The interplay of dynamic topography and eustasy on continental flooding in the late Paleozoic

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Abstract

Global sea level change can be inferred from sequence stratigraphic and continental flooding data. These methods reconstruct sea level from peri-cratonic and cratonic basins that are assumed to be tectonically stable and sometimes called reference districts, and from spatio-temporal correlations across basins. However, it has been understood that long-wavelength (typically hundreds of km) and low-amplitude (km) vertical displacements of the Earth's surface due to mantle flow, namely dynamic topography, can occur in the absence of crustal deformation. Dynamic topography can drive marine inundation or regional emergence of continents and must be taken into consideration for eustasy estimates. Our analysis indicates that the long-term trend in global-scale maximum flooding over the late Paleozoic generally correlates with global sea level curves. The first-order flooding history of North America correlates with some estimates of eustasy. The Paleozoic inundation of South America does not follow long-term sea level variations. The flooding lows during the Early Carboniferous and high during the Late Carboniferous are at odds with estimates of eustasy and can be explained by dynamic uplift and subsidence, respectively. Our dynamic topography models indicate that the Yangtze Platform of South China experienced significant dynamic subsidence during the transition from Permian to Triassic largely due to proto-Pacific subduction and its northward motion to collide with North China. The reference districts - Western New York, Oklahoma and Kansas, and West Texas in North America - were to some degree affected by dynamic uplift and subsidence associated with long-lived Panthalassa subduction zones, closure of the Rheic Ocean and large-scale upwelling above the African deep-mantle structure during late Paleozoic times. This indicates that some published global sea level curves may include non-eustatic signals such as dynamic uplift or subsidence. The interpretation of stratigraphic data gathered from these reference districts should be treated with caution to estimate global sea level variations.

Publication Details


This journal article is available at Research Online: https://ro.uow.edu.au/smhpapers1/732
The interplay of dynamic topography and eustasy on continental flooding in the late Paleozoic

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Abstract. Global sea level change can be inferred from sequence stratigraphic and continental flooding data. These methods reconstruct sea level from peri-cratonic and cratonic basins that are assumed to be tectonically stable and sometimes called reference districts, and from spatio-temporal correlations across basins. However, it has been understood that long-wavelength (typically hundreds of km) and low-amplitude (<2 km) vertical displacements of the Earth’s surface due to mantle flow, namely dynamic topography, can occur in the absence of crustal deformation. Dynamic topography can drive marine inundation or regional emergence of continents and must be taken into consideration for eustasy estimates. Our results indicate that the long-term trend in global-scale maximum flooding over the late Paleozoic generally correlates with global sea level curves. The first-order flooding history of North America correlates with some estimates of eustasy. The Paleozoic inundation of South America does not follow long-term sea level variations. The flooding lows during the Early Carboniferous and high during the Late Carboniferous are at odds with estimates of eustasy and can be explained by dynamic uplift and subsidence, respectively. Our dynamic topography models indicate that the Yangtze Platform of South China experienced significant dynamic subsidence during the transition from Permian to Triassic largely due to proto-Pacific subduction and its northward motion to collide with North China. The reference districts – Western New York, Oklahoma and Kansas, and West Texas in North America – were to some degree affected by dynamic uplift and subsidence associated with the long-living Panthalassa subduction zones, closure of the Rheic Ocean and African upwellings during late Paleozoic times. These indicate that some published global sea level curves may include non-eustatic signals that include dynamic uplift or subsidence. Therefore, the interpretation of stratigraphic data gathered from these regions should be treated with caution when used to estimate global sea level variations.
**Key words**: dynamic topography, eustatic sea level, continental inundation, mantle flow, plate tectonic model, mantle structure

1. Introduction

Eustatic sea level changes during the Paleozoic have been reconstructed by interpreting sequence stratigraphy (e.g. Vail et al., 1977; Hallam, 1992; Haq and Schutter, 2008; Snedden and Liu, 2010). Sequence stratigraphy is a semi-quantitative method that makes it possible to reconstruct relative variations in sea level from stratigraphic sections in pericratonic and cratonic basins. Correlations between sea level events recorded in sedimentary strata across several basins are used to build short-term eustatic sea level curves (Haq et al., 1987; Haq and Schutter, 2008). For the Cretaceous, long-term fluctuations in eustatic sea level are established considering the mean age of the oceanic crust, the production rate of oceanic lithosphere at mid-ocean ridges, episodes and duration of emplacement of seamounts and large igneous provinces on the seafloor, sediment input into the ocean, and continental flooding data (Haq, 2014). Of these indicators, only continental flooding data are available for the Paleozoic. A key assumption in the sequence stratigraphic approach to reconstruct past sea levels is that these chosen regions used as reference districts (Fig. 1) to establish the chronology of Paleozoic sea level changes are assumed to be tectonically stable. However, it has long been known that long-wavelength vertical deflections of the solid Earth’s surface as a result of mantle flow, called dynamic topography, may occur without any crustal thickening or thinning (Gurnis et al., 1990; Gurnis, 1993; Liu et al., 2008; Moucha et al., 2008; Spasojević et al., 2012; Flament et al., 2013). This issue has been recognised by stratigraphers who are aware of the effect of dynamic topography on long-term sea level change (Haq, 2014; Kominz et al., 2008) and on the long-term trends of maximum continental flooding. Recent developments in global tectonic reconstructions of the late Paleozoic (Matthews et al., 2016; Young et al., 2008) indicate that some of the regions (e.g. the Yangtze platform of South China during the Permian) used to reconstruct global sea level change could be affected by tectonic activity and mantle flow. Therefore, whether the estimated sea level change using this method can truly represent global sea level change is debated (Hallam et al., 1992; Cloetingh and Haq, 2015). Additionally, estimates from continental flooding derived from paleogeographic maps and used in reconstructions of past eustatic sea levels require parameterizing the evolution of the poorly constrained shapes of continents, which leads to large uncertainties (Bond, 1979; Harrison et al., 1981; Algeo and Seslavinsky, 1995; Flament et al., 2013); in addition, continental flooding is also affected by dynamic topography (e.g. Mitrovica et al., 1989; Gurnis 1990; Spasojevic et al., 2012; Müller et al., 2018a).
Published eustatic sea level curves commonly suggest a long-term decrease in global sea level over
the late Paleozoic (e.g. Vail et al., 1977; Hallam, 1992; Haq and Schutter, 2008). This is thought to
be associated with the late Paleozoic aggregation of the Pangea supercontinent leading to an
increase in the volume of ocean basins (Worsley et al., 1984; Vail et al., 1977; Hallam, 1992; Haq
and Schutter, 2008; Conrad, 2013; Guillaume et al., 2016). The sea level curve of Haq and Schutter
(2008) (hereafter HS08) has become the most widely-used model of the Paleozoic sea level history
and provides important constraints for studies in oceanic geochemistry composition (Munnecke et
al., 2010; Flament et al., 2013) and past climates (DiMichele et al., 2009; Munnecke et al., 2010) in
Paleozoic times. It was assembled from the interpretation of the sedimentary record in tectonically
stable basins largely from North America, Europe and China for the late Paleozoic (~400–250 Ma).
In this study, we use HS08 as our reference sea level curve in the main body text and discuss two
other global sea level curves – Hallam (1992), which we refer to as H92, and Algeo and
Seslavinsky (1995), which we refer to as AS95 – in the discussion section.

We estimated the flooding ratios (defined as area of continental shallow seas as a percentage of total
area of the continent at a certain time interval) of the North and South American continents
individually, as well as the flooding ratio for all continents combined, from a set of time-dependent
paleogeographic maps (Cao et al., 2017) in which the paleo-coastlines represent the maximum
transgression surfaces. We computed dynamic topography for the continents using two forward
mantle flow models based on time-dependent boundary conditions from two distinct tectonic
reconstructions by Matthews et al. (2016) and Young et al. (2018). We then compared the resulting
continental flooding ratios with modelled dynamic topography and published global long-term sea
level curves (H08, H92 and AS95). We mapped the mantle temperature along cross sections to
identify the origin of changes in dynamic topography. We predicted the evolution of dynamic
topography at the reference districts used to reconstruct the late Paleozoic eustatic curves (Fig. 1).
Our study highlights that combining digital paleogeographic reconstructions, geological
observations, plate reconstruction models and models of past mantle flow provides insights into
understanding mechanisms of continental inundation and distinguish global or regional sea level
change over deep time (e.g. Hallam, 1984; Gurnis, 1993; Spasojevic and Gurnis, 2012).
2. Methods

2.1 Paleogeography and tectonic reconstructions

We use a set of digital time-dependent global paleogeographic maps extending back to the late Paleozoic (Cao et al., 2017), which document the ancient distribution of ice sheets, mountains, land mass, shallow seas and deep ocean basins for the last 402 million years (Myr). These maps were built on a set of published global paleogeographic maps compiled by Golonka et al. (2006) using paleoenvironmental and paleo-lithofacies data sets and constructed independently from global sea level curves. They were tested and refined by the incorporation of paleoenvironmental data from the Paleobiology Database by Cao et al. (2017). In these maps, paleo-coastlines indicate the maximum transgression surfaces, so that the resulting continental flooding ratios represent maximum marine inundation. All the digital paleogeographic maps at present-day coordinates are available in the Supplementary materials of Cao et al. (2017). This facilitates the calculation of continental flooding ratios and also makes it possible to efficiently link the digital paleogeographic maps to alternative plate tectonic models.

We consider the global plate tectonic reconstructions of Matthews et al. (2016) (KM16) and Young et al. (2018) (AY18), both with continuously closing plate boundaries extending into the late Paleozoic (410–0 Ma). The reconstructions describe some possible past locations and motion velocities of plates and subduction zones through time. The reconstruction of KM16 is based on the reconstructions of Domeier and Torsvik (2014) for the period 410–250 Ma and of Müller et al. (2016) for the period 230–0 Ma, except that absolute plate motions are based on a different true polar wander-corrected reference frame (Torsvik et al., 2012). The reconstruction of AY18 is built on the reconstruction of KM16 but adopts different scenarios for the closure of the Rheic Ocean and motion of circum- Paleo-Tethys blocks and the paleomagnetic reference frame of Torsvik and Voo (2002) for the period 410-250 Ma. Plate speeds and trench migration rates during the Paleozoic are significantly lower in AY18 than in KM16 (Young et al., 2018). In addition, the AY18 reconstruction better reproduces the present-day structure of the lowermost mantle when used as a surface boundary constraint in numerical mantle flow models (Young et al., 2018).

We quantitatively estimated late Paleozoic flooding ratios for the North and South American continents individually, and for all continents combined, from the digital paleogeographic maps (Cao et al., 2017) reconstructed back in time using the tectonic models of KM16 and AY18, respectively. Ronov (1994) and Algeo and Seslavinsky (1995) used a similar metric to estimate the flooding history of the main continents over time. The calculated continental flooding ratios are the
same using the two reconstructions of KM16 and AY18 as expected, although the reconstructed locations of the continents are different (Fig. 2). In this study, we primarily focused on the North America and South America continents for the following reasons: (1) The sedimentary, paleoenvironmental and paleo-lithofacies records in these two continents in late Paleozoic times have been extensively studied and their paleogeographies are relatively well-constrained (e.g. Golonka et al., 2006; Limarino and Spalletti, 2006; Scotese, 2008, 2016; Cao et al., 2017); (2) The paleo-latitudinal constraints from paleomagnetic data for the two regions in the late Paleozoic are relatively reliable compared to smaller blocks such as North China, South China and Tarim; (3) The paleo-longitudinal location of North America is poorly constrained and very different between the two reconstructions considered here. Here, we investigate the effect of the paleo-longitude of North America on the dynamic topography predicted by numerical models of mantle flow for that continent; and (4) North America and South America were continuously affected by plate tectonic activity during the late Paleozoic, including circum-Panthalassa subduction zones and the closure of Rheic Ocean and Paleo-Tethys Ocean. These two continents are a natural laboratory to study the interaction between plate tectonic motions and deep Earth mantle processes.

2.2 Mantle flow model setup and dynamic topography computation

Convection models. We modeled thermochemical convection within Earth’s mantle in a 3D spherical shell with depth- and temperature-dependent viscosity and thermal expansivity, as in Hassan et al. (2016) and Flament (2019). We used the finite element code CitcomS (Zhong et al., 2000, 2008) to solve equations for the conservation of mass, momentum and energy under the extended Boussinesq approximation (Christensen and Yuen, 1985). Bower et al. (2015) modified CitcomS to incorporate the thermal structure of the lithosphere and of shallow slabs into the convection models using a progressive data assimilation method. The method of Bower et al. (2015) makes it possible to reconstruct the time-dependent mantle flow consistent with global plate motion models (e.g. Matthews et al., 2016; Young et al., 2018), and in turn to compare the predictions of the flow models to independent constraints. For example, the present-day temperature field can be compared to tomography models (e.g. Flament et al., 2017).

Global plate tectonic reconstructions with continuously closing plate boundaries (e.g. Gurnis et al., 2012; Domeier et al., 2014; Matthew et al., 2016; Young et al., 2018) are required to derive surface velocities that are imposed as time-dependent surface boundary condition in the mantle flow reconstructions (e.g. Hassan et al., 2015; Cao et al., 2018; Young et al., 2018). The resulting predictions of the location of past downwellings and upwellings and associated dynamic topography
from the flow models (e.g. Zhang et al., 2012; Flament et al., 2013) can be compared to the geological record of Phanerozoic marine inundation of continents (e.g. Gurnis, 1993; Spasojevic and Gurnis, 2012) and of the past vertical motion of continents (e.g. Flament et al, 2013; Flament et al., 2014; Shephard et al., 2014). Here we considered two model cases: Case KM16 is based on the global tectonic reconstruction of Matthews et al. (2016) and Case AY18 is based on the global tectonic reconstruction of Young et al. (2018).

Governing parameters. As in Young et al. (2018), the Rayleigh number, which controls the vigour of convection, is

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

and the dissipation number, which controls viscous dissipation, is

$$Di = \frac{\alpha_0 \rho_0 R_0}{C_{P_0}},$$

where $\alpha_0$, $\rho_0$, $g_0$, $\Delta T$, $h_M$, $\kappa_0$, $\eta_0$, $R_0$ and $C_{P_0}$ are the thermal expansivity, density, gravity acceleration, temperature change across the mantle, mantle thickness, thermal diffusivity, viscosity, Earth radius and heat capacity (see Table 1 for values). The viscosity of the mantle depends on temperature and depth following the law:

$$\eta = \eta(r)\eta_0\exp\left\{\left[\frac{E_\eta + Z_\eta (R_0 - r)}{R(T + T_{off})}\right] - \left[\frac{E_\eta + Z_\eta (R_0 - R_C)}{R(T_{CMB} + T_{off})}\right]\right\},$$

where $\eta(r)$ is a pre-factor varying with the depth in the lower mantle: 0.02 (<160 km), 0.002 (160-310 km), 0.02 (310-660 km) and 0.2 (>660 km). $R_C$, $E_\eta$, $Z_\eta$, $R$, $T$, $T_{off}$ and $T_{CMB}$ represent the radius, the radius of Earth’s core, the activation energy, activation volume, universal gas constant, dimensional temperature, temperature offset and temperature at the CMB (see Table 1 for values). The resulting horizontal-average, minimum and maximum present-day temperature and viscosity over depth are shown on Figure 3.

Model setup and initial conditions. We briefly outline the setup and initial conditions of our models, which are similar to those in Hassan et al. (2015) and Young et al. (2018). Earth’s mantle is represented by a spherical shell composed of approximately 13 million mesh nodes, so that the vertical resolution is ~15 km near the surface and ~27 km near the CMB. The horizontal resolution is ~ 50 km near the surface and ~ 28 km near the CMB. Our calculations are started from 410 Ma. The initial condition contains slabs that are inserted down to 425 km depth from the surface.
boundary at a dip of 45° and to 1,200 km depth at a dip of 90° (as in Flament et al., 2017). The initial model condition includes a basal thermal boundary layer (225 km thick), a surface thermal boundary layer of thickness derived from the age of the lithosphere via a thermal boundary layer cooling model as in Bower et al. (2015), and an adiabatic temperature profile (gradient: 0.3 K/km) between the two. The oceanic lithosphere is assumed to be 90 Myr (123 km thick) old for the period of interest (410-250 Ma), and the age of the continental lithosphere is defined as in Flament et al. (2014). A compositionally distinct layer (4.7 % denser than ambient mantle and 113 km thick, making its volume 2% of that of the mantle) is embedded within the basal thermal boundary layer. This layer is used to model the basal structure of the mantle (e.g. Flament et al., 2017). During the model run, slabs are assimilated down to a maximum of 350 km depth, below which the model is dynamic (e.g. Young et al., 2018).

**Dynamic topography computation.** We computed the dynamic topography with a free-slip boundary conditions and ignoring the buoyancy and lateral viscosity variations above 250 km depth (see Flament, 2019 for more detail). We computed the time-dependent air-loaded dynamic topography in 10 Myr intervals following: \( h = \frac{\sigma}{\rho_{sm}g_0} \), where \( \sigma \) is the total normal stress and \( \rho_{sm} \) is the density of the shallow mantle (e.g. Cao et al., 2018).

### 3. Results

#### 3.1 Continental flooding history from paleogeographic reconstructions

The global-scale trends indicate a broad decrease in flooding of continents throughout the late Paleozoic (orange line in Fig. 4a). Flooding continuously decreases from ~45% in the Devonian (402-359 Ma) to ~31% in the Middle Permian (285-269 Ma) with only a slight increase to ~35% in the late Permian-early Triassic (269-248 Ma). However, we note that this increase in flooding during the late Permian-early Triassic is not observed in the paleogeographic maps of Ronov (1994), Blakey (2003, 2008) and Walker et al. (2002) (Fig. 9 in Cao et al., 2017).

At the continental scale, the flooding ratios of North America consistently decrease from 45% in the Devonian (402-359 Ma) to ~23% in the late Permian-earliest Triassic (269-248 Ma) (green line in Fig. 4a). In contrast, South America experienced a more variable inundation history (green line in Fig. 5a) with an increase from 32% in the Middle Devonian (402-380 Ma) to 42% in the Late Devonian (380-359 Ma) and then a rapid decrease to 20% in the early Mississippian (359-338 Ma). In the late Mississippian (338-323 Ma), it continued to decrease down to ~15% to reach its lowest value during 338-323 Ma. This was followed by a rapid increase to 32% flooding in the
Pennsylvanian (323-296 Ma). In the early Permian (296-285 Ma), it recorded 30% flooding before experiencing less flooding (i.e. ~20%) during the rest of Permian and the earliest Triassic (285-248 Ma).

3.2 Continental dynamic topography

Late Paleozoic tectonics were characterized by the collision of Laurussia and Gondwana, by the closure of the Rheic Ocean and of the Paleo-Tethys Ocean, and by the formation of the central Pangean mountains (Matthews et al., 2016; Young et al., 2018). The active margins of western North America and South America were enduringly influenced by long-lived eastern circum Panthalassa subduction zones. The collision of Laurussia and Gondwana led to the closure of the Rheic Ocean and formation of central Pangean mountains (Fig. 2 b, d). As a result, North America experienced extensive negative dynamic topography especially along its western margins over the late Paleozoic, indicated by both time-dependent mantle flow model cases KM16 and AY18 (Fig. 2e-f). Similarly, South America also experienced a pronounced negative dynamic topography history at its western margins (Fig. 2g-h). These two continents were also affected by the African large-scale upwelling that dynamically supported topography in the eastern regions of North and South America (Fig. 2e-h). The calculated average dynamic topography for North America (Fig. 4b) and South America (Fig. 5b) over time shows that the overall dynamic topography was always negative in the late Paleozoic for both cases KM16 and AY18.

Specifically, the mean dynamic topography for North America in Case KM16 (Fig. 4b) (blue line in Fig. 4b) remained stable for the periods between 402-323 Ma (−150 to −100 m) and between 285–248 Ma (−190 to −150 m), but it experienced a fast decrease from −100 m to −210 m between 323–285 Ma. In Case AY18, there was no significant change between 402−323 Ma, then a dramatic increase from approximately −140 m to −50 m between 323–285 Ma. This was followed by a rapid decrease to less than −200 m between 285–248 Ma. The mean dynamic topography for South America over time shows distinct trends between Case KM16 and Case 18 (Fig. 5b). In Case KM16, it increased from approximately −80 m between 402–380 Ma to −40 m between 359–338 Ma and then rapidly decreased to −200 m between 296–285 Ma. It subsequently remained stable at about −170 m between 285–248 Ma. In contrast, Case AY18 indicates a rapid decrease in mean dynamic topography from approximately −20 m between 402–338 Ma to −150 m between 338–323
Ma, and then a persistent decrease to about −30 m in the late Permian-earliest Triassic (269−248 Ma).

The evolution of mean dynamic topography for all continents in cases KM16 and AY18 show different trends during the Devonian – middle Permian (402−269 Ma) but are more similar during the late Permian and the earliest Triassic (269−248 Ma) (Fig. 6). Specifically, Case KM16 (blue line in Fig. 6a) shows short-lived and minor dynamic subsidence of ~10 m during the middle Devonian (402−380 Ma) and then a stronger dynamic uplift from approximately −10 m at 380 Ma to 30 m at 338 Ma. It subsequently indicates long-term and significant dynamic subsidence until −50 m at 248 Ma. In Case AY18 (red line in Fig. 6a), the mean dynamic topography for all continents remained stable between approximately −10 m and 20 m during the whole time period except for late Permian-earliest Triassic period, marked by a modest increase of 20 m from 338 Ma to 319 Ma. However, in the late Permian-earliest Triassic, Case AY18 indicated that global continents experienced dramatic dynamic subsidence at a rate of 4 m/Myr. We note that while globally averaged continental dynamic topography only varies by ~80 m over the period of interest (Fig. 6a), these changes in the global average reflect larger changes at the scale of one continent (by about 150 m for North and South America, Figs 4b and 5b), and even larger regional changes (Fig. 2e, f, g, h).

4 Discussion

4.1 Mechanisms of continental flooding

4.1.1 North America

The first-order flooding history of North America correlates with long-term sea level changes in the late Paleozoic (Fig. 4a). The average dynamic topography of the continent remained relatively stable throughout most of the timeframe in both cases of KM16 and AY18 (Fig. 4b). This suggests that eustasy was the main driver of inundation of North America in the late Paleozoic. In addition, the flooding decrease during the Pennsylvanian (323−296 Ma) compared to global sea level increase (Fig. 4a) can be explained by dynamic uplift in Case AY18 (red line in Fig. 4b). The rates of dynamic topography change for North America during this time period show that this dynamic uplift occurred nearly all over the North American continent, but with a larger amplitude in the eastern regions (Fig. 7c). In contrast, Case KM16 predicts dynamic subsidence during that period (blue line in Fig. 4b) and the rates of dynamic topography change suggest extensive subsidence over
the continent with a larger amplitude in the northern and western regions (Fig. 7b); these predictions are difficult to reconcile with the paleogeographic constraints. We note that the main two regions of dynamic subsidence in both Cases KM16 and AY16 between 320-300 Ma correspond to flooded regions (Fig. 7). However, the regions undergoing dynamic subsidence are larger than the flooded regions indicated by the paleogeography in Case KM16, and smaller in Case AY18 (Fig. 7). The dynamic models are also not designed to predict isostatic topography, and therefore cannot reproduce the orogenic belt that is indicated by the paleogeography (Fig. 7a).

[Insert Figure 7]

4.1.2 South America

The inundation history of the South American continent does not follow long-term variations in global sea level in the late Paleozoic (Fig. 5a). Flooding was relatively low in the Mississippian but high in the Pennsylvanian compared to global sea level changes (Fig. 5a). The flooding low during the early Mississippian (359−338 Ma) may be attributed to dynamic uplift predicted by Case KM16 (blue line in Fig. 5b). The map of rate of dynamic topography change (Fig. 8b) shows that the dynamic uplift mostly occurred in southern South America and some regions in its center and north. However, the variations are small, mostly at a rate of less than 2 m/Myr, and hence the effects of the dynamic topography may not be enough to explain the flooding low during this period. In contrast, Case AY18 predicts dynamic subsidence during the time (red line in Fig. 5b), which does not match the paleogeographic constraints, and the rates of dynamic topography change show that widespread subsidence occurred over the continent (Fig. 8c). The lowest flooding during the late Mississippian (338−323 Ma) could have been caused by dynamic uplift during the time indicated from Case AY18 (red line in Fig. 5b), in contrast to the dynamic subsidence predicted from Case KM16 models (blue line in Fig. 5b). The maps of rate of change of dynamic topography for Case AY18 show that South America experienced broad dynamic uplift by up to 8 m/Myr during the late Mississippian, except for parts of its southwestern margin (Fig. 8f). However, Case KM16 shows dynamic subsidence in most regions of the continent, which is not consistent with the paleogeographic constraint, and slight dynamic uplift in its northeastern region (Fig. 8e).

[Insert Figure 8]

During the Early Pennsylvanian, South America was extensively flooded with a west-east inland seaway (Fig. 8g). The rise in global sea level during this time (purple line in Fig. 5a) contributed to
this change in regional flooding. Case KM16 predicts extensive dynamic subsidence over the whole continent, with magnitude up to −10 m/Myr along its western margins (Fig. 8h). Tectonically, this dynamic subsidence is linked to the double Panthalassan subduction zones along western South America in the reconstruction of KM16. These two subduction zones define the boundaries of the Patagonia Plate and isolate it from the western Farallon and Phoenix plates and the eastern South America Plate (Fig. 2h). The Patagonia Plate consistently subducted beneath South America until it completely vanished in the early Permian (Matthews et al., 2016). The mantle temperature along a cross section through the western margin of South America during the Early Pennsylvanian (section S2 in Fig. 2f) shows the two slabs sinking beneath the western margins of South America, leading to significant dynamic subsidence along the margin (Fig. 5c). This dynamic subsidence could have contributed to the increase in flooding over time. In contrast, Case AY18 predicts broad but small-amplitude dynamic uplift caused by large-scale upwelling above the African basal thermochemical pile during that period (Fig. 8i), instead of dynamic subsidence resulting from the subducting slabs at the western margin of the continent. This dynamic uplift for Case AY18 is not consistent with the paleogeographic constraint.

4.1.3 Global continental flooding

Conrad and Husson (2009) discussed the connection between continental dynamic topography and global sea level. They showed that a net deflection of continental areas by mantle flow will be balanced by an opposing net deflection of oceanic areas as the total of the dynamic support of the globe integrates to zero, which leads to a net offset of sea level. Our numerical models indicate that the dynamic history of global continents in Case KM16 appears to be comparable to global sea level changes for the late Paleozoic except for the late Permian-early Triassic (blue line in Fig. 6b). This scenario suggests that the global sea level curves may contain a dynamic topography signal that is not accounted for. In contrast, mantle flow in Case AY18 predicts stable dynamic history of continents except for the late Permian-early Triassic (red line in Fig. 6b), suggesting that there may not be much impact from deep Earth’s processes on global sea level changes. The variations in modelled dynamic topography between Case KM16 and Case AY18 are associated with adopting different regional tectonic reconstructions for the closure of the Rheic Ocean and circum- Paleo-Tethys blocks, and difference of the global RMS speed which is faster in the KM16 model than that in AY18 model typically for the period before ~250 Ma (Young et al., 2018). Gurnis et al. (1993) suggested that plate speeds could be an important factor controlling variations in rates of subduction and continental inundation during the Phanerozoic, with greater plate velocities leading to greater rates of subduction, and to greater subsidence and inundation at convergent margins. As an
example, the rate of dynamic topography change for North America between 320-300 Ma is much larger in KM16 than in AY18, and the faster motion of North America at an average speed of 14.2 cm/yr in KM16 than at an average speed of 9.5 cm/yr in AY18 between 370-320 Ma results in distinct trench migration and in a distinctly different mantle structure at 310 Ma (Fig. 4c and 4d).

During the late Permian – early Triassic, the dynamic topography of all continents combined indicates notable dynamic subsidence for both cases KM16 and AY18 (Fig. 6). Our time-dependent global dynamic topography models show that Asia and western North and South America underwent the most dynamic subsidence during this time period (Fig. 9). A eustatic low has been widely recognized in this period (Vail et al., 1977; Hallam, 1992; Algeo and Seslavinsky, 1995; Haq and Schutter, 2008). Guillaume et al. (2016) reviewed global sea level changes during this time and argued that the eustatic low in the late Permian - early Triassic was difficult to explain by tectonics and climatic factors and might be due to global dynamic topography. Our models in both cases predict a considerable dynamic subsidence of the continents which would have elevated relative sea levels. Therefore, our results do not indicate that the fall in global sea level during the late Permian was related to global dynamic topography. South China is the reference district during this eustatic low in the curve of HS08, and we note that both cases KM16 and AY18 suggest that South China underwent significant dynamic subsidence during the late Permian (Fig. 9d, 10e), despite the markedly different paleo-location of South China in these two reconstructions. This dynamic subsidence was due to the subduction of the Izanagi Plate (Fig. 9) and the northward motion of South China to collide with North China. The predicted dynamic subsidence of South China highlights the importance to examine whether the regions that are used to reconstruct global sea level changes are affected by mantle flow. We note that cases KM16 and AY18 do not predict a mantle plume under South China at ~260 Ma (Emeishan LIP), but that a mantle upwelling could explain the apparent sea level low for South China during the late Permian.

[Insert Figures 9-10]

4.2 Dynamic topography evolution of reference districts used to interpret eustatic sea level change

We investigate the history of dynamic topography for each of the reference districts used to establish the late Paleozoic chronology of global sea level change (Haq and Schutter, 2008; Fig. 10) for both cases KM16 and AY18. We quantified the absolute value of the maximum rate of dynamic topography change for each reference district for the duration when it is used as a reference district.
For instance, for the Western New York reference district, the time interval with the maximum change in dynamic topography in its duration as a reference district is 370–360 Ma (Fig. 11), and the maximum rate of dynamic topography change for this reference district is 7.7 m/Myr (Fig. 11). The results show that rates of dynamic topography change for the reference district in Britain are ~3 m/Myr in the KM16 case and 0.5 m/Myr in the AY18 case during the Mississippian. In contrast, South China underwent pronounced dynamic subsidence at a maximum rate of dynamic topography change of more than 14 m/Myr in the both cases during the Late Permian and the earliest Triassic.

Maximum rates of dynamic topography change for Western New York were 8 m/Myr in Case KM16 and 4 m/Myr in Case AY18, respectively, during the Late Early – Late Devonian. Oklahoma and Kansas are different for the two cases: change by 8 m/Myr in Case KM16 but by 1 m/Myr only in Case AY18 during the Pennsylvanian. West Texas shows similar results, with a rate of 7 m/Myr in Case KM16 and 6 m/Myr in Case AY18 during the Permian.

Overall, the reference district in Britain was the least affected by dynamic topography in cases KM16 and AY18, both with a maximum rate of dynamic topography change of less than 3 m/Myr during the Mississippian, indicating that it can be used as a reliable reference district for that period. Oklahoma and Kansas were also a dynamically stable reference district but only in Case AY18. South China, the reference district for the period between the Late Permian – earliest Triassic, was significantly affected by dynamic topography during that period, and as a consequence its stratigraphy for the period may reflect relative sea level change rather than global sea level change. In this context, we note that an episode of flat slab subduction between ~250-190 Ma has been proposed to explain the geological evolution of South China during that period (Li and Li, 2007).

Western New York and West Texas were also affected by dynamic topography, although to a lesser extent, for the time periods for which they are used as reference district. According to the mantle flow models, the reference districts used to reconstruct eustatic curves which are most affected by dynamic topography are those in North America and South China.

4.3 Influence of tectonic reconstruction choice on modelled dynamic topography

Our dynamic topography models strongly depend on the input tectonic reconstructions. KM16 and AY18 adopted different Paleozoic reference frames, and different scenarios for the evolution of the Panthalassa oceanic plates, the closure of the Rheic Ocean and the motion of circum- Paleo-Tethys tectonic blocks. This results in different mantle flow and dynamic topography for the late Paleozoic
Because dynamic topography averages to zero globally by definition (e.g. Conrad and Husson, 2009; Flament, 2019), regional differences in subduction-driven dynamic topography must be compensated somewhere else, leading to global differences.

The dynamic topography predicted for Case KM16 better explains some of the discrepancies between flooding and global sea level changes for North America (Fig. 4). For instance, during the Pennsylvanian, Case KM16 indicates an extensive dynamic subsidence in North America (Figs. 4, 7). The mantle temperature structure along a cross section through the western margin of North America (cross section S1 in Fig. 2) shows the Farallon slab sinking beneath the Side Mountain Ocean which is close to the western margins of North America (Fig. 4c), and the cold and dense feature driving downwelling that pulls western North America down (Fig. 4c). However, in Case AY18, the Farallon slab does not considerably result in dynamic subsidence of North America due to rapid landward trench migration (Fig. 4d). Indeed, the predicted dynamic uplift for Case AY18 during this time period (Fig. 4f) is associated with the upwelling African plume (Fig. 2f).

The dynamic topography predicted for Case AY18 better explains the continental flooding of South America than the dynamic topography predicted for Case KM16 (Fig. 5). In the plate tectonic model of AY18, there is only one subduction zone along the western margin of South America (Fig. 2h). The associated subducted slab does not significantly change the dynamic topography at the margins (Figs. 2, 5, 8). Instead, South America is extensively uplifted by the large-scale upwelling above the African basal thermochemical pile (Figs 2, 5, 8). In summary, our results suggest that neither of the considered model cases can be preferred and that different models will be required to simultaneously reconcile the Paleozoic paleogeography of both North and South American continents.

### 4.4 Comparison with other published Paleozoic eustatic curves

We compared our global flooding history and average dynamic topography of global continents with the alternative two global sea level curves, including H92 and AS95 (Fig. 12). The global sea level curve of H92 was reconstructed based on sequence stratigraphic data largely from North America and European regions. Although it suggests a first-order agreement on the late Paleozoic global sea level change with HS08, it indicates a rise in global sea level during the Mississippian and early Pennsylvanian and a fall during the rest of Pennsylvanian, contrary to what HS08 predict for the same intervals. The curve of AS95 was constructed using continental flooding and suggests less fluctuation in global sea levels – two sea level increases during the Devonian and Early and
Middle Permian and two falls in Early Carboniferous and Late Permian (black line in Fig. 12). All variations are relatively small but have large uncertainty of ~100 m (grey-shaded area in Fig. 12). The comparisons between these two global sea level curves and global average dynamic topography of continents over the late Paleozoic indicate that H92 has similar trends with the evolution of global average dynamic topography of continents only in the Middle Devonian (402–380 Ma) for Case KM16, and in some times during the Late Carboniferous and Early Permian (323–269 Ma) for Case AY18. For AS95, only in the Late Carboniferous and Early and Middle Permian (323–268 Ma) for Case KM16, and only in a period during the Carboniferous (338–320 Ma) for Case AY18. Overall, compared with HS08, H92 and AS95 indicate less similarity to the evolution of global average dynamic topography of the continents over time.

However, these published Paleozoic eustatic curves (e.g. Vail et al., 1977; Hallam, 1992; Algeo and Seslavinsky, 1995; Haq and Schutter, 2008) all have uncertainties to some degree due to the lack of geological constraints in deep geological times. It is now widely understood that estimating past global sea level change cannot be achieved by using the rock record alone (Miller et al., 2005; Haq et al., 1987; Haq and Schutter, 2008; Haq and Al-Qahtani, 2005), and time-dependent models of flow of the Earth’s mantle are crucial to establish the reference frame of past global sea level change (Flament et al., 2013; Moucha et al., 2008; Spasojević and Gurnis, 2012; Müller et al., 2008, Haq, 2014; Kominz et al., 2008). There is no systematic understanding of how mantle processes have influenced sea level globally over long geological timescales (beyond 100 million years ago) as yet. Our results illustrate that mantle flow models are also uncertain.

### 4.5 Comparison to previous geodynamic models

Gurnis (1993) and Spasojevic and Gurnis (2012) produced dynamic topography models to study the Phanerozoic marine inundation of continents and to study global sea level change and vertical motions of continents since the Late Cretaceous, respectively. These studies presented a series of instantaneous flow models, in contrast to the time-evolving mantle flow models presented here. Although Zhang et al. (2012) used time-dependent models to investigate changes in global dynamic topography since the late Paleozoic, they focused on the vertical motions of the Slave and Kaapvaal Cratons when comparing their results to geological data. The dynamic models of Gurnis (1993) were global-scale and extended further back in time than those in this study, yet the approach of geodynamic modelling in that study solved for buoyancy flux only, with no temperature variations.
and no lateral viscosity variations. The geodynamic models of Spasojevic and Gurnis (2012) were also global, but limited to the last 100 Myr because they combined forward models with backward advection models based on mantle tomography; the latter of which have limited predictive power further back in geological time (Conrad and Gurnis, 2003). The dynamic models of Müller et al. (2018a) were global and time-dependent but limited to the last 140 Myr. Gurnis (1993) and Müller et al. (2018a) carried out a global and quantitative analysis considering continental flooding, eustatic sea level and dynamic topography. Spasojevic and Gurnis (2012) developed models to predict continental flooding globally, and identified which of dynamic topography or eustasy controlled the continental flooding of North America, Eurasia and Australia over the last 100 Myr. Here, we discuss the effects of eustasy and dynamic topography for North America and South America between 400-250 Ma.

4.6 Limitations, uncertainties and mismatches between geologic observations and dynamic topography models, and future work

4.6.1 Estimates of continental flooding

The estimates of flooded continental areas may vary when using different paleogeographic maps. To check whether other published paleogeographic maps produce different flooding histories, we calculated the flooding ratios of South America continent from the static paleogeographic maps of Scotese (2008, 2016) for the late Paleozoic using the same method as we did for the maps of Cao et al. (2017). The resulting flooding ratios indicated a persistent decrease of flooding from ~60% during the Early Devonian to ~13% during the late Permian, suggesting a first-order agreement with global sea level changes. The flooding ratios were always higher than our results except for the late Permian. However, it is unclear whether the paleogeographic maps of Scotese (2008, 2016) are independent from global sea level curves. The detailed regional paleo-environmental maps of South America built based on independent sedimentary records in the late Paleozoic (Limarino and Spalletti, 2006) indicate more similar flooding history to that recorded in the paleogeographic maps adopted here (Cao et al., 2017; Golonka et al., 2006).

The steady decrease in the North American flooding ratio over time (Fig. 4a) differs from the episodicity over tens of million years documented by Sloss (1963) in the deposition of sedimentary strata in North America. In this respect, we note that the dynamic uplift of North America predicted by Case AY18 between ~310-290 Ma (Fig. 4b) is consistent with the deposition hiatus during Pennsylvanian and early Permian times (Sloss, 1963).
We used the maps of Golonka et al. (2006) as in Cao et al. (2017) because: (1) they are independent from eustatic sea level curves, (2) they are the only set of available digital global time-dependent paleogeographic maps covering the Paleozoic, (3) they have been tested and updated with the incorporation of new paleoenvironmental data sets (Cao et al., 2017) and hence represent the well-constrained late Paleozoic global paleogeography reconstructions. Incorporating with other geological data, such as stratigraphic data, paleoenvironment and paleo-lithofacies data to further constrain the paleogeographic reconstructions is needed (e.g. Wright et al., 2013; Cao et al., 2017). However, applying this approach at a global-scale and deep in geological times is a significant task.

4.6.2 Uncertainty in mantle flow models

Uncertainties in our numerical mantle flow models primarily stem from the following sources: the initial conditions of the model, the assimilated tectonic reconstructions and the rheology of the mantle. Given that Flament (2019) tested and discussed the sensitivities of the mantle flow models to the initial conditions, including boundary conditions and governing parameters, we herein briefly discuss this part. The amplitude of dynamic topography predicated by our global mantle flow models strongly depends on model setup and governing parameters (Flament et al., 2013; Flament, 2019), although it is demonstrated that paleogeographically-constrained mantle flow models compare well to time-dependent surface geological constraints (e.g. Flament et al., 2014, 2015). The predicted present-day mantle temperature can be compared to tomography models in order to test the predictive power of the forward mantle flow models (Zhong and Rudolph, 2015; Flament et al., 2017; Flament, 2019). However, the present-day structure of the mantle is not very sensitive to plate motions before ~250–275 Myr ago because most subducted slabs likely sink down to core-mantle boundary in less than ~250–275 Myr (van der Meer et al., 2010; Butterworth et al., 2014). The presented geodynamic models assume a default fixed subduction angle (45°) for all the time along all trenches, unless otherwise defined by the user (Flament et al., 2015). However, it has long been recognised that changes in slab dip angle and flat slab subduction could lead to widespread upper plate subsidence (e.g. Mitrovica et al.; Hu et al., 2018). While flat subduction scenarios have been explored with the approach used herein for South America since Miocene times (Flament et al., 2015), the effect of variable slab angles on continental flooding remains to be explored for other times and regions. The adopted plate tectonic reconstructions as boundary conditions for mantle flow models are another source of uncertainty because the global plate tectonic reconstructions determine the location of past subduction systems and plate motion history, however, their paleogeographic coordinates especially paleo-longitudes are poorly constrained. While lateral
viscosity variations are considered in the models, they are limited to three orders of magnitude, which is less than variations over up to ten orders of magnitude suggested by laboratory experiments (e.g. Karato and Wu, 1993). Achieving large viscosity contrasts in global time-dependent models of mantle flow remains a numerical challenge (e.g. Stadler et al., 2010).

5 Conclusions

The study combines geological observations, plate tectonic models and reconstructions of past mantle flow to provide insights into understanding the mechanisms of continental flooding and interactions between surface and deep Earth processes over geological time. We estimated the late Paleozoic marine inundation history of North America and South America individually, and all continents combined, from flexible time-varying paleogeographic reconstructions that are independent of eustatic sea level curves. We extracted dynamic topography for the continents from forward mantle flow models assimilating two distinct tectonic reconstructions as time-dependent boundary conditions. We compared the resulting continental flooding ratios with modelled dynamic topography and several published global long-term sea level curves. Our results indicate that the trend in global-scale flooding over the late Paleozoic generally correlates with global sea level curves. The first-order flooding history of North America correlates with global long-term sea level changes and dynamic topography can explain the second-order flooding low during the Pennsylvanian. The inundation history of South America does not follow long-term variations in global sea level. The flooding low in the Carboniferous and high in the Early Permian compared to global sea level changes can be accounted for by the dynamic uplift and subsidence predicted by mantle flow models during these times. Our global mantle flow models suggest that some estimates of global sea level changes might be affected by dynamic topography, while the eustatic low in the late Permian can be explained by dynamic topography. The reference district used to reconstruct late Paleozoic eustatic sea level changes which are most affected by dynamic topography is the Yangtze platform of South China for late Permian–earliest Triassic times. Some reference districts in North America are to some degree affected by mantle flow associated with the long-lived Panthalassa subduction zones, closure of Rheic Ocean and African upwellings. The interpretation of stratigraphic data gathered from these regions should be treated with caution when used to estimate global sea level variations.

Acknowledgments
W. Cao was supported by a University of Sydney International Scholarship (USydIS). N.F was supported by ARC grant DE160101020. R.D.M, S.Z. and S.W. were supported by ARC grant IH130200012. R.D.M and S.Z. were also supported by Alfred P Sloan Foundation grants G-2017-9997 and G-2018-11296 through the Deep Carbon Observatory. Figures were constructed using Generic Mapping Tools (Wessel and Smith, 1998; Wessel et al., 2013), GPlates (www.gplates.org) (Müller et al., 2018b) and Python 2.0. The HEALPix method (Górski et al. 2005) was used for calculation of areas of paleogeography. Mantle flow models and dynamic topography computations were carried out using resources from the National Computational Infrastructure (NCI), which is supported by the Australian Government.

Information about the supplement

We provide all the grids of the global continental dynamic topography models in both mantle reference frame and plate reference frame for cases KM16 and AY18.

The Supplement related to this article can be found at http://portal.gplates.org/portal/dt/.

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Table 1. Model parameters. Reference values are indicated by the subscript “0” indicates reference values. CMB: core-mantle boundary.

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<th>Parameter</th>
<th>Symbol</th>
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<tr>
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**Figure 1.** Cartoon illustrating the mechanisms of dynamic change – uplift and subsidence – under different scenarios. Not to scale. RD: reference districts that are used to reconstruct global sea level change. Solid lines and rectangles denote the vertical position of continents and reference districts at time $t_0$, and dashed lines and rectangles denote their vertical position at a later time $t_1$. One continent has subsided because of mantle flow, whereas the other one has undergone dynamic uplift. Global sea level may also vary between $t_0$ and $t_1$. 
Figure 2. (a-d) Paleogeography of North America (NAM) and South America (SAM) between 359-338 Ma (reconstructing age: 348 Ma) and 323-296 Ma (reconstructing age: 302 Ma) derived from Cao et al. (2017), reconstructed using two plate tectonic models of Matthews et al. (2016) (KM16) and Young et al. (2018) (AY18), respectively. (e-h) Dynamic topography of North America and South America predicted by mantle flow models using the tectonic reconstructions of KM16 and AY18 as time-dependent boundary conditions, respectively. Cross section 1 (S1) on panels f and h is anchored to North America and cross section 2 (S2) is anchored to South America. The four coloured stars on panels a-h indicate the reconstructed locations of the reference districts used to reconstruct the global sea level curve of Haq and Schutter (2008) (purple star: Western New York, red star: Britain, green star: Oklahoma & Kansas, blue star: West Texas). Black dotted lines on all panels indicate subduction zones and other black lines denote mid-ocean ridges and transform faults. Grey outlines delineate reconstructed present-day coastlines and terranes. Mollweide projection with 60°E central meridian.
Figure 3. Present-day temperature (a) and viscosity (b) as a function of depth for Case AY18. The solid line is the horizontal average, and the dashed lines are the minimum and maximum. ‘SC06’: Steinberger and Calderwood (2006).
**Figure 4.** (a) Late Paleozoic continental flooding ratios of the North America continent and all continents combined derived from the paleogeography of Cao et al. (2017), and the global sea level curve of Haq and Schutter (2008). The dashed line represents the global sea level highstands for each time interval. (b) Continental emergent ratios of the North America continent derived from the paleogeography of Cao et al. (2017), and the average dynamic topography of the continent predicted by mantle flow simulations using two tectonic reconstructions of KM16 (blue line) and AY18 (red line) as time-dependent boundary conditions. (c-d) Mantle temperature and dynamic topography sections along section 1 (anchored to North America, location shown in Fig. 2) for cases KM16 and AY18. The blue and red lines above each section are air-loaded dynamic topography (the grey line shows mean dynamic topography, which is by definition equal to zero). The numbers above the color scale denote non-dimensional temperatures and the number below the color scale denote dimensional temperatures. The solid black lines at 660 and 350 km depth denote the upper-lower mantle boundary and the depth above which buoyancy is ignored in the computation of dynamic topography, respectively.
Figure 5. (a) Late Paleozoic continental flooding ratios of the South America continent and all continents combined derived from the paleogeography of Cao et al. (2017), and global sea level curve of Haq and Schutter (2008). The dashed line represents the global sea level highstands for each time interval. (b) Continental emergent ratios of the South America continent derived from the paleogeography of Cao et al. (2017), and the average dynamic topography of the continent predicted by mantle flow simulations using two tectonic reconstructions of KM16 (blue line) and AY18 (red line) as time-dependent boundary conditions. (c-d) Mantle temperature and dynamic topography sections along section 1 (anchored to North America, location shown in Fig. 2) in cases KM16 and AY18. The blue and red lines above each section are air-loaded dynamic topography (black lines show mean dynamic topography, which is by definition equal to zero). The numbers above the color scale denote non-dimensional temperatures and the number below the color scale denote dimensional temperatures. The solid black lines at 660 km and 350 km depth denote the upper-lower mantle boundary and the depth above which buoyancy is ignored in the computation of dynamic topography, respectively.
Figure 6. (a) Comparison between global-scale flooding ratios derived from the paleogeography of Cao et al. (2017) and average dynamic topography for all continents for cases KM16 (blue line) and AY18 (red line) through time. (b) Comparison between the global eustatic curve of Haq and Schutter (2008) (purple lines) and the average dynamic topography for all continents for cases of KM16 (blue line) and AY18 (red line).
Figure 7. (a) North American paleogeography between 323-296 Ma derived from the paleogeographic maps of Cao et al. (2017). (b-c) Rates of dynamic topography change for North America between 320-300 Ma in cases KM16 (b) and AY18 (c). All maps are in the plate frame of reference. Grey outlines indicate present-day coastlines and terranes.
Figure 8. Paleogeography of South America between 359–338 Ma (a), 323–295 Ma (d) and 295–285 Ma (g) derived from the paleogeographic maps of Cao et al. (2017) and rates of dynamic topography change for South America in cases KM16 (b, e, h) and AY18 (c, f, i). All maps are in the plate frame of reference. Grey outlines indicate present-day coastlines and terranes.
Figure 9. Global continental dynamic topography at 259 Ma and 249 Ma both in the mantle frame of reference for cases KM16 (a, c) and AY18 (b, d). The red boxes highlight the locations of South China. The five coloured stars on panels e-h indicate the reconstructed locations of the reference districts used to estimate global sea level (Haq and Schutter, 2008; purple star: Western New York, red star: Britain, green star: Oklahoma and Kansas, blue star: West Texas; orange star: South China). Black dotted lines on all panels indicate subduction zones and other black lines denote mid-ocean ridges and transforms. Grey outlines delineate reconstructed present-day coastlines and terranes. Mollweide projection with 0°E central meridian.
Figure 10. Late Paleozoic dynamic topography history of reference districts − New York (a), Britain (b), Oklahoma & Kansas (c), West Texas (d) and South China (e), used to build global sea level curve of Haq and Schutter, (2008), in the KM16 (blue lines) and AY18 (red lines) cases. The shaded boxes represent the periods during which the reference districts are used to generate the global sea level curve.
Figure 11. Maximum rates of dynamic topography change for the reference districts used to reconstruct the global sea level curve of Haq and Schutter, (2008) in cases KM16 (blue) and AY18 (red), respectively. The definition of the maximum rate of dynamic topography change is described in the text.
Figure 12. (a, b) Comparison between global flooding ratios derived from the paleogeography of Cao et al. (2017) and global sea level curves of Hallam (1992, light blue lines – H92), Algeo and Seslavinsky (1995, black lines and grey-shaded range representing uncertainty – AS95). (c, d) Comparison between the average dynamic topography for all continents for cases of KM16 (blue line) and AY18 (red line) and global eustatic curves of H92 (light blue lines) and AS95 (black line and uncertainty range).