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Seismicity and tectonics of El Salvador

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ABSTRACT

The large-scale plate-tectonics framework of El Salvador was defined in the "plate-tectonics revolution" of the 1960s and 1970s, but important issues related to seismic hazards depend on details that have been only recently, or are not yet, understood. Present evidence suggests that coupling across the interface-thrust zone beneath coastal El Salvador is sufficient to produce occasional interface-thrust earthquakes as large as M ~8. The rate of such earthquakes is determined by the percentage of relative plate motion that is accumulated as elastic strain on the thrust-fault interface between the Cocos and Caribbean plates, which appears to be lower than in many other subduction zones, but is not well established. Earthquakes in the interior of the Cocos plate, such as the El Salvador earthquake of January 13, 2001, account for a significant percentage of Wadati-Benioff zone earthquakes. Separate consideration of the seismic hazard posed by, respectively, Cocos intraplate earthquakes and interface-thrust earthquakes is complicated by the difficulty of separating interface-thrust and Cocos intraplate events in earthquake catalogs. Earthquakes such as the San Vicente–San Salvador sequence of February 13–25, 2001, probably result from the motion of the Central American forearc northwestward with respect to the interior of the Caribbean plate; the geometry of the fault systems that accommodate the motion remains to be worked out. Understanding of this tectonic complexity and associated seismic hazards will be facilitated greatly by the long-term operation of high-sensitivity local seismograph networks, such as that operated by, and currently being upgraded by, the Servicio Nacional de Estudios Territoriales (SNET) of El Salvador.

Keywords: El Salvador, Central America, earthquake, earthquakes, seismicity, seismotectonics, seismic hazards.

INTRODUCTION

The high level of earthquake activity in El Salvador is a consequence of its position at the boundary of two major tectonic plates, the Cocos plate and the Caribbean plate (Fig. 1; Molnar and Sykes, 1969; Dewey and Suárez, 1991). El Salvador lies on the Caribbean plate. The subduction of the Cocos plate beneath the Caribbean plate along the Pacific coast of Central America produces several classes of damaging earthquakes that are distinguished by their positions within the tectonic plates or by their focal mechanisms (Figs. 1–5). Beneath the Pacific coast and offshore, thrust-fault earthquakes occur on the interface between the Cocos and Caribbean plates, and significant activity also occurs below the interface within the interior of the Cocos

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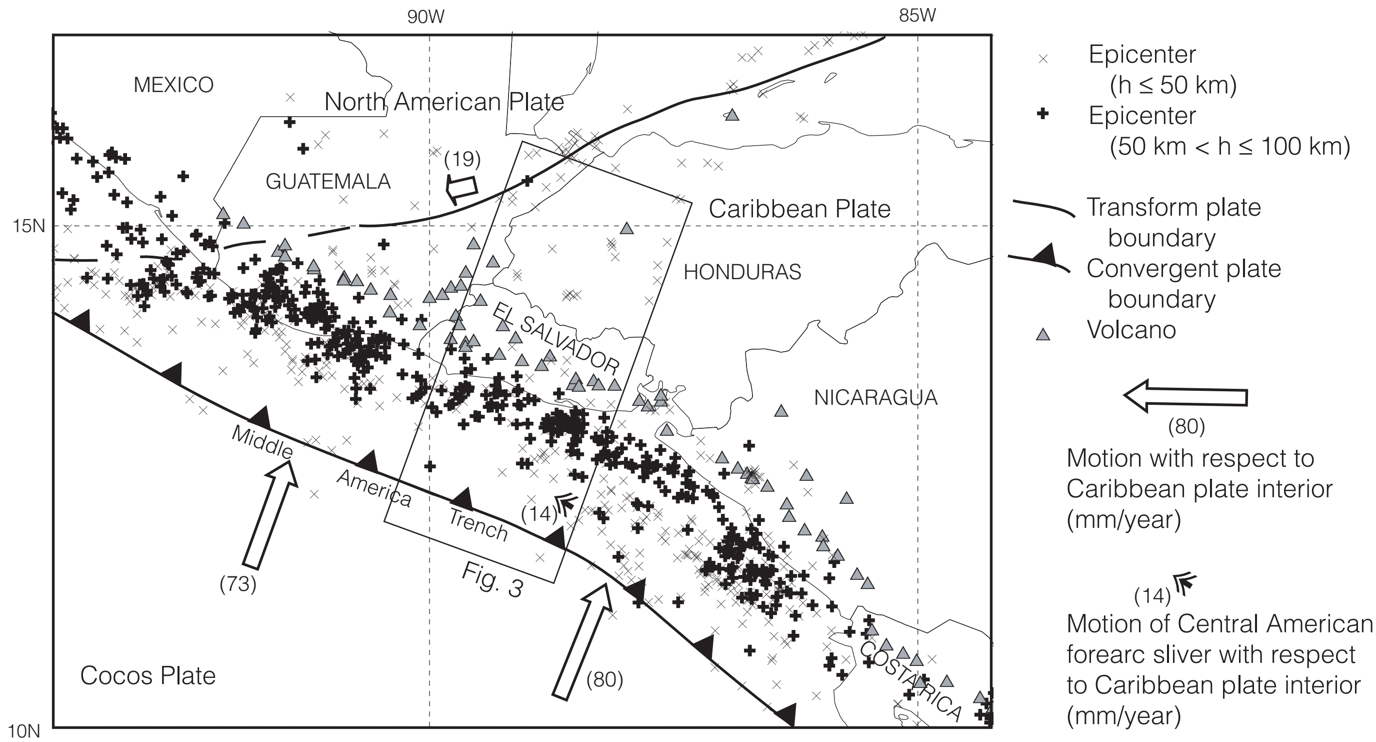

Figure 1. Tectonic plate boundaries near El Salvador, and the distribution of teleseismically recorded earthquakes calculated to have depths of 100 km and less in 1964–2000. Plate motions and motion of the Central American forearc sliver are from DeMets (2001). See Figure 2 for the location of the Central American forearc sliver. Volcanoes are from Siebert and Simkin (2002). Earthquake epicenters and focal depths are computed by the method of Engdahl et al. (1998) and were provided to the authors by E.R. Engdahl (2002, personal commun.; see text). Ninety percent of the earthquakes plotted had magnitudes of 4.2 or larger.

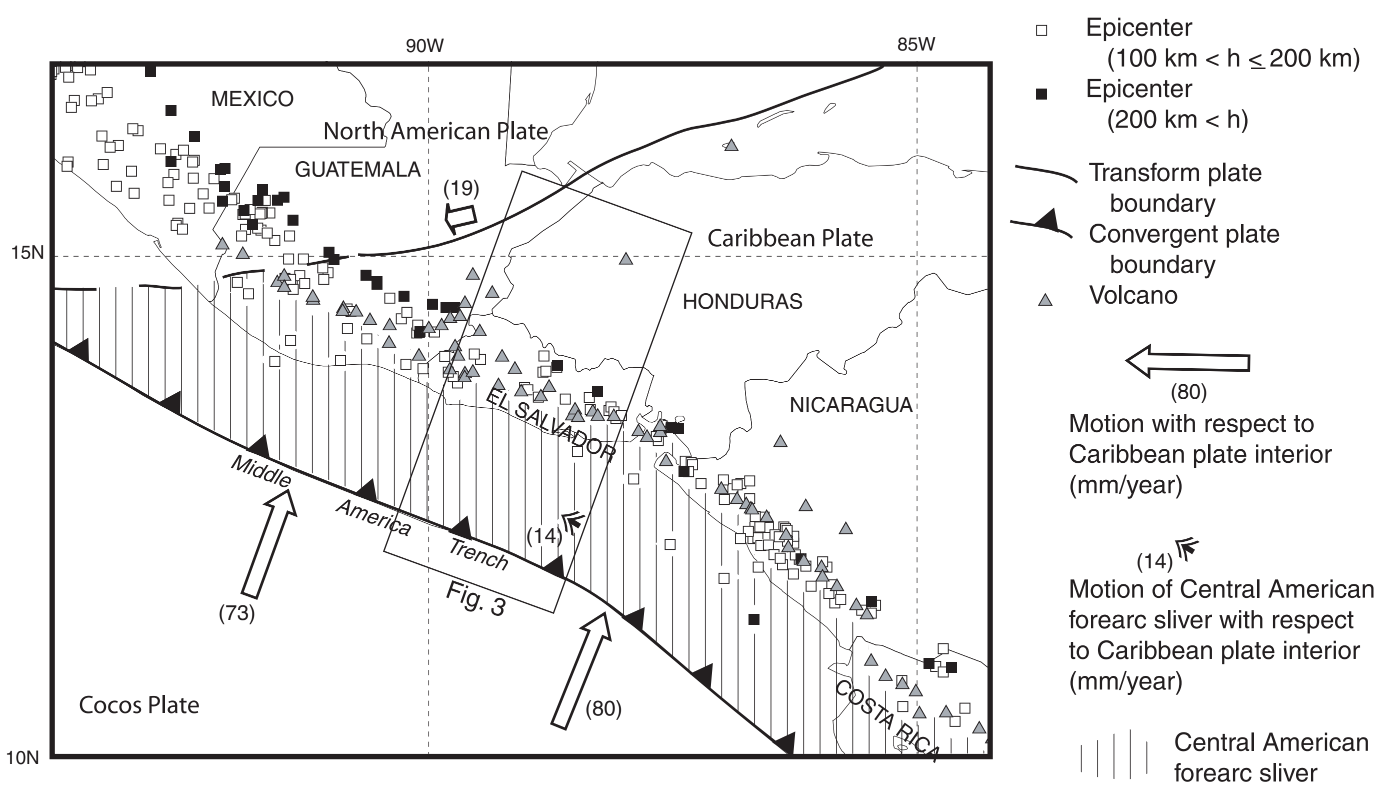

Figure 2. Distribution of earthquakes calculated to have depths greater than 100 km, for the period 1964–2000. Sources of earthquake data and information on plate boundaries, plate motions, and volcanoes are as in Figure 1.

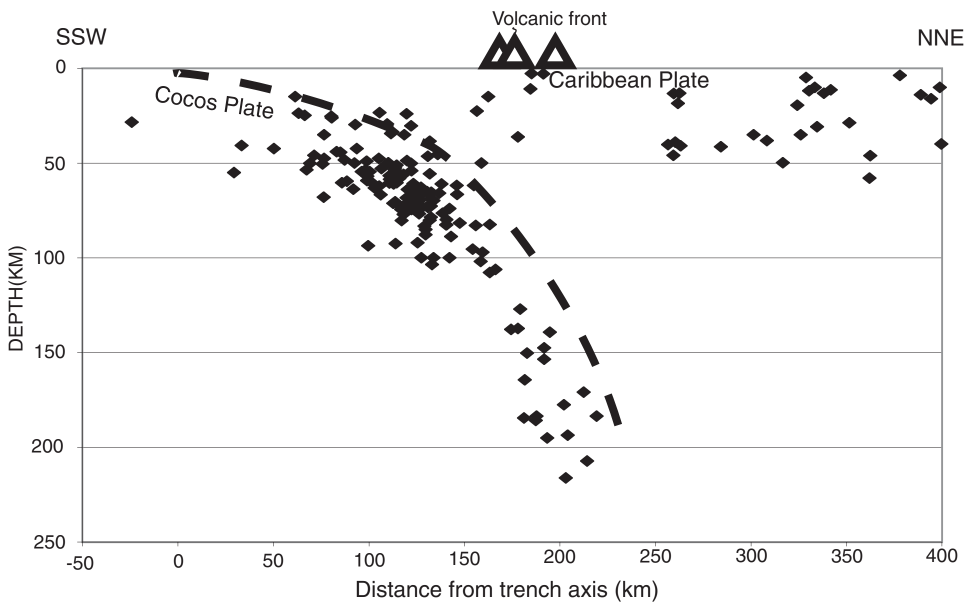

Figure 3. Cross section of seismicity beneath El Salvador, 1964–2000. Hypocenters plotted are those from Figures 1 and 2 lying within the "Fig. 3" box of those figures. Dashed curve represents the approximate location of the Cocos-Caribbean plate interface (see also Figure 5). The zone of hypocenters near and beneath the plate interface is the Wadati-Benioff zone.

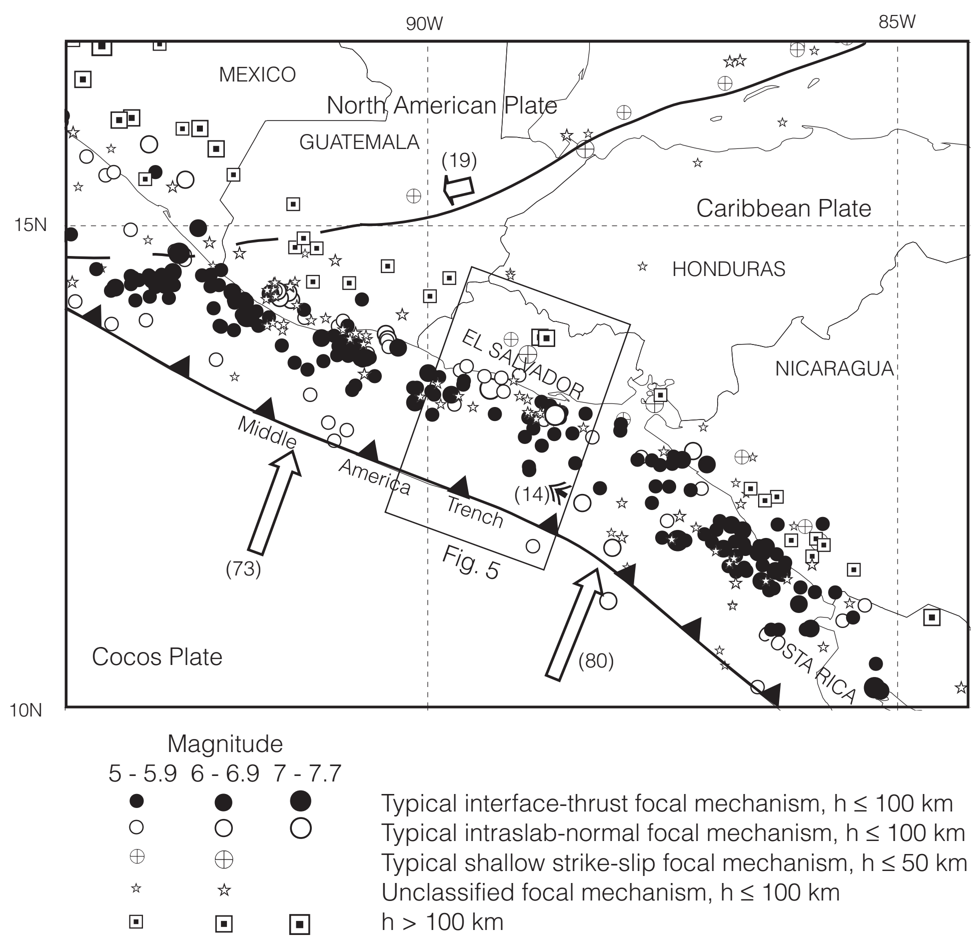

Figure 4. Larger earthquakes from El Salvador and vicinity, 1978–April 2001, classified by focal mechanism as determined by the Harvard CMT methodology (Dziewonski et al., 1981). See text for explanation of "typical interface-thrust" and "typical intraslab-normal" focal mechanisms. Epicenters plotted here are those routinely computed by the U.S. Geological Survey National Earthquake Information Center (USGS/NEIC). They are likely to differ by 5–10 km from those of the same events that are plotted in Figures 1–3. Note that far fewer events are plotted here than in Figures 1 and 2: The time span from which events are selected is smaller for this figure, and the only events plotted are those for which a Harvard CMT was determined. Plate boundaries and plate motions are as in Figure 1.

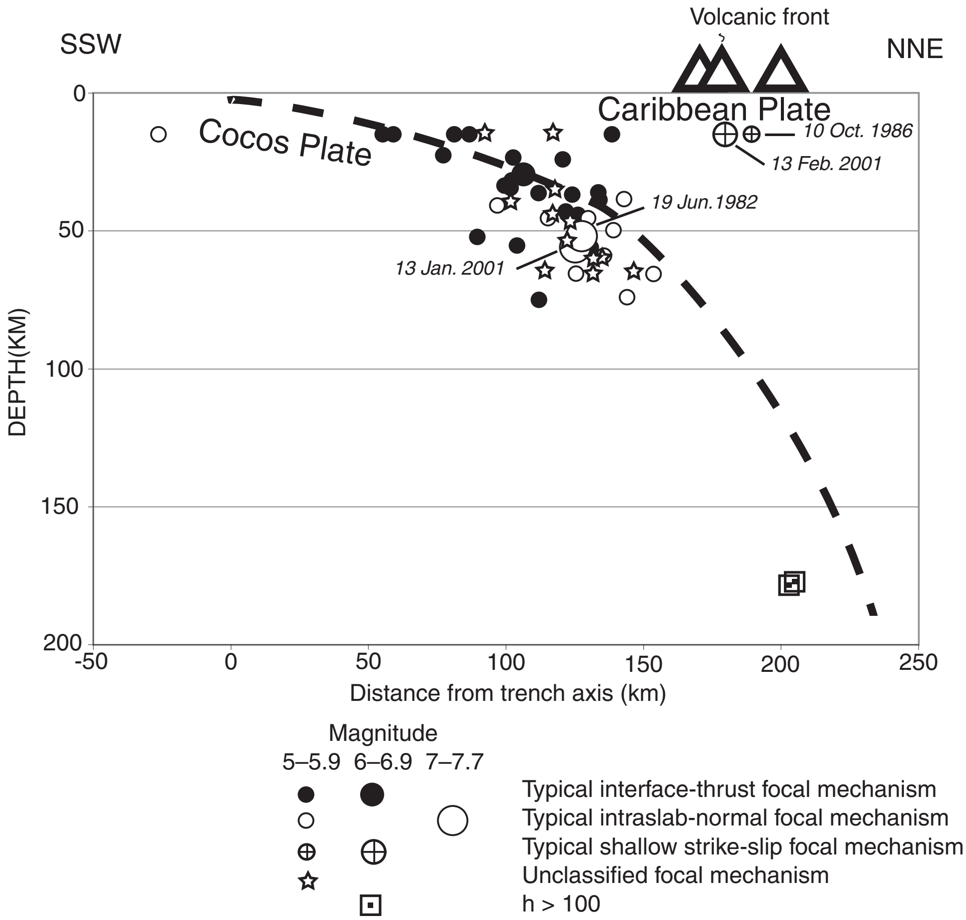

Figure 5. Cross section of earthquake hypocenters, 1978–April 2001, classified by focal mechanism as determined by the Harvard CMT methodology (Dziewonski et al., 1981). Events are those that lie within the box labeled “Fig. 5” on Figure 4. Hypocenters of the most destructive El Salvadoran earthquakes since 1978 are labeled by their dates. Shallow-focus earthquakes for which the hypocentroid cannot be accurately determined by Harvard CMT methodology are assigned default depths of 15 km.

plate. Inland from the coastline, intraplate earthquakes within the subducted Cocos plate occur to depths of more than 200 km beneath El Salvador (Fig. 3) and to more than 250 km elsewhere in Central America. A societally very significant class of earthquakes occurs at shallow depths in the overriding Caribbean plate, in the vicinity of the volcanic chain. The largest known earthquakes from El Salvador have had magnitudes of ~7.7. Shallow-focus earthquakes with magnitudes (m_b) less than 4.5 have been damaging when they have occurred directly beneath population centers.

SOURCES OF DATA ON EL SALVADOR EARTHQUAKES

Seismologists’ current understanding of the seismicity of El Salvador is based on a record of earthquake locations, sizes, and focal mechanisms that becomes generally more complete and accurate as one approaches the present. Knowledge of earthquakes occurring before the twentieth century must be based on the macroseismic effects produced by the earthquakes (e.g., White et al., this volume, Chapter 28). Seismographic data contribute usefully to knowledge of Central American seismicity from early in the twentieth century, becoming steadily more informative with the continued spread of seismographs (Ambraseys and Adams, 2001). In the early 1960s, a major improvement in the completeness of Central American earthquake catalogs resulted from the introduction of computer determinations of earthquake hypocenters and magnitudes as well as from the installation of the World-Wide Standardized Seismograph Network (WWSSN). Hypocenters and magnitudes have been routinely determined by computer since 1964 by the U.S. Geological Survey National Earthquake Information Center (USGS/NEIC) and the International Seismological Center (ISC), using data from seismographs distributed globally. Data from Salvadoran seismographic stations at San Salvador (international abbreviation SSS), La Palma (LPS), Nueva Concepcion (NCS), and Santiago de María (SDM) were able to contribute importantly to the locating of earthquakes in El Salvador and elsewhere in Central America by the USGS/NEIC and the ISC. Catalogs of earthquakes computed by the USGS/NEIC and the ISC are available online, at http://neic.usgs.gov/ and http://www.isc.ac.uk/, respectively. The installation of the WWSSN also made possible the determination of reliable focal mechanisms for larger Central American earthquakes by analysis of the geographic distribution of P-wave first motions (Molnar and Sykes, 1969; Dean and Drake, 1978). A dramatic increase in the number of digitally recording long-period and broadband seismographs in the 1970s led to the development of more sophisticated methods of describing the earthquake focal mechanism and accurately locating the earthquake source. Moment tensors of larger Central American earthquakes are independently determined and made available online by Harvard University (Dziewonski et al., 1981; http://www.seismology.harvard.edu/CMTsearch.html) and the USGS (Sipkin, 1982; http://neic.usgs.gov/neis/sopar/).

Teleseismic Hypocenters

In Figures 1–3, we have represented the seismicity of El Salvador and vicinity by use of hypocenters for earthquakes occurring in 1964–2000 that were calculated with the methodology of Engdahl et al. (1998) and provided to us by Engdahl (2002, personal commun.). The particular set of hypocenters plotted represents a subset of those determined by Engdahl and his colleagues, those for which the largest teleseismic "secondary open azimuth" (azimuthal spread between reporting seismographs that contains only one other reporting seismograph) is less than 130°. Approximately 25% of all teleseismically recorded earthquakes occurring since 1964 that are cataloged by the USGS/NEIC and ISC meet this criterion. Based on comparison of individual epicenters from Figures 1 and 3 with independently known epicenters of the same earthquakes, we think that the epicenters (but not the focal depths) of most shallow-focus inland earthquakes represented in Figures 1 and 3 are probably accurate to within 20 km. A significant fraction of location errors would be due to bias arising from the use of one-dimensional velocity models in an earth with three-dimensional velocity structure (e.g., Dewey and Algermissen, 1974). Epicenters within different parts of the Wadati-Benioff zone may also be biased by tens of kilometers, though we are not able to check for the presence of these biases without reliable independently determined epicenters. Uncertainties in focal depths shown in Figure 3 are likely to be several tens of kilometers. We do not, for example, attach tectonic significance to focal depths calculated to be greater than 25 km in the Caribbean plate at distances of 250–400 km inland from the Middle America trench (Fig. 3), although such focal depths would require a tectonic explanation if they were considered reliable. In spite of the uncertainties in the hypocenters of Figures 1–3, these hypocenters should still be on average more accurate than those routinely computed by the USGS/NEIC or the ISC (Engdahl et al., 1998). The imposition of a 130° largest teleseismic secondary open azimuth for Figures 1–3 dramatically reduces the scatter of hypocenters for earthquakes smaller than magnitude 5.5, compared to what would be obtained by plotting all USGS/NEIC hypocenters for the regions of Figures 1–3.

In Figures 4 and 5, we have represented the seismicity of El Salvador and vicinity by plotting hypocenters of earthquakes for which Harvard CMT (Centroid-Moment Tensor) focal mechanisms were calculated by the methodology of Dziewonski et al. (1981). The epicenters in Figures 4 and 5 are those computed by the USGS/NEIC. The focal depths plotted in Figure 5, however, are the hypocentroid focal depths that are computed from long-period seismograph waveforms along with the Harvard CMT mechanisms. In the case of many earthquakes for which the routine Harvard CMT methodology would calculate a focal depth shallower than 15 km, the Harvard CMT convention is to fix the depth at 15 km, in order to avoid instability in the CMT computation (Dziewonski et al., 1987). For most events, the Harvard CMT focal depths should be more reliable than the focal depths computed by the USGS/NEIC from times of first-arriving P-waves. In the absence of errors, the hypocenters computed by the methodology of Engdahl et al. (1998) (Figs. 1–3) and the epicenters computed by the USGS/NEIC would represent the points at which the earthquakes nucleated. The Harvard CMT hypocentroid depths, by contrast, correspond to the centroid of moment release (Dziewonski et al., 1981).

Teleseismic Magnitudes

Teleseismically determined long-period magnitudes, such as M_S or M_W, are currently the most useful earthquake magnitudes for seismic hazard studies in Central America (Ambraseys and Adams, 2001). Instrumentally determined M_S are available for large Central American earthquakes since the beginning of the twentieth century (Ambraseys and Adams, 2001), and M_W have been systematically determined for larger earthquakes since the late 1970s. Since 1964, teleseismic short-period body-wave magnitudes (m_b) have been assigned to earthquakes worldwide by the USGS/NEIC and the ISC. The m_b are important because they can be determined for smaller earthquakes than can M_S and M_W; they are the only instrumental measure of earthquake size for many moderate-sized earthquakes. Teleseismic m_b cannot, however, be automatically assumed to be identical to other types of magnitude (Geller, 1976). Effective use of teleseismic m_b jointly with long-period magnitudes for seismic hazard studies of Central America may require development of special, source-region specific, scaling relationships between m_b and the long-period magnitudes. Teleseismic m_b values assigned to destructive shallow-focus earthquakes along the Central American volcanic front seem disproportionately small with respect to other types of earthquake magnitude, more so than would be expected from worldwide scaling relationships (e.g., Geller, 1976) between different types of magnitude. For example, the destructive El Salvador earthquake of October 10, 1986 [M_S(USGS) = 5.4; M_W(HRV) = 5.7], had m_b(USGS) = 5.0, and the earthquake of February 13, 2001 [M_S(USGS) = 6.5; M_W(HRV) = 6.6], had m_b(USGS) = 5.5.

Local Earthquake Monitoring

Currently, monitoring of small and moderate earthquakes within the overall territory of El Salvador is conducted by seismographs run by the Servicio Nacional de Estudios Territoriales (SNET) (Fig. 6). The network of Salvadoran stations is capable of

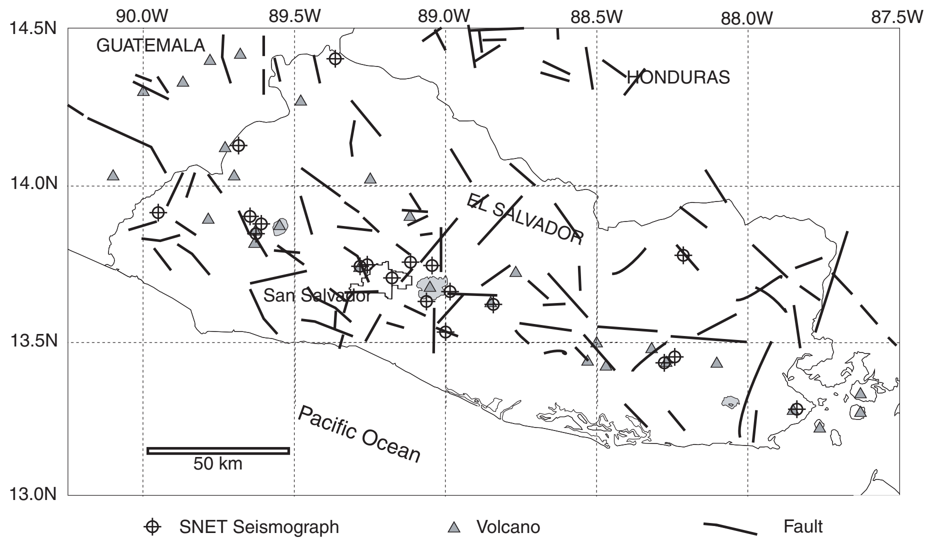

Figure 6. Seismographs operated in mid-2002 by the Servicio Nacional de Estudios Territoriales (SNET) for the monitoring of the seismicity of El Salvador, together with volcanoes (source as in Figure 1) and geologically mapped faults (Case and Holcombe, 1980).

locating earthquakes of magnitude 3.8 from throughout the area of El Salvador, and substantially smaller shocks in areas of dense instrumentation. At locations in El Salvador that are surrounded by nearby stations of the SNET network, epicenter accuracies of several kilometers can be claimed, with focal depth accuracies being somewhat poorer. Accuracies of hypocenters lying outside of the network would generally be much poorer. Expansion of the network is currently being undertaken to extend the region within which small earthquakes can be detected and earthquake hypocenters can be calculated to high accuracy. Accurate location of earthquakes on the borders of El Salvador will depend on observations from seismographs situated in neighboring countries, such as the Nicaraguan seismic network (Segura and Havskov, 1994). In addition to the network of sensitive stations used to monitor small earthquakes, several groups operate strong-motion networks in El Salvador (Shakal et al., 1987; Bommer et al., 1997). Although installed primarily to study the characteristics of the damaging ground shaking, data from the strong-motion seismographs can also provide important constraints on the locations of strong local earthquakes (e.g., White et al., 1987).

THE CENTRAL AMERICAN WADATI-BENIOFF ZONE

Earthquakes occurring in the Wadati-Benioff zone along the coast of Central America pose a significant hazard to El Salvador and other Central American countries. Earthquakes in this coastal Wadati-Benioff zone occur mostly at depths of ~100 km and less (Figs. 1–3), and most large shocks occur at depths shallower than 70 km (Fig. 5). In the twentieth century, earthquakes as large as M_S 7.9 have produced Modified Mercalli intensities as high as IX in Central America north of Costa Rica (Ambraseys and Adams, 2001). At least one coastal Wadati-Benioff zone earthquake produced a destructive tsunami.

Classification of Wadati-Benioff Zone Earthquakes by Focal Mechanism

The seismicity of the Central American coastal Wadati-Benioff zone comprises earthquakes occurring on the thrust interface between the Cocos and Caribbean plates and earthquakes occurring beneath the interface in the interior of the Cocos plate. In Figures 4 and 5 we have classified events occurring in 1978–April 2001 on the basis of their Harvard CMT focal mechanism (Dziewonski et al., 1981). Events plotted as "typical interface-thrust focal mechanism" earthquakes in Figures 4 and 5 had focal mechanisms consistent with their occurring as reverse motion on shallowly dipping planes, with slip being approximately parallel to the 30° azimuth at which slip occurs between the Cocos plate and the Caribbean forearc sliver (DeMets, 2001): These events had T-axes that plunge more than 45°, and they had one slip vector with an azimuth between 10° and 50° and a plunge less than 45°. Events plotted as "typical intraslab-normal focal mechanism" earthquakes in Figures 4 and 5 had focal mechanisms consistent with their occurring by normal faulting caused by extension approximately parallel to the downdip direction of the Wadati-Benioff zone: These events had P-axes that plunge more than 45° and T-axis azimuths between 0° and 50°. Most events in Figures 4 and 5 that are classified as "typical" interface-thrust and intraslab-normal earthquakes on the basis of their focal mechanisms had locations consistent with locations of interface-thrust and intraslab, normal-fault mechanisms in subduction zones around the world (e.g., Astiz et al., 1988), and we have defined the likely position of the Cocos-Caribbean thrust interface beneath El Salvador in agreement with the global pattern (Fig. 5).

Eventually, seismic hazard studies for El Salvador and for Central America in general will want to distinguish between the hazard caused by the coastal Wadati-Benioff zone interface-thrust and Cocos intraplate earthquakes, respectively (Bent and Evans, this volume, Chapter 29). A methodology for assessing seismic hazard that depends on physical models of strain accumulation will need to distinguish between earthquakes that release strain accumulated through different mechanisms. In addition, some studies (e.g., Youngs et al., 1997) suggest that shallow intraplate earthquakes tend to produce stronger shaking at a given epicentral distance than interface-thrust earthquakes of the same magnitude. These differences might reflect systematic differences in stress-drops between the two types of earthquakes, differences in radiation patterns from the differently oriented faults, or different focusing and attenuation along the slightly different paths from source to site. Finally, physical models for estimating the tsunami hazard to the Pacific coast of Central America may need to distinguish between interface-thrust earthquakes and Cocos intraplate earthquakes. At present, a separate accounting for interface-thrust earthquakes and Cocos intraplate earthquakes in seismic hazard studies is complicated by the difficulty of separating the two types of earthquakes in earthquake catalogs. Many of the Cocos intraplate earthquakes occur near the seismically active part of the interplate thrust interface, and it is commonly not possible to determine whether an earthquake was a Cocos intraplate earthquake or an interplate thrust-interface earthquake solely on the basis of its teleseismically determined hypocenter. Large earthquakes occurring in decades since the 1960s can be classified on the basis of their seismographically determined focal mechanism, but instrumental focal mechanisms have not been determined for most large earthquakes occurring before the early 1960s.

Seismic Gaps and Variations in Coupling across Subduction Zones

An example of a seismic-hazard assessment methodology that depends on a physical model of strain accumulation is the "seismic gap" methodology as used by McCann et al. (1979) and Nishenko (1991). In its general form, this methodology is intended to assess the probabilities of great earthquakes on plate interfaces only. The methodology is based on the assumptions that (1) strain on an interface accumulates at a rate that is proportional to the rate of relative plate motion and (2) the likelihood of a great earthquake on a particular segment of a plate interface increases with the time that has elapsed since strain was locally released by the last great earthquake on that segment. The methodology requires a catalog of past earthquakes on the interface to identify the times of the most recent great earthquakes on each segment of the plate interface and to estimate the proportion of strain on the plate boundary that is released seismically. Using the seismic gap methodology, Nishenko (1991) characterized the interface-thrust zone offshore of central and eastern El Salvador and western Nicaragua as having "significant but unknown seismic potential," and he characterized the interface-thrust fault offshore of western El Salvador and southeastern Guatemala as having a rather high probability for the recurrence of a magnitude 7.5 earthquake in upcoming decades. These characterizations were, however, based on poorly understood earthquakes of the early twentieth century and earlier. The earthquakes' focal mechanisms have not been instrumentally determined, and their hypocenters are not well enough determined that they can be definitely ascribed to rupture on the plate interface, instead of rupture in the interior of the Cocos plate. If some of the early earthquakes, instead of being interface-thrust earthquakes, were Cocos intraplate earthquakes, they would not be appropriate for use in the seismic gap methodology to assess the likelihood of future interface-thrust earthquakes.

Application of the seismic gap methodology to the Central American subduction zone is complicated by studies suggesting that interface-thrust zones worldwide differ in the sizes and frequencies of destructive earthquakes that they produce, corresponding to differences in the proportion of relative plate motion that is released respectively seismically and aseismically in the different zones. The differences in seismic behavior are broadly correlated with differences in the large-scale geotectonic characteristics of the subduction zones and with differences in the morphology of their Wadati-Benioff zones. The seismotectonic characteristics of a given subduction zone may be viewed against the characteristics of two "end-member" subduction zones—the "Chilean-type" and "Marianas-type" subduction zones (Uyeda and Kanamori, 1979; Kanamori, 1986; Jarrard, 1986). The Chilean type of subduction zone is characterized by a shallowly dipping Wadati-Benioff zone and an overriding plate that is strongly compressional; the Marianas-type subduction zone is characterized by a steeply dipping Wadati-Benioff zone and an overriding plate that is strongly extensional. The adjacent plates are tightly coupled across the interface thrusts of Chilean-type subduction zone; accumulated strain is mostly released in great thrust-fault earthquakes. The adjacent plates are loosely coupled across the interface thrusts of a Marianas-type subduction zone; relative motion between the plates is mostly accommodated aseismically, and great thrust-fault earthquakes do not occur along the margin.

The Central American subduction zone is intermediate in large-scale geotectonic style between the Chilean type and the Marianas type, and its characteristics change along strike. The part of the Central American Wadati-Benioff zone beneath Guatemala, El Salvador, and Nicaragua has a much steeper dip than Chilean-type subduction zones, with the dip slightly shallowing from south to north (Burbach and Frohlich, 1986). On a scale of 1 (active backarc spreading) to 7 (very strongly compressional), Jarrard (1986) classifies the overriding plate of the Nicaraguan section of the Central American subduction zone as a 3 (mildly tensional). The Costa Rica section of the Central American subduction zone has a tectonic style that is more Chilean-type than the rest of the Central American subduction zone; it is characterized by a much more compressive geotectonic environment and a shallower, less steeply dipping Wadati-Benioff zone (Burbach et al., 1984; Protti et al., 1994). Subduction zones that are intermediate in tectonic character and Wadati-Benioff zone morphology, between the Chilean type and the Marianas type, are expected to have seismic behavior that is in some sense intermediate between behaviors characteristic of the Chilean-type and Marianas-type zones, although the seismic behavior of such intermediate-style zones differs widely, according to the sizes and distribution of asperities (patches across which the plates are tightly coupled) on the interface thrust of each zone (Lay et al., 1982; Pacheco et al., 1993). McNally and Minster (1981) and Pacheco et al. (1993) have noted that, for the Central American subduction zone from Guatemala through Costa Rica, the historically observed rate of great earthquakes is substantially smaller than would be expected in an efficiently coupled subduction zone that accommodates ~80 mm/yr relative plate motion.

Interface-Thrust Earthquakes

Considering earthquakes occurring in 1978–April 2001 in Central America beneath or offshore of Guatemala, El Salvador, and Nicaragua, of depth less than 100 km and epicenter between 50 km and 170 km from the trench axis, we find that "typical interface-thrust focal mechanism" earthquakes accounted for ~60% of earthquakes of M_W 6 or greater (Fig. 7A). Only one of the typical thrust-mechanism earthquakes since 1978 had a magnitude approaching M ~8, consistent with the historical trend noted by McNally and Minster (1981). Even on a generally weakly coupled interface thrust, however, it is plausible that there would exist some asperities that could accumulate strain sufficient to produce interface-thrust earthquakes in the M ~8 range, and it is also possible that there could be along-strike variations in seismic coupling along the Guatemala-Nicaragua section of the Central American subduction zone. In the historical record, a good candidate for a Salvadoran M ~8 interface-thrust earthquake would be the shock of September 7, 1915, for which Ambraseys and Adams (2001) calculated a hypocenter beneath western El Salvador and an M_S of 7.7. White et al. (this volume, Chapter 28) show that P-wave first motions recorded worldwide on the sparsely distributed and relatively insensitive seismographs of 1915 are consistent with the earthquake having a thrust-faulting mechanism. The shock was followed by a number of large aftershocks (Ambraseys and Adams, 2001), a pattern more typi-

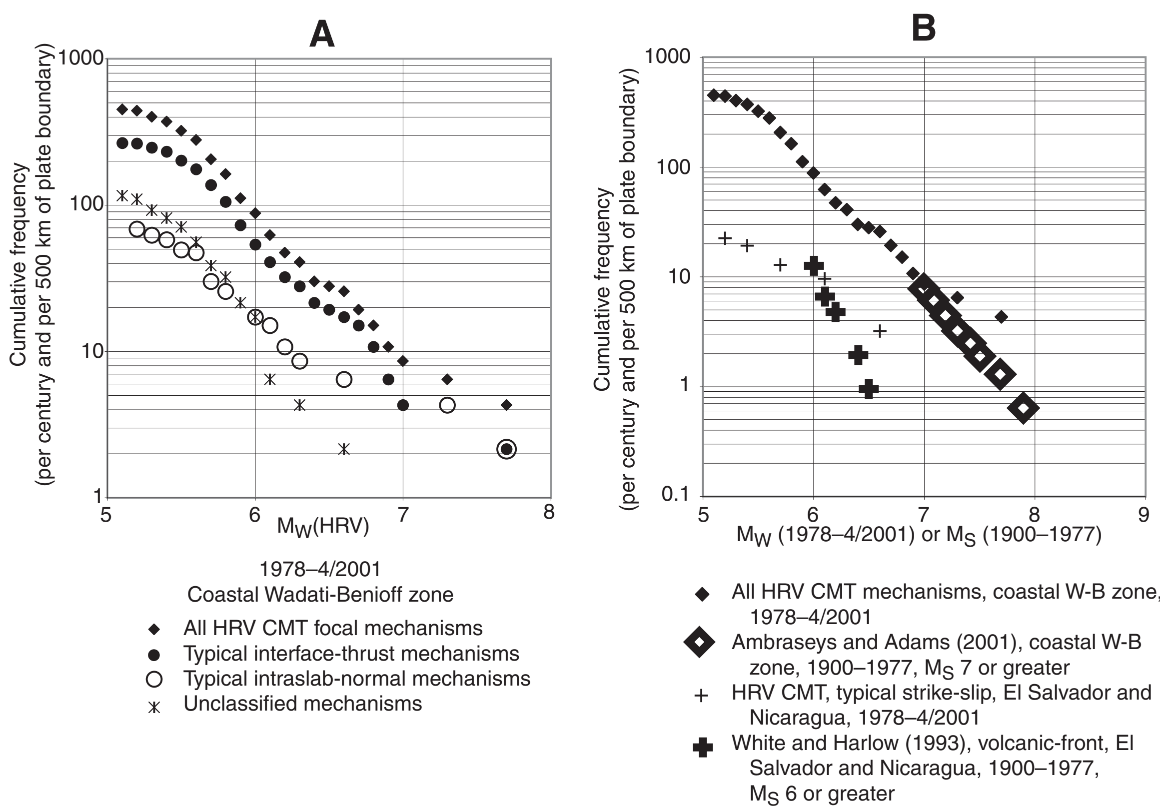

Figure 7. Cumulative magnitude-frequency plots for different sets of earthquakes distinguished by tectonic context or time of occurrence. For each set of earthquakes, the vertical axis shows the observed frequency of earthquakes with magnitudes equal to or greater than the magnitude shown on the horizontal axis. Symbols are plotted for a magnitude only if an earthquake of that magnitude is cataloged. A: Plots comparing the frequencies of different types of focal mechanism for the coastal Wadati-Benioff zone of Guatemala, El Salvador, and Nicaragua, 1978–April 2001. Events are those for which Harvard CMT focal mechanisms were determined. B: Plots comparing the frequencies for coastal Wadati-Benioff zone earthquakes and shallow-focus volcanic-front earthquakes, each determined with data from two different catalogs and time periods.

cal of interface-thrust earthquakes than slightly deeper intraplate normal earthquakes (see, however, discussion later in this paper on aftershocks to the intraplate normal earthquake of January 13, 2001). The 1915 earthquake produced Modified Mercalli intensities of VIII and higher at several locations in western El Salvador and adjacent parts of Guatemala (Ambraseys and Adams, 2001).

The largest (in terms of M_W) Central American interface-thrust earthquake in the past half-century was only lightly felt but produced a locally destructive tsunami. The Nicaraguan earthquake of 00:16 GMT, September 2, 1992, occurred as the result of slip at shallow depth on the Central American thrust interface (Satake, 1994). The earthquake nucleated in a seismic gap that had been identified by Harlow et al. (1981), but the earthquake rupture propagated well beyond the boundaries of the gap that had been identified. The local time of origin was 6:16 p.m., September 1. The shock was felt by only about half of the people in the coastal regions that were to be affected by the tsunami (Satake et al., 1993). The tsunami arrived on the coast at 8: 00 p.m., local time; runup heights exceeded 2 m along ~300 km of the Nicaraguan coastline, and the maximum runup exceeded 9 m (Satake et al., 1993; Baptista et al., 1993). An estimated 168 people died, and 13,000 were made homeless. Compared with tsunamis produced by other thrust-interface earthquakes worldwide, the tsunami produced by the Nicaraguan earthquake was disproportionately large for the earthquake's magnitude. Tsunami runup values were those typical of an M_S 8.0 earthquake, whereas the observed M_S was 7.2 (U.S. Geological Survey, 1992) and the observed M_W (HRV) was 7.7. The efficiency of tsunami generation for the Nicaraguan earthquake is attributed to the earthquake having produced unusually large displacements of the seafloor and having had an unusually long faulting duration for its M_S or M_W: These characteristics are in turn attributed to the earthquake having occurred at an unusually shallow depth, with much of the rupture process taking place in low-velocity sediments (Satake, 1994; Polet and Kanamori, 2000).

At the present state of knowledge, we have to consider that earthquakes similar to the 1992 Nicaraguan earthquake might occur elsewhere along the trenchward edge of the Central American interface thrust, and that the El Salvador coast might be subject to destructive tsunamis from this type of source. Development of procedures for issuing tsunami warnings along the Central American coast would have to consider that the light shaking produced on land from the 1992 Nicaraguan earthquake might be typical of this specific type of tsunami earthquake.

Cocos Intraplate Earthquakes

The destructive El Salvador earthquakes of June 19, 1982 [M_W(HRV) = 7.3] (Lara, 1983), and January 13, 2001 [M_W(HRV) = 7.7] (Paulson and Bommer, 2001; Lomnitz and Elizarrarás, 2001), were normal-faulting intraplate earthquakes within the Cocos plate (Bent and Evans, this volume, Chapter 29). "Typical intraslab-normal mechanism" earthquakes, as defined in this paper, account for ~20% percent of M_W 6 or greater Central American Wadati-Benioff zone earthquakes of Guatemala, El Salvador, and Nicaragua for which Harvard CMT mechanisms have been determined (Fig. 7A). Many of the "unclassified mechanisms" represented in Figures 4, 5, and 7A probably also correspond to Cocos intraplate earthquakes, with faulting styles or fault orientations differing from those of typical intraslab-normal earthquakes. Intraslab earthquakes immediately downdip of the seismogenically active interface thrust have been attributed to stresses arising from a variety of causes, such as extensional forces exerted by the deeper subducted slab or stresses caused by the distortion of the slab (e.g., Astiz et al., 1988; Bevis, 1988). We are not aware of a simple theoretical model of intraslab seismic-strain accumulation that would enable a prediction of the rate of such earthquakes that would be useful in seismic hazard studies.

The M_W 7.7 Cocos intraplate earthquake of January 13, 2001, was followed within one week by 28 aftershocks of m_b (USGS) 4.0 or greater, most of which were felt in San Salvador and the largest of which had M_W of 5.8. By comparison, the M_W 7.3 Cocos intraplate earthquake of June 19, 1982, was followed by only two aftershocks of m_b 4.0 or greater within one week of the main shock. In a global context, the 2001 aftershock sequence was exceptionally strong for an intraslab earthquake occurring well inland of the trench axis (Astiz et al., 1988). We have earlier cited the strong aftershock sequence of the earthquake of September 7, 1915, as evidence in favor of that earthquake being an interface-thrust shock rather than a Cocos intraplate shock. The aftershock sequence associated with the intraslab January 13, 2001, earthquake suggests that the presence of a significant aftershock sequence is not a conclusive basis for postulating that a historical coastal Wadati-Benioff zone earthquake occurred on the interface thrust.

Most seismic hazard associated with the Central American Wadati-Benioff zone is associated with shocks of depth less than 100 km occurring near the coast. Some large deeper shocks may also cause damage. Ambraseys and Adams (2001) favor a focal depth of ~150 km for the M_S 7.1 earthquake of May 21, 1932, which was damaging in southeastern El Salvador. Cocos intraplate normal-fault earthquakes occur near the axis of the Middle America trench, seaward of the coastal Wadati-Benioff zone (Figs. 4 and 5). Worldwide, uncommon great trench normal-fault earthquakes, such as the 1933 Sanriku, Japan, earthquake (Kanamori, 1971) and the Sumba, Indonesia, earthquake of 1977 (Spence, 1986; Lynnes and Lay, 1988), have produced devastating tsunamis on nearby coasts.

UPPER-CRUSTAL SEISMICITY IN THE CARIBBEAN PLATE

Historically, the most destructive earthquakes in El Salvador have been earthquakes occurring at upper-crustal focal depths (depth < 15 km) within ~10 km of the volcanic front in the overriding Caribbean plate (White, 1991; White and Harlow, 1993). More than half the length of the volcanic front in El Salvador experienced destructive upper-crustal earthquakes in the twentieth century—some communities more than once (Fig. 8). The largest of these earthquakes had magnitudes in the 6.5–6.7 range (White and Harlow, 1993; Ambraseys and Adams, 2001; U.S. Geological Survey, 2001), but smaller shocks have also been deadly. The M_W 5.7 San Salvador earthquake of October 10, 1986, for example, killed ~1500 people, left 100,000–150,000 homeless, and produced losses of about $1.5 billion (Bommer and Ledbetter, 1987; Olson, 1987). Earthquakes of m_b 4.5 or less, too small to be well recorded by the global network of seismographs, may cause significant alarm and minor damage (Ambraseys and Adams, 2001). Similar destructive upper-crustal earthquakes occur beneath the volcanic fronts of Guatemala, Nicaragua, and Costa Rica.

The majority of upper-crustal, volcanic-front earthquakes of Central America that have been well recorded teleseismically, and hence could have their focal mechanisms reliably determined by inversion of their seismic-wave radiation patterns, have had strike-slip focal mechanisms (White and Harlow, 1993). In Figures 4 and 5, we have plotted as "typical shallow strike-slip focal mechanism" earthquakes those for which both the P- and T-axes of the Harvard CMT solutions have plunges of less than 20° and for which the Harvard CMT depth is less than 50 km. The mechanisms of shallow strike-slip earthquakes of El Salvador and Nicaragua are consistent with the earthquakes occurring as slip on right-lateral faults that are approximately parallel to the main volcanic chain, or as slip on left-lateral faults that are approximately perpendicular to the trend of the volcanic chain.

For only a few Central American upper-crustal, volcanic-front earthquakes is there a strong basis from independent data (distribution of accurately located aftershocks, or observed fault displacement on the earth's surface) for establishing which of the two possible fault planes in the earthquake focal mechanism actually corresponds to the causative fault (White and Harlow, 1993). For additional earthquakes with two possible fault planes suggested by instrumental data, one can argue in favor of one of the two planes on the basis of observed seismic intensities, although intensities are influenced by many factors in addition to proximity to the fault plane. On the basis of its focal mechanism and aftershock distribution, the San Salvador earthquake of October 10, 1986 (Figs. 5, 8, and 9), likely occurred as the result of left-lateral faulting oriented at a high angle to the regional trend of the principal volcanic chain (Harlow et al., 1993). As we shall discuss below, the distribution of aftershock epicenters and damage from the El Salvador earthquake of February 13, 2001, seems most easily explained in terms of right-lateral slip on a fault approximately parallel to the volcanic chain. From the distribution of their intensities and focal mechanisms, the largest of the Jucuapa earthquakes of 1951 (1951a in Fig. 8) is inferred to have occurred on a left-lateral fault approximately perpendicular to the volcanic chain (Ambraseys et al., 2001), and the San Salvador earthquake of 1965 is inferred to have occurred on a right-lateral fault subparallel to the volcanic chain (White and Harlow, 1993).

Upper-crustal, volcanic-front earthquakes account for a relatively small fraction of the overall earthquake activity of the Central American subduction zone: The destructiveness of these earthquakes is a consequence of their occurring at shallow focal depth, commonly beneath heavily populated regions. In Figure 7B, we plot the average frequency of such earthquakes for El Salvador and Nicaragua, as determined by two independent catalogs spanning nonoverlapping time periods. The catalogs are the 1900–1977 portion of White and Harlow's (1993) catalog and the Harvard CMT catalog, shallow, strike-slip events only, for 1978 through April 2001. Both catalogs imply an average

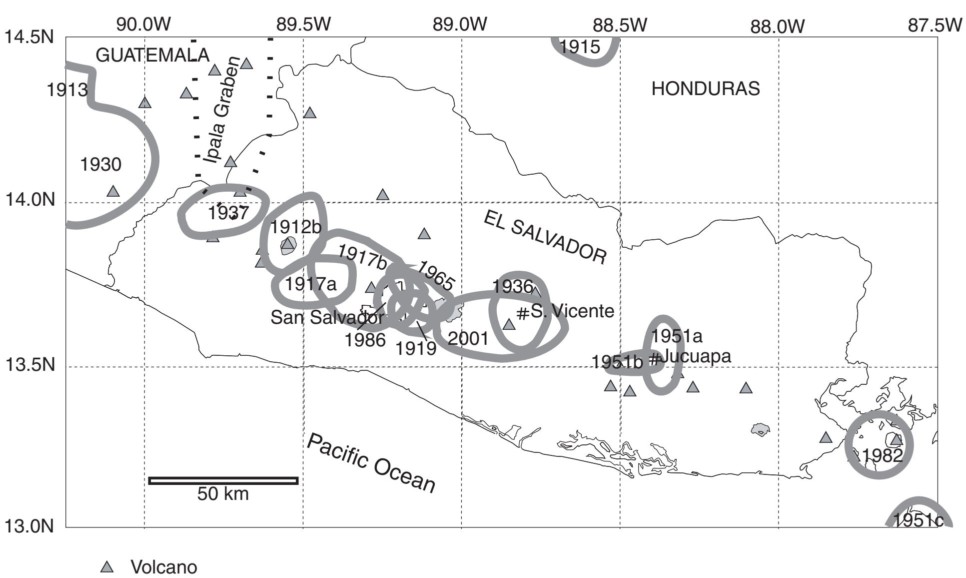

Figure 8. Regions of damage (Modified Mercalli intensities of VII or greater) from upper-crustal, volcanic-front earthquakes, twentieth century, El Salvador and vicinity. Pre-2001 isoseismals are from White and Harlow (1993) and are labeled as by White and Harlow. The Intensity VII isoseismal from the 13 February 2001 earthquake (Alvarenga et al., 2002) has been added. The boundaries of the Ipala graben are as defined by White (1991). Volcanoes are as in Figure 1.

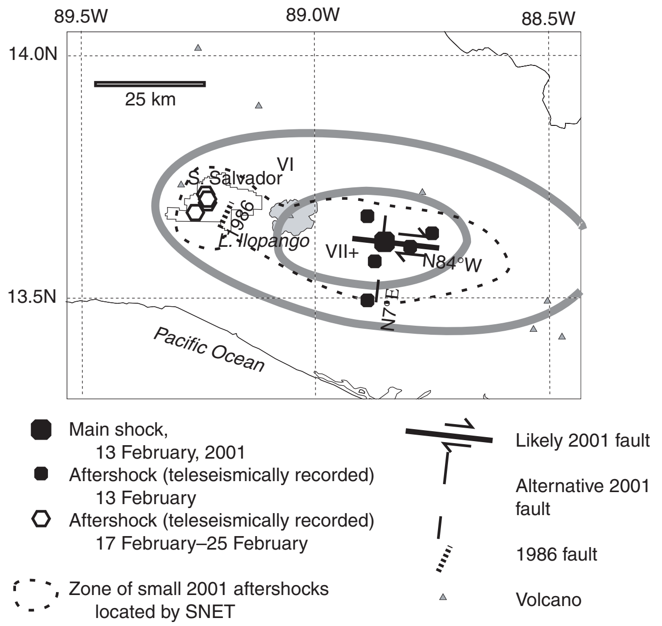

Figure 9. Sources of destructive El Salvador, volcanic-front earthquakes of October 1986 and February 2001. Location and trend of the fault that produced the earthquake of October 10, 1986, are from Harlow et al. (1993). "Likely 2001 fault" and "Alternative 2001 fault" correspond to possible fault planes of the main shock of February 13, 2001, that are implied by the Harvard CMT focal mechanism of the earthquake. They have strikes implied by the CMT mechanism and are positioned so as to pass through the main-shock epicenter and to come near to as many as possible of the teleseismically recorded aftershocks of February 13. Intensity VI and VII isoseismals are those of Alvarenga et al. (2002) for the main shock of February 13, 2001. Epicenters are those calculated by SNET using exclusively local data for earthquakes that occurred in February 2001 and that were large enough to also have been recorded at teleseismic distances. The "Zone of small 2001 aftershocks located by SNET" is generalized from maps published on the SNET Web site (http://www.snet.gob.sv/Geologia/Sismologia/sismi2001.htm) and represents the overall zone of seismicity that was activated by the February 13 main shock, from February 13 to the end of 2001.

frequency of ~10 earthquakes of magnitude greater than 6 per century per 500 km of length along the volcanic arc in El Salvador and Nicaragua. This is substantially less than the nearly 100 magnitude 6 or greater Wadati-Benioff zone earthquakes, per century per 500 km of arc-length, that is implied by the Harvard CMT catalog for 1978 through April 2001 and that is also consistent with the frequencies of large earthquakes cataloged by Ambraseys and Adams (2001) for 1900–1977 (Fig. 7B).

We note that upper-crustal, volcanic-front earthquakes of Guatemala and Costa Rica, although not represented in Figure 7B, also pose a significant hazard to the populations of those nations (White, 1991; White and Harlow, 1993). We represent earthquakes exclusively from El Salvador and Nicaragua in Figure 7B in order to estimate a regional frequency-of-occurrence of upper-crustal, volcanic-front earthquakes that is not contaminated by strike-slip events originating on the Caribbean/North American plate boundary in Guatemala, nor by strike-slip events that might be caused by stresses associated with the shallow subduction of the Cocos Ridge beneath Costa Rica. Shallow strike-slip earthquakes along the Caribbean/North American plate boundary in eastern Guatemala (Fig. 4), for example, are occurring on the faults of that transform boundary and may play a different role in regional plate tectonics than that played by the upper-crustal, volcanic-front earthquakes of El Salvador. The destructive Guatemala earthquake of February 4, 1976 (Plafker, 1976; Espinosa, 1976), occurred on the Caribbean/North American plate boundary and would not be classified as a volcanic-front earthquake within the Caribbean plate.

Tectonic Origin of Upper-Crustal Seismicity

In a broad sense, the upper-crustal volcanic-front seismicity probably reflects the northwestward motion of the forearc of the Caribbean plate, south and west of the volcanic chain, with respect to the Caribbean plate that lies to the north and east of the volcanic chain (Plafker, 1976; Harlow and White, 1985; White, 1991). The importance of motion of "forearc slivers" parallel to plate boundaries has been recognized in a number of subduction zone environments (Fitch, 1972; Jarrard, 1986). DeMets (2001) reports evidence that slip vectors of Central American interface-thrust earthquakes are systematically rotated with respect to a well-constrained independent estimate of the motion of the Cocos plate with respect to the core of the Caribbean plate; he interprets this rotation in terms of a Central American forearc sliver that is moving northwest with respect to the core of the Caribbean plate with a velocity of ~14 mm/yr (Fig. 2).

In detail, the geometry of the structures that accommodate the right-lateral motion of the Central American forearc with respect to the interior of the Caribbean plate has not been defined. The currently seismogenic faults are likely to be a subset of the faults that have defined the distribution of Quaternary volcanoes (Carr and Stoiber, 1977). Figure 6 shows mapped faults in El Salvador as compiled by Case and Holcombe (1980). Not all faults in Figure 6 are likely to be potentially seismogenic at present, however, and potentially seismogenic faults remain unmapped. White (1991) and White and Harlow (1993) have argued that the most important faults for accommodating the motion of the forearc sliver are right-lateral strike-slip faults that are parallel to the volcanic chain. La Femina et al. (2002), in contrast, emphasize the possible importance of bookshelf faulting, in which much seismogenic slip occurs on NNE-trending left-lateral faults that are within a broad WNW-trending right-lateral shear zone. Aseismic deformation may also play an important role in accommodating the relative motion of the Central American forearc sliver with respect to the interior of the Caribbean plate. In many other subduction zones, forearc slivers are separated from the interiors of overriding continental plates by geologically conspicuous regional strike-slip faults that are parallel to the trends of the arcs, but the Central American forearc sliver is not separated from the interior of the Caribbean plate by such a regionally prominent fault (Jarrard, 1986).

A better understanding of the configuration of seismogenic faults and of the role of aseismic deformation would help estimate the likelihood of future strong earthquakes in those areas of the volcanic chain in El Salvador (such as much of eastern El Salvador) that did not experience damaging upper-crustal earthquakes in the twentieth century (Fig. 8). Such knowledge would also help anticipate the maximum magnitude of upper-crustal earthquakes that are likely to occur at a given location. In general, the maximum size of earthquakes in a region will be determined by the lengths of fault segments that can rupture in a single earthquake. Considering Central America from Costa Rica through Guatemala, the maximum observed magnitude of upper-crustal, volcanic-front earthquakes is ~6.9 (White and Harlow, 1993). White and Harlow (1993) postulate that individual segments of a regional, arc-parallel, right-lateral shear zone are demarcated by the volcanic centers at their ends; the maximum sizes of earthquakes along a given section of the Central American volcanic arc would therefore be determined by the distance between the volcanic centers that bound that section of the arc. Under the model that motion of the forearc sliver is accommodated by bookshelf faulting (La Femina et al., 2002), the sizes of earthquakes would be determined by the width of the overall WNW-trending right-lateral shear zone within which the individual NNE-trending seismogenic faults are situated.

Earthquake of February 13, 2001

The El Salvador earthquake of February 13, 2001 (M_W = 6.6), and its aftershocks illustrate several important characteristics of the upper-crustal seismicity. The main shock was assigned a Harvard CMT focal depth of 15 km (U.S. Geological Survey, 2001), the default focal depth for shallow-focus shocks (Fig. 5), and locally recorded aftershocks were overwhelmingly in the uppermost 15 km of Earth's crust (http://www.snet.gob.sv/Geologia/Sismologia/sismi2001.htm). The earthquake directly or indirectly caused the deaths of over 300 people and injuries to more than 3000 (U.S. Geological Survey, 2001). The earthquake caused shaking of Modified Mercalli intensity VI or higher in an area of ~3000 km^2 (Fig. 9; Alvarenga et al., 2002). The moment magnitude (M_W) of the earthquake, together with the Wells and Coppersmith (1994) equations relating rupture length to moment magnitude for strike-slip earthquakes, suggests that the causative fault was ~20 to 30 km long. The Harvard CMT focal mechanism solution indicates that the causative fault plane was either a NNE-striking (N7°E) left-lateral strike-slip fault or a WNW-striking (N84°W) right-lateral strike-slip fault (http://www.seismology.harvard.edu/CMTsearch.html). The February 13 main shock triggered aftershocks over an ~75 km long, WNW-trending zone that includes the region of the main-shock epicenter (Fig. 9), but that is far larger than the zone of highest intensity and is far longer than the length of the main-shock rupture that would be suggested by the main shock's seismic moment. The overall 75 km long zone therefore likely reflects slip on a number of different faults that were activated by stress changes caused by the February 13 main shock. The WNW orientation of the overall aftershock zone supports the hypothesis that the February 13 earthquake occurred in a regional WNW-trending zone of right-lateral shear. The WNW elongation of the intensity VII isoseismal (Fig. 9) is most consistent with the main shock having occurred on a WNW-striking fault plane within the broader WNW-trending shear zone, and the WNW-trending plane is accordingly identified as the "likely fault" in Figure 9. The aftershock zone is, however, wide enough to leave open the possibility that the individual fault segment that produced the February 13 main shock may have been a NNE-trending left-lateral fault within the broader WNW-trending right-lateral shear zone. The NNE-trending nodal plane of the Harvard CMT focal mechanism is therefore indicated as the "alternative 2001 fault" in Figure 9.

The San Salvador Sequence of February 17–25

The western San Salvador earthquake sequence of February 17–25 may be cited as an example of a Central American upper-crustal earthquake sequence that consists of only small and moderate earthquakes but that causes great public concern. The sequence also illustrates the importance of local networks in locating such earthquakes. The sequence on February 17–25 occurred west-northwest of the source region of the February 13 earthquake. The source of the February 17–25 sequence lies on the trend of the inferred N84°W-striking fault of the February 13 main shock (Fig. 9), but it is well outside the region of strongest shaking in the February 13 main shock. The February 17–25 sequence therefore must have occurred on a different fault or fault segment than the fault segment that ruptured in the February 13 main shock. In February 17–25, 65 of these shocks were felt locally (http://www.snet.gob.sv/Geologia/Sismologia/sismi2001.htm), and several of the largest were recorded teleseismically. The first and largest earthquake, that of February 17, 20:25 GMT, had a body-wave magnitude, m_b (USGS), of 4.1 and a local magnitude of 5.1. One person was reported killed by the earthquake, and three injured (U.S. Geological Survey, 2001). The earthquakes were too small to be well recorded by the global seismographic network. Locations calculated for several of the shocks by the U.S. Geological Survey National Earthquake Information Center, using only data obtained at stations far from El Salvador, were mislocated by more than 50 km (Fig. 10). If the earthquakes had been slightly smaller, or the global seismographic network slightly less sensitive, any listing of the earthquakes in internationally compiled earthquake catalogs would have depended entirely on data from local seismographs.

Clustering of Upper-Crustal Seismicity

Upper-crustal, volcanic-front earthquakes of Central America are commonly preceded by foreshocks or occur in clusters of several damaging events closely spaced in time (White and Harlow, 1993). For example, the largest San Vicente earthquake of late 1936, that of December 20, was preceded by several weeks of smaller earthquakes from approximately the same source (Levin, 1940), and the San Salvador earthquake of May 3, 1965, was immediately preceded by foreshocks and occurred in a period of rather high activity that had begun with an earthquake swarm three months earlier (Lomnitz and Schulz, 1966). White and Harlow (1993, p. 1137) describe the clustering tendency of damaging upper-crustal, volcanic-front earthquakes as follows: "When a destructive earthquake occurs along the volcanic front after several years of calm, 43% of the time it is followed within one month by another destructive earthquake less than 60 km away." The strong 1917 earthquakes, centered just west of San Salvador (Fig. 8), occurred within 35 minutes of each other (White and Harlow, 1993), and three earthquakes of M ~6 occurred near Jucuapa on May 6–7, 1951 (Meyer-Abich, 1952; Ambraseys et al., 2001). The earthquake of February 13, 2001, was not preceded by an obvious foreshock sequence. The sequence of February 17–25, 2001, may correspond to the type of triggered seismicity that has been represented at other times and locations of the volcanic chain by temporal clusters of damaging earthquakes, although its largest event was much smaller than the main shock of February 13.

In addition to their tendency to be temporally clustered, upper-crustal volcanic-front earthquakes have over a period of decades tended to concentrate in spatial clusters. The epicentral region of the February 13, 2001, earthquake has a history of earlier destructive, shallow earthquakes. The San Vicente earthquake of December 20, 1936, M ~6.1 (White and Harlow, 1993), for example, was destructive in several of the towns and villages that were heavily damaged in the 2001 earthquake (Levin, 1940), and the region had experienced damaging earthquakes in each of the eighteenth and nineteenth centuries (Harlow et al., 1993). San Salvador has repeatedly experienced damaging earthquakes, with at least nine earthquakes having produced intensity VII effects within the city since 1700 (Harlow et al., 1993). In addition to the shocks of 1951, the region of Jucuapa had experienced two damaging earthquakes in the nineteenth century (Meyer-Abich, 1952; Ambraseys et al., 2001).

Volcano Seismicity

Most Central American earthquakes of $M_S > 5$ are not obviously associated with volcanism (White and Harlow, 1993). Their concentration near the volcanoes of El Salvador suggests that their locations may be defined in part by magmatic processes or that both seismic and magmatic phenomena may be localized by a fundamental tectonic process, but most of the earthquakes don't have obvious links to specific volcanic eruptions. An example of an exceptional, strong, upper-crustal El Salvador earthquake that was directly associated with an eruption is the M_S 6.4 earthquake that occurred on June 8, 1917, and was followed 30 minutes later by the eruption of Volcán San Salvador. The earthquake damage was centered on the northwest flank of the volcano.

Although most damaging Central American earthquakes are not associated with volcanism, volcanic eruptions are invariably preceded and accompanied by seismic activity. A dramatic local example was the eruption of Volcán Pacaya in Guatemala during 1981 and 1982 that was preceded by at least 100,000 small to

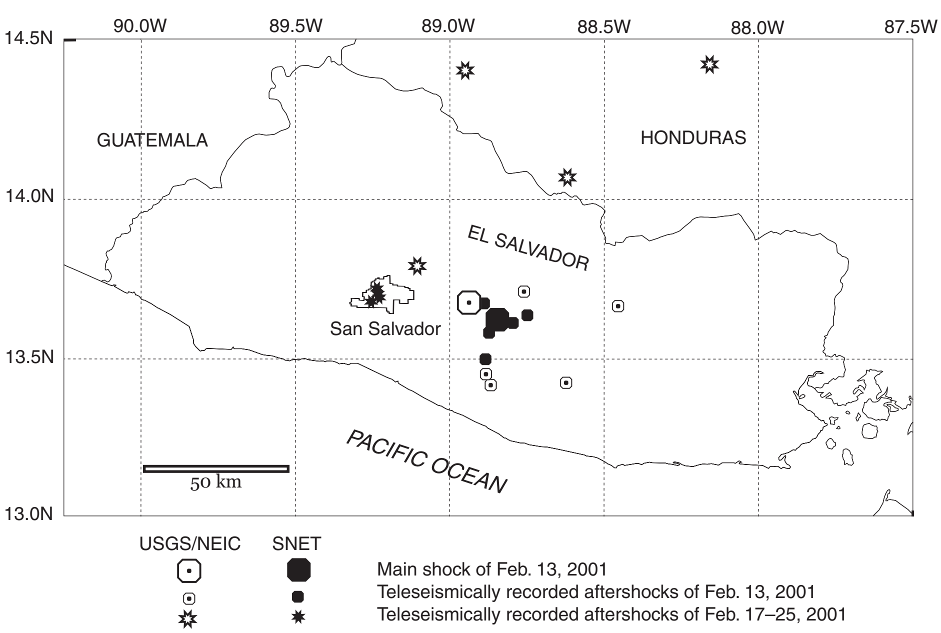

Figure 10. Comparison of epicenters calculated by the USGS/NEIC and SNET for teleseismically recorded, volcanic-front earthquakes of February 2001, illustrating the importance of locally recorded seismographic data for the accurate location of small and moderate El Salvador earthquakes. The SNET epicenters were determined with data obtained from seismographs recording in El Salvador and are accurate to within several kilometers. With the exception of the epicenter of the main shock, however, all USGS/NEIC epicenters were determined with data recorded from a limited range of azimuths by stations situated outside of El Salvador, and most are badly mislocated.

moderate earthquakes along faults southeast of the volcano (White et al., 1980). The largest of those earthquakes was the M_S 5 shock of October 9, 1979, that destroyed more than 150 houses. White and Power (2001) have shown that such earthquakes "have often been observed beneath distal regions of volcanoes weeks to years prior to eruptions." Such events, called volcano-tectonic earthquakes, are high frequency and broadband and appear to result from brittle failure along faults, resulting from magma-induced changes in stress or fault strength. However, volcano-tectonic events are often distinguishable from tectonic seismicity by their occurrence in spatial and temporal clusters, and by a tendency for the rate of earthquake activity and the maximum magnitudes of earthquakes within a cluster to increase with time. White and Power (2001) show that volcano-tectonic seismicity can sometimes be used to forecast the type and size of an eruption.

In addition to volcano-tectonic earthquakes, the other principal type of seismicity generally associated with magmatic and geothermal processes is long-period volcano seismicity. Long-period volcano seismicity is narrowband and has a dominant frequency less than 5 Hz. Long-period volcano seismicity occurs both as individual events and as extended periods of tremor lasting minutes to many days. Much of this type of seismicity is inferred to result from degassing of magma and/or boiling of the hydrothermal system (Chouet et al., 1994). Like volcano-tectonic seismicity, long-period volcano seismicity can also sometimes be used to forecast the type and size of an eruption (Chouet et al., 1994; Chouet, 1996). Such long-period volcano seismicity was used to successfully forecast the cataclysmic 1991 eruption of Mount Pinatubo in the Philippines (Harlow et al., 1996). Seismic signals are also caused by various other surface or near-surface volcanic phenomena, such as explosive eruptions, pyroclastic flows, lahars, dome collapse, and lava flows: These signals are generally of the long-period type.

Caribbean Plate Northeast of the Volcanic Front

The northwest corner of the Caribbean plate inland of the volcanic front, encompassing northern El Salvador, southeastern Guatemala, and western Honduras, has in recent decades been characterized by a moderate level of shallow-focus earthquake activity (Figs. 1 and 4), but the region has a history of occasional larger shocks. White et al. (this volume, Chapter 28) document the occurrence of a large (M ~7.5) earthquake in the northern Ipala graben of Guatemala in 1765. The Ipala graben extends into northwestern El Salvador (Fig. 8). The Honduras earthquake of 1915, whose intensity VII isoseismal extends into the region covered by Figure 8, had a magnitude (M_S) of 6.4 (White and Harlow, 1993). The relatively few Honduras events for which Harvard CMT mechanisms have been determined (Fig. 4) were characterized by predominantly normal-faulting focal mechanisms and have T-axes oriented E-W or ESE-WNW, consistent with geologic evidence for E-W extension in Honduras south of the North American/Caribbean plate boundary (Plafker, 1976). As noted in the section entitled "Sources of Data on El Salvador Earthquakes," calculated focal depths at the north-northeast end of the profile in Figure 3 are uncertain by several tens of kilometers. We are not aware of strong independent evidence that shocks within the northwest corner of the Caribbean plate have focal depths approaching 50 km, as is suggested by Figure 3. Earthquakes this deep would be unusual, though not unprecedented (Wong et al., 1984), in a region of extensional tectonism. The relatively few Harvard CMT solutions in this region (Fig. 4) have hypocentroid depths less than 25 km.

CONCLUSIONS

This paper is being written at a time when the seismographic network run by El Salvador's Servicio Nacional de Estudios Territoriales (SNET) is being significantly upgraded. The existence of a high-quality network of local seismographs, together with a network staff of local scientists who collectively have acquired experience in the interpretation of seismographic data from earthquakes in different parts of El Salvador, will greatly facilitate addressing earthquake hazard issues. Solutions of most problems will also benefit from the systematic interchange of data between SNET and other Central American seismological organizations (Lindholm et al., this volume, Chapter 26), and from the systematic interchange of data between SNET and international agencies such as the USGS/NEIC and the ISC (data shared with the USGS/NEIC are automatically forwarded to the ISC). Let us consider, for example, seismological data that would help address issues identified in this paper.

1. Removal of bias in teleseismically determined epicenters and focal depths that are cataloged by international agencies (see section entitled "Sources of Data on El Salvador Earthquakes") could be accomplished by means of calibration events accurately located with local data.

2. Calibration of the teleseismic m_b scale for use in seismic hazard studies ("Sources of Data on El Salvador Earthquakes") could be accomplished by means of local magnitudes of small and moderate teleseismically recorded shocks.

3. Accurate calculation of earthquake epicenters on the borders of El Salvador ("Sources of Data on El Salvador Earthquakes") requires exchange of data with agencies in other Central American countries.

4. Distinguishing between interface-thrust earthquakes and Cocos intraplate earthquakes (see section entitled "The Central American Wadati-Benioff Zone") on the basis of accurate hypocenters would be facilitated by the transmittal of locally recorded arrival times to international agencies.

5. Distinguishing between interface-thrust earthquakes and Cocos intraplate earthquakes ("The Central American Wadati-Benioff Zone") on the basis of their focal mechanism, for earthquakes too small to have focal mechanisms reliably determined by the Harvard CMT methodology, may be accomplished with on-scale, broadband, local recording of these earthquakes.

6. An understanding of the long-term seismic potential of the interface thrust between the Cocos and Caribbean plates

("The Central American Wadati-Benioff Zone") may require mapping of asperities on the interface using locally recorded small earthquakes.

7. Issuing tsunami warnings for offshore Central American earthquakes ("The Central American Wadati-Benioff Zone") requires exchange of data with agencies in other Central American countries and exchange of data with Pacific-wide international tsunami warning agencies.

8. Seismographic definition of seismogenic faults in and near the volcanic front, for use together with geologic observations in seismic zoning (see section entitled "Upper-Crustal Seismicity in the Caribbean Plate"), requires precisely determined, locally recorded hypocenters.

9. Ensuring the reliable location and cataloging of potentially damaging earthquakes with M < 5.5 by international seismological agencies ("Upper-Crustal Seismicity in the Caribbean Plate") requires the sharing of local data with these agencies.

10. Evaluating the temporal and spatial clustering tendencies of upper-crustal, volcanic-front earthquakes ("Upper-Crustal Seismicity in the Caribbean Plate"), and developing ways to usefully characterize this clustering for the benefit of emergency management agencies and the El Salvadoran community, requires local seismological monitoring that is sustained over a long time.

11. Recording of volcano-tectonic and long-period seismicity for use in forecasting volcanic eruptions ("Upper-Crustal Seismicity in the Caribbean Plate") requires the collection of seismological data close to potentially active volcanoes.

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