**Abstract** An island arc environment related to intraoceanic subduction or ridge subduction from the Late Carboniferous to the Early Permian has been proposed for the formation of Western Junggar, NW China, situated in the southwest of Central Asian Orogenic Belt. However, the details and consequences of the subduction remain controversial. To further identify the relics of the Late Carboniferous subduction at depth, three magnetotelluric profiles were deployed. Phase tensors, real induction arrows, and three-dimensional (3-D) isotropic resistivity models indicate the presence of a 3-D anisotropic resistivity structure in the crust. Consequently, a 3-D anisotropic resistivity model was constructed by forward modeling. The 3-D anisotropic resistivity model contains two azimuthally anisotropic middle-upper crust anomalies with minimum resistivity striking N90°E at depths of 5 to 20 km, and an azimuthally anisotropic lower crust anomaly with minimum resistivity striking N20°E at depths of 20 to 46 km. The electrically anisotropic anomalies are closely related to relatively high shear-wave velocities with significant negative radial anisotropy. The anisotropies in the middle-to-upper crust and the lower crust are related to the oceanic ridge (N90°E) subduction with a slab window in the Late Carboniferous and the remnant oceanic slab, respectively. The magnetotelluric observations support a NW70° subduction with an intraoceanic ridge-transform system in the Late Carboniferous and a remnant oceanic slab trapped in Western Junggar.

**1. Introduction**

The Central Asian Orogenic Belt (CAOB) is surrounded by the Siberian, European, Tarim, and North China Cratons (Figure 1a). It mainly formed by the progressive subduction of the Paleo-Asian Ocean and the amalgamation of terranes of diverse origins (Windley et al., 2007). The Western Junggar, which is situated in the southwest of CAOB, has vast occurrences of granitoids and volcanoes, especially ophiolitic rocks. Therefore, the Western Junggar is assumed as an ideal natural laboratory to investigate the tectonic evolution and crustal growth of the CAOB.

With regard to Western Junggar, the tectonic environments of the Late Carboniferous to the Early Permian (320–290 Ma) are subject of controversy. Two major competing models have been proposed: a postcollisional model (Chen & Arakawa, 2005) and an island arc environment related to intraoceanic subduction or ridge subduction (Geng et al., 2009; Tang et al., 2010; Tang, Wang, et al., 2012; Tang, Wyman, et al., 2012; Xiao et al., 2008; Yang et al., 2012; Yang et al., 2017; Yin et al., 2010; Zhang, Xiao, Han, Ao, et al., 2011, Zhang, Xiao, Han, Mao, et al., 2011). In recent years, the subduction model has gained increasing acceptance. So far, three mechanisms for the Late Carboniferous subduction have been proposed: (a) the Junggar Ocean northwestward subducted beneath the Karamay arc generates a back-arc basin (Geng et al., 2009; Tang et al., 2010; Yang et al., 2017; Yin et al., 2010), (b) these are double subduction systems that have a ridge-trench interaction in the Darbut area (Ma et al., 2012; Zhang, Xiao, Han, Ao, et al., 2011, Zhang, Xiao, Han, Mao, et al., 2011), or (c) two northward subduction zones are present in the eastern part of West Junggar (Yang et al., 2012).

Magnetotelluric (MT) sounding is the most effective method to investigate electrical structures in the crust and upper mantle. Electrical resistivity models that are derived from MT data provide important constraints on the thermal, mechanical, and chemical modification of the lithosphere (e.g., Jones et al., 2013). Consequently, they yield valuable information to improve geodynamic modeling. Previous three-dimensional (3-D) imaging of a MT transect with a length of 182 km indicated a fossil Paleoziel...
intraoceanic subduction system across the Darbut belt (Xu et al., 2016). This has been confirmed by other MT studies in adjacent areas (Yang et al., 2016; Zhang et al., 2017). However, the specific subduction mode and direction have not been clearly reported. Moreover, previous MT studies focused on isotropic results, although the existence of anisotropy had been suggested (Xu et al., 2016; Zhang et al., 2017).

**Figure 1.** Research region and site distribution. (a) Overview of the Central Asian Orogenic Belt (modified after Jahn et al., 2000), (b) the Western Junggar (modified after Yang et al., 2013), and (c) site setup. The yellow rectangle in (c) represents the location of regional mafic and ultramafic intrusions (Ma et al., 2012; Yin et al., 2010).
With the continuous development of geophysical observations, modeling, and high-temperature-pressure experimental techniques in recent years, evidence has been found for a multiscale anisotropic Earth. Electrical anisotropy can reflect the preferred orientation of minerals or specific geological bodies, or the imprint of rheological deformation. This provides essential data and key constraints for lithospheric fabrics and geodynamic models (Jones, 2012). So far, many MT studies have addressed the electrical anisotropy and a comprehensive overview has been given by Martí (2014).

A variety of different approaches has been used to evaluate the existence of electrical anisotropy within the subsurface, that is, phases exceeding 90° (Heise & Pous, 2003; Weckmann et al., 2003), the coincidence of a large split between phase tensor invariants and small induction arrows (Häuserer & Junge, 2011), and the coincidence of phase differences and induction arrows (Yin et al., 2014). In a MT study by Wannamaker et al. (2008), two-dimensional (2-D) isotropic inversion imaged two narrow steep conductors that are separated by steep resistive zones. These were identified as artifacts due to the presence of anisotropy. Further evidence for anisotropic features of the Earth’s crust and upper mantle were provided by combining MT results with seismic parameters, that is, obtained via shear wave splitting analyses (Frederiksen et al., 2006; Hamilton et al., 2006; Padilha et al., 2006). For quantitative interpretation, one-dimensional (1-D) inversion (Pek & Santos, 2006) and 2-D modeling (Pek & Verner, 1997) of anisotropic media were applied to real data. In recent years, significant progress was achieved in both 2-D inversion (Baba et al., 2006; Key, 2016) and 3-D modeling (Löwer & Junge, 2017; Wang & Fang, 2001; Weidelt, 1999; Xiao et al., 2018). However, no complete 2-D and 3-D electrical anisotropic inversion schemes have been developed so far.

This paper presents the results of MT investigations performed in the Darbut belt of Western Junggar. Based on the patterns of phase tensors, real induction arrows, and the results of 3-D isotropic inversion, the existence of an anisotropic conductivity structure within the crust was postulated. A 3-D electrical anisotropy model was developed by forward modeling that explains the observations. Furthermore, its significance for the architecture and geodynamic origin of the Darbut belt is discussed.

2. Geological Setting

The Western Junggar is surrounded by the Chinese Altai to the north, the Tianshan Orogen to the south, and the Junggar Basin to the east (Figure 1b). It formed by accretion, subduction, and collision between the Paleo-Asian ocean, surrounding island arcs, and microcontinental blocks beginning at the early Paleozoic (Windley et al., 2007). The Zaire mountain occupies the eastern part of Western Junggar. It is situated on the western border of the Junggar Basin and is generally proposed as a Carboniferous accretionary complex, including turbidites, tuff, greywacke, and ophiolitic series (Zhang et al., 2018). The Darbut fault (striking NE-SW with left-lateral strike slip) entirely crosscuts the Zaire mountain through its center (Figure 1c).

Today, it is considered as the key to unlocking the complex evolution of the CAOB.

There are two well-exposed NE to SW striking ophiolitic mélanges in Western Junggar: the Darbut ophiolitic mélange and the Karamay ophiolitic mélange (Zhang et al., 2018). The Darbut ophiolitic mélange occurs along the Darbut fault with an age from the middle Devonian to the early Carboniferous (Bai et al., 1995; Gu et al., 2009; He et al., 2013; Tian et al., 2015; Xu et al., 2006; Yang et al., 2012) in the Carboniferous strata. However, the dated or geologically constrained ages of the Karamay ophiolitic mélange along the Karamay-Urho fault are severely scattered, and reach from the middle-to-late Ordovician to the early Carboniferous (Geng et al., 2009; He et al., 2007; Xu et al., 2006; Yang et al., 2013; Zhang, Xiao, Han, Mao, et al., 2011). Most studies suggested that the basement of the Junggar Basin is an assembly of arcs, accretion complexes, and trapped oceanic crust (e.g., Wang et al., 2002; Xiao et al., 2008; Zhang et al., 2018).

3. Data Acquisition, Processing, and Analysis

3.1. Data Acquisition and Processing

During summer 2013, 126 broadband MT sites of five components ($B_x$, $B_y$, $B_z$, $E_x$, and $E_y$) were collected, with an average site spacing of 2 km along three profiles, 20 km apart. Each profile has an approximate length of 100 km and is orientated in NW-SE direction, that is, orthogonal to the Darbut belt. Six GMS07e systems were deployed and run for at least 20 hr at each site. The observed time series were processed by the single-station and remote reference robust method using MAPROS software (Friedrichs, 2003), yielding...
impedance tensors $Z$ and tippers $Tz$ within the period band of 0.0025–2,000 s. The data of Profile 2 has been used by Xu et al. (2016) as part of the profile with 182-km length. The research region and site distribution are shown in Figure 1.

### 3.2. Phase Tensor

The basics of the phase tensor approach are summarized in the comprehensive review by Booker (2014). Within the given coordinate system, the direction of the linearly polarized magnetic field excitation, for which the phase tensor takes the values of $\Phi_{\text{max}}$ and $\Phi_{\text{min}}$, is well defined (Caldwell et al., 2004). Furthermore, the arctangents of $\Phi_{\text{max}}$ and $\Phi_{\text{min}}$ are defined as the maximum and minimum principal phases $\phi_{\text{max}}$ and $\phi_{\text{min}}$, respectively. In a 2-D case, $\phi_{\text{min}}$ and $\phi_{\text{max}}$ refer to the MT phases of transverse electric (TE) and transverse magnetic (TM) modes, with the electric field being either parallel or orthogonal to the strike direction. Moreover, the skew angle $\beta$, calculated from phase tensor, can be used to measure the structural dimensionality (Caldwell et al., 2004). For small $\beta$ values ($|\beta| < 3$) a 2-D interpretation is valid, while for large $\beta$ values ($|\beta| > 3$) a 3-D structure should be considered (Booker, 2014; Caldwell et al., 2004).

The phase tensor can be displayed by an ellipse with the invariants $\Phi_{\text{max}}$, $\Phi_{\text{min}}$ and the skew angle $\beta$ (Caldwell et al., 2004). A circular phase tensor indicates a small phase split (the difference between $\phi_{\text{max}}$ and $\phi_{\text{min}}$), while a flat phase tensor indicates a larger split. Moreover, the phase split can be caused by the bulk anisotropic property of the material (Heise et al., 2006). In other words, the phase split can be caused by (but not only by) an anisotropic structure and is thus an unreliable sign of the anisotropic structure (Martl, 2014).

Figure 2 displays the phase tensor ellipses along the three profiles for six periods in the bandwidth 1–1,000 s. Large $\beta$ values ($|\beta| > 3$) widely appear at these periods, indicating strong 3-D subsurface structure. For the short period (0.77 s), $\phi_{\text{max}}$ in the mountain area is oriented parallel to the profiles, and toward the south a slight counterclockwise rotation occurs for the longer periods (25.9–132.2 s). In the Junggar basin, the directions of $\phi_{\text{max}}$ rotate synchronously from NNE to NW with increasing periods. It is evident that for longer periods, the directions of $\phi_{\text{max}}$ face NW 75–105° between the Darbut and Karamay-Urho faults. At periods of 674.7 and 1,161.6 s, the spatial distribution of the ellipses does not show a systematic pattern.

### 3.3. Real Induction Arrow

The real induction arrows are displayed in Figure 3, following Wiese convention (Wiese, 1962) for the same periods that were used for the phase tensors in Figure 2. In the Wiese convention, the real induction arrows point away from low-resistivity structures. In general, the larger the real induction arrow is, the larger the lateral resistivity contrast will be at a specific period. For the short period (0.77 s), the arrows show large differences on a small scale due to local lateral contrasts. However, for longer periods up to 132.2 s, the arrows display rather homogeneous patterns for all three profiles. The arrows almost vanish to the north, and point to the west at 25.9 s with decreasing size to the south. In addition, the arrows show a slight counterclockwise rotation for longer periods in the southern area. The maxima occur close to the transition between the mountainous area and the Junggar basin. The direction of the arrows is oblique to the direction of phase tensor ellipses between 25.9 and 132.2 s. This implies the existence of a 3-D conductivity anomaly. Furthermore, the size of the arrows does not change spatially along their orientation. This would be the case with increasing distance from a vertical conductivity boundary in the isotropic case, and therefore, a simple isotropic model cannot explain the observed pattern. Xu et al. (2016) discussed the existence of anisotropy in the crust of the Western Junggar basin following the analysis of real induction arrows deviating by about 30° from the major principal axes of the phase tensor ellipses to the south of Profile 2.

The following section focuses on the spatial distribution of the transfer functions in the period range between 25 and 132 s. For shorter periods, the patterns of transfer functions are either 1-D or 2-D and heterogeneous due to local effects. At longer periods, the real induction arrows are too small and too noisy for further interpretation.

### 4. The 3-D Isotropic Inversion

For the isotropic inversion, the 3-D regularized inversion code ModEM was used (Egbert & Kelbert, 2012; Kelbert et al., 2014) to fit the observed full impedance tensors $Z$ and/or the observed tippers $Tz$ (Tzx and
Tzy), at 26 logarithmically spaced periods between 0.0025 and 2,000 s. More than 20 different inversions were conducted to assess the sensitivity of the results to inversion parameters (e.g., grid resolution and model smoothing length scales), and to subsets of data (e.g., different period bands and separate inversions of Z and Tz). Four top thin layers (1, 2, 4, and 8 m) were added and 0.3 was chosen as model smoothing length, which significantly reduced the root-mean-square (rms) error.

The forward mesh consisted of 108 × 115 × 39 nodes with a lateral 1-km grid spacing within the central area (45.45°–46.30°N, 84.2°–85.55°E). The central modeling area is 94 km (N-S or x) by 101 km (E-W or y), nested within a larger domain with a dimension of 308 km × 316 km. For the vertical direction, nine layers (1, 2, 4, 8, 10, 20, 40, 80, and 100 m) were set at the top, and then, 30 layers with thicknesses of 10^2.2–10^5.2 m (~158–158,000 m) were equally spaced in the log domain. The depth of the lower boundary was 747 km, which has been assured to be sufficiently deep to avoid the boundary effects on inversion. Several poor-quality MT responses were removed from the array of 102 sites, and error floors were set to 5% of |Z_x|x Z_y|y|^{1/2} for all impedances and to 0.03 for tippers. An effective apparent resistivity for the entire array can be computed by first

Figure 2. Phase tensor ellipses for (a) 0.77 s, (b) 25.9 s, (c) 58.5 s, (d) 132.2 s, (e) 674.7 s, and (f) 1,161.6 s. The phase tensor ellipses were obtained by normalizing \( \phi_{\text{max}} \) axes. Colors represent the absolute values of \( \beta \).
averaging over modes (i.e., compute \(\frac{Z_{xy} - Z_{yx}}{2}\) for each site and each period), and then over sites. The effective apparent resistivity resulting from this process was 114.3 \(\Omega\)m. Furthermore, inversion tests were conducted for homogenous half-spaces with resistivities of 10, 100, 150, and 1,000 \(\Omega\)m (see Text S1 and Figures S1–S4 in the supporting information). Consistent with the effective apparent resistivity, the 100-\(\Omega\)m homogenous half-space provided the minimum misfit, and was used as the prior model for subsequent inversions.

The joint inversion of the Z and Tz data set terminated after 109 iterations with an overall rms error of 2.9. As presented in cross sections (Figure 4) and selected depth slices (Figure 5), the inverted 3-D resistivity model had several typical features: (1) Junggar basin sediments were clearly visible to the SE with low resistivities; (2) a highly resistive zone of more than 1,000 \(\Omega\)m with isolated conductors was identified beneath the southern part of the Zaire mountain to a depth of about 40 km; and (3) to the north, a low-resistivity anomaly of 10–50 \(\Omega\)m was found beneath 10-km depth with an E-W boundary oblique to the Darbut fault. These results are very similar to the result of the 3-D inversion of a MT transect with 182-km length, including all sites of

![Figure 3. Real induction arrows for (a) 0.77 s, (b) 25.9 s, (c) 58.5 s, (d) 132.2 s, (e) 674.7 s, and (f) 1,161.6 s.](image-url)
Profi2, which was performed in the ModEM frame by only using off-diagonal components ($Z_{xy}, Z_{yx}$) as inputs (Xu et al., 2016).

In general, the data fit is satisfactory from direct comparisons between the measured and predicted data for all profiles (Figures S5–S8). The rms varying with sites is evenly distributed, except for tippers, which have

**Figure 4.** Cross sections from the joint inversion of impedance tensors $Z$ and tippers $T_z$, shown for (a) Profile 1, (b) Profile 2, and (c) Profile 3. The inverted triangles represent the projected locations of sites along the profiles.
high misfits particularly for Profile 3 (Figure S9). Furthermore, a comparison between predicted data and observed data at representative sites indicates that poor fits of the apparent resistivity and impedance phase mainly originate from the long period (>500 s) due to the modest data quality (Figure S10). However, for the periods of 25–132 s, the $\phi_{\text{max}}$ directions of phase tensors (Figures 6a, 6c, and 6e) and real induction arrows (Figures 7a, 7c, and 7e) have large differences between Darbut and Karamay-Urbo faults. These are particularly significant for Profile 3. Moreover, the $\phi_{\text{max}}$ directions of phase tensors to the south of the Karamay-Urbo fault do not show the same feature than the real data, that is, rotating synchronously with increasing periods.

5. Evidence for Anisotropy

For a complex area, it is difficult to determine whether the underground structure is electrical anisotropy using MT data only. However, a comprehensive judgment can be made from a variety of possible phenomena or indicators that are contained in the observed data as well as from other geophysical data. Moreover, the MT data from an anisotropic model should also be interpretable by an isotropic model of arbitrary complexity (Weidelt, 1999). However, as previously pointed out by Martí (2014), this consideration does not exclude the interpretation of the MT data by structural anisotropy as a result of spatial averages over isotropic structures with preferred orientation. This might be the case of Western Junggar as evidenced by the following:

1. The isotropic model inverted from joint Z and Tz data cannot recover the $\phi_{\text{max}}$ directions of phase tensors and the real induction arrows for the periods of 25–132 s. This is particularly the case in the southeastern part, where the phase tensors and real induction arrows all show slightly counterclockwise rotations during these periods.

Figure 5. Model resistivities from the joint inversion of impedance tensors Z and tippers Tz, shown for layers at depths of (a) 1.2 km, (b) 6.1 km, (c) 12.7 km, and (d) 33.5 km.
2. The isotropic models that were inverted from different data sets all show alternating low- and high-resistivity bodies at a depth of 5 km to the lowermost crust between the Darbut and the Karamay-Urho faults (Figures 4, 5, 8, and 9; detail data fits of separate inversions can be accessed in Text S2 and Figures S11–S16). Furthermore, the depth range roughly corresponds to the penetration depths for MT data in the periods of 25–132 s. These features are strong indications for the occurrence of macroanisotropic bodies (Wannamaker, 2005), although these bodies could be interpreted as realistic structures (e.g., Zhang et al., 2016).

3. Compared to the models inverted from Z + Tz (Figures 4 and 5) and Z only (Figure 8), the model inverted from Tz only (Figure 9) shows different electrical structures below 5 km between the Darbut and the Karamay-Urho faults. Prominent differences are the resistivity contrast and spatial distribution of anomalies, which have also been reported by previous studies in the context of macroanisotropy (e.g., Kapinos et al., 2016).

Figure 6. Comparison of $\phi_{\text{max}}$ for the observed (red color) and the predicted values of the inverted isotropic model in Figures 4 and 5 (left, blue color) and the anisotropic model in Figure 11 (right, black color) for the periods of (a and b) 25.9 s, (c and d) 58.5 s, and (e and f) 132.2 s.
4. Corresponding to the area and depth range, strong seismic negative radial anisotropy is present (Figure S17), which may indicate that MT anisotropy should be considered (Ye et al., 2018). The seismic radial anisotropy \( \xi \) is determined from the velocities of SH and SV waves as 
\[ \xi = \frac{2(V_{\text{SH}} - V_{\text{SV}})}{V_{\text{SH}} + V_{\text{SV}}} \] (e.g., Luo et al., 2013). The velocities of SH and SV waves can be inverted from Love and Rayleigh wave dispersion data. A negative \( \xi \) indicates that the velocity of the SV wave is larger than that of the SH wave.

The seismic radial anisotropy in the investigated region was calculated using the data of Wu et al. (2018).

To further understand the behaviors described under (2) and (3) for MT data, a numerical test was conducted using synthetic data. Two simple 3-D models were constructed, where each isotropic/anisotropic cuboid was embedded in an isotropic half-space (Figures 10a and 10b). To model 3-D conductivity anomalies, the "RF Module-Electromagnetic Waves" of Comsol Multiphysics™ with the direct solver UMFPACK (Davis, 2004) was used. A forward finite element code that allows for anisotropy was provided by Comsol Multiphysics™ (Löwer & Junge, 2017). The Z and Tz data of 1, 2.5, 6.3, 15.8, 39.8, and 100 s were

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**Figure 7.** Comparison of the observed (red color) and the predicted real induction arrows of the inverted isotropic model in Figures 4 and 5 (left, blue color) and the anisotropic model in Figure 11 (right, black color) for the periods of (a and b) 25.9 s, (c and d) 58.5 s, and (e and f) 132.2 s.
calculated at 63 sites along three separate profiles for both 3-D models. The construction method of finite element grids was utilized, following Löwer and Junge (2017).

The MT responses of the anisotropic models are modeled by Comsol Multiphysics™, while the predicted MT responses from 3-D isotropic inversion are based on the finite difference algorithm in ModEM. The difference of forward modeling between both systems was compared, and the MT responses from Comsol are close to those from ModEM (see Text S3 and Figures S18 and S19).

For all inversions of synthetic data, ModEM was used to fit Z or Tz. The inversion mesh consisted of 93 × 93 × 40 nodes with a lateral 0.5-km grid spacing within the central area (40 km × 40 km), which is nested within a larger domain with the dimension of 329 km × 329 km. For the vertical direction, 40 layers were set with thicknesses of 30–30,000 m, which were equally spaced in the log domain. The lower boundary was set to a depth of 185 km. Error floors were set to 1% of $|Z_{xy}Z_{yx}|^{1/2}$ for all impedances and to 0.001 for tippers. A homogenous half space of 100-Ωm resistivity was used as starting model. For the isotropic model, the inversions of Z only and Tz only terminated with rms values of 1.05 and 1.04, respectively. The inverted results of Z only and Tz only at a depth of 5.2 km present a very similar pattern (Figures 10c and 10d). With regard to the anisotropic model, the inversions of Z only and Tz only terminated with rms values of 1.07 and 1.06, respectively. The inverted results of Z only and Tz only at a depth of 5.2 km present alternating low- and high-resistivity bodies (Figures 10e and 10f). Clearly, a low-resistivity anomaly was found in the central area for models of the Z-only and Tz-only inversions. However, the resistivity contrasts and positions as well as spatial scales of the alternating low- and high-resistivity bodies are very different outside the central area.

Following the line of evidence discussed above, it is necessary to explain the presented observations by taking account of a number of electrical anisotropic bodies in the crust.

Figure 8. Model resistivities from the inversion of Z only, shown for layers at depths of (a) 1.2 km, (b) 6.1 km, (c) 12.7 km, and (d) 33.5 km.
6. The 3-D Modeling Including Anisotropy

Based on the reasons discussed in the last section, this section attempts to find a 3-D anisotropic model that can reproduce the $\phi_{\text{max}}$ directions of phase tensors and the real induction arrows, especially for the periods of 25–132 s. These were considered as the main indicators of anisotropy from MT data (Liddell et al., 2016).

For the interpretation of MT data, several attempts have been made to include electrical anisotropy in resistivity models. Liddell et al. (2016) used 2-D anisotropic models to interpret strong directional dependence and large phase splits. Based on the 2-D isotropic inversion results, Brasse et al. (2009) added an anisotropic layer with three separate anisotropic anomalies and satisfactorily fitted the observed uniform induction arrows, but not the impedance tensors. Häuserer and Junge (2011) set up a 3-D model that included an anisotropic layer below the Moho, which fitted the phase tensors and induction arrows in real data. Clearly, most of the electrical anisotropic structures in complex settings are built qualitatively, since complete 2-D and 3-D anisotropic inversions of MT data are not yet available.

Based on the results of the 3-D isotropic inversion of $Z$ and $T_z$ data and the ambient noise-based tomography (Wu et al., 2018), a 3-D model with anisotropic bodies was constructed to explain the $\phi_{\text{max}}$ directions of phase tensors and the real induction arrows along the three profiles. First, the possibility of electrical anisotropic structure at the topmost 5 km was excluded. According to 3-D comparisons between MT and seismic imaging results, the mentioned possible electrical anisotropy structure roughly coincided with the seismic negative radial anisotropic area, which guided the determination of the 3-D scale of the electrical anisotropic structure. To match the different strikes indicated by phase tensor ellipses and the real induction arrows during shorter and longer periods, the anisotropic structure was divided into two layers. The top layer consists of two anisotropic bodies (A1 and A2 in Figure 11), while the bottom layer is set to an anisotropic body (A3 in

Figure 9. Model resistivities from the inversion of $T_z$ only, shown for layers at depths of (a) 1.2 km, (b) 6.1 km, (c) 12.7 km, and (d) 33.5 km.
Figure 11). The macroanisotropy model was also evaluated from the geological structures (e.g., Yang et al., 2013). After these considerations, voluminous forward calculations were conducted to obtain acceptable parameters for every part of the anisotropic structure. During the forward calculations, the low-resistivity zones at different depths (C1, C2, and C3 in Figure 11), synthesized from the isotropic resistivity model (Figures 4 and 5), became necessary to fit the observations. Figure 11 presents the best fit model and associated parameters.

Figure 10. Results of synthetic data inversions. Plan view of 3-D (a) isotropic and (b) anisotropic anomalies. The inverted model resistivities of (c) Z-only data and (d) Tz-only data modeled from the 3-D isotropic model (a) at a depth of 5.2 km. The inverted model resistivities of (e) Z-only data and (f) Tz-only data modeled from the 3-D anisotropic model (b) at a depth of 5.2 km. The 3D anomaly ($7 \times 10 \times 7.5$ km$^3$) is embedded in a 100 Ωm half space. The top surface is at a depth of 3 km. The resistivity of the isotropic anomaly in (a) is 10 Ωm. The resistivities of the anisotropic anomaly in (b) are 100, 10, and 10 Ωm, respectively.
Figure 11. The best 3-D anisotropic model. (a) Plane view of anomalies in a depth range of 5–46 km, projected on the ground and (b–d) cross sections, positions marked by AA', BB', and CC' in (a), which refer to Profile 1, Profile 2, and Profile 3, respectively. The structures above 5 km and below 46 km are similar to those of the joint 3-D isotropic inversion. The yellow area (C1) represents the crust beneath the Junggar basin with a 5–10-km depth range and a 10-Ωm resistivity. The dark blue (A1) and green (A2) areas represent the middle-to-upper crustal anisotropic structures between the Karamay-Urho and the Darbut faults with a 5–20-km depth range and a minimum resistivity in EW direction. The resistivities along the principal axes, which are rotated clockwise 90° from north, are 200, 2,000, and 200 Ωm (A1) and 10, 500, and 10 Ωm (A2), respectively. The light blue area (C2) represents the middle-to-upper crustal structure to the northwest of the Darbut fault with a 5–20-km depth range and a resistivity of 10 Ωm. The yellow contour (C3) represents the lower crustal structure to the northwest of the Darbut fault with a 20–33-km depth range and a resistivity of 30 Ωm. The black contour (A3) represents the lower crustal anisotropic anomaly within a depth range of 20–46 km. The principal axes are rotated clockwise by 20° with allocated resistivities of 10, 500, and 10 Ωm, respectively, and a minimum resistivity in SWS-NEN direction. α represents the angle of the minimum resistivity direction rotating clockwise from the north. The background resistivity is 500 Ωm.
The allocation of the resistivity to individual anomalies, as achieved with the best fit model, is justified as follows. The lower boundary of the anisotropic model was defined by the Moho in this area, as estimated from the \( P \) wave receiver function (Wu et al., 2018), with an average depth of 46 km. The top boundaries were determined by the topmost depths of high shear velocities, which are \( \sim 20 \) km in the Junggar basin and \( \sim 5 \) km in the Zaire mountain, respectively. Electrical structures above 5 km and below 46 km were adopted from the results of the 3-D isotropic inversion (cf. Figure 5a for the shallow depth). The treatment of the anisotropic model down to 46 km is meaningful (see Text S4 and Figures S20–S23; the electrical structures are similar to those in Figure 5d). Conductor C1 with a resistivity of 10 \( \Omega \)m represents sedimentary rocks of the Junggar basin up to 10 km, and its spatial extension is derived from both the \( P \) wave receiver function and the \( S \) wave velocity model (Wu et al., 2018). Conductors C2 (at depths of \( 5 \)–\( 20 \) km of 10 \( \Omega \)m) and C3 (at depths of \( 20 \)–\( 33 \) km of 30 \( \Omega \)m) represent the lower resistivity anomalies in the crust beneath the northern area of the Darbut-North Akebastao fault (hereafter E-W boundary).

To the south of the E-W boundary, the anisotropic bodies A1 (200, 2,000, and 200 \( \Omega \)m) and A2 (10, 500, and 10 \( \Omega \)m) are azimuthally anisotropic in a depth range of \( 5 \)–\( 20 \) km with a minimum resistivity direction clockwise 90° from north, that is, EW direction. The anisotropic body A3 (10, 500, and 10 \( \Omega \)m) is at a depth range of \( 20 \)–\( 46 \) km with the minimum resistivity direction clockwise 20° from the north. In general, electrically anisotropic zones are closely related to a relatively high \( S \) wave velocity (Wu et al., 2018, Figures 11 and 12) with significant negative radial anisotropy (Figure S17). The northwest boundary of the anisotropic zones apparently coincides with the Darbut Fault, because the fault interrupted the northwestward extension of the high-velocity body (Wu et al., 2018) and the pattern of alternating extremely low and high resistivity bodies. Similarly, the other horizontal extents of the shallow anisotropic blocks (A1 and A2) were determined by the spatial distribution of alternating extremely low and high resistivity bodies and the significant negative radial anisotropy (Figure S17). The anisotropic angles \( \phi \) are related to the directions of \( \phi_{\text{max}} \) for the periods of 25–132 s. The minimum resistivity directions of the shallow and the deep anisotropic bodies are mostly parallel and perpendicular to \( \phi_{\text{max}} \), respectively.

The anisotropic model has a large rms error of 3.9 (Figure S24), which is mainly due to the high misfits of impedance \( Z \). Moreover, direct comparisons between the measured data and the predicted data from the isotropic and anisotropic models (Figures S5–S8 and S10) indicate that the poor fits mainly originate from the apparent resistivity below 10 s. This can be easily expected because the anisotropic model is constructed with the aim to recover the phase tensors and real induction arrows. In reality, the fitting of \( \phi_{\text{max}} \) directions of the real data to the modeling results of the anisotropic model is better than those of the inverted isotropic model for the periods of 25–132 s (Figure 6). This is indicated by the reduction of the mean error from 25° to 17° for the periods of 25–132 s (Figure S25). The real induction arrows related to the anisotropic model also fit the real data better than those of the inverted isotropic model (Figure 7), thus reducing the rms from 3.92 to 3.62 for the periods of 25–132 s (Figure S26). Furthermore, the rotations of the real induction arrows and the \( \phi_{\text{max}} \) bars observed in the northwest Karamay-Urho fault could be well recovered by adding an electrical anisotropic structure within the crust to the 3-D model. The necessity of both shallow and deep anisotropic blocks is discussed in Text S5 and Figures S27–S30.

7. Discussion

The electrically anisotropic model suggests two azimuthally anisotropic middle-upper crust anomalies (minimum resistivity striking east-west) at depths of 5 to 20 km beneath the region between the E-W boundary to the Karamay-Urho fault, and an azimuthally anisotropic lower crust anomaly (minimum resistivity striking N20°E) at depths of 20 to 46 km south of the E-W boundary. The anisotropic resistivity model better predicts the spatial variations and rotations of observed phase tensors and real induction arrows. However, this is a macroscale preliminary model that challenges modifications by additional heterogeneities. Thus, the origin and the geodynamic implications of the electrical anisotropic bodies are not self-evident and must be further discussed. Resistivity anisotropy can arise either from microscopic phenomena or macrostructures. However, its source sensed by MT data should be a macrostructure or a volume average of microscopic anisotropy or small isotropic structures with preferred orientation, which constitutes the starting point of the following discussion.
It has widely been recognized that serpentinites are the most anisotropic rocks in active subduction zones (e.g., Katayama et al., 2009; Reynard, 2013), especially for electrical resistivity (e.g., Guo et al., 2011; Reynard et al., 2011). However, since serpentinization will not decrease the electrical resistivity of peridotites and pyroxenites below 700 °C (Guo et al., 2011; Reynard et al., 2011), the electrical anisotropy is preferably interpreted by strongly deformed amphiboles (Wang et al., 2012). Shearing can increase the anisotropy, which has been observed in antigorite and talc rocks (Guo et al., 2011). In the studied region, an abundance of amphiboles has been found in the ophiolitic complex, where the Fe-oxides exceed 10% (wt) in general (e.g., Chen et al., 2008). Thus, in a lower crust with strong shearing, the preferred orientations of Fe-bearing amphiboles may lead to electrical anisotropy with a maximum conductivity of up to 0.1–0.01 S/m.

On the other hand, Xu et al. (2016) revealed a fossil Paleozoic intraoceanic subduction system across the Darbut belt, where a remnant oceanic slab is still trapped, yielding high S wave velocity below 20 km (Wu et al., 2018) and coinciding with the anisotropic body A3 in the model developed here. Furthermore, electrical anisotropy has been successfully related to fossil subduction slabs, that is, in the Eastern Canadian Shield across the Grenville Front (e.g., Frederiksen et al., 2006) and the Great Slave Lake shear zone (e.g., Yin et al., 2014). Moreover, seismic anisotropy has been widely reported for slabs in subduction zones (Long, 2013). Seismic anisotropy is generally attributed to the fossil frozen-in anisotropy in a subducted slab that formed at a mid-ocean ridge (e.g., Zhao et al., 2016). This can present electrical anisotropy, that is, the ratio of two principle resistivities reaching ~10 beneath the East Pacific Rise of 17°S (Baba et al., 2006). Furthermore, the seismic azimuthal anisotropy of slabs is generally associated with the direction of the subduction plate motion (e.g., Hammond et al., 2010). Based on previous MT (Xu et al., 2016) and seismic (Wu et al., 2018) studies, the hidden Karamay-Urho fault (striking NE-SW) was attributed to a Late Paleozoic intraoceanic trench (a suture at present). Assuming that the minimum resistivity direction (N20°E) of the lower crust maintains the deformation imprint of the fossil oceanic slab during its underthrusting process, an ~N70°W oblique subduction may have occurred in the Late Carboniferous. This resulted in the intense deformation perpendicular to the subduction direction at the lower crust level. Moreover, the current movement of Darbut belt is NE directed according to GPS measurements (cf. Zhang et al., 2017, Figure 1). From shear-wave splitting analyses (Bao & Gao, 2017), the polarization of the fast shear wave is also NE directed. These findings indicate that the principally tensile stress to date is along the NE-SW direction. Thus, the electrical anisotropy in the lower crust may be related to the remnant oceanic slab, together with the preferred orientations of Fe-bearing amphiboles under strong shearing.

In our anisotropic model (Figure 11), a large-scale isotropic conductor below 20 km in the northwest is notable. In the previous MT study (Xu et al., 2016), the conductor was interpreted as a confluence of bilateral detachment faults resulted from upwelling of mantle melting triggered by undergoing intraoceanic subduction. However, as the MT site distribution is sparse in the NW part of the study area, the geometry of the conductor is poorly constrained. Thus, we abstain from further discussion in the present paper.

In the middle-to-upper crust, the minimum resistivity direction (EW) rotates 70° clockwise compared to that of the lower crustal anisotropic layer at large scale, which should be empirically related to different origins and geodynamic settings. The tectonic setting of the ridge subduction with a slab window in the Late Carboniferous has been widely proposed (e.g., Ma et al., 2012; Tang et al., 2010, Tang, Wang, et al., 2012, Tang, Wyman, et al., 2012; Zhang, Xiao, Han, Ao, et al., 2011). If this ridge subduction with a slab window occurred, hydrous partial melting sourced in the upper mantle would penetrate into the overriding crust through the slab window, rising within the crust via channeling. It is widely accepted that rising magma can modify channels and their surroundings, and produce metamorphic and/or metasomatic hydrous minerals in the crust expressed as subvertical low-velocity bodies. These could result in negative radial anisotropy (Jiang et al., 2018; Luo et al., 2013) and also possibly in electrically azimuthal anisotropy (Figure 12b).

Furthermore, the Late Paleozoic mafic and ultramafic intrusions in the Western Junggar are zonally distributed and associated with NE-SW and NWW-SEE faults (Yin et al., 2010; the location is marked in Figure 1c), which indicates that a subducted ridge-transform system could have existed during that time (Ma et al., 2012). Based on the aforementioned studies, the present study proposes that the subduction continued until ~300 Ma or even later (Tang et al., 2010; Xiao et al., 2008) but then failed in this region. Constrained from the ages of the dioritic dikes (~321 Ma; Yin et al., 2010) and adakites (~315–310 Ma; Tang et al., 2010), subduction of the ridge-transform system might have occurred in ~321–310 Ma. Corresponding to the bidirectional
distribution of mafic to ultramafic intrusions, the ridge axis and the strike of the transform fault should be close to the EW direction and N20°E, respectively. If so, the direction of the transform fault is the same as that of the electrical anisotropy in the lower crust, which indicates that the anisotropy of the oceanic slab

Figure 12. The suggested tectonic evolution model for the Late Carboniferous to the Early Permian in this research region. Modified after Tang et al. (2010) and Liu et al. (2019). APM = absolute plate movement, K-U fault = Karamay-Urho fault. The trench is the former K-U fault (NE45°). The direction of APM and subduction is NW70°. The ridge axis and the strike of the transform fault are close to the EW direction and N20°E, respectively, estimated from the bidirectional distribution of mafic to ultramafic intrusions (Ma et al., 2012; Yin et al., 2010). The anisotropy direction of the oceanic crust is NE20° evaluated from the electrical anisotropy in the lower crust, which is the same direction as for the transform fault. The mafic melts through the slab window cause subvertical low-velocity bodies, yielding negative radial anisotropy (V_{SH} < V_{SV}). The preferred orientation of mafic melts (along E-W direction, i.e., the direction of ridge) would generate the electrical anisotropy (p_{EW} < p_{NS}) in the middle-to-upper crust. The subduction of the ridge-transform system might have occurred in ~321–310 Ma, constrained by the ages of the dioritic dikes (~321 Ma; Yin et al., 2010) and adakites (~315–310 Ma; Tang et al., 2010). The other ages are discussed in detail by Xiao et al. (2008) and Tang et al. (2010).
might be mainly determined by the direction of the transform fault. Moreover, the shear along the transform fault may generate the preferred orientations of Fe-bearing amphiboles. At the time of subduction failed, the oceanic slab from the latest subduction can be preserved beneath the intraoceanic arc, that is, the Darbut belt at present. This constitutes the main component of the current lower crust of Western Junggar.

The suggested tectonic evolution model for the Late Carboniferous to the Early Permian is shown in Figure 12. Here the electrical anisotropy in the lower crust could be attributed to the fossil frozen-in anisotropy in the remnant oceanic slab (Figure 12a), together with the preferred orientations of Fe-bearing amphiboles under strong shearing, which significantly promoted the electrical conductivity in direction of N20°E. The electrical anisotropy in the middle-to-upper crust could be ascribed to subvertically metamorphic zones with E-W extension formed by channeling mafic intrusion during the ridge subduction, where a slab window developed in the Late Carboniferous (Figure 12b), the time slightly advanced over the subduction failed. However, the argued model does not uniquely specify the materials that contribute to the orientation-related resistivity in the middle-to-upper crust. In fact, this is a macroanisotropy and represents the configuration of a failed intraoceanic ridge-transform subduction. Moreover, the large difference of the minimum resistivity directions between the middle-to-upper crust and the lower crust suggests that they are decoupled from each other in the deformation in the Darbut belt. This feature is possibly universal in this region since a relatively weak lower crust does have been developed in the Junggar area (Deng & Tesauro, 2016). However, these argumentations are rather speculative since they are based on few constraints.

8. Conclusions

This study deployed three MT profiles in the Darbut belt of Western Junggar, NW China. The analyses of phase tensors and real induction arrows indicate a 3-D anisotropic resistivity structure. The isotropic joint inversion of Z and Tz data did not reproduce the amplitudes and significant rotation of the real induction arrows and the $\phi_{\text{max}}$ bars with increasing periods to the south of the Darbut Fault. Furthermore, the inverted models indicate patterns of rather high-resistivity contrasts below 5 km between the Darbut and the Karamay-Urholo faults. These findings corroborate the assumption of an electrically anisotropic model, which is constructed by forward modeling under the analyses of phase tensors and real induction arrows, the results of 3-D isotropic inversions, and the seismic observations. The anisotropic resistivity model predicts better the spatial variations and rotations of observed phase tensors and real induction arrows.

The presented 3-D anisotropic resistivity model includes two azimuthally anisotropic middle-upper crust anomalies (minimum resistivity striking east-west) over depths of 5 to 20 km beneath the E-W boundary to the Karamay-Urholo fault and an azimuthally anisotropic lower crust anomaly (minimum resistivity striking N20°E) at depths of 20 to 46 km south of the E-W boundary. The origins of the electrical anisotropic body in the lower crust could be ascribed to the inherited anisotropy of remnant oceanic slab along with Fe-bearing amphiboles under strong shearing. The electrical anisotropy in the middle-to-upper crust could be ascribed to subvertically metamorphic zones with E-W extension formed by channeling mafic intrusion during the ridge subduction in the Late Carboniferous. Furthermore, the minimum resistivity directions of the anisotropic middle-to-upper crust and lower crust deviate by ~70°, implying that the middle-to-upper crust in the Darbut belt is decoupled from its underlying lower crust.

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