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1 Seismic, petrologic, and geodetic analyses of the 1999 dome-forming
2 eruption of Guagua Pichincha volcano, Ecuador

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20

21 Abstract

22 Guagua Pichincha, located 14 km west of Quito, Ecuador, is a

23 stratovolcano bisected by a horseshoe-shaped caldera. In

24 1999, after some months of phreatic activity, Guagua

25 Pichincha entered into an eruptive period characterized by
26 the extrusion of several dacitic domes, vulcanian eruptions,
27 and pyroclastic flows. We estimated the three-dimensional
28 (3-D) P-wave velocity structure beneath Guagua Pichincha
29 using a tomographic inversion method based on finite-
30 difference calculations of first-arrival times. Hypocenters
31 of volcano-tectonic (VT) earthquakes and long-period (LP)
32 events were relocated using the 3-D P-wave velocity model. A
33 low-velocity anomaly exists beneath the caldera and may
34 represent an active volcanic conduit. Petrologic analysis of
35 eruptive products indicates a magma storage region beneath
36 the caldera, having a vertical extent of 7-8 km with the
37 upper boundary at about sea level. This zone coincides with
38 the source region of deeper VT earthquakes, indicating that
39 a primary magma body exists in this region. LP swarms
40 occurred in a cyclic pattern synchronous with ground
41 deformation during magma extrusions. The correlation between
42 seismicity and ground deformation suggests that both respond
43 to pressure changes caused by the cyclic eruptive behavior
44 of lava domes.

45

46 **1. Introduction**

47

48 Pichincha Volcanic Complex, located in the Cordillera
49 Occidental of the Ecuadorian Andes (Fig. 1), consists of an
50 extinct older volcano (Rucu Pichincha), and a younger and
51 still-active edifice (Guagua Pichincha), which has been
52 constructed upon the western remnants of the older edifice
53 (Hall, 1977; Barberi et al., 1992). On the basis of new
54 geological, geochemical, and geochronological data, Monzier
55 et al. (2002) defined three main eruptive phases for Guagua
56 Pichincha volcano: (1) the basal Guagua Pichincha edifice,
57 composed of andesites and dacites, which underwent a sector
58 collapse leaving the huge avalanche caldera open to the
59 south-west; (2) the Toaza edifice, nested in the avalanche
60 caldera and also composed of andesites and dacites; and (3)
61 the active dome complex, mostly of dacitic composition,
62 which has developed inside a second sector collapse open to
63 the west that affected the Toaza domes.

64

65 In the late Holocene era, and prior to the most recent
66 eruptive activity in 1999-2001, Guagua Pichincha volcano
67 experienced at least three eruptive episodes (Monzier et
68 al., 2002, 2006): ~2000 years BP, ~1100-1200 years BP, and

the last historic eruptive phase in 1566-1582 and 1660 AD. The recent eruptive activity was characterized by Plinian eruptions, which occurred during, or at the end of, dome-forming activity. The Plinian eruptions produced moderate tephra fall in Quito and accompanying pyroclastic flows that swept the western valleys (e.g. in the Rio Cristal, Fig. 1). Following the magmatic activity in 1660 AD, phreatic activity has been reported in 1830-1831, 1868-1869, 1881, and 1981-1998 (Simkin and Siebert, 1981; Hall, 1990).

Between 1981 and 1998, the phreatic activity at Guagua Pichincha was correlated with seasonal rain. This activity was characterized by persistent fumarolic emissions and small phreatic explosions affecting only the area inside the caldera and creating new craters in the intra-caldera dome complex. Marini et al. (1991) performed geochemical analyses of hydrothermal fluids from Guagua Pichincha fumaroles and springs and inferred a shallow aquifer, located at a few hundred meters beneath the surface, which was fed by magmatic fluids from depth. The vapor phase produced by boiling of this aquifer feeds the SO₂-rich fumaroles located in the crater area.

92 Guagua Pichincha volcano has been monitored by a telemetered
93 network of eleven 1 Hz seismometers, which have been
94 operated by the Instituto Geofísico (IG), Escuela
95 Politécnica Nacional (Quito-Ecuador) since 1981 (Fig. 1). A
96 portable station featuring a Lennartz seismometer with
97 bandwidth of 0.2-40 Hz and a Marslite 20-bit data logger was
98 additionally operated at Pichincha between November 1999 and
99 July 2001 (REFU, Fig. 1). Long-period (LP) and hybrid
100 events, tremor, volcano-tectonic (VT) earthquakes, and
101 volcanic explosions have been recorded by the Pichincha
102 seismic network.

103

104 In 1998, intense phreatic activity marked the start of a new
105 eruptive period of Guagua Pichincha, accompanied by shallow
106 LP and hybrid events and tremor. Intense VT and LP swarms,
107 occurring in 1999, were followed by the emplacement of
108 dacitic domes within the caldera of the volcano. Domes grew
109 recurrently and were destroyed by explosions or dome-
110 collapses, which generated pyroclastic flows that descended
111 the Rio Cristal valley (Fig. 1) and deposited ash around the
112 volcano as well as into Quito, capitol city of Ecuador.

113

114 To better understand the magmatic plumbing system beneath
115 the volcano, we performed a tomographic inversion of Guagua

Pichincha to determine the three-dimensional (3-D) P-wave velocity structure of the volcano. The 3-D P-wave velocity model was used to relocate the hypocenters of VT earthquakes and LP events. We also performed petrologic analysis of rock samples from the 1999 eruption to constrain the magma storage and ascent processes. Geodetic data were further analyzed to interpret deformation patterns during the dome extrusions. By integrating the multi-disciplinary data, we characterize the 1999 eruptive episode and present implications for the magma plumbing system beneath Guagua Pichincha volcano.

2. Overview of the 1998 – 2001 activity at Guagua Pichincha

During 1998 significant seismic activity occurred 15 km NE of the caldera. This activity began in May 1997 (Calahorrano, 2001; Legrand et al., 2002) and has been categorized as a swarm of “Distal-volcano-tectonic events” (White, 2001). Following this activity, recurrent phreatic explosions, long-duration tremor, VT earthquakes (VT-S1, Fig. 2), and swarms of shallow LP and hybrid events occurred at Guagua Pichincha volcano from September 1998 (Villagómez, 2000). Strong tremor signals lasting days to weeks were recorded from March to May 1999, accompanied by vapor-rich

140 phreatic explosions and a swarm of VT earthquakes (VT-S2).
141 Another swarm of VT earthquakes occurred in June 1999 (VT-
142 S3). This swarm was characterized by fewer events and larger
143 magnitudes (coda duration magnitudes, $M_c > 3.0$) than those in
144 VT-S2. This swarm was also accompanied by ash-rich
145 explosions. VT swarms repetitively occurred from 8 to 24
146 September 1999 (VT-S4), during which the strongest pre-dome
147 VT swarm (the daily number of 218 earthquakes) took place on
148 23 September. During this swarm more than 1000 earthquakes
149 occurred in 17 days. On 25 September, the first intense LP
150 swarm occurred in association with dome emplacement (LP-S1).
151 Following an explosion containing juvenile magma on 26
152 September, the first dacitic dome growth occurred between 27
153 and 29 September. On 30 September, LP seismicity decreased,
154 and the first dome was destroyed by Vulcanian-type magmatic
155 explosions that occurred on 5 and 7 October. The first
156 explosion produced 2 mm of ash fall in Quito.

157

158 VT swarms took place between 15 and 27 October 1999 (VT-S5),
159 in which the maximum daily number of earthquakes (444) was
160 recorded on 24 October. This activity accompanied new LP
161 activity (LP-S2), which had begun on 18 October and
162 intensified on 24 October. The second dome, characterized by
163 its spine shape, grew during this period and was destroyed

164 at the beginning of November. Subsequently, from 10 to 17
165 November, the third dome grew and collapsed. Continuing
166 magmatic activity was accompanied by intense LP swarms
167 between 17 and 26 November, during which the fourth dome
168 grew and was destroyed by a series of explosions on 26
169 November, causing the most significant ash fall in Quito.
170 Between 26 and 28 November, a new VT swarm (VT-S6) occurred
171 and was followed by intense LP swarms (LP-S3) until 17
172 December 1999. During this period, three short-lived
173 dacitic domes were extruded and destroyed by explosions
174 and/or dome-collapses. The dome destruction was accompanied
175 by ash fall around the volcano and pyroclastic flows
176 descending the Rio Cristal valley (Fig. 1).

177

178 In late December 1999, the growth of the eighth dome began,
179 accompanied by intense LP activity (LP-S4). This dome
180 partially collapsed in late June 2000. Subsequently, the LP
181 activity increased between the end of July and the middle of
182 December 2000 (LP-S5), during which another dome slowly grew
183 and still remains intact (as of December 2005) on the
184 caldera floor (Fig. 3). This activity showed some periods of
185 intense LP activity alternating with periods of low-level
186 activity. In March 2001 LP events were almost totally
187 absent.

188

189 LP swarms at Guagua Pichincha generally lasted hours to
190 several days. At the beginning of each swarm, LP events are
191 randomly spaced in time. Later, the time interval between
192 consecutive events (dt , Fig. 4) gradually decreases. When
193 events are regularly separated by only a few seconds, they
194 merge into tremor. Eventually, dt again becomes variable as
195 the swarm terminates, sometimes following an explosion.
196 Individual LP events are very self-similar, as seen in Fig.
197 5.

198

199 **3. Tomographic inversion of P-wave velocity structure**

200

201 *3.1. Method and data*

202

203 We used the tomographic inversion method of Benz et al.
204 (1996), in which first-arrival times are calculated with the
205 Eikonal equation in a finite-difference scheme. Following
206 the development of a finite-difference method in two
207 dimensions by Vidale (1988), the method was extended to
208 three dimensions by Vidale (1991) and Podvin and Lecomte
209 (1991). The finite-difference method is robust for the
210 computation of direct, diffracted, and refracted waves in
211 complex velocity structures with large velocity contrasts

and extreme topography (Benz et al., 1996). These techniques have been successfully applied to image high-resolution 3-D velocity structures in various volcanoes (Benz et al. 1996; Okubo et al., 1997; Villaseñor et al., 1998; Dawson et al., 1999; Tomatsu et al., 2001; Molina et al., 2005).

We used first-arrival times measured from VT earthquakes that occurred beneath Guagua Pichincha between 1998 and 1999. We selected VT earthquakes whose onsets were measured at 5 or more stations in the telemetered network (solid triangles in Fig. 1). 3904 P-wave arrivals from 641 earthquakes were used in our inversion. We could not measure reliable and sufficient S-wave arrival times due to the lack of horizontal-component seismometers except for two stations (PINO and TERV) (Fig. 1). For this reason we did not attempt to perform the tomographic inversion of the 3-D S-wave velocity structure in this study.

3.2. Initial model parameters

We defined the model domain with a lateral dimensions of 12 by 12 km centered on the caldera center (-0.16° S, -78.62° W) with a vertical dimension ranging between -5 and 10 km from sea level, where depth is positive downward. The

finite-difference calculations of first-arrival times and tomographic inversion were performed over a uniform grid spacing of 0.2 km. We derived the initial one-dimensional (1-D) velocity model based on previous studies. We used the 1-D velocity model for Guagua Pichincha from Villagómez (2000). Using the methods of Chapman and Bollinger (1984) and Chatelain (1978), Villagómez (2000) found that the P-wave velocity at Guagua Pichincha linearly increases with a constant P to S-wave velocity ratio down to a depth of 3 km below sea level. For greater depths we used the 1-D P-wave velocity model for Ecuador from Bernal (1996). The combined 1-D P-wave velocity model is shown in Fig. 6. We relocated the hypocenters of VT earthquakes using the inversion method of Benz et al. (1996) for the fixed 1-D P-wave velocity model. This resulted in a Root Mean Square (RMS) residual of 0.09 s. This initial RMS residual value is small compared to those of other tomographic studies at other volcanoes (e.g., Benz et al., 1996; Dawson et al., 1999; Molina et al., 2005). This is evidence that the initial model reasonably represents the average velocity structure of Guagua Pichincha. This 1-D model and associated hypocenters were used as our initial conditions for the tomographic inversion.

3.3. Results

260

261 The tomographic inversion with corrections for origin times
262 and source locations was run iteratively 10 times, during
263 which the smoothing parameter k (Benz et al. 1996) decreases
264 uniformly from 11 to 2. This resulted in increasing model
265 roughness with each iteration and stable convergence to a
266 solution (Benz et al., 1996). The P-wave arrival time RMS
267 for all events was reduced from 0.09 to 0.05 s. Fig. 7 shows
268 hypocenter locations for the VT earthquakes using both the
269 1-D and 3-D P-wave velocity models. As can be seen in Fig. 7,
270 the hypocenters in the 3-D model are more tightly clustered
271 compared to those in the 1-D model. The main cluster of
272 events is located at depths between -3 and -1 km from sea
273 level.

274

275 Fig. 8 shows plan views of the 3-D P-wave velocity structure
276 at selected depths ranging from -3.6 to -1.6 km. Fig. 8a
277 shows the layer between -3.6 and -3.4 km, in which low
278 velocities (~ 2.3 km/s) are seen in the SE rim of the caldera.
279 High velocities exist beneath the N, SE, and SW rim of the
280 caldera. In the layers between -3.2 and -3.0 and between -
281 2.6 and -2.4 km (Figs. 8b and 8c), we can clearly recognize
282 a low-velocity anomaly (~ 2.5 km/s) in the center of the
283 caldera. This low-velocity anomaly is located to the east of

the 1999 domes, nearly beneath the 1981 crater (see Fig. 3), and is surrounded by high velocities underlying the caldera rim. Such high velocities beneath the caldera rim are also seen in the layer between -2.4 and -2.2 km (Fig. 8d). The deeper layers (Figs. 8e and 8f) are dominated by high velocities of 4.5-5.5 km/s.

Fig. 9 shows N-S (A-A') and E-W (B-B') cross sections through the 3-D P-wave velocity model (profile locations are indicated in Fig. 8c). The low velocity anomaly in the center of the caldera, which are observed in the plan views (Fig. 8b and 8c), are clearly seen in the cross sections in Fig. 9. The low velocity anomaly (1.5 to 2.3 km/s) exhibits a vertical structure extending from the caldera floor to a depth of -3 km beneath the active dome complex. At this depth, the low velocity anomaly extends horizontally to the west as seen in Fig. 9b. A layer characterized by velocities of 3.5-4.0 km/s is seen in most of the volcanic edifice down to a depth of -2.0 km. This layer is underlain by a high-velocity layer characterized by velocities of 4.5-5.5 km/s.

3.4. Checkerboard tests

We conducted checkerboard reconstructions to provide estimates of the model resolution and accuracy. Synthetic P-wave arrival times were computed using the hypocenters determined with the 3-D P-wave velocity model and a checkerboard velocity model with a velocity perturbation of ± 10 per cent from the 1-D velocity model. Figs. 10a and 10b show plan views of the input checkerboard and reconstructed models, respectively, for the layer between -3.0 and -2.8 km. Cells that are not sampled by any rays are represented by blank space. The checkerboard patterns are almost perfectly reconstructed in the central region beneath the caldera in this layer. Figs. 10c and 10d display N-S (A-A') cross sections of the input checkerboard and reconstructed models respectively. E-W (B-B') cross sections of the input checkerboard and reconstructed models are shown in Figs. 10e and 10f respectively. The checkerboard patterns are reproduced in the central region down to sea level, although the recovered amplitudes become smaller as the depth increases. In the regions below sea level and beneath the southern flank, smearing occurs due to the small number of rays.

3.5. Accuracy of hypocenters

To estimate the accuracy of the hypocenters determined with our 3-D velocity model, the following simple statistical test was performed: we generated random errors within a time interval of ± 0.1 s, which were added to observed first-arrival times to generate test data for each VT earthquake. The hypocenter was then relocated by using the test data and fixed 3-D velocity model. We repeated this process 100 times and estimated the mean and standard error of the hypocenter. We selected 48 VT earthquakes to obtain uniform horizontal and vertical distributions of the hypocenters. The error interval of ± 0.1 s was chosen based upon the uncertainty in first-arrival picks made by different staff in the IG. Fig. 11 shows the mean locations and associated standard errors estimated for the 48 earthquakes. Shallow earthquakes down to sea level are located with small errors (< 1 km). On the other hand, the hypocenters of the earthquakes located below sea level have large errors (2-3 km) in both the horizontal and vertical directions. The errors are larger in the earthquakes located to the north and west of the caldera due to the poor azimuthal coverage of stations.

4. Petrologic constraints on magma storage and ascent processes

Geochemical and mineralogical analysis of dome samples from the 1999 eruption was performed to constrain the magma storage and ascent processes. Our analysis was aimed at comparing petrologic data with the results from the seismic study. A complete petrological description of Guagua Pichincha dome rock is beyond the scope of this manuscript and will be presented elsewhere. The 1999 dacitic domes at Guagua Pichincha are characterized by remarkable compositional homogeneity (64.1-65.1 wt.% SiO₂), which lies within the compositional range of other Holocene Guagua Pichincha eruptive products (Monzier et al., 2002). Hot samples were collected in pyroclastic flow deposits in the Rio Cristal valley (Figs. 1 and 12a) as well as in ash and tephra deposits from Vulcanian explosions. The dacite is porphyritic (~30-35 vol.% crystals) with phenocrysts of plagioclase (10-20 vol.%), amphibole (<10 vol.%), orthopyroxene (<5 vol.%), and Fe-Ti oxides (<1-2 vol.%). Clinopyroxene and apatite appear as accessory phases, and olivine xenocrysts are also found. The groundmass is mostly glassy with a rhyolitic composition (73.1-79.6 wt.% SiO₂) and bear microlites of the same mineral assemblage, except that clinopyroxene is absent.

378 Mineral compositions of dome rocks (sampled on 24 and 30
379 November and 10 December 1999, following a major dome
380 collapse) were analysed by electron microprobe. Amphibole is
381 the most important mafic mineral in the Guagua Pichincha
382 dacites. Following the classification scheme of Leake et al.
383 (1997), these minerals are magnesium hornblende, with some
384 tschermakite and magnesium hastingsite phenocrysts. It is
385 noteworthy that all amphibole crystals are transparent,
386 euhedral phenocrysts without any kind of reaction rims (Fig.
387 12b). These phenocrysts display a large variation in Al_2O_3
388 contents (6.2-14.7 wt.% Al_2O_3), in which two main
389 populations can be distinguished: (1) low-Al amphibole (6.2-
390 9.6 wt.% Al_2O_3), corresponding to the magnesium-hornblende
391 phenocrysts; and (2) high-Al amphibole (10.9-14.7 wt.%
392 Al_2O_3), corresponding to scarce tschermakite and magnesium-
393 hastingsite minerals (Fig. 12c).

394

395 The Al-in-hornblende geobarometer (Johnson and Rutherford,
396 1989) can be used to estimate the total pressure in the pre-
397 eruption magma storage region. The geobarometer posits that
398 the aluminum content of the amphibole is proportional to
399 total pressure in special conditions given by a specific
400 mineral assemblage composed of plagioclase, sanidine,
401 biotite, quartz, sphene, and magnetite/ilmenite. In the case

402 of Guagua Pichincha dacite, quartz is not present in the
403 equilibrium assemblage, and thus the geobarometer could
404 yield anomalously high pressures (Johnson and Rutherford,
405 1989). However, given that the interstitial glass has a
406 high-silica composition (>73 wt.% SiO_2), we consider that
407 the melt is close to silica saturation. Thus, we may apply
408 the geobarometer to the amphibole rims, which were in
409 equilibrium with the matrix glass as well as with other
410 phases. We used the PROBE-AMPH spreadsheet (Tindle and Webb,
411 1994) to determine the amphibole structural formulae (based
412 on 23 oxygens and 13 cations) as well as pressure estimates.
413 The geobarometer yields a range from 100 to 300 MPa for
414 amphibole crystallization, which corresponds to the low-Al
415 amphibole population. The corresponding crystallization
416 depth of these amphiboles ranges from 0 to 8 km below sea
417 level (assuming a bulk rock density of 2600 kg/m^3). This
418 range of pressure for amphibole crystallization in calc-
419 alkaline magmas is consistent with experimental studies in
420 siliceous andesites and dacites like Mount St. Helens
421 (Rutherford and Hill, 1993); Pinatubo (Scaillet and Evans,
422 1999; Prouteau and Scaillet, 2003), Monserrat (Barclay et
423 al., 1998), and Redoubt (Browne and Gardner, 2006).

424

To evaluate the potential effects of pressure and temperature on amphibole composition, we plotted the Al content against a temperature-sensitive component as, for example, Na+K (Scaillet and Evans, 1999). In the same diagram (Fig. 12c) we showed the region of experimentally-determined amphibole composition for the Pinatubo dacite at different pressures (Prouteau and Scaillet, 2003). The low-Al group overlaps the experimentally determined amphibole composition at 220 MPa and 785-900°C. This correlation corroborates our pressure estimation based on Al-in-hornblende geobarometer. As to the high-Al (and high Na+K) amphibole population, the most likely interpretation is that these minerals are one of the first crystallized phases, which fractionate at higher pressures (> 300 MPa) and temperatures. We note that this interpretation does not affect our pressure estimation for the low-Al amphibole group.

In a volatile-rich magma, a decrease of H₂O content produces amphibole breakdown reactions, the extent of which is a function of magma decompression or ascent rate (Rutherford and Hill, 1993; Browne and Gardner, 2006). The lack of reaction rims in low-pressure amphibole phenocrysts of the Guagua Pichincha dacitic domes suggests that they were in

equilibrium with the liquid phase or that the magma ascended relatively fast to avoid formation of a reaction rim due to devolatilisation reactions. Rutherford and Hill (1993) performed an experimental calibration of the amphibole breakdown reaction in Mount St. Helens rocks. Given the compositional similarities between St. Helens and Guagua Pichincha dacites, we use the "reaction-rim-width versus time" curves of Rutherford and Hill (1993) to estimate the Pichincha magma ascent rate. Accordingly, we conclude that the magma should have ascended in less than 4-6 days from the top of the estimated magma storage region (i.e., the low-pressure zone). This estimate suggests magma ascent rate high enough (28-42 m/h) to avoid the formation of amphibole reaction rims.

On the basis of these pressure estimates we conclude that (1) the upper magma storage region at Guagua Pichincha has a vertical extent of 7-8 km with the upper boundary near sea level (4 km below the caldera floor) and (2) that magma ascent rate from the top of the magma storage region is high enough to avoid the formation of amphibole reaction rims.

5. Ground deformation recorded by a tiltmeter

473 A cyclic pattern, with time scales similar to the occurrence
474 of VT and LP swarms and dome extrusions, is also
475 distinguished in ground deformation records. Eruption-
476 related seismicity increases and decreases contemporaneously
477 with changes in ground-surface tilt recorded by a tiltmeter
478 located at station REFU. To detect these cyclic events a
479 dual-axial platform, analog-output tiltmeter, AGI-Model 711-
480 2A with a one microradian resolution, was installed at
481 station REFU in December 1998 (Fig. 1). The data was
482 telemetered via UHF radio every 12 minutes to the IG. The
483 tilt sensor was placed in a cement hut built upon a massive
484 lava outcrop. The sensor was set within a concrete-lined,
485 1-meter deep covered hole, which helped to minimize diurnal
486 temperature effects in the data. The distance between REFU
487 station and the active dome growth is about 2.5 km. For
488 many months after installation, including a period of
489 phreatic explosions in the caldera, the instrument showed no
490 notable changes. However, between September and December
491 1999, at least five periods of inflation/deflation pattern
492 were observed (INF-1 to INF-5, Fig. 13). Inflation was
493 preceded or started almost synchronously with VT swarms: VT-
494 S4, VT-S5, VT-S6 and VT-S7, which were followed respectively
495 by LP swarms: LP-S1, LP-S2, LP-S3 and LP-S4 (Fig. 13).
496 Several of the inflationary episodes had small changes less

than 10 microradians. The most significant value of 26 microradians was associated with the period from 18 to 21 November 1999 and corresponded to the growth of the fourth dome with total volume about 0.5 million m³. With the collapse of the fourth dome on 26 November, a marked deflationary rebound occurred in the tilt data, showing that the inflation was temporary and was likely responding to the pressure buildup beneath a rigid plug of the dome.

Deflation was generally observed after explosions or possibly when the extrusion rate at the dome increased. The close correlation between seismicity and ground deformation suggests that both responded to pressure changes, especially where ascending magma caused pressurization beneath the plug of the dome. Dome destruction causes a release of pressure and corresponding deflation, a process similar to that described by Voight et al (1998) for deformation cycles at Monserrat. Similar observations have been documented at Pinatubo in 1991 (Hoblitt et al., 1996; Mori et al., 1996).

6. Discussion

6.1. 3-D P-wave velocity structure

Our tomographic inversion of P-wave velocities for Guagua Pichincha volcano reveals the following features (Figs. 8 and 9): (1) in the layer between -3.6 and -3.4 km from sea level (Fig. 8a), a low-velocity zone is located in the SE part of the caldera and two high-velocity zones are located in the eastern and southern boundaries of the Toaza caldera; (2) a vertical low-velocity zone is located in the center of the caldera at depths between -3.2 to -2.8 km (Figs. 8b and 8c and Fig. 9); (3) a horse-shoe shaped structure is open to the SW at depths between -2.8 and -2.0 km; and (4) a major velocity contrast is evident at a depth around -2.0 km.

(1) The shallow low-velocity zone in the SE part of the caldera may be associated with a very active fumarole field, known as "La Locomotora" (Fig. 1). The fumaroles in this field did not exhibit any changes in activity during the 1999 eruptions, probably reflecting an independent phreatic system. The high-velocity zones may be interpreted as massive lavas from both the Guagua Pichincha and Toaza edifices.

(2) The low-velocity anomaly located in the center of the caldera, where the hypocenters are clustered along a vertical structure, is interpreted as the active conduit zone. The low velocities, however, may also represent poorly

consolidated and highly fractured deposits filling the caldera. The horizontal extension of the low velocity anomaly, at a depth around -3 km (Fig. 9b), is suggestive of such deposits filling the caldera following the sector collapse.

(3) The low-velocity anomaly mentioned above is surrounded by high velocities, which roughly outline the caldera rim (Figs. 8b-d). This feature indicates that these high velocities are related to the pre-collapse Guagua Pichincha edifice or the Toaza edifice.

(4) The velocity contrast observed at a depth of ~2.0 km may represent the contact between the volcano-sedimentary basement and the volcanic edifice.

6.2. Temporal evolution of VT and LP activity

Fig. 14 shows the temporal evolution in the hypocentral depths of VT earthquakes during the period from July 1998 through December 1999 (blue-filled diamonds). We clearly recognize intense VT activity above sea level, which was accompanied or preceded by deeper VT activity. Based upon the data, we divide the VT activity into 4 periods. During the first period, between September and November 1998, the seismic activity increased beneath the volcano, with shallow

569 LP events and tremor and some deeper VT earthquakes (VT-S1).
570 The second period between January and March 1999 is
571 characterized by increased VT activity (VT-S2). During the
572 third period, between April and June 1999, the increased VT
573 activity (VT-S3) was accompanied by shallow hydrothermal
574 activity with energetic phreatic explosions and long-lasting
575 tremor. The fourth period started with the increased VT
576 activity (VT-S4) and culminated in the extrusion of a dome
577 at the end of September 1999. The VT activity during this
578 period began with deep VT earthquakes at depths around 10 km
579 below sea level and with the coda duration magnitude (M_c)
580 greater than 3.0. There exists a clear trend that the focal
581 depths of VT earthquakes migrated toward the surface during
582 this period. The general trend of decreasing focal depths is
583 considered significant in view of the accuracy of the
584 hypocenter determinations as estimated by our statistical
585 test (see Fig. 11). We also relocated the hypocenters of LP
586 events associated with dome extrusions during September-
587 December 1999 using the 3-D P-wave velocity model. The
588 temporal change in the focal depths of these LP events is
589 shown in Fig. 14 (red-filled triangles). These LP events
590 range from the surface to a depth of -1 km beneath the
591 caldera. We do not recognize any clear trend in the observed
592 temporal change in the focal depths of LP events.

593

594 *6.3. The 1999 eruption and magma plumbing system beneath*
595 *Guagua Pichincha volcano*

596

597 The pressure estimations from the Al-in-hornblende
598 geobarometer indicate that the upper magma storage region at
599 Guagua Pichincha has a vertical extent of 7-8 km with the
600 upper boundary near to sea level. The estimated region is in
601 good agreement with the source region of the deeper VT
602 earthquakes (see Figs. 7 and 14). This feature suggests that
603 a main magma body exists in this region and the deeper VT
604 earthquakes occurred as the brittle response of volcanic
605 rock to fracturing by magma movement. Unfortunately, our
606 tomographic image has poor resolution at such depths,
607 hindering our ability to precisely define the magma store
608 region. Fig. 15 is our interpretation of the magma plumbing
609 system beneath Guagua Pichincha volcano.

610

611 It is most likely that the magma ascent began during the
612 first period of the VT activity starting in September 1998
613 (VT-S1, Fig. 14). The intruded magma heated shallow
614 hydrothermal systems, which triggered LP and hybrid events.
615 A second magma intrusion may have occurred between January
616 and March 1999 in association with VT-S2 (Fig. 14).

617 Following VT-S2, intense phreatic activity was observed in
618 April-May 1999. This activity was characterized by several
619 vapor-rich explosions, accompanying strong LP events and
620 tremor that often saturated the seismic records for days or
621 weeks in stations located around the caldera. This intense
622 phreatic activity was also accompanied by shallow VT
623 earthquakes, which may be attributed to the fracturing of
624 volcanic rocks by magma intrusion into the shallower conduit
625 system. In April and May 1999, new deep VT activity took
626 place (VT-S3, Fig. 14). During this period explosion columns
627 were characterized by their increased ash contents. It was
628 also evident that the time durations of explosions, as
629 measured by the seismic waveforms, were significantly
630 shorter than those recorded in the preceding months. This
631 suggests a depleted hydrothermal reservoir.

632
633 8 September marked the start of a sequence of VT earthquakes
634 which ended with the first extrusion of new magma at the end
635 of September. From 8 to 24 September, VT swarms were
636 recurrent (VT-S4) and became shallower with time, eventually
637 reaching the shallow VT source located at a depth of about -
638 2 km above sea level (Fig. 14). The VT swarms were followed
639 by an inflation episode (INF-1) (Fig. 13). The VT swarms may
640 represent the fracturing of volcanic rocks associated with

magma ascent. Magma reached the shallower regions, and caused the surface deformation detected by the tiltmeter. 25 September marked the start of the first intense LP swarm (LP-S1) related to dome extrusion (Fig. 13). The LP event is interpreted as the oscillatory acoustic response of a fluid-filled resonator in a pressurized system (e.g., Chouet, 1996). Supersaturated conditions were reached in the magma that intruded into the shallower regions. The increased pressure inside the magma through decompression-induced degassing and vesicle growth may have repetitively triggered the resonance of a magmatic conduit (Neuberg et al., 1998; Voight et al., 1998) or newly-formed crack containing a particle-gas mixture (Tuffen et al., 2003; Tuffen and Dingwell, 2005) during the LP swarm. It is thought that the first dome surfaced by 29 September, when the surface deformation turns from inflation into deflation (Fig. 13). The LP swarm tapered off after 30 September, since supersaturated volatiles in the magma may have been depleted due to degassing. After 5 days of low seismic activity, the dome was destroyed by a Vulcanian explosion.

New VT swarms took place between 15 and 17 October (VT-S5), which were also associated with an inflation episode (INF-2) (Fig. 13). This feature indicates that a new batch of magma

665 intruded into shallower regions. The VT swarms continued
666 during the inflation, because the fracturing near the
667 surface may be more extensively occurred in association with
668 the dome extrusion during this period. LP swarms (LP-S2)
669 also occurred after the VT swarms. Unlike the previous LP
670 swarm (LP-S1), LP-S2 continued persistently during a
671 deflation phase (Fig. 13). Extensive dome growth and
672 collapse during this period may have continuously removed
673 magma near the surface. This resulted in sustained
674 decompression-induced degassing inside magma, and triggered
675 LP events in a continuous manner.

676

677 We observed three inflation episodes after INF-2 until the
678 end of December 1999 (Fig. 13). VT activity associated with
679 these inflation episodes were low compared to VT-S4 and VT-
680 S5 (Fig. 13). This may be because the magma path way to the
681 surface was already established and magma easily ascended
682 through this path without fracturing or VT activity. New
683 magma was supplied and extruded in the individual inflation
684 episodes. LP events were also triggered in these episodes.

685

686 The durations of the inflation episodes range from 4 to 8
687 days (Fig. 13). The fact from the petrologic analysis that
688 Guagua Pichincha amphiboles does not show reaction rims also

689 suggests a fast ascent time, less than 4 - 6 days. If we
690 assume that these durations correspond to the magma ascent
691 durations from the upper part of the magma storage zone (4
692 km beneath the caldera floor) to the surface, we can
693 estimate an ascent rate of 21-42 m/h (or equivalently 0.6-
694 1.2 cm/s). This value is in concordance with a minimum magma
695 ascent rate of 0.5 cm/s during cycles of Vulcanian eruptions
696 at Guagua Pichincha estimated by Wright et al. (2006) from
697 morphological analyses of breadcrust bombs erupted in 1999.
698 Such an ascent rate also agrees with the estimates made on
699 other dome-forming eruptions: Montserrat (Devine et al.,
700 1998), Unzen (Venezki and Rutherford, 1999; Nakada and
701 Motomura, 1999), and St. Helens (Rutherford and Hill, 1993).

702

703 Models to explain the cyclic eruptive behavior of lava domes
704 have been proposed by Denlinger and Hoblitt (1999), Voight
705 et al. (1999), and Barmin et al. (2002). Dome growth may
706 reflect complex dynamics between magma supply and flow
707 resistance (Hoblitt et al., 1996). Cyclic behavior occurs
708 when the supply rate is within a particular range; above and
709 below that range the output flux approaches the supply rate
710 without oscillation (Denlinger and Hoblitt, 1999). The model
711 of Denlinger and Hoblitt (1999) may explain the Guagua
712 Pichincha eruptive behavior. It contends that cyclic

behavior in volcanic eruptions result from Newtonian flow of compressible magma through the volcanic conduit combined with a stick-slip condition along the conduit wall. If the magma flux increases sufficiently, the shear stress exceeds the shear strength of the magma-wall contact and slip occurs. The onset of slip only slightly affects the magnitude of the shear stress along the wall, but it greatly increases the flow rate. With wall slip, the magma behaves as a Newtonian fluid but has an abnormally high flow rate for the same pressure drop in the conduit (Denlinger and Hoblitt, 1999). If the enhanced flow rate exceeds the supply rate, the conduit-pressure drops, the output flux decreases until slip ceases as the magma reattaches to the conduit wall. The pressure-drop in the conduit motivates gas to be released from the magma, providing the compressibility required. Energy can be alternately stored and released as the magma alternately sticks and slips along the conduit wall.

7. Conclusions

We estimated the 3-D P-wave velocity structure beneath Guagua Pichincha volcano using a tomographic inversion method based on finite-difference calculations of first-

737 arrival times. The 3-D P-wave velocity model was used to
738 relocate VT earthquakes and LP events associated with the
739 1999 eruption. A low velocity anomaly exists beneath the
740 caldera, where the hypocenters of VT earthquakes are
741 vertically aligned at depths between -3 and -1 km from sea
742 level. The low velocity may represent an active volcanic
743 conduit during the 1999 eruption.

744
745 VT earthquakes are also located at depths between 0 and 10
746 km below sea level. We find that the shallow VT activity was
747 accompanied or preceded by the deeper VT activity.
748 Petrologic analysis of eruptive products based on the Al-in-
749 hornblende geobarometer indicates that a magma storage
750 region beneath Guagua Pichincha has a vertical extent of 7-8
751 km with the upper boundary at about sea level. The estimated
752 region is in good agreement with the source region of the
753 deeper VT earthquakes. This feature suggests that a main
754 magma body exists in this region, and the deeper VT
755 earthquakes occurred as the brittle fracturing of volcanic
756 rock during magma ascent.

757
758 LP swarms occurred cyclically during magma emplacement in
759 the upper conduit. The LP events show similar waveforms,
760 suggesting a repetitive and nondestructive source process.

761 The cyclic pattern is also observed in ground deformation.
762 The correlation between seismicity and ground deformation
763 indicates that both respond to pressure changes caused by
764 the cyclic eruptive behavior of lava domes.

765

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767

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780

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980

Figure captions

Fig. 1. Location of seismic stations (solid triangles) at Guagua Pichincha volcano. A tiltmeter and portable three-component seismometer were installed at station REFU (star). The caldera rim is shown by dashed lines. The inset indicates the location of the Pichincha volcanic complex, Ecuador.

Fig. 2. Daily number of VT earthquakes and LP events recorded at Guagua Pichincha between July 1998 and June 2001.

Fig. 3. Views of the caldera (a) and domes (b) at Guagua Pichincha volcano, as seen in November 2003 (IG-BGR).

Fig. 4. (a) A 24-hour record of the vertical component of velocity at station REFU showing an LP swarm that occurred on 2 December 1999. (b) Enlargement of the waveform in the period between A and A'.

Fig. 5. LP waveforms observed at station REFU. These events occurred on 2 December 1999 during an LP swarm.

1005

1006 Fig. 6. The initial 1-D P-wave velocity model as a function
1007 of depth used in our tomographic inversion.

1008

1009 Fig. 7. Plan views (a, b) and N-S (c, d) and E-W (e, f)
1010 cross sections of hypocenters of VT earthquakes determined
1011 with the 1-D (a, c, e) and 3-D (b, d, f) models.

1012

1013 Fig. 8. Plan views of the 3-D P-wave velocities between -3.6
1014 and -1.4 km from sea level. Seismic stations (solid
1015 triangles), contoured topography, and caldera rim (dashed
1016 lines) are shown for reference.

1017

1018 Fig. 9. N-S (a) and E-W (b) cross sections of the 3-D P-wave
1019 velocities.

1020

1021 Fig. 10. A checkerboard P-wave velocity test model (left)
1022 and its reconstruction (right): Plan views of the layer
1023 between -3.0 and -3.2 km (a, b) and N-S (c, d) and E-W (e,
1024 f) cross sections.

1025

1026 Fig. 11. Accuracy of hypocenters of 48 selected VT
1027 earthquakes. The mean and standard errors of each earthquake

1028 are shown by open circle and error bar, respectively. See
1029 text for details.

1030

1031 Fig. 12. (a) Block-and-ash pyroclastic flow deposits from an
1032 eruption on 26 November 1999 in the Rio Cristal valley as
1033 seen on 30 November 1999. (b) Microphotograph of a selected
1034 dome sample showing a basal section of amphibole (amph)
1035 without any kind of reaction rim (pl=plagioclase). (c)
1036 Amphibole composition plotted as Na+K versus Al_{tot} (apfu =
1037 atoms per formula unit) compared to experimentally produced
1038 amphibole for the Pinatubo dacite at 220, 400 and 960 MPa
1039 (Scaillet and Evans, 1999; Prouteau and Scaillet, 2003). The
1040 arrow represents the temperature effect (from 780°C to
1041 900°C) on amphibole composition at 220 MPa. Mineral
1042 composition was analyzed by electron microprobe at
1043 Laboratoire Magmas et Volcans, Université Blaise Pascal,
1044 Clermont-Ferrand, France, using a CAMECA SX-100. The
1045 operating conditions were 15 kV acceleration voltage, 10- 12
1046 nA beam current, and 10 s counting time.

1047

1048 Fig. 13. Radial component of the tiltmeter data and the
1049 daily number of LP events and VT earthquakes during the
1050 period between 22 September and 31 December 1999. Shaded
1051 regions indicate inflation periods.

1052

1053 Fig. 14. Focal depths of VT earthquakes (blue-filled
1054 diamonds) and LP events (red-filled triangles) plotted as a
1055 function of time from July 1998 to December 1999.

1056

1057 Fig. 15. A sketch of our interpretation of the magmatic
1058 system beneath Guagua Pichincha volcano superimposed on the
1059 E-W cross section of the 3-D P-wave velocities and
1060 hypocentral locations of VT earthquakes.

1061

Figure No. 1

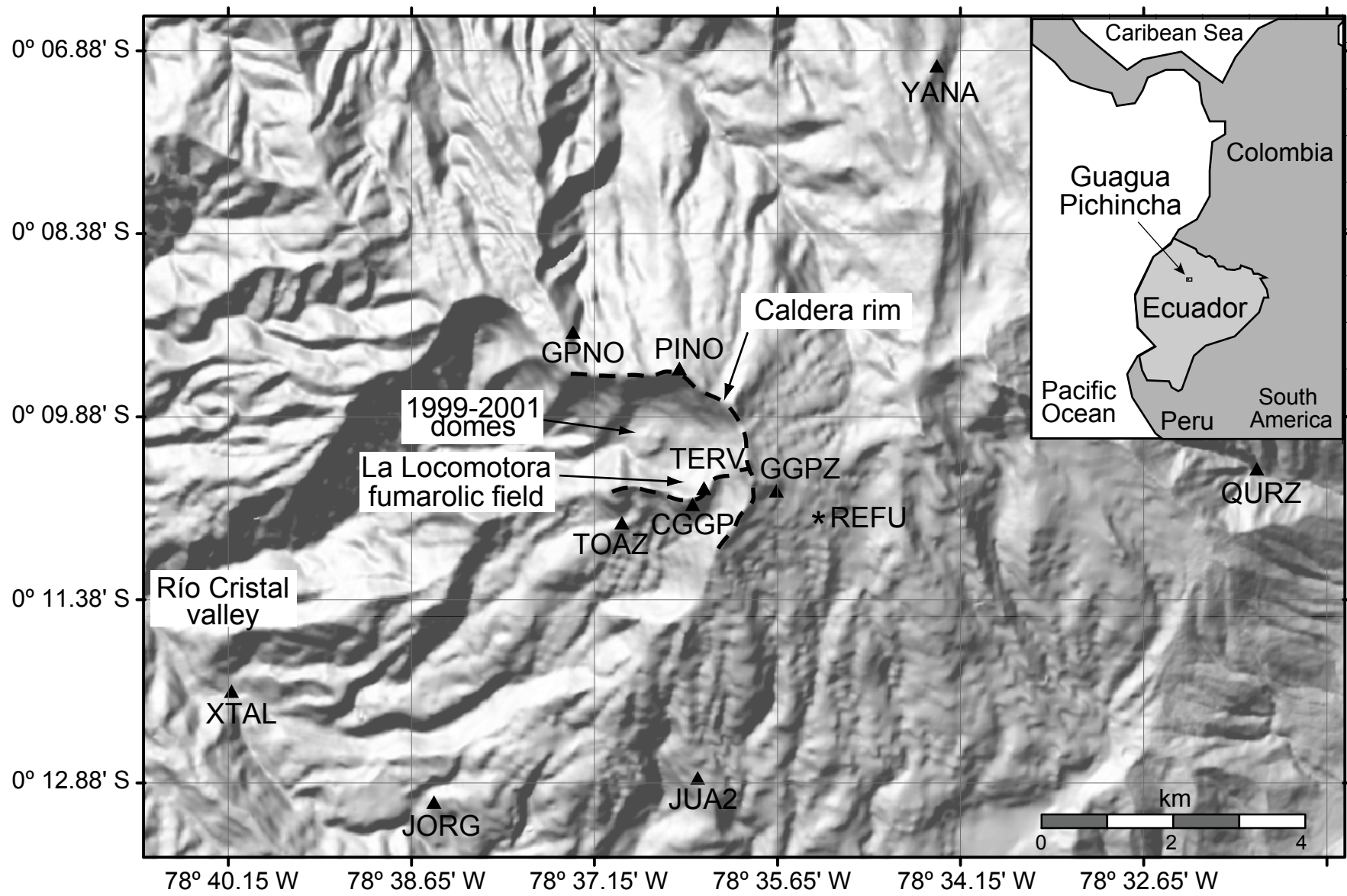


Figure 1

Figure No. 2

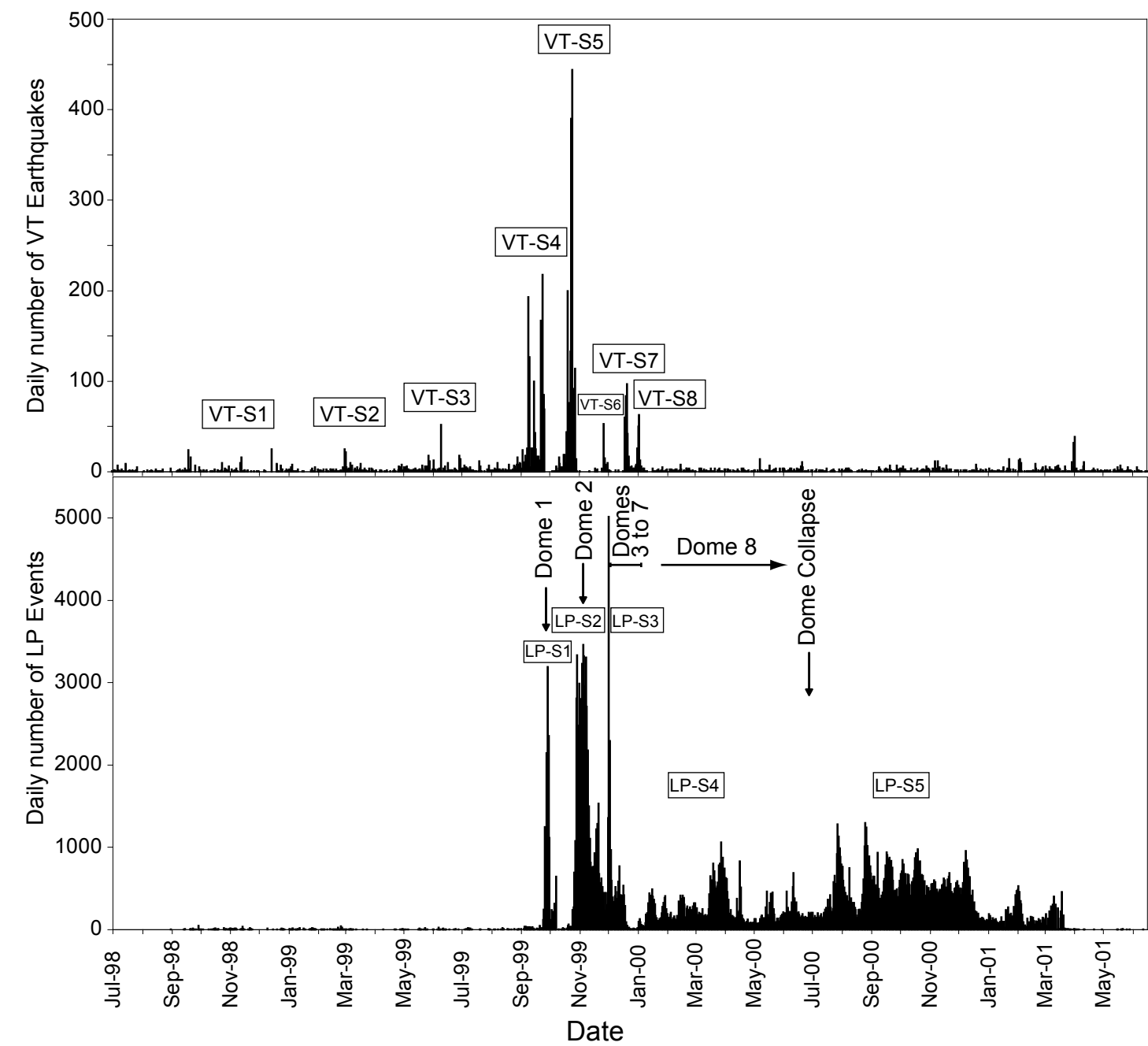


Figure 2

Figure No. 3

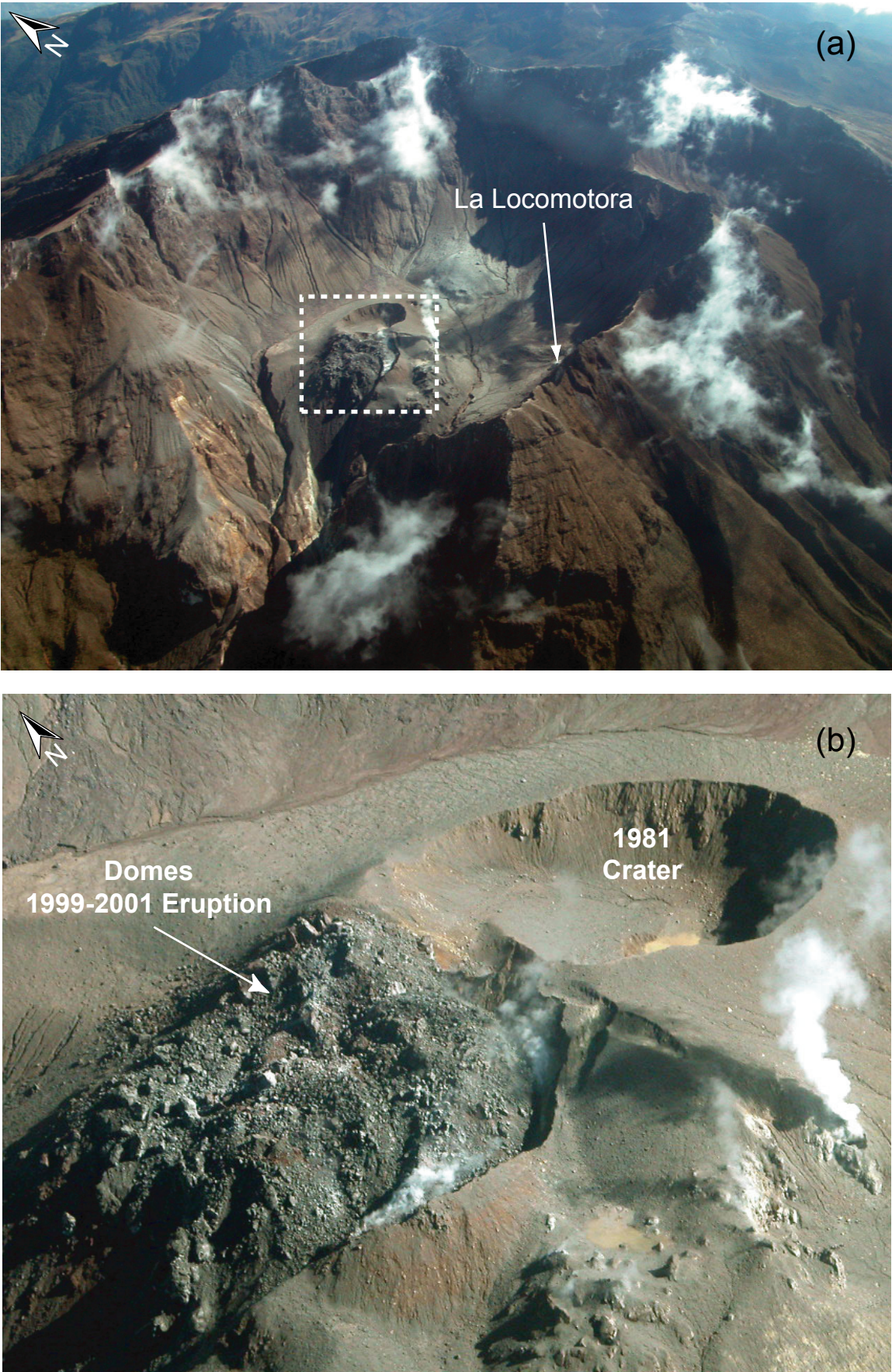


Figure 3

Figure No. 4

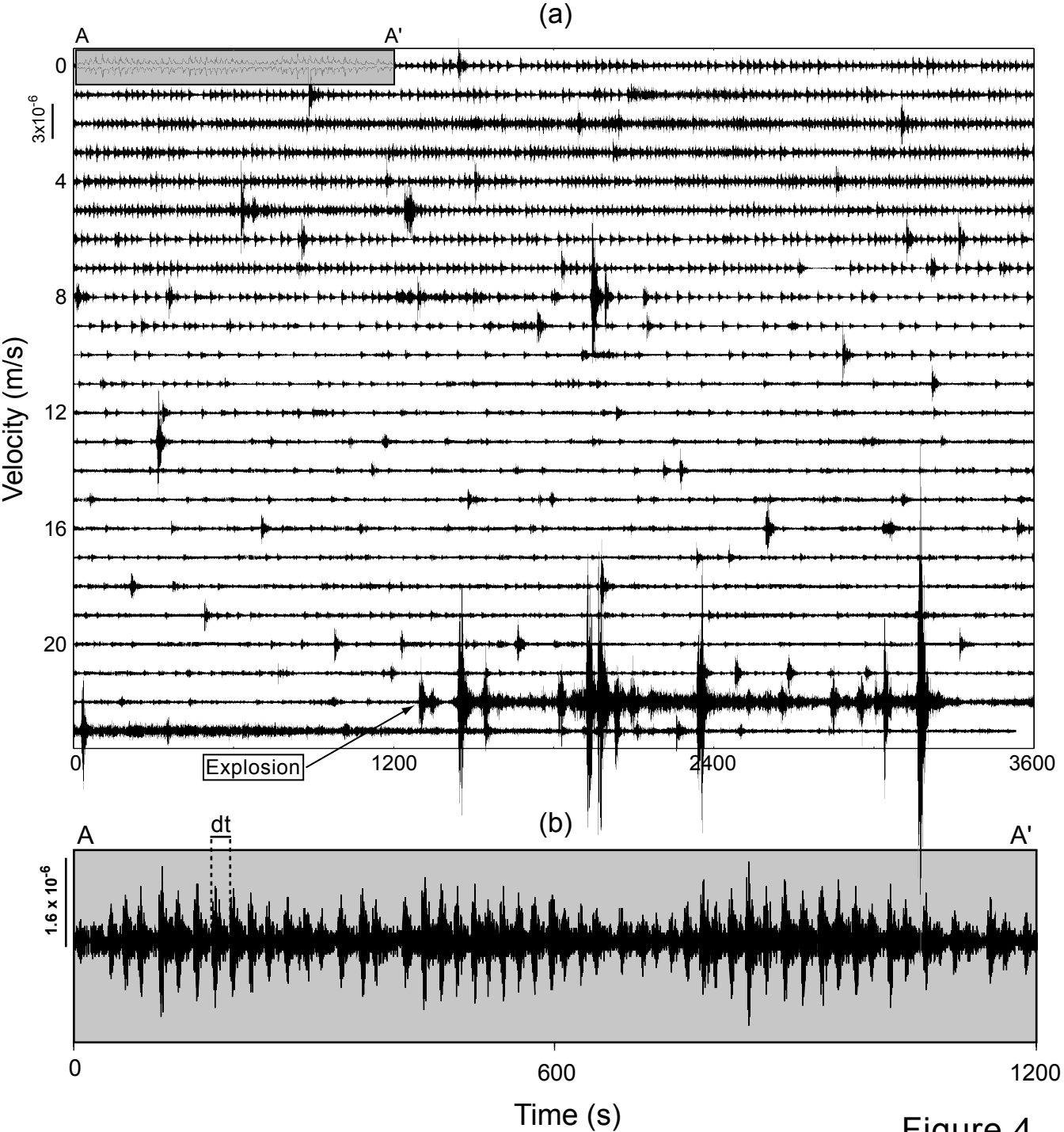


Figure 4

Figure No. 5

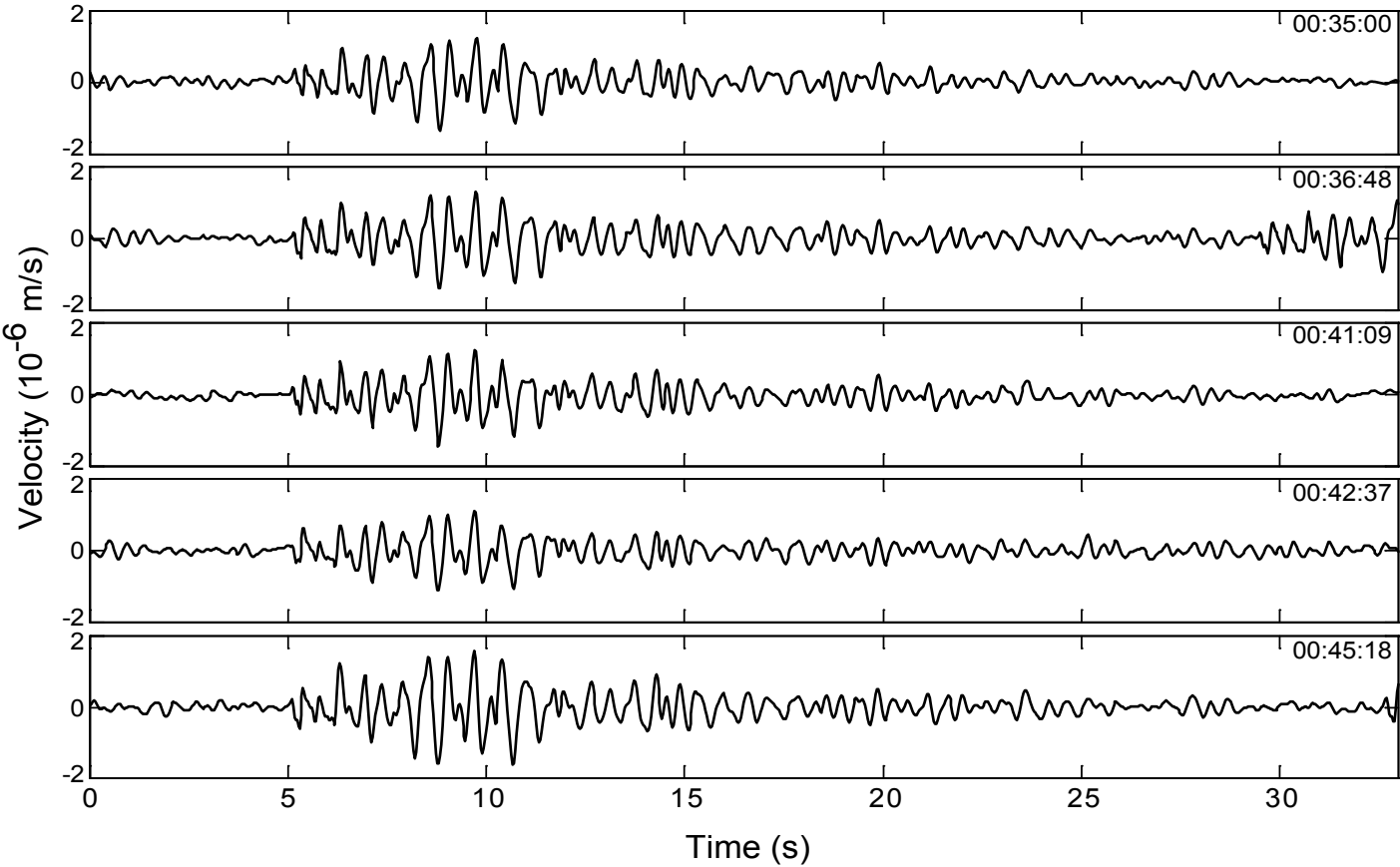


Figure 5

Figure No. 6

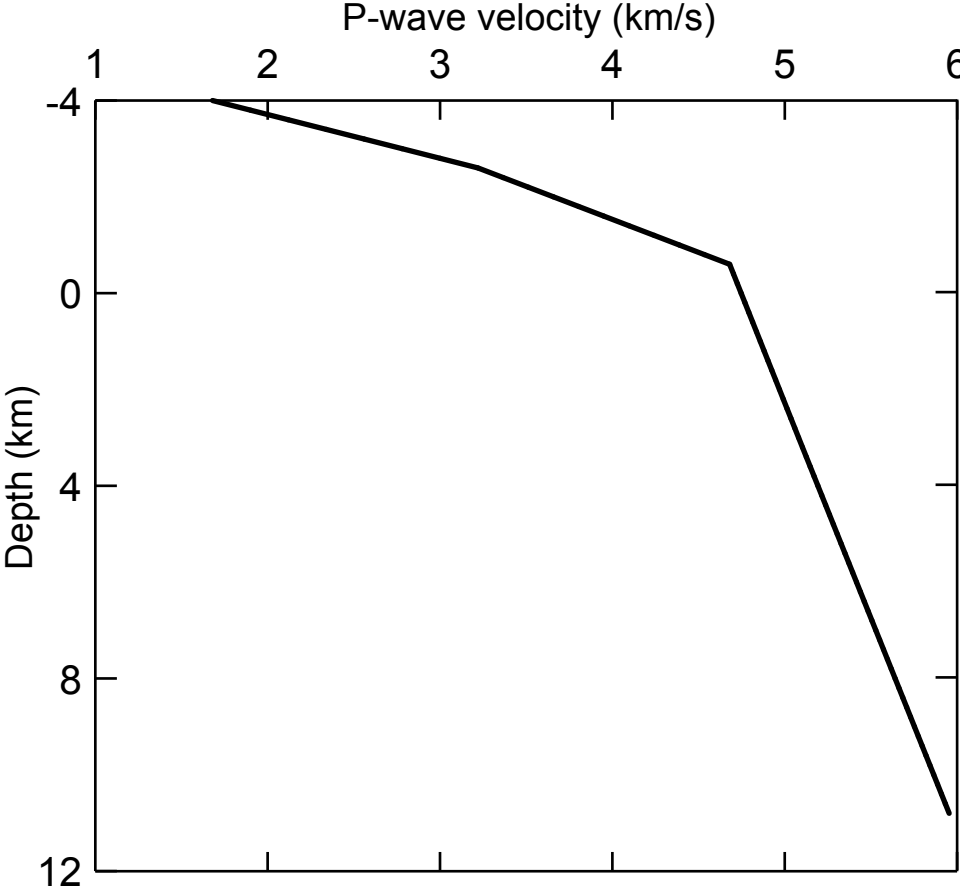


Figure 6

Figure No. 7

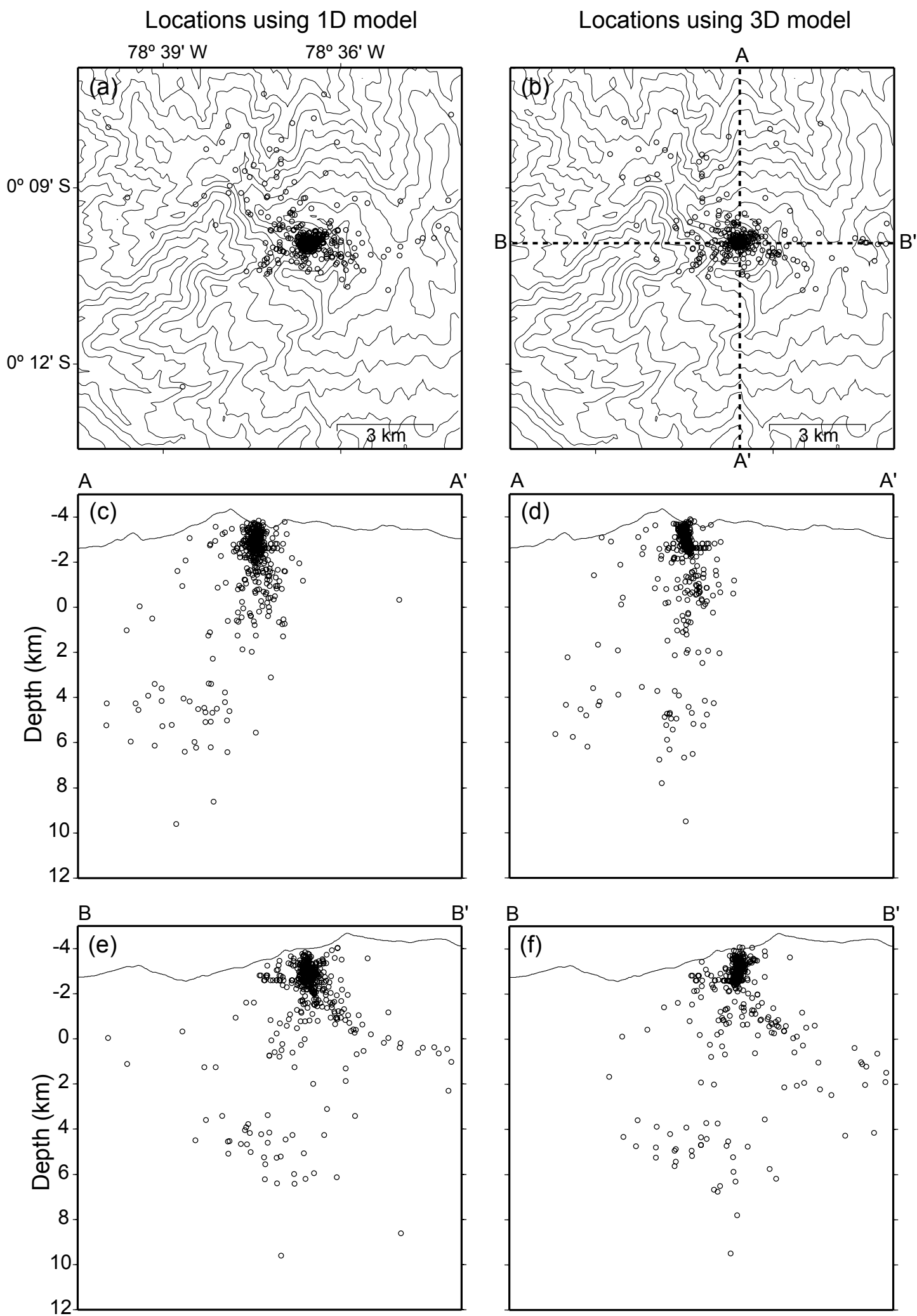


Figure 7

Figure No. 8

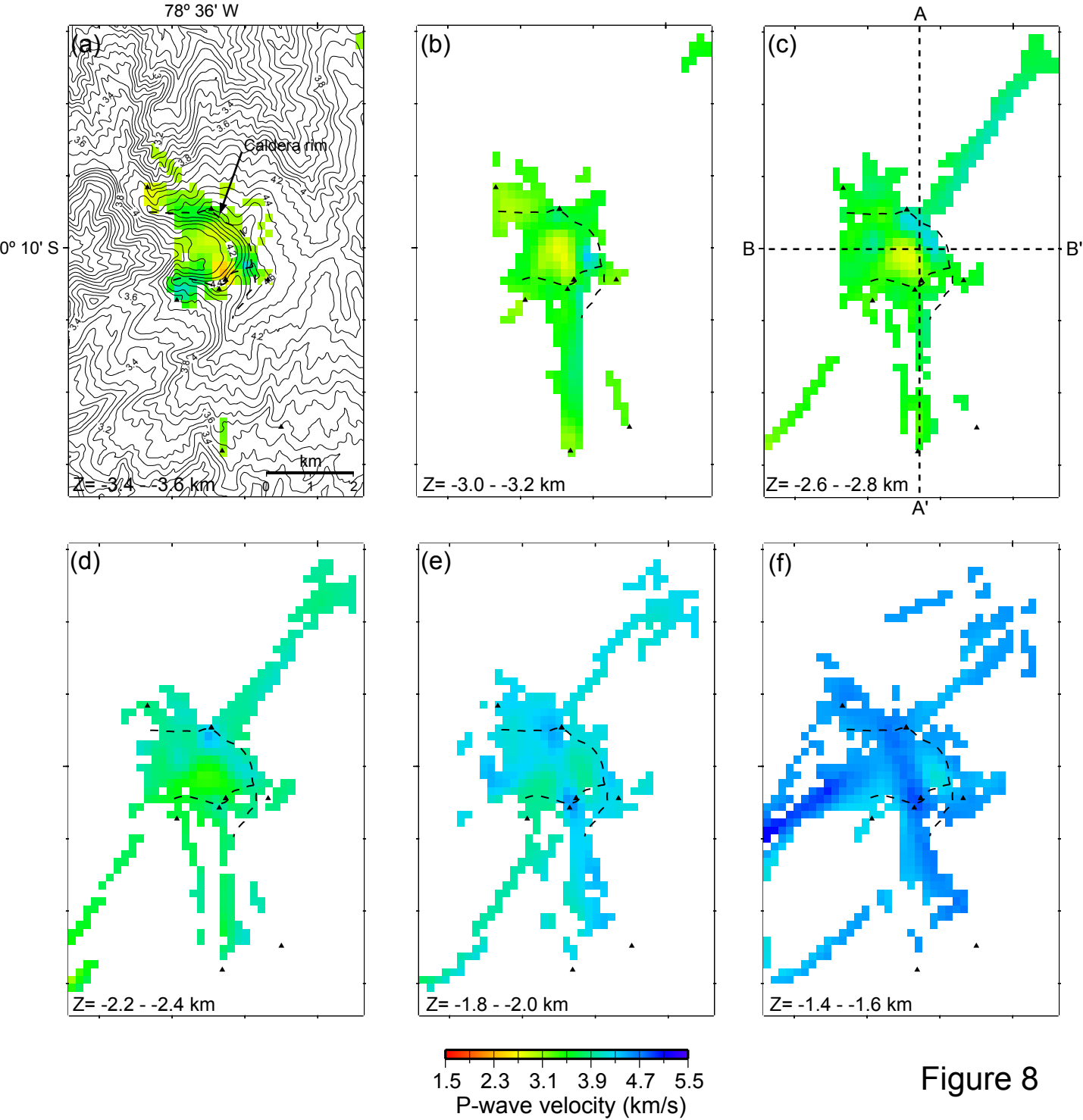


Figure 8

Figure No. 9

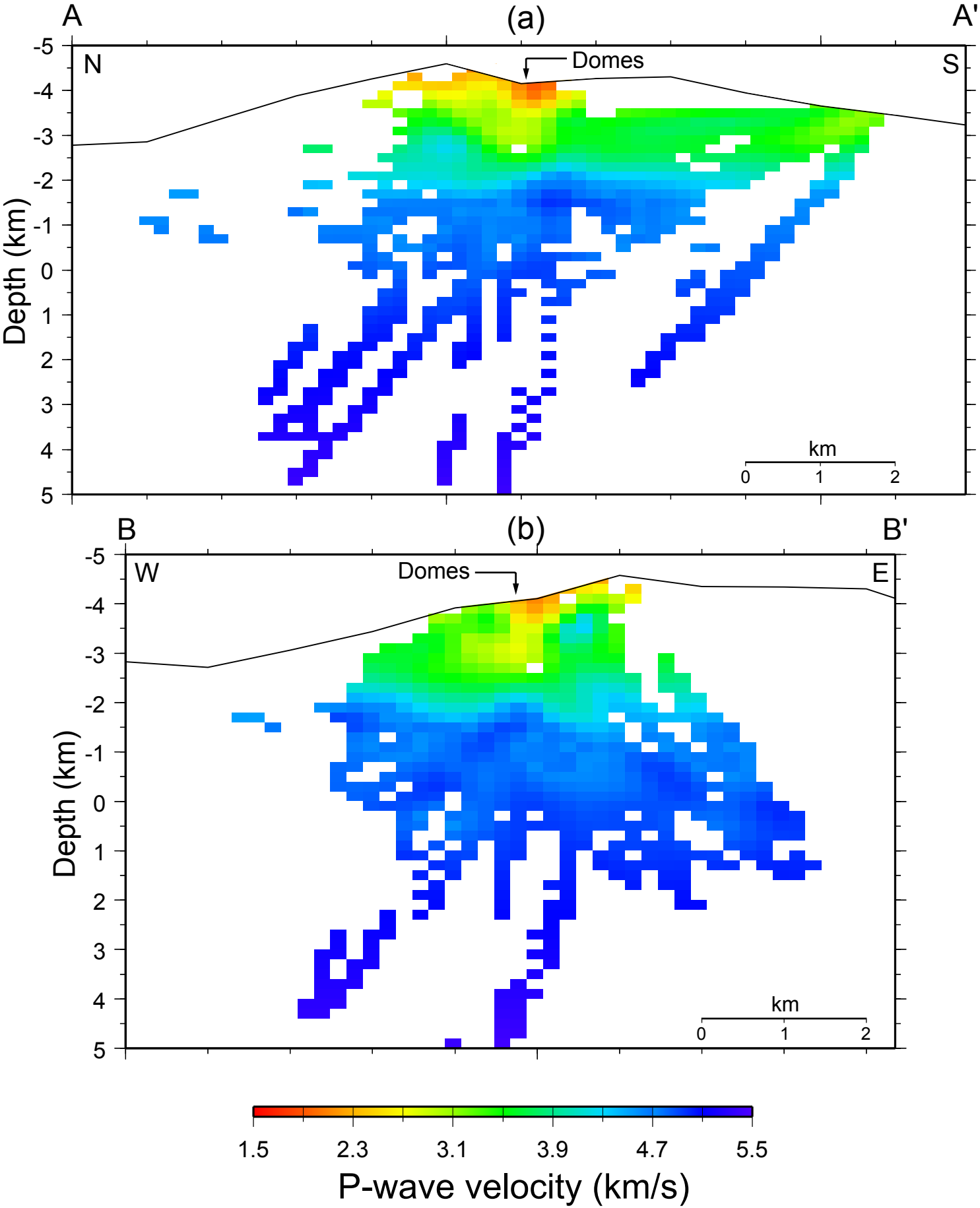


Figure 9

Figure No. 10

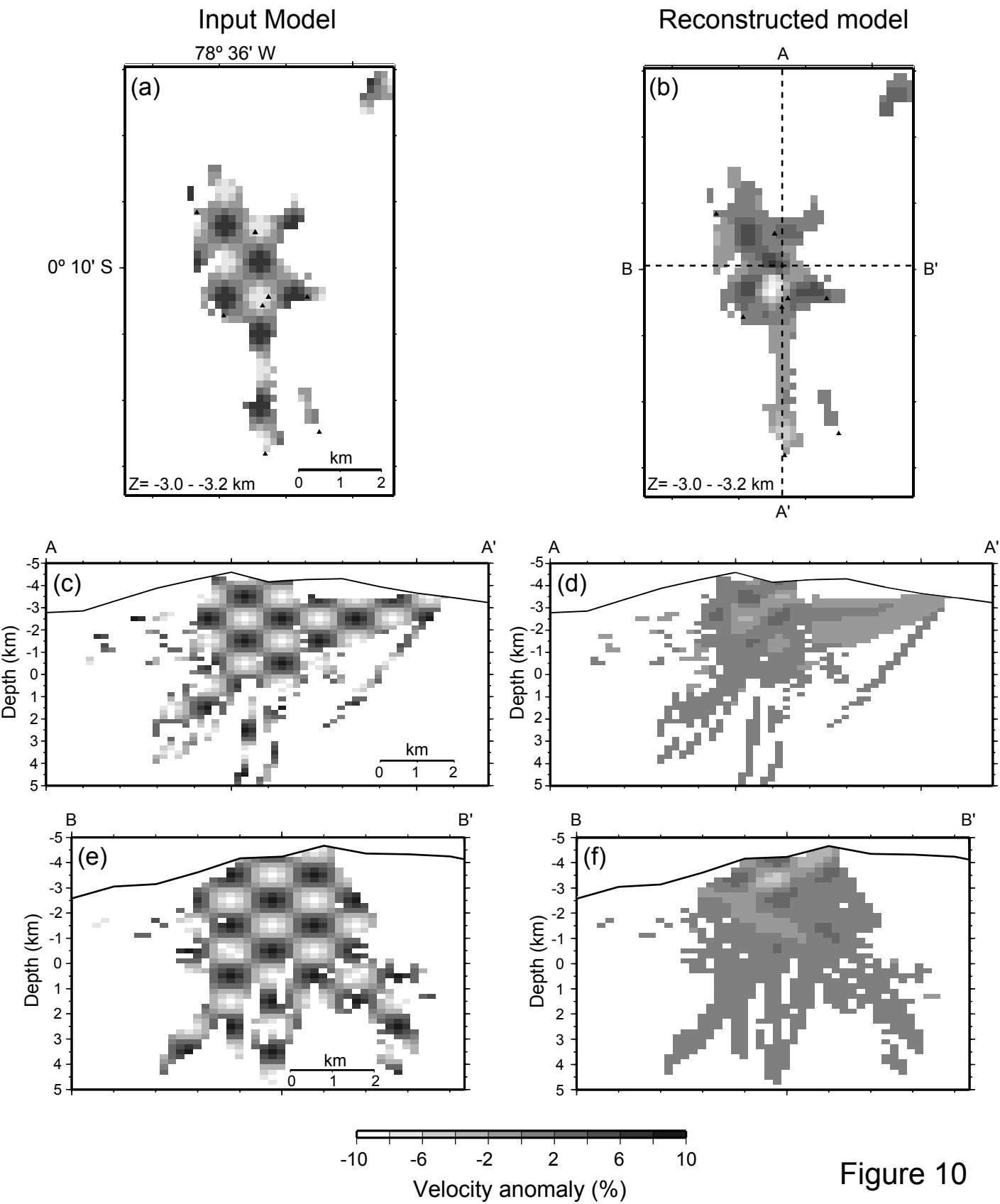


Figure 10

Figure No. 11

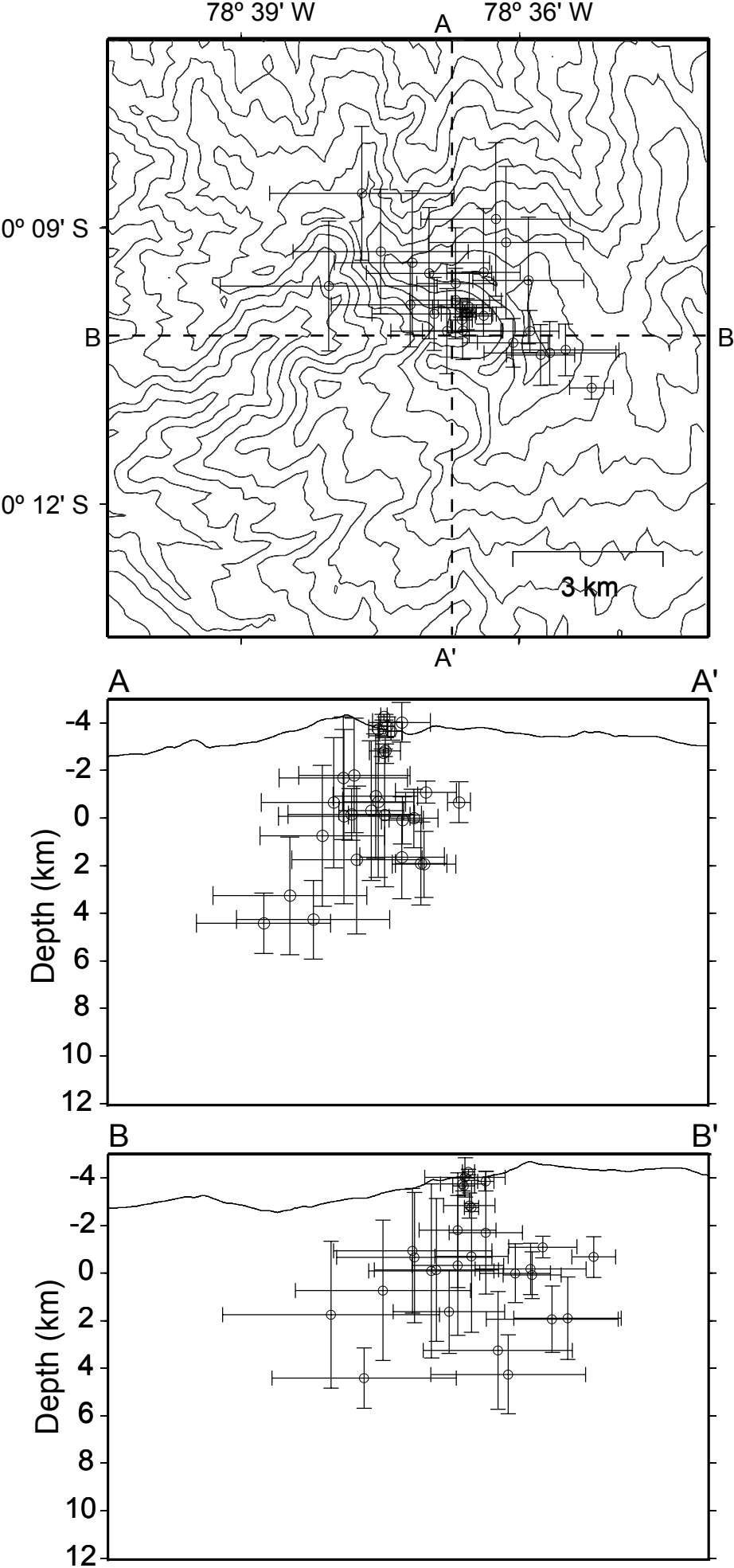


Figure 11

Figure No. 12

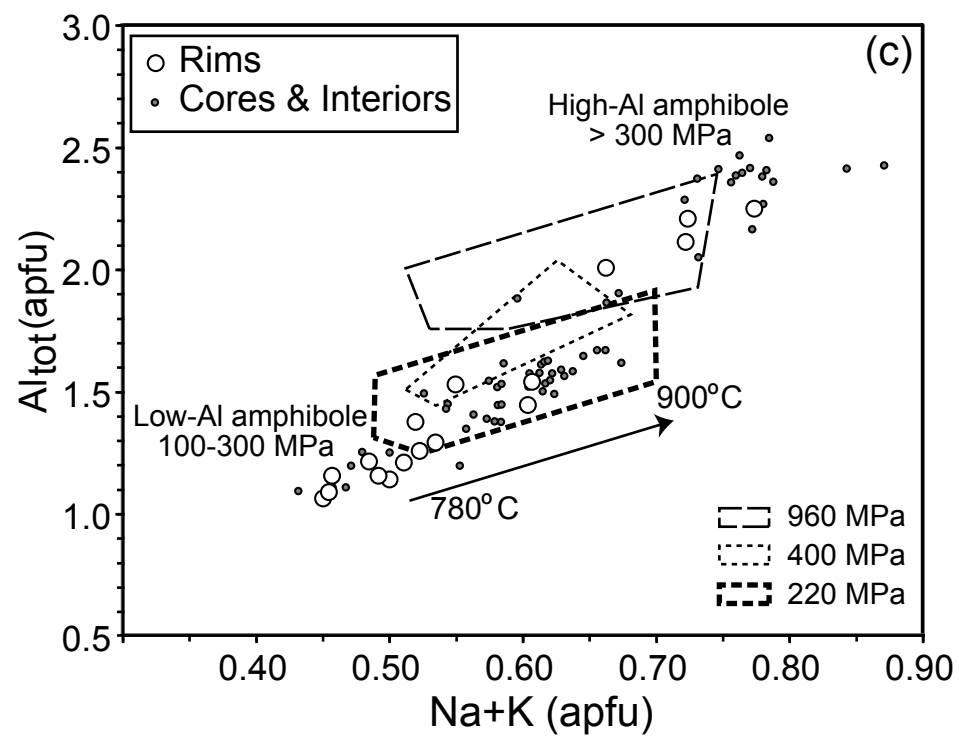
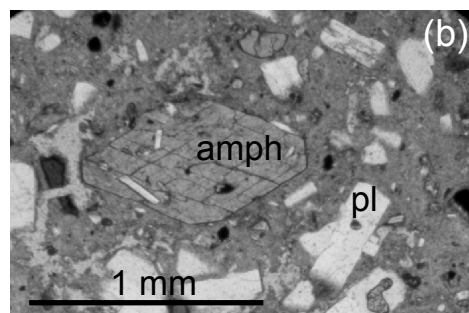


Figure 12

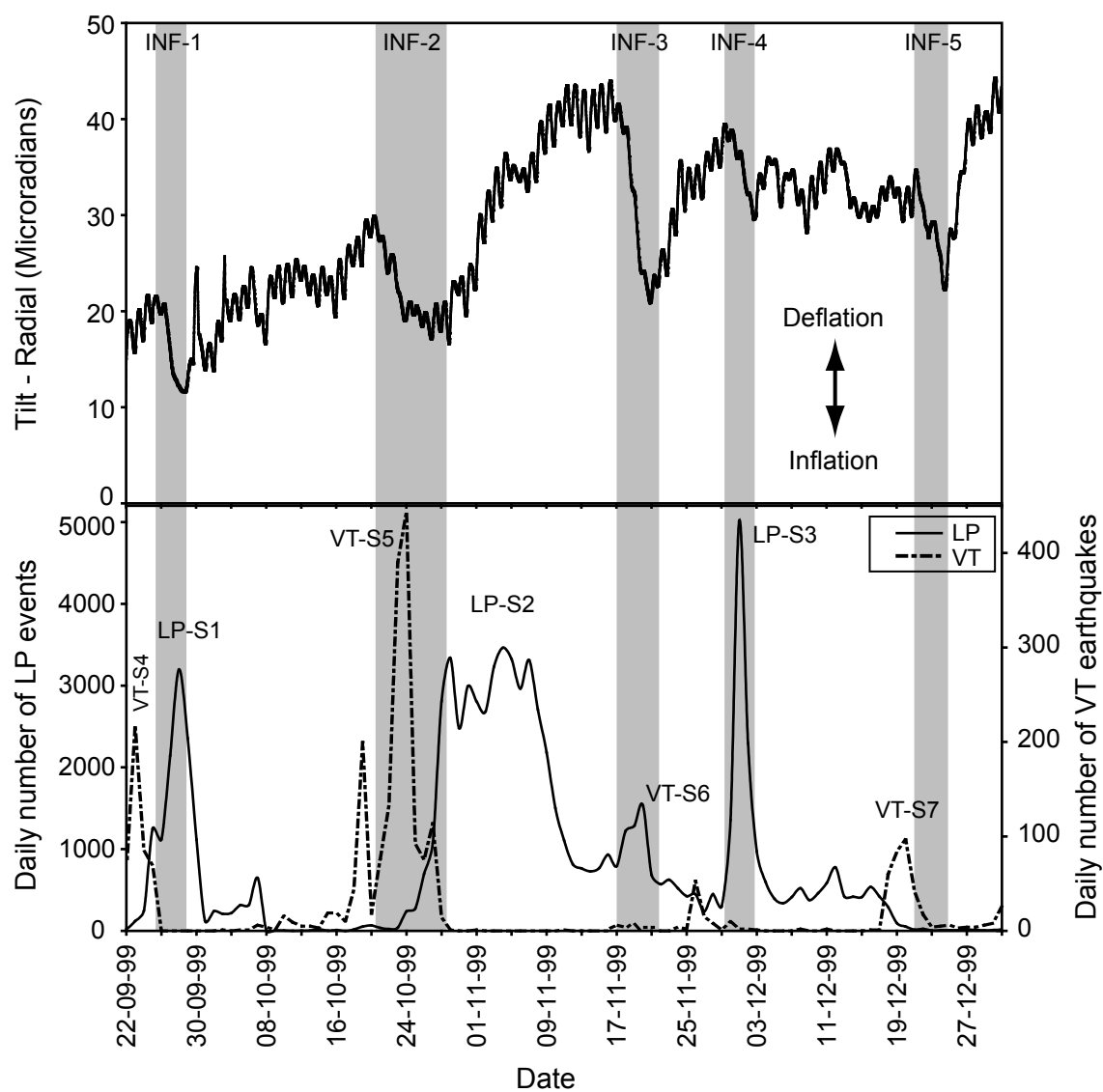


Figure 13

Figure No. 14

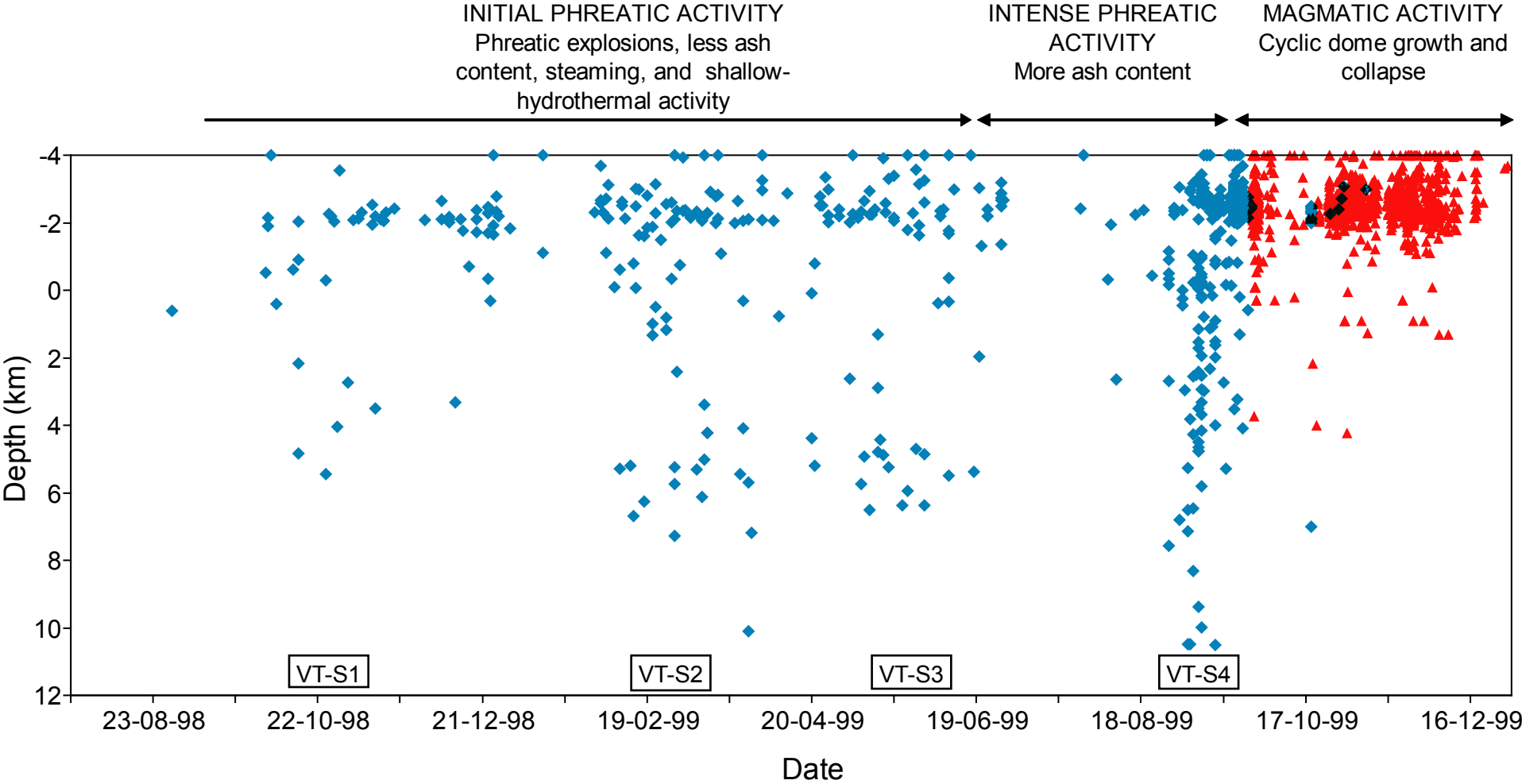


Figure 14

Figure No. 15

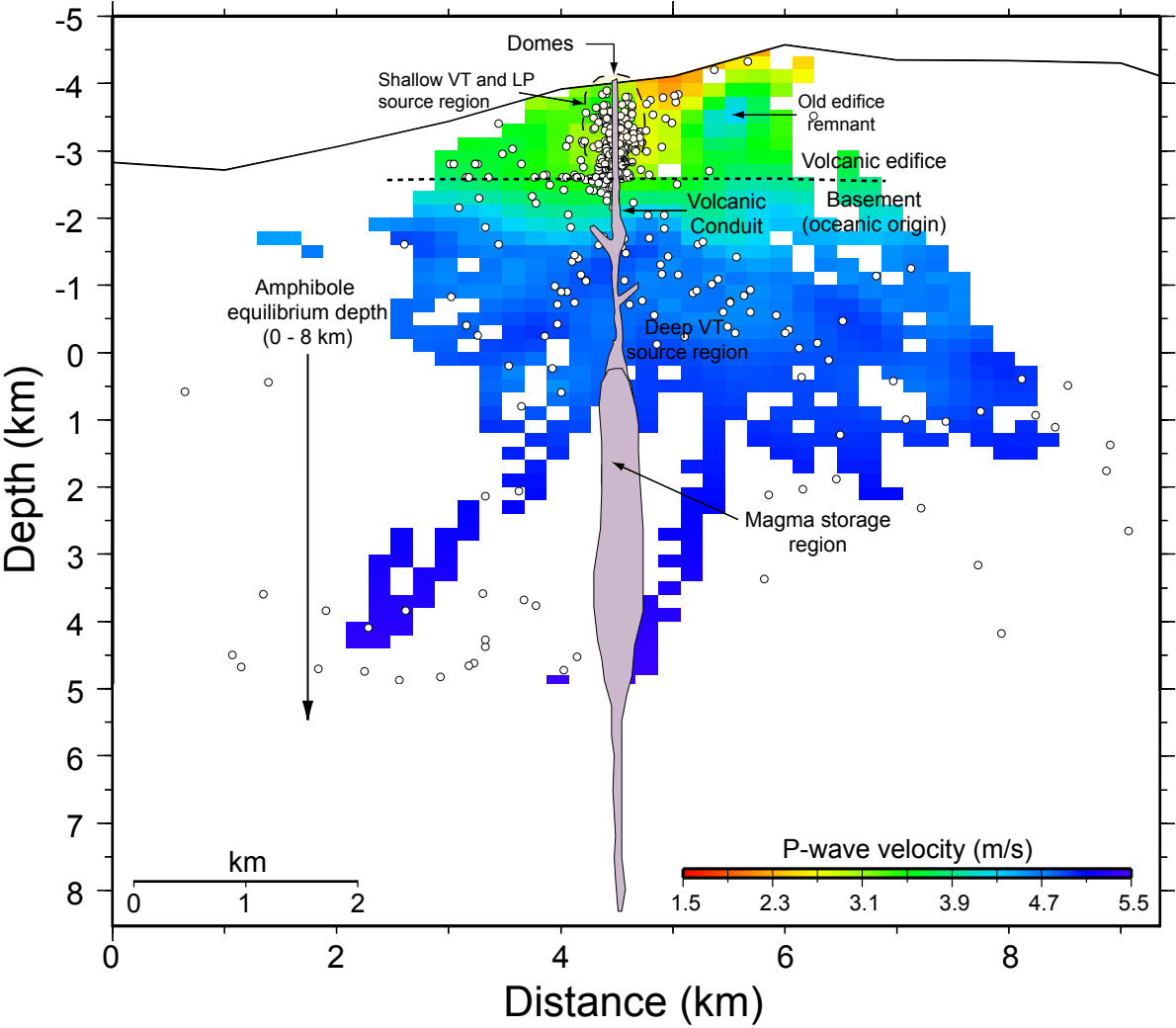


Figure 15