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Key Points:

- We jointly analyze heat flow data, tectonic subsidence, and crustal structure
- Lower crustal flow explains anomalous postrift subsidence and extension discrepancy
- The amount of postrift thinned lower crust is positively correlated with water depth

Correspondence to:

J. Zhang and L. Yu,
zhangjian@ucas.edu.cn;
yulei@lyu.edu.cn

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Quantifying Postrift Lower Crustal Flow in the Northern Margin of the South China Sea

Miao Dong^{1,2} , Jian Zhang¹, Sascha Brune^{3,4} , Shiguo Wu^{1,2}, Gui Fang¹, and Lei Yu⁵

¹University of Chinese Academy of Sciences, Beijing, China, ²Laboratory of Marine Geophysics and Georesources, Institute of Deep-Sea Science and Engineering, Chinese Academy of Sciences, Sanya, China, ³Geodynamic Modelling Section, GFZ German Research Centre for Geosciences, Potsdam, Germany, ⁴Institute of Geosciences, University of Potsdam, Potsdam, Germany, ⁵Shandong Provincial Key Laboratory of Water and Soil Conservation and Environmental Protection, College of Resources and Environment, Linyi University, Linyi, China

Abstract Postrift subsidence of sedimentary basins in the northern margin of the South China Sea exceeds more than 2,000 m, which points toward anomalous postrift crustal deformation. Previous studies have proposed lower crustal flow to explain this observation; however, this hypothesis has never been confirmed quantitatively. Here, we calculate the initial crustal structure and thermal lithospheric thickness of the northern margin of the South China Sea on the basis of recently measured heat flow data, tectonic subsidence curves, and present-day crustal structure. Crustal thinning processes during rifting and reduced thickness of the lower crust in the postrift are also calculated by the strain rate inversion method and thermal isostasy. Our results show that the initial (Early Cenozoic) crustal thickness of the northern margin of the South China Sea varies between 27.7 and 36.3 km, and the average proportion of the lower crust is 67%. The initial thermal lithospheric thicknesses decrease from ~118 to ~81 km from the shelf to the sea, which indicates that the offshore margin has high temperatures and low strength and is more easily stretched and deformed. The average stretching factor the margin during rifting is 1.24. The lower crust reduces in thickness during the postrift phase by values between 1 and 11 km, which increases gradually from the shelf to deep water. We suggest that the thicker weak crust and hot lithospheric structure therefore make positive contribution to the lower crustal flow, the direction of which is from the oceanic basin to the shelf.

1. Introduction

Rifted continental margins provide a unique record of the final stage of continental rifting and therefore serve as a window to understand the processes of continental breakup (Le Pourhiet et al., 2018; Reston & Phipps Morgan, 2004; White et al., 2003; Whitmarsh et al., 2001). The evolution and formation of rifted continental margins are accompanied by thinning and subsidence of the lithosphere, the development of rift basins as well as upwelling of hot asthenosphere (Clerc et al., 2018; Huisman & Beaumont, 2011; Peron-Pinvidic et al., 2013; Ranero & Perez-Gussinye, 2010). Although deformation and temperature changes of the crust or lithosphere in the Earth's interior cannot be directly observed, isostasy provides information regarding deep structural changes through surface observations (Hasterok & Chapman, 2007).

Total tectonic subsidence of a rift basin can be decomposed in a synrift stage and a postrift (thermal subsidence) stage. During the synrift stage, the crust undergoes rapid subsidence resulting from extension and thinning of the lithosphere. When entering the postrift stage, the lithospheric mantle begins to cool and densify, which results in comparatively slow subsidence. This behavior leads to characteristic shapes of tectonic subsidence curve: a steep slope in the synrift stage and a gentle slope in the postrift stage (McKenzie, 1978).

The instantaneous stretching model is a classic basin subsidence model (McKenzie, 1978) that has been the basis of various subsequent models, such as the stretching model (Jarvis & McKenzie, 1980), magmatic intrusion stretching model (Royden & Keen, 1980), flexural-cantilever model (Kusznir et al., 1991), and depth-dependent stretching model (Huisman & Beaumont, 2011). These models mainly focus on basin development during the synrift stage, whereas the postrift stage has mostly been studied using the

Figure 1. Regional topographic map showing major tectonic units and main reflection profiles. Circles indicate locations of sites IODP/ODP drilling sites. The white dashed line marks the continent-ocean boundary, and the pink dashed lines show the major magnetic anomalies (Li et al., 2014). Black gray line with filled triangles represent subduction zone. XI, Xisha Islands; NWSB, Northwest subbasin; ESB, East subbasin.

thermal subsidence model proposed by McKenzie (1978). However, the observed postrift subsidence of many rift basins worldwide is substantially larger than that predicted by the McKenzie model. Some studies have reported anomalous postrift subsidence and analyzed potential explanations. For example, kilometer-scale anomalous postrift subsidence was discovered along the North Atlantic margin that was considered to be the synthesis of all postrift activities (Ceramicola et al., 2005). The “excess” subsidence and high deposition rate (100 m/Ma) in the early postrift stage at the south Gabon margin (South Atlantic) were assumed to result from high thermal anomalies in the mantle (Dupré et al., 2007). In the Pattani and Malay basins of the Gulf of Thailand, the postrift sediment thickness is about 8 km, while the entire sediment thickness is 12 km, which far exceeds theoretical thermal subsidence estimates and is thought to result from lower crustal flow that significantly increased the deposition rate. The high deposition rate is further related to an increase of sediment supply, which was likely caused by the enhancement of East Asian monsoons (Morley & Westaway, 2006).

As the largest marginal sea in the western Pacific, the South China Sea is an ideal laboratory for studying the development and evolution of rift basins and rifted margin (Figure 1). In recent years, the implementation of a number of scientific projects, including the Integrated Ocean Drilling Program and South China Sea Deep Plan, has provided a range of useful data for studying the evolution of this rifted margin (Larsen et al., 2018; Li et al., 2014). Anomalous postrift subsidence is also observed in basins such as the Yinggehai, Qiongdongnan, and Pearl River Mouth basins along the northern margin of the South China Sea (Clift & Sun, 2006; Clift et al., 2015; Clift, 2015; Shi, Jiang, et al., 2017; Xie et al., 2006). After the Late Miocene, some parts of this area (e.g., Xisha) experienced large-scale inundation of carbonate platforms (Wu et al., 2016). Global sea levels at this time continued to decline, which confirms that the rapid subsidence event occurred on the northern margin of the South China Sea during the postrift stage.

Here we calculate the initial crustal thickness by the thermal isostasy principle and invert the lithospheric strain rate according to the subsidence process in the rifting stage. Our analysis is based on very recent studies on crustal structure (Bai et al., 2019) and a new compilation of heat flow measurements (Shi, Yu, et al., 2017). We then calculate the temperature change of the lithosphere during the postrift stage and obtain the theoretical subsidence curve. According to the thermal isostasy principle and lithospheric temperature structure determined by the forward calculation, the thickness of the lower crust is adjusted to make the calculated subsidence curve coincide with the actual subsidence curve derived from the backstripping analysis. Finally, we quantify the thickness and strain rate of the thinned lower crust during the postrift period.

2. Geological Setting

The northern margin of the South China Sea has been affected by plate tectonic movements of the Indochina Plate, Eurasian Plate, Pacific Plate, and Philippine Plate. Before the Late Cretaceous, the East Asian Mesozoic volcanic belt was present in the northwest of the South China Sea and the proto-South China Sea in the southeastern part. Influenced by NW subduction of the paleo-Pacific Plate, the East Asian continental margin, which contains the continental margin of the South China Sea, is strongly sheared and squeezed. Compressive stress in the continental margin of the South China Sea relaxed during the Late Cretaceous to Middle Eocene owing to the retreat of the paleo-Pacific Plate and southward subduction of the proto-South China Sea. The asthenosphere then uplifted and lithospheric extension turned the continent into a rifting margin along the NW-SE (Hall, 2002). During the Middle Eocene to Late Oligocene, the South China Sea experienced continental margin breakup and seafloor spreading. The crust in the northern margin of the South China Sea continued to thin and the scale of the basin expanded (Franke et al., 2014; Li et al., 2014; Taylor & Hayes, 1983). From west to east, there are many oil-rich rift basins on the northern margin of the South China Sea (e.g., Yinggehai Basin, Beibu-Gulf Basin, Qiongdongnan Basin, Pearl River Mouth Basin, and Taixinan Basin). Among these, the Yinggehai Basin is not only affected by rifting but also controlled by the strike-slip fault system on the west side of the basin (Clift & Sun, 2006). The other basins are all affected by the continental rift and trend NE.

During the Late Cretaceous or Paleocene to Eocene, the basins were in the synrift stage. During this period, the lithosphere in the northern margin of the South China Sea was stretched and the basins formed. The basins subsequently entered a thermal subsidence period after the Oligocene (Shi et al., 2005). Many basins, including the Yinggehai, Qiongdongnan, and Pearl River Mouth basins, experienced rapid subsidence in the postrift stage (Dong et al., 2008; Xie et al., 2006). Since the Late Miocene, the center of the postrift rapid subsidence has been concentrated in deep water areas, which are located in the central depression zone of the Qiongdongnan, Pearl River Mouth, and Taixinan basins.

Although anomalous postrift subsidence exists in the northern margin of the South China Sea, the total amount of subsidence calculated by the present-day crustal thickness is approximately equal to that from backstripping analysis (Shi, Jiang, et al., 2017). According the principle of thermal isostasy, this indicates that basement subsidence in the northern margin of the South China Sea is mainly caused by crustal thinning. The seismic profile shows that the number of fragile faults and their growth rate decreased significantly after ~23 Ma. This indicates that the upper crust did not stretch and thin thereafter (Zhao et al., 2018). Therefore, lower crustal flow was proposed to explain the rapid postrift subsidence in the margin of the South China Sea. But previous studies only focused on a single basin and did not define the amount of the lower crustal flow (Clift et al., 2015; Zhao et al., 2018).

Influenced by tectonic activities and continental margin evolution, the rift basins in the northern margin of the South China Sea are rich in sediments (3–17 km), especially in the Yinggehai Basin, the west of the Qiongdongnan Basin, and the Baiyun Sag in the Pearl River Mouth Basin (Figure 2). The maximum thickness of the Cenozoic sediment is more than 10 km (Xie et al., 2006). The distribution of the synrift sedimentary and postrift sedimentary thicknesses are different in each basin. The thickness of the synrift sediment in the Beibu Gulf Basin is much greater than that of the postrift sediment. In contrast, the thickness of the postrift sediment is substantially larger than that of the synrift sediment in the Yinggehai Basin. The synrift sediment thickness is approximately equal to the postrift sediment thickness in the Qiongdongnan and Pearl River Mouth basins.

3. Methods and Data

3.1. Backstripping Analysis and Heat Flow Data

Backstripping analysis is an important method of restoring subsidence history based on the current stratigraphic division, paleowater depth, and sea level correction. A stratigraphic decompaction and load balancing correction are performed to obtain the burial depth of the basement in each period (Sclater & Christie, 1980). We analyzed multiple sections through the northern margin of the South China Sea, as shown in Figure 1. To obtain the subsidence process that fully reflects the deep variation, we used the 1-D unloaded subsidence formula (Sclater & Christie, 1980; Zhao et al., 2018):

Figure 2. Depth section of Line 2 and Line 5 with points selected for subsidence analyses (see locations in Figure 1).

$$UTS = ST \frac{\rho_m - \rho_s}{\rho_m} + PWD \frac{\rho_m - \rho_w}{\rho_m} - \Delta BL \quad (1)$$

where UTS is the unloaded tectonic subsidence, ρ_m , ρ_s , and ρ_w are the average density of the asthenospheric mantle ($3,185 \text{ kg/m}^3$), sedimentary layer ($2,500 \text{ kg/m}^3$), and seawater ($1,030 \text{ kg/m}^3$), respectively, ST is the decompacted sediment thickness, PWD is paleowater depth, and BL is the base level change. The method of Sclater and Christie (1980) is used to calculate the decompacted sediment thickness, which is determined from the sediment lithology and relationship between porosity and depth. The porosity-depth relation curves of sandstone and mudstone differ substantially in different areas. To ensure accuracy, the porosity-depth curve used in this study is therefore combined with core data of Integrated Ocean Drilling Program expeditions and the fitting results of some deep oil wells in the northern South China Sea (Huang & Wang, 2006). The base level and paleowater depth have a significant effect on the results. Thus, according to the location of each section, different values of the PWD and BL are adopted for the calculations. Data from the Qiongdongnan Basin (Zhao et al., 2018) are used for sections on the western side of the continental margin, and Pearl River Mouth Basin data (Dong et al., 2008) are used for sections on the eastern side of the continental margin.

After more than 30 years of heat flow surveys, more than 450 heat flow points have been obtained in the northern margin of the South China Sea (Figure 3a). The data are densely distributed especially near the Qiongdongnan and Pearl River Mouth basins. In this study, we use newly published heat flow data from Shi, Yu, et al. (2017) that were obtained from borehole data analysis and seafloor heat flow probe measurements. The former are mainly distributed in areas with a water depth of less than 300 m, and the latter are mainly located in sea areas with a water depth of more than 1,000 m.

3.2. Thermal Isostasy

Isostasy theory defines the relationship between topographic alteration and changes in the thickness and density of the crust and lithosphere (Airy, 1855), which has been successfully applied to studies of basin and continental marginal evolution (McKenzie, 1978). Assuming that the continental margin is always in equilibrium, the mass above the compensation surface is always equal to that in the initial condition, either during the stretching phase or the stage of lithospheric cooling after extension (Figure 4). Therefore, the unloaded tectonic subsidence can be expressed by the following formula at any time (Hasterok & Chapman, 2007; Zhao et al., 2018):

Figure 3. (a) Heat flow measurements and heat flow isolines of the northern margin of the South China Sea (Shi, Yu, et al., 2017). The black points are heat flow measurements. The crustal thickness (b), upper crustal thickness (c), and lower crustal thickness (d) of the northern margin of the South China Sea (Bai et al., 2019).

$$\begin{aligned}
 L(0) &= \rho_{uc0} \cdot g \int_0^{UC_i} [1 - \alpha T(z, t_0)] dz + \rho_{lc0} \cdot g \int_{UC_i}^{LC_i} [1 - \alpha T(z, t_0)] dz + \rho_{m0} \cdot g \int_{LC_i}^a [1 - \alpha T(z, t_0)] dz \\
 L(t) &= \rho_{uc0} \cdot g \int_0^{UC_t} [1 - \alpha T(z, t)] dz + \rho_{lc0} \cdot g \int_{UC_t}^{LC_t} [1 - \alpha T(z, t)] dz + \rho_{m0} \cdot g \int_{LC_t}^a [1 - \alpha T(z, t)] dz \\
 S(t) &= \frac{L(0) - L(t)}{\rho_{m0} \cdot g \cdot (1 - \alpha T_m)},
 \end{aligned} \quad (2)$$

where $L(0)$ represents the lithospheric load in the initial state before the rifting, $L(t)$ represents the lithospheric load at time t , ρ_{uc0} , ρ_{lc0} , and ρ_{m0} , represent the density of the upper crust, lower crust, and mantle at 0°C , respectively, UC and LC represent the depth of the upper crust and lower crust, respectively, the subscript i represents the initial state, the subscript t represents at time t , α is the thermal expansion coefficient, $T(z, t_0)$ is the initial lithospheric temperature field, $T(z, t)$ is the lithospheric temperature field at time t , and T_m is the temperature of the thermal lithospheric bottom boundary.

Figure 4. Sketch to show the principal features of the subsidence model. a = thickness of lithosphere; T_m = temperature of asthenosphere; β = stretching factor; $G(t)$ = vertical strain rate as a function of time, t ; $v(z)$ = vertical velocity. Note that z is measured from the base of undeformed lithosphere. S_r represents the subsidence due to rifting; S_t is the thermal subsidence; S_l denotes the subsidence caused by lower crust thinning; S_f is the final subsidence.

Table 1
Values of Parameters Used in the Text

	Sediment	Upper crust	Lower crust	Lithospheric mantle
Density at 0 °C (kg/m ³)	2,500	2,700	2,900	3,330
Thermal expansion coefficient (°C ⁻¹)	0	2.4E-5	2.4E-5	3.28E-5
Specific heat (J·kg ⁻¹ ·°C ⁻¹)	1,000	1,400	1,300	1,200
Thermal conductivity (W·m ⁻¹ ·°C ⁻¹)	1.7	3.1	2	3.2
Heat generation (μW/m ³)	1.34	1.8	0.4	0.02

3.3. Stretching Factor

We use a one-dimensional finite extension model to invert the lithospheric stretching factor according to the subsidence process from backstripping analysis (Jarvis & McKenzie, 1980; White, 1994). The lithosphere is divided into four layers: a sedimentary layer, upper crust, lower crust, and mantle. We set the sedimentary thickness to 0 at first. The steps of calculating the stretching factor can be divided into the following parts: (A) gives the strain rate G for a certain period of time and solves equation (3) using finite difference methods to obtain the lithospheric temperature change; (B) the amount of subsidence in this time period is calculated according to the isostasy principle (equation (2)); and (C) by comparing with the observed subsidence curves, if the absolute value of the difference between the calculated subsidence and observed subsidence is greater than 0.1 m, we change G and repeat Steps A and B until the absolute value of difference is less than 0.1 m. Finally, we obtain the stretching factor, which is the equation of strain rate G and rifting duration.

$$\rho \cdot Cp \cdot \left(\frac{\partial T}{\partial t} + G(t)(a-z) \frac{\partial T}{\partial z} \right) = \frac{\partial}{\partial z} \left(k \frac{\partial T}{\partial z} \right) + A, \quad (3)$$

$$\beta = \exp \left(\int_0^{\Delta t} G(t) dt \right), \quad (4)$$

where ρ is the density, Cp is the specific heat capacity, T is the temperature, t is the time, a is the initial lithospheric thickness, z is the depth (the bottom of the lithosphere is 0, the direction is upward), k is the thermal conductivity, and A is the heat generation rate (Jarvis & McKenzie, 1980).

Because sedimentary thermal conductivity is lower than basement thermal conductivity, deposition results in thickening of the sediment layer, an increase of the basement temperature, decrease of the basement heat flow, and effectively delays basement cooling (Hutchison, 1985; Shi, Yu, et al., 2017). At the same time, the sediment contains radioactive elements that generate heat. The thermal blanketing effect of sediments will directly affect the calculated seafloor heat flow value. Thus, the model couples sediment deposition and compaction. In each time step, the new sediment was 0 °C, and the old sediment was compacted according to the porosity-depth relationship.

When extension ends and the lithosphere begins to cool, the temperature history can be obtained by solving equation (3) with $G = 0$. It is also necessary to consider thermal blanketing and compaction of sediments in the calculation. The theoretical thermal subsidence at the end of the rifting is obtained by equation (2). The difference between the theoretical subsidence and observed subsidence is compensated by reducing the thickness of the lower crust. The adjustment is obtained according to the isostasy principle (equation (2)). Finally, the upper and lower crustal thicknesses and heat flow can be calculated. Comparing the measured heat flow (Figure 3a) with the crustal structure derived from gravity inversion (Figures 3b–3d), the initial upper and lower crust and thermal lithospheric thickness values are adjusted until the difference between the calculated results and crustal thickness by inversion is less than 0.1 km and the difference between the calculated result and measured heat flow value is less than 0.1 mW/m².

3.4. Thermal Physical Parameters

The parameters of each layer are listed in Table 1. The values of physical parameters determine the final results, which is why we carefully chose parameters applicable to our study area. According to the drilling data in Qiongdongnan Basin (Shi et al., 2015), Yinggehai Basin (He et al., 2002), and Pearl River Basin

Figure 5. The subsidence history and the histogram of the initial crust and lithosphere. The blue dashed lines with circles are the backstripped unloaded subsidence. The red lines are the predicted subsidence history on the basis of the extension factors by assuming that the crust thinned in the synrift period follow the observed subsidence. The length of the whole cuboid is the initial lithospheric thickness. The blue cuboids represent the initial upper crust. The orange cuboids mark the initial lower crust. The orange cuboids in the shade are the thinned crust in the postrift period.

(Tang et al., 2014), the sedimentary thermal conductivities are mostly in the range of 1.4–2 W/(m °C), with an average of 1.7 W/(m °C) and the heat generations of the sediment range from 1.08 to 2.19 $\mu\text{W}/\text{m}^3$, with an average of 1.34 $\mu\text{W}/\text{m}^3$. Hence, we set the thermal conductivity and heat generation of sediments as 1.7 W/(m °C) and 1.34 $\mu\text{W}/\text{m}^3$, respectively.

So far, there are no consistent data sets for the distribution of thermal parameters in the crust of the South China Sea margins. We therefore estimate parameters through experiment-derived empirical formulas. Based on the function of temperature and crustal thermal conductivity (Whittington et al., 2009), the average values of the thermal conductivity of upper crust and lower crust are given to be 3.1 and 2 W/(m °C), respectively. Shi et al. (2002) have estimated the crustal heat generation in Xisha by the empirical formula with seismic velocity, temperature, and pressure. Results show that the heat generation of the upper crust with an average of 1.8 $\mu\text{W}/\text{m}^3$ and the heat generation at the bottom of the crust is lower than 0.4 $\mu\text{W}/\text{m}^3$ with an average of about 0.36 $\mu\text{W}/\text{m}^3$. Using measurements from exposed granulite terranes, Hasterok and Chapman (2011) estimated the average heat generation for the continental lower crust to be 0.4 $\mu\text{W}/\text{m}^3$. Integrating these results, we employ 1.8 and 0.4 $\mu\text{W}/\text{m}^3$ for the upper and lower crustal heat generation, respectively.

4. Results

4.1. Subsidence History

We select representative points to show the subsidence history on the northern margin of the South China Sea (Figure 5). The principle for selecting these points is that the selected points should cover all the geological structures shown in the section, including shelf, deep water areas, depression, and flat areas. Although the number of selected points may be limited, the results of points are sufficient to represent the region in which they locate. Moreover, some points other than P1–P17 were also selected for calculation and the results of them could be represented by P1–P17. For example, both P10 and P11 are located in shelf (Figure 2), and the subsidence curves and lower crust thinning at the points between P10 and P11 are almost equal to them. Therefore, only P1–P17 are analyzed below.

Located in the Qiongdongnan Basin, Points P1–P6 depict a slower sedimentation rate on the initial rifting stage and an accelerated sedimentation rate during the Oligocene. In the postrift stage, the subsidence rate was slow from the Early to Middle Miocene and abruptly increased from the Late Miocene to Quaternary. The synrift subsidence in the northwest and southeast of the basin are less substantial than synrift subsidence in the central depression. The anomalous postrift subsidence of the northwest continental shelf is 400–700 m, which is significantly smaller than the anomalous postrift subsidence in deep water areas that exceed 1,100 m.

Points P7–P9 are located on the west side of the Pearl River Mouth Basin. Figure 5 shows that the subsidence rate during the Oligocene is slightly higher than that during the initial rifting stage, and the synrift subsidence rates at different locations are approximately the same. The anomalous postrift subsidence ranges from 500 to 1,400 m and that of the north continental shelf is clearly smaller than in the deep water.

At the center of the Pearl River Mouth Basin, the Eocene rift subsidence rate at each point is slightly less than or approximately equal to the Oligocene rift subsidence (P10–P13). The synrift subsidence at the center of the basin is 2,000 m, which is significantly larger than the northwest and southeast of the basin (1,200–1,400 m). In the postrift stage, the subsidence rate during the Early Miocene is slower and significantly increases during the Late Pliocene. The anomalous postrift subsidence in the center of the basin and deep water in the southeast is more than 1,800 m, which is significantly larger than 500–600 m in the continental shelf area.

Points P14–P17 represent the eastern Pearl River Mouth Basin. The Eocene rift subsidence rate is less than that of the Oligocene (~600–800 m). In the postrift stage, the subsidence rate of the Early Miocene in the northwest continental shelf area is slightly greater than or equal to that from the Late Miocene to the Quaternary, and the anomalous postrift subsidence is 800–1,000 m. However, in the deep water area in the south, the subsidence rate of the Early Miocene is significantly smaller than that of the Late Miocene to Quaternary and the anomalous postrift subsidence exceeds 2,000 m.

In both the Qiongdongnan and Pearl River Mouth basins, rift subsidence in the depression center of the basin is generally larger than that in the surrounding areas of the basin. The anomalous postrift subsidence in the deep water area is significantly larger than that in the continental shelf.

4.2. Initial Crustal and Thermal Lithospheric Thickness

Previous studies usually set the crustal and thermal lithospheric thickness to 32 and 100–120 km, respectively, prior to extension in the northern continental margin of the South China Sea (Shi, Jiang, et al., 2017; Xie et al., 2006). However, the South China Sea region was dominated by an Andean-type active margin before Cenozoic continental extension, so the initial crustal and thermal lithospheric thickness are likely to have been nonuniform. Setting a uniform crustal and lithospheric thickness inevitably introduces some errors to the results. To ensure more accurate calculation results, we do not directly set a fixed value to the initial crustal and lithospheric thickness as in previous studies. Instead, we invert for initial upper and lower crustal thickness as well as the thermal lithospheric thickness based on current crustal thickness, heat flow, and observed subsidence curves.

The calculated initial crustal thicknesses of the northern margin of the South China Sea range from 27.7 to 36.3 km with an average of 31.7 km. The initial crustal thicknesses in the continental shelf area are slightly greater than or equal to the initial crustal thicknesses in the deep water area. The average initial crustal thickness on the eastern side of the continental margin is about 33 km, slightly larger than that on the west side (30.5 km). The average initial lower crustal thickness is 67% of the initial crustal thickness. Except for P17 near the Dongsha uplift, other points with a high ratio of the initial lower crust have larger synrift subsidence that exceeds 1,400 m.

The calculated average initial thermal lithospheric thickness in the northern margin of the South China Sea is 103 km. The average thermal lithospheric thickness varies little from the western to the eastern margin. However, the initial thermal lithosphere thickness of the continental shelf area is ~118 km, which is significantly larger than that of the deep water area of ~81 km.

Figure 6. (a) Thinned lower crustal thickness in the postrift stage versus the water depth. (b) The stretching factor versus ratio of the initial lower crustal thickness.

4.3. Thinned Crust in the Synrift and Postrift

We use the stretching factor to indicate the degree of lithosphere and crustal extension in the rifting period, which was derived from the inversion strain rate (equation (4)). Thinning of the crust during the rifting period leads to the formation of rift basins. The stretching factor is therefore approximately consistent with the geological structure. The average stretching factor is about 1.235 and is larger (generally >1.3) near the central depression of the basin with high-degree extension and large synrift subsidence. The stretching factor is small on the edge of the basin and shelf with low-degree extension and small synrift subsidence. The two points with the smallest stretching factor are P6 and P16. Their positions are close to the Xisha block and Dongsha uplift, both of which are hard blocks.

The thermal structure after rifting follows the cooling model and the calculated theoretical curves approach the observed subsidence curves by thinning the lower crust. The thicknesses of the lower crust reduced by thinning in postrift are shown in the black frame in Figure 5 and range from 1 to 11 km. The average thickness is 5.6 km. The amount of the reduced lower crust in the postrift is positively correlated with water depth (Figure 6a). In the continental shelf, the average thickness is 3.6 km and in deep water areas, it is 8.4 km.

5. Discussion

5.1. Model Reliability

The accuracy of the subsidence history is mainly determined by paleowater depth, base level change, and the accuracy of the decompacted sedimentary thickness, which is determined by the strata depth and relationship between porosity and depth. Abundant seismic and borehole data and previous results in the northern margin of the South China sea ensure the reliability of the backstripping analysis (Dong et al., 2008; Xie et al., 2006; Zhao et al., 2018). The subsidence history reflects the extension process of the crust and lithosphere. On this basis, the initial crustal thickness and thinned lower crustal thickness in the postrift are derived from the present-day crustal structure. We use the newly published crustal structure that has been deduced by gravity inversion (Bai et al., 2019). The root-mean-square between Moho/Conrad by gravity inversion and seismic interpretations is 1.9/2.2 km, respectively. However, relatively large upper crustal thickness discrepancies occur when there are considerable sediment thickness discrepancies from the National Geophysical Data Center global sediment thickness grid (Divins, 2004) and seismic profile. This discrepancy is mainly located in sediment basins, especially with thick sediments. In these areas, the National Geophysical Data Center sedimentary thicknesses are usually larger than those identified by the seismic profile. The crustal thickness calculated by gravity inversion is therefore larger than that identified by the seismic profile in these areas.

The subsidence history restricts the amount of thinned crust in the synrift period. A larger present-day crustal thickness leads to a larger initial thickness, which results in a smaller stretching factor. Owing to the high heat generation rate in the upper crust, a larger initial thickness of the upper crust leads to a higher calculated heat flow value. This requires a thicker initial thermal lithosphere to offset the increased thermal effects of a thicker crust. We estimate that a thermal effect produced by 1 km of thickened upper crust is necessary to be offset by an initial thermal lithospheric thickness increase of approximately 4 km. Therefore, in the depression area of the basins, the initial crustal and lithospheric thicknesses are larger and the stretching factors are smaller.

The test results show that the thermal parameters of the sediment and mantle have less than 1% influence on the final results. So we will only elaborate on the influence of the crustal thermal parameters. The thermal conductivity of the upper crust has the greatest impact on the results. If the thermal conductivity of the upper crust decreased by 10%, which is 2.8 W/(m °C), the initial thermal lithospheric thickness would minus 16%. The initial crustal thickness and stretching factor would only reduce 1.3% and 1.7%, respectively. The postrift reduced thickness of lower crust would increase by 2.8%. If the thermal conductivity of the upper crust increased by 10%, which is 3.4 W/(m °C), the initial thermal lithospheric thickness would increase by 19%. The initial crustal thickness would be 1.9% of the original value. The stretching factor would increase by 2.5% and the postrift reduced thickness of the lower crust would decrease by 4.4%.

The test results show that the initial crustal thickness, stretching factor, and the amount of reduced lower crust in the postrift were not significantly (no more than 5%) affected by the thermal parameters. Besides, the thermal parameters of the sediment and mantle have less than 1% influence on the final results. So we will only elaborate on the influence of the crustal thermal parameters on the initial thermal lithospheric thickness. The thermal conductivity of the upper crust has the greatest impact on it. Changing upper crustal thermal conductivity by $\pm 10\%$ results in changes of the initial thermal lithospheric thicknesses by more than 15%. The second major influence on the final results is the heat generation of the upper crust. A 10% change in the upper crustal heat generation would change the initial thermal lithospheric thickness by about 5%. The influences of lower crustal thermal parameters on the initial thermal lithospheric thickness are only less than 5%.

5.2. Crustal Thinning During Rifting

Rift basins and rifted margins form by the extension of the lithosphere accompanied by normal faulting and by crustal and lithospheric thinning (Brune et al., 2016). However, some rifted margins show an “extension discrepancy” where the amount of extension measured from faults is far less than that required to produce the observed crustal thinning (Reston, 2009). Although some faults may have not been identified from the seismic profile, the stretching factors determined by faults essentially represent the degree of crustal extension, at least the upper crust, in the synrift period. If the lower crust thinned in the postrift period, the

Figure 7. Schematic illustrations showing the (a) initial crustal conditions and the (b) horizontal flow of the ductile lower crustal material. The red color represents the temperature field.

present-day crustal thickness would be a composite of the crustal thinning by stretching in the synrift stage and lower crustal flow in the postrift stage that could explain certain aspects of the “extension discrepancy” problem.

Zhao et al. (2018) identified the fault distribution of a profile through the Qiongdongnan Basin and calculated the fault stretching factor to represent the stretching factor of the upper crust. The fault stretching factor (β_f) in Zone 1 at the northwest end of the profile (See Figure 4a in Zhao et al., 2018), is 1.2. β_f in Zone 4 at the middle of the profile is 1.9, and the β_f in Zone 5 at the northeast end of the profile is 1.3. Positions P1–P3 in this study roughly correspond to Regions 1, 4, and 5 in Zhao et al. (2018) and the calculated stretching factors are 1.19, 1.55, and 1.33, respectively. As discussed in section 5.1, the calculated stretching factors in basin depression areas are smaller than those derived from the seismic profile. Thus, it is reasonable that the stretching factor of the fault identified by Zhao et al. (2018) in the central depression is greater than the stretching factor at Point P2. Furthermore, the stretching factors of P1 and P3 calculated by the subsidence curve are approximately equal to the stretching factor of Zones 1 and 5 from the fault structure, which not only verifies the accuracy of the results in this study but also implies provides a new aspect on whether “extension discrepancy” exists in the northern margin of the South China Sea.

The width of the thinned crust in the western margin of the South China Sea is larger than that in the eastern margin. Hayes and Nissen (2005) proposed two end-member models to interpret this observation. Model 1 assumes that the width of the western margin was approximately equal to the width of the eastern margin and the crustal thickness on the former was ~ 1.7 times greater than the latter. During the rifting period, the western margin became more extended. Contrary to Model 1, Model 2 assumes that the crustal thicknesses were initially the same on each margin and the initial width of the rifting zone for western margin was ~ 1.7 times greater than for the eastern margin. The extensional degree of the margin on both sides was roughly equal in the rifting stage. In this study, the calculated initial crustal thickness and stretching factor of the western and eastern margins are essentially the same and more consistent with Model 2.

Figure 6b shows that the initial crustal structure determines the extension degree of the continental margin. The points where the initial lower crust ratio exceeds 70% are all located in the depressions areas. Under certain stretching forces, the greater lower crust ratio indicates that the crust is more easily stretched and vice versa. Roughly, the same crustal structure and extension degree in the synrift stage of the margins on the east and west sides of the South China Sea imply that although the starting time of stretching on both sides was different, the stretching force is essentially the same during the rifting period.

5.3. Lower Crustal Flow in the Postrift Stage

Tectonic subsidence usually occurs for the following reasons: stretching and thinning of the crust and lithosphere, cooling of the deep mantle, and/or mantle convection (Brune, 2018; McKenzie, 1978; Steinberger et al., 2001). Xie et al. (2006) calculated the dynamic topography caused by mantle convection in the northern margin of the

Table 2
Values of Parameters Used in the Calculations of the Lower Crustal Strain Rate (Sonder & England, 1986)

A_c	preexponential constant for quartz flow law	$5 \times 10^{-6} \text{ s}^{-1} \cdot \text{MPa}^{-3}$
Q_c	activation energy for quartz flow law	$1.9 \times 10^5 \text{ J/mol}$
n_c	stress exponent for quartz flow law	3
R	gas constant	$8.314 \text{ J} \cdot \text{mol}^{-1} \cdot \text{K}^{-1}$

South China Sea and found that its extent was significantly smaller than the anomalous subsidence. Mantle convection is therefore insufficient to explain the observed subsidence. Since the Late Miocene, considerable magma activity has appeared on the northern margin of the South China Sea, which indicates the absence of rapid cooling of the deep mantle (Xia et al., 2016). Shi, Jiang, et al. (2017) proposed that intrusions in the lower crust increase crustal density and accompany thermal cooling as an explanation for the rapid postrift subsidence. However, magmatic crustal thickening results in uplift rather than subsidence. In any case, there is no clear evidence of underlying magma intrusion(s) in some areas with rapid postrift subsidence, such as Baiyun Sag (Clift, 2015).

During the Early to Middle Miocene, seafloor spreading in the South China Sea continued, continental margin rifting stopped, and mantle thermal relaxation and regional thermal subsidence occurred (Franke et al., 2014; Morley, 2016). The time of the breakup of the South China Sea is around 32 Ma, but the rifting of the northern marginal continued until 23 Ma. The delay in the reflection of the strain attenuation caused by the breakup of the South China Sea suggests that the lithosphere in the margin of the South China Sea was very weak (Brune et al., 2017; Dong et al., 2018). Clift et al. (2002) estimated the lower crust viscosity of the South China Sea was 10^{18} – 10^{19} Pa·s. When the rocks' effective viscosity is less than 10^{18} – 10^{20} Pa·s, they could form a ductile channels with 10–25 km thick (Kruse et al., 1991). Although the initial crustal thickness of the continental margin of the South China Sea did not change substantially from the continental shelf to the deep water, the thickness of the initial thermal lithosphere is clearly thinner toward the sea, which indicates that the lithosphere in the offshore margin has a higher temperature and lower strength and is more susceptible to deformation (Figure 7a). On one hand, this predisposes the location of the final rupture of the South China Sea but it also provides the physical conditions for lower crustal flow.

Early subsidence of the northern margin of the South China Sea in the postrift is mainly due to thermal subsidence. While in the rift basin, owing to the large proportion of lower crust and high sedimentation rate, thick sediments with rapid accumulation promote lower crustal flow (Clift et al., 2015). After the Middle Miocene, the uplift of the Tibetan Plateau or enhanced East Asian monsoon activity strengthened the erosion of South China (Yan et al., 2009). Heavy sediments were transported from the mainland to the sea. On the one hand, rapid accumulation promotes lower crustal flow in the northern margin of the South China Sea. On the other hand, erosion on the shore also creates a regional uplift that provides a path for the lower crust to flow to the continent (Clift, 2015), causing the lower crust to flow from the sea to the land (Figure 7b), which provides the dynamic conditions for the lower crustal flow.

Assuming that the lower crust deforms by creep, the strain rate can be estimated by the relationship between the potential energy and rock layer stress (Sonder & England, 1986; Table 2):

$$\Delta PE = \int_{z_{BD}}^{z_M} \left(\frac{\dot{\epsilon}}{A_c} \right)^{1/n_c} \exp\left(\frac{Q_c}{n_c RT} \right) dz,$$

where z_{BD} is the depth of the Conrad surface, z_M is the depth of the Moho surface, $\dot{\epsilon}$ is the strain rate, A_c , Q_c , and n_c are the preexponential constant, activation energy, and stress exponent for quartz law, respectively, and R is the gas constant. Under local equilibrium compensation conditions, the potential energy difference PE can be expressed as the integral of vertical normal stress σ_{zz} (Jones et al., 1996):

$$PE = \left[\int_{-e}^L \sigma_{zz}(z) dz \right] = \left[\int_{-e}^L \left(\int_{-e}^z \rho(h) g dh \right) dz \right],$$

where L is the compensation depth, e is the altitude of the surface topography, g is gravitational acceleration, and z is the calculated depth. The average strain rate of the lower crust deformation at the postrift can be obtained by the difference of potential energy between the current crust and rift ending. With increasing water depth, the amount of lower crust thinning in the postrift increases gradually and the average strain rate increases from 10^{-23} to 10^{-16} s $^{-1}$.

6. Conclusions

The initial crustal structure and thermal lithospheric thickness in the northern margin of the South China Sea determines the degree of extension in the synrift stage. Areas with larger initial lower crust

ratios have larger stretching factors and end up becoming a depression. The postrift stage can be divided into two stages. In the first stage from Early to Middle Miocene, continental margin rifting stopped and the margin enters thermal subsidence. Only depressed areas of the basin still experience rapid subsidence. The second stage commences after the Middle Miocene when a wide range of anomalous subsidence occurred in the northern margin of the South China Sea. We find that subsidence gradually increases from the shelf to the deep water with a maximum over 2,000 m. In the initial stage, the sufficiently thick lower crust and relatively thin thermal lithosphere in the northern margin of the South China Sea provided physical conditions for lower crustal flow in the postrift stage. Areas closer to the ocean basin are more susceptible to crustal deformation. Rapid onshore denudation and sediment accumulation on the margin provide the dynamic conditions for lower crustal flow. The lower crustal flow occurs from sea to land and is responsible for the observed anomalous postrift subsidence.

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References

- Airy, G. (1855). On the computation of the effect of the attraction of mountain-masses, as disturbing the apparent astronomical latitude of stations in geodetic surveys. *Philosophical Transactions. Royal Society of London*, 145, 101–104.
- Bai, Y., Dong, D., Brune, S., Wu, S., & Wang, Z. (2019). Crustal stretching style variations in the northern margin of the South China Sea. *Tectonophysics*, 751, 1–12.
- Brune, S. (2018). Forces within continental and oceanic rifts: Numerical modeling elucidates the impact of asthenospheric flow on surface stress. *Geology*, 46, 191–192.
- Brune, S., Heine, C., Clift, P. D., & Pérez-Gussinyé, M. (2017). Rifted margin architecture and crustal rheology: Reviewing Iberia–Newfoundland, Central South Atlantic, and South China Sea. *Marine and Petroleum Geology*, 79, 251–281.
- Brune, S., Williams, S. E., Butterworth, N. P., & Muller, R. D. (2016). Abrupt plate accelerations shape rifted continental margins. *Nature*, 536(7615), 201–204. <https://doi.org/10.1038/nature18319>
- Ceramicola, S., Stoker, M., Praeg, D., Shannon, P. M., De Santis, L., Hoult, R., et al. (2005). Anomalous Cenozoic subsidence along the 'passive' continental margin from Ireland to mid-Norway. *Marine and Petroleum Geology*, 22, 1045–1067.
- Clerc, C., Ringenbach, J. C., Jolivet, L., & Ballard, J. F. (2018). Rifted margins: Ductile deformation, boudinage, continent ward-dipping normal faults and the role of the weak lower crust. *Gondwana Research*, 53, 20–40.
- Clift, P., Lin, J., & Barckhausen, U. (2002). Evidence of low flexural rigidity and low viscosity lower continental crust during continental break-up in the South China Sea. *Marine and Petroleum Geology*, 19(8), 951–970.
- Clift, P. D. (2015). Coupled onshore erosion and offshore sediment loading as causes of lower crust flow on the margins of South China Sea. *Geoscience Letters*, 2, 1–13.
- Clift, P. D., Brune, S., & Quinteros, J. (2015). Climate changes control offshore crustal structure at South China Sea continental margin. *Earth and Planetary Science Letters*, 420, 66–72.
- Clift, P. D., & Sun, Z. (2006). The sedimentary and tectonic evolution of the Yinggehai–Song Hong basin and the southern Hainan margin, South China Sea: Implications for Tibetan uplift and monsoon intensification. *Journal of Geophysical Research*, 111, B06405. <https://doi.org/10.1029/2005JB004048>
- Divins, D. L. (2004). Total Sediment Thickness of the World's Oceans & Marginal Seas. Center, N.N.G.D., Boulder, CO.
- Dong, D. D., Wu, S. G., Zhang, G. C., & Yuan, S. Q. (2008). Rifting process and formation mechanisms of syn-rift stage prolongation in the deepwater basin, northern South China Sea. *Chinese Science Bulletin*, 53, 3715–3725.
- Dong, M., Wu, S., Zhang, J., Xu, X., Gao, J., & Song, T. (2018). Lithospheric structure of the Southwest South China Sea: Implications for rifting and extension. *International Geology Review*, 1–14.
- Dupré, S., Bertotti, G., & Cloetingh, S. (2007). Tectonic history along the South Gabon Basin: Anomalous early post-rift subsidence. *Marine and Petroleum Geology*, 24, 151–172.
- Franke, D., Savva, D., Pubellier, M., Steuer, S., Mouly, B., Auxietre, J.-L., et al. (2014). The final rifting evolution in the South China Sea. *Marine and Petroleum Geology*, 58, 704–720.
- Hall, R. (2002). Cenozoic geological and plate tectonic evolution of SE Asia and the SW Pacific: Computer-based reconstructions, model and animations. *Journal of Asian Earth Sciences*, 20, 353–431.
- Hasterok, D., & Chapman, D. S. (2007). Continental thermal isostasy: 1. Methods and sensitivity. *Journal of Geophysical Research*, 112, B06414. <https://doi.org/10.1029/2006JB004663>
- Hasterok, D., & Chapman, D. S. (2011). Heat production and geotherms for the continental lithosphere. *Earth and Planetary Science Letters*, 307(1–2), 59–70.
- Hayes, D. E., & Nissen, S. S. (2005). The South China sea margins: Implications for rifting contrasts. *Earth and Planetary Science Letters*, 237, 601–616.
- He, L., Xiong, L., & Wang, J. (2002). Heat flow and thermal modeling of the Yinggehai basin, South China Sea. *Tectonophysics*, 351(3), 245–253.
- Huang, W., & Wang, P. (2006). The statistics of sediment mass in the South China Sea: Method and result. *Advances in Earth Science*, 21, 465–473.
- Huismans, R., & Beaumont, C. (2011). Depth-dependent extension, two-stage breakup and cratonic underplating at rifted margins. *Nature*, 473(7345), 74–78. <https://doi.org/10.1038/nature09988>
- Hutchison, I. (1985). The effects of sedimentation and compaction on oceanic heat flow. *Geophysical Journal International*, 82, 439–459.
- Jarvis, G. T., & McKenzie, D. P. (1980). Sedimentary basin formation with finite extension rates. *Earth and Planetary Science Letters*, 48, 42–52.
- Jones, C. H., Unruh, J. R., & Sonder, L. J. (1996). The role of gravitational potential energy in active deformation in the southwestern United States. *Nature*, 381, 37–41.

- Kruse, S., McNutt, M., Phipps-Morgan, J., Royden, L., & Wernicke, B. (1991). Lithospheric extension near Lake Mead, Nevada: A model for ductile flow in the lower crust. *Journal of Geophysical Research*, 96, 4435–4456.
- Kusznir, N., Marsden, G., & Egan, S. (1991). A flexural-cantilever simple-shear/pure-shear model of continental lithosphere extension: Applications to the Jeanne d'Arc Basin, Grand Banks and Viking Graben, North Sea. *Geological Society, London, Special Publications*, 56, 41–60.
- Larsen, H. C., Mohn, G., Nirrengarten, M., Sun, Z., Stock, J., Jian, Z., et al. (2018). Rapid transition from continental breakup to igneous oceanic crust in the South China Sea. *Nature Geoscience*, 11, 782–789.
- Le Pourhiet, L., Chamot-Rooke, N., Delescluse, M., May, D. A., Watremez, L., & Pubellier, M. (2018). Continental break-up of the South China Sea stalled by far-field compression. *Nature Geoscience*, 11(8), 605–609.
- Li, C. F., Xu, X., Lin, J., Sun, Z., Zhu, J., Yao, Y. J., et al. (2014). Ages and magnetic structures of the South China Sea constrained by deep tow magnetic surveys and IODP Expedition 349. *Geochemistry, Geophysics, Geosystems*, 15(12), 4958–4983. <https://doi.org/10.1002/2014GC005567>
- McKenzie, D. (1978). Some remarks on the development of sedimentary basins. *Earth and Planetary Science Letters*, 40, 25–32.
- Morley, C. K. (2016). Major unconformities/termination of extension events and associated surfaces in the South China Seas: Review and implications for tectonic development. *Journal of Asian Earth Sciences*, 120, 62–86.
- Morley, C. K., & Westaway, R. (2006). Subsidence in the super-deep Pattani and Malay basins of Southeast Asia: A coupled model incorporating lower-crustal flow in response to post-rift sediment loading. *Basin Research*, 18(1), 51–84.
- Peron-Pinvidic, G., Manatschal, G., & Osmundsen, P. T. (2013). Structural comparison of archetypal Atlantic rifted margins: A review of observations and concepts. *Marine and Petroleum Geology*, 43, 21–47.
- Ranero, C. R., & Perez-Gussinye, M. (2010). Sequential faulting explains the asymmetry and extension discrepancy of conjugate margins. *Nature*, 468(7321), 294–299. <https://doi.org/10.1038/nature09520>
- Reston, T. J. (2009). The extension discrepancy and syn-rift subsidence deficit at rifted margins. *Petroleum Geoscience*, 15, 217–237.
- Reston, T. J., & Phipps Morgan, J. (2004). Continental geotherm and the evolution of rifted margins. *Geology*, 32, 133.
- Royden, L., & Keen, C. (1980). Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. *Earth and Planetary Science Letters*, 51, 343–361.
- Slater, J. G., & Christie, P. A. (1980). Continental stretching: An explanation of the post-Mid-Cretaceous subsidence of the central North Sea Basin. *Journal of Geophysical Research*, 85, 3711–3739.
- Shi, X., Burov, E., Leroy, S., Qiu, X., & Xia, B. (2005). Intrusion and its implication for subsidence: A case from the Baiyun Sag, on the northern margin of the South China Sea. *Tectonophysics*, 407, 117–134.
- Shi, X., Jiang, H., Yang, J., Yang, X., & Xu, H. (2017). Models of the rapid post-rift subsidence in the eastern Qiongdongnan Basin, South China Sea: Implications for the development of the deep thermal anomaly. *Basin Research*, 29, 340–362.
- Shi, X., Wang, Z., Jiang, H., Sun, Z., Sun, Z., Yang, J., et al. (2015). Vertical variations of geothermal parameters in rifted basins and heat flow distribution features of the Qiongdongnan Basin (in Chinese). *Chinese Journal of Geophysics*, 58, 939–952.
- Shi, X., Yu, C., Chen, M., Yang, X., & Zhao, J. (2017). Analyses of variation features and influential factors of heat flow in the northern margin of the South China Sea. *Earth Science Frontiers*, 56–64.
- Shi, X., Zhou, D., Qiu, X., & Zhang, Y. (2002). Thermal and rheological structures of the Xisha trough, South China Sea. *Tectonophysics*, 351(4), 285–300.
- Sonder, L. J., & England, P. (1986). Vertical averages of rheology of the continental lithosphere: Relation to thin sheet parameters. *Earth and Planetary Science Letters*, 77, 81–90.
- Steinberger, B., Schmeling, H., & Marquart, G. (2001). Large-scale lithospheric stress field and topography induced by global mantle circulation. *Earth and Planetary Science Letters*, 186, 75–91.
- Tang, X. Y., Hu, S. B., Zhang, G. C., Yang, S. C., Shen, H. L., Rao, S., & Li, W. W. (2014). Characteristic of surface heat flow in the Pearl River Mouth Basin and its relationship with thermal lithosphere thickness. *Chinese Journal of Geophysics*, 57(6), 1857–1867.
- Taylor, B., & Hayes, D. E. (1983). Origin and history of the South China Sea basin. In *The tectonic and geologic evolution of Southeast Asian seas and islands: Part 2, Geophysical Monograph Series* (pp. 23–56). Washington, DC: American Geophysical Union.
- Wessel, P., & Smith, W. H. F. (1998). New, improved version of the generic mapping tools released. *Eos, Transactions of the American Geophysical Union*, 79(47), 579–579.
- White, N. (1994). An inverse method for determining lithospheric strain rate variation on geological timescales. *Earth and Planetary Science Letters*, 122, 351–371.
- White, N., Thompson, M., & Barwise, T. (2003). Understanding the thermal evolution of deep-water continental margins. *Nature*, 426(6964), 334–343. <https://doi.org/10.1038/nature02133>
- Whitmarsh, R. B., Manatschal, G., & Minshull, T. A. (2001). Evolution of magma-poor continental margins from rifting to seafloor spreading. *Nature*, 413(6852), 150–154. <https://doi.org/10.1038/35093085>
- Whittington, A. G., Hofmeister, A. M., & Nabelek, P. I. (2009). Temperature-dependent thermal diffusivity of the Earth's crust and implications for magmatism. *Nature*, 458(7236), 319–321. <https://doi.org/10.1038/nature07818>
- Wu, S., Zhang, X., Yang, Z., Wu, T., Gao, J., & Wang, D. (2016). Spatial and temporal evolution of Cenozoic carbonate platforms on the continental margins of the South China Sea: Response to opening of the ocean basin. *Interpretation*, 4, SP1–SP19.
- Xia, S., Zhao, D., Sun, J., & Huang, H. (2016). Teleseismic imaging of the mantle beneath southernmost China: New insights into the Hainan plume. *Gondwana Research*, 36, 46–56.
- Xie, X., Müller, R. D., Li, S., Gong, Z., & Steinberger, B. (2006). Origin of anomalous subsidence along the Northern South China Sea margin and its relationship to dynamic topography. *Marine and Petroleum Geology*, 23, 745–765.
- Yan, Y., Hu, X., Lin, G., Xia, B., Li, X., & Patel, R. C. (2009). Denudation history of South China block and sediment supply to northern margin of the South China Sea. *Journal of Earth Science*, 20, 57–65.
- Zhao, Z., Sun, Z., Liu, J., Pérez-Gussinyé, M., & Zhuo, H. (2018). The continental extension discrepancy and anomalous subsidence pattern in the western Qiongdongnan Basin, South China Sea. *Earth and Planetary Science Letters*, 501, 180–191.