Modern and ancient continental hypsometries

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Abstract: Controls on coastal hypsometries of modern continents and oceanic islands may be used to estimate palaeo-continental area-elevation distributions. The most important controls on coastal hypsometry are landmass area and coastal gradient, increases in which result in steeper area-elevation distributions. Thus, any change in sea level will flood or expose a smaller fractional area of continens which are large or have steep coastal gradients than of those which are small or have gentle coastal gradients. Coastal gradients are dictated primarily by the tectonic setting and age of continental margins. While active and transform margins generally have steep coasts, mean coastal gradients for passive margins decrease exponentially with increasing margin age. Consequently, continents composed mostly of mature passive margins are generally more 'flattened'.

These observations are applicable to a variety of palaeo-eustatic problems. Hypsometries corrected for changes in landmass area reduce variance in flooding records between different Palaeozoic continuents by more than 50%. Assuming minimum-variance superposition of flooding curves for different Palaeozoic continents permits identification of possible hypsometric anomalies. Calculated Palaeozoic sea level maxima are between +200 m and +400 m with respect to present sea level, substantially lower than previous estimates to +600 m.

Modelling evolution of coastal hypsometries due to changes in length and age of passive margins suggests that hypsometry varies cyclically in response to continental rifting and collision. Coastal hypsometries are steepened during rifting events and decrease by a factor of two to three as passive margins age. Coastal hypsometries in a 'flattened' world are steeper than those of a 'ruffled' world due to increased continent size and decreased passive margin length. Thus, sea level highstands associated with times of continental dispersal may result as part from the enhanced 'flattenedness' of ruffled continents.

Hypsometry is the distribution of area with respect to elevation for a topographic surface. Combined with continental flooding data, hypsometry can be used to determine sea level elevation (Fig. 1). The technique utilizes palaeogeographic maps which show an interpolated shoreline based on the distribution of marine and continental facies of a given age, and from which fractional land area flooded is measured directly. A landmass-specific hypsometric curve is then used to convert fractional area flooded to sea level elevation. In this manner, a continuous sea level curve can be constructed from a series of palaeogeographic maps for successive geologic epochs (Fig. 1).

Hypsometry and continental flooding data thus permit direct quantification of palaeo-sea levels. Integration of flooding data over large, relatively stable cratonic areas avoids many of the difficulties inherent in seismic stratigraphic studies due to collation of absolute sea level changes with passive margin subsidence (Pitman 1978). While hypsometry does not yield as detailed a sea level record as do seismic or sequence stratigraphic analyses, it is probably the best method for estimating the amplitudes of long-term (i.e. >10 Ma) sea level changes. Furthermore, the technique is applicable to the entire Phanerozoic, unlike methods based on passive margin seismic stratigraphy (Vail et al. 1977; Haq et al. 1987) and mid-ocean ridge volume analysis (Pitman 1978). Derivation of a reasonably accurate record of long-term Phanerozoic eustasy is critical to our understanding of such diverse phenomena as secular changes in oceanic and atmospheric chemistry (Berner et al. 1963), species diversity and extinction events (Wyatt 1987; Schufy 1974), and sedimentation patterns (Shanmugam & Moiola 1982).

The most fundamental problem inherent in hypsometric investigations of palaeo-eustasy is the absence of any record of ancient area-elevation distributions. In contrast, modern hypsometries of individual continents and of the globe as a whole are well-established (Kossina 1933; Harrison et al. 1983; Cogley 1985), and have been employed without modification in all hypsometric studies of palaeo-eustasy known to us (Peremy 1975; Bond 1979b; 1983; Harrison et al. 1981, 1983; Hallam 1984). In these studies, any change in freeboard for a given landmass is assumed to be a direct reflection of comparable eustatic change, and a temporally-invariant hypsometric curve serves only to convert fractional area flooded to sea level elevation.

In reality, continental hypsometries have probably evolved through time due to such geotectonic and geomorphological factors as rifting and suturing of landmasses, changes in type and age of continental margins, and uplift and peneplanation of cratonic interiors. While unlikely, it is at least conceivable that Phanerozoic sea level has remained approximately constant and that changes in continuous flooding have resulted primarily from changes in hypsometry. More probably, continuous flooding records the interaction of these two phenomena, and attempts to extract sea level from flooding data will be unsatisfactory unless hypsometric changes are taken into account.

What is required is knowledge of changes in continental hypsometries through time, but no direct record of such variation exists and it is necessary to find one or more proxy indicators which yield acceptable estimates of palaeo-
Sea level and hypsometry

Hypsometric slope

Harrison et al. (1981) recognized that major changes in global hypsometry may occur during the course of geotectonic cycles and attempted to quantify the effect on the global hypsometric curve of putting all landsmasses into a single continent (e.g. Perno-Triassic Pangaea). They postulated that a supercontinent would have a higher rate of change of elevation with respect to fractional land area, and that larger sea level changes would be necessary to flood or expose a given percentage of total area. We formalize the idea of the rate of change of elevation with respect to fractional area as 'hypsometric slope', which has dimensions of metres per percent area. In theory, it is an instantaneous measurement of hypsometric slope changes continuously with changes in elevation. In practice, continental areas have been determined only at elevation intervals of 100 m (Harrison et al. 1983), and their derivative hypsometric slopes are thus calculable only for a range of elevations of 100 m or more. For all continents, hypsometric slope is relatively gentle at low elevations and steepens rapidly at higher elevations (Harrison et al. 1981). Significantly, at any given elevation, there are large differences in hypsometric slope for different continents, resulting in variable rates of flooding or exposure due to eustatic change (Fig. 2).

In eustatic studies, only low-elevation areas are of significance. Examination of modern continental hypsometric curves shows that each continent exhibits a relatively uniform hypsometric slope over an elevation range from −50 m to +250 m with respect to present-day sea level (Fig. 2; Harrison et al. 1983). As a consequence, sea level changes within this range produce linear changes in area flooded for each continent at a continent-specific rate. This observation is the basis for determining a 'coastal zone' hypsometric slope for each landmass. While Phanerozoaic sea levels may occasionally have exceeded this elevation range (Halám 1984), global sea level has oscillated within a fairly narrow

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**Fig. 1.** Example of sea level determination using hypsometry and area-flooded data. (A) Percent area flooded for a given epoch measured from an equal-area palaeogographic map. (B) Percent area flooded converted to sea level elevation using a landmass-specific hypsometric curve. (C) Sea level curve constructed from a series of palaeogeographic maps.

**Fig. 2.** Hypsometric curves of present-day continents from −200 m to +800 m with respect to sea level. Note that hypsometric slope, or the rate of change of elevation with respect to fractional area, is relatively constant for each continent from −50 m to +250 m but varies substantially between continents. 'Coastal hypsometric' were determined for the elevation range 0−200 m. Data from (Harrison et al. 1983).
Fig. 4. Model hypsometric slopes resulting from various combinations of three geometric parameters. Main figure shows change in hypsometric slope due to increasing landmass area for six different values of coastal gradient (shoreline ratio is constant and equal to 1.0). Figure offset to lower right shows hypsometric slopes resulting from increasing landmass area for four different values of shoreline ratio (coastal gradient is constant and equal to 10 in km⁻¹). These parameter ranges bracket those of most present-day landmasses.

Fig. 5. Hypsometric slope versus area for modern landmasses. Oceanic islands are shown by shaded symbols (Canary Islands, circles; Hawaiian Islands, triangles; Other islands, squares). Uplifted continents are shown by solid circles and main trend continues by solid squares.

Oceanic Islands: (Canary Is.) FO, Fuerteventura; GA, Lanzarote; GC, Gran Canaria; HI, Hierro; LP, La Palma; LZ, Lanzarote; TN, Tenerife; (Hawaiian Is.) HI, Hawaii; KH, Kaho'olawe; KI, Kīlauea; LI, Lānaʻi; MI, Maui; MK, Molokai; NI, Ni'ihau; OA, Oahu; (Others) AS, Ascension; SM, Guam; HL, St. Helena; MA, Madeira; PE, poorest; PS, Puerto Santo. Continents: AF, Africa; AR, Arabia; AS, Asia; AU, Australian; BF, Baffin; BK, Bask; BR, Bonovo; CB, Cabo; EA, Eurasia; EU, Europe; FK, Falklands; HM, Hawaii; IN, India; IA, Indiana; MD, Madagascar; NA, North America; NG, New Guinea; PR, Puerto Rico; SA, South America; SG, South; SK, Sakhalin; SL, Sri Lanka; SM, Sumatra; TS, Tasmania; VC, Victoria; WOR, World. Area-elevation data from Harrison et al. (1983) and Times Atlas of the World (1985).

Empirical fit curves model hypsometric slope trends for four different landmass groups. Curve parameters: (1) Hawaiian Island Trend: $CG$ (coastal gradient) = $1.65\ m/\ km^1$, $CG \times AR$ (area); (2) Canary Island trend: $CG = 1.65\ \log_{10}\ km^{-1}$ at $400\ km^2$, $CG = AR^2$; (3) Main continental trend: $CG = -0.00\ \log_{10}\ km^{-1}$ at $1 \times 10^8\ km^2$, $CG = AR^{-2.5}$; (4) uplifted continental trend: $CG = 2.00\ \log_{10}\ km^{-1}$ at $1 \times 10^8\ km^2$, $CG = AR^{-2}$.

The dashed horizontal line represents the hypsometric slope 'floor'; no landmass with a minimum elevation less than 200 m was considered and, thus, no hypsometric slope less than 2.00 m/m² was possible.
Fig. 7. Relationship of coastal gradient to passive margin age. Mean coastal gradient decreases exponentially with increasing margin age (note that the ordinate has a logarithmic scale). The equation of the regression line (solid) is $y = -0.0062 \times x + 1.28$; its correlation coefficient is $r = 0.80$. Coastal gradient standard deviations are not age-dependent; therefore, the mean standard deviation (0.42 log km$^{-1}$) is shown as an envelope (shaded) around the regression line. Mean coastal gradients for Northern and Southern Norway (NNY and SNY) were excluded from regression calculations. The mean and standard deviation envelope for active and transform margins are shown on left side of figure. ACT, Active/transform margins; EAF, Eastern Africa; EAU, Eastern Australia; ENA, Eastern North America; ESA, Eastern South America; GCA, Gulf of California; GMS, Gulf of Mexico; IND, India; IRL, Ireland; NAF, Northern Africa; NAS, Northern Asia (Antic); NEAU, Northeastern Australia; NPY, Newfoundland; NNA, Northern North America (Antic); NNT, Northern Norway (Atlantic); NWAF, Northwest Africa; NWWA, Northwestern Australia; RED, Red Sea; SAI, Southern Australia; SNV, Southern Norway (North St.); SRI, Sri Lanka; TSM, Tsiman; WAF, Western and Southwest Africa; WAU, Western Australia. Coastal gradients measured from: Atlas of the World (1985). Passive margin ages from Eldholm & Talwani (1982), Fjeldså (1982), Kern (1982), Lowell et al. (1975), Meeusen (1985), Salvador (1987), Scrutton (1973), Sweeney (1982), and Vevers (1982).

coastal gradients along sheared passive margins in which continents separate along a transform fault rather than a spreading centre, e.g. eastern Australia (Vevers 1982) and central Norway (Eldholm & Talwani 1982), owing to sharp transitions between continental and oceanic crust; and a lack of extensional listric faulting (Scrutton 1982). In contrast, a few coastlines exhibit anomalously low coastal gradients for their ages. For example, the Libyan coastline of North Africa formed an incompletely subducted embayment during the Cenozoic collision of the African and European cratons so that its effective age may be greater than that of the youngest (i.e. Palaeozoic) rifting event. Unfortunately, such deviations are largely unpredictable for ancient continents, and statistical mean trends (e.g. Fig. 7) must be relied on in the absence of detailed knowledge of palaeo-coastal morphologies.

Hypsocmetry of ancient continents Palaeozoic sea level The effect of area on hypsometric slope is readily demonstrable for present-day landmasses (Fig. 5). This relationship is of particular significance for the determination of Palaeozoic sea levels. Continental areas during the Palaeozoic geotectonic cycle were very different from those of the Mesozoic–Cenozoic cycle (e.g. Ziegler et al. 1979). For example, Palaeozoic Gondwana is now fragmented into Africa, Antarctica, Arabia, Australia, South America, and parts of southern Europe and Asia. On the other hand, the tectonic elements of present-day Eurasia formed a number of separate plates during the Palaeozoic, the largest of which were Baltica, China, Kazakhstan, and Siberia. Palaeozoic Laurentia had roughly the same area as present-day North America.

Three Palaeozoic landmasses of markedly different sizes were chosen to evaluate the landmass area effect on sea level estimates (Fig. 8). Below, most of which was comprised of two independent microplates averaging four million km$^2$ in area (Atlas of the Palearotography of China 1985), was a Palaeozoic landmass of microcontinental dimensions. Laurentia, which was about 31 million km$^2$ in area, was a mid-sized continent, while Gondwana, roughly 105 million km$^2$ in area, was the largest Palaeozoic landmass.

These continents have widely divergent flooding records with China representing maximum (c. 30–80%) and Gondwana minimum submergence (c. 5–25%); Fig. 8A). To evaluate the sea level record of each of these landmasses correctly requires selection of continental hypsometries on the basis of contemporaneity, rather than present, landmass area. Hypsometric slope estimates based on the main continental trend (Eq. 5, Appendix) are 4.1 m/km$^{2/3}$ for India, 5.8 m/km$^{2/3}$ for Laurentia, and 7.1 m/km$^{2/3}$ for China.

These values produce sea level curves for China and Laurentia which are in good general agreement (Fig. 8B). The most salient difference in their sea level histories, the absence of a Cambrian transgressive event in China, probably reflects differing geotectonic histories. The Chinese microplates were not part of a Late Proterozoic proto- Pangaea and thus did not participate in the Cambrian rifting event that affected the Australian margin. The Chinese microplates were not part of a Late Proterozoic proto-Pangaea and thus did not participate in the Cambrian rifting event that affected the Australian margin.

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the only extant supercontinent, Eurasia, suggests otherwise (Fig. 5).

An alternative approach to the analysis of Palaeozoic sea levels is possible (Fig. 9). This method assumes (1) that continental flooding primarily records a common eustatic history and (2) that differences in the magnitude of flooding between the Palaeozoic parts of China (Fig. 9A), Laurentia (Fig. 9B), and Gondwana (Fig. 9C) are due to differing areas of landmasses. Using this approach, we can calculate regional sea-level variations for each continent and compare them with those of other regions. The results indicate that a large number of palaeo-limnographies could be reconstructed and used to determine the ages and positions of continental margins. An alternative approach is to fit values for China (5.9 m/century) and Laurentia (6.1 m/century) to the main trend and assume that

Gondwana (12.2 m/century) had an anomalously steep coastal hypsometry (Fig. 9B, right inset). The resulting sea level scales differ by a factor of 1.4 (Fig. 9B).

Comparison of sea level amplitudes based on estimated palaeo-hypsometries with those of Hallam's (1964) Phanerozoic eustatic curve reveals major differences. Hallam determined a Palaeozoic maximum of +600 m based mainly on flooding data for the Soviet Union (Vinogradov 1967-69, cited in Hallam 1977). As mentioned previously, Palaeozoic Eurasia comprised numerous smaller landmasses (Ziegler et al. 1979), and sea level estimates for these continents cannot meaningfully be based on present-day Eurasian or global hypsometry. Our analysis indicates that the magnitude of the Silurian-Ordovician highstand was between +200 m and +400 m, depending on variations in method of calculation, rather than +600 m (Figs 8 & 9). Our estimator for the Devonian-Carboniferous highstand of +100 m to +200 m is also substantially lower than Hallam's +380 m to +400 m.

This example illustrates the utility of hypsometric slopes in eustatic analysis. Much of the variance in flooding between different Palaeozoic landmasses can be accounted for through correction for changes in landmass area. Further, the resulting sea level estimates are probably superior to previous estimates due to consideration of multiple flooding records. Ideally, flooding records for a large number of palaeo-limnographies could be constructed and used to determine hypsometric adjustment ratios. These ratios could then be fitted to the main trend of an area-hypsometric slope plot (e.g. Fig. 5) and continents with anomalous hypsometries identified through their divergence from the curve.

Hypsometry and geotectonic cycles

The dependence of coastal gradient on type of continental margin has probably been a major factor...
contributing to changes in palaeohypsometry. Such changes occur as the frequency and age distribution of passive margins for each continent and for the world as a whole evolve through time. As a first approximation, this evolution is cyclic and mirrors plate tectonic phases of continental rifting and collision. During a Pangaea phase, total passive margin length (frequency) is lower than during a rifted phase. A rifting event results in lengthening of passive margins and a shift in the age distribution toward young, steep margins. As continents disperse and passive margins age, the age distribution eventually shifts back toward older, flatter margins, reducing hypsometric slopes and enhancing the ‘floodability’ of continents. Finally, continental collisions consume passive margins and their total length (frequency) decreases, initiating a new Pangaea phase. Concomitant changes in ocean basin volume through such a geotectonic cycle (not considered here) will reinforce flooding trends associated with continent margin evolution. Ocean basin volume decreases during rift-drift phases due to thinning and lateral extension of continental crust (Wyatt 1986) and to increase in size of mid-ocean ridges (Heller & Angenieux 1986), as a consequence of which sea levels are forced higher onto continent margins.

The impact of a geotectonic cycle of this type on continental hypsometry can be evaluated through calculating model hypsometric slopes based on passive margin age-coastal gradient relations. As a first approximation, coastal gradients for any coastline are log-normally distributed (Fig. 6), and mean coastal gradients for passive margins decrease exponentially with age (Fig. 7). Since coastal gradient standard deviations do not show statistically significant age-dependence, the mean standard deviation for the data set (0.42 log m km−1) is used for margins of all ages.

These data permit estimation of the most likely distribution of coastal gradients for a passive margin of any given age. Changes in coastal hypsometry were modelled through a 300 Ma plate tectonic cycle (Fig. 10). This cycle comprises an initial 50 Ma sutured phase with a fixed length of ‘active’ margin, a 200 Ma rift-drift phase during which an additional increment of passive margin is created, and a 50 Ma collision phase during which the passive margin is consumed. Hypsometric slope is considered in these phases, and a pre-rift ‘active’ margins were assigned a coastal gradient distribution representing a combination of active and mature passive margins (Eq. 8, Appendix). Coupling of active and passive margin coastal gradients (e.g. as through thermal uplift) results in an initial increase in hypsometric slope during rifting. Thereafter, hypsometric slope decreases progressively during the rift-drift phase due to the age-dependency of passive margin coastal gradients (Fig. 7). Finally, hypsometric slope returns to its pre-rift value as mature passive margins are progressively consumed (e.g. as through oblique convergence along a subduction zone).

Ranges of values for landmass area, passive margin length, shoreline ratio, and active margin coastal gradient representative of modern continents are used to illustrate the possible variation in, and rate of change of, hypsometric slope through a geotectonic cycle (Fig. 10). Correspondence between real and model hypsometric slopes is generally good. North America, South America, and Australia plot well within model limits for continents of their respective areas and passive margin age ranges (Fig. 10b). While Eurasia has a rather low hypsometric slope given its range of passive margin ages, lengthy low-gradient, supermature passive margins along its Arctic coast may skew its hypsometric slope toward unusually low values. Africa, on
The other hand, has an exceptionally high hypsometric slope, an anomaly which may be attributable to its geotectonic history of multiple rift events (Fig. 10A). Evidence that Africa is currently undergoing a rift-related thermal uplift supports this interpretation (Bond 1979b).

Such a model permits several general observations of significance. First, through the course of a geotectonic cycle, hypsometric slope changes by a factor of two to three under most combinations of model parameters. As a consequence, continental ‘flodability’, i.e. the rate at which a landmass is potentially flooded or exposed, is two to three times greater at the end of a rift-drift phase than at its beginning. Second, in any given geotectonic cycle, the range of possible hypsometric slopes for a landmass of given area is relatively large (Fig. 10A). Therefore, x is not possible to estimate hypsometric slope from position within a geotectonic cycle alone. Knowledge of hypsometric slope at some point within the cycle (e.g. the present) may permit projection of changes in hypsometric slope (e.g. into the past or future), especially if data regarding continent margin events are available. Such projection must remain tentative as vertical continental isochrons due to local tectonic, thermal, or isostatic influences can alter coastal hypsometries in an unpredictable manner on geologically short timescales.

Conclusions
The most important determinants of coastal hypsometry are landmass area and coastal gradient. Increases in either parameter result in steeper area-elevation distributions. As a consequence, any given change in sea level will flood or expose a smaller fractional area of continents which are large or have steep coastal gradients than of those which are small or have gentle coastal gradients. Coastal gradients are dictated primarily by the tectonic setting and age of continental margins. While active and transform margins generally have steep coasts, most continental coasts for passive margins decrease exponentially with increasing margin age. Consequently, continents comprised mostly of mature passive margins are generally more flodable.

These observations are applicable to several problematic aspects of palaeo-oceanography. Hypsometries torected for changes in landmass area reduce variance in flooding records between different Palaeocoeceans by more than 50%. Reconstructed Palaeocean sea level maxima are between +200 m and +400 m with respect to present sea level, substantially reducing previous estimates of +400 m. Modelling of changes in continental hypsometry due to increasing passive margins age suggests that coastal hypsometries change by a factor of two to three through geotectonic cycles. Thus, sea level highstands associated with times of continent dispersal may result in part from the enhanced ‘flodability’ of rifted continents.

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A appended: Mathematical relationships of model hypsography

A model hypsography of conical form permits mathematical definition of hypsographic slope. Whether or not the side of the cone steepens above its base (i.e. of real landmasses) is irrelevant, as we are interested only in the basal section. By definition, hypsographic slope is the rate of change in elevation with respect to fractional land area. For operational reasons, it is calculated over some elevation range of interest:

\[ \text{Hypsographic Slope} = \frac{\Delta \text{Elevation}}{\Delta \text{Fractional Land Area}} \]

where \( \Delta \text{Elevation} \) is in meters, \( \Delta \text{Fractional Land Area} \) is in m\(^2\)/%.

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