

Effects of shallow density structure on the inversion for crustal shear wave speeds in surface wave tomography

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SUMMARY

Surface wave tomography routinely uses empirically scaled density model in the inversion of dispersion curves for shear wave speeds of the crust and uppermost mantle. An improperly selected empirical scaling relationship between density and shear wave speed can lead to unrealistic density models beneath certain tectonic formations such as sedimentary basins. Taking the Sichuan basin east to the Tibetan plateau as an example, we investigate the differences between density profiles calculated from four scaling methods and their effects on Rayleigh wave phase velocities. Analytical equations for 1-D layered models and adjoint tomography for 3-D models are used to examine the trade-off between density and *S*-wave velocity structures at different depth ranges. We demonstrate that shallow density structure can significantly influence phase velocities at short periods, and thereby affect the shear wave speed inversion from phase velocity data. In particular, a deviation of 25 per cent in the initial density model can introduce an error up to 5 per cent in the inverted shear velocity at middle and lower crustal depths. Therefore one must pay enough attention in choosing a proper velocity–density scaling relationship in constructing initial density model in Rayleigh wave inversion for crustal shear velocity structure.

Key words: Interferometry; Surface waves and free oscillations; Seismic tomography; Computational seismology.

1 INTRODUCTION

The dispersion of surface wave phase velocities has long been utilized to image the crust and upper mantle shear velocity structure (e.g. Shapiro & Ritzwoller 2002; Shapiro *et al.* 2005; Yao *et al.* 2006; Lin *et al.* 2007; Bensen *et al.* 2009; Yang *et al.* 2012). Moreover, development in surface wave tomography has been accelerated by wave dispersion data sets derived from teleseismic earthquakes and seismic interferometry, that is empirical Green's functions (EGFs) of surface waves extracted from cross correlations of ambient noise (Shapiro *et al.* 2005) or diffuse earthquake coda (Campillo & Paul 2003). In particular, EGFs extracted from seismic interferometry, which serve as complementary data to earthquake recordings, have two major advantages: (1) EGFs compensate for limited data coverage due to uneven earthquake distribution; (2) EGFs avoid the source property uncertainties in earthquake studies (Yang *et al.* 2007).

Traditional surface wave tomography based on seismic interferometry generally consists of the following steps. First, Love or Rayleigh wave phase or group velocity dispersion curves are measured from cross-correlation of each station pair. Subsequently, 2-D

tomography based on ray theory is adopted to generate phase or group velocity maps in a study area. Finally, local dispersion curves at individual grids are extracted from these maps and then inverted for individual 1-D shear velocity profiles to compose 3-D velocity model. This multistep analysis has been widely employed in surface wave studies at various local or regional scales (Lin *et al.* 2007; Bensen *et al.* 2009; Yang *et al.* 2012).

Surface wave phase or group velocities are more sensitive to *S*-wave velocity (β , V_s) than *P*-wave velocity (α , V_p) and density (ρ). The sensitivity to density seems to decrease with increasing periods, and becomes nearly negligible for long-period earthquake data. Also, α and ρ are proportional to β for crustal and mantle rocks. Therefore most of the surface wave studies only invert for β and compute α and ρ by using an assumed V_p/V_s ratio and an empirical velocity–density scaling relationship derived from experiment data of crustal and mantle rocks.

Four empirical scaling relationships have been widely used, which are: (1) the constant velocity–density scaling relationship (referred to as CM in this paper), $\rho = 0.81\beta$ (Christensen & Mooney 1995; examples of studies: Bensen *et al.* 2009; Yang *et al.* 2012); (2) the Gardner's rule, $\rho = 1.74\alpha^{0.25}$ (Gardner *et al.* 1974; examples of

studies: Giancarlo 2010; Cauchie & Saccorotti 2013); (3) the Birch's law, $\alpha = 3.125 \rho - 2.400$ (Birch 1961; examples of studies: Li *et al.* 2012; Guo *et al.* 2015) and (4) the Nafe-Drake relation (Ludwig *et al.* 1970) represented by a 5th order polynomial (Brocher 2005; examples of studies: Yao *et al.* 2008; Huang *et al.* 2014):

$$\rho = 1.6612\alpha - 0.4721\alpha^2 + 0.0671\alpha^3 - 0.0043\alpha^4 + 0.000106\alpha^5. \quad (1)$$

In the remainder of this paper, we will refer to these four scale relations as the CM, Gardner, Birch and Nafe-Drake velocity–density scaling models.

The velocity–density empirical relationship has been used in dispersion data inversion in two ways. It is first used in constructing the initial density model, that is using an absolute velocity model to compute the absolute values of the density profile, which is referred to absolute velocity–density scaling here. Some studies also use the S -wave velocity updates to scale density changes during the inversion. Here since scaling is performed in terms of perturbations, that is $\delta \ln \rho = k \delta \ln \beta$, where k is a constant, we refer to it as the perturbation or relative scaling relationship. Some studies fix the density profile during the inversion (e.g. Yang *et al.* 2012), which means that $k = 0$ in the relative density–velocity scaling. The V_p/V_s scaling is also used similarly in both constructing the initial P -wave velocity model and in updating the velocity models during the inversion.

As mentioned above, traditional surface wave tomography requires intermediate steps of measuring dispersion curves, projects them onto a set of grids by using a 1-D ray path approximation, and then inverts grid-based dispersion data by assuming a 1-D layered model. Both the approximation and assumption could affect the inverted 3-D velocity structure. Recently, Chen *et al.* (2014) utilized the adjoint tomography technique that formulates the inversion by a direct matching of the surface waveform data with synthetics of 3-D models, without performing the intermediate steps. The initial density model is also computed by using the Nafe-Drake scaling, which is iteratively updated with a relative scaling of $\delta \ln \rho / \delta \ln \beta = 0.33$. Thus the issue of density–velocity and V_p/V_s scaling also exists in the adjoint surface waveform tomography here.

Recently, the advent of the interferometry method brings the short-period data into the surface wave community and these kind of data are sensitive to densities and velocities with comparable magnitude at shallow depths (Yang *et al.* 2012). Thus it is necessary to investigate whether and how different scalings could affect the final S -wave velocity model in the conventional and adjoint tomographic inversions. To do so, we first calculate the sensitivity kernels of phase velocities to the S -, P -wave velocities and density using an analytical equation derived from the variation principle (Aki & Richard 2002). We find short period phase velocities are indeed very sensitive to density structures at shallow depth. We further generate a set of synthetic dispersion and waveform data, and conduct 1-D and 3-D inversions by using different initial models with a wide range of density and P -wave velocities. We also investigate the effect of different perturbation scalings ($\delta \ln \rho / \delta \ln \beta$, $\delta \ln \alpha / \delta \ln \beta$) on the final S -wave velocity model.

We find significant trade-offs exist between shallow-depth density and mid-to-lower crust shear wave speeds in ambient noise tomography. More specifically, the absolute density–velocity scaling used in constructing the initial model is the most important scaling relationship that can significantly affect the final velocity model, while the effects of the perturbation scaling, $\delta \ln \rho / \delta \ln \beta$, and the assumed V_p/V_s ratios used in the inversion seem to be very limited.

2 FORWARD MODELLING

Our modelling region includes the eastern margin of the Tibetan plateau and the adjacent Sichuan basin (Fig. 1a). We choose this region because it contains two different types of tectonic units, plateau and basin, which have very distinct crustal and uppermost mantle structures (Yao *et al.* 2008; Yang *et al.* 2012; Chen *et al.* 2014). We set up a total of 97 virtual stations/sources in the region, which have the same geographic distribution as the broadband stations of the CEArray and IRIS/PASSCAL arrays used in a previous tomographic study (Yang *et al.* 2012; Fig. 1b). The dense ray coverage allows to conduct both 1-D and 3-D inversions to investigate the scaling effects.

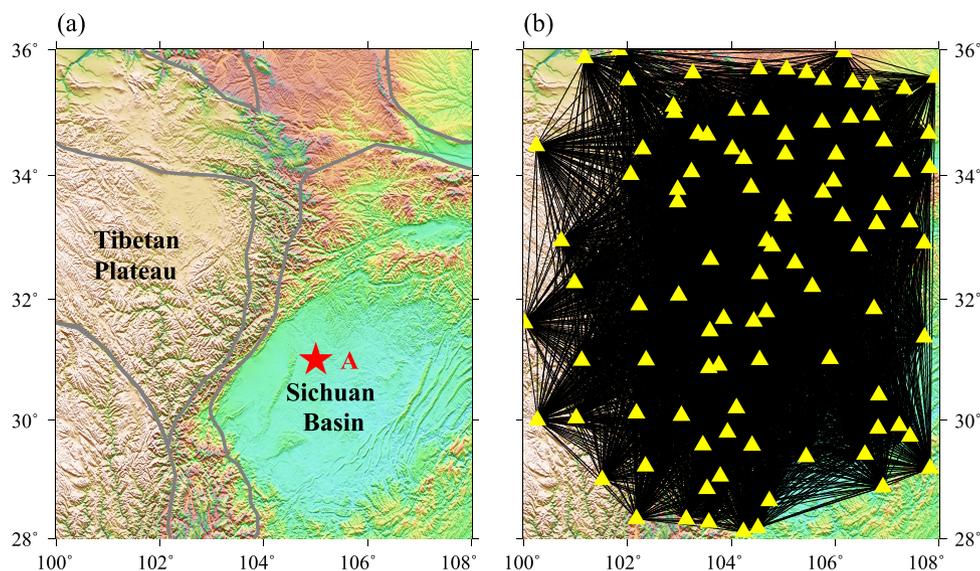


Figure 1. (a) The topography of the study area. Grey lines delineate the boundaries of tectonic blocks. Red star marks the gridpoint A at 105°E and 31°N, which is used to illustrate the density effect on Rayleigh wave phase velocity and 1-D shear wave speed inversion. (b) Station and ray path coverage for the 3-D wave speed inversion. Yellow triangles denote the 97 stations used in 3-D shear wave speed inversion. Black lines represent interstation great-circle ray paths.

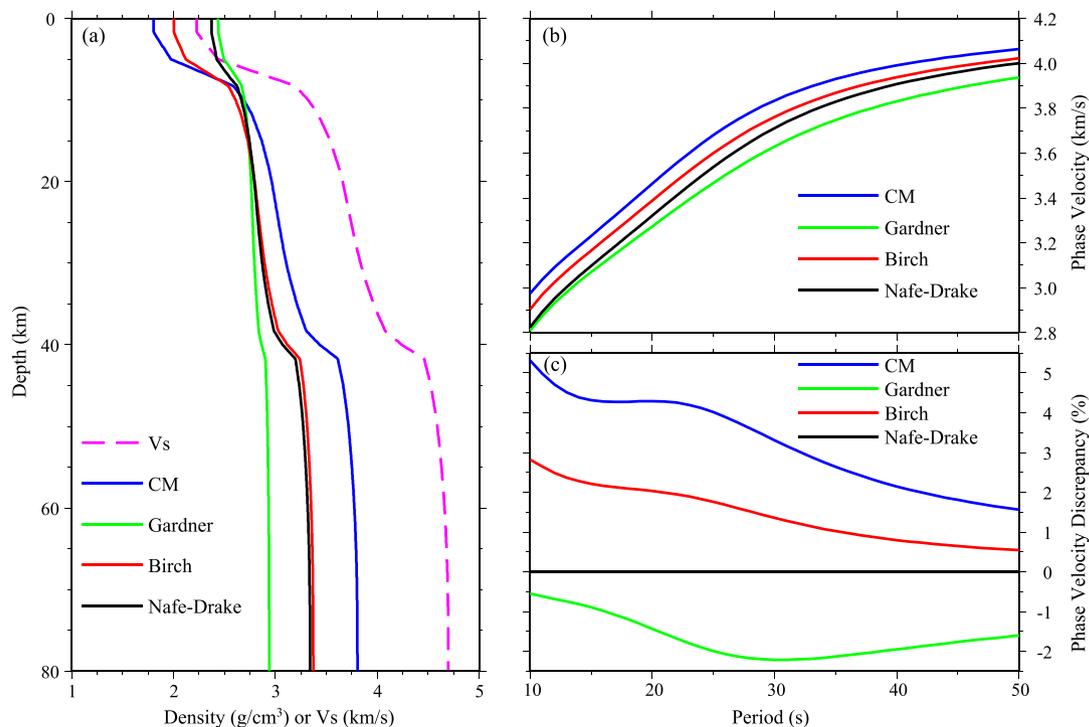


Figure 2. (a) The magenta dashed line denotes the shear wave speed profile at gridpoint A in the Sichuan basin obtained from Yang *et al.* (2012). The solid lines of different colours indicate the 1-D density depth profiles calculated from shear wave speeds using the CM, Gardner, Birch and Nafe-Drake relations. (b) Rayleigh wave phase velocities in the period range from 10 to 50 s computed with the four velocity–density models shown in (a). (c) Differences in percentage among the phase velocities shown in (b), all plotted relative to the one using the Nafe-Drake density models.

2.1 Rayleigh wave phase velocities of different 1-D density models

To probe how different density models affect Rayleigh wave phase velocity dispersion, we select the gridpoint A (105°E, 31°N; red star in Fig. 1a) in the Sichuan basin with a thick sediment layer (~8 km; Laske *et al.* 2013) as an example. Based on the shear velocity profile produced by Y12 (abbreviation for Yang *et al.* 2012), we generate four density profiles with the four empirical scaling models mentioned above (Fig. 2a). The four empirical scaling relationships were built based on rock samples with different origin depths, each scaling is thus expected to apply to certain depth range better than the others. For example, the Birch scaling model is more appropriate for deeper and denser rocks while the Nafe-Drake model is more reliable for sedimentary rocks (Maceira & Ammon 2009). The two models exhibit a ~18 per cent difference at shallow depths, but are consistent below the sediment layer. The Gardner's rule is established for studying shallow structure (eg: Giancarlo 2010; Cauchie & Saccorotti 2013), therefore it matches the Nafe-Drake model at shallow depths but has a ~12 per cent lower density in the upper mantle than the Birch predicts. Compared to the Nafe-Drake density model, the CM model has a much lower density (up to ~25 per cent) at shallow depths and a higher density (~13 per cent) in the middle to lower crust and the upper mantle. Thus, it seems that the Nafe-Drake model gives better density values over the entire depth range for the point A inside the Sichuan basin.

We adopt the analytical equations of the Thomson-Haskell method (Aki & Richards 2002) to calculate Rayleigh wave dispersion curves in the period range of 10–50 s using the above four density models (Fig. 2b). The four models have the same 1-D shear and compressional wave speed ($\alpha = 1.732\beta$) profiles extracted from

Y12. As mentioned above, the Nafe-Drake density model seems to fit the density structure beneath the Sichuan basin in the entire crustal range better than the other three models, we thus use its dispersion curve as the reference to make comparisons. Overall, the calculated phase velocities from the Birch and CM models are higher than those of the Nafe-Drake model over the entire period range of 10–50 s, while the Gardner phase velocities are generally smaller than the Nafe-Drake predictions (Fig. 2b). The corresponding percentage difference is ~+1.5 to 5.5 per cent, ~+0.5 to 3.0 per cent and ~−2.1 to −0.5 per cent for the CM, Birch and Gardner models, respectively. Interestingly, although the Birch and Nafe-Drake models have nearly the same density structure at depths >5 km below the surface, the corresponding differences in phase velocity of the two models are pretty large, especially at the short-period range (Fig. 2c). On the other hand, the short period phase velocities between the Gardner and Nafe-Drake models only differ by <1 per cent despite the large density difference between the two models in depth greater than 5 km (Fig. 2c). Based on Fig. 2, we conclude that there are substantial differences in the phase velocities calculated from the four density models, and shallow density structure appears to have significant influence on the short period phase velocities.

2.2 The depth sensitivity kernels of phase velocity to V_p and density

To better understand how shallow density structure affects phase velocities, we represent the ‘partial derivatives’ equation of Rayleigh wave phase velocity, which is shown as a function of the two Lamé constants in Aki & Richards (2002), in terms of P -, S -wave

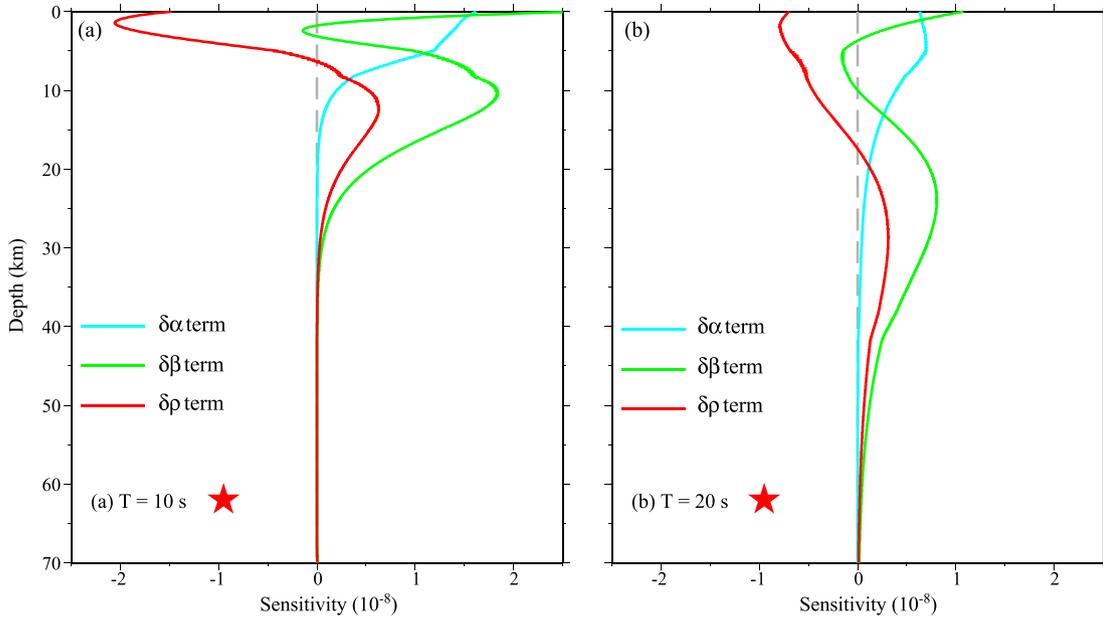


Figure 3. (a) Depth profiles of 1-D phase velocity sensitivity kernels at gridpoint A, calculated based on eq. (2) for the 10-s Rayleigh wave, with respect to $\delta\alpha$ (cyan), $\delta\beta$ (green) and $\delta\rho$ (red), respectively. Grey dashed line indicates zero sensitivity at all depths, which is used for reference. (b) Same with (a) but for the 20-s Rayleigh wave.

velocities and density:

$$\begin{aligned} \left(\frac{\delta c}{c}\right)_{\omega} = & \frac{1}{2k^2 U_c I_1} \left\{ \int_0^{\infty} \rho \alpha \left(kr_1 + \frac{dr_2}{dz} \right)^2 \delta\alpha \, dz \right. \\ & + \int_0^{\infty} \rho \beta \left[\left(kr_2 + \frac{dr_1}{dz} \right)^2 - 4kr_1 \frac{dr_2}{dz} \right] \delta\beta \, dz \\ & + \int_0^{\infty} \left[\frac{1}{2} \alpha^2 \left(kr_1 + \frac{dr_2}{dz} \right)^2 + \frac{1}{2} \beta^2 \left(kr_2 - \frac{dr_1}{dz} \right)^2 \right. \\ & \left. \left. - \frac{1}{2} \omega^2 (r_1^2 + r_2^2)^2 - 2kr_1 \frac{dr_2}{dz} \beta^2 \right] \delta\rho \, dz \right\}, \quad (2) \end{aligned}$$

where ω denotes the angular frequency, k the corresponding wavenumber, c the Rayleigh wave phase velocity, U the group velocity, (r_1, r_2, r_3, r_4) the motion–stress vector and I_1 the integral term that is defined as:

$$I_1 = \frac{1}{2} \int_0^{\infty} \rho (r_1^2 + r_2^2) \, dz. \quad (3)$$

We use the Nafe-Drake density model and the velocity profile of Y12 at gridpoint A to compute the ‘partial derivatives’, which are the sensitivity kernels of phase velocity to V_s , V_p and density (Fig. 3). At shallow crustal depths (<10 km for 10 s Rayleigh wave and <20 km for 20 s) an increase in density leads to a decrease in phase velocity (Figs 3a and b). On the contrary, at deeper depths (~ 10 – 30 km for the 10 s and ~ 20 – 60 km for 20 s Rayleigh waves) an increase in density results in an increase in phase velocity (Figs 3a and b). Meanwhile, the phase velocity of 10-s Rayleigh waves appears to be more sensitive to density perturbations than that of the 20-s wave. The former seems to be more sensitive to shallow density while the latter has a broader depth sensitivity range of 0–60 km. Roughly speaking, the 10-s Rayleigh wave is about twice as sensitive as the 20-s Rayleigh wave to shallow density perturbations, which indicates that shallow density structures have more profound effects on short period Rayleigh wave phase velocities. We further vary $\delta\rho/\rho$ from 0 to -30 per cent and use eq. (2) to compute the corresponding

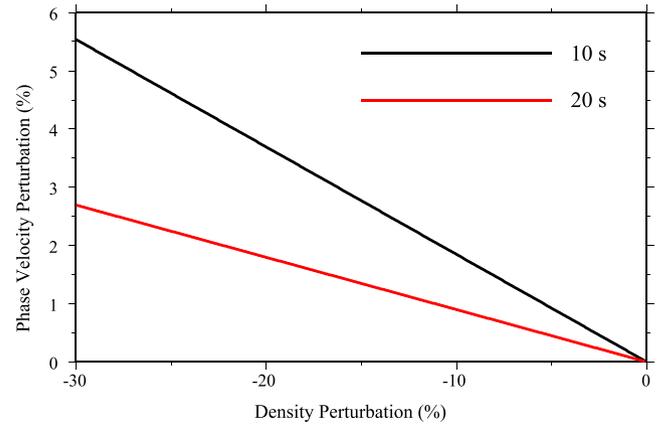


Figure 4. Phase velocity perturbations ($\delta C/C$) are shown as a function of density perturbations ($\delta\rho/\rho$) averaged over the upper 5 km at gridpoint A, for 10-s (black line) and 20-s (red line) Rayleigh waves, respectively.

$\delta c/c$ for the periods of 10 and 20 s (Fig. 4), which are found to be consistent with the phase velocity differences between the CM and Nafe-Drake models (Fig. 2b). Furthermore, the S -wave sensitivity kernels appear to peak at greater depths than the corresponding density kernels (Fig. 3). This difference in peak depth gives rise to the trade-off between the shallow density structure and the shear velocity in the mid-to-lower crustal depths.

3 1-D AND 3-D INVERSIONS WITH DIFFERENT INITIAL DENSITY MODELS

In the previous section we illustrated that shallow density structure has significant effects on Rayleigh wave phase velocities. In this section, we further quantify its effects in 1-D and 3-D shear velocity inversion used in the traditional and adjoint surface wave tomography.

In the 1-D shear wave speed inversion, we use the Monte Carlo method (Shapiro & Ritzwoller 2002) to search for suitable models that fit the phase velocity dispersion curve of a target model. In the 3-D shear wave speed inversion, we apply an iterative adjoint tomography method developed by Chen *et al.* (2014) to minimize frequency-dependent traveltimes misfits between the ‘data’ (Green’s functions produced from a target 3-D model computed with the spectral-element method; Komatitsch *et al.* 2004; Chen *et al.* 2014) and the synthetics (synthetic Green’s functions produced from an initial model and subsequent updated 3-D models).

3.1 1-D Shear wave speed inversions

We use the 1-D profile at the gridpoint A to conduct 1-D inversions of synthetic data to investigate the effect of the absolute density–velocity scaling. To set up the target profile, we take the 1-D V_s profile at gridpoint A from Y12 and use the Poisson solid assumption and the Nafe-Drake density model to calculate V_p and density, respectively. We then calculate the target phase velocity dispersion curves for this target model, which are the dispersion data for the Monte Carlo inversion. The synthetic model is hence referred to as the ‘target model’ in this section.

For each test, we start with an initial model that has exactly the same P - and S -wave velocities as the target model, but with a different density structure. In particular, we employ three initial density models, the CM, Gardner and Birch models, and perform a Monte Carlo 1-D inversion (Shapiro & Ritzwoller 2002). We first set the model depth range to be 0–150 km and divide it into three segments, the shallow depth (0.0–5.0 km), the mid-to-lower crustal depth (5.0–40.8 km), and the upper mantle depth (40.8–150.0 km). Then in each segment we parametrize the shear wave speed as a function of depth using five cubic B-splines. During the Monte Carlo inversion we perturb the five coefficients of each B-spline to produce trial models and employ the simulated annealing algorithm (Shapiro & Ritzwoller 2002) to explore the model space. During the inversion, we also update the P -wave velocity and density by using the relative scalings of $\delta \ln \alpha / \delta \ln \beta = 0.5$ and $\delta \ln \rho / \delta \ln \beta = 0.33$ (Panning & Romanowicz 2006). We will further discuss the influences of different relative scalings in the following discussion section. Vertical model smoothness is taken into account in the objective function, and the recovered model is obtained by averaging all accepted models whose objective functions are below certain threshold.

For the inversion with the CM initial model (Fig. 5a), the recovered S -velocity model is lower by ~ 4.5 per cent in the mid-to-lower crustal depths (10–40 km) when compared to the target model. This large difference between the recovered and target shear wave velocity are largely due to the difference in density at shallow depths between the CM and Nafe-Drake model (Fig. 2a). The inversion using the Birch model shows a similar but a less significant shear velocity reduction in the mid-lower crust (Fig. 5c), which corresponds to the smaller discrepancy between the Nafe-Drake and Birch density models in the shallow depths (Fig. 2a). Meanwhile, the recovered shear wave velocity with the Gardner is slightly higher only in the lower crust level (Fig. 5b), which is likely caused by its lower density in the lower crust (Fig. 2a) based on the positive density sensitivity kernel in this depth range (Fig. 3a). For validation, we conduct another inversion which uses the Nafe-Drake density model, the same as the target model, plus an initial V_s model (orange line in Fig. 5d) that is 5 per cent lower than Y12 above the Moho. The resulting shear velocity profile (black line in Fig. 5d) shows a good match to

the target model with a very subtle discrepancy, probably caused by parametrization and/or regularization in the inversion.

Combined with the results shown in the previous sections, it is clear that density structures in the upper 5 km have significant trade-off with shear wave velocity structure in the mid-to-lower crust in the above 1-D Monte Carlo inversion. In particular, if the initial model has a density ~ 25 per cent lower than the target model in the top 5 km, the recovered shear wave velocity can be ~ 4.5 per cent higher than the true model in the mid-to-lower crustal depths. Since shear wave speed anomalies at this level are significant in tomography, cautions have to be taken in choosing a proper density model at shallow depths when implementing the inversion. We will further confirm this argument with a 3-D inversion example based on adjoint tomography in the following section.

3.2 3-D shear wave speed inversions

In this application of the ambient noise adjoint tomography (Chen *et al.* 2014), we first set up a target 3-D model for the Sichuan basin and surrounding regions by taking the shear wave speeds from Y12 and using the Poisson solid assumption and the Nafe-Drake density model to calculate compressional wave speeds and density, respectively. We then generate synthetic Green’s functions (Chen *et al.* 2014) between all the station pairs (Fig. 1b) for this target model using the spectral-element method (Komatitsch *et al.* 2004). These synthetic Green’s functions are treated as observed empirical Green’s functions (EGFs) derived from ambient noise, and the target model is assumed as a known ground truth model. In the 3-D adjoint tomography, we use the CM density model to calculate the initial density model. We only invert for shear wave speeds β and scale $\delta \rho$ and $\delta \alpha$ to $\delta \beta$ based on the same scaling relations described in the 1-D inversion.

The synthetic Green’s functions (SGFs) are also calculated for the starting model and the subsequent updated models. The frequency-dependent traveltimes misfits between the EGFs and the SGFs are measured at four period bands: 10–20 s, 15–30 s, 20–40 s and 25–50 s. The overall misfit is iteratively minimized based on a preconditioned conjugate gradient method while the 3-D model gets refined using 3-D finite-frequency kernels (Chen *et al.* 2014). The inverted 3-D model is obtained after three iterations (Figs 6a–c).

Density differences between the initial model and the target model averaged over the upper 5 km (Fig. 6d) shows an amplitude of up to 25 per cent inside the Sichuan basin. We also compute phase velocities at 10 and 20 s periods using the *geopsy* software package for both the initial 3-D model and the target model. The largest phase-velocity differences (up to ~ 5 per cent) between these two models (Figs 6e and f) are located within the Sichuan basin, which are correlated with the shallow density deficit in the initial model (Fig. 6d). In order to match the observed phase velocities, the shear wave speeds in the mid-to-lower crust have to be reduced by ~ 4.5 per cent (Fig. 6b). The 1-D shear wave speed profile extracted from the inverted model at gridpoint A is also consistent with the 1-D inversion result (Fig. 5a). Therefore, the 3-D inversion results further confirm that large model errors in density at shallow depths can result in significant artificial shear wave speed anomalies at mid-to-lower crustal depths.

4 DISCUSSION AND CONCLUSIONS

In the above 1-D and 3-D inversions, we have assumed a Poisson solid ($V_p = 1.732V_s$) and a fixed perturbation scaling

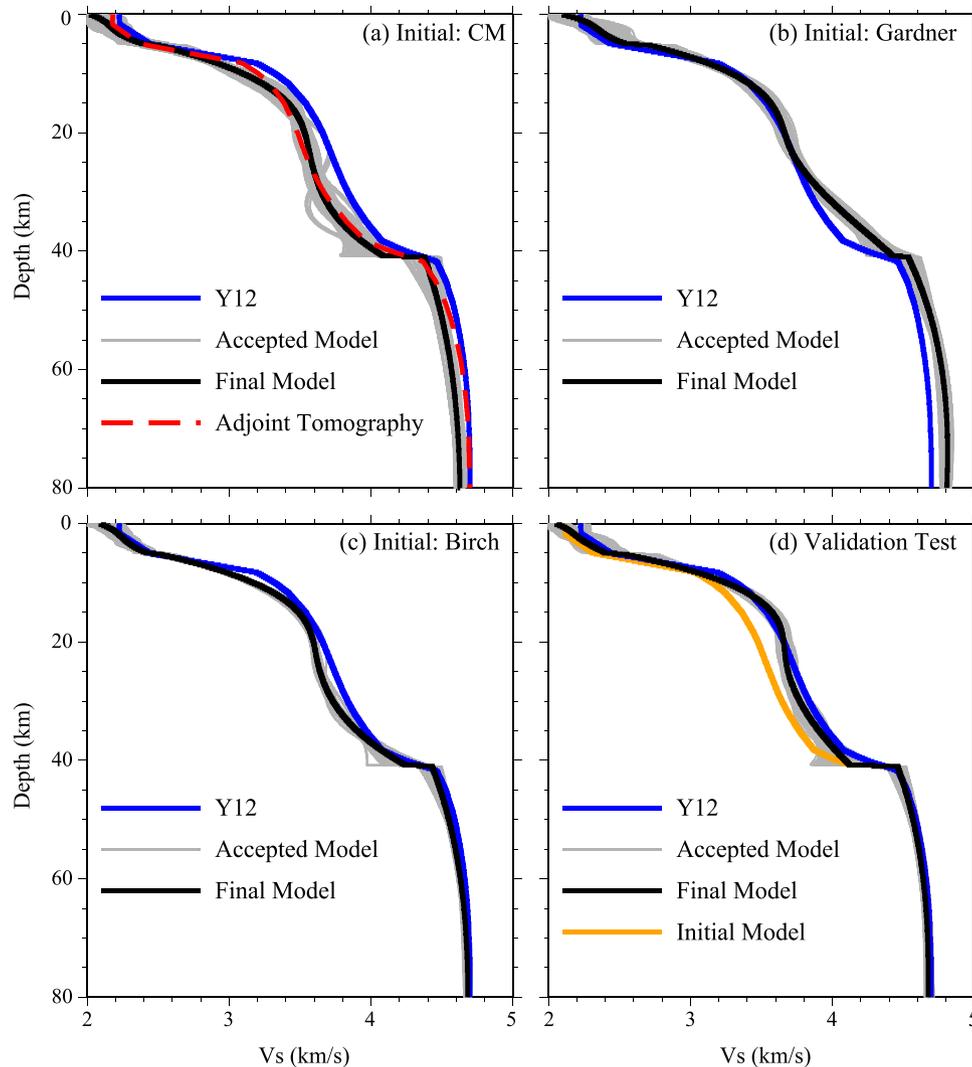


Figure 5. (a)–(c) Inverted shear wave velocity depth profiles of the 1-D Monte Carlo inversion at gridpoint A with different initial density models: (a) CM, (b) Gardner; (c) Birch. Blue line represents the target shear velocity model (Y12). Grey lines indicate all acceptable models of the 1-D Monte Carlo inversion, and the black line the finally inverted model obtained by averaging all the acceptable models. The dashed red line in (a) shows the final S -velocity model at gridpoint A derived from the 3-D adjoint tomography. (d) Same as (a)–(c) but for a validation test with a Nafe-Drake initial density model and an initial V_s model (orange line) which is 5 per cent slower than Y12 at the depths above the Moho.

($\delta \ln \alpha / \delta \ln \beta = 0.5$ and $\delta \ln \rho / \delta \ln \beta = 0.33$) in constructing the initial P -wave model and in model updating, respectively. It is well known that sediments could have very high V_p/V_s ratio, very different from that of a Poisson solid. To investigate the absolute scaling between V_p and V_s , we create an S -wave velocity model by adding a 300 m low velocity layer with $V_s = 0.5 \text{ km s}^{-1}$ on the top of the Y12 grid A profile. We then generate two sediment P -wave velocity models by using a V_p/V_s ratio of 1.732 and 4.0, respectively. Everything below the sediment is exactly the same (Fig. 7a). We compute the phase velocities of these two models, which differs by less than 0.1 per cent at periods shorter than ~ 20 s (Figs 7b and c).

In order to further evaluate how the assumed V_p/V_s scaling in the whole crust affects the 1-D inversion results, we generate another crustal P -wave model using the Nafe-Drake V_p/V_s scaling relationship (Brocher 2005; red line in Fig. 7d). We notice that the Nafe-Drake P -wave velocity model is actually not so different from the Poisson solid model. Therefore their corresponding phase velocities are very similar, with a discrepancy less than 1 per cent over the entire periods of 10–50 s (Figs 7e and f). It thus appears that the

Poisson's solid assumption used in building the initial model is valid, and the absolute V_p/V_s scaling seems to have very little influence on the final models.

In the above 1-D and 3-D inversions, we employ a relative scaling relationship of $\delta \ln \alpha / \delta \ln \beta = 0.5$ and $\delta \ln \rho / \delta \ln \beta = 0.33$ (Panning & Romanowicz 2006) in updating the density and P -wave velocity for both crustal and mantle layers. We further conduct the 1-D inversions with other three relative scalings: (1) $\delta \ln \alpha / \delta \ln \beta = 0.0$, $\delta \ln \rho / \delta \ln \beta = 0.0$; (2) $\delta \ln \alpha / \delta \ln \beta = 1.0$, $\delta \ln \rho / \delta \ln \beta = 0.0$; (3) $\delta \ln \alpha / \delta \ln \beta = 1.0$, $\delta \ln \rho / \delta \ln \beta = 1.0$. (1) Means that both V_p and density are fixed, and (2) implies for a fixed V_p/V_s ratio and a fixed density (Yang *et al.* 2012), while (3) indicates for a fixed V_p/V_s ratio and a fixed ρ/V_s ratio, such as the CM model, during the inversions. The S -wave velocity and density are kept the same in all the inversions. The final S -wave velocity models with the three perturbation scalings are shown in Figs 7(h)–(j), respectively. For comparison we also show the S -wave velocity model obtained in the previous 1-D inversion in Fig. 7(g). It seems that the final V_s model does depend slightly on how V_p is

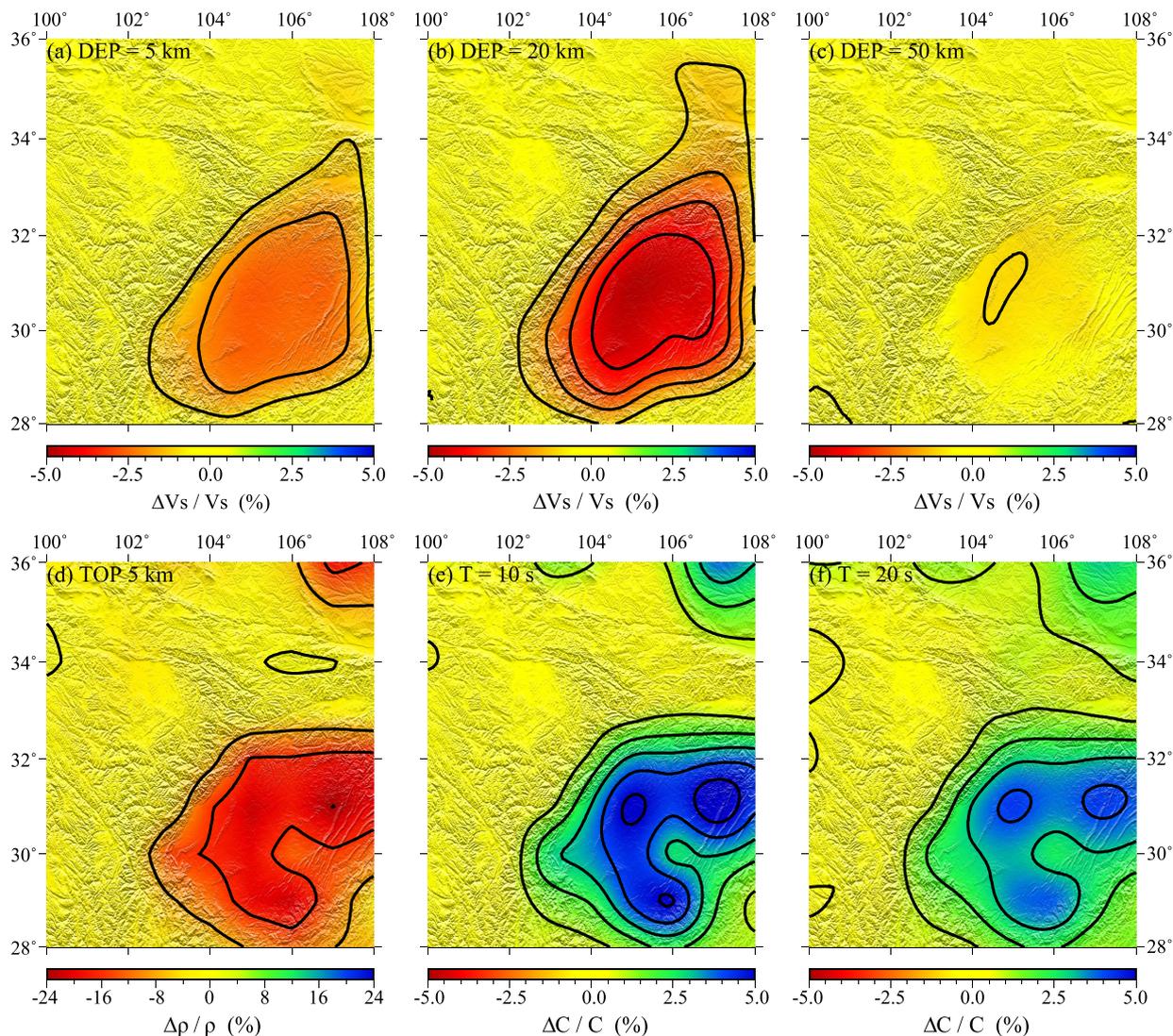


Figure 6. Maps showing the difference in percentage between the recovered and target models at 5 km (a), 20 km (b) and 50 km depths (c), obtained by the 3-D iterative adjoint tomography. The contour interval is 1 per cent. (d) Colour contour map showing the average density difference of the upper 5 km between the initial CM and the target Nafe-Drake models. The contour interval is 8 per cent. Panels (e) and (f) show the phase velocity differences between the initial CM and target Nafe-Drake model for 10-s and 20-s Rayleigh waves, respectively. The contour interval here is 1 per cent.

updated during the inversion, but the model difference is generally less than 1 per cent. Therefore, the potential bias on the inversion caused by improper perturbation scalings is almost negligible.

By conducting both 1-D and 3-D shear wave speed inversions, we demonstrate that density structures at shallow depths have great effects on the short-period surface wave dispersion curves and there is a significant trade-off between density structures at shallow depths and shear wave speed at mid/lower crust in these inversions. Improper scaling of density from shear velocity in the shallow depths can lead to spurious shear velocity anomalies in the mid/lower crust in surface wave inversion. Specifically, for sedimentary basins, such as the Sichuan basin, where sediment deposits can reach several kilometres, ~ 25 per cent deviation in density in the top sedimentary layer can introduce a 5 per cent error in shear velocity in the mid/lower crust. The artificial anomalies induced by density errors in the top layers can mislead interpretations of tomographic results. Therefore, it is very important to adopt a reasonable density model during the 1-D and 3-D inversions to obtain a robust shear wave speed model in surface wave tomography.

Recent studies (e.g. Lin *et al.* 2012) on the joint inversion of H/V ratio and phase velocity suggest that addition of ellipticity data can help constrain the density in the shallow depth of the crust. Meanwhile gravity data are always helpful in providing additional constraints on density structure. Maceira & Ammon (2009) proposed a joint inversion algorithm of surface wave and gravity data, which integrates the data with empirical relations between velocity and density. Furthermore, although suffering from the intrinsic nonuniqueness, the inversion of gravity data itself can yield smooth and geologically reasonable density model (Bear *et al.* 1995; Li & Oldenburg 1998), which can be used as the initial density model to improve the reliability of the surface wave tomography.

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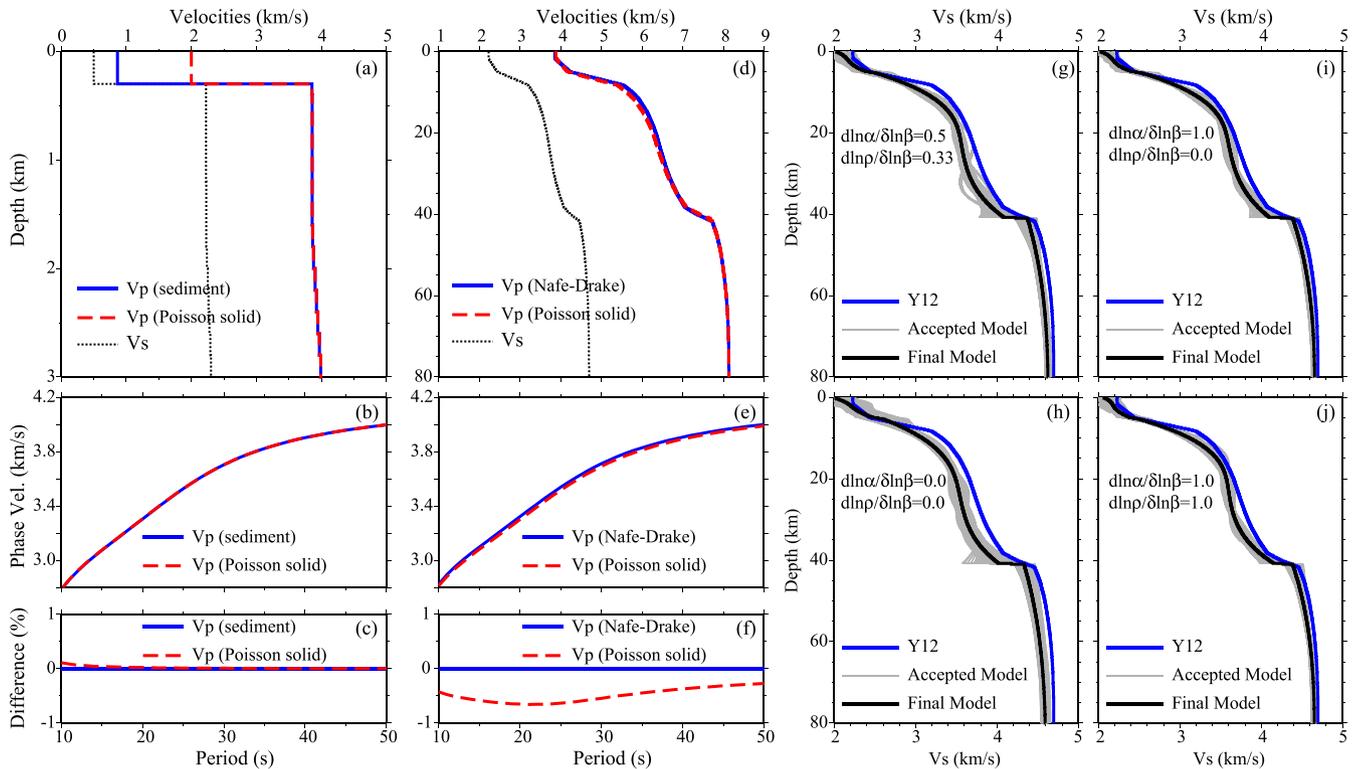


Figure 7. Influence of the absolute V_p – V_s scaling and the perturbation scalings ($\delta \ln \alpha / \delta \ln \beta$, $\delta \ln \rho / \delta \ln \beta$) on the inversion. (a) The sediment P -wave velocity model with a V_p/V_s ratio of 4.0 (blue solid line) is shown together with the Poisson solid model with a V_p/V_s ratio of 1.732 (red dashed line). The two models share with similar S -wave velocity (black dotted line) and density structures. Panels (b) and (c) show the calculated phase velocities of the two models in the period range of 10–50 s, and their difference in percentage. (d) The Nafe-Drake P -wave velocity model (blue solid line) is shown with the Poisson solid model (red dashed line) for comparison. The phase velocities of the two models and their difference are shown in (e) and (f), respectively. The S -wave velocity depth profiles of the 1-D Monte Carlo inversions at gridpoint A with different perturbation scalings: (g) $\delta \ln \alpha / \delta \ln \beta = 0.5$, $\delta \ln \rho / \delta \ln \beta = 0.33$; (h) $\delta \ln \alpha / \delta \ln \beta = 0.0$, $\delta \ln \rho / \delta \ln \beta = 0.0$; (i) $\delta \ln \alpha / \delta \ln \beta = 1.0$, $\delta \ln \rho / \delta \ln \beta = 0.0$; (j) $\delta \ln \alpha / \delta \ln \beta = 1.0$, $\delta \ln \rho / \delta \ln \beta = 1.0$. The models are plotted similarly as Fig. 5.

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