The overall appearance of a fluvial landscape is controlled by the density and structure of the valley network which drains water and sediment from the land. The channel network and the valley network are rarely equivalent. Instead, the channel head characteristically lies tens to hundreds of meters down-valley of the drainage divide (Figure 7.1 and 7.2). Consequently, the finescale topography of many hillslopes is dominated by a network of unchanneled valleys which drain directly to the channel network or to a swale upslope of the channel head, i.e. the source–basin area (Figure 7.3). This view of the landscape, emphasizing its three-dimensional character created by elaboration of a valley network, points to the need to understand: (1) where the channel head is located within the valley network; (2) how this position may change with external influences (such as climate and land use) and internal adjustments (inherent instability); and (3) what role fluctuations in channel-head position play in valley development and persistence.

It will be argued here that the channel head is a crucial landscape feature to study. The distance from the drainage divide to the channel head controls drainage density, which in turn sets the average hillslope length. Fluvial landscapes evolve as a consequence of the generation of local relief by channel incision. Instability at channel heads can generate destructive debris flows or gullying, and heavy sediment loading downstream. Because the channel-head location at any time is controlled by the relative magnitudes of sediment supply and the erosion potential due to runoff, fluctuations in climate or land use, which alter runoff rates, surface erodibility and sediment supply, cause shifts in channel-head location. Therefore, the channel head may be the element of the coupled hillslope-channel system that is most sensitive to changes in external factors. Erosion and deposition cycles which sweep through drainage basins may be largely influenced by shifts in the balance controlling channel-head locations. Hence a physically based theory for channel heads would provide the key to a large number of theoretical and practical problems in geomorphology.

The importance of understanding the controls of channel-head location was apparently first recognized as a geomorphically significant problem by Horton (1945, p. 346–8) in his pioneering study of channel networks and basin development. He argued that landscapes are progressively dissected by an expanding channel network...
FIGURE 7.1 Map of the Tennessee Valley area of Marin County, California, showing the topography and channel network observed in the field (solid heavy line) and the approximate extent of colluvium which has accumulated in valleys (stippled areas). Heavy dashed line encloses the area mapped. The large black area in the center of the figure is an artificial pond. Capital letters indicate channel heads shown in detail in Figure 7.10. Reproduced by permission of AGU from Montgomery and Dietrich (1989).
FIGURE 7.2 Sketch of the relationship between a valley head (A) and a channel head, which may exhibit a gradual (B) or abrupt (C) transition from the unchanneled valley floor or swale until hillslope lengths bordering the channel are everywhere shorter than a critical length, $x_c$, at which overland flow is unable to cause erosion and, consequently, channel initiation. Thus, in Horton’s scheme controls on $x_c$ directly influence average hillslope length and drainage density. Horton also recognized the likely role of groundwater seepage in affecting channel-head location, and reasoned that the distance to the drainage divide would be reduced below $x_c$, ultimately reaching the point where “groundwater flow is no longer effective” (p. 348, Figure 32).

In response to Horton’s initiative, studies of network structure were initiated, and thus channel heads had to be identified. This was occasionally done in the field (e.g. Morisawa, 1957; Maxwell, 1960), but more commonly some consistent criterion based on topographic map analysis was used without comparison to field observations (e.g. Langbein, 1947; Bowden and Wallis, 1964).

FIGURE 7.3 Unchanneled hillside swales draining directly to channels or to channel heads, northeastern Coast Range of California. Two of the swales (A) and (B) have recently been channelled by debris avalanches.
A resurgence of interest in this problem has been motivated by field and theoretical studies of the consequence of threshold conditions for channel initiation on landscape morphology (e.g., Schaefer, Elfriti and Barr, 1979; Dunne, 1980, 1990; Kirkby, 1980, 1987; Dietrich et al., 1986, 1987; Montgomery and Dietrich, 1988, 1989; Willgoose, 1989). The problem of defining where channels begin is also a key issue in the recent development of models that employ digital elevation data (e.g., Tarboton, Bras and Rodriguez-Iturbe, 1988). On a more practical level, channel-head studies have largely focused on gully development in response to land use (Ireland, Sharpe and Eargle, 1939; Bradford and Piest, 1977, 1985; Bradford, Piest and Spomer, 1978). These theoretical and practical studies have demonstrated that the channel head is a singularly important phenomenon deserving further research, particularly with regard to erosion mechanisms.

The focus of this chapter, then, is the channel head. We will discuss what it is, what it looks like, and what controls the location of the head within the valley network. Items (2) and (3) above, relating to the temporal aspects of channel-head position and valley development must await progress in understanding the formation of channel heads.

FIELD DEFINITION OF THE CHANNEL HEAD

A stream channel head is the upstream boundary of concentrated water flow and sediment transport between definable banks (Figure 7.2). It is not easy to provide a globally useful criterion for a well-defined bank, which may include a bedrock step or narrow zone of gradient steeper than both the neighboring hillslope and the transverse slope of the channel bed. The important criterion is that the bank must be recognizable as a morphological feature independent of the flow. For example, the edge of a stream of saturation—overland flow in a broad swale (Wilson and Dietrich, 1987, Fig. 4b) is not a bank. Nor is the edge of overland flow between microtopographic protuberances. While the steeper boundary must line the sides of the channel, the upstream end of the channel may grade smoothly into the unchanneled valley upstream, although at the heads of many channels, the topographic step continues around the upstream end of concentrated fluvial action. Such a step must be stabilized by interparticle cohesion augmented in many cases by the binding action of plant roots.

Upstream of the channel head one may observe ephemeral concentrations of sediment and oriented or trapped organic debris stranded by the declining runoff, but as the flow becomes confined between banks, the vegetation cover (if any) of the hillslope or swale is breached and the upstream end of the channel bed can usually be recognized from the presence of wash marks, small bedforms, an armored surface, or other signs of concentrated sediment transport.

Thus, our definition of a channel head is a morphological one, which is useful even in the absence of flow, and which agrees with that of Calver (1978, p. 233), who defined a fluvial channel as "an incision into the ground surface such that, if water ceased to flow, morphological evidence of its former course would, at least initially, remain apparent". The channel head is not synonymous with the stream head, which simply indicates the upstream extent of concentrated surface runoff at a particular time. For example, a stream fed by overland flow may episodically extend up an unchanneled swale above a channel head, or in dry weather the onset of streamflow may shift progressively downchannel as the outcrop of the water table shrinks (Hewlett and Hibbert, 1967; Gregory and Walling, 1968; Blyth and Rodda, 1973; Day, 1978; Dunne, 1978; Wilson and Dietrich, 1987). Throughout these fluctuations of streamflow the channel head persists as a morphological feature, and a reproducible drainage density can be defined (Maxwell, 1960, pp. 24–5). However, if there is a change in the magnitude or frequency of the forces forming the channel head, its position may change, but again it will always be definable independently of the more ephemeral and variable distribution of streamflow.
Thus, our approach to defining channel heads and therefore channel networks and their drainage density is very different from that of workers such as Tarboton et al. (1988), who define channel networks on a digital elevation model as those pixels that have an accumulated drainage area greater than some "threshold support area". These authors state (p. 1320): "If we regard a channel network as paths where water flows, it is possible to imagine . . . with higher and higher resolution, getting lower and lower orders of streams until we are literally looking at flows among the grass roots." It is our contention that observations made from the side of a channel will allow a non-arbitrary definition to be made.

Our definition of the channel head would exclude many linear depressions which convey stormflow. For example, in mountains and steep gullied terrains such as badlands, chutes are produced by selective weathering, mass failure and corrosion by sliding and rolling debris. These usually grade smoothly from an upper hillslope which does not exhibit signs of fluvial sediment transport, and they narrow downslope into a channel where water transports the fallen debris. Our definition of a channel head would exclude the peat flushes described by Burt and Gardiner (1982, Fig. 2b) from consideration as channels, although they are shallow, linear depressions and supply surface runoff to the channel network during and after storms. To judge from the description provided by Burt and Gardiner, the swales may be former surface channels or collapsed tunnel networks that have since been blanketed by peat, but subsurface investigations would be necessary to resolve this issue.

Nor is the channel head synonymous with the occurrence of a topographic planform concavity, as suggested by Kirkby (1978, 1980, 1986). There are many unchanneled swales, including those which lie upslope of channel heads (Figure 7.3). In Oregon Coast Range, Benda and Dunne (1987) found that the typical first-order drainage basin contained seven unchanneled bedrock hollows, with an average of four of them clustered around the tip of the first-order channel. Figure 7.1 is an example of a channel network determined in the field. All channel heads are found at the downslope ends of swales or hollows and many of these swales have tributary hollows. The conditions under which unchanneled valleys persist in a landscape undergoing net erosion are discussed later in this chapter and elsewhere in this book (Chapter 9).

A question exists as to whether there is a finest stable scale of channel incision. Schumm's (1956) work on drainage-network evolution on badland slopes equated channels with "permanent" features having recognizable drainage areas lying within visually distinct valleys. Rills were shown to be ephemeral features, removed each winter by frost action and formed again by summer runoff. They are distinct troughs incised in a smooth, planar surface with straight contours. Widths of troughs and of channels lying in valleys may be similar, so rills are not necessarily smaller. We follow Schumm's procedure and equate channels with features lying in valleys and therefore relatively permanent. This distinction is the same as that used by Melton (1957). Given enough time after surface smoothing, the hillslope may be eroded into valleys draining towards channels with dimensions scaled by the runoff and sediment transport from distinct drainage areas and with spacing scaled by regrading and competition for runoff and sediment.

The transition from swale to the channel is often discontinuous. In these cases a mappable channel shoals downvalley to a smooth unchanneled surface, which still further downstream is cut by another channel head. This sequence may be repeated many times, which on small or large headcuts means that the discontinuous reach can be hundreds of meters to kilometers long (Figure 7.4). Leopold and Miller (1956), Schumm and Hadley (1957), Leopold, Wolman and Miller, (1964), and Reid (1989) have provided detailed descriptions of discontinuous gullies in a range of environments. Discontinuities occur with either gradual or stepped headcuts, and the spacing of the cuts may be decimeters to decameters. It is not yet clear whether discontinuous channels are ephemeral and always evolve into continuous channels (Leopold et al., 1964), or are steady-
state sequences of headcuts which migrate upvalley but are replaced by similar incisions (Reid, 1989). Temporary discontinuous channels also occur where large macropores or tunnels collapse upslope from the head of a continuous channel (Figure 7.5).

TYPES OF CHANNEL HEAD

Little has been written about the morphology of channel heads. Here we draw largely upon our field experience and the field studies reported by others to propose a simple classification which

Figure 7.4 Discontinuous channel in a meadow in the North Forks Kings R. drainage, Sierra Nevada, California
FIGURE 7.5 Channel head formed by collapse of a tunnel eroded in a sodium-smectite-rich vertisol. The depressions upslope of the channel head are collapsed sections of the same tunnel. The channels were studied by Mitchell Swanson, who is acting as the scale, and colleagues (Swanson et al., 1989)
serves to emphasize that the head is a distinct morphological feature, and to focus our subsequent discussion on the relationship between process and form. Detailed description motivates more exact physical hypotheses, and illustrates that a simple mass-balance assumption for channel initiation is incomplete.

The transition from the upstream valley floor to a channel may occur without a significant topographic break, although commonly the channel head forms a distinct step or headcut. The channel may notch only the soil mantle or be deeply incised into alluvium, colluvium or bedrock. Processes involved in channel initiation and the dynamics of the channel head probably vary with the height of the head cut, so we have set out in Figure 7.6 a simple, self-explanatory classification of channel-head types based on depth of incision of the channel at the head. By this classification channel heads are gradual (without a measurable longitudinal break between hillslope or valley floor and channel), small or large steps, or small or large headcuts. This classification does not consider the planform of the head, but, by definition, the channel has definable banks and our field measurements indicate that height and width are crudely related, with width generally less than five times height. Typically, the faces of head cuts are concave downslope in planform, so the channel head terminates in the shape of a finger.

We have also distinguished channel-head types by the dominant runoff process generating them,

Figure 7.6 Typology of channel heads on the basis of incision depth and dominant runoff process. Sketches illustrate flow paths for Horton overland flow and subsurface flow. The smooth arrows indicate saturated flow, and the wiggly arrows indicate unsaturated percolation, including flow through macropores. Even in areas with significant Horton overland flow, the deep face of a large step or headcut can allow the emergence of erosive seepage. Saturation-overland flow drives erosion that incorporates features from both of the other runoff types. Plunge-pool erosion and slope-stability constraints become more important as the height of the feature increases.
to emphasize the relationship between runoff and the mechanics of the particular erosion process responsible for the position and form of the channel head. Such considerations will ultimately form the basis for a dynamical explanation of drainage density, but are poorly understood at present because of uncertainties about flow paths, flow mechanics, and regolith properties in the vicinity of channel heads. The two end-members in our classification are erosion driven by Horton overland flow and by exfiltrating subsurface flow, with erosion driven by saturation overland flow sharing characteristics of both. When one considers the erosion mechanics responsible for channel formation under different circumstances, it becomes clear that the field measurements and theoretical analyses required for quantitative prediction have not yet been rigorously defined or widely investigated.

Figure 7.7 includes a photograph and detailed topographic map of a gradual channel head in a weakly cohesive, sandy clay loam 500 m from the divide of an unripped convexo-concave hillslope with an average gradient of 0.015 in Amboseli National Park, South Kenya. The area, described in more detail by Dunn and Dietrich (1980), has a mean annual rainfall of approximately 250 mm, and a heavily grazed ground cover which locally attains densities of 65-80% in wet seasons, and decreases to a basal cover of 0-35% at the end of a dry season. The infiltration capacity varies with cover density and rainfall intensity between 10 and 70 mm/h; Horton overland flow and consequent rain-flow sediment transport are frequent and widespread. Figure 7.7 illustrates that the downvalley transition from hillslope to channel is smooth, although the gradient there (0.047) is twice those of the hillslope and channel. Within a distance of 2.5 m from the head the channel has deepened 7-10 cm to a well-defined, sinuous, sand-bedded channel with cohesive, convex sideslopes and a bankfull width exceeding 2 m. Figure 7.8, taken nearby, in similar conditions, shows the co-existence of channels with gradual heads (like Figure 7.6(a)) and others with abrupt steps, 8-10 cm high (like Figure 7.6(b)). Flow marks and sandy sediment converge above these steps (left side of Figure 7.8) but the depression there is only about 2 cm deep and has no obvious margin.

Large steps and small headcuts are also common forms of channel heads in East Africa, particularly where recent land-use changes have altered channel-head stability and led to upvalley propagation. Figure 7.9 shows a channel incised 2.5 m into weathered tephra since the removal of dense bush 40–50 years earlier. The vertical headcut (like Figure 7.6(i)) is retreating by tunnel erosion and plunge-pool erosion. The width-depth ratio of the channel at the headcut is approximately 2 and, typical of large steps and small headcuts, the channel width stays remarkably uniform downstream of the head. The channel is propagating up a gentle swale, presumably following the convergence of subsurface water, a portion of which is derived from Horton overland flow which enters cracks and erodes them into tunnels.

We have examined hundreds of channel heads in the coastal mountains of Washington, Oregon and California, and in the southern Sierra of California (Dietrich et al., 1986, 1987; Montgomery and Dietrich, 1988, 1989). We have also examined some channel heads studied by others in southeastern Brazil (Cocolho Netto, Fernandes and Edeyarde de Neys, 1988) and Japan (Onda, 1989). These areas have slopes generally less than 45 degrees, dense vegetation of forests, chaparral or grasslands, and a soil mantle typically 0.3–1.5 m thick. Saturation overland flow occurs on gentle slopes, but subsurface flow is responsible for the erosion of most channel heads.

Under these conditions all forms of channel heads occur. The gradual and small-step channel heads are sufficiently subtle that vegetation, particularly trees, may grow across the channel, or fallen trees and animal burrows may cross the channel, blocking the passage of sediment and causing the transition from hillslope to channel to be discontinuous. These discontinuous reaches may be hundreds of meters long, depending on the thickness of sediment along the valley axis in which the channel head periodically forms. In wetter environments the channel head is com-
Channel Network Hydrology

(a) Photograph and (b) detailed topographic map (with 0.01 m contour interval) of a gradual channel head approximately 500 m from a divide in Amboseli National Park, southern Kenya.

Dietrich, Reneau and Wilson (1987) about 30% of the channel heads were large steps and most of the others were gradual or small steps. Systematic surveys near San Francisco, California (Montgomery and Dietrich, 1988, 1989) indicated that 50% of the channel heads were large steps and the rest were gradual or small steps although a few small headcuts were also found.

Figure 7.10 illustrates channel heads formed in Tennessee Valley, Marin County, just north of San Francisco, California. The maps have the same scale and contour interval and cover only the lower parts of the unchanneled valleys. Figure 7.10(a) shows a channel head that begins in the lower right side (looking downslope) of a shallow landslide scar. In the nomenclature proposed...
here, the head is gradual but downslope the small channel quickly incises forming a distinct "V" shape in the contours. Shallow landslide scars are common at the channel heads in steep valleys. Figure 7.10(b) shows a small headcut at the upper end of a discontinuous channel in colluvium. The channel head forks, with the major channel tending in the direction of greatest drainage area, and the tributary heading towards the local maximum of topographic convergence. Figure 7.10(c) portrays a discontinuous channel beginning as a small step in colluvium. In the valleys draining to the heads shown in Figures 7.10(b) and 10(c) saturation overland flow is common, whereas it is not apparent in the steep valley above the head in Figure 7.10(a).

Some channels incise several meters into colluvium or alluvium, and their large headcuts are vertical in profile and concave in plan view (Figure 7.11). They are fed by subsurface flow which causes seepage erosion of the matrix as well as some tunnel erosion along macropores. The channel heads retreat upvalley and towards concentrations of subsurface flow emerging from unchanneled tributary valleys (Figure 7.11(b)). Their width-depth ratios are usually less than 3. Examples of these valleys in southeastern Brazil are described by Coelho Netto et al. (1988).

LOCATION OF CHANNEL HEADS

Channels generally terminate in topographically convergent areas which can range from large, low gradient alluvium-filled valleys to subtle swales on steep hillslopes. Although all analyses of channel networks depend in some fashion on the location
of channel tips, few field surveys have been reported that show channel heads located on a topographic map. The error in locating channel heads that results from the various assumptions used in defining channel tips by inspection of topographic maps has been investigated by Mark (1983) and uncertainty in drainage density and other network features associated with misidentification in the field was explored by Maxwell (1960). Published maps of field surveys of channel networks include those of: Mark (1983); Maxwell (1960); Melton (1957); Montgomery and Dietrich (1989); Morisawa (1957); and Schumm (1956).

Figures 7.1 and 7.12 show maps of field surveyed channel networks from different areas. In all cases channel heads are located at the downslope end of unchanneled valleys. The site represented by Figure 7.12(a) is in the Rock Creek basin of the Oregon Coast Range where Dietrich and Dunne (1978) described channel shortening and extension by the filling and emptying of colluvial wedges in bedrock hollows. The basin is underlain by gently dipping basaltic lavas and breccias cut by many dikes. The annual rainfall of about 2300 mm supports old-growth coniferous forest, but consistently through the basin the forest canopy shifts abruptly to alder at the channel head. Of the five channels shown, three begin as large steps due to landsliding, one begins as a small headcut in which the headcut appears to have advanced progressively upslope, and one begins gradually. Figure 7.12(b) is Maxwell’s (1960) map of the channel network in Fern Creek watershed on metamorphic and granitic rocks in Southern California. Annual rainfall of about 90 mm supports a vegetation cover that includes oak

Figure 7.9 Small headcut in weathered tephra, near Arusha, northern Tanzania. Headward retreat of the channel occurs mainly as a result of undermining due to plunge-pool erosion, tunnel erosion, and collapse.
woodland, chaparral and spruce. The proportion of the hillslope occupied by source-basin areas is much less than in the basins shown in Figures 7.12(a) and (c). Figure 7.12(c) shows the channel and hollow network in a relatively low-gradient area in the Southern Sierra Nevada of California, a site described by Montgomery and Dietrich (1988). It is underlain by granitic rocks, receives an annual rainfall of about 260 mm and is covered by oak woodland and grass. The area draining to the channel head is commonly made up of a network of unchanneled valleys. The channel heads are either gradual or small steps.

Arguing that the channel heads are located at some critical combination of runoff and valley slope, Dietrich, Wilson and Reneau (1986) showed that an inverse relationship between drainage areas above the channel and average valley slope existed for a limited set of data from northern California. Montgomery and Dietrich (1988,
FIGURE 7.11 Channel heads retreating into an alluviated and colluviated valley floor near Bananal, southeastern Brazil, described by Coelho Netto et al. (1988). The channel has bifurcated and the headcuts are retreating towards sources of subsurface flow emanating from unchanneled swales.
FIGURE 7.12 Channel networks, source basins, and hollow axes (dashed lines) in three landscapes. Maps are portrayed at the same scale with 40-foot contour intervals. (A) Channel network and source-basin areas for a portion of the Rock Creek catchment, central coastal Oregon. Due to the dense old-growth coniferous canopy the fine scale topography is not accurately mapped even in the most recent 1:24,000-scale mapping (US Geological Survey, 1984). Consequently, the contours do not reveal all unchanneled valleys. Drainage area to each channel head is delineated, not just the area of thickened colluvium as depicted in Figure 7.1. (B) Channel network and source-basin areas for a portion of the Fern Creek catchment, San Dimas Experimental Forest, southern California. Redrawn from a map made by Maxwell (1960) as part of an extensive field documentation of the channel network. (C) Channel network and source-basin areas in the southern Sierra Nevada, California.
reported strong inverse relations between drainage area and local valley gradient at the channel head for three areas represented in Figures 7.1, 7.12(c), and an area in southern coastal Oregon (Figure 7.13). They also showed that above the channel head the valley slope tended to be constant. Although other factors also varied, the drainage area to the channel head for a given slope was systematically larger for drier areas. One valuable aspect of the observations reported in Figure 7.13, is that they serve to test theoretical predictions for channel initiation over a range of climate from semi-arid to humid without the need to know specific properties of individual sites. Instead, a theory can be tested for whether it predicts the observed inverse trend.

Efforts to define an area-slope relationship have not always revealed a clear inverse relationship. Dietrich, Reneau and Wilson (1987) found no clear relationship between source area and average valley gradient over a broad range of gradient for the basins represented in Figure 7.12(a). They noted that the majority of the channel heads were found at springs emerging at the outcrops of relatively unfractured, fine-grained dikes and sills that would strongly influence ground water flow in the weathered vesicular lavas underlying the hillslopes. Hence, the fine-grained intrusive units acted like impermeable boards buried in a pile of sand, deflecting groundwater flow paths and dictating where water exfiltrated and thus the locations of channel heads. The other site which failed to exhibit a clear inverse relationship (San Pedro ridge, Marin Co., CA) had a very narrow range of average hollow gradient (Dietrich et al., 1987, Figure 3a). In addition, it is an area of relatively recent, extensive debris-flow activity and many of the channel heads have recently extended well upslope into the hollows. Also, the average hollow gradient rather than the local gradient at the channel head was used and this apparently tends to give less well-defined results.

Thus, in order to define a relationship between drainage area and local valley gradient at channel heads it is essential to obtain a broad range of slopes because there is considerable variation of critical drainage area for a given slope. It is important to document possible local effects that are natural (such as dikes) or anthropogenic (such as intense grazing, deforestation or road runoff) which influence channel-head location. To be consistent, it is essential that a specific criterion be used for a channel head, as described above.

The implication of the maps in Figures 7.12(b) and (c) and the inverse relationship between drainage area and local valley slope at channel heads in Figure 7.13 is that for a particular area...
channel tips lie at a predictable distance from drainage divides. Upslope of each of the channel heads used to construct Figure 7.13, however, lies an unchanneled valley in which thickened colluvial deposits mantle an underlying bedrock valley. In order for these unchanneled valleys to persist as the overall slope erodes, the bedrock in the valleys must lower either by dissolution processes, as suggested by Bunting (1961) for an area in southern England, or by periodic eva- sions of the colluvium and surface erosion of the bedrock, as suggested by Dietrich and Dunne (1978) for steep terrain in Oregon.

DRAINAGE DENSITY AND THE CHANNEL HEAD

For a given valley network the closer the channel head lies to the divide, the greater is the total length of channel in the catchment and consequently the greater is the drainage density. Although treated as a primary distinguishing property of catchment morphology, drainage density can vary with climatic fluctuations and land use for a particular catchment (Figure 7.14), whereas the valley density tends to be much more stable with time because its evolution requires the redistribution of a much larger volume of relatively resistant rock material than is needed for channel network evolution alone. Because of the direct linkage between channel-head location and drainage density, theories of channel-head location are an essential component of predicting landscape response to climatic and land-use change. Yet the morphology of many hillslopes, as defined by the spacing and size of valleys, is also probably controlled by processes acting to set the head location. In this case, however, the location is the minimum distance (or area) from the divide where the channel reaches with sufficient frequency to maintain a valley. Upslope of this point, infrequent periodic excursions may occur, but generally erosion is by unchanneled sediment transport. As in the case with drainage density, the less drainage area associated with the minimum channel-head location, the greater is the number of source areas to a channel and the closer the valley spacing, hence the greater the valley density. This view, then, argues that contributing adjacent slopes are everywhere too short, or not steep enough, or too divergent, for example, for conditions of channel initiation to be met. Schaefer, Elfrids and Barr (1979) and Tsuchimoto, Ohta and Nogachi (1982) have used the term “zero-order basin” for the minimum drainage area required to sustain the head of a first-order channel.

The surveys of channel-head location used to construct Figure 7.13 showed that in the wettest area, which probably has experienced the smallest climatic variation, channel heads extend to virtually every hollow on steep slopes, hence the finest scale of valley topography is probably related to the present locations of channel heads. However, given that the channel begins a certain distance from the divide, how is drainage density than predicted from this information? The following we briefly discuss studies that have linked channel heads and drainage density.

Horton (1932) proposed the use of drainage density as a quantitative measure of the degree of drainage development within a basin. He showed that for typical basins where channels are much less steep than the average hillslope gradient, the average distance between adjacent channels is simply the inverse of drainage density. Because he assumed that overland flow occurred on all slopes, he reasoned that one-half of this distance must equal the average length of overland flow. In general, however, one-half of the drainage density gives the average length between channel and adjacent divide, that is, the average hillslope length. Horton reasoned that drainage development would reach a stable network once all adjacent slopes were shorter than \( x_e \), his assumed “belt of no erosion” where it was argued that overland flow was too shallow to cause erosion. Hence, drainage density on an initially undissected surface should increase until the inverse of drainage density equalled two times the threshold value, \( x_e \). As mentioned earlier, Horton also reasoned that the channel head could be closer to the divide than \( x_e \) because of groundwater seepage (see later). In general, however, Horton's
concept of landscape evolution specifically asserted that a threshold condition for surface erosion exists, that where surface erosion occurs channels develop and, therefore, that there is a quantitative linkage between conditions controlling channel initiation and drainage density.

Schumm (1956) calculated the total length of channel for each channel order in his study basins and regressed average drainage area of each order against average channel length for that order. He found a linear relationship and the slope of the regression line he defined as the “constant of channel maintenance” which he claimed was the reciprocal of drainage density. The slope of the line is not equivalent to the drainage density as measured by sequentially calculating total channel length for progressively greater drainage areas down-valley through a
catchment. Rather it is the average drainage density for each order, and as such is strongly constrained by the drainage density of first-order channels and the topologic requirement that first-order channels connect to give second-order channels and so on. In other words, once the drainage density of first-order channels is set, even a random integration of the first-order channels (following the rules of no separation after joining) should produce the linear relationship found with a slope given by the first-order drainage density. As such, the “constant of channel maintenance” seems to be an artifact of channel network topology and bears no more fruit than the conventional definition of drainage density.

Montgomery and Dietrich (1989) argued that the relationship between drainage density and the average source area draining to a first-order channel head could be established using topologic relationships first defined by Shreve (1966). They showed that drainage density could be related to the length of the source area via four dimensionless numbers expressing the area–length relationships and length ratios for interior and exterior links. Data from Abrahams (1984) and from many different studies and direct field observations allowed Montgomery and Dietrich to set these four numbers which led to the simple result that drainage density is equal to the inverse of the source-basin length. At their study site the inverse of the mean source-basin length accurately predicted the drainage density. This implies that the average distance between adjacent hollows or first-order valleys, which is the inverse of the drainage density, also equals the distance from the divide to where the channel begins.

This may seem a surprising result because one might expect the distance to the channel head along the source basin to be shorter than the lateral distance between the valley and divide. As Horton (1945, p. 355) pointed out, however, the direction of drainage to a channel from the adjacent divide is rarely perpendicular. The less perpendicular the drainage, the more the critical length is directed parallel to the channel and the greater is the difference between that critical length and the horizontal distance from divide to channel (measured perpendicular to the channel). The channel or hollow axis and the drainage divides are roughly parallel, hence this angle of flow to the valley axis is a convergence angle. Consequently, assuming a constant critical length to channel initiation, the weaker is the convergence of flow toward the channel, the closer the channel spacing will be. Hence, to relate channel initiation to drainage density and the average spacing between channels, the relative degree of topographic convergence must be known or calculated.

**PROCESS AND FORM AT THE CHANNEL HEAD**

At the heart of the channel-head problem is the question of what happens to flow and sediment transport in the vicinity of the channel head to create an incision with well-defined banks that will persist and convey the water and sediment downstream. Three problems demand a quantitative understanding of the processes which form channel heads. First, there is simply the question of what controls the locations of channel heads within the topographic convergences we have called hollows or valleys. Second, (and really part of the first) is the question of what controls the upslope advance and downslope retreat of the channel head. The third question (and perhaps the most difficult), is what controls the size and shape of the channel head. While some research has been done on the first two issues, we know of no studies of the third.

On a qualitative level it is easy to describe the processes that act to erode the channel head and therefore control form and location. These processes are suggested in Figure 7.6, where a distinction is drawn between erosion associated with runoff caused by Horton overland flow and runoff due entirely to subsurface flow. Overland flow causes erosion either by net transport due to the shallow surface flow or by plunge-pool activity undermining the base of steps and headcuts leading to their mass failure. Subsurface flow, frequently concentrated by root, rodent and insect tunnels or other macropores, causes erosion by
seepage forces that cause liquefaction or Coulomb failure, or by scouring particles from the insides of macropores, or by lowering effective strength to trigger mass failure. Dissolution and a number of physico-chemical processes termed “slaking” may contribute to these processes. In some environments both overland flow and subsurface flow erode channel heads. Erosion of the step and headcut is also partly accomplished through ravelling and sloughing associated with drying. In all cases, upslope advance or downslope retreat depends on the rate of sediment removal through the downstream channel. All near-vertical steps or headcuts, including banks, require some cohesive strength in the headwall, whether it is due to unsaturated conditions, inherent cohesion in weathered regolith, or apparent cohesion due to roots. Figure 7.15 illustrates typical morphology, runoff, and erosion processes at gully headwalls. Although useful descriptions of the processes illustrated in Figure 7.15 (and other effects such as ravelling and spalling) have been reported, and some quantitative analyses of slope stability have been made, we are not aware of any physically based model that predicts erosion at the gully head. Instead, statistical analysis is often relied upon to identify probable gully location and migration (e.g. Thompson, 1964; Donker and Domen, 1984).

Figure 7.6 suggests one other inference. Channel heads that are large steps and headcuts must have been or are sites of upslope advance of the channel tip. In contrast, gradual transitions or small steps can be stable, retreating or advancing. Whereas Figure 7.6 could be read left to right as a temporal sequence during the up-valley advance of a channel head, the reverse does not represent the down-valley migration: large headcuts do not advance downslope and become small headcuts, and so on. Large headcuts can advance upslope while becoming progressively smaller.

**Channel-head formation by overland flow**

Whether generated through Hortonian processes or soil-profile saturation, overland flow applies a shear stress field to the particles at the soil surface, and if this field is perturbed (as it can be, for example, by microtopography or by spatial variations of roughness, infiltration capacity or erodibility) it perturbs the local sediment transport field, and under conditions defined by Smith and
Bretherton (1972, p. 1519) causes incision and channel formation. A portion of the total shear stress may be expended on immobile vegetation or other roughness elements, and so the effective shear stress and transport capacity have not yet been rigorously defined for sheetflow of either constant or spatially variable depth. However, there is a general consensus from experimental plot studies of soil erosion that gradient and average flow depth (or specific discharge) are related to transport capacity.

Saturation overland flow has some characteristics unlike those of Horton overland flow, but few (e.g., Kirkby, 1987) have yet studied the geomorphological influence of these differences. For example, saturation overland flow does not usually extend to the drainage divide, and the relationship between specific discharge and distance from divide is strongly affected by along-slope profiles of elevation, soil depth and hydraulic conductivity and by the planform concavity of the source area. If exfiltration forms an important fraction of the overland flow, then the shear force of the overland flow is augmented by a lift force associated with the emergence of water from the soil (Kochel, Howard and MacLanc, 1982). However, this force is probably not significant in many saturated source areas because of the stabilizing influence of a thick root zone and because the general increase of hydraulic conductivity as the surface is approached (e.g., Wilson and Dietrich, 1987, Figure 3) acts in the manner of a reverse graded filter to diminish the near-surface seepage force in a vegetated topsoil. Another difference is that the erosive effect of saturation overland flow is likely to depend less on the stimulation of transport by raindrop impacts, which Moss, Walker and Hutka (1979) have shown to be the dominant influence on transport by Horton overland flow. Many source areas for saturation overland flow have dense groundcovers and root mats.

**Erosion by Horton overland flow**

Smith and Bretheron (1972) proposed that if overland flow discharge increases with distance from the drainage divide, and if sediment transport increases faster than linearly with downslope increasing discharge, the surface is unstable to lateral perturbations and channel-like features should initiate and grow. They also showed that for the same conditions, the resultant constant-form hillside profile should be concave. Hence, concave profiles are unstable to lateral perturbations. Conversely, the authors showed that convex constant-form profiles are stable to all lateral perturbations.

On the basis of field observations on hillslopes in southern Kenya, Dunne (1980) suggested that, whether or not the hillslope has attained constant form, the diffusive action of rainsplash might be called upon to stabilize a surface against incision by sheetwash, and that the relative intensities of diffusive rainsplash and incisive sheetwash would determine the position of the channel head. Dunne and Aubry (1986) conducted plot experiments which confirmed this hypothesis, and Kirkby (1986, 1987) modelled the formation of channel networks using approximate expressions for the competing sediment transport processes. However, in these analyses there is nothing to distinguish whether the incision simply expands to form a valley, as we suggested earlier, or whether a channel with recognizable banks is formed within that valley. In the absence of any rigorous theory for specifying the existence of banks, the following qualitative description of channel incision and bank formation by overland flow is based largely on observations made during plot experiments conducted by Dunne and Aubry (1986) or during our observations at other field sites.

As water converges into the head of an incision, the increased discharge and local gradient cause the head to migrate upslope leaving behind a trough. Downslope of the head, flow diverges so that there are zones of low depth and discharge along the margins of the trough and the discharge
per unit length of trough margin decreases (Figure 7.16), tending to stabilize the channel margin. This lateral inflow may still be perturbed and may incise the margin of the trough, but because of the low specific discharge such incision is likely to be countered by one or more factors stabilizing the margin.

If the material is cohesionless and unvegetated, the incision will cause the sidewalls to fail to a saturated stable angle. The most effective process in stabilizing the margin against further incision by runoff cascading over it is rainsplash, which erodes back the sidewall, often producing a relatively wide, shallow channel in sandy soil. The shape of the sidewall, of course, depends not only on the intensity of rainsplash but also on the activity of the new channel bed (Figure 7.17). If this base level is degrading rapidly, because the channelized flow can transport an increasing amount of sediment as distance increases along the channel, the edge of the trough will develop into a steep convexity with a straight basal slope at its saturated stable angle (Figure 7.17(b)), and the channel will remain recognizable and will enlarge. If the channel bed remains stable or rises due to aggradation in subsequent rainstorms, rainsplash will erode the channel margin into a progressively flatter, convex form (Figure 7.17(a)).

At the upper end of the incision, the gradient of the transition is lower than that of the lateral margins because of its greater specific discharge. If the sheetwash sediment transport responsible for evolution of the knick point is described by the familiar approximation (Kirkby, 1987):

$$qs = kq^2s^2$$

(7.1)

where \(qs\) and \(q\) are the specific fluxes of sediment and water and \(s\) is the local gradient, an initial scour hole in a cohesionless substrate should evolve as illustrated in Figure 7.18. In the analysis shown therein, both the base level downstream of the incision and the valley floor upstream are stable. Also, the convergence of flow from the valley floor into the channel head is ignored; its effect would be to accelerate the smoothing. Rainsplash between runoff events would also increase the smoothing of the step. The channel

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**Figure 7.16** (a) Local discharge vectors in sheetflow converging from a short unchanneled valley floor into the head of an incision 2 m wide and 2 cm deep during steady-state runoff. The dashed curves are contours of flow depth (millimeters). (b) Shows the contours of flow depth and the discharge per unit width perpendicular to the boundary of the incision. The downvalley gradient is 0.05, the runoff rate is 15 cm/h, and the hydraulic roughness is typical of flow through a sparse grass cover. The calculations were made by W. Zhang using the numerical model described by Zhang and Cundy (1989).
No rigorous, physically based calculation of the stable source-basin length has yet been published for this case of competing sediment transport processes on a cohesionless bed, although Kirkby (1986, 1987) has reported a simulation study of drainage density using generalized expressions to represent the competing sediment transport processes. Despite the lack of a rigorous formulation of flow convergence and sediment transport, however, it is reasonable to suggest that the source-basin length will be decreased by factors which intensify sheetwash (higher rainfall intensity, lower infiltration capacity, steeper gradients and more erodible soils). Conversely, the source-basin length should be increased by factors which intensify rainsplash transport without simultaneously increasing sheetwash (large raindrops for a given intensity, high infiltration capacity). Collins and Dunne (1986) reported decreases in drainage density of rills on new tephra, partially resulting from a decrease in the intensity of sheetwash relative to that of rainsplash as infiltration capacity changed during the first post-eruption year.

Gradient favors both processes, but sheetwash appears to be more sensitive to this factor (Kirkby, 1987). Increasing ground vegetation density diminishes both processes for several reasons, and without a quantitative model it is not possible to specify which effect should dominate. Field evidence from disturbed areas (e.g. Strahler, 1956; Reid, 1989) usually confirms that removing vegetation leads to increasing drainage density, but an anecdote may illustrate how the full range of possibilities remains to be explored in a quantitative, rigorous manner.

In 1976 we conducted some plot experiments on infiltration, runoff and sheetwash transport on the Athi-Kapiti Plains, 10 km south-east of Nairobi, Kenya (Dunne and Dietrich, 1980). One of our experimental sites lay on a gradient of 0.085 underlain by a clay vertic rendzina with a groundcover density of 65%. After several years of drought the soils were cracked and severely trampled and grazed by domestic stock and wildlife. There were no channels on the hillslope. Five years later, after several wet years, the site had a 100% cover of grass more than 30 cm high, but a

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**Figure 7.17** Evolution of the banks of a channel incision under the influence of linearly diffusive rainsplash and (a) a stable base level and (b) an incising channel bed. The numerals on each curve represent arbitrary time units, the absolute magnitude of which depend on the factors governing the intensity of rainsplash. The banks of the incision eventually diffuse to the divides on both sides of the valley, leaving a gently convex slope near the original incision. The computations were made with a numerical model of hillslope evolution developed and coded for a computer by L. M. Reid, University of Washington.
dense network of channels up to 10 cm deep occurred on the steep part of the slope. Our speculative interpretation of these observations is that during the wetter period, closing of cracks in the rendzina diminished the infiltration capacity (Dunne and Dietrich, 1980, p. 50) and increased runoff while the increased grass cover removed the diffusive action of rainsplash and also decreased the smoothing action of animal trampling. Even though the thicker vegetation must have decreased the sheetwash transport, the net effect of the changes seemed to be towards the incisive action of sheetwash and away from the diffusive activity of rainsplash and trampling.

Most incisions penetrate into regolith which has some cohesion, due either to electrostatic or chemical bonding of mineral grains, or is stabilized in a quasi-cohesive manner by plant roots. In these situations, the forms of the channel head and banks depend not simply on the imbalance of sediment transport capacity but also on the rate at which sediment is released from the bed and margins of the growing incision. Unfortunately, prediction of rates at which sediment is released from a cohesive boundary has proven difficult; it is not clear how to transfer laboratory results such as those of Ashida and Sawai (1976) relating shear stress and erosion rate to the field.

Where the margins of the incision are reinforced by plant roots, the sheetflow divides around plants but at many incisions is not able to scour through the root mat. A stabilized lip is formed on the upstream margin of the incision, and water cascades over it to erode a plunge pool in the unreinforced regolith beneath the roots (Figure 7.6(c)). The same type of small waterfall occurs if unvegetated regolith is extremely cohesive and water-stable. Because of the complexity of flow patterns in a plunge pool it is difficult to calculate the distribution of effective shear stress on the margin and therefore erosion. Mason (1984) surveyed the results of model studies of scour depth below steps 0.3–2.2 m high in gravels bound by mixtures of wet clay or cement. He concluded that a good estimate of the ultimate depth of scour, \( S \) (m), below the level of the tailwater was provided by

\[
S = \frac{3.27 q^{0.60} J^{0.08} L^{0.15}}{g^{0.60} d^{0.06}}
\]  

(7.2)

where \( q \) is the water flux per unit width, \( J \) is the height of the freely falling jet above the pool, \( L \) is the tailwater depth above unscoured channel-bed level, \( d \) is the mean particle size (0.001–0.028 m) and \( g \) is the gravitational acceleration (all in SI units). Mason (1989) conducted other experiments which indicated that the weak influence of the fall height was probably due to the effect of the fall on air entrainment, which in turn affected

Figure 7.18 Evolution of the head of a 1.0 m-deep incision into a 100 m-long, cohesionless valley floor with a gradient of 0.05 under the influence of a sheetwash sediment transport equation summarized in equation (7.1). The numerals on the curves are arbitrary time units. Plunge-pool erosion is not included because there is no cohesion to sustain a step. The computation was made with a numerical model developed by L. M. Reid, University of Washington.
flow velocity and boundary shear stress in the plunge pool. As the channel incision deepens, plunge-pool activity may continue to be the dominant process that mobilizes and disperses fallen material, but gravitational stresses in the regolith of the channel head and walls become more important than surface fluid stresses in controlling the migration and form of the channel head (Bradford, Piest and Spomer, 1978; Blong, Graham and Veness, 1982; Reid, 1989).

Where the control of incision is a threshold value of shear stress required to overcome a single value of resistance (at least defined as a single value of apparent cohesion for the present discussion), it seems a reasonable approximation to calculate the source-basin length simply by balancing force and resistance. Horton (1945) did this to calculate his famous $x_c$ value. Cordova, Rodriguez-Iturbe and Vaca (1983) used the same method to stabilize their modelled channel networks. Schaefer, Elifrist and Barr (1979) used it to calculate the area of a zero-order basin required to scour a channel-head in a rainstorm with a 2-year recurrence interval. This concept of a single “criticaltractive force” resisting shear at a channel head appears to have originated in agricultural engineering design and to be well suited to short-term considerations governing, for example, the generation of rills in cohesive soil during a single storm or laboratory experiment (Mizutani, 1985), or the rapid extension of a channel network (Strahler, 1956; Schaefer et al., 1979). However, over longer periods of time, weathering, bioturbation and rainsplash can reduce the cohesion of a soil at a channel head so that the position of the channel head is once again determined by a balance between the rate of transport away from the site and the rate of loosening of material. If the channel edge is vegetated this factor will also modulate the rate of loosening and of transport to the eroding face by soil creep, and will maintain a steep edge as a balance between removal and supply.

During the retreat phase, however, the transient position and form of a channel head in cohesive material seems most often to be related to the interaction of plunge-pool erosion and mass failure of the step at the channel margins. Equation (7.2) indicates that plunge-pool depth varies relatively rapidly with runoff rate, $q$ (in the field this should probably be interpreted as some measure of the frequency distribution of flows), and therefore should diminish as the headcut retreats and decreases its source-basin area. Whether the height of the step, $H_c$, increases or decreases during retreat depends upon the relative magnitudes of the upstream valley gradient and the depositional gradient established downstream of the headcut which is related to the texture and flux rate of sediment being transported downstream by the runoff rate, $q$. If the downstream gradient is less, which appears to be the more common situation in the well-defined, hydraulically smoother channel, retreat causes headcut growth to offset, at least for some distance, the effects of decreasing flow in deepening the plunge pool. In an area of southern Brazil, near the gullies shown in Figure 7.11, we have observed headcuts fed by Horton overland flow that are more than 10 m high, eroded into steep hillslopes on which a dense groundcover has been diminished by grazing, allowing incision of a narrow channel with a gradient much lower than that of the original hillslope.

The combined step height and plunge-pool depth in equation (7.2) influence the occurrence of mass failure, which at most headcuts in cohesive materials appears to be more effective than water erosion. A headcut (Figure 7.19) fails when

$$H_c = \frac{4c}{\rho_s g} \tan \left(45 + \frac{\phi}{2}\right) - y_c$$

(7.3)

where $H_c$ and $y_c$ are defined in the figure, and $c$, $\phi$, and $\rho_s$ are the cohesion, angle of internal friction, and bulk density of the substrate (Terzaghi, 1943). Cracks often develop due to both tension and drying between storms, and the plunge pool is deepened during storms, so that failure occurs during or soon after major runoff events and is partly triggered by the percolation of water into the cracks. The exact geometry might be complicated by saturation of the material, seepage erosion or tunnel scour, and in many
localities can only be analyzed through detailed field investigations. However, it is useful to note in the simple fashion summarized above that, in principle the interaction of valley and channel gradients and the plunge-pool depth can generate, in the short term, a headcut of stable height and therefore a stable source-basin area. However, again weathering will eventually render the material cohesionless, and, unless it is reinforced by vegetation, the channel head should become more gradual and should be localized by the balance of sediment release and removal. Mass failure may also be associated with seepage, which can be significant even in areas where Horton overland flow predominates, as we will discuss later.

**Erosion by saturation overland flow**

Saturation overland flow usually (though not always) occurs over a surface stabilized by a dense groundcover and root mat. If soil transmissivity or gradient decreases downslope or if the topography converges in planform, there will be some exfiltration and a tendency for seepage erosion to augment the insertion of sediment into sheetwash. Rainsplash is strongly reduced and the sheetwash would appear to have a free rein to incise. It incises only where:

1. The total shear stress of the underloaded flow is high enough to disrupt the root mat (more than 1000 dy/cm² according to the compilation of published data summarized by Reid, 1989; or
2. The total shear stress of underloaded flow is high enough to incise bare patches of soil between vegetation, especially after the cover has been thinned by drought, fire, grazing or trampling (Reid’s, 1989, compilation indicates that shear stresses of 10–500 dy/cm² would be required); or
3. Fecile sheetwash transport through the dense groundcover is able to evacuate sediment loosened by biogenic disruption or physico-chemical release; or
4. Where seepage erosion (see later) triggers the initial incision through the vegetated surface and the bare soil is then exploited by saturation overland flow.

In the third and sometimes fourth of these cases, the slow evolution of the incision allows vegetation to colonize and stabilize the channel head and banks as they form, causing them to be steeper than their analogs formed by Horton overland flow. However, the channel head is still gradual and the convex banks relatively steep (Figure 7.6(f)). In the other cases, including some seepage erosion sites, the vegetated surface is breached suddenly to form a steep discontinuity the lip of which is stabilized by the roots or other cohesive agent (Figure 7.6(g)). Plunge-pool erosion, due either to overland flow or subsurface...
flow intercepted by the step, or to seepage erosion may then cause upvalley retreat and enlargement of the step as a sharp discontinuity (Figure 7.6(h)–(j)). The banks are also abrupt and steep unless they are degraded by animal trampling or tree throw which are sometimes concentrated in the wet areas that generate saturation overland flow.

The source–basin length required for channel initiation by saturation overland flow depends not only on the relative importance of surface shear and the seepage force of exfiltration but also on the spatial distribution of the overland flow. Several numerical models describe the generation of saturation overland flow in realistic field situations (e.g. Wilson, Dietrich and Narasimham, 1990), but they are typically too complex for use in a simplified theoretical discussion of this type. For the present purpose, the important controls on the downslope pattern of saturation overland flow can be analyzed by using a very simple model of subsurface flow in a wet soil, ignoring storage changes in the unsaturated zone.

In Figure 7.20(a) the impermeable bedrock has an elevation profile

$$y = y_0 e^{-kx}$$  \hspace{1cm} (7.4)

and is covered with a soil of constant depth ($H$) and saturated hydraulic conductivity ($K$). The elevation of the water table ($z$) = $y + h$, where $h$ is the thickness of saturated soil. Despite its well-known limitations, we assume for purposes of illustration a Dupuit–Forchheimer flow condition, so that the subsurface flow per unit width is

$$q_{sub} = -Kh \frac{dz}{dx} \hspace{1cm} (7.5)$$

Under a long-continued steady rainfall, $I$ (analogous to the average rainfall intensity over a wet season), the saturated soil thickness increases downslope and the water table intersects the surface at some distance, $x_s$, downslope of which saturation overland flow is generated. The saturated thickness at and downslope of $x_s$, is $H$, so that

$$q_{sub} = Ix_s = -KH \frac{dz}{dx} \hspace{1cm} (7.6)$$

and $x_s$ can be found by substitution into this equation which can then be solved iteratively to define the extent of the hillslope subject to overland flow. The steady-state discharge of overland flow ($q_{surf}$) is obtained through the summation of the rainfall intensity onto the saturated area and the downslope integral of the divergence of the

![Figure 7.20(a) Subsurface flow ($q_{sub}$), exfiltration ($q_{ex}$), and saturation overland flow ($q_{surf}$) on a concave hillslope during a long, steady rainfall. The soil properties are constant along the hillslope. (b) and (c) are graphs, respectively, of the alongslope variation of subsurface and surface flow for the following conditions (defined in the text): $K = 0.5 \text{ m/h}, H = 1.0 \text{ m}, I = 0.001 \text{ m/h}, k = 0.02, y_0 = 30 \text{ m}$. The width of the belt of saturation overland flow in this example is 28.4 m.]

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The Channel Head
subsurface flux, which is decreasing downslope as the gradient declines:

\[ q_{\text{surf}} = I \int_{x}^{x_0} dx + \int_{x}^{x_0} \left[ -KH \frac{dx}{dx} \right] dx \quad (7.7) \]

\[ = I(x - x_0) + KH \gamma_0 \left( e^{-kx} - e^{-kx_0} \right) \quad (7.8) \]

Typical results are shown in Figure 7.20(b) and (c). The solution for the unsteady case requires numerical computations. A similar analysis could have incorporated the effects of soil depth, hydraulic conductivity or planform concavity parameterized in terms of distance along the slope. This analysis differs from water-storage models of saturation overland flow (e.g. Kirkby, 1987), which do not include exfiltration.

Whatever the details of the spatial dependence of saturation overland flow, its specific discharge affects whether the critical shear stress for the vegetation and its root mat is exceeded. If incision occurs and the edge of the incision is reinforced by the root mat, the discharge affects the location and form of the channel head through its control of plunge-pool erosion as described above in relation to equation (7.2). Even after the step has become stabilized, the pattern of discharge also controls the surface shear stress on the upstream valley floor. The shear stress is mainly expended as form drag on vegetation in saturated zones. However, the residual effective shear stress is capable of weak sediment transport and therefore of affecting the balance of competing transport processes governing the location of channel heads, as described in the previous section. Its role is likely to be augmented by exfiltrating water in those localities where return flow is relatively abundant, although this effect as well as the simpler case of sediment transport through dense ground vegetation has received little study.

Channel-head formation by seepage erosion

Seepage erosion is the entrainment of soil or rock resulting from water flowing through and emerging from a porous medium. It requires the development of a critical body force or drag force which entrains particles individually or in bulk through liquefaction or Coulomb failure. Dunne (1990) reviewed studies of the mechanics of seepage erosion, and outlined the hydrological conditions under which it occurs and erodes channels. The fundamental requirement for seepage erosion of cohesionless materials, outlined by Kochel, Howard and MacLanc (1982) and by Iverson and Major (1986), is that in the exfiltrating subsurface flow there must be a hydraulic gradient of sufficient magnitude and direction relative to the angle of the hillslope and the density and frictional resistance of the material. On the other hand, cohesive materials may succumb to Coulomb failure as seepage emerges, but are not released as individual particles nor by static liquefaction of the failed mass. Instead, they stand in headcuts and fail in slabs unless the cohesive surface is being continually rendered cohesionless by seepage weathering (Dunne, 1990, Equation 9 and discussion). Plant roots can resist separation of material in a manner similar to interparticle cohesion and therefore can also reinforce material against seepage erosion. These considerations are important to the stabilization of channel heads against erosion by exfiltrating water.

Perturbation of the flow field of the emerging seepage causes local increases in flow rate and therefore hydraulic gradient. This convergence triggers seepage erosion and channel initiation, which for some time acts as a positive feedback on channel extension by increasing convergence of the subsurface flow field on the channel head. The channel head retreats up-valley as seepage erosion undermines the headcut and triggers toppling, slab failure and water erosion of the steepened margins.

Dunne (1980) suggested that the drainage network is eventually stabilized with a predictable drainage density because the reduction of the source-basin area which results from headward erosion decreases the supply of subsurface flow and therefore the hydraulic gradient until it can no longer trigger seepage erosion. Dunne (1990) used the Dupuit-Forschheimer flow approximation to derive an expression for the hydraulic
gradient in a conical source basin up-valley of a channel head. Unfortunately, this approximation is invalid close to the eroding face, but in this region there is also a daunting amount of complexity due to soil structure and non-Darcian flow mechanics. Despite its limitations close to the face, the expression is useful for summarizing the interaction of the hydrological variables that control the overall hydraulic gradient near the eroding face and therefore the minimum source–basin area or length required for seepage erosion. Figure 7.21 illustrates the water table beneath a conical source–basin recharged at an average rainfall rate, \( I \). The hydraulic gradient, \( \frac{dz}{dr} \), at the face is related to its controls by the equation:

\[
\left. \frac{dz}{dr} \right|_{\text{face}} = \frac{I}{2K} \left[ \frac{1}{r_s(z_s - r_s \tan \beta)} \right] - \frac{I}{2K} \left[ \frac{r_s}{z_s - r_s \tan \beta} \right]
\]

where the geometric terms are defined in Figure 7.21. The height of the seepage face, \( z_s \), can be estimated from field observations or from a calculation of the subsurface discharge from the drainage area (Dunne, 1990).

If this value for the horizontally directed hydraulic gradient is set equal to the value required for Coulomb failure of a cohesionless material (Iverson and Major, 1986) or for the lifting of individual particles (Kochel et al., 1982), one can obtain the stable source–basin length (\( r_D - r_s \)) with respect to seepage erosion:

\[
r_D^2 - r_s^2 = \frac{2K}{I} \left( r_s z_s - r_s^2 \tan \beta \right) \left. \frac{dz}{dr} \right|_{\text{crit}}
\]

Although the equation cannot be solved explicitly for source–basin length, unless an independent means is available for specifying \( z_s \) and \( r_s \), it does allow one to examine the influence of several independent variables on source–basin length. The results predict that distance from the divide to the channel head increases with increasing hydraulic conductivity, gentler slope of the underlying aquiclude or less intense recharge rates, and

are thus in general qualitative agreement with the field observations by Montgomery and Dietrich (Figure 13) where seepage erosion is active. The analysis can be generalized for other geometric approximations of source basins, such as the tipped triangular trough employed by Dietrich, Wilson and Reneau (1986). The exclusion of an upward-directed flow component from the Dupuit–Forchheimer analysis precludes the prediction of bulk liquefaction, although at real channel heads upward flow components can be estimated by field measurements, numerical solution of the groundwater flow equations, or graphical flow-net construction.

This setting of the drainage density by balancing the seepage erosion force and material resistance is seen at its simplest on permeable, cohesionless substrates in climatic regions (e.g. snowy regions) where the water influx is relatively steady and sustained for long periods. If the channel heads have stabilized with respect to their drainage areas, they are usually gradual (Figure 7.6(f )) or consist of steps reinforced by vegetation (Figure 7.6(g)). Their lateral margins are degraded by rainsplash or soil creep, and sediment-free water exfiltrates from them slowly, often through a vegetation mat which reinforces the soil surface and also acts as a reverse graded filter. Seepage-
eroded channel heads that have not yet stabilized
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with their source-basin areas are often undermined (sapped) aggressively into a headcut and vertical banks (Figures 7.6(h) and (i)). This is also true in regions where long-term climatic fluctuations cause the alternation of channel filling and then headward extension by seepage erosion.

Some regolith is sufficiently cohesive that seepage erosion cannot occur (Dunne, 1990, Equation 9a) except in material rendered cohesionless by weathering and cracking immediately around the channel head. Thus, the source-basin length in these materials approaches its stable length at a rate determined by the rate at which weathering can render the regolith mobile in the seepage forces imposed on the channel head by subsurface flow from its drainage area. In bedrock, this rate may be too slow for equilibration to occur (Figure 7.6(j)).

Howard and McLane (1988) proposed a seepage-erosion model that incorporated three interrelated zones of erosion which they had observed in laboratory experiments on unidirectional seepage (Figure 7.22). Erosion in the downstream zone was controlled by fluvial sediment transport, and was predicted with a conventional bedload equation. The transport rate in the channel set the rate of erosion in the two zones upslope. Seepage erosion was most intense at the sapping face, where both individual grain motion and intermittent shallow mass wasting occurred. In the mathematical model seepage erosion took place when this face attained a critical angle, predicted by analyzing the stability of the cohesionless material with respect to seepage. Transport rate was assumed to vary exponentially with gradient steeper than this calculated critical angle. The steepest, uppermost zone was assumed to erode by mass wasting at a rate set by the sapping zone and to retain a constant gradient. Theoretical prediction of observed channel-head migration and channel gradient in the experimental chamber shown in Figure 7.22 gave favorable results, although, to some degree, the morphology and migration rate were imposed rather than determined with the theory. Since the model is one-dimensional, it does not predict the width of the evolving feature.

Howard (1988), using a different approach, subsumed these modeled processes into a simple expression which stated that erosion at the seepage face was proportional to the water flux rate per unit width minus some critical value, and he used this assumption to predict valley development in cohesionless material. Flow was two-dimensional in plan view, steady, saturated, and analyzed using the Dupuit approximation. Perturbation of the flow field was caused by a downslope cut face and a random variation in saturated conductivity. Model results predicted headward-advancing, narrow valleys separated at roughly regular intervals. In this latter model the distinction between valley and channel is blurred.

The detailed laboratory and theoretical studies of seepage erosion have not yet been matched with equivalent studies in the field. Of course, the problems quickly compound. In the field, subtle effects of material strength, conductivity fields and upslope propagating base level changes (to name a few) must be confronted (e.g. Laity and Malin, 1985). Also, none of the available theories

![Figure 7.22](image-url)
can be used to predict downslope retreat of the channel head, or the size of the channel head.

**Channel-head formation by mass failure**

Even where seepage erosion does not cause bulk liquefaction or the removal of individual particles, subsurface flow may cause the extension of channels through landsliding. The evidence is described by Woodruff (1971, pp. 403–4) and by Dietrich and Dunne (1978), who proposed that on steep slopes the axes of unchanneled valleys (called hollows) experience a self-driven cycle of instability that causes periodic excavation of the bedrock. The convergent topography causes colluvium to be transported to the hollow where it accumulates until reaching a critical state controlled by storm-generated high pore pressures and the strength of colluvium. They proposed a critical depth of failure in cohesionless gravelly colluvium would be controlled by root strength. Okunishi and Iida (1981) also argued for development of a critical depth controlling the timing of mass failure. Reneau and Dietrich (1987) argued that lateral root strength controls the size of shallow colluvial landslides in hollows and that not just a critical depth, but a critical size, sufficient to overcome lateral root strength and basal friction, was required for instability. Once failure occurs erosion of remaining colluvium may continue through gully development and subsequent failures. Either through revegetation or accumulation of a mantle of gravel stable against overland flow, the scar stabilizes and becomes a site of net deposition until reaching critical size again. Frequency of bedrock exposure and recurrence interval of failure determined from radiocarbon dating of charcoal buried in the colluvium have been documented by Reneau (1988) and Reneau et al. (1986, 1989, 1990).

Based on these observations, Dietrich et al. (1986) and Montgomery and Dietrich (1989) proposed a theory that predicts the source-basin area and drainage density for a landscape driven by landsliding from bedrock hollows. In the model, pore pressures are estimated from an analytical subsurface flow model derived from work by Humphrey (1982) and Iida (1984), and are used to predict the requirements for Coulomb failure of a soil of constant thickness and slope. Hence, the source–basin area per unit of contour width was shown to be

\[
A = \frac{\rho_b K \sin \beta \cos \beta}{\rho_w} \left[ 1 - \frac{\tan \beta}{\tan \phi} \right]
\]

where \( \rho_b \) is the bulk density of soil, \( \rho_w \) is the density of water, \( \phi \) is the angle of the soil's internal friction and the other variables have been defined previously. The predicted relationship between drainage area to the channel head and local valley gradient is similar to that found in the field (Figure 7.23). The mechanism behind this relationship is simple: the steeper the slope for a given colluvium depth and angle of internal friction, the lower is the pore pressure required to cause landsliding. Thus, the smaller is the subsurface flow and the drainage area required to cause the scouring of a channel incision.

By using the result that drainage density is roughly equal to the inverse of the length of the source area, the slope-stability based theory can
be used to predict the variation of drainage density \((D)\) with major hydrologic and slope-strength variables:

\[
D = \frac{w_{nb} w H}{w p_a KH \sin \beta \cos \beta} \left[ 1 - \frac{\tan \beta}{\tan \phi} \right]^{-1}
\]  

(7.12)

where \(w_b\) is the source-basin width and \(w\) is either the channel width or the hollow width at the channel head. This model predicts an increase in drainage density with decreasing saturated hydraulic conductivity and frictional strength of the soil, and with increasing precipitation, local valley gradient at the channel head, and width of the source area relative to the width of the depositional zone or channel head. In some regions the development of pipes or tunnels may disturb the subsurface flow field (Pierson, 1983) sufficiently to complicate simple theoretical relationships of the kind proposed by Dietrich et al. (1986), and much work remains to be done in generating a theory of pipe-network distribution and its hydrological relationship to regularities of channel-head distribution, if any, in those regions where pipes seem to be important.

Temporal fluctuations in the position of the channel head also occur during cycles of scouring and colluviation. During the period of erosion of the hollow the channel head may extend well upslope into the valley. Figure 7.14 illustrates how the channel head position may vary with time, ranging from staying at a constant distance from the divide over long periods of time to constantly advancing or retreating along the valley. Landslides in hollows upslope of the channel head do not necessarily cause the channel head, as we have defined it here, to advance completely upslope. For example, the channel head has extended to the base of the upper slide scar but not to the crown scarp. This distinction between the crown scarp and the channel head is consistent with our definition of the latter as a feature with definable banks which contain a concentration of water and sediment transport. The concentration causes the shallow incision in the landslide scarp. Hence there can be regions upslope of the furthest extent of the channel that remain valleys by periodic landsliding. Although there is clear evidence for an inherent cycle of instability in hollows on steep slopes, radiocarbon dates of basal colluvium (see Reneau references above) indicate that evacuation of colluvium and associated upslope advance of the channel head is also climatically influenced.

**Channel-head formation by tunnel scour**

In some locations a significant proportion of subsurface flow travels through cracks of various origin, large root holes, animal burrows and other tortuous conduits. Flow along such pathways requires saturation at some shallow depth within the soil or Horton overland flow which can enter a conduit intersecting the soil surface (Figure 7.15). Thus flow in conduits is favored by low hydraulic conductivity in the matrix and high rainfall intensities as well as by processes which generate the cracks themselves. The principles governing the entry of water into such conduits or “macropores” have been reviewed by Beven and Germann (1982) and Dunne (1990), and Jones (1981, 1982) has reviewed field measurements of this runoff process. Tanaka, Yasuhara and Marui (1982) and Tsukamoto, Ofita and Noguchi (1982) have also published cursory reports of the phenomenon during large storms on steep, forested hillsides in Japan. Yair et al. (1980) and Bryan and Harvey (1985) have documented flow and sediment discharge from tunnels in arid, clay badlands.

Conduit flow is responsible for several erosion processes that eventually produce channels at the surface. For example, Pierson (1983) demonstrated with a laboratory model that the presence of a conduit disturbs the flow field in shallow groundwater on steep, colluvium-mantled hillsides, causing high pore pressures that can trigger narrow debris avalanches. Several field observers have interpreted the occurrence of debris avalanches to be caused by blockages of root conduits resulting in high pore pressures.

In other localities the erosion results from the application of a shear stress to the margins of the conduit and from consequent collapse of the walls.
and roof. Dunne (1990) proposed the term "tunnel scour" to emphasize the difference between this process and other forms of subterranean erosion, and he reviewed the mechanics of such erosion. When the flow becomes non-Darcian as it enters macropores, it accelerates, especially where hydraulic gradients are steep in mountainous terrain or perpendicular to high channel banks or heads. The important characteristics governing erosion are then the flow velocity and discharge rather than the hydraulic gradient. Because of the geometrical complexity of macropores no measurements or calculations have been made of the magnitude of the fluid shear on their margins, but sediment concentrations in the effluent indicates that some combination of shear and collapse mobilizes considerable volumes of soil particles (Jones, 1982; Tanaka et al., 1982), and enlarges the conduits (Tsukamoto et al., 1982, p. 95).

Where soils are well-aggregated, organic-rich and reinforced by roots the rate of tunnel scour appears to be relatively slow and in most cases to be balanced, at least over the ensemble of conduits in a colluvial mass, by collapse and bioturbation which tend to fill any large space in the soil. Thus, it is not clear whether a high proportion of conduits evolve into stream channels, although conduits certainly converge on the heads and banks of many channels (Jones, 1971). However, certain geologic and climatic conditions favor tunnel scour which forms intricate subterranean networks and, after their collapse, spectacular channel systems.

The critical shear stress of the boundary material may be reduced almost to zero in the presence of soluble cements or of dispersible clay minerals (Holmgren and Flanagan, 1977). Such minerals are more likely to form in subhumid climates, or in certain lithologies such as marine silts and clays or glass-rich volcanic rocks in a range of climates. Many authors have noted, for example, the association between tunnel scour and high concentrations of salts (especially sodium) in the soil (e.g. Heede, 1971; Stocking, 1971; Bryan, Yair and Hodges 1978; Imeson, Kwaad and Verstraten, 1982). The salinity causes both shrinkage cracks for the initial non-Darcian flow, if it is associated with clay minerals such as sodium-montmorillonite, and the dispersive behavior of the material. The rapidity of tunnel scour in such material, the strength of the cohesive, unweathered material and the penetration of scour below the depth of bioturbation allow the formation and survival of tunnel networks (Parker, 1963; Parker, Shown and Ratzlaff, 1964; Heede, 1971).

Enlargement of the tunnels and especially thinning of the roof eventually causes lines of pits and their coalescence into stream channels (Figure 7.5). The channels have certain distinctive features while they are being extended. The channel heads are usually vertical and the banks are steep with sharp edges and some overhangs (Figure 7.9). The channels have low gradients and tend to be straight for long reaches between sharp bends, probably localized initially by major fractures. Steps occur along some channels where bedding planes localize changes in resistance, and plunge-pool erosion occurs at these steps and at the main channel head. Hanging tributaries, sinkholes and natural bridges occur in some rapidly eroding channel networks, and the position and form of the channel head may change drastically during a few rain storms (Heede, 1974, p. 265). If the vicinity of the channel head develops sufficient scale and complexity, the topography is referred to as pseudokarst (Parker et al., 1964). The scale of headcuts, sidewalls and related features in such terrain depends not only on the depth of incision but also on the considerable cohesive strength of the undispersed, unweathered sedimentary rock.

Flow reaches the channel head through tunnels and sometimes over the surface as indicated in Figure 7.15. Sinkholes and subhorizontal tunnels form up-valley of the channel heads, and the subterranean catchment areas of neighboring heads may overlap. Channel networks formed recently by tunnel scour are widely spaced, but since the substrates are soluble or dispersible, the flow velocity, discharge, and therefore drainage area required for extending each channel tip is often very small. Thus, channel formation by tunnel erosion and collapse quickly consumes almost the entire geological unit in which it occurs, creating badlands with extremely high
drainage densities and low source-basin areas (Figure 7.24). Most studies of such channels have been carried out in geologically young terrains (Bryan and Yair, 1982; Yair et al., 1980; Löfler, 1977) or in landscapes recently disturbed by changes of base level, climate, or land use (Brown, 1961; Buckham and Cockfield, 1950; Rubey, 1928).

**GENERAL THEORETICAL ANALYSIS**

No theoretical investigation has yet examined the problem of channel-head size and shape, so we focus here on the issue of location. However, refinement of the location theories probably requires inclusion of channel-head scale because of feedback effects of channel-head form on the mechanics of the processes determining channel-head location and its temporal variations.

There are two basic views of the problem of predicting channel-head location. The first, derived from the conservation of sediment mass, emphasizes that the channel tip represents the point where sediment transport increases faster than linearly downslope. This usually requires that some kind of wash process dominate the sediment transport, although, in principle, the same effect might be achieved by some process such as creep/earthflow with the material weakening and becoming less viscous during strain along flow lines. The second view is that in many localities the channel head is not simply a point where the effect of one process dominates over that of another but instead is a point at which processes
not acting upslope become important. A simple example is a gully head above which transport occurs by creep, splash and biogenic disturbance, while erosion at the head is dominated by seepage, plunge-pool activity and slope collapse. In this case, the balance of sediment still determines whether the channel tip stabilizes at a point or migrates, but some factor provides the resistance (at least temporarily) to limit the domain of one process or another. We will summarize these two approaches briefly, and suggest how they are complementary but useful in different situations.

The sediment-transport-balance view of channel initiation emerged from the seminal work of Smith and Bretherton (1972), who examined how a long ridge, being lowered at a constant rate, would respond to lateral perturbations of the sediment transport field. They specifically examined the perturbation due to microtopography, although the disturbance could have been provided by any of the other hydrologic, hydraulic or material properties affecting capacity-limited sediment transport. Smith and Bretherton posited a sediment transport equation of the form:

\[ q_s = F[q,s] \]  
(7.13)

where \( q \) and \( q_s \) are, respectively, the volumetric fluxes of water and sediment per unit width of hillslope, \( s \) is the local gradient and \( q \) is assumed to be proportional to distance from the divide. They then showed that on a constant-form hillslope on which the sediment transport rate is \( F_0 \), the profile would be concave if

\[ F_0 < q_s \frac{\partial F_0}{\partial q_s} \]  
(7.14)

and convex if the inequality were reversed. If a particular sediment transport equation is chosen:

\[ q_s = kq^n s^m \]  
(7.15)

then equation (7.14) holds if \( n > 1 \), although parts of non-constant-form profiles evolving toward their final forms could be convex, resulting in an inflection point. Lateral topographic perturbations of the constant-form concave surface were found to be unstable, and Smith and Bretherton interpreted this to mean that channels would form on these surfaces. In these zones, combination of equations (7.14) and (7.15) yields

\[ \frac{\partial q_s}{\partial q} > \frac{q_s}{q} \]  
(7.16)

On convex constant-form profiles, all perturbations were damped out.

Smith and Bretherton (1972, p. 1520) reasoned that landscapes were composed of convex hillslope profiles, stable to lateral perturbations bounded by concave surfaces, unstable to these perturbations, and hence channelized. Channel heads would be located at inflections in the profile. The authors proposed a particular sediment transport law which would produce the requisite convexo-concave form:

\[ q_s = kq^n s^m + as \quad n > 1 \quad a > 0 \text{, constant} \]  
(7.17)

They did not formalize a perturbation analysis of equation (7.17) but reasoned qualitatively that the concave profile would tend to be channeled and the upper convex profile not, according to their criterion in equation (7.14). Hence, they reasoned that the length of the convex portion of the ridge profile is equivalent to Horton's \( x_c \), although erosion does occur upslope of that point (unlike in Horton's model) but predominantly by diffusive processes. Smith and Bretherton considered their analysis as providing a channel-formation mechanism, although it must be stressed that their "channels" are only infinitesimal perturbations on an otherwise smooth surface, rather than channels of finite dimensions. Their analysis is most appropriately applied to rills on smooth slopes of badland landscapes, as they noted in discussion of their theory.

Kirkby (1980, 1986, 1987) rewrote equation (7.16) assuming drainage area per unit contour width, \( a \), can be substituted for surface runoff, i.e.

\[ \frac{\partial q_s}{\partial a} > \frac{q_s}{a} \]  
(7.18)
and proposed models driven by valley network growth in which he argued that the channel head was located either at the transition between convex and concave slopes, or at the up-slope-most point where equation (7.18) held. His numerical model was not limited to equilibrium slope profiles, and valleys were large, rather than the infinitely small perturbations analyzed by Smith and Bretherton. Kirkby also formulated a transport law to represent erosion and sediment transport by landslides and predicted the consequent equilibrium longitudinal profile of the slope. In this case he simply assumed that the inflection from convex to concave profile was the channel-head location. Most of the channel heads we have mapped are located well downslope from the divide in distinct valleys and not at inflection points from convex to concave profiles. Hence the attempt to relate predictions from the Kirkby model to field observations regarding contributing area and local slope at channel heads must still be considered rather tenuous.

In order to test the Smith and Bretherton, and Kirkby criteria for channel-head formation, it would be necessary to establish first that a ridge was indeed eroding more or less uniformly. Then, since the theory predicts that the profile inflection and the head of incision should coincide, it would be necessary to map these features in the bedrock axes of hollows, rather than mapping channel-head locations in colluvial fills. Løwenshultz (1991a,b) has extended the work of Smith and Bretherton and of Kirkby by performing a formal stability analysis of initiation of drainage incision which identifies for the first time what might dictate wavelength selection (i.e. lateral spacing between channels). Among other findings, her theory suggests that channel heads may extend upslope of the convex to concave profile inflection and that initial short wavelength instabilities would tend to be replaced with slower growing long wavelength ones. Field work confirmed aspects of her theory by finding rills to be present on linear rather than convex portions of hillslopes and by showing the existence of multiple scales of incision.

Without question, net erosion will occur if across a channel head there is a rapid increase in net downslope sediment transport not compensated by sediment influx across the channel walls from adjacent hillslopes. This should tend to deepen and cause upslope advance of the channel head. The channel head, however, is only found at the point defined by equation (7.14) or (7.18) in the case where it is actively advancing upslope or the channel is deepening. Not all channel tips are growing upslope, nor are all channels downstream of the channel tip incising relative to the divides. Channels exist on aggrading surfaces, such as valley fills and as such their upslope tips cannot be predicted by using equations (7.14) or (7.18) as a criterion. Another criterion is necessary to specify the location of the channel head when it is migrating downvalley as illustrated in Figure 7.14(d), and although this criterion would coincide with the implication of equation (7.14) during up-valley migration or stasis, that criterion must always be the dominant reason why one can recognize a definite channel head and banks within a swale or hollow. By itself, equation (7.14) is a criterion for a valley rather than for a channel.

A mass-balance approach for transport-limited situations is obviously the simplest case to study first in modeling landscape evolution, but it is not sufficient to specify the hillslope–channel boundary. Incision, even if balanced by a diffusive process, will spread and generate valleys and swales. It will not differentiate channel-scale features with banks within those valleys, except in the case of competing rills. As discussed in the sections on processes, in order for a channel head and bank to develop in homogeneous material it is necessary that there be some effect which lowers the transport rate below the capacity value, stopping or slowing the upslope advance of the incision, and allowing a steepened slope to develop as the head and bank. The propagation of the incision is halted and therefore a bank or head is stabilized if removal of sediment from it is just balanced by the transport of sediment towards the edge of the incision. In bedrock the transport limitation is controlled by weathering rates. In a cohesive soil, the steepened slope is maintained by the resist-
ance of the interparticle forces to wash, rainsplash, soil creep or collapse, and may also be partly related to weathering. If the edges of the incision are vegetated, plant roots provide the resistance to these diffusive processes. In laboratory experiments with sand, even tension in pore water may temporarily stabilize a channel bank and head. If the incision becomes deep enough, the stability of the incision must be maintained by resistance to sliding or toppling failures. In the long-term, however, if weathering can reduce all of these earth materials to cohesionless regolith, only the resistance of vegetation remains available to halt the diffusion of an incision and stabilize a channel. Of course, some long-continued weathering can also produce cohesive regolith which can accomplish the same result.

The role of resistance to erosion has led some geomorphologists to emphasize that beyond some critical domain a channel-forming process suddenly emerges. Horton’s (1945) theory of channel incision was the first of this kind. In it, initiation occurs where the basal shear stress in overland flow exceeds the critical tractive force of soil. Thence, deeper parts of the flow exert greater shear stresses and therefore greater rates of downcutting, allowing cross-grading and micro piracy to regrade the surface and focus flow into a smaller number of channels until the hillslopes are everywhere equal to or shorter than \( x_c \), the critical distance beyond which overland flow is erosive. The model of network growth published by Cordova, Rodriguez-Iturbe and Vaca (1983), and the calculation of critical boundary shear stress for zero-order drainage-basin areas by Schaeffer, Elifrits and Barr (1979) are based on Horton’s concept. Although it may be convenient to think of this type of channel-forming process in terms of critical tractive force for the short time spans and for the design purposes required by Schaeffer et al. (1979), the continuity-based approach is more appropriate for examining the long-term evolution of channel heads where diffusion and reduction of the shear resistance by weathering are possible.

On steep, permeable hillslopes where overland flow does not occur, and possibly on gentle slopes with saturation overland flow but very resistant surfaces, Dietrich, Wilson and Renaeu (1986) showed that the sudden down-valley transition to channel erosion may be localized by landsliding. Field observations of seepage erosion by Parker (1963), Dunne (1980), and Laity and Malin (1985), and laboratory studies by Howard (1988) examined another example of a channel-forming process which requires a critical condition for the erosion. Studies of tunnel erosion by Parker (1963), Heede (1971) and others provide yet another example, although, as discussed in the section on processes, the critical resistance to tunnel erosion in some dispersible materials may be vanishingly small.

Other strong field evidence for a threshold-based channel initiation theory is the topographic analysis performed by Montgomery and Dietrich (1992). They used a digital terrain model to define the upslope drainage area and local ground slope for roughly 200 \( m^2 \) elements of the western 1.2 \( km^2 \) basin shown in Figure 7.1, a site where the channel network had been carefully mapped in the field. Channel heads were found to lie where \( 200 > (A/b)s^2 > 25 \) m, in which \( A/b \) is the drainage area, \( A \), per contour length, \( b \), and \( s \) is the local ground slope. The analysis showed that the entire channel network and nearly all of the unchanneled valley floors lie either within or above the channel-head threshold range. Hillslopes bordering the channels or lying at the heads of unchanneled valleys were consistently just below the topographic threshold defined by \( (A/b)s^2 > 25 \) m. All ridges were well below this threshold. Montgomery and Dietrich interpreted their results as evidence that the topographic conditions setting the threshold to channelization also define the limit of valley dissection and consequently, the hillslope length.

Under such conditions a channel-initiation criterion more appropriate than equation (7.14) involves a resistance condition. The relevant transport laws may thus be

\[
q_s = ks
\]  

for \( x < x_c \) or \( a/s < (a/s)_c \) and

\[
q_s = F(i, u, (\tau_s - \tau_c)) + G(q, s) + ks
\]  

for

\[
qs = \begin{cases} 
F(i, u, (\tau_s - \tau_c)) + G(q, s) + ks & \text{for } x < x_c \text{ or } a/s < (a/s)_c \\
ks & \text{for } x > x_c \text{ or } a/s > (a/s)_c 
\end{cases}
\]
for \( x > x_c \) or \( a/s > (a/s)_c \). This statement argues that there is a critical point along the valley where the resistance to channelization is overcome due to high hydraulic head gradient \((\theta)\) associated with seepage, or to high pore pressure \((u)\) causing landsliding, or due to boundary shear stress of overland flow \((\tau_b)\) greater than a critical value \((\tau_c)\) for the surface. At this point and further downslope, surface runoff becomes an important transport agent (hence the inclusion of \(G(q,s)\)).

Under these circumstances, the channel head is a point where processes change rather than where the dominance of one form of transport is replaced by the dominance of a competing form. This concept is consistent with the inference that in grasslands channels begin where some disturbance such as animal trampling or fire, allows overland flow to become sufficiently erosive that it can cut through an otherwise resistant, vegetated, cohesive soil surface (e.g. Leopold, Wolman and Miller, 1964; Reid, 1989). It should still be possible for a given set of soil properties and climate to relate mechanistically the threshold controls in equation (7.20) to critical drainage area per unit contour width, and local slope for stable channel networks.

This second approach, emphasizing the onset of a different sediment transport process at the channel head and the domain of channel erosion limited by one or other of the resistances implied in equation (7.20), is relevant to some channel tips that are stable and others that are extending aggressively. For example, channel-head retreat by seepage erosion may result in a stable drainage density where, under the given conditions of water input, hydraulic conductivity and slope, the source–basin areas are just sufficient to generate the subsurface flow to impose a hydraulic gradient which will equal that required for seepage erosion at the channel head (Dunne, 1990). The drainage network expands until competition for groundwater between channels everywhere lowers hillside slope and source–basin lengths (or more exactly, groundwater flow path lengths) below a critical value, as Horton (1945) had argued for overland flow. In regions such as the canyonlands of the Colorado Plateau, the drainage density is far from stability, and the channels begin at large headcuts of the type indicated in Figure 7.6(j). Seepage erosion of cliffs occurs as soon as weathering lowers the strength of the bedrock (Laity and Malin, 1985).

Other channel heads localized by resistance migrate up-valley if the disturbing factor (such as pore pressure) increases, and down-valley between wet periods as colluvium diffuses into the incision. In the case of channel extension by landsliding, studied by Dietrich et al. (1986, 1987), the up-valley migration is constrained where the shear strength of the colluvium is reduced sufficiently by the drainage-area-dependent pore pressure.

As the channel head migrates downvalley, its location is determined by the balance between collapse and ravelling of colluvium from the margins of the incision, and the capacity of the channel flow to transport sediment down-valley. However, this transport capacity is strongly reduced by vegetation. For example, in the channels evacuated by debris avalanches in bedrock hollows studied by Dietrich and Dunne (1978) any sediment that tumbles onto the smooth bedrock in the evacuated hollow is quickly washed away by even small flows. Only cobbles or boulders remain in the scar. In the absence of an agent to slow diffusion, collapse and soil creep would cause the side of the incision to expand to the divide. However, vegetation usually recolonizes the collapsed edge of the scar and begins to slow this process. Meanwhile, shrubs grow out from the margins of the incision, trapping sediment in their roots and colonizing the thickening colluvial wedge. Over hundreds to thousands of years the dominant runoff process during large storms evolves from Horton overland flow to saturation overland to subsurface flow with obvious implications for subsequent failures. However, because of the difficulties of incorporating the effects of vegetation into erosion models it is not yet possible to predict the location of these channel heads during down-valley migration.

Reference to the role of vegetation or other resistive agent brings up the issue of the scale, especially the width of the channel at its head. At
many channel heads localized by catastrophic processes such as landsliding, seepage erosion or plunge-pool erosion, the channel immediately downstream of the head is often narrower than the head itself. Many spring heads, for example, are alcoves, much wider than the channel. Laboratory experiments on seepage erosion show the same result (Howard, 1988). The channel width required to convey the subsurface flux of moisture which caused the mass failure or the surface flow over the hydraulically rough swale is much smaller in the smoother channel than above the channel head.

Willgoose (1989) and colleagues have formulated a model of channel network growth and hillslope evolution which explicitly incorporates their interaction. The model differs from that of Kirkby (1986) in differentiating between sediment transport in channels and on hillslopes, but the equations used for the distinction are based only loosely on qualitative physical arguments. The work has been extensively described by Willgoose, Bras, and Rodriguez-Iturbe (1991a, b, c).

The model incorporates three types of sediment transport equations. The first of these describes linear diffusion of sediment by processes such as soil creep, rainsplash and landsliding, and is represented by

\[ q_s = \alpha_1 \cdot s \]  

(7.21)

where \( \alpha_1 \) is a diffusion coefficient that characterizes the particular transport process. Simultaneously acting on all parts of the catchment, soilwash and rillwash are represented by

\[ q_s = \beta_1 \cdot q^{M+N} \quad m,n > 1.0 \]  

(7.22)

This process overwhelms the diffusive component of transport wherever the runoff rate (i.e. drainage area) is sufficiently high, given the chosen value of \( \alpha_1 \) and \( \beta_1 \) in equations (7.21) and (7.22), but the rills are not allowed to grow into channels. Instead, the model is informed of where it should impose the third sediment transport equation, which describes channelized transport. The equation for this process is

\[ q_s = \beta_2 \cdot q^m s^n \quad m,n > 1 \]  

(7.23)

where \( \beta_2 \) can be set at some fixed multiple of \( \beta_1 \) in equation (7.22) to ensure that for the same \( q \) and \( s \) channelized sediment transport rate is more intense than slopewash sediment transport. The physics of this coefficient is not discussed, but \( \beta_2 \) must exceed \( \beta_1 \) to maintain channels. In the model, channels remain fixed in place once they are formed, and cannot be removed from the landscape.

In order to specify where channelized sediment transport should take over from slopewash, the model requires that a "channel initiation function"

\[ a = \beta_5 \cdot q^{M+N} \quad M,N > 0 \]  

(7.24)

be set equal to some previously chosen value, \( a_0 \). Willgoose, Bras and Rodriguez-Iturbe (1991a, p. 240) state that "Criteria like surface velocities, shear stress, and groundwater seepage can all be cast in [the form of equation (7.24)]." However, they do not explain the physics of the connection between these hydraulic characteristics and the onset of channelization. Thus, at present, the specification of channels in the landscape is arbitrary. The authors (Willgoose, Bras and Rodriguez-Iturbe (1991a, p. 241) point out that "...the channel head used in the model only indicates that the transport rate changes at the channel/hillslope interface. The model does not indicate anything about whether the actual channel head in the field (if it can be identified) is abrupt or gradual...."

The model is applied to a surface that has some small stochastic topography, and the channel initiation function (equation 7.24) is calculated for each cell of the surface for some steady-state rate of runoff per unit area. Cells that have \( a \)-values greater than a chosen \( a_0 \) are defined to be channels and are connected to a specified outlet along local topographic gradients. The intensified sediment transport in the channels causes incision and local steepening and therefore increases in the channel initiation function around their heads. The channels extend into those localities with sufficiently high \( a \)-values. The positive feedback continues to drive headward network extension until the drainage areas which yield \( q \) in
equation (7.24) and the associated local slope cause the channel initiation function to converge on \( q_t \). Sediment mass-balance is satisfied across the channel tip by a gradient change (between \( s_h \), the hillslope gradient, and \( s_c \), the channel gradient) controlled by the ratio of the sediment rate constants, \( \beta_1 \) and \( \beta_2 \), and the exponent, \( n \). Thus, since the \( m \) values in equations (7.22) and (7.23) are equal, as are the \( n \)-values used in the current formulation, at the channel head

\[
\beta_1 q_h^n s_h^n = \beta_2 q_c^n s_c^n \tag{7.25}
\]

and

\[
q_h = q_c \tag{7.26}
\]

therefore

\[
\frac{\beta_1}{\beta_2} = \left( \frac{s_c}{s_h} \right)^n \tag{7.27}
\]

or

\[
\frac{s_c}{s_h} = \left( \frac{\beta_1}{\beta_2} \right)^{1/n} \tag{7.28}
\]

In most of the model runs by Willgoose, the ratio \( \beta_1/\beta_2 \) was set equal to 0.1, and \( n = 2.1 \). Hence the slope ratio, \( s_c/s_h = 0.33 \); that is the gradient of the channel at its head was one-third as steep as the gradient of the hillslope just up-valley from the channel tip.

As the channel initiation function approaches \( a_t \), the model calculates an index, \( Y \), which is not defined physically but which changes very rapidly from 0 for the hillslope to 1.0 for a channel. During this transition, which is accomplished by a highly non-linear, arbitrary equation (Willgoose, Bras and Rodriguez-Hurbe 1991b, p. 1672) the regions of high \( a \)-value around channel heads behave with weighted sediment-transporting properties of hillslopes and channels. This mechanism is apparently a device for stabilizing the computational behavior of the model, and its specification remains outside the limits of understanding of the physics of erosion near channel heads.

The papers by Willgoose and his colleagues employ the coupled channel network and hillslope model to explore a variety of long-standing problems concerning landscape evolution, and to suggest some critical problems for future investigation. However, the model examines the consequences of imposing a selected threshold condition on an evolving landscape. The non-physical manner in which the model specifies the location and migration rates of channels emphasizes the still unsolved problem of the mechanical basis for predicting the location of the channel heads at which the intensity of sediment transport increases dramatically, and therefore allows the incision of channel networks that drive landscape evolution.

CONCLUSIONS

The main points of this chapter are:

(1) A stream channel head is the upstream boundary of concentrated flow and sediment transport between definable banks.

(2) Channel heads are usually easy to recognize in the field and are either gradual or in the form of a small or large step (less than 1 meter), or small or large headcut.

(3) Channels lie in valley networks, but rarely extend the full length of the network to the divides; instead, they generally terminate in topographically convergent areas which range from large, low-gradient alluvium-filled valleys to subtle swales on steep hillslopes.

(4) Field surveys have revealed a strong inverse relationship between the drainage area and local valley gradient at the channel head in semi-arid and humid regions.

(5) The length of the source basin above a channel head is roughly equal to the inverse of the drainage density and, therefore, to the average spacing between channels.

(6) Processes that advance or bury channel heads and that make specific heights and widths of channel heads are only qualitatively understood.

(7) Where significant overland flow occurs,
rainsplash and erosion resistance due to cohesion and vegetation prevent sheetwash from cutting channels that would otherwise propagate to the drainage divides. Rainsplash destroys incipient emergent channel banks, whereas surface resistance can prevent upslope erosion by sheetwash. High rainfall intensity, low infiltration capacity, steep gradients and high erosion of soil will each tend to favor sheetwash erosion, causing source–basin lengths to be smaller. Cohesion causes steps and headcuts with plunge pools to form and mass failure can then become an important erosion process.

8 Seepage erosion occurs when a critical hydraulic gradient develops at the head resulting in grain displacement or mass failure. In this case, the source–basin length should decrease with lower hydraulic conductivity, high precipitation intensity, strong topographic convergence and steeper slopes.

9 Periodic landsliding at the downslope end of unchanneled valleys can control channel head location in steep areas where subsurface flow predominates. Source–basin length in this case is again expected to be smaller on hillslopes with low hydraulic conductivity, low frictional strength of the soil, high precipitation, steep slopes, and strong topographic convergence.

10 Tunnel scour due to flow in cracks, root-holes, burrows and other conduits can lead to enlargement of these conduits and eventual collapse, forming a channel. It is most effective in materials with soluble cements or dispersible clay minerals.

11 Landscape models have predicted valley development, rather than the location of a channel with distinct banks. Long-term survival of the channel against the tendency to fill in with detritus discharged from the adjacent hillslopes requires that the sharp channel bank have a much lower diffusion coefficient than the adjacent hillslope or that the channel be frequently rejuvenated by failures and rapid incision. Either condition appears to require the influence of cohesion.

Thus, whether the channel head and banks are localized by a threshold effect or through a balance of sediment transport, some form of cohesion due either to interparticle forces or vegetation is necessary. Models of channel network formation using only a mass-balance approach cannot stabilize channel banks and heads on the scale at which they occur in the field, and so current models are predicting only valley or hollow networks. Within those networks, the channel heads remain to be specified.

The problem of explaining the channel head is a difficult one. At the level of a node in a numerical model in which the channel head is a transition point, the problem can be handled in a fairly straightforward manner, once certain assumptions are made, and the work by Kirkby (1987) and Willgoose (1989) are notable contributions in this regard. However, at the level of standing at a channel head in the field and asking why the head has a certain size, or how one might predict its location in a catchment, we seem to have only a qualitative grasp of the problem. The theoretical work by Howard and MacLane (1988), although oriented toward a particular laboratory problem, illustrates how the problem might be tackled.

The tactic of seeking simple relationships that can be predicted theoretically (without extensive knowledge of details), such as the relationship between drainage area to the channel head and the local valley slope, seems a productive intermediate approach if there is evidence of a single process which controls formation of the channel head, and more field work of this kind, particularly in arid landscapes might prove instructive.

In the long run, however, it appears essential that transport "laws" for runoff and sediment transport must be tested with field data. One can (and people do) model with the conviction that the assumed transport laws are essentially correct, but what test can then be employed to validate the theory? The area–slope relationship provides only a very crude challenge to such models. What other tests of a theory of landscape evol-
tion based on a channel initiation model can be designed?

With the increasing concern about global response to climatic change, the problems of predicting the movement of the channel head takes on an even greater importance. It has generally been argued that climatic fluctuations lead to erosion and deposition cycles on hillslopes and rivers, but the mechanistic linkage between the two is not established. Behavior of the channel head is one of the links between hillslope and valley response. Uplift advance and incision of the channel leads to gullying and evacuation of stored colluvium and alluvium. Downslope retreat and net valley aggradation may cause thick deposits of colluvium to accumulate on slopes. Meis and Moura (1984) explore these ideas as they apply to south-eastern Brazil. Also, the response of a landscape to uplift and relative base-level lowering must eventually be transmitted through the channel head. The only way a particular point in a hillslope receives the signal of base-level lowering is through net erosion that travels up the channel network, perhaps locally deepening the footslopes of adjacent hillslopes, and through the channel head.

Despite the importance of the channel head, the problem has received relatively little attention. It is a difficult but vital one, holding components essential to predicting landscape evolution, morphology and response to climatic fluctuations and, possibly, to base-level changes.

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