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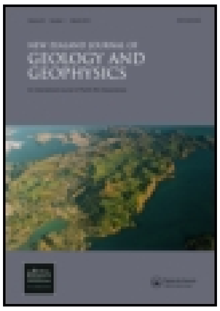

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Monitoring seismic precursors to an eruption from the Auckland Volcanic Field, New Zealand

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Abstract The Auckland Volcanic Field (AVF) in New Zealand is monitored by a network of five telemetered, vertical-component, short-period seismographs. Between 1995 and 2005, 24 earthquakes were located in the Auckland region. Ten of these were located reasonably reliably (position and depth uncertainty ≤10 km) and all of these were <15 km deep. Only one of these earthquakes occurred within the AVF. Magnitudes ranged from M_L 1.6 to 3.3, and five earthquakes of $M_L \geq 2.4$ were felt. There were few reliably located earthquakes because most were not recorded by the whole network owing to their relatively low magnitude and a high level of background noise. The Auckland earthquakes are believed to represent normal background seismicity and are not thought to be eruption precursors. All earthquakes were of high-frequency, tectonic type; no low-frequency, volcanic earthquakes were recorded. Based on seismic precursors to eruptions from historically active volcanic fields, we estimate that precursory earthquakes could occur as little as 2 weeks before an Auckland eruption and they could be as large as M_L 4.5–5.5. Based on the depth of the background seismicity in Auckland, and previous estimates of the ascent rate and source depth of AVF magmas, we calculate a precursory period as short as a few days. Our best estimate of the length of pre-eruption seismicity is therefore a few days to a few weeks. The largest precursory earthquakes could be large enough to be felt by most of the population who live in Auckland City. During a magmatic intrusion, deep long-period earthquakes might occur at c. 30 km as magma ascends into the crust. Earthquakes would probably have to be a lot shallower, perhaps only 5 km, before their epicentres might be useful for estimating the location of any eruption. Geodetic monitoring methods (GPS and InSAR) might perform as well as seismic monitoring for identifying unrest, but they have significant limitations. To better monitor and interpret precursory seismicity from the AVF, an increase in the number of seismographs and an improvement in our understanding of the local crustal structure are needed.

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Keywords seismicity; Auckland Volcanic Field; eruption precursors

INTRODUCTION

The Auckland Volcanic Field (AVF, Fig. 1) consists of 49 small basaltic volcanoes covering an area of c. 360 km^2 coincident with the city of Auckland, New Zealand (Allen & Smith 1994). The most recent eruption occurred only 600–800 yr ago and further eruptions from the AVF are likely. Auckland is New Zealand's largest city, with a population of more than one million people; it is the country's economic centre and also the site of the main international airport. The social and economic consequences of an eruption from the AVF are significant. The Auckland Regional Council (ARC), the local government authority responsible for the Auckland area, has a contingency plan for future activity (Beca Carter Hollings & Ferner 2002), has commissioned several reports on volcanic hazards, possible eruption scenarios and risks (Johnston et al. 1997; Paton et al. 1999), and has produced a range of information brochures for the general public describing the AVF and its hazards. Magill & Blong (2005a,b) have shown that tephra fall is the highest risk to people from an eruption of the AVF.

A key part of trying to mitigate the hazards associated with a future eruption from the AVF is a monitoring system that can identify when an eruption might be about to occur and can give ARC and other authorities sufficient time to implement contingency plans. Seismographs were recognised as the most practical monitoring method for the AVF (Cassidy et al. 1986) and during the late 1980s and early 1990s a network was established in the Auckland area.

In this paper we discuss the kind of seismic activity that might be expected to occur before an eruption from the AVF by examining observations from similar volcanic fields active in historical times. We describe the earthquakes in the Auckland area recorded by the monitoring network during 1995–2005 and the reliability of their calculated locations. We discuss the depth of earthquakes, the length of any pre-eruption warning period, and earthquake felt effects. We conclude by recommending how to improve our ability to interpret earthquake data from the AVF and use that to better identify precursory seismicity to a future eruption.

THE AUCKLAND VOLCANIC FIELD

The AVF has been active for c. 140 000 yr and has produced c. 3.4 km^3 of basaltic eruption products (Dense Rock Equivalent, DRE) from 49 separate vents (Searle 1981; Kermode 1992; Smith & Allen 1993; Allen & Smith 1994). Eruptions have ranged in style from phreatomagmatic (forming maars and tuff rings) to magmatic (producing lava flows and scoria cones), with individual vents often displaying several

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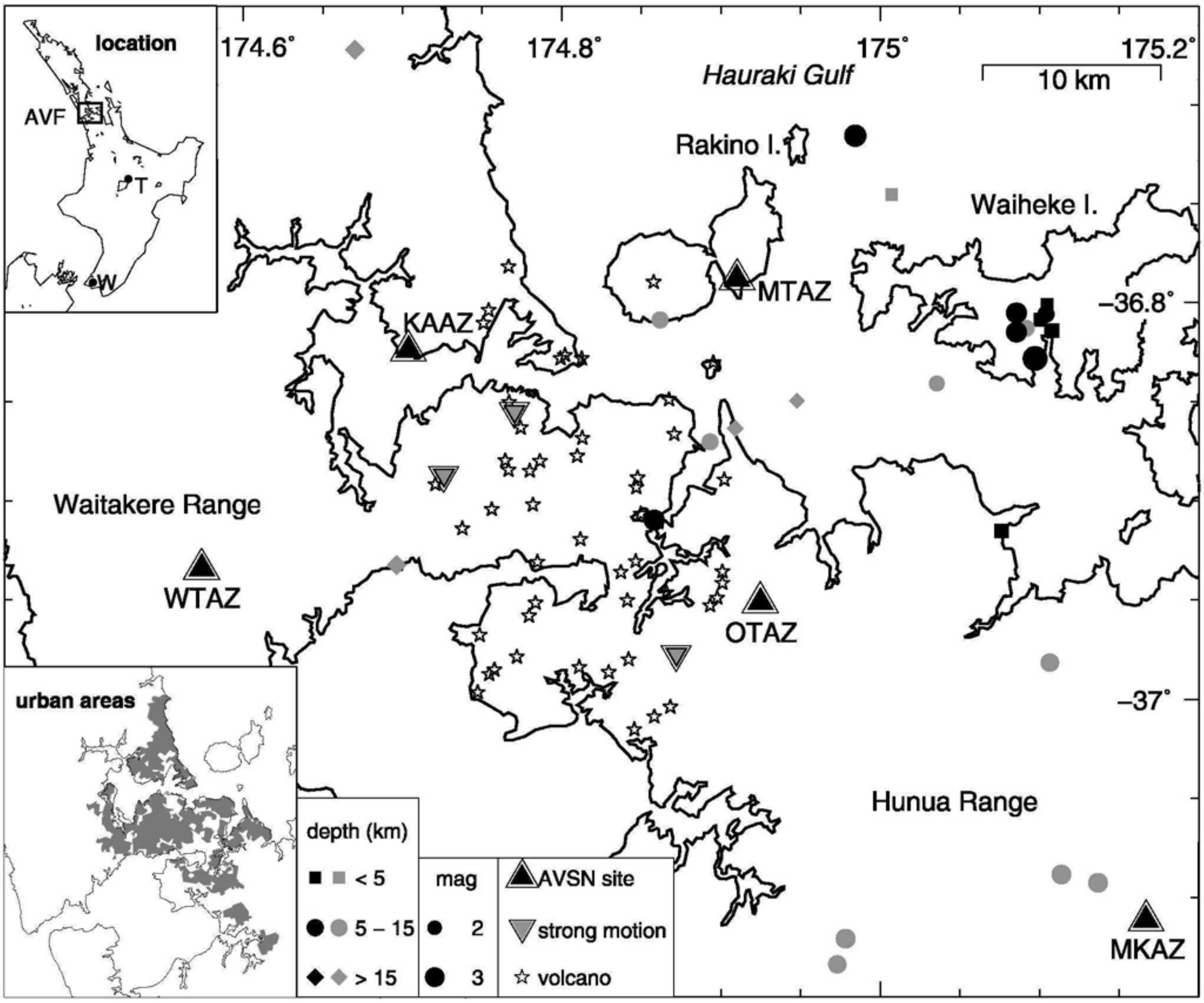

Fig. 1 A map showing the Auckland Volcanic Field (AVF), the seismographs of the Auckland Volcano-Seismic Network (AVSN), and the epicentres of earthquakes located between 1995 and 2005. Stars mark volcanic centres, triangles the location of the seismographs labelled by their site codes, and inverted triangles strong motion sensors, one of which is co-located at KAAZ. Earthquake symbols depend on depth, and the size is proportional to magnitude. Black symbols are reliably located and grey symbols poorly located. The upper inset shows the location of the AVF, with T and W marking the data recording and analysis centres at Taupo and Wellington, respectively. The lower inset shows the Auckland urban area.

eruption styles (Houghton et al. 1991). The AVF is generally considered to be a monogenetic volcanic field, with each vent erupting a single batch of magma in a discrete eruption episode that probably lasts for no more than 10 yr (Kermode 1992). However, there is growing evidence that in some parts of the field one magma batch may have fed several nearby vents over a period of less than a few hundred years (Rout et al. 1993; Allen & Smith 1994; Cassidy et al. 1999; Cassidy & Locke 2004).

Thirty-three of the volcanic centres have eruptive volumes <0.01 km^3 (DRE) and six have volumes of 0.05–0.35 km^3 (Allen & Smith 1994). The most recent eruption, Rangitoto in c. AD1350, produced c. 2 km^3 of lava and scoria and represents 59% of the total magma erupted from the AVF (Allen & Smith 1994). Although several techniques, including radiocarbon, K-Ar, thermoluminescence, and magnetic declination, have been used to attempt to date the Auckland volcanoes, the age of many remains unknown or uncertain (Allen & Smith 1994; Cassidy & Locke 2004).

Cassidy et al. (1986) noted that the basaltic magma erupted from the AVF is thought to originate within the mantle at a depth of 75–125 km. Based on the presence of mantle-derived nodules in the basalt, they inferred an ascent rate of one to several centimetres per second (cm/s). A source depth of 100 km and an ascent rate of 1–10 cm/s is equivalent to an ascent time of 10–100 days. This source depth is supported by Huang et al. (1997) who used U-Th isotopes to infer that AVF magmas originate in the upper mantle (80–140 km depth). Using a joint inversion of teleseismic receiver functions and surface wave phase velocities, Horspool et al. (2006) found a low *Vs* region at a depth of 70–90 km beneath the AVF that may represent a region with 2–3% partial melt and could be the source region for the AVF magmas.

Miocene sediments (alternating mudstone and sandstone), generally c. 500 m thick, overlie a Mesozoic greywacke basement beneath the AVF (Edbrooke et al. 2003). The basement rocks are up-faulted east of the AVF and outcrop in the Hunua Range to the southeast, where basement faults

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trend in both northeast–southwest and northwest–southeast directions. Where volcanic vents in the AVF cluster together, they are sometimes aligned northeast–southwest, possibly reflecting these basement faulting patterns. Horspool et al. (2006) estimated that the crustal thickness beneath the AVF is c. 27 km, in agreement with a value of 25 ± 2 km determined by Stern et al. (1987) using a seismic refraction profile along the Northland Peninsula, northwest of Auckland.

Historically, there has been a low level of seismicity in the AVF. From 1960 to 1983, 81 earthquakes with a depth <40 km and of $M_L \geq 4$ ($M_L \geq 3.7$ since 1982) occurred within 100 km of Auckland (Cassidy et al. 1986). Most were located in the Hunua Range southeast of Auckland or east near the Coromandel Peninsula, and only one was recorded close to the AVF. In 1891, a M_L 5.5–6.0 earthquake occurred c. 60 km south of Auckland and caused some damage (Edbrook et al. 2003).

PRECURSORY SEISMICITY FROM HISTORICAL ANALOGUE ERUPTIONS

New Zealand is believed to have been settled between c. 1250 and 1300 (Irwin & Walrond 2006), c. 100–200 yr before the Rangitoto eruption. There is no information about eruption precursors in Maori oral history, even though footprints were preserved in freshly fallen volcanic ash (McSaveney et al. 2006).

However, we can try to use observations from historically active volcanic fields similar to the AVF as a guide to possible precursors to eruptions from the AVF. Data from the Smithsonian Global Volcanism Program online database for volcanoes (www.volcano.si.edu) were searched. Only two volcanic fields have been active historically and have a record of earthquake activity—the Michoacán-Guanajuato Volcanic Field in Mexico and the Higashi-Izu Volcano Field in Japan. Information from other volcanic fields, even if they are similar to the AVF in tectonic setting and magma type, is of little use if there is no record of earthquake activity.

The Michoacán-Guanajuato Volcanic Field is in the west-central part of the Mexican Volcanic Belt (MVB). The MVB also includes cone volcanoes and calderas and is related to subduction of the Rivera and Cocos plates beneath the North American plate (Luhr & Simpkin 1993). There are >1000 small volcanic centres in an area of 40 000 km^2, with two volcanoes active historically. Jorullo volcano formed in September 1759 and continued erupting until 1774. The initial lavas were basalts, with later lavas being basaltic andesites (total SiO_2 range 52–55%, Luhr & Simpkin 1993). The eruption built a 250 m high cone and three smaller satellite cones, erupting c. 2 km^3 of magma. Earthquakes were felt in the area from 3 months before the eruption, with between 12 and 47 felt events on any given day (Yokoyama & de la Cruz-Reyna 1990).

Parícutin volcano appeared on 20 February 1943 and continued to erupt until 1952 (Luhr & Simpkin 1993). The initial composition of the Parícutin lavas was basaltic andesite (55% SiO_2), with later lavas being andesitic in composition (60% SiO_2). During its 9 yr of activity, 1.3 km^3 of ash and 0.7 km^3 of lava were erupted, building a cinder cone 450 m high. Yokoyama & de la Cruz-Reyna (1990) studied the pre-eruption earthquakes using data from a seismograph in Mexico City, 320 km east of Parícutin. They identified 22 tectonic-type earthquakes (magnitude ≥3) within the 45 days before the eruption began. No low-frequency (long-period) volcanic earthquakes were recorded at that seismograph. The largest earthquakes were two of magnitude 4.5, and 11 exceeded magnitude 4, generally increasing in magnitude towards the eruption. The first felt earthquake occurred on 5 February, 15 days before the eruption, with many small earthquakes felt in the Parícutin area between 7 and 14 February. Three hundred earthquakes were felt in the Parícutin area the day before the eruption started.

The Higashi-Izu Volcano Field consists of 62 volcanoes (cinder cones, tuff rings, maars, or lava domes) covering c. 350 km^2 on the east side of the Izu Peninsula in Japan, with many submarine volcanoes also located on the seafloor east of the peninsula (Koyama & Umino 1991). The eruption products are dominantly basalt, but also with andesite and dacite-rhyolite. Vents tend to be aligned in a northwest–southeast direction, parallel to the direction of the maximum horizontal compressive stress in the region (Koyama & Umino 1991). The latest eruption occurred on 13 July 1989, when a small submarine crater, Teishi Knoll, was formed (Oshima et al. 1991). Pumice produced by the eruption ranged in composition from 50 to 75% SiO_2 (Kobayashi et al. 1991). Some pumice had an inner core that had a rhyolitic composition (73% SiO_2) and was surrounded by material which is basaltic (51% SiO_2) and similar in composition and texture to typical Higashi-Izu basalts (Kobayashi et al. 1991). Kobayashi et al. (1991) argued that the basalt was the essential ejecta, but that it carried with it some acidic melt.

Significant seismicity in this area appears to have started in 1978, although the area is seismically very active with a notable earthquake swarm in the volcanic field in 1930 (Yoshida & Hamada 1991). Earthquakes immediately preceding the eruption began c. 5 km offshore on 30 June 1989. The number of earthquakes (almost all tectonic in character) increased greatly on 4 July and then decreased slowly before a second peak on 9–10 July, including one earthquake of M5.5 (Yamasato et al. 1991). Low-frequency earthquakes were first recorded on 5 July and frequently from 10 July. They were all located ≤3 km deep close to the eruption site, while tectonic earthquakes were 1–6 km deep. As the earthquake swarm declined on 11 July, intermittent volcanic tremor commenced, sometimes with an amplitude large enough to be recorded 300 km away (Ukawa 1993). On 13 July, water domes and stream plumes were observed 5 km offshore as a result of a submarine eruption (Oshima et al. 1991). The eruption continued for only a few hours. Observations the next day showed no sign of eruptive activity, though gas bubbles were observed on the sea surface for several months afterwards (Oshima et al. 1991). The intruded volume was only c. 10^{-3} km^3 (Yamamoto et al. 1991). Significant seismic activity and ground deformation occurred in the same area subsequent to the 1989 eruption, but there were no further eruptions.

Lead-time of precursory earthquakes

Earthquakes occurred up to 90 days before the eruption of Jorullo, 45 days before the eruption of Parícutin, and 14 days before the eruption of Teishi Knoll. This is a wide range and probably depends on several factors, including the depth of the source region, how quickly the magma ascends, and the depth of the brittle/ductile transition. These factors depend, at least in part, on the strength of the crustal rocks, and the properties of the intruding magma.

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Failed eruptions

Several earthquake swarms in the Izu area after the 1989 eruption were accompanied by ground deformation that was interpreted as shallow magmatic intrusion, but no eruptions occurred. Aeromagnetic data from the AVF show no evidence for magnetic anomalies not associated with a known volcanic centre (Cassidy et al. 1999; Cassidy & Locke 2004). However, if magma stops several kilometres below the surface, it may still generate significant seismicity, but could be too deep to produce a magnetic anomaly at the surface. Intrusion without eruption might therefore also occur in the AVF.

THE AUCKLAND VOLCANIC FIELD MONITORING NETWORK

The AVF is monitored by the Auckland Volcano-Seismic Network (AVSN), consisting of five telemetered, vertical-component, short-period (1 Hz) seismographs (Fig. 1). Four of the five sensors are installed in shallow (10–30 m deep) boreholes in an attempt to reduce the level of surface-generated ground noise. Data are digitised at a sampling rate of 100 Hz and are continuously sent to GNS Science in Taupo and Wellington (Fig. 1) for near-real-time data analysis. There are also four strong motion sensors in the Auckland area (Fig. 1). Despite most of the sensors being in shallow boreholes, the background noise level in the Auckland area is very high.

EARTHQUAKE RELOCATION AND HYPOCENTRAL QUALITY

Earthquakes in the Auckland region are routinely located by GNS Science using a standard least-squares, iterative location algorithm and a 1D velocity model (Maunder 2001). Station corrections, to help account for variations from the assumed 1D velocity structure, are not used. When there are insufficient data or the nearest seismograph is too far from the epicentre, the depth of the earthquakes is routinely fixed and only the epicentre and origin time are calculated. As the depth of an earthquake beneath the AVF may relate to the depth of a magmatic intrusion, it is important to try to estimate the depth and not to fix it at some realistic, but arbitrary, value. We have therefore relocated all the earthquakes close to the AVF between 1995 and 2005 without fixing the depth. In many cases the depth is very poorly determined, but knowing this information is important when dealing with a potential eruption.

We have used the same 1D velocity model used for routine locations, but with the computer program NonLinLoc, a probabilistic, fully non-linear earthquake location program that calculates a maximum likelihood hypocentre by a 3D search (Lomax et al. 2000; Lomax 2001; Husen et al. 2003). The hypocentre is expressed as a probability density function and includes the effects of location uncertainties due to seismic network geometry and arrival time uncertainty. The hypocentral uncertainty is displayed as a scatter diagram with the number of samples proportional to probability (Lomax 2001). This algorithm can display an irregularly shaped uncertainty region or one with more than one minimum (e.g., Husen et al. 2003). NonLinLoc also calculates an expectation hypocentre with a 68% confidence ellipsoid that corresponds to the location obtained using a conventional linearised algorithm, such as that used for routine earthquake location in Auckland.

The best located earthquakes had P-arrivals from most sites in the AVSN and at least two S-arrivals (Fig. 2). They usually have an azimuth gap (the greatest angle surrounding the epicentre without an observation) ≤180°, and the nearest seismograph is often within a distance equivalent to the depth of the earthquake. If we take a reasonably well located earthquake to be one with a location uncertainty of ≤10 km, then there were 10 reasonably well located earthquakes (Fig. 1, Table 1). Other earthquakes were poorly located with an uncertainty of as much as 15–20 km in both position and depth (Fig. 3, Table 1). These typically have fewer arrivals and an azimuth gap >180°.

The high noise level in Auckland City, the relatively small magnitude of the earthquakes (M_L 1.6–3.3), and periods of equipment failure are the main reasons for so many poorly located earthquakes. Only two earthquakes were located using data from all sites in the AVSN and only 13 earthquakes had four or more P-phase picks, though several used data from other seismographs ≥100 km away.

AUCKLAND SEISMICITY 1995–2005

Between 1995 and 2005, 24 earthquakes were located in the Auckland area (Fig. 1, Table 1). Six were in the Hunua Range southeast of the AVF, 11 offshore in the Hauraki Gulf, and one north of the AVF. The remaining six earthquakes probably occurred within the AVF. All well-located earthquakes were ≤15 km deep (Table 1). Five earthquakes were felt on islands in the Hauraki Gulf, but none in Auckland City. Four of these occurred at Waiheke Island over 2 days in November–December 2005.

There are insufficient earthquakes to estimate statistically the smallest earthquake that can be located, but the number of earthquakes between M_L 2.0 and 2.5 suggests that the network can probably locate most earthquakes c. M_L >2.0, although to be reasonably well located an earthquake probably needs to be M_L ≥2.5.

All the earthquakes were typical high-frequency, tectonic earthquakes. No volcanic earthquakes were recorded (events with a harmonic waveform, a dominant frequency content of c. 1–5 Hz, and indistinct phases; McNutt 2000).

DISCUSSION

Although six earthquakes were located within or close to the AVF, there is no evidence that they are precursors to an eruption, rather, they probably represent the normal background seismicity of the region. The earthquakes in the Hunua Range southeast of the AVF are consistent with the long-term seismicity (Cassidy et al. 1986).

Length of a pre-eruption warning period

One of the most important questions regarding the AVF is the length of any warning period before the next eruption. The AVSN is designed to recognise the first signs of unrest, but we do not know how long before an eruption earthquakes will first be recorded, nor how long after that earthquake locations may help with estimating an eruption site. The historical analogue eruptions suggest 14–90 days of precursory

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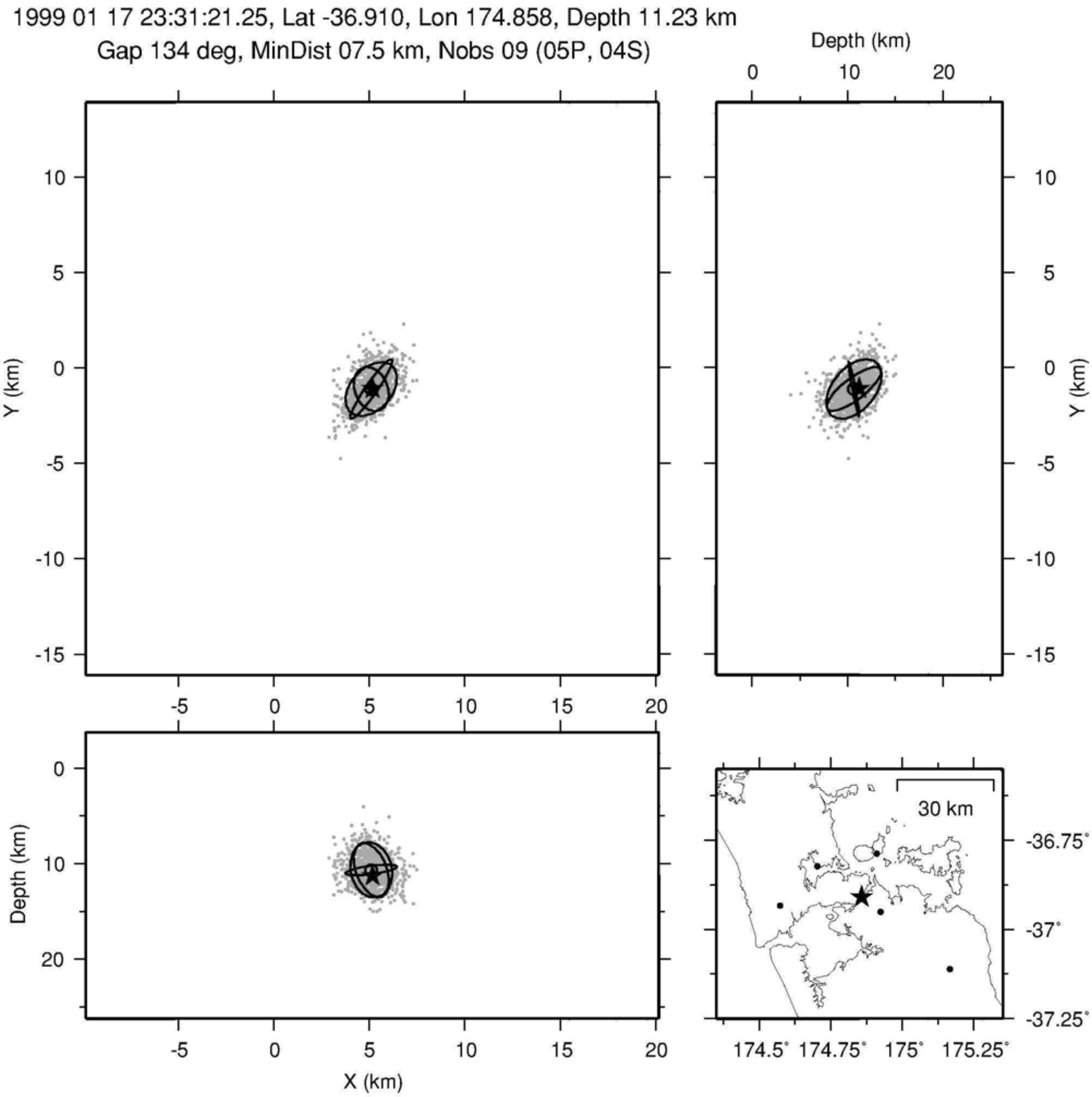

Fig. 2 Hypocentral uncertainty for a well-recorded earthquake located within the Auckland Volcanic Field (event 11 Table 1). The *x*-axis is oriented east–west and the *y*-axis north–south. Grey dots represent 1000 samples drawn from the solution probability density function. The star is the maximum likelihood hypocentre, while the open circle and ellipsoid represent the expectation hypocentre and 68% confidence limits, respectively. In this case, both hypocentral estimates are similar and the expectation hypocentre is covered by the maximum likelihood hypocentre. The lower right plot shows the maximum likelihood hypocentre (star) and seismographs that recorded the earthquake (black dots). Hypocentral parameters are shown at the top of the figure. The greatest angle without an observation (azimuth gap) is 134°, the distance to the closest seismograph is 7.5 km, and five P-phases and four S-phases were used in the location. The depth routinely calculated for this earthquake was 10 km.

seismicity, while mantle-derived nodules in the Auckland eruption products suggests that the total intrusion time from the upper mantle is 10–100 days. Another approach is to look at the depths at which earthquakes might occur.

Deep long-period (DLP) earthquakes have been observed at a number of volcanoes (Hasegawa & Yamamoto 1994; White 1996; Nakamichi et al. 2003; Ukawa 2004). These earthquakes are of lower frequency content than normal tectonic earthquakes and are typically located in the mid-lower crust or in the upper mantle where rocks would be expected to be ductile. In northern Japan, these earthquakes occurred in the lowermost crust or uppermost mantle, c. 30 km deep (Hasegawa & Yamamoto 1994); at Pinatubo they were 28–35 km deep (White 1996); at Iwate 30–35 km deep (Nakamichi et al. 2003); and at Fuji they were 10–20 km deep (Ukawa 2004). In most cases, they were displaced several kilometres from the summit of the volcano. In northern Japan, DLP earthquakes occur close to where there are low P-wave velocity zones in the upper mantle that are attributed to magma rising into the crust (Hasegawa & Yamamoto 1994). At Pinatubo and Iwate,

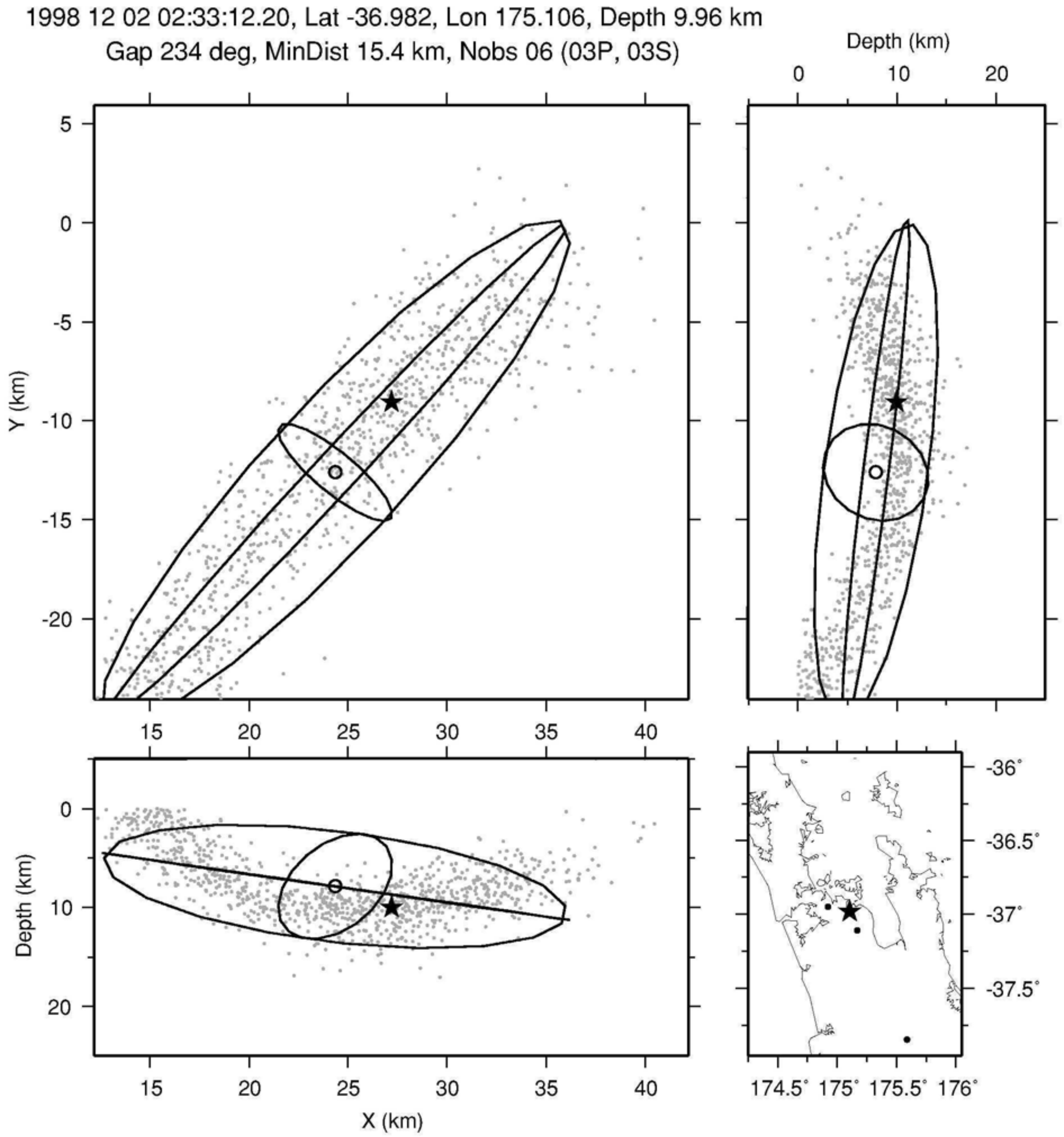

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Fig. 3 Hypocentral uncertainty for a poorly recorded earthquake located close to the Auckland Volcanic Field (event 10 Table 1). Refer to Fig. 2 for details. The greatest angle without an observation (azimuth gap) is 234°, the distance to the closest seismograph is 15.4 km, and three P-phases and three S-phases were used in the location. The depth was fixed at 5 km when this earthquake was routinely located by GNS Science. The uncertainty region for this earthquake is several times larger than that of event 11 (Fig. 2), and it is banana-shaped in cross-section.

the intrusion and/or degassing of basaltic magma has been suggested as a source for the DLP earthquakes (White 1996; Randall White pers. comm. 2000; Nakamichi et al. 2003). At Fuji, it was suggested that a change in the deep magmatic system, possibly associated with regional tectonic changes, triggered the earthquakes. Similar earthquakes were also recorded near the base of the crust on the downward extension of a seismically active fault in western Japan, distant from any active volcanoes (Ohmi et al. 2004). In this case, it was suggested that the movement of non-magmatic fluids may have produced the earthquakes. Although the mechanism of DLP earthquakes is not fully understood, we suggest that they may be the first sign of seismic unrest before an AVF eruption. Such DLP earthquakes might occur near the base of the crust, a depth of c. 30 km.

The brittle/ductile transition represents the depth (or depth range) in the crust at which rocks change the way they deform, from brittle failure (faulting) to ductile failure (creep). This depth thus represents the base of crustal tectonic earthquake activity. The transition depth depends on several factors including the mode of faulting, fault geometry, fluid pressure, geothermal gradient, rock composition, water content, and

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Table 1 Summary of earthquakes located in the Auckland region 1995–2005. All dates and times are in Universal Time, 12 h behind New Zealand Standard Time. M_L is earthquake magnitude, Gap is the greatest angle without an observation, and MinDist is the epicentral distance to the nearest seismograph that recorded the earthquake. Hypocentral uncertainties estimated from scatter plots (Fig. 2, 3) represent the maximum horizontal and vertical dimensions of the uncertainty regions.

No.	Year	Month	Day	Time	Latitude (°S)	Longitude (°E)	Depth (km)	Uncertainty position (km)	Uncertainty depth (km)	Quality	M_L	Felt	P-phase (number)	S-phase (number)	Gap (deg)	MinDist (km)	GNS ID
1	1995	Dec	20	0201	36.932	174.696	30	12	18	poor	2.3		3	3	153	11	892878
2	1996	Sep	17	0439	36.863	174.909	21	14	18	poor	1.9		4	3	188	8	1036621
3	1997	Jan	09	0918	37.093	175.137	09	17	15	poor	2.2		4	1	161	3	1070990
4	1997	Mar	06	0832	37.088	175.114	10	25	15	poor	2.3		3	2	193	5	1102260
5	1997	Oct	08	0153	37.121	174.978	11	13	16	poor	2.2		3	3	259	17	1182797
6	1998	Jan	27	2247	36.672	174.670	18	30	25	poor	2.2		3	2	337	25	1231874
7	1998	Apr	30	0847	36.870	174.893	09	15	12	poor	1.9		3	3	206	9	1265845
8	1998	May	15	1905	36.809	174.861	10	13	10	poor	1.9		3	3	280	5	1271950
9	1998	May	16	2126	36.716	174.984	06	07	10	good	2.7	Felt	6	3	126	10	1271959
10	1998	Dec	02	0233	36.982	175.106	10	30	15	poor	2.0		3	3	234	15	1336941
11	1999	Jan	17	2331	36.910	174.858	11	05	07	good	2.6		5	4	134	8	1364979
12	2000	Oct	01	1051	36.915	175.076	00	07	10	good	2.4		5	3	181	14	1625745
13	2001	Sep	28	1101	36.849	174.948	25	45	35	poor	1.6		2	2	229	22	1880914
14	2002	Sep	10	0455	37.133	174.973	09	20	10	poor	2.0		3	3	266	17	1920346
15	2004	Nov	30	1035	36.745	175.007	01	25	20	poor	2.0		3	1	279	10	2330041
16	2005	Nov	29	1035	36.805	175.085	08	06	08	good	2.5	Felt	6	2	137	16	2493467
17	2005	Nov	29	1037	36.813	175.092	06	15	10	poor	1.9		5	0	258	16	2493468
18	2005	Nov	29	1054	36.815	175.085	06	06	10	good	2.5	Felt	6	1	140	16	2493473
19	2005	Nov	29	1055	36.841	175.035	08	25	18	poor	1.5		3	1	281	13	2493474
20	2005	Nov	29	1701	36.809	175.100	05	08	08	good	2.2		6	1	141	17	2493575
21	2005	Nov	29	2208	36.828	175.097	08	04	05	good	3.3	Felt	11	3	115	17	2493672
22	2005	Nov	30	0759	36.806	175.104	06	07	10	good	1.7		5	1	166	17	2493855
23	2005	Nov	30	1045	36.801	175.105	03	06	08	good	1.9		5	1	140	17	2493907
24	2005	Dec	14	1754	36.814	175.107	00	05	08	good	2.4	Felt	5	4	162	18	2499692

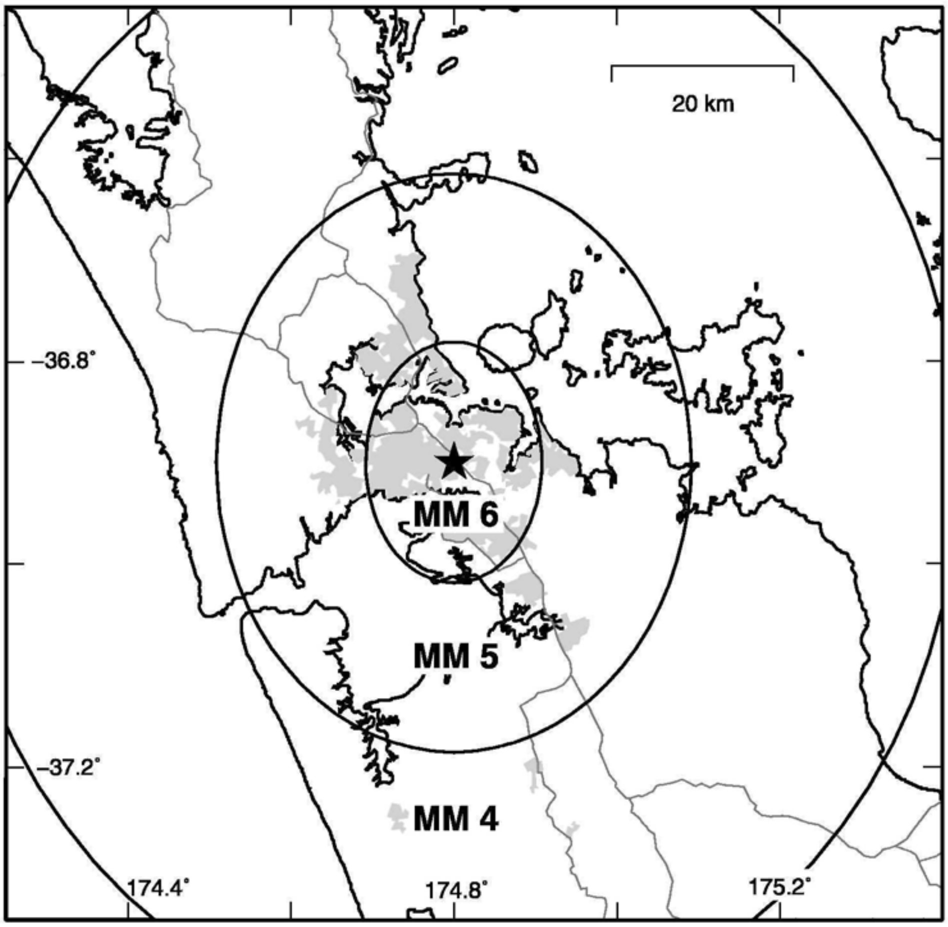

Fig. 4 An isoseismal map showing the expected felt intensity for an earthquake of M_L 4.5 at a depth of 10 km beneath the Auckland Volcanic Field based on the model of Dowrick & Rhoades (1999). The star represents the earthquake epicentre. The MM 6 isoseismal covers one-half of the Auckland urban area (grey shade) and the MM 5 isoseismal extends up to 30 km from central Auckland. Within the MM 6 isoseismal most of the population would feel the earthquake. The main roads are also shown.

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strain rate (Sibson 1984). All of the reliably located earthquakes are <15 km deep (Table 1). Heat flow data from a single well in the Auckland area suggests that a temperature of 300–350°C, representative of the temperature at the brittle/ductile transition in quartz-dominated crust (Sibson 1984), would be reached somewhere between 14 and 23 km depth (Rob Funnell pers. comm. 1999). We assume that the brittle/ductile transition in the Auckland region is c. 15–20 km.

Above the brittle/ductile transition, the stresses generated by the intruding magma should produce tectonic earthquakes by brittle failure. As earthquakes may not only occur in the immediate vicinity of an intrusion, they may have to be significantly shallower, possibly as shallow as 5 km, before they are useful for estimating the likely location of an eruption.

If we estimate the ascent time of Auckland magmas to be 10–100 days—assuming that the ascent rate is constant with depth, that DLP earthquakes may occur at c. 30 km depth, and tectonic earthquakes above 15–20 km depth—then the first seismic precursors will be c. 3–30 days before an eruption (DLP earthquakes), and tectonic earthquakes may occur c. 2–20 days before an eruption. The large range in these values is a result of the uncertainty in the ascent rate of AVF magmas. The lower ends of these estimates are significantly shorter than the observations at Jorullo, Parícutin, and Teishi Knoll, but the higher ends are within a factor of 2 or 3 of those observations. Our best estimate of the length of pre-eruption warning is therefore a few days to a few weeks.

Felt earthquakes

Observations from Parícutin and Teishi Knoll suggest that earthquakes as large as M_L 4.5–5.5 could occur before an eruption in the AVF. Modelling the felt effects of an earthquake of M_L 4.5 at 10 km depth beneath Auckland (Dowrick & Rhoades 1999) gives a maximum felt intensity of MM 6 over an area of c. 350 km^2 (Fig. 4). At MM 6, an earthquake is sufficiently strong to be felt by everyone, for objects to fall from shelves, and for appliances to move on bench or table tops (Dowrick 1996). In the one and a half months before the eruption of Parícutin, there were 11 earthquakes of $M_L \geq 4$. If a similar rate and magnitude of seismicity preceded an eruption of the AVF, then people would be acutely aware of the ongoing volcanic unrest.

Ground deformation monitoring

Ground deformation monitoring is another commonly used method of monitoring active volcanoes (Murray et al. 2000). Continuous GPS is widely used by the Geonet project (www.geonet.org.nz) to detect deformation at volcanoes in New Zealand, but its application in Auckland is difficult because the AVF is large and the location of the next eruption is unknown. Miller et al. (2001) suggested that once seismicity has been recorded, kinematic GPS measurements could be made at c. 1 km intervals to identify any deformation. If this was successful, then a network of continuous GPS receivers could be installed for real-time monitoring.

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A technique that has the potential to monitor deformation over the whole AVF with a minimum resolution of c. 3 cm is Interferometric Synthetic Aperture Radar (InSAR, Dzurisin et al. 2006; Wicks et al. 2006). Stevens et al. (2004) determined that an intruded volume of 0.01 km^3, a typical erupted volume for the AVF (Allen & Smith 1994), would be detectable by InSAR only shallower than 5 km, but that a volume of 0.1 km^3, typical of larger AVF eruptions, should be detectable at a depth of 20 km. The difficulty with InSAR is that the satellite that provides these data has a 35 day orbit period, possibly too long to collect repeat data if there is rapidly ascending magma (Stevens et al. 2004).

Better interpretation of seismic precursors

Studies of geology, eruption products, and eruption processes in Auckland have provided a reasonable understanding of the history of the AVF and what might happen once an eruption commences (Heming & Barnet 1986; Houghton et al. 1991; Huang et al. 1997; Cassidy et al. 1999; Shane & Smith 2000). However, knowledge of crustal structure is limited, and this could significantly handicap the interpretation of precursors to an eruption.

Fundamental to locating earthquakes in Auckland accurately is a realistic local velocity model. Regional-scale models exist (Stern et al. 1987; Horspool et al. 2006), but there is no model that includes near-surface geology, and this can have a significant effect on earthquake locations. There is a northeast–southwest lineation in some of the volcanic vents in the AVF, the same direction as some basement faults, and this suggests that in some cases basement structures may influence vent locations. Once magma is within a few kilometres of the surface, knowledge of the location of faults might help in estimating the eruption site. Another area needing research is the thermal structure of the crust. This is important as it is one of the main parameters that influence the maximum depth at which tectonic earthquakes will occur.

The high level of man-made ground noise in Auckland City makes operating a seismograph network difficult. The Geonet project is currently upgrading the AVSN and is considering two options to try to negate this problem. One is to increase the number of seismographs beyond the five currently used, and the other is to put sensors in deeper boreholes.

SUMMARY

The AVF is monitored by a network of five telemetered, vertical-component, short-period seismographs, which have been used to locate 24 earthquakes between 1995 and 2005 (M_L 1.6–3.3). All were high-frequency, tectonic earthquakes. Less than one-half of the earthquakes were well located, because of a high level of noise close to the city and low earthquake magnitudes. Seismic precursors to eruptions from historically active volcanic fields suggest that precursory earthquakes in the AVF could occur as little as 2 weeks before an eruption, and the largest earthquakes could be felt widely in Auckland City. By estimating the depth of the brittle/ductile transition and using previous estimates of the ascent rate and source depth of AVF magmas, we calculate that the first seismicity could occur as little as a few days before an eruption in the AVF. Our best estimate is a few days to a few weeks. Effective monitoring of eruption precursors in the AVF requires not only an effective monitoring network, but also a reasonable knowledge of the geological structure of the Auckland area, something which is currently lacking.

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