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2 Mapping the mechanical anisotropy of the lithosphere using a 2D
3 wavelet coherence, and its application to Australia

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6 **Abstract**

7 We develop a new method for imaging the spatial variations of the anisotropy of the flexural response of the litho-
8 sphere, and apply it to recent topographic and gravity data sets over Australia. The method uses two-dimensional
9 Morlet wavelet transforms, superposed in a strictly controlled geometry, to estimate the auto- and cross-spectra of the
10 two data sets in a number of different directions. The resulting wavelet coherence is a function of scale, or wavelength,
11 as well as orientation, and is inverted, at each spatial location, for the three parameters of an anisotropic, thin elastic
12 plate model, i.e., maximum and minimum flexural rigidities and the orientation of the maximum. Extensive tests of
13 the method on synthetic anisotropic, but uniform, data sets, show that it retrieves the amplitude and orientation of
14 the anisotropy with useful accuracy.

15 The results for Australia west of 143°E show a strong correlation with the shallower layers (75–175 km) of a recent
16 model of seismic SV wave azimuthal anisotropy. The ‘weak’ axes (i.e., of minimum flexural rigidity) in most cases are
17 approximately at right angles to the fast axes of the seismic anisotropy, implying that, for Precambrian Australia,
18 they arise from the same source. This is most likely deformation resulting from the most recent episode of orogeny.

19 *Key words:* Isostasy, anisotropy, lithosphere, wavelets, Australia

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21 1. Introduction

22 Estimates of the mechanical strength of the continental and oceanic lithosphere may be obtained from
23 spectral isostatic analyses. The concept of isostasy provides a physical model that relates loading on and
24 within the Earth to the compensating buoyancy of a subsurface density distribution, as in Archimedes’
25 principle. Current isostatic models and plate tectonic theory describe a mechanically strong lithosphere
26 ‘floating’ on an inviscid, higher density asthenosphere, where long-term loading is supported by flexure
27 of the lithosphere and the corresponding displacement of the asthenosphere. The degree of this flexure is
28 determined by the lithosphere’s flexural rigidity (D), a measure of its mechanical strength, which is more
29 often expressed in terms of an effective elastic thickness (T_e), the two being related by:

$$30 \quad D = \frac{E T_e^3}{12(1 - \sigma^2)} \quad (1)$$

31 (e.g., Watts, 2001), where E is Young’s modulus, and σ is Poisson’s ratio. Hence, these terms, D and T_e , are
32 used interchangeably when referring to the same physical process. The magnitude of T_e (D) depends upon
33 many factors, including the temperature, composition and state of stress of the lithosphere, and controls
34 the tectonic evolution of a region, with large-scale T_e variations known to correlate with tectonic province
35 boundaries and seismicity (e.g., Lowry and Smith, 1995).

36 The lithosphere’s thickness can be characterised in several ways. Its thermal thickness, corresponding to
37 the depth at which heat transfer mechanisms change from conduction to convection, can reach magnitudes
38 of up to 350 km (e.g., Artemieva and Mooney, 2001). Its seismogenic thickness, governed by the depth
39 to the brittle/ductile transition, has values reaching 25 km (e.g., Watts and Burov, 2003). The continental
40 elastic thickness, however, is not a physical length parameter, and does not correspond to the depth to which
41 lithospheric rocks behave elastically, as has been suggested in the past (e.g., McNutt, 1990), although there
42 may be some correlation in the oceans (e.g., Watts, 1978). Burov and Diament (1995) and Watts and Burov
43 (2003) have demonstrated that T_e represents the thickness of an equivalent *elastic plate* that best models the
44 flexural properties of the lithosphere, regardless of its actual rheology. As such, T_e is a mechanical, rather
45 than geometrical, property, and indeed, some researchers prefer to present results in terms of the flexural
46 rigidity, which avoids assumptions of the associated elastic constants E and σ .

47 In the spectral isostatic method, the Bouguer gravity anomaly and topographic signature are compared
48 via the construction of two wavelength-dependent functions, the *admittance* and *coherence*. The former is
49 essentially a transfer function from topography to gravity, while the latter is the square of the correlation
50 coefficient in the wavenumber (spatial frequency) domain between these signals. The reasoning goes that
51 large topographic features (loads) will flex even strong plates: the loads are buoyed by isostatic compensation

52 which generates large Bouguer anomalies, resulting in a coherence that approaches unity. Smaller features, in
53 contrast, will be adequately supported by the plate’s mechanical strength, generating little or no associated
54 Bouguer anomaly: the coherence at these wavelengths approaches zero. The *transition wavelength* at which
55 the coherence approximates 0.5 is indicative of the scale at which isostatic compensation begins to prevail
56 over mechanical support, as load size increases. Within a strong/thick plate, the transition wavelength has
57 large values; whereas within weak/thin plates, the coherence “rolls-over” at shorter wavelengths. In the
58 classical method, both admittance and coherence are computed in the Fourier domain (e.g., Forsyth, 1985),
59 though in recent work, their relationship has been analysed through the wavelet transform (Stark et al.,
60 2003; Kirby and Swain, 2004). By comparing predictions from theoretical loading models with the observed
61 admittance and coherence, T_e may, in principle, be determined.

62 The theoretical model most commonly used to interpret the observations is the flexure of a *thin* elastic
63 plate (e.g., Watts, 2001). Such a plate has the properties that its vertical deflections under loading are small
64 compared to its thickness, and that this thickness is small compared to the lateral extent of the plate, which
65 is true in a majority of tectonic regimes. An alternative model is that of the thick plate (Comer, 1983;
66 Wolf, 1985). However, Banks et al. (1977) and Watts (2001) conclude that the errors that arise through
67 the approximations and assumptions of thin-plate theory are not important, considering that most of these
68 errors are manifest in the short wavelengths, and not around the transition wavelengths of importance in
69 T_e -estimation. Hence, thin plate theory is adequate at modelling most tectonic environments.

70 In the past, theoretical admittances and coherences have been computed from the loading of an *isotropic*
71 and *uniform* thin elastic plate, that is, one in which the loading response is equal in all directions from the
72 applied load. However, given the highly variable temperature, compositional and stress regimes present in
73 the lithosphere, and the existence of faulting of all orientations, it is clear that the loading response of the
74 lithosphere will not generally be isotropic or uniform at all scales.

75 More recently, several researchers have attempted to detect anisotropy in the flexural rigidity by analysing
76 observed data with two-dimensional (2D) techniques (such as maximum entropy or multitaper spectral
77 estimators) and interpreting the results as evidence of anisotropy (e.g., Lowry and Smith, 1995; Simons et
78 al., 2000, 2003; Audet and Mareschal, 2004). These studies, however, have not used an anisotropic plate
79 model by which to interpret their observed coherences.

80 Using such a plate model, anisotropic estimates of T_e were computed for central Australia by Swain
81 and Kirby (2003b). In this approach, an observed 2D coherence, computed by the multitaper method,
82 was compared with a theoretical coherence predicted by loading on an anisotropic, thin elastic plate with
83 assumed values of T_x , T_y (being the anisotropic elastic thicknesses) and the orientation of weakest rigidity,

84 and based on a straightforward modification of the isotropic equations of Banks et al. (2001). The estimated
85 values of T_x , T_y and anisotropy direction for central Australia were those that minimised the misfit between
86 theoretical and observed coherence.

87 The present study forms an extension to this work. Differences with Swain and Kirby (2003b) involve
88 the replacement of the multitaper Fourier method by the wavelet transform through which to estimate
89 observed coherence (Kirby and Swain, 2004), and an anisotropic development of coherence estimation. Swain
90 and Kirby (2006) have also extended the “predicted coherence method” of Forsyth (1985) to the wavelet
91 domain for *isotropic* T_e -estimation, and while we have also tackled anisotropic T_e -estimation in this fashion,
92 the following study concerns the approach using theoretical coherence equations with an assumed ratio of
93 subsurface to surface loads. This assumption means that our anisotropy directions (but not magnitudes) will
94 be strictly comparable with those measured directly from coherence anisotropy, as in the studies by Simons
95 et al. (2003), for example, which do not depend on the thin plate model. As noted below, the differences
96 between the results of these two approaches are generally quite small, though occasionally significant, and
97 in particular, the conclusions of the present study are not changed by the use of Forsyth’s method.

98 **2. The wavelet admittance and coherence**

99 *2.1. Isotropic case*

100 In the classical method of spectral isostatic analysis, the Fourier transform is utilised to estimate the
101 frequency characteristics of the isostatic admittance and coherence, whether through the conventional pe-
102 riodogram, or more recently multitaper methods. The drawback of the isotropic Fourier approach is that
103 only one estimate of the coherence (and admittance) is achievable for a data window. These parameters are
104 dependent on wavenumber only, and all spatial information is irretrievable. The most common method used
105 to circumvent this deficiency involved the windowed Fourier transform (WFT) (e.g., Lowry and Smith, 1994;
106 Poudjom Djomani et al., 1999; Pérez-Gussinyé et al., 2004). In this approach, the coherence and admittance
107 are estimated within a moving window, smaller than the study area, as in the Gabor transform. Within
108 each window, comparison of the observed and theoretical coherences yields T_e values that can be mapped,
109 giving the variation of elastic thickness over the study area. The choice of window size depends upon the
110 strength of the plate being mapped, with larger windows needed to estimate larger plate thicknesses, due
111 to their larger transition wavelengths (e.g., Macario et al., 1995).

112 However, the WFT suffers from two shortcomings. First, owing to its fixed window size, it performs best
113 on data having a narrow wavenumber bandwidth, and cannot easily resolve broadband signals (e.g., Addison,

114 2002). Even though gravity and topography data have red spectra, they are not narrow-band signals. The
 115 second disadvantage lies in the trade-off between window size and resolution. On the one hand, use of a small
 116 window can better represent the spatial variation of T_e over a large study area, though it will not resolve any
 117 transition wavelengths longer than the window size. On the other hand, while larger windows can resolve
 118 longer wavelengths, they run the risk of averaging the elastic thicknesses within a region of highly variable
 119 T_e , thus not representing the high-frequency variations on the overall map.

120 The wavelet transform was developed by Grossman and Morlet (1984) to overcome the deficiencies of the
 121 WFT. Whereas Fourier coefficients are dependent on wavenumber only, wavelet coefficients are functions of
 122 both wavenumber and spatial coordinates. This valuable extra information is achieved through convolution
 123 of a signal with localised basis functions (the wavelet), rather than with the infinitely-extending (though in
 124 practice truncated) sinusoids of Fourier analysis.

125 In this study, we use the 2D continuous wavelet transform (CWT) in place of the Fourier transform
 126 when computing the auto- and cross-correlations of gravity and topography data needed in admittance and
 127 coherence analyses. This then gives these quantities as functions of wavenumber, geographical location, and
 128 azimuth, so that a map of the directional variations of flexural rigidity may be computed. Kirby (2005)
 129 provides a brief introduction to the 2D CWT, while a more complete discussion can be found in, e.g., Farge
 130 (1992) or Antoine et al. (2004).

131 In practice, the 2D CWT of a signal is computed via the Fourier transform, which speeds up the convolution
 132 procedure. For a 2D space-domain signal $y(\mathbf{x})$, its wavelet coefficients, $\tilde{y}(s, \mathbf{x}, \theta)$, are generated through a
 133 computational implementation of the following formula:

$$134 \quad \tilde{y}(s, \mathbf{x}, \theta) = \mathbf{F}^{-1} \left\{ \hat{y}(\mathbf{k}) \hat{\psi}_{s,\theta}^*(\mathbf{k}) \right\} \quad (2)$$

135 where, $\mathbf{x} = (x, y)$ is the space-domain position vector, for grid eastings (or longitude), x , and grid northings
 136 (or latitude), y ; s is the wavelet scale, described later; θ is the rotation parameter, determining the resolving
 137 azimuth of the wavelet; $\mathbf{k} = (u, v)$ is the 2D wavenumber, where u is the wavenumber in the x -direction,
 138 and v that in the y -direction; \mathbf{F}^{-1} is the inverse 2D Fourier transform; and $\hat{y}(\mathbf{k})$ is the 2D Fourier transform
 139 of the signal.

140 The $\hat{\psi}_{s,\theta}^*(\mathbf{k})$ are the complex conjugates of the 2D Fourier transforms of the ‘daughter’ wavelets. They are
 141 derived from a ‘mother’ wavelet, $\hat{\psi}(\mathbf{k})$, through the relationship:

$$142 \quad \hat{\psi}_{s,\theta}(\mathbf{k}) = s \hat{\psi}(s \Omega(\theta) \mathbf{k}) \quad (3)$$

143 That is, the mother wavelet is both dilated and weighted by the chosen scale, s , and also rotated through a
 144 chosen angle θ , to yield a daughter wavelet. In Eq. (3), the rotation matrix, $\Omega(\theta)$, for positive-anticlockwise

145 rotations, θ , of the mother wavelet is given by:

$$146 \quad \Omega(\theta) = \begin{pmatrix} \cos \theta & \sin \theta \\ -\sin \theta & \cos \theta \end{pmatrix} \quad (4)$$

147 The translation of the daughter wavelet over the signal is achieved by convolution in the space-domain, or
 148 equivalently via Eq. (2) in the wavenumber-domain, enabling a region-by-region analysis of the signal. The
 149 computation is performed for many values of scale and azimuth, where the former are usually chosen to span
 150 the complete bandwidth of the signal, and our choice of the latter is discussed in Sections 2.2, 5 and 6. The
 151 wavelet scale, s , determines the width (dilation) of the daughter wavelet in the space domain, and hence
 152 determines resolution. At large scales, the wavelet coefficients reveal long wavelengths in the data; at small
 153 scales they show short wavelengths. In this regard, the wavelet transform resembles a series of band-pass
 154 filters. It is the combination of this dilation with the translation which allows the frequency characteristics
 155 at each location of the signal to be revealed.

156 The choice of analysing wavelet is important with regards to a final interpretation of the CWT, particularly
 157 when computing the isostatic coherence. First, the Fourier transform of some real-valued, asymmetrically-
 158 distributed data, $y(\mathbf{x})$, is a Hermitian function, $\hat{y}(\mathbf{k})$, whose real component is an even function, and imagi-
 159 nary component is an odd function. The imaginary part, $\text{Im}[\hat{y}(\mathbf{k})]$, is shown schematically in Figure 1, and
 160 it can be seen that the lower two quadrants in the wavenumber domain contain redundant information: they
 161 are merely (negative) mirrors of the upper two quadrants for the imaginary component, and positive mirrors
 162 for the real component (not shown). Hence, if isotropic information from a signal is desired, then only the
 163 upper two quadrants need be analysed.

164 **FIGURE 1 HERE**

165 Now consider the isostatic coherence. Conventionally, the Fourier coherence between Bouguer gravity
 166 anomalies, $b(\mathbf{x})$, and topography/bathymetry, $h(\mathbf{x})$, is computed in the wavenumber domain from averages
 167 of their auto- and cross-spectra:

$$168 \quad \gamma_{\text{F}}^2(|\mathbf{k}|) = \frac{\left| \langle \hat{b} \hat{h}^* \rangle_{|\mathbf{k}|} \right|^2}{\langle \hat{b} \hat{b}^* \rangle_{|\mathbf{k}|} \langle \hat{h} \hat{h}^* \rangle_{|\mathbf{k}|}} \quad (5)$$

169 (Forsyth, 1985), where the $\langle \cdot \rangle_{|\mathbf{k}|}$ represents an averaging over annuli of similar scalar wavenumber. Alter-
 170 natively, other methods of averaging may be performed, such as through one of the multitaper spectral
 171 estimators (e.g., Simons et al., 2000, 2003), in which case the Fourier coherence is a function of 2D, and not
 172 1D, wavenumber. What is important, though, is that some form of averaging must be performed: if not then

173 the numerator and denominator in Eq. (5) cancel and the coherence is unity at all wavenumbers.

174 With the wavelet coherence, we perform an azimuthal averaging in the wavenumber domain, constrained
 175 by the geometry of the ‘fan’ wavelet (Kirby, 2005). The fan wavelet is constructed from a strictly controlled
 176 superposition of 2D Morlet wavelets which are able to exactly reproduce the Fourier power spectrum of a
 177 signal (Kirby, 2005), without further computation (e.g., Stark et al., 2003).

178 The procedure to compute the wavelet coherence involves, first, taking the Morlet wavelet transform, at
 179 some starting azimuth, θ , of the Bouguer anomaly:

$$180 \quad \tilde{b}_{s\mathbf{x}\theta} \equiv \tilde{b}^M(s, \mathbf{x}, \theta) = \mathbf{F}^{-1} \left\{ \hat{b}(\mathbf{k}) \hat{\psi}_{s,\theta}^{M*}(\mathbf{k}) \right\} \quad (6)$$

181 [c.f. Eq. (2)], where the first term indicates the convenient short-hand notation used hereafter for the Morlet
 182 wavelet coefficients. The Fourier transform of the 2D Morlet (daughter) wavelet, at some scale s and azimuth
 183 θ , is:

$$184 \quad \hat{\psi}_{s,\theta}^M(\mathbf{k}) = s e^{-[(su - |\mathbf{k}_0| \cos \theta)^2 + (sv - |\mathbf{k}_0| \sin \theta)^2]/2} \quad (7)$$

185 where $|\mathbf{k}_0| = \pi \sqrt{2/\ln 2} \approx 5.336$ (e.g., Farge, 1992). The $\tilde{b}_{s\mathbf{x}\theta}$ are then computed at a number of other az-
 186 imuths, determined according to the fan wavelet geometry in the following manner. If a total azimuthal extent
 187 of $\Delta\theta$ is required, then the necessary number of Morlet wavelet transforms is given by $N_\theta = \text{int}(\Delta\theta/\delta\theta)$,
 188 where $\delta\theta$ is the azimuthal increment between successive Morlet wavelet transforms. It was determined that,
 189 in order to avoid azimuthal over- and under-sampling, the azimuthal increment should have a value of $\delta\theta =$
 190 $2\sqrt{\ln \frac{4}{3} \ln 2/\pi} \approx 16.3^\circ$ (Kirby, 2005). This procedure is illustrated schematically in Figure 1, for $\Delta\theta = 180^\circ$,
 191 giving $N_\theta = 11$. Each set of concentric circles represents a single Morlet wavelet, and N_θ sets of gravity
 192 wavelet coefficients, $\tilde{b}_{s\mathbf{x}\theta}$, are obtained by multiplying each Morlet wavelet by the gravity Fourier transform,
 193 according to Eq. (6). [Incidentally, averaging all 11 Morlet wavelets yields the fan wavelet shown in Figure
 194 1 of Kirby and Swain (2004).] The procedure is also performed at the same azimuths on the topography,
 195 yielding N_θ sets of topography wavelet coefficients, $\tilde{h}_{s\mathbf{x}\theta}$.

196 In order to compute the wavelet coherence, the auto- and cross-spectra of the gravity and topography
 197 wavelet coefficients are averaged over the N_θ azimuths, rather than averaging the wavelet coefficients them-
 198 selves, e.g.:

$$199 \quad \left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_\theta \equiv \frac{1}{N_\theta} \sum_{n=1}^{N_\theta} \{ \tilde{b}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \}_n \quad (8)$$

200 for $\theta_n : n = [1, N_\theta]$, with Eq. (8) defining the notation used following. Then, by analogy with Eq. (5), the
 201 wavelet coherence over an azimuthal extent $\Delta\theta$ is computed from:

$$\gamma_{\text{w}}^2(s, \mathbf{x}, \theta) = \frac{\left| \left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_{\theta} \right|^2}{\left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{b}_{s\mathbf{x}\theta}^* \right\rangle_{\theta} \left\langle \tilde{h}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_{\theta}} \quad (9)$$

As can be seen, the geometry in Figure 1 is isotropic as the 11 Morlet wavelets capture all of the data in the upper two quadrants, although there is some unavoidable leakage into the lower two quadrants. Hence, if $\Delta\theta = 180^\circ$, then the wavelet coherence in Eq. (9) is isotropic and θ -independent: $\gamma_{\text{w}}^2(s, \mathbf{x})$. This quantity has been shown to be directly comparable with both theoretical and Fourier-derived coherence estimates (Kirby and Swain, 2004). However, if more than 11 Morlet wavelets were used in the superposition, cancellations of the imaginary component of the product $\hat{b}(\mathbf{k}) \hat{\psi}_{s,\theta}^{M*}(\mathbf{k})$ in Eq. (6) would begin to occur in the averaging procedure, owing to the negative reflection of $\text{Im}[\hat{b}(\mathbf{k})]$ about the axis $v = 0$, leading to loss of information. In the extreme case of a ring-shaped geometry ($\Delta\theta \rightarrow 360^\circ$), the cancellations would be total, the wavelet coefficients would have no imaginary components (Kirby, 2005), and the wavelet coherence would be unity.

To enable direct comparison with 1D coherence estimates from the conventional annular-averaging method, Eq. (5), the wavelet method can also yield a global 1D profile, through a complete averaging of the auto- and cross-spectra over the space variable at each scale. Using the notation:

$$\langle \cdot \rangle_{\mathbf{x}} \equiv \frac{1}{N_x N_y} \sum_{i=1}^{N_y} \sum_{j=1}^{N_x} \{ \cdot \} \quad (10)$$

where the data grids are of size $N_x \times N_y$, the global wavelet coherence is computed by:

$$\overline{\gamma_{\text{w}}^2}(s, \theta) = \frac{\left\langle \left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_{\theta} \right\rangle_{\mathbf{x}} \left\langle \left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_{\theta}^* \right\rangle_{\mathbf{x}}}{\left\langle \left\langle \tilde{b}_{s\mathbf{x}\theta} \tilde{b}_{s\mathbf{x}\theta}^* \right\rangle_{\theta} \right\rangle_{\mathbf{x}} \left\langle \left\langle \tilde{h}_{s\mathbf{x}\theta} \tilde{h}_{s\mathbf{x}\theta}^* \right\rangle_{\theta} \right\rangle_{\mathbf{x}}} \quad (11)$$

which is now a function of scale and azimuth only.

Finally, the scale of each daughter wavelet must be related to an ‘equivalent Fourier wavenumber’, κ , by taking the wavenumber at which the daughter wavelet has its maximum value (for that scale) to be representative of the harmonics resolved by that daughter wavelet (e.g., Kirby, 2005). The equivalent Fourier wavenumber for the Morlet wavelet at each scale is given by:

$$\kappa = \frac{|\mathbf{k}_0|}{s} \quad (12)$$

(Kirby, 2005). This enables direct comparison with both theoretical models and estimates from Fourier transform methods. Thus, we now write $\gamma_{\text{w}}^2(\kappa, \mathbf{x}, \theta)$ for the wavelet coherence, and $\overline{\gamma_{\text{w}}^2}(\kappa, \theta)$ for the global wavelet coherence.

227 2.2. *Anisotropic case*

228 As pointed out by Kirby and Swain (2004), the case of anisotropic T_e can be studied by limiting the
229 azimuthal extent $\Delta\theta$. If $2\delta\theta < \Delta\theta < 180^\circ$, then the averaged auto- and cross-spectra [Eq. (8)] become direc-
230 tional. However, the coherence formulae, Eqs (9) and (11), require averaging over at least two azimuths, since
231 if only one azimuth transform is computed, the numerator and denominator are equal, and the coherence is
232 always unity.

233 Anisotropy in the wavelet coherence can be revealed by computation at a number of central (or resolving)
234 azimuths, Θ , spanning the range $0^\circ \leq \Theta < 180^\circ$, to ensure good directional sampling. In this study we chose
235 six central azimuths from $\Theta = 0^\circ$ to 150° in increments of $\delta\Theta = 30^\circ$, each averaging $N_\theta = 5$ auto- and
236 cross-spectra over $\Delta\theta = 90^\circ$, although other values of $\Delta\theta$ and $\delta\Theta$ were tested (Section 5). Thus, the spectra
237 were computed from $\theta = -45^\circ$ to $+45^\circ$, giving $\gamma_w^2(\kappa, \mathbf{x}, \Theta)$ at $\Theta = 0^\circ$; then from $\theta = -15^\circ$ to $+75^\circ$, giving
238 γ_w^2 at $\Theta = 30^\circ$; etc. Figure 2 shows the wavelet geometry used for the anisotropic azimuthal averaging of
239 the auto- and cross-spectra at a central azimuth $\Theta = 30^\circ$.

240 **FIGURE 2 HERE**

241 Therefore, at each geographic location of the study area, \mathbf{x}_0 , we now have six anisotropic estimates of the
242 wavelet coherence, $\gamma_w^2(\kappa, \mathbf{x}_0, \Theta)$.

243 **3. The anisotropic plate model**

244 3.1. *An orthotropic plate*

245 Conventional thin plate modelling assumes a lithosphere parameterised by a single flexural rigidity, D .
246 That is, the response of the plate to loading is assumed to be isotropic. In the present study, and in Swain and
247 Kirby (2003b), we model the observed coherences with an orthotropic plate. As described in Timoshenko and
248 Woinowsky-Krieger (1959), an orthotropic plate is one whose elastic properties have at least two orthogonal
249 planes of symmetry, and where the anisotropy may be parameterised by two orthogonal rigidities, D_x and
250 D_y , being the plate strengths in these directions. However, as discussed in the Introduction, isostatic studies
251 typically have results presented in terms of the elastic thickness rather than flexural rigidity. So even though
252 it is rather counter-intuitive to speak of a “thickness in the x -direction” for example, we follow convention,
253 and relate the anisotropic rigidities to their elastic thickness counterparts through analogy with Eq. (1):

254
$$D_x = \frac{ET_x^3}{12(1 - \sigma^2)} \tag{13}$$

255 and similarly for D_y , where in this study we have assumed uniform and isotropic values of $E = 100$ GPa,
 256 and $\sigma = 0.25$. In reality, these ‘constants’ could vary geographically, and Chevrot and van der Hilst (2000)
 257 determined a range of 0.23–0.28 for Poisson’s ratio in Australia. Nevertheless, since lateral variations in
 258 these quantities should affect the coherence at a point equally in all directions, we assume uniformity. These
 259 constants could also vary directionally, and while orthotropic plate theory can account for quantities like E_x
 260 and σ_y , in the absence of such information for Australia, we assume isotropy in these constants.

261 In general, for an initial load $\ell(\mathbf{x})$, the vertical deflection, $w(\mathbf{x})$, experienced by a uniform, orthotropic,
 262 thin elastic plate overlying an inviscid fluid of density ρ_m is given by:

$$263 \quad D_x \frac{\partial^4 w(\mathbf{x})}{\partial x^4} + 2H \frac{\partial^4 w(\mathbf{x})}{\partial x^2 \partial y^2} + D_y \frac{\partial^4 w(\mathbf{x})}{\partial y^4} + \rho_m g w(\mathbf{x}) = \ell(\mathbf{x}) \quad (14)$$

264 (e.g., Timoshenko and Woinowsky-Krieger, 1959) where the last term on the left hand side is the buoy-
 265 ancy force arising from displacement of the underlying fluid, and g is the gravity acceleration. Following
 266 Timoshenko and Woinowsky-Krieger (1959) and Swain and Kirby (2003b), we can approximate H by:

$$267 \quad H \approx \sqrt{D_x D_y} \quad (15)$$

268 In general, the orthotropic plate equation [Eq. (14)] can be solved by a finite difference method, using
 269 sparse matrix techniques (e.g., Kirby and Swain, 2004). For uniform D_x , D_y , and ρ_m , Eq. (14) is far more
 270 easily solved via the Fourier transform:

$$271 \quad \left[\left(\sqrt{D_x} u^2 + \sqrt{D_y} v^2 \right)^2 + \rho_m g \right] \hat{w}(\mathbf{k}) = \hat{\ell}(\mathbf{k}) \quad (16)$$

272 using the approximation, Eq. (15). Recall $\mathbf{k} = (u, v)$. Solution of Eq. (16), through either of the two methods
 273 presented in Sections 3.3 and 3.4, yields the ‘observed’ Bouguer anomaly and topography for the model, and
 274 also the theoretical coherence and admittance.

275 To account for different directions of anisotropy, the u and v axes can be rotated through an angle β , the
 276 direction of anisotropy, defined as the (positive anticlockwise) angle from the positive u axis to the positive
 277 u' axis:

$$278 \quad \begin{pmatrix} u' \\ v' \end{pmatrix} = \begin{pmatrix} \cos \beta & \sin \beta \\ -\sin \beta & \cos \beta \end{pmatrix} \begin{pmatrix} u \\ v \end{pmatrix} \quad (17)$$

279 [c.f. Eq. (4)]. For use in the following discussions, and from Eq. (16), we define a ‘flexural operator’ in the
 280 rotated frame:

$$281 \quad \Lambda(\mathbf{k}) \equiv \left(\sqrt{D_x} u'^2 + \sqrt{D_y} v'^2 \right)^2 \quad (18)$$

282 Note that when $D_x = D_y = D$, $\Lambda(\mathbf{k}) \rightarrow D|\mathbf{k}|^4$, and Eq. (16) reduces to the conventional isotropic plate
 283 model.

284 3.2. Random fractal loads

285 Following Macario et al. (1995) [and further discussed in Swain and Kirby (2003a,b)], initial synthetic
 286 surface and subsurface loads are emplaced on and within a thin elastic plate, and the resulting synthetic
 287 topography and Bouguer anomaly are determined from solutions to the thin plate differential equation. We
 288 tested two such plate models: those of Banks et al. (2001) and Forsyth (1985).

289 We created two random fractal surfaces, $s_T(\mathbf{x})$ and $s_B(\mathbf{x})$, using the 2D spectral synthesis algorithm of
 290 Peitgen and Saupe (1988). We chose the fractal dimension of both surfaces to be 2.5. These surfaces then
 291 had their mean subtracted, and were standardised to unit variance. An example of such a surface is shown
 292 in Figure 3. Each surface was then multiplied by 100 metres, hence representing the amplitude of some load.
 293 The correlation coefficient between the surfaces [R in Macario et al. (1995)] was set to zero, to make the
 294 loads uncorrelated as required for the coherence to be diagnostic of the elastic thickness (Forsyth, 1985).

295 **FIGURE 3 HERE**

296 3.3. The plate model of Banks et al.

297 The formulation of Banks et al. (2001) in the continental case requires two loads: an initial surface load
 298 due to the topography, and an initial subsurface load representing the variable density of a thin layer. This
 299 thin layer is emplaced at depth z_l within the crust, though they also consider the gravity effect of the density
 300 contrast, $\Delta\rho$, at the Moho. Hence, in Eq. (14) we set $\ell(\mathbf{x}) = -\ell_T(\mathbf{x}) - \ell_B(\mathbf{x})$, where the loads are derived
 301 from the fractal surfaces as: $\ell_T(\mathbf{x}) = \rho_c g s_T(\mathbf{x})$ and $\ell_B(\mathbf{x}) = f \rho_c g s_B(\mathbf{x})$. The mean crustal density is ρ_c , and
 302 f is the subsurface-to-surface loading ratio, which we set to a constant value [though see Swain and Kirby
 303 (2003a) for a discussion concerning the effect on f of different fractal dimensions for the two loads].

304 To determine the ‘observed’ Bouguer anomaly and topography from the model, a value for each of T_x ,
 305 T_y and β is assumed. Eq. (13) then determines the anisotropic flexural rigidities. Rearranging the flexural
 306 equation, Eq. (16), gives the (Fourier transform of) the plate deflection as:

$$307 \quad \hat{w}(\mathbf{k}) = \frac{-\hat{\ell}_T(\mathbf{k}) - \hat{\ell}_B(\mathbf{k})}{\Lambda(\mathbf{k}) + \rho_m g} \quad (19)$$

308 which is the anisotropic version of Eq. (16) in Banks et al. (2001). Following Banks et al. (2001), it is a
 309 simple matter to generalise their Eqs (17) and (18) to the anisotropic case. The resulting surface topography
 310 after loading is the sum of the initial surface load amplitude and the resulting plate deflection:

311
$$\hat{h}(\mathbf{k}) = \frac{\hat{\ell}_T(\mathbf{k})}{\rho_c g} + \hat{w}(\mathbf{k}) \quad (20)$$

312 while the surface Bouguer anomaly is estimated from the upward continued thin layer and Moho deflections:

313
$$\hat{b}(\mathbf{k}) = 2\pi G \left(\frac{e^{-|\mathbf{k}|z_l} \hat{\ell}_B(\mathbf{k})}{g} + \Delta\rho e^{-|\mathbf{k}|z_m} \hat{w}(\mathbf{k}) \right) \quad (21)$$

314 where $\Delta\rho = \rho_m - \rho_c$. Inverse Fourier transformation of Eqs (20) and (21) then gives $h(\mathbf{x})$ and $b(\mathbf{x})$.

315 3.4. Forsyth's plate model

316 In the approach of Forsyth (1985), the two loads are an initial topography: $h_i(\mathbf{x}) = s_T(\mathbf{x})$; and relief on
 317 the Moho: $w_i(\mathbf{x}) = f\rho_c s_B(\mathbf{x})/\Delta\rho$. If values of T_x , T_y and β are assumed, the 'observed' topography and
 318 Bouguer anomaly after flexure can be determined. Again, the isotropic equations are readily converted to
 319 anisotropic ones, using Eq. (18). Forsyth's equations for ξ and ϕ become:

320
$$\xi(\mathbf{k}) = 1 + \frac{\Lambda(\mathbf{k})}{\Delta\rho g} \quad \phi(\mathbf{k}) = 1 + \frac{\Lambda(\mathbf{k})}{\rho_c g} \quad (22)$$

321 and the final surface topography and Moho relief are determined from:

322
$$\hat{h}(\mathbf{k}) = \hat{h}_i(\mathbf{k}) \left[\frac{\Delta\rho \xi(\mathbf{k})}{\rho_c + \Delta\rho \xi(\mathbf{k})} \right] - \hat{w}_i(\mathbf{k}) \left[\frac{\Delta\rho}{\Delta\rho + \rho_c \phi(\mathbf{k})} \right] \quad (23)$$

323 and

324
$$\hat{w}(\mathbf{k}) = -\hat{h}_i(\mathbf{k}) \left[\frac{\rho_c}{\rho_c + \Delta\rho \xi(\mathbf{k})} \right] + \hat{w}_i(\mathbf{k}) \left[\frac{\rho_c \phi(\mathbf{k})}{\Delta\rho + \rho_c \phi(\mathbf{k})} \right] \quad (24)$$

325 respectively [c.f. Forsyth's Eq. (18)]. An 'observed' Bouguer anomaly at the surface may then be estimated
 326 from the upward continued Moho relief:

327
$$\hat{b}(\mathbf{k}) = 2\pi\Delta\rho G e^{-|\mathbf{k}|z_m} \hat{w}(\mathbf{k}) \quad (25)$$

328 being the anisotropic version of Forsyth's Eq. (13). Inverse Fourier transformation of Eqs (23) and (25) then
 329 gives $h(\mathbf{x})$ and $b(\mathbf{x})$.

330 4. Estimation of anisotropic parameters

331 In the Forsyth (1985) method, a value for T_e is assumed and the isotropic versions of our Eqs (23) and (24)
 332 are solved simultaneously in order to estimate the initial loads from the observed topography and Bouguer
 333 gravity, and hence calculate a "predicted coherence". T_e is then estimated as the value that minimises
 334 the misfit between observed and predicted coherence. This method has the advantage that it makes no
 335 assumptions about the loading, with f essentially being determined at each wavenumber from the WFT

336 of the data. Swain and Kirby (2003b) showed how to modify this method to estimate T_x , T_y and β for a
 337 uniform orthotropic plate model.

338 In the present study we used a simpler procedure, similar to that used in Kirby and Swain (2004), and
 339 inverted the coherence for the three plate parameters using the theoretical formula for coherence (given in
 340 the Appendix). This requires the knowledge (or assumption) of the loading ratio and we simply assumed
 341 that f is uniform (spatially invariant), isotropic, and wavenumber-independent: we call this the “uniform f ”
 342 method. On this assumption, its variations should affect T_x and T_y equally. When inverting the coherence
 343 at each geographic location, we also assumed that the local wavelet spectra are independent, or “decoupled”
 344 from adjacent spectra (Stark et al., 2003). The validity of this assumption can be tested by means of non-
 345 uniform synthetic models, which we have not used in this study, though Swain and Kirby (2006) present
 346 one such (isotropic) model finding that the decoupling assumption is quite reasonable in that case.

347 It is worth mentioning here that we also tried a wavelet version of Forsyth’s method, in order to test
 348 the assumption of uniform, isotropic f . This involved extending our (isotropic) wavelet development of
 349 Forsyth (1985)’s approach (Swain and Kirby, 2006) to the case of anisotropy, which has the advantage of
 350 also allowing for the possibility of anisotropy in f . The relevant equations are identical to Forsyth’s, with
 351 the Fourier transforms of gravity and topography replaced by their wavelet transforms, $D|\mathbf{k}|^4$ replaced by
 352 $\Lambda(\mathbf{k})$, and the averaging performed over wavelet azimuth rather than scalar wavenumber. When applied to
 353 the synthetic models (see Section 5), the directions of T_e anisotropy given by this method and the uniform f
 354 method agree to within $< 10^\circ$, which is comparable with the standard deviation of the orientations of axes
 355 given in Section 5. The largest differences occur where the T_e anisotropy estimates, with either method, are
 356 small. We also used the anisotropic wavelet Forsyth method with the Australian data, obtaining a comparable
 357 agreement with the uniform f method, except for part of central Australia, where Forsyth’s method yields
 358 larger magnitudes of anisotropy, presumably because of anisotropic (east-west) loading. Because of this
 359 agreement we consider that the assumptions about loading made by the uniform f method are justified,
 360 at least for Australia and the synthetic models. The results in the rest of this paper are for the uniform f
 361 method.

362 In order to estimate the values of T_x , T_y and β for our orthotropic model at a particular \mathbf{x}_0 , we compare
 363 the observed wavelet coherence at all values of Θ with the theoretical coherences of an anisotropic plate in
 364 these directions. To calculate such a coherence, we rewrite the flexural operator in polar coordinates in the
 365 \mathbf{k} -frame:

$$366 \quad \Lambda(|\mathbf{k}|, \Theta) = \left[\sqrt{D_x} |\mathbf{k}|^2 \cos^2(\Theta - \beta) + \sqrt{D_y} |\mathbf{k}|^2 \sin^2(\Theta - \beta) \right]^2 \quad (26)$$

367 The general forms of the theoretical coherences from both Banks and Forsyth models are shown in Appendix

368 A: to compute a theoretical coherence profile, $\gamma_t^2(|\mathbf{k}|, \Theta)$, in direction Θ from Eqs (A.1) and (A.5), replace
 369 $\Lambda(\mathbf{k})$ with $\Lambda(|\mathbf{k}|, \Theta)$.

370 Starting with initial estimates of $[T_x, T_y, \beta]$, the differences at all values of Θ between the theoretical
 371 coherences, $\gamma_t^2(|\mathbf{k}|, \Theta)$, and observed wavelet coherences, $\gamma_w^2(\kappa, \mathbf{x}_0, \Theta)$, are minimised simultaneously using
 372 an iterative damped least squares algorithm similar to the one described in Swain and Kirby (2003b), except
 373 that here we assume that the uncertainty of an observed coherence estimate is proportional to $|\mathbf{k}|$, and
 374 weight the data as $1/|\mathbf{k}|$. After a number of iterations, stable values for the estimates $[T_x, T_y, \beta]$ are reached.
 375 This procedure is then repeated for each geographical location, \mathbf{x}_0 , in the study area, giving estimates in
 376 each grid cell. In a very few instances with real data we have found it necessary to use 2 different starting
 377 models, with β differing by 90° , in order to find the global minimum.

378 Single estimates of $[\overline{T_x}, \overline{T_y}, \overline{\beta}]$ for the whole study area can also be derived by inverting the global wavelet
 379 coherence, $\overline{\gamma_w^2}(\kappa, \Theta)$, using the same method.

380 5. Results for the synthetic models

381 In this study we only consider *uniform*, anisotropic models, because in such cases the plate equation [Eq.
 382 (14)] can be easily and quickly solved using the FFT, as shown above. We generated 100 pairs of random,
 383 synthetic Bouguer anomaly and topography grids using the Forsyth plate model, all with $[T_x, T_y, \beta] =$
 384 $[80, 40, 40^\circ]$ and $f = 1$. Then, as discussed in Section 2.2, the anisotropic wavelet coherence from each pair
 385 was computed at six central azimuths from $\Theta = 0^\circ$ to 150° , in increments of $\delta\Theta = 30^\circ$, each using $\Delta\theta = 90^\circ$.
 386 In order to minimise array sizes and computation time, and to make plots less cluttered, the wavelet auto-
 387 and cross-spectra were ‘binned’ into larger grid cells of size 320×320 km, prior to using Eq. (9). It must be
 388 stressed that this spatial averaging is performed only for computational convenience, and is not essential:
 389 while it has the effect of damping the noise in the coherence at high wavenumbers, it does not significantly
 390 alter the estimated values of the transition wavelength, and hence of T_x or T_y . Furthermore, to eliminate
 391 edge effects from the Fourier transform, the outer 5% at each side was discarded before implementation of
 392 Eq. (11).

393 Inverting the global wavelet coherence, Eq. (11), for all 100 pairs (using the Forsyth theoretical coherence
 394 formula, Eq. (A.5), and assuming $f = 1$), gave $\overline{T_x} = 71 \pm 9$ km, $\overline{T_y} = 45 \pm 6$ km, and $\overline{\beta} = 39^\circ \pm 10^\circ$ (arithmetic
 395 mean \pm standard deviation). The global coherences at the selected Θ values, for one model (#34), are shown
 396 in Figure 4. This particular model gave estimates: $\overline{T_x} = 79.6$ km, $\overline{T_y} = 47.6$ km, and $\overline{\beta} = 45.0^\circ$. The legend
 397 indicates the Θ value for each profile, together with an estimated elastic thickness value at each azimuth
 398 (by inversion of the observed global wavelet coherence curves). Clearly though, the anisotropy of the model

399 is being revealed.

400 **FIGURE 4 HERE**

401 **FIGURE 5 HERE**

402 Figure 5 shows the spatial variation in anisotropy for model #34. The ellipses are “ T_e -ellipses”: the length
403 of the major and minor axes indicate the relative magnitudes of $T_{\max} = \max(T_x, T_y)$ and $T_{\min} = \min(T_x, T_y)$,
404 respectively. The inclined lines within each ellipse give the orientation of maximum mechanical strength,
405 with the weak direction perpendicular; the length of each line is $2\sqrt{T_x T_y}$. The statistics for model #34 are
406 $T_x = 71 \pm 8$ km, $T_y = 41 \pm 7$ km, $\beta = 42 \pm 8^\circ$. Only two estimates (in the bottom-right) are $> 20^\circ$ in error, and
407 these points also exhibit the smallest anisotropy (i.e., $T_x \approx T_y$), so here β is inevitably poorly determined.
408 We believe that such anomalies are a product of the random process used to generate the synthetic fractal
409 models.

410 We have examined the six coherence plots at each point for some of the data sets. Although practically
411 all of them show a very clear and quite smooth transition from high to low coherence, as in Figure 4, many
412 of them also show quite separate, but large, “spikes” or “humps” in coherence at higher wavenumbers in
413 some azimuths. These also occur with real data: similar features can be seen in the 2D coherence plots in
414 Figure 11 of Simons et al. (2003). Because of the $1/|\mathbf{k}|$ weighting, mentioned in Section 4, these features
415 usually have little effect on our inversions. Without the weighting, it is possible for an inversion to find a
416 model which fits a long-wavelength transition at one azimuth together with a short-wavelength transition
417 at another azimuth, resulting in a model with too large an anisotropy.

418 To compare the wavelet method against the Slepian multitaper ($K = NW = 3$) anisotropy estimates
419 in Swain and Kirby (2003b), we also tested 100 pairs for a model with parameters $[100, 50, 0^\circ]$, using the
420 global wavelet coherence. The wavelet method returned estimates of: $\bar{T}_x = 87 \pm 14$ km, $\bar{T}_y = 55 \pm 8$ km, and
421 $\bar{\beta} = 2^\circ \pm 12^\circ$; compared with $T_x = 68 \pm 11$ km, $T_y = 43 \pm 7$ km, and $\beta = -1^\circ \pm 17^\circ$ from the multitaper
422 method. The wavelet method gives a much better agreement with the model input parameters, particularly
423 for the T_x estimate.

424 Finally, a further plate model was used to test the method, this time for a weaker plate, with parameters
425 $[10, 30, 80^\circ]$. The estimates from the inversions of the 100 anisotropic wavelet coherences were: $\bar{T}_x = 14 \pm 2$
426 km, $\bar{T}_y = 26 \pm 3$ km, and $\bar{\beta} = 80^\circ \pm 4^\circ$. Like the Fourier-based methods, it seems that the wavelet method
427 better estimates smaller, rather than larger, values of elastic thickness.

428 The choice of initial estimates of $[T_x, T_y, \beta]$ in the inversion was not found to influence the final estimates
429 of these parameters. Furthermore, computations on the 100 gravity/topography pairs using a smaller value
430 of $\delta\Theta$ (10°) in order to increase directional sampling, produced exactly the same results. Also, increasing

431 the number of wavelet scales (from 28 to 56) had a negligible effect upon the outcome, and merely gave
432 smoother coherence profiles. Finally, choosing a smaller value for $\Delta\theta$ of 45° gave a marginal increase in the
433 variance of the parameter estimates of the uniform models. Hence, we believe that our sampling density in
434 both azimuth- and scale-domains is adequate.

435 6. Application to Australian data

436 6.1. Australian gravity and topography data

437 Computations involving the Fourier transform on continental scales have to account for Earth curvature.
438 To avoid errors arising from the planar treatment of curvilinear coordinates, all data were projected onto
439 a Mercator grid, with origin at 133°E , 0°N . This projection is conformal, thereby preserving angles, and is
440 cylindrical, facilitating an intuitive visual interpretation of directions with respect to geographic parallels.
441 In geographic coordinates, the study area spans approximately 101°E to 165°E , and 2°S to 49°S .

442 For the study area, grids of the topography/bathymetry and complete Bouguer anomaly were derived
443 from the following data sets. The Australian land topography grid was derived from the GEODATA 9-
444 arcsecond DEM (digital elevation model) of Australia (Geoscience Australia, 2001). The bathymetric data
445 were taken from the GEBCO Digital Atlas, given at 1-arcminute spacing (NOAA, 2003), which are compiled
446 from ship-track data only, and, importantly, not from inverted altimetry gravity (e.g., Smith and Sandwell,
447 1997). Topography over Indonesia and Papua New Guinea were also taken from the GEBCO Digital Atlas.
448 Topography and bathymetry data were merged, and regridded at 20 km spacing on the Mercator grid.

449 Simple Bouguer anomalies over mainland Australia were taken from the 2004 release of Geoscience Aus-
450 tralia's land gravity data base. Gravimetric terrain corrections derived from the 9-arcsecond DEM (Kirby
451 and Featherstone, 1999) were added to these anomalies to form complete Bouguer anomalies over the conti-
452 nent. Over Indonesia and Papua New Guinea, free-air anomalies were derived from the EGM96 geopotential
453 model to degree and order 360: simple Bouguer anomalies were computed from these using the above GEBCO
454 topography data. Over marine areas we used free-air anomalies from the KMS02 satellite altimetry model
455 (Andersen and Knudsen, 1998) at a 2-arcminute grid spacing. A marine Bouguer anomaly was derived from
456 these by applying a complete Bouguer correction computed from the GEBCO bathymetric model using the
457 formula of Parker (1972). Again, the land and marine gravity data were merged and regridded at 20 km
458 spacing on the Mercator grid.

460 The bathymetry was first converted into a load by calculating the equivalent depth of the sea bottom
 461 if the salt water were replaced by air, i.e., multiplying the ocean water depth by $(\rho_c - \rho_w)/\rho_c$, where we
 462 used values of $\rho_c = 2800 \text{ kg m}^{-3}$ and $\rho_w = 1030 \text{ kg m}^{-3}$. Then, as for the synthetic models, the wavelet
 463 coherences were inverted, assuming $f = 1$, yielding T_x , T_y and β estimates at each grid node. For these
 464 data, however, we spatially averaged the auto- and cross-spectra into bins of size $80 \times 80 \text{ km}$. As discussed
 465 in Section 5, the bin size does not significantly alter the magnitudes of T_x , T_y , or β , and anisotropy maps
 466 generated using $160 \times 160 \text{ km}$ bins showed the same pattern of directions.

467 Figure 6 shows the results plotted as bars, or ‘axes’, in the weak direction with length proportional to
 468 the anisotropy, here defined as $(T_{\max} - T_{\min})/T_{\max}$. We only show results for the landmasses and their
 469 continental shelves. Results over the deeper ocean are very scattered, probably due to the poor sampling of
 470 the bathymetry data over these areas. Occasionally the inversion does not converge to a solution, but this
 471 appears to be confined to offshore areas.

472 **FIGURE 6 HERE**

473 Figure 6 clearly shows distinct clusters within Australia, containing homogeneous anisotropy directions.
 474 Sometimes the boundary between adjacent clusters exhibits a smooth, gradual change in direction, in other
 475 cases this change is abrupt. In addition, as previously noted by Simons et al. (2003), there is a clear tendency
 476 towards orthogonality between the axes shown in Figure 6 and the Australian coastline.

477 Figure 6 also shows the major boundaries of the Australian crustal mega-element map of Shaw et al.
 478 (1995). Although the correlation between these boundaries and the distribution of our weak axes is far from
 479 being general or precise, there are some good examples: one is the characteristic pattern over the Pilbara
 480 Craton (117°E , 23°S), mimicking its large scale structure. In other cases, such as the Kimberley (126°E ,
 481 16°S), North Australian (130°E , 18°S), and Gawler Cratons (135°E , 32°S), the anisotropy has a uniform,
 482 or slowly varying direction which changes more abruptly over their boundaries.

483 It is noticeable in Figure 6 that the axes change direction more rapidly close to the coast than over the
 484 centre of Australia. This can be explained through the result of Swain and Kirby (2006) that T_e decreases
 485 in magnitude towards the Australian coastline, and through the fact that only mid- to small-scale wavelets
 486 are important when resolving the lower transition wavelengths associated with low T_e . Such wavelets have
 487 a smaller spatial extent, and can thus resolve more rapid changes in both T_e and anisotropy direction.

488 **7. Correlations with other data**

489 It is not an aim of this paper to give a detailed review of the possible geophysical controls on T_e and its
490 anisotropy, which has been provided elsewhere (e.g., Lowry and Smith, 1995; Simons et al., 2000, 2003),
491 other than to note that their methods can all be approximated by our intrinsically anisotropic plate model
492 (Swain and Kirby, 2003b). Nevertheless, in order to give the reader the opportunity to judge the truth of the
493 anisotropy axes shown in Figure 6, we will give a comparison with some other relevant data over Australia,
494 namely: (a) previous studies of isostatic anisotropy; (b) the stress map; (c) seismic azimuthal anisotropy.

495 *7.1. Previous isostatic studies*

496 The previous studies of Australian isostatic anisotropy have all used fixed windows of varying size from
497 720×720 km (Simons et al., 2003) to 2200×2200 km (Swain and Kirby, 2003b), implying averaging over
498 areas encompassing many of the axes in Figure 6 and making comparisons difficult. The closest comparable
499 result to ours is Figure 12b of Simons et al. (2003), which shows axes of implied weakness derived from the
500 maximum transition wavelength of the coherence. Their window width is small enough to downward bias
501 estimates of both T_e and transition wavelength in Australia by more than a factor of three (Swain and Kirby,
502 2003a). This bias would affect estimates of the *amplitude* of anisotropy (which Simons et al. (2003) did not
503 make), but not of its direction. However the bias may well compromise the accuracy of the orientations.

504 Figure 7 shows a comparison of the “long-wavelength weak directions” from Figure 12b of Simons et al.
505 (2003) with the axes shown in Figure 6 calculated by averaging the wavelet spectra over their window size
506 (720×720 km). This figure shows their “good” quality (thick, white bars) and “fair” quality (thick, grey bars)
507 measurements, but omits both “bad” and “null” results (Simons et al., 2003). However, we have included
508 our axes at the locations of their missing measurements, and a majority of these (7/12) show relatively
509 small anisotropy. The agreement between their “good” measurements and ours in Figure 7 is excellent: 9
510 out of 15 are within 30° , which has a $< 2.2\%$ chance of being random. For all 28 of the Simons et al. (2003)
511 measurements (i.e., “good” and “fair”), 14 of ours agree to within 30° , implying a 2.9% chance of random
512 occurrence. Overall, we think that the agreement between the two studies is encouraging.

513 **FIGURE 7 HERE**

514 *7.2. Tectonic stress*

515 Lowry and Smith (1995) state that azimuthal variations of T_e reflect tectonic stress, because such stress
516 reduces T_e in the direction of the stress axis. Their study is of the western US Cordillera, a region of

517 extensional stress, and in several of their tectonic provinces they observe excellent agreement between the
518 minor axes of T_e ellipses (i.e., the weak directions) and directions of minimum horizontal compressive stress
519 (i.e., the direction of tectonic stress). They note, however, some exceptions to this (e.g., Yellowstone).

520 The Australian stress regime is largely compressive (Zoback et al., 1989) so the correlation should be
521 between the T_e minor axis and the direction of maximum horizontal compressive stress. The Australian
522 stress map, unlike those of most other continents, shows orientations of maximum horizontal compressive
523 stress that are variable and not generally parallel to the direction of absolute plate motion (north to north-
524 northeast) (e.g., Hillis and Reynolds, 2003). Because of the variable stress orientations, Hillis and Reynolds
525 (2003) define a number of “stress provinces” for each of which they calculate a mean stress orientation.
526 The data are very sparse and occur in concentrations, mostly within sedimentary basins since the majority
527 of them are from measurements in petroleum exploration boreholes. Thus the mean stress orientations are
528 heavily weighted towards the uppermost crust. Eleven of the twelve Australian stress provinces defined by
529 Hillis and Reynolds (2003) show statistically significant stress orientations, of which nine are from offshore
530 or coastal basins, and two from interior basins (the Cooper and Amadeus Basins).

531 Figure 8 compares the stress province data with the weak directions in our Figure 6, averaged over
532 400×400 km windows. There is little correlation between the two sets of directions. In two cases our weak
533 directions may be too scattered for a useful comparison, and four cases are offshore basins, where, as we
534 have previously noted, the T_e results are less reliable. Overall, however, it appears that tectonic stress is not
535 a major control on the lithospheric strength anisotropy that we observe. This agrees with the conclusions
536 of Simons et al. (2003), who contend that there would only be a relation between the two if present-day
537 stress and fossil strain are still related. However, another factor may be the stress magnitudes: although the
538 observed stress data do not include magnitudes, model stresses are typically less than 30 MPa (Reynolds et
539 al., 2003; Zhao and Müller, 2003). Reference to Figure 4b of Lowry and Smith (1995) shows that such stress
540 magnitudes would probably have no significant effect on T_e .

541 **FIGURE 8 HERE**

542 7.3. *Seismic anisotropy*

543 Seismic anisotropy under the continents is usually studied using shear wave splitting of near-vertical
544 core phases like SKS (Silver, 1996). For Australia such studies have tended to measure mostly weak or
545 null splitting (Debayle and Kennett, 2003). This is most likely because of the inherent lack of vertical
546 resolution of the method, usually requiring the assumption of a single anisotropic layer, which is probably
547 an oversimplification for Australia. Although it is sometimes possible to interpret shear wave splitting data

548 in terms of two anisotropic layers, this requires fairly numerous data with a range of backazimuths, which
549 have not been available for most Australian studies. An exception is the detailed study by Girardin and
550 Farra (1998) of data from the Canberra GEOSCOPE station.

551 Recently, some tomographic models derived from surface waves have included azimuthal anisotropy. Al-
552 though there is a trade-off between heterogeneity and anisotropy (e.g., Simons et al., 2002), tests show that
553 including anisotropy clearly fits the data better and it has the advantage of significant depth resolution.
554 Tomographic studies of Australian seismic anisotropy include Simons et al. (2002) and Debayle and Kennett
555 (2003) who both present maps of SV-wave azimuthal anisotropy from inversion of Rayleigh waveform data.
556 However, both of these studies were hampered by a sparsity of ray paths for Western Australia. Kennett et
557 al. (2004) show a more recent model which includes many more ray paths under Western Australia.

558 Although these models of seismic anisotropy are rather different in detail, there is general agreement
559 (e.g., Debayle and Kennett, 2003; Simons et al., 2003) that the seismic azimuthal anisotropy observed under
560 Australia at depths of $< 150 - 200$ km is quite variable, with as many E-W as N-S orientations, while
561 at depths > 200 km the orientations are consistently N-S to NNE-SSW. Simons et al. (2003) examined
562 the orthogonality between the directions of shallower seismic anisotropy and those of mechanical anisotropy,
563 concluding that they represent “frozen” deformation from the most recent episode of orogeny. There is general
564 agreement that the anisotropy at depths > 200 km reflects present-day deformation due to the northward
565 motion of the Australian Plate. Silver (1996) refers to the “frozen” deformation mechanism as ‘vertically
566 coherent deformation’ (VCD), and to the deformation due to plate motion as ‘simple asthenospheric flow’
567 (SAF), explaining them in terms of preferential alignment of olivine crystals in response to finite strain.

568 Simons et al. (2003) suggest that over most of Precambrian Australia, where compressional tectonics
569 predominate, the weak directions from isostatic analyses should lie approximately at right angles to the
570 fast axes in the shallower mantle layers of the seismic models, because for VCD the fast axes should be
571 approximately perpendicular to the compression direction, which is also the direction that has accumu-
572 lated the most deformation per unit of topographic loading. They compared the weak directions for their
573 “good” measurements, to the fast axes of their seismic model (Simons et al., 2002), finding that they are
574 approximately at right angles in $>50\%$ of cases at the shallowest depth of their model (30 km), falling to
575 30% at 200 km depth. However, the fact that this comparison was based on only 15 measurements of the
576 long-wavelength coherence anisotropy somewhat reduced the impact of this study.

577 We have obtained a recent update of the anisotropic model of Kennett et al. (2004) for 3 depths: 75
578 km, 125 km and 175 km (S. Fishwick, personal communication, 2005). The model consists of azimuth and
579 amplitude (%) of the anisotropy, and shear velocity given on a $3^\circ \times 3^\circ$ grid. In Figures 9, 10 and 11 we

580 compare the seismic anisotropy with our weak direction axes, the latter calculated by averaging the spectra
581 over a window size of 300×300 km. The length scale for the seismic anisotropy axes is $1/30$ of that of the
582 isostatic weak axes. At all three depths there is a strong tendency for the axes to lie at a large angle to
583 one another, particularly for inland Australia west of about 143°E . At 75 km, but not at the other depths,
584 there seems to be a relation between the amplitudes of anisotropy of the two data sets: their orthogonality
585 appears to improve with depth. In eastern Australia there is less correlation apparent between the two sets
586 of axes, but this might be expected since the seismic tomography of Kennett et al. (2004) shows that the
587 lithosphere is much thinner here, so the depths 125–175 km probably represent asthenosphere.

588 **FIGURE 9 HERE**

589 **FIGURE 10 HERE**

590 **FIGURE 11 HERE**

591 We have carried out tests with circular statistics on a subset of 57 pairs of axes from Figures 9, 10 and
592 11 within an area west of 143°E and between latitudes 33°S and 14°S , approximately corresponding to
593 the extent of Precambrian basement. We first applied a Rayleigh test to the hypothesis that differences
594 in orientation between the mechanically weak axes and the fast seismic axes are sampled from a uniform
595 distribution, with no preferred direction, versus the alternative hypothesis that they are from a von Mises
596 distribution - the circular equivalent of a normal distribution (Davis, 1986). Next we tested the hypothesis
597 that the differences have a mean of 90° . Since the data are orientations, they were doubled prior to the
598 calculations, and the angular results subsequently halved (Davis, 1986). The results are given as, for the
599 first test, mean resultant vectors (\bar{R}) and significance levels (α), and for the second test, mean orientations
600 (A) and 95% confidence intervals. They are as follows: for 75 km, $\bar{R} = 0.238$, $\alpha = 4\%$, $A = 101^\circ \pm 25^\circ$; for
601 125 km, $\bar{R} = 0.278$, $\alpha = 1\%$, $A = 87^\circ \pm 21^\circ$; for 175 km, $\bar{R} = 0.477$, $\alpha = < 1\%$, $A = 86^\circ \pm 11^\circ$. Thus, for all
602 three depths, we can say, with $> 96\%$ confidence, that the orientation differences have a preferred direction
603 and that the 95% confidence intervals around all three mean orientations include 90° .

604 **8. Conclusions**

605 We have described a technique for mapping the mechanical anisotropy of the elastic lithosphere which
606 yields greater detail than previously possible. The method gave good results with a wide variety of synthetic,
607 anisotropic, uniform models. Applied to new grids of Australian topography and gravity data it gave a map
608 of minimum T_e directions which is in reasonable agreement with the “long-wavelength weak directions”
609 of Simons et al. (2003). Although both studies suffer from various uncertainties which would contribute
610 to differences between the results, we think that the use in the latter study of a fixed window of width

611 comparable to, or even smaller than, the transition wavelength over much of Precambrian Australia, probably
612 results in relatively large estimation errors.

613 We have compared our results with a recent model of seismic azimuthal anisotropy (Kennett et al., 2004)
614 at depths between 75 and 175 km. Over Precambrian Australia, results of a statistical analysis carried out on
615 the differences in orientation between our weak axes and the fast seismic axes strongly support the hypothesis
616 that they are orthogonal. We are therefore in agreement with Simons et al. (2003) that the source of our
617 observed anisotropy is likely to be vertically coherent “frozen” deformation of the lithosphere (Silver, 1996)
618 due to alignment of olivine crystals. The fact that our results correlate with seismic azimuthal anisotropy
619 at depths of 75–175 km also supports the ideas that under cratonic regions crust and mantle are strongly
620 coupled and that the strength of the lithosphere resides mainly in the upper mantle (Vauchez et al., 1998).

621 Comparison of our weak axes with a map of present day stress directions (Hillis and Reynolds, 2003)
622 shows that there is no obvious plate weakening in these directions, probably because the stresses are too
623 small.

624 We have also indicated how our wavelet approach can be modified so that the predicted coherences are
625 modelled without the assumption of a constant loading ratio (Forsyth’s method). A comparison between the
626 two methods will be the subject of a further paper, but we note that our results to date indicate that using
627 Forsyth’s method will not change any of the above conclusions.

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632 reviewers, whose helpful comments greatly improved this contribution.

633 **Appendix A. Theoretical coherence from the Banks and Forsyth plate models**

634 *A.1. Theoretical coherence from the Banks model*

635 The theoretical coherence for an anisotropic Banks et al. (2001)-style plate with plate constants $[T_x, T_y, \beta]$,
636 and a wavenumber-dependent subsurface to surface loading ratio, $f(\mathbf{k})$, is given by:

$$637 \quad \gamma_t^2(\mathbf{k}) = 1 - \frac{(\alpha_T/\beta_T - \alpha_B/\beta_B)^2}{(1 + 1/A^2)(\alpha_T/\beta_T)^2 + (1 + A^2)(\alpha_B/\beta_B)^2} \quad (\text{A.1})$$

638 [c.f. Eqs (14), (11) and (12) in Banks et al. (2001)], where A , α_T , α_B , β_T , and β_B are all functions of
 639 wavenumber, \mathbf{k} :

$$640 \quad A(\mathbf{k}) = -f(\mathbf{k}) \frac{\rho_c g}{\Lambda(\mathbf{k}) + \Delta\rho g} \quad (\text{A.2})$$

641 and

$$642 \quad \alpha_T(\mathbf{k}) = \frac{-2\pi G \Delta\rho e^{-|\mathbf{k}|z_m}}{\Lambda(\mathbf{k}) + \rho_m g} \quad \alpha_B(\mathbf{k}) = \alpha_T(\mathbf{k}) + \frac{2\pi G e^{-|\mathbf{k}|z_l}}{g} \quad (\text{A.3})$$

$$643 \quad \beta_T(\mathbf{k}) = \beta_B(\mathbf{k}) + \frac{1}{\rho_c g} \quad \beta_B(\mathbf{k}) = \frac{-1}{\Lambda(\mathbf{k}) + \rho_m g} \quad (\text{A.4})$$

644 $\Lambda(\mathbf{k})$ is given by Eqs (18) or (26). Note that Banks's Eqs (19) are incorrect, and the above Eqs (A.3) and
 645 (A.4) are the corrected versions of these, updated for anisotropy. In the isotropic case $D_x = D_y = D$,
 646 $\mathbf{k} \rightarrow |\mathbf{k}|$, and $\Lambda(\mathbf{k}) \rightarrow D|\mathbf{k}|^4$, as noted in Section 3.1.

647 As mentioned in Section 4, in this study we have assumed $f(\mathbf{k}) = 1, \forall \mathbf{k}$.

648 A.2. Theoretical coherence from the Forsyth model

649 If Forsyth (1985)'s Eqs (4), (7) and (12) are substituted into his Eq. (25), the following analytical expression
 650 for a theoretical coherence can be derived, updated to account for anisotropy:

$$651 \quad \gamma_t^2(\mathbf{k}) = \frac{[\xi + \phi f^2 r^2]^2}{[\xi^2 + f^2 r^2][1 + \phi^2 f^2 r^2]} \quad (\text{A.5})$$

652 where ξ and ϕ are functions of \mathbf{k} (as is f) and are given in our Eq. (22), and $r = \rho_c/\Delta\rho$. Note that if the
 653 layer of variable density in the Banks model is placed at the Moho (i.e., $z_l = z_m$), the Banks and Forsyth
 654 theoretical coherences are identical.

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754 **Figure captions**

755 Fig. 1. The imaginary part of the Fourier transform of some real-valued data, showing the eleven con-
756 stituent Morlet wavelets that comprise an isotropic fan wavelet geometry of azimuthal extent $\Delta\theta = 180^\circ$.
757 Axes are u and v wavenumber in rad/km.

758 Fig. 2. The imaginary part of the Fourier transform of some real-valued data, showing the five constituent
759 Morlet wavelets that comprise an anisotropic fan wavelet geometry of central azimuth $\Theta = 30^\circ$ and azimuthal
760 extent $\Delta\theta = 90^\circ$. Axes are u and v wavenumber in rad/km.

761 Fig. 3. One of the random fractal surfaces (fractal dimension 2.5) used as input to form an initial load on
762 the orthotropic plate. Grey scale ± 3.5 (dimensionless).

763 Fig. 4. The global wavelet coherences computed at six indicated central azimuths (Θ), from a Forsyth-
764 style synthetic plate model (#34) with $T_x = 80$ km, $T_y = 40$ km and $\beta = 40^\circ$, with $f = 1$. The values
765 in parentheses after Θ are estimated T_e values for that azimuth after inversion of the individual coherence
766 profiles, assuming $f = 1$.

767 Fig. 5. Anisotropic elastic thickness ellipses from inversion of data for a Forsyth-style synthetic plate
768 model (#34) with $T_x = 80$ km, $T_y = 40$ km and $\beta = 40^\circ$, with $f = 1$. The relative lengths of the semi-major
769 and semi-minor ellipse axes indicate the degree of anisotropy in elastic thickness, while the orientation of
770 the major axis (shown by the inclined lines of length $2\sqrt{T_x T_y}$ at the ellipse centres) indicates the direction
771 of maximum mechanical strength. A model ellipse is shown to the right for comparison.

772 Fig. 6. Mechanical anisotropy in Australia: axes in the direction of T_{\min} and of length proportional to the
773 anisotropy $(T_{\max} - T_{\min})/T_{\max}$. Also shown are the crustal mega-elements of Shaw et al. (1995). Mercator
774 projection.

775 Fig. 7. Comparison of our axes of anisotropy in the direction of T_{\min} (averaged over a 720×720 km area)
776 (thin, black bars), with the Simons et al. (2003) “long-wavelength weak directions” (“good” data: thick,
777 white bars; “fair” data: thick, grey bars). Our axes are also shown where Simons et al. (2003) record “null”
778 or “bad” data. Mercator projection.

779 Fig. 8. Comparison of our axes of anisotropy in the direction of T_{\min} (averaged over a 400×400 km area)
780 (thin, black bars), with the mean stress orientations in the Australian stress provinces of Hillis and Reynolds
781 (2003) (thick, grey bars). Mercator projection.

782 Fig. 9. Comparison of our axes of anisotropy in the direction of T_{\min} (averaged over a 300×300 km area)
783 (thin, black bars), with the fast axes of seismic anisotropy from the tomographic model of Kennett et al.
784 (2004) at 75 km depth (thick, grey bars). Mercator projection.

785 Fig. 10. As Figure 9, but at 125 km depth.

786 Fig. 11. As Figure 9, but at 175 km depth.

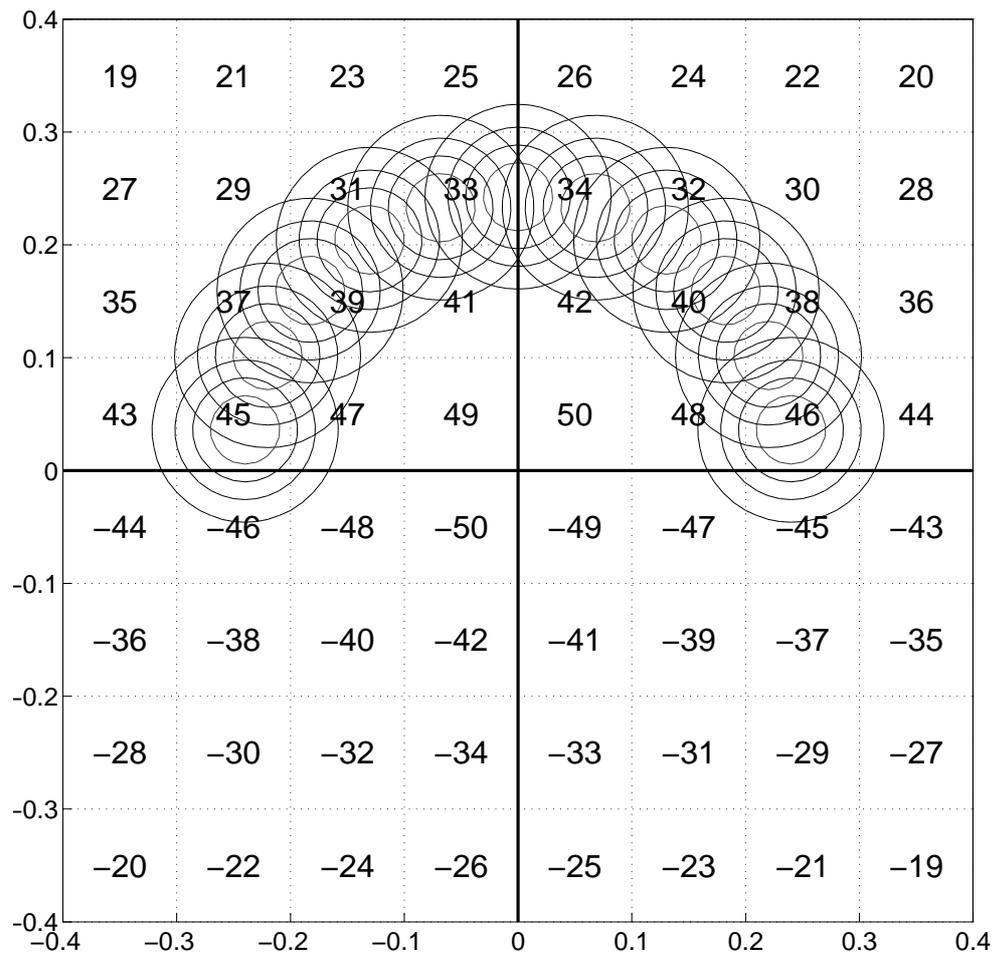


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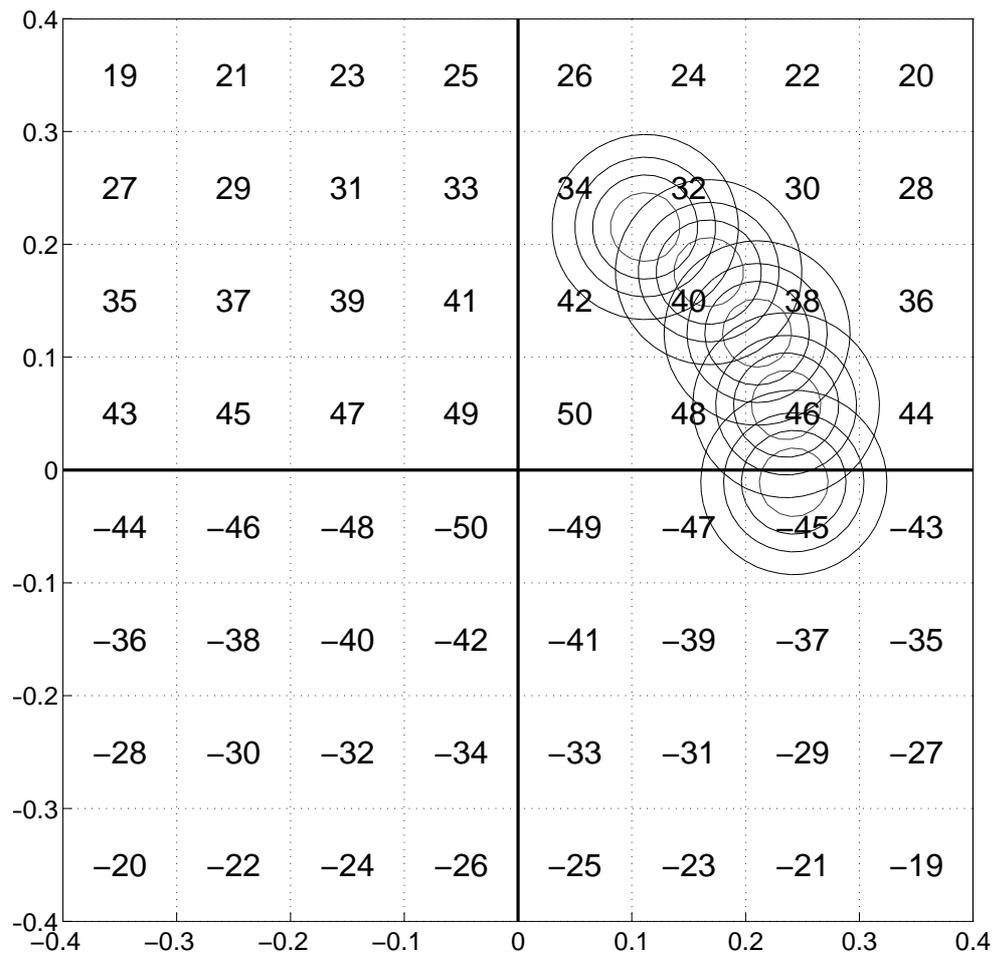


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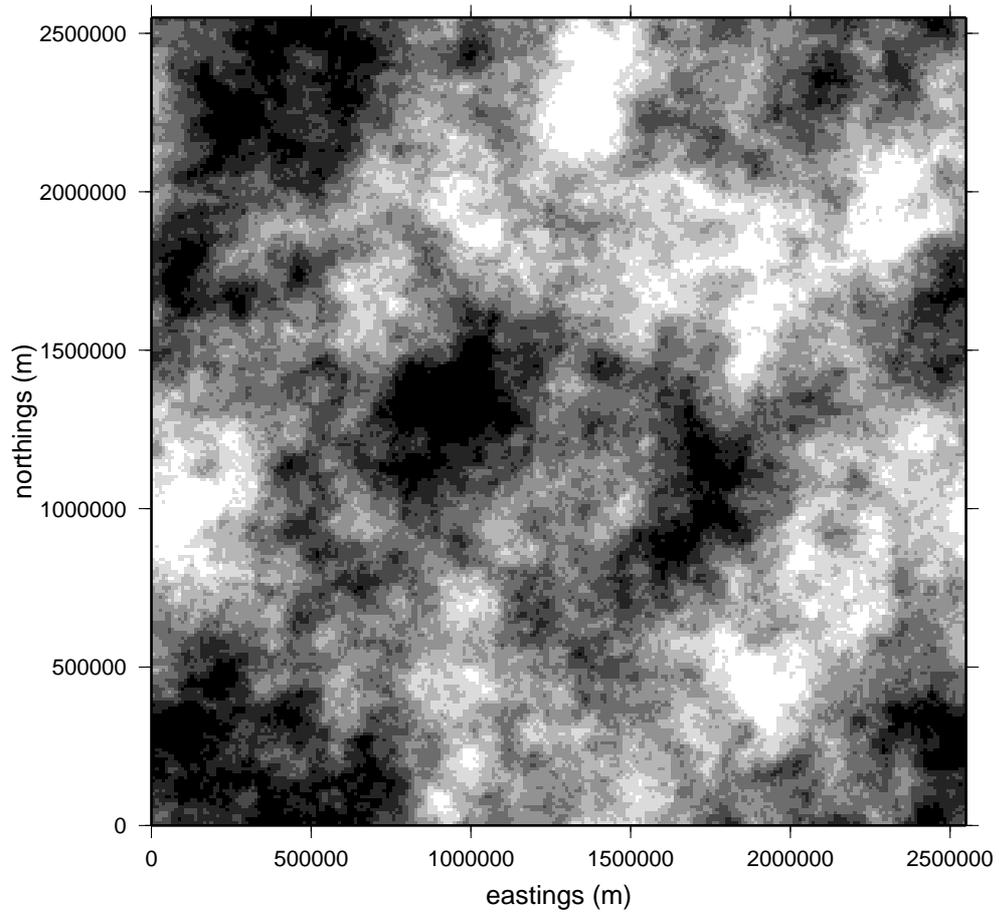


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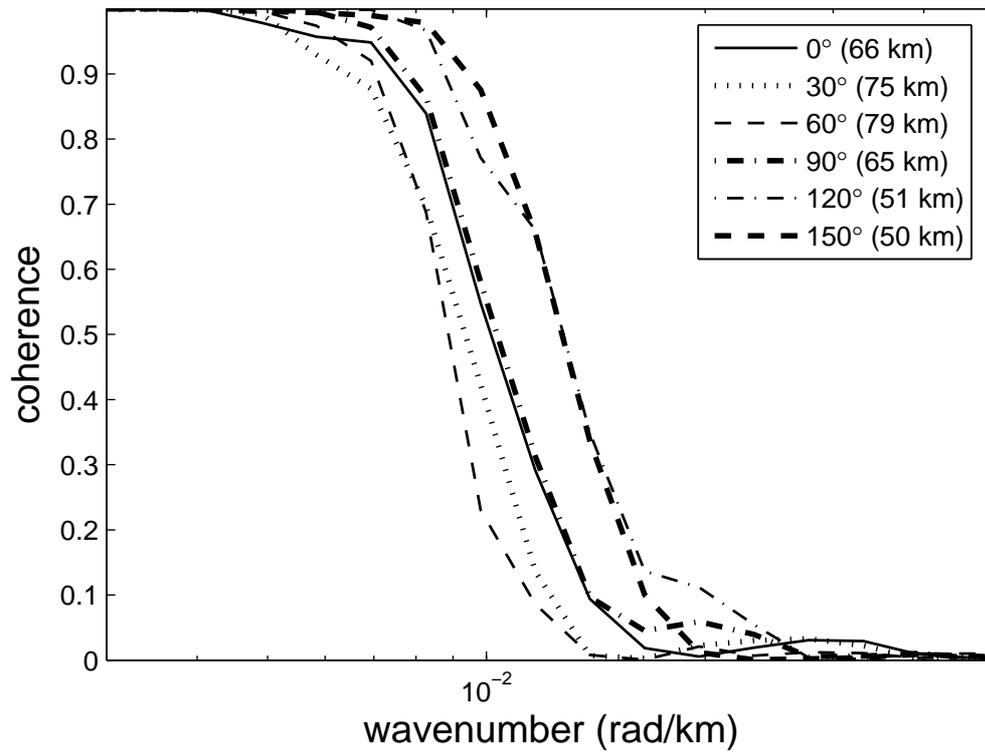


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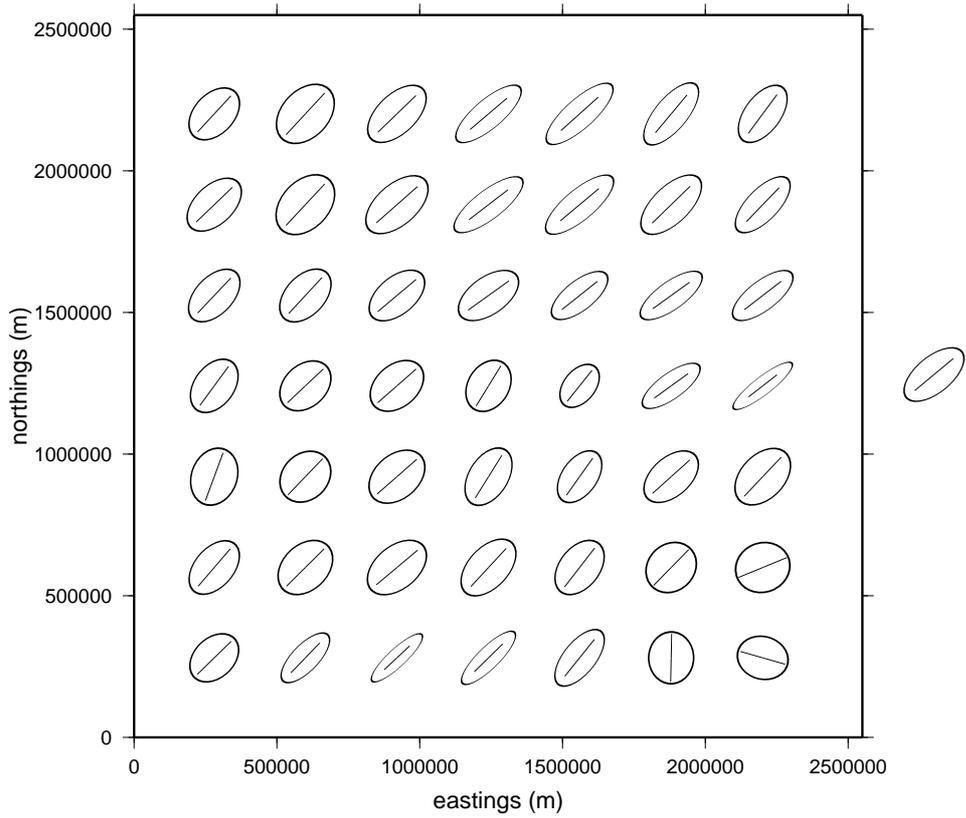


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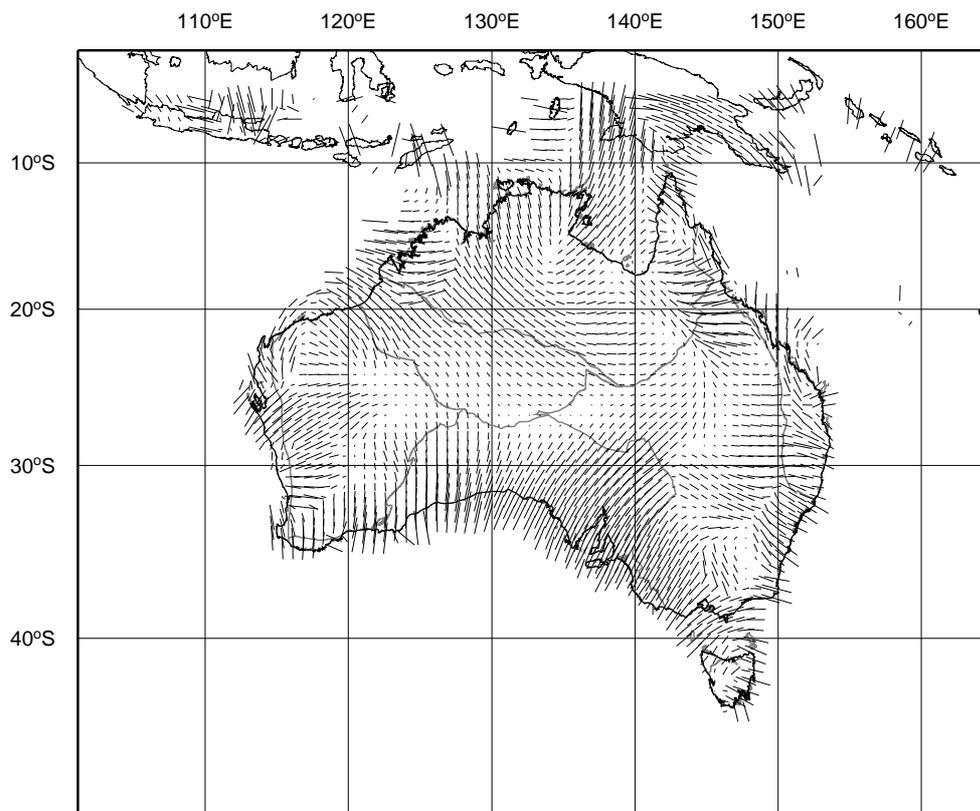


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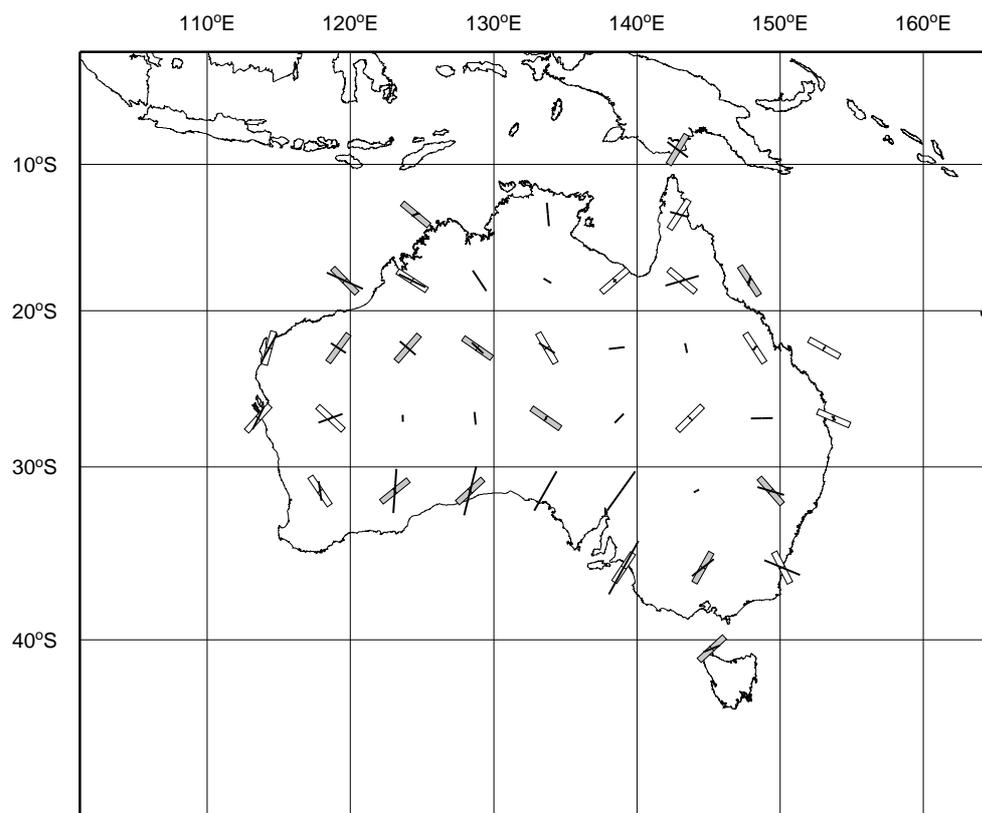


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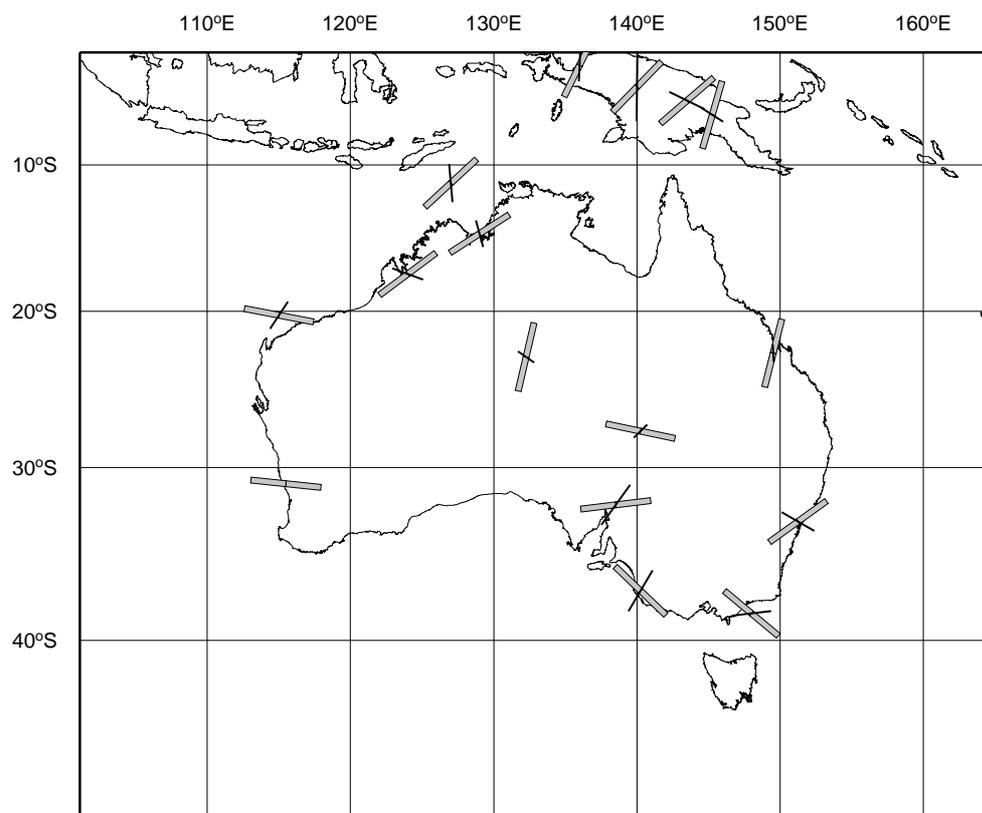


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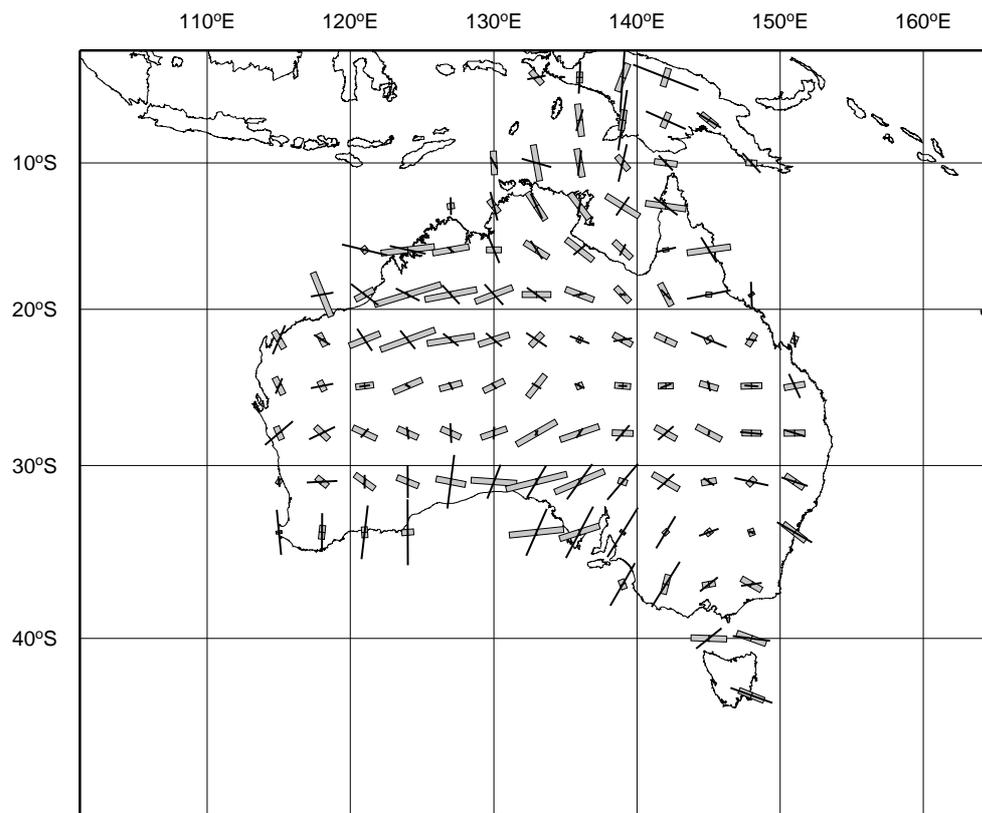


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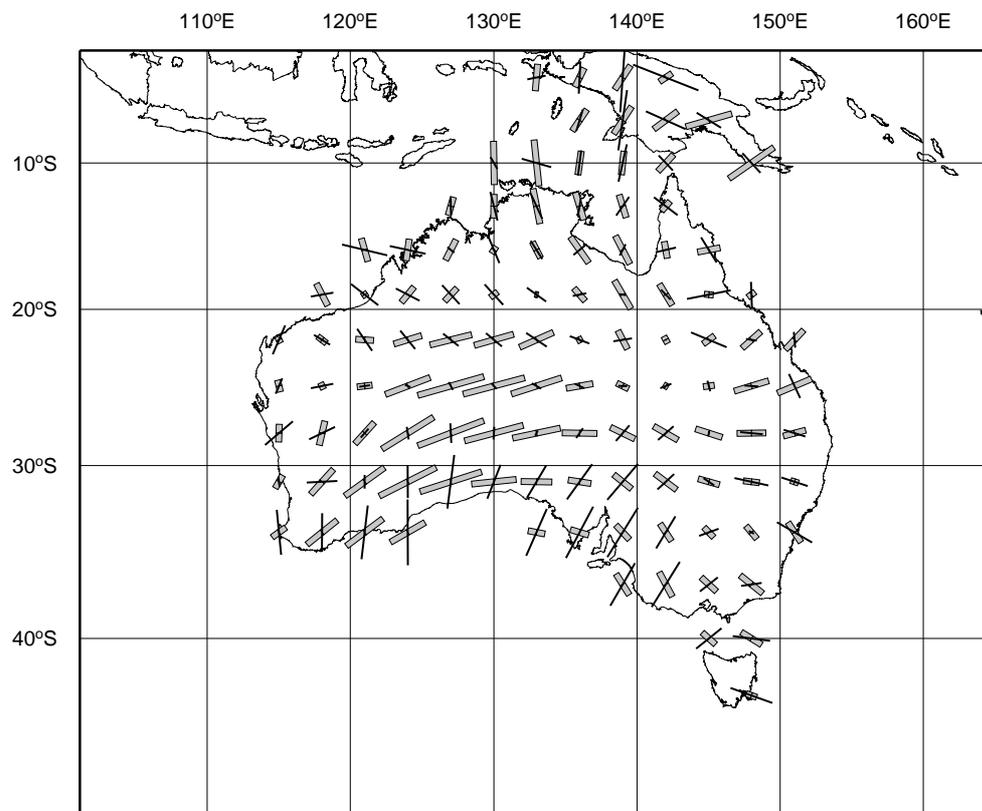


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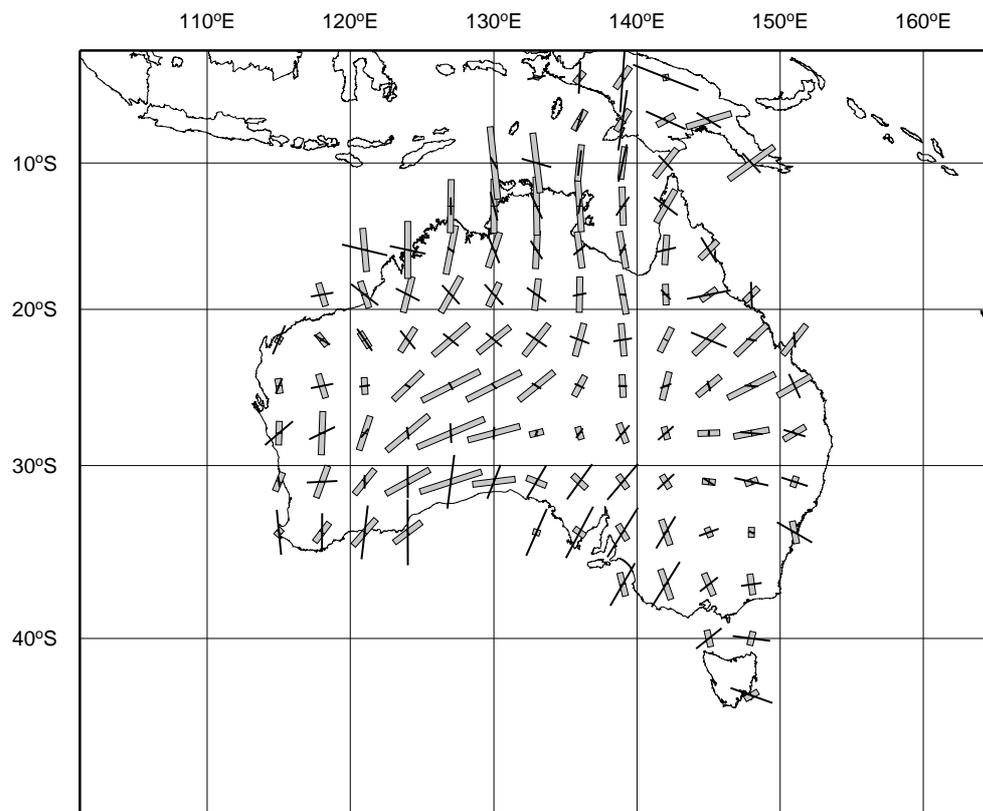


Fig. 11. As Figure 9, but at 175 km depth.