

ARTICLE

Joint inversion of ambient noise and gravity to investigate crustal structure and dynamic processes in the southern Sichuan Basin

Li Zhang¹, Guiju Wu^{1*}, Rugang Xu², Yufei Xi³, Jiapei Wang¹, and Fan Wang¹¹Key Laboratory of Earthquake Geodesy, Institute of Seismology, China Earthquake Administration, Wuhan, Hubei, China²Anhui Earthquake Agency, Hefei, Anhui, China³Institute of Hydrogeology and Environmental Geology, Chinese Academy of Geological Sciences, Shijiazhuang, Hebei, China

Abstract

The southern Sichuan Basin, particularly the Changning area, is characterized by intense tectonic activity and frequent earthquakes. However, crustal structure models derived from different geophysical methods exhibit significant discrepancies, which impede an in-depth understanding of the tectonic evolution and earthquake genesis in this region. In this study, a joint inversion integrating surface-wave dispersion and Bouguer gravity anomaly data is adopted. By employing empirical velocity–density relations, the two datasets are coupled into a unified linearized least-squares inversion framework, in which local optimization is achieved through iterative solving. This approach is utilized to obtain the crustal density structure of the southern Sichuan Basin. The joint inversion results reveal an intracrustal low-density layer in the southern Sichuan Basin, inferred to originate from plastic deformation or partial melting of felsic components at relatively high temperatures. Driven by the topographic pressure gradient, this low-density layer undergoes ductile flow toward the Sichuan Basin. Obstructed by the rigid crust of the Sichuan Basin, the material migrates along both the eastern and western margins of the basin, with a broader eastward migration range. The leading edge of the low-density layer shows a distinct special correspondence with the seismic activity concentration zone in the Changning area, a stress concentration zone jointly shaped by deep material migration and tectonic blocking. This stress accumulation serves as the key controlling factor for the frequent earthquakes in the Changning area.

***Corresponding author:**Guiju Wu
(wugjsky@126.com)

Citation: Zhang L, Wu G, Xu R, Xi Y, Wang J, Wang F. Joint inversion of ambient noise and gravity to investigate crustal structure and dynamic processes in the southern Sichuan Basin. *J Seismic Explor.* 2026;35(3):025010143.
doi: 10.36922/JSE025010143

Received: December 31, 2025**Revised:** March 23, 2026**Accepted:** March 27, 2026**Published online:** May 12, 2026

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Keywords: Joint inversion; Seismic tomography; Crustal structure; Changning earthquakes

1. Introduction

With significant improvements in seismic station coverage and the continuous accumulation and extensive application of geophysical exploration data, numerous studies on geophysical inversion have emerged, including magnetotelluric imaging, seismic body-wave tomography, surface-wave tomography, and receiver function

analysis.^{1–5} However, constrained by the limited sensitivity of single geophysical data and insufficient sampling of observational data, a specific type of geophysical data usually imposes strong constraints only on physical parameters at certain subsurface locations or of specific scales.⁶ In contrast, the joint inversion method can integrate multiple types of observational data, leverage their complementary advantages, and thus yield more reliable inversion results. The concept of joint inversion can be traced back to the work of Vozoff and Jupp,⁷ who performed joint inversion of direct current and magnetotelluric data. Subsequently, the idea of joint inversion has been applied to various fields of geophysical inversion and diverse combinations of datasets. A pioneering work in joint inversion was conducted by Julià *et al.*,⁸ who performed the joint inversion of receiver functions and surface-wave dispersion. They systematically accounted for differences in sensitivity and error characteristics across observational datasets, adjusted the constraining influence of each data type via weighting during inversion, and established a unified objective function framework for joint inversion, laying a key methodological foundation for subsequent joint inversion studies. At present, seismic surface-wave tomography and gravity inversion are the primary methods for obtaining crust-mantle shear-wave velocity and density structure, respectively. Although gravity data are highly sensitive to lateral density variations, their radial resolution is often limited. In contrast, seismic surface-wave data offer high radial resolution for shear-wave velocity but provide insufficient velocity constraints for short-distance observations. Therefore, joint inversion of surface-wave and gravity data can leverage the strengths of the two individual geophysical inversion methods, thereby improving the reliability of the inversion results.

The core of performing joint inversion across models with different physical parameters lies in establishing relationships among these physical models. Two common methods are currently used for this purpose: one is based on empirical or statistical relations between different physical properties, and the other is based on structural similarity between different physical property models, such as the cross-gradient method.⁹ Numerous studies and applications of these two methods have been carried out in recent years. Maceira and Ammon¹⁰ adopted the Nafe–Drake and Birch laws, optimized the velocity modeling process in complex basin areas by using piecewise velocity–density relations to adapt to lithological differences between shallow sedimentary layers and deep basements, and applied this method to Central Asian basins to obtain a three-dimensional (3D) shear-wave velocity model with richer shallow details and physical self-consistency. Chen *et al.*¹¹ derived a 3D shear-wave velocity model of

the lithosphere in the Sichuan area via joint inversion of gravity and surface-wave data, using the empirical relations proposed by Maceira and Ammon,¹⁰ and Brocher.¹² Wan *et al.*¹³ inverted the density model of the lithosphere beneath the Qinghai–Xizang Plateau by jointly using gravity and surface-wave phase velocity data based on the Nafe–Drake curve and Brocher’s empirical relations. Han¹⁴ developed a joint inversion algorithm for various types of seismic and gravity data based on the cross-gradient structure and applied it to regional imaging studies in different regions.

The Yangtze Craton acts as a natural tectonic barrier, effectively impeding the eastward expansion of the Qinghai–Xizang Plateau. Situated between the eastern segment of the Qinghai–Xizang Plateau and the Yangtze Craton, the Eastern Tibetan Bounding Block is a seismically active zone. Over the past two decades, a series of strong earthquakes have occurred successively within this tectonic belt and its adjacent areas; typical examples include the 2008 Wenchuan $M_s 8.0$ earthquake and the 2013 Lushan $M_s 7.0$ earthquake. Most of these earthquakes exhibit a close genetic relationship with the Longmenshan Fault Zone, thereby providing critical observational evidence for deciphering the tectonic dynamic processes governing the eastern Qinghai–Xizang Plateau.¹⁵ This study focuses on the southern Sichuan Basin (Figure 1B), a core transitional zone between the southeastern Qinghai–Xizang Plateau and the southwestern Sichuan Basin. Spanning the high-relief, steep terrain of the southeastern Qinghai–Xizang Plateau and the relatively low-lying, gentle landscape of the southwestern Sichuan Basin, this region exhibits remarkable topographic variation. As a critical zone for plate compression and block interaction, it is characterized by the development of multiple faults, including the Ebian Fault, Xihe–Meigu Fault, Mabian Fault, Guancun Fault, Wulianfeng Fault, Huayingshan Fault, and Zhaotong–Ludian Fault. Seismic activity in this region is frequent, and the processes of its earthquake preparation and occurrence are directly controlled by the properties of the crustal materials, fault characteristics, and deep tectonic stress states in the southern Sichuan Basin. Consequently, constructing a high-resolution 3D model of shear-wave velocity and density structure is a key foundation for analyzing the crust–mantle tectonic evolution and dynamic mechanisms. This model further provides essential support for investigating the correlation between the deep structure of the southern Sichuan Basin and the genesis of the Changning earthquake. In summary, there is an urgent need to develop a subsurface structure model of the southern Sichuan Basin through effective inversion methods, thereby providing key support for the genetic interpretation of seismic activity in the Changning area.

With significant improvement in the coverage density of seismic stations, research on the subsurface structure of the southern Sichuan Basin has advanced considerably. Li *et al.*¹⁶ conducted a joint inversion of receiver functions and surface-wave dispersion data, and their results revealed that the regional crustal thickness increases from 30 km in southern Yunnan to 62 km in the southeastern margin of the Qinghai–Xizang Plateau, and there are two continuous low-velocity zones in the middle-lower crust serving as core channels for crustal flow, providing key seismological evidence for the study of the lateral extrusion mechanism of the Qinghai–Xizang Plateau. He *et al.*¹⁷ systematically investigated the tectonic and geological background and mechanism of earthquakes in the Changning area by integrating multiple technical approaches, including seismic relocation, tectonic modeling, and evolutionary restoration analysis. Their findings indicated that earthquakes in the Changning area are associated with the southeastward advance of the southeastern Qinghai–Xizang Plateau, whereas moderate-to-large earthquakes in the southwestern Sichuan Basin are predominantly controlled by Cenozoic fault activity and subsurface structures. Long *et al.*¹⁸ performed 3D velocity structure inversion and demonstrated that the velocity anomalies in the Changning–Gongxian area are highly consistent with known fault zones, and small earthquake swarms mainly occur in areas with high P-wave or S-wave velocity anomalies, which are closely related to the subsurface structure. Li *et al.*¹⁹ employed double-difference seismic tomography to invert the 3D P-wave velocity structure of the upper crust in the earthquake area. Their study revealed the deep structural characteristics and seismogenic setting of the Changning M_s6.0 earthquake area, providing a detailed explanation of the tectonic background and genesis of seismic activity.

A comprehensive review of the literature indicates that numerous geophysical exploration projects have been conducted in the southern Sichuan Basin and its surrounding areas, yielding valuable insights into the subsurface structure and the depth of the Moho discontinuity in this region. However, significant debates persist regarding the fundamental causes of frequent earthquakes in the area and the corresponding seismogenic models. Therefore, this study aims to develop a subsurface structural model of the southern Sichuan Basin, identify intracrustal density anomalies and their spatial distribution, and elucidate the relationship between deep structural anomalies, tectonic stress accumulation, and frequent earthquakes in the Changning area. This study adopts a joint inversion method integrating surface-wave dispersion and Bouguer gravity anomaly data.²⁰ The empirical velocity–density relations are employed

to couple the two datasets into a unified linearized least-squares inversion framework,¹² where local optimization is achieved through iterative solving to obtain the 3D crustal density structure model of the southern Sichuan Basin. This study not only provides a more reliable and accurate 3D structural model of the southern Sichuan Basin by integrating complementary constraints from multi-physical field data, but also discusses in detail the genesis of the intracrustal low-density layer in the southern Sichuan Basin by integrating previous geological and geophysical observation results. Furthermore, it innovatively analyzes the deep-dynamic processes of earthquake preparation in the Changning area from a macroscopic perspective of deep material flow and rigid block interaction, thereby offering a novel explanatory framework for the genesis of regional strong earthquakes.

2. Data and methods

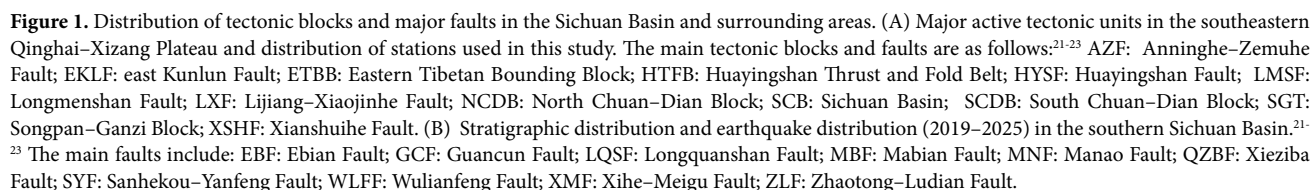
2.1. Surface-wave dispersion data

Surface-wave dispersion data were extracted from ambient seismic noise data. The noise data consist of annual waveform records collected in 2016 from 87 stations deployed in the region by the China National Seismic Network.²⁴ The station locations are illustrated in [Figure 1A](#). Ultimately, 3,127 fundamental-mode Rayleigh wave phase velocity dispersion curves and 2,167 fundamental-mode Rayleigh wave group velocity dispersion curves within the period range of 4–40 s were extracted from the ambient noise cross-correlation data.²⁵ This study utilized the data that underwent quality control of the extracted dispersion curves, as conducted by Du *et al.*²⁰ ([Figure 2](#)).

To ensure data reliability, the surface-wave ray-path coverage was evaluated. As presented in [Figure 3](#), ray-path coverage maps are shown for periods of 8 s, 12 s, 16 s, 20 s, 24 s, 28 s, 32 s, and 36 s. It can be observed that most areas are covered by stations, with dense ray distribution, and the ray paths adequately cover the entire study area, indicating that this dataset can effectively resolve the crustal structure of the area.

2.2. Bouguer gravity anomaly data

The Bouguer gravity anomaly data are derived from the Earth Gravitational Model 2008 (EGM2008), with a grid spacing set to 3'×3'. The model's primary data sources include terrestrial gravity, satellite altimetry, and satellite gravity observations.²⁶ [Figure 4](#) presents the distribution of the Bouguer gravity anomaly utilized in this study. To eliminate high-frequency disturbances caused by the dramatic topographic relief along the southeastern margin of the Qinghai–Xizang Plateau, this study processed EGM2008 gravity field data using a matched-filter method



where $T = T(\theta, \phi, \theta_0, \phi_0)$ is the arrival time of the surface-wave propagating from spherical coordinates (θ_0, ϕ_0) to (θ, ϕ) at a given period, and R is the Earth's radius. This equation can be efficiently solved using the fast marching

method.^{28,29} Given the enormous computational cost, a localization approach is adopted to calculate the dispersion map, in which the dispersion map at each point on the free surface is determined solely by the one-dimensional (1D) layered model beneath that point. Subsequently, the Thomson–Haskell algorithm^{30,31} is utilized for forward computation. With this method, surface-wave travel times for each period can be directly calculated from a given 3D model.

The radial component of the gravity field produced by the density distribution of a spherical prism can be expressed as:³²

$$g_r(r_o, \theta_o, \phi_o) = \int_{r_1, \theta_1, \phi_1}^{r_2, \theta_2, \phi_2} \frac{G \rho_s r_s^2 \sin \theta_s (-r_o + r_s \cos \delta)}{R_{os}^3} dr_s d\theta_s d\phi_s \quad (3)$$

where G is the gravitational constant, g_r is the radial component of gravity at (r_o, θ_o, ϕ_o) , R_{os} is the straight-line distance between source and receiver, and δ is the epicentre distance. For complex density distributions, the Gauss–Legendre numerical integration method can be used, such that the above integral can be expressed as:

$$g_r^n = \sum_{i=1}^I \sum_{j=1}^J \sum_{k=1}^K \omega_i \omega_j \omega_k g(r_i, \theta_j, \phi_k), n = 1, 2, \dots, N \quad (4)$$

where I, J, K denotes the order of Gauss–Legendre

integration used in each direction, $\omega_{i,j,k}$ denotes the weight coefficient in each direction, and (r_i, θ_j, ϕ_k) denotes the position of the Gauss–Legendre integration node in each direction.

Since a longitude–latitude oriented grid is utilized to solve **Equation 2**, for computational convenience, this grid is required to coincide with that adopted in gravity forward modeling. Notably, different prisms contribute unequally to gravity observations owing to their sizes and distances from the observation point, making it extremely time-consuming to use an identical number of Gauss–Legendre points in each prism. To tackle this problem, the number of grid points and the order of the Gauss–Legendre polynomial for each prism can be adaptively determined based on the distance between a grid point and its adjacent points,³³ which significantly improves computational efficiency.

2.3.2. The inverse problem

To enable joint inversion, the surface-wave and gravity inversion equations need to be unified into a single system. Surface-wave dispersion data are mainly related to shear-wave velocity, and Bouguer gravity anomaly data, after conversion between velocity and density, are also related to shear-wave velocity. During the inversion process, surface-wave travel-time residuals and gravity residuals can be correlated with shear-wave velocity perturbations via sensitivity kernel matrices. Joint inversion can assemble the surface-wave travel-time and gravity data residuals into a sparse linear system in each iteration:

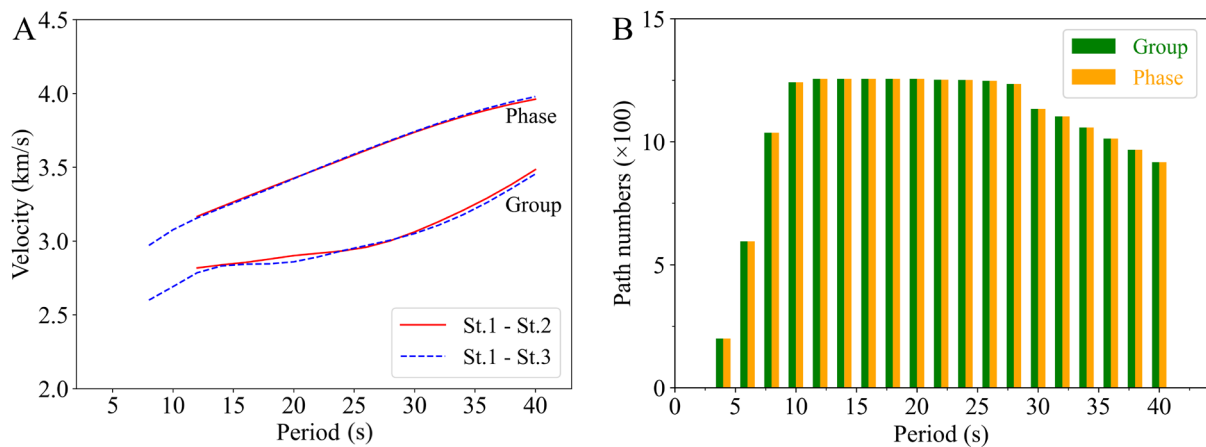


Figure 2. Surface-wave dispersion data for inversion. (A) Group velocity and phase velocity dispersion curves between station pairs St.1–St.2 and St.1–St.3 (Figure 1A); (B) number of group velocity and phase velocity paths used for each period in the inversion.

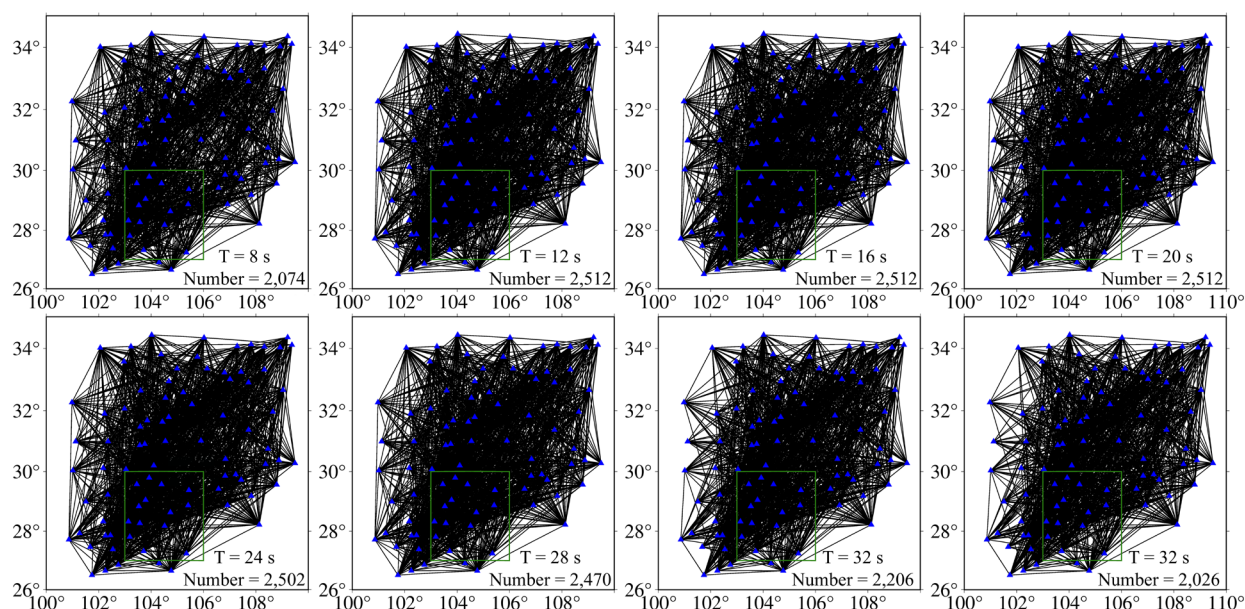


Figure 3. Ray-path coverage maps for different periods. The study area is the region enclosed by the green rectangle, and blue triangles represent stations used in this study.

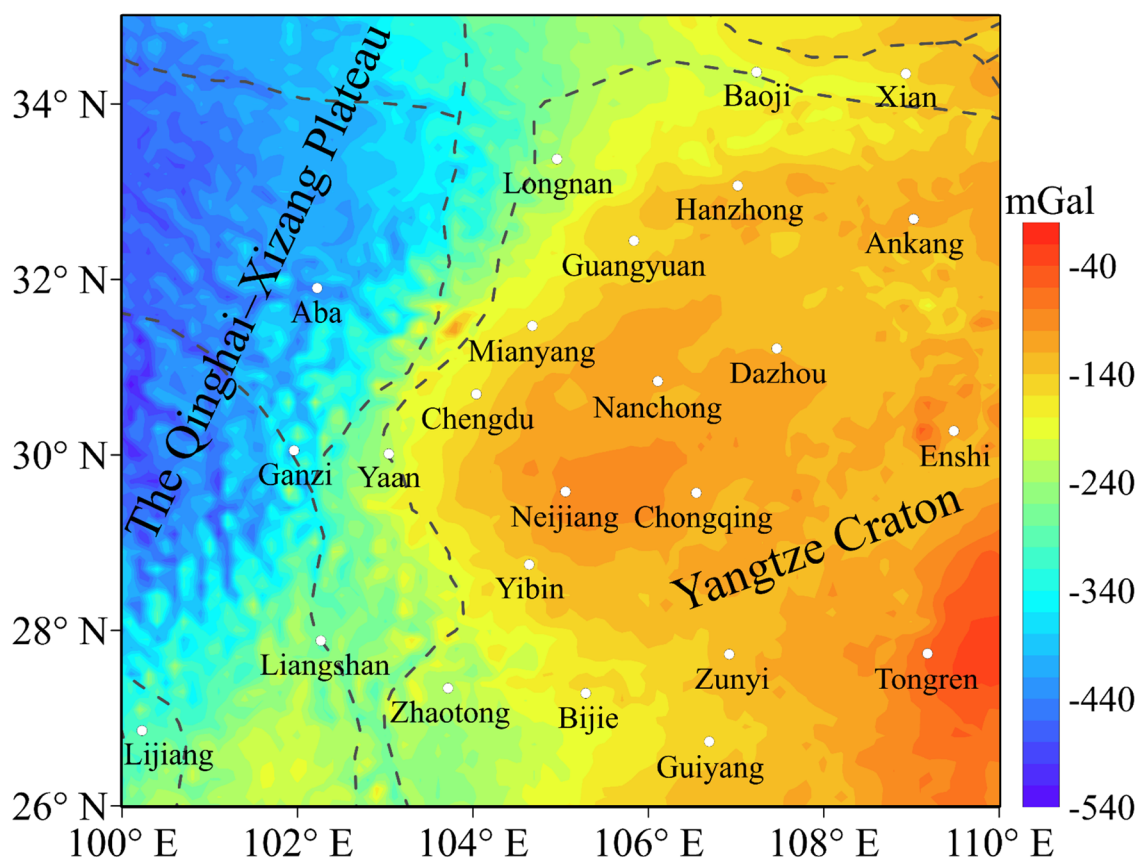


Figure 4. Distribution of Bouguer gravity anomaly data obtained from the Earth Gravitational Model 2008. Dashed lines represent the block boundaries in this region.²³

$$\begin{bmatrix} G_s \\ G_g \end{bmatrix} \delta\beta_i = \begin{bmatrix} \delta t \\ \delta g_g \end{bmatrix} \quad (5)$$

The solution to **Equation 5** can be formulated as a weighted least-squares optimization problem with second-order Tikhonov regularization and damping:^{10,34–36}

$$\delta\beta_i = \underset{m}{\operatorname{argmin}} \left\| \begin{bmatrix} W_s G_s \\ W_g G_g \\ \sigma S \\ \gamma I \end{bmatrix} m - \begin{bmatrix} W_s \delta t \\ W_g \delta g_g \\ 0 \\ 0 \end{bmatrix} \right\|_L \quad (6)$$

where G_s and G_g are the sensitivity kernel matrix of surface-wave and gravity anomaly to shear-wave velocity, respectively; σ and S are the second-order Tikhonov regularization coefficient and matrix, respectively; γ is the damping factor; I is the identity matrix; W_s and W_g are diagonalized weight matrices for surface-wave and gravity data, respectively:⁸

$$W_s = \operatorname{diag} \left\{ \sqrt{\frac{p}{N_s (\sigma_i^s)^2}}, i = 1, 2, \dots, N_s \right\} \quad (7)$$

$$W_g = \operatorname{diag} \left\{ \sqrt{\frac{1-p}{N_g (\sigma_i^g)^2}}, i = 1, 2, \dots, N_g \right\} \quad (8)$$

where p is a parameter between 0 and 1. $N_{s,g}$ are the number of data for surface-wave and gravity data, respectively, and $\sigma_{s,g}$ are the standard variations of the two datasets.

This augmented linear system is efficiently solved under the L_2 -norm using the least-squares minimum residual (LSMR) iterative method.³⁷ Based on the standard derivative property of the L_2 -norm, the gradient of the objective function $J(m)$ with respect to the variable m can be explicitly expanded as:

$$\nabla J(m) = 2 \left[G_s^T W_s^T W_s (G_s m - \delta t) + G_g^T W_g^T W_g (G_g m - \delta g_g) \right] + 2 \left[\sigma^2 S^T S + \gamma^2 I \right] m \quad (9)$$

By progressively searching along these gradient directions toward the local minimum, LSMR identifies the optimal solution.

Subsequently, the model is updated using **Equation 10** until the residuals are in a reasonable range.

$$\beta_{i+1} = \beta_i + \delta\beta_i \quad (10)$$

3. Joint inversion of the southern Sichuan Basin

3.1. Model setup and parameter selection

Based on the ray-path distribution in the study area and extensive tests, the initial velocity model was configured to divide the study area into a latitude–longitude grid with a 0.3° horizontal interval. In the vertical direction, the shallow layer is the main seismogenic layer in the study area, encompassing key structures such as sedimentary layers and active faults, and thus requiring high-resolution characterization of structural heterogeneity. In contrast, the deep crustal structure is relatively homogeneous, with lateral inhomogeneity weakening with increasing depth, and thus, sparse layering can also reflect the overall trend. To balance resolution and computational efficiency, the model is divided into 21 layers with gradually increasing grid spacing with depth, spanning 0 km to 70 km, as detailed in **Table 1**. Since linearized methods rely on the initial model, the inversion results of Xia²⁵ for this region were adopted for the crust and upper mantle in this study. In this initial model, the shear-wave velocity from these inversion results increases with depth. The shallow layer at 0–6 km is controlled by the sedimentary layers of the Sichuan Basin, where basin velocities are significantly lower than those in the surrounding mountainous areas. In the 6–20 km depth range, the extent of low-velocity anomalies inside the basin shrinks gradually, and the lateral inhomogeneity of the medium weakens markedly. Starting from 20 km depth, a remarkable reversal occurs in the velocity structural pattern: the shear-wave velocity in the Sichuan Basin becomes progressively higher than that in the eastern margin of the Qinghai–Xizang Plateau, and a distinct and prominent velocity difference forms on both sides of the basin–mountain boundary. In terms of the Moho depth, the Moho discontinuity beneath the Sichuan Basin and its adjacent areas is generally at 40–60 km, and it deepens gradually from the basin westward to the southeastern margin of the Qinghai–Xizang Plateau. This model accurately delineates the velocity differences among the thick sedimentary layers, the rigid basin basement of the Sichuan Basin, and the surrounding orogenic belts, as well as the lateral variation law of the Moho discontinuity, and is highly consistent with the regional tectonic setting. Therefore, selecting this model can provide a rational and stable initial constraint for linearized inversion, effectively reduce the non-uniqueness of inversion solutions, and simultaneously enhance the inversion process's convergence efficiency and the credibility of the final results.

For the velocity–density empirical relationship, a general relationship proposed by Brocher was employed in this study:¹²

Table 1. Three-dimensional model settings of the study

Depth/km	Thickness/km	Number of layers
0–20	2	10
20–40	4	5
40–70	5	6

$$\rho = 1.6612\alpha - 0.4721\alpha^2 + 0.0671\alpha^3 - 0.0043\alpha^4 + 0.000106\alpha^5 \quad (11)$$

$$\alpha = 0.9409 + 2.0947\beta - 0.8206\beta^2 + 0.2683\beta^3 - 0.0251\beta^4 \quad (12)$$

where ρ denotes density, α denotes P-wave velocity, and β denotes S-wave velocity. This empirical relationship is established based on global datasets of crustal borehole logging, laboratory petrophysical measurements, and seismic tomography. It covers a variety of common crustal lithologies, including sedimentary rocks, granitic rocks, and metamorphic rocks, and is applicable to seismically active continental margins, which is highly consistent with the geological and tectonic setting of the study area. Furthermore, this empirical relationship is valid for the ranges of $1.5 \text{ km/s} < V_p < 8.5 \text{ km/s}$ and $V_s < 4.5 \text{ km/s}$. Since this study focuses on the density structure of the crust and upper mantle, the empirical relationship is applicable over most depths in the research region.

To determine the weight matrices in **Equation 6**, it is necessary to estimate the measurement uncertainties of the two datasets. For the EGM2008 data, the uncertainty is typically 10.5 mGal.³⁸ For surface-wave dispersion data, the uncertainties were estimated as 1.4 s by Du *et al.*²⁰ using **Equation 13**.

$$\delta t = d \frac{\delta c}{c_o} \quad (13)$$

where δt is the surface-wave travel-time error, d is the average interstation distance, δc is the average error of the observed phase and group dispersion velocities, and c_o is the average dispersion velocity.

To accelerate convergence of the inversion, it is essential to ensure that low-frequency information in the model is recovered as much as possible in the initial iterations. Therefore, the method proposed by Roecker³⁹ was adopted. Only surface-wave travel-time data were fitted in the first 10 iterations, and gravity data were incorporated for joint inversion in subsequent iterations. In the final 10 iterations, different values of P were tested, and both datasets could be fitted with root-mean-square (RMS) residuals close to the noise level, with negligible differences. Therefore, the value of P was set to 0.5, indicating that the weights of the two datasets are determined solely by their statistical uncertainties, and no additional artificial weighting is applied to either dataset.

In the recursive solution of algebraic equations during each iteration, the damping and smoothing factors control the amplitude of velocity variation per iteration. In the numerical experiments, the effects of the damping factor are much smaller than those of the smoothing factor. Therefore, in this study, the damping factor is ultimately fixed to a very small value. To select an appropriate smoothing factor, the model regularization norm and data residual norm corresponding to different smoothing factors were calculated, and an L-curve was plotted for analysis (**Figure 5**). The parameter value was determined based on the L-curve.⁴⁰ In the L-curve, the optimal smoothing factor corresponding to the inflection point was selected as 30.

3.2. Checkerboard test

Based on the ray-path distribution in the study area and extensive tests, the size of the velocity anomaly body was finally set to $1.2^\circ \times 1.2^\circ$, with a velocity perturbation of $\pm 5\%$. In the vertical direction, intervals of 10 km were used for the 0–20 km depth range, and 20 km for the 20–70 km depth range. The forward method was applied to compute the synthetic surface-wave travel-time and Bouguer gravity anomaly data sets for this model. Then, random Gaussian noise was added to the synthetic observations, with a standard error of 1.4 s for the surface-wave travel-time data and 10.5 mGal for the gravity data. Finally, the same inversion strategy used for the real data was applied to recover the checkerboard model.

Figure 6 presents the inversion results of surface-wave-only inversion and joint inversion at four depths (6 km, 16 km, 32 km, 50 km). The surface-wave-only inversion generally recovered the velocity structure (**Figure 6A–D**). However, due to uneven ray distribution, the checkerboard pattern in edge regions with sparse coverage was not well recovered by surface-wave-only inversion, and the recovery performance of inversion deteriorated progressively with increasing depth. In contrast, the joint inversion results (**Figure 6E–H**) show improved recovery both outside

the ray-paths and across all depth ranges up to 60 km, outperforming the surface-wave-only inversion and closely reproducing the true model. This depth-dependent degradation is further confirmed by the extracted 1D velocity profile (Figure 7).

3.3. Results

3.3.1. Fitting analysis of surface-wave travel-time and gravity data

The residuals between theoretical travel times calculated from the initial model and observed travel times are distributed primarily within ± 12 s, with an RMS of 4.71 s (Figure 8A). After surface-wave-only inversion, the distribution of travel-time residuals clearly converged toward the center, mostly within ± 6 s, and the RMS reduced to 2.69 s (Figure 8B). The distribution of travel-time residuals after joint inversion was similar, also concentrated within ± 6 s. However, the RMS is 2.73 s (Figure 8C), slightly higher than that of the surface-wave-only inversion.

The initial density model was derived by converting the initial velocity model using an empirical relationship. Forward modeling of gravity anomalies was then performed on this initial density model to obtain the residuals between theoretical and observed gravity anomalies (Figure 9A). The results show that the gravity anomaly residuals range from -300 to 200 mGal, with an RMS of 86.3 mGal. The spatial distribution of residuals shows a distinct pattern: high values at the eastern and western edges and low values in the center. The gravity anomaly residuals from the surface-wave-only inversion results are smaller than those from the initial model. Compared with the residuals from the forward modeling of the initial model, larger residuals are concentrated in the Sichuan Basin, with the RMS decreasing to 83.8 mGal (Figure 9B). The gravity anomaly residuals are mainly distributed between -220 and 260 mGal and do not fit the gravity anomaly data well. In contrast, the gravity anomaly residuals after joint inversion are significantly reduced. Regions with relatively large residuals are concentrated in the Songpan–Ganzi Block and Sichuan–Yunnan Block, and the residuals are mainly confined to a small range of approximately ± 20 mGal, with the RMS reduced to 14.7 mGal (Figure 9C). Therefore, the joint inversion results can effectively fit both surface-wave travel times and gravity data.

3.3.2. Comparison of results from different inversion methods

The results of surface-wave-only and joint inversions were converted using the velocity–density empirical relationship to obtain the density structure of the study area at different

depths (Figure 10). The density structure obtained from surface-wave-only inversion and joint inversion is generally similar, but there are still some differences. At a depth of 6 km, the density difference between the two sides bounded roughly by the Wulianfeng Fault and Xieziba Fault is more pronounced in the joint inversion results, which is consistent with the findings of Xia *et al.*²⁵ At 12 km depth, the joint inversion results reveal a larger-scale high-density anomaly southwest of the epicenter of the Changning $M_s 6.0$ earthquake, with obvious density differences from other directions, exhibiting distinct lateral inhomogeneous distribution.^{18,19} At a depth of 32 km, the boundary between the Sichuan Basin and the adjacent orogen is more distinct in the tomographic results from the surface-wave-only inversion.

3.3.3. Results of joint inversion

In the horizontal direction, the density structure at different depths in the joint inversion results (Figure 11) exhibits significant lateral inhomogeneity. At a depth of 6 km (Figure 11A), the density structure is consistent with the surface geological structure, with a distinct boundary between mountainous areas and the basin.^{20,25,41} A prominent low-density anomaly develops within the Sichuan Basin, further confirming the presence of a thick sedimentary layer.

At a depth of 12 km (Figure 11B), the low-density anomaly in the Sichuan Basin gradually concentrates toward the southwest margin, and the density in the central part of the basin is higher than that in the southwest part of the basin; low-density anomalies appear on the eastern side of the Xihe–Meigu Fault and the eastern side of the Zhaotong–Ludian Fault. Additionally, the Changning $M_s 6.0$ earthquake and its surrounding area exhibit distinct lateral inhomogeneity in density structure.¹⁸

As depth increases to 20 km (Figure 11C), the density in the basin increases rapidly, and the low-density anomaly basically disappears. The Sichuan Basin exhibits a high-density anomaly, while its surrounding areas have relatively low density, in stark contrast to the density structure at 6 km depth. Within the Sichuan Basin, the density gradually exceeds that in the southeastern Qinghai–Xizang Plateau. This density reversal pattern is caused by the greater crustal thickness of the Qinghai–Xizang Plateau: at the same depth, the Qinghai–Xizang Plateau is still within the crust, whereas the Sichuan Basin has already entered the lower crust or even the crust–mantle transition zone.^{20,42,43} Furthermore, the low-density anomaly on the eastern side of the Zhaotong–Ludian Fault at 12 km depth extends northward to the southern margin of the Sichuan Basin, and then spreads along both sides of the margin after reaching it, with a larger eastward extension.⁴³

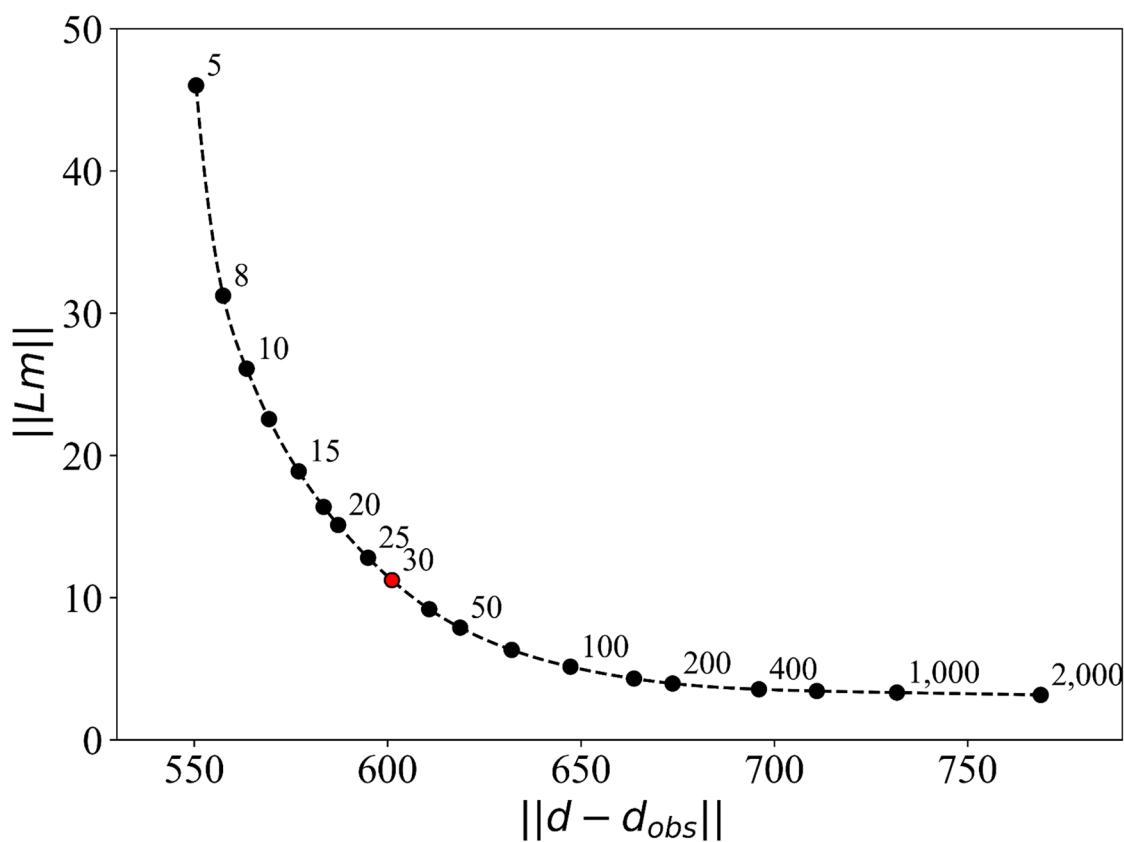


Figure 5. L-curve analysis in the joint inversion method. The red dot represents the optimal smoothing factor

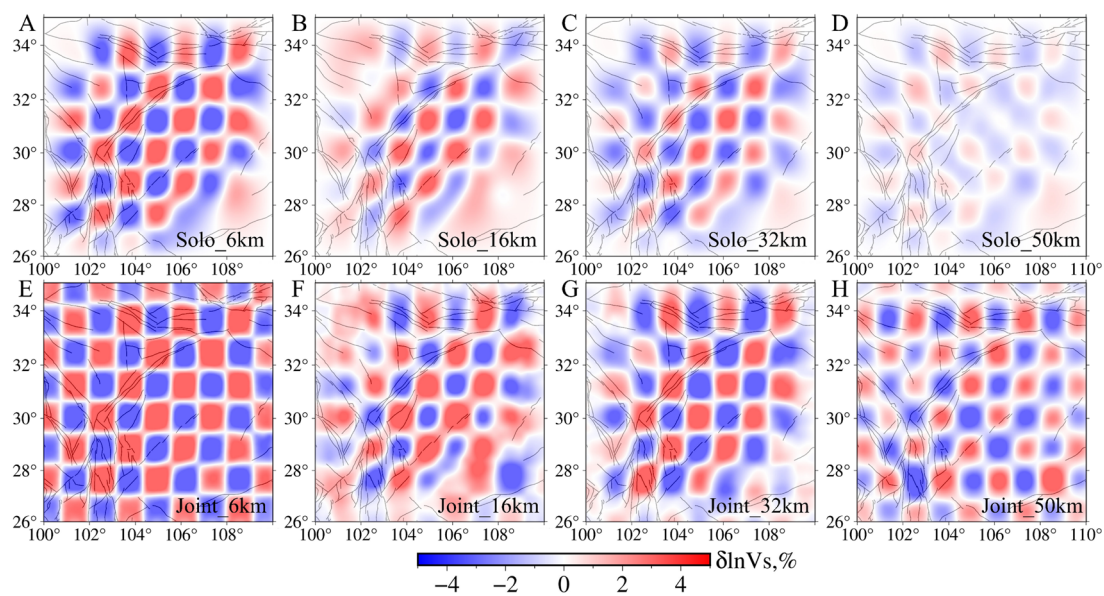


Figure 6. Horizontal slices of the inversion model at 6, 16, 32, and 50 km depth. (A–D): Results of surface-wave-only inversion; (E–H): Results of joint inversion.

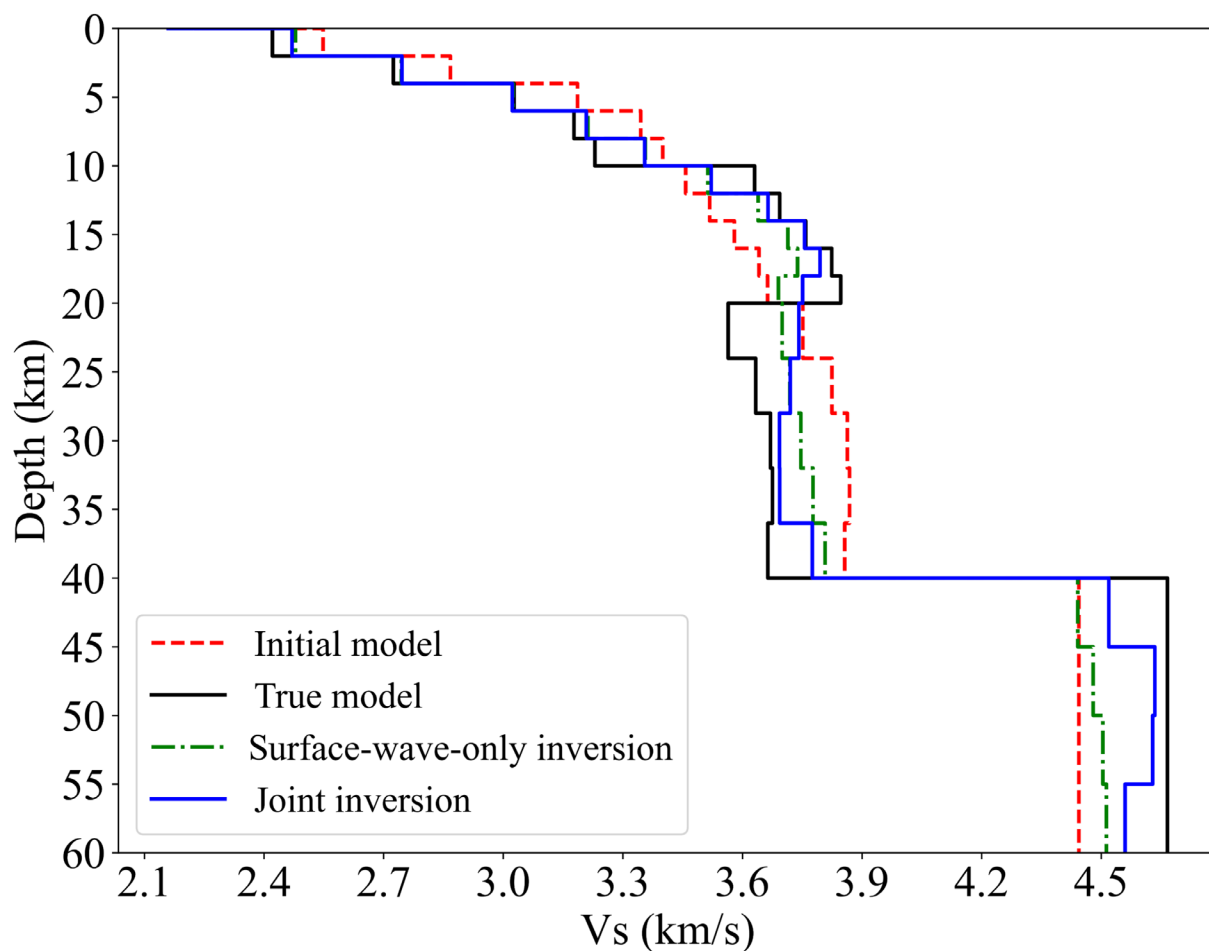


Figure 7. One-dimensional velocity profile comparison from the checkerboard test

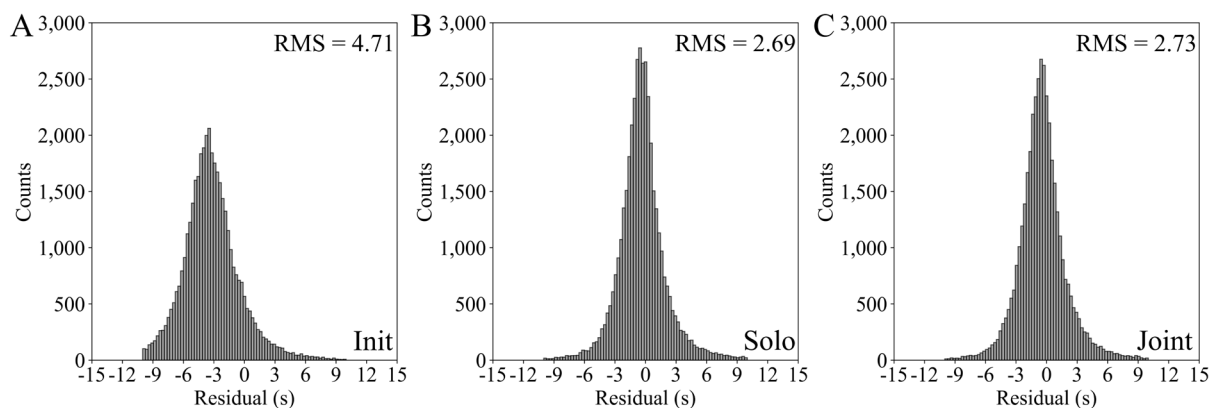


Figure 8. Histograms of travel-time residual distribution. (A) Travel-time residuals of the initial model; (B) travel-time residuals after surface-wave-only inversion; (C) travel-time residuals after joint inversion.

Abbreviation: RMS: Root mean square.

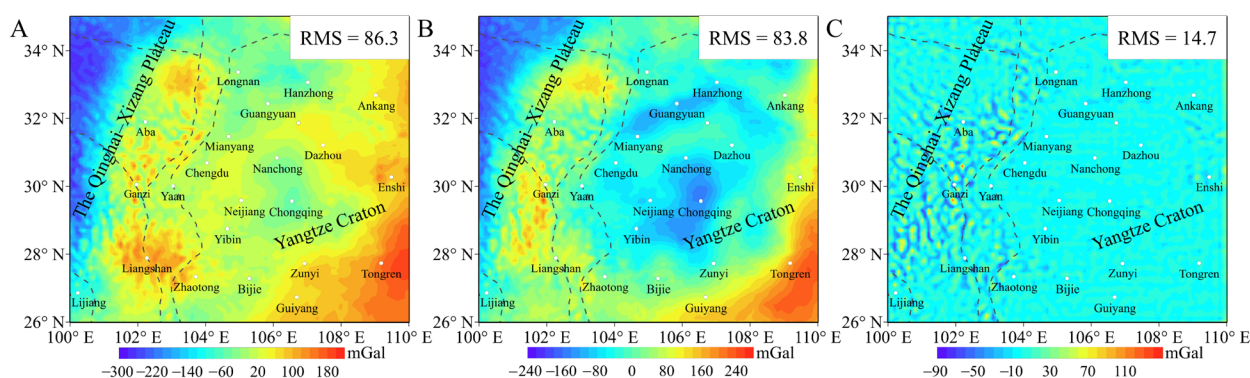


Figure 9. Residuals between observed and theoretical gravity anomalies for different models. (A) Gravity anomaly residuals of the initial model; (B) gravity anomaly residuals from surface-wave-only inversion; (C) gravity anomaly residuals from joint inversion.

Abbreviation: RMS: Root mean square.

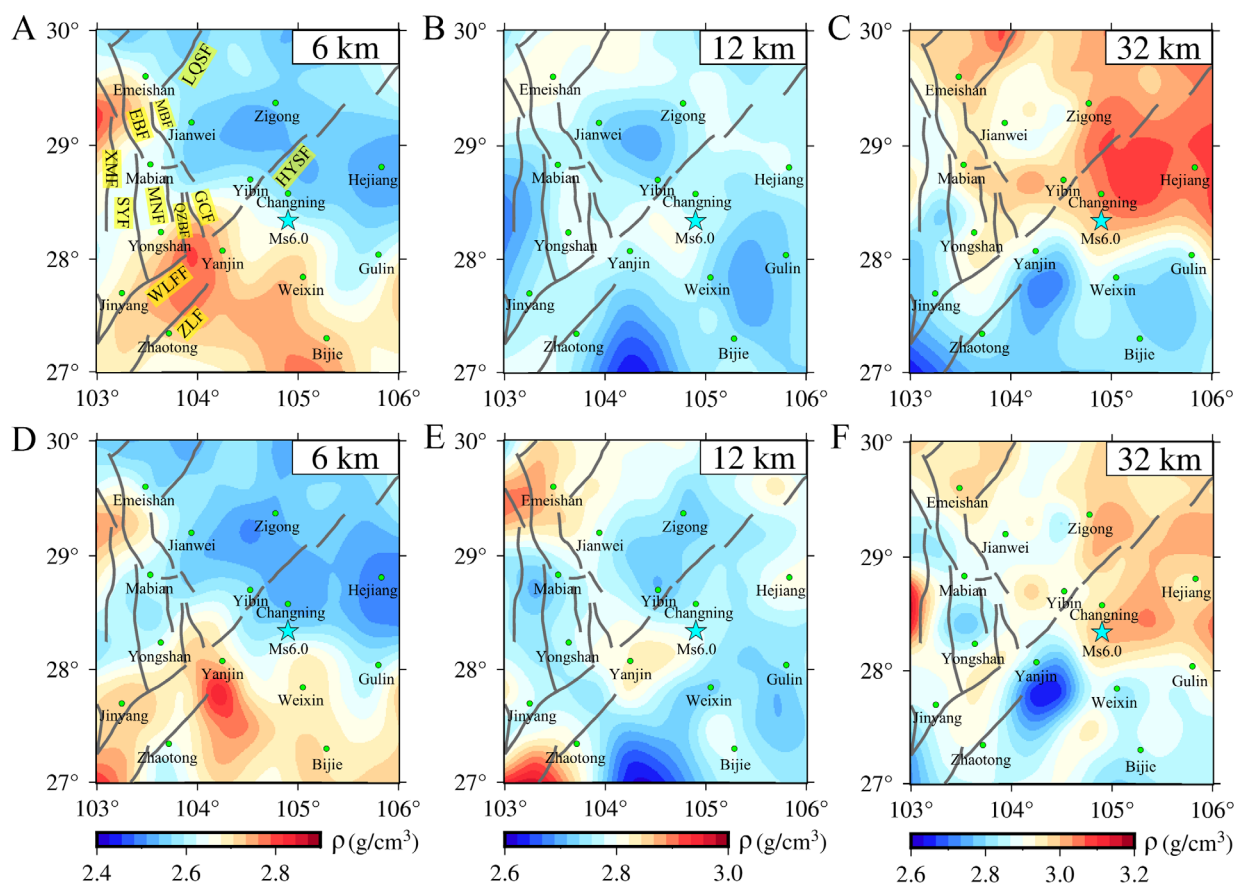


Figure 10. Comparison of tomography results from surface-wave-only inversion and joint inversion at depths of 6 km, 12 km, and 32 km. (A–C): Results of surface-wave-only inversion; (D–F): Results of joint inversion. Gray solid lines represent major active faults, and the cyan five-pointed star represents the 2019 Changning $M_s 6.0$ earthquake.

At a depth of 28 km (Figure 11D), the low-density anomaly on the eastern side of the Zhaotong–Ludian Fault extends northward, with an area smaller than that at 20 km depth, and only extends on the eastern side of the southern Sichuan Basin. The epicenter of the Changning $M_s 6.0$ earthquake is located in the gradient zone where the center of the low-density anomaly transitions into the high-density anomaly inside the basin; the low-density anomaly on the eastern side of the Xihe–Meigu Fault still persists.²⁰

At a depth of 32 km (Figure 11E), the low-density anomaly on the eastern side of the Zhaotong–Ludian Fault continues to shrink, and its extension trend almost disappears; the range of the low-density anomaly on the eastern side of the Xihe–Meigu Fault shrinks, while the high-density anomaly on its western side gradually expands compared with that at 28 km.

At a depth of 36 km (Figure 11F), the low-density anomaly on the eastern side of the Zhaotong–Ludian Fault almost disappears, while the high-density anomaly on the western side of the Xihe–Meigu Fault continues to expand eastward. However, overall lateral inhomogeneity weakens with increasing depth.^{18,42,43}

In the vertical direction, density tomography was performed along the six profiles shown in Figure 1B, producing the results displayed in Figure 12. The joint inversion results show that a prominent, continuous low-density anomaly develops in the middle-lower crust of the southern margin of the Sichuan Basin. Along the northwest–southeast direction, the low-density anomaly extends horizontally for approximately 220 km along profile CC', occurring between approximately 30 km and 250 km along the profile. It has a vertical thickness of 10–30 km and forms a large-scale continuous low-density anomaly. Before reaching profile AA', the anomaly expands both eastward and westward, with a markedly greater eastward extent. At profile BB', only the eastward extension remains, whereas the westward extension is nearly absent.

From the southwest–northeast direction, the low-density anomaly develops in the tectonic transition zone between the high-topographic mountainous area in the southeastern Qinghai–Xizang Plateau and the low-topographic Sichuan Basin. Vertically, it gradually extends toward the Sichuan Basin from a depth of 15 km, thins toward the northeast, and eventually disappears at 35 km. The anomaly is most prominently developed on Profile DD'. On Profile EE', the low-density anomaly almost disappears, while on Profile FF', the anomaly still maintains a prominent low-density anomaly and extends continuously toward the southeast direction. Furthermore, the Wulianfeng Fault and Zhaotong–Ludian Fault develop on the left side of Profile DD', which is also adjacent to

the eastern boundary of the Sichuan–Yunnan Block. In contrast, the regions on the right side of Profile DD' and where Profile FF' is located exhibit weak fault activity and limited ability to restrict the development of the anomaly, accounting for the significant differences in the degree of anomaly development across different directions.

4. Discussion and analysis

4.1. Intracrustal low-density layer in southern Sichuan Basin

A prominent, continuous intracrustal low-density layer occurs along the southern margin of the Sichuan Basin (Figure 12), exhibiting a zonal distribution mainly along the eastern side of the Zhaotong–Ludian Fault and along the southern margin of the Sichuan Basin. Its spatial location is consistent with previous studies' results.^{20,25,44–46} However, due to the limitations of research scope and imaging resolution, the detailed distribution pattern of this low-density layer has not been specifically demonstrated. The findings from this study clearly show that the low-density layer is centered on Yanjin, with its western extent extending to the Xieziba Fault and Guancun Fault, and its eastern limit terminating at the study area (Figure 11C). This low-density layer is predominantly hosted within the Upper Paleozoic strata (Figure 1B). Notably, the southern Sichuan Basin lies within the active tectonic belt at the southeastern Qinghai–Xizang Plateau. The continuous convergence between the Indian Plate and the Eurasian Plate has not only induced intense deformation of the lithosphere in the Qinghai–Xizang Plateau but also driven substantial volumes of low-density crustal material to escape toward the plateau's southeastern margin.^{47–49} Thus, the low-density layer identified in this study may be genetically linked to the crustal flow escaping southeastward from the Qinghai–Xizang Plateau. However, previous studies have shown that as the crustal material migrates southeastward, passes through the Lijiang–Xiaojinhe Fault zone, and enters the South Chuan–Dian Block, its movement is obstructed by the Emeishan Large Igneous Province. Instead of continuing southeastward, the crustal material turns to migrate southwestward.^{48,50} Furthermore, this low-density layer is separated from the low-velocity layer of the Sichuan–Yunnan rhombic block by the high-velocity body adjacent to the Anninghe–Zemuhe Fault Zone.⁴⁶ Collectively, these lines of evidence indicate that the identified low-density layer is not the ductile material escaping from the Qinghai–Xizang Plateau toward its southeastern margin.

To further elucidate its material composition and formation mechanism, this study integrates multi-source observational data to conduct a comprehensive analysis.

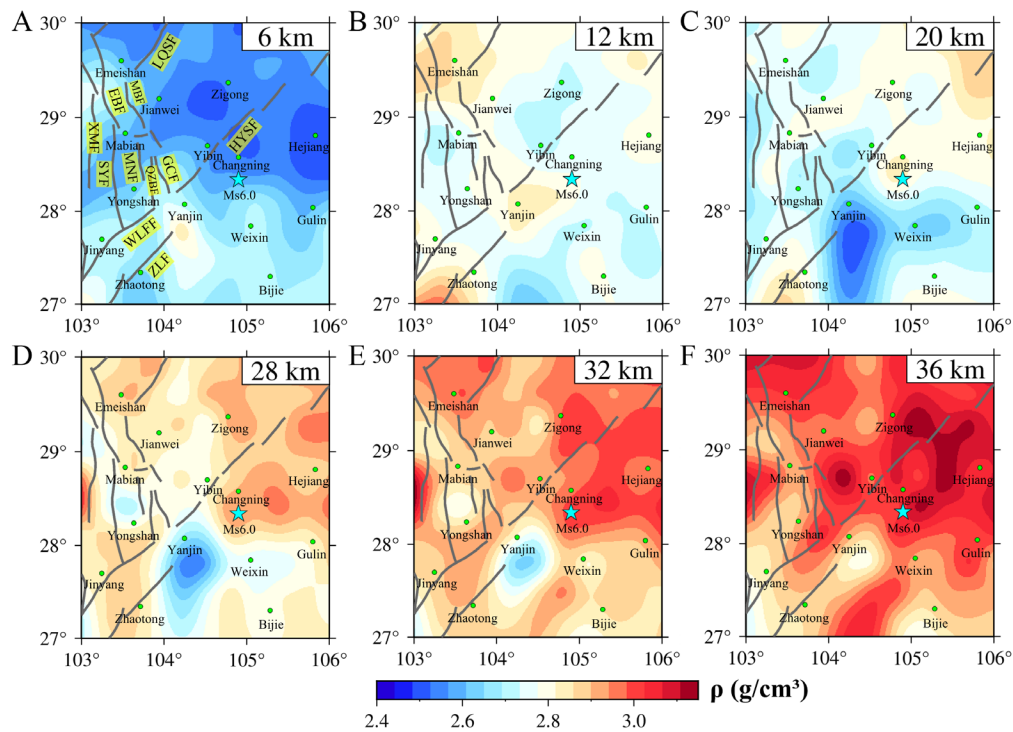


Figure 11. Density tomography results of the joint inversion at different depths. (A) 6 km; (B) 12 km; (C) 20 km; (D) 28 km; (E) 32 km; (F) 36 km.

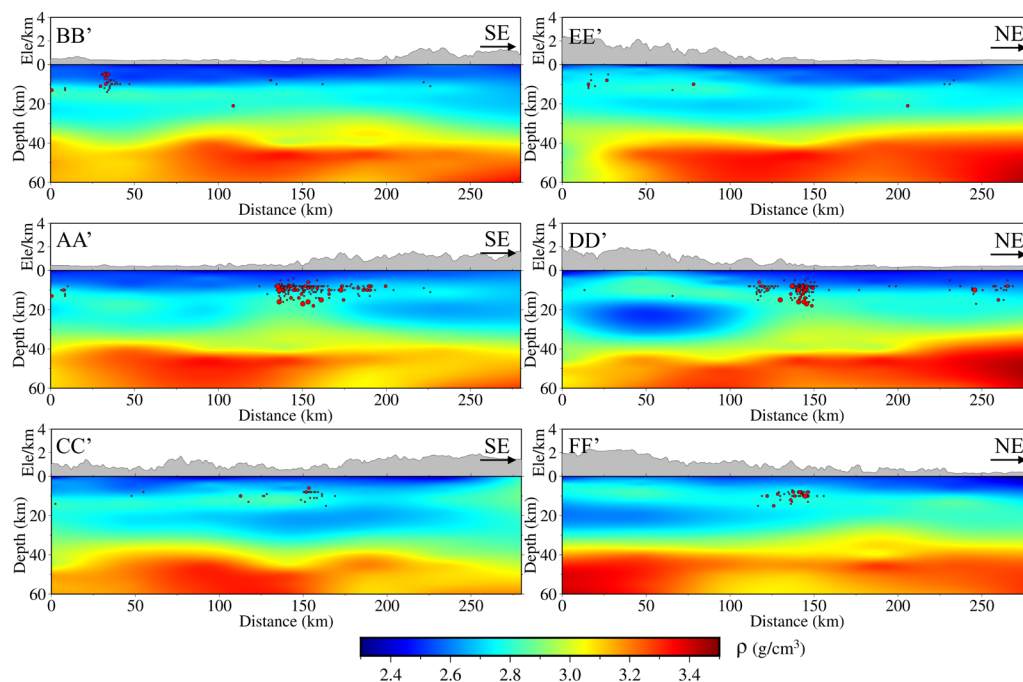


Figure 12. Density tomography results of the joint inversion along six profiles. Red dots represent $M_s \geq 3.0$ earthquakes recorded between 2019 and 2025 within 25 km of the profiles, with dot size indicating magnitude. Abbreviations: NE: Northeast; SE: Southeast.

Poisson's ratio is quite sensitive to rock composition, and the low-density layer region exhibits a low Poisson's ratio.⁵¹ Low Poisson's ratio corresponds to more felsic components, leading to the inference that the crustal components in this area are likely dominated by felsic materials.⁵¹ Felsic materials, compared with mafic materials, have lower densities, as well as relatively lower solidus temperatures and rheological strengths, making them more susceptible to melting.^{51–53} Additionally, the Moho discontinuity depth in this region indicates that the crustal thickness is moderately greater than the average continental crustal thickness. Thus, crustal thickening caused by plateau uplift may have elevated crustal temperature in this region.^{51,54} Meanwhile, seismic attenuation is more sensitive to temperature and hydrothermal activity, and Lg waves undergo strong attenuation as they pass through this region.^{48,50} Additionally, the terrestrial heat flow in this region is relatively high.⁵⁵ Magnetotelluric results also show that the crust east of the Zhaotong–Ludian Fault is characterized by low resistivity and high conductivity, suggesting the presence of fluids or partial melting.^{56,57} Based on the comprehensive analysis above, it can be inferred that the low-density layer in this area is formed by plastic deformation or partial melting of felsic crustal components at relatively high temperatures.

The presence of partial melting not only significantly reduces viscosity but also greatly weakens the crustal strength in this region, providing a prerequisite for the lateral flow of this low-density layer.^{52,53} This allows the low-density layer to undergo ductile flow driven by the topographically induced pressure gradient.⁵⁸ Located in the southern Sichuan Basin, this region forms a steep topographic transition zone with the Sichuan Basin's interior. Therefore, it can be inferred that the low-density material in the southern margin undergoes ductile flow toward the Sichuan Basin driven by the topographic pressure gradient.

4.2. Relationship between dynamic processes and seismic activity

As illustrated in Figure 1B, most seismic activity in the study area has been concentrated in the Changning area in recent years. To further analyze the relationship between seismic activity and density structure, seismic events within approximately 25 km on both sides of the profiles are projected onto the density structure, as shown in Figure 12. Profiles DD', EE', FF', and Figure 13 collectively demonstrate that seismic activity in the study area is concentrated mainly in the junction zone between the low-density layer and the edge of the Sichuan Basin, where Triassic strata dominate (Figure 1B). Meanwhile, existing studies suggest that if the weak, low-velocity zone material in the mid-lower crust is

blocked by rigid high-velocity bodies, it will result in steep topographic changes on the surface; otherwise, relatively gentle topographic changes will occur.^{47,59} The crust of the Sichuan Basin is cold, rigid, and thin, lacking developed low-velocity zones or Cenozoic magmatic activities. It has high strength and good stability, forming a rigid block that resists deformation.^{60,61} Therefore, evidence from three aspects, including seismic distribution, surface topographic relief, and lateral flow of the low-density layer, directly or indirectly proves that this low-velocity layer is expanding toward the Sichuan Basin. Blocked by the strong crustal block of the Sichuan Basin, the crustal material cannot penetrate the strong crustal block and instead flows parallel to the edge or expands around both sides of the basin⁵⁸ (Figure 13). Furthermore, during the lateral expansion of the low-density layer, the western side is affected by active faults, such as the Huayingshan Fault and block compression constraints. The shear deformation of the fault zone absorbs part of the material's momentum during migration.⁴⁷ In contrast, the eastern side is not cut by large strike-slip faults, and the crustal medium is relatively homogeneous, allowing it to extend continuously eastward along the southern margin of the Sichuan Basin, forming a broader eastward extension range (Figure 11C).

Due to the obstruction, the leading edge of the low-density layer cannot expand, resulting in strain and stress accumulation in the leading edge region of the contact zone. When the accumulated stress exceeds the fracture threshold of the overlying brittle crust, energy is released in the form of earthquakes. The middle part of profile AA' in Figure 12 is the stress concentration zone formed by the low-density layer and the edge of the Sichuan Basin. The frequent seismic activity in this region is a direct manifestation of the high regional stress concentration caused by the blocked lateral expansion of the low-density material generated by melting toward the Sichuan Basin. Based on the comprehensive analysis above, the low-density layer and stress accumulation resulting from its interaction with the rigid Sichuan Basin are the key controlling factors for the frequent earthquakes in the Changning area.

5. Conclusion

This study utilizes surface-wave dispersion data and EGM2008 Bouguer gravity anomaly data, coupled with a joint tomographic inversion of ambient noise and gravity data to obtain the subsurface density structure of the southern Sichuan Basin, yielding the following key conclusions:

- (i) The joint inversion significantly improves reliability and effectively compensates for the limitations of inversion with a single dataset. Compared with surface-

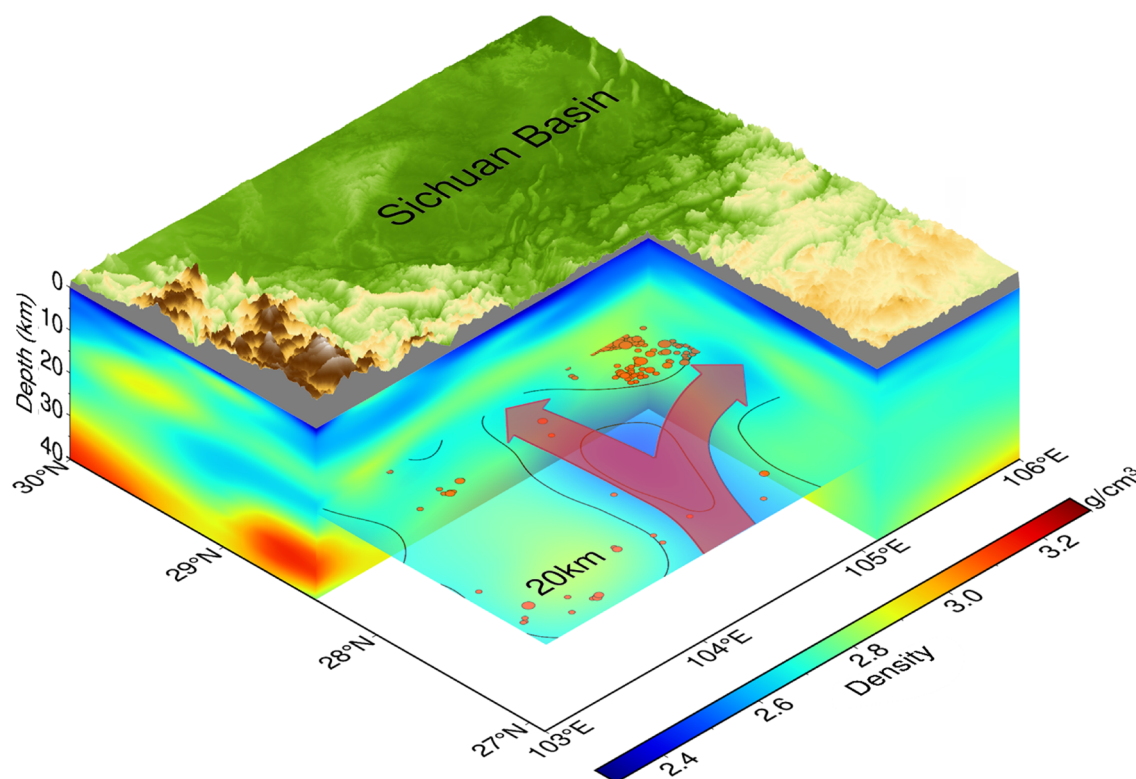


Figure 13. Three-dimensional density structure and the low-density layer in the southern Sichuan Basin. Red dots represent earthquakes. Red arrows denote the migration direction of low-density materials, which, after being obstructed by the rigid crust of the Sichuan Basin, migrate along both the eastern and western margins of the basin.

wave-only inversion, the combination of surface-wave dispersion data and Bouguer gravity anomaly data enables better recovery of the true subsurface model in edge regions with sparse ray coverage, and across all depth ranges within 60 km. Additionally, the RMS of surface-wave travel-time residuals from the initial model decreases from 4.71 s to 2.73 s, while the RMS of gravity anomaly residuals decreases from 86.3 mGal to 14.7 mGal. The results demonstrate good fitting performance with both surface-wave travel-time and Bouguer gravity anomaly observations, reducing the non-uniqueness of the inversion solutions and improving the reliability of the inverted model.

- (ii) The joint inversion reveals that an intracrustal low-density layer exists in the southern Sichuan Basin. Comprehensive analysis, combined with other geophysical data, indicates that this low-density layer is not the ductile material escaping from the Qinghai-Xizang Plateau toward its southeastern margin, but rather a low-density layer formed by plastic

deformation or partial melting of felsic components within the crust under relatively high-temperature conditions. Driven by the topographic pressure gradient, the low-density layer undergoes ductile flow toward the Sichuan Basin. Obstructed by the rigid crust of the Sichuan Basin, the material migrates along both the eastern and western margins of the basin, with a broader eastward migration range.

- (iii) The stress accumulation induced by the interaction between the low-density layer and the rigid Sichuan Basin is the key controlling factor for frequent earthquakes in the Changning area. Seismic activity in the study area is predominantly concentrated in the junction zone between the low-density layer and the margin of the Sichuan Basin, with the leading edge of the low-density layer exhibiting a distinct spatial correspondence to the seismic activity concentration zone in the Changning area. This region is a stress-concentration zone jointly shaped by deep material migration and tectonic obstruction. Blocked by

the rigid basement of the Sichuan Basin, the lateral flow of low-density material is hindered, leading to continuous accumulation of regional stress, which acts as the key controlling factor responsible for the frequent earthquakes in the Changning area.

Acknowledgments

We sincerely acknowledge Nanqiao Du from the Institute of Geology and Geophysics, Chinese Academy of Sciences, for assistance and valuable suggestions provided during the experimental process.

Funding

This research was supported by the Deep Earth Probe and Mineral Resources Exploration-National Science and Technology Major Project (2025ZD1010201), the National Natural Science Foundation of China (42574074), the Hubei Provincial Natural Science Foundation of China (2025AFB854), and the Scientific Research Fund of the Institute of Seismology, China Earthquake Administration (IS202456377).

Conflict of interest

The authors declare that they have no competing interests.

Author contributions

Conceptualization: Li Zhang, Guiju Wu, Rugang Xu

Formal analysis: Guiju Wu, Rugang Xu, Yufei Xi

Investigation: Rugang Xu, Yufei Xi, Fan Wang, Li Zhang

Methodology: Guiju Wu, Li Zhang, Fan Wang

Writing—original draft: Li Zhang, Jiawei Wang, Rugang Xu

Writing—review & editing: Guiju Wu, Li Zhang, Yufei Xi

Availability of data

The surface-wave travel times at the different periods described in this study are available at <https://github.com/nqdu/JDSurfG>, and the EGM2008 data are available at http://earth-info.nga.mil/GandG/wgs84/gravitymod/egm2008/egm08_wgs84.html.

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