Deep long-period earthquakes beneath Washington and Oregon volcanoes

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ABSTRACT

Deep long-period (DLP) earthquakes are an enigmatic type of seismicity occurring near or beneath volcanoes. They are commonly associated with the presence of magma, and found in some cases to correlate with eruptive activity. To more thoroughly understand and characterize DLP occurrence near volcanoes in Washington and Oregon, we systematically searched the Pacific Northwest Seismic Network (PNSN) triggered earthquake catalog for DLPs occurring between 1980 (when PNSN began collecting digital data) and October 2009. Through our analysis we identified 60 DLPs beneath six Cascade volcanic centers. No DLPs were associated with volcanic activity, including the 1980–1986 and 2004–2008 eruptions at Mount St. Helens. More than half of the events occurred near Mount Baker, where the background flux of magmatic gases is greatest among Washington and Oregon volcanoes. The six volcanoes with DLPs (counts in parentheses) are Mount Baker (31), Glacier Peak (9), Mount Rainier (9), Mount St. Helens (9), Three Sisters (1), and Crater Lake (1). No DLPs were identified beneath Mount Adams, Mount Hood, Mount Jefferson, or Newberry Volcano, although (except at Hood) that may be due in part to poorer network coverage. In cases where the DLPs do not occur directly beneath the volcanic edifice, the locations coincide with large structural faults that extend into the deep crust. Our observations suggest the occurrence of DLPs in these areas could represent fluid and/or magma transport along pre-existing tectonic structures in the middle crust.

1. Introduction

Many volcanoes within the Cascade volcanic chain are seismically active. While the overwhelming majority of seismicity consists of high-frequency volcano-tectonic (VT) earthquakes, other types of events including deep long-period (DLP) earthquakes (e.g., White, 1996; Power et al., 2004) also have occasionally been identified (Malone and Moran, 1997; Pitt et al., 2002). DLPs have been inferred by other investigators to represent the movement of magma and/or magmatic fluids within the mid-to-lower crust (10–50 km), and are characterized by mostly low-frequency energy (<5 Hz), emergent arrivals and long-duration codas (e.g., White et al., 1996; Power et al., 2004). Earthquakes at these depths are unusual, as such depths are generally thought to lie below the brittle–ductile transition (e.g., Sibson, 1982). The waveform characteristics of DLPs are also different than that of typical VTs. The differences are illustrated in Fig. 1, which shows seismograms and spectrograms from a DLP and a VT with similar locations near Mount Rainier. The DLP is dominated by emergent phase arrivals and low frequencies, mostly below 5 Hz, whereas the VT has more impulsive phases and broader spectral content, with frequencies between 1 and 20 Hz. Although DLPs were once considered to be rare, they are now observed in the background

seismicity of many volcanically active regions worldwide, including Hawaii (e.g., Wright and Klein, 2006; Okubo and Wolfe, 2008), Japan (e.g., Nakamichi, 2000; Nakamichi et al., 2003, 2004; Matsubara et al., 2004; Ukawa, 2005), Alaska (e.g., Power et al., 2004), California (Pitt and Hill, 1994; Pitt et al., 2002), and Washington (Malone and Moran, 1997). It is important to note that other forms of deep low-frequency seismicity, specifically deep nonvolcanic tremor composed of low-frequency (LFE) and very low-frequency (VLF) events, have been observed along major plate boundaries including the Japan and Cascadia subduction zones (e.g., Shelly et al., 2006; La Rocca et al., 2009) and the San Andreas Fault (Nadeau and Dolenc, 2005). Like DLPs, these "tectonic" deep low-frequency events occur below the brittle–ductile transition, and exhibit waveforms lacking high-frequency energy and clear impulsive phase arrivals. Although we do not discuss deep nonvolcanic tremor in this paper, it is important to note that these similarities could suggest a relation between their respective source processes.

In addition to being part of background volcanic seismicity, DLPs are especially important because some have correlated with eruptions (White, 1996; Power et al., 2004), various forms of volcanic unrest including increased shallow VT and long-period (LP) seismicity (White, 1996; Nakamichi et al., 2003), and magma intrusion with associated CO_2 emissions (Hill, 2006). Most volcanoes exhibit changes in their background seismicity pattern prior to an eruption, which sometimes include increased DLP activity. The clearest example is Mount Pinatubo, Philippines, where many hundreds of DLPs occurred
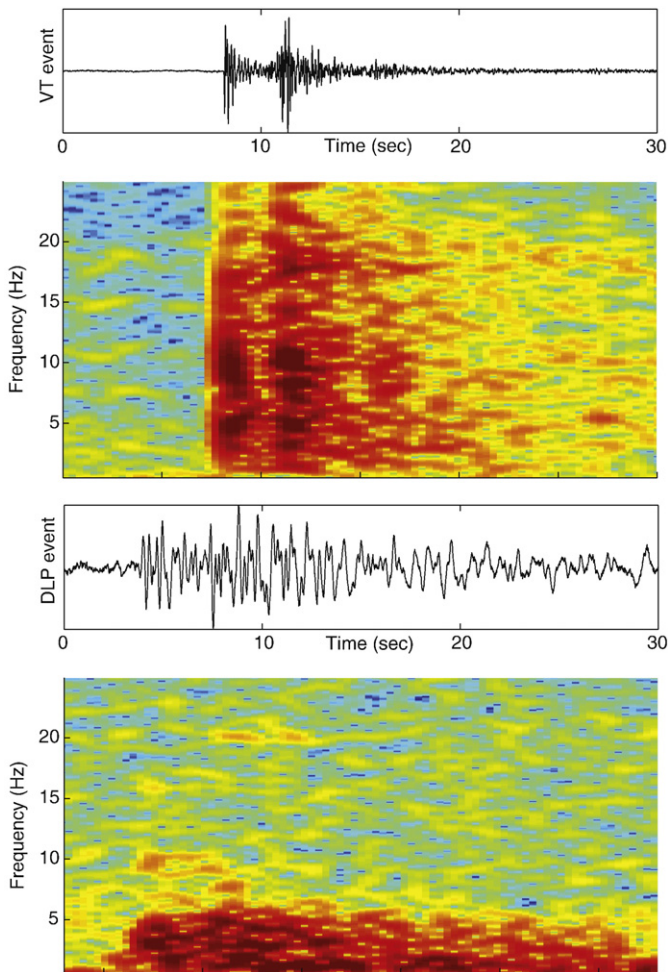
Fig. 1. Seismograms and spectrograms from two events recorded on station FMW near Mt. Rainier illustrate the difference in frequency content between a VT (top—1995/07/14 12:14) and a DLP (bottom—1996/03/05 14:09). The VT contains frequencies between 1 and 20 Hz, and the DLP contains energy mostly below 5 Hz. VT and DLP hypocentral distances are 23.2 km and 23.8 km, respectively. VT and DLP maximum amplitude counts are 1554 and 253, respectively. Spectrogram colors represent amplitude intensity and range from blue (low) to yellow (intermediate) to red (high).

increased flux of magma from lower- to mid-crustal levels in response to depressurization of the shallow magmatic system. The 1999 eruption of Shishaldin Volcano was preceded by six DLPs that occurred roughly 10 months prior (Moran et al., 2002; Nye et al., 2002) and were followed ~12 days later by the first shallow LPs. Power et al. (2004) use these temporal relations to suggest a link between magmatic activity within the lower crust, intrusive activity in the upper crust, and eruption at the surface. At Mount Redoubt, an increase in DLP activity was observed three months prior to the onset of the most recent 2009 eruption, suggesting that it may have followed the ascent of magma from the lower crust (Power et al., 2009).

At each of these volcanoes, a relation was inferred between DLPs and eruptive phenomena. Although DLPs do not necessarily signal an eruption, they may be one of the earliest indications of renewed volcanic activity, and therefore the rapid identification of DLPs could improve forecasts of future eruptions at other volcanoes. Malone and Moran (1997) previously reported a total of 11 DLPs beneath three Washington volcanoes, and indicated that additional DLPs may have been recorded, but not identified as such during routine analysis. The motivation for this study was to search for additional DLPs in the Pacific Northwest Seismic Network (PNSN) catalog in order to provide improved constraints on the conditions under which DLPs occur in the Cascades, and to better understand their potential use in forecasting eruptions at Washington and Oregon volcanoes.

In this paper we first provide a brief summary of the seismic network surrounding the volcanoes, and explain our process for systematically identifying DLPs. This is followed by a description of DLP occurrence at each of the six volcanoes where they were observed (Fig. 2). In these sections we include some background information on historical eruptions, magmatic composition, and seismicity for each volcano, followed by a description of DLP locations. Finally, we discuss our observations of DLPs and the relation between these earthquakes and magmatic transport beneath the various volcanic centers.

2. Summary of seismic network

The PNSN automatically detects and locates earthquakes in Washington and Oregon, including those occurring at 10 of the major Cascade volcanoes (Mount Baker, Glacier Peak, Mount Rainier, Mount St. Helens, Mount Adams, Mount Hood, Mount Jefferson, Three Sisters, Newberry Volcano and Crater Lake). Most instruments surrounding the various volcanic centers are short-period vertical-component stations, with a few short-period three-component seismometers and broadband seismometers near some volcanoes (Moran, 2004). The standard short-period vertical-component instruments are sufficient to detect and locate VT and impulsive LP events. The number and type of instruments at each volcano varies throughout the arc, as do the resulting detection thresholds (Table 1). Only triggered events are reported in the PNSN catalog, and prior to 2001 (when PNSN began archiving continuous data) data containing events that did not meet the network's detection threshold were discarded. Detected events are located using PNSN 1-D layered velocity models. The PNSN triggering algorithm is not well-tuned to capture DLPs due to their emergent and low-frequency nature, and those that trigger may have poor locations resulting from large picking errors. Other non-volcanic sources existing near volcanoes, such as glacier quakes and explosions, can produce similar waveforms to DLPs. For these reasons, the possibility exists for mislocated or misidentified DLPs within the existing PNSN catalog.

3. Methods

DLPs are characterized by their depth (>10 km) and by being enriched in low frequencies compared to VTs. A visual comparison between frequency spectra of previously identified DLPs and VTs from

beneath the volcano in the months preceding the major 1991 eruption (White, 1996). Some of these DLPs occurred several hours prior to increased shallow seismicity and steam emissions from the site where a lava dome later emerged. Roughly a week before the climactic eruption, DLP activity diminished with the emergence of the hybrid-andesite dome containing inclusions of basalt that had arrived from the deep crust only days to weeks earlier (Pallister and Hoblitt, 1992). The spatiotemporal development of DLP seismicity and subsequent surface activity, especially the emergence of the dome, suggests a link between the deep and shallow volcanic system. This set of observations is taken as evidence that DLPs represent the injection of deep-seated basaltic fluids into the base of the magma chamber which likely triggered the eruptive sequence at Mount Pinatubo (White, 1996).

DLPs have occurred in association with eruptions at several Alaskan volcanoes (Power et al., 2004). In 1992, Mount Spurr produced three main explosive eruptions on June 27, August 18, and September 16–17. A single DLP occurred nine days prior to the first explosion on June 27, after which the number of DLPs steadily increased, peaked just after the September eruption, and slowly declined until September 1996. Power et al. (2002) suggest that the steady increase in DLP activity after the first eruption represents an
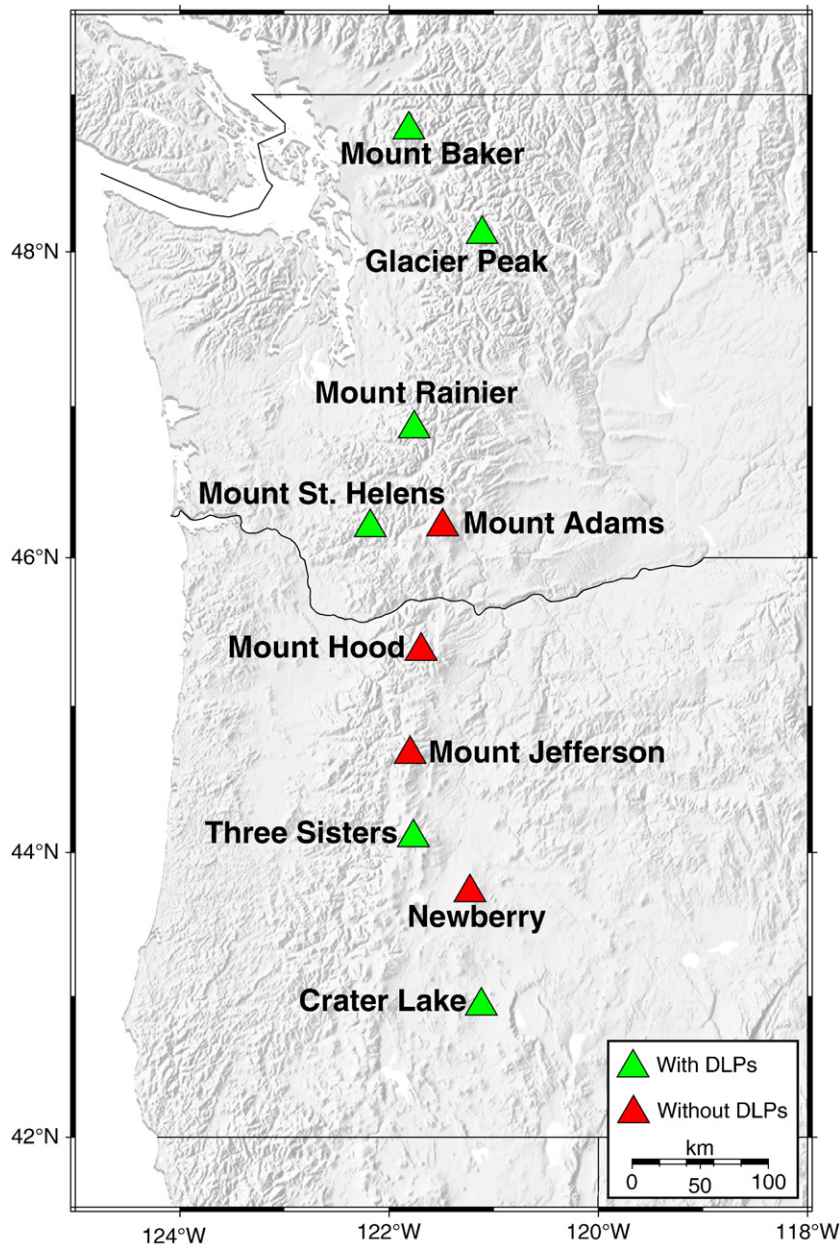
Fig. 2. Map showing the major Cascade volcanoes of Washington and Oregon. Green and red triangles represent volcanoes with and without identified DLPs, respectively.

the same regions illustrates that DLPs, unlike VTs, contain mostly low-frequency (LF) energy with peaks typically around 1–2 Hz and little high-frequency (HF) energy between 10 and 20 Hz (Fig. 3). In each case in Fig. 3, the two events are recorded on the same station, therefore the difference in frequency content is not due to site effects. Some DLPs possess HF onsets followed by LF tails, thus the difference in frequency content is not due to path effects because we would expect the HF onset to attenuate along the ray path. We employed this characteristic difference in frequency content to distinguish DLPs from the more abundant VTs and conducted a semi-automatic review of triggered earthquakes in the PNSN catalog occurring between 1980 and October 2009. By calculating a HF/LF ratio between the average energy in the 10–20 Hz and 1–2 Hz bands, we found this value to be consistently lower for a DLP than a typical VT, as would be expected. Based on the HF/LF ratios we calculated for the previously identified DLPs (Malone and Moran, 1997), we required potential DLPs to have a HF/LF ratio of ≤0.5 (threshold) on at least two stations near the

volcano. One source of false negatives is that event waveforms for potential DLPs may have a low signal-to-noise ratio, resulting in a HF/LF ratio above the threshold. Other LF signals, including glacier quakes and unidentified explosions, also sometimes have ratios below the threshold, but these false positives are rejected through the subsequent visual inspection.

In order to systematically search for DLPs, we first defined an appropriate search area around each volcano. Previously identified DLPs in the Cascades and elsewhere tend to be located horizontally within one focal depth of the volcano summit (e.g., Malone and Moran, 1997; Pitt et al., 2002; Power et al., 2004). Based on this observation, we defined search regions (roughly 40×40 km) centered on each volcano. In order to reduce the number of earthquakes at volcanoes with abundant shallow seismicity and minimize false triggers from shallow events with waveforms similar to DLPs (glacier quakes, shallow LP earthquakes, etc.), we restricted our search to events located deeper than 10 km at Mount St. Helens and Mount

Table 1
Washington and Oregon DLP Information.

Volcano	Total DLP	Depth (km)	Av EH1[a] (km)	Av EZ1[a] (km)	Av EH2[b] (km)	Av EZ2[b] (km)	Max M$_c$[c]	DT[d]
Mount Baker	31	15–42	0.9	1.2	2.4	2.4	2.6	0.4
Glacier Peak	9	10–23	0.4	1.2	1.5	2.5	2.4	0.5
Mount Rainier	9	10–22	0.8	1.0	1.9	2.5	2.5	−1.6
Mount St. Helens	9	24–32	1.2	0.8	2.0	1.4	1.5	−1.6
Mount Adams	0	–	–	–	–	–	–	0.6
Mount Hood	0	–	–	–	–	–	–	−1.1
Mount Jefferson	0	–	–	–	–	–	–	1.6
Three Sisters	1	12	0.2	0.2	1.8	1.3	1.5	0.7
Newberry	0	–	–	–	–	–	–	2.5
Crater Lake	1	32	1.6	2.6	4.3	2.2	2	1.6
Total # DLP	60							

[a] EH1 and EZ1 are formal estimates of horizontal and vertical DLP location errors, respectively. These values are based on RMS time residuals, and are likely underestimated given that DLPs often have emergent phase arrivals with significant picking uncertainties.

[b] EH2 and EZ2 are recalculated estimates of horizontal and vertical DLP location errors, respectively. These values represent the mean horizontal and vertical distance, respectively, between the final location and locations determined using a variety 1-D velocity models.

[c] M$_c$ (coda magnitude) is calculated based on signal duration. DLPs tend to have long codas, and therefore we expect these magnitudes to be slightly overestimated.

[d] DT (detection threshold) is the approximate magnitude of the smallest earthquake located within 20 km of the volcano between 1990 and 2003 (Moran, 2004). Note that additional stations have been installed since 2003 and therefore some detection threshold values have improved.

Rainier, and 5 km at Mount Hood. We also excluded known explosions identified in the PNSN catalog. Once potential DLPs were identified at each volcano, they were visually inspected in order to confirm DLP identification. An important criterion for discriminating DLPs from other sources was phase-arrival patterns across neighboring seismic stations; most glacier quakes and other shallow-focus sources record only on stations close to the source, whereas DLPs tend to be well-recorded on stations tens of km from the source. After calculating HF/LF ratios for nearly 5000 input events and visually inspecting roughly 400 triggered earthquakes, we identified 49 DLPs beneath Washington and Oregon volcanoes, in addition to the 11 previously recognized by Malone and Moran (1997).
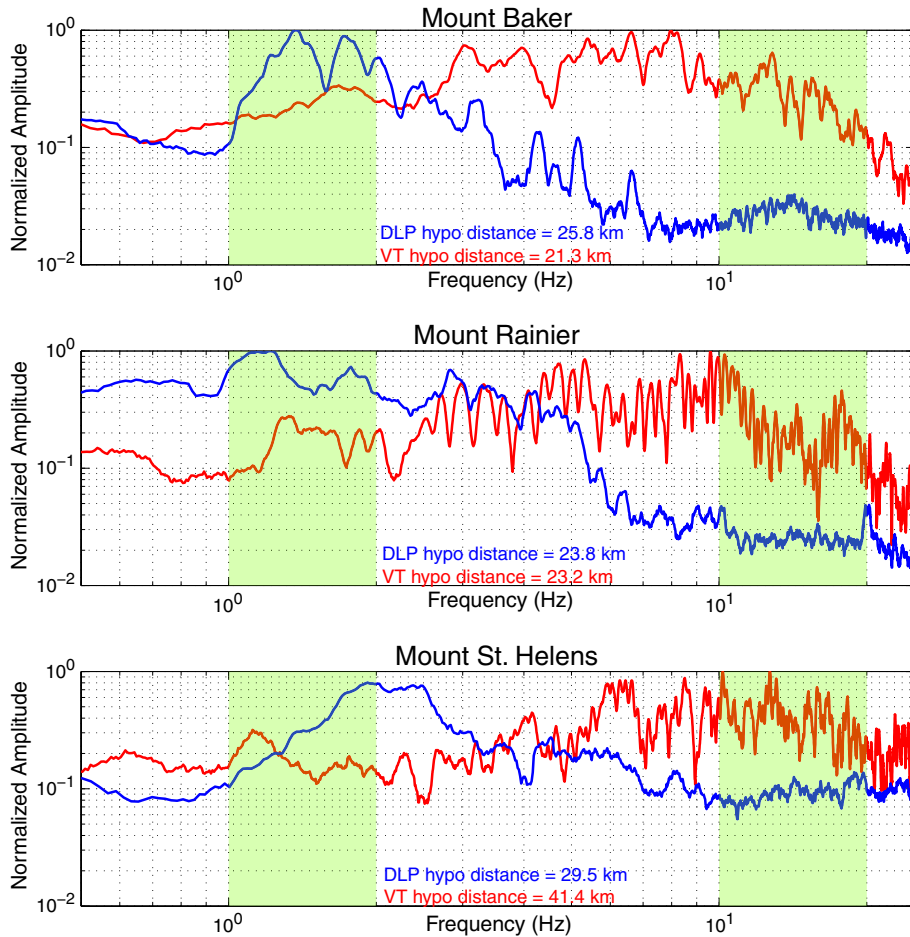

Fig. 3. Comparison of frequency spectra from several DLPs (blue) and VTs (red) beneath Mount Baker, Mount Rainier, and Mount St. Helens. Green rectangles highlight the two frequency bands (10–20 Hz; 1–2 Hz) used to calculate HF/LF ratio. Note that the VTs contain energy between both 10–20 Hz and 1–2 Hz, whereas the DLPs contain very little energy between 10 and 20 Hz.

4. DLP occurrence

4.1. Mount Baker

Mount Baker is the northernmost volcano in Washington. The present-day cone is composed mainly of andesite lava flows, and it is host to an active hydrothermal system with the greatest background flux of magmatic gases among Washington and Oregon volcanoes (Symonds et al., 2003) expressed on the surface through steaming ground and boiling fumaroles in Sherman Crater and the Dorr Fumarole field (Hildreth et al., 2003). The record of historical eruptive activity at Mount Baker consists of a phreatic explosion and several smaller phreatomagmatic eruptions between 1843 and 1891 together with increased thermal activity beginning in the 1930s (Scott and Tucker, 2003). In March 1975, geothermal activity dramatically increased in the Sherman Crater area. No earthquakes were observed in association with this increased thermal activity, although that may be due in part to poor network coverage. After the installation of additional stations shortly following the observed steaming in March 1975, no anomalous seismicity was reported in the catalog (Malone, 1979). Observations of gas geochemistry, isotopic ratios, and emission rates from 1974 to 2007 provide strong evidence for the injection of magmatic gases into the hydrothermal system by a small magmatic intrusion at mid-to-upper crustal depths (Werner et al., 2009). Gravity surveys at Mount Baker show an increase from 1977 to 2005, also consistent with the densification of a magma body emplaced in 1975 or earlier (Crider et al., 2008). Deformation data spanning the last quarter century also indicate that edifice-wide deflation has occurred, most likely in response to depressurization resulting from the densification and devolatilization of the magmatic and hydrothermal system beneath Mount Baker (Hodge and Crider, 2010). The subsequent decline in gas emissions and heat flux suggests a continuing connection between the surface and the deep volcanic system (Werner et al., 2009). Despite the low levels of background VT seismicity, we identified 31 DLPs beneath Mount Baker, more than any of the other Washington and Oregon volcanoes combined. Event locations extend from directly below the summit to ~10 km to the south and southeast of the edifice, with depths ranging from 15 to 30 km (one DLP is located at 42 km depth) (Fig. 4).

4.2. Glacier Peak

The edifice of Glacier Peak is composed principally of dacite lava dome-and-flow complexes (Hildreth, 2007). The record of historical activity consists of a very small 18th century eruption based on a thin tephra fall and corroborating Native American accounts (Wood and Kienle, 1990). Glacier Peak has produced very little seismicity, averaging only ~3 earthquakes per year. However, the volcano was poorly monitored until 2001 when a station was installed on its upper west flank (Table 1). We identified nine DLPs near Glacier Peak. All of the events are located 10 to 15 km west of the edifice near the Straight Creek fault (SCF), with depths ranging from 10 to 23 km (Fig. 5). The SCF represents the southern section of the Fraser River–Straight Creek fault system extending from Central Washington into British Columbia, which is part of a much larger 2500-km-long Late-Cretaceous–Paleogene intracontinental transform fault inferred by Price and Carmichael (1986). Although the SCF is not currently active, this major subsurface structure is interpreted to have accumulated up to ~190 km of right-lateral offset during the Eocene (Misch, 1977; Vance and Miller, 1981; Monger, 1985).

4.3. Mount Rainier

The edifice of Mount Rainier consists mainly of stacked andesite and dacite lava flows, and along with being the tallest Cascade volcano, it also supports the largest volume of snow and ice of any mountain in the contiguous United States and (Hildreth, 2007). The most recent geochemically documented magmatic eruption occurred ~1 ka, producing small-volume pyroclastic flows and subsequent lahars (Sisson and Vallance, 2009). The Electron Mudflow occurred ~600 years ago and may have been triggered by a small magmatic event, however no juvenile deposits have been found (Sisson and Vallance, 2009). Mount Rainier is the second most seismically active volcano in the Cascade Range after Mount St. Helens, averaging around 150 located earthquakes per year. Many shallow VTs are located within the edifice, mostly near sea level. These earthquakes are interpreted to be occurring in response to circulation of magmatic and hydrothermal fluids within and below the edifice, which reduce effective stress and/or weaken rock such that gravity-induced slip can occur (Moran et al., 2000). The many glaciers surrounding the summit also produce a substantial number of ice-related quakes. In addition to these events, much of the seismicity in the area occurs west of the volcano within the Western Rainier Seismic Zone (WRSZ), which is inferred to be a series of relatively small NE to NW trending buried en echelon faults that produce mostly thrust and dextral strike–slip motion (Moran et al., 1999). Stress tensors from earthquakes in the WRSZ are consistent with the regional tectonics of Washington, and yield nearly horizontal major and minor compressive stress axes (σ_1 and σ_3), with σ_1 varying from N–S to NNE–SSW and σ_3 oriented E–W (Giampiccolo et al., 1999). We identified nine DLPs near Mount Rainier. These events occur 8 to 12 km southwest of the summit along the WRSZ, with depths ranging from 10 to 22 km (Fig. 6). These DLPs also occur along the western edge of a low-velocity anomaly (Moran et al., 2000) that likely reflects a volume of hot host rock containing small pockets of magma and/or magmatic fluids (Moran et al., 1999).

4.4. Mount St. Helens

The edifice of Mount St. Helens is composed mainly of andesite and basalt lava flows, interlaced with dacite domes (Hildreth, 2007). After the eruption of May 18, 1980, several smaller explosive and dome-building eruptions occurred through October 1986, followed by nearly two decades of quiescence. In September 2004, a dome-building eruption began and ceased by January 2008. Mount St. Helens is the most seismically active volcano in the Cascade Range, averaging over 800 located earthquakes per year, with many thousands of events occurring during the 1980–1986 and 2004–2008 eruptions. Along with seismicity occurring within and below the edifice, earthquakes in the Mount St. Helens area extend north and south of the edifice along a linear feature known as the St. Helens Seismic Zone (SHZ) (e.g., Weaver and Smith, 1983). The SHZ is inferred to be a single buried dextral strike–slip fault or fault zone (Moran et al., 1999). Similar to the WRSZ, stress tensors from crustal earthquakes in the SHZ are also consistent with the regional tectonics of Washington (Giampiccolo et al., 1999). We identified nine DLPs near Mount St. Helens. These events occur 5 to 10 km southeast of the summit, along the SHZ, with depths ranging from 25 to 30 km (Fig. 7). Mount St. Helens is the only Cascade volcano to erupt since monitoring began, and we observed no temporal correlation of DLPs with the 1980–1986 and 2004–2008 eruptions. Only one station was in operation within 50 km of Mount St. Helens prior to the initial seismic activity preceding the 1980 eruption, therefore DLPs could have gone undetected before 1980. Due to the intensity of the shallow seismicity associated with the pre-eruptive sequence, it is possible that precursory DLPs could have been missed.

4.5. Three Sisters

The Three Sisters volcanic center consists of three adjacent stratocones with differing compositions: North Sister is a strictly mafic edifice, Middle Sister is a basalt–andesite–dacite edifice, and South Sister is a bimodal rhyolite–andesite edifice (Hildreth, 2007). The most recent eruption in the Three Sisters region occurred at South Sister ~2 ka, producing pyroclastic flows and tephra fallout, followed
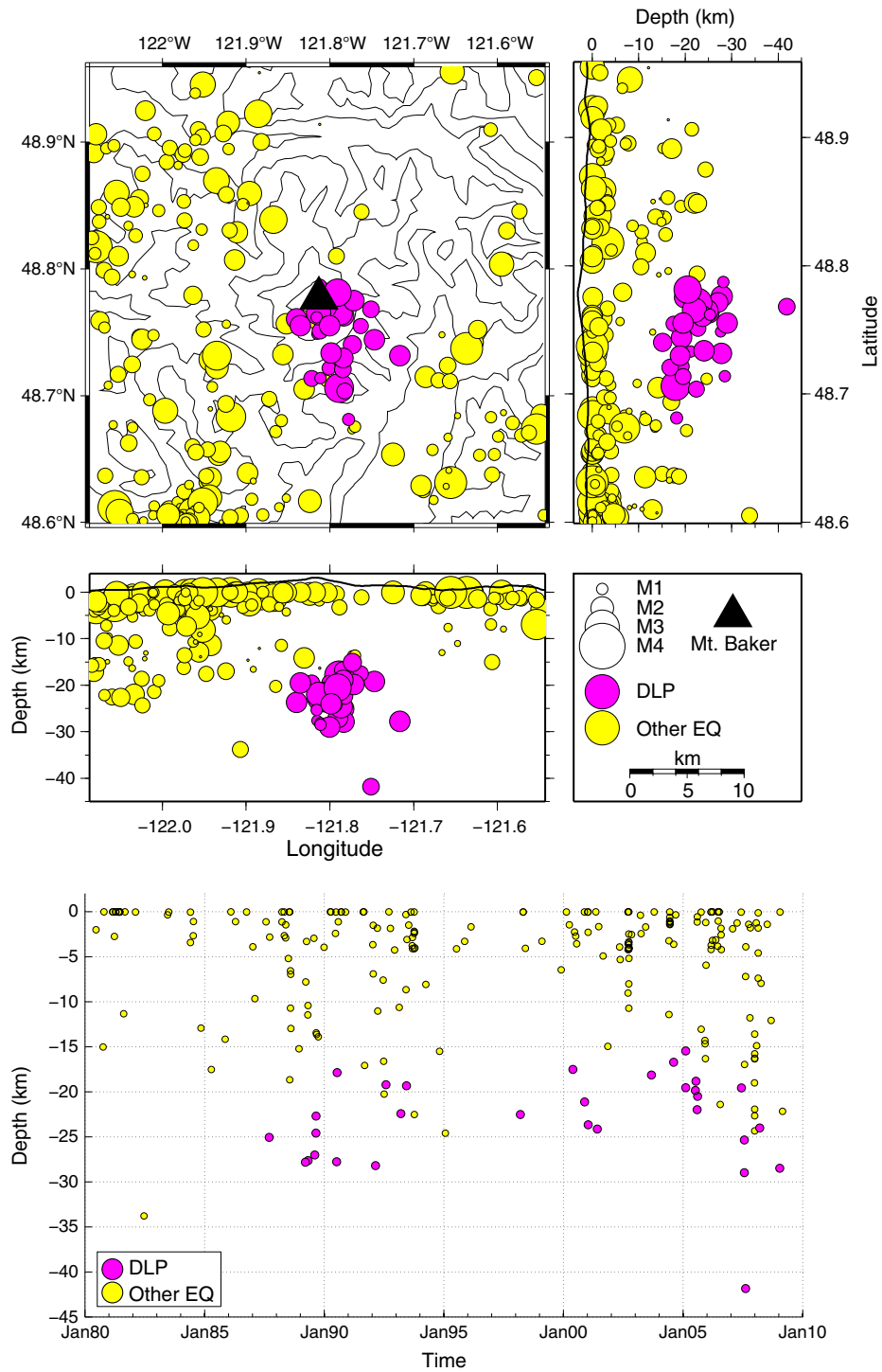
Fig. 4. Mount Baker seismicity map, E–W (bottom) and N–S (right) cross-sections, and time vs. depth series from Jan 1980 to Oct 2009. Black triangle represents the summit. Black lines atop cross-sections represent topographic profile. Note that earthquake symbols are scaled with magnitude in the map and cross-sections but not in the time vs. depth series.

by the emergence of lava flows and domes (Scott, 1987). During analysis of InSAR data in April 2001, uplift was detected by Wicks et al. (2002) in an area ~20 km in diameter centered ~6 km west of South Sister that began in September 1997 and was continuing when surveyed in August 2006 [Dzurisin et al., 2009]. A joint inversion of InSAR, GPS, and leveling data suggests that the uplift resulted from intrusion of magma, likely basaltic, located at a depth of ~5 km (Dzurisin et al., 2009). The first notable seismicity in the Three Sisters volcanic area occurred March 23–26, 2004 when a swarm of around 300 small VTs took place in the northeast quadrant of the deforming

area (Dzurisin et al., 2006). We identified one DLP near the Three Sisters region. This event occurred ~6 km west of North Sister at a depth of ~12 km, and is located within the northern section of the uplifted region and 3 to 5 km northwest of the 2004 swarm (Fig. 8).

4.6. Crater Lake

Crater Lake sits within a caldera created by the collapse of the predominantly andesitic edifice of Mount Mazama during a major plinian eruption ~7.7 ka (Bacon, 1983; Stuiver et al., 1998). The only
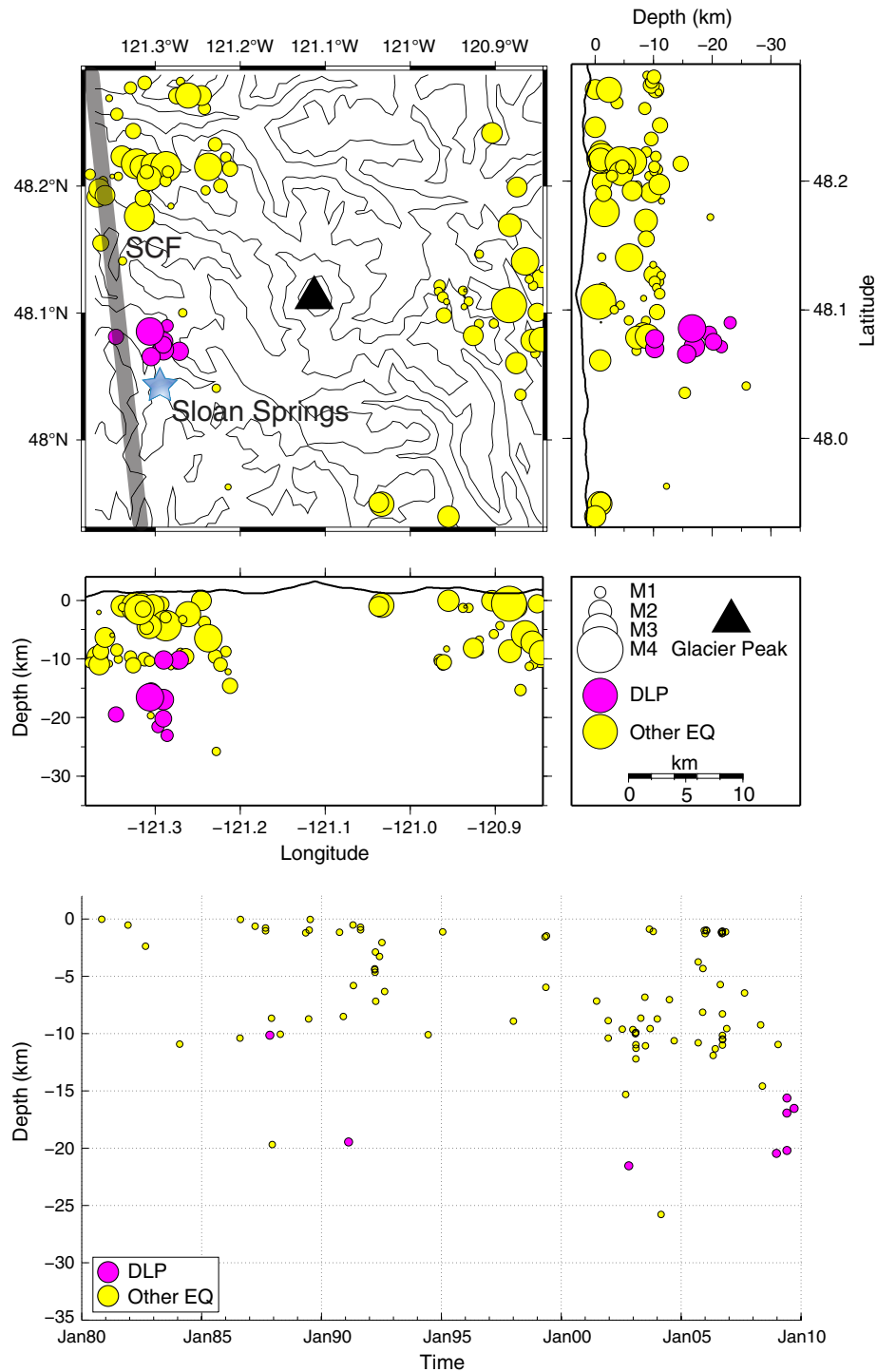
Fig. 5. Glacier Peak seismicity map, E–W (bottom) and N–S (right) cross-sections, and time vs. depth series from Jan 1980 to Oct 2009. Black triangle represents the summit. Black lines atop cross-sections represent topographic profile. Bold gray line represents approximate location of SCF translated from Haugerud and Tabor (2009). Blue star represents location of Sloan Springs (T. Cladouhos, personal communication, 2009). Note that earthquake symbols are scaled with magnitude in the map and cross-sections but not in the time vs. depth series.

volcanic eruptions since the climactic event have occurred within the caldera, the most recent of which resulted in the emplacement of a rhyodacite dome ~4.8 ka (Bacon and Lanphere, 2006). Crater Lake caldera is one of the most poorly monitored (Table 1) and least seismically active Cascade volcanoes, with only six reported earthquakes within the bounds of our search. We identified one DLP, located ~9 km west of the western edge of the caldera at a depth of 32 km. Due to the lack of earthquakes in the Crater Lake region, we do not include a seismicity plot.

5. Discussion

5.1. DLP locations

The formal estimates for vertical and horizontal errors for our DLP locations range from 0.2 to 2.6 km and 0.2 to 1.6 km (Table 1), respectively. However, these errors are likely underestimated given that the 1-D velocity models used by the location algorithm do not reflect the complex 3-D velocity structure beneath volcanoes and
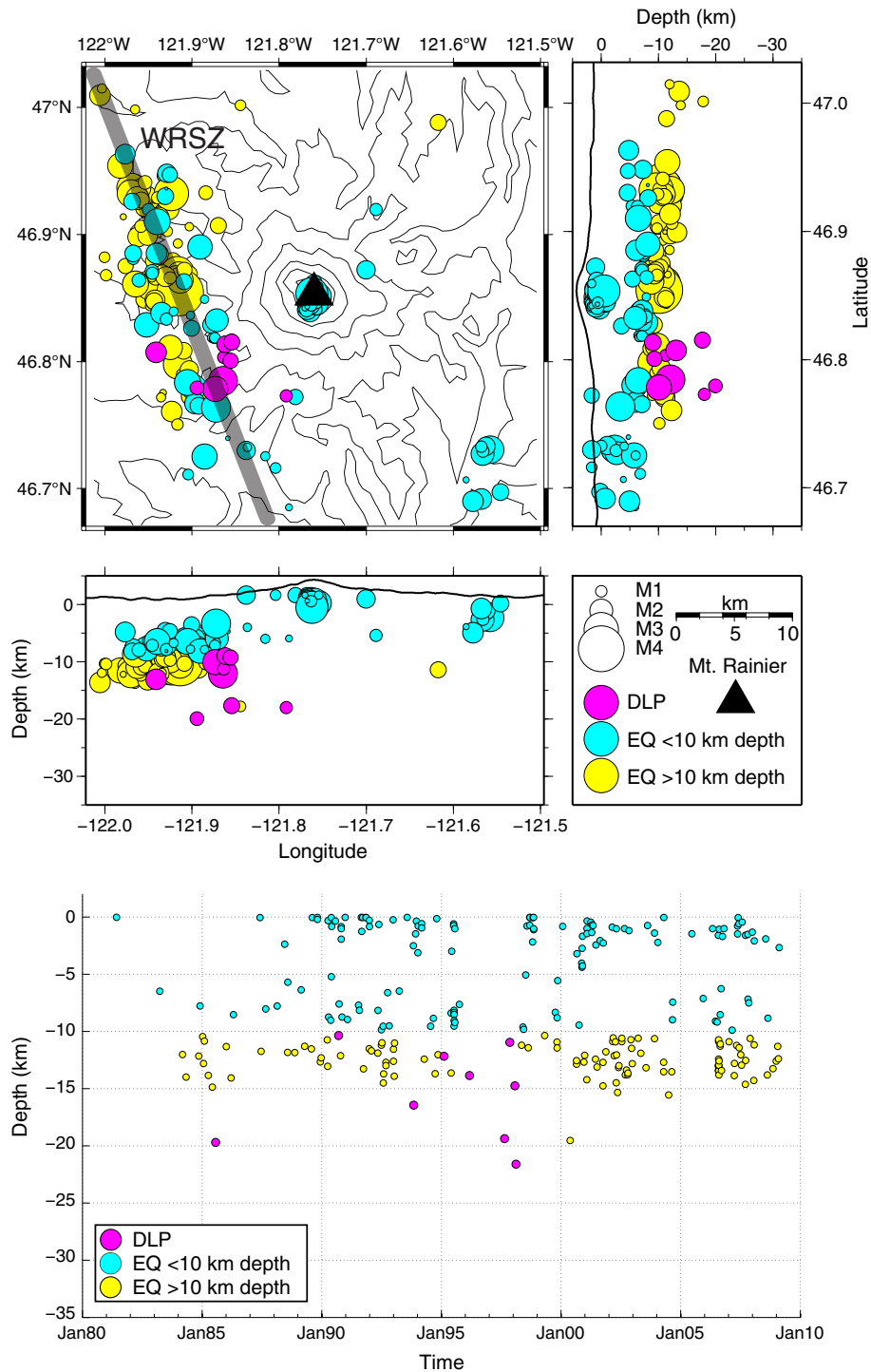
Fig. 6. Mount Rainier seismicity map, E–W (bottom) and N–S (right) cross-sections, and time vs. depth series from Jan 1980 to Oct 2009. Black triangle represents the summit. Black lines atop cross-sections represent topographic profile. Bold gray line represents approximate location of WRSZ. Note that earthquake symbols are scaled with magnitude in the map and cross-sections but not in the time vs. depth series. To increase figure clarity, we only plot highest quality earthquake locations.

emergent phase arrivals that can have significant picking uncertainties. In order to more accurately estimate our location errors, we computed the distance from our final DLP location to locations determined using a variety of 1-D velocity models, and use the mean horizontal and vertical distances as an estimate of error. Using this method, our estimates for vertical and horizontal errors range from 1.3 to 2.5 km and 1.5 to 4.3 km (Table 1), respectively. We are confident in the overall interpretation of the DLP hypocentral locations, as the DLPs occur at depths sufficient to be recorded on distant stations. In addition, we tested a variety of 1-D velocity models

and found DLP depths to be relatively invariant, suggesting that DLP depths are not significantly affected by inaccurately modeled shallow structure. We therefore interpret the loose clustering and vertical spread of our DLP locations to be reasonably accurate.

Estimates of crustal thickness beneath the Washington and Oregon Cascades range from 30 to 45 km (e.g., Johnson and Couch, 1970; Miller et al., 1997; Das and Nolet, 1998). DLPs in Washington range from 10 to 42 km deep, hence the majority of the events occur within the mid-to-lower crust. In Northern California, DLPs have also been reported at mid-crustal depths beneath two other Cascade volcanoes
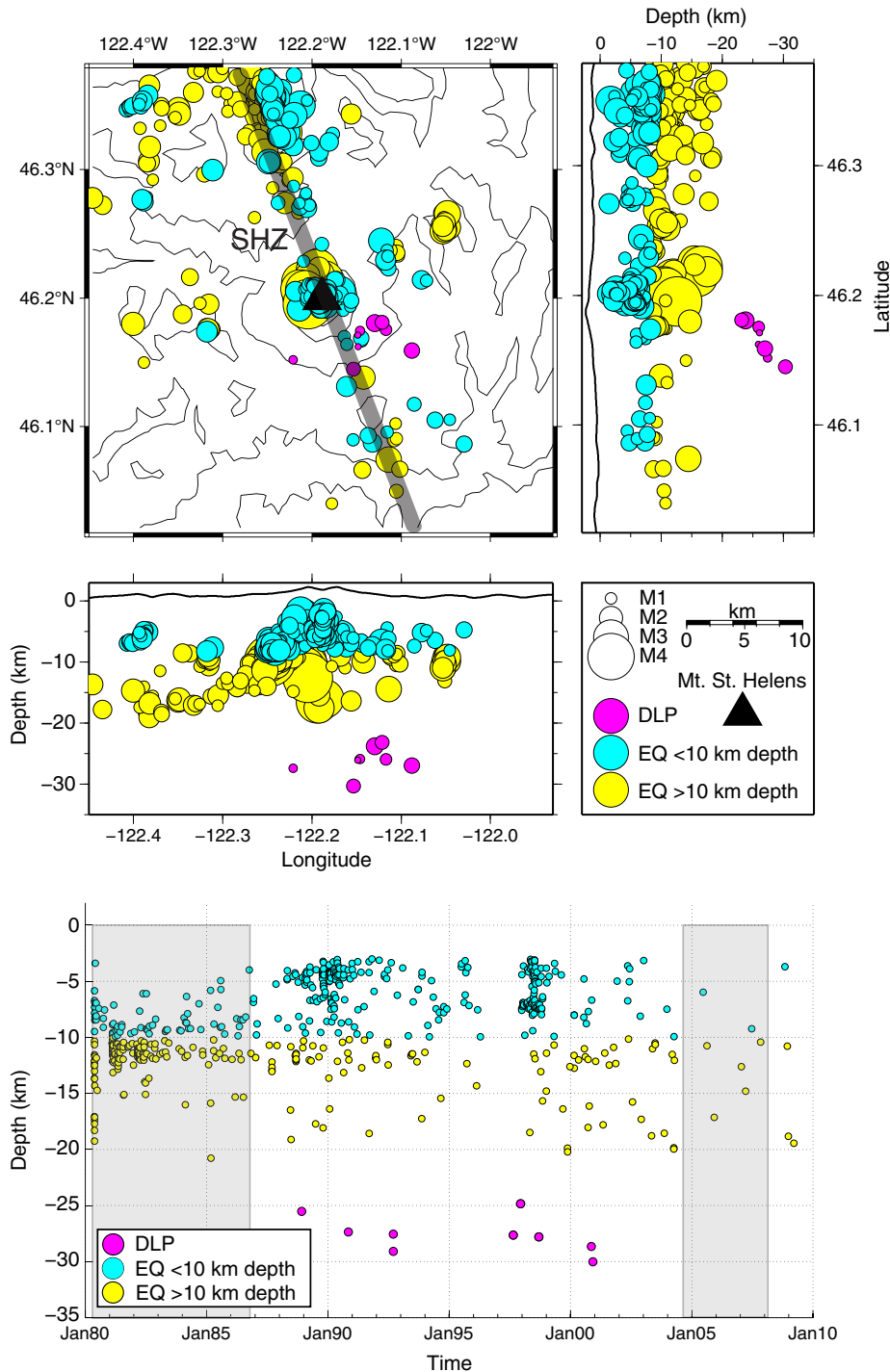
Fig. 7. Mount St. Helens seismicity map, E–W (bottom) and N–S (right) cross-sections, and time vs. depth series from Jan 1980 to Oct 2009. Black triangle represents the summit. Black lines atop cross-sections represent topographic profile. Bold gray line represents approximate location of SHZ. Shaded regions on time vs. depth series represent eruptive periods. Note that earthquake symbols are scaled with magnitude in the map and cross-sections but not in the time vs. depth series. To increase figure clarity, we only plot highest quality earthquake locations.

(Pitt et al., 2002), including two DLPs beneath Medicine Lake Volcano, only one of which could be located (15 km depth), and 35 DLPs near Lassen Peak at depths of 13 to 27 km. For comparison, DLPs near Alaskan volcanoes are also located at mid-to-lower crustal depths, typically from 10 to 35 km (Power et al., 2004). Washington and Oregon DLP epicenters range from 0 to 20 km from the vent, similar to DLP epicenters in California and Alaska where horizontal offsets are also less than the average focal depths (Pitt et al., 2002; Power et al., 2004). Washington DLPs tend to loosely cluster together, similar to

those near Lassen Peak (Pitt et al., 2002). At most Alaskan volcanoes, DLP locations are widely scattered around the active vent (within one focal depth) leading Power et al. (2004) to infer that erupted magma may be drawn from a fairly broad area of the mid-to-lower crust.

5.2. Washington DLPs

DLPs at Mount Baker are located closer to the active vent than at other volcanoes in our study (Fig. 4). In contrast to the main Cascade
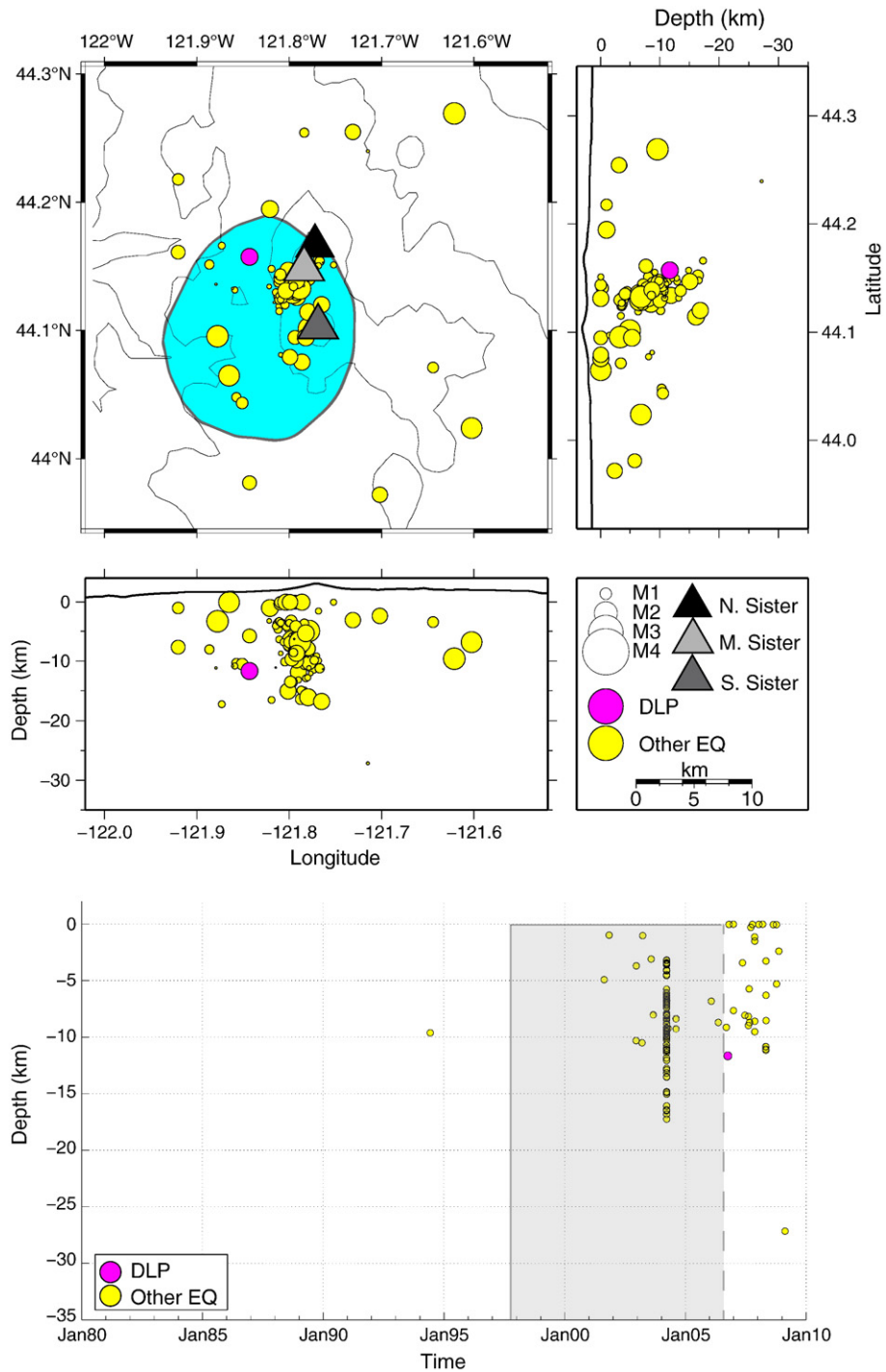
Fig. 8. Three Sisters seismicity map, E–W (bottom) and N–S (right) cross-sections, and time vs. depth series from Jan 1980 to Oct 2009. Black, light gray, and dark gray triangles represent the summits of North Sister, Middle Sister, and South Sister, respectively. Shaded region on map represents uplifted area. Black lines atop cross-sections represent topographic profile. Shaded region on time vs. depth series represents timing of inferred magmatic intrusion (Dzurisin et al., 2009). Note that earthquake symbols are scaled with magnitude in the map and cross-sections but not in the time vs. depth series.

arc from northern California to southern Washington, the northern Cascade volcanoes (including Mount Baker) tend to be more widely spaced, suggesting that perhaps crustal interception is so effective that Quaternary magmas have penetrated the crust only at a few intense or structurally favored loci (Hildreth et al., 2003). Assuming that DLPs are linked to magma movement in some fashion (e.g., White et al., 1996; Power et al., 2004), the concentration of DLPs both beneath and to the southeast of the Mount Baker edifice implies that

this is the most favorable location for deep magmas and/or magmatic fluids to enter the upper volcanic system.

DLPs at Glacier Peak are located near the SCF, which represents the southern section of the Fraser River–Straight Creek fault system that extends from Central Washington into British Columbia. Electromagnetic (EM) data from British Columbia suggest that the northern section of the fault is nearly vertical and extends through the entire crust, which may provide a conduit for escape of deeply penetrating

meteoric waters (Jones et al., 1992). Given the strike–slip geometry and extent of this fault system, we assume that the southern section near Glacier Peak is also nearly vertical, extends into the deep crust, and may also provide a path for the escape of fluids. Sloan Springs (Fig. 5) and Garland Mineral Springs, located along the SCF approximately 14 km west and 30 km southwest of Glacier Peak, respectively, provide evidence for fluid escape along the fault. Although both springs are cold, proprietary geochemical data collected from the springs yield elevated helium isotope ratios that suggest interaction with magmatic fluids (T. Cladouhos, personal communication, 2009). If the SCF penetrates the whole crust, then the concentration of DLPs near the fault suggests that magma and/or magmatic fluids also migrate along this structural boundary and interact with waters that eventually emerge at the surface through Sloan Springs and Garland Mineral Springs.

In southern Washington, seismicity and volcanic centers are concentrated near the edge of a dominant subsurface conductive body known as the Southern Washington Cascades Conductor (SWCC), an observation that has been used to infer that the SWCC influences the locations of Mount Rainier and Mount St. Helens (e.g., Stanley et al., 1987). It has been proposed that Mount Rainier and Mount St. Helens are located along block boundaries (expressed seismically as the WRSZ and SHZ), with Mount Rainier on the edge of SWCC and pre-Tertiary basement rocks to the east, and Mount St. Helens on the edge of SWCC and Siletzia rocks to the west (Stanley et al., 1996). Seismicity in the WRSZ and SHZ extends to mid-crustal levels, indicating that these structures likely penetrate much of the crust. These actively slipping and weak structural boundaries could enhance the ability of magmatic fluids to reach the surface, and the observation that DLPs at Mount Rainier and Mount St. Helens are located along the WRSZ and SHZ, respectively, provides evidence for fluid movement. As previously mentioned, the maximum and minimum compressive stress directions within the WRSZ and the SHZ are nearly horizontal (Giampiccolo et al., 1999) and are consistent with regional tectonics of Washington. However, a subset of WRSZ events located between 10 and 14 km contradicts this correlation by displaying steeply plunging (~70°) minimum stress axes. Giampiccolo et al. (1999) propose that if magma were rising at mid-crustal depths, it could increase the horizontal stresses around it and thus provide a source of east–west-oriented stress within the WRSZ, causing the intermediate stress direction to change from near vertical to near horizontal in that region. Giampiccolo et al. (1999) and Moran et al. (2000) invoke the occurrence of previously identified Mount Rainier DLPs (Malone and Moran, 1997) as evidence of magma at these depths along the WRSZ.

Further evidence of magma and/or magmatic fluids at mid-crustal depths comes from a tomographic study by Moran et al. (1999), who found a ~10-km-wide low-velocity anomaly between 5 and 18 km beneath the summit of Mount Rainier. The WRSZ occurs along the western boundary of this low-velocity region, which Moran et al. (1999) interpret to represent a region composed of hot, fractured rock with small amounts of melt or fluid. Moran et al. (2000) proposed that previously identified Mount Rainier DLPs result from movement of magma and/or magmatic fluids along the WRSZ, and we suggest this process can explain the subsequently identified DLPs at Mount Rainier as well. Like the WRSZ, the SHZ is a seismically active region of weakened, fractured rock that may provide paths of least resistance to the upper volcanic systems, and therefore we infer that DLPs at Mount St. Helens result from movement of magma and/or magmatic fluids along the SHZ.

Ascending melt from the upper mantle will encounter structural and rheological boundaries within the crust, and tectonic stresses and near-surface faults may play a significant role in the organization of volcanism (Vigneresse et al., 1999) by influencing pathways available to magma transport (Galland et al., 2007). Karlstrom et al. (2009) suggest that the presence of a volcanic system (edifices, magma chamber, and dikes) tends to cause a focusing, or "lensing" of magma transport from a broad region of the crust to a discrete volcanic center. This framework is consistent with our interpretation of DLPs tracing the flow of magma and/or magmatic fluids along fault systems. At shallow depths the magmas are then captured by the influence of the stress field created by the presence of the edifice or magma chamber, and thus erupt near the central vent.

5.3. Oregon DLPs

We identified 1 DLP each near Three Sisters and Crater Lake. Three Sisters has a similar detection threshold to Mount Baker and Glacier Peak where we observed many DLPs, and therefore we conclude that the scarcity of DLPs at Three Sisters is real. At Crater Lake, the poor station coverage may only allow identification of large DLPs, however the extremely low levels of seismicity in the area suggests the dearth of DLPs near Crater Lake may also be real. The one DLP near Three Sisters occurred in 2006 and is located within the northern section of the uplifted region and 3 to 5 km northwest of the 2004 earthquake swarm. It is possible that this DLP is related to the movement of magma and/or magmatic fluids associated with the inferred intrusion sitting ~7 km below the uplifted surface (Dzurisin et al., 2006).

5.4. Volcanoes with no DLPs

No DLPs were identified beneath Mount Adams, Mount Hood, Mount Jefferson, or Newberry Volcano. Mount Adams produces very little seismicity and has had no historical eruptions. It has a detection threshold similar to that of Mount Baker (Table 1), therefore the absence of DLPs is likely real. At Mount Jefferson and Newberry Volcano, the poor station coverage may allow identification of only large DLPs, however the sparseness of seismicity overall and lack of historical eruptions suggests the absence of DLPs at these locations may also be real. Mount Hood has relatively good station coverage (Table 1) and is seismically active (~40 earthquakes per year), thus the absence of DLPs is likely real as well. Mount Hood is the only Oregon volcano with historical eruptive activity, including the emplacement of the Crater Rock dacite dome ~200 years ago (Williams et al., 1982) along with small steam explosions and tephra fall reported by settlers from 1856 to 1865 (Folsom, 1970). Heat-flux estimates from fumaroles at Mount Hood imply that the heat loss due to the cooling of the dacite dome is not sufficient to explain the observed values, thus suggesting a deep-seated heat source (Friedman et al., 1982). Given the detection ability, level of seismicity, and the relatively recent historical eruption at Mount Hood, the absence of DLPs is notable.

As a rule, far fewer DLPs occur near Oregon Cascade volcanoes than those in Washington. We infer that this observation may in part result from differences in regional tectonic conditions. The sparse occurrence of DLPs in Oregon and the widespread occurrence of DLPs in Washington may be associated with regional extension and compression, respectively (McCaffrey et al., 2000). Mount Hood, located in the transition zone between these extensional and compressional regimes, is also positioned where the High Cascades partially override the Basin and Range extensional province. This hot, extended crust is subjected to extensional tectonic stresses as well as stresses from isolated volcanic processes (Jones and Malone, 2005) that may allow magma to move aseismically with respect to DLPs in the mid-to-lower crust. In general Oregon has a much higher heat flow than Washington (Blackwell et al., 1990), so perhaps the mid-to-lower crust is too ductile to support seismogenesis. We speculate that the combination of high heat flow and extensional conditions within the Oregon crust may allow magma to ascend more passively, while lower heat flow and compressional conditions within the Washington crust may present more resistance to ascending magma.

However, we acknowledge that this model does not explain DLP occurrence in all instances, in particular the DLPs reported at Lassen

Peak and Mammoth Mountain by Pitt and Hill (1994) and Pitt et al. (2002). The uneven occurrence of DLPs could also reflect variations in the composition of the mantle or in the subducting plate (e.g. water content). It is also possible that DLP activity is episodic, and that our 20-year dataset is not sufficient to determine which variable is most responsible for the scarceness or absence of identified DLPs near Oregon volcanoes.

6. Conclusions

We have identified a total of 60 DLPs beneath six Cascade volcanoes in Washington and Oregon from 1980 to October 2009. These events are inferred to represent the movement of magma and/or magmatic fluids within the mid-to-lower crust (10–50 km), and are characterized by mostly low-frequency energy (<5 Hz), emergent arrivals and long-duration codas. At Washington volcanoes, in cases where DLPs do not occur directly beneath the volcanic edifice, the locations coincide with large structural faults that extend into the deep crust. These features may provide paths of least resistance to the shallower volcanic system, and therefore the occurrence of DLPs within the fault zones likely highlight paths of magma ascent in the middle crust. We speculate that the scarcity of DLPs in Oregon is associated with high heat flow and extensional conditions in the crust that may allow magma to ascend passively, while the widespread occurrence of DLPs in Washington is associated with lower heat flow and compressional conditions in the crust that presents resistance to ascending magma and provides more suitable conditions for DLP generation. DLPs are an important part of background volcanic seismicity, and are especially significant because some have occurred prior to or in association with eruptions at several volcanoes. Although no DLPs were associated with volcanic activity in the Cascades, including the 1980–2086 and 2004–2006 eruptions at Mount St. Helens, we now have a better understanding of their occurrence in Washington and Oregon. This is important with respect to future eruption forecasting because DLP seismicity may be one of the earliest indications of renewed eruptive activity.

Acknowledgements

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