A STUDY OF THE SEISMIC AND VOLCANIC ACTIVITY OF MOUNT EREBUS, ANTARCTICA, 1981–1982

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Abstract: In order to monitor the long-term seismicity of Mount Erebus, Antarctica, seismic observations have been made since 1980 using radio-telemetered and local networks. The observations in the 1981–1982 field season enabled us to determine the hypocenters of 162 local events. The hypocenters can be grouped into two sets, group I with focal depths ranging from the elevation of the summit to sea level, and group II with depths ranging from sea level to 15 km. The epicenters of group I earthquakes are concentrated 1 to 2.5 km north-northeast of the summit crater, and have a nearly vertical focal depth distribution. They are accompanied by infrasonic signals of two distinct waveform types, which allowed the earthquakes to be classified into two sub-sets, called arbitrarily α- and β-type events. There are about twice as many α-type events than β-Type events. The epicenters of the deeper α-type events are more tightly clustered than those of the shallower β-type events. A comparison of the arrival times of the infrasonic signals with those of the seismic signals at the summit station indicates that infrasonic signals associated with α-type events may be shock waves traveling at a sound speed of around Mach 2. The shock waves may be generated at the surface of the lava lake that occupies an inner summit crater. The different character of α- and β-type infrasonic signals is probably the result of different partitioning of acoustic and seismic energies for the shallower β-type events and the somewhat deeper α-type events. The number of group II earthquakes is smaller by one order of magnitude compared to group I earthquakes. The hypocenters are scattered more widely and form an elongated pattern that is slightly inclined to a northeast. Group II earthquakes are not accompanied by infrasonic signals but are characterized by higher-frequency seismic signals at all stations.

1. Introduction

Since the discovery of Mt. Erebus (3794 m, 77°32′S, 167°09′E) in 1841, volcanic eruptions have been observed occasionally by Antarctic research expeditions. However, it is only since 1972 that the summit region of Mt. Erebus has been visited on a yearly basis. Since 1972, an anorthoclase-phonolite lava lake occupies the inner pit crater within the main summit crater. KYLE *et al.* (1982) summarized the lava lake development, its petrology and the seismicity observed at the summit of Mt. Erebus. The general seismicity of Ross Island and the McMurdo Sound region has been summarized by KAMINUMA (1976). Seven K-Ar ages on Erebus lavas indicate a rather young age of the volcano of 0.20–0.93 Ma. Basaltic ejecta contain mantle inclusions such as dunite (KURASAWA *et al.*, 1976). Of great interest is the tectonic setting of active Mt. Erebus, specifically with regard to the state of stress, mechanism of magma generation and the tectonic history of the McMurdo Sound area which is located at the tectonic boundary between East and West Antarctica.

In order to monitor the long-term seismicity of Mt. Erebus, radio-telemetering stations and temporary stations were installed in 1980 as part of the cooperative International Mt. Erebus Seismological Studies (IMESS) which involves the U.S.A., New Zealand and Japan. Some of the results obtained in the 1980–1981 field season have been summarized by TAKANAMI *et al.* (1982). The telemetering network in 1980 consisted of only three stations, one at the summit and two on the flank. In 1981, one station was added on the flank and another on Mt. Terror (KIENLE *et al.*, 1982). This research is mainly based on the IMESS observations from November 15, 1981 to February 15, 1982 during the 1981–1982 field season, hereafter referred to merely as this field season. Hypocenter locations of volcanic earthquakes are presented and the relationship to volcanic activity observed at the summit is discussed.

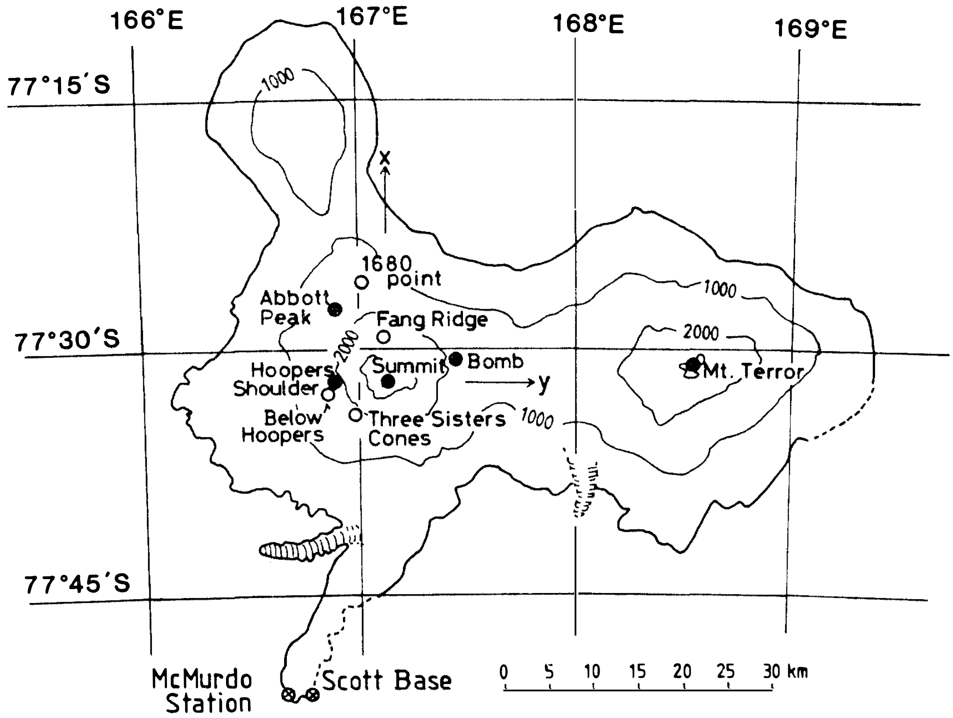

Fig. 1. Seismic network in the 1981–1982 field season. Solid circles denote radio-telemetry stations, while open circles show temporary stations.

2. Observations

Figure 1 shows the seismic network operated in this field season. Each telemetering station (solid circle) had a vertical-component seismometer with a 1 Hz natural frequency. The seismic signal was frequency-modulated for transmission to Scott Base on Ross Island and then demodulated to be recorded on a 14-channel FM data recorder as schematically illustrated in Fig. 2a. Each temporary station (open circle) had a vertical-component seismometer with 2 Hz natural frequency and the seismic signal was recorded on a 30-day direct-analog-recorder. The temporary network was operating December 1–22, 1981. Figure 2b illustrates the overall frequency responses of the telemetered seismic velocity signals. In order to prevent overloading, the amplification at the Erebus summit station, hereafter referred to merely as the summit station, is about 1/7 that of the other stations. Since the aperture of the seismic array is 3–15 km, a rather precise determination of hypocenters for the focal depth range of 3–15 km is expected.

In order to detect air pressure variations and induced geomagnetic fields which may be associated with volcanic eruptions, infrasonic sensors and an induction loop system were installed at the summit station (Dibble, 1982). Correlations of loop signals

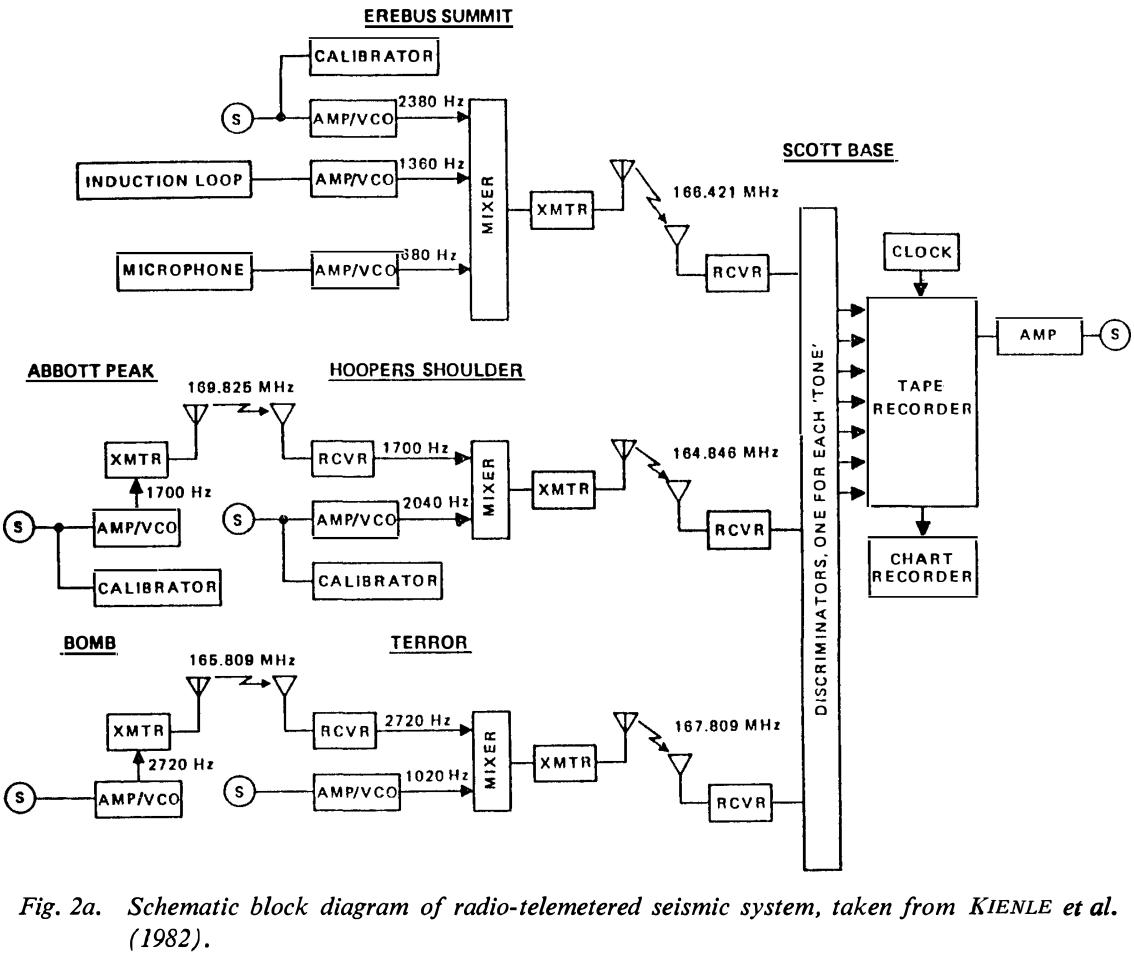

Fig. 2a. Schematic block diagram of radio-telemetered seismic system, taken from Kienle *et al. (1982).*

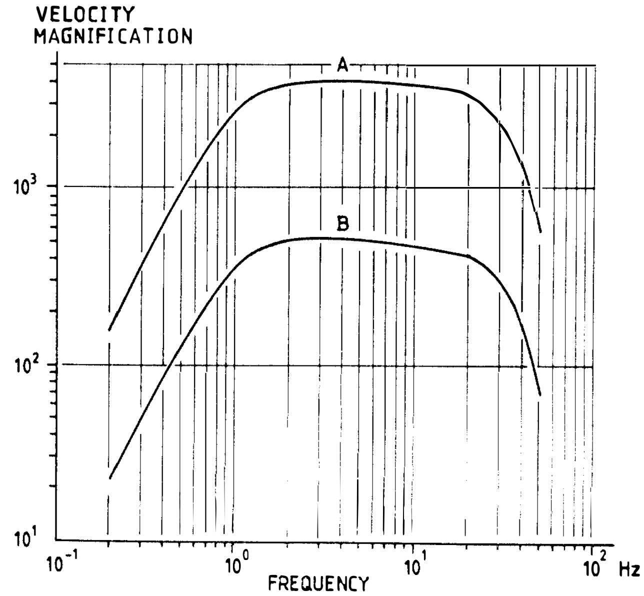

Fig. 2b. Overall velocity frequency-responses of telemetering stations. "A" for Abbott, Hoopers Shoulder and Bomb stations, "B" for Erebus summit station.

with earthquakes could not be observed in this field season. However, two distinct types of infrasonic signals were positively correlated with certain earthquakes.

3. Data Analysis

3.1. Seismicity

The lower part of Fig. 3 illustrates the daily number of events which were monitored at the Hoopers Soulder station. No seismic data was acquired for July–November 1981 because of exhausted batteries at the unmanned telemetering station. Intense swarms occurred in April, July and December. They may have been icequake swarms generated from local glaciers or snow fields on Ross Island, since most of the events were characterized by sharp onsets of higher-frequency waves followed by exponentially decaying longer period oscillations of 5–10 second duration. The daily number of events up to July 1981 was counted from pen-monitor records with variable gain. However, the events for this field season were counted using an event-triggering circuit and an analog comparator. By choosing an appropriate gain setting for the comparator, we were able to compare the seismicity at the Hoopers Shoulder station with that at the summit station (different overall gains shown in Fig. 2b). As shown on the upper right of Fig. 3, the number of events counted at the summit is greater by approximately one order of magnitude as compared to Hoopers Shoulder. The number of counts at other stations on the flank of Mt. Erebus is very similar to that at Hoopers Shoulder. There may be a quasi-periodic (1.5–2 months) change of seismicity (dotted line in Fig. 3 bottom). The long-term monitoring of seismic signals during this field season also showed several cases of prolonged higher signal levels, lasting for 2–15 hours when compared to the background noise levels (inset box in Fig. 3). These tremor episodes have a characteristic period of 1–2 s but also contain higher-frequency

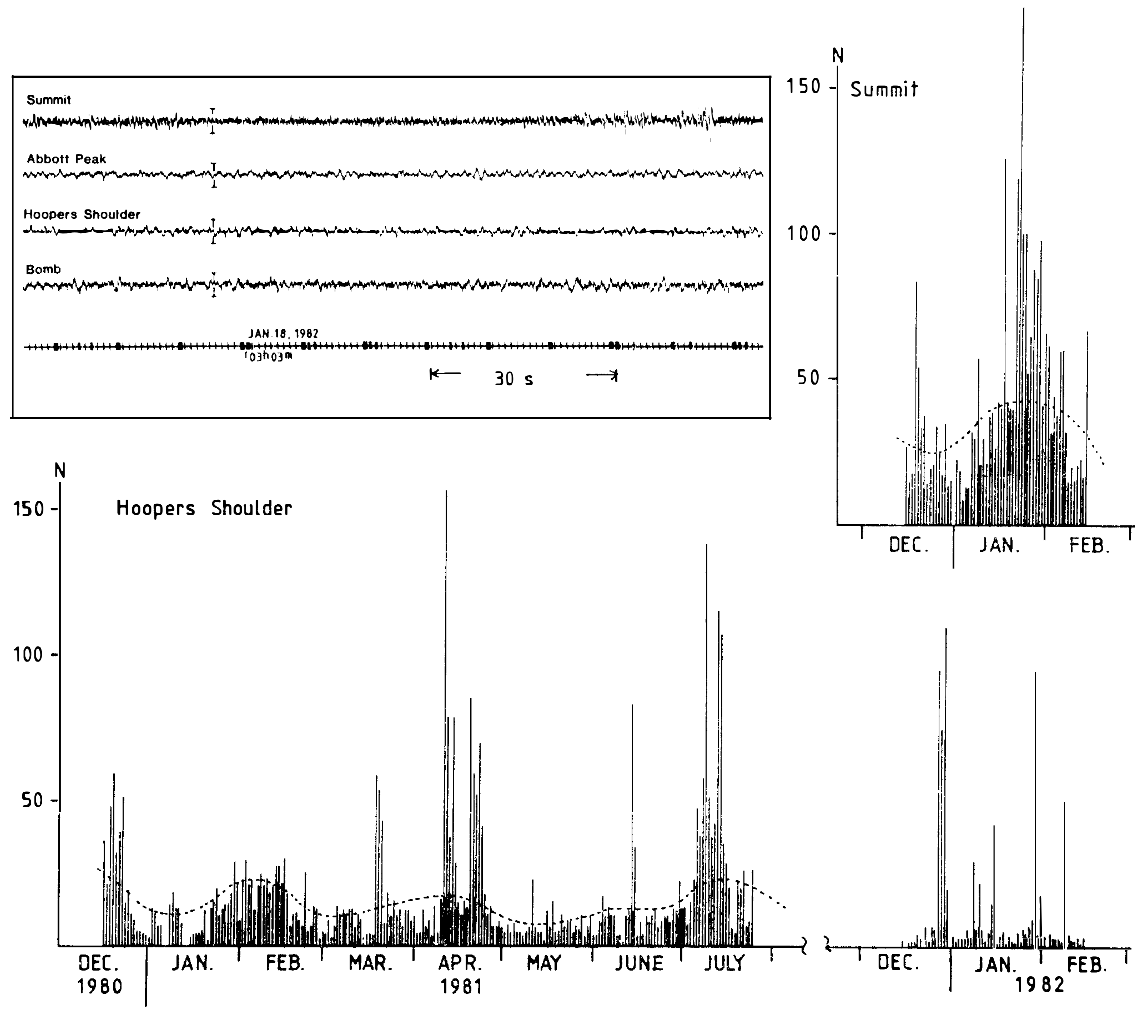

Fig. 3. Daily number of events at Hoopers Shoulder station, December 1980–February 1982 (modified from TAKANAMI *et al., 1982) and at Erebus summit station, November 1981–February 1982. The box gives an example of tremor. Short bar in the summit trace indicates a peak-to-peak level of 1.5 mkine, while bars in the remaining traces indicate 250 μkine.*

(5–10 Hz) components. The tremor episodes are most likely of volcanic origin. In contrast to the rather high level of seismicity observed at Mt. Erebus, no local events could be detected near the Mt. Terror station during this field season.

3.2. *Eruptive activity and infrasonic waves*

From November 27 to December 12, 1981, scientists stayed at the summit hut and took field notes on eruptive phenomena. They frequently heard "roaring", "rumbling", "shaking" and other explosion sounds and observed eruption clouds (Table 1). Reported eruptions correlated well with the seismograms of the summit station. Though not every field log entry (DIBBLE, 1982) had a corresponding seismic signal and not every seismic signal produced an audible or visible eruption, we note that the numbered events in Table 1 were always accompanied by infrasonic waves. Therefore, we surmise that events that are accompanied by infrasonic signals during time periods when no observers were at the summit are also associated with eruptive activity of Mt.

Table 1. Eruptive activities of Mt. Erebus.

Date	Time (UT)	Description and observers location	Duration (s)	Event No.
Nov. 29	2001	Medium long roar, in tent	5	#40
30	1943	Rumble, in tent		#41
Dec. 2	0547	Short weak roar, in crater at rim		#55
〃	1117	Short weak roar, on speaker		#58
〃	1441	Single explosion, on speaker		#60
〃	1735	Strong explosion	3	#63
3	0120	Two brief roars, 4 s apart, 1st weak, 2nd medium, on speaker	1	#70
〃	0601	Loud explosion, in Hut, on seismograph	8	#79
〃	1447	Explosion, woke observer in tent		#84
〃	1521	Loud roar, woke observer in tents	3	#85
〃	1656	Heard in tents		#86
〃	2102	Loud roar on speaker, woke observer		#87
4	0914	Dull roar, also on seismograph	4	#89
〃	1935	Short bang	2	#91
5	2311	Loud roar. 3 bombs seen above Shackleton Cairn		#93
〃	1708	Weak explosion, in tent	1	#109
7	1510	Low growing rumble at Hut, first seen on seismograph		#127
〃	1538	Long loud roar at Hut with small earthquake. No bombs seen	10	#128
〃	1820	Long loud roar	6	#131
〃	2151	Long crackly roar at Hut, with small complex earthquake on seismograph	6	#135
Dec. 8	0554	Long loud roar	17+	#150
〃	1728	Long loud roar, woke observer in tent		#158
9	0118	Little roar, on speaker	1	#173
〃	1221	Long roar on speaker, but faint outside hut		#185
〃	1523	Long loud explosion	10	#188
〃	1537	Medium explosion	2	#189
10	1046	Double roar, in and outside hut		#200
12	0702	Sharp roar, seemed to shake hut. Dark cloud arose	2	#242

Erebus. The infrasonic signals can be classified qualitatively into two types, arbitrarily called α-type and β-type in this report. The α-type signals are characterized by a sharp impulsive onset with a polarity of increasing pressure and the following high frequency pressure variations (second trace of Fig. 4a). The β-type signals are sinusoidal oscillations of lower frequency compared to α-type signals (second trace of Fig. 4b). Of the 1000 events which were detected during this field season, 175 events were found to be accompanied by infrasonic signals. Signal duration typically ranges from 5 to 20 s with a peak value of around 10–12 s for both types. α-type events occur twice as frequently as β-type events and no periodicity or patterns could be found in the daily number counts. When the long-term average of the signal level of infrasonic record became large, more seismic events were found to be detected at the the summit station, which indicates more frequent eruptions from the lava lake.

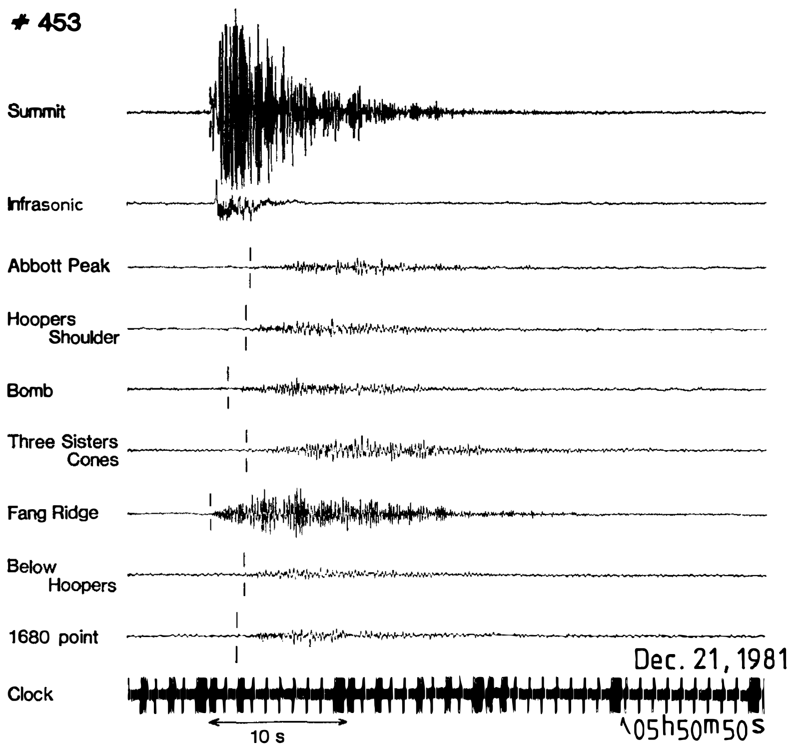

Fig. 4a.

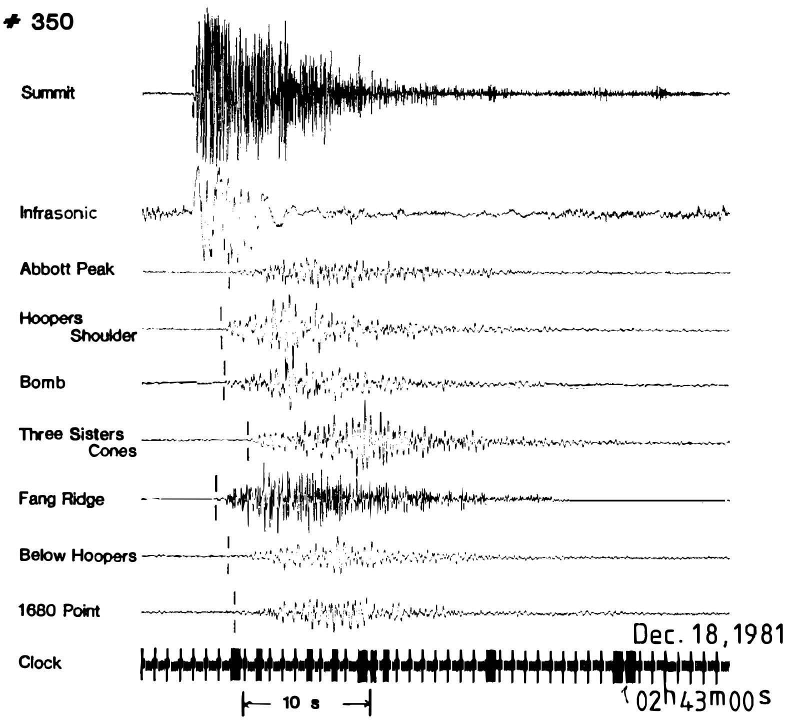

Fig. 4b.

Fig. 4. Examples of seismic events: (a) #453 with α-type infrasonic waves, (b) #350 with β-type infrasonic waves.

3.3. *Hypocenter determination*

Among 1000 events recorded in this field season, more than 300 events had clear onsets at 5 or 6 network stations. Examples are given in Figs. 4a and 4b. For this study, 210 events were selected for which the reading errors for the P-wave arrival times at all 4 Erebus stations were within ± 0.2 s. Since the 4 telemetered seismic signals were on the same clock system, time calibrations were unnecessary and IRIG-S formatted 10 Hz clock signals (see bottom traces in Figs. 4a and 4b) assured that the reading errors were less than 0.1 s. Since the crustal velocity structure beneath Ross Island and Mt. Erebus is unknown, a semi-infinite medium with uniform and homogeneous P-wave velocity of $V_p=2.1$ km/s was tentatively assumed for the hypocenter calculations. As illustrated in Fig. 1, the center of the Cartesian coordinate system was located at the summit station, where x is for longitudinal direction, y for latitudinal direction and z is positive downward from the elevation of the summit station. The station coordinates were determined form the 1:250000 topographic sheet of Ross Island. A first-order iterative approximation method for the Taylor expansion of the travel time formulae was used for the hypocenter calculations (*e.g.*, NISHI, 1971). For an iterative approximation of a non-linear equation, appropriate initial conditions

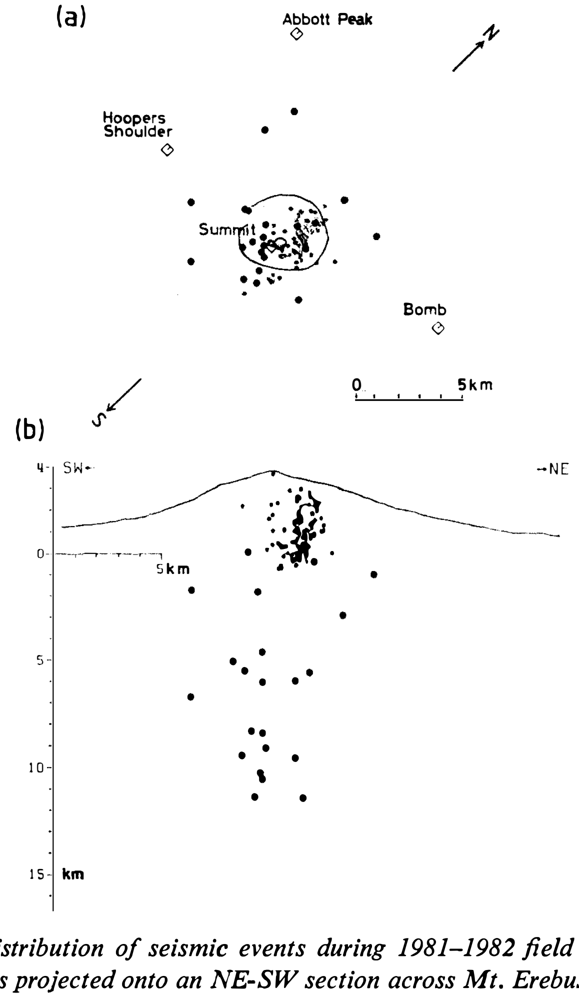

Fig. 5. Hypocentral distribution of seismic events during 1981–1982 field season, (a) epicenters, (b) hypocenters projected onto an NE-SW section across Mt. Erebus.

are essential for convergence. The initial trial hypocenter was chosen to be located beneath the summit station by selecting a most probable combination of z_0 and δt_0 in a least-squares sense for the range of $z_0=0$–4 km and $\delta t_0=0$–2 s, where z_0 denotes focal depth and δt_0 the difference between the observed P-wave arrival time at the summit station and the origin time. Since there is an equal number of 4 unknown parameters and 4 linearized observation equations, a few iterative steps must result in a unique final hypocenter solution with zero O-C (observed minus calculated) time residuals at all stations, if the solution ever exists. Since the telemetering network has an aperture of 8–15 km, the accuracy of hypocenter determinations becomes less for events with focal depths greater than 15 km and for epicenters located far outside the network. A total of 162 hypocenters were finally determined of the 210 selected events. Figure 5a shows the epicentral solutions and Fig. 5b the focal depth distribution projected onto a vertical NE-SW section across Ross Island.

The error in the hypocenter determinations comes from 200–300 m inaccuracies of station coordinates, inaccurate assumption of a 2.1 km/s average velocity and reading errors in the P-arrival times. Since most earthquakes occurred within the network, uncertainties in the station coordinates were found to have an insignificant influence on the hypocentral distribution when a re-determination was made with changed station coordinates. For lower P-wave velocities, focal depths decrease and the reverse is true for higher P-wave velocities. However, the change of focal depth can be offset by a change in the origin time, and the precise influence has to be estimated in future work. Since P-arrival data were picked for one or two probable wavelets, misidentification of the first-arriving wavelet may cause an error greater than ± 0.2 s in the observed arrival times. Therefore, a simulation test was made by re-reading arrival times. This test showed that the changes of the hypocenter parameters are within 0.2 km for the y-direction, 0.8 km for the x-direction and 1.0 km for the z-direction for 90% of the 162 events, which does not significantly alter the obtained pattern in Fig. 5.

4. Discussion

The Mt. Erebus earthquakes can be grouped into two sets based on their focal depth distributions (Fig. 5). Group I events have a pipe-like vertical distribution beneath the summit with focal depths ranging from sea level to the elevation of the summit. The hypocentral depths of group II events range from sea level to 15 km. The number of group II events is smaller by one order of magnitude compared to group I events and the hypocenters are more scattered, forming an elongated pattern that is slightly inclined to a northeast.

Most of the epicenters of group I events (Fig. 5a) are located within a small region 2 km in diameter that is displaced 1–2.5 km in a north-northeasterly direction from the main crater, and they tend to cluster in a north-south direction. Group I events can be classified into α- and β-type events (open and solid circles in Fig. 6, respectively) based on the character of the associated infrasonic waves. There are fewer β-type events than α-type events; β-type events are also shallower than α-type events. There is an indication of blank region of seismicity (dotted lines in Fig. 6b) which may correspond to a magma body beneath the lava lake at the summit.

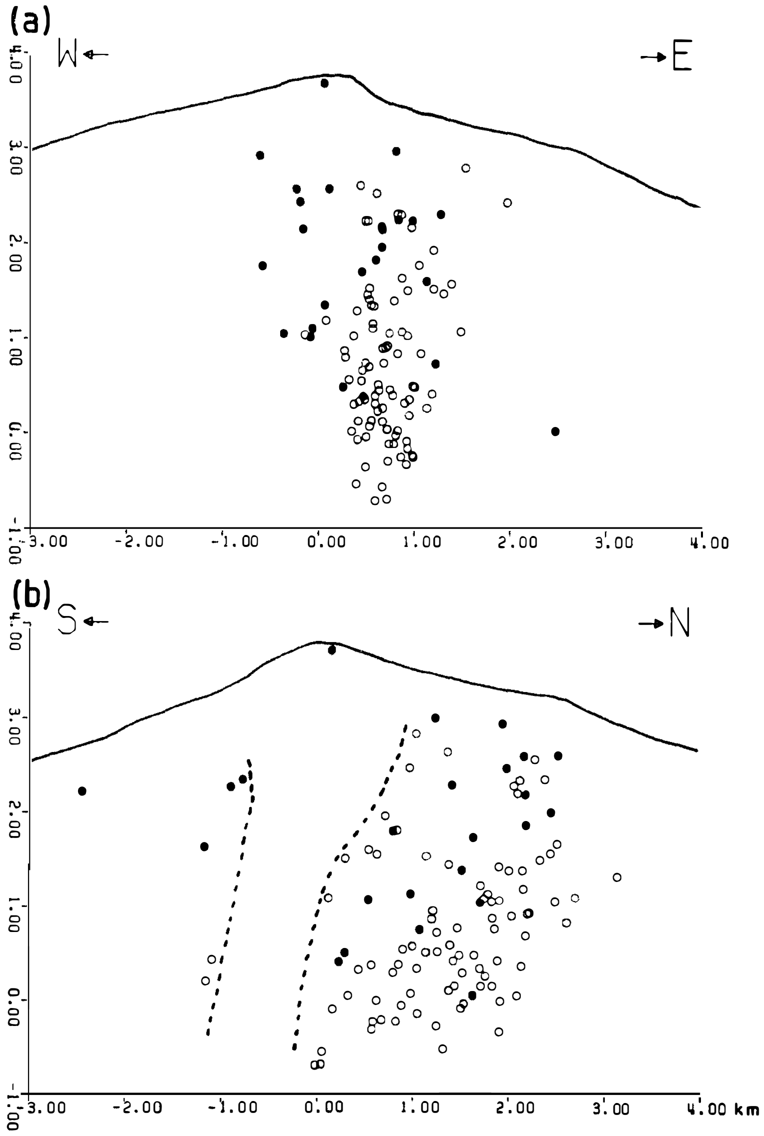

Fig. 6. Hypocentral distribution of group I events (open circles for α-type events, solid circles for β-type events), (a) E-W cross section, (b) N-S cross section. Dotted lines in Fig. 6b indicate a possible blank region of earthquakes.

For the future direction of research on the volcanic activity of Mt. Erebus, the following discussion may be useful. In order to analyze the correlation between seismic and infrasonic signals at the summit station, time differences Δt_p between the arrival time of infrasonic waves $t_{i,s}$ and that of seismic signals t_s were examined. Figure 7a shows that Δt_p $(t_{i,s}-t_s)$ ranges from 0.2 to 0.5 s for most of the α-type events, and from $-0.2\sim 0.0$ s for β-type events. Figure 7b shows ray paths for the calculation of the theoretical time difference Δt_p for α-type events. If the signals detected by the infrasonic sensor I are air pressure pulses from point L at the surface of lava lake (3550 m height), which were converted from or triggered by the transmitted seismic waves from the hypocenter H, the following equation can be obtained.

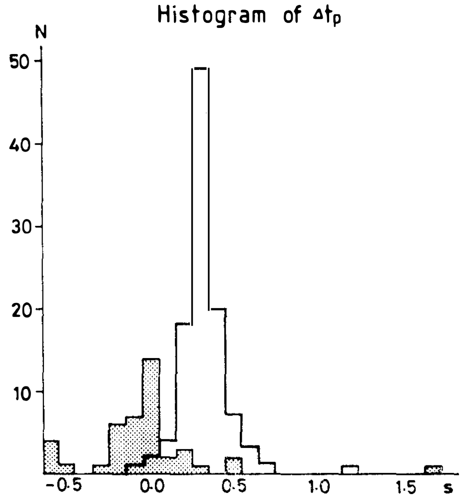

Fig. 7a. Frequency number of Δt_p (infrasonic minus P-wave arrival time) for α-type events (open histogram) and β-type events (stippled histogram).

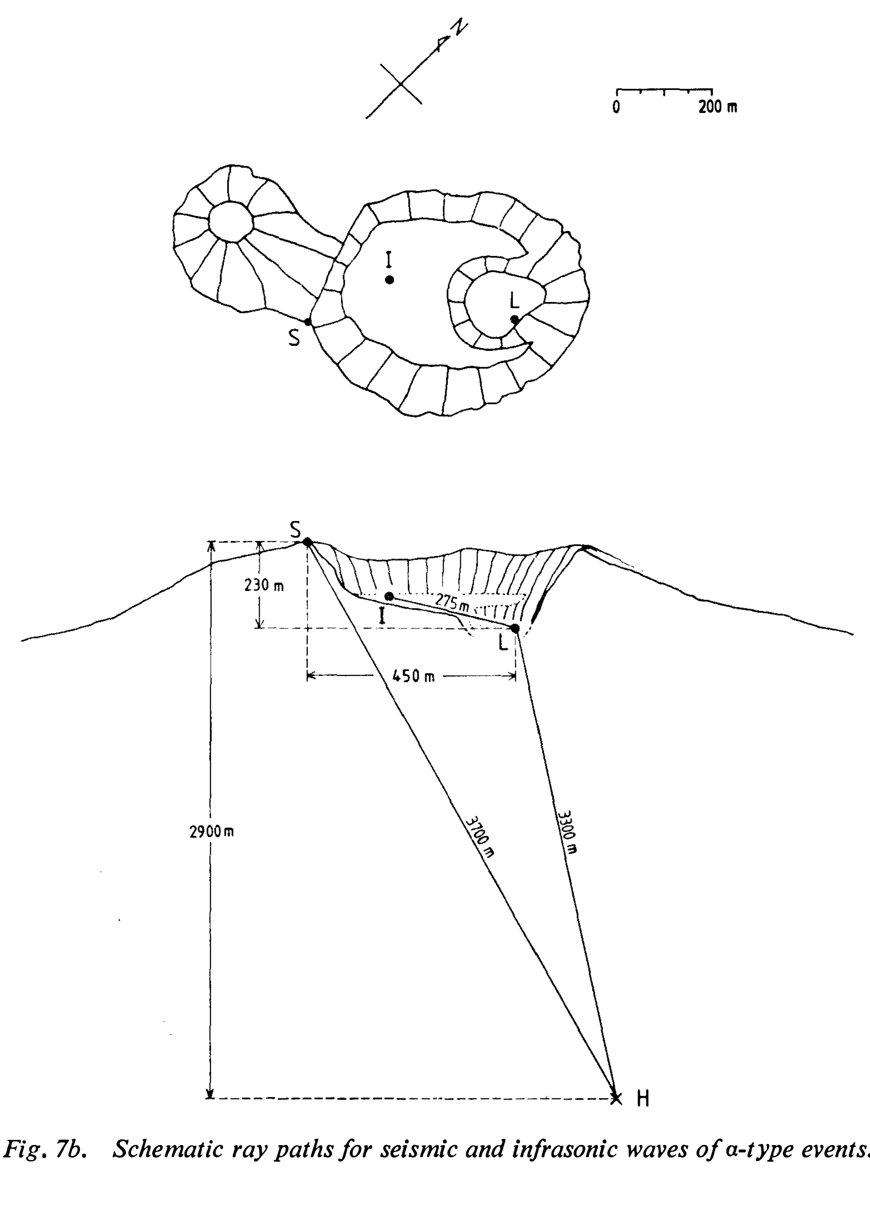

Fig. 7b. Schematic ray paths for seismic and infrasonic waves of α-type events.

$$\Delta t_p = \frac{l_3}{V_{\text{sup}}} - \frac{l_1 - l_2}{V_p}$$

or

$$V_{\text{sup}} = \frac{l_3}{\Delta t_p + \dfrac{l_1 - l_2}{V_p}}$$

where S denotes the location of the seismometer, V_{sup} does sound velocity in air and l_1, l_2 and l_3 are slant distances HS, HL and LI, respectively. The infrasonic sensor I is located on the main crater floor of 3650 m height and about 505 m distant from S on the rim of 3280 m height. The distance LI is approximately 275 m as shown in Fig. 7b. Substitution of appropriate hypocentral parameters for some events (focal depth range 1–3 km) and an appropriate choice of coordinate values of point L gives $V_{\text{sup}} \sim 590$–650 m/s for the observed value of $\Delta t_p \sim 0.3$ s and the assumed value of $V_p = 2.1$ km/s. Since the sound velocity at -20°C is 319 m/s, the obtained value of V_{sup} corresponds to Mach 1.8–2.0. In order to verify the existence of such supersonic shock waves, it would be useful to install two infrasonic sensors with an aperture of 300 m within the crater.

Negative or zero Δt_p's for β-type events in Fig. 7a cannot be explained by simultaneous generations of seismic and infrasonic signals.

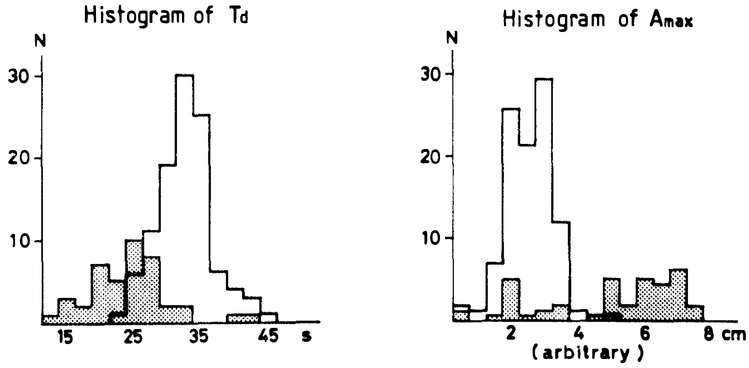

Fig. 8. Frequency number of seismic coda length T_d (left) and maximum infrasonic trace amplitude A_{max} (right). Open histogram for α-type events, stippled histogram for β-type events.

Figure 8 illustrates the statistics of duration time T_d of seismic signals (left) and the maximum trace amplitude $A_{\max}$ of infrasonic signals (right) at the summit station under fixed overall amplification conditions. Though $A_{\max}$ of β-type events is greater by a factor of 2–3 compared to that of α-type events, T_d of β-type events is smaller by 30% (10 s) compared to that of α-type events. The above relation suggests different partitioning of acoustic and seismic energies for α- and β-type events; β-type events generate more infrasonic and less seismic energy than α-type events.

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