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A LARGE NORMAL-FAULT EARTHQUAKE IN THE OVERRIDING WEDGE OF THE LESSER ANTILLES SUBDUCTION ZONE: THE EARTHQUAKE OF 8 OCTOBER 1974

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ABSTRACT

A large earthquake (M_S = 7.1 to 7.6) occurred in the overriding wedge of the Caribbean plate on 8 October 1974. Hypocenters of locally recorded aftershocks suggest that the main shock rupture extended from about 40 km, immediately above the inferred thrust interface between the Caribbean and American plates, to shallow crustal depths of about 10 km. The focal mechanism solution and aftershock hypocenters taken together imply that the earthquake occurred on a southeast-dipping normal fault that strikes NNE to NE, transverse to the regional strike of the northern Lesser Antilles arc. The location and focal mechanism of the 1974 shock are thus similar to those of small earthquakes occurring in the Adak Canyon region of the central Aleutians and studied by LaForge and Engdahl (1979). Both regions, the northern Lesser Antilles and the central Aleutians, are zones of oblique plate convergence, and it is possible that wedge normal faulting, transverse to the island arc, is a consequence of this oblique convergence. Alternatively, the large normal-fault earthquake in the northern Lesser Antilles may have been a consequence of the resistance of the aseismic Barracuda Ridge, on the American plate, to subduction beneath the Caribbean plate.

The focal mechanism of the 1974 earthquake indicates that the shock should not have released a significant amount of the elastic strain that is thought to have been accumulating along the thrust boundary of this segment of the Lesser Antilles arc since the great earthquake of 1843.

INTRODUCTION

A large earthquake (M_S = 7.1, Pasadena; M_S = 7.5, USGS; M_S = 7.6, Berkeley) occurred in the northern Lesser Antilles on 8 October 1974. The epicentral region of the earthquake is near Barbuda, north of Antigua (Figure 1). The shock is sometimes referred to as the "Antigua" earthquake of 1974 because the costliest damages were sustained on that island (Tomblin and Aspinall, 1975). Both Antigua and Barbuda are part of the limestone Caribees (Martin-Kaye, 1969), the extinct volcanic chain of the pre-Miocene Lesser Antilles arc. The presently active volcanic chain of islands in the region of the earthquake is about 100 km west of the pre-Miocene arc.

The primary tectonic features of the epicentral region are determined by the subduction of the Atlantic seafloor beneath the Caribbean plate (Molnar and Sykes, 1969), and most crustal structures strike northwest (Case and Holcombe, 1980), parallel to the Puerto Rico trench and the present-day volcanic arc. There are, however, several marked east-west offsets in the bathymetric contours that define the inner wall of the Puerto Rico trench (e.g., near 16.5°N, 17.3°N, and 18.0°N, Figure 1), and Tomblin (1972) has suggested that some of these may be caused by faults that are transverse to the island arc. The compilation of Case and Holcombe

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(1980) shows northeast-striking normal faults associated with the offset in the inner trench wall near 16.5°N. The principal conclusions of the present study are that the causative fault of the 1974 Lesser Antilles earthquake was also a northeast-striking normal fault, and that the earthquake occurred in the overriding wedge of the Caribbean plate, immediately above the thrust interface between the American and Caribbean plates.

The 1974 earthquake is also of interest because it occurred in a region that has been characterized (Kelleher *et al.*, 1973; McCann *et al.*, 1979; Dorel, 1981) as having high potential for a great earthquake in the near future. In addition to the region being an active zone of plate convergence, a great earthquake occurred here in 1843, and a large, possibly great, earthquake occurred in 1690. Kelleher *et al.* (1973) infer from the high intensities produced by the 1843 earthquake that the shock had a shallow focal depth. Also, the 1843 shock was destructive over a long region parallel to the arc; the 1843 earthquake had a meizoseismal region extending approximately from 15.25°N to 17.5°N (Robson, 1964), substantially larger than that of the 1974

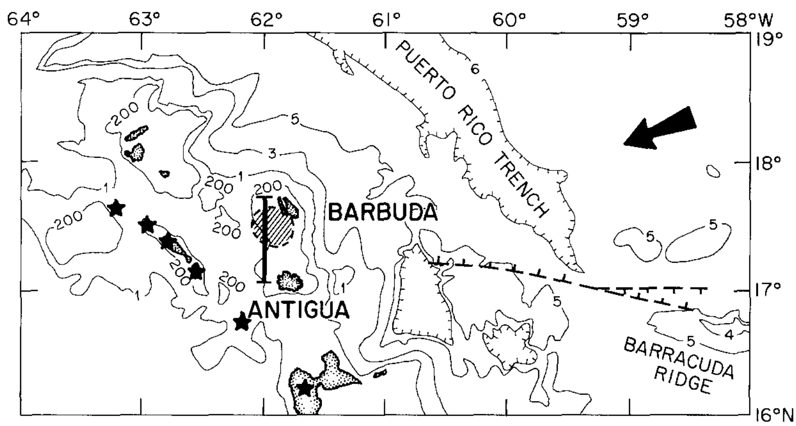


FIG. 1. Epicentral region of the Lesser Antilles earthquake of 8 October 1974. The *hatched region* is the main aftershock zone of the 1974 earthquake. Stars denote Quaternary volcanic centers. Deepest bathymetric contours are in kilometers. Two-hundred meter contour is also shown. Note offsets in the inner wall of the trench (3 km contour) near 16.5°, 17.5°, 18.0°, and 18.75°N. Thick line along 62°W is position of refraction profile shown in Figure 9. Mapped scarp of Barracuda ridge shown on inner wall of trench is from Case (1975). The broad arrow indicates the direction of motion of the American plate relative to the Caribbean plate (Sykes *et al.*, 1982).

earthquake (Tomblin and Aspinall, 1975). Kelleher *et al.* (1973) treat the 1843 earthquake as a great thrust fault earthquake. The rate of plate convergence of about 3.5 cm/yr (Sykes *et al.*, 1982) is consistent with the hypothesis that the 1843 earthquake was caused by the release of compressive strain that had accumulated on the plate interface since the earthquake of 1690 and implies that sufficient strain could have accumulated since 1843 to produce another great earthquake.

Location of Aftershocks

We shall discuss two distinct sets of aftershocks. The first set consists of small locally recorded aftershocks registered by a temporary microearthquake network installed by Lamont-Doherty Geological Observatory and by the U.S. Geological Survey (USGS) (Figure 2). The second set consists of teleseismically recorded aftershocks. Operation of the local network was plagued by difficulty in obtaining accurate absolute timing, and we decided that we could obtain the most reliable hypocenters for locally recorded shocks by using S-P time intervals rather than the customarily used P- and S-wave arrival times. The set of hypocenters determined with local data includes only aftershocks that were small enough to have S-wave arrivals clearly identifiable in the P-wave codas at the stations on Barbuda and that occurred during the time period (15 to 28 October) for which the network was most complete. The teleseismically recorded aftershocks are, in general, not as precisely located as the locally recorded aftershocks, but the set of teleseismically recorded aftershocks includes all aftershocks from the entire month following the main shock that were large enough to be recorded at teleseismic distances. We also use teleseismic locations from a broad area around the 1974 epicenter to define the regional Benioff Zone beneath the northern Lesser Antilles. Arrival times or S-P time intervals from stations of the permanent seismograph network of the University of the West Indies (UWI) were used in the location of both sets of hypocenters.

Hypocenters of locally recorded earthquakes were determined using a computer program written by R. Crosson (Crosson, 1976). In general, this program may be

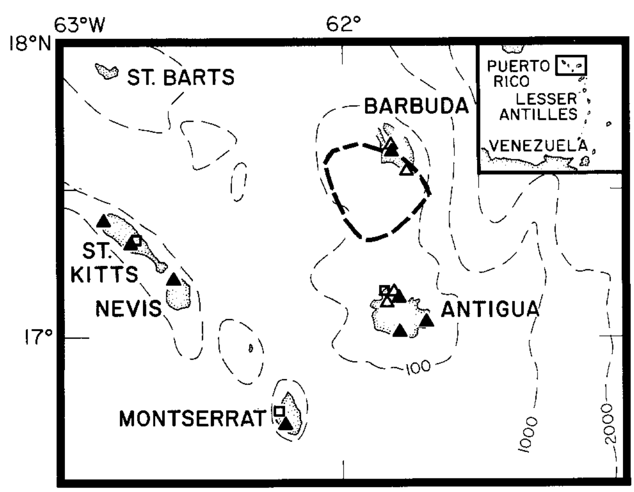


FIG. 2. Positions of local seismograph stations used in study. Solid triangles are temporary L-DGO stations; open triangles, temporary USGS stations; open squares, permanent UWI stations. USGS also operated a station off the northwest corner of the figure and at identical sites to L-DGO stations on Nevis and St. Kitts. Heavy dashed line surrounds main aftershock zone, or focal region, inferred from main shock hypocenter and locally recorded aftershocks. Contours are in fathoms (1 fathom = 1.83 m).

used to invert arrival-time data from a number of earthquakes to simultaneously obtain the earthquake hypocenters, a revised model of regional velocity structure, and station delays. In our case, all rays from the sources to a given station have nearly the same ray path, and the velocity and station delays are almost linearly dependent. We, therefore, fixed the velocity model in the computation to be that of Table 1, which is derived from a crustal velocity model proposed for this region by Officer *et al.* (1959), with a depth to the M discontinuity assumed to be 30 km. In addition, the delay at a station on the northwest corner of Antigua was constrained to −0.08 sec. This is the average S-P delay, with respect to the model of Table 1, observed for two teleseismically recorded aftershocks, the location of which will be discussed later. The 30 best-recorded aftershocks were then simultaneously located and S-P station delays computed for the remaining stations (Figure 3). In effect, we used Crosson's (1976) program as a local joint hypocenter determination program, with the S-P delay restrained at one station to make the problem nonsingular. The locally recorded hypocenters are estimated on the basis of the square root of the mean of the squares of the residuals (rms) to be precise to within 20 km at a 90 per cent level of confidence. We estimate that our best-determined hypocenters (the largest symbols in Figure 3) are precise to within 10 km at a 90 per cent level of confidence.

For teleseismically recorded earthquakes, we used the current version of the joint hypocenter determination (JHD) program of Dewey (1971) on 15 well-recorded

TABLE 1

VELOCITY STRUCTURE USED TO DETERMINE HYPOCENTERS OF LOCALLY RECORDED AFTERSHOCKS

$V_p/V_s = 1.76$

P-Wave Structure		Structure Used for *S-P* Times	
Depth (km)	Velocity (km/sec)	Depth (km)	Pseudovelocity (km/sec)
0	4.8	0	6.1
3.5	6.25	3.5	8.0
10.0	6.50	10.0	8.3
30.0	8.0	30.0	10.2

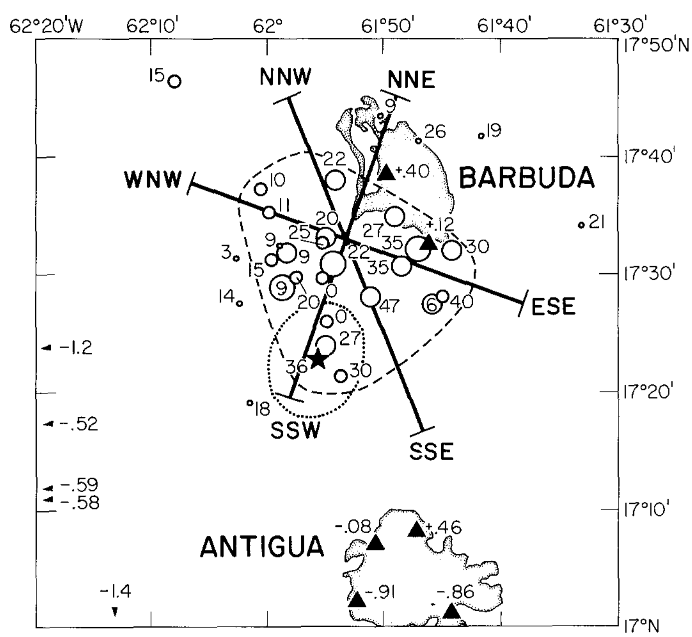


FIG. 3. Epicenters of locally recorded aftershocks. Larger symbols denote more accurate locations; numbers near these symbols are depths in kilometers. Numbers near stations (triangles) are computed *S-P* delay. Star and dotted ellipse are teleseismically determined main shock epicenter and corresponding 90 per cent confidence ellipse. The dashed line enclosing most of the aftershocks defines the main aftershock zone used in other figures.

earthquakes to estimate both the variances of the different phase types (we used *P*, *pP*, and regionally recorded *S* waves) and the station delays for each phase. Eleven of the JHD-relocated earthquakes were in the 1974 sequence and the four other shocks occurred approximately 50 km to the northeast of the 1974 sequence. The JHD-computed station delays and variances were then used in a single-event teleseismic location program to determine the hypocenters of all shocks recorded by the International Seismological Centre (ISC) in 1964 to 1974 and lying in a 300-km-long segment of the northern Antilles arc centered on the 1974 epicenter.

The calibration event for the teleseismic joint hypocenter determination was the aftershock of 14 October 1974, 040514.0 UTC, which was recorded by local stations (most crucially, by a station on Barbuda) during a time when their clock corrections were accurately known. *S* arrivals for this aftershock were not recorded by enough local stations that the shock could be located using the method described previously for locally recorded aftershocks, and its hypocenter was taken to be simply the hypocenter computed by a single-event location method using both the local and teleseismic arrival times. There is thus a possibility of some systematic mislocation between the locally recorded and teleseismically recorded earthquakes, although, as discussed below, the aftershock zones defined by the two sets of earthquakes turned out to be very similar. We note that the station delays computed in the teleseismic JHD are strictly appropriate only for sources in the region in which the JHD-

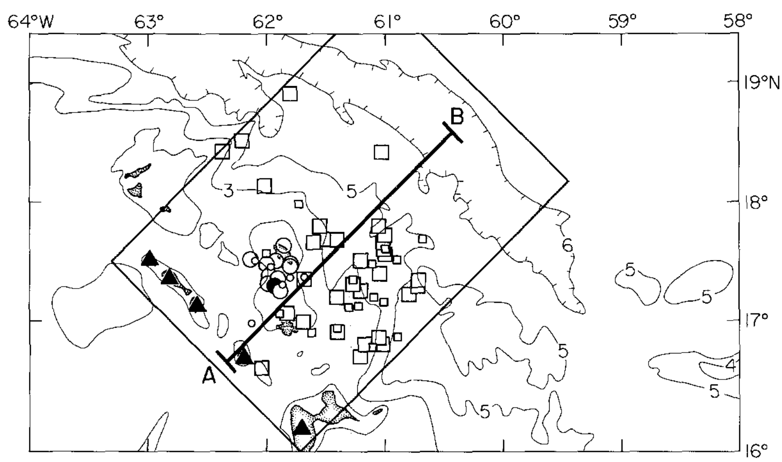


FIG. 4. Positions of events teleseismically recorded during the period 1964 to 8 November 1974 and relocated using joint hypocenter determination with estimated precision of 30 km or better. Almost all of these shocks have magnitudes (m_b, ISC) greater than 4.0. Open circles are aftershocks occurring within 31 days of the 1974 earthquake (solid circle). Epicenters denoted by large symbols are estimated to be accurate to within 20 km. Orientation and limits of vertical cross section *A-B* of Figure 5 are also shown. Contours are in kilometers with the exception of the shallowest contour, which corresponds to 200 m. We relocated all shocks in the time period that were originally assigned to within 25 km of the box. However, events relocated outside the box are not plotted.

located shocks occurred. Because of the likely rapid variations of seismic velocity in the region of the subducting slab, it is likely that location of events some distance away from the 1974 source will still be biased. For example, as discussed by Dewey and Spence (1979), the location method we used may overestimate focal depths of earthquakes occurring closer to the trench than the 1974 source.

We computed 90 per cent confidence ellipses on pairs of teleseismically determined hypocenter coordinates following the method of Evernden (1969), with the variances of the arrival time data assumed to be those computed in the teleseismic JHD. For 67 per cent of the earthquakes originally considered, the semi-axes of the confidence ellipses of the epicenter coordinates were less than 30 km in length, and these events are plotted in Figures 4 and 5. The focal depths of most events are estimated to be determined to approximately the same precision as the epicentral coordinates. The most precisely determined of the teleseismically recorded earthquakes, including the main shock of 8 October 1974, are estimated to be accurate to approximately 10 km at a 90 per cent level of confidence. (Lists of both locally and teleseismically recorded hypocenters are available from the authors upon request.)

Distribution of Aftershocks

The distribution of shocks determined with the locally recorded data is similar to the distribution of teleseismically recorded shocks (Figures 3 and 4). Because the locally recorded shocks are estimated to be more accurately located, we will base our inferences of the orientation of the 1974 fault plane principally on these shocks. We take the similarity of the distribution of locally recorded and teleseismically recorded shocks as evidence that our set of 30 locally well-recorded shocks is representative of the entire aftershock sequence.

A vertical section perpendicular to the arc shows that most of the teleseismically

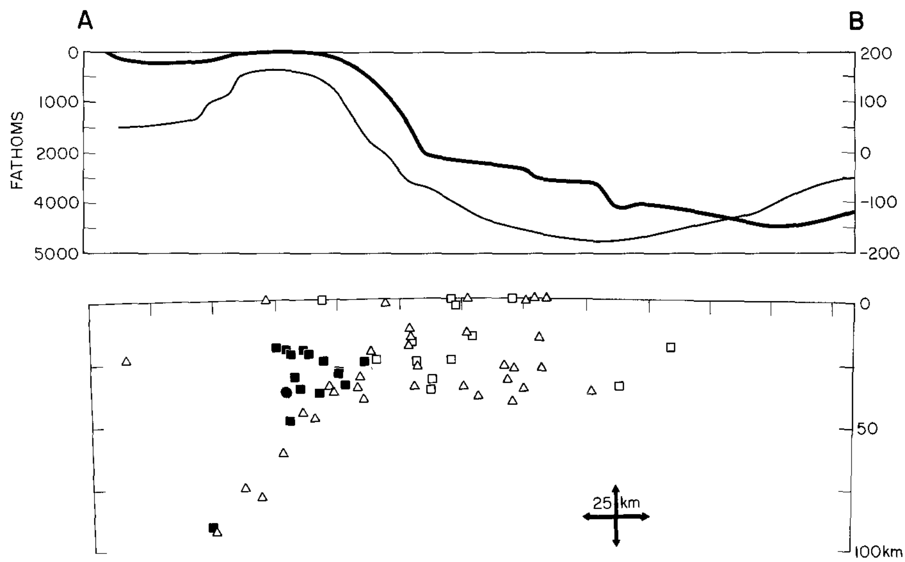


FIG. 5. Vertical cross section *A-B* of teleseismically recorded events shown in box in Figure 4. Note that aftershocks (solid symbols) generally lie above the downgoing seismic zone defined by earlier earthquakes (open) and that the 1974 earthquake occurred in a region that had been seismically quiescent for at least one decade prior to the shock. Squares are projected from northwest of line *A-B*; triangles, from the southeast. Circle is main shock. Plot above vertical cross sections shows bathymetry (heavy line) and gravity (light line) (Bowin, 1976) along center of section.

recorded (Figure 5) and locally recorded (Figure 6a) aftershocks occurred above the regional Benioff Zone defined by the teleseismically located hypocenters. The tendency for aftershocks of the 1974 zone to lie above the regional Benioff Zone has been noted previously by Tomblin and Aspinall (1975) from hypocenters determined by the ISC and UWI. The main shock and most of the aftershocks evidently occurred in the leading edge of the Caribbean plate, or what we are here calling the "overriding wedge," above the Benioff Zone, which marks the upper portion of the American plate. Because most of the shocks that we have used to define the position of the Benioff Zone occurred south and east of the 1974 source region (Figures 4 and 5), our inference on the position of the American plate beneath the 1974 source zone is based on the hypothesis that a subducted American plate does in fact exist beneath the 1974 source and is at the same depth and distance from the Puerto Rico trench as it is immediately to the southeast. The following observations support the existence and location of the American plate immediately beneath the 1974 source region:

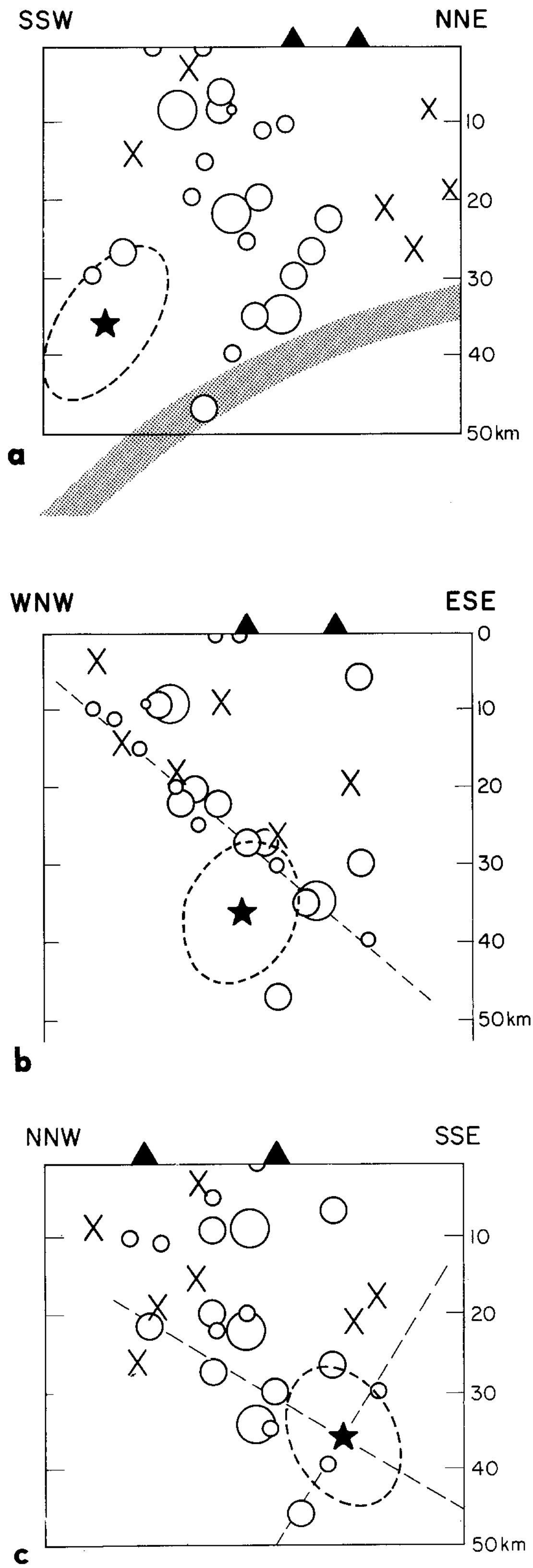
1. Seismic reflection profiles (Chase and Bunce, 1969; Marlow *et al.*, 1974) imply that the American plate is being subducted beneath the Caribbean plate trenchward of the 1974 source region. In addition, focal mechanisms of some earthquakes (Molnar and Sykes, 1969; Tomblin, 1975) are consistent with underthrusting of the American plate beneath the Caribbean plate seaward of the 1974 source region.
2. The Quaternary volcanic arc of the Lesser Antilles extends smoothly from the south to approximately 150 km northwest of the line *AB* in Figure 4, suggesting that the associated Benioff Zone continues without gross distortion for at least this distance.
3. Low-level intermediate depth seismicity has been detected northwest of the 1974 source by the Northeastern Caribbean Network (W. McCann and K. Hurst, written communication, 1981) at the same depth and distance from the arc as the Benioff Zone south and east of the 1974 source. Sykes and Ewing (1965) located several shocks deeper than 100 km that occurred in the period 1950 to 1963 northwest of the line *AB* in Figure 4. We relocated two teleseismically recorded events at a focal depth of 99 km near 17.4°N, 62.5°W in the period 1965 to 1974: the semi-axes of their confidence ellipses slightly exceeded 30 km, and the shocks were therefore not included in Figures 4 and 5. The hypocenters of all these events are consistent with the Benioff Zone in Figure 5 continuing with little distortion to the northwest of line *AB* in Figure 4.

The lower limit of aftershocks to the 1974 earthquake coincides closely with the upper surface of the Benioff Zone, as projected from the southeast (Figures 5, 6a). Two earthquakes that occurred within 1 month of the main shock are reliably located within the inferred American plate at a significant distance from the hypocenters of the aftershock sequence. These are the shocks at focal depths of 48 and 90 km in Figure 5. Because of the general scarcity of earthquakes in the subducted American plate at depths of 40 km or greater in the vicinity of the 1974 source, there is a low probability that the events at depths of 48 and 90 km would have occurred independently of the 1974 main shock in the 1-month period following the earthquake. We, therefore, think that these earthquakes are probably aftershocks of the 1974 main shock—triggered in the subducted American plate by the main shock in the Caribbean plate. Several teleseismically or locally recorded aftershocks lie both in the principal zone of aftershocks and near the inferred upper boundary of the American plate (Figure 5, 6a). Neither the hypocenters of these aftershocks nor the position of the Benioff Zone are located with sufficient precision so that we can confidently say where these shocks occurred with respect to the plate interface.

Cross sections show that the locally recorded aftershocks tended to concentrate on a SE-dipping plane (Figure 6a, b). On the basis of the focal mechanism solution to the main shock, discussed in the next section, we will take this SE-dipping plane as the fault plane of the earthquake.

Fault Plane Solution

The 1974 Lesser Antilles earthquake was a consequence of rupture on a normal fault with a northeast strike, approximately transverse to the local trend of the gross island arc structure. *P*-wave first motions and *S*-wave polarization angles are shown


SSW
NNE
10
20
30
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50km
a
WNW
ESE
0
10
20
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40
50km
b
NNW
SSE
10
20
30
40
50km
c

in Figure 7; their position on the focal sphere was determined taking the teleseismically determined focal depth of 36 km and interpolating between take-off angles computed for depths of 15 and 40 km by Pho and Behe (1972).

A NW-dipping plane is well constrained by first-motion data from North American stations. However, the plane conjugate to the NW-dipping plane is not well defined by the first-motion data.

In Figure 7, we have plotted three among many possible choices for the east or southeasterly dipping fault plane. The plane striking N20°E and dipping 40°SE is one of a family of planes consistent with the compressional first motion at the nearby station SKI, as that datum is projected on our focal sphere. This plane is also parallel to the zone defined by most of the locally recorded aftershocks. The

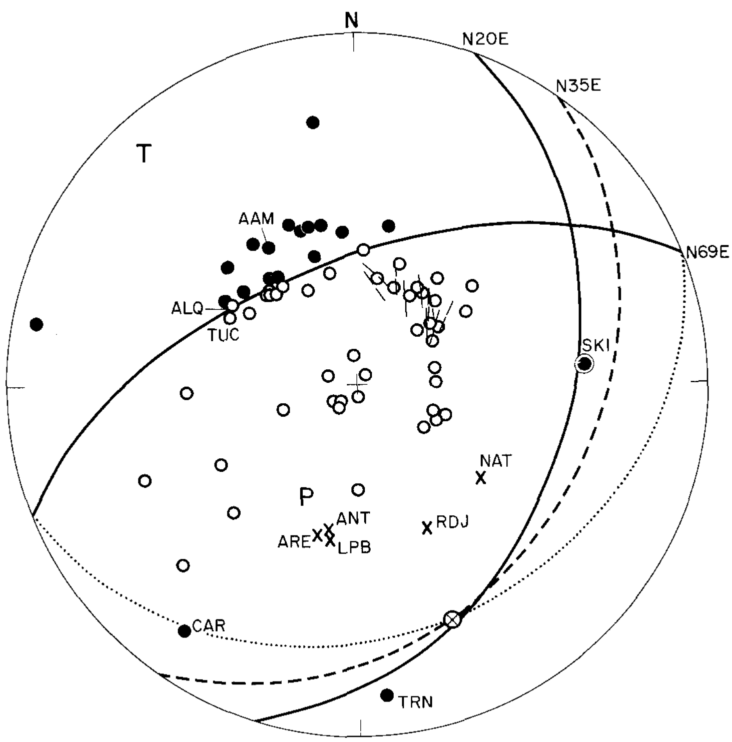

FIG. 7. Lower hemisphere plot of data used to determine focal mechanism of main event. Short lines are S-wave polarizations. Solid and open circles represent compressions and dilatations, respectively. Solid arcs are nodal planes drawn from the observed first motions (N69°E) and approximate dip and strike of the zone of aftershocks (N20°E). Dashed arc is nodal plane, orthogonal to N69°E plane, obtained inverting P-wave first motions (excepting the reading at SKI) and S-wave polarizations with the program of Dillinger *et al.* (1972). Dotted line is nodal plane corresponding to the best fit between observed and theoretical Rayleigh wave radiation patterns (Stein *et al.*, 1982a). X's represent stations at which small compressional first motions were identified that could not be reconciled with orthogonal P-wave nodal plane and with other P-wave first motions. X with circle denotes pole of N69°E nodal plane. P and T denote axes of greatest and least principal stresses, respectively, inferred from the orthogonal nodal plane solution.

FIG. 6. (a) to (c) Vertical cross sections of locally recorded aftershocks shown in Figure 3. The *stippled band* in (a) is the projection of the upper surface of the Benioff Zone inferred from Figure 5 and data from the northeastern Caribbean network (W. McCann, unpublished data). The light dashed line in (b) is the projection of the plane that strikes N20°E in Figure 7. The light dashed lines in (c) are the projections of the planes that strike N69°E in Figure 7. Sizes of symbols indicate quality of location (large, good location). X's are events outside main aftershock area in Figure 3. Star and ellipse are hypocenter and 90 per cent confidence ellipse of main shock from JHD location using teleseismic data. Triangles are stations on Barbuda. No vertical exaggeration.

plane striking N35°E and dipping 34°SE is determined if we neglect the reading at SKI (whose precise position on the focal sphere is quite uncertain because it is strongly affected by the local velocity structure), and invert both *P*-wave first motions and *S*-wave polarizations with the program of Dillinger *et al.* (1972). The plane striking N69°E and dipping 29°SE corresponds to the preferred focal mechanism of Stein *et al.* (1982a), which was based on the surface wave radiation pattern of the main shock. As the postulated strike of the SE-dipping nodal plane changes from N20°E counterclockwise to N69°E (Figure 6c), the plane is a progressively poorer fit to the distribution of aftershocks. This observation may argue against a fault striking N69°E as the cause of the earthquake, although we cannot rule out the alternate possibility that the flattened distribution of hypocenters shown in Figure 6b corresponds to a secondary tectonic structure rather than the principal fault plane of the earthquake. All three mechanisms shown in Figure 7 imply predominantly normal faulting and, assuming that most of the aftershocks occurred close to the main shock fault plane, if not right on the plane, the distribution of aftershocks favors one of the E- or SE-dipping planes as the causative fault.

All choices of a nodal plane orthogonal to the NW-dipping nodal plane are inconsistent with reliable compressional *P*-wave first motions recorded at several South American stations. In view of the consistently dilatational first motions near the center of the projection of the focal hemisphere (Figure 7), the inconsistent first motions at several South American stations do not affect our general conclusion that the Lesser Antilles earthquake involved predominantly normal faulting transverse to the trend of the island arc. However, the difficulty of fitting orthogonal nodal planes to the *P*-wave first motions is interesting for its own sake.

The *P*-wave seismograms that are inconsistent with orthogonal nodal planes are shown in Figure 8. Compressional readings at NAT and LPB seem unambiguous to us, both because of the clarity of the long-period signal and because the arrival times of the small amplitude compressional arrivals coincide with compressional *P*-wave arrivals on the short-period records. The other readings we regard as less certain. For example, our interpretation of a compressional first motion at ANT is based on a compressional first motion on the noisy short-period record arriving at the same time as the compressional "bump" on the long-period record. We could not identify on records from other distances or azimuths a counterpart to the small compressional first motion seen at the inconsistent South American stations. For example, stations in North American near the projection of the northwest-dipping nodal plane had strong, impulsive, first motions on long-period records (Figure 8) that are corroborated by impulsive arrivals on the short-period records.

Several mechanisms have been proposed to explain nonorthogonal nodal planes. We can rule out for the Lesser Antilles earthquake apparent nonorthogonality due to interference between *P*, *pP*, and *sP* (Hart, 1978) because of the earthquake's appreciable focal depth ($\approx$36 km). The distribution of inconsistent first motions differs from those predicted for the examples of dislocations in anisotropic media studied by Kawasaki and Tanimoto (1981). The difference between the small, low-amplitude, first motion at the anomalous South American stations and the strong first motions at other stations does not seem consistent with the nonorthogonality being caused by the simple superposition of an explosive component on a predominantly double-couple focal mechanism (e.g., Solomon and Julian, 1974). Two possible causes of nonorthogonality that seem most reasonable to us as explanations for the nonorthogonality of the 1974 earthquake are an earthquake source involving simultaneous rupture on fault planes with different orientations (Jackson *et al.*, 1979) and severe distortion of ray paths by lateral velocity inhomogeneities south of the epicenter (Solomon and Julian, 1974).

Tectonic Implications of the 1974 Antiguan Earthquake

The most significant seismotectonic characteristics of the 1974 Antiguan earthquake are that it occurred in the crustal wedge immediately above the thrust interface between the overriding Caribbean plate and the subducting American plate and that it occurred on a normal fault striking at a large angle with respect to the local trend of the northern Lesser Antilles arc. Many characteristics of the

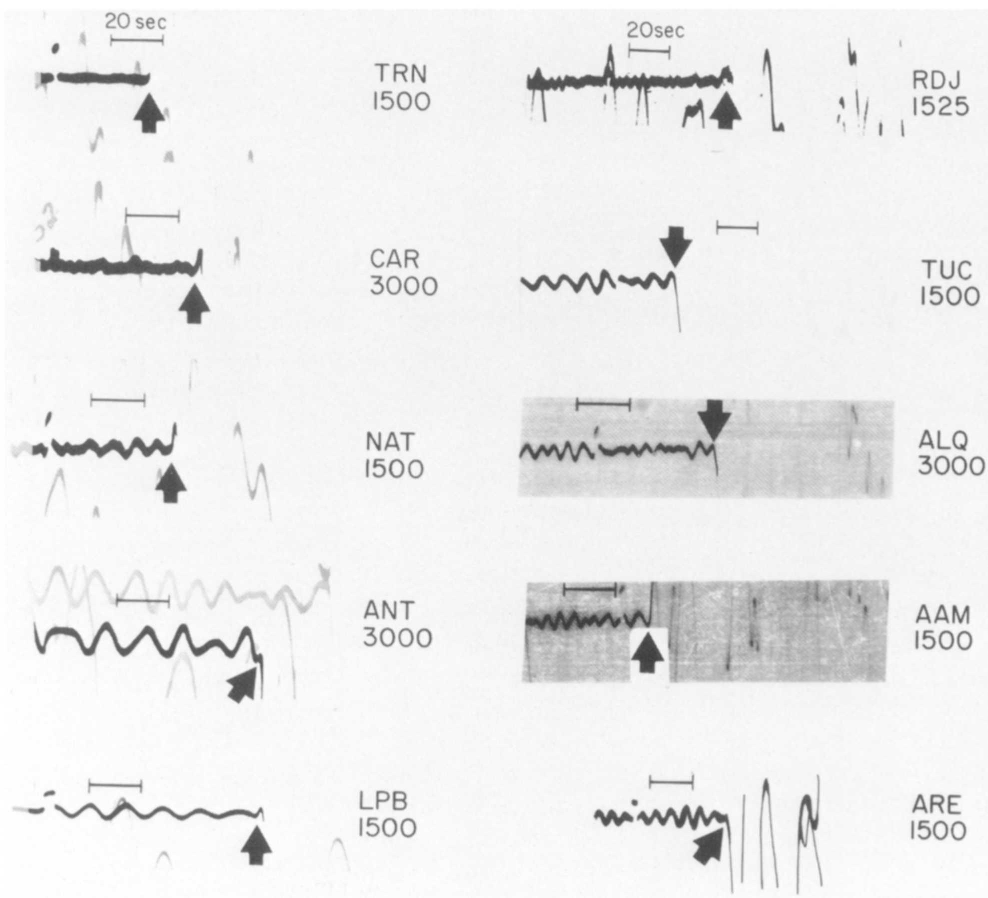

Fig. 8. CAR, NAT, ANT, LPB, ARE, and RDJ are interpreted as compressional first motions inconsistent with the orthogonal nodal plane of Figure 7. Other stations, of which TRN, AAM, ALQ, and TUC are shown here as examples, do not show an obvious counterpart to the small compressional bump seen on the inconsistent stations. Time scale for LPB is 10 sec.

Antiguan earthquake sequence are similar to those of a family of earthquakes occurring in the central Aleutian arc and discussed in detail by LaForge and Engdahl (1979).

These similarities are noted as follows.

1. Both groups of earthquakes, the Antiguan and Aleutian, occur in what we are calling the overriding wedge, immediately above the Benioff Zone at a location where the dip of the Benioff Zone changes from a shallow dip trenchward of

the earthquakes to a steeper dip beneath the island arc. In the case of the central Aleutians, normal fault earthquakes occur very near the principal thrust zone, suggesting a rapid spatial transition from the zone of thrusting to the tensional regime characterized by normal faulting. In the case of the 1974 Antigua earthquake, the principal southeast-dipping aftershock zone also extends down to the inferred Benioff Zone (Figure 6).

2. Both groups of earthquakes occur on normal faults oriented transverse to the trend of the island arc.
3. Both groups of earthquakes occur in arc segments of oblique plate convergence in which there is a significant component of plate motion parallel to the arc structure. In the region of the central Aleutians studied by LaForge and Engdal (1979), the normal to the strike of the island arc and trench trends about N10°W and the direction of the relative plate convergence trends approximately N45°W (Minster and Jordan, 1978). In the region of the Antiguan earthquake of 1974, the normal to the strike of the island arc and trench trends N135°W and the direction of relative plate convergence is about N110°W (Sykes *et al.*, 1982).
4. Both groups of earthquakes occur in regions where the local terrace between the volcanic arc and the trench is broken by transverse structural depressions or canyons with relief of several kilometers. LaForge and Engdahl associate many of their wedge normal fault earthquakes with Adak Canyon and suggest that this type of faulting was, in fact, responsible for the creation of Adak Canyon. We cannot make a strong case to attribute the 1974 Antiguan earthquake to the boundary fault of a particular structural depression. However, the major transverse structures on the southwest wall of the Puerto Rico trench near 16.5°N and near 18°N may be indicative of tensional stresses parallel to the arc structure, analogous to the stresses in the overriding plate of the central Aleutians. Within the epicentral region of the 1974 earthquake, Officer *et al.* (1959) have found a basin or graben with an inferred relief in the geologic section of about 2 km. The basin or graben structure is shown in Figure 9. Since only one section across the structure is known to us, we cannot determine its strike. It is conceivable that this structure represents a surface manifestation of the type of faulting that produced the 1974 earthquake. However, because the structure of Officer *et al.* lies south of most of our aftershocks, and because we infer that the aftershocks occurred on a SE-dipping plane, we do not consider that the earthquake can be attributed specifically to this structure.
5. In both the central Aleutians and the northern Lesser Antilles, there are fracture zones or topography on the subducting plate that may have influenced the tectonics of the overriding plate. Grim and Erickson (1969) and Spence (1977) have suggested that some transverse canyons in the Aleutians are genetically associated with fracture zones on the subducting plate, although the precise mechanism of such an association remains obscure (LaForge and Engdahl, 1979; Fisher *et al.*, 1981). In the northern Antilles, a large topographic feature, the west-northwest trending Barracuda Ridge enters the Puerto Rico trench east of Antigua. What appears to be its western extension is observed on seismic profiler records in the region east of Antigua (Marlow *et al.*, 1974) and has been extended farther west toward Barbuda and Antigua (Figure 1) by Case (1975). The northern scarp of the ridge appears to intersect the shallow portion of the arc midway between Barbuda and Antigua. McCann and Sykes (1983) provide more evidence that the Barracuda Ridge continues beneath the

accretionary wedge and then underneath the Caribbean plate. They also show that the ridge continues some 300 km to the northwest where its extension is being subducted beneath the Puerto Rico-Virgin Islands region.

We see two hypotheses that would be consistent with the above cited similarities between the seismicity of the overriding wedges in the Aleutians and the northern Antilles. One hypothesis is that the stresses producing the wedge normal-fault earthquakes are due to regional extension of the overriding wedge, in a direction parallel to the strike of the arc, caused by the oblique plate convergence. This hypothesis is essentially that of LaForge and Engdahl (1979). The occurrence of topography on the subducting plate beneath the earthquakes would not be a necessary condition for such regional extension, although the impingement of bathymetric highs on a subduction zone might have been the cause of a subduction zone change from a zone of normal plate convergence to a zone of oblique plate convergence, as suggested by Vogt *et al.* (1976).

The other hypothesis is that the stresses producing the wedge normal-fault earthquakes are generated directly by the topography on, or perhaps the buoyancy of, the downgoing plate, which would tend to lift up the overriding plate (Kelleher

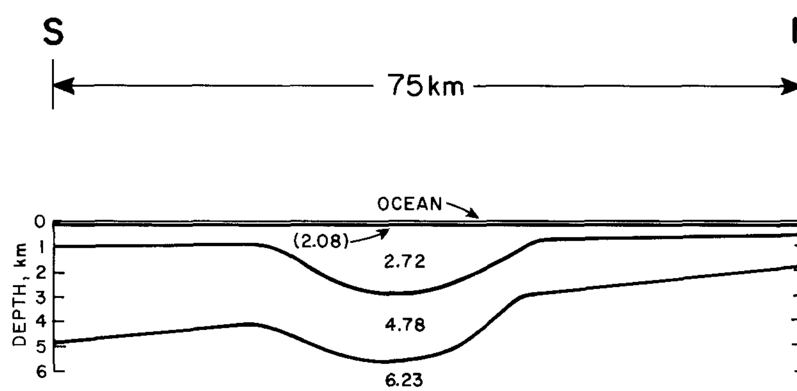

FIG. 9. Crustal structure deduced from a refraction line running through the source region of the 1974 earthquake (from Officer *et al.*, 1959). The location of the line is shown in Figure 1.

and McCann, 1976; Vogt *et al.*, 1976; Stein *et al.*, 1982a). The steep plunge of the pressure axis in the fault plane solution for the Leeward Islands event implies a large change in the direction of maximum compressive stress over a short distance, as this axis is aligned nearly horizontally for earthquakes occurring on the nearby thrust zone (Molnar and Sykes, 1969; Tomblin, 1975). Stein *et al.* (1982a) have determined focal mechanisms for several earthquakes in the overriding Caribbean plate seaward of Antigua. These mechanisms differ appreciably from each other and from the mechanism of the 1974 earthquake, further implying large differences of stress orientation within the overriding plate. Topography on the downgoing oceanic lithosphere might be the source for such rapid variations in the stress field. If a ridge entering a subduction zone, such as the Barracuda Ridge, were isostatically compensated, its buoyancy might cause the greatest principal stress axis in the overlying plate to be nearly vertical as the ridge resists subduction (Kelleher and McCann, 1976; Vogt *et al.*, 1976). The Barracuda Ridge, however, does not appear to be isostatically compensated (Birch, 1970). The topography of the ridge appears to be supported by the mechanical strength of the lithosphere. Large vertical stresses could still be transferred to the overriding plate as the subduction zone adjusts to the consumption of the topography on the underthrust plate. This adjustment could include uplift of the overthrust plate as well as internal deformation of both the overthrust and underthrust plates. If the strike of the ridge at the time it was subducted differed from the direction of motion, additional stress complexities would be introduced in the overriding plate as the locus of maximum uplift on the overriding plate-shifted position. In the case in which topograpy on the downgoing plate produces the normal-fault earthquakes, the oblique convergence of plates at a subduction zone would not be a necessary condition for the occurrence of such earthquakes although, as cited above, oblique convergence might result from the difficulty of normally subducting topography located on the downgoing plate.

On the Potential for a Great Thrust Fault Earthquake in the Northern Lesser Antilles

The 1974 Antigua earthquake occurred in the northern part of the zone of strongest shaking of the great earthquake of 8 February 1843 (Robson, 1964). The region has been hypothesized to be a potentially dangerous seismic gap in which a great thrust-fault earthquake might occur in the next several decades (Kelleher *et al.*, 1973). For this reason, the region is currently being monitored by a microearthquake network (Murphy and McCann, 1979). The postulated seismic gap is not a "gap" in the sense, sometimes used, of having been seismically quiescent in recent decades (i.e., a seismicity gap or zone of quiescence). In fact, the segment of the northern Antilles from 16.5°N to 18.0°N has experienced rather high numbers of moderate-sized earthquakes in recent decades (Figure 4). Moderate events such as these do not seem likely, however, to have released a significant fraction of the elastic strain energy that has accumulated since 1843 (Brune, 1968). The nearly vertical axis of greatest principal stress for the 1974 earthquake implies that this earthquake did not release much of the compressional elastic strain oriented nearly horizontally that is thought to be accumulating around the thrust interface between the American and Caribbean plates. From this standpoint, the 1974 earthquake did not reduce the risk of a future great thrust-fault earthquake in the northern Lesser Antilles.

Stein *et al.* (1982a) have suggested that most of the other recent, teleseismically well-recorded, earthquakes from the northern Lesser Antilles also are intraplate earthquakes, rather than thrust-fault earthquakes on the plate interface. By the reasoning of the preceding paragraph, this observation would imply that the other recent earthquakes also did not reduce the risk of a future great thrust-fault earthquake in the northern Lesser Antilles. Stein *et al.* (1982a) suggest, however, that the observed recent scarcity of thrust interface earthquakes in the northern Lesser Antilles may imply that much of the slippage between the converging Caribbean and American plates is accommodated aseismically. They (Stein *et al.*, 1982a, b) would group the Lesser Antilles with other regions, worldwide, in which old lithosphere is subducted at low velocity; these other regions do not experience a high level of thrust interface seismicity. Under this hypothesis, recurrence times of great thrust-fault earthquakes would clearly be much larger than envisaged by Kelleher *et al.* (1973) and McCann *et al.* (1979). Stein *et al.* (1982a) suggest that some of the great historical earthquakes in the northern Lesser Antilles may have been intraplate earthquakes, rather than thrust-fault earthquakes. Unfortunately, the occurrence of a normal-fault earthquake in the overriding wedge is not, of itself, sufficient evidence to rule out the hypothesis that compressional elastic strain is accumulating on the interface. Some thrust faulting is occurring seaward of the epicentral region of the Lesser Antilles earthquake, as noted previously, and the wedge normal-fault earthquakes studied by LaForge and Engdal (1979) occur in an arc segment that experiences large thrust-fault earthquakes (LaForge and Engdahl, 1979; Sykes, 1971).

Conclusions

The Antiguan earthquake of 8 October 1974 was an apparently unusual shock—a large earthquake in the overriding wedge of the Caribbean plate that occurred on a normal fault oriented transverse to the regional trend of the Antilles arc. Similar shocks, although of much smaller magnitude, have been observed in the central Aleutians arc (LaForge and Engdahl, 1979). The two regions, the central Aleutians and the northern Lesser Antilles, are both characterized by oblique plate convergence, with the component of motion parallel to the arc being slightly greater in the section of the Aleutians studied by LaForge and Engdahl. In addition, in each region it is possible to postulate that the subduction process may be perturbed by topography or structure on the subducted plate: evidence in favor of this hypothesis may be stronger in the northern Lesser Antilles. Either oblique plate convergence or subduction of topography could be responsible for the tensional stresses, oriented parallel to the arc, that produced the Antiguan earthquake. Deciding in favor of one of these processes, or formulating new hypotheses, will obviously be helped by identification of other regions of similar wedge normal-fault earthquakes. Since such identification requires accurate hypocenter locations in order to separate shocks in the overriding wedge from shocks in the underthrust plate, it is likely that wedge normal-fault earthquakes similar to those of the Aleutians and the Antilles do occur in some other regions but have not yet been distinguished from the normal-fault earthquakes occurring within the subducting plate.

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