Effective Elastic Thicknesses of the Lithosphere and Mechanisms of Isostatic Compensation in Australia

MARIA T. ZUBER
Geodynamics Branch, NASA Goddard Space Flight Center, Greenbelt, Maryland

TIMOTHY D. BECHTEL
Kurz Associates, Providence, Rhode Island

DONALD W. FORSYTH
Department of Geological Sciences, Brown University, Providence, Rhode Island

The isostatic compensation of the Australian lithosphere is examined using the coherence of the two-dimensional Fourier transforms of Bouguer gravity and topography. The isostatic model assumes regional compensation of an elastic plate that undergoes flexure in response to surface and subsurface loading. A least squares inversion of the observed coherence for continental Australia yields an effective elastic thickness of 130 km (corresponding flexural rigidity of $2.0 \times 10^{25}$ N m). The model indicates that loads with wavelengths of greater than 1500 km are locally compensated, loads with wavelengths in the range 600-1500 km are partially supported by the strength of the lithosphere, and loads with wavelengths of less than 600 km are almost completely supported by lithospheric strength. Regions within the continent with different ages and tectonic histories exhibit effective elastic thicknesses that increase with the time since lithospheric stabilization. Precambrian shields in western, northern, and southern Australia are well fit by an elastic plate with a thickness of the order of 100 km. The Phanerozoic Interior Lowlands have elastic thicknesses that range from 30 to 80 km, and the Eastern Highlands have thicknesses of 15-35 km. The theoretical model of predicted coherence, which assumes statistically independent surface and subsurface loads, does not provide a good fit to the coherence spectrum for the intracratonic basins of central Australia. The disagreement of the model and observations is most simply explained by statistically correlated surface and subsurface loads. Calculations of load distributions show that most of the elevation of the Western Shield can be represented as a locally compensated plateau with wavelengths greater than 1000 km; smaller-amplitude, shorter-wavelength relief appears almost entirely as surface loading. Approximately half of the topographic amplitude of the Southeastern Highlands is locally compensated, and both surface and subsurface loading contribute significantly to the regionally compensated topography. The low rigidity and subsurface loading in southeastern Australia are consistent with crustal underplating, igneous intrusion, and thermal perturbation of the upper mantle as mechanisms for the uplift and compensation of the highlands.

INTRODUCTION

Isostasy describes the manner in which surface topography is compensated by the subsurface mass distribution. In a flexural isostatic model [Barrell, 1914; Vening Meinesz, 1941], topographic features are treated as loads on a thin elastic plate underlain by a weak fluid, and compensation occurs on a regional basis because loads are partially supported by the lateral strength of the lithosphere. The extent to which the lithosphere can support loads is conveniently characterized by the flexural rigidity or, equivalently, the effective elastic thickness of the plate. Alternatively, in an Airy or local isostatic model [Airy, 1855], compensation is accomplished by thickening of a constant density crust. An Airy-compensated lithosphere has no finite strength and corresponds to the flexural model in the limit of zero rigidity or elastic thickness.

Transfer function relationships between topography and gravity have commonly been employed to investigate quantitatively the isostatic compensation of the lithosphere [e.g., Lewis and Dorman, 1970; Walcott, 1970; Watts and Cochran, 1974; McKenzie and Bowin, 1976; Banks et al., 1977; McNutt and Parker, 1978; McNutt, 1980, 1983; Watts et al., 1980]. Of particular note is the admittance or isostatic response function, which is the ratio of the Fourier transforms of gravity and topography [Dorman and Lewis, 1970]. The observed admittance, when compared to the admittance predicted by a flexural model of a thin elastic plate, provides an estimate of the flexural rigidity. Models employing this approach have been successful in describing the mechanical behavior of oceanic lithosphere [e.g., McKenzie and Bowin, 1976; Watts et al., 1980; Bodine et al., 1981], however, Forsyth [1981, 1985] has noted that continental flexural rigidities determined from the admittance ($10^{19}$-$10^{22}$ N m [Banks et al., 1977; McNutt and Parker, 1978; Cochran, 1980]) are consistently lower than those obtained from modeling individual features ($10^{23}$-$10^{24}$ N m) such as sedimentary basins [e.g., Watts et al., 1982], lakes [Passey, 1981; Nakiboglu and Lambeck, 1983; Bills and May, 1987], and mountain ranges [Karner and Watts, 1983]. For example, the flexural rigidity of Australia estimated from the admittance, assuming a model of an elastic plate loaded from...
above, is less than $5.0 \times 10^{20}$ N m [McNutt and Parker, 1978]. This corresponds to an effective elastic thickness of less than a kilometer and indicates a state of Airy isostasy. Since two-thirds of Australia consists of Precambrian shields (Figure 1a), this result would imply, contrary to intuition, that old, stable continental lithosphere does not have any appreciable elastic strength.

More recently it has been shown that continental flexural rigidities determined from the admittance can be greatly underestimated if loads are located within or beneath the plate instead of on the top [Louden and Forsyth, 1982; McNutt, 1983; Forsyth, 1985] or if provinces with differing rigidities are averaged together [Forsyth, 1985]. The systematic bias toward low rigidity when both surface and subsurface loading are present is the dominant of these effects [Forsyth, 1985].

Australia consists of numerous tectonic provinces (Figure 1a) that reflect an eastward progression of tectonic activity that ceased as early as the Proterozoic in the west and as late as the Mesozoic in the east [Plumb, 1979a]. The topography of the continent is subdued (Figure 2a). Forty percent of the surface area is at an elevation of less than 200 m [Palfreyman, 1984], and the dominant topographic expression occurs in a narrow band along the east coast. Gravity anomalies (Figure 2b) in many areas reflect crustal density variations [Dooley, 1977; Wellman, 1978, 1979a, 1982] as opposed to crustal thickness variations. On the basis of these observations, the unusually low flexural rigidity of Australia determined from the admittance may reflect (1) the inclusion in the study area of regions with different flexural rigidities, (2) improper weighting of the contributions of the flat, rigid
Fig. 1b. Map of effective elastic thicknesses (in kilometers) for tectonic subregions in Australia. Subregion abbreviations and parameters assumed in the least squares inversions are given in Tables 1 and 2.

Data

Two separate gravity/topography data sets were used in the analysis. The raw data in the primary data set consisted of approximately 240,000 randomly spaced elevation and gravitational acceleration values determined from land-based surveys. These data were interpolated to form gridded arrays of topography and gravity by fitting a biquadratic surface to a minimum of 10 measurements within a spherical cap surrounding a prescribed grid point. The radius of the cap was determined by performing a singular value decomposition of the normal matrix of points. The final data sets consisted of 360 × 463 equal-area Cartesian grids in which points were spaced at 10 km. Bouguer anomalies were computed at each grid point assuming a crustal density of 2670 kg m⁻³. Because gentle topography characterizes nearly all of continental Australia, terrain corrections were not applied. The second data set consisted of 0.1° × 0.1° latitude/longitude grids of corresponding elevations and Bouguer anomalies compiled by the Bureau of Mineral Resources, Canberra, ACT, Australia. This supplemented the primary data set because it included data for the continental shelves and was therefore particularly useful in isolating tectonic provinces along the coasts. Tectonic subregions common to both data sets yielded similar elastic
thicknesses, indicating that grid distortions associated with the different coordinate representations were not significant.

The gridded data were Fourier transformed to obtain the spectral amplitudes of Bouguer gravity and topography. In order to suppress edge effects while permitting consideration of the longest possible wavelengths, the data were mirrored in the north-south and east-west directions prior to transforming to produce data sets 4 times the size of the original data sets. The longest wavelengths in the Fourier transform representation of the data thus exceed the dimensions of the original study areas and were excluded from the inversions for the elastic thicknesses.

**FORMULATION**

**Observed Coherence**

The powers of topography \(E_0(k)\), Bouguer gravity \(E_1(k)\) and the complex cross spectrum \(C(k)\) were used to calculate the observed coherence. These quantities are expressed

\[
E_0(k) = \langle H(k) \cdot H(k)^* \rangle \\
E_1(k) = \langle G(k) \cdot G(k)^* \rangle \\
C(k) = \langle G(k) \cdot H(k)^* \rangle
\]
where $H$ and $G$ are the Fourier amplitudes of topography and gravity, $k = (k_x, k_y)$ is the two-dimensional wave number which is related to the wavelength $\lambda$ by $|k| = 2\pi/\lambda$, the angle brackets indicate averaging over discrete wave number bands with average value $\bar{k}$, and the asterisk indicates the complex conjugation. The observed coherence ($\gamma_0^2$) is defined as

$$\gamma_0^2 = \frac{C^2(k)}{E_0(k)E_1(k)}$$  \hspace{1cm} (4)

[McKenzie and Bowin, 1976], however, in the presence of noise an unbiased estimate of the coherence is obtained from

$$\gamma^2 = \frac{n\gamma_0^2 - 1}{n - 1}$$  \hspace{1cm} (5)

[Munk and Cartwright, 1966], where $n$ is the number of discrete wave numbers in a wave band. Note that if a wave band contains a small number of points and the coherence is small, $\gamma^2$ can be negative. The standard errors of the coherence were determined from

$$\Delta\gamma^2 = (1 - \gamma_0^2)(2\gamma_0^2/n)^{1/2}$$  \hspace{1cm} (6)

[Bendat and Piersol, 1980].

The coherence represents the fraction of the energy in the gravity field that can be predicted by a linear filter operating on the surface topography. At long wavelengths, surface or subsurface loads deflect the lithosphere creating gravity or topographic anomalies representative of the compensation of the loads. Thus at long wavelengths, Bouguer gravity and topography are correlated, and the coherence approaches unity. In contrast, at short wavelengths, loads are supported by lithospheric strength and do not deflect the lithosphere. In the short-wavelength limit, gravity and topography are uncorrelated so the coherence approaches zero (if surface and subsurface loading are statistically independent). The wavelength of the transition between coherent and incoherent wave bands indicates the flexural rigidity or elastic thickness of the lithosphere. A thicker or more rigid lithosphere undergoes the transition from high to low coherence at a longer wavelength than a thinner or less rigid lithosphere [Forsyth, 1985].

**Predicted Coherence**

Flexural rigidity ($D$), or equivalently, effective elastic thickness ($T$, where $D = ET^3/12(1 - \nu^2)$), was determined by performing a least squares fit, over a range of wave bands, of the coherence predicted from a flexural model of surface and subsurface loading of an elastic plate to the observed coherence [cf. Forsyth, 1985]. In the determination of the predicted coherence, it was assumed that loading occurred at the surface and a deep subsurface interface and, in certain cases, also at a shallow subsurface interface. For the more general case of three interfaces, relief at the surface ($H$), the shallow subsurface interface ($V$), and the deeper subsurface interface ($W$) was assumed to contain components resulting from loading at all of these interfaces such that

$$H = H_t + H_m + H_b$$  \hspace{1cm} (7)

$$V = V_t + V_m + V_b$$

$$W = W_t + W_m + W_b$$

where the subscripts $t$, $m$, and $b$ refer to load components at the top, middle, and bottom interfaces.

The amplitudes of the harmonic relief ($U(k)$) at the subsurface interfaces were determined by downward continuation of the Bouguer gravity and the slope of the power of the relief, where the relief at both subsurface interfaces was assumed to have the same spectrum as the surface topography [Bechtel et al., 1987]. The components of relief were isolated using the Fourier transform of the thin plate flexural equation

$$Dk^4U(k) + \rho_mgU(k) = q(k)$$  \hspace{1cm} (8)

where $\rho_m$ is the density of the mantle, $g$ is the acceleration of gravity, and $q(k)$ is the applied load.

At wave numbers where $Dk^4 \neq 0$ (i.e., at wavelengths where the rigidity of the lithosphere supports some or all of the loads) the system of equations for the surface and subsurface loads has a unique solution [Forsyth, 1985]. If, on the other hand, $Dk^4 = 0$, then the interface at which a load occurs is indeterminate because the topography is locally compensated. In other words, the topography created by a compensating load at another interface cannot be distinguished from the compensated relief resulting from a surface topographic load.

Assuming that the loads are statistically independent or uncorrelated, the predicted coherence ($\gamma_p^2$) expressed in terms of the Fourier components of the loads was determined from [Bechtel et al., 1987]

$$\gamma_p^2 = \frac{\langle aH_tW_t + \beta H_tV_t + \alpha H_mW_m + \beta H_mV_m + \alpha H_bW_b + \beta H_bV_b \rangle^2}{\langle H_t^2 + H_m^2 + H_b^2 \rangle \langle W_t^2 + W_m^2 + W_b^2 \rangle}$$

\hspace{1cm} (9)

where $a = 2\pi(\rho_m - \rho_c) \exp(-\kappa_z z_m)$, $\beta = 2\pi(\rho_c - \rho_b) \exp(-\kappa_z z_m)$, $\rho_m$, $\rho_c$, $\rho_b$ are the densities of the cover, crust, and mantle, and $z_c$ and $z_m$ are the depths to basement and the Moho, respectively. The assumption of statistical independence of the loads was tested on a case-by-case basis by determining the correlation between surface and subsurface loading over the range of wave bands. The assumption was found to be appropriate for all of the regions that we examined except one; that case is detailed in the discussion section.

The predicted coherence is dependent on a number of assumptions, e.g., the ratio of surface to subsurface loading, but is most sensitive to the flexural rigidity. For example, Forsyth [1985] calculated the effective elastic thickness for East Africa assuming equal amplitudes of surface and subsurface loading in determining the predicted coherence and found a best fit solution of $T = 25$-30 km. Bechtel et al. [1987] added for the analysis of East Africa using the gravity and topography fields to calculate the loads and obtained the same result ($T = 25$ km).

The elastic plate model assumed in the determination of the predicted coherence is characterized by a smooth and rapid transition from high to low coherence. In practice, however, the observed transition may be broad or show considerable scatter, indicating that a region is not uniformly rigid [Forsyth, 1985]. In this situation the result represents an average of subregions with different flexural rigidities weighted by the product of the powers of the topography and gravity anomalies. When this occurs, it is desirable to
subdivide the area further to isolate a region with uniform rigidity. However, for a rigid continent it is necessary to examine a relatively large area in order to recover long enough wavelengths (i.e., with coherence near one) to constrain the rigidity. For Australia many tectonic provinces were too small to calculate individual flexural rigidities. The rationale for choosing the sizes of subregions is discussed in greater detail in the appendix.

**Interpretation of Surface and Subsurface Loads**

On the basis of geological and geophysical knowledge of the average density structure of Australia, the subsurface interfaces \( V \) and \( W \) were chosen to correspond to basement and the Moho, respectively. The depths to these interfaces were determined from gravity and seismic data. Specifically, the depth \( z_c \) to the shallow interface \( V \) was determined from the slope of the shortest wavelength piecewise linear segment of the log gravity power spectrum [Karner and Watts, 1983; Bechtel et al., 1987], while the depth \( z_m \) to the bottom interface \( W \) was taken to be the crustal thickness determined from seismic refraction studies [Muirhead et al., 1977; Wellman, 1982; Finlayson and Mathur, 1984; Lambeck and Penney, 1984]. While we could have estimated \( z_m \) from the slope of longer-wavelength piecewise segments of the gravity power spectrum, we opted for the approach taken because it was possible to obtain reliable estimates of crustal thickness for all major tectonic subregions from the published literature. For several cases we did calculate \( z_m \) from the gravity power spectra and found results in agreement with published crustal thickness data. Values of \( z_c \) and \( z_m \) used in the determination of the effective elastic thickness for continental Australia and the tectonic subregions are listed in the caption of Figure 3 and in Table 2, respectively, while the depths used in determining the components of surface topography are given in the captions of Figures 6 and 7. The assumed densities of the cover, crust, and mantle are listed in Table 1. Although the inferred loads are sensitive to the assumed densities and thicknesses of the layers, the inferred flexural rigidity is insensitive to these assumptions.

The representation of subsurface loading as occurring on two interfaces is simply a mathematical convenience that allows for both shallow crustal and deeper loading and is not intended to imply that these interfaces are necessarily physical entities. Loads at the shallow interface may represent any subsurface density variations, such as crustal blocks of different compositions, sedimentary basins, or igneous intrusions. The deeper interface may represent loads in the form of crustal underplating, lithospheric thermal anomalies, or deeper compositional variations. Extreme caution should be exercised in the interpretation of the inferred loads, which are sensitive to the assumed depth of the interface, and should be modeled, as in any other downward continuation problem, in light of as many other geological and geophysical constraints as possible.

A shallow subsurface interface was included to characterize better the subsurface structure in areas that contained marked variations in density or for numerical stability in the determination of subsurface loads. In the latter case, downward continuation of the high-frequency component of the Bouguer gravity signal to shallow depths prevents instabilities due to noise in the signal that is progressively amplified with increasing depth [McNutt, 1983]. Because the coherence is relatively insensitive to the density structure, models with and without a shallow interface yielded similar estimates of effective elastic thickness. In the model formulation for a lithosphere loaded at the surface and a single, deep subsurface (Moho) interface, terms containing \( V \) and the subscript \( m \) vanish in equations (7) and (9).

**Continental Australia**

In order to estimate the average elastic behavior of the Australian lithosphere, we first analyzed several regions that
TABLE 2. Subsurface Interface Depths, Effective Elastic Thicknesses, and Flexural Rigidities for Tectonic Subregions

<table>
<thead>
<tr>
<th>Tectonic Subregion</th>
<th>$z_{cr}$ km</th>
<th>$z_{cm}$ km</th>
<th>$T_{m}$ km</th>
<th>$D_1$ N m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Western Shield (WS)</td>
<td>...</td>
<td>35</td>
<td>132</td>
<td>$2.0 \times 10^{25}$</td>
</tr>
<tr>
<td>Northern Craton/Northeast Orogens (NCNO)</td>
<td>...</td>
<td>37</td>
<td>92</td>
<td>$6.9 \times 10^{24}$</td>
</tr>
<tr>
<td>Nullarbor Block/Gawler Craton (NBGC)</td>
<td>1.6</td>
<td>35</td>
<td>88</td>
<td>$6.1 \times 10^{24}$</td>
</tr>
<tr>
<td>Central Blocks and Basins (CBB)</td>
<td>...</td>
<td>37</td>
<td>76</td>
<td>$3.9 \times 10^{24}$</td>
</tr>
<tr>
<td>Eromanga Basin (EB)</td>
<td>...</td>
<td>37</td>
<td>23</td>
<td>$1.1 \times 10^{23}$</td>
</tr>
<tr>
<td>Murray Basin/Southern Lachlan Fold Belt (MBFB)</td>
<td>...</td>
<td>37</td>
<td>29</td>
<td>$2.2 \times 10^{23}$</td>
</tr>
<tr>
<td>Murray and Darling basins (MDB)</td>
<td>...</td>
<td>40</td>
<td>17</td>
<td>$4.4 \times 10^{22}$</td>
</tr>
<tr>
<td>Southern Lachlan Fold Belt (SLFB)</td>
<td>0.7</td>
<td>48</td>
<td>17</td>
<td>$4.4 \times 10^{22}$</td>
</tr>
<tr>
<td>New England Fold Belt (NEFB)</td>
<td>...</td>
<td>40</td>
<td>16</td>
<td>$3.6 \times 10^{22}$</td>
</tr>
<tr>
<td>Surat Basin (SB)</td>
<td>...</td>
<td>37</td>
<td>52</td>
<td>$1.3 \times 10^{24}$</td>
</tr>
<tr>
<td>Bowen and Galilee basins (BGB)</td>
<td>...</td>
<td>37</td>
<td>64</td>
<td>$2.3 \times 10^{24}$</td>
</tr>
<tr>
<td>Bowen Basin/Northern Lachlan Fold Belt (BBFB)</td>
<td>...</td>
<td>37</td>
<td>26</td>
<td>$1.6 \times 10^{23}$</td>
</tr>
</tbody>
</table>

The tectonic development of Australia commenced in the Western Shield (WS) during the Archaen and progressed eastward through time. The primary tectonic subdivisions in the west are the Pilbara and Yilgarn blocks. These areas formed about 3500 Ma and 3100 Ma, respectively, during the period of initial crustal differentiation in Australia, and stabilized prior to 2500 Ma [Palfreyman, 1984]. Surrounding the blocks are Proterozoic orogenic belts that are overlain by a relatively undisturbed Proterozoic and Phanerozoic platform cover [Gee, 1979]. The least squares inversion of the observed and predicted coherence for the Western Shield indicates an elastic thickness of 132 km. The coherence pattern exhibits a smooth, sharp transition from high to low values (Figure 4), which indicates that this region responds to loading in the manner of a uniform plate.

Northern Craton/Northeast Orogens

The Northern Craton is composed of early Proterozoic blocks bounded by mid-Proterozoic orogenic belts and is overlain in places by a flat-lying, undeformed Paleoic-Mesozoic cover. The area has behaved as a single stable platform since approximately 1700 Ma [Plumb, 1979a]. Crustal thicknesses as determined from seismic refraction data [Wellman, 1982; Finlayson and Mathur, 1984] vary from about 30 km in the Arafura Sea on the northern continental shelf to over 50 km in the McArthur Basin and Tennant Creek-Mount Isa regions. However, in areas of greater crustal thickness the crust-mantle boundary is vertically gradational rather than distinct [Collins, 1983]. The Northeast Orogens developed adjacent to the Northern Craton and have been interpreted by Plumb [1979a, b] to be a younger extension of that province. The coherence for the combined Northern Craton and Northeast Orogens (NCNO) yields an elastic thickness of 134 km. This assumes an average crustal thickness of 40 km; however, because the coherence is relatively insensitive to the internal density stratification [cf. Forsyth, 1985], the observed major crustal thickness variations within the region do not significantly affect the estimate of elastic thickness.

Central Blocks and Basins

Central Australia consists of a series of parallel, E-W trending, intracratonic sedimentary basins separated by fault-bounded blocks of Proterozoic basement. The blocks have been uplifted and highly eroded since the Late Proterozoic [Lambeck, 1983], and as a consequence, sediments derived from the uplifted regions are more than 6 km thick. Forward gravity models [Mathur, 1976; Lambeck, 1983] and seismic travel time anomalies [Lambeck and Penney, 1984] indicate that the crust may be as much as 20 km thicker under the basins than the adjacent uplifted blocks, which has led to the suggestion [Lambeck, 1983; Lambeck and Penney, 1984; McQueen and Beaumont, 1987] and subsequent observation [Goleby et al., 1989] that the Moho is offset by deeply penetrating basin-bounding faults. The best fit elastic thickness of the Central Blocks and Basins (CBB) is 88 km. However, the coherence spectrum exhibits large standard
errors and a broad, irregular transition from high to low values, indicating that the area does not behave as a uniform elastic plate. We consider possible reasons for the poor model fit in the Discussion.

**Nullarbor Block/Gawler Craton**

The primary tectonic subprovinces in southern Australia are the Nullarbor Block and Gawler Craton. The Gawler Craton formed during the Late Archaen-Early Proterozoic over much the same time as the linear blocks of central Australia, but stabilized much earlier, at about 1400 Ma. The Nullarbor Block has been entirely covered by sediments that formed the Eucla Basin and is distinguished from adjacent tectonic provinces primarily on the basis of a distinct gravity pattern [Wellman, 1976]. This feature could either be Archaen in age or a western extension of the Gawler Craton.
The combined Nullarbor-Gawler region (NBGC) has an elastic thickness of 92 km, which is typical of other continental shields. This subregion displays reasonably uniform rigidity, as determined by the smooth falloff from high to low coherence.

**Interior Lowlands**

East of the Tasman Line, which divides Precambrian western Australia from Phanerozoic eastern Australia, lies the Interior Lowlands. This area comprises most of the Tasman Orogens shown in Figure 1a with the exception of a narrow region along the east coast. The region consists of Precambrian basement overlain by foreland basins that contain up to 6 km of folded and faulted elastic sediments. Deposition occurred during the west to east development of the Tasman fold belt system in mid- to late Paleozoic time [Doutch and Nichols, 1978; Schreibner, 1985]. The Interior Lowlands have been tectonically stable since the early Mesozoic, although marine and continental sediments have continued to accumulate in parts of the basins through the Tertiary [Doutch and Nichols, 1978]. The Eromanga Basin (EB), which is the largest and westernmost of the basins that comprise the Interior Lowlands, has an elastic thickness of about 76 km, which is similar to values obtained for the central United States [Bechtel, 1989]. Subregions containing surrounding smaller basins, i.e., the Murray and Darling basins (MDB), the Surat Basin (SB), and the Bowen and Galilee basins (BGB), have smaller elastic thicknesses that range from 30 to 65 km (Figures 1b and 4).

**Eastern Highlands**

The Eastern Highlands consist of a narrow band of erosional remnants of Paleozoic fold belts along Australia’s east coast. In the southeast, the major tectonic subdivision is the southernmost exposure of the Lachlan Fold Belt. Southeastern Australia is one of the most seismically active areas on the continent [Lambeck et al., 1984] and exhibits anomalously high heat flow [Cull and Denham, 1979; Sass and Lachenbruch, 1979; Cull, 1982] and high upper mantle conductivity [Lilley et al., 1981]. Xenolith geobarometry data [Wass and Hollis, 1983], in addition to heat flow and electrical resistivity data, indicate a high geothermal gradient indicative of a lower crustal thermal anomaly. Eastern Australia has undergone extensive Cenozoic volcanism [Wellman and McDougall, 1974a, b; Sutherland, 1978, 1981], with the most recent activity (<10 Ma) occurring in the southeast. The Southern Lachlan Fold Belt exhibits the highest topography in Australia, with a maximum elevation of over 2 km and a crustal thickness that in places exceeds 50 km [Wellman, 1982]. As in the Northern Craton, the Moho as determined from seismic refraction studies is gradational (over a 20 km depth) rather than distinct, and seismic refraction records show evidence for vertical velocity decreases within the crust that may indicate localized thermal or compositional variations [Finlayson and Mathur, 1984]. An area that encompasses parts of the Southern Lachlan Fold Belt and Murray Basin (MBFB) indicates an elastic thickness of 23 km, while a narrow coastal band that contains the Southern Lachlan Fold Belt alone (SLFB) has an elastic thickness of about 17 km.

Like the southeast, the central and northeastern sections of the Eastern Highlands are also characterized by intervening Paleozoic fold belts and basins and extensive Cenozoic volcanism. The volcanics show a progressive trend of increasing age toward the northeast that has been interpreted by Wellman and McDougall [1974a] to indicate the passage of eastern Australia over a hot spot since the late Cretaceous. Subregions containing outcrops of Paleozoic fold belt remnants in central and northeast coastal regions (NEFB and BBFB) have elastic thicknesses of 16 and 26 km.

**Discussion**

**Continental Scale Effective Elastic Thickness**

The effective elastic thickness of continental Australia determined from the least squares fit of the observed and predicted coherence (Figure 3) is about 130 km, which implies that old, stable continental lithosphere has considerable long-term strength. This result is in contrast to the findings of McNutt and Parker [1978], who determined Australia’s effective elastic thickness to be less than 1 km and concluded that long-wavelength elastic stresses in the continents viscously relax on geologic time scales. The reason for the discrepancy between the results is that McNutt and Parker [1978] employed the admittance approach, which underestimated the elastic thickness. The low value of $T$ obtained from the admittance indicates that both surface and subsurface loading are present in Australia and that tectonic subregions with varying flexural rigidities were averaged together in the estimation of the elastic thickness of the continent as a whole.

**Interpretation of Effective Elastic Thickness Variations**

Figure 1b shows that the elastic thickness of Australia is a minimum along the east coast (15–35 km), increases westward in the Paleozoic Interior Lowlands (30–80 km), and is greatest for the Proterozoic shields and orogenic belts and Archaen western Australia (>90 km). For Precambrian Australia the interpretation of the elastic thickness map is straightforward. The shields in the north, south, and west, which have been stable since at least the Proterozoic, all have average effective elastic thicknesses similar to those of Precambrian terrains on other continents. Moving eastward across the Tasman Line into Phanerozoic Australia, the overall decrease in elastic thickness can be most simply interpreted as reflecting a later (Paleozoic) time of lithospheric stabilization.

Elastic thickness variations within eastern Australia are more difficult to interpret. For example, Figure 1b suggests that the elastic thickness of the western part of the Interior Lowlands (EB) is greater than the east (BGB, SB, and MDB). This pattern may indicate a progressive decrease in elastic thickness toward the east, or an averaging of the elastic thicknesses of the basins and the much less rigid east coastal regions. The irregular transitions from high to low coherence that we observed for many different choices of subregions in central eastern Australia lead us to favor the latter interpretation.

The low elastic thicknesses along the east coast (15–35 km) may be explained by a thermally and tectonically active Phanerozoic history. Eastern Australia underwent several orogenic episodes that lasted through the Paleozoic and early Mesozoic [Plumb, 1979a]. Subsequently, the area experienced extensive erosion and deposition of sediments.
in foreland basins (summarized by Veevers [1984]). During the Mesozoic and Cenozoic, eastern Australia was punctuated with extrusive and intrusive volcanism associated with mantle thermal anomalies and offshore seafloor spreading [Wellman and McDougall, 1974a; Sutherland, 1978]. Southeastern Australia experienced significant uplift and elevated thermal gradients in the Cenozoic [cf. Wellman, 1979b; Stephenson and Lambeck, 1985b]. Evidence from lower crustal xenoliths [Ewart et al., 1980; Wass and Hollis, 1983] and seismic refraction data [Finlayson and Mathur, 1984] suggest that crustal underplating and (to a lesser extent) intrusion may have contributed up to as much as 40% (20 km) of the present crustal thickness in the southeast. Wellman [1979a] determined that southeastern Australia is close to isostatic equilibrium and is compensated by a low-density root directly beneath the highlands, which is consistent with our low apparent rigidity. Replacement of the upper mantle with 20 km of less dense ($\rho_{\text{mantle}} - \rho_{\text{root}} = 300 \text{ kg m}^{-2}$) mafic material could account for the approximately 1 km uplift of the highlands, accomplish isostatic compensation, and therefore explain the low effective elastic thickness of this region.

The rigidity along the east coast may also have been influenced by the thermomechanical structure of the adjacent continental margin. Karner and Watts [1982] found the margin in the vicinity of the Coral Sea in the northeast and Lord Howe Rise in the central east to be in a state of near-Airy isostasy ($T < 5$ km). They inferred the low effective elastic thickness to be a consequence of the relatively young age of the margin. Weisel and Karner [1984] suggested that rifting of the Tasman Sea Basin and associated heating of the lithosphere led to the Cenozoic uplift in southeastern Australia. Elastic thicknesses along the east coast of Australia are comparable to those found in the rift valley regions of East Africa (22–42 km [Bechtel et al., 1987; Ebinger et al., 1989]) and in parts of the western United States [Bechtel, 1989], both of which have been thermally active during the late Cenozoic.

The continuous elastic plate model, which assumes statistically independent surface and subsurface loads, does not provide a good fit to the observed coherence of central Australia. In an earlier study, Stephenson and Lambeck [1985a] found marked directional anisotropy in the admittance for central Australia and proposed that large admittances at intermediate wavelengths in the N-S direction could be explained if the lithosphere was treated as a viscoelastic plate subjected to N-S directed horizontal compression. Such a state of stress is suggested by mechanical models and geological information [Lambeck, 1983, and references therein; Lambeck et al., 1984]. Within the context of our model, an alternative explanation for the deviation of the predictions and observations is that current surface and subsurface loads are correlated. Figure 5 shows correlation coefficients for the initial surface and subsurface loads for the best fit elastic thickness determined for central Australia. Note that over the broad range of wave bands where the model coherence shows the poorest fit to the observations (which incidentally includes the 200-km wavelength of the basins) the apparent loads are highly correlated, as might be expected if folding involving the entire lithosphere [Lambeck, 1983] was responsible for creating the dominant features responsible for the topographic and gravity signals in this region. This correlation is a violation of one of the model assumptions and indicates that the elastic thickness may be underestimated. For comparison, Figure 5 also shows the correlation of initial loads for the best fit elastic thickness in western Australia. As for central Australia, the correlation is high (and has large standard errors) in the transition wave bands. However, in contrast to central Australia, the transition is narrow; over most of the range of wave bands the correlation is low, and the solution can be considered acceptable. Correlated loads may reflect either a real relationship between surface and subsurface loading or may indicate that the uniform plate model was applied in an area of markedly nonuniform flexural rigidity. In central Australia, the correlated loads may be a consequence of a variety of processes in the geologic history of the area such as folding, faulting, uplift, erosion, and deposition.

Figure 1b shows that the effective elastic thickness or flexural rigidity of the Australian lithosphere is greater for regions that had undergone thermal and tectonic stabilization earlier. This relationship is consistent with observed regional differences in the seismic velocity and electrical conductivity structures in the Precambrian and Phanerozoic lithospheres [Cleary et al., 1972; Goncz et al., 1975; Goncz and Cleary, 1976; Lilley et al., 1981; Finlayson, 1982] and with the continent-wide pattern of surface heat flow [Sass et al., 1989].

**Fig. 5.** Correlation between the amplitudes of initial surface and subsurface loads for (a) central Australia and (b) western Australia. For central Australia, loads are located at the surface, 1.6 km depth, and 35 km depth, while for western Australia loads are located at the surface and a depth of 35 km.
Fig. 6. Components of topography along a 900-km-long E-W transect in western Australia (latitude is 24.3243°) assuming \( T = 132 \) km. The profiles from top to bottom show the present topography, the topography due to loading at the surface \( (H_t) \) and the Moho \( (H_b) \) at depth 35 km, and the locally compensated topography \( (H_{LC}) \). Most of the total topographic amplitude appears as a locally compensated plateau that contains wavelengths >1000 km. In the assumed model of surface and subsurface loading of a continuous elastic plate, short-wavelength topographic variations due to subsurface loading are damped by the flexural strength of the plate. Components of the topography much less than the flexural wavelength therefore appear as surface loads. Profile location is marked on Figure 2.

Surface and Subsurface Loading

We consider here the development of topography due to surface and subsurface loading. Specifically, we assess the implications for the nature of isostatic compensation by examining, as end-member cases, areas with large and small elastic thicknesses.

Figure 6 plots components of the present topography along a 900-km-long E-W transect across part of the Yilgarn Block on the western Australian shield where the elastic thickness is about 130 km. (The location of the transect is shown in Figure 2.) The profiles show regionally compensated topography arising from loading at the surface and a subsurface (Moho) interface at a depth of 35 km and the locally compensated topography. The topographic components were found by inverse transforming the Fourier amplitudes of these quantities. Figure 6 illustrates that most of the total topographic amplitude consists of wavelengths greater than about 1000 km (cf. Figure 4). This is longer than the flexural wavelength of the elastic lithosphere \( (\lambda = 800 \text{ km}) \), and therefore this topography must be supported by density variations in the crust or upper mantle rather than by elastic stresses in the lithosphere. Topographic variations across the profile for the most part consist of wavelengths that are short compared to the flexural wavelength and thus

Another possible trend in effective elastic thickness can be recognized within the Eastern Highlands. The rigidity of the near-coastal regions is less in the south and central areas than in the north, as would be expected by the north-south directed propagation of a hot spot such as proposed by Wellman and McDougall [1974a]. However, it is not possible to distinguish confidently whether this trend represents an actual increase in rigidity or is simply a consequence of errors such as those associated with the choices of subregions (though we observe the trend for several different regionalizations). The calculated uncertainties in the elastic thickness of the northeast (BBFB; \( T = 26 \pm 8/5 \text{ km})

central east (NEFB; \( T = 16 \pm 6/4 \text{ km})

and southeast (SLFB; \( T = 17 \pm 6/4 \text{ km})

subregions in Figure 1b indicate that the values are not significantly different. Thus our results, though not in conflict with the hot spot propagation hypothesis, do not lend further support to it.

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Fig. 7. Components of topography along a 300-km-wide E-W profile in southeastern Australia (latitude is ~33.8739°) assuming $T = 23$ km and $15$ km. The western part of the profile contains the Murray Basin and the eastern part contains the southern Lachlan Fold Belt. Contributions to the present topography are determined for loading at the surface ($H_t$), a shallow subsurface interface ($H_m$) at a depth of 0.7 km, and deep subsurface (Moho) interfaces ($H_b$) at depths of 50 km (solid line) and 30 km (dashed line). Also shown is the locally compensated topography ($H_{LC}$). Profile location is marked on Figure 2.

appear primarily as surface loads. Subsurface loading makes a relatively insignificant contribution to the surface topography.

Figure 7 shows a map of the components of the topography in a profile that crosses the Southeastern Highlands. The flat-lying area in the western part of the profile corresponds to the Murray Basin, and the highland area in the east corresponds to the Lachlan Fold Belt. As discussed previously, the present crustal thickness beneath the highlands is approximately 50 km. This may reflect the addition of as much as a 20-km thickness of low-density material to the original crust. The topographic components in Figure 7 were calculated assuming loading at a shallow subsurface density interface at a depth of 0.7 km and two different depths ($z_m$) for the bottom (Moho) interface. The greater value ($z_m = 50$ km) represents the case of loading at the depth of the base of the present crust, while the lesser value ($z_m = 30$ km) represents loading at the presumed depth of the Moho prior to crustal underplating. Results are shown for $T = 23$ km, which is the elastic thickness of the combined basin and highlands area, and $T = 15$ km, which is the elastic thickness of the highlands only. For $T = 23$ km, approximately half of the amplitude of the present topography of the highlands arises from loads that are locally compensated. These loads have wavelengths greater than about 600 km, and it is not possible to distinguish whether they are surface or subsurface in nature. The remainder of the topography is primarily a consequence of surface and deep subsurface loading. The contribution to the topography from loading at the base of the present crust is minimal. However, loading at the Moho before underplating contributes significantly to the uncompensated topography of the highlands. For $T = 15$ km, about 60% of the topography is locally compensated. As would be expected, loading at the surface is less important and loading at the Moho more important than for the case with the thicker plate. For three of the four cases shown in Figure 7, loading at the Moho contributes 30–40% of the topographic relief produced by loads that are regionally compensated. Any nonlinear behavior of the lithosphere, such as faulting, will lead to an incorrect assignment of loads calculated assuming a continuous elastic plate, which could further increase the importance of subsurface loading in creating short-wavelength topography. Interpreted in the context of geochemical and geophysical information on the lower crustal and upper mantle structure of southeastern Australia, these subsurface loads could represent either thermal anomalies or compositional variations related to crustal underplating, igneous intrusions, or low-density upper mantle, all of which have been invoked to explain the uplift and isostatic compensation of the highland topography.

**Conclusions**

We have invoked an isostatic model for the Australian lithosphere that assumes regional compensation of an elastic plate that undergoes flexure in response to surface and subsurface loading. Using the coherence between Bouguer gravity and topography, we find that for the continent as a whole, loads with wavelengths greater than 1500 km are locally compensated; loads with wavelengths in the range 600–1500 km are partially supported by regional stresses; and loads with wavelengths less than 600 km are almost entirely supported by the strength of the lithosphere. A least squares inversion of the observed and predicted coherence indicates that the Australian lithosphere has an average effective elastic thickness of approximately 130 km ($D = 2.0 \times 10^{25}$ N m).

Subregions within the continent with different tectonic histories and ages since stabilization display different rigid-
ities. As a general rule, the flexural rigidity or effective elastic thickness of Australia increases with increasing time since lithospheric stabilization. Precambrian shields in western, northern, and southern Australia have average elastic thicknesses of the order of 100 km, while basins within the Phanerozoic-aged Eastern Interior Lowlands have a thickness of 30–80 km. Fold belts along the east coast have thicknesses in the range 15–35 km, which is similar to values obtained for thermally and tectonically active continental regions such as the East African Rift and the fold and thrust belts of the western Cordillera, but not as low as values obtained for the Basin and Range Province. We interpret the low rigidity of the east coast of Australia to be a consequence of an active thermal and tectonic history that lasted from the Paleozoic through the Cenozoic and included orogeny, uplift, erosion and deposition, crustal underplating, and thermal weakening related to magmatic intrusion and offshore seafloor spreading.

The predicted coherence for a flexural model of a continuous elastic plate does not provide a good fit to the observed coherence of central Australia. This can most simply be explained by the fact that surface and subsurface loads are statistically correlated, in violation of the model assumptions. The correlation may reflect tectonic processes such as lithosphere scale folding or faulting.

Most of the topographic amplitude of the western Australian shield consists of a locally compensated plateau that includes wavelengths greater than 1000 km; shorter-wavelength topographic relief appears almost entirely as surface loads. In southeastern Australia, roughly half of the topographic elevation of the coastal highlands is locally compensated. Regionally compensated topography contains significant contributions from both surface and subsurface loads. The importance of subsurface loading in the development of topography determined by our models is consistent with a scenario in which uplift and compensation of the southeastern highlands was accomplished by crustal underplating, igneous intrusion, and thermal perturbation of the upper mantle.

APPENDIX: RELATIONSHIP OF EFFECTIVE ELASTIC THICKNESS TO SUBREGION DIMENSIONS

The effective elastic thickness is determined from the wavelength of the transition from high ($\gamma^2 \rightarrow 1$) to low ($\gamma^2 \rightarrow 0$) coherence. As discussed in the text, a thicker or more rigid elastic lithosphere incurs this transition at longer wavelengths than a thinner or less rigid lithosphere. For example, Figure 4 shows that for an area with an elastic thickness of the order of 100 km the transition occurs at a wavelength of approximately 1000 km, while for an area with an elastic thickness of 20 km the transition occurs at a wavelength of about 500 km. The subregions in Figure 1b were chosen with the intention of isolating the smallest areas that had undergone common thermal and tectonic histories and that in addition contained long enough wavelengths to resolve fully the transitions. As a result of this approach and the fact that Australia does not contain large regions with uniformly low rigidities, the largest areas in Figure 1b have the largest elastic thicknesses. Thus the relationship between the size of a region and its elastic thickness is only apparent and not an artifact of the method. It is nevertheless useful to compare areas with the same physical dimensions that display different rigidities. The comparison will illustrate the rationale for analyzing regions of different sizes. We will then examine the implications of analyzing a large subregion of variable rigidity using the results for continental Australia as an example.

Figure A1a shows the coherence for a 700 × 700 km² area in southeastern Australia that includes parts of several subregions from Figure 1b that have elastic thicknesses ranging from approximately 15–75 km. The coherence indicates that Bouger gravity and topography were uncorrelated ($\gamma^2 \sim 0$) for $\lambda < 200$ km and highly correlated ($\gamma^2 \sim 1$) for $\lambda > 600$ km. Within the transition wavelength bands the coherence exhibits considerable scatter that indicates that the area is probably not uniformly rigid. (This is not surprising given the spatial variation of elastic thickness indicated by Figure 1b.) The transition from high to low coherence is bounded at both ends by coherences with small standard errors; therefore the best fit elastic thickness of 29 km is well-constrained. However, the result represents a weighted average of the elastic thicknesses of smaller subregions of uniform rigidity contained within the block.

Figure A1b illustrates the consequence of choosing a small subregion in an area where the lithosphere has a high rigidity. The diagram shows the coherence for a subsection of an area in western Australia that has been shown in Figure 4 to have an elastic thickness of greater than 100 km. The
region has the same physical dimensions as the area in Figure A1b; however, the observed coherence is much different. Specifically, in Figure 5b, Bouger gravity and topography are uncorrelated for \( a < 400 \text{ km} \) and are correlated only at the longest wavelengths. The longest-wavelength coherence fails in a transition wave band and displays a large standard error, so the elastic thickness of 116 km is not as well-constrained as for the previous case. However, for wavelengths greater than 200 km the coherences in all wave bands in Figure A1b are lower than those in closely corresponding wave bands in Figure A1a; thus the elastic thickness of western Australia must be greater than for southeastern Australia.

In Figure A1 neither the wavelengths of the transitions from high to low coherence nor, consequently, the best fit elastic thicknesses are controlled by the dimensions of the study areas. The figure demonstrates the practical problems encountered in (1) interpreting the elastic thickness obtained in an area of variable rigidity and (2) attempting to determine the elastic thickness of a small, rigid area. The first problem can be overcome by analyzing smaller subregions in order to isolate regions with uniform rigidities, while the second problem can be addressed by analyzing a larger area in order to obtain a better constrained result.

The solution for continental Australia (Figure 3) represents an average of many subregions with different flexural rigidities. Notice that the best fit elastic thickness \( (\approx 130 \text{ km}) \) is greater than the average value for the smaller subregions contained within it (cf. Figure 1b). This may be explained in part by positively correlated surface and subsurface loads in central Australia, which will make the plate appear weaker than it actually is [e.g., Bechtel et al., 1987]. Surface and subsurface loads are not highly correlated for the continent as a whole; correlated loads in a given area, however, may appear random when averaged on the continental scale. Because of the errors involved with including areas of variable rigidity and with correlated loads, we assign large uncertainties \( (+75/-50 \text{ km}) \) to our average best fit elastic thickness for Australia. However, even with the high uncertainties the value is quite large, of the order of 100 km.

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T. D. Bechtel, Kurz Associates, 297 Wickenden Street, Providence, RI 02903.

D. W. Forsyth, Department of Geological Sciences, Box 1846, Brown University, Providence, RI 02912.

M. T. Zuber, Geodynamics Branch, Code 621, NASA Goddard Space Flight Center, Greenbelt, MD 20771.

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