Abstract

I relocated 1073 earthquakes that occurred in the Caribbean region (5°N–25°N 50°W–90°W) between January 1913 and December 1962. *P* and *S* arrival times published by the International Seismological Summary (ISS) for stations worldwide were used in conjunction with a generalized linear inverse routine to perform the relocations. With about ten exceptions, the relocated earthquakes fall within the defined plate boundary zones of the Caribbean, Nazca, Cocos, and North and South American Plates. These boundaries are much more clearly defined by the post-relocation events, however. Plate-boundary seismicity for this time period is most intense along the Middle Americas subduction zone (Guatemala to Coast Rica segment), the Panama Fracture zone, and in eastern Hispaniola and the Puerto Rico Trench region. A somewhat less well-defined tectonic boundary is marked by shallow earthquakes south of Panama. The Colombian and Venezuelan Andes are characterized by sporadic seismicity south of 10°N. Although the Bocono Fault is clearly seismically active during this period, no events relocate to the vicinity of the Oca-Ancón fault system, Venezuela between 68°W and 65°W is devoid of all teleseismically detectable seismicity. The Lesser Antillean subduction zone is defined by isolated intermediate depth earthquakes. This area is marked by an almost complete absence of shallow seismicity from 16°N to 13°N, at the magnitude level of this catalogue (probably $m_b \geq 5.0$). The North America-Caribbean plate boundary is moderately well-defined by shallow events. Potentially intraplate earthquakes include five shallow events south of Jamaica near the Pedro Bank, two events at a location east of the Bahamas, two separate events near the Yucatan coast, and two earthquakes on the Nicaragua Rise near the Hess Escarpment.

In both the Middle Americas subduction zone and the Puerto Rico Trench region, intermediate depth seismicity is sufficient to define subducted lithosphere at an accuracy comparable to that of modern teleseismically located catalogues. The deepest earthquakes in the region occur in the Middle Americas subduction zone and attain depths in excess of 275 km. However, assuming perfect control on depth, only 15 relocated events were deeper than 200 km, all in Central America.

Introduction

Accurate earthquake locations spanning the longest possible time period are essential for correct assessment of seismic hazard. Precise determination of patterns of historical (pre-1963) seismicity is especially important in the Caribbean region where plate boundaries are geologically complex and plate velocities are predominantly slow. Thus, recurrence times for large magnitude potentially destructive earthquakes are long, and hazard assessments based on short term catalogues may be very misleading. The slow velocity of the Caribbean plate relative to its neighbors and the consequent relatively low level of seismicity along its plate boundaries has sometimes obscured definite identification of seismically active tectonic regions within the plate boundary zones. In order to rectify this situation insofar as is quantitatively possible using instrumentally recorded earthquakes, I undertook this study to relocate 1073 historic era events that occurred in the Caribbean Basin (5°N–25°N, 50°W–90°W) between January 1913 and December 1962, and in the region just to the west of the principal study region (90°W to 93°W) during the period from January 1913 to December 1945. Published studies of Caribbean seismicity that complement and supplement this work include *Sykes and Ewing*, [1965], *Molnar and Sykes*, [1969]; *Dewey*, [1972]; *Tomblin*, [1975]; *Pennington*, [1981]; *Burbach et al.*, [1984]; *LeFevre and McNally*, [1985], *Adamek et al.*, [1988]; *McCann and Pennington*, [1990]; *Protti et al.*, [1994], and *Malave and Suarez*, [1995].

Although the historic seismicity data set is of limited value, the study of instrumentally recorded historic earthquakes has several intrinsic advantages. Because of the sparseness of the global seismometer network prior to the installation of the WWSSN (World Wide Standardized Seismic Network) in 1963, and given the highly variable sensitivity of the seismometers themselves during this time, the events that are clearly recorded at a sufficient number of stations to allow location teleseismically are necessarily events of fairly large magnitude. These are potentially the most damaging earthquakes and are therefore crucial to studies of seismic hazard. Large magnitude events are also likely to best represent long-term tectonic displacements, and thus these earthquakes will also be important events for neotectonic interpretation. The earthquakes I have studied were already located with some precision by the ISS (International Seismological Summary), who tabulated arrival times as well as locations. Thus, relocation is facilitated by the fact that the available data were contemporaneously gathered, collated into events, and their quality was assessed indirectly through the ISS location procedure. These procedures were carried out by a single agency who thereby produced a fairly homogeneous data set over many decades. However, the ISS locations were based only on *P* first arrivals, included only a rudimentary assessment of event depths in most cases, and suffer in general from a lack of precision in epicentral parameters. Thus, the data set as it existed was of limited value for tectonic and seismic hazard analyses. This work is an attempt to render these data more useful for such purposes.

Caribbean Region Motions, Plate Boundaries, and Known Active Faults

Results of global and regional plate motions studies [*Jordan*, 1975, Minster and Jordan 1978, *Sykes et al.*, 1982; *Stein et al.* 1988, *DeMets et al.* 1990, 1994, *Deng and Sykes* 1995] generally agree that the motion of the Caribbean plate relative to its two large neighbors, North and South America, is eastward at a rate between 1 and 2.5 cm/yr (see also *Rosencrantz et al.*, [1988]). Thus, far-field relative motions between stable North America (NA) and the undeforming Caribbean (Ca) interior is primarily sinistral [*Lundgren and Russo*, 1996a], and far-field motions between South America (SA) and Ca is largely dextral [*Lundgren and Russo*, 1996b]. Debate about components of far-field motion normal to the generally east-west trending (Fig. 1) NA-Ca and SA-Ca plate boundary zones continues; however space geodetic techniques (e.g., GPS) are beginning to yield results that will likely resolve this issue [*Farina et al.*, 1995, *Lundgren and Russo*, 1996a]. Within the wide NA-Ca plate boundary zone, at least one microplate, called the Gonave microplate (Fig. 1c) by *Rosencrantz and Mann*, [1991] appears to be present. The Gonave microplate is bounded on the west by the Cayman Spreading Center and its eastern boundary probably lies within Hispaniola. To the north and south it is bounded by strike-slip faults (see further discussion below). Relative motions between the Gonave platelet and NA are near 1.7 cm/yr sinistral slip; along its southern boundary with the Ca plate rates are much less, around 2-3 mm/yr [*Lundgren and Russo*, 1996a].

Motions between the Caribbean plate and adjacent portions of the Cocos (Co) plate (Fig. 1) are currently better defined; Co-Ca convergence rates reach 8-10 cm/yr in a generally northeasterly direction [*Lundgren and Russo*, 1996b]. This convergence is taken up, with attendant deformation in a fairly wide plate boundary zone, at the Middle Americas subduction zone. Relative motions are somewhat less rapid (6-7 cm/yr) along the north-striking Panama Fracture Zone, the boundary between the Cocos and the northernmost Nazca plates, and between Co and a small microplate (Fig. 1) whose existence was first proposed by *Hey*, [1977], and which was named the Coiba microplate by *Adamek et al.* [1988]. The Coiba microplate appears to move with largely sinistral strike-slip relative to both the Nazca plate to its south and the Panama Block to its north, although there is also evidence for compression along the latter boundary [*Silver et al.*, 1990; *Mann and Corrigan*, 1990, *Russo and Villaseñor*, 1996]. Thus, eastward motion relative to SA increases from nearly 4 cm/yr at the Panama Block to 6 cm/yr in the Coiba microplate, to 7 cm/yr in the northernmost Nazca plate [*Freymueller et al.*, 1993, *Lundgren and Russo*, 1996b]. Along the Panama Block's northern boundary, where it abuts the Caribbean plate, motions are sinistral strike-slip and compression at relative rates around 2.5 cm/yr. Within the Northern Andes region, motions of continuum and on faults derived from finite element modeling based on far-field and geodetic data [*Lundgren and Russo*, 1996b] are between 1 and 2 cm/yr. Much of this region (Fig. 1) appears to be moving relative to SA. Finally, in the eastern Caribbean convergence at the Lesser Antilles arc appears to be relatively slow at around 2 cm/yr.

Major active faults within the plate boundary zones surrounding the Ca plate include (Fig. 1): the Swan Islands [*Mann et al.*, 1991], Oriente [*Calais and Mercier de Lépinay*, 1991] and Walton faults [*Rosencrantz and Mann*, 1991] in the NA-Ca plate boundary west of Hispaniola, the Septentrional fault [*Mann et al.*, 1984; *Prentice et al.*, 1992, *Russo and Villaseñor*, 1995, 1996a], the Mona Passage graben [*Masson and Scanlon*, 1991], the 19° fault [*Speed and Larue*, 1991], and the Anegada Passage fault system [*Frankel et al.*, 1980; *Jany et al.*, 1990] in and east of Hispaniola; the Lesser Antilles trench, the Central Range fault system of Trinidad [*Smith*, 1993]; the El Pilar fault system of NE Venezuela [*Russo et al.*, 1993]; the San Sebastian fault along the north coast of Venezuela [*Suárez and Nábělek*, 1990]; the Bocono fault of the Venezuelan Merida Andes [*Schubert*, 1982]; the East Andes Frontal fault zone [*Lundgren and Russo*, 1996b], the Romeral fault system [*Ego et al.*, 1995], and the Atrato-Sinu fault [*Lundgren and Russo*, 1996b] in the Colombian Andes; the Oca-Ancón fault system of northern Colombia and Venezuela [*Audemard*, 1996]; the Santa Marta fault [*Kellogg and Bonini*, 1982] of northern Colombia; the Panama Fracture Zone [*Adamek et al.*, 1988], the Nicaraguan Depression fault system

[*White*, 1991], and the Polochic-Motagua fault system that is the onland NA-Ca plate boundary fault zone [*Langer and Bollinger*, 1979, *Guzman-Speziale et al.*, 1989]. Regions of active deformation associated with active faulting include the North and South Panama Deformed belts [*Pennington*, 1981; *Adamek et al.*, 1988, *Kolarsky and Mann*, 1995; *MacKay and Moore*, 1990, *Mann and Corrigan*, 1990, *Moore and Sender*, 1995], Muñoz, 1988; *Silver et al.*, 1990, *Westbrook et al.*, 1995], the East Panama-Colombia collision zone [*Mann and Kolarsky*, 1995], the Panama-Costa Rica deformation zone [*Montero and Dewey*, 1982, *Fisher et al.*, 1994], the Merida Andes [*De Toni and Kellogg* 1993], the Venezuelan fold and thrust belt [*Speed* 1985 *Russo and Speed*, 1992]; the Muertos Trough subduction zone of SE Hispaniola [*Ladd and Watkins*, 1978, *Byrne et al.*, 1985], and the Lesser Antilles accretionary prism complex [*Speed et al.*, 1984, 1989]. Motion senses and, in some cases, estimates of relative motion rates on the faults and within the deforming belts mentioned above are shown in Fig. 1c, derived primarily from *Lundgren and Russo*, [1996a,b]

The relative motions between the the Caribbean region plates, microplates, and deforming regions, and in particular the rates of such motion bear directly on the subject of this paper Caribbean region seismicity before the advent of the World Wide Standardized Seismic Network (WWSSN) in 1963 This is true because the rates of relative plate motion are a strong factor in determining the frequency and intensity of seismicity in associated plate boundary zones. The generally slow NA-Ca and SA-Ca relative motion rates results in less frequent seismicity and longer recurrence times than, for example, along the Middle Americas subduction zone, but not necessarily to less hazardous large magnitude earthquakes. The latter factor is important because a reliable estimation of the earthquake cycle in these plate boundary zones [*Dixon*, 1993, *Lundgren et al*, 1993]. requires the longest possible reliable catalogue of earthquakes

Method

The relocations are a significant advance over the ISS locations in that the data set for each event is expanded by including S arrival times in the location, and I solve for event depth simultaneously Relocations were obtained via a generalized linear inverse code [*Wysession et al*, 1991; *Russo et al.*, 1992]. The code minimizes, in a least-squares sense, the time residual between observed and calculated travel times from the projected earthquake hypocenter to the recording stations. The method is iterative and involves updating the model parameters to improve the fit to the data. Thus, a data vector, **d**, is related to a model vector, **m**,

$$\mathbf{d} = \mathbf{A}\mathbf{m} \tag{1}$$

where the data vector, **d**, is a list of arrival times of P and S waves at the available stations. The model vector, **m**, is composed of the model parameters to be solved for, namely the event focus (latitude, longitude, depth) and origin time. The matrix **A** incorporates information of the Earth s velocity structure, in this case in the form of derivatives of travel time with respect to the four model parameters, interpolated from the Jeffreys-Bullen Earth model tables (1958) These derivatives are estimated using the standard first order Taylor's series expansion of the dependencies of travel time on the model parameters. Derivatives for S-wave travel times are downweighted assuming a Poisson's ratio of 0.25 to account for differences in P and S intrinsic velocities The arrival time is the sum of the (unknown) origin time, t, and the travel time through the Earth, $\mathbf{T}(x,x_i)$ The travel time is a function of the earthquake focus, x, and the station locations, x_i The arrival at the ith station is,

$$d_i = \mathbf{T}(x,x_i) + t \tag{2}$$

We guess a reasonable solution for the hypocenter and origin time (usually based on the P arrival time at the station nearest the ISS location) to forward calculate the four model parameters We then minimize the difference between calculated and observed travel times resulting from the guessed solution and update the model parameters via generalized linear inversion:

$$m_{i_{calc}} - m_{i_{obs}} = (\mathbf{G}^{\mathrm{T}}\mathbf{G})^{-1}\mathbf{G}^{\mathrm{T}}\, d_{i_{calc}} - d_{i_{obs}} \tag{3}$$

and matrix **G** relating changes in the model and data vectors is:

$$\mathrm{G}_{ij} = \left[\frac{\partial d_i}{\partial m_j}\right] \tag{4}$$

evaluated at the model parameters for the iteration step. For any relocation, depth can be constrained to a given value, which is useful in case the data do not control depth (often the case for shallow focus events). All calculated travel times are corrected for Earth ellipticity and station elevation. Data for travel paths that include the core are not used and are automatically suppressed if updated event locations move station-event distances to greater than 102°. However, no other formal procedure is included for automatically distinguishing and suppressing spurious arrivals (e.g., *SKS* phases at cross-over distances mistakenly identified as *S* by station operators or ISS personnel; such phases are usually identifiable by their large and systematic residuals and can be suppressed manually during inversion). At any point in the inversion procedure, arrival times for any station can be suppressed, if necessary.

For each iteration, individual station travel time residuals and data importances are calculated, as is the total standard deviation, in seconds, of the data. In general, I considered individual station residuals less than or equal to 5 s to be acceptable, although I accepted larger residuals in certain cases in the early data set (e.g., 1913-1920). Larger residuals were considered suspect for reasons having to do with one or a combination of along-path departures from the JB radial Earth model, station timing errors, incorrect phase identification, or incorrect association of arrival times with the event in question (i.e., two events in different locations arriving at subgroups of stations in the same time window). The data importances, calculated from the matrix of model parameter-travel time dependency partial derivatives, were used primarily to identify which stations in the overdetermining data set were controlling the location and which were relatively redundant. In general, I used the residual standard deviation, the individual station residuals, and the data importances as guides in identifying and suppressing grossly bad arrival times.

For events with depths unconstrained by the available data, I systematically fixed the event depth, varying it between typically, 50 km and surface focus, and determined corresponding locations and residuals for a set group of stations. I compared the results to find the approximate minimum in the standard deviation of the residuals, if such a minimum was obvious, and chose the corresponding event hypocenter and origin time as the best result for entry into the final data set. In cases where the data set had no clear minimum standard deviation, I fixed the depth at 10 km and used the corresponding hypocenter and origin time; fixed-depth relocations appear in Table 1 with 'depth code' = 9. For all events for which available data did not initially constrain depth, I re-assessed the depth constraint after the relocation using the subset of consistent arrival times remaining after suppression of obviously inconsistent arrival times. These events appear in the table with depth code 2. Although I have not determined a statistical estimate of the accuracy of the relocated hypocenters explicitly, I consider that they are probably accurate to ±10 km in epicenter and ±20 km in depth, similar to error estimates in modern routinely located seismicity catalogues. Variability in accuracy can be gauged qualitatively by comparison of standard deviations reported in Table 1.

Because I used the ISS data set and not original seismograms, it was not possible to determine other important parameters such as event magnitudes or focal mechanisms for these events. I refer readers interested in event magnitudes to studies by *Gutenberg and Richter*, 1965, *Sykes and Ewing* 1965; *Abe*, 1981; *Pacheco and Sykes*, 1992; and *Ambraseys and Adams*, 1996). Readers interested in focal mechanisms of some events during the study period are referred to work by *Molnar and Sykes* [1969], *Camacho* [1991], *Russo et al.* [1992], *Doser and VanDusen* [1996a,b] and *Russo and Villaseñor* [1995; 1996b].

Results and Discussion

Results of the relocation procedures are shown in Figures 2 and 3 and detailed in Table 1. A large majority of the relocated events lie within the wide Caribbean plate boundary zones, in keeping with what is known about more recent (1963-1996) seismicity [*McCann and Pennington*, 1990]. Likewise, a large majority of the events (858 vs. 215) were shallow focus ($h < 75$ km), and the events that were deeper than 75 km almost all occurred within the Middle Americas, Antillean, or South American subduction zones.

Within the plate boundary zones surrounding the Caribbean plate proper, certain faults are reasonably well defined by the relocated seismicity (Fig. 2a). These include the Oriente and Swan Islands Faults in the Cayman Trough, and the Cayman Spreading Center itself. The Panama Fracture Zone, forming the eastern boundary of the Cocos plate, is also very well defined. The northernmost Colombia

Trench, forming the eastern limits of the Coiba microplate [*Adamek et al.*, 1988; *Russo and Villaseñor*, 1996b] and the northernmost Nazca plate is also visible in the seismicity pattern. Shallow (h < 75 km) seismicity associated with the principal subduction zones of the region is somewhat more heterogeneous. Thus, although the Middle Americas subduction zone was quite active seismically during 1913-1962, as was the Antillean subduction zone from eastern Hispaniola to Guadeloupe, the Lesser Antilles south of Guadeloupe were largely seismically inactive at the resolving magnitude level of the relocated events. The latter relative inactivity of the southern Lesser Antilles is a long-standing feature of the seismicity in this region [*Tomblin*, 1975; *Wadge and Shepherd*, 1984, *Russo et al.*, 1992; 1993]. Shallow seismicity is also present in Colombia and western Venezuela, where there is a concentration of events in the Venezuelan Merida Andes and the northeasternmost Colombian Cordillera Oriental, and scattered seismicity elsewhere. Panama is surrounded by a rough halo of shallow events, but its interior regions are devoid of seismicity. Finally, note that the northern Colombia Basin particularly the region of rough bathymetry between the Nicaragua Rise and the Hess Escarpment includes a fair number of shallow events.

Seismicity deeper than 75 km (Fig. 2b) is primarily restricted to the above mentioned subduction zones, with a few exceptions. Most notable regions of frequent intermediate depth seismicity are the Middle Americas subduction zone and the westernmost portion of the Antillean subduction zone beneath eastern Hispaniola. Apparently high seismic activity at the latter site is primarily due to the many intermediate depth aftershocks of the large magnitude Aug. 4, 1946 earthquake (see Table 1, *Russo and Villaseñor* [1995; 1996a]). Intermediate depth seismicity in the Lesser Antilles arc is somewhat more active than shallow seismicity in this region. Note the concentration of events near the Paria Peninsula (NE Venezuela) associated with northwestward subduction and slab tearing of oceanic South America beneath the Caribbean [*Perez and Aggarwal*, 1981, *van der Hilst*, 1990, *Russo et al.*, 1993, 1996]. Note also the few scattered earthquakes present in western Venezuela and Colombia, associated with subduction of the Nazca and Caribbean plates beneath South America [*Dewey*, 1972, *Pennington*, 1981, *van der Hilst and Mann*, 1994, *Malave and Suarez*, 1995], and activity within the Bucaramanga nest seismic region [*Schneider et al.*, 1987; *Frohlich et al*, 1996]. Intermediate depth events not within these regions, with the possible exception of the three such events near the Azuero Peninsula in Panama, are most likely depth mislocations. I treat these separately in the next section.

Causes of Location Errors

The best evidence that the relocations are subject to the same types of errors that commonly plague teleseismic earthquake locations is the intermediate depth seismicity not associated with subduction zones, visible in Fig. 2b. These events most likely are assigned incorrect depths during the generalized linear inverse because there is a nearly complete trade-off between event origin time and depth for earthquakes located without nearby stations. The latter was frequently the case for times early in the catalogue when the global seismometer network included few instruments, and for events in the westernmost portion of the study region, where the vast and station-free Pacific Ocean stretches nearly 90° to the southwest. This problem is manifest in the five or six intermediate depth events relocated to the oceanic Cocos plate (Fig. 2b): if these events are fixed to shallow depth, their epicenters do not change significantly, but their origin times become systematically earlier. Lack of stations in the southwestern quadrant probably accounts for the rather scattered intermediate depth seismicity in the Middle Americas region even late in the study period, thus, although the events show clear evidence for northeastward dip of the Cocos slab beneath Central America and southern Mexico, better data for estimating slab dip and morphology exist [e.g., *Protti et al*, 1994].

Other sources of location error include poor timing inherent in the early days of global seismology, resulting in gross (and often obviously inconsistent) arrival time errors. Another important factor, frequently noticeable for nearby stations located above the Middle Americas slab, is the effect of large departures of local velocity structure from the Jeffreys-Bullen radial earth model. The latter can bias arrival times at nearby stations by tens of seconds. Finally, since the data are arrival times from the ISS Bulletin and not seismograms, phase misidentification is a frequent and intractable problem, especially at local and regional source-receiver distances. Misidentification of head waves and triplications can introduce errors of 5 to 30 seconds, and even at teleseismic distances, for shallow events confusion of direct and surface reflected phases (e.g., *pP*, *sP*, *pS*, *sS*) is possible and detrimental to accurate location

The effects of timing errors and poor station distribution are greatest for small events recorded at few stations. For such events, a gross error in the arrival time for a nearby station can dominate the relocation completely leading to large errors in hypocentral estimation. In order to gauge this effect, I estimated the minimum number of stations necessary to ensure a fair degree (qualitative) of accuracy, while seeking to retain the largest number of event relocations (see also the discussion in *Cahill and Isacks*, [1992]). I concluded that events located with 20 or more arrival times were likely to be sufficiently accurate for most tectonic or seismic hazard analyses. Thus, I suppressed events relocated with fewer arrival times, and plot the remainder in Figures 3a and 3b. There is an inherent trade-off in such procedures because it is clear that in suppressing all such events, some small-magnitude well located events are inevitably discarded while some bad locations are retained. However, there is no simple was to distinguish one from the other, so the somewhat arbitrary cutoff of 20 arrival times was adopted.

Comparison of Figures 2 and 3 demonstrates the resulting improvement in the relocations' correlation to known structures and slab morphology. In Fig. 3a, the Caribbean plate boundaries are better defined, although by fewer events. The Oriente and Swan Islands Faults are still shown to be active along most of their lengths, and the earthquakes that define them fall more nearly on their bathymetrically defined traces. The width of the Panama Fracture Zone is reduced by nearly a factor of two relative to that in Fig. 2a. However, much of the seismicity visible in Fig. 2a in western Hispaniola, around Jamaica, and in the Nicaragua Rise-Hess Escarpment region has been suppressed. The potential danger is that the slower-moving, less seismically active regions of the plate boundary zones have been preferentially removed.

In Fig. 3b, the northeastward dip of the Middle Americas slab is clearer than in Fig. 2b. However, the remaining subduction zones are now very poorly defined relative to Fig. 2a. The number of intermediate depth events suppressed is proportionally greater than the shallow events so treated. This most likely reflects the well-known diminution of event magnitude with increasing depth. Implicit in such an interpretation is that the number of consistent arrival times is a moderately strong function of event magnitude, which I have not attempted to show explicitly because most of the events in the relocated catalogue lack any kind of magnitude estimation.

Comparison to NEIC hypocenters, 1963-1993

In an effort to demonstrate the accuracy of the relocated events relative to modern (i.e., post-1963) routinely located seismicity of the region, I show seismicity occurring between January 1963 and December 1993 within the study region, located by the National Earthquake Information Center (NEIC), in Fig. 4. For the reasons outlined above, I have only retained hypocenters for events recorded by at least 20 stations.

Several observations are immediately apparent. First, the modern catalogue is much more complete (many more events, especially at smaller magnitudes), both at shallow (Fig. 4a) and intermediate (Fig. 4b) depths, than the 1913-1963 relocated catalogue. This is primarily due to increases in the number of seismic stations globally and, especially, regionally in the Caribbean, and improvements in the sensitivity of seismometers with time. Thus, more arrivals from smaller earthquakes are recorded, more phases are associated in recognizable events, and locations are possible. Second, as do the relocated events, the modern earthquakes are almost exclusively restricted to Caribbean plate boundary zones: there is little evidence for internal Caribbean plate deformation. Finally, the primary improvement over the 1913-1963 catalogue visible in the modern catalogue is in depth control of the hypocenters (compare Figs. 3a and 4a, 3b and 4b). Benioff zones are much better defined in the modern data set, although some obvious errors of the types discussed above still occur, e.g., events assigned depths of 75-200 km at the Middle Americas trench (Fig. 4b) rather than in the Benioff zone. One notable disadvantage of the modern catalogue is the practice of fixing the depth of shallow earthquakes to 33 km, which has a visible effect on scatter in subduction zones (see Fig. 4a) and, undoubtedly, an unresolved entropic effect on events located elsewhere. The 1913-1963 relocations are not free of this type of effect, but by fixing depths at 10 km (or shallower in some cases) I avoid introducing obvious scatter in the shallow portions of subduction zone at least.

Contours of Subducted Lithosphere

I used the combined relocated 1913-1963 and the NEIC 1963-1993 data sets to estimate the positions and depths of the Middle Americas and Antillean slabs. For both estimates (Fig. 5) I used the culled data sets (no. of arrival times greater than 20 for the pre-1963 data, and no. of stations greater than 20 for the modern data). The data were insufficiently numerous to estimate slab orientation or gross morphology in northwestern South America, however. Interested readers are referred to studies focused on these subduction zones for more details [*Dewey*, 1972, *Pennington*, 1981, *Schneider et al.*, 1987; *Malave and Suarez*, 1995, *Frohlich et al.*, 1996]. In northeastern South America, the southernmost end of the Lesser Antilles slab west of Trinidad was mapped where the seismicity sufficiently defines the active portion of the slab (see also *Perez and Aggarwal*, [1981], *Russo et al.*, [1993]). I incorporated the results of travel-time tomography in this region *Russo et al.*, 1995, *VanDecar et al.*, 1996] to extend the slab to its full (aseismic) extent. It was also possible to estimate the Muertos Trough slab orientation beneath southeastern Hispaniola [*Byrne et al.*, 1985].

In the Middle Americas subduction zone, where several estimations of slab location and morphology have been made [*Burbach et al.*, 1984; *LeFevre and McNally*, 1985, *Protti et al.*, 1994; 1996], the contours shown on Fig. 5 are comparable and differ only in details. In general, the more focused studies of slab structure like those of *Protti et al.* [1994, 1996] are probably more reliable where they are valid than the contours shown in Fig. 5. In the Antillean subduction zone, the infrequency of shallow subduction related seismicity [*Stein et al.*, 1982; *Russo et al.*, 1993; *Doser and VanDusen*, 1996] made it impossible to estimate the shallow (< 70 km) morphology of the slab, although more detailed studies [*Dorel*, 1981, *Girardin and Gaulon*, 1983] and studies using local networks [*Shepherd and Aspinall*, 1983; *Wadge and Shepherd*, 1984] give reliable indications of shallow slab structure in portions of the Lesser Antilles arc.

Comparison of Relocated Events with Surface Structures (Gravity Field)

In an attempt to match shallow seismicity from the relocated (1913-1962) and NEIC (1963-1993) data sets with potential surface faults in the marine portions of the study region, I plotted epicenters of events shallower than 50 km deep on the recently published free-air gravity anomaly map of *Sandwell and Smith* [1996]. The free-air gravity map (Fig. 6) is suitable for this purpose because free-air anomalies generally track topography and bathymetry closely, and the clear superiority of the resulting bathymetry proxy (at 2' by 2' gridding) over, for example, etopo5 (5' by 5' gridding; used for bathymetry in Figs. 1-5) makes for a more reliable correlation between bathymetric features and superposed seismicity. The entire data set of relocated events plus NEIC earthquakes recorded by 20 or more stations is shown in Fig. 6a. In Fig. 6b, I show the same NEIC events but only historic-era relocations based on 20 or more arrival times.

Aside from correlations of seismicity with faults mentioned above (e.g., Swan Islands and Oriente Faults), seismicity in Fig. 6a correlates to some degree with regions of sharp variation in bathymetry that probably reflect structural differences. For example, the few events within the North American plate tend to lie on ocean-continent transitions such as along eastern Yucatan, around Cuba, and northeast of the Bahamas. Structural heterogeneities like ocean-continent transitions are marked by variable bathymetric relief and can function as stress concentrators. Hence, the associated infrequent seismicity. Other notable correlations visible in Fig. 6a include six events on the Pedro Bank, SW of Jamaica, a scattering of events on the Nicaragua Rise and near the Hess Escarpment, some events along the accretionary complex north of Colombia's Caribbean coast, and a few events at the edges of the Cariaco Trough north of the Venezuelan coast. The South American continental shelf east of Trinidad is the site of some apparently deeper (> 25 km) earthquakes in the modern catalogue, but not in the pre-1963 data set. On the Pacific side of Central America, the Cocos Ridge appears to variably seismically active in several spots along its length within the study region.

In the culled data set (Fig. 6b), many of the correlations visible in the wider data set are absent (for example, the correlation between events and ocean-continent boundaries within the North American plate). However, others persist or are perhaps even enhanced. Seismicity at the Pedro Bank persists, as does that north of the Caribbean coasts of Colombia and Venezuela, and along the Cocos Ridge. Seismicity in between the Nicaragua Rise and Hess Escarpment is greatly diminished, but the two events that remain fall precisely on the eastern margin of the Nicaraguan Rise where an apparently deep bathymetric trough exists. Nevertheless, the reduction of seismicity within North America, in particular, in Fig. 6b I take to be evidence that the somewhat arbitrary cut-off in the data set at 20 arrival times is misleading where low strain-rate, slower, and less seismically active processes are concerned. On the other hand the latter result is a confirmation of the general validity of the plate boundary zone postulate for the Caribbean region.

Conclusions

Relocations of 1073 earthquakes that occurred in the Caribbean region (5°N–25°N, 50°W–90°W) between January 1913 and December 1962 confirm that during this period seismic activity in the Caribbean region is largely restricted to the wide plate boundary zones of the Caribbean, Nazca, Cocos, and North and South American Plates. These boundaries are better defined by the post-relocation events, and some apparently lower magnitude seismicity associated with low strain-rate regions outside the plate boundaries exists. Plate-boundary seismicity for this time period is most intense along the Middle Americas subduction zone (Guatemala to Coast Rica segment), the Panama Fracture zone, and in eastern Hispaniola and the Puerto Rico Trench region. Seismicity is more scattered in the Colombian and Venezuelan Andes during the study period. The southern Lesser Antilles region is marked by an almost complete absence of shallow seismicity between 1913 and the end of 1962 in the relocated data set. Low levels of seismic activity in the region persists in modern times, but at a rate higher than that in the relocated catalogue, probably reflecting establishment of regional networks and improved seismometers. Major faults of the North America-Caribbean plate boundary, like the Oriente, Swan Islands faults, and the Cayman Spreading Center, are fairly well-defined by shallow events in both the pre- and post-1963 events. Intraplate seismicity occurs south of Jamaica near the Pedro Bank, east of the Bahamas, near the Yucatan coast, and along the Cocos Ridge.

Intermediate depth seismicity in the Middle Americas and Greater Antilles subduction zones defines subducted lithosphere at an accuracy only slightly less good than that of modern teleseismically located catalogues. Data were insufficient, however, to improve understanding of subduction in the northern Andes. The deepest earthquakes in the region occur in the Middle Americas subduction zone and attain depths in excess of 275 km.

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