The seismicity of the 2009 Redoubt eruption

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1. Introduction

Redoubt Volcano is a 3108 m high stratovolcano in the Cook Inlet region of south-central Alaska that has erupted three times since the mid 1960s (Schaefler, 2012; Bull and Buurman, 2013). Given its recent eruptive history, its location near communities, oil platforms, an oil storage facility, and its potential impact to air traffic routes, Redoubt Volcano was closely monitored when unrest began in summer 2008. The seismic network at that time consisted of 5 single-component and 2 3-component L-4 and L-22 model telemetered short-period seismometers within 25 km of the vent, operated by the Alaska Volcano Observatory (AVO) (Fig. 1). As the level of unrest increased the network was augmented: two additional telemetered broadband Guralp 6TD instruments and a telemetered single-component short-period L-4 seismometer were installed in late February 2009 and 4 campaign-style broadband Guralp 6TD seismometers with on-site recording were deployed in the 2 days prior to the magmatic explosions that occurred in March 2009.

In this paper we present an overview of the seismic activity that was associated with the 2009 eruption of Redoubt Volcano. There are many aspects to the seismicity both prior to and during the eruptive episode, including swarm activity, tremor episodes, seismicity from explosion signals and lahars and variations in the background hourly earthquake rates. When referring to explosion events, we follow the numbering scheme used by Schaefer (2012) who numbers them 0–19. Our objective is to place each set of seismic patterns in the context of other geological and geophysical observations. As a result, this paper encompasses a wide variety of seismic signals that were generated by a range of volcanic processes. For organisational simplicity we include brief discussion and speculation of the seismic sources within the individual sections instead of in a lengthy discussion section at the end, and close with a brief eruption summary that encompasses the major conclusions drawn from the seismic record. We begin with a short eruption overview to provide context for our seismological interpretations.

2. Eruption overview

Retrospective analysis of continuous GPS data indicates that inflation began as early as May 2008 (Grapenthin et al., 2013), but the earliest signs of unrest at Redoubt Volcano recognised by AVO were reports by field geologists working on the edifice of H2S odors from fumaroles near the ice-covered 1990 lava dome in July 2008. Brief bursts of tremor in the 2–6 Hz range were recorded in September 2008 coincident with reports from local part-time residents of explosion-type noises in the vicinity of the summit,
and in late September crevasses began to expand in the upper Drift Glacier (Bleick et al., 2013; Schaefer, 2012). Continued enlargement of these ice fractures, combined with increased and anomalous gas emissions (Werner et al., 2013) prompted AVO to increase the Volcano Alert Level and Aviation Color Code to advisory/yellow below the edi.

Earthquakes began in December 2008 at depths between 28 and 32 km (Werner et al., 2013) prompting AVO to increase the criteria discussed in Section 3.

Table 1

<table>
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<th>Swarm</th>
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<th>Total earthquakes</th>
<th>Maximum earthquake rate (per h)</th>
<th>Maximum repeating earthquake rate (per h)</th>
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<td>7470</td>
<td>191</td>
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Fig. 1. Map of the Redoubt seismic network that was operational during the 2009 eruption. The inset shows the location relative to Alaska. Stars indicate campaign-style seismometers with on site recording, and circles indicate the telemetered seismic stations used for monitoring during the eruption.
July 2009, when the eruption was declared over (Bull et al., 2013; Diefenbach et al., 2013).

3. Data

We combine earthquake information from multiple catalogues to get the best qualities from each. Discussions that rely on hypocenter location are based on an analyst-reviewed catalogue (Dixon et al., 2010). This catalogue provides the highest quality depths and locations and has a magnitude of completeness of 0.4. Most of the earthquake analyses in this paper are based on bulk processing of hundreds to thousands of events. For these analyses we prefer the temporal completeness of an algorithm-based (i.e., automated) catalogue with a lower threshold for inclusion. Though the errors of the hypocenter solutions are much larger in the automated catalogue, the algorithm-based approach identifies 37,000 earthquakes between January and May 2009 compared to the 3766 analyst-reviewed solutions during the same period.

The automated catalogue is based on traditional single channel earthquake detections using short- to long-term signal ratios. These detections are compared across nearby stations and associated into events when they are consistent with P phase travel times from a pre-computed grid of trial locations. The short- to long-term signal ratios are computed in two frequency bands (0.8–5 Hz and 3–25 Hz) to detect both low frequency and volcano–tectonic earthquakes. We require events to register P-wave arrivals at 4 stations for inclusion in the catalogue. A thorough description of the methodology can be found in Thompson and West (2010). The eruption caused significant outages at RSO — a station critical to both earthquake catalogues. When RSO was not operational, most notably from March 23 to April 16, the smallest earthquakes registered only on 3 stations and as a result did not meet the criteria for inclusion in the catalogue. At all times, however, there is sufficient station coverage that events of magnitude —0.9 are generally included in our analyses.

Our analyses include classifying earthquakes as ‘repeating events’. We assess this using cross correlation-based clustering techniques applied to waveforms from one representative data channel. For each event we segment a 7 s seismogram beginning 1 second before the P-wave and filtered using a 4-pole Butterworth filter between 0.5 and 25 Hz. Each event is then cross correlated against all other events. We use hierarchical clustering to group the cross correlations into event families. Within each family all events have an average cross correlation value with all other events of 0.75 or greater. This method is described further in Buurman and West (2010). Because of the large number of earthquakes, we cross-correlate the catalogue in groups of 500 consecutive events. If an event is part of a family of four or more members we consider it a ‘repeating event’, in line with other studies (Buurman and West, 2010; Thelen et al., 2010).

4. Swarms

4.1. Method

In this section we analyse earthquakes using methods that allow direct comparison between the different swarms. Earthquake swarms are defined as increases in earthquake rates within a given volume over a relatively concentrated period of time without a single outstanding shock (Mogi, 1963). We quantify this rather loose definition using our own criteria, identifying swarms as episodes during which the hourly rate of earthquakes exceeds 50, or when the hourly rate of repeating earthquakes exceeds 20 (Fig. 2). Swarm onsets are identified as the time when the hourly rate of earthquakes exceeds the previous six-hour average, and swarms are considered over when the hourly rate returns to the stable mean background rate for the following 6 h. We find that a six-hour average accounts for natural fluctuations in seismicity. Six swarms are identified using these criteria. We refer to each of the six swarms by the UTC date when the activity began. For analyses that rely on waveform characteristics, we use data from short-period station REF unless otherwise noted. This is one of the closest stations to the vent and it operated throughout the entire period of unrest. Because of its close proximity to the vent, this station recorded lower amplitude activity in the summit region but was susceptible to clipping during the most energetic seismicity. Most of the analyses that follow do not require unclipped data, although where data are clipped we refer to broadband station RDWB.

4.2. February 26 swarm

The first swarm began on February 26, 2009, 25 days before the first magmatic explosion. The onset was sudden and occurred just 6 h following the end of a 3-week long tremor episode. Initial earthquake activity peaked at 91 events per hour but this high rate
<table>
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<th>Lahar number</th>
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<th>RDE time since last explosion (min)</th>
<th>Follows explosion?</th>
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<th>Onset error (min)</th>
<th>DFR end time (UTC)</th>
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<th>Follows explosion?</th>
<th>Max amp (nm/s)</th>
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Max amp: mean amplitude in a 2 minute window around the peak signal to noise ratio.
Onset: time when the signal to noise ratio begins to increase again following the explosion.
End: time when the signal to noise ratio drops below 2.
Onset error: visual estimate.
End error: Number of minutes for the signal to noise ratio to drop from 2 to 1.
quickly diminished and the activity continued to fluctuate around 30 events per hour for the remainder of the swarm, which lasted 29 h in total. The activity ended abruptly on February 27; however, the background rate of earthquakes that followed remained higher than it had been through most of January (Fig. 2). This elevated rate was sustained until the onset of the next swarm on March 20.

The majority of earthquakes during the February 26 swarm are small (<Ml 1), although there is some scatter in size particularly early in the swarm and again during the last 12 h evident in Fig. 3A. Most of the events are dissimilar, and only 6% meet our criteria for repeating events (described in Section 3) (Fig. 4A). The waveforms vary greatly in shape, with some events exhibiting P- and S-wave arrivals and others showing more emergent onsets. For all events, however, the majority of energy in the spectrum occurs between 1 and 7 Hz at station REF. These waveform characteristics are not typical for volcano-tectonic earthquakes, which usually have impulsive P-wave arrivals and are dominated by frequencies between 5 and 15 Hz (Lahr et al., 1994). While volcano–tectonic earthquakes are regularly recorded at Redoubt Volcano, they rarely occur as swarms; waveforms with lower dominant frequencies and more emergent P-wave arrivals such as those recorded during the February 26 swarm are historically more typical (Lahr et al., 1994).

The February 26 swarm marked the transition from 3 weeks of seismicity dominated by tremor to a further 3 weeks of seismicity dominated by a higher rate of detected earthquakes, suggesting that a large-scale change had occurred below or within the edifice. Our interpretation for this change is that the gas movement that had been generating the tremor became blocked, leading to higher pressures and gas-driven cracking. The closed volcanic system then produced a higher rate of background seismicity as the fluids continued to enter the system but could not reach the surface. Werner et al. (2013) reported the lowest measurements of SO2 emissions during the precursory build-up to the eruption on February 27. Although these data lack the temporal resolution to confirm the closing of the gas system, their results support this model.

4.3. March 20 swarm

The second swarm began on March 20 and culminated in a series of magmatic explosions (events 1–6). The swarm lasted 66 h and had two pulses: a first smaller pulse that peaked at 45 events per hour between March 20 and 21; and a second pulse beginning late on March 21 which peaked at 82 events per hour. The first pulse increased gradually over 10 h, while the second pulse had a rapid onset (Fig. 3B) that prompted AVO to increase the Volcano Alert Level and Aviation Color Code to watch/orange. Volcanic tremor occurred towards the end of the second pulse, increasing in amplitude until it dominated the seismic record shortly before the magmatic explosions.

The majority of earthquakes during both pulses are small (Ml < 0.8), repeating earthquakes that cluster into three main families...
The smallest of these families occurs exclusively during the first pulse of activity, and is characterised by waveforms with peak frequencies around 5 Hz, broad spectral content and visible P- and S-wave arrivals (Fig. 3B). The second and largest family has low rates during the first pulse but later increases significantly to dominate the second pulse of seismicity, dying out with the onset of the last family. One of the most striking features of this second family is the waveform evolution that occurs over its duration, where the

(Figs. 3B and 4B). The smallest of these families occurs exclusively during the first pulse of activity, and is characterised by waveforms with peak frequencies around 5 Hz, broad spectral content and visible P- and S-wave arrivals (Fig. 3B). The second and largest family has
waveforms gradually change shape, dropping from a peak frequency of around 5.1 Hz to 4.4 Hz. The third family differs considerably, exhibiting peak frequencies around 2.5 Hz with no clear S-wave arrivals (Fig. 3B). This type of earthquake is often referred to as a long-period or ‘LP’ event, and is characteristic of the seismic swarms during the 1989–90 eruption (Chouet et al., 1994; Stephens and Chouet, 2001; Power et al., 2013). The magnitudes of these events are also distinct: where the maximum amplitudes vary widely within the first two families, the amplitudes in the last family are tightly clustered, growing steadily until they become dwarfed by the volcanic tremor in the final hours before the eruption onset (Fig. 3B).

The presence of different earthquake families during the swarm indicates that several different processes were occurring within the edifice in the final hours before the eruption. The clear P- and S-wave arrivals present in the first two families indicate that they were generated by brittle failure within the edifice, and their repetitive nature suggests that they originated in approximately the same location. Given the timing of these earthquakes prior to the eruption onset, it is likely that they were generated from the incremental opening of cracks ahead of the rising magma body. When a crack opens slowly the shift in hypocenter location (i.e. the migration of the crack tip) is small and will not necessarily be obvious in the earthquake locations. The waveforms in the last earthquake family are much lower in frequency and were generated by a very different source. The timing of this family coincided with the growth of the first lava dome: the family emerged around 1700 UTC and the dome was observed in satellite images at 2000 UTC on March 22 (Bull et al., 2013). It is therefore likely that a process related to dome growth generated this family of earthquakes, and that the onset of the family occurred once the conduit had been widened enough to allow the magma to move through the shallow edifice. We speculate that as the earthquakes increased in amplitude, the rate of extrusion increased which in turn generated tremor through rapid degassing. Once the extrusion rate became rapid enough that the magma could not equilibrate through degassing, the explosive phase of the eruption began.

Fig. 4. Cross correlation plots for the 6 Redoubt swarms. Each pixel represents a cross correlation pair, where the colour represents the maximum cross correlation value between the two events. Time progresses top to bottom and left to right in each panel, and in each plot the diagonal is equal to 1 (the auto-correlation, highlighted in A). A) February 26 swarm, B) March 21 swarm with the 3 largest families circled, C) March 27 swarm, D) March 29 swarm, E) April 2 swarm, F) May 2 swarm.
4.4. March 27 swarm

The third swarm of the eruptive sequence lasted only 8.5 h and occurred prior to the 9th explosion of the sequence on March 27 (Schaefer, 2012), less than six hours after the previous explosion on March 26. Activity increased suddenly within the hour shortly before 00:00 UTC on March 27 and the event rate remained around 50 events per hour for the majority of the swarm, reaching a peak of 92 events per hour in the 2 h before the explosion. This sequence represents the most powerful swarm of the 2009 eruption. The majority of earthquakes have local magnitudes of 1.5, and the largest event of the sequence is a Ml 2.6 earthquake that occurred late in the swarm. Although almost all the events in the sequence clipped at the summit stations, the events remained on scale at broadband station RDWB, 10 km to the west of the vent (Fig. 1). The swarm was dominated by a single event family characterised by an impulsive P-wave arrival and a definitive S-wave arrival (Figs. 3C; 4C). Event amplitudes steadily increased over the first 6 h of the swarm from Ml 0.7–1.6, before rapidly decreasing in magnitude. As the event size decreased, so did the time between events until they merged into tremor. During the final minutes before the explosion the frequency of the tremor glided exponentially as a function of time from less than 1 Hz to 10 Hz. Hotovec et al. (2013) model these gliding tremor episodes as regularly repeating stick–slip earthquakes that are tightly clustered in time and space. The episode of gliding tremor that followed the swarm was the first of several instances of this phenomenon that occurred during the explosive activity on March 27–29 (events 9–18; Hotovec et al., 2013). Fee et al. (2013) noted a difference in the infrasonic explosion signal characteristics during this same period and attributed it to a change in eruptive style, from the subplinian-type activity observed during the early phase of the eruption to vulcanian activity. Wallace et al. (2013) also observed a change in the tephra componentary, finding much finer grain sizes indicative of more explosive activity following the March 27 swarm. These observations each support a change in the eruptive behaviour around March 27, and the timing of the March 27 swarm strongly suggests that it heralded the shift in activity. Hotovec et al. (2013) examined the March 27 swarm closely and concluded that the location and focal mechanism solution for the repeating events indicated stick–slip behaviour along the conduit walls. A change in the viscosity of the ascending magma could explain why this type of activity was not observed earlier in the eruption, and could also account for the differences observed in the explosion characteristics compared to the earlier sequence on March 23–24. However this cannot be corroborated with other evidence.

4.5. March 29 swarm

This hour-long swarm occurred just hours after event 18 on March 29 and preceded an episode of high amplitude tremor and continuous but weak eruptive activity (Schneider and Hoblitt, 2013). In addition to the 37 events of sufficient size for the automated catalogue, there are a few hundred smaller events that can be observed on summit stations (Ketner and Power, 2013). The events are generally small, lack clear S arrivals (Fig. 3D), and have a high degree of waveform similarity (correlation > 0.9, Fig. 4D). Satellite data indicate that the third lava dome emerged on March 29 (Bull et al., 2013). Our interpretation for this swarm is similar to our interpretation for the 3rd main family of waveforms in the March 20 swarm: directly related to dome growth. As a fresh plug of magma ascended, it is possible that friction in the conduit created stick–slip earthquakes. The decreasing amplitude of the earthquakes suggest that once the conduit was widened, the shear mechanism weakened as dome growth became continuous and, eventually, aseismic. Although the waveforms between the March 23 events and the March 29 swarm are dissimilar, the 18 large explosions that separated these two episodes may have changed the conduit geometry enough that the same process occurred in a different location, resulting in the dissimilarity between the two swarms.

4.6. April 2 swarm

This swarm preceded the April 4 explosions that produced the highest ash column of the eruption (Schaefer, 2012). The swarm had a rapid onset (~2 h) and lasted 43 h. The peak of 107 events per hour occurred in the middle of the swarm and declined over the next day before the explosion (Fig. 3E). This decreasing event rate contributed to a decision by AVO staff to downgrade the Volcano Alert Level and Aviation Color Code to warning/orange on April 3. 77% of the events were repeating, dominated by a family of emergent low amplitude waveforms that evolved considerably over the course of the swarm (Fig. 4E). During the swarm the dominant spectral peak at 2.7 Hz became gradually stronger while energy above 7 Hz became weaker. Although the swarm contained nearly 2000 earthquakes, none of the events were of sufficient size to be located in the analyst-reviewed catalogue.

Visual observations indicate that dome growth stalled during the swarm (Bull et al., 2013; Diefenbach et al., 2013). The change in magma effusion rate could have been caused by a change in the magma supply rate, an increase in viscosity or an equilibration of the pressure in the conduit. This transition is seen in the steady evolution of seismic waveforms. In light of the stalled dome growth, the evolving seismic signature probably reflects changing material properties near the source of the earthquakes, though we also cannot rule out a changing source location or a change in the bulk properties of the edifice.

4.7. May 2 swarm

The final swarm was the longest-lived (five days), and occurred well after the explosive activity had ceased during a period of continuous dome growth. Earthquake amplitudes are markedly smaller than the other magmatic swarms, with 7400 events of sufficient size for the automated catalogue (Fig. 3F). Ketner and Power (2013) examine the larger population of events too small to be recorded beyond the summit stations. Though the earthquakes are generally small, they occur at higher rates (up to 191 events per hour) than any of the other swarms (Table 1). Waveforms throughout this swarm are highly repetitive and emergent with dominant frequencies near 3.5 Hz (Fig. 4F). Early on May 6, two additional waveform families were detected, concurrent with the increase in event rate and amplitude of the main repeating family. The families all share waveform characteristics, and likely originate in the same region. Several measures point to a change in the volcanic system in conjunction with the May 2 seismic swarm. Gas measurements show an increase in SO2 and CO2 around May 4 (Lopez et al., 2013; Werner et al., 2013). Lava dome samples show a change in vesicularity and texture (Bull et al., 2013), and the effusion rate increased following the May 2 swarm (Diefenbach et al., 2013). These factors together suggest that there may have been an influx of a different batch of magma. The earthquakes could have been the result of failure around the edges of the extruding lava dome as new magma entered the system below, similar to the model proposed by Iverson et al. (2006) during the Mount St Helens eruption in 2004–2008. Alternatively they may have been generated in response to the influx of different magma into the upper conduit. The different lava textures observed in the samples indicate that the new material may have had different properties that caused it to move differently through the conduit. The swarm may have reflected slip–stick at the edges of the magma
plug as the dome adjusted to the new extrusion rate. There is insufficient data with which to be able to distinguish between these models, however, and we instead leave them both as possible sources for the May 2 swarm.

5. Volcanic tremor

The first episode of tremor associated with the 2009 eruption occurred in late September 2008, when several bursts were recorded at the summit stations. These events had durations less than 2 min, and were dominated by frequencies between 1 and 4 Hz. Aside from these brief events in September, no more tremor was detected until January 2009.

Beginning in late January, tremor featured regularly in the seismic record. In lieu of a lengthy chronology, we examine the notable styles of tremor that occurred during the eruption of Redoubt Volcano. Our waveform analyses are based on station REF for its high fidelity and continuity of operation (Fig. 1).

Tremor amplitudes are reported using surface-wave reduced displacement denoted as $D_{RS}$ (Fehler, 1983). Each $D_{RS}$ value is the root-mean-square displacement of a one-minute window measured on the vertical component of station REF, then corrected for geometrical spreading (assuming surface waves, and a wavelength of 1 km). No corrections for attenuation or site effects are made. Note that in Fig. 2 we plot a downsampled version of this 1-minute $D_{RS}$ timeseries: we plot the median value for each hour. We do this because, in the figure, we wish to emphasize the (continuous) tremor amplitude and filter out transient signals including earthquakes and noise spikes. As a result, any tremor bursts which lasted less than 30 min cannot be seen in Fig. 2. Also note that McNutt et al. (2013) use a different $D_{RS}$ methodology more suited to the (transient) explosion signals.

5.1. Late January: high amplitude precursory tremor

Tremor activity began in earnest at 1810 UTC on January 24, 2009 as a 4-minute burst of broadband tremor (energy up to 10 Hz) with an amplitude of 16 cm$^2$. At 1000 UTC on January 25 there was a gradual onset of tremor, which increased from 0.4 cm$^2$ to >2 cm$^2$ at 1047 UTC. Amplitudes of 2–5 cm$^2$ were sustained until around 1500 UTC and prompted AVO staff to increase the Volcano Alert Level and Aviation Color Code from advisory/yellow to watch/orange (Schafer, 2012). The January 25 tremor (Fig. 5A) was dominated by frequencies between 2.5 and 6 Hz, although there was some energy as high as 18 Hz. Examination of the continuous data at the summit stations showed that this tremor consisted of closely spaced low frequency earthquakes which contained some energy as high as 18 Hz. Examination of the continuous tremor with amplitude 1–3 cm$^2$ lasting 3 h.

There were several more bursts of tremor over the following days. The most significant were 3 bursts between 1930 and 2200 UTC on January 30 exceeding 5 cm$^2$. This episode was strong enough to be recorded at the seismic networks on the nearby Iliamna (54 km) and Spurr (94 km) volcanoes. The final burst was followed by continuous tremor with amplitude 1–3 cm$^2$ lasting 3 h.

Shallow volcano–tectonic earthquakes began suddenly on January 25 coincident with the tremor episodes, Werner et al. (2013) also report an increase in $SO_2$ flux during late January and early February. We interpret these notable changes in seismicity and gas flux as evidence for magma intrusion into the shallow crust. Inflation beginning in May 2008 indicates that a small volume of magma moved into the mid crust between 5 and 13 km below sea level (Grapenthin et al., 2013). This ascent appears to have been aseismic, since there were no changes in the seismic record at that time. The upick in activity in January can be interpreted as the ‘renewed’ ascent of magma from the mid crust to shallower depths between 3 and 6 km. This model accounts for the rapid onset of seismicity in January by the arrival of magma in the shallow crust. The earthquakes are generated by the fracturing of rock below the edifice in response to the intruding magma and the tremor is a result of the increased degassing of the shallow magma body.

5.2. February 2–March 20: sustained tremor

A new tremor sequence beginning on February 5 had a very different character. The spectrum was dominated by frequencies between 2.5 and 5 Hz with a peak at 2.9 Hz, representing a much narrower spectrum than the tremor in late January (Fig. 5B). The February tremor varied from a steady, low-amplitude-type tremor with $D_{RS}$ 0.2 cm$^2$ to 12 h periods of slightly more broadband (frequencies of 1–5 Hz), higher amplitude activity peaked between 3 and 4 Hz with $D_{RS}$ up to 1 cm$^2$. This style of tremor continued for 20 days (Fig. 2) and ended abruptly on February 26 with a vigorous (10 cm$^2$) 5 min burst of broadband tremor containing energy above 10 Hz.

We interpret the source of the sustained tremor as hydrothermal. Leet (1988) showed that low-amplitude (<5 cm$^2$) tremor can be generated by steam bubble growth in water as a result from heat transfer from the surrounding rock. McNutt (1992) showed that the amplitude of volcanic tremor scales with eruption intensity, and that the lowest tremor amplitudes (between 0.05 and 5 cm$^2$) can be attributed to hydrothermal activity. Given that no magma had arrived at the surface, and that the values of reduced displacement (0.2–1 cm$^2$) are relatively low compared to other volcanic settings (volcanic tremor can reach extreme amplitudes of 100,000 cm$^2$ according to McNutt, 1992), it is likely that the sustained precursor tremor was generated by boiling in the shallow hydrothermal system. Although the January tremor was much higher in amplitude and contained higher frequencies than the February tremor, the episodes shared similar spectral peaks at 2.9 and more weakly at 1.9 Hz.

At 2100 UTC on March 15 there was a 3 h episode of low amplitude (0.3 cm$^2$) tremor that coincided with the first phreatic explosion of the eruptive sequence at 2123 UTC ($D_{RS}$ ~3 cm$^2$).

5.3. Explosion tremor

The magmatic explosions of late March and early April were accompanied by high amplitude tremor that remained sustained for periods of hours to days. The explosion tremor was more broadband than the sustained precursor tremor, with energy spread across the spectrum up to 15 Hz during the vigorous episodes (Fig. 5C), and up to 9 Hz during quieter periods. The high amplitude explosion tremor had a broad spectrum with the majority of energy concentrated between 1.5 and 7 Hz, and had two main peaks at 1.8 and 2.8 Hz.

The first and most vigorous episode of explosion tremor followed the closely spaced explosion events 4 and 5 on March 23. This episode produced sustained $D_{RS}$ of 2–5 cm$^2$ for a period of 9 hours, after which the activity became more spasmodic in character. The spasmodic tremor continued for 5 h before ceasing abruptly prior to event 6. This explosion tremor followed explosive eruptions, suggesting that the tremor was generated by the vigorous degassing that also followed the eruptions. If this is true, then this tremor was a direct manifestation of the degassing of the magma which remained in the conduit after the explosions. Similar models have been proposed by Neuberg et al. (2000) who examined tremor associated with explosions at Soufriere Hills Volcano.
5.4. Pseudo-explosion tremor

The final episode of high amplitude tremor followed the March 29 swarm. We label this tremor as ‘pseudo-explosion tremor’ because it occurred within the explosive episode of the eruption but did not follow explosive activity. More spasmodic in character, this tremor was similar in amplitude and frequency content to the March 23 explosion tremor (Fig. 5D). It generated $\Delta V$ in the range 0.5–6 cm$^3$ and lasted for 20 h, before changing in character to short-lived bursts lasting less than 30 s at much lower amplitudes for a further 48 h before the onset of the April 2 swarm. The pseudo-explosion tremor spectrum was much more sharply peaked than the explosion tremor spectrum. It also shared several peaks with the late January tremor. Based on the other available data during this period, it is unclear if there was any ongoing volcanic activity that might have generated the pseudo-explosion tremor. $SO_2$ emission was relatively low (Lopez et al., 2013) and there was no identifiable infrasound signal (D. Fee, personal communication). However, 2 of the 3 lahars that did not follow explosive activity were observed during this episode of tremor, indicating that there was enough activity occurring at the vent to generate a debris flow. It is also notable that growth of the 3rd lava dome was first recorded during this period (Bull et al., 2013), and the emergence of the new lava dome may have melted ice from the crater glacier that triggered the lahar events. Lightning was detected on two occasions around the time that the lahars were generated, suggesting low-level ash emission was occurring (Behnke et al., 2013). These observations suggest that magma and/or gas were actively venting at this time, and that pseudo-explosion tremor is probably a manifestation of that process.

5.5. Swarm tremor

5.5.1. The March 20 swarm: increase in continuous background tremor

Tremor was associated with 2 of the 6 seismic swarms during the 2009 unrest period, but had a very different character during each episode. The first occurrence of ‘swarm tremor’ occurred near the end of the March 20 swarm (Fig. 5E). As the swarm progressed there was a gradual increase in continuous background tremor as well as occasional bursts of higher amplitude tremor that lasted several minutes. The tremor progressively increased and became

Fig. 5. Spectrogram examples of the different types of tremor recorded at station REF during the eruption. The location of REF is shown in Fig. 1.
5.5.2. The March 27 swarm: earthquakes merging into tremor

The March 27 swarm produced a different type of swarm tremor. Instead of a gradual increase in background tremor, the volcanic–tectonic earthquakes in the swarm became progressively closer in time, merging into a continuous signal (Fig. 5F). The tremor spectrum for this episode is broad (like the earthquakes that comprise it), contains energy between 1.5 and 9 Hz and is characterised by sharp peaks of similar amplitude, many of which were shared with the different episodes of tremor examined in this section. After 10 min of this steady tremor, the dominant frequencies rose exponentially with time to 10 Hz before abruptly stopping. A minute-long pause followed the end of the tremor before explosion event 9 occurred. This marked the first of several periods of ‘gliding tremor’ prior to explosive eruptions between events 9 and 18. Hotovec et al. (2013) examine these sequences of events in detail, modelling them as accelerating failure at the edges of an ascending magma plug in the shallow conduit.

5.6. Comparing tremor with explosion signals

Fig. 5G shows the explosion on April 4 for comparison with the different types of tremor. Explosion signals are distinguished from tremor signals primarily through their high amplitude and broad frequency content. Energy between 1 and 9 Hz dominates the spectra although significant energy continues above 20 Hz. These signals are generated during continuous ash emission. It is likely that the variations in the signal strength reflect variations in the rate of sustained emission (McNutt and Nishimura, 2008), although we do not observe this directly. McNutt et al. (2013) and Schneider and Hoblitt (2013) compare the explosion signals to plume height, infrasound and lightning in order to examine different aspects of the eruptive activity. We include an example here to illustrate the differences between tremor and explosion signals.

6. Explosion seismicity

Several authors have addressed details of the explosive eruptions at Redoubt Volcano (Fee et al., 2013; Haney et al., 2013; McNutt et al., 2013). The seismic signals associated with these explosions vary greatly. Our objective here is to distil the explosions to simple parameters that can be put in context with the swarms, tremor and lahars. To accomplish this we use total seismic energy, seismic energy in high and low frequency bands, peak amplitude from reduced displacement, and duration.

Energy is estimated from the broadband three-component records of station RDWB (Fig. 1). We calculate a relative measure of seismic energy from the trace of the covariance matrix of the three component displacement waveforms in a moving time window (see Montalbetti and Kanasewich (1970) and Ereditato and Luongo (1994) for examples). We sum this measure over the duration of the explosion to get total seismic energy. Energy in high and low frequency bands is calculated with the same technique using waveforms filtered on 0.3–25 Hz and 0.033–0.3 Hz, respectively. Separating the energies at 0.33 Hz segregates earthquakes and most tremor into the higher band. The ratio of low to high frequency energy shows the relative contributions of each to the total seismic energy (Table 2). The duration of the explosion is measured at station SPU on Mount Spurr, located 85 km northeast of Mount Redoubt (see Power et al., 2013). We also consider the maximum ash cloud height (Schneider and Hoblitt, 2013).

The first explosion occurred on March 15 (Event 0) and is the smallest explosion in several respects: it has the least seismic energy, the lowest plume height and it did not register at station SPU. This event was phreatic in nature as it contained no juvenile material and deposited only a small amount of ash at the vent (Wallace et al., 2013). This, as well as its timing shortly before the onset of the magmatic activity, suggests that this event was an explosion of gas that had sufficient pressure to break a narrow pathway to the surface but was not accompanied by magma.

On March 23 six magmatic explosions occurred over 22 h (explosions 1–6 in Table 2). The first three were closely spaced in time and had progressively longer durations. Explosions 4–6 had the greatest seismic energy of the 2009 eruption, some of the longest durations and had two ash plumes exceeding 18 km ASL. With the exception of number 6, these explosions contained a smaller fraction of low frequency energy.

The next sequence of explosions (events 7–18) occurred between March 26 and March 29. The majority of these events produced large ash plumes that exceeded heights of 12 km ASL. They also had shorter durations than events 1–6, and many were preceded by gliding tremor (Hotovec et al., 2013). Beginning on March 28, the explosions had a much greater fraction of low frequency energy than the first explosion sequence. Haney et al. (2013) take advantage of this low frequency energy to derive a volumetric source depth of 1.9 km below the crater floor for event 12. Fee et al. (2013) note that the infrasonic pulses associated with these later events were more impulsive and shorter in duration than earlier events. The nature of the deposits was also different during this period, exhibiting much finer grain-sizes than deposits from the March 23–24 explosion sequence, although the chemical composition remained unchanged (Wallace et al., 2013). These observations indicate that the style of the eruptive activity changed during the March 26–29 sequence. It is possible that the viscosity of the magma increased, resulting in a greater build-up of pressure and material behind the magma plug in the conduit producing the more explosive events with correspondingly larger low frequency components.

After a period of lava effusion and dome growth, the final large explosion of the eruption occurred on April 4 (event 19) after a 5-day earthquake swarm (See Section 4.6). This explosion was the longest in duration, producing an ash cloud above 15 km ASL and destroying the lava dome that had been growing since March 29. However, it was of modest energy and contained relatively little low frequency energy. The explosion contained two main pulses and several smaller pulses that are examined in detail by Fee et al. (2013) and Schneider and Hoblitt (2013). The proximal deposits from this sequence contained a large amount of material that was derived from the lava dome, suggesting that dome collapse may have played a role in generating the explosion sequence (Bull et al., 2013; Wallace et al., 2013). We speculate that the longer duration, more gradual onset and relative lack of low frequency energy of this explosion were all due to the influence of the collapsing lava dome. The presence of the dome may have inhibited the final ascent of magma in the shallow conduit, causing the initial phase of the explosion to be weaker. Once enough of the dome had collapsed and/or the explosion had removed enough of the dome to clear the vent area, the explosion was able to progress in the fashion typical of the earlier explosion events.

7. Lahars

The steep sided, heavily glaciated edifice of Redoubt Volcano makes it an ideal setting for pyroclastic flows, debris avalanches and...
lahars. The latter is of particular relevance due to the Drift River Marine Terminal at the mouth of the Drift River (Fig. 1) — a well-documented hazard prior to the 2009 eruption. During the 1989–90 eruption the larger lahars reached the oil terminal, prompting operations at the terminal to be suspended.

Seismic records from lahars share similar frequency content and duration to pyroclastic flows and can be difficult to distinguish without additional information (e.g. Marcial et al., 1994; Nye et al., 1995; Huang et al., 2007). Visual observations exist from time-lapse cameras in the Drift River Valley, but the photos rely on daylight and good weather and are therefore sporadic (Bull et al., 2013; Waythomas et al., 2013). Given the location of the seismic stations along the main lahar channel, their more distal locations from the edifice, and the long durations of these seismic signals following the explosions, it is likely that the majority of flow-type seismic signals were due to lahars. Without visual confirmation, however, we cannot conclusively discriminate between pyroclastic flow and lahar signals, and instead we refer to these signals as ‘flow events’.

7.1. Quantifying the lahar seismic record

Flow signals were identified by visually scanning the data for sustained seismic activity on stations RDE and DFR, which were located on either side of the Drift River valley. Most of the flow events followed explosions and shared some common signal characteristics to the explosions, including durations greater than 10 min and energy up to 25 Hz. However, the majority of the energy in the explosion signals was concentrated below 5 Hz, whereas the energy in the flow signals was more broadly distributed in frequency. Fig. 6 compares the frequency spectra between station NCT, DFR, RDE and RDN. Located close to the vent, the seismic record at station RDN was dominated by the explosion. Stations NCT, DFR and RDE are located at similar distances from the vent and recorded the explosion signals, however only stations DFR and RDE were located near Drift River valley and recorded flow signals. While the explosion and flow signals appeared as discrete events at station DFR, the transition between explosion signal and flow signal at RDE was less clear. Also of note in Fig. 6 is an apparent discrepancy in the concentration of energy from 8 to 10 Hz between stations RDE and DFR. Regardless of whether this is a site response or sensor noise, it appears consistent throughout the eruption.

We filter waveform data between 5 and 10 Hz to emphasize flow signals over explosion signals. The timing of the flow events is determined from the signal to noise ratio (SNR) of the filtered waveforms at stations RDE and DFR (Fig. 6B,D,F,H). We define the noise for each flow event by averaging the signal over a 2 min window prior to the onset of the activity at the vent. The SNRs of these events were typically double peaked, with the first peak due to the explosion signal and the second peak due to the flow passing close to the station. We define the flow onsets at the SNR minimum between the explosion and lahar peaks. Flow event end times are defined as the time when the SNR drops below 2. Although subjective, these definitions enable a quantitative comparison of the flow seismic records. Errors within the onset times are estimated visually, and range between 1 and 10 min largely because of variations in the explosion signals that preceded them. Errors in the end times are based on the transition from SNR 2 to SNR 1.

To estimate the relative location and properties of each flow we examine the seismic amplitude at stations RDE and DFR on opposite sides of the Drift River Valley. We calculate the maximum amplitude of the flow by taking the mean amplitude of the filtered signal in a two-minute window around the peak SNR. The mean amplitude ensures that high amplitude spikes from earthquakes did not contaminate the data.

7.2. Lahar comparisons

A total of 20 events are identified, 19 of which are recorded at station RDE and 17 at DFR (Table 3). All but three of the flow events

Fig. 6. Lahar on March 24 recorded at 4 stations on the Redoubt network. Panels A, C, E and G show the spectra of the signals at stations RDN, RDE, DFR and NCT respectively. Panels B,D,F and H show the waveforms for the same event in grey, filtered between 5 and 10 Hz, and the signal to noise ratio (SNR) in black. Lahar durations are shown in black horizontal bars for stations RDE and DFR. The explosion onset is indicated by the dashed vertical line.
followed the major explosions listed in Table 2. The three remaining flows followed significant summit seismic events. These seismic events may have been very small explosions or gravitational collapses of loose material. The time delay between summit events and the onset of the seismic signal at stations DFR and RDE varied between 4 and 26 min. This time difference is influenced by the flow path, the volume of the flow, and the duration of the explosion signal that masks the calculated onset time.

The majority of the flow signals recorded during the March 23–24 explosions had larger seismic amplitudes on both RDE and DFR than those recorded later in the sequence. There were 2 exceptions (Table 3): 1) the first weak flow event (flow event 1) recorded on March 23 was notably smaller than the flows that followed, and 2) the flow following event 19 on April 4 had larger amplitudes than the flows earlier in the sequence. With the exception of the April 4 flow, only the March 23–24 flow events reached the Drift River Marine Terminal (Schaef er, 2012; Waythomas et al., 2013). This suggests that the earlier flows were volumetrically larger than most of the later flows, and that the seismic amplitudes are proportionate to the flow volume.

Most flow events had the highest amplitudes at station RDE, which is closer to the Drift River valley both in distance and elevation making it more sensitive to flows in the main channel. Field observations confirm that the majority of flows, particularly early in the eruption, flowed predominantly down the south side of the river valley close to RDE (Waythomas et al., 2013). Comparison of the maximum amplitudes at RDE and DFR shows some variation between flows. The ratios of the maximum amplitudes on each station are shown in Table 3 and vary in general between 1 and 2. Flows that migrated further north in the valley have higher maximum amplitude ratios. The maximum amplitude ratio is a useful metric from a monitoring perspective, as it indicates a relative location of the flow within the complex channel system.

Higher amplitudes correspond to longer durations in all but one case (flow event 11), where heavy tephra fall was recorded in the Drift River valley and no change in Drift River discharge was observed from the Dumbell Hills camera (Bull and Buurman, 2013), suggesting that this event may have been a pyroclastic flow rather than a lahar. The lower seismic amplitude of flow event 11 may reflect weaker coupling between the pyroclastic flow and the ground, or that the pyroclastic flows are less energetic than the lahars at those distances. Most recent events had earlier onsets at station DFR, located up-valley of RDE, although 4 events appeared earlier at station RDE. Those events that were seen at RDE first had lower maximum amplitudes at DFR, suggesting the flow was mostly restricted to the southern part of the valley and lacked the energy needed for the early part of the flow to appear at station DFR.

8. Background seismicity

In this section we examine the seismic event detection rate (EDR) of the automated catalogue outside of the swarms identified using the criteria described in Section 4. This “background” seismicity includes signals generated by rockfalls and glacial quakes (glaciers cover 80% of the upper volcanic edifice), as well as high- and low-frequency volcanogenic earthquakes. Because the event detection rate is sensitive to many types of seismic activity it is a good qualitative metric of overall unrest. Many volcanoes have increased rates of small earthquakes prior to eruption. Rockfalls demonstrate instability in the upper portions of the edifice that has been shown to be precursory at times (Deroin and McNutt, 2012), and glacier ice is highly sensitive to temperature and deformation near the vent. In the absence of tremor the EDR can, at times, be the primary seismic metric by which to assess unrest. We examine the EDR chronologically, dividing it into 3 sections: 1) prior to the onset of explosive activity, from January 1 through March 20; 2) during the explosive activity, from March 23 through April 2; and 3) following the explosive activity during steady dome growth from April 4 through May 31.

8.1. Precursory EDR: January 1–March 20

The first major increase in EDR occurred on January 27, two days after the first significant tremor. This followed a spike in EDR on the January 25 that was coincident with the first episode of high amplitude precursory tremor. Previous to the onset of tremor, the EDR fluctuated mostly between 2 and 6 events per hour (Fig. 2). This activity represented the true background seismicity at Redoubt Volcano outside of any eruptive activity, and it was largely dominated by small amplitude events originating from the heavily glaciated edifice. Following the onset of tremor the EDR increased to rates of 3–10 events per hour. Both volcano–tectonic and low-frequency earthquakes were located in the analyst-reviewed catalogue, and while the volcano–tectonic activity was confined to the summit region, earthquakes with lower frequencies exhibited scatter below the edifice down to depths of 4 km. This elevated EDR dropped on February 6 concurrent with the increase in tremor (likely due to the decrease in detection capabilities due to the tremor signal), and remained low until the February 26 swarm. Once the February 26 swarm ended, the seismicity rates returned to the same elevated levels that had persisted during late January and early February.

The onset of tremor indicated an increase in fluid movement in the shallow portions of the magmatic system below Redoubt Volcano. The seismicity that followed was likely also generated by the reactivated shallow hydrothermal system. During the tremor episodes the background noise level at the summit stations was higher which decreased the signal to noise ratio of the earthquake activity and resulted in fewer earthquake detections. This was also the reason for the apparent delay in the increase in the EDR following the first burst of tremor on January 25.

8.2. EDR during explosive activity: March 23 – April 2

The seismic record in late March was dominated by explosive activity, episodes of tremor and several seismic swarms — all of which masked the EDR for significant periods of time. In addition, the summit station RSO was destroyed during an explosion on March 23 which reduced the number of smaller earthquake detections.

Following the explosion sequence on March 23–24, the EDR was slightly lower than it had been prior to the March 20 swarm, and was comparable to the background prior to late January (Fig. 2). Rates remained low through the second explosion sequence on March 27–29, and picked up again following the March 29 swarm. The increase in EDR following the March 29 swarm coincided with renewed dome growth which was observed during March 29 and April 2 (Bull et al., 2013). However dome growth was also observed between March 24 and March 27 during a period of lower EDR, which suggests that dome growth was not the only source of seismicity during this period.

8.3. EDR during steady dome growth: April 4–May 31

The EDR was highest between the April 2 and May 2 swarms. Immediately following the April 02 swarm the EDR was greater than prior to the swarm, and increased further on April 17 when summit station RSO was repaired (Fig. 2). Rates remained fairly steady until the onset of the May 2 swarm. We attribute the high EDR during April and May to the growing lava dome. Dome growth at other volcanoes is often characterised by high rates of low-frequency seismicity and rockfall signals (e.g. Soufriere Hills Volcano (Luckett et al., 2008); Augustine Volcano (Power and Lalla, 2010)). The EDR declined slightly after the May 2 swarm, when changes
in the dome facies and eruption rate were observed (Bull et al., 2013). It is likely that the lava dome extruded then had a different seismic character, producing fewer events as the dome grew and cooled.

A notable feature of the background seismicity during this period was the presence of several families of repeating earthquakes which began towards the end of the April 2 swarm and persisted through April and May. These earthquake families were dominated by high frequencies around 10 Hz, had impulsive P- and S-wave arrivals and showed S-P times of 1 s at summit station REF. The fact that these earthquakes were dominated by such high frequencies further increased the significance of their high cross correlation values (>0.75), since high frequency earthquakes recorded at volcanoes are not commonly found to repeat due to their destructive brittle failure source mechanisms. Some of the earthquakes from this repeating family were large enough to be included in the analyst-reviewed catalogue, where their locations are scattered around 4 km below sea level. Given that they first appeared immediately prior to the last and most voluminous explosion on April 4, it is likely that these events were generated from the stress adjustment around the magma reservoir. This was observed during the 1989 eruption of Redoubt Volcano (Power et al., 1994), as well as at other volcanoes including Mount St Helens in 1980 (Moran, 1994), Augustine in 2006 (Power and Lallia, 2010) and Pinatubo (Mori et al., 1996). This type of seismicity was seen much earlier on during the 1989 eruption than was observed during the 2009 eruption (Power et al., 2013). We speculate that, until the April 4 explosions, not enough material had been removed from the deeper (greater than 4 km depth) magma reservoirs to allow for any stress adjustment. In addition the last explosion on April 4 may have had the effect of finally establishing an open conduit system that was able to support prolonged and stable dome growth that lasted through the end of the eruptive period. These repeating earthquake families were then produced by the relaxation of the conduit system behind the last of the ascending magma, which continued to erupt as a stable lava dome.

9. Summary

The progression of magma through the conduit system during the 2009 unrest at Redoubt Volcano can be readily tracked by the seismicity. The deep LP earthquakes in December gave the first indications that magma was moving at depth. By late January it had ascended to depths where gas and heat could easily escape to the surface, generating the precursory tremor and swarm episodes. The first pulse of magma to reach the surface produced a lava dome on March 23 and was preceded by a seismic swarm, which was generated as the shallow conduit system was opened. Dome effusion lasted only 10 h before a series of magmatic explosions occurred which produced high amplitude tremor that continued for several hours on March 24. The explosions melted much of the ice that formed the upper Drift glacier, generating voluminous lahars that were recorded seismically as they flowed down the Drift River valley to the coast. The next pulse of magma erupted on March 26–30 as a series of powerful explosions, many of which were preceded by tremor that was composed of closely spaced earthquakes. This batch of magma had slightly different properties, producing finer-grained deposits and more impulsive explosion signals. The explosions were followed by a short-lived swarm that accompanied the onset of renewed lava dome growth that continued steadily for several days before stalling, concurrent with a 43-hour seismic swarm. During the final hours of the swarm a family of deeper repeating earthquakes began and persisted through the end of May, reflecting relaxation around the mid-crustal magma storage area in response to the evacuation of the magma. The final and longest-lasting explosion of the eruption followed the seismic swarm on April 4, generating a large lahar which reached the coast. Lava continued to erupt following the explosive activity and the steady dome growth lasted through June, accompanied by high rates of summit seismicity which was likely generated by the extrusion of the lava dome. A change in the lava properties in early May was accompanied by a long-lived but low amplitude swarm of repeating earthquakes, but no explosive activity followed and the dome continued to grow, largely uninterrupted.

The variety of signals present in the seismic data reflect the variety of volcanic processes which generated them. These processes are often closely linked, as is evident from the interplay between the tremor and swarm sequences, the lahar and explosion signals and the EDR with different types of eruptive activity. Our interpretations of these signals benefited greatly from the numerous other datasets collected during the eruption. Rarely was there volcanic activity that did not manifest itself in some way seismically, however, resulting in a remarkably complete eruption chronology within the seismic record of the 2009 eruption.

It is clear from our preliminary overview that much work remains to be done with the seismic dataset from the 2009 eruption. Many areas within the seismic dataset remain poorly understood, including the sources of the various tremor episodes, the relationship of the seismic signals from the lahars to the physical properties of the flows, and the cause of the waveform evolution observed during several of the swarm episodes. As with the 1989 Redoubt eruption, we expect to see studies emerging from the seismic record for years to come.

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