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SEISMIC AND RELATED PHENOMENA AT ACTIVE
VOLCANOES IN NEW ZEALAND,
HAWAII, AND ITALY

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Abstract

Recordings of the seismic power (rate of radiation of seismic energy) and seismic power spectral density (frequency distribution of seismic power) of volcanic vibrations at volcanoes Ruapehu and White Island in New Zealand, Kilauea in Hawaii, and Stromboli in Italy are reported, and related to visible volcanic activity.

At the andesite volcano Ruapehu, volcanic tremor of dominant frequency 2 Hz and seismic power < 1 kW occurs about 20% of the time. In 1961, 64, 66, 68 and 69, tremor increased (≤ 20 kW) and developed stoppages followed by powerful recommencements of tremor (≤ 230 kW) which resembled the B-type volcanic earthquake of Minakami. Brief eruptions of mud, water, ash, and pumice from the crater lake followed this activity in 1966, 68 and 69, the largest eruption (22 June 1969) having seismic power $= 8 \times 10^6$ W, kinetic power of eruption $= 10^{12}$ W, total seismic energy $= 7 \times 10^8$ J, and total kinetic energy $= 7 \times 10^{12}$ J. The volume of the eruption was $\geq 10^6$ m³. Ninety percent of the observed eruptions were forewarned within 12 days by the concurrence of volcanic tremor of ≥ 1 kW, and volcano-

seismic events of ≥ 20 kW, but only about 50% of these warning indications were followed by eruptions. The behaviour is intermediate between that reported at Aso volcano, Japan, where eruptions occur after tremor (without powerful recommencements) has decreased, and Asama Volcano, Japan, where B-type earthquakes rather than tremor precede eruptions.

The smoothed thermal power output of Ruapehu, calculated from the temperature of crater lake, increased during strong tremor, and exceeded 10^9 W in 1964, 66 and 68. Between 1965 and 1970, the depth of the lake decreased from 300 to 65 m below overflow level, and the volume decreased from 10×10^6 to 7×10^6 m³ due to infilling with tephra or lava.

At the andesite volcano White Island, volcanic tremor of dominant frequency 1.5 Hz, together with small swarms of A-type micro-earthquakes, occurred during long periods of non-violent ash eruption in 1966-67 and 1968. Tremor was most powerful (≤ 3 kW) and variable (stoppages and powerful recommencements ≤ 10 kW) when the erupting tephra had maximum fresh magmatic content but a comparatively low temperature of 445°C (maximum was 900°C). Powerful recommencements of tremor were closely followed by pulses in the ash emission, and from recorded times, the depth of origin was inferred to be 0.8 km.

At the basalt volcano Kilauea, the eruption and lava lake at Mauna Ulu vent on the upper east rift zone was the most powerful source of tremor during the study between January and April 1971. From mid February on, tremor with either a fairly flat spectrum between 1 and 5 Hz, or a dominant frequency of 1.3 Hz, occurred, while lava flowed steadily from Mauna Ulu through tubes to the SE. The tremor had stoppages, powerful recommencements and pulses which corresponded with cycles of quiet inflation of the lava lake and deflation by initially vigorous ebullition (fountaining).

Rockfall earthquakes in the lake had spectra similar to tremor, and often caused ebullition and tremor to start or stop. Rapid decay of their coda indicates a Q value of 2 - 6 for vesicular lava, in which the P-wave velocity could reach as low as 17 m s^{-1} .

The onset of ebullition was often influenced by surface events, and it is suggested that ebullition is the self-exciting collapse of a layer of lava foam which is continuously being formed at the top of the lava column, and which has minimum stability at a shallow depth due to high order inverse variations of viscosity and foam life with depth. It appears that the tremor recorded by the

author was the seismic noise of ebullition, modified by the response of bodies of rock, lava and gas, and by absorption. The latter apparently caused peak near 1.3 Hz on recordings made at any appreciable distance. Narrow spectral peaks were sometimes present, especially when ebullition involved the regular bursting of bubbles of several metres diameter. Change of spectra with recording distance gave a Q value of very approximately 30 for the upper layers of Kilauea volcano. Between 1 and 10 km, RMS amplitude of tremor was inversely proportional to distance.

At the trachybasalt volcano Stromboli, tremor of dominant frequency 1.6 - 2.9 Hz was almost continuous, and accompanied the ejection of spatter and discharge of short bursts of gas from the vents at intervals averaging 0.7 to 2 second. Seismic power of tremor varied from 0.3 - 6 kW, and its spectrum was not inconsistent with the pulse spectrum of audible discharge of gas from the vents. Once, tremor increased in power and changed in spectrum coincident with an explosion and partial collapse of a vent, which allowed more vigorous eruption. Similar slow changes of spectrum occurred during the 20 days recordings in July 1971. Between 6 and 13 explosion earthquakes per hour

occurred, with remarkably few earthquakes of other types.

Comparing the four volcanoes, stoppages and powerful recommencement of tremor occurred at all except Stromboli. At all volcanoes, tremor usually correlated with the eruption of gas (steam?) and tephra or spatter rather than lava flow, and this was best observed at Kilauea. The data are consistent with gas release from vesicular lava as a common cause of tremor, B-type earthquakes, and explosion earthquakes. Rockfall earthquakes can also resemble B-type earthquakes however, and other types of tremor undoubtedly exist.

The seismic power of eruption tremors during the study was greatest at Ruapehu (7.5 MW), followed by Stromboli (350 kW), Mauna Ulu (30 kW), and White Island (12 kW). The seismic energy per unit volume of eruption (noisiness) was greatest at Stromboli ($2 \times 10^5 \text{ J m}^{-3}$) followed by White Island (10^5 J m^{-3}), Ruapehu ($3 \times 10^3 \text{ J m}^{-3}$) and Mauna Ulu (10^2 J m^{-3} on 10 February 1971). The seismic power spectra of eruption tremor were similar at all volcanoes and were broadly consistent with modified white noise. Some peaks in the spectrum are shown to be special to the transmission path, but others appear to be features in the generated spectrum.

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CHAPTER 1.

Introduction

The seismometric study of volcanic activity has two declared purposes: to increase the knowledge and understanding of volcanic mechanism and behaviour, and to search for warning indications of potentially disastrous eruptions. The eruption itself is a brief event, often over before the researchers even hear about it, but the volcano is geophysically and geochemically active over a much larger period of time and volume of rock than is evident in the eruption. Continuous studies are therefore necessary and justifiable, but they do not come to fruition in the absence of eruptions. During a study of limited duration there is a considerable risk of obtaining an insignificant negative result. For this reason, the author feels extremely fortunate in being able to make seismic studies of eight eruptions at four volcanoes in a period of seven years. In fact the quantity of data collected is temporarily an embarrassment. Four of the eruptions occurred between 1964 and

1969 at Ruapehu, which is an andesite strato-volcano in the middle of the North Island of New Zealand (Fig 1.1) Ruapehu was also the only volcano for which continuous seismic recordings covering premonitory, eruptive and post eruptive activity were available to the author, and the study of these forms the largest part of the thesis.

Next in order of length of observation period was White Island volcano, an andesite strato-volcano 50 km off the coast of the Bay of Plenty, New Zealand (Fig 1.1). Volcano-seismic studies were made after ash eruptions began in November 1966 and January 1968. The remaining two volcanoes were studied seismometrically while the author was on Sabbatical leave in Hawaii and Italy during 1971, and were not originally intended for inclusion in the thesis. They broadened the author's experience so greatly that it was agreed they should be included. Nearly 3 months recordings at 2 stations were obtained during the 1969-71 Mauna Ulu eruption of Kilauea Volcano, Hawaii, (Fig 1.2), and nearly one month of recordings was obtained at Stromboli Volcano, Lipari Islands (Fig 1.3). At both these volcanoes eruptive

Fig 1.1: Map of the central region of the North Island of New Zealand, showing the locations of Ruapehu and White Island volcanoes.

Fig 1.2: Map of the island of Hawaii and the Hawaiian Islands (inset) showing the location of Kilauea volcano.

Map of the Island of Hawaii,

Fig 1.3: Map showing the position of Stromboli volcano in the Tyrrhenian Sea.

activity was continuous, and comparatively simple, but at Hawaii the complete course of the eruption was under intensive study by the Hawaiian Volcano Observatory, and the author was indeed privileged to be able to make a contribution. Only one volcano - Etna in Sicily - failed to erupt while the author was making recordings near by. These recordings, made in 1971 during the quiescence following nearly 20 years of activity, are almost unique data on the non-eruptive vibrations of Etna, but have not been included in the thesis owing to pressure of time and space.

The study has necessarily been episodic in execution, because the eruptions, and the opportunities for travel have been episodic. The organization of the thesis perforce follows the same lines. It is welded together by a continuity of purpose, and a general consistency in methods of observation and analysis, which is aided by the restricted range of volcanic activity which occurred. All four volcanoes are habitually active to some degree and commonly erupt non catastrophically and without felt earthquakes.

Previous Work on Instrumental Volcanic Seismology.

The instrumental study of volcanic seismology began with the classic studies by Omori (1911, 1914, 1920) of the eruptions of the Japanese volcanoes Usu-san in 1910, and Asama in 1910-1913. Omori recognised the seismic differences between earthquakes accompanied by eruption, and those not accompanied by eruption, and correctly interpreted them as due to difference in focal depth. As early as 1894 he correctly measured both the kinetic and seismic energy of an eruption (In Yokoyama 1957 a). He also published reports of earthquakes preceding the eruption of Usu-san in 1663 and 1822, and of Fuji in 1707 and Osen-dake in 1792, and it was at once obvious that warnings of eruptions might be possible from seismograph records.

Contemporaneously, seismological studies were begun at Kilauea in 1912 (Wood 1913), and improved at Osservatorio Vesuviano by the installation of Omori-Alfani seismographs (Malladra 1914). These replaced the undamped pendulum seismoscopes which Perret (1924) observed to resonate with volcanic tremor during the 1906 eruption of Vesuvius.

Interest was maintained by the earthquakes preceding the major eruption of Sakura-jima in 1914. Omori (1916) recorded the post-eruption activity on one of his seismographs, but it was left for Kizawa (1963) to interpret the Gray-Milne 3 component recordings of the earthquakes preceding the eruption in terms of upward migration of foci. In Hawaii, Wood (1915) recognised earthquakes and tremor as the prelude to an eruption of Mauna Loa.

At this stage volcano-seismic studies declined, a contributing factor being no doubt, a lack of convenient eruptions. In the period 1914 to 1928, only the explosive eruption of Halemaumau in 1924 (Jaggar 1924, Finch 1924) was of volcano-seismic importance. To the accompaniment of earthquakes along the east rift zone, a major submarine flank eruption apparently occurred, and drained Halemaumau to great depth, so that the unsupported superheated walls exploded.

At this stage in the review, the progress in volcanic seismology of geographically united groups of researchers will be followed separately, and interest will be concentrated on active volcanoes which have permanent seismographs close by so that the review can be more relevant to the author's work. As Minakami (1959) concluded, "Eruption of

highly viscous lava is preceded by large earthquakes. Eruption of andesitic lava sometimes is accompanied by large earthquakes and sometimes by microseisms (in a few volcanoes where seismometric observations have been made). Eruptions in fluid basaltic lava are almost without exception not accompanied by quakes". These generalisations are especially true of the active andesitic and basaltic volcanoes with open vents, which form the subject of this dissertation. Such was the stature of Omori's initial work however, that the relative importance of smaller and more continuous volcanic vibrations has often been overlooked, especially by researchers seeking to understand not so much the mechanism of the volcano itself, but the reason for its existence at that time and place.

Hawaii

At Hawaii, one of the early seismological successes was the recognition of volcanic tremor (Wood 1913) and its intimate association with molten lava. Jaggar (1920a) named the more regular tremor of frequency 1.25 - 1.7 Hz "harmonic tremor", and the erratic tremor of frequency 5Hz, "spasmodic tremor".

The difficulties of interpretation were soon apparent however. Whereas Wood (1913) said that volcanic vibrations clearly increase in energy in a general way when the magma

surface is high, is seething violently, and is sending up many magma fountains, Jaggar (1920a) noted that in December 1919, when thousands of fountains were playing in the Halemaumau lava lake with a roar audible for miles, there was no increase in the tremors on the seismograms. Then in a comment written in 1938, Jaggar (1947) said "this micro tremor is a volcanic vibration that fluctuates at Kilauea with the increases and decreases in lava fountaining at Halemaumau. The subsequent classification of short period tremors (0.2 to 0.5 seconds) at the Observatory indicated that those due to continuous lava fountaining or frictional rush of gas upward through the lava column, produced a uniform sinusoidal autograph --- we called harmonic tremor".

Finch (1949) was not convinced however, especially because harmonic tremor sometimes occurred when no molten lava was visible in Halemaumau, but was believed to be moving underground. He proposed a deep seated origin such as surging against the walls of the feeding conduit, and Omer (1950) theorised convincingly that it was caused by longitudinal vibration of lava laminae freed by differential tilting around the volcano, and excited by pulsating flow past one end. The model predicts a correlation between frequency of tremor and tilting which has never been

demonstrated, although Jaggard (1920a) found a correlation between amplitude of tremor and tilt.

In 1953 the upgrading of the Hawaiian Volcano Observatory seismograph network was begun by Eaton (Macdonald and Eaton 1955), with emphasis on more accurate focal determinations of earthquakes over a greater range of size than previously possible. In 1955 the first eruption along the east Puna rift zone in 115 years was studied by Macdonald and Eaton (1964). Earthquakes and harmonic tremor occurred first under Halemaumau and then along the rift zone.

In a general review Eaton and Murata (1960) reported the shallow earthquakes on the Honuapo-Kaoiki and Hilina fault systems, the small earthquakes around Kilauea caldera during swelling and shrinking, the shallow origin of harmonic tremor, the dike-splitting earthquakes preceding flank eruptions, and the great swarms of sharp earthquakes at 45 - 60 km depth which were accompanied by spasmodic tremor, but not clearly connected with eruptions. They also reported the remarkable eruption of Kilauea Iki in 1959, when the first observations of massive drainback of lava into a vent were made. Of $110 \times 10^6 \text{ m}^3$ of lava erupted, $70 \times 10^6 \text{ m}^3$ poured back underground, a

phenomenon which has not yet been adequately explained. The paper presents one of the clearest and most comprehensive models of a volcano and its activity ever published, but it was not clearly tied to the data published in the Hawaiian Volcano Observatory Summary and in Bulletins and Professional Papers of the U.S. Geological Survey.

Eaton (1962) consolidated the study, noting that at Hawaii every earthquake results from elastic strain release and not from explosions. Strong harmonic tremor correlated in 1959 with lava fountaining and drainback at Kilauea Iki although moderate tremor occurred at other times also (Eaton 1967). On three occasions, (Richter and Eaton 1960, Eaton 1962, Richter et al 1964) tremor was stated to originate at depths greater than 10 km, although no supporting data or references were given.

The papers on Hawaiian volcanoes published between 1963 and 1966 have been briefly reviewed by Decker (1967). Intrusive dike injections along rift zones, with or without surface eruptions were recognised by Moore and Krivoy (1964) from deflation of the summit of Kilauea accompanied by volcanic tremor, shallow earthquake swarms, and surface deformation along the rift zones. Earthquakes and faulting on the Koae fault zone in 1963

were intensively studied by Kinoshita (1967). An important and unique study of volcanic tremor was made in 1963 by Shimozuru, Kamo and Kinoshita (1966), who reviewed the previous tremor studies at Kilauea, made tripartite studies of propagation, and produced amplitude spectra of tremor showing dominant peaks between 0.5 and 2 Hz. Tremor of deep, intermediate and shallow origins were recognised on the basis of amplitude variation across the summit region of Kilauea. Deep tremor was of frequency 0.5 Hz and was very strong in the vertical component. Charts of tremor amplitude versus time were prepared for different recording sites, and tremor isoseismals at different times were drawn. The isoseismals were interpreted as outlining the migration path of magma from the summit of Kilauea to the eruption site (Napau crater) along the rift zone. The publication reveals the difficulties of interpretation. Moore and Koyanagi (1969) observed that subsidence of the summit region on 5 October 1963, indicating a movement of magma into the east rift zone, started 10 minutes before strong harmonic tremor and shallow earthquakes began. An air shock wave apparently marked the commencement of the eruption at Napau crater, and harmonic tremor increased markedly.

Since 1955, eruptions from the magma chamber under the summit have tended to occur along the rift zone. Kinoshita et al (1969) described the Halemaumau eruption of 1967 - 1968. Harmonic tremor of 2 Hz and of shallow origin preceded the eruption by about 1 hour, and accompanied all eruptive phases in proportion to the rate of lava fountaining.

On 24 May 1969 the Mauna Ulu eruption began with high fountaining on the upper east rift zone. An earthquake was felt (Koyanagi et al 1970), and harmonic tremor began and was monitored on a portable seismograph. During lava fountaining phases, tremor was strong, and between phases it was weak, and was correlated by Koyanagi et al (1970) with the "sloshing and bubbling noises" which could be heard down in the vent. The eruption area was the main source of tremor, but immediately after a major fountaining phase a secondary tremor source was obvious near the summit, and generally correlated with rapid summit re-inflation.

The value of tremor as an indicator of eruptive activity is clear in most of the Hawaiian Volcano Observatory publications, but few absolute measurements other than duration in minutes per day have been made. The

present author therefore included this aspect in his study of the relatively simple and stable activity at Mauna Ulu vent between January and April 1971.

Aso

The prehistoric eruptions of Aso volcano in Kyushu, Japan, covered large areas with ignimbrite of hypersthene-augite-andesite composition, and resulted in the formation of an impressive caldera. The overlapping cones now within the caldera have produced about 90 eruptions of incandescent lava fragments and scoria in historic time. According to Shimozuru (1961) the lava contains 53% silica. Sassa (1935,36) recognised four different kinds of volcanic tremor at Aso during the activity between 1928 and 1933, although he did not list them in any obviously logical order. They were:

1st kind, 1 Hz; Love waves from internal eruption
of gas,

2nd kind, 0.15 - 0.3 Hz, Oscillations of the magma
chamber.

3rd kind, 2 Hz: Rayleigh waves from internal
eruptions of gas,

4th kind, 5 Hz; Vibrations due to surface eruptions.

Sassa found that tremor could be used as a warning of eruptions. He plotted the ground amplitude and period of tremors against time, and showed that the amplitude increased suddenly to a maximum, was maintained for a period, and then rapidly decreased towards the time of eruption. This is similar to the pattern the author has observed at Ruapehu, N.Z., and Kubotera and Yoshikawa (1963) found it was still the best indication of impending eruption at Aso. Sassa was apparently the first person to publish a chart of tremor variations which could be calibrated in seismic power.

After a long period of quiescence, Aso killed 5 sightseers in 1953 and 10 in 1957, and these disasters stimulated scientific studies. Shimosuru et al (1958) and Yoshikawa and Kitsunozaki (1959) made new observations of tremors and pulsations, and studies were begun using tripartite seismograph arrays, Fourier analysis, and autocorrelation techniques. Wada and Ono (1963b) found that the tremor of the 1st kind travelled at 1.2 km s^{-1} , by determining the frequency at which the phase of the wave changed by 180° in crossing a tripartite net. Usually the wave was not strong at that frequency and noisy cross

correlation curves were obtained. Kamo (1962) and Kubotera (1964), using tripartite recordings and Fourier analysis, confirmed that the 1st and 3rd kinds of tremor were Love and Rayleigh waves respectively, with velocities of about 1.2 km s^{-1} .

Kubotera (1963; with Yoshikawa, 1963) used the frequency and damping factor of tremor of the second kind, to calculate the radius (2 - 4km) and damping ($h = 0.1$) of the resonant magma chamber assumed to be the source of this tremor. Tremor of the 3rd kind they considered to be caused by the ground acting as a narrow band filter for the Airy phase of Rayleigh waves, and showed that the P wave train (PL) of a distant earthquake was similarly affected, so that its dominant frequency was 2 Hz. Their data and analysis are insufficient to prove this however.

Kikuchi (1962,1964) found that the 4th kind of volcanic tremor consisted of dispersed Love and Rayleigh waves, and from observed frequency versus velocity curves, deduced that a surface layer 17 m thick, with P and S velocities of 1.5 and 0.235 km s^{-1} , was underlain by a 40 m layer, with P and S velocities 2.1 and 0.84 km s^{-1} below the tripartite array. Kikuchi and Sako (1966, 1968) studied

microearthquakes near the crater and found that they excited decay trains of the 1st, and 3rd kinds of tremor. It is not yet necessary to accept their conclusion that tremor is probably caused by micro earthquakes however. Their work typifies the difficulties of accurate location of foci in a noisy volcano, even when high gain - high frequency recordings are made on tape at short distances, and arrival time differences are determined by cross correlation techniques. Wada and Sudo (1967) found that volcanic earthquakes giving dilational first motions preceded and accompanied the eruption of 1965. Previously, Sassa (1935, 1936) found that explosion earthquakes at Aso had a quadrant pattern of compressional and dilational first motion.

Minakami (1963) has noted that at Mt. Aso, as at other volcanoes which erupt in Strombolian fashion, tremor instead of separate earthquakes usually precede eruptions, and he interprets tremor as small B-type earthquakes occurring in a continuous train.

Recently, Wada et al (1970) developed a digital energy accumulator which totals the energy of tremor and earthquakes each day. This is about 10^{12} erg per day

during the quiescent state of Aso. They assumed that the energy was propagated spherically as body waves in rock of density 2.0 Mg m^{-3} and velocity 1.0 km s^{-1} . This velocity is slightly below that found for tremor by Sassa (1935) and Wada et al (1965), and considerably below the P-velocity of 3.1 km s^{-1} found by Kamo and Kikuchi (1969) from seismic refraction studies in the caldera. Partly for this reason, the energy per day must be multiplied by 2.5 (and converted to Watts) to make it comparable with the seismic power of volcanic vibration measured by Dibble (1966, 1969) at Ruapehu volcano in New Zealand.

Asama

The smooth gentle silhouette of this pyroxene andesite volcano in Honshu, Japan, both misleads and warns us of its dangerous vulcanian activity. The studies of its earthquakes and explosions by Minakami, and indeed all his writings on volcano physics, are remarkable.

There was a long pause after Omori's classic study (1914) of the 1910-13 eruption which he recorded at 2 km distance. The Asama Volcano Observatory was established in 1933, and by 1947, over 600 explosions in the crater had been recorded. Minakami (1950) found great similarity

in the seismograms of different explosion earthquakes, and inferred that their mechanism and locations remained constant. He related the kinetic energy of eruption to the horizontal amplitude of surface waves, and to ground tilt, using the latter to determine a statistical forecast function which was successfully used to predict the explosion on 14 August 1947.

While studying the 1944 eruption of Usu and construction of the Showa Shinzan lava dome, Minakami, Ishikawa and Yagi (1951) divided volcanic earthquakes not accompanied by eruptions into three types:

- (i) A-type earthquakes, which were of normal tectonic appearance, and wave frequency nearly independent of distance.
- (ii) B-type earthquakes, which were small shallow shocks with indistinct S waves, and predominant surface waves of a frequency which decreased as distance increased.
- (iii) C-type earthquakes, which all had nearly identical waveforms with large amplitudes at the beginning and end.

The A and B-types were also recognised at Asama, and with explosion earthquakes and volcanic tremor (known as volcanic microseisms or micro tremor in Japan) make up the

four types which Minakami (1959, 1963) and Minakami et al (1959, 1960) have described as applying to Asama, Sakura-jima and other volcanoes. They showed that at Asama the relation between eruptions and a 5 day running total of daily frequency of B-type earthquakes was sufficient to predict the eruptions in October 1958, but that swarms which were slightly deeper than usual were not always followed by eruptions. A-type earthquakes and tremor are comparatively unimportant at Asama. Minakami et al (1957) concluded that B-type earthquakes and explosive eruptions shared the same energy source, and that earthquakes near the crater were an important clue to eruptions.

Kubotera (1965) reviewed the Japanese literature on volcanic earthquakes and found agreement with Minakami's classification of earthquake types.

In recent years Asama has been quiescent, and Minakami et al (1963, 1969 b, 1970 b) have consolidated their work on Asama, and studied many other volcanoes. Minakami (1969 c) summarised the results of four papers (Minakami et al 1969 a, 1970 a) on the earthquake swarms at Kirisima volcanoes, Kyushu, in 1968 - 69. Strong A-type earthquakes were followed by B-type earthquakes and continuous volcanic

tremor, and the earthquake foci moved from 3 - 9 km deep under Kakuto caldera to 0.2 km deep under the volcanic cones, 15 km further to the SE. No eruption occurred. Practically all first motions were downward, whereas Minakami (1969 c) states that explosion earthquakes give upward first motion. The first motion direction of B-type earthquakes are still unknown because they are so small and emergent.

Minakami et al (1969 b) made short seismometric surveys at 7 Japanese volcanoes and at Sotara in Columbia, and found that volcanic tremors are a common pre-eruption phenomenon at most volcanoes, but continuous observations over long periods of time are necessary to establish a pattern at any one volcano.

A useful method of classifying sequences of earthquakes was found by Ishimoto and Iida (1939) from recordings of earthquakes at Tokyo. Their coefficient m relates frequency of occurrence $N(a)$ of earthquakes to their amplitude a , and is conventionally written:

$$N(a) da = k a^{-m} da$$

where k is the constant of proportionality. Sakuma and Nagata (1957) report that m is 1.8 - 1.9 for ordinary earthquakes, and 2.5 - 4 for earthquakes accompanied by

outburst, or of very shallow origin. Minakami et al (1969 b) investigated the relation of m with depth by considering A-type earthquakes with S - P interval <3 s and >3 s separately at Kusatsu-Shirane volcano near Asama. Values of m were 2.8 and 1.8 respectively, although the earthquakes were clearly a single group. Later (1969 c) he assembled many data on m and depth for different areas, and showed clearly that m was related to depth. Mogi (1963) showed m increased as heterogeneity of material was increased, and Warren and Leatham (1970) showed that cracking due to large thermal gradients had a large m value. Scholz (1968) found the same for non uniform stress at normal temperatures. All these mechanisms could occur around a volcano, and cause variation of m with depth.

Sakura-jima

At the pyroxene andesite volcano Sakura-jima in Kyushu, Omori (1916) and Minakami (1950 a) made seismic studies of the effusive flank eruptions prior to 1955 which were sometimes accompanied by strong A-type earthquakes. Minakami et al (1957), Minakami and Mogi (1959), and Minakami et al (1960 a) studied the explosive summit

activity which commenced historically on 13 October 1955. Kubotera and Yoshikawa (1963) report that more than 1000 explosions had occurred by 1963. Explosive eruptions were often preceded by swarms of B-type earthquakes 10 - 15 days beforehand, and volcanic tremor was common. Graphs of eruption energy, daily earthquake frequency, and tremor were given by Minakami and Mogi (1959).

Yasui (1963) reported that change in daily frequency of B-type earthquakes and reduction of S - P interval of A-type earthquakes could be used to forecast the actual day of an eruption of Sakura-jima with good success, and Yoshikawa (1963) showed that the cumulative explosion earthquake energy between 1955 and 1960 was increasing at an accelerating rate. He accumulated the square of the maximum ground amplitude at an unstated distance from the volcano.

Yoshikawa et al (1963 a, English version Wada et al 1963 a), observed the 1962 eruptions with a 3 component seismograph and tripartite net, and measured the velocity of P and Rayleigh waves as 2 and 0.8 km s⁻¹ respectively. Love waves were very weak.

Tanaka (1967) made a statistical analysis of smoke, air waves, and earthquakes accompanying eruptions, and

found a close relation between frequency of explosion earthquakes and total earthquake energy. This reached a maximum in 1964. His logarithmic graph of number of explosion earthquakes against energy was convex upward, indicating a limiting size of explosion, thought to be due to the limited size of the explosive vent. Tanaka (1970) found that the magnitude of an eruption was unrelated to the length of the preceding period of quiescence.

Kamchatka

Excellent Russian studies have been made of the powerfully active Hornblende-andesite volcanoes of the Kurile - Kamchatka zone. Gorshkov (1959, 1960, 1961 a & b) found that the eruption of Bezymianny on 22 October 1955 began during a rapid increase in the rate of release of earthquake energy. The earthquakes were 50 km deep, and of unusual form and long period (2.5 - 3 s). On 30 March 1956, when the eruption was apparently over and the rate of seismic energy release was falling, a gigantic explosive eruption occurred, and decapitated the volcano.

The seismic studies of the Bezymianny eruptions enabled Tokarev (1964, 1967) to anticipate the gigantic eruption of Shevuluch volcano on 12 November 1964 after

a swarm of A-type earthquakes occurred accompanied by periods of 2 Hz tremor. He observed (1966) a sharp increase in earthquakes 80 - 270 km below the volcano at the time of the eruption. Gorshkov and Dubik (1970) reported that 479 earthquakes of total seismic energy 2×10^{20} erg preceded and accompanied the Sheveluch eruption which began with a gigantic explosion, and lasted about 70 minutes. Volcanic tremor of 2 Hz began 13 minutes after the main explosion, intensified and changed to 0.67 Hz, and stopped when the eruption stopped. It was recorded to distances of 430 km, and totalled 10^{16} ergs in seismic energy (300 KW for 1 hour). The source of this tremor could be 90 - 120 km deep on the basis of its amplitude distribution. The kinetic energy of the eruption was 10^{24} erg. Zobin (1970) found that the strongest earthquakes associated with the eruption had strike-slip mechanisms on a vertical plane of rupture.

Tokarev (1963b, 1966, 1971) suggested a method of prediction based on the Benioff strain release parameter, (seismic energy) ^{$\frac{1}{2}$} , for the preliminary earthquakes. For the eruptions of Bezymianny volcano in 1958, 1959, 1960, 1961, Sheveluch volcano in 1964, and Tokati volcano, Hokkaido, in 1962, the cumulative strain release

curves were approximately hyperbolic, and the eruptions occurred between 0 and 5 days before the time at which the cumulative strain release would have tended to become infinite. Seven to 10 days before the eruption, the hyperbola was sufficiently well defined by the preliminary earthquakes to allow prediction correct within 5 days. The mass of the eruption can also be predicted.

A different pattern preceded the 1951 eruption of the adventive craters on Kliuchevskaya. Gorshkov (1960) and Tokarev et al (1968) observed that the eruption began during a sharp decrease in preliminary earthquakes. The same occurred in 1966 when a new chain of adventive craters broke out (Gorshkov and Kirsanov 1968). The dike-splitting earthquakes stopped and volcanic tremor began as the eruption started. Balesta and Farberov (1968) mapped the lava underground by seismic refraction survey and obtained $750 - 850 \text{ m s}^{-1}$ as the velocity in molten lava using parametric methods. Very few data on velocity in molten lava in situ can be found in the literature, and these have been calculated rather than measured directly.

The eruptions of the Karymsky volcano in 1962 and 1965 were reported by Gorel'chik (1964) and Tokarev et al (1969). Volcanic earthquakes of type II, explosion earthquakes, and volcanic tremor were recorded, the latter holding

promise for the prediction of eruptions of this volcano. Graphs of numbers of earthquakes versus energy on log-log scales were parabolic for explosion earthquakes but linear for type II earthquakes, behaviour which resembles that reported at Sakura-jima by Tanaka (1967).

Several studies of volcanic tremor have been made. Gorshkov (1954) observed that weak tremor preceded the eruption of Bylinkina in 1951 by 24 hours, and became strong (15 micrometre at 22 km and 0.75 Hz) during the eruption. Shteynberg and Shteynberg (1964) pointed out that tremor is a stationary process, and that the only stationary process in volcanic activity is the emission of gas. Shteynberg (1965) was critical of Omer's (1950) explanation of tremor, and proposed instead the natural vibration of the lava column and branch channels above the magma chamber. At Klyuchevskaya volcano Faberov and Balesta (1966) found tremor was generated in the centre of the cone, and propagated as Rayleigh and Love waves at different group velocities. Frequencies up to 10 Hz were recorded, and anomalous attenuation was ascribed to geological inhomogeneity at the southern foot of the volcano. Gorshkov (1971) has recently reviewed work

on the prediction of eruptions and the detection of magma chambers, which depends mainly on high absorption of S waves.

Averko and Balesta (1970) suggest that assumptions made about shadow zones for S cast by magma chambers may be theoretically unsound.

Italy

In Italy, interest in volcanic seismology has been spasmodic. No substantial seismic studies at Etna have been found in the literature search. Three papers deal with the installation of seismographs on Etna, yet the 1971 eruption went almost entirely unrecorded. Latter (1967) recorded vulkanian type activity at Etna for 5 days in 1966, and found the hourly release of seismic energy in explosion earthquakes was 4×10^{16} erg. When this is expressed as an average seismic power of 1 MW, it appears suspiciously high. Latter (1967, 1971a) and Blot (1970 or 1971) studied the volcano-seismic activity of the Lipari Islands in 1966. At Vulcano during quiet solfataric activity, unusually perfect sinusoidal tremors were ascribed to the resonance of explosions in a gas filled cavity. At Stromboli the average number of earth-

quakes per hour was 10, of which 7 correlated with visible explosions in the crater. The seismic energy of each was about 3.5×10^{14} erg. Peterschmitt and Tazieff (1962) discovered feeble premonitory shocks of uniform appearance at Stromboli, which preceded surface eruptions and their vibrations by 15 to 40 second, depending perhaps on the height of the lava column in the conduit.

Imbó et al (1968) have reviewed volcanic seismology at Vesuvius. Harmonic tremor is seen to be closely linked with degassing of the lava column and the height of the column (Imbó 1935, 1952 a), and its regular nature is controlled by "quasi rhythmic lava motion amid episuperficial layers" (Imbó 1949). Spasmodic tremors are linked with explosive activity and have P and L waves. Imbó (1952 b) observed that the time interval L - P varied with the height of the lava column in the crater. Earthquakes tend to occur in groups during times of low activity. The stronger shocks are felt (Signore 1934). There are two types of volcanic earthquake:

a-type, of ordinary appearance and dominant frequency 10 Hz, and

b-type, with obscure phases but regular amplitude variations due to beating of two frequencies of about 2 and 3 Hz.

Imbó (1952) reproduced many seismograms of the 1944 eruption.

New Guinea - Solomon Is.

Although the hornblende-andesite volcano Mt. Lamington in Papua was not being monitored in any way at the commencement of its Pelean type eruption in 1951, the late stages of eruption were seismically recorded and described by Taylor (1958). Strong volcanic tremor and frequent earthquakes accompanied the rapid dome building phase and the secondary eruption of nuee ardentes of 18 February 1951. Taylor compares the earthquakes with those recorded at Usu and Asama By Minakami et al (1951) and presents a graph, equivalent to daily strain release (Benioff 1951) versus date, obtained by summing the amplitudes of earthquakes per day. This relates well with the eruptive activity. Taylor (1963) was also privileged to study a Pelean type eruption of the basaltic volcano Manam in New Guinea during 1960. The eruption was preceded by felt earthquakes and continuous volcanic tremor of 1.7 Hz.

Miscellaneous Studies

A number of valuable seismic studies have been made by expeditions to continuously active volcanoes.

Shimozuru (1960) recorded tremor of frequency 2.5 Hz at the basaltic volcano, Nyiragongo in the Congo. From variation of amplitude with distance he deduced the tremor source was at shallow depth under the south end of the lava lake. The attenuation coefficient was approximately $1 \times 10^{-3} \text{ m}^{-1} \text{ Hz}^{-1}$. Shimozuru compares this with apparently unpublished values of $1 \times 10^{-3} \text{ m}^{-1} \text{ Hz}^{-1}$ at Mihara and $0.3 \times 10^{-3} \text{ m}^{-1} \text{ Hz}^{-1}$ at Sakura-jima.

Shimozuru (1969) also studied Merapi Volcano in Indonesia where 5 types of volcanic earthquakes were recognizable by their spectra and shape of seismogram. B-type was the most frequently occurring type. Merapi volcano is unusual for its alternatively growing and avalanching lava dome.

Shimozuru's global experience is evident in his observations (1961) that the frequency of volcanic tremor is related to the percentage of silica in a volcano, but the exact relations he produces between frequency of tremor, and seismic velocity in magma to obtain a length of 500 m for the inferred resonating lava column in every case, may be accidental.

There is still much room for review and synthesis of the work on volcano physics, and volcanic seismology. The encyclopedia article by Sakuma and Nagata (1957) is

outstanding and a lot of the research since then has only served to confirm it, but it is conspicuously silent on the physical properties of rocks and magma within a volcano. Without knowledge of these, the internal geometry of the volcano cannot be determined, and critical models cannot be set up.

Energy Studies

Yokoyama (1956, 1957 a, 1957 b) was the first to make comprehensive studies of the energy release in volcanoes, 60 years after Omori had pointed the way. Working at the Strombolian type volcano Mihara, Ooshima Is., Yokoyama considered energy released in the following forms: volcanic tremors; local earthquakes; disruption of lava; potential, kinetic, and thermal energy of eruption; air and sea waves. He calculated the energy of the 16 largest tremors each day, assuming Rayleigh wave propagation, and by simple proportion to the total number of tremors, he obtained totals up to 10^{17} erg per day, (a power of 10^5 W), which he plotted against the date. His value is probably too high because he assumed all tremors equal to the 16 largest, but he was apparently the first to publish a graph of seismic power. He

restricted the use of the word power to the work per second done in raising the magma to the surface. By making calculations of energy for the historically great eruptions, using mainly the mass and type of ejecta, Yokoyama found great eruptions were as energetic as great earthquakes.

Gorshkov (1959, 1961b) independently considered the various forms of energy release in his work on the Bezymianny eruption, and presented a graph of energy of earthquakes each 10 days over the eruption period. The peak rate of earthquake energy release of 10^{21} erg per 10 days, corresponds to an average power of 100 MW over that period. Gorshkov estimated the total thermal, kinetic, airwave, seismic, and explosion earthquake energies as being in the ratio 10^6 : 3×10^4 : 10^3 : 80: 3.

Nakamura (1965) estimated that the average thermal energy of eruption was 1.4×10^6 J kg⁻¹ of erupted material, and the average energy of gas expansion (including seismic and kinetic energy) was 1.5×10^5 J kg⁻¹. The concept of power of a volcano was still largely neglected, but Yokoyama (1964) gave the kinetic power of the eruption of Tokati volcano in 1962 as 46 MW and the

thermal power as 4×10^6 MW. It appears the seismic power of volcanic tremor was first reported by Dibble (1965, 1966).

Shimozuru (1968) collected some published data on volcanic energy release in known time, and expressed it in ergs per second so that the different forms of activity at volcanoes Kilauea, Mihara and Nyiragongo could be compared. He emphasised that energy release in volcanic eruptions should be stated as energy per unit time, but it is not yet widely appreciated that this is power, and should be expressed in Watts. Most values are still being given as ergs over miscellaneous time periods which may or may not represent the period of measurement and are therefore misleading.

New Zealand

The potential value of volcanic seismology in New Zealand first became apparent on 10 June 1886, when an eruption of mud and basaltic lapillae at Tarawera, near Rotorua, killed 150 people. A series of felt earthquakes commenced an hour before the eruption, became violent during the eruption, and died away during the next few days (Hutton 1887, Roche 1948). As a result of this, and a much smaller hydrothermal eruption at Waimungu

on the Tarawera rift, which killed 2 people on 1 April 1917 (Grange 1937), Jaggar (1920b) advocated setting up a volcanological observatory at Waimungu, equipped with seismographs and other instruments. His proposal gained support from the Taupo earthquakes and fault movements of June - July 1922 (Grange 1932) which were long regarded as a typical earthquake swarm of possibly volcanic origin (Bullen 1947, Richter 1958), but which Eiby (1966) has reinterpreted as a major earthquake with aftershocks.

Grange (1937) also advocated an observatory, and in November 1945, during an ash and lava eruption of Ruapehu, volcano-seismic recordings were begun (Beck 1950), unfortunately too late to yield any data. No premonitory earthquakes were felt, and apart from explosions which cored out the tholoid which had temporarily displaced the crater lake, only one volcanic earthquake (on 27 September 1945) was felt during the eruption. Activity increased suddenly at that time (Beck 1950).

Jones (1947) showed that the P difference method would be well suited to locating epicentres at Ruapehu if 4 well distributed seismograph stations were available, and the first permanent station, Tongariro, was installed 9.2 km

from the craters of both Ruapehu and Ngauruhoe volcanoes in October 1952, just prior to the 1952-53 eruption of Ngauruhoe. Volcanic tremor of dominant frequency 1 Hz, and Vulcanian type explosions in the crater were recorded. Subsequently the magnification of this seismograph - a vertical component Jones - was insufficient to record volcanic vibrations, but when the Chateau Volcanological Observatory of Geophysics Division D.S.I.R. was built and equipped with a Willmore seismograph in November 1959, volcanic tremor of dominant frequency 2 Hz began to be recorded about 10% of the time.

Ruapehu was identified as the source of tremor after November 1960, when a slow motion tape seismograph (Dibble 1964, 1965) began recording by wire telemetry from 3 geophones in a line from the Observatory towards Ngauruhoe. By May 1961 the line extended to the summit of Ngauruhoe, where very small B-type earthquakes and moderate tremor were detected during weak activity in 1962. In 1964, the power of volcano-seismic noise from Ruapehu increased greatly, and the slow motion seismic tape recorders were transferred from Ngauruhoe to Ruapehu, which proved more profitable to study (Dibble 1965, 1966, 1969).

Other parts of the volcanic region were more profitable to the national economy however, and in August 1962 the Wairakei seismograph station (Willmore Z, $T_0 = 1$ s, $T_g = 0.25$ s, magnification 300 ca, code name WNZ) of Geophysics Division D.S.I.R. began operation, partly to safeguard the Wairakei Geothermal power station, and partly for regional coverage. In December 1964 the Wairakei seismograph and 3 portable seismographs were the closest recorders to a major earthquake swarm at Taupo (Eiby 1966, Ward 1972). No volcanic activity accompanied this swarm, but the swarm at Raoul Island in November 1964, which was initially recorded by a Willmore Seismograph (Z, $T_0 = 1$ s, $T_g = 0.25$ s, magnification 9400 at 5 Hz), installed by Geophysics Division D.S.I.R. in 1957 with no thought of volcanology, was followed by an eruption of mud and rock to a height of 800 - 900 m (Adams and Dibble 1966). The swarm was accompanied by volcanic tremor, and climaxed 10 days before the eruption. Eiby (1966) lists 12 swarms and possible swarms in New Zealand between 1922 and 1964, and on a proportional basis, perhaps 20 or more swarms unaccompanied by eruptive activity occurred between the pre-eruptive swarms of 1886 (Tarawera) and 1964 (Raoul Is.).

Perhaps by chance, the underwater volcano Rumble III was also unusually active in November and December 1964. Kibblewhite (1966) detected and located it by means of sea-borne volcanic noise as being 240 km ENE of Great Barrier Island, New Zealand. During an hour of its most intense activity, the acoustic power output averaged 30,000 MW. Subsequently, Kibblewhite (1967) discovered Rumble IV, 200 km E of Great Barrier Island. These volcanoes mark the extension of the Tonga-Kermadec Volcanic arc into New Zealand.

The tape recording techniques for studying volcano-seismic noise proved useful in mapping hydrothermal noise, which had previously been a problem in seismic prospecting for steam. Clacy (1968) proposed that the noise originated where water in contact with hot rock was boiling and recondensing. Whiteford (1970) found hydrothermal noise was essentially continuous, with a marked absence of discrete events. This, rather than the dominant frequency which was between 1.9 and 2.3 Hz at half the sites, distinguished hydrothermal noise from volcano-seismic noise.

The above review of previous work on instrumental volcanic seismology indicates the background against which the work reported in the present thesis was carried out. The author has paid special attention to the seismic power parameter, and whilst the greater part of the study is concerned with New Zealand volcanoes, especially Ruapehu, the author has also enjoyed several opportunities to apply his techniques to the observation of important eruptions in other parts of the world.

CHAPTER 2.
OBSERVATIONAL TECHNIQUES

Introduction

The basic instrumental technique used in the study has been that of seismology, but in an effort to understand the relation of volcanic seismology to eruptive activity, the author also carried out visual observations, audio observations and recordings, temperature measurements, and surveys of temperature, depth and chemistry of the water in the Ruapehu Crater Lake. At White Island, Kilauea, and Stromboli volcanoes, the author's seismic recordings and visual observations of eruptions were concurrent over limited periods, but this was never the case at Ruapehu. Here the author was heavily dependent on observations and photographs of eruptions made by the public, the Rangers of Tongariro National Park, and members of the New Zealand Geological Survey.

In collecting observations relating to the seismic recordings, the two immediate reporting systems to which the author belongs have been invaluable. In New Zealand, the system for volcanology involves Geological Survey, Geophysics Division and Chemistry Division, all of the New Zealand Department of Scientific and Industrial Research

and the Geology Department of Victoria University of Wellington. It is an entirely voluntary co-operative effort to disseminate all observations in the shortest possible time. Of wider extent, but of less use for the author's purpose are the event cards of the Smithsonian Centre for Short Lived Phenomena. It was C S L P event card 656 reporting rhythmic degassing from the Bocca Nuova (literally "new mouth"), at Etna which prompted the author to write to Dr. Haroun Tazieff, who arranged for the author to make seismic recordings there and at Stromboli.

Seismic Techniques

The recording techniques used by the author have centred around the slow motion tape recording seismograph (fig 2.1) originally invented by H. Benioff (1952) to aid in magnitude determinations of earthquakes, and developed by Dibble (1964) into a low cost portable unattended seismograph station, suitable to be left in the field for a month between visits. The tapes are played back at normal speeds such as 95 mm s^{-1} , but of course hundreds of times faster than they were recorded. The seismic vibrations then become audible, and can be

Fig 2.1: Slow motion tape recording seismograph,
1970 model, as used at Kilauea and Stromboli volcanoes.
The seismograph will operate unattended for 1 month
from a 12 V car battery.

classified by their distinctive sounds. Sutton and Pomeroy (1963) have demonstrated that the playback output is very suitable for connection to analog frequency analysers. It is also ideal for power level recordings and cumulative energy analysis, but not for visual readings of seismograms. Filming by oscilloscope and camera and reading in a viewer is very slow compared to reading conventional seismograms.

SEISMIC RECORDING EQUIPMENT

The Chateau Telemetry Seismograph

This slow motion tape seismograph was installed at the Volcanological Observatory of D.S.I.R. at the Chateau, 9.2 km from both Ruapehu and Ngauruhoe volcanoes, in November 1960, and was connected by wire telemetry to a line of 3 Willmore Mk 1 vertical seismometers extending initially from the Chateau to the foot of Ngauruhoe, and reaching the summit of Ngauruhoe in April 1961. The seismometer sites are referred to as Chateau 1, 2, 3, and 4, in Chapter 4, and are shown on fig 4.1. (Due to the geographic range of the observations, maps have been placed close to the text for each volcano).

The author has not used the recordings since 1964. The recorder is an ordinary household tape recorder (Philips E L 3516) modified by replacing the motor by a geared synchronous motor running 395 times slower, the tape head by a 4 channel head, and the record amplifier by 3 low frequency seismic prospecting amplifiers. Minute time marks and radio time signals, accurate to 0.1 s, are fed to the fourth channel. The tape speed is 0.24 mm s^{-1} , giving 26 days recording on a 549 m reel of 6.35 mm wide tape. The recordings have nominally uniform velocity sensitivity between 1.5 and 25 Hz, and the metal formers of the 100 kilohm seismometer coils ensure adequate damping of the 1.3 Hz seismometers.

Radio Telemetry Seismograph

By 1965 it was becoming obvious that a capacity for interpreting volcano-seismic recordings in real time was necessary. The Taupo earthquake swarm of 1964-65 (Eiby 1966) raised the possibility of an ignimbrite or pumice eruption, and rapid epicentre and depth determinations could have revealed any migration of foci towards the surface, and allowed the relocation of seismographs

for optimum accuracy. Considering that 2000 earthquakes a day commonly occur in swarms, normal seismological methods were inadequate.

To overcome the problem, the author built a portable 4-channel radio telemetry tape seismograph and analog epicentre computer in 1966-67. It was designed to take advantage of both the skill and art of a seismologist in the visual interpretation of seismograms, and the versatility of electronic playback of a tape record, and to allow rapid determination of epicentres in a continuous or intermittent process any time more than 20 minutes after the earthquakes occurred. A block diagram of the equipment is shown in fig 2.2, and one telemetry channel and the recorder are shown in fig 2.3.

Four Willmore Mk 1 seismometers are arranged in a triangle with a centre point, so as to give clear lines of sight from any one (the base station) to the other three. The recorder and analog computer (if required) are set up at the base station, and the radio seismometers at the other stations. All the seismometer signals are recorded on a four channel slow motion tape recorder, the tape from which is fed into a storage bin. Anytime more than 20 minutes after an earthquake has

Fig 2.2: Block diagram of radio telemetry seismograph and analog epicentre computer.

ANALOG EPICENTRE COMPUTER

Fig 2.3: Radio Telemetry Seismograph Equipment.
Upper: Willmore Mk 1 Seismometer, Transmitter,
Antenna and Batteries for one month's operation.
Lower: Antenna, Receiver, Slow Motion Tape
Recorder, and Battery.

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been recorded, the tape can be taken from the other side of the bin, and replayed 8 times per second at fast speed by means of a circular tape head rotating against the stationary tape.

The playback signals from the four seismometers are displayed in accurate time relationship as a stationary seismogram on a four channel oscilloscope whose sweep is synchronised by the rotating tape head. The vertical positions of the oscilloscope traces are controlled by the position of a marking point on a plotting table by means of cords connecting the marking point, through holes in the table representing the geophone positions, to vertical shift potentiometers. The earthquake can then be located in position and depth by moving the marking point across and away from the plotting table, until the P and S wave arrivals displayed on the oscilloscope are correctly aligned on slopes which represent the velocities in the area, and which intersect at the zero distance axis (giving the origin time). Permanent records of the recordings, epicentre, and quality of fit are made by photographing the oscilloscope and plotting table.

Specifications for the radio telemetry seismograph
are as follows:

Seismometers: Willmore Mk 1 Z, period 0.8 s, $h = 0.3$,
coil resistance 500 ohm, generated
EMF $100 \text{ V m}^{-1} \text{ s}$

Seismic Amplifiers: Teledyne EA 310.

Voltage gain; 400,000 max.

Frequency range; 0.5 - 12.5 Hz.

Noise level; 0.05 microvolt RMS.

Common mode rejection; -100db

Transmitters:

R-F output power 1 Watt.

Frequencies, 162.825, 162.850,
162.875 MHz, Pulsed transmission, 320
pulses per second. Pulse width modu-
lation by varying the mark/space ratio
from a median of 1:2 at the seismic
signal frequency and amplitude. E.g.
ground motion upwards, mark/space
ratio increases; ground motion down-
wards, mark/space ratio decreases.

Power requirements 12,0,-12 V, 3 W.

Antennae: Yagi 6 element array on 5 m mast.

Receivers: Frequencies 162.825, 162.850, 162.875
MHz.
Standard R-F and I-F sections.
Audio frequency section restores the
pulses to square shape and constant
amplitude.
Low pass filter and balanced output
stage reproduces the seismic signal.

Recorder: Modified BSR tape deck, type TD 10
for 6.35 mm tape.
Speeds 0.1, 0.2, 0.4 mm s⁻¹.
Nortronics 4 channel tape head type 5601,
Bias frequency 1 kHz.

Rotating Tape Head: Nortronics 4 channel head type 5602.
Mounted in drum of diameter 57.5 mm.
Nominal speed 8.4 rev/second, 1.52 m s⁻¹.
Connection by 4 slip rings and earth
return.

Oscilloscope: Six inch tube with 4 separate guns.
Frequency range 50 Hz to 100 kHz.
Triggered sweep, $0.01 - 6 \text{ s mm}^{-1}$.
External vertical shift controls.
Vertical deflection sensitivity
 0.7 mm mV^{-1} max.

Plotting Table: Area 760 x 690 mm.
Height range 0 - 400 mm.

The equipment was operated in the Bay of Plenty district through the period February - April 1967, when one of the radio seismometers was installed at Crater Bay (site S 1) White Is. (fig 9.2), during the study of that volcano, and the signal recorded at Whakatane, 51 km away (fig 1.1). Excellent recordings were obtained over this link. Recordings from the two other radio seismometers at Motu and Haroharo (fig 1.1) suffered from transmission paths which were not quite line of sight, but most of the tectonic earthquakes in the region were able to be located. They do not form part of this study however. In 1969, a radio seismometer was installed at Ski Patrol and recorded at the Chateau (fig 7.6) after the eruption on 22 June 1969 had damaged the seismograph

installation on Ruapehu. These recordings are reported in Chapter 7.

The Dome Seismograph

The 1966 eruption of Ruapehu emphasised the necessity for processing and interpreting volcano-seismic data in real time. Although seismic activity increased 9 days prior to the eruption, the Park Rangers, the researchers, and the public who were at the volcano were unaware of it at the time. Not only did this increase the danger of the situation, but it also resulted in a lack of observational data over this important period. To overcome the problem, the author proposed a co-operative programme between University of Wellington, Tongariro National Park Board and Geophysics Division D.S.I.R. which had three aims:

- (1) Better volcanic research data:
- (2) Real time analysis and display of volcano-seismic data for the Park Ranger:
- (3) Stimulation of interest in volcano physics.

These aims were achieved by recording the mean square value of ground velocity near the active crater of

Ruapehu, on a low speed ink chart recorder at Park Board Headquarters (here after referred to as PBHQ), and simultaneously recording the seismic waveform on a slow motion tape recorder (fig 2.4).

The seismometer was buried in tephra at the Dome, 700 m north of the centre of Ruapehu Crater Lake. The site was chosen to maximise the ratio of volcanic signals (from the crater) to noise (from wind, people and vehicles), and also to provide the best data on depth of activity below the crater. The pre-amplifier (fig 2.5) was installed in the Dome Shelter, a few metres from the seismometer, and connected to PBHQ 9 km away via a spare pair of wires in the emergency phone line. Good temperature stability and tolerance of the pre-amplifier to lightning surges was achieved by using 300 volt silicon transistors, and excellent signal to noise ratio on the telephone line resulted from locating the power supply and load resistance of the output state at PBHQ, and the output transistor in the pre-amplifier, thus allowing high signal power in the line without battery replacement problems. A serious disadvantage however was that breaks and partial shorts in the telephone line caused large pseudo signals to be recorded, and caused two false alarms as to eruptions.

Fig 2.4: Dome seismograph recorder at Park Board Headquarters.

Centre: Chart recorder for seismic power of Ruapehu volcano. Five days recordings including the eruption on 27 April 1968, are visible.

Top: The slow motion tape seismograph recorder.

Fig 2.5: Preamplifier of the Dome Seismograph, Ruapehu.

Upper: Preamplifier at top left of emergency phone box in the Dome Shelter. Lightning discharge tube, batteries and spare amplifier below.

Lower: Circuit diagram of preamplifier.

Lightning protection was by ballast resistors and a zener diode in the pre-amplifier, and 3-electrode gas discharge tubes connected between the telephone line and earth.

At PBHQ, the signal from the line is fed directly to the tape head, and also to an amplifier and rectifier with approximately logarithmic response (fig 2.6.). The rectified output is smoothed, and recorded on the chart against a logarithmic scale of Watts, the calculation and significance of which is explained in Chapter 3. The chart speed of 5 mm per hour enabled the last 8 days recording to be displayed below the recorder, and yet separated most seismic events along the chart. The equipment commenced operating on 29 Feb 1968.

Specifications for the Dome seismograph are as follows:

Seismometers: Willmore Mk II, Z , $T_0 = 2$ s, $h = 0.8$;
Coil resistance 380 ohm;
Damping resistor 1500 ohm;
Generated EMF $200 \text{ V m}^{-1} \text{ s}$.

Fig 2.6: Circuit diagram of the logarithmic amplifier and rectifier which drive the seismic power recorder of the Dome seismograph.

Amplifier: Voltage gain 500;
Frequency range 0.5 - 10 Hz;
Output impedance 500 ohm;
Battery life 9 months.

Tape Recorder: BSR TD 10 tape deck.
Channel 1 seismic signal, channel 3
time marks.
Tape speed 0.2 mm s^{-1} , width 6.35 mm;
Bias frequency 200 Hz.

Time Marks: Supplied by Geophysics Division D.S.I.R.;
Second marks from quartz clock;
Radio time signals at exact hours.

Logarithmic Amplifier:
Frequency range 0.5 - 20 Hz;
AC input range 5 - 0.5 - 0.05 - 0.005 V
peak;
DC output range 3.3 - 2.2 - 1.2 - 0.3 V;
Smoothing filter time constant, 2.5 s.

Recorder: Norma ink recorder 43-110-012;
Chart width 140 mm;
Moving coil meter 5500 ohm;
0.6 mA, 3.3 V, full scale deflection;
Response time 2 s;
Chart speed 5 mm per hour.

Portable Tape Seismograph, 1970 Model

The earlier versions of this instrument previously referred to (Dibble 1964), were usually available to the author on loan from Geophysics Division D.S.I.R. after 1966, but were becoming obsolescent. Improved amplifiers, and coding of time marks with the time of day were required for the work at Stromboli and Hawaii. The lack of coded time marks in the previous instruments forced the manual counting of minute time marks over periods up to hours. Automatic counters during playback were unreliable. However available coders were larger than the whole recorder and used more power, and new coders were built with RCA Cos/mos integrated circuits which allowed simple reliable design, and used negligible power.

The code consists of pulses each second, the minute pulse being enlarged 10 times, and the pulses on the seconds numbered the same as the hour and minute of the day being enlarged 3 times.

An example of the code is shown below.

The hour count was the one which did not change during any 2 minute section of the recording.

This code had 3 advantages:

1. The coding circuit was quickly and cheaply built.
2. The resultant pulses at exact minutes, exact hours and exact days increase progressively in height, and provide a visible time scale in a power chart recording (fig 10.5).
3. The code is simple to set and to read, and during fast playback the change of hours is audible as a change in harmonics.

The disadvantages may be as powerful:

- (i) The code is incompatible with any standard code or available code readers.
- (ii) It has very limited capacity. Day number, recorder, and site identification, cannot be added intelligibly.

The author's design is shown as a wiring diagram in fig 2.7 (pocket). Division from the crystal frequency (6.4kHz) through seconds, minutes and hours to 1 cycle per day was by series connected flip flops with gated reset signals, thus preserving true binary code.

Two 6-bit parity gates were connected between

(i) the seconds and minutes counters,

(ii) the seconds and hours counters,

so that each time either pair were at identical counts, a pulse 3 times larger than the seconds pulses was recorded. All pulses were 0.05 s in duration, and were superimposed on the seconds pulses.

Setting to time was done by releasing a reset button at an exact minute, and then pressing minute advance and hour advance buttons for the number of seconds corresponding to the number of the minute and hour during the day.

Specifications for the 1970 model portable tape seismograph, for which a wiring diagram is shown in fig 2.8 (pocket), are as follows:

Seismometer:	Willmore Mk 1, $T_0 = 0.8$ s, $h = 0.3$,
	Coil resistance 500 ohm, generator
	EMF $100 \text{ V m}^{-1} \text{ s}$.
	Damping resistor 220 ohm.

Recorder Size and Weight: 370 X 320 X 190 mm, 10 kg.

Tape Deck: Modified BSR model TD 10; speeds
0.1, 0.2, 0.4 mm s⁻¹; Nortronics 4
channel tape head type 5601, for 6.35
mm wide tape;
spool diameter 145 mm.

Amplifier: Teledyne EA 310; noise 0.05 microvolt
RMS; common mode rejection -100 db;
attenuator 0 to -42 db in 6 db steps;
range of level response, 0.1 - 25 Hz;
gain of channel 1 = 30,000,
channel 2 = 1000.

Internal Calibration: Symmetrical antiphase square-wave
signals of 0.5 Hz, at levels of 10, 10²
10³, 10⁴, and 10⁵ microvolt peak.

Radio: Range 0.53 - 1.6 MHz, with 1000 Hz
filter and rectifier feeding channel 3.

Time Marks: Minutes and seconds pulses coded with
the time of day, fed to channel 4;
rate less than 1 s per day.

Dynamic Recording Range: 80 db for channels 1 and 2
combined.

Crosstalk: -40 db between adjacent channels.

Endurance/Frequency: Frequencies up to 15 Hz can be recorded for 47 days on 822 m of triple-play, or 31 days on 549 m of double-play tape on 145 mm spools.

Power Consumption: 0.7 W at 6, 0, -6 Volt.

A small 12 V auto battery (44AH) will run the recorder for 1 month.

The performance of the two recorders at Kilauea, and Stromboli was very pleasing. They both operated in the field for nearly 4 months, during which only 28 instrument visits were made, and only 3 days of record were totally lost. At first, about 2 hours per day of high gain recording at Kilauea were lost on one recorder due to excessive solar heating, but a sun shade cured this. Another problem at Hawaii was that no time signals were receiveable on the broadcast band, to which the built in radio was restricted, and surprisingly great difficulty was experienced in receiving WWVH, transmitted from only 200 km away, on special portable receivers. As a result, absolute times on the Hawaiian tapes are only accurate to a few seconds. At Stromboli, the time signals on the Italian National programmes were well received but are not of guaranteed accuracy, and the

Neuchatel signals which are accurate, could only be well received with an elaborate installation.

It became evident that for field installations, the New Zealand station 2YA (570 kHz 250 kW) which broadcasts 18 time signals per day with normal accuracy better than 30 ms, provides superior service and coverage to that in many other countries.

PLAYBACK EQUIPMENT

The playback tape recorder normally used was a Philips stereo recorder, model EL 3534; which was fitted with a Nortronics 4-channel tape head, type 5602, and a patch panel channel selector allowing any two channels to be played simultaneously. The recorder had four speeds, 190, 95, 47.5, and 23.75 mm s^{-1} allowing signals in a wide range of frequencies and durations to be changed to the best range for processing or aural study, and had outputs which were independent of tone, volume, and balance controls, and therefore very suitable for calibrated playback into cameras, power level recorders, and spectrum analysers. Frequency response as determined by playing the EMI test tape SRT 13 previously used by

Dibble (1964), was satisfactory (fig 2.9), but the square wave calibration signal from the 1970 model tape seismographs was of distorted shape except when played at 47.5 mm s^{-1} . Normally, seismic tapes recorded at 0.2 mm s^{-1} were played back at 95 mm s^{-1} , a speed-up of 475 times.

In Hawaii, a Crown 4-channel professional tape recorder with Nortronics head type 5651 was used (fig 2.10). This 3 speed machine (380, 190, and 95 mm s^{-1}) had superior frequency response (fig 2.11) and allowed power level charts of all four channels to be made simultaneously. A serious fault was that it damaged tapes when stopping after fast wind. Both tape recorders could achieve a dynamic range of 50 db during playback.

Filters

It was often desirable to enhance a particular signal during playback by means of band-pass filtering. In New Zealand a Krohn-Hite continuously variable band-pass filter (model 330M) was used. It had R-C characteristics and a cut off slope of 24 db per octave. In Hawaii both Krohn-Hite type 330A and Rockland filters

Fig 2.9 Playback response of Philips EL 3534 type recorder to test tape SRT 13

Fig 2.10: Equipment at Hawaii Institute of Geophysics, used to process the authors slow motion tape seismograph recordings.

Left: Four channel rack mounted Sanborn AC level recorder.

Centre: Sona-graph spectrum analyser.

Right Centre: Crown portable 4 channel tape recorder.

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kHz on SRT 13 ; Hz on seismotapes recorded at 0.2 mm sec⁻¹

Fig 2.11 Playback response of Crown tape recorder to test tape SRT 13

were used. The Rockland filter has continuously variable high and low pass cut-off frequencies, and in addition, a choice of cut-off slopes (6,12,18,24 db per octave) and of Butterworth or R-C characteristics. It also served as an oscillator for use with the Willmore bridge when seismograph B was calibrated.

Oscilloscope and Camera

Normal wiggle-trace seismograms with a scale of 54 mm per minute were made from the slow motion tape recordings by playing them at 95 mm s^{-1} into a 4 beam oscilloscope with 150 mm screen, and recording on 35 mm film with a Shacklock automatic recording camera operating at a film speed of approximately 432 mm s^{-1} . The sweep of the oscilloscope was switched off during filming. Many other speed combinations were possible, and some seismograms were made at a scale of 13.75 mm per minute. Only two channels were filmed at a time because the playback tape recorder had only 2 channels. An alternative method of obtaining a seismogram of short length was to play the tape on the rotating tape-head of the analog-epicentre computer, and again into

the 4 beam oscilloscope and photograph the repeating stationary display with a Polaroid camera. This method was fast and gave a 4 channel seismogram if required, but the quality of the photos was less than with 35 mm film.

Power Level Re-recording Equipment

The concept of seismic power is defined on page 99. To obtain graphs of power level versus time from the slow motion tape recordings, audio frequency level recorders such as the Bruel and Kjaer AC level recorder at Physics and Engineering Laboratory N.Z. D.S.I.R. were used. This instrument records on waxed paper with a diamond stylus, and has a writing speed of up to 1600 db s⁻¹, and a range of 50 db (5 decades of power).

Its logarithmic scale is accurate to 0.5 db, and it can record the peak power level of volcanic earthquakes correctly provided the tape recordings are not played at more than 475 times the recording speed. Time scales of 22.5 to 180 mm per day along the chart proved convenient. Accurate time marks from the tape cannot be added to the B and K record because it is a single

channel instrument, but approximate time signals were impressed manually with the event marker. At the Hawaii Institute of Geophysics a Sanborn 4-channel audio frequency level recorder was used (fig 2.10). It also had a 50 db range and logarithmic scale but recorded with a hot stylus on heat sensitive paper 10 inches wide. Its accuracy was not as high as the B and K recorder, but the recordings were more useful because a channel was available for the coded time marks on the seismic tapes, and these gave recognisable changes of level at exact hours and days. Also the low sensitivity channel on the seismic tapes, and a filtered narrow-band version of the high sensitivity channel could be recorded, thus allowing recognition of extraneous noise and overloading on the power level graphs.

Frequency Analysis Equipment

Frequency analysis is discussed on page 114 . The machines used by the author to obtain seismic spectra were the Sona-graph machine manufactured by the Kay Electric Co., and the Rayspan spectrum analyser manufactured by the Raytheon Co. Both are audio frequency

machines very suitable for analysing speeded up seismic signals recorded on the slow motion tape seismograph and were made available by Physics and Engineering Laboratory, N.Z. D.S.I.R. However, until the author visited Hawaii Institute of Geophysics in 1971, calibrated and quantitative power spectra could not be obtained. At Hawaii, Dr. R. Norris and Dr. A. Popper generously made their Sona-graph machines available (fig 2.10), the second being a new machine of high reliability and accuracy. On returning to New Zealand the author was able to borrow and recondition the Sona-graph machine from Physics and Engineering Laboratory, D.S.I.R., and obtain quantitative power spectra for Stromboli volcano. The following brief description of this machine is abstracted from the instruction book:

The Sona-graph analyses a complex signal as a function of both frequency and time. The record, known as a sonagram, displays frequencies between 85 and 8000 Hz along a 4 inch vertical axis, a 2.4 second period of time along a 12.5 inch horizontal axis, and intensity by the darkness of the pattern. A signal can be analysed in about 5 minutes. In operation, the signal or a part of it is first recorded on a magnetic disc. Then it is

reproduced repeatedly at a speed 3.33 times the recording speed. On each repetition, a slightly different portion of the signal spectrum is scanned by either a 45 Hz ("narrow") or 300 Hz ("wide") band-pass filter. The output of the filter is recorded as a line of varying darkness on facsimile paper that is fastened around a drum rotating in synchronism with the magnetic disc. The recording stylus shifts gradually along the frequency scale in step with the scanning heterodyne oscillator. High signal intensity at a particular time and frequency shows up as a dark area on the sonagram, and a range of 40 db can thus be displayed qualitatively. A quantitative display known as a section can also be made. This presents, at any pre-selected point in time, a graph of spectral intensity on a scale of 1 db mm^{-1} in the horizontal direction versus frequency in the vertical direction. The intensity is integrated over a 5 ms period (i.e. 0.025 inches along the chart) prior to the instant of the section. The range of the section display is 35 db.

For random noise type signals, sections are difficult to relate to each other, even if 12 are packed on to one record. To aid interpretation, the author has superimposed the sections on the variable density display.

The relation between the peaks on the sections and the spectral trends along the sonagram are then clear. The machines in Hawaii had two additional features:

- (i) a contour display which gave a quantitative sonagram;
- (ii) frequency-scale-expansion which enabled any frequency interval between 85 Hz and 16 kHz to be expanded to fill the 4 inch vertical range.

The latter allowed the time resolution of the seismic sonagram to be increased by playing the seismic tapes more slowly, without compressing the frequency scale. However, frequency resolution is reduced. The contour display gave up to 7 contours at 6 db intervals, and when used with the wide-band (300 Hz) filter, gave very attractive sonagrams. However it proved impossible to calibrate the contour levels accurately. While reconditioning the rather elderly machine from Physics and Engineering Laboratory, a frequency-scale-expansion of two was incorporated by adding two series resistors (10 and 5 kilohm) on opposite sides of the stylus slider potentiometer, and an original feature - adjustable integration time for spectral sections - was added

to enable statistical fluctuations of the spectrum in time to be smoothed out of the section display. As built, the section sampling circuit had a time constant of 1.6 ms (equivalent to 5 ms at Sona-graph record speed). By adding a 2 megohm variable resistor, as shown below, the time constant could be varied between 1.6 and 42 ms.

The sections are then the integrated spectrum over 5 to 138 ms of recorded signal prior to the section instant. As discussed on page , spectral sections made with the narrow band filter and 138 ms integration time have 26 degrees of freedom, and with 90% confidence, are within 2.0 db of the true mean values. For 5 ms integration time they are within 6 db of true mean value.

The author's experience was that special care was necessary to obtain reliable spectra with a Sona-graph machine. Recording gain was kept low to avoid distortion due to overloading, especially when analysing

events too brief to indicate correctly on the VU meter, and frequent adjustment to the writing stylus and zero setting, and cleaning of heads and magnetic disc were performed. The positioning of sections on the sonagram was critical. The integration period had to lie within the event or period of stationary tremor which was being analysed.

The Rayspan analyser was used to analyse the White Island recordings and the Ruapehu recordings in 1968, and to give a compact display of spectrum versus time for recordings of one month's duration. It consists of 420 narrow band filters at equal spacing which cover the audio frequency range (via a heterodyne frequency changer) and are connected in parallel to the audio frequency signal being analysed. The outputs of the filters are scanned by a commutator at the rate of 62.5 or 125 scans per second, and the pick-off voltage applied to the brightness control of an oscilloscope, the sweep of which is synchronised to the commutator. This display is photographed on continuously moving 35 mm film, producing a picture similar to a sonagram, but of indefinite length. The nominal bandwidth of the filters is 32 Hz, and their spacing 25 Hz, so the

frequency resolution of a Rayspan film is similar to that of a sonagram. However there is no provision for a quantitative display in the New Zealand instrument, and the time resolution is poor. When analysing a seismotape at a speed-up of 475 times, and a commutator speed of 125 scans per second, the scans are 3.8 seconds apart in real time. Brief high frequency earthquakes can be lost from the record. Usually a speed-up of 950 and a film speed of 0.25 mm s^{-1} were used, giving a compact display with scales 0 - 10 Hz across, and 700 mm per month along the film. Long term spectral changes could be easily seen and the film served as a summary and index of volcanic tremor.

CALIBRATION

Two basic techniques of calibration have been employed:

- (i) the recording and playback of calibration signals through dummy seismometers.
- (ii) the use of the Willmore bridge technique to calibrate seismometer, recorder and playback equipment.

It was rarely possible to use the second technique, and most of the quantitative results depend on the first technique, and the assumption that the seismometers conform to their manufacturers' specifications.

Calibration by Dummy Seismometer Technique

In this method, the calibration signal from a signal generator is applied through a high resistance (e.g. 1 Megohm) to a 10 ohm resistor, and the voltage across this is fed to the seismograph amplifier via a resistor equal to the seismometer impedance. Usually the common lead of signal generator, 10 ohm resistor, and amplifier is grounded as shown below.

The advantages of the techniques are:

- (1) The recorder does not appreciably load the 10 ohm source. Therefore the voltage across it can be treated as a calibration EMF.

- (2) The calibration EMF is easily determined from the generator output voltage and resistance division.
- (3) Hum and noise from the dummy seismometer are always negligible.
- (4) Variations in the input impedance of recorders due to transistor characteristics, temperature, or choice of damping resistor, are automatically included in the calibration.
- (5) Subsequent calculations involving the seismometer EMF are simplified.

The early model portable slow motion recorders, and the Dome Seismograph have always been calibrated by this method. Initially this was done in the laboratory using mains powered signal generators but after several instances where recorders appeared to change sensitivity during long periods in the field, an ultra portable battery operated calibrator was built so that more frequent calibrations could be made.

In this device, a primary cell supplies a constant current through a large selectable resistance to a chain of ten 0.75 ohm resistors connected across 11 neighbouring copper strips on a piece of circuit board, so as to

give a voltage gradient across the board. The gradient is scanned by a 2-pole commutator, the brushes of which contact the strips along a circular path of diameter 11 strips wide, and pick off approximately sinusoidal antiphase voltages which are fed out through a resistor equal to the seismometer resistance. The voltage across the commutator represents the seismometer EMF, and is selectable from 1 microvolt to 100 mV peak to peak in a 1, 3.3, 10 sequence. The calibrator is connected in place of the seismometer and the commutator turned at about 2 Hz by whirling the device around the commutator shaft which serves as a handle, and as the lead in for the wires.

The calibration so obtained is reliable for all conditions of the seismograph except those involving faults in the seismometer. Other faults such as unbalanced resistance to ground of the telemetry circuit, which cause one seismometer input terminal to be more sensitive than the other, cause no errors when the portable calibrator is used owing to its fully floating antiphase output.

A problem with sine wave calibration methods is the excessive time required to calibrate a slow motion tape recorder over its full frequency range so that each frequency has sufficient duration during high speed play back. This was solved in the 1970 version of the recorder by means of built-in calibration using square-wave signals, the spectrum of which covered most of the recording frequency range. Two antiphase symmetrical square wave signals of 0.5 Hz and 5 V peak amplitude were taken from the quartz clock, and separately attenuated and switched to provide dummy seismometer signals of 10 , 10^2 , 10^3 , 10^4 , and 10^5 microvolt (peak) EMF. For calibration of seismograms and power level charts at 0.5 Hz, the peak voltage, which is also the RMS voltage, is used. For calibration of power spectra, and for calibration curves of sensitivity versus frequency it is necessary to know the amplitude of the individual harmonics in the square wave.

These can be found by elementary Fourier analysis as follows:

Let the square wave pass through zero amplitude at $t = 0$. As the sum of the harmonics must equal the whole wave form $f(t)$, we can assume:

$$f(t) = B_0 + \sum_{n=1}^{n=\infty} B_n \sin \frac{2\pi n t}{T} \quad 2.1$$

where B_n is the peak amplitude of the n^{th} harmonic;

B_0 is the DC offset, = zero for a symmetrical square-wave;

t is the time;

T is the period of the square-wave.

Then the amplitude of the n^{th} harmonic is

$$B_n = \frac{2}{T} \int_0^T f(t) \sin \frac{2\pi n t}{T} dt \quad n = 1, 2, 3, \dots \quad 2.2$$

Let $f(t) = A$ for $0 < t < T/2$ where A is the peak
 $= -A$ for $T/2 < t < T$ amplitude of the square-wave.

$$\begin{aligned} \text{Then } B_n &= \frac{2}{T} \left[\int_0^{T/2} A \sin \frac{2\pi n t}{T} dt + \int_{T/2}^T -A \sin \frac{2\pi n t}{T} dt \right] \\ &= \frac{A}{\pi n} \left[-\cos \pi n + 1 + \cos 2\pi n - \cos \pi n \right] \quad 2.3 \end{aligned}$$

Because n is a positive integer $\cos 2\pi n = 1$

$$\therefore B_n = \frac{2A}{\pi n} (1 - \cos \pi n) \quad 2.4$$

To evaluate this, put $A = 1$, $T = 2$ s,
and take n in the range 1 to 41, noticing that for n
even, $B_n = 0$.

The amplitudes of B_n , in db relative to the
amplitude (A) of the square-wave, are given in table 2.1.
The power in each harmonic (proportional to $B_n^2/2$), in
db relative to the power of the square-wave (proportion-
al to A^2), is also given.

TABLE 2.1: Frequency and Amplitude of Harmonics of a Symmetrical Square Wave of 0.5 Hz and Unit Peak Amplitude

n	Frequency Hz	Peak Amplitude	Relation of harmonic amplitude to square wave amplitude db	Relation of harmonic power to square wave power db
1	0.5	1.273	1.07	- 1.9
3	1.5	0.425	-7.5	-10.5
5	2.5	0.255	-11.9	-14.9
7	3.5	0.182	-14.8	-17.8
9	4.5	0.141	-17.0	-20.0
11	5.5	0.116	-18.7	-21.7
13	6.5	0.098	-20.2	-23.2
15	7.5	.085	-21.4	-24.4
17	8.5	.075	-22.5	-25.5
19	9.5	.067	-23.5	-26.5
21	10.5	.0607	-24.4	-27.4
23	11.5	.0554	-25.2	-28.2
25	12.5	.0509	-25.9	-28.9

n	Frequency Hz	Peak Amplitude	Relation of harmonic amplitude to square wave amplitude db	Relation of harmonic power to square wave power db
27	13.5	.0472	-26.5	-29.5
29	14.5	.0439	-27.2	-30.2
31	15.5	.0411	-27.7	-30.7
33	16.5	.0386	-28.3	-31.3
35	17.5	.0364	-28.8	-31.8
37	18.5	.0344	-29.3	-32.3
39	19.5	.0327	-29.7	-32.7
41	20.5	.0311	-30.2	-33.2

Calibration by Willmore Bridge Technique

The theory of the technique is given by Willmore (1959), and the calibration circuit is shown in fig. 2.12.

The method is to set up the seismometer coil as the unknown impedance of a Maxwell bridge circuit, and to use the seismic recorder as the detector. The bridge is driven by an oscillator at a frequency for which calibration is required, and balanced with the seismometer clamped.

On unclamping the seismometer, a signal is recorded with an amplitude equivalent to that which would be produced by a ground acceleration proportional to the current which flows through the seismometer. The constant of proportionality is determined by switching the output of the drive oscillator from the main drive input of the bridge to the substitution drive input, and again recording the resulting signal.

It is assumed that the suspended mass M of the seismometer is known. The seismometer period and the overall velocity sensitivity at each test frequency are calculated from the measurements.

Fig 2.12: Willmore bridge calibration circuit, as used to calibrate tape seismograph B.

Fig 2.13: Frequency response of tape seismograph B and Willmore Mk 1 seismometer 161713 to constant ground velocity, when tapes are played back on the Crown and Philips tape recorders.

The following values were measured -

R_R, R_B	the bridge arm resistances at balance; ohm
f_T	the oscillator frequencies at which the seismometer is calibrated: Hz
θ_2	the playback level (db) for main drive input at each frequency, at seismograph amplifier gain -36 db, and known playback and level recorder gains.
θ_4	the playback level (db) for substitution drive input at each frequency, at the same gain settings as for θ_2 .
Ed	the drive voltage measured at the input to the bridge; volts peak.

The following were computed -

f_0	the seismometer natural frequency, = f_T Hz when $\theta_4 - \theta_2$ db is a minimum. From a graph of $\theta_4 - \theta_2$ versus f_T .
R_C	the seismometer coil resistance, found from $R_C = \frac{R_R R_B}{R_D} \quad \text{ohm,}$

where $R_D = 7482$ ohm was the fixed arm of the
bridge.

G a nondescript quantity found from $G^2 =$ the slope of the graph $\left[f_T - \frac{f_0^2}{f_T} \right]^2 \text{ Hz}^2$ versus the power ratio corresponding to $\theta_4 - \theta_2 \text{ db; Hz}^{-1}$.

K the seismometer voltage for unit relative velocity between seismometer mass and frame,

$$= \left[\frac{R_R (R_B + R_C)}{(R_S + R_R)G} 2\pi M \times 10^{-3} \right]^{\frac{1}{2}} \text{ V m}^{-1} \text{ s}$$

where $R_S = 9710 \text{ ohm}$, was the series resistance in the substitution drive circuit,

$M = 4.26 \text{ Kg}$ is the suspended mass within a Willmore Mk 1 seismometer.

SGA the simulated ground acceleration.

$$= \frac{K}{M} I_c \text{ m s}^{-2}$$

$$= \frac{K}{M} \frac{Ed}{(R_B + R_C)} \text{ m s}^{-2}$$

where I_c is the current (amp, peak) through the seismometer coil for main drive input.

SGV the simulated ground velocity

$$= \frac{SGA}{2\pi f_T} \text{ m s}^{-1}$$

$$\text{Velocity Sensitivity} = \frac{\theta_2}{SGV} \text{ db m}^{-1} \text{ s}$$

Measured and computed values for slow motion tape seismograph B for the calibration made in Apr 1971 are listed in table 2.2.

The measurements were made at a seismograph amplifier gain of -36 db, but in table 2.2 the velocity sensitivity is given at a seismograph gain of 0 db and refers to the db scale of channel 1 of the Sanborn AC level recorder at Hawaii Institute of Geophysics when the level recorder is set to 30 db gain, and fed from channel 1 of the Crown portable tape recorder (speed 190 mm s^{-1} , equalisation M, gain setting 6.9), through a Rockland filter set to Butterworth characteristic, gain 0 db, and cut-off slope 0 db/octave. The resulting response curve (Fig 2.13) strictly applies only under these conditions but to enable it to be used in other circumstances, test tape SRT 13 was played and recorded on the level recorder under the same conditions. The "one kHz for level" signal on SRT 13 measured 40 db on the same scale as for θ_2 and θ_4 , and the relation of other frequencies is given in Figure 2.11.

Using also Figure 2.9, a response curve was thus drawn for tape seismograph B and the Philips EL 3542 playback equipment, and is shown at arbitrary level on Figure 2.13.

TABLE 2.2: Calibration of Portable Slow Motion Tape Seismograph B (1970 Model)
by Willmore Bridge Technique in April 1971

R_R	R_B	f_T	Ed	θ_2	$\theta_4^{-\theta_2}$	$\theta_4^{-\theta_2}$	$\theta_4^{-\theta_2}$	f_0	$\left[\frac{f_T - f_0}{f_T} \right]^2$	G	R_C	K	SGA	SGV	Velocity Sensitivity at 0 db gain
Ω	Ω	Hz	V peak	db	db	Power Ratio	Hz	Hz	Hz ²	Hz ⁻¹	Ω	V m ⁻¹ s	$\mu\text{m s}^{-2}$	$\mu\text{m s}^{-1}$	db $\mu\text{m}^{-1}\text{s}$
30.1 117112															
		0.4	0.50	17.5	8	6.3	0.98		4.00	1.30	471	86.3	861	343	2.8
		0.6	0.52	24.0	2	1.6			1.00	1.25			895	238	12.4
		0.7	0.51	26.0	-0.5	0.89			0.451	1.41			878	200	16.0
		0.8	0.50	29.7	-6.5	0.22			0.160	1.65			861	171	21.0
		0.9	0.50	31.7	-17	0.02			0.028				861	152	24.1
		1.0	0.50	31.7	-20.5	0.01			0.0015				861	137	25.0
		1.2	0.50	29.2	-5.5	0.28			0.160				861	114	24.1
		1.5	0.50	25.0	0.5	1.12			0.739	1.23			861	91.4	21.8
		2.0	0.52	21.5	6	4.0			2.31	1.32			895	71.2	20.4
		3	0.52	17	9.5±1	8.9			7.18	1.12 ca			895	47.5	19.5
		4	0.51	13	14	25.1			14.1	1.33			878	35.0	18.1
		6	0.51	9	17.5	56			34.1	1.28			878	23.3	17.7
		8	0.6ca	9ca	19	79			62.1				1030ca	20.5ca	18.8ca
		10	0.50	4	21.5	141			98.1				861	13.7	17.3
		12	0.50	3.2	21.5	141			142				861	11.4	18.1
		15	0.50	0.7	23	200			223				861	9.1	17.5
		20	0.51	0.5	20.5	112			398				878	7.0	19.6

The internal calibration signal of tape seismograph B was also included in the comparison, and gave the following result:-

Square-wave calibration EMF, microvolts peak.	Sanborn Chart level, db
10	< 4
10^2	9.5
10^3	28.0
10^4	47.0

The main results of the Willmore Bridge calibration are that the velocity response of the 1970 model portable tape seismograph B at Hawaii, was constant within ± 1 db between 3 and 20 Hz, but had a 6.5 db peak at the seismometer natural frequency. The measured value of K was 13.7% less than the manufacturers specification.

PERMANENT SEISMOGRAPH STATIONS

At two of the four volcanoes studied in this thesis, permanent seismograph stations existed, and their seismograms have been used by the author. The stations are at Ruapehu Volcano, New Zealand, and Kilauea Volcano, Hawaii, and have specifications as follows:

Ruapehu:

Two permanent seismograph stations were operated by Seismological Observatory, Geophysics Division, D.S.I.R. during the study and the author enjoyed continuous free access to the seismograms. This invaluable assistance is gratefully acknowledged. The station specifications are given in Table 2.3.

Hawaii:

A telemetry network of 17 seismograph stations recording on a single Develocorder and 6 or more smoked paper recorders was being operated by the Hawaiian Volcano Observatory, U.S. Geological Survey during the period of the present study, 21 January - 13 April 1971, and the author is appreciative of the opportunity to inspect the records during his necessarily brief field visits.

TABLE 2.3: Permanent Seismographs at Ruapehu.

Name	Code	Lat. S			Long. E			Alt m	Seismograph	Compt	To		Tg	Magnification		Period of Operation; Remarks
		D	M	S	D	M	S				s	s		Mag.	Hz	
Chateau	CNZ	39	12	00	175	32	51	1116	Willmore I	2	1.0	0.25	36750	5		Nov. 1959 to present.
Tongariro	TON	39	12	10	175	32	17	1120	Wood-Anderson	NW- SE	0.8	--	2800	5		Oct. 1962 - Early 1966. Component direction is in the direction of Ruapehu.
													1800	1		
													600	0.5		

A complete set of photo copies of an observed rockfall earthquake was supplied by Dr. D.W. Peterson, Chief Scientist, and is gratefully acknowledged. Dr. John D. Unger (personal communication) supplied the station specifications given in Table 2.4.

TABLE 2.4: Permanent Seismographs at Hawaii.

Name	Code	Latitude N		Longitude E		Alt	Recorder D = Develocorder S = Smoked paper	Compt.	Remarks
		D	M	D	M				
Ahua	AHU	19	22.40	155	15.90	1070	D S	Z	
Cone Peak	CPK	19	23.70	155	19.70	1038	D	Z	
Desert	DES	19	20.20	155	23.30	815	D S	Z	
East Koeae	EKO	19	21.80	155	15.30	1009	D S	Z	
Hilina Pali	HLP	19	17.96	155	18.63	707	D	Z	Above average magnification?
	HPU	19	46.85	155	27.50	3396	D	Z	
Kahuku	KHU	19	14.90	155	37.10	1939	D	Z	
Kipuka Nene	KPN	19	20.10	155	17.40	924	D	Z	
Makaopuhi	MPR	19	22.07	155	09.85	881	D	Z	
Mauna Loa	MLO	19	29.80	155	23.30	2010	D S	Z	
Lower Mauna Loa	MLX	19	27.60	155	20.70	1475	D	Z	
North Pit	NPT	19	24.90	155	17.00	1115	D	Z	
Outlet	OTL	19	23.80	155	16.94	1038	D	Z	
Pahoa	PHO	19	28.90	154	53.40	215	D	Z	
Puu Huluhulu	PHH	19	22.45	155	12.66	988	D	Z	Above average magnification?
Uwekahuna	UWE	19	25.40	155	17.60	1240	D	Z	
West Pit	WPT	19	24.70	155	17.50	1115	D S	Z	

CHAPTER 3.
METHODS OF ANALYSIS
Introduction

Five principal methods of interpreting the physical data have been employed. They are:

- 1) Interpretation of ground velocity amplitudes of seismic waves in terms of the nominal seismic power of the source. Seismic power is defined as the time rate of radiation of seismic energy in all directions from the source.
- 2) Interpretation of the time series of ground velocity amplitude in terms of the power spectral density of the seismic power of the source.

Power spectral density of a time series $X(t)$ is defined as the frequency distribution of its squared value $X^2(t)$ (Bendat and Piersol 1966 p 79).
- 3) Interpretation of temperature measurements in crater lakes in terms of the thermal power of the volcano.
- 4) The estimation of Q , the quality factor of the medium in which vibration is generated and propagated.

- 5) Determination of the Ishimoto - Iida coefficient (m), which describes the relation between size and frequency of occurrence of earthquakes.

The methods will be described in this order, which is the order of importance in this dissertation.

Seismic Power

The concept of the seismic power of a vibration source is closely related to the concept of the magnitude of an earthquake as defined by Richter (1935). In fact seismic power could be defined as the rate of accumulation of magnitude. Richter developed the magnitude scale as an easily obtained measure of the importance of the earthquake as an energy source. He used only the distance and the maximum amplitude on a Wood Anderson seismograph (two components if available), and from an empirical graph of amplitude versus distance, the amplitude expected at 100 km was obtained. Magnitude was originally defined as the logarithm of this amplitude in micrometres. It turned out to be simply related, at least in elementary theory, to the energy radiated as

seismic waves from the source of the earthquake, but it only works because the earthquakes at their source are brief compared with the recorded duration of the seismic waves, and because there is sufficient similarity between earthquakes, and sufficient homogeneity in the structure of the earth.

It is evident that for volcanic earthquakes and tremors which cover an indefinite range of duration and form, that these latter have to be taken into account in determining seismic energy, and that for this purpose, and in order to follow the progress of volcanic activity, the seismic power, or seismic energy released per unit of time must be determined. This has been done by Yokoyama (1956, 1957a), Gorshkov (1961), Dibble (1965, 1966, 1969), Latter (1967), Gorshkov and Kirsanov (1968), Shimozuru (1968), and Wada, Kamo and Sudo (1970). The term seismic power had been used previously by Evison (1957) in connection with the power radiated by vibrators used as seismic signal sources.

In deciding on a suitable unit of seismic power, it was noted that the arbitrary nature of Richter's unit of magnitude has been both its weakness and its strength;

a weakness because the first Richter magnitude was of no more value than that on any other magnitude scale, so that more than 10 years elapsed before Richter's scale became universally adopted. A strength because the values on the scale have not had to be much changed as knowledge of propagation effects has increased. The relation between magnitude M and energy E (ergs) however has been changed considerably. The formula $\log E = 12 + 1.8 M$ (Gutenberg and Richter 1954) gives E for the largest earthquakes 1,000 times larger than Richter's (1958) formula $\log E = 11.4 + 1.5 M$.

Knowing the difficulties of acceptance of arbitrary scales, and the lack of credibility of constantly changing scales, the author has aimed for a simple uniform method of determining nominally absolute values of instantaneous seismic power (averaged over about 5 s) in the SI unit of power, the Watt, which is a rate of energy transfer of 1 Joule per second, or 10^7 erg per second. When instantaneous values are disadvantageous, average powers in Watt over longer periods are preferred to units such as Joule per day, which are not SI units.

The methods of calculating seismic power investigated by the author are essentially those described by Bullen (1947) for measuring the seismic energy of earthquakes. The energy per unit volume (energy density) in the travelling wave which passes through unit area parallel to the wave front at the observation point each second is the areal power density and is integrated over the total area of the wave front and the duration of the wave to get the seismic energy of the source. To get the seismic power of the source, the integration over time is omitted. The areal power density is:

$$Pd = \frac{1}{2} \rho c V^2 \quad \text{W m}^{-2} \quad (3.1)$$

where ρ is the density of the medium, kg m^3 .

c is the velocity of propagation, m s^{-1} .

V is the peak particle velocity, m s^{-1} .

This equation applies to any travelling sinusoidal plane wave, and will be sufficiently accurate for moderately curved wave fronts. To find the seismic power of the source, we must define the shape of the wavefront, and the type of seismic wave. Three categories are necessary:

- 1) Spherical propagation of body waves.
- 2) Cylindrical propagation of body waves.
- 3) Surface waves.

Most volcano-seismic events involve all three, but in the routine measurement of seismic power it is not feasible to separate waves arriving by different propagation methods and at different wave velocities, and a single category and formula must be selected for all. The equations for each category are as follows:

1) Spherical propagation.

As we are dealing with very shallow seismic sources, integration of areal power density over a hemisphere of radius equal to the source distance (d) is appropriate. For a uniform half space model of the earth, the areal power density is uniform and the seismic power (P_s) of the source is:

$$P_s = \pi \rho c d^2 v^2 \quad W \quad (3.2)$$

For a deep source beneath the observation point, (d) is the ray path distance and the result is half the correct value because spherical rather than hemispherical propagation is involved.

2) Cylindrical propagation.

If most of the seismic energy is confined by reflection and refraction to a uniform surface layer of depth H , much less than the distance between source and seismograph, the areal power density is approximately constant

over the cylindrical area, and with sufficient accuracy the seismic power is:

$$P_s = \pi \rho c d H V^2 \quad W \quad (3.3)$$

Errors will be small provided $H \leq d$.

3) Rayleigh wave propagation.

For a homogeneous half space model of the earth, the penetration of Rayleigh wave motion into the half space is approximately equal to the wavelength (λ), and Bullen (1947, p 232) derived a formula for energy in Rayleigh waves which is equivalent to substituting $H = 1.1\lambda$ in formula 3.3 for cylindrical propagation. Replacing λ by cT , and considering also the components of surface ground motion, the seismic power is:

$$P_s = 1.63 \rho c^2 d T V_z^2 \quad (3.4)$$

where V_z is the vertical component of surface ground velocity.

T is the wave period.

As a basis for the selection of a single formula, calculations of the seismic energy of an explosion (Fig 3.1) of 10 kg of gelignite AN 60 fired 5 m deep

Fig 3.1: Seismograms recorded at Chateau (CNZ).

Lower: Explosion of 10 kg of Polar AN 60 gelignite,
3 m deep in Lower Tama Lake (distance 5.3 km) on 20
April 1963.

Upper: Powerful recommencements of tremor at Ruapehu
volcano (distance 9.2 km) on 6 September 1961.

in Lower Tama Lake at a distance of 5.3 km from the Chateau (CNZ) and Tongariro seismograph stations were made.

The recorded data were:

Parameter	First Part of record	Second Part of record
Ground Velocity	$V_z \quad 4 \times 10^{-7} \text{ m s}^{-1}$	$2 \times 10^{-6} \text{ m s}^{-1}$
Wave Period	$T \quad 0.4 \quad \text{s}$	$0.5 \quad \text{s}$
Wave duration	$t \quad 3.5 \quad \text{s}$	$12 \quad \text{s}$

Rock properties along the ray path were:

Density $\rho = 2.3 \text{ Mg m}^{-3}$ (from gravity profiles).

P wave velocity $c = 4.0 \text{ km s}^{-1}$ (from time distance graph of powerful recommencements of tremor in 1964).

L wave velocity $c = 1.2 \text{ km s}^{-1}$ (assumed).

Depth to basement rocks = 2 km (estimated from delay times of 0.5s for shots fired on basement rocks 22 km distant, and residual gravity anomalies of -20 mgal).

Calculations were made for four simple models, as follows:

- (i) Hemispherical propagation of all energy as body waves with the velocity of P.
Result; $E = 3.9 \times 10^4 \text{ J}$
- (ii) Cylindrical propagation of all energy as body waves with the velocity of L.
Result; $E = 4.4 \times 10^3 \text{ J}$
- (iii) Propagation of all energy as Rayleigh waves.
Result; $E = 6.9 \times 10^2 \text{ J}$
- (iv) Hemispherical propagation of the first part of the recorded vibration as body waves with the velocity of P and Rayleigh wave propagation for the second part.
Result; $E = 1.5 \times 10^3 \text{ J}$

The total potential energy in gelignite AN 60 is about $2.6 \times 10^6 \text{ J kg}^{-1}$, (Byerly 1946) and for a 10 kg charge, the potential energy is

$$E_p = 2.6 \times 10^7 \text{ J} \quad (3.5)$$

The seismic efficiency of explosions in water is given by O'Brien (1967) as 0.32% for a charge of 4.5 kg and an upper frequency limit of 10 Hz. A lower efficiency would not be expected for the shot in Lower Tama Lake, because the water depth was adequate and hard rock outcropped at the shot point. Therefore the expected seismic energy is:

$$E_e = 8.3 \times 10^4 \text{ J} \quad (3.6)$$

This value is higher than the energies calculated from the seismogram for any of the models, but it is in best agreement with the seismic energy calculated on the assumption that all energy is propagated hemispherically as P waves. As this assumption is also the simplest and allows calculation of seismic power without knowledge of the vibration frequency, the author has adopted it for all routine calculations, even though Yokoyama (1956, 1957a) adopted the Rayleigh wave formula, and a number of workers (Sassa 1935, Kikuchi 1963, 1964, Kubotera 1963, 1964, Wada and Ono 1965) have produced evidence that volcanic tremor is propagated as Love and Rayleigh waves. Others allow that body waves are important (Sakuma and Nagata 1957, Shimozuru 1966), and Wada et al (1970)

have made assumptions similar to the author's in calibrating their "digital energy accumulator" which records the energy per day of micro-tremors at Aso volcano, Japan.

It is shown later that attenuation of tremor with distance follows the law for hemispherical spreading more closely than that for cylindrical or surface spreading, possibly owing to absorption and scattering of surface waves in heterogeneous volcanic ground, and so the adopted equation should give sufficiently accurate results, irrespective of the method of propagation.

As a further step towards uniformity the values of wave velocity and rock density used have been standardised at 4.0 km s^{-1} and 2.5 Mg m^{-3} respectively and the resulting values are regarded as nominal seismic powers.

It may be argued that values of velocity and density at the seismograph itself should be used, yet obviously the vertical motion in a thin pocket of ash on top of massive lava is not significantly different from the motion of the lava. Part of the energy incident at the lava-ash interface has been reflected instead of causing the motion in the ash to increase. The effect of a pronounced continuous velocity gradient such as a weathered layer is less clear, but O'Brien (1967)

states this tends to lengthen a seismic pulse and slightly reduce its amplitude. It appears therefore that average properties of the rock along the transmission path will result in more accurate values of seismic power than the properties of the ground at the seismometer.

Other factors to be taken into account in determining seismic power to good accuracy are discussed by O'Brien (1967) and Werth (1967) in connection with seismic refraction prospecting. The components of surface ground velocity are functions of the angle of incidence, of wave type, and of plane of polarisation of the incident ground velocity. Knopoff et al (1957) gives families of curves for these functions.

The vertical surface motion of incident P waves varies smoothly with angle of incidence, and is twice the incident motion at vertical incidence, zero for horizontal incidence, and equals the incident motion for incidence angles of $60 - 65^{\circ}$ depending on the value of Poisson's ratio. Normally there is a sufficient gradient of velocity at the surface to ensure that the surface motion is between factors 1 to 2 of the incident motion. The factor varies in the opposite direction to that due to depth of the source. For a source beneath the

the seismograph, the surface motion is twice the incident motion, and the seismic power is increased by a factor of 4, but propagation is spherical instead of hemispherical, reducing the error to a factor of 2. In general, however, the depth of source is less than the epicentral distance and both may be unknown. The distance of the crater must be used for (d), and consequently the seismic powers of sources deep below the crater are likely to be underestimated rather than overestimated, because (d) is underestimated.

Another factor considered was the resonance between seismometer and ground which has been investigated by Bycroft (1957). For a Willmore Mk 1 seismometer, the mass of the seismometer frame is 5 kg, and the radius of the base is 60 mm. Assuming 2.3 Mg m^{-3} and 300 m s^{-1} as the density and shear velocity of the ground, Bycroft's equations and curves show that magnification due to resonance with the ground is less than 1.1 for frequencies less than 100 Hz. In line with Bycroft's recommendation, and to minimise wind vibrations and damage to the installation, the author has always buried seismometers left in the field.

Although the absolute errors in seismic power may exceed an order of magnitude at some sites, relative errors when positions of activity and seismograph are fixed, will be small if the mean square level of vertical ground velocity is recorded by the seismograph, and will be independent of the complexities of wave form and spectrum. This can be done quickly and easily electronically, compared to the eye readings and computations required with conventional seismograms.

A chart (Fig 3.2) enables the seismic power equivalent of a given calibration signal recorded at a given distance from the source of volcano-seismic activity to be read off immediately. A calibration constant has also been calculated for the CNZ seismograph which has a velocity sensitivity of 3.5 mm per micrometre/second between 1 and 4 Hz, and is 9.2 km from Ruapehu volcano. Values of instantaneous seismic power are given by:

$$P_s = 190 A^2 \quad W$$

where A is the peak trace amplitude in mm. This method was used at Ruapehu in 1966 when the author temporarily had no electronic seismographs, and also to obtain correct relative levels of seismic power of Ruapehu

Fig 3.2: Seismic power versus recording distance for values of seismometer EMF (peak), equivalent to the square-wave calibration signals of the 1970 model tape seismograph, when a Willmore Mk 1 seismometer with 500 ohm coil is used. Hemispherical propagation at 4 km s^{-1} in rock of density 2.5 Mg m^{-3} is assumed. The sinusoidal ground velocity equivalent to the square-wave calibration signal is shown.

volcano over the period 1964 to 1969, for which CNZ was the only continuously operating seismograph station.

Power Spectral Analysis

The power spectral density $P(f)$ of a time series $X(t)$ describes the frequency distribution of its squared value $X^2(t)$. It is equivalent to the square of the Fourier spectrum of $X(t)$, and both have the same shape when plotted on logarithmic scales, and can be obtained using either digital or analog techniques. In seismology the most suitable time series for power spectral analysis is that of particle velocity $V(t)$ because the spectral densities $P_v(f)$, are directly proportional to the true power per Hz bandwidth of seismic body waves of frequency (f) , passing through unit area of wave front at the recording site. More simply, and in a form consistent with the author's use of seismic power as a measure of volcano-seismic activity, spectral densities can be expressed as $P_s(f)$, the apparent seismic power of the source, per Hz at frequency (f) , using the formula:

$$P_s(f) = \pi \rho c d^2 P_v(f) \quad \text{W Hz}^{-1} \quad (3.7)$$

where

ρ is rock density, Mg m^{-3}

c is seismic velocity, m s^{-1}

d is ray path distance to source, m

$P_v(f)$ is the power spectral density of particle velocity, $\text{m}^2 \text{s}^{-2} \text{Hz}^{-1}$.

The formula assumes hemispherical propagation of all seismic energy as body waves, and takes no account of frequency dependent transmission. It is thus subject to the same errors as seismic power P_s . P_s is related to the seismic power spectrum measured at the same time and place by

$$P_s = \int_{f_1}^{f_2} P_s(f) df \quad (3.8)$$

where f_2, f_1 are the frequency limits of the data.

Methods of estimating power spectra have been classified by Blackman and Tukey (1958) using the following possible properties of time series $X(t)$:

(i) Physical Phenomena.

a/ Aperiodic. $X(t)$ never repeats or repeats at random intervals.

b/ Periodic. $X(t)$ repeats at regular intervals and has a harmonic spectrum.

(ii) Spectral Frequencies.

a/ Continuous. An infinite number present.

b/ Discrete. A finite number present, giving a line spectrum.

(iii) Time Series Function.

a/ Statistical. Cannot be uniquely determined, except by the time series itself.

b/ Unique. Uniquely determined by a mathematical expression.

It is apparent from the spectra and the sound of speeded up seismic tapes, that volcano-seismic vibrations are normally aperiodic phenomena with continuous frequency distribution and statistical function. These are best treated by the methods of Blackman & Tukey (1958) as though they were stationary random gaussian time series; that is, time series with amplitude probability density

functions independent of time, and representable as $e^{-(\text{quadratic})}$.

The methods have been extensively discussed (Bendat and Piersol 1966, Parzen 1961) and deal with a fundamental problem; namely that the spectral density of a stationary random time series has statistical variations which are functions of both frequency and time. These variations are called spectral instabilities and, as they are of little value, and may be dangerously misleading, they are reduced by smoothing or averaging. Smoothing is performed over both frequency and time, but usually it is minimised over frequency to preserve frequency resolution, and maximised over time, so as to make efficient use of the stationary nature of the time series.

DIGITAL TECHNIQUES

The digital techniques of power spectral analysis, now widely used, involve direct Fourier analysis of pieces of the time series by fast Fourier transform, followed by squaring and smoothing of the Fourier coefficients. This technique is the mathematical equiva-

lent of the analog filter techniques used by the author. However the necessary expressions for spectral instability and resolution were worked out by Blackman and Tukey (1958) with their mathematically easier but indirect technique of power spectral analysis using the auto-covariance function $C(\tau)$.

This is defined by:

$$C(\tau) = \lim_{T_n \rightarrow \infty} \frac{1}{T_n} \int_{-T_n/2}^{T_n/2} X(t) X(t + \tau) dt \quad (3.9)$$

where

T_n is the length of the time series $X(t)$.

t is time from the centre of the series.

τ is the lag or time difference between sampling points, and takes incremental values between $+T_m$ and $-T_m$

T_m is the maximum value of $\tau \leq T_n$

$C(\tau)$ is an even function of τ . That is, it is symmetrical about $\tau = 0$. Its full length is twice that of the time series $X(t)$.

The Fourier transform of $C(\tau)$ into the frequency domain is power per Hz, as a function of frequency f , and is therefore the power spectral density $P(f)$. Because $C(\tau)$ is an even function, the sine transform is zero and we can write:

$$P(\pm f) = \lim_{T_m \rightarrow \infty} \int_{-T_m}^{T_m} C(\tau) \cos 2\pi f \tau \, d\tau \quad (3.10)$$

where

$P(\pm f)$ is the power spectral density at positive and negative frequencies, and is half the spectral density at real positive frequencies.

In practice T_n is finite, and often quite short, and the auto-covariance function is said to be truncated. Also T_m must be considerably less than T_n in order to provide sufficient products for summation at the longest values of τ ; the effective length of record T_n is less than T_n because there are fewer possible products for $\tau = T_m$ than for $\tau = 0$.

The truncated auto-covariance function is called the apparent auto-covariance function $C_{oo}(\tau)$, and its transform is no longer the true power spectra, but is called a "raw" estimate of the spectra, and has spurious positive and negative peaks, especially at low frequencies. When the apparent auto-covariance function is tapered towards its ends by multiplying by $D_i(\tau)$, a suitable even function of τ called the "lag window" (Fig 3.3), its transform is a "refined" estimate of the smoothed spectral density, averaged now over frequency as well as time. Let the apparent auto-covariance function modified by the lag window be defined by:

$$C_i(\tau) = D_i(\tau) C_{oo}(\tau) \quad (3.11)$$

Then the refined estimate of spectral density $P_i(f)$ at positive frequencies is the cosine transform of $C_i(\tau)$:

$$P_i(f) = 4 \int_0^{T_m} C_i(\tau) \cos 2\pi f\tau \, d\tau \quad (3.12)$$

By averaging estimates of $P_i(f)$ from enough pieces of the time series, the instabilities will become negligible.

Fig 3.3: Auto-covariance functions versus the lag τ .
Upper: Apparent auto-covariance function $C_{00}(\tau)$ for a sample time series $X(t)$ of limited length $T_n \geq T_m$.
Centre: Lag window $D_i(\tau)$.
Lower: Tapered auto-covariance function obtained by multiplying values of the two functions above.

The relation to the true (unsmoothed) power spectra $P(f)$ is then:

$$\text{Ave} \left[P_i(f) \right] = Q_i(f) * P(f) \quad (3.13)$$

where $Q_i(f)$, the transform of $D_i(\tau)$, is an even function of f called the spectral window and $*$ indicates convolution in the frequency domain, which is the transform of multiplication in the time domain. More directly, $\text{Ave} \left[P_i(f) \right]$ is the weighted running average of $P(f)$ over frequency when $Q_i(f)$ is the weighting and averaging function. The spectral window is mathematically equivalent to a narrow band filter with a power transfer function $H_i(ff_1)$, in which case:

$$\text{Ave} \left[P_i(f) \right] = \int_0^{\infty} H_i(ff_1) P(f) df \quad (3.14)$$

The spectral window has a width W_e , defined below, which is analogous to the bandwidth of a filter.

Frequency Resolution of Power Spectra

Blackman and Tukey (1958 p 147) define frequency resolution as the minimum frequency separation at which spectral density estimates are effectively independent. They adopt $1/T_m$ Hz as the optimum frequency resolution because for this case, the covariance between the estimates (with a moderately flat spectra) is less than 5% of either variance. The effective frequency resolution is the larger of the optimum frequency resolution and the width W_e of the spectral window. Reducing W_e below $1/T_m$ does not improve the resolution, but causes increased spectral instability.

Spectral Stability

The stability or lack of instability of estimates of power spectral density at a given frequency from a single record can be measured by comparing estimates from many pieces of the same time series, and computing the mean square estimates and the variances (mean square errors) of spectral density. The stability can also be

predicted by probability theory, for stationary random time series having a gaussian distribution, and can be expressed as the number of degrees of freedom of the chi square variate having the same distribution of values about the mean as the spectral estimates.

If $Y_1, Y_2, \dots, Y_k$ are independently distributed according to a standard normal distribution (i.e. a gaussian distribution with zero average, unit variance and unit standard deviation), then

$$X_k^2 = Y_1^2 + Y_2^2 + Y_3^2 + \dots + Y_k^2 \quad (3.15)$$

follows by definition a chi square distribution with k degrees of freedom. The coefficient of variation of X_k^2 , is defined as

$$\frac{\text{R.M.S. Deviation from the Average, of } X_k^2}{\text{Average value of } X_k^2} = \left(\frac{2}{k}\right)^{\frac{1}{2}} \quad (3.16)$$

As k increases, X_k^2 becomes relatively less variable. Blackman and Tukey (1958) give the following table relating k to spectral stability:

% Confidence that estimates   Spread of Values in db
will lie within given spread: for each confidence limit:

60	$10 (k-1)^{-\frac{1}{2}}$
80	$16 (k-1)^{-\frac{1}{2}}$
90	$20 (k-1)^{-\frac{1}{2}}$

Thus if it is required that 90% of the spectral density estimates will lie within $2\frac{1}{2}$ db of the true average value (spread 5 db), the required value of k is 17.

To find k for spectral estimates, we substitute in equation 3.16;

- a) $\text{Ave}[\text{Pi}(f_1)]$ as in equation 3.14;
- b) $\text{RMS deviation of } \text{Pi}(f_1) = [\text{Var } \text{Pi}(f_1)]^{\frac{1}{2}} ;$

Blackman and Tukey (1958) give $\text{Var } \text{Pi}(f_1)$ for a moderately flat spectra as:

$$\text{Var } \text{Pi}(f_1) \doteq \frac{1}{T_n} \int_0^{\infty} \left[\text{Hi}(ff_1) P(f) \right]^2 df \quad (3.17)$$

Then equation 3.16 can be rewritten as:

$$k = \frac{2 \left[\text{Ave} \left\{ P_i(f_1) \right\} \right]^2}{\text{Var} \left\{ P_i(f_1) \right\}} \quad \text{def} \quad \frac{W_e}{\Delta f} \quad (3.18)$$

$$\text{where } W_e \quad \text{def} \quad \frac{\left[\int_0^\infty H_i(ff_1) P(f) df \right]^2}{\int_0^\infty \left[H_i(ff_1) P(f) \right]^2 df} \quad (3.19)$$

is called the equivalent statistical width of the spectral window, and is a function of the shape of the spectrum, and:

$$\Delta f \quad \text{def} \quad \frac{1}{2 T_n} \quad (3.20)$$

is called the elementary frequency band, and is considered to have a single degree of freedom.

If the spectrum is flat, $P(f)$ is constant for all f , and cancels out of the expression for W_e .

$$W_e = \frac{\left[\int_0^\infty H_i(ff_1) df \right]^2}{\int_0^\infty \left[H_i(ff_1) \right]^2 df} \quad (3.21)$$

Although only exact for flat spectra, the expressions for k and W_e are satisfactory approximations in most cases, but do not apply to line spectra. Blackman and Tukey show that $k = 2$ if $P(f)$ is a single sharp peak of width $\ll \frac{1}{Tn'}$ and frequency $\gg \frac{1}{Tn'}$. The low value of k reflects the decreased opportunity for smoothing over frequency and the more obvious beating of frequencies as the peak becomes very narrow.

RELATION OF ANALOG TO INDIRECT DIGITAL TECHNIQUES

To enable a more general discussion, the key points of similarity and difference between direct analog and indirect digital analysis are summarised below.

- (i) The effective record length Tn' and the elementary frequency band $\Delta f = \frac{1}{2 Tn'}$, have similar significance in both methods.
- (ii) The maximum lag Tm in the auto-covariance function has no analog equivalent, and the optimum frequency resolution $1/Tm$ must be approximated by $1/Tn'$ in the analog method.

(iii) The equivalent statistical width W_e is applicable to both methods, and can be calculated from equation (3.21) using the digital power transfer function of the window $H_i(f\Delta t)$ or the analog power response of the filter (i.e. voltage response $M(f)$ squared).

(iv) The stability of estimates, expressed as $k = \frac{W_e}{\Delta f}$ degrees of freedom is applicable to both methods.

(v) The process by which absolute values of spectral density are obtained, differs for the two methods:

The indirect digital method gives spectral density in mean square value of X (for time series $X(t)$) per Hz directly, because the lag window and spectral window are smoothing functions which preserve the units. For instance the 'hamming' spectral window weights, 0.23, 0.54, 0.23, applied at successive frequency increments, total unity.

In the direct analog method, the filter window has approximately unit height but a variable effective bandwidth B_n , defined as the width of

an ideal rectangular window of unit height, and area equal to that under the response curve of the filter. The power transferred by the filter is the power in bandwidth B_n , and if the analyser integrates this over time T , then the spectral density is:

$$P_i(f) = \frac{1}{B_n T} \int_0^T X^2(t, f, B_n) dt \quad \text{power Hz}^{-1} \quad (3.22)$$

where $X(t, f, B_n)$ is the filtered time series.

B_n can be calculated (Bendat and Piersol 1966 page 265) from:

$$B_n = \frac{\int_0^{\infty} [M(f)]^2 df}{M_m^2} \quad (3.23)$$

Where $M(f)$ is the voltage response function of the filter, and M_m is the peak response.

Bias Errors

At dips or peaks in the spectrum, the averages of refined estimates of spectral density may not converge on the true spectral densities due to the averaging over frequency. The differences are called bias errors. Bendat and Piersol (1966) give the bias error b at frequency f , as:

$$b = \text{ave} \left[P_i(f) \right] - P(f) \div \frac{W_e^2}{24} \cdot P''(f) \quad (3.24)$$

Where $P''(f)$ is the second derivative with respect to frequency of the true spectral density $P(f)$ at frequency f - a difficult quantity to measure.

The best way to avoid bias errors, and also errors due to side lobes of spectral windows is to pre-whiten (flatten) the spectrum by passing it through known filters of approximately inverse shape to the spectrum before performing the spectral analysis. However, if the spectrum is complex and changeable this is impracticable. Forlifer (1964) has shown that if the filter bandwidth is $\leq \frac{1}{4}$ the bandwidth of the narrowest peak

in the spectrum, the bias error will be less than 0.2 db or 3%. However, some of the spectra in this thesis show peaks of width comparable with the filter bandwidth. For the idealised case when the shapes of both the analysing filter response and true spectral density of a time series are identical isosceles triangles, it is easily shown that the bias error in the estimated peak spectral density is 1.8 db or 33% low.

When a harmonic calibration signal is analysed, however, the true power spectrum consists of one or more lines of finite power but of nominally infinite power density, because they have nominally zero bandwidth. The analysing filter (bandwidth B_n) passes all the power, N watts, in the spectral line on which it is centred and indicates a spectral density.

$$P_i(\text{cal}, f) = \frac{N}{B_n} \quad \text{W Hz}^{-1} \quad (3.25)$$

This enables correct calibration, even though the bias error is infinite in this case.

Analysis of Non-Stationary Time Series

Non-stationary time series are those with amplitude probability density functions which vary with time.

The types of non-stationary behaviour which are common in volcanic seismology are as follows:

- (i) Rapid changes in the seismic power of tremor, which are not synchronous at all frequencies.
- (ii) Sudden changes in power level and spectral shape as in earthquakes.
- (iii) Step shifts in the frequency of spectral peaks.
- (iv) Gliding in frequency of spectral peaks.

The effects of these on estimated spectra can be discussed under two headings:

- a/ The effects on true spectral density.
- b/ The effects on estimates of spectral density.

The effects on true spectral density can be qualitatively described following Peterson and Dobson (1966). Types (i), (ii), and (iii) of non-stationary volcanic vibrations can be considered to consist of individual frequency bands truncated at individual times.

The truncation causes broadening of each band and addition of side lobes, in amounts varying with the sharpness of truncation. Thus the spectrum of an impulsive onset is smoother than that of the vibrations following the onset.

The effects on estimates of spectral density in analog measurement of time-averaged power spectra are discussed by Bendat and Piersol (1966 p 361). In this technique the output of the analysing filter is squared and averaged in an integrating device before being recorded. This is described by:

$$P_i(f,t) = \frac{1}{B_n T} \int_0^T x^2(t,f,B_n) dt \quad (3.26)$$

where T is the averaging time.

The errors in a single estimate of time-averaged power spectra are statistical errors and bias errors as for stationary spectral analysis, and in addition a time-interval bias error, defined as the difference between the true non-stationary spectra at a given instant and the spectral estimates averaged over an interval including that instant. The time-interval bias error can be reduced

by choosing parts of the time series which are approximately stationary, or by reducing T , but nothing is gained by reducing T to less than the filter rise time, which is approximately $\frac{1}{B_n}$. The Sona-graph and Rayspan analysers used by the author are both of the time-average type, with T equal to the sum of the filter rise and ring times. Bendat and Piersol (1966 p 364) show that time-averaged power spectra and auto-covariance functions of non-stationary time series are related by the Fourier transform, and state:

"Hence one is justified in employing results available for stationary processes in studying the time-averaged power spectrum".

The time resolution of time-averaged power spectra, defined as the minimum time interval between independent estimates of spectral density at a given frequency, is T second for true integration averaging. For the time resolution with RC averaging the author has adopted

$$\text{Time resolution} = 3 T_c \quad \text{s} \quad (3.27)$$

where T_c is the time constant, because after this time the effect of any change in true spectral density will have decayed to 5% of its initial value. For filter rise time averaging, the author has adopted

$$\text{Time resolution} = \frac{3}{B_n} \quad \text{s} \quad (3.28)$$

CHARACTERISTICS OF THE SONA-GRAPH ANALYSER

This machine was described in Chapter 2. Its characteristics as an analyser are given in Tables 3.1 and 3.2.

When producing a qualitative display of spectral density versus time and frequency, the machine marks the paper with density averaged only over the rise or ring time of the filter. This is the effective record length T_n' as far as optimum frequency resolution and stability of the display is concerned, and can be estimated from sonagrams of switching transients.

The actual frequency resolution is the equivalent statistical bandwidth W_e of the filter, which is greater than the optimum resolution. It and the equivalent bandwidth B_n were evaluated by summation of the power response curve of the filter over 20 Hz intervals. The values of time and frequency given in Table 3.1 apply to Sona-graph record time, and must be altered by the speed-up factor of the seismic tape recording if they are to apply in real time to analyses of volcanic vibrations.

Frequencies must be divided and times multiplied by the speed-up factor. For instance, with speed-up 480, and narrow band analysis we have in real time:

Frequency resolution	$W_e = 0.19 \text{ Hz}$
Effective bandwidth	$B_n = 0.13 \text{ Hz}$
Time resolution	$= 23 \text{ s}$

The latter is the minimum time between independent values of spectral density, but successive seismic phases arriving 9 or more seconds apart could be recognised.

TABLE 3.1: Time and Frequency Characteristics for Qualitative Analysis with 1955 Model Sona-Graph

Characteristic	Equation No.	Data Source	Measured Value	
			Narrow Band Analysis	Wide Band Analysis
Nominal bandwidth	-	Sona-graph Handbook	45 Hz	300 Hz
Tn' effective record length	-	Ring time for transients	0.02 s	0.01 s
Δf , elementary frequency band	3.20	$1/2Tn'$	25 Hz	50 Hz
Optimum frequency resolution		$1/Tn'$	50 Hz	100 Hz
W_e , actual resolution and equivalent statistical bandwidth.	3.21	Filter response, Handbook	93 Hz	530 Hz
Bn, effective bandwidth	3.23	Filter response, Handbook	62.5Hz	417 Hz
Time resolution qualitative display.	3.28	$3/Bn$	48 ms	7.1 ms

TABLE 3.2:

Spectral Stability and Time Resolution of the 1955 Model Sona-graph Section Display for Different Values of Added Resistance R in the RC Averaging Circuit, and for 45 Hz (N) and 300 Hz (W) Bandwidth Analysis.

Resistance Added Megohm	Time Resolution at Sona-graph Record Speed		Number of Degrees of freedom		Statistical Error, with 90% confidence	
	N	W	N	W	db	db
0	60	30	3.7	10.6	6.1	3.2
0.14	80	50	5.0	18	5.0	2.4
0.28	100	70	6.3	25	4.3	2.0
0.57	160	130	10	46	3.3	1.5
1.15	270	240	17	86	2.5	1.1
1.7	380	350	24	126	2.0	0.9
2.0	440	410	27	156	1.9	0.8

For the section display, the figures for frequency resolution in Table 3.1 still apply. Values for the time resolution, and the stability of the quantitative spectral densities are given in Table 3.2 for different values of the resistor R (which the author added to increase the time constant of the RC circuit for time averaged spectra), and for narrow and wideband analysis.

The averaging time T_c (referred to the recording time of the Sona-graph) which is also the effective record length T_n' , is taken as:

$$T_c = T_n' = 3.33 RC + 0.02 \text{ s for } 45 \text{ Hz bandwidth} \quad (3.29)$$

$$= 3.33 RC + 0.01 \text{ s for } 300 \text{ Hz} \quad " \quad (3.30)$$

where $C = 0.02$ microfarad is the capacitor in the averaging circuit.

The number of degrees of freedom k , and the time resolution were calculated using equations 3.18, and 3.27.

Table 3.2 shows that the Sona-graph as modified can produce spectra of high stability. Most of the sonagram sections were made before the modification became possible, and these have very modest stabilities. However, the background qualitative display and the multiple sections on each sonagram usually allow a section of better than average accuracy to be chosen.

Comparison between sonagrams made in Hawaii and New Zealand show that in frequency resolution and bias errors the Hawaii narrow bandwidth sonagrams are superior, because the crystal filter of nominal 45 Hz bandwidth was better than average. The spectral peaks were usually wider than the filter bandwidth and as shown above, the bias errors are then less than about 2 db. On the New Zealand sonagrams and the Hawaii wideband sonagrams, the spectral peaks were often narrower than the filter bandwidth, and bias errors may occur up to an estimate of 18 db at the frequencies of the narrowest peaks and troughs in the spectrum.

Calibration of Sonagram Sections

The Sonagrams for Hawaii and Stromboli were calibrated in units of $P_s(f)$, the nominal seismic power spectral density, using the 3rd and 9th harmonics (1.5 and 4.5 Hz) of the square wave calibration signal. The power in these harmonics is -10.5 db and -20 db respectively of the seismic power taken from Fig. 3.2. Dividing the seismic power in each harmonic by the effective analysis bandwidth B_n for real time seismic recording the power spectral densities of the harmonics in $W Hz^{-1}$ are obtained. These were reduced by the heights of the peaks on the sonagram, averaged, and rounded to obtain the nominal seismic power spectral density in $W Hz^{-1}$ corresponding to zero db on the spectral profiles at mid-band frequencies. Where a seismograph calibration curve for a flat seismic power spectrum is shown on the profile, its level in $W Hz^{-1}$ at 4 Hz can be obtained from the profile, and measurements at other frequencies made relative to it.

In most other cases where it has been possible to put scale values of nominal seismic power spectral densities on the sonagram profiles, this has been done by first assuming that zero db corresponds to $1 W Hz^{-1}$. The profile

is integrated to find the total power (equation 3.8), which is compared with the chart of seismic power, and the assumed zero db value is then scaled in order to make the two powers equal.

The values of nominal seismic power spectral density $P_s(f)$ suffer the same sources of uncertainty as values of seismic power, and in addition, the errors in power spectral analysis. This total could exceed 16 db. By dividing by $\pi \rho c d^2$ (equation 3.7), values of $P_v(f)$, the power spectral density of ground velocity in $m^2 s^{-2} Hz^{-1}$ are readily obtained with less uncertainty, but also with less significance.

Thermal Power of Ruapehu

The Ruapehu Crater Lake can be regarded as a water calorimeter with which to measure the heat output of the volcano. The lack of hot springs, steaming ground or even snow-free ground attributable to volcanic heat anywhere except on the narrow shore of the lake shows that almost the entire heat flow is released from the lake surface and can be calculated from lake temperature

measurements as described by Dawson (1964). Considering heat lost by evaporation, conduction, and radiation, but not by overflow, we have:

$$F_s = F_e (1 + R) + F_b \quad W \quad (3.31)$$

where F_s is total heat loss from the calm lake surface

F_e is heat loss by evaporation

R is Bowen's ratio

$$= \frac{\text{Heat loss by conduction \& diffusion}}{\text{Heat loss by evaporation.}}$$

F_b is net loss by radiation.

These can be calculated using the following formula:

$$F_e = 1.782 LA (0.031 + 0.0033 W^{0.76} D^{-0.12}) \times (p_s - p_d) (101325 p_1^{-1}) 10^2 \quad W \quad (3.32)$$

$$R = 6.1 p_1 (T_s - T_a) (p_s - p_a)^{-1} \times 10^{-4} \quad (3.33)$$

$$F_b = 0.955 C (T_s^4 - T_a^4) A \quad W \quad (3.34)$$

where

L is the latent heat of water (2.406 MJ kg^{-1} at 40°C)

A is the area of lake, ($200,000 \text{ m}^2$)

D is diameter of lake, (500 m)

W is wind speed (m/s)

p_s is vapour pressure of water (Pascal) at temperature of lake surface.

p_d is pressure of water vapour in atmosphere (Pa)

p_l is atmospheric pressure (Pa)

p_a is vapour pressure of water (Pa) at temperature of atmosphere.

T_s is absolute temperature of lake surface (K)

T_a is absolute temperature of atmosphere (K)

C is the Stefan-Boltzmann constant = 5.67×10^{-3}

This follows Dawson (1964) except that the term involving wind velocity W has been changed from $0.0135W$ to $0.0033 W^{0.76} D^{-0.12}$, where D is the diameter of the pool. The original term was correct for water in small pans, but not for large lakes because the time for a complete change of air over the lake is then large, and the air can become saturated during its passage. The new term results in a calculated rate of evaporation which agrees with Dawson's measured value for a pool of 2.5 m diameter with 1.5 m s^{-1} wind velocity, and the powers of W and D are as given by Penman (1948).

The meteorological data needed in the calculation (Table 3.3) were estimated by the New Zealand Meteorological Service for free air conditions at 2515 m altitude in the position of Ruapehu. The presence of

the mountain would increase the wind speed and possibly lower the air temperature, but Crater Lake is in a sheltered position, and conditions there appear to be similar to those estimated for free air.

TABLE 3.3   Estimated Meteorological Conditions in Free Air for the Position of Ruapehu Crater Lake.

	Mean Annual	Mean January	Mean July
Mean Air Pressure	74592 Pa		
Mean Air Temperature	+0.9°C	+6.2°C	-3.5°C
Mean Wind Speed	12.0 m s ⁻¹	11.3 m s ⁻¹	11.7 m s ⁻¹
Mean Relative Humidity	45%	42%	48%

Calculations have been made for four different meteorological conditions: Mean annual, Mean January, Mean July, and for zero wind speed with mean annual conditions otherwise, and a constant lake temperature of 40°C (Table 3.4). It can be seen that the heat loss is not strongly dependent on the meteorological conditions. The vapour pressure of water at the temperature of the lake surface is the controlling factor. Calculations were made for mean annual conditions at a range of lake temperatures and plotted in Figure 3.4. This curve is used in Chapter 8.

TABLE 3.4:

Calculation of Heat Loss from Ruapehu Crater
Lake at 40°C for Mean Annual, Mean January and
Mean July Conditions.

Quantity, Units	Mean Annual Conditions	Mean January Conditions	Mean July Conditions	Mean Annual but zero wind speed
P_s	7377	7377	7377	7377
P_d	293.3	398.0	226.5	293.3
$P_s - P_d$	7084	6979	7150	7084
P_a	651.8	947.5	471.7	651.8
$P_s - P_a$	6725	6429	6905	6725
$T_s - T_a$	39.1	33.8	43.5	39.1
$T_s - T_a$	3.9697×10^9	3.5213×10^9	4.3228×10^9	3.9697×10^9
$.0033 W^{0.76} D^{-0.12}$	0.01009	0.00964	0.0099	0
F_e	459	447	461	345
R	0.2643	0.2392	0.2866	0.2643
$F_e (1 + R)$	580	554	594	437
F_b	43	38	47	43
F_s	623	592	641	480

Values in this table have been taken from the Smithsonian Meteorological
Tables (1951).

Fig 3.4: Heat loss from Ruapehu Crater Lake versus near surface lake temperature. Evaporation, conduction and radiation are included, but not overflow or seepage.

Determination of Q

For a sinusoidal wave train propagating through a medium, Q is 2π times the ratio of the peak elastic energy stored in unit volume, to the energy dissipated in unit volume per cycle. It is the same quantity as the quality factor in electrical circuit theory, and is related to the logarithmic decrement δ by:

$$Q = \pi/\delta \quad (3.35)$$

where δ is defined as $\ln \frac{A_1}{A_2}$ and

A_1, A_2 --- are the amplitudes of successive crests (or troughs) of an oscillatory decay train.

The decay train can be taken as either decay with time at a given point, or decay of a continuous propagating plane wave with distance. An expression for the latter is given by Knopoff and MacDonald (1958), which views the wave train as it passes (x constant), or while travelling with it ($kx - \omega t$ constant).

$$U = A_0 e^{-\alpha x} \cos (kx - \omega t + \beta) \quad (3.36)$$

Where

- U is the displacement at distance x and time t
 A_0 is the initial displacement.
 α is the dissipation coefficient per unit distance at frequency ω .
 x is the angular distance from the initial displacement.
 k is the wave number ($2\pi/\text{wave length}$)
 ω is the circular frequency
 β is the initial phase angle.
 t is time

Now for crests at x_2 and x_1 a wavelength apart at a given instant:

$$\begin{aligned}
 x_2 - x_1 &= \lambda \\
 &= \frac{2\pi c}{\omega}
 \end{aligned}$$

where c is the wave velocity;

λ is the wave length.

The ratio of amplitudes of the crests is:

$$\begin{aligned}
 \frac{U_1}{U_2} &= \frac{e^{-\alpha x_1}}{e^{-\alpha x_2}} = e^{\alpha(x_2 - x_1)} \\
 &= e^{2\pi\alpha c/\omega}
 \end{aligned} \tag{3.37}$$

$$\text{Therefore} \quad \delta = \ln e^{2\pi\alpha c/\omega} = \frac{2\pi c\alpha}{\omega} \quad (3.38)$$

$$\text{and} \quad Q = \pi/\delta = \frac{\omega}{2c\alpha} \quad (3.39)$$

Knopoff and MacDonald (1958) have reviewed the experimental data on dissipation of acoustic and seismic waves in solids, and find that for many rocks and solids other than ferro-magnetic materials and certain organic plastics, the specific dissipation function $1/Q$ is independent of frequency. Then α is proportional to frequency and the dissipation is a constant per wavelength:

$$\text{Therefore} \quad \alpha = qf \quad \text{m}^{-1} \quad (3.40)$$

$$\text{and} \quad \alpha\lambda = qc \quad (3.41)$$

where q is defined as the dissipation coefficient per unit distance and frequency.

Knopoff and MacDonald show that the result cannot be accounted for by any plausible linear mechanism of attenuation. They suggest frictional dissipation as the cause of the non-linearity. In fluids on the other hand, α is approximately proportional to frequency squared, and Q is approximately inversely proportional to frequency. This may be the case for magma.

At least four methods are available by which Q can be calculated from seismological data.

- (i) The Q of the rock along the transmission path can be determined from the dissipation coefficient found from attenuation with distance of a seismic wave, provided the propagation law, velocity, and frequency of the wave are known. The relation is of the form:

$$A = A_0 x^{-n} e^{-\alpha x} \quad (3.42)$$

where n is 0, 0.5, 1, or 2 depending on whether propagation is plane, two dimensional, spherical, or by critical refraction.

$$\text{then } Q = \frac{\pi f}{c\alpha} \quad (3.43)$$

The author has not used this method because c and n are both difficult to determine for volcanic tremor.

- (ii) The Q of the rock along the transmission paths can be found from the change in the spectrum with distance. From equation 3.42 the attenuation in db of a plane wave is

$$20 \log_{10} \frac{A_0}{A} = 20\alpha x \log_{10} e \quad \text{db} \quad (3.44)$$

Using equation 3.40

$$20 \log_{10} \frac{A_o}{A} = 8.6863 \text{ xfq db} \quad (3.45)$$

Therefore when Q is independent of frequency the attenuation in db due to dissipation increases linearly with distance and frequency. At zero frequency there is zero absorption. Attenuation due to geometrical spreading, on the other hand, increases at the same rate at all frequencies.

The effect on profiles of spectral density (db) versus frequency on a linear scale is easily seen if the spectrum of the vibration at its source has a flat profile shape.

With increasing distance the profile shape tilts increasingly in the direction of higher frequencies, but remains a straight line. The change in slope of the profile, measured in $\text{db Hz}^{-1} \text{m}^{-1}$ is the attenuation due to dissipation and is equal to $8.686q$, irrespective of whether geometrical spreading has occurred or not.

$$\text{Therefore } q = \frac{\text{change of slope in db Hz}^{-1} \text{m}^{-1}}{8.686} \quad (3.46)$$

$$\text{and} \quad Q = \frac{\pi}{c\eta} = \frac{\pi}{\alpha\lambda} \quad (3.47)$$

Knowledge of c but not of n is required. Results using the method are given in Chapter 10.

- (iii) A maximum value of the Q of the vibration source can be determined from the logarithmic decrement (δ) of the decay train of a seismic event. As the Q of the material, the Q of the mechanism, and the propagation conditions are involved, a wide range of interpretations could be put on values obtained in this way. A stick-slip earthquake in very high Q material could give a low value of Q if frictional dissipation across the fault were high. On the other hand a dike-splitting earthquake in the same material could show high Q if wedging and gas pressure were to pull the split surfaces apart and allow them to ring. Fig. 3.5 allows Q to be found from the damping ratio ϵ between amplitudes of successive half cycles.

Fig 3.5: Quality factor Q versus damping ratio ϵ . $Q = \pi/\lambda$, $\lambda = 2 \log_e \epsilon$.

(iv) The Q of a vibration generator or filter can be found from the width of the spectral peak that it produces. A well known result (Terman 1943 p 138) is:

$$Q = \frac{\text{Peak Frequency (Hz)}}{\text{Width of peak (Hz) } 3 \text{ db below peak level}} \quad (3.48)$$

Q determined in this way has been used in the description of the spectra of volcano-seismic vibrations.

Ishimoto-Iida Coefficient (m).

Two relationships between the occurrence of earthquakes and their size have been used by the author.

The first, due to Ishimoto and Iida (1939), relates the number of earthquakes in unit amplitude interval with the amplitude (the density of occurrence with respect to amplitude), for a given time period, and is

most clearly written as:

$$\frac{N(A \pm \frac{\delta A}{2})}{\delta A} = kA^{-m} \quad (3.49)$$

where A is the earthquake amplitude (usually recorded amplitude);

δA is the amplitude interval actually used;

$N(A \pm \frac{\delta A}{2})$ is the number of earthquakes within interval δA centered on Amplitude A , in a given time;

k is proportional to the total number of earthquakes in the period;

m is the coefficient defining the size-occurrence relation.

The author has used this relationship when the data comprises recorded amplitudes. Coefficient (m) was determined from the slope of a log-log graph of density of occurrence against amplitude. In general, (m) has

values between 1.5 and 2.5. Values of m above 2.5 indicate an excess of small earthquakes compared with the normal, and may be diagnostic of earthquakes leading to an eruption (Sakuma and Nagata 1957).

The second relationship is due to Gutenberg and Richter (1954) and relates the number of earthquakes (N) per year, above a given magnitude (M), with the magnitude. It is best known by:

$$\text{Log } N = a + b (8 - M) \quad (3.50)$$

where a is a constant specifying the seismicity;

b is a constant defining the size - occurrence relationship.

The author has used a more convenient form given by Richter (1958):

$$\text{Log}_{10} N' = a' - b M \quad (3.51)$$

where N' is the number of earthquakes per unit magnitude interval, and is therefore the density of occurrence with respect to magnitude M ;

b is the same constant as in (3.50);

a' is the new constant specifying the seismicity.

The Richter relation is convenient when earthquakes are recorded as spikes against a linear db scale of vibration level, because $M = \log_{10} A$ minus a constant, when the recording distance is constant. Thus 20 db is one unit on the magnitude scale. The author has determined b from the slope of the graph of number of earthquakes in a constant db interval against the db level, using log scales for number of earthquakes and a linear db scale for amplitude level.

The relation between m and b can be found by integrating the density of occurrence given by the Ishimoto-Iida relationship (3.49) to find the number of earthquakes N'' of amplitude $\geq A$.

$$N'' = \int_0^{\infty} \frac{N(A \pm \frac{\delta A}{2})}{\delta A} da = (1 - m) k A^{-(m-1)} \quad (3.52)$$

Taking logs:

$$\text{Log } N'' = \log (1-m) k - (m-1) \log A \quad (3.53)$$

This is of the form

$$\text{Log } N'' = a'' - (m-1)M$$

which is similar to (3.51)

$$\text{therefore} \quad b = m-1 \quad (3.56)$$

Values of m are determined in Chapters 4, 7 and 10.

CHAPTER 4

RUAPEHU VOLCANO 1959 - 1964

Introduction

Ruapehu is an active andesitic strato-volcano (fig. 4.1) near the centre of the North Island of New Zealand.

It rises 2797 m above sea level, and 1900 m above its surroundings, and stands at the southern end of the chain of volcanoes along the Tonga - Kermadec - New Zealand island arc structure. Ruapehu originated during the Quaternary period. The lower Pliocene marine sediments it overlaps contain no volcanic material (Fleming and Steiner 1951) but early Quaternary conglomerates south of Ruapehu contain andesite pebbles. The activity of Ruapehu prior to 1960 has been described by Gregg (1960). Two important historic eruptions have occurred: an ash eruption in 1861 and an ash and lava eruption in 1945. In the latter, a tholoid entirely displaced the water in the Crater Lake. By the end of the eruption the tholoid had been destroyed by periodic explosions,

Fig 4.1: View of Ruapehu Volcano from the east, during a brief hydrothermal eruption from the crater lake at 1105 NZST on 4 June 1969. The seismic power of this event was 50 kW. The height of the eruption above the crater lake is 700 m. A base surge can be seen.

Photo by M. Aldred.

leaving a hole about 300 m deep which filled with water to form a lake remarkably similar to the one which had existed before. Numerous other minor eruptions of steam, ash, and water from the crater lake have also occurred, so that the volcano was generally thought of as interesting but not seriously dangerous. The normally warm crater lake of area 0.2 km^2 and altitude 2510 m "acts as a steam condenser, and hides as it were to some extent the activity of Ruapehu" (Friedlander 1898). The lake also acts as a calorimeter, chemical collecting apparatus, and measuring platform, and allows the integrated volcanic activity to be easily measured. Its temperature has varied from freezing to at least 58.4°C . The lake and its outlet tunnel have been described by O'Shea (1954) and Odell (1955). It has been responsible for numerous lahars, including that on 24 December 1953 which swept away the Tangiwai railway bridge minutes before a train dived into the icy torrent, causing the loss of 151 lives.

Observations, 1959 - 1963

After the Chateau Volcanological Observatory of Geophysics Division D.S.I.R. was built and equipped with a Willmore seismograph (code name CNZ, site 1, fig 4.2) in November 1959, volcanic tremor of low amplitude, and dominant frequency 2 Hz began to be recorded about 10% of the time. Enhanced volcano-seismic activity occurred between August and December 1961. In August, the Crater Lake was unusually warm (estimate 43°C) and was bubbling vigorously around the edge (Mr. Christopher Bell, personal communication). Volcanic tremor was more incessant, and was recorded at sites 1, 2, 3, 5 and 16. When the slow motion tape seismograph recordings from these sites (except 5) were played back at about 400 times recording speed, a warbling roar (volcanic tremor) was audible which would occasionally stop suddenly, and after a silent period of up to 15 minutes real time, would recommence with a single bang. The author refers to these as powerful re commencements of tremor. Their speeded up sound is characteristic and unique, and different from tectonic earthquakes or the A-type volcanic earthquakes of Minakami (1961) which have two bangs

Fig 4.2: Map of the Tongariro National Park volcanoes, showing the seismograph locations (numbered), the epicentre of powerful recommencements of tremor in 1964, and the distance arcs from the seismographs.

(P and S), but similar to the seismic explosion (fig 3.1) which was fired in Lower Tama Lake (fig 4.2) to test the seismograph equipment.

Volcano-seismic activity at Ruapehu was low in 1962 and 1963 and apparently zero for considerable periods. Figure 4.3 shows power spectra of ground velocity at the Chateau obtained by Bradner et al (1964) in September 1963. At that time, ground noise power density was a minimum in the frequency range of volcanic tremor. The high densities below 1 Hz were caused by ocean micro-seisms. The peak at 6 Hz is probably the Chateau generating plant.

Observations in 1964

VOLCANIC ACTIVITY

From 20 April 1964 onwards, increased fumarole and hot spring activity was obvious at the northwest shore of Ruapehu Crater Lake. Several metres off shore a 1 m high fountain of hot gassy water was sometimes playing, and during May the lake surface was agitated even when there was no wind. The lake temperature at the shore rose from a fairly average value of 22-28°C on March

Fig 4.3: Profiles of power spectral density at the Tongariro seismograph station (site 5) in September 1963 when the volcanoes were quiescent. High densities below 1 Hz are caused by sea microseisms and instrument noise.

FIG. 4.3 The solid line indicates $.01 (\mu/\text{sec})^2/\text{cps}$. The unlabelled ordinate scales show tic marks every 10 db. (After Bradner et al, 1964)

1964 to 41-54°C on 26 May (J. Healy, personal communication), corresponding to an increase in thermal power output from 200 to 700 MW. From 6 May until at least 4 September, steam was visible above the volcano on those occasions when the top of the volcano was visible, and the relative humidity not too low.

VOLCANO-SEISMIC RECORDINGS, 1964

Volcanic tremor of 2 Hz was unusually strong and incessant between April and September, with a climax in May, when the seismic power reached peaks of 20 kW. Figure 4.4 is a chart of seismic power of tremor constructed from amplitudes read from CNZ seismograms until 25 April, and then from log scale chart recordings of playback level of the seismic tape recordings. The vertical downward lines are stoppages before powerful recommencements of tremor (not all present due to slow chart recorder response). Figure 4.5 shows the recording periods at each seismograph site. Figure 4.6 is a graphical summary of the seismic activity within 50 km of the Chateau. Five phases of activity can be recognised:

Fig 4.4: Seismic power of volcanic tremor from Ruapehu during 1964. Downward lines are stoppages of tremor. Most upward peaks failed to record due to slow pen response. Wind may cause values up to 10 W, but the grassy trace in late July is due to pulsing tremor.

Fig 4.5: Chart of recording periods at the seismograph sites used in 1964.

Fig 4.6: Chart of daily seismic activity within 50 km of site 1 (CNZ) during 1964. A-type earthquakes with prominent P phases are shown by open circles.

1) Preliminary activity.

A swarm of 12 similar small A-type earthquakes, possibly of volcanic significance, occurred between 3 and 7 February. Between 5 and 26 February, 31 isolated periods of volcanic tremor occurred, each of 1 to 10 minutes duration and 0.4-2 kW seismic power. The sonagram of one of these periods (fig 4.7) recorded at site 1 shows a dominant frequency of 2 Hz in a band between 0.3 and 3.5 Hz, and no change in the frequency of spectral peaks as the tremor starts and stops.

2) Period of increasing activity

From 31 March to 28 April the number of hours of tremor a day, and its seismic power increased erratically, and the tremor usually started and stopped gradually.

3) Period of maximum activity.

On 28 April tremor became continuous except for sudden and complete stoppages of up to 20 minute, most of which ended with a powerful recommencement of tremor. The tremor spectrum (fig 4.8) was similar to that of the preliminary isolated periods

Fig 4.7: Sonagram of an isolated period of tremor recorded at site 1 in February 1964.

Fig 4.8: Sonagram of a stoppage of tremor at 7.5 minute, and a powerful recommencement of tremor at 11.8 minutes. The spectra of these are somewhat similar to each other, but different from that of the P-wave of a distant earthquake at 03.7 minute. Recorded at site 1 in May 1964.

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of tremor but the spectrum of the powerful recommencements extended from 0.2 to 7 Hz with maximum density in the range 1.5 to 3 Hz. Maximum activity occurred on 4 May when there were 12 stoppages and powerful recommencements of tremor.

Many of the powerful recommencements of tremor had remarkably similar ground motion at any one seismograph as shown in Fig 4.9. However, not all tremor began with identical powerful recommencements during the period of maximum activity. Almost an equal number started gradually or with a formless burst of strong tremor.

On 23 May the sudden stoppages ceased, and the frequency of spectral peaks (fig 4.10) and the seismic power of tremor (fig 4.4) began to decrease, disappearing entirely for a short period on 26 May, and increasing again between 27 May and 2 June, when stoppages and powerful recommencements resumed. Although the frequency of lines in the tremor spectrum sometimes wandered within the width of the main bands, this is the only instance of change in the frequency of the bands themselves during the 1964

Fig 4.9: Seismograms of powerful recommencements of tremor recorded at site 5 in April - May 1964, and arranged in order of size. Their similarity indicates that all had the same source and mechanism.

Fig 4.10: Rayspan spectrum analysis of tremor recorded at site 8 (Ski Patrol), showing concurrent decreases of frequency and spectral density on 25 - 26 May and 28 - 29 May 1964, and increases on 30 May. On 23 - 24 May a pair of absorption lines, spaced about 2 Hz apart, descend through the spectrum.

activity. It was recorded at sites 8 and 9 and is not of instrumental origin.

On 6 June the tremor became very strong and erratic, and the greatest number of powerful recommencements of tremor in one day (14) occurred. At times the tremor seemed to consist largely of incipient recommencements (fig 4.11. The dominant frequency of 1 Hz in Fig 4.11 was characteristic of number 11 recording site while at other sites it was 2 Hz). At 1800 NZST on 6 June, tremor suddenly ceased entirely for two hours during which time a number of ultra-micro-earthquakes were recorded, and the erratic character of the activity ceased.

4) Middle period of low activity.

From 6 June, tremor decreased smoothly, and had almost ceased by 10 June. On 19 June it began to switch on and off fairly regularly every few minutes. The spectrum (fig 4.12) was fairly constant in shape and density during each pulse. The relation between average seismic power (P_{av}), seismic power of a pulse (P_p), pulse duration (t_p), and pulse occurrence rate (f) is approximated by

$$P_{av} = P_p \cdot t_p \cdot f.$$

Fig 4.11: Sonagram of tremor with many incipient stoppages and recommencements recorded at site 11 on 6 June 1964. The dominant frequency of 1 Hz was characteristic of site 11.

Fig 4.12: Sonagram of pulsing tremor recorded at site 8 in late July 1964. Pulse spacing varied more than pulse duration.

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To a first approximation (P_{av}) and (t_p) were constant and pulse power (P_p) was inversely proportional to occurrence rate (f), which varied mainly by change in the interval between pulses. The seismic power (fig 4.4) was low while the tremor was pulsing on and off. On 29 July it started to increase and by 31 July pulsations had ceased.

5) Final period of tremor stoppages and powerful recommencements.

Relatively infrequent stoppages recommenced on 31 July, and moderate activity continued until 8 September when the stoppages ceased again.

A new feature was that the stoppages were not always complete. Instead, the seismic power of tremor decreased suddenly without obvious shift in frequency spectrum, and remained small up to the powerful recommencement. The end of the 1964 activity occurred between 12 September and 31 October when no seismographs were operating.

Summary of Spectral Information

- (1) Most sonagrams and rayspan spectragrams show tremor is a broadband non-coherent ground noise.
- (ii) At any one site the band of high spectral densities lies between 0.8 and 6 Hz, and shows peaks and troughs which are normally constant in frequency.
- (iii) The spectra at different sites for the same time period differ in detail, the frequencies of many of the peaks and troughs being characteristic of the sites themselves.
- (iv) Peaks at 1 and 2 Hz are often recognisable at all sites and are apparently generated by the volcano. No clear harmonic at 3 Hz has been recognised, and at most sites, 2 Hz is the dominant frequency, and possible harmonics at 4 and 6 Hz may be present. At one site (11) the dominant frequency was 1 Hz.
- (v) On one occasion the main peaks centred at 2.2 and 4.4 Hz in the tremor spectrum shifted gradually down by half an octave in 6 hours, to 1.6 and 3.1 Hz, as tremor was dying away (fig 4.10). Peaks became troughs and vice versa. When tremor increased again

the peaks rose again to 2.2 and 4.4 Hz in 13 hours. This was recorded at sites 8 and 9 on opposite sides of the volcano. Almost as rarely, absorption lines descended asymptotically to a new frequency, but these events were recorded on only one seismograph at a time.

- (vi) At times of strong tremor, spectral lines sometimes formed within the normal tremor frequency band (fig 4.15), and could sweep quickly over a range of about 0.3 octave.
- (vii) Powerful recommencements of tremor have the same main spectral peaks as tremor at any one site.

Summary of Stop-Start Behaviour of Tremor

Data relating the duration of a tremor stoppage to the amplitude of powerful recommencement are given in table 4.1. From the initially available data (asterisk), there is a weak positive correlation (coefficient 0.15) between these quantities. The values of duration and amplitude squared ($\propto$ seismic energy) were well distributed and had similar ranges, the largest being about 20 times

the smallest for this typical group of data. The ratio of the amplitude of a powerful recommencement to the average tremor amplitude before the stoppage, approximated 5 in 1964 when temporarily stronger tremor often followed the powerful recommencement, compared with 30 in 1961 when the tremor following was not enhanced (fig 3.1).

A cumulative energy analysis of recordings from site 8 (fig 4.13) was made to see if the combined energy of a powerful recommencement and temporarily stronger tremor immediately following, equalled the energy conserved during the preceding stoppage of tremor. In seven of the curves (2,5,7,8,17,20,21) this was so, but in the other 14 curves the energy in the powerful recommencement exceeded the conserved energy by factors up to 15 (curve 1). Curves 3,4,5,11,12 and 16 also show other large changes of slope (seismic power) which do not correspond with stoppages or recommencements.

Fig 4.13: Cumulative seismic energy versus time, for sequences of tremor (slopes), stoppages (flats) and powerful recommencements of tremor (steps), recorded at site 8. The date, and time of powerful recommencements are shown for each sequence. On average, the energy released in a powerful recommencement exceeds that which could be stored due to the stoppage.

TABLE 4.1: Data for Powerful Resummencements of Tremor at Ruapehu in 1964. Recorded on Wood Anderson Seismograph (NW-SE $T_0 = 0.8s$, $h = 0.8$, Magnification 2800) 9.1 km from the crater.

Date	Time	Trace	(Amplitude) ²	Duration of	Remarks
1964	G M T	Amplitude	mm ²	stoppage	
		mm		minute	
30 Apr	0737	1.1	1.2	14.0	Emergent
30 "	1716	0.6 *	0.4	2.9	
1 May	0001	0.35	0.1	1.4	
1 "	0306	1.1 *	1.2	4.7	
1 "	1324	1.5 *	2.2	10.5	
1 "	1558 ca	0.6	0.4	-	Emergent
1 "	1716	1.0 *	1.0	9.5	
2 "	0105	1.1 *	1.2	10.0	
2 "	0750	0.9 *	0.8	11.0	
3 "	1554	1.2	1.4	-	
3 "	1935	1.3 *	1.7	9.6	
3 "	2033	1.0	1.0	-	Emergent
3 "	2334	1.8 *	3.2	2.5	

Date	Time	Trace	(Amplitude) ²	Duration of	Remarks
1964	G M T	Amplitude	mm ²	stoppage	
		mm		minute	
4 May	0124	1.5 *	2.2	4.5	
4 "	0236	0.5	0.2	4.5	
4 "	0251	1.8 *	3.2	4.5	
4 "	0914	1.5	2.2	-	
4 "	1050	3.5	12.2	08.6	
4 "	1140	2.8	7.8	9.8	
4 "	1238	2.6	6.8	6.0	
4 "	1847	0.3	0.1	3.9 ?	
5 "	0054	2.6	6.8	11.5	
5 "	0318	0.9	0.8	2.9	
5 "	0416	3.3	10.9	4.1	
5 "	0505	1.9	3.6	5.3	
5 "	0613	0.7	0.5	4.4	
5 "	0811	2.3	5.3	4.1	
5 "	0915	1.5	2.2	1.5	
5 "	1031	0.3	0.1	1.1 ?	
5 "	2109	0.4	0.2	2	Emergent

Date	Time	Trace	(Amplitude) ²	Duration of	Remarks
1964	G M T	Amplitude	mm ²	stoppage	
		mm		minute	

---

5 May	2325	0.8	0.6	2.6	
6 "	0034	1.1	1.2	1.6	
6 "	0106	0.4	0.2	2.0	
6 "	0133	0.7	0.5	0.6	
6 "	0415	0.2	0.0	2.1	Doubtful
6 "	0855	1.0	1.0	5 ca	Emergent
6 "	1031 *	0.6	0.4	4.5	
6 "	1236 *	0.7	0.5	6.9	
6 "	1724	0.5	0.2	2.4	
6 "	2110	0.9	0.8	4.4	
8 "	0046	1.0 *	1.0	3.5	
9 "	1234	2.1 *	4.4	8.3	
9 "	1648	1.4 *	2.0	5.3	
13 "	1627	1.7 *	2.9	1.0	
13 "	2124	2.0 *	4.0	7.0	
19 "	2233	0.9 *	0.8	7.1	
20 "	0057	1.5	2.2		
21 "	0335	2.5 *	6.2	15.5	

Date	Time	Trace	(Amplitude) ²	Duration of	Remarks
1964	G M T	Amplitude	mm ²	stoppage	
		mm		minute	
21 May	0432	0.8 *	0.6	19.5	
21 "	0634	1.5 *	2.2	14.0	
21 "	1213	1.3 *	1.7	21.5	
3 June	1323	2.2	4.8	-	
3 "	1509	0.6	0.4	2.0	
3 "	1534	0.7	0.5	3.5 ca	Emergent
3 "	1637	1.4	2.0	9.5 ca	
3 "	1739	1.4 *	2.0	9.6	Emergent
3 "	1819	1.7	2.9	13.7	
3 "	1915	0.9	0.8	2.6	
3 "	1922 ?	1.4 *	2.0	13.7	
5 "	0907	0.3	0.1	0.8	
5 "	1902	1.4 *	2.0	1.9	
5 "	2120	0.8 *	0.6	2.8	
5 "	2153	2.5 *	6.2	10.0	
5 "	2338	1.5	2.2	4.7	Emergent
6 "	0107	2.1	4.4	9.0	

Date	Time	Trace	(Amplitude) ²	Duration of	Remarks
1964	G M T	Amplitude	mm ²	stoppage	
		mm		minute	

---

6 June	0150	0.7	0.5	3.0	
6 "	0453	0.2	0.0	2.3	Doubtful

* Initially available Data. Used to find correlation between amplitude of recommencement and duration of stoppage.

Several varieties of recommencements have been observed. Quite frequently there is just strong tremor at the end of a stoppage. Some powerful recommencements occurred which were followed by tremor lasting only a few seconds (fig 4.14). Others were multiple (they stuttered). Fig 4.15 shows a powerful recommencement, followed after a brief minimum by very strong tremor in which the frequencies of spectral peaks rose 1/7th octave in 1.25 minute. Strong tremor can become a continuous series of incipient stoppages and powerful recommencements (fig 4.11). This phenomenon resembles the tremor in which B-type "volcanic earthquakes are often observed in the form of volcanic tremors of continuous train" (Minakami 1969). The powerful recommencements of tremor at Ruapehu in 1964 also resemble the B-type earthquakes described by Minakami. However, the Ishimoto - Iida (1939) coefficient m , based on the 50 largest powerful recommencements in 1964, is only 2.2 ± 0.4 (fig 4.16), distinctly less than the values 3.1 - 4 reported by Minakami (1969) for B-type earthquakes at Sakurajima, Asama, and Kilauea Iki. This

Fig 4.14: Sonagram of abortive powerful recommencements of tremor recorded at site 8 on 22 May 1964. Tremor stopped after half a minute.

Fig 4.15: Sonagram of a powerful recommencement of abnormally powerful tremor, recorded at site 8 in May 1964. Spectral peaks rose 1/7th octave in 1 minute (note harmonics). A spectral line at 1.3 Hz developed at 13 minutes, and shifted to 0.9 Hz at 14.5 minutes.

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Fig 4.16: Number of powerful recommencements of tremor in each amplitude range versus amplitude. From the eye-fitted curve, the Ishimoto-Iida coefficient m is 2.25 ± 0.4 .

might be explained by the tendency at Ruapehu for stoppages and powerful recommencements not to occur at low tremor levels, and by the correlation between amplitude of tremor and of accompanying powerful recommencements apparent on fig 4.6. Less plausibly, Minakami's (1969) graph of m versus depth could indicate that the powerful recommencements are about 5 km deep.

The Epicentre of Powerful Recommencements of Tremor

At most volcanoes the source of volcanic tremor is difficult to locate precisely, but at Ruapehu it is assumed to be the same as for powerful recommencements of tremor. However, no single powerful recommencement was well recorded at enough seismograph stations to give a reliable epicentre. It was necessary to assume that all had the same focus, as shown by the remarkable similarity of many of the recordings at any one site (fig 4.9), and to average the time differences for P at each pair of stations for all the powerful recommencements.

which were well recorded by that pair. The averages of these time differences were added, to give P-time differences between each station and station 9 which recorded the earliest P times:

Station number	8	11	1	2	15	16
P-time difference	0.1	0.75	1.20	1.73	2.23	3.63

(second) with respect

to station 9.

From the time-distance graph for a preliminary epicentre a velocity of 3.98 km s^{-1} was obtained. The final epicentre is at the north side of Crater Lake, and was determined using this velocity and the time differences between stations 9, 8 and 11 which were closest to the epicentre. The epicentre and the arcs for all stations are shown on Fig 4.2, and the time distance graph is shown in Fig 4.17. Fig 4.18 shows the best seismogram for each site arranged on a time-distance diagram. The reliability of the epicentre can be estimated if it is accepted that the volcano has a radially symmetrical structure to the 3 nearest seismographs which were all on andesite flows forming the lower slopes of the volcano. The error sources are then

Fig 4.17: Time distance graph for P of powerful recommencements of tremor. Data has been averaged on the assumption that the foci of all the recommencements in 1964 were the same.

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Fig 4.18: Seismograms of powerful recommencements of tremor arranged on a time-distance diagram. The best recorded event at each station was photographed with optimum amplitude. Note the lack of phases of consistent velocity lower than that of P.

the apparent velocity which is uncertain by 0.5 km s^{-1} (the difference between the apparent velocities to the east and to the north west of Ruapehu) and the time differences between station pairs, individual values for which differ by up to 0.25 s from the mean values. The epicentre based on stations 8, 9 and 11 would be shifted 0.37 km by an error of 0.5 km s^{-1} in velocity, and not more than 1.0 km by a 0.25 s error in timing. Inconsistencies between the distance arcs from stations are also about 1 km (Fig 4.2).

The depth cannot be determined without a station near the epicentre, or an independent measurement of the seismic velocity, but the linearity of the time distance curve suggests the depth is less than 5 km .

Magnitude and Energy of the Most Powerful

Recommencement of Tremor

The most powerful recommencement of tremor occurred at 2250 NZST on 4 May 1964. Its magnitude was determined from the trace amplitude (3.5 mm) recorded on the Wood Anderson seismograph at site 5, 9.1 km from the epicentre. Richter's (1958) table for local magnitude applies to

earthquakes of about 16 km depth, but it is possible to use it for local very shallow earthquakes if it is assumed that equal magnitude earthquakes at 16 km epicentral distance and zero depth, and at zero epicentral distance and 16 km depth are recorded with equal amplitude.

$$\text{Then: } M_L = \log_{10} A_{(16,0)} - \log_{10} A_{O(0,16)}$$

Where $A_{(16,0)}$ is the trace amplitude in mm of a surface earthquake recorded at 16 km epicentral distance.

$A_{O(0,16)}$ is the trace amplitude in mm of a zero magnitude earthquake of depth 16 km, recorded at the epicentre.

$$\text{and } -\log A_{O(0,16)} = 1.4 \text{ (Richter 1958)}$$

In this way, the amplitudes are compared at the same distance along ray paths. Assuming Wood Anderson trace amplitudes proportional to d^{-2} the amplitude at 16 km which is equivalent to 3.5 mm at 9.1 km is $A_{(16,0)} = 1.1$ mm, and the magnitude for the most powerful recommencement of tremor in 1964 is:

$$M_L = 1.5$$

Richter (1958) gives the relation between M_L and the seismic energy E as:

$$\text{Log}_{10} E = 9.9 + 1.9 M_L - 0.024 M_L^2 \text{ erg}$$

Then the seismic energy for the most powerful recommencement of tremor in 1964 is:

$$E = 5 \times 10^5 \text{ J}$$

The charge weight of a TNT explosion equivalent to the most powerful recommencement of tremor in 1964 can be estimated from the data in table 4.2 collected by the author during unpublished crustal structure investigations in the area by Geophysics Division of D.S.I.R.

TABLE 4.2: Data for Comparison of Seismic Explosions near Ruapehu and Powerful
Recommendations of Tremor.

Date	Seismograph	Site No.	Peak Trace Ampl.	Shot Size TNT	Distance	Azimuth of shot	Remarks	Analysis for A ∝ d ⁻¹		Analysis for A ∝ d ⁻²	
			mm	kg	km	°E		Equiv W-A Ampl at 9.1 km mm	kg TNT per mm at 9.1 km	Equiv W-A Ampl at 9.1 km mm	kg TNT per mm at 9.1 km
1.12.60	Willmore Z	1 (CNZ)	0.08	46	28.1	139.4	2 m deep in Moawhango R.	0.025	1860	0.076	600
4.12.60	"	"	0.57	460	"	"	"	0.182	2530	0.58	790
5.12.60	"	"	1.0	925	"	"	"	0.32	2890	1.02	905
13.12.60	"	"	3.33	2780	"	"	3 m deep in Moawhango R.	1.065	2610	3.40	820
4. 9.61	"	"	7.0	635	22.5	82.5	5 m deep in Tongariro R.	1.73	370	4.28	148
20. 4.63	"	"	9.5	10	5.3	97	3 m deep in Lower Tama Lake	0.55	18	0.32	31
20. 4.63	Wood Anderson	5	0.8	10	"	"	"	0.47	21	0.27	37
4. 5.64	"	"	3.5	Volcanic	9.1	169	Most powerful recommence-ment of tremor at Ruapehu	3.5		3.5	
4. 5.64	Willmore Z	1 (CNZ)	Unread-able	"	"	"					

Because no single seismograph adequately recorded both seismic explosions and the powerful recommencement of tremor the amplitudes recorded at CNZ were converted to equivalent Wood Anderson amplitudes by dividing by 10 (the observed magnification ratio for small events). The Wood Anderson amplitudes expected if the seismic explosions had been recorded at 9.1 km were then calculated assuming proportionality to d^{-1} and d^{-2} . The charge weight per mm amplitude on a Wood Anderson seismograph at 9.1 km distance was found (table 4.2) using the linear relation between trace amplitude and charge weight shown by the 1960 recordings. O'Brien (1967) also found this relation, but not for such large charges.

The results for $A \propto d^{-2}$ are the more consistent and have been adopted. The apparent correlation of results with distance is explainable by reduced seismic efficiency at greater distance due to larger shots but less effective tamping. Taking the geometric mean of 160 kg of TNT per mm of Wood Anderson trace amplitude at 9.1 km distance, the charge equivalent to the most powerful recommencement of tremor in 1964 is within a

factor of 5 of 560 kg of TNT. The seismic efficiency of the most powerful recommencement of tremor inferred from this result is:

$$\frac{\text{Seismic energy from magnitude} = 5 \times 10^5 \text{ J}}{\text{Chemical energy in 560 kg TNT} = 2.4 \times 10^9 \text{ J}} = 0.5\%$$

Local Earthquakes

The author searched the seismograms from the Tongariro and Chateau (CNZ) stations prior to 1964 and found no earthquakes (excluding powerful recommencements of tremor) which appeared to be connected with the volcanic activity of Ruapehu; 1964 was no exception.

The CNZ seismograms were searched for earthquakes with S minus P intervals less than 10 s, and these were plotted on fig 4.6. Two types of earthquakes were recognised, those having a prominent short sharp initial movement in the P phase, which is often larger than S, and those having a normal P phase of increasing or constant amplitude up to the arrival of S which is larger than P. Histograms of occurrence against S minus

P interval for each type of earthquake are shown in fig 4.19. The relation between S minus P and slant distance in the area is approximately 5 km per second of S minus P interval. The theoretical histogram for uniform distribution of earthquakes with frequency of occurrence increasing by a factor of 8 for each unit decrease in magnitude, and for radius of detectability derived from the magnitude scale, is also shown on the figure.

Apart from an adjacent peak and trough which are probably not significant and can be smoothed out, the earthquakes with normal P agree closely with the calculated normal histogram. The histogram of earthquakes of the prominent P type has a pronounced peak at S minus P between 4 and 5 corresponding to a slant distance between 20 and 25 km. Many of these earthquakes, including those between 3 and 7 February 1964, are very similar and one such on 14 June 1964 at 1353 was recorded at 7 stations (1,2,3,8,9,10,15) and was located 16 km south-east of Ruapehu at a depth of 19 km.

However for this earthquake, the prominent P phase was only recorded at station sites 1 and 8 and as Ruapehu

Fig 4.19: Histogram of number of earthquakes against S - P interval at CNZ (site 1) for:

- (a) Earthquakes of ordinary (tectonic) appearance.
- (b) Hypothetical uniformly distributed earthquakes.
- (c) Earthquakes with prominent P arrivals.

lies between these sites and the epicentre, it may be that the effect is caused by the waves passing through the magma system. There is no evidence of a concentration of shallow earthquakes under the active craters, apart from the powerful recommencements of tremor.

CHAPTER 5

RUAPEHU VOLCANO 1965 - 1966

Introduction

In this period the author had no access to seismograph equipment, but the Chateau (CNZ) seismograms were kindly made available to the author by the Superintendent of the Seismological Observatory, Geophysics Division D.S.I.R. The opportunity was taken to probe the Crater Lake of Ruapehu, and during this programme, an eruption occurred on 24 July 1966, which was well recorded on CNZ. The writer's understanding of (and respect for) Ruapehu increased considerably as a result.

Temperature, Bathymetric, and Chemical

Surveys of Ruapehu Crater Lake

The calculations of the thermal power output from Ruapehu Crater Lake described in Chapter 3, depended on the assumption that lake temperatures measured at the shore, could be applied to the whole area of the lake.

The surveys in 1965 and 1966 were designed to test this, and also to provide bathymetric data with which future surveys, as well as photographs and sketches of the empty crater after the 1945 eruption (Gregg 1960), could be compared. This, it was hoped, would help answer the question: "Does tremor indicate movement of lava?"

Field work at the lake occupied the periods 2 - 4 September 1965, 7 - 8 September 1965, 14 - 17 May 1966, 18 - 19 June 1966, and 26 - 30 July 1966. The initial two surveys were concerned with the temperature and depth of the whole lake, and the positions of stations were determined from difference angles between compass bearings to four marker stakes at easily recognised points on the shore (Fig 5.1). Equipment for the initial surveys consisted of a 4 m fibre-glass row boat, Kelvin sounding winch with 1.5 mm galvanised steel wire, metering pulley, bathythermograph and mercury thermometer, all kindly loaned by the New Zealand Oceanographic Institute, Department of Scientific and Industrial Research (D.S.I.R.).

Twenty six stations were occupied, the positions of which are shown in Fig 5.2 with uncertainties of about 8 m. At each station the temperature about 70 mm below

Fig 5.1: Vertical aerial view of Crater Lake, showing the positions of the survey markers (dots) on the shore. The photo was taken on 17 November 1967. An ice cliff later built up at the NNW shore of the lake.

Photo by N.Z. Aerial Mapping Limited

Fig 5.2: Station positions (o) and marker positions (O) at Crater Lake. The grid lines are taken from the Wharepu map, NZMS 2, Sheet N122/2.

the surface was measured to 0.1°C , and the depth of the lake was measured to 0.5 m (Table 5.1). At the first 6 stations, the bathythermograph was used as the sounding weight, but the wire then fractured near the end due to local embrittlement caused by the hot acid lake water. In the deepest part of the lake, the lower end of the wire dissolved completely away in about 20 minutes. A further difficulty at depth was that although the weight was invariably free when first placed on the bottom (it was lifted and replaced several times to 'feel' the bottom at each site), it was often irretrievably stuck when re-winding was attempted. Over the shallower parts the bottom was light grey mud.

In the next two surveys, maximum temperatures and water samples were obtained each 25 m in depth at the station (21) over the deepest part of the lake, position being maintained by pre-cut lines to the shore. A 2.5m dingy, marked reel of Ulstron line, maximum thermometer, water sampler, marked reel of electric cable, and a thermistor thermometer were used. The Ulstron line failed due to chemical alteration at 230 m depth, while the insulation of the cable failed before 50 m.

TABLE 5.1:

Station Data for the Survey of Ruapehu Crater

Lake on 4 and 7 September 1965.

Station No.	Bearings to Markers 1 2 3 4 Degrees East of Mag- netic North.	Surface Temperatures °C	Depth Metres	Remarks
1	Markers obscured	24.5	28	Bathythermogram 2 Station in rising gas
2	Markers obscured	23.5	83½	Bathythermogram 3
3	Markers obscured	23.5	62	Bathythermogram 4
4	Omitted			
5	273 163 90 52	25.5	23	Bathythermogram 5 Station over upwell- ing water
6	291 167 76 32	23.8	74	Bathythermogram 6
7	295 188 52 9	23.8	139½	Bathythermogram 7
8	293 213 25 352	23.6	77½	
9	295 233 3 344	23.6	36½	
10	302 232 13 351	23.9	15	
11		23.5	½	Outlet Stream
12	208 159 108 79	24.8	6	

Station No.	Bearings to Markers 1 2 3 4 Degrees East of Mag- netic North.				Surface Temperatures °C	Depth Metres	Remarks
13	218	166	114	83	23.5	17	
14	226	173	123	89	23.9	17½	Very muddy bottom
15	243	183	126	55	24.0	15½	Photo of Marker 4
16	256	189	116	13	23.9	44	
17	270	197	95	351	23.9	65	
18	274	203	62	345	23.9	64½	
19	281	202	52	352	23.6	138½	
20	302	213	27	358	23.5	67½	
21	289	192	57	5	23.7	298	
22	275	188	78	8	23.9	-	
23	319	181	38	12	23.3	30	
24	327	144	52	23	24.0	35½	
25	319	147	72	39	23.8	45½	
26	313	146	83	52	24.0	-	

The final survey was intended to complete the temperature with depth curve, using stainless steel wire, but the lake erupted on 24 July 1966 and destroyed our boat on the shore, so a survey of the immediate effects of the eruption was made instead.

SURVEY RESULTS

Bathymetry

Fig 5.3 shows the depth contours at 50 m intervals. Control on the contours for depths greater than 100 m is poor, and the lake could be more than 300 m deep. The shape of the lake floor is shown by cross sections in Fig 5.4. The average profile resembles an exponential horn, with shallow gently sloping mud bottom near the shore, and a very steep sided hole near the centre. As far as can be judged from a photo of the empty crater shortly after the 1945 eruption, (Gregg 1960), no great changes have taken place since then. The lake volume, computed from Figs 5.3, 5.4 is 10^7 m^3 .

Fig 5.3: Bathymetric contours at 50 m intervals in Crater Lake.

Fig 5.4: Profile of the bathymetry of Crater Lake in the WSW - ENE direction.

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Surface Temperatures

During the September 1965 survey the surface temperatures over most of the lake lay between 23.3°C and 24.0°C . The temperature in the outlet stream (23.5°C) was also in this range. Where gas bubbles or obvious currents of water were rising, the temperatures were 24.5 to 25.5°C . The only measurement below 23.3°C was 21.5° made at arms length from the shore, (covered to the water's edge with snow) near marker position No. 1. All the rising gas and upwelling water observed was in the NE part of the lake, and this was the position of the greatest fumarole activity in 1964. In June 1966 a conspicuous line of upwellings extended from a point on the shore 50 m north of marker No. 1 across the lake nearly to marker No. 4. The temperatures in September 1965 were taken while the lake was heating up. The maximum of 47°C was reached in March 1966. In May and June 1966 it was cooling down again, and the surface temperatures in the middle of the lake were 33.5° and 31.0°C respectively.

Lake Temperature at Depth

All the bathythermograph charts (Fig 5.5) show an upper layer of water with a low variable temperature gradient. For the sites at which the depth is 74 m or less, the uniform layer reaches to the bottom, but the two sites with greater depth show rising temperatures below a depth which is smaller where the depth to the bottom is greater.

Chart No. 7 shows an average temperature gradient of 1°C per 6 m between 50 and 140 m together with several small regions of negative gradient, the one between 110 and 130 m depth being especially clear.

Greater depths were reached in June 1966 and the temperature-depth curve shown in Fig 5.6 was obtained. The temperatures were everywhere higher than in September 1965, and the average gradient is higher, being 1°C per 2 m. Measurements in June 1966 were made every 25 m and cannot show detail, but a region with a low or negative gradient was revealed between 175 and 200 m.

Fig 5.5: Bathythermograph charts for Crater Lake, made on 4 and 7 September 1965. Station positions for charts 2, 3 and 4 are unknown, but those for 5, 6 and 7 are shown similarly numbered on Fig 5.2. The lake bottom was reached in all cases.

Fig 5.6: Temperature-depth curves in Crater Lake measured on the dates indicated. The theoretical curve is for the assumptions of single phase turbulent convection with heat input at the deepest point in the lake, and does not agree with the observed curves.

Lake Density and Chemistry at Depth

After the highly destructive conditions in the lake at depth had been discovered, a water sampler was devised by Professor H.W. Wellman, Dr. R.A. Hoare and R.C. Ross, which was cheap, simple, and would also collect the effervescent gases released when the sample was brought up. It consisted (Fig 5.7) of a bar of aluminium which was not rapidly attacked by the lake (Tombs 1960) supporting a thin polythene tube 1 m long in a straight vertical position, so that the water would pass through it as the sampler was lowered. The top half of the tube was fastened to a weight, which was held in its position at the top of the aluminium bar by a glass rod. When this was broken by sending a messenger down the line, the tube folded so that it pinched in the middle and the ends hung down to the same level. With care the sampler could be lifted into the boat, and the tube sealed with clamps and removed from the sampler without losing any of the sample. A fresh tube was used to collect and store each 10 cm^3 specimen.

Fig 5.7: The water sampler, thermometer and winding reel in use at Crater Lake on 19 June 1966. At sampling depth, a messenger released the top end of the polythene sampling tube which folded, and thus retained the samples.

Photo by R.G.P. Zachariadis

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The samples were separated into gas and turbid liquid fractions, and the relative densities of the turbid liquids were measured with a pycnometer at 15°C to obtain the curve of increasing density with depth shown in Fig 5.8. The accuracy of each density determination in relation to the others is ± 0.0005 . The density of the sample from 75 m is probably too high because the sampler was lowered too far and approached this depth from below. The lag in the flow through the tube may have caused the effective depth from which this sample was collected to be greater than 75 m. The density curve has a low gradient not significantly different from zero above 50 m depth, and below this the average density gradient is $4.3 \times 10^{-2} \text{ kg m}^{-4}$.

The samples were then allowed to settle, and the supernatant liquids were analysed by Dr. R.B. Glover and W.A.J. Mahon of Chemistry Division D.S.I.R. and the compositions plotted against depth (Fig 5.9). From the graph, and by inspection of the samples, it is apparent that the density gradient in the lake is due not to dissolved material, but to increasing turbidity with depth. The only definite variation in the composition of the liquid with depth, occurs at 125 and 150 m.

Fig 5.8: Density of Crater Lake water versus depth on 18 - 19 June 1966. The increase in density reflects the increase in turbidity of the samples. In situ densities at sample collection temperatures were calculated from the NTP densities.

Fig 5.9: Chemical composition of the liquid fraction of lake water versus depth of collection on 18 - 19 June 1966. Strong acids were measured by neutralising to Methyl Orange. Weak acids were measured by the difference in neutralising to Methyl Orange and Phenolphthalein. Acid concentrations are in milli-equivalents/litre.

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At 125 m, Na and K are lower than average, and Ca, F and HBO_2 are higher, while at 150 m, the variations are the reverse of this. One possible explanation is the inflow of water at these depths.

Chemistry Division, D.S.I.R. also analysed the gas portion of the samples with a chromatographic column. It was found that the samples were essentially Nitrogen and Oxygen, but as there was insufficient sample for more than one column the relative proportions of these could not be determined. The proportion of minor constituent gases were:

<u>Depth</u>	<u>Hydrogen</u>	<u>Carbon Dioxide</u>	<u>Methane</u>
150 m	Not detectable	Present	Not detectable
175	Not detectable	0.05%	0.05%
200	0.5%	0.05%	0.06%
225	0.5%	0.05%	0.1%

The most surprising thing is the absence of sulphurous gases, but as 35 days elapsed between collecting the samples and analysing them, it is possible that diffusion largely replaced the sample by air.

The percentage of volume of gas in the water samples at NTP is shown in Fig 5.10.

Fig 5.10: Volume percentage of gas in Crater Lake water on 18 - 19 June 1966, measured at NTP. This amount of gas reduces the effective boiling point of the lake by only a small amount.

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INTERPRETATION OF SURVEY RESULTS

Convection must be occurring in the lake, because the turbid lake water clears completely when it is allowed to stand. Also the calculated heat loss from the lake surface for a temperature of 25°C , is 1 kW m^{-2} , which is 4000 times greater than could be sustained by conduction alone. However, the shape of the temperature gradient curve is unusual. A possible explanation for the increase in gradient with depth is provided by the funnel shape of the lake bottom. Assuming that the heat enters at the bottom of the lake, and is transferred through the lake by turbulent convection, the average heat flux H at any depth z is inversely proportional to the area of cross section A at that depth. According to Malkus (1963).

$$H = K_t U (R/R_{\min}^c)^{1/3} \quad (5.1)$$

where K_t is the thermal diffusivity
 U is the vertical temperature
 gradient = $\frac{\Delta t}{\Delta z}$

R is the Rayleigh number

$$R = \frac{g \gamma \Delta t (\Delta z)^3}{\nu K_t} \quad (\text{Veronis 1965}) \quad (5.2)$$

- t is the temperature
 g is the acceleration of gravity
 Y is the thermal expansion coefficient
 (volumetric)
 Δz is the depth of water across which the
 temperature difference is Δt .
 ν is the kinematic viscosity.
 $R_{\min}^C$ is the minimum value of R at which
 Bénard convection can be maintained;
 about 10^3 for a homogeneous fluid
 (Knopoff 1964)

In a given system $R \propto \Delta t \propto U$, and $R_{\min}^C$ is constant;

therefore $\frac{R}{R_{\min}^C} \propto U$

Substituting in (1) $H \propto U^{4/3}$

And as $H \propto 1/A$ at depth z

Then $U = k A^{-3/4}$ at depth z (5.3)

Where k includes other variables which
 in a Boussinesq fluid are independent
 of depth.

Therefore $t = \int_0^z k A^{-3/4} dz$ (5.4)

Temperature t was evaluated by plotting $A^{-3/4}$ against depth, and integrating graphically. The temperature scale of t was obtained by fitting the integral at $t = 31^{\circ}\text{C}$ for $z = 0$, and $t = 32^{\circ}\text{C}$ for $z = 60$ m. The theoretical curve so computed is plotted with the observed temperature-depth curve (Fig 5.6). It appears that this mechanism alone is inadequate to explain the change in temperature gradients within the top 100 m of the lake.

The observed increase in turbidity (and NTP density) with depth may assist the above mechanism. The turbidity gradient will increase the minimum Rayleigh number at which convection can occur, and assuming that suspensions behave like solutions over periods much less than the settling time, the minimum Rayleigh number is given by Veronis (1965) for Bénard convection as follows:

$$R_{\min}^C = \frac{27\pi^4}{4} + \frac{R_s K_t}{K_s} \quad (5.5)$$

Where R_s is the Rayleigh number for a solute gradient,

$$R_s = \frac{g B \Delta s (\Delta s)^3}{\nu K_t} \quad (5.6)$$

B is the fractional density change per unit change of chemical concentration at constant temperature and pressure.

s is the difference in chemical concentration over depth z.

$\frac{B\Delta s}{\Delta z}$ = W is the fractional density change (NTP) per m of depth between 100 and 200 m, = $5 \times 10^{-5} \text{ m}^{-1}$ for Crater lake.

K_s is the chemical diffusivity, $2 \times 10^{-13} \text{ m}^2 \text{ s}^{-1}$ for a suspension in water at 60°C of particles of diameter 5 micrometre (Perrin 1910). Cf $10^{-9} \text{ m}^2 \text{ s}^{-1}$ for salt in solution at 20°C (Coulson and Richardson 1964).

Using also data: $\Delta z = 2 \times 10^2 \text{ m}$; $U = 5 \times 10^{-1} \text{ K m}^{-1}$, and values from the Handbook of Chemistry and Physics for water at $60 - 65^\circ\text{C}$: $v = 4.4 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$,

$$K_t = \frac{\text{Thermal conductivity of water}}{\text{Specific heat} \times \text{density}} = 1.6 \times 10^{-7} \text{ m}^2 \text{ s}^{-1},$$

$$Y = 5.5 \times 10^{-4} \text{ K}^{-1}, \quad g = 9.8 \text{ m s}^{-2};$$

we find for the region between 100 m depth and the bottom of the lake:

$$R_{\min}^c = 10^{25}$$

$$\text{and } R = 6 \times 10^{19}.$$

Since $R \ll R_{\min}^C$, neither Bénard nor turbulent convection can occur in this region, and the simple explanation of lake bottom shape for the increases in temperature gradient with depth cannot apply.

To check the stability against ordinary convection, the relative densities of the samples at their in-situ temperatures were calculated (Fig 5.8). They show the lake is gravitationally stable between 50 and 125 m depth, but between 125 and 175 m, where the in-situ density gradient may be zero, we depend on the Rayleigh number result:

$$R_{\min}^C = 4 \times 10^{22} \gg R = 3 \times 10^{17}$$

to show that ordinary Bénard convection is not occurring.

Veronis (1965) shows that overstable convection motion similar to an internal seiche is possible if:

$$R > R_{\min}^O = \frac{27\pi^4}{4} + \frac{\sigma}{1 + \sigma} R_s \quad (5.7)$$

where $\sigma = v/K_t$

For crater lake below 100 m, $R_{\min}^O = 5 \times 10^{18} < R$ and overstable motion should occur, transmitting heat at a rate about three times the purely conductive heat flux (Veronis 1965). This is insignificant compared with

the observed heat flux, but the overstable motions may explain the seiches of amplitude up to 75 mm and period 15 to 31 s observed on 3 and 11 August 1966 (E.F. Lloyd personal communication). Seiching is not normally obvious however.

The discussion so far implies that while ordinary convection may be occurring in the upper part of the lake where temperature and density gradients are low, it is not occurring in the lower part. Another mechanism is required to explain the turbidity and heat flux, and as volcanic tremor and the eruption of vesicular andesite (collected by E.F. Lloyd 3 Aug. 1966) indicate temperatures above boiling below the lake, a reasonable mechanism is two-phase convection. It is suggested that steam bubbles entering the bottom of the lake, rise progressively higher as they heat more water, entraining and accelerating a column of very hot but dense muddy water sufficiently to throw it up through the central gravitationally stable part of the lake. The highest level to which successive columns rise before disappearing by mixing is marked by a sharp change in temperature gradient, and this level becomes shallower as the lake

heats and becomes more turbid, until the columns cause periodic geyser-like eruptions as observed in the centre of the lake during August 1966. Heat flow across the boundary between the stable central layer, and the convecting upper layer occurs by settling of the sediment, so that hotter water joins the convecting layer.

The first part of the process can be demonstrated in the laboratory. If a beaker of water with sugar crystals on the bottom is heated, steam bubbles rise, vanish, and propel 'smoke rings' of syrup into the water above. These slow, disperse suddenly, and sink again, producing temperature and chemical gradients of the required type.

The Eruption of 24 July 1966

DESCRIPTION OF THE ERUPTION

At 10 hours NZST on 24 July 1966, Crater Lake erupted, and sent a mushroom head of steam 2500 m above the top of the volcano. At 1230 NZST Chief Ranger J. Mazey visited the crater by helicopter and saw that the lake was agitated and steaming vigorously, especially along a line from Pyramid Peak to Paretetaitonga Peak,

and that about 570 litre/second of water was emerging from the lake outlet tunnel instead of the normal 25 to 50 litre/second. Water had been thrown up about 10 m all round the crater sides, melting the snow ($6 \times 10^3 \text{ m}^3$) and washing away boulders at least 0.6 m in diameter, and a jet of muddy water had been ejected, leaving a line of discoloration on the snow to the rim of the crater, 150 m above, and 230 m west of the lake. Two climbers in the Dome Shelter about 400 m north of the lake were unaware of the eruption, but had seen half metre high waves and mild ebullition at the centre of the lake the day before. On 25 July, the lake was still agitated and overflowing at 570 litre/second (R.B. Glover personal communication 1966) but on 26, 28 and 30 July it was quite calm. On 28 July the temperature at the north eastern beach was 53.5°C and on 30 July the overflow was 30 litre/second at a temperature of 54.3°C . Fragile sulphur bubbles about 3 mm in diameter were found on the beach near the outlet stream. At 0650 NZST on 2 August, park rangers observed a plume of steam rising to 1200 m above the lake, and on 3 August, blocks of fresh vesicular andesite up to 1 m across were found near the outlet stream, some floating, and others cast

up by 1 m waves from an eruption after 30 July (E.F. Lloyd, personal communication 1966).

After 3 August the lake fell progressively further below the outlet level until December 1966 (E.F. Lloyd and R.B. Glover, personal communications 1966). On 11 August small eruptions began to be observed close to the centre and deepest part of the lake, and about two a day probably occurred until the end of August. Domes and geyser-like columns of very muddy upwelling water up to 15 m high and 30 m across were erupted, usually with a sighing sound or a noise like a waterfall, but at least once with a bang, and sometimes waves up to 0.5m high reached the shore. Each eruption lasted less than a minute and released a head of steam as in Fig 5.11.

THE SEISMIC OBSERVATIONS

The records at the Chateau Seismograph and reports from Rangers and Ski Instructors indicate that volcanic activity began to increase on 16 July. The seismic data is summarised in Fig 5.12. Six different types of seismic activity were recognised.

Fig 5.11: An upwelling of hot black water from the depths of Crater Lake on 7 April 1968 which was similar to those observed during August 1966. Cells of bottom water were thrown up through the gravitationally stable section of the lake due to acceleration and heating by two phase convection.

Photo by N.C. Simpson

Fig 5.12 Seismic power at Ruapehu in July-August 1966.

Upper: Nominal peak power and seismic energy of the following types of event:

X A-type earthquakes	O Powerful recommencements of tremor
Δ B-type earthquakes	+ Vibrations of hydrothermal eruption type.
	■ Vibrations accompanying dike injection?

Lower: Seismic power of volcanic tremor.

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- (i) Small A-type earthquakes of tectonic character with high frequencies, and S minus P intervals of 2 to 3 s. Three occurred between 2 and 4 July.
- (ii) B-type volcanic earthquakes (Fig 5.13) with frequencies of 1 - 2 Hz, and intervals of 4 - 4.5 s between P and the maximum phase (L?). Four occurred on 6, 8 and 16 July, when volcanic tremor was becoming obvious.
- (iii) Volcanic tremor of dominant frequency 2 Hz. This became persistent on 16 July, and was very strong for seven hours on 21 July. Thereafter it declined irregularly and had practically ceased by 5 August.
- (iv) Tremor stoppages of 1 to 30 minute, ending in powerful recommencements of tremor. Eleven occurred on 17 - 18 July, and six on 21 July. The stoppages were usually partial, and more than one tremor source may have been active.
- (v) Irregular outbursts of volcanic noise during the period of maximum visible activity. Nine occurred between 23 and 26 July, and that at 2059 NZST on 23 July (Fig 5.14) was the most energetic event during the whole active period, and had a dominant

Fig 5.13: Seismogram of a B-type earthquake recorded at CNZ (site 1) on 16 July 1966.

Fig 5.14: Volcano-seismic events at 1900 and 2059 NZST the night before the main observed eruption of Crater Lake. Note the lack of distinction in form between beginnings and endings. The event at 2059 NZST was the most energetic volcano-seismic event recorded in 1966.

frequency of 1.75 Hz. The recordings lack clear distinction between onset and ending. They look the same at each end, and are similar to vibrations which Dibble (1969) reported as accompanying strong eruptions of mud, ash and water from the lake on 26 - 28 April, 1968.

- (vi) Vibration accompanying the eruption at 1000 NZST on 24 July, (Fig 5.15). The dominant frequency of 1.6 Hz, interval of 13 s between onset and maximum phase, and the coda of over 3 minute duration, give this event a distinctive appearance. A very similar vibration at 0630 NZST on 2 August probably accompanied the eruption of the blocks of vesicular andesite found by Lloyd on 3 August.

DISCUSSION OF THE ERUPTION

Although the data are insufficient to show definitely how the eruption occurred, a useful analysis is possible. The average heat output from the volcano which caused the lake temperature to increase from 31°C to 53°C in the eight days before the eruption is 1200 MW plus the

Fig 5.15: Volcano-seismic event accompanying the eruption of Crater Lake at 1000 NZST on 24 July 1966. Note the interval of 13 s between the onset and the maximum amplitude, and also the long coda similar to volcanic tremor. The only other event of this form occurred at 0630 NZST on 3 August 1966, and probably accompanied the emission of blocks of fresh pumice into the lake.

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heat loss by evaporation. If the heat output was constant, this loss would average 700 MW giving a total heat output of 1900 MW.

When the surface temperature is 53°C the author's hypothesis for heat transfer through the lake indicates that the region of high temperature gradient almost reaches to the surface, and that all the water below about 100 m depth is above 100°C , and a considerable quantity is near boiling point for its depth. The energy of superheat relative to 100°C would be 2×10^{14} J if the 10^6 m^3 of water below 100 m averaged 50°C below boiling point for its depth. This energy is more than enough to cause the main eruption, and could accumulate in about 25 hours.

Because the vibration during the main eruption was similar to that on 2 August when vesicular andesite was apparently emitted, it is suggested that convective overturn and release of super heat in the eruption of 24 July was triggered by the injection of magma into a fissure in the lake floor. The line of vigorous steaming across the lake after the eruption also suggests a fissure, but apparently no lava entered the lake on this occasion.

The water balance in the lake (Fig 5.16) plays an important part in the analysis. The net water inflow rate balances the sum of the visible outflow rate and the evaporation rate, the curve for which also gives the rate of heat loss by evaporation which is the major part of the total loss. The curves are based on lake temperatures, levels, and overflow rates, generously supplied by Lloyd and Glover prior to publication. A feature which requires explanation is the large overflow during the day after the eruption. This could not all be condensed steam or the lake would have continued to heat up at 3.5°C per day. Neither could it be due to thermal expansion of the lake as this would have required a temperature increase of over 100°C . Most likely it was ground water displaced by a growing volume of steam under the floor of the lake, but the possibility of upward movement of part of the lake bottom has not been eliminated.

Between 26 July and 2 August, the evaporation and net inflow curves remained above normal, and the heat output from the volcano averaged 900 MW. Following the eruption of vesicular andesite into the lake on 2 August, the heat loss and evaporation fell to about twice normal,

Fig 5.16: Water balance and heat loss curves for Ruapehu crater lake during 1966.

and allowed water to re-invade the steam filled rock below the lake, so that lake level fell below the outlet level, and the net inflow fell close to zero. By 11 August, however, the heat output and net inflow had increased again to 700 MW and 255 litre second⁻¹ respectively, and small eruptions began to occur as columns of muddy super-heated water reached to the surface. The bang and sudden up-doming of the lake at the start of some of the eruptions indicate that explosive boiling of the upwelling water was possible in spite of its muddy nature. The eruptions stopped when the heat output of the volcano began its final decline to normal.

CHAPTER 6
RUAPEHU VOLCANO 1966
Foreword

The activity of Ruapehu in 1968 was the subject of the following paper (Dibble 1969), published in the Journal of Geophysical Research. The values of seismic power reported in this paper are measured from the Dome power recorder, and are approximately 10 times greater than the values from CNZ which have been used in the remainder of the thesis.

It is believed that this resulted from a fault in the amplifier at the Dome which increased the effective gain by 3, but which was not suspected until after the seismograph and telemetry cable had been damaged by lightning in August 1968, and repaired in May 1969 after the snow had melted.

Calibration then showed that the indicated values on the Dome seismic power recorder were 20% of the theoretical values taken from Fig 3.2.

Seismic Power Recordings during Hydrothermal Eruptions from Ruapehu Crater Lake in April 1968

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Ruapehu is an andesite volcano in the North Island of New Zealand whose activity is being monitored by two new instruments: a seismic power recorder described in this paper, and a differential magnetometer described in a companion paper. On April 5, 18, and 26, May 29, and June 9, 1968, episodes of activity commenced with volcanic tremor of frequency 2 c/s and radiated seismic power up to 100 kw. Between 10 hours and 7 days after the commencement, when the tremor power had decreased, strong hydrothermal eruptions occurred in the crater lake. The largest eruption, which occurred at 0749 NZST on April 27, 1968, had a peak seismic power of 9 Mw and a seismic energy of 10^{15} ergs; it ejected nearly 10^6 cubic meters of mud and water from the lake.

INTRODUCTION

In April, May, and June 1968, moderately large hydrothermal eruptions occurred in Ruapehu crater lake (area 0.2 km²) and covered 1 km² of the mountain with mud from the bottom of the lake (Figure 1). This caused a small lahar differing only in scale from those which have built the lahar plains on the east and south sides of the volcano [O'Shea, 1954]. Muddy rain left a noticeable deposit at Waiouru, 24 km from the crater. This was the greatest activity the volcano had shown since the ash and lava eruption of 1945, although lesser periods of activity had occurred in 1964 and 1966.

INSTRUMENTATION

Two new recording instruments were operating in 1968: a differential magnetometer [Johnston and Stacey, 1969] and a seismic power recorder developed by the author. This seismograph consists of a vertical Willmore seismometer (period 2 sec) and amplifier situated at the Dome Shelter (arrow, Figure 1) 400 meters north of the crater lake, and connected by telephone cable to a strip chart recorder and slow motion tape recorder [Dibble, 1964], which are housed in the display room of Tongariro National Park Board Headquarters near the Chateau, 9 km from the lake.

TABLE 1. Hydrothermal Eruption Vibrations in April 1968

Date	N.Z. Time	Peak Seismic Power, Mw	Total Seismic Energy, Ergs	Duration, Sec	Remarks
9.4.68	0525	0.6	10^{14}	70	Doubtful eruption
20.4.68	0313	0.1	10^{13}	50	
20.4.68	0900	0.8	10^{14}	73	
25.4.68	0626	0.5	10^{14}	60 ca.	Steam clouds observe P. Otway at Taupo from 0830 to 1100 NZST
25.4.68	0634	0.3	10^{13}	60 ca.	
27.4.68	0134	6	10^{15}	120	Muddy rain observe Steam cloud observ Eruption observed at Crater Lake
27.4.68	0749	9	10^{15}	130	
27.4.68	0935	4	10^{15}	120	
27.4.68	1335	0.05	10^{13}	50	
27.4.68	1529	0.3	10^{14}	60	
28.4.68	0447	0.3	10^{14}	70	

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Fig. 1. Ruapehu volcano from the East, the day after the hydrothermal eruption on April 27, 1968. Note the small lahar, and the position of the seismometer (arrow), which is 700 meters from the center of the crater lake.

Over the frequency range of powerful volcanic vibrations (1–10 c/s) the output voltage from the amplifier is proportional to ground velocity, and therefore to the square root of the seismic power of body waves per unit area of wave front. The strip chart recorder incorporates a square law amplifier, rectifier, and smoothing circuit, and records the total seismic power on a logarithmic scale of watts (10^7 ergs/sec). The scale calibration is correct for *P* waves radiated spherically from a source 1 km from the seismometer, assuming that the seismic velocity is 3.7 km/sec, the density 2.3 g/cc, and that the vertical ground velocity is equal to the incident velocity.

SEISMIC AND VOLCANIC ACTIVITY

This paper presents the seismic power chart for the period March 31 to May 4, 1968 (Figure 2), during which the magnetometer [Johnston and Stacey, 1969] was recording unusual activity. As visible volcanic activity is confined

Fig. 2. (*Opposite*) The seismic power of volcanic vibrations from Ruapehu, between March 31 and May 4, 1968. *G* denotes gong-like vibrations; *e*, tectonic earthquakes; *E*, eruption vibrations; *B*, vibrations like explosions; *H*, small eruptions observed but not recorded; *I*, vibrations possibly accompanying lava movement; Bars, periods of stronger tremor at the Chateau but not at Ruapehu summit.

Fig. 3. Some volcano-seismic events detected at Dome Shelter, Ruapehu. *G*: A gong-like vibration; 0822 NZST, April 3, 1968, peak seismic power 50 kw. *B*: An explosion-like vibration; 2257 NZST, May 29, 1968, 10 kw, tremor partially ceased during the preceding minute. *I*: 0958 NZST, April 19, 1968, 200 kw; a similar vibration preceded the appearance in August 1966 of fresh pumice on Crater Lake. *E*: A hydrothermal eruption vibration; 1353 NZST, May 30, 1968, 150 kw.

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Fig. 4. Vibrations (*E*) with equally gradual rates of growth and decay that accompanied the largest hydrothermal eruptions of Ruapehu crater on April 27, 1968. Times are NZST. Chateau seismograph, code CNZ (Willmore, T_0 0.7 sec, T_1 0.25 sec), operated by the Department of Scientific and Industrial Research.

within Crater Lake, and the bottom of the lake below the position of greatest activity is 300 meters deep and nearly 1 km from the seismometer, the power cannot be greatly overestimated for the shallowest activity but is underestimated by an order of magnitude or more for vibration sources more than 3 km deep. The Chateau seismograph station (operated by the Department of Scientific and Industrial Research under code name CNZ), which is 9 km from the crater, has recorded volcanic vibrations from what are believed to be relatively deep sources because they were so weakly recorded at the summit. The times of these vibrations are shown with bars immediately above the time scale on Figure 2.

The seismic activity was classified by listening to the slow motion seismograph tapes while playing them back 480 times faster than they were recorded, and also by inspecting the CNZ seismograms. In order of occurrence the types of vibration were as follows:—

1. Between February 29, 1968, when the Dome Shelter seismograph was installed, and April 6, almost continuous vibration occurred with a narrow frequency band whose center frequency varied between 3 and 10 c/s. Characteristically, the frequency rose slowly and dropped suddenly, and the seismic power varied gradually up to 3 kw. These vibrations were not recorded at CNZ and therefore had a very shallow origin.

2. Until April 20, short sharp events were occasionally recorded, which, when played back at 120 times the recording speed, sounded like a gong being struck, and these are labelled *G* in Figure 2. Their onsets contained several discrete frequencies in the range 5–10 c/s, followed immediately by decaying vibrations with a consistent frequency of 3 c/s. Although the largest (Figure 3) reached a momentary power level of 50 kw, they were barely recorded at CNZ, and must have had a shallow origin. They had not been detected before the seismograph was installed at the Dome Shelter.

3. On April 5, tremor with dominant frequency 2 c/s, bandwidth 0.5–6 c/s, and with fluctuations causing a grass-like appearance of the trace in Figure 2, became obvious. This phenomenon had already been recorded during activity in 1964 and 1966, and in 1968 the

conspicuous increases in power on April 5 (10 kw), April 18 (100 kw), April 25 (100 kw), and May 29, (10 kw) caused the Chief Ranger, Mr. J. Mazey, to alert the author and the New Zealand Geological Survey promptly so that observations of the crater could be made or attempted.

4. As the power decreased from the maxima, the tremor became unstable, and sudden stoppages of up to 10 min duration occurred at intervals of the order of an hour. Tremor recommenced impulsively with vibrations like those of explosions (Figure 3), which were well recorded at CNZ, and are labelled *B* on Figure 2. Similar events in 1964 had their foci below the crater lake.

The momentary power of explosion-like vibrations exceeded 1 Mw on April 26, 1968, but the total seismic energy of each did not exceed 10^{14} ergs, and none of the vibrations that occurred while the lake was under observation on April 6 and 7, 1968, correlated with the small geyser-like eruptions in the center of the lake that ejected black water to a height of 30 meters (E. F. Lloyd, personal communication). The source of the explosion-like vibrations was too deep for them to affect the lake surface, whereas the small eruptions, which were almost undetectable on the Dome Shelter seismograph tapes, were probably caused by surface boiling of rising cells of superheated muddy water.

5. After the strong tremor and explosion-like vibrations had decreased, very strong vibration events of about 1 min duration occurred, each of which started gradually, reached a maximum power about half way through, and decreased at about the same rate as it began. Figure 4 is a copy of the Chateau seismogram of the events on April 27, 1968. At the times of nearly all the events, the volcano was obscured by cloud, or it was night, but one of the smaller events at 1335 NZST on April 27, correlated with an eruption from the lake that showered P. Otway and W. M. Crafar (personal communications) with pellets of mud at a distance of 400 meters from the shore. Another at 1105 NZST on June 4, 1968, with a peak seismic power of 0.5 Mw, correlated with an eruption of mud and water to a height of about 600 meters above the lake, and so these vibration events are classified as hydrothermal eruption vibrations, and are labelled *E* on Figure 2. Tape

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recordings played back 240 times the recording speed sound like the momentary release of compressed gas.

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CHAPTER 7.

RUAPEHU VOLCANO 1969

The Eruption on 22 June 1969

From July 1968 to June 1969, Ruapehu was quiet and apparently inactive. At about 0025 NZST on 22 June, the Chief Ranger Mr. J. Mazey awakened, and sensed a feeble explosion followed by a shock wave (personal communication). At approximately 0030 he heard a low long rumbling sound, and about this time 3 other people reported flashes of light from an eruption, and one of them reported 'sparks' from the crater (N.Z. Herald 23 June). At about 0033 a hail of ash and scoria fell on the skiing grounds, 4 km NNW of the crater. Mr. M.W. Rea (personal communication) reported that it lasted about 5 minutes, and that a large black cloud above the crater was silhouetted against the night sky. A second huge cloud shot up, and a little while later more ash fell. Dr. E.B. Jenner (personal communication) noted that ash ceased to fall at the skiing grounds and noise diminished at 0050 NZST. The only time given definitely by an eye-witness was that of 0025.

Reconnaissance flights and inspections by members of the N.Z. D.S.I.R. and V.U.W. showed a lobe of tephra (Fig 7.1) 2-6 km wide extending 16 km north of the crater (E.F. Lloyd, personal communication). The coarser material near the crater included juvenile pumice bombs, while the fine tephra contained between 18 and 95.6% material similar to disintegrated pumice (Dr. P.C. Wood, personal communication). Sizeable lahars had swept down the Whakapapanui, Whakapapaiti (Fig 7.1), Mangaturuturu (Fig 7.2), and Whangaehu valleys (Fig 7.3), the first damaging the ski field kiosk, and the last surging over a 100 m high ridge which stood in its path. Blocks of old lava up to 0.25 m across had been thrown about 1 km onto Dome Ridge at a height of 70 m above Crater Lake, and 1.6 km onto the Whakapapanui Glacier about 100 m below the lake (M.R. Johnston, pers. comm.), while blocks up to 2½ m (Fig 7.3) had been thrown onto the rim. Aided by a strong SSE wind, a base surge had blasted a mantle of ice about 1 m thick off the Dome, ripping up the telephone cable, and completely demolishing the Dome Shelter (Fig 7.4) which had contained the seismograph amplifier. Surprisingly the Crater Lake appeared unchanged (Fig 7.5) except for an apparent drop in level of 2½ m. Although Mr. J. Healy (personal communication) considered it possible that

Fig 7.1: The lobe of tephra from the crater, seen from the north the day after the eruption on 22 June 1969. The Whakapapanui and Whakapapaiti lahars run left of the lobe and right of the centre of the lobe respectively.

Photo by J.J. Whalen

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Fig 7.2: Lahar down the Mangaturuturu Glacier and Gorge seen from the north the day after the eruption on 22 June 1971. Note the spray below the two waterfalls.

Photo by J.J. Whalen.

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Fig 7.3: The lahar down the Whangaehu Glacier seen from the east the day after the eruption on 22 June 1969. The lahar surged over a 100 m high ridge (bottom) after crossing the Whangaehu River. An arrow (top) marks the 2½ m boulder which was thrown onto the Dome during the eruption.

Photo by J.J. Whalen

Fig 7.4: The Dome Shelter, 700 m north of the crater centre, the day after the eruption on 22 June 1969. It housed the preamplifier of the Dome seismograph which stopped recording at 0032.9 NZST. The shelter and ground were covered by about 1 m of hard snow before the eruption.

Photo by L. Homer

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Fig 7.5: Ruapehu crater viewed from the west the day after the eruption on 22 June 1969. The outlet and ice tunnel at right show the lake is about $2\frac{1}{2}$ m below normal.

Photo by J.J. Whalen.

the eruption was of hot ash flow type, this could not be proved by the ground surveys. In particular, no charring of the timber or contents of the Dome Shelter was found.

Seismographs Operating in 1969

In 1969, prior to the eruption, there were 3 seismographs within 16 km of Ruapehu, positions and details of which are given in Fig 7.6 and Table 7.1. Of these, Chateau (CNZ) operated correctly throughout the eruption. Dome operated correctly until the preamplifier in the Dome Shelter and the upper half mile of the telemetry cable were destroyed by the eruption, and Mangatepopo Hut (MH) operated extremely poorly between 16 and 23 June due to battery failure. The Chateau and Dome seismographs were calibrated before the eruption in March and May 1969 respectively, and their relative sensitivities approximately confirmed by recordings of earthquakes about 150 km distant on 20 June 1969. After the eruption, seismographs were temporarily installed at Ski Patrol (site 8 in 1964), Alpine Club Hut, and Mangawhero Falls (site 11 in 1964). The maximum number of seismographs operating at one time was four.

Fig 7.6: Map of Ruapehu, Ngauruhoe and Tongariro volcanoes, showing geophone locations (double circles) in 1969. Dome (telemetry), Chateau and Mangatepopo were operating during the eruption on 22 June. Ski Patrol (Telemetry) replaced Dome on 3 July and Mangawhero was added in August 1969.

TABLE 7.1: Seismographs within 16 km of Ruapehu during 1969

Station Name	Seismometer	Recorder	Recording Range: Ground Velocity; Seismic Power.	Recording Periods in 1969	Remarks
Dome	Willmore Mk 2, Z, To = 2 s h = 1	Tape Recorder, speed 0.2 mm s ⁻¹ Chart recorder for seismic power	1 - 180 μ m s ⁻¹ 30 - 10 ⁶ W	20 April - 22 June	Telemetered by cable to Whakapapa Village
Ski Patrol	Willmore Mk 1, Z, To = 0.7 s h = 0.25	Same as for dome	0.3 - 13 μ m s ⁻¹ 60 - 2 x 10 ⁶ W	1 July - 13 Aug	As for Dome
Alpine Club Hut	Willmore Mk 1 Z, To = 0.7 s h = 0.3	As for dome	0.1 - 30 μ m s ⁻¹ 3 - 3 x 10 ⁵ W	13 August onwards	As for Dome
Chateau (CNZ)	Willmore Mk 1, Z, To = 0.9 s h = 0.5	Photographic drum speed 1 mm s ⁻¹ Tg = 0.25 s	0.4 - 17 μ m s ⁻¹ (4 - 7 x 10 ⁵ W)	Almost all 1969	Operated by Seismological Observatory D.S.I.R.
Mangatepopo Hut	Willmore Mk 1, Z, To = 0.7 s h = 0.25	Portable Tape Recorder speed 0.2 mm s ⁻¹ (Dibble 1964)	0.04 - 400 μ m s ⁻¹ (10 - 10 ⁹ W).	20-23 Jan, 25 Feb-28 Mar, 4 Apr - 5 May, 23 May - 23 Jun, 2 - 17 Jul, 12 Aug - 14 Sept, 28 Sept - 21 Nov.	Field Installation
Mangawhero Falls	Willmore Mk 1, Z, To = 0.7 s h = 0.25	Portable Tape Recorder speed 0.2 mm s ⁻¹	0.04 - 400 μ m s ⁻¹ (2 - 2 x 10 ⁸ W)	15 - 23 Sept. 29 Sept - 21 Nov.	Field Installation

Seismic Activity Prior to 22 June

The activity chart (Fig 7.7) shows that prior to 8 May 1969, only a few small seismic events per month were occurring within 50 km of the Chateau seismograph, and that volcanic tremor did not exceed 1 kW in seismic power. On 8 May, two earthquakes of magnitude $4\frac{1}{2}$ were felt at the Chateau, and were accompanied by one foreshock and 30 after shocks of $M_L \geq 1.8$, up to 15 May. Epicentres determined by Seismological Observatory D.S.I.R. are:

- (1) 8 May 1969, origin time = 10 41 49.7 $\pm$ 0.03 NZST
latitude = $38.97 \pm 0.02^\circ$ S, longitude = $175.75 \pm 0.02^\circ$ E, h = 12 km ca, $M_L = 4.5$
- (2) 8 May 1969, origin time = 12 04 54.4 $\pm$ 0.1 NZST,
latitude = $38.89 \pm 0.01^\circ$ S, longitude = $175.77 \pm 0.01^\circ$ E, h = 12 km ca, $M_L = 4.7$

These epicentres are within the volcanic region but are 43 and 51 km NNE of Ruapehu, and unlikely to be connected with the eruption.

Between 15 May and 11 June, there were 113 local seismic events. Of these, 103 which occurred between 22 May and 5 June resembled explosions in appearance

Fig 7.7: Seismic activity through April - July 1969, within 50 km of the Chateau (CNZ) seismograph. The upper graphs show peak ground velocity of seismic events (left scale) and their nominal seismic power (right scale) assuming they originated under Ruapehu. The lower graphs show seismic power of volcanic tremor from Ruapehu.

(Fig 7.8) and speeded up sound, and appear to be B-type volcanic earthquakes. Most of these were recorded at Mangatepopo Hut, and only 8 were recorded at CNZ, and none were recorded at the Dome. The frequency spectrum at Mangatepopo Hut was similar for all the earthquakes, with a dominant peak between 1.1 and 1.6 Hz, and lesser peaks at 2.3, 2.7, 3.0, 3.5 and 4.1 Hz. Readings for the 5 best recorded events are given in Table 7.2. The average P-time difference ΔT_p between CNZ and Mangatepopo Hut is 1.9 s. As the distance between the seismographs is 5.45 km, the velocity determined by the average ΔT_p cannot exceed 2.9 km s^{-1} . This low apparent velocity indicates a very shallow focus, and unless the actual velocity is less than that, the mean epicentre must lie along the projection of the line from CNZ to Mangatepopo Hut. The epicentres could then be in the region of Ngauruhoe and Tongariro, but must be at least 12.3 km from Ruapehu. The remaining 10 earthquakes (only one of which occurred between 5 and 21 June), cannot be located with the available data, but were not close to the Dome seismograph. In fact, during the 3 months prior to 22 June 1969, no seismic events (excluding tremor) that can reasonably be ascribed to Ruapehu have been found.

Fig 7.8: B-type volcanic earthquakes recorded on tape at Mangatepopo on 3 and 4 June 1969, but not at Dome. They probably originated under Tongariro or Ngauruhoe. Note the prominent surface wave $3\frac{1}{2}$ sec. after P, and the lack of S.

TABLE 7.2: Seismograph Readings of the Best Recorded Probable Volcanic Earthquakes through 3 - 5 June 1969. P Times are Reliable to 0.2 s.

Date	Chateau Seismograph (CNZ)				Mangatepopo Hut Seismograph (MH)				P-time Diff.		Remarks
	Phase	NZST	Ground Vel.	Freq.	Phase	NZST	Ground Vel.	Freq.	ΔT_p		
		h m s	$\mu m s^{-1}$	Hz		h m s	$\mu m s^{-1}$	Hz		CNZ - MH	
3.6.69	P	01 07 24.1	0.3	2.6	i P	01 07 22.1	4	3		2.0	CNZ record marred by glass strip which secured record to drum.
	e	27	0.2	1.7							
	e	30	0.6	2.0	e L	24.7	6	1.2			
	e	34	0.5	1.2							
4.6.69	i P	05 46 50.9	0.6	3	e P	05 46 48.5	0.9	4		2.4	Onset at Mangatepopo confused by ghost recordings on the tape, 6 s ahead of the true recording.
	e L	54	1.7	1.6	i	49.8	3	3			
					e L	51.3	11	1.2			
4.6.69	e P	05 53 03.4	0.1	2.6	i P	05 53 01.7	0.6	3.5		1.7	
	e (L)	06.3	0.1	1.7	e L	04.0	3	1.2			
	e L	09	0.3	1.3							
4.6.69	i P	07 07 52.4	0.4	2.8	e P	07 07 50.4	1	3.5		2.0	
	e	55.4	0.2	1.4	i	51.6	3	3			
	(L)	08 00.5	0.8	1.7	e L	53.6	7	1.2			
5.6.69	P	22 44 47.4	0.4	3	P	22 44 46.0	2	3		1.4	
	(L)	53	0.6	1.0	e L	49.2	5	1.2			

Seismic Recordings of the Eruption on 22 June 1969

Readings of seismic events at CNZ and the Dome associated with the eruption are given in Table 7.3, and seismic power and power spectra versus time recorded at the Dome are shown in Figures 7.9 and 7.10. The foci of events cannot be located from recordings at only two stations, but to simplify discussion it has been assumed that all lay vertically below Crater Lake, and nominal depths (Fig 7.11) have been calculated (Table 7.4) from S minus P intervals, P-difference times, and ground velocity ratios between CNZ and Dome (Fig 7.12).

Volcano-seismic activity began to record at both CNZ and Dome at 0004.0 NZST, and by 0006 six small A-type (Minakami 1961) earthquakes of magnitude 0.9 to 1.6 had occurred at nominal depths between 13 and 45 km. They had prominent P phases with spectral peak 3 - 4 Hz, and apparent S phases at CNZ, but no S at Dome, due perhaps to absorption in intervening magma. During these earthquakes, and partly obscuring them, volcanic tremor with spectral peak 1.5 - 3.2 Hz began gradually, and reached a seismic power of 200 W. Between 0006.9 and 0010 NZST,

TABLE 7.3 SEISM

Durations are the total
the first, when they ref
1 s for the Dome se
for both recorders are 1

Chateau Seismograph (CNZ)						Phase Time	
Phase	Time	NZST	Ground velocity	Duration	freq.	Phase	Time
	h	m	s	$\mu\text{m s}^{-1}$	s	Hz	h
iP?	00	04	00.0	0.5	9	3 ca	
e(S)			04.1	0.4		1.7	
i	00	04	10.9	0.8	5	2.4	e 00
P	00	04	23.0	0.8	11	3.5	eP 00
eS			26	1.8		2.6	
iP?	00	04	45.8	3.5	12	4 ca	eP 00
eS			49.5	1.7		2½ ca	
eP?	00	04	59.9	0.9	16	?	P 00
eS		05	03.5	2.3		2	
iP?	00	05	58.5	4	18	4	e? 00
e(S)		06	04.3	1		1.5	
iP	00	06	56.3	3.7	25	1.8	iP 00
P	00	07	42.3	5	25	?	iP 00
							i
							i
i(P)?	00	09	20.0	0.8	-	1.6	(P) 00
(P)			22.9	1.2			
P?	00	10	38.5	0.5			eP? 00
(P)			39.9	2.3	19	2 ca	e
e(P)?	00	12	30	1	?	2 ca	e(P) 00
P	00	13	29.3	3½	16	2	eP 00
							e
eP ?	00	14	21.6	0.7	27		iP 00
(P)			23.5	7	25	?	
							(P) 00
iP?	00	20	03.9	≥11	32	?	iP 00
							eL
P	00	24	58.3	≥10	38	?	P 00
							i
e	00	31	10.3	1	-		e? 00
			32.0	2.3			e
i	00	32	30.0	>40	?		i 00
							00

Fig 7.9: Seismic power recording from the Dome, for the eruption on 22 June 1969. Levels below 0.1 kW are affected by interference. The peaks of 2-20 kW through 0007 to 0025 are volcanic earthquakes. Tremor increased at 0007, decayed again after 0012, and the major increase began at 0030, and was overloading the seismograph when it was damaged by the eruption.

Fig 7.10: Sonagrams of volcanic earthquakes and tremor, recorded at Dome during the eruption on 22 June, 1969.

Times are minutes after 0000 NZST.

Upper: Dominant frequencies are 2.5 - 4.5 Hz before 0006 NZST, 1 - 3 Hz during 0006 - 0031, 1.5 - 7 Hz during 0031 - 0032.5. Seismograph overloaded to 0032.5, and was damaged at 0032.9 by eruption. Frequency resolution 0.2 Hz. Lines at 5.3 and 3 Hz are spurious.

Lower: Profiles of stability 6 db with 90% confidence, 0 db $\equiv$ 1 W Hz⁻¹ until 0026 NZST, 16 W Hz⁻¹ after 0026.

Fig 7.11: Change in nominal depth with time, for the volcanic earthquakes and onsets of tremor immediately preceding the eruption on 22 June 1969.

TABLE 7.4: Seismic Evidence of Changing Focal Position of the Earthquakes Immediately Preceding the Eruption of 22 June, 1969.

Time of Event	S - P Interval at CNZ	Ratio of Ground Vel.	Difference in Onset Time CNZ - Dome	Slant Distance from S - P at CNZ	Difference in Distance $\Delta T_p \times V_p$, km	Apparent Velocity Assuming Surface Focus at Crater Lake	Depth Below Lake Bottom, Assuming Foci Vertically Below	Remarks
hr min	s	CNZ Dome	s	km	Vp=6km/s, 4km/s	Va = $\frac{8.51}{\Delta t}$ km/s	km	
00 04.0	4.1	-	-	32	-	-	32	Doubtful
04.1		0.4	4.7	-	-	1.8	zero?	Volcanic Tremor
04.4	3.0	0.36	0.2	23	1.2	42.5	21.5	
04.8	3.7	1.75	0.1	29	0.6	85.1	28	
05.0	3.6	0.46	0.4	28	2.4	21.2	13-27	
06.0	5.8	2.7	-	45			45	
06.9	No S	0.1	2.3	-	13.8	3.7	zero	
07.6	"	0.16	2.7	-		3.2	zero	
09.3		0.16	5.4 (or 2.5)	-	10	1.6	zero?	Volcanic tremor
10.6	"	0.26	1.7 (or 0.1, 3.1, 1.5)	-	6.8?	5.0	2?	Doubtful
12.5	"	0.1	3.7?	-		2.3	-	Doubtful
13.5	"	0.19	1.6 (or 4.9)		6.4	5.3	2.5?	
14.4	"	0.13	1.4 (or 3.3)	-	5.6	6.1	3.5?	
20.1	"	>0.18	1.3	-	5.2	6.6	4.2	
25.0	"	>0.16	2.0	-	8.0	4.3	0.5	
31.2		0.1	5.4 (or 8.4)	-		1.6	zero	Volcanic tremor
32.0		0.08	-	-		-	-	"
32.5			6.2			1.4	zero	"

274"

Fig 7.12: Evidence of changes in focus of the earthquakes and onsets of tremor leading to the eruption on 22 June 1969. Assuming that only depth was varying, greater depth is indicated by a larger ratio of ground velocity, and a smaller onset time difference between Chateau (CNZ) and Dome recordings.

two B-type volcanic earthquakes ($M_L = 1.2$ at CNZ) occurred at zero nominal depth. Their spectral peak of 1.5 - 3.2 Hz was different from that of the A-type earthquakes, and they reached 55 kW (from CNZ) in seismic power. At 0009.3 NZST the seismic power of tremor increased suddenly to 2 kW (CNZ), and its apparent velocity of 1.6 km s^{-1} (Table 7.4) indicates surface wave propagation and a very shallow origin. Between 0010.6 and 0025 NZST, a further 6 B-type volcanic earthquakes (M_L 0.9 to >1.7 , spectral peak 1.5 - 2 Hz) occurred at nominal depths between 0.5 and 4.2 km, and with seismic powers up to >500 kW. Meanwhile the seismic power of tremor decreased to 200 W (CNZ) or less (Dome). Tables 7.4 and 7.5 tabulate the data on the B-type earthquakes which might enable any explosion earthquakes among them to be distinguished.

There followed a very quiet period at both CNZ and Dome between 0026 and 0030. No appreciable eruptions or local lightning occurred in the period because the Dome seismograph and connecting cable have proved very sensitive to both, and neither were recorded.

At 0031.1 NZST, tremor of seismic power 12 kW (CNZ) began at zero nominal depth, and at the Dome had an essentially flat spectra between 1.5 and 6 Hz. At 0032.5

TABLE 7.5: Diagnostic Parameters for Distinguishing Between B-Type
and Explosion Earthquakes on 22 June 1969.

Time NZST	$\frac{\text{Amplitude of L}}{\text{Amplitude of P}}$	Air Wave Recorded at CNZ?	Nominal Depth km	$k=A_t^{-1}$ at CNZ	Seismic Energy ⁴ $J \times 10^4$
0006.9	2	No	0	0.153	24
0007.7	2	No	0	0.176	46
0010.6	1.7	No	2?	0.163	7
0012.5	?	No	?	-	1
0013.5	2.5	No	2.5?	0.198	14
0014.4	3.1	No	3.5?	0.165	99
0020.1	1 ca	Doubtful	4.2	0.171	300
0025.0	4.5	No	0.5	0.154	280

the tremor increased sharply and completely overloaded all the seismographs except Mangatepopo, which was insensitive due to battery failure. This seismograph recorded the tremor, but not the preceding volcanic earthquakes. By relating the 3 records, the seismic power of tremor after 0029.5 was estimated as in Table 7.6.

TABLE 7.6: Seismic Power and Frequency of
Volcanic Tremor, 0029.5 - 1900 NZST
22 June 1969

Time NZST	Seismic Power as from CNZ,kW	Frequency Hz	Remarks
0029.5-0031.1	0.05 - 0.2		Gradual Increase
0031.1-0032.5	12	1.5 - 7.5	Power from CNZ
0032.5-0032.75	7500	1 - 10	Power from MH
0032.9			Dome seismograph damaged by eruption
0033	650		Power from MH
0034	4500		Power from MH
0034 $\frac{1}{4}$ - 0035	650		Power from MH
0035 - 0036 $\frac{1}{2}$	1450		Power from MH
0036 $\frac{1}{2}$ - 0038	500		CNZ and MH
0039 - 0041	120		CNZ
0041 - 0042	1000		MH
0042	2500		Briefly; MH
0042 - 0043	2500 - 60		Decreasing; CNZ
0043 - 0048	60 - 5		Decreasing; CNZ
0049 - 0100	5		Constant; CNZ
0100 - 0300	5 - .05	1.52	Decreasing; CNZ
0300 - 1330	0.05	1.6 and 2 $\frac{1}{2}$	Constant; CNZ
1330 - 1745	0.2 - 1.2	2	Fairly constant CNZ
1745 - 1900	0 - 10	1.8	Bursts and stoppages; CNZ
1900 onward	0.2 - 0	2	Decreasing; CNZ

The tremor was proved to be accompanied by eruptions by the damage to the Dome seismograph at 0032.9. The tremor spectrum was broader and higher after 0031.1 NZST than earlier, indicating a change in the generation of tremor, consistent with an eruption. After 0048 NZST, the tremor recorded at CNZ was less than 5 kW in seismic power, and had a fairly narrow spectral band width, and varied in frequency. It did not appear to be the vibration of lahars, and appeared to have greater seismic power than that of the lahar which followed the breakout of $2 \times 10^6 \text{ m}^3$ of water from Crater Lake on 24 December 1953.

ANALYSIS OF DATA ON B-TYPE EARTHQUAKES

The ratio of L-wave amplitude to P-wave amplitude given in Table 7.5 is probably the most sensitive means of distinguishing explosion earthquakes (those accompanying surface explosions) from B-type earthquakes. Minakami (1950) states that this ratio is 5 - 10 for explosion earthquakes at Asama volcano. At Ruapehu the earthquake at 0025.0 NZST (Fig 7.13) most nearly reaches this range, and that at 0020.1 NZST is farthest below it.

The other parameters single out the latter as also the deepest earthquake and the only one with possible (but probably spurious) traces of an air wave. At the time, the seismometer of CNZ was in a garage in sight of Ruapehu, and was sensitive to wind vibrations. It should have recorded any air waves strong enough to be felt at Whakapapa Village. Reports of B-type or explosion earthquakes being felt are evidence of explosive eruptions, because Minakami (1969) has found that the ground vibrations of B-type earthquakes are never felt but the air waves of explosion earthquakes often are.

The value k for a series of earthquakes, (where $k = A^m t^{-1}$, A is the maximum amplitude of each earthquake, t is the duration, and m is a constant), was investigated for earthquake swarms of volcanic significance by Kizawa (1965) and is used here (with A = double trace amplitude in mm on CNZ) to reveal differences in duration of B-type and explosion earthquakes. It shows that the earthquake at 0013.5 NZST has an anomalously short duration, and is less likely than the others to be an explosion earthquake. Finally, Minakami (1969) reports that ground amplitudes of B-type earthquakes at Asama Volcano have never exceeded

10 micrometre at 4.1 km from the epicentre. This corresponds to between 2.3 and 5 micrometre at the distance of CNZ, depending on whether amplitude varies inversely with distance or with distance squared. Figure 7.13 shows that the amplitude of the earthquake at 0025 NZST was 1 micrometre on CNZ, and therefore this event could be either a B-type or an explosion earthquake.

Briefly the analysis of the B-type earthquakes reveals none which are very likely to be explosion earthquakes; one at 0020.1 NZST which is most unlikely to be; five which are probably not, and one at 0025 NZST which may be an explosion earthquake. This latter appears to be two earthquakes separated by 3.3 second (Fig 7.14).

THE ENERGY OF THE ERUPTION ON 22 JUNE

The total seismic energy of volcanic earthquakes and of tremor have been estimated from seismic powers and durations at CNZ (Tables 7.3, 7.5 and 7.6). Results are:

Total seismic energy of		
volcanic earthquakes	=	$8 \times 10^6 \text{ J}$
Total seismic energy of		
volcanic tremor	=	$7 \times 10^8 \text{ J}$

Fig 7.13: Seismograms from CNZ of the Ruapehu eruption on 22 June 1969.

(a) The event at 0025 NZST may have been an explosion earthquake which started the eruption. The trace has been reinforced. Undoubted eruption vibrations are visible in the background.

(b) The beginning of strong tremor at 0031 and climactic tremor at 0032.5 NZST. The onsets, which are outlined, are seen against a background of decreasing tremor recorded 15 minutes later.

0025

Fig 7.14: Seismograms from the Dome for the eruption on 22 June 1969.

Upper: The initial shallow volcanic earthquakes at 00 06 54 and 00 07 40 NZST, followed by tremor. No eruption observed.

Middle: Volcanic earthquake at 00 24 56 NZST which may have awakened the ranger, and been accompanied by eruption. Note the quiet background.

Lower: Climactic increase of tremor at 00 31 05 and 00 32 24, and cessation of recording at 00 32 56, presumably when the Dome Shelter was destroyed by the eruption.

The kinetic energy E of the eruption can be found from $E = \frac{1}{2} M V_0^2$ if the mass M and initial velocity V_0 of ejecta are known. The volume of ejecta can be estimated from data collected by the N.Z. Geological Survey (Table 7.7, Fig 7.15) from which graphs of tephra thickness and angular width of lobe versus distance can be drawn (Fig 7.16). Fortuitously the survey route up the volcano lay near the eastern edge of the lobe, and it can be assumed that the average thickness across the lobe is equal to at least the thickness measured along the route. Results were:

Volume of tephra deposited within

$$34 \text{ km of the crater} = 9 \times 10^5 \text{ m}^3$$

Volume of water apparently lost

$$\text{from Crater Lake (area } 0.2 \text{ km}^2) = 5 \times 10^5 \text{ m}^3$$

Total mass M ejected, assuming a

$$\text{mean bulk density of } 1 \text{ Mg m}^{-3} = 1.4 \times 10^9 \text{ Kg}$$

The result is not of good accuracy, partly because the true mass of water ejected is unknown. Dr. R.B. Glover reports (personal communication) that the chloride content of the lake decreased by 5% in the eruption, probably by dilution with snow-water melted by the ejecta. The volume of melt water required is 5% of the lake volume

TABLE 7.7: Measurements of the Ejecta from the
Eruption of Ruapehu on 22 June 1969.

Distance from Crater, km.	Location	Thickness mm	Remarks	Observer	Date
0	Crater Lake	2500	Drop in lake level	R.B. Glover	26 June 1969
0.5	Nth side of crater.	300	Coarse pumiceous ash, unfrozen, variable.	M.R. Johnston D.L. Homer	23 June
0.7	Head of Whakapapaiti Glacier	300	Ash	E.F. Lloyd	26 June
1.6	Restfull Rocks	90-120	Ash	W.M. Crafar	26 June
2.5	N122; 051684	25	Ash	E.F. Lloyd	25 June
2.7	N122; 052686	20-25	Ash	D. Rishworth	25 June
2.9	Knoll Ridge	50-70	Ash	W.M. Crafar	26 June
3.1	Lower Knoll Ridge	20	Ash layer snow covered, Sample 44% water by weight	R.B. Glover	26 June
3.2	Alpine Club Hut	15	Ash	E.F. Lloyd	25 June
3.2	Alpine Club Hut	10-20	Ash	D. Rishworth	25 June
3.7	2nd Chair Lift	10-50	Ash	W.M. Crafar	26 June

Distance from Crater, km.	Location	Thickness mm	Remarks	Observer	Date
3.9	Track to A.C.Hut	7-8	Ash	D. Rishworth	25 June
4.0	R.S.C. Lodge	8-9	Ash	W.M. Crafar	26 June
4.1	Staircase Kiosk	5-6	Ash	D. Rishworth	25 June
4.5	1st Chair Lift	5-8	Ash	W.M. Crafar	26 June
5.1	Top 0' the Bruce	8	Fine dk grey ash, frozen pisolites, _<10 mm diam.	E.F. Lloyd	22 June
5.1	Top 0' the Bruce	4-5	Fine ash with sub- ordinate silt and minor fragments up to 20 mm, partly pelletised.	D. Rishworth	25 June
16	Rangipo-National Park Highway at Mangate- popo Hut turn off.	0.2?	Fine grey ash detectable on road	E.F. Lloyd	22 June

Fig 7.15: Map of qualitative distribution of tephra
from the eruption of 22 June 1969. Drawn by J. Healy,
N.Z. Geological Survey.

ERUPTION OF RUAPEHU, JUNE 22, 1969

Area with complete cover of dark grey ash & debris

Area with almost complete dark grey cover

Area with thin ash deposit

Lahar courses and deposits

1 0 1 Miles 2

Drawn by J. Healy,
July, 1969

Fig 7.16: Thickness and angular width of the tephra lobe deposited during the 22 June 1969 eruption, plotted against distance from the crater. Data is indicated by dots and range bars.

(10^7 m^3 in 1966), $= 5 \times 10^5 \text{ m}^3$. An equal amount of lake water must have been lost from the lake, additional to the apparent drop in volume, while a further similar quantity was no doubt ejected but ran back in. The velocity of ejection was calculated assuming that the large (0.25 m diameter) rocks found on snow at or above crater elevation and furthest from the crater (1 km), had an initial angle of ejection of 45° , and trajectories unaffected by air or wind.

$$\text{From } V_o^2 = \frac{Dg}{\sin 2 \times 45},$$

where V_o is initial velocity, D is range, and g is gravity, the result is:

$$V_o = 99 \text{ m s}^{-1}$$

Then:

$$\text{Kinetic energy of the eruption} = 7 \times 10^{12} \text{ J.}$$

This may be a maximum estimate because although M may be underestimated, the average velocity of ejection (which is squared) may have been less than that of large blocks. On the other hand, Fudali and Melson's (1972) curves of initial velocity versus range in still air give 190 m s^{-1} for the initial velocity of rough round blocks of diameter 0.25 m and density 2.5, thrown 1 km

from the vent. The strong wind in nearly the direction of the blocks invalidates this figure however. The rate and time of release of kinetic energy during the eruption may be estimated from the seismic data. Minakami (1950) has determined from observations of over 600 explosions at Asama volcano in Japan that the kinetic energy of explosive eruptions is proportional to the seismic energy of the accompanying explosion earthquake. The ratio
$$\frac{\text{Seismic energy of explosion earthquake}}{\text{Kinetic energy of explosive eruption}}$$
 is given as 10^{-2} at Aso volcano by Shimozuru (1968), and 10^{-2} to 10^{-4} at Asama and Sakurajima volcanoes by Sakuma and Nagata (1957). The values are in accord with the seismic efficiency of large (200 kg) seismic explosions in rock for frequencies below 10 Hz, given as 0.1% by O'Brien (1967, p. 163).

The seismic energy of the earthquake at 0025 NZST was about 3×10^6 J, and so the kinetic energy of the possible eruption at that time was in the range 3×10^8 to 3×10^{10} J, which is 0.01% to 1% of the total kinetic energy of eruption. If all the B-type earthquakes were accompanied by eruptions their combined contribution

would be at most 2% of the total kinetic energy. This implies that nearly all of the eruption occurred during the strong tremor. Alternatively, the total kinetic energy may have been over-estimated. A check is available from the ratio

$$\frac{\text{Seismic energy of volcanic tremor}}{\text{Kinetic energy of eruption}}$$

which Shimozuru gives as 10^{-5} for Aso volcano, which in several ways behaves similarly to Ruapehu. The seismic energy of volcanic tremor in the 1969 Ruapehu eruption was of the order of 7×10^8 J, and so the kinetic energy would be 7×10^{13} J if the relation for Aso applies to Ruapehu. It appears probable that the total kinetic energy calculated from the ejecta has not been over-estimated, and that practically all of it was released during the period of extremely strong tremor between 0032.5 and 0038 NZST. The average power of kinetic eruption in this period was about 10^{10} W, and the peak period at 0032.7 NZST scaled from the peak seismic power was of the order of 10^{12} W.

Seismic Activity after 22 June

8 July 1969

Seismic activity after the eruption was low until 1716.1 NZST on 8 July 1969, when tremor (1.9 Hz) and B-type volcanic earthquakes (Fig 7.7, 7.17) equalling those of 22 June began. There were no associated A-type earthquakes. Activity can be summarised as follows:

8d	1716.1-1717.5 NZST	Tremor increases to a maximum of 1 kW (1.6Hz)
	1717.9	Principal B-type earthquake $M_L = 1.6$ (CNZ)
	1718.8-1908.6	30 B-type earthquakes or pulses of tremor, M_L 0.1 to 1.5, during which background tremor decreases to zero.
	1908.6-1916	Tremor increases to ≥ 9 MW within 0.8 minutes.
	1916 - 2140	Tremor varying between 10 and 30 kW.

Fig 7.17: Volcanic tremor and B-type earthquakes from Ruapehu on 8 July 1969, recorded at Ski Patrol.

Upper: Earthquake at 17 29.9 NZST which occurred minutes after tremor first began.

Lower: The last 2 of a series of volcanic earthquakes or pulses of tremor which followed the earthquakes above, and ended at 19 08.6 NZST in prolonged tremor reaching 9 MW. No eruption was observed.

8.9d - 9.0d	Tremor varying between 2 and 10 kW
9.0d - 10.9d	Tremor varying between 0.6 and 10 kW
10.9d - 13.2d	Tremor varying between 0.1 and 9 kW
13.2d - 16.9d	No appreciable seismic activity.

From the relation between seismic and observed activity of Ruapehu in 1966 and 1968, an eruption should have accompanied the activity, but none was reported.

16 July 1969

At 2336.4 on 16 July, a further brief period of strong seismic activity (Fig 7.7) began as follows:

16d	2336.4-2337.4	Tremor increases to 35 watt.
	2337.4-2339.1	2 B-type Volcanic earthquakes $M_L = 0.9$ and 1.3 (Fig 7.18).
	2341.0-2346.3	Volcanic Tremor, 1.4kW, 1.6 Hz.
	2346.3-2349.3	Tremor increases to 2 MW (Fig 7.18)
	2351 - 2354	Tremor decreases almost to zero.
17d	00.0 - 01.0	6 B-type Volcanic earthquakes, $M_L = -0.1$ to 0.5 .

Fig 7.18: Volcanic tremor and B-type earthquakes from Ruapehu on 16 July 1969 recorded at Ski Patrol.

Upper: The impulsive earthquake at 23 39.1 NZST which occurred 2.7 minutes after tremor began. It has no clear S phase.

Lower: Tremor increased to 2 MW after 2346.2 NZST, and decreased to normal 6 minutes later.

Again no actual eruption was reported.

The readings of the 5 best recorded volcanic earthquakes and the onsets of strong tremor during 8 and 16 July 1969, are given in Table 7.8. Only 4 of the earthquakes had clear onsets at 2 or more stations, and times for these were earliest at Ski Patrol, and latest at Mangatepopo Hut, suggesting that the epicentres were in the vicinity of Ruapehu. As the data are insufficient to determine the epicentres, the possibility of their all being at Crater Lake was tested by examining the consistency of the apparent velocities which would result (Table 7.9). The range of 2.4 to 5.5 km s^{-1} for V_p , and the lack of trends explainable by change of depth alone, suggests that the epicentres were not all at Crater Lake.

The large increase of tremor on 8 July 1969, had a very low apparent velocity of 0.6 km s^{-1} , which is not inconsistent with surface waves originating at the crater. That on 16 July was less equivocal, with apparent velocities from 0.4 to 2.1 km s^{-1} . This may be due to errors in picking the increases, or both body and surface waves may have been involved.

TABLE 7.8

Seismograph Readings of the Best Recorded Volcanic Earthquakes
and Tremor Onsets During 8 and 16 July 1969.

SKI PATROL (SP)			CHATEAU (CNZ)			MANGATEPOPO HUT (MH)			Remarks
Date	Phase	Time NZST	Vel. Ampl.	Freq. Hz	Phase	Time NZST	Vel. Ampl.	Freq. Hz	
July		h m s	$\mu\text{m s}^{-1}$			h m s	$\mu\text{m s}^{-1}$		
8	i P	17 17.8	>13	1.4	i P	17 17 53.7	13	2.5	i P 17 17 55.0 9 1.7
									Clear earthquake. Times at Ski Patrol inaccurate.
8	e P?	17 29 53.8			P	17 29 56.2	7	2.2	P 17 29 57.3 5 1.4
	(P)	54.2	>12	2					Fairly clear earthquake.
8	P	18 16 08.2	11	1.5	i P	18 16 10.3	4.4	1.5	P 18 16 12.1 3.6 1.4
									Low freq. earthquake.
8	e P	19 07 34±1	1.7	2.5	e P	19 07 37.5	0.7	2	e 19 07 37.8 0.8 1.8
	i	36±1	9	2.5	i	40.7	1.7	2.5	Earthquake with stepped onset. Times at Ski Patrol inaccurate.
8	e?	19 08 27.5	2.5	1.5	e?	19 08 34.0	1.0	1.5	e? 19 08 36.0 2.2 1.3
	i	29.5	>13	1.5	e	41.0	4.3	1.7	Definite increase in tremor 50.3 5 1.6
					e	52	9	?	Further increase in tremor.
	e	09 00	>13	2	e	09 06	>9	?	i 09 14.8 >10 1.6
									Renewed increase in tremor

SKI PATROL (SP)

CHATEAU (CNZ)

MANGATEPOPO HUT (MH)

Date	Phase	Time NZST	h	m	s	Vel. Ampl.	Freq. Hz	Phase	Time NZST	h	m	s	Vel. Ampl.	Freq. Hz	Phase	Time NZST	h	m	s	Relative Ampl.	Freq. Hz	Remarks
July						$\mu\text{m s}^{-1}$	Hz						$\mu\text{m s}^{-1}$	Hz								
16	i P	23 39 07.6				>13	1.6	i P	23 39 08.5				8	1.6	e P	23 39 10.1				1.5	1.6	Clear earthquake
16	e	23 46 14.8				1.5	1.5	i?	23 46 17.8				0.6	2	e?	23 46 23.6				0.2	1.8	Onset of strong tremor
	e	19.2				3	1.7	e	21.6				1	1.5	e	25.5				0.3	1.8	Increase in tremor
	i	23.2				13	2.1	e	28.2				2.5	1.7								Further increase in tremor
	i	53.3				13	2.0	e	55				2.2	1.7								Renewed increase in tremor
	e	47 28.3				>13	2.0	i	47 41.5				5	?		47 54				1	1.5	Renewed increase in tremor.
	e	48.04				>>13	1.7	i	48 09				>14	?	e	48 19.6				2.5	1.5	Big increase in tremor
																51				>10		Maximum tremor.

TABLE 7.9:

Arrival Time Differences and Apparent Velocities of Volcano-
Seismic Waves on 8 and 16 July 1969, Assuming an Epicentre at
Ruapehu Crater Lake.

Date	Phase	Time		CNZ - SP		MH - CNZ		MH - SP	
		h	m	ΔT	$Va = \frac{4.95}{\Delta T}$ km s ⁻¹	ΔT	$Va = \frac{5.45}{\Delta T}$ km s ⁻¹	ΔT	$Va = \frac{10.8}{\Delta T}$ km s ⁻¹
July				s		s		s	
	i P	17	17.8			1.3	4.2		
	P	17	29.9	2.0	2.5	1.1	5.0	3.1	3.5
	P	18	16.1	2.1	2.4	1.8	3.0	3.9	2.8
	e (L)	19	08.5	11.5	0.4	9.3	0.6	21	0.5
	e (L)	19	09.1	6	0.8	8.8	0.6	14.8	0.7
	i P	23	39.1	0.9	5.5	1.6	3.4	2.5	4.2
	e (L)	23	46.2	3	1.6	5.8	0.9	9.8	1.1
	e (L)	23	46.3	2.4	2.1	3.9	1.4	6.3	1.7
	e (L)	23	47.4	13.2	0.4	12.5	0.4	25.7	0.4
16	e (L)	23	48.0	5	1.0	10.6	0.5	15.6	0.7

13 July 1969

On 13 July 1969, between the above two periods of tremor and volcanic earthquakes, there was an earthquake swarm which began at 1558 NZST and was of 56 minutes duration. In that time at least 212 earthquakes occurred, in which the P wave was one to three times as large as S and often only 0.5 s in duration. The dominant frequency was 5 Hz, and S was sometimes very clear (Fig 7.19) but more often vague. Magnitudes ranged from -1.6 to + 1.7, and the Ishimoto-Iida parameter m had the value $m = 1.9$. This is a normal value for earthquakes not associated with volcanic outbursts, and not of extremely shallow origin.

The S minus P intervals were close to 1.7 s at Mangatepopo and 2.5 s at CNZ for those earthquakes with a clear S, and for the 6 best recorded earthquakes, P arrived 1.0 to 1.2 s earlier at Mangatepopo Hut than at CNZ. Using the usual relation $T_p = 1.3 (S - P)$, and a P velocity of 4 km s^{-1} determined in 1964, the ray path distances were calculated as 3.8 km from Mangatepopo, and 13 km from CNZ. Bearing in mind that the value of

Fig 7.19: Earthquake swarm on 13 July 1969, recorded at Mangatepopo Hut. The majority were similar to that at the second minute mark, and had no apparent S, but the estimated location of the swarm under Tongariro is based on the earthquakes which have clear P and S, as shown here.

m excludes extremely shallow depths, these distances define a focal surface up to 7 km deep below the area of Tongariro Volcano.

Thus the swarm is notable not for its nearness to Ruapehu but because it was a rare event in the area, and occurred close to the date of the Ruapehu eruption.

CHAPTER 8

SUMMARY OF RUAPEHU ACTIVITY FROM 1961 TO 1970

Patterns of Activity

Enhanced volcanic and seismic activity occurred in August - December 1961, April - October 1964, July 1966, April - June 1968, and June 1969. The average interval between enhancements in this period was 2 years, and there was a tendency for enhancement to occur during the winter months. However, the lists of activity by Gregg (1960) show that this is not a permanent feature of the activity. Each enhancement consisted of one or more spasms which were versions of the following synthesized sequence of phases:

- (1) A very few B-type volcanic earthquakes.
- (2) A notable increase in the seismic power of 2 Hz volcanic tremor to 1 kW or more.
- (3) Tremor mechanism becomes unstable, with many stoppages and sudden powerful recommencements of tremor, dying to a period of quiet during which micro-earthquakes occur within the volcano. Heat output through the crater lake increased to 800 MW or more during this phase.

- (4) Eruptions of mixed secondary and primary ash through the crater lake, having kinetic energies up to 10^{12} J, and accompanied by vibration with roughly equally gradual growth and decay, and seismic power up to 7.5 MW. The proportion of undoubted primary ash is small. Individual eruptions during this phase occur without individual premonitory activity.
- (5) Harmonic tremor of frequency 1.5 to 2 Hz followed some eruptions, and possibly was accompanied by upward movement of a lava plug.

Over the years 1961 to 1969, the trend was for phases 4 and 5 to develop and intensify, while the total duration of a sequence decreased. In the 1969 sequence, phases 2 through 4 were of only 20 minutes duration. Only in this sequence were clearly associated A-type earthquakes recognised, and they were very small.

Variations in Thermal Power Output

Measurements of the temperature of the lake until 1970 are listed in Table 8.1. The heat loss from the lake at these temperatures has been determined using

TABLE 8.1: Temperature Measurements in Ruapehu Crater Lake,

1886 to 1970

Date	Temperature °C	Source	Remarks
1886 Jan	0	Gregg (1960) reporting Park,	Lake frozen
1895 Apr 5	53	" " " Dunnage,	Following an eruption
1924 Jan	10	" " " Blyth,	
1926 Aug	0	" " " Cowan, Thompson	Frozen
1930	<37	" " " Grange, Williamson	Below bloodheat
1945 July	58.4	Oliver (1945)	During eruption
1954 Oct 20	36.5	J. Healy (pers.comm.)	
1956 Mar 3	22½	" " "	
1956 Dec 19	33	" " "	
1959 Jun 12	32	" " "	At 1.5 m depth
1959 Dec 16?	36	" " "	Highest measured value
1962 Jan 21	26-30	" " "	Measured at outflow and other points

Date	Temperature °C	Source	Remarks
1964 Mar 20	22-28	J. Healy (pers.comm.)	Measured at outflow and other points
1964 May 26	41-54	" "	Measured at several points
1965 June 18	16.6	" "	
1965 Sept 4,7, 24		R. Dibble	Average over lake surface
1965 Dec 5	29.5	J. Healy (pers.comm.)	
1965 Dec 24	32	J.W. Cole (pers.comm.)	At NW beach
1966 Jan 27-28	40	J. Healy "	
1966 Jan 15	45	G.R.T. Clacy "	At NW beach
1966 Feb 1	40	N. Blackbourne (pers.comm.)	
1966 Feb 15	45-50	D. Dickinson "	At NW beach
1966 Mar 9	46.6	T. Hatherton "	
1966 Mar 24	47	" "	
1966 May 15	33.5	R.R. Dibble	At lake centre

Date	Temperature °C	Source	Remarks
1966 June 18	31.0	R.R. Dibble	At lake centre
1966 July 24	53	J. Healy (pers.comm.)	
1966 July 25	53	R.B. Glover "	At outlet
1966 July 28	53.5	R.R. Dibble	At NE beach
1966 July 30	54.3	" "	At outlet
1966 Aug 3	53	E.F. Lloyd (pers.comm.)	" "
1966 Aug 11	48.8	" "	" "
1966 Aug 19	50.5	R.B. Glover "	" "
1966 Aug 21	48.5-50.5	C.A.Y. Hewson "	At NE beach
1966 Aug 22	50.5	R.B. Glover "	
1966 Aug 27	49.2	J. Purvis "	At 10 mm depth
1966 Sept 1	49.5	E.F. Lloyd "	At outlet
1966 Sept 13	44	R.B. Glover	" "
1966 Sept 27	41	E.F. Lloyd (pers.comm.)	" "
1966 Oct 18	45-46	R.B. Glover "	" "

Date	Temperature °C	Source	Remarks
1966 Nov 11	44.3	E.F. Lloyd (pers. comm.)	At outlet
1966 Dec 15	31.7	" "	" "
1967 Jan 11	39	R. Annabell	NE beach, 6 m off shore.
1967 Jan 24	39.5-40	R.B. Glover	At outlet
1967 Feb 14	39	E.F. Lloyd	" "
1967 Mar 15	39.5	R.B. Glover	At outlet
1967 Apr 14	37.5	E.F. Lloyd	At outlet
1967 June 13	33	L. Klyen	" "
1967 Aug 9	41.5	E.F. Lloyd	" "
1967 Dec 12	26.5	" "	" "
1968 Jan 22	36	" "	" "
1968 Mar 1	38	" "	" "
1968 Apr 7	53.5	" "	At SW cliffs
1968 Apr 29	60	C.A.Y. Hewson	" "
1968 May 7	51-60	W.M. Crafar	At outlet and off shore

Date	Temperature °C	Source	Remarks
1968 May 23	43-48.5	W.M. Crafar (pers.comm.)	At outlet and off shore
1968 June 6	53-55	" "	At outlet
1968 June 21	48-51.5	C.A.Y. Hewson	NE beach
1968 July 5	32.5-34	W.M. Crafar	" "
1968 Aug 8	27.3-29.5	C.A.Y. Hewson	At outlet
1968 Aug 21	33.5-33.7	W.M. Crafar	At outlet and off shore
1968 Sept 13	20.6	C.A.Y. Hewson	At outlet
1968 Nov 1	25.5	W.M. Crafar	NE beach
1968 Dec 11	29	L. Klyen	At outlet and off shore
1969 Jan 20	33	R.R. Dibble	NE beach
1969 Feb 8	33	B. Jefferies (pers.comm.)	" "
1969 Mar 20	29.8	E.F. Lloyd	At outlet and off shore
1969 June 26	43	R.B. Glover	Off shore

Date	Temperature °C	Source	Remarks
1969 July 4	35.5	R.R. Dibble	At NE beach
1969 July 12	38	E.F. Lloyd (pers.comm.)	At outlet
1969 Aug 14	33	" "	" "
1969 Aug 25	36	B. Jefferies	" "
1969 Sept 6	40-44	" "	" "
1969 Oct 4	38	P. Otway	" "
1969 Nov 6	36.5	E.F. Lloyd	At outlet
1969 Dec 6	40	C.A.Y. Hewson	NE beach
1969 Dec 23	38.5-39	D. Rishworth	At outlet
1970 Jan 5	40.5	B. Jefferies	NW beach
1970 Jan 24, 26	38	R.B. Glover	Near centre of lake
1970 Apr 12	33-35	G.W. Grindley	Over area of lake

Fig 3.4, and is plotted against time on Fig 8.1.

Between 1956 and 1961 it is reasonably certain that no large peaks occurred in the graph. The peak in 1961, which may have reached 500 MW, was accompanied by tremor with stoppages and powerful recommencements. In 1964 the heat output increased during strong tremor and powerful recommencements, and early in 1966 a smaller increase of heat output and tremor occurred. Apparently there were no eruptions up to this stage. Most of the increase in temperature at about the times of the eruptions of July 1966 and April 1968 probably occurred during the tremor which preceded them, because the temperature rise during the large eruption of 22 June 1969, which was preceded by little tremor, was rather small. The peak apparent values of heat output in 1964, 1966 and 1968 lay between 900 and 1100 MW, while the average seismic power of tremor measured at CNZ at the same times approximated 1 kW.

The ratio
$$\frac{\text{Thermal power}}{\text{Seismic power of tremor}}$$

is about 10^6 at Ruapehu, and occupies a middle position in the range 10^{11} for Kilauea, 10^5 for Mihara, and 10^6 for Nyiragongo reported by Shimozuru (1968).

Fig 8.1: Thermal power output through Ruapehu crater lake between 1955 and 1970. Calculated from lake temperature on the assumption that all heat is lost from the lake surface.

Changes within the Crater Lake

In January and April 1970, the author co-operated with D.S.I.R. in a resurvey of the crater lake, and it was found that the depth of the lake had decreased from 300 to 65 m below overflow level, and the volume from 10×10^6 to $7 \times 10^6 \text{ m}^3$. The shape of the bottom of the crater had been profoundly altered (Fig 8.2) either by infilling with low density debris derived from the lake floor or by upward movement of the lava column. If the latter occurred it may have been during the slightly different tremor which followed the outbursts on 27 April 1968 and 22 June 1969. It could also explain why the lake level seemed unchanged by these outbursts, but the probability that just enough lava movement would occur to restore the lake level to normal is very small. It is more probable that the lake would overflow, as it did when the tholoid was being erupted in 1945 (Oliver 1945). It may have been a flow of lava into the bottom of the lake which caused the overflow, agitation, and continued heating of the lake during the day following the eruption on 24 July 1966, but if so, the flow was not accompanied by notable volcanic tremor.

Fig 8.2: Cross section of Ruapehu crater lake showing changes in bathymetry between 1965 (heavy dotted profile) and 1970 (solid bottom profile). Depth decreased from 300 to 65 m, and volume from 10×10^6 to 7×10^6 m³.

Diagram after G. Grindley (personal communication).

The subsequent high heat flow and fall in lake level might then be due to conductive heating and evaporation of the lake by this lava.

Evaluation of a Hypothetical Warning System

The premonitory indications of the seismic power recorder at Ruapehu, prior to the eruptions of the 1968 series, raised hopes that a reliable warning system could be based on the concurrence of volcanic tremor of seismic power ≥ 1 kW, and volcanic earthquakes of seismic power ≥ 20 kW, (as measured from CNZ recordings). To investigate this on a uniform basis, the seismic powers of volcano-seismic events which were, or might have been associated with eruptions, were remeasured from the CNZ seismograms. The largest typical amplitude of volcanic tremor preceding each apparent group of eruptions, and half the peak to peak amplitude of volcano seismic events was used as described in Chapter 3. Table 8.2 lists the maximum seismic power of each type of event preceding the first proven eruption in each spasm of activity, or if no visible eruption was

TABLE 8.2:

Volcano-Seismic Power, Heat Loss from Crater Lake, and
Visibly Observed Eruptions of Ruapehu during the Periods
of Increased Activity in 1964, 1966, 1968 and 1969.

Dates of Proven Eruptions	Date of Main Associated Seismic Event	Visible Activity	Thermal Power MW	Seismic Power Preceding any				Observed Eruption		
				Observed CT	Observed PR	Observed BQ	Preceding EV	Preceding IV	Seismic Power kW	Vibration Type
D. M. Y.	D. M. Y.			kW	kW	kW	kW	kW		
-	1. 5.64	Steam	?	5	230	-	-	-	-	None
-	14.5.64	Steam	?	2.5	70	-	-	-	-	None
-	20.5.64	Steam	800	1	150	-	-	-	-	None
-	3.6.64	Steam	?	1	130	-	-	-	-	None
-	Mar. 66	Steam	650	< 5?	?	?	?	?	-	None
24. 7.66	23.7.66	Steam, water erupted	1900	7	62	30	275	-	150	IV
2. 8.66	23.7.66	Steam, Pumice erupted	900	0.1	17	-	150	-	170	IV
Aug. 66	-	Steam, Geysers	800	0.01	-	-	-	-	-	-
6-7.4.68	-	Steam, Geysers	850	1	10	37	-	-	-	-
25. 4.68	20.4.68	Steam	?	3	20	-	100	20	30	EV, IV
27. 4.68	27.4.68	Steam, Mud erupted	1200?	3	110	110?	1000	-	>1000 1000 7	EV EV EV
4. 6.68	31.5.68	Mud erupted, Steam	850	2	30	-	300	-	50	EV

Dates of Proven Eruptions D. M. Y.	Date of Main Associated Seismic Event D. M. Y.	Visible Activity	Thermal Power MW	Seismic Power Preceding any				Observed Eruption	
				CT	PR	BQ	EV	Seismic Power kW	Vibration Type
9. 6.68	9. 6.68	Mud erupted, Steam	850	0.3	-	-	-	200	EV
22. 6.69	22. 6.69	Mud Ash and Bombs erupted, Steam	500	2	-	400	-	7500	EV?
-	8. 7.69	?	?	1	-	290	9000	?	?
-	16. 7.69	?	?	4	-	200	2000	?	?

observed, the most powerful vibration of eruption type. It should be realised that the first proven eruption in a spasm may have occurred after a bigger eruption which was not observed. Table 8.3 lists whether warning indications were followed by eruptions or not, whether proven eruptions were preceded by warning indications, and whether inferred eruptions, defined as recorded vibrations of EV and IV type similar to those accompanying proven eruptions, were preceded by warning indications. Table 8.4 lists the intervals between warning indications and proven eruptions or the larger inferred eruptions.

Symbols used in the tables are as follows:

- CT : Continuous tremor.
- PR : Sudden powerful recommencement of tremor.
- BQ : B-type volcanic earthquake of Minakami;
doesn't affect tremor.
- EV : Vibration resembling those accompanying
observed eruptions in 1968.
- IV : Vibration resembling that accompanying the
injection of pumice into Crater Lake on 2
August 1966.

TABLE 8.3:

Performance of Hypothetical Warning System Based on Occurrence

Together of - (i) Continuing Tremor $\geq$ 1kW (CNZ ampl. 2.3 mm)(ii) Volcano-Seismic Events $\geq$ 20kW (CNZ ampl. 10 mm)Number of Warnings, Followed
within 7 Days by:Number of Observed Eruptions (large
steam-heads, mud or ash) within 7
Days after:Number of Inferred Eruptions
(EV or IV $\geq$ 10 kW)
with No Visual Observations,
following:

Observed Eruptions	Inferred Eruptions, EV, or IV $\geq$ 10 kW	No Eruptions, EV, or IV	Warning	Sizeable Eruptions	No Warning	Warnings (within 7 days)				No Warning
						$\geq$ 10kW	$<$ 100 kW	$>$ 100 kW	$\geq$ 10 kW $<$ 100kW	
5	3	5	7	3	0	24	9	4	0	
Warning Dates D. M. Y.	Warning Dates D. M. Y.	Warning Dates D. M. Y.	Eruption Dates D. M. Y.	Eruption Dates D. M. Y.		Inferred 'Eruption' Dates D. M. Y.	Inferred 'Eruption' Dates D. M. Y.	Inferred 'Eruption' Dates D. M. Y.	Inferred 'Eruption' Dates D. M. Y.	
20. 7.66	6. 4.68	1. 5.64	24. 7.66	2. 8.66		20-26.7.66(3)	23-26.7.66(2)	18.7.66(4)		
18. 4.68	8. 7.69	14. 5.64	25. 4.68				19.4.68(1)			
26. 4.68	16. 7.69	20. 5.64	26. 4.68			9.4.68(1)	20.4.68(1)			
29. 5.68		3. 6.64	27. 4.68	27. 4.68		20.4.68(2)	27.4.68(1)			
22. 6.69		10. 5.68	27. 4.68	9. 6.68		27-28.4.68(7)	30-31.5.68(3)			
			4. 6.68			30-31.4.68(10)	8.7.69(1)			
			22. 6.69				16.7.69(1)			

TABLE 8.4: Intervals Between Earliest Possible Warning and Visible or Inferred (EV or IV > 100kW) Eruptions.

Interval:	0.3 hours	2 hours	0.3 days	0.7 days	2-4 days	6-7 days	11-12 days
Date of	22. 6.69	8. 7.69	27. 4.68	19. 4.68	23. 7.66	25. 4.68	2. 8.66
Proven or				30. 5.68	9. 4.68		9. 6.68
Inferred Eruptions.					16. 7.69		

Eruption: The ejection of tephra, lava, water, or large steam heads from the crater.

Geyser: Geyser type eruptions caused by the rise of super-heated water to shallow depth in the lake. These are not (normally) recorded on the seismographs.

Evaluation of hypothetical performance is confused by lack of visual observations of eruptions. Forty seven vibrations of eruption type (EV or IV) with seismic power ≥ 10 kW were recorded from 1964 through 1969, but only 10 eruptions were observed. Darkness and snow storms undoubtedly obscured many, and two climbers in the Dome Shelter 0.7 km from the eruption on 24 July 1966, failed to observe that eruption or its results. It can only be estimated that perhaps 35 eruptions dangerous to persons at the crater lake occurred. Hypothetical warning evaluation is then:

Number of hypothetical warnings 1964-1969	13
Percentage of probable eruptions covered by warnings of 12 days existence:	90%
Percentage of warnings not followed by eruption:	46%

It is evident that the hypothetical system is far from perfect even when no errors or failures are allowed. Also the warning indication for the most dangerous eruption (22 June 1969) was marginal and preceded the eruption by 20 minutes or less, while the most powerful of all seismic indications of eruption (8 July 1969) was not otherwise observed and could not have been accompanied by a large eruption.

A broad spectrum of research and monitoring techniques is essential, and indications of internal pressure such as ground deformation, and piezomagnetic field variations (Stacey et al 1965, Johnston and Stacey 1969), are specially valuable. However, volcanic seismology may be expected to give the earliest evidence that eruptive activity has actually begun.

RAYMOND RUSSELL DIBBLE

SEISMIC AND RELATED PHENOMENA AT ACTIVE
VOLCANOES IN NEW ZEALAND,
HAWAII, AND ITALY

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CHAPTER 9.

WHITE ISLAND VOLCANO 1966 - 1969

Introduction

White Island is an andesite volcano of diameter 2 - 2.4 km and altitude 321 m, which is situated in the Bay of Plenty, 51 km off the coast of the North Island of New Zealand. It is uninhabited, but often visited by fishing boats, and its perpetual plume of steam (Fig 9.1) is widely observed from the mainland.

The Island, and its history, volcanology, pedology, and biology have been described by Hamilton et al (1959).

The first seismic recordings on the Island were made by D.S.I.R. on 19 November 1966, six days after a large cloud of ash and steam was seen to rise to about a kilometre above the Island (Healy 1967). A new vent 60 m in diameter and 120 m deep, which was named Gilliver, had formed in the bench at the rear of the crater (Fig 9.2), and during the two day visit, was erupting steam with an occasional burst of ash. Mr. G.R.T. Clacy kindly made the tape seismograph recording available for inclusion in this study.

Fig 9.1: Aerial photo of the crater of White Island from the east in February 1967. The steam head is rising from Gilliver vent, Rudolf fumarole is emitting steam from the cliff below and to the left of Gilliver.

Fig 9.2: Map of White Island Crater, showing Gilliver and Rudolf vents, the Crater Bay seismograph station S 1, and contours of ground particle velocity (micrometre/second) measured at points S 2 - S 19, on 4 October and 1 November 1967.

The writer began seismic recordings on 28 February 1967, while Gilliver was erupting ash. Recording continued intermittently as the eruption from Gilliver ceased, and Rudolf fumarole, which had been at red heat during the Gilliver eruption, changed to an ash vent, built a small cone, ejected rocks, and decayed to a steaming condition. Clark (1970) has described the chronology of the eruption and the accompanying level changes, and Duncan and Vucetich (1970) have described the stratigraphy and petrography of the eruptions.

The seismic program had 3 objectives.

- (i) Correlation between visible activity and seismic activity during research visits.
- (ii) Monitoring volcano-seismic activity between visits.
- (iii) Location of volcanic noise sources within the crater by mapping seismic intensity.

Access difficulties dictated a single recording site (S1) for objectives (i) and (ii), on rock at the west side of Crater Bay (Fig 9.2). It was handy to the landing place; in line of sight with Whakatane; protected from eruption blast; and a quiet site except for surf noise.

Two types of seismograph were used; portable slow motion tape seismographs (Dibble 1964) and a radio telemetry seismograph described in Chapter 2. The latter was

used between White Island and Whakatane, 51 km away, from 28 February to 12 April 1967, and reception from the 1 watt transmitter was excellent at all times, in spite of corrosion due to acid gas and ash.

Seismic and Volcanic Activity

The data for each of the 130 days of seismic recording between 19 November 1966 and 20 June 1968 are summarised in Fig 9.3. Between November 1966 and April 1967 when Gilliver was erupting tephra, the seismic power of tremor was moderate (10 W ca) and sizeable local earthquake swarms occurred. In July, October and November 1967 when only fumaroles were active, tremor was very weak (0.3 W) and swarms diminished. In February and March 1968 when Rudolf was vigorously erupting tephra, tremor was moderate (10 W) and similar to that early in 1967, except that earthquakes were essentially absent, but in May and June 1968, during the end stages of the Rudolf eruption, tremor was strong and erratic (10 - 3000 W), and a few local earthquakes occurred each day. Seismic power was calculated as described in Chapter 3, but, owing to erratic behaviour of the seismographs, the values obtained give only the order of magnitude of seismic power.

Fig 9.3: Seismic data recorded at Crater Bay, White Island, for daily intervals between 19 November 1966 and 20 June 1968.

Six types of volcano-seismic activity were recognised as follows:

- (i) Volcanic tremor of dominant frequency 1 - 2 Hz;
- (ii) Stoppages and powerful recommencements of tremor;
- (iii) Pulsing tremor;
- (iv) A-type volcanic earthquakes;
- (v) B-type volcanic earthquakes;
- (vi) The volcanic earthquake accompanying the ash eruption of 20 November 1966.

These are described below.

TREMOR OF DOMINANT FREQUENCY 1 - 2 Hz

This was the strongest continuous tremor on all recordings from Crater Bay. In 1966 and 1967, its uncorrected spectrum (Fig 9.4) usually showed a single narrow band of high density, which was sometimes accompanied by 2 or 3 peaks between 2 and 4 Hz, the relative intensities of which changed several times per day. At times of low activity, the tremor appeared to consist of individual pulses (>500 per day) each of symmetrical envelope, frequency 1 - 2 Hz, and duration about 3 second. Occasionally, larger tremor pulses or groups of pulses (Fig 9.5) of this type appeared as events in

Fig 9.4: Rayspan spectrogram of tremor at Crater Bay (S 1) on 11 March, 1967. The dominant frequency was 1 - 2 Hz and the apparent frequency range was 0.5 - 8 Hz. The relative intensity of the peaks in the tremor of 2 - 4 Hz which commenced at 04 hours NZST, changed several times a day.

Fig 9.5: Seismogram of tremor of dominant frequency 1 - 2 Hz. This was sometimes observed to consist of separate symmetrical pulses, as in this recording made at Crater Bay (S 1) on 31 March 1967.

the tremor. Tremor composed of individual pulses has previously been reported by Minakami (1969) who states that frequent lava explosions or B-type earthquakes "are often observed in the form of volcanic tremors of continuous train".

The strong tremor in 1968 gave no evidence of consisting of overlapping events, and had a wide spectrum (Fig 9.6) covering and possibly incorporating the earlier intermittent 2 - 4 Hz band. Principal data are given in Table 9.1.

TABLE 9.1   Seismic Power and Spectral Peaks
of Tremor at White Island.

Date	Dominant Frequency Hz	Minor Spectral Peaks Hz	Typical Seismic Power, Watt
Nov 1966	1.3	0.6, 2.6, 3.1, 3.8, 4.2	10
Feb-Apr 1967	1.3	2.75, 3.0, 3.6, 7.0	10
Feb-Mar 1968	1.7	2.1, 2.6, 3.8, 6.0	30
Apr-Jun 1968	1.7	2.9, 3.5, 4.7, 7.6	3000

Fig 9.6: Rayspan spectrogram of a series of 20 brief stoppages and powerful recommencements of tremor on 29 April 1968. High frequencies in the recommencements have been lost by over-loading in this high gain recording from Crater Bay.

Fig 9.7: Seismogram of a powerful recommencement of tremor at 20 h 10 m 53 s followed by a high frequency vibration (8 - 22 Hz) with maximum at 11 m 10 s on 29 April 1968. Recorded at Crater Bay on the low gain channel which was noisy (equivalent to 2 micrometre/second). Filter 1.25 - 25 Hz.

The relations between seismic power of tremor, temperature of eruption (Clark 1970), and the parameters of the tephra eruption (Duncan and Vucetich 1970) are given in Tables 9.2 and 9.3. Tremor power was strongest (and most variable) when the tephra being erupted had a maximum magmatic component but comparatively low temperature. No simple consistent relations between tremor power and volume and violence of eruption are apparent.

TABLE 9.2   Relation between Smoothed Seismic Power of Tremor, and the Parameters of Eruptions from Different Vents.

Seismic Power	Vent Erupting Tephra	Fresh Magmatic Component of Tephra	Average Rate of Tephra Eruption	Recording Period
Watt		%	m ³ /day	
0.3	None	None	None	July October 1967
3-30	Gilliver	10	≥ 6000	Nov, 1966 Mar-Apr 1967
30-3000	Rudolf	30-50	≥ 3000	Feb and May 1968

TABLE 9.3   Relation between Smoothed Seismic Power of Tremor, and the Temperature and Violence of Continuous Tephra Eruptions from Rudolf Vent.

Seismic Power	Temperature in Vent	Violence of Rudolf Eruption	Date of Observation
Watt	$^{\circ}\text{C}$		
10-30	850 - 900	Noisy, Unapproachable	9 February 1968
100-3000	445	Quiet, easily approached	29 April, 18 May 1968

STOPPAGES AND POWERFUL RECOMMENCEMENTS OF TREMOR

These events were recorded between 24 February and 15 May 1968, when the tephra being erupted from Rudolf had a maximum magmatic component (Table 9.4). They were most numerous when the eruption temperature was low.

TABLE 9.4: Rate of Occurrence of Stoppages of Tremor, Compared with the Magmatic Content and Temperature of Tephra being Erupted from Rudolf.

Stoppages of Tremor per day	Magmatic Content of Tephra %	Eruption Temperature °C	Averaging Period
0	10	?	Nov 1966-Apr 1967
0.15	30 - 50	850 - 900	9 - 29 Feb 1968
4.3	30 - 50	445	28 Apr - 16 May 1968

Typically, tremor would stop suddenly (Fig 9.6) for periods between 0.5 and 90 minute (median 3 minute) then recommence with intensity 10 - 20 db above the prevailing level, and return to normal in about 30 s. The powerful re commencements were followed (Fig 9.7) by emergent vibrations of dominant frequency 16 Hz and range 8 - 22 Hz which reached maximum power about 16 s after the onset and lasted a quarter of a minute. On 29 April 1968, the author observed the activity during powerful re commencements at 0956 and 1521 NZST. Nothing was noticed while tremor was stopped, although Prof. A. Wilson was very close to Rudolf vent,

but at about the time of the powerful recommencement of tremor, the ash eruption rapidly increased and was audible as a low rumble for about a quarter of a minute.

It appears reasonable to correlate the initiation of the eruption pulse with the recommencement of tremor, and the audible rumble with the recorded 16 Hz vibration. If it is assumed that the latter was generated as edge tones at Rudolf Vent, the 16 s delay between recommencement and maximum 16 Hz vibration could be the time for the eruption pulse to travel from its primary source to the surface. The velocity of the eruption pulse up the vent is estimated as 50 m s^{-1} from the rate of rise of ash above the vent (20 m s^{-1}) observed by the author and Prof. A.T. Wilson on 29 April, and measurements averaging 68 m s^{-1} made by Prof. D.A. Christoffel (personal communication) with a pitot tube inserted 2 m into the vent on 13 May 1968. Thus the depth of origin of the eruption is very approximately estimated as 800 m.

PULSING TREMOR

On 23 May 1968 after tremor had decreased, a new pattern appeared which was best recorded on 4 June (Fig 9.8). Regular pulses of 1 - 2 Hz tremor, 0.5 - 1 minute

Fig 9.8: Rayspan spectrogram of a sequence of pulses of 1 - 2 Hz tremor leading up to a strong burst of tremor at 17.2 NZST on 4 June 1968. High frequencies in the burst were lost by overloading in this high gain recording from Crater Bay, but reappear in the coda.

in duration, occurred at about 7 minute intervals for several hours, and ended with a 10 kW burst of tremor lasting 5 minutes. There were no visual observations of these bursts (the tremor pulses on 6 June stopped without one), but their power and wide spectrum (0.5 - 24 Hz) with strong high frequencies not dissimilar to those of the powerful recommencements of tremor, suggest they accompanied the final climaxes of the 1968 ash eruption, which ended about this time (Clark 1970).

A-TYPE VOLCANIC EARTHQUAKES

These earthquakes had clear P and S arrivals (Fig 9.9) and the S minus P intervals ranged from 1 to 2 s corresponding to 5 - 10 km in slant distance. Their frequency range was 1 - 15 Hz with a peak at 10 Hz, and they usually occurred in swarms of a few hours duration at rates of up to 80 earthquakes per hour. The dates of the swarms are shown in Fig 9.3.

On 4 March 1967, 2 earthquakes were strong enough to be located with the aid of 3 other telemetry seismometers on the mainland. They were at shallow depth, 1.5 km east of the island, with an uncertainty of 6 km. No

Fig 9.9: Seismograms of A-type earthquakes on 29 February and 17 March 1967. S minus P interval 1.1 s. These occurred in swarms of a few hours duration at rates up to 80 per hour.

340 a

connection was obvious between the earthquakes and the other seismic or eruptive activity, and the swarm on 12 July 1967 was not felt by the research party visiting the Island.

B-TYPE VOLCANIC EARTHQUAKES

B-type earthquakes were not a clear or consistent type at White Island and have not been studied intensively. The larger of the individual pulses in the 1 - 2 Hz tremor are assigned to this type. They were rather formless emergent events of duration 5 - 20 second, and a spectra similar to the tremor spectra, which were common in 1966 and early 1967, but not in 1968. Several occurred while the volcano was under observation, but no change in activity was reported.

Smaller events which were shorter, more impulsive and of higher frequency than those above were also present. They had similarities with the initial parts of the powerful recommencements of tremor, and with the A-type earthquakes except that only a single phase was evident.

THE VOLCANIC EARTHQUAKE ACCOMPANYING THE ASH
ERUPTION OF 20 NOVEMBER 1966.

At 1050 NZST, Healy (1967) observed at about 300 m distance a brief and silent eruption from Gilliver vent which initially spread laterally, and deposited up to 150 mm of hot powdery white ash over an area of 25000 m². The accompanying earthquake (Fig 9.10a) had an impulsive onset, and nearly constant particle velocity during its duration of 10 s. The dominant frequency was high at the beginning and end, and low in the middle (Fig 9.10b). This may be due to dispersion but the velocity versus frequency curve would be unusual in having about 4 different velocities at frequency 2.5 Hz. Another interpretation is that the source was prolonged, and its dominant frequency varied with time.

No other events at White Island were sufficiently similar to be labelled as further eruptions of this type.

Fig 9.10:

- (a) Seismogram of an event of dominant frequency 2 - 3 Hz, recorded at Crater Bay concurrently with a brief silent eruption from Gilliver vent at 1050 NZST on 20 November 1966.
- (b) Wave frequency versus time for the same event.

Isoseismals of Tremor Intensity

On 4 October and 1 November 1967 during the quiescence between the Gilliver and Rudolf eruptions, ground velocity was recorded at 19 sites (Table 9.5) and contoured (Fig 9.2). Sites near obvious noise sources were avoided, and the two maximum values have no apparent cause, but the general area of high values included the main hydrothermal area containing Donald and Noisy Nellie fumaroles, and corresponds roughly to the middle section of the crater. The comparatively low values near Gilliver and Rudolf vents are consistent with their low activity at the time.

TABLE 9.5   Peak Ground Velocities (wide band) at
 Sites on the Crater Floor of White Island,
 on 4 October and 1 November 1967.

4 October 1967		1 November 1967	
Site No.	$\mu\text{m s}^{-1}$	Site No.	$\mu\text{m s}^{-1}$
S 1	0.4	S 12	1.8
S 2	0.7?	S 13	3.3
S 3	1.0	S 14	2.2
S 4	4.0	S 15	3.1
S 5	2.4	S 16	3.1
S 6	1.2	S 17	6.5
S 7	1.3	S 18	1.5
S 8	1.9	S 19	1.1
S 9	3.6		
S 10	1.5		
S 11	<1.5 (strong wind)		

To quantitatively investigate the source of vibration in the area of high values, theoretical curves for spherical propagation (with absorption) from : (a) a shallow disk source and (b) a point source at depth were fitted to a profile of ground velocity between site S1 and the centre of the area of high values (Fig. 9.11). Curves for (a) were obtained by computer from the expressions:

$$v^2 = 2V_o^2 \int_{\Delta-R}^{\Delta+R} x(x^2+h^2)^{-1} \arccos \frac{x^2+\Delta^2-R^2}{2x\Delta} e^{-2\alpha(x^2+h^2)^{\frac{1}{2}}} dx$$

for points outside the disk, and

$$v^2 = 2V_o^2 \int_{R-\Delta}^{R+\Delta} x(x^2+h^2)^{-1} \arccos \frac{x^2+\Delta^2-R^2}{2x\Delta} e^{-2\alpha(x^2+h^2)^{\frac{1}{2}}} dx \\ + 2\pi V_o^2 \int_0^{R-\Delta} x(x^2+h^2)^{-1} e^{-2\alpha(x^2+h^2)^{\frac{1}{2}}} dx$$

for points inside the disk, where:

V is ground velocity at distance Δ from the centre:

V_o is the ground velocity at unit distance from each assumed point source per unit area of disk;

Δ is horizontal distance of observing point from centre of disk;

Fig 9.11: Profile of measured ground particle velocity (circles) against distance towards Crater Bay from the centre of the area of high values. Theoretical curves are also shown for (a) a shallow disk source; radius 250 m, depth 10 m, absorption coefficient $\alpha = 4 \times 10^{-4} \text{ m}^{-1}$; and (b) a point source of depth 200 m with absorption coefficient $\alpha = 9.5 \times 10^{-4} \text{ m}^{-1}$. These are best fitting curves for the 2 models.

- h is depth of disk below observation plane;
 R is radius of disk source;
 α is absorption coefficient per unit ray path distance.

For the best fitting curve (a in Fig 9.11), the values were:

$$R = 250 \text{ m}, \quad h = 10 \text{ m}, \quad V_o = 1 \text{ } \mu\text{m s}^{-1}$$

$$\alpha = 4 \times 10^{-4} \text{ m}^{-1}$$

Taking the wave velocity as 2 km s^{-1} and the dominant frequency as 4 Hz , then $\alpha\lambda = 0.2$, and $Q = 16$, which seems a reasonable value for water saturated tephra (O'Brien 1967).

Curve (b) for a point source (to which all sufficiently deep sources approximate) is of the form;

$$V = V_o (\Delta^2 + h^2)^{-\frac{1}{2}} e^{-\alpha(\Delta^2 + h^2)^{\frac{1}{2}}}$$

Values used were:

$$V_o = 7500 \text{ } \mu\text{m s}^{-1}$$

$$h = 200 \text{ m}$$

$$\alpha = 9.5 \times 10^{-4} \text{ m}^{-1}$$

$\alpha\lambda = 0.46$, and $Q = 7$ (the corresponding amplitude absorption factor is $B = 4 \text{ db per wavelength}$).

In spite of the low accuracy of the contour map, the shallow disk model is clearly to be preferred, and although the depth is not reliably determined, it is certainly less than 100 m. Fig 9.12 shows that the seismic spectrum versus time function in the area of high values, differs from that on Crater Bench near Rudolf, in that it is more continuous and of higher frequency, and shows less of the 1 - 2 Hz pulses believed to be associated with the activity of Gilliver and Rudolf. Thus it appears that the two areas of activity are separate.

Fig 9.12: Rayspan spectrograms of seismic noise between sites S 7 (Gilliver) and S 1 (Crater Bay) recorded on 4 October and 1 November, 1967. Sites S 5 to S 13 lie in the high noise area which contains Donald and Noisy Nellie fumaroles. Frequencies here were between 2 and 8 Hz. Sites S 19, S 6 and S 18 are near Rudolf vent, and show pulses of 1 - 2 Hz tremor.

CHAPTER 10.

KILAUEA VOLCANO 1971

Mauna Ulu Eruption 1969-1971

The 1969-71 eruption of Mauna Ulu has been described by Swanson et al (1971) as the longest and most voluminous flank eruption in Kilauea's recorded history. It began on 24 May 1969 and ceased in mid-October 1971. The author recorded eruption-related seismic activity, and intermittently observed eruptive activity between 21 January and 13 April 1971.

HISTORY OF MAUNA ULU 1969-1970

Swanson et al (1971) state that prior to 24 May 1969, tiltmeter and geodimeter data showed that the summit at Kilauea (Fig 10.1) was highly distended, in spite of 4 eruptions totalling $24 \times 10^6 \text{ m}^3$ of lava during the previous 18 months. On 24 May 1969 lava fountains broke out along several en echelon segments of a new fissure 4 km long and striking 65° E of N on the upper east rift zone. After a few hours, activity became localised along

Fig 10.1: Topographic Map of Kilauea Volcano showing seismograph stations.

(o) Stations occupied by the author.

(O) Stations of the Hawaiian Volcano Observatory.

The author's principal stations were Puolani and Puu Huluhulu (PHH) which was also an H.V.O. station.

the segment equidistant from Aloi and Alae Craters (Fig 10.2), and centred there for the rest of the eruption, the magma being supplied from the summit region via more or less horizontal passages along the rift zone (Moore and Krivoy 1962). During the first 6 days, $8 \times 10^6 \text{ m}^3$ of lava were erupted. A second spasm of fountaining began on 12 June, producing $4 \times 10^6 \text{ m}^3$ of lava in 21.5 hours, and forming a flow 11 km long, reaching within 1 km of the sea.

During the rest of 1969, activity was cyclic, with periods of several hours of voluminous high fountaining separated by days or weeks of subdued activity. The fountains reached 540 metres maximum height, and the average lava output during high fountaining was 10^6 m^3 per hour. Each episode ended abruptly (in less than 4 minutes), allowing lava pooled around the vent to pour back down while residual gas billowed forth. The lava erupted last during each major fountaining phase was relatively rich in olivine (Swanson et al 1971). The phenomena apparently involved the formation of a column of gas - enriched pyromagma, through which olivine crystals could settle. The column ejected itself by vesiculation, forming pumice, spatter fragments, and

Fig 10.2: Topographic map of Maunu Ulu and the upper east rift zone. The author's seismic noise survey sites are shown: O. Station 44 is at the edge of the Mauna Ulu vent and lava lake. Site 20 was also the author's Puu Huluhulu seismograph station, and was 100 m east of station PHH of the Hawaiian Volcano Observatory.

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occasionally reticulite (Fig 10.3), a basaltic pumice with a bulk density of about 0.1 Mg m^{-3} , and leaving the vent nearly empty and exhausted. Most of the time however, molten lava was visible a few tens of metres below the rim of the 100 m long eruptive fissure, and was rising and falling in level, normally each 10 - 15 minutes, in a cyclic degassing process (Swanson et al 1971).

The fissure was divided into two main compartments by a septum of solid lava which extended downward 50 - 100 m below the vents, (D.A. Swanson personal communication). Lava almost always started to rise first in the larger compartment, followed within a few minutes by the smaller. The rate of rise in the smaller compartment was the greater so that when they stopped rising, (usually simultaneously), the lava stood several metres higher in the small compartment. Suddenly, and within a few seconds of each other (never more than 20 s), the thin black crusts over the compartments would break, and with vigorous bubbling and spatter action, accumulated gas would escape and the lava columns would drop several tens of metres in 2 to 4 minutes. The volume change was 5000 to 25000 m^3 (Swanson et al 1971).

Fig 10.3: Reticulite, also known as limu, or thread-lace scoria (Perret 1950). A basaltic pumice sometimes formed during high fountaining at Mauna Ulu.

Scale: approximately full size.

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By the beginning of 1970 the vent had enlarged to 130 x 35 m and high fountaining ceased, leaving only the rise and fall degassing cycle. Periodically lava spilled over, and gradually buried the spatter cone formed during fountaining, with a very gassy, cavernous, fragile lava so as to form the dome of Mauna Ulu (growing mountain) which by July 1970 was 1 km wide and 100 m high (Fig 10.4). At that time a new radial fissure opened on the eastern slope of the Dome, and lava began to flow away in tubes towards the sea at an average rate of about $3 \times 10^5 \text{ m}^3$ per day (Swanson, 1972). The rate showed no apparent correlation with the rise and fall cycle of the lava lake, (D.A. Swanson personal communication) and tiltmeter and geodimeter readings showed that the state of tumescence of Kilauea was not changing. Swanson (1972) suggests that a steady state discharge of magma from the mantle of the earth was occurring. This was still the state of affairs when the author began his observations on 21 February 1971, using the two 1970 model slow motion tape seismographs described in Chapter 2.

Fig 10.4: Oblique aerial view of Mauna Ulu from the west, on about 19 February 1971.

Right centre: main vent and lava lake. A major rock-fall from the right quadrant of the vent wall occurred on 18 February 1971.

Top left: the trench marking the line of the initial eruptive fissure.

Lower right: a partially collapsed lava tube. Vehicle tracks can be seen leading to cable way across the near end of the lake.

Photo by Dr. Wendell A. Duffield
U.S. Geological Survey

MAIN FEATURES OF SEISMIC RECORDINGS AND VOLCANIC
ACTIVITY, 22 JANUARY - 13 APRIL 1971.

On 22 January seismograph B was installed at Puu Huluhulu (PHH, Fig 10.1) 0.7 km west of Mauna Ulu vent, and seismograph A was installed at Puolani Rd, 6.7 km north of Mauna Ulu, a site almost equidistant from all the craters in the chain between Halemaumau and Napau. The seismographs operated almost continuously until 13 April 1971, and charts of seismic power versus time (Figs 10.5, 10.6, 10.7, 10.8), power spectra versus time, and seismograms were prepared for study. Tremor from Mauna Ulu predominated at both the Puolani and Puu Huluhulu sites, and as only 5.5 hours of suspected intermediate to deep tremor occurred (R. Koyanagi, personal communication; recognised by similar amplitudes over most of the Hawaiian Volcano Observatory seismic network), it appears that Mauna Ulu was the most powerful source of tremor within Kilauea during the recording period.

Until 28 January, continuous weak ebullition in the Mauna Ulu lava lake was occurring, and lava was flowing towards the south-east in tubes at a moderate rate

Fig 10.5: Nominal seismic power from Mauna Ulu during January 1971. Recordings from Puu Huluhulu (upper) and Puolani (lower) in frequency ranges 1 - 4 Hz (channel 1) 0.5 - 1 Hz and 4 - 20 Hz (ch 2 and 3, see Fig) are calibrated as if Mauna Ulu was the vibration source. Channel 4 shows time marks. Vibrations from sources other than Mauna Ulu are evident at Puolani. Note the decrease in power late on 28 January, when eruption from fissures SW of Mauna Ulu began.

Fig 10.6: Nominal seismic power from Mauna Ulu during February 1971. Details as for Fig 10.5. Rockfall earthquakes are marked at top: F - probable or observed; F? - possible; EF - earthquakes possibly followed by rockfalls; Number of associated events follows. Days on which possibly deep tremor occurred are marked D below Ch 2, Puolani. Note decrease of power on 10 February when eruption switched from SW to east, and stoppages and pulses after 15 February, corresponding to cyclic degassing of the lava lake.

Fig 10.7: Nominal seismic power from Mauna Ulu during March 1971. Details as for Figs 10.5 and 10.6. Note reappearance of longer periods of tremor on 7 March, and the change in the pattern of power variation on 28 March after a major observed rockfall.

Fig 10.8: Nominal seismic power from Mauna Ulu during April 1971. Details as for Figs 10.5 and 10.6. A stable pattern of longer periods of tremor separated by periods of stoppages and pulses of tremor has developed.

(Swanson et al 1971). The seismic power of tremor (Fig 10.5) was low (10 to 100 W) and fairly steady, so that the spectrum of variations in tremor level versus time (Fig 10.9) was of low density, and did not agree well at the two seismograph stations. This type of spectrogram was obtained by rectifying and smoothing the initial high-speed playback signal, re-recording the D-C level variations on a slow motion tape recorder, and playing the new tape at high-speed into a Sona-graph analyser. The range of variation frequencies displayed was then 0.1 to 5 cycles per hour.

The ordinary power spectrum of tremor was moderately flat between 0.5 and 5 Hz (Fig 10.10), and there was sometimes a peak at 1.3 or 4 Hz. During an electrical storm and typhoon on 29 January 1971, the tape seismographs recorded a swarm of very local earthquakes accompanying the opening of a new eruptive fissure on the west flank of the Mauna Ulu dome. Low fountains (1-5 m) and voluminous lava flows containing little gas emerged from this fissure until 10 February 1971 (Swanson et al 1971). The chart of seismic power from Puu Huluhulu (Fig 10.5), recorded 0.5 km from the new fissure, shows that the swarm began at about 15.2 HT on 28 January, and reached peaks of intensity

Fig 10.9: Sonagrams of the variations in seismic power of tremor from Mauna Ulu. The original seismic recordings were rectified, smoothed, and speeded up 1.8 million times for analysis. Results for Puolani (upper) and Puu Huluhulu (lower) agree closely after 15 February. The spectra of stoppages and short pulses of tremor which correspond with cyclic degassing of the lava lake is broad band with maximum at 3 - 4 cycles per hour. Longer periods of tremor sometimes have narrow rising bands between 0.1 and 0.7 cycles per hour as on 4 - 6 April 1971.

Puolani

Puu Huluhulu

Fig 10.10: Sonagram of tremor at PHH at about 02 hours HT on 25 January 1971. Contours are at 6 db intervals. Seismic power spectral density of the darkest contour is 400 W Hz^{-1} ca.

at 17.0 and 17.7 HT, the latter being the more intense. At 19.3 HT a burst of volcanic tremor occurred which briefly reached a power of 3 kW, and is interpreted as the break-out of the fissure and release of gas at the surface.

The Ishimoto-Iida (1939) coefficient m was 2.5 ± 0.1 (Fig 10.11) for the 249 shocks which were recorded clearly. This is at the bottom end of the range 2.5 - 4 ascribed to volcanic earthquakes. The seismograms (Fig 10.12) show A-type earthquakes, and earthquakes with no apparent S phase. They had a maximum duration of 12 s and the logarithmic decrement of their decay trains was about 0.1, corresponding to an apparent Q of 30. The S minus P interval of A-type earthquakes was approximately 0.5 s, the outside limits being 0.2 to 2 s. No clear trend of change in S minus P interval can be discerned. The swarm did not cease at the time of apparent break out as is usually the case (Eaton and Murata 1960), but died out at about 21.0 HT, about an hour after park rangers discovered the outbreak.

Surprisingly, the intensity of tremor at Puu Huluhulu was lower during the first 5 hours of the new outbreak than before, in spite of its proximity. Although this may be partly due to a decrease in wind microseisms

Fig 10.11: Number of earthquakes versus relative magnitude in the swarm associated with the beginning of the fissure eruption on the west flank of Mauna Ulu on 28 January 1971.

Fig 10.12: Seismograms of the earthquake swarm on 28 January 1971. The onset of tremor at 1918 is assumed to mark the breakout of the eruption fissure, and release of gas at the surface. Amplitude scale: $1 \text{ mm} \equiv 2.5 \text{ } \mu\text{m s}^{-1}$.

after the storm, it is quite obvious that flow of lava to the new fissure, and fountaining of gas-poor lava was not a powerful source of tremor. An increase of tremor with a dominant frequency of 4 Hz (Fig 10.13) began at about 00 Hours HT on 29 January, but no reported visual observations correlate with this. It reached a peak power about 24 hours later. At about 07.5 HT on 30 January, mild variations of the power of tremor began at a rate (frequency) of 0.2 - 0.7 per hour. Concurrently, there was sporadic fountaining at the upper end of the new fissure (D.W. Peterson, personal communication). From 3 February, tremor became strong and rather uniform (Fig 10.6) and the lava lake was reported to be continuously active and within a few metres of the rim. By this time, flow through the lava tubes to the east had dwindled and stopped.

At 1510 HT on 10 February 1971 there was a momentary burst of volcanic tremor reaching 10 kW, followed by 4 B-type volcanic earthquakes peaking to 16 kW. Flows from the western fissure stopped abruptly and lava recommenced flowing through the tubes towards the east at a rate of 10^6 m^3 per day, quickly filling Alae crater and

Fig 10.13: Sonagram of tremor at PHH on 29 January 1971. The dominant frequency is 4 Hz. Contours are at 6 db intervals. Seismic power spectral density of the darkest contour (4 Hz at 4.5 min) is 1.6 kW Hz^{-1} ca.

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spreading widely from there (Swanson et al 1971). Again tremor decreased with the increase of lava flow, and concurrently, the level of the lava lake fell to its more usual 15 - 30 m below the rim.

At 08.0 HT on 11 February sudden variations of tremor power began to develop, and were accompanied by increasingly vigorous cyclic ebullition of the lava lake (D.W. Peterson personal communication). Until 07 hours HT on 15 February, the tremor variations were increases above the normal level, but after that time, stoppages or decreases to below 50 W occurred, as well as increases in an extremely variable but recognisable pattern. Continuous tremor of seismic power between 0.1 and 1 kW would occur for a few hours, and then stop suddenly and a series of 1 to 40 pulses of strong tremor, each of several minutes duration and with a powerful onset, would occur at an average rate of 2 to 3 per hour, the last onset inaugurating another few hours of continuous tremor.

Over a number of days the pulses became stronger (10kW) and the series longer at the expense of the periods of continuous tremor, culminating in a series of 40 pulses in 14 hours on 5 - 6 March (Fig 10.7). Then the number

and size of pulses in a series would suddenly decrease again. This occurred on 13 and 19 February, and 7 March 1971. Between 7 and 19 March the pattern was confused, but after that it was rather regular, and the number of pulses in a series rarely exceeded 10. Cyclic ebullition with rise and fall of the lava lake was observed during most visits to the vent during the period of tremor stoppages and pulses, and lava continued to flow in tubes to the south east at a rate of $3 \times 10^5 \text{ m}^3$ per day (Swanson 1972).

The spectrum (versus time) of variations in tremor level (Fig 10.9) revealed some unexpected features. Firstly the variation spectra from the two seismographs matched in great detail after the variations increased on 15 February. Secondly, there was almost no sign of a dominant peak in the spectrum corresponding to the repetition rate of the tremor pulses, but instead a very roughly flat band of variable width. Its upper boundary varied from 0.5 to 8 cph, and its lower boundary varied from 0.4 to 0.8 cph. The only signs of frequency trends within the band were between 28 February and 4 March when a zone of high densities descended from 4 to 3 cph, and 20 - 22 March and 4 - 5 April when vague upward trends appeared. Apparently the time of occurrence of pulses was almost random.

Thirdly, spectral densities between 0.1 and 0.5 cph increased after 9 March. Peaks which corresponded to the repetition rate of the longer periods of continuous tremor appeared to shift upwards from about 0.15 to 0.4 cph over a day or two, and then step down again. On 5 April, clear peaks rose to 0.8 cph at a rate of 0.3 cph per day. It appears that the occurrence of periods of continuous tremor was not random.

OBSERVATIONS OF THE LAVA LAKE AND CONCURRENT
SEISMIC ACTIVITY, 21 JANUARY - 13 APRIL 1971.

Opportunities for visual observations of activity at Mauna Ulu and subsequent correlation with the seismic record (Table 10.1) were severely limited by the author's teaching duties at University of Hawaii, and the rulings of the Chief Ranger of Volcano National Park. The author's observations are in basic agreement with those made from time to time by Hawaiian Volcano Observatory personnel (D.W. Peterson, D.A. Swanson, and R. Koyanagi, personal communication), and show that most of the time, degassing

TABLE 10.1 Visual Observations of Mauna Ulu Lava Lake, and Accompanying Seismic Activity

VISUAL OBSERVATIONS SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time h m	HT	Location	Description	Dimensions	Time h m s	Description	Power kW	Dominant Frequency Hz
21 Jan	Midday		NW & E Walls	Constant mild ebullition.	2 m high		Seismographs not yet installed.	-	-
18 Feb	11 37.3 ±0.5		NW Wall	Vigorous ebullition begins while moderate ebullition at E end continued.	NW unknown E, 5 m high	11 37.4	Tremor level, measured on north rim of crater with seismic power meter, suddenly increases by 8 db.	1	?
18 Feb	19 00		NW Wall	Moderate ebullition proceeding. Convection E to W.	3.5 m high	-	Seismograph insensitive.	erroneous	
18 Feb	19 07		Mid N Wall	Small 'bubbler' begins at wall.	?	-	"	"	
18 Feb	19 10		Mid N Wall	Moderate ebullition.	4½ m high	-	"	"	
18 Feb	19 20			Ebullition and convection stop, lake begins to rise.	zero	-	"	"	
18 Feb	19 34½		NW Side	Ebullition begins with a line of bubble, moves to the NW Wall, and becomes very vigorous. Convection E to W begins.	7 m high	19 34 13	Small gradual increase of tremor.		
						19 34 59	Tremor increases rapidly.	0.8	1.3
18 Feb	19 38		W end Wall	Second vigorous centre of ebullition forms.	4½ m high	19 37 43	Tremor increases slightly.	0.8	1.1
18 Feb	19 40			Ebullition decreases. Lake level has fallen 3 m. Moderate ebullition continues at W end.		19 39 25	Tremor decreases.	0.1	1.1
18 Feb	19 45		Mid N Wall	About 1 m ³ of wall collapses into lake.	1 m ³ ca	19 44 55	Tremor begins.	?	0.9-5

VISUAL OBSERVATIONS

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time h m	HT	Location	Description	Dimensions	Time h m s	Description	Power kW	Dominant Frequency Hz
18 Feb	19 47		Mid N Wall	About 2 m ³ of wall collapses into the lake and provokes vigorous ebullition.	2 m ³ ca	19 46 27	Event recorded in strong tremor.	2	0.9-5
18 Feb	19 57		E Wall	Vigorous ebullition begins after brief pause.	7 m high	19 56 44	Tremor increases after 2 min of low power.	0.9	0.7-5
18 Feb	20 03		N Wall	About 5 m ³ of wall collapses into lake followed by a brief upwelling 10 m high.	5 m ³ 10 m high	20 02 56 20 03 23	Weak event, followed after 0.5 minute by increased tremor.	0.8	1.4
18 Feb	20 10		E Wall	Moderate ebullition begins.	-	20 11.9	Tremor increases slightly.	0.25	0.7-5
18 Feb	20 18		E Wall	" "	-		Uniform tremor.	0.25	0.7-4
18 Feb	20 32		SW Wall	At least 12,500 m ³ of wall collapses into the lake, causing a cloud of lava spray, followed by feeble ebullition over the whole area of the lake.	Slab 100 m long x 5 m wide x >25 m high	20 32 40.5 44.5 49.4 eP20 i Max	Volcanic earthquake with forerunner (beginning of collapse?) about 4½ s before main shock; Tremor decreased 6 db immediately following the event.	500	1.4
18 Feb	20 35		-	Lake calm and quiet.		e?20 35 41 e 41.8 i 43.3	Type B earthquake of short duration, and rather flat spectra.	30	0.6-17
18 Feb	20 40		W Wall	About 1 m ³ of wall collapses into lake. Party leaves, unnerved.		20 40 20	Vague event followed 2 minute later by a stronger sharper one. Tremor very weak.	0.1	
18 Feb	21 28		?	Great increase in light and noise from the vent observed from a distance of 0.5 km. Duration about 3 minute.		21 27 21	Powerful recommencement of tremor. For the first 6 s, the power was restricted to the frequencies around 1.3 Hz, but after that the spectrum extended to 17 Hz. In profile the seismogram is cigar shaped.	10	1.3

VISUAL OBSERVATIONS

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time HT	Location	Description	Dimensions	Time HT	Description	Power kW	Dominant Frequency Hz
	h m				h m s			
19 Mar	11 30-53	E Wall	Moderate ebullition, view from W end, W wall not visible. Lake 27 m below rim.			Uniform tremor with strong spectral peak at 1.4 Hz.	0.2	1.4
19 Mar	11 53	E Wall	Moderate ebullition causing swells of sound at 38 per min.			Uniform tremor.	0.2	
19 Mar	12 00-07	E & W Walls	Moderate ebullition continuing, convection W to E.			Uniform tremor.	0.2	
19 Mar	12 07.6	NW Wall	Moderate ebullition begins at NW wall.			"	0.2	
19 Mar	12 08.3	NW & E Walls	Audible bang, nothing seen. Moderate ebullition continues NW & E wall.		12 08 24	Small local event of power 0.3 kW, and ampl. about twice the uniform tremor background.	0.3	
19 Mar	12 09.1	E Wall	Ebullition ceases at NW Wall; but continues at E end. Lake level had fallen about 1.5 m.			Uniform tremor.	0.2	
19 Mar	12 10-34	E Wall	Moderate ebullition continues at E wall.			Uniform tremor.	0.2	
19 Mar	12 34-38	E Wall	Vigorous ebullition at E Wall.			"	0.2	
19 Mar	12 44.3	E Wall	Sudden noise; nothing seen. Vigorous ebullition E Wall.			"	0.2	
19 Mar	12 54½	E Wall	Vigorous ebullition, smoke.			"	0.2	

VISUAL OBSERVATIONS

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time h m	HT	Location	Description	Dimensions	Time h m s	Description	Power kW	Dominant Frequency Hz
19 Mar	13 16		E Wall	Vigorous ebullition.			Uniform tremor.	0.2	
19 Mar	13 23		E Wall	Vigorous ebullition, with clearly visible bursting of gas bubbles.			"	0.2	
19 Mar	13 27		E Wall	Vigorous ebullition. Rates of bubble bursting: 75; 72; 67; 70 per minute.			"	0.2	
19 Mar	13 50.3		E End	Minor rock fall into lake.			"	0.2	
19 Mar	13 56½		E End	About 1 m ³ of wall fall into lake near E Wall ebullition point. Photographed resulting upwelling.	Ebullition 5 m high Upwelling 6 m high	13 56 26	Event of 0.6 kW. Tremor begins, then decreases.	0.6	1.3
19 Mar	13 57-59½		E End	Ebullition decreases and stops. Lake level ½ m lower than aver- age.		13 58.5	Tremor reaches a minimum.	0.06	1.0 2.7 4.2.
19 Mar	14 05			Lake risen about 3 m Skin sinking along a line SE of lake centre causes short- lived line of bubb- lers. Traces of out- ward radial convection.			Brief tremor?	0.15?	
19 Mar	14 11.25			Pockets of gas form- ing under skin. Some- times skin folds back exposing quiescent lava when bubble bursts.			Quiet	0.06	Flat 2-4.3

VISUAL OBSERVATIONS

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time h m	HT	Location	Description	Dimensions	Time h m s	Description	Power kW	Dominant Frequency Hz
19 Mar	14 12		E End & NE Wall	Line of bubblers forms in middle of E end. Ebullition rapidly becomes violent and noisy and for a few seconds a vortex appears to be present. Activity moves to NE Wall as convection begins. Two photos by 1413.	7 m high	14 11 22	Powerful recommencement of tremor, of 2½ minute duration followed by 1 minute of weak tremor and then uniform tremor of 0.2 kW.	1	1.4 3, 4.4
19 Mar	14 14.7		SE Wall	Moderate ebullition begins at SE Wall while vigorous ebullition at NE Wall continues.			Uniformly weak tremor.	0.3?	1.4
19 Mar	14 15			Lake level fallen to initial level.			Weak tremor.	0.1	
19 Mar	14 16		E Wall	Activity now normal with ebullition at E Wall.		14 15.8	Tremor increases during next 0.5 minute.	0.2	
19 Mar	15 53½		E Wall	Vigorous ebullition, Vibration felt?		15 40 51	Very local earthquake initiates gradual increase of tremor with strong spectral line at 1.3 Hz.	9 0.3	4.7 - 8.3? 1.3
19 Mar	16 02		E Wall	Vigorous ebullition, with very large bubbles bursting at 45 per minute.			Tremor increasing.	1	1.3
19 Mar	16 05-6		E Wall	Bubbles bursting at 53 per minute.			" "		1.3

VISUAL OBSERVATIONS

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

Date 1971	Time HT	Location	Description	Dimensions	Time HT	Description	Power kW	Dominant Frequency Hz
19 Mar	16 08		Earthquakes felt? Vigorous ebullition E Wall.			Tremor increasing.		1.3
19 Mar	16 9.8	E end	About 2 m ³ of wall falls into lake, and causes vigorous ebullition of whole E end.		16 09 32	Fall not recorded, but tremor reaches maximum value. Spectral peak dev- elops at 8½ Hz.	5	1.4 3.1 8.5
19 Mar	16 14	E End	Ebullition decreases, lake level has fallen 1½ m.		16 12 59	Tremor drops quickly to 0.3 kW. Continues slow decrease.	0.3	1.4
19 Mar	16 45	E End	Moderate ebullition, level normal.			Uniform tremor.	0.2	
13 Apr	16 02½	N Wall	Line of bubblels form when skin starts sinking NE of lake centre; moderate ebullition develops and moves to N Wall, only 50 m from seis- mograph. Fine spray formed.		16 03 16	Tremor increases suddenly after a brief minimum of 0.004 W. At the edge of Ulu Vent increase occurs over entire band- width, but at Puu Huluhulu, tremor below 1 Hz does not increase.	0.3	
13 Apr	16 23	N Wall	Moderate ebullition.			Tremor increasing, except below 1 Hz at Puu Huluhulu.	0.1	1 - 2
13 Apr	16 33	E End	Moderate ebullition in progress, 150 m from seismograph.		16 25	Tremor below 1 Hz jumps 7 db at Puu Huluhulu, during gradual increase at edge of Ulu Vent. A double spectral peak is developing.	0.3	1.2 3.5

SEISMIC RECORDINGS, PHH TAPE SEISMOGRAPH

VISUAL OBSERVATIONS

Date 1971	Time h m	HT	Location	Description	Dimensions	Time h m s	HT	Description	Power kW	Dominant Frequency Hz
13 Apr	16 34		E End	Very vigorous ebullition.				Tremor above 1 Hz increasing.	0.3	1.2, 3.5
13 Apr	16 39		E End	Very vigorous ebullition with large bubbles bursting into large fragments.				Tremor above 1 Hz increasing, but with a sharp cut off at 12 Hz.	0.3	1.2 3.5
13 Apr	16 47		E End	Extremely vigorous ebullition, strong convection W to E. Lava surging in waves against the E end wall, but not reflected, suggesting an outlet lava tube there.				Tremor 1 - 12 Hz increasing.	0.5	1.2, 3.5

of the lava lake in the main vent (Fig 10.14) was by ebullition at the east end wall or north west wall and that the whole lake was slowly convecting east or west as a single cell, with surface movement towards the dominant ebullition site. Tremor always accompanied the ebullition but was not always in apparent proportion to it.

Occasionally ebullition and large scale convection died away within 10 to 50 s, and the lake level began to rise, stretching and splitting the skin to expose incandescent lava which quickly formed a new skin. In places the old skin, being denser than the gas-charged lava beneath it, began to sink, dragging a raft of skin towards the line of sinking (Fig 10.15). Tremor was always low (<50 W) when the lake was quiet, and at 1358 HT on 19 March it subsided within 1 minute of the cessation of visible degassing. That is to say, within the accuracy of time comparison.

Up to an hour later, bubbling would begin along a line, either where skin was sinking (1603 HT, 13 April) or somewhere around the walls (1907 HT, 18 February), and would rapidly grow into an initially vigorous fountain

Fig 10.14: Typical ebullition and degassing at the east end of Mauna Ulu lava lake, viewed from the west end on 18 February 1971. A can of wet sand is being prepared to drop in the lake.

Fig 10.15: Lava skin sinking in the lake on the evening of 18 February 1971. The skin is folding downwards and exposing incandescent lava around it as it sinks.

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of ebullition (Fig 10.16 a,b). The lake level would then begin to fall, and convection would begin towards the fountain, which moved to the nearest wall if not already there (1412 HT, 19 March), and also tended to move along the wall towards the preferred ebullition sites. After several minutes, ebullition stopped again (Fig 10.17a) or decreased to a more normal rate (Fig 10.17b) in which case an hour or more would elapse before the next stoppage. Tremor appeared to increase proportionally to the initially strong fountaining, and on all 4 observed occasions (1137 and 1934 HT on 18 February; 1412 HT on 19 March; 1602 HT on 13 April) the onset times of tremor and ebullition agreed to within a minute.

Rockfalls from the vent walls into the lake often affected the activity of ebullition and tremor. Of 7 observed rockfalls, activity increased after 4 (18 February 1945, 1947, and 2003 HT, 19 March 1610 HT), and decreased after 2 (18 February 2032, 19 March 1356). The rockfall of about 5 m³ at 2003 on 18 February which was the second largest observed, was immediately followed by an upwelling about 5 m high (Fig 10.18), and accompanied by a weak seismic event of dominant frequency 1.4 Hz. At 2032 on

Fig 10.16: Lava fountains of ebullition in Mauna Ulu lava lake during their initially vigorous stage, after the lake level had risen due to gas accumulating within it.

- a/ Fountain at the NW wall at about 1936 HT on 18 February 1971. The surface layer near the fountain has begun to move inwards and downwards, rifting the skin further away. The fountain originated at the wall at 1934.5 HT.
- b/ Fountain at the east end at about 1413 HT on 19 March 1971. The fountain is moving towards the wall from its origin clear of the wall at 1412 ca.

a

b

Fig 10.17:

- a/ The deflated lava lake at 1940 HT on 18 February 1971, following very vigorous degassing. The high lava mark about 3 m above the new level is shown by the incandescent walls. View to the NE from the middle of the south wall.
- b/ The lava lake at low level, at 1356.5 HT on 19 March 1971. The level is several metres below that shown in Fig 16b. Normal steady ebullition is occurring on the right. A 1 m^3 rockfall has just occurred on the left. Ebullition ceased altogether at about 1359.

386 a

Fig 10.18: Upwelling in Mauna Ulu lava lake following a 5 m³ rockfall from the middle of the north wall at 2003 HT on 18 February 1971. This rockfall did not induce a second centre of ebullition to accompany that at the east end.

387 a

18 February, the south-west quarter of the crater rim, 100 m long, 5 m wide and at least 25 m high, on which the observing party led by Dr. D.W. Peterson had been standing, peeled off and disappeared into the lake. Its absence is noticeable on Fig 10.4. A member of the party felt the block begin to move, noticed glowing fissures opening, and gave warning, so that when the block fell we were 15 m away and accelerating, and so observed little more than the time.

The seismic power of this event exceeded the scale range of Fig 10.6, but from the low gain channel of the Puu Huluhulu tape seismograph, a nominal peak seismic power of 4 MW is obtained. The seismic power spectrum as recorded at Puu Huluhulu (Fig 10.19), has peaks at the same frequencies as the tremor, the seismic power of which decreased by 6 db during the rockfall and a further 7 db soon after. A second smaller earthquake (Fig 10.19) followed the rockfall by 3 minutes - a time when the lake was again under observation. Small bubbles were quietly bursting all over, but no second rockfall was observed.

Fig 10.19: Seismic recordings at PHH of the major observed rockfall at 2032 on 18 February 1971.

- a/ Sonagram. 0 - 2 minutes: tremor with a spectral line at 1.3 Hz. 2 - 3 minutes: rockfall earthquake at 2032, with spectrum similar to preceding tremor. 3 - 5 minutes: tremor with reduced spectral density except at 1.2 Hz. 5 - 6 minutes: B-type earthquake. 6 - 10 minutes: weak tremor.
- b/ Profiles of seismic power spectral density with frequency resolution 0.2 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 4 W Hz⁻¹.
- c/ Upper: High gain recording of rockfall.
Scale 1 mm $\equiv$ 10 μ m s⁻¹.
Centre: Low gain recording of rockfall.
Scale 1 mm $\equiv$ 150 μ m s⁻¹.
Lower: High gain recording of the B-type earthquake. Scale 1 mm $\equiv$ 10 μ m s⁻¹.

Degassing resumed an hour later with an especially strong recommencement of tremor (Fig 10.20) recognised by R. Koyanagi (personal communication), from the cigar shaped envelope of the seismogram, as similar to vibration accompanying drainback of lava into a vent after an episode of fire-fountaining. The onset had a narrow spectral peak at 1.3 Hz, which broadened greatly during the first 12 s. From a distance this event was accompanied by a great output of light and audible noise.

ROCKFALL EARTHQUAKES AT THE LAVA LAKE

Rockfalls and the vibrations they cause have been observed at Kilauea from the time scientific studies began (Perret 1950, Jaggar 1947) and have been reported in the Volcano Letter and the Hawaiian Volcano Observatory Summary. As an object for seismometric study they have the advantage of known location and the disadvantage of emergent low frequency arrivals. Dr. D.W. Peterson kindly supplied the author with copies of the Hawaiian Volcano Observatory seismograms (Fig 10.21) for the rockfalls on 18 February 1971, together with the following

Fig 10.20: Sonagram of powerful recommencement of tremor recorded at PHH at 2128 HT ($\approx$ 12 minute on Fig) on 18 February 1971, when degassing resumed after the major rockfall. Vibration recommenced first at 1.3 Hz ca. Frequency resolution 0.6 Hz; stability 3.2 db with 90% confidence; contour interval 6 db; seismic power spectral density of darkest contour 3 kW Hz^{-1} .

391a

Fig 10.21: Seismograms of the Hawaiian Volcano Observatory network for the rockfall earthquake at 2032 HT on 18 February 1971 (upper) and the B-type earthquake at 2035 HT (lower). The Station Code names, and distances from Mauna Ulu vent correspond with trace numbers as follows:

No.	Code	Dist., km	No.	Code	Dist., km
1	MLO	24.0	10	PHH	0.8
2	MLX	17.6	11	KPN	9.8
3	AHU	6.4	12	MPR	4.2
4	HPU	60 ca	13	CPK	13.3
5	OTL	8.5	14	HLP	13.7
6	DES	19.5ca	15	UWE	11.0
7	NPT	9.6	16	PPL	-
8	WPT	10.2	17	PHO	36 ca
9	EKO	5.4	18	KHU	45 ca

report by R. Koyanagi:

"The two events at 2032 and 2035 HST, 18 February 1971 showed the type of seismic recordings characteristic of rockfalls that occur at Mauna Ulu. The arrival times at the various stations are poor, and so precise determination of epicentres is difficult. However, the records showed high attenuation of amplitude and predominant long-period seismic waves (0.5 - 1.0 sec) at increasing distances from Puu Huluhulu (station PHH) so that these events appear very shallow and near Mauna Ulu. The event at 2032 measured an approximate magnitude of 0.9, and the one at 2035 measured an approximate magnitude of -0.1."

The times of the rockfall earthquake read from Figs 10.19, 10.21 and also from the smoked paper records from Hawaiian Volcano Observatory, were plotted against distance from the rockfall site (Fig 10.22). The time distance curves determined by Eaton (1962) and Hill (1969) have been superimposed on the data. It is clear that two P waves about 4 s apart were radiated by the rockfall, and that there are sizeable anomalies in the travel times. Clear arrivals at MPR and AHU are about 1 s late, and those at NPT and WPT

Fig 10.22: Arrival times versus distance from Mauna Ulu vent for the rockfall earthquake at 2032 HT on 18 February 1971. Where two recordings from the same seismometer were available, they were read and plotted independently. It appears that 2 P-waves were radiated about 4 s apart. The P-wave curves adopted by Hill (dotted) and Eaton (solid) have been fitted by eye to the data.

are about 0.5 s early. These anomalies cannot be substantially reduced by altering the epicentre or depth.

There is little to choose between the curves of Eaton and Hill, but it does appear that the 1.8 km s^{-1} surface layer which Hill inferred from long range refraction surveys along the Puna coast line may not be present near the summit of Kilauea. Possibly the 3.1 km s^{-1} layer reaches the surface here.

The maximum phase at PHH recorded by the tape seismograph 5 s after the second P arrival is not easily explained. It is probably a surface wave, but if it originated with the second P arrival, its velocity would be only 0.13 km s^{-1} , much lower than the velocity of 0.4 to 0.6 km s^{-1} for the maximum amplitudes at AHU, WPT, and UWE.

Possibly the fall was a three stage event, the first two stages being fracturing which generated P waves but negligible L waves, and the third stage being the arresting of the block in the lake which caused surging and generated L waves but negligible P waves. The minimum time for the block to fall 25 m into the lake is 2.3 s and it is suggested that the origin time for the maximum phase was about 4.5 s later than that for the second P. This time is possible because the block initially extended

below the lake surface, and keeled over as it fell, and this would increase the time of the fall.

The multi-stage origin facilitates explanation of the differences in the recordings of other rockfalls. These were obviously occurring, because the flat ground of area 2500 m^2 and volume 10^5 m^3 , which in early March separated the lake from the trench to the east (Fig 10.4), had completely disappeared by 13 April 1971.

If the seismic energy of the rockfall is generated in the two main stages clearly separated on the PHH tape seismogram, it is worth while considering separately their seismic energy (E). Assuming that they were separate P and Rayleigh waves, the energy of each was obtained by summation of seismic power, calculated using equations 3.3 and 3.5 respectively, over the duration of each phase. The data were:

Quantity	Value	Source
<u>P Wave</u>		
Peak Ground Velocity	$350 \text{ } \mu\text{m s}^{-1}$	PHH Tape, Ch 2
Duration	1.5 s	PHH Tape, Ch 2
Wave Velocity	3.1 km s^{-1}	Fig 10.22
Distance	0.7 km	Fig 10.2
Density	2.5 Mg m^{-3}	Assumed

Quantity	Value	Source
<u>Rayleigh Wave</u>		
Peak Ground Velocity	1000 $\mu\text{m s}^{-1}$	PHH Tape, Ch 2
Duration	1.3 s	" " "
Wave Period	0.72 s	" " "
Wave Velocity	1.5 km s^{-1}	Assumed

The results were:

$$\text{Energy of main P wave} = 2 \times 10^6 \text{ J}$$

$$\text{Energy of maximum phase} = 6 \times 10^6 \text{ J}$$

The average seismic power of the maximum phase over 1.3 s is then 5 MW, in good agreement with the nominal peak seismic power of 4 MW from the power chart.

The gravitational energy released by the rockfall, assuming an average drop of 12.5 m for the 12,500 m^3 section initially above lake level, was $4 \times 10^9 \text{ J}$. Thus the apparent seismic efficiency was 0.2%.

The seismic energy was also calculated from the magnitude $M_L = 0.9$, and the relation between M_L and E given by Richter (1958).

$$\text{Log}_{10} E = 9.9 + 1.9 M_L - 0.024 M_L^2$$

From this $E = 4 \times 10^4$ J, which is 1/200 of the value from the PHH tape seismogram. The reliability of the magnitude is unknown, but if it is correct, then seismic power values on Figures 10.5, 10.6, 10.7 and 10.8 would be too high by a factor of about 200, and the seismic efficiency would be 0.001%, which is an unusually low value.

Other unobserved rockfalls were identified on the recordings by using the following characteristics of the rockfall earthquake observed on 18 February:

- (i) Spectrum similar to the tremor spectrum.
- (ii) Concurrent change in tremor power and/or spectrum.
- (iii) Power level of earthquake (db) relative to accompanying tremor nearly independent of distance or frequency band of the recording.
- (iv) Lack of ordinary P and S phases.

Not all the events chosen satisfied all the criteria, and even those that did, cannot be proved to be rockfalls, because none were identical to the one observed on 18 February.

It was necessary to list them in 3 Tables, as probable rockfalls (Table 10.2), possible rockfalls (Table 10.3), and earthquakes possibly accompanied by rockfalls (Table 10.4).

TABLE 10.2

Seismic Events Probably Caused By Observed or Inferred Rockfalls

at Mauna Ulu

Date 1971	Time HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases	Assoc Events	Q	Effect on Tremor	Remarks on Speeded-up Sound
	h m	$\mu\text{m s}^{-1}$	Hz	s	J	Max-P Vel. Ampl.	No. Lag; Lead	Minute	Power	Freq.
11 Feb	20 40.1	45	1.0	1.5	5×10^4	0.8	2 lag	1.8 2	Increases	Rises
11 Feb	20 41.9	73	1.3	1.	9×10^4	None	2 lead	1.8 4	Increases	Falls
18 Feb	20 31.8	1200	1.5	1.3	3×10^7	(9 4.6)	2 lag	3 2½	Decreases	None
										Rockfall observed volume 3 12500 m ³ .
22 Feb	06 40.6	250	1.6	2	2.1×10^6	4.7	2 lag	3 2	Starts	None
3 Mar	10 49.9	290	1.25	1.5	2.1×10^6	3 ca	2 lag	1.4 3	Stops	" "
12 Mar	21 49.8	54	1.2	1.7	8×10^4	1	3.6 1	- 3½	Stops	" "
15 Mar	03 44.5	81	1.6	1.2	1.3×10^5	None?	1 -	- 3½	Increases then stops	"
18 Mar	06 58.0	730	2.1	2	1.8×10^7	None	1 -	- 3	Decreases	"
18 Mar	13 33.9	170	1.4	1.5	7×10^5	None	1 -	- 6	Decreases slightly	"
19 Mar	02 52.5	130	2.3	1	3×10^5	None	2 lag	1.5 6	Decreases	"
19 Mar	02 54.0	180	1.5	1.5	8×10^5	2	2 lead	1.5 6	Stops	"
22 Mar	12 15.4	110	1.1	1	2.1×10^5	1	2 lag	0.25 4½	Decreases	"
22 Mar	12 15.6	100	1.25	1	1.7×10^5	0.5	2 lead	0.25 4	"	"
22 Mar	21 54.3	50	1.1	2	9×10^4	None?	1 -	- 5	"	"
24 Mar	17 19.6	250	1.4	1	1.1×10^6	None	2 lead	1.3 4½	Increases then decays	Changes
										Knock sound.

Date 1971	Time HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases		Assoc Events	Q	Effect on Tremor		Remarks on Speeded-up Sound
						Max-P	Vel. Ampl.			Power	Freq.	
	h m	$\mu\text{m s}^{-1}$	Hz	s	J	s	$\mu\text{m s}^{-1}$	No. Lag;Lead				
								Minute				
25 Mar	01 15.2	76	1.4	3	3×10^5	4	-	2 lag 2.1	4	Increases	Changes	Knock sound.
25 Mar	01 17.3	55	1.4	4	2.1×10^5	None	-	2 lead 2.1	>6	"	"	"
28 Mar	20 16.8	100	2.8	2.5	4×10^5	-	-	2 lag 0.65		None	Obscure	Double onset, the 1st of higher freq.
28 Mar	20 17.4	1240	7	2.5	6×10^7	4,9,10		2 lead 0.65		Increases	Marked rise	Rockfall observed by HVO personnel, triple onset.

TABLE 10.3: Seismic Events Possibly Caused by Rockfalls at Mauna Ulu

Date 1971	Time HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases		Assoc Events	Effect on Tremor	Remarks on Speeded-up Sound
						Max-P	Vel. Ampl.			
		h m	Hz	s	J	s	$\mu\text{m s}^{-1}$	No. Lag;Lead	Power	Freq.
12 Feb	01	11.3	68	1.5	1.2×10^5	1.2	16	1	None	Decreases Mushy knock sound
13 Feb	04	14.7	14	4	1.3×10^4	None	-	2 lead 0.8	None	Possibly P and S of deep earthquake.
17 Feb	17	53.8	39	4.5	1.2×10^5	None?	-	1	Decreases	Knock with vague double onset and strong coda of 1 minute duration.
18 Feb	11	49.1	100	2.5	4×10^5	None	-	1	Decreases 3min later	Mushy knock. Observers reported no rockfalls.
19 Feb	09	48.1	45	2.5	9×10^4	None?	-	2 lag 2.9	Decreases	Mushy knock with high freq.onset.
19 Feb	19	41.3	38	4.7	1.1×10^5	2	6	1	Increases and then dies away	" " " "
20 Feb	20	35.5	110	3	6×10^5	1.5	25	1	Increases	Mushy Knock.
21 Feb	17	38.5	14	2.5	8×10^3	2.7	3.8	1	"	Bang with high freq. onset.
24 Feb	09	43.4	120	1.5	4×10^5	None	-	1	None	Sharp knock with 1.5 min. coda.
24 Feb	23	33.5	57	1.5	8×10^4	None	-	1	Decreases	High freq. knock with 0.5 min. coda
25 Feb	13	02.6	46	3	1.1×10^5	2.5	16	2 lead 0.4	Increases in coda then decreases & stops.	Bang with high freq.onset follows vague event 0.4 min.earlier.
27 Feb	10	49.4	66	2.5	1.8×10^5	3.2	24	2 lag 2.4	Decreases to zero	Mushy knock with high freq.onset

Date 1971	Time HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases	Assoc Events	Effect on Tremor	Remarks on Speeded-up Sound
	h m	$\mu\text{m s}^{-1}$	Hz	s	J	Max-P s	Vel. $\mu\text{m s}^{-1}$	Power	
							No. Lag;Lead Ampl.	Freq.	
27 Feb 11	32.1	90	1.7	2	3×10^5	2	8.5 2? lag 1.5?	" "	"
1 Mar 10	15.6	22	1.2	8	7×10^4	None	- 1	Decreases	Mushy bang with high freq. onset.
2 Mar 14	23.6	63	1.8	1.5	1×10^5	2	21 1 -	Restart tremor	Mushy knock with 0.7 minute coda.
5 Mar 12	53.7	22	1.2	2	1.7×10^4	None	- 1 -	Decreases and stops	Bang with vague double onset.
9 Mar 16	19.7	32	1.3	1.5	2.6×10^4	None	- 1	None	Knock with 0.8 min. coda.
17 Mar 08	32.2	33	2.1	2	4×10^4	None	- 1	None	Knock partly masked by tremor.
24 Mar 06	12.1	27	3.0	2	2.5×10^4	None	- 5 lag 0.6 lead 2.3, 2.5, 2.7	Increases	Knock with decreasing freq. and 2 minute coda
24 Mar 06	12.6	60	1.2	1.5	9×10^4	None	- 5 lead 0.6, 2.9, 3.1, 3.3	Increases	High freq. knock with low freq. coda.
25 Mar 19	07.6	410	2.0	2.5	7×10^6	3	5.5 1 -	Change	Knock of nearly uniform freq.
27 Mar 19	07.1	50	1.0	2.5	1.1×10^5	None	- 1 -	Increases	Double onset; double rockfall or P & S.
28 Mar 05	52.0	160	1.8	1.2	5×10^5	2	21 2 ?	Begins decrease	Mushy knock with 0.5 minute coda.
30 Mar 06	51.8	91	1.3	2.5	4×10^5	4.5	39 1 -	" "	Sharp knock with 0.4 minute coda.
									Double knock, the 1st of higher freq.

Date 1971	Time HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases	Assoc Events	Effect on Tremor	Remarks on Speeded-up Sound
	h m	$\mu\text{m s}^{-1}$	Hz	s	J	Max-P s	No. Lag; Lead Minute	Power Freq.	
2 Apr	20 53.6	1180	1.0	1.5	4×10^7	7	2 Lag 5.1	None (Tremor Stopped)	Double knock, the 1st weak & of higher freq.
2 Apr	20 58.7	75	1.0	2	1.9×10^5	2	2 lead 5.1	None (Tremor Stopped)	Mushy knock, with high freq. onset.
7 Apr	03 25.1	20	2.3	2	1.4×10^4	None	1	Decreases after event	Knock in moder- ately strong tremor.

TABLE 10.4: Apparent A-Type Earthquakes Which Affect Tremor at Mauna Ulu.

Date 1971	Time	HT	Velocity Ampl.	Dominant Freq.	Duration of Large Ampl.	Nominal Seismic Energy	Prelim Phases	Assoc Events	Effect on Tremor	Remarks on Speeded-up Sound
	h	m	$\mu\text{m s}^{-1}$	Hz	s	J	S - P M - P	No. Lag;Lead Minute	Power Freq.	
12 Feb	22	39.7	39	2.4	2.5	6×10^4	16 3½	1 -	Marked Increase	None
22 Feb	21	18.6	62	2.3	2.5	1.6×10^5	2 -	2 lead 0.4	Begins Increase	None
26 Feb	01	11.6	390	4	1	2.6×10^6	2 -	2 lead 1.6	Stops	None
7 Mar	21	18.6	25	3.1	2	2.1×10^4	1.7 -	1 -	Begins decrease	None
20 Mar	06	19.4	79	4	3.5	4×10^5	- 2.3	1 -	decreases	None
21 Mar	02	03.1	110	1.9	3.6	7×10^5	3 -	4 Several small after shocks.	Starts Tremor	None
23 Mar	09	17.6	49	1.4	1.5	6×10^4	3.1 -	1	Begins Stoppage	None
24 Mar	06	09.8	220	1.4	2.5	2.1×10^6	3 6	4 lead 0.4 lag 2.3, 2.8.	Decreases	Increases
25 Mar	09	23.3	59	2.5	5	3×10^5	2.8 -	1 -	Starts Tremor	None
2 Apr	00	02.0	86	1.3	6	8×10^5	1.5 4	1 -	Dies away in coda	None

In the latter the rockfalls are imagined to be triggered by nearby earthquakes. Examples of seismograms and sonograms chosen from each group are shown in Fig 10.23.

Considering the data as a whole, criterion (iv) is the least reliable indication. Presumably (ii) may fail because rockfalls occurring when the lava lake is quiescent and depleted in gas do not cause degassing to start (Fig 10.23b), and those occurring diametrically opposite the degassing point may not cause it to stop. Or the change in tremor may be so gradual as to raise doubts as to its cause. Criterion (iv) may fail because the shear failure of the collapsing wall may be prolonged, as it fortunately was on 18 February (Fig 10.19), and so cause vibration similar to P and S of a more distant earthquake, or the rockfall may be precipitated by an A or B-type earthquake, and appear as a modification of its coda (Fig 10.23c).

Two interesting results can be found from the seismograms of rockfalls. The maximum value of the Q of the source area can be calculated from the logarithmic decrement measured on the Puu Huluhulu seismogram. Presumably some rockfalls are brief simple seismic sources, consisting of an upward rebound of the surviving wall as the falling block breaks free, followed by a downward pulse as the block stops in the lake.

Fig 10.23: Seismic recordings at PHH of probable rockfall earthquakes at 0252 and 0254 on 19 March 1971.

Upper: Sonagram showing similarity of tremor and earthquake spectra, and decrease of tremor after each earthquake.

Centre: Profiles of seismic power spectral density with frequency resolution 0.2 Hz, stability 6 db with 90% confidence, $0 \text{ db} \equiv 2.5 \text{ W Hz}^{-1}$. Curves of uniform spectral density are shown with each profile.

Lower: Low gain seismogram: amplitude scale $1 \text{ mm} \equiv 21 \text{ } \mu\text{m s}^{-1}$.

Fig 10.24: Seismic recordings at PHH of a possible rockfall earthquake at 2053 HT on 2 April 1971.

Upper: Sonagram showing that the earthquake occurred 3.5 minutes after tremor stopped, and did not restart it. Spectral densities initially increase above 3 Hz as in a P phase, and then become very strong at similar frequencies to the preceding tremor.

Centre: Profiles of seismic power spectral density. Frequency resolution 0.2 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 4 W Hz⁻¹. Response curves for constant spectral density are shown.

Lower: Low gain seismogram, scale 1 mm $\equiv$ 120 μ m s⁻¹. A weak P onset is visible 7 s before the main onset.

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Fig 10.25: Seismic recordings at PHH of earthquakes accompanied by a rockfall at 2017 HT on 28 March 1971.

- a/ Sonagram showing similarity of spectra of tremor and the coda of the earthquakes, and the increased spectral density of tremor above 5 Hz after the earthquakes.
- b/ Profiles of seismic power spectral density with frequency resolution 0.2 Hz, stability 6 db with 90% confidence, and $0 \text{ db} \equiv 4 \text{ W Hz}^{-1}$.
- c/ High gain seismogram showing apparent P and S phases with S - P interval 4.5 s.
Scale $1 \text{ mm} \equiv 14 \text{ } \mu\text{m s}^{-1}$.
- d/ Low gain seismogram showing typical rockfall vibrations in the coda of the second earthquake.
Scale $1 \text{ mm} \equiv 170 \text{ } \mu\text{m s}^{-1}$.

If so, the oscillatory wave train is the response of the source region, but it could be prolonged by release of energy stored in the lava lake, thus reducing the logarithmic decrement, and giving an erroneously high value of Q . Values listed in Table 10.2 are between 2 and 6. These very low values may be explained by the high damping viscous lava provides for seiches, shear waves, and surface waves. The other result (Fig 10.24) is the cumulative seismic energy of the collapses, which when combined with the relation between the seismic energy and volume of the observed rockfall on 18 February, gives a measure of the rate at which rockfalls were occurring. The total cumulative volume of collapse calculated from the seismic data is about 72500 m^3 , which is close to the volume of 10^5 m^3 estimated by eye to have collapsed at the east end, but may be only half the volume from the whole circumference (excluding volumes below the surface of the lake).

The trend in the rockfall volume rate had a maximum nominal value of $3 \times 10^3 \text{ m}^3$ per day between 17 March and 5 April. The cause of the increase is uncertain. Presumably the walls were being undermined by melting, or they would not have remained vertical when they collapsed.

Fig 10.26: Cumulative seismic energy versus date of probable and possible rockfall earthquakes, and of earthquakes possibly accompanied by significant rockfalls at Mauna Ulu.

The degassing cycle sometimes changed at the time of a collapse, as at 2017 on 28 March. Probably this was a result of the collapse rather than a cause, but the rise and fall cycle of the lava must have had some loosening effect on the walls. The level of the lava lake was noticeably lower on 13 April than it had been on 19 March, and this would have allowed rockfalls to occur more easily. The reduced rate after 5 April merely indicates that the least stable sections of the wall had already fallen.

VARIATION IN TREMOR LEVEL AND SPECTRA WITH DISTANCE

On 17 February and 13 April 1971, recordings averaging 15 minutes duration were made at 24 sites using seismograph A, while seismograph B recorded continuously at Puu Huluhulu. Differences in level A minus B db (bandwidth 0.5 - 20 Hz) versus distance Δ from the degassing point in Mauna Ulu lava lake (Fig 10.25) lie on a curve which for distances between 0.05 and 1 km, approximately fits the relation:

$$(\text{Particle velocity})^2 \propto \Delta^{-1}$$

Fig 10.27: Change of tremor intensity A - B db
(difference in RMS amplitude between portable seismo-
graph A and fixed seismograph B) versus distance from
site of degassing in Mauna Ulu lava lake. A curve
together with a line corresponding to ground velocity
 $\propto (\text{distance})^{-1/2}$ have been eye-fitted to the data.

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which applies to two-dimensional propagation as in a surface layer, while between 1 and 10 km, the data approximates the relation:

$$(\text{Particle velocity})^2 \propto \Delta^{-2}$$

which is the theoretical relation for spherical propagation in a homogeneous medium, and which the author assumes in routine calculation of seismic power.

It cannot be claimed that either type of propagation applies uniquely however. The percentage errors in the distances close to Mauna Ulu could be large enough to allow of spherical propagation from a source at a depth of 0.35 km below Mauna Ulu, or the propagation could be entirely two-dimensional with absorption explaining the increasing attenuation with distance. More probably a mixture of two-dimensional, spherical and refraction propagation at 3 or 4 different velocities is occurring with appropriate absorption in each case, and in proportions which cannot be determined from the data.

The tremor spectrum at Puu Huluhulu was very consistent on 17 February when most of the mobile recordings were made, and at about half the mobile sites, tremor from Mauna Ulu could be reliably identified by its simultaneous variation with tremor at Puu Huluhulu. By analysing the simultaneous high power tremor, errors due to vehicles and natural sources other than Mauna Ulu were reduced.

The power spectral densities (Fig 10.26) relative to a constant power level at Puu Huluhulu show that the tremor bandwidth becomes narrower at increasing distances, and that the spectral peak is consistently 1.3 Hz and attenuates slowly. The slope of the low frequency flank of the spectral peak does not vary appreciably with distance, because it is controlled by the cut off slope of the seismometer. The slope of the high frequency flank of the peak (db Hz^{-1}) is roughly linear and the smoothed gradient over 1.3 to 8 Hz appears to increase linearly with distance up to 9.5 km from Mauna Ulu (Fig 10.27) at a rate of $1.1 \pm 0.5 \text{ db Hz}^{-1} \text{ km}^{-1}$.

The gradient of the tremor spectrum at zero distance is close to zero when corrections for seismograph response are applied. That is, tremor from Mauna Ulu lava lake approximates a white noise at its source.

Possible errors will now be considered.

- (i) The spectral analyses may be in error. The sonagraph analyses were made with 11 degrees of freedom, and with 90% confidence should be within 3.2 db of the true mean values. However

Fig 10.28: Profiles of power spectral density of volcanic tremor at increasing distances from Mauna Ulu, (see also Fig A.3). Stability 3.2 db with 90% confidence. Recording site numbers are (from left) 1, 5, 2, 6, 3, 4, 8, 14, 15, 9, 16, 17, 12, 18, 19.

VOLCANO - SEISMIC NOISE SPECTRA AT INCREASING DISTANCES FROM MAUNA ULU LAVA LAKE
VERTICAL COMPONENT RECORDED 17 FEB 1971. FILTER BANDWIDTH 0.33 HZ. POWER LEVELS RELATIVE.

Fig 10.29: Spectral profile slope versus recording distance of tremor. Data at distances >1 km are the slopes in db Hz^{-1} above 1.3 Hz of profiles in Fig 10.28. Data at distances <1 km were recorded on 13 April 1971 and analysed with different equipment.

Dashed line: eye-fitted curve assuming the data is homogeneous.

Solid lines: eye-fitted curve to preferred data (squares) at distances >1 km, plus a parallel line through data at <1 km.

the Sona-graph machine at Hawaii Institute of Geophysics, used for the distance range 1.4 to 13.7 km, was not in good condition, and errors up to ± 5 db may be present due to instability in its frequency response. Subsequent analyses for shorter distances were made on an excellent machine made available by Dr Art Popper of the Fish Research Laboratory, but there was no opportunity to repeat the previous work.

- (ii) The spectra may be contaminated. At Hilina Pali (site 19), 13.7 km from Mauna Ulu, the vibration level did not follow the tremor variations at PHH, and it appears that microseisms from the coast 3.7 km away were the strongest vibrations.

They also gave a peak at 1.3 Hz in the uncorrected spectrum (Fig 10.26). Therefore data at this distance was omitted on Figure 10.27. To further reduce the effects of contamination, the data on Figure 10.27 at distances 1.44, 1.87, 2.64, 4.1 (lower), and 5.9 km, where the tremor to microseism ratio was very good, and which were obtained with the same Sona-graph machine, were considered separately.

They define a rate of change of spectral gradient with distance of $0.58 \text{ db Hz}^{-1} \text{ km}^{-1}$, and reveal a step of approximately 0.7 db Hz^{-1} between the values obtained with the two sona-graphs.

The value of $0.58 \text{ db Hz}^{-1} \text{ km}^{-1}$ is considered more reliable than $1.1 \text{ db Hz}^{-1} \text{ km}^{-1}$ and will be adopted as a minimum estimate in the calculation of Q .

The change of spectral shape with distance can be caused by absorption, which in the solid earth is proportional to frequency (Knopoff and McDonald 1958), and by discrimination in a resonant transmission path, but in the latter, a linear change would not be expected. For the first case, Q can be calculated using equations 3.51 and 3.52. Accordingly:

$$\begin{aligned} q &= \frac{0.58}{8.686} \text{ km}^{-1} \text{ Hz}^{-1} \\ &= 6.67 \times 10^{-5} \text{ m}^{-1} \text{ Hz}^{-1} \end{aligned}$$

To calculate Q we need the seismic velocity of tremor, which at Kilauea is not well defined. Shimosuru et al (1966) found that it apparently varied from 3.18 to 0.3 km s^{-1} as frequency changed from 0.9 to 3 Hz, and they found evidence that frequency was a complex function of depth and place of origin. A medium velocity of

1.5 km s^{-1} will be adopted. Then:

$$\alpha\lambda = qc = 0.1$$

$$Q = 32 \quad \text{within a factor of 4.}$$

This value agrees with that calculated from the decay trains of the dyke splitting earthquakes on 28 January 1971, and apparently pertains to solid rock, whereas the values as low as 2, calculated from the decay trains of rockfalls into the lava lake, apparently relate to the liquid lava.

THE TREMOR SPECTRUM AND ITS VARIATION WITH TIME

The spectrum of volcanic tremor at Puu Huluhulu could be classified by means of the following six features visible on the sonagrams:

- (i) Broad dominant band of approximately constant power spectral density between 1.3 and 5 Hz.
- (ii) Clearly dominant peak at 1.3 ± 0.2 Hz.
- (iii) Dominant frequency clearly above 2 Hz.
- (iv) High spectral densities below 1 Hz.
- (v) Narrow persistent spectral troughs.
- (vi) Narrow persistent spectral lines.

These features will now be described and illustrated.

Feature (i) The lower edge of the 1 - 5 Hz band was determined by the natural frequency of the seismometer, and the upper edge by an inclined step decrease of 5 - 10 db to a lower level above 5 Hz. Sometimes the band consisted of overlapping peaks at 1.3, 3 and 5 Hz. Figure 10.28 shows this feature on 3 March 1971. The event at 1049.9 HT (4.7 minutes on the figure) was probably a rockfall earthquake which caused tremor to decrease.

Feature (ii) This can be seen on the sonagram (Fig 10.29) of the vibration during the switching on 10 February 1971 of the lava flow from the west flank of Mauna Ulu to the tubes leading south east. The 1.3 Hz peak at 02 and 09 minute is considerably larger than the peak in the seismograph response near that frequency.

Feature (iii) At 04.5 minute on Fig 10.29, (equivalent to 1456.5 HT on 10 February 1971) a burst of tremor of dominant frequency 3.7 Hz shows feature (iii). It was accompanied by micro-earthquakes of lower power than the tremor, which are seen as vertical striations. Figure 10.13 shows continuous occurrence of feature (iii) on 30

Fig 10.30: Sonagram showing the 1 - 5 Hz band of high spectral density which was the most common spectral feature of tremor recorded at PHH. The event at 5 minutes is a probable rockfall at 1050 HT on 3 March 1971. Frequency resolution 0.19 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 1.6 W Hz⁻¹.

Fig 10.31: Sonagram showing two common spectral features of tremor at PHH: a dominant spectral peak of 1.3 Hz between 09 and 16 minutes, and of 3.5 Hz at 5 minutes, Zero minutes $\equiv$ 1452 HT on 10 February 1971. Frequency resolution 0.1 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 4 W Hz⁻¹ ca.

January during the eruption of lava fountains up to 5 m high on the west flank of Mauna Ulu.

Feature (iv) Fig 10.31 shows that the dominant frequency was 0.5 Hz at 23.0 HT on 1 February 1971. The Q of the peak was at least 6. It is evident that the 0.5 Hz peak had a separate origin to the rest of the spectra, because the spectrum of the earthquake at 27 minutes was involved with the latter, but not with the 0.5 Hz peak.

Feature (v) Fig 10.32 shows a trough of frequency 1.1 Hz and width 0.07 Hz, at the time the seismograph was calibrated at 11.0 HT on 19 March 1971. The clarity of the trough on the display between profiles shows it was not a statistical accident, whereas the profile with stability 6 db at the 90% confidence level cannot prove this. Ten minutes after the time displayed on the sonagram, steady ebullition was observed at the east end of the Mauna Ulu lava lake.

Feature (vi) Fig 10.33 shows spectral lines at 1.0, 1.15 and 1.3 Hz, recorded between 1540 and 1620 HT on 19 March 1971, when ebullition was occurring at the east end of the lava lake. Figs 10.30 and 10.31 also show spectral lines.

Fig 10.32: Sonagram showing a dominant frequency of 0.5 Hz in tremor at PHH (cf the curves of uniform seismic power spectral density). The volcanic earthquake at 27 min (23.0 HT on 1 February 1971) is involved with the tremor spectrum above 1 Hz, but not at 0.5 Hz. Resolution 0.05 Hz, stability 6 db with 90% confidence, 0 db $\equiv 6.5 \text{ W Hz}^{-1}$.

Fig 10.33: Sonagram showing a narrow spectral trough at 1.1 Hz, in tremor at PHH. At 17.5 minute ($\equiv$ 23.0 HT on 19 March 1971), the 0.5 Hz square-wave calibration signal of 100 μV peak amplitude is analysed. Vertical ticks show the theoretical spectra. Resolution 0.05 Hz, stability 6 db with 90% confidence, 0 db $\equiv 1.2 \text{ W Hz}^{-1}$.

Fig 10.34: Sonagram showing spectral lines at 1.0, 1.15, and 1.3 Hz in tremor at PHH. Zero minutes $\equiv$ 1540 on 19 March 1971. Spectral density of tremor and ebullition at the east end of the lava lake reached a maximum at 1610 HT. Resolution 0.05 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 1.2 W Hz⁻¹.

It is evident that the combination of spectral features, which make up the tremor spectrum, changes with time. To investigate these changes, and, where possible, the visible activity accompanying them, recordings were selected for Sona-graph analysis using three criteria:

- (1) Their sound during high speed playback was interesting.
- (2) They corresponded to interesting activity observed between the author's visits.
- (3) They corresponded with visual observations of the lava lake, and seismic recording surveys made by the author.

Even with these restrictions it has not been possible to make a complete study of spectral changes with time. With one exception only the longer periods analysed as sequences of overlapping sonagrams have been described. As the descriptions have been done in some detail, they are presented as Appendix A.

Summary of Spectral Results

The following generalisations can be made from the sequences of sonagrams described in Appendix A, and from the other single sonagrams which make up the 130 prepared for study.

Feature (1) was the normal spectral feature of tremor during the recording period, especially after 10 February. It was observed to occur during ebullition anywhere in the lake except the south side, where ebullition was not observed.

Feature (ii) was common throughout the recording period, and was observed to occur during ebullition at the north and north west sides of the lake. Although the frequency of the peak is near the natural frequency of the seismometer, its reality is shown by its appearance and disappearance over short time intervals, and its level on Figures 10.29, 10.32, 10.33, of 12 to 18 db above the average spectral densities at other frequencies.

Feature (iii) occurred only during the eruption from the rift west of the main vent between 28 January and 10 February 1971, when the dominant spectral peak was usually 4 Hz, and reached 6 to 12 db above the average spectral densities away from the peak.

Feature (iv) appeared in three forms:

- (a) a strong spectral line at 0.5 - 0.6 Hz;
- (b) a strong line at 0.8 - 0.9 Hz;

(c) downward widening of the peak at about 1.3 Hz.

The first form was found to be only loosely associated with the remainder of the spectrum, as though due to a separate resonator which was excited by tremor, but it sometimes appeared to be the frequency of greatest spectral density. The second form was less well observed. The third form was detected on the charts of seismic power by comparing the relative power in the band 0.5 - 1 Hz with that in the band 1 - 4 Hz (Figs 10.5 to 10.8). Power at 0.5 - 1 Hz was relatively high for periods exceeding 30 minutes on 22, 26, 28 January; 7, 12, 17 February; and 3, 23, 24, 31 March. Seismic power was also high at these times.

Features (v) and (vi) were frequent prior to 10 February, and at other times when the seismic power of tremor was uniform, or increasing smoothly (to a crescendo). They were rare during periods of tremor pulses and cyclic degassing. The commonest line was 1.3 Hz, the Q of which reached 15 (Fig 10.33). Next was 0.5 - 0.6 Hz, for which Q reached 8 (Fig 10.31). These values of Q are limited by the bandwidth of the Sona-graph filter (0.065 Hz between

-3 db points for the standard sonagram) to 15 at 1 Hz and 8 and 0.5 Hz. Lines with Q of 8 or more were rare above 2 Hz at Puu Huluhulu, but were recorded at the lava lake (site 44).

The most persistent trough was a broad one at 2.2 Hz. Many other temporary lines and troughs appeared, and it was generally impossible to determine which had real causes, (such as a trough caused by an energy absorber) and which were residual features (such as a trough between spectral lines due to resonators).

On two occasions this question appeared to be answered by peaks changing into troughs and vice versa. To regard these apparent reversals as accidental would be special pleading. On one occasion the position of ebullition was observed to change at about the time the spectrum changed, but other similar changes in ebullition were not accompanied by spectral reversals.

On another occasion, a spectral peak glided down through the spectrum and up again, and was reinforced where it coincided with fixed peaks. This indicates that two systems in series were controlling the spectrum. Probably the variable frequency system was the more closely linked with the lava-gas system.

When ebullition⁴ in the lake was stopped, residual tremor sometimes had maximum spectral density at 2.5 - 3 Hz and sometimes at 1.3 Hz. The recommencement was sometimes simultaneous at all frequencies, as might be expected if ebullition was initiated by a triggering event such as a rockfall. At other times it occurred first at 1.3 Hz and progressively later at higher and lower frequencies.

Mechanisms of Degassing and Tremor
at Mauna Ulu, 1971

SUMMARY OF OBSERVATIONS

The author's observations, confirmed by valuable discussions with Drs. D.W. Peterson, D.A. Swanson, and W.A. Duffield of Hawaiian Volcano Observatory, can be summarised as follows under headings of:

(i) time, (ii) position, (iii) dynamics.

(i) Time:

Ebullition and strong tremor were observed to begin and end within a minute of each other; this period is assessed to be the accuracy of observation. Convection increased and decreased at about the same time as ebullition.

(ii) Position:

Ebullition began anywhere around the vent walls, or where falling rock or lava skin was sinking within the area of the lake. Hawaiian Volcano Observatory staff were sometimes able to induce it by throwing suitable

material into the lake (W.A. Duffield, public lecture 12 May 1972). The ebullition site appeared to be superficially controlled. The tremor spectrum was related to the ebullition site.

(iii) Dynamics:

Ebullition commenced and decreased quickly, and usually consisted of the bursting of bubbles with diameters up to several metres at an average rate of the order of 1 per second. The average size of spatter fragments decreased and their number increased with increasing violence of ebullition. Surface convection of the lake was towards the dominant ebullition site and Hawaiian Volcano Observatory staff determined that the lake sloped down in this direction (D.A. Swanson personal communication). The ebullition site moved from its initial position towards the nearest wall, and then usually towards preferred positions at either end of the lake. The lake level fell during stronger than average ebullition, rose when ebullition stopped, and was fairly steady during continuous average ebullition.

PROPOSED MECHANISM OF EBULLITION

The commencement of ebullition at almost any irregularity in the surface of the lava lake shows that ebullition can originate in a near surface layer. The convection and slope which develops towards the new centre shows that ebullition is not an upwelling of fresh gas charged magma from great depth, or the ascent of trains of growing bubbles from great depth. Apparently it is the release of gas bubbles from a near surface layer of vesicular lava. The feasibility of separating gas in large bubbles from a layer of lava foam is demonstrated by the fragile cavernous lava, which had overflowed from the lake to build the Mauna Ulu shield in 1970. The highly vesicular crust about 0.1 m thick frequently collapsed into an almost continuous cavern up to 0.3 m high as one walked across it. It is proposed that during ebullition, a near surface layer of lava foam collapses into ascending gas and descending denser magma, and is replaced by inward flow of the collapsing layer.

Lava foam and vesiculation is an idea of long standing in volcanology. Perret (1950 p 43) and Jagger (1947) both recognised its great importance, but neither specifically

considered the collapse of foam as an explanation for the fountaining activity which they described in Halemaumau lava lake between 1911 and 1924. This activity was in many ways similar to, although more variable than, the ebullition in the lava lake of Mauna Ulu. Both preferred explanations involving vortex like tubes extending to considerable depths, and ^{gas} vents in the bottom of the lava lake. Perrett (1950 p 43) said that gases rise from deep origins (undefined) and make room for the surface layer to plunge down, and Jaggar (1947 p 49, 101) considered that the surface lava plunged down because it was cooler and therefore denser than underlying lava, and that gas was released from it in the plunging process. It is with considerable trepidation that, on the basis of only eight hours observations, the present author proposes a different explanation. The idea of a shallow origin of ebullition to explain the convection direction was suggested to the writer by Dr. D.A. Swanson while we were observing the lake and is gratefully acknowledged. It is beyond the scope of this thesis to solve the problems of thermodynamics, bubble nucleation etc. involved in completely explaining the ebullition process, but the following main results can be deduced from the literature.

(1) Input of water to the lava lake

If Swanson's (1972) concept of a steady flow of 10^7 m^3 per month of rather uniform tholeiitic basalt (Swanson et al 1971) from the mantle of the earth to the lava lake is adopted, the water content of the lava should be close to the primary water content of 0.5% estimated by MacDonald (1963) and Moore (1965). In this case, 18 kg s^{-1} of water was entering the lake with the magma, and following Swanson's (et al 1971) statement that the lava emerging from the tubes south of the lake had lost most of its gas, about 10 kg s^{-1} of water must have been given off from the lake. The volume of steam at 0.5 MPa and $1165 - 1170^\circ\text{C}$, which was consistently the ebullition temperature of the lava (Swanson et al 1971), would have averaged $12 \text{ m}^3 \text{ s}^{-1}$ (extrapolating the specific volumes given for 1 MPa and 1000°C by Kennedy 1957). This can be compared with the average rate of gas release during cyclic degassing in 1969, estimated as $8 - 28 \text{ m}^3 \text{ s}^{-1}$ at 1170°C and a pressure of several atmospheres (from Swanson et al 1971). The power in the steam relative to atmospheric conditions would have been about 60 MW.

(2) Vesiculation

Bubbles can grow bigger than about 10 times their nucleating radius (10^{-6} m ca), only by gaining volatiles from the liquid by diffusion, or by coalescing, (Verhoogen 1951).

The diffusivity is:

$$D = \frac{R T}{6 \pi \eta r N}$$

where   D   is diffusivity
              R   is the gas constant
              T   is the absolute temperature
               η is the viscosity
              r   is the radius of diffusing particles
              N   is Avogadro's number

Diffusion decreases with decreasing temperature and increasing viscosity. The rate of increase in radius of gas bubbles by diffusion is slow, and decreases as the $2/3$ power of the radius (Verhoogen 1951).

(3) Variation of equilibrium state with depth

McBirnie (1963) shows that basalt with 1% water content, which is just saturated at 1100°C and 50 MPa pressure-depth, would expand to a gas/liquid ratio >3 (density $<0.7 \text{ Mg m}^{-3}$) and cool adiabatically to 800°C

if it could reach equilibrium at the surface. In changing from 1200°C with 1% dissolved water to 1100°C with 0.5% dissolved water, the viscosity of the lava would increase by a factor of about 3 (McBirnie and Murase 1971).

An estimate of the changes in bulk density occurring in Mauna Ulu lake can be obtained from the differences in behaviour of the lava in the two compartments of the eruptive fissure in 1969 described at the beginning of this chapter. Difference of 1 - 2 m in height of columns 50 - 100 m high above the confluence of the compartments, indicates approximately 2% difference in the average density of inflated lava above that level. Probably the lava was inflated to greater depths than the bottom of the septum, or the columns would have been more independent of each other. The several tens of metres rise and fall of columns probably more than 50 - 100 m deep indicates a maximum difference of 20% between the density of the inflated and deflated lava. The latter however could still have contained considerable exolved gas.

(4) Velocity of Rise of Bubbles

The velocity of rise of small bubbles in Hawaiian lava can be calculated using Stokes law:

$$V = \frac{2g\rho r^2}{9\eta}$$

Where V is upward velocity m s^{-1}
 r is radius of bubble m
 ρ is density of liquid kg m^{-3}
 g is gravity m s^{-2}
 η is viscosity of liquid $\text{N m}^{-2} \text{ s}$

Using $\rho = 2.75 \text{ Mg m}^{-3}$ (MacDonald 1963 Peck 1966), and $\eta = 10^2 \text{ N m}^{-2} \text{ s}$ (McBirnie and Murase 1971), we find values:

$r, \text{ m}$	10^{-1}	10^{-2}	10^{-3}
$V, \text{ m s}^{-1}$	0.6	6×10^{-3}	6×10^{-5}

If viscosity was 10 times greater near the surface, the velocities would be a tenth of these. For bubbles large enough to have the shape of a spherical cap, Davenport et al (1967) give:

$$V \doteq g \text{ Re}^{\frac{1}{2}}$$

where Re is the radius of the equivalent volume sphere.

Taking $\text{Re} = 1 \text{ m}$

Then $V = 2.1 \text{ m s}^{-1}$

Small spherical bubbles and large cap-shaped bubbles rise in a stable manner, but intermediate sized bubbles have ellipsoidal shape, and may oscillate and rise in a helical manner (Davenport et al, 1967).

There is apparently no adequate theory to predict the velocity of rise or oscillatory behaviour of ellipsoidal bubbles (Davenport et al 1967). If bubbles of 1 m equivalent radius are ellipsoidal in the Mauna Ulu lava lake, their velocity will lie between 2.1 m s^{-1} and the value given by Stokes law (60 m s^{-1}), and in any case, will be less than the maximum velocity of ejection of lava fragments from the lake. From the observations in Table 10.1, this was 12 m s^{-1} .

(5) Collection of Foam

It is evident from 1 - 4 above that if fresh magma is brought from depth into a lava lake, a foam will tend to form in which temperature increases with depth. On the other hand, viscosity, size of bubbles, and gas/liquid ratio will decrease with depth, with viscosity having the greatest rate of change.

(6) Stability of Foam

The stability of a foam against coalescence varies directly with the elasticity and thickness of the film between the bubbles (Osipow 1962).

(7) Elasticity and surface tension of lava films

Film elasticity has been defined by Gibbs (1948) as:

$$E = 2A \left(\frac{d\sigma}{dA} \right)_{T, N_1, N_2}.$$

where A is the surface area of a film

σ is the surface tension

T is the temperature

N_1, N_2 are the chemical components.

For a film containing an adsorbed component which reduces the surface tension (a surface active agent), stretching of the film will decrease the concentration of the adsorbed component and thus increase the surface tension. High elasticity increases the stability of the foam by reducing expansion and distortions due to mechanical, thermal and chemical shocks.

Silica has been identified as a surface active agent by Cooper and Kitchener (1959), and McBirnie and Murase (1971). The latter also identified water and other volatiles as surface active agents, and measured the surface tension of Columbia River basalt as 0.34 N m^{-1} at 1170°C . The elasticity of magma films has not been measured, but should be less for basalt than for other magmas with higher silica and volatile content.

(8) Factors controlling film thickness

The thickness of the films is reduced by:

- (i) Gravitational drainage of liquid.
- (ii) Expansion and distortion of bubbles.
- (iii) Suction of liquid into film intersections, where curvature and surface tension reduce the pressure.
- (iv) Collision of bubbles rising at different rates.

The reduction of thickness is opposed by:

- (i) High viscosity. This also dissipates shear shocks.
- (ii) Electrostatic repulsion of the film surfaces.
- (iii) Spreading of adsorbed components and liquid into thin regions of film.

Osipow's (1962) graph for water plus sodium laurate and lauryl alcohol show foam life increasing proportionally to nearly the square of viscosity.

(9) Investigations of foaming slags

Observations of the stability of foams of silicate melts have been frequently made by the steel industry, because foaming of slags occurs with disadvantage during smelting. Slag foam in blast furnaces has much greater

stability when the viscosity is greater (Ondescin 1949, Cooper and Kitchener 1959, Popel 1960, Kozakevitch 1969) and when the slag contains a dispersed solid phase (Popel 1960, 1963, Kozakevitch 1963, 1969).

Factors which tend to destroy slag foams are increase of temperature (Evans 1944, Kozakevitch 1963, Lingorski 1965), and increase of the basicity ratio (CaO/SiO_2) to 1.7 or higher.

DISCUSSION OF THE MECHANISM OF EBULLITION AND TREMOR

Although data is inadequate, it appears that viscosity, and especially foam life have higher order variations with depth than the other variables in Mauna Ulu lava lake, and cause the lava foam within a metre or so of the surface to be stable. Below this the foam becomes less stable with time, due to growth and coalescence and rise of bubbles, and drainage of liquid from the foam, and when degassing is not occurring, a critically stable layer of increasing thickness develops below the stable surface layer. Collapse and wholesale coalescence of bubbles in the critically stable foam can begin at a wall due to shearing motion,

or within the lake due to skin or rocks sinking through it, and the vibration of collapse and rise of bubbles partially collapses the surface layer in a small area by physical and thermal shock. Parts of the disrupted bubble walls may be erupted as spatter and the remainder falls into the underlying foam, possibly extending the collapse downwards.

It is inferred that it was the vibration of bursting bubbles, both below and at the surface, the vibration of spatter landing, and the oscillatory motion of large bubbles rising rapidly through descending degassed magma, which caused the tremor which the author recorded at Mauna Ulu.

The reduction of pressure in the degassing collapse column allows inward flow of foam, aiding its collapse by distortion, and causing a ring of depression around the ebullition site. A hydraulic gradient develops across the lake and the whole foam layer moves towards the ebullition point, and is replaced by fresh stable foam from depth. Critically stable foam continues to develop over the whole area of the lake, and provided collapse and surface motion are not too fast, it will continue to arrive at the

ebullition site at sufficiently low stability to collapse, so that ebullition can be continuous.

In these conditions, a rockfall some distance from the ebullition may encounter foam too stable to be collapsed, as in Fig 10.18. A rockfall close to the ebullition centre (Fig 10.17 b) may start a second ebullition centre, and thus consume critically stable foam at a faster rate than it is forming, so that ebullition stops at both centres. It is apparent that ebullition should stop rapidly if it is maintained by its own vibration and thermal shock until after the critically stable foam is consumed. Similarly it should not begin until foam with less stability than necessary has built up, causing a powerful recommencement of ebullition and convection which may quickly exhaust the collapsible foam.

The resulting relaxation oscillation could be the cause of cyclic degassing and the short pulses of tremor, and the randomness of its periodicity previously mentioned may be due to randomness in the nature and location of the feature which triggers the collapse, and in the distribution of critically stable foam. It is interesting to note that the periodicity was 10 - 15 minutes in 1969 when the lava lake measured 100 by 5 - 10 m (Swanson et al 1971) and had only increased to 15 - 25 minutes in March 1971 when the

lake was about 10 times larger in area. The rise and fall cycle was of about 10 times greater amplitude in 1969 (Swanson et al 1971) than in 1971, and presumably the maximum foam thickness was also 10 times greater, but apparently it took nearly the same time to become critically stable.

A possible explanation of the pattern of short pulses of tremor between periodic longer periods of tremor is provided by the observation of E.F. Lloyd (personal communication, and C.S.L.P. card 1193) at Erta'ale (a basaltic shield volcano in Ethiopia), that lava fountains moved slowly around the circular 80 m diameter lava lake in an anti-clockwise direction, with the surface motion always towards the fountain. No doubt the consistent migration around the lake is due to the foam being oldest and least stable in that direction.

At Mauna Ulu the lake is not circular; the fountains starting along the sides have only the narrow width of foam across the lake to consume before encountering young stable foam, and stopping again. The next fountain may well begin in a consistent direction around the lake, and periodically will begin at or close to one end, and have the foam of the full length of the lake available to it,

and so have a longer duration. The writer was not at the lake long enough during short period cyclic degassing to observe whether a regular progression occurred, but certainly fountains observed along the sides were all short-lived, and all long-continued fountains observed were at the ends of the lake.

It is less easy to explain the step increases and gradual decreases in the durations of the longer periods of tremor, which were revealed by the spectrum of variation in tremor power. In April 1971 when these features were best developed, the short period cycling occurred at times of minimum smoothed tremor power, and the longest continuous period on 6 Apr was a time of maximum smoothed power (0.5 kW). If seismic power of tremor can be used as an indication of rate of release of gas, then the steps and gradual decreases in the duration of longer periods of tremor reflect steps and decreases in the rate of release of gas.

On 18 February 1971, the lake was observed to deflate by about 45000 m^3 in 5.5 minutes, during which the seismic power of tremor averaged 1 kW. This is equivalent to $137 \text{ m}^3 \text{ s}^{-1}$ of gas per kW of tremor. The long term average gas release of about $15 \text{ m}^3 \text{ s}^{-1}$ corresponds to a long term

average seismic power of approximately 0.2 kW. This is equivalent to $75 \text{ m}^3 \text{ s}^{-1}$ of gas per kW of tremor.

FACTORS WHICH AFFECT THE TREMOR SPECTRUM

It has not proved possible to determine parameters of the ebullition mechanism from spectral data, but a number of factors have been identified:

(i) The seismic velocity of magma. This is not well known. Gorshkov (1958) obtained values of $1.6 - 1.8 \text{ km s}^{-1}$ for P waves passing through subcrustal magma chambers, and deduced that the bulk modulus K was 10^{10} N m^{-2} . Shimosuru (1961) deduced 1.58 km s^{-1} and $7 \times 10^9 \text{ N m}^{-2}$ for basaltic magma at 1190°C . Sassa (1935) calculated 0.79 km s^{-1} at Aso.

The velocity (c) in a lava foam can be calculated with even less certainty from the mean bulk modulus and mean density, using an expression by Wood (1946):

$$c^2 = \frac{K_1 K_2}{x K_2 + (1 - x) K_1} \quad \frac{x \rho_2 + (1 - x) \rho_1}{1}$$

where

- K_1 is bulk modulus of lava
- K_2 is bulk modulus of air
- ρ_1 is density of non-vesicular lava
- ρ_2 is density of compressed air

x is volume proportion of liquid

$1-x$ is volume proportion of gas

Using values $K_1 = 10^{10} \text{ N m}^{-2}$ (Gorshkov 1966); $K_2 = 1.2 \times 10^5 \text{ N m}^{-2}$ (Wood 1946); $\rho_1 = 2.73 \text{ Mg m}^{-3}$ (Macdonald 1963); $\rho_2 = 0.01 \text{ Mg m}^{-3}$ (assumed); $x = 0.8$ (corresponding to the 20% vesicularity for solidified lava flows adopted by Swanson 1972), the resulting velocity is:

$$c = 17 \text{ m s}^{-1}.$$

This must be confirmed before it is believed, even as a possible minimum velocity. Refraction seismic measurements on the surface of the lava lake might be possible, and velocity as a function of depth could be obtained.

(ii) The free vertical oscillation of the lava column has a natural frequency and damping dependent on the vertical gradient of velocity and Q in the magma. There are no data on these gradients, but as Q appears very low, it is expected that the resulting spectrum would be broad and fairly flat. Shimozuru (1961) interpreted the dominant frequency of tremor at various volcanoes including Kilauea by means of this model, and concluded that the oscillating column was 500 m long (vertically).

(iii) The spectrum of underwater explosions (Bancroft 1966) is strongly affected by interfering reflections from the surface (phase reversal), and the bottom and sides of the liquid body. The same should occur in foamy lava which must have a very low seismic velocity due to its high compressibility. For surface reflections the fundamental resonance frequency will be $c/4h$ where h is the depth of the bubbling, and for side reflections the fundamental frequency will be $c/2d$, where d is the distance to the side. As P-wave velocities of the order of 100 m s^{-1} or lower appear to be possible in lava foam, spectral peaks within the seismic recording range might be caused in this way. For instance, the 1.3 Hz peak could be caused by multiple reflections across the width of the lava lake. If ebullition originated at a depth of 25 m, the surface reflections could give a broad spectrum above 1 Hz, with overlapping rising tones as each bubble rose. This may explain the occasions when tremor commences at 1 Hz and extends during some seconds to higher frequencies. Sonagrams do not show tremor to generally consist of rising tones however.

(iv) Resonant bubbles. Bubbles in a liquid subjected to intense sound at their natural frequency are violently active, and tend to destroy any structure within the liquid by their to and fro motion. In particular they could collapse a liquid foam surrounding them. Minnaert (1933) deduced their natural frequency to be:

$$f_o = \frac{1}{2\pi r} \left(\frac{3\gamma p}{\rho} \right)^{\frac{1}{2}}$$

Where r is the radius of bubble,
 γ is the ratio of specific heats,
 p is the gas pressure in the bubble,
 ρ is the density of the liquid.

It is evident that for bubbles rising rapidly through the lake, p decreases and r increases, so that f_o decreases. Overlapping falling tones have not been observed.

As a numerical example, put $f_o = 2$ Hz,
 $p = 0.3$ MPa, $\rho = 2.73$ Mg m⁻³ (Macdonald 1963),
 $\gamma = 1.33$ (assumed by Verhoogen 1951):

Then $r = 1.6$ m. This seems large but not impossible when compared to the bubbles observed at Mauna Ulu.

(v) Frequency of bursting of bubbles. The rate of bursting of bubbles at the surface controls the lower cut off frequency of vibration produced at the surface. The

rate was always observed to lie close to the seismic spectral peak, but showed no control over the peak.

(vi) Frequency of landing of lava fragments. The number and size of spatter fragments per second, landing on the lake surface, determines in part the seismic spectral density at the higher frequencies. Its commencement will always be delayed a second or two by the time of flight. The seismic power from spatter will be a maximum during the highest fountaining, and may partly explain the relatively higher power between 4 and 20 Hz in the strong recommencements of tremor, and at the power peak about half way through the longer periods of tremor (Figs 10.7, 10.8).

CHAPTER 11

STROMBOLI VOLCANO 1971

Introduction

Stromboli volcano is an island about $4\frac{1}{2}$ km in diameter, and 924 m high, situated approximately 60 km north of Sicily, and west of Calabria, Italy. It is the most northerly of the Aeolian Islands, and is much larger and more complex than is visually apparent (Fig 11.1), for it stands in water 2 km deep, and the lavas which are exposed, range in composition from rhyodacites (the oldest) through andesites, trachyandesites, and finally the olivine trachybasalts now being erupted (Rittman 1962 p 150). This is the reverse of the usual order of progression.

The active crater is 200 m below and NW of the central summit peak Pizzo o Sopra la Fossa (Fig 11.2), which provided the author with an excellent observation point. This peak is part of the older andesitic volcano, Serra Vancori, the NW sector of which was downfaulted (Williams 1941) or blown out (Scrope 1872) and is being rebuilt by the present eruptions. Although recent activity has included 8 periods of lava flow totalling about 200 days between 1949 and 1967

Fig. 11.1: The Island of Stromboli seen from the southwest. The village of Ginostra can be seen on the left. The active crater and the Sciara del Fuoco are beyond the left skyline.

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Fig 11.2: Topographic map of Stromboli, showing the seismic recording sites.

After Grainge, Standring and Edwards

(Silvestri 1967) and a major explosion in 1930 which hurled large boulders distances of 3 km (Rittman 1962, p 37), the term "Strombolian activity" has come to mean the ejection of incandescent scoria and lava fragments by mild explosions and gas jets. This was the type of activity the author observed and recorded on slow motion tape seismograph B, which was the only seismograph on the island at the time.

The seismograph commenced recording at site 1 (Fig 11.2) at 0030 IST on 3 July 1971, and recorded without interruption at tape speed 0.2 mm s^{-1} until 23 July 1971. The 1.2 Hz seismometer was buried in an abandoned Olive grove beside the path up the volcano. No radio time signals were recorded, but the crystal clock in the seismograph and an Accutron watch were compared by eye and ear, with the time signals on the Italian National Programme from time to time, thus allowing comparison of times of visible and recorded events to about 1 second accuracy. From 2028 IST on 23 July to 1400 IST on 24 July, tape speed was increased to 0.4 mm s^{-1} , and recordings were made at sites 1 to 5. Previous seismic recordings on Stromboli have been reported by Peterschmitt and Tazieff (1962) and Latter (1967).

Visual Observations

Between 1130 and 1445 on 3 July, 2055 and 2401 on 4 July, and 2227 and 2417 on 23 July, activity was observed from the summit observation point (Tables 11.1, 11.2, 11.3). The positions of the vents (Figs 11.3, 11.4) in the crater were substantially the same as described by Bullard (1962) and Grainge, Standring, and Edwards (1969), and have indeed remained similar since about 1600 AD (Bullard 1962) in spite of the major explosion of 1930, which lowered the level of the vents by between 50 and 100 m (Rittman 1962 p 130). No consistent names or symbols have been given to them however, and the author numbered them from south west (left) to north east (right), along the elongated crater, and described them on 3 and 4 July 1971 as follows:

Vent 1. A pit of about 20 m diameter containing a lava lake which glowed and emitted vapour continuously. About 5 times per hour (irregular) powerful explosions would throw bombs between 100 and 250 m above the vent (Fig 11.5), followed by a secondary upsurge of lava within the pit. The irregular pulsing noise of gas discharge from the vent (80/minute) ceased for some seconds after each explosion, but otherwise activity

TABLE 11.1 Observations of Crater Activity and Seismic Recordings at Stromboli, 3 July 1971.

Time of Observation		Vent No.	P-time Site 1	Gas Pulses per minute	Ground Velocity Site 1	Height Bombs Rose	Description
h	m	s	IST h m s	IST m s	$\mu\text{m s}^{-1}$	m	
11	30			60ca			
11	33	10	1		20		Explosion
11	37	ca	1	78			Pulse duration 1/5 s ca
11	44	21ca	1		15		Explosion, many bombs.
11	46	06-31	4		7		Fountain.
11	50	ca	1	94			
11	50	16	4		12		Explosion
11	51	40ca	1		15		Photo 1-1/2 s after explosion.
11	54	16	4		11		Powerful roar.
11	55	ca	3 or 4	32			
12	00	08	1		14		Explosion.
12	02	06	1		30		Explosion obscured by smoke.

Time of Observation		Vent No.	P-time Site 1		Gas Pulses per minute	Ground Velocity Site 1	Height Bombs Rose	Description
h	m	s	h	m	s	μm s ⁻¹	m	
12	03	21-36	1					Slide within vent? heard but not seen.
12	04	48-69	4	12	04	36.5(e)	7	Fountain and roar.
12	11	48	2	12	11	47.5(e)	12	Sharp explosion and cool reddish ash cloud
12	19	49	1	12	19	47.5(i)	20	Explosion
12	30	46	1	12	30	43(i)	21	Explosion
12	35	21	1	12	35	20(i)	19	Explosion
12	51							
12	50	41	4			5		Noise increases
12	51	21	1	12	51	19.5(e)	20	Explosion
12	53	01	2		NR	7		Dull explosion. Ash cloud.
12	53	16	3 or 4		NR	?		Powerful roar.
13	01	51	1	13	01	49	25	Explosion. Felt vibration.
13	05	46	3	13	05	41.5	9	Roar; glow from vent.
13	24	21-36	3 or 4		NR	6		Roar; crater obscured.
13	32	25	1	13	32	22	30	Explosion.
13	37						90	Close direct sound.

Time of Observation			Vent No.	P-time Site 1		Gas Pulses per	Ground Velocity Site 1	Height Bombs Rose	Description
h	m	s		h	m	s	$\mu\text{m s}^{-1}$	m	
13	39		3 or 4				28		Distant echoing sound.
13	46	31	1	13	46	31			Explosion
13	53	00	3	13	52	55.5(e)	20		Roar, bombs erupted.
13	53	40	4		NR		10		Dull explosion
14	02	09	4		NR		5	200	Violent fountain
13	03	25	4		NR				Noisy
14	04	13	1	14	04	12(i)	17		Explosion
14	04	50ca	1	14	04	47.5	23	200	Explosion.
14	14	42	1?	14	14	40.5	16	200	Explosion
14	16	35	(1 (4	14	16	31.5	16		Explosion Fountain
14	25	ca	1	14	24	35	21	200	Explosion.Gas pulse noises ceased for about 10 s.
14	36	37	4		NR				Brief roar
14	40	ca	1	14	39	25	40	200	Explosion. Gas pulse noises ceased briefly.
14	41	05-25	4	14	40	49.5	10	200	Fountain.
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TABLE 11.2

Observations of Crater Activity and Seismic

Recordings at Stromboli, 4 July 1971

Time of Observation	Vent No.	P-time Site 1		Gas Pulses per minute	Ground Velocity Site 1 $\mu\text{m s}^{-1}$	Description
		IST h m s	IST h m s			
20 54 ca	1				15	Explosion
21 07 20	3	21 07	17.5		26	Explosion, causes partial collapse of spatter cone, followed by increased spatter rate and noise.
21 13						Audio tape recording begins
21 13 17	3	21 13	20.5		13	Explosion
21 21 ca	1				21	Explosion.
21 35 12ca	1	21 35	08		30	Explosion.
21 37 22	4		NR		6	Fountain.
21 45				60		Noise pulses correlate well with ejection of spatter.
21 50 ca						Audio recording 2nd side begins.
21 55 14	1	21 55	15(i)		21	Explosion.
21 56 ca	4					Fountain.

Time of Observation	Vent No.	P-time Site 1		Gas Pulses per minute	Ground Velocity Site 1 $\mu\text{m s}^{-1}$	Description
		IST h m s	IST h m s			
22 02	3			54		
22 12 47	1	22 12	46.5 (i)		15	Explosion
22 17 37	1	22 17	37.5		19	Explosion.
22 21 05	1	22 21	02.5 (i)		13	Explosion.
22 22 52- 23 12	4	NR			5	Fountain
22 35 02ca	1	22 34	57.5 (i)		22	Explosion
22 42 07	2	NR			7	Quiet explosion. Dark ash followed by incandescent bombs.
22 43 22	1	22 43	21		19	Explosion.
22 44 27-52	4	NR			5	Fountain
22 51 42	3	22 51	40.5		8	Explosion.
23 02 ca	1				27	Explosion.
23 14±						3 Telephotos of vents.
23 21 52- 22 17	4	NR			7	Fountain from 2 vents.
23 22 46ca	1	23 22	39.5 (i)		12	Explosion

Time of Observation	Vent No.	P-time Site 1	Gas Pulses per	Ground Velocity Site 1	Description
IST h m s		IST h m s	minute	$\mu\text{m s}^{-1}$	
23 34	1			14	Explosion
23 39 ca	4			5	Fountain
23 39.8ca	1	23 39 39.5(i)		32	Explosion. Lava fingers surge up in vent some seconds after explosion.
23 44 ca	1		81		Low intensity pulses.
23 44 51	2	22 44 49.5(e?)		7	Quiet explosion, cool ash, then hot bombs.
23 46 11-46	4	NR		8	Fountain
23 47 16	3	23 47 10.5		10	Explosion
23 49 ca	3		72		Moderate intensity.
23 53 21ca	2	23 53 18		9	Quiet explosion, mostly hot.
23 53 46	3	23 53 47.5(e)		10	Explosion.
23 54 16	1	23 54 16.5		16	Explosion.
23 59 31	1	23 58 32.5(i)		22	Explosion followed by more intense pulsing noise.

TABLE 11.3:

Observations of Crater Activity, and Seismic

Recordings at Stromboli, 23 - 24 July 1971.

Time of Observation	Vent No.	P-time Site 1		Gas Pulses per minute	Height Bombs Rose	Description
		h	m s			
IST						
h m s						
22 22 3/4-23	4					Violent fountain.
22 32.5	1					Gas pulsing ceases.
22 33 $\frac{1}{4}$	1					Gas pulsing starts.
22 34 39 $\frac{1}{2}$ -47	1	22	34 39(e)			Fountain
22 36 ca	1					Gas pulsing ceases.
22 37 13	1	22	37 09.5		300	Explosion plus ash
22 37 56	1		NR			Gas pulsing resumes with medium eruption of bombs.
22 41.5	1					Gas pulsing ceases.
22 42 26	2	22	42 27			Brief fountain.
22 45 51	2	22	45 52(e)			Moderately large fountain.
?	3					Only vent active.
23 00 27	1	23	00 29.5			Large but quiet explosion ejects bombs and ash.

Time of Observation	Vent No.	P-time Site 1	Gas Pulses per minute	Height Bombs Rose	Description
h m s		IST h m s		m	
23 01.5	1				Gas pulsing starts, Lava well down in vent.
23 02 58ca	3	23 02 57(e)		100	Explosion
23 06-07	1		41		Ordinary gas pulsing
23 09-11	3		48		Ordinary gas pulsing
23 11 16-28	4	23 11 22.5(e!)		225	Fountain
23 11 59ca	1	23 11 58(i)			Very brief fountain
23 14 39	2	23 14 40(i)			Small very brief fountain.
23 17 00-13	4	23 17 01(e)		200	Noisy fountain, lava con- tent very variable.
23 18 31-39	1	23 18 29.5		200	Noisy fountain
23.5 ca	3				Lava level much higher than in other vents.
23 31 39ca-52	1	23 31 34(e)		250	Fountain, directed slightly NE.
23 40 05	1	23 40 02.5			Fountain starts.
23 40 06-11	1	23 40 46.5(e)		100	Fountain continues.

Time of Observation	Vent No.	P-time Site 1	Gas Pulses per minute	Height Bombs Rose	Description
IST h m s		IST h m s		m	
23 40 45-46	4	23 45 10(e)		200	Noisy fountain.
23 41.5	1	23 49 18			No gas pulsing
23 45 16-26ca	4			300	Noisy fountain.
23 48.5	1			>200	Fountain
23 49 21ca	3			100	Explosion
24 July 1971					
00 08ca	1			200	Fountain
00 12ca	1			250	Fountain

Fig 11.3: View of the SW half of the crater of Stromboli at 13 32.3 IST on 3 July 1971, the instant of an explosion of vent 1 (centre). Vent 2 (right) is marked by a patch of steam.

Fig 11.4: View of the NE half of the crater of Stromboli, showing vent 2 steaming at the extreme left and vent 3 (centre left) which at this stage consisted of two pipe-like orifices. Vent 4, further right and closer to the camera, is hidden by the crater rim.

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Fig 11.5: A time exposure of an explosion of vent 1
at 2121 IST on 4 July 1971.

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was constant, with frequent small bombs cast within the pit and no warning of the large explosions. The audible pulsing was strong and clear on 3 July, but quiet on 4 July - hardly audible over the noise of Vent 3.

Vent 2. A shallow pit of cinders which erupted about once per hour, once with a sharp bang, but usually with very little noise. A mushroom cloud of reddish ash arose, partly obscuring the glowing bombs which followed.

Vent 3. On 3 July this consisted of 2 small steep sided spatter cones like pipes projecting from the sloping back wall of the crater area. These blew smoke rings and occasionally ejected lava and made muffled reverberatory noises about 30 times/minute. On 4 July they had transformed into one rapidly growing cinder cone which emitted vigorous pulsing sounds and spatter at an average rate of 70 - 90/minute, and produced about 2 explosions per hour.

Vent 4. This was partially obscured by the crater wall, but was characterised by fire fountaining with much gas and little lava. On 3 July the fountains were very vigorous and spectacular,

reaching 200 m in height and playing for 10 - 30 second. On 4 July they were weak, and occurred simultaneously from several points many metres apart. An average of 2.5 fountains and 0.3 explosions per hour were produced.

The significance of the change in vent 3 between the observations of 3 and 4 July became apparent when it was observed to explode at 21 h 07 m 41 s on 4 July, just after the new cone had been photographed (Fig 11.6). During the explosion the cone collapsed to a flatter shape (Fig 11.7), and the noise of gas pulsing and volume of ejecta increased markedly. However, when the author observed the crater again on 23 July the cone was not much larger than on 4 July although its rate of growth still appeared to be considerable. Destruction by explosions, or by collapse and engulfment is indicated, and considering the apparent constancy of Stromboli in shape, size and activity during historic time, collapse and engulfment is the more probable explanation.

Soon after the collapse and increase in eruption at vent 3, the audible noise of eruption was recorded at the summit on a Philips cassette recorder loaned by Mr. J. O'Meara. The recording between 2114 and 2220 on 4 July is interrupted between 2123 and 2130, and has an

Fig 11.6: View of vents 1, 2 and 3 (left to right) at 2059 IST on 4 July 1971. The two pipe-like orifices of vent 3 have disappeared, but the vent is still a fairly steep sided cone, with possibly a fissure down the front right side. View from Pizza o Sopra La Fossa.

Fig 11.7: View of vent 3 at 2118 IST, after an explosion at 2107.3 caused an apparent collapse of the cone to a flatter shape, and a further enlargement of the vent. Photo from the same point on Pizzo o Sopra La Fossa as for Fig 11.6.

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uncalibrated change in gain at this time, but is otherwise excellent.

On 23 - 24 July 1971, activity was rather less explosive than on 3 and 4 July. Vent 1 produced about 1 muffled explosion and 4 fountains per hour, vent 2 produced 2 minor fountains per hour and no explosions, vent 3 produced one moderate explosion per hour, and ejected spatter continuously, and vent 4 produced about 2.5 large fountains per hour. Although activity was less sharp and eventful, bombs were thrown to greater heights (300 m) than on 3 - 4 July.

Volcano-Seismic Recordings

VOLCANIC TREMOR

Volcanic tremor was essentially continuous over the recording period. At site 1, 1.62 km from the crater, the average particle velocity was 3 micrometre/second, and the average seismic power was 1 kW (Fig 11.8). Variations occurred in two time ranges. Over periods between 0.2 and 20 minutes, variations from 0.1 to 6 kW occur, giving an appearance of grass on the record. Over periods greater

Fig 11.8: Chart of seismic power from Stromboli, recorded at site 1 between 2100 ($\equiv$ 0 hours) and 2350 IST on 4 July 1971. Observed explosions and high fountains are labelled with the vent number. Tremor increased during the explosion of vent 3 at 2107 IST. Fountains rarely exceeded continuous tremor in seismic power.

than an hour the range of smoothed power is about 0.5 to 2 kW, with some uncertainty about the lower limit due to periods of spurious low level probably caused by particles of oxide from the edge of the tape temporarily lodging on the tape head. Most of these can be recognised by their absence from the low sensitivity channel.

At the time of the collapse of vent 4 and increase in audible noise and eruption, the seismic power of tremor jumped 8 db from 1 kW to 6 kW (Fig 11.8), and by 2220, the end of the audio tape recording, had fallen back to 1 kW again. The audible power level was compared with the seismic power to see if they correlated in time. No clear parallel downward trend in the audio power level is evident, but on examining the seismic power recording in detail, it is seen that the power had fallen 3 db before the audio recording began, and it fell a further 3 db between 2123 and 2130 during which the gain change was made, leaving only 2 db to be correlated, which is less than the likely errors.

EXPLOSION EARTHQUAKES

Superimposed on the tremor were the vibrations of explosions and fountains at the vents. Those which were visually observed ranged in seismic power from 1 to 90 kW. They threw bombs to heights of 300 m, and the most powerful explosions were felt at the summit observation point. Even more powerful explosion earthquakes were recorded but not visually observed on 15 July at 17 hours (300 kW) and 17 July at 07 hours (350 kW). These had a greater proportion of high frequencies than the average. The duration of large amplitudes was about 10 s, and the total seismic energy was about 10^6 J.

The frequency of occurrence of explosion earthquakes did not vary greatly during the recording period. Counts during the first 2 hours of each day are as follows:

3 July	15	12 July	16	21 July	17
4	16	13	18	22	15
5	16	14	17	23	16
6	14	15	21	24	23
7	23	16	14		
8	12	17	26		
9	13	18	26		
10	19	19	16		
11	16	20	20		

In many sections of the record the explosion earthquakes seem to be periodic, but the pattern changes often and is very imperfect.

The percentage of earthquakes other than explosion earthquakes was extremely low during the recording period. During a total of 8 hours visual observations, only four seismic events were recorded for which there are no noted visual observations. All appeared to be explosion earthquakes, and occurred at times when the author was unable to take notes. Rather than confirming Latter's (1967) observation that only 70% of shocks could be correlated with explosions, the author found that some quite impressive gas fountains from vent 4 caused no increase in vibration level. In particular the fountain at 14 h 02 m 09 s IST on 4 July was audibly noisy, and threw (or blew) bombs 200 m above the vent, but had seismic power less than the tremor level of 1 kW.

The premonitory shocks which Peterschmitt and Tazieff (1962) observed to precede the explosion earthquakes and gas jet vibrations by between 12 and 44 s were not observed. Commencement times of visible activity at the summit usually agreed with seismic arrival times at site 1 (Tables 11.1, 11.2, 11.3) within 3 s, the average difference "audible onset time - seismic onset time" being 1.0 s. This compares favourably with the calculated time for sources in the vents at slant distances 340 m in air from the observation point (travel time 1.0 s) and 1710 m in rock from seismograph site 1 (travel time 0.4 s at nominal velocity 4 km s^{-1}).

POWER SPECTRA

Spectra at Stromboli were studied under four headings:

- (i) Long term trends at recording site 1.
- (ii) Rapid changes at recording site 1.
- (iii) Comparison between seismic spectrum and audibly recorded gas pulse spectrum.
- (iv) Variations between recording sites 1 to 5.

(i) Long Term spectral variations

To investigate long term trends, the seismic tape recording between 0000 and 0020 IST each day was analysed with a Sona-graph machine. Spectral sections with 25 degrees of freedom, and a 3 db resolution bandwidth of 0.1 Hz were made, and redrafted as power spectrum versus date (Fig 11.9). The same minor peaks and troughs are present on all the spectra, but the broad shape changes with time.

Early in July the dominant peak was at 1.6 Hz and the smoothed curve of decreasing spectral densities above 1.6 Hz was roughly linear. Gradually but somewhat erratically the peak broadened, and shifted upwards, reaching 2.8 Hz by the middle of July, giving a broad

Fig 11.9: Profiles of seismic power spectral density versus date, made between 0000 and 0020 IST each day from recordings at site 1, Stromboli. The parallel lines mark the consistent narrow peaks in the spectrum, and the heavy dashed line traces the shift of the dominant frequency. The dotted flanks connect each profile to its date line cum zero db line. Frequency resolution 0.2 Hz, stability 1.9 db with 90% confidence, 0 db $\equiv$ 0.6 W Hz⁻¹. A response curve of the seismograph to constant ground velocity is also shown.

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Date, July 1971

rounded shape of spectrum with the minor peaks and troughs superimposed on it. Then the time trend reversed and by 23 July when the author returned to Stromboli, both spectrum and visual activity were similar to when he left on 5 July.

(ii) Short term spectral variations

A rapid spectral change similar to the long term one was observed at 2107.3 IST on 4 July, when vent 3 partially collapsed, and the noise and rate of eruption of spatter increased. The sonagram for this event (Fig 11.10) has spectral sections with 25 degrees of freedom which show that the dominant frequency of tremor shifted from 1.6 to 3 Hz at the time of the event. Each spectral section represents the average spectral density over the preceding 3 minutes of tremor.

Between 2030 and 2053 IST, just prior to the author's arrival at the observation point, the seismic spectrum had a broad peak between 1.7 and 3 Hz, and seismic power was decreasing from 2 to 0.6 kW. At 2053, about the time of an explosion of vent 1, spectral densities increased in a band about 2 Hz wide, and centred on 1.6 Hz, and gave the spectrum a saw tooth shape, with peak at 1.6 Hz. At the same time the

Fig 11.10:

(a) Sonagram of tremor and explosion earthquakes at site 1, Stromboli, on 4 July 1971. The change at 12 minutes (2107.3 IST) accompanied a change in shape and activity of vent 3 during an explosion.

(b) Profiles of seismic power spectral density corresponding to (a). Frequency resolution 0.2 Hz, stability 1.9 db with 90% confidence, 0 db $\equiv$ 0.6 W Hz⁻¹.

Fig 11.11: Spectrum of variations in the level of audible noise from vent 3, recorded at Pizzo o Sopra Fossa. Frequency resolution 0.1 Hz, stability 1.9 db with 90% confidence.

seismic power recovered to 1 - 1.6 kW. The next change was at 2107.3 IST when vent 3 partially collapsed. Immediately, spectral densities increased at all frequencies, and especially between 3 and 4 Hz where the increase was 10 db. At 1.6 Hz the increase was less than 2 db, and as a result, the dominant frequency became 3 Hz for the next 20 minutes. During this time the seismic power was between 2 and 8 db above average. After 2200 IST the spectra and seismic power were closely similar to those before 2050.

Interpretation of this event is complicated by the increase in spectral density at 1.6 Hz prior to the collapse of the vent 3 cone. Possibly this was a separate spectral change connected with the explosion and subsequent activity of vent 1, but it is also possible that it was a premonitory change in vent 3, and that the partial collapse of vent 3 had no causal relationship with the increased tremor and eruption, or with the spectral changes. What is certain is that increased activity at vent 3 is a possible explanation for the long term spectral change (Fig 11.10c).

(iii) Comparison between seismic and audible gas pulse spectra

The apparent relation between seismic power of tremor, and vigour of spatter eruption, and the obvious source of seismic energy in gas pulsing, encouraged a closer look at the rough agreement between rate of gas pulsing and the tremor spectrum. The sound of gas pulses resembles the discharge of a safety valve. It appears that the gas bubbles in the lava are vertically elongated, or are in close spaced trains, and that most discharges end prematurely when the orifice closes at the surface. Each gas pulse is imagined to cause a downward force on the ground as the gas escapes, after the manner of a rocket partly embedded downwards in the ground. Spatter contributes some extra seismic energy when it lands.

The spectrum of the downward force is controlled by the statistical variations in duration and shape of each gas pulse and in the intervals between them and can be obtained approximately by analysing the envelope of intensity of audible sound which the pulses produce. The cassette recording of the sound, was played into a rectifying and smoothing network, and the very low frequency level variations re-recorded on a slow motion tape recorder. This tape was played at 1000 x recording

speed into the Sona-graph machine. The spectrum (Fig 11.11) has a single peak at 0.5 Hz, and a nearly linear decrease of density (db) with frequency above and below the peak. As would be expected from the hypothetical mechanism there appears to be a considerable spectral density at zero frequency, which corresponds to the average downward force on the ground.

The spectra of particle velocity of seismic P-waves can be calculated from the spectra of pressure following Blake (1952):

$$V_{\omega} = \frac{P_{\omega} K}{\rho c^2} \frac{\omega a^3}{r} \left[\frac{\left(\frac{\omega}{c}\right)^2 + \left(\frac{1}{r}\right)^2}{1 + (1-2K) \left(\frac{\omega a}{c}\right)^2 + K^2 \left(\frac{\omega a}{c}\right)^4} \right]^{\frac{1}{2}}$$

- Where
- V_{ω} = peak particle velocity at angular velocity ω
  - P_{ω} = Amplitude of radial stress at surface of equivalent radiator (at angular velocity ω .)
  - K = $(1 - \sigma) / 2 (1 - 2\sigma)$
  - σ = Poisson's ratio
  - ρ = density
  - c = P wave velocity

a = radius of equivalent radiator

r = distance from centre of radiator.

To find the relation between V_ω and P_ω with varying frequency we adopt approximate values as follows:

ω = $2\pi f$ Taking $1 < f < 10$

c = 2000 m s^{-1}

r = 1620 m

σ = 0.4

K = 1.5

a = 1 m

The term in large brackets then approximates $0.003f$, and we have:

$$V_f \propto P_f f^2 \quad \text{where } V_f \text{ and } P_f \text{ are equivalent to } V_\omega \text{ and } P_\omega.$$

Applying the f^2 term as a correction to P_f in db (set to zero for $f = 1 \text{ Hz}$), we have the calculated spectrum of ground velocity $V_f(\text{calc})$. This is compared with the observed seismic spectrum of ground velocity $V_f(\text{observed})$ which has been corrected for seismograph response.

f	0.25	0.5	1	2	4	6	8 Hz
P_f	27	32	28	21	13	6	1 db (arbitrary level)
Correct- ion	-24	-12	0	12	24	31	36 db
V_f (calc)	3	20	28	33	37	37	37 db (arbitrary level)
V_f (observed)		20	22	33.5	36	25	21 db (arbitrary level)

The observed and calculated spectral levels agree well except at the highest frequencies where the observed values are lower. This can be explained by absorption, but the data are insufficient to lift the model out of the working hypothesis class.

(iv) Variation of seismic power and power spectra of tremor with distance.

Seismic power was calculated at sites 1, 2, 3, and 5 which cover a range of distance from the crater of 1.62 to 2.53 km. Because only one seismograph was taken to Stromboli it was not possible to compare power or spectra at different sites simultaneously.

To obtain a comparison between sites 1 to 5 it is necessary to assume that tremor did not vary over the 4 hour survey period. The seismic power at each site (except 4 where the seismometer mass rested on its stop due to the heat) is given in Table 11.4.

TABLE 11.4: Seismic Power of Tremor and Explosion Earthquakes Measured at Various Distances from the Crater on 24 July 1971.

Site No	Distance km	Seismic Power kW			
		Tremor		Maximum Explosion	
		Min	Max	Typical	Earthquake
1	1.62	0.4	4	1	160
2	1.91	0.37	0.92	0.6	37
3	2.00	0.25	0.4	0.33	6.3
5	2.53	0.25	0.8	0.4	16

It appears that the indicated power decreases as distance increases at Stromboli, as would be the case if ground velocity was inversely proportional to distance to a higher power than the first. This could be due to

absorption, or refraction propagation or to the wave motion tending towards horizontal, or it could be a real change of power with time. In any case the error would be a minimum at the recording site nearest the source, station 1.

Spectral profiles of volcanic tremor at sites 1 to 5 on 24 July 1971 are shown in Fig 11.12. The effect of the site on the spectra is very obvious. The dominant frequency ranges from 3.7 Hz to 0.9 Hz or less over a distance of only half a kilometre, and the peaks which were so consistent at site 1 are not present at the other sites. In view of this, the overall distance range of recordings was barely great enough to allow the absorption and Q value to be determined. The average spectral slope (by eye) above 3 Hz was $1.8 \text{ db Hz}^{-1} \text{ km}^{-1}$ at sites 1, 2, and 5, whereas the source spectra, calculated from the audible gas pulse spectra on 4 July, was apparently flat between 4 and 8 Hz. Assuming it was also flat between 3 and 7 Hz on 24 July 1971, and that the seismic velocity was 1.5 km s^{-1} (as at Hawaii) then from equations 3.51 and 3.52:

$$\alpha\lambda = c\eta = 0.31$$

$$Q = 10$$

This value has a large indeterminate uncertainty.

Fig 11.12: Profiles of seismic power spectral density of tremor on 24 July 1971, recorded at various distances from the crater of Stromboli. Frequency resolution 0.2 Hz, stability 1.9 db with 90% confidence, 0 db $\equiv$ 1 W Hz⁻¹ at 1.62 km, 1.91 and 2.00 km, $\equiv$ 0.6 W Hz⁻¹ at 2.53 km. A response curve for constant ground velocity is shown.

CHAPTER 12

COMPARISONS BETWEEN RUAPEHU, WHITE ISLAND
KILAUEA AND STROMBOLI VOLCANOES.

Morphology and Petrology

Ruapehu, White Island, and Stromboli volcanoes are composite cones, built of interstratified layers of lava and tephra. Their flanks slope at between 20 and 30°. Kilauea is a shield volcano made up almost entirely of flows of very fluid lava. As a result its flanks slope at only 2 - 3° except at the coast where they have been steepened by faulting. With the exception of Stromboli the slopes are clearly related to the chemistry of the rocks. The labradorite andesite of Ruapehu has hypersthene as its dominant heavy mineral and contains 60.1% silica (Clark 1960). The augite-hypersthene labradorite andesites of White Island with 60 - 62.5% silica (Black 1970) are not too different. The recent lavas at Stromboli are Olivine trachybasalts with only 49.8% silica (Abbruzzese and Cavallaro 1956), much closer to the olivine labradorite basalts of Kilauea with 48.35% silica (Macdonald and Katsura 1964), than to the steeper volcanoes it morphologically resembles.

The Variation Patterns in Volcanic Tremor.

These will be discussed in relation to the patterns originally observed at Ruapehu.

Continuous volcanic tremor. This was observed at all the volcanoes. At White Island it could at times be resolved into overlapping pulses resembling Minakami's B-type earthquakes of continuous train. At Ruapehu, White Island, and Mauna Ulu, tremor sometimes appeared and disappeared gradually. At Stromboli tremor was practically continuous.

Sudden stoppages of strong tremor. These were observed at Ruapehu, White Island, and Mauna Ulu, but not at Stromboli. At Mauna Ulu they were sometimes triggered by an earthquake or rockfall. Often weak tremor persisted during the stoppage, but at Ruapehu and Mauna Ulu, apparently complete stoppages were also observed. Durations of sudden stoppages ranged from about a minute up to about an hour at each volcano.

Powerful recommencements of tremor. These also occurred at Ruapehu, White Island and Mauna Ulu. At Stromboli, sudden temporary increases in tremor sometimes occurred during explosion earthquakes, and these may be a related phenomena. At Ruapehu, powerful recommencements

resemble B-type earthquakes.

At White Island the powerful recommencements were immediately followed by a very high frequency vibration apparently produced at the vent by a pulse-like increase in the ash eruption. At Mauna Ulu, powerful recommencements were usually emergent and corresponded to the onset of degassing of the lava lake. The few impulsive ones appear to be triggered by rockfall earthquakes.

Pulsing tremor. This was observed at Ruapehu, White Island, and Mauna Ulu. At Ruapehu there were two types: (i) a type which persisted for many days and was of low intensity (as in July 1964). The pulses became larger and less frequent as it was stopping. (ii) A high intensity type of several hours duration, such as followed the last eruption (9 June) of the 1968 series. This comprised tremor with brief stoppages at decreasing intervals of time. At White Island, 22 brief pulses in 3 hours led up to one of the last seismometrically identifiable outbursts of ash in the 1968 eruption of Rudolf. At Mauna Ulu, pulsing tremor (1 to 5 pulses per hour) corresponded with cyclic inflation and degassing of the lava lake, and occurred in groups of pulses separated by periods of continuous tremor and degassing.

Volcanic Earthquakes

A-type volcanic earthquakes. These were recorded at all four volcanoes, and were most common at Kilauea, where by definition, all earthquakes were volcanic. Typically about 15 earthquakes with dominant frequency above 4 Hz were recorded at the Puolani site each day. At White Island an average of about 3 very local A-type earthquakes occurred per day, often in brief intense swarms. At the other 2 volcanoes, local A-type earthquakes were not noticeably more frequent than is usual in active tectonic regions. Only the brief swarm at Tongariro volcano on 13 July 1969 and the weakly recorded initial shocks of the 22 June 1969 eruption of Ruapehu could be classified as volcanic with any confidence.

B-type earthquakes. At Ruapehu, events closely resembling those illustrated by Minakami (1961,1969) occurred occasionally, and during high speed playback, were heard as a single low frequency knock, as on wood. Many of the powerful recommencements of tremor at Ruapehu began with similar sound, and at Mauna Ulu, earthquakes

accompanying the switching in direction of lava flow on 11 February 1971, and some of the events interpreted as rock falls into the lava lake sounded similar during high speed playback. No sufficiently similar sounds were heard on the tape seismograms from White Island. Rather formless pulses in the tremor, and small impulsive earthquakes with no apparent S phase occur instead. At Stromboli B-type earthquakes were not observed.

Explosion earthquakes. In the narrow sense of the term these were positively identified only at Stromboli. There, they differed from the B-type earthquakes of Ruapehu and Kilauea in having a greater proportion of high frequencies, and a coda of rising frequency, due either to dispersion or perhaps to coupling of the air wave to the ground. The earthquake accompanying the brief and silent ash eruption at White Island on 20 Nov 1966 also had a coda of rising frequency, but if it was an explosion earthquake it was deep down the vent, and neither of the above explanations would apply. At Ruapehu the earthquake at 0025 on 22 June 1969 was probably an explosion earthquake because Chief Ranger J. Mazey reported a shock wave. No explosion earthquakes were recognised at Kilauea.

Eruption vibrations. These are necessarily defined differently on different volcanoes. At Ruapehu they were typically emergent, and symmetrical about their mid point in the time direction, and were usually of about 1 minute duration, although that on 22 June 1969 was continuous for 6 minute. The accompanying eruptions look like explosions but are of longer duration, and are not audible at any distance. These characteristics may be caused by the inertia of the crater lake. It was notable in 1968 that impulsive re-commencements of tremor graded through hybrid types into eruption vibrations, as though all were vibration sources of explosion type, but that they gradually broke through into the lake and became seismically muffled. At White Island and Mauna Ulu eruption vibrations were powerful commencements of tremor and continuing tremor, and were related to the eruption of gas carrying some ash or spatter. At Stromboli the eruption vibrations were explosion earthquakes, and tremor.

Seismic Power of Volcanic Vibrations

A comparison of seismic power at the four volcanoes is made in Table 12.1. For volcanic tremor, the values of typical, minimum, maximum over 1 hour, maximum powerful recommencement, and overall maximum power of tremor are compared. For vibrations accompanying proven eruptions, the maximum seismic powers, rates of eruption, and the ratio of these - the seismic energy of eruption vibrations per unit volume of erupted material, were compared. The rates of eruption were estimated as follows:

(i) Ruapehu. In the eruption on 22 June 1969, the ejecta totalled $1.4 \times 10^5 \text{ m}^3$, and was erupted in about 7 minutes. This is a rate of $3 \times 10^3 \text{ m}^3 \text{ s}^{-1}$.

(ii) White Island. From a sample of the ash cloud collected at Rudolf vent on 29 April 1968 at the onset of a pulse in the eruption, Duncan and Vucetich (1970) estimate 54 kg s^{-1} of tephra of density 1.7 Mg m^{-3} was being erupted. This is a volume rate of $0.03 \text{ m}^3 \text{ s}^{-1}$.

Seismic recordings began 20 minutes later while the eruption was still in progress. The power of tremor was then peaking to 3 kW.

TABLE 12.1: Comparison between Seismic Power and Eruption Rate;
Ruapehu, White Island, Mauna Ulu and Stromboli.

	Ruapehu	White Is.	Mauna Ulu	Stromboli
Typical Power of tremor	kW 0.1-1	0.01-0.03	0.1-1	1
Date of Observation above	Apr. 68	Feb. 67	Mar. 71	Jul. 71
Minimum power of tremor	kW 0.01	10^{-4}	0.03	0.3
Date of observation above			25. 3.71	
Maximum power of tremor over a 1 hour period.	kW 20	3	2.5	4
Date of observation above	27. 4.68	5. 5.68	25. 2.71	22. 7.71
Maximum power in powerful recommencement of tremor	kW 230	10	30	
Date of observation above	1. 5.64	2. 5.68	25. 2.71	
Maximum power in volcano seismic events.	kW 9000	12	4000	350
Date of observation above	8. 7.69	3. 6.68	18. 2.71	17. 7.71
Maximum power of proven eruption vibrations, kW	kW 7500	3	30	90
Date of observation above	22. 6.69	29. 4.68	25. 2.71	4. 7.71

	Ruapehu	White Is.	Mauna Ulu	Stromboli
Rate of eruption at time of max power	$\text{m}^3 \text{s}^{-1}$ 3 x 10^3	0.03	3	0.4
Date of observation above	22. 6.69	2. 5.68	Feb 71.	4. 7.71
Seis Energy per unit volume erupted.	J m^{-3} 3 x 10^3	10^5	10^4	2×10^5

(iii) Mauna Ulu. Swanson's (1972) steady rate of $9 \times 10^6 \text{ m}^3$ per month has been used.

(iv) Stromboli. The volume erupted in a large explosion of vent 1 was estimated from a photograph (Fig 11.5 field of view 40° , distance of vent 340 m approx). The radius of cross section of the solidly illuminated hemisphere above the vent is approximately 20 m. If the fragments had all been stopped and held at the boundary of solid illumination, they would at least be equivalent in volume to half a solid ring of radius 20 m, and of cross sectional area equal to the average cross sectional area of a lava bomb. Estimating the average diameter of these as 0.2 m, the volume of ejecta from the explosion is found to be 2 m^3 .

An independent value was obtained by counting bomb trajectories over a sample area of the photograph. The total number of bombs was estimated as 1000. If the average diameter is 0.2m, the total volume is 4 m^3 . This value was adopted. The duration of eruption was taken to be the approximate duration of large amplitudes on the seismogram (10 second).

It is seen from the table, that Ruapehu was seismically by far the most powerful volcano during the observation period. White Island was the least powerful volcano in almost all respects. The values of seismic energy of eruption vibration per unit volume of erupted material show that Mauna Ulu erupted relatively quietly, followed by Ruapehu and White Island, and with Stromboli erupting with the greatest seismic noise per m^3 . The ranges of values at Mauna Ulu and Ruapehu were very large during the period of observation. For short periods, Mauna Ulu erupted lava quite silently. Ruapehu showered observers 400 m away with mud when the eruption vibration was only 7 kW in seismic power, and left no trace when it was 9000 kW.

Frequency Spectrum of Tremor

Several difficulties stand in the way of accurate comparisons of the power spectra at the four volcanoes.

Firstly, the spectral shapes varied with time and with the stage of volcanic activity. This variability has been minimised but not eliminated by comparing only the spectra of strong tremor which accompanied vigorous eruptions.

Secondly, the spectrum of tremor has been shown to vary with distance. This has been minimised by using recordings made over the smallest available range of distances 0.7 to 1.6 km.

Thirdly, the recording and analysis equipment has of necessity changed considerably with time. Differences in the analysis equipment have been eliminated by making a few new analyses, on the PEL Sona-graph machine and Philips EL 3534 playback equipment, of periods of tremor selected as the best solutions of the first two difficulties. Differences in the seismic recorders and seismometers remain. The Willmore Mk I seismometer is not sufficiently stable to allow spectral shapes to be compared below 1.3 Hz. It was ascertained that the peak in the velocity response curve could reach 2 db above the value determined by calibration in Hawaii, if the seismometer was appreciably colder than the temperature at which it was adjusted. At White Island, Kilauea and Stromboli, the opposite occurred, while at Ruapehu, the recordings selected for comparison were made with a Willmore Mk II seismometer ($T_0 = 2$ s, $h = 0.8$) which was quite reliable. All the recorders had a nominally flat response; that is, within about 4 db

between 0.5 and 10 Hz. Details of the recordings selected for comparison are given in Table 12.2. The sonagrams are shown in Fig 12.1 and the spectral data are compared in Table 12.3.

TABLE 12.2: Recordings Selected for Comparison of Spectra at
 Ruapehu, White Island, Mauna Ulu and Stromboli.

Volcano	Date	Local Time	Distance Km	Seismograph	Activity
Ruapehu	27. 4.68	1230	0.7	Dome	Tremor preceding mud and ash eruption.
White Island	12. 2.68	0515	0.9	DSIR No. 4	Tremor accompanying eruption of crystalline tephra containing 30 - 50% fresh magmatic component.
Mauna Ulu	13. 4.71	1625	0.7	1970 B	Tremor accompanying vigorous degassing at east end of lava lake.
Stromboli	3. 7.71	1240	1.6	1970 B	Tremor accompanying noisy degassing of lava lake in vent 1.

Fig 12.1: Sonagrams of tremor during eruptions at each volcano, made under as nearly the same conditions (Table 12.2) as possible.

- a: Ruapehu, as tremor was commencing on 26 April 1968.
0 db $\equiv$ 4 W Hz⁻¹.
- b: White Island, on 12 February 1968. A distant earthquake begins at 14 minutes. 0 db $\equiv$ 40 mW Hz⁻¹.
- c: Mauna Ulu, as ebullition shifts to the east end of the lake at 1622 HT on 13 April 1968 ($\equiv$ 2 minutes on fig). 0 db $\equiv$ W Hz⁻¹.
- d: Stromboli, during normal activity on 3 July 1971.
Zero minutes on fig $\equiv$ 1230.4 IST. 0 db $\equiv$ 0.6 W Hz⁻¹.

501a

TABLE 12.3: Comparison of Spectra of Recordings Listed in Table 12.2

Volcano	Dominant Frequency Hz	Maximum Spectral Density W Hz^{-1}	Upper Roll Off Freq. Hz	Roll Off Rate $\text{db, km}^{-1} \text{Hz}^{-1}$	Seismometer Cut-off Freq. Hz
Ruapehu	2.0	4000 ca	3.0	3.6 ca	0.6 ca
White Island	1.5, 2.2	60 ca	2.2	2.2	1.2 ca
Mauna Ulu	1.3, 3.0	950	4.5	2.2	1.0
Stromboli	1.6, 2.9	210	2.9	2.0	1.1

It can be seen from Fig 12.1 and Table 12.3 that:

- 1/ Tremor is a broad band noise at each volcano.
- 2/ Above 4.5 Hz, spectral densities decrease fairly uniformly with increasing frequency at all volcanoes.
- 3/ There is no evidence of a rapid decrease in spectral density below 1 Hz at any volcano.
- 4/ The frequencies of maximum spectral density of seismic power, assuming tremor is propagated as body waves, appear to lie between 1.3 and 3.0 Hz at all volcanoes, but dominant peaks below 1.3 Hz cannot be excluded. For the assumption of propagation as Rayleigh waves, on a uniform half space, spectral densities should be divided by frequency to show the approximate shape of the spectra of seismic power. Then the maximum spectral densities would probably lie at or below the natural frequency of the seismometer, and be indeterminate from the author's data.
- 5/ The direct relation between dominant frequency of tremor and silica content of erupted lava given by Shimozuru (1961) is not evident in the present comparison. Stromboli has fairly high frequency tremor and low silica, and White Island has lower frequency tremor and considerably higher silica content.

- 6/ The values of maximum spectral density differ considerably between the volcanoes. This is partly the result of choosing examples on criteria other than the power level. The range of maximum observed seismic power is smaller than the range of spectral densities which were compared, but still Ruapehu was the noisiest and White Island the quietest volcano during the period of study.

Conclusion

The comparison of volcano-seismic phenomena at four volcanoes, the activity of which covers roughly half of the known range of eruption types, reveals no fundamental differences between them. At times of eruptions involving at least some juvenile ejecta, there is sufficient similarity in the variation patterns of seismic power and spectra to warrant the suggestion that volcanic tremor is generated in essentially the same way at all four volcanoes. At Mauna Ulu and Stromboli, it is very probable that tremor is caused by the observable release of gas bubbles

from the lava. This will occur most easily in the absence of containment pressures, but it is also possible in shallow chambers and conduits inside the volcano. The process cannot be rejected just because it cannot be seen, and it does not exclude resonance of contiguous bodies of rock or gas as agents in the shaping of the spectra. The assumption of separate predominantly gas-filled, liquid-filled, and solid-filled spaces within the volcano, as well as spaces containing mixtures such as vesiculating magma, allows internal activities, and goes far towards explaining the inconsistent correlations between seismic and eruptive phenomena.

Future Possibilities for Volcano-Seismic Research

This thesis would hardly be complete without consideration being given to future trends in volcano-seismic research. There is no doubt that further studies of eruption cycles at a variety of volcanoes are needed, in order to establish the norm and the range of behaviour patterns. These studies naturally form part of surveillance

programmes for the purpose of public warnings, and involve charting the variations in time and space of volcanic earthquakes and tremor. For this purpose, the measurement of seismic power and power spectra seem to the author to be essential. Monitoring of eruptive activity by time lapse photography, or infra red radiation meter (Shimozuru 1971) and microbarometer recording is also necessary as the studies at Ruapehu and Mauna Ulu revealed.

Efficient use of the data, and especially of the seismic spectra depends on the setting up and testing of models of each volcano, and here much remains to be done. The seismic velocity, absorption coefficient, density, and resistivity of live magma, and the variation of these quantities with depth are only approximately known, and will be difficult and expensive to obtain. Measurement in lava lakes such as Mauna Ulu offer the easiest starting point.

It has yet to be proved beyond doubt that volcanic tremor, powerful recommencements of tremor, B-type earthquakes, and explosion earthquakes have a common essential origin - the release of hot gas. First motion studies

and propagation studies can provide the answer provided high gain seismographs are sited close enough around the focus to give unequivocal directions of first motion.

APPENDIX A

DETAILS OF SPECTRAL VARIATIONS AT
PUU HULUHULU RECORDING SITE,
KILAUEA VOLCANO

Introduction

Sequences of sonagrams corresponding to the author's observations of activity at Mauna Ulu vent, and other periods of interest will be described, and figured in terms of the spectral density of seismic power.

2 February 1971: 04.8 - 19.0 HT

This sequence of sonagrams (Fig A.1) has stability 6 db and resolution 0.06 Hz, and shows tremor recorded at Puu Huluhulu during the voluminous but quiet eruption from the fissure on the south west flank of Mauna Ulu, 0.5 km from the tape seismograph. No visual observations of the lava lake correspond with the recording, but active fountaining and 'drainback' were reported on 1 and 3 Feb (D.W. Peterson, personal communication; drainback is interpreted to mean a short quick fall of the lake level).

Fig A.1: Sequence of sonagrams of tremor recorded at PHH between 04.8 and 19.0 HT on 2 February 1971. The time scale is accurate to 0.5 minute. Spectral lines appear while tremor is increasing, and disappear at the maximum. A partial reversal of peaks and troughs occurs as tremor decreases. Frequency resolution 0.05 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 2.5 W Hz⁻¹ ca.

The recording was selected as a good example of changes in tremor spectrum. The sequence begins when the seismic power of tremor was 0.08 kW, (Fig 10.6) and spectral features (i), (v) and (vi) were present.

There were spectral lines at 1.35 Hz, 0.8 and 0.6 Hz, and troughs at 0.7, 2.3, 3.9 and 5.0 Hz.

This spectral shape persisted through short pulses of tremor to 1 kW at 04.8 and 05.1 HT. At these times, spectral densities increased suddenly and simultaneously at all frequencies except 0.6 Hz, and decayed gradually. The line at 0.6 Hz disappeared for 15 minutes after the pulse at 05.1 HT. There followed a gradual but somewhat erratic increase in seismic power. At 06.4 HT, the spectral line at 0.6 Hz became the dominant frequency, but by 08.0 HT when seismic power levelled off at 1.6 kW, the line at 0.6 Hz had died away, and all other fine structure had disappeared behind a broad-band noise which was audibly uneven (lumpy) in the time domain, and which had spectral feature (i) with shallow troughs at 2.2 and 3.8 Hz. As seismic power decreased, spectral feature (ii) appeared, but with a secondary broad peak at 2 - 3 Hz where previously there had been a trough, and a new trough at 1.5 - 2 Hz.

The maximum power spectral densities were still between 0.8 and 1.5 Hz, but the peak was at 1.1 Hz, with a small trough at 1.2 Hz, Audibly this tremor was smooth.

At 08.9 HT, a pulse of tremor (2 kW) occurred with a different spectrum showing features (i) and (iii), and having a dominant frequency of 4.1 Hz. At 09.0 HT tremor almost ceased briefly, and then continued with dominant frequency 4.1 Hz.

6 February 1971, 08.8 HT

While listening to the tape recording from Puu Hulu-Hulu, a gliding frequency was heard to descend and rise in the tremor, and a sonagram of this was made (Fig A.2). The feature occurred near the end of some days of strong tremor with a dominant peak near 4 Hz. The eruption from the fissure on the west flank of Mauna Ulu was continuing and the lava lake was very active (D.W. Peterson, personal communication).

Until the time of 10 minutes shown on Figure A.2a, there was a spectral peak at 0.5 Hz. At 13 minutes a vague peak appeared gliding down through 0.65 Hz. As it reached 0.5 Hz its spectral density increased greatly. Spectral profiles with resolution 0.1 Hz show that the Q of the peak was then 5 or more. The peak rose again,

Fig A.2:

(a) Sonagram of tremor with gliding frequency, recorded at PHH on 6 February 1971 at 08.8 HT. The narrow peak glides from 0.65 Hz at 13 minutes to 0.5 Hz at 17 minutes and then up to 1.0 Hz after 30 minutes.

(b) Profiles of the above (shifted slightly left). The dominant frequency is 4.2 Hz. Frequency resolution 0.05 Hz, stability 6 db with 90% confidence, $0 \text{ db} \equiv 2.5 \text{ W Hz}^{-1} \text{ ca.}$

a

b

and at 30 minutes, appeared to level off at 1.0 Hz. Its progress was marked by higher densities where it crossed the fixed spectral lines. Where it crossed troughs it disappeared from the sonagram, but during audible playback its progress is more continuous.

A number of other diagonal and curving trends can be seen on the diagram. These may be of a statistical nature, and ought not to be confused with the gliding peak described above.

17 February 1971, 05.5 - 19.0 HT

This sequence of sonagrams (Fig A.3) has stability 3.2 db, and resolution 0.3 Hz, and shows tremor recorded at Puu Huluhulu while the other tape seismograph was recording at numerous different sites. The sequence was made to enable changes of spectra with time to be separated from changes with distance. It shows groups of pulses of tremor alternating with periods of continuous tremor. Lava was flowing steadily south east through tubes from Mauna Ulu at this time, and vigorous cyclic degassing was occurring in the lava lake.

During the first 2 pulses of tremor (8 kW), spectral feature (i) was dominant.

Fig A.3: Sequence of sonagrams of tremor, contoured in 6 db steps of seismic power spectral density. Recorded at PHH on 17 February at 05.5 - 10.9 HT. Short pulses and longer periods of tremor are occurring. Usually spectral densities are fairly uniform between 1 and 5 Hz, but sometimes (e.g. 0700 - 0750) there is a peak at 1.3 Hz. Frequency resolution 0.3 Hz, stability 2 db with 90% confidence, 0 db $\equiv$ 0.2 W Hz⁻¹. Profiles of spectral density at times and sites marked along the top of the sonagrams are shown on Fig 10.28.

The next two pulses (3 kW) at 06.8 HT had a combination of features (i) and (ii), and this shape continued through a succession of pulses separated by moderate tremor (0.3 kW) and part way through a period of low tremor (.02kW) which began at 08.0 HT. At 08.2 HT, feature (i) became dominant when densities increased above 1.5 Hz, and this shape continued through 3.2 hours of tremor which rose to a seismic power of 3 kW and declined, and then into the next group of pulses and stoppages. The third of these pulses (12.7 - 13.1 HT) saw a brief return of the combination of (i) and (ii), after which feature (ii) was absent.

18 February 1971, 19.4 - 21.7 HT

This sequence of sonagrams (Fig A.4) has stability 6 db and resolution 0.12 Hz, and shows spectra of tremor and rockfall earthquakes recorded at Puu Huluhulu during cyclic degassing of the lava lake. The activity was observed by the author in a party led by Dr. D.W. Peterson. The sequence begins during a stoppage of tremor and of ebullition in the lava lake. Tremor of 3 Hz began to reappear at 1934.3 HT. Then frequencies of 0.8 - 5 Hz increased suddenly at 1934.9, followed about 10 second later by frequencies above 5 Hz. Spectra feature (ii) was dominant, and there were minor

Fig A.4: Sequence of sonagrams of tremor and rockfall earthquakes recorded at PHH on 18 February 1971 from 19.4 - 21.7 HT. Seismic power spectral densities for tremor are approximately constant from 1 - 5 Hz, except at 1940, and during the major rockfall at 2032 and the powerful recommencement of tremor at 2128, when there was a dominant peak at 1.3 Hz. At 2128 densities increased first at 1.3 Hz. Frequency resolution 0.10 Hz, stability 6 db with 90% confidence, $0 \text{ db} \equiv 4 \text{ W Hz}^{-1}\text{ca.}$

secondary peaks at 4.7, 7, and 7.8 Hz, and a trough at 5 - 6.5 Hz. Violent ebullition of the lake began at the north west wall at about 1934.5 HT. At 19.8 HT, the spectral feature (i) became dominant when a minor rockfall from the middle of the north wall caused temporarily vigorous ebullition there. A trough developed at 2.3 Hz at the same time. Feature (i) continued through pulses of tremor (0.9 kW) separated by weak tremor (0.09 kW) which accompanied periods of ebullition at the east end, and at the north wall following a rockfall of about 5 m³ there at 2003 HT, but feature (ii) was dominant in the earthquake caused by the large rockfall of 12500 m³ from the south wall at 2032 HT, and during the following 4 minutes of reduced tremor (0.1 kW) when subdued surface bubbling occurred all over the lake. In the quiet period between 2036 and 2128 HT there was weak tremor with dominant frequency 2.5 - 4.5 Hz. In the powerful recommencement of tremor at 2128 HT, spectral densities increased first at 1 - 1.2 Hz, and progressively later at higher frequencies. Feature (ii) was initially dominant, but feature (i) became dominant as seismic power decreased again.

The trough at 2.3 Hz disappeared after 2128 HT. It also disappeared briefly at 2003 HT during a rockfall

similar to the one in which it first appeared, but during the rest of its life, after 2032 HT, it appeared to widen from 0.3 Hz to 1.5 Hz.

19 March 1971, 1130 to 1645 HT

This period of visual observations is covered by a sequence of sonagrams having profiles with stability 2 db and resolution 0.19 Hz (not reproduced here), and several single sonagrams of stability 6 db and resolution 0.06 Hz (Fig 10.33, 10.34). All are from recordings at Puu Huluhulu. For most of the observing period, moderate ebullition was occurring at the east end of the lava lake, and the lava flow through the tubes leading southeast from Mauna Ulu was continuing.

In the spectrum of the steady accompanying tremor of power 0.16 kW, features (i) and (vi) were dominant, but a line at 1.35 Hz with a Q of 13 (~ 3 db bandwidth 0.1 Hz) was the dominant frequency. There were weaker lines at 1.0 and 1.6 Hz, a broad peak at 3 Hz and a shallow trough at 2 Hz, but the higher stability/lower resolution sonagrams show nearly constant spectral densities of seismic power between 0.5 and 5 Hz. At 12.1 HT when moderate ebullition occurred briefly at the northwest wall of the lake, the peak at 1.35 Hz increased and feature (ii)

became dominant. At 13.4 HT, ebullition consisted of the bursting of large bubbles, at average rates over 4 separate minutes of 1.1 to 1.25 Hz, which approximates the dominant frequency, but does not explain the spectral lines.

At 1357 HT, a rockfall of about 1 m^3 fell into the east end of the lake, and the spectrum of the seism resembled the tremor spectrum. Ebullition stopped within 2 minute of the rockfall, and the spectral line at 1.3 Hz disappeared. Tremor decreased to 0.06 kW, and had a broad spectral peak between 2 and 5 Hz, and a weak line at 2.4 Hz. At 14.2 HT, when ebullition recommenced vigorously about 20 m from the northeast side of the lake, tremor with dominant spectral feature (i) increased rapidly, but for the first 2 minutes there were broad dominant peaks at 3.1 and 4.4 Hz, and a secondary peak at 8 Hz. This may have been the time taken for the ebullition centre to move to the northeast wall. Then lines appeared at 1.0, 1.3, and 1.6 Hz and the spectrum remained constant until about 1540 HT, when spectral densities began to increase smoothly, especially in the range 1.0 to 1.6 Hz. An additional and very narrow line at 1.2 Hz ($Q = 20$) and a trough at 3.5 Hz developed, and the flank below 1 Hz became very steep,

densities falling 12 db between 0.8 and 0.7 Hz. Ebullition at the east end of the lake became more and more vigorous, in parallel with the tremor, and very large bubbles were bursting at rates from 45 to 53 per minute or 0.82 Hz, which is on the steep cut-off slope of the spectrum. Then at 1610 HT about 2 m³ of the east wall collapsed into the lake, and within 4 minutes both tremor and ebullition returned to their pre-crescendo state (Fig 10.34).

13 April 1971. 16.0 - 17.0 HT

The sequence of sonagrams covering the visual observations of Mauna Ulu during this period (Fig A.5) is from a seismic recording at site 44 on the edge of Mauna Ulu vent. Tape seismograph (A) was used and has been calibrated, not by the Willmore bridge method, but by its internal system. As the author's seismometers were nominally identical, the calibration curves for the other seismograph (B) have been applied. The sonagram profiles have stability 6 db and resolution 0.1 Hz. The sequence begins during the stoppage of ebullition prior to 16.0 HT when tremor was very weak but had a low spectral peak at 1.3 Hz. This increased slightly in height during the minute preceding the recommencement at 1603 HT of tremor

Fig A.5: Sequence of sonagrams of tremor, recorded at site 44 beside Mauna Ulu vent on 13 April 1971 from 16.0 to 17.0 HT. Ebullition and tremor began at 1603, and changed at 1625 HT. The spectrum partially reversed at 1625 HT and lines developed as ebullition increased in vigour and size of bubbles. Frequency resolution 0.1 Hz, stability 6 db with 90% confidence, 0 db $\equiv$ 5 mW Hz⁻¹ ca. The spectral analysis of the 100 μ V and 1 mV calibration signals are shown.

and ebullition near the north side of the lava lake. Otherwise, tremor began simultaneously at all frequencies and spectral feature (ii) was dominant, and initially lines at 0.55 and 0.9 Hz, and a peak at 11.2 - 12 Hz were present. The first and last were nearly of equal spectral density to the dominant peak at 1.2 Hz, and the last had $Q = 21$ (3 db bandwidth 0.56 Hz). At 1625 HT the spectrum began to change, apparently when the ebullition site changed to the east end of the lake, and began to increase in vigour. Lines at 1.0, 1.4 and 1.6 Hz joined that at 1.2 Hz, forming a very steep sided dominant band, and an equally strong peak developed at 3.5 Hz. Fig A.5 shows that to a degree, the spectra was reversed relative to that prior to 1625 HT. Spectral densities at 3.5 Hz had been low prior to the development of the peak, and densities at 2.25 Hz had been high prior to the development of the trough at that frequency. The geometrical changes which accompanied the spectral change are:

- (i) Azimuth; Prior to 1625 HT the seismograph was on the rock side of the lava/rock boundary at which ebullition was occurring. After 1625 HT it was on the lava side.

(ii) Distance of the seismograph from the ebullition site was 50 m before 1625 HT, and 150 m afterwards.

In addition, ebullition of the lake was very spectacular after 1625. Bubbles about 4 m in diameter were bursting in the lake and lava was surging in large waves against the east end wall, without sign of their being reflected.

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