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Estimation of the near-surface velocity structure of the Yasur-Yenkahe volcanic complex, Vanuatu

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ABSTRACT

Small-aperture array measurements of seismic noise at seven sites around Yasur volcano (Vanuatu) are performed to estimate the V_P and V_S velocities of the shallow structure. The spatial autocorrelation (SPAC) and the frequency–wavenumber (f–k) methods are used to determine Rayleigh phase velocity dispersion curves. Phase velocities computed with the SPAC method vary between 580 m/s and 960 m/s at 1 Hz and between 270 m/s and 420 m/s at 15 Hz. F–k dispersion curves show velocities of 300–340 m/s and 800–940 m/s at 1 Hz and 200–230 m/s at 15 Hz. An inversion technique based on the use of the neighbourhood algorithm has been used to calculate the one-dimensional velocity model at each site. Velocity models reach 200 m deep and mainly contain two layers and a half-space. For sites close to the Siwi caldera rims, comparisons with geology and hydrothermal system studies suggest that the two layers highlighted in models may correspond to two large pyroclastic sequences related to caldera collapses based on the flank of an old volcano. Results obtained for the other three sites, located inside the caldera, show the influence of the hydrothermal system on P- and S-wave velocities. For these sites, fluid circulation inside the volcanic deposits causes lower velocities at depth. To obtain a near-surface velocity model of the volcanic structure, each 1D velocity model is spatially extrapolated according to the surface geology. Results highlight four distinct areas, the Siwi caldera edges with high velocities and the resurgent block, the ash plain and the Yasur edifice with lower velocities at depth.

1. Introduction

Yasur is a small scoria cone located on the south-western part of Tanna Island, in the south of Vanuatu archipelago. The edifice belongs to an important volcanic system consisting of a large caldera (5×8 km) with a resurgent block growing in its centre. The current morphology of this system was shaped through successive phases of eruptive activity with variable intensities. The Pliocene to Pleistocene period was marked by several important phases of volcanic activity in the northern part of Tanna Island (Green Hill centre) then in the southern part (Tukosmeru centre), generating significant pyroclastic deposits (Carney and MacFarlane, 1979). The last known major eruptive event was the collapse of the Siwi caldera during which approximately 1 to 2 km^3 of material was ejected, forming the Siwi ignimbrites (Carney and MacFarlane, 1979; Robin et al., Mar., 1994). After the collapse, the floor of the caldera began to uplift, giving rise to the present Yenkahe resurgent block. The uplift rate of the Yenkahe horst was determined using the emerged reef terraces by Chen et al. (September, 1995). They estimated a rate of ~156 mm/yr using $^{230}Th/^{234}U$ dating. The large difference between the regional rate estimated from samples coming from the north–west coastof Tanna of ~1 mm/yr and the rate obtained for the Yenkahe horst highlights an anomalous uplift below the caldera centre (Chen et al., September, 1995). This ground deformation was related to the emplacement of a magma reservoir below the Yenkahe resurgent block, at shallow depths (Nairn et al., 1988; Chen et al., September, 1995; Métrich et al., 2011). Since ~1400 yr B.P., the eruptive activity has been focused at Yasur volcano (Métrich et al., 2011). It is characterised by Strombolian to Vulcanian explosions, with a frequency of one to three explosions per minute during very active phases of activity and one explosion every few minutes during quiet phases, ejecting pyroclastic materials as bombs, scoriae and ash plumes in a radius mainly limited to the crater and sometimes to the flanks. However the presence of pyroclastic deposits linked to the Yasur activity north of the caldera indicates that the volcano has known at least one important eruptive phase, identified as a sub-Plinian eruption (Nairn et al., 1988).

The various types of eruptive activity have led to a complex volcanic structure with discontinuities of various origins: lithology, tectonics and fluid saturation. The lithological discontinuities are mainly related to the eruptive type, i.e. small, moderate or strong explosive or effusive eruptions, giving rise to various pyroclastic materials (bombs, scoriae, ashes and lava flows). The tectonic discontinuities are related to the caldera collapse (ring faults) and to the Yenkahe uplift.

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Knowledge of the near-surface volcanic structure is crucial to understand the geological, geochemical and geophysical processes that occurred inside the edifice. It is mostly based on the study of the geology and hydrothermal systems allowing us to distinguish major faults, fracturing areas and discontinuities that may cause instability zones (Chevallier and Bachèlery, 1981; Finizola et al., 2006; Merle et al., 2010). It also goes through the determination of the seismic velocity model which allows the improvement of seismic signal locations by avoiding the errors on their positions and geometries (Bean et al., 2008).

In this paper, we study the near-surface volcanic structure from seismic data to highlight major discontinuities and velocity anomalies. Various techniques have been proposed on volcanoes to determine velocity structures. The most used one is seismic tomography with the inversion of P- and S-wave arrival time data from teleseismic, regional or local earthquakes (Lees, 2007). The velocity model is built by contouring areas with similar properties, i.e. low or high velocity anomalies, determined from the differences between the observed travel-times and the predicted travel-times computed using a velocity model. The anomalies are generally on the order of one to several kilometres and can be detected a few tens of kilometres deep (De Luca et al., 1997; Patanè et al., 2006; Lees, 2007; Prõno et al., 2009). P-wave velocity model can also be determined using first arrival time data from an active seismic survey (Zollo et al., 1998; De Matteis et al., 2000; Onizawa et al., 2007; Aoki et al., 2009). Battaglia et al. (2008a) propose combining earthquake data sets and recordings from shots to realise a precise P- and S-wave tomography at Campi Flegrei caldera. For these techniques, layers or anomalies with distinct velocities can be highlighted to a few kilometres depth (<5 km) and have a size on the order of a few hundreds of metres to one kilometre. Other techniques, based on the inversion of the dispersion curves, are also employed. De Barros et al. (2008) extracted the fundamental mode of the Rayleigh wave from regional and teleseismic earthquakes recorded by a large-aperture broadband network to deduce the dispersive curves. The inversion of these curves allowed them to highlight the average crustal depth below and around Popocatépetl volcano (Mexico) and to establish a velocity structure down to 50 km depth. Although these techniques are widely used to compute volcano velocity models, they are not appropriate for studying near-surface structures because of the lack of resolution for the detection of thin objects and layers. Techniques based on cross correlations using small-aperture arrays are increasingly employed on volcanoes to compute dispersion curves. These methods provide an estimation of surface wave phase velocities and thus allow us to calculate 1D velocity models of the structure for the first hundreds of metres below the surface according to the array's aperture.

The spatial autocorrelation (SPAC) method, developed by Aki (1957), is the method most applied on volcanoes. It was first used on Kilauea volcano (Hawaii) by Ferrazzini et al. (1991). They computed the 1D velocity model down to 200 m depth by using data recorded by a 120-metre-radius circular array. Using the same approach, Métaxian et al. (1997) determined the 1D compressional and shear wave velocity models down to 900 m at Masaya volcano (Nicaragua) using circular arrays with 60-metre and 120-metre-radius. Later, Chouet et al. (1998) used two circular arrays with 60-metre and 140-metre-radius to estimate the 1D P- and S-wave velocity models to 300 m at Stromboli volcano (Italy). More recently, Saccorotti et al. (2003) and Mora et al. (2006) estimated the velocity models of Kilauea volcano and Arenal volcano (Costa Rica) to 1 km and 450 m using the SPAC method.

The frequency–wavenumber (f–k) method (Capon, 1969; Lacoss et al., 1969), generally used to separate the different types of waves which compose the wave field, can also be used to calculate the dispersion curve and thus to determine the 1D velocity structure. It has been employed by Petrosino et al. (2009) to compute the shear-wave velocity model of Solfatara volcano (Campi Flegrei, Italy). These studies led to 1D velocity models that allow the building of a homogeneous structure with constant layer thicknesses.

To obtain a P- and S-wave velocity model of Yasur and its surroundings, we propose to use the SPAC and the f–k methods on a data set recorded by several small-aperture arrays close to the volcano. For each seismic array, the dispersion curves are computed independently, and then jointly inverted to obtain one-dimensional P- and S-wave velocity profiles. Afterwards, each profile is spatially extrapolated on an area defined or partly defined by the information provided by the surface geology.

This paper starts with a brief description of the surface geology of the active volcanic system and a presentation of the SPAC and f–k methods. We follow with a description of the seismic arrays and data set used for the calculation of the 1D velocity models, which have been employed to build the structure of Yasur volcano and Siwi caldera. Finally, we discuss the velocity ranges obtained and propose an interpretation for the main discontinuities highlighted in this study.

2. Surface geology

The geological map of the south-eastern part of Tanna Island (Fig. 1; Carney and MacFarlane, 1979), shows three main pyroclastic formations, with basaltic andesite to andesite compositions, dating from Pliocene (5.33 Ma) to Holocene (<0.01 Ma). The two oldest, the Green Hill Group and the Tukosmeru Group, consist of ash- and scoria—flow deposits associated with tephra fallouts covered by indurated pumice flows, highlighting phreatomagmatic features (Robin et al., Mar., 1994). The younger volcanic sequence, the Siwi Group, comprises lavas, scoria cones, lapilli tuff and the Siwi Serie (Allen, 2005). The Siwi pyroclastic Series, emplaced during the caldera collapse (Carney and MacFarlane, 1979), consists of a basal phreatomagmatic deposit covered by two flow units, each formed by a welded layer and a non-welded layer of ash-flow deposits (Robin et al., Mar., 1994; Allen, 2005). Late Pleistocene–Holocene volcanic activity has formed two scoria cones inside the caldera, Mounts Ombus and Yasur on the south and west edges of the Yenkahe resurgent block respectively. These edifices were built from pyroclastic apron ejected during the explosions. The only effused lava flows coming from the Yasur activity, the last probably occurring in 1878, are found in the south-eastern part of the Siwi caldera (Aubert de la Rüe, 1960). The north-western part of the caldera, named the Siwi plain, is filled with ash. The presence of the former Siwi Lake and several rivers coming from neighbouring reliefs makes this area a zone of alluvial filling.

3. SPAC and f–k methods

The SPAC and f–k techniques have been used to determine the autocorrelation and dispersion curves of the surface waves for each array.

The principle of the SPAC technique is based on Aki (1957). The spatial auto-correlation function between two sensors is defined by:

$$C(\xi) = \frac{1}{T}\int_0^T \upsilon_0(t)\upsilon_\xi(t)dt \quad (1)$$

where υ_0 and υ_ξ are the signals recorded during T seconds at two stations separated by a distance ξ. If the signals are filtered in a narrow frequency band around ω, the auto-correlation ratio can be expressed as:

$$\rho(\xi,\omega) = \frac{C(\xi,\omega)}{C(0,\omega)} \quad (2)$$

This ratio is computed for all pairs of seismometers. For a given distance ξ, Aki (1957) demonstrated that the azimuthal average of $\rho(\xi,\omega)$ can be defined by:

$$\overline{\rho(\xi,\omega_0)} = J_0\left(\frac{\omega_0\xi}{c(\omega_0)}\right) \quad (3)$$

with J_0 the Bessel function of zero order and $c(\omega_0)$ the phase velocity of waves in the medium for the pulsation ω_0, i.e. the dispersion curve.