Topography of mountain belts controlled by rheology and surface processes

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It is widely recognized that collisional mountain belt topography is generated by crustal thickening and lowered by river bedrock erosion, linking climate and tectonics[1–4]. However, whether surface processes or lithospheric strength control mountain belt height, shape and longevity remains uncertain. Additionally, how to reconcile high erosion rates in some active orogens with long-term survival of mountain belts for hundreds of millions of years remains enigmatic. Here we investigate mountain belt growth and decay using a new coupled surface process[5,6] and mantle-scale tectonic model[7]. End-member models and the new non-dimensional Beaumont number, Bm, quantify how surface processes and tectonics control the topographic evolution of mountain belts, and enable the definition of three end-member types of growing orogens: type 1, non-steady state, strength controlled ($Bm > 0.5$); type 2, flux steady state[8], strength controlled ($Bm \approx 0.4–0.5$); and type 3, flux steady state, erosion controlled ($Bm < 0.4$). Our results indicate that tectonics dominate in Himalaya–Tibet and the Central Andes (both type 1), efficient surface processes balance high convergence rates in Taiwan (probably type 2) and surface processes dominate in the Southern Alps of New Zealand (type 3). Orogenic decay is determined by erosional efficiency and can be subdivided into two phases with variable isostatic rebound characteristics and associated timescales. The results presented here provide a unified framework explaining how surface processes and lithospheric strength control the height, shape, and longevity of mountain belts.

Mountain belt evolution in collisional settings comprises crustal thickening and surface uplift, followed by tectonic quiescence and isostatic rebound that may include extensional collapse. Surface processes shape mountain belt surface morphology by counteracting tectonic growth and by causing topographic decay. End-member collisional mountain belt types[9] include (1) active, narrow orogens with high rock uplift and erosion rates, such as those in Taiwan and the Southern Alps of New Zealand (SANZ), (2) active, wide orogens with orogenic plateaus and overall low erosion rates, such as in Himalaya–Tibet and the Andes, and (3) inactive orogens with slowly decaying topography surviving tens to several hundreds of millions of years, such as the Urals or the Appalachians. High erosion rates in small orogens, co-existence on Earth of large and small orogens of variable height, and the long-term survival of orogenic topography raise fundamental questions about the factors controlling the width, height and longevity of mountain belts during their growth and decay phases. In non-glaciated mountain belts, rainfall and river incision control the erosional efficiency, denudation rate and sediment yield[2,10,11], implying that climate may set the width, height and relief in growing orogens[2,3,12–18]. Erosional efficiency is also thought to control the longevity of mountainous relief[19,20]. However, others have shown[21–24] that finite crustal strength may be the main factor limiting the maximum elevation of orogens under some circumstances, demonstrating the need for a proper representation of tectonic deformation to study the effect of the erosional efficiency on mountain growth and decay, which includes isostasy, mantle lithosphere subduction, discrete faulting, and proper earth-like rheologies.

Coupled tectonic–surface process model

We use the thermo-mechanical tectonic model FANTOM[7,25] coupled to the landscape evolution model FastScape[5,6], resolving the interaction between upper mantle scale tectonic deformation and surface processes at high resolution. FANTOM computes deformation of earth-like materials with frictional-plastic and non-linear thermally activated viscous flow, whereas FastScape solves for river erosion, hillslope processes, sediment transport and deposition (Methods; Extended Data Fig. 1). The erosional efficiency depends mostly on fluvial erodibility K_f, which spans several orders of magnitude owing to its dependence on rainfall, rainfall variability, lithology, fracturation, vegetation cover, and abrasive agents[14,20,26–29]. We present three end-member models with low (model 1), high (model 2), and very high (model 3) fluvial erodibility (Figs. 1 and 2).

End-member model results

In all models, shortening is accommodated by one-sided subduction of the strong lower crust and lithospheric mantle, and mountain building

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