Magnetotelluric evidence for distributed lithospheric modification beneath the Yinchuan-Jilantai rift system and its implications for Late Cenozoic rifting in western North China

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Key Points:

- Rift-related, fluid-rich fault zones are imaged as subvertical, crustal-scale conductors
- Enhanced conductivity in the uppermost mantle is attributed to volatile-enriched melts
- Inherited lithospheric rheology and the India-Eurasia collision induced far-field forces control the rifting structures
Abstract

The Yinchuan-Jilantai rift system (YJRS) is a prominent Cenozoic intracontinental rift zone located along the northwestern margin of the Ordos Block in western North China. Although the absence of volcanism indicates a passive origin, the tectonic driving forces and rift-related deep processes remain poorly understood. Here we use newly obtained broadband magnetotelluric (MT) data to image the lithospheric electrical structure along a ~500-km-long profile across the YJRS. At middle–lower crustal levels, the resulting model reveals several subvertical, crustal-scale conductors that spatially correlate with the rift-parallel, high-angle normal faults. We attribute these features to a combined effect of saline fluids and partial melt due to recent supply of heat and volatiles into the crust from below. The crustal-penetrating normal faults that were reactivated during extension serve as permeable pathways for deep fluid migration. In the uppermost mantle, a ~400-km-wide zone of enhanced conductivity is present and requires the presence of partial melt. We interpret this feature as evidence for lithospheric modification through upwelling and decompressional melting of the volatile-enriched mantle. When compared with the narrow (<100-km-wide) mantle conductors imaged beneath the Shanxi rift system along the eastern margin of the Ordos Block, such a feature indicates the YJRS has experienced a larger extent of lithospheric modification. Combined with additional geologic and geophysical observations, we attribute this essential difference to the inherited lithospheric heterogeneity and the eastward decreasing far-field impacts from the India-Eurasia collision.

Plain Language Summary

Throughout the Cenozoic, the western North China Craton (NCC) has undergone widespread extensional deformation, forming large-scale rift systems around the stable Ordos Block (OB). Although these rifts are active today, the tectonic driving forces and related deep processes remain poorly constrained. We use magnetotelluric (MT) data, which are highly sensitive to deep fluid distribution, to investigate the rift-related fluid dynamics beneath the Yinchuan-Jilantai rift system (YJRS) along the northwestern margin of the OB. In deep crust, we find several isolated, subvertical high-conductivity features that spatially correlate with the rift-parallel normal faults, likely representing pathways for deeply-sourced fluids. Looking deeper, we identify a ~400-km-wide zone of enhanced electrical conductivity in the uppermost mantle, which is interpreted as
evidence for lithospheric modification through low-degree melting of the volatile-enriched mantle. By comparing our results with previous MT observations, we find that the lithosphere of the YJRS has undergone a larger extent of modification than that of the Shanxi rift system (SXRS) along the eastern margin of the OB. Such a difference matches with the surface features of the two rift systems and can be attributed to the inherited lithospheric heterogeneity and the eastward decreasing far-field impacts from the India-Eurasia collision.

1 Introduction

Despite the large distances (> 2000 km) from active plate boundaries, the western North China Craton (NCC) in the interior of the Eurasian continent has undergone widespread extension during the Cenozoic, forming a series of elongated rift systems around the tectonically stable Ordos Block (Figure 1) (Zhang et al., 1998; Shi et al., 2020). These rifts share some characteristics typical of active continental rifts, such as elongated sedimentary grabens, elevated rift flanks, syn-depositional normal faults and high seismic activity (Figure 1a) (Olsen, 1995). However, the circum-Ordos rifts are largely amagmatic, except for small outcrops of Quaternary volcanic rocks distributed along its northeastern margin. Such a lack of magmatism has been regarded as a strong indicator of passive rifting in response to lithospheric extension driven by far-field tectonic loading stresses (Ye et al., 1987; Xu & Ma, 1992), yet it remains controversial which tectonic processes have powered the extensional forces and whether active mantle processes are occurring beneath the rifts (e.g., Molnar & Tapponnier, 1977; Liu et al., 2004; Schellart & Lister, 2005; Bao et al., 2011).

Two end-member driving mechanisms have been proposed to account for the Cenozoic extension in North China (Su et al., 2021). Some authors have suggested that this extensional event is entirely driven by the far-field effects of the India-Eurasia collision and continued convergence, through either lateral extrusion of crustal fragments along significant strike-slip faults (Molnar & Tapponnier, 1977; Tapponnier et al., 1982), or lateral flow of the asthenospheric mantle (Liu et al., 2004). At a broader scale, the formation of the Baikal rift zone in Siberia (Yin et al., 2000) and even the back-arc basins along the East Asia margin (e.g., Japan Sea, Okhotsk Sea) has also been regarded as a direct result of extrusion tectonics related to the India–Eurasia collision (Worrall et al., 1996; Xu et al., 2014). By contrast, other studies have questioned the role of India-Eurasia convergence in extensional regions outside of the Tibetan Plateau but instead argued that the
subducting Pacific plate is driving the Cenozoic extension in East Asian (Northrup et al., 1995; Schellart & Lister, 2005; Schellart et al., 2019). In such a scenario, the widespread normal and strike-slip faulting in East Asia originated from back-arc extension in the overriding lithosphere, driven by the progressive slab rollback and associated mantle flow (Schellart & Lister, 2005). Some recent studies attempt to reconcile the two competing mechanisms and suggest that the Early Cenozoic extension around the Ordos Block is mainly associated with Pacific subduction from the east, whereas the extension since the Late Cenozoic is primarily due to the extrusion tectonics of the Tibetan Plateau (Shi et al., 2020; Su et al, 2021).

The generally NE-striking Yinchuan-Jilantai rift system (YJRS), which lies along the northwestern margin of the Ordos Block, is one of the most prominent Cenozoic continental rifts and earthquake belts in China (Figure 1a). The YJRS has been struck by several devastating earthquakes in the past (RGAFSO, 1988; Hao et al., 2020), including the M 7.6 1739 Pingluo earthquake occurred within the Yinchuan Graben (Middleton et al., 2016a). It consists of a system of subparallel, northeast-southwest elongated grabens and horst-type mountain ranges bounded by normal faults with strike-slip components (Figure 2). As one of the earliest rift structures developed around the Ordos Block, the YJRS is filled with anomalously thick (> 5 km) Cenozoic sedimentary formations, making it an ideal place to study the Cenozoic rifting evolution in the western NCC (Shi et al., 2020). Comprehensive fault kinematic analysis and geochronological results suggest the YJRS has formed in response to two main phases of extension, initiated in the Eocene and then followed by a further extensional phase from the Late Miocene to the present (Shi et al., 2020). Both geological and geodetic measurements show that the YJRS is currently undergoing intensive NW–SE extension and right-lateral shearing (Middleton et al., 2016b). However, the absence of exhumed samples from magmatism prevents insights into the deep dynamic processes occurring beneath the rift. Therefore, geophysical imaging of the subsurface structure beneath the YJRS is key to understanding the rifting processes around the Ordos Block.

Among various geophysical imaging techniques, the magnetotelluric (MT) method is susceptible to electrical conductivity anomalies associated with the presence of partial melt and free aqueous fluids in the lithosphere and has therefore been extensively used to investigate the deep fluid and magmatic processes beneath continental rift zones (e.g., Wannamaker et al., 2008; Rippe et al., 2013; Dong et al., 2014; 2020; Feucht et al., 2017;2019). However, for the western NCC, previously reported dense MT observations are largely restricted to the Shanxi rift system.
(SXRS) on the eastern side of the Ordos Block (Yin et al., 2016, 2017; Zhang et al., 2017). In this study, we present results from a MT survey along a profile crossing from the western Ordos Block in the east, over the entire YJRS, into the Alxa Block in the west. We employed a 3-D inverse modeling approach to determine the crust and uppermost mantle’s electrical resistivity structure, and examined the possible origins of the major resolved electrical features and their tectonic implications. By a thorough exploration of our results and a comparison with previous MT observations from the nearby regions, we intend to give new insight into the dynamic processes that control the Late Cenozoic rifting in the western NCC.

2 Geologic Background

The Ordos Block is surrounded by the Qinling Orogen to the south, the Yinshan-Yanshan Orogen to the north, the Qilian Orogen and Alxa Block to the west and the Trans-North China Orogen to the east (Figure 1a). This block has an early Precambrian basement similar to the eastern NCC, which is rarely exposed due to thick (4–6 km) Mesoproterozoic to Phanerozoic sedimentary cover (Wan et al., 2013). Unlike the eastern part of the NCC, where the lithosphere underwent dramatical modification and thinning during the Late Mesozoic (Xu et al., 2009; Zhu et al., 2012), the western NCC is believed to be much less involved in this process and its thick cratonic lithosphere remain largely preserved (Menzies et al., 2007; Chen et al., 2009; Bao et al., 2013). During the Cenozoic, intense extensional deformation has occurred around its periphery, but its interior remains stable and intact (Shi et al., 2020). The stability of this block is supported by the absence of modern earthquake activity and Phanerozoic magmatism within the block (Figure 1a) and is consistent with seismic data showing a high-velocity anomaly extending down to the upper mantle (Huang & Zhao, 2006; Bao et al., 2013). However, some seismic and MT surveys reveal prominent low-velocity and low-resistivity anomalies in the crust and upper mantle beneath the Ordos Block (Tian et al., 2009; Jiang et al., 2013; Dong et al., 2014; Ye et al., 2020), implying lithospheric thinning and reworking probably also occur beneath the block, particularly beneath its marginal areas.

The Alxa Block is located between the northeastern Tibetan Plateau and the Ordos Block. This block is largely covered by Mesozoic–Cenozoic sedimentary deposits, with Precambrian basement rocks being sporadically exposed in its eastern and southern parts (Wu et al., 2014; Dan et al., 2016). Traditionally, the Alxa Block was considered as the westernmost part of the NCC,
and collided with the NCC during the Proterozoic (Zhao et al., 2005). However, some recent studies argued that the two blocks were independent, at least in the Early Paleozoic (Dan et al., 2016; Yuan & Yang, 2015). During the Late Paleozoic, the northern portion of this block has been extensively reworked by the southward subduction of the Paleo-Asian Ocean (Xiao et al., 2018). The by-products of the subduction–collision processes include the Enger Us and Qagan Qulu ophiolitic mélanges and the widespread magmatic rocks along the NE-trending Shalazha and Bayan Nuru mountain ranges (Figure 2) (Xiao et al., 2018). These two mountain ranges, plus the Bayanwula mountain to the east, represent the unloaded and tilted footwall blocks of the Cenozoic rift system (Zhang et al., 2021). In contrast to the Ordos Block, the Alxa Block exhibits a high level of seismicity, suggesting that the fault systems within the block remain active nowadays (Figure 1a). Geodetic measurements of present-day deformation field show that the Alxa Block is currently undergoing a NE–SW compression and moves northward at a rate of ~3 mm/a relative to the Ordos Block (Figure 1a) (Wang & Shen, 2020; Hao et al., 2021).

Between the Ordos and Alxa blocks, the NNE-trending Helan Shan is a horst undergoing a NE–SW-directed compression. Geological studies suggest that both bounding were thrust faults during the Late Jurassic and inverted to normal faults during Cretaceous to Cenozoic times (Yang & Dong, 2018). Rock units exposed in the Helan Shan include high-grade metamorphic Archean to Proterozoic basement rocks and Paleozoic cover sequences (Figure 2; Yang & Dong, 2018). Fission track analysis suggests the latest rapid uplift of the Helan Shan began after the Late Miocene (10–12 Ma), accompanied by subsidence of the Yinchuan and Jilantai grabens (Liu et al., 2010).

Bounded by the eastern Helanshan fault (EHF) and the Yellow River fault (YRF), the Yinchuan Graben east of the Helan Shan is approximately 60 km wide and 160 km long (Figure 2). Except for the two bounding faults, the graben is cut by two major blind faults as indicated by seismic reflection data (Huang et al., 2016; Liu et al., 2017). These normal faults, which are in accordance with a NW–SE directed extension, control the Cenozoic sedimentation in the graben (Shi et al., 2020). According to the seismic data, the YRF dips westward and extends to 60 km depth, while the EHF dips eastward and meets the YRF at a depth of ~20 km (Huang et al., 2016; Liu et al., 2017). It has been suggested that the EHF might host the M 7.6 1739 normal-faulting Pingluo earthquake (Middleton et al., 2016a). The graben is filled with ~7-km-thick Paleogene–Neogene sedimentary rocks and 1.6 km thick Quaternary sediments (Huang et al., 2016; Shi et al.,

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Both geologic and geodetic measurements are consistent with a maximum extension rate of ~1 mm/a and a significant component of right-lateral strike-slip displacement across the graben (Middleton et al., 2016b; Wang & Shen, 2020).

West of the Helan Shan, the Jilantai Graben is bounded by the west Helanshan fault (WHF) to the east and the Langshan-Bayanwula fault (LBF) to the northwest (Figure 2). Like the Yinchuan Graben, extensive NW–SE extension occurred during the Cenozoic, resulting in thick sediments (~6 km) and a series of syn–depositional normal faults within the graben (Shi et al., 2020). Unfortunately, unlike the well-studied Yinchuan Graben, the geometries of these faults are largely unknown due to the lack of high-resolution seismic constraints.

South of the Helan Shan, the NS-trending Liupan Shan separates the stable Ordos Block from the eastern edge of the actively deforming northeastern Tibetan Plateau. To the west of the Liupan Shan, a series of active strike-slip and thrust faults accommodate the continuing expansion of the northeastern Tibetan Plateau. Fission-track data from the Liupan Shan suggest a rapid cooling event occurred at ~8 Ma, which has been regarded as a sign of the onset of crustal shortening in response to far-field effects of the India-Asia collision (Zheng et al., 2006).

3 Magnetotelluric Method and Data

Magnetotellurics is a passive geophysical method for imaging the electrical resistivity of the subsurface by using the naturally occurring electric (E) and magnetic (B) fields measured at the Earth’s surface (Chave & Jones, 2012). The electric (E_H) field is related to the horizontal magnetic (B_H) field by a second-rank frequency-dependent complex tensor that is called the impedance tensor (Z), which can be mathematically described by

\[ E_H = Z \cdot B_H \] (1)

The four components of Z can be further represented as apparent resistivity (\( \rho_a \)) and phase (\( \phi \)) responses as a function of period at each site:

\[ \rho_a = \frac{|Z|^2}{\omega \mu_0} \] (2)

\[ \phi = \tan^{-1}\left(\frac{\text{Imag}[Z]}{\text{Real}[Z]}\right) \] (3)
The vertical magnetic transfer function (T), also known as tipper, is a dimensionless, complex vector that describes the relationship between the horizontal magnetic field ($B_H$) and the vertical magnetic field ($B_Z$)

$$B_Z = T \cdot B_H$$  \hspace{1cm} (4)

Unlike the impedance tensor, the tipper data are free of electric galvanic distortion since they are independent of electric fields. Graphically, tipper can be represented by induction vectors whose real part points toward good conductors (in the Parkinson convention) and its magnitude represents the strength of vertical magnetic fields (Parkinson, 1962). At each sounding, the depth of investigation is a function of both the period (or its inverse, the frequency) and subsurface resistivity distribution, with longer periods and a more resistive subsurface allowing for a greater investigation depth (e.g., Feucht et al., 2017).

3.1 Data collection and processing

In May and June 2019, new broadband MT (BMT) data were collected at 40 locations along a ~500-km-long transect starting on the western Ordos Block, crossing the entire YJRS into the Alxa Block (Figures 1 and 2). The typical recording time was 24 hr at each of these BMT sites. During the whole field campaign, we deployed a remote reference site that is located ~260 km away from the center of the transect. At each of these sites, five components of the MT time-series (two electric fields and three magnetic fields) were synchronously recorded with five Phoenix MTU-5A instruments, following standard BMT field procedures (see Chave & Jones (2012) for details). These data were transformed into the frequency-domain impedance tensors (Z) and vertical magnetic transfer functions (T) using standard remote-reference, robust time series processing techniques (e.g., Jones et al., 1989; Varentsov, 2015). Since the cultural noise in the study area is generally low, smooth response curves with minor uncertainties were obtained to at least 1000 s for most sites. Figure 3 shows example sounding curves from three sites at different geological units.

3.2 Phase tensor analysis

Phase tensor analysis provides an efficient tool to examine the general electrical feature of the subsurface, with the advantage that the results are unaffected by galvanic effects (Caldwell et al., 2004). It is defined as the ratio of the imaginary part of the impedance tensor to the real part
(Φ= \text{Imag} [Z]/\text{Real} [Z]). Figures 4a and 4b show pseudo-section plots of normalized phase tensor ellipses colored with the skew angle and mean phase values respectively. Except for stations in the Helan Shan, our data show low skew values (≤4°) at short periods (< 1 s), indicating a generally low-dimensional shallow conductivity structure beneath the study area (Figure 4a). Moreover, circular ellipses are present in regions covered by young sediments (e.g., the Ordos Block and the Yinchuan and Jilantai grabens), reflecting a one-dimensional (1-D) resistivity structure. At longer periods (> 1 s), the data show varying levels of dimensionality along the profile. High skew values (≥4°) are observed for stations adjacent to the Helan Shan and west of the Bayan Nuru Shan, suggesting that the geoelectric structure in these regions is three-dimensional (3-D) at depth. Elsewhere, the skew values are generally low and consistent with a relatively simple crustal structure.

The geometric mean of the maximum and minimum phases (Φ2) shown in Figure 4b reflects the vertical conductivity gradient (see, for example, Hill et al., 2009). Phase values lower than 45° for periods between 3 and 100 s occur within the Ordos Block, corresponding to an increase in resistivity from shallow sediments to deeper crystalline basement. The Φ2 values rise again in this region at the longer periods (> 100 s), indicating that the resistivity decreases as the depth increases. Other smaller regions of low phase are present within the Helan Shan and the Alxa Block at different period ranges, implying the presence of high-resistivity anomalies with varying spatial scales. By contrast, a broad region of high phase values occurs in the middle section of the profile to a maximum period of >1000 s, suggesting a large high-conductivity area is present in the middle-lower crust and upper mantle. Also note that, at long periods, the highest phase values occur in the Yinchuan Graben and in the region between the Bayan Nuru Shan (BNS) and Bayanwula Shan (BWS), indicating the high-conductivity features are more pronounced at these locations.

3.3 Induction vectors

Induction vectors, which are primarily sensitive to lateral resistivity variations and are also insensitive to distortion by near-surface heterogeneities, are plotted for three different periods (50 s, 316 s, 1585 s) in Figure 5. In the Parkinson convention used here, the real induction vectors point away from resistive regions and toward zones of higher conductivity (Parkinson, 1962; Rippe et al., 2013). At a period of 50 s, the induction vectors are mainly affected by local conductivity
anomalies in the upper crust. Vectors at stations adjacent to the Jilantai and Yinchuan grabens point toward the center of the grabens; this likely represents the thick and conductive Cenozoic sediments as imaged in our resistivity models. At a period of 316 s, the influence of the two grabens (especially the Jilantai Graben to the west) is enhanced and the induction vectors for almost all stations point toward the basins. In general, the magnitude of these vectors decreases gradually with increasing distance from the grabens. One exception is the three stations (marked by the dashed black circle in Figure 5b) in the westernmost Ordos Block, which exhibit a smaller magnitude than those at either side and have a prevailing eastward component. For the 1585 s period, the vectors in the eastern section of the profile rotate counterclockwise toward the north and show a consistent increase in their magnitude, most likely reflecting the lower crustal and upper mantle conductors previously imaged beneath the Hetao Graben and the northern part of the Ordos Block (Dong et al., 2015; Ye et al., 2020).

4 3-D Inversion and Sensitivity tests

The above phase tensor and induction vector analysis illustrate the substantial 3-D effects in the MT data, particularly in the long period range of 100–2,000 s. In addition, the analysis of the induction vectors suggests off-profile structures significantly influence the MT responses. In this situation, applying a 2-D inversion approach to the MT profile data may lead to a resulting resistivity model that contains false anomalies (Ledo, 2005). Therefore, 3-D inversions of the profile data were required and undertaken using the 3-D MT inversion package ModEM that employs a limited-memory quasi-Newton algorithm to minimize the objective function during the inversion process (Egbert & Kelbert, 2012). To assess the resolution of model features and derive an optimal model with a reasonable misfit value, multiple inversion runs were conducted with different model setups, data subsets, and inversion parameters. We focus on the inversion models obtained from different starting/prior half-space models, model covariance and data components, which have been proven to have significant influences on the outcome of 3-D inversion (Robertson et al., 2020).

3-D inversion results derived from various starting models with different prior background resistivities are summarized in Figure 6. Both the impedance and tipper data were used to generate these models. Other parameters are the same as those used to obtain the preferred model as described below. The results show that the starting/prior model significantly influences the final
converged model, with a general trend towards more resistive final models when higher starting resistivities were used (Robertson et al., 2020). Also, there is a strong tendency for the crustal high-resistivity bodies to extend into greater depths for the models with higher starting resistivities. These findings are generally in agreement with the results of Robertson et al. (2020), which highlight the importance of choosing an appropriate starting/prior model. Nevertheless, these models consistently show some common first-order features, although their exact resistivity values and spatial extents could differ. For instance, zones of reduced resistivity in the uppermost mantle are evident in all models but with relatively higher resistivities and smaller sizes in the more resistive starting models. There are also some regions where the imaged resistivities are very similar to the prior model resistivity in these models. One prominent example is the upper mantle of the Alxa Block at the northwestern end of the profile. As suggested by Robertson et al. (2020), features within these regions should be treated with caution since the data might poorly constrain them.

While these tests cannot give a definitive answer to which model is best, the 200 Ω·m model can be precluded since it yields substantially higher initial and final root mean square (RMS) misfit values than others. On the other hand, although the 25 Ω·m model has the lowest overall RMS of all tests, the imaged crustal and mantle resistivities are too low for an intraplate region that exhibits moderate heat flow and lacks active magmatic-hydrothermal processes at the surface (Jiang et al., 2019). For the two remaining models with 50 and 100 Ω·m prior resistivities, which both fit the data well and seem geologically plausible, we prefer the latter for two major reasons: (1) a starting resistivity of 100 Ω·m is approximate to the average apparent resistivity (130 Ω·m) of all data points across all periods and sites, which has been considered as an important criterion to determine the appropriate background resistivity (Meqbel et al., 2014; Robertson et al., 2020); and (2) it is more reasonable to compare the 100 Ω·m model to other previously published resistivity models in the study area and other rift zones (e.g., Yin et al., 2016; 2017; Zhang et al., 2017), given that most of these models are also derived from a 100 Ω·m homogeneous half-space.

We also conducted several inversions to assess the impact of the model covariance parameter on the inversion solutions. In brief, this parameter controls the behavior of the model norm, larger (smaller) covariance values result in smoother (rounder) models (Robertson et al., 2020). Figure 7 summarizes the final inversion models for various covariance values (0.1, 0.2, 0.3,
0.4 and 0.5). In all of the resulting inversion models, the major electrical features in the crust and uppermost mantle are consistently present; however, their extents and resistivity values vary depending on the choice of the covariance values. Applying higher covariance values in horizontal and vertical directions generally produces more pronounced conductive features at lower crustal and uppermost mantle depths (Robertson et al., 2020). The resistive features at shallower depths were also markedly enhanced with higher covariance values. We prefer the model with a covariance of 0.3 (default covariance value used in ModEM) since it is geologically acceptable and has the lowest overall RMS value.

Shown in Figure 8 are the models calculated from the individual components of the observed MT data. These models were obtained from the 100 Ω·m starting model with the same parameters as those used to generate the preferred model. The impedance-only and tipper-only models show many common features, especially at crustal depths. However, the crust in the tipper-only model appears to be more heterogeneous than the impedance-only model. Besides, most uppermost mantle features presented in the joint inversion model are not shown in the tipper-only model, suggesting the impedance data mainly constrain them. The significant differences between the two models are not unexpected, given that the tipper data are only sensitive to lateral conductivity contrasts. Previous numerical and field experiments have demonstrated that inversion of tipper data alone can only constrain the horizontal distribution and lateral contrasts of subsurface resistivity structures, but not their depth extents and absolute resistivity values (Siripunvaraporn & Egbert, 2009; Padihla et al., 2015; Miensopust, 2017). For this reason, the qualities of the tipper-only models are generally inferior to those obtained from inverting the impedance tensor alone (Siripunvaraporn & Egbert, 2009). Nonetheless, the impedance data are potentially affected by galvanic distortion caused by near-surface inhomogeneities, whereas the tipper data are not. Thus, a joint inversion of impedances and tippers provides an optimal strategy for improving the reliability of geoelectrical models.

Figures 5 and 10e depict horizontal and vertical sections extracted from the center of the preferred 3-D resistivity model. To obtain this model, we inverted the full impedance tensor and tipper data at 24 periods in the range from 0.01 s to 5000 s. The impedance error floors were set to 5% $\sqrt{|Z_{xy} \times Z_{yx}|}$, and a constant absolute error of 0.02 was applied to the tipper data. According to the above tests, a starting resistivity of 100 Ω·m was used with a model covariance factor of 0.3
in all directions. To make the inversions less computationally expensive, the array of sites was rotated 45° counterclockwise to align with the model grid, and the data were rotated clockwise according to the grid axis (Jiang et al., 2019). Our model domain consists of a volume of 1,500 km × 1,500 km × 1400 km in the x, y, and z directions, divided into 37 × 136 horizontal cells and 70 vertical layers, respectively. Cells in the core of the model have horizontal dimensions of 5 km, and increase by a factor of 1.5 toward the outer parts of the model. Layer thicknesses start with 50 m at the surface and then increase logarithmically as a function of depth. After 109 iterations, a final RMS misfit value of 1.41 was reached, which represents a 91% reduction in data misfit relative to the starting model that has a RMS value of 15.90. The similarity between the measured and modelled data suggests that the model fits the impedance and tipper data well (Figure S1).

We conducted a series of resolution tests via a constrained inversion approach to investigate the robustness of the imaged resistivity structures in the lower crust and upper mantle. We modified the original preferred model by replacing its deep parts below specific threshold depths with a fixed resistivity of 200 Ω·m and we then restarted the inversion to assess the influence of these imposed constraints on the individual and overall RMS misfit values (Figure 9). The larger the RMS misfit increases, the higher sensitivity of the data to the removed structures. With few exceptions, significant increases in RMS occur at almost every site for the shallowest testing depth of 42 km, clearly demonstrating that the data require structures below this depth. As expected, sites above the crustal conductive regions have substantially smaller resolution depths than those located within the resistive regions. Increasing the testing depth to 61 and 78 km results in a relatively subtle but distinct increase in RMS value for most sites, suggesting the mantle structures deeper than such depths might not be tightly constrained but are still required by the data. Note that sites at the northwesternmost profile segment consistently show small, negligible changes in RMS values for the two deeper depths (i.e., 61 and 78 km). This confirms that our data poorly constrain the upper mantle structure in this area. These tests suggest that most MT sites on the profile can resolve the uppermost mantle structures with varying degrees of resolution. Estimates of penetration depth derived from a Niblett-Bostick transformation (Niblett & Sayn-Wittgenstein, 1960; Bostick, 1977) further confirm that most sites on the profile have penetration depths deeper than the base of the crust (Figure S2).
5 Results and Discussion

5.1 Upper crustal structure

The upper crustal resistivity structure (<15 km) has a good correlation with the surface geological features along the profile (Figures 10d and 10e). Generally, resistivities beneath young sedimentary basins along the profile are low due to the high-porosity materials within the basins (Chave & Jones, 2012). In western Ordos Block, a thin, continuous low-resistivity layer is observed atop the electrically resistive Precambrian basement and is interpreted to represent the Cenozoic and Mesozoic sedimentary cover of the Ordos Block. Although the thickness cannot be tightly constrained due to the conductivity-thickness trade-off and potential static shift of the MT data, the imaged low-resistivity sedimentary layers appear to be substantially thinner at the southeast end of the profile. Such a pattern is consistent with evidence from seismic receiver functions showing that the sediment thickness of the Ordos Block decreases gradually from west to east (Wang et al., 2017a, b). Within the Yinchuan and Jilantai grabens, where thick (5–10 km) Cenozoic sedimentary deposits are present (Shi et al., 2020), low-resistivity layers are also found at near-surface depths in our resistivity model, although their bases are unable to be accurately delineated due to the presence of deeper conductors in the middle and lower crust. Several small-scale, isolated conductors are found in the Alxa Block at depths shallower than 5 km, coinciding with the locations of young sedimentary basins between the mountain ranges. Besides, a relatively deep-seated (~10 km), isolated conductor is imaged beneath the Bayan Nuru Shan and perhaps reflects an upper-crustal fracture zone that was overprinted by faulting in the Cenozoic. In contrast to the conductive sediments, the upper crustal structure in regions with exposed bedrock is mainly characterized by high resistivity. For example, the upper crust is resistive beneath the two tilted and exhumed footwall blocks that bound the Jilantai Graben (Zhang et al., 2021): one is beneath the Bayanwula Shan to the northwest and the other one beneath the Helan Shan to the southeast. In northern Alxa Block, two zones of high resistivity are imaged beneath the Late Paleozoic–Triassic granitoid belts along the Bayan Nuru and Shalazha Shan respectively, which are interpreted as two cooled magmatic intrusions that formed during the southward subduction of the Paleo-Asian oceanic plate in the Late Paleozoic (Xiao et al., 2018).
5.2 Middle-lower crustal structure

Our resistivity model shows some high-conductivity anomalies at middle–lower crustal depths. Below the conductive sediments, the crust of both the Yinchuan and Jilantai grabens is featured by elevated electrical conductivity. Each of these two conductors is required by both the impedance and tipper data (Figure 8). In agreement with their surface expressions, the conductor imaged beneath the Jilantai Graben is broader than that beneath the Yinchuan Graben. Between the two grabens, the Helan Shan is imaged as a sub-vertical high-resistivity block that extends from the near-surface into the uppermost mantle. This feature agrees well with seismic tomographic images showing a crustal high-velocity anomaly beneath the Helan Shan (Cheng et al., 2016; Wang et al., 2017a). A similar but less prominent resistor is also imaged beneath the Bayanwula Shan. As shown in Figure 10a, both mountain ranges are also characterized by relatively thicker crust than adjacent areas (Wang et al., 2017b). These features indicate the presence of intact, thick crustal roots below the two mountains, which likely isostatically support their high surface topography.

To the west, the model shows two less distinct conductors beneath the Alxa Block, roughly coincident with the locations of normal faults on the northwestern side of the Bayanwula and Bayan Nuru Shan respectively. The two conductors appear to dip northwestward, and are laterally bounded by high-resistivity regions associated with the Late Permian granitic plutons. The conductor located northwest of the Bayan Nuru Shan is primarily required by the impedance data, whereas the conductor situated northwest of the Bayanwula Shan is required by both the impedance and tipper data (Figure 8).

To the east, the Ordos crust below the conductive sediments is generally of high resistivity and exhibits significant lateral electrical heterogeneity. The imaged resistivities beneath the southeastern end of the profile are substantially higher than those adjacent to the Yinchuan Graben, which might reflect lateral variations in the crust’s compositional structure and thermal state during Cenozoic rifting. Also, note that this change in resistivity is possibly in part due to the screening effect of the conductive sedimentary layer (Feucht et al., 2017), given that the sedimentary layer within the Ordos Block is thicker in the northwest. Between sites 05-07, a local conductor is robustly resolved in the lower crust, with its top located at a depth of ~20 km (Figure 10e). The presence of this conductor is most evident in the tipper data, as manifested by the behavior of the induction vectors at nearby sites (Figure 5). Its location at the western edge of the Ordos Block
coincides with the north-south trending terrane boundary between the western thrust-fault belt and the Tianhuan Depression (Zhang et al., 2015), which is marked by a subtle change in topographic relief and the exposition of elongated Mesozoic strata at the surface (Figure 10d). This feature likely reflects a fractured, weakened fault zone in the crust that has undergone multiple reactivation episodes during the Phanerozoic.

High electrical conductivities in the middle-lower portion of the crust have been widely reported in active extensional areas, and most plausibly explained by free saline fluids, partial melt or a combination of both (e.g., Wannamaker et al., 2008; Rippe et al., 2013; Meqbel et al., 2014; Feucht et al., 2017; 2019). Given that saline fluids have extremely high conductivity values up to ~100 S/m (Sakuma & Ichiki, 2016; Sinmyo & Keppler, 2017), or even higher (~400 S/m, Guo & Keppler, 2019), a small fraction of hypersaline fluids would be sufficient to produce the high conductivities observed in the deep crust (Dong et al., 2020). In active transtensional regimes like the YJRS, the presence of free saline fluids could occur because of recent and/or ongoing supply of deeply sourced fluids to the base of the brittle–ductile transition, which serves as an impermeable cap that prevents fluids from migrating into shallower crustal levels (Hyndman & Shearer, 1989; Ritter et al., 2003). These fluids could be derived from cooling melts, or from high-temperature metamorphic reactions of previously hydrated minerals (Frost & Bucher, 1994; Feucht et al., 2017). In our case, syn-rift normal faults and pre-existing fractured zones within and adjacent to the rifts might act as permeable pathways for deep fluids to migrate from the deeply-situated source areas into higher levels of the crust. Moreover, continued shearing along the faults could have increased the horizontal connectivity of the fluids and further enhance the conductivity beneath the YJRS (Marquis & Hyndman, 1992; Rippe et al., 2013). Direct evidence for the presence of rift-related high-salinity fluids near the study area comes from the fluid inclusions trapped in the Late Cenozoic hydrothermal uranium minerals from the Hangjinqi deposit, which is located ~200 km northeast of the Yinchuan Graben (Zhang et al., 2017).

Although our MT data cannot differentiate between hydrous fluids and partial melt, partial melting in the lower crust is also likely to occur beneath the YJRS. First, several lines of evidence indicate that temperatures in the lower crust of the YJRS are high enough to permit water-present melting, including (1) The western NCC has an average heat flow value of 62 ± 9.5 mW/m², which is much higher than those measured in typical Archean cratonic regions (≤50 mW/m²) (Jiang et al., 2019); (2) The relatively shallow Curie point depths (<30 km) estimated from the satellite
magnetic data (Figure 10b) (Li et al., 2017) gives a clear indication of elevated geothermal gradients between the Ordos and Alxa blocks; and (3) Modelling the measured heat flow data in this area produces temperatures of 600–800 °C at depths between 20 and 40 km (Figure 10c) (Sun et al., 2013), which are generally above the water-undersaturated solidus temperature of crustal rocks (Thompson & Connolly, 1995). Moreover, recent receiver function analysis based on dense seismic arrays shows prominent low-velocity zones and high Poisson's ratios in the middle to lower crust beneath the YJRS, consistent with the presence of partial melting in this region (Wang et al., 2017b). Additional seismic evidence for partial melt below the YJRS has also been provided by seismic amplitude tomography showing strong crustal attenuation around the Ordos Block (Hearn et al., 2008). Such melts could be derived either from basaltic injections or lower-crustal melting triggered by the introduction of heat and volatiles from below (Feucht et al., 2017). However, given the absence of Cenozoic magmatism within the YJRS, large volumes of crustal melt are unlikely to occur since they would be extracted from the source area and unavoidably tend to rise towards the surface (Brown, 2007). Given these arguments, we attribute the enhanced conductivities in the middle-lower crust to a combination effect of partial melt and saline fluids, with partial melt mainly concentrating in the lower crust. Such a two-phase mode has also been invoked to explain crustal conductors observed in other continental rift zones (Li et al., 2003; Wannamaker et al., 2008; Feucht et al., 2017).

5.3 Uppermost mantle structure

Except for the two resistive regions at either end of the profile, our inversion models consistently show zones of elevated conductivities (5~30 Ω·m) at uppermost mantle depths beneath the YJRS (Figure 10). Due to the limited period range of BMT data and the presence shallower conductors in the crust, the actual resistivity values and the lower boundaries of these conductors may not be well constrained. However, depth sensitivity tests have demonstrated that high conductivities are required at such depths to explain the measured data and should extend down to depths over ~80 km (Figure 9). Significantly, these anomalies appear to be interconnected to form a massive conductive feature that is mainly located on the northern side of our profile, as manifested by the northward-pointing induction vectors at periods longer than 1000 s (Figure 5). Synthetic modeling tests (Figure S3) indicate that our profile can partly resolve this off-profile feature, even if its actual conductance value is only moderately high (5000 S). This feature likely represents the southernmost part of the upper mantle conductor beneath the northern Ordos Block.
and Hetao Graben, which previous MT surveys observed in the western NCC (Dong et al., 2017; Ye et al., 2020).

The primary factors that have been invoked to explain the enhanced conductivity in the subcontinental lithospheric mantle include graphite films on grain boundaries, water in nominally anhydrous minerals, and partial melt (Chave & Jones, 2012). Although the temperature has also been proven to have a substantial effect on mantle resistivity (Ledo & Jones, 2005), the resistivity values observed in our model certainly require additional conductivity mechanisms since dry olivine behaves quite resistively (> 1000 Ω·m) at realistic temperature conditions (Constable, 2006). Given the estimated temperatures (>700 °C) at the uppermost mantle depths beneath the YJRS (Figure 10c) (Sun et al., 2013), graphite film can also be excluded since they are suggested to be unstable at such high temperatures according to recent laboratory results (Zhang & Yoshino, 2017). The remaining candidates are water and partial melt, both of which have been widely identified beneath active extensional areas (e.g., Wannamaker et al., 2008; Feucht et al., 2017, 2019).

Based on published experimental results, Naif (2018) estimated the upper-bounds on the electrical conductivities of the hydrated mantle and argued that hydrous olivine is inadequate to explain resistivities less than ~10 Ω·m. If true, mantle resistivities imaged beneath the YJRS are sufficiently low to require some level of partial melt in the uppermost mantle. However, since the mantle temperatures determined from surface heat flow (Figure 10c) are substantially lower than the nominally dry peridotite solidus (Hirschmann, 2000), some mechanisms are needed to reduce the peridotite solidus below the estimated mantle temperatures effectively. Experimental studies have demonstrated the large effect of CO\textsubscript{2} and H\textsubscript{2}O on partial melting properties in the Earth’s upper mantle (Dasgupta et al., 2013; Falloon & Green, 1989). These studies emphasize that the solidus temperature of hydrous carbonated peridotite reduces dramatically to < 1000 °C at a depth of ~60 km (Green, 2015). Given the estimated mantle temperatures (900~1000 °C at 60 km depth) beneath the YJRS (Figure 10c), these experimental results are consistent with the presence of small amounts of CO\textsubscript{2}- and H\textsubscript{2}O- rich melts at uppermost mantle depths. Because the addition of both CO\textsubscript{2} and H\textsubscript{2}O can significantly increase the melt conductivity (Sifré et al., 2014), small amounts of CO\textsubscript{2}- and H\textsubscript{2}O-bearing melts would be sufficient to explain the observed mantle resistivities. Given the tectonic context, the source region for these volatile-rich melts could be associated with
a metasomatized lithospheric mantle enriched by volatile-bearing fluids and melts released from the southward subducted Paleo-Asian oceanic slab during the Paleozoic (Dai et al., 2019). Owing to the enhanced transport properties of carbonated melts, such melts would rise to shallower levels and liberate gaseous CO$_2$ through decarbonation reactions. In extensional regimes like the YJRS, the interconnected network of normal faults, and extensional fractures and veins are expected to provide permeable pathways for these mantle-sourced volatile-rich fluids (Tamburello et al., 2018). This well explains the spatial correlation between elevated CO$_2$ emission and crust-penetrating normal faults within the Yinchuan graben (Cui et al., 2019).

A variety of independent geophysical observations support the existence of a melt-modified lithosphere beneath the YJRS. Early teleseismic S-receiver functions have shown that the lithosphere beneath the circum-Ordos rifts is considerably thinner (~80 km thick) compared to the thick cratonic lithosphere (>200 km) beneath the Ordos Block (Chen et al., 2009; Tian et al., 2009). Within the YJRS, the existence of a notably thinned lithosphere is further supported by recent deep seismic reflection profiling, which reveals a dome-shaped upper mantle reflector (UMR) at a depth of 82–92 km beneath the Yinchuan Graben (Figure 10e) (Liu et al., 2017). Rayleigh wave tomography based on dense seismic networks identifies a prominent low-velocity anomaly extending from the uppermost mantle to a depth of ~200 km beneath the Yinchuan and Hetao grabens, consistent with upwelling and melting of the mantle in this region (Li et al., 2017). Likewise, Cheng et al. (2016) found high Poisson's ratios in the uppermost mantle beneath the conjunction area between the Ordos and Alxa blocks, spatially coincident with low-velocity anomalies and likely requires some level of partial melt. Additionally, joint inversion of seismic and gravity data identified a distinct low-density anomaly in the lithospheric mantle beneath the YJRS, which has been interpreted to reflect partial melting induced by thermo-mechanical erosion (Wang et al., 2014).

Beneath the Ordos Block, another remarkable feature in our model is the resistive mantle lithosphere imaged at the southeastern end of the profile (Figure 10e). These resistivities are consistent with a melt-free lithospheric mantle, but still substantially lower than those expected for Archean cold, stable cratons where highly electrically resistive (>10$^3$ Ω·m) are usually observed (Eaton et al., 2009). The reduced resistivity values could be explained by the addition of a water content of < 200 p.p.m at 1000 °C (Jones et al., 2012). This contrasts with the low resistivities (<30 Ω·m) imaged beneath the westernmost Ordos Block adjacent to the Yinchuan Graben, which
are consistent with a hydrated/melted composition as previously suggested. This southeastward increase of lithospheric resistivity likely reflects an eastward decrease in the degree of rift-related lithospheric modification as the distance from the YJRS increases. To the northwestern end of the profile, the upper mantle beneath the northern Alxa Block also appears to be generally resistive. We omit discussion of this feature, given that its actual resistivity is not well resolved owing to the masking effect of the lower crustal conductor.

5.4 Implication for Late Cenozoic rifting processes

From a petrological point of view, free saline fluids are gravitationally unstable in the deep crust and would be rapidly consumed by retrograde metamorphic reactions of anhydrous high-grade rocks (Frost & Bucher, 1994; Yardley & Valley, 1997), or dissolved in the partial melt under high temperature conditions (Holtz et al., 2001; Le Pape et al., 2015). Although salt-rich solutions (e.g., NaCl, KCl) are expected to lower H$_2$O activity and prevent the lower crustal rocks from melting (Aranovich & Newton, 1997), recent studies suggest that they could also enhance the rates of retrograde metamorphism and thus further reduce the residence time of deep fluids (Yardley et al., 2014). Similarly, without a continued supply of heat and/or mantle-derived basalts from below, melt in the deep crust would cool and solidify quickly and eventually becomes electrically resistive (Feucht et al., 2017). Numerical modeling suggests that the thermal effects of repetitive basalt intrusions on the crust would be fully decayed within several millions of years after the cessation of basaltic injections (Annen & Sparks, 2002). Due to the relatively short residence times of fluids in the deep crust, the sources of deep fluids in modern rift zones have been often attributed to recent and/or ongoing tectonic processes that have supplied heat, volatiles, and/or basaltic melts into the crust (e.g., Rippe et al., 2013; Feucht et al., 2017; 2019). Within the YJRS, these fluids are most likely associated with the Late Cenozoic rifting activity as indicated by petrological and geochemical data (Zhang et al., 2017). This argument is also compatible with modern geodetic observations showing that the YJRS is tectonically active nowadays, with both extensional and shearing deformation distributed regionally across the whole rift system (Figures 11a and 11b) (Wang & Shen, 2020).

Combined geochronological, structural, and fault kinematic investigations show that the YJRS has been dominated by a transtensional stress regime since the Late Miocene, as a result of the northeastward expansion of the Tibetan Plateau (Shi et al., 2020). During this period, continued
NW-SE-directed extension and subsequent thinning of the lithosphere are expected to induce decompression partial melting of the passively upwelling mantle beneath the rifts (Olsen, 1995). However, it should be noted that the low magnitude of extension across the YJRS seems difficult to reconcile with the degree and lateral extent of partial melting observed in the present-day lithospheric mantle. Modern geodetic data show that the entire YJRS is actively extending at a rate of ~2 mm/a (Figure 11a), which is about twice the geologically determined Late Quaternary extension rate across the Yinchuan Graben (Middleton et al., 2016b). Numerical models for passive continental rifts predict that, syn-rift cooling would predominate when the lithosphere is extending at such low rates and the volume of extension-induced melt would be suppressed due to conductive and convective heat diffusion of the lithosphere during the extension (McKenzie & Bickle, 1988). In this case, the stress-induced extension of the lithosphere alone is unlikely to produce a broadly distributed melting area as we observed below the YJRS, and some form of active mantle convection might be required. By themselves, our MT data cannot identify the nature and style of convection in the mantle. Seismic receiver functions along a profile across the Hetao Graben (Tian et al., 2011) show relatively flat 410- and 660-km discontinuities, indicating convection in the nearby region is unlikely driven by a deep-sourced mantle plume. The absence of a mantle plume beneath the YJRS also agrees with seismic tomographic results showing that the upper-mantle low-velocity anomaly terminates at a depth of ~200 km (Li et al., 2017). Given the regional tectonic context, the most likely candidate is the collision-induced asthenospheric mantle flow from underneath the high Tibetan Plateau, which has long been regarded as a possible mechanism to explain the widespread Cenozoic rifts in eastern China (Liu et al., 2004). This view is supported by recent seismic observations suggesting mantle flow from below Tibet has reached the gap between the Ordos and Alxa blocks, and contributed to rift formation and the occurrence of large earthquakes (Shen et al., 2017).

Comparing our resistivity model with published MT results gives us a better understanding of the Cenozoic rift evolution around the Ordos Block. At the eastern margin of the Ordos Block, Yin et al. (2017) found a prominent conductive feature that extends from the upper crust into the upper mantle depths beneath the southernmost segment of the SXRS (Figure 11d). A similar lithosphere-scale conductor beneath the northern segment of the SXRS was also identified by the MT survey of Yin et al. (2016). Similar to the conductive features we observed beneath the YJRS, these features were interpreted as asthenosphere-derived fluids upwelling through pre-existing
lithospheric weak zones that have been reactivated during the Cenozoic extension. However, in contrast to the broadly distributed (~400 km wide) high conductivities imaged beneath the YJRS, both of the two conductive, fluid-bearing zones appear to be confined within a relatively narrow region (<100 km wide). If the widths of these features reflect the lateral extents of partial melt generated during rifting, it would suggest that the lithosphere of the YJRS has undergone a more significant extent of melting and modification compared to that of the SXRS. Note that the surface expressions of these two rifts also exhibit a similar pattern in terms of their widths, and modern geodetic observations show higher levels of extensional and shearing rates over a broader region across the YJRS (Figures 11a and 11b; Wang & Shen, 2020). All these observations suggest that, compared to the narrow rifts formed along the eastern side of the Ordos Block, the YJRS at the northwestern side of the Ordos has experienced a more distributed extension.

What caused the different styles of rifting on the two sides of the Ordos Block? Among various factors that exert control on the structural style of continental rifts, the depth-dependent rheology of the lithosphere is believed to play a central role in determining their width (Buck, 1991; Brun, 1999; Gueydan et al., 2008). Both analogue and numerical models predict that distributed continental rifting occurs when the lower crust and lithospheric mantle are mechanically weak, while localized rifting requires a high strength lithospheric mantle (Brun, 1999; Gueydan et al., 2008). If this is the case, the initial lithospheric strength of the YJRS should be considerably lower than that of the SXRS. Geological data suggest that, unlike the SXRS which has been formed primarily during the Late Cenozoic, the YJRS underwent significant extension in the Early–Middle Cenozoic (Zhang et al., 1998; Shi et al., 2020). Although the stretched lithosphere will become strong again as the rift-induced thermal anomaly decays, it can be permanently weakened by deep-reaching normal faults and be more prone to transtensional reactivation during subsequent tectonic periods (Ziegler & Cloetingh, 2004). Therefore, the pre-existing lithospheric weakness of the YJRS can be inherited from the early stages of Cenozoic extension, possibly associated with the northwestward subduction of the Pacific plate (Shi et al., 2020). During the Late Cenozoic, such an inherited weakness makes the YJRS being a locus of transtensional deformation in response to the northeastward extrusion of the Tibetan lithosphere. Moreover, in the presence of such a pre-existing weak zone, the collision-induced mantle flow from beneath the Tibetan Plateau would be preferentially channeled into the YJRS (Shen et al., 2017), leading to further weakening and thinning of the extended lithosphere. In comparison,
although the collision-induced far-field stresses and mantle flow could also reach the SXRS (Yu & Chen, 2016), their magnitudes and potential effects on the less extended lithosphere should be far less profound due to the increased distance from the high plateau. Therefore, we speculate that differences in both the inherited lithospheric rheology and the magnitude of the collision-induced far-field effects may both contribute to this abrupt change in rift style.

Our MT image, combined with additional geological and geophysical evidence, allows us to build a coherent view of the Late Cenozoic evolution of the YJRS (Figure 12). In the Late Miocene, the far-field effect of propagating extrusion induced by the India-Eurasia collision eventually reached the northeastern Tibetan margin, as evidenced by the rapid exhumation of the Liupan Shan circa ~8 Ma (Zheng et al., 2006). In response to the NE-directed compressional stresses from the south, the Alxa Block moves northwestward relative to the Ordos Block (Wang & Shen, 2020; Hao et al., 2021), leading to broadly distributed right-lateral shearing and crustal extension in the conjunction zone between the two blocks. This transtensional stress regime is responsible for the reactivation of a set of pre-existing faults and basement fabrics and the continued subsidence of the rift basins that initially formed during the Early Cenozoic extension (Shi et al., 2020). This process has also led to the rapid uplift of the Helan Shan and other adjacent host-type mountain ranges that are bounded by high-angle normal faults (Liu et al., 2010). Below the surface, continued lithospheric extension has allowed for thinning and melting of the volatile-rich lithospheric mantle, generating basaltic melts that buoyantly rise and concentrate at the base of the crust. In addition, weak and hot asthenospheric material could be extruded from beneath the Tibetan Plateau into the pre-existing weak zone beneath the YJRS and contribute to the extensive lithospheric modification. The heat and volatiles transferred from basaltic melts could induce partial melting of the lower crust. As the melts cool, saline magmatic fluids are exsolving from the melts and migrating upward into the shallow crust through the reactivated fault zones. Although similar processes might also occur along the eastern border of the Ordos Block (Yin et al., 2016; 2017), their magnitude and spatial extent are presumably less significant than those observed along its western margin, owing to the lack of inherited lithospheric weakness and the increased distance from the Tibetan Plateau.
6 Conclusions

This study presented an electrical resistivity model of the crust and uppermost mantle through 3-D inversion of broadband MT profile data across the Yinchuan-Jilantai Rift System (YJRS) along the northwestern margin of the Ordos Block. We identified several prominent anomalies in the model and examined the nature and origin of the respective anomalies. We further compared our resistivity model with previously reported geo-electrical models and integrated these results with surface geologic and geodetic observations to better understand the Late Cenozoic rifting processes in western North China. The main conclusions are as follows:

1. The shallow resistivity structure along the profile correlates well with known structural and lithologic features. Young sedimentary basins are generally characterized by low resistivities, while areas with exposed bedrock exhibit high resistivities.

2. Crustal-scale, subvertical high-resistivity blocks were imaged beneath the Helan Shan and Bayanwula Shan, implying these horst-type mountain ranges are supported by thick crustal roots through Airy compensation.

3. In the middle-lower crust, several discrete, subvertical conductive zones correlate with the active rift-parallel normal fault zones. We attributed these features to a combination of saline fluids and partial melt, both of which are likely associated with the recent and ongoing supply of heat, volatiles and melts from the mantle.

4. The laterally extensive zone of elevated conductivity observed in the uppermost mantle requires the presence of partial melt and was interpreted as an indication of decompression melting and thinning of the volatilized lithospheric mantle during extension.

5. In comparison to the lithosphere of the Shanxi rift system along the eastern margin of the Ordos Block, the lithosphere of the YJRS has undergone a larger extent of melting and modification, owing to its inherited lithospheric weakness and closer proximity to the India-Asia collision zone.

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Figure captions

**Figure 1.** (a) Tectonic setting of the Ordos Block and surrounding areas. Yellow squares show the locations of the collected MT soundings. The yellow star shows the location of the remote reference site deployed during the whole field campaign. Black solid lines denote the major active faults modified from Deng et al. (2003) and Zhang et al. (2021). Red circles mark the earthquakes from the China Earthquake Data Center (http://data.earthquake.cn). Aqua arrows show the GPS velocity field relative to the Ordos Block (Wang & Shen, 2020). Abbreviations: YJRS: Yinchuan-Jilantai rift system; SXRS: Shanxi rift system; HTG: Hetao Graben; WHG: Weihe Graben. (b) Simplified tectonic background of the eastern Eurasian continent. The black rectangle shows the location of Figure 1a.

**Figure 2.** Simplified geologic map of the western North China Craton modified from Xiao et al. (2018) and Zhang et al. (2020). Abbreviations: YCG: Yinchuan Graben; JLTG: Jilantai Graben; HTG: Hetao Graben; HLS: Helan Shan; BWS: Bayanwula Shan; BNS: Bayan Nuru Shan; SLZS: Shalazha Shan; LPS: Liupan Shan; YRF: Yellow River Fault; EHF: East Helanshan Fault; WHF: West Helanshan Fault.

**Figure 3.** Observed (circles & squares) and predicted (solid lines) MT data for three selected sites (02, 24, and 40) from different geologic units. These data include apparent resistivity (App. Res.) and phase responses obtained from the full impedance tensor (Zxx, Zxy, Zyx and Zyy) and the real and imaginary parts of the tipper vector (Tx and Ty). Red squares in Figure 2 mark the locations of these sites.

**Figure 4.** Pseudo-sections of normalized phase tensor ellipses for all periods and sites along the profile. The top panel shows the topographic variation along the MT transect. Ellipses in Figures 4a are colored by skew angle, which reflects the subsurface dimensionality. High skew values (≥ 4°) are indicators of 3-D structures. Ellipses in Figure 4b are colored by the geometric mean of the maximum and minimum phase tensors representing the vertical conductivity gradient; high values of Φ2 (≥ 45°) indicate increasing conductivity with depth.
Figure 5. Real induction vectors in Parkinson convention (Parkinson, 1962) at three different periods (50, 316, 1585 s). Horizontal resistivity slices of the preferred inverse model were also shown at three depths approximately corresponding to the penetration depths of the induction vectors. Note these vectors point away from resistive blocks and toward zones of higher conductivity. The black dashed line in (c) denotes the location of vertical resistivity sections shown in the following figures.

Figure 6. Comparison of vertical resistivity sections through the center of the inversion models with different initial resistivities (25, 50, 100 and 200 Ω⋅m). The location of the sections is shown as a black dashed line in Figure 4c. In addition, relevant information about each inversion run (Initial resistivity, iteration number, initial and final RMS values) was labelled in the corresponding vertical section.

Figure 7. Comparison of vertical resistivity sections through the center of the inversion models obtained with different covariance values (0.1, 0.2, 0.3, 0.4 and 0.5). The location of the sections is shown as a black dashed line in Figure 4c.

Figure 8. Comparison of vertical resistivity sections through the center of the models obtained from inversion of different data components (Impedance (Z) only, Tipper (T) only and joint inversion of Z and T).

Figure 9. Resolution tests for the imaged resistivity features at mantle depths. (a) Comparison of the site-by-site RMS deviations computed as differences between the RMS values of the original unconstrained model and those of the three constrained models. (b-e) Vertical resistivity sections through the center of the original unconstrained model and the three constrained models. The three constrained models were obtained by inversion runs with model resistivity fixed to a constant resistivity value of 200 Ω⋅m below 42, 61, and 78 km.

Figure 10. (a) Seismically defined Moho depth profile along the survey line (Wang et al., 2017b); (b) Curie depth profile from the Global Curie Depth Model (GCDM) (Li et al., 2017); (c) Temperature profiles at different depths (20, 40 and 60 km) extracted from the thermal model of
Sun et al. (2013); (d) Geological provinces with the same color scheme as in Figure 2. The black dashed line denotes a buried fault inferred from this study. The red solid line shows the location of the deep seismic reflection profile (Liu et al., 2017); (e) Vertically exaggerated resistivity section through the center of the preferred inversion model (V:H=1.5: vertical axis has been exaggerated by a factor of 1.5, compared to the horizontal axis). Black solid lines show the geometry of the fault system within the Yinchuan Graben (YCG) according to the interpretation of seismic reflection data (Liu et al., 2017). Black dashed lines mark our inferred structural geometries. The red dashed line marks a strong upper mantle reflector (UMR) revealed by deep seismic reflection data (Liu et al., 2017).

**Figure 11.** (a) GPS velocity profiles across the Ordos Block. Red and blue squares show the components of velocity parallel (N45°E) and normal (S45°E) to the bounding faults of the Jilantai Graben (JLTG), respectively. (b) Map showing the GPS velocity field relative to the Ordos Block. The GPS velocities in both (a) and (b) are derived from Wang & Shen (2020), in which the original GPS velocity field is relative to a stable Eurasian reference frame. The locations of the GPS stations used to transform the reference frame are shown in Figure S4. (c) and (d) show resistivity sections along the profiles A-B and C-D (green dashed lines) shown in (b). The resistivity section across the Shanxi rift system (SXRS) is extracted from a 3-D resistivity model with an initial background resistivity of 100 Ω·m (Yin et al., 2017).

**Figure 12.** Schematic interpretation of resistivity features related to the recent and present-day rifting processes on the western and eastern sides of the Ordos Block. Yellow and red arrows at the surface indicate the extensional and shearing directions inferred from geologic and geodetic observations. Note that both the magnitude and lateral extent of surface extension and shearing rates are much more prominent on the western side of the Ordos Block. Blue arrows and ellipses indicate upward migration of deep-sourced, volatile-rich fluids into shallower crustal levels through the crustal-penetrating normal faults (solid black lines) that were reactivated during the extension. Red ellipses indicate basaltic underplating and injection into the lower crust and low degree of lower-crustal melting induced by the addition of heat and volatiles from below. Red dots indicate volatile-assisted decompression melting of the upwelling mantle.
Induction arrow magnitude

Period: 50 s
Depth: 2.2 km

Period: 316 s
Depth: 21.1 km

Period: 1586 s
Depth: 39.4 km

Alxa Block

Ordos Block

HLS
BWS
BNS
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Lateral asthenospheric flow from beneath the Tibetan Plateau

Eastward decrease in the magnitude of collision-induced far-field effects

Mantle upwelling

Thick cratonic lithospheric mantle

Locally modified lithospheric mantle

Broadly modified lithospheric mantle

Crust

Lithospheric mantle

Asthenosphere

Thick crustal root

Yinchuan-Jilantai rift system

Ordos Block

Shanxi rift system

Alxa Block

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LFG

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