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SEISMICITY OF THE LOWER EAST RIFT ZONE OF KILAUEA VOLCANO, HAWAII, 1960 TO 1980

By

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1. Introduction

In mid-February 1970, seismometer PAX at Kaniahiku Village in the lower east rift of Kilauea (fig. 1) detected an increasing number of microearthquakes. To confirm and define the source area of the activity, we installed an additional station at Puu Honuaula (PHO) and operated a portable seismograph at various localities nearby. Signature characteristics and the relative number of small events recorded from the various stations indicated a concentration of activity near the 1955 eruptive vents between Puu Honuaula and Puulena. Epicenter determinations made from P-arrival times picked on the temporary and permanent stations of the HVO net indicated an epicentral area that extended south to Pohoiki and southwest to Kaimu at depths of mostly less than 10 km, although some of the scatter of hypocenters may be due to errors in locating earthquakes near the edge of the seismic net. The frequency of microearthquakes detected in February and March fluctuated from about 20 per day to about a thousand per day with peak activity on February 21 and March 2. During two months of activity, nearly 6,600 events of −1.0 to 3.9 magnitude (M) were detected. Nearby residents felt many of the apparently shallow earthquakes of $M \gtrsim 1.5$.

Such swarms of microearthquakes have occurred intermittently and, in general, more after eruptions in the eastern and central parts of the east rift in 1960 and 1961. Since the east rift eruption in 1961, however, lower east rift swarms did not correlate with specific volcanic events, although those in the upper east rift zone typically continued to do so. Nevertheless, there is repeated documentation showing evidence that the seismic activity at Kilauea is due to changes in stress conditions ultimately caused by volcanism.

This report is intended to summarize seismic and other geophysical studies of the lower east rift zone near the exploratory geothermal well HGP-A, and to describe the pattern of seismicity there primarily from instrumental data collected over the past two decades.

2. Geological and Geophysical Setting

Kilauea, the southeasternmost volcano of the Hawaiian Archipelago, is a broad shield with linear eruptive fissures extending from the summit caldera eastward along the east rift zone and southwestward along the southwest rift zone. This volcano, comprised of numerous lava flows of tholeiitic basalt, is continually studied at the Hawaiian Volcano Observatory. Deformation, seismic, and geoelectrical data collected and evaluated enhance scientific understanding of the volcanic processes.

From seismic and tilt data, Eaton (1962) developed a structural model of Kilauea where magma traced to nearly 60 km in depth rises to fill a small storage system a few kilometers beneath the summit. His data further indicated that magma is transported laterally through shallow rift zones to feed eruptions along these linear fissure systems.

The mechanics of magmatic movement in the rift zones were later studied in detail by Fiske and Jackson (1972). The laboratory experiments they conducted using gelatin models demonstrated how stresses from magmatic pressures could create extensive swarms of thin magma-transporting dikes within the rift zones. Also studied in their investigation was how the orientation of shallow rift zones could be influenced by gravitational loading of existing volcanic masses.

Koyanagi and others (1972) described the asymmetric distribution of crustal earthquakes along the east rift zone to support the concept that stresses generated by magmatic intrusion are relieved on the seaward, free slopes of the volcano; the north flank is buttressed against the massive Mauna Loa volcano and remains stable and relatively immobile. Epicenter maps prepared in summary bulletins from HVO since 1962, such as in Koyanagi and others, 1978a, 1978b, 1978c, show the high concentration of crustal earthquakes on the seaward slopes of Kilauea, particularly in 1975-77 during the intense aftershock activity following a 7.2-magnitude earthquake on November 29, 1975 (fig. 2). The aftershocks clustered in the south flank along an elongate zone that extended from the southwest rift eastward to about Puu Honuaula on the east rift. The areas north of the east rift, as well as east of Puu Honuaula are contrastingly aseismic. Swanson and others (1976) verified the crustal instability and showed the trend of southward movement of Kilauea's south flank with detailed deformation surveys. They contend that magma forcefully intrudes the rift zones by forming many thin and near vertical dikes. The intrusion forces the wallrocks apart normal to the direction of intrusion ultimately causing the seaward displacement of the south flank.

Zablocki (in Keller and others, 1977) summarized these concepts by putting into perspective his findings from geoelectric surveys of the east rift zone, with particular emphasis in the lower part of the subareal rift southeast of Pahoa in an area of known microearthquake localization. He described a positive thermal anomaly elongated transverse to the major axis of the rift zone at a place that corresponds to an apparent left-lateral offset of the principal volcanic features that define the surface of the rift zone. He suggested that the buttressing effect of Mauna Loa north of the rift lessens with increasing distance away to the east, and eastward from Puulena the more symmetrical topography and deformation pattern accordingly indicate symmetry in the distribution of stress normal to the rift zone. Such changes in the stress field may cause structural offsets to develop in the rift and may cause magma fed laterally from the summit to accumulate, forming secondary storage zones. These findings formed the primary scientific basis in determining the location of the first Hawaii geothermal drill site.

Furumoto (1978), using seismic and gravity data and information gathered from the geothermal well, proposed a crustal model for the geothermal area. He described the core of the rift as a broad zone that extends downward from a depth of 4 km. This hot zone is overlain by a 2 km thick dike swarm. A small geothermal reservoir was defined above the dike complex and at about 1 km below the surface.

3. Instrumentation and Data Quality

The Hawaiian Volcano Observatory currently maintains a 45-station island-wide network of seismometers (Koyanagi and others, 1978c). Signals are primarily telemetered by radio and collected at the observatory where earthquake phase arrivals are picked to 0.05-second accuracy from Develocorder microfilms. Seismometric coverage is best in the summit area of Kilauea and becomes marginal in the lower east rift zone near the edge of the seismic net. Since 1955, when a Loucks-Omori seismograph operated at Pahoa detected increase in seismicity associated with the eruption in the lower east rift zone, more sensitive instruments were gradually placed closer to Puulena and Puu Honuaula improving our microearthquake detection capability there. Timing improved after 1967 with the expansion of the telemetry network and installation of the Develocorder. Stations added in the east rift zone in the late 1970's improved our capability to locate earthquakes in that area. Seismic refraction studies in Hawaii (Ryall and Bennett, 1964; Hill, 1969) formed necessary bases for the refinement of crustal velocity models. Simple-layered crustal models and manually-calculated earthquake foci were replaced by various multi-layered velocity models and computer-generated earthquake solutions. Currently, Hawaii earthquakes are determined at HVO with the computer program HYPOINVERSE using a linear gradient velocity model with statistically derived P-delays mainly from Kilauea earthquakes developed by Fred Klein (Klein, 1978; Koyanagi and others, 1978c; Klein, 1981). The earthquake solutions in this report were essentially prepared by Klein using his velocity model and location program. Figure 2 illustrates typical epicenter plots prepared for the HVO annual summaries.

Magnitude determinations for local earthquakes were based on trace amplitudes read from Wood-Anderson seismographs (M_A). Amplitudes measured from other instruments of higher sensitivity were corrected to correspond to Wood-Anderson amplitude and used to derive magnitude comparable to the Richter local magnitude. Coda length magnitudes for small lower east rift earthquakes were derived from a statistical correlation of amplitude magnitude and signal duration in a semi-logarithmic relation:

$$M_{F-P} = -2.91 + 3.31 \log (F-P),$$

where F-P represents coda length in seconds from a lower east rift station and M_{F-P} is the coda magnitude. The nonlinearity of the magnitude-frequency relation of events located during 1962-77 indicates that our data set is generally complete for all events of $M_A \geq 2.5$. The magnitude-frequency relation of earthquakes is discussed in more detail in a later section.

During episodic swarms of activity in the lower east rift, portable seismographs with magnification of about 50-100 K peaking at frequencies of about 5-10 H_Z were operated to supplement normal instrumental coverage.

Our increasing capability to detect and locate earthquakes in the lower east rift zone with the advent of instrumental advancement and expansion over the years are graphically outlined in figures 3a and 3b. Of 804 lower east rift earthquakes (mainly M_A 1.5 to 4.5) picked to be located in 1962-77, more than 50 percent were located with standard errors of 2.5 km or less. The continued expansion of the network and improvement in detection made it possible for more than 90 percent of the earthquakes to be located with a standard error of 2.5 km or less in 1975-77. Essentially all the events of $M_A \geq 3.0$ in 1975-77 were located within these constraints.

4. General Seismicity of Kilauea

Strain accumulated over the years of repeated volcanic activity would be expected to lead to a catastrophic event and be extensively relieved as it did when the 7.2-magnitude earthquake occurred on November 29, 1975. Because this was a recent event and well recorded instrumentally, the data collected during the aftershock period was ideal to examine the focal distribution (figs. 4a and 4b) and strain release (fig. 5) patterns at Kilauea that presumably would outline the volume subjected to stresses over prior decades.

The epicenter and depth plots of figure 4 include well-recorded events of $M_A \geq 3.0$ located with standard errors of about 2.5 km or less. The data for well-determined earthquakes of $M_A \geq 3.0$ from December 1975 to December 1977 is relatively complete. The epicenters of these events relate to a zone within the south flank of Kilauea that extends from the southwest rift zone east-northeastwards along the east rift zone to near Puu Honuaula and Puulena. The earthquakes center in the south flank adjacent to parts of the rift zone that were active with intrusions and eruptions in the past two decades. Consequently, the seismic concentration is highest in parts adjacent to the upper parts of the east rift. Most of the large events occur at 5-10 km depth near the base of the volcanic pile (Swanson and others, 1976). Earthquakes decrease sharply at depths below 10 km, except beneath the south summit area of the volcano, where earthquakes occur deeper and in the mantle down to nearly 60 km.

Although the east rift is known to extend many tens of kilometers seaward beyond the east cape of the island (Moore, 1971), crustal earthquakes do not persistently occur beyond Puu Honuaula. The activity ends there but extends normal to the rift zone to the south. Sections across the mid- and lower east rift zone (fig. 4b) further illustrate the concentration of earthquakes at 5-10 km depth south of the rift zone, and during times of eruptions, shallow earthquakes commonly occur beneath the rift nearly vertically from the surface to about 5 km in depth. There are few earthquakes below 10 km.

The pattern of strain release for the Kilauea area for the time period from December 1975 to December 1977 was examined from earthquakes of $M_A \geq 3.0$ and compiled in contour maps as shown in figure 5. The earthquakes were assigned to map grid squares 0.4 km on a side (divisions of one minute of latitude and longitude). Energy release values were calculated from magnitudes of earthquakes using Gutenberg-Richter's formula:

$$\text{Log } E = 9.9 + 1.9\, M_L - 0.024 M_L^2$$

where M_L is local earthquake magnitude, and E is energy in ergs (Richter, 1958). The square root of the energies calculated for individual earthquakes in each grid square were summed to determine the strain release value. The compilations were subdivided into three depth categories of 0-5 km, 5-10 km, and 10-15 km. Using the log values of the strain release sums in each grid and each depth category, the strain release quantities were contoured and mapped after a method used by Ryall and others (1966) to describe strain release patterns in the western United States.

The strain release plots complement the hypocenter plots with the added parameter of earthquake magnitude for the quantitative evaluation of seismic energy release. The relatively large earthquakes of $M \geq 2.5$ tend to describe the tectonic setting of the region, and they outline the most intensive seismic belt to be south of the rift zones and deep in the crust of the flank at 5-10 km depths. Shallow seismicity is localized southeast of the caldera where swarms of earthquakes associated with magmatic intrusions repeatedly occur. On the summit and rift zones of Kilauea, volcanic activity is commonly accompanied by swarms of small earthquakes within the shallow crust at about 0 to 5 km depths. Adjacent to these centers of volcanic activity and deeper in the crust beneath the south flank of the volcano, however, earthquakes occur more widely apart in time, magnitude, and space. The summit and rift microearthquakes associated with intrusive or eruptive activity are often referred to as "volcanic earthquakes," and their counterparts deeper in the south flank that accommodate long-term relief of stress generated by magmatic intrusions are alternatively called "tectonic earthquakes." Consequently, the treatment of small earthquakes in hypocenter maps and count plots without magnitude distinction tend to emphasize "volcanic" seismicity where large numbers of tiny earthquakes locally relate to shallow intrusions and eruptions. In contrast, strain release plots emphasize the tectonic aspect since they take into consideration the size of each earthquake and are heavily influenced by a few large earthquakes.

5. Lower East Rift Seismicity

A. Chronology of volcanic activity and earthquakes

The last volcanic eruption in the lower east rift zone took place in early 1960, and the related high seismicity is indicated in figure 6. Since 1960, Kilauea erupted numerous times, but always west of Heiheiahulu. During these times of repeated eruptive activity, seismicity on Kilauea's eastern flank occurred episodically, with an average rate of strain release at roughly 40×10^7 ergs$^{1/2}$ per year calculated from $M \geq 2.5$ earthquakes that happened there. The energy rate involved is about equivalent to one 4.0-magnitude earthquake per year. For nearly five years from 1971 to November 1975, the rate of strain relase was very low. The pattern then changed dramatically with the occurrence of a major 7.2-magnitude earthquake on November 29, 1975, in the western part of the zone. The aftershock activity covered nearly the entire subareal south flank of Kilauea from the summit to Puu Honuaula and caused the lower east rift strain release rate to increase by a factor of 3 for two years up to the time of the mideast rift eruption in September 1977. Since the eruption, seismic activity in the lower east rift assumed a lower rate of strain release nearly comparable to that of 1960-70.

During the 20-year period from 1960 to 1980, at least seven notable periods of high activity were detected in the lower east rift near Puu Honuaula, specifically in January-November of 1962, August-December of 1964, July-February of 1966-67, February-July of 1970, August-October of 1973, November-May of 1975-76, and May-September of 1980. Although such swarms were more frequent in 1960-68, our seismometer in Pahoa (PHA) was too distant to detect many of the tiny shocks. When the station was relocated 5 km southeast on the rift zone (PAX) in 1968-70, more of the small events were detected. During a swarm in 1970, a seismometer was operated simultaneously at Puu Honuaula (PHO) where, to our surprise, nearly an order of magnitude more of the micro-events were picked up (fig. 7). Station PHO later replaced PAX on the HVO seismic net. Portable seismographs operated at various localities confirmed that many small events of the swarm occurred near Puu Honuaula at depths of about 5 km or less. Moderate-sized earthquakes ($M \geq 2.5$) appeared to extend to greater depths (~10 km) and seaward over a wider area. This activity in early 1970 was characterized by pronounced 2-day to 2-week long swarms when detectable microearthquakes of 0.1 to 3.9 magnitude numbered nearly a thousand per day. The episodic nature of the activity could also be observed within intervals of several hours so that four distinct bursts of activity occurred in the two days of peak activity during the swarm in April. These microearthquake swarms were not immediately related to major magmatic movement because sustained harmonic tremor or deformation changes typically recorded during magmatic intrusions were not detected during these lower east rift swarms.

b. Magnitude-frequency of earthquakes

The magnitude-frequency parameter b was caculated according to the Gutenberg-Richter formula, $\log N = a-bM$ (where N is number of earthquakes, M is earthquake magnitude, a and b are constants). The b-value describes the relative number of earthquakes at different magnitude intervals, and was determined to be 0.93 from a total of 312 lower east rift earthquakes of $M \geq 2.0$ recorded inclusively between 1962 to 1975 (fig. 8). A comparative determination of b-value was made from crustal earthquakes beneath the entire Kilauea area. From a total of 3,056 earthquakes of $M \geq 2.0$ from the entire crustal area of Kilauea recorded for the same time period in 1962 to 1975, a slightly higher b-value of 0.97 was determined. Similar values ranging from .9 to 1.0 were found for other Hawaiian earthquakes (Furumoto, 1966; Koyanagi and others, 1966; Kinoshita, 1967).

Episodic swarms in the lower east rift consisted of many small shocks detected by an insufficient number of stations to be normally located. To obtain enough samples for a reliable b-value determination during such a swarm as that in February-May 1970, we extended our data set by including small earthquakes with magnitudes computed from signal duration measured from the nearest east rift stations PAX and PHO. A statistical fit of coda lengths and amplitude magnitude for small eastern rift earthquakes is shown in figure 9. The data set thus extended to include the smaller events of -1.0

to 2.0 magnitude during the peak of the 1970 swarm was used to determine a b-value of 0.71. The relatively low b-value may be speculated to represent high stress conditions in the limited focal zone near Puu Honuaula during the localized activity there.

6. Discussion

Seismic data collected in recent years, particularly during the intense activity following the major earthquake of November 1975, suggest that the lateral movement of magma from the summit of Kilauea into the rift zone takes place within a conduit complex at 5-10 km depth. The persistence of 5-10 km deep earthquakes in the south flank adjacent to parts of the rift zones where eruptions were most numerous over the past two decades imply a gradual lateral transport of magma near the base of the rift zones. Episodic intrusions and eruptions along the rift zones are fed nearly vertically and rapidly through local fissures formed from this linear storage complex below as evidenced by precursory swarms of shallow seismicity and complementary pattern of radial ground tilting in the locality of the volcanic outbreak.

Earthquake swarms in Hawaii have been typically associated with volcanic activity, in contrast to aftershock sequences that generally occur in peripheral parts of the volcano. The swarms of earthquakes in the lower east rift from 1968 to 1975, however, did not correlate with intrusive activity as typically determined by deformation techniques and conspicuous harmonic tremor. These swarms were probably related to magma movements in a secondary storage zone near Puu Honuaula, temporarily isolated from the principal magma conduit of Kilauea's east rift, and apparently of too small a size and intensity to perpetuate eruptions. Stresses generated by persistent volcanism to the west and from gravitational loads caused by the massive Mauna Loa volcano to the northwest could differentially stress the rift zone sufficiently to ultimately create a dynamic structural offset characterized by episodes of seismic activity. The seaward extension of the rift zone has remained aseismic during the past 15-year interval; it appears dislocated and not subjected to the normal stresses that apply to the upper rift zone. We agree with Fiske and Jackson that rift zone orientations are dictated by gravitational stresses, and along with Zablocki, we propose that the offset of eruptive features near Puulena and Puu Honuaula represents a structural disruption in the east rift where magma tends to accumulate and where changes in the stress pattern takes place. The seismic swarms could represent incremental southeastward displacements of the rift zone west of Puu Honuaula accompanied by the shifting of the stored magma there. The small earthquakes appear to initiate near Puu Honuaula and typically occur in swarms, whereas many moderate events that follow extend farther and deeper into the unstable south flank. The seaward component of the rift east of Puu Honuaula is farther from the central Kilauea activity, and being unaffected by Mauna Loa buttressing, may remain relatively stable until volcanic activity occurs in the area. Stresses induced by intrusion of magma are relieved immediately and not transmitted to the east, so that crustal seismicity east of Puu Honuaula may not persist. In summary, the intensity and frequency of microearthquake swarms at Puu Honuaula are dictated by the state of instablity of the structure caused by (1) refilling of magma from renewed intrusive activity there, (2) stresses generated from repeated magmatic activity along Kilauea's upper flanks, and (3) from constant gravitational loading due to Mauna Loa buttressing from the northwest.

Although seismic and self-potential surveys clearly indicate a dynamic structure and thermal gradient several kilometers beneath the locality south of Puu Honuaula (as substantiated by the experimental geothermal well HGP-A), deformation surveys and preliminary geologic mapping do not provide conclusively near surface evidence for the implied transform faulting along the east rift zone there. The eventual results from detailed geologic mapping of the area currently undertaken by the U.S.G.S. (R. Moore, personal communication) should add to the understanding of the volcanic and tectonic processes.

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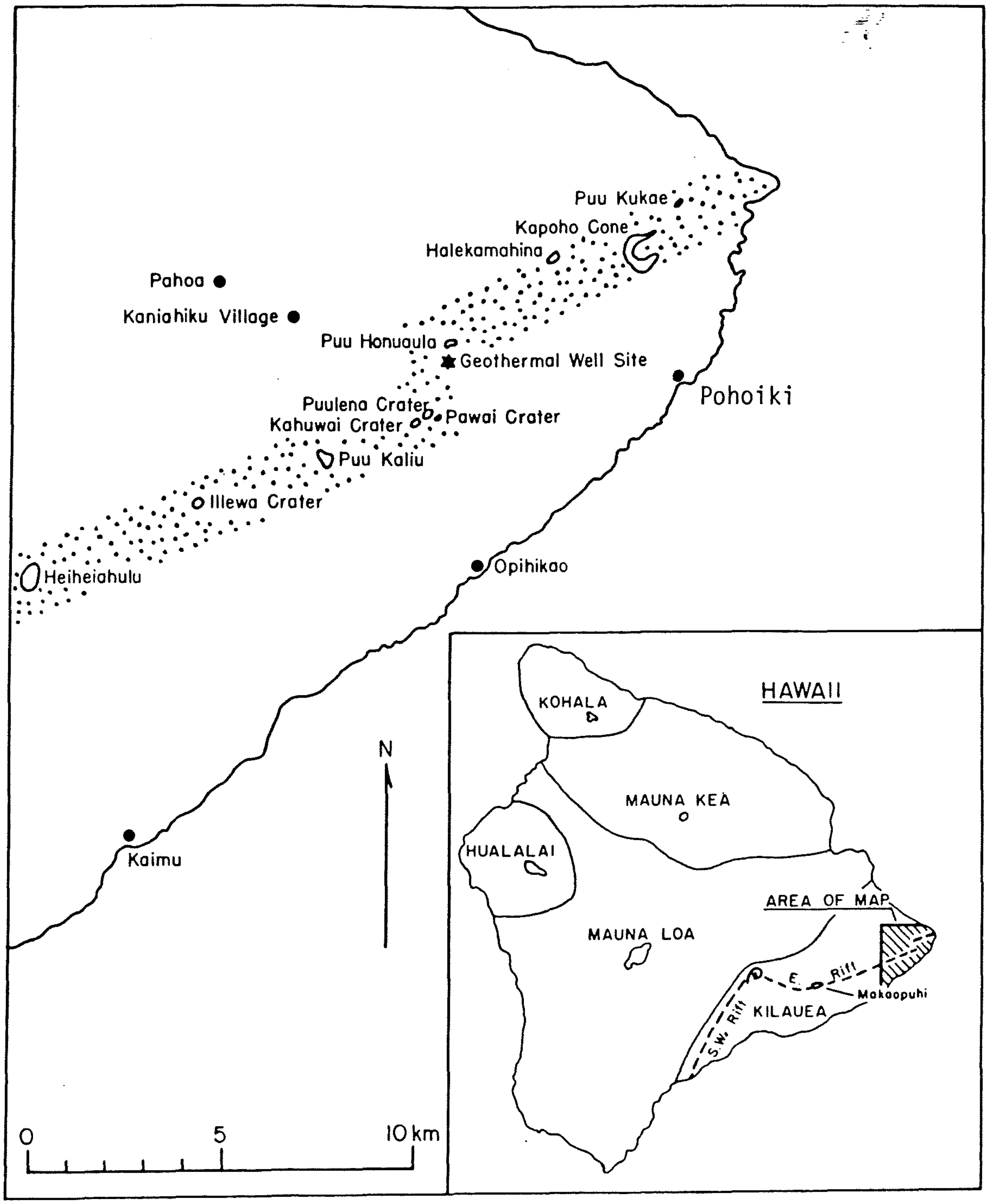


Figure 1. Map of the Puna district of Hawaii with volcanic features of Kilauea with the topographic crest of the rift zone delineated by the dotted area, the geothermal well site, and nearby villages. The inset map of the island of Hawaii in the lower right corner includes the summit and bounding areas of the 5 volcanoes; in the southeast part of the island at Kilauea Volcano, the east and southwest rift zones are shown in dashed lines, Makaopuhi crater is labelled, and the principal area of our report is hatchured.

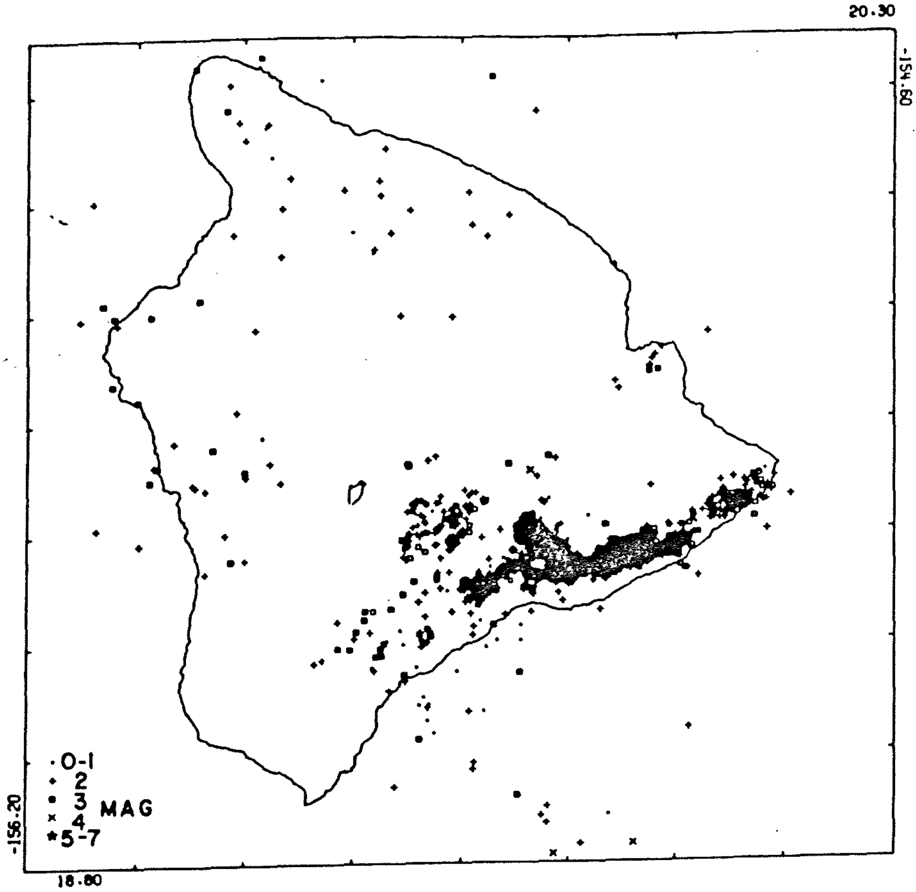
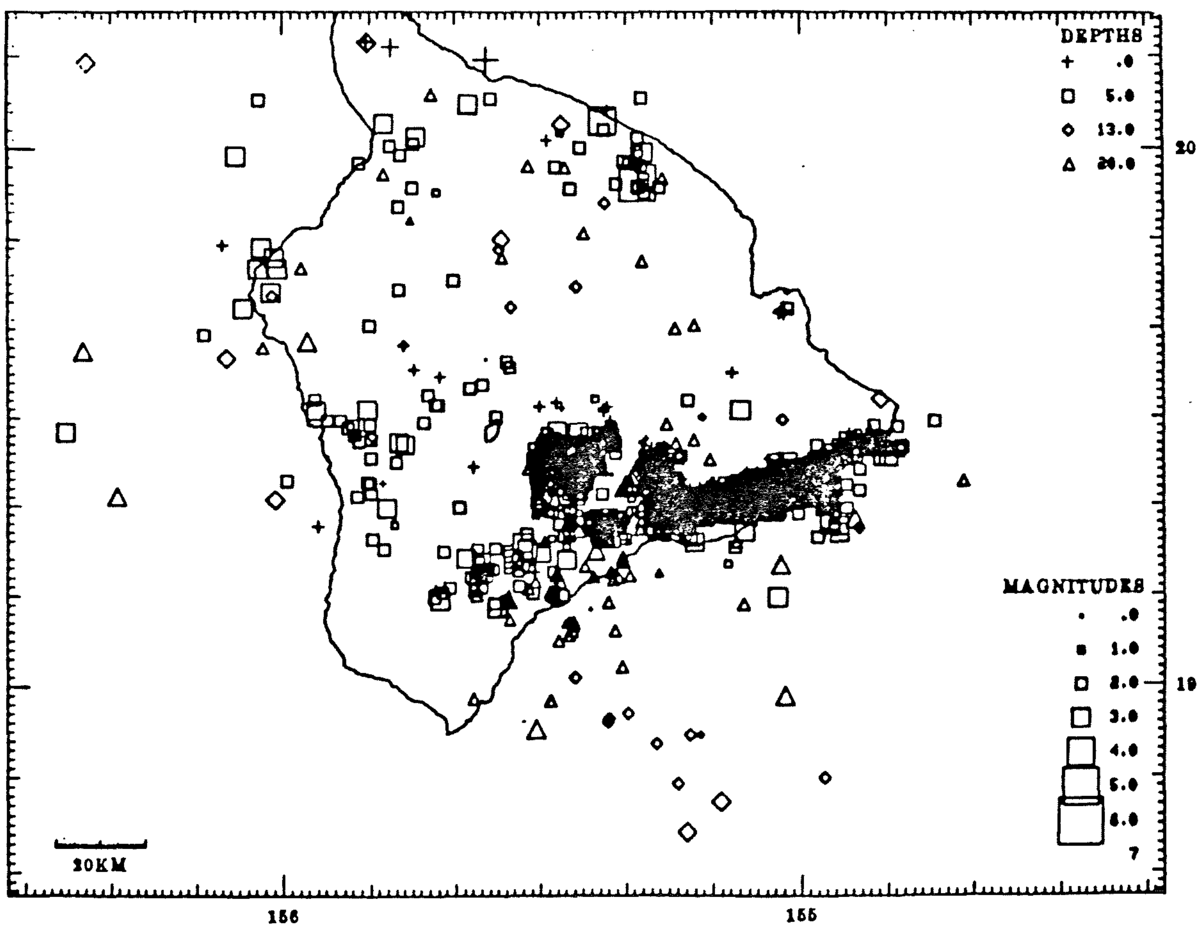


Figure 2. Epicenter plots presented in Hawaiian Volcano Observatory summaries 76 and 77 showing the concentration of earthquakes in the south flank of Kilauea along a belt that extends from the subareal southwest rift eastward to about Puu Honuaula on the east rift. The seismicity at Kilauea was particularly high during these years following the 7.2-magnitude earthquake in southeastern Kilauea on November 29, 1975, and when several volcanic episodes occurred.

Epicenter plot of all events located for the year 1976.



Epicenter plot of all events located for the year 1977.

Figure 3a. Histogram indicating the quality of focal determination for lower east rift earthquakes in terms of standard errors of epicenter and depth. The plot shows percentages of the total number of earthquakes located for 5 consecutive time periods indicated in years. The number of earthquakes sampled are in parenthesis beneath each sequential time periods, and mainly include 1.5- to 4.5-magnitude events recorded on 6 or more stations on the seismic net.

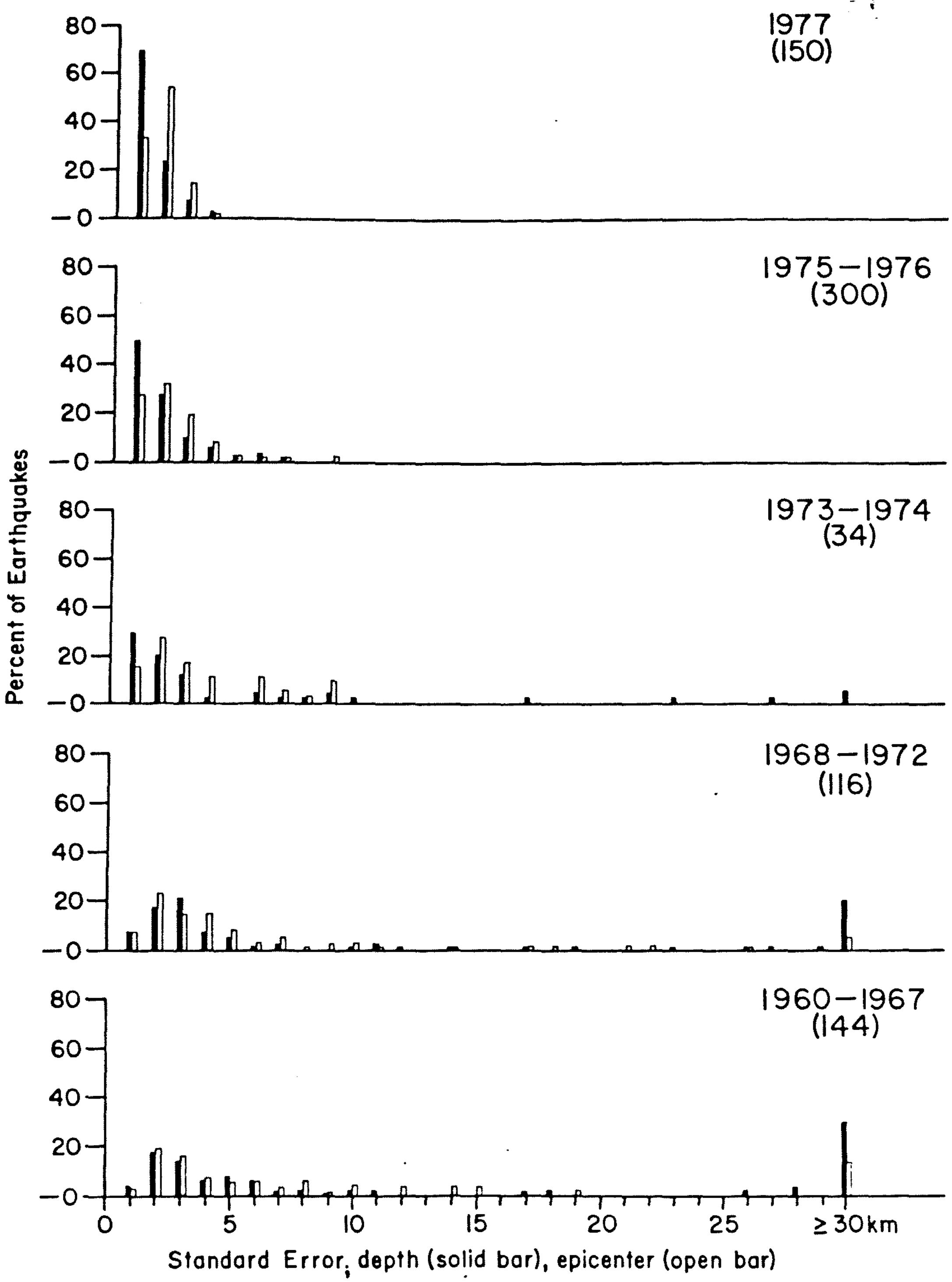
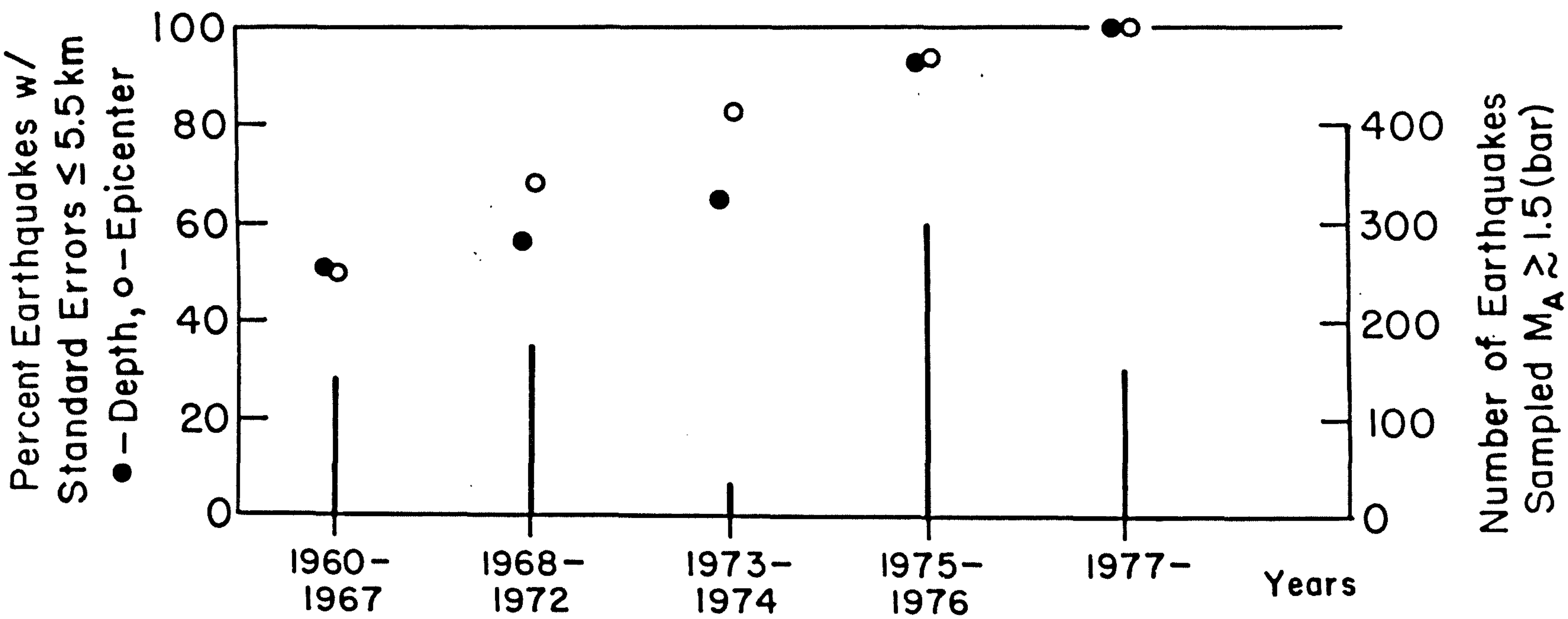


Figure 3b. Percentage of earthquakes located with standard errors of 5.5 km or less, and the number of earthquakes located for different time periods between 1960-1977. Open circles represent standard errors in epicenter, and dots represent standard errors in depth. Number of earthquakes sampled are indicated by vertical bars. Of the 804 lower east rift earthquakes of about 1.5 to 4.5 magnitude picked and located at HVO in 1962-1977, more than 50% were located with standard errors of about 5.5 km or less. About 80% of these were located with standard errors of 2.5 km or less; these were selected to represent earthquake focal distribution in the lower east rift zone.



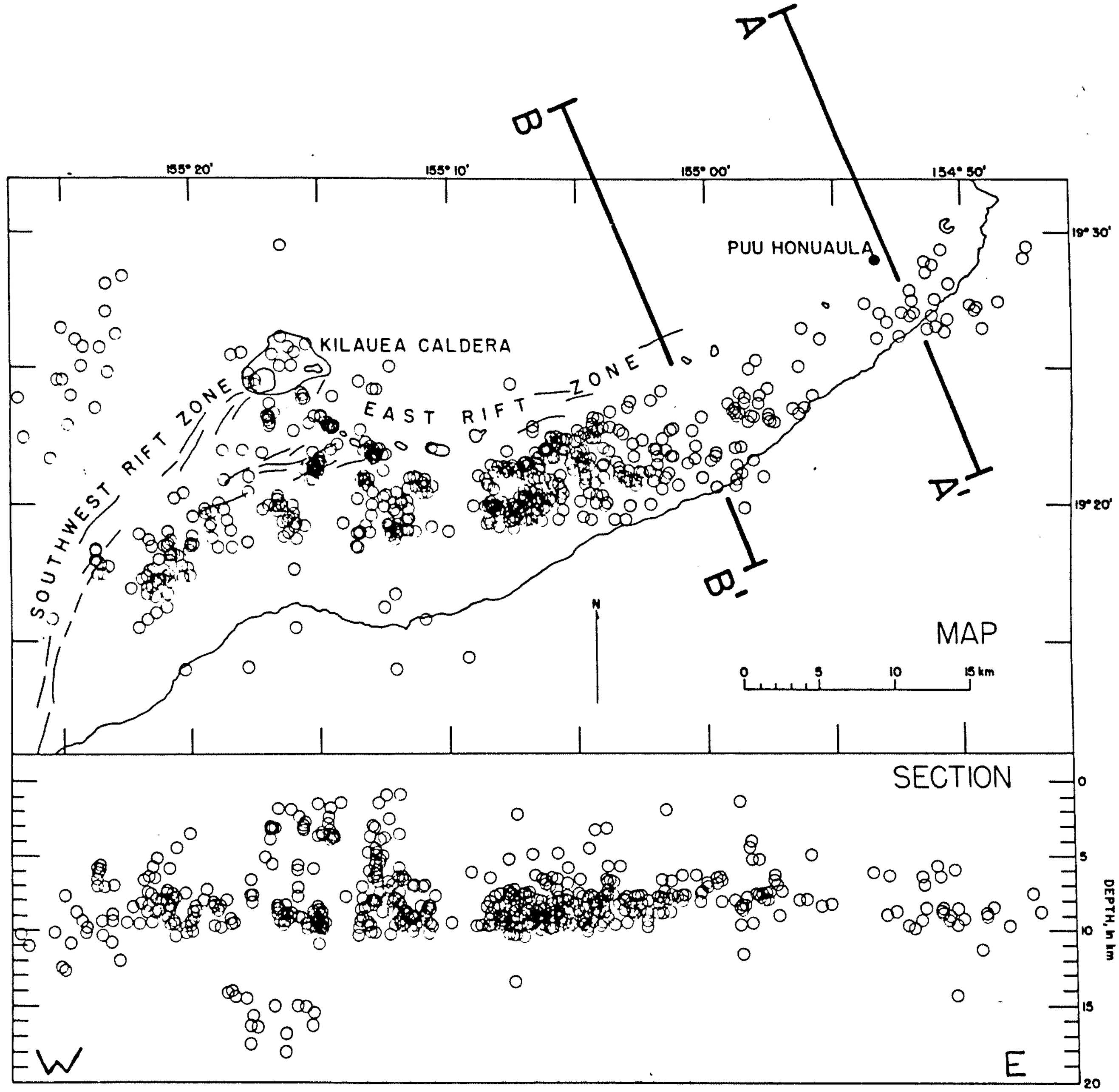


Figure 4a. Locations of $M_A \geq 3.0$ earthquakes with focal depth of 0-20 km at Kilauea from December 1975 to December 1977. The quality of location for most of the events plotted are less then 2 km in standard errors of epicenter (upper plot) and depth (lower plot). A-A' and B-B' are center lines for cross-sectional hypocenter plots prepared in the following figure 4b.

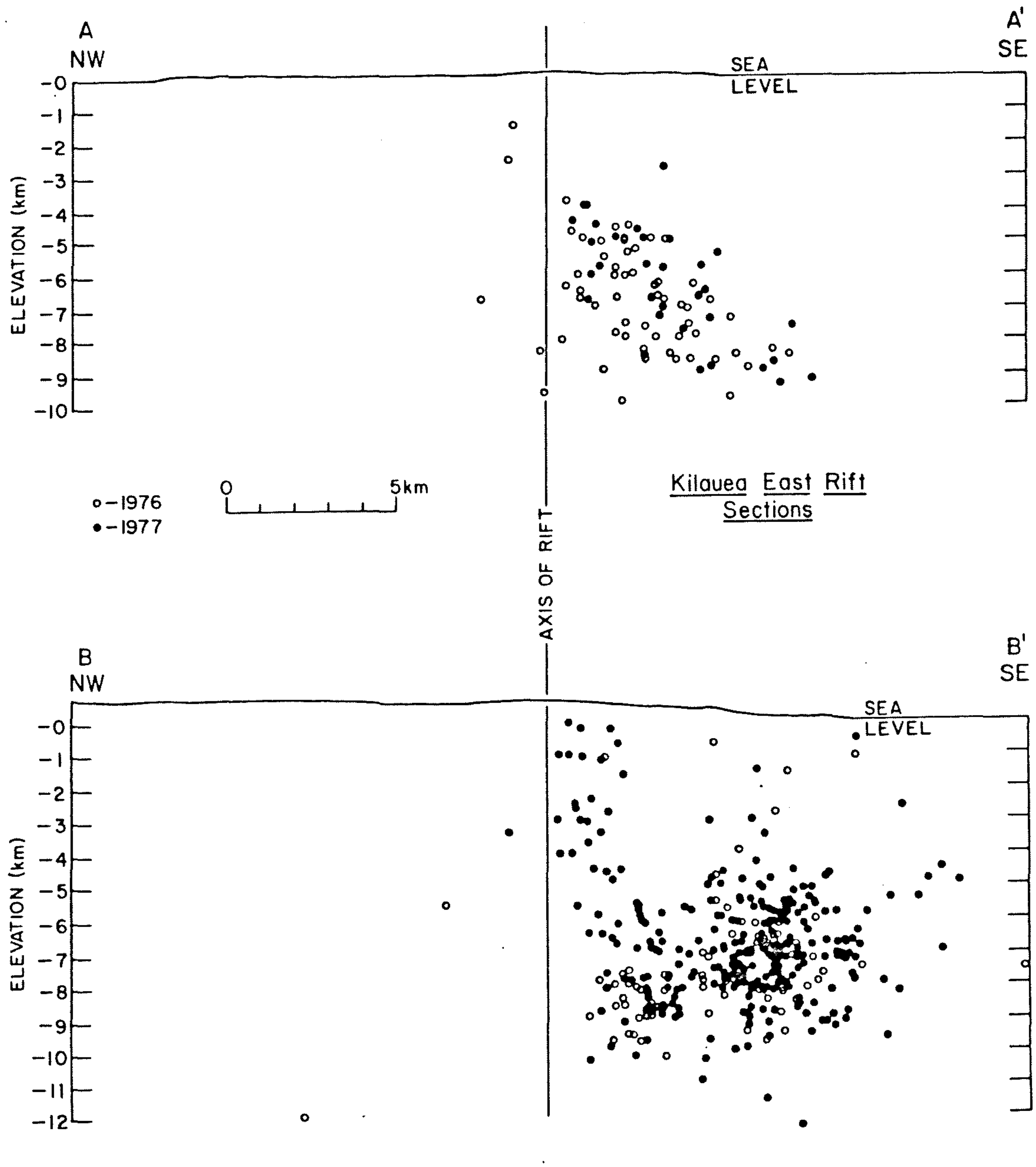


Figure 4b. Earthquake hypocenters plotted within 10-km wide sections centered on lines A-A' and B-B' oriented normal to the rift zone as shown in figure 6a. Earthquakes of all magnitudes and with standard errors in epicenter and depth less than 2.5 km are included. Hypocenters for earthquakes in 1976 are marked with open circles and those in 1977 are indicated by black dots. The column of progressively shallow earthquakes immediately south of the axis of the rift in the middle east rift zone in section B-B' mainly occurred about the time of the September 1977 mid-east rift eruption.

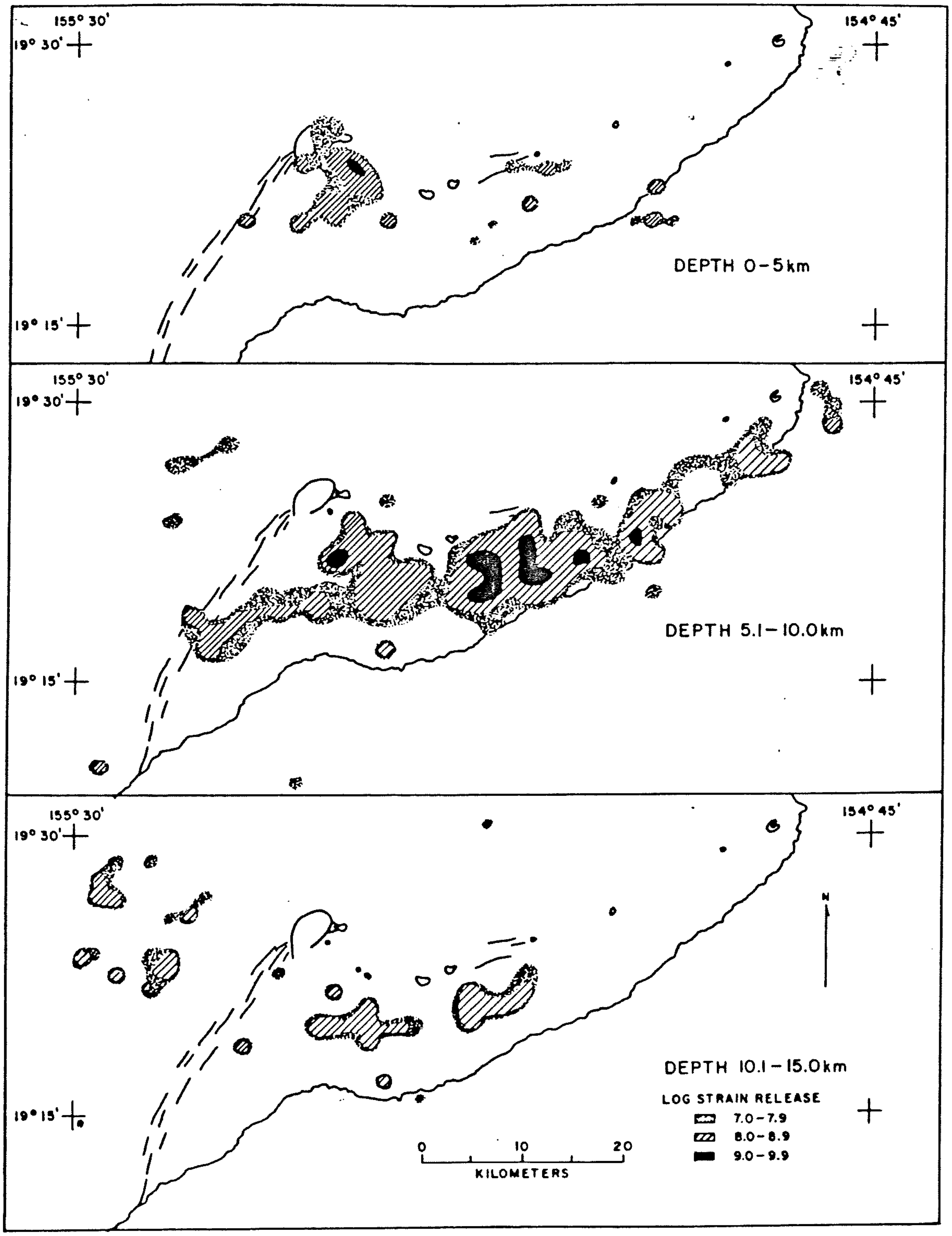


Figure 5. Strain release maps at 3 depth intervals prepared from earthquakes of $M_A \geq 3.0$ at Kilauea, December 1975 to December 1977.

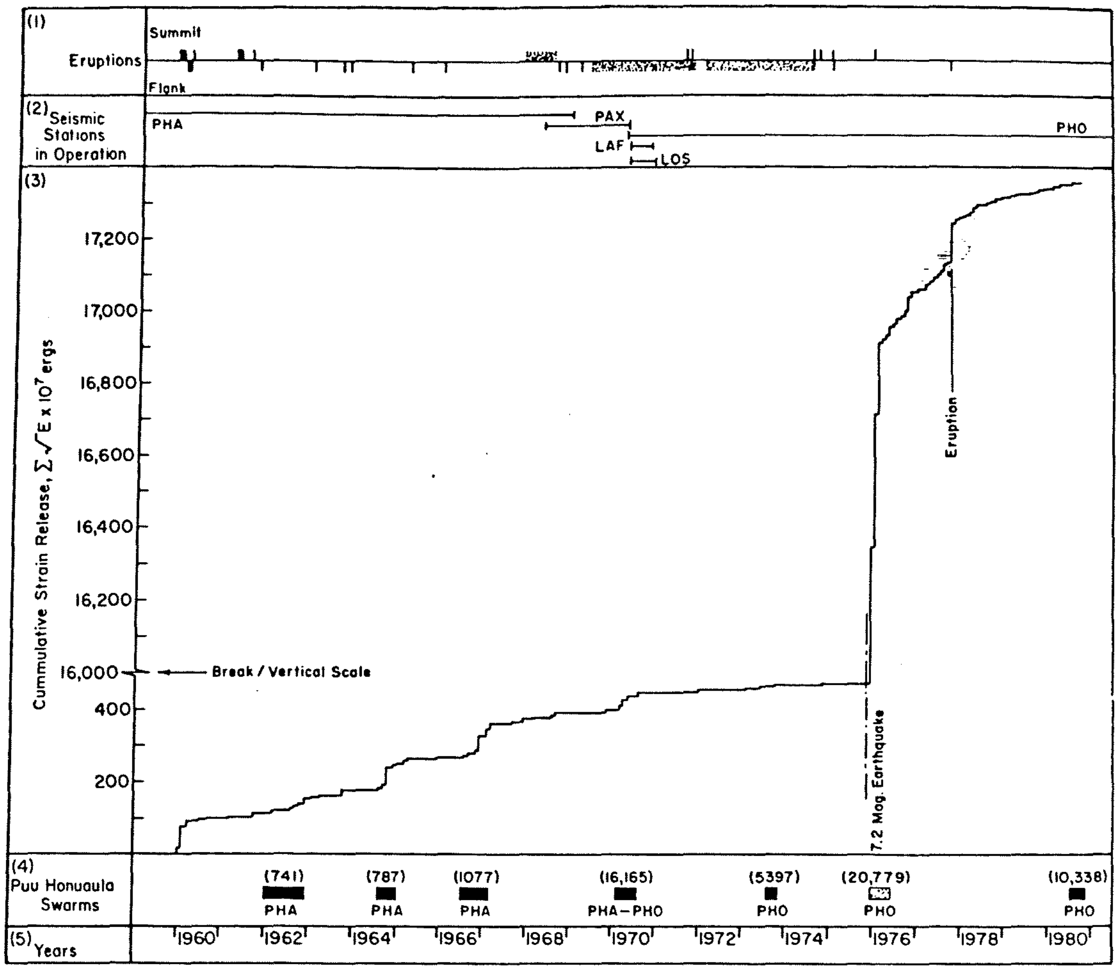


Figure 6. Chronological Summary of Kilauea volcanic events and microearthquake activity recorded from January 1960 to September 1980.

(1) Kilauea volcanic events where solid bars represent times of well-defined eruptions, and dotted bars represent times of prolonged eruptions characterized by numerous phases of activity. Summit outbreaks are shown above and flank outbreaks are shown below.

(2) Seismic stations operated in the lower east rift zone: PHA was an optical drum-recorded seismograph operated at Pahoa High School; PAX at Kaniahiku Village was telemetered to the Observatory and recorded on Develocorder microfilm; PHO at Puu Honuaula was telemetered to the Observatory, and recorded on Develocorder microfilm and magnetic tape; LAF and LOS near Puu Honuaula were telemetered to the Observatory and recorded on Develocorder microfilm.

(3) Strain release plotted cumulatively from earthquakes in the lower east rift zone extending from about Heiheiahulu to Cape Kumukahi. Seismic energy release was determined for each crustal earthquake of $M_A \geq 2.5$ in the area using the energy-magnitude relation Log E = $9.9 + 1.9M_L - 0.024M_L^2$ (Richter, 1958). The square root of the calculated energy release for each earthquake was determined, and the quantities for all the earthquakes were summed and plotted cumulatively at successive monthly intervals to represent the rate of strain release from 1960 to mid-1980. For convenience in plotting, a break was made in the vertical scale to accommodate the relatively high strain release from the 7.2-magnitude earthquake on November 29, 1975.

(4) Times of increase or swarms of shallow microearthquakes in the lower east rift local to Puu Honuaula (1962-1980) are blocked in, and in parentheses above are indicated the approximate number of microearthquakes detected at the particular stations then in operation; the notable periods of high activity were January-November in 1962, August-December in 1964, July-February in 1966-67, February-July in 1970, August-October in 1973, November-May in 1975-76 that continued for many months more during an intense aftershock activity (dotted area), and June-September in 1980.

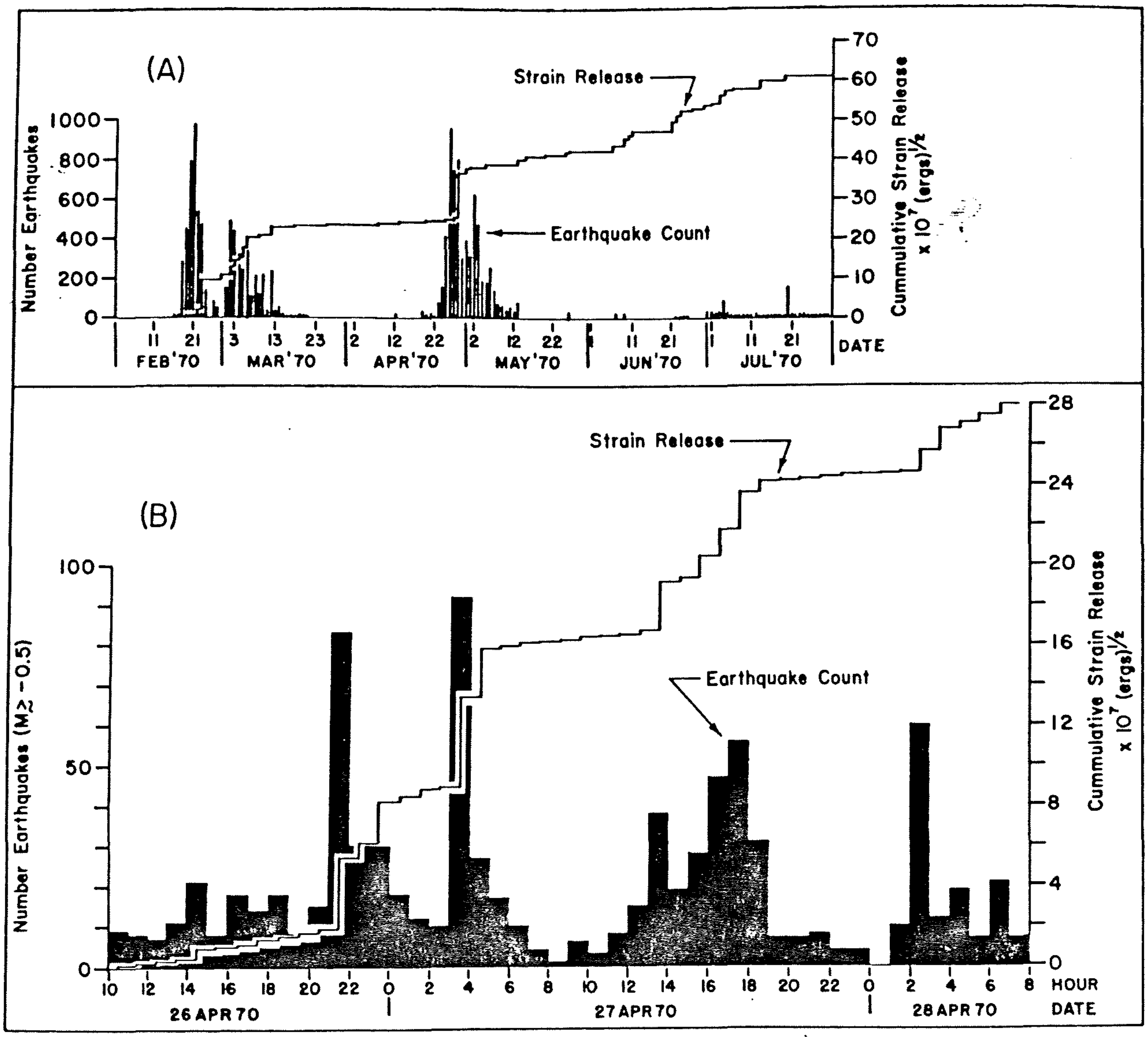


Figure 7. Earthquake frequency and strain release in the lower east rift during the swarms in 1970 characterized by discrete bursts of activity.

(A) Daily plot of earthquake frequency (vertical bars) and cumulative strain release (continuous line) describes the two 3-week-long swarms between February to May.

(B) Expanded hourly plot of earthquake frequency (vertical bars) and cumulative strain release (continuous line) for the peak period of activity on 26-28 April during the second swarm reveals short-term episodic strain release at 6- to 10-hour intervals.

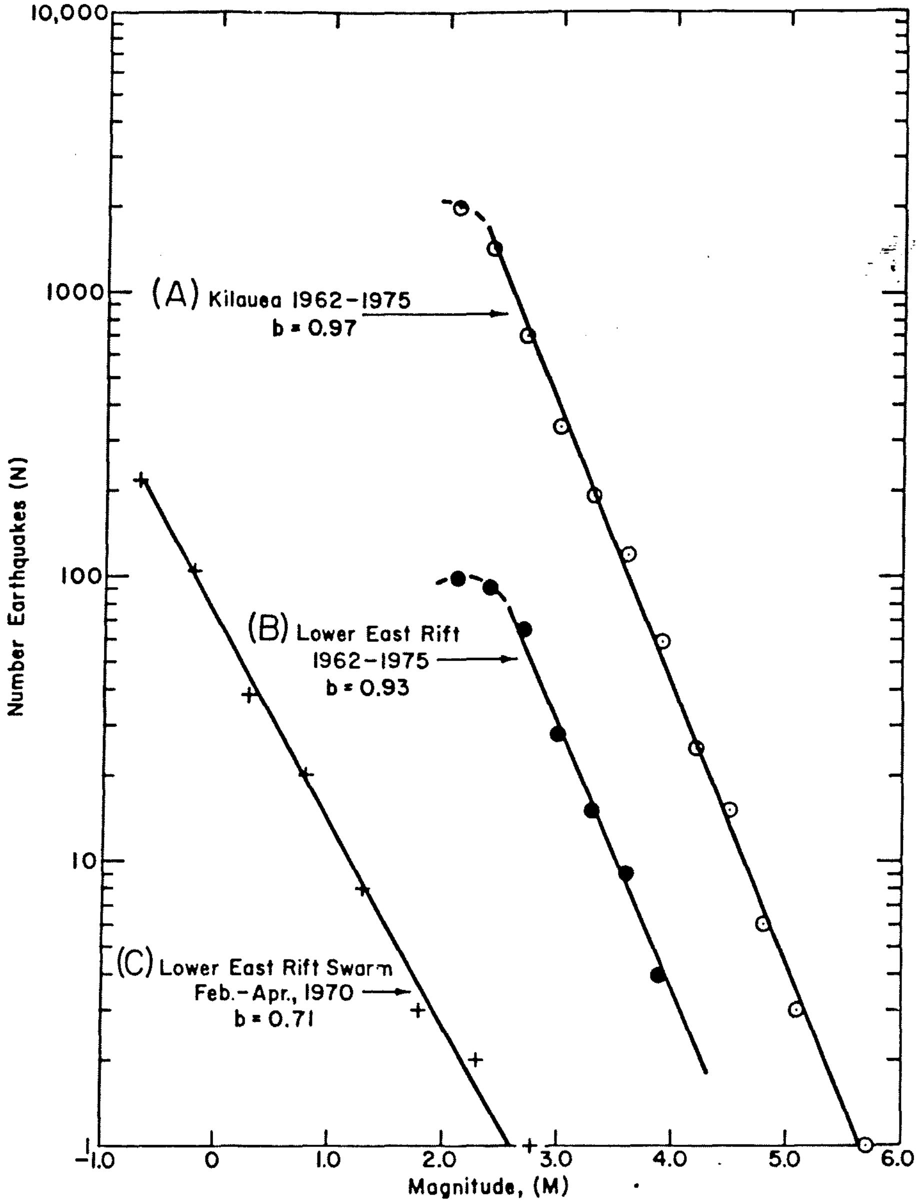


Figure 8. Magnitude-frequency relation for three categories of Kilauea earthquakes of crustal depths: (A) for all earthquakes in 1962 to 1975, before the 7.2-magnitude earthquake on November 29, 1975, the b-value was determined to be 0.97; (B) for earthquakes located near and east of Heiheiahulu in the lower east rift zone during the same time period as above, the b-value was determined to be 0.93; (C) for earthquakes from a lower east rift swarm near Puu Honuaula in February-April 1970, the b-value was determined to be 0.71. For the A and B groups, well-recorded earthquakes routinely picked and located at HVO were included, and calculation of magnitude for these earthquakes was based on trace amplitude measured from the Wood-Anderson seismograph (M_A). For the C group, the data set was extended to include earthquakes that were mostly too small for routine focal determination, and magnitude calculation for these earthquakes was based on signal duration measured on the local stations PHO or PAX (M_{F-P}). The earthquakes were compiled sequentially at 0.3-magnitude intervals for the A and B groups and at 0.5-magnitude intervals for the C group, and fitted a line for each plot by least squares. For the routinely processed events in A and B, the magnitude threshold of the data set below which the count drops off was indicated by dashed lines; in A, the data is relatively complete for earthquakes of $M_A \gtrsim 2.3$, and in B, the data is relatively complete for earthquakes of $M_A \gtrsim 2.5$.

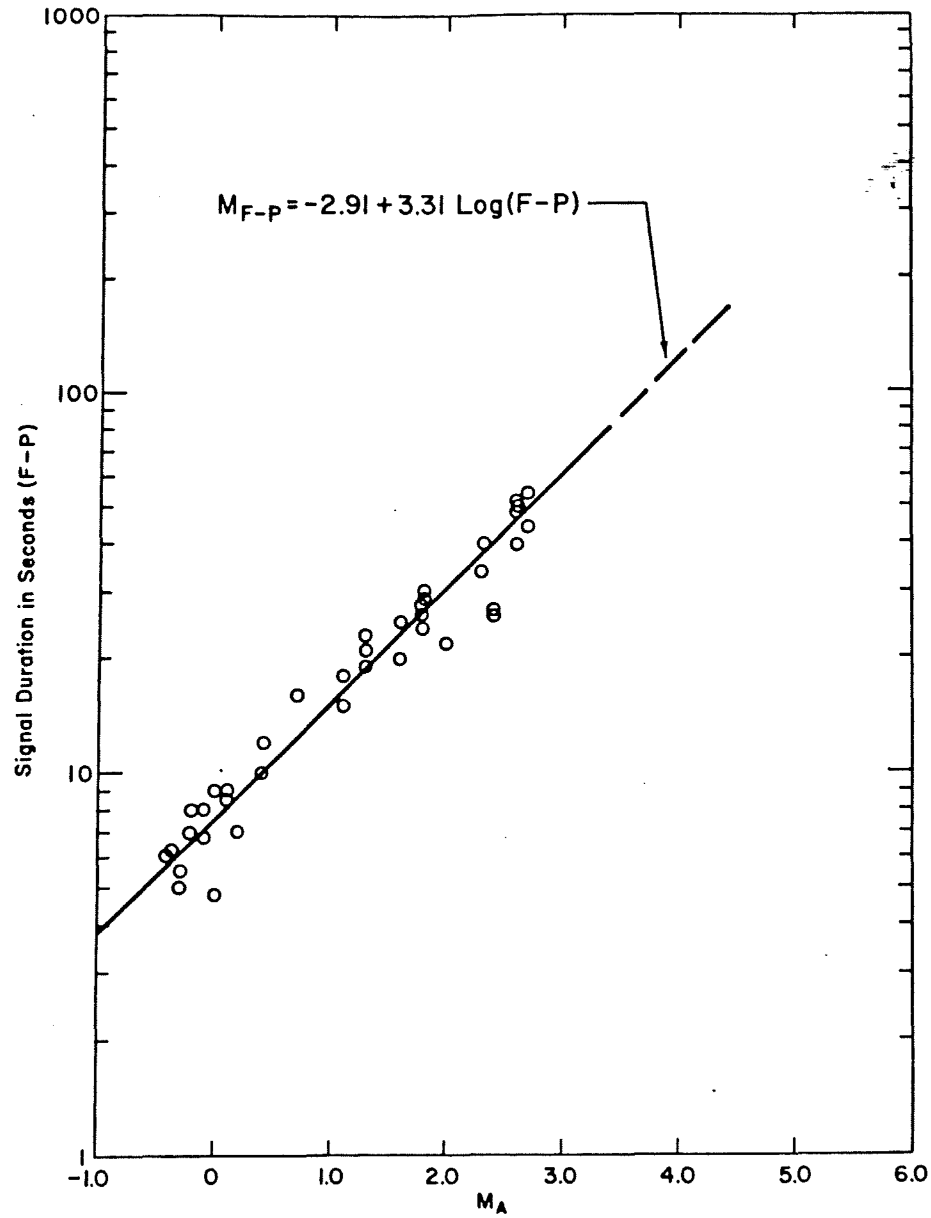


Figure 9. Magnitude scale for lower east rift earthquakes derived from a logarithmic relation of signal duration and amplitude magnitude. Corrections based on trace amplitudes from Wood-Anderson seismographs were applied to the more sensitive local stations PHO and PAX to determine amplitude magnitudes for the smaller rift earthquakes. Focal distances necessary for the amplitude-magnitude determination of the events too small to be located by the usual procedure using P-wave times were, alternatively, estimated from the time interval of P and S waves. Thirty-eight selected lower east rift earthquakes ranging in amplitude-magnitude from -0.4 to 2.7 were plotted against the logarithm of their respective signal duration in seconds. The line through the locus of the points gives the relation $M_{F-P} = 2.91 + 3.31 \log (F-P)$; where F-P is the signal duration in seconds, and M_{F-P} is the magnitude determined by signal duration. This relation was used to calculate magnitude for the rift earthquakes too small for normal processing and used as a means to expand the quantity of earthquakes required for magnitude-frequency and strain release assessments detailed for short increments of time.