SEISMICITY, CRUSTAL DEFORMATION, AND SEISMIC RISK FROM EARTHQUAKES ON MAUI, MOLOKAI, AND LANAI

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MASTER OF SCIENCE

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ABSTRACT

Four seismometers have been installed on Maui by the Hawaii Institute of Geophysics to monitor local seismicity and crustal deformation. During the monitoring period from mid-May 1980 until mid-August 1981, this array detected one hundred twenty-four local events. Twenty-six of these were quarry blasts. Twenty-five of the earthquakes yielded reliable hypocentral locations with residuals less than 0.50 seconds and standard deviations less than twenty kilometers. The locations indicate that there are two main areas of seismicity near Maui: beneath Haleakala Volcano and between Molokai and Lanai. Seismic activity was also detected between Maui and the island of Hawaii, north of Maui, and north of Molokai. Most seismic activity seems to be related to the loading of the crust. The Molokai Fracture Zone does not appear to be seismically active. The seismicity study results, as well as data obtained from a geodetic network established by the Hawaii Institute of Geophysics, suggest that seismic crustal deformation is not geodetically measurable in the Maui-Molokai-Lanai region. Current and historical seismicity are analyzed to determine the seismic risk from earthquakes with source locations near Maui, Molokai, and Lanai. The 6 3/4 magnitude 1938 Maui earthquake has been relocated to 21.02° N. and 156.09° W., just 0.2° south of its 1938 location. Magnitudes of 6.8 and 6.9 were calculated. The revised depth estimate (about 20 km based upon P and pP wave arrivals) is considerably less than the 1938 estimate of 65 miles. For the Gutenberg-Richter relation (Log N = A - bM) the respective A and b values obtained are 4.28 and 0.50 for the monitoring

period and 4.31 and 0.67 for the past twenty year period. The b values for the recent seismic study and the past twenty years are significantly lower than known Hawaiian Island b values, suggesting that the Maui-Molokai-Lanai region is under stress. Strain release data for the last twenty years show that the Maui region releases strain with earthquakes of magnitude 3 to 5 every 2 to 5 years. This result may be of use in predicting the possible magnitude of future earthquakes. Earthquake recurrence rates are calculated for the monitoring period and the past twenty years. Recurrence rates for the monitoring period are unreliable due to a strong dependence on the March 5, 1981 magnitude 5.0 Molokai earthquake and its aftershocks. The recurrence rates for the past twenty years indicate a magnitude 7 earthquake could occur once every 250 years. Evaluation of the 1938 Maui earthquake and the magnitude 7.1 1871 Molokai earthquake indicate that seismic risk for Maui, Molokai, Lanai, and Oahu has been underestimated. Maui, Molokai, and Lanai should be within seismic risk Zone 3 and Oahu within Zone 2.

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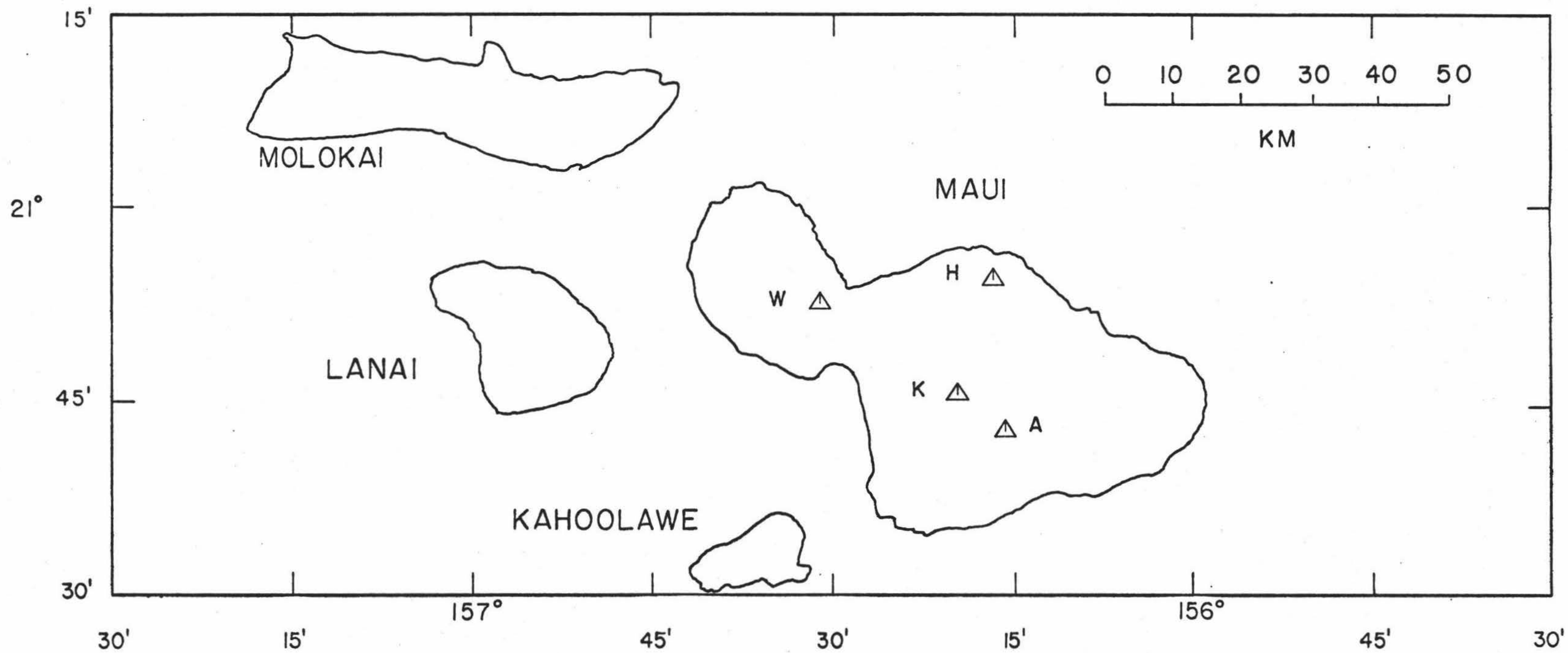
CHAPTER I

INTRODUCTION

The National Aeronautic and Space Administration began to operate the Lunar Ranging Experiment (LURE) in 1965 (Bender, et. al., 1973). One of the experiment's purposes has been to measure the relative movement of lithospheric plates (Bender and Silverberg, 1975; Carter et. al., 1977). Two stations, McDonald Observatory in Texas, and Haleakala Lunar Ranging Observatory on Maui were selected for the project because their separation allowed polar motion studies and the measurement of relative plate movement between the North American and Pacific plates (Bender et. al., 1973; Bender and Silverberg, 1975; Berg et. al., 1978). Distances were measured from these two observatories to retroreflectors left on the moon by the Apollo astronauts in order to obtain the geocentric location of each observatory. The secular change in either observatory's geocentric position would be a function of plate movement and crustal deformation (Bender and Silverberg, 1975; Carter et. al, 1977). In order to obtain the rate of plate movement alone, corrections for local and regional crustal deformation at both stations needed to be determined. The Hawaii Institute of Geophysics (HIG) in 1976 began to establish a coordinated system of geophysical observations designed to monitor local and regional crustal deformation near Haleakala Observatory. The system consisted mainly of high precision (0.3 ppM) laser distance measurements, forming a geodetic network linking Haleakala Lunar Ranging Observatory with the islands of Maui, Molokai, and Lanai (Berg, et. al., 1978; Berg, et. al., 1981). Gravimetric surveys, tiltmeters, tide gauges, and seismometers comprised the remaining portion of the system (Berg, et. al, 1978). Four seismometers near Wailuku, Haiku, Kula, and the summit of Haleakala (Fig. 1) provided data for an investigation of the microseismicity near Maui.

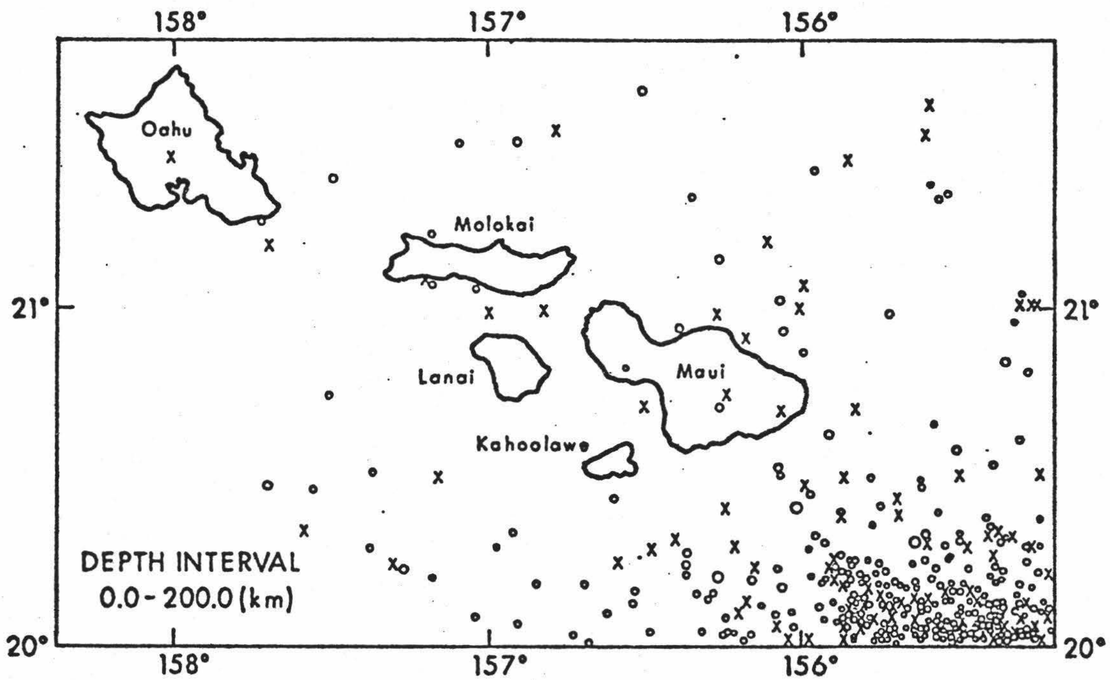
This is the first study of microseismicity exclusively confined to the Maui-Molokai-Lanai region. Furumoto and others (1973) examined seismic risk in the Hawaiian Islands. They described and listed some of the larger earthquakes that have occurred near Maui, but most of the study focused on the island of Hawaii. Another study of Hawaiian seismicity (Estill, 1979), briefly treated earthquakes in areas other than the island of Hawaii (henceforth referred to as the Big Island). Using earthquake epicenters from the Hawaiian Volcano Observatory (HVO) seismic network and data gathered from an ocean bottom seismometer array, Estill noted particular areas of seismic activity between Maui and the Big Island, Molokai and Lanai, north of Molokai and Maui, and on Maui (Fig. 2). Estill stated that events north of Maui and Molokai may be related to the Molokai Fracture Zone or to submarine landslides. Earthquakes beneath the channel separating Molokai and Lanai may also be related to the Molokai Fracture Zone. He also cited magma movement as a possible cause of earthquakes on or near Maui, and stated that events south of Maui may be be caused by slumping along offshore fault scarps. Moore (1964) also proposed that some events north of Molokai may be

Figure 1. The Locations of the HIG Maui Network Seismometers
A= ARPA, K= Kula, H= Haiku, W= Wailuku.



15'
21°
45'
30'
157°
156°
30'
15'
45'
30'
15'
45'
30'
0 10 20 30 40 50
KM
MOLOKAI
LANAI
KAHOOLAWE
MAUI
W
H
K
A

Figure 2. Historical Earthquake Epicenters Near Maui, Molokai, and Lanai. The Epicenters With an "o" Were Located With Residuals Less than 0.5 Sec. The Epicenters With an "x" Were Located With Residuals Greater than 0.5 sec. (After Estill, 1979).

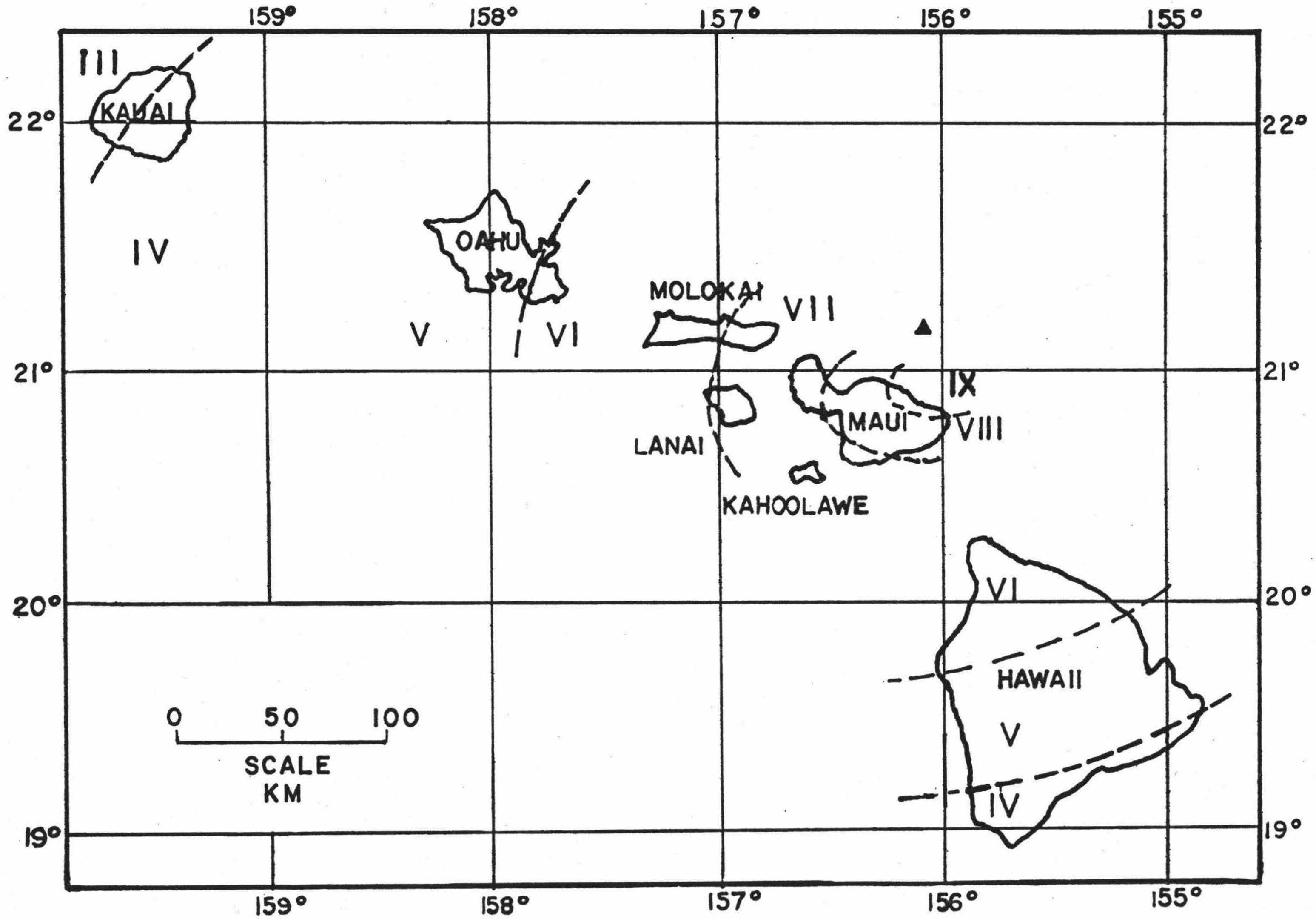

158°
157°
156°
21°
20°
Oahu
Molokai
Lanai
Maui
Kahoolawe
DEPTH INTERVAL
0.0 - 200.0 (km)

caused by submarine landslides. J. A. Carter (1978) suggested that earthquakes in the Maui region could be caused by magma movement or by loading the crust with the mass of the Hawaiian Islands.

Offsets along faults or in response to volcanic activity near Maui may alter the geocentric position of the Lunar Ranging Observatory at the summit of Haleakala Volcano. The most recent eruption of the volcano occurred in 1790 and, although it is now considered to be dormant, another eruption is not unlikely (Carter, et. al., 1977). Another source of significant crustal activity could be the possible seismicity associated with the Molokai Fracture Zone, which passes north of Maui. The Molokai Fracture Zone may have been associated with large historic events such as the 7.1 magnitude 1871 Molokai earthquake and the 6 3/4 magnitude 1938 Maui earthquake (Furumoto, et. al., 1973). As stated in a paper concerning the tectonic setting of Haleakala Observatory, "Vertical and horizontal displacements of several centimeters occur even at some distance from such shallow focus events." (Carter, et. al., 1977). Also, a major (7.1 magnitude) earthquake occurred in 1975 on the Big Island, just 150 km from the observatory. As crustal movement caused by earthquakes could modify the observatory location relative to the Pacific plate, the amount of local and regional crustal deformation must be determined. The results of a study of local seismicity from mid May-1980 until mid-August 1981, coupled with historical seismicity data, are used to define zones of earthquake activity near the Lunar Ranging Observatory.

Damaging earthquakes that have occurred outside of the Big Island, such as the 1871 Molokai earthquake and the 1938 Maui earthquake pose a significant threat to densely populated areas and tourist resorts on Maui, Molokai, Lanai, Oahu and the Big Island (Furumoto, et. al., 1973). A significant result of this study is the determination of earthquake frequency and seismic risk from earthquakes near Maui, Molokai, and Lanai using the results from the 1980-1981 seismic study and HVO earthquake data for the past twenty years. The 6 3/4 magnitude 1938 Maui earthquake remains one of the more significant seismic events in the history of Hawaii. Extensive damage occurred on Maui and Molokai. Many people on Maui, Molokai, Lanai, Oahu, and Kauai during the quake commented that it was the strongest tremor they had ever experienced (Volcano Letter, 1938); yet, there has been no formal study of the earthquake in the technical literature. Further, the hypocenter of the earthquake (Volcano Letter, 1938) seems suspect: the epicenter (21.2° N. and 156.1° W.) lies outside the area of maximum intensity (Fig. 3) and the estimated depth of 65 miles appears abnormally deep for Hawaiian earthquakes, which usually have maximum depths of 40-50 km (Estill, 1979; HVO seismicity summaries). The reports and descriptions of the 1938 Maui earthquake have been compiled and the earthquake has been relocated using the ATD relocation model (Spence, 1980).

Figure 3. Isoseismal Map for the 1938 Maui Earthquake. The Epicenter is Indicated by a Solid Triangle. Note that the Epicenter Lies Outside the Highest Intensity Isoseismal.


159°
158°
157°
156°
155°
22°
21°
20°
19°
III
KAUAI
IV
OAHU
V
VI
MOLOKAI
VII
IX
MAUI
VIII
LANAI
KAHOOLAWE
VI
HAWAII
V
IV
0 50 100
SCALE
KM

CHAPTER II

HYPO71 AND THE CRUSTAL MODEL FOR EARTHQUAKE LOCATION

HYPO71 (Revised), hereafter referred to only as HYPO71, is a computer program for determining hypocenter, magnitude, and the first motion pattern of local earthquakes (Lee and Lahr, 1975). For hypocenter locations HYPO71 requires P (and S) wave arrival times, seismometer locations, and a layered, laterally homogeneous crustal model that approximates the actual crustal structure. This chapter primarily treats the third requirement. The crustal model used was constructed using results from quarry blast data, geophysical experiments performed near Maui, and studies of the geology of Maui. A short summary is given below on the operation of HYPO71 and the options used to locate the earthquakes and quarry blasts.

The Operation of the Earthquake Location Program HYPO71

HYPO71 was used to locate all of the earthquakes and quarry blasts in this study. The program uses Geiger's Method of earthquake location to compute local event hypocenters and origin times (Lee and Lahr, 1975). Using arrival times and a layered, laterally homogeneous crustal model, HYPO71 locates events by minimizing travel time residuals through a step-wise multiple regression (Lee and Lahr, 1975).

HYPO71 allows the user many options in locating events. Among the most important options are station corrections, two types of crustal models, trial and alternate epicenters, trial and alternate focal depths, Vp/Vs ratio, and the weighting of arrivals. The program assumes that all receivers are on the datum plane. For a station above or below the datum a correction can be added to account for the elevation difference. HYPO71 also allows one to select either of two crustal models: the station delay model or the variable first layer model. For the station delay model the station correction is simply added to the calculated travel time to each station, while the variable first layer model converts the correction to an equivalent first layer thickness using the first layer velocity (Lee and Lahr, 1975). However, the station corrections may not consist entirely of topographic effects; the velocity above the datum plane may be different from the first layer or there may be variations in the local geology near a seismometer. For these reasons the station delay model was chosen. One of the most critical options deals with the initial epicenter. One can choose a trial epicenter from which the program iterates to a solution or the user can allow the program to start at a default epicenter approximately 0.3 km from the station with the earliest arrival (Lee and Lahr, 1975). Of particular importance for quarry blast studies, the program can fix the known origin time and hypocenter of a blast, so that theoretical travel times to various stations can be determined and compared with the data. The expected depth of earthquakes near Maui varies from a minimum of 1 km, for earthquakes caused by submarine slumping, to a maximum of

40 km for earthquakes beneath Haleakala (Moore, 1964; HVO seismicity summaries). Since most earthquake depths are expected to cluster between 5 and 15 km, the trial focal depth was set at 5 km. The Vp/Vs ratio, which is used to calculate the S wave velocity, was initially assumed to be 1.78. The results of a quarry blast study (more fully discussed in a later section of this chapter) confirmed this was a reasonable choice. The arrivals used to calculate hypocenters are weighted according to user input and program statistics (Lee and Lahr, 1975). For this study the P wave was given full weight and the S wave half weight.

The Tectonic Setting and Geology of Maui

The island of Maui is part of a volcanically and seismically active island chain on the oceanic Pacific plate far from active plate boundaries. The tectonic activity is believed to be caused by the Pacific plate passing over the "fixed" Hawaiian hotspot now located near the southeastern portion of the island of Hawaii (Wilson, 1965a; Suyenaga, 1977). Maui was formed by two volcanoes: West Maui Volcano and Haleakala. The island is still seismically active (HVO seismicity summaries; Estill, 1979), with most seismicity concentrated beneath Haleakala. The Molokai Fracture Zone in the region of the Hawaiian chain is another area of earthquake activity and passes north and west of Maui. It is thought to be the source of the 1938 Maui earthquake (Furumoto, et. al., 1973).

The varied near-surface geology of Maui has a definite effect on the shallow layer average velocities. In order to insure that the model's first layer P wave velocity approximated that of the actual first layer, a detailed study was made of the velocity effects caused by the geology beneath each element of the seismic array. Maui consists of six eruptive series from two volcanoes (Table 1). The formations below the model datum (sea level) that affect the first layer velocity are the Wailuku and Honomanu Series. The Kula Series, on which three of the four seismometers rest, consists of homogeneous hawaiite flows and lies almost entirely above sea level. It's velocity effects are included in the station corrections. The Wailuku Series forms most of western Maui. Mainly composed of olivine basalt, it has some layers of tuff near the surface. The Honomanu Series forms the base of Haleakala. Its main rock type is also olivine basalt, but aa clinkers, breccias, tuffs, and weathered surfaces are interspersed throughout the formation (Stearns and McDonald, 1942).

The station corrections were based on elevation, crustal studies done in the Hawaiian Islands, and a vertical angle of emergence. Seismic refraction experiments on the islands of Hawaii and Oahu (Furumoto and Woollard, 1965; Furumoto, et. al., 1968; Hill, 1969; Furumoto et. al., 1970; Zucca and Hill, 1979;) and the inversion of Hawaiian Volcano Observatory earthquake travel time data (Crosson and Koyanagi, 1979) indicate there are three distinct near surface layers. The uppermost layer on land extends approximately 100 m below the surface and has an average velocity of 1.9 km/sec. Beneath lies a layer

Table 1

A List of Eruptive Series on Maui

	Series	Age* (m. y.)
	Hana	Recent
Haleakala	Kula	0.46
	Honomanu	0.84
	Lahaina	1.05
West Maui	Honolua	1.16
	Wailuku	1.29

* (McDougall, 1964)

100 to 200 m thick with velocities ranging from 2.7 to 3.3 km/sec. The layer below has velocities between 3.5 and 4.2 km/sec. It is assumed that Maui has nearly the same average surface structure as Hawaii and Oahu. The station corrections (Table 2) were constructed by multiplying a 100 m thick layer by 1.9 km/sec, a 150 m thick layer by 3.0 km/sec, and the remaining thickness by the model's first layer velocity. The first layer velocity was initially assumed to be 4.0 km/sec based on studies of the geology of Maui (discussed subsequently); however, it was decreased to 3.7 km/sec on the basis of geophysical studies near Maui (discussed subsequently). Local variations in the geology near each seismometer may alter each correction. The seismometer near Haiku is mounted at the base of a cinder cone. Loose and unconsolidated cinders do not propagate seismic waves as well as thick and continuous basalt flows, so there may be an additional delay.

HYPO71 assumes a layered and homogeneous crustal model. Velocity variations within a layer will increase or decrease the velocity of the dominant rock type. The combined effects yield an average velocity for the layer. The velocity effects of each formation below sea level must be assessed and weighted in order to obtain an average first layer P wave velocity. In general, Hawaiian olivine basalts have P wave velocities between 4.5 and 5.5 km/sec (Manghnani and Woollard, 1965). The presence of any inhomogeneities such as tuffs tends to reduce the velocity. The Wailuku Series contributes little to the overall first layer velocity, as it indirectly affects only those earthquakes that originate in West Maui or further west. The layers of tuff near the

Table 2

The Locations and Station Corrections
for the HIG Maui Seismic Network

Station	Location Lat N	Location Long W	Elevation (m)	Sta Corr (sec)
ARPA	20° 42.71′	156° 15.73′	2980	.85
Kula	20° 45.57′	156° 19.78′	880	.28
Haiku	20° 54.46′	156° 16.73′	270	.14
Wailuku	20° 52.63′	156° 31.07′	210	.08

The locations and elevations were taken form USGS 7 1/2′ quadrangle maps surface are too few and thin to be of consequence. Almost all of the first layer velocity variations originate in the Honomanu Series, because clinkers, breccias, and weathered surfaces combine to lower the olivine basalt velocity. Based on the geological data alone, an average velocity of 4.0 km/sec was assumed for the first layer of the crustal model.

Geophysical Studies on Maui

The results of a seismic refraction study done near Maui in 1962 were used to refine the first layer velocity and determine the remainder of the crustal model. As part of the Mohole site selection process Shor and Pollard (1964) performed a seismic refraction experiment in the Hawaiian Islands. Reversed refraction lines running northwest-southeast were shot near Maui (Fig. 4). Table 3 lists the results.

A comparison of both sites indicates that the uppermost layers of each is composed of what appears to be alternating layers of basalt and limestone and mud (Vp 2.69 km/sec) covering low velocity basalt (3.65 km/sec) (Shor and Pollard, 1964). The results of a seismic refraction and a well-logging experiment done in 1965 (on the Ewa Beach Plain, Oahu) are in agreement with the velocities and lithologies of the offshore crustal layers near Maui. In the Ewa Beach experiment the topmost layer consisted of alternating layers of basalt and limestone with an average velocity of 2.4 km/sec. The layer below consisted of alternating layers of basalt, tuff, and clinkers with an average

Figure 4 Two Sites from the Shor and Pollard (1962) Seismic Refraction Experiment. The Site North of West Maui is at 21° 17′ N and 156° 35′ W and the Site North of East Maui is at 20° 53′ N and 155° 56′. The Line Was Shot in Both Directions by Two Ships Alternately Shooting and Receiving.

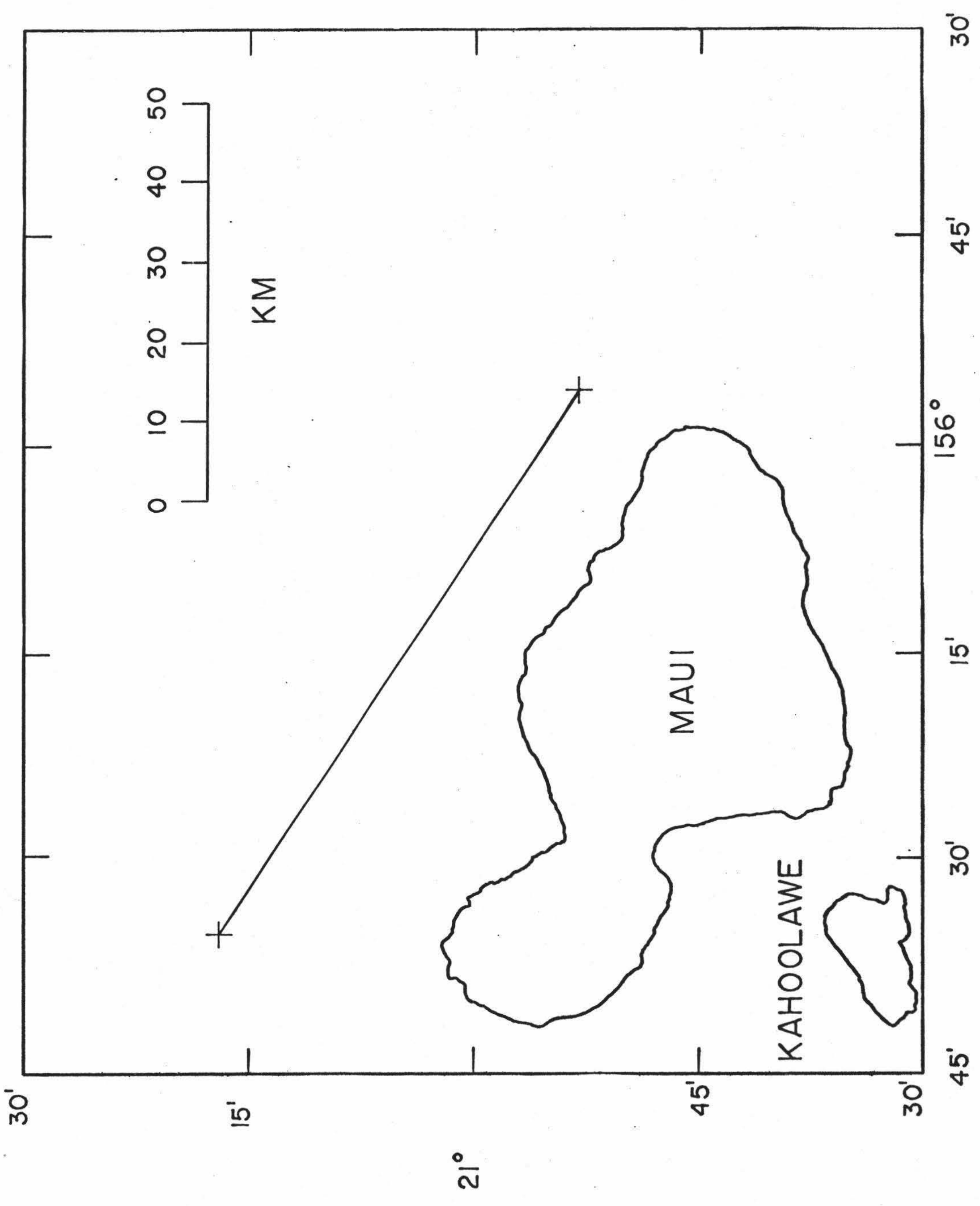

KM
0
10
20
30
40
50
MAUI
KAHOOLAWE
30'
15'
21°
45'
30'
45'
30'
15'
156°
45'
30'

Table 3

Two Velocity Depth Profiles North of Maui
(After Shor and Pollard, 1964)

West Maui Site			East Maui Site		
21° 17′ N 156° 35′ W			20° 53′ N 155° 56′ W		
Vp (km/sec)	Depth to Bottom of Layer (km)		Vp (km/sec)	Depth to Bottom of Layer (km)	
1.50	Water	.52	1.50	Water	1.20
2.68	a	1.35	2.69	a	2.22
3.65	b	2.55	3.65	b	2.31
4.96	c	7.05	4.96	c	10.1
			7.10	d	15.5
8.10	e		8.05	e	

velocity of 3.5 km/sec (Furumoto, et. al., 1970). This last layer of volcanics is lithologically similar to the Honomanu Series. Based on these results the first layer P wave velocity for the model was reduced to 3.7 km/sec. The 3.1 km/sec velocity used in the station correction was not lowered to 2.7 km/sec, because the layers above sea level do not contain appreciable amounts of coral and mud that would lower P wave velocity.

The remaining layers of the crustal model were determined by the Shor and Pollard results. At both sites the layer beneath the low velocity basalt appears to be an average velocity basalt (4.96 km/sec) that extends to a depth of 10 km. A problem exists for the next layer. At the refraction site northwest of Maui there were Moho velocities of 8 km/sec at a very shallow depth of 8 km. Shor and Pollard attributed this phenomenon to large scale faulting in the area. Other researchers (Malahoff and Woollard, 1968) suggested that the shallow mantle depth was the result of an intrusion of high velocity mantle material into the crust. The other site northeast of Maui offered a more plausible apparent velocity of 7.1 km/sec for the next layer, but the layer's velocity and upper limits were poorly defined. The bottom of the layer at 15.5 km was clearly defined by a Moho velocity of 8.05 km/sec. The depth to mantle of 15.5 km appears reasonable.

The velocity-depth profile obtained by Shor and Pollard in 1962 at the site northeast of Maui forms the basis for the crustal model used in HYPO71. The crustal model is layered, homogeneous, and extends infinitely in all directions.

Velocity (km/sec)	Depth to Top of Layer (km)
3.7	0.0
5.0	2.3
7.0	8.0
8.1	15.5

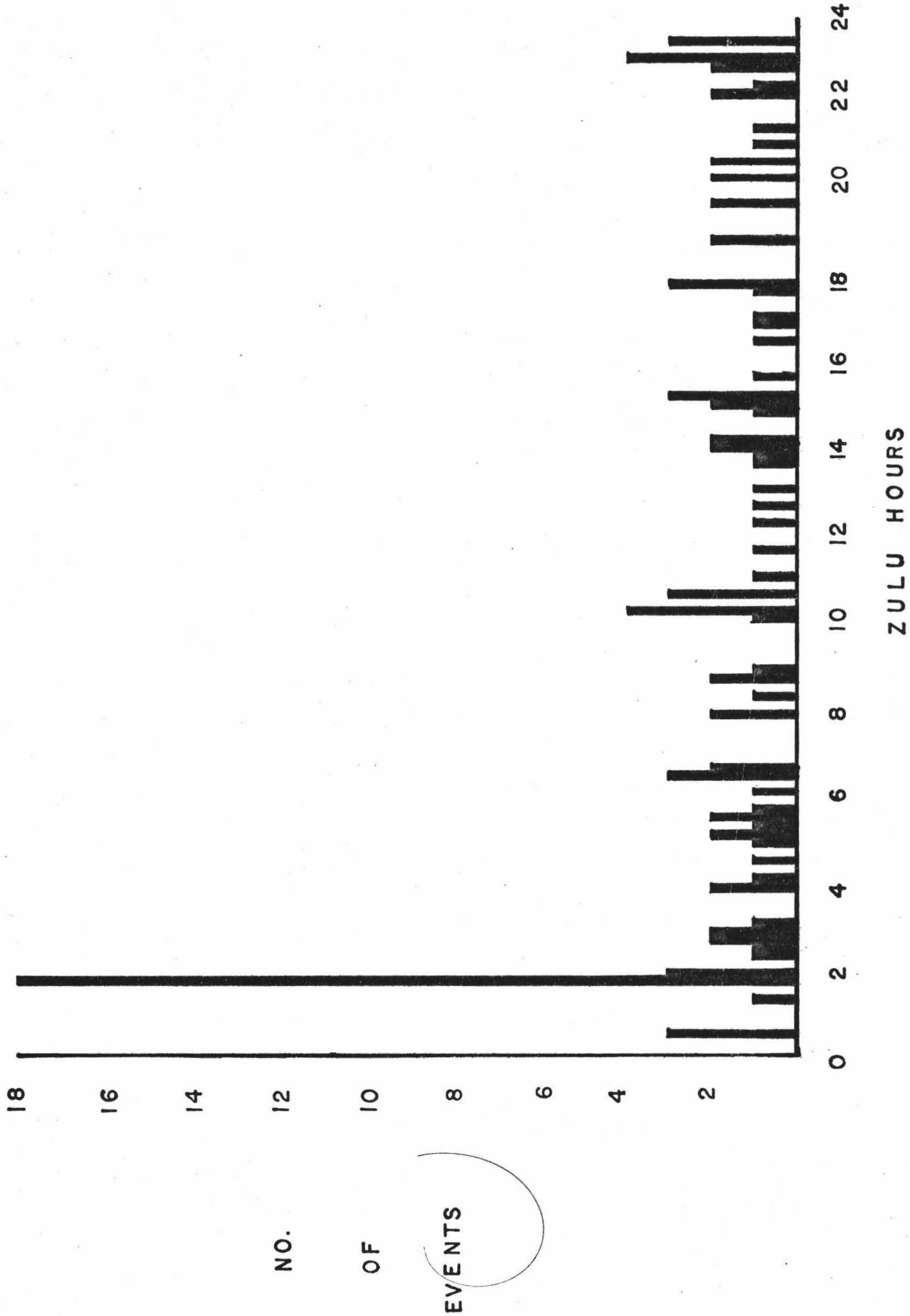
The profile was altered for the purposes of this study in the following manner. The layers of water and sediment were removed and in their place was substituted a layer of equal thickness and velocity 3.7 km/sec. Also, all velocities were rounded to the nearest tenth.

The geology and velocity structure of the Hawaiian Islands is varied and complex. Layer velocities tend to decrease with age along the Hawaiian Ridge (Furumoto, et. al., 1968; Hill, 1969; Furumoto, et. al., 1970). Different loads on the ridge depress the Moho discontinuity to different depths. Near Oahu the depth to Moho is 21 km, Maui 15.5 km, Kahoolawe 18 km, and the island of Hawaii 12 km (Furumoto and Woollard, 1965; Furumoto, et. al., 1968; Furumoto, et. al., 1970; Watts, 1978). The depression of the Moho is greatest under the islands. Away from the ridge the Moho depth is closer to the surface. These departures from homogeneity should not affect the locations of local earthquakes that occur within the crust. The crustal model may bias the hypocenters of near-surface events that originate farther from the array. because of the different surface layer velocities and Moho depths.

Quarry Blasts and the Crustal Model

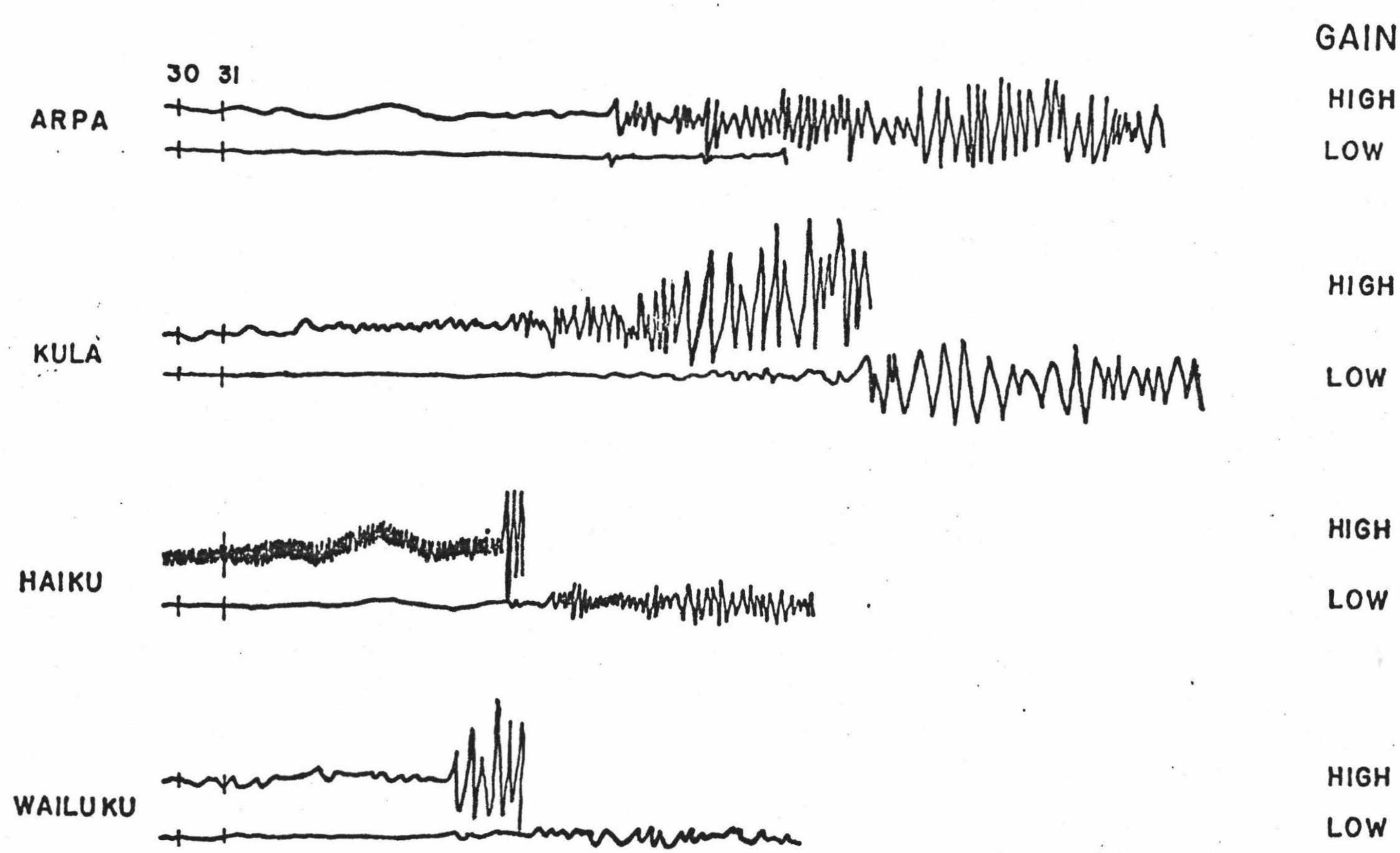
Quarry blasts occurred throughout the period June 1980 to August 1981 and were used to determine seismic velocity and verify station corrections. The nature of the events that occurred at or near 0200Z about every two to three weeks (Fig. 5) was not determined until the project had ended, thus no accurate origin time is available. The Hawai State Office of Occupational Safety and Health Administration in Honolulu supplied a list of possible companies, construction sites, and quarries that conducted blasts on Maui. A subsequent search lead to a construction company near Kahului that was responsible for the blasts. According to company officials, the near-surface blasts used nitrate fertilizer and Tovex packed into 50 to 100 holes, 2 to 3 inches in diameter, drilled 20 to 36 feet deep, and spread over an area of 5000 to 10000 square feet. An average blast contained 4000 to 5000 pounds of explosive and was detonated with 5 to 15 delays spaced milliseconds apart. Blasts from other quarries and construction sites occurred, but were not recorded, as either less explosives or more delays were used. A company supplied list showed that a total of thirty-two blasts occurred during the period of the seismic survey. Twenty-six of these provided good records.

Figure 5 Event Frequency vs. Time of Day (GMT). Hawaiian Standard Time = Greenwich Mean Time - 10 Hours. Note the Large Concentration of Events Near 0200Z.


NO. OF EVENTS
18
16
14
12
10
8
6
4
2
0
2
4
6
8
10
12
14
16
18
20
22
24
ZULU HOURS

Since quarry blast origin times were not accurately known, it was necessary to locate the blasts as events with known hypocenter and unknown origin time. When the location was fixed and the origin time was allowed to float, HYPO71 yielded hypocentral solutions with residuals of 1 second or greater. When allowed to freely iterate hypocenters and float origin times, HYPO71 located the blasts far from the actual blast site. The repicking of P and S arrivals showed the original picks appeared correct. To further study this problem, each blast record was traced (Fig. 6) and compared to others recorded by the same station. This procedure was done to see if arrivals were consistent for each blast. Inconsistency might be caused by noise or emergent arrivals from small blasts. Matching of the waveforms for each blast showed that some blasts appeared to arrive 0.5-1 second earlier than others. The P waves for the earlier arriving blasts were consistent within 0.2 seconds for each other. The blasts that apparently arrived earlier had the largest magnitudes (1.5), while the rest had smaller magnitudes (0.5-1.0). Also, the arrivals for the larger blasts were more impulsive, than the smaller ones which had emergent arrivals. Noise masked the first arrivals in some cases. Assuming the blast locations did not change appreciably and therefore the arrival times at each station would remain approximately the same for each blast, the arrivals for all records were normalized to the arrivals for the larger blasts. Analysis of the tracings, travel time curves, and S-P times agreed with a first layer P wave velocity of 3.7

Figure 6 Quarry Blast Traced From a Geotech Viewer Screen.


GAIN
30 31
ARPA
HIGH
LOW
KULA
HIGH
LOW
HAIKU
HIGH
LOW
WAILUKU
HIGH
LOW

km/sec and a Vp/Vs ratio near 1.78. The Vp/Vs ratio yields a Poisson's ratio of .27, which is in agreement with values for Hawaiian olivine basalts and hawaiites (Manghnani and Woollard, 1965).

The reworked data provided excellent blast epicenters from the four station array. When the location was not fixed and the origin time was allowed to float, the computer program iterated to a point within 0.4 km of the center of the quarry, with residuals generally less than 0.1 second (Fig. 7, Table 4). Company officials stated blasting occurs in a zone 1/4 mile from the quarry center, so the locations are close to the error limit. In contrast, the freely iterated epicenters yielded by the linear three station sub-array of ARPA, Kula, and Wailuku were displaced 4 km southwest of the quarry (Fig. 7, Table 4). This result indicates the geometry of this particular sub-array does not allow the proper location of the quarry blasts, which is not surprising since all three stations are in a line and the quarry is 10 km northeast of the line.

In summary, the crustal structure model for HYPO71 was determined mainly from the Shor and Pollard velocity depth profile. The water and sediment layers were removed and a layer with velocity 3.7 km/sec was substituted in their place. Geological and geophysical studies as well as quarry blasts helped to confirm the velocity of the upper layer and establish station corrections.

Figure 7 Maui Quarry Blast Epicenters. The Quarry is Marked by a Triangle. The Blasts Located by the Three Station Linear Sub-array (ARPA, Kula, and Wailuku) are Located Southwest of the Quarry.

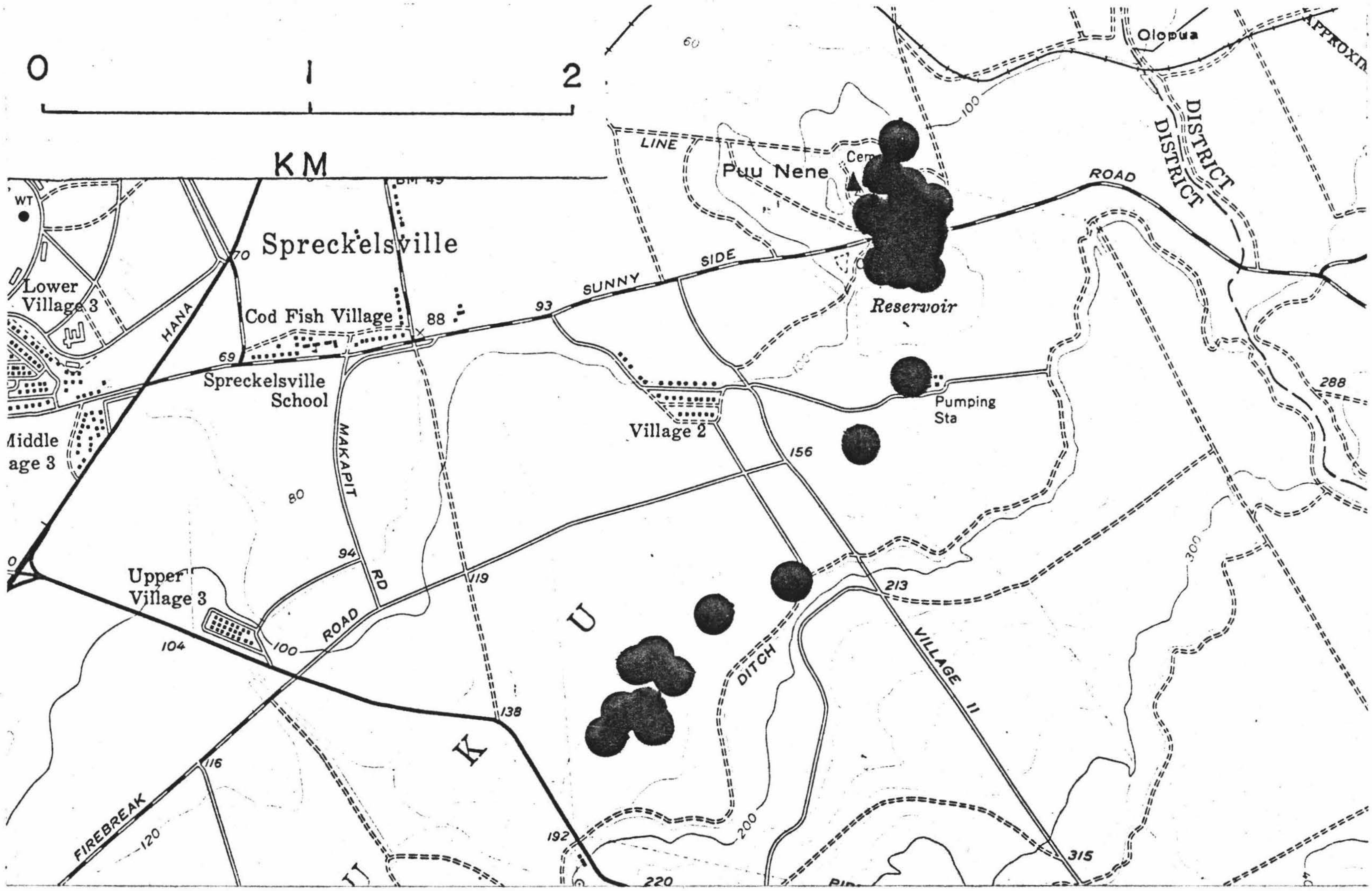

0
1
2
KM
WT
Spreckelsville
Lower Village 3
Cod Fish Village
Spreckelsville School
Upper Village 3
HANA
MAKAPIT RD
ROAD
SUNNY SIDE
ROAD
LINE
Puu Nene
Reservoir
Pumping Sta
Village 2
DITCH
VILLAGE 11
FIREBREAK
DISTRICT
DISTRICT
Olopua
U
K
69
70
88
93
94
100
104
116
119
120
138
156
192
200
213
220
288
300
315

Table 4

Quarry Blasts Located
by the HIG Maui Seismic Network:
June 1980 to July 1981

DATE	ORIGIN		LAT N	LONG W	DEPTH	MAG	GAP	DMIN	RMS	ERH	ERZ	R
800611	145	58.59	20-53.94	156-23.90	0.00	1.45	187	12.5	0.06	0.3	0.4	
800621	152	58.53	20-53.02	156-24.44	0.00	1.49	235	11.5	0.06	4.2	4.1	*
800626	143	27.43	20-53.02	156-24.44	0.00	1.32	235	11.5	0.06	4.2	4.1	*
800719	2300	4.38	20-54.08	156-23.91	0.00	1.39	189	12.5	0.02	0.1	0.1	
800730	142	31.39	20-53.94	156-23.90	0.00	1.47	187	12.5	0.06	0.3	0.4	
800807	143	49.99	20-53.94	156-23.90	0.00	0.90	187	12.5	0.06	0.3	0.4	
800809	2258	44.09	20-53.94	156-23.90	0.00	1.20	187	12.5	0.06	0.3	0.4	
800814	229	25.79	20-54.00	156-23.91	0.00	1.17	188	12.5	0.04	0.2	0.2	
800819	148	0.90	20-53.88	156-23.89	0.00	0.95	186	12.5	0.05	0.2	0.3	*
800821	144	11.08	20-53.25	156-24.37	0.00	1.19	238	11.7	0.04	3.1	3.1	*
800826	142	42.37	20-53.01	156-24.50	0.00	1.58	235	11.4	0.04	0.8	0.8	*
800829	143	9.78	20-53.25	156-24.37	0.00	1.15	238	11.7	0.04	3.1	3.1	*
800903	143	53.98	20-53.25	156-24.37	0.00	0.97	238	11.7	0.04	3.1	3.1	*
800913	144	0.83	20-53.02	156-24.44	0.00	1.32	235	11.5	0.06	4.2	4.1	*
801002	144	38.98	20-53.25	156-24.37	0.00	1.09	238	11.7	0.04	3.1	3.1	*
801128	2329	9.19	20-53.94	156-23.90	0.00	1.42	187	12.5	0.06	0.3	0.4	
801218	2259	32.09	20-53.94	156-23.90	0.00	0.98	187	12.5	0.06	0.3	0.4	
810109	2328	40.89	20-53.94	156-23.90	0.00	1.20	187	12.5	0.06	0.3	0.4	
810120	2328	31.09	20-54.00	156-23.91	0.00	1.10	188	12.5	0.04	0.2	0.2	
810131	028	46.03	20-53.02	156-24.44	0.00	1.32	235	11.5	0.06	4.2	4.1	
810223	2329	10.40	20-53.74	156-24.09	0.00	1.35	244	12.3	0.06	2.2	1.5	*
810310	030	16.11	20-53.26	156-24.28	0.00	1.60	238	11.8	0.05	3.6	3.6	*
810331	145	58.39	20-53.94	156-23.90	0.00	1.71	187	12.5	0.06	0.3	0.4	
810623	144	7.60	20-54.01	156-23.92	0.00	0.91	188	12.5	0.09	0.4	0.6	
810717	143	12.69	20-53.92	156-23.89	0.00	0.89	186	12.4	0.03	0.2	0.2	
810724	145	55.99	20-53.94	156-23.90	0.00	0.82	187	12.5	0.06	0.3	0.4	

* Quarry blasts located by linear sub-array: ARPA, Kula and Wailuku

MAG is the duration magnitude calculated using a bilinear relation derived by the Hawaiian Volcano Observatory.

GAP is the largest azimuthal separation in degrees between stations.

DMIN is the epicentral distance in kilometers to the nearest station.

RMS is the root mean square error in seconds of the arrival residuals.

Table 4 (continued)

Quarry Blasts Located
by the HIG Maui Seismic Network:
June 1980 to July 1981

ERH is the standard error of the epicenter in kilometers.

ERZ is the standard error of the focal depth in kilometers.

R is a remark.

CHAPTER III

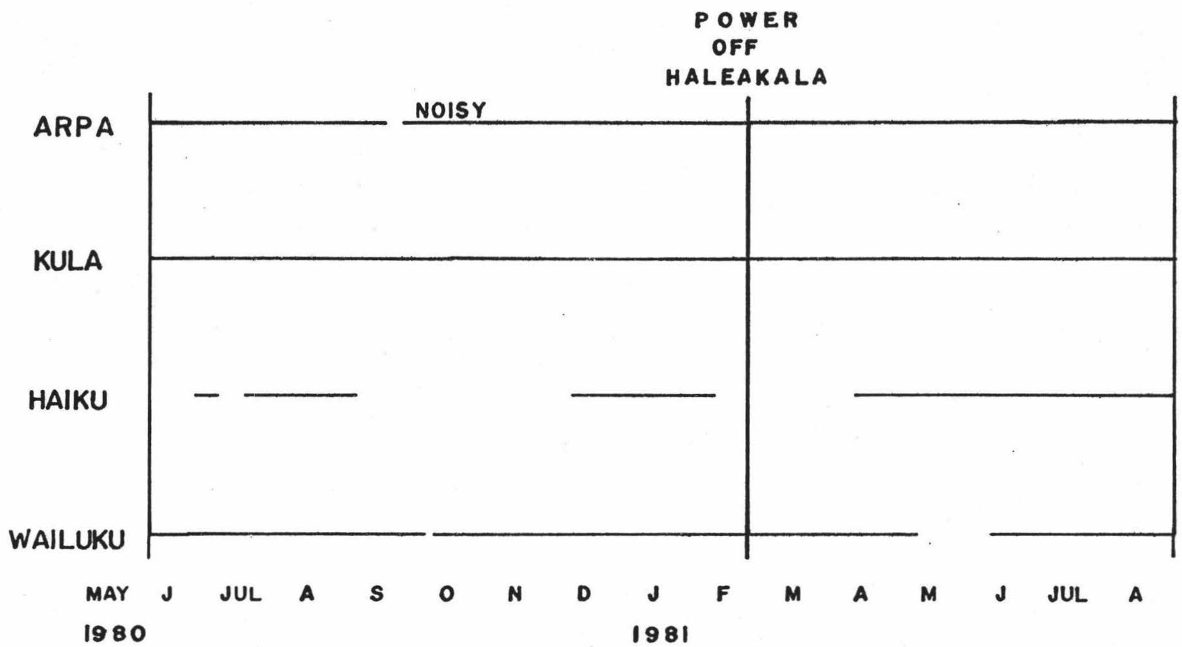
CURRENT MAUI SEISMICITY NEAR MAUI, MOLOKAI, AND LANAI

Data Acquistion and Reduction

In 1978 the Hawaii Institute of Geophysics installed a four station seismic array on Maui that was fully operational in 1980. The array's primary purpose was to monitor seismicity and crustal deformation in the Maui region for NASA's Lunar Ranging Experiment (LURE). The array consisted of three vertical seismometer stations and one tiltmeter station, which acted as a horizontal seismometer. The instruments were located near Wailuku, Haiku, Kula, and the summit of Haleakala. The tiltmeter at Kula also measured the local tilt of Haleakala. Seismic signals were telemetered from Maui to Oahu and recorded on film by a Develocorder at the Hawaii Institute of Geophysics. Seismic data were recorded between mid-May 1980 and mid-August 1981. Figure 8 shows the operational dates for each seismometer.

Earthquakes with S-P times of 15 seconds or less (indicating an epicentral distance of 120 km or less) were selected for this study from drum recorder seismograms. HVO seismicity summaries were consulted to eliminate any earthquake located on the island of Hawaii. One hundred twenty-four events with three or more stations reporting were chosen for study. Arrivals and durations were picked from a Geotech Viewer. The duration is defined as the time interval between the onset of the P wave and the point where the signal decays to

Figure 8 Dates of Operation for Seismometers in the HIG Maui Seismic Network. The Solid Line Indicates When a Seismometer Was Functioning.



106 EVENTS LOCATED

BY THE MAUI NETWORK

NO. OF EVENTS	ARPA	KULA	HAIKU	WAILUKU
SEISMOMETER WAS OPERATIONAL	102	106	54	100

the noise level before the earthquake. The duration magnitude was calculated using a bilinear least squares relation derived by HVO (Tanigawa, et. al., 1980).

$$M = -5.2 + 3.891\log(D) + 0.013(Z) + 0.0037(\Delta) \quad D < 210 \text{ sec.} \qquad (1)$$

$$M = -0.905 + 2.026\log(D) + 0.013(Z) + 0.0037(\Delta) \quad D > 210 \text{ sec.} \qquad (2)$$

where

D = Duration

Z = Focal depth

Δ = Epicentral distance

The first motion was also noted but not used as there were not enough stations for a focal mechanism study. As stated earlier, all events were located using HYPO71.

Earthquake Location Study

The four station Maui seismic network is essentially a triangular array, with two stations near one corner (Fig. 1). Many researchers (Francis, et. al., 1977; Lilwall and Francis, 1978; Urhammer, 1980; Bosher, 1981) have done studies on the the efficiency and errors associated with this type of network. Epicentral error is least inside the array and increases away from the array. The depth error is a minimum at the outer edges of the array. Inside the array there is poor depth control. At a distance twice the array radius from the center, depth control rapidly diminishes (Lilwall and Francis, 1978; Urhammer, 1980). The epicenter is least constrained along the line connecting the greatest concentration of seismic instruments. For the Maui network this is a line running northwest-southeast through ARPA, Kula, and Wailuku stations.

A check of Hawaiian Volcano Observatory monthly seismicity summaries shows there were 12 earthquakes located by HVO that were also located by the four station Maui seismic network. All earthquakes in common were outside the Maui network and within or near the extensive HVO network. The epicenters located by the Maui array are generally displaced 5 to 40 km from the HVO locations. The calculated depths are 3 to 30 km different from the HVO estimates. These differences were expected since all the earthquakes in common were outside of the Maui network where there is little accuracy. The Maui network duration magnitudes were 0.1 to 0.8 units less than the HVO magnitudes. The error is directly attributable to differences in calculated focal depths (see equations 1 & 2 above). When the same depths are assigned to the same earthquakes, the Maui and HVO magnitudes agreed within .1 to .2 units. This comparison shows that there is internal consistency in the estimated epicenters and magnitudes between the Maui network and HVO.

For parts of the monitoring period, Haiku was not operational leaving a three station linear sub-array composed of ARPA, Kula, and Wailuku. The three station linear sub-array provided problems in hypocentral location. More than 75% of the initial epicenter iterations failed because the location errors exceeded the limit set by the computer program. In order to surmount this problem it was necessary to provide an alternate hypocenter near the estimated epicenter. The program computed a solution with a small RMS residual; however, the standard errors were usually large (> 50 km) and in many cases the trial depth of 5 km did not change.

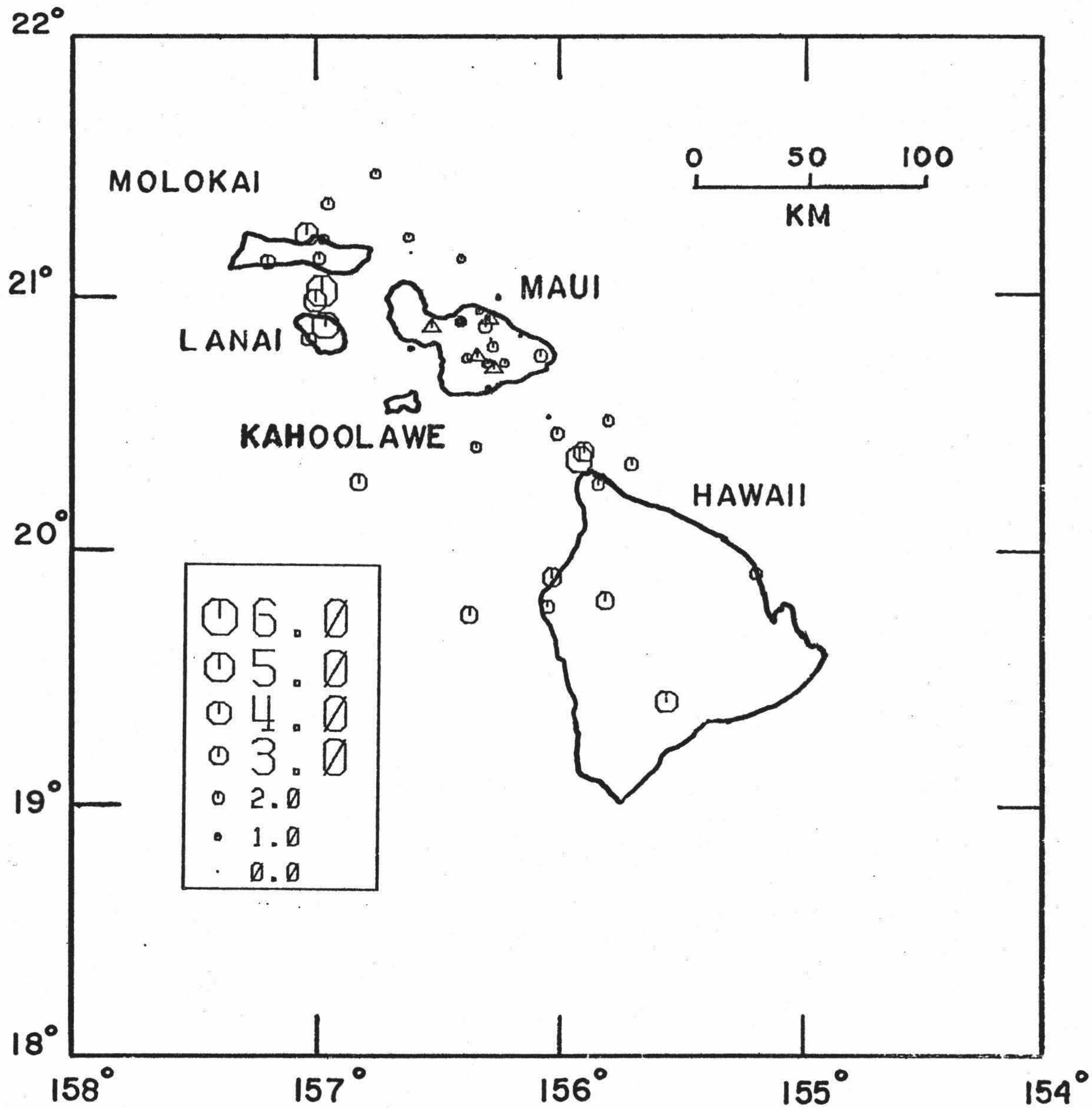
A study was made of location problems with the linear sub-array. Trial epicenters were assigned to each earthquake and quarry blast located by the linear sub-array. The hypocenter was allowed to float in order to obtain an estimate of hypocentral error. For initial depths of 1, 5, 10, 15, and 25 km, over 40% of the final hypocenters retained the initial depth. Epicenters for the deeper focal depths are closer to the network than those with shallow depths. Earthquake locations from the four station array were compared to those from the four station array without Haiku. Epicenters from the linear sub-array were displaced 4 to 30 km northeast or southwest of the four station location. Sometimes the relocated epicenters were on the same side of the linear sub-array and sometimes they were displaced on the other side. The focal depth differences between the two varied from 4 to 20 km. This study shows that there are many possible estimated solutions for a given earthquake located by the linear sub-array of ARPA, Kula, and Wailuku. Based on the location studies none of the earthquake or quarry blast locations computed with the linear sub-array of ARPA, Kula, and Wailuku can be accorded any weight in this paper.

In general, the earthquakes and quarry blasts located by the other array combinations provided results similar to the four station network. Using earthquakes and quarry blasts from the four station array epicenters for the sub-arrays ARPA, Haiku, and Wailuku (delete Kula) and Kula, Haiku, and Wailuku (delete ARPA) were within 1 to 10 km of the four station location. The residuals and standard errors for the sub-arrays were slightly greater than the four station residuals and errors. The exception was when Wailuku was deleted and the sub-array ARPA, Kula, and Haiku remained. The epicenters varied by 5 to 20 km and the focal depths by 5-10 km. The residuals and standard errors were also larger. The reason for this exception is that Kula and ARPA are relatively close (8 km) relative to Haiku (20 km). For earthquakes outside of the Maui seismic network, this sub-array behaves almost like a two station array.

Hypocenters from the Maui Network

Twenty-six of the 124 selected events were identified as quarry blasts. Eighty of the remaining ninety-eight suspected earthquakes were located with RMS residuals less than or equal to a cutoff value of 0.50 seconds. Residuals larger than the cutoff generally indicate a poor hypocenter. The duration magnitudes of these earthquakes ranged from a minimum of -1.15 to a maximum of 5.0. Thirty-eight of the remaining earthquakes were eliminated because Haiku was missing from the data. The remaining earthquakes and quarry blasts are illustrated in Figure 9.

Figure 9 Earthquakes and Quarry Blasts Detected by the HIG Maui Seismic Network: June 1980 until July 1981, RMS < 0.50 sec. Note, All Earthquakes and Blasts Located by the Three Station Linear Sub-array (ARPA, Kula, and Wailuku) Have Been Deleted.


22°
21°
20°
19°
18°
158°
157°
156°
155°
154°
0 50 100
KM
MOLOKAI
MAUI
LANAI
KAHOOLAWE
HAWAII
6.0
5.0
4.0
3.0
2.0
1.0
0.0

The remaining earthquakes are listed in Table 5. Seven of these earthquakes were located beyond the area of this study in the central and southern portion of the island of Hawaii and thus are not germane. Ten solutions have epicentral or depth standard errors greater than twenty kilometers so their hypocenters were discarded. These final deletions left a total of twenty-five reliably located earthquakes in the Maui, Molokai, Lanai region with low residuals and standard errors (Fig. 10).

Two zones of seismicity were detected by the Maui seismic array: beneath Haleakala and between Molokai and Lanai. There is a diffuse area of small earthquakes (magnitude 2 or less) 20 to 40 kilometers below Haleakala. During the period of observation there were no swarms or harmonic tremor that would indicate magma movement. The second zone of seismicity was located between Molokai and Lanai. The three large earthquakes in this area are related to the March 5, 1981 manitude 5.0 Molokai earthquake. Lesser levels of activity were detected in other areas. One earthquake was located near West Maui Volcano. Three earthquakes were located between Maui and the Big Island at shallow to medium depths (5-15 km). One earthquake had a magnitude of 3.2, while the other two were less than 2.0. Four earthquakes were located north of Maui and Molokai. They occurred at shallow to medium depths (5-15 km) and all four had magnitudes less than 2.0.

Table 5

Earthquakes Detected
by the HIG Maui Seismic Network
RMS < 0.50 sec:
June 1980 to August 1981

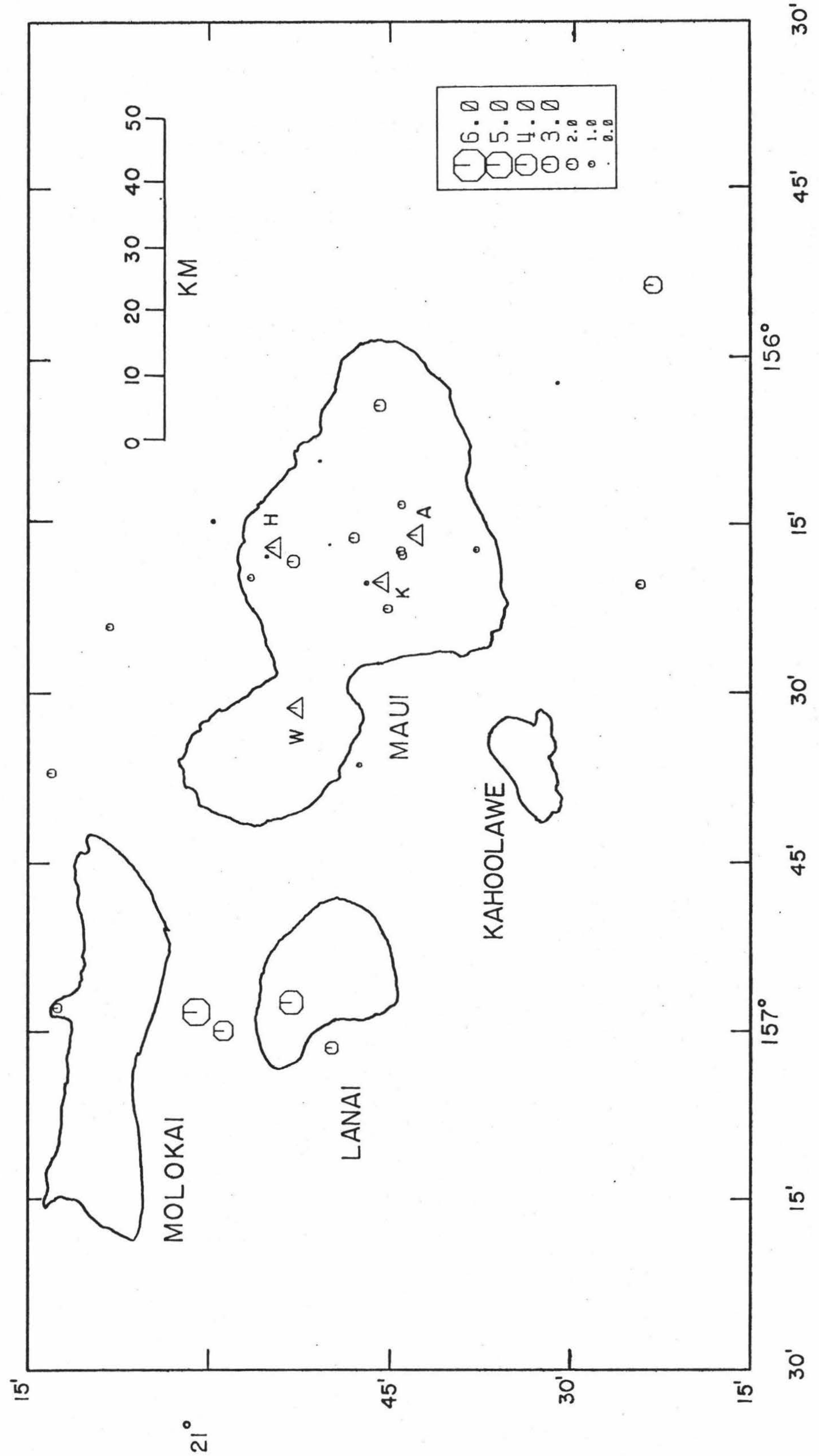
DATE	ORIGIN	LAT N	LONG W	DEPTH	MAG	GAP	DMIN	RMS	ERH	ERZ	R
800609	23 4 16.47	19-44.67	156-21.68	5.00	2.69	347	107.6	0.19	127.3	99.4	#
800616	646 13.44	19-46.54	156- 2.38	5.00	1.95	349	106.2	0.15	243.8	190.6	#
800702	1945 2.81	21-13.02	156-36.84	13.50	1.52	329	38.9	0.12	17.3	18.9	
800703	837 58.25	20-44.06	156-17.09	27.56	1.52	164	3.4	0.33	5.2	3.8	
800703	852 53.36	20-43.94	156-17.41	28.77	1.45	176	3.7	0.30	4.8	3.4	
800704	21 1 23.28	21- 8.20	156-24.99	1.02	1.23	310	29.1	0.21	2.9	3.2	
800710	1349 32.65	20-30.08	155-47.42	5.00	1.72	344	54.4	0.36	29.3	10.6	&
800715	1014 35.49	20-47.53	156-36.15	29.16	0.78	299	12.9	0.20	3.9	2.9	
800716	1259 44.69	20-45.16	156-22.22	16.17	1.50	200	4.3	0.27	3.3	3.5	
800812	642 50.66	20-21.35	155-54.58	7.78	4.26	345	53.9	0.09	22.4	1.3	&
801208	1018 17.21	20-46.91	156-19.88	27.37	0.69	123	2.5	0.33	4.4	3.9	
801212	4 2 4.44	19-47.85	155-48.13	2.50	2.91	350	112.0	0.39	352.3	275.5	#
801213	1220 13.93	20-49.96	156-16.51	7.62	-0.07	177	8.3	0.35	15.1	11.1	
801216	2046 9.42	20-56.45	156-19.36	35.27	1.06	238	5.9	0.03	0.9	0.6	
801219	17 9 36.08	20-16.07	156-49.01	13.51	2.70	335	74.3	0.29	72.0	121.7	&
801220	1039 20.77	20-59.59	156-14.50	33.53	0.58	314	10.2	0.06	1.6	1.0	
801228	2133 6.19	20-31.05	156- 2.18	12.53	0.37	339	31.9	0.17	4.2	1.4	
801229	154 43.43	20-24.08	156-20.03	14.92	1.54	328	35.2	0.13	16.5	17.5	
810103	13 4 40.62	20-23.04	155-53.55	5.00	3.24	345	53.0	0.14	5.9	3.1	
810107	528 17.00	20-37.79	156-16.97	19.19	0.91	305	9.3	0.25	4.2	2.9	
810114	629 8.41	21- 7.40	157-11.49	2.50	2.22	349	75.2	0.37	577.4	451.5	&
810114	1418 57.63	20-20.17	155-41.62	2.50	2.05	349	72.4	0.14	999.9	999.9	&
810116	037 13.38	19-24.27	155-32.94	5.00	3.67	353	162.8	0.27	846.0	663.2	#
810122	1339 3.44	19-53.28	156- 1.26	9.73	3.12	348	94.6	0.21	142.1	243.3	#
810305	14 9 43.86	21- 0.95	156-58.16	17.22	5.00	176	49.4	0.45	11.1	14.1	
810305	1416 16.92	20-58.71	156-59.80	7.80	3.60	168	51.1	0.41	12.0	3.3	
810306	243 37.79	20-53.09	156-57.31	5.00	4.34	318	45.5	0.41	19.3	14.3	
810328	1947 43.66	20-49.77	157- 1.32	2.50	2.24	344	52.7	0.36	13.9	9.1	
810331	16 6 37.51	20-47.94	156-15.94	29.39	1.71	184	8.0	0.29	4.4	3.4	
810401	10 0 38.06	21-13.80	157- 1.95	5.00	3.83	348	66.2	0.34	999.9	999.9	&
810405	611 58.77	21-12.52	156-57.75	5.00	1.60	347	59.0	0.23	15.6	10.6	
810406	1853 37.31	21- 8.21	156-58.77	13.56	1.89	347	55.9	0.21	795.9	999.9	&
810427	1044 53.13	20-45.82	156- 4.13	8.51	2.09	308	20.9	0.22	1.9	48.8	
810505	15 1 4.75	21-21.05	156-56.71	7.73	1.89	350	84.9	0.24	546.4	427.1	&
810526	2221 13.41	20-27.15	156- 0.00	2.50	1.97	345	39.6	0.15	91.6	92.8	&
810613	1456 44.40	20-55.13	156-17.51	36.62	0.46	233	1.8	0.10	2.5	1.6	
810616	1756 24.80	21-28.34	156-44.92	13.02	1.49	342	70.1	0.25	110.0	186.9	&
810620	23 4 54.38	20-50.76	156- 9.22	17.75	0.35	280	14.7	0.19	4.8	6.8	
810706	1750 32.76	20-52.96	156-17.94	35.05	2.18	131	3.5	0.29	4.8	3.5	

Table 5 (continued)
Earthquakes Detected
by the HIG Maui Seismic Network
RMS < 0.50 sec:
June 1980 to August 1981

DATE	ORIGIN	LAT N	LONG W	DEPTH	MAG	GAP	DMIN	RMS	ERH	ERZ	R
810723	1139 6.72	20-44.01	156-13.13	6.10	1.36	260	5.1	0.37	2.7	1.4	
810802	1848 16.36	20-15.66	155-49.92	4.50	1.90	347	67.1	0.31	259.8	202.9	&
810806	23 0 18.66	19-54.15	155-10.68	5.00	1.85	354	144.4	0.20	999.9	999.9	#

\# Earthquakes located outside of the Maui-Molokai-Lanai region
& Earthquakes with epicentral or depth errors > 20 km

Figure 10 Reliably Located Earthquakes from the HIG Maui Seismic Network: June 1980 until August 1981. RMS < 0.50 sec, Epicentral and Depth Errors Less than 20 km. The Seismometer Locations are Indicated by Open Triangles.



KM
0 10 20 30 40 50
6.0
5.0
4.0
3.0
2.0
1.0
0.0
MOLOKAI
LANAI
MAUI
KAHOOLAWE
W
H
K
A
30'
15'
157°
45'
30'
15'
156°
45'
30'
15'
21°
45'
30'
15'

The largest earthquake recorded during the current monitoring period and the largest to occur near Maui in the last 24 years (Furumoto, et. al.,1973) was the manitude 5.0 March 5, 1981 Molokai earthquake. The U.S.G.S. (Preliminary Determination of Epicenters, 1981) located the earthquake at 20.994°N. and 156.995° W., at a depth of 15 km. Later, the U.S.G.S. (Earthquake Data Report, 1982) relocated the earthquake at 21.017° N. and 156.988° W. at a depth of 10 km. The Maui network linear sub-array was operating at the time so arrival times from Oahu (Pacific Tsunami Warning Center: Kipapa) and the Big Island (HVO: Uwekahuna and Ahua) were used to locate the mainshock and two large aftershocks. The epicenter obtained for the mainshock (21° 0.95′ N. and 156° 58.16′ W.) was similar to that obtained by the U.S.G.S. Furumoto (personal communication) has located the earthquake north of Molokai near 21° 20′ N. and 157° 00′ W. Fifteen aftershocks were detected until April 6, 1981, when activity apparently ceased. The largest aftershock (20° 53.09′ and 156° 57.31′ W., 5 km depth) occurred at 02:43 March 6, 1981 and had a magnitude of 4.3. Only two aftershocks could be reliably located. Consequently, the fault area affected by the earthquake could not be determined.

Possible Earthquake Mechanisms in the Maui-Molokai-Lanai Region

Plate tectonics theory has been developed over the last twenty years and it now provides a unifying framework for the earth′s geological processes (Vine and Matthews, 1963; Wilson, 1965c; Sykes, 1967; Isacks, et. al., 1968; Morgan, 1968b). One of the findings of plate tectonics is that, in general, seismicity is concentrated at plate boundaries, such as spreading ridges, transform faults, and subduction zones (Wilson, 1965c; Sykes, 1967; Isacks, et. al., 1968). As geologists and geophysicists better understand the causes of seismicity near plate margins, they are focusing their attention on intraplate earthquakes that occur far from plate boundaries (Sykes and Sbar, 1973; Sykes, 1978; Richardson, et. al., 1979; Okal, et. al., 1980).

Several explanations have been advanced to account for the causes of earthquakes away from plate margins. Some possible earthquake mechanisms that may affect Maui, Molokai, and Lanai are the following:

1). Mid-plate volcanism (Carter, 1978; Estill, 1979; Okal, et. al., 1980).

2). Reactivation of pre-existing zones of weakness (Sykes, 1978).

3.) Lithospheric loading (Unger and Ward, 1979; Endo and Rogers, 1977).

4). Lithospheric stress transmission (Carter, 1978, Okal, et. al., 1980).

5) Plate deformation due to latitudinal changes on an ellipsoidal earth.

All of these mechanisms or any combination of them may cause earthquakes. Each one of these mechanisms is evaluated as a possible cause of earthquakes near Maui, Molokai, and Lanai.

In 1974, Turcotte advanced a specialized theory of plate tectonics. The theory, called membrane tectonics, states that lithospheric plates must deform as they move latitudinally across the earth's elliptical surface. A plate moving away from the equator, such as the Pacific plate near Hawaii, will be subjected to compressional stresses in the plate interior and tensional forces along the plate edge. Significant stresses on the order of kilobars can build so as to fracture the lithosphere (Turcotte, 1974). Near Hawaii, membrane stresses are much smaller and on the order of 100 to 300 bars (Turcotte, 1974). The stresses associated with membrane tectonics may be a minor cause of earthquakes in the Maui-Molokai-Lanai region.

Lithospheric stress transmission may also be a significant cause of earthquakes. Carter (1978), using b values from circum-Pacific plate earthquakes, showed that stress from large earthquakes may be transmitted across the Pacific Plate. Okal and others (1980) believe that regional stress fields control many of the earthquakes that occur south of Tahiti.

> The correlation between the azimuths of compressional axes and the directions of plate motion has been noted elsewhere and attributed to the significance of "ridge-push" in the force balance equation of the plates...Our data are consistent with this inference...

A regional stress field may be responsible for some earthquakes in the Maui-Molokai-Lanai region.

Large volcanic loads can lead to lithospheric bending and fracture. Bending stresses from loads of 1 kbar are near the breaking point of rock (Walcott, 1970). For the Hawaiian Islands the maximum bending stresses are close to 1.3 kbar (Watts, et. al., 1980). Unger and Ward (1979), using a focal mechanism solution, have interpreted the 46 km deep, magnitude 6.2 Honomu earthquake on April 26, 1973 was the result of lithospheric bending stresses. Rogers and Endo (1977) have determined that the greatest compressive stress axes for many mantle earthquakes near the island of Hawaii are radially distributed. These studies suggest that lithospheric loading plays an important role in upper mantle Hawaiian earthquakes.

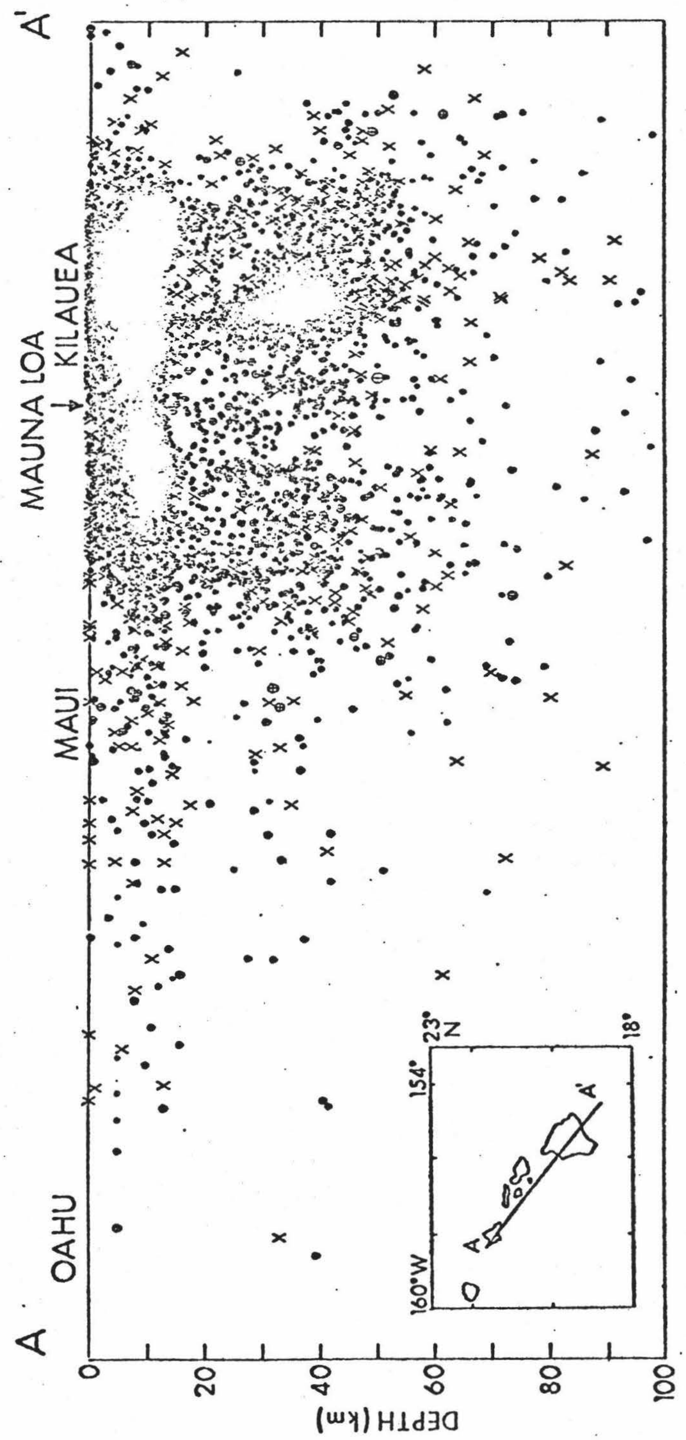
Sykes (1978) made a study of intraplate earthquakes in continental areas. He concluded that continental earthquakes tend to be concentrated along pre-existing zones of weakness, such as fault zones, fracture zones, suture zones and failed rifts. These zones of weakness were reactivated during the latest episode of continental rifting. Earthquakes in the Gulf of Guinea near Accra may be associated with the eastern edge of the Romanche Fracture Zone. Seismicity from South Carolina to New Madrid, Missouri in the United States may be related to a westward extension of the Blake Fracture Zone. Although it is not in a continental area, Hawaii may have intraplate earthquakes occurring along a reactivated pre-existing zone of weakness: the Molokai Fracture Zone. A possible scenario for reactivation might be similar to the following: Three million years ago, as the Pacific plate was moving northwest, the stable Molokai Fracture Zone was loaded with volcanics from the Hawaiian "hotspot". The new load reactivated a pre-existing zone of weakness. The Molokai Fracture Zone readjusted to the new state of stress by means of earthquake activity. As the Molokai Fracture Zone moved away from the "hotspot", earthquake activity decreased and the fracture zone adjusted to its "loaded" condition.

Mid-plate volcanism is responsible for many of the earthquakes detected on the Pacific plate (Estill, 1979, Okal, et. al., 1980). In volcanically active areas, magma movement is the main cause. Manifestations of magma movement are earthquake swarms and harmonic tremor (Bosher, 1981). The filling or draining of magma chambers can also trigger events (Estill, 1979). Minor shocks can also come from the cooling of magma deposits or chambers (Chouet, 1979). All of these mechanisms might be occurring presently in the Maui-Molokai-Lanai region.

It is difficult to estimate the effects of membrane tectonics and lithospheric stress transmission near the island of Hawaii. Stress associated with volcanism at Kilauea and Mauna Loa dominates all others that may be present (Carter, 1978). Either of these mechanisms could be a source of earthquake activity near the less volcanically active Maui-Molokai-Lanai region.

Recent findings have raised serious doubts as to the seismic activity of the Molokai Fracture Zone. Francis and others (1978) showed that numerous earthquakes occur on a ridge and transform fault near the eastern end of St. Paul's Fracture Zone; the fracture zone itself is inactive. Okal and others (1980) located only one out of one hundred earthquakes on or near a fracture zone southeast of Tahiti. In both of these studies fracture zone seismic activity was practically nil. Figure 2 from Estill (1979) shows the historical seismicity for the Maui-Molokai-Lanai region. If the Molokai Fracture Zone were active, one would expect clear linear trends north of Maui and Molokai, where the fault area is thought to be located; there are many possible trends. Figure 11, also from Estill (1979) shows a cross section of Hawaiian seismicity along a line connecting Kilauea and Oahu. The greatest concentration of earthquakes is below Kilauea and Mauna Loa. Seismicity decreases with distance from the Big Island, until there is next to none near Oahu. If there were any activity associated with the western portion of the Molokai Fracture Zone, one would expect a large concentration of events north of Maui; there is no such concentration. Perhaps the most conclusive evidence of the seismic inactivity of fracture zones is offered by Sandwell and Schubert (1982) who conclude that fracture zones do not slip over periods of 10^7 years. Scarp heights in the fracture zones are equal to the initial offsets at the ridge-transform fault intersection. Bending stresses are approximately 1 kbar and the maximum shear stresses are about 300 bars. These shear stresses are more than twice as large as the stress drops of large

Figure 11 Cross-section of Earthquake Hypocenters from Oahu to Southeastern Hawaii. Note, There is No Concentration of Events North of Maui, Where the Molokai Fracture Zone is Supposedly Located (After Estill, 1979).


A
A'
OAHU
MAUI
MAUNA LOA
KILAUEA
DEPTH (km)
0
20
40
60
80
100
160°W
154°
23°
N
18°
A
A'

earthquakes (Brune, 1968). If fracture zones were areas of weakness, they would slip in order to relieve these large stresses. Instead, fracture zones are welded together and they represent the remnant surface expression of the transform fault and spreading ridge. According to Sandwell and Schubert, the Mendocino Fracture Zone, between the Hawaiian-Emperor Seamount Chain and the Mapmaker Seamounts, has not slipped significantly in over 130 million years. The implication is that when this area was being loaded by the hotspot 40 million years ago, it did not become seismically active and slip. One is lead to the conclusion that the Molokai Fracture Zone is seismically inactive.

The dominant causes of earthquakes near Maui, Molokai, and Lanai appear to be directly or indirectly related to Hawaiian "hotspot" volcanism. Earthquakes may be caused by thermal stresses, loading of the crust with volcanic products, or the slumping of large, unstable blocks. Thermal stresses could include compressional ones caused by crustal uplift or tensional ones associated with the cooling of a magma chamber. Although Haleakala has not erupted in nearly 200 years, it possible that there is magma movement under the volcano. Earthquakes with calculated depths below the Moho may be crustal adjustments caused by lithospheric loading. Another possible source is submarine slumping. Offshore areas near Maui, Molokai and Lanai are the most likely locations for these types of earthquakes.

CHAPTER IV

CRUSTAL DEFORMATION ON MAUI, MOLOKAI, AND LANAI

A Brief History of the Lunar Ranging Experiment and the Maui Crustal Deformation Project 1976-1981

The secular change of a point in a geocentric coordinate system can be used to measure plate movement and crustal deformation. One of the most accurate techniques to determine the location of a point on the Earth's surface involves laser ranging from "fixed" observatories to retroreflectors left on the moon by the Apollo astronauts. Plate motion can be determined by calculating the secular change of geocentric position of two or more observatories and subtracting any change in observatory location caused by local or regional deformation (Berg, et. al., 1978). Since 1965, the National Aeronautic and Space Administration has operated the Lunar Ranging Experiment (LURE) with observatories in Texas and Hawaii (Bender, et. al., 1973). One of the purposes has been to measure relative movement of the Pacific and North American plates (Bender and Silverberg, 1975). The Hawaii Institute of Geophysics and the Institute for Astronomy were supported by NASA for the portion of the program undertaken in Hawaii (Berg, et. al., 1978).

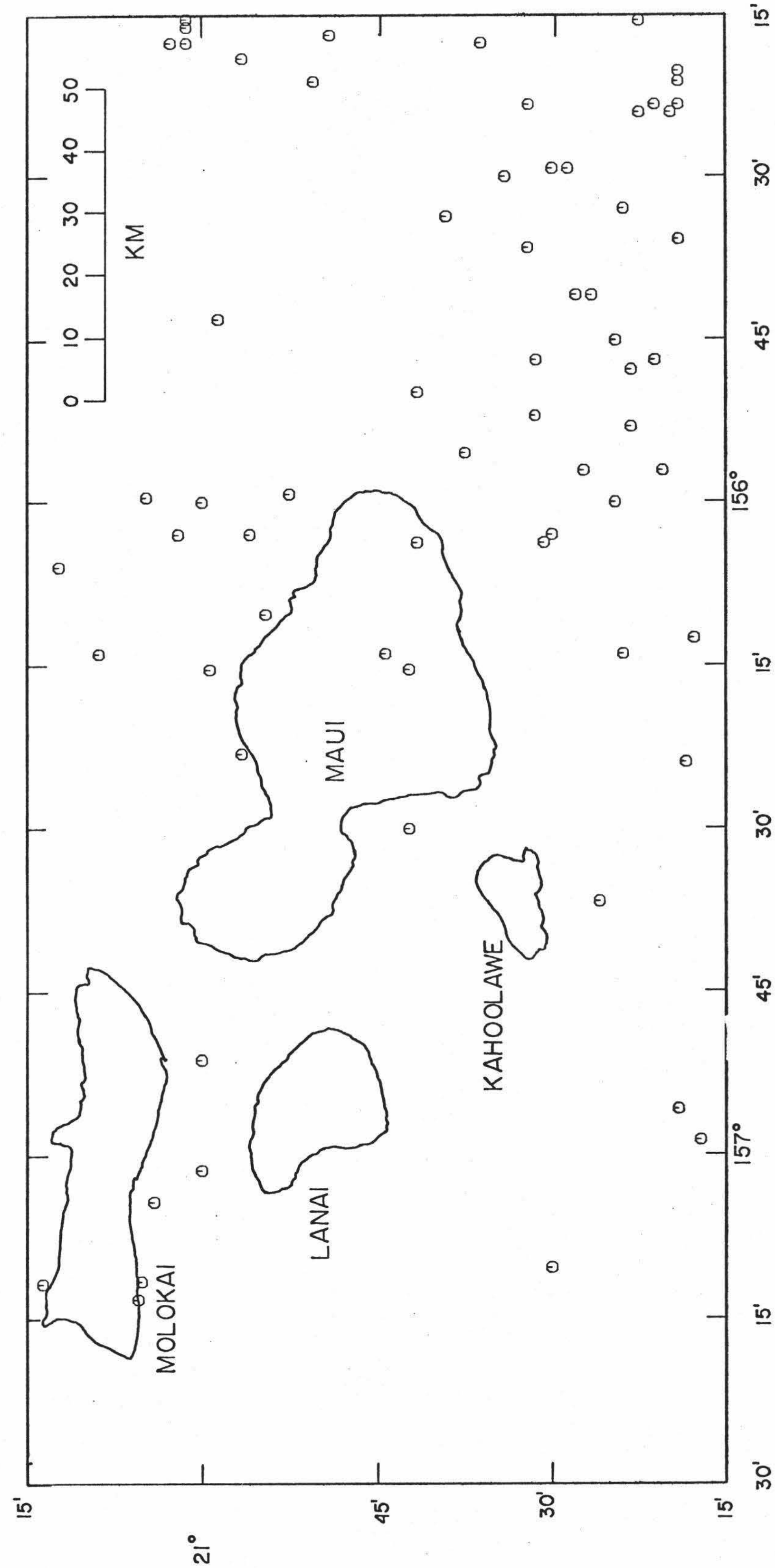
The "fixed" observatories, McDonald Observatory in Texas and Haleakala Lunar Ranging Observatory on Maui made repeated laser surveys to the moon. Ranges from Haleakala Observatory were obtained for only a brief period during 1977 (Lou Macknik, personal communication). Ranges were taken three times a night, weather permitting, throughout the lunar month (except during the new moon). The instrument used was a neodymium-YAG laser, which shot 0.25 nsec pulses of a laser beam three times every second. A timing system, capable of measuring time to 100 psec, measured the transit time of a pulse to the moon and back. A run was defined as 50-300 shots fired over a 5-20 minute period, in which a significant number of return signals were received. With the known wavelength of the laser pulses (5320 A^{o}), the speed of light, and the transit time, one can calculate the distance to the moon. The estimated resolution for one measurement is $\pm$ 15 cm. For a set of observations the uncertainty decreases to $\pm$ 3 cm (Laurila, 1976). The distance to the moon along with astronomical corrections allows the determination of geocentric position (Bender, et. al, 1973). Haleakala Lunar Ranging Observatory was not operational long enough to obtain a fix on geocentric position or to compute relative movement of the North American and Pacific plates (Lou Macknik, personal communication).

Maui, although in the middle of the Pacific plate, is part of a volcanically and tectonically active area, due to its proximity to the Hawaiian hotspot (Carter, et. al., 1977). As stated in a paper on the deformational environment of the Haleakala Observatory, "Short term, rather large variations on the order of tens of centimeters should be expected in the three dimensional location of the Haleakala Lunar Ranging Station", (Berg and Sutton, 1977). In order to eliminate this motion as a possible source of error in plate movement calculations, the Hawaii Institute of Geophysics in 1976 began a study to provide support for the LURE program by monitoring local and regional crustal deformation on Maui. The program principally consisted of repeated distance measurements that linked Haleakala Observatory to points on Maui, Molokai, and Lanai. Other parts of the program included gravimetric surveys, tide gauges to tie sea level variation with level surveys on Maui, tiltmeters to measure variations in the local vertical, and seismometers to monitor local crustal activity (Berg, et. al., 1978). One of the aims of the project was to correlate possible deformation on Maui detected by seismometers to that detected by changes in line lengths.

Measurable Crustal Deformation Detected by Seismic Means

The current seismicity study shows two areas of activity that may be sources of measurable crustal deformation in the Maui-Molokai-Lanai region. These are Haleakala Volcano and the area between Molokai and Lanai (Fig. 10). Historical earthquake data (Estill, 1979; HVO seismicity summaries) indicate 2 additional sources: between Maui and the Big Island, and north of Maui (Fig. 12). Earthquakes beneath Haleakala during the monitoring period had low magnitudes (-0.07 to 2.18), were relatively deep (20 to 40 km), and scattered in space.

Figure 12 Earthquake Epicenters Near Maui, Molokai, and Lanai (After Estill, 1979).

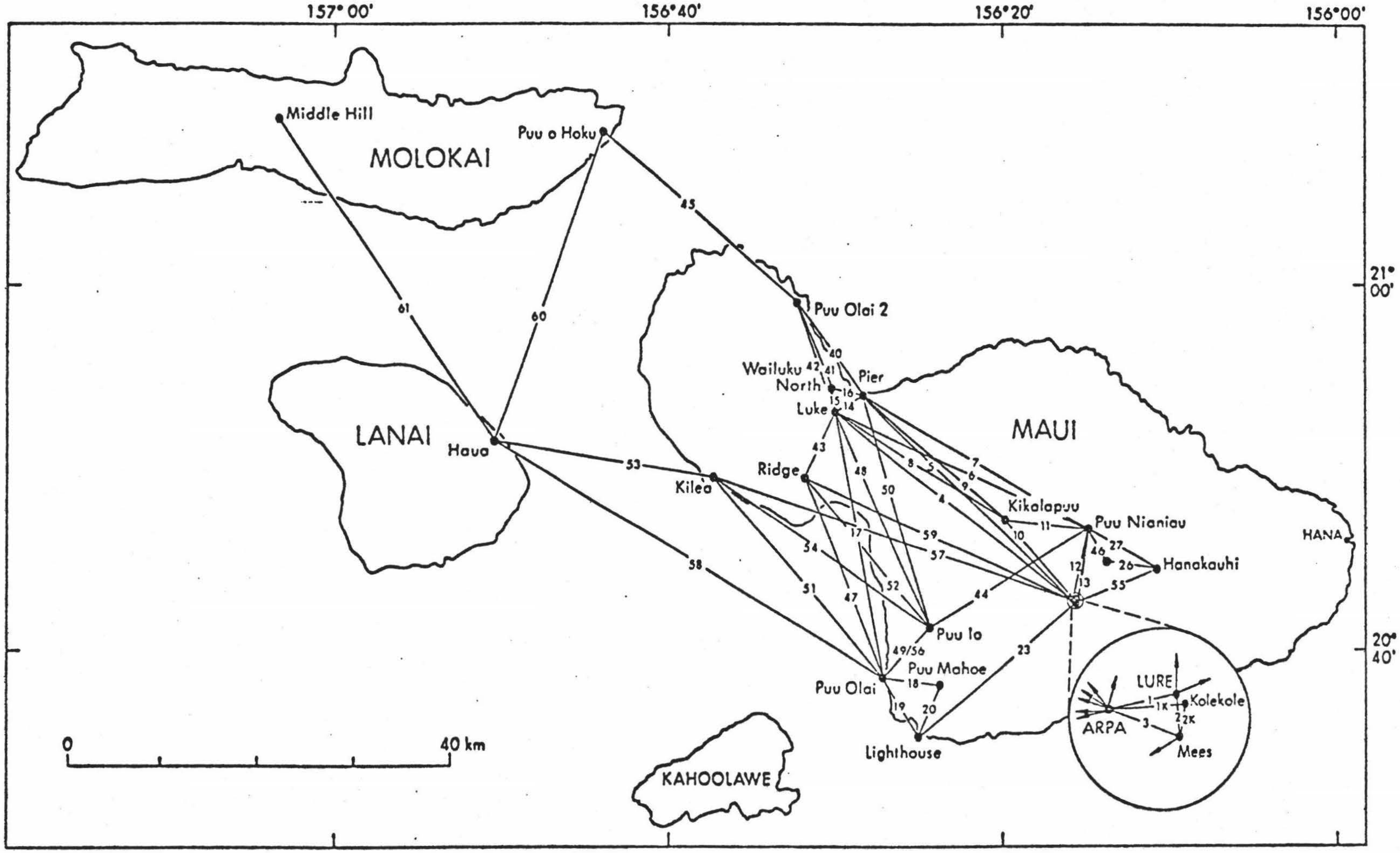

MOLOKAI
LANAI
MAUI
KAHOOLAWE
0 10 20 30 40 50
KM
30' 15' 157° 45' 30' 15' 156° 45' 30' 15'
15' 21° 45' 30' 15'

They do not appear to be associated with any particular tectonic feature. Although the microseismicity appears to be volcanically related, the earthquakes monitored were not part of any swarm activity or harmonic tremor that would accompany inflation caused by magma injection. Because these events are small in magnitude and number, this area is not likely to be a region of measurable crustal deformation. Only four earthquakes (all less than magnitude 1.5) were detected north of Maui and Molokai during the current seismicity study. Neither of these two areas appears to be a source of observable crustal deformation. Deformation associated with earthquakes is more likely to be detected beneath the channels between Maui and the Big Island and between Molokai and Lanai. Earthquakes located in the areas were larger (magnitude 3-5) and shallower (5-15 km), than the earthquakes beneath Haleakala. Also, more earthquakes greater than magnitude 4 have been detected historically between Maui and the Big Island than beneath Haleakala (Estill, 1979; HVO seismicity summaries).

Comparison of Measurable Crustal Deformation Detected by Seismic and Geodetic Means

A geodetic network on Maui, Molokai, and Lanai (Fig 13), designed to measure local and regional crustal deformation, was established over a four year period from 1976 to 1980. Geodetic surveys were performed with a helium-neon laser with the distance between the laser and retroreflectors determined by multiplying the velocity of light in a vacuum by the number of frequency modulated wavelengths and dividing by

Figure 13 The Maui-Molokai-Lanai Geodetic Network (After Berg, et. al., 1981).


157° 00'
156°40'
156°20'
156°00'
21° 00'
20° 40'
Middle Hill
Puu o Hoku
MOLOKAI
45
61
60
Puu Olai 2
LANAI
Haua
MAUI
Wailuku North
Pier
Luke
Ridge
Kilea
53
58
Kikalapuu
Puu Nianiau
Hanakauhi
HANA
Puu Io
Puu Mahoe
Puu Olai
Lighthouse
KAHOOLAWE
LURE
Kolekole
ARPA
Mees
0
40 km

an assumed index of refraction. After a survey, the distance was adjusted using the actual index of refraction calculated from atomospheric data recorded during the survey. Lines longer than five kilometers generally had a helicopter fly along the line to gather the atomospheric data; otherwise temperature and pressure data were recorded at the base and reflecting stations. Line lengths were reduced and fitted into a least squares Mercator plane projection. The standard of accuracy desired for line length measurements was 1 part in 10^7 (Berg, et. al., 1978). The final accuracy obtained was 3 parts in 10^7 (Berg, et. al., 1981).

There are five geodetic lines near possible areas of measurable crustal deformation detected by seismometers (Fig 14, Table 6). The lines interconnect Maui, Molokai, Lanai. They are Puu Olai 2 to Puu O Hoku (Maui to Molokai), Kilea to Haua (Maui to Lanai), Puu Olai to Haua (Maui to Lanai), Haua to Puu O Hoku (Lanai to Molokai), and Haua to Middle Hill (Lanai to Molokai). Not all lines were surveyed in the same year and the number of individual measurements used to determine line length varied for each line (see Table 6 for details). Line length data from 1978 are suspect due to an unremovable 1 mbar barometer calibration error (Berg, et. al., 1981). The error may not seem significant until one realizes that a 1 mbar error in atomospheric pressure can cause an error of 3×10^{-7} in line length. Data from 1979 and 1980 are not subject to the error.

Figure 14 Five Geodetic Lines with Possible Measurable Crustal Deformation as Indicated by Seismicity (After Berg, et. al., 1981).

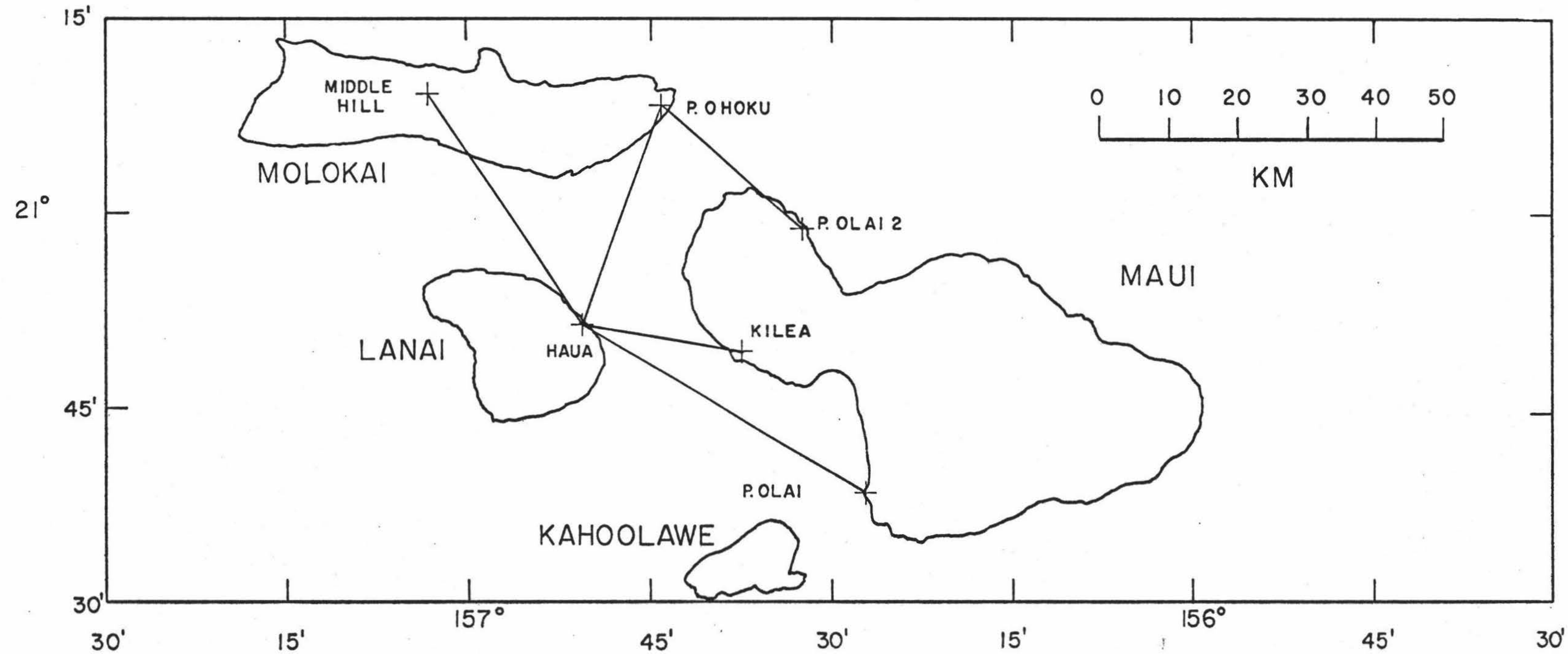

15'
21°
45'
30'
30'
15'
157°
45'
30'
15'
156°
45'
30'
MIDDLE HILL
P. OHOKU
MOLOKAI
P. OLAI 2
MAUI
LANAI
HAUA
KILEA
P. OLAI
KAHOOLAWE
0 10 20 30 40 50
KM

Table 6

Maui Geodetic Network Lines
with Possible Measurable Crustal Deformation
(after Berg, et. al., 1981)

From	To	Year	Av Line Length	Tot # Obs	Av Std dev Between yrs (mm)	Av Line Length Diff Between yrs (mm)		Twice Av Std Dev (mm)
P Olai 2	P OHoku	1978	26 849.119	101	5.0	2	<*	12
			.121	157	7.0			
Kilea	Haua	1978	23 293.411	91	3.7	19	>	7
		1979	.430	142	3.2	10	>	5
		1980	.420	120	1.3			
P Olai	Haua	1979	47 196.076	26	5.4	44	>	10
		1980	.120	14	4.3			
Haua	P OHoku	1979	33 097.869	182	9.0	9	<	13
		1980	.878	45	4.4			
Haua	Middle Hill	1980	39 557.655	189	7.4			

* If the average line length difference between years is less than twice the average standard deviation between years, crustal deformation may not be geodetically measurable. If the average line length difference between years is greater than twice the average standard deviation between years, crustal deformation may be geodetically measurable.

Possible crustal deformation was determined in the following manner. The observed change in average line length over time was chosen as the criterion for crustal deformation. Measurable crustal deformation could be expected if the average line length difference in successive years exceeded twice the line's average standard deviation for those years. If line length change were less than the cutoff, measurable crustal deformation might still exist, provided there were systematic tendencies in the data. The yearly average distances, standard deviations, line length differences, and standard deviations between years are listed in Table 6.

None of the five lines appears to display measurable crustal deformation, even though two lines exceed the cutoff. The line from Haua to Middle Hill cannot be considered because it was surveyed only once. The distance between Puu Olai 2 and Puu O Hoku appears to be constant, but some of the data is from 1978 and the line was surveyed only two years, therefore the results remain in doubt. The line Haua to Puu O Hoku also falls below the cutoff. There are only two years of data for this line also, so no trend could be established. Even though the line Puu Olai to Haua exceeds the deformation criterion, the data are inconclusive, because this line was also surveyed for only two years. The line from Kilea to Haua also exceeds the criterion; however, there are data from 1978 and there is no clear pattern in the line length differences. These factors cloud any conclusion on crustal deformation.

Comparisons of the seismic and geodetic crustal deformation data are not conclusive. The seismic network operated from May 1980 until August 1981. Geodetic line measurement started in 1977 and ended in 1980. Thus, the seismic and geodetic networks were never operational for two successive years. For valid comparisons, it is necessary to assume that the rate of crustal deformation remained unchanged during the years 1977 through 1981. Clearly, this is not the case. The magnitude 5.0 March 5, 1981 Molokai earthquake was the largest tremor in the Maui region since 1957. If one ignores the 5.0 magnitude earthquake and its aftershocks and assumes the remainder of the earthquakes detected during the monitoring period are representative of the years 1977-1980, then earthquakes in the Maui-Molokai-Lanai region were not a likely source of measurable crustal deformation during the geodetic surveys.

CHAPTER V

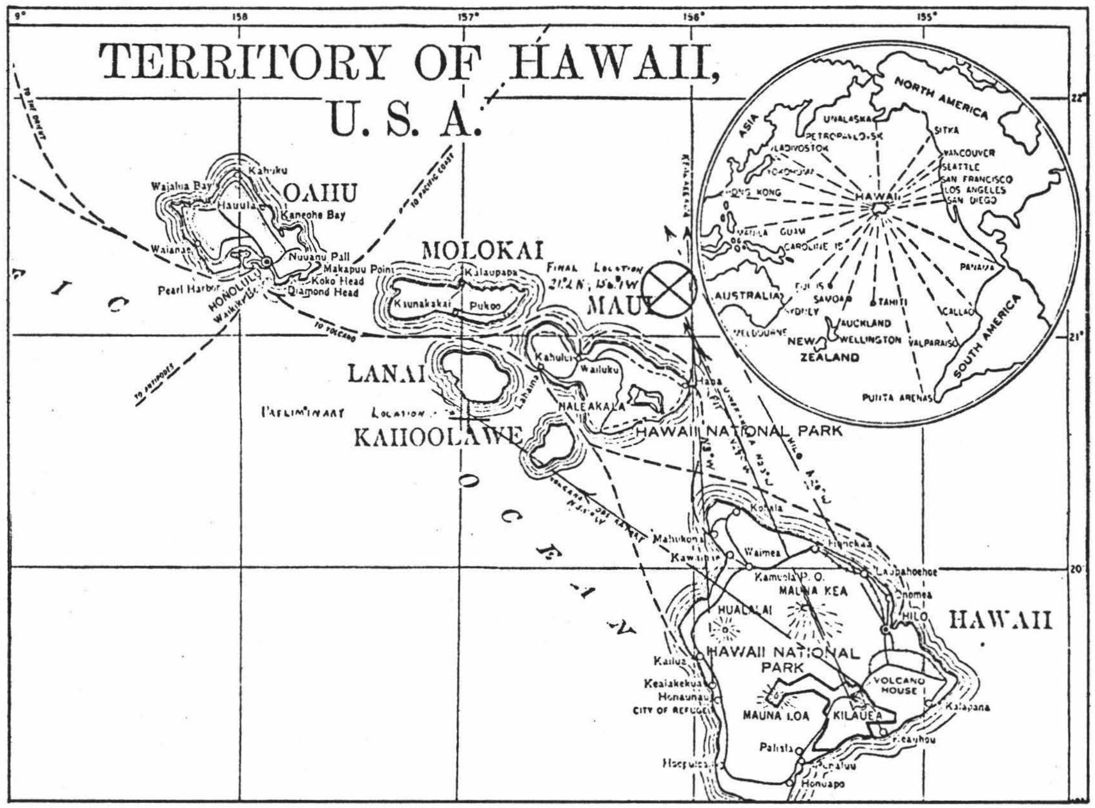
THE RELOCATION OF THE 1938 MAUI EARTHQUAKE

Description of the 1938 Maui Earthquake

On January 23, 1938, at approximately 08:30 GMT (January 22,1938, 10:00 HST), a strong earthquake (magnitude 6 3/4) shook the Hawaiian Islands. The Hawaiian Volcano Observatory placed the hypocenter twenty-five miles northeast of Pauwela Point, Maui at 21.2^{o} N. and 156.1^{o} W. (Fig. 15) and at a depth of sixty-five miles. The origin time was determined to be 08:32:47 (Volcano Letter, 1938). HVO used the arrivals on Oahu and the island of Hawaii to calculate the hypocenter. Teleseismic arrivals confirmed the computed epicenter (Volcano Letter, 1938).

Compilations of newspaper reports and eyewitness accounts published by HVO (Volcano Letter, 1938) and the Seismological Society of America (Seismological Notes, 1938) describe the reactions to the Maui earthquake. Eastern Maui experienced the most severe damage. A crevice 15 feet long and 10 feet wide opened and underground water pipes were broken. The shock also caused landslides to block roads, stone walls to crumble, and water tanks to topple. Extensive, but less severe damage occurred on Western Maui, Molokai, and Lanai: a crack in a concrete building in Lahaina represented the area's worst damage. Although Oahu received minor shaking causing little property damage, people in

Figure 15 The Original HVO Location of of the 1938 Maui Earthquake.



TERRITORY OF HAWAII, U. S. A.
OAHU
Kahuku
Wajalua Bay
Hauula
Kaneohe Bay
Waianae
Nuuanu Pali
Makapuu Point
Koko Head
Diamond Head
Pearl Harbor
HONOLULU
TO PACIFIC COAST
TO VOLCANO
TO ANTIPODES
TO THE ORIENT
MOLOKAI
Kalaupapa
Kaunakakai
Pukoo
FINAL LOCATION
MAUI
Kahului
Wailuku
Lahaina
HALEAKALA
Hana
LANAI
PRELIMINARY LOCATION
KAHOOLAWE
HAWAII NATIONAL PARK
Kohala
Mahukona
Waimea
Kamuela P. O.
MAUNA KEA
Laupahoehoe
Onomea
HILO
HUALALAI
HAWAII
Kailua
Kealakekua
Honaunau
CITY OF REFUGE
MAUNA LOA
KILAUEA
VOLCANO HOUSE
Kalapana
Pahala
Honuapo
NORTH AMERICA
ASIA
UNALASKA
SITKA
VLADIVOSTOK
VANCOUVER
SEATTLE
SAN FRANCISCO
LOS ANGELES
SAN DIEGO
HONG KONG
HAWAII
GUAM
CAROLINE IS
PANAMA
AUSTRALIA
SAMOA
TAHITI
SYDNEY
AUCKLAND
CALLAO
NEW ZEALAND
WELLINGTON
VALPARAISO
SOUTH AMERICA
PUNTA ARENAS
158
157°
156°
155°
22°
21°
20°

downtown Honolulu and Waikiki panicked and rushed from hotels and theaters. The inhabitants of Kauai reacted more calmly; yet, some of the older residents described it as the worst shock they could remember. Most people on the island of Hawaii felt the earthquake, but dismissed it as being no more severe than the tremors they usually experienced. The northern portion of the Big Island incurred the most damage on the island. The quake was strong enough to dismantle all the seismometers on the island, except for the low magnification instrument at the HVO. The damage reports from all the islands indicate that the 1938 Maui earthquake was a major earthquake.

The damage reports and descriptions yielded the following intensities for each area. (See Appendix I for a listing of the Modified Mercalli Scale, 1956 revised, used to estimate the intensities.)

East Maui (Keanae)	IX
East Maui (Kahului)	VIII
West Maui (Lahaina)	VIII
Molokai	VII
Lanai	VII
E. Oahu (Honolulu)	VI
W. Oahu	V
Kauai	IV
N. Hawaii	VI
Central Hawaii	V
Southern Hawaii	IV

An isoseismal map shows an intensity greater than IX on the north coast of East Maui; yet the estimated epicenter falls outside the highest intensity isoseismal (Fig. 3).

The 1938 Maui earthquake is an excellent candidate for relocation, since the estimated epicenter lies beyond the area of maximum intensity. Also, the computed depth of sixty-five miles seems excessive for Hawaiian earthquakes. Average Hawaiian earthquake depths are on the order of five to forty kilometers (Estill, 1979; HVO seismicity summaries). The Arrival Time Difference Earthquake Relocation Method (Spence, 1980), which uses P wave arrival time differences to relocate an unknown earthquake with respect to a nearby reference earthquake, may provide a better hypocenter than that determined in 1938. An explanation of the method is given in Appendix II.

Data Acquisition and Reduction

The reference earthquake chosen was the 7.1 magnitude November 29, 1975 Kalapana earthquake. The location of the epicenter was 19.334^{o} N. and 155.024^{o} N. (Earthquake Data Report, 1976). The 2 1/4^{o} separation of the 1975 Kalapana earthquake and the 1938 Maui earthquake location (21.2^{o} N. and 156.1^{o} W.) satisfies the ATD model criterion of a separation of less than 2 1/2^{o}. Finally, the earthquake provided clear teleseismic arrivals throughout the world (Earthquake Data Report, 1976).

Teleseismic (>30°) P wave arrival times from sixteen stations operating in 1938 and 1975 provided the data for the relocation process (Table 7). When possible, the arrival times, corrected for clock error, were picked from seismograms. Six seismic observatories supplied seismograms, others (that did not have records) supplied only arrival times. If requests for seismograms went unanswered, arrival times published by the International Seismological Summary were used. The number of obsevations is less than desired because many of the stations operating in 1938 closed before 1975. The number of stations was also reduced by selecting arrivals for the 1938 earthquake with residuals of ten seconds or less in the International Seismological Summary. This step helped to insure that the P wave arrival was properly picked. To further minimize error, any station locations that were different in 1975 from those in 1938 had a station correction added to the 1975 arrival. The correction was calculated by dividing the difference of the 1938 epicentral distance to the two station locations by the average P wave velocity for the ray path. The arrival times for the reference earthquake in 1975 were primarily taken from the Earthquake Data Report published by the USGS. Only Riverview (RIV) and Hong Kong (HKC) were picked from seismograms. The ATD relocation was solved using singular value decomposition (explained in Appendix III) and a computer program called DECOM (listed in Appendix IV) calculated the relocation.

Table 7

Relocation Data
for the 1938 Maui Earthquake

STA[*]	LOCATION	Δ°	AZ°	P Arrival 1938 (m s)	P Arrival 1975 (m s)	P -P 38 75 (s)	DEPTH (km) P wave JB[&]	DEPTH (km) P wave H[#]	DEPTH (km) pP
PAS	34° 08′ 54.0" N 118° 10′ 18.0" W	35.8	60.7	7 02	7 3.6	-1.6	S	0	30 km
BRK	37° 52′ 24.0" N 122° 15′ 36.0" W	33.6	52.8	6 45	6 49.6	-4.6	S	0	30 km
HUA	12° 02′ 18.1" S 75° 19′ 22.1" W	86.0	104.4	12 40	12 35.5	4.5	S	15	
LPZ	16° 29′ 43.0" S 68° 07′ 57.7" W	94.1	106.0	13 25	13 9.0	16.0	S	0	
TOK	35° 41′ 12.0" N 139° 45′ 30.0" E	57.3	299.6	9 43	10 3.6	-20.6	33	50	
COL	64° 54′ 0.0" N 147° 47′ 36.0" W	44.0	5.1	8 07	8 20.6	-13.6	66	50	
BRS	27° 28′ 42.0" S 153° 01′ 54.0" E	69.1	227.6	11 07	11 05.3	1.7	33	0	
RIV	33° 49′ 47.5" S 151° 09′ 30.0" E	74.3	223.5	11 37	11 32.8	4.2	33	0	
VIC	48° 24′ 49.0" N 123° 19′ 26.0" W	37.7	36.1	7 18	7 29.6	-11.6	S	0	

Table 7 (continued)

Relocation Data
for the 1938 Maui Earthquake

SIT	57° 03′ 25.0" N	38.9	17.9	7 32	7 40.2	-8.2	S	0
	135° 19′ 28.0" W							
TAC	19° 24′ 18.0" N	53.2	81.1	9 22	9 12.6	9.4	S	0
	99° 11′ 37.0" W							
TUO	32° 14′ 48.0" N	41.6	64.9	7 52	7 49.4	2.6	S	0
	110° 50′ 06.0" W							
WES	42° 23′ 4.9" N	72.3	50.7	11 28	11 32.3	-4.3	S	0
	71° 19′ 19.5" W							
MHC	37° 20′ 31.2" N	33.9	54.0	6 48	6 47.7	0.3	S	0
	121° 38′ 31.2" W							
BUT	46° 00′ 48.0" N	43.2	44.5	8 08	8 09.9	-1.9	S	0
	112° 33′ 48.0" W							
HKC	22° 18′ 12.8" N	82.0	290.1	12 25	12 31.0	-6.0	S	0
	114° 10′ 18.8" E							

* Standard three letter code for seismographic stations

& Jeffreys-Bullen Seismological Tables depth estimates

\# Herrin Seismologiacal Tables depth estimates

@ Surface Earsthquake, Depth < 15 km

Discussion

The relocated hypocenter for the 1938 Maui earthquake is 21.02° N. ± 25 km and 156.09° W. ± 20 km, with a depth of 20 km ± 15 km. The relocated origin time was 08:32:47 ± 3 sec (Fig. 16). The RMS for the arrival time difference residuals is 2.38 seconds and the weighted standard error of the residuals was 2.46 seconds (Table 8). The weighted standard error is large and is due to the approximation in station location and errors in time-keeping (Dumas, et. al., 1980). P wave arrival times for twelve of sixteen stations indicated a depth of 15 or less. Arrival time differences for pP-P for Pasadena and Berkeley indicate a depth of less than 30 km. The body wave magnitudes calculated for the short period records at Pasadena and Berkeley are 6.9 and 6.8 respectively. Using a surface wave with a period of ten seconds, the magnitude for Berkeley is 6.9. These magnitudes are in agreement with the intensity of IX on the Modified Mercalli Scale 1956 Revised Version (Nuttli, et. al., 1979).

The relocated epicenter is 0.18° south of its previous location of 21.2° N. and 156.1° W. The relocated epicenter seems to be an improvement over the 1938 estimated epicenter, because it is within the highest intensity isoseismal and within a zone of historical seismicity (Fig. 17) (Estill, 1979; HVO seismicity summaries). The depth estimated in 1938 has been reduced from 65 miles to 20 km and is more in agreement with the known depths of Hawaiian earthquakes. The relocated origin time is 2 seconds earlier than in 1938.

Figure 16 Isoseismal Map of the 1938 Maui Earthquake. The Relocated Epicenter is Indicated by a Square and the 1938 Estimated Epicenter by a Triangle.

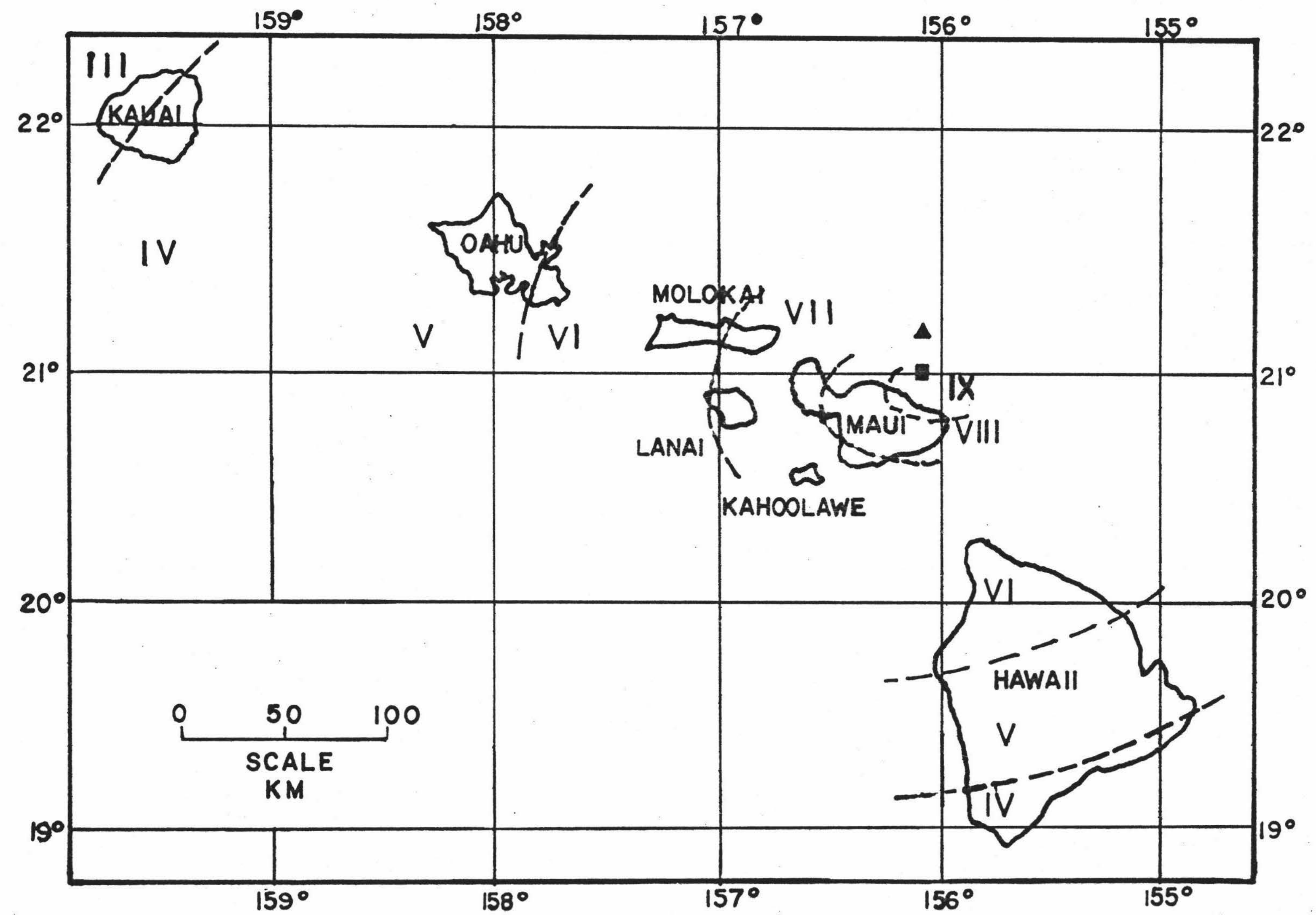

159°
158°
157°
156°
155°
22°
21°
20°
19°
III
KAUAI
IV
OAHU
V
VI
MOLOKAI
VII
LANAI
KAHOOLAWE
MAUI
IX
VIII
HAWAII
VI
V
IV
0 50 100
SCALE
KM

Table 8

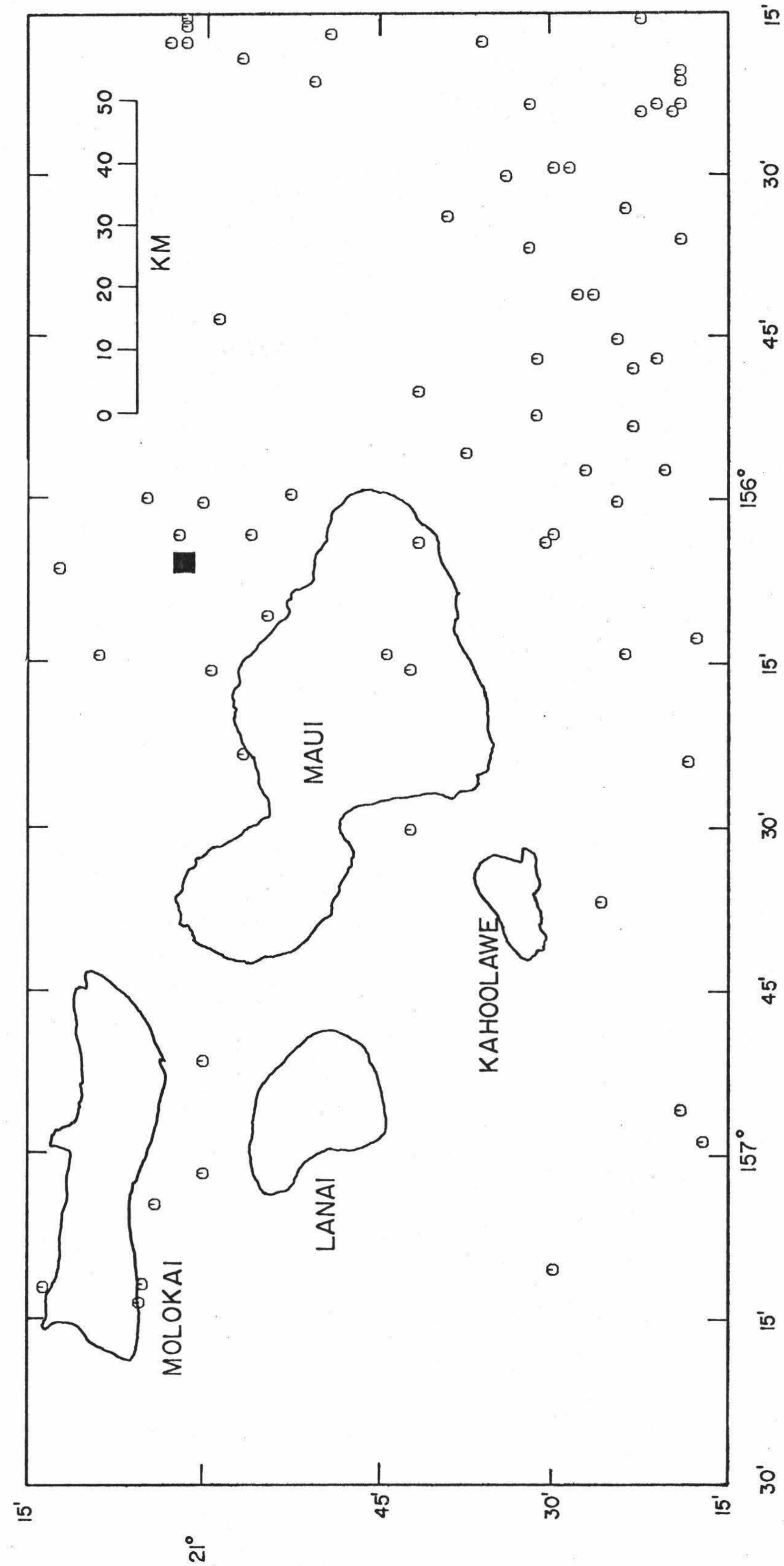
Summary Statistics
for the Relocation
of the 1938 Maui Earthquake

STA	DELTA° 1938	DELTA° Reloc	AZIMUTH° 1938	AZIMUTH° Reloc	RESIDUAL (sec) 1938	RESIDUAL (sec) Reloc	WT
BRK	33.59	33.95	52.79	52.58	0	-0.32	1.000
MHC	33.87	33.98	53.98	53.77	+1	2.06	.952
PAS	35.79	35.88	60.65	60.43	-1	0.01	1.010
VIC	37.73	37.88	36.07	35.94	-1	1.39	.973
SIT	38.93	39.10	17.94	17.87	+3	2.14	.942
TUO	41.65	51.72	64.92	64.74	+1	1.37	.973
BUT	43.17	43.30	44.51	44.37	+4	2.16	.942
COL	44.01	44.18	5.08	5.06	-4	-4.57	.747
TAC	53.19	53.22	81.10	80.97	0	2.34	.932
TOK	57.32	57.41	299.57	299.66	-9	-4.50	.788
BRS	69.07	68.95	227.58	227.63	-2	-0.17	1.000
WES	72.27	72.38	50.68	50.63	-1	-0.58	.993
RIV	74.33	74.20	223.49	223.52	-4	-0.62	.984
HKC	82.00	82.07	290.84	290.86	+2	1.21	.983
HUA	85.96	85.92	104.42	104.41	-3	-1.15	.962
LPZ	94.09	94.04	105.95	105.96	-1	4.84	.788

Origin time 1938: 08:32:47 ± 3.47 sec RMS 3.32 sec

Relocated origin time: 08:32:45 ± 2.87 sec RMS 2.46 sec

Figure 17 The 1938 Maui Earthquake Original and Relocated Epicenters Superimposed on Historical Seismicity. Note that the Relocated Epicenter Lies Near an Area of Seismicity (After Estill, 1979).


KM
0 10 20 30 40 50
MOLOKAI
LANAI
MAUI
KAHOOLAWE
157°
156°
15'
30'
45'
21°

The relocation errors are similar to those obtained by Spence (1980) in his relocation of nuclear blasts in the Aleutians, but the relocation of the 1938 Maui earthquake may be a poor one. Spence (1980) recommended using at least 35 stations in order to insure stable relocations. Only sixteen P wave arrival time differences could be used in this study. Further, the stations used are mostly confined to North America and Australia. Only two stations (HUA and LPZ) represent South America and TOK is the only one northwest of Hawaii. There are none to the south. The solution, then, may be biased by the narrow azimuthal ranges of the stations used.

CHAPTER VI

EARTHQUAKE FREQUENCY, STRESS, AND SEISMIC RISK

FROM EARTHQUAKES ON MAUI, MOLOKAI, AND LANAI

The 1938 Maui earthquake and the 7.1 magnitude 1871 Molokai earthquake are ample evidence that large Hawaiian earthquakes can occur in places other than the island of Hawaii. Damaging earthquakes such as these pose an especially serious threat to populated areas on Maui, Molokai, Lanai, and Oahu. In order to estimate the dangers posed it is important to be able to predict the expected magnitude and recurrence rates of large earthquakes.

b values and Regional Stress

It has been shown that the number of earthquakes (N) occurring over a given time period is related to the magnitude (M) by the Gutenberg-Richter relation $\log N = A - bM$, where A and b are constants (Richter, 1958). For most seismic areas in the world b values are remarkably constant. They vary between 0.5 and 1.5 and average 1.0 (Furumoto, et. al., 1973).

Two studies were made of b values for earthquakes near Maui. One was for the short term (15 month) period mid-May 1980 to mid-August 1981 and the other was for the long term period October 1961 to August 1981. Seventy earthquakes from the current Maui seismicity study were used for the short term study, while HVO seismicity summaries until June 1980 coupled with the current Maui seismicity study provided 36 earthquakes greater than magnitude 3 for the long term (Fig. 18, Table 9). Records for the period January 1960 until September 1961 have not been processed by HVO and data before 1959 were not used because HVO published only general locations (e.g. 14 mi east of Hana) and qualitative magnitudes.

In order to calculate meaningful b values a cutoff magnitude is needed below which the Gutenberg-Richter relation fails. The cutoff for the Maui seismicity study was 3.0 because of incomplete sampling below that magnitude. The cutoff for the historical data was 3.5. The cutoff for the historical data was higher than the current seismicity study, because the HVO network is located primarily on the island of Hawaii. Any smaller distant event from Molokai might be too weak to detect.

Six earthquakes with magnitudes greater than 3.0 occurred between May 1980 and August 1981 and seventeen earthquakes with magnitudes greater than 3.5 occurred between October 1961 and August 1981. The b values were calculated using a computer program based on cumulative frequency and the maximum likelihood method (Carter, 1978; Aki, 1965). The advantages to this method are (1) accurate b values can be obtained with small data sets and (2) undue weight is not given to poorly represented large magnitude earthquakes (Carter, 1978).

Figure 18 HVO Epicenters for the Maui-Molokai-Lanai Region with Magnitudes > 3.0 (October, 1961 to May 1980).

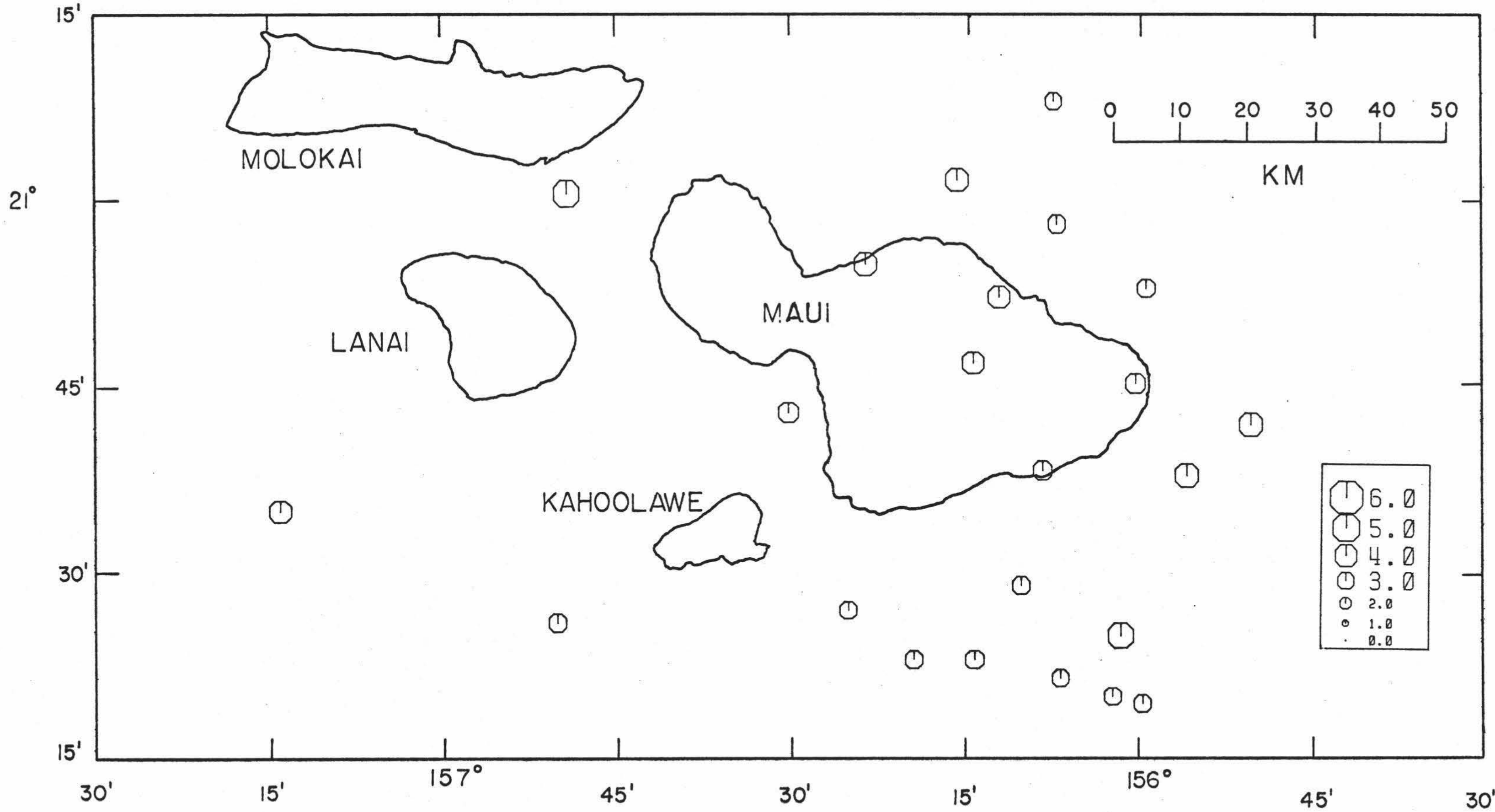

MOLOKAI
LANAI
MAUI
KAHOOLAWE
0 10 20 30 40 50
KM
6.0
5.0
4.0
3.0
2.0
1.0
0.0
15'
21°
45'
30'
15'
30'
15'
157°
45'
30'
15'
156°
45'
30'

Table 9

Historical Earthquake Locations from HVO:
October 1961–May 1980

DATE	TIME HR	MIN	SEC	Lat N	Long W	DEPTH	MAG
9-28-62	12	44	43.7	21° 27′	156° 47′	13	3.3
3-11-63	17	45	28.3	21° 08′	156° 07′	13	3.0
3-25-63	7	18	35.0	20° 47′	156° 14′	35	3.9
6-28-63	22	50	50.1	21° 20′	157° 05′	13	3.1
8-25-63	22	41	43.0	21° 23′	156° 29′	13	3.1
9-21-63	00	17	48.0	20° 27′	156° 25′	13	3.1
10-22-63	7	35	45.2	20° 23′	156° 14′	13	3.3
1- 9-64	9	47	44.5	20° 20′	156° 02′	13	3.1
2-20-64	22	31	44.5	20° 42′	155° 50′	13	4.3
4-29-65	10	08	30.1	20° 35′	157° 14′	13	4.0
4-29-65	17	22	29.2	20° 35′	157° 14′	13	4.0
7- 8-65	23	26	2.0	20° 43′	156° 30′	5	3.6
8-30-65	16	43	15.0	20° 29′	156° 10′	8	3.2
10-14-65	16	27	37.1	20° 19′	156° 12′	13	3.1
2- 3-67	22	07	54.1	20° 53′	155° 59′	8	3.2
6-20-68	4	20	9.8	20° 44′	157° 05′	13	3.0
4- 9-69	7	30	56.1	20° 59′	155° 43′	13	4.2
6-21-69	3	48	24.0	20° 26′	156° 50′	8	3.3
5-22-70	21	56	32.8	20° 21.5′	156° 6.6′	8.0	3.2
10-25-70	9	55	28.0	21° 0.6′	156° 49.1′	7.5	4.8
1-19-71	10	08	44.4	20° 38.3′	156° 8.1′	8.0	3.5
3-27-72	6	01	1.4	21° 1.7′	156° 15.3′	38.1	4.1
9-20-73	18	07	8.9	20° 52.3′	156° 11.8′	63.0	3.9
10-13-73	6	11	11.6	20° 37.9′	155° 55.6′	14.7	4.3
7-29-75	12	53	4.85	20° 19.45′	155° 59.47′	22.73	3.1
2-20-76	19	51	14.91	20° 24.95′	156° 1.29′	31.75	4.6
5-23-76	23	24	7.63	20° 54.93′	156° 23.40′	49.09	4.2
1-31-77	21	20	51.98	20° 58.15′	156° 6.76′	7.75	3.3
3- 3-78	20	19	4.24	20° 23.00′	156° 19.22′	14.26	3.2
8-11-78	15	08	51.86	20° 45.36′	155° 59.94′	15.95	3.6

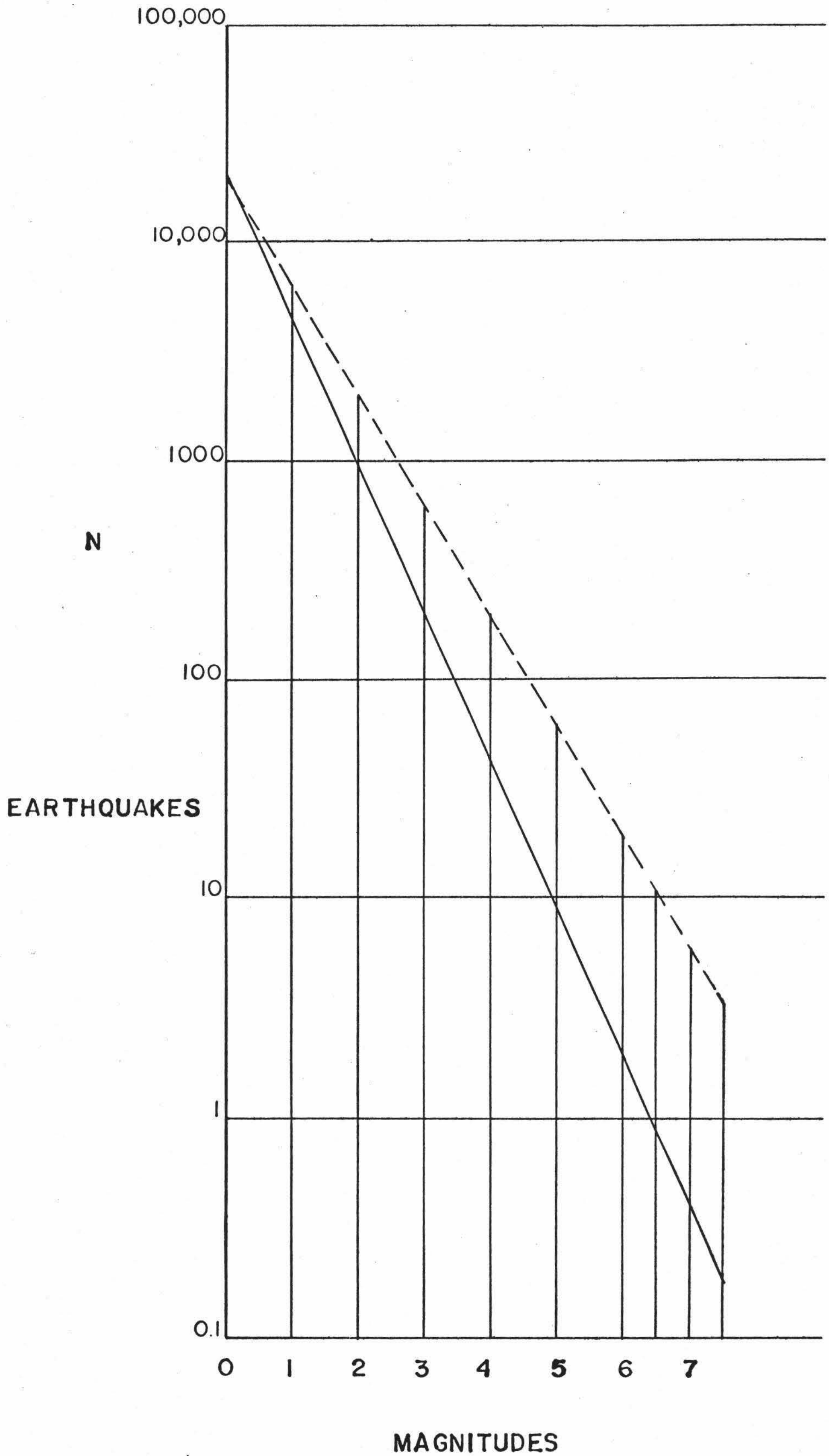
The b values for the one year and twenty year periods are similar and significantly below previously determined Hawaiian b values. The current seismicity study b value is 0.50 ± .19 and the long term b value is slightly higher at 0.67 ±.16. On the other hand, Furumoto and others (1973) calculated a b value of 1.05 ± .14 for the Hawaiian Islands. Carter (1978) and Estill (1979) both obtained b values of 0.93 ± .10 for the Hawaiian Ridge. Bosher (1981) computed an overall b value of 0.94 ± .07 for the December 1965 Kilauea earthquake swarm. The first three studies contained at most 1/4 to 1/2% Maui-Molokai-Lanai magnitude data. Consequently, these b values are more representative of the Big Island than the Maui region. The b value for the the current seismicity study was computed from data that does not represent the long term. The small data set (6 earthquakes) is based almost entirely on the magnitude 5.0 Molokai earthquake and its aftershocks. The b value for the twenty year period may be erroneous, although the data set is larger (17 earthquakes) and more consistent over a long time span. The large difference between the previous b value studies and these suggest that Maui, Molokai, and Lanai comprise a different earthquake regime separate from the Big Island.

It has been determined experimentally and empirically that b values vary inversely to stress (Mogi, 1962b; Scholz, 1968; Carter, 1978). A b value of 0.67 for the past twenty years indicates a high stress level (Scholz, 1968). An even lower b value of 0.50 for current seismicity may indicate a high stress level before or during the 5.0 magnitude Molokai earthquake. The b value for the current seismicity study may be representative of only a small portion of the Maui region, mainly from the zone near the 5.0 magnitude Molokai earthquake. The b value for the last twenty years, on the other hand is based on earthquake data from the entire Maui region.

The Time Distribution of Earthquakes in the Maui-Molokai-Lanai Region

The Gutenberg-Richter relation was used to determine earthquake recurrence rates. The values of A for the short and long term studies were calculated using least squares and the b values computed earlier. The value of A for the current seismicity study was $4.28 \pm .71$ and for the 20 year study it was $4.31 \pm .62$. Both values of A were normalized to a 100 year period. The recurrence rates for the short term study (Fig. 19) show that large earthquakes occur more frequently than indicated by the long term data (Fig. 19). The unusually high earthquake frequency predicted by the short term seismicity is not surprising, since it is almost entirely based on a mainshock and aftershock sequence. Thus the small and uncharacteristic sample would predict that large magnitude earthquakes should occur more often than they actually appear to occur. Recurrence rates for the long term are much more in agreement with historical records for magnitude 4 and 5 earthquakes. According to HVO data, magnitude 4 earthquakes occur every 2 to 3 years. These are the values predicted by the Gutenberg-Richter

Figure 19 The Occurrence-Magitude Relation (Log N=A-bM) for the Maui-Molokai-Lanai Region. The Solid Line Represents the Recurrence Rates for HVO and HIG Maui Seismic Network Data from 1962-1981. The Stippled Line Represents the Recurrence Rates Calculated from the HIG Maui Seismic Network Data (June 1980 to August 1981). Both Recurrence Rates Have Been Normalized to a 100 Year Period.


100,000
10,000
1000
100
10
1
0.1
N
EARTHQUAKES
0
1
2
3
4
5
6
7
MAGNITUDES

relation for the long term data. The last similar magnitude Maui-Molokai-Lanai earthquake occurred in 1957: the magnitude 5.6 shock, which was located 45 km east of Hana, Maui. Again, the recurrence rates are on the order predicted by the Gutenberg-Richter relation.

Another way to determine the time distribution of earthquakes is the elastic strain release method proposed by Benioff (1951). Benioff defined elastic strain release as the square root of earthquake energy, J, with energy related to the magnitude M by the equation: $\text{Log } J = 9 + 1.8M$. According to Benioff, if the cumulative strain release is plotted over time, a pattern should develop which can give a rough idea of expected earthquake magnitudes. If there is a pattern, the cumulative strain release must eventually rise to the value indicated by an envelope drawn to the curve.

The data for the strain release studies came from the same sources used in the b value and earthquake frequency studies. For the Maui seismicity study seventy earthquakes formed the data set. Only 7 earthquakes with magnitude 2.8 or greater yielded strain releases of $1 \times 10^7 \text{ (erg)}^{1/2}$. Thirty-six samples comprised the long term seismicity data set. Thirteen earthquakes had magnitudes of 3.9 or greater that gave strain releases of $1 \times 10^8 \text{ (erg)}^{1/2}$.

The strain release plot for the current seismicity (Fig. 20) shows little activity before December 1980. Twenty-three earthquakes were detected from May until December, but their cumulative strain release added to only $2x10^7$ $(erg)^{1/2}$. Two of the 7 earthquakes occurred south of Maui during December 1980 and January 1981. The other 5 earthquakes are the main and after shocks of the magnitude 5.0 Molokai earthquake. After early April 1981 the strain release curve resumed its previously quiet state. This curve is obviously dominated by the 5.0 Molokai earthquake and is not indicative of the long term trend.

Figure 21 shows the Maui-Molokai-Lanai region long term strain release pattern. The strain release is about $2\text{-}2.5x10^8$ $(erg)^{1/2}/yr$. For the last 20 years a magnitude 4 earthquake has occurred about every 2 to 3 years. The exception is the period February 1976 until March 1981. During this interval no earthquake greater than magnitude 3 occurred in the region. Apparently strain increased until it was finally released by the 5.0 Molokai earthquake. Based on the observed regular recurrence of magnitude 4 or greater earthquakes the elastic strain release method offers a valuable tool in predicting the possible magnitude of future earthquakes in this region.

Figure 20 The Maui-Molokai-Lanai Elastic Strain Release for the Period May 1980 until August 1981. Elastic Strain Release is Defined as the Square Root of Earthquake Energy. Earthquake Energy is Related to Magnitude by the Equation,

$\log J = 9 + 1.8M$

Note the Dominant Effect of the Molokai Earthquake and its Aftershocks.

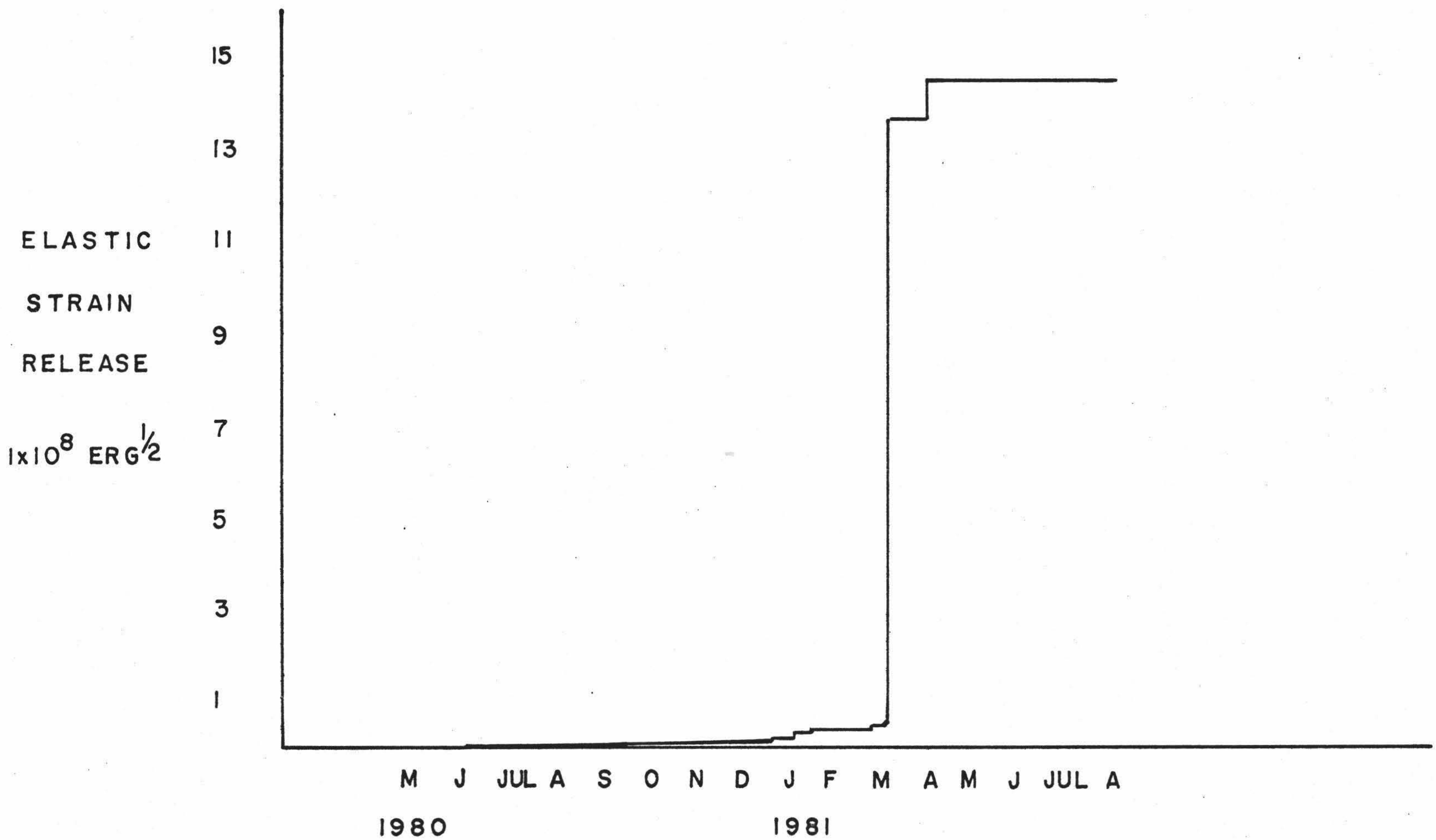
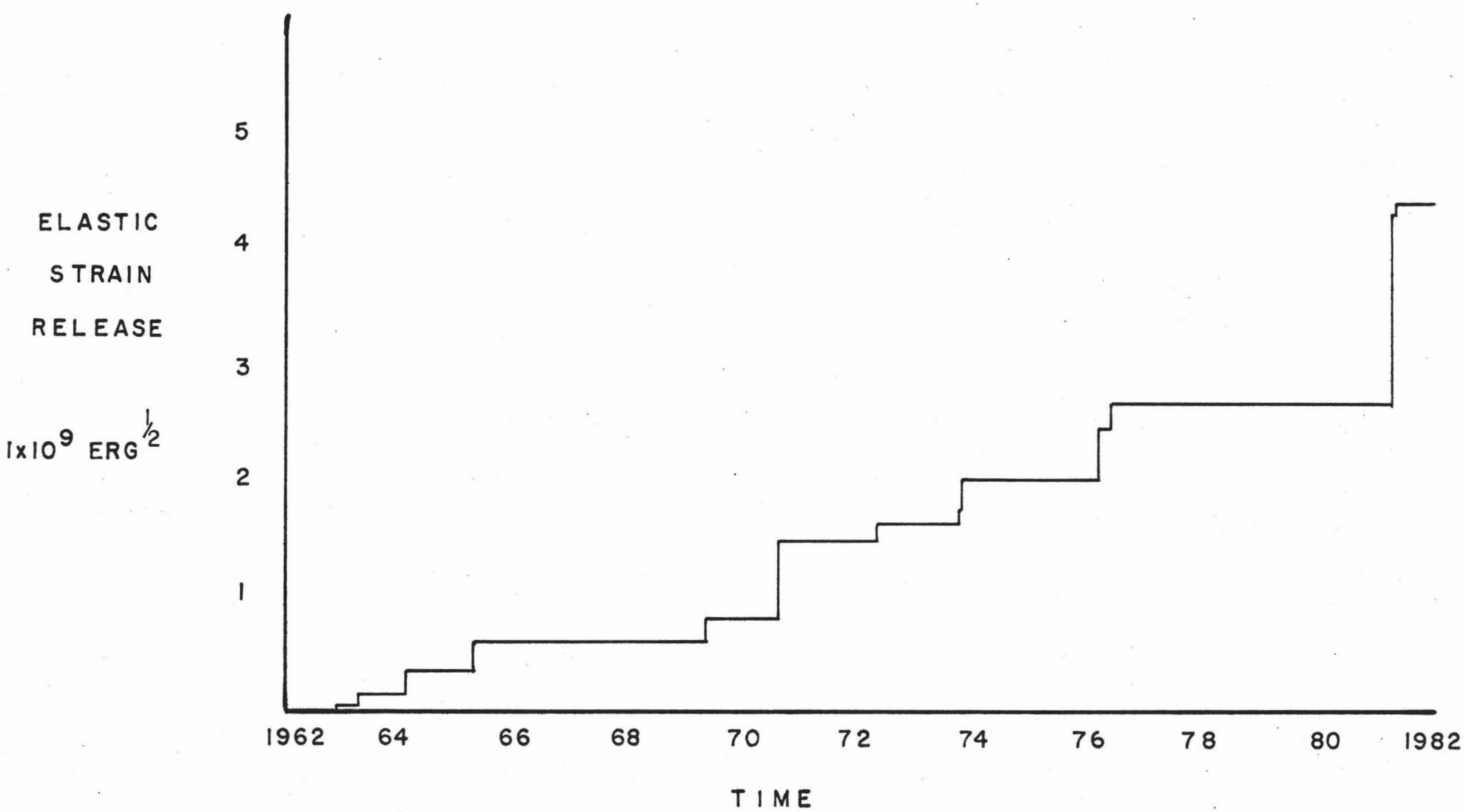

ELASTIC
STRAIN
RELEASE
1×10^8 $ERG^{1/2}$
15
13
11
9
7
5
3
1
M J JUL A S O N D J F M A M J JUL A
1980
1981

Figure 21 The Maui-Molokai-Lanai Elastic Strain Release from October 1961 to August 1981. Note the Regular Strain Release Pattern.



ELASTIC
STRAIN
RELEASE
1x10^9 ERG^½
5
4
3
2
1
1962
64
66
68
70
72
74
76
78
80
1982
TIME

Seismic Risk Zoning for Earthquakes from Maui, Molokai, and Lanai

Seismic risk maps indicate areas of approximately equal risk. The ratings are based on damage from past earthquakes and do not consider earthquake recurrence rates. The U. S. Coast and Geodetic Survey established four zones of seismic risk (shown below) and issued a seismic risk map for the conterminous United States in 1949 (Roberts and Ulrich, 1950).

Zone 0	No damage
Zone 1	Minor damage
Zone 2	Moderate damage
Zone 3	Major damage

In 1949 the International Conference of Building Officials adopted the seismic risk map into the Uniform Building Code, which is the source of many of the construction regulations for the United States (Housner and Jennings, 1974). The main effect of seismic risk zoning is to increase the level of shear forces a structure should be designed to withstand in earthquake prone areas. The requirements for Zone 2 are half those of Zone 3 and those for Zone 1 are half of those for Zone 2, etc. The map was amended in 1950 (Robert and Ulrich, 1951) and withdrawn (due to possible misinterpretation) in 1952 (Roberts and Ulrich, 1953). In 1969 the U.S.C.G.S. revised the 1949 map for the conterminous United States and included a quantitative assessment of earthquake damage (Algermissen, 1969).

Zone 0	No damage	MM Intensity < V
Zone 1	Minor damage	MM Intensity V & VI
Zone 2	Moderate damage	MM Intensity VII
Zone 3	Major damage	MM Intensity > VII

The 1970 Uniform Building Code adopted the revised U. S. seismic risk map and included seismic risk maps for Alaska and Hawaii.

The seismic risk map presently in effect for Hawaii is shown in Figure 22. It is based on data from the April 2, 1868 earthquake on the island of Hawaii, which was the largest historical earthquake in Hawaii. The map shows Hawaii in Zone 3, Maui and Kahoolawe in Zone 2, Lanai, Molokai, and Oahu in Zone 1, and Kauai in Zone 0.

Furumoto and others (1973) demonstrated that the seismic risk map for the state of Hawaii has been underestimated by not including the effects of the 1871 Molokai and the 1938 Maui earthquakes. Furumoto and others proposed changes for the state of Hawaii seismic risk map (Fig. 23) that would upgrade Maui and Kahoolawe from zone 2 to Zone 3, Molokai and Lanai from zone 1 to zone 3, Oahu from zone 1 to zone 2, and Kauai from zone 0 to zone 1.

In general, the findings of this report support the suggestions offered by Furumoto and others (1973). The relocation of the 1938 Maui earthquake does not change the estimated epicenter so as to reduce seismic risk for the Hawaiian Islands. The isoseismal map for the 1938 Maui earthquake shows Maui within an intensity VIII isoseismal.

Figure 22 The Current Earthquake Zoning Map for Hawaii. The Highest Risk Area is Zone 3; the Lowest is Zone 0. (After Furumoto, et. al., 1973).

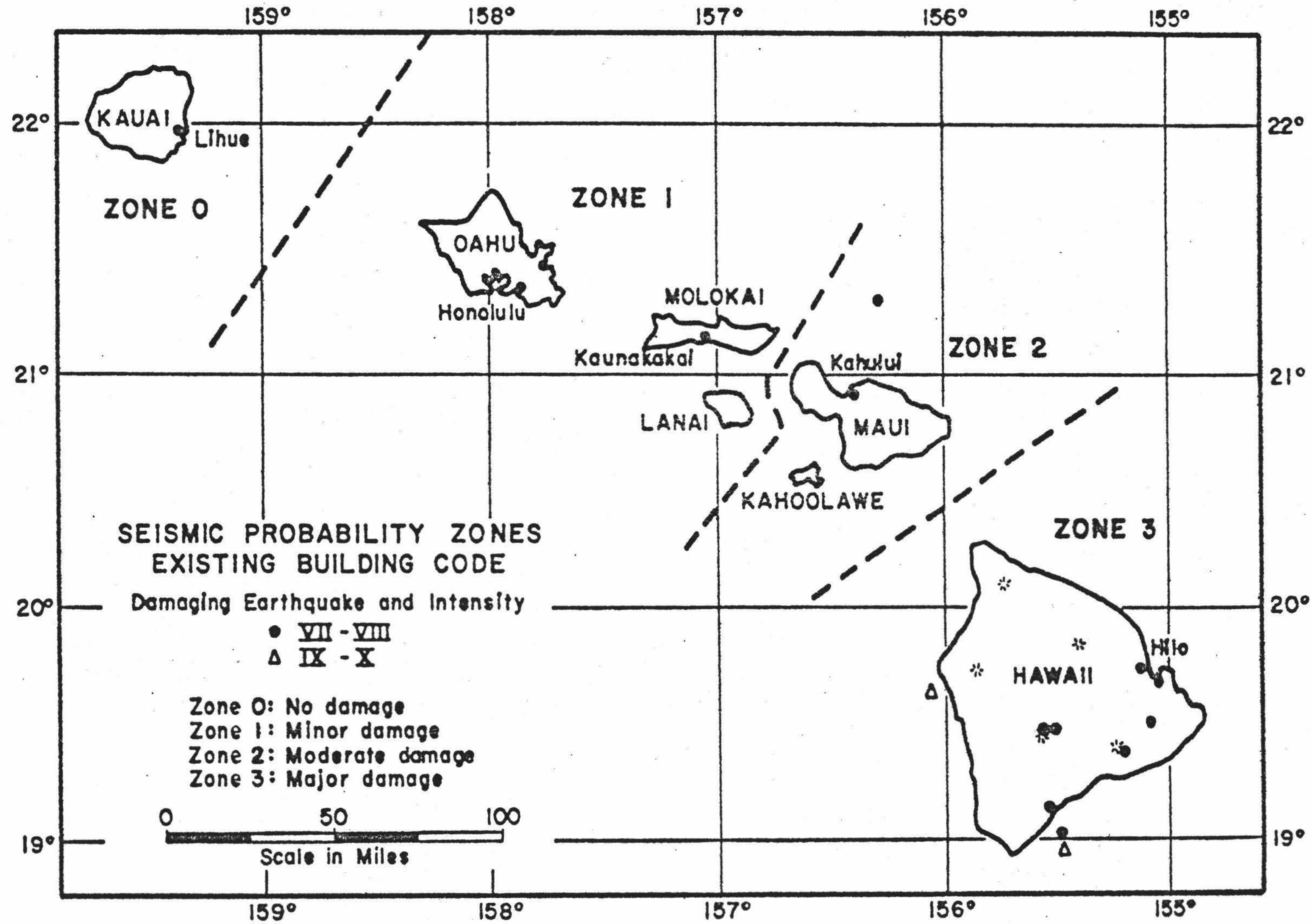
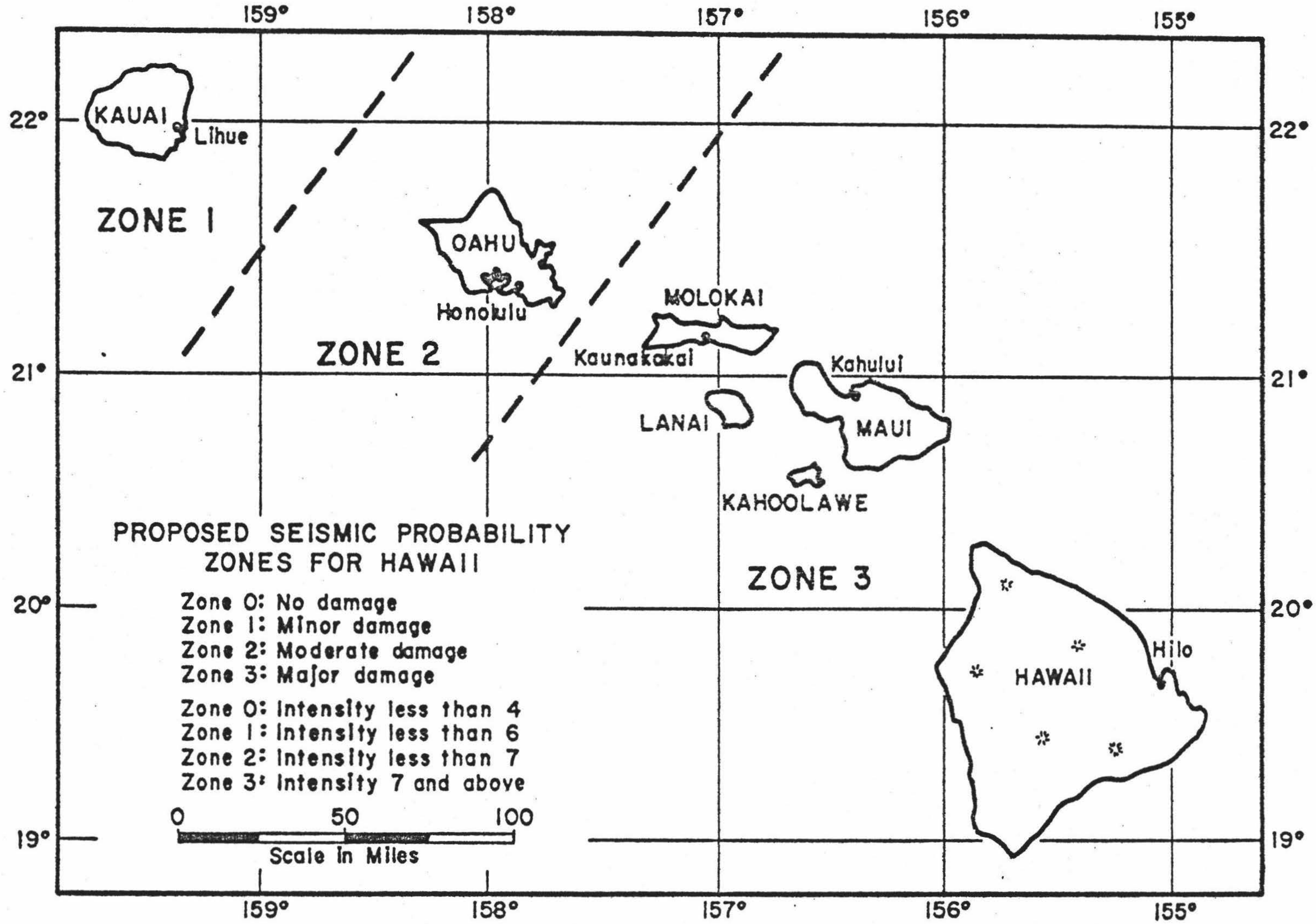

SEISMIC PROBABILITY ZONES
EXISTING BUILDING CODE
Damaging Earthquake and Intensity
● VII - VIII
Δ IX - X
Zone 0: No damage
Zone 1: Minor damage
Zone 2: Moderate damage
Zone 3: Major damage
0 50 100
Scale in Miles
ZONE 0
ZONE 1
ZONE 2
ZONE 3
KAUAI
Lihue
OAHU
Honolulu
MOLOKAI
Kaunakakai
LANAI
MAUI
Kahului
KAHOOLAWE
HAWAII
Hilo
159°
158°
157°
156°
155°
22°
21°
20°
19°

Figure 23 The Revised Earthquake Zoning Map for Hawaii Proposed by Furumoto and Others (1973). The Revised Map Places Maui, Molokai, and Lanai in Zone 3. Oahu is place in Zone 2 and Kauai in Zone 1 (After Furumoto, et. al., 1973).


159°
158°
157°
156°
155°
22°
21°
20°
19°
KAUAI
Lihue
ZONE 1
OAHU
Honolulu
ZONE 2
MOLOKAI
Kaunakakai
Kahului
LANAI
MAUI
KAHOOLAWE
ZONE 3
HAWAII
Hilo
PROPOSED SEISMIC PROBABILITY
ZONES FOR HAWAII
Zone 0: No damage
Zone 1: Minor damage
Zone 2: Moderate damage
Zone 3: Major damage
Zone 0: Intensity less than 4
Zone 1: Intensity less than 6
Zone 2: Intensity less than 7
Zone 3: Intensity 7 and above
0 50 100
Scale in Miles

Thus, Maui merits a Zone 3 rating. If the relocated epicenter of the 1938 Maui earthquake were translated to the supposed epicenter of the magnitude 7.1 1871 Molokai earthquake, then Molokai and Lanai would be within an intensity VIII or higher isoseismal and would also warrant a Zone 3 rating. The major population centers of Oahu would be within an intensity VII isoseismal; thus, Oahu merits a zone 2 classification. The intensity on Kauai does not exceed intensity IV, so it appears that Kauai should not be upgraded to Zone 1; however; Furumoto (personal communication) based the upgrading for Kauai on a strong earthquake that occurred south of the island in 1967.

SUMMARY

The salient points of this thesis can be summarized as follows:

1) The HIG Maui seismic network detected low levels of seismicity beneath Haleakala and between Molokai and Lanai during the monitoring period mid-May 1980 until mid-August 1981. Lesser areas of activity were detected between Maui and the Big Island, north of Maui, and north of Molokai. The Molokai Fracture Zone appears to be seismically inactive. The strongest earthquake during the monitoring period was a magnitude 5.0 shock south of Molokai. Possible mechanisms include adjustments to loads, thermal contraction, crustal uplift, magma movement, or submarine landslides. There was no harmonic tremor or swarm activity, indicating active volcanism, during the seismic period.

2) The crustal deformation caused by earthquakes in the Maui-Molokai-Lanai region does not appear to be measurable geodetically.

3) The 1938 Maui earthquake is relocated to a point near 21.02° N. 156.09° W. Independent depth phase information places the depth at 20 km. The magnitude calculated was 6.8. The relocated epicenter is within the highest intensity isoseismal and within a zone of historical seismicity. The relocated depth is considerably less than the original estimate of 65 miles and is more in agreement with the depths of other large Hawaiian earthquakes.

4) The b-values obtained for the current Maui seismicity study and the last twenty years are 0.50 and 0.67 respectively. The b values are significantly lower than those calculated for the island of Hawaii. The low b values suggest that the Maui-Molokai-Lanai region is under greater stress than the island of Hawaii.

5) The short term earthquake recurrence rates do not reflect data for the last twenty years and are biased by the anomalously high seismic activity from the 5.0 magnitude March 5, 1981 Molokai earthquake. The long term recurrence rates seem more reasonable. A magnitude 6 earthquake could occur once every 50 years and a magnitude 7 earthquake could occur once every 250 years.

6) Regional strain is regularly released every 2 to 3 years in the form of 3 to 5 magnitude earthquakes. Strain release diagrams appear to offer a method of predicting the possible size of future earthquakes.

7) The available intensity data for large earthquakes near Maui, Molokai and Lanai indicate Maui, Molokai and Lanai should be included in seismic risk Zone 3. Oahu should be included in Zone 2.

SUGGESTIONS FOR FUTURE STUDY

1) A seismic refraction experiment using the Maui quarry blasts could improve the northern Maui crustal structure model.

2) For the present array of HIG seismometers to provide accurate hypocentral locations Haiku must be operational at all times.

3.) ARPA, the seismometer at the summit of Haleakala, should be moved because of noise problems from nearby trucks and cars.

4.) The geodetic line between Haua and Middle Hill should be resurveyed in order to verify any measurable line length change after the March 5, 1981 magnitude 5.0 Molokai earthquake.

5.) The seismic array on Maui does not make maximum use of the available seismometers. A new seismic network should be established that includes the short period seismometer operated by HVO at Haleakala National Park. Data from the new network should be shared in a cooperative effort between HIG, HVO, and the Pacific Tsunami Warning Center in order to better monitor Hawaiian seismicity.

If the seismicity were more frequent near Haleakala it would be sensible to have such a tightly spaced array as ARPA, Kula, and Haleakala National Park; however, there are not enough earthquakes to justify three seismometers only eight kilometers apart. The array should be spread apart in order to better locate earthquakes west of Maui. The two seismometers Wailuku and ARPA should be moved to Molokai and Lanai. Coverage of the area near Haleakala should continue. The seismometer at Haleakala National Park should remain at its present location. The seismometer at Kula should be moved to eastern Haleakala and the one at Haiku should remain in its present location. If possible, seismometers should be installed on West Maui and Kahoolawe. All data should be relayed by telemetry to HVO where it could be recorded and processed along with the presently recorded network.

APPENDIX I

MODIFIED MERCALLI INTENSITY SCALE, 1956 REVISED VERSION

I. Not felt. Marginal and long period effects of large earthquakes.

II. Felt by persons at rest, on upper floors, or favorably placed.

III. Felt indoors. Hanging objects swing. Vibration like passing of heavy trucks. Duration estimated. May not be recognized as an earthquake.

IV. Hanging objects swing. Vibration like passing of heavy trucks; or sensation of a jolt like a heavy ball striking the walls. Standing motor cars rock. Windows, dishes, doors rattle. Glasses clink. Crockery clashes. In the upper range of IV wooden walls and frames creak.

V. Felt outdoors; direction estimated. Sleepers wakened. Liquids disturbed, some spilled. Small unstable objects displaced or upset. Doors swing, close, open. Shutters, pictures move. Pendulum clocks start, stop, change rate.

VI. Felt by all. Many frightened and run outdoors. Persons walk unsteadily. Windows, dishes, glassware broken. Knickknacks, books, etc., off shelves. Pictures off walls. Furniture moved or overturned. Weak plaster and masonry D cracked (See below for explanation of masonry types). Small bells ring (church, school). Trees, bushes shaken (visibly or heard to rustle).

VII. Difficult to stand. Noticed by drivers of motor cars. Hanging objects quiver. Furniture broken. Damage to masonry D, including cracks. Weak chimneys broken at roof line. Fall of plaster, loose bricks, stones, tiles,cornices, (also unbraced parapets and architectural ornaments). Some cracks in masonry C. Waves on ponds; water turbid with mud. Small slides and caving in along sand or gravel banks. Large bells ring. Concrete irrigation ditches damaged.

VIII. Steering of motor cars affected. Damage to masonry C; partial collapse. Some damage to masonry B; none to masonry A. Fall of stucco and some masonry walls. Twisting, fall of chimneys, factory stacks, monuments, towers, elevated tanks. Frame houses moved on foundations if not bolted down; loose panels thrown out. Decayed piling broken off. Branches broken from trees. Change in flow or temperature of springs and wells. Cracks in wet ground or on steep slopes.

IX. General panic. Masonry D destroyed; masonry C heavily damaged, sometimes with complete collapse; masonry B seriously damaged. (General damage to foundations). Frame structures, if not bolted, shifted off foundations. Frames cracked. Serious damage to reservoirs. Underground pipes broken. Conspicuous cracks in ground. In alluviated areas sand and mud ejected. Earthquake fountains, sand craters.

X. Most masonry and frame structures destroyed with their foundations. Some well-built wooden structures and bridges destroyed. Serious damage to dams, dikes, and embankments. Large landslides. Water thrown on banks of canals, rivers, lakes, etc. Sand and mud shifted horizontally on beaches and flat land. Rails bent slightly.

XI. Rails bent greatly. Underground pipelines completely out of service.

XII. Damage nearly total. Large masses displaced. Lines of sight and level distorted. Objects thrown into the air.

Explanation of masonry types.

Masonry A. Good workmanship, mortar and design; reinforced, especially laterally, and bound together by using steel, concrete, etc.; designed to resist lateral forces.

Masonry B. Good workmanship and mortar; reinforced, but not designed to resist lateral forces.

Masonry C. Ordinary workmanship and mortar; no extreme weaknesses like failing to tie in at corners, but neither reinforced nor designed against lateral forces.

Masonry D. Weak materials, such as adobe; poor mortar; low standards of workmanship; weak horizontally.

Reference

Richter, C. F. (1958). Elementary Seismology, W. H. Freeman and Co., San Francisco and London, 768 pp.

APPENDIX II

THE ARRIVAL TIME DIFFERENCE EARTHQUAKE RELOCATION METHOD

The Arrival Time Difference Model

The Arrival Time Difference (ATD) earthquake relocation model uses a set of teleseismic P wave arrival time differences to relocate an unknown earthquake with respect to a nearby reference earthquake. It is a first term (linear) Taylor expansion of the Geiger Model of earthquake location and is analogous to the epicentral refinement method explained in Appendix X of Richter (1958) (Spence, 1980). The model relates the arrival time differences of unknown and reference events to the location parameters origin time, latitude, longitude, and depth. The use of P wave arrival time differences provides ray path redundancies for closely spaced events and thus minimizes the effects of lateral velocity heterogeneities (Spence, 1980).

Spence noted that the model's use has its limitations. Second order terms of the Taylor expansion become significantly non-zero when the epicentral distance does not greatly exceed the separation of the unknown and reference earthquakes. Spence determined that the maximum separation must be 2 $1/2^{\circ}$ or less and that the minimum distance from the epicenters of the reference and unknown earthquakes to the receivers must be 30° or greater, otherwise second order terms become significant. Further, the method requires independent depth phase information, such as pP waves, because a linear dependence develops between the origin time and depth when the epicentral distance is much greater than the separation of the two earthquakes. The dependence occurs because the partial derivative of the arrival time with respect to the depth varies extremely slowly for greater than 30^o. Finally the ATD method is unstable for a small number of observations, particularly if the observations are within a narrow azimuthal sector. If properly used, this method can yield relative locations equivalent to those achieved by a master event method, but without a suite of master residuals (Spence, 1980). Using the ATD model Spence (1980) obtained relocations of nuclear blasts in the Aleutians within 20-30 km of the blast locations.

The Relocation Model

The seismic equation of condition for the Geiger model of earthquake location is:

$$A_k = f(H, \phi, \lambda, h) \tag{3}$$

where

A_k = Arrival Time at Station k
H = Origin Time
ϕ = Geocentric Colatitude 0-180°
λ = East Longitude 0-360°
h = Depth

A Taylor expansion yields a first order guess, y, of the hypocenter:

$$A_k = f(H,\phi,\lambda,h) + \frac{\partial A_k}{\partial H}\bigg|_{\phi_y,\lambda_y,h_y} \delta H + \frac{\partial A_k}{\partial \phi}\bigg|_{H_y,\lambda_y,h_y} \delta\phi \quad (4)$$

$$+ \frac{\partial A_k}{\partial \lambda}\bigg|_{H_y,\phi_y,h_y} \delta\lambda + \frac{\partial A_k}{\partial h}\bigg|_{H_y,\phi_y,\lambda_y} \delta h + \text{Higher Order Terms}$$

By the use of spherical trigonometry, the chain rule, and a rotation of axes, equation 4 can be simplified to an equation in terms of a reference and unknown earthquake.

$$\delta_{tk} = (A_{xk} - A_{rk}) = \delta H + \frac{\partial A}{\partial \Delta} \cos Z_{rk}\, \delta\phi \quad (5)$$

$$- \frac{\partial A}{\partial \Delta} \sin\phi \sin Z_{rk}\, \delta\lambda + \frac{\partial A}{\partial h}\, \delta h$$

where

Z_{rk} = The Azimuth from the Epicenter of the Reference Earthquake to the Observing Station
δ_{tk} = The Arrival Time Difference at Station k
x = The Unknown Earthquake
r = The Reference Earthquake

$\frac{\partial A}{\partial \Delta}$, $\frac{\partial A}{\partial h}$, and Z are computed for each ray path

Spence (1980) used Jeffrey's method of uniform reduction, an iterated, weighted least squares method to solve equation 5. The solution yields the parameter corrections $\delta H, \delta\phi, \delta\lambda, \delta h$. Their addition to reference earthquake location parameters gives the relocated hypocenter and origin time. The method used to solve equation 5 is outlined in Appendix III.

References Cited

Spence, W. (1980). Relative epicenter determination using P wave arrival time differences, Bull. Seism. Soc. Am. 70, 171-183.

APPENDIX III

THE METHOD OF SINGULAR VALUE DECOMPOSITION AND NOTES ON THE COMPUTER PROGRAM 'DECOM'

The general form of the equation to be inverted is:

$$d = f(m) \tag{6}$$

where d is a vector of N observations equal to a function of a vector m composed of M parameters. The seismic equation of condition for the Geiger Model of earthquake location (Geiger, 1910) is an instance of equation 6.

The first step in solving for the parameter vector m is to linearize equation 6, using the first term of a Taylor expansion about a first guess of the parameters.

$$f(m_o) = f(m_o) + \frac{df}{dm} m_o (m-m_o) \tag{7}$$

If the first guess is reasonably close to the solution, then the following equation can be written.

$$\Delta d = G \cdot \Delta m \tag{8}$$

where $G = \frac{df}{dm} (m_o)$

$$\Delta d = d - f(m_o)$$

$$\Delta m = m - m_o$$

The arrival time difference model equation is the first term Taylor expansion of the seismic equation of condition. The matrix G consists of the ATD model coefficients.

From equation 8 a computer prgram called DECOM (Appendix IV) begins to iterate a relocation for an earthquake. DECOM uses a process called singular value decomposition (e.g. Jackson, 1972; Wiggins, 1972) to minimize the residuals of the observed and predicted arrival time differences. This process has two advantages over the least squares method: data with large errors can be given low weight and the partial derivative matrix can be manipulated so that it is never ill-conditioned.

This explanation of the program is partially based on lecture notes from a class on inverse theory at the University of Hawaii (Frazer and Lienert, 1981). The computer routine works as follows: After the data are read, the inversion routine in DECOM normalizes the observations and parameters, so as to remove their dimensionality. This step prevents extra weight being given to numerically large values. The normalization process involves dividing the observations and parameters by their standard deviations. A problem occurs because the standard deviation of the data can be estimated, but that of the parameters cannot be. This difficulty can be circumvented by assigning a fractional error to each parameter. The fractional error is corrected later when the actual parameter standard deviation is calculated. The normalization yields:

$$\Delta d^* = G^* \cdot \Delta m^* \tag{9}$$

where $\Delta d_i^* = \dfrac{\Delta d_i}{\text{std dev } (d_i)}$

$$\Delta m_j^* = \frac{\Delta m_j}{m_j \text{ [fractional error of } (m_j)]}$$

$$G_{ij}^* = \frac{df_i}{dm_j} \; \frac{m_j \text{ [fractional error } (m_j)]}{\text{std dev } (d_i)}$$

After normalization the next step is to obtain an inverse of G^*, that will yield the parameter vector Δm^*. The inverse can be determined by computing the eigenvectors of the MxM matrix $\tilde{G}^* G^*$, where $\tilde{G}^*$ is the complex conjugate transpose of G^*.

The generalized inverse of G^* is

$$H = \sum_{k=1}^{p} (1/\lambda_k) \; v_k \; \tilde{u}_k \tag{10}$$

where p is the number of eigenvectors of $\tilde{G}^* G^*$ used to construct the inverse
v_k is the kth eigenvector of $\tilde{G}^* G^*$
λ_k is the positive square root of the kth eigenvalue of $\tilde{G}^* G^*$
$\tilde{u}_k$ is the kth eigenvector of $G^* \tilde{G}^*$

The vector u_k can be calculated from v_k by the formula

$$u_k = (1/\lambda_k)\ G^* \cdot v_k \tag{11}$$

This method of obtaining the inverse of G^* can lead to a problem with the variance of the parameters, as discussed by Jackson (1972) and Wiggins (1972). The inverse H depends on the value of p. If p equals M, the problem is fully determined in which case:

$$H \cdot G = I \tag{12}$$

Multiplying equation 9 by the left-hand side of equation 12 gives:

$$H \cdot \Delta d^* = (m - m_o)^* \tag{13}$$

The parameter resolution is perfect, but the inverse may be highly unstable and the parmeter variance may be large. To see why this is so, note that if the data are statistically independent and have unit variance, then the parameter variance can be shown to be (Jackson, 1972):

$$\mathrm{var}(\Delta m_j^*) = \sum_{k=1}^{p} (v_{jk} / \lambda_k)^2 \tag{14}$$

The variance can be reduced by omitting eigenvectors with small eigenvalues from the sum in equation 10. A reasonable procedure is to successively eliminate eigenvectors, until the variance of each parameter is less than 100% of its value. Then, equation 12 no longer holds, as the off-diagonal values on the left-hand side are, in general, non-zero. This matrix $H \cdot G^*$ is called the resolution matrix.

$$R = H \cdot G^* \tag{15}$$

and equation 13 becomes

$$H \cdot \Delta d^* = R \cdot \Delta m^* \tag{16}$$

Now the parameter corrections are no longer unique. Instead they are linear combinations of each other determined by the rows of the resolution matrix. The total number of linear combinations of parameters is p , the number of eigenvectors used to create the inverse. If p is limited the resolution suffers, but the inverse is more stable, and the variance of the parameter corrections is reduced.

The parameter corrections filtered by the resolution matrix are then added to the location of the reference earthquake. If the criteria for termination are not met, the program iterates another set of corrections based on the new relocated hypocenter. Termination of the iterative process occurs when (1) the standard deviation of the relocation residuals is less than the standard deviation of the arrival time differences and (2) the magnitude of the parameter correction vector tends to zero.

References Cited

Frazer, L. N. and B. R. Lienert (1981). Introduction to inverse theory, (Lecture notes), University of Hawaii, Fall semester.

Geiger, L. (1910). Herd-bestimmung bei Erdbeben aus den Ankunftzeiten, K. Gessell. Wiss. Goett 4, 331-349, English trans. (1912). Bull. St. Louis U. 8, 60-71.

APPENDIX IV

THE COMPUTER PROGRAM ´DECOM´

```
C*********************************************************************
C                                                                    *
C                                                                    *
C                                                                    *
C  ´DECOM´ IS A PROGRAM DESIGNED TO RELOCATE EARTHQUAKES USING THE ATD *
C  RELOCATION METHOD (SPENCE, 1980, BSSA V.70, 171-183. THE MAIN     *
C  PROGRAM READS IN DATA AND CALCULATES GEOCENTRIC COLATITUDE,       *
C  DISTANCE IN DEGREES AND KILOMETERS, AND AZIMUTH.  INVERSION IS DONE *
C  BY THE SUBROUTINES ´INVERSE´ AND ´EIGEN´.  OUTPUT IS THE RELOCATED *
C  HYPOCENTER.                                                       *
C                                                                    *
C  THE INPUT VARIABLES ARE:                                          *
C                                                                    *
C                                                                    *
C  EPLAT= THE LATITUDE OF THE REFERENCE EARTHQUAKE EPICENTER         *
C  EPLON= THE LONGITUDE OF THE REFERENCE EARTHQUAKE EPICENTER        *
C  TRLAT= THE INITIAL LATITUDE OF THE UNKNOWN EARTHQUAKE EPICENTER   *
C  TRLON= THE INITIAL LONGITUDE OF THE UNKNOWN EARTHQUAKE EPICENTER  *
C  ID   = THE STANDARD THREE LETTER CODE FOR SEISMOGRAPHIC STATONS   *
C  LAT1, MIN1, SEC1 = THE LATITUDE OF THE SEISMOGRAPHIC STATION      *
C  LON1, MIN2, SEC2 = THE LONGITUDE OF THE SEISMOGRAPHIC STATION     *
C  MTX,STX = THE ARRIVAL TIME FOR THE UNKNOWN EARTHQUAKE             *
C  MTR,STR = THE ARRIVAL TIME FOR THE REFERENCE EARTHQUAKE           *
C  DADDEL  = THE PARTIAL DERIVATIVE OF THE ARRIVAL TIME WITH         *
C            RESPECT TO THE DISTANCE, DELTA                          *
C  DADH    = THE PARTIAL DERIVATIVE OF THE ARRIVAL TIME WITH         *
C            RESPECT TO THE DEPTH, H                                 *
C                                                                    *
C                                                                    *
C  INTERMEDIATE PROCESSING VARIABLES ARE:                            *
C                                                                    *
C                                                                    *
C  RLAT, RLON = THE LATITUDE AND LONGITUDE IN DEGREES OF THE         *
C               SEISMOGRAPHIC STATIONS, REFERENCE AND UNKNOWN        *
C               EARTHQUAKES                                          *
C  GCOLAT     = THE GEOCENTRIC COLATITUDE OF THE SEISMOGRAPHIC       *
C               STATIONS, REFERENCE AND UNKNOWN EARTHQUAKES          *
C  ELON       = THE EAST LONGITUDE OF THE SEISMOGRAPHIC STATIONS     *
C               REFERENCE, AND UNKNOWN EARTHQUAKES.                  *
C  AZIM       = THE AZIMUTH FOR EACH STATION WITH RESPECT TO THE     *
C               REFERENCE AND UNKNOWN EARTHQUAKES                    *
C  DIST1      = THE DISTANCE IN DEGREES BETWEEN THE UNKNOWN          *
C               EARTHQUAKE TRIAL EPICENTER AND THE SESMOGRAPHIC      *
C               STATIONS                                             *
```

```
C  DIST2       = THE DISTANCE IN KILOMETERS BETWEEN THE REFERENCE         *
C                  EARTHQUAKE AND SEISMOGRAPHIC STATION                    *
C  DIST3       = THE DISTANCE IN KILOMETERS BETWEEN THE UNKNOWN           *
C                EARTHQUAKE AND THE SEISMOGRAPHIC STATIONS                 *
C  COEF1, COEF2 = THE COEFFICIENTS FOR LATITUDE AND LONGITUDE             *
C                  PARAMETER CORRECTIONS FROM THE ATD RELOCATION           *
C                  EQUATION                                                *
C GG           = THE COEFFICIENT MATRIX THAT IS TO BE INVERTED            *
C SIGMA        = THE STANDARD DEVIATION OF THE ARRIVAL TIME DIFFERENCE    *
C                DATA                                                      *
C                                                                          *
C                                                                          *
C  THE OUTPUT VARAIBLES ARE:                                               *
C                                                                          *
C                                                                          *
C  XOR, XLAT, XLON, XDEP = THE RELOCATED EARTHQUAKE'S ORIGIN TIME         *
C                          LATITUDE, LONGITUDE, AND DEPTH.                 *
C                                                                          *
C        CHARLES W. HOLMAN     JULY 1982                                   *
C                                                                          *
C*************************************************************************
      NAME DECOM
      COMMON G(16,4),ATD(16),SDD(16),CM(4),DPRED(16),STDM(4)
      COMMON N,M,DELMX,STDMX
      DIMENSION LAT(16),MIN1(16),MIN2(16),SEC1(16),SEC2(16),
     +DADDEL(16),DADH(16),RLAT(17),RLON(17),ELON(17),GCOLAT(17),
     +AZIM(16),GLAT(17),MTX(16),MTR(16),STX(16),STR(16),COEF1(16),
     +COEF2(16),ID(17),GG(16,4),DIST2(16),DIST3(16),E(16),AZIMU(16)
      REAL LON(16)
      COMMON ID,EPLAT,EPLON,TRLAT,TRLON,RLAT,RLON,AZ,DIST,DIST1
C    READ DATA
      READ(5,30) N,M,DELMX,STDMX,NTAP,IOUT,EPLAT,EPLON,TRLAT,TRLON
   30 FORMAT (2I5,2F5.1,2I5,4F8.3)
      READ (5,31) (ID(I),LAT(I),MIN1(I),SEC1(I),LON(I),MIN2(I),SEC2(I)
     +, I=1,N)
   31 FORMAT (A5,2I5,F5.1,F5.0,I5,F5.1)
      READ (5,32) (MTX(I),STX(I),MTR(I),STR(I),DADDEL(I),DADH(I),I=1,N)
   32 FORMAT (2(I2,F5.1),2F6.2)
      READ (5,33) (CM(J),J=1,M)
   33 FORMAT (4F5.2)
C    CHANGE LATITUDE, LONGITUDE FROM DEGREES, MIN, SEC TO DEGREES
      FOR I=1,N
      IF (LAT(I) .LT. 0)
      RLAT(I)=LAT(I)-(MIN1(I)+(SEC1(I)/60.))/60.
      ELSE
      RLAT(I)=LAT(I)+(MIN1(I)+(SEC1(I)/60.))/60.
      END IF
      IF (LON(I) .LT. 0.)
      RLON(I)=LON(I)-(MIN2(I)+(SEC2(I)/60.))/60.
      ELSE
```

```
      RLON(I)=LON(I)+(MIN2(I)+(SEC2(I)/60.))/60.
      END IF
      END FOR
      RLAT(17)=EPLAT
      RLON(17)=EPLON
      L=N+1
      PI=3.14159265
      R=57.2957795
C     COMPUTE GEOCENTRIC COLATITUDE
      FOR I=1,L
      GLAT(I)=ATAN(TAN(RLAT(I)/R)*0.99328)
      IF (RLAT(I) .LT. 0.)
      GCOLAT(I)=(PI/2. + ABS(GLAT(I)))*R
      ELSE
      GCOLAT(I)=(PI/2. - GLAT(I))*R
      END IF
      IF (RLON(I) .GT. 0.)
      ELON(I)=RLON(I)
      ELSE
      ELON(I)=360. - ABS(RLON(I))
      END IF
      END FOR
C CALCULATE ARRIVAL TIME DIFFERENCES
      FOR I=1,N
      ATD(I)=0.
      END FOR
      FOR I=1,N
      IF (MTX(I) .LT. MTR(I)) STR(I)=STR(I)+60.
      IF (MTX(I) .GT. MTR(I)) STX(I)=STX(I)+60.
      ATD(I)=STX(I)-STR(I)
      END FOR
      FOR I=1,N
      IDEN=ID(I)
      HYPLAT=EPLAT
      HYPLON=EPLON
      REFLAT=RLAT(I)
      REFLON=RLON(I)
C    CALCULATE AZIMUTH, DISTANCE IN DEGREES AND KILOMETERS FOR
C    REFERENCE EARTHQUAKE AND SEISMOGRAPHIC STATIONS
      CALL DISTAZ (IDEN,HYPLAT,HYPLON,REFLAT,REFLON,DIST,DIST1,AZ)
      DIST3(I)=DIST
      AZIM(I)=AZ
      END FOR
      FOR I=1,N
      IDEN=ID(I)
      HYPLAT=TRLAT
      HYPLON=TRLON
      REFLAT=RLAT(I)
      REFLON=RLON(I)
C     CALCULATE AZIMUTH, DISTANCE IN DEREES AND KILOMETERS FOR
```

```
C     THE UNKNOWN EARTHQUAKE AND SEISMOGRAPHIC STATIONS
      CALL DISTAZ (IDEN,HYPLAT,HYPLON,REFLAT,REFLON,DIST,DIST1,AZ)
      DIST2(I)=DIST
      AZIMU(I)=AZ
C     CALCULATE STATION CORRECTION
      E(I)=(DIST2(I)-DIST3(I))/9.5
      IF(DIST1 .GE. 50.) E(I)=(DIST2(I)-DIST3(I))/11.5
      IF(DIST1 .GE. 70.) E(I)=(DIST2(I)-DIST3(I))/12.5
      END FOR
C     CALCULATE COEFFICIENTS FOR MATRIX GG
      FOR I=1,N
      COEF1(I)=DADDEL(I)*COS(AZIM(I)/R)
      COEF2(I)=-(DADDEL(I)*SIN(GCOLAT(I)/R)*SIN(AZIM(I)/R))
      GG(I,1)=1.
      GG(I,2)=COEF1(I)
      GG(I,3)=COEF2(I)
      GG(I,4)=DADH(I)
      END FOR
      SUM=0.
C STANDARD DEVIATION OF DATA
      SQ=0.
      FOR I=1,N
      SQ=SQ+(ATD(I)-E(I))**2
      END FOR
      SIGMA=SQRT(SQ/N)
C CALCULATE PREDICTED VALUES FOR ARRIVAL TIME DIFFERENCES
      K=0
   39 K=K+1
      XOR=0.
      XDEP=5.
      ALAT=GCOLAT(17)
      ALON=ELON(17)
      FOR I=1,N
      SDD(I)=2.
      DPRED(I)=0.
      FOR J=1,M
      G(I,J)=GG(I,J)
      END FOR
      END FOR
      FOR I=1,N
      FOR J=1,M
      DPRED(I)=DPRED(I)+G(I,J)*CM(J)
      END FOR
      END FOR
C     INVERT MATRIX GG AND CALCULATE NEW CORRECTION PARAMETERS
      CALL INVERT(G,ATD,SDD,DPRED,CM,STDM,SIGMA,N,M,DELMX,STDMX,
     +SIGMAE,SIGMAD,DELMM,NTAP,IOUT)
C     CALCULATE RELOCATED HYPOCENTER
      XOR=XOR+CM(1)
      ALAT=ALAT+CM(2)
```

```
      ALON=ALON+CM(3)
      XDEP=XDEP+CM(4)
      XLON=ABS(ALON-360.)
      XLAT=ATAN(TAN(-ALAT/R+PI/2.)/.99328)*R
      WRITE(6,119) XOR,XLAT,XLON,XDEP
  119 FORMAT('0','ORIGIN TIME W.R.T. 08:32:47 IS',F9.2,'SEC',//,
     +'LOCATION OF EARTHQUAKE  LATITUDE',F9.2,1X,'DEGREES N.',1X,
     +'LONGITUDE',F9.2,1X,'DEGREES W.',/,
     +'DEPTH OF EARTHQUAKE'F9.2,1X,'KM',/)
C     CONDITIONS FOR TERMINATION OF THE ROUTINE
      IF (SIGMAD .LT. SIGMA .AND. DELMM .LT. 0.01) STOP
      IF (K .LT. 10) GO TO 39
      STOP
      END
C***********************************************************************
C                                                                      *
C  SUBROUTINE TO INVERT A PARTIAL DERIVATIVE MATRIX,G(I,J)             *
C  OF N OBSERVATIONS,D(I), W.R.T. M PARAMETERS, CM(J).                 *
C  THE SUBROUTINE WILL PERFORM ONE ITERATION OF A LINEARIZED           *
C  INVERSION, RETURNING THE CORRECTED PARAMETERS IN CM(J).             *
C                                                                      *
C  INPUT PARAMETERS ARE:                                               *
C      G(I,J)   - NXM PARTIAL DERIVATIVE MATRIX                        *
C      N        - NO OF OBSERVATIONS (MAX=100)                         *
C      M        - NO OF PARAMETERS (MAX=50)                            *
C      D(I)     - INPUT DATA                                           *
C      SDD(I)   - INPUT DATA ERRORS                                    *
C      CM(J)    - PARAMETERS (1ST GUESS)                               *
C      SIGMA    - PROBLEM STD DEVIATION                                *
C      DELMX    - MAX FRACTIONAL PARAM. CHANGE                         *
C      STDMX    - MAX ALLOWABLE PARAM. STD DEVIATION                   *
C      DPRED(I)- PREDICTED VALUES OF DATA FOR CM(J)                    *
C      NTAP     =0...SHARP CUTOFF                                      *
C               =1...TAPERED CUTOFF                                    *
C      IOUT     =0...NO OUTPUT                                         *
C               =1...OUTPUT ON UNIT 6                                  *
C               =2...RESOLUTION MATRIX                                 *
C               =3...   "          "      PLUS INFORMATION MATRIX      *
C                                                                      *
C  OUTPUT PARAMETERS                                                   *
C      CM(J)    - CORRECTED PARAM. VALUES                              *
C      STDM(J) - PARAM. STD DEVIATIONS                                 *
C      SIGMAE  - ERROR OF FIT                                          *
C      SIGMAD  - DATA MISFIT                                           *
C      DELMM   - MAGNITUDE OF PARAM. CHANGE VECTOR                     *
C                                                                      *
C  USE SIGMAD<SIGMA AND DELMM<<1 AS CONVERGENCE CRITERIA               *
C                                                                      *
C                                                                      *
C         BARRY R. LIENERT   SEPTEMBER 1981                            *
```

```
C                                                                       *
C************************************************************************
C
      SUBROUTINE INVERT(G,D,STDD,DPRED,CMM,STDM,SIGMA,N,M,DELMX,
     1STDMX,SIGMAE,SIGMAD,DELMM,NTAP,IOUT)
C
C  G MUST BE DIMENSIONED THE SAME AS IN THE CALLING PRORAM
C
      DIMENSION G(16,4),D(1),CMM(4),STDD(1),STDM(1),DPRED(1)
C
C THESE DIMENSIONS ARE FOR NMAX=100 AND MMAX=50
C
      COMMON DELD(100),AA(5000),EIG(50),V(2500),U(10000),
     1BB(100,100),DELP(50),DELM(50)
      INTEGER PC,PCMX
C
C   NORMALIZE G
C
      DO 1 I=1,N
      DO 1 J=1,M
1     G(I,J)=G(I,J)*CMM(J)*SIGMA/STDD(I)
C
C   DIFERENCES BETWEEN OBSERVED AND PRED. DATA
C
      DO 2 I=1,N
2     DELD(I)=(D(I)-DPRED(I))/STDD(I)
      IF(IOUT.GE.0)WRITE(6,130)
130   FORMAT(//,10X,"DATA DIFFERENCES")
      IF(IOUT.GE.0)WRITE(6,110)(DELD(I),I=1,N)
C
C    FORM AA = GT * G
C
C
      IC=1
      DO 8 J=1,M
      DO 8 K=1,J
      SUM=0.0
      DO 7 KK=1,N
7     SUM=SUM+G(KK,J)*G(KK,K)
      AA(IC)=SUM
8     IC=IC+1
C
C  GET EIGENVECTORS V, AND EIGENVALUES (STORED IN DIAGONAL OF AA)
C
      CALL EIGEN(AA,V,M,0)
      II=1
      DO 6 I=1,M
      EIG(I)=SQRT(ABS(AA(II)))
6     II=II+I+1
      IF(IOUT.GE.1)WRITE(6,150)(EIG(I),I=1,M)
```

```
150   FORMAT(//,10X,'EIGENVALUES: ',/,10F12.4)
      II=1
      IF(IOUT.LT.2)GO TO 88
      WRITE(6,185)
185   FORMAT(//,10X,'EIGENVECTORS')
      DO 87 I=1,M
      I1=(I-1)*M+1
      I2=I*M
87    WRITE(6,110)(V(J),J=I1,I2)
88    CONTINUE
C
C
C   DETERMINE MAX DEGS FREEDOM, PC
C
      VAR=(STDMX/SIGMA)**2
      PC=M
      DO 15 I=1,M
      SUM=0.0
      DO 9 J=1,M
      II=(J-1)*M+I
      IF(EIG(J).EQ.0.)GO TO 10
      SUM=SUM+(V(II)/EIG(J))**2
      IF(SUM.GE.VAR)GO TO 10
9     CONTINUE
10    IF(PC.GT.J-1)PC=J-1
      PCMX=J-1
15    CONTINUE
      IF(IOUT.GE.1)WRITE(6,112)PC
112   FORMAT(//,10X,'DEGREES OF FREEDOM = ',I5)
      IF(NTAP.EQ.0)GO TO 226
C
C  CALCULATE EQUIVALENT TAPERED CUTOFF CONSTANT,XK
C
      XKMAX=FLOAT(PC)
      DO 220 K=1,1000
      XK=FLOAT(K)*0.001
      SUM=0.0
      DO 218 J=1,PCMX
      XX=EIG(J)**2
218   SUM=SUM+XX/(XX+XK)
      IF(SUM.LE.XKMAX)GO TO 221
220   CONTINUE
221   WRITE(6,1801)XK
1801  FORMAT(" TAPERED CUTOFF CONSTANT = ",F12.4)
      IF(IOUT.GE.1)WRITE(6,1802)
1802  FORMAT(" **** TAPERED CUTOFF ****")
      PC=PCMX
226   CONTINUE
C
C  CALCULATE PARAM. STD DEVIATIONS
```

```
C
      DO 77 I=1,M
      SUM=0.0
      DO 77 J=1,PC
      II=(J-1)*M+I
      XX=EIG(J)
      IF(NTAP.EQ.1)XX=EIG(J)+XK/EIG(J)
      SUM=SUM+(V(II)/XX)**2
77    STDM(I)=SQRT(SUM)*SIGMA
      IF(IOUT.GE.1)WRITE(6,2000)
2000  FORMAT(//,10X,'PARAMETER VARIANCES')
110   FORMAT(1H ,10F8.2)
      IF(IOUT.GE.1)WRITE(6,110)(STDM(I),I=1,M)
1800  FORMAT(//,10X,'RESOLUTION MATRIX')
C
C  RESOLUTION MATRIX = V * VT
C
      DO 70 I=1,M
      DO 70 J=1,M
      SUM=0.0
      DO 61 K=1,PC
      II=(K-1)*M+I
      JJ=(K-1)*M+J
      XX=1.0
      IF(NTAP.EQ.1)XX=1.+XK/(EIG(K)**2)
61    SUM=SUM+V(II)*V(JJ)/XX
70    BB(I,J)=SUM
      IF(IOUT.LT.2)GO TO 63
      WRITE(6,1800)
      DO 62 I=1,M
62    WRITE(6,110)(BB(I,J),J=1,M)
63    CONTINUE
C
C  CALCULATE EIGENVECTORS U(I,J) = V(L,K) * G(I,K) / EIG(L)
C
      DO 89 I=1,PC
      DO 89 KK=1,N
      SUM=0.0
      DO 79 K=1,M
      II=(I-1)*M+K
79    SUM=SUM+V(II)*G(KK,K)
      JJ=(I-1)*N+KK
89    U(JJ)=SUM/EIG(I)
      IF(IOUT.LT.3)GO TO 83
      WRITE(6,1900)
1900  FORMAT(//,10X,'INFORMATION MATRIX')
C
C  INFORMATION MATRIX = U * UT
C
      DO 80 I=1,N
```

```
      DO 80 J=1,N
      SUM=0.0
      DO 81 K=1,PC
      II=(K-1)*N+I
      JJ=(K-1)*N+J
81    SUM=SUM+U(II)*U(JJ)
80    BB(I,J)=SUM
      DO 82 I=1,N
82    WRITE(6,110)(BB(I,J),J=1,N)
83    CONTINUE
14    CONTINUE
C
C   CALCULATE DELP(I) = U * DELD / EIG(I) I=1..PC    (WIGGINS,1972)
C
      DO 20 I=1,PC
      SUM=0.0
      DO 16 K=1,N
      II=(I-1)*N+K
16    SUM=SUM+U(II)*DELD(K)
      XX=EIG(I)
      IF(NTAP.EQ.1)XX=EIG(I)+XK/EIG(I)
20    DELP(I)=SUM/XX
      SUM=0.0
      DO 21 I=1,PC
21    SUM=SUM+DELP(I)*DELP(I)
C
C    CALCULATE PARAMETER CORRECTIONS DELM = VT * DELP
C
      DO 40 I=1,M
      SUM=0.0
      DO 31 J=1,PC
      II=(J-1)*M+I
31    SUM=SUM+DELP(J)*V(II)
C
C    MAX PARAM. CHANGE LIMITED TO DELMX
C
      IF(ABS(SUM).GT.DELMX)SUM=DELMX*SUM/ABS(SUM)
40    DELM(I)=SUM
      IF(IOUT.GT.0)WRITE(6,140)
140   FORMAT(//,10X,'PARAMETER CORRECTIONS')
      IF(IOUT.GT.0)WRITE(6,110)(DELM(I),I=1,M)
C
C   CALCULATE ERROR OF FIT, SIGMAE AND MISFIT, SIGMAD
C
      SUM1=0.0
      SUM2=0.0
      DO 50 I=1,N
      SUM=0.0
      DO 41 J=1,M
41    SUM=SUM+G(I,J)*DELM(J)
```

```
      SUM=SUM-DELD(I)
      SUM1=SUM1+SUM*SUM
50    SUM2=SUM2+DELD(I)*DELD(I)
      SIGMAE=SQRT(SUM1/FLOAT(N-PC))
      SIGMAD=SQRT(SUM2/FLOAT(N-PC))
      DO 60 I=1,M
      WRITE(6,3002) CMM(I)
      CMM(I)=CMM(I)*(1.+DELM(I)*SIGMA)
      WRITE(6,3002) CMM(I)
   60 CONTINUE
 3002 FORMAT('0','CM=',F10.3)
C
C  CALCULATE MAGNITUDE OF PARAM. CHANGE VECTOR, DELMM
C
      SUM=0.0
      DO 66 J=1,M
66    SUM=SUM+DELM(J)*DELM(J)
      DELMM=SQRT(SUM)
      IF(IOUT.GE.1)WRITE(6,2200)SIGMAD,SIGMAE,SIGMA,DELMM
2200  FORMAT(1H ,/,"SIGMAD = ",F8.4," SIGMAE = ",F8.4,
     1" SIGMA = ",F8.4," DELMM = ",F8.4,/)
      RETURN
      END
C***********************************************************************
C                                                                      *
C        SUBROUTINE EIGEN                                              *
C                                                                      *
C        PURPOSE                                                       *
C           COMPUTE EIGENVALUES AND EIGENVECTORS OF A REAL SYMMETRIC   *
C           MATRIX                                                     *
C                                                                      *
C        USAGE                                                         *
C           CALL EIGEN(A,R,N,MV)                                       *
C                                                                      *
C        DESCRIPTION OF PARAMETERS                                     *
C           A - ORIGINAL MATRIX (SYMMETRIC), DESTROYED IN COMPUTATION. *
C               RESULTANT EIGENVALUES ARE DEVELOPED IN DIAGONAL OF     *
C               MATRIX A IN DESCENDING ORDER.                          *
C           R - RESULTANT MATRIX OF EIGENVECTORS (STORED COLUMNWISE,   *
C               IN SAME SEQUENCE AS EIGENVALUES)                       *
C           N - ORDER OF MATRICES A AND R                              *
C           MV- INPUT CODE                                             *
C                   0   COMPUTE EIGENVALUES AND EIGENVECTORS           *
C                       DIMENSIONED BUT MUST STILL APPEAR IN CALLING   *
C                       SEQUENCE)                                      *
C                                                                      *
C        REMARKS                                                       *
C           ORIGINAL MATRIX A MUST BE REAL SYMMETRIC (STORAGE MODE=1)  *
C           MATRIX A CANNOT BE IN THE SAME LOCATION AS MATRIX R        *
C                                                                      *
```

```
C        SUBROUTINES AND FUNCTION SUBPROGRAMS REQUIRED                 *
C           NONE                                                        *
C                                                                       *
C        METHOD                                                         *
C           DIAGONALIZATION METHOD ORIGINATED BY JACOBI AND ADAPTED BY *
C           VON NEUMANN FOR LARGE COMPUTERS AS FOUND IN 'MATHEMATICAL  *
C           METHODS FOR DIGITAL COMPUTERS', EDITED BY A. RALSTON AND   *
C           H.S. WILF, JOHN WILEY AND SONS, NEW YORK, 1962, CHAPTER 7  *
C                                                                       *
C.......................................................................*
C                                                                       *
      SUBROUTINE EIGEN(A,R,N,MV)
      DIMENSION A(1),R(1)
C                                                                       *
C.......................................................................*
C                                                                       *
C        IF A DOUBLE PRECISION VERSION OF THIS ROUTINE IS DESIRED, THE *
C        C IN COLUMN 1 SHOULD BE REMOVED FROM THE DOUBLE PRECISION     *
C        STATEMENT WHICH FOLLOWS.                                       *
C                                                                       *
C     DOUBLE PRECISION A,R,ANORM,ANRMX,THR,X,Y,SINX,SINX2,COSX,         *
C    1                 COSX2,SINCS,RANGE                                *
C                                                                       *
C        THE C MUST ALSO BE REMOVED FROM DOUBLE PRECISION STATEMENTS   *
C        APPEARING IN OTHER ROUTINES USED IN CONJUNCTION WITH THIS     *
C        ROUTINE.                                                       *
C                                                                       *
C        THE DOUBLE PRECISION VERSION OF THIS SUBROUTINE MUST ALSO     *
C        CONTAIN DOUBLE PRECISION FORTRAN FUNCTIONS. SQRT IN STATEMENTS*
C        40, 68, 75, AND 78 MUST BE CHANGED TO DSQRT.  ABS IN STATEMENT*
C        62 MUST BE CHANGED TO DABS. THE CONSTANT IN STATEMENT 5 SHOULD*
C        BE CHANGED TO 1.0D-12.                                         *
C                                                                       *
C        BARRY R. LIENERT   SEPTEMBER 1981                              *
C                                                                       *
C************************************************************************
C
C        GENERATE IDENTITY MATRIX
C
    5 RANGE=1.0E-7
      IF(MV-1) 10,25,10
   10 IQ=-N
      DO 20 J=1,N
      IQ=IQ+N
      DO 20 I=1,N
      IJ=IQ+I
      R(IJ)=0.0
      IF(I-J) 20,15,20
   15 R(IJ)=1.0
   20 CONTINUE
```

```
   15 R(IJ)=1.0
   20 CONTINUE
C
C        COMPUTE INITIAL AND FINAL NORMS (ANORM AND ANORMX)
C
   25 ANORM=0.0
      DO 35 I=1,N
      DO 35 J=I,N
      IF(I-J) 30,35,30
   30 IA=I+(J*J-J)/2
      ANORM=ANORM+A(IA)*A(IA)
   35 CONTINUE
      IF(ANORM) 165,165,40
40    ANORM=1.414*SQRT(ANORM)
      ANRMX=ANORM*RANGE/FLOAT(N)
C
C        INITIALIZE INDICATORS AND COMPUTE THRESHOLD, THR
C
      IND=0
      THR=ANORM
   45 THR=THR/FLOAT(N)
   50 L=1
   55 M=L+1
C
C        COMPUTE SIN AND COS
C
   60 MQ=(M*M-M)/2
      LQ=(L*L-L)/2
      LM=L+MQ
   62 IF( ABS(A(LM))-THR) 130,65,65
   65 IND=1
      LL=L+LQ
      MM=M+MQ
      X=0.5*(A(LL)-A(MM))
   68 Y=-A(LM)/ SQRT(A(LM)*A(LM)+X*X)
      IF(X) 70,75,75
   70 Y=-Y
   75 SINX=Y/ SQRT(2.0*(1.0+( SQRT(1.0-Y*Y))))
      SINX2=SINX*SINX
   78 COSX= SQRT(1.0-SINX2)
      COSX2=COSX*COSX
      SINCS =SINX*COSX
C
C        ROTATE L AND M COLUMNS
C
      ILQ=N*(L-1)
      IMQ=N*(M-1)
      DO 125 I=1,N
      IQ=(I*I-I)/2
      IF(I-L) 80,115,80
```

```
   80 IF(I-M) 85,115,90
   85 IM=I+MQ
      GO TO 95
   90 IM=M+IQ
   95 IF(I-L) 100,105,105
  100 IL=I+LQ
      GO TO 110
  105 IL=L+IQ
  110 X=A(IL)*COSX-A(IM)*SINX
      A(IM)=A(IL)*SINX+A(IM)*COSX
      A(IL)=X
  115 IF(MV-1) 120,125,120
  120 ILR=ILQ+I
      IMR=IMQ+I
      X=R(ILR)*COSX-R(IMR)*SINX
      R(IMR)=R(ILR)*SINX+R(IMR)*COSX
      R(ILR)=X
  125 CONTINUE
      X=2.0*A(LM)*SINCS
      Y=A(LL)*COSX2+A(MM)*SINX2-X
      X=A(LL)*SINX2+A(MM)*COSX2+X
      A(LM)=(A(LL)-A(MM))*SINCS+A(LM)*(COSX2-SINX2)
      A(LL)=Y
      A(MM)=X
C
C        TESTS FOR COMPLETION
C
C        TEST FOR M = LAST COLUMN
C
  130 IF(M-N) 135,140,135
  135 M=M+1
      GO TO 60
C
C        TEST FOR L = SECOND FROM LAST COLUMN
C
  140 IF(L-(N-1)) 145,150,145
  145 L=L+1
      GO TO 55
  150 IF(IND-1) 160,155,160
  155 IND=0
      GO TO 50
C
C        COMPARE THRESHOLD WITH FINAL NORM
C
  160 IF(THR-ANRMX) 165,165,45
C
C        SORT EIGENVALUES AND EIGENVECTORS
C
  165 IQ=-N
      DO 185 I=1,N
```

```
      IQ=IQ+N
      LL=I+(I*I-I)/2
      JQ=N*(I-2)
      DO 185 J=I,N
      JQ=JQ+N
      MM=J+(J*J-J)/2
      IF(A(LL)-A(MM)) 170,185,185
  170 X=A(LL)
      A(LL)=A(MM)
      A(MM)=X
      IF(MV-1) 175,185,175
  175 DO 180 K=1,N
      ILR=IQ+K
      IMR=JQ+K
      X=R(ILR)
      R(ILR)=R(IMR)
  180 R(IMR)=X
  185 CONTINUE
      RETURN
      END
C************************************************************************
C                                                                       *
C     'DISTAZ'  COMPUTES THE DISTANCE AND AZIMUTH BETWEEN ANY TWO       *
C     POINTS ON THE GLOBE.  THE FORMULAS ARE TAKEN FROM                 *
C     THOMAS, P. D. (1966). GEODESIC ARC LENGTH ON THE REFERENCE        *
C     ELLIPSOID TO SECOND ORDER TERMS IN THE FLATTENING, JGR            *
C     70, 3331-3340.                                                    *
C     THE COMPUTATIONS INCORPORATE ALL FIRST AND SECOND ELLIPTICITY     *
C     CORRECTIONS.                                                      *
C                                                                       *
C     GEOGRAPHIC LATITUDES ARE ENTERED IN DEGREES; '+' FOR NORTH AND    *
C     '-' FOR SOUTH.  GEOGRAPHIC LONGITUDES ARE ENTERED IN DEGREES;     *
C     '+' FOR EAST LONGITUDE AND '-' FOR WEST LONGITUDE                 *
C                                                                       *
C     JOSEPH GETTRUST          AUGUST 1978                              *
C                                                                       *
C************************************************************************
      REAL LAT1,LON1,LAT2,LON2
      R=57.2957795
      LAT1=LAT1/R
      LAT2=LAT2/R
      LON1=LON1/R
      LON2=LON2/R
   50 CALL DISAZ(LAT1,LON1,LAT2,LON2,AZ,BB,DIST)
      DIST1=DIST/111.199
      AZ=AZ*R
      BB=BB*R
      RETURN
      END
      SUBROUTINE DISAZ (FLAT1,FLON1,FLAT2,FLON2,AZ,BB,S1)
```

```
C***********************************************************************
C      'DISAZ' COMPUTES DISTANCE AND AZIMUTH.                            *
C      THE COMPUTATIONS INCORPORATE                                      *
C      ALL FIRST AND SECOND ELLIPTICITY CORRECTIONS.                     *
C                                                                        *
C      GEOGRAPHIC LATITUDES ARE ENTERED IN RADIANS. '+' FOR NORTH AND    *
C      '-' FOR SOUTH.                                                    *
C      LONGITUDES ARE ENTERED AS RADIANS. '+' FOR EAST LONG.             *
C      '-' FOR WEST LONG.                                                *
C                                                                        *
C      JOSEPH GETTRUST    AUGUST 1978                                    *
C                                                                        *
C***********************************************************************
      REAL LAT1,LON1,LAT2,LON2
      FLAT=0.0033900753
      R=57.2957795
      A=6378.2064
      PI=3.14159265
      LAT1=FLAT1
      LAT2=FLAT2
      IF (FLON1) 5,5,10
    5 LON1=-FLON1
      GO TO 20
   10 LON1=2.*PI-FLON1
   20 IF (FLON2) 25,25,30
   25 LON2=-FLON2
      GO TO 40
   30 LON2=2.*PI-FLON2
   40 NCON=0
      IF(LON1-LON2) 35,35,31
   31 SAVE=LON1
      SAVE1=LAT1
      LON1=LON2
      LAT1=LAT2
      LON2=SAVE
      LAT2=SAVE1
      NCON=1
   35 IF(LON2-LON1-PI) 50,50,36
   36 SAVE=LON1
      SAVE1=LAT1
      LON1=LON2
      LAT1=LAT2
      LON2=SAVE+2.*PI
      LAT2=SAVE1
      IF(NCON-1) 37,38,38
   37 NCON=1
      GO TO 50
   38 NCON=0
   50 PHIM=.5*(LAT1+LAT2)
      DPHIM=0.5*(LAT2-LAT1)
```

```
      DL=(LON2-LON1)
      DLM=0.5*DL
      CDPHI=COS(DPHIM)
      SDPHI=SIN(DPHIM)
      CDLM=COS(DLM)
      SK=SIN(PHIM)*COS(DPHIM)
      BK=SIN(DPHIM)*COS(PHIM)
      BH=COS(DPHIM)**2-SIN(PHIM)**2
      BL=SIN(DPHIM)**2+BH*SIN(DLM)**2
      CDEL=1.-2.*BL
      DEL=ACOS(CDEL)
      T=DEL/SIN(DEL)
      U=2.*SK*SK/(1.-BL)
      V=2.*BK*BK/BL
      BE=60.*COS(DEL)
      X=U+V
      Y=U-V
      BD=8.*(6.+T*T)
      BA=4.*T+(16.+BE*T/15.)
      BB=2.*BD
      BC=2.*T-.5*(BA+BE)
      DELTA=-(FLAT/4.)*(T*X-3.*Y)
C
C     DISTANCE GIVEN IN KILOMETERS.
C
      S1=A*SIN(DEL)*(T+DELTA+(FLAT**2/128.)*(BA*X+BB*Y+BC*X**2+BD*X*Y+
     1BE*Y**2))
C
C     AZIMUTH GIVEN AS EAST OF NORTH.  NOTE THAT THOMAS GIVES AZIMUTH
C     AS WEST OF SOUTH.
C
      SINDL=SIN(DL)
      DA2PD=-.5*FLAT*BH*(T+1.)*(BK*SINDL/BL)
      DA2MD=-.5*FLAT*BH*(T-1.)*BK*SINDL/(1.-BL)
      CD2=COS(DEL/2.)
      ARG1=SDPHI*CDLM/SIN(DEL/2.)
      ARG2=COS(PI/2.-PHIM)*SIN(DLM)/CD2
      A2PA1=ASIN(ARG1)+3.14159265
      A2MA1=3.14159265-ACOS(ARG2)
      A12=A2PA1+A2MA1+DA2PD-DA2MD
      A21=A2PA1-A2MA1+DA2PD+DA2MD
      IF(NCON-1) 61,60,60
   60 AZ=A21
      BB=A12
      GO TO 70
   61 AZ=A12
      BB=A21
   70 RETURN
      END$
```

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