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Separation of scattering and intrinsic attenuation at Asama volcano (Japan): Evidence of high volcanic structural contrasts

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ABSTRACT

In this study we show 2D intrinsic- and scattering-Q images of Asama volcano obtained by analyzing 2320 waveforms from active data. Observed energy envelopes were fitted to the diffusion model and separate intrinsic- and scattering-Q images were produced using a back-projection method based on a Gaussian-type weighting function. Synthetic tests indicate robustness and reliability of the results. Areas of high scattering attenuation coincide with the volcanic edifice and the summit at which recent eruptions took place. The intrinsic dissipation pattern shows a strong contrast between the east and west side of the volcanic structure with the low values observed in the west interpreted as solidified magma bodies. Our results demonstrate a strong relationship between structural heterogeneities and attenuation processes in volcanic areas and confirm the effectiveness of the present technique, which can be used as an imaging tool complementary to conventional techniques.

1. Introduction

Studying the heterogeneity of volcanic regions gives more insights into the magma pathway, which is under structural controls (e.g. Aoki et al., 2013). The structural complexity of volcanoes is one of the most important factors controlling the attenuation parameters of seismic energy in these regions, spanning a wide variety of cases associated with different geological conditions and volcanism (e.g. Del Pezzo, 2008; Sato et al., 2012). Seismic attenuation is driven by intrinsic dissipation and scattering effects, the first of which is controlled by the rheology, while the other by geological inhomogeneity. Both effects are present in volcanoes, so it is essential to measure the amount of intrinsic dissipation with respect to scattering attenuation to better understand the volcanic structure. For example, high intrinsic attenuation can be due to an elevated temperature of rocks, while high scattering attenuation may be generated by heterogeneity due to the unconsolidated volcanic deposits accumulated in years of activity. In addition hydrothermal activity may produce a rheological alteration of some geological layers which would increase both the intrinsic dissipation and inhomogeneities beneath the volcano (see e.g. Prudencio et al., 2015), in turn increasing scattering attenuation too.

In volcanic areas, interpretation of estimated attenuation coefficients becomes more difficult by steep topography, which produces surface wave scattering which severely affects the seismogram shape (see e.g. Del Pezzo et al., 1997; O'Brien and Bean, 2004; O'Brien and Bean, 2009; Bean et al., 2008; Lokmer and Bean, 2010). A main conclusion of the previous studies cited above is that topographical contrasts are one of the main scattering sources in volcanoes. Numerical simulations (Lokmer and Bean, 2010) indicate, however, that the waves scattered from topographical contrasts modify only the very early coda of seismograms, leaving the coda spectral content almost unaffected at longer lapse time. This property can be utilized in developing robust techniques to separate intrinsic and scattering attenuation coefficients from coda waves in order to obtain a spatial map of such parameters. In turn a useful joint interpretation of the spatial distribution of seismic velocities can be obtained with conventional seismic tomography and other geological/petrological/geophysical evidence.

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In literature, there are a few methods such as the well known Multiple Lapse Time Window Analysis (MLTWA) (e.g. Hoshiba, 1991; Hoshiba, 1993; Akinci et al., 1995; Del Pezzo et al., 1995) that allow a separate estimation of intrinsic and scattering attenuation coefficients. However, the MLTWA method cannot be used for single paths, which forbids obtaining regional maps of attenuation parameters. Wegler and Lühr (2001) and Wegler (2003), using bandpass-filtered seismogram envelopes and a diffusion model, inverted each seismogram separately in the time domain and they estimated attenuation coefficients from the shape of the envelope at Merapi and Vesuvius volcanoes. These authors have demonstrated that a diffusion model can be used to model envelopes of observed waveforms, showing that this single-station technique permits intrinsic and scattering attenuation coefficients to be obtained separately with reasonably small uncertainties when multiple scattering dominates as in volcanoes (Sato et al., 2012; Del Pezzo, 2008), and therefore, the regional distribution of the attenuation parameters in small scale regions.

Imaging regional distribution of attenuation coefficients has previously been based on very simple assumptions: assigning the Q-values to the position of the seismic station (Carcolé and Sato, 2010) or to the mid point between the source and station (e.g. Pujades et al., 1990; Canas et al., 1995). In both cases standard averaging procedures were used to obtain the distribution maps. Xie and Mitchell (1990) described a more accurate representation method using a back-projection method by applying the single scattering model and assuming that the obtained attenuation parameter represents the average seismic attenuation inside the scattering ellipse. Calvet and Margerin (2013) and De Siena et al. (2014) applied similar assumptions to their studies of the Pyrenees and Mount Saint Helens. Based on previously described assumptions, Prudencio et al. (2013a) developed a new representation technique, which is a new way of spatial averaging of attenuation coefficients with a back-projection method using a Gaussian weighting function. They demonstrated that this representation technique is robust and provides an improvement in imaging resolution comparing to standard averaging methods.

In the present work we study the seismic attenuation of Asama volcano (Japan) using the abovementioned techniques to provide 2D horizontal images of both scattering and intrinsic seismic attenuation. These images are jointly interpreted with the recent seismic velocity tomography obtained by Aoki et al. (2009a,b) and other volcanological evidence. These new results help us to provide a unified model of the magma plumbing system of Asama volcano that can be used to better interpret and constrain its dynamics.

2. Asama volcano and data

Asama volcano (2568 m), located about 160 km from Tokyo, is one of the most active volcanoes in Japan (Aramaki, 1963). This andesitic volcano erupted many times in historical time with frequent vulcanian eruptions, such as the 2004 eruption. The largest Plinian eruptions occurred in 1108 and 1783 (VEI 5) (Miyazaki, 2003) and its most recent eruption took place in June 2015. Due to this potentially explosive behavior of Asama volcano around 20 million people live under an evident volcanic risk. For this reason Asama volcano has been continuously monitored since 1911 (Omori, 1914). The vulcanian eruption that occurred in 2004 revealed some open questions regarding the magma pathway and the inner structure beneath Asama volcano. In order to solve those questions, an active seismic experiment was performed in October 2006 (Aoki et al., 2009a,b) aiming at obtaining seismic velocity images. During the experiment 464 portable seismic stations were deployed and seismic data was generated by five dynamite sources of 250–300 kg (Fig. 1). Mark Products L22-D (natural frequency of 2-Hz) and GeoSpace GS-11D (natural frequency of 4.5 Hz) seismometers were deployed with an average spacing of 100–150 m and 250 Hz sampling rate. The

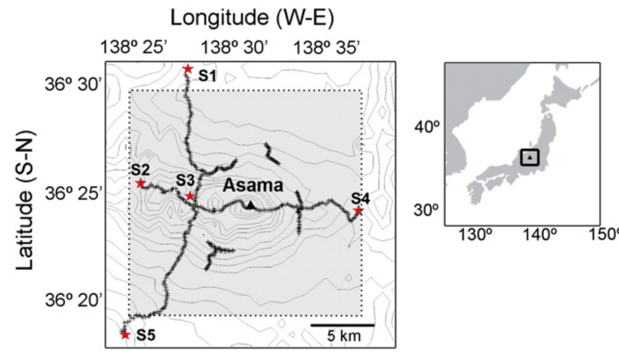

Fig. 1. Regional settings and location of the Asama volcano in Japan. (right) Map of the active seismic experiment carried out in Asama volcano during October 2006. Position of dynamite shots appear as red stars and the positions of seismic stations are shown as black crosses. The position of the Asama crater is marked with the black triangle. The region shown in the results is highlighted in gray.

configuration of the experiment was designed to delineate the seismic velocity structure of the dike intrusion area of 2004 eruptions. The experiment revealed the existence of a body with high seismic velocity 4 km west of the summit, which was interpreted to be the solidified magma intrusion. Combining this interpretation with seismic and geodetic observations (Takeo et al., 2006; Aoki et al., 2013) suggested that magma intrusions have occurred repeatedly at similar locations to the west of the summit. In the present work, we use the same data-set used in Aoki et al. (2009a,b).

3. Data processing

In the present work, we used the diffusion model, which is the asymptotic approximation of the transport equation in case of high density of scatterers (Sato et al., 2012) to obtain intrinsic and scattering attenuation values separately. For each source-receiver pair, intrinsic and scattering attenuation coefficients have been estimated by fitting the squared envelope of filtered seismograms to the diffusion model. The fitting procedure in this study is the same as the one applied by Wegler and Lühr (2001) and Wegler (2003), who applied this method to Merapi and Vesuvius volcanoes. The entire procedure is fully described in Prudencio et al. (2013a) for Tenerife Island and recently updated by Prudencio et al. (2015) who applied this method to an active source seismic dataset of Stromboli volcano. Then, 2D images were obtained using a back-projection method based on a Gaussian-like weighting function which is described in the next section.

The diffusion model (Eq. (1)) describes the seismogram energy envelope (E[r,t]) as a function of source-receiver distance, r, and lapse time (measured from origin time), t, as:

$$E[r,t] = \frac{E_0}{(4\pi dt)^{-p/2}} \exp\left[-\frac{r^2}{4dt} - bt\right] \tag{1}$$

where E_0 is the source energy, d is the diffusivity, b is the intrinsic attenuation coefficient and p represents the geometrical spreading term ($p = 2$ for surface waves and $p = 3$ for body waves) (Dainty and Toksöz, 1981). Coefficients d and b are directly related with intrinsic- and scattering quality factors, Q_i and Q_s, respectively, through the following equations:

$$Q_i = \frac{2\pi f}{b} \tag{2}$$

$$Q_s = \frac{2\pi f p d}{v^2} \tag{3}$$