

Morphotectonic Evolution of Landscape

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The Volcano Colima is located approximately 100 kilometers from the Pacific coast in southwestern Mexico. It measures 3860 meters and has a five-million-year history of volcanic activity. Colima is assessed even as one of the most dangerous volcanoes worldwide. In recent years, the volcano has erupted frequently and spat in recent months have often an ash cloud in the sky. Now the surrounding villages had to be evacuated again.

Plate tectonics has been on the curriculum for a while now. Most people are fairly familiar with the theory; the Earth's crust - or lithosphere - is broken up into different plates which move around on the very hot, viscous substance beneath. This movement is driven by the heat of the Earth's core, causing the viscous substance beneath the Earth's crust to flow and move the plates above, like pieces of toast floating in a massive bowl of hot beans. Plates can move closer together or further away from each other. Over millions of years, these movements define our continents. Plates coming together can form mountain ridges, whereas plates moving away from each other can form oceans. On a smaller timescale, plate tectonics is responsible for earthquakes, volcanoes and tidal waves. Now, scientists at the Scripps Institution of Oceanography have identified a new force driving plate tectonics: plumes of hot magma swelling up from deep inside the Earth's core, up to 2,500 kilometres deep. Published in the journal Nature, the research shows that these hotspots, or "mantle plumes", are responsible for the movement of whole continents. These plumes of incredibly hot, molten rock push against continental plates and drive their movement. Around 70 million years ago, the tectonic plate that now includes the Indian subcontinent lay northeast of Madagascar. Suddenly, it started moving incredibly quickly - by geological standards - at 10 centimetres per year. Around the same time, a spate of huge volcanoes occurred at the Deccan Plateau, sited in the area that is now India. Molten lava was thrown over around 1.5 million square kilometres and the volcanoes coincided with the mass extinction of dinosaurs, two events which some scientists think are related. Steve Cande and Dave Stegman, who led the study for Scripps Institution of Oceanography, tracked movements of continental plates throughout Earth's history. Their research suggests that the Indian subcontinent tectonic plate sat over a powerful mantle

plume which began around 70 million years ago, around what is now the Reunion Islands. This rising mass of hot rock hit the Earth's crust and spread out. The pushing force of the mantle plume sent the Indian plate hurtling towards what is now Asia. The Reunion mantle plume is also thought to be responsible for the mass volcanism at the Deccan Plateau. Cande and Stegman also think that the Reunion mantle plume caused the African tectonic plate to slow down for around 5 million years. After the plume subsided, the Indian tectonic plate slowed to a more normal geological movement of around a few centimetres each year, whereas the African plate sped up. The movement of these two plates in sync, always countering each other, provides strong evidence that a powerful mantle plume was responsible for their motion. The area around Reunion Islands is still a hotspot for volcanic activity, even though the plume has now spluttered out. Could mantle plumes be responsible for more of our present mountain ranges, volcanoes and continents? Whilst not all scientists agree that mantle plumes are responsible for the movement of whole continents, the latest research should shed new light on the way we think about the Earth's geological features.

Abstract: Research in landscape evolution over millions to tens of millions of years slowed considerably in the mid-20th century, when Davisian and other approaches to geomorphology were replaced by functional, morphometric and ultimately process-based approaches. Hack's scheme of dynamic equilibrium in landscape evolution was perhaps the major theoretical contribution to long-term landscape evolution between the 1950s and about 1990, but it essentially 'looked back' to Davis for its springboard to a viewpoint contrary to that of Davis, as did less widely known schemes, such as Crickmay's hypothesis of unequal activity. Since about 1990, the field of long-term landscape evolution has blossomed again, stimulated by the plate tectonics revolution and its re-forging of the link between tectonics and topography, and by the development of numerical models that explore the links between tectonic processes and surface processes. This numerical modelling of landscape evolution has been built around formulation of bedrock river processes and slope processes, and has mostly focused on high-elevation passive continental margins and convergent zones; these models now routinely include flexural and denudational isostasy.

Major breakthroughs in analytical and geochronological techniques have been of profound relevance to all of the above. Low-temperature thermochronology, and in particular apatite fission track analysis and (U-Th)/He analysis in apatite, have enabled rates of rock uplift and denudational exhumation from relatively shallow crustal depths (up to about 4 km) to be determined directly from, in effect, rock hand specimens. In a few situations, (U-Th)/He analysis has been used to determine the antiquity of major, long-wavelength topography. Cosmogenic isotope analysis has enabled the determination of the 'ages' of bedrock and sedimentary surfaces, and/or the rates of denudation of these surfaces. These latter advances represent in some ways a 'holy grail' in geomorphology in that they enable determination of 'dates and rates' of geomorphological processes directly from rock surfaces. The increasing availability of analytical techniques such as cosmogenic isotope analysis should mean that much larger data sets become possible and lead to more sophisticated analyses, such as probability density functions (PDFs) of cosmogenic ages and even of cosmogenic isotope concentrations (CICs). PDFs of isotope concentrations must be a function of catchment area geomorphology (including tectonics) and it is at least theoretically possible to infer aspects of source area geomorphology and geomorphological processes from PDFs of CICs in sediments ('detrital CICs'). Thus it may be possible to use PDFs of detrital CICs in basin sediments as a tool to infer aspects of the sediments' source area geomorphology and tectonics, complementing the standard sedimentological textural and compositional approaches to such issues.

One of the most stimulating of recent conceptual advances has followed the considerations of the relationships between tectonics, climate and surface processes and especially the recognition of the importance of denudational isostasy in driving rock uplift (i.e. in driving tectonics and crustal processes). Attention has been focused very directly on surface processes and on the ways in which they may 'drive' rock uplift and thus even influence sub-surface crustal conditions, such as pressure and temperature. Consequently, the broader geoscience communities are looking to geomorphologists to provide more detailed information on rates and processes of bedrock channel incision, as well as on catchment responses to such bedrock channel processes. More sophisticated numerical models of processes in bedrock channels and on their flanking hillslopes are required. In current numerical models of long-term evolution of hillslopes and interfluves, for example, the simple dependency on slope of both the fluvial and hillslope components of these models means that a Davisian-type of landscape evolution characterized by slope lowering is inevitably 'confirmed' by the models. In numerical modelling, the next advances will require better parameterized algorithms for hillslope processes, and more sophisticated formulations of bedrock channel incision processes, incorporating, for example, the effects of sediment shielding of the bed. Such increasing sophistication must be matched by careful assessment and testing of model outputs using pre-established criteria and tests.

Confirmation by these more sophisticated Davisian-type numerical models of slope lowering under conditions of tectonic stability (no active rock uplift), and of constant slope angle and steady-state landscape

under conditions of ongoing rock uplift, will indicate that the Davis and Hack models are not mutually exclusive. A Hack-type model (or a variant of it, incorporating slope adjustment to rock strength rather than regolith strength) will apply to active settings where there is sufficient stream power and/or sediment flux for channels to incise at the rate of rock uplift. Post-orogenic settings of decreased (or zero) active rock uplift would be characterized by a Davisian scheme of declining slope angles and non-steady-state (or transient) landscapes. Such post-orogenic landscapes deserve much more attention than they have received of late, not least because the intriguing questions they pose about the preservation of ancient landscapes were hinted at in passing in the 1960s and have recently re-surfaced. As we begin to ask again some of the grand questions that lay at the heart of geomorphology in its earliest days, large-scale geomorphology is on the threshold of another 'golden' era to match that of the first half of the 20th century, when cyclical approaches underpinned virtually all geomorphological work.

Keywords: long-term landscape evolution; tectonics; plate tectonics; passive margin; convergence zone; orogenic; post-orogenic; topography; climate; isostasy; low-temperature thermochronology; cosmogenic isotope analysis

Introduction

The earth sciences in general, and the science of landforms and landscapes ('geomorphology'), were in effect founded when, in the late 18th century, John Hutton realized the great depth of time available in Earth history for landscapes to evolve and he understood the centrality of slow non-catastrophic fluvial processes in this landscape evolution. The intellectual landscape was then opened up for the notion of biological evolution and all that has flowed from this (Chorley *et al.*, 1964). The grand, or even 'heroic', scale of the science of geomorphology is captured by the big ideas that have underpinned much of geomorphology from these earliest days, concepts such as base-level, grade, the concave-up equilibrium fluvial long profile, the law of accordant tributary junctions, and so on (Chorley *et al.*, 1964). Those who follow the modern literature on long-term landscape evolution will recognize at once the centrality of these ideas to the most recent research in long-term landscape evolution, which is the subject of this overview.

Landscape evolution over long timescales (millions to tens of millions of years) is inextricably tied, at least in the Anglophone literature, to W. M. Davis and his 'geographical cycle' (or 'cycle of erosion') (for example, Chorley *et al.*, 1973; Bishop, 2004). Despite early criticism (e.g. Tarr, 1898; Shaler, 1899), Davis and his approach dominated the discipline for more than half a century. The reasons for this persistence are not our concern here (for example, Judson, 1960; Chorley, 1965; Chorley *et al.*, 1973; Bishop, 1980), but the persistence *is* relevant to our present task to the extent that it ultimately triggered such a strong and negative reaction, with Davis coming to be caricatured 'as an old duffer with a butterfly-catcher's sort of interest in scenery' (Mackin, 1963, p. 136). The dissatisfaction was embodied in Strahler's (1952) call for radical change and the embracing of a new approach and underpinning concepts, ultimately taking the discipline into spatial and temporal scales much reduced from the grand vision and sweeping canvas of Davis and his disciples. Curiously, and as is outlined further below, Strahler's call is now being heard in long-term landscape evolution as geomorphology embraces quantitative and geochemical analytical approaches to the sorts of questions that Davis sought to address.

Strahler's call to abandon Davis was not met by an immediate adoption of the 'reduced-scale', quantitative approaches he advocated, with the Davisian approach (or at least an approach based in part on reconstruction of uplifted planation surfaces) initially persisting (e.g., Wooldridge and Linton, 1955). Cyclical approaches slowly faded, nonetheless, and long-term landscape evolution was pushed into something of a backwater, despite volumes such as that of Gardner and Scoging (1983). This side-lining and marginalizing of long-term landscape evolution was no doubt aided by the viewpoint that landscapes cannot be much older than the Tertiary and are probably no older than the Pleistocene (e.g., Thornbury, 1969). Viewpoints such as Thornbury's were derived by a simplistic dividing of the average elevation of the Earth's surface by measured rates of surface processes, thereby ignoring both the role of isostasy in relief maintenance and the spatial heterogeneity in erosion rates that modern techniques, such as cosmo-genic nuclide analysis, have quantified (see below for development of both of these points). In any event, the Thornbury analysis used measured erosion rates that were almost certainly too high because of anthropogenic disturbance (e.g., Judson and Ritter, 1964; Schumm, 1963). Even when more 'reasonable' long-term erosion rates were used and it was concluded that the

Daviesian cycle required between 10 and 25 Myr to run its full course (e.g., Gilluly, 1955; Schumm, 1963; Judson and Ritter, 1964; Melhorn and Edgar, 1975), this only stretched the timescale just beyond the Neogene.

Throughout this period of marginalization (or at least under-emphasis) of long-term landscape evolution as a central research issue, Twidale (1976), Crickmay (1975) and a few others championed the importance of ancient landscapes. Probably more important in the initial reinvigoration of long-term landscape evolution, however, was the fact that, as W. M. Davis was being supplanted by a systems approach and the quantitative revolution in geomorphology, another revolution was proceeding apace, namely, plate tectonics.

Plate tectonics is an elegant system that integrates rock uplift, surface uplift and surface processes, and the importance of plate tectonics for the re-emergence of long-term landscape evolution cannot be over-emphasized. The links between plate convergence and the development of the high elevation parts of the Earth's surface were recognized very early in the plate tectonics revolution (e.g., Dewey and Bird, 1970), followed by the postulation of tectonic events and topographic development that accompany the lithospheric extension and rifting that can culminate in continental break-up and the formation of high-elevation passive continental margins (e.g., Falvey, 1974). The importance of plate tectonics to geomorphology led Ollier to comment in the early 1980s that 'to play a part in the problems of geology over the next few decades geomorphologists must forget their trivial catchments and see mega forests instead of trees' (Ollier, 1981, p. 300). Early papers attempting to link a cyclical approach to plate tectonics, such as in the 1975 Binghamton volume on *Theories of Landform Development* (Melhorn and Flemal, 1975), were of varying success. The attempt by Melhorn and Edgar (1975) to integrate world-wide correlation of erosion surfaces into the emerging plate tectonic framework was unsuccessful and highlighted a 'split personality' in the discipline as it attempted to integrate a Daviesian viewpoint into the plate tectonics framework. Judson's (1975) discussion of the evolution of Appalachian topography likewise finds some inspiration in Davis's work but three decades later seems more successful as it is more wholly set within a plate tectonic (passive margin) framework, as was Falvey's paper noted above. More enduring impact has come from a paper from the mid- to late 1970s, elegantly linking the locations of the world's large rivers to their plate tectonic settings (Potter, 1978). Potter (1978) observed that the world's 28 largest rivers drain to passive continental margins, which themselves have the 25 largest deltas on Earth. Large deltas would not be expected on a convergent margin because of the latter's narrow continental shelf and deep offshore trench to which sediment is lost.

The emphasis on passive continental margins in much of the early work exploring links between plate tectonics and topography (or between geomorphology and global tectonics, the title of the book edited by Summerfield (2000)) continues. This focus was widened in the 1990s with the beginning of work on geomorphological evolution in active (convergent) plate settings. Thus, many major conceptual breakthroughs have fruitfully exploited the framework, questions and concepts generated by plate tectonics. Other advances have built on important methodological and analytical developments over the last two to three decades, including cosmogenic isotope analysis and low-temperature thermochronology, enabling the direct determination, from rocks and sediments at the Earth's surface, of the rates of denudation of land surfaces (and, in restricted situations, of the ages of these land surfaces). In many ways, these latter breakthroughs represent a kind of geomorphological 'holy grail' because they enable us for the first time to determine the rates of erosion of the ground surfaces that make up topography.

This review paper examines the last several decades of research in long-term landscape evolution. I build the review around seven major themes that can be discerned in this research. These themes are: post-mid-1950s theoretical insights outside plate tectonics; a more coherent analysis incorporating plate tectonics; rock uplift, surface processes, surface uplift and isostasy; new techniques; passive continental margins; orogenic settings; and steady-state landscapes and post-orogenic settings. The review is brought together by a forward look to the emerging research challenges.

Post-mid-1950s Theoretical Insights outside Plate Tectonics

Equilibrium models for long-term landscape evolution

The importance of plate tectonics in the re-emergence of research in long-term landscape evolution over the last decade or two is explored in more detail below. Before this exploration, however, I examine a range of other insights from the last quarter of the 20th century. Probably the most important of these major new theoretical insights came from John Hack, in two broad conceptual areas. Hack's earliest work, notably his 1957 paper on bedrock river long profiles, is still widely referenced, remaining highly influential in current bedrock river research. As in all of Hack's work, there is a strong empirical element in this bedrock river research, in which he explored relationships between relatively simple fluvial morphometric variables, such as channel width, gradient, mean grain size, downstream distance and so on, at about the same time that Leopold and colleagues were exploring these issues in alluvial systems (e.g., Leopold *et al.*, 1964). Hack's 1950s empirical research on bedrock rivers highlighted many of the relationships that still lie at the heart of bedrock long profile research (e.g., Willgoose, 1994; Whipple and Tucker, 1999, to highlight but a couple of the many papers in this burgeoning research area). In the case of the more recent work, the morphometric relationships have generally been derived from theory, and Hack's empirical results have been used to provide parameter values for these theoretically derived relationships. This is not to diminish Hack's contributions in these areas, because he interrogated his empirical data in ways that foreshadowed many current analyses (e.g. slope-area plots in bedrock river long profile analyses: Willgoose, 1994; Whipple and Tucker, 1999; Stock and Dietrich, 2003). Hack *did* ultimately rely on simplified statements of the relationships he had identified in his 1957 paper, however, and subsequent work seems to have lost some of the complexity and subtlety of the 1957 paper (e.g., Hack's influential 1973 paper; see the fuller exploration of these issues by Goldrick and Bishop, 2007).

One of Hack's key findings is that bedrock rivers on more resistant lithologies exhibit steeper gradients, which he argued was because these lithologies generate coarser sediment requiring steeper gradients to transport these clasts. This was an equilibrium relationship between lithological resistance and gradient, which had the logical and essentially inescapable implication that for a given lithology and other factors, such as climate and tectonic setting, remaining unchanged, river gradient must be essentially constant and cannot decline with the passage of time, as Davis (1899) had argued. It was then a relatively small step to argue that landscapes attained dynamic equilibrium and that, also contrary to what Davis (1899) proposed, landscapes did not evolve through the geographical cycle's sequence of time-dependent morphological stages of youthful, mature and old age forms. Hack's major statement on dynamic equilibrium in landscape evolution is in his 1960 paper, and it was reiterated 15 years later (Hack, 1975).

The notions of dynamic equilibrium and steady-state landscapes remain prominent in the bedrock rivers and landscape evolution literature. Adams (1985) applied it to the changes in styles of landform evolution as the Pacific Plate is progressively uplifted as it is carried towards the Alpine Fault that marks the zone of plate convergence in the Southern Alps (New Zealand). He argued that two broad zones can be identified on the Pacific Plate on the South Island. In the east, a zone of uplifted ancient erosion surfaces is characterized by time-dependent landforms, whereas once a critical surface elevation is reached as landscapes are uplifted as they are carried westwards towards the Alpine Fault and the equilibrium slope angles intersect in 'spiky' peaks, time-independent landscapes emerge with constant slope angles. Within this zone of more rapid uplift close to the Alpine Fault, these spiky mountains have the same time-independent morphology irrespective of their distance from the fault and the zone of maximum uplift (Adams, 1985). Burbank *et al.* (1996) have shown how slope morphology and process is closely adjusted to lithology in very rapidly uplifting areas in the Himalayas, but the adjustment is to the strength of the rock rather to a more conventional measure of erodibility (such as the strength or erodibility of the regolith, as Hack (1960) and Adams (1985) had argued).

Hovius (2000) has disagreed with the detail of Adams' (1985) interpretation but the important point here is that the notion of equilibrium landscapes is once again a major research issue. Indeed, the notion of steady-state landscapes underpins a very large majority of recent research in long-term landscape evolution that uses numerical models (see below) and physical models (e.g., Bonnet and Crave, 2003). Testing any models for long-term landscape evolution, under conditions of dynamic

equilibrium (or steady state) or not, is a major challenge. Numerical and physical models are highly dependent on their structure and parameterization (e.g., Beven, 1996), and contingency complicates field-based model testing at landscape evolution timescales (e.g., Church, 1996; Hoey *et al.*, 2003). The pitfalls associated with testing a model such as Hack's dynamic equilibrium model are highlighted by attempts to test it in even well studied settings such as the SE Australian highlands where rates of landscape evolution and accompanying morphological evolution can be reconstructed with confidence. Bishop *et al.* (1985) argued that catchment evolution under conditions of Hack-type dynamic equilibrium was demonstrated by (i) the gross equivalence between Neogene rates of river incision and Neogene rates of sediment flux to basins (i.e., catchment-wide downwasting) and (ii) the apparently close adjustment between long profile gradient and lithology (i.e. steeper bedrock channels associated with resistant lithologies, as Hack had proposed is also characteristic of dynamic equilibrium). Subsequent analysis has shown, however, that the steeper reaches on more resistant lithologies are disequilibrium reaches (knickpoints) caught up on these lithologies, highlighting the difficulties (Bishop and Goldrick, 2000; Goldrick and Bishop, 2007).

Non-equilibrium models for long-term landscape evolution

Other landscape evolution systems proposed in the last quarter century as alternatives to both Davis and Hack include Crickmay's (1975) hypothesis of unequal activity, in which it is posited that fluvial erosional energy (and, for that matter, but for different reasons, glacial erosional energy in some situations - Sugden, 1968) becomes progressively concentrated in valley bottoms, whereas lowering of divides and upstream areas slows because of lower erosion rates in stream headwaters. River channels continue to incise and become de-coupled from slopes, the evolution of which slows considerably. This situation results in landscape evolution with increasing relief amplitude (Crickmay, 1975; Twidale, 1976, 1991). Equilibrium landscapes do not develop in this landscape evolution system: the landscape consists of a complex mosaic of elements evolving at their own individual rates and along their own individual trajectories.

Representative rates of long-term landscape evolution

Young and McDougall (1993) emphasized complex pathways of landscape evolution, in particular highlighting the persistence of disequilibria, such as knickpoints and over-steepened slopes, throughout tens of millions of years of landscape evolution. They also highlighted very slow rates of landscape evolution. A decade earlier, Young (1983) had pointed out that the denudation rates reported by Thornbury (1969) and almost claimed as being 'normal' were not representative of the full spectrum of natural rates operating on the Earth's surface. Indeed, Young (1983) argued that analyses such as Thornbury's were too reliant on (Northern Hemisphere) settings that had experienced Pleistocene glaciation, the impacts of which are likely still being felt as glaciated landscapes continue to adjust to interglacial conditions, or, as argued above, were anthropogenically disturbed settings (or both). Thus, denudation rates from glaciated terrains are likely to represent the high end of the spectrum of possible rates and non-glaciated, tectonically quiescent, undisturbed settings would be expected to evolve at much lower rates (Young, 1983). Representative rates of landscape evolution in such tectonically stable, non-glaciated settings are of the order of 5-10 m Ma⁻¹ (e.g., Bishop 1985; Bishop and Goldrick, 2000; Bierman and Caffee, 2001, 2002; Cockburn *et al.*, 2000; Matmon *et al.*, 2002; Reusser *et al.*, 2006). Even these relatively low rates are one to two orders of magnitude more rapid than rates in the tectonically stable, hyper-arid Dry Valleys of Antarctica, from where Nishiizumi *et al.* (1991) and Summerfield *et al.* (1999a, 1999b) have reported rates as low as 0-14 m Ma⁻¹.

These extremely low rates prompt, in passing, the following question: how low can rates of landscape evolution be? It seems unlikely that it is possible for landscapes to have survived unaltered at the Earth's surface for the lengths of time (500 Myr) that have been argued by Stewart *et al.* (1986) and Ollier *et al.* (1988) for landforms that they claim have been exposed at the Earth's surface since the Cambrian. Even at the vanishingly low rate of denudation of 0-14 m Ma⁻¹ reported from arid Antarctica, a land surface would still have been lowered 70 m in the roughly 500 Ma since the Cambrian and much more than that at more usual erosion rates. Indeed, Belton *et al.* (2004) used fission track analysis and cosmogenic isotope analysis to show that the Cambrian floodplain reported by Stewart *et al.* (1986) must have been buried prior to and during the Mesozoic, to be then exhumed

in a phase of kilometre-scale exhumation that was largely complete by the beginning of the Cenozoic. Given what we now know about surface erosion rates, it is more reasonable to invoke such exhumation of ancient unconformities and land surfaces to explain apparently extremely ancient landscapes, rather than by hypothesizing extremely low denudation rates that have never been shown to exist. Twidale (1998) has provided many examples from the large literature on exhumed unconformities (e.g. exhumed Precambrian, sub-Cambrian and sub-Mesozoic surfaces in Sweden (Lidmar-Bergstrom, 1987, 1996), the exhumed sub-Torridonian surface of the Lewisian Gneiss in NW Scotland (Sissons, 1967, p. 9) and the exhumed Cretaceous valleys of northern Australia (Nott, 1995)).

An obvious implication of rates of landscape evolution that are lower than had previously been thought likely is that the timescales of landscape evolution must be extended, whatever the precise sequence of landforms - declining slopes and relief, parallel slope retreat or increasing slopes and relief etc - during this evolution. Ollier (1979) took this viewpoint to a logical conclusion with his notion of evolutionary geomorphology, arguing that in less humid areas with low rates of tectonic activity, cyclical or equilibrium approaches are inappropriate and that landscape evolution should be considered much more as a directional process (i.e. as 'evolutionary'). Thus, it is argued that the starting point for landscape evolution in areas that experienced Quaternary glaciation is this glaciation. This approach may not be applicable to all glaciated landscapes, such as when glaciation does not create the landscape *ab initio* (for example, in Scotland, where the consensus (e.g., Sissons, 1967) is that Quaternary glaciation mostly modified a pre-existing fluvial landscape), and it is most certainly inappropriate for areas outside the Quaternary glaciated area. In many of these latter areas, such as southern Africa and southeastern Australia, the 'glaciation' starting point for landscape evolution is likely to be the Permian glaciation, 250 million years ago. As landscape evolution extends back into deeper geological time, the controls on geomorphological processes change, and even such relatively simple issues as the fact that the grasses emerged only in the Cretaceous have profound implications for pre-Cretaceous geomorphological processes.

At a more general level, it is striking how the various insights summarized in this section and leading to the extending of the timescales of landscape evolution have strong associations with the ancient landscapes and researchers of the Gondwana continents, including southern African, Antarctica and Australia. We return to this point below, but we can foreshadow here one of our main conclusions, namely, the need to reconcile landscape evolution over different timescales and in different tectonic settings.

A More Coherent Analysis Incorporating Plate Tectonics

The preceding section distils, and attempts to draw common threads from, a disparate range of long-term landscape evolution literature from the last quarter of the 20th century. It is probably fair to judge, however, that this literature lacks sufficient coherence to justify a claim that it constitutes the main reason for the full re-emergence of landscape evolution in the mainstream geological and geomorphological literature. This full re-emergence of long-term landscape evolution in mainstream research - that is, with a coherent research agenda, extending beyond both the fairly straightforward and generalized linking of plate tectonics and macrogeomorphology/long-term landscape evolution, and the disconnected, post-1950s theoretical insights from outside plate tectonics outlined above - can be argued to reflect four important and inter-related factors:

1. the fundamental (re-)realization that landscape evolution reflected the balance between surface processes and rock mass flux and therefore that landscape evolution could be used to test plate tectonic models, with a corresponding need for greater understanding of surface processes as they operate on essentially bedrock landscapes over long time periods (Figure 1);
2. the development of a suite of techniques to measure rock mass flux over a range of temporal and spatial scales (especially low-temperature thermo chronology and cosmogenic isotope analysis);
3. the realization that there may be a two-way linkage between tectonics/surface uplift and climate and
4. the ongoing increase in computing power contemporaneously with the formulation of numerical models of long-term landscape evolution, and the consequent capability to explore tectonics and long-term landscape evolution. Before addressing recent research in long-term

landscape evolution, I briefly discuss some important concepts and methods that underpin this work.

Rock Uplift, Surface Processes, Surface Uplift and Isostasy

Earth surface morphology reflects the interaction between Earth surface geomorphological processes and the rock, regolith and soil at the Earth's surface. Earth surface processes generally act to decrease the elevation of the Earth's surface (the distance of the Earth's surface above some reference plane, such as sea level or the geoid or the centre of the Earth), and surface uplift acts to increase surface elevation. The distinction between surface uplift and rock uplift (also termed 'crustal uplift') is an important one that has not always been made clearly (England and Molnar, 1990; Summerfield, 1991). Rock uplift is the upward movement of the rock column with respect to some datum, arising from the vertical mass transfer of crust (tectonics). Rock uplift equates to surface uplift only if there is no denudation during the rock uplift. If rock uplift is matched by denudation, there is no surface uplift (the classic steady-state landscape alluded to above), and if rock uplift is less than denudation, the surface elevation is lowered. Fundamental to considerations of rock uplift and the generation and maintenance of topography is the concept of isostasy. This is the notion (from the Greek for 'equal standing') that the rigid (or semi-rigid) lithosphere (the crust and upper mantle) float on the underlying, deformable (or plastic) part of the mantle, the asthenosphere; Summerfield (1991) and Burbank and Anderson (2001) provide introductions to isostasy. If the lithosphere is in isostatic equilibrium (i.e. free to float on the underlying asthenosphere and neither held down nor supported by other forces), the 'height' at which the lithosphere 'floats' depends on its thickness and its density. Thus, oceanic lithosphere is thinner and denser than continental lithosphere and so 'floats' at the lower elevations that characterize the ocean

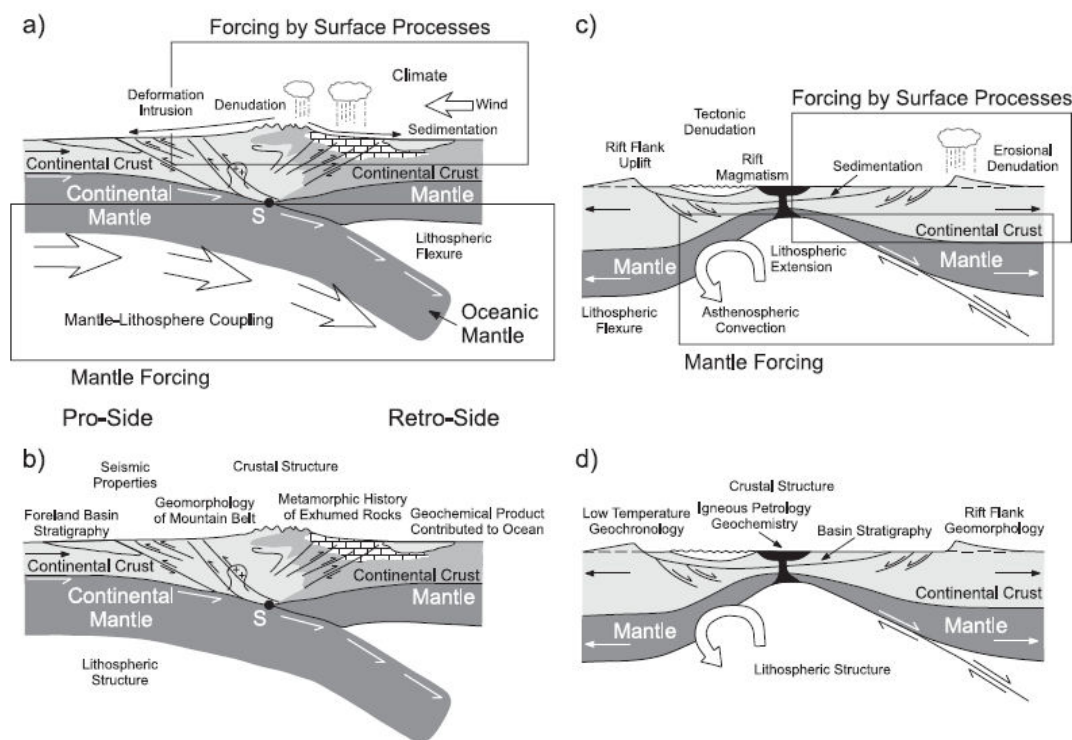


Figure 1. The processes driving landscape evolution in (a) convergent settings and (c) extensional (rifting) settings. (b) and (d) highlight the various observations that can be used to test models of the structure and evolution of the two tectonic settings. Note that geomorphology is an important observational element in both settings (Beaumont *et al.* (2000)

One of the key issues in understanding the long-term landscape evolution of high-elevation continental areas, and particularly the persistence of such areas, is generating the lithospheric thickening that underpins or supports such high elevations. In effect, the issue is generating the horizontal and vertical fluxes of rock required to produce high elevations and to sustain them, and the

elegance of plate tectonics in supplying these mass fluxes of lithosphere is one of the fundamental reasons that plate tectonics re-vivified long-term landscape evolution studies. It is also important to note that the lithosphere has strength and will bend and flex when a load is applied, including when a negative load is applied as a result of denudation. This flexure of the lithosphere in response to a load is termed flexural isostasy, and the strength of the lithosphere (or its flexural rigidity) determines the lateral extent (or horizontal distance) over which the loading has an effect.

Denudational isostatic rebound, the notion that the lithosphere floats up in response to denudational unloading, is generally a central part of all modern interpretations of long-term landscape evolution. For areas in isostatic equilibrium, denudational isostasy means that, for every metre that is removed from the Earth's surface by denudation, the Earth's surface is uplifted by about 0.8 m (0.8 being the approximate ratio of the densities of the crust and sub-lithospheric mantle). The importance of such denudational isostasy in maintaining relief had been realized by the 1830s, and the notion was applied sporadically throughout the 20th century to suggest that mountain peaks may be uplifted isostatically as a result of the unloading caused by the incision by major rivers flowing between the mountain peaks (e.g., Jeffreys, 1931; Wager, 1937; for more detail see Bishop, 2007). King (1955) and Pugh (1955) argued that the crustal unloading associated with escarpment retreat on continental margins such as southern Africa would likewise lead to isostatic uplift of the margin. Molnar and England (1990) revived the insight of Jeffreys (1931) and Wager (1937) and suggested that regional isostatic response to valley incision may lead to mountain peak uplift. They took this insight much further, however, by then arguing that climate change can drive tectonics and surface uplift. In summary, the argument states that global cooling leads to an increase in glacial erosion, resulting in isostatic uplift of peaks. This uplift results in further increases in glacial erosion and isostatic uplift of peaks, further glacial erosion, and so on. Raymo and Ruddiman (1992) proposed the opposite causal linkages in this 'chicken or egg' question as to whether climate can drive (or 'lead') tectonics. They proposed that mountain peak uplift leads to an increase in glaciation and glacial erosion, liberating large quantities of very fine-grained glacial sediment, the weathering of which leads to CO₂ drawdown, global cooling, increased glacial erosion, and so on.

The hypothesis by Molnar and England (1990) seems to face several fundamental problems. Numerical and physical models suggest that a change to a more erosive climate leads to a reduction of relief (not an increase) (Whipple *et al.*, 1999; Bonnet and Crave, 2003). Brozovic *et al.* (1997) have likewise argued that a change to a more erosive glacial climate leads to a reduction in relief, and numerical modelling to assess the notion of mountain peak uplift as a result of localized valley incision points to limits to this effect (Montgomery, 1994; Gilchrist *et al.*, 1994b). Gilchrist *et al.* (1994b), for example, showed that it seems unlikely that the volumes of the valleys in such major mountain belts as the Alps are sufficient to generate the mountain peak altitudes envisaged by Molnar and England (1990) (cf. Whipple *et al.*, 1999).

These limitations notwithstanding, the (re-)recognition of denudational isostasy and flexural isostasy has had important implications for understanding long-term landscape evolution. At the most general level, as we have already seen, denudational isostasy means that surface uplift may not necessarily be tectonically induced, but may be the regional isostatic response to localized denudation (Small and Anderson, 1995; Pelletier, 2004). Climate and erosion can be drivers of mountain peak uplift; moreover, two-way links between tectonics and topography are reasonable and expected (see below). Second, in situations where denudation is more regional and not restricted to rapidly incising valleys, denudational isostasy significantly increases the longevity of mountain belt topography in its post-orogenic (tectonically quiescent) phase (Baldwin *et al.*, 2003; Matmon *et al.*, 2003a, 2003b). Thus, in situations of catchment-wide downwasting, where denudation has maximum spatial extent, denudational unloading, and hence denudational isostasy, are also maximized. Maximizing of denudational isostasy means that the surface is lowered at only at about 20% of the denudation rate, and this effect occurs even at the low rates of denudation that characterize tectonically stable intra-plate settings (Bishop and Brown, 1992). In short, the fundamental principle of isostasy means that denudational rebound must follow as a matter of course, thereby providing an ongoing source of potential energy for the surface processes of landscape evolution. Third, denudational rebound is achieved by passive rock uplift (unloading), meaning that deeper levels of the crust are progressively brought to the surface over time by this rock uplift, even though surface elevation is declining (albeit more slowly than the rate of denudation). Research in long-term landscape evolution has been

increasingly relying on techniques, such as low-temperature thermochronology, that are used to track rocks as they move from depth to the surface by denudation. We now turn to brief reviews of these and other important recent techniques.

New Techniques

A suite of exciting new techniques has revolutionized research in long-term landscape evolution. Three of these are now reviewed briefly: low-temperature thermochronology; the analysis of *in situ* terrestrial cosmogenic nuclides (cosmogenic isotope analysis) and numerical modelling of landscape evolution.

Low-temperature thermo chronology

Thermo chronology is the investigation of the cooling history of a mineral as the mineral's host rock moves from depth in the crust to the surface. In effect, the technique provides the mineral's time-temperature history corresponding to its trajectory to the surface via denudation. If the geothermal gradient is known (or can be reconstructed), thermo chronology provides data on denudation rates and the amount of rock section removed by denudation. A brief overview of low-temperature thermo chronology follows; fuller reviews have been provided by Gallagher *et al.* (1998), Gleadow and Brown (2000), Ehlers and Farley (2003) and Reiners and Ehlers (2005).

Thermo chronometers are based on absolute radiometric dating systems, which measure in a mineral the amount of a daughter product produced by radioactive decay, relative to the amount of the parent element. These absolute dating systems become thermo chronometers when the temperature at which the daughter product is retained in the mineral is known (Figure 2). This temperature - the closure temperature - is converted into a depth in the crust via the geothermal

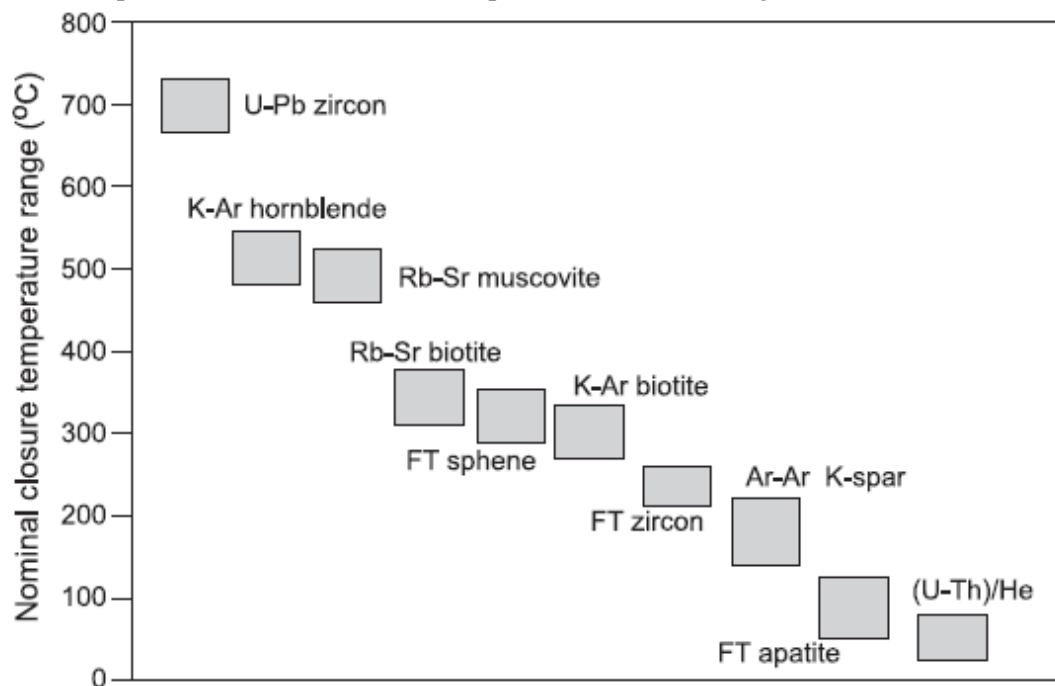


Figure 2. Nominal closure temperatures of various thermo chronometers. (Image from Roderick Brown.)

Gradient. With typical geothermal gradients of $25\text{-}30\text{ }^{\circ}\text{C km}^{-1}$, these closure temperatures correspond to denudation from crustal depths of kilometres. These depths have probably not typically been considered 'relevant' to geomorphology. However, in tectonically active zones, for example, where rock uplift, driven by both active tectonic processes (crustal thickening) and passive denudational rebound attendant on very high rates of denudation, is extremely rapid, higher-temperature thermo chronometers will return very young 'ages', which are similar to the ages obtained from lower-temperature thermo chronometers, reflecting that fact that rocks have travelled extremely rapidly through the closure temperatures of the various thermo chronometers on their trajectory to the surface (e.g., Zeitler *et al.*, 1993, 2001).

The main thermo chronometers used in long-term landscape evolution are apatite fission track thermo chronology (AFTT) (Gallagher *et al.*, 1998; Gleadow and Brown, 2000; Reiners and Ehlers, 2005) and apatite (U-Th)/He analysis (apatite-He) (Ehlers and Farley, 2003; Reiners and Ehlers, 2005). The simple concept of an 'on-off' closure temperature is, in fact, inaccurate and closure should be thought of as a temperature zone - a partial annealing zone in the case of AFTT (i.e. a temperature zone that is warm enough to anneal or repair the minute damage tracks in the apatite that are produced by fission of ^{238}U) and a partial retention zone in the case of apatite-He (i.e. a temperature zone that is warm enough to cause some of the ^4He produced by decay of U and Th to be lost from the apatite grain by diffusion). This partial annealing zone in AFTT is $\sim 120\text{-}60^\circ\text{C}$ ('FT apatite' in Figure 1), and the partial retention zone for apatite-He is $80\text{-}40^\circ\text{C}$ ('(U-Th)/He' in Figure 1). AFTT data include track lengths as well as the number of tracks (the latter being used to calculate AFTT age). Monte Carlo simulation of AFTT age and track length distribution provides a family of possible time-temperature histories of the mineral's trajectory to the surface as a result of denudation (Gallagher, 1995; Ketcham and Donelick, 2003; Gallagher *et al.*, 2005). These AFTT-based time-temperature histories are then interrogated to select those that reproduce apatite-He ages (Balestrieri *et al.*, 2005; Persano *et al.*, 2005).

Low-temperature thermo chronology has many applications in understanding long-term landscape evolution, both in tectonically active settings and more stable settings, both of which are considered below. In tectonically active settings, it is noteworthy that even the much higher-temperature thermo chronometers in Figure 1 have application in understanding landscape evolution: the same thermo chronological ages for higher and lower temperature systems means very rapid rock uplift and denudation (e.g., Zeitler *et al.*, 1993, 2001). The low-temperature systems can also be used to date the longevity of major, long-wavelength topography. This is done by exploiting the way in which long-wavelength topography perturbs the thermal structure of the shallow crust, deforming the shallow isotherms (Figure 3). House *et al.* (1998), for example, used this approach to conclude that the major valleys and ridges of the Californian Sierra Nevada mountain front date from at least the Cretaceous. A cautionary note in such an application, and indeed in the application of low-temperature thermo chronology in general, has been sounded by Dempster and Persano (2006), in respect of whether the geothermal gradient in the shallow crust is linear, as is commonly assumed. Mitchell and Reiners (2003) have sounded a further cautionary note in reporting that bushfires ('wildfires') may cause helium loss in surface samples.

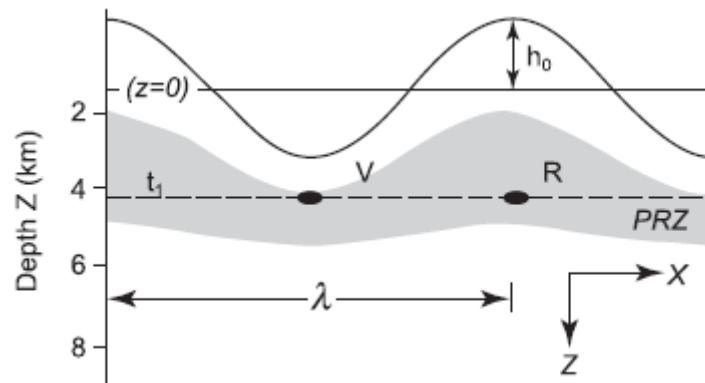


Figure 3. Illustration of the principle of using low-temperature thermo chronology to date the antiquity of topography (from House *et al.*, 1998, Figure I). The crust's shallow thermal structure is perturbed by long-wavelength topography and so the apatite-He partial retention zone (PRZ) sub-parallel the surface topography. As the surface is lowered by denudation, samples V ('valley') and R ('ridge') come to crop out where they may be sampled. In the situation illustrated here, sample V will have an older apatite-He age than sample R (because sample V cooled below the PRZ temperature earlier than did sample R). This outcome demonstrates that the topography was already in existence at the age of sample V. If the long-wavelength topography post-dated the passage of V and R through the partial retention zone, then V and R will have the same apatite-He ages.

Analysis of *in situ* terrestrial cosmogenic nuclides (cosmogenic isotope Analysis)

Another development that is revolutionizing the study of long-term landscape evolution is the analysis of terrestrial cosmogenic nuclides (TCNs), springing from the work of Lal and Peters (1967) and Lal (1988, 1991), following the seminal paper of Schaeffer and Davis (1956) and the first substantial data sets of Kurz (1986). The technique, also known as cosmogenic isotope analysis, is introduced only briefly here; more information is given in reviews by Bierman (1994), Cerling and Craig (1994), Bierman and Nichols (2004), Gosse and Phillips (2001), Niedermann (2002) and Cockburn and Summerfield (2004). Watchman and Twidale (2002) provided cautionary notes on methodological issues associated with the use of cosmogenic isotope analysis for the absolute age dating of surfaces and the need for thoughtful, carefully structured and rigorous sampling programmes in such attempts at dating. The reservations of Watchman and Twidale (2002) actually highlight the difficulties in being confident that a surface has not eroded and that the cosmogenic isotope concentrations in the rock surface can be interpreted as an exposure age of the rock (i.e. as 'dating' the rock surface).

The essence of the analysis of TCNs in landscape evolution studies is that cosmogenic radiation from outer space bombards rocks and sediments at, and within the upper few metres of, the Earth's surface. This radiation interacts with elements in minerals in the rocks and sediments to produce nuclides that are either stable (e.g. ^3He and ^{21}Ne) or unstable (subject to radioactive decay: e.g. ^{10}Be , ^{14}C , ^{26}Al and ^{36}Cl). These nuclides accumulate in the minerals at rates that can be calculated and estimated empirically (e.g., Nishiizumi *et al.*, 1989; Lal, 1991) and the concentration of a particular isotope in a sample can therefore be interpreted as an exposure age for the sample if (and only if) the sampled surface has not eroded since initial exposure. Alternatively, an isotope concentration can be interpreted in terms of an erosion rate of the sampled surface if the sampled surface is undergoing steady and uniform erosion. In many cases, field observations cannot strictly determine the exposure and erosion histories of a rock surface and thus limiting model exposure ages and erosion rates are often calculated and interpreted. An important breakthrough has been the application of the technique to sediments at a catchment outlet to determine catchment-wide erosion rates (e.g., Bierman and Steig, 1996; Brown *et al.*, 1995; Granger *et al.*, 1996; Bierman *et al.*, 2001). In special cases, application of multiple isotopes (either two unstables with different half-lives or one stable and one unstable) enables the identification of complex exposure histories, including burial, although the utility of this approach is often limited by measurement precision (Gillespie and Bierman, 1995). Isotopic concentrations are measured by conventional noble gas mass spectrometry in the case of stable isotopes, and accelerator mass spectrometry in the case of the radio-isotopes (e.g., Gosse and Phillips, 2001).

Numerical modelling of landscape evolution

Numerical modelling of landscape evolution is the development and implementation of computer-based models that attempt to represent and to 'marry' tectonic processes of rock mass transfer (including isostasy) and surface geomorpho-logical processes, in order to simulate and visualize landscape evolution over millions to tens of millions and hundreds of millions of years. Recent commentary and reviews of numerical modelling of landscape evolution by Oreskes *et al.* (1994), Beaumont *et al.* (2000), Coulthard (2001), Willgoose (2005) and Codilean *et al.* (2006a) treat in more detail the points summarized here. The first of these reviews addresses methodological issues, the second pays attention to both tectonic models (TMs) and surface process models (SPMs), and the remainder focus more on the SPMs. Two of the earliest papers attempting to couple TMs and SPMs were those by Koons (1989) and Beaumont *et al.* (1992). The SPM components have all built on the pioneering work of Howard and Kerby (1983), Howard (1994) and Kirkby (1986).

It is not possible, with our current understandings of the physics of tectonics and surface processes, to implement full, physical-law-based formulations of tectonics and landscape development processes (including weathering, runoff, sediment entrainment and transport, and so on). Rather, numerical simulation has generally used a simplifying approach in formulating and implementing the process laws that underpin the models, thereby adopting the recommendation by Kooi and Beaumont (1994) that numerical modelling of tectonic and surface processes be based on 'simple relationships cast in terms of large-scale, long-timescale quantities' (p. 12 192); the reviews referenced above provide more details on this matter. The surface processes represented in a numerical model commonly consist of bedrock weathering, slow diffusive hillslope processes, rapid

hillslope processes (landsliding), bedrock channel incision and fluvial sediment transport and deposition; the algorithms used to represent these processes are generally implemented in one or two dimensions (i.e. at a point or in plan view). The tectonic models represent mass flux of crust in three dimensions and it is not always possible easily to couple the planform operation of an SPM to a two-dimensional (vertical section) tectonic (geodynamic) model (Beaumont *et al.*, 2000).

These simplifications demand special awareness of issues of temporal and spatial resolution in the model results, not least because the grid resolution of a large SPM (typically 1 and 5 km for, say, a high-elevation passive continental margin) may mean that the physical features it represents, such as rivers, valleys and landslides, are unrealistic in size. Also, the temporal resolution (time step) commonly used in SPMs means that model time steps are much longer, and generally seek to encompass (and therefore represent) much larger 'events', than those for which there are field data to generate parameters. For example, each time step in the *Cascade* model notionally represents 100 years of landscape evolution (Braun and Sambridge, 1997). In other words, SPM parameters often cannot be derived from - or calibrated by - field measurements and model parameterization may serve to ensure that the model produces 'realistic' output, with the parameters perhaps having no direct physical significance outside the model context (van der Beek and Braun, 1998).

The principal uses of numerical modelling of long-term landscape evolution are to test generic conceptual models of landscape evolution (e.g., the modelling by Kooi and Beaumont (1996) of classical models of landscape evolution) and/or to model the landform evolution and denudation history of specific regions (e.g., the modelling by van der Beek *et al.* (1999) of long-term landscape evolution of the SE Australian passive continental margin). The first of these uses includes the heuristic use of numerical models for sensitivity analysis and the asking of 'what if' questions (Merritts and Ellis, 1994; Oreskes *et al.*, 1994; for more detail see Codilean *et al.*, 2006a, and Hoey *et al.*, 2003). Identifying the appropriate tests or evaluation of model outputs is a non-trivial issue. These various issues notwithstanding, numerical models have been at the forefront of the last decade's re-energization of research in long-term landscape evolution, building on the stimulus to this research that was provided by the plate tectonic revolution. Numerical models have figured prominently in investigations of the geomorphic evolution of two major plate tectonic settings, namely high-elevation passive continental margins and convergent orogens. We now turn to these.

Passive Continental Margins

Passive continental margins are the continents' trailing edges that result from continental rifting, breakup and sea-floor spreading. These are the classic 'Atlantic' type coasts first recognized by Suess in the 1880s (and published in English translation in the early 20th century - Suess, 1904, 1906; see also Gregory, 1913). They are 'passive' because their formation is not associated with active convergent tectonics. Rather, they are the outcome of a combination of the extensional and rifting processes that precede continental breakup and the post-breakup tectonic processes (see the introductory summary by Summerfield, 1991). They have been a focus of much research since the plate tectonic revolution because they are major, first-order topographic features of the Earth's surface that can be clearly related to plate tectonic processes and setting, as was recognized by Wegener (1929) in the early 20th century and reiterated by Du Toit (1937) (Bishop, 2007). Second, the major issues surrounding the formation and evolution of high elevation of passive continental margins relate to the spatial and temporal distribution of denudation and sediment flux, both of which are issues fundamental to the evolution of major hydrocarbon fields on passive continental margins (e.g. the UK North Sea oil and gas fields, the western African margin oil field and the northwestern Australian gas field).

Passive continental margins (PCMs) may be low elevation (e.g. eastern India, the central and northern stretches of western Africa, southern Australia) or high elevation (e.g. southern Africa, western India, southeastern Australia) (Figure 4). Some margins vary along their length between high elevation and low elevation (e.g. the opposing ['conjugate'] Atlantic margins of South America and Africa, and the southern and southeastern margins of Australia -Figure 4). The classic morphology of high-elevation PCMs consists of a coastal plain of varying width, backed by a steep, often wall-like escarpment and a low-relief plateau surface inland of the escarpment lip. This morphology has long

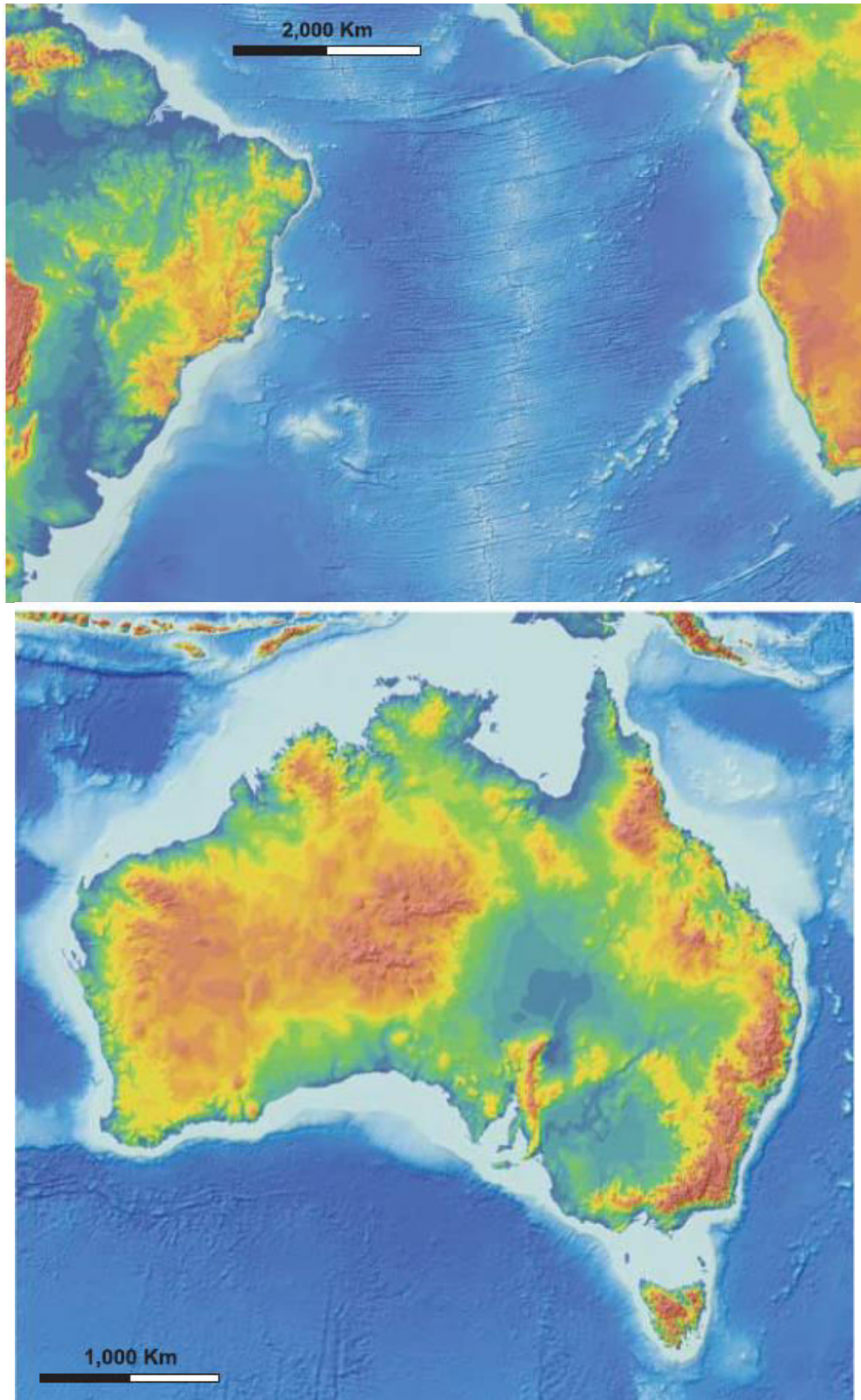


Figure 4. (a) DEMs of eastern South America and central and southern Africa. Note how elevations vary along the Atlantic margins of both continents. Note also that, if South America and Africa are rejoined by closing the South Atlantic Ocean, the conjugate (opposing) margins are not necessarily mirror images of each other in having similarly high or low elevations; Brown *et al.* (2000a) discussed these morphologies in more detail. (b) DEM of Australia. Note the low elevation of the southern margin (conjugate with Antarctica) and the high elevation of the eastern margin (conjugate with Lord Howe Rise in the southeast). Note also the contrasting widths of the continental shelves of these two margins.

been recognized as a major morphological feature of continental margins (e.g., King, 1955, 1962; Ollier, 1985a) and was important in the emergence of plate-tectonic-based interpretations of large-scale geomorphology (e.g., Ollier, 1982, 1985b; Summerfield, 1985, 1991) (Figure 5). The escarpment has traditionally been interpreted as having retreated steadily landward into the upland plateau surface since breakup (e.g., Ollier, 1982, 1985a); more recent views are discussed below. The continental drainage divide on high-elevation PCMs either coincides with the escarpment lip or, less commonly, lies 'inboard' (landward) of the escarpment lip, in which case the seaward-flowing rivers flow across the upland plateau surface inboard of the escarpment lip before plunging across the escarpment in deep gorges that are eroding back into the highland mass and crossing the coastal plain.

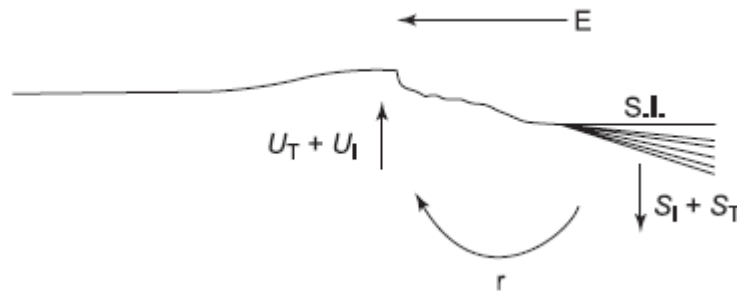


Figure 5. Summary of some of the major tectonic factors controlling the morphological evolution of rifted passive margins: U_T is thermally driven uplift; U_I is denudational unloading isostatic uplift; S_T is thermally driven subsidence; S_I is isostatic subsidence driven by sediment loading; r is 'rotation' of the margin driven by U_I and S_I ; E is escarpment retreat and S.I. is sea level (from Summerfield, 1991,

It is widely agreed that the extensional and rifting processes that precede breakup are associated with vertical tectonics of the rift shoulders (e.g., Buck, 1986, building on the important paper by McKenzie, 1978). One of the continuing challenges is to explain why the rift shoulder persists to form this classic high-elevation PCM morphology, long after the essentially transient thermal effects associated with extension and rifting have ceased and/or, once sea-floor spreading starts, the margin moves away from the high heat flows of the mid-ocean ridge spreading centre and thermal buoyancy effects on the rift shoulder lithosphere. If the lithosphere is not thickened to support the rift shoulder surface uplift, isostasy demands that the rift shoulders subside back to pre-extension elevations. The asymmetric rifting models of Lister *et al.* (1986) and Lister and Etheridge (1989) propose underplating (i.e. the permanent addition of material) at the base of the lithosphere, thereby thickening the lithosphere and providing the isostatic 'root' for ongoing support of the high elevation topography (Young, 1989; Matmon *et al.*, 2002). Support of the high elevation of some PCMs may also be provided by a rotational flexural effect of sediment loading on a wide continental shelf after breakup (Summerfield, 1991) (Figure 5). This effect is a function of the flexural moment of the continental shelf, depending on the shelf's width, the magnitude of sediment flux and consequent shelf loading, and the flexural rigidity of the lithosphere (its 'effective elastic thickness'). Pazzaglia and Gardner (1994, 2000) have argued that such a mechanism can explain a pulse of offshore Miocene sedimentation on the Atlantic margin, corresponding to between 35 and 130 m of Neogene rock uplift in the Appalachians; a somewhat similar mechanism has been proposed for the Western Ghats in India (Gunnell and Fleitout, 2000). A lower impact of the offshore loading effect is demonstrated in the case of post-15 Ma sediment loading in the Amazon delta. The maximum thickness of post-15 Ma sediment in the Amazon delta is about 5-3 km. This load has depressed the crust by a maximum of about 2-3 km and generated a maximum of 40 m of flexural uplift inland of the mouth of the Amazon, perhaps influencing drainage patterns in this area (Driscoll and Karner, 1994). In short, the impact of such offshore sediment loading in terms of flexural uplift of the margin's hinterland is of the order of tens to hundreds of metres of rock uplift (and surface uplift to the extent that this rock uplift is not matched by denudation). Likewise, flexural effects associated with denudational unloading as the coastal plain and the associated escarpment are formed may act generally to enhance or at least maintain the high elevation of the PCM highland belt (Figure 6) and specifically to generate the upwarp that is observed at some escarpment lips (Figure 7).



Figure 6. The passive margin highland escarpment at Point Lookout in northern New South Wales, highlighting the plateau surface and the steep fall at the escarpment that culminates in the coastal plain (out of picture to the right).

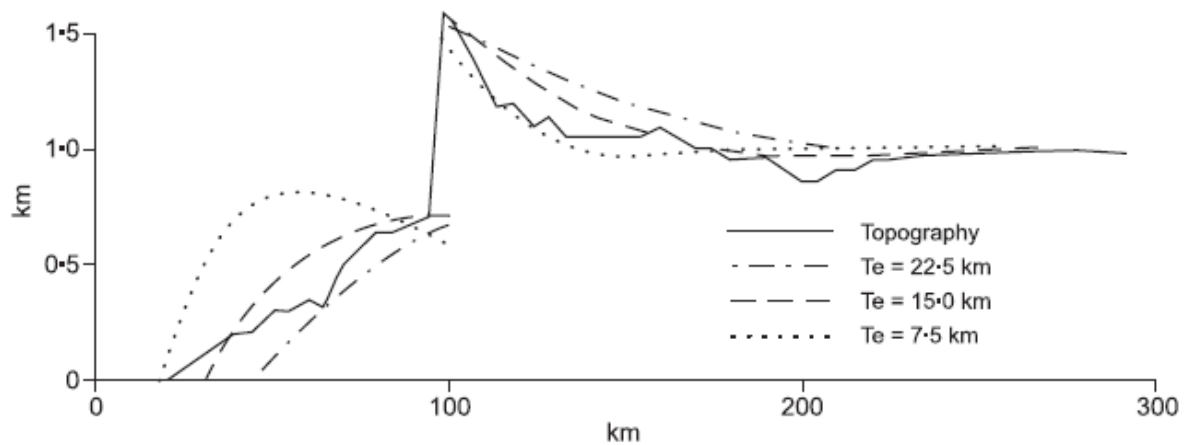


Figure 7. Results of numerical model showing the escarpment lip upwarp in SW Africa that is the flexural response of the plateau edge to the denudational unloading associated with the formation of the coastal plain by escarpment retreat or in-place excavation of the escarpment, seaward (to the left of) the escarpment (see text). The solid line represents the topography. The escarpment lip flexes up in response to the unloading on the adjacent coastal plain. The different model outcomes, represented by the variously dotted and dashed lines, correspond to model 'runs' employing different lithospheric flexural rigidities (different values of the effective elastic thickness, T_e). Higher values of T_e correspond to more rigid lithosphere and flexural effects over a correspondingly longer distance. Thus, the unloading associated with coastal plain formation has the longest-distance flexural effect on the plateau surface for the maximum T_e value (here 22.5 km). Low values of T_e (i.e. very flexible lithosphere) mean that the flexural effect is transmitted the shortest distance inland of the escarpment lip, and are also associated with a very pronounced updoming on the coastal plain.

The wall-like escarpments on high-elevation PCMs (shoulder-type margins in the terminology of Matmon *et al.*, 2002) may be classified according to whether they are simply etched into a highland belt that is associated with the original rifting (type-1 escarpment; e.g. southern Africa; southeastern Australia) or are etched into the seaward flank of a post-rift flexural bulge on a margin in which post-rift offshore sediment loading and onshore flexure dominate the post-breakup evolution of the margin (type-2 escarpment; e.g. the US Atlantic passive continental margin - Pazzaglia and Gardner, 2000). As noted above, the lips of such wall-like escarpments tend to be a drainage divide, either the main continental divide or a subsidiary but still major divide, in both cases maintained by flexural effects associated with denudational unloading associated with coastal plain formation (Gilchrist and Summerfield, 1990; Gunnell and Fleitout, 1998).

To complete the two-part typology by Pazzaglia and Gardner (2000) of escarpments, a type-3 escarpment may be identified, being the gorge-like, embayed escarpment associated with rivers that rise at a continental divide inboard of the escarpment lip and plunge across the escarpment before flowing across the coastal plain (the arch-type margins of Matmon *et al.*, 2002). These rivers are eroding back into the wall-like escarpment to produce its characteristic gorge-like, embayed morphology (e.g., Nott *et al.*, 1996; Weissel and Seidl, 1998). The apparent antiquity of the continental divide inboard of the escarpment lip (e.g., Bishop, 1986; Bishop and Goldrick, 2000) remains difficult to explain, but numerical modelling of PCM evolution has been helpful in this matter (see below).

Conceptual models for evolution of high-elevation PCMs

Two broad, general models have been proposed to explain the post-breakup evolution of high-elevation PCMs (Figure 8). The first, the down warping model, argues that the edge of the margin was tectonically lowered by down warping and/or faulting soon after breakup and that the escarpment retreated into this down warped/down-faulted plateau (Ollier and Pain, 1994, 1997; Seidl *et al.*, 1996) (Figure 8(a)). The second group of models does not accept post-breakup tectonic down warping of the margin, suggesting instead either that the escarpment has retreated across a coastal plain (Figure 8(b)) or that the escarpment has been excavated in place by down wearing (Figure 8(c)), with both of the latter accompanied by flexural isostatic rebound (Gallagher *et al.*, 1998). Greater thicknesses of crust are removed in the second group of models (Figure 8(b), (c)) in response to the enhanced denudation that is triggered by continental breakup and the establishment of a new base level (i.e. sea level) adjacent to the new continental margin with its rift shoulder topography immediately landward of the new coastline. Denudation to form the coastal plain is a maximum at the coast (and minimum at the escarpment foot) and this denudation plus flexural denudational isostasy bring deepest levels of the crust to the surface along the coast. The rapidity and depth of the incision in models (b) and (c) mean that low-temperature thermochronological ages are youngest at the outer edge of the coastal plain (where the maximum crustal thickness is removed) and are approximately the same age as rifting/breakup, whereas the downwarping model predicts old low-temperature thermochronological ages closest to the coast, younging inland to the foot of the escarpment corresponding to the greatest depth of denudation (Gallagher *et al.*, 1998) (Figure 8).

Evolution of passive continental margins

The conceptual models for the evolution of high-elevation PCMs have been tested using a combination of low-temperature thermo chronology and cosmogenic isotope analyses, and numerical modelling.

Low-temperature thermochronology. Low-temperature thermo chronology is well suited to testing the various conceptual models of post-breakup evolution of high-elevation PCMs because the two broad models have very different implications for the amount of post-breakup denudation and hence thermo chronology. Early compilations of AFTT data from the SE Australian high-elevation PCM clearly demonstrate kilometre-scale denudation of the margin during rifting and breakup, especially along the coastline on the new continental margin, adjacent to the new base level (Moore *et al.*, 1986; Dumitru *et al.*, 1991). This pattern of kilometre-scale post-breakup denudation on the coastal plain, between the escarpment and the coast, and a marked episode of accelerated denudation broadly coincident with continental breakup, has now been broadly confirmed for several margins (e.g. southern Africa, Brown *et al.*, 2000a, 2000b; Eritrea, Balestrieri *et al.*, 2005), as well as being

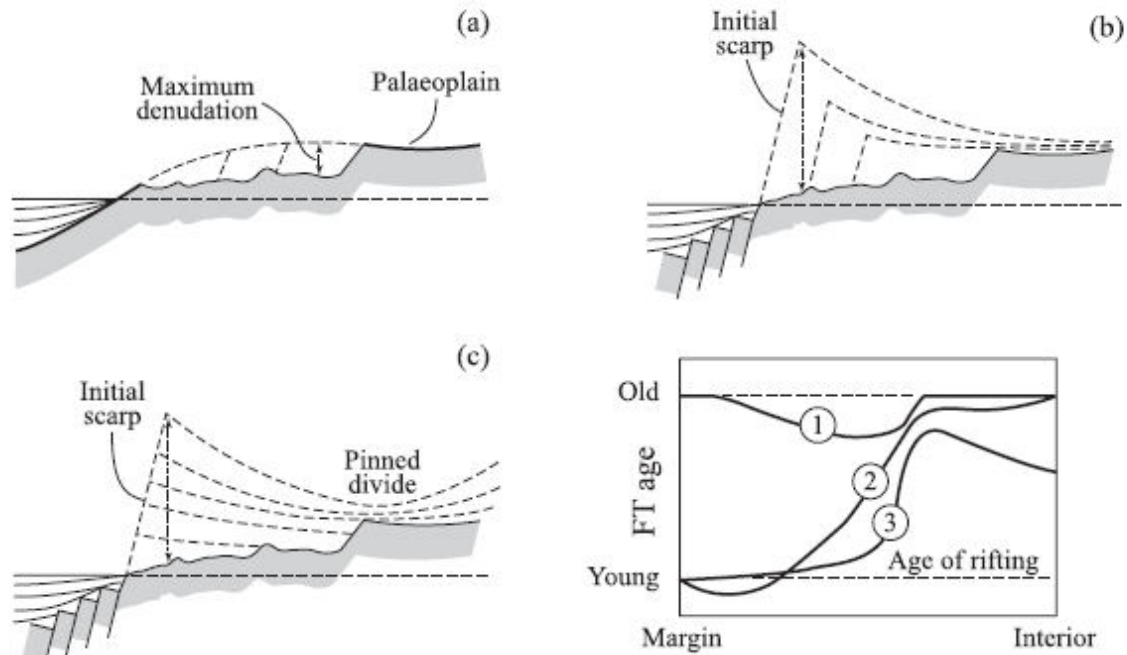


Figure 8. (a) Model for the post-breakup evolution of a high-elevation PCM involving escarpment retreat into a downwarped continental margin. Flexural denudational rebound of the margin is not invoked and so remnants of the ancient down warped plateau surface are postulated along the outer edge of the coastal plain. (b), (c) Models for the post-breakup evolution of a high-elevation PCM involving flexural denudational rebound following (b) escarpment retreat or (c) excavation-in-place of the escarpment. In (b) and (c) the dotted lines represent the total thickness of crust removed as a result of coastal plain formation and denudational rebound. The fourth panel gives a diagrammatic representation of the low-temperature thermo chronological ages predicted by the three models (model (a) = 1, model (b) = 2 and model (c) = 3) (Gallagher *et al*, 1998,)

very clearly evident in more recent data for the SE Australian margin (Persano *et al.*, 2002, 2005). All these various data unequivocally confirm the critical role of flexural isostasy in passive margin evolution, and make it extremely unlikely that the downwarp model is sustainable on the southern African and southeastern Australian PCMs. On the other hand, a major, margin-parallel monoclinical structure on the northern part of the western Indian margin (the Western Ghats) does appear to be consistent with a downwarping model (Widdowson and Cox, 1996; Widdowson 1997) but the lithology of this part of the margin (Deccan Trap lavas) means that it is not possible to test this hypothesis using the apatite-based low-temperature thermo chronology, and others have expressed reservations about it (e.g. Gunnell and Fleitout, 2000).

The most recent thermal modelling of AFTT and apatite-He data from several high-elevation PCMs convincingly points to either very rapid erosion of at least the bulk of - and in some cases virtually all of - the full width of the coastal plain with its backing escarpment within 10 Myr or so of rifting and breakup, and an essentially stable escarpment thereafter (e.g., Brown *et al.*, 2000b; Cockburn *et al.*, 2000; Persano *et al.*, 2002, 2005). These low rates of escarpment retreat after the escarpment was formed are confirmed by Late Cenozoic denudation rates of the escarpment and the coastal plain based on terrestrial cosmogenic nuclide (TCN) analysis (southern Africa, Fleming *et al.*, 1999; Brown *et al.*, 2000b; Bierman and Caffee, 2001; Cockburn *et al.*, 2000; Van der Wateren and Dunai, 2001; southeastern Australia, Heimsath *et al.*, 2000). The rapid formation of the escarpment essentially in its present location, and its considerable stability subsequently, are also consistent with (and thereby provide an explanation for) geomorphological and geological data on the coastal plain, including the weathered mantles on the coastal plain in southeastern Australia that date from very soon after breakup (Bird and Chivas, 1988, 1989, 1993) and mid-Tertiary lavas and sediments on the coastal plain only a few kilometres from the foot of the escarpment (Young and McDougall, 1982;

Nott *et al.*, 1991). On the other hand, modelling of AFTT data from part of the Western Ghats margin formed in basement rocks points to a 35-40 Myr delay between breakup and the onset of major escarpment erosion and coastal plain denudation at 30 Ma (Gunnell and Fleitout, 2000). This delay is difficult to explain but is taken by Gunnell and Fleitout (2000) to indicate that 'the lateral distance of headward erosion at ~30 Ma BP had reached a threshold whereby cumulative unloading since rifting had generated the first major signal of flexural response in the history of the margin' (p. 328).

In summary, low-temperature thermo chronological and TCN data, as well as other data from PCMs, are generally not compatible with a model of constant retreat of a type-1 or type-2 escarpment from an initial position near the present coast (cf. Ollier, 1982; Ollier and Pain, 1994; Seidl *et al.*, 1996; Weissel and Seidl, 1998); this conclusion is also consistent with geophysical data (Matmon *et al.*, 2002). Where there are sufficiently detailed AFTT and apatite-He data, it is clear that these margins have formed by rapid post-breakup river incision seaward of a pre-existing drainage divide (or perhaps, but less likely, by rapid escarpment retreat to about its present position) (e.g., Brown *et al.*, 2000b; Persano *et al.*, 2005). The AFTT and apatite-He data are not yet able to distinguish between these two models. In most situations, the escarpment lip is pinned to a drainage divide, either the main continental divide between seaward- and landward-flowing rivers (e.g. the Drakensberg escarpment in southeastern Africa - Brown *et al.*, 2002) or a secondary, but still major, drainage divide between coastal rivers and rivers that initially flow inland before turning back out to the sea (e.g. the escarpment in far southeastern Australia - Persano *et al.*, 2002, 2005). On type-3 escarpments, various data, including TCN measurements, point to high rates of gorge extension where the coastal rivers cross the escarpment (Nott *et al.*, 1996; Weissel and Seidl, 1998). Thus, we can conclude that escarpments reach their long-term location at the landward edge of the coastal plain very soon after breakup. On type-1 and 2 margins the escarpment then becomes, in gross terms, essentially stable, whereas on type-3 (archlike) margins escarpment evolution proceeds via gorge-head retreat and increasing embaying of the escarpment (Matmon *et al.*, 2002). *Numerical modelling of PCMs.* The early 1990s saw a concerted effort in the numerical modelling of high-elevation PCMs culminating in the publication of several of these models in the *Journal of Geophysical Research* special issues on tectonics and topography in 1994 (Merritts and Ellis, 1994) (Figure 9). These models focused on the gross evolution of the high-elevation PCM escarpment, while also exploring the ways in which the simulated landscape evolution varies with model parameterization and formulation, and with the inherent characteristics of the landscape, including climate, lithological resistance and flexural isostatic properties (e.g., Kooi and Beaumont, 1994; Tucker and Slingerland, 1994). Gilchrist *et al.* (1994a) used a similar approach to exploring escarpment evolution in the specific locality of SW Africa. Overall, numerical models have been able to generate reasonable PCM morphology over reasonable model timescales, using relatively simple process laws and denudational flexural isostasy.

More recent PCM models, following the place-specific approach of Gilchrist *et al.* (1994a), have used the *Cascade* surface process model by Braun and Sambridge (1997) to explore the post-breakup evolution of southern Africa and southeastern Australia, in both cases calibrated by field data from these well studied margins (e.g., van der Beek and Braun, 1998). The modelling indicates that the apparently anomalous drainage patterns of SE Australia, which are supposedly evidence for post-breakup downwarping and have long been the subject of debate (e.g., Ollier and Pain, 1994, 1997; cf. Bishop, 1995; Bishop and Goldrick, 2000), are an expected element of post-breakup drainage evolution and require no downwarping (van der Beek *et al.*, 1999). The modelling by van der Beek *et al.* (1999) also showed that a drainage divide could be maintained inboard of the escarpment lip, as is the case in the southeastern Australia arch-type margin, if the seaward slope of the plateau surface between the divide and the escarpment lip lies in a very narrow range of values. Also in the southeastern Australian context, but with much more general implications, van der Beek *et al.* (2001) have used a numerical model to predict different escarpment evolution scenarios (one similar to the escarpment retreat scenario, and one to the plateau downwearing mode of development) from only slightly different initial conditions in the model runs. A low (100 m) pre-existing inland drainage divide leads to plateau downwearing, whereas an initially horizontal plateau leads to escarpment retreat.

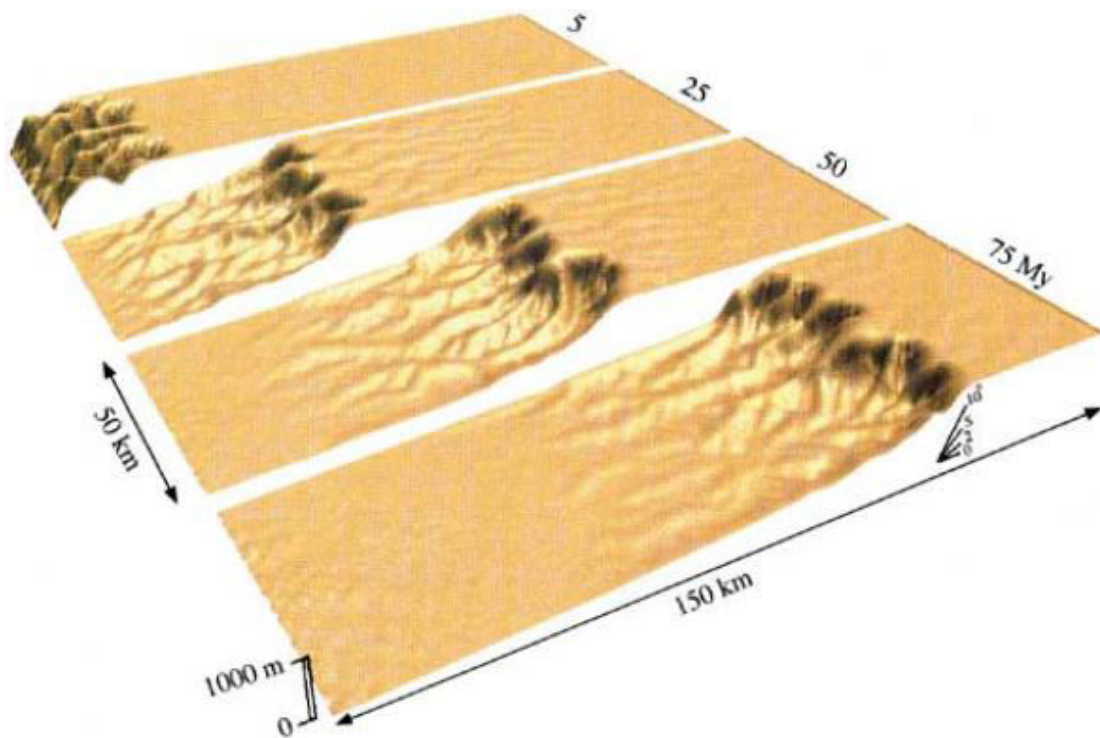


Figure 9. Numerical model of escarpment evolution under a relatively arid climate, highlighting the formation of the coastal plain in the wake of the inland-retreating escarpment. Note how the escarpment retreats at a steady average rate following breakup (Kooi and Beaumont, 1994)

van der Beek and co-workers have also used numerical modelling to predict the pattern of low-temperature thermo chronological data along various margins (e.g., van der Beek *et al.*, 1999, 2002). The models can generally reproduce the observed pattern of AFTT and apatite-He data, likewise reproducing the kilometre-scale denudation along the coastal plain. This approach has been widened to explore in generic ways the initial conditions required for, and the patterns of low-temperature thermo chronological data to be expected in, plateau downwearing and escarpment retreat models (Braun and van der Beek, 2004). The value of these numerical modelling approaches is clear: the simulations show that plateau downwearing and escarpment retreat produce very similar morphologies (Figure 10), but the predicted distributions of apatite-He ages are very different, thereby enabling testing of the two scenarios. Second, Braun and van der Beek (2004) have provided guidance as to the spatial sampling strategy to maximize the information content from a suite of apatite-He ages collected to test competing models for PCM evolution. Even for apatite-He ages of samples collected prior to the setting out of this strategy, the numerical modelling approach is still valuable in confirming the rapidity of the escarpment formation, while not being able to discriminate between the two broad scenarios of PCM evolution (cf. the analysis by Braun and van der Beek (2004) of the data of Persano *et al.* (2002)).

The important conclusions to be drawn from the foregoing include the following. The evolution of high-elevation PCMs, a major first order topographic feature of the Earth's surface, reflects the interaction of rifting tectonics, post-breakup tectonics (including flexural isostasy driven by onshore denudational unloading and offshore sedimentary loading) and surface processes. The considerable field-based, geo- and thermo-chronological and numerical modelling efforts to understand PCMs over the last decade or so mean that much is known about their evolution; major questions remain, however.

It is inescapable that flexural isostasy in response to denudational unloading plays a fundamental role in post-breakup PCM evolution and that downwarping is unlikely to be a major mode of such evolution, but determining whether escarpment evolution is via in-place excavation (plateau degradation) or via escarpment retreat may not be possible with current techniques (Braun and van der Beek, 2004).

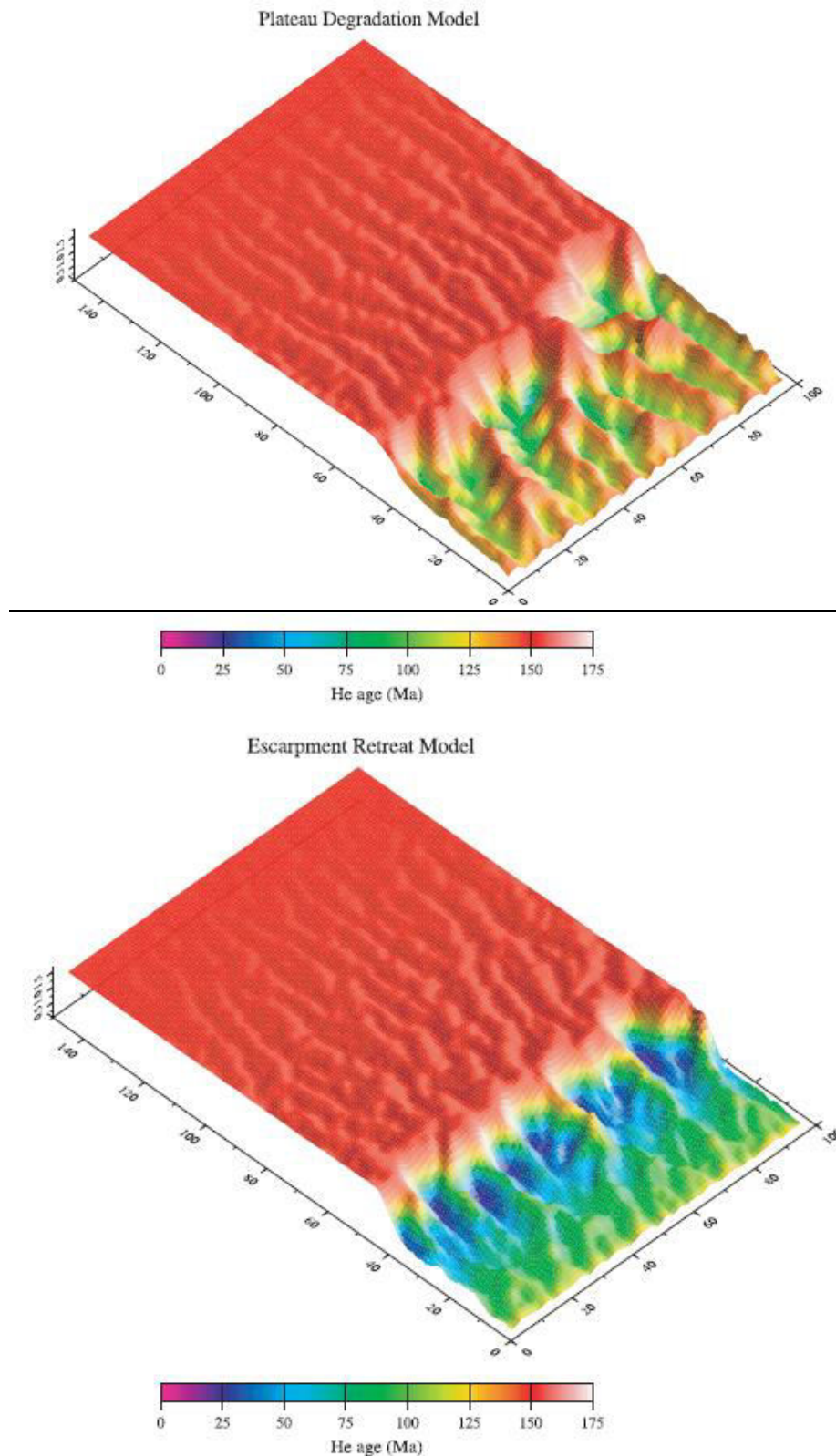


Figure 10. Passive continental margin morphology and predicted apatite-He ages for a 100 Ma breakup age and post-breakup margin evolution by (a) plateau downwearing and (b) escarpment retreat. (b) shows a clear younging of ages from rifting/breakup ages near the coast to younger ages at the base of the escarpment, corresponding to the age when the escarpment stabilized at its present-day position. In the plateau downwearing model (a), post-breakup evolution by river incision and denudation without a clear pattern of escarpment migration mean that all ages are relatively old and close to the 100 Ma age of rifting (Braun and van der Beek, 2004,)

Indeed, to polarize the processes of escarpment formation as either in-place excavation or escarpment retreat may be inappropriate, as a mix of both processes is likely: plateau degradation (excavation) along drainage lines and adjacent hillslopes, with escarpment retreat on interfluvial areas that are distant from drainage lines and/or are formed on resistant lithologies that support an escarpment (such as the basalts of the Drakensberg escarpment, Brown *et al.*, 2002, and the Triassic sandstones of the escarpment in the Sydney Basin, Nott *et al.*, 1996). Flexural isostasy is an integral part of escarpment formation but the narrowness of many coastal plains (that is, the short distance between the escarpment and the coast) means that lithosphere of very low flexural rigidity is required if kilometre-scale denudation has occurred along the coast during rifting and breakup, and much lower amounts of denudation are recorded just a few tens of kilometres inland at the foot of the escarpment (e.g., Braun and van der Beek, 2004; Persano *et al.*, 2005). Persano *et al.* (2005), for example, have argued that such low flexural rigidities (very low values of effective elastic thickness) are unlikely in the case of southeastern Australia, concluding that there must be (as-yet-unmapped) brittle failure (faulting) on the coastal plain to accommodate the amount of denudation and accompanying flexure at the coast and/or higher heat flows along the coast during rifting. A higher geothermal gradient along the coast means that the rocks now cropping out at the coast and sampled for, say, apatite-He analysis have come from shallower depths and that there is a smaller difference between the amounts of flexure at the coast and at the escarpment foot. These various issues are not just geophysical matters of little consequence to an understanding of Earth surface processes and landforms. Low flexural rigidity (a low value of T_e - Figure 7) means that the response to denudational unloading is spatially more restricted. Second, faulting on the coastal plain might be expected to have impacts on geomorphological evolution and drainage patterns.

Orogenic Settings

If the evolution of PCMs reflects the interaction of surface processes and the essentially passive lithospheric responses to these surface processes, the evolution of orogenic settings, the areas of active rock uplift associated with the zones of plate convergence, reflects the interaction of surface processes and active rock uplift, with denudational isostatic feedbacks increasing the amounts of this rock uplift. Moreover, if landscape responses occur over timescales of 100s Myr in PCM settings, they may be geologically instantaneous in actively uplifting mountain settings: tectonic rates and landscape response rates in orogenic settings are at the maximum values recorded on Earth. However (and superficially counter-intuitively), the landscapes of these tectonically active areas may experience little 'evolution', exhibiting forms that remain essentially constant through time, reflecting a balance between rock strength, rock uplift rates and processes (e.g., Burbank *et al.*, 1996; Tippett and Hovius, 2000). In other words, the active landscapes of orogenic belts may be dominated by extreme process rates but may experience little overall morphological change (or evolution). In effect, very high rates of rock uplift feed the crust through a landscape of essentially constant morphology. The Himalayas are one of the Earth's archetypical orogenic landscapes. It was realized during Imperial Britain's great early 19th century survey along the length of India that mountain masses such as the Himalayas must be underlain by a root of material of similar density to the continental crust (Keay, 2000). In the early 20th century Wegener interpreted Suess's 'Pacific' type continental margins in terms of continental convergence and Holmes in the 1920s and Du Toit in the 1930s highlighted the continental over-riding and crustal thickening associated with continental convergence (see Bishop, 2007, for more detail). Dewey and Bird (1970) revived this viewpoint, highlighting the plate tectonics significance of the relationship between high elevation and plate convergence, and for two decades or so this relationship was essentially accepted as reasonable and almost self-evident. Much effort in the last decade or so has been spent in attempting to understand convergent settings, particularly in terms of the relationships between tectonic processes and surface geomorphological processes (Beaumont *et al.*, 2000). The move to include surface processes in such thinking, and in the numerical models that were developed to explore this thinking, was initially triggered, not so much by a desire to understand Earth surface processes and landforms *per se*, but rather from the need to understand the ways in which the ways in which rock uplift responds via isostasy to rapidly eroding topography and thereby influences a rock's trajectory to the Earth's surface and hence the duration of the pressure and temperature conditions, and thus metamorphic regimes, associated with different depths in the crust (e.g., England and Richardson, 1977; Burbank, 2002; Finlayson *et al.*, 2002). The oft-quoted

comment by Hoffman and Grotzinger (1993) neatly captures the critical point: 'Savour the irony should those orogens most alluring to hard-rock geologists owe their metamorphic muscles to the drumbeat of tiny raindrops' (p. 198); likewise, the title of the paper by Zeitler *et al.* (2001) includes the marvellous phrase: 'the geomorphology of metamorphism'. Pysklywec (2006) highlights another important impact of surface geomorphological processes on deeper crustal processes, concluding that active surface erosion is necessary for stable, subduction-like plate consumption, and that, without surface erosion, subduction is inhibited by accumulating crust. The modelling suggests, in effect, that the surface-erosion-driven flux of crust through the Earth's surface means that crust does not accumulate and a 'log jam' of crust is avoided.

The pioneering work of Koons (1989, 1994) has led to the development of numerical models that explore other aspects of the interaction between surface processes and tectonics (rock flux), especially in terms of the direction of the rain-bearing air masses relative to the polarity of subduction (e.g., Willett, 1999; Willett *et al.*, 2001; Beaumont *et al.*, 2000; Reiners *et al.*, 2003). These models highlight the ways in which the crustal flux leads to different patterns of rock uplift and erosion depending on whether orographic precipitation is on the pro- or retro-side of the accretionary orogenic wedge (Figure 11). These models thus indicate that climate drives patterns of rock exhumation and the distribution of strain (and therefore metamorphic grade) with depth. Moreover, the interaction between rock mass flux and erosion driven by orographic rainfall also influences the gross topographic structure of the mountain belt in terms of the location of the crest (topographic divide) of the orogen vis-a-vis the detachment point of the subducting plate (Figure 11(b)). It has also been argued that very high rates of local erosion may also lead to enhanced denudational isostatic rock uplift and hence localized rock deformation within the orogen, reviving Wager's (1937) argument for the development of river anticlines in the Himalayas (e.g., Dahlen and Suppe, 1988; Whipple and Meade, 2004; Montgomery and Stolar, 2006). In extreme cases, such as beneath extremely rapidly incising rivers such as the Tsang Po where it passes the Namche Barwa massif in the easternmost Himalaya, the mantle is bowed upwards beneath the areas of highest stream power and most rapid incision (Zeitler *et al.*, 2001). This bowing upwards is taken to reflect the extreme rates of denudational isostatically driven rock uplift attendant on the extreme rates of river incision. These very high rates of rock uplift mean also that the major knickpoint where the Tsang Po crosses this 'tectonic aneurysm' (Zeitler *et al.*, 2001) is essentially stable in location and not migrating headwards. In effect, the very high rates of isostatic rock uplift as a result of gorge incision downstream of the knickpoint prevent headward migration of the knickpoint.

Thiede *et al.* (2005) found evidence for such erosionally controlled rock uplift in the pronounced spatial correlation between very young apatite fission track ages and areas of high precipitation and river and sediment discharges in the Sutlej Valley of the Himalaya. Similar links between climate, discharge and rock uplift have been likewise proposed by, for example, Wobus *et al.* (2003) and Grujic *et al.* (2006). Others (e.g. Burbank *et al.*, 2003) are less convinced that surface processes (i.e. erosion) exert a fundamental control on rock uplift rates, but the observations of Thiede *et al.* (2005) and Zeitler *et al.* (2001), as well as numerical modelling results (e.g., Dahlen and Suppe, 1988; Whipple and Meade, 2004), do point strongly to this effect.

These sorts of deep-seated impact of surface processes on rock uplift and metamorphic history are only possible when the rates of surface processes are very high and lithospheric properties are such that the rock eroded at the surface is continually replaced isostatically by rock uplift from beneath. Rapidly eroding, tectonically active settings are characterized by massive transfers of crustal material, both as rock is tectonically advected into the orogen, and as rock (sediment) is advected out of the orogen by surface processes; these fluxes provide one of the great attractions of research in these areas. Thus sediment fluxes out of active mountain belts correspond to rates of river incision of 2-12 mm a⁻¹ in the Himalayan Indus (Burbank *et al.*, 1996; Hancock *et al.*, 1998), 10-20 mm a⁻¹ in the Southern Alps (New Zealand) (Tippett and Hovius, 2000), 2-6 mm a⁻¹ in the Spanish Sierra Nevada (Reinhardt *et al.*, in review) and 3-7 mm a⁻¹ (and locally up to 60 mm a⁻¹) in Taiwan (Dadson *et al.*, 2003, 2004; Schaller *et al.*, 2005). These rates, which for steady-state landscapes correspond to rates of landscape lowering, may be compared with rates of <0.01 mm a⁻¹ in the PCM settings described above. The incision rates in active orogens are being driven by some of the highest rainfalls on Earth (e.g. 15 m of precipitation at the crest of the Southern Alps, Tippett and Hovius, 2000; 2-5 m mean annual precipitation in Taiwan with an average of four typhoons per year, Dadson *et al.*, 2003, 2004).

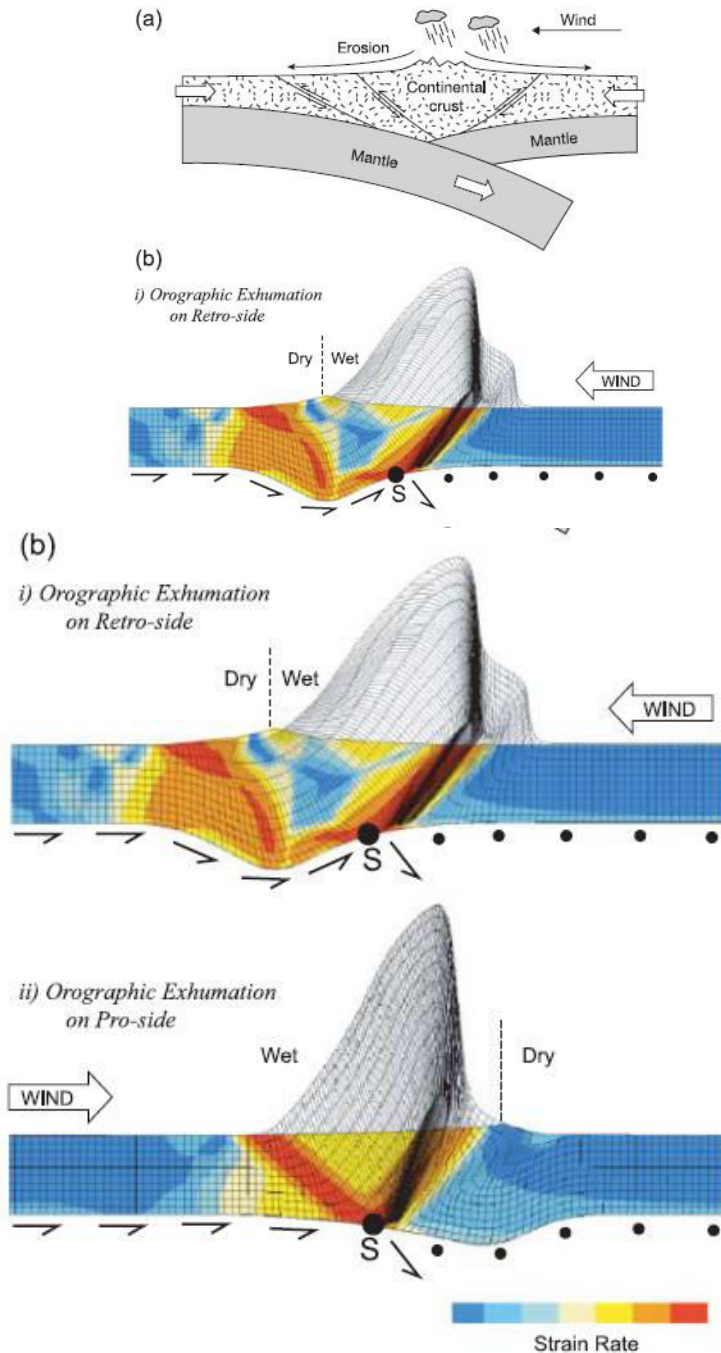


Figure 11. (a) Diagrammatic representation of convergent plates with continental crust overlying subducting mantle. The plate on the left is subducting beneath the plate on the right and the moisture-bearing air masses are coming from the left ('Wind'). Many major orogens, including the Himalayas, the Southern Alps and Taiwan, have this strongly polarized approach of air masses. The thin normal-headed arrows show the flux of eroded material out of the orogen, and the thin half-headed arrows indicate faulting resulting from the convergence. (b) Finite-element numerical model of the interaction of precipitation, rock uplift, denudation and subduction in a convergent orogen with a subducting plate. (i) The results when air masses approach from the opposite side to the subducting plate (i.e. on the retro-side of the orogen) and (ii) those when the subduction and weather systems have the same polarity. The topography is the solid line above the coloured/shaded parts of the model; note the way in which the topographic divide and therefore the drainage distribution vary with the polarity of the weather systems. The deformation of the mesh indicates the crustal deformation (the colours/shading highlight the strain rates, which are of relevance to metamorphism); the mesh shown above the surface is a measure of the crust removed by denudation. (a) is a reversed version of Figure 4a of Dietrich and Perron (2006). (b) is from Beaumont *et al.* (2000).

The erosive potential of these high to extreme, high-intensity precipitation rates is considerably enhanced by seismic shaking and concomitant mass movement (Dadson *et al.*, 2003, 2004). These rates alone make these settings extremely attractive for any geomorphologist: amounts of annual bedrock river incision are directly measurable (e.g. up to 10 mm in one year in Taiwan, equivalent to 10 km Ma^{-1} - Hartshorn *et al.*, 2002; up to 4 mm in one year in the Indus, equivalent to 4 km Ma^{-1} - Hancock *et al.*, 1998), and incision processes can be assessed by judicious placement of monitoring sites (e.g., Hancock *et al.*, 1998).

These tectonically active settings are also extremely important for understanding the large sediment fluxes to the world's oceans that are sourced in such settings: Taiwan represents about 0-02% of the world's sub-aerial area but contributes about 2% of the world's suspended sediment discharge to the oceans (Dadson *et al.*, 2003). We return to these points below in the next section. A third attraction of these convergent settings is that mass balance studies show that the amount of crust being advected into the orogen approximately matches that being lost by erosion (e.g. Southern Alps, Adams, 1985; Tippett and Hovius, 2000; Taiwan, Dadson *et al.*, 2003, 2004). These orogens have therefore been treated as being in steady state and used to assess competing controls on the development of steady-state topography. Before turning to this question, we explore briefly the landscape couplings and linkages that underpin the very high rates of sediment production and export in orogenic belts.

Bedrock river response to rock uplift: 'top-down' and 'bottom-up' processes

The very high rates of denudation in orogenic belts and the corresponding sediment fluxes that are exported from such mountain belts depend on very efficient coupling between hillslopes and channels, and very efficient transport of sediment down the drainage net once the sediment has entered the channel from the adjacent hillslopes. A large proportion of recent research documenting bedrock river incision and channel-hillslope linkages has focused on large rivers flowing through seismically active orogens experiencing rapid rock uplift and very high-magnitude, monsoonal or typhoon-driven storm events, such as in the Himalaya or Taiwan. These large, sediment-laden bedrock rivers are able to incise bedrock rapidly, at rates that are generally understood to keep pace with rock uplift (Southern Alps, Adams, 1980, 1985; the Indus, Burbank *et al.*, 1996; Hancock *et al.*, 1998; Taiwan, Hartshorn *et al.*, 2002; Schaller *et al.*, 2005), and to maintain valley sides at the critical slope angle for rock failure (Burbank *et al.*, 1996; Hovius *et al.*, 1997; Hovius, 2000; Dietrich *et al.*, 2003; Roering *et al.*, 1999, 2005) (Figure 12). In numerical models, steady-state topography is attained in a time that scales with the uplift rate, width of the mountain belt and physical parameters of the erosion model (Kooi and Beaumont, 1996; Willett *et al.*, 2001). Valley side and channel are closely linked in these steady-state topographies, justifying the common treatment in SPMs of close-coupling channel and hillslope, with the hillslope, in effect, 'tracking' the bedrock channel at its foot and lowering at the same rate as this channel (cf. Howard *et al.*, 1994). This tracking depends on channel bed lowering only to the extent that the incising channel provides the accommodation space and sediment transport capacity for hillslope-derived sediment to be delivered to the channel and moved downstream. In these situations, the processes that deliver hillslope-derived sediment to the channel can be thought of as 'top-down' processes, whereby seismically shaken hillslopes experience very intense monsoonal or typhoon-driven rainfall events and actively deliver sediment to bedrock channels (Dadson *et al.*, 2004). This sediment and the high river discharges then combine to incise the river bed.

Many bedrock rivers, however, do not respond to perturbation by the geologically instantaneous lowering of the channel bed and hillslopes at the rate of surface uplift. Instead, such channels develop one or more knickpoints, which are steplike steepenings in the channel profile, migrating upstream in response to a base-level change (e.g., Gardner, 1983; Hayakawa and Matsukura, 2003; Bishop *et al.*, 2005; Crosby and Whipple, 2006).

The bottom-up process of headward propagation of knickpoints is a communication link between base level, the drainage net and the whole catchment including hillslopes (Whipple and Tucker, 1999; Crosby and Whipple, 2006).



Figure 12. Downstream view of the middle reaches of the Rio Torrente, a rapidly incising stream at the western end of the Sierra Nevada mountain block, southern Spain. Note how the hillslopes are dominated by shallow bedrock landsliding, which delivers sediment very efficiently to the Torrente. The two dams, built in 1981, were completely filled with sediment in less than 20 years. See the work of Reinhardt *et al.* (in review) for more detail. (Photograph of Liam Reinhardt.)

A catchment with a propagating knickpoint is segmented into the reach downstream of the knickpoint, with steep slopes that are well connected to the channel, and the reach upstream of the knickpoint, which may exhibit subdued, unrejuvenated topography if it has been a long time since knickpoints last propagated through the drainage net (Schoenbohm *et al.*, 2004; Clark *et al.*, 2005; Reinhardt *et al.*, in review). In these settings dominated by knickpoint propagation, the processes that deliver hillslope-derived sediment to the channel can be thought of as 'bottom-up' processes, whereby a knickpoint, triggered by surface uplift, propagates headwards up a trunk stream, triggering knickpoints in tributaries and steepening hillslopes. Such a landscape is not in steady state and the response to rock uplift is time transgressive, and the principal factors that determine the duration of the transient state are the rates and styles of propagation of knickpoints, as well as the rates and styles of adjustment of hillslopes to channel bed lowering as the knickpoints propagate through the drainage net, past the bottoms of slopes. Maintenance and headward retreat of a steplike knickpoint in the channel can communicate a large proportion of the relative base-level fall to the whole drainage net. On the other hand, if the knickpoint dies out by backwards rotation (that is, by Gardner's (1983) processes of inclination or replacement), the knickpoint diffuses away and either the relative base-level fall is largely accommodated in the vicinity of that relative base-level fall and the full magnitude of the base-level fall may not be communicated to the catchment, or the base-level fall does propagate headwards but at a much lower rate than for the propagating knickpoint (Whipple and Tucker, 1999).

As a first approximation, it is not unreasonable to treat hillslopes as closely linked to channels and as 'tracking' the bedrock channel as the channel lowers, be that via top-down or bottom-up processes or channel bed incision (e.g., Howard *et al.*, 1994; Whipple and Tucker, 1999). This treatment is less appropriate for transient topographies in which slope steepening as a result of knickpoint propagation triggers time-transgressive responses on hillslopes depending on hillslope morphology and materials. The rate of sediment flux on convex hillslopes, for example, is linearly related to hillslope angle (Gilbert, 1909; Anheer, 1976, 1987; Fernandes and Dietrich, 1997; Roering *et al.*, 1999, 2001), and such hillslopes respond to increases in the rate of base-level lowering by

slowly increasing the rate of sediment production and hillslope convexity and/or through lateral hillcrest migration (Fernandes and Dietrich, 1997; Mudd and Furbish, 2005).

A developing mountain belt will begin to be sculpted by top-down processes of bedrock river incision driven by high discharges and high sediment fluxes only after the development of sufficient topography to generate the high precipitation rates, high discharges, steep slopes and high sediment fluxes necessary for the development of topographic steady-state landscapes (see below). While the mountain belt is growing prior to the development of topographic steady state (Montgomery, 2001), bottom-up processes of knickpoint propagation are the key way in which the 'information' that relative base level is falling is communicated to the rising mountain block. In some situations, mountain blocks may never pass from the transient landscapes characterized by propagating knickpoints to the full steady-state landscape dominated by top-down processes. Smaller streams, even in tectonically active areas, and/or in regions normally experiencing lower rainfall and discharge than in, say, the Himalaya or Taiwan, may never develop steady-state conditions. In the case of the smaller streams, lower sediment fluxes and lower discharges may mean that top-down processes are less capable of incising channel beds and lowering hillslopes, even though hillslope (valley side) and channel are closely coupled. Hillslope supply of sediment may exceed the transport capacity of the channels, leading to shielding of the bed and the inhibiting of incision (e.g., Sklar and Dietrich, 1998). Knickpoint propagation then becomes the fundamental control on landscape response times and these landscapes dominated by bottom-up processes will exhibit transient conditions as the norm. Rates of bedrock channel knickpoint propagation determine the duration of these transient states and have been examined by, for example, Weissel and Seidl (1998), Hayakawa and Matsukura (2003), Bishop *et al.* (2005), Crosby and Whipple (2006) and Reinhardt *et al.* (in review).

Steady-state Landscapes and Post-orogenic Settings

The preceding section highlights the ways in which landscapes in orogenic belts are not necessarily in steady state. Nonetheless, considerations of steady-state topography have dominated research in active orogens, often without clear definition of what is meant by steady state and how it is attained (Willett *et al.*, 2001). As Willett *et al.* (2001) have noted, at steady state, erosion must be in balance with the full tectonic (rock advection) velocity field, which has both vertical and horizontal components (Figure 11). Horizontal displacements or velocities are significant in essentially all active tectonic settings including convergent mountain belts, and, in fact, are typically larger than the vertical component by about an order of magnitude (Willett *et al.*, 2001). This point has not been fully appreciated, and it is common to see topographic steady state defined in the literature as the condition that rock uplift rate is equal to erosion rate. Willett and Brandon (2002) defined four different types of steady state.

- *Flux steady state*, where rock flux into an orogen is equal to the erosional flux out. This does not necessarily equate to steady (or constant) topography, as spatial variations in erosion can result in temporal variations in landscape form while erosion flux remains constant.
- *Topographic steady state*, where the surface elevation at every point of the region of interest remains constant. In this situation, erosion at every point must balance both the rock uplift and the horizontal advection of rock. Numerical models indicate that this is readily achieved under conditions of vertical rock uplift only, but is less likely when the crust is being advected horizontally.
- *Thermal steady state*, where the temperature field in the crust is time invariant.
- *Exhumation steady state*, where ages obtained for a specific low-temperature thermochronometer from rocks in the region of interest are constant.

In terms of our interests here, it is important to remember that virtually all SPMs and physical ('sand-box') models of tectonically active orogenic settings (e.g., Bonnet and Crave (2003) for a physical model) are allowed to run until they are in flux steady state and, generally, topographic steady state. At this stage of the simulation exercise, the models exhibit time-invariant (or time-independent) landforms, which can then be assessed for the influence of the various forcing factors that determine landscape morphology (such as climate, lithology, rates of vertical and horizontal tectonics, and so on) (Figure 13).

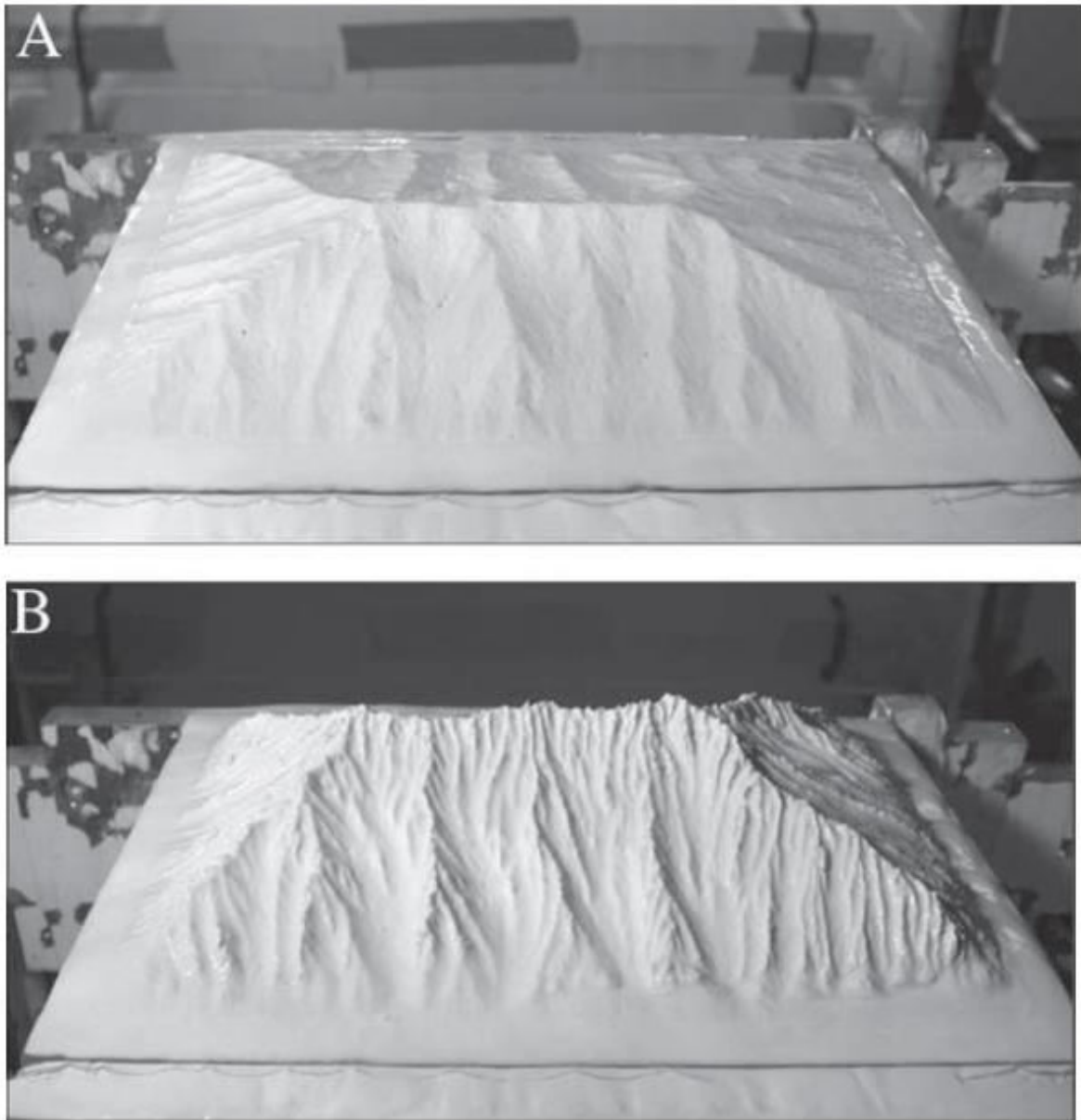


Figure 13. Physical model by Bonnet and Crave (2003) used to explore the impacts of changes in precipitation on the morphology of the topographic steady state. The model consists of a silica paste in a box. The block of silica paste is uplifted and 'rained' on (by misting water) at constant rates until a topographic steady state is achieved (A). In this model run, a decrease in 'rainfall' results in a new steady-state topography at a higher elevation (B). The lower 'rainfall' (and runoff) requires steeper slopes to re-establish the erosion rate that achieves steady state with the rate of rock uplift. Steeper slopes for a given spacing of large channels equate to higher-elevation peaks, as in (B). This result is consistent with numerical modelling results in which the erosiveness parameter is decreased and elevations increase (Whipple *et al.*, 1999). The figure is Figure 3 of Bonnet and Crave (2003)

The numerical models by Willett *et al.* (2001) indicate that topographic steady state is achieved more rapidly for models with no horizontal crustal advection (i.e. with vertical rock uplift only). With horizontal rock advection, topographic steady state is achieved only at the scale of the entire mountain range, with even the first order drainage basins being unstable over time in the presence of horizontal shortening. Numerical modelling suggests that, with the horizontal rock advection typical of convergent mountain belts, the resulting mountain range is asymmetric. The small, active mountain ranges in Taiwan, New Zealand, and the Olympic Mountains (Washington state) all exhibit asymmetric topographic form with the asymmetry consistent with the polarity of

subduction (Willet *et al.*, 2001). The consistent presence of this asymmetry suggests that horizontal tectonic motion is an important determinant of the macro-geomorphic form of these convergent mountain ranges. This asymmetry is likely enhanced by the relative stream powers of the rivers draining either side of the mountain belt, particularly in settings, such as described above, where precipitation is highly asymmetric. Concentration of orographically forced precipitation on the pro-side of the orogen (the side experiencing rapid rock uplift - Figure 11(b)(ii)) may enhance the asymmetry inherent in the mountain block as a result of the convergent tectonics with a substantial component of horizontal rock advection (Beaumont *et al.*, 2000; Willett *et al.*, 2001; Reiners *et al.*, 2003; cf. Hovius, 2000).

The relative roles of climate and tectonics in fashioning landscapes are, in effect, in balance in steady-state landscapes. A 'standard' interpretation would argue that tectonics (rock uplift) drives surface elevations upward, triggering increased orographic precipitation and weathering and ultimately glaciation, all of which act to lower the landscape until a balance between rock uplift and denudation is achieved and steady-state topography is formed (e.g., Willett and Brandon, 2002). Moreover, even if a full steady-state topography is not formed, the increased intensity of surface processes with surface uplift acts to limit the height to which mountain peaks will grow (e.g., Brozovic *et al.*, 1997; Whipple *et al.*, 1999; Roe *et al.*, 2002). These results mean that the mechanism of climatically driven enhanced valley incision and associated isostatic uplift of mountain peaks, as proposed by Molnar and England (1990), can generate only modest amounts of surface uplift of peaks (Gilchrist *et al.*, 1994b; Montgomery, 1994). The related debate, now polarized into two camps, was touched on briefly above and recently summarized in Molnar's (2003) commentary on papers by Burbank *et al.* (2003), Dadson *et al.* (2003) and Reiners *et al.* (2003). In brief, one camp argues that denudational isostasy and orographic exhumation mean that climate-associated erosion is a major 'driver' of rock uplift (e.g., Montgomery *et al.*, 2001; Reiners *et al.*, 2003; Thiede *et al.*, 2005). The other camp argues that there is no evidence that rock uplift rates reflect precipitation rates (e.g., Burbank *et al.*, 2003).

Flux and topographic steady states are features of numerically and physically modelled landscapes, and indeed probably of the real landscapes they seek to simulate (e.g., Pazzaglia and Brandon, 2001). As we have seen, however, such steady state is generally achieved in the modelled landscapes under conditions of constant tectonic and climate-process forcing. The Earth does not exhibit such constancy of forcing, especially in climate in the Cenozoic, and it is therefore important to know whether real landscapes can ever achieve flux and topographic steady states. The numerical experiments of Beaumont *et al.* (2000) show that for slow forcing (that is tectonic and/or climatic forcing with a long period) the landscape evolution is very close to steady state, and as the period of the forcing decreases (i.e. the fluctuations in forcing become more rapid) the landscape response lags and is attenuated. For high-frequency forcing (i.e. rapidly fluctuating climate and/or pulsed rock uplift), the landscape response approaches an average condition, with the rapid variations in forcing strongly attenuated. Numerical modelling, parameterized by the morphology of the Central Range of Taiwan, which is often cited as a landscape in topographic steady state (e.g., Suppe, 1981), has been used to investigate the response times of such a system to perturbations in climate (Whipple, 2001). Estimated response times generally range from 0-25 to 2-5 Ma, depending on the precise formulation and parameterization of the model and the magnitude and type of climatic perturbation. Thus, it is reasonable (as is intuitively apparent) to argue that steady-state topography and denudation are likely to prevail during periods of climatic stability, but the rapid climatic fluctuations of the Quaternary appear to preclude the attainment of steady-state conditions in modern orogens (Whipple, 2001). Active orogens have thus been the focus of considerable geomorphological, thermochronological and tectonics research over the last decade or so, probably reflecting the perceived tractability of working in a supposed steady-state system. However, such orogenic settings constitute only a minor proportion of the Earth's surface (albeit, as we have seen, contributing disproportionately large amounts of sediment to the world's oceans), and there remain vast areas of the Earth that have been largely ignored in the recent re-focussing of attention on landscape evolution. These vast continental (intra-plate) areas are largely distant from tectonic activity and pose intriguing questions about landscape longevity. We have seen above how it is unlikely that landscapes can have survived from, say, the Cambrian, but considerable areas of the Earth's surface have almost certainly survived from the Mesozoic at least (e.g., Twidale, 1976, 1998; Nott, 1995).

Recalling Thornbury's (1969) opinion that landscapes on Earth are largely probably no older than the Pleistocene thus prompts the following question: how can landforms persist sub-aerially for the demonstrably long periods that they have? Sugden's (1968) seminal work on selective linear erosion in glacial terrains accompanied by the preservation of ancient upland surfaces by cold-based plateau ice points to appropriate mechanisms for preservation of ancient landscapes in glacierized settings, and these ideas have now been confirmed by detailed field and morphological analyses (e.g., Kleman, 1994) and by cosmogenic isotope analyses (e.g., Bierman *et al.*, 1999; Briner *et al.*, 2003, 2006; Fabel *et al.*, 2004; Phillips *et al.*, 2006; Staiger *et al.*, 2005). A few workers, including Twidale (1976, 1998), Crickmay (1975) and Young and McDougall (1993), have addressed this and related issues in non-glaciated terrains but by and large these matters remain to be examined in detail. The numerical modelling by Baldwin *et al.* (2003) points to several important factors, concluding that a change from detachment-limited to transport-limited conditions through time as slopes and stream power decline, changing magnitude-frequency characteristics of erosional and sediment transporting events, and denudational isostasy may all act to increase landscape longevity of post-orogenic settings; to these factors can be added declining precipitation as elevation declines. Lithological controls, whereby drainage net rejuvenation driven by denudational isostasy is stalled or 'caught up' on resistant lithologies, must also play a part in slowing the transmission of base-level changes to the whole catchment (Twidale, 1998). Thus, there is an inherent inertia in post-orogenic landscapes in that, as elements of the landscapes get older, they are progressively less likely to be eroded (i.e., more likely to persist): 'The ability of a landscape to resist impulses of change tends to increase with time' (Brunsden, 1990). This inertia must reflect several factors (Migon and Goudie, 2001). First (and perhaps most obviously), extreme aridity may slow landscape evolution to very low rates (e.g. Namibia, Cockburn *et al.*, 2000; Bierman and Caffee, 2001; Van der Wateren and Dunai, 2001; Peru, Dunai *et al.*, 2005; Dry Valleys of Antarctica, Nishiizumi *et al.*, 1991; Summerfield *et al.*, 1999a, 1999b), but this cannot strictly be taken as a cause of inertia in the landscape. The reporting by Bierman and Caffee (2002) of denudation rates in the Eyre Peninsula in Australia that are lower than the rates they reported from Namibia, which is much drier than the Eyre peninsula, likewise shows that aridity cannot be the full explanation for landscape 'inertia'. Second, Twidale's and Crickmay's arguments that inequality of activity may cause parts of the landscape to become separated from erosional processes that respond to base-level changes mean that those parts of the landscape isolated in this way may in effect cease to evolve. Third, magnitude-frequency considerations mean that, as the duration of persistence increases, it is reasonable to postulate that the landscape will have been exposed to ever-higher-magnitude events. Having resisted these (perhaps due, say, to being formed on highly resistant lithologies or to the fact that channel incision means that slopes are disconnected from the fluvial system), the landscape is unlikely to evolve further. Of course, the foregoing does raise the issue of what is precisely meant by a landform and a landscape. The high-level surfaces that are widespread on the Earth's surface and have universally been taken since Davis's time as relict old surfaces are a case in point (Figure 14).

Research Challenges

Tectonics has always been a consideration in landscape evolution research, whether it is assumed that the landscape evolution has proceeded under conditions of tectonic quiescence, as Davis did (albeit an initial brief period of surface uplift being necessary to initiate the cycle of erosion), or that the landscape evolution was concurrent with tectonic activity, as other schemes of landscape evolution have assumed (e.g., Penck, 1953). The interactions between climate and tectonics ('chickens and eggs') remain a fascinating and challenging research area (Molnar and England, 1990). It seems unlikely that increased valley erosion as a result of climate change can drive sufficient mountain peak uplift to have an appreciable impact on relief (and hence further climate change) because the volumes of valley erosion seem insufficient to drive substantial peak uplift (Gilchrist *et al.*, 1994b) and in any event the increased 'orographic erosion' and glaciation induced by such peak uplift soon limits the peak uplift (Whipple *et al.*, 1999). But whether climate-driven processes lead to enhanced *rock* uplift (rather than enhanced *surface* uplift) is still a matter of debate. In steady-state topography, there may



Figure 14. The summit plateau and the Shoalhaven River gorge in the southeast Australian passive margin highlands, New South Wales.

be no topographic signature of greater rock uplift as a result of increases in climate-driven erosion, but such enhanced rock uplift may be recorded by low-temperature thermochronological systems (e.g., Zeitler *et al.*, 2001). This viewpoint is critical to understanding sediment fluxes in the context of high sediment fluxes from steady-state topography and also means that the climate-tectonics question can be re-framed in terms of when major erosive climate regimes were established. Recent cosmogenic isotope data confirm the impact of changing climates on changes in rates of river incision (e.g., Schaller *et al.*, 2005; Reusser *et al.*, 2006).

The rates of landscape response to such changes in climatic and tectonic forcing (and the concomitant issue of landscape sensitivity) constitute a major research question. Brunsdon and Thornes (1979) explored these issues nearly three decades ago, and the issues must again be the focus of research, building on the research on steady-state landscapes that has been so fruitful. Moves towards addressing transient landscapes (e.g., Whipple *et al.*, 2007) are therefore highly welcome, and recent work shows that judicious sampling for low-temperature thermochronological analyses has the potential to enable the determination of the timescales of post-orogenic erosional decay of orogenic topography (e.g. a decrease of the orogenic topography of the Dabie Shan by a factor of 2-5 to 4-5 during the last 60-80 Myrs -Braun and Robert, 2005). The bedrock river - 'the backbone of the erosional landscape' (Hovius, 2000, p. 88) - lies at the heart of the recent resurgence of interest in long-term landscape evolution and interests in landscape response rates. I have not reviewed bedrock rivers explicitly here but the large body of research over the last two decades or so on bedrock rivers, building on the seminal work of Howard and Kerby (1983) and Kirkby (1986), underpins a major proportion of our discussion (e.g., Whipple and Tucker, 1999; Snyder *et al.*, 2000; Whipple, 2004). Indeed, without the bedrock river stream-power framework that Howard and Kerby (1983) provided, much of the exploration of the relationships between tectonics and climate and topography would probably have been more difficult. That said, there remains the issue of better formulation of the algorithms used to understand bedrock river incision, given the variable capability of current algorithms to simulate known long profile evolution (e.g., Stock and Montgomery, 1999; van der Beek and Bishop, 2003). To date, the bedrock fluvial incision algorithms have generally been variants of the stream power law, modified in various ways to take more explicit account of sediment

transport. Recent work is now attempting to incorporate sediment transport more rigorously, including, for example, shielding of the bed (e.g., Sklar and Dietrich, 1998, 2001). Moreover, it is now important that the interactions and influences of various other fluvial parameters, such as width, depth and sinuosity, be elucidated for bedrock rivers. The latter issues were addressed decades ago for alluvial rivers and now it is necessary to do the same for bedrock rivers, not least to improve the formulation of the process 'laws' in numerical models (e.g., Finnegan *et al.*, 2005). Moreover, cosmogenic isotope analyses now enable the determination both of rates of incision, which should increasingly be used to test fluvial incision algorithms, and of the timing of incision, by dating strath terraces (e.g., Hancock *et al.*, 1998; Reusser *et al.*, 2006).

The representations of slope processes in SPMs also remain relatively under-developed, and range from the straightforward lowering of a slope at the same rate as the bedrock channel at its foot (Howard *et al.*, 1994) to representations based on simple functions of the slope angle or slope curvature and a diffusion constant (see Codilean *et al.*, 2006a, for more detail). This diffusion constant is still poorly constrained (Martin, 2000), and in any event the relationship between 'real-world' values of such coefficients and the values they take within numerical models is not straightforward (van der Beek and Braun, 1998). Indeed, the values of model parameters may have no 'real-world' significance, simply taking values that 'tune' the model to produce 'realistic' output (Anderson, 1994; van der Beek *et al.*, 1999). At least three areas of slope-related 'process' research, additional to ongoing investigation of diffusive slope processes *per se* (e.g., Roering *et al.*, 2001), can be identified as requiring attention if understanding of long-term landscape evolution (and its numerical modelling) is to continue to advance. These are the following: the controls on rates and depths of soil production (Heimsath *et al.*, 1997, 2000, 2001; Wilkinson *et al.*, 2005), hitherto represented in numerical models in a rather elementary fashion (e.g., Tucker and Slingerland, 1994); the role of debris flows in landscape evolution (Stock and Dietrich, 2003) and the assessment of rates at which bedrock channel rejuvenation is communicated from the channel to the adjacent hillslopes (channel-hillslope connectedness).

The fuller specification of processes in SPMs and further validation of these models will permit better assessment of the spatial extent of channel and hillslope responses to rejuvenation, and the rates of these responses, and hence of landscape relaxation times, as well as better quantification of the sediment fluxes from the landscape (along with the isostatic unloadings and loadings associated with these fluxes). The overall framework of 'top-down' and 'bottom-up' processes presented here may be useful in understanding why different landscapes may have different relaxation times. Orogens like Taiwan and the Southern Alps, drained by rivers with high discharges and high sediment concentrations, appear to be in topographic and flux steady states driven by 'top-down' processes, but such orogens must have experienced a period during which propagating knickpoints transmitted the rejuvenation signal upstream via 'bottom-up' processes. Once, however, such orogens reach sufficient elevation for impinging typhoons to have a major impact on discharge and sediment production (especially when seismic shaking is followed by typhoons), then 'top-down' processes will drive landscape evolution.

The recent emphasis on steady-state landscapes experiencing high rates of rock uplift has somewhat glossed over the widespread occurrence of large areas of relict landscapes on the Earth's surface, often in the upstream parts of the steady-state landscapes themselves (e.g., Epis and Chapin, 1975; Abbott *et al.*, 1997; Gubbels *et al.*, 1993; Sugai and Ohmori, 1999; Clark *et al.*, 2005). These relict landscapes are of major interest for several reasons and deserve attention (Figure 14). First, their presence and preservation prompt questions as to their antiquity: do they represent ancient surfaces? It has long been argued that Scotland's Cairngorm mountain-top plateaux, for example, are Tertiary-age erosion surfaces (e.g., Sissons, 1967), but recent cosmogenic isotope analyses have not yielded any pre-Quaternary ages on the tops of tors on the Cairngorm 'surface' (Phillips *et al.*, 2006). The apparently 'ancient' morphology of an apparently relict upland surface may not match the age of the surface. In other words, these data imply that an ancient plateau morphology may be retained as the surface is lowered (by, for example, solution, etch and other chemical weathering processes), but that the 'age' of the 'present' surface may be much younger than the age of formation of the original surface from which the present surface has 'descended'. Herein lies an excellent example of the power of new techniques such as cosmogenic isotope analysis in bedrock settings. Perhaps more importantly, the time is ripe to re-visit the more general issues of age and origin of the classic low relief upland

surfaces that formed such a focus of Davis's work (Phillips, 2002). Cosmogenic isotope analysis has passed through the pioneering phase and is now a well established technique. The application of the technique must increasingly involve larger numbers of determinations and the analysis of probability density functions of ages (or rates) to enable hypotheses to be tested in more sophisticated ways (e.g., Briner *et al.*, 2006; Dunai *et al.*, 2005; Reusser *et al.*, 2006). Such probabilistic approaches can now be extended into several areas. Parker and Perg (2003) have presented a probabilistic treatment of the impact of temporal and spatial variations in elevation and erosion rate on average catchment-wide erosion rates derived from the average cosmogenic isotope contents of sediments. They highlight the lengths of time required for new cosmogenic steady-state conditions to develop after the perturbation. The increasing 'ease' of cosmogenic isotope determinations, especially in the case of stable isotopes such as ^{21}Ne and ^3He (which also both require small sample volumes compared to the radio-isotopes), means that determination of the cosmogenic isotope concentrations of individual clasts is now a realistic goal. Such individual clast concentrations have the potential to provide hitherto unavailable data on the history of detachment and transport of individual clasts and their transport velocities. This possibility derives from the close relationship between cosmogenic isotope production rates and elevation, and the strong dependence of detrital cosmogenic isotope concentrations on the duration of transport and storage of the sediment (i.e. the grain 'velocity' through a catchment). The simulation of cosmogenic isotope production in individual clasts as they are tracked through a catchment in a surface process model will enable the development of individual cosmogenic isotope concentrations and probabilistic approaches to the 'raw' cosmogenic isotope concentration data. Such a probabilistic approach means that multiple determinations of detrital cosmogenic isotope concentrations should offer the possibility of reconstructing drainage basin geomorphology and grain velocities from individual clast concentrations (Codilean *et al.*, 2006b). This approach will mean that average concentrations will provide average erosion rates, with PDFs of individual clast concentrations providing valuable data on catchment geomorphology. It is even tempting to suggest that such PDFs of individual clast concentrations from shallow basin sediments may provide data on the geomorphology of the sediment's source area, complementing the traditional measures of sedimentological textural and compositional maturity.

Returning to the fuller specification of the geomorphological processes in surface process models: this fuller specification will permit the development of sophisticated numerical models that are more process based, and include stochastic rainfall-runoff events and grain-size-based variations in sediment transport and storage. Numerical modelling and quantitative observation point to the reality of steady-state, time-independent landscapes, albeit that these landscapes' slopes are controlled not by regolith properties, as Hack (1960) suggested, but by the properties of the bedrock (Adams, 1985; Schmidt and Montgomery, 1995; Burbank *et al.*, 1996; Tippett and Hovius, 2000). Equally, numerical modelling indicates that Davis's geographical cycle is an appropriate description of long-term landscape evolution (Kooi and Beaumont, 1996). Thus, both the Davis and the Hack approaches appear to be applicable, but in different tectonic contexts - Hack for ongoing rock uplift with high stream discharges, and Davis for post-orogenic settings (in effect, just as Davis had posited). Of course, the dependency on slope of both the fluvial and hillslope components of these numerical models inevitably means that a Davisian-type of landscape evolution is 'confirmed' by the models, but the very fact that the 'Davisian' question is posed and addressed by Kooi and Beaumont (1996) means that we are yet to pass into a new post-Davisian paradigm, despite the extensive literature that suggests the opposite. Moreover, even if the geographical cycle was abandoned (I am suggesting it was not), this abandonment was not a paradigm shift, because when the geographical cycle was 'abandoned' it was not replaced by a new paradigm, but just a new approach to gathering data, the so-called quantitative revolution (Sherman, 1996). A feature of the Davisian scheme was the difficulty of testing it (e.g., Tarr, 1898; Shaler, 1899; Chorley, 1965; Bishop, 1980), and hypothesis testing remains a fundamental concern in long-term landscape evolution, not least because, as length- and timescales increase, so does contingency, the particularities of each place's history (Church, 1996). It remains a constant challenge to formulate adequate tests for numerical models but the co-application of multiple low-temperature thermochronometers along with cosmogenic isotope analyses offers many opportunities for increasing the adequacy of model testing. Moreover, as length scales increase to the sub-continental scales of, for example, high-elevation passive continental margins, the Earth gives us fewer individuals to research. Landscape evolution research then approaches historical

narratives about these individuals (Simpson, 1963; Kitts, 1963; Frodeman, 1995; Church, 1996; Bishop, 1998).

Dewey and Bird (1970) highlighted the plate tectonics significance of the relationship between high elevation and plate convergence 35 years ago, but much of the detail is still required. Indeed, John Dewey in his reflections on the plate tectonic revolution has observed 'Perhaps the most pressing problem that remains in tectonics, and one that we scarcely understand, is the tectonic and structural evolution of plate boundary zones, particularly in relation to topography. How exactly do mountains grow in areas of crustal compression? What controls the rate of deformation and uplift? A great challenge now is to use seismic data . . . to determine the distribution of strain along plate margins and relate those quantitatively to the growth of topography' (Dewey, 2001, p. 238).

Dewey (2001) was probably not thinking precisely of the issues that are our focus here, but his message remains valid. Summerfield (2005) recently noted that 'a few years back it would have been difficult to imagine *Nature* publishing three geomorphological papers in a single issue (Burbank *et al.*, 2003; Dadson *et al.*, 2003; Reiners *et al.*, 2003)' (p. 779). These papers relate to macroscale processes and landforms, and signal, a half-century after Strahler's call to abandon descriptive, qualitative approaches to the science of geomorphology, a return to the macroscale questions that the discipline was posing a little more than a century ago when Gilbert and Davis were leading figures in geomorphology. The re-emergence of geomorphology at the large spatial scale and long timescale has been made possible by the marriage of the approaches that Strahler (1952) advocated and the questions that Davis (1899) (and Du Toit (1937), Penck (1953), King (1962) and others) asked. We are at an exciting time in the science of Earth surface processes and landforms.

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