Bulletin of the Seismological Society of America, Vol. 72, No. 4, pp. 1241-1253, August 1982

VARIABLE RUPTURE MODE OF THE SUBDUCTION ZONE ALONG THE ECUADOR-COLOMBIA COAST

By Hiroo Kanamori and Karen C. McNally

ABSTRACT

Three large earthquakes occurred within the rupture zone of the 1906 Colombia-Ecuador earthquake (M_w = 8.8): in 1942 (M_S = 7.9); 1958 (M_S = 7.8); and 1979 (M_S = 7.7). We compared the size and mechanism of these earthquakes by using long-period surface waves, tsunami data, and macroseismic data. The 1979 event is a thrust event with a seismic moment of 2.9×10^{28} dyne-cm, and represents subduction of the Nazca plate beneath South America. The rupture length and direction are 230 km and N40°E, respectively. Examination of old seismograms indicates that the 1906 event is also a thrust event which ruptured in the northeast direction. The seismic moment estimated from the tsunami data and the size of the rupture zone is 2×10^{29} dyne-cm. The 1942 and 1958 events are much smaller (about $\frac{1}{5}$ to $\frac{1}{10}$ of the 1979 event in the seismic moment) than the 1979 event. We conclude that the sum of the seismic moments of the 1942, 1958, and 1979 events is only $\frac{1}{5}$ of that of the 1906 event despite the fact that the sequence of the 1942, 1958, and 1979 events ruptured approximately the same segment as the 1906 event. This difference could be explained by an asperity model in which the fault zone is held by a discrete distribution of asperities with weak zones in between. The weak zone normally behaves aseismically, but slips abruptly only when it is driven by failure of the asperities. A small earthquake represents failure of one asperity, and the rupture zone is pinned at both ends by adjacent asperities so that the effective width and the amount of slip are relatively small. A great earthquake represents failure of more than one asperity, and consequently involves much larger width and slip.

Introduction

Many recent studies indicate that a long segment of a subduction zone sometimes ruptures in a single great earthquake, but at other times it breaks in a series of smaller earthquakes abutting to each other. One of the best examples is seen for the subduction zone off the coast of Ecuador-Colombia. A great earthquake occurred in 1906 along the coast of Ecuador-Colombia (M_S = 8.7, estimated M_w = 8.8) (Figure 1). Kelleher (1972) estimated the rupture length to be about 500 km on the basis of the macroseismic data. Abe (1979) estimated the tsunami magnitude M_t to be 8.7 which is consistent with Kelleher's estimate of the size of the rupture zone. Approximately this same segment ruptured again during the last 37 yr in three large earthquakes which occurred in 1942 (M_S = 7.9), 1958 (M_S = 7.8), and 1979 (M_S = 7.7). Although there is some uncertainty in the interpretation of the old events, the evidence is strong that this segment of the Ecuador-Colombia subduction zone behaved differently from sequence to sequence.

Similar examples are found for southwest Japan along the Nankai trough (Imamura, 1928; Ando, 1975; Seno, 1977) and along the Aleutian Islands (Sykes *et al.*, 1980). However, in these examples, no instrumental data are available for the older events, and the details of the rupture mode are unknown.

The purpose of the present paper is to investigate the nature of this type of variable rupture behavior by studying the Ecuador-Colombia sequence for which instrumental data are available.

THE 1979 EVENT

Kanamori and Given (1981) made a detailed analysis of this event by using 15 Rayleigh waves recorded at seven IDA (International Deployment of Accelerograph) stations. Since the details are given in Kanamori and Given, we briefly summarize the results in the following.

Kanamori and Given (1981) inverted the Rayleigh-wave spectra at the period of 256 sec by using a moment tensor source placed at a depth of 33 km. The moment tensor thus obtained was decomposed into the major and the minor double couple (Gilbert, 1980). The seismic moment of the minor double couple is 0.2 per cent of

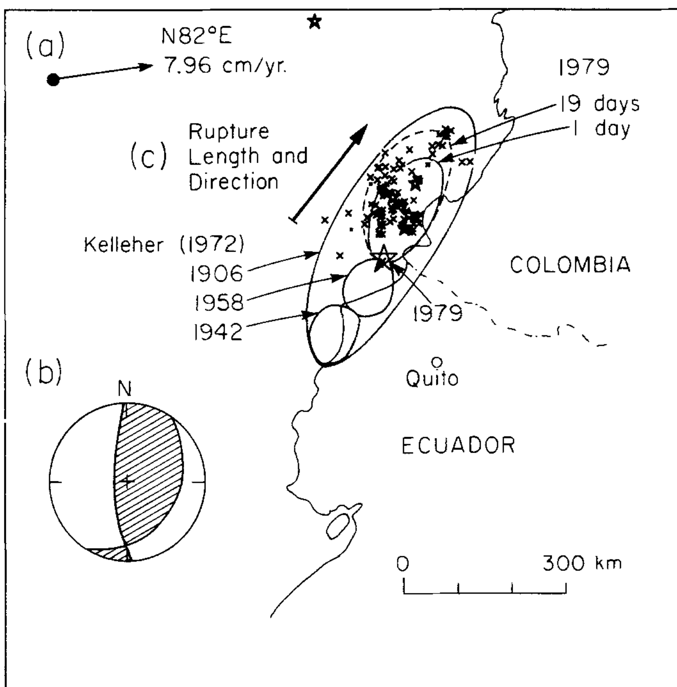

FIG. 1. Rupture zones of the 1906, 1942, and 1958 Colombia-Ecuador events (Kelleher, 1972). Aftershock zones 1 day and 19 days after the 1979 event (asterisk) are shown. The rupture length and the direction (c) are determined from azimuthal variation of group arrival times of Rayleigh waves (Kanamori and Given, 1981). The mechanism diagram (b) is from Kanamori and Given (1981). The lower focal hemisphere is shown. *Hatched areas* show compressional quadrants. The arrow shown in (a) indicates the convergence vector between the Nazca and South American plates at the epicenter of the 1979 event (after Minster and Jordan, 1978).

that of the major double couple, and is considered negligible. As shown by Figure 2, the fault geometry of the major double couple is consistent with the first-motion data obtained from the WWSSN long-period seismograms. The strike of the low-angle plane is parallel to the trench axis, and the mechanism is consistent with subduction of the Nazca plate beneath South America. A similar mechanism has been reported by Herd *et al.* (1981).

If the low-angle plane dipping towards SE is taken to be the fault plane, then the slip direction is in nearly EW direction and is consistent with the motion of the Nazca plate with respect to the South American plate determined by Minster and Jordan (1978). A seismic moment of 2.9×10^{28} dyne-cm ($M_w = 8.2$) is obtained.

Since the size of the rupture zone is critical for the present discussion, we made a special effort to determine it by using the observed long-period Rayleigh waves. Usually, the size of the aftershock area expands as a function of time, which results in the uncertainty of the estimate of the rupture zone. The directivity method developed by Ben-Menahem (1961) is often used for the determination of the rupture length, but this event is not large enough to bring the directivity spectral holes in the period range with high signal-to-noise ratio. We, therefore, used the azimuthal variation of group arrival times of Rayleigh waves to determine the rupture length.

Figure 3 compares band-passed synthetic seismograms computed for the seven IDA stations used with the band-passed observed records. The band-pass filter is centered at about 270 sec, and the synthetics are computed for a point source placed at the epicenter. It is seen that the waves which propagated in the SW azimuth (R_2

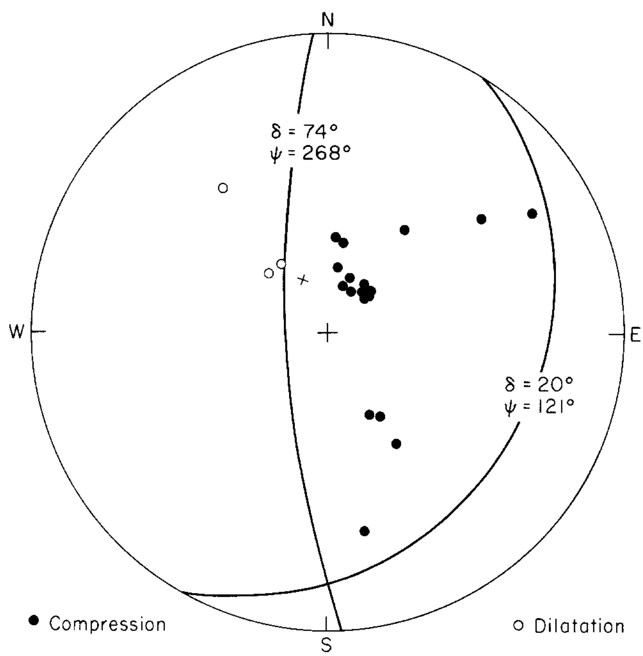

FIG. 2. Focal mechanism (stereographic projection of the lower focal sphere) of the 1979 Colombia earthquake determined by a moment tensor inversion (Kanamori and Given, 1981). The P-wave first-motion data are obtained from the WWSSN records. δ is the dip angle, and ψ is the dip direction.

at HAL and ESK, R_3 at TWO) were delayed by about 65 sec with respect to those propagated in the NE azimuth (R_3 at HAL and ESK, R_4 at TWO). The waves which propagated in NW and SE azimuths do not show significant delays (CMO and SUR). This pattern of group delays clearly indicates rupture propagation in the NE direction. Also, the observed trains are delayed by 58 sec on the average with respect to the synthetics computed for a point source. For a unilateral fault with the rupture length L, the phase delay at a station in the azimuth θ from the rupture direction is given by (Ben-Menahem, 1961)

$$X = \frac{\omega L}{V}\left(1 - \frac{V}{C}\cos\theta\right)$$

where V is the rupture velocity, C is the phase velocity, and ω is the angular frequency. The group delay time is then obtained by

$$\tau_g = \frac{dX}{d\omega} = \frac{L}{V}\left(1 - \frac{V}{U}\cos\theta\right) \tag{1}$$

where U is the group velocity. Using (1), the range and azimuthal average of τ_g can be written respectively by

$$\Delta\tau_g = 2\,\frac{L}{U} \tag{2}$$

and

$$\bar{\tau}_g = \frac{L}{V}. \tag{3}$$

Since $U = 3.6$ km/sec at $T = 225$ sec, we estimate $L = 230$ km from (2), and $V = 2$ km/sec from (3) using the observed values of $\Delta\tau_g$ and $\bar{\tau}_g$. The best-fit rupture direction is N40°E. The rupture length and the rupture direction are compared with the aftershock area in Figure 1. The rupture length determined from the group delay

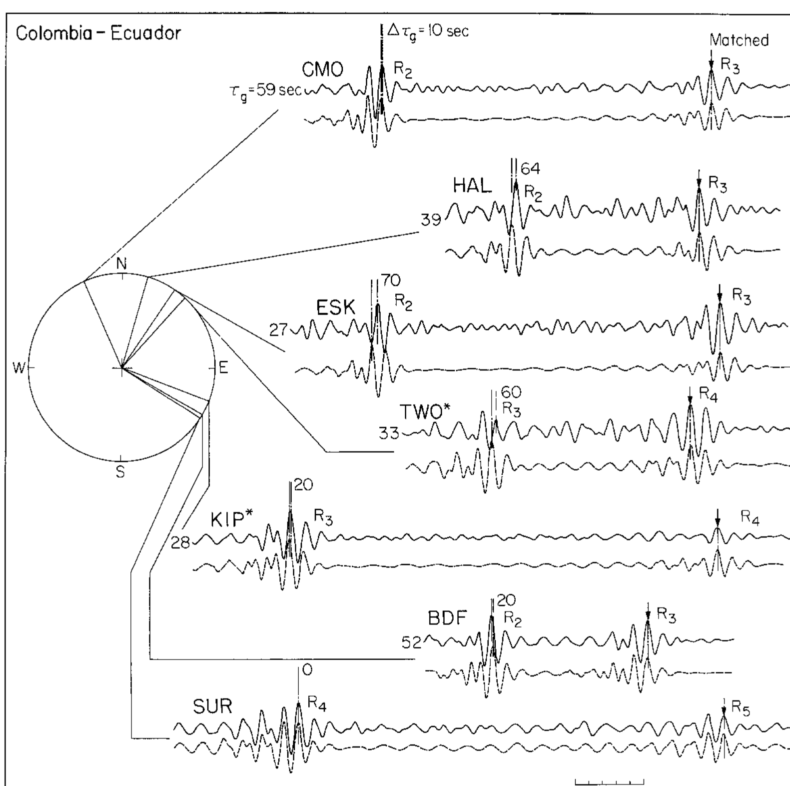

FIG. 3. The band-pass (150 to 1500 sec)-filtered observed (solid curve) and synthetic (dashed curve) seismograms of the 1979 event. The observed and synthetic traces are matched at the point indicated by an arrow, and τ_g is the delay time of the observed trace with respect to the synthetics. $\Delta\tau_g$ is the relative delay time of the first Rayleigh wave train on the observed trace with respect to the second (Kanamori and Given, 1981).

times is in good agreement with the extent of the aftershock area 1 day after the main shock, but is slightly shorter than that 19 days after the main shock. Thus, in this case, the 1-day aftershock area which is often used to estimate the fault area, appears to be a good approximation of the size of the rupture zone.

Comparison of Tsunami Data

Table 1 compares the tsunami data for the 1906 and 1979 events. No tsunamis at teleseismic distances are reported for the 1942 and 1958 events. Abe (1979) estimated the tsunami magnitude M_t of the 1906 event to be 8.7. Using Abe's method, $M_t = 8.2$ is obtained for the 1979 event. Thus, the tsunami data clearly indicate that the 1906 event is substantially larger than the 1979 event. The absence of reports of far-field tsunamis for the 1942 and 1958 earthquakes suggests that they are even smaller than the 1979 event.

TABLE 1

Maximum Tsunami Height H in Meters

Tide Station	1906	1979
Honolulu	0.2	0.04
Hilo	3.6	0.40*
Hakodate	0.18	0.09
Ayukawa	0.22	0.13
Kushimoto	0.29	0.10
Hosojima	0.19	—

* This value is given by H. G. Loomis (written communication, 1980).

Forty centimeters is reported as peak-to-peak amplitude in the NEIS monthly listing of earthquakes.

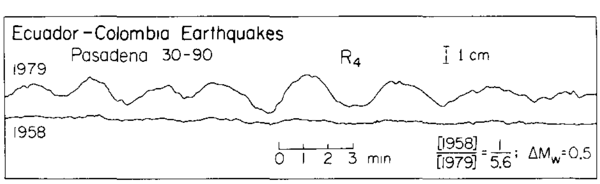

FIG. 4. Comparison of R_4 recorded by a Press-Ewing seismograph at Pasadena between the 1958 and 1979 events.

The 1942 and 1958 Events

The 1958 event was recorded by a Press-Ewing seismograph (30 to 90 sec) at Pasadena which also recorded the 1979 event. If we assume that these two events have approximately the same mechanism, we can estimate the seismic moment of the 1958 event from the amplitude ratio of long-period Rayleigh waves of the 1958 event to the 1979 event. As shown in Figure 4, the amplitude ratio is about 1:5.6 which would give a seismic moment of 5.2×10^{27} dyne-cm ($M_w = 7.7$) to the 1958 event.

Kelleher (1972) estimated the aftershock area of the 1958 event by relocating many of the aftershocks. The aftershock area determined by Kelleher (1972) is shown in Figure 1 and listed in Table 2. The empirical relation between the aftershock area and the seismic moment (e.g., Kanamori, 1977) suggests a seismic moment of 2.8×10^{27} dyne-cm ($M_w = 7.6$) which agrees reasonably well with that estimated from the Rayleigh-wave amplitude. We prefer the value 5.2×10^{27} dyne-cm estimated from the Rayleigh-wave amplitude but in any case, this event is signficantly smaller than the 1979 event.

No long-period seismogram is available for the 1942 event. However, the size of the aftershock area determined by Kelleher (1972) suggests that this event is of about the same size as the 1958 earthquake. We used Kelleher's aftershock area to estimate the seismic moment which is listed in Table 2. Although this estimate is indirect and is subject to some uncertainty, it is reasonable to conclude that this event is also significantly smaller than the 1979 event.

THE 1906 EVENT

The epicenter of this event was located at 1°N and 81.5°W by Gutenberg and Richter (1959). Since the location of the epicenter is critical for the determination of the rupture direction, we examined the original data used by Gutenberg and Richter which are now available in the form of microfiche (see Goodstein *et al.*, 1980).

TABLE 2

COLOMBIA-ECUADOR EARTHQUAKES SUMMARY

Event	Rupture Area* S_r (km^2)	M_S	M_0 (10^{27} dyne-cm)	M_w	M_t	$\bar{D}$, m
1906	(1.14×10^5)†¶	8.7‡	(200)	(8.8)	8.7§	(5.20)
1942	7.1×10^3¶	7.9‡	(3.2)	(7.6)	—	(1.30)
1958	6.6×10^3¶	7.8‖	5.2#	7.7	—	(2.30)
1979	2.8×10^4**	7.7††	29‡‡	8.2	8.1§§	(2.70)

* $S = Sr/1.75$ is used for the moment calculation through the relation $M_0 = 1.23 \times 10^{22} S^{3/2}$ dyne-cm (e.g., Kanamori, 1977).

† The values in the parentheses are obtained indirectly.

‡ Geller and Kanamori (1977).

§ Abe (1979).

¶ Kelleher (1972).

‖ Rothé (1969).

\# Relative to the 1979 event.

** The aftershock area for the period 12 to 31 December 1979.

†† National Earthquake Information Service (NEIS).

‡‡ Kanamori and Given (1981).

§§ Determined from tsunami height at Hilo and Japanese stations.

For the data in 1906, it is probably best to use the *S-P* times. Figure 5 shows loci of "constant *S-P*" distance for five stations: Münich; Baltimore; Tacubaya; Victoria; and Göttingen. The loci for Münich, Baltimore, and Victoria intersect each other near (within 200 km) the Gutenberg-Richter epicenter. The loci from Göttingen and Tacubaya overshoot it by several hundred kilometers. Since the data are incomplete, the result is inconclusive. Nevertheless, the three closely located intersections (between Münich and Baltimore, Baltimore and Victoria, and Münich and Victoria) are very close to the Gutenberg-Richter epicenter which is near the southwestern end of the rupture zone. It is possible that Gutenberg and Richter (1959) determined their epicenter by more or less the same reasoning.

Kelleher (1972) estimated the rupture zone of the 1906 event on the basis of macroseismic data that include reports of diminution of water level in the harbors of Manta (59′S) and Buenaventura (3°54′N), and a broken submarine cable found near Buenaventura (see Figure 6).

Rudolph and Szirtes (1911) made a detailed account of the macroseismic effects

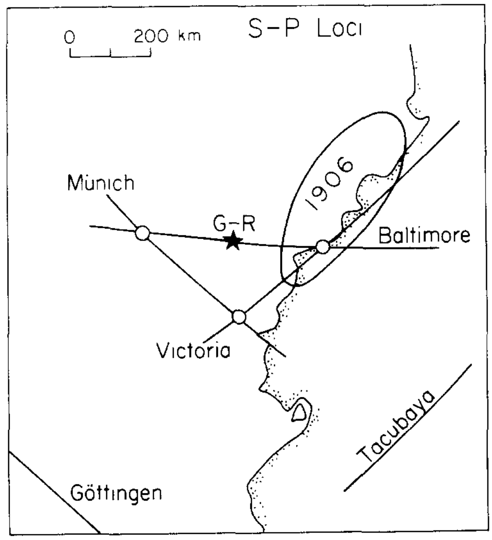

FIG. 5. *S-P* loci (loci of a point corresponding to a constant *S-P* time) for the 1906 Colombia-Ecuador earthquake. The asterisk indicates the epicenter determined by Gutenberg and Richter (1959). The rupture zone of the 1906 event estimated by Kelleher (1972) is shown.

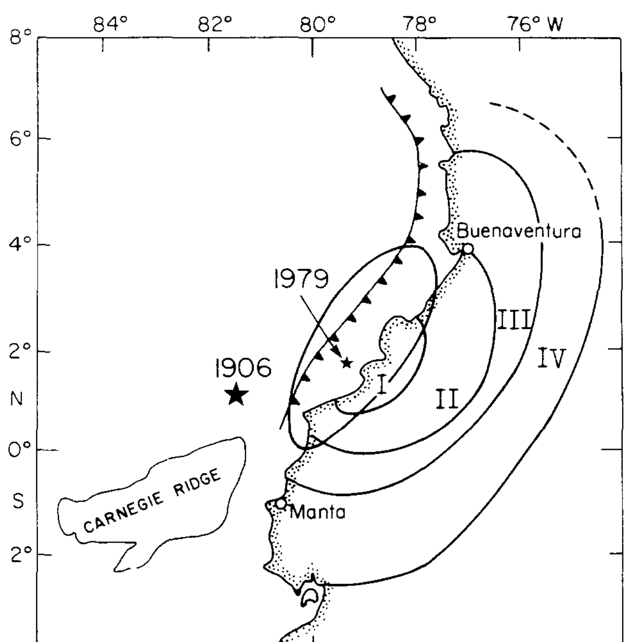

FIG. 6. Intensity distribution described by Rudolph and Szirtes (1911). For the explanation of zones, I to IV, see the text.

of this earthquake. They estimated the intensities inland to be from *V* to *X* on the Rossi-Forel scale, and divided the affected area into four zones (I, II, III, and IV) in a decreasing order of the strength of shaking as shown in Figure 6. In zone I, many buildings were completely destroyed, and many lives were lost. In zone II, destructions were limited to masonries including churches and public buildings. Zone IV suffered only very minor or no damage. Very long-period ground shakings were felt in zone IV.

Although Kelleher (1972) considers that the evidence from which the end points of the rupture zone are determined is marginal, the intensity distribution shown in Figure 6 strongly suggests that the rupture zone did not extend much further beyond Kelleher's end points.

Furthermore, the aseismic Carnegie Ridge intersects the Colombia trench at about 0° latitude (Figure 6), and the rupture propagation probably did not extend southwest past this intersection. Concerning the northeast end, the Colombia trench bends sharply from the NE–SW trend to the N–S trend at about 4°N, where the chain of active volcanoes along the coast is interrupted. Probably the 1906 event did not rupture past this sharp bend.

The seismic moment of this event is estimated to be 2×10^{29} dyne-cm (Kanamori, 1977) from Kelleher's (1972) estimate of the rupture zone, and is inevitably subject to some uncertainty. However, the value of the corresponding M_w is in close agreement with Abe's (1979) tsunami magnitude M_t, suggesting that it is reasonably accurate.

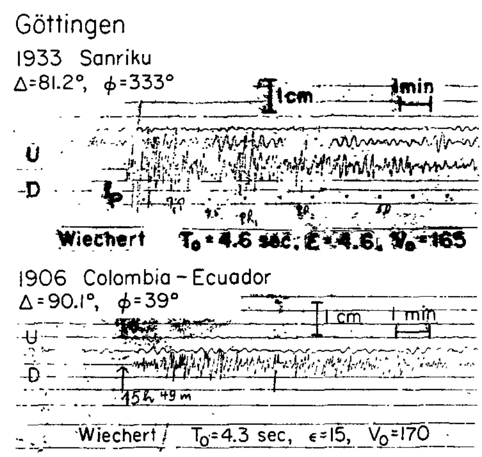

FIG. 7. Comparison of the P waveform of the 1906 Colombia-Ecuador earthquake with that of the 1933 Sanriku earthquake recorded by a vertical-component Wiechert seismograph at Göttingen, Germany. T_0 is the natural period of the pendulum, ϵ is the damping constant, and V_0 is the static magnification. Note the sharp onset and the large amplitude of the 1933 earthquake compared with the 1906 event.

Concerning the nature of the 1906 event, a possibility remains that it was a large normal-fault event near the trench axis such as the 1933 Sanriku (Kanamori, 1971) and the 1977 Indonesian earthquake (Stewart, 1978; Given and Kanamori, 1980). Kelleher (1972) dismissed this possibility on the basis of severe destruction which occurred well over 100 km inland (see Figure 6). In the case of the 1933 Sanriku earthquake, the intensity inland was mostly VI on the JMA (Japan Meteorological Agency) scale (equivalent to VI to VII on the Modified Mercalli scale) and no severe destruction was caused. Comparison of this observation for the Sanriku earthquake with the intensity distribution for the 1906 event shown in Figure 6 does suggest that the 1906 event is not a trench normal-fault event.

Since this problem is crucial for estimating the seismic recurrence rate in this region, we examined some old seismograms to resolve this problem. Unfortunately, we could collect only a few seismograms for this event. However, a Wiechert seismogram recorded at Göttingen, Germany, provides key information. One of the characteristic features of the large normal-fault events is a very sharp onset of body waves (see Kanamori, 1971). Figure 7 compares the waveform of the 1933 Sanriku and 1906 Colombia-Ecuador earthquake recorded by a Wiechert seismograph (vertical component) at Göttingen. The characteristics of the seismograph are almost identical for the two events. The record of the Sanriku earthquake shows a very sharp onset followed by a large *P*-wave train. On the other hand, the 1906 event shows a very gradual onset typical of subduction-zone thrust events. Also, the first motion is up for the 1906 event while it is down for the 1933 event. In view of the geometry of the trench and the location of the station, it is very unlikely that the Göttingen station is located near the node of the radiation pattern, and the upward motion indicates a thrust event.

The N-S-component Weichert seismograph at Göttingen registered a clear G_3 wave as shown by Figure 8. Although the amplitude is very small (peak-to-peak = 1 mm), it is clearly above the noise level and the group velocity is appropriate for the long-period Love wave (i.e., 4.3 to 4.4 km/sec). However, no clear long-period arrival is found at the time corresponding to the G_2 wave, as shown by Figure 8. If a point source is assumed, the amplitude of G_2 should be about 3 times larger than G_3. This observation suggests a northeastward rupture propagation which is consistent with the location of the epicenter at the southwestern end of the rupture zone.

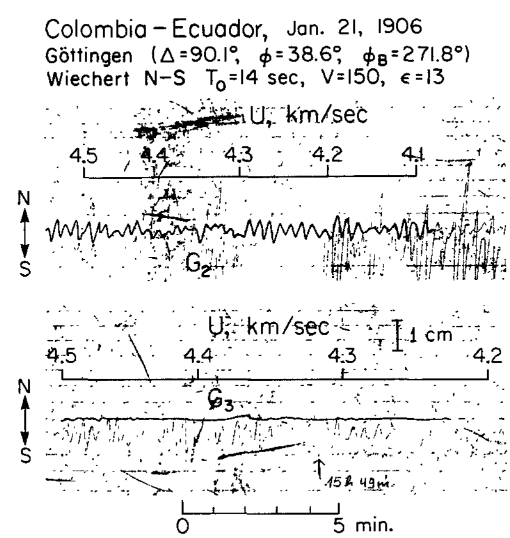

FIG. 8. G_2 and G_3 waves from the 1906 Colombia-Ecuador earthquake recorded by a N-S-component Wiechert seismograph at Göttingen, Germany.

We computed a synthetic seismogram assuming the same mechanism as the 1979 event and a rupture length of 500 km with a rupture velocity of 2 km/sec in the NE direction. Comparison of the amplitude of the synthetic seismogram with the observed gives a seismic moment of 8×10^{28} dyne-cm, which is about $\frac{1}{3}$ of the estimate given in Table 2. A most likely cause for this discrepancy is the solid friction between the stylus and the recording paper of the seismograph. Since this type of seismograph was designed for recording seismic waves with a period of up to 100 sec or so, the effect of friction becomes very serious for very long-period waves with an extremely small amplitude. We, therefore, consider this value a lower bound.

Discussion and Conclusions

Various source parameters for the four Colombia-Ecuador earthquakes are summarized in Table 2.

The average values of slip are estimated from the seismic moment and the rupture area, and are less accurate than the seismic moment. Nevertheless, the factor of about 2 difference between the 1906 and other events is probably larger than the uncertainty involved in this calculation.

The sum of the seismic moments of the 1942, 1958, and 1979 events is 3.7×10^{28} dyne-cm and is only $\frac{1}{5}$ of the seismic moment of the 1906 event estimated from the rupture area and the tsunami data. This difference is clearly larger than the uncertainty in the moment calculations. Thus, we conclude that the sum of the 1942, 1958, and 1979 events is not equivalent to the 1906 event, despite the fact that the sequence of the 1942, 1958, and 1979 events ruptured approximately the same segment of the subduction zone as the 1906 event. If the entire segment of the Colombia trench from the intersection with Carnegie Ridge to the sharp bend broke in 1906, the rupture length could have been as large as 600 km. Even if this were the case, the conclusion that the seismic moment release per unit rupture length is larger for the 1906 event than for the 1942, 1958, and 1979 events would remain unchanged. If this is the general characteristic of the rupture behavior, the amount and extent of the coseismic displacement in a great earthquake would be significantly larger than those for a series of smaller earthquakes which occur along the same segment abutting to each other. Sykes and Quittmeyer (1981) suggest this kind of behavior on the basis of a rupture model in which the stress drop increases with the rupture length.

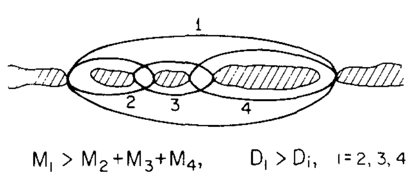

FIG. 9. An asperity model which explains the larger moment per unit rupture length and slip of a great earthquake (event 1) than the smaller events (events 2 to 4). M_i and D_i are the seismic moment and the amount of slip of the ith event.

Since the accurate size and the geometry of the rupture zones could not be determined very well, the question of whether the stress drop increases with the rupture dimension or not cannot be resolved in this study. Here, we attempt to explain the present results by using a simple asperity model such as the one described by Lay and Kanamori (1981) (see also Lay *et al.*, 1982). In this model, the fault zone is held by a discrete distribution of asperities and the zone between them is considered weak. The weak zone normally behaves aseismically, but slips abruptly only when it is driven by failure of the asperities (Figure 9). A recent study by Ruff and Kanamori (1980) demonstrates that this type of asperity distribution is inferred from complexities of body waveforms of large earthquakes. In the framework of this simplified model, a small earthquake represents failure of one asperity, and the rupture zone is pinned at both ends by adjacent asperities. Therefore, the effective width and the amount of slip are relatively small. On the other hand, a great earthquake represents failure of more than one asperity, and consequently involves much larger width and slip. Although whether this model is mechanically feasible or not must await further studies, it appears to explain qualitatively many of the observed features of great earthquakes (e.g., complexity of the waveform).

In the case of the Colombia-Ecuador sequence studied here, the entire zone may be modeled by three asperities, a large one in the northeastern end, and two smaller ones in the southwestern end of the rupture zone. As Kelleher (1972) suggests, the 1942 and the 1958 events seem to have ruptured from the southwest to the northeast end of the individual rupture zone. The 1979 event ruptured from the northeastern end of the 1958 rupture zone toward northeast. On the basis of the location of the main shock, Kelleher (1972) suggests that the 1906 event ruptured also in the northeast direction. The asymmetric radiation pattern of G_2 and G_3 demonstrated by Figure 8 supports Kelleher's suggestion.

In terms of the asperity model, triggering occurred instantaneously in 1906, resulting in a single great earthquake, while in the sequence from 1942 to 1979, there were pauses of about 20 yr between the successive events. Then the question is why the trigger pattern changed from sequence to sequence. Using the model by Lay and Kanamori (1981), two extreme cases can be considered.

1. Each asperity has its own characteristic repeat time. The asperities normally behave more or less independently, but occasionally they synchronize. If the characteristic repeat times are 36, 52, and 73 yr for the 1942, 1958, and 1979 zones, respectively, they may synchronize in about 210 yr. During the 210 yr, the 1942, 1958, and 1979 zones would break about 6, 4, and 3 times, respectively. In this case, the 1906 earthquake represents a relatively rare event.
2. The degree of mechanical coupling between the asperities is very strong so that when one asperity breaks, it tends to trigger the adjacent one even if the latter is not quite ready to fail by itself. Only when the stress in the adjacent asperity is very much lower than its strength, triggering fails to occur. In this case, the 1906-type event is the norm for the subduction zone.

Historic record in Colombia is not complete enough to test these two hypothetical cases against data. In the catalog for the period 1575 to 1915 compiled by Ramirez (1933), no event similar to the 1906 event is reported before 1900. A very large earthquake is reported in 1882, but this event seems to be located near the northern end of the 1906 event. As far as this catalog indicates, it appears that the 1906-type event is relatively infrequent.

Whether events similar to the 1942, 1958, or 1979 events occurred during the time period covered by Ramirez's catalog is more difficult to determine because the historical data which are mainly based on intensity data on land are considered less complete for the events off shore.

If the actual situation is close to case (1), long-term prediction of seismic activity can be made relatively accurately, but if triggering controls the sequence, as in case (2), prediction of recurrence time at a given point of the subduction zone would be more difficult.

Detailed studies on the waveforms of the events in these rupture zones may be able to determine the mechanical conditions there, which in turn may provide clues to the behavior of this subduction zone in the future.

The result that the amount of displacement at a given point along a subduction zone depends on the rupture length has an important bearing on risk analysis. Recently, some attempts have been made to predict the nature of strong ground motions and tsunamis which would be excited by failure of certain seismic gaps. In these studies, various parameters such as the intensity, the size of the affected area, the observed tsunami height, and the magnitude of crustal deformation of the previous events are used to constrain the models for the earthquake in the future. However, if the behavior of the seismic gap varies from sequence to sequence as is demonstrated for the Colombia-Ecuador earthquake sequence, these parameters should be scaled appropriately for the dimension of the gap which is expected to break.

In the case of the Tokai, Japan, gap (Ishibashi, 1981), the length of the predicted event is about 100 km, while the previous two events in 1854 and 1707 seem to have a much larger (about 500 km) rupture length. Although the situation in the Tokai area may be different from the Colombia-Ecuador case, some caution should be exercised in using the data on the old events for risk estimate of future events.

Acknowledgments

We thank John Kelleher for providing us with valuable comments and information on the 1906 events. Jim Brune suggested that we should examine the possibility that the 1906 event is a normal-fault event similar to the 1933 Sanriku earthquake. We thank the directors of seismographic stations at Göttingen, Germany, and Uppsala, Sweden, for kindly sending us the records of the 1906 Colombia earthquake. Katsuyuki Abe, Tokutaro Hatori, Harold Loomis, and Jack Fancher provided us with valuable information on the tsunami height of the 1906 and 1979 events. George Purcaru and Ines Cifuentes sent us useful information on the 1979 and 1906 events, respectively. Fumiko Tajima translated for us some key sentences in Rudolph and Szirtes (1911). This research was supported by the U.S. Geological Survey Contracts 14-08-0001-19265 and 14-08-0001-19755 and also by the National Science Foundation Grant EAR78-05352.

References

Abe, K. (1979). Size of great earthquakes of 1837–1979 inferred from tsunami data, *J. Geophys. Res.* **84,** 1561–1568.

Ando, M. (1975). Source mechanisms and tectonic significance of historical earthquakes along the Nankai Trough, Japan, *Tectonophysics* **27,** 119–140.

Ben-Menahem, A. (1961). Radiation of seismic surface waves from finite moving sources, *Bull. Seism. Soc. Am.* **51,** 401–435.

Geller, R. J. and H. Kanamori (1977). Magnitude of great shallow earthquakes from 1904 to 1952, *Bull. Seism. Soc. Am.* **67,** 587–598.

Gilbert, F. (1980). An introduction to low-frequency seismology, in *Physics of the Earth's Interior*, A. M. Dziewonski and E. Boschi, Editors, North-Holland Publishing Co., Amsterdam, 41–81.

Given, J. W. and H. Kanamori (1980). The depth extent of the 1977 Sumbawa, Indonesia earthquake (abstract), *EOS* **61,** 1044.

Goodstein, J. R., H. Kanamori, and W. H. K. Lee (1980). Seismology microfiche publications from the Caltech Archives (announcements), *Bull. Seism. Soc. Am.* **70,** 657–658.

Gutenberg, B. and C. F. Richter (1959). *Seismicity of the Earth*, Princeton University Press, Princeton, New Jersey, 310 pp.

Herd, D. G., T. L. Youd, H. Meyer, J. L. Arango C., W. J. Person, and C. Mendoza (1981). The great Tumaco, Colombia earthquake of 12 December 1979, *Science* **211,** 441–445.

Imamura, A. (1928). On the seismic activity of central Japan, *Jap. J. Astron. Geophys.* **6,** 119–137.

Ishibashi, K. (1981). Specification of a soon-to-occur seismic faulting in the Tokai district, Central Japan, based on seismolectonics, in *Earthquake Prediction*, D. W. Simpson and P. G. Richards, Editors, American Geophysical Union, Washington, D.C., 297–332.

Kanamori, H. (1971). Seismological evidence for a lithospheric normal faulting—The Sanriku earthquake of 1933, *Phys. Earth Planet. Interiors* **4,** 289–300.

Kanamori, H. (1977). The energy release in great earthquakes, *J. Geophys. Res.* **82,** 2981–2987.

Kanamori, H. and J. W. Given (1981). Use of long-period surface waves for fast determination of earthquake source parameters, *Phys. Earth Planet. Interiors* **27,** 8–31.

Kelleher, J. A. (1972). Rupture zones of large South America earthquakes and some predictions, *J. Geophys. Res.* **77,** 2087–2103.

Lay, T. and H. Kanamori (1981). An asperity model of large earthquake sequences, in *Earthquake Prediction*, D. W. Simpson and P. G. Richards, Editors, American Geophysical Union, Washington, D.C., 579–592.

Lay, T., H. Kanamori, and L. Ruff (1982). The asperity model and the nature of large subduction zone earthquakes, *Earthquake Prediction Research* (in press).

Minster, J. B. and T. H. Jordan (1978). Present-day plate motions, *J. Geophys. Res.* **83,** 5331–5354.

Ramirez, J. E. (1933). Earthquake history of Colombia, *Bull. Seism. Soc. Am.* **23,** 13–22.

Rothé, J. P. (1966). The seismicity of the earth 1953–1965, UNESCO, 336 pp.

Rudolph, E. and S. Szirtes (1911). Das Kolumbianskche Erdbeben am 31 January 1906, *G. Beitr.* **11,** 132–199, 207–275.

Ruff, L. and H. Kanamori (1980). Variation of asperity size inferred from the body waves of great earthquakes, *EOS* **61,** 1026–1027.

Seno, T. (1977). The instantaneous rotation vector of the Philippine Sea plate relative to the Eurasian plate, *Tectonophysics* **42,** 209–226.

Stewart, G. S. (1978). Implications for plate tectonics of the August 19, 1977, Indonesian decoupling normal-fault earthquake (abstract), *EOS* **59,** 326.

Sykes, L. R., J. B. Kisslinger, L. House, J. N. Davies, and K. H. Jacob (1980). Rupture zones of great earthquakes, Alaska-Aleutian arc, 1784–1980, *Science* **210,** 1343–1345.

Sykes, L. R. and R. C. Quittmeyer (1981). Repeat times of great earthquakes along simple plate boundaries, in *Earthquake Prediction*, D. W. Simpson and P. G. Richards, Editors, American Geophysical Union, Washington, D.C., 217–247.

Seismological Laboratory
Division of Geological and Planetary Sciences
California Institute of Technology
Pasadena, California 91125
Contribution No. 3603

Manuscript received 6 November 1981