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Historical Seismicity of the Southeastern Caribbean and Tectonic Implications

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Abstract—We have relocated the twenty-eight largest magnitude ($4.3 \leq M_s \leq 7.3$) historical (1922–1963) earthquakes of the southeastern Caribbean. We also present new focal mechanisms for seven of these events. The relocations are based on reported ISS *P* and *S* arrival times that we analyzed using generalized linear inversion techniques. The new focal mechanisms were constrained by first motion *P* polarities as reported by the ISS and as picked by us where records were available, and by the polarities and ratios of *SH* and *sSH*, and *SV* and *sSV* arrivals that we determined from seismograms. The results of the relocations are commensurate with the distribution of seismicity observed in the recent era: hypocenters are shallow and intermediate in depth (0–200 km), and the events occur almost exclusively in areas known to be currently seismic. The frequent seismic activity in the vicinity of the Paria Peninsula, Venezuela, is clearly a persistent feature of the regional earthquake pattern; intermediate depth earthquakes indicative of subduction beneath the Caribbean plate occur here and along the Lesser Antilles arc. The Grenadines seismic gap is confirmed as an area of low seismic moment release throughout the historical era. Trinidad and the eastern Gulf of Paria were also largely quiescent.

The new focal mechanisms, despite being a sparse data set, give significant insight into both subduction processes along the Lesser Antilles arc and into the shallow deformation of the Caribbean–South America plate boundary zone. The largest earthquake to have occurred in this region, the 19 March 1953 event ($M_m = 7.01$), is a Lesser Antilles slab deformation event, and another earthquake in this region of the Lesser Antilles is probably a rarely-observed interplate thrust event. Shallow deformation in the plate boundary zone is complex and, near the Paria Penninsula, involves mixed southeastward thrusting and dextral strike-slip on east-striking faults, and secondarily, normal faulting. Bending of the subducting Atlantic–South American plate also seems to generate seisms. The rather high ratio of intraplate deformation to interplate deformation observed along the Lesser Antilles subduction zone in the more recent era seems to have been operative in the historical era as well.

Key words: Historical earthquakes, southeastern Caribbean, relocations, focal mechanisms.

Introduction

The southeastern Caribbean has long been recognized as an area of frequent strong seismicity. Shallow (0–70 km) and intermediate (70–200 km) depth earth-

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quakes of the region have been recorded at teleseimic distances since the early decades of this century, and extensive local networks have been maintained in the Lesser Antilles since the 1950s (TOMBLIN, 1975; SHEPHERD and ASPINALL, 1983), and in Martinique, Guadeloupe, and Venezuela from even earlier (DOREL, 1981; FIEDLER, 1988). We have examined this wealth of historical seismicity data (instrumentally-recorded events between 1922 and 1963) and relocated and determined focal mechanisms for earthquakes of the region with the goal of answering several questions concerning the region's tectonics.

First, the region includes the Caribbean–South America plate boundary (Figure 1). Plate velocities along this boundary are low (around 1 cm/yr at Trinidad; DEMETS *et al.*, 1990), the recent (i.e., post-1962 henceforth) seismicity patterns are highly heterogeneous, and deformation within the plate boundary zone is complex (RUSSO *et al.*, in press). Although there is a consensus that the Caribbean is generally moving east relative to South America (MOLNAR and SYKES, 1969; SYKES *et al.*, 1982; SPEED, 1985; ROBERTSON and BURKE, 1989), the magnitude and direction of north-south motion in this plate boundary is disputed, and the range of possibilities includes pure strike-slip, oblique collision, and oblique extension. Thus, it is very important to understand the mode of seismic strain over a time

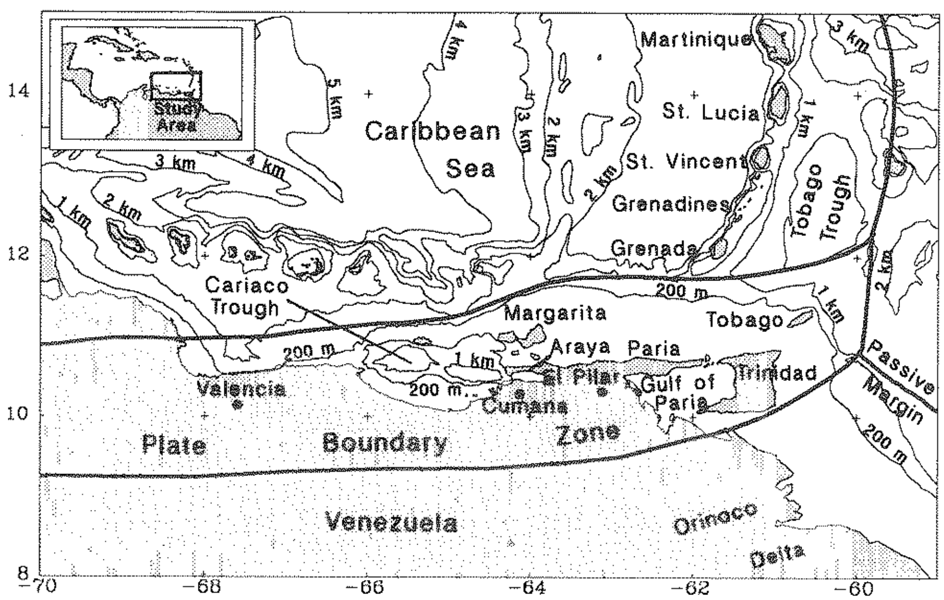

Figure 1

Caribbean–South America plate boundary zone. Atlantic oceanic lithosphere moving with either North America or South America subducts beneath the Caribbean plate along the heavy line east of the Lesser Antilles (Martinique to Grenada). The area bounded by the heavy lines is the approximate Caribbean–South America plate boundary zone. Localities mentioned in the text and bathymetry are also shown.

period longer than the WWSSN era. The potential for identifying seismic gaps, for differentiating between large magnitude plate boundary slip events and secondary seismic deformation, for isolating the most important components of deformation within the plate boundary zone, and the implications of these characteristics of the regional seismicity for the relative motions of South America and the Caribbean require a careful analysis of this dataset.

Second, throughout the recent period, subduction of Atlantic oceanic lithosphere beneath the Caribbean plate, represented by the Lesser Antilles island arc, has been characterized by a slow rate of subduction and a low level of moment release (STEIN *et al.*, 1982, 1983, 1988; SYKES *et al.*, 1982), by the paucity of large magnitude interplate thrusting earthquakes relative to large magnitude intraplate deformation events in the Lesser Antilles subduction zone (STEIN *et al.*, 1982), and by the obvious seismic gap between St. Lucia and Grenada (Figure 1), beneath and around the Grenadines (WADGE and SHEPHERD, 1984). Thus, either the interplate slip in this subduction zone, in the recent period, is taking place largely aseismically, or perhaps this motion causes only very modest earthquakes, or, in the case of the Grenadines gap, perhaps no such motion is occurring at all. An examination of the historical earthquakes is warranted to determine whether or not this pattern of seismic strain has been of long duration. If confirmed, the presence of the seismic gap may require reappraisal of estimates of seismic hazard for the area in question. The lack of large magnitude interplate slip earthquakes and the high frequency of intraplate deformation earthquakes, if persistent, would require some tectonic explanation.

Determining the longer term mode of strain release for this region takes on added importance when one considers that mechanisms of northern Lesser Antilles earthquakes have been used to constrain Caribbean–North America relative motions (SYKES *et al.*, 1982), and that the resulting Euler pole is very different from other Caribbean–North America relative motion estimates based on the Cayman Trough ridge–transform system (JORDAN, 1975; MINSTER and JORDAN, 1978; STEIN *et al.*, 1988; DEMETS *et al.*, 1990). Predicted motions based on the Lesser Antilles focal mechanisms involve dextral extension in the Caribbean–South America plate boundary zone, whereas the Cayman Trough-based prediction entails dextral collision between the two plates. The latter group of workers have discounted the northern Lesser Antilles earthquake slip vectors, proposing that these earthquakes reflect more slab and Caribbean intraplate deformation than true interplate slip. Thus, two important questions are whether or not historical earthquakes in the southern portion of the Lesser Antilles subduction zone are clearly identifiable as interplate or intraplate, and whether the Grenadines (Figure 1) seismic gap, mentioned above, has been a persistent feature of the arc seismicity. Although the catalogue of reported earthquakes in the southeastern Caribbean–northern Venezuela region extends back to the 1500s (ROBSON, 1964), we use the term "historical" earthquakes throughout this paper to mean instrumentally recorded events occurring between 1922 and 1963. This usage is in keeping with standard terminology in the sub-discipline (LEE *et al.*, 1987).

Recent Seismicity of the Plate Boundary Zone

NEIC (National Earthquake Information Center) hypocenters of recent (1963–1990) seismicity of the southeastern Caribbean are shown in Figure 2. A brief discussion of the recent earthquakes in the region is necessary because the recent data set is far more complete than the historical data set, and because we are interested in how the historical earthquakes, which are almost all larger magnitude events, fit into and refine the tectonic inferences that we derived from the recent hypocenters (RUSSO *et al.*, in press). Our discussion centers on the distribution of recent earthquakes and on the displacements delineated by focal mechanisms of the larger magnitude recent events.

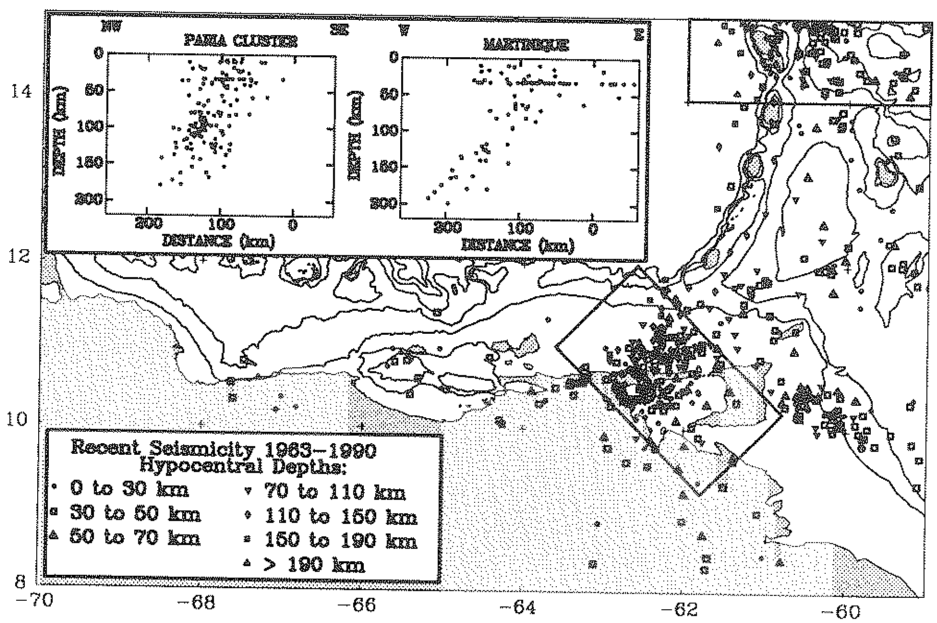

Figure 2

Map of the recent (1963–1990) seismicity of the southeastern Caribbean. Hypocenter depths are given in the key (inset). The E-W trending box at the northeast of the area encompasses the hypocenters projected onto a vertical plane (Martinique inset) showing the well-developed Benioff zone at Martinique and St. Lucia. The NW-SE trending box similarly includes data used to produce the Paria Cluster vertical section (inset). Note the decrease in seismic activity in the region between the two regions (Grenadines gap). The many events to the east and ESE of Trinidad are the 10 March 1988 earthquake and its aftershock series.

At the northern boundary of the study area (Figure 2), Atlantic oceanic lithosphere subducts beneath the Caribbean plate. A well-defined Benioff zone is apparent in the recent seismicity (Figure 2): earthquakes deepen westward from the lithospheric subduction trace inferred at 60°W (SPEED *et al.*, 1984) to depths of 200 km beneath and immediately west of the Lesser Antillean islands of Martinique and St. Lucia (DOREL, 1981; GIRARDIN and GAULON, 1983; WADGE and SHEPHERD, 1984). Focal mechanisms of Benioff zone events indicate that the large magnitude earthquakes here are almost exclusively caused by slab deformation or sinking. Similarly, shallow earthquakes of the area are related to deformation of the downgoing lithosphere. There are few large magnitude interplate slip events (STEIN *et al.*, 1982).

South of St. Lucia the number of both shallow (0–70 km) and intermediate depth (70–200 km) earthquakes diminishes. This area of reduced seismicity is the Grenadines seismic gap. The continuity of subduction between the Benioff zone at St. Lucia, and Grenada, at the southern end of the Grenadines gap, is inferable from the occurrence of deeper earthquakes in the gap; from the unbroken continuity of the Lesser Antilles arc platform and the Bouguer gravity minimum associated with the subduction zone (SPEED *et al.*, 1984); from deformation observed in forearc sediments (TORRINI and SPEED, 1989); and from the recent volcanism of St. Vincent (SHEPHERD *et al.*, 1979; SHEPHERD and SIGURDSSON, 1982) and Grenada (ARCULUS, 1976). The eastern portion of the Grenadines gap, where frequent shallow earthquakes are expected but not observed, is coincident with a thick accretionary prism.

South of Grenada, the number of earthquakes increases precipitously to a maximum of seismic activity centered on the Paria Peninsula of northeastern Venezuela (Figure 2). We designate these earthquakes the Paria Cluster. This patch of earthquakes forms a fairly broad NE trending rectangle (70 km wide by 200 km long) extending from near the Venezuelan town of El Pilar, in the southwest, to the southwestern corner of the Tobago Trough, in the northeast. The concentration of recent-era hypocenters decreases markedly from southwest to northeast, and dies out as a traceable trend at approximately 12°N, at a point some 50 km east of Grenada. The Paria Cluster does not include the many shallow earthquakes south of the Gulf of Paria. The Paria Cluster earthquakes form a Benioff zone dipping approximately 60°NW (Figure 2; SHEPHERD *et al.*, 1990; RUSSO *et al.*, in press) and indicate that lithosphere attached to, or once attached to, South America is subducting beneath the Caribbean. Focal mechanisms of intermediate depth Paria Cluster events indicate that this slab is sinking and in down-dip tension, and is also deformed by along-strike compression which causes lateral bending of the slab. The subducting slab at Paria is continuous with lithosphere subducting beneath the Caribbean plate, and with Atlantic ocean lithosphere that is joined to South America at its eastern passive margin. The Paria Cluster earthquakes terminate abruptly to the west and southwest, where seismicity drops off to low levels. The intermediate depth events associated with subduction cease completely. RUSSO and SPEED (1992) have proposed that southwest of the Paria Cluster the subducting lithosphere has detached completely from South America, and, therefore, the slab is sinking without seismic deformation into the mantle.

Shallow earthquakes occur both west and south of the Paria Cluster. The events to the west are almost certainly part of the Caribbean–South America plate boundary zone deformation. Unfortunately, because of their small magnitudes, they are difficult to analyze fully, and thus the motions they represent are largely unknown. The more diffuse earthquakes south of the Paria Cluster may be related to sediment loading of the Guyana Shield by the Orinoco River delta. The reported focal depths of some of these events, as deep as 50–70 km, may be mislocations: there are no near stations to constrain depths well.

Analyses of all available focal mechanisms of recent shallow earthquakes of the Caribbean–South America plate boundary zone (RUSSO *et al.*, in press) reaffirm what is evident from studies of the regional surface geology: deformation, both seismic and otherwise, is complex, but a strong component of active, south-southeastward thrusting and folding is apparent (ALVAREZ *et al.*, 1985; ROSSI *et al.*, 1987). Shallow (0–15 km) thrust earthquakes with north-dipping ENE-striking nodal planes occur within the Venezuelan thrust belt indicating that it is currently active and that this deformation is probably thin-skinned. The thrust events occur in the Araya-Paria isthmus and the western Gulf of Paria, but are not known to occur elsewhere.

Coincident with the thrust earthquakes, geographically, dextral strike-slip events occur on shallow, E-striking faults in a restricted, but fairly wide zone (40 km), of northeastern Venezuela, but do not occur in the eastern Gulf of Paria or Trinidad. The geologic evidence of strike-slip faults in this region is equivocal (SCHUBERT, 1979), but there is some indication in the focal depths of the strike-slip earthquakes (12–35 km) that this deformation may be accommodated at depth below the thrust belt. In addition, focal mechanisms with mixed thrust and strike-slip components of earthquakes in the western Gulf of Paria may indicate that there is a west-to-east gradient in partitioning of the two types of motion: in the western active zone, strike-slip and thrusting are fully partitioned; in the western Gulf of Paria the two motions are coupled; and in the eastern Gulf and Trinidad this deformation is aseismic. This observation is commensurate with the previously noted west-to-east diachronous development of thrust belt structures and timing of deformation (SPEED, 1985).

The absence of strike-slip earthquakes, or really of any larger magnitude events east of the western Gulf of Paria is also highly significant because of postulates that a strike-slip fault of recent large-displacement, the El Pilar fault, extends through the Gulf and across Trinidad into the western Atlantic (ROBERTSON and BURKE, 1989). We thus confirm the observation of SHEPHERD *et al.*

(1990), who have noted based on 25 years of Trinidad Local Network data (1964–1989), that the supposed east extension of the 'El Pilar' is largely devoid of earthquakes of any magnitude. It is, of course, possible that in the recent period the eastern Gulf of Paria and Trinidad are a seismic gap, and that analysis of the historical earthquakes might demonstrate that large magnitude earthquakes, perhaps with east-striking dextral-slip mechanisms, have occurred in this area. Thus, the locations and mechanisms of the historical era earthquakes are crucial to both seismic hazard analysis and to understanding the tectonics of the plate boundary.

The magnitudes of thrust and strike-slip earthquakes within the plate boundary zone are comparable ($m_b = 4.9$–5.7), and thus it is difficult to argue that one type of event represents primary, plate motion slip, and the other is caused by secondary, consequent deformation. Nevertheless, the addition of mechanisms of large historic earthquakes is highly desirable, because the numbers of both populations we have discussed are low. In addition to the thrust and strike-slip earthquakes, several shallow normal faulting earthquakes have occurred in the southeastern Caribbean. These events seem to be related to bending of the subducting Atlantic or South American lithosphere.

Relocations of Historical Earthquakes

We have relocated the twenty-eight earthquakes listed in Table 1 using arrival time data from the relevant ISS (International Seismological Summary) Bulletins. The earthquakes are from the period 1922–1962 and 18 of them have reported magnitudes measured at Pasadena (M_{PAS}) or Palisades (M_{PAL}) greater than 5.5. Such historical magnitudes have been shown by GELLER and KANAMORI (1977) to be equivalent to present-day M_s for shallow earthquakes. The largest magnitude event is the 19 March 1953 St Lucia event, with a reported $M_{\mathrm{PAS}} = 7.3$. Thirteen of the events have reported magnitudes between 6.0 and 7.0. An earthquake is listed in the NEIC (global database) on 13 April 1938 (13:53:13.0; $M_{\mathrm{PAS}} = 5.6$), but we could find no trace of this event in the ISS catalogue or in the BCIS catalogue (Bureau Central International de Séismologie). The twenty-eight events therefore include all the events which occurred in the southeastern Caribbean between 1922 and 1962 with at least one reported magnitude reaching 5.5. They also include a number of smaller magnitude events like foreshocks and aftershocks (see below). These earthquakes represent the majority of the seismic moment released in the region during the historic period.

We used an iterative least-squares location inversion which seeks to minimize travel time residuals calculated on the basis of the hypocentral estimate and the Jeffreys-Bullen earth model. We took care to include S arrival times in our solutions wherever possible, as these data are often crucial in adding constraints

Table 1

Relocation parameters

Date	Relocation: Time GMT	Relocation: Epicenter °N	Relocation: Epicenter °W	Relocation: Depth (km)	Relocation: Magnitude	Original Location: Time GMT	Original Location: Epicenter °N	Original Location: Epicenter °W	Original Location: Depth (km)	Original Location: Magnitude
11 May 1922	06:45:28.6	11.33	60.05	5		06:45:25	11.80	60.50	0	6.0 PAS
08 Aug 1923	12:01:22.0	11.00	62.76	5		12:01:27	10.60	65.60	191	6.5 PAS
01 Feb 1926	01:17:47.6	10.88	62.40	147		01:17:33	10.60	65.60	191	6.5 PAS
27 Sep 1928	00:44:09.8	11.58	59.49	30		00:44:00	11.80	60.50	0	6.5 PAS
17 Jan 1929	11:45:42.0	10.35	63.98	20		11:45:34	10.60	65.60	0	6.9 PAS
10 Apr 1935	22:32:35.1	10.64	62.77	102		22:32:29	11.00	62.60	0	6.5 PAS
11 Aug 1939	Unconstrained by available data					Undetermined Shock				
14 Oct 1939	06:02:21.1	10.67	63.17	20		06:02:22	10.70	63.80	0	5.6 PAS
27 Feb 1940	12:12:41.7	8.51	60.96	5		12:12:40	8.30	60.80	0	
23 Jun 1940	18:59:36.3	10.08	68.19	20		18:59:33	10.00	68.00	0	5.6 PAS
06 Jul 1940	03:40:20.2	13.31	61.48	160		03:40:15	13.50	61.50	96	6.5 PAS
06 May 1942	21:18:01.9	10.91	64.56	10		21:18:03	10.80	64.90	0	
06 May 1942	22:50:13.0	10.85	64.93	10		22:50:12	10.80	64.90	0	6.0 PAS
06 Feb 1944	18:40:33.0	10.50	62.94	50		Undetermined Shock				
23 Dec 1945	08:09:56.1	10.20	61.95	28		08:09:59	10.20	61.70	65	6.5 PAS
21 May 1946	09:16:44.6	14.78	60.40	49	6.0 M_s	09:16:43	14.70	60.50	0	7.0 PAS
31 Jul 1946	00:29:01.1	12.07	59.91	56		00:28:54	12.20	59.80	0	6.0 PAS
21 Aug 1949	20:33:39.0	10.52	62.69	162		20:33:32	10.50	62.60	96	
20 Apr 1951	22:54:35.2	10.62	62.37	10		22:54:33	10.50	62.60	0	
23 Dec 1951	06:57:20.6	14.88	61.28	177		06:57:21	15.10	61.30	160	
24 Aug 1952	15:00:59.9	11.81	61.81	170		14:59:43	12.00	61.00	0	
31 Dec 1952	01:38:20.4	11.53	59.23	40		01:38:15	11.70	59.00	0	
19 Mar 1953	08:27:58.9	14.00	61.24	133	7.01 M_m	08:27:52	14.00	61.20	128	7.3 PAS
25 Jun 1953	21:49:07.8	11.12	62.41	121		21:49:07	10.50	62.60	128	
04 Dec 1954	18:31:10.9	10.79	61.36	46	6.1 M_s	18:31:14	10.90	61.40	65	6.5 PAS
02 Oct 1957	12:27:55.2	10.88	62.79	10	5.5 M_s	12:27:57	10.91	62.79	0	$6^5/_8$ PAS
04 Oct 1957	05:26:03.4	10.86	62.77	6	6.7 M_s	05:26:06	10.97	62.79	10	$6^3/_4$ PAS
06 Oct 1957	00:54:07.7	10.88	62.68	10		00:54:06	10.85	62.66	0	5.1 PAL
25 Dec 1957	16:26:02.5	10.46	62.55	22	5.8 M_s	16:26:01	10.53	62.52	0	5.9 PAL

to locations, and they were not utilized in the ISS locations (WYSESSION *et al.*, 1991). Hypocentral parameters converged fairly rapidly for the majority of the earthquakes we relocated, but for some we were forced to constrain depths in order to achieve convergence. We did this by reperforming the location for a series of constrained depths and then choosing a best constrained-depth location based on the respective travel time residuals. See the Appendix for more details of the relocations. A Monte Carlo linear least-squares relocation scheme described in full by WYESSION *et al.* (1991) was used to test the relocations and to give an estimate of the directions of maximum and minimum epicentral control due to the distribution of recording station. We used 400 locations, each involving three inversion iterations about a Gaussian perturbed set of arrival times, to determine a statistically significant distribution of locations and a covariance matrix which defines a 95% confidence ellipse. The size of the ellipse depends on the standard deviation of the Gaussian error distribution, and the orientation of the ellipse's major and minor axes are functions of the recording station distribution, thus indicating directions of maximum and minimum epicentral control. The relocation results, along with confidence ellipses are shown in Figures 3 and 4. Interpretation of the results is relegated to the Discussion section below.

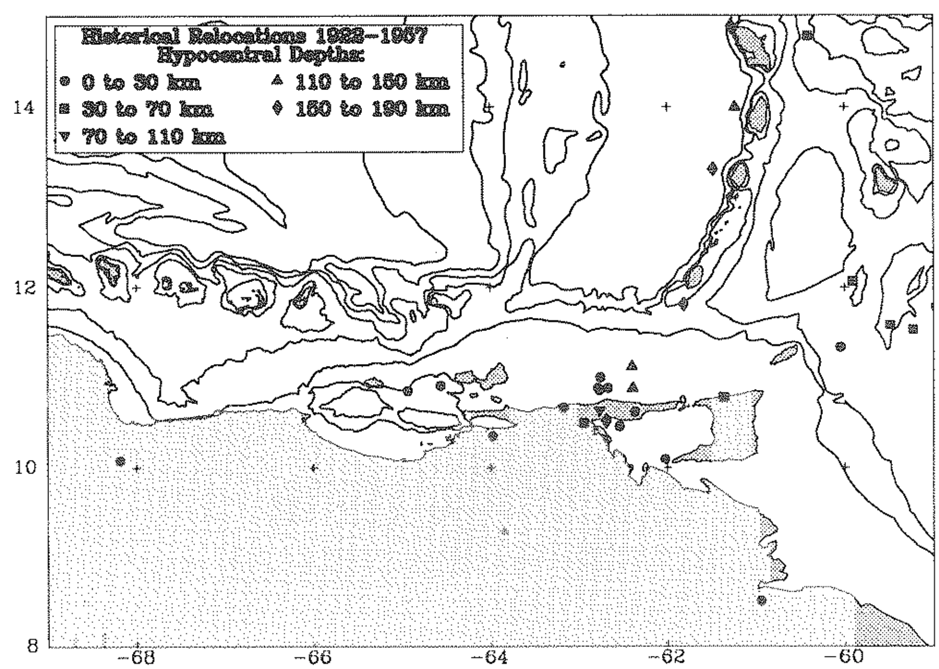

Figure 3
Map of the relocated hypocenters (1922–1963) of this study. Depths are given in the inset key.

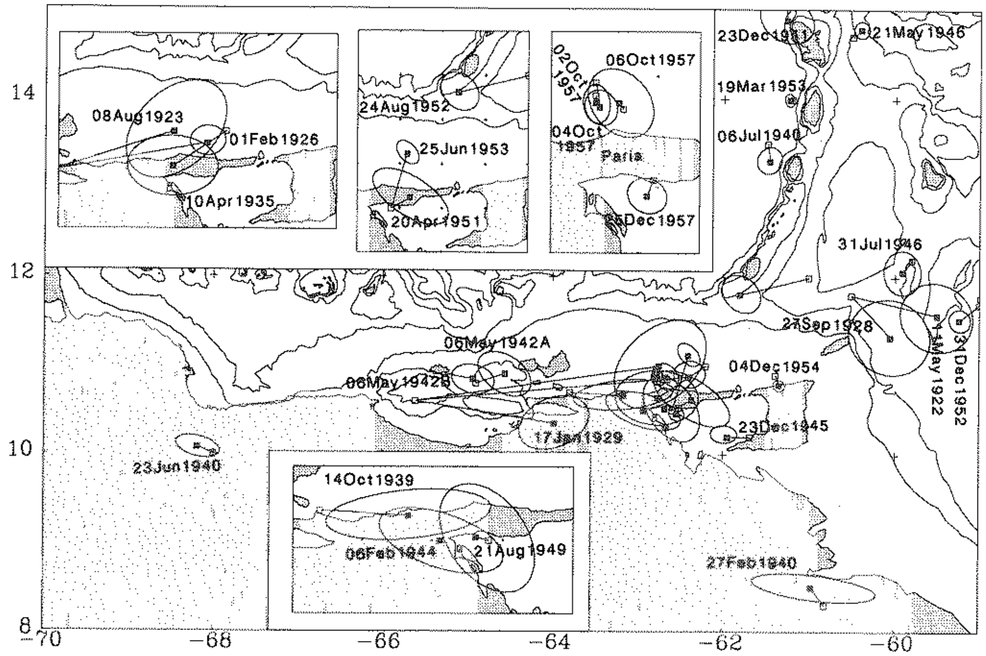

Figure 4

Map of 95% confidence ellipses of the relocated epicenters. Original ISS locations are unfilled squares. New locations are filled squares. Tie-lines join the old and new epicenters. The many events that occurred in the Paria Cluster region are shown in groups in the insets for clarity.

Historical Earthquake Magnitudes

When records of sufficient quality were available, we also reassessed the magnitudes of the events. This was performed primarily by measuring the amplitudes and periods of Rayleigh waves, yielding an M_s. In several cases, we found surface wave magnitudes that were substantially lower than those published. The magnitudes we have calculated are listed in Table 1. The 19 March 1953 event was energetic enough at long periods to allow a computation of the mantle magnitude, M_m (OKAL, 1992a).

Focal Mechanisms of Historical Earthquakes

We have produced new focal mechanisms for seven of the twenty-eight earthquakes we relocated. The mechanisms were constrained using a combination of ISS and BCIS first motion picks, first motion picks made by us from seismograms recorded at both local Caribbean stations and at stations at telesismic distances,

Table 2

Focal Mechanism Parameters

Date	Lat	Lon	Depth	Magnitude	Strike	Dip	Slip
21 May 1946	14.78	−60.40	49	6.0 M_s (TUC) 5.8 M_s (UPP)	325 ± 15	70 ± 2	85 ± 15
19 Mar 1953	14.00	−61.24	133	7.01 M_m (PAS)	260 ± 5	41 ± 5	290 ± 5
04 Dec 1954	10.79	−61.36	46	6.1 M_s (TUC)	190 ± 5	58 ± 1	253 ± 31
02 Oct 1957	10.88	−62.79	10	5.5 M_s (PAS)	47 ± 5	86 ± 2	225 ± 7
04 Oct 1957	10.86	−62.77	6	6.7 M_s (PAS 30-90)	75 ± 5	45 ± 5	139 ± 5
06 Oct 1957	10.88	−62.68	10		132 ± 6	44 ± 12	240 ± 30
25 Dec 1967	10.46	−62.55	22	5.8 M_s (PAS)	215 ± 5	87 ± 3	100 ± 20

and our determinations of *SV* and *SH* first motion polarities and ratios at several stations at teleseismic stations. Data pertinent to the seven earthquakes, their mechanisms, and the types of data used to constrain each are listed in Table 2 and in the Appendix.

As is frequently the case when dealing with historical events, we were faced with processing very limited datasets, occasionally involving instruments with dubious responses. In this respect, waveform inversion can usually provide a solution, but the critical evaluation of its quality in terms of confidence limits is difficult. Rather, as in our previous work (OKAL, 1984), we prefer to analyze each earthquake based on a number of robust constraints directly defined on the available seismograms, and body wave forward modeling. Each of the following constraints must be based on a datum of unquestionable quality, so as to warrant elimination of any mechanism violating the constraint.

1. Polarities of *P*, *SV* and *SH* body waves. These must correspond to impulsive arrivals, preferably read by one of us on long-period instruments. Conventions for *S* polarities are those of KANAMORI and STEWART (1976).
2. Polarities of depth phases such as *pP*, *sSV*, or *sSH*. Same remark as above.
3. *SH*/*SV* amplitude ratios. Such constraints are used only when one of the two components (for example, *SH*) is clearly much larger than the other one (typically by a factor of 2). We take a conservative approach by requiring the mechanism to predict an amplitude ratio $SH/SV > 1$ at the relevant station, and we reject all mechanisms for which the ratio would be less than 1. In using *SV* constraints, we must be careful to take into account the surface response function at the receiver. The latter, identically a factor of 2 for *SH* in a half-space, varies with the incidence angle j for *SV*, and with crustal structure and the frequency for realistic layered models (HASKELL, 1960, 1962; OKAL, 1992b). In practice, we use *SH*/*SV* ratios only at large distances ($j < 30°$) and long periods, for which the response functions for *SH* and *SV* remain comparable.

4. Amplitude ratios of depth phases to direct phases, such as pP/P or sSH/SH. The approach and the conservative limits are the same as in 3 above, but the surface response effects can be neglected, since the two waves reach the station under practically equal incidences.

A systematic computer search is then carried out by looping through all possible focal geometries (with increments of a few degrees in all three angles), and retaining only those compatible with the designated constraints. This procedure provides good insight into the allowable range of solutions compatible with the dataset.

As an example of our procedure, we outline the constraints on the focal mechanism of the 19 March 1953 earthquake. The Trinidad Local Network was not in place at the time of the earthquake. We inspected all the available records of the event at the NEIC Data Center in Denver, and identified polarities of P first motion arrivals at stations HUA, PAS, TUC, UPP, and LPB (all were dilatational). Reported first motion arrivals for 42 other stations, including the near stations FDF ($\Delta = 0.8°$) and BEC ($\Delta = 18.6°$), were taken from the ISS bulletin. These were then plotted on a lower focal hemisphere assuming a focal depth of 133 km and near-source P velocity of 8.2 km/s. As is clear from Figure 5a (and also noted by

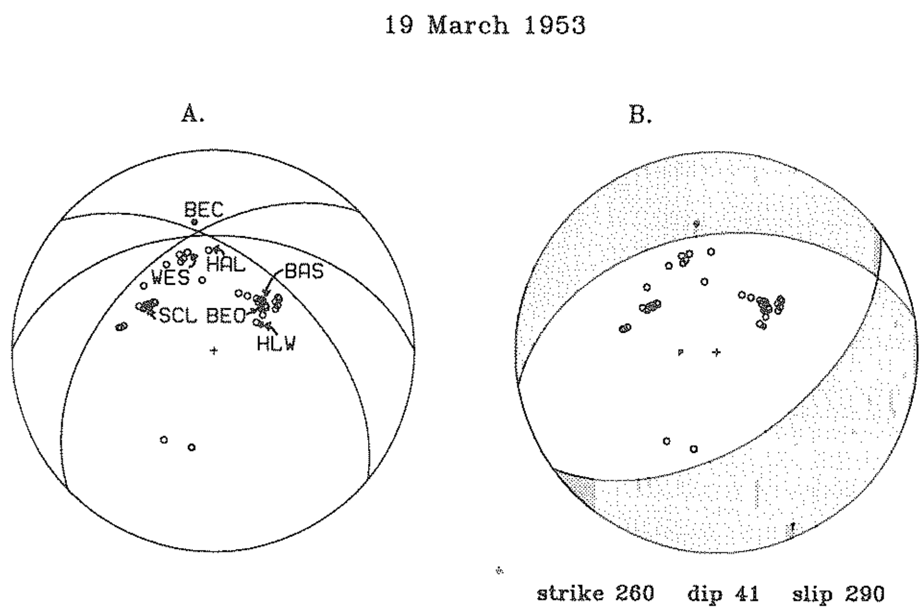

Figure 5
Available first motion P constraints for the 19 March 1953 event, as described in the text. Dilatations are unfilled circles, compressions are filled. A.: Inconsistent ISS P arrivals (SCL, BAS, BEO, and HLW) are labeled; arrivals at BEC (compressional) and WES and HAL (dilatational) marginally constrain one plane of the mechanism. B.: Final mechanism after constraints from other body-wave phases and body wave modeling are included.

SYKES and EWING, 1965), the focal mechanism is unconstrained, but entails normal faulting. Reported compressional arrivals from several stations (SCL, BAS, BEO, and HLW) are inconsistent with the many dilatations at other stations in the same distance-azimuth range, and we suppressed these arrivals. The resulting data yields some constraint on one nodal plane, which must lie between the compressional arrival at BEC, and dilatational arrivals at HAL and WES.

In order to get additional constraints on the focal mechanism, we examined the S arrivals at the stations for which we had records, in particular, PAS, TUC, UPP, and LPB. The data at two stations (PAS and UPP) are shown in Figure 6. The depth of the earthquake is confirmed by the 44 s separation of P and sP at PAS, and the 57 s separation between SH and sSH at UPP. Both stations being close to naturally polarized, we infer SH strongly to the North and SV strongly to the East (i.e., both negative using KANAMORI and STEWART's 1976, conventions) at PAS; and SV strongly to the West (again negative) at UPP. We add the constraint $SH/sSH > 1$ at PAS but refrain from using the ratio of SV and SH at PAS, because the incidence angle ($j = 33°$) is large enough affect the surface response of SV. At the more distant station UPP ($j = 26°$), on the other hand, we can impose $SV > SH$

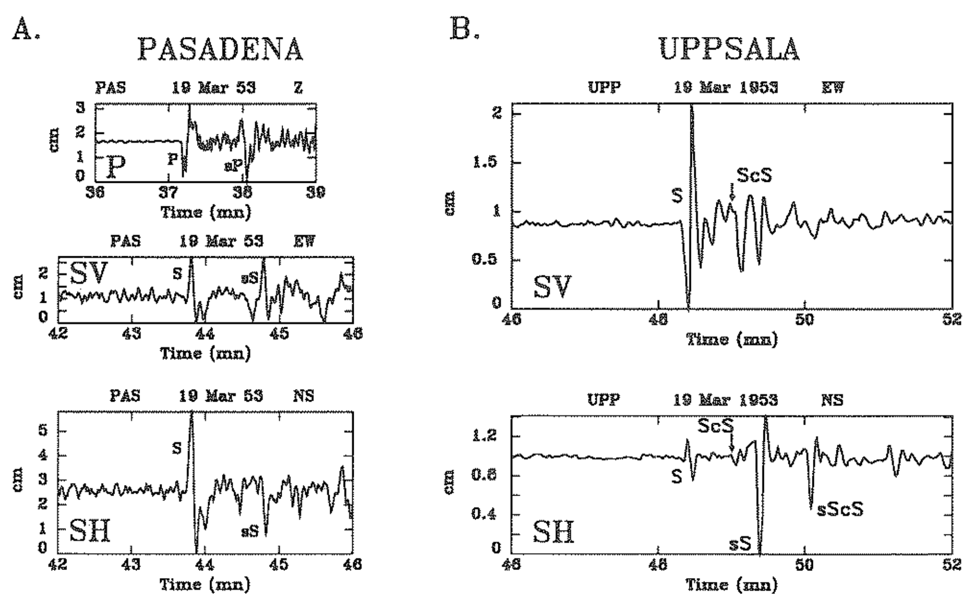

Figure 6

A.: P and S waves of the 19 March 1953 earthquake recorded at Pasadena (PAS) on the 1-90 Benioff seismographs. The 44 sec time separation of P and sP (vertical component) constrains the depth at 133 km. Polarities of SV and SH, and sSV and sSH are described in the text. B.: S-wave constraints on the 19 March 1953 earthquake mechanism recorded at Uppsala (UPP) on the Wiechert instrument. See text for further explanation.

and $sSH > SH$; however we do not use the polarity of SH because the arrival is itself small.

We acquired similar ocnstraints from S and sS arrivals at TUC and LPB, and used the S determinations in conjunction with the trustworthy P first arrivals to filter out inconsistent focal mechanism geometries by performing a search through all possible mechanisms. Source geometries which violated the P or S and sS arrival polarities or which resulted in inappropriate magnitudes of the ratios of SH, SV, and sSH and sSV excitation were discarded. This procedure significantly restricted the range of possible focal mechanism: while they cannot control strike and dip, the polarities of the S waves require the slip angle to exceed 270°, and in most cases 280° (assuming the focal mechanism is described using the plane with an azimuth between 230 and 310°, as shown on Figure 5). We fine-tuned the mechanism by waveform modeling of P and SH at PAS and UPP (see Figure 7). The final result ($\phi = 260°$; $\delta = 41°$; $\lambda = 290°$) is shown in Figure 5b. Waveform modeling is sensitive to variations in the focal angles on the order of 5°, and we take this number as the precision of our proposed mechanism. Thus, we have constrained the compression axis orientation for this event, which was not possible using P-wave first arrivals alone.

The average moment obtained from body waveform modeling is 4.5×10^{26} dyn-cm. This is about a factor of 2 smaller than suggested by the moment magnitude of the event, and expresses the classical trade-off between source time function and

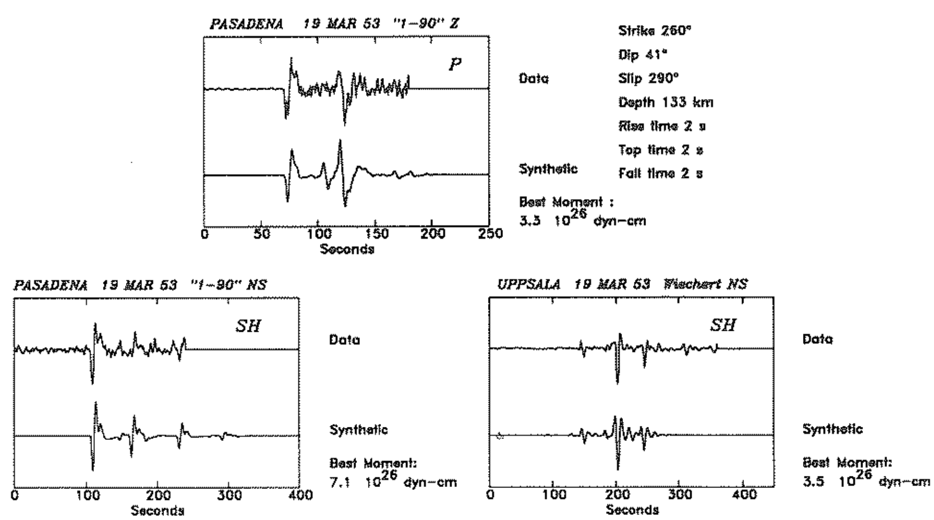

Figure 7

Body wave modeling of the 19 March 1953 event near St. Lucia. Each box compares the original seismogram (top trace) to the body wave synthetic in the final geometry (bottom trace). The Pasadena NS record has been reversed to conform to KANAMORI and CIPAR'S (1974) orientation conventions.

moment: it is probable that the use of a longer source time function would reconcile the two moments.

Through a procedure similar to that described above, we found mechanisms for the other earthquakes listed in Table 2. We were able to constrain one plane well in five of the remaining six mechanisms, but the constraints on second planes were generally not as strong. In several cases, given a range of possible orientations for the second plane, we made a choice of a best mechanism by bisecting the possible range and taking the median value orientation.

Discussion

Relocations

The most striking feature of the distribution of relocated historical earthquakes is the concentration of earthquakes which occurred within the Paria Cluster area (Figure 3). In particular, note that two early events (08 August 1923 and 01 February 1926) located by the ISS in the Cariaco Trough are moved 350 km east into the Paria Cluster. As noted above, in the recent period this region has been the locus of frequent shallow and intermediate depth earthquake activity. The geographic coincidence of the concentration of relocated historical events with the recent seismicity of the Paria Cluster is a clear indication that this feature of the plate boundary zone activity has persisted since earthquakes have been instrumentally recorded. The fifteen historical era earthquakes which relocate to this area also have shallow and intermediate depths (5–170 km), and there seems to be no reason to think they are not part of either the subduction process that the Paria Cluster represents, or the shallow deformation of the plate boundary zone. In particular, the earthquakes of 01 February 1926 (147 km), 10 April 1935 (102 km), 21 August 1949 (162 km), 24 August 1952 (170 km), and 25 June 1953 (121 km) most probably occurred within the slab subducting beneath the Caribbean, and, by analogy with the recent seismicity, are caused by intraplate deformation within it. Shallower events, such as the 08 August 1923 (5 km), 14 October 1939 (20 km), 06 February 1944 (50 km), 23 December 1945 (28 km), 20 April 1951 (10 km), the 02–04–06 October 1957 (10, 6, and 10 km, respectively), and the 25 December 1957 (22 km) earthquakes are more likely representative of the plate boundary zone deformation.

One earthquake (04 December 1954, $h^{*} = 46$ km, $M_{PAS} = 6.1$) occurred on the north coast of Trinidad. This is the only large magnitude earthquake to have occurred in this area in the historic or recent period. Smaller magnitude earthquakes are likewise infrequent (five since 1963, four in the historical period). Two recent era events have similar depths and also have mechanisms, and, as discussed below, the depths and mechanisms of these events make it unlikely that the earthquakes are related to surface faults such as the North Coast fault (ROBERTSON and BURKE, 1989).

Four of the relocated earthquakes are associated with the central and southern Lesser Antilles arc. Three of the events, 06 July 1940 (160 km), 19 March 1953 (133 km), and 23 December 1951 (177 km) are deep enough to be considered slab earthquakes, and all three lie west of the Lesser Antilles island chain. The 23 December 1951 event occurred in the Benioff zone below the northwestern corner of Martinique. The 19 March 1953 earthquake, with a mantle magnitude $M_m = 7.01$ is the largest southeastern Caribbean earthquake to have occurred in the historical era, and occurred just west of the north end of St. Lucia, on the southern limit of the area where recent seismicity defines a Wadati-Benioff zone (TOMBLIN, 1975; STEIN *et al.* 1982; WADGE and SHEPHERD, 1984).

The 21 May 1946 earthquake ($h = 49$ km; $M_s = 6.0$) occurred 50 km east of Martinique, in the region where one would expect the active interplate thrust faulting between the overriding Caribbean plate and the downgoing Atlantic lithosphere to be taking place. As discussed below in the section on focal mechanisms, this seems to be indeed the case, making this earthquake one of the few such events recorded in either the historic or recent era in the Lesser Antilles subduction zone.

The 06 July 1940 earthquake occurred west of St. Vincent, and was therefore clearly within the Grenadines seismic gap. Thus at least one large magnitude ($M_{\mathrm{PAS}} = 6.5$) earthquake of significant depth (160 km) occurred in the Grenadines gap in the historical era. Nevertheless, the total seismic moment released in this region, in both the historic and recent period, appears to have been substantially less than areas both to the north and south. Typically, the few earthquakes that have occurred along this segment of the Lesser Antilles arc, like the 06 July 1940 event, have been of intermediate depth (70–190 km), indicating that deformation of a subducting slab is ongoing. The recent and active volcanism of St. Vincent (SHEPHERD *et al.*, 1979; SHEPHERD and SIGURDSSON, 1982) and Grenada (ARCULUS, 1976), and the unbroken continuity of the arc platform and associated gravity anomalies (SPEED *et al.*, 1984) all indicate subduction is an active or recently defunct process. The almost complete absence of large magnitude earthquakes within this zone belies these observations, however. It is possible that convergent motions are aseismic because the interplate interface is lubricated by the subduction of accretionary prism sediments, of which there is an abundance in this zone. Another possibility is that convergence between the arc and the Atlantic oceanic lithosphere underlying the accretionary prism of the Tobago Trough has ceased and that the earthquakes, active volcanoes, and gravity-anomaly inducing arc morphology are maintained only by the slow destruction of the already subducted and recently detached oceanic slab. If this is the case, then it is likely that the long-sought Caribbean–North America–South America triple junction is at the north end of the gap, near St. Lucia or Martinique. This would be required by the differences in arc-normal motion between the actively converging northern segment and the nonconvergent southern segment. Furthermore, the Tobago Trough lithosphere must be a microplate in motion relative to continental South America (SPEED and WALKER, 1991) to accommodate active convergence between the arc and continental South America.

The area northeast of Tobago was the site of four large magnitude earthquakes (11 May 1922, $M_{PAS} = 6.0$; 27 September 1928, $M_{PAS} = 6.5$; 31 July 1946, $M_{PAS} = 6.0$; 31 December 1952) in the historical era. All four of the earthquakes occurred at shallow depths, 5–56 km. Recently, several large magnitude shocks have occurred just north of the area, along the Barbados Ridge, and one earthquake with a published CMT mechanism [11 February 1984 (DZIEWONSKI *et al.*, 1984)] occurred essentially at the same epicenter as the 31 July 1946 shock. The mechanisms and depths of the recent large events [05 September 1980 (DZIEWONSKI *et al.*, 1988a); 02 March 1981 (DZIEWONSKI *et al.*, 1988b); 11 February 1984, 05 October 1984 (DZIEWONSKI *et al.*, 1985); 11 April 1986 (RUSSO *et al.*, in press)] indicate that they are related to flexural bending of the Atlantic oceanic lithosphere as it moves toward the Lesser Antilles subduction zone (RUSSO *et al.*, in press). That is, the shallower earthquakes are normal faulting events whose nodal planes are parallel to the local strike of the Lesser Antilles subduction zone, and one deeper event is a similarly oriented thrust. It seems probable that the four historical era shocks, although none has a constrained mechanism, share this flexural bending as their cause. If this is true, then the flexure of the plate could be caused either by the load of a slab subducting beneath Grenada and the Grenadines, or by the load of the accretionary prism. Intuitively, one might suspect that the bending moment of the slab load might more easily cause the observed thrusting at depth, and that the prism load would result only in high-angle normal faulting (Figure 8), but neither can be denied as a possibility.

The remaining large magnitude relocated earthquakes are not sufficiently numerous to show significant patterns. Four shallow earthquakes (10–20 km) occurred west of the Paria Cluster, one near Cumana in NE Venezuela (17 January 1929; $M_{PAS} = 6.9$), two in the Cariaco Trough (06 May 1942, $M_{PAS} = 6.0$, and precursor), and one far to the west, near the Venezuelan city of Valencia (23 June 1940, $M_{PAS} = 5.6$). The earthquakes near Cumana, Margarita, and the Gulf of Cariaco are most likely related to deformation within the Caribbean–South American plate boundary zone, but in the absence of focal mechanisms and given the complexity of recent seismic deformation of this region, little else can be stated usefully. The earthquake near Valencia may be associated with deformation of the Merida Andes, perhaps with motion along the La Victoria fault (Fig. 11; SCHUBERT, 1981). The 06 May 1942 earthquakes may be associated with observed normal faulting in the Cariaco Trough (SCHUBERT, 1982). The 17 January 1929 event was the subject of a report by PAIGE (1930), who examined coseismic deformation in sediments in the environs of Cumana shortly after the earthquake

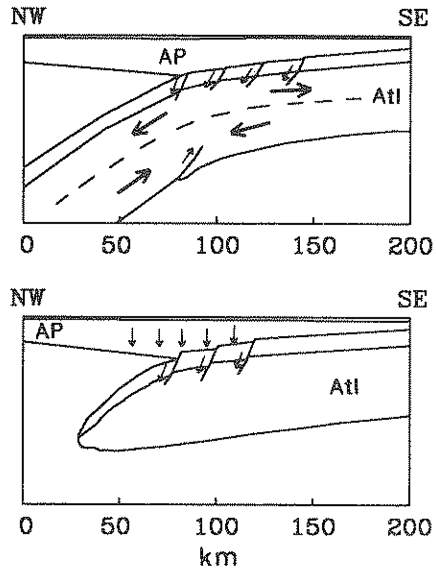

Figure 8

Schematic models of seismogenic faults in the Tobago Trough–Barbados Ridge forearc region. Top: plate bends as it subducts, causing extension and compression above and below, respectively, some neutral surface (dashed) in the plate. Bottom: loading by the accretionary prism causes normal faulting in the detached Atlantic lithosphere. The active thrust earthquakes noted in the recent seismicity (RUSSO *et al.*, in press) indicates that the plate bending model is perhaps more likely.

occurred. Paige was able to trace an east trending surface rupture 4 km along the south coast of the Gulf of Cariaco. Slip-sense indicators he observed indicate that both N-S extensive normal faulting and dextral strike-slip may have occurred during the earthquake. These motions may be consistent with observed recent dextral strike-slip and normal faulting in this portion of the Venezuelan thrust belt.

One shallow earthquake occurred south of the Paria Cluster beneath the Orinoco River delta. This earthquake (27 February 1940; $h = 5$ km; $M_{PAS} = 6.0$) is one of a series of events, previously noted in the recent seismicity of the region (RUSSO, 1990), which have occurred beneath the Orinoco River and delta. Some of these earthquakes show a clear alignment with the Orinoco River bed, which itself marks the northernmost exposures of the Guyana shield (FEO-CODECIDO *et al.*, 1984). It is possible that the earthquakes represent reactivation of an old fault zone.

Focal Mechanisms

As described above, we have produced seven new focal mechanisms of historical earthquakes in the southeastern Caribbean. Five of these earthquakes occurred in the Caribbean–South America plate boundary zone, within the Paria Cluster (04

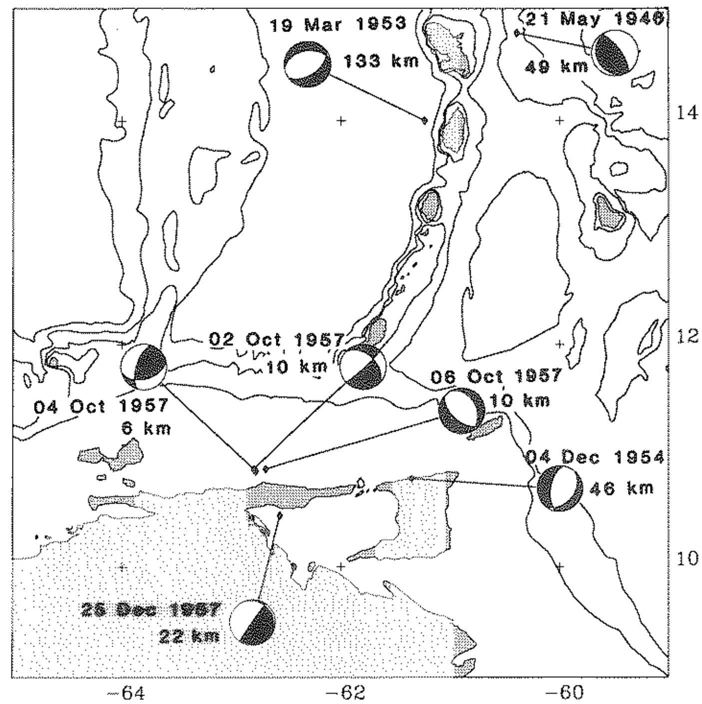

Figure 9
Map of focal mechanisms of seven historical earthquakes.

December 1954; 02–04–06 October 1957 earthquakes; 25 December 1957). The other two earthquakes with constrained mechanisms occurred in the Lesser Antilles subduction zone east of Martinique (21 May 1946) and west of St. Lucia (19 March 1953).

The 21 May 1946 earthquake has a thrust mechanism with a small and poorly determined component of strike-slip (Figure 9, Table 2). That is, the slip is such that small components of dextral or sinistral strike-slip are permissible on either fault plane, given the available constraints. Furthermore, there is significant ($\pm 15°$) range possible in the strike of the fault planes as well. Therefore, we restrict our interpretation of this earthquake to that mentioned above: the event demonstrates that interplate thrusting occurs east of Martinique, but the event cannot be used to further constrain plate motions. Given the sense of subduction along the Lesser Antilles arc, it is likely that the shallowly SW dipping plane is the active fault plane.

The focal mechanism of the 19 March 1953 earthquake (Figure 5) involves normal faulting on nodal planes that both strike at a high angle to the local strike of the Lesser Antilles subduction zone. As described above, the hypocentral depth ensures that the event occurred in the subducting Atlantic slab, and thus, the earthquake represents intraplate deformation of the downgoing lithosphere. The mechanism's geometry is commensurate with the extensional regime of lateral plate bending (CARDWELL and ISACKS, 1978; MCCAFFREY *et al.*, 1985). The fact that the largest magnitude earthquake to have occurred along the arc is an intermediate depth slab deformation event, rather than a shallow interplate thrust earthquake, supports the observation (STEIN *et al.*, 1982) that most of the recent interplate slip taking place along this subduction zone must occur aseismically, and implies that the observation is valid for the historic era as well. The meaning of this unusual partitioning of large magnitude seismic strain is not clear, and two hypotheses have been put forth to explain the frequency of intraplate earthquakes relative to interplate events: aseismic slip along the interplate zone facilitated by subduction of sediments derived from the thick accretionary prism, and aseismic interplate slip caused by decoupling of the subduction zone due to the large negative buoyancy of the old (80–100 Ma) lithosphere being subducted and the low convergence rate (2 cm/yr).

One of the few large magnitude earthquakes to have occurred in or near Trinidad in historic or recent times is the 04 December 1954 earthquake ($h = 46$ km, $M_{TUC} = 6.1$. It has a clear normal faulting mechanism. One plane is well constrained by data ($\phi = 190 \pm 5°$, $\delta = 58 \pm 1°$), but the slip on this plane is poorly constrained, and can range from 222° to 285°. Assuming that the plane constrained by the data is the fault plane, the earthquake's slip represents largely extension in the E-W direction at approximately 50 km depth. This could be caused by flexure of subducting South American transitional or oceanic lithosphere. The event's hypocentral depth precludes correlation with surface and near-surface faults, and in any case, no shallow normal faults with N or NNE strikes are observed near the epicenter. However, two recent era earthquakes with similarly oriented nodal planes and locations have been observed (Figure 10; RUSSO *et al.*, in press), although these earthquakes are both thrusting events. The similarities in location and fault strike of these three earthquakes, and their opposite senses of slip provide evidence for the reasonable interpretation that these events are plate bending earthquakes.

The 02–04–06 October 1957 and 25 December 1957 earthquakes all occurred on shallow (6–10 km) faults within the southern Caribbean plate boundary zone. The four earthquakes are among the largest magnitude events to have occurred in this portion of the plate boundary (see Table 2). As such, they deserve careful attention because of their potential for delineating primary seismic deformation and plate boundary motions, as opposed to earthquakes caused by secondary deformation.

Clearly, the three October 1957 events are related, but even a cursory examination of the three mechanisms (Figure 9) reveals that these earthquakes represent a very complex deformation. Two lines of inquiry are warranted: an examination of possible relationships among the four earthquakes, and among the earthquakes and

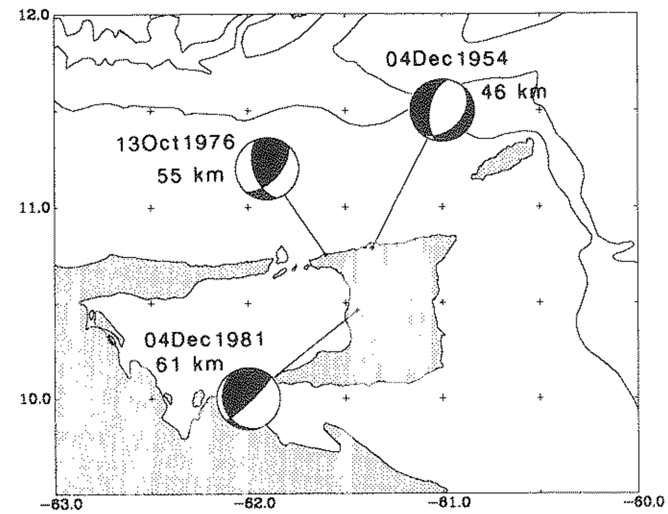

Figure 10
Mechanisms of recent Trinidad earthquakes with a similarly oriented fault plane to the 04 December 1954 event. Like the events in the Tobago Trough forearc, these may all be caused by plate bending.

other recent era earthquakes; and an analysis of the relationships between the earthquakes and coincident known near-surface faults and structures.

The three mechanisms of the October series are glaringly dissimilar, comprising as they do a mixed thrust and strike-slip mechanism (04 October), a mixed strike-slip and normal mechanism (02 October), and a normal mechanism (06 October), each with differently oriented nodal planes (Figure 11). The 02 October earthquake's mechanism exhibits pure strike-slip on a nodal plane striking NW and dipping approximately 45°NE, and a combination of strike-slip and normal displacement on a NE-striking, nearly vertically dipping nodal plane. The NE-striking nodal plane is parallel to several shallow faults identified in the region (see Figure 11; SPEED *et al.*, 1989), but its epicenter lies 20 km south of these faults, and the sense of displacement on this plane is dissimilar to that recognized on these faults—i.e., the faults have north side downthrow, but the earthquake mechanism shows a combination of dextral strike-slip and north side upthrow. Farther north in this area, there is one NE-striking fault with southward normal displacement, but it seems unlikely that the earthquake occurred on this fault, because of the discrepancy in their locations. Althouth the NW-striking plane is approximately parallel to several faults identified within other areas of the Caribbean–South America plate boundary zone (the Urica, San Francisco, El Soldado, and Los Bajos faults; see Figure 11), there is no fault with NW strike known in this area, and besides, these faults are believed to have dextral displacements (WILSON, 1968; BELLIZZIA, 1976; PEREZ and AGGARWAL, 1981; MUNRO and SMITH, 1984; SPEED,

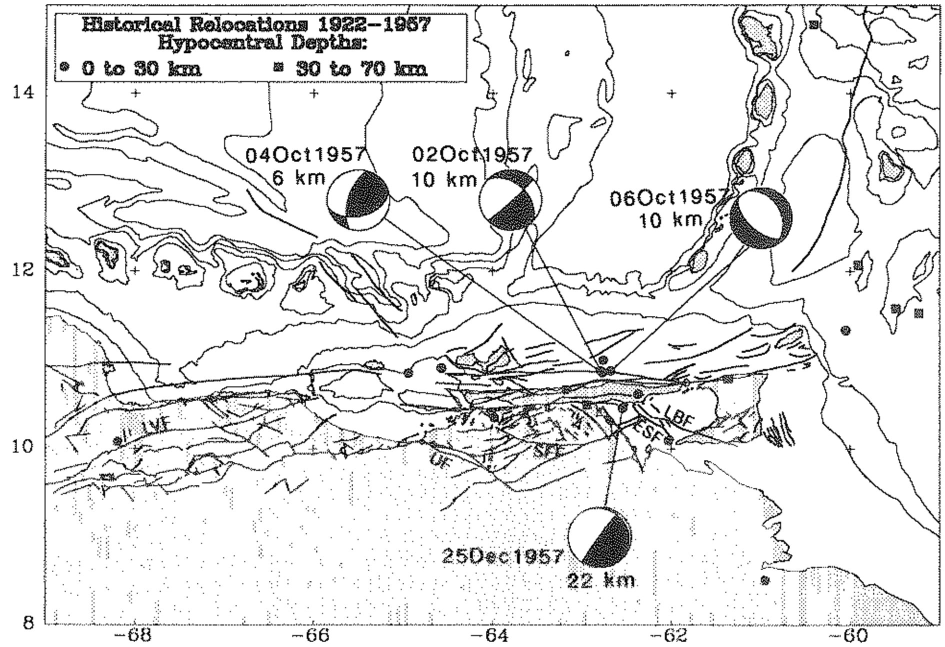

Figure 11
Shallow mechanisms and epicenters superposed on mapped faults of the plate boundary zone. Faults mentioned in the text are shown (LVF = La Victoria Fault; UF = Urica Fault; SFF = San Francisco Fault; ESF = El Soldado Fault; LBF = Los Bajos Fault). The 02–04–06 October 1957 sequence apparently occurred on or near on E-trending fault shown in the Figure.

1985; Rossi *et al.*, 1987), whereas the displacement on the NW-striking plane of the 02 October mechanism is clearly sinistral. The incompatibility of the senses of strike-slip between the mechanism and the near-surface faults is more difficult to reconcile than is the inconsistency of normal faulting dips, which may be explained by horst and graben type block faulting. Therefore, it seems more likely that the 02 October 1957 earthquake occurred on a NE-striking normal fault, and included a significant component of dextral strike-slip displacement.

The 04 October 1957 earthquake (Figure 11), the largest of the three shocks under consideration, occurred at essentially the same location as the 02 October event, west of the 06 October earthquake. Its mechanism exhibits a combination of strike-slip and thrust displacements. One nodal plane strikes ENE and dips moderately south, the other strikes approximately NNE and dips approximately 60°W. Slip is dextral on the ENE striking plane and sinistral on the NNE plane. The ENE plane is nearly parallel to a fault just north of the Paria Peninsula coastline identified on reflection seismic lines (Speed *et al.*, 1989). No faults parallel to the

NNE nodal plane are known. Thus, we choose the ENE plane as the active fault plane. Motions on this plane, dextral strike-slip and a component of northward directed thrusting, are compatible with the analysis of RUSSO *et al.* (in press) which demonstrated that this region is the locus of E-W dextral strike-slip and NW-SE directed thrusting at shallow crustal levels. This result is important because the earthquake is the largest magnitude shallow earthquake in the plate boundary zone, and, as such, its displacements are most likely to represent the true long-term plate boundary displacements. On the basis of this mechanism and other smaller magnitude earthquake mechanisms, we can say that such motions are clearly collisional with a strong component of southeastward (dextral) obliquity.

The much smaller magnitude ($M_{PAL} = 5.1$) 06 October 1957 earthquake occurred to the east of the other two events. The earthquake's mechanism is the most poorly constrained of the seven new mechanisms. The event is a normal faulting earthquake, with one plane fairly clearly constrained to have NW strike and moderate SW dip, but the strike of the second plane can vary from NW (pure normal faulting) to NNE (mixed normal and strike-slip). All the possible nodal planes of the 06 October event strike at considerable angles to both the chosen active fault planes of the 02 and 04 October earthquakes, and to recognized faults in the local near-surface geology. As a result, choosing an active fault plane is not possible. This earthquake may be understood, however, in that it does have a relationship with the main shock of the series, its immediate predecessor, the 04 October event. The main shock occurred to the west, with dextral motion along a plane striking slightly north of east. This slip could reasonably be expected to cause extension parallel to the fault within material at the eastern terminus of the fault region (see Figure 12). Calculations of stress trajectories induced by strike-slip displacement (CHINNERY, 1966) and field and seismic observations (LANGER and BOLLINGER, 1979) of strike-slip faults confirm this possibility.

Thus, the October 1957 earthquake series may have proceeded as follows (Figure 12). The 02 October earthquake failed with a combination of normal and dextral slip on a NE-striking fault within an area of the plate boundary zone known to have both recently active normal faulting and dextral strike-slip. The displacement associated with this motion most likely caused extension in a zone of rock at the western terminus of the active rupture surface. This extension allowed the release of ambient plate boundary zone stresses, which resulted in the large 04 October shock. This earthquake is typical of mixed thrust-and-strike-slip earthquakes that occurred more recently in the western Gulf of Paria, as mentioned above. These events reflect unpartitioned Caribbean–South America plate boundary zone deformation. The 04 October event likewise caused extension along the southside of its eastern terminus, as discussed above, and the 06 October earthquake occurred as a result.

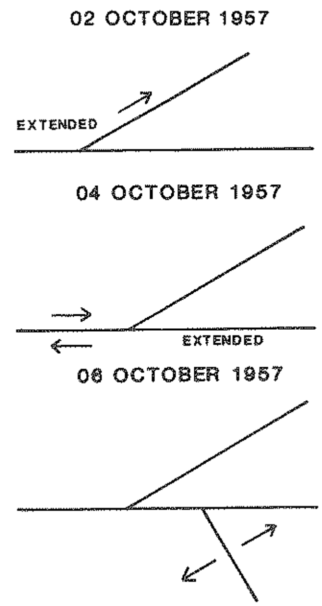

Figure 12
Hypothetical mechanism of the three events. The 02 October event occurs on a splay of the main E-trending fault. Extension in the region to the west of the slipped zone combines with ambient plate boundary compressive stresses to cause the main shock, 04 October. The 06 October event is a much smaller magnitude normal faulting event related to extension at the eastern terminus of the main shock slipped zone.

The 25 December 1957 earthquake ($h = 22$ km, $M_{PAS} = 5.8$; Figure 11) took place considerably south of the three October 1957 earthquakes, in the northwestern Gulf of Paria. This is also a very seismically active area in recent times. The earthquake's mechanism exhibits a combination of strike-slip and dip-slip, similar to that of the 02 October 1957 earthquake, but for this event the dip-slip component is paramount relative to a minor component of strike-slip displacement. Potential nodal planes strike NNE and NW-SE. The dip of the NW-SE plane is somewhat unconstrained, but dips shallowly NE; the NNE-striking plane is close to vertical. Faults with NW-SE strike are known within the area (BELLIZZIA, 1976; ROSSI *et al.*, 1987), but are generally assumed to dip steeply. If we assume that the NNE-striking plane (the plane best-constrained by data) is the fault plane, then this earthquake may represent brittle normal faulting associated with loading of South America by overriding Caribbean terranes (SPEED and FOLAND, 1991; RUSSO *et al.*, in press). A second possibility is that the earthquake occurred on a reactivated normal fault in the continental South America basement underlying this area. Such faults are presumed to have formed during Triassic rifting of North and South America (KLITGORD and SCHOUTEN, 1986).

Conclusions

Our study of the historical seismicity of the southeastern Caribbean and the South America–Caribbean plate boundary zone has yielded the following conclusions.

The frequent shallow and intermediate depth earthquakes in the vicinity of the Paria Peninsula, Venezuela, is a persistent feature of the Caribbean–South America plate boundary zone. Mechanisms of earthquakes in this zone show that a complex combination of dextral strike-slip, southeast directed thrusting, and normal faulting, recognized in the recent seismicity, is also characteristic of the historical earthquakes. The earthquakes demonstrate that oblique collision between the Caribbean and South American plates is occurring, and that the plate boundary zone may have a strong deformation gradient between the Araya-Paria isthmus of Venezuela and the island of Trinidad. In particular, this gradient can be observed in the degree of strain partitioning, which decreases from west to east, and in the activity of dextral strike-slip on E-striking faults, which also diminishes from west to east. However, the deformation gradient is even more apparent when we note that Trinidad and the eastern Gulf of Paria have not been subjected to large magnitude or frequent earthquakes throughout the period of seismic recording. Thus, either this region is a seismic gap, or tectonic motions are not causing earthquakes here.

A seismic gap exists along the Lesser Antilles arc between St. Vincent and Grenada throughout the time of seismic recording. Infrequent intermediate depth (70–150 km) earthquakes, recent and active volcanism, gravity anomalies, and deformation observed at exposed portions of the arc all indicate the presence of a subducting or recently detached Atlantic oceanic slab beneath this portion of the arc, however. Thus, it is unclear whether this gap is due to lubrication of the interplate interface by subducted accretionary prism sediments, to a change in convergence rate or direction along the arc, or to intraplate deformation of the arc and backarc which diffuses the expected interplate slip.

During the historical and recent eras, subduction along the Lesser Antilles arc from north of Martinique to south of Grenada has been characterized by frequent relatively large magnitude slab and intraplate deformation earthquakes, and by a corresponding paucity of large magnitude interplate slip earthquakes. This mode of seismic strain release is unusual for a subduction zone, indicates that the plates are relatively decoupled, and may be related to the slow rate of convergence between the Caribbean plate and subducting Atlantic lithosphere, or to the great age of the subducting lithosphere, which should have a large negative buoyancy.

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Appendix

• *11 May 1992*

The final solution (11.33°N; 60.05°W) is based on a data set of 17 stations ($\sigma = 3.3$ s) with the depth constrained at 5 km. All efforts at using a greater depth yielded a significantly larger residual.

• *08 August 1923*

This event is given as intermediate depth by both the NEIC (110 km) and the ISS (191 km). In the absence of any available record, we used a data set of 18 *P* times, and found an excellent solution ($\sigma = 2.07$ s) at a very shallow depth (5 km). Constraining the depth to its ISS value increased σ to 4.07 s, but more significantly, the first arrival at SJG is then more than 12 seconds early. We conclude that the earthquake is probably shallow. Our relocated epicenter is almost three degrees east of the ISS location.

• *01 February 1926*

Inversion of a data set of 14 *P* and *S* arrival times yielded a depth of 147 km ($\sigma = 1.6$ s) and moved the epicenter 3°E with respect to the ISS location. The relocated hypocenter is in the Paria Cluster, northwest of Trinidad.

• *27 September 1928*

The quality of this relocation (using 19 *P* and *S* times) is good ($\sigma = 2.8$ s). The earthquake epicenter moved approximately 100 km to the southeast. We verified that the event could not be deep; the minimum-residual depth is 30 km.

• *17 January 1929*

We relocate this earthquake on the basis of 26 arrivals (21 P and 5 S), with a resultant σ of 2.85 s. The arrival times at FDF, the nearest station, are inconsistent with an intermediate or deep focus. The data do not constrain the depth, however. The minimum-residual depth is 20 km.

• *10 April 1935*

This event relocates 70 km southwest of the ISS location, within the Paria Cluster at a depth of 102 km. The quality of the solution is good ($\sigma = 2.2$ s on 22 times).

• *11 August 1939*

This event is listed as an "undetermined shock" in the ISS, but there exists a CGS listing (11.8°N; 67.2°W). We were unable to find a solution given the available dataset of 26 arrival times reported by the ISS.

• *14 October 1939*

This earthquake relocated approximately 100 km to the east of the ISS location, into the Paria Cluster. The earthquake could not be deep.

• *27 February 1940*

This event occurred to the south of our study area, in the Orinoco River delta. There is some discrepancy regarding its longitude between the ISS location (60.8°W) and GUTENBERG and RICHTER's (62°W). On the basis of 15 P and S arrivals, we prefer the ISS location.

• *23 June 1940*

Inversion of 14 P and S times gives an excellent solution at the extreme western end of the study area, in basic agreement with the ISS location. The earthquake cannot be deep.

• *06 July 1940*

Although the epicenter of this event is not significantly altered, its depth is increased substantially (ISS depth = 96 km) to 160 km on the basis of 32 P times and 10 S times. The earthquake clearly occurred in the Lesser Antilles Benioff zone west of St. Vincent.

• *06 May 1942*

Two earthquakes occurred on 06 May 1942 at 90 minutes interval. The ISS lists them at the same location. However, the first event relocates 150 km east of the ISS location, to the Carupano Basin east of Margarita. The second earthquake relocates at the approximate ISS epicenter in the Cariaco Trough. We constrain the depths of both earthquakes to 10 km on the basis of minimizing σ. The events are probably unrelated.

• *06 February 1944*

This event is given as an "undetermined shock" in the ISS, although there exists a CGS entry (10°N; 62°W). The best solution is in the Paria Cluster at a depth of 50 km.

• *23 December 1945*

This event is reported by the ISS at 65 km depth in the Gulf of Paria. We find a very similar epicenter, but the data set of 39 P and 20 S arrival times yields a hypocentral depth of 28 km. The solution is very good, with $\sigma = 1.8$ s.

• *21 May 1946*

The relocation does not move the epicenter significantly and suggests a shallow focus ($d = 49$ km), controlled by the arrival at FDF.

Focal mechanism: ISS listings constrain one plane (Figure A1). We use the following constraints UPP: $SV > SH$, $SV > 0$; TUC: SV > 0, SV > SH; BUR: $SV > 0$ to obtain the preferred mechanism $\phi = 325 \pm 15°$; $\delta = 70 \pm 2°$; $\lambda = 85 \pm 15°$.

M_s measurements at TUC (6.0) and UPP (5.8) are significantly lower than published $M_{\mathrm{PAS}} = 7.0$ (GUTENBERG and RITCHER, 1954).

• *31 July 1946*

Inversion of 37 P and S times moved the earthquake southwest approximately 20 km, and the depth converged to 56 km.

• *21 August 1949*

On the basis of 26 P and S times, the relocation converges in the Paria Cluster at a depth of 162 km, with a residual of less than 2 s.

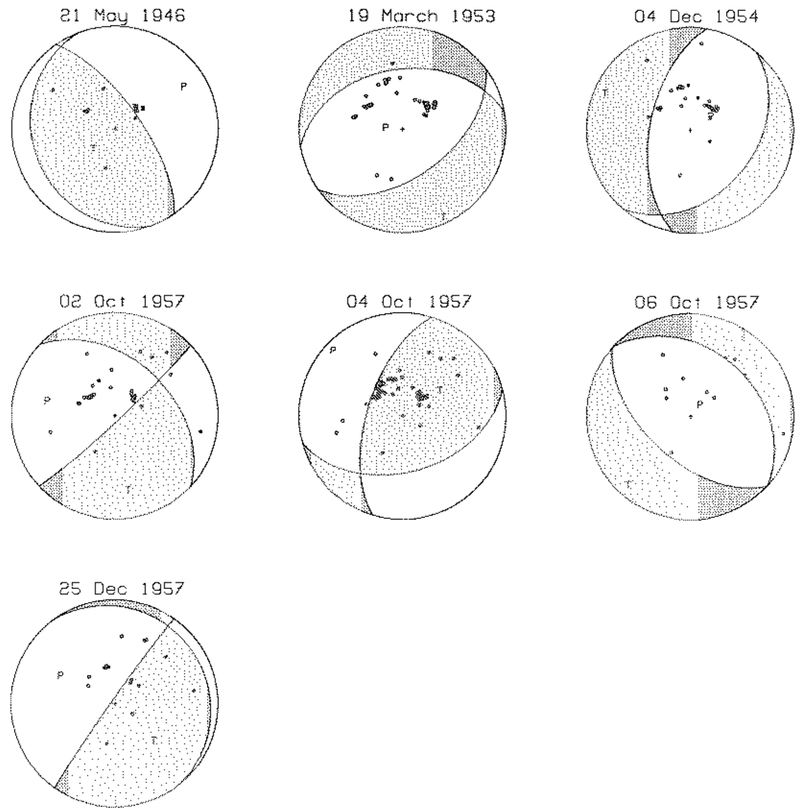

Figure A1

First-motion polarity data for the focal mechanisms constrained in this study. Data are as reported by ISS and also from picks made by the authors. Compressional arrivals are filled circles, dilatations unfilled. Equal-area lower hemisphere projections, compressional quadrants shaded; P = pressure axis, T = tension axis.

- *20 April 1951*

Relocation did not alter this epicenter substantially: a location in the Paria Cluster at a depth of 10 km optimized the residuals.

- *23 December 1951*

We relocate this event into the Benioff plane near the northern tip of Martinique. The dataset of 43 P and S times converged to 177 km depth with a residual of 1.9 s.

• *24 August 1952*

The relocation moves the epicenter 90 km west-southwest to a position at the southern end of the Lesser Antilles arc platform south of Grenada. The hypocentral depth of 170 km indicates that the earthquake occurred in the Lesser Antilles Benioff zone.

• *31 December 1952*

We propose a depth of 40 km for this event, on the basis of 39 P and S times ($\sigma = 1.96$ s). The epicenter is moved 30 km southwest of the ISS location. This depth is upheld by estimates from $pP - P$ at CLE, OTT, SHF and FAY ranging from 43 to 54 km.

• *19 March 1953 St. Lucia*

Relocation using 164 P and S times listed in the ISS Bulletin converges to 14.00°N; 61.24°W at a depth of 133 km, with a standard deviation $\sigma = 2.14$ s. This depth is confirmed by our observations of $[sS - S]_{SH}$ and $sP - P$ at PAS, and $[sS - S]_{SH}$ and $pP - P$ at UPP, which indicate a hypocentral depth of 133 km.

Focal mechanism: As mentioned by SYKES and EWING (1965), this earthquake clearly involved normal faulting, but the first motions reported to the ISS fail to constrain its mechanism, leaving the azimuth of the tension axis indeterminate. We obtained the mechanism shown on Figure 9 and in Figure A1: ($\phi = 260 \pm 5°$; $\delta = 41 \pm 5°$; $\lambda = 290 \pm 5°$) by adding to the first motion data the following constraints: LPB: $SV < 0$; PAS: $SV < 0$, $SH < 0$, $SH > sSH$; UPP: $SV < 0$, $sSH < 0$, $SV > SH$, $SV > sSV$, $sSH > SH$; TUC: $SH < 0$, $sSH > 0$.

The mantle magnitude at PAS is $M_m = 7.01$, which, corrected for focal mechanism and depth, leads to a seismic moment $M_0 = 1.6 \times 10^{27}$ dyn-cm.

• *25 June 1953*

We relocate this earthquake approximately 80 km north of its ISS location. Forty-two P and S arrival times indicate a hypocentral depth of 121 km, very similar to the ISS listing of 128 km. The standard deviation, σ, is 1.46 s.

• *04 December 1954*

Relocation of this event does not move its epicenter significantly. Inversion of 89 P and S arrivals yields a depth of 46 km. This is in fair agreement with our observations of pP phases at PAS and SJG, which indicate a depth of 57 km. The standard deviation for this event is 2.17 s.

Focal mechanism: The mechanism (Figure A1) was constrained using local data from the Caribbean network (dilatations at TRN and SLI). The reported compressional arrival at KIM (unavailable for inspection) is incompatible with the rest of the data. While the strike and dip of one fault plane are well constrained ($\phi = 190 \pm 5°$; $\delta = 58 \pm 1°$), the slip vector may range from 222 to 285°.

The TUC horizontal records yield $M_s = 6.1$.

• *02 October 1957*

Relocation of this event does not move its epicenter significantly.

Focal mechanism: ISS listings are grossly inconsistent (Figure A1). Local stations of the Trinidad local network (TRN, GRE, SVT compressional; BRB dilatational) constrain the mechanism to $\phi = 47 \pm 5°$; $\delta = 86 \pm 2°$; $\lambda = 225 \pm 7°$. A magnitude $M_s = 5.5$ was measured at PAS.

• *04 October 1957*

Relocation of this event does not move its epicenter significantly. The dataset of 79 P and S arrival times indicates a shallow depth of 6 km, with a standard deviation of 1.51 s.

Focal mechanism: One fault plane ($\phi = 195 \pm 5°$; $\delta = 62 \pm 1°$) is relatively well constrained by compressions at TUC and dilatations at TAC and SJG (Figure A1). Despite good local data (FDF, SVT, GRE and TRN compressional; UAV dilatational), the slip remains largely unconstrained, and can vary between 40 and 100°. The only usable S constraints ($SV > 0$ and $SV > SH$ at PAS) have only marginal resolution in this geometry.

$M_s = 6.65$ was measured on the PAS 30-90 instrument.

• *06 October 1957*

Relocation of this event does not move its epicenter significantly.

Focal mechanism: Compressional arrivals at near stations (TRN, GRE, SVT, BRB) and dilatational arrivals at SJG and HUA, among others, clearly demonstrate that the mechanism is normal faulting, but do not constrain even one plane well (Figure A1). Additional constraints from S arrivals at SJG ($SH < 0$) and HUA ($SV < 0$; $SV/SH \geq 1$) help somewhat. The resulting ranges of focal angles are: $\phi = 132 \pm 6°$, $\delta = 44 \pm 12°$, $\lambda = 240 \pm 30°$. The possible mechanisms are considerably different from those of the previous events in the October 1957 series.

• *25 December 1957*

The epicenter is not moved significantly by the relocation; the hypocentral depth is 22 km, resulting from inversion of 58 P and S times, with a standard deviation of 1.91 s.

Focal mechanism: First arrivals constrain one plane ($\phi = 215 \pm 5°$; $\delta = 87 \pm 3°$) (Figure A1). Additional resolution is available only from the S waves at PAS: $SV > 0$; $SH < 0$; $SV/SH \geq 1$. The constraint on the slip vector is not strong, and results in a possible range of slip, $\lambda = 100 \pm 20°$. This event is relatively large for the region, ($M_{\mathrm{PAL}} = 5.9$; we measured $M_s = 5.8$ at PAS).

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