Earth and Planetary Science Letters 421 (2015) 27–38

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Hydrothermal and magmatic reservoirs at Lazufre volcanic area, revealed by a high-resolution seismic noise tomography

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A R T I C L E   I N F O

Article history:
Received 23 December 2014
Received in revised form 17 March 2015
Accepted 18 March 2015
Available online 11 April 2015
Editor: P. Shearer

Keywords:
Lazufre
Lastarria
uplift
ambient noise tomography
hydrothermal reservoir
magma chamber

A B S T R A C T

We determine here for the first time the geometry and location of the hydrothermal and magmatic reservoirs in the Lazufre volcanic area. This furthers the understanding of the origin of one of the largest worldwide volcanic uplift regions, both in space and amplitude. The exact locations and shapes of the sources generating a double-wide uplift region in the Lazufre found by past deformation data (InSAR and GPS) and generating hydrothermal and magmatic fluids found by geochemical gas analysis have not been well-delimited. In this study, we use seismological data to perform a 3-D high-resolution S-wave velocity model, which allows defining better the locations and shapes of the sources of the deformations and the hydrothermal and magmatic reservoirs. We find three anomalies. Two of them (with S-wave velocity of about 1.2–1.8 km/s) are located below the Lastarria volcano. The shallow one (<1 km below the volcano base) has a funnel-like shape. The deeper one is located between a depth of 3 and 6 km below the volcano base. Both are strongly elliptical in an EW direction and separated by a 2–3 km thick zone with Vs of ∼1.5–2 km/s. As far as these anomalies are located under the hydrothermal activity of Lastarria volcano, they are interpreted as a double hydrothermal (the shallow part) and magmatic source (the deeper part). The latter can feed the former. This double hydrothermal and magmatic source is in agreement with previous geochemical, deformation (GPS and InSAR) and magneto-telluric studies. In particular, it explains the double origin of the gases (hydrothermal and magmatic). The third low-velocity zone (with S-wave velocity of about 2.3 km/s) located at 5 km depth and deeper is centered beneath an area of surface uplift as determined by InSAR data. We compare the seismic tomography and InSAR results to propose that this low-velocity zone is at the top of a large reservoir, hosting hydrothermal fluids and possibly also magma.

1. Introduction

Knowledge of volcanic-reservoir geometry, location and strength is of major importance in monitoring and understanding volcanic unrest. Volcanic reservoirs include magmatic and hydrothermal storage zones and may feed volcanic eruptions when molten material and/or gases reach the surface or a secondary magma reservoir. The depth of a magma reservoir is controlled by a complex association of factors such as the regional stress regime, the magma density, viscosity, volatile content, crystal content or the local crustal structure. In extensional, transtensional, transpressional or strike-slip contexts, the magma reservoirs are generally shallow (between the sub-surface and about 5 km depth) whereas in compressional settings, the magma reservoirs are found to be deeper, without shallow magma reservoirs (e.g., Pritchard and Simons, 2004; Chaussard and Amelung, 2012). These factors also influence the behavior of a volcanic eruption since they control the pressure–temperature condition of the magma reservoir (Dzurisin, 2006; Chaussard and Amelung, 2012). Hence it is important to know if andesitic volcanoes can have shallow reservoirs and we take the Lastarria volcano as a study example.

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http://dx.doi.org/10.1016/j.epsl.2015.03.042
0012-821X/

Direct and convincing seismic images of the geometry and properties of volcano reservoirs are rare (Marsh, 2000). This is due to the fact that traditional seismic tomography, based on earthquake data, is not adequate for obtaining high resolution images of the shallow crust where small magma chambers, dykes or sills are expected (e.g., Lees, 2007). It is particularly true when the number of earthquakes used is small or when the seismograms are difficult to read due to strong path/site effects.

The problem of having a fewer earthquakes can be easily solved with another technique: the ambient seismic-noise tomography (ANT). The ANT technique can produce high-resolution images of the upper crust (e.g., Shapiro et al., 2005), without earthquakes. The technique allows the retrieval of Green's function between pairs of seismometers by cross-correlating the ambient noise recorded at each of them (Shapiro and Campillo, 2004). ANT techniques have successfully been applied to reveal different geological structures at global, regional and local scales using only a few hours to a few months of continuous seismic noise (e.g., Shapiro et al., 2005; Brenguier et al., 2007; Bensen et al., 2007; Mordret et al., 2013). This technique has also been successfully applied to obtain images of volcanic structures, being especially promising for imaging volcano reservoirs at unprecedented resolution, for instance at: the Piton de la Fournaise volcano (Reunion Island; Brenguier et al., 2007; Mordret et al., 2014), the Okmok volcano (Alaska, U.S.A; Masterlark et al., 2010), the Toba volcano (Sumatra, Indonesia; e.g., Jaxybulatov et al., 2014), the Uturuncu volcano (Bolivia; Jay et al., 2012), Mount Asama (Japan; Nagaoka et al., 2012) and at the Colima Volcano (Mexico; Spica et al., 2014).

The Lazufre (an acronym for Lastarria and Cordón del Azufre) area (Pritchard and Simons, 2002) is one of the largest uplift deformation areas in the world (Ruch et al., 2008), located in the Altiplano–Puna Plateau in the central Andes (Chile–Argentina). An area of $\sim$2000 km^2 started inflating between 1997 and 2000 (Pritchard and Simons, 2002; Froger et al., 2007; Ruch et al., 2009) related to an over-pressurized source at depth. Basaltic volcanoes generally show such uplifts before eruptions (e.g., Wicks et al., 2002; Lu et al., 2010), but it is not clear it is the case for andesitic volcanoes (Pritchard and Simons, 2004; Fournier et al., 2010; Chaussard and Amelung, 2012). A few questions are still unsolved. Geodetic data showed the existence of two sources of deformation (a shallow one and a deep one), but with inaccurate shape and location. The depth of the deep source is not well constrained since it is modeled somewhere between 7 and 18 km (Pritchard and Simons, 2004; Froger et al., 2007; Ruch et al., 2008; Anderssohn et al., 2009; Henderson and Pritchard, 2013; Pearse and Lundgren, 2013) and re-estimated to be between 2 and 14 km by Remy et al. (2014). Hence, the depth of this source may be between 2 and 18 km. These errors are due to the trade-off between the pressure, the shape of the source and the depth. The shallower source was supposed to be unique, at a depth of about 1 km and located just beneath the Lastarria volcano (Froger et al., 2007; Ruch et al., 2009). We show in this study that this source double. Furthermore, these geodetic data cannot discriminate between a hydrothermal and a magmatic system below the Lastarria volcano, as is suggested by geochemical studies.

In this study, we perform a high-resolution 3-D ANT, using data from 26 mainly broadband seismic stations recorded at two different seismic networks deployed at Lazufre. The location and geometry of hydrothermal and magmatic reservoirs below the Lazufre volcanic area are deduced through S-wave velocity tomographic images obtained from the ANT. Results are compared to source inversions from InSAR and GPS data (Pritchard and Simons, 2002, 2004; Froger et al., 2007; Ruch et al., 2008, 2009; Anderssohn et al., 2009; Pearse and Lungren 2013; Remy et al., 2014). As the geometry and the depth of the sources of these deformations cannot be well and uniquely determined with only the InSAR and GPS deformation field, it is important to image these sources with independent data, such as seismicity.

2. Data and methods

2.1. Data

The data used in this study come from two temporary seismic networks installed during two distinct time periods. Network 1 (red triangles in Fig. 1) was deployed from 1 February to 26 March 2008 by the GFZ (Germany) and the DGF (Chile). It was composed of 18 seismometers: 17 broadband (12 Guralp CMG3ESP -60 s- and five Trillium T40 -40 s-) and one short period (LE-3D -1 s-), covering the main deformation zone at Lazufre. Network 2 (blue triangles in Fig. 1) was deployed from November 2011 to March 2013 by the University of Alaska Fairbanks in the framework of the PLUTONS project, but we used only the data from January to March 2012 for this network. Network 2 was composed of eight broadband seismometers (six CMG3T -120 s- and two CMG6TD -30 s-). All seismometers (Fig. 1) were GPS-time synchronized.

The next paragraphs describe the procedures of how the Rayleigh-wave group velocities were obtained from continuous seismic noise records in order to perform an S-wave tomography later.

2.2. Reconstruction of Green's functions from seismic noise

The following steps were applied on the vertical components of each individual continuous seismic data: (1) a removal of the mean and the trend of the signal; (2) a down-sampling to 20 Hz; (3) an instrumental response correction; (4) a 1–30 s band-pass filter; (5) a temporal (1-bit) normalization; and (6) a spectral (whitening) normalization. Steps (5) and (6) were applied in order to diminish the influence of earthquakes and/or non-stationary noise sources at the vicinity of the seismometer. These normalizations allowed using a larger frequency band (e.g., Bensen et al., 2007) and diminishing the influence of heterogeneous distribution of noise sources. Then, cross-correlation functions (CCFs) were calculated for the vertical components of all concomitant station pairs (181) on 200 s time windows. The duration of 200 s was chosen because we found that if a particular station would have a punctual instrumental failure, only a 200 s time window of signal would have been lost. We then stacked all available 200 s CCFs for a given station-pair and added the positive and negative parts of the CCF to enhance the signal-to-noise ratio and to reduce the effect of the heterogeneous distribution of the sources (e.g., Sabra et al., 2005; Bensen et al., 2007). Each folded and stacked CCF converged towards the estimated Green's function (EGF) between each pair of seismometers. Only the EGFs with a signal-to-noise ratio greater than 8 (value that has been found to give the best associated dispersive curves) were used for further analysis. The signal-to-noise ratio was calculated as the ratio of the maximum amplitude of the signal over the root-mean square of the noisy part of the EGF. Since only the vertical components of the ground motion were used in this study, the EGFs are dominated by Rayleigh surface waves. The Love waves were harder to extract from the seismic noise analysis than the Rayleigh waves at Lastarria volcano, so we performed the following analysis using only the Rayleigh waves. Fig. 2 depicts some examples of EGFs with respect to station LGG01 for different azimuths corresponding to different stations.

2.3. Group velocity measurements

The Rayleigh-wave fundamental mode dispersion curves were determined from each EGF via a frequency–time analysis (FTAN;