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Lithosphere Mantle Density of the North China Craton

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Key Points:

- STP lithospheric mantle density of the NCC ranges from 3.32 to 3.40 g/cm³
- Most of the NCC lithosphere lost cratonic characteristics
- Lithosphere density anomalies do not follow geological boundaries

Supporting Information:

- Supporting Information S1

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Abstract We constrain the lithospheric mantle density of the North China Craton (NCC) at both in situ and standard temperature-pressure (STP) conditions from gravity data. The lithosphere-asthenosphere boundary (LAB) depth is constrained by our new thermal model, which is based on a new regional heat flow data set and a recent regional crustal model NCcrust. The new thermal model shows that the thermal lithosphere thickness is <120 km in most of the NCC, except for the northern and southern parts with the maximum depth of 170 km. The gravity calculations reveal a highly heterogeneous density structure of the lithospheric mantle with in situ and STP values of 3.22–3.29 and 3.32–3.40 g/cm³, respectively. Thick and reduced-density cratonic-type lithosphere is preserved mostly in the southern NCC. Most of the Eastern Block has a thin (90–140 km) and high-density lithospheric mantle. Most of the Western Block has a high-density lithospheric mantle and a thin (80–110 km) lithosphere typical of Phanerozoic regions, which suggests that the Archean lithosphere is no longer present there. We conclude that in almost the entire NCC the lithosphere has lost its cratonic characteristics by geodynamic processes that include, but are not limited to, the Paleozoic closure of the Paleo-Asian Ocean in the north, the Mesozoic Yangtze Craton flat subduction in the south, the Mesozoic Pacific subduction in the east, the Cenozoic remote response to the Indian-Eurasian collision in the west, and the Cenozoic extensional tectonics (possibly associated with the slab roll-back) in the center.

1. Introduction

The lithosphere, comprising the crust and the lithospheric mantle (LM), is one of the most fundamental layers in the Earth's evolution. Old cratonic continental LM is expected to be made of low-density melt-depleted peridotite so that the cratons remain stable over billions of years (Carlson et al., 2005; Jordan, 1978, 1988; Lenardic et al., 2003). The composition of the cratonic LM plays a significant role in the long-term evolution of continents (Artemieva et al., 2002; Griffin et al., 2003; Lee et al., 2011; Lenardic et al., 2003; Sleep, 2005). It is constrained by laboratory studies of mantle-derived xenoliths with sparse localized horizontal and vertical sampling (Griffin et al., 2003; Lee, 2006), which questions its representativeness (Artemieva et al., 2019).

In general, the mineral and chemical composition, thickness, temperature, and density of the LM depend on the lithosphere tectonic-thermal age (Artemieva & Mooney, 2001; Djomani et al., 2001; Kaban et al., 2003). Density depends on the composition of the LM and reflects multiple mantle modifications through geological time. Most Archean cratons have a thick (>150 km) LM with Fe-depleted composition (Mg# ~ 91.5–94), which implies its low density of 3.33 ± 0.02 g/cm³ at standard temperature and pressure (STP) condition (Djomani et al., 2001; Lee, 2003). Compared to Archean sections, the Proterozoic LM is often thinner (80–160 km), denser (3.35 ± 0.02 g/cm³ at STP condition), and less depleted (Mg# ~ 90–91.5). The Phanerozoic LM has fertile composition similar to asthenospheric mantle (Mg# ~ 88–90); it is commonly less than 80 km thick with STP densities of 3.38 ± 0.02 g/cm³ (Djomani et al., 2001). However, due to limited sampling, xenoliths cannot provide data for the entire heterogeneous LM, but instead provide a selective, and probably biased, Nature's sampling (Artemieva et al., 2019).

The cratonic lithosphere, with large volumes formed in the Archean, has remained relatively stable where it has undergone only minor modification of the LM since its formation (Lenardic et al., 2003). The

Figure 1. (a) Tectonic sketch of the North China Craton and adjacent regions with topography in background. Geological boundaries are after Zhao et al. (2001) with modifications. Blue numbers—tectonic ages of each block. Symbols—xenolith locations (triangles, squares and stars for Cenozoic, Mesozoic, and Paleozoic emplacement ages, respectively); black numbers—ages of the host volcanics (Chu et al., 2009; Dai et al., 2018; J.-G. Liu et al., 2011; Y. Xiao et al., 2010; Ying et al., 2006; Zheng et al., 2007, and references therein). WB: Western Block; TNCO: Trans-North China Orogen; EB: Eastern Block; CAOB: Central Asian Orogenic Belt; NCC: North China Craton; YC: Yangtze Craton. (b) Free-air EGM-2008 gravity anomalies (Pavlis et al., 2012).

sedimentary cover and the occurrences of magmatic rocks emplaced since cratonization hold a record of the tectonic and geologic evolution of the lithosphere. However, recent studies suggest that many cratonic blocks (e.g., the Kaapvaal and Wyoming Cratons) have experienced several lithosphere modification events (J.-S. Hu et al., 2018; Levander et al., 2011; Youssof et al., 2015). Different mechanisms may modify the LM, including hot spot-lithosphere interaction such as below the Western Gondwana craton (J.-S. Hu et al., 2018) and collisional and subduction processes such as beneath the Wyoming Craton (Levander et al., 2011).

The NCC, surrounded by terranes of Paleozoic to late Mesozoic ages, provides an ideal opportunity to probe the processes that modify the LM (Figure 1a). The existing data on the lithosphere structure of the NCC are limited. Geological studies provide only surface information, xenoliths have a limited geographical coverage

Figure 2. Surface heat flow (G-Z. Jiang et al., 2019) with borehole measurements deeper than 200 m. (a) Point data and xenolith locations and (b) interpolated values (kriging with an interpolation radius of 1°). Color scales in (a) and (b) are the same. Background in (a)—shaded topography; crosses in (b)—heat flow boreholes. Red numbers in (a)—locations of xenoliths used to construct P-T arrays (Figure 3) (the first letter corresponds to the emplacement age: C: Cenozoic, M: Mesozoic, and P: Paleozoic).

(Figure 2a), seismic tomography results may be affected by anisotropy and fluids, in particular in the eastern NCC affected by the Mesozoic Pacific subduction, and thermal studies are significantly uncertain. The Archean Eastern Block (EB) of the North China Craton (NCC) has high surface heat flow with a mean average value of $>60 \text{ mW/m}^2$ (Figure 2), which suggests a thin lithosphere (He, 2015; S.-B. Hu et al., 2000). Analysis of sparse kimberlite and garnet-xenolith samples (chiefly based on two Paleozoic mantle-derived xenoliths in the eastern NCC, Figure 2a) suggests that the subcontinental LM beneath the eastern margin of NCC may have extended down to 200 km depth (Fan et al., 2000; Griffin et al., 1998; Menzies et al., 1993; Y.-G. Xu, 2001) in the Paleozoic (Q.-L. Li, Wu, et al., 2011; Lu et al., 1998) when it had typical Archean cratonic mantle geochemical characteristics (Griffin et al., 1998). In contrast, mantle xenoliths carried by Mesozoic-Cenozoic magmas indicate a thin lithosphere ($<80 \text{ km}$) with fertile composition (Griffin et al., 1998; Zheng et al., 2007) and lithosphere reworking is usually explained by the Mesozoic Pacific

subduction (Fan et al., 2000; Gao et al., 2002; Griffin et al., 1998; Menzies et al., 1993; Y.-G. Xu, 2001) through lithosphere delamination (Godey et al., 2004; Wu et al., 2002), thermochemical erosion (Y.-G. Xu, 2001), water metasomatism (Q.-K., Xia et al., 2010), or a combination of these three components in relation to a possible ridge subduction (Ling et al., 2013).

While the presence of thin lithosphere in the EB is supported by various seismic studies, detailed information on the LAB (lithosphere-asthenosphere boundary) depth beneath the NCC is still controversial (Chen et al., 2006; Huang et al., 2009; S.-J. Wang et al., 2014; Y.-Y. Zhang et al., 2019). *S* receiver functions indicate that the LAB depth ranges from 70 to 100 km in most of the EB, reaching down to a 140–160 km depth in its southwestern part (Chen et al., 2006; Y.-Y. Zhang et al., 2019); but controversy remains if the imaged converter corresponds to the LAB (Romanowicz, 2009; Thybo, 2006). Similarly, a regional surface wave tomography model was interpreted in terms of a 70–130 km thick lithosphere with local anomalies down to 160 km (Huang et al., 2009), and an ~1,500 km long W-E seismic refraction profile across the NCC indicates that the LAB is at a depth of 75–90 km (S.-J. Wang et al., 2014).

Recent detailed geophysical and geological studies suggest that the LM of the Western Block (WB) and the Trans-North China Orogen (TNCO) may have also experienced modification/reworking (Chen, 2010; Guo & Chen, 2017; M.-M. Jiang et al., 2013; Wan et al., 2013; Zang et al., 2005). The presence of Cenozoic-Mesozoic basalts with mantle xenoliths at the northern edge of the WB and at the TNCO implies regional LM modification (J.-G. Liu et al., 2011; Y.-J. Tang et al., 2008; Y.-G. Xu et al., 2008). New high-quality borehole measurements provide evidence for a relatively high surface heat flow (>65 mW/m²) in the WB of the NCC (He, 2015; G.-Z. Jiang et al., 2019), consistent with regional surface wave tomography (An et al., 2009; Bao et al., 2013), which indicates that the seismic LAB is at a depth of less than 150 km. *S* receiver function studies also suggest that the LAB is at a depth of 120–160 km; that is, the lithosphere is 50–100 km thinner than in typical cratonic environment (Chen, 2010; Y.-Y. Zhang et al., 2019). The electrical resistivity structure of the WB mantle as constrained by magnetotelluric data shows a low resistivity anomalous zone between 90 and 150 km depths and indicates that the LM may have experienced regional modification (Dong et al., 2014). Joint inversion of topography, gravity, and geoid indicates that the LAB depth varies from 120 to 175 km in most of the WB and down to 205 km at its western part (Y. Xu et al., 2016).

In this study, we focus on the structure of the LM in the whole NCC. We first analyze the thermal regime of the lithosphere based on a new regional heat flow data set and mantle xenoliths data. Next we constrain the LM density from gravity data to determine the amplitude and the extent of the LM reworking beneath the NCC. Residual mantle gravity anomalies (RMGA) is calculated from the free air gravity anomaly (Figure 1b) by subtracting the gravitational effect of topography, sedimentary cover, the crystalline crust, the Moho, and the LAB undulations. Our recent high-resolution crustal model NCcrust (Xia et al., 2017), as well as a new thermal model for the lithosphere thickness provide constraints for the calculation. LM densities are calculated at both in situ and standard-temperature-pressure (STP) conditions ($T = 20^\circ\text{C}$, $P = 1$ atm) under the assumption that all RMGA anomalies are caused by density anomalies in the LM. STP densities do not depend on temperature but only on the composition of the mantle, which reflects the long-term evolution of the region.

2. Geological Background

Precambrian: The NCC is composed of a Neoproterozoic to Paleoproterozoic metamorphic basement and a Proterozoic to Cenozoic sedimentary cover. While the EB is Neoproterozoic, the basement of the WB consists of Paleoproterozoic sedimentary rocks without information on the Archean component (Wan et al., 2013). During the Paleoproterozoic (1.95–1.85 Ga), the amalgamation of the EB and WB along the SW-NE trending TNCO resulted in the assembly of the NCC.

Paleozoic-early Mesozoic: The NCC was magmatically quiescent until the eruption of kimberlites in the Paleozoic (ca. 480 Ma) (Q.-L. Li, Wu, et al., 2011; Lu et al., 1998). In the north, the Paleozoic Central Asian Orogenic Belt (CAOB) between the Siberian craton and the NCC records the closure of the Paleo-Asian Ocean (Windley et al., 2006; W.-J. Xiao et al., 2003). Garnet-xenoliths with olivine-Mg# > 92 suggest that the Archean LM still existed beneath the EB during the Paleozoic (Zheng et al., 2005), while the Re-Os age of these garnet xenoliths suggests minor modification of the LM before the Paleozoic (Godey et al., 2004). In the south, the Paleozoic (440–430 Ma) and Mesozoic (230–220 Ma; S.-G. S.-G. Li

et al., 1993) collision of the Yangtze and the NCC cratons resulted in northward subduction of the Yangtze Craton beneath the NCC and led to the formation of the ultrahigh-pressure Qinling-Dabie orogenic belt (Figure 1a).

Late Mesozoic-Cenozoic: The Mesozoic subduction of the Pacific plate produced a widely distributed magmatic activity through the entire late Mesozoic to Cenozoic with peak at 130–110 Ma (Wu et al., 2005). Xenoliths carried by Mesozoic and Cenozoic basalts indicate that by that time the LM of the EB had acquired a fertile composition (Griffin et al., 1998). The western edge of the region affected by the Mesozoic subduction is usually placed at a sharp gradient in the Bouguer anomalies (the so-called North-South Gravity Lineament) ~100 km west from the pronounced topographic ramp at the border between the EB and the TNCO (Figure 1a) (e.g., Chen, 2010). Limited xenolith studies from the TNCO and the eastern margin of the WB suggest that the lithosphere beneath the central western NCC has been also partially modified by multiple metasomatic enrichments in the Cenozoic (J.-G. Liu et al., 2011).

3. Input Parameters for Gravity Calculations

3.1. Free-Air Gravity

We use free-air gravity anomalies from the EGM2008 model (Figure 1b), defined on a $5' \times 5'$ grid (Pavlis et al., 2012) to calculate residual gravity anomalies, from which we calculate the LM density. The magnitude of the free-air gravity anomalies is smaller than ± 20 mGal in most parts of the EB, which suggests that this region is close to isostatic equilibrium. The TNCO clearly stands out in free-air gravity anomalies by a heterogeneous pattern with values between -80 and $+100$ mGal. The WB is close to isostatic equilibrium with free-air gravity values in the central part within ± 40 mGal, which increase in amplitude to ± 80 mGal to the north and south.

3.2. Crustal Structure

To calculate the gravitational effect of the crust, we use a recent regional crustal database NCcrust (Xia et al., 2017), which includes the seismic crustal structure based on a four-layer model (sedimentary cover, upper crust, middle crust, and lower crust). It is constrained only by seismic data (reflection and refraction seismic profiles, as well as receiver functions), which makes it ideal for gravity corrections for the crustal structure. NCcrust includes the Moho depth as well as thickness and P wave velocity of each internal crustal layer.

The crustal model is used to

1. constrain densities of the sedimentary cover and the crystalline crust by converting V_p to density (Figure S1 in the supporting information);
2. calculate the gravity effect of the sedimentary cover, the crystalline crust and the Moho depth variation (section 5.3); and
3. calculate the thermal lithospheric thickness based on a new model of surface heat flow (section 4.1).

We calculate density of each crustal layer from seismic velocity by using an experimental relationship (Brocher, 2005) between the P wave velocity (v_p) and density (ρ):

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

The formula is valid for the range of velocity values between 1.5 and 8.5 km/s and therefore includes the whole range of velocities in our NCcrust model for sedimentary sequences, crystalline crust and uppermost mantle.

In the northern part of the TNCO and in the eastern part of the EB, where Quaternary sedimentary sequences are present, the density of the sedimentary cover ranges from 1.8 to 2.2 g/cm³ (Figure S1a). The remaining part of the NCC, dominated by hard sediments, has average density of 2.2 to 2.5 g/cm³ in the sedimentary cover.

The average density of the entire crust (including sediments) is calculated as the weighted average of all crustal layers (Figure S1b):

$$\rho = \frac{\sum \rho_i h_i}{\sum h_i} \quad (2)$$

The average density of the entire crust ranges from 2.70–2.72 g/cm³ in the northern part of the TNCO and the EB with thick low-density sediments to 2.76–2.78 g/cm³ in the southern part of the TNCO and the EB where the sedimentary cover is ~2 km thick. The WB with a thick sandstone sequence has an intermediate average crustal density of 2.74–2.78 g/cm³.

3.3. LM Thickness

Calculation of RMGA requires knowledge on the gravitational effect of the crust, which can be calculated from the crustal density model NCcrust (Xia et al., 2017). However, the next step, calculation of mantle density anomalies, requires additional information on the thickness of the layer to which the density anomalies are confined. Since the effect of crustal density heterogeneity has been taken into account for calculation of RMGA, the top of the mantle layer with density heterogeneity corresponds to the Moho depth. Geochemical studies of mantle-derived xenoliths from cratonic settings worldwide indicate a gradual decrease with depth in mantle depletion within the lithosphere mantle (Griffin et al., 1998), such that the cratonic mantle composition gradually approaches the fertile composition of the asthenospheric mantle in the bottom part (~50 km thick) of the LM. This ~50 km difference in the adopted base of the layer responsible for RMGA leads to significantly different values of LM density (Cherepanova & Artemieva, 2015). In some cratons, geochemical data indicate a layered structure of the LM, with a depleted upper portion and a fertile bottom portion of the LM (Griffin et al., 1999; Kopylova & Card, 2004; Lehtonen et al., 2004).

In the absence of xenolith-based information on mantle depletion in the interior parts of the NCC, it is somewhat arbitrary where to place the bottom of the mantle layer, which produces RMGA. However, it is unlikely to be deeper than the LAB, and as an end-member model we assume that the base of the layer that produces RMGA corresponds to the LAB. Therefore, the LAB depth is a parameter required for the gravity calculations, and we determine the lithosphere thermal thickness in the study region by calculating lithosphere geotherms.

4. Lithosphere Thermal Model

4.1. Lithosphere Geotherms

To calculate the lithosphere geotherms and the thermal LAB, we follow a standard approach (Artemieva & Mooney, 2001) constrained by a new regional surface heat flow compilation (Figure 2; G.-Z. Jiang et al., 2019). For simplicity and given the uncertainty of thermal parameters, we define the thermal LAB by the 1300°C isotherm, which introduces only a minor difference to results based on a definition of the LAB by the intersection of geotherm with a 1300°C mantle adiabat. Our assumption on a 1-D steady-state heat transfer is justified by a relatively long-term stability of most of the region (which is therefore close to a steady-state regime) and the sparse distribution of borehole heat flow data (so that 3-D thermal effects cannot be taken into account; Jaupart, 1983; Petitjean et al., 2006; Stephenson et al., 2009). Note that the assumption of a steady-state heat transfer may not be entirely valid for the eastern part of the region affected by the Mesozoic subduction at 110–130 Ma. The time delay for a thermal front associated with a thermal perturbation at depth z to reach the surface and to become reflected in the surface heat flow is

$$t \sim z^2/\chi, \quad (3)$$

where χ is thermal diffusivity, typically 1 mm²/s (Turcotte & Schubert, 2014). It means that over the time of 110–130 Ma, the thermal front associated with a mantle temperature anomaly has propagated upward by 60–65 km, which is less than the lithosphere thickness in the EB. However, both the dip angle of the Pacific subduction and its lateral extent beneath the NCC are unknown, and it is possible that various parts of the EB has either reached the steady state (in places where the depth to the slab was 60–65 km, as flat subduction implies, Ling et al., 2013) or did not experience the steady-state disruption in Mesozoic at all, as may be the case in the western EB. In case steady state is not yet reached in parts of the EB, heat flow may not be representative of the Mesozoic thermal anomaly (is too low), and our model will overestimate the LAB depth. However, we show below (section 4.3) that our calculated xenolith pressure-temperature (P-T) arrays for Cenozoic and Mesozoic xenoliths fall on the steady-state

geotherms that match the observed heat flow in the EB and the TNCO and therefore justify our assumption.

To calculate lithosphere geotherms, we use a standard iterative approach by applying Formula 4a to each lithosphere layer (sediments, upper, middle, and lower crust, and LM) iteratively downward from the surface (Furlong & Chapman, 2013; Turcotte and Schubert, 2014):

$$T(z) = T_{\text{top}} + zQ_{\text{top}}/k - z^2 A/2k, \quad (4a)$$

where z is depth within the layer, T_{top} and Q_{top} are temperature and heat flow at the top of the layer (for the top layer $T_{\text{top}} = 0^\circ\text{C}$ and Q_{top} is measured heat flow), and A and k are heat production and thermal conductivity within the layer. At each step this equation is used for calculating temperature at the base of the current layer (that is the top of the next layer used in the iteration) by replacing z in the equation above by the layer thickness h ; heat flow at the base of the current layer Q_{base} is calculated by subtracting radiogenic heat flow generated in the layer from heat flow at the top of the layer Q_{top} (which is surface heat flow for the sedimentary layer):

$$Q_{\text{base}} = Q_{\text{top}} - Ah, \quad (4b)$$

where A is average heat production in the current layer (Artemieva & Mooney, 2001; Furlong & Chapman, 2013).

Additionally, the Moho temperature (T_{Moho}) is needed for calculation of average temperature in the LM (T_{LM}): $T_{\text{LM}} = (T_{\text{Moho}} + T_{\text{LAB}})/2$, where $T_{\text{LAB}} = 1300^\circ\text{C}$. We use T_{LM} to calculate the reference LM density (section 5.1) and to apply temperature correction to in situ LM density in order to calculate STP density (section 5.6)

The thermal model is, therefore, constrained by the following parameters: surface heat flow, heat production, and thermal conductivity in each lithospheric layer. In our calculations, we exclude shallow (<200 m) heat flow measurements because we generally expect that they are subject to large uncertainty, for example, due to either a possible groundwater circulation (Pollack et al., 1993) or possible climate variations (Balling, 1995; Guillou-Frottier et al., 1998).

Various studies based on local and regional measurements of thermal parameters on rock outcrops (Joeleht & Kukkonen, 1998), including the exposed sections of deep crust (Ashwal et al., 1987; Fountain et al., 1987; Pinet & Jaupart, 1987), show large short-wavelength variations in thermal parameters of crustal rocks. However, such studies remain very sparse in the NCC. In the absence of a regional-scale database, we follow a commonly adopted strategy (Artemieva & Mooney, 2001; Balling, 1995; Furlong & Chapman, 2013; Goes et al., 2020; Jaupart & Mareschal, 1999; Mareschal, 1991). For each crustal layer (their thickness is constrained by a high-resolution regional crustal model NCcrust; Xia et al., 2017) and the LM, we adopt standard values of heat production (Artemieva & Mooney, 2001; Ashwal et al., 1987; Balling, 1995; Fountain et al., 1987; Gard et al., 2019; Jaupart & Mareschal, 1999; Pinet & Jaupart, 1987; Rudnick et al., 1998) and thermal conductivity (Cermak & Rybach, 1982; Clauser & Huenges, 2013; G.-Z. Jiang et al., 2019; Schatz & Simmons, 1972; Seipold, 1992; Y. S. Xu et al., 2004) compiled from the literature and regional studies (Table 1), including the temperature and pressure dependence of thermal conductivity (Cermak & Rybach, 1982). Therefore, our thermal model is a regional update of a global continental model (Artemieva & Mooney, 2001) based on a new regional heat flow compilation (Figure 2; G.-Z. Jiang et al., 2019).

Alternatively, the lithosphere thermal structure could potentially be calculated from seismic velocity models (Goes et al., 2000; Priestley & McKenzie, 2006). However, separation of thermal effects from seismic anomalies reveals significant LM compositional heterogeneity (Artemieva, 2009), further supported by xenolith data (Griffin et al., 1998) and by studies that show complex effects of mantle partial melting (that forms LM as the restite of mantle melting) on mantle seismic velocities and density (Afonso & Schutt, 2012). Nonetheless, seismic velocity to temperature conversions either neglect compositional heterogeneity of the upper mantle (which is the goal of the present study) (Goes et al., 2000; Priestley & McKenzie, 2006) or make a priori assumptions on compositional differences between continental lithosphere of different ages (Godey et al., 2004; Goes & van der Lee, 2002). Other factors, such as anisotropy, the presence of fluids and

Table 1
Thermal Parameters Used in Calculations

	$K(z)$ (W/m/K)	Reference	$A(z)$ ($\mu\text{W}/\text{m}^3$)	Reference
Sedimentary cover	2.4 ^a	G.-Z. Jiang et al. (2019)	1.2	Artemieva and Mooney (2001)
Upper crust (0–10 km)	3.0	Cermak and Rybach (1982) and Clauser and Huenges (1995)	1.0	Gao et al. (1998) ^b and Gard et al. (2019) ^c
Upper crust (>10 km)	3.0	Cermak and Rybach (1982) and Clauser and Huenges (1995)	0.4	Artemieva and Mooney (2001)
Middle crust	2.5	Clauser and Huenges (1995) and Seipold (1992)	0.3	Artemieva and Mooney (2001)
Lower crust	2.0	Clauser and Huenges (1995) and Seipold (1992)	0.1	Artemieva and Mooney (2001)
Lithospheric mantle	4.0	Schatz and Simmons (1972) and Y. S. Xu et al. (2004)	0.01	Artemieva and Mooney (2001); Rudnick et al. (1998)

^aRegional mean thermal conductivity 2.4 ± 0.4 W/m/K with variations from 0.9 to 4.9 W/m/K. ^bCalculated from crustal composition in East China. ^cRegional mean heat production 1.0 ± 0.5 $\mu\text{W}/\text{m}^3$ with values in the range from 0.2 to 3.2 $\mu\text{W}/\text{m}^3$.

melts, and grain size variations, have also strong effect on seismic velocities (Faul & Jackson, 2005), but there is no direct information on their possible contribution to seismic velocity variations at the NCC mantle or elsewhere. We therefore do not use seismic models as a proxy for lithospheric temperatures, since they also seem to be poorly correlated to thermal lithosphere structure (Goes et al., 2000). In principle, we could use Curie temperature depths to constrain the lithosphere structure where surface heat flow measurements are sparse. However, regional aeromagnetic studies show that there is no direct relationship between the Curie temperature depth and surface heat flow (Xiong et al., 2016), similar to global observations (C.-F. Li et al., 2017). Therefore, it will introduce unacceptable error and uncertainty if we constrained the thermal structure from the Curie depth (Zang et al., 2002). We therefore, do not include the Curie depths into our thermal model.

4.2. Sensitivity Analysis

The true range of lateral and depth variations of thermal parameters used in the thermal calculations is unknown, and we perform a sensitivity analysis by changing the input parameters by a fixed value within physically reasonable limits (Table 2). The largest uncertainty is associated with the surface heat flow, which arises from three factors: (i) use of interpolated values in the absence of borehole data in many regions (especially in the TCNO, the Dabie Orogen, and the northern part of the WB; Figure 2a); (ii) the unknown quality of borehole data, which we partially compensate by excluding shallow boreholes; and (iii) potential delay of the thermal front associated with the Mesozoic subduction below the EB, which may not be fully reflected in the measured heat flow. Factors (i) and (ii) are not likely to introduce regional-scale heat flow errors greater than 10%, which would change the LAB depth by $\sim 20\%$ and will introduce a <0.01 g/cm³ change in the lithosphere mantle density (Table 2). Factor (iii) may be essential only along the Pacific coast (discussed in sections 4.1 and section 4.3). In case the thermal front associated with the Mesozoic subduction has not yet reached the surface, heat flux may be $\sim 20\%$ underestimated, which will result in $\sim 40\%$ overestimated LAB depth and ~ 0.02 g/cm³ underestimated lithosphere mantle density (Table 2). This shift will not change the results qualitatively.

The choice of heat production values in the crust, in particular in sediments and upper crust, also has strong effect on the thermal model (Artemieva & Mooney, 2001; Goes et al., 2020). This information is largely unavailable for the NCC. A global compilation of heat production in granitic rocks of the upper crust (Artemieva et al., 2017) includes no measurements from our region, while other databases dominated by mafic rocks

Table 2
Sensitivity Analysis for Thermal Calculations

	Regionally average value in the model	Test model, absolute value	Parameter change in %	Resultant change in lithosphere thickness	Resultant change in LM density
Average crustal heat production ($\mu\text{W}/\text{m}^3$)	0.6	0.5	−20%	<14% thicker	<0.01 kg/m ³ reduction
Average crustal conductivity (W/m/K)	2.6	2.1	−15%	<10% thinner	<0.01 kg/m ³ increase
Lithospheric mantle conductivity	4	3.5	−13%	<9% thinner	<0.01 kg/m ³ increase
Surface heat flow (mW/m^2)	65	80	+23%	<45% thinner	<0.02 kg/m ³ increase

Figure 3. (a) Regional geotherms (colored lines) calculated from heat flow data and xenolith P-T arrays in the NCC (gray area enlarged in panels b–d), compared with reference geotherms of Pollack and Chapman (1977) (black lines labeled by surface heat flow). For each heat flow value, we show three geotherms that correspond to the average crustal structure of the EB (solid lines), WB (dashed lines), and TNCO (dotted lines). (b–d) P-T pairs calculated for spinel xenoliths with Mesozoic and Cenozoic emplacement ages in three subblocks of the NCC (numbers correspond to locations in Figure 2b). Colored lines—geotherms as in (a). All three tectonic subblocks follow a 65–75 mW/m^2 geotherm, with two locations in the Sulu Belt (C5 and M2) following a 80 mW/m^2 geotherm, in overall agreement with the present-day heat flow (Figure 2b). Xenolith data do not show any age progression in lithosphere temperatures from Mesozoic to Cenozoic, possibly except for the M4 location in the TNCO, which also plots close to a 60 mW/m^2 geotherm as the adjacent C6 xenolith (Figure 2a).

have very sparse data in our region (Gard et al., 2019). We therefore recognize that while our thermal model is basic in the absence of regional heat production data in sediments and the upper crust, its resolution is sufficient for gravity calculations of STP LM density (Table 2).

4.3. Xenolith P-T Arrays

To examine our thermal lithosphere model, we also calculate xenolith P-T arrays (equilibrium temperatures and pressures of mantle xenoliths, Figure 3) based on our compilation of all available major element composition data (Chu et al., 2009; Dai et al., 2018; Lin et al., 2019; H.-G. Liu et al., 2011; Y.-J. Tang

et al., 2008; Y. Xiao et al., 2010; X.-S. Xu et al., 1998; Ying et al., 2006; Zheng et al., 2001, 2007; Zou et al., 2016). The equilibrium temperature is constrained by the Ca-in-Opx method (Brey & Kohler, 1990) at an assumed pressure of 15 kbar, which corresponds to a depth of ~50–80 km (Figure 3), which is the normal preset pressure for spinel faces xenoliths. The equilibrium pressure is calculated using the Cr-in-Cpx barometer (Nimis & Taylor, 2000) for spinel peridotite.

A comparison of P-T arrays for xenoliths with different emplacement ages shows that while Paleozoic P-T data plot along a slightly elevated cratonic geotherm, which corresponds to a surface heat flow of 50–55 mW/m², Mesozoic and Cenozoic xenoliths mostly plot between ~65 and 75 mW/m² geotherms and without any temporal or spatial patterns (Figures 3b–3d and S2). This difference of ~20 mW/m² between the presubduction and postsubduction xenolith geotherms justifies our choice of the model parameters in the sensitivity analysis (Table 2). Furthermore, the warmest xenolith geotherms (~80 mW/m²) correspond to two locations at the Sulu Belt, east of the Tan-Lu fault (C5 and M2 in Figure 2a), where the measured heat flow is also ~80 mW/m² (Figure 2b). The agreement between the present-day measured heat flow and heat flow constrained by both Mesozoic and Cenozoic xenoliths geotherms indicates the representativeness of the present heat flow data, which shows no delay in the upward propagation of a thermal anomaly associated with the proposed Mesozoic Pacific subduction.

Importantly, all known Paleozoic xenoliths are from the near-coastal corridor in the eastern part of the EB (along the major Tan-Lu fault zone). Therefore, xenolith data do not provide evidence for a change of lithosphere thermal regime at circa 130–110 Ma in any of the blocks of the NCC west of the Tan-Lu zone because they are absent there. We note that our estimate of the Paleozoic heat flow (~55 mW/m², Figure 2a) is slightly higher than reported previously (40–50 mW/m² based on global thermal model, Griffin et al., 1998).

4.4. Lithosphere Thermal Thickness

Lithosphere geotherms (Figure 3a) define the base of the lithosphere, the LAB. They imply that the thermal lithosphere is <100 km thick along the eastern part of the NCC (Figure 4a) and increases to >160 km in the northern and southern parts of the TNCO where surface heat flow is low as in other cratonic regions. Most of the WB has an 80 to 120 km thick thermal lithosphere. This result is in agreement with continental reference geotherms (Pollack & Chapman, 1977), which predict the thermal LAB at 80–130 km depth for regions with surface heat flow of 60–70 mW/m² (Figure 3a) as observed in the WB.

In the EB, the LAB depth shows significant variations with <80 km in the north and >140 km in the south. However, the transition from thick to thin lithosphere does not follow the geological boundaries (Figure 4a). Our LAB model suggests that the westward extent of the region affected by the Pacific slab may not correspond to the topographic ramp and that the region instead may extend to inside the EB. We explore this hypothesis by calculating the density structure of the LM in the NCC. We further discuss our model of the lithosphere thermal thickness in section 6 together with the results for the lithosphere mantle density.

5. LM Density

5.1. Reference Model for Gravity Calculations

To calculate mantle lithosphere residual gravity and density anomalies, we use a 120 km thick lithosphere reference model with a crustal thickness of 38 km, which corresponds to the average crustal thickness in the NCC (Xia et al., 2017; Table 3).

We adopt a reference mean STP density $\rho_{M0} = 3.38 \text{ g/cm}^3$ of fertile upper mantle based on densities of mantle-derived Phanerozoic xenoliths (Djomani et al., 2001). Since the gravity signal is produced at in situ P-T conditions, we calculate reference in situ mean lithospheric and asthenospheric mantle densities (ρ_{LM} and ρ_{AM}) by introducing a temperature correction to the STP value ρ_{M0} :

$$\rho_{LM} = \rho_{M0} \times (1 - \alpha \times (T_{\text{Moho}} + T_{\text{LAB}})/2) \quad (5a)$$

$$\rho_{AM} = \rho_{M0} \times (1 - \alpha \times T_{\text{LAB}}) \quad (5b)$$

where the thermal expansion coefficient $\alpha = 3.5 \times 10^{-5} \text{ K}^{-1}$. The calculated Moho temperature T_{Moho} ranges from 450°C to 750°C (Figure 4b), which results in an in situ reference LM density of 3.23–3.26 g/cm³ for an in situ asthenospheric mantle density of 3.21 g/cm³.

Figure 4. (a) Thermal lithosphere thickness and (b) Moho temperature.

5.2. Gravity Calculations

We first calculate the residual LM gravity anomaly Δg_{rmg} (Figure 5) by removing the gravitational effect of the crust and the LAB from the free air gravity anomaly (Figure 1b). This is done by subtracting the gravitational effect of the crustal, Δg_{crust} , and LAB models, Δg_{LAB} , from the observed gravity Δg_{obs} and adding the gravity response of the reference model Δg_{ref} :

$$\Delta g_{rmg} = \Delta g_{obs} - \Delta g_{crust} - \Delta g_{LAB} + \Delta g_{ref} \quad (6)$$

Table 3

Reference Model Parameters for Gravity Calculations

	Depth (km)	In situ density (g/cm ³)
Crust (incl. sediments)	0–38	Variable (2.68–2.2.79) ^a
Lithospheric mantle	38–120	Variable (3.23–3.26) ^b
Asthenospheric mantle	>120	Constant, 3.21

^aDepends on crustal V_p structure, based on V_p -density conversion (Figure S1b). ^bDepends on Moho temperature.

We use the Bouguer plate approximation $\Delta g = 2\pi G h \rho$ for calculation of Δg_{crust} , Δg_{LAB} and Δg_{ref} , where Δg is the calculated gravity anomaly for a layer with density ρ and thickness h , and G is the gravitational constant. Since the calculation is made on a $1 \times 1^\circ$ grid, the gravitational effect of the neighboring cells for the expected density heterogeneity in the crust and LM is small (<20 mGal) and can be neglected. Furthermore, the crust in the study region has a smooth Moho without sharp changes and a

Figure 5. (a) Gravity anomaly from surface to the bottom of the sedimentary sequences, (b) gravity effect from the variation in the Moho depth, (c) LAB topography contribution to gravity anomaly, and (d) residual mantle gravity anomalies.

nearly uniform average crustal velocity and density (Xia et al., 2017), so that the difference between the Bouguer plate approximation and the full 3-D solution is also small (Shulgin & Artemieva, 2019). Due to the small density contrast across the LAB and its large depth, the 3-D gravitational effects from the LAB are insignificant.

5.3. Gravity Effect of the Crust and the LAB

The low-density sedimentary cover and the Moho topography provide the largest contributions to the gravity anomalies (Figures 5a and 5b). The gravity effect from the surface (including topography) to the base of the sediments varies from -120 mGal in the east to $+80$ mGal in the west, since the topography produces positive and sediments produce negative gravity anomalies (Figure 5a). For example, in the central parts of the EB, the low surface topography and the thick (3–5 km) low-density (<2.2 g/cm³) sedimentary cover produce a gravity anomaly of -120 to -40 mGal, while in the southern and northern parts of the NCC with a thin (0–2 km) sedimentary cover, the joint contribution of the sedimentary cover and the topography to the gravity field is -40 to -20 mGal. The WB with a relatively thick (up to 6 km) and dense (2.4 g/cm³) sedimentary cover and a high topography produces a gravity anomaly of $+40$ to $+80$ mGal.

The gravity effect of the Moho depth variation (Figure 5b) follows the pattern of the crustal thickness with a systematic reduction from the east ($+50$ mGal) to the west (-200 mGal, Figure 5b). The thin crust (thinner than the reference crustal thickness of 38 km) results in a positive gravity anomaly, and the EB with the Moho depth of 30–34 km has $+60$ to $+120$ mGal gravity anomaly caused by the variation in the Moho depth. Due to the thick crust (40–50 km), the effect of the Moho topography in the WB is -70 to -175 mGal locally.

The contribution of the LAB boundary to the gravity field is important and follows the overall pattern of the LAB topography (Figure 5c). The largest contributions between $+140$ and $+220$ mGal correspond to the

130–170 km thick lithosphere in the northern and SE parts of the TNCO and at the southwestern margin of the WB, while areas with the thermal LAB <120 km produce LAB gravity anomaly of +80 to +120 mGal.

5.4. RMGA

RMGA are caused by density variations below the Moho (Figure 5d). They range from −50 to +100 mGal in the areas with a thin (<120 km) lithosphere and a relatively thick low-density sedimentary cover. In regions with a thin (<2 km) sedimentary cover as in the southern part of the EB, the residual gravity anomalies are ~ −100 mGal. In the northern and southern parts of the TNCO with the thick thermal lithosphere (>140 km), the residual gravity anomalies range from −200 to −100 mGal.

5.5. Uncertainty Analysis

The depth distribution of density anomalies that produce RMGA is unknown, and we assume that they originate from the LM, that is, the layer between the Moho and the LAB (section 3.3). LM density calculated from RMGA provides a vertically averaged density value for the entire LM column. The uncertainty in the calculated mantle density comes from several factors (see also Herceg et al., 2016):

- i. the uncertainty in the LAB depth that we estimate to be ~20 km based on the agreement between the thermal model based on heat flow and regional Meso-Cenozoic xenoliths geotherms (sections 4.1–4.3). This uncertainty propagates twofold: through the RMGA values due to the effect of the LAB undulations (causes the uncertainty of ~50 mGal) and through the conversion of the RMGA values (Figure 5d) to LM density (Figure 6): for example, an RMGA anomaly of 50 mGal corresponds to a 0.01 g/cm³ difference in mantle density between the models with a 130 and 150 km thick lithosphere;
- ii. the uncertainty associated with calculation of the RMGA that mostly arises from the uncertainty in the crustal correction; this uncertainty cannot be directly assessed and we estimate that it does not exceed 20 mGal (which corresponds to realistic 0.05 g/cm³ uncertainty in average crustal density and 2 km uncertainty in the Moho depth); a 40 mGal RMGA uncertainty corresponds to a 0.01 g/cm³ uncertainty in mantle density for a 150 km thick lithosphere.

Another possible source of uncertainty is associated with the choice of the reference model (Table 4). The choice of different values in the reference model will only produce a systematic shift in the calculated LM density. Our choice of reference model values is justified by adopting regionally representative values (Table 5) for the crustal thickness (Xia et al., 2017) and the LAB depth (Figure 4a) and by the fact that the calculated LM density values fall within the expected range predicted by geochemistry analysis (Djomani et al., 2001). Uncertainties related to the crustal correction, including the V_p to density conversion and the crustal model uncertainties (the thickness and V_p in each crustal layer), have been analysed in detail previously (Herceg et al., 2016). The results show that the largest uncertainties are related to the V_p to density conversion, which leads to LM density uncertainties significantly smaller than 0.02 g/cm³.

We also test our results by calculating isostatic topography predicted from our models for thermal LAB and LM density (Figure S3) and for the crustal density structure used in the gravity calculations and based on the NCcrust model (Xia et al., 2017). The discrepancy between the predicted and observed topography in most of the study area is 200–400 m (with the regional average close to 0). Although this discrepancy may transform to a 0.02 g/cm³ error in our model of the LM density, the real misfit is smaller since a basic isostasy model does not account for lithosphere flexure, for example, in relation to the India-Eurasia collision and significant parts of the NCC are not in the isostatic equilibrium.

5.6. LM Density

The overall pattern shows noncratonic LM densities in most of the North China (NC) with low, cratonic-type in situ values (3.22–3.24 g/cm³) in the southern part of the EB, northern and southern parts of the TNCO, and high in situ (>3.25 g/cm³) background LM densities elsewhere, overprinted by numerous small anomalies with an amplitude of 0.01–0.02 g/cm³ (Figure 6a). The largest high-density anomaly is at the NW part of the Yangtze Craton; it may indicate the presence of some eclogitic material in the LM, which may be the cause of the Sichuan basin subsidence.

In situ densities cannot be compared with laboratory and xenolith data on rock densities; therefore, we apply a temperature correction to convert LM density to laboratory, STP conditions (Equation 5a). The overall

Figure 6. Lithospheric mantle density at (a) in situ condition and (b) STP condition.

pattern is the same as for in situ density with variations from 3.32 to 3.40 g/cm³, which shows a highly heterogeneous composition of the LM of the NCC (Figure 6b).

High-density LM (>3.37 g/cm³) is locally present along the Pacific coast of the EB. The WB has a high-density LM with values of >3.36 g/cm³ comparable to the density of Phanerozoic fertile mantle, interlaced with blocks of very high density values (>3.39 g/cm³). High density is also observed in the central and

Table 4
Sensitivity Analysis for Lithosphere Mantle Density Calculations

For LM density model	Reference model	Alternative value in reference model	Change to reference model	Resultant change in LM density	
Reference crustal thickness (km)	38	34	−4	−12.50%	−0.02 kg/m ³
Reference thickness of Lithospheric mantle (km)	82	112	+30	+62.70%	−0.02 kg/m ³

Table 5
Summary of Lithosphere Structure in the North China Craton

	The North China Craton		Eastern Block		Trans-North China Orogen		Western Block	
	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.
Moho depth (km) ^a	39	6	33	1	37	2	46	3
Average V_p of bulk crust (km/s) ^a	6.3	0.04	6.33	0.02	6.33	0.04	6.37	0.04
Bulk crustal density (g/cm ³) ^a	2.8	0.02	2.74	0.02	2.75	0.02	2.76	0.01
Surface heat flow (mW/m ²) ^b	62	23	63	12	57	15	64	10
Moho temperature	644	145	614	105	619	138	793	140
Lithosphere thermal thickness (km)	149	33	148	32	165	37	135	27
Density of lithospheric mantle								
In situ	3.25	0.01	3.24	0.01	3.23	0.01	3.25	0.01
STP	3.34	0.02	3.34	0.02	3.33	0.02	3.36	0.02

^aXia et al. (2017). ^bG.-Z. Jiang et al. (2019).

the coastal parts of the EB. A low-density LM (<3.34 g/cm³) typical of depleted, Archean LM is present beneath the southern and northern parts of the TNCO, the southern EB.

6. Origin of the LM Heterogeneity Beneath the NCC

The highly heterogeneous structure of the LM density in the NCC suggests that most of the cratonic LM has been reworked to attain density typical of fertile LM, whereas only a minor part of the craton includes a preserved low-density depleted LM typical of Archean cratons. Locally, extremely high LM density with values of >3.39 g/cm³ may mark the presence of eclogite in the LM. In the following discussion of the possible geodynamic origin of the LM compositional heterogeneity; all density values refer to STP condition.

6.1. EB

With a high surface heat flow of 50 to 80 mW/m² (Figure 2) and thin crust, the thermal lithosphere of the EB is 80–150 km thick, which is significantly thinner than typical Archean cratonic lithosphere (>200 km) which has low surface heat flow (<50 mW/m²). Our model for the LAB depth implies that the southern and northern parts of the EB presently includes ~120 and ~50 km of LM, respectively, which means that they may have lost >40–70% of the original Archean LM. A significant part of this loss was possibly associated with the Pacific subduction, which means that subduction-related magmatism should have modified the composition of the preserved part of the LM.

Indeed, the EB has a heterogeneous LM density structure with STP values ranging from 3.35 g/cm³ in the southwest to 3.38–3.39 g/cm³ along the eastern margin and in the western central part with 3.40 g/cm³ locally (Figure 6 and Table 4). Therefore, in most places the LM has significantly higher density than typical Archean cratonic LM with densities of 3.30–3.34 g/cm³ (Artemieva et al., 2019; Artemieva & Vinnik, 2016; Djomani et al., 2001). Low LM density (~3.33–3.35 g/cm³) typical of depleted Archean lithosphere mantle is preserved only in the SW part of the EB where surface heat flow is low (<55 mW/m²), and the lithosphere is relatively thick. Regions with thin lithosphere have high-density LM (>3.37 g/cm³) typical of fertile Phanerozoic LM; these regions also usually have a thick sedimentary cover, suggesting that densification of the LM may have contributed to compositional subsidence. An extreme LM density locally in the western central part of the EB (~3.40 g/cm³), slightly higher than expected for primitive mantle (3.39 g/cm³), may indicate the presence of eclogitic material. Alternatively, it may have a composition similar to high-T lherzolite in the bottom part of the Kaapvaal LM, which has density of 3.38–3.44 g/cm³ (James et al., 2004).

Our results are in overall agreement with seismic models for the LAB depth and xenolith data on mantle densities. The presence of a <120 km thick lithosphere in the central part of the EB, which is significantly thinner than in most stable cratons, is reported in numerous seismic studies; but the lithosphere thickness in the southern part of the EB may be controversial. Shear wave tomography indicates that the lithosphere is only 70–100 km thick in most of the EB (Y.-H. Li et al., 2013). *S* receiver functions results suggest that the lithosphere is 60 km thick beneath the Tan-Lu Fault zone (Chen et al., 2006) and 100–120 km thick in other parts of the EB (Chen et al., 2008) and therefore may underestimate the lithosphere thickness by ~20 km as

Figure 7. Sketch map showing our interpretation of lithospheric mantle composition based on density values (Figure 6). Triangles, squares, and stars mark locations of Cenozoic, Mesozoic, and Paleozoic kimberlites, correspondingly. Numbers close to symbols mark the average olivine-Mg # in xenoliths and color of symbols shows average olivine-Mg # in xenoliths (Chu et al., 2009; Dai et al., 2018; Lin et al., 2019; J.-G. Liu et al., 2011; Y. Xiao et al., 2010; Ying et al., 2006; Zheng et al., 2007, and references therein).

compared to our thermal LAB model. A more detailed *S* receiver functions study with a dense station spacing indicates a nearly uniform LAB depth of 60 to 80 km in the entire EB (X.-C. Wang et al., 2016) although these estimates may be biased by detecting the midlithospheric discontinuity rather than the LAB (Karato et al., 2015; Selway et al., 2015; Thybo, 2006; Thybo & Perchuc, 1997).

Mantle-derived xenoliths carried by kimberlites and basalts at the margins of the EB (for locations see Figure 2) indicate that LM largely lost its cratonic signature between Paleozoic and the Mesozoic time (Griffin et al., 1998; Zheng et al., 2007). In particular, most xenoliths with Mesozoic to Cenozoic emplacement ages have an olivine-Mg# of 89–90 (Griffin et al., 1998, and references therein) typical of fertile mantle (Figure 7) and not observed in depleted Archean mantle (Carlson et al., 2005). In agreement with xenolith data, our calculated mean LM density at the eastern margin of the EB is typically 3.37–3.38 g/cm³, which corresponds to fertile mantle (Djomani et al., 2001).

Xenoliths from the southern part of the EB with low surface heat flow (<50 mW/m²) have mostly olivine-Mg# > 92 (Zheng et al., 2001) in the region where our model predicts the presence of a low-density depleted mantle. Only few samples from the SE part of the EB have low olivine-Mg# (89–90) (Y.-J. Tang et al., 2008), which indicates fertilization of cratonic LM, in agreement with our density model which also shows a high LM density toward the coast (Figures 6 and 7). A complex heterogeneous pattern of LM composition recorded by xenolith data beneath the southwestern margin of the EB has been interpreted as a mixture of Archean cratonic mantle with a minor component of newly added fertile accreted lithosphere (Zheng et al., 2001).

The lithosphere structure beneath the north central part of EB with high LM density (>3.37 g/cm³) and thin (80–120 km) lithosphere can be explained by refertilization caused by intruded basaltic material, possibly associated with the Mesozoic subduction. In contrast, the southern part of the EB may still preserve depleted cratonic LM with minor chemical modification, as indicated by low LM density (<3.35 g/cm³), low surface heat flow (<50 mW/m²), and limited xenolith evidence (Mg# > 92).

Importantly, we do not observe any correlation between the LM densities and the topographic ramp, and therefore, our results do not support models with the western edge of the Pacific slab located at the western edge of the EB. Partial destruction of the lithospheric keel with significant lithosphere thinning and strong metasomatic reworking of the preserved upper part of the LM is largely limited to the coastal corridor, with a

high-density anomaly extending westward to the TNCO boundary only in the central part of the EB. We note that in the *S* receiver function model, the LAB depth also does not follow the topographic ramp (Y.-Y. Zhang et al., 2019). We discuss in the next section a possible origin of the low-density block in the southern part of the NCC, which we ascribe to the Mesozoic Yangtze Craton subduction and conclude that an interplay of two subduction systems of different ages shaped the area that was subject to decratonization by the later, Mesozoic Pacific subduction.

6.2. TNCO

The TNCO, a Paleoproterozoic (1.95–1.85 Ga) belt formed during the amalgamation of the EB and WB, has LM dominated by a depleted low-density cratonic composition, particularly preserved in its northern and southern parts. This pattern is similar to observations from the Siberian Craton, where a highly depleted LM has been reported for the Proterozoic Akitkan mobile belt formed by the collision of two Archean blocks (Cherepanova & Artemieva, 2015). The TNCO LM density structure shows a high contrast with values of 3.32–3.33 g/cm³ in the southern and northern parts cut by a block with high-density heterogeneity (3.33–3.38 g/cm³) in the central part of the TNCO (Figures 6 and 7).

Our results, supported by geochemical studies of mantle-derived xenoliths, indicate that the cratonic LM under the central part of the TNCO experienced strong reworking and has characteristics of fertile Phanerozoic LM, while the northern and southern parts still preserve a largely unmodified depleted cratonic lithosphere (Figures 6 and 7). The boundary between the depleted northern and reworked central block approximately corresponds to the southern extent of the outcropping Archean basement in the TNCO. The boundary between the reworked central and depleted southern block does not seem to be reflected in surface geology but corresponds to the area with a thin lithosphere, which has been subject to Cenozoic rift-ing (Figure 7). It is therefore likely that partial destruction of the cratonic lithosphere in the central TNCO was, at least in part, caused by lithosphere extension and the associated mafic magmatism, possibly associated with roll-back of the Pacific slab.

A striking feature is the presence of a low-density LM beneath the southern part of the EB and the TNCO; this low-density block extends northward by ~500 km from the Qinling-Dabie orogen, which was formed by the Mesozoic collision of the NCC and the Yangtze Cratons. It has long been recognized from large differences in isotopic composition of Mesozoic alkaline intrusive complexes that the northward subduction and the subsequent subduction of the Yangtze Craton beneath the NCC should have affected the composition of the NCC lithosphere over a distance of ~450 km northward from the Qinling-Dabie orogen (H. F. Zhang et al., 2005), which approximately corresponds to the northward extent of the low-density LM anomaly (Figure 6). Our results are supported by the composition of peridotites from mantle-derived xenoliths in Cenozoic basalts in the southern part of the TNCO, which indicate the presence of a thick lithosphere as also imaged seismically (Y.-Y. Zhang et al., 2019) and depleted LM with mainly high olivine-Mg # of 92.0 ± 0.9 (J.-G. Liu et al., 2011; Zheng et al., 2005; Figure 7). The high olivine-Mg# (>92) harzburgites are similar to old refractory LM and support our conclusion that the Archean LM has not been significantly modified or replaced in the southern part of the NCC. We speculate that flat subduction of the Yangtze Craton precluded subduction-related magmatism and decratonization of the southern NCC and instead preserved a slightly modified cratonic LM with a deep LAB, which may possibly correspond to lithosphere doubling. Our results for the thermal LAB (140–180 km) are in agreement with an *S* receiver function study, which shows that the southern part of the TNCO lithosphere is >160 km thick (Y.-Y. Zhang et al., 2019), and with surface wave tomography which indicates a LAB depth of 130–150 km (Huang et al., 2009).

In the northern part of the TNCO, Cenozoic xenoliths have an average olivine-Mg # of 90–91, typical of Proterozoic-to-Phanerozoic mantle, and whole-rock T_{RD} age of ~1.8 Ga suggests either Paleoproterozoic modification/reworking of the original Archean LM (Griffin et al., 1998) or a Paleoproterozoic age (J.-G. Liu et al., 2011; Y.-G. Xu et al., 2008). For this region, our results show the presence of a low-density LM (~3.31 g/cm³) and are in contrast with xenolith Mg# (Figure 7), which we explain by local lithosphere modification at the xenolith locations.

6.3. WB

Globally, xenolith studies indicate that STP density of the Archean, Proterozoic, and Phanerozoic LM is 3.31 ± 0.02 , 3.35 ± 0.02 , and 3.38 ± 0.02 g/cm³, respectively (Djomani et al., 2001). Our results constrain

LM density of $>3.35 \text{ g/cm}^3$ in most of the WB and therefore indicate that the WB LM has lost its Archean cratonic characteristics (Figure 6 and 7 and Table 4). Our results suggest a contrast in the lithosphere structure between different parts of the WB, which may explain differences in subsidence history.

Recent borehole heat flow measurements in the WB yield $45\text{--}75 \text{ mW/m}^2$ with an average value of $64 \pm 8 \text{ mW/m}^2$ (Figure 2; He, 2015; G.-Z. Jiang et al., 2019), which is higher than in Archean and Proterozoic regions with heat flow typically $<50 \text{ mW/m}^2$ (Artemieva & Mooney, 2001). Our thermal model predicts that the lithosphere is 110–130 km thick only in the southern part of the WB and is on average 100–110 km thick in most of the WB (Figure 4a). In support of our results, regional surface wave tomography (Guo & Chen, 2017; Huang et al., 2009; Y.-C. Tang et al., 2013; X.-C. Wang et al., 2017) and a long-range refraction seismic profile which reach the LAB suggest that most of the WB has a relatively shallow LAB (S.-L. Li, Lai, et al., 2011; S.-J. Wang et al., 2014).

Dense LM, high surface heat flow, and thin lithosphere are consistent with a heavily reworked LM beneath the WB. Mantle xenoliths are nearly absent in the WB, and our conclusion is supported by a low olivine-Mg # of 89–90 (Figure 7) in mantle xenoliths in Cenozoic to early Mesozoic basalts from the northern margin which indicate a fertile LM composition (Dai et al., 2018; J.-G. Liu et al., 2011). In the north, LM reworking may possibly have taken place during the closure of the Paleo-Asian Ocean and the formation of the Central Asian Orogen Belt (Y. Liu et al., 2010). At the southwestern margin of the WB, close to the edge of the Tibetan Plateau, LM reworking is possibly related to Cenozoic tectonics (Figure 7).

6.4. Mantle Reworking in the NCC

The olivine Mg # measured in xenolith samples from LM of the NCC varies between 89.1 (reported for most of the NCC and indicative of fertile mantle; Zheng et al., 2007) and 92.0 in the TNCO (J.-G. Liu et al., 2011; Zheng et al., 2005) indicative of (almost) unmodified Archean depleted mantle (Griffin et al., 2003). Except for the northern TNCO, these values agree with in situ and STP density determined by our gravity calculations (Figure 6 and 7 and Table 4). By combining our models for the thermal LAB and LM density with xenolith data, we propose a model of the LM reworking and the NCC decratonization as illustrated by the sketch map in Figure 7.

In our interpretation, only minor parts of the NCC LM retain Archean characteristics, whereas most of the cratonic lithosphere has been partly or fully reworked. Therefore, our model is in contrast with many other models of NCC reworking, which generally include decratonization of only the EB or its easternmost parts in response to Mesozoic subduction of the Pacific plate (Zhu et al., 2012). Our interpretation includes a series of tectonomagmatic processes of various ages that contributed to the reworking of the cratonic LM from the east, west, south, and north.

1. In the northwest, the reworking of cratonic LM may be related to Paleozoic collisional tectonics that formed the Central Asia Orogenic Belt (Y. Liu et al., 2010).
2. In the east, the decratonization may be caused by processes related to Mesozoic subduction of the Pacific Plate, as often proposed (Wu et al., 2018; Zhu et al., 2012). However, the westward extent of the reworked lithosphere does not correspond to the topographic ramp and Bouguer gravity gradient zone (which is also related to the topography ramp) but shows a complex pattern, possibly caused by an interplay between the late Mesozoic Pacific and the Mesozoic Yangtze subduction systems.
3. In the south, the Mesozoic subduction of the Yangtze Craton, in case of flat subduction, may have preserved a depleted composition of the cratonic mantle with only minor metasomatic reworking and preserved a thick lithosphere, possibly caused by lithosphere doubling.
4. In the center (central TNCO), strong reworking of the LM may be caused by Cenozoic rifting, possibly associated with roll-back of the Pacific slab. Basaltic magmatism associated with lithosphere extension caused lithosphere thermal erosion, mantle metasomatism and LM densification, which may have led to partial delamination of the lower portions of the LM.
5. In the southwest, the causes for the LM reworking are speculative and we propose that lithosphere deformation caused by far field processes arising from the Meso-Cenozoic collision between the Indian and Eurasian plates may have affected the mantle lithosphere (Chang et al., 2009).

We conclude that the reworking of the NCC has involved a series of diverse processes that operated through most of the Phanerozoic and included the Mesozoic collisional tectonics in the northwest and the Yangtze

subduction in the south, the late Mesozoic Pacific subduction in the east, the Cenozoic rifting in the center, and the remote response to the Indian-Eurasian collision in the southwest.

7. Conclusions

Based on the EGM2008 gravity model (Pavlis et al., 2012), the NCcrust seismic model for regional crustal structure (Xia et al., 2017), and a new regional heat flow database (G.-Z. Jiang et al., 2019), we calculated lithosphere thermal thickness and LM density of the NCC. On this basis, we conclude the following:

1. The NCC has been subject to a series of major Phanerozoic tectonomagmatic processes, which caused decratonization of most of the craton. The coastal and northern parts of the EB, the central segment of the TNCO and almost the entire WB have high surface heat flow and a thin thermal lithosphere (100–130 km thick), which is atypical for cratonic regions. Low cratonic heat flow ($<45 \text{ mW/m}^2$) and thick thermal lithosphere ($>160 \text{ km}$) are largely restricted to the northern part of the TNCO and the southern EB-TNCO region. Lithosphere geotherms are consistent with regional xenoliths P-T arrays, which we calculated for regional data.
2. The density structure of the subcontinental lithosphere beneath the NCC is heterogeneous. The in situ LM density varies from 3.22 to 3.28 g/cm^3 and STP density from 3.32 to 3.38 g/cm^3 (locally up to 3.40 g/cm^3). We interpret the typical density of $>3.36 \text{ g/cm}^3$ in most of the NCC by significant compositional modification of the cratonic LM.
3. The low-density ($<3.34 \text{ g/cm}^3$), thick LM in the southern part of the NCC retains characteristics of depleted Archean LM; a lithosphere block with similar cratonic characteristics is also present in the northern part of the TNCO. The presence of a thick, low-density mantle in the southern NCC, supported by high Mg# in Mesozoic mantle xenoliths, suggests flat subduction of the Yangtze Craton beneath the NCC in Mesozoic, which hampered decratonization by limiting slab-related magmatism.
4. Most of the WB, the central part of the TNCO-EB, and the coastal zone of the EB have a high-density LM with STP values ($>3.37 \text{ g/cm}^3$) typical of fertile LM, which suggests that the original Archean lithosphere has been lost through replacement and metasomatic modification. Local very high density ($\sim 3.40 \text{ g/cm}^3$) anomalies beneath the WB and the west central EB may mark the presence of eclogites in the LM, which may have contributed to compositional subsidence and basin formation. The western edge of the area affected by Pacific subduction does not correlate with the topographic and gravity anomalies and shows a complex pattern, possibly caused by an interplay between the Pacific and the Yangtze subduction systems.
5. Our results indicate that most of the NCC lithosphere has been reworked since the Archean. The heterogeneity in LM density does not follow any known geological or topographic boundaries. The Mesozoic Pacific subduction can only explain significant reworking of the LM at the eastern part of the NCC. The Mesozoic flat subduction of the Yangtze Craton has affected the lithosphere structure of the southern NCC over $\sim 500 \text{ km}$ north of the Qinling-Dabie orogeny, where it has contributed to maintaining a thick lithosphere. The Paleozoic subduction/collision associated with the Central Asia Orogenic Belt may have played an important role in LM modification/reworking in the northern parts of the NCC. High-density LM in the southwestern part of the WB may reflect a remote response to the Indian-Eurasian collision, and a block with a thin ($<100 \text{ km}$) and very dense LM in the central TCNO may result from Cenozoic rifting, possibly caused by roll-back of the Pacific subduction system.

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Data Availability Statement

Data used in calculations are based on published data and are available as referenced in Xia et al. (2017), G.-Z. Jiang et al. (2019), Y.-Y. Zhang et al. (2019) Chen (2010), and Pavlis et al. (2012). The updated NCcrust model (Xia et al., 2017) is available at Zenodo (<https://zenodo.org/record/3546583#.XdNbFiJfhjU>).

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