Chapter 42

Landform development

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INTRODUCTION

Subsurface water affects near-surface processes and landforms in a wide variety of ways. Its essential role in weathering, soil development, slope failure, and karst topography has long been acknowledged. However, its importance in other aspects of the development of landforms has just begun to be recognized in the last decade or so. In this chapter we discuss some major aspects of ground-water geomorphology and ways in which subsurface water can shape the Earth’s surface. In the next chapter, D. C. Ford, A. N. Palmer, and W. B. White discuss the special conditions of karst landforms; in this chapter we discuss other aspects of the role of underground water in landform development.

This chapter is organized as follows: first, some effects of water in the vadose or unsaturated zone above the water table, with emphasis on weathering and soil development and their influence on landscape (by Pavich), mass wasting and slope failure (Dietrich and Rogers), and hillslope hydrology (Dunne), with its influence on piping and pseudokarst (Parker) and on development of hillslopes and gully heads (Higgins). Next, the role of water in the saturated zone at or beneath the water table, and its effects on permafrost and pseudokarst (Sloan), land subsidence (Péwé), spring sapping and valley network development (Baker), submarine landforms (Robb), sea cliffs (Norris), scarp retreat (Higgins), and surface stream channels (Keller). Finally, ways are considered in which regional geomorphology can control ground-water behavior and hydrology (Coates).

WATER ABOVE THE WATER TABLE

Weathering and soil development

Water is essential for weathering, soil development, and life. It is not only the medium for chemical weathering of minerals,
but it is also vital for most mechanical weathering, such as "frost action" or hydration cracking, needle-ice action, salt-crystal wedging, and all types of biological weathering. Weathering is but the first step in soil development, in which these and other water-dependent processes further modify the regolith.

The soil occupies the interface of the atmosphere and lithosphere. Due to its unique textural, mineralogical, and structural properties, the soil forms a special zone of the hydrosphere. This zone has a higher mineral/solution ratio than other hydrospheric bodies such as streams, lakes, or the ocean. Within the soil hydrosphere, many processes occur, including infiltration of fluids into micropores, transmission of fluids along macropores, mineral-solution reactions—such as hydration, hydrolysis, congruent and incongruent dissolution—and dehydration of pores during periods of intense evapotranspiration.

The first-order dependence of soil on climate is a reflection of the importance of rainfall in soil-forming processes. In a general sense, hot-humid environments have higher rates of chemical alteration of rock to soil than do cold-arid environments. However, even in high-latitude cold climates, water is critical in mineral-alteration reactions.

At the temperate latitudes of North America, seasonal variation of rainfall and evapotranspiration related to climate (temperature, wind, solar insolation) is important in determining soil drying, development of macropores, and recharge of ground water to the subsoil through macropores. Under seasonal humid climates, the soil hydrosphere may control the volume and timing of ground-water recharge, so that the underground water not only affects soil development, but the soils themselves influence infiltration rates, thereby controlling further soil formation and landform development.

An example of soil processes as they affect landform in the United States is found in the southeastern Appalachian Piedmont, where one of the major hydrologic functions of clay-rich ultisols beneath relatively flat drainage divides is to partition precipitation. As much as 60 percent of total annual precipitation in Piedmont drainage basins is lost through evapotranspiration. Roughly half of the remaining precipitation runs off, owing to the relatively low infiltration capacity of the soil. The relatively small remainder recharges ground water in the regolith. The soil itself thus limits the volume of water that can infiltrate and reach the regolith, and thereby acts to control the rates of subjacent rock weathering.

On the Piedmont, it appears that the flattened upland landscape is a result of ongoing physical and chemical denudation processes that are in part controlled by soil hydrology. Desilification in soils and volume reduction during alteration of saprolite or sediment to ultisols are ongoing processes of mass and volume reduction that have lowered land surfaces and upland divides through Quaternary time. In this way, pedogenic processes can produce broad, flat divides of the type classically described as penepines. These processes are partly responsible for the morphology of large tracts of landscape.

Virtually all erosion is regolith erosion—that is, erosion of soil, saprolite, or weathered rock—because fresh unweathered rock is little affected by most erosional agents. Thus, erosional landscapes must strongly reflect the results of differential weathering and pedogenesis, which in turn are strongly dependent on subsurface water.

Over time, beneath stable geomorphic surfaces where erosion is minimal, depletion of soluble minerals and concentration of insoluble minerals can result in aluminous, siliceous, and/or ferric duricrusts. Although not recognized in the United States, duricrusts on old, cratonic rocks elsewhere in the world exercise control over subsequent landscape evolution. In some areas of Africa and Australia, a change from humid to arid climate over geologic time has left remnants of Tertiary and possibly older duricrusts that retard erosion and serve as caprocks for extensive uplands and plateaus. An extensive discussion of the relationship between soils and landforms can be found in Twidale (1984).

**Mass wasting and slope failure**

In recent years, engineering geology and geotechnical studies have advanced our ability to quantify the role water plays in controlling slope stability and evolution. Net movement occurs when shear stress associated with an inclined surface causes the rock or soil to strain slowly, or ultimately to yield, sometimes with catastrophic consequences. Ground water plays a critical role in these mass-wasting processes.

Shear strength of materials generally comprises two components—intrinsic cohesion and frictional resistance between grains of a rock or soil. In rock masses possessing joints, shear strength is derived from friction on opposing sides of the joint planes and from cohesion of the rock itself. The presence of water at less than saturation has been recognized to provide some cohesion through surface tension, which allows uncemented sand and silt to support a low vertical cliff. In argillaceous materials such as shales, however, saturation may cause a decay in pure cohesion and as much as a two-thirds reduction in shear strength (Rogers and Pyles, 1979). A further reduction in frictional resistance results from the buoyancy effect of submergence, which reduces the normal stress between grains or fracture planes. Saturation of the material also increases the downslope driving force. As shown in Figure 1, consequent rotational failure tends to reestablish equilibrium, although inertia involved in such movements often takes the rotation far beyond that which is theoretically predicted. Such rotational slope failures can involve as much as 4 km³ of material.

Failures in rock usually are associated with discontinuities in strength and permeability caused by stratification or jointing. For example, massive slump blocks in the cliffs of the Colorado Plateau are controlled by weak basal shales that yield when saturated. Jointing reduces the overall strength of the rock by diminishing or eliminating its cohesive strength and by providing frictional, but permeable boundaries where high pore pressures can develop.

Shallow soil failures commonly occur during periods of intense rains when high pore pressures are generated from shallow
subsurface flow perched on the soil-bedrock boundary. Stratification within the underlying bedrock may also bring ground-water flow up into the soil under sufficiently high pressures to cause instability. It is now becoming widely recognized that shallow soil slides typically occur in small colluvium-mantled bedrock hollows (Dietrich and others, 1986). Hollows can divide hillslopes into small unchannelized basins and focus the drainage of water and the discharge of debris from slopes to channels. Deeper ground-water flow also tends to re-emerge in the lower portions of the larger hollows and contribute to slope instability.

Static liquefaction (Seed, 1983) may be responsible in large part for the remarkable fluidity of debris flows generated in coarse soils, sometimes referred to as “blow outs.” Liquefaction begins during the initial slab failure of the debris, when rapid strain occurs under effectively undrained conditions. Excessive pore pressures leading to liquefaction may also occur where deep ground water is forced to the surface or where the B-horizon of the surface soil is much less permeable than the underlying coarse colluvium or weathered bedrock. In sharp contrast to this highly fluid mode of failure, clay-rich material tends to have a complex, time-dependent rheology as it responds to ground-water accumulation by deforming as a slow-moving earth-flow. Substantial recent progress has been made in both the monitoring and modeling of these flows (Keefe and Johnson, 1983). Results from these studies may well help in understanding the mechanics of solifluction.

The least understood but perhaps most pervasive set of physical transport processes influenced by water on hillslopes is that of slow mass movement caused by creep and biogenetic transport. Over large parts of soil-mantled landscapes, these processes appear to be the dominant transport mechanism. Some clay-rich soils subjected to strong seasonal drying show a strong time-dependent response to infiltrating water by slowly swelling, flowing downslope several millimeters, and then stopping, despite continued saturation (Fleming and Johnson, 1975).

**Hydrology of subsurface flow processes**

The model of runoff processes that until recently has been the most influential in geomorphology is that describing the generation of overland flow when rainfall intensity exceeds the infiltration capacity of the soil (Horton, 1945). Runoff of this kind (Path 1 in Fig 2) has been described thoroughly in other literature. However, over large areas of the Earth’s surface, infiltration capacities almost always exceed rainfall intensity, and most runoff travels beneath the soil surface (Hursch, 1944; and others).

When *Horton overland flow* does occur, the soil surface becomes saturated, and if the soil is vertically homogeneous, a wave of saturation extends downward into the soil. If the rainfall intensity is less than the saturated hydraulic conductivity of the soil, the soil surface will remain unsaturated, but its moisture content will increase until its hydraulic conductivity is high enough to pass water downward at the applied rainfall rate. After
the end of the rainstorm, soil water may continue to drain downward, reducing the moisture content of the near-surface soil. As water moves to any level in the regolith, it will either be stored in pores, raising the moisture content and therefore the hydraulic conductivity until the flux rate can be accommodated, or if that moisture content already exists, the water will continue its downward path, eventually reaching the water table (Path 2 in Fig. 2). It then follows a relatively long path to a stream channel, where it supplies the base flow of the stream. A simple flow path of this kind usually occurs in deep, permeable, and isotropic materials.

In many landscapes, however, the hydraulic conductivity of the near-surface zone decreases from the topsoil to the undisturbed parent material. This decrease may be gradual or abrupt, and is not necessarily monotonic, though for the sake of simplicity it will be treated as such here. As the infiltrating rainwater encounters a zone of diminishing vertical hydraulic conductivity, the moisture content of overlying layers is raised by storage. At some depth, moisture content will be raised to saturation, and the hydraulic conductivity to its saturated value. At this depth a perched zone of saturation develops over an obvious impeding horizon, or simply at some depth (varying between rainstorms) at which the saturated conductivity is less than the rainfall intensity (Fig. 3).

In the perched saturated zone, water is diverted downslope along the hydraulic gradient. If the soil and rock are sufficiently deep and permeable to conduct all the infiltrated water to the stream channel, the flow path will lie entirely below the ground surface (Dunne and Black, 1970). The pressure distribution in this subsurface flow influences the effective strength of the soil and rock, and may lead to mass failure or to other forms of seepage erosion described elsewhere in this chapter.

Under two circumstances, illustrated in Figure 2, the subsurface water may emerge at the ground surface, where it then has a potential for surface erosion. In one case, the water may emerge at a free face, such as a stream bank, gully head, or cliff (Path 3 in Fig. 2). In the other, the subsurface flux is augmented by infiltration as it travels downslope, until at some point on the hillside it may exceed the conveyance capacity of the soil because of reduced gradient, hydraulic conductivity, or soil thickness. When supply from upslope exceeds drainage downslope, the water content of the soil increases until the perched water table reaches the ground surface. Further excess of supply over drainage generates a pressure gradient that drives water to the surface, in a process referred to as exfiltration. The pressure gradient and associated water exert a stress on the soil particles, which may thus be entrained and carried downslope by the resulting saturation overland flow (Path 4 in Fig. 2). The soil and topographic conditions under which such emergence occurs have been documented by Dunne (1978) and others.

In most soils and sediments, subsurface water flows slowly enough and through pores that are small enough that the flow is dominantly laminar and obeys Darcy's Law. However, in some soils and sediments there are passages large enough to convey nonlaminar fluxes of water that are significant from both a hydrologic and geomorphic point of view. Such passages are formed by mechanical and biogenic processes, and include root holes, the burrows of soil fauna, shrinkage cracks, tension cracks between landslide blocks, solution openings, and tectonic joints. The origin and hydraulic characteristics of these "macropores" have recently been described by Jones (1981) and Beven and Germann (1982).
who review research on the subject. Macropores provide paths by which water can bypass slow flow through the soil matrix and, especially important, move through horizons of very low conductivity. Even thin, discontinuous zones of low conductivity in the soil may force water into macropores such as root holes and worm holes. Such locally concentrated flow promotes entrainment and transport of soil particles, both underground along the walls of the macropores, and at seepage faces, where saturated flow emerges at the surface.

Subsurface flow can erode mechanically in at least two ways—through the development of a critical seepage force that acts to entrain particles in water emerging from a porous medium, and through the application of a shear stress to the margins of a macropore with consequent removal of particles from the walls. The former process is “seepage erosion,” following Hutchinson (1968, p. 691), and the latter is “tunnel erosion” (Buckham and Cockfield, 1950). Either or both of these processes can lead to development or enlargement of pipe-like or tunnel-like openings underground. We use the term “piping” for the effects of the formation and enlargement of such openings. Also, through either of these processes, hillslopes or stream headcuts may be undermined, with consequent failure and retreat of the slope. The term “sapping” traditionally includes the results of such undermining.

**Seepage erosion.** Seepage erosion is the entrainment of grains by water emerging from a porous medium. Terzaghi (1943) defined the critical conditions required for the simplest case of instantaneous lifting, or “heave,” of a relatively large mass of cohesionless material by vertical outflow beneath a horizontal surface. The magnitude of the vertical component of the pore-pressure gradient at the surface must be large enough for the upward drag force on a particle or aggregate at the surface to exceed the immersed weight of the grain.

Zaslavsky and Kassif (1965) extended the treatment of seepage forces to the case of cohesive materials on a sloping surface. In their analysis, erosion occurs as a tensile failure when the vectorial sum of the fluid drag and the normal component of the soil weight averaged over particle or aggregate exceeds the cohesion of the soil (Fig. 4). These forces in turn depend upon several hydrologic and lithologic factors, as indicated by the following equations.

The fluid drag ($F_d$) on a grain or aggregate is assumed to be proportional to the macroscopic drag force averaged over many particles,

$$F_d = -c_1 V ho_f g \nabla \phi \cdot \hat{n} = c_1 V ho_f g Q \cdot \hat{n} \frac{K}{K},$$

where $c_1$ is a shape factor, $V$ is the volume of the soil element, $\rho_f$ is the fluid density, $g$ is the gravitational acceleration, $\nabla \phi$ is the gradient of hydraulic head in the fluid near the surface, $\hat{n}$ is a unit vector normal to the surface, $Q$ is the specific flux vector of water, and $K$ is the saturated hydraulic conductivity of the porous medium near the surface.

The normal component of the soil’s weight is

$$F_g = -V (\rho_s - \rho_d) (1-p) g \cos \theta,$$

where $\rho_g$ is the density of the mineral phase, $p$ is the porosity, and $\theta$ is the slope angle. The negative sign in this case indicates that the force acts downward.

The vectorial sum of $F_d$ and $F_g$ is resisted by the magnitude of the cohesion,

$$F_c = c_2 D^2 \sigma_t,$$

where $c_2$ is another geometric coefficient relating the particle diameter, $D$, to the projected area of the grain or aggregate normal to the resultant of the driving forces $F_g$ and $F_d$, and $\sigma_t$ is the tensile strength of the porous medium.

Seepage entrainment occurs when locally

$$F_c = |F_g| F_d + F_g,$$

and the associated velocity of flow is great enough to transport the grains or aggregates that are mobilized.

Where such a condition occurs along the base of a hillside, seepage erosion saps the slope and the hillside retreats. Where seepage is concentrated at one or a few points, a reentrant forms in the slope (Fig. 5). Such a disturbance of the boundary of the porous medium causes convergence of flow at the head of the reentrant, increasing discharge and therefore the rate of erosion and retreat. Where the soil is sufficiently cohesive, an arch and vertical walls can survive around the site of the initial seepage entrainment. Continued headward erosion may then produce a subterranean conduit or pipe.

**Tunnel erosion.** Once storm runoff has penetrated into a macropore, the water may then travel underground in sufficient quantity and with sufficient velocity to shear sediment from the walls of the conduit, which may then be eroded and enlarged. In a general way the fluid shear is proportional to the hydraulic gradient, the diameter or width of the conduit, and its hydraulic roughness. The shear stress can be particularly large beneath a steep hillslope or close to the bank of a stream or gully.
Where the shear stress is sufficiently large, grains are eroded from the margins of the conduit. The subsequent fate of the eroded grains depends upon the velocity of flow and whether the conduit has an outlet from which water can drain freely. Where there is an open connection to the atmosphere and sufficient head gradient, the velocity may be great enough to transport the eroded sediment as bedload or in suspension.

The shear stress required to erode the margins of a tunnel is radically reduced where the soil or sediment consists of a dispersible clay. All that is then required is a flow velocity sufficient to transport the resulting colloidal suspension. The susceptibility of a soil or sediment to dispersion is enhanced by high concentrations of monovalent cations, particularly sodium, in the pore water and on the clay particles. Water traveling along conduits is drawn into pores and cracks in response to capillary and osmotic potential gradients. There it causes swelling, fracturing into aggregates, and dispersion of particles. The physico-chemical principles governing the processes of dispersion are summarized in a collection of papers edited by Sherard and Decker (1977).

Under favorable circumstances, dispersion can also reduce the critical pore-pressure gradients required for seepage erosion. For example, where dispersible silty sediments with a conductivity sufficiently high to allow significant saturated outflow are connected with a source of ground water, dispersive seepage erosion can be quite rapid. However, in the experience of Sherard and Decker (1977, p. 5), most dispersive clays are so impervious that no significant seepage entrainment occurs, so that in order to initiate erosion it is necessary to provide a path for concentrated leakage. The low conductivity then forces flow from the surface or from a shallow, permeable horizon into cracks and other macropores, which are then rapidly enlarged by dispersion-accelerated shear. Many of the field occurrences of piping described in the papers edited by Bryan and Yair (1982) appear to be of this type.

**Piping and pseudokarst**

Although it occurs in many humid areas (e.g., Jones, 1981), piping is a major factor in the erosional process in arid and semiarid lands, where it affects alluvial valley fills, gully walls, hillslopes, and even barren hillslopes (Fig. 6).

Piping occurs commonly in unconsolidated silt, clay, and loess. It is also a destructive agent in certain shales, claystones, and chemically altered volcanic ash and tuff. In such rocks the geomorphic effects may simulate solution erosion of calcareous rocks. Where piping is highly developed, erosion of the underground drainage conduits produces a host of karstlike features, such as sinkholes, caves, ragged gullies where pipe roofs have
collapsed, and natural bridges. The result is a “pseudokarst” topography, complete with ephemeral subsurface streams and a sinkhole-scarred landscape (Fig. 6). However, although the forms produced by piping resemble karst landforms, they are generally not so large nor expansive in area, and their duration as landforms is relatively short. Whereas a true karst may maintain its physical features for centuries, pseudokarst may develop or change radically as the result of a single heavy rainstorm.

Some of the largest pseudokarst features result from enlargement of soil pipes and can approach karst features in size. Officers’ Cave, in the John Day Formation of Oregon, consists of a 200-m-long system of underground drainage tunnels fed by sinkholes. It includes several large cave rooms, and the entire system is drained by a master cave stream. However, despite its appearance of permanence, parts of the cave increased in volume about seven times between 1914 and 1962 (Parker and others, 1964), and additional extensive changes have occurred between then and 1984. Some soil pipes that have developed in irrigated fields recently reclaimed from sagebrush near Phoenix, Arizona, have been found to extend as much as 75 m inland from a nearby arroyo into which they discharge at the base of the arroyo wall. However, such large pseudokarst features are exceptional. Most pipes in valley fill appear to be less than 10 m long and begin in desiccation or stress cracks paralleling the arroyo walls. Such is the situation pictured in Figure 7, where piping in the Quaternary valley fill of Aztec Wash has produced an extensive subsurface drainage system. There runoff locally concentrated by highway culverts has exacerbated the erosional undermining and threatened the highway itself.

Pseudokarst such as these may show marked changes wrought by the erosion of a single heavy rainstorm. The intense summer storms of drylands in the western U.S. may create flash floods in the desert arroyos. The ragged, vertical piped walls of these arroyos, such as those shown in Figure 7, crumble and
slump into the arroyos. This forces the thalweg to shift to the opposite wall, and renewed undercutting occurs there. In Aztec Wash, some arroyo channels have doubled in width in only a few hours as a consequence of a single flash flood.

Piping is most common in denuded heavy soils and bedrock with a high proportion of montmorillonite and illite (Parker and Jenne, 1967). Such materials crack widely and deeply when they lose their moisture by evaporation in the hot summer sun and the drying winds. The resulting stress-desiccation cracks are deepest in a strip along each side of a deep arroyo, and their size and depth diminish away from the arroyo. Surface runoff flows into these wide, deep cracks; where they are drainable and the hydraulic gradient is sufficient to cause tunnel erosion, the walls of the cracks erode readily, especially at the base, where a widened floor is created. Eventually, the upper walls become wetted enough to swell shut, but in the meantime a horizontal pipe has developed at the base, bridged by an arch of cohesive soil. This pipe then becomes a locus for subsequent drainage, enlarging with each rainstorm, and in the course of time may become a cave large enough to afford range animals respite from summer heat or winter snow and frost. As the pipe is further enlarged, or as the land surface above it is lowered, it eventually collapses, often in sections, and thus evolves from a subterranean passage to a series of conduits with alternating suberial reaches, and then to a rapidly growing continuous gully. In a few years, such erosion may destroy a large part of a valley fill or even entire hillsides.

The role of seepage erosion in hillslope and gully development

Many soils do not form soil pipes. Instead, diffuse or concentrated outflow of soil water from porous granular materials may entrain and carry away particles at sites where saturation overland flow emerges at the surface. Such seepage entrainment and erosion undermine hillslopes and gully heads. Then slope failure causes the oversteepened slope to retreat. Such sapping is similar to the spring sapping produced by outflow of ground water, described below, except that the flows are intermittent—even rare in some cases—and are generally of much smaller discharge. Nevertheless, the effects may rival or even overshadow those of running water in the development of many hillslopes and gully systems. This role of shallow subsurface water in landform development has just begun to be recognized in the last two decades (Lepold and others, 1964; Dunne, 1978; Higgins, 1984; Howard and McLane, in press).

WATER AT OR BENEATH THE WATER TABLE

The geomorphic effects of ground water in the saturated zone tend to be more pronounced and recognized than those of water table partly because the reservoir is larger, the circulation is more continuous, and it has been more intensively studied. The features that result vary according to many factors, some of which are discussed in the following sections.

Influence of ground ice on landforms of cold regions

Permafrost is ground that has a temperature colder than 0°C continuously for 2 or more years (Péwé, 1974), but it is not necessarily frozen (Sloan and Van Everdingen, this volume). It forms and is maintained in a climate where the mean annual air temperature is 0°C or colder. About 20 percent of the land area of the Earth is underlain by permafrost, including about one-half of Canada and nearly 85 percent of Alaska.

The thermal regime of permafrost is dependent on the quantity of heat affecting the permafrost and the overlying active layer that freezes and thaws annually. Permafrost can build up (aggrade) and thaw (degrade) in numerous ways, all controlled by the thermal regime. Changes in the regime may depend on climatic, geomorphic, and vegetational factors.

Frozen ground water, or ground ice, tends to be concentrated in the upper levels of permafrost. The occurrence of ground water in the permafrost region is discussed by Sloan and van Everdingen in this volume. Ground ice may take many forms, some of the more common being pingo ice, ice lenses, and ice wedges. Ground water influences the geomorphology of permafrost regions in two distinct ways. First, the formation and growth of ground ice produces landforms such as ice-wedge polygons, palsas, and pingos. Second, the thaw of ground ice produces thermokarst features such as thaw lakes, alas, and beaded streams (Fig. 8). The resulting hummocky ground surface resembles karst topography found in limestone areas.

Ice-wedge polygons, the most extensive cold-region landforms created by freezing of ground water, have an ice wedge coincident with their borders. The borders tend to be raised or depressed with respect to the central areas. During the thaw season, low-centered polygons often contain ponds in the central areas, whereas high-centered ones hold water in the bordering depressions. Palsas are mounds or more irregular forms about 1.5 to 6 m high that generally have peat as an important constituent, contain perennial ice lenses, and occur in bogs. Palsas are characteristic of the subarctic, where they commonly occur in areas of discontinuous permafrost. In contrast to palsas, pingos are large perennial ice-cored mounds that tend to be more or less circular in form, 3 to 60 m high, and 15 to 500 m in diameter (Fig. 8). Pingos are necessarily associated with permafrost, and like ice-wedge polygons are key indicators of polar or subpolar environments.

Thermokarst comprises karstlike topographic features produced by the melting of ground ice and the subsequent settling or caving of the ground. One of the most conspicuous kinds of thermokarst is the thaw, or thermokarst lake. The basins of thaw lakes form or are enlarged by the thawing of ice-rich frozen ground.

Beaded streams consist of series of small pools connected by short watercourses. The pools result from the thawing of ice masses, commonly at ice-wedge polygon intersections. The connecting drainage generally lies along thawing ice wedges and therefore tends to comprise short, straight sections separated by
angular bends. Where ice-wedge polygons are degrading, pronounced intertongue mounds as much as 3 to 15 m in diameter and 0.3 to 3 m high can be left standing in relief.

Permafrost features are reviewed briefly by Black (1976), and a more comprehensive review of periglacial processes is by Washburn (1980). Brown (1973) has discussed ground ice as an initiator of landforms in permafrost regions.

**Land subsidence as a consequence of ground-water withdrawal**

Land subsidence is caused by a variety of natural and man-made processes. Surface sinking or collapse may follow thawing of ground ice, as mentioned above. It may also result from subsurface solution of soluble rock masses. Effects owing to solution of carbonate rocks are discussed in Ford and others (this volume). Leaching of salt may produce similar effects. A number of karst-like features of the Texas panhandle can be related to dissolution of subsurface Permian evaporites; local dissolution of as much as 150 m of salt at depths as great as 500 m is largely responsible for regional topographic development there, including the location of the Pecos and Canadian rivers (Gustavson and Finley, 1985).

Subsidence can also be caused by withdrawal of subsurface fluids. At some sites it is caused by pumping of petroleum. This was first noted at the Goose Creek oilfield of Texas in 1925, but the best-known example is the area of Long Beach, California, where as much as 9 m of subsidence has taken place. Extraction of petroleum may also play some part in local subsidence in the Houston-Galveston area, although most of this has been attributed to withdrawal of ground water. Indeed, excessive pumping of ground water, with associated artesian-head decline or water-table lowering, is a major cause of land subsidence in many parts of the world. Subsidence caused by ground-water extraction was recorded as early as 1933 in the Santa Clara Valley, California, and has since been recognized as a serious problem in many areas, notably Venice, Italy; Mexico City, where subsidence now exceeds 7.5 m; and parts of the western San Joaquin Valley, California, where an area of more than 13,000 km² has subsided as much as 9 m (Poland and Davis, 1969).

A spectacular result of land subsidence in Arizona is the formation of hundreds of earth fissures (Péle, 1984). These long, narrow, eroded tension cracks occur in unconsolidated sediments, typically near the mountains, along the margins of the basins where ground-water levels have dropped from 60 m to more than 150 m. Some fissures are 1 to 2 km long. They are generally perpendicular to drainage, intercepting surface water, which then erodes hairline-width cracks into gullies as much as 3 m wide and 6 m or more deep. Many, if not most, fissures form as a result of

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Figure 8. Collapsed pingo on Mackenzie Delta near Tuktoyaktuk, northwestern Canada. Ice-wedge polygons in foreground; oriented thaw lakes in background. Photo 18 August 1954 by Troy L. Péle.
differential subsidence and compaction over buried bedrock hills, ridges, or fault scarps. Some may occur where there are variations in type and thickness of alluvium, and variations in water-level decline. Economic losses caused by this land subsidence and earth-fissure formation are increasing rapidly.

A less pronounced but significant kind of subsidence results from "hydrocompaction," where unconsolidated low-density sediments are saturated for the first time, generally by irrigation water. This has accounted for as much as 4.5 m of settling in part of the San Joaquin Valley, California.

Spring sapping and valley network development

Where ground water follows a deep route, like that of Path 2 in Figure 2, and where the flow is concentrated it may emerge in a spring. Spring outflow may directly entrain and remove soil or rock particles by seepage erosion, or it may help to concentrate erosion at the site by enhancing rock weathering. In either case the slope locally becomes steepened and undermined so that sapping occurs. This can contribute to the development and headward extension of valleys.

Spring sapping can contribute both to channel and valley development. The relationship to channel networks has been reviewed by Dunne (1980). In a summary of field-experimental and theoretical analyses, he outlined a model of headward channel growth and branching by sapping (illustrated here in Fig. 5) that contrasts with channel network development by overland flow. Models for the latter predict network development by piracy and cross grading (Horton, 1945) or by headward growth through abstraction (Abrahams, 1984). Abrahams finds that networks developed by overland flow can be differentiated from those developed by spring sapping by certain aspects of their morphology.

Unfortunately, geomorphic evidence for channel network processes is rarely discernible in the complex hillslope-channel relationships of valleys. Valley processes involve a considerable component of nonfluvial degradation. Although fluvial incision may drive the hillslope systems of valleys, the picture can be complicated by the action of greatly enhanced past processes that result in a relict valley morphology. One reason that sapping has been underappreciated as a geomorphic process is that lowered water tables or desiccating climatic conditions have reduced its influence in many Holocene valleys (Higgins, 1984). In other valleys the results of spring-sapping processes may be obscured by modification of valley forms by nonsapping morphogenetic processes.

Excellent examples of valleys formed by sapping have long been known from the Colorado Plateau, where massive sandstone units are eroded by perched water emerging from bedding-plane boundaries (Stetson, 1936; Laity and Malin, 1985). These valleys show numerous distinctive attributes (Fig. 9), including the following: elongate basin shape, low network drainage density, low degree of interfluvе dissection, widely spaced and short tributaries to main trunk valleys, theaterlike valley heads, prominent structural control of networks, irregular junction-angle relationships, local examples of long and narrow interfluvеs between adjacent valley segments that join at unusually acute angles, steep-sided valley walls meeting valley floors at a sharp angle, irregular variation in valley width as a function of valley length, relatively high drainage densities in upstream portions of basins, and local examples of hanging valleys. Many of the attributes of these valleys can be seen in miniature valley systems that can be observed forming entirely by ground-water outflow sapping on some beaches during falling tides (compare Fig. 9 with Fig. 11 or with other photographs in Higgins, 1984).

Permeable volcanic rocks, especially basalts, are also dissected by valleys that may involve spring-sapping processes. The large springs of the Snake River Plain, Idaho, have a probable relation to headward valley growth where seepage-induced weathering concentrates erosion and sapping at the springheads (Johnson 1939). In the Hawaiian Islands, valley development by overland flow is inhibited by the extremely high permeability of the lava flows that comprise the individual shield volcanoes. Nevertheless, less-permeable volcanic ash blankets on the older shield volcanoes (such as Mauna Kea, Kohala, and Haleakala) support the development of long, parallel, V-shaped valleys. Where deep incision encounters the underlying permeable lava flows, the valley morphology is transformed to U-shaped cross section. These steep-walled flat-bottomed valleys are the result of enhanced

Figure 9. Distinctive valley morphology associated with spring-sapping processes in massive sandstones of southeastern Utah. Steep valley sides and theater-like valley heads are created by headward valley growth as sandstone caprocks are undercut by ground-water outflow.
weathering at the water table (Wentworth, 1928) and development and recession of cirque-like valley heads. Perennial flow is maintained by large springs. Hanging valleys have developed as the deep U-shaped valleys recede headward and capture the more shallowly incised V-shaped valleys.

Perhaps the most exotic valley systems that have been attributed to sapping are the ancient networks that dissect heavily cratered uplands on the planet Mars (Pieri, 1980; Baker, 1982). Nirgal Vallis (Fig. 10) illustrates a Martian valley typical of many that formed early in the history of the planet, when liquid water was an important agent of denudation (Baker, 1985). The network shows strong structural control, theater-headed tributary valleys, and numerous other attributes of sapping morphogenesis.

A variety of small-scale analogs have been invoked to understand the processes involved in drainage system development by sapping. Because of material, time, and scale differences these studies must be considered suggestive of process operation rather than definitive in identifying the precise genesis of complex large-scale valleys. For example, intertidal beach-face channels (Higgins, 1984) provide models of network development as water emerges from sand during falling tides (see Fig. 11).

It is clear that spring sapping has profound implications for morphogenesis on both Earth and Mars. Paleohydrologic, paleoecologic, and denudational studies of valleys on the two planets must give special consideration to the operation of this fundamental geomorphic process.

**Submarine geomorphic effects of ground-water processes**

Processes related to ground-water discharge and pore pressures also modify underwater landscapes. Submarine artesian spring sapping and ground-water-induced mass wasting were proposed many years ago as processes on continental margins by Stetson (1936) and Johnson (1939).

Beneath continental shelves, particularly along passive margins, pressure gradients can be generated subaerially recharged aquifers that extend to the continental slopes. Pressure gradients may also be caused by eustatic or local relative lowerings of sea level that increase the head. Pore-water overpressures may also result from rapid deposition of sediment or from compression by tectonic forces. Ground water discharging from the sediments can modify submarine terrain by such processes as seepage-face erosion and sapping, by enhancing or triggering of mass wasting, or by solution.

Since large amounts of sediments accumulate along continental margins, where conditions conducive to outflow are most commonly found, submarine erosion, mass wasting, and solution by ground water may help explain extensive submarine unconformities that developed on ancient continental margins during drops in sea level (Vail and others, 1977). A large volume of ground water is stored in sediments within the zone of sea-level variation. Numerical modeling shows that sea-level drops are more significant in creating long-term pressure differentials on continental slopes and fresh ground-water outflow in deep water than is subaerial recharge of artesian aquifers during high, stable sea levels (Leahy and Meisler, 1982; H. Meisler, personal communication, 1984).

An example of submarine erosion by discharge of ground water may be found on the lower continental slope off New Jersey, where recently acquired sidescan-sonar images and observations from a submersible vessel show valleys in outcrops of truncated coastal plain strata (Robb, 1984). These valleys, located in water depths between 1,500 m and 2,100 m, have theaterlike heads, steep walls, generally flat floors, and basins along their courses. Fragile clastic dikes protruding from cliff faces and excavated trace fossils within tafoni-like cavities suggest that erosion in those places took place by some particle-by-particle mechanism such as seepage-face erosion, and not by bottom-current scour or mass wasting. Features resembling kluftkarren, rillenkarren, and solution pits further suggest that solution of chalky Eocene strata may have resulted from discharge of fresh or mixed waters.

Solution and mass wasting related to ground-water discharge may also have contributed to the removal of large volumes of carbonate rocks to form great erosional escarpments like those bordering the east and west sides of the Florida plateau. Active submarine springs, subsurface (Miocene) sinkholes, and surface sinkholes relict from periods of low sea level (Popenoe and others, 1984; Hebert, 1985) are evidence of relatively shallow outflow around the Florida peninsula. Active deep-water outflow of ground water there at depths greater than 3,000 m at the foot of the West Florida escarpment in the Gulf of Mexico was recently discovered by Paull and others (1984).

Another region where groundwater-influenced processes
may modify subocean topography was pointed out by Arthur and others (1980), who suggested that dewatering of sediment accumulations compressed by underthrusting could increase rates of mass wasting in trenches. Complex dendritic canyon systems have been found at some plate boundaries in parts of the oceans that are isolated from the continents, where turbidity-current erosion related to active terrigenous deposition is minimal. Dewatering of sediments, coupled with fracturing, and mass wasting triggered by earthquakes, may play a part in the formation of such networks.

**Sea-cliff erosion by ground-water outflow**

Ground water can contribute to sea-cliff retreat by seepage erosion, enhanced weathering, or solution of carbonate minerals (Norris, 1968). Where coastal cliffs are weakly consolidated, emerging ground water readily weakens the rocks by removing intergranular cement, by dislodging grains by seepage pressure, and by encouraging the growth of vegetation near the discharge point. Vegetation adds weight and organic acids to the cliff face and also wedges rocks apart as root growth takes place. In areas where outflow is concentrated, spring sapping may produce alcoves, often with relatively lush patches of vegetation. It is probable that many short, steep-sided canyons cutting the sea cliffs of southern California are due at least in part to intermittent spring sapping.

Along some calcareous coasts, sea-cliff retreat results largely from the development of intertidal nips or notches. Much of this undercutting is caused by the activities of intertidal organisms, but in some places it may be caused largely by solution. Where coastal limestone is porous and permeable, slow effluent seepage of ground water all along the water line may dissolve the rock, especially where fresh ground water has mixed with sea water. Such mixing of waters, even where both are saturated with respect to carbonate minerals, can produce an unsaturated solution, as described by Back and others (1984). Where outflow is locally concentrated in coastal springs, brackish mixed waters floating on sea water may corrode nips along the shore (Higgins, 1980).

The role of ground water in contributing to the development of coastal landforms is more widespread and certainly more important than commonly acknowledged, and should be considered routinely in all studies of coastal erosion.

**Seepage weathering and cliff sapping**

Just as concentrated outflow of ground water in springs can initiate the recession of steep-sided alcoves and valley heads, so too can diffuse seepage from the base of a slope promote its wearing away and undermining so that the slope retreats more or less uniformly, maintaining a steep front. As this effect has no formal name, we shall call it cliff sapping. Its progression is well illustrated on some beaches where outflow of ground water during falling tides produces miniature landforms that result solely or mainly from seepage erosion and sapping and that may serve as models of much larger landforms (Higgins, 1984).

In the example illustrated in Figure 11, an extended steep-headed valley is developed in the upper left, where sapping by ground-water outflow was concentrated by a partially buried boulder. Elsewhere in the scene, seepage erosion by relatively nonconcentrated outflow has formed an "escarpment" with shallow alcoves. Characteristic braided outflow channels, terraces, and fallen masses of sand along the miniature cliffs of this example are also typical of larger landscapes that can be attributed to ground-water seepage and cliff sapping. For example, compare the escarpment in Figure 11 with that shown in Figure 12, an aerial photograph of an area at least two orders of magnitude larger. Both of these scenes resemble a view of a much larger embayed escarpment near Cameron, Arizona, reproduced by Stetson (1936, Fig. 2), who proposed that both it and the similar appearing submarine southern face of Georges Bank were formed by sapping by ground-water seepage at the base. Similar escarpments, still larger by several orders of magnitude, border the "chaotic terrain" of Mars (Sharp, 1973), which is thought to have been formed by liquefaction and outflow following widespread thawing of ground ice.

Simple entrainment and removal of erodible grains suffice to sap low cliffs in loose sediments, but this mechanism cannot, by itself, account for the erosion of consolidated rock in larger escarpments. There, cliff sapping must result from a group of related processes. These may include intensified chemical weathering, leaching, and solution within the seepage zone, coupled with enhanced physical weathering of the periodically damp or saturated rock face by granular disintegration or flaking owing to wetting-drying (Smith, 1978), salt-crystal wedging, root-wedging, rainbeat, and, especially in temperate to cold regions, congelification, including needle-ice wedging, which Sharp (1976) found effective even in southern Oklahoma. The combined effect of these processes can be termed seepage weathering. With sufficient hydraulic gradient, seepage outflow may at times be forceful enough to entrain the particles loosened by seepage weathering, but it is unlikely that diffuse seepage can transport the grains very far. Actual removal of the weathered products must then owe to other agents, such as wind, slopewash, and streamwash, or in the case of sea cliffs, to wave action. The combined effect of all these is to sap the base of a cliff so that it retreats by parallel backwasting.

In most places where cliffs have been formed by sapping and seepage weathering, seepage is no longer active and the landforms are now relict. The lack of present activity provides one reason why seepage weathering and erosion have not been generally recognized in the origin of such cliffs and of the extensive land-scapes that they front. Another reason is that there have been few studies of this process in the places where it does occur. Preliminary results of one such study—of recession of the Ogallala escarpment of the Southern High Plains—have been published by Osterkamp and Wood (1984). However, more studies are needed of seepage weathering and cliff sapping, which in some places are
among the most important contributors to the denudation of the Earth's surface and the shaping of its landforms.

**Effects of ground water on channel form and process**

Discharge of ground water may not only initiate the development of valleys, and even of entire drainage networks, but can influence stream channel form and process in the valleys that result. For example, complex interactions between riparian vegetation, shear strength of channel banks, and channel morphology result from changes in ground-water level.

Redwood Creek, near Orick, California, drains 720 km² of mixed forest land varying from Douglas fir in the upper part of the watershed to Redwood in the lower. In recent years the watershed has experienced accelerated erosion resulting from high-magnitude winter storms and from timber harvesting. Accelerated erosion has resulted in channel aggradation and subsequent rise in ground-water level adjacent to the main stem of Redwood Creek. In the upper part of the watershed, the local rise in water table has killed Douglas fir trees that had previously stabilized large alluvial terraces. Loss of these trees has contributed to accelerated erosion and widening of the channel (Nolan and Janda, 1979).

The Carmel River drains 660 km² in coastal Monterey County, California. Two major episodes of channel widening in the Carmel River have occurred in this century (Kondolf, 1983): one was produced by major flooding, and the second was associated with a lowering of the water table, which killed riparian vegetation stabilizing the channel banks. The 1978 to 1980 erosional episode, which locally increased channel width by more than three times, occurred in response to floods that were of only moderate intensity. This is in marked contrast to the high-magnitude flood of 1911. Detailed studies by Kondolf (1983) clearly suggest that those reaches of the Carmel River that suffered the greatest bank erosion in 1978 to 1980 were the same where the die-off of riparian vegetation due to drawdown of the water table from pumping was also the greatest. These examples suggest that either rising or lowering of the ground-water level may affect the bank vegetation and consequently initiate significant bank erosion, channel widening, and a tendency to braid.

Ground water and surface water are intimately related in the fluvial system, and base flow is often provided by point sources as well as more diffuse ones. Variability of base-flow discharge commonly results from a change in ground-water level, and may produce several effects in stream channels.
Harrison and Clayton (1970) reported that field observations in a small Alaskan stream suggest that seepage of water into or out of the stream bed seems to alter the stream’s competence. They reported that gaining reaches, characterized by effluent ground water, had considerably greater competence than losing reaches, characterized by influent ground water. Simons and Richardson (1966) had predicted such an effect by suggesting that effluent ground water should produce a seepage force that would reduce the effective weight of bed material and consequently decrease the stability of the bedload, leading to easier transport. Harrison and Clayton (1970) designed laboratory experiments to test the hypothesis, but the results were conflicting. Later flue studies by Richardson and Richardson (1985) confirm that in sand-bed channels, seepage of water into the channel can increase stream power and sediment transport for relatively low flow events. In bedrock channels of some streams in Indiana, presumed solutional processes create grooves, notches, and large boulders, apparently during low flow. This aspect of the role of base flow discharge in altering channel morphology is largely unstudied.

Effects of ground-water discharge on channel form and process remain a part of fluvial studies that has been little researched. What work has been done is speculative, controversial, and has led to contrasting ideas. In spite of this, future endeavors to explore relationships between ground water and channel process should be fruitful, provided interactions between the ground water and surface water can be measured and evaluated.

**GEOMORPHIC CONTROLS ON GROUND-WATER HYDROLOGY**

Just as subsurface water affects surface processes and landforms, so too the landscape and regional geomorphic history affect the behavior of ground water.

**Topography and general terrain conditions**

The landscape consists of a series of interconnected hillslopes. The variability of hillslope length, steepness, and shape are reflected in variable water-table configurations, which in turn translate into other aspects of ground-water conditions.

In the crystalline terrane of Statesville, North Carolina, LeGrand (1954) found that topographic slope is the most important control for well yield. The greatest well production there occurs in valleys and broad ravines, with lowest yield where wells are at or near the crests of narrow hills. Also he found that wells drilled in the centers of flat upland areas produce more water than wells farther down on the hillslope.

Water chemistry can also be influenced by topography. Ground water in the lowlands and lower parts of hills commonly contains more dissolved solids than does water upgradient. Because ground-water flows from the upper slopes and discharges at lower elevations, ground water near hilltops and upper slopes is younger and has had a shorter time to dissolve minerals than the waters near the base of hills.

The conditions that determine the position of the water table in valleys and basins include topographic relief, drainage basin size, amount and composition of the valley fill, permeability distribution, and climate. In all terrains the depth to the water table is greatest at higher elevations and least in valleys and basins. Humid areas also have shallower water tables than do dryland environments. Although effluent systems (with ground water recharging streamflow) tend to occur in humid regions, geomorphic factors are also important. For example, in humid areas it is unusual for second- or third-order basins to have perennial streams, and in fine-grained bedrock it usually requires at least fourth-order basins to have strong base-flow conditions. In arid areas the water table is usually so deep that only the major through-flowing rivers have permanent flow with drainage basin order of at least seven. Topographic slope, water-table position, and ground-water hydrology are also closely related in coastal and small island environments where there is an interface between salt and fresh water (see Hunt and others, this volume; Back, this volume).

**Hydrologic properties of sediments**

The geomorphic environment determines the character of unconsolidated sediments during their erosion, transportation, and deposition. When these sediments become aquifers, their history and properties can influence the behavior of the ground water. Thus sediments, which may also include regolith, alluvium, and colluvium, play important roles in the hydrologic realm by determining the recharge-discharge system and the nature and rates of infiltration, transmissivity, throughflow, and underflow. Geomorphic controls are especially significant in controlling ground-water hydrology in fluvial and glacial terranes. Ground water can also occur in significant quantity in loess, dune deposits, and beach deposits.

The caliber and fabric of fluvial sediments are a function of the flow regimes of the depositing river. High-energy streams produce coarse granular materials, whereas quiet waters and lacustrine conditions yield finer grained silts and clay. Although the latter sediments possess high porosity, they form poor aquifers because of their low permeability. Such deposits typically originate as overbank and back-swamp materials of floodplains. Coarse sands and gravels form in point bars and lateral accretion sites. Yields of wells in these sediments are higher because of greater permeability.

In similar fashion, glaciaton has a significant impact on determining ground-water hydrology in glaciated regions. Deposits of till generally have extremely low permeability, in contrast to much more permeable glacial outwash. Thus any search for new ground-water supplies in fluvial or glacial terranes may have to reconstruct the geomorphic history of the sites in order to locate the areas where conditions permitted the greatest amounts of sand and gravel to be deposited.

Bedrock can also be an important consideration in geomorphic and hydrologic studies because it influences both topographic characteristics and drainage density. Carlston (1963) studied
Figure 13. The relation of drainage density to base flow and floods. A, Base flow ($Q_b$); B, Mean annual flood ($Q_{2.33}$). From Carlston (1963).

the hydrologic relations of 13 small monolithologic basins in nonglaciated areas of eastern United States. He found that drainage density, surface water discharge, and ground-water movement are all part of a unified physical system. His data (Fig. 13) show that ground-water discharge into streams, as reflected by base-flow conditions, is directly related to drainage density. Furthermore, because the transmissibility of the terrain also controls the proportion of precipitation that infiltrates the ground, as contrasted with that which flows off the surface, surface water or overland discharge is inversely related to drainage density.

In a hydrogeomorphological comparison of basins of the Catskill Mountain Section of the glaciated Appalachian Plateau with those of the Susquehanna Section, Coates (1971) also found drainage density to be an important indicator of base flow in streams. The Catskill basins are underlain primarily by sandstone, whereas the Susquehanna basins are largely of shale and siltstone. Mean base flow is nearly twice as great in the 12 sandstone basins with the lowest drainage density, as compared with the 13 shale-siltstone basins.

Landforms

Specific landforms can be a predominant control of ground-water hydrologic conditions. Although alluvial fans and pediments are not restricted to arid regions, they are much more prevalent there, as in the American Southwest. Superficially these landforms seem quite similar, but they differ greatly in the availability of ground water. The chance of developing wells on pediment surfaces is slight because bedrock there occurs at shallow depths, whereas the deep sands and gravels of alluvial fans can yield much larger ground-water supplies.

Buried valleys in glacial terranes can be very important aquifers. The largest buried valley system in the United States is that of the ancestral Teays River. Prior to glaciation this river flowed west through the states of Ohio, Indiana, and Illinois. Norris and Spieker (1966) have provided a detailed analysis of this buried valley near Dayton, Ohio, where it provides one of the principal aquifers for the city. The Binghamton, New York, metropolitan area has several buried preglacial valleys of the Susquehanna and Chenango rivers. The most important of these underlies an 8 km² area in the urban core, and is one of the highest producing aquifers in the region (Randall, 1977).

Kaye (1976) provides an interesting case history of what happens when there is a failure to identify landforms and to recognize the character of the materials that compose them. Engineers who designed several new buildings in the Beacon Hill area of Boston assumed that the hills there were drumlins and that the sediment that composed them was clay-rich glacial till. It had been interpreted that this till would be largely impermeable and so would not present any serious water-flow problems. Instead, excavations at the sites encountered large ground-water flows that required different and costly construction methods and delayed the projects many months. Instead of drumlins, the Beacon Hill area is part of an end moraine with ice-contact deposits of clay, sand, and gravel that locally contain abundant ground water.

Slope aspect and orientation

The Appalachian Plateau has often been described as a maturely dissected upland terrain with well-rounded hills interspersed among rivers with floodplains. In the Susquehanna Section, Coates (1971) has called attention to the large-scale asymmetry of the upland bedrock hills, which reflects the thickness of till. On south-facing slopes, till averages about 30 m deep, but the depth to bedrock on the steeper north-facing slopes averages only about 3 m. These hills have been called till shadow hills (Fig. 14). Because till does not yield sufficient water for household wells, water supply in this region must come from bedrock aquifers. In the underlying shale and siltstones it usually requires well penetration into rock of at least 50 m to produce 5 to 10 gallons (17.5 to 35 liters) per minutes. These ground-water conditions have important financial implications. Well yields on north-facing slopes are greater because of increased secondary porosity in the upper bedrock zones and steeper water tables. Thus, homeowners there can save about $2,000 on wells because they do not have to be drilled through so much till or contain so much casing as their counterparts on the south slopes.

Landscape trends can influence hydrologic conditions in
other ways. For example, in the Northern Hemisphere, southern and southwestern slopes receive greater insolation than northern and northeastern slopes. This leads to differences in soils, vegetation, and water retention and infiltration rates. Such factors create feedback systems that affect the amount of water that enters the ground-water zone.

Directional trends are also produced in glaciated terrains. In a study of the glaciated part of the Susquehanna River basin, Coates (1971) showed that an important relationship exists between the orientation of tributary streams and base-flow discharge. He identified seven stream azimuths that are associated with different levels of base flow. The most dramatic base flow discrepancies occur when south-flowing streams are compared to their north-flowing counterparts. On average, south-flowing streams deliver more than ten times as much ground-water discharge as do north-flowing streams. The reason for this great discrepancy stems from the composition of the valley-fill materials, which can be attributed to the deglaciation history of the valleys (Fig. 15). During recession of the glaciers, mostly fine-grained sediments were emplaced into proglacial lakes imponed by the ice margin. These sediments settled out on what would later become the north-facing slopes. However, as recession continued and the lakes were drained, the glacial meltwater fines were washed out of the south-facing slopes, leaving lag deposits of sand and gravel. These coarser sediments on south-facing slopes are now much more permeable, provide better aquifers, and produce larger surface and ground-water flow rates than do the finer materials on the north-facing slopes.

These examples give some idea of how the geomorphology of an area can help determine its ground-water hydrology. Topography, surface processes, Earth materials, regional geomorphic history, slope steepness and aspect, and even such features as drainage density all can be important variables that affect ground-water behavior as well as the base flow of surface streams. In such cases it is important to understand the geomorphic history and nature of the landscape in making hydrogeological inferences.

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