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Seismic structure beneath the Gulf of California: a contribution from group velocity measurements

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SUMMARY
Rayleigh wave group velocity dispersion measurements from local and regional earthquakes are used to interpret the lithospheric structure in the Gulf of California region. We compute group velocity maps for Rayleigh waves from 10 to 150 s using earthquakes recorded by broadband stations of the Network of Autonomous Recording Seismographs in Baja California and Mexico mainland, UNM in Mexico, BOR, DPP and GOR in southern California and TUC in Arizona. The study area is gridded in 120 longitude cells by 180 latitude cells, with an equal spacing of 10 × 10 km. Assuming that each gridpoint is laterally homogeneous, for each period the tomographic maps are inverted to produce a 3-D lithospheric shear wave velocity model for the region.

Near the Gulf of California rift axis, we found three prominent low shear wave velocity regions, which are associated with mantle upwelling near the Cerro Prieto volcanic field, the Ballenas Transform Fault and the East Paciﬁc Rise. Upwelling of the mantle at lithospheric and asthenospheric depths characterizes most of the Gulf. This more detailed ﬁnding is new when compared to previous surface wave studies in the region. A low-velocity zone in north-central Baja at ∼28°N which extends east–south–eastwards is interpreted as an asthenospheric window. In addition, we also identify a well-deﬁned high-velocity zone in the upper mantle beneath central-western Baja California, which correlates with the previously interpreted location of the stalled Guadalupe and Magdalena microplates. We interpret locations of the fossil slab and slab window in light of the distribution of unique post-subduction volcanic rocks in the Gulf of California and Baja California. We also observe a high-velocity anomaly at 50-km depth extending down to ∼130 km near the southwestern Baja coastline and beneath Baja, which may represent another remnant of the Farallon slab.

Key words: Surface waves and free oscillations; Seismic tomography; Dynamics of lithosphere and mantle; Crustal structure.

1 INTRODUCTION
The Gulf of California is a young, obliquely divergent plate boundary where the processes affecting continental breakup and the onset of seafloor spreading can be directly studied. With the deployment of the NARS-Baja (Trampert et al. 2003; Clayton et al. 2004), a set of 14 broad-band seismic stations located in Baja California and the western Mexico provinces (Fig. 1), large-scale studies of the Gulf lithosphere along the entire ~1100-km length of the rift axis became possible. Here, we present a surface wave dispersion study of regional earthquakes recorded by the NARS-Baja seismic stations and five other stations (Fig. 1). The focus of this study is to compare our seismic velocity structure, which is determined at a map resolution of approximately 10 × 10 km to the gross lithospheric structure based on regional volcanism suggested from geological and geochemical studies (e.g. Castillo 2008; Calmus et al. 2011).

In addition, we compare our results to previous surface wave studies in this region to provide further detail, specifically on the dimensions and depth of slab remnants of the Farallon Plate beneath the Baja California Peninsula, and possible regions of mantle upwelling along the rift axis. To a ﬁrst order, these results are important for constraining lithospheric properties, such as regions of partial melting or magmatism, and thinned crust in models seeking to explain the geodynamic evolution of the Gulf, the unique volcanism recorded in Baja California (e.g. Negrete-Aranda et al. 2013) and the variation in styles of rifting along the Gulf axis. Based on numerical models, Bialas & Buck (2009), conclude that higher sedimentation rates in the northern Gulf has led to a rapid transition
from core complex to narrow rift mode compared to the southern part of the Gulf (Alarcón Basin, Fig. 1). Dorsey & Umbhoefer (2012) further suggest that the high rate of sediment input from the Colorado River plays a major role in delaying the creation of new magnetized oceanic crust in the northern Gulf and Salton Trough regions compared to the central Gulf (Guaymas Basin). The influence of sediment input at both early and late stages of rifting may be what distinguishes the northern and central-southern rift segments. Our results give insight into the current crustal and upper-mantle structures beneath the Gulf along the full length of the rift. These
results can be used in future numerical models to establish the possible trade-off between the thermal state of the lithosphere and sediment input in determining the observed variations in rift architecture.

In this study, we first measure the dispersion of Rayleigh waves from short to long periods to create a series of 2-D maps of the isotropic group velocities, which we then invert for shear wave velocities beneath the study area. In contrast to previous large-scale seismic studies of lithospheric structure in this region, we estimate the error in our group velocity maps using the bootstrap method in an effort to produce qualitative controls on the reliability of our maps. We further calculate the azimuthal distribution of ray paths in the study region to verify whether anisotropic group velocity coefficients can be reliably determined. We find that our ray coverage though very dense does not provide enough azimuthal coverage for a meaningful determination of anisotropic coefficients. Furthermore, other workers have stated that the inclusion of anisotropic terms made no difference in the lateral velocity variations (Wang et al. 2009). Our study is therefore restricted to the discussion of features in our Rayleigh wave isotropic velocity results though anisotropic effects are expected in the study region. Finally, the shear wave velocity inversion is carried out with two different starting velocity models as described in Section 4. The final step demonstrates the robustness of our approach, since our results prove to not be dependent on the initial velocity model.

We want to point out that, although several studies on surface waves were published in the last decade, most of them were based on phase velocity measurements. Our use of Rayleigh wave group velocities over a broad period range distinguishes this study from previous work in this region. There are many advantages in using group velocity instead of phase velocity measurements. In general, source effects can be ignored for most group traveltime measurements (Levshin et al. 1999). In addition, surface wave group velocity is more sensitive to shallow structure than phase velocity at corresponding periods. At an active plate boundary with long transform faults such as the Gulf of California, significant lateral variations in lithospheric structure and composition are expected. As a consequence, group velocity data between 20 and 100 s provide good and reliable constraints on the crustal and upper-mantle structure, which is also fundamental for understanding the propagation of waves at the regional scale and for locating seismic events. Group velocity maps can also be used to calculate group velocity correction surfaces (Levshin & Ritzwoller 2001) that are widely used in phase-matched filtering routines to extract low signal-to-noise wave packets from a seismogram. Finally, accurate group velocity maps can greatly lower the detection threshold for small events, which is essential for catalogue completeness.

In addition to previous surface waves studies, other related investigations that have used NARS-Baja as well as other data sets to determine the large-scale tectonic features in this region include receiver function studies of Moho depth (Persaud et al. 2007), as well as SKS (Long 2010) and SKS-receiver function studies of anisotropy (Obrebski & Castro 2008; van Benthem et al. 2008). We analyse our detailed results in light of these studies.

## 2 GEOLOGICAL SETTING

Extension at the Pacific–North America Plate boundary in the Gulf of California region started after 12 Ma along a 24–12 Ma subduction related volcanic arc (Gastil et al. 1979; Sawlan 1991). About 6 Ma, the Pacific–North America Plate boundary shifted ∼250 km inland to its current location in the Gulf of California, roughly parallel to the coastline (Oskin et al. 2001) and marine sediments were widely deposited in the Gulf (Helenes & Carreño 1999; Miller & Lizarralde 2013).

Crustal deformation in the Gulf varies from classic ridge-transform structures in the south to diffuse deformation in the north, where a large number of shallow small-offset normal faults exist (Persaud et al. 2003; Lizarralde et al. 2007). Variations in the crustal structure and composition along the Gulf and beneath Baja are found in several studies. Based on receiver function analysis, Persaud et al. (2007) found a thinning of the crust from western Baja towards the Gulf, and for the southernmost stations (NE76 down to NE79 in Fig. 1) relative to the northern stations (north of NE75 in Fig. 1). Also oceanic-like lithosphere, based on seismic velocities exists beneath the eastern Gulf coastline. This becomes continental lithosphere in Mexico and in Baja (Savage & Wang 2012).

The spatial distribution of seismicity along the plate boundary reflects the fact that the seismic activity is accommodated mostly on the transform fault segments, whereas extension across the plate boundary is accommodated aseismically (Sumy et al. 2013). Widely distributed seismicity located off the ridge axis suggests that a component of Pacific–North America Plate motion is accommodated by deformation within Baja California and the offshore regions (Sumy et al. 2013).

During the last 24 Ma, several types of volcanic lavas were erupted in Baja, the Gulf and mainland Mexico that shed light on the tectonic setting at the time of emplacement. We give here a brief description of the main lavas in the study region following Calmus et al. (2011). Among mafic rocks, regular calc-alkaline lavas are evident mostly from 32° to 24° N in eastern Baja, in the Gulf beneath Isla Tiburon and also close to the coast of central Sonora (Fig. 1). Mid-oceanic ridge basalts (MORBs) are found mainly in the Gulf, beneath the northern sector of the Guaymas Basin and they probably derive from depleted asthenospheric sources. Late Miocene tholeiitic basaltic andesites are found in central western Baja, between 26° and 25.5° N as very fluid flows overlying sedimentary rocks. Similar tholeiitic lavas dated to 6 Ma are also common in northern Baja near ∼31.3° N. Neogene–Quaternary tholeiite to transitional basaltic volcanic rocks are mapped in Sonora, mainly beneath the Pinacate volcanic field just north of the Gulf (∼31.8° N, −113.5° E, Early Miocene), whose origin is probably in the mantle with the involvement of lithospheric material, in the central-western and coastal Sonora (Middle and Upper Miocene); and in the Pinacate and Mocetzuma (∼30° N, −109° E), Quaternary volcanic fields.

Numerous Plio-Quaternary bajasites are found from the Jaraguary (∼29.5° N, −114° E; including Miocene) volcanic field to La Purisima (∼26° N, −115° E; Holocene) in Baja California, with an NNW–SSE trend for ∼500 km. These rocks are magnesianandesites, although their chemical and isotopic composition varies along the Peninsula implying that they may not have a single source. Adakites, igneous rocks rich in silica, are found in Baja from the late Miocene, in the Santa Clara (∼27.5° N, −113° E) volcanic field, also located in the Gulf on Isla Santa Rosalia (∼27.3° N, −112° E) and in Isla Margarita (∼24.5° N, −112° E) in the Pacific, southwest of Baja. Pliocene or Quaternary adakites are found in Isla San Esteban (∼29° N, −113° E), and north and south of Isla Santa Rosalia in the Gulf. Finally, at the same sites as the adakites in Baja, the unusual niobium enriched basalts (NEBs) are found. These NEBs are different from the typical arc basalts because of their high content of niobium and silica. The origin of the listed volcanic rocks is still a matter of discussion, particularly the presence of adakites and NEBs in Baja, whether or not they have the same source since they...
are closely located and if their source is in the asthenosphere or in mantle melts. To summarize, high-magnesium andesites, NEB and adakites are found along the axis of the peninsula, tholeiites and post-subduction calc-alkaline volcanic rocks are found in the Cerro Prieto volcanic field in northern Baja and at other locations along the axis of the peninsula.

In the following sections, we present our group velocity tomographic results and relate them to the regional tectonics and volcanism. We then discuss our interpretation of the shear wave velocity maps derived from the inversion of dispersion data. Finally, we compare our results and interpretations to those published in recent papers.

3 GROUP VELOCITY TOMOGRAPHY

3.1 Method

From the NEIC bulletin, we selected 76 regional events with $4.2 \leq M \leq 6.5$ recorded by the NARS-Baja array during the time period 2002 October 29–2007 October 23 (Fig. 1). In order to improve the ray coverage, we also used recordings at stations from the Southern California Seismic Network (BDR, DPP and GOR in Fig. 1), UNM in Mexico (19.3°N, −99.2°E) and TUC in Arizona (Fig. 1). We did not include other stations, such as the Earthscope stations which for the same time period were located to the north of our study area, because it would not have improved the azimuthal gaps in the ray coverage. In addition, for the same time period as the NARS-Baja stations, only limited data, less than 20 per cent were available from most Earthscope stations considered.

We first computed the dispersion curves for the surface wave paths crossing the region. The period interval of each dispersion curve depends on the magnitude of the earthquake and the source to receiver distance, with longer periods being better recorded from larger events and at longer distances.

The group velocities of the surface wave fundamental mode are computed applying a multiple filter analysis technique (Herrmann (1973) and references therein) to the three-component displacement seismograms at different periods. Herrmann (1973) developed a measurement technique which allows for the separation of the fundamental mode of the surface wave from the higher modes by applying iteratively phase matched filters to the selected waveform. In order to measure group velocity from the three-component seismograms, the quality of the recording is first evaluated in terms of signal-to-noise ratio at each filter period; after demeaning and detrending the waveform, the instrument response is deconvolved from the seismograms to obtain displacements. To minimize errors due to mislocations, we use a minimum ray path length of ~400 km (Hazler et al. 2001).

The determination of the predicted dispersion curves is affected by errors such as noise, earthquake origin times, path coverage and azimuthal anisotropy. Uncertainties due to the density of wave paths and azimuthal coverage can be taken into account a posteriori by applying a smoothing filter to the tomographic maps, as was done in this study. Ritzwoller & Levshin (1998) proved that random mislocations have a minimal effect on the group velocity maps. Although the effect of mislocations is stronger for closer events, it can be reduced when there are many paths crossing the region, which do not originate in the same source region (Ritzwoller & Levshin 1998), as in our case. Besides, when the number of crossing paths is large the smearing of the tomographic inversion is greatly reduced with respect to areas of poor ray path coverage (Pasyanos & Walter 2002).

Group velocity maps are azimuthal averages of group velocity at each gridpoint, but since the azimuthal coverage of rays at each point is not uniform, a correction for azimuthal anisotropy should be done. Several authors studied the seismic anisotropy in the Gulf of California. Obrebski & Castro (2008), using teleseismic receiver functions and shear wave splitting analysis, found that the percentage of anisotropy in the crust is ~10 percent in northern Baja, specifically beneath stations NE71 (+--+) and NE75 (--), as well as beneath NE81 (--~) in the Mexican mainland. Measuring SKS splitting, van Bentheim et al. (2008) obtained little or no anisotropy in the upper mantle beneath southern Baja, except under NE79, at the tip of the Gulf, where delay times of ~1.3 s were observed corresponding to a ~150-km thick anisotropyc layer. Based on phase velocity dispersion measurements in the period range 10–100 s, Zhang et al. (2009) pointed out differences in the azimuthal anisotropy between the northern and southern Gulf. In the Zhang et al. (2009) study, the azimuthal anisotropy is <0.2 per cent at most periods, with higher values in northern Baja, where lower phase velocities are found. In order to determine the reliability of an azimuthal anisotropy calculation in the case of our study, we calculate the azimuthal distribution of ray paths at each gridpoint. For this purpose, following Barmin et al. (2001), ray paths are distributed across a fixed number $n$ of azimuthal bins from 0 to 180°. The resulting histogram of the azimuthal distribution of rays is then used to calculate the function $\chi = \sum_{i=1}^{n} \frac{f_i}{\max f_i}$, where $f_i$ is the density of azimuths within the $i$th bin; $\max f_i$ is the maximum of $f_i$ at a given point, which may be in any of the $n$ azimuthal bins. However, setting $n$ to an arbitrarily fixed value can lead to misinterpretation, since the function $\chi$ would vary greatly according to the value chosen for $n$, independently of the actual number and distribution of rays within the cell. To provide a better estimate of the anisotropy, we slightly modify this relation by setting $n$ to the number of rays in the cell. In this manner, uniform ray coverage (i.e. $\chi = 1$) is reached only when the $n$ rays are equally spaced across $n$ bins, which is, that $f_i = 1/n$ in any bin. Considering that the possible range is $[1/n, 1]$, a perfectly uniform ray path distribution implies $\chi = 1$, whereas $\chi = 1/n$ implies all rays are concentrated along one direction. Barmin et al. (2001) assume that values of $\chi < 0.3$ do not provide a reliable estimate of anisotropy. In their study, $\chi$ is used to damp azimuthal anisotropy in regions with poor azimuthal coverage based on the ray paths at a given point.

In Fig. 2(a), we plot the distribution, $f_i$ of ray paths as a function of azimuth for period 26 s, where the total number of rays is 475. The blue histogram (Fig. 2a) represents the distribution corresponding to the cell with the highest $\chi$ value, despite a moderate number of rays. In contrast, the red histogram shows the distribution in the cell with the highest number of rays, yet the corresponding $\chi$ function is much lower, which indicates less uniform ray coverage (Fig. 2a). In Fig. 2(b), we report the geographical distribution of the maximum values of $\chi$ across all periods (we consider only cells with a minimum number of 10 ray paths). It is clear that over most of the study area $\chi$ does not reach the recommended value of 0.3, except in a few locations shown in yellow in Fig. 2(b). The value of 0.3 is subjective and based on the value used in Barmin et al. (2001). Considering that the azimuthal density is clearly non-uniform (Fig. 2a) and that $\chi$ is well below 0.3 (Fig. 2b), we did not take anisotropic effects into account in the surface wave tomography.

Rayleigh wave group velocities were computed on the vertical component of the displacement waveform. In Fig. 3, we show the number of source to receiver paths versus period for the computed
Rayleigh wave speeds. Even though the number of measurements at very short (10 s) and very long periods (above 100 s) drops below 200, the number of paths at periods between 12 and 80 s (Fig. 3) is suitably dense, and therefore provides bounds on the periods at which the group velocity maps are reliable.

The along-path group velocity measurements for multiple periods are converted into tomographic images using kernels which vary in off-path width as a function of the average velocity and of the period. Off-path rays are weighted according to a cosine function, with a maximum along the master path and a null weight at the boundaries of the kernel band. Because the bandwidth is period-dependent, off-path rays become increasingly effective at longer periods. The study area is gridded in 120 longitude cells by 180 latitude cells, with an equal spacing of 10 × 10 km. This spacing was chosen based on the minimum overall bandwidth, which was measured at the 10-s period. The tomographic reconstruction is done defining the norm
\[
\| \delta g_{p,q} \|_2 = \| G' - G \|_2
\]
with
\[
\delta g_{p,q} = \frac{\sum_k k_i \delta g_i}{\sum_k k_i + \gamma}, \quad \text{and} \quad \delta g_i = g_i - \text{tr}[K_i G_T].
\]
where $G$ is the group velocity matrix, $K_i$ is the kernel weight matrix for the ray $i$, $g_i$ is the group velocity for the ray $i$, $\gamma$ is the damping, $p$ and $q$ are the matrix indices. The matrix $G$ is iteratively updated with $G' = G + \delta G$ until $\|\delta G\| < \varepsilon$ with $\varepsilon$ a user-limit convergence limit. The elements of the kernel weight matrix $K$ are calculated for each ray as

$$K_i = \frac{1}{2} \left[ 1 + \cos \left( \frac{\pi d_i}{\omega} \right) \right].$$

where $d_i$ is the distance from the ray path $i$ to the cell centre, the kernel bandwidth $\omega$ is defined as $\omega = \alpha \cdot T \cdot \bar{g}$ with $T$ the period and $\bar{g}$ the mean group velocity across the mapped area, the term $\alpha$ is a user-defined quantity, that should be smaller than the features we want to resolve, which in turn depend on the ray path distribution. In our inversion, we fix $\alpha = 0.25$. In Fig. 4(a), a schematic representation of the parametrization used in this study and described above is shown, for a couple of stations A and B and two events $i$ and $j$.

We used the least-square algorithm from Paige & Saunders (1982) with lateral smoothing and norm damping to derive tomographic maps at several periods. The amount of smoothing and damping is subjective, so that a good compromise between them can explain the data (Deschamps et al. 2008). Since the velocity distribution is unknown a priori and may differ substantially from the normal distribution, it is important to assign an error on the statistical estimators of the velocity, specifically the mean and standard deviation, in order to give a statistical significance to the velocity field. We computed the mean group velocity (solid lines) and the relative rms (dashed lines) as a function of period at different damping values.

![Figure 3.](image-url) Number of source-to-receiver paths used in the tomographic inversion as a function of period for Rayleigh waves. The horizontal line indicates the period interval, 12–80 s, with the larger number of ray paths.

![Figure 4.](image-url) (a) Schematic representation of the parametrization used in the tomographic inversion of group velocity measurements. (b) Mean group velocity (left axis, solid lines) and relative rms (right axis, dashed lines) as a function of period at different damping values.
It can be observed that while the mean group velocity does not change much, the rms shows large variations with damping. We set the damping value to 0.1 corresponding to the rms curve in blue in Fig. 4(b). This value also allowed reasonable converging times on the tomographic reconstruction, with respect to lower damping values. In addition, the choice of damping and smoothing factors mainly depend on how good they fit the observed data (Deschamps et al. 2008).

It should be noted that the ray coverage may be poor in some regions, in which case little or no statistical confidence can be established. To address this issue, a common practice consists in using the bootstrap method (Efron & Tibshirani 1993), which creates bootstrap samples by randomly sampling and replacing data from the original data set. This is a Monte-Carlo type of algorithm that proceeds in three steps: (1) draw a large number of bootstrap samples from the ray data set using a random number generator, which means that a given ray can be selected once or multiple times, especially for small data sets; (2) for each bootstrap sample, perform the tomographic inversion of the velocity and (3) calculate the mean value and standard deviation across all the reconstructed velocity maps. According to Efron & Tibshirani (1993), 50 to 200 bootstrap samples are adequate. They also demonstrate that little improvement is found past 100, while generally a number of 25 samples give reasonable results. We set our number of bootstrap inversions to 25. In Fig. 5, we report the velocity mean distribution (top panels) obtained from the tomographic inversion of the source-to-receiver ray data set (middle panels). The relative error shown in the middle of Fig. 5 represents the standard deviation recovered from the tomographic inversion of 25 bootstrap samples, expressed in terms of percentage of the mean velocity. This quantity is an estimate of the statistical error on the mean velocity, reaching a maximum on the order of 10–12 per cent at lower periods. With increasing period, the error tends to decrease as a consequence of the larger population of source-to-receiver paths, as demonstrated in the bottom panel of Fig. 5, which reports the number of rays on a per-cell basis.

To evaluate the depth sensitivity of the tomographic maps in Fig. 5, we calculate the Fréchet derivatives of the fundamental-mode Rayleigh waves group velocities with respect to S-wave velocities at different depths (Fig. 6). The Rayleigh wave sensitivity kernels are computed following the method of Herrmann & Ammon (2002) for the Savage & Wang’s (2012) I-D model at the indicated periods. Sensitivity kernels peak at deeper depths with increasing periods, because seismic waves sample progressively deeper structure at increasing periods (for instance, ~50 km at 40 s and 80 km at 80 s). Considering the curves in Fig. 6 and the dense ray coverage of the region shown in Fig. 3, we assume that the average elastic properties of the crust and upper mantle would be well constrained. Keeping in mind the depth sensitivity (Fig. 6), interesting and coherent features observed in the Rayleigh wave group velocity maps (Fig. 5) can be tied to particular tectonic features.

### 3.2 Results: group velocity tomography

In Fig. 5 (top panel), group velocity maps are shown at different periods: 15, 30 and 50 s along with the uncertainties (middle) and the distribution of crossing rays in each cell (bottom). We show the results only in the Gulf region (Gulf of California and Baja, Sinaloa and Sonora regions), excluding areas with few or non-crossing paths. A study by Zhang et al. (2009) has determined a 3-D velocity model for the Gulf of California using the Rayleigh wave phase velocity measurements. Although group and phase velocity are different, because the former refers to the velocity of constructive wave packets while the latter emphasizes the velocity of each harmonic component, they are related in the sense that the group velocity \( U \) depends on the phase speed \( c \) and the variation of the phase speed with the wave number \( k \), \( U = c + k \frac{d}{d k} \). With this in mind and considering the general rule of thumb that at a given period group velocities are sensitive to shallower depths than phase velocities, we compare dispersion maps in Fig. 5 to Rayleigh wave phase velocity anomalies in fig. 4 of Zhang et al. (2009). We note that the depth sensitivity kernels are not shown in Zhang’s et al. (2009) study and we therefore cannot make a more direct comparison to their results.

At 15 s (Fig. 5), the tomographic maps reflect the crustal structure in the study region. At this period, group velocities in the northern and central Gulf are in general slower than in the southern Gulf, where relatively high velocities are indicative of a relatively thinner crust, as also pointed out from the PESCADOR experiment (Lizarralde et al. 2007). We also found low velocities along most of the Baja California Peninsula as well as onshore Sonora, while beneath Sinaloa the group velocities are overall higher. Our results are also in good agreement with the phase velocity maps obtained by Zhang et al. (2009) at 14 s.

At 30 s (Fig. 5), beneath the study area tomographic images confirm the complexity of the deeper lithospheric structures which are in fact less continuous than at shorter periods. At these periods, relatively high velocities are found beneath northern and central Baja and at similar latitudes in mainland Mexico. Starting from the north, a low-velocity zone is present at the border with California and west of the Cerro Prieto Fault (Fig. 1 for locations). A significant low-velocity patch is observed at about 28°N onshore and offshore, near the Guaymas Basin and the Ballenas Transform Fault (BTB; Fig. 1 for locations). Low-velocities zones are also noted in southern Baja, onshore and offshore Sinaloa as well as near the southern coast of Sonora. The tomographic image at 30 s (Fig. 5) is different from the one obtained by Zhang et al. (2009), although there is correspondence with the slow regions we described above (Fig. 5). Using phase velocity data, Zhang et al. (2009) found that the study region at 30 s is homogeneously slow. This contrasts with our results, which show a distinctively heterogeneous lithosphere (Fig. 5). From periods of 30–60 s (Fig. S1 in Supporting Information), fast velocity zones are observed in central-western Baja and beneath the central-southern Gulf.

In general at 50 s (Fig. 5), which samples the upper mantle, the low-velocity areas are more continuous. A distinct low-velocity zone is observed in the centre of the Baja California Peninsula at ~28°N and extends eastwards into the Gulf. An isolated slow patch is found in western Baja centred at ~25°N. This wedge-shaped low is present at all periods and is particularly prominent at longer periods. Also the tip of Baja as well as the southern Gulf in the Alarcón Basin and the Sinaloa coast are slow. At 50 s, the plate boundary in the northern Gulf is relatively slow, while its segment between the Guaymas and Pescadero basins appears fast. There is a significant high-velocity zone roughly between 25° and 27°N beneath the central-southern portion of the Gulf. Also a less pronounced high-velocity region is noted along the southwestern Pacific coast of Baja, just south of 25°N latitude. The map at 50 s in Fig. 5 is in good agreement with the one in Zhang et al. (2009) at corresponding periods, except for the distinct low velocities we observe in western Baja at ~25°N.
Figure 5. Tomographic group velocity maps (top panels) for Rayleigh waves, group velocity uncertainties (middle panels) and the number of crossing rays in each cell (bottom panels) at corresponding periods as indicated in the bottom left corner of each panel. Black triangles are the seismic stations and the thick black line is the plate boundary. Note the change in the group velocity colour scale at each period (top panel).

Going deeper into the mantle (60–110 s in Fig. S1), the velocity pattern becomes relatively slow in the southern part of the study area, from the Farallon to Alarcón basins, although the resolution in this region is reduced due to the fewer ray paths to the west of the East Pacific Rise (EPR). The roughly NW–SE oriented slow region at ~28° N is evident through the whole range of periods. In all cases, it is closely associated with the BTF, the longest active transform fault in the Gulf. Low velocities are also observed at these periods.
beneath northern Baja near the southern California border. The fast velocity region beneath western Baja is again evident at 90 s and longer periods.

We conclude that the major features at a period of 30 s are similar, but in the range 30–50 s our results (Fig. 5) are more complex than those in Zhang et al. (2009), reflecting the laterally varying lithospheric structure in the depth range 30–60 km throughout the study region. In order to give an estimate of the spatial resolution of dispersion maps in Fig. 5, we perform a checkerboard test for the region at all periods. We created a non-standard checkerboard with rectangular (1.5° × 2.25°) checkers (Fig. 7, top panel). The cell size is 0.75°. The longest side of the +5 per cent checkers (blue) is aligned in the latitudinal direction and shortest side of the −5 per cent checkers (red) is aligned in the longitudinal direction. This creates a pattern that is not aligned with the Gulf coastline and provides a reliable test of the lateral resolution of our model both parallel to and perpendicular to the coastline. Results of the tomographic inversion (Fig. 7 middle panels) show what is expected from the ray path distribution (bottom panels). There is a good resolution where the number of ray paths is denser, while in regions with a few crossing rays the resolution is reduced especially at shorter periods, for instance in the Vizcaino Peninsula (central-western Baja, ~28° N, −115.5° E), beneath the EPR, and to the east of the Cerro Prieto Fault (Fig. 1 for location).

4 SHEAR WAVE VELOCITY FROM THE INVERSION OF THE DISPERSED SPEEDS

4.1 Method

For each period, the tomographic maps shown in Fig. 5 are used to solve for group velocity variations by assuming that each gridpoint is laterally homogeneous. Following Rodi et al. (1975), the group velocity is a function defined as

\[ U = f \left( c, \frac{dc}{dT}, \omega, m \right) , \]

where \( m \) is a model parameter that can take the values \( V_p, V_s, \) layer thickness or inverse attenuation \( Q, \omega \) is the angular frequency, \( T \) is the period. Considering that \( c(\omega, m) \) is a continuous function, the required partial derivative with respect to period of the group velocity depends on the partial derivative of the phase velocity, which in turn depends on the partial derivative with respect to model parameter \( m \).

We use the linearized least-squares inversion method by Herrmann & Ammon (2002), based on a 1-D velocity model derived from the 3-D model in Savage & Wang (2012), specifically developed for the Gulf of California Extensional Province. This model includes information on the sedimentary and crustal layers determined in recent studies (see Savage & Wang 2012 for details). The initial crustal model as well as the thickness of sediments varies laterally depending on whether the crust is continental or oceanic. The inversion procedure is based on taking the starting velocity model, calculating the predicted dispersion curve, and comparing it to the observed dispersion. The starting model is then iteratively refined until it fits the observations. Iterations are controlled to avoid artificial features such as spurious low-velocity zones. We invert for both the layer thickness and velocity and each layer is equally weighted.

The inversion code inverts for the S-wave velocity and then updates the P-velocity using the \( V_p/V_s \) ratio of the initial model; the new density is computed from the new \( V_p \) using the Nafe-Drake relation. We did not consider any physical dispersion corrections.

The least-squares inversion includes a damping factor that limits the range of variations in the model between two subsequent iterations. The damping factor affects how fast the model converges and minimizes residuals between the observed and predicted values. In this study, we found that a damping factor of 3 gives the most stable results with a reasonable number of iterations (20 maximum) when inverting the Rayleigh wave group velocities.

The errors associated with the group velocity tomographic maps described in Section 3.1 reduce the accuracy of the shear wave velocity and crustal thicknesses obtained from this inversion, but a careful selection of the data set helps in minimizing these effects. In this inversion, we neglect the effects of anisotropy (see Savage & Wang 2012 for details) in favour of determining the average lateral variations in velocity at the regional scale.

In determining the crustal structure from the surface wave tomography, there is a significant trade-off between depth and velocity (Pasyanos & Walter 2002). In order to reduce this trade-off, the group velocity measurements need to be accurate, systematic errors over a broad range of periods have a larger effect than random errors that vary rapidly with period (Lebedev et al. 2013). In addition, \textit{a priori} constraints on crustal and mantle structure, and a large number of measurements at short and long period across the study region, will reduce the velocity–depth trade-off (Lebedev et al. 2013). With this in mind, we think that the use of the velocity model specifically for the Gulf of California region by Savage ...
& Wang (2012) as a starting model for the inversion of the group velocity data is a good choice to reduce the trade-off. Finally, following Pasyanos & Walter (2002), we note that the average crustal velocity is better constrained than the velocity in individual layers, and also surface waves are sensitive to average shear wave velocities so they do not adequately resolve sharp or step-like discontinuities.

In order to show that shear wave velocity maps do not depend on the initial velocity model, we plot in Fig. 8(a) an example of
Figure 8. (a) Fitting of the dispersion data to the predicted Rayleigh wave group velocities (right-hand panel) using two different initial models, CRUST2.0 (Bassin et al. 2000) over IASP91 (Kennett & Engdahl 1991) for the mantle structure (black) and Savage & Wang’s model (2012, red). In the left-hand panel, the obtained (solid) shear wave velocities versus depth are plotted with respect to the initial models (dashed). (b) Resolution matrix of Rayleigh wave group velocities computed in the period range 5–150 s for the 1-D S-wave velocity model shown in Fig. 8(a) (Savage & Wang 2012). The velocity model consists of 39 layers each 5-km thick over a half-space and the damping in all the 30 iterations was 3. The linearized inversion was performed using Herrmann & Ammon’s code (2002, see text for more details). In Fig. 8(b), left-hand panel, the final inverted model is shown; in Fig. 8(b), right-hand panel, each column corresponds to a specific layer so that the last column of the resolution matrix is referred to the bottom half-space layer.
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4.2 Results: shear wave velocity maps

In Fig. 9, we show maps of the shear wave velocities averaged over different depth intervals. In the range 10–20 km, the overall shear wave velocities are lower than 4 km s$^{-1}$ along the peninsula and the western coast of Mexico. There are two main low-velocity regions beneath Baja which are also present in the group velocity maps at 15 s: the circular (~200 km in diameter) and more prominent low-velocity patch is centred at ~28°N and extends across the peninsula; a secondary slow region is observed in the central-southern Baja (~26°N, ~112°E). Along the Gulf, the S-wave velocities increase from north to south and reach the highest values at the mouth of the Gulf and beneath the EPR. This pattern of high velocities beneath the Gulf could likely reflect the differences in the sediment thickness in the northern basins with respect to the southern ones.

In the depth interval 20–30 km (Fig. 9, upper right panel), we are likely imaging crustal velocities and an overall correspondence can be found with the group velocity tomographic map at 30 s in Fig. 5. Although in the Gulf generally shear velocities are lower than at shallower depths, two main low-velocity areas are noted: the northernmost one, which underlies the northern Gulf (Wagner Basin–Salton Trough), but is also located along the Cerro Prieto Fault and east of the Agua Blanca Fault, a shear zone that extends across the northern peninsula towards the Pacific (Fig. 10 for location). The second one underlies the BTF starting just south of Isla Angel de la Guarda and Tiburon Island and reaches into the Guaymas Basin from the north (Fig. 10 for location). This may indicate hotter or weaker crust north of the Guaymas Basin. This low-velocity anomaly in the crust extends at least to 90 km and again from ~130 to ~170 km (Fig. S2). At intermediate depths, in the range ~40–120 km (Fig. 9), this low-velocity patch turns into an NW–SE trend and with increasing depths, shifts northwards along the BTF, west of Tiburon Island, reaching the eastern coast of the peninsula. This as well as another low-velocity region are also imaged in a NW–SE cross-section (Fig. 11) extending from Baja towards Sinaloa, and roughly parallel to the BTF (red dashed line in Fig. 10). In Fig. 11, the two main slow regions beneath Sinaloa and the BTF extend to, respectively, 80 and 110 km, and their borders are 300–400 km apart at a depth of 40 km.

At mantle depths (>30 km, middle panels in Fig. 9), low S-wave velocities characterize the edges along the southern Gulf, along coastal Sonora and Sinaloa. There are three distinct elongated velocity highs located in the southern basin and Range of Sonora at depths of 30–40 km (left-central panel in Fig. 9). These N–NW trending features are also present in the 15 and 30 s group velocity maps (Fig. 5). The anomalously low velocities at the mouth of the Gulf beneath the Alarcón and Pescadero basins, in and off-ridge axis, extends to the EPR, is L-shaped at depths greater than 50 km and has the largest areal extent of the above-mentioned slow regions (Fig. 11). These slow patches are also shown in the E–W cross-sections of Fig. 12 at different latitudes, in the depth interval 40–70 km. Previous workers (Wang et al. 2009) have interpreted low shear velocities spaced ~250 km apart beneath the Gulf as dynamic, buoyancy-driven upwelling and melting. Their anomalies were however, located beneath the Wagner, Delfin and Guaymas basins, with no coverage in the southern Gulf, and depending on the depth, the anomalies were off the rift axis. We find a similar anomaly for the Wagner Basin and beneath the northern part of Delfin Basin at 20–30 and 50–90-km depths. Our central Gulf anomaly is centred more on the southern segment of the BTF than on the Guaymas Basin in the depth range 20–40 km. In addition, anomalies beneath the BTF and beneath the Alarcón and Pescadero basins are spaced roughly 300–400 km apart (Fig. 11).

At 30°N and southwards to 24°N, a fast velocity zone is distinct in the central-southern Baja Peninsula or offshore Pacific at depths larger than 30 km (Fig. 9). More precisely, it does not reach the tip of the peninsula and the Gulf, and is located between the two northernmost slow patches discussed above. This fast velocity is observed at most depths until at least ~90 km and is less pronounced at ~90–130-km depth (bottom panels, Fig. 9). A relatively high-velocity anomaly was also found by Zhang et al. (2009) at depths between 120 and 160 km. Considering the distribution of volcanic rocks in Baja, Wang et al. (2013) interpret this fast velocity anomaly as a fossil slab attached to the unsubducted Magdalena microplate. The fast velocity zone is also evident in the east–west profiles in Fig. 12 at depths between 40 and 70 km.

We also want to point out the prominent high-velocity region (Fig. 9) in the depth interval 50–90 km west of the Baja Peninsula and south of ~25°N which extends down to 90–130-km depth. Based on our coverage, this feature is not continuous with the fossil slab we interpret to the north, it is deeper than that and smaller in lateral extent. Using active source seismic reflection and wide-angle seismic refraction profiles across southwestern Baja California (~24.5°N), Brothers et al. (2012) infer an oceanic slab detachment beneath the western margin of Baja ~40 km landwards of the extinct Farallon–North America trench. While we do not have good resolution as far west as Brothers et al. (2012), we do observe a high-velocity anomaly starting at around ~50-km depth near the western Baja coastline and beneath Baja (Fig. 12) that may represent another remnant of the Farallon slab in this region.

5 DISCUSSION

For a better understanding of the significance of the observed anomalies in the study region, it is important to take into account the distribution of the volcanism. The localized low-velocity
Figure 9. Average shear wave velocity maps obtained from the inversion of the dispersion data using Savage & Wang’s model (2012) in different depth intervals as indicated at the top of each panel. Note the difference in colour scale for each panel. Black triangles are the seismic stations and the black line is the plate boundary. The fossil slab in central western Baja and the slab detachment south of 25°N beneath Baja are outlined in blue and the slow shear wave velocity zones in red in the right-central panel.
organized in an extension-orthogonal (along strike) direction. In particular, tholeiitic volcanism or MORB are associated with the Cerro Prieto volcanic field, the Guaymas Basin (Isla Tortuga and Isla Esteban, which are close to the Guaymas Basin) and the EPR.

The more anomalous volcanic rocks shown in yellow and green in Fig. 10 are for the most part approximately found above the eastern edge of the interpreted slab remnant (blue area in Figs 9, 10 and profiles at 25.8°N and 26.8°N in Fig. 12), which is located at shallow depths in the mantle. Previous studies (Zhang et al. 2009; Wang et al. 2012) found the presence of small-scale melt pockets. The eruption of these patches in Baja (Fig. 9, 50–90 km) could be associated with the presence of small-scale melt pockets. The eruption of these pockets, likely triggered by local changes (for example, regional stress changes), caused the formation of the unusual post-subduction Baja California lavas such as magnesian andesites, adakites, tholeiitic lavas and high NEBs (Negrete-Aranda & Canón-Tapia 2008; in green in Fig. 10). The larger low shear wave velocity regions in central-eastern Baja and the Gulf near the northern portion of the BTF (Fig. 10) coincide with the area occupied by mainly calc-alkaline and MORB-like lavas, respectively, as described above. Previous tomographic studies (van der Lee & Frederiksen 2005; Wang et al. 2009; Zhang et al. 2009; Savage & Wang 2012) found the low shear wave velocities lower than the global averages beneath the Gulf with respect to reference models. This is also confirmed in our study and in particular at deeper depths (~90–130 km, bottom panels in Fig. 9) the shear wave speeds are well below 4.5 km s⁻¹, which is the shear wave velocity at 130-km depth in IASP91 model (Kennet & Engdahl 1991). Low velocities are also observed at depths at least down to ~170 km (Fig. S2), although the reliability of those results is limited due to the reduced number of ray paths at periods longer than 100 s.

The large low-velocity zones beneath the Gulf likely reflect the upwelling of asthenospheric material in the upper mantle (Lizarraide et al. 2007) and this is predominant in the central and southern Gulf (Figs 9, 11 and 12, profile at 24.2°N). Numerical models of continental break-up (Corti et al. 2003) demonstrate that this process is characterized by a first stage of extension affecting the entire rift length and a second stage in which asthenospheric upwelling occurs in regularly spaced well-confined regions propagating in an extension-orthogonal (along strike) direction. In particular, tholeiitic volcanism or MORB are associated with the Cerro Prieto volcanic field, the Guaymas Basin (Isla Tortuga and Isla Esteban, which are close to the Guaymas Basin) and the EPR.

Figure 10. Schematic map showing the main Neogene and Quaternary volcanic rocks from Negrete-Aranda & Canón-Tapia (2008): high-magnesium andesites, Niobium-enriched basalts and adakites in green; tholeiites and post-subduction calc-alkaline volcanism in yellow. The question mark in the NW corner of the interpreted fossil slab (shaded in blue) indicates that in our study we cannot establish the lateral continuity of the slab, as well as whether or not it extends past the peninsula into the Gulf. Similarly the question mark in the area marked south of 25°N (the southernmost blue area) indicates a possible slab remnant beneath Baja. Low shear wave velocity regions are shaded in red. Dashed red line is the NW–SE profile in Fig. 11, whose coordinates are (30°N, −115°E) and (22.5°N, −106°E), while the dashed blue horizontal lines indicate the four profiles shown in Fig. 12. ABF = Aqua Blanca Fault. Other main tectonic features are labelled as in the caption of Fig. 1. Black triangles are the seismic stations.
Figure 12. Average shear wave velocities along east–west profiles (dashed blue lines in Fig. 10) at different latitudes as indicated in left bottom of each panel in the depth interval 40–70 km. The average is taken over three grid cells in and out of the plane of the profile. Contours are at 3.8, 3.9, 4.0, 4.1, 4.2, 4.3, 4.4, 4.5 and 4.6 km s\(^{-1}\). Region boundaries are marked along the top of each profile.

well as whether or not it extends past the peninsula into the Gulf is not unambiguous in our model hence a question mark in Fig. 10. We however note the extent of this high-velocity feature in the E–W profiles shown in Fig. 12 (top three panels). A magnetotelluric survey at a latitude of 28°N across the peninsula imaged the slab at depths of 30–40 km (Romo-Jones 2002). Receiver function analyses also suggested a possible slab top close to the Moho beneath station NE75, with an anisotropic layer at around ~100 km (Persaud et al. 2007). In addition, numerical models show that shearing in a mantle wedge above the slab beneath the peninsula could produce the unique compositional variation in post-subduction volcanism found in Baja (Negrete-Aranda et al. 2013). More importantly, the timing of melt production from these numerical models matches geochemical observations.

Calmus et al. (2011) suggest that the spatial and temporal distribution of volcanism in Baja is consistent with the development of a slab-tear evolving into a 200-km wide slab window parallel to the trench extending from the Pacific coast across Baja to coastal Sonora. Tholeiites and alkali basalts of subslab origin rose through this window, while the adakites were derived from the partial melting of its upper lip, close to the trench. Pallares et al. (2007) report that the magmatic activity continued after the cessation of the subduction (11.5–7.5 Ma) in a 600-km long and ~100-km wide array, parallel to the trench, emplacing adakites, NEBs and andesites. Castillo (2008) instead proposes that the Pacific asthenosphere was the direct source for post-subduction magmas that erupted in Baja California, considering the chronology of magmatism and the isotopic composition of lavas. A low-velocity zone beneath Baja at ~29°N (Fig. 9, 50–90-km depth, and Fig. 12, profile at 29°N) which may connect with the low associated with the BTF could be interpreted as a slab window as shown in Fig. 2 in Pallares et al. (2007).

The three velocity highs in Sonora at depths of 30–40 km (Fig. 9, left-central panel) may be associated with mafic intrusions at lower crustal depths that were emplaced in the wide core complex belt post 30 Ma and have since solidified. Alternatively, they may reflect differences in lower crustal composition of separate terranes.
or suture zones as they appear to be aligned with the second vertical derivative of Magmat anomalies used to map terrane boundaries in the region (Campos-Enriquez et al. 2005). We note that the lower resolution of the Magsat data (150–300 km) relative to our shear wave velocity model does not allow all components of the lows and highs in that data set to be distinguished. In addition, there is only very limited constraint on crustal thicknesses in this region. With these limitations in mind, following on the work of Campos-Enriquez et al. (2005), we interpret the lower crust of the Caborca terrane (Fig. 10) of northwestern Sonora as composed of three domains. Two are associated with the northernmost velocity highs in northern Sonora at 30–40-km depth (Fig. 9). The third represents a low between these two velocity highs, possibly related to the Mojave-Sonora Megashear (MSM; Fig. 10) or a suture zone. Noteworthy is the MSM correlates roughly with the boundary between a high and low in the second vertical derivative of the Magsat data (Campos-Enriquez et al. 2005). The presence of Caborca-type basement 40-km north of the MSM has been documented by Amato et al. (2009) and supports our interpretation above, although the specific location and geological significance of the MSM remains controversial in regional geology. Station NE80 (Fig. 1 for location), which has a Moho depth of 32.1 km (Persaud et al. 2007) is located at the northern end of the easternmost velocity high which extends up to shallower crustal levels of ~20 km near latitude 30°, suggesting that the velocity anomaly may indeed be a part of the lower crust rather than the mantle. At depths of 20–40 km, we interpret the southernmost velocity high in Sonora as associated with the Seri Terrane or southern segment of the Caborca Block. It should, however, be noted that station NE82 located at the southern end of this anomaly has a Moho depth of 25.9 km (Persaud et al. 2007). In this case, the deeper shear wave high would instead be associated with the upper mantle. The low velocities extending from the eastern edge of the Gulf across Sinaloa (left-central panel in Fig. 9) correlate with a minor high in the Magsat data around the same location (Fig. 3 in Campos-Enriquez et al. 2005). This low-velocity feature is discontinuous below ~45–50-km depth where it instead becomes a velocity high (bottom two panels in Fig. 12) in Sinaloa, but continues to depths below 70 km at the eastern border of the Gulf near the Alarcón Basin. We expect that the high in the Magsat data is associated with the high velocities below ~45 km. In the Gulf, we interpret this low-velocity feature as a region of mantle upwelling as previously described above.

6 CONCLUSIONS

We have shown that Rayleigh wave group velocity measurements allow us to determine important information on the Earth structure in the tectonically complex Gulf of California region. We have found heterogeneous lithosphere in the group velocity maps at 30 s, but complex tectonic features are also mapped at lithospheric and asthenospheric depths (periods greater than 60 s). The main results of this study are: overall shear wave velocities slower than the average in reference models; low-velocity regions in the Gulf of California associated with mantle upwelling at lithospheric and asthenospheric depths; a low-velocity zone in north-central Baja at ~28°N and extending east–south–eastwards to the coast of Sonora, likely associated with the presence of an asthenospheric window; three main velocity highs in Sonora in the depth range 30–40 km; and a well-confined high-velocity zone in the upper mantle beneath central-western Baja California related to a fossil slab.

This study differs from previous studies in the same area because we use group instead of phase velocity data, and these are more sensitive to shallow structure than phase velocity at corresponding periods. Since variations in crustal structure greatly influence the propagation of regional phases and consequently earthquake locations, it is very important to have accurate information on crustal shear wave velocity. Our study contributes a reliable velocity model for the Gulf of California region. Complexities in the lithospheric structure beneath the study region are pointed out in our shear wave velocity maps as well as their association with the main types of volcanic lavas in the area.

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**SUPPORTING INFORMATION**

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

**Figure S1.** Tomographic dispersion maps for Rayleigh waves at periods from 60 to 110 s. Corresponding periods are labelled in the bottom left corner of each panel. Black triangles are the seismic stations and the black line is the plate boundary. Note the different colour scale for different periods.

**Figure S2.** Shear wave velocity at several depth intervals as indicated at the top of each panel. Black triangles are the seismic stations and the black line is the plate boundary. (http://gji.oxfordjournals.org/lookup/suppl doi:10.1093/gji/ggu338/-/DC1).

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