Crustal and upper-mantle structure below Central and Southern Mexico

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Key Points:

- A three-dimensional full seismic waveform model of Central and Southern Mexico for 5–50 s periods was constructed via adjoint methods
- The shear wave velocity model indicated two continuous anomalies in the shallow upper mantle and the lower crust
- These anomalies were associated with the ultra-slow seismic velocity zones and the high conductivity regions where slow earthquakes have been identified.

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This article has been accepted for publication and undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process, which may lead to differences between this version and the [Version of Record](#). Please cite this article as [doi: 10.1029/2020JB020906](#).

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Abstract

The Mexican subduction zone is a complex convergent margin characterized by a slab with an unusual morphology, an abnormal location of the volcanic arc, and the existence of slow earthquakes (slow slip events and tectonic tremors). The number of seismic imaging studies of the Mexican subduction zone has increased over the past decade. These studies have revealed the seismic wave velocity structure beneath Central and Southern Mexico. However, the existing tomographic models are either confined to specific areas or have coarse resolution. As a consequence, they do not capture the lateral and finer-scale variations in the mantle and the crust structure. Here, we fit frequency-dependent traveltime differences between observed and synthetic seismograms in a three-dimensional model of Central and Southern Mexico to constrain crustal and upper mantle seismic velocity structures jointly. We use $\sim 3,300$ seismic records, filtered between 5 and 50 s, from 74 regional earthquakes. We perform point spread function tests to assess the resolution and trade-offs between model parameters. Our tomographic model reveals several seismic wave velocity features. These features include the geometry of the Cocos slab, the partial melting zone beneath the Trans-Mexican Volcanic Belt, and the Yucatan slab diving below Los Tuxtlas Volcanic Field that are consistent with previous studies. We also identify horizontal structures in the lower crust and the shallow upper mantle that were not previously identified. These structures include slow seismic wave velocity anomalies that may be correlated with ultra-slow seismic velocity zones and high conductivity regions, where slow earthquakes have been identified.

1 Introduction

Along the Middle American Trench (MAT), the Cocos Plate subducts beneath the North American Plate producing large seismic events, slow slip events (SSE), tectonic tremors (TT) and an unusual orientation of the Trans-Mexican Volcanic Belt (TMVB) (Manea et al., 2013); (Fig. 1). The subducting Cocos slab is complex due to the variability in age, the convergence rate and dip angle along the MAT. Pardo and Suárez (1995) first described the unusual sub-horizontal subduction beneath Central Mexico based on relocated hypocenters of local and regional earthquakes. They were unable to determine the geometry of the slab beneath the TMVB, however, due to the scarce seismicity in this region.

To improve understanding of the subduction of the Cocos plate under the North American Plate, new temporal seismic networks were deployed in Mexico. As a result, new crust and upper mantle models were obtained by different seismological methodologies. In Central Mexico, teleseismic converted phases studies (Pérez-Campos et al., 2008; Song et al., 2009; Kim et al., 2010, 2012) and seismic tomography studies (Husker & Davis, 2009; Iglesias et al., 2010; Huesca-Pérez & Husker, 2012) presented the first image of the Cocos slab, which showed that the slab dives horizontally for ~ 250 km from the MAT. Simultaneously, Husker and Davis (2009) presented a P-wave tomography of the subducted Cocos slab underneath the TMVB. They found that the slab plunges into the mantle just below the TMVB with an angle of $\sim 75^\circ$. Moreover, they observed that the high-velocity anomaly associated with the slab continues to a depth of ~ 450 km, which supports the slab detachment hypothesis by Ferrari (2004). Based on surface wave tomography, Iglesias et al. (2010) determined the S-wave velocity structure of the Cocos slab. They observed a low-velocity zone below the active part of the TMVB that is associated with the mantle wedge. This low-velocity zone can be attributed to a high-attenuation lower-crust region (Chen & Clayton, 2009), which correlates with the low-resistivity region obtained by Jödicke et al. (2006). On the other hand, Chen and Clayton (2012) imaged the slab geometry near the Isthmus of Tehuantepec, in Southern Mexico. They reported two dipping high-velocity anomalies. One of them associated with the Cocos slab, which subducts northward with a dip angle of about 30° to 35° . The other anomaly subducts southward from the Gulf of Mexico and is associated with the Yucatan slab that was pro-

Figure 1. Tectonic setting of Central and Southern Mexico. The red square delimits the area of study. The depth contours show the depth to the top of the subducted slab (Hayes et al., 2018). The red shaded regions show the locations of tectonic tremors (TT) that have been found in previous studies (Huesca-Pérez & Husker, 2012; Brudzinski et al., 2010). The blue outlines indicate the slow slip event (SSE) regions (Graham et al., 2015; Radiguet et al., 2012). The cyan bars indicate the regions of high conductivity (from Jödicke et al. (2006)). The volcanic provinces, Trans-Mexican Volcanic Belt (TMVB) and Los Tuxtlas Volcanic Field (LTVF) are delimited in orange outlines, and the locations of volcanoes are indicated by orange triangles (www.ngdc.noaa.gov/hazard). EPR: East Pacific Rise; RFZ: Rivera Fracture Zone; OFZ: Orozco Fracture Zone; TR: Tehuantepec Ridge.

duced by the collision of the Yucatan block with Mexico during the Miocene (Kim et al., 2011).

Apart from the unusual subduction geometry beneath Mexico, occurrences of slow slip events (SSEs) and tectonic tremors (TTs) along the subduction zone have been observed (Graham et al., 2015; Radiguet et al., 2012; Huesca-Pérez & Husker, 2012; Brudzinski et al., 2010). In general, most SSE and TT events occur at transitions in frictional properties from stick slip to stable sliding. To date, SSEs and TTs have been located in isolated places (Figure 1). In Central Mexico, SSEs extend about 150 km inland from the coast, and in Southern Mexico, they are limited within 125 km near to Oaxaca (Radiguet et al., 2012). The locations of TTs are concentrated in two clusters along dip in Central Mexico: the up-dip cluster in the transition zone and the down-dip cluster, often called

the “sweet spot” (Villafuerte & Cruz-Atienza, 2017). In Southern Mexico, TT activity is close to the down-dip end of the slow-slip zone beneath Oaxaca (Brudzinski et al., 2010).

The triggering mechanisms of SSEs and TTs are not yet well understood, but one mechanism that may be common to both is the presence of fluids (Schwartz & Rokosky, 2007). In Mexico, the location of SSEs has been associated with the existence of an ultra-slow velocity layer (USL) on top of the Cocos slab. This layer is related to the hydrated oceanic crust, originating a high-pore-fluid pressure zone (Song et al., 2009). The TTs are correlated with the fluids released from the Cocos slab, that interact with the lower continental crust (Manea et al., 2013; Brudzinski et al., 2010).

In summary, the complex morphology of the Cocos slab and the presence of slow earthquakes (SSEs and TTs) make this region one of the most studied in Mexico. Even though the number of studies of the seismic wave velocity structure in Central and Southern Mexico has increased in the past decade (Castellanos et al., 2018; Spica et al., 2016; Chen & Clayton, 2012), these studies have all utilized a specific part of the seismic recordings (body or surface waves), which constraints the seismic wave velocity model to a particular type of seismic waves. In this study, we improve the resolution of the tomographic models beneath Central and Southern Mexico by applying full-waveform inversion, which allows us to exploit the full seismic record. We are therefore able to capture the lateral and finer-scale variations of the shallow upper mantle and crust structure, leading to clearer understanding of newly identified anomalies.

2 Data

We used data from three different networks (Fig. 2): MASE (Meso America Subduction Experiment), a temporary network of 100 stations operated from 2005 to 2007, extended perpendicularly from the trench passing through Acapulco, Mexico City and Tampico; VEOX (Veracruz Oaxaca line), another temporary network of 45 stations located near to the Isthmus of Tehuantepec and operated from 2007 to 2009; Servicio Sismológico Nacional (SSN), a permanent Mexican seismic network distributed along Central and Southern Mexico. In this study, we analyzed 74 earthquakes distributed along the modeling area, with magnitudes, M_w , ranging from 4.5 to 5.5 (Fig. 2). We selected earthquakes that were simultaneously recorded from two seismic networks, either MASE - SSN or VEOX - SSN, which optimized the number of recordings per earthquake. With a maximum number of stations of 67 and a minimum of 15, the complete data set for this study contains more than 3,300 three-component waveforms.

The earthquakes were modelled as point sources. For each earthquake, we retrieved the origin time, the hypocentre and the moment tensor from the Servicio Sismológico Nacional (SSN) catalog. We performed several tests to determine the optimal source time function and the moment tensor accuracy for every earthquake.

3 Methods

To retrieve a new three-dimensional (3D) model of Central and Southern Mexico, we applied Full-Waveform Inversion (Tarantola, 1984; Fichtner et al., 2009; Tape et al., 2010; Simuté et al., 2016). Further details regarding the forward modeling, initial model, data selection, misfit function, and the inverse problem are provided in the following section.

3.1 Initial Model and Forward Modeling

Our 3D initial model, $\mathbf{M}_{00}$, was built based on a previous study of surface wave tomography by Spica et al. (2016). The model was parametrized in terms of shear wave velocity, β , compressional wave velocity, α , and density, ρ . We then assumed a constant

α/β ratio of 1.73 for the whole model and determined the density from the compressional wave velocity (Gardner et al., 1974). The model parameters were smoothed by a Gaussian function with a radius of 15 km to facilitate their representation in the discrete grid (Figures S1 and S2). The modeling area dimensions are from 16° to 21°N and -102° to -94°E (Fig. 1).

Shown in Figures S1 and S2 in the supporting information, we plotted vertical cross-sections and map views of the initial velocity model, M_{00} . In Figure S1, the initial velocity model is close to a 1D velocity model with two layers, which vary smoothly in the vertical direction. The map views of the initial velocity model (Figure S2) describe the long-wavelength lateral structure of compressional wave velocity α and shear wave velocity β .

Figure 2. Area of study. The black dashed lines indicate isodepths to the top of the subducted slab (Hayes et al., 2018). The inverted triangles show the locations of the stations used in this study: MASE array (magenta), VEOX array (yellow), and SSN network (blue). The cyan dots indicate the earthquake epicenters employed in the inversion. The TMVB is delimited in red, and red triangles show the locations of volcanoes.

For the forward modeling, we implemented elastic wave propagation by a finite-difference time domain code with a fourth-order scheme in space and a second-order scheme in time for computational efficiency (Martin & Komatitsch, 2009; Kristek et al., 2002; Moczo et al., 2002; Hicks, 2002; Pitarka, 1999). The simulation model has a spatial grid size of 568 X 364 X 96 with a grid spacing of 1.5 km, resulting in 10 nodes per minimum wavelength. This finer discretization ensured that the numerical errors were negligible compared to the differences between the observed and the synthetic data. The period range for the wave simulations and inversion was from 5 to 50s.

3.2 Inverse problem

To solve the inverse problem, we employed adjoint methods, which allowed us to compute the gradient of any misfit function in an efficient way. This technique was successfully applied at many different scales in exploration seismology (Luo & Schuster, 1991; Pratt et al., 1998; Virieux & Operto, 2009), regional and global seismology (Fichtner et al., 2009; Simut et al., 2016; Bozda et al., 2016).

3.2.1 Selection of time windows and misfit measures

To compute the gradient, we define a suitable misfit function to avoid nonlinearities, which may occur when we assimilate the observed data. We followed a two-step procedure to quantify the difference between the observed and synthetic data. In the first step, we employed an automated algorithm PYFLEX (Krischer & Casarotti, 2015; Maggi et al., 2009) that is a python-based workflow window selection. The objective of PYFLEX is to find a suitable time window within which there is a waveform similarity (amplitude and phase) between the observed and synthetic data. PYFLEX requires several user parameters which have to be adjusted for a given data set. The window selection algorithm was applied to the three components to assimilate body waves and surface waves. To ensure the misfit measurements quality, we used a conservative window acceptance criterion, selecting high values of acceptance for the window selection algorithm's parameters. The selection of windows was performed at each iteration so that the number of windows evolves during the inversion. For a detailed description of the PYFLEX parameters involved in this study we refer to Table S1 in the supporting information.

In the second step to quantify the difference between the observed and synthetic data, we quantified the misfit within the windows. We consider two different misfits according to the type of waveform and window length: a traveltime difference (Luo & Schuster, 1991),

$$F_i^T(\mathbf{m}) = \left[\frac{\Delta T_i(\mathbf{m})}{\sigma_i} \right], \quad (1)$$

and a frequency-dependent traveltime (Tape et al., 2010),

$$F_i^T(\mathbf{m}) = \int_{-\infty}^{\infty} \frac{h_i(w)}{H_i} \left[\frac{\Delta T_i(w, \mathbf{m})}{\sigma_i} \right] dw, \quad (2)$$

where $\mathbf{m}$ is a model vector, $\Delta T_i(w, \mathbf{m}) = T_i^{obs}(w) - T_i(w, \mathbf{m})$ and $\Delta T_i(\mathbf{m}) = T_i^{obs} - T_i(\mathbf{m})$ denote the traveltime difference frequency-dependent and frequency-independent, respectively. The traveltime measurements, $T_i(w, \mathbf{m})$, are computed using a multi-taper method (Percival et al., 1993), and $h_i(w)$ is a frequency domain window which is normalized with a constant, $H_i = \int_{-\infty}^{\infty} h_i(w) dw$, (Tape et al., 2010). In the case of a frequency-independent measurements, $\Delta T_i(\mathbf{m})$, is a cross-correlation travel time difference. For both misfits we have to compute the uncertainty, σ_i , associated with each window i . We followed Tape et al. (2010, Appendix A) to estimate σ_i . To stabilize equation (1) and equation (2) when σ_i becomes very small, we have to define a minimum value, σ_0 , such that $\sigma_i \geq \sigma_0$. For this work, we selected $\sigma_0 = 1$ s, to include possible uncertainties in the source parameters (hypocenter or origin time).

3.2.2 Optimization Method

Once we computed the forward simulations, misfit function and adjoint source, we calculated the gradient employing adjoint techniques (e.g., Tarantola, 1984; Tromp et al., 2005). To reduce the misfit, we performed 19 iterations using the quasi-Newton method

L-BFGS, which is based on a low-dimensional quadratic model of the misfit function, resulting in faster convergence to the minimum (Liu & Nocedal, 1989; Modrak & Tromp, 2016). To avoid cycle skipping and a local minimum, we followed a multi-scale approach (Bunks et al., 1995) that consisted of gradually increasing the range of the period band. We started with long periods between 10 to 50 s. Then, we performed several iterations until we reached a minimum in the misfit function. In the next stage, we increased the period band once again from 7 to 50 s, and finally 5 to 50 s. In each period band we restarted the L-BFGS algorithm.

3.2.3 Line search and Model Update

To update the velocity model in every iteration, we perform a line search to find the step length, γ , such that the updated model $\mathbf{m} + \gamma \mathbf{p}$ satisfies the decrease and curvature conditions (Wolfe conditions) (Nocedal & Wright, 2006), where $\mathbf{m}$ is the current model and $\mathbf{p}$ the search direction (gradient). The L-BFGS algorithm with proper initialization returns a well-scaled search direction, where a step length, $\gamma = 1$, is an appropriate first approximation. Most of the time, through the inversion, a unit step length satisfies the decrease and the curvature conditions. When that is not the case, we have to perform one or more trial steps. In this study, we perform a backtracking line search (Modrak & Tromp, 2016), which allows us to compute subsequent trial steps by interpolating backwards, towards zero, within the unit interval. To do the interpolation, we use the quadratic and cubic interpolation algorithms given by Nocedal and Wright (2006).

In general, traveltime measurements are more sensitive to wave velocity than density which is usually constrained by normal modes (Ishii & Tromp, 1999). Thus, in our inversion, we updated the compressional wave velocity, α , and the shear wave velocity, β . We used a scaling relationship between the relative perturbation in density, $\delta \ln \rho$, and the relative shear wave velocity perturbation, $\delta \ln \beta$, to update density:

$$\delta \ln \rho = 0.2 \delta \ln \beta. \quad (3)$$

This scaling factor of 0.2 was suitable for the study area and remained in the assumed range of validity within the Earth from -0.1 to 0.3 (see e.g., Karato & Karki, 2001; Simmons et al., 2009; Moulik & Ekström, 2016).

3.2.4 Regularization

In adjoint tomography, we do not include a regularization term in the misfit function. Instead, we smoothed the gradient in every iteration via convolution with a 3D Gaussian function, which may be considered as a regularization procedure (Tape et al., 2010). In this study, we employed a 3D Gaussian function with different radius in the horizontal and vertical directions. The radius of the Gaussian function was defined by the minimum resolvable wavelength for each period band in the multi-scale approach. For the initial stage of the inversion, we chose a conservative 3D Gaussian function with a ~ 25 km radius to smooth the misfit gradient. By conservative, we mean that the Gaussian function's diameter is larger than the minimum resolvable wavelength for this initial stage. The radius is gradually decreased from ~ 25 km to ~ 15 km as we include shorter periods in the data, resulting in small scale features contained in the misfit gradient.

4 Results

After performing 19 iterations, we obtained a new model for Central and Southern Mexico. We present our new model, $\mathbf{M}_{19} = [\rho, \alpha, \beta]_{19}$, on both absolute and relative scales. The relative scale is represented by the model update, $\ln(\mathbf{M}_{19}/\mathbf{M}_{00})$, representing element-by-element operations. Given that the first order approximation of the

natural logarithm is the formula for a percent difference, $\delta \ln(\alpha) \equiv \ln(\alpha_{19}/\alpha_{00}) \approx \frac{\alpha_{19}-\alpha_{00}}{\alpha_{00}}$, this is the presentation of the tomographic model as a relative perturbation from the reference model ($\mathbf{M}_{00}$). The absolute scale is represented by $\mathbf{M}_{19}$, which describes the shear wave velocity, β , and the compressional wave velocity, α , in units of $km\ s^{-1}$.

In Figure 3, we show the relative perturbations in the compressional and shear wave velocities at different profiles. The compressional wave velocity results show strong heterogeneities, reaching relative perturbations from -4% to 6% . We identified a positive anomaly in each profile, that dipped northeast from the coast. Specifically, in the B-B' profile, the high-velocity structure is sub-horizontal, with an extension of 200 km from the coast. In the same profile, we observed a shallow negative anomaly with a varying thickness, that increased in depth towards the northeast. For E-E' and F-F' profiles, two important structures with high compressional velocity we identified, one dipping northeast from the Pacific coast and the other close to the Gulf of Mexico and is dipping southwest.

The shear wave velocity results exhibit a great variation across the study area, with relative velocity perturbations ranging -13% to 11% . We clearly observed two important anomalies in B-B' and D-D' profiles. The first anomaly is a sub-horizontal high-velocity structure, extending 200 km from the coast. The second anomaly is a shallow low-velocity perturbation with a variable thickness, that increases towards the northeast.

In Figure 4, we show map views of relative compressional and shear wave velocity perturbations at different depths. We label the important features observed in every cross section. In the lower crust (35 km depth), we identified a low-velocity anomaly parallel to the MAT in the compressional wave velocity perturbations (labeled A), which is ~ 200 km away from the trench. This negative anomaly is located where the horizontal subduction occurs, with a lateral extent of ~ 600 km from the western part of Guerrero to Oaxaca. In the upper mantle (45 and 60 km depth), we observed the same low-velocity anomaly but now it is fragmented in three structures (labeled A1, A2 and A3) and where the magnitude of the perturbations decreases with depth.

We identified four important shear wave velocity features at different depths (Figure. 4). In the lower crust (35 km depth), two strong low-velocity perturbations are observed (V1 and V2). These anomalies are correlated in location with those observed in the compressional wave velocity perturbations (A1, A3), which are associated with a high conductivity zone and the location of TT's (Husker et al., 2019; Payero et al., 2008; Jödicke et al., 2006). There is another weak low-velocity anomaly located in the boundary between Oaxaca and Veracruz (labeled VB). This anomaly is vertically continuous from the surface to the lower crust where its magnitude decreases with depth. Finally, T1 is a low-velocity anomaly continuous throughout the entire depth range of our model, extending along the TMVB near Mexico City.

4.1 Resolution Analysis

Checkerboard tests to estimate the resolution of the tomographic inversion are standard in seismic traveltime tomography. However, for 3D Full-Waveform Inversion (FWI) is computationally expensive because the same number of iterations as in actual structural inversion have to be performed. To alleviate this problem, Zhu et al. (2015); Luo et al. (2013); Tape et al. (2010) successfully performed such experiments in two dimensions (2D). They deduced that the resolution is controlled by the ray coverage, as we might expect. A reasonably good approximation of the resolution is the ray density map (Luo et al., 2013). On the other hand, to estimate the quality of the models and the trade-off between parameters, we perform the point spread function test (PSF) proposed by Fichtner and Trampert (2011). To perform the PSF test, we used a finite-difference approximation to compute the action of the Hessian on the localized model perturbation,

$$\mathbf{H}\delta\mathbf{m} \approx \mathbf{g}(\mathbf{m} + \delta\mathbf{m}) - \mathbf{g}(\mathbf{m}), \quad (4)$$

where $\mathbf{H}$ denotes the Hessian and $\delta\mathbf{m}$ refers to a localized model perturbation with respect to the current model, $\mathbf{m}$. Therefore, we have to evaluate the gradient, $\mathbf{g}$, for both models, $\mathbf{m}$ and $\mathbf{m} + \delta\mathbf{m}$, an operation that requires only two gradient calculations. Because, we already have the gradient for $\mathbf{g}(\mathbf{m})$ from the last iteration in the inversion, we only have to compute the gradient for the localized perturbation, $\mathbf{g}(\mathbf{m} + \delta\mathbf{m})$. Considering that we are computing the action of the Hessian on a model perturbation, $\mathbf{H}\delta\mathbf{m}$, the PSF test allows us to estimate the degree of blurring of the induced model perturbation and to identify trade-offs between parameters.

Using the PSF test, we perform a resolution analysis at six different locations parallel to the coast. A -2% 3D-Gaussian model perturbation with a radius of 25 km was utilized. In Figure 5, the velocity model is perturbed (shear wave velocity, β) at a depth of 45 km with respect to the last model obtained from the inversion. The model perturbations were reasonably well retrieved in Central Mexico, although we observed modest horizontal smearing at the eastern and western parts of the modeling area. From the vertical cross-sections (Figure 6), the model perturbations are reconstructed without much bias or smearing in the shallow upper mantle, which supports the resolution of the observed features in the model, $\mathbf{M}_{19}$. Although there are some small spurious anomalies at the location of the perturbation (profiles A-A' and B-B'), they are not strong enough to overshadow the main perturbation.

Shown in Figures S4 and S5, we performed two additional PSF tests in different locations. In Figure S4, the model perturbation is localized beneath Oaxaca, at a depth of 50 km. The shear wave velocity perturbation is well located and reconstructed, although it exhibits reasonable horizontal and vertical smearing. The second PSF test (Figure S5) is beneath Veracruz at a depth of 50 km. Even though there are many small spurious perturbations, the main 3D-Gaussian perturbation in the compressional wave velocity is reconstructed correctly. Based on the PSF analysis (Figures 5-6 and Figures S4-S5), we conclude that the anomalies observed in our inverted model are robust, and the trade-offs between wave velocities (α and β) are fairly weak.

Finally, following Zhu et al. (2015); Luo et al. (2013) we used an approximation of the Hessian diagonal to compute the ray coverage in the region of study. To compute the ray coverage map at different depths, we followed the same procedure utilized to obtain the gradient of the misfit function but replacing the traveltime measurement weight for each adjoint source (i.e., using a unit traveltime measurement). The approximation of the Hessian diagonal has maximum sensitivity close to the surface and decreases with depth. Because the diagonal of the Hessian is an approximation, the ray coverage map only helps to evaluate the model sampled areas by the dataset qualitatively. From Figure 7, we observe the ray coverage from the upper crust to the shallow upper mantle. Low (close to zero) and high values (close to four) indicate poor and good coverage, respectively.

4.2 Misfit Analysis

As mentioned in Section 3.2.2, we started the inversion with long periods and then we successively increased the period band three times, 10-50 s, 7-50 s, and 5-50 s. In Figure 8, we show the evolution of the misfit from the initial 3D Model, $\mathbf{M}_{00}$, to $\mathbf{M}_{19}$ for all three inversion stages. The misfit is normalized to 1 for every period band. For the first stage, the misfit reduction is about 25%; for the second stage, it is 35%; and for the last stage, it is approximately 15%. A representative example of the misfit reduction for individual stations is presented in Figure 9, where we compare the fit of the synthetics for the initial model, the synthetics for the final model, and the observed data.

To assess the quality of the new model, $\mathbf{M}_{19}$, we utilized the waveform difference, $F_i^w(\mathbf{m})$,

$$F_i^w(\mathbf{m}) = \frac{\int_T w_i(t)[d(t) - s(t, \mathbf{m})]^2 dt}{\left[\int_T w_i(t)[d(t)]^2 dt \int_T w_i(t)[s(t, \mathbf{m})]^2 dt\right]^{\frac{1}{2}}} \quad (5)$$

proposed by Tape et al. (2010). In equation (5), $d(t)$ denotes the observed data, $s(t, \mathbf{m})$ the synthetic data, $w_i(t)$ the time window, and i the measurement window index. The waveform difference allows us to compare waveforms from the initial and final models, which is not possible when the traveltime difference is due to the degree of the similarity required (equation 1). Thus, we compute the waveform difference using the time windows that were selected from the final model, $\mathbf{M}_{19}$. In Figure 8, we show the misfit analysis histogram for the initial and final tomographic models for all three period bands (10-50 s, 7-50 s, 5-50 s). We see that the misfit distribution for the final model shrinks around $F_i^w(\mathbf{m}) = 1$, reducing the observed misfit tail in the initial model distribution from 5 s to 3 s.

5 Discussion

5.1 Comparison with previous studies

The flat subduction in Central Mexico is unique due to its occurrence at a relative shallow depth of (~ 45 km), in comparison to other flat subduction zones, Chile (~ 100 km), Peru (~ 100 km) and Japan (~ 60 km) (Gutscher et al., 2000). Although the peculiar behavior of the subduction in Central Mexico is not well understood, there are several hypotheses that seek to explain the slab's geometry. Song et al. (2009); Kim et al. (2010) suggested the presence of an ultra-slow velocity layer (USL) on top of the Cocos slab that explain the decoupling of the flat slab from the overriding plate, which may be responsible for the flat subduction. The origin of the USL is not known but it may be related either to fluid-saturated oceanic slab (Song et al., 2009) or hydrous minerals (Manea et al., 2013; Kim et al., 2012).

In the B-B' profile (Figure 10), we observed a high-velocity perturbation in the shear and compressional wave velocity. This anomaly is identified as the Cocos slab, diving horizontally perpendicular to the trench. The subduction geometry observed in our results correlates to previous studies (Pérez-Campos et al., 2008). On top of the slab, we identified a low-velocity perturbation that may be related to dehydration of the oceanic crust (Manea et al., 2013). This low-velocity structure is associated with the region of high conductivity found by Jödicke et al. (2006). In our results, we did not resolve the USL due to the low vertical resolution. Instead, we may observe a smeared version mapped in the slow velocity anomaly atop of the slab. This assumption is possible because some of the fluids released by the slab may accumulate and then partially infiltrate the lower crust (Ferrari et al., 2012).

In Southern Mexico near the Isthmus of Tehuantepec (Figure 11, F-F' profile), we observed a slow compressional velocity anomaly located under the Veracruz Basin (VB) near the LTVF, which has an extension of ~ 150 km and reaches a depth of 50 km. Chen and Clayton (2012) found a similar structure that correlates in location and extension with our study. On the other hand, Jödicke et al. (2006) found a high conductivity zone that is also beneath the VB that may be associated with the low-velocity anomaly identified in our study. The origin of the anomaly may be connected to the VB in the upper crust (Chen & Clayton, 2009). However, we observed a smeared version of the anomaly covering the upper and lower crust due to the lack of vertical resolution.

For the same vertical profile, we identified two high-velocity structures, one the Cocos slab, subducting northeast with a dip angle of $\sim 30^\circ$. At a depth of ~ 100 km, the

Cocos slab is truncated by the other observed high-velocity structure, which subducts southwest from the Gulf of Mexico. This unusual high-velocity structure beneath LTVF was also found in previous tomographic and receiver function studies (Chen & Clayton, 2012; Kim et al., 2011). The origin of this high-velocity structure close to the Gulf of Mexico may be related to the Yucatan slab, which was caused by the collision between Mexico and the Yucatan Block during the Miocene (Kim et al., 2011). The presence of the high-velocity anomaly in the upper crust may explain the absence of the partial melting zone beneath the volcanic field (Chen & Clayton, 2009). One alternative interpretation for this high-velocity structure is the relatively shallow Moho (~ 30 km) observed in the area (Castellanos et al., 2018; Melgar & Pérez-Campos, 2011).

From the absolute velocity models (Figures 10 and 11), we estimated the depth of the Moho in Central and Southern Mexico. In Central Mexico (B-B' profile), the depth of the Moho reaches ~ 45 km in the back-arc and ~ 50 km below the TMVB. Conversely, in Southern Mexico, we estimated that the Moho is shallow, ~ 45 km near to the coast, ~ 40 km beneath the Veracruz Basin, and ~ 30 km under LTVF. The depths of Moho inferred from our models are consistent with previous studies on the area (Castellanos et al., 2018; Espíndola et al., 2017; Melgar & Pérez-Campos, 2011).

5.2 Lateral slow-velocity features

In Figure 4, our last model shows critical features that may be related to the location of the USL and the dehydration zone of the Cocos slab. We observed two specific low-velocity regions that appear in both the shear and compressional wave velocities (A1, A3, V1 and V2), located at the shallow upper mantle (45 km). These anomalies correlate with the locations of TT and SSE activity, inferred from previous studies (Graham et al., 2015; Radiguet et al., 2012; Brudzinski et al., 2010; Payero et al., 2008). Furthermore, at the same depth, the α/β ratio is high in the same location of the observed V1 and V2 anomalies (Figure 4, yellow rectangles). These structures have also been identified by Jödicke et al. (2006), who located high conductivity zones, suggesting the presence of fluids in these regions.

From the same figure, in the lower crust (35 km depth), a large low-velocity structure is observed (A), which extends along the strike in the subduction zone. We suggest that this structure may be related to the identified anomalies in the upper mantle (A1, A2, A3). On the contrary, in the shear wave velocity perturbations, we see larger anomalies (V1 and V2) separated by a faster anomaly (S1) that may be associated with change in the slab dip (Hayes et al., 2018; Dougherty & Clayton, 2014; Pardo & Suárez, 1995). At the same location (anomaly S1), the α/β ratio is low, suggesting the absence of fluids in this region opposite to what has been observed by Dougherty and Clayton (2014) (USL region). This difference may be related to the low vertical resolution of this study, which does not detect the USL zone. An alternative interpretation is that the fluids that are associated with the USL are trapped in the interface. This hypothesis is also supported by Husker et al. (2018), who found a gabbroic lower crust that acts as a seal, trapping the fluids from the subduction zone within the slab.

To observe with more clarity where the low-velocity anomalies are located over the Cocos Plate, we projected the velocity variations along the geometry of the subducted slab from the 3D velocity perturbation models. The 3D slab geometry was obtained from Slab2 (Hayes et al., 2018).

In Figure 12, the low-velocity and high α/β ratio anomalies related to slow earthquakes (SSEs, TTs) lie in the sub-horizontal section of the subduction (V1) and in Southern Mexico near Oaxaca (V2). The values obtained from the α/β ratio close to the interface are similar in magnitude as those found in Japan where slow earthquakes were observed ($\alpha/\beta > 1.8$) (Shelly et al., 2006). With the tomographic results, specifically the α/β ratio, it is not possible to determine if the two regions where slow earthquakes

occur are connected as the observed anomaly suggests (Figure 4, labeled A). A possible interpretation of the presence of the large low-velocity perturbation parallel to the MAT (labeled A) is a combination of increased pore fluid pressure and temperature. Vanorio et al. (2005) found that the compressional velocity, α , depends more on variations in pressure and temperature than shear velocity, β , when a rock is fluid saturated. The evidence of high conductivity regions ~ 200 km away from the trench suggest the presence of fluids, which increase the pore pressure (Jödicke et al., 2006). Therefore, as suggested by Huesca-Pérez and Husker (2012), the increase in pressure and temperature with distance from the coast explains why the compressional velocity, α , decreases with distance from the coast. An alternative interpretation is that anomalies V1 and V2 may be related, forming a wider anomaly. This interpretation implies that we could observe slow earthquakes over a wider area. The same hypothesis has also been proposed by Maury et al. (2018), who found that slow earthquakes in Guerrero and Oaxaca have similar behavior (tremor duration, focal mechanism and SSEs magnitudes), and the origin of slow earthquakes gap between Guerrero and Oaxaca is associated with the lack of stations in this region.

From Figure 4 and Figure 12, we observed that anomalies V1 and V2 are located at the along-strike transitions from flat to normal subduction. Dougherty and Clayton (2014); Dougherty et al. (2012) propose that the slab tends to tear along the sharp transitions in slab dip. Alternatively, Fasola et al. (2016) suggest that the transition from flat to steeper subduction in Oaxaca is associated with a sharp flexure of the slab. Our study results further indicate a variation in the velocity structure and α/β ratio around those locations that tries to accommodate the transitions in the slab dip.

In Figure 4, we also identified another significant anomaly (labeled T1), which is a low-velocity anomaly that is continuous throughout the model and lies beneath the TVMB. Although α/β ratio generally depends on complex relationships of pressure and temperature conditions as well as rock physical (such as matrix composition, fluid saturation, degree of consolidation, differential pressure, porosity, and many other factors such as pore geometry) and chemical properties (mineral composition), in the particular context of volcanic region T1, the increase of α/β ratio can be associated to high temperature and melting below the TVMB (Figure 4, red rectangle). We found this high α/β ratio area as a piece of supporting evidence to the hypothesis proposed by Ferrari et al. (2012), which suggests that this region is associated with a partial melting zone beneath the volcanic field, a theory that is also supported by Chen and Clayton (2012), who found a high seismic attenuation anomaly below the TVMB. This well-defined low shear wave velocity and density region, labeled as T1, located just below the active back-arc zone of the TVMB, also contributes to the isostatic balance of the area.

6 Conclusions

We assimilated the frequency-dependent traveltime differences between observed and synthetic seismograms in a fully 3D model of Central and Southern Mexico that constrained crustal and upper mantle structures jointly. Our seismic wave velocity model exhibits anomalies that were found in previous regional and local studies. We also identified new seismic velocity structures that increase understanding of the Mexican subduction zone's complexity. Along the 2D vertical profile, the tomographic model identifies a high-velocity structure associated with the Yucatan slab, suggesting the absence of partial melting beneath the volcanic field below the LTVF. Moreover, the results show two important features observed in the lower crust (V1 and V2). These low-velocity and high α/β ratio anomalies may be correlated with ultra-slow seismic velocity zones and high conductivity regions, where slow earthquakes have been identified. Although our model helps to increase understanding of the crustal and upper-mantle structures below Central and Southern Mexico, we note that we cannot determine if the slow veloc-

ity anomalies (A1, A2, A3, V1, and V2) are related to either USL or the high hydrated slab (high conductivity regions).

Acknowledgments

This research was supported by the Extreme Computing Research Center (ECRC) at KAUST under the project on HPC for large physics-based earthquake simulations (KSL K1319). SSN data were obtained from the Servicio Sismológico Nacional (México). We are grateful for station maintenance, data acquisition and data distribution. MASE and VEOX data are available at the IRIS DMC under the TO network code. The final wave velocity model is available in the Repository KaustDATA (<https://doi.org/10.25781/KAUST-1A96Z>)

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Figure 3. Vertical cross-sections of relative wave velocity perturbations. (Top) Location of the vertical cross-sections. (Bottom left) Relative perturbations in compressional wave velocity, $\delta \ln(\alpha)$. (Bottom center) Relative perturbations in shear wave velocity, $\delta \ln(\beta)$. (Bottom right) Relative perturbation of the $\delta \ln(\alpha/\beta)$ ratio. The black dashed lines show the depth to the top of the subducted slab (Hayes et al., 2018). The yellow regions show the locations of tectonic tremors (TT) that have been found in previous studies (Husker et al., 2019; Brudzinski et al., 2010). The green squares indicate the slow slip event (SSE) regions (Song et al., 2009).

Figure 4. Map view of relative wave velocity perturbations at different depths. (Left) Relative perturbations in compressional wave velocity, $\delta \ln(\alpha)$. (Middle) Relative perturbations in shear wave velocity, $\delta \ln(\beta)$. (Right) Relative perturbation of the $\delta \ln(\alpha/\beta)$ ratio. Depths are shown in the right top corner. The black dashed contours indicate the depth to the top of the subducted slab.

Figure 5. Resolution analysis computed from the last model obtained in the inversion. (Left) 3D-Gaussian model perturbations (shear wave velocity, β) at a depth of 45 km. (Right) Point spread function (PSF) reconstruction.

Figure 6. Resolution analysis: (Left) Vertical cross-sections of the 3D-Gaussian model perturbation (shear wave velocity, β), (Right) Vertical cross-sections of the point spread function reconstruction.

Figure 7. Approximate Hessian computed from the final model at different depths. Depths are shown in the right top corner. These maps may be used to quantify the ray coverage; low and high values depict areas with poor and good coverage, respectively.

Figure 8. (Top) Evolution of the normalized phase misfit during the inversion. (Bottom) Waveform misfit analysis, $F_i^w(\mathbf{m})$, of the initial and final tomographic models at different period bands.

Figure 9. (Top) Map showing the distribution of stations used to compare the overall fit between synthetic and observed data. (Bottom) Comparison between synthetic from the initial model (blue dashed line), from the final model, M_{19} (red continuous), and the observed data (black) for an event ($M_w=4.8$) in northwest Guerrero that occurred on 11 August 2006 (cyan dot).

Figure 10. Vertical cross-sections. Left: (Top) absolute compressional velocity model, (Bottom) relative compressional wave perturbations. Right: (Top) absolute shear velocity model, (Bottom) relative shear wave velocity perturbations. Distribution of seismicity (black and white dots) projected for every section from the SSN catalog for 2010-2019. Black and white dashed lines indicate depth to the top of the subducted slab. Yellow dashed lines show the depth of the Moho as indicated by our study.

Figure 11. Vertical cross-sections. Left: (Top) absolute compressional velocity model, (Bottom) relative compressional wave perturbations. Right: (Top) absolute shear velocity model, (Bottom) relative shear wave velocity perturbations

Figure 12. 3D relative velocity perturbations along the top of the slab's surface. (Top) Relative compressional velocity perturbations, $\delta \ln(\alpha)$. (Middle) Relative shear velocity perturbations, $\delta \ln(\beta)$. (Bottom) Relative perturbation of the $\delta \ln(\alpha/\beta)$ ratio.