10

Water

> All indurated rocks and most earths are bound together by a force of cohesion which must be overcome before they can be divided and removed. The natural processes by which the division and removal are accomplished make up erosion. They are called disintegration and transportation. Transportation is chiefly performed by running water.
>
> …A portion of the water of rains flows over the surface and is quickly gathered into streams. A second portion is absorbed by the earth or rock on which it falls, and after a slow underground circulation reissues in springs. Both transport the products of weathering, the latter carrying dissolved minerals and the former chiefly undissolved.
>
> G. K. Gilbert, *Geology of the Henry Mountains* (1880)

The Earth's surface is dominated by landforms that have been carved by running water. Fluvial landforms are usually apparent in even the driest deserts. Running water is such an effective agent of erosion because of its density: Almost 1000 times denser than air, it exerts greater shear stress, buoys the weight of entrained particles, and is driven more forcefully by gravity than an equivalent volume of air.

Where rainfall is possible, even small amounts of water trump any other agent of erosion. Although rain is not possible on Mars under current conditions, its landscape plainly bears the scars of rainfall in the distant past. Some things are different: Mars has seen enormous floods that are comparable to the largest floods known on Earth, and groundwater sapping plays (or played) a far larger role than it does on our planet. We still do not understand how Mars' floods originated or how such large volumes of water came to be suddenly released. Nevertheless, once released, the water followed the same laws as water on Earth and produced landforms for which terrestrial analogs exist.

Recent exploration of Titan reveals that active fluvial landscapes are not unique to Earth: Although the fluid on Titan is liquid methane at 95 K and the rocks are hard-frozen water ice, Titan's landscapes of stream valleys, riverbeds, and enormous lakes are eerily familiar. The materials are different but the processes are similar and can be analyzed by the same methods as the Earth's water-carved surface. In the following chapter, as in its title, the word "water" is used to keep the discussion focused, but most of the concepts apply to any fluid interacting with a solid substrate.

As we explore fluvial processes in this chapter we will emphasize general laws and relationships that hold where any fluid interacts with a solid surface. Whether on Earth, Mars, or Titan, similar physical laws lead to similar landforms. Process trumps contingency here.

10.1 "Hydrologic" cycles

It is a familiar concept that rain falling on the land suffers a number of possible fates, returning eventually to the sea or evaporating into the air to continue the cycle that made rivers seem eternal to our ancestors. Rain that falls on the surface can infiltrate into the ground, where it joins the volume of groundwater beneath the surface, evaporate immediately back into the air, or run over the surface to collect into streams and possibly large bodies of standing water. On Earth, transpiration by plants is also an important factor, but plants, so far as we know, play no role on either Mars or Titan.

The relative quantities of these factors, along with the movement of water underground, are the subject matter of the field of *hydrology*. We will be mainly interested in that part of the cycle which results in runoff and can, thus, perform work on the landscape.

10.1.1 Time, flow, and chance

The climates of the Earth are usually characterized by their temperature and annual rainfall. The annual rainfall statistic specifies how much water arrives as rain in an "average" year. However, as everyone knows, the amount of rainfall in a given year can fluctuate greatly. Moreover, the pattern in which the rainfall arrives is also important – clearly, 30 cm of rain per year that arrives in several large, brief thundershowers is more effective in eroding the landscape than 30 cm of rain arriving in gentle, continuous showers.

In general, most geologic work is done by large, relatively infrequent events. Most mountain streams run clear over large boulders that they obviously cannot move. Such boulders are transported only during extreme rains when the clear rivulet may become a muddy, churning torrent for a few hours. The level of water in streams may reach as high as their banks only once a year, but that annual flood determines the form of their channel. Much less frequent floods may sculpt their higher floodplains.

It is important to realize the significance of fairly rare events in surface processes. Consider the example shown in Figure 10.1. The upper panel depicts a typical relationship between streamflow (which is proportional to the friction velocity) and the amount of sediment moved (e.g. Equation (9.15) for wind-blown sand: Sediment in streams obeys a similar equation). The middle panel is a probability distribution showing the likelihood of a given streamflow. It peaks at the mean discharge. Discharges both lower and higher than average are less likely. The curve becomes increasingly less certain as the discharge increases: Very rare events that are, nevertheless, possible may never have been observed within the time frame over which records are available. The product of the two curves is shown in the lower panel. This yields the probability of a given sediment discharge.

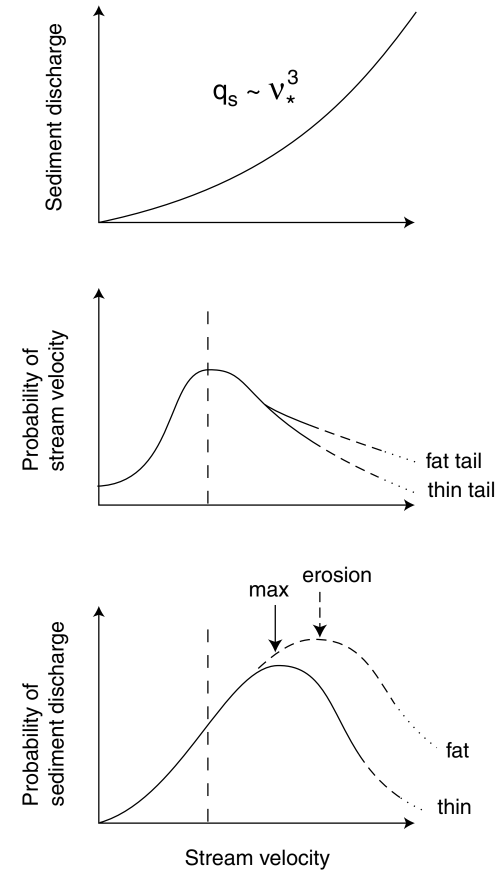

Figure 10.1 The amount of geological work done by flowing water depends on the cube of the velocity, as shown in the top panel. The probability of finding a given velocity peaks at some stream velocity between zero and infinity, as shown in the middle panel, although the high-velocity behavior is somewhat uncertain, depending on whether the probability falls exponentially (thin tail scenario) or as a power-law function of the velocity (fat tail scenario). The lower panel shows that the product, the probability of a given sediment discharge, peaks at a higher stream velocity than the maximum velocity.

Because the amount of sediment moved by a given flood is a strongly increasing function of the streamflow, the peak of the lower curve is displaced away from the mean flow, toward a higher than average streamflow. Thus, the event that is most likely to move a large amount of sediment, and so, cause a large amount of geologic change, is not the annual average but a rare event that corresponds to an unusual flood.

The extreme end of the probability curve is the subject of major debates at the moment. Because it corresponds to very rare events, there is not a lot of data to tie down the precise form of the right-hand end of the curve, and these uncertainties are greatly magnified by the large amount of sediment moved in extreme events. This debate is connected with the general "fat tail" problem that haunts large, rare events. If one assumes that the shape of the probability curve is Gaussian, the probability of events far from the mean drops exponentially fast with distance from the mean (a "thin tail"). However, if the curve falls as a power law (a "fat tail"), then the probability drops much more slowly. In an extreme case, if the curve falls as a power law less steep than $1/v^3$, where v is the stream velocity, then there is no maximum at all and erosion is dominated by the most rare, but largest events! Water floods do not seem to work this way, although meteorite impact events do, because there is no meaningful upper limit to the size of a potential impactor.

As an example, the mean annual discharge of the Columbia River is about 7800 m^3/s. It fluctuates during an "average" year from a low of about 2800 m^3/s to a high of about 14 000 m^3/s in the late spring. However, there are wide fluctuations about these means: The lowest flow ever recorded is about 1000 m^3/s, while the peak is about 34 000 m^3/s. Nothing in these records, however, prepares one for the fact that between 15 000 and 13 000 yr ago the Columbia River drainage suffered a series of catastrophic floods that increased the discharge to about 10^7 m^3/s for a few days. These floods were associated with the collapse of ice dams holding back the waters of Glacial Lake Missoula. The outpouring of water from these floods scarred the surface of eastern Washington, digging huge channels, transporting blocks of basalt tens of meters across, and leaving enormous deposits of gravel in a series of catastrophic floods that did more geologic work in a few days than millions of years of normal erosion could accomplish (Baker and Nummedal, 1978).

Geologist Gene Shoemaker reached a similar conclusion in 1968 after he repeated Powell's historic trip down the Colorado River through the Grand Canyon. Shoemaker reoccupied 150 sites that Powell's photographers had recorded in 1871–1872 and compared the images of the canyon taken almost 100 yr apart. In looking at these images (Stephens and Shoemaker, 1987), one is struck by the fact that, apart from predictable changes in vegetation, very few differences are seen in most images. Individual rocks can be recognized in the same places as 97 yr previously. However, in a small number of comparisons, the scene has changed utterly: almost nothing has remained the same. These drastic changes are due to unusual floods originating in side canyons of the river that cleared away massive heaps of rock and deposited new piles of rock debris from further upstream.

The lesson Shoemaker learned about stream erosion is that it is episodic: little may change for a long time until an unusual event comes along that makes sudden, large alterations. The fluvial landscape does not change gradually, but evolves in a series of jumps whose effects accumulate over time.

This kind of evolution is called catastrophic: The strict definition is that a catastrophe is a large event that causes more change than all smaller events combined. Floods fulfill this definition within limits: For floods, the curve in the bottom panel of Figure 10.1 eventually turns over (although the Lake Missoula floods do cause some concern about the validity of this claim). This is not the case for meteorite impacts. Similar probability curves have yet to be established for many other processes.

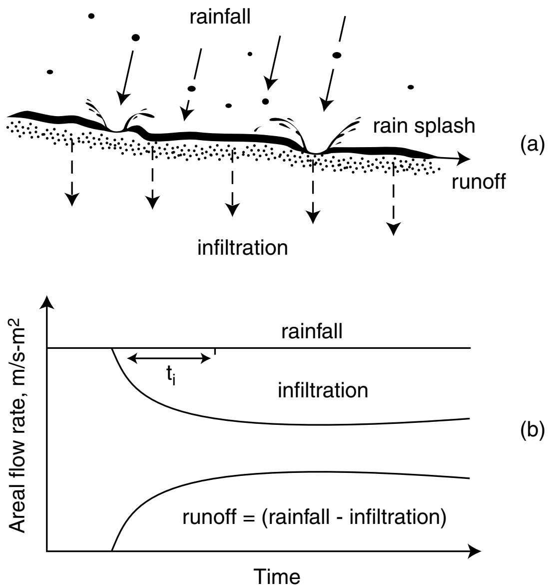

Figure 10.2 When rain begins to fall on a porous surface it initially all soaks into the ground. Once the capacity of the surface to absorb water is saturated, however, the water collects in a thin film on the surface and runs off. Panel (a) shows this process schematically, as well as the disturbance of the granular surface by the small impacts of raindrops. Panel (b) indicates the volumetric division of rainfall into infiltrated water and runoff, which reaches a steady state after an infiltration relaxation time t_i.

10.1.2 Rainfall: infiltration and runoff

Rainfall is ubiquitous on the Earth wherever temperatures permit liquid water to exist, although some regions may receive more, others less. Rain also seems to have fallen on the ancient cratered highlands of Mars to create valley networks and, although it has not been observed directly, must be occurring in near-current times on Titan (except that it is methane drops that fall there, not water).

The first drops of water that fall on the ground surface are trapped in small surface depressions and irregularities – no runoff is produced. However, as more rain falls, the capacity for depression storage is usually exceeded after some period of time. This moment is easily observed while watching rain fall on a bare soil surface, for as soon as a connected sheet of water forms the entire surface seems to suddenly glisten and water begins to flow down whatever slope is present. Some water soaks into the ground: The amount depends upon the permeability of the soil, among other factors. The remainder begins to flow along the surface. This runoff is responsible for most geologic work (Figure 10.2a).

Runoff is equal to the difference between the rate at which rainwater arrives on the surface and the rate at which it infiltrates into the surface (Figure 10.2b). If the infiltration capacity is high enough or the rain gentle enough, there may be no runoff at all. However, as the rainfall intensity increases, the amount of infiltration remains approximately constant while the runoff necessarily increases.

The process of infiltration has received much study, and there are entire books devoted to it (Smith *et al.*, 2002). A great deal of this study has practical ends: It is often considered desirable to increase infiltration to both reduce erosion and increase groundwater supplies. Because infiltration involves only the partly saturated flow of water into loose soil that already contains air, the process is very complex and depends on many factors. Chief among them is the intensity of the rainfall (volume of water per unit time per unit area) and the duration of the rain, along with its integrated volume. Raindrops falling on a bare surface usually beat down and compact the soil, reducing the infiltration capacity as time goes on. Equilibrium between rainfall and infiltration is reached after a relaxation time t_i, after which the rate of infiltration becomes approximately constant, whatever the rainfall intensity or duration might be.

One factor that most geologists take for granted is the ability of water to "wet" silicate minerals. Capillary forces play an important role in infiltration and the surface contact energy between water and minerals is crucial to the ability of water to flow into the pores between mineral grains in the soil. But what about methane on Titan? Does liquid methane flow into the pores between grains of cold ice? If methane beads up on the ice surface like water on a well-waxed car, Titan might not have a subsurface hydrologic cycle at all. Fortunately, recent experiments (Sotin *et al.*, 2009) show that not only does liquid methane wet ice, it soaks into the tiniest cracks and pores on contact. Quantitative data on surface energy is still lacking, but the viscosity of liquid methane is only about 10% that of water (Table 9.2), so that liquid methane may readily infiltrate into the Titanian surface.

On Earth, the infiltration capacity depends upon the condition of the soil surface. Vegetation plays a big role here, as does the type of soil. Clays, for example, are very impermeable and, thus, have a low infiltration capacity, which promotes runoff and hence surface erosion. Gullying on clay-rich badlands is intense. On the other hand, coarse sands and loose volcanic cinders on the sides of fresh cinder cones are highly permeable and may entirely suppress runoff. A frequent observation is that fresh cinder cones stand for long periods of time without any sign of gullies, in spite of being composed of loose, often cohesionless, volcanic lapilli. However, with time, weathering eventually converts the glassy lapilli to impermeable clay and wind-blown dust settles between the lapilli, filling the pores between them. When this has gone far enough, runoff finally begins and the cone is removed in a geologic instant: In a field of cinder cones it is common to see ungullied fresh cinder cones, but rare to see gullied cones. Instead, one finds lava flows whose original vents lack cinder cones – they have been removed by fluvial erosion. Cinder cones in volcanic fields on Venus are far more abundant than on Earth, presumably because of the greater efficacy of fluvial erosion on our planet.

On Earth, other highly permeable deposits may display a strong resistance to erosion, not because of intrinsic strength but because of their ability to soak up rainfall and prevent runoff. Highly fractured lava plains often have very little runoff and may, thus, be very long-lasting. Even gravel may be highly resistant to erosion for this reason (Rich, 1911).

Water that infiltrates into the soil may percolate down to join the water table, flow laterally to seep out as springs or into streams (meanwhile initiating sapping erosion that may undermine the slopes out of which it flows), or evaporate from the surface, depending upon the climate, geologic structure beneath the surface, and rock permeability. The fate of rainfall is, thus, complex and depends upon many factors, so that simplified models of the type presented here may not be realistic: Much current work in hydrology and geomorphology is focused on overcoming the drawbacks of overly simplistic models.

10.2 Water below the surface

On the Earth, every vacant space in rocks underground is occupied by fluid: water, brine, oil, natural gas, or, very close to the surface, air. Most of these fluids reside within the upper few kilometers of the surface, but significant porosity and permeability may extend to depths of tens of kilometers. When piezometric pressures differ in these fluids they tend to flow from areas of high pressure to low pressure. Underground flows can be as important as flows above ground in moving dissolved substances, contaminants, and potential resources. On Mars, where surface water is not stable under the present climate, a global underground hydrologic cycle has been proposed to explain sapping and catastrophic outbursts of water on the surface. Seeping brines may account for some of the modern-day gullies on crater walls.

An understanding of how underground water moves and interacts with the surface was first achieved in 1856 by French engineer Henry Darcy (1803–1858), who investigated the water supply of Dijon, France. Percolation of fluids through fractured rocks has been vigorously studied ever since, driven largely by economic concerns with water supply, oil exploration, and, at present, pollution remediation. In more modern times, M. King Hubbert (1940) made vital contributions to this field.

10.2.1 The water table: the piezometric surface

When a well is drilled into the surface of the Earth, water is usually not immediately encountered. Although some water is often found adhering to mineral grains, the zone in which all the pores are filled with water is only reached at some depth. The top of this saturated zone is known as the water table. It is defined by the level at which water stands in an open well. Above this fully saturated zone is a thin, partially saturated layer (the “capillary fringe”) in which capillary forces raise water a small distance against gravity.

If ground water is stagnant and suffers neither gain nor loss, the water table beneath the surface is level, conforming to a surface of constant gravitational potential (Figure 10.3a). However, when ground water is recharged by infiltration or drained by discharge, its upper surface becomes complex, crudely reflecting the overlying topography in regions where the permeability is uniform (Figure 10.3b).

To describe how water moves underground it is necessary to define the force that makes it move. Simply citing a pressure is not enough, for the pressure varies with depth in a

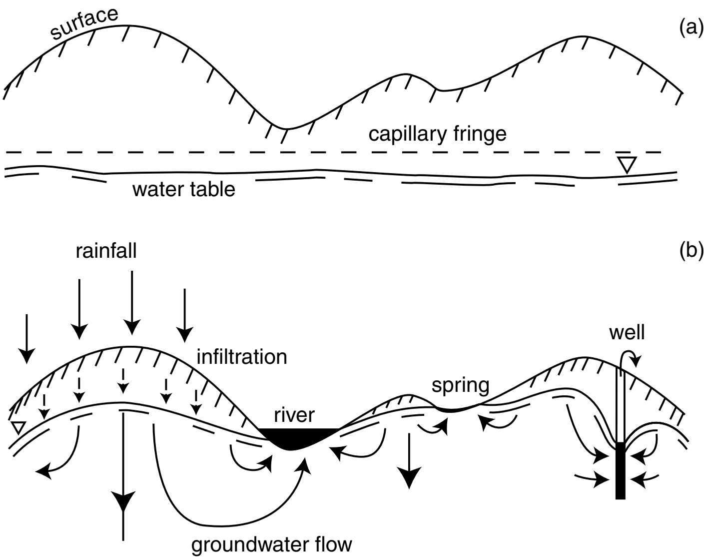

Figure 10.3 If there is no recharge of the groundwater, as in panel (a), the water table with its capillary fringe relaxes to parallel an equipotential surface, whatever the topographic complexity above. When rainfall recharges the groundwater, as in (b), the water table tends to mimic the topographic variations. Springs and streams of water emerge from the ground where the water table intersects the surface. Extracting water from a well, shown on the right of this figure, locally draws down the water table and creates a "drawdown cone."

stagnant fluid because of the weight of the fluid. The most meaningful measure of the tendency of a fluid to move is its total energy per unit mass, which is the sum of its gravitational potential energy, kinetic energy, and the work done by the pressure on the fluid volume. Energy is defined only up to an arbitrary constant, Φ_0. Assuming that we can treat water as incompressible, the total energy of a unit mass of water underground is:

$$\Phi - \Phi_0 = gz + \frac{P}{\rho} + \frac{v^2}{2} \tag{10.1}$$

where z is elevation above some datum (measured positive upwards), P is the pressure, ρ the (constant) density, and v the fluid velocity. Because water underground usually moves slowly, the last term can be dropped and the energy equation is written:

$$\Phi - \Phi_0 = gz + \frac{P}{\rho}. \tag{10.2}$$

In a stagnant fluid Φ is constant and the pressure $P = P_0 - \rho g z$, increasing as z becomes more negative (deeper).

Hydrologists usually do not deal directly with fluid energy, but instead work with the *hydraulic head*, h, which is the level above a datum to which a fluid would rise in a tube inserted at the point of interest. By definition, $h = (\Phi - \Phi_0)/g$. Variations in head express variations in the potential energy capable of driving fluid flow. The head is constant if the fluid is stagnant, but if the head varies laterally, the fluid tends to flow from regions with higher head to lower head.

A piezometric surface is the hypothetical surface to which water would rise in a well drilled into an aquifer. It coincides with the water table if the aquifer is unconfined, that is, not closed in by impermeable layers. However, if impermeable layers do exist the piezometric surface can be very different from the surface topography.

When two fluids are present in the same aquifer, for example, oil and water or fresh water and brine, their potential energies are different even at the same elevation and pressure because their densities in Equation (10.2) are different. Thus, oil will displace water downward because it is less dense and brine will displace fresh water upward because it is more dense. The density of a fluid may also vary for other reasons, such as the content of a solute or temperature. Potentials thus exist for driving convective motions of the fluid due to either compositional or thermal differences. There are numerous specialized treatments for these cases and we do not consider them further in this book.

10.2.2 Percolation flow

It is important to distinguish the porosity of a rock from its permeability. Porosity measures the fraction of void space in a rock. It is usually designated by the symbol ϕ. Porosity ranges from 0 (a perfectly dense rock) to 1 (a hypothetical "rock" that is all void). If the density of a non-porous rock is ρ_0, then the density of a rock with porosity ϕ is $(1-\phi)\,\rho_0$. Permeability, on the other hand, measures how readily a fluid can move through a rock. A rock must have some porosity to be permeable, but it is possible for a rock to be porous without having any permeability at all – it depends on how well connected the pores are.

Henry Darcy first determined the relation that is now named after him. He measured the rate at which water flows through a cylinder full of sand as a function of the difference in head between the water at its entrance and exit and wrote an equation very similar to Fourier's law for heat conduction:

$$\mathbf{Q} = -\frac{k\,\rho}{\eta}\nabla\Phi = -\frac{k\,\rho\,g}{\eta}\nabla h = -K\,\nabla h \tag{10.3}$$

where $\mathbf{Q}$ is the vector volume flux of water (which has the same units as velocity), η is its viscosity, and k is the permeability, which has dimensions of $(\text{length})^2$. Permeability k is *defined* by the rate at which a fluid (water) moves through a rock. The permeability is, thus, a kind of fluid conductivity. $K = k\rho g/\eta$ is known as the hydraulic conductivity. The equations describing the percolation of a fluid underground can be put into the same form as the heat conduction equation, which immediately makes a host of solutions to the heat conduction equation applicable to the problem of fluid flow (Bear, 1988; Hubbert, 1940).

The magnitude Q of the volume flux of water in Equation (10.3) is not equal to the velocity of the water in the rock, despite having the same units. Sometimes called the "Darcy velocity," it is the average rate at which a volume of water flows through the rock. However, because water fills only a fraction of the total rock's volume, the water must move faster than Q to deliver the observed flux. If all of the porosity ϕ contributed to the permeability, the actual velocity of water in the rock would equal Q/ϕ. In practice, not all of the porosity contributes to the permeability and the local velocity may be much higher. The local velocity may become sufficiently high in some circumstances that the flow becomes turbulent, in which case the permeability depends on the flow rate in a complicated manner.

Equation (10.3) can be elaborated to describe the unsteady flow of fluids as well as co-transport of heat and dissolved substances. Combined with other relations describing the storage of fluid in the rock and the rate of chemical reactions, many geologic problems can be related to the transport of water, or other fluids, through fractured rock (Lichtner *et al.*, 1996). Problems of this type are important for understanding the origin of ore bodies, underground motion of contaminants, and the transformation of sediments into rock (diagenesis and metamorphism).

Computation of the permeability of a given rock is a very complex affair and many different equations for estimating the permeability of a rock exist. Indeed, much of the uncertainty in the field of hydrology centers about permeability and its distribution. Permeability depends strongly on the size and distribution of the connected passages through rock. A simple model in which the rock is filled with a cubic array of tubes of diameter δ spaced at distances b apart can be solved to give a permeability (Turcotte and Schubert, 2002):

$$k = \frac{\pi}{128}\frac{\delta^4}{b^2}. \tag{10.4}$$

Note the strong dependence of permeability on the size of the narrowest passages through the rock. This is a typical behavior: Permeability is a very strong function of the grain size of the rock. Thus, it is large for coarse sands, but becomes very small for fine-grained silts and clays. Table 10.1 lists some "typical" values for the permeability of various rock types and indicates the enormous range of permeability found in nature. Note that permeability is commonly listed in the more convenient units of "darcys": 1 darcy = 9.8697×10^{-13} m^2.

It is very important to note that, although most of the values for permeability in Table 10.1 apply to "intact" rocks, the actual permeability of a rock mass may be very different (higher) because of fractures or even macroscopic cavities. Permeability is usually measured for small specimens that are often selected for their integrity. A rock mass in nature, however, is inevitably cut by large numbers of joints and faults that can have a major effect on the permeability of the mass as a whole: open joints allow fluids to flow along them easily, while faults may be lined with fine-grained gouges that inhibit fluid motion. Extrapolations of permeability from a small sample to an entire rock mass may, thus, give highly misleading results.

Where possible, permeability is measured directly in the field. The first such method was developed by Charles Theis (1935) and involves measuring the level of water in observation wells adjacent to a test well from which water is actively pumped. Other methods involving suddenly displacing the water in a single well by a cylindrical "slug" and

Table 10.1 *Permeability of rocks*

Rock type	Permeability k (m^2)
Gravel	10^{-9}–10^{-7}
Loose sand	10^{-11}–10^{-9}
Permeable basalt[a]	10^{-13}–10^{-8}
Fractured crystalline rock[a]	10^{-14}–10^{-11}
Sandstone	10^{-16}–10^{-12}
Limestone	10^{-18}–10^{-16}
Intact granite	10^{-20}–10^{-18}

After Turcotte and Schubert (2002), Table 9–1.
[a] Lichtner *et al.* (1996)

measuring the recovery of the water level, or by injecting water between pressurized packers in a borehole.

The porosity and permeability of rocks generally decrease with increasing depth. Near the surface, wide variations of permeability are common, which lends hydrology much of its variety and complexity. However, with increasing depth, overburden pressure rises and pores are crushed closed. Temperature also increases with depth and this promotes pore closure by viscous creep, so that porosity and permeability in terrestrial rocks are essentially absent at depths greater than a few kilometers in most places.

On the Moon and planets such as Mercury and Mars, large ancient impacts initially fractured the surface rocks to great depth, but overburden pressure has closed most pores at a depth that depends on the resistance of the rock to crushing. Seismic measurements on the Moon suggest that most lunar porosity disappears at a depth of about 20 km, or at a pressure of about 0.1 GPa. Under Martian surface gravity this implies that most porosity is crushed out at a depth of about 6.5 km. The fractured Martian regolith is estimated to be capable of holding the surface equivalent of between 0.5 and 1.5 km of water.

10.2.3 Springs and sapping

When water percolating through rock reaches the surface it leaves the fractures and pores of the rock and flows out onto the surface, either joining a saturated body of water already there (a streambed, lake, or the ocean), or flowing out onto the land surface as a spring. Because of the low permeability typical of rocks, the flow of water underground is usually slow and springs may continue to flow long after rains have ceased (see Box 10.1 to estimate how long a streamflow may continue without recharge). Water underground constitutes a reservoir that buffers changes in surface runoff. The total volume of groundwater on the Earth is only about 1.7% of the total surface water (this includes the oceans), but 30% of the Earth's fresh water resides underground. Other planets may have substantial inventories of groundwater: The principal uncertainty in estimates of how much water is

Box 10.1 **How long can streams flow after the rain stops?**

The question of how streams can continue to flow even when they are not being fed directly by rainfall is as old as the science of hydrology. Pierre Perrault (1608–1680) was the first to show that the rain- and snowfall in the basin of the Seine River is about six times larger than the river's discharge and could, thus, more than account for the flow from all of the springs and streams in the region (he neglected evaporation and transpiration by vegetation, which balances the net water supply).

Some streams, particularly those in arid climates, do not flow unless they have been recently filled by rain. The water in such streams rapidly infiltrates and percolates downward to a deep water table. In more humid climates, however, streams may continue to flow for weeks or months between rains. Such streams are fed by groundwater seeping into their beds. Because groundwater moves slowly, it may take a considerable period of time for groundwater to move from the area where it infiltrates to its emergence in the bed of a stream, during which time the stream continues to flow.

Using Darcy's law for a uniform permeability, we can estimate the relaxation time over which an elevated water table continues to feed a stream or spring from which the water drains. Let h be the height of the water table above its outflow point, located a horizontal distance L from the groundwater divide (Figure B10.1.1). The total volume of water contained above the outflow point is of order $hL\phi$ per unit length along the stream, where ϕ is the permeability of the aquifer. If the width of the zone along which water seeps out is w, and the groundwater discharge is Q per unit area, then water is discharged from the groundwater reservoir at a rate Qw per unit length of the stream (perpendicular to the plane of Figure 10.B1.1). The rate at which the height of the water table declines is thus $\dot{h}L\phi = \mathrm{Q}w$, from conservation of volume. The discharge Q is, thus, related to the height of the water table (which is equal to the head in this case) by $\mathrm{Q} = \dot{h}L\phi/w$. This expression can now be inserted into Darcy's Equation (10.3) to give:

$$\mathrm{Q} = \frac{\dot{h}L\phi}{w} = -\frac{k\rho g}{\eta}\frac{h}{L} \qquad \text{(B10.1.1)}$$

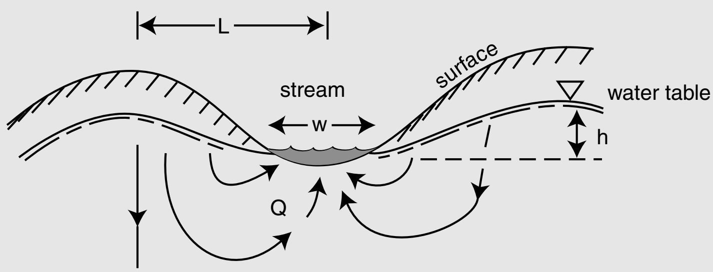

Figure B10.1.1 The capacity of subsurface water to support a surface streamflow depends on the quantity of water available, measured by the height of the water table h above the elevation of the stream and the horizontal extent of the water supply, L. This water flows to the stream, discharging a volume flux Q over a stream of width w. The rate of flow is regulated by the average permeability of the rock beneath the surface.

Box 10.1 **(cont.)**

where the gradient ∇h has been approximated by *h/L*. This equation can be rearranged into a first-order differential equation for the height of the water table *h*:

$$\dot{h} = -\left(\frac{k\rho g w}{\eta\phi L^2}\right)h. \tag{B10.1.2}$$

The solution to an equation of this type is well known:

$$h = h_0 e^{-t/\tau_R} \tag{B10.1.3}$$

where the *relaxation time*, τ_R, is given by:

$$\tau_R = \frac{\eta\phi L^2}{k\rho g w}. \tag{B10.1.4}$$

Inserting some typical values, $L = 1$ km, $w = 10$ m, $\phi = 0.1$, $k = 10^{-10}$ m^2 (sand), and the density and viscosity of water at 20°C, we find a relaxation time of about 4 months. It, thus, appears that even for relatively permeable aquifers like sand, the time for discharge of groundwater may be a large fraction of a year and streams may continue to flow even after months of drought. Conversely, recharge of a depleted aquifer may take a similarly long period. The timescales for groundwater flow may be very long because of the small permeability of many rocks that form aquifers.

present on Mars is its subsurface inventory of ice and liquid water. Ice appears to lie within tens of centimeters of the surface at high latitudes, suggesting that much of Mars' crustal porosity may be saturated with water. Similarly, the volume of methane observed in Titan's lakes might be only a fraction of the total residing below the surface. The Huygens lander detected methane that evaporated from the regolith beneath the warm lander, suggesting a methane table just below the surface of the landing site.

Water flowing from springs may undermine the mechanical stability of the surface. Just beneath the surface, the exit flow creates a pressure gradient that is connected to the fluid discharge by Darcy's law (10.3). This extra pressure adds to the pore pressure in the rock and may greatly weaken it, for reasons described in Section 8.2. In addition, the flow of water through the surface may carry fine-grained silt and clay out of the rock matrix, further weakening it. This selective removal of fine-grained constituents or dissolution of grain-binding minerals may disintegrate the rock or soil and excavate tunnels through which the water flows out readily. Known as piping, such tunnels develop in regions of heavy rainfall and the cavities it creates can undermine the soil surface (Douglas, 1977). Piping also plays a major role in undermining dams through which water is seeping. Water seepage may, thus, cause wholesale collapse of the rock face from which it flows. This erosion process is known as sapping.

Sapping may produce large-scale landforms. Such landforms are rare on Earth, where overland flow and stream transport usually cause erosion. However, in restricted locales where a permeable layer is underlain by an impermeable zone, surface runoff is limited and most of the outflow occurs at the interface between the permeable and impermeable layers. This situation is common, for example, on the USA's Colorado Plateau where permeable eolian sandstones overlie clay-rich fluvial formations (Howard *et al.*, 1988). Former interdune areas in the sandstone units also create extensive impermeable layers. Springs form everywhere at the contact between permeable and impermeable layers.

Long-continued spring flow at the base of cliffs undermines the overlying rock by dissolving the minerals that cement the sand grains into sandstone. The constant seepage at the base of cliffs also encourages chemical weathering that further weakens the rock. Alcoves form where the cliffs collapse and spring-fed streams carry the fallen debris away. As time passes, the alcoves grow deeper because they serve as drains for the subsurface water flow and, thus, intensify the flow at their heads as they lengthen. Given sufficient time, a canyon system develops that is characterized by stubby, sparsely branched tributaries with steep amphitheater-like headwalls. Canyon de Chelly is a terrestrial example of this kind of development. Many canyons on Mars also appear to have formed by groundwater sapping, of which the Nirgal Vallis is a prime example.

Groundwater sapping is of particular interest for Mars because it does not require overland flow initiated by rainfall. However, excavation of a canyon does require some process that removes collapsed material from the headwall. Streamflow cannot presently occur on the surface of Mars and so even these valleys may be relicts from a time when Mars possessed a higher atmospheric pressure.

10.3 Water on the surface

The terrestrial hydrologic cycle is dominated by the flow of water over the Earth's surface. Water falls as rain or snow, of which a portion flows over the surface from high elevations to low. In the process, liquid water entrains solid material, transports it, and eventually deposits it at lower elevations. Because of the prevalence of this process, undrained depressions on the Earth's surface are rare and, where they occur, draw immediate attention as indicators of an unusual geologic situation.

The processes and products of fluvial erosion, transport, and deposition are so familiar to Earth-bound geologists that it comes as a shock to realize that no other body in our Solar System, with the possible exception of Titan, is so completely sculptured by fluvial erosion as the Earth. The "rock cycle" as envisioned by James Hutton is essentially a fluvial cycle, in which rock debris is removed from the land by rainfall, washed to the oceans by streams, and converted back into rock after deep burial by other sediments. No other planet in the Solar System experiences a cycle of this type. Because of the importance of fluvial processes to terrestrial geology, each part of this cycle has been examined in great detail by many researchers. Most texts that deal with surface processes on the Earth devote one or more chapters to each portion of the fluvial cycle. This book attempts to put surface processes into a broader Solar System context, so our coverage of fluvial processes is necessarily more cursory, being compressed into a single chapter. References to more detailed treatments are given at the end of this chapter.

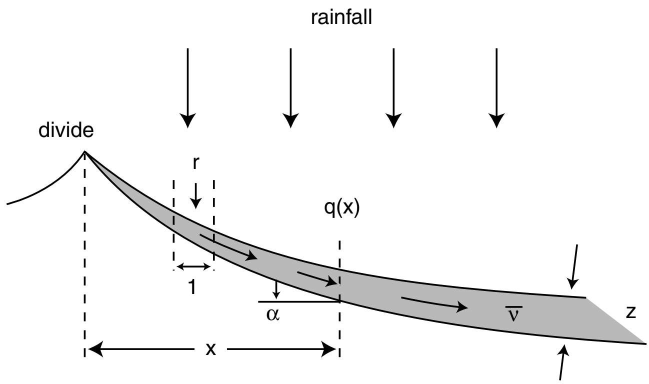

Figure 10.4 Runoff from rainfall on a sloping surface accumulates as an increasing function of distance x from the divide. As the volume $q(x)$ of runoff increases the flow both moves faster and becomes deeper. The flow velocity is regulated by an equation that incorporates both these factors.

10.3.1 Overland flow

When rain falls on a slope, that portion that does not infiltrate or evaporate flows off the surface as runoff. An important characteristic of runoff is that it increases steadily with distance downhill from the crest of the slope. The slope crest is an important location for fluvial processes and in this context it is called a divide because runoff moves in different directions on either side of it. If the rate at which runoff is generated per unit area (projected onto a horizontal surface) is r, measured in units of m^3/s per m^2, or m/s, then the volume flux of water flowing off the slope per meter of contour distance q (m^2/s) is given by:

$$q(x) = r\,x = zV \tag{10.5}$$

where x is distance from the divide measured along the slope gradient, z is the average depth of the flow (measured perpendicular to the slope), and V is the mean velocity of the flow (Figure 10.4). The runoff r is roughly equal to the rate of rainfall minus the infiltration rate. The volume flux of water, or slope discharge, increases with distance from the divide because the amount of water passing each contour of the slope includes all of the runoff generated between that contour and the divide. Because of the increasing volume of water, the potential for erosion also increases downslope. This is one of the distinguishing characteristics of fluvial erosion: The farther one travels from the divide, the greater the ability of water to erode. The consequences of this relationship for slope profiles will be explored in more detail in Section 10.4, but it should already be clear that it leads to the concave-up profile typical of landscapes dominated by fluvial erosion (see Figure 8.6).

Fluvial erosion. The process of sediment entrainment by overland flow was first analyzed by Robert Horton (1875–1945) in a landmark paper published just months before his death and is now known as “Horton overland flow” (Horton, 1945). Horton assumed that runoff begins as a thin sheet of nearly uniform thickness before concentrating into rilles farther downslope. Although thin sheets of water may flow in this way, during heavy rainfall on smooth slopes the flow organizes into waves, now known as roll waves, that develop when the Froude number, $\overline{v}^2/gd$, exceeds approximately 4. Under the wave crests, erosion proceeds faster than expected for a uniform flow. Horton described such waves, but he neglected the process of rain splash, which can mobilize soil particles in the zone that Horton considered to be erosion-free as well as generating turbulence in the flowing sheet of water. Nevertheless, Horton's model agrees qualitatively with observations and is still widely used to explain the onset of water erosion.

Horton supposed that runoff close to a divide would form a sheet too thin to entrain surface material and so a "belt of no erosion" would develop. Such belts are observed in badlands that lack vegetation, although rain splash may actually entrain some material in the flow as well as redistributing it through creep. Carson and Kirkby (1972) point out that natural surfaces are seldom smooth enough to permit an actual sheet of uniform depth to flow over the surface and that the flow very rapidly becomes concentrated into small rilles and channels. When the flow is thin, however, rain splash and surface creep constantly rearrange these channels so that persistent rilles do not form. Farther downslope, with increasing discharge, rilles do form that evolve into drainage networks.

Nevertheless, as Horton described, there is a finite distance between a divide and the location of the first permanent rille. This distance depends upon the infiltration capacity of the surface: It is larger for materials with high infiltration capacity, such as sand and gravel, and small for materials with low infiltration capacity, such as clay or silt.

The capacity of the flowing water to entrain surface material and erode the slope is proportional to the shear stress exerted on the surface, as discussed in Section 9.2.1. The shear stress τ exerted by a sheet of flowing water of depth z on a slope α is given by:

$$\tau = \rho g z \sin \alpha \tag{10.6}$$

where ρ is the density of the fluid (water). Flowing water begins to erode its substrate when the shear stress exceeds a threshold that depends on the grain size and cohesion of the surface material as well as on the properties of the fluid. Modern research on this topic has centered on the semi-empirical Shields criterion, originally proposed by A. Shields in 1936 (Burr *et al.*, 2006). This criterion can be expressed in a non-dimensional form that is applicable to any planet. The threshold shear stress is given in terms of a non-dimensional parameter θ_t:

$$\tau_t = (\sigma - \rho)\, g d \theta_t \tag{10.7}$$

where σ is the density of the entrained material and d is the particle diameter. Defining a friction velocity v_* in terms of the shear stress as in Equation (9.6), and a friction Reynolds number Re_* as in Section 9.2.1, the Shields threshold is given by (Paphitis, 2001):

$$\theta_t = \frac{0.188}{1+\mathrm{Re}_*} + 0.0475\left(1 - 0.699 e^{-0.015\,\mathrm{Re}_*}\right). \tag{10.8}$$

This expression is good up to $\mathrm{Re}_* \approx 10^5$. This curve lacks the steep upturn at small Re_* described by Bagnold (discussed in Section 9.2.1) and by Shields in his original work. This

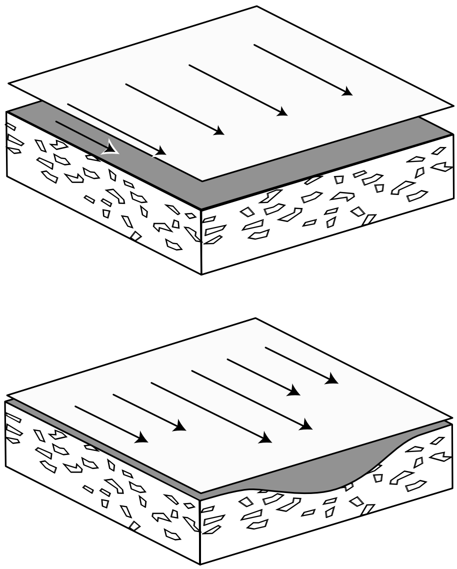

Figure 10.5 Overland flow of a thin sheet of water over an erodible bed is not stable. Panel (a) shows a uniform sheet flowing over the surface of the soil. Panel (b) indicates that any small, accidental, increase of depth concentrates the flow, increasing the shear stress under the deeper flow (and at the same time thinning the flow over the adjacent regions), which increases the erosion rate beneath the deeper portion, producing a positive feedback that quickly concentrates the flow into channels.

is reportedly because this equation takes into account rare but intense turbulent fluctuations in the boundary layer. It seems unclear at the moment whether this equation applies to eolian transport as well as water transport: The data on which it is based becomes sparse and somewhat contradictory below Re_* less than about 0.1, although it is stated to be valid down to $Re_* = 0.01$. The experimental material may also have contained a mixture of grain sizes. It seems that this equation should be applied with caution to very small grains until this situation is clarified.

When the threshold stress is exceeded, water begins to entrain material and carry it away, eroding the underlying slope. Erosion is enhanced by any factor that increases the shear stress: Deeper flows and increased slopes both contribute to the erosion rate, in addition to the erodibility of the material.

Rille networks. As runoff moves downslope it collects into channels, and the channels merge into larger channels. This process has been observed both on natural slopes exposed to gullying by vegetation removal and in experimental rainfall plots. Horton described the evolution of rilles into a drainage network in detail. He began by asking the question "Why, then, do rille channels develop?" He supposed that accidental variations in slope topography first concentrate sheet flow into slightly deeper than average proto-channels. Because these concentrations are deeper than average, they also exert greater than average shear stress on their beds and, thus, grow still deeper. That is, a flowing sheet of water of

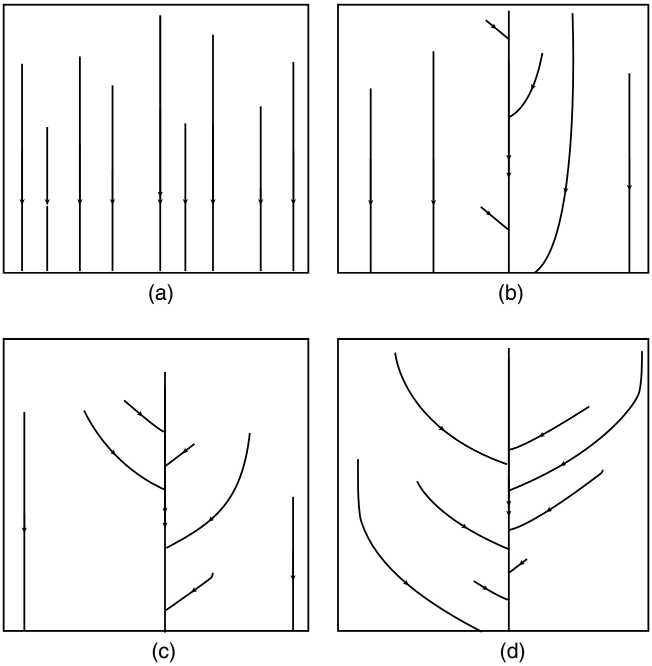

Figure 10.6 Straight, parallel rilles flowing off of an initially uniform slope are themselves unstable. Side branches develop that collect the water from adjacent rilles, deepening the main channel while starving the adjacent rilles. This process, called "cross-grading," operates by the principle that "the rich get richer while the poor get poorer." A dendritic stream network is the result of this kind of natural selection, which can be seen as a variety of stream piracy. Redrawn after Carson and Kirkby (1972, Figure 8.1).

uniform thickness on an erodible bed is unstable: Any slightly deeper pocket erodes faster than the adjacent regions and becomes still deeper, soon concentrating all of the flow in one channel (Figure 10.5). Horton observed, however, that the actual evolution of a system of rilles is more complex, because deepening channels initiate lateral flow that diverts water from adjacent rilles, "pirating" their headwaters in a process Horton termed "cross-grading." This lateral evolution continues until a network of initially parallel rilles becomes a branched network, illustrated in Figure 10.6.

The process of rille development and cross-grading proceeds at progressively larger scales as the landscape evolves. Undrained hollows are filled in, steep scarps are eroded down, and the stream network adjusts itself until, as described in 1802 by John Playfair (1964),

Every river appears to consist of a main trunk, fed from a variety of branches, each running in a valley proportioned to its size, and all of them together forming a system of [valleys], communicating with one another, and having such a nice adjustment of their declivities that none of them join the principal valley either on too high or too low a level.

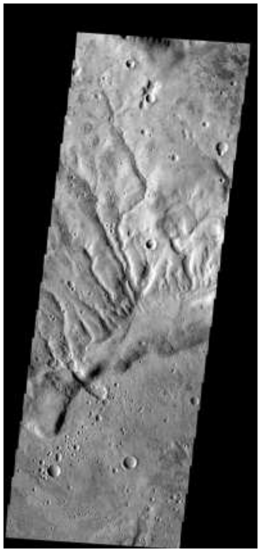

Figure 10.7 Dendritic stream network on Mars. Warrego Valles at 19.1° S and 244.0° E. This is part of the densest drainage network on Mars and strongly suggests that it was created by surface runoff from precipitation. Image is 24.8 km wide. THEMIS image, PIA05662. NASA/JPL/ASU.

Because of the tendency of water to collect in channels, the landscape becomes dissected into a fractal pattern of valleys and hillslopes on many different scales (Figure 10.7). In strong contrast to soil creep, which tends to simplify contours and smooth slopes, fluvial erosion sharpens irregularities.

Drainage basins. As stream networks develop, the landscape becomes divided into a hierarchical set of units known as drainage basins. In a fluvial landscape, drainage basins are the most natural geographic unit from the viewpoint of water supply or transport of sediment and solutes. The drainage basin associated with some point on a stream or river is that area of the land that contributes runoff to the stream. Divides – hillcrests down which water flows in different directions – separate drainage basins. The size of a drainage basin varies with the size of the stream: Small rilles may have drainage basins of only a few hundred square meters, while major rivers have drainage basins that encompass most of a continent.

One of the most regular quantitative relationships in fluvial geology relates drainage basin area A_b to the length L_b of the basin. Over a range of 11 orders of magnitude, from tiny rilles to continental drainages, the relation states (Montgomery and Dietrich, 1992):

$$L_b \simeq \sqrt{3A_b}. \tag{10.9}$$

Drainage networks are self-similar over this enormous range. That is, a map of a drainage network does not allow one to infer the actual scale of the image. However, this regular relationship breaks down at the smallest scale where the "belt of no erosion" asserts itself at a drainage length of a few tens of meters. Channels disappear below this scale. Another way of looking at this is to note that there is a threshold for channelized stream erosion that must be exceeded before network processes become important. Once this threshold is crossed, however, the processes that shape drainage networks become self-similar and any hint of a scale disappears from the system.

10.3.2 Streamflow

After his success in understanding eolian processes, Ralph Bagnold, our hero of Chapter 9, spent the rest of his career working on sediment transport in streams, applying ideas of fundamental physics to streamflow. He wrote a series of classic papers that still form the basis of our understanding of how fluids affect their beds (Bagnold, 1966). Before Bagnold, G. K. Gilbert approached the problem of sediment transport in streams from an experimental perspective, having constructed an enormous flume on the campus of the University of California at Berkeley to understand how streams transport sediment (Gilbert, 1914). Gilbert's interest had been piqued by the very practical problem of how to deal with the fluvial debris produced by hydraulic gold mining in California's Sierra Nevada. By 1905, that debris was advancing down California's rivers and had begun blocking the mouth of San Francisco Bay, creating a classic confrontation between the interests of gold miners, farmers along the river margins, and ocean shipping (Gilbert, 1917). Although Gilbert's meticulous flume experiments produced data that are still cited today, he failed to come up with any simple laws describing the interaction of sediment transport with the flowing water. His results were summarized by a large number of empirical correlations that, to a large extent, still characterize this field.

Sediment transport. Once established, steams continue to erode their beds and transport sediment delivered to them from upstream tributaries. The load of material transported by a stream is divided into several components. The dissolved load is composed of chemically dissolved species or colloids that are uniformly mixed with the water. The bedload consists of coarse material that slides, rolls, or saltates along the bed. Finer sediment moves as a "suspended load" that is concentrated near the bed but may be found higher in the water column, while the "washload," composed of still finer sediment, is uniformly mixed with the water. The criterion that distinguishes these categories is based on the dimensionless ratio ζ between the terminal velocity v of the sediment (Section 9.1.1) and the friction velocity of the flow, v_*:

$$\zeta = \frac{\text{terminal velocity}}{\text{friction velocity}} = \frac{v}{v_*}. \tag{10.10}$$

The transition between bedload and suspended load can be taken at $\zeta = 1.0$ to 1.8, while that between suspended load and washload occurs about $\zeta = 0.05$ to 0.13 (Burr *et al.*, 2006). Figure 10.8 illustrates these transitions and the threshold of motion using the modern version of the Shield's curve, Equation (10.8).

Stream velocity and discharge. A frequently asked question is, "What is the discharge (or velocity) of a stream or river given its depth, width and slope?" This question has engaged

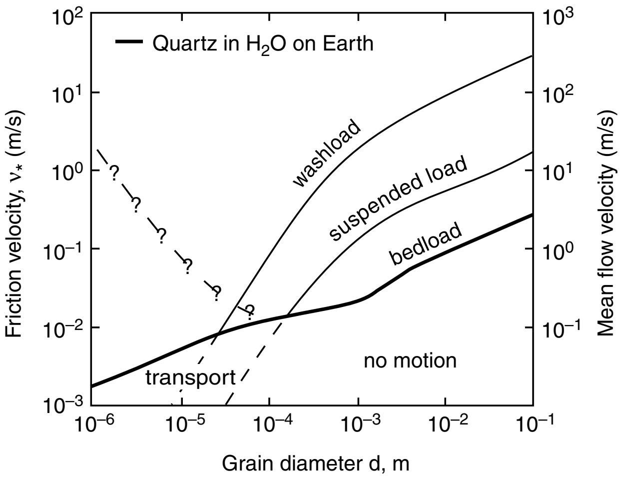

Figure 10.8 The relation between friction velocity and grain diameter for quartz grains in water on Earth. The heavy curve for grains moving along the bed has no minimum, although the queried extension at small grain sizes indicates much uncertainty in this conclusion, because the data from which the heavy curve was compiled may include sediment with a range of sizes and, thus, represents the impact threshold. Other curves show the thresholds for suspending grains in the lower water column or mixing it entirely through the water mass. The limits for either transport or no bed motion at low velocities are also shown. Greatly simplified after Figure 3 in Burr *et al.* (2006).

hydraulic engineers for centuries and several widely used equations can be found in the literature. French engineer Antoine de Chézy (1718–1798) gave the first useful answer to this question in 1775. His formula, as written today, is:

$$V = C\sqrt{RS} \qquad (10.11)$$

where V is the mean velocity of the flow (equal to the discharge Q divided by the cross-sectional area A of the stream), S is the slope of the channel, equal to tan α, and R is the hydraulic radius (equal to the area A divided by the perimeter of the wetted surface P). C is a dimensional constant that Chézy determined by comparing the velocity of one stream with that of another. Although this equation was adequate for Chézy's canal-design efforts, it contains many empirical constants and it is unclear how to scale this to a planet with a different gravitational field and to fluids other than water.

The next major improvement in an equation for streamflow came from the Irish engineer Robert Manning (1816–1897) in 1889. The "Manning" equation that we now write was neither recommended nor even devised by Manning himself, who actually did include the acceleration of gravity in his original formula. As usually written, the equation is:

$$V = \frac{k_M}{n} R^{2/3} S^{1/2}. \qquad (10.12)$$

Where k_M is a dimensional factor equal to 1.49 $ft^{1/2}$/sec or 1.0 $m^{1/2}$/sec.

Table 10.2 *Manning roughness for terrestrial rivers*

Bed material	Grain size (mm)	Manning roughness n ($m^{1/6}$)
Sand	0.2	0.012
Sand	0.4	0.020
Sand	0.6	0.023
Sand	0.8	0.025
Sand	1.0	0.026
Gravel	2–64	0.028–0.035
Cobbles	64–256	0.03–0.05
Boulders	>256	0.04–0.07

Data from Arcement and Schneider (1989).

The factor n, called the "Manning roughness," has dimensions of $(\text{length})^{1/6}$. This factor is widely tabulated for different channel conditions (smooth concrete, gravel, rock, etc.; usually in units of $ft^{1/6}$). It, like the Chézy coefficient, does not indicate how to scale to other fluids or planetary surface gravities. Table 10.2 gives some representative values of n for terrestrial rivers in metric units.

The most fundamental approach to this problem makes use of the balance between driving forces and resisting forces through the Darcy–Weisbach coefficient f. Unfortunately, this generality comes with the price of an equation that cannot be expressed as a simple analytic formula. Julius Weisbach (1806–1871) was a German engineer who focused on fundamental equations in hydraulics. In 1845 he published his major work on fluid resistance.

Consider a straight section of a stream channel that we will, for the moment, take to be rectangular in section with depth h and width w. The weight of the water per unit length of channel is $\rho g w h$. The component of the force acting downstream is $(\rho g w h)\sin\alpha$, where α is the channel slope. The shear stress on the bed and sides of the stream is just this force divided by the area of the streambed and wetted banks, equal to the perimeter $P = (w + 2h)$. The stress τ_b on the wetted bed of the stream is, thus:

$$\tau_b = \frac{\rho g w h \sin\alpha}{w+2h} = \rho g\left(\frac{wh}{w+2h}\right)\sin\alpha = \rho g\left(\frac{A}{P}\right)\sin\alpha = \rho g R \sin\alpha \qquad (10.13)$$

where R is, again, the hydraulic radius. Note that for a stream much wider than its depth the hydraulic radius is nearly equal to its depth. Weisbach related this shear stress, the driving force, to the flow resistance, which he expressed in terms of the mean velocity V of the stream:

$$\tau_b = \frac{f}{4}\frac{\rho V^2}{2}. \qquad (10.14)$$

Equating (10.13) and (10.14), then solving for V, we obtain the Darcy–Weisbach equation for the mean flow velocity:

$$V = \sqrt{\frac{8}{f} g R \sin\alpha}. \tag{10.15}$$

If we ignore the difference between sin α and tan α for small angles, comparison of (10.15) and (10.11) shows that the Chézy coefficient is:

$$C = \sqrt{\frac{8g}{f}}. \tag{10.16}$$

This indicates how the average stream velocity depends on the acceleration of gravity, but we still lack an expression for the Darcy–Weisbach friction coefficient f. Determination of this coefficient requires the solution of a transcendental equation, for which the reader is referred to a clear and detailed discussion in the book by Rouse (1978). The friction coefficient is a function of the Reynolds number of the streamflow, $\mathrm{Re} = \rho Vh/\eta$. For Re >> 1000 the usual practice is to relate f to the widely tabulated Manning roughness n and to use these empirical tables to calculate f:

$$\frac{1}{\sqrt{f}} = \frac{\mathrm{k_M}\ R^{1/6}}{\sqrt{8g}\ n}. \tag{10.17}$$

It may come as a surprise that, although this expression depends explicitly on the acceleration of gravity, it does not contain the viscosity of the liquid. The flow velocity does, in fact, depend somewhat on the fluid viscosity, but only through the Reynolds number Re. At very low Reynolds number this equation becomes equivalent to the expression for the velocity of a flowing viscous liquid, Equation (5.15), which depends on the inverse viscosity, $1/\eta$, but at high Reynolds number the viscosity of the fluid does not matter much because the resistance to fluid flow depends mainly upon the exchange of momentum by turbulence in inertial flow.

Floods. The discharge of a stream or river varies enormously over a seasonal cycle and from season to season. Furthermore, catastrophic floods have occasionally scoured the surface of both Earth and Mars. Most sediment transport takes place during floods. There is no simple rule that relates sediment concentration to discharge, but rivers do typically carry more sediment in flood than during average flows. However, as illustrated in Figure 10.1, the peak sediment transport takes place during greater than average floods because the transport capacity increases non-linearly with increasing flow velocity. The morphology of a river valley is, thus, controlled by large, rare events, a fact that makes many fluvial features difficult to understand unless this is taken into account.

The level of water carried in a river channel varies with the discharge. It is difficult to estimate the depth of a natural flow of water from a given discharge without detailed

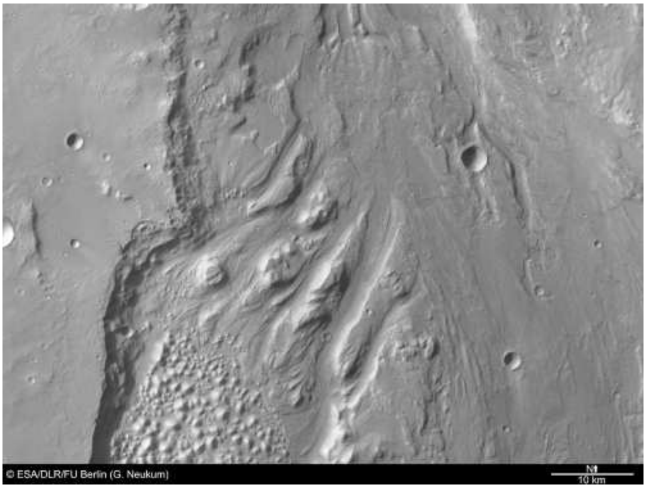

Figure 10.9 Ares Vallis is one of the large outflow channels on Mars. This image shows the transition between the Iani Chaos region to the lower left and the plains of Xanthe Terra to the top (north). The spurs between the individual channels have been shaped into crude streamlined forms by massive floods of water. 10 km scale bar is at lower right. Mars Express images by ESA/DLR/FU Berlin. (G. Neukum). See also color plate section.

knowledge of the topography. The Darcy–Weisbach equation, (10.15), can be written in terms of the total discharge $Q = VA$ and a geometric factor $A\sqrt{R}$:

$$Q = A\sqrt{R}\,\sqrt{\frac{8}{f} g \sin\alpha}. \tag{10.18}$$

As the discharge increases in a flood, the product of channel area and the square root of hydraulic radius increases, but to make definite statements about the depth of the flow we have to know how the depth and width vary with each other. In a wide, shallow channel, $A\sqrt{R} \approx wh^{3/2}$, but unless the width is known we cannot solve for the depth as a function of the discharge. In general, the width of the channel increases as its depth does, so all we can say is that increased discharge leads to increased depth.

Catastrophic floods. It is somewhat easier to estimate the discharge in the aftermath of a flood when we know both the depth achieved by the flow and its width. This method has been applied to the outflow channels on Mars to estimate the discharge during the height of the floods. Some of these channels are hundreds of kilometers wide (Figure 10.9) and, if they were once completely filled, must have carried enormous volumes of water. Martian channels present the problem that the depths of the flows are not known, but estimates based on the elevation of water-modified surfaces adjacent to the channels suggest that discharges ranged from 10^7 to 10^9 m^3/s, compared with about 10^7 m^3/s for the largest terrestrial floods (Carr, 1996). The total volume of water is estimated from the volume of sediment removed. Assuming a maximum sediment/water ratio (typically about 40%), one can estimate the water volume. Combined with the peak discharge rate, this gives the duration of the flood. For example, the Ares Vallis flood was estimated to have moved 2×10^5 km^3 of material. If the flow was 100 m deep, then the flood must have lasted 50 days. If it was 200 m deep, then it lasted 9 days.

Floodplains. Apart from catastrophic floods of the type that created the Martian outflow channels or the Channeled Scabland on Earth, much smaller floods occur regularly on terrestrial rivers and streams. During times of larger than average discharge the water spills over the banks and floods the adjacent terrain, putting the excess water into temporary storage. Unless steep canyon walls confine the channel, the water spreads out over a broad area, where its velocity decreases. Sediment previously in suspension settles out into a thin deposit of fine-grained material adjacent to the channel. The grain size of the deposited sediment falls off rapidly with distance from the channel, grading laterally from coarse near the former banks to fine silt farther away.

Repeated overbank flooding eventually builds up a gently sloping plain adjacent to the channel, known as the floodplain. Gentle ridges and swales, traces of former channels, usually curve across floodplains as their parent rivers shift. "Oxbow" lakes (abandoned former channels) and wetlands occur locally. The material that underlies floodplains is generically referred to as "alluvium." Alluvium is generally fine-grained material: gravel, sand, and silt, that is only temporarily at rest. Deposited by floods, its fate is to eventually become re-mobilized as the channel shifts and move on downstream, traveling inexorably toward the sea by slow leaps and bounds.

Levees. After repeated cycles of floods, the relatively coarse deposits close to the channel build up broad natural levees that stand above the level of the surrounding floodplain and tend to confine the water to the main channel. In subsequent floods the water rises higher in the channel and, when it eventually breaks through a levee, causes more violent floods. The location where the water breaks through a levee is known as a crevasse and the fan-shaped deposit of coarse material laid down after a breakthrough is known as a splay.

Floods are a normal part of the hydrologic cycle, although humans who have built structures on the floodplain often treat them as a major calamity. The floodplain is an active and necessary part of a river system, but its operation is unfortunately sporadic, making it difficult for many people to appreciate its essential role in the river system.

Alluvial fans. When a sediment-laden stream debouches onto a land surface, as opposed to a body of water, the flow spreads out, slows down, and the sediment is deposited in a conical heap known as an alluvial fan. In plan view, the contours of alluvial fans form circular arcs that center on the mouth of the stream. At any given time the stream flows down a radial channel over the surface of the fan, but as sediment is deposited and the channel builds up, the active stream eventually shifts to another direction, in time covering the entire conical pile. On alluvial fans in California's Death Valley one can easily recognize multiple deposits of different ages by the different degrees of desert varnish on the fan debris. The slope of an alluvial fan is typically steep at its head and gentler with increasing distance downslope, in the concave-upward pattern characteristic of fluvial landforms.

The area of an alluvial fan A_f is related to the area A_b of the drainage basin of the stream that feeds it by a simple equation (Bull, 1968):

$$A_f \approx cA_b^{0.9} \tag{10.19}$$

where the dimensional coefficient c varies with the climate and geology of the source area. This equation has been established both in the field and in small-scale laboratory experiments. Alluvial fans are often fed by debris flows and, high up on the fan surface, one can often recognize boulder-strewn debris flow levees. Lower down, the surface is covered with finer silt where muddy splays of water separated from the boulders as the fan slope decreased. When many alluvial fans form close together against a mountain front they may merge into a sloping surface known by its Spanish name, a bajada.

10.3.3 Channels

Channels develop where the flow of water over the surface persists over long periods of time. The morphology of the channel itself is often distinctive and, thus, indicates the action of a fluid flowing over the surface and excavating the channel, although the nature of the fluid itself is often unclear. When the Mariner 9 images of Mars first revealed giant outflow channels in 1971, many planetary scientists did not believe that they could have been cut by water, because Mars' atmospheric pressure is too low to sustain liquid water on the surface. Many different fluids were proposed, ranging from low-molecular-weight hydrocarbons to ice or mixtures of ice and water. Although it is still not possible to entirely rule out ice or brines, in the wake of the discovery of channel networks that seem to require overland flow, most planetary scientists now accept the likelihood of liquid water on the surface of Mars under climatic conditions that differ greatly from those prevailing today. Channels on Titan (Figure 10.10) were probably cut by liquid methane and channels on Venus were formed by lava flowing over its hot surface.

Channel features. Streamflow over a granular bed produces a variety of distinctive features from the interaction of the fluid and the deformable bed. Streamflow in a fixed channel is difficult enough to analyze at high Reynolds number because of the complex nature of fluid flow, but when coupled with the additional complexity of a deformable bed it poses problems that have yet to be completely solved. Nevertheless, extensive experimental study by many researchers using flumes and field observations of streamflow and its consequences has produced some understanding of how flow affects the bed of a stream and what features are produced by the interaction of the streamflow and its bed.

The major factor in the formation of bed features such as ripples, dunes, and larger accumulations of sediment is grain size. In addition, some measure of the velocity of the flow is needed. Much research has shown that to understand channel features the best measure of the flow is stream power (Allen, 1970), given by the shear stress on the bed τ_* multiplied by the mean velocity V, $\tau_* V$, which is measured in W/m^2. Low-power streams do not transport sediment at all, but as the stream power increases characteristic fluvial features such as

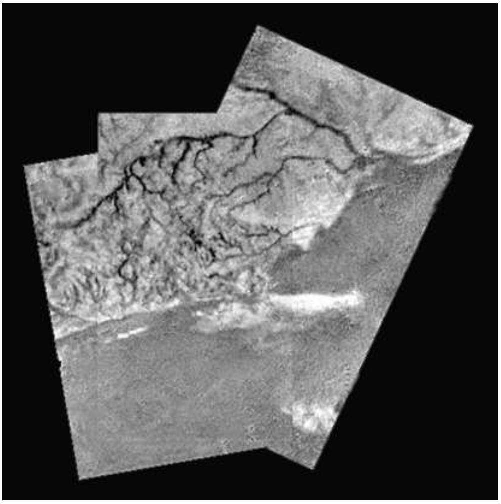

Figure 10.10 Mosaic of three images from the Huygens Descent Imager/Spectral Radiometer showing a dendritic drainage system on the surface of Titan. Each panel of the mosaic is about 7.5 km wide. Image PIA07236 NASA/JPL/ESA/University of Arizona.

ripples and dunes develop as a function of both grain size and stream power (Figure 10.11). A peculiar finding is that, at certain combinations of stream power and grain size, the bed is flat, even though intense sediment transport may be occurring. There is a small such plane bed field for intermediate stream power and large grains, and a much larger field at high stream powers for all grain sizes. These are designated, respectively, the lower and upper plane bed regimes.

Antidunes are unique to shallow, rapid flows. Unlike dunes, antidunes move upstream as flow proceeds. Although the form itself moves against the flow, sediment continues to move downstream: Only the wavelike shape of this feature moves counter to the current direction. Antidunes form when the Froude number of the flow, $V/\sqrt{gh}$, approaches 1. Their wavelength is approximately $2\pi h$, where h is the depth of the flow.

Ripples, dunes and antidunes can be distinguished in sedimentary deposits by means of cross-bedding. The ability to "read the rocks" and infer the nature of a flow, whether fast or slow, unidirectional or alternating, shallow or deep, is an important tool in the kit of a sedimentary geologist (Allen, 1982; Leeder, 1999). Observations of apparent cross-bedding in Martian sedimentary deposits by the Opportunity rover have led to the important inference of the former presence of shallow lakes on the surface of Mars (Grotzinger *et al.*, 2006b).

Bedforms develop even in streams that flow over solid rock. Channels often excavate into the underlying bedrock by quarrying away small joint blocks through differential pressures and cavitation behind obstacles. Gravel and cobbles carried along the bed may, over time, gradually erode the bed by abrasion. The result is flutes and, where circular motion is maintained over long periods of time, potholes. Potholes reach impressive depths of tens of meters as they are ground into the bed by swirling cobbles and debris below the streambed.

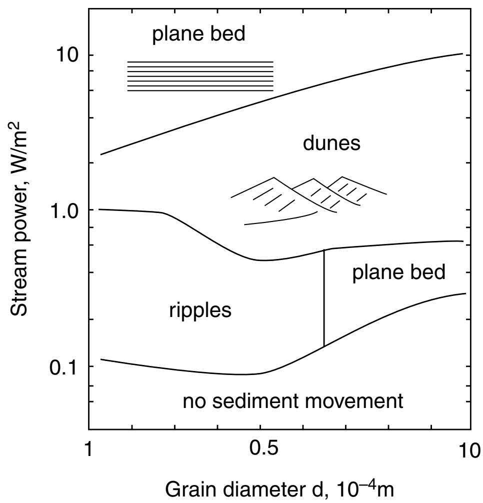

Figure 10.11 The type of bedform that develops beneath flowing water depends upon both stream power and the grain size of the sediment. This figure applies to water flowing over quartz sand on Earth, but it shows the variety of forms that can develop. No movement occurs below the lowest line. Note that the bedform pattern is not unique: Plane beds occur for both high stream power (upper regime flow) and low stream power over coarse sediments (lower regime flow). Simplified after Allen (1970, Figure 2.6).

In addition to the abrasion of the bed, the material carried by a river or stream is itself abraded as it travels downstream. Angular grains of sand become rounder, gravel becomes finer and cobbles are smoothed, rounded and reduced in size as they move downstream. Such downstream variations in the size of transported sediment make it difficult to associate a particular grain size with fluvial processes, because grains that might begin at the threshold of transportation move into suspension as they are broken and abraded while large, initially immovable rocks are broken down and begin to slide and roll along the bed.

Streamlined forms. Large floods create characteristic bedforms. One that is considered particularly diagnostic of floods are teardrop-shaped, streamlined islands that develop behind obstacles that divert the flow. These features are bluntly rounded in the upstream direction and taper to a point downstream. Once diverted by the obstacle, the flow closes back in around it downstream, but because this takes some time, the obstacle shields a tapered triangle from the flow, creating this shape. The faster the flow, the more gradual is this closure and the longer the island becomes. Streamlined islands in the Channeled Scablandsof Washington State are typically about three times longer than their maximum width, whereas streamlined forms on Mars are a little longer, about four times their width (Baker, 1982).

Hydraulic geometry. A widely used metric relates quantitative descriptors of the channel to its discharge. Discharge is chosen as the independent variable because it is believed that, while a stream may adjust its width, depth, or velocity by moving sediment from one

Table 10.3 *Hydraulic geometry of selected rivers (Leopold et al., 1964, p. 244).*

Location	Type	b	f	m
Midwest, USA	Fixed station	0.26	0.40	0.34
Semiarid, USA	Fixed station	0.29	0.36	0.34
Rhine River	Fixed station	0.13	0.41	0.43
Midwest, USA	Downstream	0.5	0.4	0.1
Semiarid, USA	Downstream	0.5	0.3	0.2

place to another, the discharge is determined by the climate and area of the drainage basin and so cannot be adjusted by the interaction between the stream and its bed. The method is entirely empirical: Data are collected from many rivers and streams, plotted against log-log axes, and lines are fit to the data that typically form fuzzy linear arrays. These fits must thus be regarded as approximate, but they do indicate general trends. Because straight lines on log-log plots indicate power laws, the relations for stream width w, depth h, and velocity V are written:

$$\left.\begin{aligned} w &= aQ^b \\ h &= cQ^f \\ V &= kQ^m \end{aligned}\right\} \tag{10.20}$$

where a, c, k are dimensional fitting parameters and b, f, m are dimensionless exponents. There are two constraints on these parameter sets because $Q = whV$: $a\,c\,k = 1$ and $b + f + m = 1$. Because river discharge even at one location is not constant over time, there are two ways in which these parameters are compiled: either at a fixed location as a function of time, or at different locations downstream on the same river. The exponents, b, f, m are considered the most significant parameters and they are tabulated in Table 10.3 for a small number of river systems.

Ratios illustrate the utility of this parameterization. For example, the aspect ratio of a river is the ratio between its depth and width. The aspect ratio h/w is proportional to Q^{f-b}, so at a fixed location the aspect ratio equals $Q^{0.14}$ for rivers in both the Midwest and semiarid USA. Thus, as discharge increases the relative depth increases slowly at a given station. On the other hand, going downriver the aspect ratio decreases slowly with discharge as $Q^{-0.1}$ for Midwestern rivers: Near its mouth the Mississippi is much shallower for its width than it is upstream. But, despite folklore, the Mississippi is not actually a "lazy" river – its velocity continues to increase with discharge either downriver or as a function of time at one location.

Meandering rivers. The tendency of rivers and streams to deviate from a straight course has long fascinated observers. Rivers flowing without constraint over an erodible bed quickly develop a meandering course. In an initially straight channel, the meanders begin as gentle bends as the water swings from side to side. The meanders grow in amplitude until they develop such extreme hairpin curves that they loop back on themselves until the water finally cuts through the narrow neck, temporarily shortening the course of the river. The abandoned meander loop forms an oxbow lake on the floodplain adjacent to the new main channel. The wavelength of a meander is a function of the channel size. A careful regression of meander wavelength λ_m and channel width w shows that they are nearly (but perhaps not exactly) proportional to one another (Leopold *et al.*, 1964):

$$\lambda_m = 10.8\, w^{1.01} \tag{10.21}$$

where the wavelength and width are both in meters. Although meander loops on a river floodplain are continually growing and being cut off, there is correlation between meander amplitude A_m and width as well:

$$A_m = 2.4\, w^{1.1}. \tag{10.22}$$

In addition to meandering laterally, rivers also meander vertically: Rhythmic variations of depth develop as deep pools alternate with shallow riffles with the same periodicity as the lateral meanders. The pools develop on the outside of meander bends, while the riffles form between them. These rhythmic depth variations develop even when the channel is confined between rocky walls that suppress lateral meanders, such as in the Colorado River confined within its canyon in Arizona.

One occasionally finds a river channel meandering through a valley that itself meanders on a much larger scale. In such cases one can infer that the smaller stream (called an “underfit” stream) carries a much smaller discharge than its former counterpart. This relationship is often observed in channels that once drained the meltwater from retreating Pleistocene ice sheets on Earth. A similar relationship, but in a sinuous lava channel, is observed in Schröter’s Rille on the Moon.

The outside of meander bends is usually a steep bank that is often undercut and is obviously undergoing erosion. A gently sloping bar on which sand and fine gravel is deposited, called a point bar, occupies the inside of the bend. As the channel shifts laterally, the floodplain is consumed at the outer part of the bend and rebuilt on the inner bend. Cross-sections of the migrating channel show coarse-grained material (often gravel) at the former channel floor, fining upward into sands where the point bar is deposited, then silts where former point bars are buried by floodplain silts. Such fining-upward sequences, when they can be recognized in ancient fluvial deposits, provide a direct indication of the depth of the former river channel (Figure 10.12). From the depth, the correlations of hydraulic geometry yield an estimate of the discharge of the ancient river that created the floodplain.

Meanders do not form simply because water flowing in the straight sections between meanders impinges on the outside of the bend. Many authors, including James Thomson (William Thomson’s brother) in 1876 and Joseph Boussinesq in 1883 independently discovered the helical flow of water in meander bends. However, the most famous re-discoverer of this effect was Albert Einstein, who perceived the effect while stirring a cup of tea (Einstein, 1954). Originally publishing in 1926, Einstein noted, as many others have

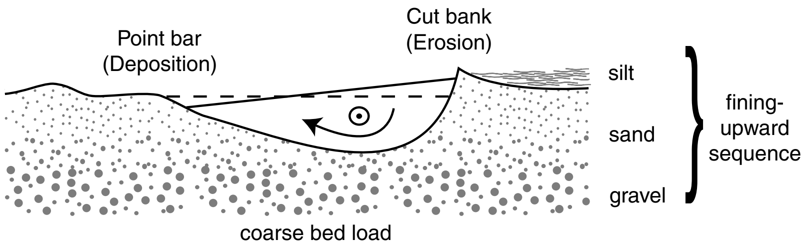

Figure 10.12 Water on the outside of meander bends rises higher than the water on the inside of the bend because of centripetal acceleration. This drives a secondary circulation that moves water from the outside of the bend toward the inner bend to create an overall helical circulation as the water moves downstream. Sediment is eroded from the outside of the bend and deposited on point bars on the inside. At the same time, coarse material remains near the bed while sand is deposited higher up on the point bar. Silt is deposited on top of the sand during overbank floods to produce a fining-upward sequence of sediment sizes.

done, that the tealeaves in the bottom of a stirred cup of tea gather together at the center of the cup. He inferred that the rotating liquid must create a helical flow in addition to its rotation. This flow descends along the outside of the cup and ascends in its center, sweeping the tealeaves into the center as it flows. Einstein realized that this flow could account for river meanders and point bars by transporting sediment down along the outside of the meander bends and depositing it on the inner point bar. The reality of such helical flows in river bends has now been demonstrated many times and this spiral flow is an accepted part of river hydraulics.

The helical flow is due to two factors. The first is centrifugal force. As the water flows around the bend, it rises higher along the outside of the bend. By itself, this would not cause a secondary flow if the water in the river were in rigid-body rotation. However, friction reduces the water velocity along the contact between the water and the outside wall of the bend and the excess pressure caused by the elevated water surface drives a flow downward along the outside wall.

Meanders are not restricted to just rivers flowing over granular material. They are commonly observed in glacial meltwater streams flowing over solid ice and in lava channels (they are called "sinuous rilles" in lava feeder channels on the Moon). In these cases one must presume that the same helical flow enhances channel migration on the outside of bends, perhaps by thermal erosion as the flow brings hotter material to bear against the outside of bends, but the precise mechanism is currently unclear.

Braided rivers. Steeply sloping streams heavily loaded with coarse sediment do not flow in well-defined channels. Instead, the flow divides into a complex network of shallow short-lived channels that diverge and rejoin many times as the water and sediment move downstream. Such channels are known as braided rivers.

Unlike rilles on a slope, the constantly shifting channels in the bed of a braided river do not unite to create channels progressively more capable of carrying the available load of sediment. Such streams are sometimes described as "overloaded," in the sense that the union of two sediment-laden channels is less capable of transporting the load than the individual channels, so some of the load is deposited when channels join, creating a temporary bar that eventually diverts the flow to a new location.

The factors that decide whether a stream channel is meandering or braided are still poorly understood (Schumm, 1985). Several rivers have been observed to alternate in style between meandering and braided or vice versa in historical time, a process termed "river metamorphosis" by fluvial geologist Stanley Schumm. For example, the channel of the South Platte River in Nebraska changed from braided in 1897 to meandering in 1959 in response to a large decrease in mean annual discharge due to irrigation projects that extracted water from the river (Schumm, 1977, p. 161). Likewise, the lower reaches of the Pleistocene Mississippi River were braided because of its greater slope down to the lowered sea level during the ice ages, as well as to the greater discharge it carried as the ice sheets melted. As the sea level rose and the ice sheets disappeared, its channel changed from braided to meandering.

An often-cited criterion that divides meandering from braided rivers on Earth is expressed in terms of channel slope S and bankfull discharge Q_{bf}, where the discharge is in m^3/s (Schumm, 1985). The river is usually braided when the average channel slope exceeds:

$$S \geq 0.0125\, Q_{bf}^{-0.44}. \tag{10.23}$$

Thus, any factor that increases either slope or discharge favors braiding over meandering. The frequent observation that braided rivers typically carry coarse debris is implicit in Equation (10.23) through the connection between average channel slope S and the grain size of bedload: Rivers that carry coarse debris are steeper than those that carry fine sediment.

The paleohydrologic hypothesis. Schumm also noted an apparent connection between channel stability and the presence of vegetation. Vegetation growing on islands in shifting channels tends to stabilize them. Root mats add cohesion to channel banks and hold fine sediment in place until undercut by powerful currents. Both of these factors tend to favor meandering channels rather than braided. Schumm noted the lack of evidence for meandering river channels before the Late Paleozoic era when vegetation first covered Earth's land surface. He elaborated a "paleohydrologic hypothesis" that suggests that all river channels were braided before the evolution of land plants. If this connection between vegetation and channel form is correct, we should not expect to find meandering river channels on Titan or Mars. On Titan, present resolution is too poor to be sure, but meandering channels do appear to cross the surface of Xanadu. On the other hand, the boulder-strewn surface at the Huygens landing site is consistent with a braided river channel (Figure 7.16d). On Mars, there are now HiRISE images of indisputable meandering channels in Aeolis Planum (Burr *et al.*, 2009), Figure 10.13. It must, thus, be possible for meanders to develop in the absence of vegetation (Howard, 2009), perhaps because of cohesion from clay or permafrost that binds the sediment together. Channel meanders, while evidently not requiring the presence of vegetation, may, nevertheless, indicate special mechanical conditions in the sediment adjacent to the channel.

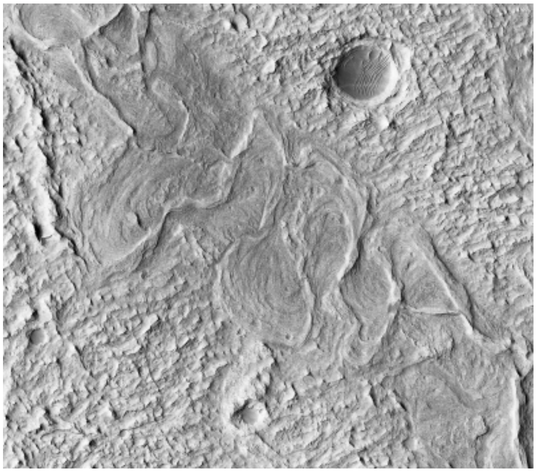

Figure 10.13 A meandering channel in Aeolis Planum, Mars, that belies the proposal that meanders develop only when banks are stabilized by vegetation. These highly sinuous meanders actually stand as ridges at the present time, an example of inverted topography. Gravel in the channel presumably made it more resistant to erosion than fine-grained surrounding material. Portion of HiRISE image PSP_006683_1740. Image is approximately 2.3 km wide. NASA/JPL/University of Arizona.

River terraces. Gently scalloped scarps are often found parallel to the active floodplains of large river systems. The downstream slope of the relatively flat surface behind such scarps is similar to that of the active floodplain. These surfaces, which often look like treads on a giant staircase stepping up away from the river, are known as river terraces. Because of their distinctive appearance and their importance for land use, terraces have received a great deal of attention in the terrestrial geologic literature (e.g. Ritter, 1986).

River terraces are abandoned floodplains of a river system that has eroded deeper into its valley. Geomorphologists distinguish paired terraces, which appear at the same elevation on opposite sides of the river, from unpaired terraces. Terraces are the result of the lateral migration of the river channel back and forth across the floodplain as the channel slowly erodes downward into the floor of the river valley. Terraces are important because they indicate changing conditions, although they are not usually diagnostic of exactly what conditions are changing. For example, the erosion they record could be due either to tectonic uplift of the rock underneath the stream, or increased downcutting of the stream. Downcutting can be due to increasing water supply, decreasing sediment load, or a lower base level at which the river discharges.

Although it is sometimes stated that the existence of discrete terraces indicates that downcutting must be episodic rather than steady, this is not necessarily true. Because a long interval of time may separate the impingement of the main channel on one valley wall during its slow lateral swings, the change in the level of the stream between two terrace-cutting events reflects the accumulated erosion between cutting events. Discrete terraces form even if the rate of downcutting is uniform because of this interval between terrace-cutting events. Although the ages of river terraces on the Earth can now be determined through the measurement of cosmogenic isotopes, it is still extremely difficult to discriminate episodic versus steady downcutting from such data.

Tributary networks. The most familiar pattern of drainage networks is one in which smaller channels join into larger ones that, in turn, join still larger ones, forming a network formally called a tree or dendritic pattern (Box 10.2). This tributary pattern persists on the Earth's land surface over most of its area because of the increasing capacity of downstream water to carry sediment. Rivers that branch downstream and then rejoin do occasionally occur on Earth, but they are relatively rare. Such non-tree-like patterns are more common in Martian channels, for reasons not currently understood.

The junction angle in tributary networks is such that the acute angle between links usually occurs upstream of the junction. This is presumably because the momentum of the joining currents tends to carry both in the same direction – it is unusual for a tributary to discharge its water upstream into the channel it joins.

Distributary networks. When a stream or river can no longer carry its sediment load, due either to loss of water by infiltration into a substrate (as on an alluvial fan), or because of a decreasing gradient (as when it encounters a lake or the ocean), its sediment is deposited and a system of dividing channels develops. The branching pattern of such a distributary network may resemble the tree-like form of a tributary network, but the slope in this case is reversed: The largest channels are upslope of the smallest channels. The acute angle of the junctions is downstream in this case, again tending to preserve the momentum of the dividing channels. Similar networks develop among the channels actively feeding lava flows spreading over flat terrain.

Unusual networks develop where the flow direction alternates, such as in tidal marshes where the surface is alternately flooded and drained. In this case the same channels serve alternately as a distributary and a tributary network. In such networks the channels tend to divide and rejoin frequently and junction angles are typically close to 90°, perhaps because of the frequent collisions between incoming and outgoing streams of water.

Venusian channels. No one seriously expected to find fluvial channels on Venus. The surface temperature is far too high to permit liquid water to flow over its surface. However, images returned by the Magellan radar (Figure 10.14) reveal channels that are remarkably similar to those of terrestrial river systems. Meandering channels with natural levees, streamlined islands within the channel, even crevasse splays and abandoned meanders can all be identified in the images. On Venus, however, we must certainly be looking at channels that once carried lava, not water. Because of Venus' high surface temperature and large eruption volumes, lava cools relatively slowly compared to terrestrial lava flows; furthermore, the flows may have continued over such long periods of time that "fluvial" features developed. Lava, like water, is capable of eroding its bed by plucking and of transporting "sediment" in the form of more refractory minerals, so this may be a case in which similarity of physical process promotes similarity of form, even though the materials involved are very different.

Box 10.2 **Analysis of stream networks**

Rilles and gullies join to form larger streams, which join again to form still larger streams, and so on up to large rivers. The result is a branching or *dendritic* (tree-like) network that extends from the smallest rilles to the largest trunks: rivers like the Mississippi or the Amazon. Water and sediment eroded from the land are flushed down these channels, eventually to be deposited in the oceans.

Most sediment is fed into the network from overland flow at the level of rilles. Larger and larger streams mainly serve to transport it. Most erosion, thus, occurs on the scale of small drainage basins, grading into transportation at larger scales, although deeply incised rivers such as the Colorado may be fed large amounts of material directly along the main channel by mass movement.

Robert Horton (1945) brought order to stream network analysis by proposing a simple numbering scheme, which has been slightly improved by other authors. Horton assigned the smallest recognizable rilles to order number 1. When two order 1 channels join they become a channel of order 2. The union of two order 2 channels is of order 3, and so on. When a channel of lower order joins a channel of higher order, the order of the higher channel does not change. Thus, when an order 3 channel is joined by an order 2 channel, it remains of order 3. See, for example, the fourth-order network shown in Figure B10.2.1.

To appreciate the success of Horton's idea of numbering from the smallest rilles to the largest river, consider the opposite scheme in which the main trunk of a large river is assigned order 1. Proceeding upstream to smaller and smaller tributaries, we would fined that the last recognizable rilles in most basins had a different order, even though their function in the

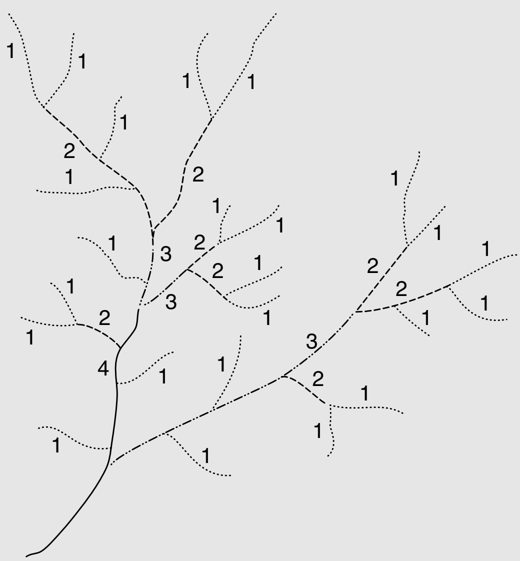

Figure B10.2.1 A typical stream network, which illustrates the ordering of stream segments. This network is of order 4. After Figure 10.1 of Morisawa (1968).

Box 10.2 **(cont.)**

hydrologic system is identical. Apparently similar rilles on opposite sides of a divide would, in general, be assigned different orders.

Some North American examples of high-order river systems are the Mississippi of order 10, the Columbia of order 8, the Gila of order 8, and the Allegheny of order 7. Unfortunately, tables of the orders of all the world's rivers seem to be hard to find, although ordering is so suitable for computer computation that the ArcGIS program includes a tool that assigns orders to streams.

The major problem with this scheme is that the definition of the first-order rille is uncertain: Depending on the map scale, this could be the smallest recognizable rille (as Horton supposed), or the smallest perennial stream (which means its assignment depends on climate). Inadequate resolution caused a problem with the initial ordering of Martian channel networks: At the lower resolution of Mariner 9 and Viking images, Martian valley networks appeared to have much lower drainage densities than terrestrial networks, for which a variety of causes were cited (Carr, 1996). However, once higher-resolution Mars Orbiter Camera (MOC) images became available, true first-order rilles could be recognized and it was realized that Martian and terrestrial networks have similar densities (Carr, 2006). Given this history, no one has seriously tried to assign orders to Titan networks (except at the Huygens landing site) because of the low resolution at which they are currently seen.

Stream ordering would be an amusing but mechanical pastime, except that ordering clarifies the statistical properties of networks for practical applications in flood wave prediction and sediment yield estimates (as only two examples). There are useful quantitative relations between landscape properties, such as average slope or stream discharge, as a function of order. It was also originally hoped that the statistical properties of stream orders might be diagnostic of network origin.

Some properties of a fourth-order drainage basin are listed in Table B10.2.1. Several characteristics are clear upon inspection: The number of streams decreases sharply with increasing order. In addition, the average channel slope decreases regularly with increasing order, the average channel length and basin area increases with order, and the drainage density is nearly independent of order.

Drainage density is defined as the sum of the lengths of all channels in a basin divided by its area. Its inverse is approximately the distance between streams in the basin. For first-order streams this is also the width of the belt of no erosion, as defined by Horton. The drainage density

Table B10.2.1 *The Mill Creek, Ohio, drainage network (Morisawa, 1959, Table 12).*

Stream order	Number of streams	Average length (m)	Average basin area (km^2)	Average channel slope	Stream density (km/km^2)
1	104	111	0.065	21.6°	3.39
2	22	303	0.313	7.01°	4.36
3	5	1046	1.505	2.23°	3.77
4	1	1915	6.941	0.57°	3.52

Box 10.2 **(cont.)**

is, thus, closely related to infiltration capacity: Large drainage densities imply a low infiltration capacity, as often seen in badlands, whereas low densities imply that water sinks in readily.

Plots of the logarithms of different characteristics of drainage networks versus stream order generally form straight lines. This suggests power-law relationships among the different quantities. For example, the number of streams of a given order p, expressed as $N(p)$, can be written:

$$N(p) = r_b^{(s-p)} \tag{B10.2.1}$$

where s is the order of the main stream in the network. The constant r_b is known as a bifurcation ratio. In the fourth-order network of Table B10.2.1, $s = 4$. There are five third-order streams, and $s - p = 4 - 3 = 1$, so $r_b = 5$. So far, this is nothing new. But now note that for $p =$ 2 this formula predicts that there should be 5^2 or 25 second-order streams (there are really 22). For $p = 1$ the formula predicts 5^3 or 125 first-order streams (there are really 104). The fit is not perfect, but it is fairly close.

Similar power-law formulas can be constructed for other quantities, such as stream length and slope. Over large, high-order drainage basins fits can be adjusted by least squares to obtain best estimates for each of these quantities, which then describe the branching properties of a drainage network.

This type of fitting was popular between 1945 and about 1970 when it was believed that such fits and bifurcation ratios reveal important information about how a drainage network develops and could, in some way (no one knew quite how) be related to the fluvial processes that created the network. Unfortunately, geomorphologist Ron Shreve dashed most of these hopes in 1966 when he showed that relations of this type develop in *any* dendritic network, including networks generated by random walks in a computer (Shreve, 1966b). Shreve's arguments have been confirmed and extended by later work (e.g. Kirchner, 1993). Nevertheless, many workers are convinced that stream networks statistics indicate *something* about the organization of fluvial processes and attempts have been made to assign a kind of entropy to stream networks and show that actual networks maximize that entropy (Rinaldo *et al.*, 1998) or expend the least work. Other, more complex, numbering schemes have been developed that claim to have genetic significance, but their success is presently unclear.

Statistical descriptions of drainage networks do have practical value in estimating the numbers and lengths of links at different levels without actually having to measure the entire network, but it is not easy to relate the parameters to genetic processes. Networks developed by sapping seem to be less branched and possess shorter links than those developed by overland flow, but this type of distinction can probably be made without the aid of detailed statistical analyses.

10.3.4 Standing water: oceans, lakes, playas

Running water tends inexorably downhill. When it reaches the lowest possible level it accumulates into a body of standing water that may range in size from tiny temporary ponds to global oceans. The most important geologic fact about running water is its ability to transport sediment from higher levels to lower. When it enters a large accumulation

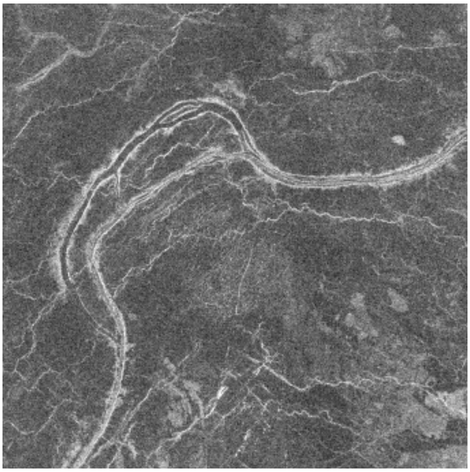

Figure 10.14 Sinuous lava channel on the plains of Venus. The overall channel flows from Fortuna Tessera in the north, south to Sedna Planitia. Channel is about 2 km wide and is interrupted by streamlined islands. The channel pattern illustrates the formation of an alternative channel during flow. Frame width is 50 km. Magellan F-MIDR 45N019;1, Framelet 18. NASA/JPL.

of water its velocity drops (although currents are never completely absent in any body of water: Underwater gravity currents are discussed in Section 8.2.3) and its sediment is dropped somewhere near the shore. In contrast to the land surface, bodies of standing water are the locales of sediment accumulation rather than erosion.

Standing water, however, possesses its own distinctive ability to move material. This ability depends on the action of waves, so that sediment transport occurs mainly at the level of the water surface. Wave action produces distinctive landscape features that remain even long after a body of water disappears.

Aside from the prevalence of coastlines on the Earth, ancient Mars may have possessed extensive bodies of water whose former shorelines, if found, would demonstrate their presence and dimensions. Titan is now known to possess ephemeral lakes of liquid methane and ethane, making beach and lake processes of prime interest to planetary geologists.

An appreciation of the landforms created by waves began with eighteenth-century British geologists who initially attributed most former geologic activity to the waves that they observed crashing around the edges of their sea-girdled isle. In the nineteenth century, American geologist G. K. Gilbert took a large step forward with his study of the now nearly extinct Lake Bonneville in Utah (Gilbert, 1890). The present Great Salt Lake in Utah is a small remnant of a much greater lake that existed during the Pleistocene. When it drained about 14 500 yr ago, it exposed the beaches, deltas, spits, and bars that formed during its brief existence of about 17 000 yr.

Gilbert was impressed that most of these features are the consequence of waves breaking against the shore. His research, as does much modern research, therefore focused on the generation, propagation, and interaction of waves with the shore.

Waves on water. The generation of waves has received a great deal of attention for its own sake and we can only touch on the basics in this book. The reader who wishes to go further should consult the treatise by Kinsman (1965). Waves on the ocean, lakes, or even ponds are created by wind blowing over the surface. William Thomson (who became Lord Kelvin) and German physicist Hermann von Helmholtz were the first to understand how wind can generate water waves by an aerodynamic instability, now called the Kelvin–Helmholtz instability. The interface between two fluids, such as air and water, cannot remain flat if the fluids move with different velocities. Waves develop on the interface, beginning with small, short-wavelength waves for which the restoring force is surface tension, then growing into much larger waves that are dominated by the weight of the water, called gravity waves. The overall wave-generation process transfers energy from the wind to waves on the water surface.

Wind must blow over the surface of the sea for some time, and continue over some distance, before a fully developed set of waves develops. The size and wavelength of water waves, thus, depend on the speed of the wind, its duration, and the distance, or fetch, over which the wind acts. Higher wind speeds develop higher waves of longer wavelength and period. These simple facts permitted Gilbert to understand why the beaches of Lake Bonneville are best developed along the Wasatch Mountain front on the eastern side of the former lake: The prevailing wind blows from west to east, to the extent that shoreline features are hardly recognizable on Lake Bonneville's western side where few waves ever beat.

Although the shape of a water wave may move at high speed over the water surface, anyone observing the motion of an object floating in the water knows that the water itself moves very little. There are two velocities relevant to waves. They both depend on the period T, wavelength L, and water depth H, in addition to the acceleration of gravity g for a gravity-dominated wave. In general, wave speed $c = L/T$. The first important speed is the phase speed, c_p, which is the speed at which some part of the waveform, its crest or trough, moves across the water. For waves of small amplitude (compared to their wavelength) the general expression for this phase speed is:

$$c_p = \frac{gT}{2\pi} \tanh \frac{2\pi H}{L}. \tag{10.24}$$

In deep water, $H >> L$, this simplifies to $c_p = gT/2\pi$. Similarly, in shallow water this equation simplifies to $c_p = \sqrt{gH}$. Note that the speed of deep-water waves depends on their period, so after traveling some distance long-period waves arrive earlier than short-period waves. This explains why long, slow waves are the first to arrive at the shore after a distant storm, followed by shorter, choppier waves. The dependence of wave speed on period is called dispersion: Wave packets tend to disperse as they propagate, spreading out and changing shape.

The other velocity associated with waves is called the group velocity c_g. This velocity determines how fast the energy associated with a packet of waves propagates. It can be derived from the phase velocity by a simple derivative:

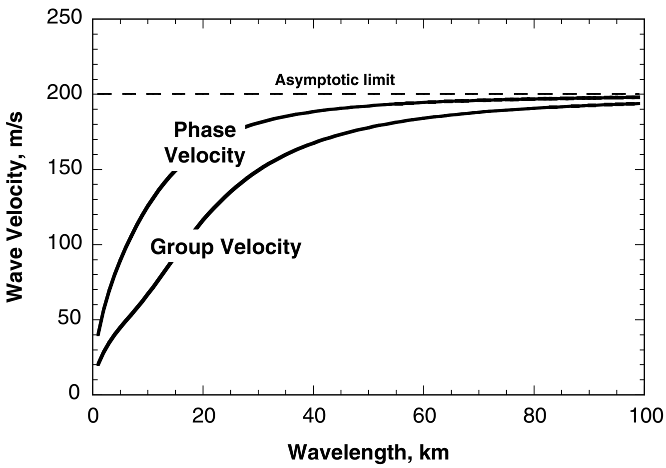

Figure 10.15 Group velocity and phase velocity for water waves on the Earth in water 4 km deep, equal to the average depth of the Earth's oceans. Long-wavelength waves approach a "shallow-water wave" limit when their length is much longer then the depth of the water. Short-wavelength wave velocities are increasing functions of their length in the "deep-water wave" limit.

$$c_g = \frac{c_p}{2}\left[1 + \frac{\frac{4\pi H}{L}}{\sinh\left(\frac{4\pi H}{L}\right)}\right]. \tag{10.25}$$

Figure 10.15 illustrates the relations between the phase velocity, group velocity, and wavelength for waves in 4 km deep oceans on the Earth. The important feature to note is that both wave velocities are highest in deep water, and that the group velocity is always less than the phase velocity at a given wavelength. Thus, the energy from a disturbance on the ocean propagates more slowly than the leading waves. Waves from deep water thus slow down on approaching the shore. Because energy is conserved, energy piles up in shallow water. More energy means higher waves, so we can deduce that the wave height must increase as the water shoals.

The energy E (per unit area of ocean surface) in a wave of amplitude A_0 (one-half of the vertical distance from crest to trough) is made up of equal contributions of the gravitational potential energy and the kinetic energy of motion. Its magnitude is given by:

$$E = \frac{1}{2}\rho g A_0^2. \tag{10.26}$$

Because energy propagates at speed c_g, the energy flux P in a wave is given simply by $P = c_g E$.

The path of a particle of water as a wave passes by is approximately a circle of radius A_0 near the ocean surface. At greater depths below the ocean surface the amplitude of the

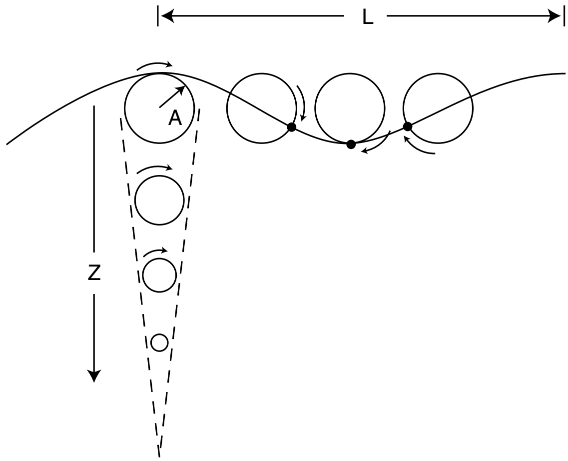

Figure 10.16 The orbits of particles of water in deep-water waves are nearly circular. The amplitude of this motion decreases exponentially with greater depth. Although the wave appears to progress from left to right, the water itself only moves in small circles, whose maximum diameter at the surface is equal to the distance between the crest and trough of the wave.

circle decreases exponentially. If z is the depth below the surface, the wave amplitude $A(z)$ in deep water decreases as:

$$A(z) = A_0 e^{-\frac{2\pi z}{L}}. \tag{10.27}$$

Figure 10.16 illustrates the rapid decline in wave amplitude with increasing depth. In water shallower than the wavelength, $H < L$, the velocity of the water on the bottom is still appreciable. The orbits of the water particles in shallow water are ellipses, not circles, which degenerate to a straight line parallel to the bottom on the seabed itself.

The rapid decrease of the amplitude of water oscillation with increasing depth below the surface defines the geologic concept of "wave base." It is well known that a submarine can ride out even the most violent storm by submerging to a depth comparable to the wavelength of the largest seas in the storm. In a similar manner, the seabed below some critical depth is unaffected by waves that may break up rocky shorelines. The ability of waves to erode the shoreline is, thus, limited in depth. The short-lived island of Surtsey off the southern coast of Iceland provides a fine example of the limited power of the waves. Surtsey was built by a series of volcanic eruptions in 1963. Well observed and widely reported in the news media, Surtsey was immediately attacked by the waves and within a few years most of its original area had disappeared below the waves. The eroded base of the island is still there, but it was planed off by waves to a depth of about 30 m below the surface, a depth that represents the effective wave base at this location. In a similar manner, volcanic Graham Island appeared in the Mediterranean in 1831. Its ownership was hotly disputed by Britain, Spain, and Italy, but wave erosion cut it down below the sea surface by 1832. Normal waves can move sand down to a depth of about 10 m, so the concept of wave base is somewhat fuzzy: The exact limit of erosion depends on the frequency of storms, the wavelength (and, hence, the exposure to wave-generating winds), and the wave amplitude. The important concept is that waves act only close to the surface of a body of standing water.

As waves approach the shoreline the water shoals and, as mentioned before, the waves increase in height, eventually breaking. Wave breaking is a complex phenomenon for which many theories have been proposed. A good summary can be found in the book by Komar (1997). A simplified way of looking at wave breaking is that it occurs when the velocity of the water particles at the crest of the wave exceeds the phase velocity. When this happens, the steepness of the wave front exceeds the vertical and the wave crest cascades over its front, dissipating much of its energy as turbulence. There are several ways in which waves break, each type distinguished by the steepness of the beach face. In order of increasing beach slope, these styles are called spilling, plunging (the iconic breaking wave is a plunging breaker), collapsing, surging, and, in the case of vertical seawalls, a reflected wave that does not break at all.

The principal consequence of wave breaking is that the wave energy, ultimately originating from the wind, is focused on the beach, where it is dissipated in turbulence. Where the waves impinge directly on rocky cliffs, the hydraulic pressures generated by the breaking waves may drive air or water into joints, loosening joint blocks or abrading the rock by dashing smaller boulders and sand against it. Wave action moves sand up and down the beach face and alternately offshore into bars, then onshore onto the beach again. Beach sand is suspended by each breaking wave and becomes vulnerable to transport by longshore or rip currents. Overall, wave energy makes the beach a highly dynamic environment in which erosion, deposition, and sediment transport are all active processes.

Coastal processes have received a great deal of study and limited space prevents a detailed treatment in this book: The interested reader is referred to a number of excellent texts on this subject in the further reading section at the end of this chapter. For our brief survey here the only other processes of major importance are wave refraction and longshore drift, as these are chiefly responsible for building beaches that might be seen from orbiting spacecraft.

Wave refraction. Wave refraction refers to the bending of wave fronts in water of varying depth. Once waves begin to "feel bottom" at a depth H equal to about $L/2$, the phase speed is proportional to the square root of the depth, $c_p = \sqrt{gH}$. Thus, the shallower the water becomes, the more slowly the wave fronts move. Consider a linear wave approaching a uniformly sloping shoreline at an oblique angle (Figure 10.17a). Because the wave moves more slowly in shallow water, the oblique wave gradually rotates to become more parallel to the shoreline as it approaches. It cannot turn exactly parallel to the beach, but the angle it makes with the beachfront is greatly decreased before it reaches the beach and breaks.

An oblique wave arrival means that the momentum transported in the wave is not completely cancelled when the wave breaks on the beach. A component of this momentum remains and generates a current, the longshore drift, which moves sediment in the direction of the acute angle between the wave front and the beach.

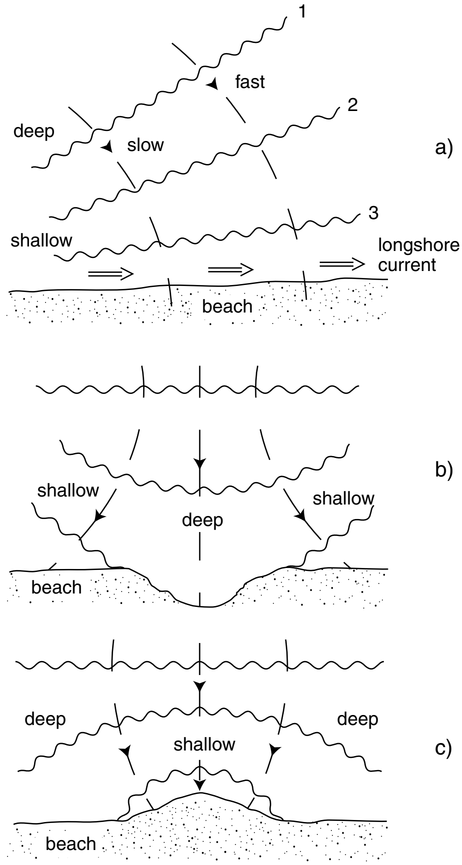

Figure 10.17 Behavior of wave crests approaching a shoreline. Panel (a) illustrates an oblique approach of the wave crests (wavy lines) to the shore. As the water becomes shallower, the wave velocity decreases so that wave crests near the shore travel more slowly than those farther out in deeper water. The net result is that the wave crests rotate and tend to become parallel to the beach as they approach. The oblique convergence also transfers a component of momentum along the beachfront to produce a longshore current. The dashed lines indicate the direction of energy flow perpendicular to the wave crests. Panel (b) illustrates the refraction of wave energy away from an offshore trough. The wave crest over the deep water in the trough moves ahead of the wave crests over its shallow flanks, turning the wave crests away from the axis of the trough and directing energy away from the trough and onto the adjacent portion of the beach. Panel (c) illustrates the opposite effect, when the waves approach over a ridge perpendicular to the shoreline. In this case the waves move more slowly over the shallow ridge and the wave energy is concentrated on the ridge crest. The combination of the focusing actions shown in (b) and (c) tends to even out submarine irregularities near the shore by wave erosion of ridges and filling of troughs.

If the bottom is not uniform, wave refraction acts to fill in hollows and erode protuberances. Figure 10.17b shows how waves approaching a shore are refracted over a submarine canyon running perpendicular to the beach. Because the canyon is deeper than the surrounding seafloor, waves moving down its axis travel more rapidly to the shore. Waves

Figure 10.18 Liquid methane lakes near the North Pole of Titan imaged by the Cassini synthetic aperture radar. Dark regions are smooth lake surfaces and brighter regions are the surface. Intermediate brightness levels near the lake shores indicate some radar return from the lake bottoms. Image is centered near 80° N and 35° W and the strip is 140 km wide. Smallest details are about 500 m across. The radar strip was foreshortened to simulate an oblique view from the west. Image PIA09102. NASA/JPL/USGS. See also color plate section.

to either side move more slowly and refract the wave fronts away from the canyon axis. Because wave energy flows perpendicular to the wave crests, the wave energy is refracted away from the canyon axis toward its edges. The margins of the canyon are, thus, more heavily eroded by wave action than its axis and so sediment tends to accumulate in the canyon, evening out the bottom contours parallel to the shoreline.

On the other hand, a submarine ridge, which might reflect the presence of a headland on shore, tends to focus wave energy over its crest, as shown in Figure 10.17c. The approaching waves collapse onto the shore right over the ridge, leading to intense wave action and erosion of the ridge, again evening out the bottom contours parallel to the shoreline.

Wave refraction, thus, tends to straighten out complex shorelines, focusing wave energy on promontories and diverting it from inlets. A newly flooded landscape, such as might be created by erecting a dam at the mouth of a river (Lake Powell on the Colorado River is a good example), presents a fractal shoreline of great complexity with protruding spurs and deep inlets everywhere. However, if the water level remains constant, in time these spurs and inlets are battered back and filled up, leading to a much more even shoreline.

The highly convoluted, fractal shorelines of the methane lakes on Titan constitute a puzzle (Figure 10.18). These shorelines show no sign of wave action; no bars, no spits, or eroded headlands. On the other hand, an observation of sunlight reflected from the surface of one lake appears to be perfectly mirror-like, showing no sign of the glitter typical of reflections from wave-ruffled liquid surfaces. Do lakes on Titan lack waves? If so, why? Titan's equatorial region is notable for its broad expanses of sand dunes, so winds must exist. Or are the levels of methane in the lakes constantly changing so that there is no time for erosion to straighten out the shorelines? At the moment the answers to these questions are unknown.

Longshore drift. Longshore drift is another powerful force that tends to straighten out shorelines. Beachgoers often confound longshore drift with the along-beach motion of sand particles in the back-and-forth swash of waves breaking on the beach face. This motion does drive some sand in the general direction of the longshore current illustrated in Figure 10.17a, but the current that flows parallel to the beach just offshore transports a far larger flux of sand. This current is driven by the uncompensated component of the momentum of the waves parallel to the beach. It is localized near the beach within the breaker zone. Ocean bathers are often unaware of its existence until they suddenly notice that they are far down the beach from where they thought they should be.

Sediment suspended by waves breaking in the surf zone is caught up in the longshore drift and transported parallel to the beach. This sediment-laden current is a true "river of sand" that moves large volumes of material along the shore. The direction of the longshore drift varies with the shoreline topography and the local direction of approaching waves. Coarse sediment deposited by rivers flowing into a lake or the sea is often caught up by the longshore drift and moved "down drift" to nourish beaches and build bars across inlets or spits out from headlands. G. K. Gilbert noted many gravely bars and spits created during the high stands of Lake Bonneville. These bars and spits are now dry, level ridges standing in the Utah desert to bear witness to the former existence of a large lake.

Many other currents and interactions occur close to the beachfront. Rip currents develop outside of the surf zone to return water pushed up onto the beach by shoaling waves that, unlike deepwater waves, transport water in addition to energy and momentum. Rip currents are often spaced periodically along the beach, their spacing determined by the excitation of edge waves, a variety of trapped wave that can exist only along the beach face. Beach cusps are rhythmically spaced beach features whose origin is still debated, but appear to be related to standing edge waves. All of these currents have complex interactions with tides and the material that makes up the beach. For more information, however, the reader should refer to the references at the end of the chapter.

Playas. Playas are shallow, ephemeral lakes that develop in regions dominated by interior drainage. Playas are flooded after rain falls in the drainage basins that discharge into them. The flowing water carries fine silt and dissolved minerals into the lake basin and then evaporates, leaving this non-volatile material behind. Playas are, thus, accumulation points for evaporite minerals. These minerals often form concentric rings around the center of the basin, ranging from the most soluble minerals in the center (usually halite and other chlorides on Earth) to less soluble minerals at the edges (typically carbonates on the outside and sulfates in an intermediate position between carbonates and chlorides). The edges of playas grade upslope into alluvial fans, which trap most coarse sediment moving from adjacent mountain fronts.

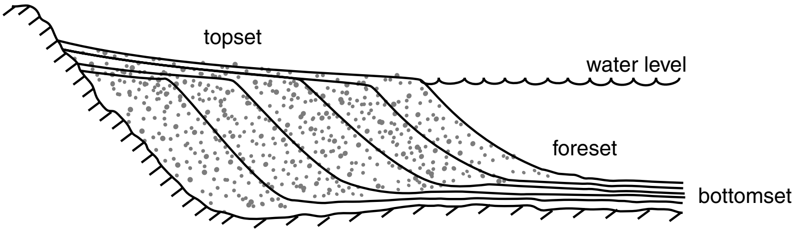

Figure 10.19 Formation of a delta near the mouth of a river or stream discharging into a larger body of water. As the stream loses its momentum in the lake or a sea it deposits much of the sediment it carries. This initially produces steep foreset beds of coarse material close to the shore and thin beds of finer sediment farther away in deep water. As the delta continues to build outward, topset beds are laid down on a shallow slope, while the foreset and bottomset beds build farther from the mouth of the stream. After Figure 14 in Gilbert (1890) and based on Gilbert's observation of deltas left behind when Lake Bonneville drained.

Because the water that floods playa lakes is often only a few tens of centimeters deep, the surfaces of playas deviate only very slightly from an equipotential surface – large playas form some of the most level (but not flat!) surfaces on Earth. Playa surfaces are usually devoid of rocks or coarse sediment, except in circumstances where other processes move rocks across them. A famous example is Racetrack Playa in Death Valley, whose surface is criss-crossed by the trails of boulders weighing many kilograms that somehow move across the level playa. No one has yet seen these boulders in motion, but they do shift between repeated surveys, perhaps during winter storms with high winds when the playa surface is wet and slick.

Playas may also serve as sources of fine dust, as they do in the American southwest. High winds drive sand grains and mud chips across their surfaces, raising dust that may be exported in suspension from their immediate vicinity.

Deltas. Deltas form where a river transporting sediment enters a larger body of water, decreases its velocity, and drops its sediment load near the shoreline. If this sediment is not carried away immediately by longshore drift, it builds up into an accumulation called a delta. The eponymous delta is that of the Nile River, which is triangular in plan like the Greek letter Δ. Because of interaction with waves and currents, deltas can be of many different shapes and sizes ranging from a few meters across to hundreds of kilometers, but all are sediment accumulations built out into a body of standing water.

The sedimentary layers that compose a delta are divided into three general types: Bottomset beds that underlie the delta, foreset beds that compose most of its interior and topset beds that cap it near water level (Figure 10.19). Bottomset beds, as their name implies, are laid down at the foot of the delta. They are typically composed of fine-grained sediment that formerly traveled in suspension and may be deposited by density currents that carry their sediment load far out into the body of water. Bottomset beds are usually thin and their sediments are graded from coarse at the bottom of each bed to fine near the top. Foreset beds are laid down in more steeply dipping sets. They are composed of coarse material originally carried as bedload that avalanches down the front of the delta. The dip of foreset beds may approach the angle of repose in small lakes, but in the deltas of major river systems they may dip as little as 1°. Topset beds are extensions of the floodplain. They come to overlie the foreset beds as the delta advances into the lake or ocean. Topset beds are composed of sand- and silt-sized material typical of the floodplain and typically show cut-and-fill channel features.

Sediment deposition in Earth's oceans is complicated by the mixing of fresh water from rivers with salt water in the oceans. Fine-grained sediments such as clay particles flocculate upon mixing with salt water and settle out more rapidly then they would if they had entered fresh water.

Turbidity currents. Upon arriving in a large body of quiet water, the suspended load from rivers usually forms a mixture of water and sediment that is denser than the surrounding water. If the time required for the sediment to settle out of the mixture is long compared to the time for the mixture to flow down the face of the delta, it moves downslope as a density current, called a turbidity current. Turbidity currents act somewhat as underwater rivers: They gouge underwater channels that may possess levees and create distributary networks on the lower parts of deltas or deep-sea fans.

Deep submarine canyons that head on the continental shelves off the mouths of major rivers, such as the Hudson River of New York, were initially thought to require enormous fluctuations in sea level when they were first discovered. Only after a great deal of research was it realized that turbidity currents, not subaerial rivers, cut these canyons.

Turbidity currents usually flow episodically. After a period of accumulation near a sediment source, a threshold is passed and the sediment pile collapses, mixing sediment with water and generating a muddy, underwater density current. Storms and earthquakes may also trigger the release of large and powerful turbidity currents.

The deposits of turbidity currents are called turbidites. Turbidites are layered accumulations of sediment that grade from coarse at the bottom of each layer to fine near the top. They often show evidence of high-velocity deposition, such as incorporation of rip-up clasts and upper-regime planar bedding. The thickness of individual turbidite beds is usually highly variable, reflecting statistically random triggering processes. Although turbidites are usually deposited in deep water, many of them incorporate shallow-water fossils and other debris acquired in the near-surface source area, before being carried to much greater depths.

10.3.5 Fluvial landscapes

Long-continued fluvial transportation and erosion create distinctive landscapes. We have discussed branching channel networks, but fluvial processes have additional characteristics. Undrained depressions are rare on fluvial surfaces: Lakes and other depressions are quickly filled by transported sediment. Terrestrial geomorphologists regard any undrained depression as an anomaly needing explanation. Even our ocean basins are anomalous: If it were not for plate tectonic recycling, all land surfaces would eventually be cut down below the level of the sea (to wave base) and the ocean basins would be partially filled.

Timescales are important when considering fluvial erosion. Annual floods and large multi-annual inundations adjust the forms of the channel and floodplain, but the landscape as a whole responds on a much longer timescale. This timescale can be estimated by comparing the volume of material that can be eroded from a drainage basin to the rate at which sediment is carried out of the basin. The sediment discharge has been measured for many watersheds on Earth and the result is about 10 cm of land surface (averaged across the entire basin) per thousand years for the pre-industrial Earth – present erosion rates are much higher. These rates depend upon relief in the basin as well as climate, so this is a very rough average. If we take the average elevation of the continents to be a few hundred meters, the terrestrial erosion timescale is thus a few million years. This is roughly the time required for fluvial erosion to strongly affect the Earth's topography.

Base-level control. The base level for large fluvial systems on Earth is mean sea level. This is the level to which long-term fluvial erosion tends to reduce the land, because erosion below the level of the sea is very slow. Of course, base-level control on the Earth has been complicated by hundreds of meters of sea-level change during the glacial cycles of the past 3 Myr, a fact that must be taken into account when interpreting modern fluvial landscapes. Local base levels may develop above long-lived lakes or other obstructions to fluvial downcutting. The base level may occasionally change drastically in extraordinary events, such as the nearly complete evaporation of the Mediterranean Sea about 6 Myr ago, which caused rivers such as the Rhone, Po, and Nile to excavate kilometer-deep gorges that are now buried by modern sediment.

The base level changes whenever a dam is built; we now have considerable experience with the changes that such disturbances engender. Aside from artificial dams, landslides and lava flows create natural dams that may block an existing drainage, creating a temporary lake and inducing changes in river flow that gradually work their way upstream.

Older discussions of fluvial erosion supposed that after some change takes place, conditions remain constant for a long period and the landscape has time to adjust to the change. It has become clear that the Earth's landscape is too dynamic for such long-term equilibrium. Changes usually occur on timescales that are short compared to the equilibration time, and so the landscape is constantly adjusting to perturbations.

Graded rivers. The concept of the graded river was famously introduced by J. Hoover Mackin (1948). Mackin proposed that, over the long term, the slope of a river is adjusted to the volumes of both the water and the sediment it carries. The volume of water increases downstream (for perennial rivers) as more and more tributaries feed their water into the main trunk river. At the same time the size of the sediment carried usually decreases downstream, permitting more of the load to travel in suspension. Mackin's idea was that the river strives toward a balance between the load to be carried and the water that carries it, such that the "long profile" of the river (its elevation as a function of distance from its mouth) tends toward a final state that is steep near its sources and gentle near its mouth.

Mackin proposed that the long profile of a river is self-adjusting: If some reach of a river should suffer a decreased slope for any reason, the sediment that was formerly in transit is deposited, building up the bed of the river upstream while the water downstream, relieved of its sediment load, erodes into its bed (this is currently happening around many artificial dam sites). Both processes tend to increase the slope and oppose the original perturbation of the profile. Similarly, if any reach becomes steeper for some reason, the capacity of the river to erode its bed increases and the river cuts into its bed, forming a step in the long profile that gradually propagates upstream until an equilibrium is again established.

A change in the nature of the sediment carried by a river has analogous effects. As described above, hydraulic mining in California's Sierra Nevada in the mid-1800s added a large volume of coarse debris to the rivers flowing into the Pacific (Gilbert, 1917). The rivers responded by steepening their gradients until the coarse gravel could move downstream. The extra load of gravel built up the riverbeds downstream, causing the rivers to overflow their previous banks and deposit gravel on the adjacent farmland. If the injection of coarse gravel into the headwaters had continued, the net result would have been a river system with a much steeper gradient from the ocean to the mountains, although this would have required burial of most of the interior valley of California – an outcome considered highly undesirable by the residents of the Golden State, which is why hydraulic mining is now strictly banned.

Landscape evolution. Theories of how the Earth's landscape evolves under the influence of fluvial processes are as old as geology itself. James Hutton in 1795 attributed river valleys to erosion by the streams that flow in them, an idea that did not sit well with his contemporaries or even with his intellectual heir Charles Lyell nearly a century later. Even after the fluvial origin of landscapes was accepted, ideas on how they evolved were qualitative and made few testable predictions. American geographer William Morris Davis (1850–1934) is widely remembered for his classification of landscapes as young, mature, and old, based on a presumed rapidity of tectonic movements, which create initial landscapes that are later dissected by fluvial erosion during an era of tectonic quiescence. Davis supposed that, following a long period of erosion, landscapes are reduced to a surface of low relief near the base level that he christened a "peneplain." Davis and German geologist Walther Penck (1888–1923) engaged in a bitter but somewhat fuzzy controversy over whether landforms "wear down" (Davis), with slopes everywhere declining as time passes, or "wear back" (Penck) who suggested that steep slopes retreat from their initial positions while retaining their steepness.

Classical ideas on fluvial processes, to which the ubiquitous G. K. Gilbert made many contributions, especially in his report on the geology of the Henry Mountains (Gilbert, 1880), divided the fluvial system into erosion, transportation, and deposition. Erosion takes place mainly at the level of valley sides and first-order rilles, which yield sediment that is carried through the stream system and is finally deposited in an alluvial fan or body of standing water.

Research in the modern era has focused on quantitative descriptions of each of the parts of this cycle. Until recently, most effort has gone into understanding individual processes, such as Bagnold's work on the physics of sediment entrainment and transportation, or have focused on hillslope processes, of which the book by Carson and Kirkby (1972) is a fine example. "Process geomorphology" has now become so large a subject, and the synthetic landscape evolution models of Davis and Penck have such a reputation for imprecision, that many recent textbooks avoid the topic of landscape evolution altogether.

Most recently, perhaps driven by the immense increase in computer power, quantitative syntheses of fluvial processes into landscape models have become possible and are now achieving impressively realistic results. These results are being subjected to quantitative tests through our recent ability to date landscapes through cosmogenic isotope methods. The first quantitative model of fluvial evolution of this kind was constructed by geophysicist Clem Chase (1991), who built a model of landscape evolution on a two-dimensional grid that evolved by simple rules suggested by fluvial processes. Simple as this pioneer model was, it produced very realistic landscapes that evolved in ways similar to those inferred for actual landscapes. Models of this kind are now reaching a high level of sophistication: The time has clearly come for our detailed understanding of individual processes to be synthesized into descriptions of how entire landscapes evolve. A recent review of progress in this area is by Willgoose (2005).

Although the evolution of fluvial landscapes has seemed a quintessentially terrestrial process (Martian fluvial landscapes are clearly not highly evolved), Cassini images of integrated drainage networks on Titan have created a new field for application of these models. Many of the basic parameters that control fluvial processes on Titan are unknown: How much precipitation falls and how it varies with time, what materials are being eroded, how Titanian "bedrock" (very cold water ice) weathers to sand-sized particles, and many other important facts are presently unknown. However, landscape evolution models themselves may shed some light on these questions. For example, how much sediment must be moved before an integrated drainage network forms? How much erosion does it take to turn a densely cratered landscape into a fractal landscape of connected drainage basins?

Further reading

Hydrology is an enormous subject on its own. One of the founding papers that can still be read with profit is Hubbert (1940). This paper takes a quantitative analytical approach that is difficult to find even in the modern literature. A comprehensive look at the older literature that still has much of value is Meinzer (1942). A full treatment of the percolation of fluids through porous media is found in Bear (1988). Students looking for a quick but mathematical overview of flow through porous media and its application to geodynamic problems can do no better than to consult Turcotte and Schubert (2002).

Fluvial processes are also the subject of an enormous literature, although many of the modern treatments focus more on societal problems such as pollution and water supply than on fundamental science. Two of the classic, science-oriented treatments are Schumm (1977) and Leopold *et al.* (1964). A thorough examination of the role of water in all surface processes, not just rivers and streams, is by Douglas (1977). The fluid mechanics of flow in open channels and an in-depth discussion of the various resistance formulas can be found in Rouse (1978). The interaction between flowing water (and other fluids) and its channel is covered by Allen (1970), while the best treatment of the physics of sediment entrainment is still Bagnold (1966). The sedimentological aspects of transport by water are well reviewed in the massive book by Leeder (1999). Coastal and wave processes in general are lucidly discussed in Komar (1997), while the more geomorphological aspects of shorelines and coasts are treated by Bird (2008). The modern synthetic approach to landscape evolution is too new to have texts describing it: The interested reader is referred to the short review of Willgoose (2005), but readers wishing insights into the history of landform analysis will be delighted by Davis (1969), or for a shorter and more comprehensive introduction by Kennedy (2006).

Exercises

10.1 Underground plumbing on Titan

During a particularly hot, dry spell on Titan, when temperatures rose to a balmy 97 K, over a period of one Earth year about a meter of methane evaporated from a (hypothetical) 100 km wide lake. In spite of the methane loss, no change in the level of the lake was observed (to a precision of ± 10 cm). If no surface methane flowed into the lake at this time, the loss must have been compensated by subsurface flow. Estimate the minimum permeability required in the subsurface to supply this loss. Comparing this to permeability of rocks on Earth (Table 10.1), what can you infer about Titan's subsurface?

10.2 Take Manning to new worlds

Show how the Manning equation (10.12) scales with the acceleration of gravity g by assuming that the Darcy–Wiesbach coefficient f is independent of gravity. Using this equation, compute the flow velocity of a flood on Mars that moved down a channel 30 km wide and scoured hills up to 100 m high above the channel floor. MOLA measurements indicate that the slope in this region was about 1 m/km. Assume that the material of the valley floor was large cobbles more than 20 cm in diameter. Use the same method to calculate the velocity of a methane stream on Titan that flowed in a channel 100 m wide and 3 m deep down a slope of 0.3°. Estimate the grain size of the material using available data/reasonable guesses.

10.3 Swing wide, lazy river

A meandering river on Earth flows through a channel 100 m wide and moves along at an average velocity of 3 m/s. It swings through a meander bend with an inner radius of 0.5 km. The centripetal acceleration in moving around the bend raises the level of the water on the outside of the bend relative to that inside. What is this elevation difference and is this large enough, in your opinion, to measure?

10.4 Entrenched meanders on Mars?

One of the major problems of analyzing ancient rivers is figuring out how much water once flowed through them. Nirgal Vallis on Mars is a sinuous trough nearly 500 km long and about 8 km wide in its lower reaches. In this area the sinuous undulations and alcoves along the walls have a wavelength of about 15 km. Although Nirgal has been attributed to sapping, suppose that its undulations were formed by a meandering river that gradually eroded down into the surface. Use the information given here to estimate the original width of its channel. Given this width, how might you estimate the discharge of the river that cut this gorge?

10.5 Low-gravity surfing

Evidence is accumulating for the former presence of lakes and perhaps oceans on Mars. Discuss, so far as you are able, how the acceleration of gravity affects waves on the surface of standing bodies of water (considerations of wave velocity and energy transport are particularly relevant here). In particular, how do you think low gravity might affect shoreline processes such as erosion and transport on Mars or Titan?