Introduction

The unrelenting competition between tectonic processes that tend to build topography and surface processes that tend to tear them down represents the core of tectonic geomorphology. Anyone interested in the Earth’s surface has wondered why it has the shape it does and what forces are responsible for that shape. For more than a century, this natural curiosity has inspired numerous conceptual models of landscape evolution under varied tectonic and climatic regimes. In the past, our ability to assign reliable ages to geomorphic and tectonic features was commonly very limited. In the absence of a chronological framework, testing competing concepts of landscape evolution was nearly impossible. As a consequence, these unquantified models were often viewed skeptically and treated as speculative notions.

During the past few decades, innovative applications of new techniques for determining the ages of landscape features, for assessing the mechanisms and rates of geomorphic processes, and for defining rates of crustal movement, have helped revitalize the field of tectonic geomorphology. It is now possible to measure at the scale of millimeters how rapidly a given site is moving with respect to another and how those rates of relative convergence or divergence are partitioned among various faults and folds. Similarly, we can now quantify how rapidly rivers and glaciers are incising into bedrock and the rates at which landslides are stripping mountain slopes. Clearly, the merger of such data sets can underpin a new understanding of the balance between the rates at which crustal material is added at a given site and the rates at which this material is eroded away. Defining this balance and interpreting the landscape that results from this competition represents a major component of modern tectonic geomorphology.

One of the remarkable attributes of tectonic geomorphology is the breadth of fields it often encompasses. Few other geological fields easily blend as disparate topics as seismology, Quaternary climate change, geochronology, structure, geodesy, and geomorphology. In fact, such breadth makes this field particularly exciting, as new data and ideas emerge from fields as diverse as paleobotany and fault mechanics. This diversity also presents a formidable challenge, because successful studies commonly require blending of appropriate data from specialized fields that were traditionally considered to be unrelated. Certainly, specialists in well-established disciplines like structural geology or stratigraphy often make important contributions to tectonic geomorphology. Fundamental leaps in our understanding, however, more commonly emerge from an integration across several disciplines.

This book is intended primarily for readers who are familiar with basic geomorphological and structural concepts and terms. Although we...
define many terms and review basic concepts and models throughout the book, we assume that our audience understands many commonly used geomorphological and structural terms, such as base level or conjugate faults. For those readers without such a background, occasional reference to a basic geomorphological or structural text will probably be helpful.

It is beyond the scope of this book to cover the contributions of each of the subdisciplines of tectonic geomorphology. Instead, we focus on some of the key tools, approaches, and concepts that have served to advance tectonic geomorphological studies during the past few decades. Several building blocks underpin many such studies, including a knowledge of how the Earth deforms both during earthquakes and between them, what sorts of features can be used to track deformation in the past, and what types of techniques are useful for dating features of interest.

Initially, we introduce geomorphic markers: landscape features, such as marine or fluvial terraces, that can be used to track deformation. Prior to deformation, the surfaces of these markers represent planar or linear features with a known geometry that can be predictably tracked across the landscape. Faulting or folding can subsequently deform such markers. Documentation of any departure from their unperturbed shape can serve to define the magnitude of deformation. Consequently, the recognition and measurement of such displaced or deformed markers is critical to many tectonic-geomorphic studies. These deformed markers are the raw data that cry out for interpretation.

In order to delineate rates of deformation, both the timing and amount of deformation must be defined. Fortunately, dozens of techniques have been used successfully to define the age of displaced features. In a chapter on dating, we present examples of some of the more commonly used techniques, and we try to convey a sense of the situations in which use of each technique would be appropriate. For each technique, we provide its conceptual underpinnings, discuss some of its limitations, and describe what data are actually collected in the field.

The oft-used term “tectonic processes” is a grab-bag expression that encompasses all types of deformation, including the motion of tectonic plates, slip on individual faults, ductile deformation, and isostatic processes. We concern ourselves here primarily with those processes that are most relevant to relatively localized deformation of the Earth's surface. In this book, we have generally chosen to ignore deformation and surface processes related to volcanism. Many of the concepts that are developed here are descriptive of landscape responses to volcanic processes, but the way in which material is added to the system (sometimes from above or only from below), as well as the time and spatial scales, are commonly different between volcanic and non-volcanic settings.

Because the accumulated movements of individual faults and folds have built many landscapes, we describe some current concepts from seismology and structural geology concerning coseismic rupture, the scaling of fault slip, fault-related folding, and geometries of deformation in different tectonic settings (compressional, extensional, and strike-slip environments). This deformation generates the fundamental topographic morphology on which erosive forces act. Much has been learned about ongoing deformation of the Earth's surface through geodetic studies: detailed surveys that delineate regional to local crustal displacements, often at time scales of a few years. During the past two decades, the increasingly widespread application of new techniques, such as the Global Positioning System (GPS) and radar interferometry, has yielded spectacular new images of current crustal motions. In a chapter on geodesy, we describe these techniques and some of the insights gained from them. In many ways, it is in the context of this greatly improved understanding of modern deformation that we now can make the largest strides in interpreting past deformation and landscape evolution.

Knowledge of current rates of crustal deformation leads to a host of provocative (and socially relevant) questions: Have these rates been steady over time? Do individual faults rupture at regular intervals and produce earthquakes of similar magnitude through time? Do groups of
faults accelerate synchronously at the expense of other faults? How large are the earthquake ruptures that accommodate the stress that is built by the incessant deformation of the crust? In a chapter on paleoseismology, we discuss some of the techniques for delineating the past behavior of faults, defining recurrence intervals and paleoslip rates, and assessing whether earthquakes occur with temporal regularity or randomness or in clusters.

Naturally, discovering ways to measure the amount and rate of erosional losses from an area and to define changes in the height of the surface of that same area through time lies at the heart of many tectonic-geomorphic studies. Armed with new tools for dating and with digital topographic databases, we can now address these topics more accurately than in the past. In a chapter on erosion and uplift, we describe several strategies for quantification of rates of erosion, surface uplift, and rock uplift, and we illustrate how these can be synthesized to examine the balance between building and tearing down of topography.

In the latter half of this book, we step away from a focus on the key building blocks and instead introduce concepts and examples of landscape evolution and tectonic interpretations at different time scales. Moreover, rather than examine specific geomorphic environments or topics, such as the ways in which rivers and fluvial features can provide insights on tectonic processes, we focus on more integrated studies. The key issues can be encompassed by two questions: “What information from the Earth’s surface improves our understanding of the nature of the interactions between tectonics and geomorphological processes?” and “How do we interpret preserved geomorphic features in order to reveal rates and patterns of tectonic deformation in the past?” The more recent record is often best suited to answer the first question, whereas great value comes from studies that are able to answer the second question for intervals in the distant past.

There is a natural, time-dependent progression in the nature of tectonic-geomorphic landscapes and the insights that can be gained from them. Perhaps surprisingly, part of this progression is dictated by past climate changes (Fig. 1.1).

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**Fig. 1.1** Schematic illustration of climate change scaled at orders of magnitude (Holocene, one glacial cycle, Middle–Late Pleistocene).

The relative stability of the Holocene climate is atypical of a full glacial–interglacial, 100-kyr cycle. Similarly, the sawtooth pattern of change during past glacial–interglacial cycles does not show oscillations of similar magnitude, as is typical of the past million years.

At Holocene time scales, we are confined to a post-glacial era in which climatic conditions have varied relatively little. At time scales of 100 000 years or more, complete or multiple glacial–interglacial cycles have occurred.
Large-scale changes in sea level, the size of glaciers, and the discharge of rivers at these time scales leave strong imprints on the landscape. These climate cycles are often the “drivers” that generate the geomorphic markers, in the form of moraines or marine and fluvial terraces, that are so important in the field of tectonic geomorphology. At time scales of a million years or more, numerous climatic cycles can create a more “time-averaged” landscape.

These different time scales also often create a natural segregation of the focus of tectonic-geomorphic studies. At Holocene time scales, cause and effect can be examined rather directly. For example, if faulting causes the land surface to tilt in a given direction, the timing and rate of diversion of rivers in response to that tilt can be measured. Present rates of crustal deformation can be compared against the sediment load of rivers. Commonly, geomorphic markers, such as river terraces, are almost untouched by erosion and can function as pristine recorders of Holocene deformation. The strength of Holocene studies is that the record is often most complete, the dating is most precise, rates of many competing processes can be directly measured, and their interactions can be examined. At least three potential disadvantages can limit Holocene studies. First, the rates of tectonic processes may be sufficiently slow or the occurrence of deformational events, such as earthquakes, may be sufficiently rare that the tectonic signal is obscure. Second, rates of geomorphic processes during Holocene (interglacial) times may not be very representative of long-term rates. Third, the time it takes for geomorphic systems to respond to a change imposed by tectonic forcing can be longer than the Holocene. Thus, the geomorphic system may be in a state of transition with respect to tectonic perturbations imposed upon it.

As described in subsequent chapters, the limited precision of most dating techniques means that, as one delves beyond the Holocene and farther into the past, it becomes progressively more challenging to associate specific events in time. Thus, it can be difficult to define direct responses to individual forcing events. On the other hand, major climate changes often have created robust geomorphic markers, such as river terraces, that persist as recognizable features and provide a lengthy time framework within which to document patterns of deformation. At short time scales, tectonic forcing is commonly unsteady, because it results from discrete events, such as earthquakes, that are widely separated in time. At time scales of more than $10^4$ yr, this unsteadiness is commonly smoothed out, and average rates of deformation can be defined. Because many geomorphic markers persist in the landscape at time scales of single glacial–interglacial cycles, this is an ideal interval within which to document past rates of tectonic forcing. At still longer time scales, erosion has typically removed most markers that are more than about $10^5$ yr old. Yet, it is at this time scale that large-scale landscape responses to sustained tectonic forcing can be clearly seen. Typically, landscapes must be treated at a coarse spatial scale in order to examine the products of tectonic-geomorphic interactions, such as the topographic characteristics of a collisional mountain belt or the broad swath of deformation that occurs as a continent passes over a hot spot.

We conclude with a chapter on numerical modeling of tectonically active landscapes. In contrast to efforts aimed at directly measuring tectonic-geomorphic processes, the interactions between deformation and surface processes can also be studied theoretically. If we could write numerical rules that represent phenomena such as the displacement of the crust due to faulting or the erosion and redistribution of mass due to surface processes, we could proceed to investigate interactions among these processes. Consider, for example, displacement on a normal fault that bounds the front of a mountain range. The topographic offset will change the local gradient of any river crossing the fault, and a numerical rule for river incision could predict how that reach of the river would respond and how that response would be propagated upstream.

Rather complex models for landscape evolution in different tectonic environments have recently been formulated. It is not our intent to describe or compare these in detail. Instead, we describe several of the basic building blocks that could go into a numerical model, and we illustrate some of the predictions of these models.
In our view, the usefulness of numerical models is not in creating a reproduction of some actual landscape. Instead, models provide a means to explore potential interactions within a landscape, so that we have a firmer basis for understanding how variables as diverse as crustal rigidity, susceptibility of bedrock to landslides, and the distribution of precipitation may interact with each other, perhaps in unexpected ways. Such models are meant to develop our insight into the complex interactions among the processes, and they can serve to point toward measurements one might make in the field that would most efficiently constrain the process rates.

**Energetics**

Energy drives the interactions between tectonics and surface processes. In order to build topography, work must be done against gravity. The energy needed to accomplish this work comes ultimately from the conversion of a small fraction of the energy involved in the horizontal motions of the lithospheric plates that constitute the more rigid exterior of the planet. The energy driving plate tectonics comes from primordial heat associated with building of the planet, from the decay of radioisotopes, and from phase changes in the interior of the Earth.

It is perhaps surprising to cast the energy expenditure represented by plate motion in everyday terms. Consider, for example, a simple calculation of the kinetic energy of that part of the Indian Plate lying south of where it is colliding with mainland Asia and building the Himalaya. We may specify its approximate dimensions as 3000 km wide by 7000 km long by 50 km thick. If we assume a mean density of 3000 kg/m³, this yields a mass of approximately $3 \times 10^{21}$ kg. Taking a mean plate velocity of 5 cm/yr yields a kinetic energy ($\frac{1}{2}mv^2$) for the plate of about $4 \times 10^{23}$ J (joules), equivalent to about 1/200 of a “Snickers” bar ($10^8$ J)! On the other hand, the rate of energy expenditure required to move the Indian Plate (force × distance/second) is equivalent to approximately $10^{13}$ J/s or $10^7$ “Snickers” bars per second! Over the past 50 million years, some tiny fraction (~1/100 of 1%) of that energy has been expended (or converted, really, to potential energy) in elevating the crust of the Himalaya and Tibet far above the geoid and thereby creating the world's largest topographic anomaly, averaging about 5 km above sea level.

The energy for driving surface processes derives from a combination of gravitational potential energy and solar energy. All rocks that have been elevated above the geoid have a potential energy equivalent to the product of their mass, gravitational acceleration, and their height above the geoid ($PE = mgh$). At the top of the atmosphere, the Earth receives solar energy equivalent to about 1300 W/m². The solar energy received at the Earth's surface is several orders of magnitude greater than the energy that leaks out from the Earth's internal heat engine (~40 mW/m²). Solar energy evaporates water and heats the air that carries the moisture with it as the air rises, expands, and cools. The water attains its maximum potential energy at the top of its atmospheric trajectory. When the vapor condenses and falls as precipitation, it converts some of its potential energy to kinetic energy that it delivers to the surface of the Earth with the force of its impact. The potential energy that is represented by the mass of water at some elevation above the geoid is then available to be expended doing geomorphic work or to be lost through heat dissipation or frictional processes. Solar energy is also an important factor in chemical weathering processes, especially in any temperature-dependent reactions, as well as in mechanical weathering processes, such as freeze-thaw cycles.

**Active tectonics and models of landscape development**

Sharp contrasts in the appearances of landscapes in a given climatic or tectonic regime inspired geologists in the past to devise schemes to explain those contrasts. One of the most prominent such geologists was William Morris Davis, who in the late 1800s and early 1900s developed the well-known geomorphic models (Davis, 1899) showing a progression from “youth” to “maturity” to “old age” (Fig. 1.2). Living in the wake of the revolutionary ideas of
Charles Darwin, it was perhaps natural to conceive of landscapes as evolving from one stage to the next. In Davis's models, tectonic forcing is an impulsive phenomenon that occurs at the beginning of the “geomorphic cycle.” All of the building of topography occurs at the start of the cycle (Fig. 1.2). Subsequently, geomorphic processes attack and degrade the topography, with the end-result predicted to be a peneplain. This view of landscape evolution prevailed through much of the 1900s and is still widely cited by introductory geology textbooks.

A new theory of landscape development that opposed Davis's models was promoted by Walther Penck in the 1950s (Penck, 1953). Rather than having all of the tectonic deformation at the beginning of a cycle, Penck suggested a more wave-like pattern of tectonic forcing through time. In this scheme, the magnitude of deformation gradually increases toward a climax and then slowly wanes away. Instead of calling upon an impulsive building of topography followed by long intervals of erosion, Penck's model invoked steadily increasing rates of deformation that would accelerate rock uplift and gradually build the topography toward a maximum state of topographic relief. Geomorphic processes were conceived as attacking the uplifting region throughout the period of mountain building, so that the resulting landscape could be interpreted as a product of this competition between deformation and erosion. As the

Fig. 1.2 Classical models of tectonic forcing and landscape responses. Based on the landscape response theories of (top) Davis (1899), (middle) Penck (1953), and (bottom) Hack (1975), each panel is characterized by differences in the duration and rate of tectonic forcing (rock uplift) and by the topographic response engendered by these different styles of rock uplift. Modified after Summerfield (1991).
rate of mountain building waned, erosion was proposed to overtake deformation rates, thereby causing a gradual reduction in the residual topography (see Fig. 1.2).

A third approach to landscape development was proposed by Hack (1975), who suggested that, when rates of deformation and rates of erosion are sustained for long intervals, landscapes will come into a sort of balance or dynamic equilibrium. Given the finite strength of rocks,Hack realized that topography could not increase without limit, even if rates of tectonic forcing persisted for very long periods. At some point, as topography grew and grew, the relief on hillslopes would create forces exceeding the rock strength, and they would collapse. With continued uplift of the bedrock, additional slope failures would limit the height that the topography could attain. Eventually, the topography would enter into a rough steady state or a dynamic equilibrium. Consequently, in this model, rates of mountain building and rates of erosion would come into a long-term balance (see Fig. 1.2). Unlike the models of Davis or Penck, there is no need for rates of deformation to become negligible, or even to wane at all. Similarly, after attaining a maximum sustained topography, there is no need for the landscape to “evolve.” Instead, it simply fluctuates around an equilibrium topography until such time as the rates of tectonic forcing change.

One might wonder whether any of these theories are applicable to real landscapes. One way to consider this problem is to compare the time scales of deformation and of geomorphic responses. For example, an earthquake is a very impulsive event that “instantaneously” creates topographic change, say in the form of a fault scarp. If the time between earthquakes is long relative to the time it takes geomorphic processes to cause significant topographic evolution, the scarp will be degraded before the next earthquake. This topographic change could be thought of as a small-scale version of a Davisian scheme for landscape evolution. On the other hand, if the recurrence interval between earthquakes is short, then the scarp will be continuously refreshed and will not degrade in a predictable fashion toward a low-relief surface. In fact, some balance might be expected to develop between the renewal of the scarp by faulting and erosion of it during interseismic periods. Thus, even though the total displacement grows through time, the scarp itself might attain a “steady-state” shape.

It is also useful to consider the response times or inertia of different elements in a geomorphic system. Consider, for example, a drainage basin in a mountain belt (Fig. 1.3). The basin consists of river channels, hillslopes, the crests of interfluves, and the drainage divide that defines the shape of the catchment. Some of these elements will respond more rapidly to changes imposed on them than will others. Suppose that, near its outlet from the mountains, this basin is bounded by a normal fault. How will the various basin elements respond to displacement on that fault? The channel that crosses the fault will “see” the displacement first. The local base level will have fallen abruptly, causing the river gradient to steepen locally and erosion of the channel to intensify. At this point, all other positions in the catchment will have been undisturbed by the faulting event, so the faulting elicits no initial geomorphic response. Subsequently, a wave of erosion of the channel bed can be expected to

Fig. 1.3 Geomorphic elements of a drainage basin. Diverse geomorphic components of a catchment have a hierarchy of inertia to change, with the channel network as most responsive, and the drainage divide as least responsive.
propagate up the fluvial system. As this channel erosion passes the base of an adjacent hillslope, it causes the hillslope gradient to increase and will initiate a change in the rate of hillslope processes. These hillslope processes could eventually affect the interfluvies and finally the shape of the entire catchment. Large geomorphic elements, such as the catchment shape, are unlikely to be affected by individual seismic events. Thus, a clear hierarchy of response times exists (rivers respond sooner than hillslopes, etc.), and an analogous hierarchy of topographic inertia is evident (catchment shape is resistant to change, whereas river gradients are susceptible to small perturbations).

Consideration of the scaling of elements in a given geomorphic system, of response times or inertia of those elements with respect to imposed changes, and of the rates, magnitude, and duration of different styles of tectonic forcing suggests a way in which the apparently incompatible landscape evolution concepts of Davis, Penck, and Hack can be reconciled. In fact, recent numerical models of tectonically perturbed landscapes have explicitly addressed this problem. Using a surface-process model that links channel incision, sediment transport, and hillslope erosion, Kooi and Beaumont (1996) developed a model that predicts a lag between the onset of deformation and the response of the geomorphic system to that deformation (Fig. 1.4). The overall response of the geomorphic system and the magnitude of the lag depend strongly on the nature of the tectonic forcing. If the forcing is impulsive (à la Davis), then the topography is rapidly created and simply degrades through time. If the deformation increases to a maximum through time and then wanes (à la Penck), topography gradually builds in the face of progressively increasing rates of erosion. The maximum topographic expression occurs slightly after the rate of deformation begins to wane, because the rock uplift still outpaces the rate of erosion. Finally, in the latter half of the cycle, the topography wanes, despite gradually diminishing rates of erosion. If the tectonic forcing is continuously sustained, then the Kooi and Beaumont model predicts that, after an initial interval of building of topography, rates of rock uplift and erosion will become balanced (à la Hack), and the topography will attain a persistent dynamic equilibrium (Fig. 1.4). A change in the rate of tectonic forcing would push the system toward a new equilibrium, whereas cessation of deformation would return the system to an almost Davisian state in which the topography is systematically degraded.

\[ \begin{align*}
\text{Impulsive} & \quad \text{Process-Response Models} \\
\text{Varying} & \\
\text{Sustained} & \\
\end{align*} \]

Fig. 1.4  Tectonic versus sediment flux process–response models.

Duration and magnitude of rock uplift (the vertical tectonic flux) are compared with the erosional sediment flux from uplifted mountains. The three scenarios (impulsive, varying, sustained) are analogous to the models of Davis, Penck, and Hack (Fig. 1.2). Note that the time lag between tectonic forcing and sediment response is variable. Modified after Kooi and Beaumont (1996).
These predictions are the output of a numerical model that represents a great simplification of both tectonic and geomorphic processes. In fact, the physics of many of these processes are still poorly understood. Nonetheless, it is satisfying to think that the thoughtful, yet apparently contradictory, landscape development models put forth during the past 100 years are indeed reconcilable and that simple numerical models can help enhance our understanding of tectonically active landscapes.

The new world

Over the past two decades, technological advances have dramatically improved our view of the Earth. As the diversity and availability of digital topography has expanded, our ability to visualize the shape of the Earth’s surface has become much easier and more accurate. For 11 days in February 2000, NASA’s Space Shuttle used an active radar system to map the topography of the world between 60°N and 60°S (Farr et al., 2007). The digital elevation model (DEM) derived from this mission now provides almost complete elevation coverage with a spatial resolution of 90 m. Still more recently, a higher-resolution, 30-m DEM that covers the world between 83°N and 83°S has been developed from satellite stereoimages and is freely available (http://www.gdem.aster.ersdac.or.jp/). The success of Google Earth in merging such topographic data with remote sensing images has provided tectonic geomorphologists with an unprecedented opportunity to explore the Earth’s surface. As high-resolution (≤1 m) imagery is increasingly incorporated into Google Earth, individual fault scarps, uplifted marine terraces, and channels on actively growing folds in previously inaccessible areas can now be visually explored from a computer almost anywhere in the world!

In the temperate mid-latitudes, however, where forests blanket much of the landscape, even high-resolution imagery does not commonly permit a clear view of the actual land surface. But, this restriction is also changing. A new technology, lidar (light detection and ranging), which is also known as laser scanning (Carter et al., 2007), uses concentrated pulses of light emitted from an airborne instrument to penetrate through openings in a forest canopy. By measuring the return time and direction of the light, the vector to the ground can be calculated from the “last returns” (those that were not reflected by above-ground vegetation). Via an integration of all of the last returns and with precise knowledge of the position of the aircraft, a bare-Earth topographic image of the Earth’s surface can be created. The resolution of the image depends on how high the aircraft is flying and the nature of the light beam, but commonly DEMs are created with 1-m spatial resolution and topographic uncertainties of a few centimeters. The stunning success of lidar in revealing previously unknown fault scarps beneath the canopy of dense northwestern US forests (Fig. 1.5) made converts of skeptics and launched widespread efforts to acquire lidar data over diverse tectonic and geomorphic targets. Tripod-mounted laser scanners are also gaining increasing popularity. Based on the same physical principles, these instruments permit topographic reconstructions with millimeter- to centimeter-scale uncertainties to be created of individual hillslopes, fault scarps, or river beds; and, with a series of topographic scenes over time, the details of landscape changes can be recorded and quantified. The vastly improved accuracy of these new topographic views, ranging from global topography to individual hillslopes, has underpinned a new view of the Earth and the processes that mold its surface.

Some modern controversies

At present, many lively controversies animate studies in tectonic geomorphology. Some of these provide an interesting backdrop for reading subsequent chapters of this book. For example, the history of Cenozoic cooling of the rocks within many mountain ranges has been interpreted to suggest that the ranges experienced accelerated rates of uplift during late Cenozoic times. With increased recognition that this apparent increase was approximately coeval with the onset of the “Ice Ages,” it was commonly
believed that mountain uplift led to cooling and helped to precipitate the Ice Ages. In 1990, Molnar and England challenged this entrenched idea and suggested nearly the opposite: as climate changed to more glacial conditions in late Cenozoic times, enhanced rates of erosion within mountain belts caused increased rates of valley incision, which in turn incited isostatic uplift of the residual peaks (Molnar and England, 1990). Although isostatic uplift can produce ranges with higher summits, it does not implicate tectonic uplift as the cause of climate change.

So, how do we tell whether the climate caused uplift of the summits or whether surface uplift of the ranges caused changes in climate? Potential resolutions to this quandary require many ingredients, including the nature and magnitude of changes in mean elevation of mountains, in rates of erosion, in climate, and in elevations of summits and valley bottoms. Moreover, we would like to know when and how rapidly changes took place. If you think carefully about any one of these ingredients, you quickly realize the reason why tectonic geomorphological studies are often interdisciplinary in nature. Consider, for example, the concept of mean elevation of a range. Using digital topographic data, the current mean elevation is straightforward to calculate. But how do you determine mean elevations in the past? As described in subsequent chapters, approaches to estimating former altitudes range from paleobotanical to isotopic studies. Or, consider the effects of changes in climate. During Ice Age times, was there more precipitation or less? Did the expansion of glaciers lead to enhanced rates of erosion? There is intense interest in, and considerable argument about, whether glaciers are effective agents of erosion in comparison to rivers. The controversy has spawned a flood of recent research into the physical and
chemical processes involved in erosion of bedrock by both fluvial and glacial processes. Can enhanced erosion drive uplift of mountain summits? Yes, but only if there is an increase in topographic relief associated with the erosion. In essence, there has to be less erosion of the summits than there is of the valley bottoms. Think about an ice cube floating in a drink (Fig. 1.6). Its upper surface is at a mean elevation equivalent to \(-1/10\)th of its total thickness, equivalent to \(1 - \frac{\rho_{\text{ice}}}{\rho_{\text{water}}}\), where \(\rho = \) density. If you were to cut canyons into the upper surface of the cube, the mean elevation of the cube's surface would decrease and the base of the cube would bob upward in the water in order to maintain the isostatic balance (maintaining \(1/10\)th of the total mass above the water surface). If you could cut the canyons without “eroding” other parts of the upper surface of the cube, these remnants would actually rise higher than their original height in response to the lowering of the mean elevation.

Although melting of an ice cube demonstrates that peak uplift can occur due to enhanced erosion, has this commonly occurred in the past, is it related to a more “erosive” climate, and what are the magnitudes of the uplift of peaks involved? The common view has been that increased rates of erosion tend to increase topographic relief and, therefore, would promote the uplift of summits. On the other hand, theoretical studies of river profiles suggest that increased erosive power causes river gradients to decrease (Whipple, 2004). If this is true, then, in order to generate an increase in topographic relief in the landscape, hillslopes would have to be concurrently lengthened and steepened. Tectonic-geomorphic studies that document the temporal evolution of both the valley bottoms and the adjacent hillslopes are needed to resolve these issues.

Another modern controversy in tectonic geomorphology revolves around earthquake prediction. Over the past two decades, a substantial research effort has focused on answering such questions as: “Which fault is most likely to rupture in the next large earthquake?”, “When is that event likely to occur?”, and “How large an earthquake can we expect to occur?” Many scientists would maintain that the best way to make such predictions is to understand the past history of faulting. Such topics fall into the realm of paleoseismology with its focus on the reconstruction of past earthquakes in terms of their distribution in space and time, coseismic displacements, and interactions among faults.
If paleoseismological studies were to indicate that successive ruptures of a particular fault occurred at regularly spaced intervals in time, it would then be possible to define a “recurrence interval.” If we also know when the last rupture occurred, we could predict when the next earthquake was likely to happen. But, do faults display regularity in their rupture histories? The known paleoseismological record commonly extends back through only a few earthquakes, making it difficult to define a statistically significant recurrence interval.

One popular model among paleoseismologists suggests that many faults can be said to experience “characteristic earthquakes,” implying that each successive rupture mimics the displacement of the previous event. If true, this implies that, along any particular section of a fault, the amount of displacement that results from a single earthquake will be duplicated in subsequent ruptures (Fig. 1.7). Imagine the predictive power that would result if faults were shown to generate both characteristic earthquakes and to have predictable recurrence intervals! City planners and paleoseismologists would all be delighted. For most faults, however, this seductively attractive concept has yet to be demonstrated. Consequently, much current debate revolves around whether or not certain faults display characteristic earthquakes and predictable recurrence intervals. Or, could some faults display a very irregular behavior in both the time between ruptures and displacement variations in successive events? Far from being resolved, this debate lies at the core of paleoseismology, and has spawned much recent work both in the field and in numerical models of the faulting process.
Many paleoseismological studies rely on trenches that are dug across faults to reveal the slip in past earthquakes. Measurements of displaced strata serve to define the magnitude of displacement in an earthquake, whereas dating of those strata can determine when the rupture occurred. But do such measurements capture the full slip history? Might they overestimate the slip? Several recent discoveries complicate the interpretation of trenches. As faults approach the Earth's surface, slip commonly decreases, so trench exposures at the surface may tend to underestimate the total coseismic slip. This underestimate is especially likely for “blind” thrust faults that do not rupture the surface. Even where faults do break the surface, is it possible that significant strain occurs as diffuse deformation between faults? Although difficult to quantify, such broadly distributed strain may represent a significant fraction of the total deformation. In such conditions, displacement in trenches would be underestimates of the integrated deformation across an area. The advent of extensive geodetic networks (mostly relying on GPS) has provided new insights on slip during and between earthquakes. Recent recognition of “slow earthquakes” that occur over hours to days (Heki et al., 1997) indicates that slip during an earthquake may be equaled by slow slip of nearly equal magnitude. If such slip were propagated to the surface, it would be impossible to recognize as a slow slip event in an exposure within a paleoseismic trench.

The rates of convergence between tectonic plates, as well as measured rates of local deformation, indicate that, in many active mountain ranges, rocks are moving upward with respect to sea level at rates of several millimeters per year. Recall that a rate of 1 mm/yr is equivalent to 1 km in a million years. Thus, in the absence of erosion, vertical rock uplift rates of several mm/yr would build very high mountains in only a few million years. Clearly, mountains do not grow indefinitely: only 14 peaks poke more than 8 km above sea level. But what controls their ultimate height? Is there a limit to the energy available to lift the mass of rock and increase its potential energy? Does rock strength set the height limits, or do changing rates of erosion determine the topography of ranges? Such questions lie at the core of another current controversy in tectonic geomorphology: Can the concept of “dynamic equilibrium” be applied in active orogens at the mountain range scale? Dynamic equilibrium implies that, on average over time, the landscape maintains a steady-state form, whereby the height of the summits, the steepness of the valley walls, and the topographic relief fluctuate around long-term mean values. If the mean height of the mountains stays the same through time, this persistence implies that rates of rock uplift (vertical movement of rocks with respect to sea level or the geoid) are balanced by rates of erosion. How is this accomplished? Are surface processes capable of eroding at several mm/yr? Which processes are responsible (river erosion, landsliding, glacial erosion, conversion of rock to soil) and do these processes operate in different ways in different mountain belts? Or, is the traditional idea correct that rapid rates of rock uplift are commonly compensated, not by geomorphic agents at all, but by events of tectonic denudation (extensional faulting) that efficiently lower the regional height of the landscape?

In order to answer such questions, we have to be able to document rates of modern and past erosion, to quantify rates of rock uplift and changes in the mean elevation and topographic relief of mountain ranges (Fig. 1.8), and to document the role of extension within orogens. Also, because climate undergoes large glacial–interglacial fluctuations at 100,000-year intervals, rates of erosion that are responsive to climate (most erosional processes are) will also vary strongly through time. Consequently, when thinking about the problem of dynamic equilibrium and steady-state topography, it is most useful to consider time spans that exceed a full glacial–interglacial cycle, so that average rates can be determined.

Such constraints present some great challenges to researchers in tectonic geomorphology. It is inadequate to document only modern rates (which are difficult enough to measure accurately!). Rates from intervals throughout a climate cycle or rates that integrate an entire
cycle are needed. This integration requires clever ways to measure quantities of material removed from or added to a landscape. Moreover, researchers must somehow inject a reliable “clock” into the rock record, because rate calculations can be no more precise than the time interval across which they are measured. Such measurements are difficult to make. An alternative approach would be to determine the operational “rules” by which various surface processes erode the landscape, and to incorporate them into a theoretical model of landscape evolution. For example, how does the rate of bedrock erosion at the base of a glacier relate to the speed at which the glacier is sliding, its thickness, the steepness of its bed, freezing and thawing at the ice–rock interface, and the resistance of the bedrock beneath it? Only if we can both define these rate relationships and determine how glaciers have extended and retreated in the past, will we be able to model the mean rates of erosion within this portion of the landscape.
In recent years, provocative linkages between climate, erosion, and tectonics have been proposed. Even at the scale of orogens, the large-scale organization of major faults and rock uplift has been attributed to patterns of differential erosion that are in turn controlled by climate.

Underpinning these interpretations is the observation that rapid erosion both removes rock mass, thereby perturbing the stress field on the underlying rocks, and compresses the geothermal gradient, thereby heating and weakening the underlying rocks. Together, these

Fig. 1.9  Proposed climate–tectonic coupling in the Himalayan orogeny.
A. Geological observations and conceptual model that connect partial melting in Tibet and lower crustal flow to climatically driven erosion and major faults in the Himalaya. Monsoon-driven erosion along the southern flank of the Himalaya is proposed to weaken the underlying crust and to draw the channel flow toward the surface. GHS: Greater Himalayan Sequence; LHS: Lesser Himalayan Sequence; MCT: Main Central Thrust; MHT: Main Himalayan Thrust; STD: South Tibetan Detachment. B. In the absence of erosion, numerical modeling suggests simple outward lower crustal flow within a confined channel (tunneling). C. With intense erosion localized in a specific region, modeling suggests lower crustal flow is drawn upwards toward the zone of high erosion. Note that this geometry predicts relative motion along the channel's margins that is consistent with the slip on the major Himalayan faults (MCT and STD in A). Modified after Beaumont et al. (2001).
effects are interpreted to guide patterns of deformation and metamorphism.

Consider, for example, four attributes of the Himalayan–Tibetan orogen (Fig. 1.9A) and how they might be linked. First, thickening of the crust beneath the Tibetan Plateau appears to have caused partial melting of at least some of the lower crust (Nelson et al., 1996), which then becomes weak and tends to flow outward as a channel confined above and below by stronger rock. Second, the rocks of the Greater Himalaya are bounded by two large faults that have been active at the same time: below the Greater Himalayan rocks, the Main Central Thrust is an up-to-the-south thrust, whereas above them, the South Tibetan Detachment is a down-to-the-north normal fault. Third, every summer the Indian monsoon dumps about 4 m of rainfall on the southern flank of the Greater Himalaya, but less than 10% of that amount penetrates into Tibet. Across this decreasing rainfall gradient, current erosion rates also diminish by three- to four-fold (Gabet et al., 2008). Fourth, combining these observations, some geologists suggest that the intense, monsoon-driven erosion on the southern flank of the Himalaya weakened the underlying crust and “drew” the partial-melt channel toward the surface, thereby determining where the big faults bounding the channel are located. Numerical modeling of lower crustal flow tends to reinforce the importance of erosion (Beaumont et al., 2001, 2004). In the absence of erosion, lateral tunneling is predicted as channel flow remains confined within the lower crust (Fig. 1.9B), whereas when erosion is introduced, the flow migrates upward toward the region of high erosion (Fig. 1.9C). A striking aspect of these numerical models is that, if the pattern of rainfall were reversed so that rain was focused on the opposite side of the orogen, the models predict that the orientation of the large faults could also flip direction! Whereas more questions than answers remain about potential linkages between climate, erosion, and tectonics, their proposed interactions set an exciting stage for future exploration of the role of climate in mountain building.

**Looking ahead**

Resolutions to these controversies are beyond the scope of this book, but they provide a framework for thinking about many of the topics discussed in the subsequent chapters. These controversies illustrate some of the breadth of modern tectonic geomorphological studies. Any serious consideration of potential solutions to these controversies quickly reveals the interdisciplinary nature of the research required to address them. Although certainly not unique in its demands for interdisciplinary work, tectonic geomorphology attains much of its current vibrancy from the cross-pollination that is occurring between specialists of many disciplines who are coming together to address major unresolved issues. It is our intent that the following chapters provide some insight into the tools, approaches, and interpretational techniques that are currently used in tectonic geomorphological studies. We hope to convey the striking innovation and creativity of past researchers, upon whose shoulders future advances will be made.