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Transition zone structure under a stationary hot spot: Cape Verde

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ABSTRACT

We report on a two-year seismic deployment in the Cape Verde Islands, one goal of which was to study the upper mantle to determine its structure under a hot spot that is stationary in the hot spot reference frame. We find from analysis of P-to-S receiver functions estimated from broadband seismic recordings that, within uncertainty, the time separation between the 410 and 660 km discontinuities is normal compared to radial earth models. Thus, to exist, even stationary hot spots do not require vertical thermal anomalies from deep melting sources anchored in the lower mantle or at the core–mantle boundary or their anomalies are narrower than ~250 km in the upper mantle.

1. Introduction

Oceanic hot spots are an old concept motivated by observational characteristics: long-term volcanism with an age progression, anomalously shallow seafloor, and regional gravity or geoid and heat flow anomalies (Wilson, 1963; Morgan, 1972). The idea has survived to the present [though with the last criterion discarded due to its difficulty to show; Stein and Stein, 1993; Harris and McNutt, 2007], but the physical process of hot spot creation and sustenance is still controversial. One idea is that deep-seated melting anomalies in the lower mantle or at the core–mantle boundary buoyantly deliver a flux of material to hot spots (Morgan, 1971; Duncan and Richards, 1991; Davies and Richards, 1992). This model envisages vertical thermal anomalies associated with a conduit connecting the lower-mantle source with the surface features.

The hypothesis has testable consequences. On account of the thermal anomaly, reckoned to be ~200 K (Sleep, 1990), the phase transitions that demark the seismic discontinuities at 410 and 660 km should be perturbed, and their depths changed by 20–30 km (Helffrich, 2000). A vertically-coherent warming of the transition zone should deepen the 410 and shoal the 660, leading to a thinner transition zone (TZ, the separation between the 410 and 660). Past studies used this effect to test whether particular hot spots had deep sources. Shen et al. (1998) found the TZ 20 km thinner below Iceland; Li et al. (2000) found it thinned by 40–50 km below Hawaii; Humphreys et al. (2000) found it 20 km thinner under the slow-moving Yellowstone hot spot; and Suetsugu et al. (2007) reported thinning up to 30 km under a south Pacific hot spot. These studies all support the idea of a lower mantle melting origin.

In view of this consistent verdict of a deep source, one further hot spot study seems superfluous. We argue that a study of Cape Verde, a hot spot located 400 km off the western-most point of Africa, is not, for two reasons. Firstly, the upper mantle beneath it has not yet been characterized seismically. Secondly, unique among hot spots, it is stationary in the hot spot reference frame (Gripp and Gordon, 2002). Thus the deep-sourced material flux should be located directly beneath the islands, as distinct from hot spots on moving plates (Hawaii, Galapagos) or near ridges (Iceland) or plate boundaries (Azores, Easter) where the conduit position could be affected by plate motion or ridge melt flux. Consequently, the transition zone under Cape Verde affords a unique view into hot spot formation processes. Our results indicate that the TZ is not thinned under Cape Verde, suggesting that no deep-seated thermal anomaly exists under the archipelago. The mean apparent 410 km discontinuity depth is deeper than whole-earth models predict and suggests that a slow upper mantle but an unperturbed transition zone exists under Cape Verde.

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2. Data and methods

The data were collected during a temporary seismic deployment in Cape Verde which lasted from July 2002 to September 2004. The project was collaborative between institutions in the UK, Cape Verde, Portugal and Japan. Instruments were broadband Guralp 3Ts deployed on six of the islands and augmented by the GSN station SACV located on Santiago (Fig. 1). Data was recorded on local disks that were serviced semi-annually. Under funding agency agreements, the data is deposited with the IRIS DMC.

The portable network data was originally collected at 50 sps but decimated to 20 sps for receiver function (RF) estimation. We selected events with $M_w \geq 5.8$ and with distance Δ $30° \leq \Delta \leq 95°$ for P-wave study and $60° \leq \Delta \leq 180°$ for PP. The PP arrivals were generally emergent and therefore not further analyzed. There were 41 events with suitable P that occurred during the deployment and the longer operational period of the GSN station. Most portable stations recorded 16–18 events except for two where instrument failures reduced the yield to 8; the GSN station yielded 30. We calculated P-to-S receiver functions around the P arrival times using two independent methods, a multi-taper frequency domain method (ETMTRF; Helffrich, 2006) and a time domain method (Lígorria and Ammon, 1999). Records were classified as qualities A, B, or C depending on data and receiver function characteristics. A trace was category C if the P arrival was not visible in the raw data, category B if the ETMTRF RF derived from it was acausal or if its largest radial peak was not the leading one in the record, and category A if the RF had no significant acausality and the pulse train was dominated by the zero-lag peak. Category B (32 traces) and category C (8 traces) were discarded; the remaining category A data resulted in 82 estimates network-wide from 39 unique events whose sources are shown in Fig. 1. P-wave slownesses for the sources ranged from 4.62 to 8.82 s/°. Each method uses a high-frequency cutoff to limit the complexity of the RF estimates. We calculated RF suites with cutoff frequencies of 0.1, 0.25, 0.5, 1 and 2 Hz and stacked the results at each frequency, migrating each trace to a common slowness of 6.5 s/° before stacking. The stacking is linear and unweighted. Stacks for 0.5 Hz are shown in Fig. 2 for the two methods used. A limit of 150 peaks was imposed for the time domain method. This resulted in a recovery of >85% of the power in the original seismogram; lower frequency cutoffs yielded higher values, closer to 95%.

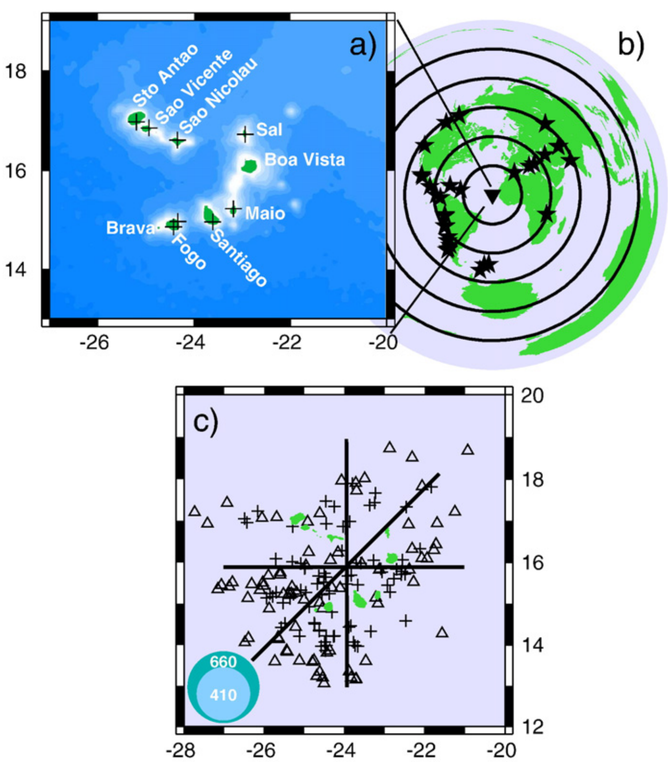

Fig. 1. a) The Cape Verde archipelago consists of 10 major islands, 7 of which were occupied (crosses) with a portable seismic network. Station on Santiago is the permanent GSN station SACV. Bathymetric shading delineates two chains: Sal-Brava and Santo Antão-São Nicolau. Island ages are not firmly established, but physiography suggests that eastern islands in each chain are oldest and western youngest. Fogo last erupted in 1995. b) Location map of Cape Verde in the Atlantic Ocean basin. The azimuthal-equidistant projection also shows the locations of the earthquakes (stars) used to analyze the transition zone under the archipelago. Range rings spaced every 30°. c) Piercing points of P-to-S conversions studied from sources shown in (b). Crosses indicate Ps conversions at 410 km depth, and triangles at 660 km depth. 0.5 Hz $\lambda/2$ Fresnel zones at 410 and 660 km shown at bottom left. Lines indicate positions of cross-sections shown in Fig. 4.

Our choice of methods, and the use of stacks at different frequencies were motivated for a variety of reasons. The main problem with RF estimation at ocean island stations is high microseismic noise which has a broad peak at 0.2–0.3 Hz (Bromirski and Duennebier, 2002). Though multi-taper methods are designed to defeat narrow-spectrum noise, the use of frequency cutoffs below the peak assures the influence in mitigated. A secondary motivation is that the discontinuities may be broadened due to thermal and chemical effects (Helffrich and Bina, 1994), leading to high-frequency estimates further being contaminated by noise if the discontinuity was broadened. Finally, by using time domain and frequency domain methods, we can verify similarity of results. This provides a performance cross-check against noise and validation of the implementation of the methods.

The result confirms that the two methods yield similar estimates despite the high microseismic noise levels present in ocean island environments. The stacks feature prominent peaks at around 46 s and 70 s associated with the TZ discontinuities at 410 and 660 km, respectively. Fig. 1 also depicts the piercing points at 410 and 660 km for all RFs, which shows a fairly complete coverage of the archipelago.

We also calculated RF synthetics to check the resolving power of the experiment. We used Frazer and Sen's (1985) Kirchhoff-Helmholz method to calculate noise-free seismograms containing Ps conversions from the 410 and 660 topographically perturbed by the thermal anomaly associated with a rising plume under the archipelago. Using Bina and Helffrich's (1994) estimates of 410 and 660 depth perturbations from a plume-characteristic ~200 K thermal anomaly with a parabolic profile expected from conduit flow leads to 410 depression of ~20 km and 660 elevation of ~13 km. The response depends on conduit diameter, so we calculated synthetics for diameters 1, 2, 3, 4 and 5°. The same source–receiver geometry and slownesses were used in order to mimic the data coverage along with the same time windows and methods for the data RFs at a cutoff frequency of 1 Hz, and linearly stacked the synthetics migrated to a common slowness of

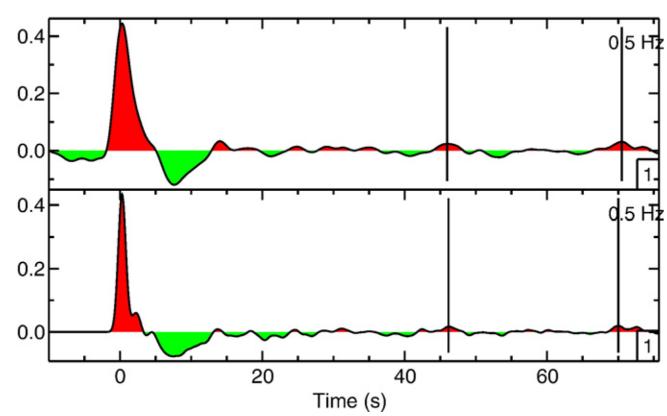

Fig. 2. Migrated and stacked receiver function estimates using independent methods. 0.5 Hz high-frequency cutoff applied during RF estimation. Frequency domain multi-taper method (top) and time domain method (bottom) both feature peaks at ~46 s and ~70 s (vertical marks), corresponding to 410 and 660 km conversions. The reference slowness is 6.5 s/°.