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Crustal structure below Popocatépetl Volcano (Mexico) from analysis of Rayleigh waves

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Abstract

An array of ten broadband stations was installed on the Popocatépetl volcano (Mexico) for four months between October 2002 and February 2003. 26 regional and teleseismic earthquakes were selected and filtered in the frequency time domain to extract the fundamental mode of the Rayleigh wave. The average dispersion curve was obtained in two steps. Firstly, phase velocities were measured in the period range [2–50] s from the phase difference between pairs of stations, using Wiener filtering. Secondly, the average dispersion curve was calculated by combining observations from all events in order to reduce diffraction effects. The inversion of the mean phase velocity yielded a crustal model for the volcano which is consistent with previous models of the Mexican Volcanic Belt. The overall crustal structure beneath Popocatépetl is therefore not different from the surrounding area, and the velocities in the lower crust are confirmed to be relatively low. Lateral variations of the structure were also investigated by dividing the network into four parts and by applying the same procedure to each sub-array. No well-defined anomalies appeared for the two sub-arrays for which it was possible to measure a dispersion curve. However, dispersion curves associated with individual events reveal important diffraction for 6 s to 12 s periods which could correspond to strong lateral variations at 5 to 10 km depth.

Keywords: volcano seismology; Popocatépetl Volcano; Rayleigh waves; crustal structure

1. Introduction

Popocatépetl is a large andesitic strato-volcano, located 60 km south-east of Mexico City and 40 km West of Puebla (Fig. 1a). Part of the Trans-Mexican Volcanic Belt (MVB), Popocatépetl's cone is the second highest summit of Mexico (5452 m above sea level) with an ellipsoidal 600–800 m-wide crater.

The present active period began on December 21st 1994. Since 1996, an andesitic to dacitic dome grows cyclicly into the crater, accompanied by bursts producing high plumes of gas and ash (Arcieniega-Ceballos et al., 2000; Wright et al., 2002). More than 100 000 persons could potentially be directly affected by a major eruption and ashes could affect an area with more than 20 million people (De La Cruz-Reyna and Siebe, 1997; Macías and Siebe, 2005).

The overall crustal structure beneath the MVB is relatively well studied (Campillo et al., 1996; Valdes et al., 1986; Shapiro et al., 1997). However, the crustal seismic structure beneath Popocatépetl itself is not well known. Receiver-functions analysis by Cruz-Atienza et al. (2001), using 4 events from South America, indicates that a low-velocity zone may be present beneath a station located 5 km north of the crater.

The aim of this paper is to improve the knowledge of this complex volcano structure, and particularly to determine if the

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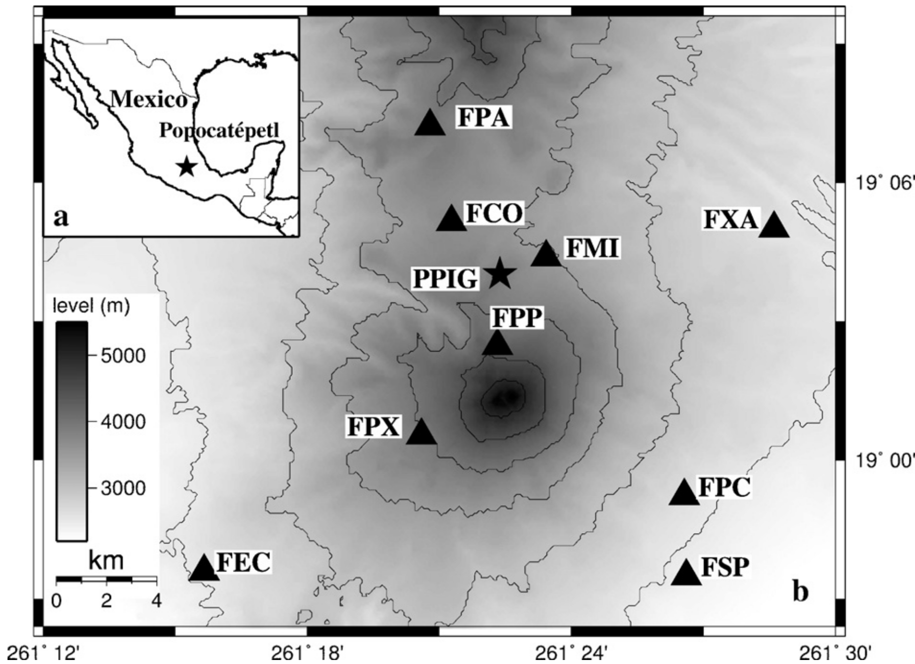

Fig. 1. (a) Location of the Popocatépetl Volcano and (b) array geometry used in the analysis. PPIG is a permanent station used by Cruz-Atienza et al. (2001) and is not used in this study.

entire crust beneath the volcano is significantly different from the rest of the MVB. The first few kilometers of crust beneath several volcanoes have been studied (e.g. Benz et al., 1996; Dawson et al., 1999; Laigle et al., 2000). Typical volcanic anomalies are low-velocity zones, attributed to the presence of partial melt, or high-velocity zones, due to solidified magmatic intrusions.

We concentrate on the S-wave structure, as S-wave velocities are very sensitive to temperature changes and to the presence of even small amounts of partial melt. The easiest way to get an overall picture of S-wave velocities is through surface-wave analysis. However, the traditional 2-stations methods can not be used in this rather diffractive environment, because measurements would possibly be strongly biased due to local and regional diffraction (Wielandt, 1993; Friederich and Wielandt, 1995). An alternative approach, therefore, is to use array analysis. Such methods have been used on other volcanoes, particularly for tremor source location (Métaxian et al., 2002; Almendros et al., 2002) or for shallow structure study (See for example Saccorotti et al., 2001). More details can be found in Chouet (2003), who presents a review of the state of the art of volcano seismology.

The dispersion curve had to be measured over a wide frequency range (0.02–1 Hz) to study the overall crustal structure beneath Popocatépetl. The array configuration, which was strongly influenced by topography and logistical constraints issues was such that we could not use spatial Fourier transforms outside a very narrow frequency range. Consequently, methods based on wavenumber decomposition were excluded. The use of time-domain methods was problematic as we needed a good frequency resolution.

These considerations led us to use the procedure of Pedersen et al. (2003) to measure phase velocities across the array. The assumption behind this method is that the records are constituted by one single-plane wave which propagates through the array. Even though this hypothesis is most probably wrong for most individual events, it may be corrected by averaging out unwanted waves (diffraction effects, non-plane waves, etc.) using events from different directions. The variability between different events will also provide an error estimate on the dispersion curve. To increase frequency range and azimuthal coverage we used both teleseismic and local events.

After a short description of the data and the processing methods used, we present and discuss the main results, with a comparison of the overall crustal structure beneath the volcano to that of the MVB.

2. Data

An array of nine stations (Guralp CMG 40 T) with three-components broadband sensors (30–60 s cut-off period) was installed in October 2002 on Popocatépetl Volcano and continuously recorded four months of seismic events. Fig. 1b shows the array geometry. The station altitudes were between 2500 and 4300 m above sea level. The reference altitude used in this study corresponds to an average level of 3500 m a.s.l.

To obtain dispersion curves in a period range of 2 to 50 s, we chose to use both teleseimic and local events with epicentral distances between 200 and 15000 km (see Fig. 2). We selected vertical components of events with a good signal-to-noise ratio and with well-developed Rayleigh waves. The usable frequency range for the two types of events overlapped; however, the long period part of the dispersion curve was mainly calculated using