INSTRUMENTALLY RECORDED SEISMIC ACTIVITY PRIOR TO THE MAIN EVENT

By David H. Harlow

INTRODUCTION

Instrumental recording of earthquakes in Guatemala began in 1925 with the installation of a three-component Wiechert seismograph at the Observatorio Nacional in Guatemala City (Vassaux, 1969). No other permanent seismographs were operated in Guatemala until early 1973, when three radio-telemetered high-gain seismic stations were installed near Guatemala City and Pacaya and Fuego Volcanoes as part of a cooperative project between the U.S. Geological Survey and government agencies of Guatemala, El Salvador, and Nicaragua to monitor seismic activity and ground tilt at active volcanoes in Central America. This project is described in detail by Ward and others (1974). The cooperating agency in Guatemala was originally the Instituto Geográfico Nacional and since early 1975 has been the Observatorio Nacional.

In March 1973, seismographs were temporarily installed at Chiantla and Chiquimula in cooperation with Dartmouth College, the Instituto Geográfico Nacional, and the U.S. Geological Survey. The purpose of these instruments was to monitor earthquake activity on an east-west system of faults that cuts across central Guatemala and consists mainly of the Polochic, Motagua, and Joctán faults (Quittmeyer, written commun., 1974).

Three additional radio-telemetered stations were added in February 1975, bringing the total to six. The purpose of this seismic network is to monitor earthquake activity associated with faults and active volcanoes.

The locations of the permanent and temporary seismic stations are plotted in figure 7, and station data are listed in table 1. Seismometers with a natural frequency of 1 Hz are employed at each station. The seismic signals from the permanent stations are relayed to the Observatorio Nacional and recorded on drum recorders at a paper speed of 60 mm/min. A description of the instrumentation and

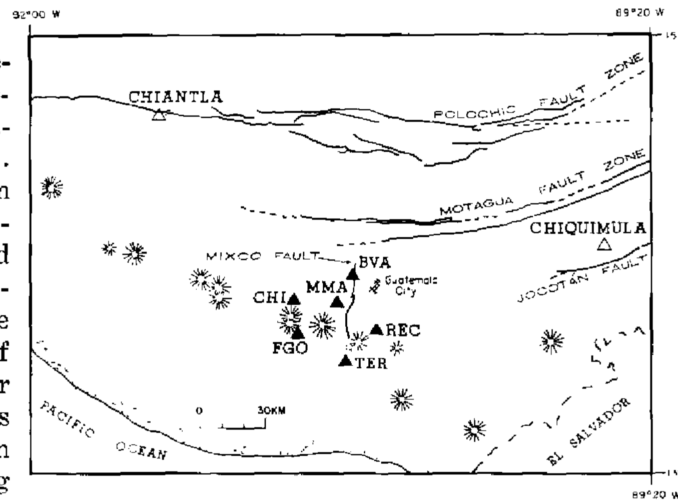

Figure 7.—Seismic station locations. Temporary seismic stations operated during 1973 are shown by open triangles, and the existing telemetering stations are shown by solid triangles. Volcanic cones are also shown.

the response curve can be found in Ward and others (1974). For the Chiantla and Chiquimula stations, recording was done on site.

SEISMICALLY ACTIVE AREAS IN GUATEMALA

Molnar and Sykes (1969) studied the regional tectonics of the Caribbean and Central America by using earthquake hypocenters and focal mechanisms for earthquakes larger than magnitude 4.0 recorded by the WWNSS between 1954 and 1967. The most prominent seismic feature is a zone of earthquakes that dips northeastward beneath Central America and appears to be the result of the convergent plate motion and underthrusting of the Cocos plate beneath the Caribbean plate. Hypocenters for earthquakes within this zone range from near-surface depths at the Middle America Trench to depths of approximately 100 km beneath the line of active volcanoes and then to maximum depths of about 250 km (fig. 3*B*). During this time inter-

TABLE 1.—*Coordinates and magnifications of seismograph stations*

Stations [1]	Latitude (°N.)	Longitude (°W)	Installation date	Magnification at 25 Hz
Present locations:				
FGO	14° 26.74′	90° 50.43′	2/73	120,000
CHI	14° 35.69′	90° 51.62′	2/75	60,000
MMA	14° 32.28′	90° 40.89′	2/75	60,000
BVA	14° 40.00′	90° 38.24′	2/73	120,000
REC	14° 26.25′	90° 31 36′	2/73	120,000
TER	14° 18.25′	90° 41.01′	2/75	240,000
1973 locations:				
Chiquimula	14° 47.4′	89° 33.6′	3/73	120,000
Chiantla	15° 21.6′	91° 27.0′	3/73	60,000

[1] Station codes are listed in the Glossary.

val, however, only a few shallow-focus earthquakes locate along the fault system through Central Guatemala, which marks the boundary between the North American and Caribbean plates (Malfait and Dinkelman, 1972; Jordon, 1975). Included in this fault system is the Motagua fault (fig. 7) that was the source of the February 4 earthquake.

Figure 8 is a seismicity map of Guatemala based on 30 years of data recorded by the Wiechert seismograph at the Observatorio Nacional. The gain of this seismograph is 35, and thus it is sensitive to earthquakes larger than roughly magnitude 2.0 in the vicinity of Guatemala City and larger than magnitude 5.5 at a distance of 200 km. Earthquake epicenters are located only approximately by using the S−P-wave difference to calculate distance and P-wave amplitudes on the two horizontal components to determine azimuth. In addition, felt earthquake reports sent to the Observatorio Nacional aid in verifying the locations. Therefore, this group of earthquakes is approximately equivalent to what would have been located by the WWNSS. Even though the epicenters are subject to large errors, these long-term data provide important data on the features of seismically active zones in Guatemala and their relative level of seismic activity over three decades.

The greatest concentration of seismic activity shown in figure 8 is along the Pacific coast of Guatemala and is probably associated with the Benioff zone that dips northeastward beneath the country. Recent results, from a seismic network in Nicaragua (Aburto, 1975), show that, in addition to the very active dipping Benioff zone, a separate but seismically less active zone of shallow earthquakes with depths less than 20 km occurs along the Nicaraguan chain of active volcanoes. Data from the six-station seismic net suggest that a similar group of shallow earthquakes occurs in the vicinity of the volcanoes near Guatemala City (fig. 10). Historically this shallow seismic zone is the source of moderate-sized locally destructive earthquakes in Central America (Carr and Stoiber, 1976, unpub. data) including, most recently, the Managua earthquake of December 23, 1972, that damaged much of the capital city of Nicaragua. Therefore, the earthquakes in Guatemala that occur near the comparatively less seismically active northern boundary of the large hatched area shown in figure 8 may also be associated with shallow seismicity along the line of volcanoes.

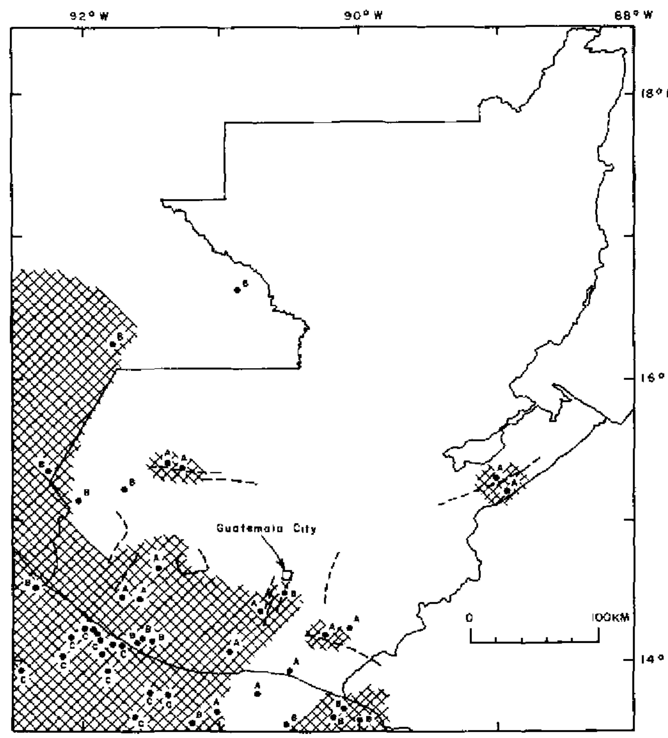

FIGURE 8.—Seismicity map for Guatemala (1945–75) compiled by J. Vassaux, Observatorio Nacional, Guatemala. Hatched areas are regions of frequent seismic activity. Dots indicate the number of earthquake series or swarms at specific areas during this period: A, fewer than 5 times; B, between 6 and 10 times; and C, more than 10 times.

Two other small areas of relatively low-level seismic activity are located in figure 8 in west-central and east-central Guatemala. These areas lie along the fault system that runs across central Guatemala, and the eastern area occurs on the Motagua fault. Previous data (Molnar and Sykes, 1969), the extensive ground breakage on the Motagua, and its tectonic similarity to the San Andreas fault indicate that the earthquakes in these areas occur at shallow depths.

Thus, there are three sources of magnitude 4 or larger earthquakes in Guatemala, and their relative levels of seismic activity appear to have been constant for the last 30 years. The main source of seismicity is the zone of earthquakes that dips northeastward beneath the country. The other two sources are shallow and seismically less active. One lies along the chain of active volcanoes, and the other along the fault system that crosses central Guatemala. Historically, however, these shallow, relatively less active zones are the sources of the majority of moderate-sized locally destructive earthquakes. Although the deep shocks are more numerous, their depths, which are in the range of 50–250 km under the more heavily populated regions, lessen their threat by placing them at considerable distances from cities and towns.

DISTRIBUTION OF S–P TIMES

Distributions of S – P times from the station at Fuego Volcano and the two stations temporarily installed along the fault zone in central Guatemala are plotted in figure 9. Approximately 10 percent of the recorded earthquakes could not be used, either because the arrival times were unclear or because the earthquake waves were large enough to exceed the dynamic range of the instruments, thereby making the S – P time unreadable for S – P intervals of less than 15 or 20 s (a unit S – P interval is equivalent to a focal distance of about 8 km). For earthquakes within 5 km, the lowest detectable magnitudes are roughly estimated (Brune and Allen, 1967) to be 0.5 for Fuego and Chiquimula and 0.8 for Chiantla.

Events with S – P times of less than 5 s were recorded at all stations, indicating that, at each site, local faults are active. For microearthquakes within 40 km of the stations, an average of 5.5 events per day were recorded at Fuego (Harlow and Ward, unpub. data, 1976), and about 1.0 event per day was recorded at Chiquimula and Chiantla. The higher seismicity at Fuego is probably related to its high level of recent volcanic activity. Local

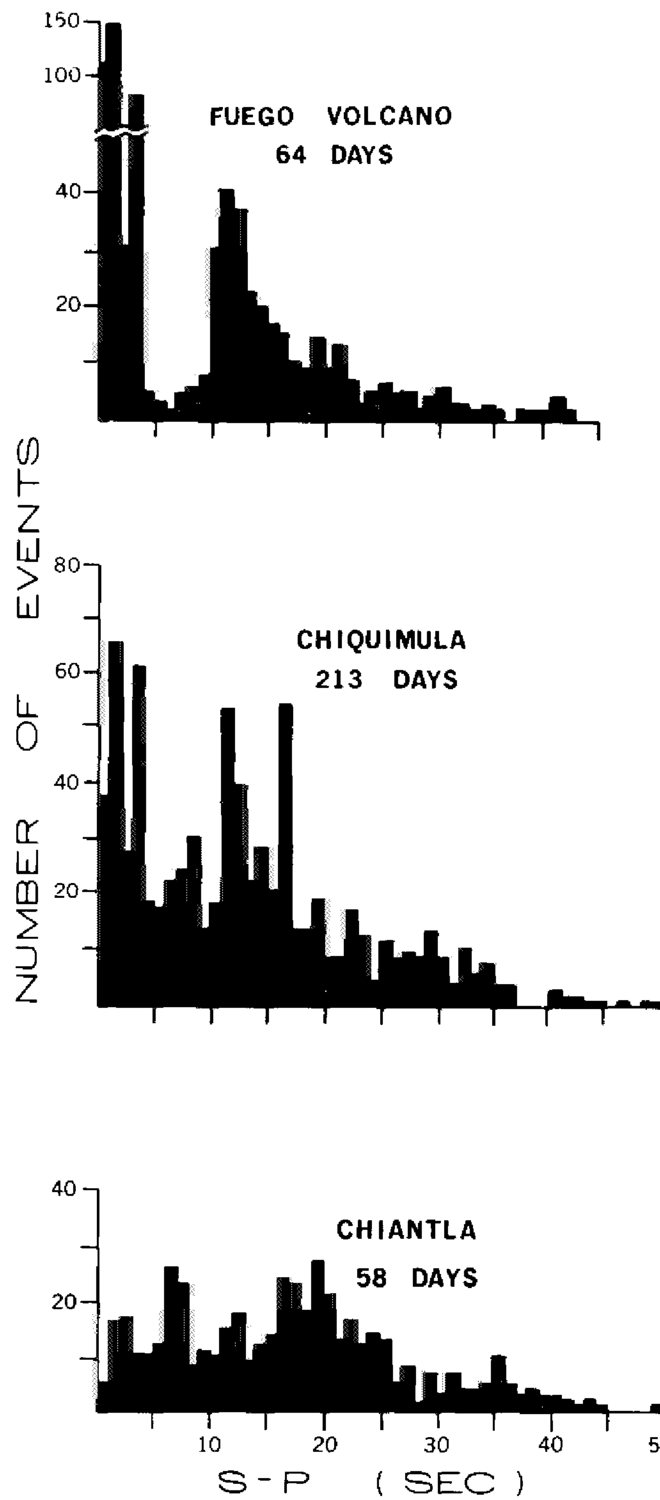

FIGURE 9.—Distribution of S–P times in seconds during 1973 at seismic stations in Guatemala. Useful recording days are noted for each station.

seismicity near Fuego originates both on local faults and at the volcano. The Chiquimula station is 8 km north of the Jocotán fault and 25 km south of the Motagua fault. Thus, events with 1- to 2-s S−P times at Chiquimula likely occur on the Jocotán fault or on north-south-trending fractures in the Ipala graben just to the south of the Jocotán fault in this area (Carr, 1974). The numerous events with S−P times of 3 to 5 s could occur either on the above faults or on the Motagua fault. The data suggest that the Polochic fault is also seismically active in the vicinity of Chiantla. The events with S−P times of 7 and 8 s and greater at Chiantla could originate on the western end of the Motagua fault. Thus, the occurrence of microearthquakes within 40 km of these seismograph stations indicates that the Polochic, Motagua, and Jocotán faults are probably seismically active.

Intermediate-depth earthquakes with magnitudes greater than 1.5 to 2.0 originating in the Benioff zone beneath the stations would be expected to produce peaks in the histograms at S−P times of 10 to 14 s. Peaks in this range occur at the Fuego and Chiquimula stations. There are no clear peaks in the distribution of S−P times at Chiantla, however, possibly because of the relatively short recording time or the diffuse seismic activity in southeastern Mexico (Molnar and Sykes, 1969; see also fig. 8).

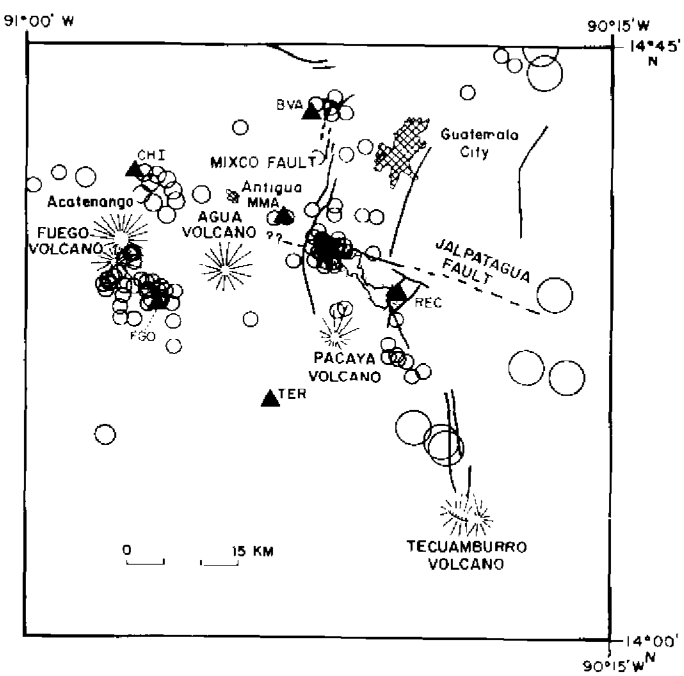

FIGURE 10.—Epicenter locations and known faults in the region of the six-station seismic network (solid triangles). The size of the circles reflects the estimated error in the calculated location. Station code is in the Glossary.

SEISMIC ACTIVITY IN THE VICINITY OF GUATEMALA CITY

Epicenters of shallow (less than 15 km deep) microearthquakes near Guatemala City are plotted in figure 10. These data include 29 events recorded by the three-station network during 65 useful recording days in 1973 and 75 events recorded by the six-station network from March to September 1975. For earthquakes that occur inside or near the seismic network, magnitude 1 events are the smallest that can be located. The most intense seismicity occurs within 10 km of Fuego Volcano, which has had large eruptions in 1971, 1973, and 1974, plus minor eruptive activity in 1975. These events, then, are probably related to eruptive processes at Fuego. Epicenters show that the Jalpatagua and Mixco faults are seismically active (ground breakage on the Mixco fault zone was caused by the February 4 earthquake). Other epicenters lie on or near other known faults in this area. Thus, the shallow seismicity indicates that there is a complex pattern of active faults near Guatemala City.

Events that may originate on the Motagua fault have not been systematically located because they are outside the network and their epicenters are therefore poorly determined. A check of all regional earthquakes recorded by the six-station network from March 1975 through January 1976 was made to determine how many events might have originated on the fault zone across Guatemala represented by the Polochic, Motagua, and Jocotan faults. Criteria used to identify regional events occurring at shallow focal depths north of the seismic net are: (1) relative arrival time at the seismic stations to roughly determine azimuth and (2) apparent velocities across the network of less than 10 km/s. The second criterion passes shallow earthquakes at distances up to approximately 175 km from the seismic net and discriminates against events that occur north of the seismic net but at depth on the Benioff zone. The lowest detectable magnitude for these earthquakes is about 1.5 for events on the Motagua fault nearest to Guatemala City and 4.0 for events at a distance of 150 km. The results, listed by month in table 2, show that, at most, only 11 percent of the regional earthquakes could have originated on the Motagua fault during this period. In addition, no unusual activity such as swarms was observed. This, together with previous data from other sources, suggests that, although the Motagua fault is an historic source of large and damaging earthquakes, it is not continuously the most seismically active tectonic feature in Guatemala.

TABLE 2.—*Number of events each month originating from the central fault system compared with the number of all regional events*

Date		Possible central fault system events	Total regional events
1975	March	9	76
	April	14	77
	May	28	90
	June	---	---
	July	[1] 5	[1] 43
	August	9	49
	September	8	104
	October	14	119
	November	6	127
	December	6	123
1976	January	8	124
	Total	107	932

[1] The numbers of events for July are extrapolated from 22 useful recording days

SUMMARY

Seismic data collected over an interval of about 30 years before the earthquake of February 4 reveal three main sources of seismic activity in Guatemala. The most intense source of seismic activity is a zone of earthquakes that dips northeastward beneath Guatemala and results from the Cocos plate being thrust under the Caribbean plate. The danger of this seismic zone is greatly reduced by the depths, and therefore the distance, of these events, which range from 50 to 250 km under the more heavily populated regions. During the interval of the last 30 years, the second and third source regions have consistently generated fewer shocks. The second source of earthquakes occurs at shallow depths along the chain of active volcanoes. This zone is historically the source of many moderate-sized, locally damaging earthquakes and, at least in the vicinity of Guatemala City, consists of many complexly related active faults. The third source is the fault system that cuts across central Guatemala and includes the Motagua fault, which was the source of the February 4 earthquake. During 1973, about 1.0 local (within 40 km) microearthquake per day was recorded at two high-gain seismograph stations installed on this fault system, suggesting that the Polochic, Motagua, and Jocotán faults are seismically active. During the 11 months preceding the February 4 earthquake, however, only 11 percent of all regional earthquakes recorded at a seismic network near Guatemala City could have originated from this fault system. Thus, the seismic zone that produced the most destructive earthquake in the recent history of Guatemala has exhibited a level of seismicity over the last 30 years that is lower than the prominent seismic activity that occurs on the deep seismic zone.

THE GUATEMALAN EARTHQUAKE OF FEBRUARY 4, 1976,
A PRELIMINARY REPORT

MAIN EVENT AND PRINCIPAL AFTERSHOCKS FROM TELESEISMIC DATA

By WAVERLY PERSON, WILLIAM SPENCE, and JAMES W. DEWEY

The hypocenter of the main event of the Guatemala earthquake was determined by using stations available to the National Earthquake Information Service (NEIS) throughout the world. The preliminary hypocenter and origin time parameters are:

Origin time:	09 01 43.3 UTC (03 02 43.3 local time)
Latitude:	15.32° N.
Longitude:	89.08° W.
Depth:	5 km (constrained)
Magnitude:	$M_S = 7.5$

The main event is located near Los Amates about 157 km northeast of Guatemala City on the Motagua fault. It should be emphasized that the hypocenter of the main event represents the point of the initial rupture. The fault break extends at least 160 km westward towards Guatemala City and 80 km towards the northeast (Plafker and others, this report).

Data from 90 stations were used in locating the main event, including readings from Guatemala. The wide distribution of these stations gives us reasonable confidence in the epicenter solution. Travel-time anomalies, however, could conceivably be producing a location bias of tens of kilometres; the preliminary location of the Managua, Nicaragua, earthquake of December 23, 1972, for example, was biased 25 km to the northeast of the true epicenter. In the case of the Guatemala earthquake, a component of location bias in the direction of the Motagua fault would be difficult to detect.

There were neither reliable teleseismic depth phases nor stations close enough to the epicenter of the Guatemala main event to enable us to estimate hypocentral depth with confidence. The hypocenter was restrained to a shallow depth, 5 km, because of the surface faulting that accompanied the earthquake and because the depths of aftershocks located by Langer, Whitcomb, and Aburto (this report) were in the range 0–12 km. Computation with no depth restraint of the hypocentral parameters of the main event always yielded focal depths in the shallow crust, but these results could be fortuitous.

The magnitude is based on the average of surface-wave data from several stations. $M_S = 7.5$ is consistent with amplitudes of 100-s G-waves measured by Dewey and Julian (this report).

Short-period P-wave arrivals of the February 4 main event and subsequent aftershocks are generally emergent. The teleseismic records strongly suggest that the main event was a multiple rupture.

The February 4 main event was followed by damaging aftershocks. The two largest aftershocks (both m_b 5.8), as of March 7, 1976, occurred soon after the main event. These aftershocks occurred near Guatemala City, possibly on the north-south-trending Mixco fault (see fig. 7). Other

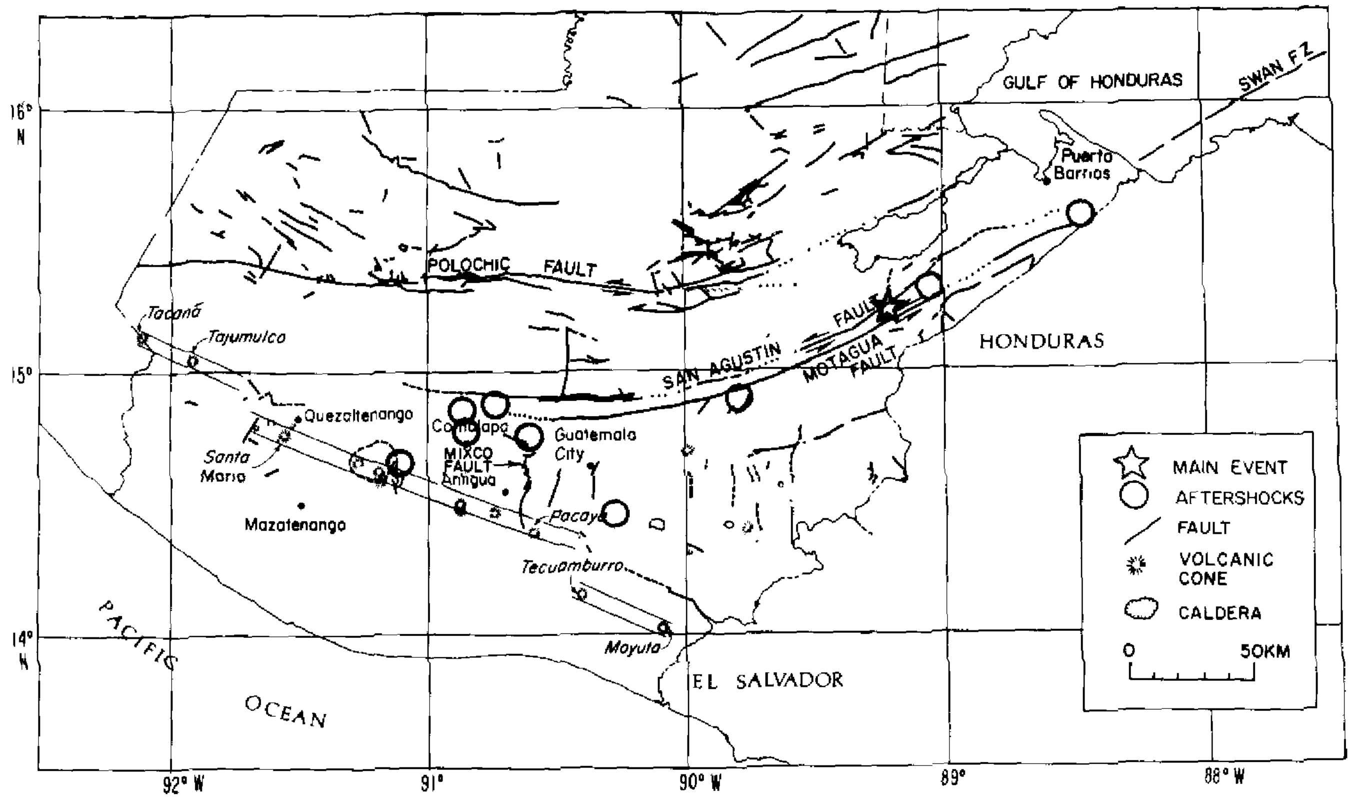

FIGURE 11.—Epicenters of main event and the principal aftershocks through March 7, 1976. These locations are based on data available as of April 5, 1976, and are subject to slight revision. Light parallel lines represent volcanic linears. (Base map modified from Bonis and others, 1970, and Plafker and others, this report.)

TABLE 3.—*Epicenter location of aftershocks*

Date	Origin time	Latitude (°N.)	Longitude (°W.)	Depth [1] (km)	Magnitude
Feb. 4	09 30 28.3	14.7	90.6	5	5.8 m_b
6	04 11 03.3	14.6	91.1	5	4.8 m_b
6	18 11 59.2	14.3	90.2	5	5.2 m_b
6	18 19 17.7	14.7	90.6	5	5.8 m_b
8	08 13 51.9	15.7	88.5	5	5.7 M_s
9	11 44 46.7	15.3	89.1	5	5.1 m_b
10	06 17 43.0	15.0	89.7	5	4.7 m_b
Mar. 7	02 54 05.4	14.9	90.9	5	4.8 m_b
7	03 15 40.3	14.7	90.5	5	4.9 m_b

[1] Constrained.

damaging aftershocks could have occurred immediately after the main event, and their seismic signatures could be buried in the coda of the main event. The M_s=5.7 aftershock of February 8 is near the eastern end of the surface-fault rupture. The two aftershocks in table 3 that are not alined with the Motagua fault zone lie immediately north of the central Guatemalan volcanic lineament shown in figure 11.

Table 3 is a list of preliminary origin times, epicentral coordinates, and magnitudes of the principal aftershocks occurring through March 7, 1976, as located by NEIS. Focal depth was restrained to 5 km in the aftershock hypocenter determination. Figure 11 shows epicenters of the main event and principal aftershocks. There were no foreshocks located by the NEIS, and we have no foreshocks recorded at high-gain stations at teleseismic distances.

MAIN EVENT SOURCE PARAMETERS FROM TELESEISMIC DATA

By JAMES W. DEWEY and BRUCE R. JULIAN

INTRODUCTION

The Guatemala earthquake occurred on the Motagua fault, a strike-slip fault thought to be part of the boundary between the Caribbean plate and the North American plate. The characteristics of the focal mechanism of the Guatemala earthquake are of considerable interest to U.S. seismologists, because there may be similarities between this fault and strike-slip faults in California that are also associated with plate boundaries. Conversely, the history of strike-slip earthquakes in California and other regions may help anticipate the future course of the aftershock sequence of the Guatemala earthquake.

FOCAL MECHANISM

The P-wave first-motion pattern of the main event is shown in figure 12. The east-northeast-striking nodal plane corresponds to the fault plane of the earthquake. Motion across this plane is left-lateral strike-slip. The strike of the fault plane is well determined and is about N. 65° E. Possible dips vary from 84° N. to 82° S. The motion is of almost pure strike-slip character. The fault plane in figure 12 agrees well with the geologically mapped fault trace.

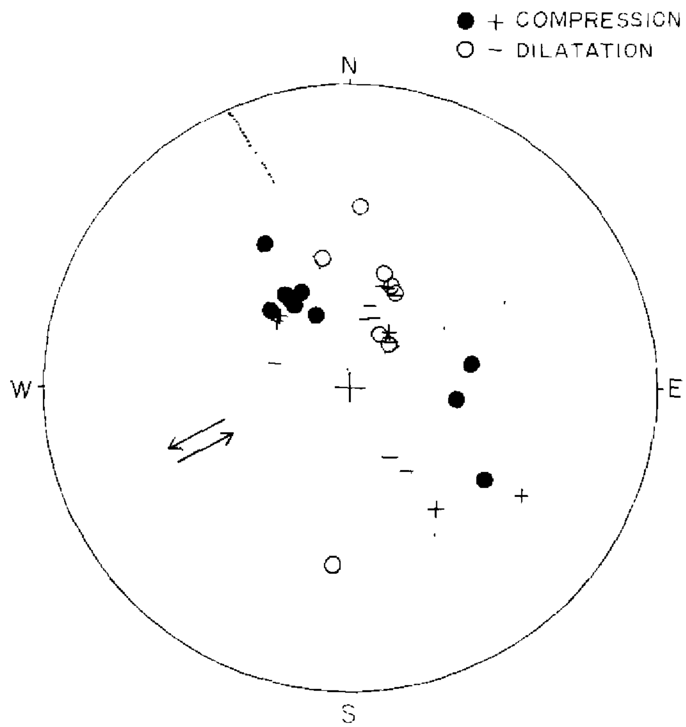

FIGURE 12.—Stereographic projection of P-wave first motions for the main event. Circles represent first motions especially measured for this study. Plus and minus represent first-motion data reported to the NEIS. The solution has been chosen to be consistent with all the readings made by us and, in addition, to be as consistent as possible with the first motions reported to the NEIS. The northeast-striking nodal plane corresponds in strike and sense of displacement to the Motagua fault.

SEISMIC MOMENT

The Guatemala earthquake produced mantle Love waves that had periods of around 100 s and that were well recorded by many seismographs of the WWNSS, from which the seismic moment of the earthquake can be determined. For this preliminary report, we estimated the displacement spectral density of the ground motion by multiplying the measured pulse amplitudes by 70 s for those phases with periods near 100 s (Brune and Engen, 1969). Amplitudes were normalized to an epicentral distance of 90° (see Brune and Engen, 1969) with Q, the seismic quality factor, taken to be 107 and the group velocity of the G-wave taken to be 4.4 cm/s (table 4 and fig. 13). These spectral densities were then adjusted for the orientation of the focal mechanism, assuming a vertical strike-slip fault as the earthquake source. More accurate determinations based on Fourier analysis of digitized records are planned.

TABLE 4.—*Distance, azimuths, and spectral density for G-waves*

Station [1]	Phase [2]	Δ (degrees) [3]	Θ (degrees) [4]	Spectral density [5] at 90° (cm-s)	Spectral density [6] corrected for focal mechanism
SJG	G2	337.7	14.5	7.11	5.18
SJG	G4	697.7	14.5	5.06	3.68
[7] KIP	G3	425.1	41.8	2.88	
[7] PRE	G2	239.1	46.1	1.98	
ALQ	G3	385.0	79.8	6.65	4.52
DUG	G3	392.2	80.0	5.09	3.45
LON	G3	401.5	81.0	5.10	3.41
GOL	G3	388.1	88.0	6.39	4.08
GIE	G2	334.1	118.8	5.40	6.42
[7] WES	G3	391.1	141.2	2.48	
[7] KIP	G2	294.9	221.8	2.41	
[7] PRE	G3	480.9	226.1	1.78	
[7] GUA	G2	240.7	230.4	1.49	
ALQ	G2	335.0	259.8	8.65	5.88
ALQ	G4	695.0	259.8	13.35	9.07
DUG	G2	327.8	260.0	5.21	3.53
DUG	G4	687.8	260.0	8.69	5.89
LON	G2	318.5	261.0	4.38	2.93
LON	G4	678.5	261.0	5.10	3.41
GOL	G2	331.9	268.0	5.30	3.38
COL	G2	297.8	270.9	5.89	3.75 (median)
COL	G4	657.8	270.9	5.86	3.73
GIE	G3	375.9	298.8	7.80	9.27
[7] GEO	G2	334.1	317.2	2.46	
[7] WES	G2	328.9	321.2	2.99	
PTO	G2	286.8	346.2	5.11	3.67
PTO	G4	646.9	346.2	7.42	5.33

[1] Station code listed in Glossary.
[2] Phase: defined in Glossary.
[3] Distance traveled by G-wave from source to station.
[4] Azimuth from source to station of the phase in question, measured clockwise from the direction of fault rupture propagation (S. 65° W.).
[5] Spectral density at 90°, estimated by multiplying the ground displacement by 70 s and applying the distance correction factor of Brune and Engen (1969).
[6] The correction factor, assuming a vertical strike-slip fault, is $[\pi/2 \cos 2\Theta]^{-1}$.
[7] These stations were within 15° of a G-wave nodal plane and have not been used in the computation of moment.

Several of our stations lay near nodes of the theoretical Love-wave radiation pattern (fig. 13). Because the radiation-pattern correction for these stations is subject to large uncertainty, we used in the computation of seismic moment only those stations that were well removed (more than 15°) from the theoretical nodal planes (table 4). The mean of the adjusted displacement spectral densities is 4.77±0.43 cm-s; the median is 3.75 cm-s. The mean seems unduly influenced by a few large values, and we take the median as more representative of the sample as a whole. The seismic moment corresponding to a displacement spectral density of 3.75 cm-s is 2.6×10^{27} dyne-cm (Brune and Engen, 1969).

The provisional surface-wave magnitude assigned to the main event was $M_S=7.5$ (Person and others, this report). The amplitudes of 100-s surface waves do not support a major adjustment of this magnitude; the moment computed from these amplitudes falls in the middle of the "cloud" of data points in Brune and Engen's (1969) graph of 20-s surface-wave magnitude versus moment.

SOURCE DIMENSIONS, DISPLACEMENT, STRESS DROP, AND DIRECTION OF FAULT PROPAGATION

The zone of the largest (M>5.0) best located aftershocks of the main event, occurring during the first week, is about 250 km long (Person and others, this report). This zone of the largest aftershocks coincides very closely with the zone of surface faulting mapped after the earthquake (Plafker and others, this report). Possibly an additional 50 km of fault rupture could be postulated on the basis of small aftershocks recorded after the earthquake (Langer and others, this report) and from high damage west of Guatemala City (Espinosa and others, this report). For the purpose of the analysis that follows, we shall take the fault length as 300 km.

The relationship between seismic moment (M_o), fault length (L), width (w), and average displacement ($\overline{D}$) is (Aki, 1966)

$$M_o=\mu L w \overline{D}, \qquad (1)$$

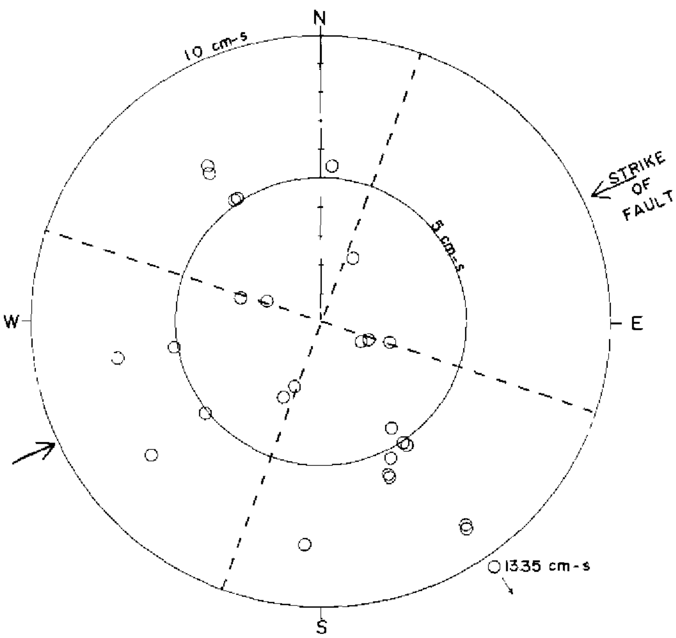

FIGURE 13.—Azimuthal variation of the displacement spectral density normalized to an epicentral distance of 90°. The dashed lines are theoretical nodal lines of the Love-wave radiation pattern for a vertical strike-slip fault, striking N. 65° E.

where μ is the rigidity of the faulted medium, here taken to be 3×10^{11} dynes cm^{-2}. The average displacement, $\overline{D}$, is inferred from geologic observation to be 100 cm (Plafker and others, this report). Together with our estimate of seismic moment, a fault length of 300 km and an average displacement of 100 cm implies

$$w=29 \text{ km}.$$

This fault width is apparently greater than those of earthquakes on California's San Andreas fault; the seismogenic fault width associated with the California earthquake of 1906, for example, is thought to be 10 km (Thatcher, 1975).

If we had chosen M_0 equal to the mean rather than the median of the moments observed at individual stations, if we had taken $L=250$ km rather than $L=300$ km, and if we had taken $\overline{D}$ less than 100 cm, the discrepancy between the fault width of the Guatemala earthquake and that of the 1906 California earthquake would be even greater. There are alternative explanations for this discrepancy:

1. Displacement on the fault at depth may be larger than surface-fault displacement, so that 100 cm would be significantly smaller than the actual $\overline{D}$. The difficulty with this explanation is that the displacements observed at the surface are quite uniform over long distances (Plafker and others, this report); it is hard to visualize a process acting uniformly over 100 or more kilometres that would retard surface-fault slippage relative to slippage at depth. In fact, one might make the contrary argument—that seismic-fault displacement on a long strike-slip fault will have a tendency to decrease with depth from the free surface.
2. Seismic rupture on the Motagua fault may actually have extended to several tens of kilometres in depth. Such rupture would have to produce a large amount of long-period energy in order to significantly affect amplitudes of 100-s G-waves. However, the fault rupture at depth need not necessarily have produced a large amount of short-period energy. Likewise, the shallow depths of aftershocks recorded by Langer, Whitcomb, and Aburto (this report) do not preclude fault rupture extending several tens of kilometres into the crust, since such rupture could occur comparatively slowly in a medium that is incapable of producing high-frequency strike-slip earthquakes.

The stress drop, $\Delta\sigma$, for the main event may be estimated from

$$\Delta\sigma=\frac{2}{\pi}\mu\left(\frac{\overline{D}}{w}\right) \text{(Knopoff, 1958).} \quad (2)$$

With $\mu=3\times10^{11}$ dynes cm^{-2}, $\overline{D}=100$ cm, and $w=29$ km,

$$\Delta\sigma=6.6 \text{ bars}.$$

A stress drop of 6.6 bars is less than the worldwide average for interplate earthquakes; Kanamori and Anderson (1975) find that 30 bars is typical for such earthquakes. If, as discussed above, w were less than 29 km, the stress drop would be correspondingly larger.

The epicenter of the main event lay about 90 km from the eastern end of the inferred 300-km-long zone of fault rupture, a position that suggests that the fault rupture propagated from northeast to southwest. The level of shaking near the western end of the fault might be expected, under such circumstances, to be higher than the level of shaking at the eastern end of the fault because of constructive interference of waves from the propagating source. The mantle-wave observations tend to support such a conclusion, the amplitudes being larger for waves leaving the event to the southwest

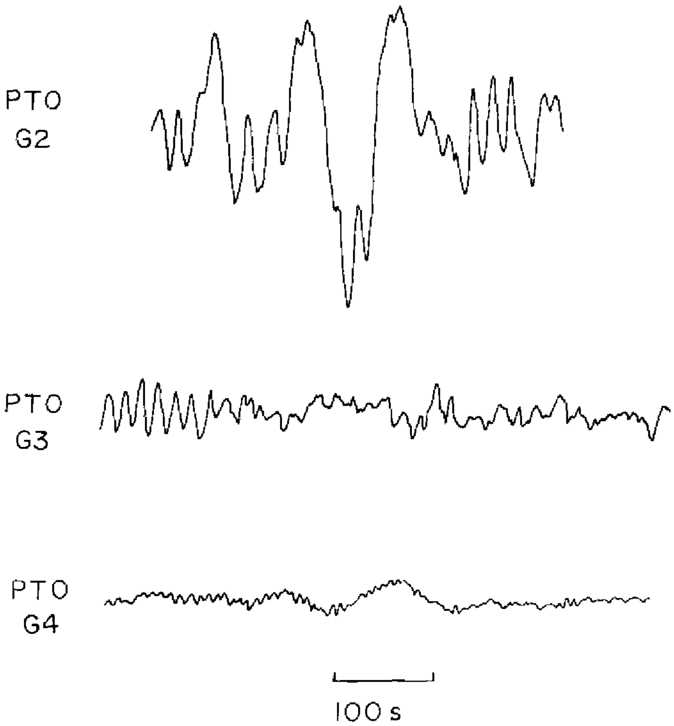

FIGURE 14.—Effect of source propagation on G-wave amplitudes at station PTO. G2 and G4 left the source at an angle of about 15° from the direction of fault rupture. G3 left the source at an angle of 165° from the direction of fault rupture. Note that the amplitude of G4 is at least equal to that of G3, although G4 has traveled 213.6° farther than G3.

than for those leaving toward the northeast. Figure 14 shows a striking case of this phenomenon in the records from Pôrto, Portugal. The high intensities observed near the western end of the Motagua fault (Espinosa and others, this report) may thus be, in part, an effect of source propagation.

FUTURE EARTHQUAKES ON THE MOTAGUA FAULT SYSTEM

At the time this paper was written (April 1976), Guatemala had experienced several moderate ($5<M\leq6$) aftershocks. In order to anticipate the future activity of the Motagua fault, we may study the seismic history of other major continental strike-slip faults. We shall consider the North Anatolian fault system, the seismicity of which one of us (Dewey, 1976) has recently studied. The following conclusions seem consistent also with the history of large earthquakes on California's San Andreas fault; they may therefore be valid for other major continental strike-slip faults like the Motagua fault:

1. Major earthquakes, involving hundreds of kilometres of fault rupture, do not tend to recur on the same segment of a strike-slip fault within a short period of time. This conclusion is based on the three Anatolian earthquakes comparable in size to the Guatemala earthquake: those of December 26, 1939, November 26, 1943, and February 1, 1944, none of which have yet been followed by earthquakes of comparable size on the same section of the fault. Likewise, neither of the great San Andreas earthquakes of January 9, 1857, or April 18, 1906, has yet been followed by another great earthquake on the same section of the fault.
2. Large sections of a strike-slip fault ruptured in a large earthquake will not produce aftershocks of magnitude greater than 5. This conclusion is based on the characteristics of aftershocks of the Anatolian earthquakes; it is consistent with reports of aftershocks to the San Andreas earthquake of 1906 (Dewey, 1976). To date, this conclusion seems to be valid for the Guatemala earthquake.
3. Those moderate aftershocks that do occur will tend to be concentrated near the ends of the fault rupture. This has thus far been the case with the Guatemala earthquake; the largest aftershocks ($M\geq5.5$) have occurred near Guatemala City, at the western end of the fault break, and near the eastern end of the fault break (Person and others, this report).
4. Regions near the ends of the fault rupture may continue to experience moderate earthquakes for some years following the main event. It seems apparent that occurrence of the large shock does not significantly reduce the level of tectonic strain in the regions near the ends of the fault rupture. In fact, on theoretical grounds, the occurrence of the major earthquake should produce high tectonic strain near the extremities of the fault rupture (Chinnery, 1963).

For Guatemala, conclusions 1–4 imply that most of the Motagua fault ruptured by the earthquake will be seismically inactive during the next decades. The regions near the ends of the main fault rupture, near Guatemala City and south of Puerto Barrios, may, however, have several moderate earthquakes ($5<M<6$) in the coming decade.

There is also the possibility that the occurrence of the main event could induce, in the next several decades, a major earthquake on a segment of the Motagua fault adjacent to the fault ruptured by the February 4 earthquake. Such a migration of seismic sources over a period of several decades seems to have occurred on the North Anatolian fault and has also been postulated for the San Andreas fault (Savage, 1971).

There seems to be ample fault length to generate a major earthquake east of the rupture of February 4 where the Motagua fault trends into the tectonically similar Swan fracture zone in the Gulf of Honduras (Spence and Person, this report). In the west, the likelihood of a future major earthquake, similar to this earthquake but centered to the west of it, may depend on whether the Motagua fault persists for hundreds of kilometres as a continuous fault west of the fault rupture of February 4, 1976.