The role of deep Earth dynamics in driving the flooding and emergence of New Guinea since the Jurassic

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The role of deep Earth dynamics in driving the flooding and emergence of New Guinea since the Jurassic

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Abstract

The paleogeography of New Guinea indicates fluctuating periods of flooding and emergence since the Jurassic, which are inconsistent with estimates of global sea level change since the Eocene. The role of deep Earth dynamics in explaining these discrepancies has not been explored, despite the strongly time-dependent geodynamic setting within which New Guinea has evolved. We aim to investigate the role of subduction-driven mantle flow in controlling long-wavelength dynamic topography and its manifestation in the regional sedimentary record, within a tectonically complex region leading to orogeny. We couple regionally refined global plate reconstructions with forward geodynamic models to compare trends of dynamic topography with estimates of eustasy and regional paleogeography. Qualitative corroboration of modelled mantle structure with equivalent tomographic profiles allows us to ground-truth the models. We show that predicted dynamic topography correlates with the paleogeographic record of New Guinea from the Jurassic to the present. We find that subduction at the East Gondwana margin locally enhanced the high eustatic sea levels from the Early Cretaceous (~145 Ma) to generate long-term regional flooding. During the Miocene, however, dynamic subsidence associated with subduction of the Maramuni Arc played a fundamental role in causing long-term inundation of New Guinea during a period of global sea level fall.

Key words: New Guinea, inundation history, mantle flow, dynamic topography, paleogeography
1. Introduction

The New Guinea margin is arguably one of the most tectonically complex settings in the world, comprising a diverse assemblage of accreted arc terranes, continental fragments, and obducted ophiolite belts (Baldwin et al., 2012). The geodynamic evolution of New Guinea in the post-Pangea period has been dominated by the long-term convergence between the Australian Plate in the southwest and the Pacific Plate in the northeast (Baldwin et al., 2012; Dow, 1977; Hill and Hall, 2003). The rapid north-northeast motion of the Australian Plate relative to the Pacific Plate since the Eocene, and the interaction with the Sunda continental promontory has resulted in oblique, arc-continent collisions and the slow growth of the island through successive accretionary episodes (Baldwin et al., 2012; van Ufford and Cloos, 2005). Such episodes include the accretion of ribbon terranes, which can be continental or oceanic such as the Torricelli-Finisterre Arc accreted in the middle to late Miocene, or a composite continental-oceanic terrane such as the Sepik terrane accreted during the Eocene-Oligocene (Zahirovic et al., 2016b; Zahirovic et al., 2014). These accreted terranes are typically ~100 km across, and more than ~1000 km long. Northern New Guinea has also undergone periods of rifting and lithospheric rupture to form ocean basins including the Sepik back-arc ocean basin in the Late Mesozoic supported by syn-rift sedimentation in the Early-Mid Jurassic followed by a breakup unconformity (Davies, 2012; Zahirovic et al., 2016b; Zahirovic et al., 2014). The region has also experienced intra-oceanic subduction and proposed subduction polarity reversal episodes, including those associated with the consumption of the Sepik back-arc in the latest Cretaceous (Baldwin et al., 2012; Hill and Hall, 2003). This subduction history has resulted in a lack of preserved seafloor, compounded by poor outcrop due to weathering, vegetation cover and inaccessible terrain on the continent, which
results in significantly uncertain tectonic reconstructions (Hill and Hall, 2003; van Ufford and Cloos, 2005; Zahirovic et al., 2014). Geologically, the northern half of the island overlies Mesozoic crystalline basement of ocean crust with arc affinities derived from the Pacific basin (Hill and Hall, 2003; van Ufford and Cloos, 2005), whilst the southern portion comprises Mesozoic and Tertiary passive margin strata underlain by Australian continental crust (Dow, 1977). The mountainous spine of the island comprising the highly deformed Mobile Belt delineates the north and south of New Guinea (Fig. 1).

In addition to the changing tectonic framework of New Guinea, the island also experienced alternating periods of short-term (related to Milankovitch cycles) and long-term (related to eustasy and mantle processes) inundation and emergence that remain preserved in the sedimentary record (Fig. 2). Throughout Mesozoic times, the continent was almost entirely inundated by shallow to deep seas with shelf-type and deep-marine sediments dispersed across the island (Dow, 1977). In contrast, the late-Eocene to early-Oligocene record is characterised by almost a total cessation in sedimentation, with a distinct lack of lower to middle Oligocene fossils throughout the oceanic crust and island arc terranes, particularly ordinarily pervasive foraminifera (Dow, 1977). This is likely attributable to the combined effects of global sea level fall (Haq, 2014; Haq et al., 1987) as well as the uplift and erosion corresponding to the late Eocene-early Oligocene orogeny (Dow, 1977; van Ufford and Cloos, 2005). The manifestation of this in the geological record is a pervasive, regional unconformity (Fig. 2) (Norvick, 2001) except for a belt of mixed-grade metamorphics, in southeastern Papua New Guinea including the Owen Stanley and Emo metamorphics which have been dated to between the Late Cretaceous and early Eocene (Worthing and Crawford, 1996). This tectonic uplift is further supported by the late Eocene...
intrusions in the north Sepik region along the marginal trough, and further south by
the vertical displacement of the Papuan Ultramafic Belt, along the Owen Stanley
Fault System (Davies, 2012; Dow, 1977). This non-depositional environment is
geologically short-lived with sedimentation resuming in the late Oligocene. Shallow-
water carbonates varying between 500 and 1500 m in thickness suggests flooding
peaked during the early- to mid-Miocene, despite inconsistencies with global sea level
estimates (Haq, 2014; Haq et al., 1987). From the Pliocene to the present, the flooding
somewhat retreated due to the combined effects of medium to short term sea level
change (Haq et al., 1987) as well as the shedding of debris from the accretion-related
mountains into the surrounding shallow shelf. The result today is a topographically
diverse and predominantly emergent island of New Guinea. Whilst the island likely
experienced short- and medium-term inundation patterns following regional sea-level
variations, the focus of this study is comparing long-term eustatic sea level (Haq,
2014; Haq et al., 1987) and long-term inundation and flooding indicated by
paleogeographic reconstructions.

Despite the well-documented paleogeographic record within New Guinea
(Dow, 1977; Norvick, 2003), there remain inconsistent correlations between the
mapped inundation patterns and global sea level curves (Haq, 2014; Haq et al., 1987).
Most previous work in the region has focused on unravelling the immense tectonic
complexity of New Guinea (Baldwin et al., 2012; Davies, 2012; Hill and Hall, 2003;
van Ufford and Cloos, 2005), with little focus on the role of deep Earth processes in
shaping the tectonic and topographic evolution of the northern Australian continental
margin. Such processes include subduction-driven mantle flow and its surface
expression of dynamic topography (Flament et al., 2013). Unlike crustal deformation
which occurs on spatial scales of 100 – 200 km resulting in elevations as large as
8 km, mantle-driven topography occurs over wavelengths of hundreds or thousands of kilometres with amplitudes typically no greater than 1.5 km (Winterbourne et al., 2014). Recent studies of Southeast Asia including that by Zahirovic et al. (2016a) and Clements et al. (2011) have highlighted the importance of considering dynamic topography to better understand the vertical motion of continents and their interaction with eustatic sea level change. In Sundaland for example, Clements et al. (2011) and Zahirovic et al. (2016a) linked regional dynamic uplift to plate tectonic history, attributing its Late Cretaceous-Eocene emergence to the collision of Gondwana-derived terranes and the associated subduction hiatus along the Sunda active margin.

Similar to Southeast Asia, over the last 160 Myr, the New Guinea margin has been part of the complex convergence zone between Eurasian, Indo-Australian and Pacific plates, which today manifests as a slab burial ground underlying much of the northern half of Australia (Heine et al., 2010; Li et al., 2008; Ritsema et al., 2011; Sandiford, 2007). Consequently, New Guinea represents an important case study with which to investigate the contribution of mantle flow to patterns of flooding and emergence. This study extends upon the work by Husson et al. (2014) and to an extent that of Flament et al. (2015) who explored the interplay of the plate-mantle system in tectonically complex orogenic settings.

We use global plate reconstructions with regional refinements that are applied as boundary conditions to forward geodynamic models to extract mantle evolution and predicted dynamic topography. The dynamic topography trends are compared to those interpreted from regional paleogeography and global sea level estimates. The mantle flow models are qualitatively compared to the mantle structure inferred from P- and S-wave seismic tomography.
2. Methods

2.1 Estimating flooding from paleogeography

Regional paleogeographic maps of Papua New Guinea from Dow (1977) allowed us to identify shifting paleo-environments and constraints on flooding history from the Jurassic to the present. Dow’s (1977) compilation includes eight maps of Papua New Guinea detailing patterns of sedimentation with a range of classifications including land, marine shelf sediments (probable and outcropping), marine trough (probable and outcropping), as well as basic delineations of outcropping volcanics and metamorphics. These maps were digitised, georeferenced and where necessary supplemented by the paleogeographic maps of Norvick (2003) to allow complete coverage of the New Guinea continental margin (Fig. 3). Norvick (2003) and Dow’s (1977) facies maps were simplified to two discrete groupings of land and marine to estimate flooding through time. To calculate the evolution of the extent of flooded areas, the paleogeographic polygons were reprojected into a cylindrical equal area coordinate system, with a central meridian of 145˚E and a standard parallel of 5˚S that are appropriate for New Guinea. From this, total and percentage values of inundation and emergence were calculated at the eight time intervals provided by Dow (1977), and subsequently used for comparison to eustatic sea level curves and predicted dynamic topography.

2.2 Plate tectonic reconstructions

Global plate motion models have evolved over many years of research with the latest reconstructions representing a synthesis of previous tectonic models refined and constrained through the accumulation of geological data. Due to the uncertainty associated with poorly constrained regions, we tested two alternate plate
reconstructions for New Guinea presented in Zahirovic et al. (2014) (Model A) and
Zahirovic et al. (2016b) (Model B). These reconstructions are differentiated by
regional refinements that make it possible to test end-member tectonic scenarios (see
Table 1). Importantly, these scenarios are used as surface boundary conditions for
mantle flow models, providing the evolution of plate boundaries, plate velocities,
thermal lithospheric thicknesses for oceans and continents and slab buoyancy flux
(Bower et al., 2015). The vertical component of topography resulting from the
modelled mantle convection is extracted to generate dynamic topography predictions
(see Section 2.3).

In the New Guinea region, the major differences between the two plate
reconstructions primarily concern the timing of tectonic events. In the model of
Zahirovic et al. (2014) (Model A) the rifting on northern New Guinea, which resulted
in the separation of the Sepik terrane, occurs in Late Cretaceous times (Hill and Hall,
2003), whilst the latest model by Zahirovic et al. (2016b) (Model B) places this event
in latest Jurassic times (~172 Ma) (Table 1) (Davies, 2012). Consequently, the
opening of the Sepik ocean basin occurs significantly later in Model A compared to
Model B, which uses the supra subduction zone (SSZ) ophiolites in the Central Irian
Ophiolite Belt (likely of latest Jurassic age) to mark this opening at ~157 ± 16 Ma
(Table 1) (Permana, 1998). Regional volcanism in the Early Cretaceous and the
Kondaku Tuffs (Dow, 1977) support an active New Guinea margin by Early
Cretaceous times, as well as latest Jurassic SSZ ophiolites, which are likely a remnant
of the long-lived East Gondwana active margin (Zahirovic et al., 2016b).

It remains difficult to determine the longevity of this ocean basin as minimal
seafloor spreading history has been preserved. Model A uses a late Paleogene age of
~35 – 31 Ma for the onset of north-dipping subduction along the Sepik as proposed by
Pigram and Davies (1987), whilst Model B uses the presence of ~68 Ma high-temperature metabasites in the West Papuan Ophiolite as suggested by Davies (2012) to support a significantly earlier age of subduction initiation at ~71 – 66 Ma (Table 1). The New Guinea margin subsequently experienced two collisional phases; one in the late Eocene to early Oligocene, the evidence for which remains preserved solely in eastern New Guinea (Crowhurst et al., 1996), and a second, island-wide collisional phase in Miocene times (Hill and Hall, 2003). In regards to the former, Zahirovic et al (2014) (Model A) interpreted cooling histories derived from K-Ar thermochronology in the New Guinea Mobile Belt to place the docking of the Sepik terrane between 27 – 18 Ma (Crowhurst et al., 1996). This is compared to Zahirovic et al.’s (2016b) (Model B) slightly earlier interpretation of 35 – 31 Ma, that was based on the Ar-Ar amphibolite age of Emo metamorphics, assuming the Sepik terrane docking was contemporaneous with compression in the Papuan Peninsula (Table 1) (Worthing and Crawford, 1996). It is likely that rather than representing alternate timing scenarios for collision, these different age interpretations reflect the diachronous collision east-west along the New Guinea margin.

Following this accretion, both reconstructions mark the presence of a south-dipping subduction zone to the north of New Guinea that accounts for the ~18 – 8 Ma Maramuni arc volcanism (Hill and Hall, 2003). The models differ in timing however, with Model A placing the subduction between 15 – 5 Ma, compared to an earlier date of ~20 – 10 Ma used in Model B (Table 1). The final phase of collision involved the accretion of the Halmahera-Torricelli-Finisterre Arc. Model A uses apatite fission track geochronology to mark the collision at ~6 Ma as proposed by Hill and Raza (1999), whilst Model B uses an earlier collision age of ~14 Ma as evidenced through K-Ar thermochronology (Table 1) (Crowhurst et al., 1996). As apparent, the complex
tectonic history of New Guinea has resulted in unresolved uncertainties in the plate reconstructions, and thus testing multiple kinematic reconstructions allows us to generate geodynamic scenarios that can be compared with the available surface geology and mantle seismic tomography.

2.3 Geodynamic modelling

We apply methods developed by Bower et al. (2015) that incorporate plate reconstructions into numerical models of mantle convection to predict past mantle flow. Viscous incompressible mantle flow was computed using the Boussinesq Approximation in the convection modelling code CitcomS (Zhong et al., 2008). To calculate mantle flow, surface boundary conditions were imposed including global plate velocities and lithospheric thicknesses derived from the plate reconstructions. The surface boundary conditions such as the thermal structure of the lithosphere and slabs above 350 km depth are progressively assimilated at 1 Myr time intervals following Bower et al. (2015). The implications of assimilating slab and lithosphere structure is that the upper boundary layer is no longer dynamic (i.e., imposed), which modifies the slab flux entering the mantle. However, due to the complex tectonic history in this region and the deep-time nature of the evolution (i.e., since Jurassic times), backward advection models (Conrad and Gurnis, 2003) are not suitable. In addition, the focus of this study is on long-wavelength dynamic topography, which primarily results from whole-mantle convection. All numerical models were computed from 230 Ma to the present to capture post-Pangea break-up and allow the flow models to reach a dynamic equilibrium from the initial conditions (Flament et al., 2014). Initial conditions at 230 Ma include a slab insertion depth of 1400 km with a dip angle of 45° above a depth of 425 km, and a dip angle of 90° below 425 km.
Further, to account for advective thickening (i.e., thickening due to an increase in viscosity), the slabs are initially twice as thick in the lower mantle as in the upper mantle in the initial condition at 230 Ma. The models are agnostic of mineral physics, and include an initial basal thermochemical layer that is 113 km thick above the core-mantle boundary, representing 2% of the volume of the mantle, consistent with the seismically inferred value (Hernlund and Houser, 2008). The material in this layer is 3.6% denser than the ambient mantle, corresponding to a buoyancy ratio of 0.5. This model set up supresses mantle plume formation thereby allowing for the subduction-driven dynamic topography signal to be isolated (Flament et al., 2014). The implications of this, as well as applying a non-adiabatic radial temperature profile and disregarding internal heating, results in a lower mantle that is somewhat colder than expected, with an overestimation of slab volumes.

The Rayleigh number ($Ra$) determines the vigour and style of convection, and is defined by:

$$Ra = \frac{\alpha_0 \rho_0 \gamma_0 \Delta T h_M^3}{\kappa_0 \eta_0}$$

where $\alpha$ is the coefficient of thermal expansivity, $\rho$ the density, $\gamma$ the acceleration due to gravity, $\Delta T$ the temperature difference between the surface and the CMB, $h_M$ the thickness of the mantle, $\kappa$ the thermal diffusivity, $\eta$ the dynamic viscosity, with the subscript “0” indicating reference values (see Table 2). We varied the viscosity profile based on stress and temperature, following

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

where $\eta_0(r)$ is a depth-dependent and pre-defined value with respect to a reference viscosity, $E_\eta$ is the activation energy ($E_{UM}$ in the upper mantle and $E_{LM}$ in the lower mantle), $T$ is the temperature, $T_\eta$ is a temperature offset, $T_b$ is the ambient mantle
temperature, and $R$ the universal gas constant. Whilst post-glacial rebound studies reasonably constrain the viscosity of the upper mantle (Fjeldskaar et al., 2000; Mitrovica and Forte, 2004), viscosity estimates of the lower mantle are less accurate and thus our models are designed to test a range of possible scenarios (see Fig. 4). Cases 1 to 3 are based on the plate reconstruction presented in Zahirovic et al (2014) (Model A) with varying viscosity profiles, whilst a fourth case utilises the reconstruction by Zahirovic et al. (2016b) (Model B). In each case, with respect to the reference viscosity ($\eta$), the depth-dependent viscosity $\eta_0(r)$ is multiplied by a factor of; 1 above 160 km; either 0.1 or 1 between 160–310 km depth (with and without an asthenosphere respectively); 1 between 310–660 km depth; and either 100 or linearly increasing from 10 to 100 in the lower mantle between 660 km and the core–mantle boundary (CMB) (Fig. 4) (Steinberger and Calderwood, 2006). This radial viscosity pre-factor is also applied to the assimilated slab material in the lower viscosity asthenosphere. However, due to the temperature-dependent viscosity used in these models, the slabs retain a larger relative viscosity than the asthenosphere. The present-day volume-averaged viscosity for Case 4 is $41.6 \times 10^{21}$ Pa s. Here, it is important to test a range of viscosity scenarios for the lower mantle as it plays a significant role in the observed dynamic topography trends, accounting for on average 58% of the observed signal as derived from Case 4 (Supp. Fig. 1). 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 near the CMB, and $\sim 40 \times 40 \times 100$ km in the mid-mantle. With this model setup, we hope to quantify some of the uncertainty in the region by capturing possible end-member scenarios for the New Guinea margin.
We computed time-dependent dynamic topography ($h$), at intervals of 10 Myr from 230 Ma to the present following:

$$h = \frac{\sigma_{rr}}{\Delta \rho g_0},$$

where $\sigma_{rr}$ and $\Delta \rho$ are the radial component of stress and the density difference between the shallow mantle ($\rho_{\text{UM}} = 3340 \text{ kg m}^{-3}$) and sea water ($\rho_{\text{w}} = 1030 \text{ kg m}^{-3}$) respectively (see Table 2 for other parameters). Water-loaded dynamic topography is calculated from the total normal stress resulting from mantle flow but excludes buoyancy and lateral viscosity variations above a depth of 350 km, which is the maximum depth to which slabs are inserted using time-dependent upper boundary conditions (Bower et al., 2015). This results in a relatively low amplitude of dynamic topography, as convection closer to the upper thermo-chemical boundary layer generates stronger dynamic topography signals. However, due to the synthetic insertion of slabs down to 350 km, the procedure to exclude these shallow depths is necessary. The output dynamic topography and mantle evolution is then coupled with the aforementioned plate reconstructions, to present a series of modelled snapshots and vertical profiles from the latest Jurassic to the present (~160 – 0 Ma) (Fig. 5). In addition, we extracted the predicted dynamic topography at specified points in New Guinea (Figs 5 and 6) to obtain the point-specific evolution of dynamic topography for all four cases from 160 Ma to the present (Fig. 7).

3. Results

3.1 Comparison of predicted present-day mantle structure to seismic tomography

We qualitatively compare vertical profiles of predicted mantle temperature (Fig. 5) to seismic tomography models (Fig. 8). Model temperature anomalies are
compared to seismic tomography velocity anomalies, assuming the latter result primarily from thermal perturbations (Becker and Boschi, 2002; Grand, 2002). We use a combination of P- and S-wave tomographic models, with the former providing high resolution mantle imaging beneath subduction zones and continents (Romanowicz, 2003), and the latter providing a more uniform global coverage of the mantle and more equal sampling of the lower mantle (Grand, 2002). Whilst the seismic tomography models share first-order similarities, they differ on scales smaller than several hundreds of kilometres (Fig. 8) and this is due to model resolution, data collection biases including the earthquake sources used and the earthquake relocations applied as well as crustal correction and model parameterisation (Grand, 2002; Romanowicz, 2008). Tomographic models have generally been shown to correlate poorly with geodynamic models for spherical harmonic degrees ≥ 5 (Becker and Boschi, 2002) and thus here we visually concentrate on long-wavelength correlations. The high resolution P-wave models MIT-P08 (Li et al., 2008) (Fig. 8b) and GAP-P4 (Obayashi et al., 2013) (Fig. 8c) are utilised in conjunction with the S-wave models S40rts (Ritsema et al., 2011) (Fig. 8d) and MontelliS (Montelli et al., 2006) (Fig. 8e).

Our numerical models exhibit reasonable compatibility with the positive seismic velocity anomalies in P- and S-wave tomographic models (Fig. 8), with Case 4 arguably generating the best match to mantle structure. In the mid-mantle, notable discrepancies are observed between the four modelled cases, particularly regarding the position of the Sepik slab which in Case 4 (Fig. 8) is located between 25°S and 30°S at a depth of approximately 1000 km. Though slightly underestimating its volume, Case 4 reproduces good estimates for the depth and lateral position of the subducted Sepik slab, compared to Cases 1–3 that display distinct lateral offsets. These differences are likely attributable to the placement and timing of the associated
Sepik subduction zone, with the results supporting an earlier initiation of subduction at ~70 Ma as modified in Model B by Zahirovic et al. (2016b). Similarly, Case 4 again produces better predictions of the Caroline slab in regard to both its depth and geometry with correlations observed in both the P- and S-wave seismic tomography. The Maramuni slab and northern New Guinea subduction do not fare as well, with all cases seeing lateral offsets and incorrect estimations of slab volumes. These offsets are likely a function of the model setup, particularly errors in the imposed subduction history (Table 1), as well as the choice of radial viscosity (Fig. 4). Comparatively, Model B (Zahirovic et al., 2016b) better predicts the depth of the Maramuni slab; however, the interpretations regarding the northern New Guinea slab suggest further plate reconstruction refinements are required. Overall, we note the better performance of Case 4 in its reproduction of the mantle structure, with these observations providing reasonable confidence in the associated dynamic topography predictions, but also direction for future modifications to Model B, the plate tectonic reconstruction by Zahirovic et al. (2016b).

3.2 Comparison of time-dependent dynamic topography to paleogeography

The reasonable visual agreement between the mantle structure produced by the flow models and seismic tomography models encourage us to analyse the time-varying prediction of dynamic topography trends, which are likely to provide insight into the dynamic uplift and subsidence of the region. The modelled evolution of dynamic topography is characterised by periods of subsidence and uplift that are similar to the alternating periods of flooding and emergence preserved in the regional sedimentary record (Fig. 6). The dynamic topography signals varied only minimally between the four tested scenarios, with the observed differences attributable primarily
to relative plate motions as well as variations in radial viscosity. In the models, subduction history plays a key role in determining dynamic topography trends. The paleogeographic record (Figs 2, 3 and 6a) provides a constraint with which to compare our topography predictions. The mantle flow models present a period of dynamic subsidence from the latest Jurassic to Early Cretaceous times (Fig. 6b) likely associated with the descending slabs of the East Gondwana active margin. This subsidence is consistent with the regional paleogeography, which records the deposition of shallow and deep marine sediments (Fig. 2) and according to Dow (1977) and Norvick (2003) represents a long period of continental inundation (Fig. 6a). The gradual dynamic uplift that follows can be linked to slab roll-back of the same East Gondwana subduction zone whereby dynamic subsidence associated with the subducted slabs decreases, resulting in relative uplift of the surface (see Digital Supplement). Regional dynamic subsidence was subsequently re-established due to the onset of the north-dipping subduction of the Sepik oceanic basin. Here, Cases 1–3 show a timing lag relative to Case 4, attributable to the later onset of the Sepik subduction, initiating at ~ 35 Ma in the plate reconstruction of Zahirovic et al. (2014) (Model A) compared to ~ 71 Ma in the plate reconstruction of Zahirovic et al (2016b) (Model B). This timing offset is similarly observed in the final period of dynamic subsidence throughout Miocene times, which in all cases is linked to the south-dipping Maramuni Arc subduction zone on the northern margin of New Guinea. In the latest plate motion model the Maramuni subduction is initiated at 23 Ma compared to a later onset of 15 Ma in the earlier reconstruction, and this is manifest in dynamic subsidence from 18 Ma in Case 4 compared to 8 Ma in Cases 1–3. This period of dynamic subsidence is again validated using the paleogeographic record,
which preserves a history of widespread continental inundation throughout Miocene times (Fig. 6).

We also compare modelled dynamic topography at specified points in New Guinea, namely Irian Jaya (P1), central New Guinea (P2) and Papua New Guinea (P3) (Fig. 1 and 5), to study the regional variation of subsidence and uplift trends (Fig. 7). Moreover, with eastern and western New Guinea experiencing greater amplitudes of dynamic change, these trends provide possible end-member scenarios on the signals derived from central New Guinea, which is influenced by the Southeast Asian Sunda slabs to the northwest and the New Guinea and Pacific slabs to the east (Figs 5 and 6).

Figure 7 highlights that whilst the general trends and timing are the same across all point locations, eastern and western New Guinea experienced an opposite net dynamic movement over time, with the downward continental tilt reversing from eastward at 160 Ma, to westward at present. This long-term signal superimposed beneath the temporally shorter dynamic topography trends highlights the complexity and spatio-temporal variation of mantle dynamics influencing regional topography. Furthermore, it must be emphasised that in a global context, New Guinea is a relatively small continent located in a long-wavelength dynamic topography low (Fig. 5) and that the dynamic topography trends presented here primarily reflect the motion of the continent over individual slabs associated with regional subduction systems.

4. Discussion

Our analysis suggests that inundation patterns through time in New Guinea could be controlled by an interplay of deep Earth and surface processes. Our results suggest a likely influence of dynamic topography on the long-term continental
flooding patterns of New Guinea since the Jurassic. Due to the difficulty in constraining the amplitude of dynamic topography our analysis focuses on trends rather than absolute values (Fig. 6). Of significance is the correlation between the two periods of widespread flooding in New Guinea representing Cretaceous and Miocene times (Figs 2, 3 and 6), and the modelled periods of dynamic subsidence. The flooding of New Guinea from 145 – 90 Ma is influenced by high global sea levels (Haq, 2014; Haq et al., 1987) amplified by dynamic subsidence linked to the East Gondwana slab (Figs 5 and 6) (Zahirovic et al., 2016b). Gurnis et al. (1998) supported the origin of this subducted slab based on the existence of a converging margin between the Pacific plates and Gondwanaland since at least ~200 Ma. Similarly, they explored the surface expression of this subducting slab focusing on the anomalous vertical motion of eastern Australia throughout the Cretaceous (Gurnis et al., 1998). Notably however, their modelled East Gondwana trench was a simplification of the much more extensive Gondwanaland-Pacific margin, which they noted may have extended further towards New Guinea. Here, we incorporate the northwestward extension of this converging margin (Fig. 5), and link the contribution of this slab to the continental subsidence of New Guinea.

During the Miocene, mantle dynamics hold greater significance with dynamic subsidence trends alone correlating with the widespread inundation observed in New Guinea (Fig. 6). During this time, all four scenarios present dynamic subsidence corresponding to the subducted Maramuni slab, in contrast to global eustatic sea level trends which are falling from ~35 Ma (Haq, 2014; Haq et al., 1987). Our models suggest that sinking slabs are drawing the continent down faster than long-term sea level is falling, causing regional flooding of New Guinea, similar to Southeast Asia (Zahirovic et al., 2016). The models suggest this slab now resides in the lower mantle
at depths of between ~850 – 1350 km beneath the Gulf of Carpentaria (Fig. 8). This is compatible with the work of Heine et al. (2010) and Sandiford (2007) who correlated Australia’s northward tilt since mid-Miocene times to a slab graveyard beneath northern Australia.

In contrast, during the Paleogene, the predicted dynamic subsidence associated with the subduction of the Sepik back-arc basin exhibits notably weaker correlations with continental inundation as recorded by the paleogeography (Dow, 1977; Norvick, 2003). At times contemporaneous to the Sepik subduction and in the millions of years following, New Guinea is characterised by regional uplift and a non-depositional environment (Figs 3 and 6), as opposed to continental flooding. We suggest that tectonic processes, including terrane collision and accretion associated with the docking of the Sepik terrane in the late Eocene to early Oligocene, are likely responsible for the observed margin-wide unconformity.

Indeed, the highly active tectonic setting of the region also demands the consideration of tectonic processes in comprehensively isolating the mechanisms behind regional flooding. For example, lithospheric flexure resulting from orogenic loading is a plausible tectonic mechanism for continental subsidence (DeCelles and Giles, 1996). Since Jurassic times, New Guinea has experienced two main periods of orogenesis associated with the collision and accretion of Sepik terranes in early Oligocene times (~35 – 31 Ma) and the accretion of the Halmahera-Torricelli-Finisterre-Arc in mid-Miocene times (~14 Ma) (Zahirovic et al., 2016b). However, in both cases the paleogeographic record provides clear evidence for the onset of flooding prior to orogenesis, with marine sediments present from as early as the ~66 Ma preceding the first collision, and ~25 Ma preceding the second collision (Fig. 2) (Dow, 1977; Norvick, 2003). This timeline suggests orogenic loading associated
with New Guinea’s fold and thrust belts could not have initiated the observed flooding. Work by Abers and Lyon - Caen (1990) on the limited extent of eastern New Guinea’s foreland basin further supports this hypothesis. Their investigation reveals that whilst the Australian plate underlying the foreland is relatively strong with flexural rigidities of $10^{24}$ to $10^{25}$ Newton metres (Nm), localised plate weakening beneath the eastern highlands results in a small and shallow foreland basin. Abers and Lyon - Caen (1990) suggested this weakness could be related to a combination of thermal and mechanical processes resulting from Quaternary volcanism and thick-skinned midcrustal faulting throughout the thrust belt. The overall result is that the Australian lithosphere saw little deflection under the loading of New Guinea’s eastern highlands, with the foreland basin extending no more than 20 – 60 km from the mountain front at present day (Abers and Lyon - Caen, 1990). Others, including Pigram et al. (1989), have investigated the development of the foreland basin since the Oligocene and argue for a significantly larger flexural basin up to 600 -700 km wide, south of the Fold and Thrust Belt. However, their analysis failed to incorporate elastic thickness, which varied from 40 - 60 km in eastern Papua New Guinea to 70 - 80 km in the west (Haddad and Watts, 1999), which is incompatible with such flexural basin widths. In the India-Eurasia collision zone for example, the elastic thickness of the downgoing craton is estimated to be ~70-125 km, (ignoring the effect of dynamic topography), with a foreland basin width of ~300 - 400 km (Jordan and Watts, 2005; Tesauro et al., 2012). It is therefore unlikely that the large foreland basin width on Papua New Guinea suggested by Pigram et al. (1989) is solely flexural, and can rather be explained spatially and temporally with the addition of a dynamic topography component from the south-dipping subduction related to the Maramuni Arc, much like the manifestation of the Cretaceous-age epicontinental Western
interior Seaway (USA) and the Eromanga Sea (Australia). The study by Husson et al. (2014) provides a precedent for this argument with their results suggesting the uplift history of the Himalayas and the subsidence of its foreland basin cannot be explained without considering the effects of dynamic topography. In light of this, Husson et al. (2014) emphasises the need for revising estimates of elastic thickness to incorporate the effects of dynamic topography in regions particularly effected by subduction-driven mantle flow. In this regard, whilst lithospheric flexure likely played a minor role in generating continental inundation, the regional flooding observed during the Miocene (Figs 2, 3 and 6) cannot be explained without considering the effect of mantle-driven dynamic topography during a time of global sea level fall. Through the development of models that incorporate elastic thickness and foreland basin flexure, the comparative contribution of these processes can be more realistically ascertained.

Despite the promising results presented here, these models, like all numerical experiments, are inherently limited by their input parameters and simplifying assumptions of complex Earth processes. For example, our current methods use rigid plate motion models that do not incorporate continental deformation including the key periods of orogenesis that characterise the tectonic history of our study area. Future work should incorporate deforming plate reconstructions for the New Guinea margin into forward geodynamic models. This would allow for the consideration of the effects of crustal thickening associated with such collisions as that of the Sepik terrane and the Halmahera-Finisterre-Torricelli Arc (Baldwin et al., 2012) on mantle flow and dynamic topography. Moreover, future tectonic models should see refinements to both the absolute plate motions in deep time and regional plate motions.
towards present day that aim to correct for the lateral offsets and incorrect estimation of slab volumes (Fig. 8).

5. Conclusions

Here we investigate the interplay of the plate-mantle system and its impact on the vertical motion of continents in complex areas of orogeny. Using a case study of the northern Australian continental margin we couple global plate reconstructions with forward geodynamic models to predict the influence of the mantle flow on ancient patterns of flooding and emergence. Our results provide support for subduction driven dynamic uplift and subsidence from the Jurassic to the present, with our model predictions being in agreement with the paleogeographic record. We predict subduction at the East Gondwana margin prior to the Cretaceous provided a positive feedback with higher eustatic sea levels to generate long-term regional flooding. During the Miocene, however, subduction that produced the Maramuni arc played a fundamental role in causing the widespread inundation, with evidence suggesting contemporaneous long-term sea-level fall. As for the subduction of the Jurassic-Cretaceous age Sepik back-arc basin, local collision and terrane accretion masks its subsidence effect on New Guinea. Further research could explore the potential surface expression of this slab in the tectonically quiescent region of Australia.

To more confidently evaluate the role of tectonic processes in generating the observed subsidence, further research may include the development of models that consider orogenesis, elastic thickness and foreland flexure. Our analysis demonstrates that deep Earth dynamics can be coupled to paleogeographic reconstructions of New
Guinea, providing new insights into the contribution of long-wavelength mantle flow to the vertical motion of continents in areas of orogeny.

6. Acknowledgements

This research was undertaken with the assistance of resources from the National Computational Infrastructure (NCI), which is supported by the Australian Government. SZ and RDM were supported by Australian Research Council grant IH130200012 and DP130101946. NF was supported by Australian Research Council grant DE160101020. Figures were constructed using Generic Mapping Tools (Wessel and Smith, 1998; Wessel et al., 2013), GPlates (www.gplates.org) and ArcGIS.

Digital supplementary files

We provide an animation for the evolution of predicted dynamic topography from Case 4.
References


Zahirovic, S., Matthews, K.J., Flament, N., Müller, R.D., Hill, K.C., Seton, M., Gurnis, M., 2016b. Tectonic evolution and deep mantle structure of the eastern Tethys since the latest Jurassic. Earth-Science Reviews 162, 293-337.


Figure 2. a) Simplified chronostratigraphic cross-section for New Guinea from the Middle Jurassic to the present, synthesised from Norvick (2001) and van Ufford and Cloos (2005). The stratigraphic sequence extends laterally from the north to the south of New Guinea (A-D) and is sub-divided based on the geographic boundaries Bird’s Head, Irian Jaya and Papua New Guinea. This schematic highlights the Late Eocene to Oligocene regional unconformity, attributed to eustatic sea level fall and tectonic uplift resulting from the collision of the Australian and Pacific plates. Regional inundation throughout the Miocene is also apparent, with the geologic record preserving shallow marine carbonates, indicative of widespread shallow seas. b) Simplified map of New Guinea showing the A-D locations from which the stratigraphic sequence is delineated.
Figure 3. Paleo-environments of New Guinea from 160 – 0 Ma. The paleogeographies were digitised from Dow’s (1977) and Norvick’s (2003) patterns of sedimentation and attached to the plate reconstruction by Zahirovic et al (2016b). This paleogeographic reconstruction highlights the widespread inundation and sedimentary deposition throughout Miocene times, as well as the uplift and erosion environment that dominated during Oligocene times.
Figure 4: Horizontally averaged present-day a) mantle temperature and b) viscosity for Cases 1 – 4.
Figure 5. Dynamic topography and mantle temperature as predicted by Case 4. The evolution of modelled dynamic topography is presented within the tectonic framework of the region displaying subduction zones (red) and plate boundaries (brown). White stars indicate locations from which dynamic topography values were extracted throughout time (see Fig. 7). A great circle (thick black line, with white markers every 5 degrees) intersecting central New Guinea and eastern Australia has been reconstructed with the plate reconstruction of Zahirovic et al. (2016b). The evolution of mantle temperature is presented along this vertical profile from the surface to the core mantle boundary. In the Late Jurassic, tectonics in the northern Australian continental margin are dominated by the East Gondwana (EaG) active margin, with the first appearance of the EaG slab evident at ~150 Ma (a). This temperature anomaly dominates the mantle structure between New Guinea and Australia until the onset of north-dipping subduction of the Sepik (SEP) at ~70 Ma followed by south-dipping subduction of the Maramuni arc from ~23 – 15 Ma. MS – Maramuni Slab. CS – Caroline Slab.
Figure 6. a) Continental inundation of New Guinea (blue) and long-term sea level derived from Haq et al (1987) and Haq et al (2005) (respectively lightpink and green). The comparative trends between sea level and flooding history highlight the discrepancies in correlating eustasy to inundation patterns. This is particularly evident during Early Miocene times when despite long-term falling sea levels, the continent is approximately 90% flooded. Such observations suggest the presence of another process influencing the continental inundation of New Guinea. b) Modelled dynamic topography signal of Cases 1–4 from central location in New Guinea (P2 in Fig.1). The trends of dynamic subsidence and uplift indicate a link to flooding and emergence patterns where eustasy explanations are lacking. The light blue shading that denotes regional flooding is generally correlated with dynamic subsidence (decreasing dynamic topography).
Figure 7. Modelled dynamic topography from three locations across New Guinea as depicted in Figures 1 and 5 for Case 4 (Zahirovic et al 2016b). The dynamic topography signals at Irian Jaya (west), central New Guinea and Papua New Guinea (east) display the same peaks and troughs over time, yet showcase an opposite net dynamic movement, with the downward continental tilt reversing from eastward at 160 Ma to westward at present.
Figure 8. Comparison of geodynamic model predictions with seismic tomography models. The seismic tomography profiles are taken along the present-day transect depicted in a) encompassing central New Guinea and eastern Australia. We use P-wave tomographic models provided by Li et al. (2008) (b) and Obayashi et al. (2013) (c), and S-wave tomographic models provided by Ritsema et al. (2011) (d) and Montelli et al. (2006) (e). The overlying slab contours represent temperature anomalies from Cases 1–4, with the contours demarcating mantle 10% colder than ambient mantle temperature. Case 4 generates an overall better reproduction of the mantle structure, notably matching the Sepik slab (SEP) and the Caroline slab (CS) in both the P- and S-wave models. MS – Maramuni slab, NNG – Northern New Guinea slab.
Table 1. Comparison of Plate Tectonic Reconstructions

<table>
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<tr>
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<tbody>
<tr>
<td>Rifting of the northern New Guinea margin</td>
<td>Late Cretaceous times</td>
<td>Late Jurassic times (~172 Ma)</td>
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<tr>
<td>Opening of the Sepik ocean basin</td>
<td>~80 Ma</td>
<td>~157 ± 16 Ma</td>
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<tr>
<td>Subduction polarity reversal and onset of north-dipping Sepik ocean basin subduction</td>
<td>~35 to 31 Ma</td>
<td>Maastrichtian times (~71 to 66 Ma)</td>
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<tr>
<td>Sepik terrane accretion to New Guinea</td>
<td>27 to 18 Ma</td>
<td>~35 to 31 Ma</td>
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<tr>
<td>South dipping subduction to the north of New Guinea</td>
<td>15 to 5 Ma</td>
<td>23 to 15 Ma</td>
</tr>
<tr>
<td>Halmahera-Torricelli-Finisterre Arc collision</td>
<td>~6 Ma</td>
<td>~14 Ma</td>
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Table 2. Parameters common to all model cases. Subscript “0” denotes reference values.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Value</th>
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<tbody>
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<td>Rayleigh number</td>
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<td>Thermal expansion coefficient</td>
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<td>Density</td>
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<td>Gravity acceleration</td>
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<td>Temperature change</td>
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<tr>
<td>Temperature offset</td>
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<td>Background mantle temperature</td>
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<tr>
<td>Mantle thickness</td>
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<tr>
<td>Earth radius</td>
<td>$R_0$</td>
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<tr>
<td>Universal gas constant</td>
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<td>J mol$^{-1}$ K$^{-1}$</td>
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<tr>
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<td>kJ mol$^{-1}$</td>
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Supplementary Figure 1: Modelled dynamic topography from three locations across New Guinea for Case 4 (Zahirovic et al 2016b). The solid lines show the original dynamic topography signal as depicted in Fig. 7, whilst the dashed lines represent the dynamic topography signal from the lower mantle only, that is, from beneath the 660 km threshold. The figure shows that whilst the lower mantle controls much of the dynamic topography trends observed, it has a lower amplitude than that of the shallower signal.