The 1973 Hawaii earthquake: a double earthquake beneath the volcano Mauna Kea

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Summary. The 1973 Hawaii earthquake occurred north of Hilo, at a depth of 40 to 50 km. The location was beneath the east flank of Mauna Kea, a volcano dormant historically, but active within the last 4000 yr. Aftershocks were restricted to a depth of 35–35 km. The event and its aftershock sequence are located in an area not normally associated with the seismicity of the Mauna Loa and Kilauea calderas. The earthquake was a double event, the epicentres trending NE–SW. The events were of similar size and faulting mechanism. The fault plane solutions obtained by seismic waveform analysis are a strike-slip fault striking EW and dipping $55^\circ$ S, the auxiliary plane a NS vertical plane with a faulting plunge of $35^\circ$. The axis of maximum compressive stress is aligned with the direction of the gravity gradient associated with the island of Hawaii. The fault plane striking EW parallels a surface feature, the Mauna Kea east rift zone. The earthquakes were clearly not associated with volcanic activity normally associated with Mauna Loa and Kilauea and may indicate a deep seated prelude to a resumption of activity at Mauna Kea.

In the beginning

At the leading edge of a linear chain of volcanoes extending from Midway Island lies the island of Hawaii, largest of the Hawaiian Islands. The island of Hawaii is composed of five shield volcanoes: Mauna Loa, Kilauea, and Hualalai have erupted in historic time, Mauna Kea and Kohala have not. On 1973 April 26 at 20:26:30.6 GMT a magnitude ($M_3$) 6.2 earthquake occurred 15 km north of Hilo, Hawaii, at a focal depth of 48 km (Unger & Ward 1979). The earthquake was located at 19.90° N, 155.13° W beneath the east rift zone of Mauna Kea. This event is the largest known subcrustal earthquake in the central Pacific and the second largest event in the Hawaiian Islands since 1951 (Furumoto, Nielsen & Phillips 1973). The earthquake and aftershock locations from the local Hawaii array (Unger & Ward 1979) are shown in Fig. 1; the location of a second shock occurring with the mainshock was determined from an analysis of seismic records from WWSSN and Canadian stations. The events lie beneath the area of the east rift zone of the volcano Mauna Kea. Concurrent
increases in other seismicity were noted in two areas noted for sporadic seismic activity: the upper crustal regions beneath the east slope of Mauna Kea volcano and the summit caldera region of the very active Kilauea volcano (Unger \& Ward 1979).

**The focal mechanism**

To obtain a faulting mechanism for the 1973 Hawaii earthquake, an analysis was undertaken of the waveforms of compressional \((P)\) phases and shear \((S)\) phases from the event recorded on long-period WWSSN and Canadian seismometers. \(P\)-waves at 33 stations and \(S\)-waves at 29 stations were collected, twice digitized, and averaged to minimize digitization errors. \(S\)-waves were rotated into radial \((SV)\) and transverse \((SH)\) components of motion; only the \(SH\) component was used in the analysis of the shear wave data. The \(P\)-wave data are typically of high quality. The \(SH\) data vary in quality. \(SH\) data for azimuths east of the event to North and South America are of high quality and relatively low noise, \(SH\) data to the south-west are relatively poorer in quality. \(SH\)-waves travelling to the south-west pass or reflect beneath the island of Hawaii. Lateral heterogeneity beneath the island probably complicate the \(SH\) waveforms.

In modelling the Hawaii earthquake, synthetic seismograms were generated for a point shear dislocation in a layered elastic medium (Langston \& Helmberger 1975). The effect of body wave attenuation was generated by convolving the synthetic seismogram with a Futterman (1962) attenuation operator with \(T/Q\) (the ratio of seismic travel-time to path average \(Q\)) = 1 for \(P\)-waves and \(T/Q = 4\) for \(S\)-waves. The observed \(P\) and \(SH\) waveforms were
modelled by a combination of trial and error searching and formal body waveform inversion (Langston 1976; Burdick & Mellman 1976). The trial and error searching obtained appropriate zeroth-order starting models satisfying the first motions of direct and surface reflected energy; the starting models were then used in a formal body waveform inversion procedure. Only seismic data at teleseismic distances greater than 30° geocentric angle were used in the analysis. The crust-mantle velocity model in the study was adapted from Hill (1969) and Crosson (1976) and is shown in Table 1. The shear velocity was initially at Poisson's ratio of $\alpha/\beta$ of 1.73, but a check of the relative timing of the surface reflected phases $pP$, $sP$ and $sS$ suggested a ratio of $\alpha/\beta$ of 1.84.

The first motions of $P$- and $SH$-waves from the 1973 Hawaii earthquake are plotted in Fig. 2 on an equal area projection of the lower focal hemisphere of the event. Selected waveforms of $SH$ and $P$ are shown in Figs 3 and 4, respectively. For the $P$-wave data in Fig. 4 a secondary arrival consistently having the same direction of motion as the first motion is observed. The relative timing between the initial and second $P$-wave arrivals varies in a

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Table 1. Seismic velocity model.

Figure 2. The $P$ and $SH$ first motions recorded at seismograph stations are shown on an equal-area projection of a lower hemisphere centred at the earthquake source. The first motion data (compression or dilation for $P$-waves, clockwise or counter-clockwise about the earthquake source for $SH$-waves) are plotted relative to the azimuth and take-off angle of the $P$- and $SH$-waves from the source region. The composite symbol indicates different $SH$ first motions for the main shock and a second coseismic found in the analysis. Two possible fault plane solutions, 1S and 1P, are determined for the main shock. The fault plane solution of the second coseismic, labelled 2, is shown. Seismic stations are denoted by their three letter code names.
Figure 3. The upper trace to the right of the station code plots observed transverse (SH) component of the shear wave. The lower thin trace plots synthetic SH-wave for the seismic station using Hawaii source models 1S and 2. The geocentric angle $\Delta$ and azimuth $Az$ of the station measured relative to the earthquake clockwise from north is indicated as per station KBS. The seismic moment $M_0$ is calculated from models 1S and 2, and is in units of $10^{23}$ dyne cm. Note S motion at stations KBS, Kings Bay, Spitsbergen; BLA, Blacksburg, Virginia; SHK, Shiraki, Japan; MAT, Matsushiro, Japan.

Figure 4. The upper trace to the right of the station code plots observed vertical component of the P-wave. The distance $\Delta$ and station azimuth as noted in Fig. 3 are shown as per station COL, College, Alaska. Seismic moment was calculated using Hawaii source models 1P and 2. Note ‘shoulder’ on first arriving P motion at all stations.
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systematic azimuthal fashion from 2.5 s for stations in the United States to about 5 s in the Philippine Islands. The relative amplitude of the first and second arrivals also varies. I attempted to model the second arrival as reflected energy from a sharp velocity transition near the earthquake source, but without success. The timing and waveform characteristics of the second arrival indicates that the 1973 Hawaii earthquake was actually a double event with both sources having similar mechanisms and the second source situated north-east of the first.

Three source mechanisms are noted on the first motion diagrams in Fig. 2. Fault mechanism 1S and 2 are determined from the analysis of SH waveforms and represent a best fit to the SH data for the double earthquake. The SH solution for the first source, 1S, does not fit the P-wave first motion for station KBS at Kings Bay, Spitsbergen. The P-wave first motions allow a wide range of focal mechanisms, but it was found in a systematic study of the P and SH waveforms that no solution for the first source could be found which simultaneously fit both the P and SH data. The P-wave solution for the first source, 1P provided a good fit to the P waveforms when used in conjunction with the SH determined second source, '2'.

To determine the far field displacement time functions for the two sources, the WWSSN instrument response was deconvolved from the P-wave observations. Symmetrical triangular time functions were fit to the pulse shapes. Station MAT, Matsushiro, Japan, in Fig. 4 was particularly suited for this measure. The first source was well fit with a symmetric triangle 3.3 ± 0.3 s in duration. The second source was less constrained, but could be matched with a symmetric triangle 2.5 ± 0.8 s in duration. Further details of the source, including source directivity, were obscured by the interference between the two sources.

Evidence for laterally heterogeneous and dipping structure

A depth of 48 ± 7 km for the first shock was obtained using the local 29 station seismic array on Hawaii (Unger & Ward 1979). A depth was determined from the teleseismic observations using the relative arrival times of the surface reflected phases to the direct arrivals recorded on WWSSN seismometers at distant stations. In modelling the P-wave observations an immediate difficulty presented itself in matching the amplitudes of the reflected phases pP and sP. This problem is illustrated with four P-wave observations in Fig. 5 and source models 1P and 2 from Fig. 2. The waveforms of the direct P-waves are well fit, as are the arrival times and directions of motion of pP and sP, assuming source depths of 42 and 32 km for the two sources. An arrival noted at COL, College, Alaska and observed at other stations to a northern azimuth may be modelled as a reverberation in the ocean, pwwP. However, the amplitude of sP at stations in the United States and Canada is much larger on the synthetic seismograms than is observed. A systematic check of source mechanisms fitting the direct P-wave arrivals revealed this to be a consistent problem. The problem is effectively resolved noting that the free surface and crust-water interface are not flat, but have substantial dips. The topography and bathymetry dip north-east between 3° and 20° in the area of sP reflections. The effect of planar dipping structure (Langston 1977) is illustrated in Fig. 6. Basically, the effect of the dipping surface is to change the point on the focal sphere from which sP is radiated. Including a 5° dipping island surface moves the source of sP radiation for QUI, Quito, Ecuador toward a sP nodal line, reducing the amplitude of sP relative to the direct P-waves. Uncertainties in the correction for dip compromise the use of the surface reflected phases in the determination of the source mechanisms of the two events. Timing information from the surface reflected phases is relatively preserved, however. A source depth of 42 km for the first event and 32 km for the second event.
Figure 5. Observed and synthetic seismograms of vertical component of $P$-wave motion are shown. Synthetic seismograms were calculated using Hawaii source models 1P and 2. At COL, College, Alaska two initial $P$ arrivals are followed 25 s later by $ppwP$, a multiple reflection within the ocean. At OTT, Ottawa, Canada and QUI, Quito, Ecuador the synthetic seismogram produces too large $pP$ and $sP$ surface reflections — each trace is normalized to the largest arrival. At RIV, Riverview, Australia $P$-wave arrivals $P_1$ and $P_2$ are well separated in time, $pP$ and $sP$ follow with proper amplitudes on the synthetic.

Figure 6. The nodes of $sP$, an upward travelling $SV$-wave converting to $P$ on reflection at the Earth's surface, are shown (a) on an equal area projections of the upper focal hemisphere for Hawaii source model 1P. A surface interface dipping at $5^\circ$ to the NE moves the effective location of QUI, Quito, Ecuador, on the focal hemisphere toward a smaller value near the $sP$ nodal line. The smaller amplitude of $sP$ for a $P$ synthetic calculated including a dipping ocean bottom relative to a synthetic calculated assuming a flat layered structure fits the observed QUI, Quito, Ecuador record better.

provides the best qualitative fit to the $pP$, $sP$, and $sS$ arrivals. The shallower depth of the second source allowed the arrivals of the surface reflections to overlap, producing a better fit to the observations. Uncertainties in the crust–mantle velocity model, sources, and the effect of dipping structure suggest this point is perhaps not resolvable. The depth of 42 km for the first source is within the error bounds of the local Hawaii network determination. Unger & Ward (1979) used a mantle velocity of 8.2 km s$^{-1}$ in determining the location of the event. If I adjust the mantle velocity in Table 1 to this lower value, the $pP–P$ times yield a
depth of 45 km for the first event. Without the additional control of $pP$ and $sP$ relative amplitudes, the direct $P$-waves cannot constrain the source mechanisms as the information is essentially limited to the $P$ first motions in Fig. 2. The timing variation between the direct $P$ arrivals of the two sources was fit if the second source initiated 3 s after and was located 15 km north-east of the first source. This assumes a 10 km depth difference between the two sources. If the two sources are placed at the same depth, the time lag between the sources is 4 s.

Records were obtained from strong-motion accelerographs in Honolulu and Kilauea. The Kilauea record at an epicentral distance of 55 km from the earthquake showed maximum horizontal accelerations of 0.17g N 30° W and 0.11g N 60° W with a maximum vertical acceleration of 0.07g (Nielsen et al. 1977). The strong ground motion arrives primarily in two bursts of roughly 4 and 3 s duration separated by a 0.5 s hiatus. If the strong ground motion instrument was triggered by the $P$ arrival, then the two strong ground motion arrivals probably correspond to shear arrivals from the two earthquakes.

Fixing the source depths, source time functions, and relative locations, the $SH$ waveforms in the eastern azimuth to the North American plus BAG, SHK, and MAT in the west were inverted to determine the best fault plane solution for each source for the $SH$ data. Only the direct $SH$ ray for each source was modelled as the earthquake was sufficiently deep such that surface reflected energy does not interfere with energy directly leaving the source. The solutions obtained from the inversion of the $SH$ data are shown in Fig. 3. The waveform comparison is quite satisfactory. Fig. 4 shows the fit of the $SH$ solution to selected $P$-wave observations, noting that the $SH$ solution violates the $P$-wave first motion of KBS in Fig. 2. While the $P$ waveform of COL is not well fit in Fig. 4, it is well matched in Fig. 5 using the fault plane solutions 1P and 2. This $P$-wave solution does not adequately fit the $SH$ data. Several inversion runs were tried using a joint $P$ and $SH$ data set, but the fit to COL could not be improved without substantially degrading the fit to the $SH$ data — particularly the $SH$ observation at BLA, Blacksburg, Virginia. The focal parameters favoured for the 1973 Hawaii earthquake are given in Table 2; the convention of fault angle parameters follows Langston & Helmberger (1975). Fig. 7 shows the match between observed and synthetic seismograms for the shear phase $SS$ radiated by the earthquake. $SS$ is a shear wave which reflects at the Earth's free surface at the midpoint of its propagation between earthquake and receiver. Unlike the direct shear wave $S$ and $sS$, the reflection above the earthquake, each of which are minimum time phases, $SS$ is a maximum time phase. The distortion that $SS$ undergoes at an internal caustic in its propagation is effectively modelled by a Hilbert transformation (Butler 1979a; Choy & Richards 1975).

The solution 1S and 1P in Fig. 2 are not so dissimilar as to cause much concern, but some thought is in order as to the nature of the apparent incompatibility of the $P$ and $SH$ data. Four possibilities present themselves: (1) too much confidence is placed upon the $SH$ data near the $SH$ nodes; (2) the assumed $P$ and $S$ velocities at the source are incorrect and distort the true focal solution; (3) the velocity structure at the source is somewhat laterally heterogeneous; (4) the $P$- and $SH$-waves are radiated from different places in the source region, an example of the $Z$ phenomenon (Bullen 1963).

| Table 2. Source parameters for 1973 Hawaii double earthquake. |
|-------------------|----------------|----------------|
| Source | 1S | 1P | 2 |
| Dip | 81° | 94° | 95° |
| Rake | 152° | 141° | 145° |
| Strike | -9° | 0° | -14° |
Figure 7. Transversely polarized SS are shown. SS is a shear wave which reflects off the Earth's surface at the midpoint of its propagation. The effect of an internal caustic in the propagation of SS is mimicked by Hilbert transformation of S. Corrections for geometric spreading and attenuation are included. Hawaii source models 1S and 2 for the 1973 Hawaii earthquake provide good agreement between observed and synthetic SS-waves.

The S-wave data detailing the SH nodal lines are high quality, low noise, and the azimuths between station and earthquake lie within 10° of being naturally rotated into the transverse component of motion, SH. Changing the mantle velocity at the source by 4 per cent shifts all stations radially inward or outward on the focal sphere by less than 1.5°. This effect is small and does not significantly change the relative relationship of P and SH data. Differentiation between lateral variation in the source region and complexities in the source is not easily resolved. However, an analysis of SH travel times and P and SH amplitudes together with the observed waveform complications for ray paths passing beneath the island of Hawaii suggest that lateral heterogeneity is important. Precise shear wave travel times from the earthquake were measured by a waveform correlation technique comparing the observed SH with synthetics generated using the well-constrained SH source model. Shear wave travel times of rays leaving the earthquake source to the north-east between 0° and 60° azimuth are about 1 s slower than Jeffreys-Bullen (1958) tables and exhibit low scatter (Butler 1979b). Shear travel times for azimuths greater than 60° from the southern United States to South America average 6 s slower than Jeffreys-Bullen tables and show greater scatter. A similar change was found for SH and P amplitudes at about a 50° azimuth in Figs 3 and 4. In calculating the seismic moments of the double earthquake, the moment of the second source is approximately one-third of the first, although uncertainties of the durations of the sources can change this estimate by 50 per cent. The P-wave moments for azimuths between 0° and 50° average at 3.5 x 10^{25} dyne cm. Past 50° the apparent P amplitudes grow significantly to values at South American stations of a factor of 5—7 greater than at the stations to the north-east. The average moment of the SH-waves between 0° and 50° is 5.3 x 10^{25} dyne cm. SH data at azimuths greater than 60° average a factor of 2 greater than the amplitude data to
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A source mechanism for the 1973 Hawaii earthquake

Although difficulties with lateral heterogeneities and dipping structure have hindered the resolution of the source parameters, the overall source mechanism of the 1973 April 26, Hawaiian earthquake is fairly well constrained by the $P$ and $SH$ waveforms and first-motions. The event is found to be a double earthquake with both events having similar strike-slip mechanism. The focal diagrams in Fig. 2 describe two possible orthogonal fault planes: left lateral motion on an east—west striking plane dipping southward 55°; right lateral motion on a near vertical north—south striking plane with a slip angle plunging 35° north. As no surface rupture or source directivity effects were observed, the body wave analysis cannot distinguish between the active and auxiliary fault planes. The approximate location of the second event fits fairly well with an observed trend of aftershocks north-east of the main shock in Fig. 1. However, the azimuth of the second source does not lie on strike with either fault plane of the events, indicating the two sources did not occur on a common fault. En echelon faulting in a multiple earthquake has been observed in the 1967 Venezuela earthquake (Rial 1978) and could also have occurred in these Hawaii events. The east—west trend of aftershocks about the mainshock suggests the east—west nodal plane was the active plane, but sufficient scatter in the distribution of aftershocks decreases the certainty of this conclusion. The aftershocks range 17 km east—west and lie between 35 and 55 km depth. Assuming an east—west fault dipping 55°, the fault area of the main event is roughly 415 km². Using a mean seismic moment of $4.5 \times 10^{25}$ dyne cm from the $P$ and $SH$ data to the north and north-east azimuths, the event stress drop $\Delta \sigma$ is calculated to be about 13 bar. This stress drop is comparatively lower than the average stress drop of 100 bar determined for intra-plate earthquakes (Kanamori & Anderson 1975). The average displacement on the presumed fault may be calculated from $M_0 = \mu DS$, where $M_0$ is the seismic moment, $\mu$ the rigidity, $D$ the average displacement, and $S$ the fault surface area. For $M_0 = 4.5 \times 10^{25}$ dyne cm, $\mu = \rho \nu G = 7.5 \times 10^{11}$ dyne cm², and $S = 415$ km², the average fault displacement is 14 cm. These values of stress drop $\Delta \sigma$ and displacement $D$ are considerably less than the stress drop of 93—42 bar and fault displacement of 5.5—3.7 m obtained in the study of the shallow, magnitude $(M_s) = 7.1$ Hawaii earthquake of 1975 November 29 associated with Kilauea (Ando 1979).

Unger & Ward (1979) report a fault plane solution for the main shock determined from $P$-wave first motions from the local short-period network and world-wide stations: nodal planes oriented N33°E dipping 77° NW and N70°W dipping 61° SW. The solution is primarily constrained by the local network. This local solution is similar to the solution presented in Fig. 2, but with a clockwise rotation of 33°. This discrepancy is considerably greater than the difference between solutions 1P and 1S in Fig. 2. The teleseismic $SH$ data are clearly incompatible with the local data. Two explanations may be invoked to explain the discrepancy. The local solution was found using a simple one layer crust over the mantle. Lateral heterogeneity and dipping internal structure beneath the island can change the apparent location on the focal sphere of the locally determined $P$ nodal line in much the same way as the nodal line of $sP$ was affected by the dipping surface of the island in Fig. 6. Further, the local stations constraining the locally determined N33°E nodal plane lie on the flanks of the active volcanoes Mauna Loa and Kilauea opposite to the 1973 earthquake, such
that the upward travelling P-waves will interact with the cores of the volcanoes. Inversion of local (Crosson & Koyanagi 1979) and teleseismic (Ellsworth & Koyanagi 1977) P arrival time data are interpreted to indicate a low-velocity zone at the base of the crust, a crust—mantle transition depressed 3°–4° in a conical configuration about the volcano Mauna Loa, and 11–18 per cent faster velocities in the crust and mantle beneath the Kilauea summit and rift zones than surrounding shields. Given the geometric configuration of the local Hawaii array in relation to the 1973 Hawaii earthquake epicentre, a moderate 4 per cent velocity anomaly along the path to key Hawaii stations can account for differences between the local and teleseismic solutions. Focal mechanisms of mantle earthquakes beneath Hawaii determined by the local array may be substantially biased due to uncertainties in the velocity structure. An alternative view of the discrepancy is that the P first motions recorded by the local array was actually a foreshock a second or so before the main event. A magnitude 4.5 foreshock could effectively mask the first motions of the local data and yet go undetected teleseismically. However, studies of the local seismograms by Koyanagi, Endo & Ward (1976) and Unger & Ward (1979) did not offer a foreshock to the 1973 Hawaii earthquake, and therefore tend to preclude this possibility.

Earthquakes occurred in the crust and mantle shortly following the 1973 April earthquake beneath the north-east flank of the volcano Mauna Kea and the summit caldera region of the very active Kilauea volcano (Unger & Ward 1979). The nodal solution in Fig. 2 striking north—south is somewhat aligned with mantle earthquakes beneath Kilauea and 35 km SSW of Kilauea caldera (Koyanagi & Endo 1971). Swarm earthquake activity has occurred at depths of 5 to 20 km beneath Mauna Kea in 1969 and 1967 and sporadic activity was noted in 1965 (Koyanagi 1968, 1969; Koyanagi & Endo 1971). A magnitude $M_L = 4.5$ earthquake in 1961 was located at a depth of 35 km at essentially the epicentre of the 1973 Hawaii event (National Earthquake Information Service, Golden, Colorado).

The 1973 Hawaii earthquake radiated a strong downward S pulse which was observed on the high gain long-period seismograph at Kipapa, Oahu, 3°Δ distant as multiple ScS reflections up to ScS6 (Best, Johnson & McEvilly 1974). These shear waves which reflect nearly vertically off the Earth’s core yield accurate estimates of the total crust—mantle shear travel time and average attenuation beneath the Hawaiian ridge. The relative strength of downward S for the focal mechanism in Fig. 2 is about 60 per cent of the maximum for the focal sphere.

The orientation of the axis of maximum principal stress for the two earthquakes trends N 45° E and is inclined about 25°, the axis of least principal stress trends S 45° E and is also inclined at 25°. These directions do not agree with the hypothesis of the origin of the Hawaiian chain as a propagating tensional fracture (Betz & Hess 1942; Green 1972; Turcotte & Oxburgh 1973) nor with other proposed stress directions in the Pacific plate (Jackson & Shaw 1975). The orientation of the maximum principal stress is qualitatively consistent with gravitational loading effects from the island of Hawaii upon the lithosphere. The intermediate and least principal stress axes may be controlled by lithospheric plate stress, magma injections and transport, or pre-existing zones of weakness. Uncertainties in the rheology, state of stress, and temperature in the earthquake source region do not allow a more definite statement.

Relationships between the 1973 Hawaii earthquake and the volcano Mauna Kea

The epicentre of the 1973 Hawaiian earthquake lies 5 km north of the east rift zone of the volcano Mauna Kea (Fig. 1). The east—west striking southward dipping nodal plane in Fig. 2 parallels the east—west strike of the rift zone. The depth of the event, about 45 km, is also
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the approximate distance to the central core of Mauna Kea. Mauna Kea is a volcano in the last stage of principal volcanism with an andesitic cap almost completely covering the original tholeiitic shield volcano (Stearns 1966; McDonald & Abbot 1970). An average of six carbon-14 dates of tephra indicates that a series of eruptions occurred along the south rift zone of Mauna Kea in about 2500 BC (Porter 1971, 1973). The age of the youngest ash layer in a lake near the summit of Manua Kea, inferred to be 3600±300 yr old, 1650 BC, probably records the most recent eruption of the volcano (Porter 1973; Woodcock, Rubin & Duce 1966). Mauna Loa and Kilauea on the island of Hawaii are active, young volcanoes. The volcanoes Hualalai on Hawaii and Haleakala on the island of Maui, are each of comparable age or older than Mauna Kea (McDougall 1964) and each has erupted in recent times: Haleakala in about 1790 and Hualalai in about 1801 (Stearns 1966; McDonald & Abbot 1970). The progression of volcanic shields in the Hawaiian chain has been explained by a melting spot roughly 300 km in diameter in which simultaneous tholeiitic eruptions occur (Jackson, Silver & Dalrymple 1972). The trend of the Hawaiian chain suggests the centre of the melting spot lies 40–50 km north of the island of Hawaii, nearest to the volcano Mauna Kea.

Rift zones are a feature associated with all the volcanoes of the Hawaii archipelago (Stearns 1966; McDonald & Abbot 1970). In the very active Mauna Loa and Kilauea volcanoes, the rift zones are persistent lateral distributaries — usually in imperfect communication with the central reservoir — which tap magma from the reservoir and possibly from the upper portion of the conduit feeding the reservoir from depth (Eaton 1962). Rift zones for individual volcanoes tend to lie on the trend of the archipelago, whereas volcanoes butressed by neighbours have rift systems of more varied orientations (Fiske & Jackson 1972). The rift zones may result from dikelike injection of fluids following the maximum principal stress direction (Fiske & Jackson 1972). The shallow extent of the Hawaii rift zones inferred from studying fluid injection in gelatin models (Fiske & Jackson 1972) depends critically upon the assumption that the magma reservoirs and rift zones radiating from them are contained largely or completely within the volcanoes themselves. However, inversion of teleseismic P-wave arrival times for the three-dimensional structure beneath seismograph stations located on Kilauea is interpreted to indicate that large volume concentrations of magma are absent to depths of at least 40 km (Ellsworth & Koyanagi 1977). Seismic activity beneath the active Kilauea east rift zone between 1965 and 1970 was shallower than 15 km depth (Fiske & Jackson 1972). Mantle earthquakes to depths of 60 km are located beneath the summit of Kilauea and Mauna Loa (Koyanagi, Swanson & Endo 1972). Eaton (1962) states 'From their heavy concentration on and around the island of Hawaii and its active volcanoes, it seems clear that even those earthquakes not intimately related to specific eruptions have an ultimate cause which is closely linked to volcanism: that is, the slow adjustment of the crust to the growing load of the volcano rising upon it and to changes within the crust and mantle resulting from the removal of magma from the depths and its migration to the surface.'

As noted earlier, lateral variations were found in SH travel times and P and SH amplitudes. Measuring in a clockwise fashion from north, rays leaving the source region (downward) show a marked change in character at an azimuth of 60°. Rays travelling east and south-east have slower travel times and are amplified relative to rays travelling north and north-east. The rays travelling east and south-east pass through the depth projections of the Mauna Kea east rift zone. If this is a region of lower velocity, a slowing and amplification effect through focussing can result. The shallow summit cores and rift zones of Mauna Loa and Kilauea have higher crustal velocities than the flanks of the volcanoes, while below 30 km depth low velocities appear beneath the summits of Mauna Loa and Kilauea.
Although the travel-time and amplitude data can be qualitatively explained by a slowing, of focusing structure — perhaps associated with the east rift of Mauna Kea — a more definitive, quantitative conclusion cannot be reached from the data at present.

Conclusions

In this study of the 1973 Hawaii earthquake the relationship of the volcano Mauna Kea has entered on several occasions. The question may be posed as to the nature of the connection between the double earthquake and the dormant volcano. The events and aftershocks occurred beneath the east rift zone of Mauna Kea, at a depth roughly equal to the lateral distance from the core of Mauna Kea. The east–west fault planes of the double events parallel the trend of the east rift zone. Shortly following the main shock an abrupt increase in earthquake activity was noted beneath the summit of the active volcano Kilauea. Amplitude and travel-time anomalies qualitatively correlate with the trend of the east rift zone of Mauna Kea. The earthquakes may have been caused by stress adjustment from magma movement in the mantle beneath Mauna Kea or from the gravitational load of the island not directly related to volcanism. The data cannot differentiate between these possibilities. Several studies mentioned previously lend circumstantial support for an origin of the 1973 earthquake related to a reactivation of volcanism at Mauna Kea. The volcano Mauna Kea is dormant, not extinct — having erupted in about 1650 bc (Porter 1973; Woodcock et al. 1966). From the trend of volcanic loci of the Hawaiian archipelago, the melting spot originating the volcanism lies north or east of Hilo (Jackson et al. 1972), within the proximity of Mauna Kea. Neighbouring dormant volcanoes — Hualalai on Hawaii and Haleakalā on Maui — have erupted within the last two hundred years (Stearns 1966; McDonald & Abbott 1970). These volcanoes are of comparable age to Mauna Kea (Porter 1971, 1973).

The 1973 Hawaii earthquake is the largest mantle earthquake to have been recorded in the archipelago. Detailed seismological studies of Hawaiian earthquakes before 1962 are practically non-existent. Prior to the deployment of large numbers of seismometers on the island of Hawaii in the 1960s the seismicity of Mauna Kea essentially was not recorded, and thus a fundamental gap in our knowledge of Mauna Kea must remain. In their present configuration on the island of Hawaii seismometers are primarily deployed on Mauna Loa and Kilauea. Deployment of more seismometers on Mauna Kea would improve detection of background crust and mantle earthquake activity beneath Mauna Kea and offshore. Epicentral locations of possible future mantle earthquakes and aftershocks would also be improved.

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References

The 1973 Hawaii earthquake


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