recognizable spatial pattern may be observed before large thrust events as in Japan (SHIMAZAKI, 1976, 1978).

Since the advent of the World-Wide Standardized Seismograph Network in 1962, a large body of epicentral data has been collected for small and moderate size shocks. The spatial-temporal variations in this data set should be examined for each region of high seismic potential. Marked quiescence for small and moderate size shocks has also preceded many large shocks; examination of this data set may well reveal a similar behavior for some of the gaps in Fig. 1. A search of this and other types of premonitory phenomena should be made, especially for the regions assigned to category 1. One set of data that is easily researched is the level of volcanic activity for regions near seismic gaps. Variations in eruptive activity and in the level of lava lakes (NAKAMURA, 1975; KIMURA, 1976; CARR, 1977; KIMURA, 1978) were observed before some great and major earthquakes. More research is needed to properly document the strength of known volcanic eruptions so as to reduce the large amount of what appears to be noise in the historic record of volcanic activity.

### (c) *Collection of new data in seismic gaps*

A concentrated effort is recommended to collect a variety of possible premonitory data for regions assigned to category 1 in Fig. 1. This would undoubtedly increase our chances of 'catching' pre-seismic changes associated with large earthquakes. Detailed studies may also help us to identify those regions in category 1 that are highly stressed.

*In-situ* stress measurements using the hydrofracturing technique and determinations of stress drops for moderate size shocks could provide data pertinent to understanding the state of stress in seismic gaps or of possible stress concentrations near the edges of seismic gaps. HOUSE and BOATWRIGHT (1979) infer stress drops of about 500 bars for two moderate size shocks in the seismic gap near the Shumagin Islands and the Alaska Peninsula. Their study is another attempt to ascertain the seismic potential for large shocks from the state of stress in major seismic gaps (see WYSS and BRUNE, 1971).

Many of the regions assigned to the highest categories of risk in Fig. 1 are situated largely in submarine areas along active continental margins. There is a great need to develop the capability of measuring crustal movements (both co-seismic and premonitory) and strong ground motion as well as to record small shocks on the ocean floor in some of the critical seismic gaps of Fig. 1.

Many seismic gaps, including those in category 1, are only sparsely instrumented. Several strong-motion instruments could be deployed and level lines constructed in many of the areas designated as category 1 with a relatively small expense.

### (*d*) *Regions of particular interest*

Several of the regions assigned high seismic potentials in Fig. 1 (categories 1 to 3) are of particular interest since some observations indicate that these regions may be exhibiting some types of long-term premonitory behavior which may be indicative of future large shocks.

*Sumatra.* The last great earthquake along the Sumatran coast occurred in 1861. It was associated with a seawave that destroyed coastal villages for several hundred kilometers. Since the last great earthquake occurred more than 100 years ago, the area is assigned to category 1. During this century large ($M \geq 7$) shallow shocks along the coast of Sumatra clustered into two groups separated by some 500 km. The northern half of the zone of quiescence that lies between these clusters was the site of moderate (i.e., $6 < M < 6.9$) events until 1955. A single event recently occurred in this area in 1970. The southern portion of the quiescent region has not had any shallow events of magnitude 6 or above since at least the 1930s and possibly much longer. Small earthquakes (i.e., $M < 6$) recorded since 1962 have distributed themselves rather evenly along the shallow portion of the seismic zone.

A zone of quiescence bounded by a region of high seismic activity has been observed prior to large thrust and strike-slip type earthquakes (MOGI, 1969). The observations of the earthquake activity along the Sumatran coast are similar in several respects to those of previous researchers but there are some differences. The zone of quiescence for moderate and large shocks has only persisted since 1955; for larger shocks the quiescence extends back to the beginning of instrumental coverage (KRAUSE *et al.*, 1978). Shallow shocks of the last 15 years appear to cover the region rather evenly despite the clustering of the larger events. Thus, more work is needed to fully understand the significance of the zone of quiescence for large shocks between 0° and 5°S, near the coast of Sumatra, as the spatial pattern of occurrence of large and moderate earthquakes is similar to those observed prior to great earthquakes.

*Northern Lesser Antilles.* Great earthquakes ruptured an extensive portion of the northern Lesser Antillean arc in 1690 and 1843. In the absence of other data it is not unreasonable to take the time interval between these events as an indication of a recurrence time of about 150 years. In 1969 and 1974 large intraplate events occurred on the edges of the rupture area of the 1843 event. An increase of shallow interplate activity within Japan was observed prior to several large earthquakes along the nearby plate boundary (SHIMAZAKI, 1976, 1978). These events in Japan may be indicative of higher stress within the overriding plate prior to the occurrence of large interplate shocks. The intraplate shocks in the Lesser Antilles may represent a similar phenomenon.

Grand Soufrière on Guadeloupe in the northern Lesser Antilles has been active during the past few years. CARR (1977) suggests that this increase in volcanic activity may be indicative of a forthcoming great earthquake. That volcano erupted a few years before the last great shock in 1843.

*Shumagin Islands, Alaska.* This region has an unclear seismic history but may be the site of a future large shock. As discussed previously this region experienced events with high stress drops as inferred from the ratio $M_s/M_b$ and from information derived from local strong-motion records. This area was cited by KELLEHER (1970) as the next in a space–time progression of events that has moved north along the southeast coast of Alaska and then west along its southern coast. The volcano Pavlof (near 162°W, on the Alaskan Peninsula), which is adjacent to the gap, has shown increased eruptive activity in the last few years. Thus, despite the incomplete seismic history of this region there are several observations that could be interpreted as indicators of a high level of tectonic stress.

## 6. *Summary*

Plate tectonics has given us a fuller understanding of the mechanism by which strain energy is stored on plate boundaries and then suddenly released as large earthquakes. This understanding has led to successful forecasts of the sites of several large shocks along convergent and strike-slip boundaries. In this paper simple, major plate boundaries of the Pacific, Caribbean, Sunda and South Sandwich regions are classified into six categories of seismic potential for large earthquakes for the next few decades. These categories reflect our knowledge and assessment of the historic record for large earthquakes, our understanding and speculations about tectonic regimes and the length of time since the last large earthquake. Several hypotheses are used to infer the seismic potential for regions that have an ambiguous history of great shocks. Since categories 3 and 5 are based on hypotheses and not on proven fact, the assignment of regions to those categories should be regarded as tentative. The degree of seismic potential is more certain for areas that have experienced large earthquakes, i.e., for categories 1, 2 and 6. Several of the areas assigned the highest risk are poorly instrumented and obviously deserve high priority for intense study.

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### *Note added in proof*

Since the original manuscript for this paper was completed in May 1978, several large shocks have occurred in seismic gaps. They are:

23 August 1978; Costa Rica, 7.0
5 November 1978; Oaxaca, Mexico, 7.8
29 November 1978; Solomon Islands, 7.5
28 February 1979; Southern Alaska, 7.7
14 March 1979; Guerro, Mexico, 7.6

The events, although not thoroughly studied as yet, indicate that at least portions of several seismic gaps have ruptured recently and thus should be added to the list of successful forecasts.

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