can have many different effects (Wilcock, 1983). The power-law *hydraulic geometry*, discussed in Chapter 5, suggests that stream velocity can increase by the increase in channel depth, increase in channel slope, or decrease in channel roughness elements, and by any combination of these effects. Hence, interdependent relationships in geomorphic systems make the prediction of *cause and effect* difficult indeed.

THRESHOLDS

Thresholds denote a threshold of change in the process activity or of the stability of geomorphic form. Reynolds number and Froude (pronounced "frood") number are best-known thresholds in *hydraulics*. The *Reynolds number* differentiates laminar and turbulent flows by viscous and inertial forces, and the *Froude number* distinguishes tranquil and rapid flows by the gravity effect on the fluid flow. A few other thresholds commonly referred to in geomorphology may also be mentioned. *Static* and *dynamic thresholds* respectively quantify the limiting condition for entrainment and continuous transport of sediments by the wind, the *angle of internal friction* just exceeding the slope inclination provides a limiting value for the stability of loose sediments on slopes, and *thermal limits* govern the cold region processes and provide limiting conditions for the distribution of Holocene coral worldwide.

Thresholds are extrinsic and intrinsic to the geomorphic systems (Schumm, 1981). *Extrinsic thresholds* refer to the external effect of climate change or tectonism or both on the functional energy and balance of forces in geomorphic systems. Most landform attributes, however, do not differentiate the climatic signature from the tectonic signal (Derbyshire, 1999).

Intrinsic thresholds (geomorphic) are internal to the process dynamics. They describe the limiting condition of shift between system states. Several geomorphic thresholds, like bankful discharge, channel slope, sediment load, and resistance to flow, have been suggested at which *channel patterns* change in the plan view. At constant discharge, laboratory channels change from straight to meandering and from meandering to braided at two thresholds of *sediment load* in the system (Figure 1.2). Channel patterns also change with a change in the *resistance to flow* on account of long-term variations in the sediment transport rate and, consequently, the evolution of bedform morphology. Coleman and Melville (1994) observe that the channel pattern in experimental alluvial streams changes at a critical height of bedform features. In certain conditions, the *basin area* upstream of gully heads becomes a threshold for the trenching of discontinuous gullies in valley fills (Figure 1.3), and so does the *biomass density*, reaching a low of 5 kg m^{-2} of the channel floor in sand-bedded streams (Graf, 1982). Geomorphic thresholds also exist at the microscopic level of landscape organization (Culling, 1988). The *creep deformation* of soil particles, discussed in Chapter 4, occurs at a threshold of activation energy in the system.

CLIMATE AND PROCESSES

Process domains broadly derive identity from the climatic regime. Hence, a certain set of climatic conditions support particular process domains and landscape characteristics that are different from the geomorphic landscapes of other climatic

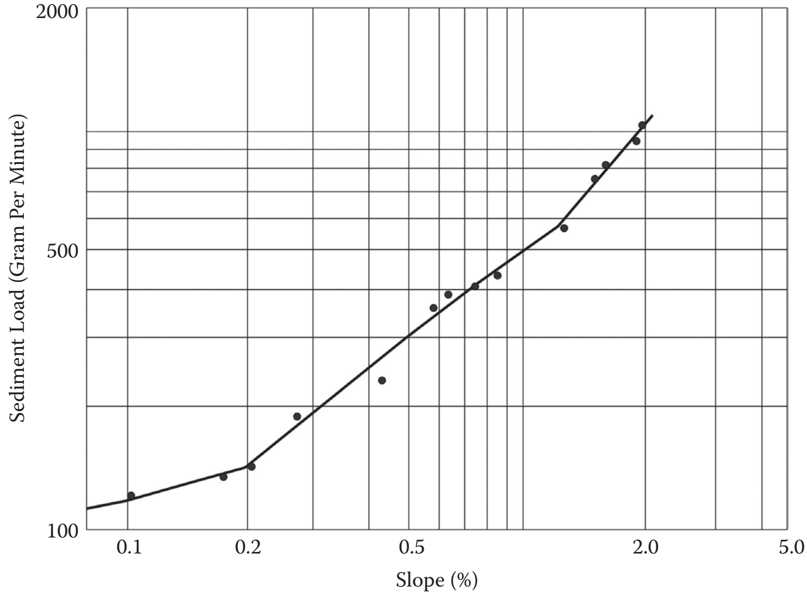

FIGURE 1.2 Relation between rate at which sand was fed into experimental channels and slope of surface on which channels were formed. Breaks in slope of line indicate threshold values of sediment load or slope on which channel pattern changed. (Source: Figure 2 in Schumm, S. A., and Khan, H. R., "Experimental Studies of Channel Patterns," *Nature* 233 [1971]: 407–9. With permission from Macmillan Magazines Ltd.)

conditions. On this analogy, Peltier (1950) proposed nine idealized climate-process systems called *morphogenetic regions* by the amplitude of mean annual temperature and mean annual rainfall amount (Figure 1.4). In essence, morphogenetic regions define limiting conditions for the existence and intensity domains of the earth surface processes. The concept of morphogenetic regions thus implies that landscape characteristics are in dynamic equilibrium with the contemporary processes and process rates (Table 1.1).

Climate, however, is only one among the variables that distinguishes regional landscape characteristics. Several related studies have shown that tectonic, structural, and lithologic aspects of terrain are variously more significant to the evolution of landscape characteristics of certain regions than are the climate-driven processes and process rates (Carson and Kirkby, 1972; Bradshaw, 1982). Hence, the relationship between climate and landscape evolution is imperfect at local and regional scales. The *relict landforms* similarly suggest out-of-phase association of the elements of landscape with the climate of areas in which they exist.

Geomorphic Response to Climate

Although certain aspects of landscape characteristics are broadly related to the present-day climate, the manner of geomorphic response to climate and climate change is far from understood (Graf, 1977; Eybergen and Imeson, 1989; Rinaldo et al., 1995;

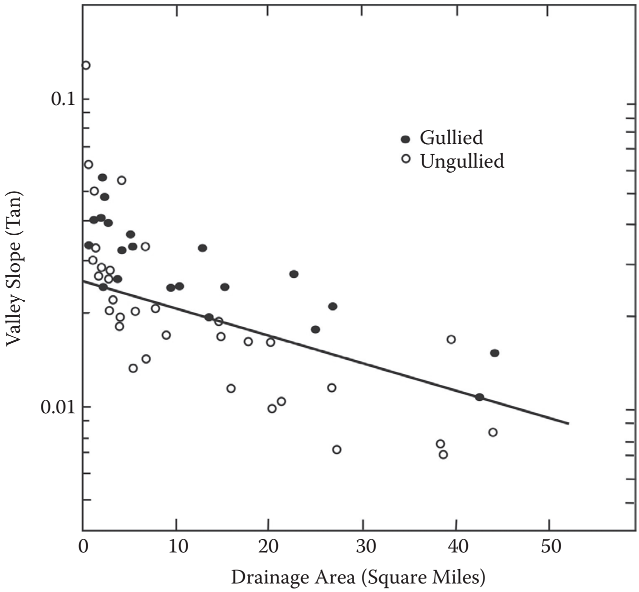

FIGURE 1.3 Relation between valley slope and drainage area, Piceance Creek basin, Colorado. Line defines threshold slopes for this area. (Source: Figure 1 in Schumm, S. A., "Geomorphic Thresholds and Complex Response of Drainage Systems," in *Fluvial Geomorphology*, ed. M. Morisawa [London: George Allen & Unwin, 1981], 299–310. Originally from Patton [unpublished MS thesis, 1973]. With permission from George Allen & Unwin.)

Whipple et al., 1999). Drainage density is one such geomorphic property that varies more with the climatic regime or *erosive potential* of climate than with other factors of terrain (Peltier, 1962). In general, the *drainage density* is highest in semiarid climates, decreases in humid temperate zones, and increases somewhat in humid tropical regions in close agreement with the pattern of mean annual effective rainfall (see Figure 6.10). In statistical terms, 93% variation in the drainage density is due to climate alone (Gregory, 1976), provided climate and terrain remain stable for a sufficiently long period of time (Rinaldo et al., 1995). Over shorter timescales, however, the dependence of form elements on erosive potential of climate becomes obscure and even reversed in semiarid and arid environments (Derbyshire, 1999).

Hillslope geometry is also governed by the erosive potential of regional climate (Toy, 1977). Thus, hillslopes that are shorter and steeper with smaller radii of curvature of the convex segment in arid climates become progressively longer and flatter toward the humid climatic regime (Figure 1.5). Studies subsequently referred to in the following section, however, suggest that the erosive potential of climate itself becomes a negative factor in the development of relief in high-energy alpine environments.

In certain environments, the relationship between *climate change* and evolution of geomorphic systems is indirect or even suspect. Studies from the recently deglaciated landscape of Alaska suggest that progressive dilution of nutrients in the

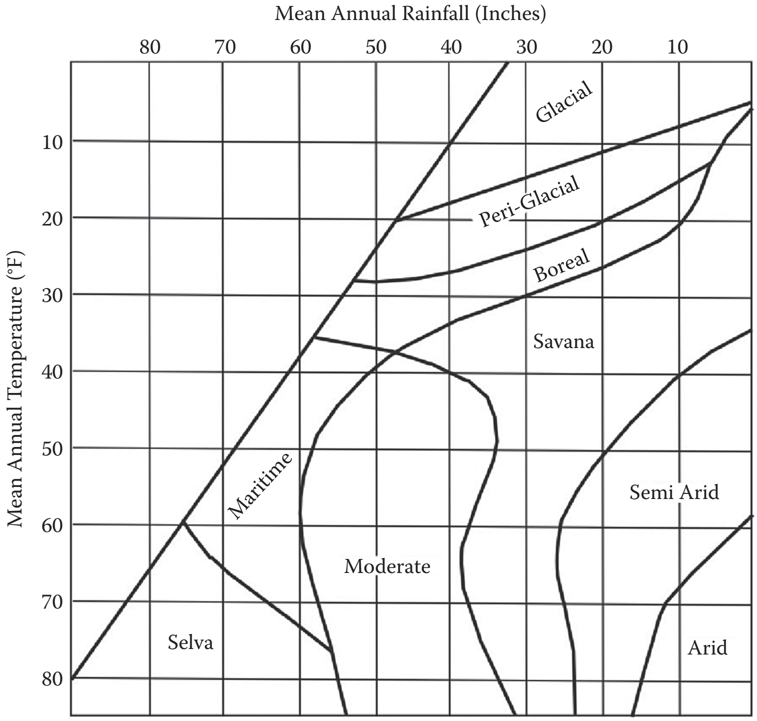

FIGURE 1.4 Climatic controls of morphogenetic regions. (Source: Figure 7 in Peltier, L. C., "The Geographic Cycle in Periglacial Regions as it is Related to Climatic Geomorphology," *Ann. Assoc. Am. Geogr.* 40 [1950]: 214–36. With permission from Routledge Publishers.)

water column of 10,000- to 12,000-year-old freshwater boreal lakes is related more to *successional changes* in vegetation and soil characteristics than to the warming of Holocene climate (Engstrom et al., 2000). Hence, the evolution of *aquatic systems* depends more on the physical and biological manifestation of climate in terrestrial environments (King, 2000). In contrast, the *pollen record* from eastern Amazon suggests that the Holocene warming of climate has had little effect on the *succession rate* of vegetation communities (Bush et al., 2004). Hence, natural systems do not always appear to embody identical signatures of a gradual environmental change.

In recent years, attempts have been made to theoretically evaluate the effect of climate-driven *process rates* on the evolution of landscape characteristics. Molnar and England (1990) predict that a glacial climate is more erosive than the humid climate. Hence, *isostatic compensation* resulting from *offloading* accentuates the mountain relief that furthers the erosive potential of climate. Other theoretical models to the contrary, however, suggest that a humid climate is more erosive than the glacial climate (Summerfield and Kirkbride, 1992), and that glacial erosion does not accentuate relief (Whipple et al., 1999) or add to the cooling of climate (Small, 1999). There is unanimity, however, that the greater erosive potential of climate irreversibly dissects hillslopes to optimum-sized smaller segments, restricting the development of relief in high-energy alpine environments (Schmidt and Montgomery, 1995;

TABLE 1.1
Climatic Controls and Morphologic Attributes of Morphogenetic Regions

Morphogenetic Region	Range of Mean Annual Temperature (°F)	Range of Mean Annual Rainfall (in.)	Process Activity
Glacial	0–20	0–45	Glacial erosion, nivation, wind action
Periglacial	5–30	5–55	Strong mass movement, moderate to strong wind action; weak effect of running water
Boreal	15–38	10–60	Moderate frost action, moderate to slight wind action, moderate effect of running water
Maritime	35–70	50–75	Strong mass movement, moderate to strong action of running water
Selva	60–85	55–90	Strong mass movement, slight effect of slope wash, no wind action
Moderate	35–85	35–60	Maximum effect of running water, moderate mass movement, slight frost action in colder parts, no significant wind action except on coasts
Savanna	10–85	25–50	Strong to weak action of running water, moderate wind action
Semiarid	35–85	10–25	Strong wind action, moderate to strong action of running water
Arid	55–85	0–15	Strong wind action, slight action of running water and mass movement

Source: Peltier, L. C., "The Geographic Cycle in Periglacial Regions as It Is Related to Climatic Geomorphology," *Ann. Assoc. Am. Geogr.* 40 [1950]: 214–36. With permission from Routledge Publishers.

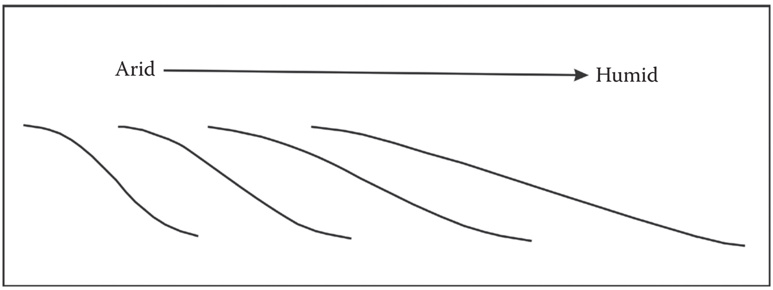

FIGURE 1.5 Conceptualized hillslope profiles. (Source: Figure 4 in Toy, T. J., "Hillslope Form and Climate," *Bull. Geol. Soc. Am.* 88 [1977]: 16–22. With permission from Geological Society of America.)

Whipple et al., 1999). Hence, geomorphic controls override the effect of climate on the development of relief.

FREQUENCY CONCEPT OF GEOMORPHIC PROCESSES

Earth surface processes are characteristic of a certain frequency and magnitude of occurrence, such that high-magnitude events are infrequent and low-magnitude events reoccur frequently in space and time. General observations suggest that high-magnitude, low-frequency events as stream floods seemingly perform a larger amount of work than frequent channel discharges below the bankful stage. In fluvial systems, the work done by geomorphic processes is conveniently expressed by the amount of solute and clastic load transported in the fluvial flow.

The geomorphic work of events is tied down to the magnitude of opposing driving and resisting forces in the system. Sediments are entrained in the fluid flow when the magnitude of fluid force or shear stress just exceeds the resisting force due to the bed and bank materials. This quantity is called *effective stress*. Leopold et al. (1964) predict that the quantity of sediment moved by the fluid flow varies with the lognormal effective stress as

$$Q = s(\tau - \tau_c)^n$$

in which Q is the sediment transport rate, s is the shear strength of materials, τ is the shear (or fluid) stress on the mobile bed, τ_c is the threshold stress for initiation of grain movement, and n is the exponent. The term $\tau - \tau_c$ then is the effective stress. The above relationship suggests that the sediment transport rate increases manifold with a small increase in the effective stress. Hence, high-magnitude stream flow events transport a large amount of sediment per unit of time. By comparison, the low-magnitude events are so frequent that their *cumulative work* exceeds the work of high-magnitude events of low recurrence interval. Leopold et al. (1964) present data on the temporal distribution of discharge and sediment load in selected streams of the United States, which similarly suggest that the cumulative volume of sediment moved is higher for frequent low and intermediate channel discharges than for infrequent overbank flows.

The statistical recurrence interval of extreme events of specified magnitude can be predicted, provided observations are spread over a long period of time. The *recurrence interval* is given as

$$T = (N + 1)/M$$

in which T is the average recurrence interval, N is the period of observation, and M is the rank of extreme event in a series of observations arranged in descending order of magnitude. The recurrence probability of events is given as 1/T, which suggests that high-magnitude events are infrequent and low-magnitude events are frequent in occurrence (Table 1.2).

TABLE 1.2
Probability of Instantaneous Peak Floods at Masani Barrage on the Inland Sahibi Basin in Southern Haryana, India (1965–1984)

Discharge Magnitude (cumec)	Rank Interval (T)	Recurrence (1/T)	Probability
3028.00	1	21.0	0.048
827.00	2	10.5	0.095
657.13	3	7.5	0.143
451.52	4	5.2	0.190
434.12	5	4.2	0.238
433.09	6	3.5	0.286
268.86	7	3.0	0.333
258.05	8	2.6	0.381
254.70	9	2.3	0.429
231.03	10	2.1	0.476
216.32	11	1.9	0.524
189.02	12	1.7	0.571
142.01	13	1.6	0.619
140.79	14	1.5	0.667
139.57	15	1.40	0.714
119.54	16	1.31	0.762
84.11	17	1.23	0.810
60.35	18	1.16	0.857
40.39	19	1.10	0.905
35.23	20	1.05	0.952

Source: Parmar, A. S., "Flow Patterns in Sahibi Basin, Haryana" (MPhil dissertation, Kurukshetra University, 1986).

The frequency concept provides a framework for comparing the work of a known process domain in diverse geomorphic environments. *Water hardness* data from the karst terrain of different climatic environments (Chapter 10) suggest that the rate of limestone/dolomite dissolution is highest in the tropics than in other environments. For equal magnitude of discharge, however, the dissolution rates do not differ much between climatic provinces (Smith and Atkinson, 1976). Rapid evolution of the tropical landscape was similarly once attributed to the high rate of *desilication* (Chapter 3) of terrain. By the equal magnitude of stream flow, however, the rate of silica loss is nearly the same in the tropics and other environments (Douglas, 1969). Hence, differences in the state of landscape development across lithologies and climatic regimes require an alternative explanation. The frequency concept also finds applications in the evaluation of *process thresholds* (Ahnert, 1987; Derbyshire, 1999), and in river training and site planning projects.

ENVIRONMENTAL CHANGE

Components of landscape and landscape characteristics evolve over a long period of time. Hence, landscape attributes inherit the imprints of past process domains and process rates. Biotic components of the landscape that evolve through *natural selection* and *succession* also affect the biochemical processes and, thus, the evolution of landscape through time. Interpretation of the landscape, therefore, requires appreciation of the environmental change. Environmental change, in general, refers to a broad spectrum of temporal changes in climate, sea surface level, continents and seafloor, and flora and fauna (Goudie, 1977). These changes can be reconstructed with confidence to the beginning of the *Quaternary period*.

Interpretation of the condition of the geologic past seeks to establish the cause and magnitude of change, and the geologic date for the initiation and cessation of the given environmental change. Signatures of the environmental change are preserved in *environment-sensitive* indices of certain biotic and abiotic components of the geomorphic systems. The theory, application aspects, and regional significance of some such *proxy indicators* for the interpretation of the environmental change are outlined in this section. Reliable *documentary evidence* for a specified period of written history also provides data of interpretative significance to the study of proximal earth surface events (Nash, 1996; Derbyshire and Goudie, 1997).

Dendrochronology

Absolute chronology of proximal climatic events extendible back in time to the age of trees can be reconstructed from the *tree-ring calendar*. Tree rings evolve by the addition of woody material to the xylem each year. The woody material develops large-sized cells at the beginning of the growth season, which progressively become smaller through time and eventually terminate as thick-walled minute cells at the end of the growth season. This sudden change in the annual appearance of wood tissues constitutes a tree ring. The ring counts in tree stems thus equal the age of trees. The ring widths, however, are sensitive to moisture surplus and deficiency, and thermal amplitude of the environment during the growth cycle of trees. In a general term, larger ring widths denote more optimum moisture and temperature conditions for the growth of trees than do the rings of smaller width.

Dendrochronology has been applied to date various kinds of archaeological remains, such as the pottery made by Pueblo Indians in the Rio Grande Valley of southwestern United States (Leopold et al., 1964). The technique, though, finds best applications in the interpretation of *landform change* and *variability of climate* for the last 3000 to 4000 years at the most. The ring width data from trees along the bed of a gully of accelerated rate of erosion in the Denver area of Colorado suggests that landform change follows the *rate law* (Graf, 1977), discussed in the section on perturbation of geomorphic systems. The ring width and wood density data on *Pinus sylvestris L.* (Scot pine) from northern Sweden suggest that summers were cooler by 0.5°C, and −1°C below the average thirteen times during the last millennium (Briffa et al., 1990). The air temperature calibrated from the ring widths of *Pinus wallichiana* (Blue pine) in the sub-Himalayan environment of northwestern India suggests a positive *ice budget* in 1970–1976, 1981–1984, and 1989 AD for otherwise retreating glaciers of the region (Bhattacharyya and Yadav, 1996). The *oxygen isotope* record in the woody material of trees over ecologically distinct sites in high mountains of northern Pakistan suggests that the last 1000 years of the twentieth century had been the wettest phase of climate on account of human interference with the environment (Treydte et al., 2005).

Pollen

Pollen grains are microspores of seed plants. They are practically indestructible and remain indefinitely preserved in conventional and extreme environments as signatures of the past climate. Pollen-related reconstruction of the environment is based on the assumption that airborne pollen suites are derived from local vegetal associations in the vicinity of ideal accumulation sites in lakes, swamps, and bogs. The frequency distribution of pollen sequence through the depth of the deposit yields data on the paleoenvironment and *habitat* of vegetation communities of the area.

Palynology is ideal for the reconstruction of the *Quaternary climate*. The generic profile of the pollen sequence in the lakebed sediments of southern Thar Desert unveils a continuous record of the variability of Holocene climate for the northern plains of India (Singh, 1971). The pollen data suggest that the climate was intensely arid about 10,000 BC, moderately arid between 7500 and 3000 BC, and severely arid between 800 and 500 BC. Further, short humid phases of climate between 8000 and 7500 BC and in 1800 BC, recording respectively 25 and 50 cm more rainfall than the present-day amount, punctuated the long sequence of Holocene aridity in the northern plains of India.

The pollen preserved in aquatic systems holds immense potential for the reconstruction of *Late Pleistocene climate*. A few pollen-based studies from the Himalayan realm suggest a deglaciation date of 9500 BC at an elevation of 3120 m above the msl (mean sea level) at Toshmaidan in Kashmir Valley (Singh and Agrawal, 1976). The climatic signal of pollen suites from Tso Kar (Kar Lake) at 4535 m above the msl in central Himalayas of northwestern India suggests a warm and moist climate at 30 to 28 thousand years (ka), 22 to 18 ka, 16 ka, and 10 ka for the region (Bhattacharyya, 1989). The moist phase of climate at 10 ka for the northwestern Himalayas, however, coincides with the intense dry phase of the climate in the northern plains of India. The asynchronous moist and dry phases of the climate to the north and south of the Himalayas probably suggest fluctuation in the moisture-bearing monsoon advective system, rather than a change of climate per se (Benn and Owen, 1998; Owen et al., 2001).

The pollen profile from Kashiru Swamps, Burundi, unveils a paleotemperature record for the tropics (Bonnefille et al., 1990). The climatic signal of pollen sequence suggests a *late glacial advance* 21,500 years ago in east Africa, and a temperature 4 ± 2°C below the freezing point for the event, as opposed to the accepted average temperature change of −5 to −7°C required for the onset of glaciation.

Radiometric Methods

Radiometric methods utilize the decay property of unstable natural and induced radioactive elements, and ratios of stable chemical isotopes of the same element for

TABLE 1.3
Half-Life of Primary Radioactive Elements

Element	Half-Life (year)
Natural Radioactive	
Uranium-238	4.5×10^9
Uranium-235	7.1×10^8
Thorium-232	1.39×10^{10}
Rubidium-87	5×10^{10}
Potassium-40	1.3×10^9
Induced Radioactive	
Carbon-14	$5{,}730 \pm 40$

the reconstruction of a paleoenvironment. *Natural radioactive elements* emit radiation of their own, and *induced radioactive elements* acquire radiation from the transmutation of nonradioactive elements (Table 1.3). *Chemical isotopes* fractionate by the physiochemical processes of enrichment that involve a vapor-liquid-solid state of water transformation in the environment.

Natural Radioactive Elements

Natural radioactive elements possess *unstable nuclei*. They spontaneously decay by emitting α and β particles. The decay decreases the number of atoms of radioactive elements, evolving disintegration products that behave chemically in a manner different from preceding and succeeding radioactive products. The disintegration eventually reaches the end product of stable nonradioactive elements. The time for half of the atoms of a given radioactive element to decay into the next disintegration product is called *half-life period*, which, depending upon the radioactive species, varies from a few days to millions of years. The radiometric dating of rocks and sediments is based on the molar ratio of the daughter to the parent nuclide in the specimen.

Several radioactive series are used for the age determination of rocks and the reconstruction of ancient environments. Among these, the uranium-thorium series has been widely used for calibration of the age of the earth. The method has also been applied to sediments of diverse geomorphic environments for the interpretation of paleoenvironment. Among others, the desert varnish on lag deposits of Colorado Plateau is 300,000 years old, suggesting a drier and windier environment for the Late Pleistocene (Knauss and Ku, 1980). The uranium-thorium dating of older marine sediments suggests the midpoint of 13,500 ± 250 years BP for the penultimate deglaciation in the northern hemisphere, a period of time consistent with the tilt and wobble of the earth's orbit around the sun and coral-based reconstruction of the sea level estimate (Henderson and Slowey, 2000).

Radiocarbon Method

Radiocarbon (^{14}C) is continuously produced in the *upper atmosphere* by the bombardment of nitrogen (^{14}N) with high-energy cosmic neutrons (n) as

$$^{14}N + n = {}^{14}C + H^1$$

The radioactive carbon incorporates into the atmospheric carbon dioxide and becomes a part of a wider *food chain system*. The intake of ^{14}C by living plants and animals, however, ceases when they die or become isolated from the atmosphere, such as by burial. The stored ^{14}C in organic remains decays thereafter with the emission of β radiation, evolving a stable nonradioactive nucleus of nitrogen as

$$^{14}C \rightarrow {}^{14}N + \beta$$

The radiocarbon decay has a *half-life period* of 5730 ± 40 years. The age of organic materials, such as wood, charcoal, bone, and shell, is determined from the amount of radioactivity they retain. The radiocarbon method is suitable for the reconstruction of events generally not more than 30 ka old (Deevey, 1952).

The carbon-14 method of environmental reconstruction is ideal for dating Late Quaternary climatic events. The radiocarbon *deglaciation date* of 9160 ± 70 years BP from the base of a peat deposit in the northwestern Himalayas (Owen, 1998) is nearly consistent with the 9600 years BP degalciation date established by other methods on marine sediments in the Gulf of Mexico (Emiliani et al., 1975). The organic matter in *morainic deposits* is 23,000 and 16,000 years old in Tibet (Derbyshire et al., 1991) and 18,100 to 15,700 years old in the northern Karakoram Mountains to the south of Tibet (Li and Shi, 1992), suggesting a phase of late glacial advance in this part of the world. Organic remains in valley fills (Chapter 6) likewise provide radiocarbon dates for the environment favoring channel aggradation.

Chemical Isotopes

Physiochemical processes of enrichment and depletion of geochemical isotopes, and their application for dating climatic events, are discussed in several publications (Hamilton and Farquhar, 1968; Ferronsky and Polyakov, 1982; Bowen, 1991). In general, temperature-related processes of evaporation and condensation in the seawater, and sublimation and solidification of the ice, determine the chemical isotopic composition of oxygen and hydrogen in the seawater and the glacier ice. Thus, the isotopic ratios of two elements are sensitive to the *ambient temperature* of the environment. They are also an appropriate measure of the thermal amplitude of the environment. Oxygen exists as isotopes of ^{16}O, ^{17}O, and ^{18}O, and hydrogen as isotopes of hydrogen (^{1}H), deuterium (^{2}H), and tritium (^{3}H). Records of $^{18}O/^{16}O$ and $^{2}H/^{1}H$ in the seawater and the glacier ice are proxy for *paleotemperatures* of the earth's environment. The isotope method is applicable for dating climatic events generally not more than 30,000 years old.

Global temperature fluctuations during the *Great Ice Age* affected isotopic ratios of oxygen and hydrogen in the seawater and the glacier ice. Experimental studies suggest that the concentration of ^{18}O relative to ^{16}O decreases with decreasing temperature in ice sheets but increases in carbonate-secreting marine organisms called *foraminifera*. The $^{18}O/^{16}O$ data from several sources suggest that the Pleistocene ice grew to massive thickness by slow accretion over thousands of years but thawed quite abruptly within a decade (Bowen, 1991) about 9600 BC (Emiliani et al., 1975), oceans held 4 to 5% more volume of water in the interglacial than glacial phases of the Ice Age climate (Bowen, 1991), the sea level fluctuated by about 100 m during the glacial-interglacial phases of the Pleistocene climate (Shackleton and Opdyke, 1973), and continental ice of the last phase of Pleistocene glaciation 500,000 years ago readvanced five times with a periodicity of about 10,000 years (Schrag, 2000). The ocean-based oxygen isotope paleotemperature record suggests that the extent of Pleistocene aridity and thickness variation of high-latitude loess varies directly with the expanse of continental ice (Shroder et al., 1989; Kemp et al., 1995; Derbyshire, 1999). The record of $^{2}H/^{1}H$ from ice cores in Greenland and Antarctica suggests that the climate of the last glacial maxima warmed 1000 to 2500 years earlier in Antarctica than in Greenland (Blunier et al., 1998). Hence, the asynchronous pattern of Pleistocene climatic change for the two hemispheres requires reassessment of the dynamic behavior of the global climatic system (Steig et al., 1998; Henderson and Slowey, 2000; Scharg, 2000).

Thermoluminescence and Related Dating Methods

Thermoluminescence (TL) is a powerful research tool for the reconstruction of the environment. It is based on interpretation of the record of natural and induced radioactivity in the earth materials. McDougall (1968) and Hurford et al. (1986) discuss the source of TL in the earth materials, laboratory procedures for calibration of the age of rocks and Quaternary sediments, and application aspects of the technique for interpretation of the environment.

Igneous rocks are naturally thermoluminescent. They had trapped uranium and thorium in the structure of their imperfect mineral lattices at the time of formation. This *metastable* locked radiation escapes the mineral structure, and emits glow, when rock specimens are laboratory heated to a very high temperature. The glow intensity varies with the activation energy and duration of heating, producing thermoluminescent *glow curves* that are diagnostic of the age and thermal history of rocks.

Most recent thermal, crystallization, and sun-bleached events also lend to the TL dating. *Contact heating* of the Quaternary sediments by lava flows and firing events of archaeological pottery de-trap the geologic TL. The sediment and pottery, if subsequently buried, acquire radioactivity from the environment of the burial site. *Crystallization events*, like precipitation of tufa and gypsum, begin with *ab initio* zero radioactivity. These chemical sediments, till excavation, also acquire radioactivity from the surrounding environment. *Sun-bleached* weathered and transported sediments rapidly de-irradiate. The sediments, if subsequently buried, likewise acquire radiation from the surrounding environment.

TL-related optically simulated luminescence (OSL), infrared simulated luminescence (IRSL), and electron spin resonance (ESR) methods of environmental reconstruction apply to a variety of sediments and geologic environments. The OSL employs argon laser, IRSL uses an infrared source, and ESR utilizes a microwave signal in a high magnetic field to stimulate luminescence from clastic sediments. In comparative terms, the OSL technique is particularly useful for dating sun-bleached events, the IRSL method is ideal for dating sediments of diverse geologic environments, and the ESR technique is suitable for dating chemical sediments and organic deposits older than Quaternary.

The OSL technique is routinely used for calibrating the age and paleoclimatic environment of aeolian sand, coastal sand, loess, and lacustral deposits. The dated stabilized dunes of the western desert of India are Late Pleistocene in age, and suggest evolution by a slow rate of sedimentation in the Holocene (Singhvi et al., 1982). The coastal sand overlying the emergent continental shelf of eastern China is Early Pleistocene (2.4 million years [Ma]) to most recent in age (2500 years BP). This sand correlates well with the periods of Quaternary sea level oscillation, sequences of marine transgression, and the age of high-latitude loess of eastern China (Mulin, 1986). The TL-dated reworked peri-desert loess of the northwestern Himalayas is 25 to 18 ka old (Owen et al., 1992), and correlative with the Late Pleistocene multiple glaciation of the region (Richards et al., 2000). The OSL-dated glaciogenic sediments in the Himalayas suggest three major glacial advances, of which the 60 to 30 ka Late Pleistocene glacial advance had been the most extensive in the region (Owen et al., 1992; Sharma and Owen, 1996; Owen, 1998). The glaciolacustrine deposits on either side of the Indus Valley in the Gilgit region of the northwestern Himalayas are 38.1 ± 2.6 ka old (Shroder et al., 1989), suggesting ice damming of the channel at 1000 m above the msl (Richards et al., 2000).

The IRSL method has been used for dating sand, loess, and alluvial deposits. The IRSL dates on loess-paleosol sequences of Tadijikistan, and the aeolian sand of Indian Thar Desert, are consistent with the OSL dates on similar sequences of two areas (Mavidanam, 1996). The IRSL-dated Indo-Gangetic alluvium is 20 to 1 ka old across its lateral extent (Mavidanam, 1996).

SUMMARY

Process geomorphology studies the motion and behavior of geomorphic processes at and near the surface of the earth, and explains the evolution of landscape and component landforms in terms of the process dynamics. Geomorphic processes comprise of the activity of wind, water, and ice. These agents of change are considered fluids, as their motion can be analyzed in terms of the fluid mechanics. Processes are quantified in the field, few activities that present difficulty of instrumentation are calibrated in test conditions, and the dynamics of still others is understood in theory only. The work of geomorphic processes varies directly with the process intensity and recurrence frequency. The scale of observation essentially affects the identity of landforms and hypotheses on the evolution of landforms.

An open systems approach provides an ideal framework for the explanation of time-dependent evolution of landscape and time-independent development of spatially contiguous form elements of geomorphic landscapes. The landscape and its form elements evolve in equilibrium with the through flow and continuous exchange of energy and material in geomorphic systems. The state of equilibrium is cyclic, graded, and steady with reference to time, and quasi-equilibrium with respect to the rate of energy expenditure among the interdependent variables of drainage systems. The attributes of geomorphic landscape and its components form elements that are governed by feedback mechanisms, optimum magnitude, and equifinality in process-form relationships.

External effects of climate change, tectonics, and human activities disturb the established order of exchange of energy and material in geomorphic systems. Disturbed geomorphic systems, therefore, pass through reaction and relaxation thresholds, called extrinsic thresholds, before reaching the steady-state equilibrium. The rate of process-form adjustment in systems disturbed by the external stimulus of energy is exponential in nature. Geomorphic thresholds are intrinsic to the process activity. The intrinsic thresholds are due to internal causes that affect the shear stress and, thereby, process rates in geomorphic systems.

Regional climate broadly regulates the earth surface process domains. Hence, geomorphic landscape is visualized to evolve in equilibrium with the contemporary process domains of respective climatic regimes. The hypothetical concept of morphogenetic regions discusses a broad-based climate-process-form relationship for nine hypothetical regions. However, the manner of geomorphic response to climate and climate change is far from simple. Most often, physical and biochemical controls of landscape override the effect of climate on process rates. Theoretical studies on process-form relationships also suggest that climate itself becomes a limiting factor in relief development of the high-energy alpine systems.

Landscape and component landforms evolve over a long period of time, during which process domains may have changed altogether or process rates may have varied considerably due to a change in the intensity of process activity. Hence, an explanation of landforms also requires evaluation of the environmental change. Several environment-sensitive proxy indicators are used for interpreting the type, magnitude, and duration of environmental change. Depending upon the nature of indicator source, dendrochronologic, palynologic, radiometric, chemical isotopic, and themoluminescence dating techniques find applications in the reconstruction of the Quaternary environment. Dendrochronology interprets the ring width data for deducing the hydrologic and thermal amplitude of the proximal environment. A depth profile of pollen sequences in lakebed sediments and peat deposits unveils a continuous record of the magnitude of moisture and temperature change in the environment. A record of the radiocarbon in organic remains enables interpretation of the environment and environmental change to within the safe limit of 30 ka before present. The chemical isotope ratio of ^{18}O and ^{16}O in the ice and skeletal remains of deep sea organisms, called foraminifera, is a proxy for the Quaternary ambient temperature, sea level change, and chronology of climatic events. The isotope ratio of ^{2}H and ^{1}H in the seawater and the glacier ice is a measure of the amplitude of paleotemperature of the earth's past environment. The property of luminescence due to geologic and induced radioactivity in rocks and sediments finds applications in calibrating the age and paleoclimatic environment of the earth materials of diverse origins.

REFERENCES

Abrahams, A. D. 1968. Distinguishing between the concepts of steady state and dynamic equilibrium in geomorphology. *Earth Sci. Rev.* 2:160–66.

Ahnert, F. 1967. The role of the equilibrium concept in the interpretation of landforms of fluvial erosion and deposition. In *Evolution des Versants*, ed. P. Macar, 23–41. Liege: L'Université de Liege.

Ahnert, F. 1987. An approach to the identification of morphoclimates. In *International geomorphology*, ed. V. Gardiner, 159–88. Vol. 2. Chichester: John Wiley & Sons.

Benn, D. I., and Owen, L. A. 1998. The role of Indian summer monsoon and mid-latitude westerlies in Himalayan glaciation: Review and speculative discussion. *J. Geol. Soc. Lond.* 155:353–63.

Bhattacharyya, A. 1989. Vegetation and climate during the last 30,000 years in Ladakh. *Palaegeogr. Palaeobot. Palaeoecol.* 73:25–38.

Bhattacharyya, A., and Yadav, R. R. 1996. Dendrochronological reconnaissance of *Pinus wallichiana* to study glacial behaviour in western Himalaya. *Current Sci.* 70:739–43.

Blunier, T., Chappellaz, J., Schwander, J., Dällenbach, A., Stauffer, B., Stocker, T. F., Raynand, D., Jouzel, J., Clausen, H. B., Hammer, C. U., and Johnsen, S. J. 1998. Asynchrony of Antarctic and Greenland climate change during the last glacial period. *Nature* 394:739–43.

Bonnefille, R., Roeland, J. C., and Guiot, J. 1990. Temperature and rainfall estimates for the past 40,000 years in equatorial Africa. *Nature* 346:347–49.

Bowen, R. 1991. *Isotopes and climate*. London: Elsevier Applied Science.

Bradshaw, M. 1982. Process, time and the physical landscape: Geomorphology today. *Geography* 67:15–28.

Briffa, K. R., Bartholin, T. S., Eckstein, D., Jones, P. D., Karten, W., Schweingruber, F. C., and Zetterberg, P. 1990. A 1,400-year tree-ring record of summer temperatures in Fennoscandia. *Nature* 346:434–39.

Brunsden, D. 1988. Slope instability, planning and geomorphology in the United Kingdom. In *Geomorphology in environmental planning*, ed. J. M. Hooke, 105–19. Chichester: John Wiley & Sons.

Brunsden, D., and Thornes, J. B. 1979. Landscape sensitivity and change. *Trans. Inst. Br. Geogr.* 4:463–84.

Bull, W. B. 1991. *Geomorphic response to climatic change*. New York: Oxford University Press.

Bush, M. B., Silman, M. R., and Urrego, D. H. 2004. 48,000 years of climate and forest change in a biodiversity hot spot. *Science* 303:827–29.

Carson, M. A., and Kirkby, M. J. 1972. *Hillslope form and processes*. Cambridge: Cambridge University Press.

Chorley, R. J. 1962. *Geomorphology and general systems theory*. U.S. Geological Survey Professional Paper 500-B. U.S. Geological Survey.

Chorley, R. J. 1978. Bases for theory in geomorphology. In *Geomorphology: Present problems and future prospects*, ed. C. Embleton, D. Brunsden, and D. K. C. Jones, 1–13. Oxford: Oxford University Press.

Church, M., and Slaymaker, O. 1989. Disequilibrium of Holocene sediment yield in glaciated British Columbia. *Nature* 337:452–54.

Coleman, S. E., and Melville, B. W. 1994. Bed-form development. *J. Hydraul. Eng.* 120:544–60.

Culling, W. E. H. 1987. Equifinality: Modern approaches to dynamical systems and their potential for geographical thought. *Trans. Inst. Br. Geogr.* 12:57–72.

Culling, W. E. H. 1988. A new view of landscape. *Trans. Inst. Br. Geogr.* 13:345–65.

Deevey, E. S., Jr. 1952. Radiocarbon dating. *Sci. Am.* 206:24–28.

Derbyshire, E. 1999. Landforms, geomorphic processes and climatic change in drylands. In *Palaeoenvironmental reconstruction in arid lands*, ed. A. K. Singhvi and E. Derbyshire, 1–26. New Delhi: Oxford & IBH.

Derbyshire, E., and Goudie, A. S. 1997. The drylands of Asia. In *Arid zone geomorphology of desert environments*, ed. S. G. Thomas, 487–506. 2nd ed. Chichester: John Wiley & Sons.

Derbyshire, E., Shi, Y., Li, J., Zheng, B., Li, S., and Wang, J. 1991. Quaternary glaciation of Tibet: The geological evidence. *Quat. Sci. Rev.* 10:485–510.

Douglas, I. 1969. The efficiency of humid tropical denudation systems. *Trans. Inst. Br. Geogr*. 46:1–16.

Emiliani, C., Gartner, S., Eldridge, K., Elvey, D. K., Huang, T. C., Stipp, J., and Swanson, M. F. 1975. Palaeoclimatological analysis of Late Quaternary cores from the northeastern Gulf of Mexico. *Science* 189:1083–88.

Engstrom, D. R., Fritz, S. C., Almendinger, J. E., and Juggins, S. 2000. Chemical and biological trends during lake evolution in recently glaciated terrain. *Nature* 408:161–66.

Eybergen, F. A., and Imeson, A. C. 1989. Geomorphological processes and climatic change. *Catena* 16:307–19.

Ferronsky, V. I., and Polyakov, V. A. 1982. *Environmental isotopes in the hydrosphere*, English trans. S. V. Ferronsky. Chichester: John Wiley & Sons.

Goudie, A. S. 1977. *Environmental change*. Oxford: Clarendon Press.

Graf, W. L. 1977. The rate law in fluvial geomorphology. *Am. J. Sci*. 277:178–91.

Graf, W. L. 1982. Spatial variation of fluvial processes in semi-arid lands. In *Space and time in geomorphology*, ed. C. E. Thorn, 193–217. London: George Allen & Unwin.

Gregory, K. J. 1976. Drainage networks and climate. In *Geomorphology and climate*, ed. E. Derbyshire, 289–315. London: John Wiley & Sons.

Hack, J. T. 1960. Interpretation of erosional topography in humid temperate regions. *Am. J. Sci.* 258-A:80–97.

Hamilton, E. I., and Farquhar, R. M., eds. 1968. *Radiometric dating for geologists*. London: Interscience Publishers.

Henderson, G. M., and Slowey, N. C., 2000. Evidence from U-Th dating against Northern Hemisphere forcing of penultimate deglaciation. *Nature*, 404, 61-66.

Howard, A. D. 1982. Equilibrium and timescales in geomorphology: Application to sand-bed alluvial streams. *Earth Surface Processes and Landforms* 7:303–25.

Howard, A. D. 1988. Equilibrium models in geomorphology. In *Modelling geomorphological systems*, ed. M. G. Anderson, 49–72. Chichester: John Wiley & Sons.

Huggett, R. J. 1980. *Systems analysis in geography*. Oxford: Clarendon Press.

Hurford, A. J., Jäger, E., and Ten Cate, J. A. M., eds. 1986. *Dating young sediments*. CCOP Technical Publication 16. Bangkok: CCOP.

Kemp, R. A., Derbyshire, E., Meng, X. M., Chen, F. H., and Pan, B. T. 1995. Pedosedimentary reconstruction of a thick loess-palaeosol sequence near Lanzhou in north-central China. *Quat. Res*. 43:30–45.

King, G. W. 2000. A lake's life is not its own. *Nature* 408:149–50.

Kirkby, M. J. 1990. Landscape viewed through models. *Z. Geomorph.* 79:63–81.

Knauss, K. G., and Ku, T. 1980. Desert varnish: Potential for age dating via uranium series isotopes. *J. Geol*. 88:95–100.

Knighton, D. 1984. *Fluvial forms and processes.* London: Edward Arnold.

Lamberti, A. 1992. Dynamic and variational approaches to the river regime relation. In *Entropy and energy dissipation in water resources*, ed. V. P. Singh and M. Fiorentino, 505–25. Dordrecht: Kluwer Academic Publishers.

Langbein, W. B., and Leopold, L. B. 1964. Quasi-equilibrium states in channel morphology. *Am. J. Sci*. 262:782–94.

Leopold, L. B., and Langbein, W. B. 1962. *The concept of entropy in landscape evolution*. U.S. Geological Survey Professional Paper 500-A. U.S. Geological Survey.

Leopold, L. B., Wolman, M. G., and Miller, J. P. 1964. *Fluvial processes in geomorphology*. San Francisco: W. H. Freeman and Co.