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Wetlands Regulatory Assistance Program (WRAP)

The Upstream Extent of a River Network

A Review of Scientific Knowledge of Channel Heads

Ellen Wohl

June 2018

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The Upstream Extent of a River Network

A Review of Scientific Knowledge of Channel Heads

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Abstract

A river channel is a linear feature with definable bed and banks created by erosion of water concentrated into persistent flow paths. A river starts at the channel head, which is the upstream-most point of concentrated water flow and sediment transport between definable banks that are spatially continuous downslope. The locations of individual channel heads are difficult to predict because of hillslope-scale differences in gradient, infiltration capacity, porosity and permeability, and cohesion, each of which influences flow paths and erodibility of near-surface materials. Understanding the state of knowledge regarding the initiation point of channels can be useful in a management context when assessing what features on the landscape constitute river channels. Therefore, the primary objective of this report is to concisely summarize the existing state of scientific knowledge regarding where river channels and channel networks begin in the landscape. The report draws on published studies from diverse regions within the United States and around the world but focuses on research summarized in English. This report introduces terms, reviews the processes related to channel initiation in different landscape settings, discusses field and remote identification of channel heads, and outlines the research needs for channel-head identification and for improving channel delineation.

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Preface

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Acronyms and Abbreviations

CPOM	Coarse Particulate Organic Matter
CRREL	Cold Regions Research and Engineering Laboratory
CWA	Clean Water Act
DEM	Digital Elevation Model
ERDC	Engineer Research and Development Center
FPOM	Fine Particulate Organic Matter
GPS	Global Positioning System
MAP	Mean Annual Precipitation
OHWM	Ordinary High Water Mark
S-A	Slope-Area
USACE	U.S. Army Corps of Engineers
WRAP	Wetlands Regulatory Assistance Program

1 Introduction

1.1 Background

The Ordinary High Water Mark (OHWM) is defined by federal regulation as "that line on the shore established by the fluctuations of water and indicated by physical characteristics such as a clear, natural line impressed on the bank, shelving, changes in the character of soil, destruction of terrestrial vegetation, the presence of litter and debris, or other appropriate means that consider the characteristics of the surrounding areas" (U.S. Congress 1986). The OHWM determines the lateral extent of federal jurisdiction of non-tidal waters of the United States in the absence of adjacent wetlands, according to Section 404 of the Clean Water Act (CWA) (U.S. Congress 1977). Therefore, accurate, consistent, and repeatable delineation of the OHWM is essential for proper implementation of the CWA.

As part of implementing the CWA, the U.S. Army Corps of Engineers must identify the longitudinal extent of streams along with the lateral extent. Where the transition between stream and hillslope processes is gradational over some distance, it can be challenging to distinguish between streams and non-stream conveyances. Channel-head indicators, along with the mechanisms for channel initiation, are variable across different regions, landforms, and land uses, creating additional difficulty in consistent identification of these features. Additionally, the discontinuous nature of stream channels, at their initiation point, can make it difficult to delineate the OHWM.

1.2 Objectives

The primary objective of this report is to concisely summarize the existing state of scientific knowledge regarding where river channels and channel networks begin in the landscape. Understanding the state of knowledge regarding the initiation point of channels can be useful in a management context when assessing what features on the landscape constitute river channels.

1.3 Approach

The report draws on published studies from diverse regions within the United States and around the world but focuses on research summarized in the English language. The report is written to be accessible to nonspecialist readers interested in understanding river process and form. The report is broken into six sections. The first section introduces terms and gives an overview of the report. Section 2 and 3 focus on the processes related to channel initiation in different landscape settings. Sections 4 and 5 focus on field and remote identification of channel heads. The final section outlines the research needs for channel-head identification and for improving channel delineation.

1.4 Definitions

The glossary at the end of this report provides a more comprehensive list of definitions, but it is useful to define several basic terms before reviewing current knowledge of where channels begin. A *river channel* is typically defined as a linear feature with a definable bed and banks at the ground surface created by erosion of water concentrated into persistent flow paths. This definition implies nothing about size and can encompass a channel less than 1 m wide or more than 1 km wide. Ephemeral and intermittent channels may include longitudinal segments in which flow becomes very diffuse and a bed and banks are not readily defined. In this scenario, the presence of a channel is typically assumed based on more readily definable segments up- and downstream.

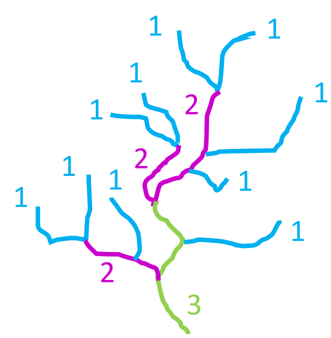
A river channel is created predominantly by flowing water and is thus distinguished from similarly appearing surface features created by debris flows, dry ravel, rockfall, or other colluvial processes that are created during the downslope movement of dry sediment or sediment-water mixtures dominated by sediment. In this report, *river* includes all channels created by water and subsumes terms such as *stream*, *creek*, *brook*, *gully*, *wash*, and *arroyo*.

A river does not have to contain flowing water at all times. A river can be ephemeral and flow only during and soon after precipitation inputs and have no groundwater inputs or base flow (USACE 2012). A river can also be intermittent with continuous flow only at certain times of the year when the water table intersects the surface along the channel course, such as when the river receives water from a spring or a surface source such as melting snow (Osterkamp 2008). Dry channel segments alternate with flowing channel segments to create longitudinally discontinuous flow along intermittent rivers during periods of low flow (Reynolds et al. 2015). Intermittent rivers are also known as temporary rivers because of the lack of longitudinally and temporally continuous flow (Arthington et al. 2014). Even in wet climates, small rivers in particular can be ephemeral or intermittent. Finally, a river can be perennial, with continuous surface flow at all times.

A river starts at the *channel head*, which is the upstream-most point of concentrated water flow and sediment transport between definable banks (Montgomery and Dietrich 1988, 1989). The channel head does not necessarily coincide with the *stream head*, which is the upstream-most extent of perennial flow within a river (Jaeger et al. 2007). A channel head represents a transition from diffusive hillslope processes to a distinct channel (Dietrich and Dunne 1993) that persists more than briefly during and shortly after precipitation such as rainfall or snowmelt. This persistence distinguishes a river channel from a rill, which also has definable bed and banks created by concentrated water flow but is a transient feature present only during and briefly after precipitation.

Although a channel head can be present on a planar slope, channel heads typically occur within hillslope concavities that serve to concentrate surface and subsurface water flow sufficiently to create channel erosion. The presence of a slope concavity can also result in accumulation of sediment moving downslope via diffusive processes such as sheetwash. A concavity that accumulates sediment upslope from a channel head is known as a *colluvial hollow* (Dietrich and Dunne 1978) or *zero-order basin* (Tsukamoto et al. 1982). Downstream from the channel head, the river is a first-order channel until joined by a tributary. In the most commonly used ordering system, a river increases in order when joined by a channel of equal order (i.e., two first-order channels create a second-order channel, two second-order channels create a third-order, etc. [Strahler 1952], as shown in Figure 1).

Figure 1. Illustration of stream order after Strahler (1952), with different colors representing different stream orders.



The upstream-most segments within a river network are known as *head-water channels*. Different authors refer to first-order (Nadeau and Rains 2007), first- and second-order (Meyer et al. 2007), and first- through third-order channels (Adams and Spotila 2005) as headwaters; but the most common usage for *headwaters* is for first- and second-order channels.

2 Process Domains within a Watershed

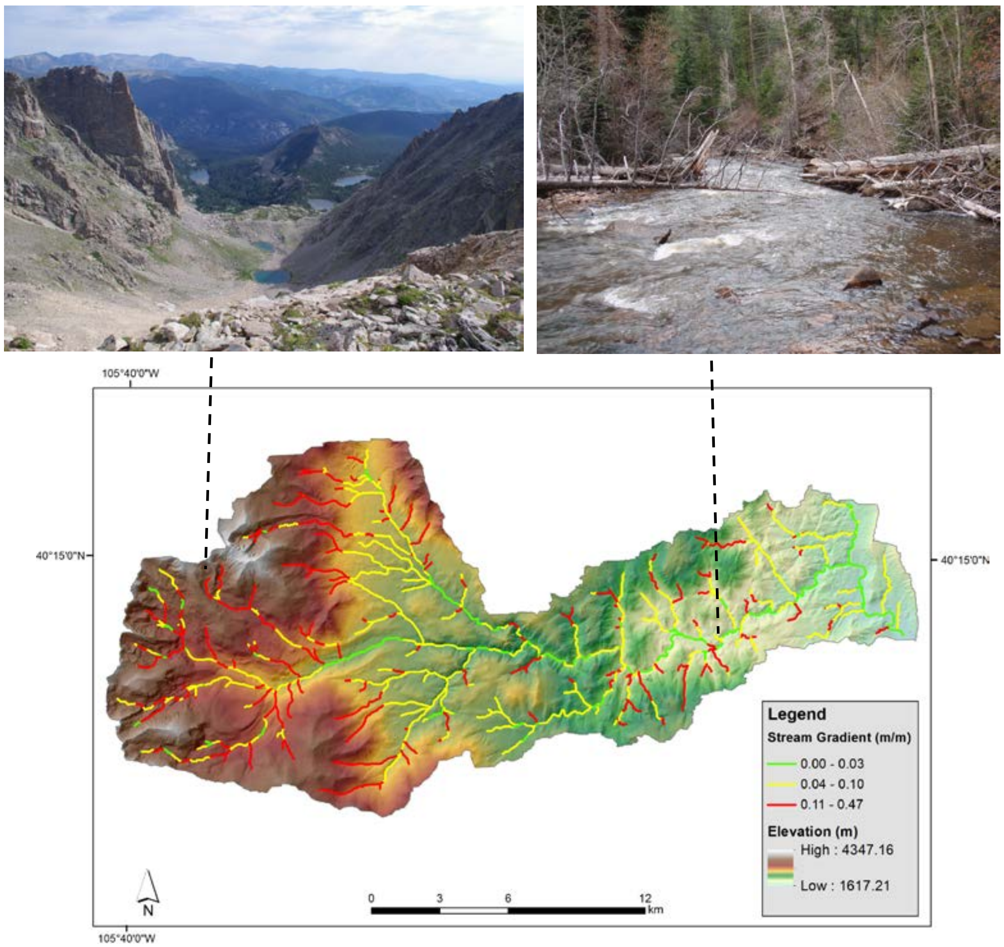
2.1 Concepts of spatial zonation of process and form

A *process domain* is a spatially identifiable area characterized by a distinct suite of geomorphic processes (Montgomery 1999). The process domain concept implies that landscapes and river networks can be divided into discrete regions in which biotic community structure and dynamics respond to different disturbance regimes. Process domains are useful because they reflect spatial differences in disturbance regimes and dominant geomorphic processes shaping distinct portions of a landscape, as illustrated by two examples.

In mountainous river network A, Pleistocene glaciers created broad, low-gradient valleys now covered in alpine vegetation and drained by the headwaters of the river network. River flow is dominated by seasonal snowmelt. Rockfalls recurring at intervals of hundreds of years introduce abundant coarse sediment to the channel network. Downstream portions of the river network flow through lower elevation forests that introduce large wood to the channels. Fires occurring at intervals of a few decades followed by high-intensity rainfall from convective storms result in flash floods that effectively redistribute the wood and sediment along the river network. In this example, upstream portions of the river network have a lower magnitude and frequency of disturbance, a different flow regime, and distinctive morphology relative to lower portions of the river network (Figure 2).

One portion of lowland river network B is composed of intermittent and perennial channels that head at springs because this portion of the drainage network is underlain by relatively porous and permeable sandstone bedrock. Another portion of the network underlain by impermeable shale facilitates abundant surface runoff during precipitation but supports only ephemeral channels. In this example, the intermittent and perennial channels have a different morphology and support different aquatic and riparian biota than the ephemeral channels in the river network.

Figure 2. An example of distinct process domains within a drainage network or region and a map of reach-scale stream gradient segments and watershed elevation within the North St. Vrain Creek watershed in the Colorado Front Range. The photos illustrate high-elevation, glaciated valleys (at *upper left*, this photo is from an adjacent watershed) and lower-elevation, fluvial valleys with forest cover and instream large wood (at *upper right*).



2.2 Hillslope and channel process domains

As noted above, the channel head represents a distinct transition from diffusive movement of water, solutes, and sediment down slope to concentrated flow within a channel. As such, channel heads create a boundary between hillslope and fluvial process domains. *Hillslope* in this context does not necessarily refer to a steeply sloping surface. In regions of low relief, an unchanneled hillslope may have a gradient of only one to two degrees.

The boundary marked by a channel head can vary through time. Disturbances such as fire, landslides, or extreme storms can increase surface runoff and sediment mobility, for example, causing channel heads to move upslope abruptly (Kirkby et al. 2003; Wohl 2013). Depending on site-specific conditions, channel heads can gradually move downslope again over a period of years. This is a critical point to understand: the location of any channel head over a period of years to decades is best understood as an average location that can and does change through time. The magnitude of change through time in the location of a channel head will reflect the disturbance regime of the hillslope—a channel head located on a stable slope will exhibit relatively little change through time—and the speed with which hillslope-channel processes recover following disturbance. Channel heads may recover quickly following disturbance, for example, at sites where vegetation regrows quickly and thus stabilizes downslope movement of water and sediment (Wohl and Scott 2017).

2.3 Colluvial versus fluvial channels

In regions with moderate to high topographic relief, colluvial and fluvial channels can be present. *Colluvial channels* are those in which sediment transport and channel form are dominated by nonfluvial erosion; predominantly, colluvial channels are dominated by debris flows. The distinction is important because debris flows can transport much greater quantities of sediment and larger-sized sediment than water flows because of the greater viscosity of debris flows (Coussot and Meunier 1996). Because debris flows do not exhibit the same type of flow mechanics as water flows, indirect methods of fluvial discharge estimation that use erosional and depositional features to infer flow magnitude will produce inaccurate results if applied to debris-flow channels.

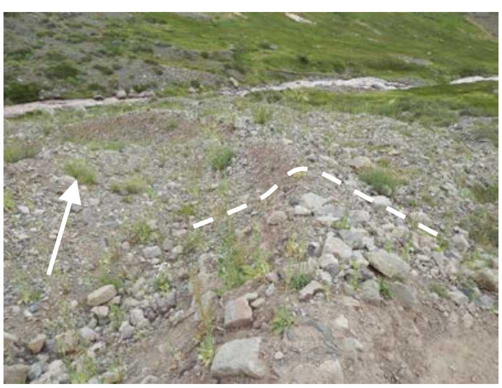
Channels dominated by debris flows deposit coarse-grained levees that tend to be sharp crested rather than flat crested as in flood bars and levees (Figure 3). Debris-flow deposits lack the bedding, sorting by grain size, and imbrication of large sediment clasts characteristic of water-deposited sediments. Debris flows also result in lobate deposits in portions of the river network where downstream gradient decreases and/or the channel becomes less laterally confined, such as at mountain fronts.

Recognition of colluvial channels is important because the river network, if defined as including only those channels formed by fluvial processes, may start well downstream from individual channel heads. These distinctions are very difficult to apply in the field. A particular channel may be formed by fluvial processes occurring every year, for example, but may then be substantially modified by a debris flow that does not recur for another century. Judging the relative importance of debris flows versus fluvial processes in creating and maintaining a channel head or river segment can be subjective and influenced by the time elapsed since the last debris flow.

Figure 3. (*A*) Examples of flood bars with relatively flat tops, here along Skin Gulch in the South Fork Cache la Poudre River watershed of Colorado. These deposits were created by flooding following a wildfire in the summer of 2012. (*B*) Debris flow deposits. At *left* is a downstream view of a sharp-crested debris-flow levee along an unnamed ephemeral channel tributary to Clear Creek in Colorado. The *white arrow* indicates flow direction in the channel, and the *dashed line* highlights the shape of the levee. At *right* is a view of a debris-flow deposit exposed in a road cut along the Cache la Poudre River. Note the abrupt changes in grain size vertically and laterally within the exposure and the lack of imbrication (alignment of clasts parallel to flow) or bedding. The *white bar* indicates a 1 m length.

A

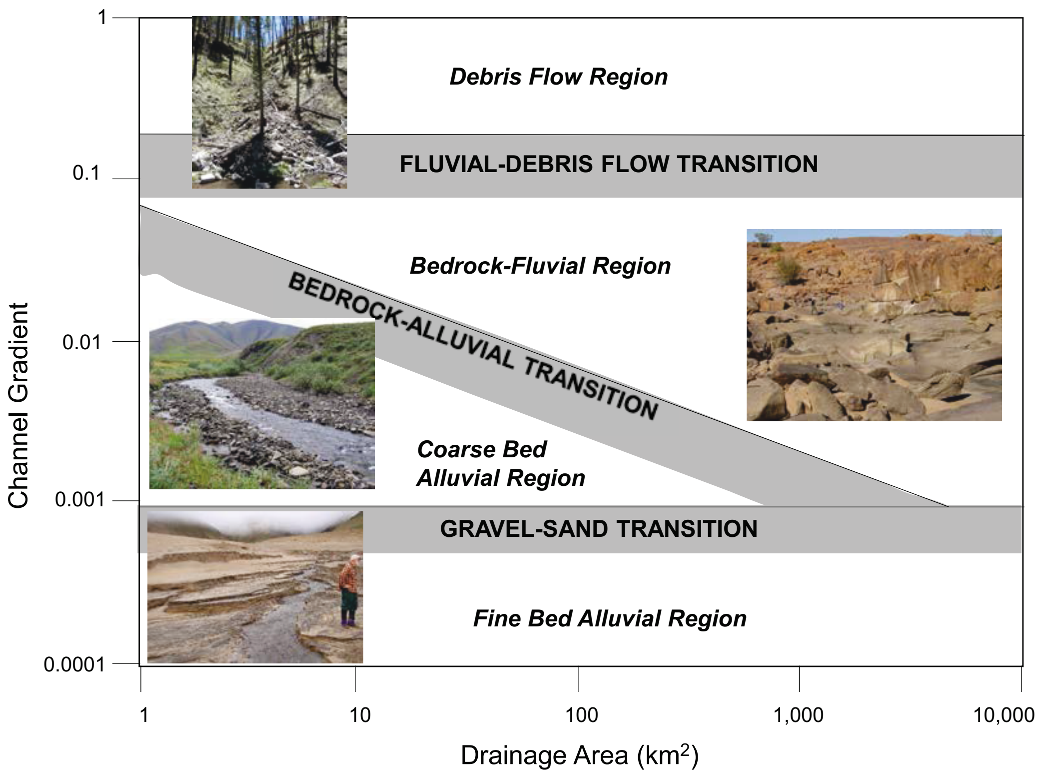
B



2.4 Process domains defined by grain size

In low-relief river networks, the dominant grain size of the channel substrate may change very little downstream. In moderate- to high-relief networks, however, dominant grain size is likely to vary downstream in correlation with channel gradient. As a generalization, the progression from bedrock to boulder bed, gravel bed, sand bed, and silt/clay in the channel bed reflects progressively lower stream gradients (Figure 4). Although stream gradient typically declines downstream, exceptions occur where a river crosses a more resistant substrate. Similar downstream alternations in valley geometry and channel gradient are common in mountainous river networks (Wohl 2010).

Figure 4. Idealized plot of bed substrate versus reach-scale gradient, with inset photos illustrating debris flow and fluvial-substrate types. Plot is after Sklar and Dietrich (1998, Fig. 1), but subsequent research (e.g., Livers and Wohl 2015) indicates that transitions between substrate types are not as clearly differentiated based on slope–area relations as this figure implies. Note the logarithmic scale on both axes.



Distinguishing process domains on the basis of substrate grain size is useful because differences in substrate equate to differences in channel stability and sediment transport. The substrate of sand-bed channels is typically mobile during relatively frequent flows whereas the bed of a boulder-bed channel may become mobile and result in substantial sediment transport only during rare, high-magnitude floods. In the context of channel heads, a channel head formed in sandy a substrate may change its location more readily than one formed in cohesive silt and clay.

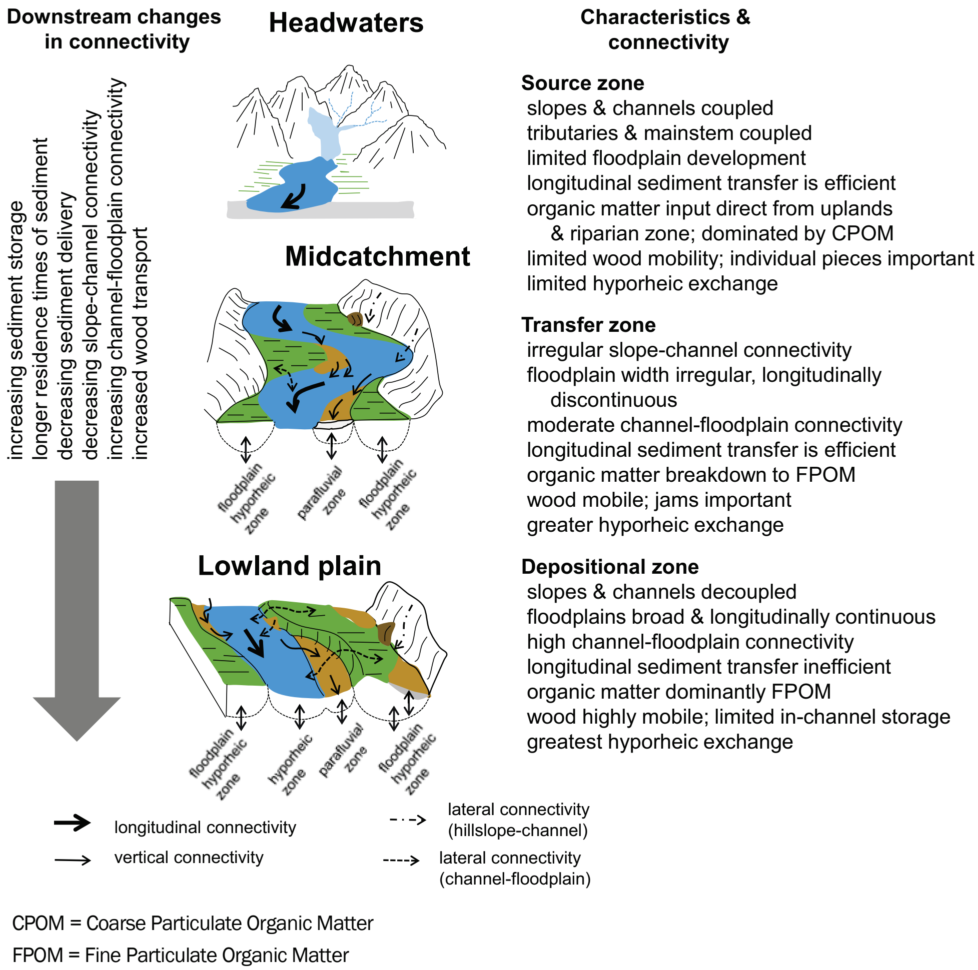
2.5 Glacial and fluvial process domains

Distinguishing portions of river valleys formed only by river erosion and those formed by past glacial erosion and contemporary river erosion can be useful for understanding variations within a watershed that influence channel-head formation. Glaciers typically erode a much wider valley cross section than a river and create steeper sidewalls at elevations below the surface of the ice (Amerson et al. 2008). This valley morphology may be modified only very slowly (over several thousand years) following glacial retreat, so that the river channel flows in an inherited valley morphology with different patterns of gradient and channel morphology than channels with similar drainage area in fluvial valleys (Livers and Wohl 2015). The distinction between glacial and fluvial process domains can be important in the context of channel heads because the average downslope location of a channel head can vary between these process domains (Henkle et al. 2011).

2.6 River styles

Many of the concepts described in sections 2.1–2.5 come together in the River Styles Framework (Brierley and Fryirs 2005; Fryirs and Brierley 2013) first proposed for Australian rivers but applicable to any region of the world. The River Styles Framework is a set of procedural guidelines that can be used to document the geomorphic structure and function of river segments within a watershed context. The guidelines emphasize contemporary river character and behavior (Figure 5) but also evolutionary trajectory and recovery potential. Although the River Styles Framework does not focus on channel heads, the process-based understanding of channels and emphasis on watershed context are also very useful in considering the location and characteristics, through time and space, of channel heads. Downslope movements of water and sediment can differ consistently among river and valley segments differentiated within the River Styles Framework, for example, leading to different local slope–drainage area thresholds for channel-head formation.

Figure 5. River Styles diagram illustrating idealized downstream changes in valley geometry, river characteristics, and connectivity between the river corridor and uplands, as well as connectivity within the river corridor (after Brierley and Fryirs 2005, Fig. 2.10).



3 Factors Influencing Where Channels Start

This portion of the text reviews the erosive processes that can create a channel head and the regional- to local-scale influences on these processes. The predominant distinction is between surface and subsurface processes.

3.1 Surface processes

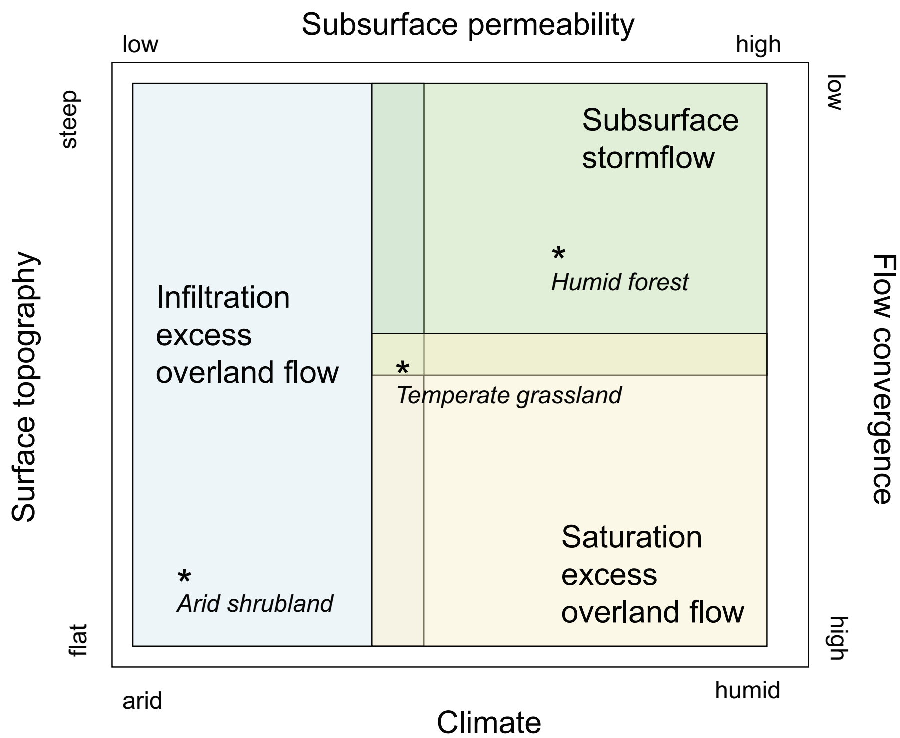
3.1.1 Downslope movement of water

Water can flow downslope at the surface as overland flow, which is surface flow outside the confines of a channel. Overland flow can take the form of Hortonian overland flow, also known as infiltration excess overland flow (Horton 1945), if the infiltration capacity of the hillslope surface is low relative to precipitation intensity. Hortonian overland flow is uncommon in natural watersheds but can occur where vegetation cover is sparse, hillslope gradients are steep, soil is thin or of low permeability (e.g., clay), and precipitation intensities are high. All of these characteristics can occur in arid or semiarid regions and in watersheds where land use compacts the soil or causes erosion of permeable, near-surface soil layers. Hortonian overland flow is quite common on paved surfaces.

Water can also flow down hillslope surfaces as saturation overland flow, also known as saturation excess overland flow (Dunne and Black 1970a, 1970b). Saturation overland flow results from the combined effects of direct precipitation onto saturated areas and return flow from the subsurface as saturation occurs. Saturation overland flow is more likely to reflect antecedent soil moisture and subsurface transmissivity than hillslope steepness (Montgomery and Dietrich 2002). Conditions that favor saturation overland flow include high permeability near the surface, a humid climate with high cumulative water input, and gentler slopes with shallow soils that cannot drain as easily as steep slopes (Kampf and Mirus 2013) (Figure 6).

Saturation overland flow rarely occurs outside of convergent flow zones such as hillslope concavities (Dietrich et al. 1992) but can occur as a result of topographic breaks, permeability contrasts (e.g., roads or pavement), low subsurface storage capacity, geologic structures that promote rapid saturation (e.g., layered basalt flows) (Mirus et al. 2007), exclusion from frozen soil, or snowmelt over saturated soil (Kampf and Mirus 2013). Saturation overland flow commonly occurs first in downslope portions of a hillslope and then expands upslope (Dunne 1978).

Figure 6. Environmental controls on dominant runoff-generation mechanisms. Runoff-generation mechanisms are represented by colored fields; overlapping fields indicate that no single runoff mechanism dominates the hydrologic response. Asterisks represent approximate conceptualizations for example environments (after Kampf and Mirus 2013, Fig. 10).



Overland flow starts when water accumulates on the surface to sufficient depth to begin flowing downslope. Surface roughness created by microtopography and vegetation causes flow resistance that influences the pathways of overland flow (Bergkamp 1998), especially in humid environments that have dense vegetation and organic litter. Overland flow can also modify surface roughness by moving sediment and organic litter, and vegetation influences overland flow by altering infiltration. Clumps of grass or shrubs in semiarid regions, for example, facilitate infiltration and generate less overland flow than intervening bare ground, creating downslope flow pathways that are strongly coupled to vegetation patterns (Dunne et al. 1991).

3.1.2 Downslope movement of sediment

Outside of channels, sediment moves downslope through mass movements and gradual diffusive processes. Aggregates of grains move together during mass movements, which include slides, flows, and creep. Diffusive processes such as rainsplash and overland flow involve movement of individual grains. Both mass movements and diffusive sediment processes can influence the location of channel heads.

Mass movements are typically strongly seasonal as a function of moisture availability and freeze–thaw processes (Hales and Roering 2009). Slides occur when a mass of unconsolidated material moves without internal deformation along a discrete failure plane that can be curved or relatively straight. Slides typically result from either (1) a decrease in the shear strength of the soil as a result of weathering, increased water content, seismic vibrations, freezing and thawing, or land uses such as deforestation and road construction or (2) an increase in shear stress caused by additions of mass or removal of lateral or underlying support. Slides commonly transition downslope into flows, which occur when the moving mass is sufficiently liquefied or vibrated to create substantial internal deformation during downslope movement. Slides or flows recur frequently in many high-relief terrains and transport the majority of sediment to or along low-order stream channels (Jacobson et al. 1993; Guthrie and Evans 2007), thus exerting substantial influence on the location of channel heads and the morphology of first- and second-order channels. Creep occurs when sediment particles displaced by bioturbation (physical displacement via organisms) and in wetting–drying or freeze–thaw cycles move downslope under the influence of gravity (Kirkby 1967). Creep is greatest in the upper meter of soil and is proportional to surface gradient (Selby 1982; McKean et al. 1993).

Gradual diffusive processes of sediment movement can take the form of rainsplash or overland flow. Rain falling on a surface can loosen or detach individual particles, making the particles more susceptible to entrainment by overland flow (Furbish et al. 2009; Dunne et al. 2010). Overland flow is most likely to be capable of eroding measurable quantities of sediment where unvegetated, unfrozen hillslope surfaces are exposed (Dingwall 1972; Rustomji and Prosser 2001). Thread flow occurs when overland flow goes around individual roughness elements, such as small pebbles. Sediment can be stripped evenly from a hillslope crest and upper zone during sheet flow that submerges individual roughness elements and forms a relatively continuous sheet of water across the hillslope. Erosion by sheet flow is most effective where interparticle cohesion has been reduced by needle ice, trampling, or disturbance to vegetation (Selby 1982). Conversely, microbiotic soil crusts, very coarse particles at the surface, or vegetation cover substantially reduce sediment detachment and erosion by rainsplash or sheet flow (Uchida et al. 2000). Seasonally or permanently frozen soil can enhance overland flow and soil erosion during the melt season by impeding infiltration (Ollesch et al. 2006).

At some distance downslope, surface irregularities concentrate overland flow into slight depressions that then enlarge as increasing water depth increases the shear stress acting on the substrate at the base of the flow. This can give rise to a channel head. Downslope movement of sediment, however, tends to be highly spatially and temporally variable as a result of local changes in hillslope gradient, ground cover, vegetation, and microtopography (Saynor et al. 1994). Most hillslopes and the location of most channel heads are shaped through time by some combination of multiple processes (Jimenez Sanchez 2002).

3.2 Subsurface processes

Subsurface flow commonly dominates hillslopes with full vegetative cover and thick soils (Dunne and Black 1970a). Infiltrating water that remains in the subsurface can flow downslope in the unsaturated (vadose) zone above the water table as throughflow or in the saturated (phreatic) zone below the water table as groundwater. In either case, subsurface water flowing through small, interconnected pores will have low-velocity laminar flow (Kampf and Mirus 2013). When the void space is filled with water under saturated conditions, the hydraulic conductivity (ease of water flow through the medium) reaches a maximum. When the void space is not completely filled, connectivity of pore space and hydraulic conductivity decrease.

Hillslopes are typically unsaturated at the ground surface, and water movement is predominantly vertical during infiltration. Subsurface moisture is redistributed through vertical and lateral movements that are influenced by interactions between infiltration and evapotranspiration (Kampf and Mirus 2013) and can take the form of diffuse wetting fronts, fingered flow paths, or preferential flow along conduits such as macropores or pipes (Dunne 1980; Jones 1981; Wang et al. 2003). Diffuse throughflow and fingered flow paths reflect the general porosity and permeability of the unsaturated zone. Macropores are openings sufficiently large that capillary forces have an insignificant effect on the water flowing through the pores (Germann 1990), leading to relatively rapid, turbulent flow. Macropore flow is likely triggered at a threshold wetness level (Beven and Germann 1982) and is particularly widespread on densely vegetated hillslopes in steep terrain where dense biological activity (burrowing animals and plant roots) creates high concentrations of macropores.

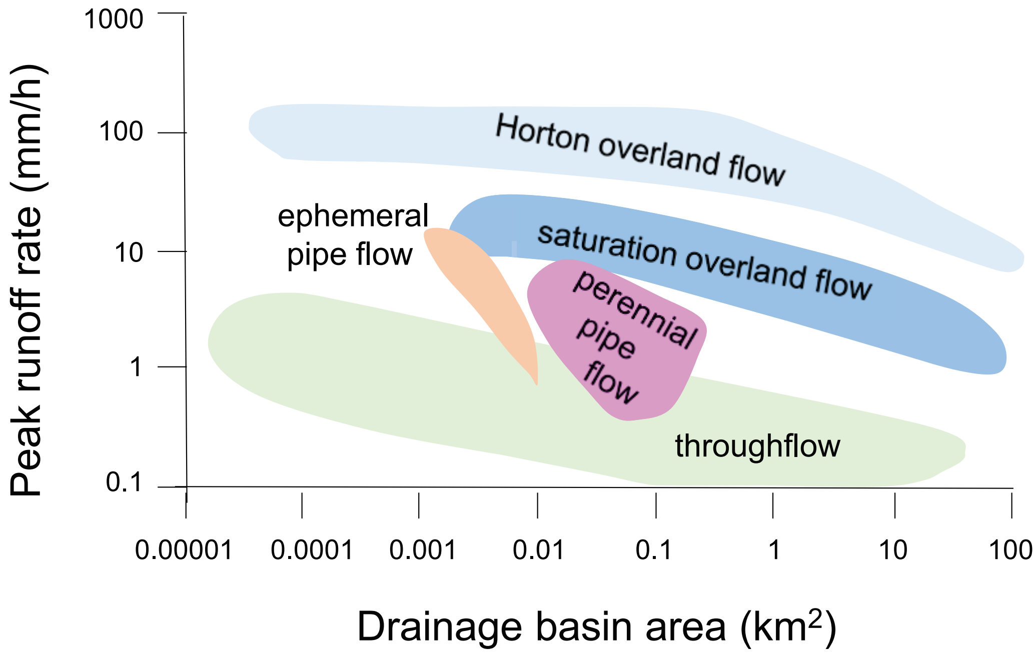
Substantial preferential flow through macropores can facilitate the formation of soil pipes. Pipes are larger than macropores but can vary in size from only a few centimeters in length or diameter to more than 2 m in diameter and hundreds of meters in length (Selby 1982). Pipes typically form just above a zone of lower porosity and permeability or along a cavity created by animal burrowing or the decay of plant roots. Piping can occur anywhere but is particularly associated with arid and semiarid regions. Piping can contribute nearly half of stormwater flow in some catchments (Jones 2010). Pipe networks can exist at multiple levels in the subsurface with each level being activated by precipitation of different magnitudes (Gilman and Newson 1980; Kim et al. 2004). Hillslopes can “turn on” when sufficient water infiltrates to activate rapid lateral flow via pipes (Uchida et al. 2001; McDonnell 2003), allowing water stored in the subsurface to be rapidly released during storms (Kirchner 2003). This can create abrupt changes in hydrologic response and water delivery to channels as hillslope wetness state changes (Kampf and Mirus 2013).

Downslope water movement below the water table can also be highly complex as a function of spatial variations in depth and rate of groundwater movement. The water table responds separately in riparian and hillslope zones (Seibert et al. 2003), for example, so that upslope and downslope groundwater dynamics can be asynchronous.

3.3 Summary

Compared to the other types of flow discussed here, water moving downslope via Hortonian overland flow typically moves most rapidly (50–500 m/h), followed by saturation overland flow, throughflow, and groundwater flow, which can move as slowly as 1×10^{-8} m/h (Selby 1982). Preferential throughflow in pipes or macropores is also capable of moving rapidly (Figure 7).

Figure 7. Generalized graph of hillslope hydrologic processes in relation to the size of the contributing area (after Jones 2010, Fig. 7b). The drainage basin area in this figure reflects the total contributing area of the channel upstream from the point at which runoff enters a channel but does not necessarily reflect the continuous extent of the area generating runoff. A 100 km² drainage basin area, for example, does not mean that Hortonian overland flow has to persist for tens of kilometers before entering a channel.



The distribution of water moving downslope via different pathways can alter in relation to precipitation magnitude, intensity, or duration during a storm or on a regular annual basis in strongly seasonal climatic regimes. Dominant runoff processes also change with spatial scale from a soil column to the hillslope scale (McDonnell et al. 2005). Hillslope flow paths occur along a spectrum from predominantly vertical to predominantly lateral (Elsenbeer 2001), but vertical and lateral flow are both highly nonlinear and exhibit threshold behavior that influences downslope connectivity (Sidle et al. 2001; Hopp and McDonnell 2009).

3.4 Special conditions

Karst terrains and cold regions underlain by permanently frozen ground form distinctive subsets in terms of regional patterns of downslope pathways of water. Karst terrains underlain by carbonate bedrock (limestone and dolomite) or evaporites (e.g., gypsum) have distinctive landforms and drainage processes because of greater bedrock solubility in the presence of natural water. Surface flow in karst terrains can move abruptly downward to the groundwater via swallow holes, which are open cavities where surface flow passes from rocks of low solubility onto carbonate rocks. Precipitation falling on karst terrains can percolate to the groundwater via diffuse infiltration (this is known as *vadose seepage*) or can move rapidly downward via a highly permeable zone created by vertical joints in the bedrock or cylindrical solution openings (this is known as *vadose flow* or *internal runoff*) (Ritter et al. 2011).

The extent of frozen soil strongly influences downslope pathways of water in cold regions where average annual temperature is low enough to allow permafrost, or permanently frozen ground. An active layer that varies from 15 cm to 5 m thick in different locations is present at the surface. The active layer, which thaws seasonally, overlies the permafrost. Because frozen soil impedes infiltration and limits percolation, a large portion of warm-season precipitation moves downslope as overland flow and is quickly delivered to channels (Vandenberghe and Woo 2002). In regions with very low relief, however, permafrost can result in extensive bodies of standing water and limited integration of the drainage network. As the active layer thaws, the depth and importance of infiltration can change; but the presence of permafrost ultimately limits deep infiltration and groundwater flow.

3.5 The role of disturbance

Disturbances, such as wildfire or land use, that alter the characteristics and extent of vegetation and the permeability of the soil can disrupt surface and subsurface downslope pathways of water and sediment. The most common scenario is that the disturbance reduces vegetation cover and decreases soil infiltration capacity, causing channel heads to migrate upslope and initiate with smaller contributing areas. Wildfire (Wohl 2013), timber harvest (Montgomery et al. 2000), grazing, and cropping (McNamara et al. 2006) are all documented to have this effect whereas urbanization can increase the contributing area of channel heads, primarily because headwater streams are buried or piped to accommodate urban growth (Roy et al. 2009). Paved or unpaved roads are especially effective at altering downslope flow pathways, and both channel heads and hillslope mass movements typically initiate at roads (Montgomery 1994; Larsen and Parks 1997).

Where the land use change persists, the alteration of channel-head locations is also likely to be persistent. In the case of an episodic disturbance such as wildfire, the duration of the change in channel heads depends on the rate at which vegetation regrows and soils recover permeability. Documenting changes in channel heads after a 2012 wildfire in the Colorado Front Range, Wohl (2013) found that channel heads migrated upslope and formed at minimum drainage areas two orders of magnitude smaller than prefire minimum drainage areas (Figure 8, top). However, contributing areas for channel heads had mostly returned to prefire locations within four years of the fire (Figure 8, bottom).

Wohl and Scott (2017) distinguished three populations of channel heads in the study area: (1) permanent channel heads on distinct hillslope concavities, with the channel head fixed in place by a persistent slope discontinuity such as a bedrock outcrop (the location of these channel heads was not affected by the fire); (2) transient channel heads on straight and convex hillslopes, which formed immediately after the fire but then completely disappeared within a few years as diffusive downslope movement of sediment and regrowth of vegetation obliterated the channel head; and (3) mobile channel heads, typically located on concave portions of the hillslopes, which moved upslope immediately after the fire but then gradually moved back downslope to a prefire location.

Disturbance can also take the form of an extreme precipitation input (Montgomery and Dietrich 1992), such as a convective storm or a dissipating tropical storm. Analogous to the distinctions drawn by Wohl and Scott (2017), Kirkby et al. (2003) distinguished ephemeral channel heads produced in swales in response to a single storm from permanent channel heads that represent the cumulative impact of the distribution of storms over a period of decades to centuries. The location of the ephemeral channel heads depends on the magnitude of the storm and, although both populations of channel heads have an inverse relationship between hillslope gradient and drainage area, the ephemeral channel heads typically form upslope from the permanent channel heads.

Figure 8. *Top*: Examples of channel heads that moved rapidly upslope following a wildfire in a semiarid conifer forest. The location of each channel head is *circled*, and an arrow indicates the flow direction. In each case, the channel head is only 10–20 cm tall; but evidence of channelized flow in the form of an eroded channel is continuous downslope from this point. Upslope areas have sheetwash and discontinuous small rills. *Bottom*: Views of a hillslope immediately after a June 2012 fire (*left*) and four years later (*right*), showing recovery of understory vegetation, including woody shrubs and tree saplings.

3.6 Regional- to local-scale controls on channel heads

As noted earlier, channel initiation is a threshold phenomenon in which surface or subsurface flow concentrates and persists sufficiently to create a channel head that separates the process domains of hillslopes or colluvial hollows from channel networks (Dietrich and Dunne 1993) (Figures 9–13). The stream head (Figure 14), which is the start of perennial flow, does not necessarily coincide with the channel head. Channel segments of ephemeral and intermittent flow can be present downslope from the channel head and upslope from the stream head even in wet regions, and this pattern is the norm in dry regions.

Figure 9. Example of a gradual channel head created by surface runoff on a desert surface in Arizona, USA. The channel grows wider and deeper toward the back of the photo (flow is from foreground to rear in this view).

Figure 10. An unchannelized, colluvial hollow along a hillslope in the drainage of the Wulik River in Alaska, USA.

Figure 11. In this view of a tributary channel head along the Wulik River in Alaska, USA, the channel starts well down the slope. The crest of the slope forms the horizon in this view.

Figure 12. Upslope view in a subalpine meadow of Rocky Mountain National Park, Colorado, USA. The channel head starts very subtly in a swale; channelized surface flow begins at the lower portion of this view and is about 20 cm wide.

Figure 13. Channel network formed primarily by piping and sapping, rather than surface erosion, in the Pawnee National Grassland of northeastern Colorado, USA. The channel heads at lower right are approximately 30 m across at the top, and 6 m deep.

Figure 14. The stream head—the start of surface flow—along a tributary of the Wulik River in Alaska, USA (*left*), and as a spring at the base of a hillslope in the Lake District of England (*right*).

The locations of individual channel heads can have substantially different drainage areas even over relatively small portions of a watershed as a result of hillslope-scale differences in gradient, infiltration capacity, porosity and permeability, and cohesion, each of which influences surface and subsurface flow paths and erodibility of near-surface materials. The distribution of channel heads can also reflect primarily surface or subsurface flow, some combination of the two, or mass movements. Regardless of the dominant mechanism, flow convergence facilitated by topography and/or stratigraphy promotes the concentration of flow that initiates channels (Dunne 1990; Dietrich and Dunne 1993).

Channel heads that reflect primarily surface processes of Hortonian or saturation overland flow are associated with the development of rills. Rills can develop nearly simultaneously across a terrain and then integrate into a network (Dunne 1980), although arid and semiarid regions commonly have discontinuous headwater networks of short, actively eroding channel reaches separated by unchanneled or weakly channeled, vegetated, stable reaches (Tucker et al. 2006). Alternatively, channels can extend downslope during slow warping or intermittent exposure of new land on a rising land surface, or channels can extend upslope in response to an increase in slope gradient or the lowering of base level (Dunne 1980). Erosional hot spots occur where topographic constrictions or locally steep gradients amplify hydraulic forces sufficiently to overcome surface resistance and initiate headcuts (Tucker et al. 2006), which may coincide with the channel head or occur downstream from the channel head. Headcuts that coincide with the channel head are vertical faces that separate upslope unchanneled environments from downslope channels. Headcuts downstream from the channel head separate upslope, presently stable channel segments from downslope, recently incised channel segments.

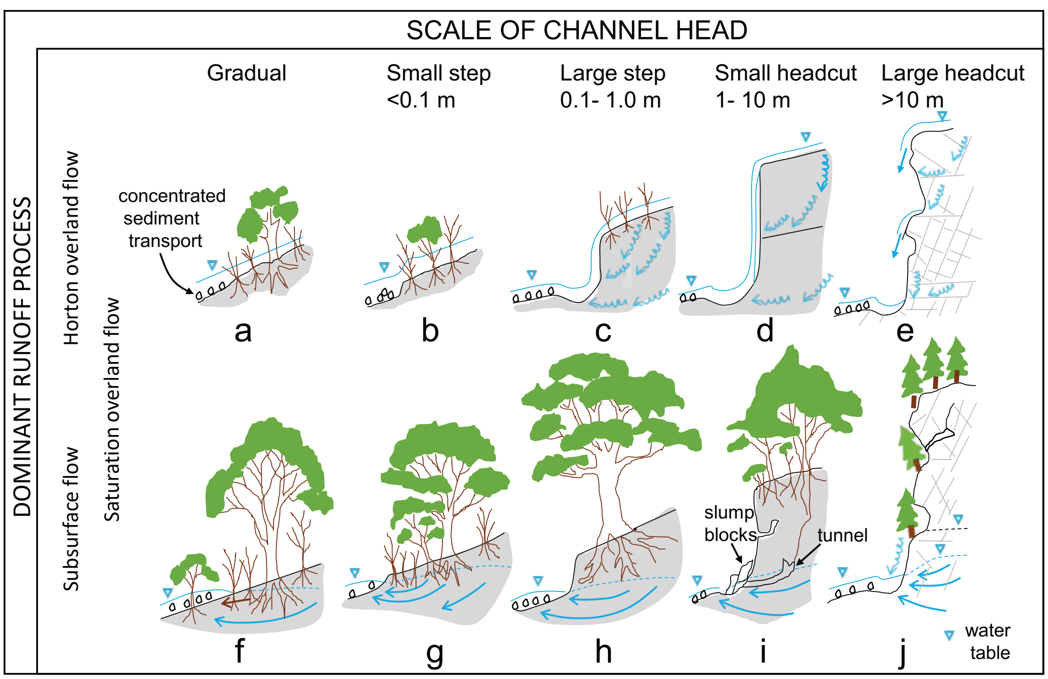
Channel heads that reflect primarily subsurface processes can form via piping or sapping or from shallow landsliding on steep slopes that creates a topographic low where subsurface flow can begin to exfiltrate (Montgomery et al. 2002; Kampf and Mirus 2013). Piping occurs in the unsaturated zone, typically in unconsolidated materials, when preferential flow in conduits erodes or dissolves subsurface materials. Sapping occurs in the saturated zone, which can intersect the surface to form a spring or seep, and can occur in unconsolidated material or bedrock.

4 Identifying Channel Heads in the Field

4.1 Characteristics used

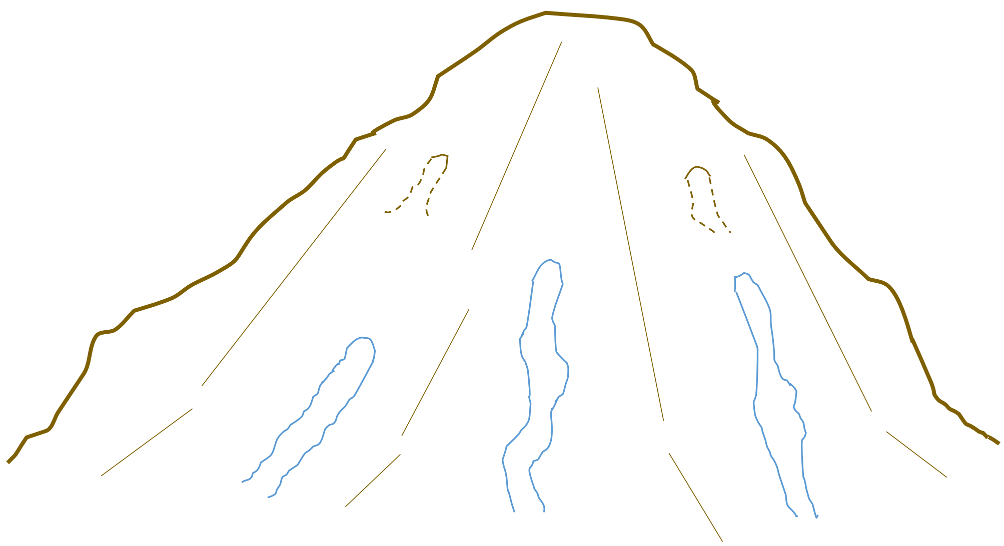
A channel head can be a relatively diffuse feature, the identification of which is subjective, or a very discrete and prominent break in the surface associated with a headcut (Figure 15). The channel downstream from the channel head can be identified based on the evidence of sediment transport (wash marks, small bedforms, and armored surfaces) and observable breaks in slope that define the banks (Dietrich and Dunne 1993). These banks "must be recognizable as morphological features independent of the flow" (Dietrich and Dunne 1993, 178).

Figure 15. Classification of channel heads based on incision depth and the dominant runoff process. Sketches indicate flow paths for Hortonian overland flow and subsurface flow. *Smooth arrows* indicate saturated flow; *wiggly arrows* indicate unsaturated percolation, including flow through macropores. Even at sites with substantial Hortonian overland flow, the face of a large headcut can allow the emergence of erosive seepage. Saturation overland flow drives erosion that includes features from both of the other runoff types. (Adapted by permission from Dietrich and Dunne 1993, Fig. 7.6.)



The most difficult component of identifying a channel head can be deciding which feature to designate as the channel head if several potential channel heads are present along a preferential flow path on a hillslope. In this scenario, it is most appropriate to designate as a channel head the feature farthest upslope that has a longitudinally continuous channel downslope (Jaeger et al. 2007; Jefferson and McGee 2013; Wohl 2013) (Figure 16). As noted earlier, channel heads can also move upslope or downslope through time in response to temporal variations in controlling variables (Kirkby et al. 2003; Wohl and Scott 2017). Distinguishing ephemeral and permanent channel heads during a single site visit to a watershed, however, can be difficult.

Figure 16. Schematic illustration of discontinuous channels and multiple channel heads. In this view toward a steep hillslope, the channel heads that do not give rise to a longitudinally continuous channel downslope are indicated by *brown, dashed lines*. Channel heads that give rise to a continuous channel are indicated by *blue, solid lines*. *Straight brown lines* indicate the trend of the hillslope surface.



4.2 Differentiation of a channel head from other features

Features related to channel heads and that might be confused with channel heads in the field include rills, debris flows scars, and what are designated *erosional features* in a regulatory context. As noted earlier, a rill is a channel with a definable bed and banks created by concentrated water flow, but a rill is a transient feature present only during and briefly after precipitation. Although rills can be confused with a channel head immediately after a storm, particularly an intense storm, a subsequent visit to the field site could help to discern rills from more persistent channel heads. Debris flows can create channel heads, but the highly viscous flow occurring during a debris flow typically leaves distinctive depositional features, including sharp-crested levees along the sides of the channel; deposits in which coarse clasts are not imbricated, sorted, or stratified; and lobate deposits at the downstream end of the flow (Costa and Jarrett 1981;

Waythomas and Jarrett 1994). In contrast, floods are more likely to create flat-topped levees and bars; deposits in which clasts are imbricated, sorted, and stratified; and less-distinct lobate features at the downstream end of the flood zone (Figure 3).

In connection with the Rapanos Supreme Court decision, swales and erosional features (e.g., rills, gullies, and small washes characterized by low volume, infrequent, and short duration flow) are generally not considered waters of the United States in a jurisdictional context because they are not tributaries of larger channels or they do not have a significant nexus to downstream channels traditionally defined as navigable waters. The start of channelized flow with definable bed and banks that does not become tributary to another channel downstream and integrate into a persistent channel network could be defined as an erosional feature. It is not uncommon to find discontinuous channelized segments upslope from a longitudinally continuous channel. The start of each of these discontinuous channelized segments would not be designated as a channel head but rather as an erosional feature.

5 Predicting and Mapping Channel-Head Locations

5.1 Equations

The great majority of investigations focusing on the quantification of channel initiation have been conducted in humid regions. Pioneering, field-based studies of network development indicate that the source area above the channel head decreases with increasing local valley gradient in steep, humid landscapes (5°–45° slopes) with soil cover (Montgomery and Dietrich 1988). For hillslopes of equal gradient, source area can vary in relation to total precipitation or precipitation intensity, as these characteristics influence concentration of runoff (Henkle et al. 2011); drier regions tend to have larger source areas.

Montgomery and Dietrich (1989) used field-mapped channel initiation points in northern California, USA, to develop empirical equations relating source basin length (L), local valley slope (θ, now commonly designated S), and contributing drainage area (A):

$$L = \lambda \tan \theta^{-0.83}, \text{ where } \lambda = 67 \text{ m}$$

$$A = \lambda \tan \theta^{-1.65}, \text{ where } \lambda = 1978 \text{ m}^2$$

$$A = 0.46\ L^{1.99}$$

$$L = 1.48\ A^{0.50}$$

Montgomery and Dietrich (1989) noted an inverse relationship over a wide range of slopes between A and θ at channel heads. Subsequent work continues this trend of developing empirical, site-specific relations between topographic parameters, typically using bounding equations to quantify the range in channel-head locations (Montgomery and Dietrich 1992; Dietrich et al. 1992; Prosser and Abernethy 1996). Some of these investigators also find an inverse relationship between A and θ (Roth et al. 1996; Roth and La Barbera 1997) although the nature of this relationship varies between low-gradient hollows with convergent topography and seepage erosion and steeper topography where channel initiation is more likely to reflect saturation or Hortonian overland flow or landsliding (Montgomery and Dietrich 1989; Montgomery and Foufoula-Georgiou 1993; Prosser and Abernathy 1996).

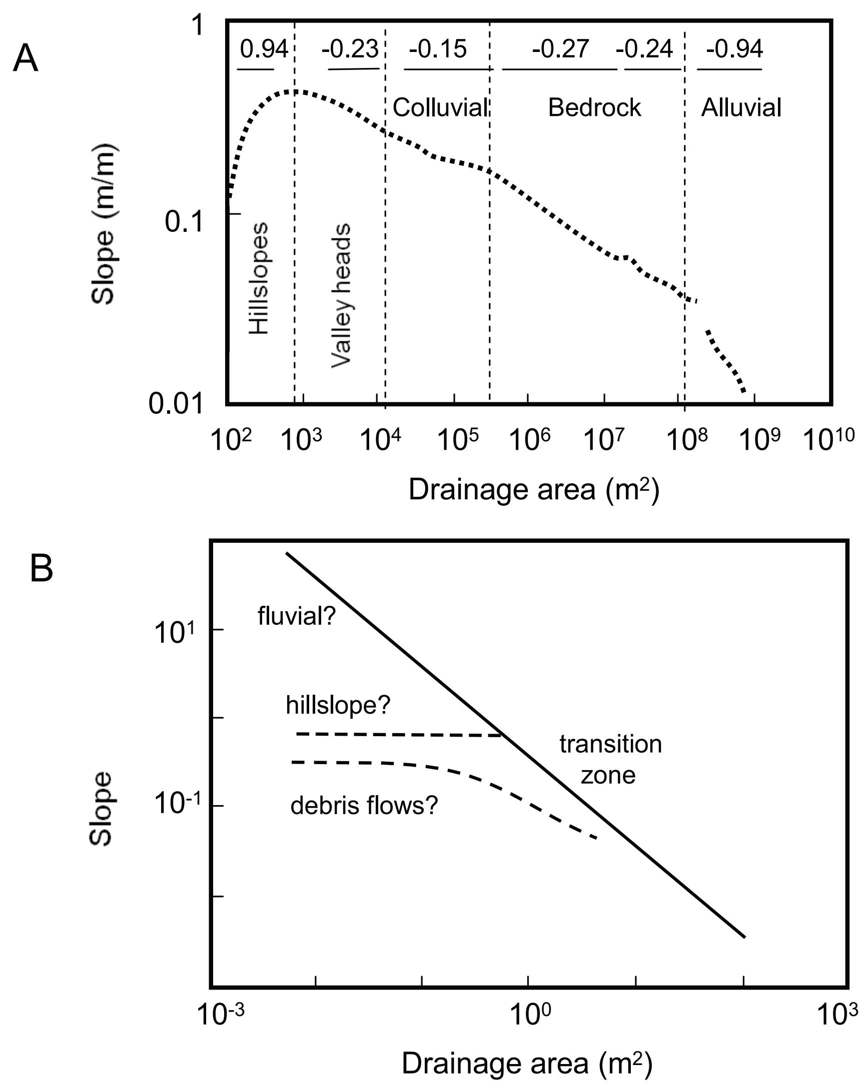
The inverse slope-area (*S-A*) relationship does not always hold for diverse environments (Bischetti et al. 1998; Adams and Spotila 2005) because of differences in runoff processes. Low-gradient hollows with convergent topography and seepage erosion can differ from steeper topography where channel initiation is more likely to reflect Hortonian or saturation overland flow or even landsliding (Montgomery and Dietrich 1989; Montgomery and Foufoula-Georgiou 1993). In terrains with substantial flow through fractured bedrock, bedrock topography is likely to exert a greater influence on channel head locations than does surface topography (Anderson et al. 1997; McDonnell 2003; Adams and Spotila 2005; Jaeger et al. 2007). The role of deeper subsurface flow in fractured bedrock may be particularly critical in arid or semiarid mountainous headwaters where peak runoff can originate from snowmelt over saturated or frozen ground or from rainfall runoff, but base flow comes from groundwater flow paths below the shallow, typically dry soils. Hattanji and Matsushi (2006) found that the *S-A* relation grows less consistent with larger relative groundwater contribution. Heterogeneities in the bedrock, such as spatial variation in joint density, can influence both the location of channel heads and the spatial distribution of channels within a river network, with channels tending to follow more densely jointed bedrock (Loye et al. 2012).

If channel heads form where saturation overland flow exerts a boundary shear stress that exceeds the critical value for substrate erosion, the channel initiation threshold, C, can be expressed as the product of contributing catchment area, A, and hillslope gradient, S (Dietrich et al. 1992; Dietrich and Dunne 1993)

$$AS^{\alpha} \geq C$$

Substantial variability in values of A and S reflects the influence of factors such as vegetation, slope aspect, surface versus subsurface flow paths, and substrate grain size (Montgomery and Foufoula-Georgiou 1993; Prosser et al. 1995; Istanbulluoglu et al. 2002; Yetemen et al. 2010) (Figure 17). Rivenbark and Jackson (2004), for example, find that A varies from 4 to 13 ha in the southern Appalachian Mountains, USA. Limited, unpublished data from an arid region indicate that changes in critical shear stress (for channel initiation via overland flow) arising from differences in ground cover may dominate the proportionality constant between A and θ (Montgomery and Foufoula-Georgiou 1993).

Figure 17. (*A*) Plot of log-bin averaged drainage area versus slope relationship for the Olympic Mountains of Washington, USA. Plot shows the mean slope of individual 10 m grid cells for each 0.1 log interval in the drainage area. Numbers at the top are the exponent for a power function regression of values in the segments of the plot indicated by *horizontal lines* below. *Dashed vertical lines* divide the plot into areas considered to reflect different geomorphic zones of the landscape, or process domains. (Adapted from Montgomery 2001, Fig. 5A). (B) Hypothetical topographic signatures for hillslope and valley processes. Area and slope are measured incrementally up the valley mainstem to the valley head. (Adapted from Stock and Dietrich 2003, Fig. 1a). (Reprinted by permission from Wohl 2010, Fig. 2.4.)

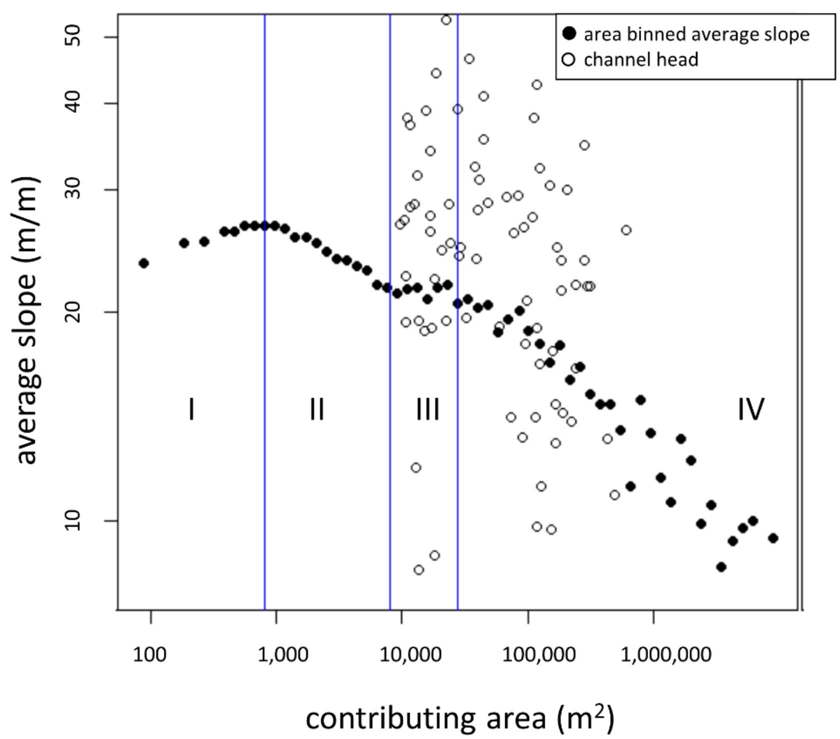


5.2 Remote predictions of channel-head locations

Field mapping of channel head locations is time and labor intensive, particularly in forested terrain where channel heads commonly are not visible except from very close proximity. Consequently, investigators have sought ways to remotely predict the location of channel heads by using either numerical simulations or the average hillslope profile for a watershed or portion of a watershed. A commonly used approach is to extract channel-head locations from digital elevation models (DEMs) by assuming that channel heads correspond to the transition from convex to concave hillslope profiles (Kirkby 1971, 1980; Tarboton et al. 1991, 1992). This transition commonly coincides with the change from divergent to convergent topography and may be more likely to reflect the start of valley development rather than channel heads (Montgomery and Foufoula-Georgiou 1993), as suggested by the substantial scatter in the actual locations of channel heads relative to the location of reversals or inflections in averaged hillslope profiles. Montgomery and Foufoula-Georgiou (1993) proposed that the commonly observed reversal in slope gradient represents the transition from hillslopes to unchanneled valleys and channels dominated by debris flows whereas the next inflection downslope represents the start of alluvial channels. Ijjasz-Vasquez and Bras (1995) designated four regions in hillslope profiles and interpreted the transition between regions I and II to represent the location of channel heads (Figure 18). Despite Montgomery and Foufoula-Georgiou's (1993) warning that "acquisition of even limited field data is recommended," many subsequent investigators have assumed in the absence of field data that channel heads are located at the reversal in slope gradient.

Among the studies that have tested this assumption are Tarolli and Dalla Fontana (2009) and Henkle et al. (2011). Tarolli and Dalla Fontana mapped the location of 30 channel heads in the eastern Italian Alps. They found that channel heads are mostly confined to region II, which supports the assumptions of Ijjasz-Vasquez and Bras (1995), although assuming that channel heads are exactly at the transition between regions I and II underestimates the actual minimum contributing area and overestimates the length of channel networks. The 78 channel heads in the Colorado Front Range mapped by Henkle et al. (2011) plot at the threshold between regions II and III although some extend into region IV (Figure 18). Most actual contributing area values for channel initiation in this semiarid region with snowmelt runoff are thus an order of magnitude larger and located in a significantly different portion of the average hillslope profile than commonly assumed.

Figure 18. Example slope-area plot for hillslopes in the Colorado Front Range, showing the four regions designated by Ijjasz-Vasquez and Bras (1995), who proposed that channel heads are located at the transition between regions I and II, and the location of actual channel heads mapped in the area by Henkle et al. (2011). *Solid circles* indicate average hillslope characteristics in the study, and *open circles* represent mapped channel-head locations. *Vertical lines* indicate transitions between regions denoted by inflections in the hillslope curve. (Adapted by permission from Henkle et al. 2011, Fig. 3.)



Numerical simulations of downslope water and sediment movement can also be used to estimate channel-head locations. Some studies have adapted the widely used TOPMODEL (Sun and Deng 2003; Kim and Lee 2004). Other investigators have developed new modeling approaches such as coupled hydraulics and sediment transport (Simpson and Castelltort 2006; Smith 2010). Lin et al. (2006) and Kim and Kim (2007) developed methods for extracting channel-head locations from DEMs based on *S-A* relations. These studies have limited field verification, instead focusing on sensitivity analyses of differences that result from different thresholds for channel extraction or different assumptions built into a model. Limited validation makes it difficult to evaluate the accuracy of modeling or remote extraction of channel heads against field observations, but the limited field verifications suggest that numerical simulations do not yet accurately reproduce the variation in actual channel-head location observed in field studies. High-resolution, lidar-derived digital terrain models can substantially improve both remote predictions and field mapping of channel heads.

5.3 Field mapping

Field mapping can rely on high-resolution, remotely sensed images or lidar-based, high-resolution DEMs (Heine et al. 2004), where these data exist; but most field mapping requires ground-based mapping and measurements of channel heads. This can be particularly strenuous in high-relief terrain because channel heads can start at different positions on the hillslope, requiring that the field investigator repeatedly walk up- and downslope to locate channel heads. Repeated up- and downslope traverses may also be needed to ensure that a particular channel head has a longitudinally continuous channel downslope. Most field mapping relies on GPS (Global Positioning System) surveying units to locate the geographic coordinates of the channel head (e.g., Jaeger et al. 2007; Henkle et al. 2011; Julian et al. 2012; Wohl 2013) and either field surveys of local hillslope gradient or calculation of slope gradient from DEMs.

5.4 Summary of regional studies where these exist in the U.S.

Only a handful of field-based channel-head studies exist for the United States (Table 1). Montgomery and Dietrich's (1988) pioneering study using field data from northern California and Oregon demonstrated that the source area above the channel head decreases as local valley gradient steepens over a range of 5 to 45 degree slopes. This study also suggested that drier regions tend to have larger source areas for the same hillslope gradient. Montgomery and Dietrich (1989) distinguished between (1) channel heads on steep slopes, which tend to be controlled by subsurface flow-induced instability of the colluvial fill; (2) abrupt channel heads located on gentle slopes and controlled by seepage erosion; and (3) gradual channel heads on gentle slopes governed by saturation overland flow.

Table 1. Field-based studies of channel-head locations in diverse regions of the United States.

Location	Description	Range of Contributing Area (m^2)	Reference
California and Oregon Northern California	Multiple sites; diverse vegetation and lithologies; n = 71 Grasslands; metamorphic lithologies; MAP* = 760 mm; surface and subsurface flow; *n* = 63	1200–40,100	Montgomery and Dietrich (1988, 1989)
Idaho Batholith	Conifer forest; granitic bedrock; MAP = 1000 mm; surface flow; *n* = 27	Not provided in paper	Istanbulluoglu et al. (2002)
Flint Hills, Kansas	Grassland; limestone and shale bedrock; MAP = 820 mm; flow type not specified; *n* = 20	(mean 17,495)	Heine et al. (2004)
Southern Appalachians	Deciduous forest; lithology, climate, and flow type not specified; *n* = 16	26,709–75,272	Rivenbark and Jackson (2004)
Washington	Coniferous forest; sandstone and basalt bedrock; MAP = 2300 to 2800 mm; surface and subsurface flow; *n* = 81	637–60,978	Jaeger et al. (2007)
Northcentral Colorado	Pine and spruce-fir forest; crystalline rocks; MAP = 430 to 1000 mm; surface and subsurface flow; *n* = 78	10,000–600,000	Henkle et al. (2011)
Southern Ohio	Southern Ohio around Cincinnati; diverse land cover, from urban to mature eastern deciduous forest; sedimentary bedrock (sandstone, shale, siltstone, limestone) and glacial till; MAP ≈ 1000 mm; surface and subsurface flow; *n* = 241	300–272,900	Roy et al. (2009)
Mid-Atlantic (5 physiographic provinces)	Predominantly deciduous forest; diverse lithologies; surface and subsurface flow; *n* = 253	2700–793,400	Julian et al. (2012)
North Carolina Piedmont	Mixed hardwood-conifer forest; crystalline lithologies; MAP = 1140 to 1180 mm; surface and subsurface flow; *n* = 100	1000–33,000	Jefferson and McGee (2013)
Northcentral Colorado	Pine forest; crystalline rocks; surface and subsurface flow; *n* = 50	10,400–608,600	Wohl (2013)
Western Colorado	Pinyon-juniper and pine forest; sandstone and shale lithologies; MAP = 290 mm; surface and subsurface flow; *n* = 38	500–494,400	Garrett (2016)

* MAP is mean annual precipitation

Whereas the work in the Pacific Northwest emphasized the influence of hillslope topography on channel initiation, the next field-based channel-head study in the U.S. came from a much drier region and emphasized the influence of local grain size on channel heads. Working in the Idaho Batholith, Istanbulloglu et al. (2002) found that median grain size at each channel head explained a significant portion of the observed variability of *S-A* relations, suggesting that local surface erodibility influences channel-head development. They found that a gamma probability distribution provides a reasonable match to the distribution of *S-A* thresholds measured at the field sites.

Rivenbark and Jackson (2004) and Jaeger et al. (2007) focused on stream heads (the upstream-most location of perennial flow) rather than channel heads. Each study found a range of values for contributing area at the stream head, and Jaeger et al. (2007) noted the lack of systematic *S-A* relations, which they attributed to the controlling influence of bedrock springs. Similarly, Henkle et al. (2011) found only weak *S-A* relations, which they attributed to the mixed population of surface- and subsurface-initiated channel heads and the strong influence of local controls such as bedrock outcrops. Henkle et al. (2011) also compared elevational trends in channel-head locations in the mountainous Colorado Front Range, where increasing elevation equates to greater precipitation and a greater proportion of precipitation in the form of snowfall. They found that channel heads at lower, drier elevations have smaller contributing areas, which is the opposite trend to that noted by Montgomery and Dietrich (1988). This may reflect the fact that the drier sites in Henkle et al. (2011) receive high-intensity convective rainfall that is more likely to create overland flow and thus create channel heads with smaller contributing areas, analogous to the effect of wildfire on allowing channel heads to form at smaller contributing areas (Wohl 2013).

Comparing *S-A* regression lines for data from western Colorado sites with channel heads initiated by surface and subsurface flow and data from studies in other regions, Garrett and Wohl (2017) found significant differences in *S-A* relations between surface and subsurface sites from the same region (Figure 19). However, they found no significant differences in *S-A* relations for channel heads initiate by surface runoff in very different locations (the Appalachian Plateau, central California, western Colorado, north-central Colorado, the mid-Atlantic region, the North Carolina Piedmont, and sites in Australia) (Figure 20). This is surprising given the range of precipitation and hillslope surface characteristics across the diverse study sites. This finding suggests that channel-head locations can be predicted with reasonable accuracy where surface runoff dominates channel initiation, although the envelope curves that define maximum and minimum *S-A* thresholds for each region are not as consistent (Figure 21). A challenge remains in that subsurface processes strongly influence the locations of many channel heads, and the *S-A* relations for these channel heads are much less consistent and predictable.

Figure 19. Plot of contributing area regressed on the local gradient for channel heads in western Colorado. *White triangles* represent channel heads with evidence of subsurface flow initiation; *black circles* represent channel heads with evidence of surface runoff initiation. The *solid line* represents the regression equation for the complete data set (adjusted R^2 = 0.4613); the *dotted line* represents the regression equation for channel heads with subsurface flow (adjusted R^2 = 0); the *dashed line* represents the regression equation for channel heads with surface runoff (adjusted R^2 = 0.5253). (Reprinted by permission from Garrett and Wohl 2017, Fig. 3.)

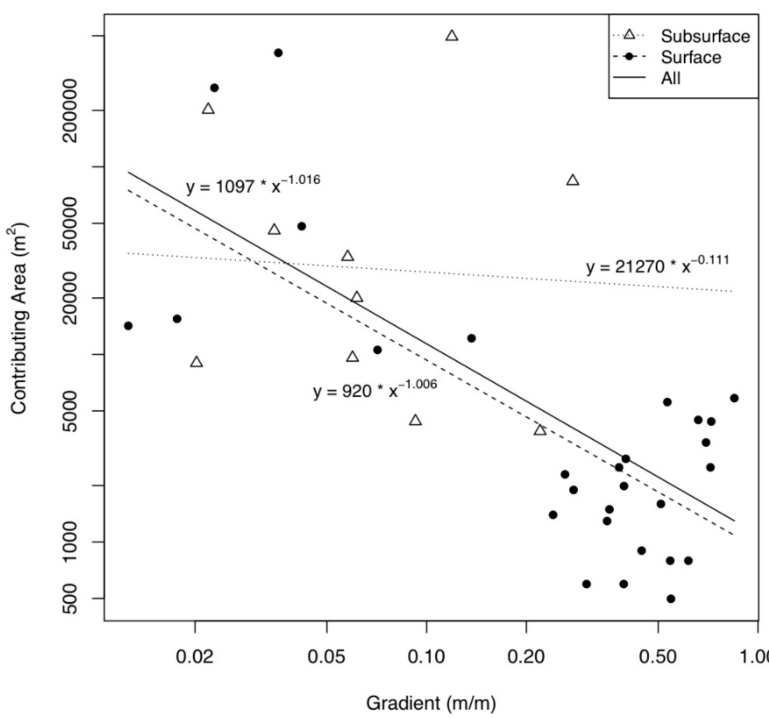


Figure 20. Plot of contributing area regressed on the local gradient for channel heads from multiple datasets. For western Colorado channel heads, *white triangles* represent channel heads with evidence of subsurface flow initiation; *black circles* represent channel heads with evidence of surface runoff initiation. The *solid lines* represent regression equations for all western Colorado channel heads, for channel heads with subsurface flow initiation, and for channel heads with surface flow initiation. (Reprinted by permission from Garrett and Wohl 2017, Fig. 4.)

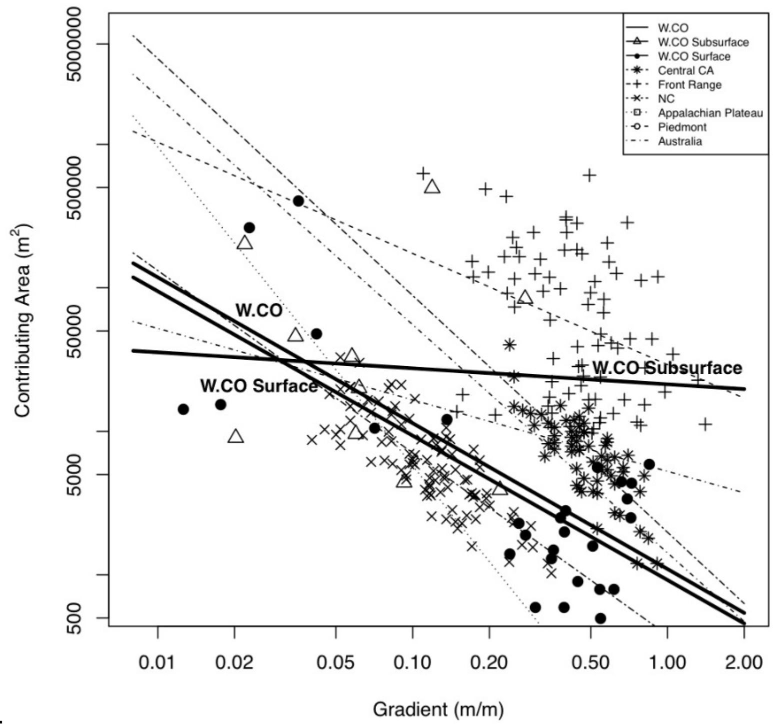
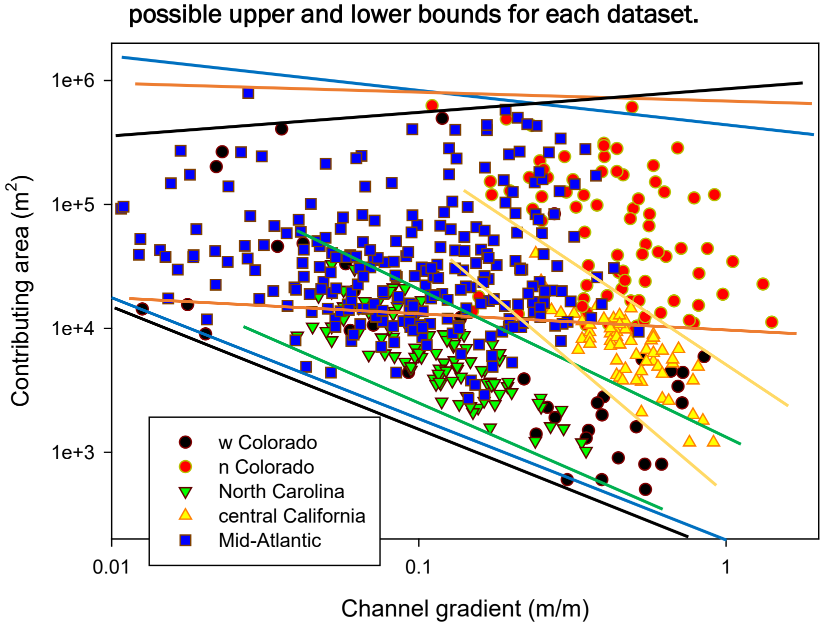


Figure 21. Using the same data as Fig. 20, this plot illustrates possible upper and lower bounds for each dataset.



6 Research Needs

The broad ranges of values for contributing area within even a geographically limited region (Table 1) suggest that indirectly estimating the location of channel heads, rather than mapping each individual channel head in the field, will always result in some uncertainty with respect to channel-head location and contributing area. However, several aspects of channel head identification and the prediction of channel-head locations could be improved with additional research. These include the following:

- Evaluation of the consistency of channel-head identification in the field—No studies have compared the location of channel heads as identified in the field or from remote imagery by different investigators.
- Evaluation of the ability to identify channel heads in relation to spatial resolution of lidar imagery—As relatively high-resolution lidar imagery becomes increasingly common and available, identification of channel heads is likely to increasingly rely on this imagery, but it remains unclear how well channel heads can be identified from meter- versus sub-meter-scale imagery, for example.
- Prediction of the regional characteristics likely to result in channel heads created by predominantly surface versus subsurface flow—The strong association between lithology and flow path (subsurface flow in regions underlain by sandstone and surface flow in regions underlain by shale) in the Garrett and Wohl (2017) study and significant differences in average contributing area in relation to lithology in the Jaeger et al. (2007) study suggest that existing data layers, such as bedrock geology and topography, might be useful in distinguishing which portions of a watershed or landscape are likely to be dominated by surface versus subsurface processes and which portions of a watershed have significant differences in average contributing area.
- Development of envelope curves for different regions—Although the best-fit line for *S-A* relations for different regions in Figure 20 has a consistent slope, the envelope curves that define maximum and minimum *S-A* thresholds for each region are not as consistent (Figure 21), suggesting that additional field research would be useful in constraining maximum and minimum values for distinct regions.

  Targeting specific regions in which little work has been done—Such regions include arid and hyperarid environments, karst and permafrost terrains, and grassland and savanna environments.

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Appendix A: Glossary

Bedding: Depositional units within sediments, may be horizontal or dipping at an angle from the horizontal (cross-beds); presence indicates sediment transport by a current of water or wind

Bioturbation: Physical displacement of sediment as a result of the presence of organisms (e.g., rodent burrowing, growth of plant roots, or tree fall that pulls the tree's root-ball from the ground)

Channel head: The upstream-most point of concentrated water flow and sediment transport between definable banks

Clast: A mineral grain of any size although typically used for gravel size ($\geq$2 mm)

Colluvial hollow: Unchanneled valleys upslope of the channel network

CPOM (coarse particulate organic matter): Particles larger than 1 mm in diameter

FPOM (fine particulate organic matter): Particles 45 µm to 1 mm in diameter

Hyporheic zone: Water beneath a channel and that originates in the channel and returns to the channel with residence times varying from seconds to days

Imbrication: Alignment of clasts with the long axis of each clast parallel to the primary flow direction (edge of clasts sometimes overlap when one end of the long axis dips downward in the upstream direction)

Process domain: A spatially identifiable area characterized by a distinct suite of geomorphic processes

Stream head: The upstream-most point of perennial flow within a river

Topographic relief: The maximum difference in elevation within a defined region, such as a drainage basin (a high-relief area has a larger difference in elevation than a low-relief region)

Zero-order basin: An unchannelized hollow with convergent contour lines

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14. ABSTRACT

A river channel is a linear feature with definable bed and banks created by erosion of water concentrated into persistent flow paths. A river starts at the channel head, which is the upstream-most point of concentrated water flow and sediment transport between definable banks that are spatially continuous downslope. The locations of individual channel heads are difficult to predict because of hillslope-scale differences in gradient, infiltration capacity, porosity and permeability, and cohesion, each of which influences flow paths and erodibility of near-surface materials. Understanding the state of knowledge regarding the initiation point of channels can be useful in a management context when assessing what features on the landscape constitute river channels. Therefore, the primary objective of this report is to concisely summarize the existing state of scientific knowledge regarding where river channels and channel networks begin in the landscape. The report draws on published studies from diverse regions within the United States and around the world but focuses on research summarized in English. This report introduces terms, reviews the processes related to channel initiation in different landscape settings, discusses field and remote identification of channel heads, and outlines the research needs for channel-head identification and for improving channel delineation.

15. SUBJECT TERMS
Channel heads, Hydrology, Ordinary high water mark, River channels--Identification, Stream heads, Watersheds, Wetlands

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