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Global tectonic and climatic control of mean elevation of continents, and Phanerozoic sea level change

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Abstract

Mean elevation of all continents, as well as mean elevation of individual continents, should be at the fundamental level controlled by the global tectonic and climatic systems. I propose a first-order model considering the interplay of the two factors in controlling mean elevation of continents. The model is able to account for the positive correlation between the present-day mean elevation and area of individual continents (except for Antarctica). Furthermore, it can also explain the low sea level during the times of supercontinents. Finally the model is used to evaluate the variation of continental crust thickness with time.

Keywords: continental crust thickness; continental height; eustasy; freeboard; hypsometry; ocean depth; sea level; coupling between global tectonic and climatic systems

1. Introduction

At present, total land area covers 29.2% of Earth's surface. Total continental area (including continental shelves and slopes) covers about 40% of Earth's surface [1–3]. Estimates of average thickness of continental crust range from 38 to 41 km [4,5]. At the level of individual continents, mean elevation increases with the area of a continent except for ice-covered Antarctica [1–3]. These first order observations have not been addressed before. What controls the mean elevation and thickness of all continents? What controls the mean elevation and thickness of each individual continent? Has mean thickness of continental crust stayed more or less the same over the Phanerozoic, or has it varied significantly with time? For example, is it possible that the mean continental thickness has doubled or halved (and total continental area halved or doubled) in the past? If not, what is keeping the continent–ocean system in balance?

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Changes in continental area or thickness would lead to changes in ocean area, ocean depth, and sea level [6]. There are two prominent sea level lows in the relative sea level curve, corresponding to the times of supercontinents, Rodinia and Pangea [7,8]. Worsley et al. [9] explained the mean elevation of individual continents by the effect of thermal isolation of continents (but without quantification), and developed a model for sea level changes. Gurnis [10] used a mantle flow model to calculate the effect of dynamic topography on marine inundation of continents as a function of age. His model matches the geographical pattern of flooding, but predicts a high fractional inundation of continents when actual data shows low fractional inundation during the times of Pangea.

I propose an alternative model for the controlling of mean continental elevation by coupling between global tectonic and climatic systems. With this model, it is possible for the first time to explain the relationship between the present mean continental elevation and continental area, as well as low sea levels during the times of supercontinents. Nevertheless, the goals of this semi-mechanistic and quantitative model are limited. (i) Only the mean elevation and thickness of individual continents are considered. The full hypsometry (frequency distribution between area and elevation) is beyond the scope of this work; McElroy and Wilkinson [11] have investigated the shape of the continental hypsometry curve. (ii) Only Phanerozoic sea level variation is considered. The more distant past (Precambrian) is beyond the scope of this work. Other authors [6,12,13] have modeled the relationship between and evolution of freeboard (mean continental elevation relative to sea level), continental crust thickness, volume of continental crust [14,15], volume of ocean water [16], oceanic crust thickness [17], and tectonic uplift rate [12].

2. A first-order model

Many factors affect the elevation and thickness of a continent. These factors can be broadly classified into two categories: internal (tectonic) and surficial (mostly climatic). Tectonic processes affecting elevation and thickness of a continent include: continental collision (such as India–Eurasia collision), subduction of oceanic lithosphere (such as Nazca Plate subducting under South American Plate), rifting (such as East African Rift), subsidence due to cooling (such as subsidence of continental margins), tectonic age (elevation of old cratons is lower than young tectonic belts), delamination [17], and dynamic topography [10,18]. Surficial processes include erosion and sedimentation, which are related to climatic system and tend to bring continental surface to sea level. Although some tectonic processes such as extension and subsidence can lower the elevation of a continent, the net effect of tectonic forces must be to uplift continents. If there were no net uplift, there would be no reason for the mean elevation of continents to be above sea level because erosion in billions of years would bring continents to sea level. To rigorously model continental elevation would require quantification of all processes and evaluation of feedbacks among them. Unfortunately, such quantification and evaluation are not available yet. For example, the sophisticated quantification of dynamic topography [10] leaves major features such as the low sea level during Pangea on the sea level curve unexplained. I propose a simple model as a first-order approximation, in which the coupling between tectonic and climatic systems controls mean elevation of continents (and hence mean thickness of continents assuming isostacy). The simple model serves to elucidate the interplay between tectonic and climatic systems, although it does not predict the full sea level curve and the inundation pattern. The assumptions, approximations and developments are summarized below:

(a) It is assumed that the net effect of tectonic processes (including collision, subduction, extension, and dynamic topography) is to uplift continents, and surficial processes are mainly erosional. Hence only two processes are considered: net uplift and erosion [19].

(b) The individuality of continents is ignored. That is, a continent is assumed to be large enough so that different geologic history, landforms, erosional behavior, tectonic age and units, and thermal regimes are averaged sufficiently so that all continents behave similarly (it will be seen that Antarctica is an exception). Continent-specific behavior is ignored. In other words, continent-specific behavior is treated as scatter. Simple calculations show that a single feature, even if it is striking, does not change the mean height of a continent in a major way. For example, the imposing

Tibet Plateau with a mean height of 4.0 km and 2 million km^2 only adds 0.12 km to the mean height of Eurasia (54 million km^2), and low-lying Europe only reduces the mean height of Eurasia by 0.11 km.

(c) Isostatic equilibrium is assumed so that the mean continental elevation (H) is related to the crustal thickness (h) through the following [20]:

$$h = h_0 + \beta H + \alpha \Delta z, \quad (1)$$

where β is the isostatic factor and equals $\rho_{\text{mantle}}/(\rho_{\text{mantle}} - \rho_{\text{cc}}) \approx 7$ in which ρ is density and subscript "cc" means continental crust, h_0 is the mean continental crust thickness for $H=0$, Δz is the change in ocean depth (or absolute sea level change), and $\alpha = (\rho_{\text{mantle}} - \rho_{\text{water}})/(\rho_{\text{mantle}} - \rho_{\text{cc}}) \approx 4.8$. The unit of h, H, and Δz is km. When considering the present-day mean crustal thickness of different continents, Δz is zero because there is a common sea level. However, when considering the variation of mean crustal thickness in the past, Δz may be non-zero. In the presence of ice cover, H in Eq. (1) should be the equivalent rock topography [21], i.e., a 0.33-km layer of rock should replace a 1-km layer of ice.

(d) It is assumed that all continents have a similar shape (again ignoring the individuality of continents) so that

$$L = \gamma A^{1/2}, \quad (2)$$

where L and A are the perimeter and area of the continent, and γ is a dimensionless proportionality factor. Montgomery and Dietrich [22] showed that basin length and drainage area are related by the above equation.

(e) Erosion is assumed to behave similarly to mass transport ([23–26], references therein), in which both "diffusion" and "convection" (mass flow) play a role. For example, hill slope mass wasting may be modeled as diffusion, and river transport may be viewed as mass flow. In order to affect the mean height, erosion must bring mass to the oceans. Hence, the mass flux ($\mathbf{J}$) per unit length of continent circumference is

$$\mathbf{J} = -D(\nabla H)_{\text{at boundary}} \approx DH/\delta, \quad (3)$$

where D is the erosion coefficient, $(\nabla H)_{\text{at boundary}}$ is the elevation gradient at the continental boundary, which is approximated by H divided by the "boundary-layer width" δ. Near the continental boundary, the elevation gradient is small and non-linear diffusion [25] can be ignored. The concept of boundary-layer width is also borrowed from mass transport theory, in which (i) a steady-state boundary layer may form when both diffusion and convection are present, and (ii) mass transport across the boundary layer is through diffusion. The parameter δ is assumed to be the same for all continents. The land volume above sea level (HA) would thus change with time as:

$$[\text{d}(HA)/\text{d}t]_{\text{erosion}} \approx -\mathbf{J}L/\beta = -LDH/(\beta\delta), \quad (4)$$

where t is time. The above relation that the erosion rate is proportional to mean elevation has been proposed by Harrison [19]. The isostasy factor is included because removal of mass equivalent to a continental height of ΔH would only decrease continental height by $\Delta H/\beta$ due to isostatic rebound. The value of L depends on the length scale of measurement. Because the transition from continental to oceanic crust involving a change of about 30 km in crustal thickness usually occurs over 100 km horizontal distance [27], L should be measured at a scale of hundreds of km. Combining Eqs. (2) and (4) leads to:

$$[\text{d}(HA)/\text{d}t]_{\text{erosion}} = -\gamma DHA^{1/2}/(\beta\delta). \quad (5)$$

The above equation for volume erosion rate is similar to that for volume erosion rate by rivers. Whipple and Tucker [28] showed that sediment volume transport rate of a river is proportional to $A^m S^n$ where A is the drainage area, S is the slope, n is between 2/3 to 5/3, and $m \approx n/2$. Taking the mean value of n to be about 1, volume erosion rate by rivers would be proportional to $A^{1/2}S$, equivalent to Eq. (5) in which H/δ is proportional to S. That is, Eq. (5) captures the mass erosion rate in a broad sense. Hence, the detailed mechanisms of erosion are not critical to this first-order model.

Eq. (5) is for volume (or mass) erosion rate $\text{d}(HA)/\text{d}t$, from which the height reduction rate ($\text{d}H/\text{d}t$) can be derived. Although A (the area of a continent) depends on h and hence on H, the relative variation of A is much smaller than that of H. Hence, for simplicity, $\text{d}A/\text{d}t$ is assumed to be zero. By moving A out of the differential, and then dividing both sides by A, Eq. (5) is simplified to:

$$(\text{d}H/\text{d})_{\text{erosion}} \approx -D\gamma H/\left(\beta A^{1/2}\delta\right). \quad (6)$$

According to the above equation, the height reduction rate of a continent is roughly inversely proportional to the square root of its area A. That is, larger continents are more difficult to erode. One way to rationalize it is as follows. The larger the continental area, the smaller the L/A ratio, meaning less boundary per unit area to bring mass to the oceans. Another way to rationalize it is as follows. The larger the continental area, the more difficult it is to erode the interior of a continent by rivers because the interior is farther away from the oceans, leading to smaller mean fluvial gradient.

In Eq. (6), $\beta A^{1/2}\delta/(D\gamma)$ can be viewed as the characteristic time scale for the mean height to decrease to 1/e of the initial mean height. Although the parameter D depends on mean annual temperature, precipitation, ice cover, and hence should vary from one continent to another, the variation is difficult to quantify and is ignored in this paper.

(f) To model net uplift by tectonic processes, it is assumed that: (i) averaged over a continental scale, any unit area has the same probability to be uplifted. Because tectonic processes do vary considerably, only large masses such as continents may be approximated by this assumption. (ii) Without erosion, uplift would bring continental height asymptotically and exponentially to a critical height H_c above which a continent would collapse, either because of gravitational pull, or because of softened (hot) continental root as it thickens, or because of glacier cutting. With the assumptions, a first order approximation for uplift rate is:

$$(\mathrm{d}H/\mathrm{d}t)_{\mathrm{uplift}} = (H_c - H)/(\beta\tau), \tag{7}$$

where $\beta\tau$ is a time scale for uplift, and H_c is the critical height for the continent to remain stable. H_c is allowed to vary from 4 to 8.8 km (the elevation of Mount Everest) to 15 km.

Combining Eqs. (6) and (7), the elevation of a continent would vary as follows:

$$\begin{aligned}\mathrm{d}H/\mathrm{d}t &= (\mathrm{d}H/\mathrm{d}t)_{\mathrm{uplift}} + (\mathrm{d}H/\mathrm{d}t)_{\mathrm{erosion}} \\ &= (H_c - H)/(\beta\tau) - D\gamma H/\left(\beta A^{1/2}\delta\right).\end{aligned} \tag{8}$$

(g) The above differential equation can be solved to investigate the transient behavior, but more input information, such as continental size as a function of time, H_c, τ, and $D\gamma/(\beta A^{1/2}\delta)$, would be needed. Only the steady state ($\mathrm{d}H/\mathrm{d}t=0$) is considered in this work. The steady-state solution of Eq. (8) is:

$$H_{ss} = \frac{H_c}{\dfrac{D\gamma\tau}{A^{1/2}\delta} + 1}, \tag{9}$$

where H_{ss} is the steady-state mean continental height. In the above equation, the parameter $D\gamma\tau/(A^{1/2}\delta) = (\beta\tau)/[\beta A^{1/2}\delta/(D\gamma)]$ is the ratio of the characteristic time scale for uplift, $\beta\tau$, to that for erosion, $A^{1/2}\beta\delta/(D\gamma)$. By solving Eq. (8), the characteristic time scale to reach the steady-state elevation is smaller than either one of the two time scales and is $1/[1/(\beta\tau) + D\gamma/(A^{1/2}\beta\delta)]$.

Although the justification for some of the above assumptions is not available because there is no quantitative understanding of some processes, it is of interest to investigate whether this simple model can account for various observations.

3. Present-day mean elevation versus area of continents

Eq. (9) shows that at steady state, the mean continental height H increases with continental area A, a prediction roughly consistent with observations [1–3]. More quantitatively, the relation between present-day values of H versus A can be fit using Eq. (9) by adjusting two parameters, H_c and $D\gamma\tau/\delta$. It turns out that only one of the two parameters can be constrained from the H versus A data of five continents. Hence, H_c is allowed to vary independently from 4 to 15 km, and $D\gamma\tau/\delta$ is obtained by a least-squares fit (Fig. 1).

Fig. 1 shows that my model closely reproduces the general trend of mean elevation versus area for all major continents, except for Antarctica. Fig. 1A fits land area versus land mean height [1,2], and Fig. 1B fits continental (including continental shelves) area versus mean height [3]. The relation in Fig. 1A is significantly better than that in Fig. 1B, probably because my erosion model does not adequately account for elevation variation with time for areas below sea level (continental shelves). Previous workers often treated Eurasia as two continents, Europe and Asia (e.g., [1,2,9]). In Fig. 1, Eurasia is treated as one continent because Eurasia is one landmass and

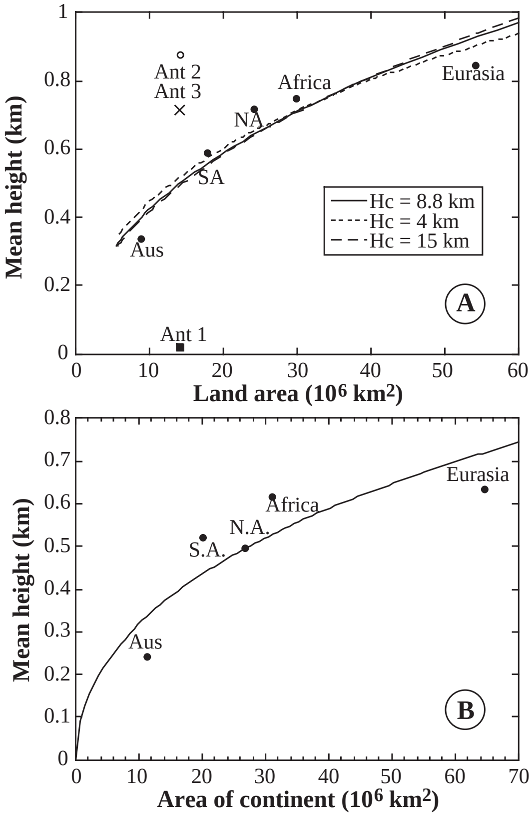

Fig. 1. Mean height vs. area of continents: data and model. A. Land area of each continent versus its mean height; and B. Continental area (including continental shelves; data from [3]) versus its mean height. Solid dots: mean elevation of continents [1,2]; curves are fits to data of five continents (Eurasia, Africa, Aus = Australia, NA = North America, SA = South America) with different assignment of H_c. To fit data in A, $D\tau\gamma/\delta = 7751H_c - 5813$, where $H_c > 2$ km, and both $D\tau\gamma/\delta$ and H_c are in km. Mean elevation of Antarctica at 2.2 km is not shown so as not to compress the vertical axis, but that of rock surface is shown as filled square (Ant 1), that after isostatically removing ice [2] is shown as open circle (Ant 2), and that with equivalent rock topography [20,29] is shown as cross (Ant 3). Different shelf breaks [9] and difference between long- and short-term (such as ice formation) sea levels [8] are not considered.

has been together for long enough time, even though treating it as two separate continents would have significantly increased the quality of the fit (the mean land height of Asia is 0.96 km and that of Europe is 0.34 km; [1]). For Antarctica, the predicted elevation of 0.5 km does not match the observed mean elevation of 2.2 km, nor the mean elevation of rock surface of about 0.0 km, nor a 0.88 km model elevation by isostatically removing ice cover [2], nor a 0.72 km model elevation by replacing ice using equivalent mass of rock [20,29]. The inability to predict the mean elevation of Antarctica is not surprising because of ice cover on Antarctica. The thickness of ice cover can vary within a short period of time (less than 1 million years) and hence its thickness does not mean much in the context of my steady-state model. On the other hand, the rock surface is the meaningful surface. A mean rock surface elevation of about 0.0 km implies that ice is very efficient in eroding the rock surface.

Ignoring Antarctica, small misfits of other continents are attributed to the individuality of continents, such as tectonic and thermal regimes, erosion rate, dynamic topography, Greenland ice cover, and whether or not steady state is reached.

From the fitting parameter $D\gamma\tau/\delta$, another parameter, $D\gamma\tau/(A^{1/2}\delta)$ (the ratio of the characteristic time scale for uplift to that for erosion), can be estimated. For $H_c = 8.8$ km, the ratio ranges from 9 for Asia to 25 for Australia; for $H_c = 4$ km, the ratio ranges from 4 for Asia to 10 for Australia. Hence the characteristic time for uplift is about an order of magnitude greater than that for erosion. For example, if the characteristic time scale for erosion is taken to be 50 Myr [19], that for uplift would be of the order 500 Myr, and that for reaching the steady state would be smaller than either of the two, about 45 Myr. The reason for such a long time scale for tectonic uplift is as follows: the uplift time scale is for the whole continent, whereas in a tectonic cycle, only part of a continent is uplifted. Hence the time scale for uplifting a plateau or an orogenic belt is much shorter.

4. Mean thickness of continental crust during the Phanerozoic

Eq. (9) can also be used to predict mean continental height during Phanerozoic time, over which the total volume of the continental crust (V) and that of ocean water are assumed to be constant. (It has been shown that ≤10% change in continental volume and/or seawater volume can be roughly modeled as constant volumes [6].) The total area of continents would vary with the mean crustal thickness as V/h. For N equal-sized continents, the area of each continent is $A = V/(Nh)$. Using Eq. (9), area and mean elevation of continents as a function of N are derived iteratively as follows. Given N, in the first iteration, Δz is taken to be zero, and A is estimated from the present-day continental area divided by N. Then H is calculated using Eq. (9) and (H_c, $D\gamma\tau/\delta$) values from Fig. 1. Then h is calculated from isostasy considerations (Eq. (1)). Then total continental area is calculated as V/h. Then, new A, new oceanic area, new ocean depth, and new Δz are all calculated. The new A would be slightly different from the initial estimate. Then H, h, total continental area, total ocean area, new ocean depth, Δz, and A are estimated again iteratively until convergence.

Calculated results of mean continental height and Δz as a function of N are shown in Fig. 2. It shows that if there was only one supercontinent, the mean continental height would increase by 0.54 km compared to the present, mean continental crust thickness would increase by 3 km, and total continental area would decrease by 7%. Oceanic area would increase by 4% and oceanic depth would drop by about 0.2 km. That is, absolute sea level (as measured by mean ocean water depth) would be lowest during the time of supercontinent. Although absolute sea level defined here is not identical to relative sea level as measured by transgression and regression, the two most prominent lows in the relative sea level curve correspond to

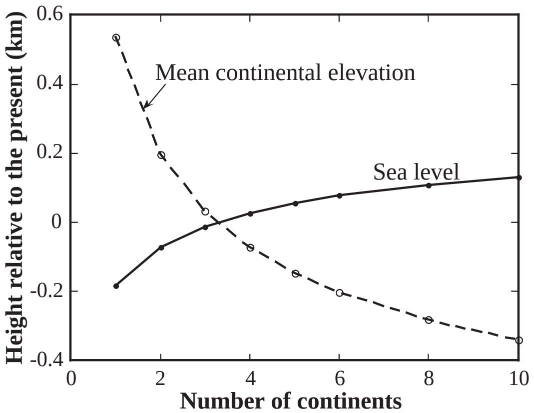

Fig. 2. Calculated mean elevation of continents (open circles) and sea level (solid circles) as a function of the number of continents. Present-day elevation and sea level are defined to be zero. Hence, a negative mean elevation in this figure does not mean that the continents are below sea level. Present mean continental height and crustal thickness correspond to those for 3.3 (not 6) equal-sized continents, because the present six continents are not of equal size.

the two periods of supercontinents, Pangea at about 220 Ma and Rodinia at about 560 Ma [7–9]. In the context of this model, I suggest that, contrary to the model of Worsley et al. [9], the lower sea level during supercontinents is not due to thermal insulation of the continents, but due to the inherent difficulty in reducing the mean height of a large continent by erosion.

A corollary of this model is that total mass erosion rate of continents (eroded mass per unit time) decreases when the number of continents decreases. The estimation of the effect is as follows. If a supercontinent formed from 4 equal-sized continents, before new steady state is reached, the total mass erosion rate, $ND\gamma A^{1/2}H/(\beta\delta)$ (the right-hand side of Eq. (5) times N), would be only 50% of that for four equal-sized continents assuming $D\gamma H/(\beta\delta)$ stays the same. When new steady state is reached for the supercontinent, H would increase, and calculation shows that the erosion rate would still be smaller by 20% than that for four equal-sized continents. The change in global erosion rate would lead to variations in some radiogenic isotopic ratios in seawater. For example, assuming constant oceanic hydrothermal flux, a smaller erosion rate would lead to a smaller seawater $^{87}Sr/^{86}Sr$ ratio during Pangea, consistent with the seawater Sr isotope curve [30]. To quantify the effect of continent size on past seawater $^{87}Sr/^{86}Sr$ ratio is impossible at present because there is not even an agreement on the current seawater Sr budget [31].

It should be noted that consideration of the number of continents alone can only explain some features of the sea level curve (such as low sea level during Pangeas), but cannot explain the whole sea level curve (such as high sea level during Cretaceous). Other factors such as seafloor spreading rate and dynamic topography must be considered to explain the whole sea level curve.

5. Conclusions

The topography of continents is controlled by the global tectonic and climatic systems. A semi-mechanistic and quantitative model considering such controls is able to account for the following observations: (i) the present-day relation between the area and mean elevation of individual continents (except for Antarctica), (ii) two prominent lows (by about 0.2 km) during the times of Pangea and Rodinia in the sea level curve, and (iii) lower $^{87}Sr/^{86}Sr$ ratio during the times of Pangea. Because rearrangement of continents takes a long time, short-term (high-frequency) features in the sea level curve cannot be explained by this model. When the model is applied to investigate Phanerozoic variations of mean thickness of continental crust, it is predicted that continents thicken during the times of supercontinents, but the variation on the mean thickness of continental crust amounts to only about $\pm 10\%$. In summary, because of interplay between tectonic and climatic systems, the number of continents (or the size of continents) plays a significant role in sea level, continental elevation, and continental erosion rate. It may be possible to couple this simple model with sophisticated geodynamic models [10] for a better understanding of marine inundation and sea level. Mean thickness of continental crust has not changed significantly in the Phanerozoic.

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