Preface

Many people enjoy scenery, and see landscapes as beautiful. Landscapes have inspired painters and photographers, even poets and composers of music. What is it about scenery that inspires people? For some it is clearly related to a scene with a 'natural' vegetation cover, for others it might relate to a mood created by the weather, or atmospheric conditions. For most, the physical landscape (mountains, hills, rocks, rivers, the sea) is the basis of 'scenery'. Geomorphology is the science that deals with landforms and physical landscapes. As a study it lies between the traditional disciplines of physical geography and geology, draws from both, and contributes to both.

The purpose of this book is to introduce the reader to the science of geomorphology. The book is not intended as a textbook; there are many of these at every level. I make no bibliographic references in the text, but in the end pages I make suggestions for further reading. In writing I have tried to present a broad and reasonably comprehensive view of the conceptual basis of the subject. I have tried to avoid mathematical treatments and to keep the level of physics and chemistry to a minimum.

Geomorphology inevitably involves a great range of spatial scales, from the global scale (continents and mountain systems) to the regional scale (individual mountain and hill ranges, river basins) to the local scale (conventional scenery: rivers, hillslopes, beaches, glaciers) and the micro-scale (weathering phenomena, sedimentary details). Partly related to spatial scales are timescales (geological time – millions of years; the timescales of the ice ages – half a million to tens of thousands of years; modern timescales – the last ten thousand years; timescales of individual events, e.g. landslides, floods – hours or days). The form of the earth's surface, at all scales, results from the interplay between two sets of forces, though the relative importance of each varies with scale. The two sets of forces are internal (essentially geologically driven) and external (essentially climatically driven).

This book is organised from the 'top down', initially introducing the concepts related to spatial and temporal scales and the two main drivers of landform evolution (internal and external forces). Then the bulk of the material is organised by spatial scale, dealing first with global and regional scales, then with local and (to some extent) micro-scales. There are two final short chapters, one dealing with the integration of timescales and landscape evolution, the other dealing with interactions between human society and geomorphology.

Note: all terms highlighted in **bold** are defined in the Glossary at the end of the book.

Acknowledgements

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Introduction to geomorphology

In this chapter I introduce some fundamental concepts related to spatial and temporal scales of landforms and some basic material on the primary forces driving landform evolution: internal (geological) and external (climatic) forces.

1.1 What do we mean by 'landforms'?

Geomorphology is the scientific study of the landforms of the surface of the Earth. These forms encompass a range of scales from that of, for example, the Earth's major plains, plateaux and mountain ranges to that of small-scale forms, such as a beach or a river bank. Landforms of various scales are nested within one another so that within a mountain range, for example, there are individual mountain ridges and valley systems; within valley systems there are valley-side slopes and river channels; and within river channels there are gravel and sand bars. Studying landforms and the processes that create them inevitably involves study over a range of temporal and spatial scales. The processes that create the landforms include the creation of the relief itself and its modification by erosion and deposition. The temporal scales range from the short-term timescales at which some erosion processes operate, to the longer-term timescales of Earth history. The spatial scales range from large scales related to the global distribution of the relief forms of the land surface and the sea floor to local scales of, for example, individual hillslopes or river channels.

As a science, geomorphology lies between the traditional disciplines of physical geography, the study of the natural environment, and geology, the study of the solid Earth. As the Earth's surface forms part of the natural environment, geomorphology interacts with the other sciences that deal with environmental systems: climatology, hydrology, pedology (soil science) and ecology. It also interacts with several subdisciplines of geology, especially with tectonics and structural geology in relation to deformation of the Earth's crust, with sedimentology in relation to the properties of sediments, the products of erosion at the Earth's surface, and with stratigraphy, the account of Earth history.

This interface between geology and environment influences the spatial and temporal scales relevant to the study of geomorphology. In the simplest terms, internally driven (geological) forces generally tend to operate over large spatial and long time scales. They create the gross or **available relief** of the Earth's surface by deformation of the Earth's crust. Externally driven forces, ultimately controlled by climate, modify this surface by erosion and deposition. These external processes can be described as the 'sediment cascade' and they create the more detailed landforms that form the heart of the study of geomorphology.

1.2 What we mean by spatial scales

At global and continental scales geomorphology deals with the major features of the surface of the Earth (e.g. continents, mountain systems). At a regional scale it deals with intermediate forms (e.g. individual mountain and hill ranges, river basins). At a local scale it deals with what could be described as individual features of conventional 'scenery' (e.g. rivers, hillslopes, beaches, glaciers), and at a micro-scale with the detail of the surface itself and its constituent materials (e.g. weathering phenomena, sedimentary details).

This book is organised around these themes, with the main chapters devoted to global scales, regional scales, with local and micro-scales treated together. Figure 1.1 shows how different features are apparent at different scales. On the Google Earth satellite image of western Canada (Figure 1A), features related to global/continental scales are most apparent. The NNW–SSE alignment of the major structures of the Rocky Mountains dominate the west of the image. These structures relate to the plate tectonics context (*see* below) of the North American continental plate, convergent with the oceanic Pacific plate. In contrast, the Canadian Prairies present a much more uniform land surface, formed over a much more stable zone of the Earth's crust. In addition to the geological features the vegetation cover reflects major continental-scale climatic contrasts. The grassland of the southern Prairies contrasts with the forest cover further north and with the forested zones within the Rocky Mountains. Coming down a scale to that of a major valley within the Rockies (Figure 1B), the NNW–SSE alignments of the major mountain ridges and of the Saskatchewan and Bow river valleys are still apparent, but the mountain slopes dominate the image. Coming right down to the local scale (Figure 1C), the characteristics of the braided river channel dominate the photograph.

1.3 What we mean by temporal scales

The three spatial scales represented by the views shown in Figure 1.1 all relate to different timescales. The geological processes that

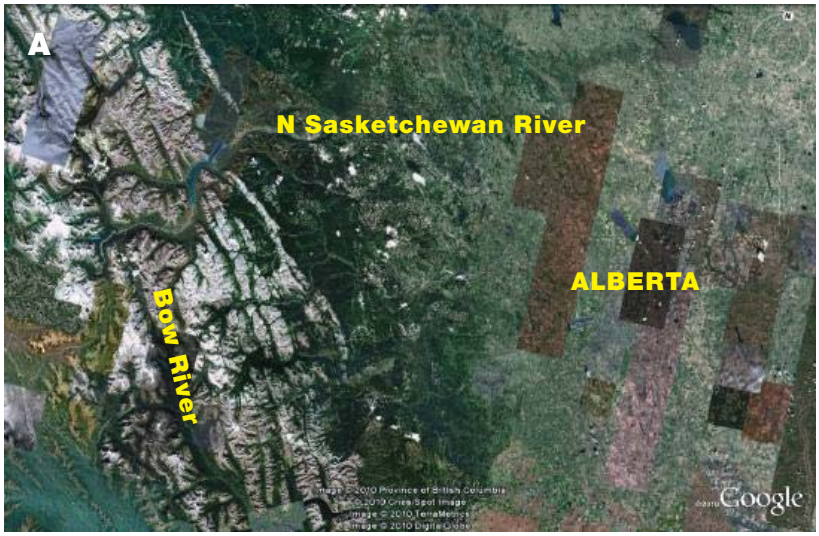
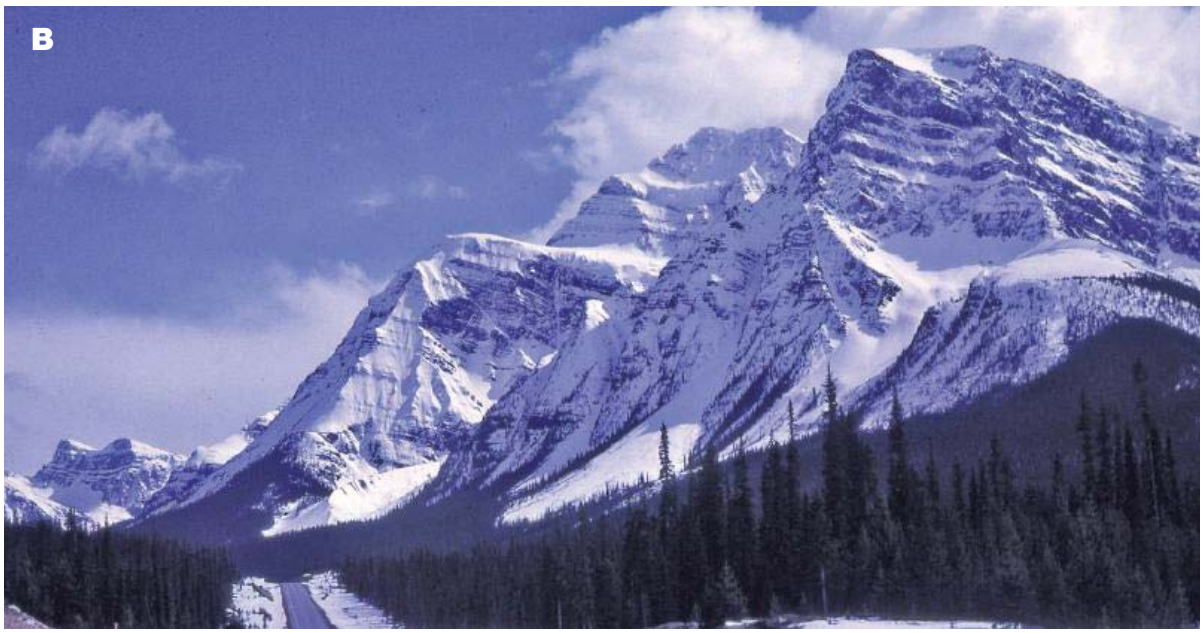


Figure 1.1 Examples of scale in geomorphology, illustrated by the Canadian Rocky Mountains. **A.** Continental scale, represented by a satellite image – ©Google Earth image of the Rocky Mountains and adjacent areas of the Canadian Prairies. The timescale related to the development of the physical features depicted on the image is one of millions of years. Note the contrast between the structurally complex mountain chain and the structurally stable area to the east, where the plains are underlain by near-horizontal bedrock. Note also how the satellite image brings out the vegetation contrasts between the grasslands of the southern



Prairies and the forested areas further north and in the mountains. **B.** Regional (landscape) scale photograph looking south along the Bow Valley within the Rocky Mountains. At this scale the straight alignment of the main valley is clear, as is the mountain morphology to the west. The westerly dip of the sedimentary rocks forming the mountains is clear, as are the glacial erosional features on the mountains and the scree slopes at the base. The timescale related to the development of the landforms visible in the photo is one of tens of thousands of years. **C.** Detailed photograph of the bed of the braided river channel of the North Saskatchewan River within the Canadian Rockies. Note the milky colour of the water, due to suspended sediment, released by melting glaciers within the catchment of the river. Note also the gravel bars on the bed of the river. This river has only existed for the last 8000 years or so, and its detailed form is modified at least annually by snowmelt floods.

formed the structures of the Rockies have been operating over 30–50 million years, but their **tectonic** uplift to form a mountain chain has probably occurred only over the last 10 million years or so. In the second view we are looking at a landscape that was periodically covered by ice sheets during the last half million years, then by huge valley glaciers that deepened the main valleys. The small glaciers in the Rockies today are merely diminutive remnants of much bigger bodies of ice that melted only 8–10 thousand years ago. The third view shows a river that has only existed for maybe 8000 years, but its detailed morphology is modified following every flood generated by summer storms or the annual snowmelt flood in the spring.

To grasp this wide range of timescales we need to know a little about geological time, and especially about the particular timescales relating to approximately the last two million years. The Earth as a planet is about 4000 million years old. These vast ages have been determined on the basis of the decay rates of radioactive elements contained within the rocks of the crust. The later 15% of geological time (the Phanerozoic; *see* Figure 1.2A) is divided on the basis of fossil evidence into a system of eras and periods, related to the evolution of life forms on the Earth. The landforms of the surface of the Earth are young in relation to geological time. Most of the detailed form has developed only over the last 1.6 million years or so during the Quaternary (the Pleistocene and the Holocene; Figure 1.2B). The present (Alpine) system of mountain ranges dates largely from the mid-Cenozoic (*c*.25 million years) and the modern pattern of continents and ocean basins dates largely from the early Cenozoic (65 million years). Only in parts of Australia and Africa can Mesozoic landform patterns be recognised to any great extent. However, older mountain systems can be recognised in the structural patterns and to a certain extent in the relief of all the continents. It is with Quaternary (*see* Figure 1.2B) timescales (the last 1.6 million years) that the geomorphologist is most concerned.

The modern basis for subdivision of the Quaternary is climate. Over the last 1.6 million years there have been numerous climatic oscillations. Interglacial conditions (such as the present time) have alternated with global glacial periods, when lower global temperatures allowed large ice sheets to form over much of the northern continental areas, in addition to the more permanent ice sheets over Greenland and Antarctica. These oscillations are caused by cyclic variations in the Earth's orbital characteristics (**Milankovitch cycles**: named after the Yugoslav mathematician who discovered them). We shall deal with the effects in Chapter 2. Here we deal with the modern basis for Quaternary chronology. The chronology is based on the oxygen isotope record preserved in ocean-floor sediments and in the glacial ice of Greenland and Antarctica. The two isotopes of oxygen (O^{16} and O^{18}) have different temperature-related potentials for evaporation. Hence, atmospheric concentrations of each isotope are enhanced and seawater concentrations diminished, or vice versa, as the result of fluctuations in the global temperature, and particularly in the volume of water in the oceans, which reflect global glacial/interglacial cycles. Oxygen isotope ratios preserved within the calcium carbonate ($CaCO_3$) of foraminifera shells (small marine protozoans) or within ice crystals therefore vary with the global glacial/interglacial cycles. Figure 1.2B summarises the oxygen isotope

EON	ERA	PERIOD		DATE (Ma) approx
PHANEROZOIC	Cenozoic (Tertiary)	Quaternary	Holocene	0.01
			Pleistocene	2
		Neogene	Pliocene	6
			Miocene	25
		Palaeogene	Oligocene	40
			Eocene	65
	Mesozoic	Cretaceous		135
		Jurassic		200
		Triassic		240
	Palaeozoic (Upper)	Permian		280
		Carboniferous		370
		Devonian		415
	Palaeozoic (Lower)	Silurian		445
		Ordovician		515
		Cambrian		590
(PRE-CAMBRIAN)				
	PROTEROZOIC			2500
	ARCHAEAN			4000

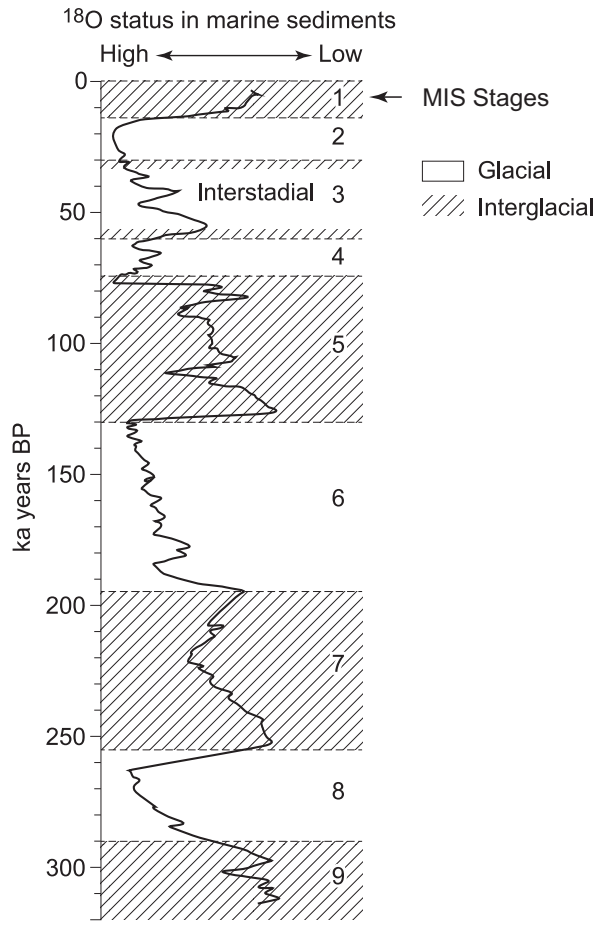


Figure 1.2 Geological timescales. **A.** *(Above)* Geological timescale. **B.** *(Right)* Timescale for the last 300,000 years of the Quaternary: based on the oxygen isotope record from foraminifera in marine sediments. The nomenclature for the stages here labelled MIS (Marine Isotope Stages) replaces the earlier OIS (Oxygen Isotope Stages) nomenclature.

record for last 300,000 years of the middle and late Quaternary derived from the marine record. The warm interglacial or mild interstadial phases are given odd numbers increasing in age from the modern Holocene, Marine Isotope Stage 1 (MIS 1). MIS 3 was an interstadial, not as warm as the Holocene, and the last major interglacial (MIS 5) occurred around 125,000–90,000 years ago. The intervening cold or glacial phases of the Pleistocene are numbered with even numbers increasing in age from the last glaciation (MIS 2). This and earlier glaciations during MIS 6 and MIS 8 were particularly important for geomorphology (*see* Section 1.4 below). The oxygen isotope-based chronology and notation has largely replaced the Alpine, Northern European or American regional terminology previously used to define glacial phases. The modern approach to Quaternary chronology is based on the global climatic sequence rather than on local and incomplete stratigraphic sequences.

The last 10,000 years (The Holocene: MIS 1) have been important in modifying the effects on the landscape of the environments of the last global glacial (MIS 2). They have also been important for another reason. For the latter half of the Holocene humans have had an

increasing impact on the landscape. For the majority of this timespan the natural landscape was progressively modified by human settlement and the development of agriculture. The evidence for geomorphic change is intimately linked with that for vegetation change and with archaeological evidence for the development of human societies. Over the last 200 years or so the human impact on geomorphic systems has accelerated enormously, both indirectly through radical changes to the properties of the Earth's surface, especially through vegetation change, and more directly through engineering interventions in the sediment cascade. We will consider timescales of landform evolution in more detail in Chapter 5 and human impact on geomorphology in Chapter 6.

There is still another approach to timescales that is important in understanding geomorphic processes (Chapter 4). The effectiveness of individual geomorphic events (e.g. floods, landslides) tends to increase with their rarity. For example, a flood that occurs on average once every 100 years (ie. with a **recurrence interval** of 100 years) will bring about much more erosional change than a flood that occurs on average once every five years. However, because we might expect there to be 20 five-year floods for every 100-year flood, the overall and cumulative effects of the five-year floods may be greater. This concept, referred to as the **magnitude and frequency concept**, was developed in the 1960s in a classic paper by Gordon Wolman and John Miller (*see* Figure 1.3), where they demonstrated that in active landscapes the greatest cumulative amount of geomorphic work done (erosion, sediment transport, deposition) was carried out by events of moderate magnitude and frequency. Active landforms, especially in river systems, tend to adjust to such events by erosion and deposition. For example, river channels (*see* Chapter 4) tend to be maintained by a balance between erosion and deposition at a size related to such moderate events.

Two major elaborations of this concept were developed particularly in the concept of **geomorphic sensitivity** by Dennis Brunsden and John Thornes, and in the **geomorphic threshold** concept by Stanley Schumm. A landscape or geomorphic system is said to be sensitive if recovery from a major disturbing event, such as a major flood, is long-drawn-out in relation to the frequency of the disturbing event. In such landscapes there is a relatively

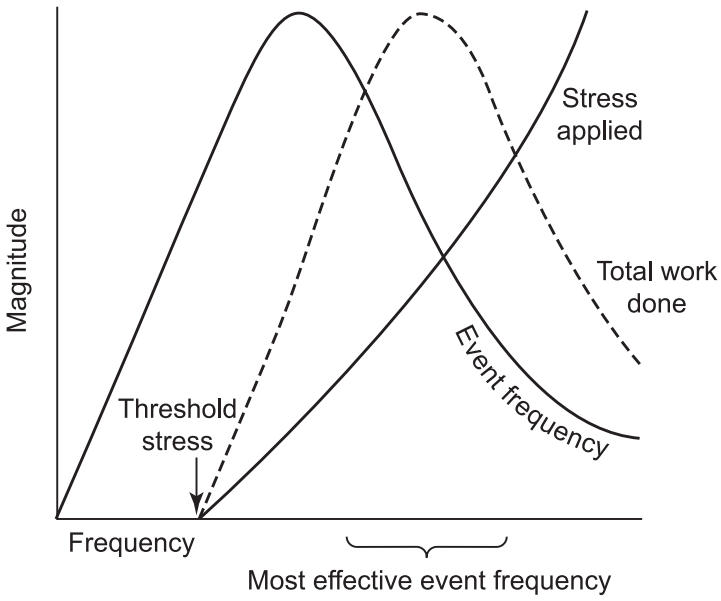


Figure 1.3 Relationships between event magnitude and frequency and geomorphic work done (modified from the classic work of Wolman and Miller).

high probability of further disturbance before the system has recovered from the previous event, for example if re-vegetation of eroded slopes is slow. On the other hand, landscapes that recover quickly from disturbance are said to be robust.

At a larger scale, a landscape may undergo sudden and rapid changes from one state to another, for example from stable to gullied hillslopes or from a meandering to a braided river channel. Such a landscape is said to have crossed a geomorphic threshold. These thresholds may be related to internal properties of the geomorphic system or may be brought about by external stresses; for example, by tectonic, climatic or human-induced changes in the environment. One research problem is to differentiate between externally and internally induced thresholds, a problem exacerbated by the common occurrence of a **complex response** to threshold-related changes to the geomorphic system. Robust landscapes are better able to withstand such stresses, but sensitive landscapes may be more susceptible to externally induced threshold changes. A modern consideration would be to identify landscapes that may be particularly vulnerable to change in response to global warming.

1.4 What we mean by the fundamental driving forces

The world's landforms are the result of the interaction between internal (geologically-driven) and external (climatically-driven) forces. The vast majority of the Earth's surface form is the result of climatically-driven erosional and depositional processes, operating on the surface of the planet, whose structure and composition are the result of geologically-driven internal processes.

1.4.1 Internal (geological) forces

Huge advances in our understanding of the Earth as a planet occurred in the late 1960s and 1970s through the development of the plate tectonics concept. In this model the Earth's crust is seen to be composed of a series of relatively stable rigid (lithospheric) plates separated from one another by less stable plate boundary zones.

The crust is lighter than the underlying mantle and rests on it in **isostatic** equilibrium; in other words, at an elevation proportional to the crustal thickness and density. The crust itself is of two sorts: lighter, thicker **continental crust**, composed dominantly of **granitic** rocks, and denser, thinner **oceanic crust**, composed dominantly of **basaltic** rocks, so the continents sit at higher elevations than the ocean floor. Most of the lithospheric plates include both oceanic and continental portions.

There are three types of **plate boundary**: constructive, destructive and conservative boundaries. Below the crust the upper mantle (the asthenosphere) is sufficiently near its pressure melting point to allow heat transfer by convection, and deformation by slow flowage. Convection within the upper mantle creates linear zones of upward heat transfer that partially melt the upper mantle (peridotite) rocks, releasing **basalt** lavas into the crust. These zones of upwelling are known as **constructive plate boundaries** as this is where new crust is formed by the extrusion of basalt lava. They form **mid-oceanic ridges**, separating two tectonic plates. The basalt lava is added to the inner margin of each plate, creating new oceanic crust, and thus widening the ocean basin in the process known as sea-floor spreading (Figure 1.4A). This process is slow; for example, the modern North Atlantic

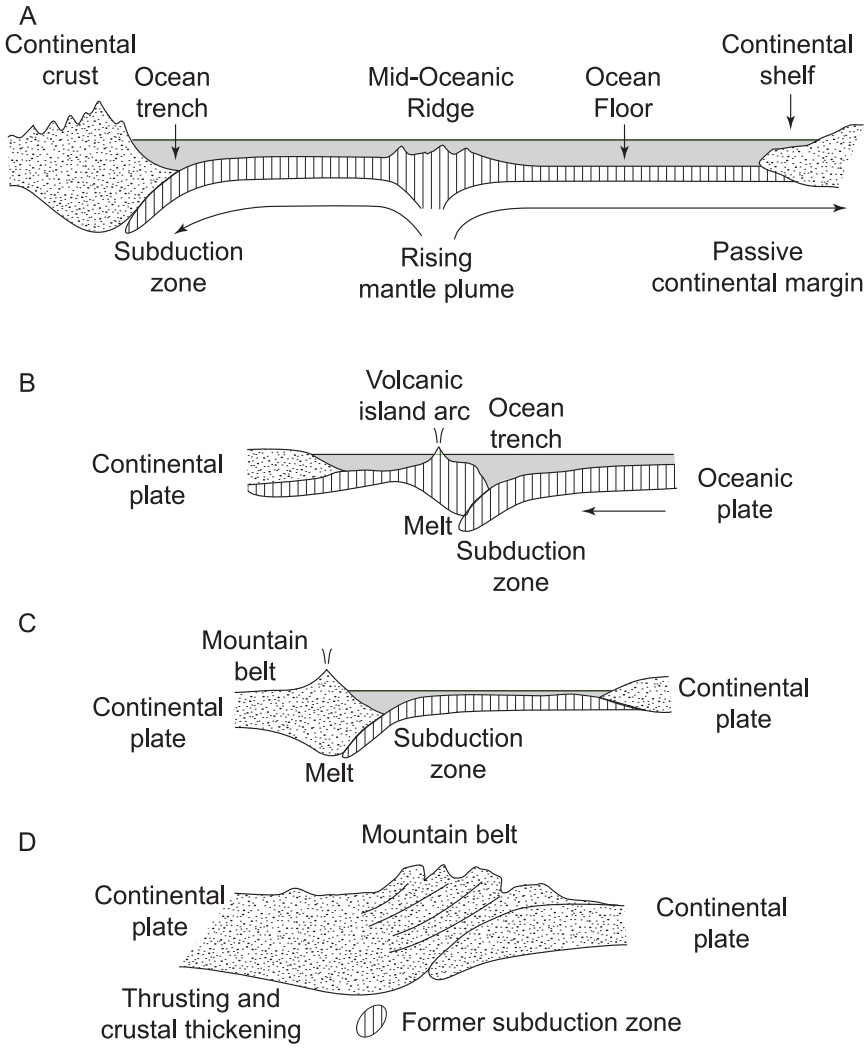


Figure 1.4 Schematic sections of plate boundaries: **A.** Constructive boundary: sea-floor spreading zone; **B.** Subduction zone: island arc setting; **C.** Subduction zone: continental margin setting; **D.** Continent to continent collision zone.

Ocean has developed over the last 70 million years or so. An early stage in this process is crustal rifting. The Red Sea rift system is beginning to separate the African and Arabian plates. Mid-oceanic ridges and rift systems are zones of volcanic activity.

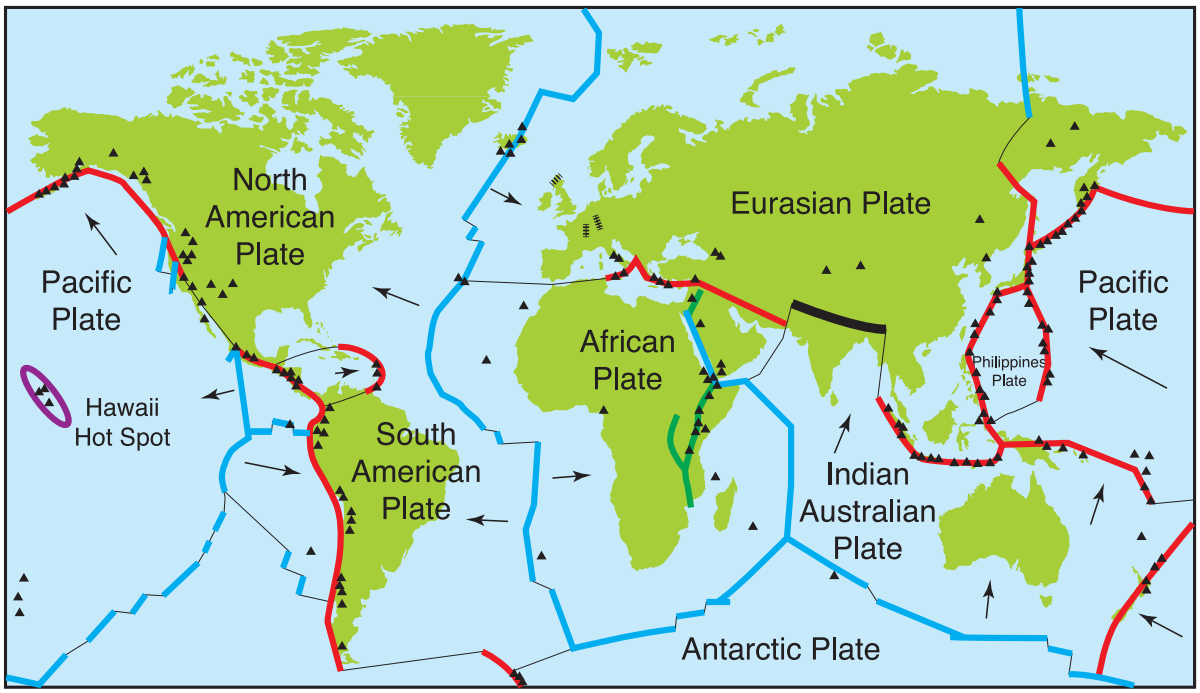
At the same time as the ocean basins develop by sea-floor spreading, the opposite plate margin undergoes compression as the plate converges with a neighbouring plate This type of plate boundary is called a **destructive plate boundary**, because it ultimately involves the destruction of crust and its absorption into the upper mantle. There are three types of convergent (destructive) plate boundary. In the first case the two plates are of oceanic crust, (Figure 1.4B) and the more mobile plate is forced below the other plate. In the second case, the opposing plate is formed of thicker continental crust (Figure 1.4C). The thinner, denser oceanic plate is forced under the continental plate. In both cases this process is

called **subduction**. The descending plate undergoes partial melting, causing volcanic activity at the surface. The third case occurs when two continental plates collide (Figure 1.4D). Both of the latter cases involve thicker continental crust and so considerable crustal thickening occurs, some due to subduction itself, but also as a result of the compressional tectonic setting forcing bodies of rock to be folded and to be thrust over one another.

Destructive plate boundaries have major topographic expression (Figure 1.5). Those involving oceanic crust are characterised by deep marine trenches and volcanic island arcs. Zones of crustal thickening are isostatically elevated to form mountain chains.

Continental-margin boundaries coincide with some of the Earth's major mountain chains, the 'young fold mountains', including those that more or less encircle the Pacific Ocean. Subduction below such mountain chains produces volcanic activity. Continent to continent collision zones result in the greatest amounts of crustal thickening, the greatest amounts of crustal isostatic uplift, and therefore form the highest mountains of all, the Himalayas.

Again, the timescales involved are enormous; for example, the evolution of the Western Cordillera of North America relates to westward movement of the Americas plate over the same timescale as the widening of the



Zones of historically active (subaerial) volcanism
Continent/continental collision
Subduction
Sea-floor spreading
Rifting
Major hot-spot
Former rifting
Transform faults

Figure 1.5 Global plate-tectonic patterns.

Atlantic Ocean (*see* above). Similarly the Eurasian Alpine/Himalayan mountain system has developed over a similar period as the result of the southern continental plates (Africa, India) encroaching on Eurasia and the closure of what was once an intervening Ocean (Tethys), roughly in the position of, but much larger than, the modern Mediterranean Sea.

A third type of plate boundary is a **conservative boundary**, where crust is neither created nor destroyed, but one plate moves laterally against the margin of another. These are major **transform fault** zones, and although earthquakes are common on all plate boundary types, transform faults are the sites of some of the most powerful earthquakes on Earth. Classic examples include the San Andreas Fault in California and the Anatolian Fault in Turkey.

The plate tectonics model provides a mechanism for the much earlier continental drift theory, within which, over geological time, the continents 'moved' in relation to one another. The plate tectonics model also provides the basis for interpreting previous geological patterns, as well as for understanding the modern patterns of gross spatial and elevational characteristics of the Earth's surface.

We will deal with the topographic expression of global-scale plate tectonics in more detail in Chapter 2, but there are also implications at the regional and local scales. At the regional scale (Chapter 3), the distribution of volcanic activity closely reflects the plate tectonics context together with the location of 'hot spots' above mantle plumes (Figure 1.5). The intensity of structural rock deformation by folds and faults also reflects modern and past plate tectonic activity. At the local scale the plate tectonics context is expressed by the location of individual volcanoes and local tectonic patterns.

1.4.2 External (climatic) forces

Gross relief (mountains, plains, etc.), can be related to plate tectonics, but the transformation of that gross relief into landforms is the result of processes largely generated by the climate system. Surface geomorphic processes (*see* Chapter 4), which can collectively be described as the **sediment cascade**, are driven largely by the climate system. These processes involve **weathering**, the breakdown of rock by mechanical and chemical processes, dependent on moisture and temperature; then erosion, transport and deposition of rock debris by various geomorphic systems, driven by gravity, flowing water including waves and currents, wind, and glacial ice. Apart from gravity, all of these are dependent on the climate system.

We will deal with the implications for global geomorphology in Chapter 2, but here we will focus on some of the climatic mechanisms important for geomorphic processes, particularly those related to temperature and moisture.

Temperature is important, both directly and in relation to moisture availability. A high diurnal temperature range affects heating and cooling of rock surfaces and therefore mechanical weathering. Similarly, the frequency of freeze-thaw activity affects the mechanical weathering regime. Sustained high temperatures and high moisture content accelerate chemical weathering processes. Mean annual temperature range is also important. Mean annual temperatures below 0 °C may result in a permanently frozen subsoil and bring about a whole range of soil and slope processes

characteristic of **periglacial** environments. Sustained winter temperatures below 0 °C allow accumulation of an annual snowpack, which on melting in spring or summer may lead to heavy annual flooding. Sustained temperatures throughout the year below 0 °C allow perpetual snow accumulation and its conversion to glacial ice.

Moisture availability is also fundamental. Important is the relation between annual precipitation and annual potential **evapotranspiration**, which differentiates humid from sub-humid, semi-arid and arid environments. Where precipitation is in excess, soil moistures are maintained, groundwater recharge takes place and perennial rivers are sustained. In dry regions rivers are often ephemeral, but prone to flash floods from occasional storms. In humid regions on free-draining sites, high soil moistures promote rapid soil development, and the development of soil profiles is dominated by downward movement of moisture through the soil, whereas in dry regions soil formation is slower and there is much less downward movement. In humid areas, both temperate and tropical, soil formation tends to be faster than removal by erosion, leading to the general condition of soil-mantled landscapes. In arid areas the reverse is more common, with characteristically bare, eroded landscapes. Rainfall itself is an important geomorphic agent, especially high intensity rainfall, leading to erosion by run-off, to flood conditions, and often to the initiation of landslides.

Perhaps the most effective way of considering the impact of the climatic system on geomorphic processes is to consider the hydrological cycle. At a global scale the hydrological cycle involves transfer of water by evaporation, precipitation, glacial melt and river flows between the major water storage zones on the planet, the sea, the atmosphere, the land surface, underground, and glacial ice. During the Pleistocene ice ages, reduced global temperatures increased the proportion of the world's water stored in glacial ice, reducing that stored in the oceans, thereby reducing global sea levels (*see* Chapter 2).

However, it is at the regional and local scales, that of the drainage basin, that the hydrological cycle is most relevant for consideration of geomorphic processes. In Figure 1.6 the major stores of the drainage basin hydrological cycle are indicated by boxes, the flows by arrows, and the mechanisms controlling the flows by diamonds, with those processes that are of particular importance for geomorphology highlighted in bold.

Precipitation falls from atmospheric storage onto the land surface. If it falls as snow, it may be stored on the surface for some time as a snowpack, before it melts. If seasonal melting is rapid it can contribute significantly to rapid run-off and river flooding. If it falls as rain, the intensity of the rainfall is important in determining its subsequent behaviour. Vegetation will act as an umbrella, protecting the land surface below by **interception**, its effectiveness depending on the vegetation type and decreasing with increasing rain duration and intensity. The soil will absorb incoming rainfall at a particular rate, known as the **infiltration capacity**, depending on soil properties and on the antecedent soil moisture. In most cases infiltration capacity will be higher than the effective rain intensity, in which case almost all the incoming rain will be absorbed by the soil. Only when rain intensity exceeds the infiltration capacity

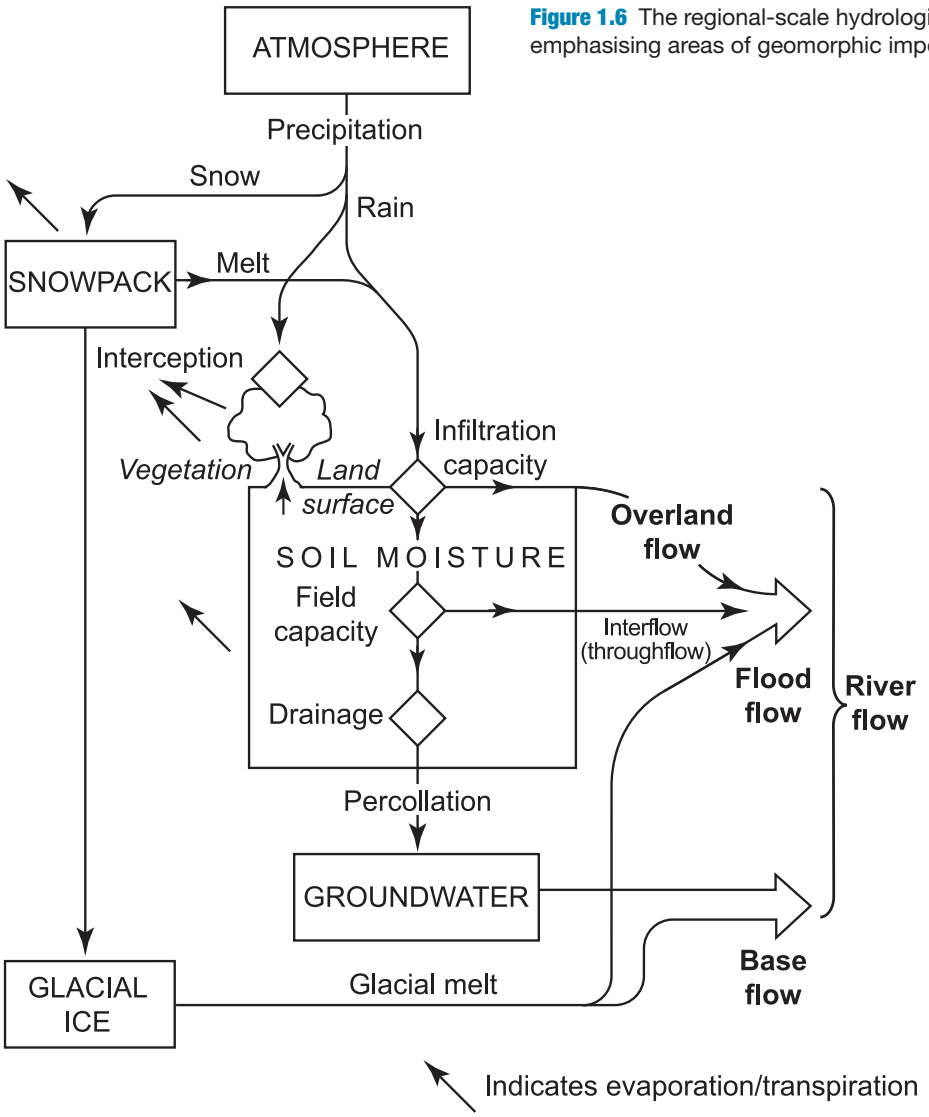


Figure 1.6 The regional-scale hydrological cycle, emphasising areas of geomorphic importance.

will run-off (overland flow) take place. Some surfaces will have very low infiltration capacities (e.g. already saturated soils, frozen soils, very clayey soils, some bare rock surfaces, artificial concrete surfaces), in which case all incoming rain will run off. **Run-off** is an important geomorphic agent capable of hillslope erosion (*see* Chapter 4). It is also a major contributor to the **floodflow** component of the **hydrograph**, especially for flash floods. This run-off and erosion model is most applicable in arid regions, where infiltration capacities are low and storm rain intensities tend to be high. In humid regions floodflows are more commonly fed by **saturation overland flow**, run-off from saturated soils. Such soils are usually located near the channel on low-angle, well vegetated slopes, and

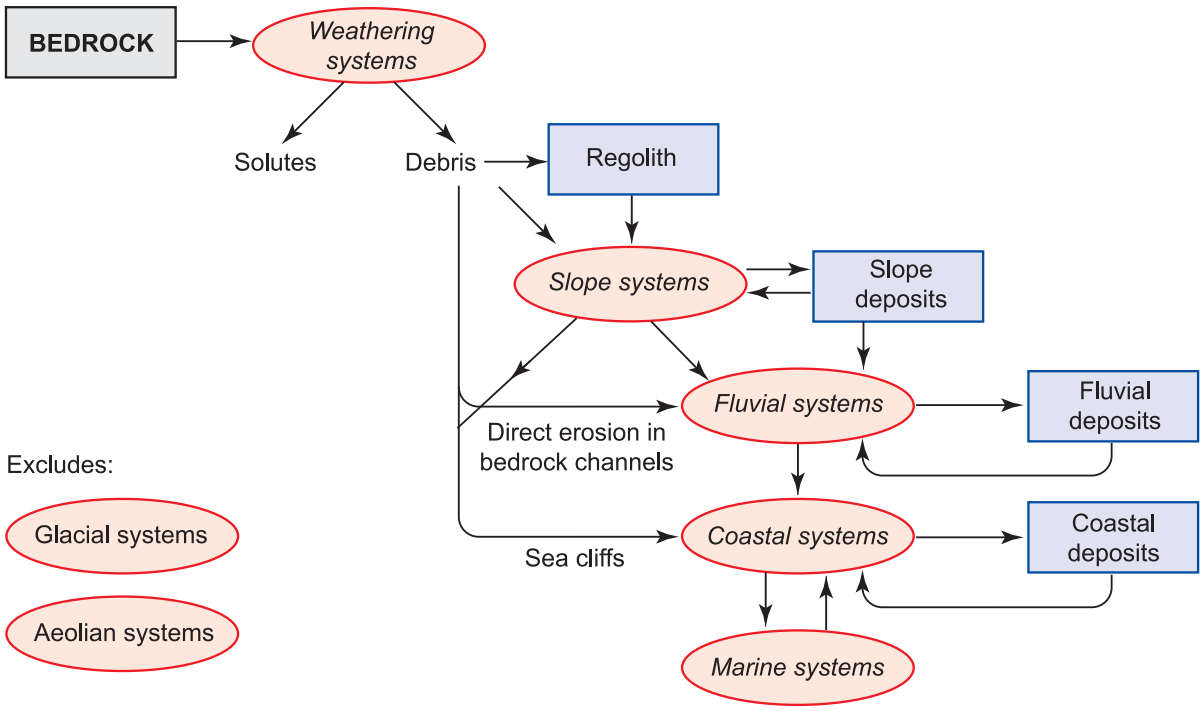


Figure 1.7 Schematic representation of the sediment cascade as applied to drainage basins (Note: for simplicity glacial and aeolian systems are excluded).

therefore yield much less sediment.

The fate of the soil moisture depends on another soil property, the **field capacity** of the soil, the amount of soil moisture that can be retained in the soil by **capillarity**, rather than draining downwards through the influence of gravity. If the infiltrating water does not bring the soil moisture up to field capacity, there will be no further movement; soil moisture will simply be used by plants or evaporate during dry weather. If, however, soil moisture exceeds field capacity, it will drain if it can, or if drainage is impeded, the increased water content will lead to a saturated soil. High soil moisture contents greatly reduce the stability of the soil, and may lead to shallow landsliding (*see* Chapter 4). Where drainage from the soil is possible, the excess water will drain either laterally by what is known as **interflow**, or in free-draining soils vertically by percolation into the underlying bedrock to the water table as **groundwater**. Rapid interflow may be an important constituent of river floodflow in humid areas. Groundwater will eventually drain through springs to river systems, sustaining the dry-weather flow (**baseflow**) of rivers.

The hydrological cycle operates very differently in different global climatic regions. In humid areas during rainy weather most of the cycle may operate. In arid areas, interception is limited and soils are shallow, limiting infiltration capacity; hence, during storm conditions run-off may be heavy. Also, soil

moistures rarely reach field capacity, limiting groundwater recharge, as well as influencing pedological processes. There are therefore fundamental differences in geomorphic processes between humid and arid areas. Similarly, temperatures have a profound effect on the operation of the hydrological cycle, especially when freezing is involved. The seasonal presence of a snowpack or of frozen ground is of importance to geomorphology, as is the presence of long-term water storage as glacial ice. These processes again create major differences between global climatic regions, in terms of their geomorphic regimes.

1.5 Different approaches to the study of geomorphology

In studying geomorphology we are trying to do two things; first to explain how the landforms of the Earth developed (evolutionary geomorphology), and second, how landform processes operate (process geomorphology). In both cases it helps to consider geomorphic phenomena as systems that respond to energy and material inputs, in ways first suggested in the 1970s by Dick Chorley and Barbara Kennedy. Systems respond to variations in energy and material inputs by adjustments of their internal structure and morphology. In the first approach (evolutionary geomorphology) real timescales are an essential part of the system; the geomorphic system responds to real events (e.g. tectonic activity, glaciation, climatic change). Any concept of equilibrium is timebound, for example the initial high rates of erosion following tectonic uplift tend to lessen over time. In the second approach (process geomorphology) absolute time is less relevant. More important are aspects of time that relate to how the system responds to changes in inputs of energy or materials. Under these conditions equilibrium concepts are less timebound. They relate to a balance between input and output and the internal adjustment of the system that maintains such a balance. For example, a river channel undergoes both erosion (sediment output) and deposition (sediment input). If the two are in balance then the channel morphology is likely to be maintained in a form of **dynamic equilibrium**. In other words, the channel morphology adjusts to variations of flood magnitude (energy input) and sediment supplied (material input).

There are two ways of considering such process systems; the first by considering the passage of energy and material through the system (as a cascade), the second by considering the internal structure of the system (as a morphological system). A cascading system comprises a series of flows between stores within the system. The hydrological system (*see* above, Section 1.4.2; Figure 1.6) is one such system relevant to geomorphology. The sediment cascade (*see* above, Section 1.4.2; Figure 1.7) is another. Mechanisms within the system control the flows from store to store. Sets of mechanisms can be viewed as cascading subsystems. Equilibrium conditions within any of the subsystems relate to a balance between inputs and outputs. An important aspect of understanding cascading systems is the awareness not only of the internal mechanisms, but also of the magnitude and frequency characteristics of the inputs to the system (*see* above, Section 1.3).

Morphological systems define the internal structure of a system and are usually described by relationships between the system components. These may be expressed statistically, but

ideally relate to underlying causal relationships. For example, the **hydraulic geometry** of stream channels (*see* Section 4.3) expresses the relationships between flow and channel variables (e.g. by the response of channel width, depth, water velocity, and sediment transport properties to variations in water discharge). In many cases a network of causal relationships can be identified within morphological systems, which 'feed back' on one another. These can radically affect the equilibrium tendencies of the system. Some **feedback** relationships (known as negative feedback) tend to balance out the effects of disturbance. For example, in a river channel, bank erosion will tend to widen the channel, reducing depths, velocities and the erosional stresses on the banks, thereby reducing the likelihood of further erosion. On the other hand some feedback relationships (positive feedback) are self-reinforcing, and therefore tend to destabilise the system. For example, mountain glacial erosion will deepen the **cirques** (*see* Section 4.5.2), which form the main gathering grounds for snow and ice, thus increasing the erosional potential. Ultimately only a major climatic change or complete erosional destruction of the lip of the cirque can break this vicious circle, in other words by the crossing of a major geomorphic threshold (*see* Section 1.3).

The two basic systems approaches (cascades and morphological systems) can be linked together in what is known as a process response system. Many of the cascade mechanisms can be treated as variables in morphological systems, allowing such linkages, and allowing the system as a whole to evolve through time. Such systems are of relevance not only for understanding process geomorphology, but also, because absolute timescales are involved, they can be applied to questions related to evolutionary geomorphology.

Global-scale geomorphology

In this chapter we consider global-scale geomorphology in relation both to the plate tectonics context and to global climatic patterns, present and past. Though dealing with global phenomena, we shall necessarily come down in scale when we consider examples of how these global phenomena are expressed in landforms. It is at this scale that the effects of plate tectonics on the form of the Earth's surface can be seen most clearly.

2.1 The plate tectonics context

2.1.1 The ocean floor

The form of the ocean floor is almost wholly determined by plate tectonics. Mid-oceanic ridges encircle the globe (Figure 1.5). They are formed by the injection of a magma chamber into the crust below zones of seafloor spreading. Away from the zone of seafloor spreading the oceanic crust forms extensive **abyssal plains**. Isolated seamounts and volcanic islands stand above the abyssal plains. These sit above stationary mantle plumes that form hot spots in the crust, triggering volcanic activity. The plumes themselves are geostationary, but the plates move over them, resulting in volcanic activity spanning a range of ages. For example, the Hawaiian islands sit over a hot spot in mid Pacific. The volcanoes in the north-western islands in the chain are now extinct, but Mauna Loa in the south-east is active, reflecting the north-westerly movement of the Pacific plate over the hot spot.

At the leading edge of a plate the ocean floor is modified by subduction. In oceanic subduction, a deep-ocean trench (Figures 2.1, 2.2A) is formed, beyond which lie island arcs, a situation found in many areas in south-east Asia and in the Caribbean. Continental margin subduction may also produce ocean trenches. On this type of continental margin, such as that on the west coast of the Americas, the coastline is relatively straight and there is no significant **continental shelf**.

At trailing-edge continental margins such as those around most of the Atlantic Ocean, the continental crust forms continental shelves, bounded by the **continental slope** down to the abyssal plain. At times of low global sea levels during Pleistocene glacial phases (Section 1.4.2), much of the continental shelf was exposed as dry land over which rivers flowed. Where the larger rivers (e.g. the Amazon, the Hudson) flowed over the continental slope some of them incised, cutting deep canyons. River-fed, sediment-charged water is funnelled down these submarine canyons as turbidity currents, spreading their sediment onto the sea floor as submarine fans. Near to land on the continental shelf, especially around the North Atlantic Ocean, there may be a relict terrestrial topography submerged by the post-glacial rise in sea level. This may be an erosional topography where the tops of the partially submerged ridges form islands (e.g. the Hebrides, off western Scotland), or a depositional topography, especially a glacial

depositional topography, adding complexity to the morphology of the sea floor.

2.1.2 Global-scale continental landforms

At the global scale the topography of the continental areas also strongly reflects the influence of plate tectonics, but in a more complex way than that of the ocean floors. The young, high mountain ranges of the Alpine/Himalayan system and of the Pacific rim coincide with active or recently active destructive plate boundaries (compare Figures 1.5. and 2.1), their high elevation being the result of crustal thickening and sustained crustal isostatic uplift (*see* above, section 1.4.1). In many areas the detailed structures of these mountain ranges are far from simple. The broadly continent-to-continent collision that created the European Alpine system and the changing plate tectonics setting of the western United States have produced incredibly complex mountain systems at the regional scale (*see* Chapter 3).

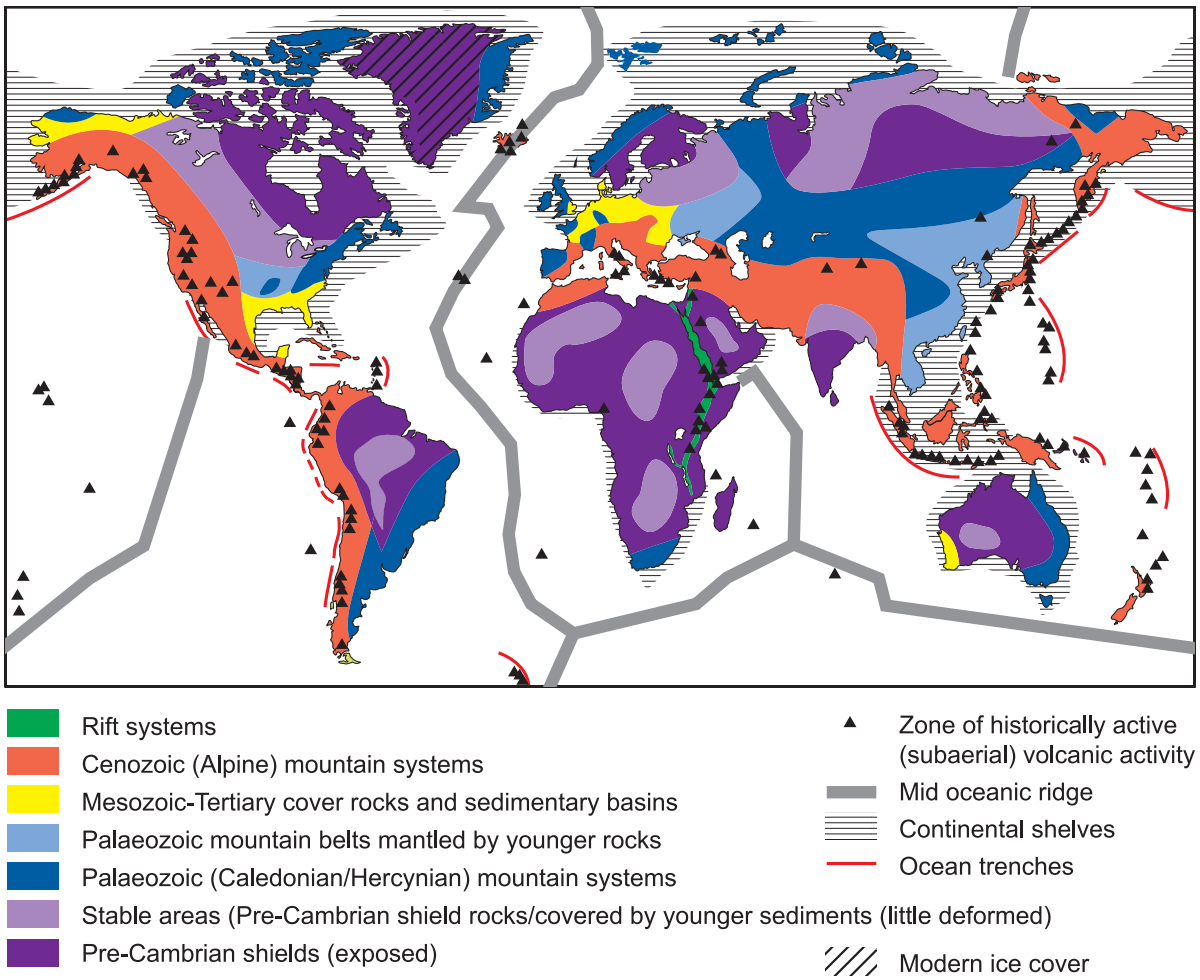
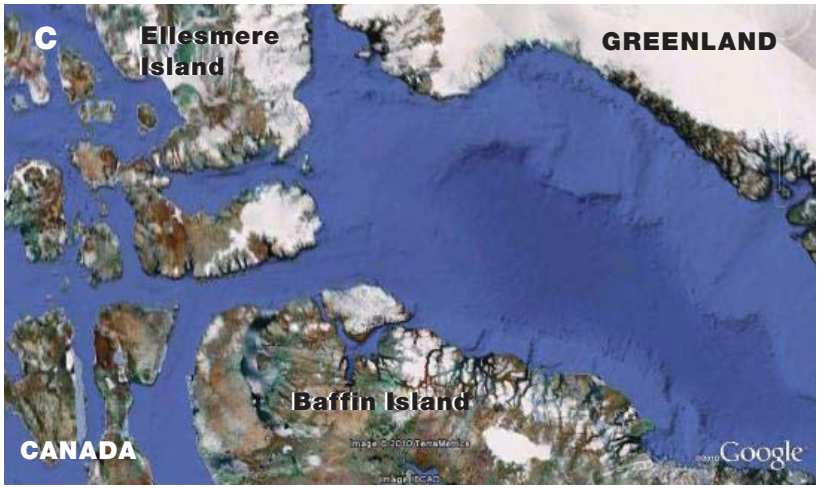


Figure 2.1 Global relief patterns and structural units.

Figure 2.2 Examples of satellite images of global/continental-scale features (©Google Earth).
A. The ocean floor off Indonesia: this is a subduction zone. Note the absence of a continental shelf, and the ocean trench off the west coast of Indonesia.

B. Deserts in central Australia. To the east are the folded rocks of the Flinders Ranges; in the centre are the salt flats of dry Lake Torrens; elsewhere are Pleistocene dune systems.



C. Parts of Arctic Canada and western Greenland. Note the extensive ice sheet cover, not only over Greenland to the east but also the smaller ice sheets on the Canadian Arctic islands.

In addition to young mountain systems created by the modern plate tectonics setting, the remnants of older mountain systems form lower, less dramatic mountains on every continent. These mountain systems relate to destructive plate boundaries that are no longer active. They owe their present relief in part to crustal isostatic uplift, related to the original crustal thickening, but their modern topography is purely erosional and related to the presence of older, harder, more erosionally resistant rocks. Two systems can be identified (Figure 2.1). The Hercynian/Variscan system dates from a major tectonic phase during Permian times, 300–250 million years ago (Figure 1.2). At that time the Atlantic Ocean did not exist and a major mountain system developed along the southern margin of a proto-Eurasian continent with an arm extending along the Urals. It extended westwards into what are now the Appalachians in the eastern part of the United States. In the southern hemisphere a similar system can be traced through eastern Australia, southern Africa and into southern Argentina. Today, only in a few places (e.g. the Appalachians, the Urals, eastern Australia) does the Hercynian system represent more or less continuous mountain chains. Elsewhere, especially in Europe, the Hercynian structures and their constituent rocks have been incorporated into younger Alpine structures, or the remnants of the Hercynian mountain system have been fragmented, by post-Hercynian faulting, into discrete upland blocks (e.g. the Massif Central in France), between which areas have subsided and been buried by younger sedimentary rocks.

The Caledonian mountain system relates to the closure of an ocean between the North American and Eurasian plates during the late Silurian and Devonian periods (*c*.400–350 million years ago, Figure 1.2). Today their remnant rocks and structures form the Scandinavian mountains, the upland areas of north-west Britain and much of Ireland. They can be traced across the Atlantic into North America, where they have been incorporated into the Appalachian structures. Another fragment is present in South America. Again, their elevation relates in part to the original crustal thickening, but their present relief forms are entirely erosional, related primarily to rock resistance.

Within each continent are zones that have been far away from plate boundaries and have been tectonically stable throughout the Phanerozoic. These are the cratonic '**shield**' areas (Figure 2.1), composed of Precambrian, mostly metamorphic, rocks older than 550 million years. Over extensive areas (e.g. the Baltic Shield, the Canadian Shield) the Precambrian rocks are exposed at the surface, but away from the core areas (under the Russian platform, under the North American plains) the Precambrian rocks are mantled by a little-deformed cover of younger sedimentary rocks. The shield areas of the northern continents (the Baltic Shield, the Canadian Shield) were heavily glaciated during the Pleistocene (*see* below, Section 2.2.2), therefore their detailed landforms are relatively young. However, the shield areas of the southern continents escaped glaciation during the Pleistocene, therefore they preserve ancient land surfaces dating back well into the Tertiary and possibly beyond.

The major lowlands of the continents are areas of less deformed younger sedimentary rocks that lie between shield areas, remnants of old mountain chains and modern

mountain chains. In some cases these form distinct Mesozoic or Cenozoic sedimentary basins (e.g. the Paris Basin), but elsewhere they simply comprise relatively undeformed Mesozoic or Cenozoic sedimentary rocks burying older structures.

One other feature that can be related to modern and ancient plate tectonics patterns are rifts. These form over zones of mantle upwelling and may be the precursors of sea-floor spreading and the development of ocean basins. The East African rift system and its extension along the Dead Sea Rift constitute a modern zone of rifting, which eventually may mean the splitting of the African plate. In Europe an aborted discontinuous rift system, dating from the Miocene, can be identified in the Auvergne (France), and in the Rhine rift valley on the Franco-German border. The Eocene volcanic rocks of western Scotland and the Inner Hebrides may represent another early (Tertiary) aborted rift system that was replaced by the mid-Atlantic rift.

2.2 The global climatic context

2.2.1 Climatic geomorphology

The world's weather and climate patterns control the distribution of heat and moisture, which in turn control the distribution of geomorphic processes. We will deal with geomorphic processes in more detail in Chapter 4, but here we identify how different climatic regimes favour the operation of various geomorphic regimes. Box 2.1 summarises how different climatic regimes favour the operation of different geomorphic processes.

Climatic geomorphology, the basis of which is outlined in Box 2.1, deals with the climatically controlled distribution of geomorphic

Box 2.1 The geomorphic effectiveness of global climates

1. Arctic and Antarctic (Glacial) climates

In these areas mean annual temperatures are well below 0 °C.

Year by year snow accumulation converts to glacial ice.

Glacial processes are dominant.

2. Sub-arctic (Periglacial) climates

In these areas mean annual temperatures are below 0 °C.

The subsoil remains frozen (permafrost) throughout the year, but there is sufficient summer warmth to cause snowmelt floods, and to thaw the surface layers of the ground: slope processes (solifluction) are effective. Frequent freeze-thaw cycles and **nivation** are very effective in mechanical weathering.

3. Humid temperate climates

These climates cover a wide range of actual climates, but they have several characteristics in common. Precipitation exceeds potential evapotranspiration: soils are commonly moist, there is recharge of groundwater, sustaining perennial rivers.

Rainfall may be all-year (e.g. Western Europe), or seasonal with summer dominance (e.g. continental climates; American east-coast climates), or seasonal with winter dominance (the more humid Mediterranean climates). Summer temperatures may range from cool (e.g. Scotland) to hot (e.g. Italy, north-eastern USA) and winter temperatures may range from mild (e.g. Brittany) to cold

(e.g. eastern Canada). Temperature and moisture conditions mean that the natural vegetation of this zone is forest, but much of this area is now used for agriculture. Under most 'natural' circumstances the rate of surface erosion is less than the rate of weathering and soil formation, so most landscapes are soil-mantled. The most effective geomorphic processes are slope (mass movement) and fluvial processes.

4. Dry climates

These climates again cover a wide range of conditions, but the critical aspect is that annual precipitation is markedly less than the high potential evapotranspiration. This results in dry soils, little groundwater recharge, and usually a scant vegetation cover. Summer temperatures are invariably hot; winters may range from mild to cold. Dry climates include semi-arid Mediterranean climates (e.g. Spain, Israel), interior continental steppe and dry grassland areas, desert margins and the truly arid climates of the great deserts of the world. Rainfall tends to occur in occasional heavy convectional storms, resulting in rapid run-off; therefore slopes are dominated by surface erosion rather than by mass movement processes. Rivers tend to be ephemeral, dry for much of the year, but responding rapidly to heavy rainfall by flash floods. Weathering is slow, but produces characteristic soils and surfaces. In truly arid areas, precipitation may be rare, and geomorphic processes may be dominated by wind action.

5. Tropical climates

In these areas temperatures are high all year round; freezing is virtually unknown, and precipitation increases towards the Equator from the great desert areas. Precipitation may be all-year as in the truly equatorial regions, but may be strongly seasonal, especially on the drier margins of the tropics (e.g. in the Sahel), and in Monsoon regions. In Monsoon regions wet-season rainfall may be exceptionally heavy. High temperatures and high precipitation favour rapid weathering and the development of very deep soil profiles. Natural vegetation ranges from grassland and dry woodland in the seasonally dry areas to tropical rainforest. In undisturbed areas geomorphic activity reflects the rainfall regime, with seasonally high river flows in the seasonally wet/dry regions to high perennial flows in the true humid tropics. Because of a naturally deep weathering mantle and the high rainfall, these regions may be prone to considerable human-induced disturbance.

6. Mountain climates

Because of their elevation, high mountain environments may experience very different climates from surrounding lowlands, generally with cooler temperatures and higher precipitation. Glacial conditions exist in high mountain areas in temperate and even in tropical latitudes. Mountain areas within deserts may well support forest vegetation. The geomorphology of mountain regions is made distinct not only by the climatic regime, but also by the presence of steep slopes, which tend to accelerate most geomorphic processes.

processes. Although modern processes essentially determine the gross landscape types (e.g. Deserts, the Arctic: Figure 2.2B, C,), in most regions the landforms carry a legacy of past processes, especially of those processes that were active during the Pleistocene.

2.2.2 Quaternary climatic change – glaciation

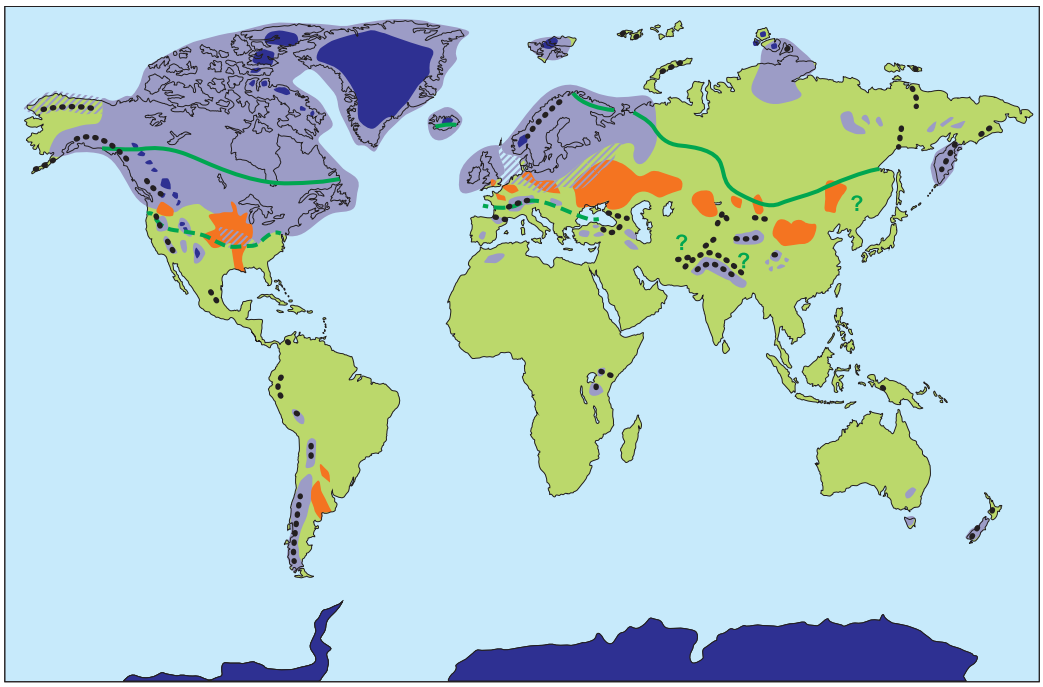
Over the last half million years or more the Earth's climates have oscillated between global glacial and interglacial conditions (*see* Section 1.3). In addition to the semi-permanent ice caps over Greenland and Antarctica, during the Pleistocene glaciations large continental ice sheets formed over large areas of the northern hemisphere, and glaciers in many mountain areas elsewhere became much more extensive. The limits of glaciation are fundamentally important for geomorphology in that they determine the spatial extent of past glacial and related processes. Two sets of limits are important; those related to the last glaciation (MIS 2) reaching its maximum limits about 20,000 years ago, and those related to the maximum Pleistocene glacial extent that occurred during MIS 6 or MIS 8, >150,000 years ago.

Two huge continental-scale ice caps were formed in North America, the Laurentide and Cordilleran ice caps (Figure 2.3). At maximum glaciation the two ice sheets met in western Canada and their limits extended south to the latitudes of New York, the Ohio Valley, and in the west to somewhere south of the Canadian border. Separate mountain ice caps were present over some of the higher mountain ranges in the west, including parts of the American Rocky Mountains and the Sierra Nevada. Ice spread out from the centre of the Laurentide ice cap, scouring bedrock to produce the intensely eroded terrain on the Canadian Shield. Deposition, both by glacial ice and by temporary lakes during the melting of the ice, took place around the margins to produce the glacial depositional terrain of southern Canada and the Midwest of the USA. Similarly the Cordilleran ice cap and the smaller ice caps and glaciers in the other mountain areas tended to be erosional near-source in the mountains and depositional in the neighbouring lowlands.

In Europe there was a similar situation at maximum glaciation (Figure 2.3). The Scandinavian ice sheet was coalescent with the Scottish ice cap. Ice extended south to the Bristol Channel, London, across Holland into north Germany, Poland and around the Baltic. In and near the source areas, the Scottish Highlands, the Norwegian mountains and across the Baltic Shield, the ice was primarily erosional. Further from the source areas of the ice, across the English Midlands and the North European Plain, deposition was dominant. Further south, smaller ice caps and mountain glaciers occurred in the Alps and the Pyrenees.

In Asia there were ice caps over the Himalayas and in the mountains of north-east Asia. In the Southern Hemisphere, glaciation was much less extensive, but ice caps and mountain glaciers were present in the Andes, Tasmania and New Zealand.

Globally, the maximum glacial limits relate to a glaciation earlier than the last glaciation (Figure 2.3). This has important implications for the preservation of glacial landforms and their modification by non-glacial processes. This is nowhere better illustrated than in Britain (Figure 2.4, Box 2.2), where four zones can be identified.



MODERN FEATURES:

- Modern ice caps
- Main mountain ranges with modern glaciers
- Modern limit of discontinuous permafrost

PLEISTOCENE FEATURES:

- Ice cover at LGM
- Ice cover at maximum Pleistocene glaciation (where known and different)
- Major Pleistocene loess sheets
- Southern limit of features related to Pleistocene permafrost (where known)
- ? Extent of Pleistocene permafrost uncertain

Figure 2.3 Global extent of modern and Pleistocene cold-climate phenomena.

Box 2.2 Zonation of British landscapes in relation to the glacial limits

The first zone is southern England. With the exception of a small area on the north Devon coast, and possibly of the Scilly Isles, the area south of the Bristol Channel and of the Thames valley has never been glaciated. In this area there is a greater preservation of landscape patterns inherited from the late Tertiary (*see* Section 1.3). River system development continued throughout the Pleistocene, without interruption from glaciation. However, this area experienced **periglacial** processes (hillslope processes influenced by the presence of permafrost; *see* Section 4.2) through all the glacial phases of the Pleistocene. The hillslopes have been extensively smoothed by **solifluction** (Figure 2.5A), and great thicknesses of **head deposits** (*see* Figure 4.7C) have accumulated, particularly on the valley sides.

The second zone is that between the maximum glacial limit and that of the last glaciation, essentially midland and eastern England. This area was glaciated, but more than 150,000 years ago; hence

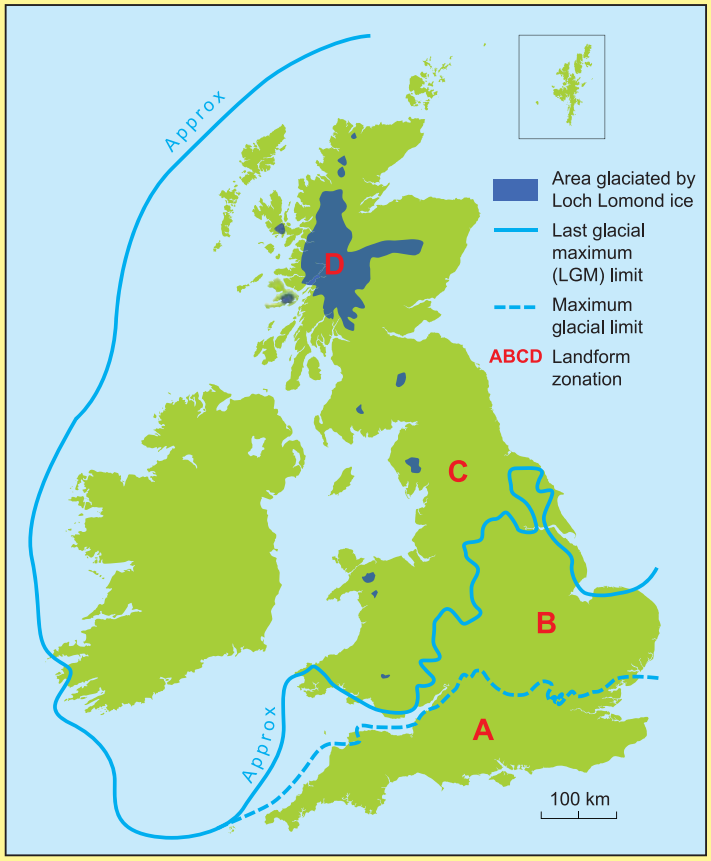


Figure 2.4 Zonation of British landscapes in relation to Pleistocene glacial limits. Zone A has never experienced glaciation. Zone B was glaciated >150 ka ago, but was ice free throughout the last glaciation, so was affected by periglacial processes throughout the last glacial period. Zone C was glaciated at the Last Glacial Maximum (LGM), at 20 ka, but the ice had melted completely by 15–13 ka. The glacial forms were then affected by periglacial processes during the late Pleistocene. Zone D experienced glacial ice cover for about 500 years during the Loch Lomond (Younger Dryas) glaciation, the ice melting rapidly as climate warmed into the temperate Holocene 10 ka ago.

the drainage pattern has been deranged by glaciation. The predominantly glacial depositional topography was modified by periglacial processes throughout the last glaciation and preserves little of the original depositional form.

The third zone is most of the area within the last glacial limit, including most of Wales, everywhere north of a line from Shropshire to the Humber area being glacial, with the exception of the Peak District of Derbyshire and the North York Moors. Glaciers reached this limit at the Last Glacial Maximum (LGM) about 18,000 years ago. The ice melted last of all in the Spey valley of Scotland about 13,000 years ago. In most of this zone glacial erosional landforms (*see* Chapter 4) characterise the highest areas, with glacial depositional landforms elsewhere. The glacial landforms are relatively fresh, though they have been modified by varying durations of periglacial processes that occurred during cold conditions after the melting of the ice cap but before the end of the Pleistocene 10,000 years ago (Figure 2.5B).

There is one other limit that we need to take into account, which defines the fourth zone. This is the limit of a small ice cap that developed at the very end of the last glaciation around 10,000 years ago, and persisted for around 500 years, the so-called Loch Lomond Readvance (the

Figure 2.5 British landscapes (representative of the zones defined in Figure 2.4). **A.** *Zone A:* a periglacial landscape in an area never glaciated, the Quantock Hills, Somerset. Note the strongly convex hillslopes, characteristic of periglacial solifluction processes (*see* Section 4.2). Note also the Holocene incision into the valley floor. **B.** *Zone C:* a landscape glaciated during the LGM, but ice free since 13 ka. Yarrow Valley, Southern Uplands of Scotland. This landscape bears the imprints of both glaciation (deep scour of the valley into the upland plateau; deposition of glacial till – *see* Section 4.5) and limited periglacial processes during the last several thousand years of the Pleistocene. Note the extensive solifluction surface in the centre of the photo, trimmed at the base by Holocene fluvial activity. **C.** *Zone D:* A landscape glaciated during the Loch Lomond stage 10 ka, with little **periglacial** modification, but modification under **paraglacial** conditions by slope and fluvial processes during the Holocene. Glencoe, Western Highlands of Scotland.

Younger Dryas period, in European terminology). It occupied the southern and western Grampian Highlands, with other glaciers in the Cairngorm Mountains and the north-west Highlands, plus small glaciers in the Southern Uplands, the English Lake District and Snowdonia in north Wales. The Loch Lomond phase ended with a rapid climatic warming, so there was no periglacial modification of the glacial topography. Within the Loch Lomond limits the glacial erosional and depositional topography is fresh (Figure 2.5C).

2.2.3 Other changes to the climatic system during the Pleistocene

Outside the global glacial limits there were other climatic changes related to changes in atmospheric circulation. During glacial periods there had been a marked reduction of global temperatures, allowing the development of permafrost and the operation of periglacial processes southwards into the USA and into Europe (Figure 2.3). Even where there was no permafrost, in such areas as the American West, and Spain, freeze-thaw activity was much more effective than it is today. Another effect, in the periglacial regions around the margins of the continental ice sheets, was deposition of windblown silt (**loess**). Loess blankets the topography in some parts of the American Midwest, and across northern Europe in a belt through Belgium and Germany, but the greatest thickness of loess deposits is in the Loess Plateau in central China (Figure 2.3). Interbedded with the Chinese loess deposits are a series of interglacial **palaeosols**, the whole sequence preserving a complete palaeoclimatic record for the whole of the Pleistocene.

With the reduction in global temperature there was also a reduction in evaporation and a consequent reduction in precipitation. This produced a cold, dry steppe climate in the Western Mediterranean region, and greater aridity in some of the world's dry regions, probably so in parts of Australia and southern Africa. Elsewhere some of today's deserts were cooler and more humid, allowing rivers and lakes to exist in, for example, the American south-west and the Sahara. Indeed, desiccation of the Sahara occurred only in the mid-Holocene, well into the present interglacial.

2.2.4 Quaternary sea-level change

During global glacials a much higher proportion of the world's water was stored within the continental ice sheets rather than in the oceans, allowing world-wide sea levels to fall by about 100 m (**eustatic sea-level change**). The continental shelves were exposed; many islands (including Britain) became continental peninsulas; 'land bridges' were created (e.g. between Alaska and Siberia). Around the great ice sheets, the picture was complicated by the crust being depressed under the weight of ice (**isostatic sea-level change**).

On deglaciation, global eustatic sea level rose from its lowest point about 18,000 years before present (BP), coincident with timing of the last glacial maximum, until about 6000 BP when the last ice melted from the Laurentian ice sheet (Figure 2.6). Hence modern coasts are young features – less than 6000 years old. Adjacent to the glaciated areas the picture was complicated by the isostatic rebound of the crust as the weight of the ice sheets was removed (Figure 2.7). In such areas eustatic

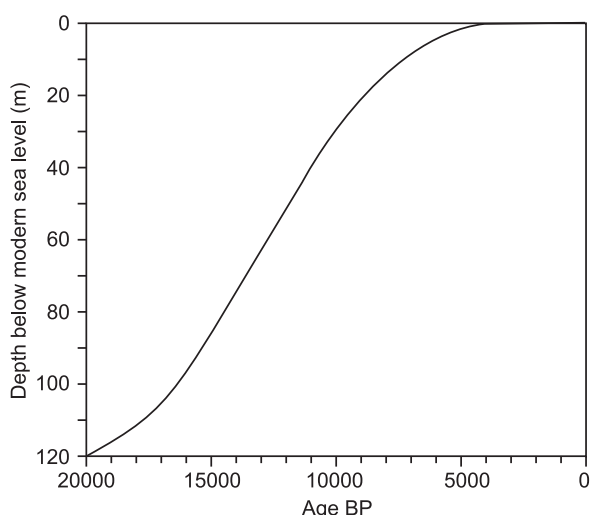


Figure 2.6 Post-glacial eustatic sea-level curve.

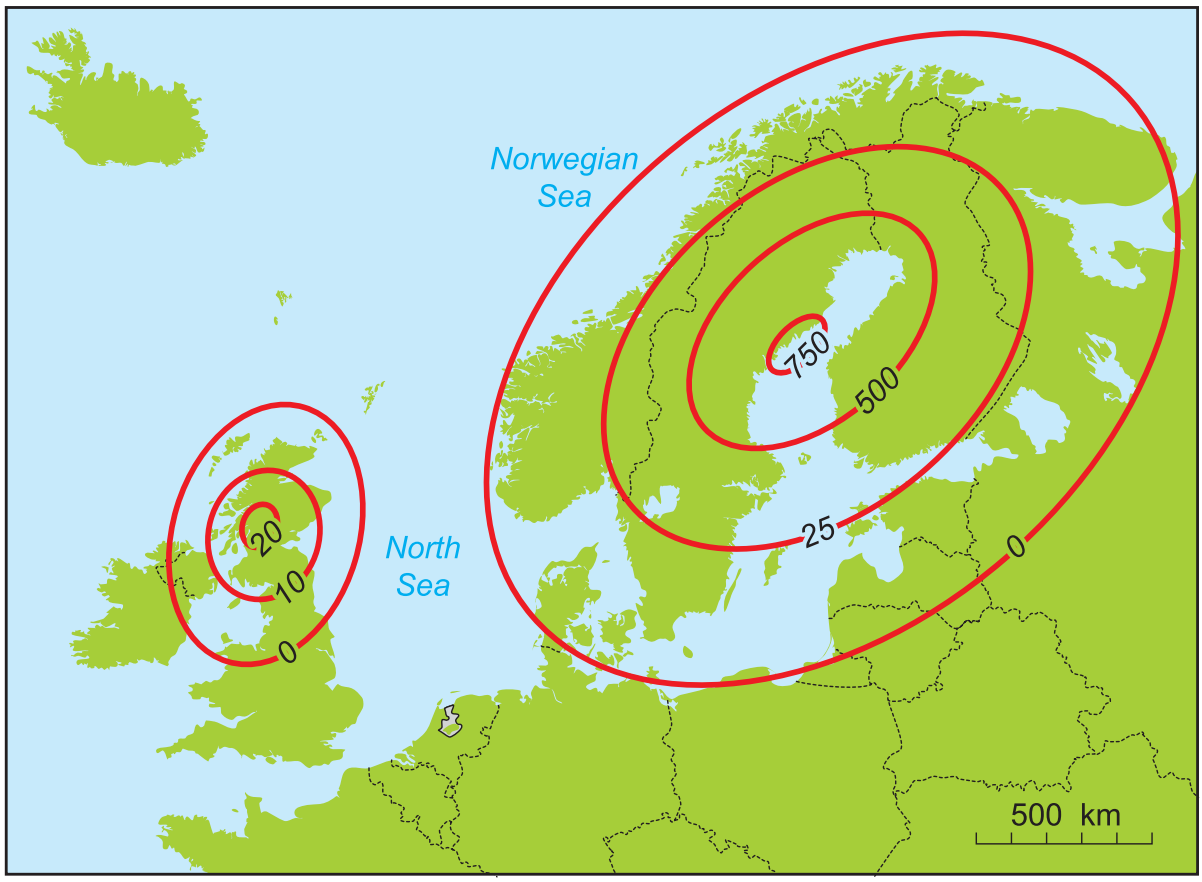


Figure 2.7 Europe: extent of glacio-isostatic rebound (contours in metres) since the melting of the ice sheets from the Last Glacial Maximum.

and isostatic processes interacted. Eustatic inundation was an immediate response to the water released from the melting ice sheets, but isostatic rebound is still going on today. Late-glacial beaches were formed on the west coast of Scotland (Figure 2.8A) as sea level rose and inundated the still depressed coastline. They became raised beaches as the rate of isostatic rebound overtook the rate of sea-level rise. A most impressive sequence of such raised beaches is around Hudson's Bay, depressed under the centre of the Laurentide ice sheet (Figure 2.8B). There is a complex sequence around the Baltic Sea and a similar one in eastern Canada. There, as the Laurentide ice sheet melted, the sea flooded the St Laurence Lowland at a time when eustatic sea levels were still about 50 m below modern sea levels, but the crust was depressed below that level. Since then, isostatic rebound has elevated the saline clay sediments deposited by the 'Champlain Sea' to about 200 m above modern sea levels.

In summary: over the last 500,000 years or so, glacials have tended to last longer than interglacials. We live during an interglacial (the Holocene) which has lasted roughly for the last 10,000 years – and in the absence of

Figure 2.8 Glacio-isostatic modification of shorelines. **A.** Late Pleistocene raised coastal platforms, Loch Linnhe, western Scotland. The flat surfaces of the two islands in the centre of the photo represent two late Quaternary raised coastal platforms. **B.** Satellite image (from ©Google Earth) of the shore of Hudson's Bay, Canada. Note the sub-parallel beach ridges inland from the present shoreline.

any human-induced global warming, could be expected to last for maybe another 10,000 years or so.

What is important, from the point of view of geomorphology, is that present conditions have persisted for approximately only 10,000 years. Most of our landscapes bear the imprint of past conditions, either directly of glaciation or of the effects of previous climates. This has a two-fold importance for geomorphology. Many landforms are relict features, only partially adjusted to present day conditions. However, the landforms and their constituent sediments may preserve evidence for past climatic regimes.

2.3 Global-scale interactions between tectonic and climatic forces

In this chapter so far, we have seen how global plate tectonic patterns and global climatic patterns, past and present, have influenced global geomorphology. One way in which the interaction between global tectonic and climatic forces is expressed is in the relation between uplift and denudation rates. In zones of modern or past destructive plate margins considerable crustal thickening occurs. The relatively light continental crust rises isostatically, giving mountain ranges their high elevation (*see* Section 1.4.1). The high elevations create steep

gradients, stimulating rapid incision by the drainage network. This in turn stimulates high erosion rates, which reduce the mass of the mountain range, stimulating further isostatic uplift. Recent studies of crustal uplift rates, involving complex geophysical methods, have demonstrated a general relationship between uplift rates and estimates of denudation rates, derived from river sediment load data (*see* below). Uplift continues, albeit at a diminishing rate, long after plate-tectonic activity has ceased, because the crustal thickening persists. This 'post-orogenic' process is known as **epeirogenic** uplift and affects large regions, particularly former mountain systems. There are other possible mechanisms, related to mantle processes, that cause epeirogenic uplift. For example, it is thought that some of the relief patterns on the African continent, a long way from past or present plate boundaries, are related to mantle processes. Similarly the sustained uplift of the Colorado Plateau in the American south-west is thought to be due to mantle processes. Indeed, epeirogenic uplift, driven isostatically by erosional offloading is now seen as a possible mechanisms for the continued uplift of continental areas, not only in mountain areas with significant crustal thickening, but also to a

Table 2.1

The "top ten" of the worlds river systems, ranked a) in terms of drainage area, b) river discharge, c) total river sediment load and d) specific sediment load (load /drainage area), the latter set of rivers taken only from large rivers with drainage areas > 500,000 km2 (data source – D Higgitt).

BY DRAINAGE AREA (10^6 km^2)	BY DISCHARGE (km^{3}yr^{-1})	BY TOTAL SEDIMENT LOAD (10^6t yr^{-1})	BY SPECIFIC SEDIMENT LOAD (t km^{-2} yr^{-1})
1. Amazon 6.15	Amazon 6307	Amazon, 1150	Yellow 1102
2. Congo 3.70	Congo 1290	Yellow 1080	Irrawaddy 888
3. Mississippi 3.34	Parana 1101	Ganges 524	Brahamaputra 852
4. Nile 2.72	Orinoco 1101	Brahamaputra 520	Magdelena 846
5. Parana 2.60	Yangtze 899	Yangtze 480	Ganges 535
6. Yenisei 2.58	Mississippi 580	Mississippi 400	Indus 260
7. Ob 2.50	Yenisei 561	Irrawaddy 364	Yangtze 247
8. Lena 2.43	Lena 511	Indus 250	Mekong 198
9. Yangtze 1.94	Mekong 470	Magdelena 220	Amazon 187
10. Amur 1.86	St. Lawrence 451	Mekong 160	Pearl 174

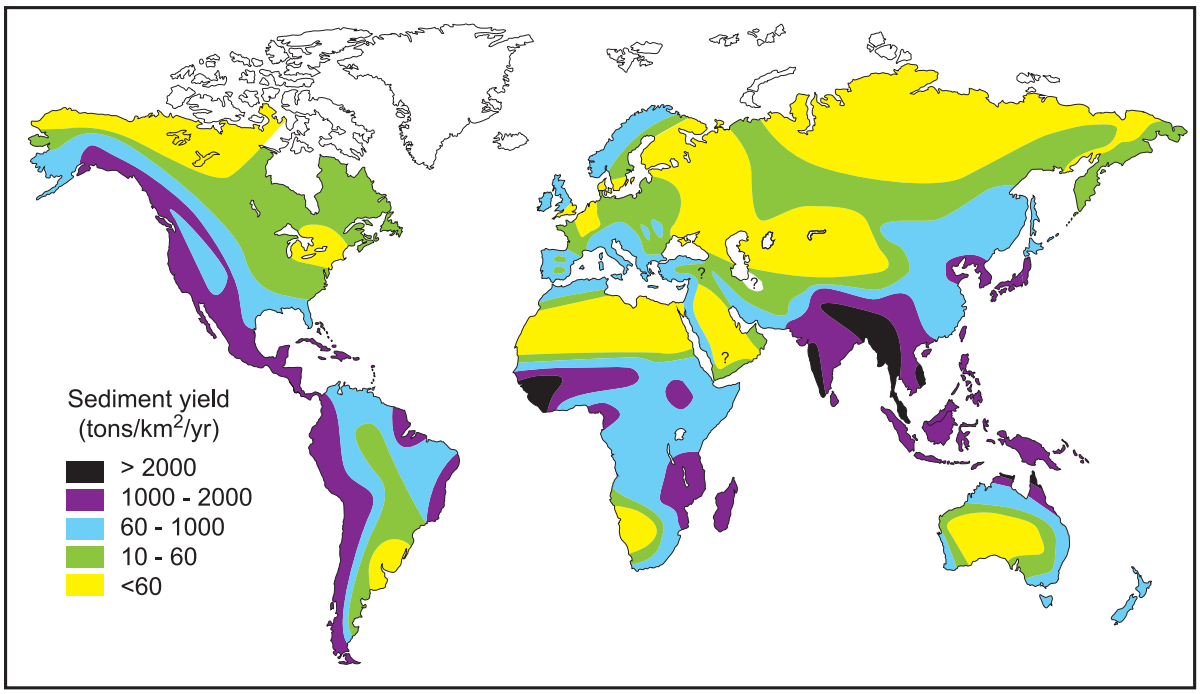


Figure 2.9 World sediment yields.

lesser extent away from mountain areas. The uplift is compensated for by flow of material in the lower crust below the uplifted areas. The reverse appears to take place under crustal loading in subsiding sedimentary basins.

Another way in which the interaction between global tectonic and climatic forces is expressed is in the global patterns of river sediment yield from the continents to the oceans. River sediment loads are a rather crude measure of net erosional amounts within their drainage basins, disregarding within-basin sediment storage. However, they do throw some light on the interactions between the two sets of driving forces. Obviously, in general the largest rivers tend to deliver the largest amounts of sediment to the oceans (Table 2.1). There are, however, some interesting anomalies that reflect the interplay of tectonic and climatic forces. Very high sediment yields are indicative of high rates of erosion. For example, the Yellow River drains the highly erosional and deeply dissected loess plateau of central China, a reflection of the interplay between uplift and Pleistocene climatic conditions. When sediment yields are considered in relation to drainage basin areas (Table 2.1; Figure 2.9), the rivers of southern Asia stand out. These rivers drain the Himalayan mountain belt within a broadly monsoonal climate. High erosion rates here are produced by a combination of rapid tectonic uplift with its associated deep dissection, and climatically-led factors producing intense seasonal precipitation.