Crustal structure of the central Betics associated with a STEP fault from reverse time migration of teleseismic converted phases.

A. Molina-Aguilera¹,², X. Shang³, M. V. de Hoop⁴, FdL Mancilla¹,², R. D. van der Hilst⁵, J. Morales¹,², D. Stich¹,², B. Heit⁶

¹Instituto Andaluz de Geofísica, Universidad de Granada, Campus de Cartuja, 18071, Granada, Spain
²Departamento de Física Teórica y del Cosmos, Facultad de Ciencias, Universidad de Granada, Campus de Fuente Nueva, 18071, Granada, Spain
³Shell Oil Company, Houston TX 77079, USA.
⁴Department of Computational and Applied Mathematics and Department of Earth Science, Rice University, Houston TX 77005, USA
⁵Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts, USA.
⁶Deutsches GeoForschungsZentrum, GFZ, Telegrafenberg, 14473, Potsdam, Germany

Corresponding author: Flor de Lis Mancilla (florlis@ugr.es)

Key Points:

- Reverse Time Migration of teleseismic converted phases reveals a sequence of Moho offsets along a high-density seismic array, Central Betics
- A distinct block of Iberian crust adsorbs most of the collision and lateral roll-back tearing deformation
- A STEP fault of ~15 km vertical Moho offset occurs through an inherited weakness zone in the continental-transitional Iberian crust
Abstract

Reverse Time Migration (RTM) of teleseismic P-to-S-converted phases reveals a sequence of abrupt Moho offsets under the Central Betics, at the northern edge of the Westernmost Mediterranean subduction region. RTM exploits the entire seismic wavefield and, with sufficiently dense sampling it produces high-resolution images of the crust. We used data recorded at two high-density seismic arrays (with inter-station distance of ~2 km and ~10 km) perpendicular to the collisional front. The RTM image shows substantial Moho topography, including vertical offsets and variations in slope. The Moho offsets define a distinct block of Iberian crust in the collision area between the Iberian foreland and the Alboran domain on the overriding plate. Its southern boundary is formed by a STEP (Subduction-Transform-Edge-Propagator) fault with ~15km vertical Moho offset. This block may be formed by reactivation of an existing weak zone in the continental-to-transitional Iberian crust.

Plain Language Summary

The westernmost Mediterranean region has undergone a complex tectonic evolution where slow WNW-ESE oblique convergence of Iberian and African plates coexists with a rapid westward rollback of a subducting slab. High-resolution images of the crustal structure are essential to understand its tectonic evolution and how the deformation produced by these processes is distributed among the involved geologic domains. To obtain images of crustal structure, we use the wavefields from large, distant earthquakes recorded at a dense seismic station deployment in Southern Spain, and propagate them back into a model of the Earth’s interior. This approach shows where the interactions between different types of seismic phases took place, and thereby, images the discontinuities in the crust. We observe a particularly complex topography of the discontinuity between Earth’s crust and mantle, showing several vertical offsets and variations in slope between distinct crustal blocks in the collision area. The most prominent feature is an abrupt, ~15km high, vertical offset identified with the fault along which the subducting slab detached from the continental crust during rollback.
With this aim a high-density seismic array were deployed in central Betics located in the northern edge of the interest region. To exploit the entire seismic wavefield of teleseismic P-to-S seismic phases we apply a Reverse Time Migration (RTM) technique. The RTM image shows substantial topography in the Crustal-Mantle discontinuity (Moho discontinuity), including vertical offsets and variations in slope. These Moho offsets define a distinct block of Iberian crust in the collision area between the Iberian foreland and the Alboran domain on the overriding plate. Its southern boundary is formed by a STEP (Subduction-Transform-Edge-Propagator) fault with ~15km vertical Moho offset. This block may be formed by reactivation of an existing weak zone in the continental-to-transitional Iberian crust.

To obtain images of crustal structure, we use wavefields from large, distant earthquakes recorded at a dense seismic station deployment in Southern Spain, and play them back into a model of the Earth’s interior. This approach shows where the interactions between different types of seismic waves took place, and thereby depicts heterogeneity in Earth. We observe a particularly complex topography of the limit between Earth’s crust and mantle, showing several offsets and variations in slope between distinct crustal blocks in the collision area. The most prominent feature is an abrupt, ~15km high vertical offset, identified with the fault along which the subducting slab detached from the crust during rollback.

1 Introduction

The Central Betics are located at the northern edge of the westernmost Mediterranean region. Since Miocene times, its tectonic evolution has been characterized by the interplay between slow convergence of Iberia and Africa (with current WNW-ESE oblique direction at a rate of ~4-5 mm/yr, e.g. Nocquet and Calais, 2004), as well as rapid slab rollback process in approximately perpendicular direction associated with the western Mediterranean subduction system (Fig. 1A, e.g. Lonergan and White, 1997; Rosenbaum 2002; Faccenna et al., 2004; Spakman and Wortel 2004; Jolivet et al 2009; Hinsbergen et al 2014, Palomeras et al. 2014). This subduction system is currently in its last evolution stage with the subduction process is no longer active (e.g. Stich et al 2006, Chertova et al. 2014, Spakman et al. 2018). Slab tearing along the subduction edges, where continental crust started to be involved, was expected (e.g. Faccenna et al., 2004; Duggen
et al. 2005; Govers and Wortel, 2005; Garcia-Castellanos & Villaseñor, 2011) and observed as a Moho offset perpendicular to the tearing direction (red line in fig. 1A, Mancilla et al. 2015a,b).

At the edges of a subduction system, lateral tearing of the subducting slab occurs through STEP faults (Subduction-Transform-Edge-Propagator, Govers and Wortel, 2005) as a necessary geometrical element to enable the advance of the subduction/roll-back. While STEP faults at the subducting slab have been observed and described largely during the last years (e.g. Govers and Wortel, 2005; Pierce et al 2012; Agostinetti et al, 2015), the pattern of deformation produced around them is still poorly understood (e.g. Baes et al., 2011; Özbakir et al, 2013). High-resolution images of crustal structures associated with a STEP fault will improve our understanding on how the deformation is distributed in the subducting and overriding plates (Özbakir et al, 2013, Mancilla et al. 2018).
Figure 1. A) Topography map with the boundary of the main tectonic units of the Betic domain and foreland. The continuous red line marks the edge of the Iberian crust (slab tear fault) imaged by the P-receiver functions (Mancilla et al. 2015a) and the red dashed line indicates the presumably current tip position of the tear fault (Heit et al. 2017). The purple transparent area delimits the positive velocity anomaly at ~75 km depth described in the tomography study of Bezada et al. (2013) associated with the oceanic Alboran slab. Red triangles mark the first leg (~2 km inter-station distance, HIRE-I) and blue triangles the second leg (~10 km inter-station distance, HIRE-II). B) Geologic map including the different geologic terrains. The black rectangle defines a distinct deformed block observed in this study.

Dense, north-south trending profiles of seismic broadband stations were deployed across the central Betics (triangles in Fig. 1, Heit et al. 2010) to study the crustal structure of the different geological units involved in the collision front in southern Iberia. This seismic profile probes two
main tectonic units involved in this collision (Fig. 1B): the Alboran domain (overriding plate) and the Iberian domain (Iberia foreland). The Iberian domain (Iberian Massif) belongs to the European Variscan orogeny and forms a major part of the Iberian Peninsula. Its crustal structure in the study area is rather homogeneous with a flat Moho discontinuity (e.g. Mancilla et al., 2015b). The Alboran Domain (comprising the so-called Internal Zones) is composed mainly of Paleozoic and Mesozoic meta-sedimentary rocks with varying metamorphic being divided into the Nevado-Filabride (IZNV), Alpujarride (IZAL) and Malaguide complexes (e.g. Balanyá and García-Dueñas, 1987). Only the first two complexes are observed at the surface along the profile. The External Zones, located between the Alboran and Iberian domains, consist of the Mesozoic to Miocene sedimentary rocks that were deposited onto the rifted Iberian paleomargin prior to the collision and formation of the Betics orogen (Fig 1B, Prebetic and Subbetic units, Garcia-Hernandez et al. 1980).

Common conversion point stacking (CCP) images of P-wave receiver functions from this seismic array provided first indications of the complex crustal architecture of this oblique collision including a sharp Moho step of ~17 km below the northern edge of the Alboran domain (Mancilla et al. 2018). The assumption that P-to-S conversions occur at (locally) horizontal interfaces limits the lateral resolution of CCP stacks. The inferred structural complexity and the availability of waveform data from dense arrays in this area motivate the use of full waveform techniques. In this study we consider Reverse Time Migration (RTM) of teleseismic-converted phases (Shang et al. 2012). With synthetic and real data, Shang et al. (2012, 2017) have shown that with sufficiently dense sampling, RTM of scattered wavefields performs better than migration of CCP stacks of traditional receiver functions. We use RTM of teleseismic-converted phases to improve the resolution and clarity of images of the crustal structure across the central Betics, and in particular the small-scale variations of the Moho topography.

2 Data and Methods

Reverse time migration (RTM) of teleseismic P-to-S converted phases recorded at dense seismograph arrays has been developed for high resolution imaging of crustal and upper mantle discontinuities (Shang et al. 2012, 2017). This wave equation prestack depth migration technique is based on direct back propagation of elastic waves and circumvents simplifying assumptions
such as the locally flat interfaces that form the base for common conversion point stacking (CCP). RTM accounts for single scattering processes in wave propagation, including diffractions, reflections and refractions. Subsurface imaging with time reversal techniques typically involves correlations between the adjoint wavefield and a synthetic forward wavefield (e.g. Tromp et al., 2005; de Hoop et al., 2006; Stich et al., 2009), which is impractical for the teleseismic short period waves that form the base of receiver-functions. Instead, vertical and radial components of the waveform section are back-propagated into the medium, either with the P-wave or the S-wave speed. Due to the difference in P and S wavespeeds, transmitted and converted wave fronts will reach the same location only at the time of phase conversion. This condition yields the conversion point, the location of which is found by integrating for all time steps a cross-correlation function between P and SV wavefields (Brytik et al., 2012, Shang et al. 2012).

Dense profiles of seismic stations are needed to exploit the imaging capabilities from waveform coherence among consecutive receivers. Wave equation migration methods require a restrictive condition on regular and high-density spatial sampling (Liu and Sacchi, 2004; Xu et al., 2005). Then, we resample the recorded wavefield on a fine regular grid and interpolate it where the original data set is too sparse. Synthetic tests with interpolated wavefields have shown that RTM can reconstruct dipping and vertically offset interfaces even in the presence of complex wave phenomena (Shang et al. 2012, 2017)

2.1 Data

We process data from teleseismic seismograms recorded at the 50 seismic broadband stations array deployed across the Sierra Nevada mountain in central Betics depicted in Fig. 1 (Heit et al., 2010). The stations were deployed in two legs operating in different periods of time: the first leg, recording for 14 months, consists of 40 broad-band seismic stations with average inter-station distance of ~2 km, amounting to ~80 km length (red triangles, Fig 1); the second leg which recorded for 22 months includes 10 stations (blue triangles, Fig 1). The inter-station distance in this second leg is ~10 km. The total length of the profile is ~180 km.

We consider earthquakes with magnitude larger than 5.5 in the distance range between 30° and 90° and with signal-to-noise ratio larger than 2. We include in our analysis events recorded by a
large majority of the stations (< 70%), to warrant a dense enough receiver sampling that is
prerequisite for wavefield interaction in RTM. Because of the different recording time of the two
legs the analysis of their data is done separately. We retain for the last stacking step 50
earthquakes for the first leg, and 39 events for the second leg both with good coverage (Fig. 1B-
C). The back-azimuth and incidence angle distribution guarantees that the wavefield impinges
the target region from different directions providing a good illumination of the structures (Shang
et al. 2012).

2.2 Processing

We follow the RTM workflow described in Shang et al. (2017), consisting of (1) extraction of
the signature of the source time function, (2) wavefield regularization, (3) deconvolution of the
source signature, (4) backward propagation of the wavefield, (5) application of the imaging
condition, and (6) stacking of images from individual events. To estimate the source signature,
we extract coherent features of the vertical component waveforms across the seismograph array
through Principal Component Analysis (PCA, Rondenay et al., 2005), using the open-source
software package Crazyseismic (Yu et al., 2017). After visual inspection of the seismograms in
the frequency range from 1 Hz to 0.03 Hz, we eliminate noisy and/or ringing traces that can
perturb both the estimation of the source signature with PCA and the interpolation of the
wavefields. We use a time window (-10, 30) s around the theoretical P onset to ensure the
inclusion of the Moho converted phases (Pms). After alignment, we retain the first principal
component, corresponding to the largest singular value obtained in PCA, to estimate the coherent
common source signature in all vertical seismograms (Rondenay et al., 2005).

Wavefield regularization involves interpolation of the radial and vertical components and
resampling the recorded wavefield on a sufficiently fine regular grid, in order to avoid aliasing
and meet the homogeneous sampling requirement. To perform the wavefield reconstruction we
interpolate in the curvelet domain (e.g. Herrmann and Hennenfent 2008; Naghizadeh and Sacchi
2010). The algorithm minimizes a cost function and finds the best trade-off between the data
misfit and the $l^1$-norm of the solution with a Basis pursuit approach (see Shang et al. 2017 for
details). In order to avoid artifacts in the interpolated images, we balance temporal and spatial
resolution. The temporal resolution is controlled by the sampling and filtering of the traces, and
the spatial resolution by the interstation distance. A threshold of the interpolation distance is
determined by the resolution in terms of the Rayleigh diffraction limit (that is, $\lambda/4$, where $\lambda$ is the
incident wave wavelength) and the interstation distance to avoid spatial aliasing (Shang et al.
2012). Stable images of the interpolated vertical and radial wavefields (Fig. 2a) are obtained after
bandpass filtering the signal between 0.8 and 20 s and using an interpolation distance of 500 m
and 1000 m for the first and second legs, respectively. We penalize wave packets associated with
near-horizontal wave propagation using a mask matrix in the curvelet domain with threshold for
apparent S velocity of 4 km/s.

The deconvolution of the source signature then yields the estimated radial and vertical Green’s
functions (Fig. 2b). We use iterative time-domain deconvolution (Ligorria and Ammon, 1999)
with an approximate pulse width of 1 s ($a = 2.5$) and apply a reverse time continuation to back
propagate the estimated Green’s functions for each earthquake. The snapshots of the elastic wave
field are reconstructed from the recorded multicomponent array data, solving the elastic wave
equation in the time domain using a staggered grid finite difference scheme (Virieux, 1986). For
each time step P and SV modes are separated by polarization decomposition (Brytik et al., 2011,
Shang et al. 2012). In order to save computational cost, we restrict the model domain to 92 x 80
km (along the profile x depth) and 125x80 km, for the first and second legs, respectively. We use
a total backward propagation time of 52 s to guarantee that the P wavefront reaches the
boundaries of the numerical domain. We use a 1D earth model for propagating the wavefield,
combining the IASP91 reference model (Kennett and Engdahl, 1991) with crustal information
from local refraction profiles (Diaz and Gallart, 2009). The imaging condition is implemented as
the integral for all time steps of a cross-correlation function between the radial and vertical
wavefields for each earthquake (eq. 4 in Shang et al 2012). Finally, to enhance the signal-to-
noise ratio and the illumination of structures along the profile, we stack over the normalized
images from all individual events.
Figure 2. A) A back-azimuth (top panels) and incidence angle (bottom panels) distributions in the data of both legs. B) Example of the interpolated vertical and radial components at a regular 500-m-grid for one earthquake recorded by the Hire-I seismic array (~2 km interstation distance); C) Green’s functions obtained after deconvolution of the source signature from the traces displayed in B).

3 Results and Discussion

In figure 3, we display the RMT images for HIRE-II (in the north) and HIRE-I (in the south). The RTM images show clearly the topography of the Moho discontinuity (dashed lines at the bottom panels, Fig. 3). Waveform coherence increases with increasing depth and decreasing interstation distances. Various intra-crustal discontinuities are imaged, albeit with less lateral extension than the Moho discontinuity. As expected from the difference in station spacing we resolve less complexity in the subsurface structures below HRII than below HRI.
Figure 3. (a) and (b): RTM images along Hire-I and Hire-II legs, respectively. At the top we display the topography along the profile, the location of the seismic stations and the contact at the surface of the different geological units (vertical dashed lines; IM: Iberian massif; EZP: External Zones Prebetics; EZS: External Zones Subbetics; IZAL: Internal Zones Alpujarride; IZNV: Internal Zones Nevado-Filabride). The light blue circle labeled GB encloses the location of the Guadix Basin. At the bottom panels, the figures are the RTM images from above panels including interpretation. Dashed black (Iberian domain) and white (Alboran domain) lines mark the Moho discontinuity whose lateral variations in depth have been labeled N1, N2, and N3 in Hire-II (a), and S1, S2, S3 and S4 in Hire-I (b). The light blue circle labeled GB encloses the location of the Guadix Basin. The profiles coordinates are (37.60,-3.08)-(36.72,-3.01) for Hire-I, and (38.65,-3.00)-(37.40, -3.08) for Hire-II.

The observed Moho discontinuity presents clear variations in depth and slope along the profile, including a series of rather abrupt offsets. For discussion, we labeled distinct segments with N1 to N3 and S1 to S4 (Fig. 3). The crustal thickness of ~30-32 km observed at the northern side segment N1 is in agreement with previous receiver function studies (Mancilla et al. 2005b). The Iberian massif, in its southern part, is characterized by a homogenous crustal structure and an almost flat Moho discontinuity (e.g Palomeras et al. 2009, Mancilla and Diaz 2015). Towards the south, there appears to be a gentle transition to segment N2 with deeper Moho (crustal thickness of ~ 35 km). However, the lack of data from HR46 produces a gap in data coverage and prevents the observation of a continuous Moho. Further south, the depth of the Moho increases gradually along segment N3, it reaches its largest depth (45-50 km) under the Guadix Basin (segment S4).
The change in the Moho dip from N2 to N3 occurs over a very short distance along the profile (<5 km) and is well constraint.

The RTM image of the southern leg (Fig. 3b) shows the continuation of the N3 segment toward the south and the change in Moho depth towards S4 segment with an offset of ~6-7 km. Lateral variation in the velocity structure along the profile (including sedimentary basin) are not considered in the wave-field propagation. Using the same 1D model for the whole profile may introduce artificial variations in the Moho topography. For example, slow wave propagation through the Guadix sedimentary basin may produce a larger Moho depth in the segment S4 with respect to the adjacent segments. The thickness of the sedimentary cover is less than 1000 m in its thickest part, which is located above the transition from segment N3 to S4 (Sanz de Galdeano et al., 2007). Assuming average Vp and Vs for the sediments of 3.8 km/s and 2.0 km/s, respectively, the expected bias on crustal thickness is less than 2 km. Accordingly, we do not interpret the small Moho offset between segments S4 and S3 (~3km) as structural, because it could be significantly influenced by inaccuracy of the applied earth model. However, the total Moho offset between N3 and S4 of ~6 km, cannot be plausibly explained by lateral variations of the velocity structure.

The most prominent feature of the Moho topography is a Moho step with ~15 km vertical offset, located at 37.2° latitude, between S2 and S3 segments (Fig. 3b). This Moho step was already identified in CCP images of receiver functions in the same profile (~17 km in CCP image, Mancilla et al. 2018) and was interpreted as produced by a STEP fault that allowed the westward roll-back of the Alboran slab along its northern edge. The RTM confirms the sharpness of this Moho step.

An apparent Moho offset between segments S1 and S2 inside the Alboran domain coincides with a substantial change in altitude along the profile, reaching 2400 m near the center of S2 segment (Sierra Nevada mountain). Since we do not account for topography in back-propagation, we cannot rule out that the transition between S1 and S2 actually corresponds to a more gentle decrease of crustal thickness towards the Mediterranean coast than suggested by Fig. 3b.

The change in the Moho dip from N2 to N3 and the ~15km vertical offset near latitude 37.2° may mark the boundaries of a distinct ~60 km wide crustal block, located in between the Iberian
foreland to the north and the Alboran domain (overriding plate) to the south (empty black square in Fig. 1). Its thickness and complex internal structure (with three distinct Moho segments, N3, S4 and S3, Fig. 3) suggest that this block is heavily deformed and internally fragmented.

The large crustal thickness of this deformed Iberian block is laterally offset from the highest topography (further south at segment S2) (Fig. 3b) but coincides with the minimum in the Bouguer anomaly (~100-120 mGal, Fig. 4a, Bureau Gravimétrique International, http://bgi.omp.obs-mip.fr). Both observations point to the lack of isostatic compensation of the Sierra Nevada Mountains, in agreement with the putative removal of Iberian lithosphere along the STEP fault (Mancilla et al. 2018). North of this block, the crust is formed by undeformed continental crust of 30-35 km thickness that possibly continued to a more transitional crust. In general, STEP faults tend to propagate along the ocean-continent boundary (e.g. Govers and Wortel, 2005, Gallais et al., 2013). However, our RTM images, the continuity of Iberian crustal structure up to the STEP fault rules out a spatial coincidence of the tearing with the continent-ocean boundary. Instead, we observe that the tearing occurs within continental or transitional crust, and suggest that the tearing process probably used a pre-existent weakness zone to evolve, possibly inherited from rifting of the continental paleomargin (Subbetics, García-Hernandez et al. 1980). The location of this tearing, together with the other Moho offsets and variations of Moho dip seems to be the result of the reactivation during the collision and roll-back processes of previous deformation features.

Roughly, the observed crustal thickness values and its variation in the study area are in agreement with previous receiver function studies of individual stations located nearby the profile (Mancilla and Diaz, 2015) and CCP images (Mancilla et al. 2015b, Mancilla et al. 2018). In Figure 5, we compare both methodologies building images for the same profile using the same set of earthquakes, and the same stations with exception of station HR46, which was equipped with short period sensors and excluded from RTM (Fig. 1). In the CCP images (Fig. 4d), we back-projected the P-wave receiver functions along their incident ray path using the same 1D earth model as for RTM (for details in the calculation of the P-wave receiver functions see Mancilla et al. 2018). In this back-projection, we take into account the increase of the width of the Fresnel zone with depth. We stack in the cross-section all receiver function amplitudes with
piercing points within 20 km distance from the profile.

Fig 4. Topography and Bouguer anomalies (blue line) along the combing profile showing the location of the seismic stations (red triangles, Hire-I, and blue triangles, Hire-II) and the contact at the surface of the different geological units (see Fig. 4 caption for details); b) Combined RTM image; c) Same as b) including interpretations lines of tectonic character. Dashed black (Iberian domain) and white (Alboran domain) lines mark the Moho discontinuity; d) CCP images with interpretation lines. The profile coordinates are (38.65, -3.00)-(36.72, -3.01)

For comparison with the CCP, we create a merged RTM profile stacking the images resulting from the analysis of both legs. The region in which both profile intersect is emphasized by a transparent-grey square ~20 km long (Fig 4a-b). In rough outlines, the topography and depth of the first order discontinuities are traced quite similar with both RTM and CCP approaches. The principal Moho step at 37.2° latitude is recovered from both methodologies, although the depth offset is lightly smaller in the case of RTM. The CCP stacks suggest a discontinuous Moho with an overlap between the Alboran domain and Iberian domain (between segments S3 and S2). This overlap is an artifact of the assumption in the CCP that interfaces are locally horizontal.
However, the RTM image does not show this overlap and supports a sharp separation between both domain at crustal scale. This observation suggests that underthrusting of the Iberian lithosphere under the Alboran domain is minor or absent in this sector.

The continuity of the Moho along the profile can be traced more clearly using the RTM method. This is observed especially in zones where the CCP method shows an ambiguous Moho such as the section from 37.2° to 37.6° (segments N3, S4 and S3) where the thickest crustal thickness is found. Furthermore, RTM methodology resolves finer the Moho slope and we can identify variations in depth that are not visible in the CCP cross section where the discontinuity appears blurred (i.e. transitions from S3 to S4, and to N3). Intracrustal structures exhibit some coincidences using both approaches such as those ones located in the segments S2 and N2. However, the RTM image shows less complexity at intracrustal depths. Note that with the station spacing used in this study the very shallow structures could be not considerably enhanced with the RTM approach.

4 Conclusions

We apply RTM to the teleseismic converted phases recorded in a high-density seismic profile to create a high-resolution image of the crustal structure, and in particular of the Moho discontinuity. RTM produces clearer images of the structure and better resolution of lateral changes than CCP stacking. The profile probes the crust of the Central Betics around the complex contact region between the Iberian and Alboran domains.

The main observations include a sequence of rather abrupt Moho offsets, one of them associate with a STEP fault, as well as variations in Moho dip. We propose that the principal Moho irregularities define a deformed block of ~60 km width formed by continental/transitional Iberian crust that concentrates most of the deformation due to the collision and roll-back processes. We suggest that the main variations in the topography of the Moho discontinuity, and the geometry of the STEP fault in Central Betics are driven by inherited weaknesses within the Iberian paleomargin.
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https://figshare.com/s/9edd19ff4956ec60f4e6

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