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A LARGE, DEEP HAWAIIAN EARTHQUAKE—THE HONOMU, HAWAII EVENT OF APRIL 26, 1973

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ABSTRACT

The largest subcrustal earthquake ever recorded from the Hawaiian Island chain (magnitude 6.2) occurred at a depth of 48 km on April 26, 1973. The proximity of the Hawaiian Volcano Observatory's extensive seismograph network and our knowledge of the crustal and upper mantle structure beneath the island made it possible to calculate accurate hypocenters for both the main shock and 57 aftershocks. The earthquake may have triggered swarms of small, shallow earthquakes at two different locations on the island: one 25 km and the other 50 km from the epicenter of the earthquake. The polarity of the *P*-wave arrivals for the main shock and most of the aftershocks, as recorded by the local network and worldwide stations, define nodal planes oriented N26°E dipping 77°W and N70°W dipping 61°S. Comparison of the inferred directions of the greatest and least principal stresses derived from these data with the stress direction within the Pacific plate assumed for various hypotheses of the formation of the Hawaiian island chain show closest agreement with the concept that the orientation of the archipelago is aligned parallel to the direction of the maximum shear stress and is not perpendicular to the orientation of the least principal stress.

INTRODUCTION

On April 26, 1973, an earthquake of body-wave magnitude 6.2 occurred at a depth of 48 km beneath the northeast coast of the island of Hawaii (Figure 1). This event was the largest to occur in Hawaii since 1962 and is the largest subcrustal earthquake ever recorded in the archipelago. This earthquake is of special interest because it is located in a region of the lithosphere that is virtually aseismic compared to the very active areas nearby (Koyanagi, 1965; Koyanagi and Endo, 1971). As will be discussed in the following sections, this earthquake has provided us the opportunity to learn more about the structure and tectonics of Hawaii but has left us with many unanswered questions.

Damage from the earthquake, which was estimated by county officials to be $5.6 million, was considerable along the northeast coast of the island, where 11 people were injured; most of the structural damage occurred in the city of Hilo. Maximum intensity was reported to be about VIII (Modified Mercalli Scale) in the epicentral region and at least VI over virtually all of the island of Hawaii (Nielsen *et al.*, 1977). The zone of intensities I to III extended to the island of Kauai.

A record of the earthquake was obtained from a Kinemetrics SMA-1 strong-motion accelerograph located near the Hawaiian Volcano Observatory (HVO), about 55 km from the epicenter of the main shock. Analysis of this record showed maximum horizontal accelerations of 0.17 *g* in a direction N30°W and 0.11 *g* at N60°W and a maximum vertical acceleration of 0.07 *g* (Nielsen *et al.*, 1977).

LOCATION OF THE MAIN SHOCK

At the time of the earthquake, the seismic network on the island of Hawaii consisted of 29 short-period seismograph stations (Figure 1). These instruments utilize either a model EV-17 or L4-C, 1 Hz vertical geophone. The data are telemetered to HVO and recorded on one of two microfilm Develocorders. Complete system amplification averages about 10^4 at 1 Hz (Koyanagi *et al.*, 1974). In addition, data were used from three, self-contained, three-component, Wood-Anderson stations located in Hilo and on the islands of Maui and Oahu. Errors in reading arrival

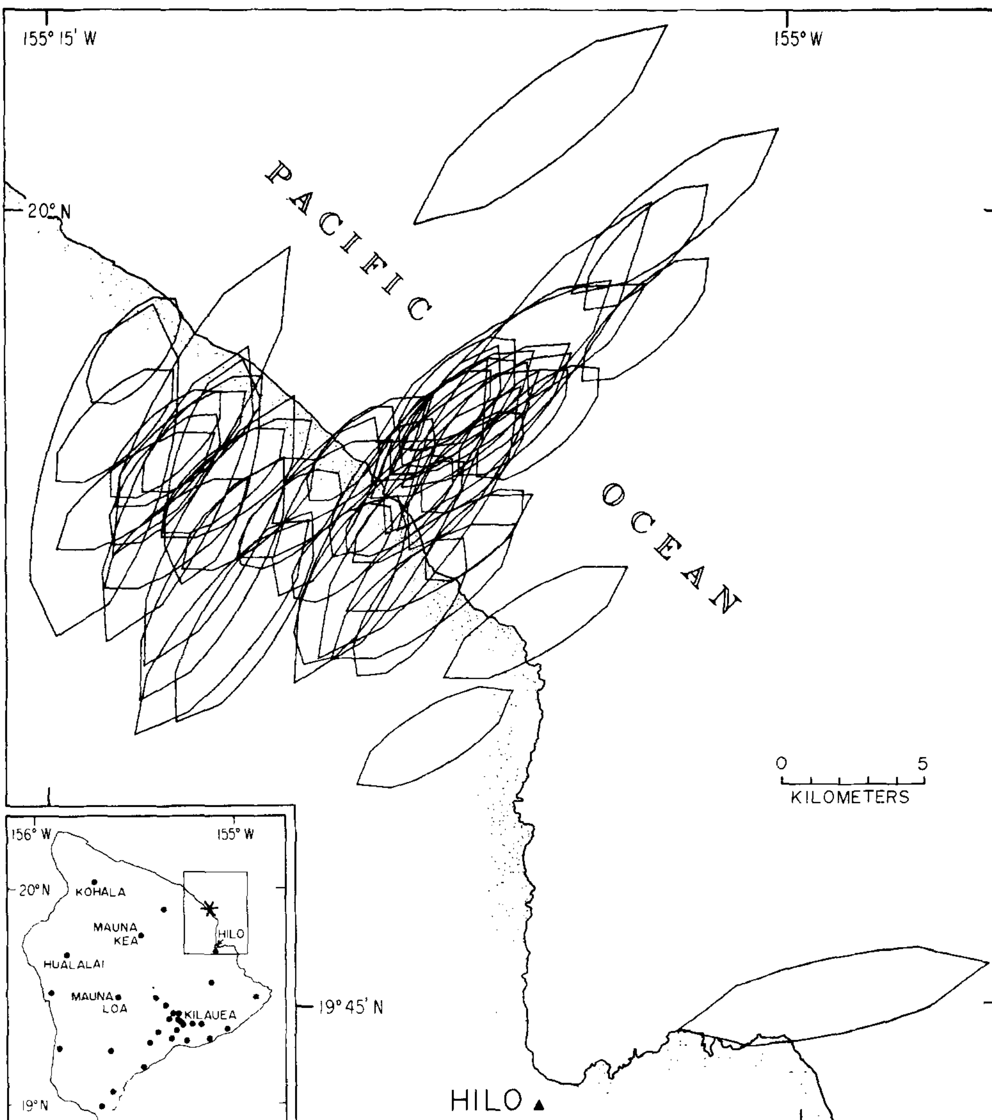

FIG. 1. Map of a portion of the northeast coast of the island of Hawaii showing the epicentral region and the distribution of aftershocks (ellipses). The size of the ellipses indicates the uncertainty in the epicenters of the aftershocks. The small map in the *lower left* corner (*inset*) shows the locations of the seismographs in the Hawaiian Volcano Observatory network (dot) and the epicenter of the main shock (asterisk).

times, in picking first motions of events in the magnitude range discussed here, or in determining the clock corrections are typically 0.05 sec and rarely exceed 0.1 sec. The computer program HYPOELLIPSE (Lahr and Ward, unpublished computer program, 1974) was used to locate the earthquake hypocenters. A unique feature of this location routine is that it allows calculation of the ellipsoid of standard error around the hypocenter rather than the standard errors in the geographic coordinate system. These ellipsoids are projected as ellipses onto the plane of view in the maps and cross sections. The ellipses plotted in this paper are represented as 18-sided polygons, for computational efficiency.

Many different models of the *P*-wave velocity structure of the crust under Hawaii have been derived by various investigators (for example, see Ellsworth and Koyanagi, 1977). The thickness of the crust in Hawaii is known to vary by many kilometers (Ryall and Bennett, 1968; Hill, 1969). Furthermore, lateral variations in velocity are known to exist within the crust (Ward and Gregersen, 1973). Intrusives with *P*-wave velocities on the order of 7 km/sec are thought to occur at several kilometers shallower depth beneath the volcanic rift zones than elsewhere on the island.

Some of the effects of lateral variations in the crust and the wide range of the elevations of the seismic stations (sea level to 4 km) can be accounted for by calculating corrections for the individual stations. Compensation for the effects of inhomogeneities is easier for subcrustal earthquakes located in a fairly tight cluster than for shallow earthquakes because: (1) the waves travel at relatively steep angles of incidence through the crust, and (2) the various travel paths from the events in the aftershock zone to any individual station are almost identical.

Accordingly, two corrections were calculated for each station: First, an elevation correction was determined by dividing the station elevation by 6.7 km/sec. This velocity was chosen because the higher stations are located over the centers of the volcanoes, where crustal layers with velocities of 6 to 7 km/sec appear to be relatively thick.

Then, a crustal thickness correction was determined by utilizing the Bouguer gravity anomolies that have been calculated throughout the island (Kinoshita, 1965), and by comparing these values with the crustal thicknesses that have been determined by refraction surveys along the coast of the island. A family of travel-time curves was calculated for a series of crustal structures identical to that reported by Ward and Gregersen (1973), except allowing the thickness of the 6.7 km/sec layer to vary from 0.8 to 6.8 km and thus the total crustal thickness to vary from 10.2 to 16.2 km. These travel-time curves were compared to the ones observed by Hill (1969) during refraction studies around the island, and a crustal thickness was assigned to the center of those refraction lines that were long enough to sample the mantle. The Bouguer anomaly was determined for the same general region and plotted versus the crustal structure (Figure 2). The error bars represent the uncertainty in choosing a typical value for the Bouguer anomaly in each area. A line was drawn subjectively through the four points. A travel-time delay was calculated for waves traveling through the upper mantle, and each of these crustal thicknesses was compared to the standard crustal model used in the location routine. The Bouguer anomaly at each seismic station was determined and a corresponding delay calculated from the relationship shown in Figure 2.

These two corrections are probably not sufficiently large in regions with thick deposits of relatively low-velocity material, such as under stations located far out on the flanks of the volcanoes, and may be too large for stations located on the summit areas or rift zones, which are underlain by relatively high-velocity material.

The eight, best-recorded aftershocks and the main shock were located using the above corrections and their travel-time residuals averaged for each station to determine a total station correction. The main shock and all of the aftershocks were then relocated using this composite station correction. Not only did this procedure minimize the root-mean-square of the travel-time residuals, but the final corrections agree quite well with our intuitive feelings about the variations in crustal structure thoughout the island.

We chose a mantle velocity of 8.2 km/sec in our model because work by Ellsworth (1977) indicated that a mantle velocity in the range 8.07 to 8.27 km/sec with an average of 8.2 km/sec gave the best fit for the inversion of arrival-time data from 60 earthquakes located beneath different parts of the island at depths from 20 to 50 km. Previous refraction studies (Shor and Pollard, 1964; Woolard, 1966) have reported that mantle velocities may vary from island to island within the archipelago and with the azimuth of propagation. Because of these complexities, we thought it best to use the velocities that were relatively well-determined for the island of Hawaii and to disregard values for other areas.

The location of the main shock determined by using different station corrections and velocity models is given in Table 1. The first solution listed is the preferred

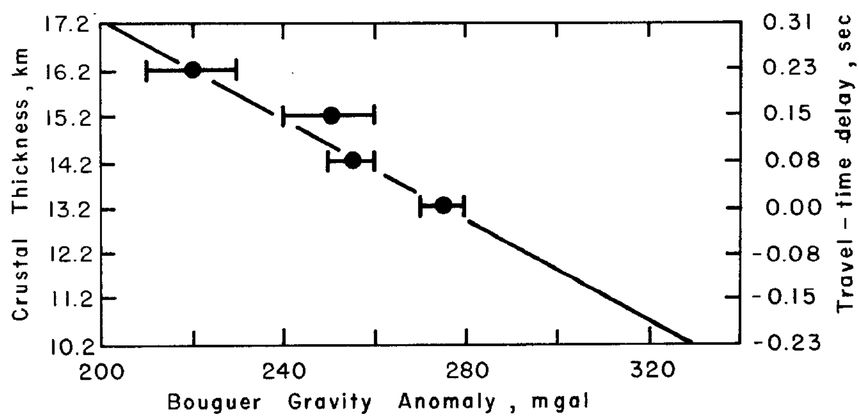

FIG. 2. Relationship between crustal thickness or the corresponding travel-time delay and the Bouguer gravity anomaly for the island of Hawaii.

TABLE 1

A COMPARISON OF HYPOCENTRAL LOCATIONS OF THE MAIN SHOCK USING DIFFERENT STATION CORRECTIONS, MANTLE VELOCITIES, AND CRUSTAL VELOCITIES (THE FIRST LOCATION IS THE ONE PREFERRED BY THE AUTHORS)

Station Corrections	Crustal Velocity Model	Mantle Velocity (km/sec)	Latitude		Longitude		Depth (km)	Origin Time (sec)	Root-Mean-Square Residuals
			Deg N	Min	Deg W	Min			
Yes	*	8.2	19°	54.2	155°	7.8	48	30.6	0.04
No	*	8.2	19°	54.1	155°	5.9	51	30.4	0.15
Yes	*	8.3	19°	54.2	155°	7.8	46	30.8	0.04
Yes	†	8.2	19°	54.2	155°	7.9	48	30.8	0.04
(Ellsworth, see text)	†	8.2	19°	52.8	155°	9.0	42	31.6	0.11

* Crustal model after Ward and Gregersen (1973).

† Constant crustal velocity of 6.0 km/sec, crustal thickness of 13 km.

location of the main shock. The 95 per cent confidence limits of this location, assuming a possible error in the arrival times of ±0.1 sec, are ±5.1 km in the N50°E direction, ±1.8 km in the N40°W direction, ±7 km in depth, and ±0.1 sec in origin time. These error estimates are rather liberal and seem to appropriately reflect the uncertainty in the choice of crustal structure, mantle velocity, and station corrections, with the possible exception of origin time, which could perhaps vary by a few tenths of a second if the corrections and velocities chosen turn out to be unreasonable.

The large change in the location of the main shock shown in Table 1 when no station corrections are used can be explained by the absence of stations in the semicircle of azimuths northeast of the epicenter, and by the fact that the stations to the west are located at high elevations and in geologically complex regions. When the station corrections are used, the hypocenter is less sensitive to the deletion of some station arrivals, which suggests that the station corrections are at least internally consistent and, therefore, probably reasonable.

Ellsworth (1978) used teleseismic *P*-wave arrivals across the HVO seismic network to study the mantle beneath Hawaii. During the course of that investigation he derived a set of mean residuals for the stations in the HVO network, which were based on the travel times calculated from either the Herrin or Jeffreys-Bullen tables or by using deviations from the best-fitting plane wave. Because these residuals essentially represent an externally derived set of station corrections, we recomputed the hypocenters of the main shock and the eight test aftershocks using Ellsworth's residuals as station corrections (Table 1). Although the earthquake locations differ significantly from and are of lower quality than those computed using the corrections described earlier in this section, the statistics related to the precision of these

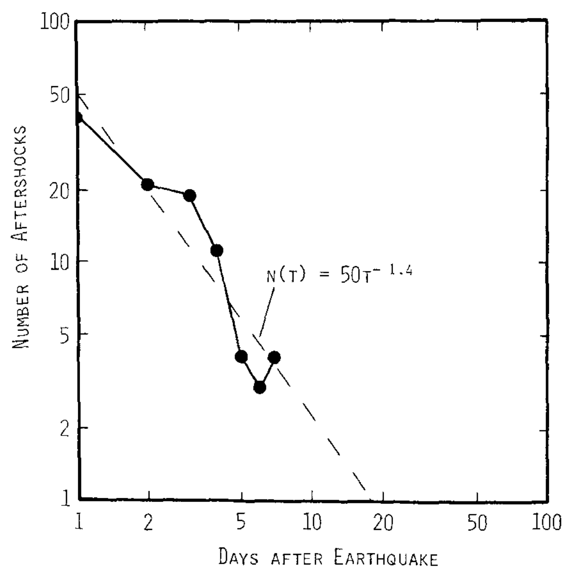

FIG. 3. The number of aftershocks recorded by the local seismograph network plotted against time after the main shock.

locations clearly indicate that they are a definite improvement over using no station corrections to locate the earthquakes.

One possible explanation for the differences between the corrections derived in this paper and those calculated by Ellsworth (1978) is that the teleseismic *P* waves used to generate the latter set traversed the HVO network at much steeper angles of incidence than the *P* waves from the earthquakes discussed in this paper. Furthermore, any lateral variations in mantle structures deeper than 50 km would be reflected only in the corrections derived from the teleseismic arrivals.

AFTERSHOCKS

The main shock damaged the commercial power system on the island so that the microfilm recorders did not operate properly for several hours. However, within 1 day after the main shock the network was recording about 40 aftershocks per day. Five days after the main shock, only about four events per day were recorded. The activity during this 5-day interval died off at an exponential rate (Utsu, 1961) with an exponent of about 1.4 (Figure 3). This level of activity seems low when compared to the thousands of aftershocks reported after shallow events of comparable size (e.g., Page, 1968). Because of the attenuation of seismic waves with distance, however, the number of events recorded at 50 km from the hypocenter, which is the distance to the closest station in this study, would be on the order of 40 times less than the number recorded by the same instruments for events only 5 km away (Ward and Bjornsson, 1971). Thus, the number of aftershocks recorded and the decay in the aftershock activity actually may be comparable to aftershock sequences from similar sized crustal earthquakes.

Fifty-seven aftershocks were recorded clearly enough at 15 or more stations to allow calculation of reasonably accurate locations, which are shown in Figure 1 and in the cross-sections in Figure 4, respectively. The ellipses show the 66 per cent

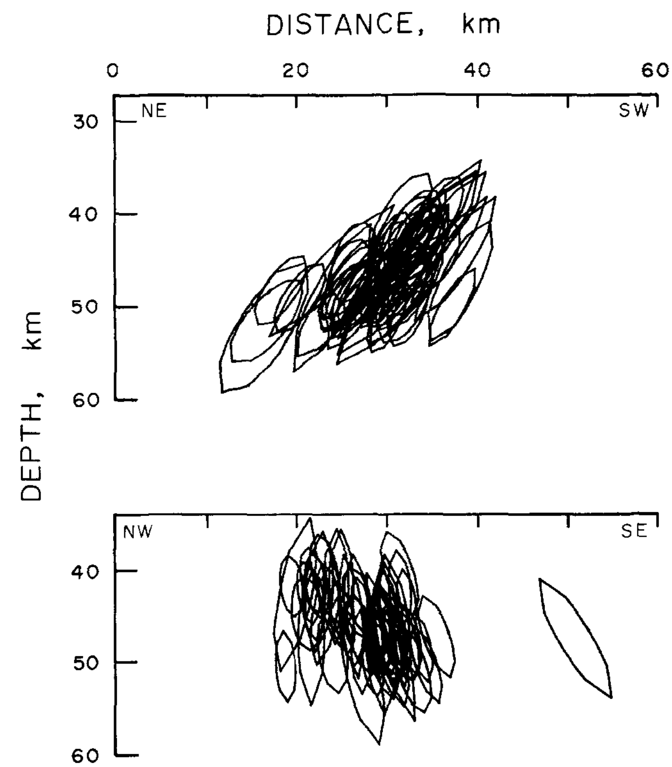

FIG. 4. Projection of the aftershock hypocenters onto two cross sections, oriented northeast-southwest and northwest-southwest, to show the distribution of the aftershocks with depth.

confidence limits of the error in location and were calculated assuming a possible 0.1-sec error in reading each arrival time. The data for each location were reëxamined carefully, and the records for any station with a large travel-time residual were reread. If the reading was unclear, the arrival time was disregarded. The root-mean-square of the travel-time residuals for the aftershocks ranged from 0.03 to 0.08 sec when the station corrections discussed in the previous section were used.

The hypocenters, with the exception of the one to the southeast, occur in an area about 15 km wide and appear to extend along a west-northwest trend and a northeast trend. If the northeast trend is only an apparent one caused by the elongation in the ellipses, which in turn is due to the uneven distribution of stations around the focal region, then the fault plane active during the aftershock sequence could be the one corresponding to the N70°W trending nodal plane of the main shock (Figure 5).

This assumption seems reasonable, but the aftershock distribution contains enough scatter so that fault plane cannot be unambiguously determined.

The depths of the aftershocks range from 40 to 50 km, and the area of aftershock activity is of the order of 150 km^2, which is roughly comparable to the area of fault planes active during well-located aftershock sequences from shallow earthquakes of similar size.

Possible Triggering of Other Earthquake Activity

Shortly following the main shock, a substantial increase was observed in the number of small earthquakes located in the upper crust in two areas known for sporadic seismicity: the first lies about 25 km west-northwest of the main shock beneath the east slope of Mauna Kea volcano, and the second lies in the summit

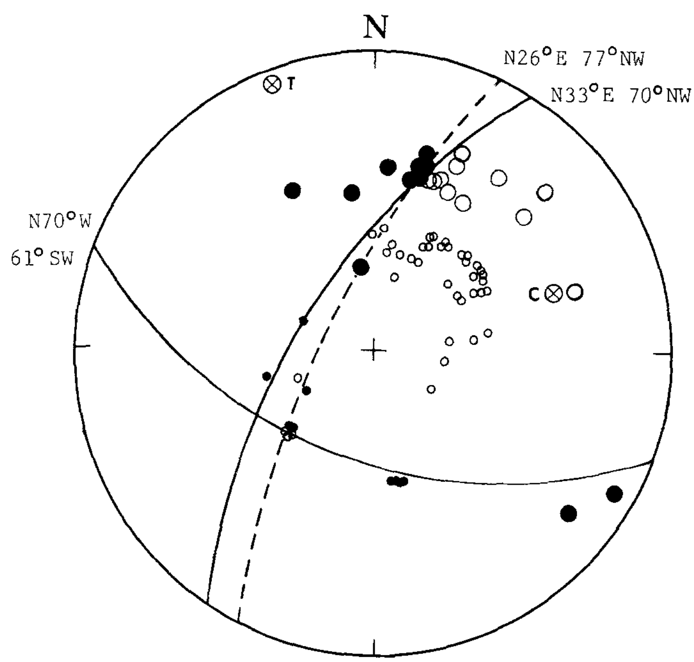

Fig. 5. The first motions of the main shock as observed on both the local Hawaiian network and on worldwide seismographs are plotted on a lower hemisphere projection. The smaller circles represent first motions from worldwide stations and the larger circles are from the local network. The open circles represent down first motions and solid circles up first motions. The directions of the greatest and least principal stress are indicated by the C and T, respectively.

caldera region of the very active Kilauea volcano about 50 km south-southwest of the earthquake. Normally, only a few small earthquakes per week are recorded from the east flank of Mauna Kea; however, several earthquake swarms have been recorded over the past two decades. The largest previously recorded swarm consisted of only about 30 events per day during the peak of the activity (Koyanagi and Endo, 1971). The first day following the main deep earthquake, however, about 80 events were recorded on the nearby station (KKU), and the activity remained high for about 5 or 6 days (Figure 6). The earthquake activity in the summit caldera region is highly variable and ranges from a few tens of earthquakes to as many as several thousand shallow events per day (Koyanagi, 1965). This activity increased by a factor of more than six following the main shock (Figure 6). Such abrupt changes have occurred at other times and are usually associated with volcanic processes

(Eaton, 1962), but only five similar occurrences were observed during the preceding year. Thus, the possibility that these events were triggered by deep earthquake activity several focal dimensions (of the aftershock zone) away cannot be totally ruled out.

Nodal Plane Solution and Regional Stresses

The first motions of the *P*-wave arrivals at stations in the HVO network and around the world are plotted on a stereographic projection in Figure 5. These data show predominately strike-slip motion with the axis of greatest principal stress inclined about 30° and striking approximately N73°E and that of the least principal stress about horizontal and striking nearly N22°W. The first-motion data from most of the aftershocks are consistent with this solution. Five aftershocks, however, had a large number of compressional first motions in the northwest quadrant, but sufficient data were not available to determine nodal planes for these events.

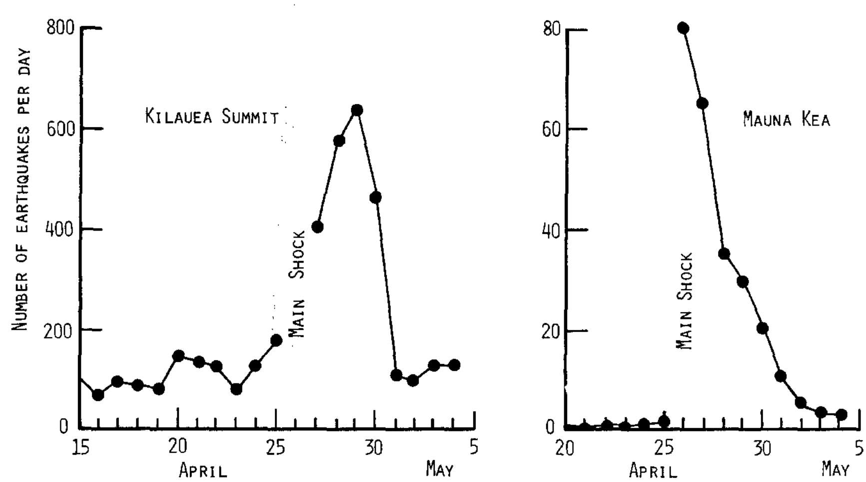

Fig. 6. The number of earthquakes recorded per day at the summit of Kilauea volcano (*left*) and on the east flank of Mauna Kea (*right*) for a time period that includes the main earthquake. Note that the vertical axes have different scales.

Similar nodal planes and sense of motion were determined for the magnitude 6.1 Kaoiki earthquake, which was located at a depth of 8 km about 14 km west of the summit of Kilauea volcano (Koyanagi *et al.*, 1966). Endo (written communication, 1973) found comparable solutions for many earthquakes in the upper mantle beneath Hawaii in 1968. Ward and Gregersen (1973) determined roughly analogous principal stress axes for four out of five mantle earthquakes located in Hawaii during 1967. Sykes and Sbar (1974) determined a similar nodal plane solution for an earthquake in the Pacific plate east of Hawaii at 12°N and 131°W. The majority of the nodal plane data available, with the exception of those shallow earthquakes closely related to magma intrusion and eruptions on Kilauea (Endo, 1971), imply that the maximum principal stress in the central part of the Pacific plate strikes nearly east-northeast, while the least principal stress strikes about north-northwest.

The general trend of the Hawaiian archipelago from Midway to Kilauea is about N65° to N70°W and roughly coincides with one of the nodal planes in Figure 5. This observation agrees with the suggestion of Betz and Hess (1942) that the Hawaiian Islands may lie along a zone of strike-slip faulting, even though such faults have never been observed. Menard (1959) noted that the minor lineations along the

Hawiian ridge might be explained by horizontal displacement along the ridge, and showed that a mirror image of the Hawaiian chain and its associated minor lineations compare closely in size, trend, and complexity with the San Andreas fault and the Basin and Range province. This analogy implies left lateral movement along the Hawaiian Ridge and would be in agreement with the nodal plane solution shown in Figure 5.

The orientation of maximum and minimum stresses derived from the first motions do not agree either with the suggestion by Green (1971) and by others that the volcanoes lie along a propagating tension crack, or with the stress directions postulated by Jackson and Shaw (1975). The best agreement of these derived stress directions is with the stress field proposed by Turcotte and Oxburgh (1973). They suggest that en-echelon fractures underlying the arcuate segments of the Hawaiian ridge are formed at an angle of 45° to the minimum principal stress, which therefore must strike about N13°W, or about 13° more northerly than the direction shown in Figure 5. It is not clear what the orientation of the lithosphereic stresses would be if the Hawaiian chain is formed in response to such mechanisms as thermal plumes (Wilson, 1963; Morgan 1971, 1972a, b) or to gravitational anchors (Shaw and Jackson, 1973).

We can find no obvious relationship between the nodal planes and geological or tectonic features on the island of Hawaii. The east rift zone inferred on Mauna Kea (Macdonald and Abbott, 1970; Fiske and Jackson, 1972) strikes nearly east-west rather than N70°W. The Kaoiki fault system (Macdonald and Abbott, 1970) between Mauna Loa and Mauna Kea coincides closely with an extension of the N30°E nodal plane, but little geological or seismic evidence exists to suggest that the Kaoiki faults extend westward to Hilo. Malahoff and Woolard (1968) infer a possible sinuous extension of the Molokai fracture zone, which is roughly parallel to the N70°W nodal plane, in the general region of this earthquake, but no direct evidence for such an extension has been found. Furthermore, the sense of motion along the fracture zone is inferred to be normal rather than strike-slip.

SUMMARY

The only mechanism that is generally accepted for subcrustal earthquakes in Hawaii is one associated with the generation and/or transport of magma from depths of about 60 km through the mantle and into the crust. In fact, this model has only been proposed for those deep events associated with Kilauea volcano, where the relationship between the crustal and mantle earthquakes is the clearest (Eaton, 1962; Koyanagi and Endo, 1971). Subcrustal earthquakes, however, occur relatively uniformly beneath the entire island, and no clear mechanism or cause has been accepted for those events not clearly related to Kilauea. Endo and Rogers (in preparation) have recently examined the epicentral distribution and focal mechanisms of upper mantle earthquakes in Hawaii and have hypothesized that subcrustal earthquakes such as the one discussed in this paper could be caused by crustal loading on the lower lithosphere.

In the previous section, we pointed out that the nodal planes of the Honomu earthquake apparently are not related to any identifiable structures on the surface. Even more important is that, with the exception of the deep events related to the magma plumbing system beneath Kilauea, almost all of the seismicity in Hawaii, that is related either directly or indirectly to structures such as volcanic centers, rift zones, or fault zones, does not appear to extend deeper than about 10 to 15 km (Swanson *et al.*, 1976). Therefore, we cannot necessarily expect the stress system inferred by this event to be correlated with any surface structures, either. For the present time, until our understanding of the tectonic processes in the mantle beneath Hawaii improves, we can only speculate as to what the cause of this event and many other mantle earthquakes might be.

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