Studying the 410-km and 660-km discontinuities beneath Spain and Morocco through detection of $P$-to-$s$ conversions

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SUMMARY

In this study, we analyse the 410-km and 660-km upper-mantle transition zone discontinuities as seen from seismic $P$-to-$s$ wave conversions beneath the Eurasian–African Plate boundary at south Spain and north Morocco. For this purpose we use teleseismic events recorded at 43 broad-band seismic stations deployed mainly by the TopoIberia project. The conversions from the upper-mantle discontinuities arrive in the $P$-wave coda together with other signals and are usually identified on stacked receiver functions. We build a new processing approach which is leaned on receiver functions and which is based on cross-correlation and stacking techniques to efficiently detect and extract signals by means of their coherence, slowness, traveltime and polarity. In order to add consistency and robustness to the detections, our final results are based on a joint analysis of two different cross-correlation functionals and receiver functions. This permits to assess errors and to bridge observation gaps due to the breakdown of any of the techniques inherent to signal and noise characteristics. Finally, discontinuity depths are determined using time corrections obtained from a 3-D velocity model. We present topography maps for the 410-km and 660-km discontinuities which show a thickening of the transition zone beneath the plate boundary towards Morocco. The transition zone thickness is about global average beneath south Spain (240–250 km) and is thicker beneath east Morocco (250–275 km). This is mainly due to a deeper 660-km discontinuity, while the topography of the 410-km discontinuity is smaller. In the Alboran Sea we find an up to 25 km deflection of the 660-km discontinuity which suggests that the Alboran Sea heterogeneity or slab is still sufficiently cold to depress the post-spinel phase transition. We finally discuss the results in order to add new constraints on temperature and composition to seismic velocity anomalies observed in the transition zone beneath south Spain and north Morocco.

Key words: Time-series analysis; Body waves.

1 INTRODUCTION

The western part of the boundary between the African and European plates comprises a complex tectonic setting still not completely understood. In the Atlantic Ocean the plate boundary separates oceanic lithosphere while in the Alboran Sea (western Mediterranean) the contact is not well defined (e.g. Vernant et al. 2010) and involves a continental collision. The convergence of Africa with respect to Europe initiated in the Cretaceous between 120–83 Myr (Rosenbaum et al. 2002), and the current convergence rate is of 2–6 mm yr$^{-1}$ (Argus et al. 1989; Demets et al. 1990). Present-day seismic activity occurs from shallow to intermediate depth in a diffuse band which extends to both sides of the Strait of Gibraltar. Earthquake activity stops at 150 km depth and reappears in a small area south of Granada at about 630–660 km depth (Buforn et al. 2011; Bezada & Humphreys 2012). The origin of these deep isolated events is still an open question. Most tomography studies have revealed a positive $P$-wave velocity anomaly beneath the Alboran Sea and southeast Spain which is visible down to the base of the upper mantle transition zone (TZ) (Spakman 1990; Blanco & Spakman 1993; Wortel & Spakman 2000; Piromallo & Morelli 2003; Spakman & Wortel 2004; Faccenna et al. 2004; Garcia & Villaseñor 2011). There is a long debate on the origin of this anomaly (Platt & Vissers 1989; Royden 1993; Seber et al. 1996; Wortel & Spakman 2000; Gutscher 2002; Faccenna et al. 2004; Bokelmann et al. 2011; Verges & Fernandez 2012, among others) as it is an important clue to understand the geodynamic state of the western Mediterranean. Most explanations of the positive velocity anomaly involve different types of continental delamination, convective removal or subduction processes. SKS splitting (Buontempo et al. 2008; Díaz et al. 2010) and $P$-wave dispersion analyses (Bokelmann & Maufroy 2007) are consistent with the presence of a subducted lithosphere. Nevertheless,
there is still room for different interpretations. Seismic analyses to characterize the volumetric upper-mantle heterogeneities revealed in tomographic images become crucial to provide additional constraints, such as thermal and compositional information. The study of the 410-km and 660-km discontinuities is most probably one of the best approaches, since these discontinuities are globally observed mineral phase transitions which, as function of composition, respond with depth and thickness variations to temperature anomalies.

The 410-km and 660-km discontinuities (hereafter referred to as 410 and 660, respectively) limit the TZ. In an upper mantle of pyrolitic composition (60 per cent olivine with additional garnet and pyroxene), the 410 is the result of olivine to wadsleyite transition, while the 660 is the dissociation of ringwoodite into perovskite + magnesiowustite. Temperature anomalies in the mantle move the phase changes to different pressures (depths) according to the Clapeyron slopes (e.g. Bina & Helffrich 1994). Due to the opposite sign in the Clapeyron slopes of the phase changes responsible for each discontinuity, the 410 and 660 depths changes are anticorrelated as response to a thermal anomaly. While the 410 becomes shallower in colder regions and deeper in hotter ones, the 660 behaviour is opposite. As a consequence, the TZ becomes thicker near subducted slabs and thinner beneath plumes or high temperature regions due to small scale mantle convection (Vidale & Benz 1992; Helffrich 2000; Collier et al. 2001; Lawrence & Shearer 2006). However, the presence of other transforming or non-transforming mantle components can change the characteristics of the phase transitions and makes the interpretation of the corresponding seismic discontinuities more difficult (Stixrude 1997; Weidner & Wang 1998; Wang et al. 2004; Deuss et al. 2006; Tauzin et al. 2008; Thomas & Billen 2009; Cao et al. 2011).

In this paper, we present maps of the upper-mantle TZ discontinuities beneath south Spain and north Morocco to provide additional and independent constraints to aid and strengthen the interpretation of the observed seismic velocity anomalies. The interpretation of these anomalies has direct implications for the understanding of the geodynamic state of the western Mediterranean. The discontinuities are not resolved with seismic tomography and their study requires the detection and identification of seismic body waves which directly interact with the discontinuity through a reflection and/or wave type conversion (from $P$ to $S$, or $S$ to $P$). Reflection/conversion coefficients are typically smaller than 5 per cent and the corresponding small-amplitude signals are concealed in a multitude of other scattered waves. This makes it difficult to identify the signals on individual records. Upper-mantle discontinuities are most commonly studied using the receiver function technique (RF) (Phinney 1964; Vinnik 1977; Langston 1979; Ammon 1991) which enhances the $P$-to-$s$ conversions/reflections from discontinuities below the recording stations. Here, we build a new processing approach which is leaned on RFs and which is based on cross-correlation and stacking techniques. The instantaneous phase coherence obtained from analytic signals forms the backbone of one of the cross-correlation approaches and stacking used. We focus on $P$-to-$s$ conversions which we detect and extract based on their coherence, slowness, traveltime and polarity and which we use to map the discontinuities. In order to add consistency and robustness to the detections, our final results are based on a joint analysis of two different cross-correlation functionals and RF. This approach further permits to assess errors and to bridge observation gaps due to detection failure of any of the approaches.

Our data set consists of a vast volume of seismograms thanks to the IberArray of the Spanish Topolberia project (Diaz et al. 2009). For our analysis we use teleseismic events recorded from 43 three-component broad-band stations (Fig. 1). The analysis extends previous RF studies of the TZ discontinuities beneath our study area (van der Meijde et al. 2005; Dündar et al. 2011), which did not have access to the new data volume. Furthermore, the new data volume permits to resolve new TZ topography.

In what follows, we explain the methodologies employed and show their performance using theoretical and real data. Then, we discuss the data processing approach and signal extraction using the Topolberia data. The upper-mantle conversions are corrected for 3-D heterogeneities and mapped to depth. Finally, we present the maps of the estimated 410 and 660 depths and discuss five north–south profiles located in the vicinity of the Alboran Sea and the Gulf of Cadiz.

2 METHOD

The TZ discontinuities are generally studied through the detection and identification of different body wave phases present in the seismic records (Shearer 1991, 2000). The converted waves at the 410 and 660 arrive in the $P$-wave coda together with multiple reflected and scattered waves. Coda phases are characterized by low amplitudes; this makes the identification of coda phases difficult within the multitude of different other phases in individual records. However, with individual stations these coda signals can be detected whenever they are coherent with a template or pilot, like the direct $P$ waveform. The coda phases are expected to be coherent with the waveform of the first arrival for conversions/reflections at discontinuities which are thinner than one fourth of the wavelength (Richards 1972; Paulsen 1988; Bostock 1999). Taking advantage of this property we apply coherency measurement tools to detect coda signals by their waveform similarity as function of lag time. We apply different cross-correlation techniques between components of teleseisms recorded at individual stations. Finally, the result cross-correlograms are stacked and signals are identified by the measured traveltime, slowness and polarity.
2.1 Theoretical background

In order to determine the waveform similarity between the coda phases and the P phase, we apply two cross-correlation techniques which are based on different strategies: the classical cross-correlation geometrically normalized (CCGN) and the phase cross-correlation (PCC) presented by Schimmel (1999). To enhance coherent signals and to suppress incoherent noise we use the phase weighted stack (PWS) (Schimmel & Paulsen 1997). In the following sections, we briefly explain the cross-correlation approaches and the PWS.

2.1.1 Cross-correlations

In analogy to the classical cross-correlation, the PCC measures the waveform similarity between two signals as function of lag time. It is based on the instantaneous phase similarity of the corresponding analytic traces. Given a seismic trace $s_i(t)$, the PCC detects the signals included in $s_i(t)$ that are coherent with a reference or pilot wavelet, $s_p(t)$. For this purpose, wavelet $s_p(t)$ is shifted in time and compared with the corresponding portion of the seismic trace $s_i(t)$. The PCC expression is given by

$$PCC_i(t) = \frac{1}{2\tau^2} \sum_{t=0}^{\tau-1} [e^{\phi_1(t)} e^{\phi_2(t)}] - |e^{\phi_1(t)} - e^{\phi_2(t)}|^v,$$

where $e^{\phi_1(t)}$ and $e^{\phi_2(t)}$ are the amplitude-normalized analytic signals, while $\phi_1(t)$ and $\phi_2(t)$ are the instantaneous phases of the seismic trace $s_i(t)$ and the pilot $s_p(t)$, respectively. $T$ is the pilot length in samples, $t$ is the lag time and $\tau_p$ is the start time of the correlation window. The normalization term $1/(2\tau^2)$ ensures that $|PCC_i(t)| \leq 1$, with $PCC_i = 1$ in case of perfect correlation and $PCC_i = -1$ for anticorrelation. The sharpness of the transition between similarity and dissimilarity is controlled by the power $v$. We use PCC with $v = 1$ throughout this paper.

In addition to the technique described above, we employ the CCGN, which is the classical cross-correlation normalized by the geometric energy of the traces. This measure varies between $-1$ and $+1$, where $+1$ corresponds to perfect sign coherence, and $-1$ corresponds to perfect coherence of signals of different polarity. The CCGN expression is given by

$$CCGN(t) = \frac{\sum_{t=0}^{\tau-1} s_i(t + \tau) s_p(t)}{\sqrt{\sum_{t=0}^{\tau-1} s_i(t + \tau)^2} \sqrt{\sum_{t=0}^{\tau-1} s_p(t)^2}}.$$

CCGN and PCC are independent approaches which are based on different strategies. PCC is amplitude unbiased and is more sensitive to waveform coherence than CCGN and therefore, well suited for the detection of coherent weak amplitude signals. CCGN is based on the sum of signal amplitude products and is therefore less sensitive to waveform coherence. The decreased sensitivity may favour signal detection when there is less waveform similarity due to waveform distortion.

2.1.2 Phase weighted stack

The PWS suppresses signals that do not stack coherently. This technique uses the phase stack (PS) as a time-dependent weight of the linear stack (LS). PS measures the phase coherence based on the similarity of the instantaneous phases and is obtained by summing up the envelope normalized analytic signals. The PWS expression is given by

$$PWS(t) = LS(t) P(t) = \frac{1}{N} \sum_{j=1}^{N} |\sum_{k=1}^{N} e^{i\phi(t)}|^v,$$

where $s_i(t)$ is the $j$th seismic trace and $\phi(t)$ is the instantaneous phase of the respective analytic signal. Each sample in the LS is weighted by the coherence of the instantaneous phases obtained from all individual traces $s_i(t)$. Thus, small amplitude signals which are coherent are enhanced through the attenuation of incoherent noise. The PS acts as a phase similarity filter. The parameter $v$ controls the sharpness between phase similarity and dissimilarity. The LS is retrieved with $v = 0$. In what follows, we use the PWS with $v = 2$ for synthetic and real data.

2.2 Detection of P-coda phases using cross-correlation

In order to detect the weak amplitude converted and reflected phases at the upper-mantle discontinuities, we combine cross-correlation (CCGN, PCC) and stacking (PWS) techniques. Under the assumption of lateral homogeneity and considering teleseismic earthquakes, most of the P-wave energy arrives on the vertical (Z) component and the SY-wave energy, such as P-to-s conversions, on the radial (R) component. Therefore, we only consider the R and Z components of teleseismic recordings from individual stations. First, we extract a pilot wavelet ($P_{PZ}$) from the Z component, which contains the P phase and part of its coda with the later arriving depth phases. Then, we perform the PCC and the CCGN between the pilot $P_{PZ}$ and the R and Z components. For each cross-correlation method we obtain an R and Z correlogram (hereafter referred to as PCCR or CCGNR and PCCZ or CCGNZ). The maximum amplitudes are obtained for lag times where $P_{PZ}$ and a particular coda segment on the R or Z component show waveform similarity. Fig. 2 shows an example of signal detection using the correlation (PCC and CCGN) of a pilot $P_{PZ}$ with the R and Z components. The example uses a synthetic seismogram for an event at 55° distance (the synthetic data generation is explained in Section 2.4.1). $P_{PZ}$, R and Z are shown at the top and the corresponding correlations at the bottom of Fig. 2. The correlation maxima of PCCR and CCGNR at about 45 and 70 s are due to the waveform similarity of the $P_{PZ}$ with the P410s and P660s conversions, which are larger than for other coda signals. The maximum at zero lag is due to the recorded P phase on R. In analogy to PCCR and CCGNR, correlation minima on PCCZ and CCGNZ show the topside P-wave reflections $P_{P410p}$ and $P_{P660p}$ from the 410 and 660 upper-mantle discontinuities. The negative correlation is due to the polarity change of the reflections with respect to the pilot $P_{PZ}$. The figure shows that PCC and CCGN provide relative traveltimes with respect to the P phase through their correlation maxima and minima.

Depth phases, such as $pP$, $sP$ and their near source multiples ($pmP$, $smP$) are included into $P_{PZ}$ since they are similarly affected by receiver site discontinuities. The inclusion of depth phases aids the detection of receiver structure since the respective lag time for the depth phases conversions and reflections is the same as for the direct P wave. This is in analogy to the teleseismic source function in RF studies. Phases related with the source depth (such as $pP$, $pmP$, etc.) may also correlate with the P phase. However, the use of a pilot with depth phases decreases the correlation with respect to a pilot which consists of only the direct P wave. The final stacking over different events enhances the signals which arrive consistently,
such as near receiver conversions and reflections, and attenuates spurious arrivals and source site reverberations.

Fig. 3 shows an example where the approach has been applied using teleseisms from epicentral distances of 65°–95° registered at the Spanish station CART (Fig. 3a). Blue crosses show the piercing points as obtained from P-to-s conversions at a discontinuity at 510 km depth to give an idea of the sampled TZ by the P410s and P660s phases. Move out corrected radial correlograms are shown in Fig. 3(b). We obtained these correlograms using a P7 of 100 s length, a fixed relative slowness parameter of −0.1 s/km, and a reference distance of 80°. P410s and P660s phases are consistently seen in the CCGNR time-backazimuth section, while the PCCR correlograms show intermittently the same phases with a lower signal-to-noise ratio (SNR). We attribute this difference to the loss of waveform coherence due to noise contamination. The unambiguous detection of P410s and P660s phases is obtained in the relative time-slowness domain, where the reference is the P phase and where the stack is performed using a range of slowness values. This approach is also known as slant-stack and permits the identification of phases in time and slowness. The signals are detected with respect to the P phase which means that the obtained traveltimes and slowness values are relative values with respect to the values of the P phase. The PWS of PCCRs and CCGNRs correlograms from Fig. 3(b) are displayed in Figs 3(c) and (d). These figures show a clear detection of P410s and P660s phases close to the reference values of relative time and slowness, and with a positive coherence value (black contours). For a reference distance of 80°, as the one used in the stacking of Fig. 3(e), the P410s arrives at a relative time of 42.8 s and the P660s at 65.8 s, both with a positive coherence value and a negative relative slowness. Figs 3(e) and (f) show the slant-stacks of PCCZs and CCGNZs for events registered at station CART with epicentral distances between 70° and 120°. These figures show a clear detection of Pp410p at the expected relative time and slowness and with a negative amplitude (red contours). However, there is no evidence for a Pp660p detection. Detection of topside reflections from the 410 and 660 is a more difficult task due to the great variability of these phases caused by the two extra trajectories through the upper mantle and crust. Ppdp phases (where d is the discontinuity depth) are therefore more sensitive to lateral heterogeneities than the P phase which can affect the waveform similarity and the traveltimes.

2.3 Relation with Receiver Function method

Discontinuities beneath seismic stations are commonly studied through the detection of P-to-s conversions in receiver functions (RFs). These are based on deconvolution of the Z component from the R, which is equivalent to spectral division in the frequency domain. This process removes the source wavelet and complexities, while it enhances P-to-s conversions from the receiver site discontinuities (an example of RFs for the station CART is given in Fig. S1 in the supplementary material). The division of small amplitudes (spectral holes), however, makes the spectral division unstable and regularization of the deconvolution is required. A common regularization approach is the water level technique (Clayton & Wiggins 1976) which nevertheless may cause artefacts. Also the presence of high frequency noise is known to impair the deconvolution (Clayton & Wiggins 1976) and often handled through the multiplication of a Gaussian window during the spectral deconvolution or through the application of a low-pass filter.

Our processing approach (presented in Section 2.2) is based on cross-correlation, which is equivalent to spectral multiplication in the frequency domain, and thus, there is no need of regularization. However, as we suppress the amplitude information, we only retrieve the kinematic response of the Earth while RFs retrieve the dynamic response. The RF can be expressed in terms of cross-correlation (Clayton & Wiggins 1976; Ammon 1991; Galetti & Curtis 2012), its analytic expression in the spectral domain is

\[ RF(w) = \frac{Z^*(w) R(w)}{Z^*(w) Z(w)}, \]

where \( Z^* \) is the complex conjugate of Z. The numerator of eq. (4) is the cross-correlation between Z and R in the spectral domain, while the denominator is a positive real number which functions as a frequency dependent normalization factor. In our approach we avoid division and compute only the nominator of eq. (4) when using the classical cross-correlation. Using eq. (4) the classical cross-correlation can be expressed as

\[ Z^* R = [Z^* Z] RF, \]

which shows that our approach (cross-correlation between Z and R) is equivalent to a RF which is multiplied by the auto-correlation of Z, that is, a symmetric function in the time domain, peaked and
2.4 Robustness analysis

In what follows, we present a synthetic data analysis to show the robustness of the correlations and RFs technique (water level deconvolution) towards different levels of noise contamination. This analysis is meant to be a proof of concept test and it is not our intention to compare the synthetic data results with the real data.

2.4.1 Generating test data

We built the synthetic data by computing the Green’s function of the $R$ and $Z$ components with the WKBJ algorithm (Chapman et al. 1988) and using the AK135 velocity model (Kennett et al. 1995). The synthetic seismograms consist of $P$, $PpP$, and $sP$ and their respective $P$-to-$s$ conversions and $P$ reflections at the Moho, 210, 410 and 660 discontinuities. The Green’s functions were computed for three different source depths (0, 20 and 35 km) and epicentral distances every degree from 55 to 65, and convolved with different source functions for each event. The source functions were 10–12 s long and generated from random number sequences filtered in different frequency bands with central frequency at 0.4 Hz (bandwidth...
As a result a data set of 33 events (11 for each source depth) was obtained. The corresponding time-series were contaminated by adding random noise (bandpassed from 0.02 to 1 Hz) of different noise level. We used for each seismogram and noise level 21 different noise realizations to permit a statistical analysis of the robustness of the detections as a function of noise level. The noise level was defined as a percentage of the P phase maximum amplitude in the Z component, ranging from 2 to 30 per cent. This way, theoretical test data was generated for nine different noise levels.

2.4.2 Processing and results

Using the test data we computed for each of the nine noise levels (2–30 per cent) radial correlograms (PCC and CCGN), as explained in Section 2.2. The pilot wavelets were 12 s long and started with the P phase. Further, RFs were computed using a frequency domain deconvolution with a water level regularization of 0.1 to avoid the division with numbers smaller than 10 per cent of the maximum power |Z|₂. The radial correlograms or RFs for each of the 21 noise realizations and each noise level were stacked using PWS with the reference slowness for the P660s phase at 60° distance and corresponding linear time corrections. At the end we obtained 21 stacks for each technique (PCC, CCGN, RF) and noise level from which we determined the traveltime of the P660s phase. The respective mean and standard deviations are shown in Fig. 4(a) relative to the noise free arrival times. The increasing error bars (standard deviations) manifest a higher detection variability due to the increasing noise contamination. It can be seen that for small noise PCC often shows the smallest time variations which we attribute to the high wave-form sensitivity inherent to its phase coherence approach. However, for a large noise contamination the waveforms are more corrupted and PCC traveltime have a larger variation around the mean than CCGN and RF. Nevertheless, the most stable mean traveltimes are obtained with PCC, which are centred at the expected time for all noise levels. It seems that CCGN and RF are differently affected by the large amplitude noise since these approaches are less waveform sensitive than PCC. This is why in Fig. 4 the CCGN and RF mean traveltime at large noise level diverges from the mean value for noise free data. In the RF the presence of noise in the components may breakdown the deconvolution due to amplification of the high frequency components of noise which is controlled by the application of a frequency low-pass filter. Anyhow, noise may destroy signal waveforms at any frequency which destabilizes the deconvolution through the loss of signal coherence. This is independent of the water level which avoids instabilities due to the division of small numbers. Using lower values for the water level (e.g. 0.02) causes mean lag time variations already at lower noise levels.

Fig. 4(b) shows the PWS of PCCRs, CCGNRs and RFs for each noise level and only one of the 21 data sets. This figure shows the decreasing amplitudes (coherence values) obtained for the increasing noise levels. What is more, it can be seen that the maxima are slightly shifted in time, which is manifested in the lag time variability shown in Fig. 4(a). It is visible that the PCC maxima are smaller than the CCGN and RF maxima. This is due to the fact that PCC measures a lower coherence due to its higher sensitivity to waveform perturbations. This is not a problem to our approach since we do not use the absolute values of the coherence.

This robustness analysis demonstrates that with reasonable noise level the performance of PCC, CCGN and RF is similar. Thus, the three independent approaches can be used together to add consistency to the results and to bridge observation gaps due to breakdown of any of the methods inherent to the data characteristics. This study was performed with the P660s phase and random noise without loss of generality. For conversions at other depths the results are expected to be similar for similar signal amplitudes. Besides random noise, signal destruction may also happen due to the interference with other signals which has not been investigated here.

3 Real data processing and results

We employ data from the IberArray seismic network to investigate the 410 and 660 discontinuities beneath South Spain and North Morocco. The data set belongs to the first phase of the IberArray seismic network deployment of the Spanish Topolberia project (Diaz et al. 2009). The map of Fig. 1 shows the location of the 43 stations that were used in this study and which were located on a grid with approximately 50-km station spacing.

In what follows, we describe the real data processing and application of our methodologies to detect P410s and P660s phases beneath Spain and Morocco. Then, we analyse, map and discuss the detected 410 and 660 upper-mantle discontinuities and the TZ thickness (TZE).

3.1 Processing

Our data set consisted of 793 teleseismic earthquakes from 3 yr of continuous recording (from 2007 January to 2009 December), with mb ≥ 4.9 and epicentral distances between 30° and 90°. The epicentral distance range was selected to avoid the 410 and the 660 triplication and the P-wave path through the core–mantle boundary
and D′ region. Due to the sparse distribution of event–station pairs below 70°, we only used event–station pairs with distances between 70° and 90°.

The three component seismograms were rotated into the ZRT coordinate system. The records were filtered in the frequency band 0.03–0.2 Hz and decimated to lower the sample frequency to 10 sps. A STA/LTA algorithm (Withers et al. 1998) was used to perform a quality control and for an approximate first arrival picking on the Z components to define the start time of the pilots. We used an STA window of 10 s and an LTA window of 100 s. Data with \( \{\text{STA/LTA}\}_{\text{max}} \leq 4 \) were rejected, where \( \{\text{STA/LTA}\}_{\text{max}} \) is the maximum amplitude ratio. The number of earthquakes after the quality control reduced to 177 (mostly with \( mb \geq 5.1 \)). Fig. 5(a) shows the distribution of the earthquakes used in this study. The starting time of the pilot was defined as the time where the STA/LTA amplitude reaches 80 per cent of \( \{\text{STA/LTA}\}_{\text{max}} \). As the approaches are based on cross-correlation techniques a very accurate start time of the pilot is not required since it does not affect the position in time of the correlation maxima. Further, a 100 s pilot was extracted automatically from the Z components and the quality of the R components were inspected visually. Then, PCC and CCGN were performed between the pilot and the R component. No quality control was performed over correlograms. However, the quality control over Z and R components is extremely important to avoid processing traces that are composed only of noise.

In addition, RFs were computed for the same data set using the water level deconvolution with a water level parameter of 0.1. A cosine taper filter was used to attenuate artefacts due to discontinuities at the ends of the pilot. Deconvolution was performed in the spectral domain. Two quality controls were applied over these RFs. The first control was based on the SNR which was defined as the rms ratio of the absolute amplitudes in a 15 s window before and after the \( P \) phase arrival. RFs were discarded whenever the SNR ≤ 1.5. The second condition was that the RF must have an amplitude maximum in the vicinity of 0 s. The first pulse in the RF is a spike that corresponds to the first arrival or its P-to-s conversion at a shallow discontinuity. A total of 1171 RFs were obtained. For a direct comparison of the three techniques, only data that pass the receiver functions quality controls were used.

Finally, the cross-correlation functions and receiver functions were stacked using the PWS. Stacking was performed either by using bins of epicentral distance or bins of common piercing point region. Our way of proceeding is explained with more details in the following subsections.

### 3.2 Converted phases at a regional scale

First, we analyse the detected conversions at a regional scale before mapping the discontinuities at a more local scale. For this purpose, we divide the stations into two groups: 23 stations in south Spain and 20 in north Morocco. For each group of stations we compute regional stacks using all correlograms or RFs. Piercing points as obtained from a P-to-s conversion at a 510-km deep discontinuity are shown in Fig. 5(b) to provide an idea of the extension of the sampled TZ.

Figs 6(b), (e) and (h) show the PWS computed for PCCRs, CCGNRFs and RFs at a reference distance of 80° for events recorded at Spanish stations. The blue crosses mark theoretical relative traveltimes and slowness values for the \( P410s \) and \( P660s \) phases in model AK135. As can be seen from this figure the detected mantle conversions arrive close to the reference value of time and slowness with expected positive amplitude for the three techniques. Figs 6(a), (d) and (g) show the corresponding correlations and RFs as a function of epicentral distance. Each trace is the PWS for all events in a 2° distance bin with 1° step interval. Each bin contains between 20–80 traces. The blue lines mark the theoretical traveltimes for \( P410s \) and \( P660s \). It is visible that both phases are detected throughout the entire distance range for all three techniques with positive blue filled amplitude and negative relative slowness, that is, the larger the epicentral distance, the smaller the traveltime difference to reference \( P \).

Similarly, Figs 6(c), (f) and (i) show the PWS for PCCRs, CCGNRFs and RFs for events recorded at Moroccan stations, respectively. These figures show a clear \( P660s \) signal which, however, arrives delayed by about 1–1.5 s for the three methods. Time delays can be caused by discontinuity depression or by seismic low-velocity anomalies along the propagation trajectory in the upper mantle. Furthermore, the figures show two signals with opposite
amplitude polarity nearby the predicted P_{410}s arrival. The signal with positive amplitude is closest to the theoretical P_{410}s and is only clearly seen in the PCCR slant-stack. Thus, it seems that this signal is small in amplitude and mainly detected due to its waveform coherence. The negative amplitude signal is detected with the three techniques although it is less well observed with the RFs. There are two possible explanations for this signal: (1) the positive relative slowness may indicate that this is a reverberation at a subcrustal discontinuity; or (2) this signal corresponds to a P-to-s conversion at a discontinuity atop the 410 with topography or lateral anomalies to explain the observed slowness deviation. The first explanation is less preferred than the second since recent P RF studies did not reveal subcrustal discontinuities beneath Morocco (Mancilla et al. 2012). The second explanation, however, invokes an inclined discontinuity or the presence of some other lateral heterogeneity to deflect the waves such that the incidence angle becomes larger than for a spherical symmetric Earth model in concordance to the observed slowness deviation. Besides, a negative amplitude signal before P_{410}s was also observed in other areas (Revenaugh & Sipkin 1994; Kawamoto et al. 1996; Vinnik & Farra 2002; Vinnik et al. 2010) and it was explained as a P-to-s conversion at a low-velocity layer atop the 410. Recently, it has been proposed that this layer is a worldwide feature which is not associated with a particular tectonic environment (Taudzin et al. 2010).

The traveltime differences for the detected P_{410}s and P_{660}s phases do not exceed 0.3 s for the three different techniques. It means that the 410 and 660 beneath south Iberia and the 660 beneath north Morocco are clearly and robustly detected. Nonetheless, the 410 is only detected with PCC beneath north Morocco, which may indicate that the P_{410}s phases are identified more by their coherence than by their amplitude.

### 3.3 Smaller scale detection and robustness control

Furthermore, we investigate the 410 and 660 discontinuity depths along five north–south profiles. Their locations are shown in Fig. 7.

The traveltime differences for the detected P_{410}s and P_{660}s phases do not exceed 0.3 s for the three different techniques. It means that the 410 and 660 beneath south Iberia and the 660 beneath north Morocco are clearly and robustly detected. Nonetheless, the 410 is only detected with PCC beneath north Morocco, which may indicate that the P_{410}s phases are identified more by their coherence than by their amplitude.

Figure 7. Map with the five profiles used in Figs 8 and 9. Lines indicate the central longitude of each profile: −2.5° (orange), −3.5° (green), −4.5° (red), −5.5° (blue) and −6.5° (dark grey).
Figure 8. Slant-stack examples for the longitude $-5.5^\circ$ (blue profile in Fig. 7). The first three columns show the PWS of PCCs, CCGNs and RFs, respectively. The centre latitude of each bin is specified in the last column. For a reference, theoretical relative traveltime and slowness values are marked with blue crosses for $P_{410}s$ and $P_{660}s$. Yellow crosses mark other phases that are expected to arrive in the same time interval.

We perform the phase detection in the relative time-slowness domain. Fig. 8 shows the slant-stacks for the three techniques and the longitude $-5.5^\circ$. The first column shows PCC results, the second and the third column are for CCGN and RF, respectively. The centre of each bin latitude is specified in the last column. For many bins clear $P_{410}s$ and $P_{660}s$ phases can be identified. However, some of the stacks are noisy, with other signals which also stacked coherently. In order to assess the robustness of the $P_{410}s$ and $P_{660}s$ phases we perform a bootstrap resampling of 21 repetitions for each

stacked for groups of common piercing points. Therefore, we define common conversion point areas using bins of $2^\circ$ width in latitude and longitude and centred with $1^\circ$ step interval along each profile. We perform the phase detection in the relative time-slowness domain. Fig. 8 shows the slant-stacks for the three techniques and the longitude $-5.5^\circ$. The first column shows PCC results, the second and the third column are for CCGN and RF, respectively. The centre of each bin latitude is specified in the last column. For many bins clear $P_{410}s$ and $P_{660}s$ phases can be identified. However, some of the stacks are noisy, with other signals which also stacked coherently. In order to assess the robustness of the $P_{410}s$ and $P_{660}s$ phases we perform a bootstrap resampling of 21 repetitions for each
piercing point bin. This way, time standard deviation values are obtained to measure variations in the observed relative traveltimes. On average, time standard deviations for PCC, CCGN and RF are 0.29, 0.28 and 0.23 s, respectively. Finally, to convert the estimated traveltimes to depth, the detected phases should pass two quality controls. First, the amplitude of the signal in the stack should be larger than twice the mean absolute amplitude of the stack in the time interval 30–80 s (and arbitrary slowness range -4 s/° to 4 s/°) and second, the standard deviation values should be smaller than 0.7 s. Whenever both quality controls are satisfied, we perform a visual control over the time-slowness plots. We discard the data when a clear identification of phases is not possible.

3.3.1 Time-to-depth conversions

Absolute 410 and 660 depths are conditioned by the velocity model used to perform the depth conversion of the estimated mean relative traveltimes. Thus, in order to convert relative traveltimes to discontinuity depths accurate \( v_p \) and \( v_s \) models are needed for the upper mantle beneath the studied area (the lower mantle anomalies do not affect the depth estimates due to the reference phase with common wave path below the discontinuity). Seismic tomography models can be used to estimate time corrections, which, however, should be used with caution. Blurred not well localized anomalies and inversion artefacts will introduce errors to the traveltimes corrections. Furthermore, seismic velocity anomalies are often underestimated inherent to the regularization of tomographic inversions.

We use the \( P \)-wave tomography model by Villaseñor et al. (2003) to account for seismic velocity anomalies and to correct the estimated relative times before depth conversion. For this purpose we compute an average 1-D \( P \)-velocity profile for each bin. Corresponding \( S \)-velocity profiles are derived from the \( P \)-velocity anomalies, \( \delta v_p \), by employing a constant factor \( \delta v_p / \delta v_s = 1.5 \). This factor typically ranges between 1.5 and 2 in the upper mantle (Ritsema & Van Heijst 2002) depending on the type of anomaly (thermal and/or compositional). The scaling of localized \( P \)-wave or \( S \)-wave velocity models is often used to perform \( RF \) time corrections (e.g. Duerck & Sheehan 1997; Li et al. 2002; van der Meijde et al. 2005). We obtain time residuals (between the traveltimes predicted by the local velocity model and AK135) in the order of \(-0.6 \) to 0.8 s for the P410s and P660s phases, which are translated to depth corrections of about \(-6 \) to 8 km. Changing the applied constant velocity perturbation ratio \( \delta v_p / \delta v_s = 1.5 \) to 2 leads to high-end depth corrections for the \( P \)-to-\( s \) conversions which are in the order of \(-9 \) to 12 km. A more accurate depth correction is not performed due to the not yet totally resolved upper-mantle velocity structure beneath Spain and Morocco.

3.3.2 410 and 660 depths

The 410 and 660 depth values obtained with the three techniques along the five profiles (in Fig. 7) are depicted in Figs 9(a), (b), (c), (d) and (e) for the longitudes \(-2.5°, -3.5°, -4.5°, -5.5° \) and \(-6.5° \), respectively. Different symbols are used to discriminate the results from the different techniques: PCC (circles), CCGN (triangles) and RF (squares). Thin black lines at 410 and 660 km indicate the 410 and 660 nominal depths, respectively. Some of the profiles show discrepancies in the estimated 410 and 660 depth values. These differences are expected for real data where the small amplitude coda phases are often obscured by other signals and noise, and where signals are less coherent due to the presence of structural heterogeneities. As a consequence, the observed differences can be attributed to the different strategies of the employed methods which differently handle the detection of less coherent signals. Nevertheless, it can be seen from Fig. 9 that these discrepancies are usually small and that the detection of the three methods follow a clear trend. It is also appreciated that at some places only one of the methods provides a detection which helps to bridge observation gaps.

Differences in the traveltime measurements of the coda phases indicate data complexities and also indicate that at least one of the approaches is less suited to accurately detect the signals due to the present signal and noise characteristics for this detection. It is often not possible to determine which of the approaches provide the better time measurement and we therefore understand the variations in the time measurements as an indicator of inconsistencies due to data complexity. That is, the smaller the traveltime variations among the three methods are, the more confidence we get into a correct coda phase detection. What is more, the mean depth can stabilize against systematic and non-systematic errors. Note that a small standard deviation of traveltime (depth) measurements obtained for one single approach can not reveal systematic time errors due to some strange noise. Therefore, we merge the independent \( Pds \) \((d = 410 \) or 660\)) detections from each bootstrap repetition for each of the three techniques. This way, new mean and standard deviation values are obtained for the 410 and 660 depths at each latitude-longitude bin. The results are displayed in Figs 10(a) and (b). Opposite colour codes are used to represent the 410 and 660 topography and the symbol size is proportional to the depth standard deviation. Additionally, the different symbols show the techniques that contribute to estimate the mean depth value and standard deviation at each location. The mean depth estimations are interpolated and plotted in Figs 10(c) and (d). The interpolation is performed by adjusting a continuous curvature surface to the corresponding discontinuity depth values in Figs 10(a) and (b). The green stars mark the bins which contribute to the interpolated map.

The P660s phases are consistently detected in the five profiles along the whole range of latitudes, which permit to map the 660 topography with deflections of about \(-10 \) to 30 km, that is, the 660 discontinuity is locally about \(-10 \) to 30 km deeper than global average values (Shearer 1991; Revenaugh & Jordan 1991; Vidale & Benz 1992; Shearer 1993; Lawrence & Shearer 2006). This deflection presents a clear trend which is related to the geographical boundary. In general, we find that the 660 is deeper beneath Morocco and is close to global averages beneath Spain. This is clearly seen in the profiles which cross the western part of the Alboran Sea, the Strait of Gibraltar and the Gulf of Cadiz. Another interesting result is the downward deflection of the 660 in the Alboran Sea. The easternmost profiles at \(-2.5°, -3.5° \) and \(-4.5° \)(Figs 9a, b and c and 10d) also show a downward deflection of the 660, with maximum depth of 680–690 km at latitudes 35°–36°, below the Alboran Sea. This region coincides with the area where the tomographic images demonstrate the presence of a positive \( P \)-wave velocity anomaly down to the base of the upper mantle.

From Fig. 10(c) it is clear that the topography of the 410 is smoother than for the 660, however, the 410 is less well constrained, especially in the profiles that cross the Alboran Sea (at \(-2.5°, -3.5° \) and \(-4.5° \)). In this region, at latitudes between 34° and 37°, a P410s detection gap is observed (Figs 9a, b and c, and 10a). The origin of the gap is either due to the lack of P410s detections, or due to detections of incoherent low-amplitude signals which do not satisfy the employed quality criteria. Most of the clear P410s detections are around global averages. Maximum 410 depth reflections from the reference value are in general less than 10 km. The 410 is deeper
Figure 9. Discontinuity depths along five north–south profiles (Fig. 7) at longitudes: (a) −2.5°, (b) −3.5°, (c) −4.5°, (d) −5.5° and (e) −6.5°. Circles, triangles and squares mark the obtained discontinuity depths for PCC, CCGN and RFs, respectively. Thin lines indicate the nominal depths of 410 and 660 km for the discontinuities.

beneath eastern Morocco (Figs 9a, b, and 10c), while it seems to be slightly deflected to shallow depths beneath western Morocco. Below Spain, the 410 appears near the expected nominal depth of 410 km.

Our 410 and 660 depth values indicate a TZT which is in general close to global average beneath south Spain (about 240–250 km) (Flanagan & Shearer 1998; Chevrot et al. 1999; Lawrence & Shearer 2006) and thicker beneath the southwest corner of the sampled Moroccan region (250–275 km). The TZT in the Alboran Sea is unknown due to the 410 detection gap. The results for the TZT are summarized in Fig. 11.

4 DISCUSSION

4.1 Processing approach

We have used a new processing approach based on cross-correlation and stacking techniques to detect weak amplitude phases that arrive in the P-wave coda. As cross-correlation functional we have used the CCGN which is the classical approach to measure waveform similarity and the PCC which is based on instantaneous phase coherence and which has not been widely explored before (Schimmel et al. 2011). We have proposed to use these cross-correlations (PCC and CCGN) together with RFs to stabilize the detections against errors and to bridge observation gaps. Without noise or with low to reasonable noise level the results of the three techniques are similar but inherent to their different strategies the signals are detected differently. Similar results obtained through different methods add robustness and confidence to the detections and interpretations. Varying results or non-detections are expected for more difficult data, depending on the signal and noise characteristics. Therefore, the variations of the final results and the amount of approaches which lead to an independent detection can be used as a quality indicator. Our procedure is a step forward to stabilize against detection problems and to identify more ambiguous detections. Another advantage is that one automatically bridges observation gaps by one or the other method. Each difficult detection is a special case due to the non-stationarity of signals and noise and none of the methods will lead satisfying detections for all cases. Furthermore, the increasing volume of data makes it less feasible to manually investigate the measurements.

PCC is amplitude unbiased and more sensitive to waveform coherence than CCGN and RF. Coherent signals are therefore accurately detected even in the vicinity of other larger amplitude signals which may bias the detections with CCGN (figs 3 and 4 in
410-km and 660-km discontinuities

Figure 10. Topography of TZ discontinuities. (a) Estimated 410-km discontinuity depth from the detection of $P_{410}$ phases using PCC, CCGN and RF. Symbol sizes are proportional to the standard deviation values associated with each mean depth. Different symbols mark the different approaches: circles for PCC, triangles for CCGN and squares for RF. The grey box denotes the minimum and maximum standard deviation. (b) Same as (a) but for the 660 km discontinuity. (c) 410 topography obtained by adjusting a continuous curvature surface to the mean 410 depth estimations in (a). (d) 660 topography obtained by adjusting a continuous curvature surface to the mean 660 depth estimations in (b). Green stars mark bins with estimated depth values.

Figure 11. Transition zone thickness along the five north-south profiles (Fig. 7) at longitudes: $-2.5^\circ$ (orange), $-3.5^\circ$ (green), $-4.5^\circ$ (red), $-5.5^\circ$ (blue) and $-6.5^\circ$ (grey).

Schimmel 1999, fig. 1 in Schimmel et al. 2011). Our synthetic data tests (Fig. 4) have shown that PCC may even provide more stable results than CCGN and RF at high-noise levels. Of course, PCC fails when signals cannot be detected by their coherence. In these cases CCGN is usually the better approach. In practice, for our real data we see that the detection with PCC is a more difficult task, which is attributed to the loss of waveform coherence with the reference phase. This can be deduced from the small number of circles (PCC detections) in Figs 10(a) and (b), compared to the other symbols. Nonetheless, it is also seen that in some occasions PCC is the only approach which provides detections and therefore bridges detection gaps.

Cross-correlations have been used before to detect upper mantle conversions (Paulssen 1985, 1988; van der Lee et al. 1994; Schimmel 1999). However, while we use a 100-s pilot wavelet, these studies use a shorter one that comprises only the $P$ phase or part of it. In Paulssen (1985, 1988) and van der Lee et al. (1994) the pilots are about 5–6 s long. Besides, they normalize the correlograms with respect to the autocorrelation of the $P$ phase, and linear stacks are employed to identify the signals. Shearer (1991) uses the cross-correlation between components to detect $P$-to-$s$ conversions.
and other phases using a global data set. For the $P_{ds}$ phases a 41-s reference wavelet is used and correlation peaks are plotted without previous stacking. Similar to this study, our approach can be used to detect other phases such as $S_{ds}$, $P_{dp}$, $S_{dp}$, etc. with an adequate definition of the pilot.

### 4.2 TZ discontinuities

We have used a vast volume of data thanks to the IberArray of the Topolberia project (Diaz et al. 2009). This, has permitted us to study in detail and with high resolution a previously undersampled portion of the upper-mantle TZ discontinuities beneath south Spain and north Morocco. Our results locate the 410 and 660 discontinuity within the expected depth range as obtained by global studies (Revenaugh & Jordan 1991; Shearer 1991; Vidale & Benz 1992; Shearer 1993; Lawrence & Shearer 2006), and are further in good agreement with a large scale $P$ RF study that covered the entire Mediterranean region (van der Meijde et al. 2005). In van der Meijde et al. (2005) two stations are located in central Spain and north Morocco which show the same TZT variations as observed in our study. We localize the observed thickening of the TZ towards Africa as a narrow and steep transition between the two continents. This transition is not seen by Dündar et al. (2011) who studied the 410 and 660 discontinuities beneath the Alboran Sea and close surroundings using $P$ RFs and available stations before the IberArray deployment. Their RFs are stacked for two piercing point areas, the Alboran Sea and surrounding areas (south Spain, north Morocco, Gulf of Cadiz), respectively, and show a good agreement of the conversions from the different RF groups, within 1 s of the theoretically expected arrivals. Their results do not contradict the presence of topography which likely is not seen due to the averaging in the large piercing point area. Our analysis shows systematic variations for independent data from many smaller piercing point areas and for the different techniques. Also our regional slant stacks for the Spanish and Moroccan data Fig. 6 show a clear traveltime delay of the $P_{660s}$ phases beneath Morocco.

The maximum observed TZT beneath Morocco (275 km) is about 25–35 km thicker than global averages of 240–250 km (Flanagan & Shearer 1998; Chevrot et al. 1999; Lawrence & Shearer 2006), which is within the expected TZT variations for global studies (maximum TZT perturbations are about 35 km in active subduction environments). Under the assumptions of a 250-km TZ thickness for reference, a pyroelastic mantle composition, and Clapeyron slopes of 4 MPa K$^{-1}$ for the olivine–wadsleyite transition (410) (Katsura et al. 2004) and $-1.3$ MPa K$^{-1}$ for the post-spinel transition (660) (Katsura et al. 2003), the TZ thickening of 25 km beneath Morocco can be translated into an approximate temperature decrease of 160 K (Helffrich 2000). A 10 per cent change of the Clapeyron slopes leads to a variation of about 10 K. Tomographic images (e.g. Wortel & Spakman 2000; Villaseñor et al. 2003; Hansen et al. 2012), so far, have not revealed a positive velocity anomaly in relation to such a hypothetical thermal anomaly. The thicker TZ beneath the southwest corner of the Moroccan region is mostly due to a deeper 660 discontinuity which shifts downwards by as much as 20 km while the 410 shows less variations and stays on average at its expected nominal depth of 410 km. That is, the 410 discontinuity shows no similar anticorrelated depth deflections as expected for a large cold anomaly which affects the entire TZ.

Alternatively, the observed traveltime difference $t_{P_{660s}} - t_{P_{410s}}$, which determines the TZT, may be altered by the presence of an isothermal negative velocity anomaly inside the TZ. If this is the situation beneath Morocco, then we would expect a time shift only for the $P_{660s}$ phase, as it is observed in our data. Beneath Morocco, at latitude $-5.5^\circ$ and longitude $34^\circ$, we estimate an average value of 25.4 s for $t_{P_{660s}} - t_{P_{410s}}$, which is 2.4 s larger than the reference value (23 s for a reference distance of 80 '). The time corrections that we compute from the tomographic models (Villaseñor et al. 2003) only explain 0.2 s of the time deflection from the reference value. At TZ depths (400–700 km) the $P$-wave tomographic model by Villaseñor et al. (2003) shows negative velocity anomalies (with respect to AK135) of about $-0.7$ to 0.4 per cent (with a mean value of $-0.2$ per cent) beneath Morocco, which could be underestimated due to missing resolution and seismic velocity anomaly smearing. A not considered negative velocity anomaly of 5–6 per cent is needed to explain the corrected traveltime difference $t_{P_{660s}} - t_{P_{410s}}$. We can not exclude an overestimation of the 660 discontinuity depth due to underestimated time corrections. Nevertheless, realistic velocity anomalies can only explain part of the 20–30 km deflection.

If, however, a thermal anomaly rises the TZ temperature and if other components are present in the mantle composition, then another hypothesis may better explain the observations. In hot mantle anomalies and in the presence of aluminum within garnet, the garnet to perovskite transition is expected near the 660 with a positive Clapeyron slope (Wang et al. 2004). As a consequence of the positive Clapeyron slope, the post-garnet related 660 discontinuity deflects downwards in a hotter mantle. This transition is observed in different tectonic environments (e.g. Thomas & Billen 2009; Cao et al. 2011). Up to our knowledge, no strong temperature anomalies have been revealed within the TZ beneath Morocco and right now more constraints are needed to support one of the hypothesis. The increasing seismic station coverage and data volume will...
better constrain the mantle structure and finally help to understand the observed differences in the TZ between south Spain and north Morocco.

The position of the 660 depressions beneath the Alboran Sea is consistent with the positive P-wave velocity anomaly due to the Alboran slab (Wortel & Spakman 2000; Piromallo & Morelli 2003; Villaseñor et al. 2003). Fig. 12 shows a cross-section with the 410 and 660 topography along the longitudes $-3.5^\circ$ and $-4.5^\circ$ with a section of the tomography model to demonstrate the relation of the 660 depression with the high-velocity anomaly. The cold slab material cools the 660 and shifts the post-spinel transition to greater depth. In our study we observe a maximum depression in the order of 25 km (Fig. 10d). 660 deflections typically range between 10 to 30 km in subduction related areas (Helffrich 2000). Interpreting the maximum 25 km deflection in terms of temperatures implies a minimum thermal anomaly of $-640$ K for the slab at 660 km depth. However, part of the 660 discontinuity depression could be due to a high water content in ringwoodite (Litasov et al. 2005). Thus, the temperature anomaly of $-640$ K should be considered as a lower bound. Fig. 12 demonstrates that the localized depression of the 660 includes an area of isolated deep earthquakes beneath Granada ($37^\circ$ N $3.5^\circ$ E approximately) with similar focal mechanism, magnitudes ($M_0$) from 4 to 7, and a recurrence time of about 20 yr (Buforn et al. 2004, 2011; Bezada & Humphreys 2012). The occurrence of these events is not yet understood but may indicate that the anomaly is cold enough to produce earthquakes.

The 410 has been less well observed in the Alboran Sea area which leads to an observation gap (Fig. 10a). A detection gap of converted or reflected energy from the 410 in subduction zones has been reported in other studies, for example, for near source conversions (Collier & Helffrich 2001; Titi & Wiens 2005), $\alpha$CS precursors (Tono et al. 2005) and SS precursors (Conteniti et al. 2012). Fig. 12 shows that the 410 gap coincides with the position of the velocity anomaly of the tomographic image. We believe that structural complexities cause the 410 observation gap due to defocusing or loss of coherence and destructive interference with other coda waves. Furthermore, the seismic visibility of the 410 is often decreased in subduction areas owing to its thermodynamic properties which may lead to smaller conversions/reflections in colder than normal mantle and additional discontinuity broadening in the presence of water (Collier & Helffrich 2001).

5 CONCLUSIONS

We have shown that our approach is useful to detect coda signals which mostly are weak and hidden by a multitude of other signals. Our analysis is suited to process large data volumes and has been focused on the detection of P-to-s conversions from the upper-mantle TZ discontinuities beneath north Morocco and south Spain. The joint usage of three independent techniques (PCC, CCGN and RF) permits us to stabilize the detections against measurement errors, to use the measurement variability as robustness indicator and to bridge observations gaps. We thus achieve consistency and robustness on the estimated depth values of the 410 and 660 discontinuities. None of the three methods is considered to be the best and the different detection strategies determine their performance depending on data characteristics.

We have used our approach to map the 410 and 660 discontinuity beneath north Morocco and south Spain and we have found a thickening of the TZT at the Iberia–Africa Plate boundary. The TZT is around global average beneath south Spain (240–250 km) and thicker beneath the southwest corner of the sampled Moroccan region (250–275 km), which is mainly due to a depressed 660, while the 410 topography shows less variations. The explanation for the TZ thickening beneath southwest Morocco is ambiguous. A plausible cause can be a not considered isothermal velocity anomaly, which increases the P660s traveltime and which leads to an apparent TZT, thicker than expected. Alternatively, the thickening of the TZ may be explained by an Al rich Garnet related 660 discontinuity, which in hotter environments is expected at greater depth than the post-spinel transition.

The decreased visibility of the P410s phase in the Alboran Sea does not allow us to retrieve TZT values for this region. The observation gap is possibly due to structural complexities and the thermodynamic properties of the olivine to wadsleyite phase change in colder than normal mantle. In this region the 660 is observed depressed to larger depth by up to 25 km, which suggests that the Alboran Sea heterogeneity, demonstrated in tomography images, is still cold enough to induce downward deflection of the post-spinel transformation. This may further constrain the origin of the deep earthquakes beneath Granada.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article:

Figure S1. Real data example for the Spanish station CART. (a) Move out corrected radial RFs sorted by their backazimuths. White lines indicate P410s and P660s theoretical relative travel-times. Note that P410s and P660s are consistently seen in contiguous tracks for certain backazimuths. (b) PWS of RFs shown in (a). Yellow crosses mark the expected P410s and P660s conversions using model AK135. Normalized amplitudes larger than 0.5 and smaller than −0.5 are contoured in black and red, respectively. Note the clear detection of the P410s and P660s phases. (http://gji.oxfordjournals.org/lookup/supp1/doi:10.1093/gji/ggt129/-/DC1)

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