Formation of Australian continental margin highlands driven by plate-mantle interaction

R. Dietmar Müller^{a*}, Nicolas Flament^a, Kara J. Matthews^a, Simon E. Williams^a, and Michael Gurnis^b

^aEarthByte Group, School of Geosciences, The University of Sydney, New South Wales, 2006, Australia

^bSeismological Laboratory, California Institute of Technology, Pasadena, CA, USA

*Corresponding author: dietmar.muller@sydney.edu.au

Abstract

Passive margin highlands occur on most continents on Earth and play a critical role in the cycle of weathering, erosion, and atmospheric circulation. Yet, in contrast to the well-developed understanding of collisional mountain belts, such as the Alps and Himalayas, the origin of less elevated (1-2 km) passive margin highlands is still unknown. The eastern Australian highlands are a prime example of these plateaus, but compared to others they have a well-documented episodic uplift history spanning 120 million years. We use a series of mantle convection models to show that the time-dependent interaction of plate motion with mantle downwellings and upwellings accounts for the broad pattern of margin uplift phases. Initial dynamic uplift of 400-600 m from 120-80 Ma was driven by the eastward motion of eastern Australia's margin away from the sinking eastern Gondwana slab, followed by tectonic quiescence to about 60 Ma in the south (Snowy Mountains). Renewed uplift of ~700 m in the Snowy Mountains is propelled by the gradual motion of the margin over the edge of the large Pacific mantle upwelling. In contrast the northernmost portion of the highlands records continuous uplift from 120 Ma to present-day totalling about 800 m. The northern highlands experienced a continuous history of dynamic uplift, first due to the end of subduction to the east of Australia, then due to moving over a large passive mantle upwelling. In contrast, the

southern highlands started interacting with the edge of the large Pacific mantle upwelling ~40-50 million years later, resulting in a two-phase uplift history. Our results are in agreement with published uplift models derived from river profiles and the Cretaceous sediment influx into the Ceduna sub-basin offshore southeast Australia, reflecting the fundamental link between dynamic uplift, fluvial erosion and depositional pulses in basins distal to passive margin highlands.

Keywords: plate tectonics, geodynamics, dynamic topography, margin highland, uplift, subduction

1. Introduction

A wide range of mechanisms has been suggested to account for the formation and evolution of passive margin highlands. A recent review (Japsen et al., 2012) pointed out that most passive margin plateaus are too wide to be a product of plate flexure (Weissel and Karner, 1989) and they frequently exhibit renewed post-rift uplift, whereas rift-shoulder uplift would only produce one phase of uplift, accompanying continental rifting. Japsen et al. (2012) favoured lithospheric-scale folding at craton boundaries as a universal explanation for highland uplift, but observations supporting this idea are scarce, including a lack of evidence for lithospheric-scale folding in these regions, and some prominent highlands such as the eastern Australian highlands have formed entirely in relatively young Phanerozoic crust (Lambeck and Stephenson, 1986). Deformation and significant uplift due to intraplate stress changes through time can be accompanied by significant crustal deformation, expressed by thrust and reverse faults, which is readily observed in regions where it is the primary driver of uplift, for instance along the Flinders Ranges in South Australia (Clerier et al., 2005; Dyksterhuis and Mller, 2008). No such deformation is observed along the eastern highlands of Australia, driving the search for alternative hypotheses to explain their uplift. It has alternatively been argued that the eastern highlands represent an ancient remnant of the Paleozoic Lachlan foldbelt, and have experienced no post-Paleozoic tectonic uplift at all, in which case the Mesozoic-

Cenozoic uplift of the highlands would be entirely driven by erosional unloading (Lambeck and Stephenson, 1986; Stephenson and Lambeck, 1985) and the highlands should currently be in regional isostatic equilibrium. In contrast, Czarnota et al. (2014) suggested that the eastern highlands are not in isostatic equilibrium, but dynamically supported, based on their large positive gravity-topography admittance ratio. The idea that mantle dynamic topographic support may account for the elevation of passive margin highlands has been explored for the Brazilian Highlands (Flament et al., 2014), the South African escarpment (Braun et al., 2014; Flament et al., 2014; Gurnis et al., 2000; Lithgow-Bertelloni and Silver, 1998) and the Ethiopian and East African plateaus (Moucha and Forte, 2011). These cases all invoke interaction of continental margins with the African superswell, which can account for a single phase of uplift. However, many marginal highlands have experienced more than one uplift event. The Eastern Australian highlands display an initial uplift phase in the Late Cretaceous, followed by a mid-late Cenozoic renewal in uplift, with the timing and magnitude of uplift differing along strike (Czarnota et al., 2014). The age of the onset of uplift is constrained by elevated marine Cretaceous sedimentary rocks in the north and by uplifted Triassic deltaic sequences of the Sydney Basin in the south (Wellman, 1987). Renewed Cenozoic uplift has been particularly well documented in the southernmost portion of the eastern highlands. Uplifted Cenozoic fluvio-lacustrine deposits on the East Highlands in Victoria suggest deposition in lowland paleovalleys, incised into a widespread paleoplain surface that had a maximum paleorelief in the Eocene of less than 600 m, in contrast with their current location at elevations greater than 1400 m (Holdgate et al., 2008).

The successive phases of uplift and exhumation of the eastern highlands have resulted in erosion, transport and deposition of large volumes of sediments. South of the Snowy Mountains, the “Bassian Rise” within the Bass Strait is estimated to have been unloaded by 2.1 km (O’Sullivan et al., 2000). The southern coalfield of the Sydney Basin records 1.5 km of unroofing (Faiz et al., 2007), while the Bathurst batholith along the crest of the eastern highlands in New South Wales is estimated to have been uplifted between 1-2 km (O’Sullivan et al., 1995). Further north the Cooper-

Eromanga Basin records unloading by ~200-1000 m (Mavromatidis, 2006). In the Early Cretaceous the transcontinental Ceduna River is thought to have transported sediment from Queensland and New South Wales into the Ceduna sub-basin, forming a large delta (Lloyd et al., 2015; Norvick et al., 2008), whereas others (MacDonald et al., 2013) have argued that most Ceduna delta sediments are relatively locally-derived. The products of the Cenozoic uplift of the southern portion of the eastern highlands have been linked to sedimentary sequences in the Murray, Otway, Bass and Gippsland basins (Jones and Veevers, 1982).

Analysis of apatite fission track data has demonstrated rapid Early Cretaceous uplift within the eastern highlands in New South Wales (O'Sullivan et al., 1995), as well as rapid Cenozoic uplift around the Bass Strait at the southernmost extent of the eastern highlands (O'Sullivan et al., 2000), incompatible with the model of Lambeck and Stephenson (1986). O'Sullivan et al. (2000) speculate that exhumation is triggered by the opening of the Tasman Sea, whereas Czarnota et al. (2014) argue for a dynamic support mechanism driving the uplift, identifying temperature anomalies in the lithosphere as its potential cause. Here we provide a plausible mechanism for both exhumation events, linking plate motions to mantle convection through time, building on the work by Gurnis et al. (1998), and we test competing hypotheses for the distal versus proximal origin of Ceduna Delta sediments. We demonstrate that the time-dependent interaction of the Australian continental margin with a sinking Cretaceous eastern Gondwana slab and the large-scale mantle return flow in response to circum-Pacific subduction can explain most of the observed spatio-temporal uplift history.

2. Methods

We model global mantle flow based on the subduction history predicted by topologically-evolving plate boundaries (Seton et al., 2012), including a deforming plate model that accounts for the

extension of the continental margins south and east of Australia, leading to an improved reconstructed geometry of plate boundaries through time. The location of subduction along the eastern margin of Australia through time is a key parameter in the models, as it determines how the overriding plate has interacted with mantle convection since the Cretaceous. The subduction zone location depends on the existence or absence of back-arc basins through time, and on how we account for deformation during the Cretaceous rifting and breakup of East Gondwana. The Cretaceous was a time of major change in the geodynamic setting of the East Gondwana margin, with the previous long-lived active margin post-dated by rifting, continental breakup, and initiation of seafloor spreading. The rifting resulted in the formation of a largely submerged continent (the northern part of Zealandia) separated from Australia by oceanic crust of the Tasman Sea (Gaina et al., 1998) and from Antarctica by the Amundsen Sea (Larter et al., 2002). The switch to dominant extension is dated to around 105 Ma based on evidence from New Zealand and West Antarctica (Mortimer, 2014; Siddoway, 2008; Tulloch et al., 2009).

How Australia interacted with sinking slabs through time also depends on the absolute plate motion model. Here we use a hybrid absolute reference frame that minimizes net lithospheric rotation (Shephard et al., 2014) and test two alternative regional subduction models. The first tectonic model (Fig. 1, Reconstruction A) includes a large (~1000 km width at its maximum extent) Early Cretaceous (140-120 Ma) back-arc basin east of the Lord Howe Rise, representing the now subducted South Loyalty Basin which may have formed due to eastward rollback of the long-lived west-dipping eastern Gondwana subduction zone (Matthews et al., 2011); the alternative scenario (Fig. 1, Reconstruction B) is based on the premise that west-dipping subduction is continuous to the east of the Lord Howe Rise between 140-85 Ma, without a large intervening back-arc basin, and the South Loyalty Basin opens as a back-arc basin from 85-55 Ma, after which it is consumed by subduction (Matthews et al., 2015).

The plate models incorporate revised reconstructions of the eastern margin of Gondwana that include quantitative estimates of crustal deformation during East Gondwana breakup. The relative motion between Australia and Antarctica during the Cretaceous is constrained by restoration of extended continental crust (Williams et al., 2011) combined with additional kinematic constraints along the plate boundary described by Whittaker et al. (2013). For the rifting between eastern Australia and northern Zealandia (Lord Howe Rise), we build on previous studies focussed on the seafloor spreading history (Gaina et al., 1998) by using estimates of crustal thickness within the extended crust of Zealandia (Grobys et al., 2008) and the eastern Australian passive margin (Aitken, 2010). We adapt the method of Williams et al. (2011) to restore the crust within northern Zealandia to its pre-rift extent, thus defining the geometry of the margin of the East Gondwana continent in the early Cretaceous. We assume that extension occurred along small circles orthogonal to the trend of the continental ribbon and to the extensional basins formed within it, and consistent with the subsequent opening of the Tasman Sea defined by fracture zones within oceanic crust. The crustal restoration approach requires assuming a value for the average crustal thickness before the onset of extension, for which an appropriate value is poorly constrained. We tested a range of plausible values for crustal thickness along an active margin (Mantle and Collins, 2008), with the inferred locations of the restored East Gondwana margin reconstructed along small circles (Fig. 2). Our preferred reconstruction uses an average initial crustal thickness of 45 km, following previous suggestions that the present-day Chilean margin is an analogue for the Early Cretaceous margin of northern Zealandia (Collot et al., 2009; Veevers, 2006), and considering that the crust beneath New Zealand is estimated to have reached thicknesses exceeding 50 km around 120 Ma (Mantle and Collins, 2008).

Viscous mantle flow is driven by thermal convection with plate velocities applied as surface boundary conditions, extracted in 1 Myr intervals from the plate reconstructions (Bower et al., 2015). Our calculations begin at 230 Ma, but we only analyse mantle evolution from the Early Cretaceous

(~150 Ma) since it takes ~50 Myr for the models to reach an equilibrium from the initial condition (Flament et al., 2014). The initial condition includes a basal thermochemical layer 113 km thick just above the core–mantle boundary (CMB) that consists of material 4.2% denser than ambient mantle. This setup prevents the formation of upwelling mantle plumes, making it possible to study the interaction of moving continents with subduction-driven mantle downwellings and the associated large-scale mantle return flow in the absence of individual plumes. The thickness of the thermal lithosphere, derived from the age of the oceanic lithosphere and tectonothermal age of the continental lithosphere, is assimilated in the dynamic model. We use a modified version of the finite element code CitcomS to obtain one-sided subduction, in which the shallow portion of subducting slabs is imposed to a maximum depth of 350 km, below which mantle convection is fully dynamic (Bower et al., 2015). Air-loaded dynamic topography is calculated from the vertical stress resulting from mantle flow in restarts of the main model run in which the surface boundary condition is no-slip and the 250 km uppermost part of the mantle do not contribute to the flow. Key model parameters are listed in Table 1.

The Rayleigh number that determines the vigour of convection is defined by

$$Ra = \frac{\alpha_0 \rho_0 g_0 \Delta T h_M^3}{\kappa_0 \eta_0},$$

where α is the coefficient of thermal expansivity, ρ the density, g the acceleration of gravity, ΔT the temperature change across the mantle, h_M is the depth of the mantle, κ the thermal diffusivity, η the viscosity, and the subscript “0” indicates reference values. Values are listed in Table 2.

The viscosity varies by 1,000 due to temperature-dependence following

$$\eta = \eta_0(r) \exp\left(\frac{E_\eta}{T+T_\eta} - \frac{E_\eta}{T_b+T_\eta}\right),$$

where $\eta_0(r)$ is a pre-factor defined with respect to the reference viscosity η_0 for four layers (Table 1) E_η is the non-dimensional activation energy (E_{UM} in the upper mantle and E_{LM} in the lower mantle), T is the temperature, T_η is a temperature offset, and T_b is the ambient mantle temperature (values are

listed in Table 2). The average model resolution, obtained with $\sim 13 \times 10^6$ nodes and radial mesh refinement, is $\sim 50 \times 50 \times 15$ km at the surface, $\sim 28 \times 28 \times 27$ km at the core–mantle boundary (CMB), and $\sim 40 \times 40 \times 100$ km in the mid-mantle.

3. Dynamic surface topography evolution

We extract time-dependent surface dynamic topography since 150 Ma from our global coupled plate-mantle forward model. The best-fit Model 1 (Table 1), based on tectonic model B (Fig. 1) with Andean subduction along the Lord Howe Rise during the entire Cretaceous period, aligns the Cretaceous East Gondwana slab, which Australia overrides due to its time-progressive eastward motion, with the Australian-Antarctic Discordance (AAD) at present-day (Fig. 3). This is supported by a variety of geophysical and geochemical observations and previous geodynamic modelling (Gurnis et al., 1998; Ritzwoller et al., 2003). This model also captures the geometry of the East Gondwana and Micronesian slabs in the lower mantle west of the large Pacific mantle upwelling (Fig. 4). In this model, the eastward motion of Australia between 130 Ma and 100 Ma (Fig. 1) results in eastern Australia overriding previously subducted slab material sinking in the mantle, drawing the surface down and leading to continental inundation during this time, creating the Early Cretaceous infill of the Eromanga Basin (Gurnis et al., 1998). After ca. 100 Ma, subduction undergoes a hiatus along the eastern Gondwana margin and Australia stops moving eastwards (Rey and Müller, 2010). These changes lead to a gradual sinking of remnant slab material under eastern Australia, resulting in a broadening and amplitude reduction of the dynamic surface anomaly, causing gradual uplift and reflected in the disappearance of the Eromanga Sea after 100 Ma (Veevers, 1984), similar to the scenario proposed by Gurnis et al. (1998). Dynamic rebound and eastern Australian surface uplift continues for tens of millions of years, gradually slowing through time until about 40 Ma (Fig. 5a). After 40 Ma, a new NNW-SSE-oriented positive topographic surface anomaly appears along northeastern Australia, which gradually sweeps inland and eventually includes the entire eastern

Australian highland region at present-day (Fig. 5a). In our model this uplift is related to the mid-late Cenozoic north-northeastward motion of Australia, which gradually moves eastern Australia over the perimeter of the large mantle upwelling in the Pacific that occurs in response to circum-Pacific subduction. Because the eastern Gondwana slab, now straddling the Australian-Antarctic Discordance and central Australia (Fig. 3) marks the long-lived boundary between the Indian and Pacific ocean domains in terms of deep mantle structure (Gurnis et al., 1998; Ritzwoller et al., 2003), the large passive upwelling in the Pacific extends to the mantle underneath eastern Australia today, as opposed to being limited to the west by the post-Eocene Tonga-Kermadec subduction zone.

Our alternative Model 4 is based on the same mantle rheology as Model 1, but assimilates our tectonic model A (Fig. 1) as time-dependent surface boundary condition, assuming that a back-arc basin formed east of the Lord Howe Rise in the Cretaceous. The opening of such a back-arc basin substantially increases the distance between the eastern Gondwana subduction zone and eastern Australia. When Australia starts moving to the east after 130 Ma eastern Australia does not override recently subducted slab material in the upper mantle (which is located much farther to the east; Fig. 6). As a consequence, this model does not produce a substantial downdrawing of eastern Australia between 130 and 100 Ma, which would not lead to significant inundation of eastern Australia during this time, contrary to what is observed in the geological record (Veevers, 1984). The absence of substantial Early Cretaceous dynamic drawdown of eastern Australia in a model that places a subduction far east of the Lord Howe Rise was also observed by Gurnis et al. (1998). The Late Cenozoic post-30 Ma history of eastern highlands dynamic uplift is generally similar to Model 1, emphasizing that the geometry of the large-scale Pacific mantle return flow is not dependent on the details of ephemeral back-arc basin geometries (Fig. 6). Models 2 and 3 are equivalent to models 1 and 4 and only differ in terms of mantle rheology (Table 1). Their modeled dynamic surface topography mainly differs from models 1 and 4 in terms of amplitude, but not in terms of the spatial distribution of dynamic topographic anomalies (Figs 5, 6).

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230 Models 1 and 2, constructed using tectonic models without a Cretaceous back-arc basin east of the
231 Lord Howe Rise, match river profile inversions better than models 3 and 4, in which the Cretaceous
232 subduction zone is farther to the east. Model 1 produces a total uplift of ~400-600 m in the interval
233 between 120 and 90-80 Ma that roughly matches recent estimates from river profile inversion for
234 the Snowy Mountains, New England, the Central Highlands and the Atherton Tablelands (Czarnota et
235 al., 2014; Figs. 5-7). In our model the driving mechanism for this phase of uplift is rebound from the
236 eastward motion of Australia over a sinking slab, first leading to transient subsidence, followed by
237 rebound and uplift, as in the model by Gurnis et al. (1998). Model 1 predicts a slowing down of uplift
238 from 90-50 Ma in the Snowy Mountains and 80-60 Ma in New England, followed by renewed uplift of
239 500-700 m (Fig. 7). In the Central Highlands we find continuing, but distinctly slower uplift from
240 80 Ma to the present, totaling ~500 m. In Model 1 the mechanism for this second phase of uplift
241 represents the gradual motion of eastern Australia over the edge of the large passive Pacific mantle
242 upwelling (Figs. 1, 5). The Atherton Tablelands and Central Highlands (Fig. 7a, b) experienced the
243 influence of the perimeter of the mantle upwelling first, due to their more northerly location, more
244 proximal to the edge of the upwelling, resulting in a continuous history of uplift since the mid-
245 Cretaceous, whereas the Snowy Mountains started interacting with the edge of the mantle upwelling
246 ~40-50 Myr later, resulting in a distinct break in uplift, as also suggested by river profile inversion
247 (Czarnota et al., 2014).

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249 4. Discussion

250 The Atherton Tablelands (Fig. 7a) provide an excellent match between river profile inversion and
251 geodynamic models, resulting in continuing uplift at a rate of about 6.5 m Myr^{-1} since 120 Ma,
252 totaling ~800 m (Fig. 7a). In the Central Highlands (Fig. 7b) river profile inversion suggests a total
253 amount of ~600 m of uplift, whereas the magnitude of uplift of our preferred Model 1 is

overpredicted to be around 900 m. In New England (Fig. 7c) river profile inversion (Czarnota et al., 2014) suggests a distinct, phase of accelerated uplift from 80-50 Ma. In contrast, in our favoured models 1 and 2 the equivalent phase of uplift is predicted to occur between 120 and 80 Ma. However, the uplift magnitude matches well at about 800 m for both river profile inversion and geodynamic model (Fig. 7c). The mismatch in timing may reflect uncertainties in modeled subduction zone geometry and absolute plate motions. For instance, if the subduction zone were located farther east than modeled here at this latitude, then the continent would override it later, as shown previously (Gurnis et al., 1998). In the Snowy Mountains (Fig. 7d) our preferred Model 1 predicts an initial phase of uplift from 120 to 90 Ma, precisely as suggested by river profile inversion (Czarnota et al., 2014), and at a similar magnitude. The uplift magnitude for this time period is about 250 m from river profile inversion and 350 m from our best-fit geodynamic Model 1 (Fig. 7d), followed by a slowing of uplift, also seen in both models. Subsequently, a major phase of Cenozoic uplift occurs, totaling 700 m based on both river profile inversion (Czarnota et al., 2014) and our Model 1 (Fig. 7d). A distinct phase of post-Eocene uplift is also suggested by a range of geological and geomorphological observations (Holdgate et al., 2008).

The distinct two-phase uplift history for the eastern highlands implied by river profile inversion (Czarnota et al., 2014) and by our results are consistent with the thick accumulation of Cretaceous and Cenozoic sediments in the Ceduna Sub-Basin (Totterdell and Bradshaw, 2004; Fig. 8, see seismic profile location in Fig. 5). Sediment input into the Ceduna Sub-basin accelerated substantially after 100 Ma, with the deposition of the Albian/Cenomanian White Pointer Supersequence (Totterdell et al., 2000; Fig. 8). We relate this event to the dynamic rebound of Eastern Australia after the eastward motion of Australia and long-lived subduction along eastern Gondwana ceased. This process led to slab breakoff and the gradual sinking of slab material into the lower mantle, slowly broadening the area affected by dynamic rebound but decreasing its amplitude through time (Fig. 5), similar to the model by Gurnis et al. (1998). A characteristic of this process is its long time-constant, resulting in

continuing uplift over ~30-50 Myr, the uplift rates depending on the location (Figs 5, 6). This modeled long-term uplift is reflected in the sedimentation history in the Ceduna Sub-Basin, comprising the White Pointer, Tiger and Hammerhead sequences ranging from late Early Cretaceous to latest Cretaceous in age (~100 - 65 Ma; Fig. 8). These sediments are thought to be derived from a transcontinental Ceduna River (Norvick et al., 2008) based on zircon U-Pb geochronology and zircon Lu-Hf isotopes (Lloyd et al., 2015), and this is well-supported by our model. However, in a competing hypothesis detrital zircons had been interpreted to suggest that the upper part of the Ceduna Sub-basin sequence (Santonian–Maastrichtian, ~85-65 Ma) was sourced largely from basement eroded from the continental margin proximal to the basin (MacDonald et al., 2013), challenging the idea of a distal provenance of >2000 km from the eastern margin of Australia. A purely local origin of the post-85 Ma Ceduna sediments would rule out any large-scale post-85 Ma uplift of eastern Australia, as any such widespread uplift would inevitably have contributed substantially to sedimentation in the eastern Great Australian Bight, where the Ceduna sub-Basin is located, based on the estimated geometry of river systems at that time (Norvick et al., 2008). In contrast our preferred Model 1 produces continuing uplift of large parts of eastern Australia throughout the entire Late Cretaceous period (~100-65 Ma), consistent with the distal provenance model.

The cessation of this phase of uplift is represented by a 5–7 Myr hiatus in the earliest Paleocene (Totterdell et al., 2000), followed by deposition of a thin Paleocene-early Eocene carbonate sequence (Wobbegong Formation), which is unrelated to eastern highland uplift, as is the overlying Dugong Formation (Fig. 8). The renewed uplift of southeastern Australia, related to movement over the western rim of the large Pacific mantle upwelling (Figs. 1, 5), instead led to the deposition of clastic sequences in the Murray, Otway, Bass and Gippsland basins along the periphery of the southeastern highlands, a link made by Jones and Veevers (1982). The well-documented exhumation of the eastern highlands by up to ~2 km, over regions exceeding 1000 km in width, would have resulted in overall subsidence, as the wavelengths involved (> 500 km)

far exceed wavelengths characteristic of lithospheric flexure (Czarnota et al., 2014). Airy-isostatically compensated exhumation reduces crustal thickness and lowers surface height, as the gravitational potential of the lithosphere decreases (England and Molnar, 1990). This process would have attenuated the tectonic uplift driven by the normal stresses below the base of the lithosphere predicted by our model. Our dynamic uplift estimates are therefore minimum estimates of surface uplift.

5. Conclusions

We use a plate-mantle convection model to illustrate how the interaction of a moving continental edge with subduction, subduction cessation and slab breakoff, and the subsequent movement over the edge of a large-scale mantle upwelling can drive the successive, dynamic uplift of margin highlands. The initial dynamic uplift of the Australian eastern highlands (after 120-100 Ma) was driven by rebound due to the motion of Eastern Australia over and away from a sinking slab in the mantle, enhanced by the cessation of long-lived East Gondwana subduction around 100 Ma. Renewed dynamic uplift in the Cenozoic reflects the gradual motion of the margin over the edge of the large Pacific mantle upwelling. The model accounts for distinct spatial differences in the two-phase uplift history mapped by river profile inversion (Czarnota et al., 2014) and matches the associated two-phase sediment influx into the Ceduna Basin offshore southeast Australia. Ceduna Sub-basin stratigraphy reflects mid to Late Cretaceous Eastern Highland uplift that gradually slows as the dynamic rebound after cessation of East Gondwana subduction becomes broader and smaller in amplitude, ultimately resulting in a cessation of sedimentation and an unconformity in the Ceduna Sub-basin in the early Paleocene. Renewed Cenozoic eastern highland uplift, particularly pronounced in the Snowy Mountains and associated with deposition in the Murray, Otway, Bass and Gippsland basins can be linked to southeastern Australia overriding the edge of the large Pacific mantle upwelling. Geophysical and geological observations combined with geodynamic models provide a strong argument against the idea that the post 200 Ma uplift of the region was solely

driven by passive isostatic rebound due to erosion (Lambeck and Stephenson, 1986). Our model demonstrates that the time-dependent interaction of continental margins with both sinking slabs and deep mantle passive upwellings can contribute to elevating passive margin highlands.

Acknowledgments

R.D.M., K.J.M. and M.G. were supported by ARC Discovery Grant DP130101946 and R.D.M, N.F. and S.E.W. were supported by ARC ITRP grant IH130200012; M.G. was also supported by the NSF (through EAR-1161046 and EAR-1247022). This research was undertaken with the assistance of resources from the National Computational Infrastructure (NCI), which is supported by the Australian Government. We thank Sabin Zahirovic for help with drafting Figure 1, and Joel Potter for drafting Fig. 2. We are grateful to Alan Collins and an anonymous reviewer for their detailed comments, which helped improve the manuscript substantially.

Figure Captions

Fig. 1. The global relative plate motion Reconstruction A and associated plate boundary topologies (left column) are from Seton et al. (2012) with northern hemisphere modifications from Shephard et al. (2013) (not visible here), and southeast Asia modifications from Zahirovic et al. (2014). Reconstruction A includes a large (~1,000 km width at its maximum extent) Early Cretaceous (140-120 Ma) back-arc basin east of the Lord Howe Rise (a), representing the now subducted South Loyalty Basin (SLB) which may have formed due to eastward rollback of the long-lived west-dipping eastern Gondwana subduction zone. Reconstruction B includes continuous west-dipping subduction to the East of the Lord Howe Rise between 140-100 Ma, without a Cretaceous back-arc basin forming, followed by only brief subduction cessation after 100 Ma until subduction is re-initiated, perhaps due to cordilleran mountain belt collapse (Rey and Müller, 2010); the South Loyalty Basin is opening as a back arc basin from 85-55 Ma in this scenario (c), which is subsequently consumed by

subduction. Mid-ocean ridges and boundaries of rift systems are shown as black lines, whereas subduction zones are shown as toothed black lines; absolute plate velocities arrows are shown as black arrows. See text for additional details of relative and absolute plate motion models used. PAC – Pacific Plate, PHO – Phoenix Plate, IZA – Izanagi Plate, LHR – Lord Howe Rise, TAS – Tasman Sea.

Fig. 2. Methodology used for determining the restored location of the East Gondwana margin east of Australia prior to Cretaceous rifting and separation of Zealandia. The crustal thickness across the East Australian margin and northern Zealandia, the Lord Howe Rise (LHR) continental ribbon, is restored to its pre-extension shape, testing a range of values for the average crustal thickness before extension, with the restored boundaries between continental and ocean crust (RCOB) represented by filled coloured circles, with red circles representing our preferred model. The dashed white lines define the extent of stretched continental crust (as opposed to oceanic crust, or continental crust not disturbed during this rifting event). Small circles are derived from an Euler pole consistent with describing the early opening of the Tasman Sea (Gaina et al., 1998). Grey-shaded background image represents satellite-derived gravity anomalies, highlighting the crustal tectonic fabric (Sandwell et al., 2014).

Fig. 3. Geodynamic evolution of Australia since the Cretaceous period. A GPlates (www.gplates.org) visualisation of modelled mantle structure showing Australia and surrounding continents (translucent gray) at 120 (a), 80 (b), 40 (c) Ma and the present (d), with plate boundaries shown as thin white lines, active lithospheric extension indicated by meshes and mantle temperature illustrated by cold, sinking slab material coloured by depth (yellow and magenta in upper mantle and blue in lower mantle) and large mantle upwellings rising from the core-mantle boundary (red). AAD is Australian-Antarctic Discordance, CGS – Cretaceous East Gondwana Slab, JGS – Jurassic East Gondwana Slab, SPM – South Pacific mantle upwelling, Tonga – Tonga Slab.

Fig. 4. Modelled present-day mantle thermal structure (a), and seismic P-wave velocity anomalies from Li et al. (2008) (b) and S-wave anomalies from Grand (2002) on a map view showing lower mantle anomalies at 1920 km depth (c) and along a W-E oriented profile straddling northern Australia and the westernmost extension of the southwest Pacific mantle upwelling (b, d). Note that the modelled sinking Micronesian slabs (10% colder than ambient mantle) in the uppermost lower mantle and the Cretaceous Gondwana slabs (a) correspond broadly to imaged fast velocity regions in both P- (b) and S-wave (d) anomalies.

Fig. 5. Air-loaded dynamic topography of eastern Australia from 150 Ma to the present in 10 Myr steps from Model 1 (Table 1), in a fixed Australian reference frame. This model is based on Andean subduction along the eastern Gondwana margin in the Cretaceous (Fig. 1). The four stars correspond to four locations representing modelled uplift from river profile inversion (Czarnota et al., 2014) from south to north corresponding to the Snowy Mountains, New England, the Central Highlands, and the Atherton Tablelands. Note the tilting of eastern Australia towards the East Gondwana subduction zone from 150 to 100 Ma; the associated region of negative dynamic topography broadens towards 100 Ma due to East Gondwana's progressive eastward absolute plate motion from 130 to 100 Ma (Fig. 1). Cessation of subduction and slab breakoff after 100 Ma leads to rebound of eastern Australia and a phase of broad uplift and exhumation, continuing from 100 to ~40 Ma. After ~40 Ma the maps illustrate a stepwise uplift of eastern Australia, starting in the north, and gradually migrating to the south, as Australia migrates to the NNE, gradually overriding the edge of the large Pacific mantle upwelling, resulting in a second phase of eastern highlands uplift.

Fig. 6. Air-loaded dynamic topography of eastern Australia from 150 Ma to the present in 10 Myr steps from Model 4. This model is based on a Cretaceous back-arc basin forming east of Gondwana in the Cretaceous after 130 Ma (Fig. 1). Symbols as in Fig. 5.

Fig. 7. Modelled dynamic topography (coloured lines) from four alternative geodynamic models versus the uplift history (black lines) obtained from river profile inversion (Czarnota et al., 2014). Models 1 and 2 (blue and green), corresponding to tectonic models with Andean-type subduction along the eastern Lord Howe Rise in the Cretaceous, without a Cretaceous back-arc basin east of the Lord Howe Rise (Reconstruction A), match river profile inversions better than Models 3 and 4, in which the Cretaceous subduction zone is farther to the east (Reconstruction B), resulting in a lack of dynamic drawdown of Australia during the Cretaceous, not supported by geological observations which document widespread Cretaceous flooding of eastern Australia (Gallagher and Lambeck, 1989). Our modelled uplift amplitudes generally underestimate the uplift modelled from river profiles, especially in SE Australia, but the inflection points of uplift curves from our model show a good match with changes in uplift rates predicted by river profile models, especially in predicting two distinct uplift phases in the Snowy Mountains.

Fig. 8. Ceduna Sub-Basin stratigraphy. Interpreted NE-SW-oriented seismic section across the Ceduna Sub-Basin (see Fig. 5 for location), modified from Totterdell and Bradshaw (2004). The bulk of the sediments in the Ceduna Sub-Basin were deposited after ~100 Ma, starting with the Blue Whale Formation (dark blue). The long-sustained period of sediment input into this basin through the Late Cretaceous, reflected in the Blue Whale, White Pointer, Tiger and Hammerhead sequences, reflects the prolonged period of uplift and exhumation of eastern Australia predicted by our mantle convection model (Fig. 5, 6) during this period. The thin Paleocene–early Eocene Wobbecong Supersequence was deposited during the Paleocene–early Eocene after a hiatus of 5–7 Myr that produced an unconformity with the underlying Hammerhead succession (Totterdell et al., 2000). The overlying mid Eocene to Pleistocene Dugong Formation reflects a renewed increase in sedimentation rates, reflecting the renewed uplift of the eastern highlands due to Australia progressively overriding the edge of the large Pacific mantle upwelling (Figs. 5, 6).

Table 1: Tectonic boundary conditions and depth-dependence of viscosity of the models referred to in this study.

Name/Acronym	Plate Reconstruction	Viscosity profile
Model 1	B	1,1,1,100
Model 2	B	1,0.1,1,100
Model 3	A	1,0.1,1,100
Model 4	A	1,1,1,100

Models 1-4 refer to geodynamic model runs with two alternative plate tectonic reconstructions A (including a Cretaceous back-arc basin east of the Lord Howe Rise) and B (modelling Andean subduction along the eastern Cretaceous Lord Howe Rise, Fig. 1) and two different mantle viscosity profiles, either including a low-viscosity asthenosphere (Models 2 and 3) or not (Models 1 and 4). The viscosity profile is given by applying a factor to the reference model viscosity (1×10^{21} Pa s) above 160 km depth (lithosphere), between 160 and 310 km depth (asthenosphere), between 310 and 670 km depth (upper mantle) and below 670 km depth (lower mantle) in order to obtain depth-dependent viscosity in addition to temperature-dependent viscosity.

447 **Table 2:** Parameters common to all model cases. Subscript “0” denotes reference values.

Parameter	Symbol	Value	Units
Rayleigh Number	Ra	7.8×10^7	
Thermal expansion coefficient	α_0	3×10^{-5}	K^{-1}
Density	ρ_0	4000	$kg\ m^{-3}$
Gravity acceleration	g_0	9.81	$m\ s^{-2}$
Temperature change	ΔT	2825	K
Mantle thickness	h_M	2867	km
Thermal diffusivity	κ_0	1×10^{-6}	$m^2\ s^{-1}$
Viscosity	η_0	1×10^{21}	Pa s
Activation energy (upper mantle)	E_{UM}	100	$kJ\ mol^{-1}$
Activation energy (lower mantle)	E_{LM}	33	$kJ\ mol^{-1}$
Temperature offset	T_η	452	K
Background mantle temperature	T_b	1685	K
Radius of the Earth	R_0	6371	km

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Figure 1

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