Effective elastic thickness of Zealandia and its implications for lithospheric deformation

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ABSTRACT

Zealandia was once located at the active Gondwana margin and experienced complicated tectonic evolution from continental breakup to subduction and oblique slipping along the plate boundary. Here, we investigate the lithospheric thermal and rheological properties via spatial variations in effective elastic thickness (\(T_e\)). Using the topography, sediment thickness and gravity data, the fan wavelet coherence method is employed to recover \(T_e\) values. The effects of dynamic topography and gravity in the subduction zone on the estimated \(T_e\) are considered and removed. The final results reveal that the \(T_e\) variation pattern generally coincides well with the tectonic regimes. Relatively high \(T_e\) values up to 22 km are observed in North Island and the southernmost part of South Island, while central and northern South Island are dominated by smaller \(T_e\) values (<15 km) similar to those of the surrounding submerged plateaus. We interpret the relatively high \(T_e\) values as indicating the combined lithospheric strength due to the contacting plates in the subduction zone. To the north along the Tonga-Kermadec subduction zone, the lithosphere surprisingly exhibits anomalously high \(T_e\) values, ranging from 22 to 36 km, which contradicts the geological evidence of normal faults developed in the incoming Pacific Plate and the magmatism in the overlying Australian Plate that generally contribute to low lithospheric strength. We speculate that the high-\(T_e\) zone is related to the large bathymetry and gravity anomalies in the bending plate and controlled by non-isostatic factors. Finally, we find widespread low \(T_e\) values covering the entire arcs and oceanic basins in northern Zealandia, which may indicate that regional lithospheric thinning associated with delamination weakened the integrated strength of the lithosphere.

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1. Introduction

Zealandia, which is located in the Southwest Pacific region, is a submerged geological continent, except for the subaerial portion of New Zealand, which separated from Gondwana during the Cretaceous (Mortimer et al., 2017). The development of the Australian, Pacific and Antarctic plates since the Mesozoic has led Zealandia to evolve into a variety of tectonic units: ocean-ocean and ocean-continent subduction zones along the Tonga-Kermadec Trench and Hikurangi Trench, an oblique continental collision zone along the Alpine Fault, and arcs and oceanic basins in northern Zealandia, as well as submerged plateaus surrounding New Zealand. The collisions and rifting between these plates provide an ideal laboratory to study lithospheric deformation and rheology under long-term plate subduction and loading associated with rifting. Although various relevant studies have been carried out in this region using several datasets and techniques, e.g., the joint inversion of passive and active seismic source data (Eberhart-Phillips and Bannister, 2002) and GPS measurements (Beavan et al., 2002; Wallace et al., 2004), either the regional scale is too limited to cover all of Zealandia or the time scale is far from reflecting geological time scales due to the dataset itself.

Lithospheric mechanical strength is a fundamental property of the lithosphere that controls deformation processes in response to the long-term (>10^9 yr) geological loads (Burov and Diament, 1995). As this property is difficult to be measured directly, effective elastic thickness (\(T_e\)) is used as a proxy for lithospheric strength. \(T_e\) equals to the thickness of an idealized elastic beam that would bend similarly to the actual lithosphere under the same applied loads (Watts, 2001). In the last several decades, many researchers have concentrated on the determinations of \(T_e\) and found the connection of this parameter with the lithospheric evolution. Watts (1978) calculated the flexural rigidities of the Hawaiian-Emperor seamount chain and showed that the \(T_e\) values

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of oceanic lithosphere reflect its thermal age. Watts and Burov (2003) reported that continental lithosphere has a multi-layer rheological structure and that its $T_v$ values are primarily controlled by the thermal state, composition and thickness of the lithosphere. Moreover, they found that these values in continental lithosphere vary over a wide range, from <10 km in tectonically active continents (Lowry and Smith, 1995) to over 100 km in stable craton (Zuber et al., 1989). In contrast, the $T_v$ values in the oceanic lithosphere, which has a single-layer rheological structure, are generally within 50 km. Pérez-Gussinyé et al. (2009) employed $T_v$ variations in Africa to analyze the lithospheric structure and different tectonics. Based on their finding of a low-$T_v$ values zone along the rifted conjugate margins of India and Madagascar, Ratheesh-Kumar et al. (2015) proposed a plate construction of the India-Madagascar paleo-fit. These observations demonstrate that $T_v$ can provide valuable information on the lithospheric structure and deformation.

Over decades of $T_v$ research, multiple methods have been proposed and developed for recovering its value. In some current mainstream approaches (Lowry and Smith, 1995; Kirby and Swain, 2011; Pérez-Gussinyé et al., 2009), $T_v$ is generally estimated in different domains according to topography and Bouguer gravity anomalies, incorporating information on loading and the resulting generated downward flexure. A few methods complete the calculation in the spatial domain (Watts et al., 1975; Braitenberg et al., 2002), but most invert $T_v$ in the frequency domain based on various measures (e.g., Lowry and Smith, 1995; Kalnins and Watts, 2009; Pérez-Gussinyé et al., 2009; Kirby and Swain, 2011), producing high-resolution $T_v$ values over the continental and oceanic lithosphere. For example, Kalnins and Watts (2009) estimated the spatial variations in $T_v$ by combining the window technique and admittance method, while Pérez-Gussinyé et al. (2009) and Kirby and Swain (2011) recovered these variations using Forsyth's (1985) method in the Fourier and wavelet domains, respectively.

In this study, we employ the fan wavelet coherence method (Kirby and Swain, 2011) to present a new map of spatial variations in $T_v$ over Zealandia based on the satellite-derived gravity model, the topography and sediment thickness grid data. By analyzing the evolutionary processes in this region, we interpret the lithospheric structure and deformation based on the implications of the $T_v$ values. The following two topics are the primary focus on: (I) the cause of high $T_v$ values along different types of plate boundaries and (II) the manner of lithospheric deformation in arcs and oceanic basins in northern Zealandia.

2. Tectonic setting

Reconstruction studies (Eagles et al., 2004) have suggested that the Chatham Rise, Campbell Plateau and Bounty Trough (Fig. 1) were connected with the Antarctic Plate before the Cretaceous breakup of Gondwana and were located close to the Pacific margin of Gondwana at that time. The Chatham Rise once overrode the subducting Phoenix Plate and then separated from Marie Byrd Land at approximately 90 Ma, changing from a convergent to a divergent plate setting. Studies of magnetic anomalies indicate that the Campbell Plateau separated from West Antarctica at 83 Ma (Larter et al., 2002). During the Eocene and Oligocene, the Pacific Plate began to subduct obliquely beneath the Australian Plate and then formed the modern Tonga-Kermadec Trench in the north (Schellart et al., 2006; Collett et al., 2008). The very large trench system extends to the southern Hikurangi Plateau, changing from ocean-ocean to ocean-continent (North Island) subduction. Along the subduction margin, the downwelling Pacific Plate is extremely steep in morphology, exceeding 10,000 m in depth, and it becomes flat in the Hikurangi Plateau. This change in the dip of subduction is related to the great buoyancy of the Hikurangi Plateau (Wood and Davy, 1994). This plateau is a remnant large igneous province that separated from other plateaus during the Cretaceous (Wood and Davy, 1994; Davy et al., 2008). Seismic studies show that the Hikurangi Plateau has a lithospheric thickness of approximately 73 km (Stern et al., 2015). On South Island, the subduction system transforms into strike-slip faulting; for example, the Marlborough fault system (Fig. 1b) comprises five major strike-slip faults, and the Alpine fault next to the South Island accommodates the relative motion of these two plates (Wallace et al., 2004; Seebeck et al., 2014). The polarity of subduction is reversed in the southernmost Macquarie Ridge Complex, where the Australian Plate subducts beneath the Pacific Plate.

Based on multiple geophysical techniques, e.g., 2-D travel-time inversion and joint 3-D inversions of gravity and received earthquakes and active sources, extensive studies on crustal and lithospheric structures in New Zealand have been conducted. Eberhart-Phillips and Bannister (2002) constructed a 3-D crustal model based on seismic and gravity data, and the results detected a low-velocity zone in the Alpine fault, exceeding 15 km in depth. Similarly, Scherwath et al. (2003) detected low-velocity zones near the western coast of South Island (25 km in width) at approximately 10 km depth and on the eastern front of the Alpine fault (50 km in width) at approximately 15 km depth, which are related to the cracks created due to plate bending and fluid pressure, respectively. In the Southern Alps Orogen, several studies found a thickened crust of up to 44 km and an asymmetric crustal root with a larger gradient on the western side of the mountains than on the eastern side (Eberhart-Phillips and Bannister, 2002; Scherwath et al., 2003). North Island is characterized by a thinned crust in the Taupo Volcanic Zone (TVZ, Fig. 1b), and the crustal thickness increases toward both sides to ~26 km (Stratford and Stern, 2006). For the submerged plateaus surrounding New Zealand, they belong to the extended continental crust and have a uniform crustal thickness of 20–24 km (Grohs et al., 2008).

The subparallel N-S-trending arcs and oceanic basins are located in northern Zealandia. From west to east, this system consists of the Lord Howe Ridge, the New Caledonia Basin, New Caledonia and the Norfolk Ridge. The Lord Howe Rise is one of the largest submarine plateaus, and its spatial distribution is >1500 km in length and 500 km in width. Conventional ideas interpret the rise as a fragment that detached from the eastern Gondwana supercontinent before its breakup, which was formerly connected to eastern Australia (Gaina et al., 1998; Schellart et al., 2006). The New Caledonia Basin is a rifted trough bounded by the Lord Howe Rise in the west. The time and mechanism of its formation are debated. Many authors have suggested that crustal extension and thinning beneath the New Caledonia Basin were associated with the Gondwana breakup in the Cretaceous (Uruski and Wood, 1991; Lafay et al., 2005). However, Collot et al. (2008) suggested that the northern New Caledonia Basin subsided in the Oligocene. Thus far, direct evidence of the crustal nature of the New Caledonia Basin has not been obtained. Based on the constructed gravity model, Uruski and Wood (1991) showed that this basin is underlain by thinned continental crust. In contrast, Sutherland (1999) favored the idea of oceanic crust according to linear magnetic anomalies. The crustal structure in this area has been studied via seismic refraction and gravity modeling (Klingelhofer et al., 2007). The crustal thickness changes from 23 km in the northern Lord Howe Rise to 10 km in the New Caledonia Basin and then increases again to 17 km beneath the Norfolk Ridge.

3. Methodology and data

3.1. Fan wavelet coherence method

The lithosphere, with its different flexural rigidities, deflects variously under a specific load and produces the corresponding observed gravity anomaly. Consequently, the coherence between the topography and the Bouger gravity anomaly can be used to measure the flexural degree and properties of the mechanical strength of the lithosphere. Forsyth (1985) stated that coherence is a function of the wavelength. Theoretical analysis shows that the coherence tends toward one at long wavelengths, where an elastic plate is isostatically compensated (Airy isostasy). However, the coherence tends toward zero for short
wavelength loads, indicating that the loads undergo mechanical support. The specific wavelength band at which the coherence changes dramatically from 0 to 1, i.e., the “transition wavelength”, increases as the lithospheric flexural rigidity increases (Watts, 2001). These transition wavelengths in the coherence curve provide a fundamental basis for measuring $T_e$ values. In Forsyth’s (1985) method, the initial surface...
and subsurface loads are built based on an elastic plate model with an assumed $T_e$ value using the observed topography and gravity anomaly. The initial loads produce the predicted topography and gravity anomaly, and the corresponding coherence is calculated and further compared with that of the observed topography and gravity anomaly. The assumed $T_e$ value changes, and when the predicted coherence best matches the observed coherence, the $T_e$ value is obtained.

Kirby and Swain (2004, 2011) developed the wavelet coherence technique, which is a combination of coherence and wavelet transformation and has been demonstrated to accurately recover spatial variations in $T_e$ values with high resolution. To date, this method has been successfully applied in several oceanic and continental lithosphere studies (e.g., Mao et al., 2012; Jiménez-Díaz et al., 2014; Ji et al., 2017). Thus, the wavelet coherence technique is employed to complete the calculation of auto-spectra and cross-spectra between gravity and topography in terms of coherence across the study region.

To avoid anisotropic bias when estimating $T_e$ values, the fan wavelet was developed by Kirby and Swain (2004), presenting a superposition of a series of 2-D Morlet wavelets. The fan wavelet can produce isotropic and complex wavelet coefficients. The 2-D Morlet wavelet in the wavenumber domain can be expressed as

$$\hat{w}_w(k) = e^{-|k-k_0|^2/2}$$

where $s$ and $\theta$ are the scale and azimuth of the Morlet wavelet, respectively, and $k_0 = (|k_0| \cos \theta, |k_0| \sin \theta)$ is the central wavenumber. Kirby and Swain (2011) proved that Eq. (1) is valid when $|k_0|$ is $>5$. In contrast, for lower $|k_0|$ values, Eq. (1) does not produce the needed zero-mean values for the wavelet (Addison et al., 2002). In this case, replacement by the complete Morlet wavelet is required, which restores its nature of zero-mean values. The complete Morlet wavelet in the wavenumber domain is

$$\hat{w}_w(k) = e^{-|k-k_0|^2/2} - e^{-(|k|^2 - |k_0|^2)/2}$$

The 2-D wavelet coherence between the observed gravity and topography at scale ($s$) and location ($x$) can be calculated by summing the wavelet cross- and co-spectra over different azimuths through the following equation (Kirby and Swain, 2011):

$$\gamma^2_{obs}(s, x) = \left| \frac{\text{Re} \left( \hat{g}_w h^*_w \right)_{s,x}}{\hat{g}_w \hat{h}_w^{*}_{s,x}} \right|^2$$

where $\hat{g}_w$ and $\hat{h}_w$ are the wavelet coefficients of the Bouguer gravity anomaly and topography obtained using the wavelet transform. The asterisk denotes the complex conjugation, and the angular brackets represent averaging over the azimuths to ensure that the wavelet coefficients are isotropic.

### 3.2. Data

According to Eq. (3), the equivalent topography and complete Bouguer gravity anomaly data are required to estimate values of $T_e$. In this study region, the data selected include the bathymetry/topography derived from the ETOPO1 (Fig. 1) 1′ × 1′ global digital elevation model (Amante and Eakins, 2009), the free-air gravity anomaly extracted from the global gravity anomaly model version 23.1 (Sandwell et al., 2014) with 1 arc-minute resolution over both land and ocean (Fig. 2a) and the sediment thickness from the National Geophysical Data Center (NGDC) sediment database (Divins, 2003) with a grid spacing of 5 arc-minutes (Fig. 2c).

Considering that most of the study area is dominated by ocean, we transformed the seawater and the sediments buried beneath the seawater into the equivalent topography $H$ (Fig. 3a). This process means that the bathymetry ($h$) and sediment thickness ($s$) need to be multiplied by the factors $(\rho_s - \rho_w)/\rho_c$ and $(\rho_s - \rho_c)/\rho_c$, respectively, as follows:

$$H = h(\rho_s - \rho_w)/\rho_c + s(\rho_s - \rho_c)/\rho_c$$

where $\rho_w$, $\rho_s$, and $\rho_c$ are the densities of the seawater, sediment and crust, respectively (Table 1).

The Bouguer correction is computed using the F22BOUG program, in which Bouguer slab, curvature and terrain corrections are included (Fuller et al., 2008), Shi et al. (2017) and Kaban et al. (2018) underlined the effect of sediments on flexural analysis and suggested that whether sediments are considered could significantly affect the estimation of $T_e$. Hence, we further obtain the gravity effect (Fig. 2d) due to the sediments using the Parker algorithm (Parker, 1972). During this process, the sediment compaction model is utilized to control the density changes with depth. The complete Bouguer gravity anomaly is generated by subtracting the gravity effects of land, seawater and sediments from the observed gravity anomaly (Fig. 3b). Segev et al. (2012) also calculated the complete Bouguer gravity anomaly of the Southwest Pacific region, in which the gravity effect of sediments was not considered. Similar to other flexural studies (Kalninns and Watts, 2009; Pérez-Gussinyé et al., 2007, 2009), we hypothesize that subsurface loads occur at the Moho, which is extracted from the CRUST1.0 model (Laske et al., 2013). Furthermore, the CRUST 1.0 model can satisfy the needs for other information on the structure of the lithosphere.

### 4. Results

#### 4.1. $T_e$ values and gravitational noise

Kirby and Swain (2011) proved that the central wavenumber is a critical parameter that controls the spatial resolution of the resulting $T_e$ values. Larger wavenumber values yield a better resolution in the wavenumber domain, a lower space-domain resolution and a higher accuracy, and vice versa (Addison et al., 2002). To ensure both the accuracy of $T_e$ and its spatial variation features, the values of the central wavenumber chosen in this study are 3.081 and 5.336. In the process of recovering $T_e$ values, larger grid data are used to avoid edge effects. The variations in $T_e$ values recovered are shown in Fig. 4a and Fig. 4b, and the trends and ranges between the two figures are largely consistent. Fig. 4a, with a lower central wavenumber, appears to show a higher spatial resolution by exhibiting the variations more sharply, and $T_e$ values with a larger central wavenumber (Fig. 4b) are more accurate. For example, the amplitudes of the estimated $T_e$ values in the SNHT (South New Hebrides Trench) and eastern Australian continent in Fig. 4a are larger and their wavelengths are shorter.

Considering that the $T_e$ results with a larger wavenumber are more accurate, Fig. 4b is used in this study to describe the spatial variations in $T_e$ values. The $T_e$ values in the study region vary from 2 km to 36 km, and their pattern of variation correlates well with lithospheric structures at the regional scale and with major tectonic boundaries. The eastern Australian continent is characterized by moderate $T_e$ values, which change between 15 km and 30 km, indicating that the lithosphere beneath the Tasman Orogen is not strong. The $T_e$ values decrease gradually seaward, which is similar to the results of Zuber et al. (1989). The $T_e$ values along the Australian coastline are lower than 15 km, and this feature continues to Tasmania in the south, reflecting the low mechanical strength in southeastern Australia. New Zealand is another continental block and exhibits $T_e$ values with similar amplitudes, which indicates that lithosphere deformation occurred under the influence of collision between the Australia Plate and Pacific Plate. Relatively high $T_e$ values occur in North Island, at up to 22 km, whereas in South Island, these values decrease to approximately 15 km, except in the southernmost part, where a certain degree of mechanical strength still remains. In addition, the submerged continental lithosphere surrounding New Zealand, such as the Campbell Plateau and Chatham Rise in
the east and the Challenger Plateau in the northeast, shows low to intermediate $T_e$ values (~20 km) and records the lithospheric thinning since the breakup of Gondwana (Gaina et al., 1998; Larter et al., 2002; Grobys et al., 2008).

The feature of high $T_e$ values observed in North Island does not extend to the nearly N-S-trending arcs and oceanic basins in northern Zealandia, including the Lord Howe Rise, New Caledonia Basin, New Caledonia Basin and Norfolk Ridge. Surprisingly, no variations in $T_e$ are apparent in this region, and a low mechanical strength of the lithosphere ($T_e < 9$ km) occurs across the entire region, suggesting possible regional isostatic compensation. However, the extremely low-$T_e$ zone (3–5 km) along the west margin of the Lord Howe Rise (along the red dashed lines in Fig. 4a) could still mark the tectonic boundary and record its rifting history.

A prominent feature—distinctively high $T_e$ values (22–36 km)—occurs along the Tonga-Kermadec Trench, covering the trenchward downgoing Pacific Plate and the subducted area behind the trench, which is consistent with the findings of earlier studies of a global $T_e$ map based on a combination of the admittance algorithm and moving window technique (Kalnins, 2011). Additionally, Shi et al. (2017) and Bai et al. (2018) found similar high $T_e$ values over the Sunda Trench and Mariana Trench. The band of high $T_e$ values with widths of 700–1300 km is aligned basically parallel to the Tonga-Kermadec Trench, and the $T_e$ values in the southern part (north of 30°S) are larger than those in the northern part. Although the high-value zone covers the two sides of the trench, it is located mostly in the subducting Pacific Plate. This pattern dominates in the entire Tonga-Kermadec Trench and abruptly decreases to the level of the surrounding oceanic lithosphere to the south, which exactly corresponds to the area where the Tonga-Kermadec Trench and Hikurangi Trench separate.

McKenzie and Fairhead (1997) and McKenzie (2003) stated that erosion and sedimentation can generate topographically unexpressed initial internal loading, which is termed “gravitational noise”, affecting the observed coherence and then biasing the $T_e$ results. To assess the

Fig. 2. (a) Free-air gravity anomaly from the satellite-derived gravity model Version 23.1 (Sandwell et al., 2014). (b) Bouguer anomaly obtained by removing the topography correction from (a). (c) Sediment thickness (Divins, 2003) and (d) calculated gravity effect due to sediments.
effect of gravitational noise on the $T_e$ estimates, the maximum values of the free-air normalized squared imaginary component (free-air NSIC) around the transition wavelength (Kirby and Swain, 2009) are provided in Fig. 5a and b. Most studies (e.g., Kirby and Swain, 2009; Mao et al., 2012; Chen et al., 2015) assume that the $T_e$ values are possibly contaminated by gravitational noise when the corresponding free-air NSIC values are larger than 0.5. Therefore, we mask this noise out in Fig. 5c and d. Although Kirby and Swain (2009) noted that calculated noise is inclined to appear in regions of subdued topography, such as the wide and flat oceanic seafloor and hinterland of eastern Australia in Fig. 5a, our results also show that significant noise is present in the Tonga, Kermadec and South New Hebrides subduction zones. The shaded areas in Fig. 5c present the biased $T_e$ results. The sizes of the affected areas in the subduction zones and on the seafloor decrease for higher $|k_0|$ values, but that in hinterland of eastern Australia remains for different $|k_0|$ values. The $T_e$ errors are calculated (Fig. S1 in the Supplementary Material), and they vary mostly within 3 km, with the maximum error (5 km) dominating in the ocean-ocean subduction zones. We attribute the noise in the subduction zone to the extra tectonic force that sabotages the lithospheric isostatic equilibrium mechanism and contaminates the coherence of the surface and subsurface loads. The large noise and error is observed in previous studies of the subduction zones as well, e.g., the Peru-Chile Trench, Central American Trench and Mariana Trench (Tassara et al., 2007; Jiménez-Díaz et al., 2014; Bai et al., 2018).

4.2. $F$ values

Fig. 6 shows the internal load fraction ($F$) as a function of the wave-number near the flexural wavelength corresponding to the estimated $T_e$ values (Tassara et al., 2007). This parameter can be calculated based on the load deconvolution method and refers to the ratio of subsurface to total loads. $F$ is interchangeable with $f$, which is defined as the ratio of the subsurface and surface loads and expressed as $f/(1 + f)$ (Mckenzie, 2003). Consequently, $F$ is usually used to assess the distribution of surface and subsurface loading (Chen et al., 2015; Shi et al., 2017). For example, the area where $F$ is below 0.5 is dominated by surface loads, e.g., the Louisville Ridge and purely oceanic areas, which is consistent with the results found in other studies (Jiménez-Díaz et al., 2014; Shi et al., 2017). However, the values in the Tonga-Kermadec subduction zone are excluded from the analysis because the corresponding estimated $T_e$ values may be biased, as discussed later. Here, we represent the areas in which $T_e$ values are unrealistic based on the distribution of gravitational noise and $T_e$ errors, shown in Fig. 5c.

Large $F$ values are observed in most of the arcs and oceanic basins in northern Zealandia (Fig. 6), including the thinned submerged crust, e.g., the Lord Howe Rise and New Caledonia Basin, and the subaerial South Island and North Island, suggesting that these regions are controlled by subsurface loading. Furthermore, eastern Australia is dominated by the largest $F$ values, which extend to Tasmania and South Tasman Rise in the south. A high-velocity and high-density body is shown in the lower crust beneath the northern New Caledonia Basin (Klingelhofer et al., 2007). Grobys et al. (2007) also found a massive high-velocity body beneath the Campbell Plateau and suggested that an underplated body may extend to nearby plateaus, such as the Lord Howe Rise and Challenger Plateau. In addition, the subsurface loads

| Table 1: Parameters used for the gravity correction and coherence technique. |
|-----------------|-----------------|-----------------|
| Parameter       | Value           | Units           |
| Density of the sea water | 1030           | kg·m$^{-3}$     |
| Density of the upper crust | 2670           | kg·m$^{-3}$     |
| Density of the crust     | 2800           | kg·m$^{-3}$     |
| Density of the mantle     | 3300           | kg·m$^{-3}$     |
| Young’s modulus         | 100            | GPa             |
| Poisson’s ratio          | 0.25           |                 |
| Gravitational constant   | 6.67×10$^{-11}$| m$^2$·kg$^{-1}$·s$^{-2}$ |
| Gravitational acceleration| 9.80           | m·s$^{-2}$      |
may include the high-density subducting Pacific slab, which may also affect the North Island. For the larger $F$ values in South Island, we speculate that they are related to the deep crustal root beneath the Southern Alps Orogen, where the distinctively thickened lower crust increases the sizes of the subsurface loads (Scherwath et al., 2003; Ball et al., 2016).

### 4.3. Effect of subduction zone dynamics on $T_e$ estimates

During the process of subduction, an asthenospheric wedge induces sinking stress that alters and deepens the bathymetry of the overlying plate (Husson, 2006). Moreover, the subducting slab that is colder and has a higher density than that of the surrounding materials of the asthenosphere inevitably provides a positive gravity effect above the surface. Accordingly, additional perturbations are present on the equivalent topography and Bouguer gravity anomaly along the Tonga-Kermadec-Hikurangi trench system, which may change the observed coherence and contaminate the $T_e$ results. Bai et al. (2018) referred to these perturbations as an effect of dynamic topography and gravity.

To remove these effects, we need to restore the input data to situations in which the topography and gravity represent only the deflections caused by surface loadings. The downward force on the surface of the overlying plate can be assessed using the Stokes stream function and is expressed as follows (Husson, 2006):

$$ F = \frac{3\Delta\rho_i v_i g z_i^2}{\pi \mu_i} $$

where $\Delta\rho_i$ and $v_i$ denote the density anomaly and its volume of the subducting slab, respectively; $g$ is the gravitational acceleration; $z_i$ is the depth of the slab unit beneath the surface and $r_i$ is the distance between the unit and the calculation point at the surface. Then, the relationship between the dynamic topography ($h_d$) and the pulling force ($\sigma$) is expressed as

$$ \sigma = (\rho_m - \rho_i)gh_d. $$

When the surface is free of stress, these two forces are equal. The final dynamic topography caused by the slab mass can be calculated as follows:

$$ H = \sum_j \sum_i \frac{3\Delta\rho_i v_i g z_i^2}{\pi \mu_i (\rho_m - \rho_i)} $$

where $\rho_m$ indicates the mantle density and $\rho_i$ represents the density of air or seawater, which are shown in Table 1. The density anomaly used is converted according to the shear wave velocity model (SL2013sv, Schaeffer and Lebedev, 2013) based on the relationship between them. The corresponding equation and parameters have been clearly illustrated by previous studies (see Eq. 14 of Bai et al. (2018)). By taking the downgoing slab as a cold plate with a constant thickness of 100 km (Husson, 2006; Pérez-Gussinyé et al., 2008; Bai et al., 2018), the dynamic topography (Fig. 7a) can be calculated using Eq. (6). Finally, based on the assumed density contrast between the downgoing plate and surrounding materials, the corresponding dynamic gravity (Fig. 7b) is further obtained by the application of the Parker algorithm (Parker, 1972).

The $T_e$ values are then re-estimated after removing the dynamic topography and gravity from the equivalent topography and Bouguer gravity anomaly and are shown in Fig. 8. However, as seen in Fig. 8, $T_e$ varies little in the subduction zone after correcting for the dynamic effects or the size of the noise-affected zone, suggesting that the gravity effect due to the slab mass and the downward topography caused by subduction do not contaminate the observed coherence. Following Pérez-Gussinyé et al. (2008), we consider the circumstance in which the dynamic topography on the overlying lithospheric surface is unexpressed because of the lower viscosity in the asthenospheric wedge and correct for only the gravity effect. The test result shows that the differences between the $T_e$ values before and after considering the

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*Fig. 4.* The spatial variations in the $T_e$ values of Zealandia and surrounding regions estimated using the fan wavelet coherence method with $|k_0|=3.081$ (a) and 5.336 (b). To better orient the reader, topographic shaded relief is superimposed. For the abbreviations and symbols, please refer to Fig. 1.
dynamic effect are still negligible. Bai et al. (2018) performed the dynamic correction over the Mariana Trench and found that the high $T_e$ values along the trench axis decreased to approximately 50 km from the original values that were larger than 100 km. Despite the strong depression of the band of high-amplitude $T_e$ values, the large mechanical strength was preserved, indicating that the dynamic effect is not the key factor affecting the observed band of high $T_e$ values along the subduction zones.

Fig. 5. The maximum values of the free-air normalized squared imaginary component (free-air NSIC) around the Bouguer transition wavelength for $|k_0|=3.081$ (a) and 5.336 (b). The corresponding estimated $T_e$ values (c and d) marked by the gray areas have free-air NSIC values larger than 0.5, which may imply the possible influence of gravitational noise. The shaded areas in Fig. 5c indicate biased $T_e$ values.
4.4. Comparison with previous results

A few different methods have been employed to estimate the $T_e$ values in New Zealand and its surrounding area. Holt and Stern (1991) utilized forward modeling to determine the mechanical properties of the western passive margin of North Island and recovered the $T_e$ as 20–28 km, which basically agrees with the $T_e$ values of 13–25 km in this study and reveals moderate lithospheric strength. Stern et al.
(1992) subsequently found that the suitable $T_c$ value is approximately 10 km in the adjacent Wanganui Basin, which is lower than that derived in the present study (25 km in $T_c$); the difference can be attributed to the low $T_c$ feature of the overlying Australian Plate captured by the algorithm in the previous study. The Wanganui Basin lies in the Hikurangi subduction zone; compared with the overlying Australian Plate, the subducting Pacific Plate contributes more mechanical strength because of interplate coupling (Pérez-Gussinyé et al., 2008), which results in the larger lithospheric strength observed in Fig. 4. Moreover, Hackney et al. (2012) modeled the flexural rigidities in the New Caledonia Basin and achieved a constant $T_c$ of 10 km. Richards et al. (2018) recovered the lithospheric rigidity of the Tasmanid Seamounts and exhibited that the inverted $T_c$ values vary from 0 km and 11 km. These $T_c$ values are consistent with our results, but most of these studies generated a constant $T_c$ value, rather than spatial variations, in their study regions.

In the Louisville Ridge, Cazenave and Dominh (1984) first estimated the variations in $T_c$ along the ridge using gravity and bathymetry data, showing that it gradually increases from 15.5 km in the southeast to 22.5 km in the northwest and significantly contradicting the study results of Watts et al. (1988). Then, based on the 3-D method, Lyons et al. (2000) recalculated the variations in $T_c$ along the Louisville Ridge. Their results are consistent with those of Cazenave and Dominh (1984) and suggest that the contradiction in the flexural rigidities may have resulted from the dimensionality of the mountain and the quality and spatial coverage of data input (Watts et al., 2006). In a later study, Contreras-Reyes et al. (2010) used the 2-D velocity structure of the crust that intersects the chain at 27.6°S in the northern part as a constraint to calculate the lithospheric flexural rigidities ($T_c=10\pm2$ km), which is similar to the result (6 km in $T_c$) estimated by Hwang and Kim (2016) but lower than those of previous studies (>20 km in $T_c$, Cazenave and Dominh, 1984; Lyons et al., 2000; Watts et al., 2006). Based on the relationship between $T_c$ and the age of the oceanic lithosphere when loading occurs, Contreras-Reyes et al. (2010) speculated that the larger estimates of the previous studies correspond to the $T_c$ feature of the outer rise, instead of that of the Louisville Ridge. Our results show that the low $T_c$ values (8–15 km) in the southeast agree with those in previous studies (Cazenave and Dominh, 1984; Lyons et al., 2000; Watts et al., 2006). However, unfortunately, we could not assess the $T_c$ in the northwest part of the Louisville Ridge (Contreras-Reyes et al., 2010; Hwang and Kim, 2016) due to the biased band of high $T_c$ values.

Fig. 8. The re-estimated spatial variations in the $T_c$ values after removing the effects of the dynamic topography and gravity anomaly calculated in Fig. 7a and Fig. 7b from the observed topography and the complete Bouguer gravity anomaly.

When studying the strength of the incoming oceanic plate in the Tonga-Kermadec subduction zone, Billen and Gurnis (2005) and Arredondo and Billen (2012) found that the lithospheric strength decreases obviously (>15 km in $T_c$) near the trench axis and ascribed this decrease to the plate changing from elastic to viscous deformation. Many studies on the flexure of global subduction zones, including the Tonga-Kermadec subduction zone, have been conducted and found a similar reduction feature when approaching the trench axis (Hunter and Watts, 2016; Zhang et al., 2014, 2018, 2019). Several researchers (Zhang et al., 2014; Zhou et al., 2015) suggested that the reduction in $T_c$ is primarily controlled by bending-related faulting. Nevertheless, unlike these study results, our results for $T_c$ in the trendward plate are opposite, exhibiting a band of high $T_c$ values (22–36 km) in the Tonga-Kermadec subduction zone, which is consistent with the results of some other studies (Kalnins, 2011; Hu et al., 2016). We present a detailed discussion of this contradiction in Section 5.1 and suggest that it is result of the interference of non-isostatic factors that affect our flexural model.

In general, except for the biased Tonga-Kermadec subduction zone, compared with those of previous studies, our $T_c$ map provides detailed information on $T_c$ variations over a larger region, with the exception of the biased values for the ocean-ocean subduction zones (Fig. 5c), instead of fixed values of flexural rigidity within a limited area. Our $T_c$ map yields important insights into the lithospheric deformation in this study region.
5. Discussion

5.1. Subduction zones

To date, several studies have suggested that the strength of a subducting oceanic plate is weakened as the plate approaches the trench axis. Related studies using numerical modeling show that significant extensional faults related to the downward deflection of the incoming plate form at the outer rise (Zhou et al., 2015; Zhang et al., 2018). Along the Tonga-Kermadec subduction zone, many trench-parallel normal faults can be found in the downgoing Pacific Plate (Fig. S2 in the Supplementary Material), which could reduce its mechanical strength (Ranero et al., 2003; Zhang et al., 2014; Garcia et al., 2019). In addition, back-arc spreading and earthquakes generally accompany the subduction of a plate, such as in the Lau Basin (Taylor et al., 1996). Therefore, all of the above geological events seem to support the idea of reduced integrated strength over the subducting plate. Nevertheless, this idea contradicts our results (Fig. 4), which show that the subducting Pacific Plate is characterized by higher $T_e$ (22–36 km) values and old oceanic crust (Fig. 1, Müller et al., 2008) in the Kermadec subduction zone.

It is worth noting that in this study, the band of high $T_e$ values dominates in the Tonga-Kermadec subduction zone and abruptly disappears at the Hikurangi Plateau, exactly where the extremely steep topography and gravity anomaly terminate, which are the key input data to estimate $T_e$ values. Accordingly, we hypothesize that the steep topography and gravity features along the trench may be the cause for the band of high $T_e$ values. We can see that the bathymetry (Fig. 1) in the Tonga-Kermadec Trench, with a maximum of ~10 km, is 5 km deeper than that in the adjacent abyssal plains in the east, and this difference disappears in the Hikurangi Plateau. Another good example is the Himalayan-Tibetan orogeny studied by Chen et al. (2015) who used the same method employed in the present study. In this orogeny related to the subduction of Indian Plate, steep features of topography and gravity are absent, and no anomalous band of extremely high $T_e$ values appears. In addition, in studying the lithospheric strength over the steep subducting plate, Contreras-Reyes and Osses (2010) utilized the finite difference method to solve the flexural equation and successfully found a reduced-$T_e$ feature along the Chile trench. In their flexural model, the vertical force must be added at the trench axis to ensure that the calculated profiles fit the observed profiles (Hunter and Watts, 2016; Zhang et al., 2014, 2018, 2019), which is a prominent difference from Forsyth’s (1985) theory used in this study. In our flexural model, gravity and topography are expressed as the flexural isostatic response of the elastic lithosphere to the surface and subsurface loadings without being affected by other extra forces. Malservisi et al. (2003) studied the origin of the observed high elevation and Bouguer gravity of Fiordland, which is subduction margin along the South Island, and found that they are controlled by plate boundary forces that provide non-isostatic support and generate the current elevated topography. Therefore, we further propose that in the Tonga-Kermadec subduction zone, lithospheric bending due to the pulling force introduces the non-isostatic factors to the anomalously steep bathymetry and gravity anomaly, resulting in the band of high $T_e$ values obtained when using the approach of Forsyth (1985). This explanation also applies to the high $T_e$ values recovered using the admittance method of Kalninns and Watts (2009). For the southern Hikurangi subduction zone, because of the absence of sharp variation of the bathymetry and gravity anomaly in the subducting Hikurangi Plateau, the inverted $T_e$ values are unbiased.

5.2. Arcs and oceanic basins in northern Zealandia

The values of $T_e$ in the parallel arcs and oceanic basins between the Tasman Sea in the west and the South Fiji Basin in the east change little, averaging ~9 km and suggesting that the entire lithosphere is very weak. The tectonic boundaries between these basins can barely be seen based on the $T_e$ results with a larger spatial resolution (Fig. 4a). The earlier flexural studies demonstrate that the $T_e$ values in the oceanic lithosphere are mainly determined by the plate and load ages (Watts, 1978; Watts and Burov, 2003). The low $T_e$ values observed in the young Tasman Sea and South Fiji Basin (Fig. 1) are easy to understand. Although the crustal type of the New Caledonia Basin is uncertain (Uruski and Wood, 1991; Sutherland, 1999), the multiple phases of crustal thickening that happened here (Klingelhoefer et al., 2007; Sutherland et al., 2010) are likely to have weakened the underlying lithosphere. Watts and Burov (2003) suggested that stable continental blocks are generally characterized by high $T_e$ values. Additionally, previous studies pointed out that variations in $T_e$ reveal different tectonic provinces (Mao et al., 2012; Pérez-Gussinyé et al., 2009; Chen et al., 2017). The remaining ridges are interpreted as thick continent (Schreckenberger et al., 1992; Klingelhoefer et al., 2007), such as the Lord Howe Rise, New Caledonia and Norfolk Ridge, and they have different evolutionary histories from that of the thin New Caledonia Basin. However, Fig. 4a and b show that these ridges are dominated by uniformly low $T_e$ values as well, which lead us to hypothesize that they underwent a tectonic event during the Cenozoic time after the Gondwana breakup that decreased the lithospheric strength broadly. $T_e$ studies show that a series of tectonics, such as rifting, hot spot activity and continental breakup, could contribute to the low $T_e$ values in local areas (e.g., Pérez-Gussinyé et al., 2009; Ratheesh-Kumar et al., 2015). We speculate that a tectonic event decreased the lithospheric strength on a large regional scale, resulting in widespread low $T_e$ values across the entire region of arcs and oceanic basins in northern Zealandia.

A model of lithospheric delamination associated with the Cenozoic inception of Tonga-Kermadec subduction has been proposed based on a previous study of seismic reflection and rock data (Sutherland et al., 2010), and it can explain the evidence of the observed tectonics, including the Eocene-Oligocene uniformity beneath the New Caledonia Basin. The key consequence of this delamination involves the removal of the lower crust, which reduces the thickness of the lithosphere and its integrated mechanical strength over arcs and oceanic basins in northern Zealandia. Moreover, although the entire Zealandia experienced Late Cretaceous Gondwana extension and crustal thinning, the region of arcs and oceanic basins is the only place where Paleogene subduction occurred (Sutherland et al., 2010; Mortimer et al., 2017). Therefore, the delamination model coincides with the regional estimated low $T_e$ values and can be applied to the northern arcs and oceanic basins. Such large-scale lithospheric delamination seems most likely to be the reason why $T_e$ values vary little between these blocks. The lithospheric thickness is the primary factor controlling the lithospheric strength in this region.

5.3. New Zealand

The collision and strip slipping of the Australian Plate and Pacific Plate created significant lithospheric deformation in New Zealand (Beavan et al., 2002; Scherwath et al., 2003), and the surrounding submerged plateaus experienced crustal thinning (Groby et al., 2007) and high-temperature metamorphism (McFadden et al., 2010; Schulte et al., 2014) upon separating from West Antarctica. These tectonics indicate that the lithosphere over this region is weak and that the variations in $T_e$ are small. However, changes in the lithospheric rigidity can still be seen in Fig. 9a, and they correlate well with the main geological features.

High lithospheric strength dominates North Island and the southernmost part of South Island, where the plate boundaries are characterized by subduction (Nicol et al., 2007), while the remaining area of New Zealand displays low lithospheric strength. Moreover, the distinct ribbon of low $T_e$ values extends along the plate boundary, which is marked by oblique slipping (Walcott, 1998; Beavan et al., 2002). The $T_e$ values in New Zealand (Fig. 9) appear to be related to the boundary types: larger strength is apparent over the overlying plate in the subduction zone and lower strength dominates along the area of oblique slipping. Pérez-
Gussinyé et al. (2008) noted that during subduction the contact between the overlying and subducting plates may make the estimated $T_e$ values reflect the total strength of both lithospheric plates. The TVZ (Fig. 9) in the central North Island is characterized by a thinned crust of 15–20 km (Stratford and Stern, 2006), contains $> 10,000 \text{ km}^3$ of magma (Wilson et al., 1995) and exhibits high heat flow (Bibby et al., 1995). Pulford and Stern (2004) interpreted the distinct surface uplifting in central North Island to be the underlying thermal anomaly. These geophysical features indicate that the lithospheric strength of North Island is not as strong as observed in this study. Considering that subduction occurs along the Hikurangi Plateau and Alpine fault, concluding that the observed relatively high $T_e$ values in North Island and the southernmost part of South Island is a consequence of the combined lithospheric strength of the upper continental and lower oceanic plates is reasonable. The fine structure of the merged plates across the Hikurangi Trench can be imaged in the 3-D velocity structure (Reyners et al., 2006; Eberhart-Phillips et al., 2010). As described above, the mechanical strength of the lithosphere is governed by its thickness and thermal structure. The oceanic crust in the Hikurangi Plateau is much thicker ($\approx 15 \text{ km}$) than normal oceanic crust (Davy et al., 2008; Sutherland et al., 2009). Furthermore, the oceanic age here is also the oldest (120–130 Ma, Müller et al., 2008, Fig. 1) along the Tonga-Kermadec-Hikurangi trench system, and the age is $> 70 \text{ Ma}$ along the subduction zone of the southernmost part of South Island; these features appear to favor a high mechanical strength of the subducting oceanic plate. Consequently, we speculate that the relatively high estimated $T_e$ values in North Island and in the southernmost part of South Island result mainly from the subducting plate having a high integrated strength.

For the remaining part of South Island, the low $T_e$ values similar to those of the surrounding submerged plateaus indicate that the lithosphere there is weak. This interpretation is consistent with the low-resistivity zone detected by Wannamaker et al. (2002) and reduced seismic velocity anomalies (Van Avendonk et al., 2004; Eberhart-Phillips and Bannister, 2002; Scherwath et al., 2003). The strike-slip and collision of continental blocks created extensive strike-slip faults and reverse faults that significantly decreased the lithospheric strength (Burov et al., 1998) and shortened the crust beneath the Southern Alps (Norris et al., 1990; Walcott, 1998), indicating that intense lithospheric deformation occurred in this region. As such, the low $T_e$ values in the central and northern South Island are controlled by tectonic activity, i.e., oblique collision.

6. Conclusions

We estimate new $T_e$ maps for Zealandia based on the fan wavelet coherence method. To consider the spatial resolution of the recovered $T_e$ values and their accuracy, two different central wavenumbers, $|k_0| = 3.081$ and 5.336, are applied (Kirby and Swain, 2011). The analysis of the $T_e$ variations leads us to reach the following key findings:

1. The high-$T_e$ zone (22–36 km) occurs along the Tonga-Kermadec subduction zone. We interpret our results to indicate that the topographic expression is disturbed by the non-isostatic factors in the subduction zone, causing steep bathymetry and gravity anomalies. The contaminated coherence between these parameters creates the biased high-$T_e$ zone. Therefore, coherence method should be used with caution in the subduction zone.

2. We speculate that the relatively high strength in North Island and the southernmost part of South Island may represent the combined lithospheric strength of the Pacific Plate and Australian Plate. The low $T_e$ values of approximately 15 km in central and northern South Island are the results of the strike-slippping and collision of continental blocks.
Largely uniform T values (6–9 km) are observed beneath the arcs and oceanic basins in northern Zealandia, and variations can barely be seen between these blocks, indicating that the whole lithosphere is weak. Along with the area of arcs and oceanic basins being the only place in Zealandia where Paleogene subduction initiated, our results are consistent with a lithospheric delamination model (Sutherland et al., 2010) associated with the broad removal of the lower crust. The reduced lithospheric thickness on a large regional scale may explain the uniformly low lithospheric strength.

Credit author statement
Fei Ji prepared, processed and interpreted the data, made the figures. Qiao Zhang conceived the study, proposed the main idea and jointly wrote the manuscript. Xin Zhou, Yong Liang Bai and Yong Dong Li assisted with the data processing. The rest co-authors, Xia Zhou, Yong Liang Bai, Y., Dong, D., Kirby, J.F., Williams, S.E., Wang, Z., 2018. The effect of dynamic topography and gravity on lithospheric effective elastic thickness estimation: a case study. Geochem. Geophys. Geosyst. 19, Q01005.

Appendix A. Supplementary data
Supplementary data can be found online at https://doi.org/10.1016/j.gr.2020.05.008.

References