

HILLSLOPE TO FLUVIAL PROCESS DOMAIN TRANSITIONS
IN HEADWATER CATCHMENTS

by

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I can do everything through Christ, who gives me strength. Philippians 4:13

TABLE OF CONTENTS

1. INTRODUCTION	1
Description of the Study Area	9
Dissertation Organization	12
References Cited	21
2. HILLSLOPE AND FLUVIAL PROCESS DOMAIN TRANSITIONS AND SCALING RELATIONS	25
Contribution of Authors and Co-Authors	25
Manuscript Information Page	26
Abstract	27
Introduction.....	28
Methods	34
Contributing Area-flowpath Length Relation.....	38
Contributing Area-unit Stream Power Relation.....	39
Contributing Area-slope Relation.....	39
Results.....	40
Discussion.....	42
Conclusion	49
References Cited	77
3. DETECTION OF HOLLOWES AND CHANNEL HEADS USING SPECTRAL ANALYSIS OF DEMS.....	81
Contribution of Authors and Co-Authors	81
Manuscript Information Page	82
Abstract	83
Introduction.....	84
Methods	88
Field Mapping of Hollowes and Channel Heads.....	88
DEM Pre-processing.....	88
Two-dimensional Discrete Wavelet Transform Filter.....	89
Two-dimensional Discrete Fourier Transform Filter	91
Magnitude-frequency Distributions of Curvature.....	94
Magnitude-frequency Distributions of Derivative of Aspect (dthet)	94
Results.....	95
Discussion.....	98

TABLE OF CONTENTS-CONTINUED

Conclusions.....	102
References Cited.....	143
 4. IN-CHANNEL PROCESS DOMAINS: IDENTIFICATION OF COLLUVIAL TO ALLUVIAL TRANSITIONS USING WAVELET TRANSFORMS.....	 146
Contribution of Authors and Co-Authors	146
Manuscript Information Page	147
Abstract.....	148
Introduction.....	149
Methods	154
Study Area	
DEM Generation.....	157
Hydraulic modeling	158
Wavelet Analysis	160
Planform Analysis	
Results.....	161
Discussion.....	164
Conclusions.....	168
References Cited.....	193
 5. SUMMARY	 196
 REFERENCES	 206

LIST OF TABLES

Table	Page
3.1: Range of wavelengths applied in high pass filter.	94

LIST OF FIGURES

Figure	Page
1.1: Conceptualized headwater catchment (Hack and Goodlett, 1960).....	15
1.2: Shaded relief map of western Montana showing locations of study areas.	16
1.3: Site geologic map, Second and Third Yellowmule Creeks, Big Sky, Montana.	17
1.4: Schematic stratigraphic column, Second and Third Yellowmule Creeks,.....	18
1.5: Site geologic map, Stringer and Spring Park Creeks, TCEF, Montana.....	19
1.6: Schematic stratigraphic column, Stringer and Spring Park Creeks, TCEF, Montana.	20
2.1: Conceptual model of Hack's intra-basin law, contributing area, A versus flowpath length, L (after Dodds and Rothman, 2000). Hillslope is denoted by region m1, where the hillslope length, L1 is the contributing area, A1. Weakly convergent flow of hollows characterizes region m2. Addition of tributaries and change in basin size transition from hollow to colluvial and alluvial channel in regions m3 and m4, respectively.	52
2.2: Conceptual model of rate of change in unit stream power, ω , with increasing contributing area, A (after Brummer and Montgomery, 2003). Transition in rate of change of stream power corresponds to process domain transitions from hillslope to fluvial domain.....	53
2.3: LiDAR-derived hillshade map of Stringer and Spring Park Creeks,.....	54
2.4: LiDAR-derived hillshade map of Second and Third Yellowmule Creeks, ...	55
2.5: Total annual snow water equivalent (SWE), in inches, versus SNOTEL elevation for SNOTEL stations near the study areas.	56
2.6: Field mapped channel heads and hollows, Stringer and Spring Park Creeks (STR, SPC), TCEF, Montana.	57

LIST OF FIGURES-CONTINUED

Figure	Page
2.7: Field mapped channel heads and hollows, Second and Third Yellowmule Creeks (YM2, YM3), Big Sky, Montana. Geologic units: Kmdt, Muddy sandstone and Thermopolis Formation; Km, Morrison Formation; Kk, Kootenai Formation; Kdap, Dacite porphyry.....	58
2.8: Contributing area versus flowpath length, TCEF and Big Sky catchments. m1, m2, and m3denote hillslopes, transitional hollow region, and transitional colluvial and alluvial channel region.	59
2.9: Contributing area versus flowpath length and mapped hollows and heads (STR, SPC, YM2, YM3).....	60
2.10: Contributing area versus unit stream power, TCEF and Big Sky catchments.....	61
2.11: Contributing area versus unit stream power and mapped hollows and heads (STR, SPC, YM2, YM3).	62
2. 12: Contributing area versus slope, TCEF and Big Sky catchments.	63
2.13: Contributing area versus slope and mapped hollows and heads (STR, SPC, YM2, YM3).	64
2.14: Process zones as delineated by flowpath length - contributing area relation, STR.	65
2.15: Process zones as delineated by flowpath length - contributing area.....	66
2.16: Process zones as delineated by flowpath length - contributing area relation, YM2.....	67
2.17: Process zones as delineated by flowpath length - contributing area relation, YM3.....	68
2.18: Process zones as delineated by unit stream power - contributing area relation, STR.	69
2.19: Process zones as delineated by unit stream power - contributing area relation, SPC.	70

LIST OF FIGURES-CONTINUED

Figure	Page
2.20: Process zones as delineated by unit stream power - contributing area relation, YM2.....	71
2.21: Process zones as delineated by unit stream power - contributing area relation, YM3.....	72
2.22: Landscape regions as delineated by slope - contributing area relation, STR.	73
2.23: Landscape regions as delineated by slope - contributing area relation, SPC.	74
2.24: Landscape regions as delineated by slope - contributing area relation, YM2.	75
2.25: Landscape regions as delineated by slope - contributing area relation, YM3.	76
3.1: Example of symmetric (sine) and asymmetric (wavelet) waveforms.....	105
3.2: Normalized spectra of SPC, STR, YM3, YM2 (clockwise).	106
3.3: Hillshade of portion of DEM, Stringer Creek. From left to right: unfiltered DEM, wavelet filtered	107
3.4: Hillshade of portion of DEM, Spring Park Creek. From left to right: unfiltered DEM, wavelet.....	108
3.5: Hillshade of portion of DEM, Second Yellowmule Creek. From left to right: unfiltered DEM, wavelet.....	109
3.6: Hillshade of portion of DEM, Third Yellowmule Creek. From left to right: unfiltered DEM, wavelet.....	110
3.7: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, STR catchment.	111
3.8: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, SPC catchment.	112

LIST OF FIGURES-CONTINUED

Figure	Page
3.9: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, YM2 catchment.....	113
3.10: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, YM3 catchment.....	114
3.11: Magnitude-frequency distributions of dthet for unfiltered, high pass filtered, and wavelet filtered DEMs, STR catchment.	115
3.12: Magnitude-frequency distributions of dthet for unfiltered, high pass filtered, and wavelet filtered DEMs, SPC catchment.	116
3.13: Magnitude-frequency distributions of dthet for unfiltered, high pass filtered, and wavelet filtered DEMs, YM2 catchment.....	117
3.14: Magnitude-frequency distributions of dthet for unfiltered, high pass filtered, and wavelet filtered DEMs, YM3 catchment.....	118
3.15: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, STR catchment.....	119
3.16 : Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, STR catchment.....	120
3.17: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, STR catchment.	121
3.18: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, SPC catchment.....	122
3.19: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, SPC catchment.....	123
3.20: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, SPC catchment.	124
3.21: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, YM2 catchment.	125

LIST OF FIGURES-CONTINUED

Figure	Page
3.22: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, YM2 catchment.	126
3.23: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, YM2 catchment.	127
3.24: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, YM3 catchment.	128
3.25: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, YM3 catchment.	129
3.26: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, YM3 catchment.	130
3.27: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, STR catchment.....	131
3.28: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, STR catchment.....	132
3.29: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, STR catchment.	133
3.30: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, SPC catchment.	134
3.31: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, SPC catchment.	135
3.32: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, SPC catchment.	136
3.33: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, YM2 catchment.....	137
3.34: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, YM2 catchment.....	138

LIST OF FIGURES-CONTINUED

Figure	Page
3.35: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, YM2 catchment.	139
3.36: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, YM3 catchment.....	140
3.37: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, YM3 catchment.....	141
3.38: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, YM3 catchment.	142
4.1: Fluvial system spatial zones (Wohl, 2010).	172
4.2: Channel types in mountain watersheds (Wohl, 2010; Montgomery and Buffington, 1997b).....	173
4.3: Analysis sequence for generation of longitudinal distributions of unit stream power and wavelet analysis.....	174
4.4: Location of Stringer (STR) Creek surveyed reach and inset map showing HEC-GeoRAS cross-sections.....	175
4.5: Location of Spring Park (SPC) Creek surveyed reach and inset map showing HEC-GeoRAS cross-sections.....	176
4.6: Locations of YME and YMW surveyed reaches and inset map showing HEC-GeoRAS cross-sections.	177
4.7 Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, SPC. RAS model derived from total station survey data only. RAS river stationing in meters.....	178
4.8: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, SPC. RAS model derived from 1 m. DEM only. RAS river stationing in meters.....	179

LIST OF FIGURES-CONTINUED

Figure	Page
4.9: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, STR RAS model derived from last returns and total station survey data, downstream-most 100 meters only. RAS river stationing in meters.	180
4.10: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, STR RAS model derived from 1 m. DEM only. RAS river stationing in meters.....	181
4.11: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, YME RAS model derived from last returns and total station survey data. RAS river stationing in meters.....	182
4.12: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, YMW RAS model derived from last returns and total station survey data. RAS river stationing in meters.	183
4.13: Power spectra of planform (deviations from valley centerline), SPC. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.....	184
4.14: Power spectra of planform (deviations from valley centerline), STR. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.....	185
4.15: Power spectra of planform (deviations from valley centerline), YME. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.....	186
4.16: Power spectra of planform (deviations from valley centerline), YMW. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.....	187

LIST OF FIGURES-CONTINUED

Figure	Page
4.17: Planform series: centerline deviation versus valley distance, cubic fit overlaid.	188
4.18: Power spectra of planform (deviations from valley centerline) after cubic detrending, SPC. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.	189
4.19: Power spectra of planform (deviations from valley centerline) after cubic detrending, STR. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.	190
4.20: Power spectra of planform (deviations from valley centerline) after cubic detrending, YME. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.	191
4.21: Power spectra of planform (deviations from valley centerline) after cubic detrending, YMW. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.	192
5.1: Conceptual model of the formation of the study headwater catchments, Big Sky and TCEF, Montana.	205

ABSTRACT

The landscape is partitioned into hillslopes and unchanneled valleys (hollows), and colluvial (hillslope controlled) and alluvial (self-formed) channels. The key issue for any study of headwater catchments is the rational distinction between these elements. Accurate identification of process domain transitions from hillslopes to hollows, hollows to colluvial channels and colluvial to alluvial channels, are not obvious either in the field or from topographic data derived from remotely sensed data such as laser derived (LIDAR) digital elevation models. The research in this dissertation investigates the spatial arrangement of these landforms and how hillslope and fluvial process domains interact in two pairs of headwater catchments in southwest and central Montana, using LIDAR data. This dissertation uses digital terrain analysis of LIDAR-derived topography and field studies to investigate methods of detection, modeling, and prediction of process transitions from the hillslope to fluvial domains and within the fluvial domain, from colluvial to alluvial channel reaches. Inflections in the scaling relationships between landscape parameters such as flowpath length, unit stream power (a metric of the energy expended by the channel in doing work), and drainage area were used to detect transitions in flow regimes characteristic of hillslope, unchanneled valleys, and channeled landforms. Using the scale-invariant properties of fluvial systems as a threshold condition, magnitude-frequency distributions of curvature and the derivative of aspect were also used to detect hillslope, fluvial, and transitional process domains. Finally, within the classification of channeled landforms, the transition from colluvial to alluvial channels was detected using the presence/absence of repeating patterns in the power spectra of fluvial energy and channel form parameters. LIDAR-derived scaling relations and magnitude-frequency distributions successfully detected and predicted locations of mapped channel heads and hollows and spatial regions of process transitions. Subreaches of arguably alluvial channel conditions were also identified in power spectra. However, extrinsic forcing limits ability to detect a clear transition from colluvial to fully alluvial conditions. Headwater catchments present a mosaic of process domains, in large determined by local structure and lithology. However, process domain transitions appear detectable and statistically, though not deterministically, predictable, irrespective of setting.

CHAPTER 1

INTRODUCTION

Headwater catchments and headwater streams have been described by multiple definitions based upon their physiographic, hydrologic, or ecologic characteristics. These definitions rely upon attributes such as perennial versus ephemeral flow, fish-bearing versus non-fish bearing, and steep versus lower gradient topography, and have largely been developed in the temperate and coastal forests of Washington, Oregon, and northern California (Benda et al., 2005; Danehy and Ice, 2007). The characteristics and dynamics of headwater systems of the intermountain West are distinct due to differences in precipitation patterns, vegetation communities, physiography, and geomorphic history, and consequently are more poorly understood.

Nevertheless, there are definitions and attributes of headwater systems common to both the Pacific Northwest and intermountain West. The term "headwaters" implies upstream-most extent or "fringe" (Richardson and Danehy, 2007) of the channel network. From this perspective, headwater catchments begin at the drainage divide separating one watershed from another, and are comprised of an assemblage of four landscape units: hillslopes, zero-order basins, "temporary" or discontinuous channels and perennial first- and second-order streams (Hack and Goodlett, 1960). This definition draws upon the Strahler-Horton stream ordering system to define basins and channels. In this system (Strahler, 1957), order is defined by the number of tributaries entering a given link of the network. If there are no tributaries upstream of the link, it is termed first-order. The

channel link downstream of the confluence of two first order links is termed second-order. The hollow or unchannelized valley upstream of a first-order channel is termed a "zero-order" basin.

The upper bound of the headwater catchment is easier to define than its downstream extent. Current definitions rely on stream order (Benda et al., 2005), catchment size and bankfull width (Richardson and Danehy, 2007), aquatic-terrestrial interactions (Nakano and Murakami, 2001), channel morphology (Montgomery and Buffington, 1997a; Hassan et al., 2005), annual streamflow and bankfull width (Danehy and Ice, 2007), fish presence (Richardson et al., 2005), vulnerability to disturbance regimes (Swanson et al., 1998) and dominant geomorphic processes (Whiting and Bradley, 1993).

Process-based classifications are arguably the most appropriate means of delineating the downstream extent of headwater catchments. Whiting and Bradley (1993) use the degree of interaction between hillslopes and channels and the mode of sediment transport to define process domains such as debris flow susceptibility and hillslope stability. Schumm (1977) partitions the watershed into sediment production, transfer, and deposition zones. Schumm's Zone 1 corresponds to the upper portion of the watershed system and functions as the sediment supply for downstream reaches. Production is predicated on the ability of disturbances (mass wasting, biological) to be readily transmitted from the hillslopes to the valley and channel. However, headwater catchments also supply water and organic and woody debris as well as sediment due to the high degree of coupling between hillslopes and channels.

Other definitions rely upon the character of the headwater channels to define the downstream extent of the headwater catchment. However, a headwater stream is not a smaller version of the downstream channel. Headwater streams are distinguished by their strong interaction with the adjacent landscape. Within the headwater system, both the channel and valley morphology is influenced by hillslope processes. In humid, temperate forests, the dominant hillslope processes are debris flows and landslides, in part because the inherited valley morphology ("V"-shape) dictates a high degree of hillslope and valley coupling. However, in semi-arid regions characteristic of the intermountain West, hillslope process is often characterized by slower processes of creep, and fire-facilitated debris flows. Snow avalanches may also activate hillslope erosion sources as they share attributes of steep slopes and topographic convergence with other mass wasting processes. In the intermountain West, processes are modulated by high elevation climatic and vegetation drivers such as accumulation, duration, and runoff of snowpack and the type and extent of the alpine vegetation assemblage. Geomorphic history also modulates hillslope and channel interaction through an often inherited glacial ("U"-shape) valley morphology. Relict glacial landforms can drive channel initiation and channel type by imposing local slope and the degree of hillslope coupling (Brardinoni and Hassan, 2007). Brardinoni and Hassan (2007) found that the arrangement of cirques, hanging valleys, and valley steps in headwater catchments of southwestern British Columbia dictated local slope and consequently where channels begin, as well as the type of channels (colluvial versus alluvial) likely to be found. Continental glaciation also leaves a strong topographic signature found in headwater catchments. In Massachusetts, Brooks and

Colburn (2011) found headwater catchments exhibiting a deranged network pattern with channels periodically interrupted by wetlands or ponds and incised channel reaches alternating with reaches of wetland and/or subsurface flow zones. The sediment production properties characteristic of headwater catchments are partially determined by primary controls of lithology and structure. Bedrock resistance and the orientation of units was found to exert first-order control on channel initiation and network topology in the Appalachian Mountains (Golden and Springer, 2006; Kavage-Adams and Spotila, 2005). Subsurface properties such as bedrock joint density, shallow subsurface flow, and weathering rate may also be primary drivers on initiation of headwater streams. The interaction of subsurface properties and surface topography was found to be the dominant control on channel initiation in the Colorado Front Range (Henkle et al., 2011). In non-mountainous regions, subsurface effects such as seepage erosion and groundwater sapping can dominate the development of headwater streams. Laity and Malin (1985) hypothesized that tributary canyons to the Colorado River formed entirely by subsurface means: infiltration of precipitation, collection on aquitard interbeds, and exfiltration at scarps on the interbeds. Exfiltration of subsurface water promotes weathering and mass wasting of the scarp, followed by transport of weathered debris. Headwater streams in Missouri frequently appear as low-lying grassy fields or swales lined with organic debris (Zale et al., 1989). These features flow intermittently often by groundwater exfiltration, and act as a conduit for the transport of organic material and nutrients. Because streams are widely spaced in prairie systems, headwater streams provide critical riparian habitat (Zale et al., 1989).

Headwater channels begin where surface runoff is sufficiently concentrated to cause scour and distinct banks (Dietrich and Dunne, 1993). Channels are distinct from hillslope rills in their occurrence in convergent topography and their spatial persistence (MacDonald and Coe, 2007). In humid environments, they begin in hollows, which drain an upslope source area (Figure 1.1) (Hack and Goodlett, 1960; Montgomery and Dietrich, 1989). Both channels and hollows are convergent landforms, but channels have banks defined as "steepened" (Montgomery and Dietrich, 1989), "well-defined" and "measurable" (Montgomery and Dietrich, 1989), and "defined" (Gomi et al., 2002; Richardson and Danehy, 2007; Bryant et al. 2007). A definable bank "must be recognizable as a morphological feature independent of the flow" (Dietrich and Dunne, 1993, p. 178) with characteristics such as "a narrow zone of gradient steeper than both the neighboring hillslope and the transverse slope of the channel bed" (Dietrich and Dunne, 1993, p. 178). Typical in-channel sediment transport signs include bedforms and armored surfaces (Dietrich and Dunne, 1993).

The channel head has a unique function as the transition between hillslope and fluvial process domains. However, these transitions are spatially and temporally heterogeneous, creating a patchy landscape of hillslope-dominated and fluvially-dominated process domains. Both hillslope and fluvial process domains may create channels by debris flow or landslide initiation, surface flow, or subsurface flow.

The channel head is a threshold feature defined as the "upslope limit of erosion and concentration of flow within steepened banks" (Montgomery and Dietrich, 1989, p. 1909). This definition is based upon morphology rather than the variable characteristics

of flow. The channel initiates where driving forces of flow-induced shear stress or mass wasting-driven scour exceed the strength of the underlying soil and colluvium.

Intrinsic properties of the catchment such as slope, source area, infiltration capacity, soil, regolith, and bedrock permeability and porosity exert control on the point of channel initiation. These factors interact to define the dominant runoff mode, from overland flow to subsurface saturated overland flow. Lithology also controls channel initiation by determining the type and rate of weathering, predisposing channel initiation by either surface erosion or subsurface-facilitated spring sapping. The point of initiation is also variable in space and time, dependent on the probability distribution of disturbances including large floods and mass wasting events. External factors such as land use and management and climate can also effect the location of the channel head, causing either erosion or deposition, which can propagate to the remainder of the watershed (Dietrich and Dunne, 1993)

The mode of channel initiation and the consequent channel head morphology results from the combined modes of hillslope runoff generation and hillslope erosion. Hillslope runoff generation may take the form of Hortonian overland flow, subsurface flow, or saturated overland flow, dependent on lithology, vegetation, weathering mechanism, and precipitation form. At the headwaters, runoff is less likely to be buffered in riparian areas and hence is more variable. The mode of hillslope erosion associated with channel initiation has been correlated with gradient and contributing area, as well as mode of runoff generation (Dietrich et al., 1992). Although channel heads are typically downslope of hollows, they can be related to larger controls of lithology and

bedrock structure. Channel heads may occur at the contact of geologic units of differing permeability, or at the convergence of shallow, macropore flowpaths.

Headwater streams are, by definition, at the upstream extent of the channel network and represent a transition in physical processes, morphologic characteristics, and ecological communities (May, 2007). They comprise 60-80% of the total stream length in a watershed (Schumm, 1956; Shreve, 1969) consequently the topology and density of the channel network profoundly influences how and when catchments route water, sediment, and wood debris to downstream reaches. The rate and magnitude of water, sediment, and wood debris routing has a significant influence on the physical habitat and water quality of downstream fish-bearing reaches (Danehy and Ice, 2007).

Downstream of the channel head, the channel may be discontinuous, ephemeral, or perennial. The terms "ephemeral" and "perennial" refer to the constancy of flow throughout the year while "discontinuous" refers to maintenance of channel form. Discontinuous channels alternate between aggradation and degradation (Bull, 1997). These zones are typically separated by a channel fan (Bull, 1997), a thin and narrow accumulation of alluvium. Locations of aggradation and degradation zones are temporally and spatially variable, reflecting changes in upstream sediment production and transport.

The downstream extent of headwater channels has been defined as the colluvial-alluvial transition zone, where alluvial processes become the dominant control on channel morphology (MacDonald and Coe, 2007). Montgomery and Buffington (1997a) describe the downstream extent of headwater systems in terms of properties of the channels.

Colluvial, cascade, and step-pool channel types are often imprinted by hillslope process domain as hillslopes are their dominant sediment source (Montgomery and Buffington, 1997a). Headwater channels are typically supply-limited with a high transport capacity, and found in colluvium-filled valley segments.

In headwater catchments, channel morphology is influenced by the hillslope (mass wasting) domain, fluvial domain, or both. Channels that lie in material derived from hillslopes (colluvium) are termed “colluvial.” Colluvial channels are characterized by randomly deposited rock and wood debris delivered from hillslopes by mass wasting. They lack the fluvial energy to create their own bedforms and architecture. Lower in the watershed, alluvial streams lie in their own deposits, which thus can be transported to create forms. With increasing drainage area, increasing fluvial energy enables alluvial channels to form. Alluvial channels dissipate energy transporting and rearranging their own deposits to create distinctive patterns, or repeating signatures in either bedform architecture or planform.

The transition from the hillslope to the fluvial domain is a critical threshold for drainage basin evolution but quantitative prediction of this transition is limited by the interaction of multiple variables and processes. Basin morphometry and lithology define intrinsic controls such as slope, source area, infiltration capacity, porosity, and weathering products. These in turn influence the mode of surface and subsurface processes of runoff and surface erosion. Extrinsic controls of climate and land use further modulate surface and subsurface processes. How colluvial channels transition to alluvial channels in headwater catchments is subject to multiple processes and controls.

The mass wasting and erosive processes that create channels have been identified and observed in the field (Dietrich and Dunne, 1993; Benda et al., 2005), but how these newly created channels transition from hillslope forcing to fully alluvial conditions is poorly understood.

This dissertation uses digital terrain analysis and field studies of topography to investigate methods of detection, modeling, and prediction of process transitions from the hillslope to fluvial domains and within the fluvial domain, from colluvial to alluvial channel reaches. Accurate identification of these transitions is key to understanding the formative processes of headwater channels.

Description of the Study Area

Four headwater catchments in two study settings in southwest and central Montana were used in this study: near Big Sky, Montana (45.185° N, 111.421° W) and in the U.S. Forest Service Tenderfoot Creek Experimental Forest (TCEF) (46.933° N, 110.880° W) (Figure 1.2). One pair of catchments in each study setting was selected to compare the range of natural variability within adjacent catchments sited in the same physiographic, climatic, and lithologic region. The catchment pairs were also selected to compare the variation in process domain transitions between two different settings with similar lithology but different climate and aspect.

In southwest Montana, the headwaters of Second and Third Yellowmule Creeks (upper approximately 4 sq. km. of each catchment) were studied. Elevation within the study areas ranges from 2600 to 2800 m. Second and Third Yellowmule Creeks are

located in the Madison Range, south of Big Sky, Montana and are tributary to the West Fork of the Gallatin River. These catchments have a north aspect, and flow from south to north. Their headwaters are formed by the Buck Creek anticline, which trends northwest. The Buck Creek anticline is genetically related to the Buck Creek fault, immediately to the south. This structure was likely formed during the Precambrian and reactivated during Laramide time (Tysdal et al., 1986). The upper catchments form on the gently dipping (5-10 degrees) northern limb of the anticline. This area may have supported small alpine glaciers but was certainly subject to periglacial activity during Pleistocene time (Locke, pers. comm.). Several post-glacial landslides and debris flows are observed along the steep east and west margins of the catchments

During Cretaceous time (~100 Mya), the region was covered by the Western Interior Seaway. Carbonate (limestones) and shale units were deposited in deep water environments. The shorelines varied with tectonic activity and deposition of sediments shed from uplifted terrain, creating variable shallow water, lotic, and deltaic environments. Limestones, sandstones, and siltstones were deposited in the shallow water environments. The bedrock of Second and Third Yellowmule Creeks consists of sandstones, siltstones, mudstones, and shales deposited in this variable environment (Figures 1.3 and 1.4). Specifically, the catchments are primarily Kootenai Formation: shale, mudstone, siltstone, and ledge-forming sandstone deposited in braided to meandering alluvial systems and fresh to brackish water lakes on a coastal plain (Suttner et al., 1981). The upper portion (10-15 m.) of the Kootenai Formation has been mapped as oolitic limestone (Kellogg and Williams, 2005) but none was observed in the study

area. The catchments are rimmed by Muddy Sandstone and Thermopolis Formation, clayey sandstone, and fissile, poorly indurated shale, deposited in deeper water environments (Kellogg and Williams, 2005).

In central Montana, the headwaters of Spring Park and Stringer Creeks (upper approximately 4 sq. km. of each catchment) were used as study areas. Elevation within the study areas ranges from 2170 to 2330 m. These catchments are part of the U.S. Forest Service Tenderfoot Creek Experimental Forest (TCEF), located north of White Sulphur Springs, Montana, in the Little Belt Mountains. Both streams are tributary to Tenderfoot Creek, which is tributary to the Smith River.

These catchments have a south aspect, and flow from north to south. During Pleistocene time, this area was subject to periglacial activity. Post-glacial-age debris flows have since occurred (Locke, pers. comm.) and floor the valley. Both catchments are underlain by a nearly flat-lying Cambrian sedimentary rock sequence consisting of Flathead Formation below Woolsey Shale (Reynolds, 1995). Portions of the uppermost part of the catchments are capped by Eocene quartz monzonite (Reynolds and Brandt, 2007), which imparts the greatest degree of relief to the catchments (Figure 1.5).

The Flathead Formation is pebbly quartz sandstone deposited in shallow, near-shore environments adjacent to the middle Cambrian seaway. Woolsey Shale was deposited in deeper water environments. In this part of the Little Belts, the Flathead Formation consists of a thin sandstone member overlying a "fissile, shaly" member, overlying another sandstone member (Weed and Pirsson, 1899) such that the formation has three

members. The study areas are primarily underlain by the upper sandstone and middle shale members of the Flathead Formation (Figure 1.6).

Although both the southwest and central Montana catchment pairs are characterized by gently dipping to flat shale units, the strength of the shale units is quite different. Cretaceous shales in southwest Montana have not been buried deeply enough to destroy all expandable clay layers, and tend to be softer, more prone to water absorption, and more friable. Cambrian shales have less expandable clay layers, and have been buried deeper and longer, rendering them more indurated due to burial near-metamorphism (Schmitt, pers. comm.) Both Cretaceous and Cambrian shales may still produce debris flows but Cambrian shales are less prone to fail and more likely to resist fluvial incision.

Dissertation Organization

This dissertation is organized into three chapters. Chapter 2, "Hillslope and fluvial process domain transitions and scaling relations", examines the use of landscape scaling relations to identify where hillslopes transition to hollows and where hollows transition to channels. Specifically, the log-log relations between contributing area and flowpath length, unit stream power, and gradient, as extracted by digital elevation models (DEMs), are found to exhibit inflections that are hypothesized to correspond to hillslope and process domains. The signatures of diffusive flow, concentrated flow, and mass wasting-generated incision are assumed to have a topographic signature, detectable in topographic data. This chapter investigates the use of DEMs over an entire catchment to generalize process domain transitions in individual mapped hollow and channel heads of

the headwater network. This approach relies on the applicability of generalized relations (by log-bin averaging) to specific topographic and lithologic conditions.

Chapter 3, "Detection of hollows and channel heads using spectral filtering of DEMs", attempts detection of hollows and channel heads using power law behavior of topographic parameters derived from spectrally filtered DEMs. This chapter uses the spectral signature of channelized and convergent landforms to filter unchannelized and divergent portions of the landscape from the catchment DEMs. Curvature and aspect of the remaining DEM cells are calculated and used to generate magnitude-frequency distributions. The transition to power law behavior in these magnitude-frequency distributions implies a fractal structure, or scale-invariance, commonly associated with channel networks (Birniir et al., 2007). This property is used to detect process domain transitions.

Chapter 4, " In-channel process domains: identification of colluvial to alluvial transitions using wavelet transforms", extends the previous two chapters by examining the transition from colluvial to alluvial process in headwater channel reaches. The transition from colluvial to alluvial process can be suggested by the presence of patterns, defined as widespread, repeating signatures in topography (Perron, 2011), in alluvial reaches and randomness in colluvial reaches. Similar to the association of power law behavior with fractal structure and scale-invariance, organization, as defined as the presence of patterns, is associated with alluvial process.

Chapter 5 summarizes the main findings of each chapter of this dissertation and discusses their implications. This dissertation addresses transitions in process domains in

headwater catchments, an important question for landscape evolution and watershed management. Each chapter demonstrates a different lens through which to view process domain transitions, despite the multiplicity of processes and controls at work in headwater catchments.

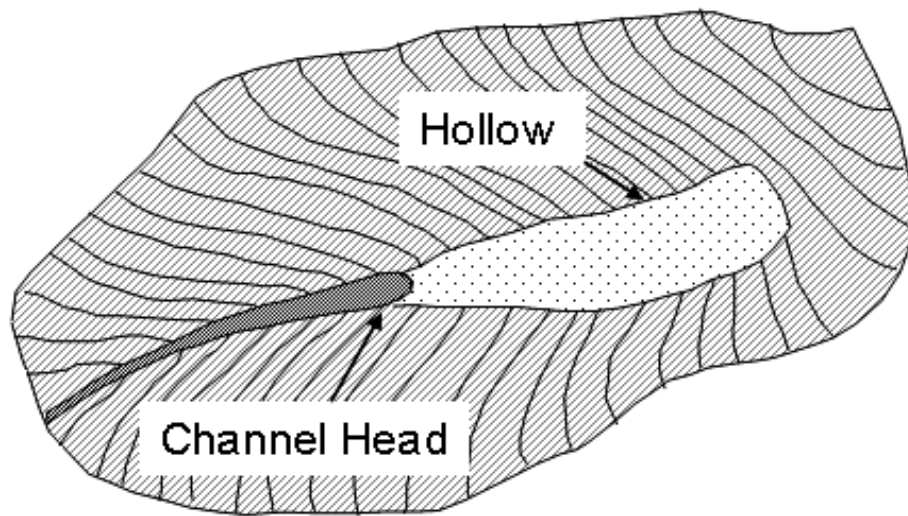


Figure 1.1: Conceptualized headwater catchment (Hack and Goodlett, 1960).

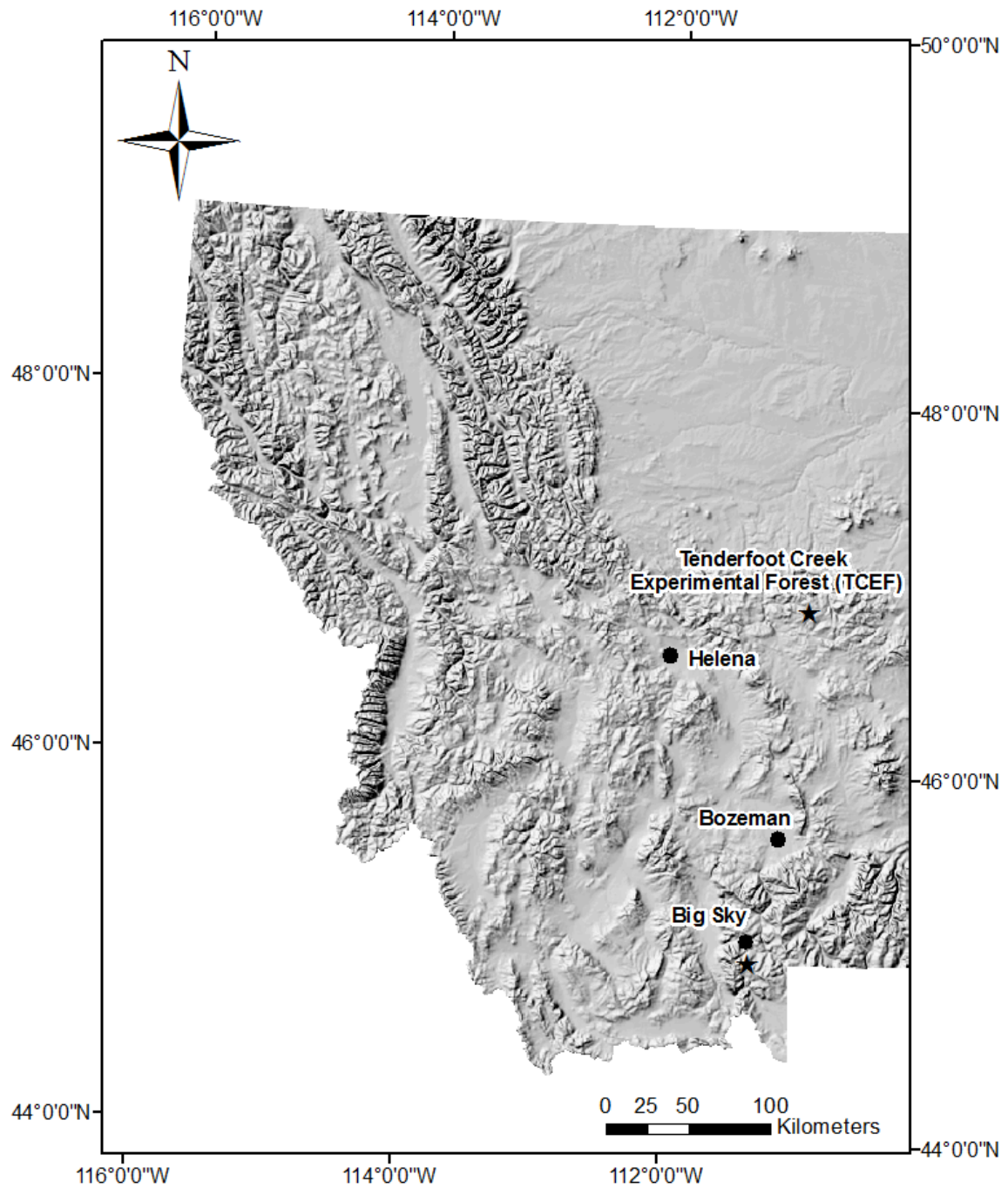


Figure 1.2: Shaded relief map of western Montana showing locations of study areas, marked by stars.

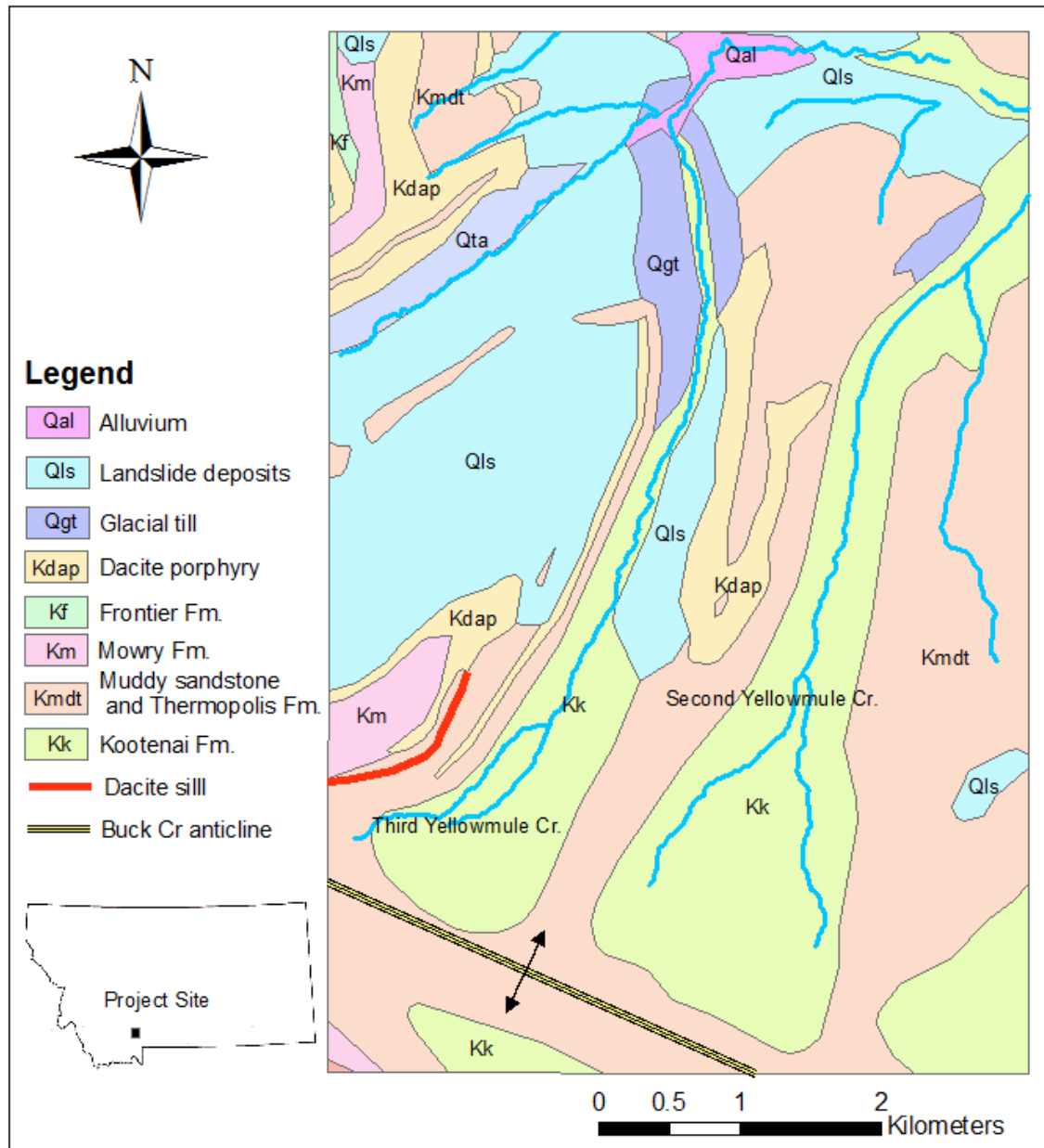


Figure 1.3: Site geologic map, Second and Third Yellowmule Creeks, Big Sky, Montana (Kellogg and Williams, 2000).

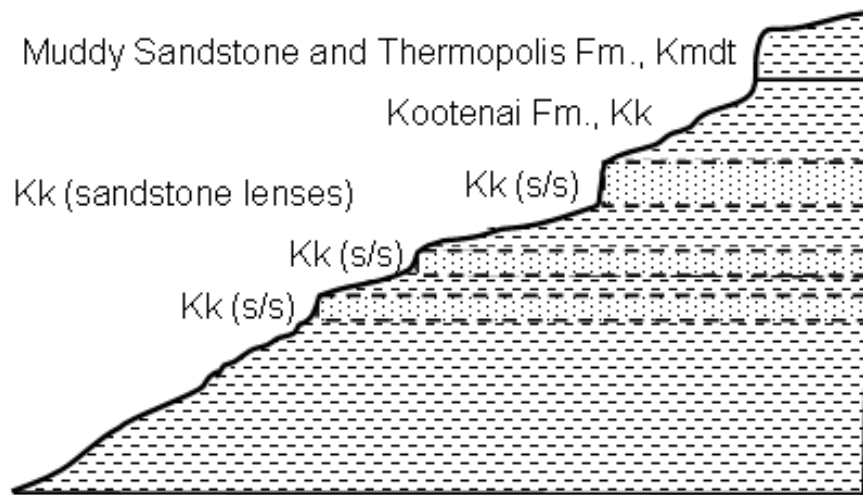


Figure 1.4: Schematic stratigraphic column, Second and Third Yellowmule Creeks, Big Sky, Montana

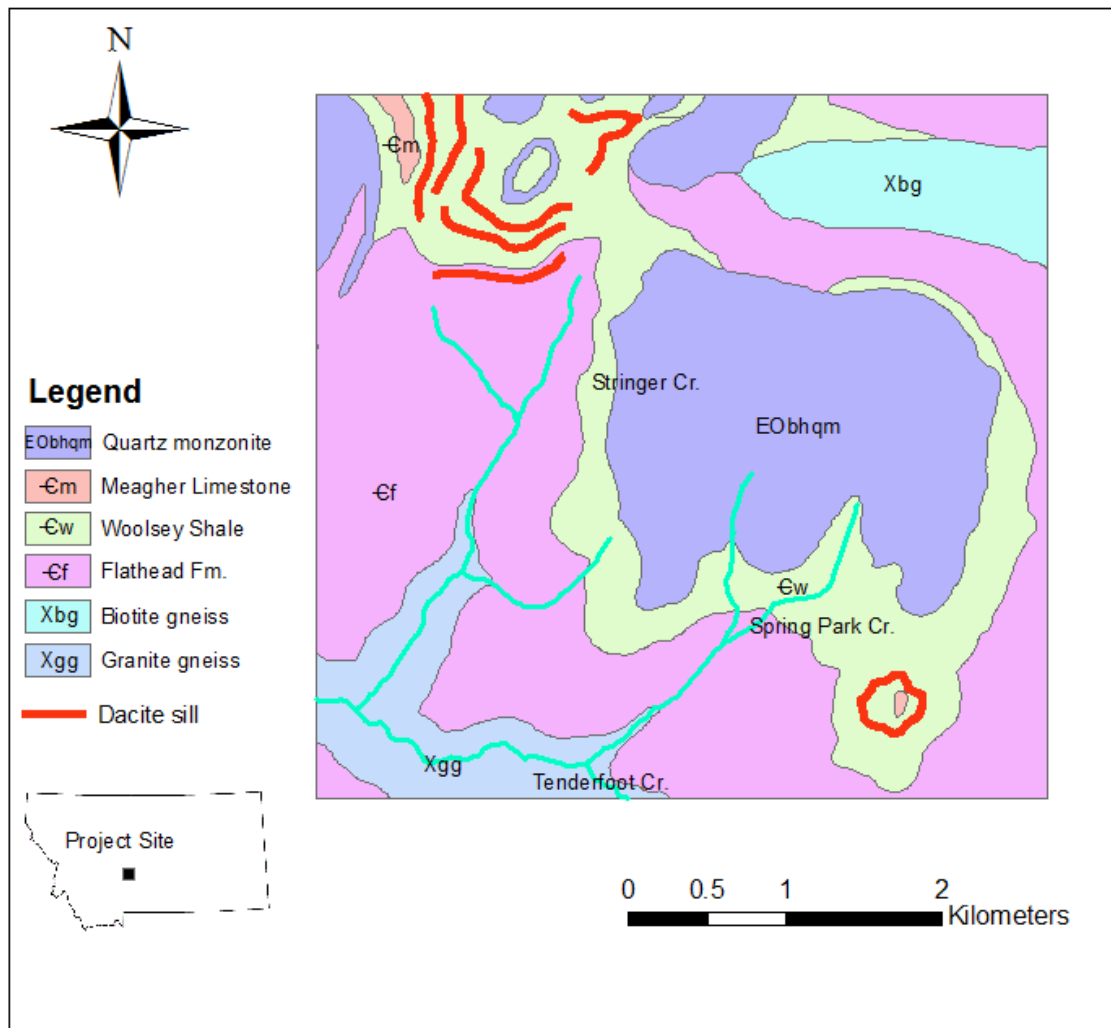


Figure 1.5: Site geologic map, Stringer and Spring Park Creeks, TCEF, Montana (Reynolds and Brandt, 2007).

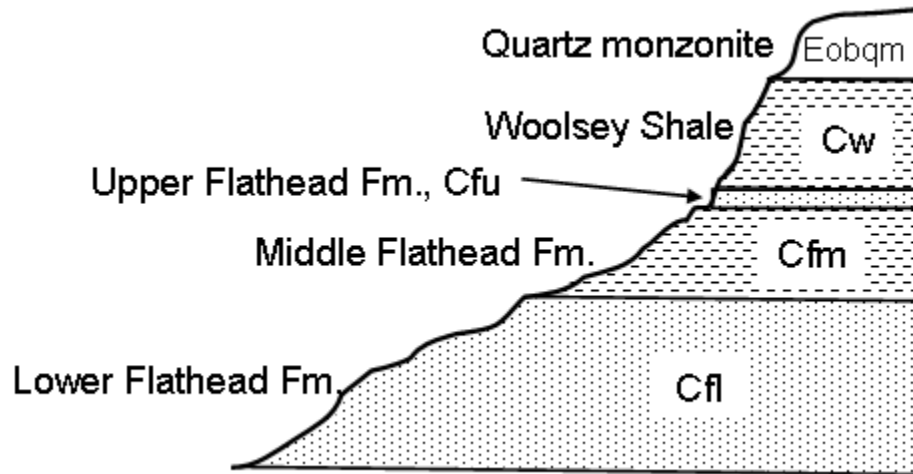


Figure 1.6: Schematic stratigraphic column, Stringer and Spring Park Creeks, TCEF, Montana.

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CHAPTER TWO

HILLSLOPE AND FLUVIAL PROCESS DOMAIN TRANSITIONS
AND SCALING RELATIONS

Contribution of Authors and Co-Authors

Manuscript in Chapter 2

Author: Karen M. Williams

Contributions: Conceived the study, collected and analyzed output data, and wrote the manuscript.

Co-author: William W. Locke III

Contributions: Assisted with study design and discussed the results and implications and edited the manuscript at all stages.

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Abstract

Accurately tessellating the landscape into hillslopes, hollows and channels is important for calibration and verification of hydrologic runoff and sediment production models. Methods of detection of topographic features using high resolution topographic data are increasing in number and use, facilitated by the increasing availability of airborne Light Detection and Ranging (LIDAR) data. We test two types of landscape scaling relations to detect and predict hillslope to fluvial domain transitions using LIDAR-derived digital elevation models (DEMs) of a pair of headwater catchments in southwest Montana and a second pair of catchments in north-central Montana. Inflections in scaling relations of contributing area and flowpath length in a single basin (a modified Hack's law) and contributing area and unit stream power were used to identify process domain transitions. In this use of Hack's law, the exponent of the power law is hypothesized to transition from near one at the hillslope scale to transitional values, ranging from near 0.4 to 0.6, as colluvial and alluvial channels begin. Similarly, the rate of change of unit stream power with contributing area is hypothesized to become constant and then decrease at the hillslope to fluvial domain transition. Locations of process domain transitions identified from scaling relations were validated by field mapping of hollow and channel heads. Results suggest that the inflections in scaling relations roughly correspond to process domain transitions, and regions between slope breaks indicate landform type and their spatial extent. Hillslopes and hollows and their

formative processes (creep) are indicated by flowpath length as a scaling parameter. Unit stream power scaled against contributing area, a proxy for cumulative discharge, better defines fluvial landforms. As scaling relations are derived from DEM populations, they provide a statistical, rather than deterministic, prediction of the process domain transitions.

Introduction

Headwater streams have an importance which belies their size. Most of the channel network is comprised of low order streams (Shreve, 1956), which define the topology of the network and consequently how sediment and water are routed to downstream channel networks. These streams also provide nutrients, sediment and organic debris that form aquatic habitat (Gomi et al, 2002). They also have ecological characteristics which separate them from larger streams such as providing amphibian refuge and habitat (Olson et al., 2007). Their strong control on in-situ and downstream habitat was instrumental in the development of mapped riparian management zones to exclude timber harvest (USDA, 1994).

Headwater channels (Strahler stream order of two or less) in most networks are too numerous to efficiently characterize by field studies but the pervasive spatial distribution of headwater channels lends itself well to study by remote sensing techniques. However, few remote sensing techniques can capture both the broad spatial extent and small size of fingertip tributaries. Aerial photography and spectral imagery have been used to produce digital elevation models (DEMs) for topographic analyses but

typical DEM scales of 10-30 meters are inadequate for analysis of headwater channels which are commonly a meter wide and a decimeter deep, or even smaller. Only airborne laser swath mapping (LIDAR) data has been able to produce DEMs with the lateral extent (square km) and the horizontal resolution (~ 1 m) required to study headwater catchments.

Given a DEM, common methods of identifying channel initiation rely on arbitrarily defined threshold values of contributing area and slope-area product (Montgomery and Dietrich, 1989; Tarboton et al., 1989). These methods assume homogeneous change from hillslope to fluvial process domains irrespective of varying lithology and soils. The method of runoff generation and how precipitation is mediated through the hillslopes is also frequently heterogeneous. Runoff generation is particularly non-uniform in headwater catchments because of the frequent lack of buffering in poorly developed or absent riparian areas.

Landscape scaling characteristics have been used to establish the scale invariance of landscape organization (Horton, 1945; Hack, 1960; Rodriguez-Iturbe and Rinaldo, 1992, Rinaldo and Rodriguez-Iturbe, 1998), as well as determine its characteristic scale (Weissel et al., 1994; Dietrich and Montgomery, 1998; Birnir et al., 2001; Smith, 2010). The fractal nature of river network topology (Rodriguez-Iturbe et al., 1992, Rigon et al., 1996; Rodriguez-Iturbe and Rinaldo, 1992) and the power law behavior of hydraulic geometry of streams (Leopold and Maddock, 1953) show that the fluvial process domain imparts a scale invariant signature to the channeled portion of the landscape. However, the intervening hillslopes have a characteristic length, limiting further headward

extension of the network. It is this transition from scale dependence to scale invariance in topographic attributes that is utilized in this analysis to delineate hillslopes from hollows from channels. No single process domain is sufficient to explain the nature and distribution of landforms in headwater catchments. But, topographic scaling characteristics may be used to identify process transitions by finding the transition from a characteristic hillslope length to a scale-invariant channel network. Here, we test the hypothesis that scaling relations can distinguish hillslopes from hollows and hollows from channels using the characteristic scaling properties of the landscape.

Hack's law is an empirical scaling relationship between channel length and drainage area, where channel length is measured starting at the drainage divide to at any point along the channel (Hack, 1957). This relation has the power law form

$$L = 1.4A^{0.6} \quad (2.1)$$

where L is channel length in miles and A is drainage area in square miles. The original measurements used to derive Hack's law were inter-basin, that is, using separate basins. Here, we use Hack's law in an intra-basin sense, that is, all measurements from one catchment, to look at how the slope of the relation changes as hillslopes transition to hollows and then to channels. A hypothesized form of the intra-basin scaling relationship (Figure 2.1) exhibits multiple regimes, as flow structure transitions from diffusive overland flow to channelized flow (Dodds and Rothman, 2000; Birnir et al., 2007, 2008). The value of slope (exponent, m ,) may also vary with age and size of the basin (Birnir, 2008). At the scale of the hillslope, non-convergent flow dictates that the "basin" (A_1) is the flow path (L_1) such that the exponent on the scaling relationship is equal to one

($m_1=1$). As flow becomes convergent, the scaling relationship exhibits an inflection as the slope flattens from a slope of 1 to m_2 . This inflection represents the confluence of hillslope rills to form first order channels. As they coalesce, total contributing area increases with little additional length of the primary channel. The slope of the scaling relationship is then variable through a "crossover" region (m_3), until a scale-invariant network is realized (m_4). Values of $m_2 - m_4$ are hypothesized to vary from 0.4 -0.6 driven by the development of random networks and ultimately geologic control (Dodds and Rothman, 2000). Here, we test the ability of this conceptual model to predict where hollows and channels begin.

A second scaling relation (Figure 2.2), between unit stream power and contributing area, has been hypothesized to delineate the transition from a debris flow dominated domain to a fluvial process domain (Brummer and Montgomery, 2003). This relation follows earlier work characterizing the downstream variation of stream power with downstream distance (Knighton, 1999; Lecce, 1997). In large (fifth order) basins, stream power was found to exhibit peaks at approximately mid-basin, subject to convexities in the longitudinal profile resulting from lithology, structure, or past geomorphic process. Brummer and Montgomery (2003) examine how unit stream power varies in the downstream direction in mountain headwater catchments, and hypothesized that the transition zone between mass wasting process and fluvial process domains seen in the inflection of area-slope relations has a correlative transition zone in area- unit stream power relations.

This correlation arises from the formulation of unit stream power as a function of contributing area. Unit stream power is defined as the rate of energy expenditure per unit area of the channel bed (Eqn. 2.2), where ρg is the unit weight of water, Q is discharge, S is slope, and w is channel width. Unit stream power is a proxy for work done forming the channel and transporting sediment. Brummer and Montgomery (2003) derive a formulation of unit stream power as a function of contributing area by substituting discharge, channel width, and slope as power law functions of contributing area. In Equations 2.3 and 2.4, the coefficients e and c are empirical constants. The exponent d has a known range of values, typically 0.7 to 1.0 (Dunne and Leopold, 1978; Leopold, 1994). The exponent b is commonly reported as 0.5 for alluvial channels (Leopold and Maddock, 1953; Knighton, 1998). Using regression equations of discharge versus drainage area and width developed for southwest and central Montana (Parrett and Johnson, 2004), we use values of 0.8 and 0.56 for d and b , respectively. Here, we substitute Equations 2.3 and 2.4 in Equation 2.2 but retain slope as an independent variable, resulting in unit stream power as a function of contributing area and slope (Eqn. 2.5).

$$\omega = \frac{\rho g Q S}{w} \quad (2.2)$$

$$Q = e A^d \quad (2.3)$$

$$w = c Q^b = c A^b \quad (2.4)$$

$$\omega = \rho g \left(\frac{e}{c} \right) A^{(d-b)} S \quad (2.5)$$

This analytical formulation is supported by field studies of mountain channels which suggest that bed surface grain size increases downstream, reaching a maximum at the transition from mass wasting-dominated channels to fluvial dominated channels (Montgomery and Buffington, 1993; Lambert et al., 1996). Decreasing grain size would suggest that less geomorphic work, measured via unit stream power, is done performing sediment transport such that bed material distributions become finer. Here, this hypothesized scaling relation is tested as an additional means of detecting process transitions, albeit in catchments that may be too small to show the entire spectrum of behaviors.

Regions subject to diffusive (hillslope) flow have been distinguished from concentrated (fluvial) flow based upon the inflection to power law form in the relation of channel slope to contributing area (Eqn. 2.6).

$$S = kA^{-\theta} \quad (2.6)$$

where S = slope, ft/ft, k and θ are empirical coefficients representing steepness and concavity of the longitudinal profile, respectively (Hack, 1957; Sklar and Dietrich, 1998). The inflection in this relation has been used as a means of defining the transition in dominance from hillslope and mass wasting processes to fluvial process in channels (Montgomery and Foufoula-Georgiou, 1993; Stock and Dietrich, 2003). It has also been proposed that debris flows have a topographic signature based upon a characteristic trend of flat to decreasing slope in contributing area-slope plots (Stock and Dietrich, 2003). However, Brummer and Montgomery (2003) suggest that this region is actually a transitional zone to the fluvial domain. This analysis tests the hypothesis that hillslope

and fluvial process domains, as well as transitional zones, have distinct signatures in scaling relations of flowpath length and unit stream power with contributing area.

Methods

Two study areas in the Little Belt Mountains and Madison Range, Montana were selected based on physiographic attributes and availability of LIDAR data. The Little Belt study area is located in the U.S. Forest Service Tenderfoot Creek Experimental Forest (TCEF). The second study area is located in the Madison Range, south of Big Sky, Montana. The study areas are comprised of two paired catchments: Stringer (STR) and Spring Park Creeks (SPC) in TCEF (Figure 2.3) and Second and Third Yellowmule Creeks near Big Sky (YM2, YM3) (Figure 2.4).

The channels STR and SPC are north-south trending with a south aspect, sharing a common watershed divide. The study area consists of the upper approximately 4 sq. km. of each watershed because this extent includes all of the first order channels. Both catchments are second order basins. Elevations within the study area range from 2170 to 2330 m. Bedrock consists of gently dipping Cambrian sedimentary rock, largely Flathead Formation and Woolsey Shale, with some outcrops of Eocene quartz monzonite capping the upper parts of the catchments (Reynolds and Brandt, 2007). The Flathead Formation in this area is described as having a thick middle shale layer beneath a thin and above a massive basal sandstone layer (Weed and Pirsson, 1899). Most of the headwater channels begin in shale (Woolsey or middle Flathead) but the main stem valley floor is mantled with debris flow lag of sandstone flagstones rafted by failing shale layers. The

lowest part of the study area is underlain by the lower unit of the Flathead Formation, a resistant sandstone unit. At this change in lithologic type, the channel transitions from a pool-riffle morphology to cascade and step-pool morphology.

Hillslope hydrology of the TCEF area has been well studied (Jencso et al., 2009; Payn et al., 2009). These works have established that stream heads (the initiation point of perennial stream flow) is well-correlated with the upslope accumulating area of hillslopes, and that large accumulating areas on hillslopes drives water connectivity between hillslopes and channels (Jencso, 2009). However, channel heads form by lithologic control, where the relatively impermeable shale and permeable sandstone meet, or by threshold exceedence of surficial critical shear stress.

The YM2 and YM3 channels are also north-south trending, but with a north aspect, and they also share a common watershed divide. The YM2 and YM3 study areas (approximately 3.0 sq. km. each) include most of the first order channels in these third order basins. Elevations in the study area range from 2600 - 2800 m. These catchments are underlain by Morrison Formation, Cretaceous Muddy Sandstone and Thermopolis Formation and Kootenai Formation, thin sandstone layers separated by thicker shale layers (Kellogg and Williams, 2000). The basins form on the gently north-dipping north limb of an east-west-trending asymmetrical anticline, the axial crest of which underlies the southern drainage divide. of the West Fork. The east and west sides of the catchments are steep and subject to failure of sandstone on the shale layers. The lowest part of the study areas encounters a thicker sandstone unit of the Kootenai Formation. Headwater channels initiate by either gradual accumulation of surface flow in convergent

topography or as seepage erosion-generated spring heads at the interface of lithologic units. Channel heads are also observed at contacts with interbedded mudstones and sandstones.

Both catchments pairs are high elevation and runoff generation is dictated by annual snow-melt. It is during this annual runoff period that geomorphically effective work is performed in the channels. Large snowdrifts as late as July have been observed as direct source areas for first order channels in the Big Sky catchments. In TCEF, peak runoff occurs when hillslopes with high upslope accumulating area are connected to the streams by routing shallow subsurface water through riparian areas (Jencso et al., 2009). Though the catchment pairs generate runoff through snowmelt, the channels of the Big Sky catchments appear to have greater geomorphic effectiveness due to less resistant bedrock and greater runoff potential resulting from larger available snow water equivalent (Figure 2.5). The result is high drainage density in the Big Sky catchments compared to the low drainage density TCEF catchments.

The TCEF and Big Sky study areas were selected to compare the range of natural variability within the same physiographic, climatic, and lithologic region, as well as the variation in process domain transitions between two different settings with similar lithology but different climate and aspect. A cursory inspection of the bare earth images (Figures 2.3 and 2.4) shows the striking difference in drainage density between the catchment pairs, despite similarities in lithology and structure. The headwaters of both catchment pairs are underlain by near flat lying units, the most extensive of which is a shale layer. Yet, the Big Sky catchments exhibit a greater drainage density than the

TCEF catchments. The origin of the difference is not certain, but may be due to differences in permeability of the soil or rock, amount of snow and rate of melt, or another factor not investigated.

The delineation of hollows from hillslopes and channels from hollows relies on LIDAR-derived topographic data, verified by field data collection. Last return signals were classified by NCALM, who was contracted to perform the Lidar flight, using Terrascan software (Soininen, 2004) to produce 1 m DEMs. These DEMs (in ESRI GRID format) for TCEF and Big Sky were used as the starting point for this analysis.

Analytical data was verified against Global Positioning System (GPS) field mapping of hollow and channel heads in the four catchments (Figures 2.6 and 2.7). Mapped locations were post-corrected to an accuracy of less than 0.5 meters. Though channels and hollows are convergent landforms, channels have banks defined as "steepened" (Montgomery and Dietrich, 1989). For field identification, hollow heads were defined as the upstream-most extent of a valley, surrounded by concave-upwards topography on three sides, or concave-upwards plan curvature. Channel heads were defined as an abrupt upstream termination of a persistent incision in the valley floor, either discontinuous or continuous, with definable streambanks. A definable bank "must be recognizable as a morphological feature independent of the flow" (Dietrich and Dunne, 1993, p. 178) with characteristics such as "a narrow zone of gradient steeper than both the neighboring hillslope and the transverse slope of the channel bed" (Dietrich and Dunne, 1993, p. 178).

Numerous hollows and channel heads in TCEF and Big Sky (Figs. 2.6 and 2.7) are stratigraphically controlled, i.e.; heads emerge at the interface of differing lithologic units. In TCEF, this is typically the interface of the permeable upper Flathead sandstone and the underlying relatively impermeable middle Flathead shale, as well as the interface of the upper Flathead sandstone and the overlying Woolsey shale. These heads typically have no upslope hollow. Similarly in Big Sky (Figure 2.7), numerous channel heads are controlled by the interfaces of the Morrison Formation and Muddy sandstone and Thermopolis Formation, and the Muddy sandstone and Thermopolis Formation and the Kootenai Formation. These heads do not have an upslope hollow and are produced by subsurface saturated flow as evidenced by seepage erosion scarps. Subsurface flowpaths may be related to unconformities between the lithologic units and relative permeability.

Contributing Area-flowpath Length Relation

Changes in flow regime, and hence landforms (convex to concave, concave to incised), may be indicated by inflections or rollovers in the slope of Hack's intra-basin law (Figure 2.1). To determine the relation between channel length and drainage area, the length of the flow path or channel from the watershed divide to each cell of the DEM, and the contributing area of each cell, was calculated. Values were generated using the Basic Grid analysis toolset in TauDEM (Tarboton et al., 1989, 1991). The maximum flowpath length (plen) and the D-infinity contributing area (sca) rasters were exported to MATLAB and used to create a log-log plot of channel length versus contributing area.

This analysis relies on accurate determination of contributing area to each grid cell in the DEM, as well as a continuous flowpath from each cell to the watershed divide.

However, landscape texture ensures that some cells will have very limited flowpaths and not participate in the channel network. These cells, identified by a grid order of one, were excluded from the analysis. The grid order of each cell is calculated as a raster (gord) in TauDEM. The population of flowpath length and contributing area values at each grid cell, of order greater than one, were binned and averaged to create log bin average relations of flowpath length and contributing area. Convergent landforms with a small contributing area may also exist yet not be an active part of the channel network. These features were excluded by only calculating flowpath length for cells with a contributing area greater than 10 sq. m. The values of contributing area and flowpath length associated with each mapped head and hollow were extracted by spatial analysis in ArcGIS.

Contributing Area-unit Stream Power Relation

Unit stream power as a function of contributing area and slope was calculated by (Equation 2.5). Unit stream power was calculated using contributing area and local slope at each cell of a 1 m. DEM with a cell-based grid order of greater than one. The grid order of each cell is calculated as a raster (gord) in TauDEM. The values of contributing area and unit stream power associated with each mapped head and hollow were extracted by spatial analysis in ArcGIS.

Contributing Area-slope Relation

Using the rasters generated in the previous analyses, log bin average plots of area-slope relations were produced for each of the four catchments. The values of

contributing area and slope associated with each mapped head and hollow were extracted by spatial analysis in ArcGIS.

Results

This analysis investigates the ability of scaling relationships to detect transitions in process domains. These relations are verified against field mapping of process transition points, namely hollow heads and channel heads of continuous and discontinuous channels.

Log bin averaged contributing area-flowpath length relations of the four catchments exhibit relations similar to the general form described by Dodds and Rothman (2000). Flowpath length-area relations show several discrete domains in the Big Sky and TCEF catchments, defined by clear breaks in slope (Figure 2.8). The first slope inflection defines a transition from hillslopes to hollows where the rate of increase of contributing area exceeds the rate of increase of flowpath length. Flowpath length-area relations are internally consistent within the Big Sky and TCEF catchments. The flowpath-area relations of both catchment pairs show an initial slope near one, sharply declining in the m2 region to approximately 0.42. The catchment pairs differ in the transition from the m2 to m3 region, with the Big Sky catchments exhibiting an increase to a slope of roughly 0.5 and the TCEF catchments showing a decrease to an approximate slope of 0.37.

Field mapping of hollows and channel heads shows a clustering mostly in the m2 region in the Big Sky catchments (Figure 2.9). Hollows and heads in TCEF are more

scattered in the m1 and m2 regions, as well as a few outliers in the m3 region (Figure 2.9). Mapped TCEF hollows tend to have longer flowpaths than mapped heads. These heads are stratigraphically controlled and generally emerge as springs, at the interface of a flatlying, permeable sandstone unit underlain by a relatively impermeable shale unit (Figure 2.6). Hollows also exhibit some geologic control, often occurring at the interface of the sandstone unit and the overlying Eocene unit. In the Big Sky catchments, the dependency of channel heads on lithologic interfaces is also evident from surficial mapping (Figure 2.7) and the lack of order in the mapped hollows and heads. However, despite strong lithologic controls, only one channel head is mapped in the hypothesized fluvial m3 region.

Log bin averaged contributing area-unit stream power relations for all four catchments (Figure 2.10) increase, flatten, and then show both increasing and decreasing values of unit stream power with increasing contributing area. The form is similar to that observed by Brummer and Montgomery (2003). The downstream trend in decreasing unit stream power values may be interrupted by high stream power values at high resistance, bedrock controlled reaches. In TCEF, the lowest part of the study area is underlain by the lower unit of the Flathead Formation, a resistant sandstone unit, transitioning to Precambrian granite gneiss. In the Big Sky catchments, the lowest part of the study areas encounters a thicker sandstone unit of the Kootenai Formation.

Field-mapped hollows and channel heads (Figure 2.11) in the Big Sky catchments plot exclusively on the rising limb of the curve, but exhibit no order between hollows and heads. In STR, hollows generally have higher values of unit stream power than channel

heads. The STR hollows and heads also plot on the rising limb of the relation, however the channel heads of SPC are scattered throughout rising and falling limbs of the relation. The channel heads of SPC most strongly show the influence of lithologic control on channel initiation.

Log bin averaged contributing area-slope relations of the four catchments (Figure 2.12) display the general form found in mountain channel networks by Stock and Dietrich (2003) and Brummer and Montgomery (2003). Contributing area-slope relations define two discrete domains in the Big Sky and TCEF catchments. All hollows and nearly all channel heads plot in the mass wasting domain of the contributing area-slope relation (Figure 2.13). Most channel heads plot nearer to the inferred transition to the fluvial domain at the slope break.

Discussion

This analysis tests statistical prediction of the location of hollows and channel heads and the spatial extent of hillslope and fluvial process domains using scaling relations derived by remote sensing techniques, specifically LIDAR-derived topographic data. Statistical classification, rather than deterministic prediction, is predicated on the assumption that the hillslopes, hollows, and channels have distinctive scaling regimes in either the length or contributing area scale. Hillslope hydrologic process and hillslope processes, largely creep, in the study areas, likely drive transitions in length scale relations, and fluvial process and channelized flow likely drive transitions in contributing area relations.

The scaling relation of Hack's law was used in an intra-basin sense to detect transitions between the landscape elements of hillslopes, valleys, and channels. In Big Sky and TCEF, the hypothesized m1 and m2 regions have similar values of slope (Figure 2.8). In the m1 region, flowpath lengths of 15- 20 m. and 20- 30 m. at a contributing area of 10 sq. m. in Big Sky and TCEF, respectively, may indicate the length of convex to planar hillslopes in the catchments with parallel flowpaths of nonconvergent wash flow. In this region, surface evolution is dominated by diffusion, wash, and other processes that tend to smooth the ground surface. This smoothing "fills in" low spots and removes high spots, by definition creating planar to convex hillslopes lacking persistent convergence. The slope of this segment is hypothesized to be equal to or greater than one.

In the m2 region, contributing area increases faster than flowpath length and slope of the segment flattens to approximately 0.42. This region corresponds to convergent hillslopes or hollows where mapped hollows and channel heads may initiate. The m2 process zones of the landscape in TCEF (Figures 2.14 and 2.15) show subparallel flowpaths on the hillslopes on the order of several hundred meters in length. The order of magnitude of the TCEF hillslope length is set by the uplift of flat lying bedrock and its relative erodibility in Pleistocene climatic regimes. Resistant Eocene quartz monzonite maintains relief and underlying massive Flathead sandstone limits the development of valley dissection to the lower part of the headwaters where Woolsey Shale and middle Flathead shale units can be incised by relatively low stream power, and where mass wasting in weak shale can provide a hollow to maintain a convergent flowpath. In the Big Sky landscape (Figures 2.16 and 2.17), the m2 process zones are much shorter,

truncated by interfluvies between channels. The relatively short convergent flowpaths result from a well-dissected landscape of weak bedrock and relatively high available stream power.

In the m3 region, there is increased variability in flowpath length as larger tributaries are added to the total contributing area and the flowpaths are more sinuous. Channeled flow begins but these first order channels are limited to increasing their contributing area by increasing flowpath length due to their rigid sub-parallel structure. As their stream order increases, tributaries may enter, increasing contributing area without proportional increases in flowpath length. In the m3 region, the slope of the Big Sky flowpath relations increases relative to the m2 region, but decreases in the TCEF catchments. This difference reflects the continued coalescence of multiple incipient channels in the Big Sky catchments, increasing flowpath length, relative to contributing area.

Three distinct regions were also seen in flowpath length scaling relations in three headwater catchments of the Eel River (Gangodagamage et al., 2011). These regions were hypothesized to correspond to different flow path topologies marking geomorphic transitions. Gangodagamage et al. (2011) posit that flowpath length, rather than contributing area, is better able to distinguish the transition from sub-parallel flow to convergent flow. The contributing area metric, on the other hand, is better able to distinguish transitions in channeled flow. However, Gangodagamage et al. (2011) did not verify the predicted regions with mapped hollows and channel heads.

In the TCEF study area, the locations of mapped channel heads are generally unrelated to the spatial boundaries of the m2 and m3 process zones (Figures 2.14 and 2.15) though most hollows occur near the upstream-most extent of the m3 process zone. However, the m3 process zones in TCEF spuriously group hollows and channels in the same process zone. Field mapping verifies that m3 process zones predicted upslope of the mapped channel heads are actually hollows. In the Big Sky catchments, channel heads are near or at the upstream-most extent of the m3 process zone, but generally the m3 process zone does not extend to the hollows. The m3 process zone generally accurately portrays the spatial extent and discontinuous nature of the Big Sky channels except where it maintains channels at the crossing of sandstone lenses, clearly visible in hillshade imagery. Many of the Big Sky channels do not persist across these units and reinitiate downslope. The Big Sky flowpath relation exhibits scatter at the highest values of contributing area. This scatter is interpreted as corresponding to changes in flow regime as channels become less sub-parallel and show the influence of tributaries and increasing flowpath length due to greater sinuosity as channels transition from colluvial to alluvial.

The overall pattern of unit stream power in Big Sky and TCEF (Figure 2.10) is consistent with previous work utilizing unit stream power (e.g., Brummer and Montgomery, 2003; Knighton, 1999; Lecce, 1997) which found unit stream power increases downstream in headwater channels before reaching an inflection where unit stream power begins to decrease. There is scatter in the values of unit stream power at the greatest values of contributing area but these can be attributed to the encounter of

bedrock-controlled reaches, as the channels flow downsection from shale and shale-dominated units to massive, high resistance sandstones. Apart from scatter, both study areas reach a peak in unit stream power at roughly 50,000 to 150,000 sq. m. of contributing area. The rising limb of the Big Sky relation steepens as sandstone lenses are encountered. Brummer and Montgomery (2003) found similar behavior of multiple peaks in unit stream power in some of their headwater catchments. They attribute the disruption of the declining trend in unit stream power to inherited glacial landforms.

The contributing area at the inflection zone is greater than that at mapped hollows and channel heads in all catchments except SPC (Figure 2.11). The cluster of mapped hollows and heads in Big Sky and STR reflects the transition from hillslope to convergent and channelized landforms. The considerable scatter in unit stream power values in SPC again reflects the strong lithologic control on channel initiation. In STR, the unit stream power values at hollows equals or exceed unit stream power at channel heads due to the location of hollows within high relief, resistant Eocene quartz monzonite and the emergence of channel heads within the flat-lying middle Flathead shale unit. The TCEF transitional and fluvial process zones (Figures 2.18 and 2.19) generally correctly delineates channels as fluvial zones and hollows as transitional zones. One notable exception is the delineation of the west fork of STR as a transitional, hollow zone. This fork of STR is a perennial tributary to STR. Short segments of transitional process zone also appear within delineated fluvial zones of STR. This may result from the channel sporadically flowing through debris flow lag of immobile sandstone flagstones rafted to the valley floor by failing shale. In SPC, the fluvial zone is also incorrectly delineated

upstream of the channel heads and the channels associated with some channel (spring) heads are delineated as transitional or hollow zones. However, the delineated transition zones generally extend just up to the mapped hollows.

In the Big Sky catchments, transitional process zones begin at or near most mapped hollows (Figures 2.20 and 2.21). The intermittent transitional process zones correctly depict the loss of channels crossing resistant sandstone lenses. There is also a repeated shift between the transitional and fluvial process zones in YM2. In YM3, multiple low order channels coalesce to several dominant channels all of which are delineated in the fluvial zone. In YM2, drainage density remains high through most of the study area, diffusing contributing area to multiple channels such that power to maintain fluvial conditions is limited. Consequently, these multiple channels shift back and forth between transitional and fluvial process zones. Figures 2.10 and 2.11 show deviations from the hypothesized unit stream power scaling relation at contributing area values greater than one square kilometer, particularly in the Big Sky catchments. These regions are represented in Figures 2.20 and 2.21 as black channels. These channels have increasing unit stream power due to increases in slope resulting from flow over resistant bedrock.

The topographic signature of the transition from hillslope process dominated channels to fluvially dominated channels has been identified as an inflection in the relationship between contributing area and channel slope (Montgomery and Foufoula-Georgiou, 1993; Stock and Dietrich, 2003). This inflection is evident in the Big Sky and TCEF catchments at roughly 800,000 and 200,000 sq. m., respectively (Figure 2.12).

The convergence of flowpaths in TCEF into a limited number of channels facilitates an earlier transition to fluvial process conditions. In Big Sky, weak shale facilitates the development of high drainage density, delaying the fluvial process transition until multiple channels confluence. This is reflected in the appearance of hollows and heads at order of magnitude smaller values of contributing area (Figure 2.13). In TCEF, heads and hollows are located much closer to the inflection to fluvial conditions. In STR, hollows appear closer to the fluvial transition than mapped heads owing to the emergence of most channel heads as springs with relatively small upslope areas. Numerous workers have tested the slope-area relation but recent work has tested this relation in the intermountain West (Colorado Front Range), (Henkle et al., 2004). They found that channel heads mapped in the channelized zone, past the point where they predicted based on slope-area relations. This discrepancy with predicted behavior was attributed to substantial control on channel initiation exerted by subsurface flow, lithologic structure, jointing, and strength. Differences in rock strength may explain why the TCEF catchments require greater contributing area to initiate channel heads than the Big Sky catchments.

The location of fluvial domains in SPC continues to exhibit a lack of predictability given its dominant lithologic and structural controls. Fluvial process zones in TCEF predicted by slope-area relations (Figures 2.22 and 2.23) show a partial fluvial process zone in STR and overpredict the fluvial process zone in SPC. The intermittency of the fluvial domain in STR may reflect intermittent colluvial and alluvial conditions as the channel encounters immobile debris flow lag. In the Big Sky catchments, the fluvial

process zone is predicted far downstream of the presence of channels (Figures 2.24 and 2.25). However, alluvial conditions may be predicted by the fluvial process zone.

Conclusion

Our analyses of two pairs of headwater catchments in north-central and southwest Montana tests the efficacy of statistically derived scaling relations to detect transitions in hillslope and fluvial process domains. We used inflections in scaling relations of flowpath length, unit stream power, slope and contributing area to delineate where hillslope process spatial domains end and fluvial process spatial domains begin. Typical means of defining the network fingertips rely upon constant thresholds of contributing area or slope. Here, we suggest using statistical distributions of topographic and geomorphic parameters (flowpath length and unit stream power) in scaling relations as an alternative means of predicting where hollows and channels begin, as well as a means of assessing the topology of the network. Flowpath length versus contributing area assesses the type of flow and arrangement of flowpaths. Unit stream power versus contributing area evaluates the downstream trends in geomorphic work in the fluvial domain. These relations represent a shift from simple measures of contributing area to measures that tie topographic and geomorphic signatures to landforms.

Both the flowpath length and unit stream power relations more accurately predicted the locations of hollows and heads, and the spatial extent of hillslope and fluvial process zones than slope-area relations. Flowpath length and unit stream power relations provided the best predictions in the Big Sky catchments. The location of

hollows, heads, and the transitions between process zones were best detected using the unit stream power relation. This relation was able to detect the shifts between hillslope-dominated and fluvial-dominated process zones, as well as accurately detecting the intermittency of channels through resistant sandstone lenses. Flowpath length scaling relations produced similar results but predicted continuous channels where they do not exist. In TCEF, flowpath length scaling relations were better able to detect hollows and the spatial extent of the network topology. Unit stream power relations produced similar results with respect to detecting hollows. However, detecting the presence of channel heads by statistical scaling relations was relatively unsuccessful in TCEF. The SPC catchment, in particular, was the least amenable to accurate prediction of hollows, heads and network topology by any of the means employed. Slope-area relations produced the least accurate prediction of process zones in both study areas. However, this relation can be augmented and verified using an intra-basin version of Hack's law and unit stream power scaling relations. Area-length relations have the advantage of defining the characteristic hillslope length. The evident scatter in the mapped hollow and channel heads shows that deterministic predictions of hollow and channel heads are not possible, particularly where channel heads may not be the result of an upslope zero order basin. Discrete channel head emergence may be driven by underlying structure, aspect, lithology, and water table connectivity of hillslopes to channels, rather than by surface topography and surface water hydrology. However, corroborating use of all three scaling relations enables high resolution detection of the topographic signatures of hillslope, hollow and channeled landforms and quantification of their spatial distribution. Although

locations of stream heads in STR and SPC are well-correlated with upslope accumulated area, we argue that prediction of channel heads is equally important. Headwater streams, whether perennial or intermittent, occupy a pivotal position in the landscape, between hillslope and fluvial processes, and upland and riparian landscapes. They are the initiation point for the downstream transmittal of runoff, nutrients, sediment and organic matter, forming an ecological niche separate from both upland and aquatic habitats. Discrimination of their inception points is critical to habitat protection and delineation of riparian buffers. The intermittent or ephemeral reaches between the channel head and the stream head have important riparian values, serve as source areas for sediment, wood, and nutrients, and provide uncommon amphibian habitat. Knowledge of channel heads locations can then inform watershed management. The proposed scaling relations suggest an improvement in detection of the limits of valley dissection and colluvial/alluvial transitions.

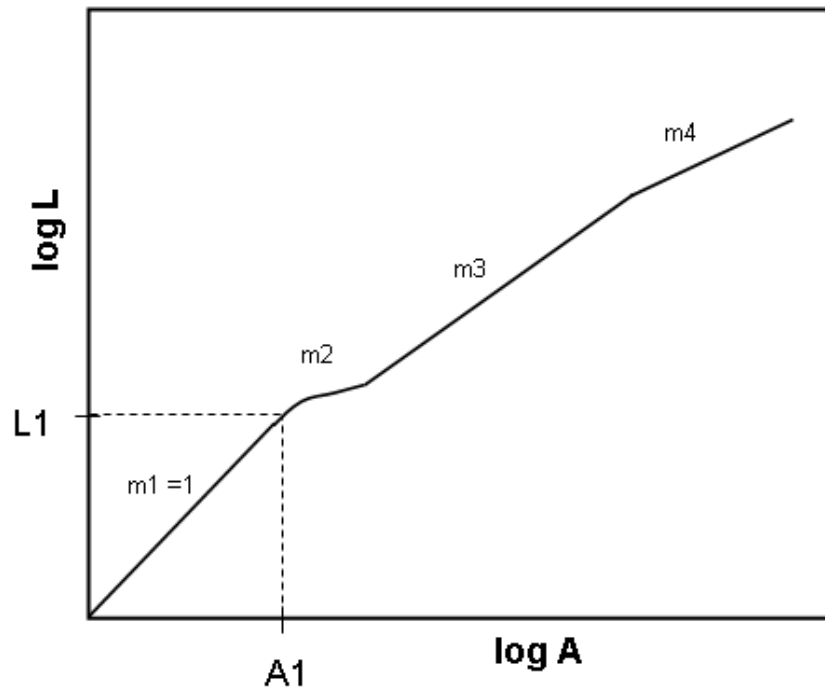


Figure 2.1: Conceptual model of Hack's intra-basin law, contributing area, A versus flowpath length, L (after Dodds and Rothman, 2000). Hillslope is denoted by region $m1$, where the hillslope length, $L1$ is the contributing area, $A1$. Weakly convergent flow of hollows characterizes region $m2$. Addition of tributaries and change in basin size transition from hollow to colluvial and alluvial channel in regions $m3$ and $m4$, respectively.

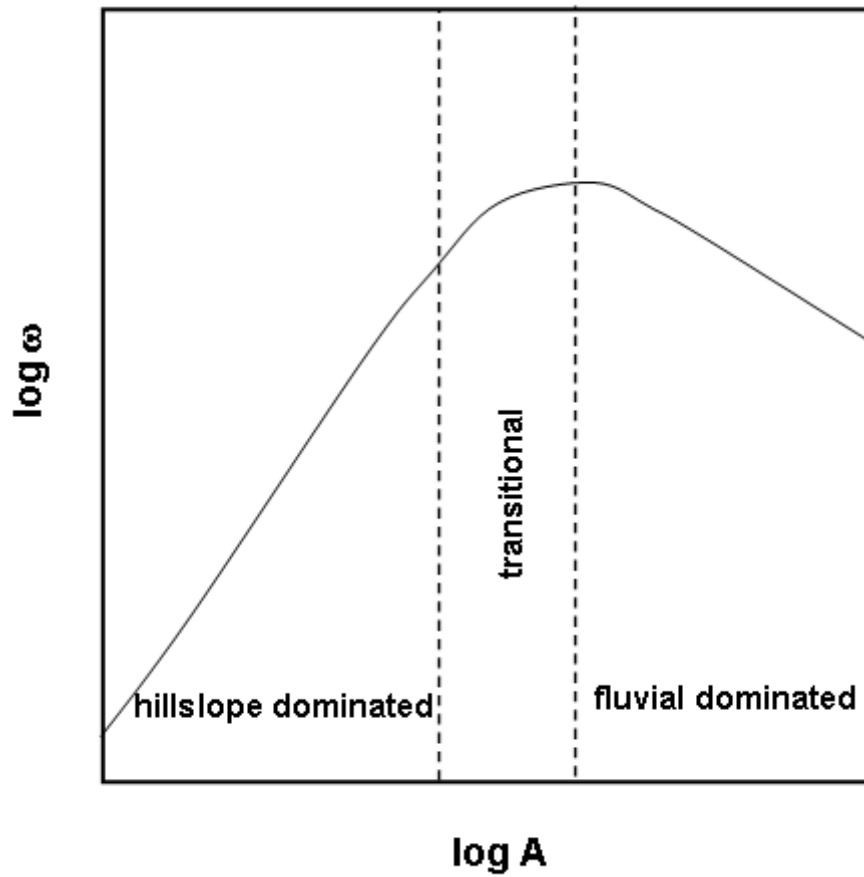


Figure 2.2: Conceptual model of rate of change in unit stream power, ω , with increasing contributing area, A (after Brummer and Montgomery, 2003). Transition in rate of change of stream power corresponds to process domain transitions from hillslope to fluvial domain.

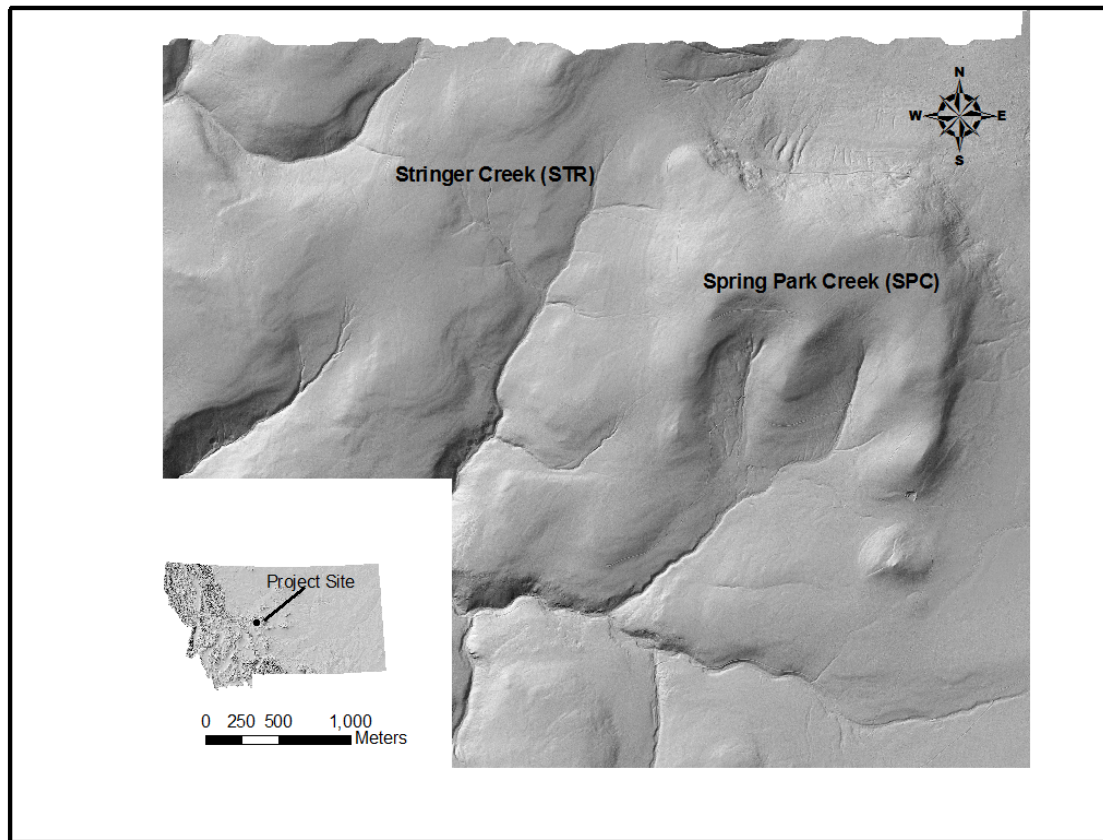


Figure 2.3: LIDAR-derived hillshade map of Stringer and Spring Park Creeks, TCEF, Montana.

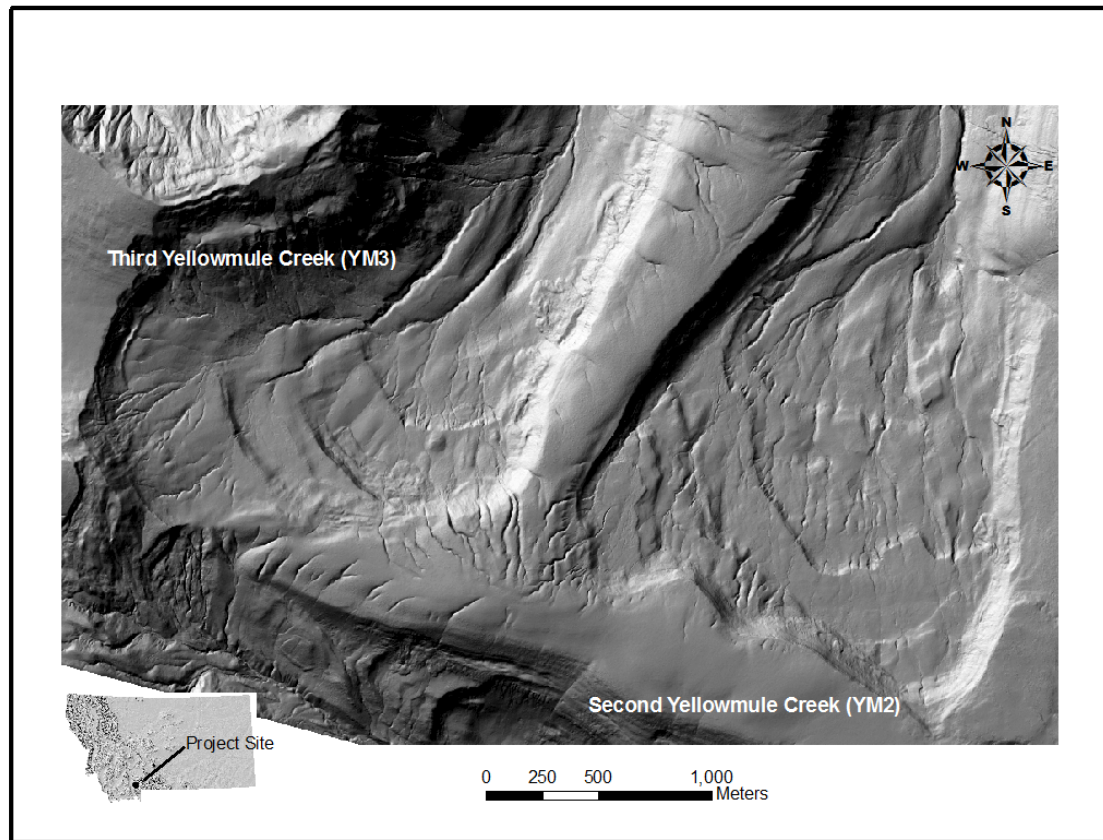


Figure 2.4: LIDAR-derived hillshade map of Second and Third Yellowmule Creeks, Big Sky, Montana.

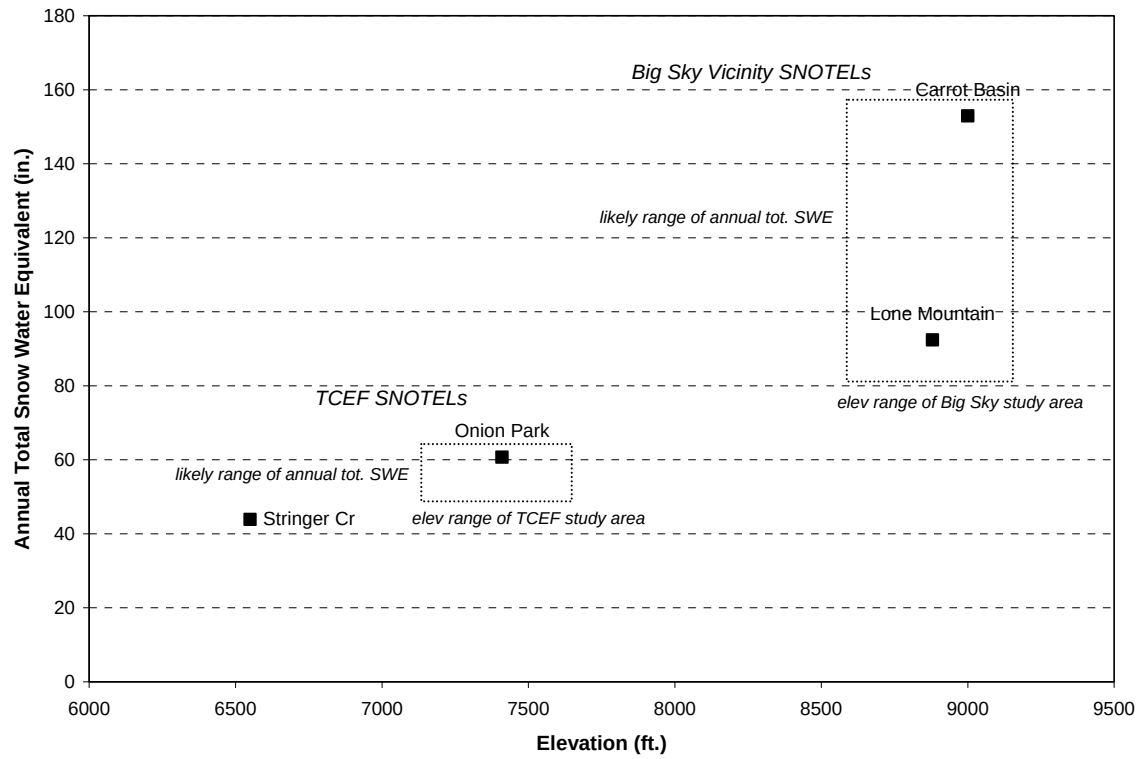


Figure 2.5: Total annual snow water equivalent (SWE), in inches, versus SNOTEL elevation for SNOTEL stations near the study areas.

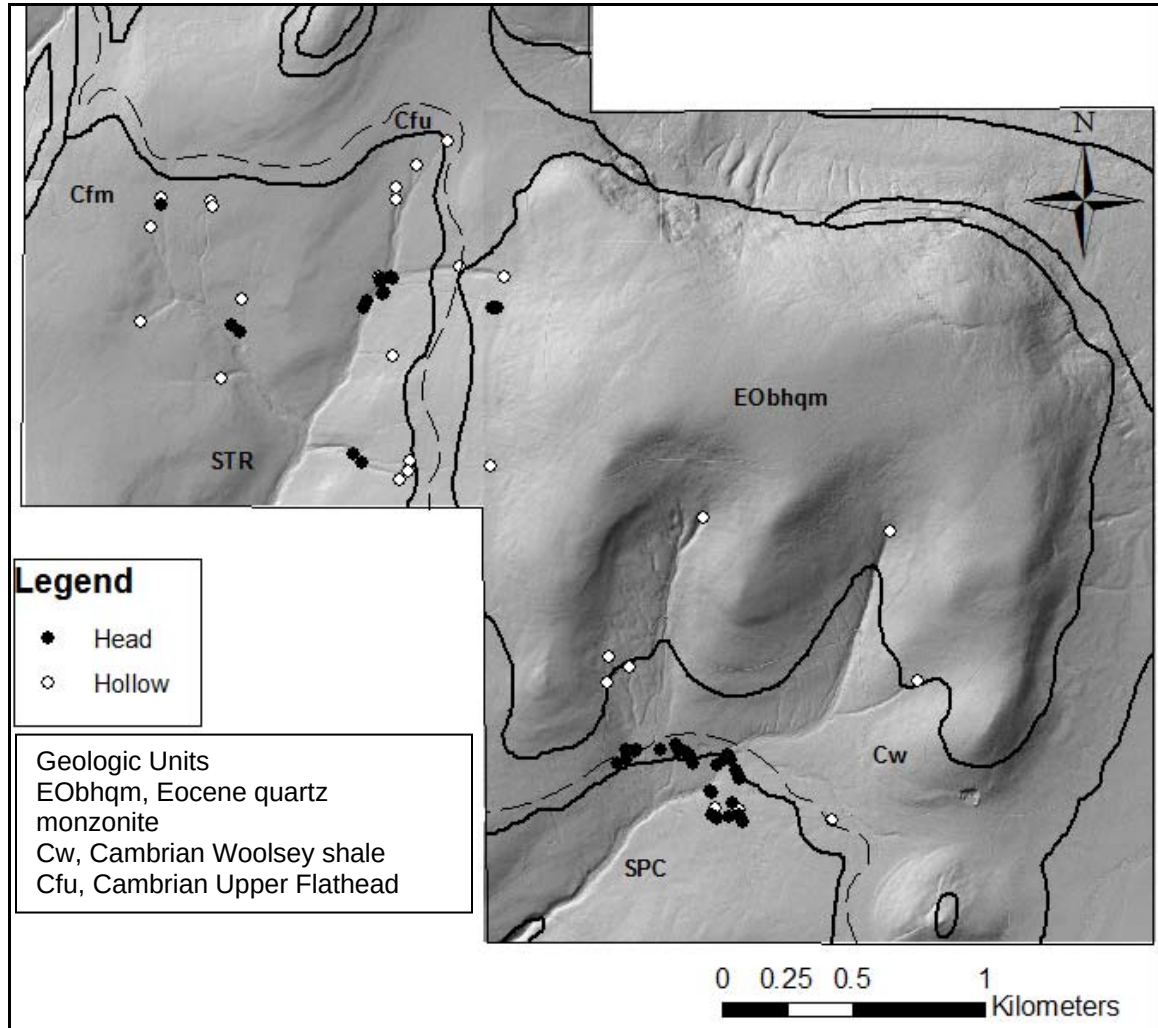


Figure 2.6: Field-mapped channel heads and hollows, Stringer and Spring Park Creeks (STR, SPC), TCEF, Montana (Reynolds and Brandt, 2007).

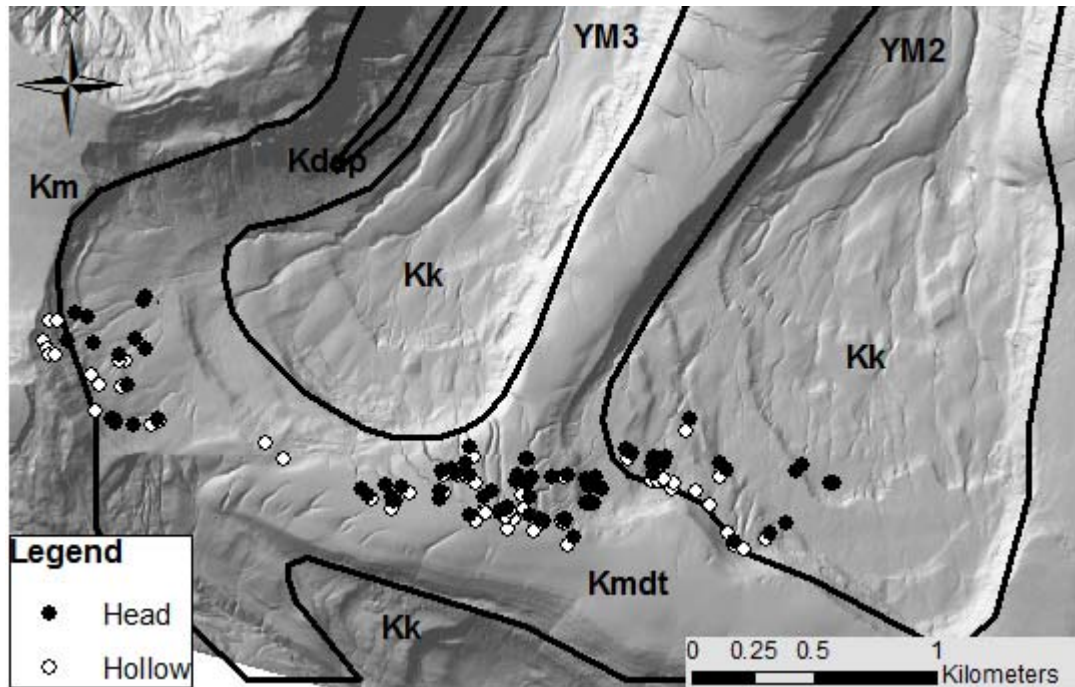


Figure 2.7: Field-mapped channel heads and hollows, Second and Third Yellowmule Creeks (YM2, YM3), Big Sky, Montana. Geologic units: Kmdt, Muddy sandstone and Thermopolis Formation; Km, Morrison Formation; Kk, Kootenai Formation; Kdep, Dacite porphyry (Kellogg and Williams, 2000).

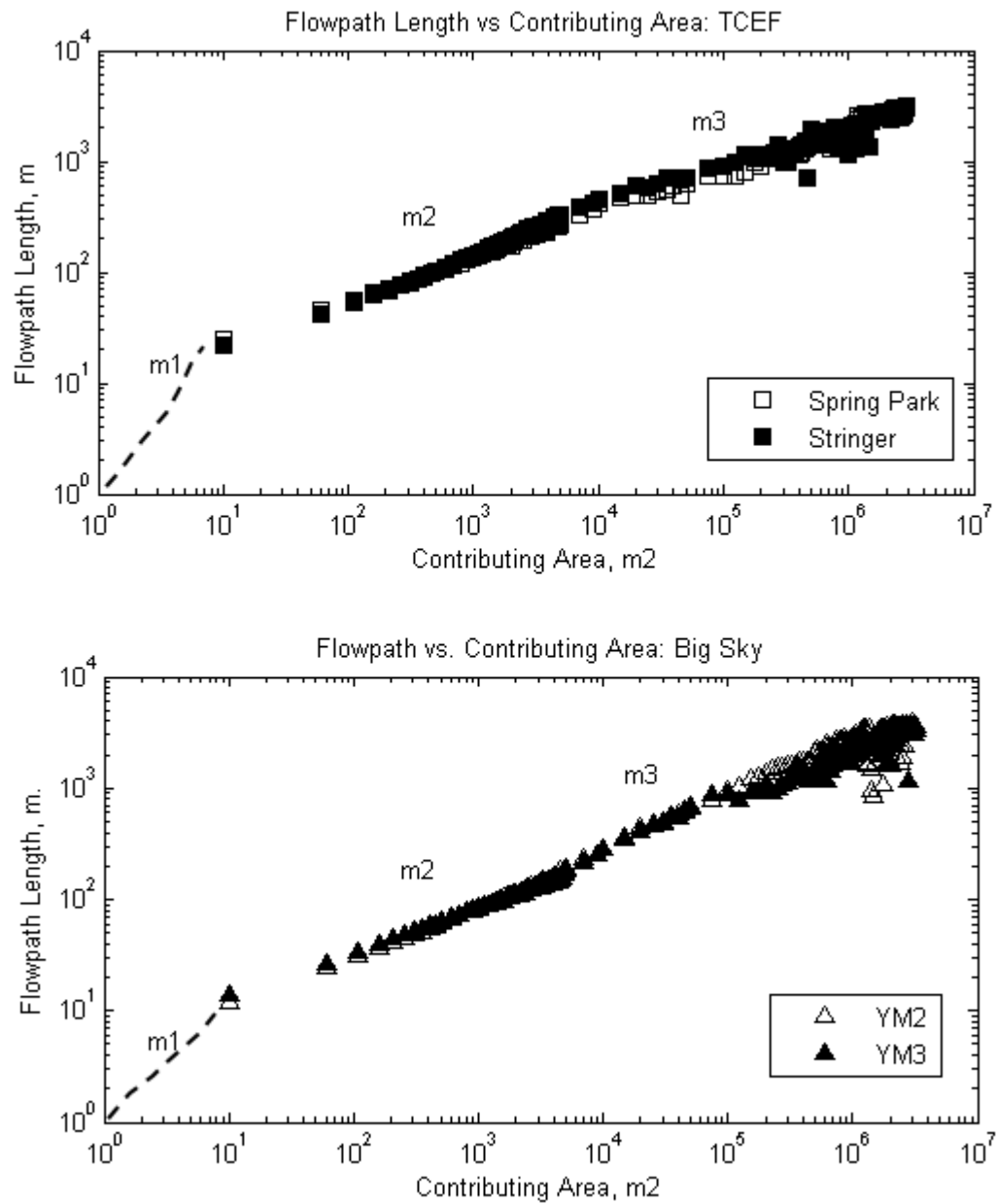


Figure 2.8: Contributing area versus flowpath length, TCEF and Big Sky catchments. m1, m2, and m3 denote hillslopes, transitional hollow region, and transitional colluvial and alluvial channel region.

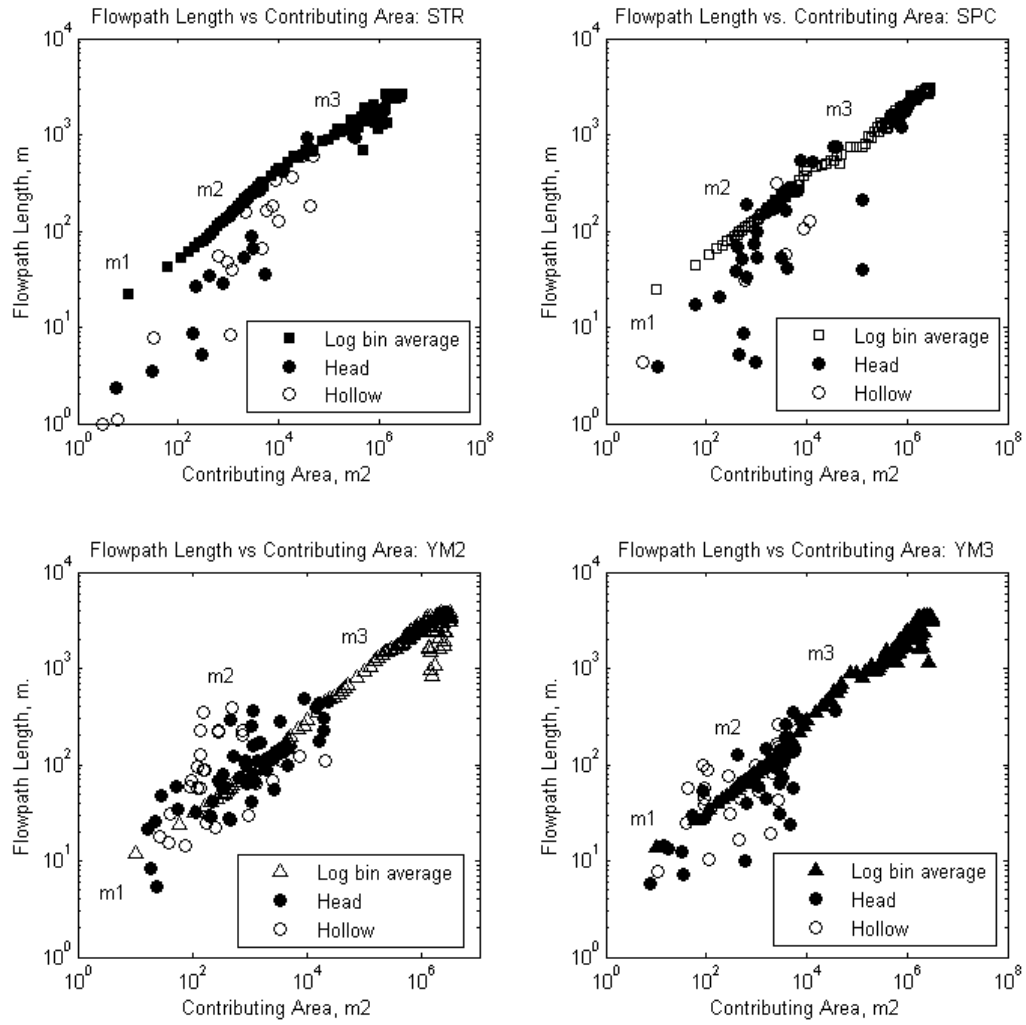


Figure 2.9: Contributing area versus flowpath length and mapped hollows and heads (STR, SPC, YM2, YM3).

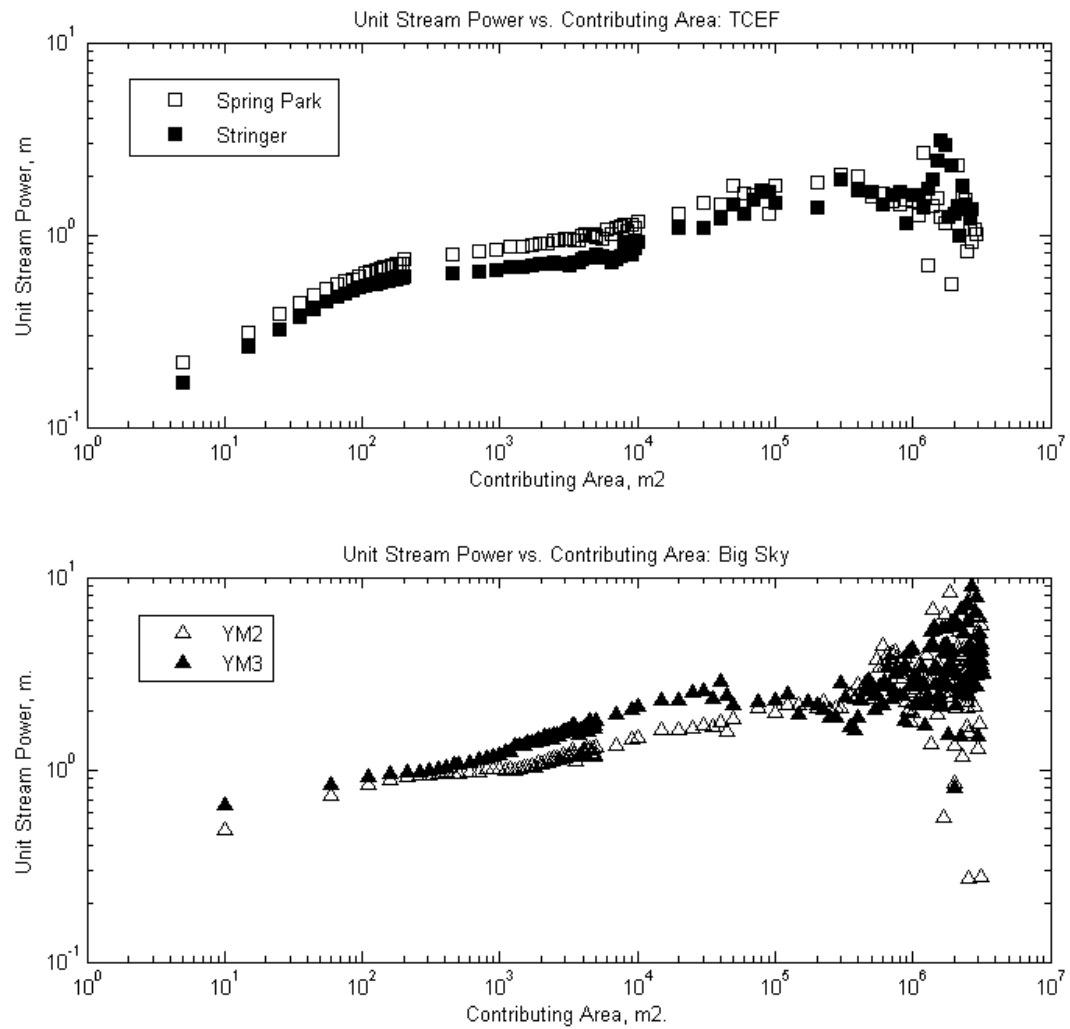


Figure 2.10: Contributing area versus unit stream power, TCEF and Big Sky catchments.

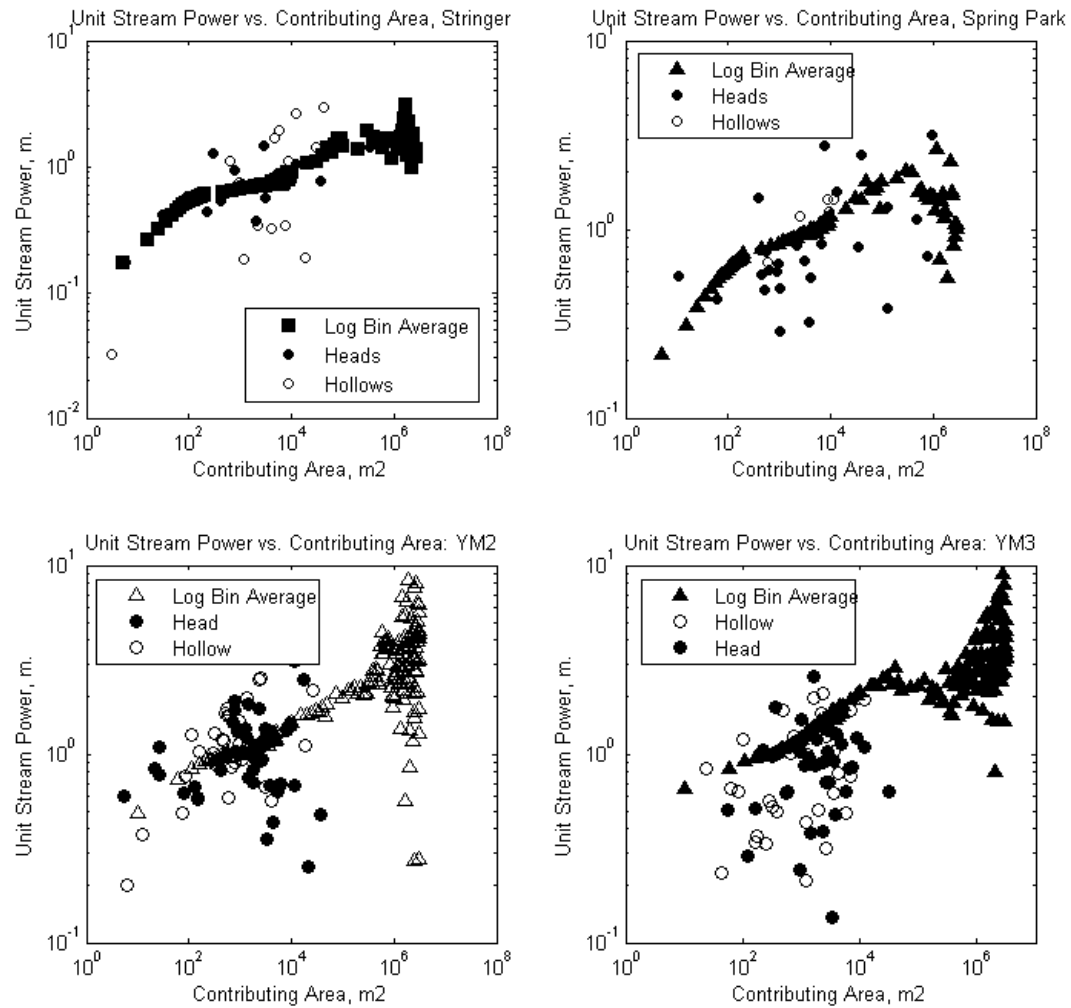


Figure 2.11: Contributing area versus unit stream power and mapped hollows and heads (STR, SPC, YM2, YM3).

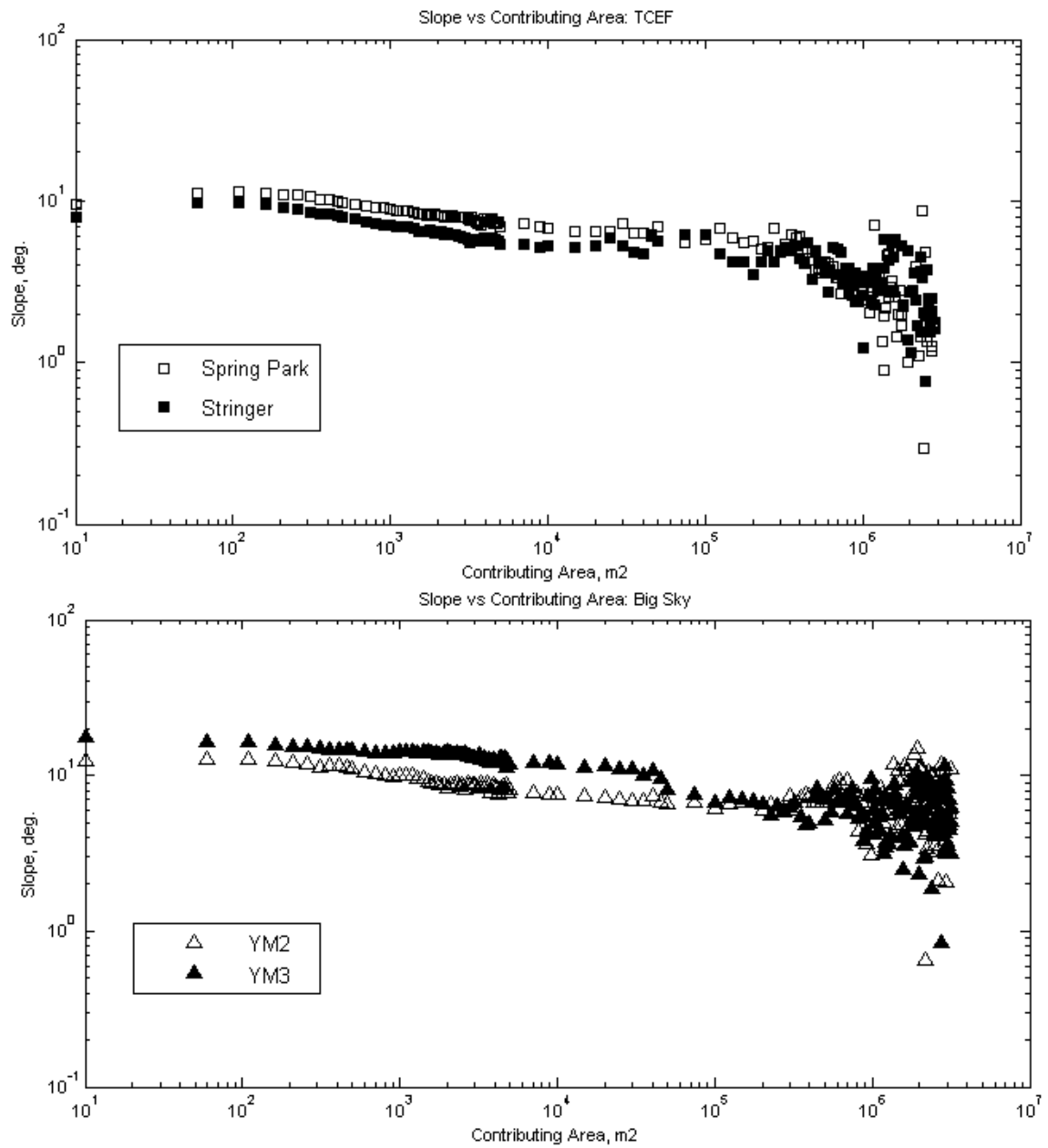


Figure 2.12: Contributing area versus slope, TCEF and Big Sky catchments.

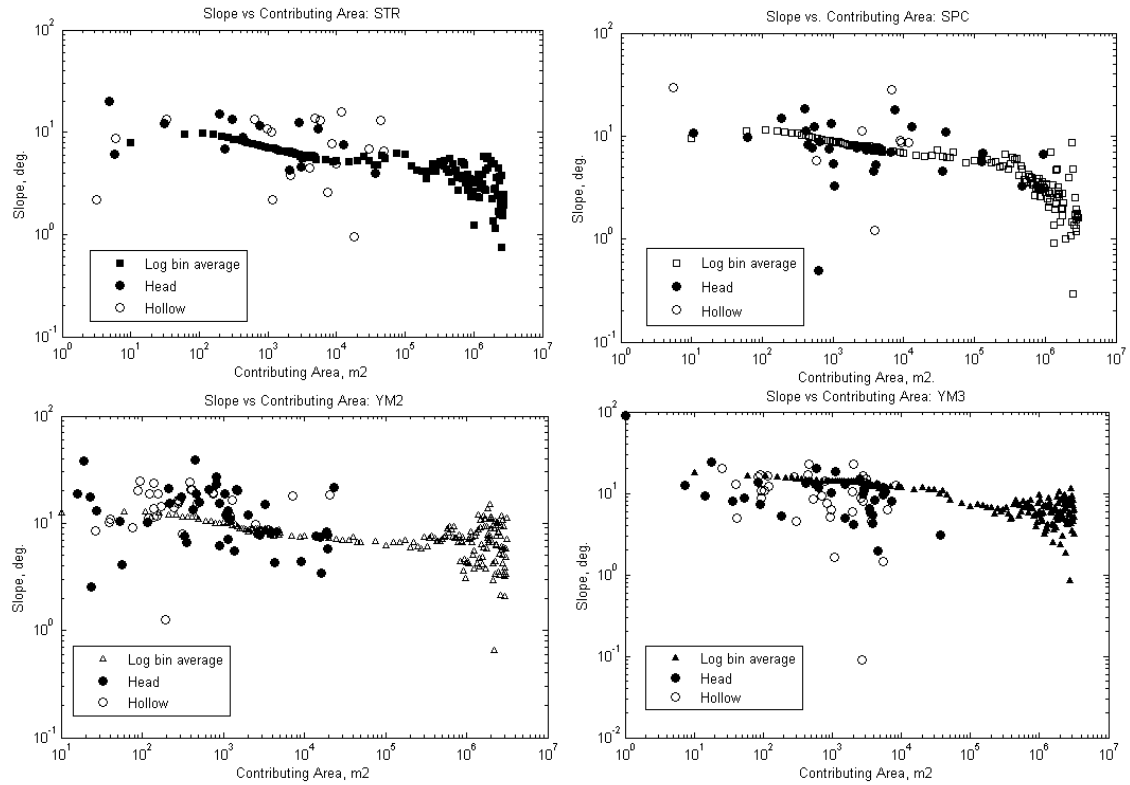


Figure 2.13: Contributing area versus slope and mapped hollows and heads (STR, SPC, YM2, YM3).

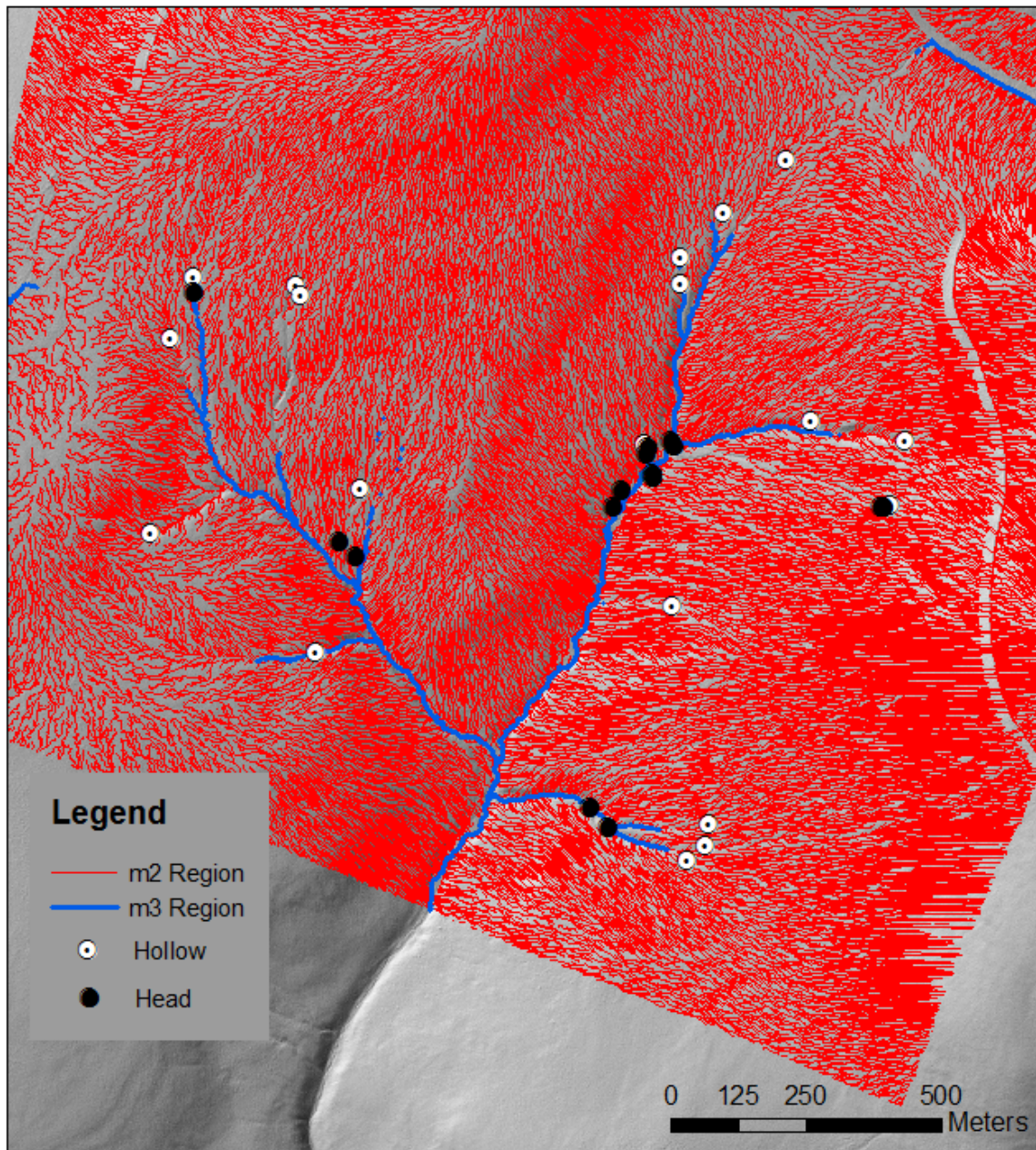


Figure 2.14: Process zones as delineated by flowpath length - contributing area relation, STR.

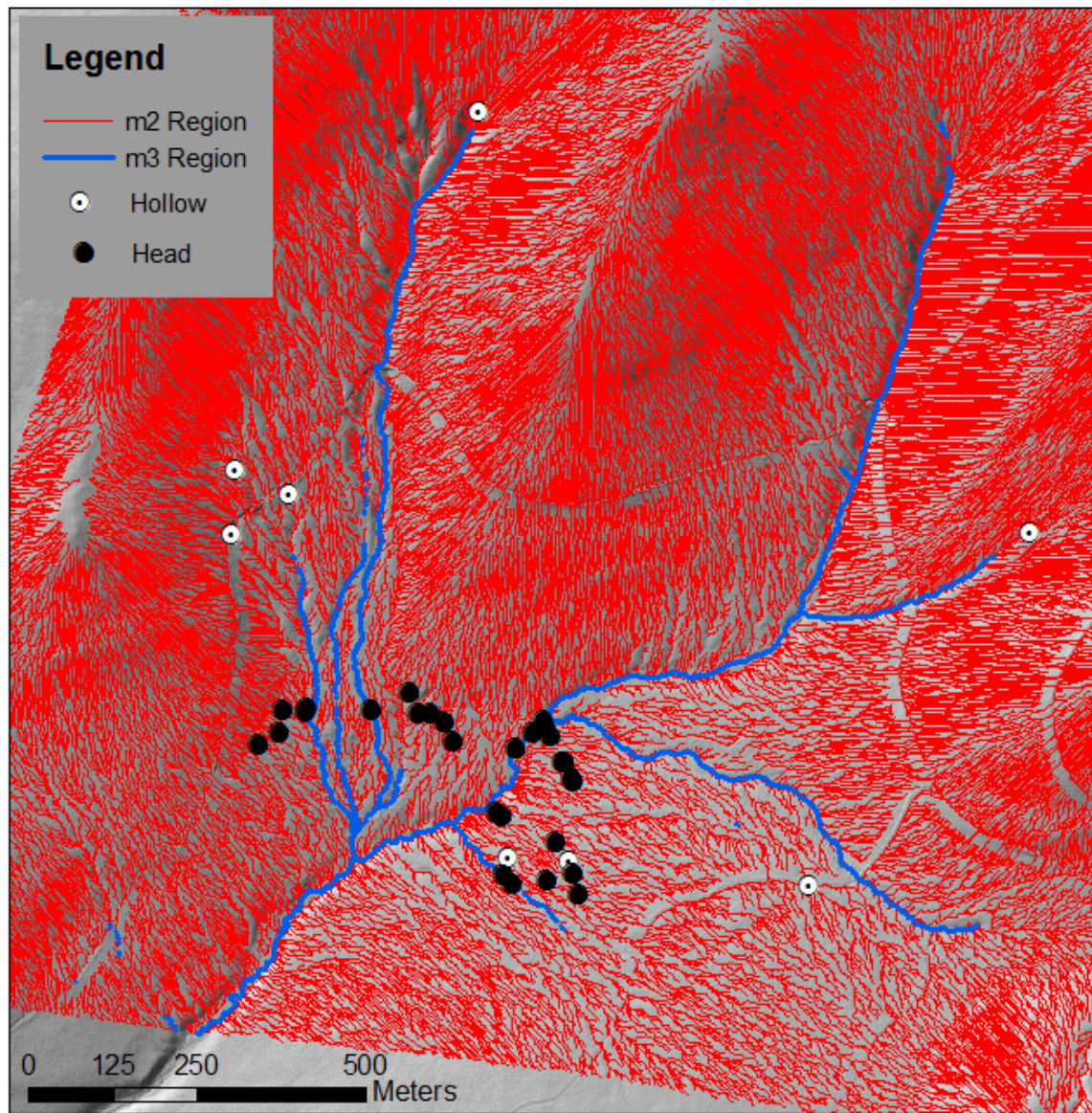


Figure 2.15: Process zones as delineated by flowpath length - contributing area relation, SPC.

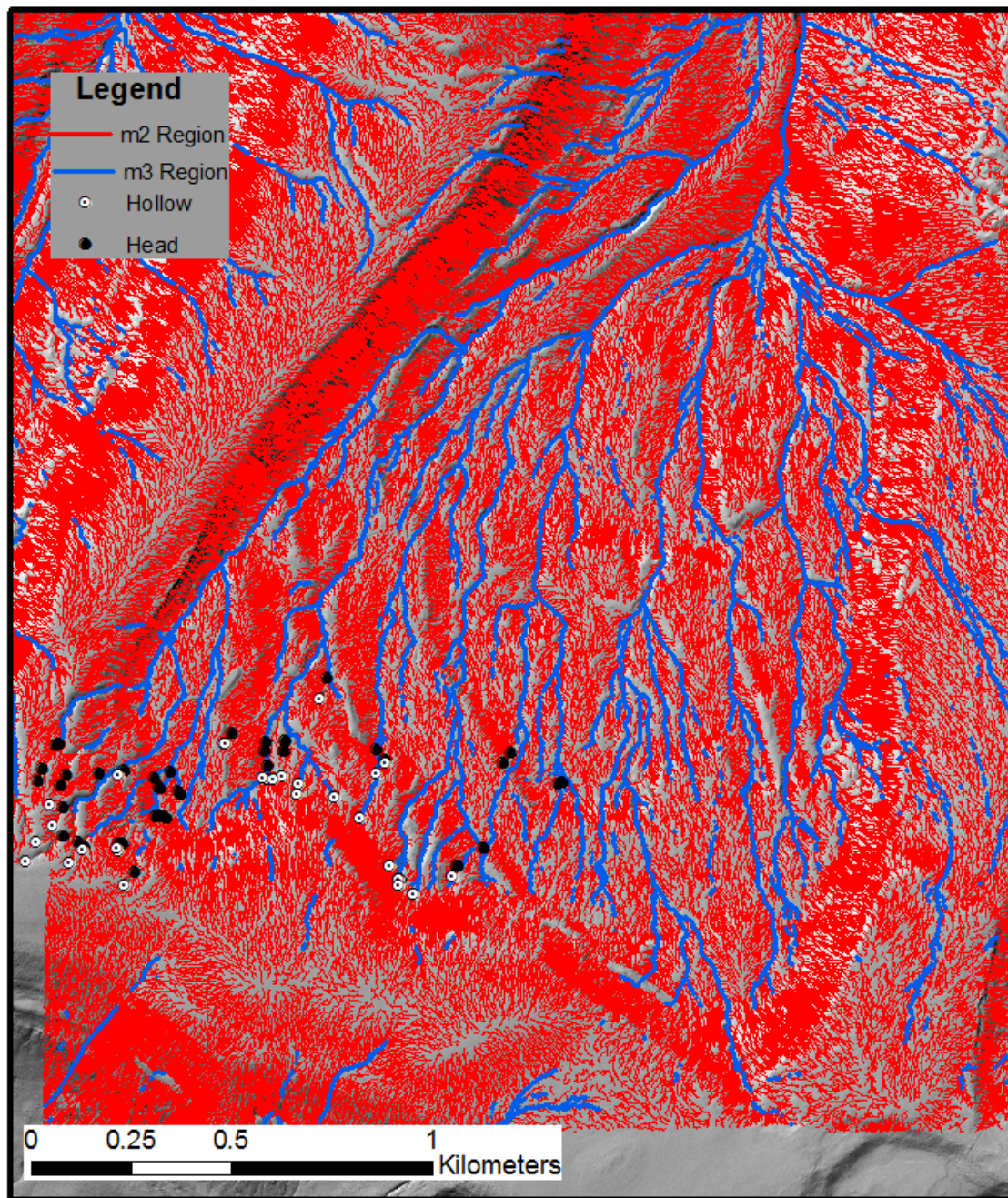


Figure 2.16: Process zones as delineated by flowpath length - contributing area relation, YM2.

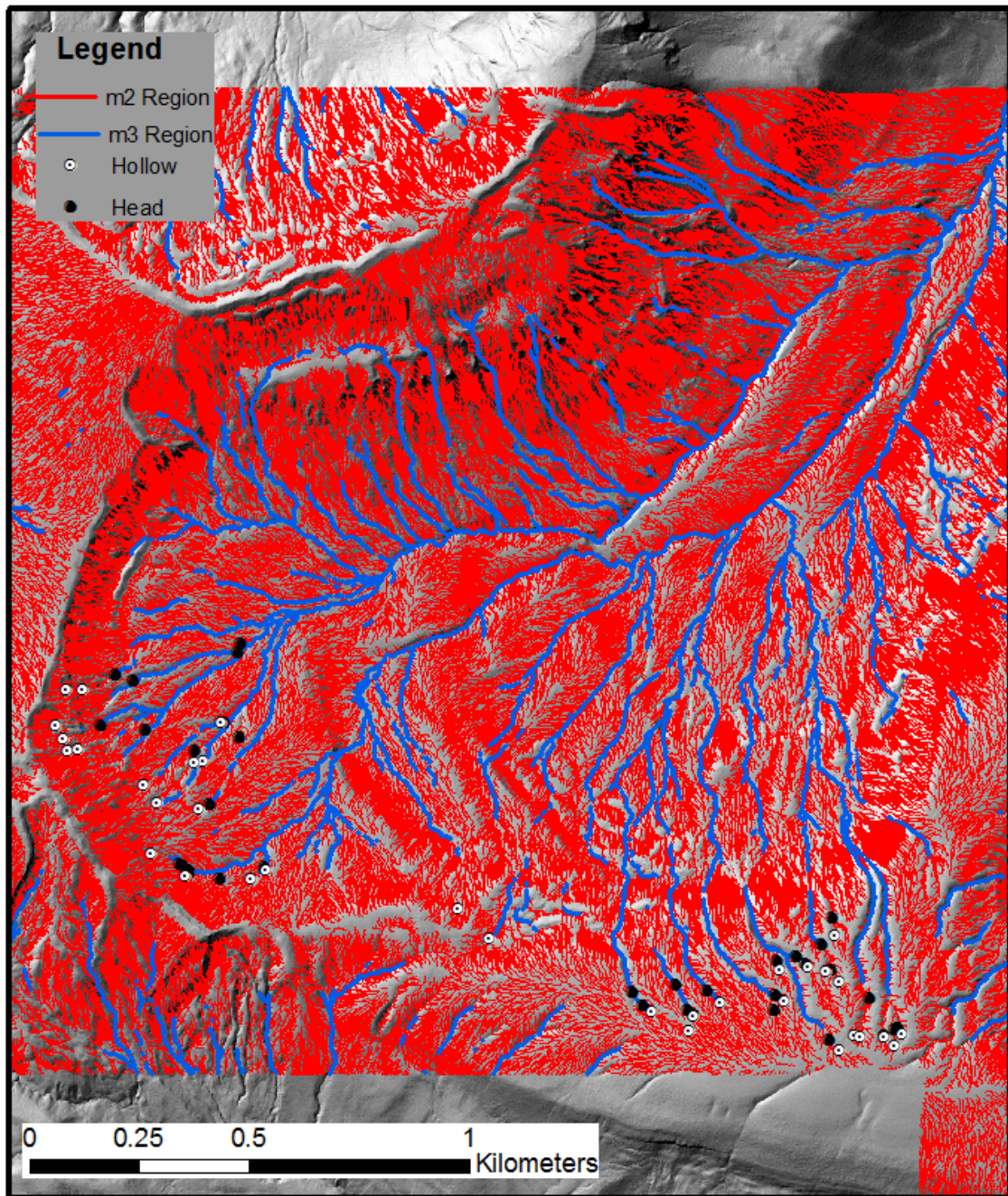


Figure 2.17: Process zones as delineated by flowpath length - contributing area relation, YM3.

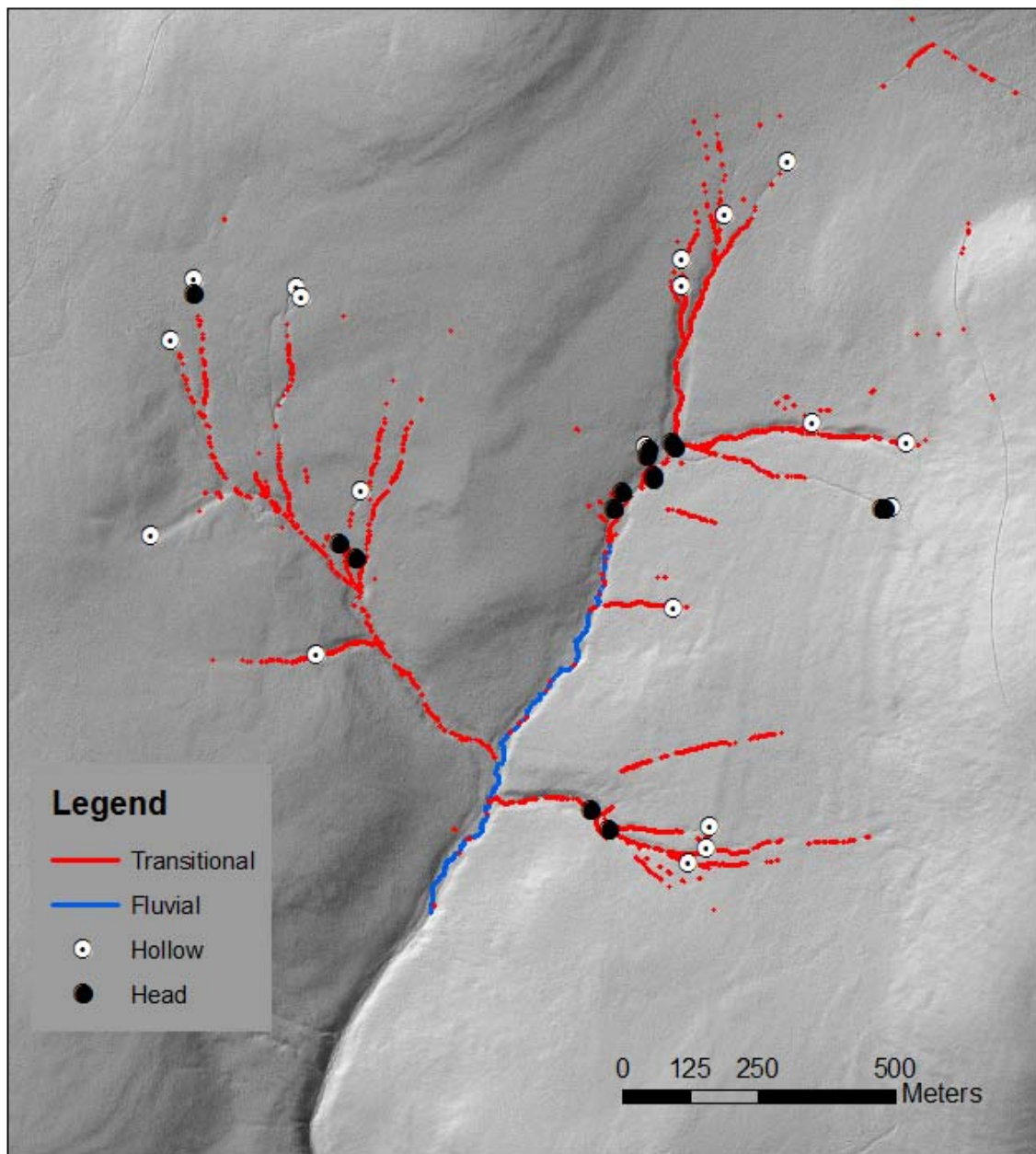


Figure 2.18: Process zones as delineated by unit stream power - contributing area relation, STR.

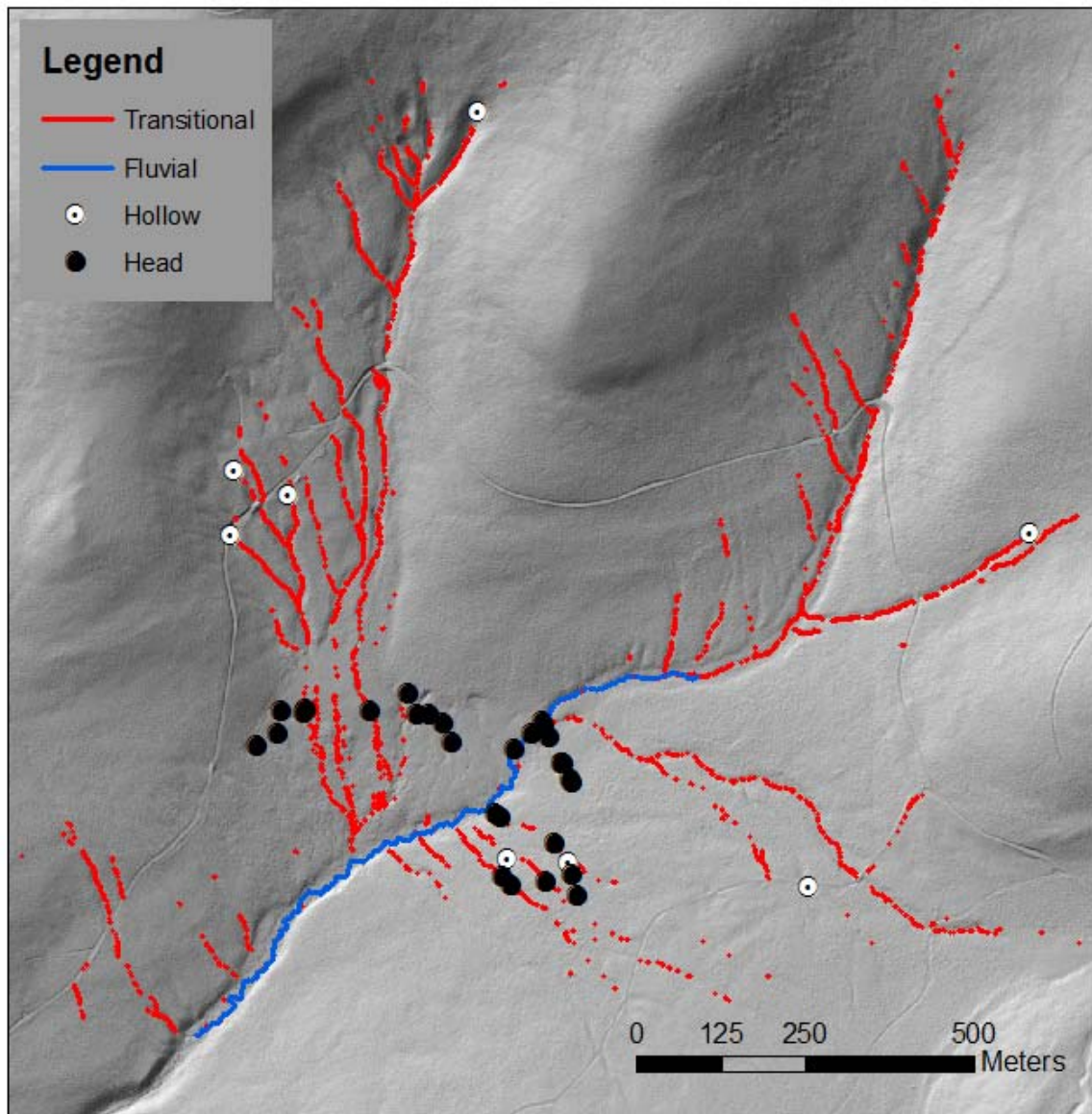


Figure 2.19: Process zones as delineated by unit stream power - contributing area relation, SPC.

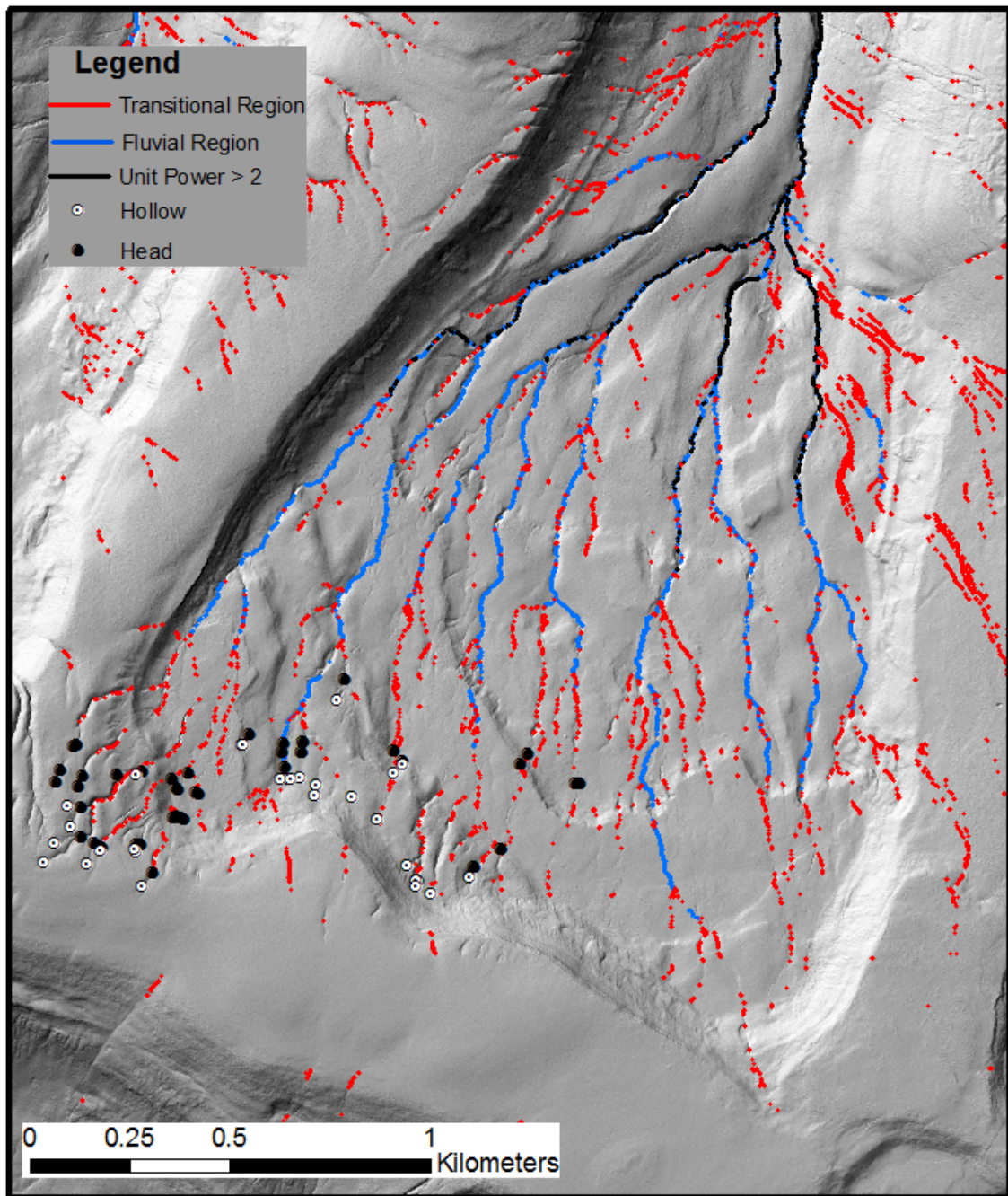


Figure 2.20: Process zones as delineated by unit stream power - contributing area relation, YM2.

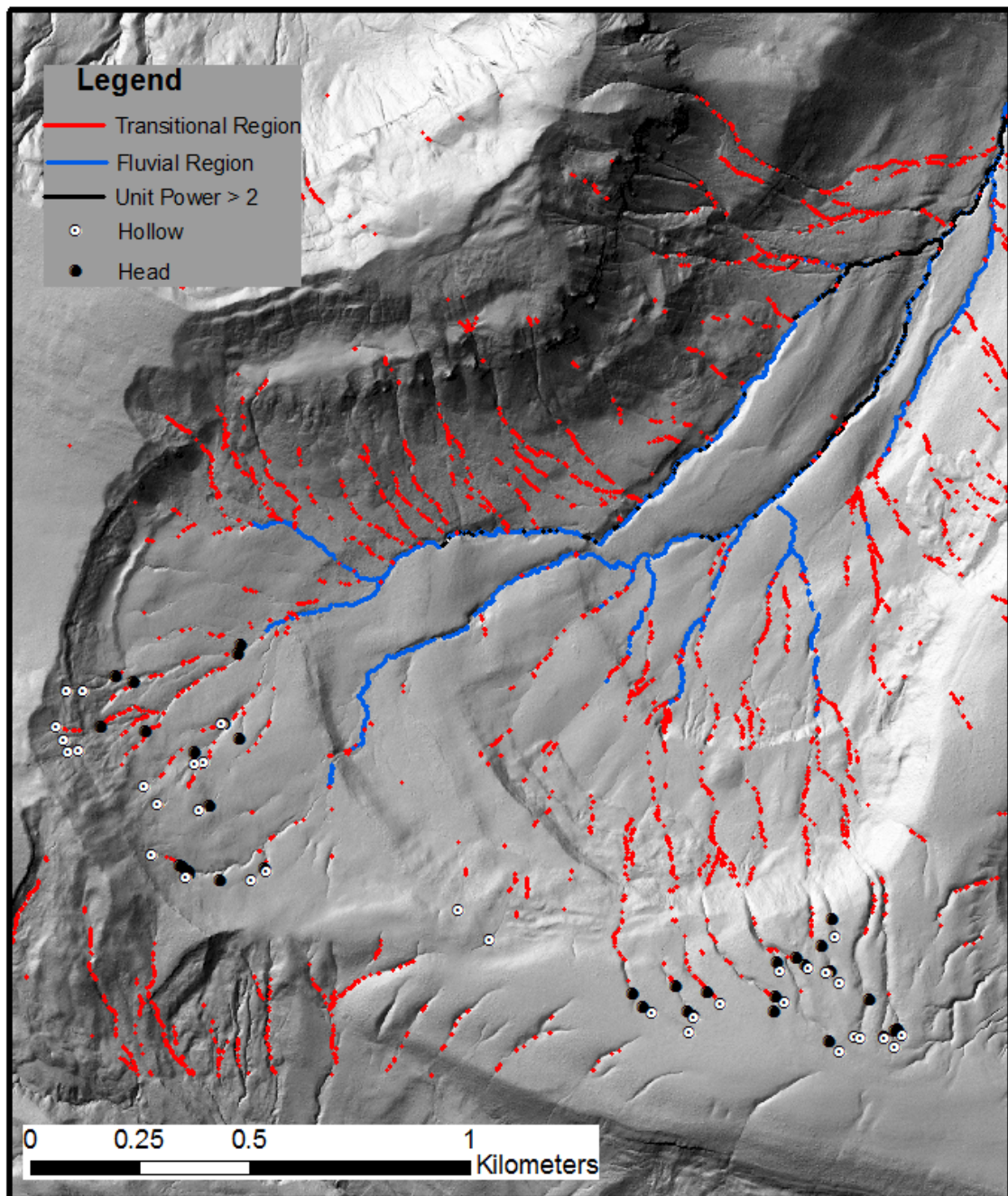


Figure 2.21: Process zones as delineated by unit stream power - contributing area relation, YM3.

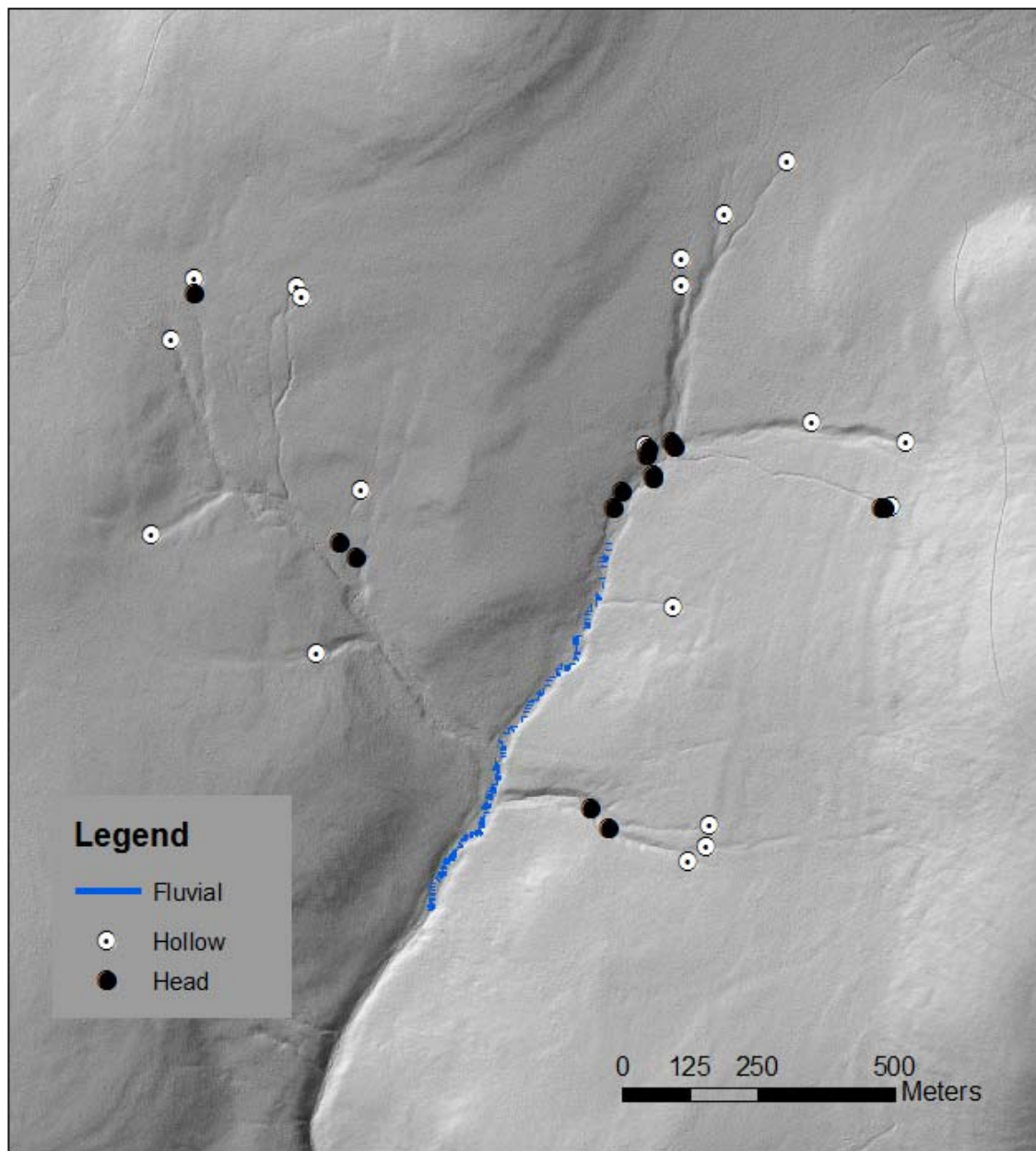


Figure 2.22: Process zones as delineated by slope - contributing area relation, STR.

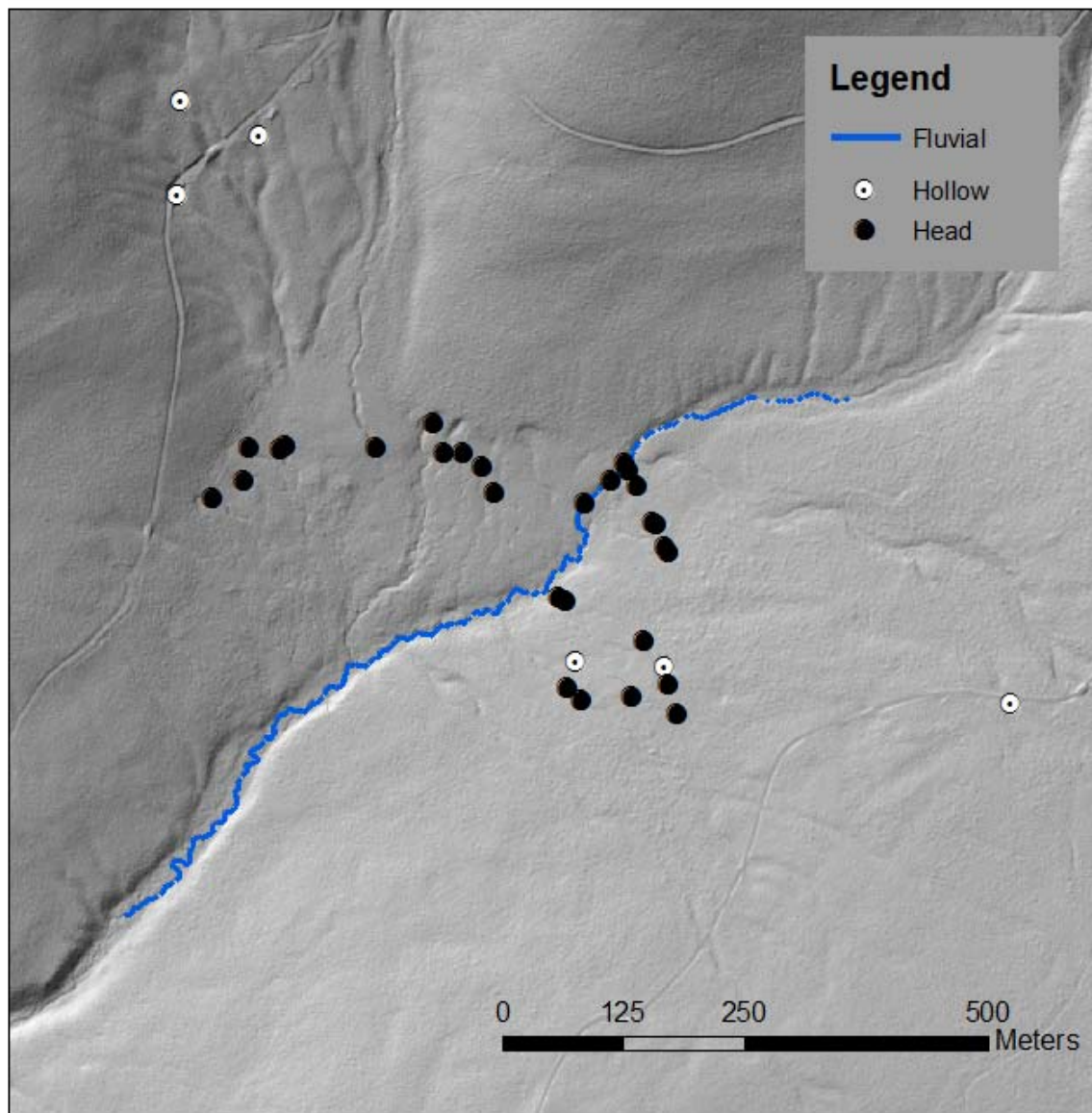


Figure 2.23: Process zones as delineated by slope - contributing area relation, SPC.

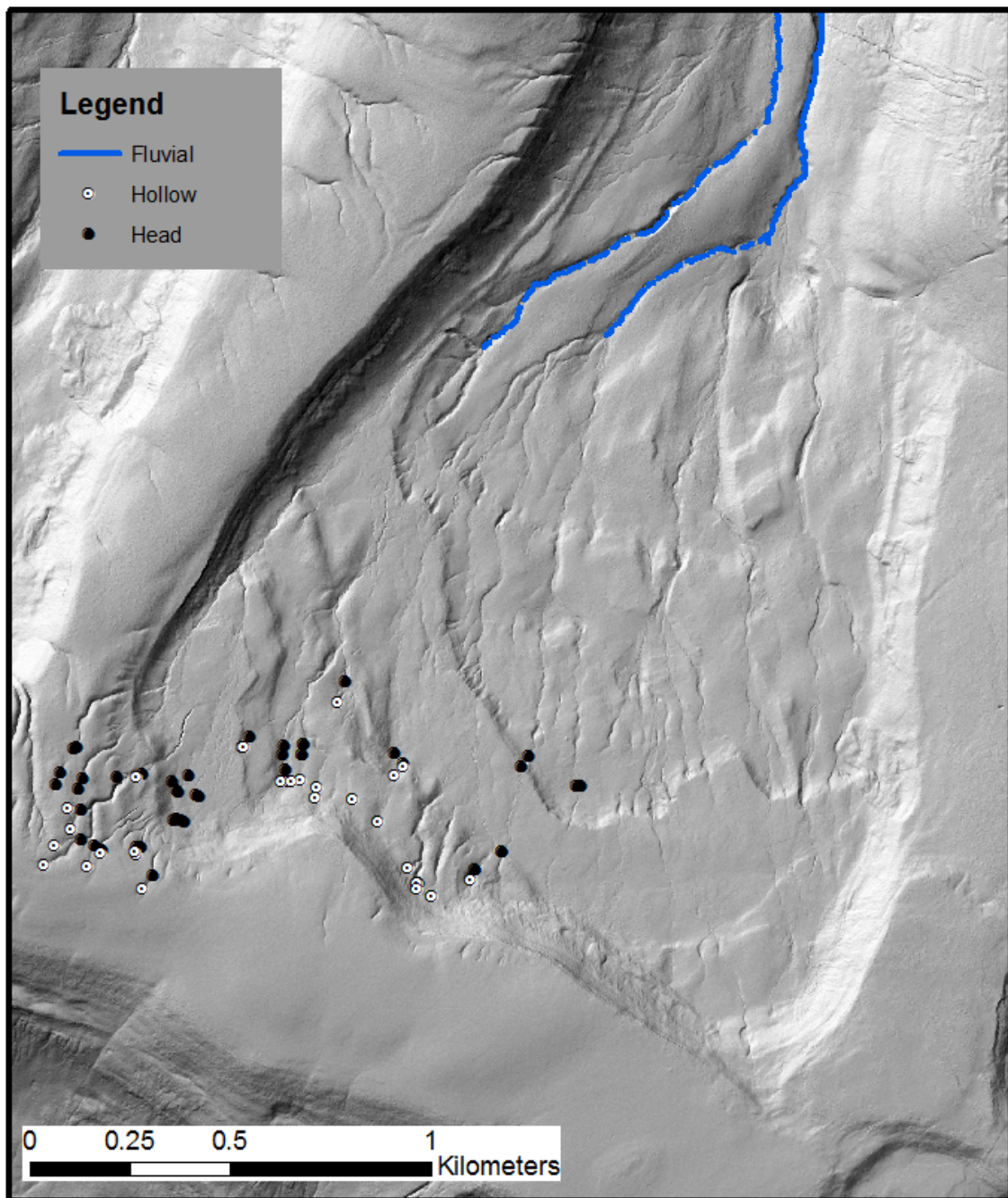


Figure 2.24: Process zones as delineated by slope - contributing area relation, YM2.

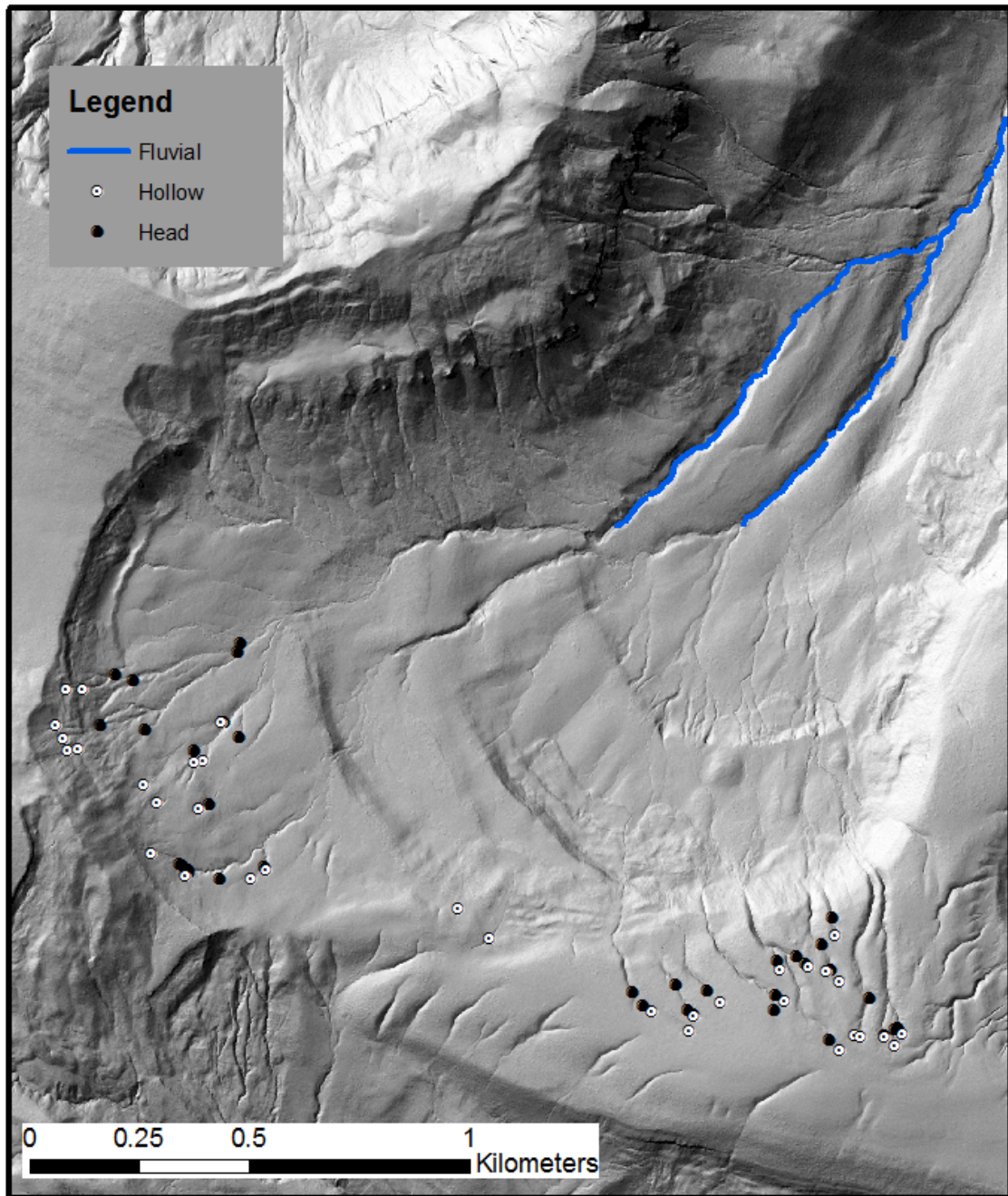


Figure 2.25: Process zones as delineated by slope - contributing area relation, YM3.

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CHAPTER THREE

DETECTION OF HOLLOWS AND CHANNEL HEADS USING SPECTRAL
ANALYSIS OF DEMS

Contribution of Authors and Co-Authors

Manuscript in Chapter 3

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Contributions: Conceived the study, collected and analyzed output data, and wrote the manuscript.

Co-author: William W. Locke III

Contributions: Discussed the results and implications and edited the manuscript at all stages.

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Abstract

The advent of high resolution LIDAR-derived topography has facilitated the detection of landforms and their spatial patterns. Spectral analysis of topography can identify wavelengths of fine- and coarse-scale landforms as well as their underlying periodic structure. Periodic structure is revealed by spectral analysis of the strength of periodic components of a signal at different frequencies. Two types of spectral analysis (Fourier and wavelet transforms) are applied to 1 m. DEMs of two pairs of headwater catchments in central and southwest Montana. Using the resulting energy spectrum, portions of the landscape with long wavelengths, corresponding to coarse-scale landforms, are filtered from the DEMs. The remaining portions of the landscape represent fine-scale features such as headwater channels. Magnitude-frequency distributions of curvature and the derivative of aspect were derived from the filtered DEMs. These plots exhibit a rollover to power law behavior that is hypothesized to represent the transition from scale-dependent hillslope process to scale-invariant fluvial process. This hypothesis is tested using field-mapped hollow and channel heads. Results of spectral filtering indicate that wavelet filtering sharpens the rollover to power law behavior in magnitude-frequency distributions. Most field-mapped hollows and heads have values of curvature and derivative of aspect near the rollover zone. The association of heads and hollows with this rollover zone may be indicative of a change in process domains. However, while prediction at the DEM scale may be possible,

deterministic prediction at individual locations is limited by local interactions of lithology, slope, and hillslope hydrology.

Introduction

Methods of detection of landforms using high resolution topographic data are increasing in number and use, facilitated by the increasing availability of airborne Light Detection and Ranging (LIDAR) data. The high resolution offered by LIDAR enables quantitative analysis of fine-scale landforms associated with hillslope and fluvial processes. However, the classification and processing of last return data has intrinsic error that results in an imperfect digital elevation model (DEM) of the ground surface. In order to minimize the effect of error on subsequent topographic and geomorphic analyses, signal processing techniques such as spectral filtering are becoming popular means of treating topographic data as a signal. Landforms also have different spatial scales requiring different scales of analysis (Schmidt and Andrew, 2005) and spectral filtering can be used to select the spatial scale appropriate to the landform of interest. Spectral analysis is the transformation of data from the spatial domain to the frequency domain, where it can be interpreted on the basis of inherent periodicities. Here, wavelet and Fourier transforms are used to emphasize fingertip tributaries and remove larger scale surface features.

The spectral characterization of land and sea surfaces has been used to identify morphotectonic lineaments (Jordan and Schott, 2005) and landslides (McKean and Roering, 2004; Glenn et al., 2006; Booth et al., 2009), and delineate channel networks

(Lashermes et al., 2007). In addition to identification of landforms, periodic structures can be detected such as spatial patterns of seafloor faults (Little, 1994), ocean currents (Zhang and Lu, 2005), channel morphology (McKean et al., 2008), and harmonics of topography (Balmino, 1993; Harrison and Lo, 1996). High and low pass spectral filtering of DEMs has also been used to define the wavelength of landforms such as ridges, valleys, and channel banks (Perron et al., 2008) and characterize the morphology of mima mounds (Reed and Amundson, 2011).

Evans (2012) suggests identifying channel networks by delimiting forms based upon slope position, where breaks in gradient, aspect, or curvature are considered. This approach has also been used to identify glacial features including drumlins (Saha et al., 2011). Recent work (Lashermes et al., 2007; Perron et al., 2008; Tarolli et al. 2009; Passalacqua et al., 2010) has utilized statistical behavior of populations of local curvature and aspect to directly detect and extract channel networks and landforms from DEMs. For example, Lashermes et al. (2007) perform a wavelet analysis of DEMs to calculate curvature and aspect. The deviation from a Gaussian distribution of these parameters is hypothesized to indicate incipient hollows and channel heads. Tarolli et al. (2009) used multiples of the standard deviation of curvature to define threshold values for channel initiation. Both of these methods utilize landform curvature i.e.; the Laplacian of elevation. The Laplacian is the Laplace operator is the divergence (or flux) of the gradient. In Cartesian coordinates, the Laplacian is the sum of second partial derivatives of a function (here, elevation) with respect to the independent variables, (here, x- and y-coordinates). Passalacqua et al. (2010) followed a similar procedure to Lashermes et al.

(2007), first smoothing the DEM data by filtering. The geometric curvature rather than the Laplacian curvature was also used to normalize curvature values.

Another method of assessing populations of curvature and aspect values is magnitude-frequency distributions. These relations are commonly log-log plots of the frequency of an event or parameter versus its magnitude or size. Magnitude-frequency distributions of many natural phenomena such as landslide size, earthquake intensity, flood size, and wildfire size exhibit an inverse power law relationship, that is, high frequency of small events and low frequency of large events. A power law relationship in magnitude-frequency distributions indicates scale-invariant behavior, and is often associated with self-organized criticality (SOC), where a system is in dynamic equilibrium near a threshold condition (Phillips, 1999). In numerical models, the evolution of channel networks has been associated with SOC systems (Rinaldo, 1993) and the degree of SOC of a channel network has been associated with catchment morphology and shape and the degree of connectivity within the channel network (Coulthard and Wiel, 2007). Channel networks exhibit a known scale-invariance (Rodriguez-Iturbe and Rinaldo, 1992) but the landscape retains scale dependence at the hillslope length. The characteristic scale will vary with inherited controls such as relief, lithology, and catchment shape but in every catchment a "specific scale emerges in the landscape at the transition from the river to the hillslope domain" (Dietrich and Montgomery, 1998, p. 41). Properties of channel networks such as hydraulic geometry relations have been shown to exhibit scale-invariant properties (Leopold and Maddock, 1953) which break down at the scale of the hillslope (Wohl, 2004).

Magnitude-frequency distributions are frequently used in studies of erosion and sedimentation associated with landslides and debris flows (Pelletier et al., 1997; Guzzetti et al., 2002; Brardinoni and Church, 2004). These studies have observed a "rollover" zone where the magnitude-frequency distribution begins to exhibit a power-law form in log-log space. The initiation of power-law scaling indicates the characteristic scale of one process and the limits of scale-invariance of another. Scale-specificity, or breaks in magnitude-frequency distributions has been used to identify limits of scaling in magnitude-frequency distributions (Brardinoni and Church, 2004). One explanation of this rollover is that it is related to the transition between hillslope length and the development of channel networks (Guzzetti et al., 2002), as rainfall and snowmelt driven landslides depend on slope length, and channel network topology, and how hillslopes generate runoff.

Here, the rollover to power law behavior in magnitude-frequency distributions of curvature and the derivative of aspect is hypothesized to indicate the transition between the hillslope and fluvial domains by linking fluvial process to scale-invariance and hillslope process to scale-dependence. Spectral filtering is applied to the DEM to limit the population of curvature and aspect values, as much as possible, to cells belonging to the channel network. Limiting values to the channel network may "sharpen" power law scaling of the magnitude-frequency distributions. Portions of hillslopes may have curvature values in common with channelized portions of the landscape, but they do not represent the organized, convergent areas associated with valley dissection.

Methods

Field Mapping of Hollows and Channel Heads

The locations of a subset of hollows and channel heads were mapped using a mapping-grade Global Positioning System (GPS) unit in the four catchments. Mapped locations were post-corrected to an accuracy of less than 0.5 meters. For field identification, hollow heads were defined as the upstream-most extent of a valley, surrounded by concave-upwards topography on three sides, or concave-upwards plan curvature. Channel heads were defined as an abrupt upstream termination of a persistent incision in the valley floor, either discontinuous or continuous, with definable streambanks. A definable bank "must be recognizable as a morphological feature independent of the flow" (Dietrich and Dunne, 1993, p. 178) with characteristics such as "a narrow zone of gradient steeper than both the neighboring hillslope and the transverse slope of the channel bed" (Dietrich and Dunne, 1993, p. 178).

DEM Pre-processing

The TCEF and Big Sky project areas were mapped in 2005 by NCALM (National Center for Airborne Laser Mapping) using an Optech 1233 Airborne Laser Terrain Mapper mounted in a Piper Chieftain aircraft. For each project site, ground-based GPS reference stations, and on-board inertial measurement unit data were processed by the National Geodetic Survey Laboratory to produce the best estimate aircraft trajectory. For the Big Sky project site, the standard deviation of the differences between the nearest neighbor laser point and the ground truth points was roughly 7 cm. This trajectory and the

raw laser range data were used by NCALM to produce a laser point dataset consisting of the northing, easting, height, and intensity of first and last laser returns. Last return points represent the lowest surface intersected by the laser pulse and consequently were the only signals retained for further processing. Ground classification of last return points using iterative development of a triangulated surface model was performed by NCALM using Terrascan software (Soininen, 2004) to produce a triangulated surface model.

The resulting ground points were gridded to 1 m. grid cell size DEMs (in ESRI GRID format). These 1 m. DEMs for TCEF and Big Sky were used as the starting point for this analysis. The point density, classification and processing of last return data has intrinsic error that results in a slightly inaccurate DEM which may not capture small landforms such as headwater channels. In order to minimize the effect of DEM error on subsequent analyses, the TCEF and Big Sky DEMs were filtered using two-dimensional discrete wavelet transform (2D DWT) and discrete Fourier transform (2D DFT) to retain small features.

Two-dimensional Discrete Wavelet Transform Filter

A wavelet is a family of functions of irregular and asymmetric waveform, with limited duration and zero mean (Figure 3.1). The wavelet transform decomposes a spatial signal into energy or amplitude over given spatial frequencies, at each position in the data. Wavelet analysis adjusts its scale to the frequency of interest such that fine

scale spatial resolution is used at high frequencies and coarse scale resolution at low frequencies.

The continuous wavelet transform converts a signal, here, elevation as a function of position in the x, y plane, $z(x, y)$, into a function $C(s, a, b)$ of scale parameter, s , and position (a and b) by convolving $z(x, y)$ with scaled and shifted versions of ψ , the wavelet function. The coefficient, a , compresses or stretches the wavelet function ψ and the coefficient, b , translates it. The general form of the 2D continuous wavelet transform for elevation $z(x, y)$ and wavelet scale parameter s is

$$C(s, a, b) = \frac{1}{s} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} z(x, y) \psi\left(\frac{x-a}{s}, \frac{y-b}{s}\right) dx dy \quad (3.1)$$

where ψ is the wavelet family (Kumar and Fouloula-Georgiou, 1994). In the DWT, a wavelet is scaled and translated only by discrete values, usually a power of two. The resulting wavelet is of the form

$$\psi\left(\frac{x-2^j}{s}, \frac{y-k2^j}{s}\right) \quad (3.2)$$

where j and k are integers.

The decomposition of a signal in a DWT results in a low frequency component (approximations), and a high frequency component (details). The decomposition process can be iterated, with successive approximations being decomposed in turn, so that one signal is broken down into many lower resolution components. The number of decompositions is referred to as the number of levels. Every level of decomposition results in a set of approximation coefficients and a set of detail coefficients. The low and

high frequency components can be reassembled in a process called reconstruction, by performing an inverse discrete wavelet transform.

The discrete reconstruction of the original signal has the form

$$z(x, y) = \sum_j \sum_k C(j, k) \psi_{j,k}(x, y) \quad (3.3)$$

If only the high frequency components of the signal are of interest, the detail coefficients may solely be used to reconstruct the signal.

For this analysis, the DEMs of YM2, YM3, STR and SPC were decomposed using a 2D DWT in MATLAB using the Haar wavelet. The Haar wavelet (a step function) was selected because the channel banks are likely to resemble the discontinuous form of this function. The detail coefficients were then used to reconstruct the DEMs, without the low frequency components of topography, to emphasize fine scale features of the landscape.

Two-dimensional Discrete Fourier Transform Filter

The Fourier transformation is a mathematical means of resolving or decomposing a signal into its constituent frequencies, and is usually performed on a detrended, one dimensional time series of data, though space can be substituted for time. The Fourier transform takes a signal, or series of data, from the time (spatial) domain to the frequency domain, where it is represented by harmonics and their frequencies. The result is the amplitude of topographic features over a range of spatial frequencies. The Fourier transform can be applied to either discrete data or a continuous series or function. In the

case of elevations in a matrix (i.e.; a DEM), these data are discrete and require a two dimensional Fourier transform.

The DFT of a DEM, consisting of N_y rows and N_x columns of elevation values is written (Priestley, 1981; Perron et al., 2008) as

$$Z(k_x, k_y) = \sum \sum z(m\Delta x, n\Delta y) e^{-2\pi i \left(\frac{k_x m}{N_x} + \frac{k_y n}{N_y} \right)} \quad (3.4)$$

where k_x and k_y are wavenumbers in the x- and y-directions, Δx and Δy are grid spacing intervals, and m and n are indices in Z , the elevation array.

The resulting periodogram is an estimate of the power spectrum, or variance of elevation with frequency.

$$V_{DFT} = \frac{1}{N_x^2 N_y^2} |Z(k_x, k_y)|^2 \quad (3.5)$$

which has units of amplitude squared. The periodogram is a means of assessing how much of the variance (or power) of elevation is found over a range of frequencies. The resulting two dimensional array contains the amplitudes at the frequency components of elevation, z . Because we are working in units of length, frequency is 1/length.

The 2D DFT of the DEMs of all four catchments was taken using MATLAB scripts (Perron, 2010). A one dimensional periodogram was generated by converting the power spectrum and frequency matrices to vectors. Multiple peaks are observed but to assess their significance, an estimate of the background spectrum was made. A representative mean spectrum was generated by using a power law fit to the binned average values. The observed spectrum is divided by the mean spectrum to obtain a

normalized spectrum (Figure 3.2). A mean spectrum is biased by the observed peaks but the normalized spectrum is used to isolate a range of relevant high frequencies corresponding to channelized features, rather than a single frequency. The frequencies identified in the normalized spectrum were used to construct a Gaussian high pass filter, which by definition keeps high frequency components of the signal while removing low frequency components. The detrended DEM is filtered using a MATLAB script implementing a high pass filter, F_{high} (Perron, 2010).

$$F_{\text{high}} = \begin{cases} \exp\left(\frac{-(f - f_2)^2}{2\sigma^2}\right) & f < f_2 \\ 1 & f \geq f_2 \end{cases} \quad (3.6)$$

where f is frequency, f_2 is the frequency at the center of the Gaussian function, and σ is standard deviation.

This process of normalizing each periodogram and selecting the range of frequencies at which to filter was repeated for each DEM. Based upon inspection of each periodogram, two transition frequencies (wavelengths) were selected to define the edges of the filter. The lower frequency is the frequency at which the high pass filter starts to increase above zero. The higher frequency is the frequency at which the high pass filter reaches one. These wavelengths (inverse of frequency) are tabulated below.

Table 3-1: Range of wavelengths applied in high pass filter.

Catchment	High transition wavelength (m)	Low transition wavelength (m)
STR	8	5
SPC	8	6
YM2	10	7
YM3	20	5

Magnitude-frequency Distributions of Curvature

Using the unfiltered and filtered DEMs, curvature was calculated at each 1m. DEM cell in ArcGIS 9.3 and MATLAB. Local curvature is calculated as the second derivative of elevation, or the Laplacian of the function defining the ground surface. Convergent and persistent areas of cells of positive curvature values are upwardly concave, likely occurring in valleys. Grid cells of convex curvature values were replaced by null values. Hollow and channel heads were also field-mapped using a mapping grade Global Positioning System (GPS) receiver.

Magnitude-frequency Distributions of Derivative of Aspect (dthet)

Aspect, θ_a , is the steepest downslope direction from each DEM cell to its neighbors. It can also be thought of as the slope direction or the compass direction that a slope faces. Aspect can be measured clockwise in degrees from 0, due north, to 360, again due north, coming full circle. A significant change in aspect of adjacent cells may indicate a sharp change in slope or incision in the land surface, characteristic of a channel. The change, or derivative, of aspect could detect zones of incision however

aspect measurement in degrees is problematic at due north. Near the origin, the derivative of aspect could be anomalously large, due to a large change from one side of the origin (0° to 360°) to the other. To circumvent this singularity at due north, aspect was calculated in ArcGIS twice, using an origin of due north and an origin of due south. The derivative of aspect was calculated on each elevation raster in ArcGIS using the "Slope" function, which calculates the steepest change between a center grid cell and its eight neighbors. A composite derivative of aspect raster was constructed using the minimum value of the derivative at each cell, selected from the two rasters of different origins. Using this composite raster of the derivative of aspect, cells with the largest aspect derivative value, $d\theta$, are more likely to occur in channelized or incised zones.

Using the methods described above, magnitude-frequency distributions of curvature and the composite derivative of aspect raster, $d\theta$, were calculated in MATLAB for each of the four catchments. Three 1 m. DEMs were used for each catchment: unfiltered, high pass filtered, and wavelet filtered. The filtered DEMs were used to emphasize small topographic features. The values of curvature and $d\theta$ associated with each mapped head and hollow were extracted by spatial analysis in ArcGIS.

Results

Hillshade imagery of the unfiltered and filtered DEMs was created for qualitative comparison. A portion of each hillshade image is given in Figures 3.3 - 3.6. The unfiltered DEMs are highly textured compared to the filtered DEMs. The wavelet and

high pass filtered DEMs are considerably smoother. The high pass filtered DEMs show accentuated channel banks. The wavelet filtered DEMs show an added rectilinear artifact in the hillshade image. The expected result is that concave-upwards grid cells will be more likely to lie on hollows and channels.

Using the unfiltered, high pass filtered, and wavelet filtered DEMs, log bin averaged magnitude-frequency distributions of curvature (Figures 3.7-3.10) and d_{thet} (Figures 3.11-3.14) were produced for each catchment. Magnitude-frequency distributions of curvature show distinctive signatures between the three types of DEMs, but the forms of the signatures are similar within the paired catchments and between TCEF and Big Sky. The unfiltered and wavelet filtered DEMs exhibit a rollover to power law scaling at higher curvature values, but the rollover in the wavelet filtered case is delayed to higher values of curvature. The wavelet filtered case also exhibits a sharpened rollover region. These differences may be due to the filtering process retaining more cells of higher curvature. The unfiltered and high pass filtered distributions share the same rollover region and then diverge at higher curvature values. This may indicate that the effects of high pass filtering is confined to portions of the DEM with greater curvature. The mapped channel heads generally plot on the power law portion of the unfiltered DEM distribution and the rollover portion, rather than the power law portion, of the wavelet filtered distribution.

The wavelet filtered DEM has fewer low curvature values due to filtering of large amplitude, long wavelength features like ridges. Magnitude-frequency distributions of the unfiltered and high pass filtered DEMs are the same at low curvature values but

diverge at higher curvature values. The high pass filter begins allowing features with a wavelength of 8-20 m. (depending on the catchment) and passes all features with a wavelength of 5-7 m. These thresholds may cause the sharp breaks in slope at higher curvature values. Mapped heads generally have higher curvature values than mapped hollows.

Magnitude-frequency distributions of d_{thet} (Figures 3.11 - 3.14) exhibit a rollover to power law scaling until a maximum change in aspect value is reached. Unfiltered and high pass filtered distributions are nearly identical except for the power law region. In this region, the high pass filtered distribution has the same shape as the unfiltered distribution, but has more values for the same d_{thet} . The majority of channel heads plot within the power law region, and most channel heads have higher values of d_{thet} than mapped hollows.

The wavelet filtered distributions have fewer cells at low d_{thet} values. Low values of d_{thet} indicate little change in local aspect. The wavelet distribution does not exhibit power law scaling. At the rollover, the wavelet distribution increases then sharply decreases for all four catchments. This increase may be due to greater changes in aspect between adjacent cells due to the rectangular artifacts introduced by reconstruction of the elevation signal without the long wavelength component. The form of the wavelet filtered distribution is similar for all four catchments: at the rollover, frequency values increase and then decrease as the maximum d_{thet} value is reached.

Discussion

In unfiltered magnitude-frequency distributions of curvature, power law scaling in TCEF begins at values of approximately 0.2 and extends to values of 0.9. In Big Sky, power law scaling begins at values of roughly 0.35 and extends to 2.0. Curvature values at Big Sky hollows and heads also generally exceed curvature values at TCEF hollows and heads. This may be due to the mode of channel initiation. In TCEF, channel initiation is dominated by exfiltration of hillslope water with few upslope hollows, facilitated by juxtaposition of lithologic units of quite different permeability and composition. In Big Sky, though numerous heads are driven by similar lithologic juxtaposition, the difference in permeability between units is not as great. Here, many channels also initiate by threshold exceedence of the critical shear stress of regolith and colluvium, facilitated by upslope-delivery of discharge and sediment in zero order hollows.

The initiation of power law scaling may result from how landforms in each study area "become curved." The interaction between climate, relief, bedrock strength and structure, and geomorphic history may result in a mosaic of landforms with a certain degree of curvature. In Big Sky, the bedrock geology is considerably weaker than in TCEF, as well as having more available relief than in TCEF. The synergy of abundant runoff and weak shales facilitates valley evolution and channel incision such that greater curvature may be more characteristic in this setting. The maximum value of curvature in Big Sky is nearly twice that calculated in TCEF, implying a land surface more prone to erosion by mass wasting and fluvial processes.

In the TCEF catchments, power law scaling begins at an approximate d_{thet} value of 10 and ends when the maximum value of 100 is reached. In Big Sky, power law scaling begins at a d_{thet} value of 15-20 and ends at a maximum value of 100. This difference in initial power law value may indicate that in Big Sky, when hollows and channels begin, the initiation is a threshold-driven, mechanistic process, or channel head initiation zones are relatively "youthful" and not well modified by diffusive process. In the TCEF catchments, a lower value of d_{thet} at the power law rollover may indicate a gradual accumulation of flow or perhaps a well-worn mass wasting scarp.

Curvature, or the slope of the slope, describes the pattern of surface flow upon the landscape, from nonconvergent over convex-upwards surfaces to convergent over concave-upwards surfaces. The slope of the slope may be increased by erosional processes until channelized flow results. The range of curvature values at the mapped heads and hollows is roughly an order of magnitude, approximately between 0.1 and 1.0. This variation may be explained by the variation in modes of channel initiation, from threshold exceedence of surface critical shear stress to exfiltration of subsurface saturated flow. In either case, the range of values is similar irrespective of differences in slope length, degree of dissection, and lithologic composition. This may indicate that a global driver is also influencing the development of channels, and that driver imparts a scale-invariant signature to the population of curvature values.

The fluvial process zones predicted by the power law regions of the magnitude-frequency distributions of curvature, of the unfiltered DEMs, include most of the channels and hollows mapped in TCEF (Figures 3.15 and 3.18). Grid cells on the

hillslopes are also identified with the fluvial process zone, but they represent random, disorganized, convergent patches. When the DEMs are high pass filtered (Figures 3.16 and 3.19), these misidentified cells are largely absent but fluvial process zones no longer predict the locations of mapped hollows as well as some heads. In SPC, in particular, many of the mapped heads are not predicted. This is not unexpected given the nature of channel initiation, as spring heads, and the lack of characteristic channel morphology (curved banks, upslope hollow) of these heads. The reconstruction of the wavelet filtered DEMs (Figures 3.17 and 3.20) creates a rectilinear artifact which disrupts the model of the surface topography. Bjorke and Nilsen (2003) also used wavelets to filter landscape features DEMs. Their results indicate that reconstruction of a wavelet filtered DEM can result in "block effects" unless a threshold procedure is used to determine which wavelet coefficients to retain in the reconstruction. Consequently, none of the wavelet filtered analyses successfully predicted fluvial process zones.

In the Big Sky catchments, magnitude-frequency distributions of curvature of the unfiltered DEMs (Figures 3.21 and 3.24) successfully predicted fluvial process zones up to the mapped channel heads as well as the upslope hollows. The discontinuous nature of the channels is also accurately predicted. However, grid cells on hillslopes and the sandstone lenses are misidentified as fluvial process zones, due to landscape texture. High pass filtering (Figures 3.22 and 3.25) largely removes the effects of landscape texture but at the expense of losing detection of some channel heads and most hollows. The multiple peaks exhibited by the normalized periodogram indicates that more than one type of landform has a significant signal at high frequencies. The results of the high pass

filtering indicate that the range of removed wavelengths is likely too close to the signature of small channels and hollows, and contains some of the same frequencies as the channels and hollows. In other words, the headwater fluvial features here do not have a distinct signature that can be isolated by the employed normalization technique. Predictions of fluvial process zones in wavelet filtered DEMs remained unsuccessful (Figures 3.23 and 3.26).

A sharp change in aspect between a DEM grid cell and any of its eight neighbors may arise for a variety of reasons besides representing a ground surface of opposing channel banks or the edge of a channel bank. The ground surface may be highly textured due to bedrock outcrop, non-diffusive geomorphic process, or human manipulation. Landscape texture may also be introduced in the filtering process. Despite these deficiencies, values of d_{thet} within the power law portion of the unfiltered DEM magnitude-frequency distributions were able to distinguish hollows and channel heads in the Big Sky catchments (Figures 3.33 and 3.36). In TCEF, values of d_{thet} within the power law region correctly identified hollows and channel heads in STR (Figure 3.27), but only hollows in SPC (Figure 3.30). However, the unfiltered cases also identified ridgetops (TCEF) and sandstone lenses (Big Sky) as channel cells. In the TCEF high pass filtered case, misidentified channel cells on ridgetops are removed, but the entire DEM has a uniform background "noise" introduced (Figures 3.28 and 3.31). However, the d_{thet} values within the power law region are now largely confined to channels and hollows. In the Big Sky high pass filtered case, more channelized terrain is identified, particularly those initiating on the east and west divides (Figures 3.34 and 3.37).

However, more channel cells are incorrectly identified on the transverse sandstone lenses. High pass filtering appears to increase landscape texture in the Big Sky catchments more than the TCEF catchments, and consequently increase the change in aspect between grid cells. But in TCEF the increased texture generally does not cause grid cells to be incorrectly classified as channels, as it does in the Big Sky catchments. The wavelet filtered case was unsuccessful in identifying either hollows or channel heads. For all catchments, removing the long wavelength component of the landscape introduced significant noise to the resultant DEM. Multiple attempts at wavelet filtering using different portions of the reconstructed signal, or a different wavelet family, did not yield better results. Both high and low wavelength landforms may also have a mix of ongoing and relict hillslope and fluvial signals that make it difficult to remove one signal alone. Hillslope and fluvial processes may be operating in the same landforms, blurring the distinction between fluvial and hillslope dominated landforms.

Conclusions

Spectral filtering was applied to 1 m. DEMs to quantify the topographic expression of headwaters channels and to examine the statistical behavior of curvature and the derivative of aspect within these landforms using their magnitude-frequency distributions. Hillshade imagery of spectral filtered 1 m. DEMs shows that fluvial landforms can be isolated by removal of hillslope-dominated landforms. However, headwater channels in these catchments do not exhibit a frequency unique to fluvial features. Consequently, filtering partially removes channel and hollow features, as well

as longer wavelength features. Filtering also alters the topographic attributes (curvature, derivative of aspect) that are of interest in characterizing where hollow and channels begin.

Spectrally filtered and unfiltered DEMs were used to create magnitude-frequency distributions of topographic parameters (curvature and the derivative of aspect) characteristic of fluvial landforms. These distributions exhibit a rollover to power law scaling. The values of curvature and derivative of aspect at the mapped channel heads and hollows fall within the power law region of the magnitude-frequency distributions of the unfiltered DEMs. This behavior appears to indicate the onset of fluvial landforms is coincident with the onset of scale-invariance in these topographic parameters. Power law scaling is associated with the scale-invariance of channel networks, and its appearance may be related to the initiation of the channel network within the headwaters.

In general, high pass filtering of the DEMs removes misidentified fluvial process zone grid cells from hillslopes but also loses detection of fluvial zones at some mapped heads and hollows. Magnitude-frequency distributions of high pass filtered DEMs exhibit a similar rollover as unfiltered DEMs but lose the power law scaling form as a result of the filtering process. Magnitude-frequency distributions of wavelet filtering exhibit a sharper rollover than unfiltered DEMs and largely retain the power law scaling form but poorly predict fluvial process zones.

The derivative of aspect appears to be a better predictor of fluvial process zones than curvature, particularly in the Big Sky catchments. This may be due to the deeply incised nature of these channels relative to the TCEF channels. The values of the

derivative of aspect representative of fluvial features are also limited to landforms subject to non-diffusive processes.

In summary, channelized landforms initiating at mapped hollows and channel heads appear to exhibit a scale-invariant, topographic signature in their magnitude-frequency distributions of curvature and derivative of aspect. The onset of this scale-invariance correlates with field mapping of hollows and channel heads, regardless of differences in physiography, geology, and climate of catchments. Spectral filtering may aid in isolating channelized terrain but may require an accurate estimation of the background spectrum and a unique spectral signature unmodified by hillslope process. Landforms such as drumlins or mima mounds may be better candidates for these techniques.

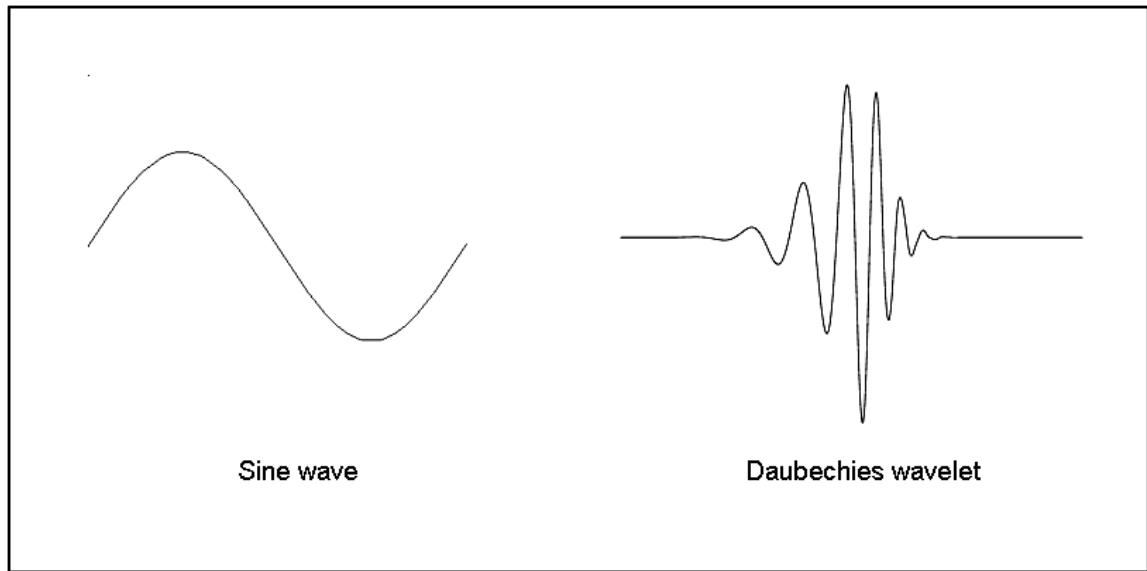


Figure 3.1: Example of symmetric (sine) and asymmetric (wavelet) waveforms.

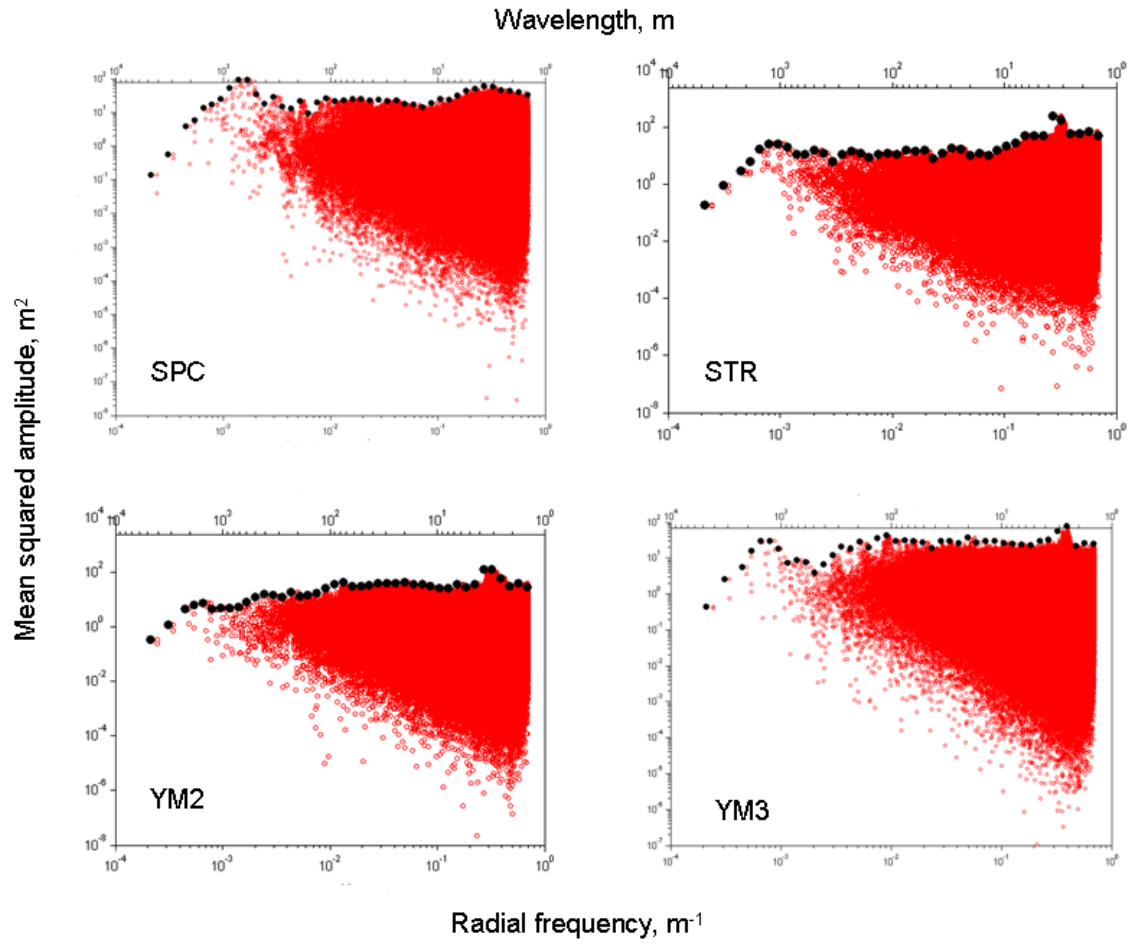


Figure 3.2: Normalized spectra of SPC, STR, YM3, YM2.

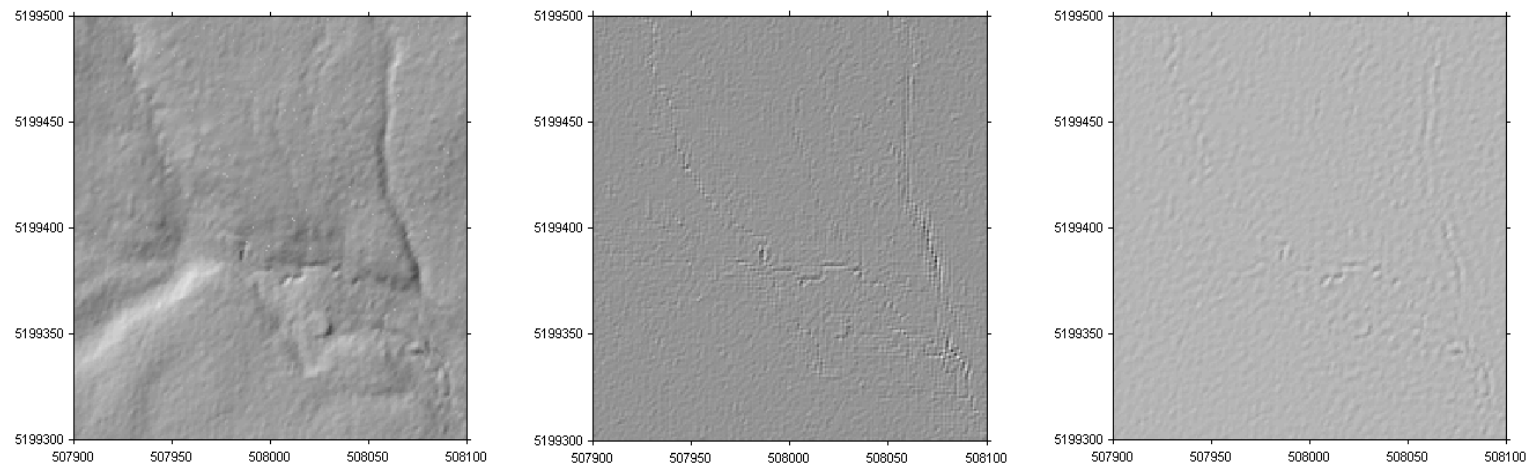


Figure 3.3: Hillshade of portion of DEM, Stringer Creek. From left to right: unfiltered DEM, wavelet filtered DEM, high pass filtered DEM. Scale: 200 m.

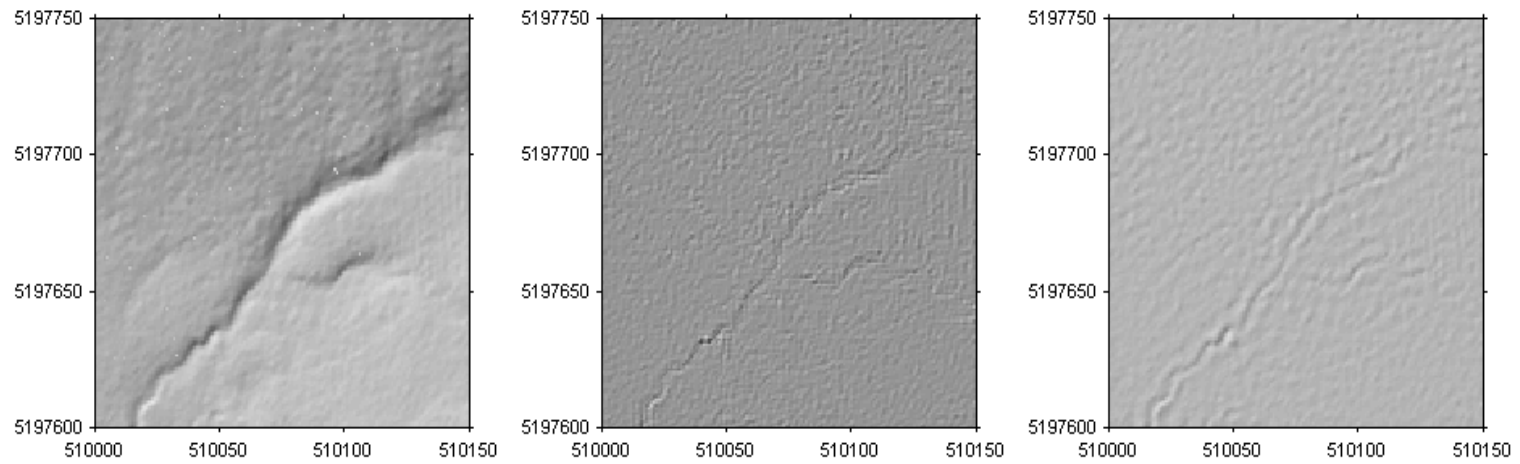


Figure 3.4: Hillshade of portion of DEM, Spring Park Creek. From left to right: unfiltered DEM, wavelet filtered DEM, high pass filtered DEM. Scale: 150 m.

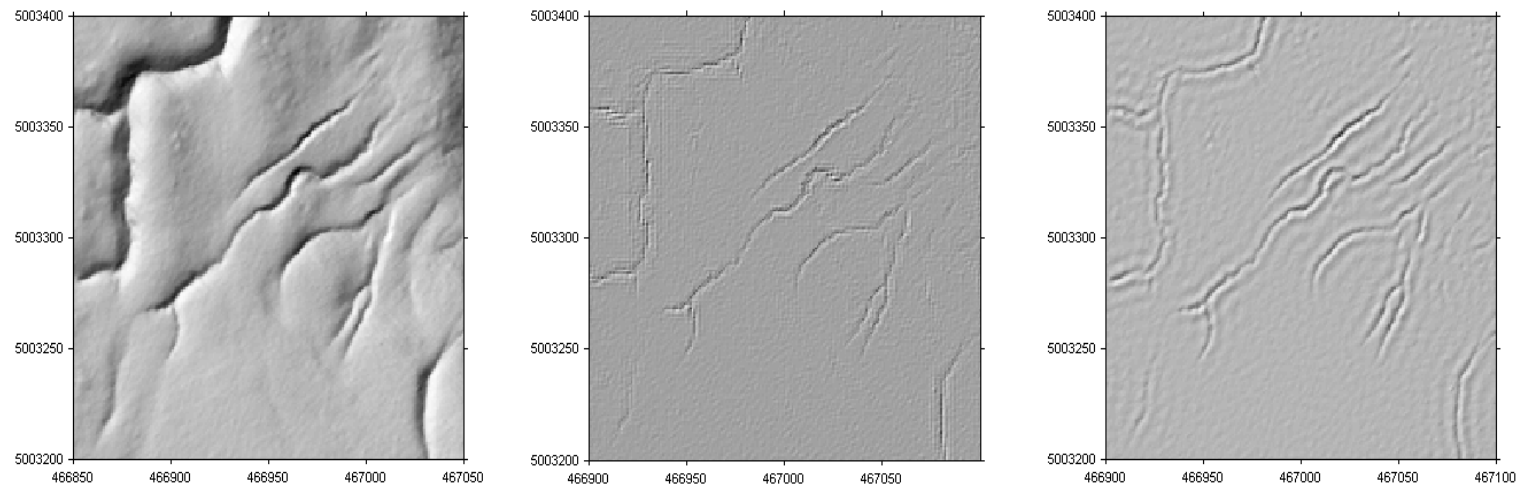


Figure 3.5: Hillshade of portion of DEM, Second Yellowmule Creek. From left to right: unfiltered DEM, wavelet filtered DEM, high pass filtered DEM. Scale: 200 m.

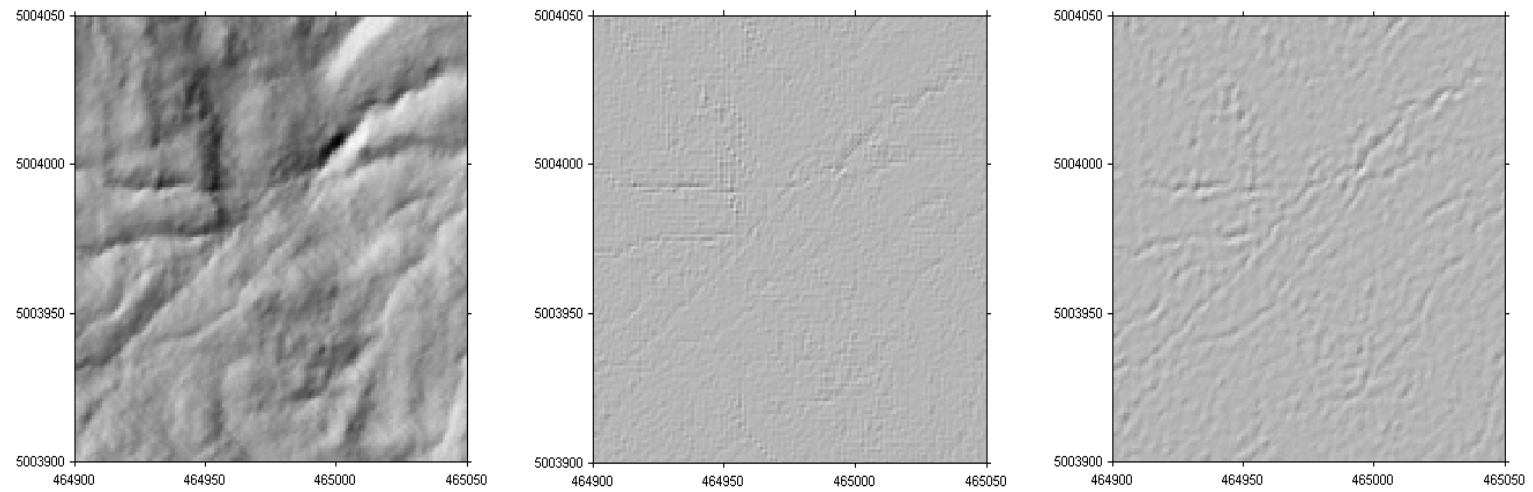


Figure 3.6: Hillshade of portion of DEM, Third Yellowmule Creek. From left to right: unfiltered DEM, wavelet filtered DEM, high pass filtered DEM. Scale: 200 m.

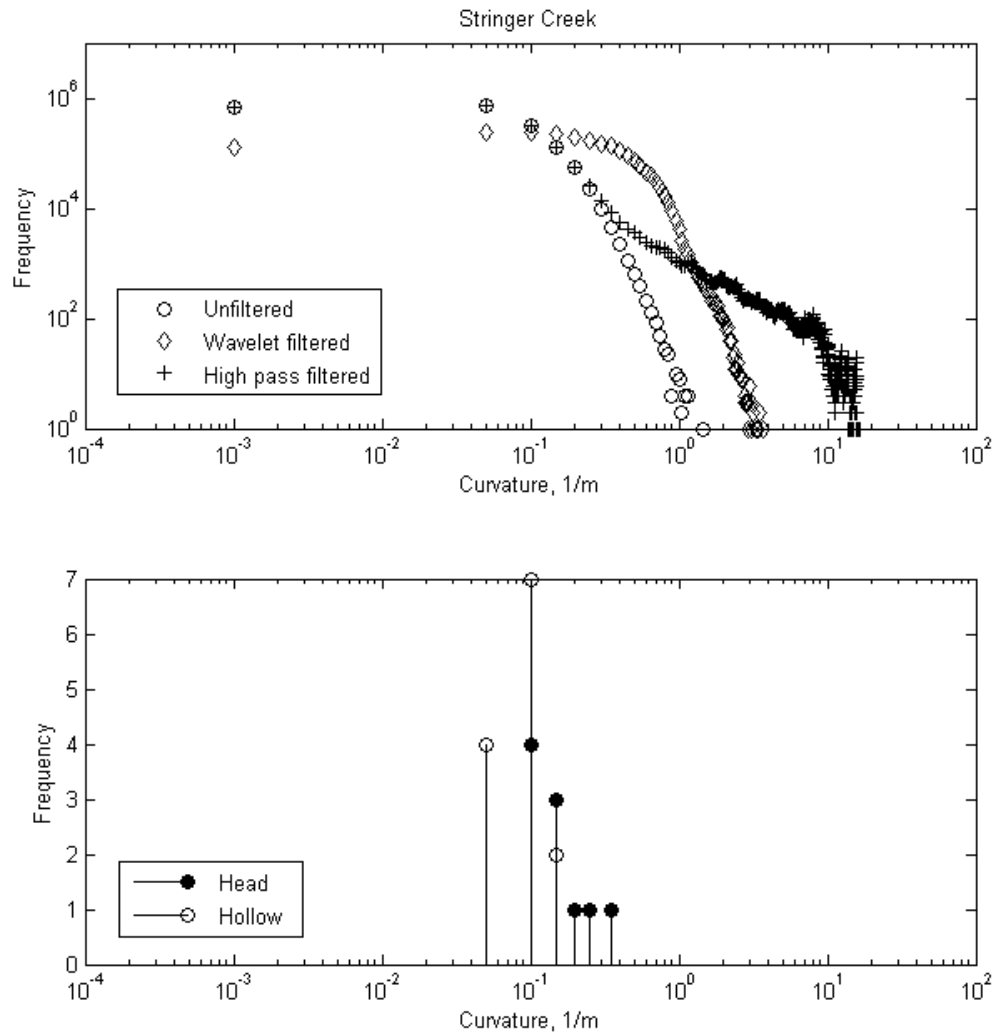


Figure 3.7: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, STR catchment.

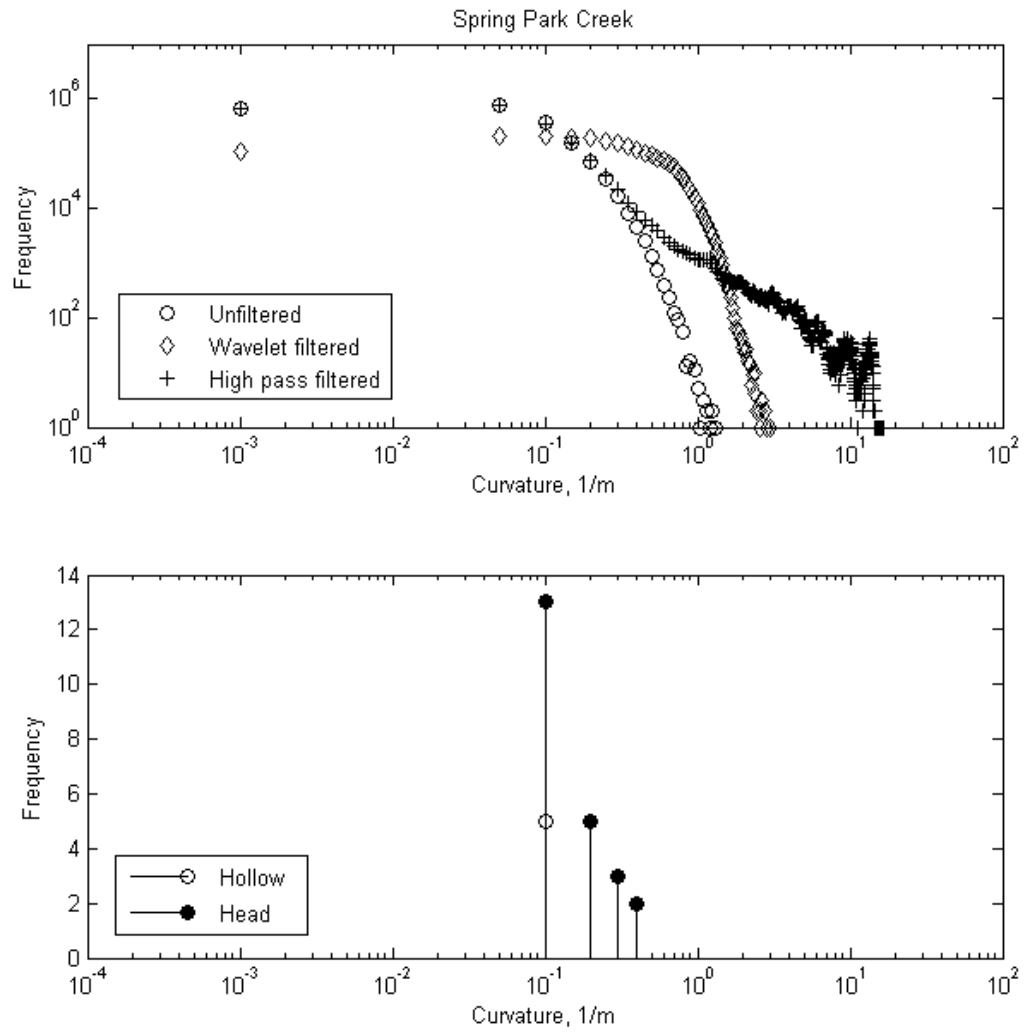


Figure 3.8: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, SPC catchment.

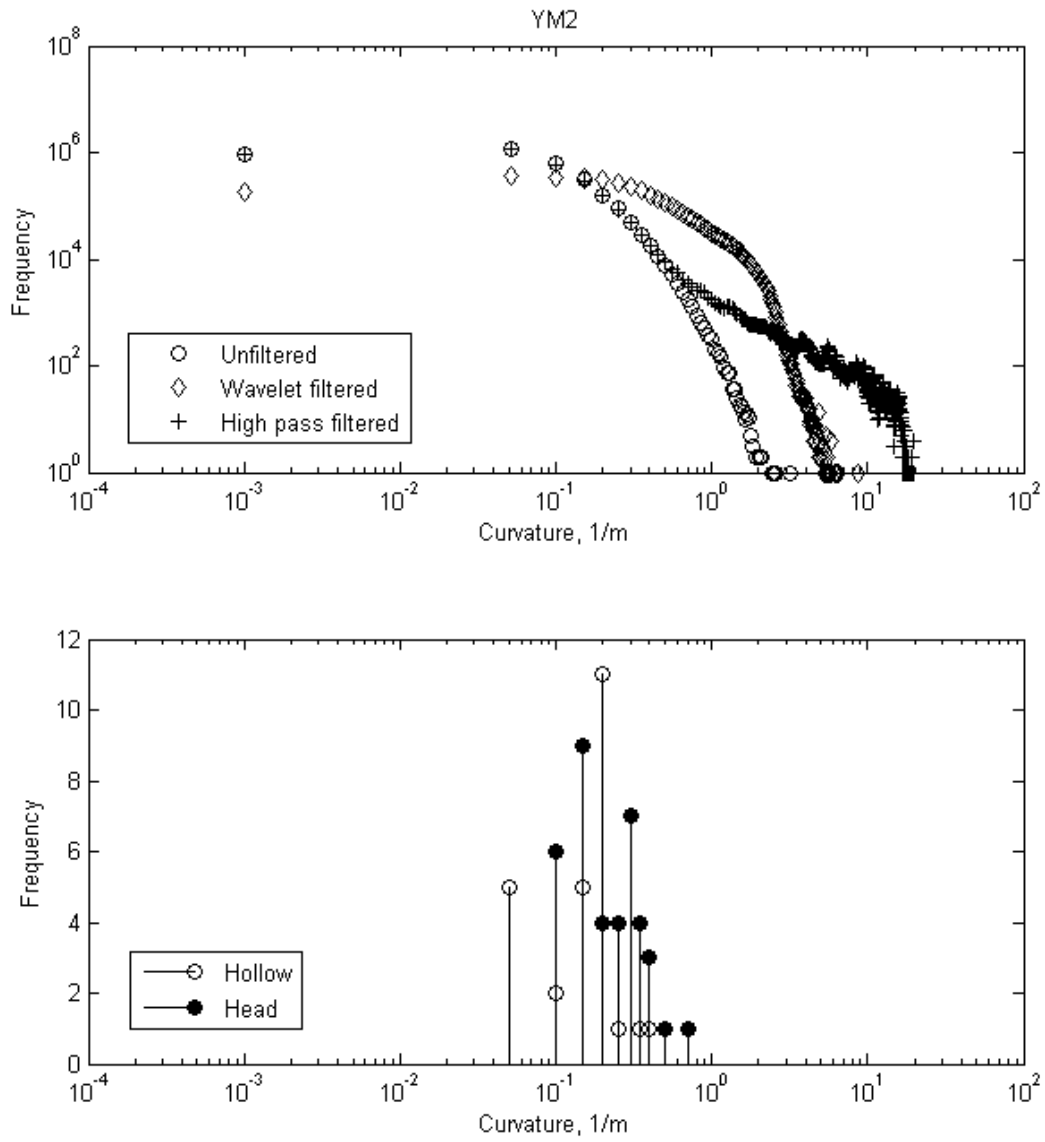


Figure 3.9: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, YM2 catchment.

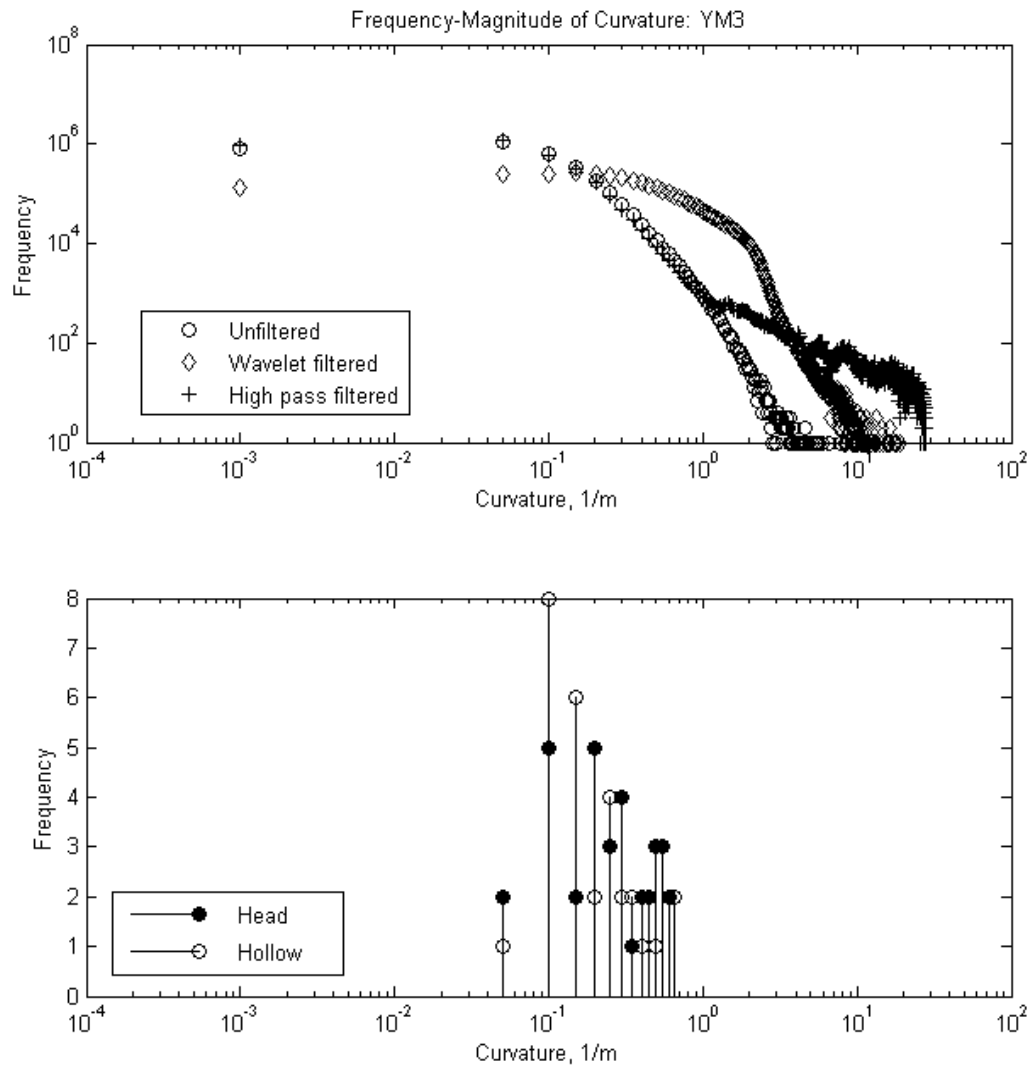


Figure 3.10: Magnitude-frequency distributions of curvature for unfiltered, spectral filtered, and wavelet filtered DEMs, YM3 catchment.

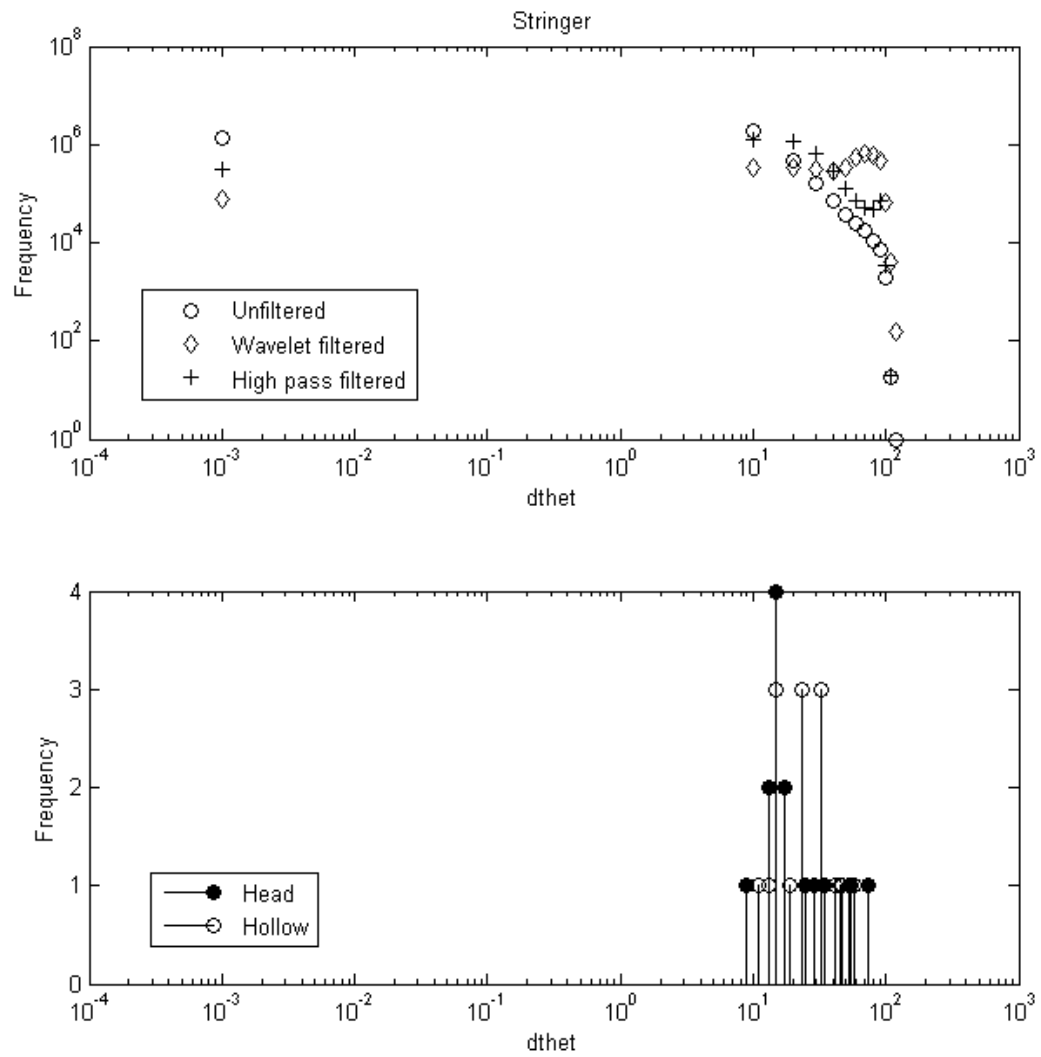


Figure 3.11: Magnitude-frequency distributions of dthet for unfiltered, high pass filtered, and wavelet filtered DEMs, STR catchment.

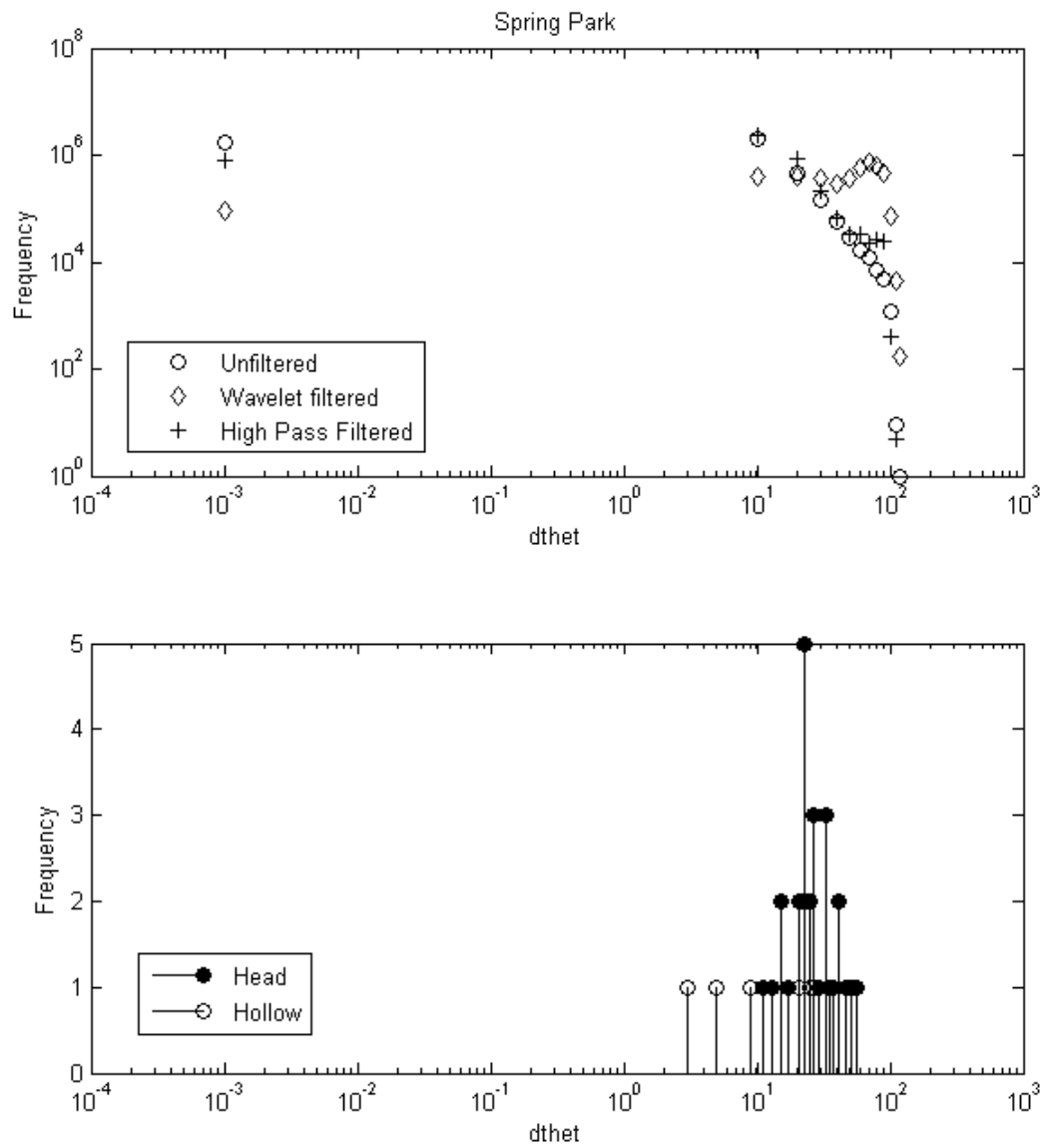


Figure 3.12: Magnitude-frequency distributions of d_{thet} for unfiltered, high pass filtered, and wavelet filtered DEMs, SPC catchment.

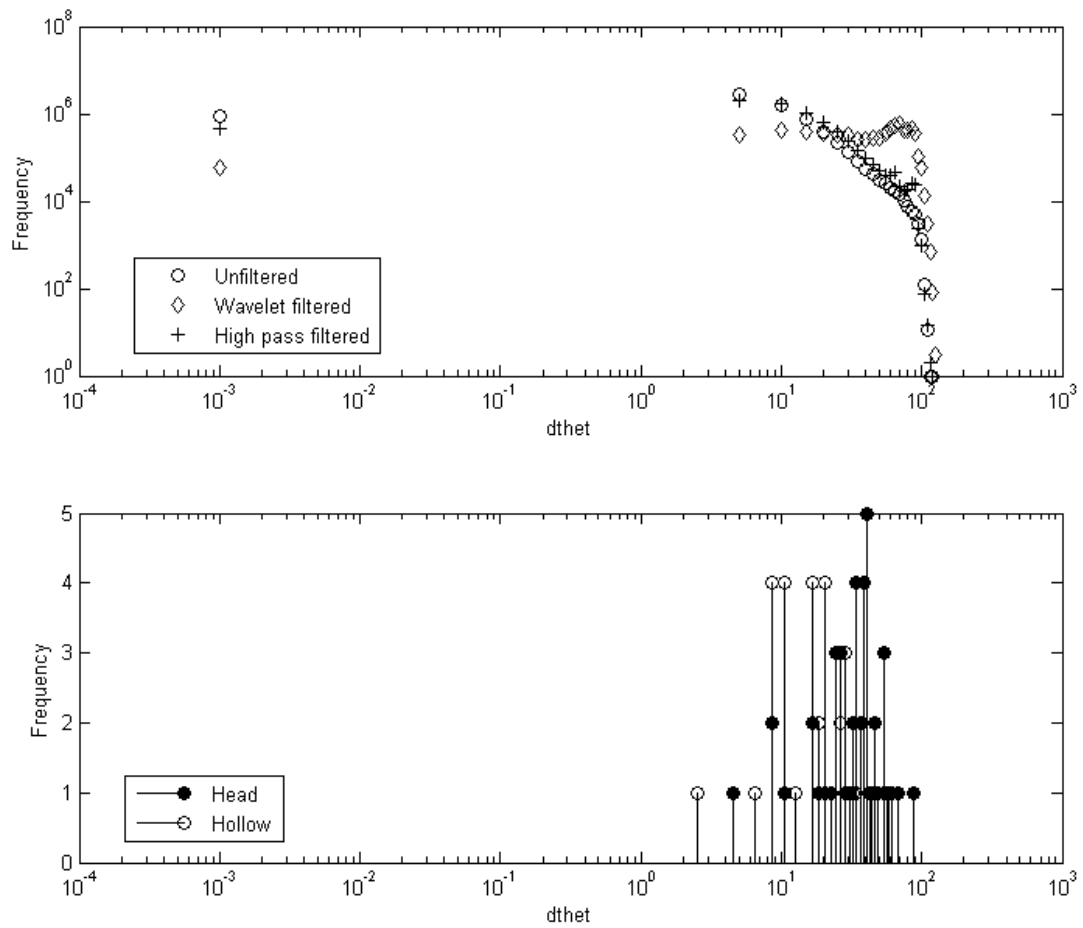


Figure 3.13: Magnitude-frequency distributions of d_{thet} for unfiltered, high pass filtered, and wavelet filtered DEMs, YM2 catchment.

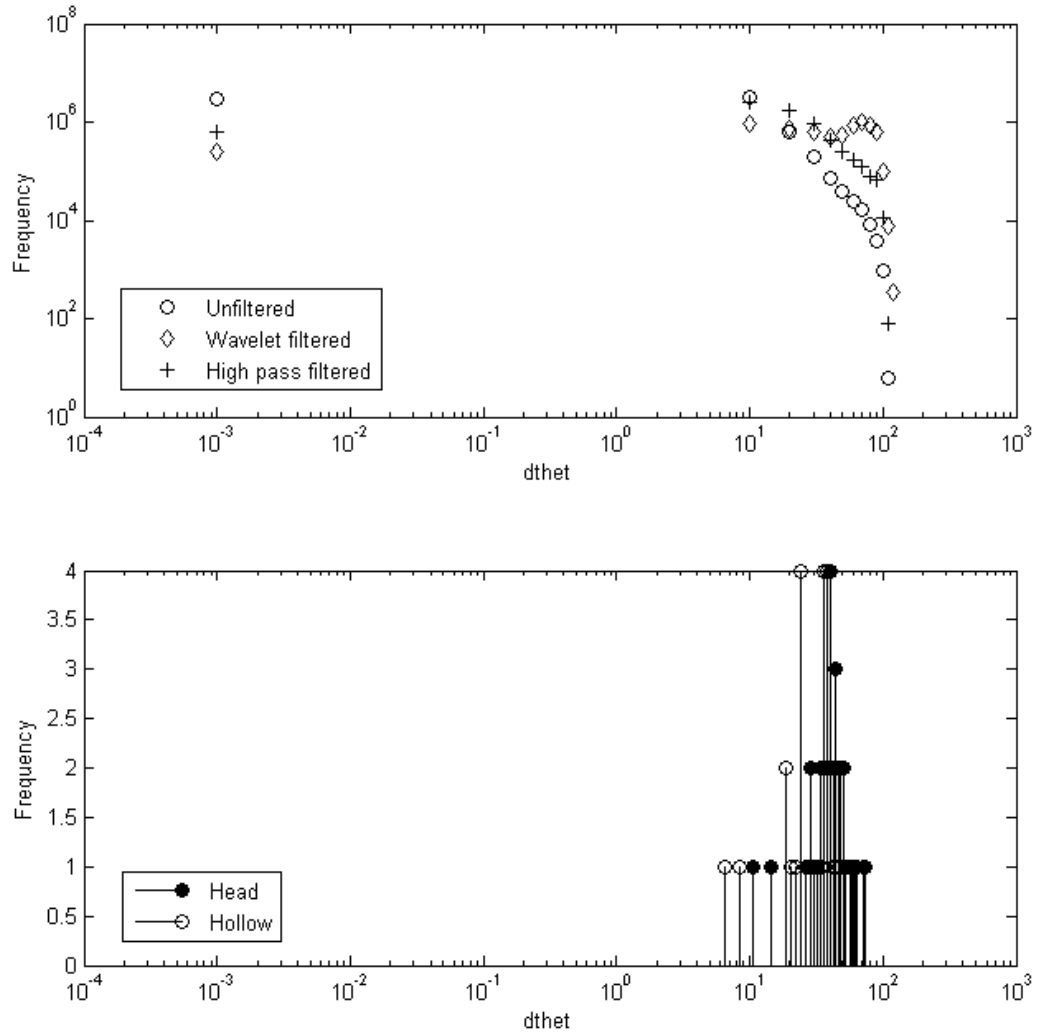


Figure 3.14: Magnitude-frequency distributions of d_{thet} for unfiltered, high pass filtered, and wavelet filtered DEMs, YM3 catchment.

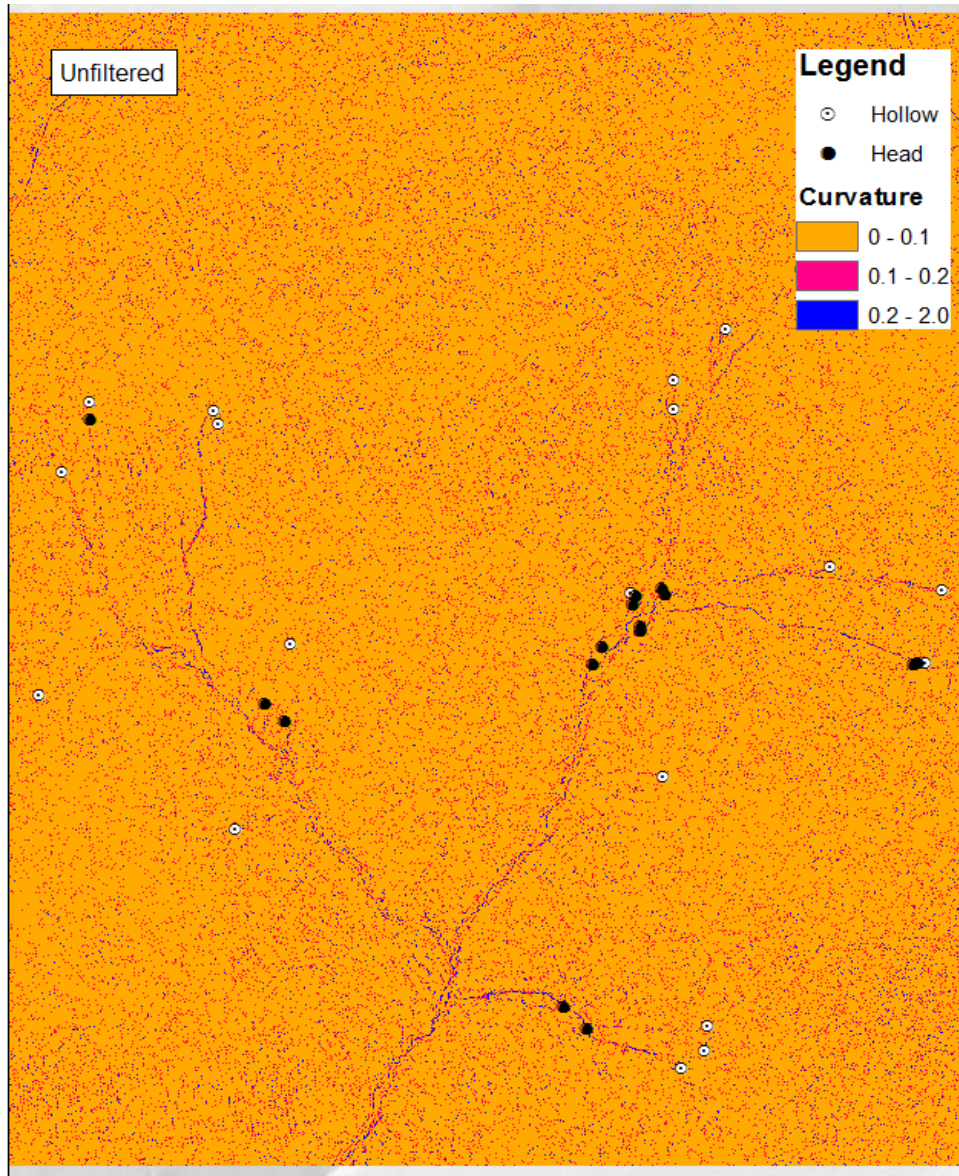


Figure 3.15: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, STR catchment.

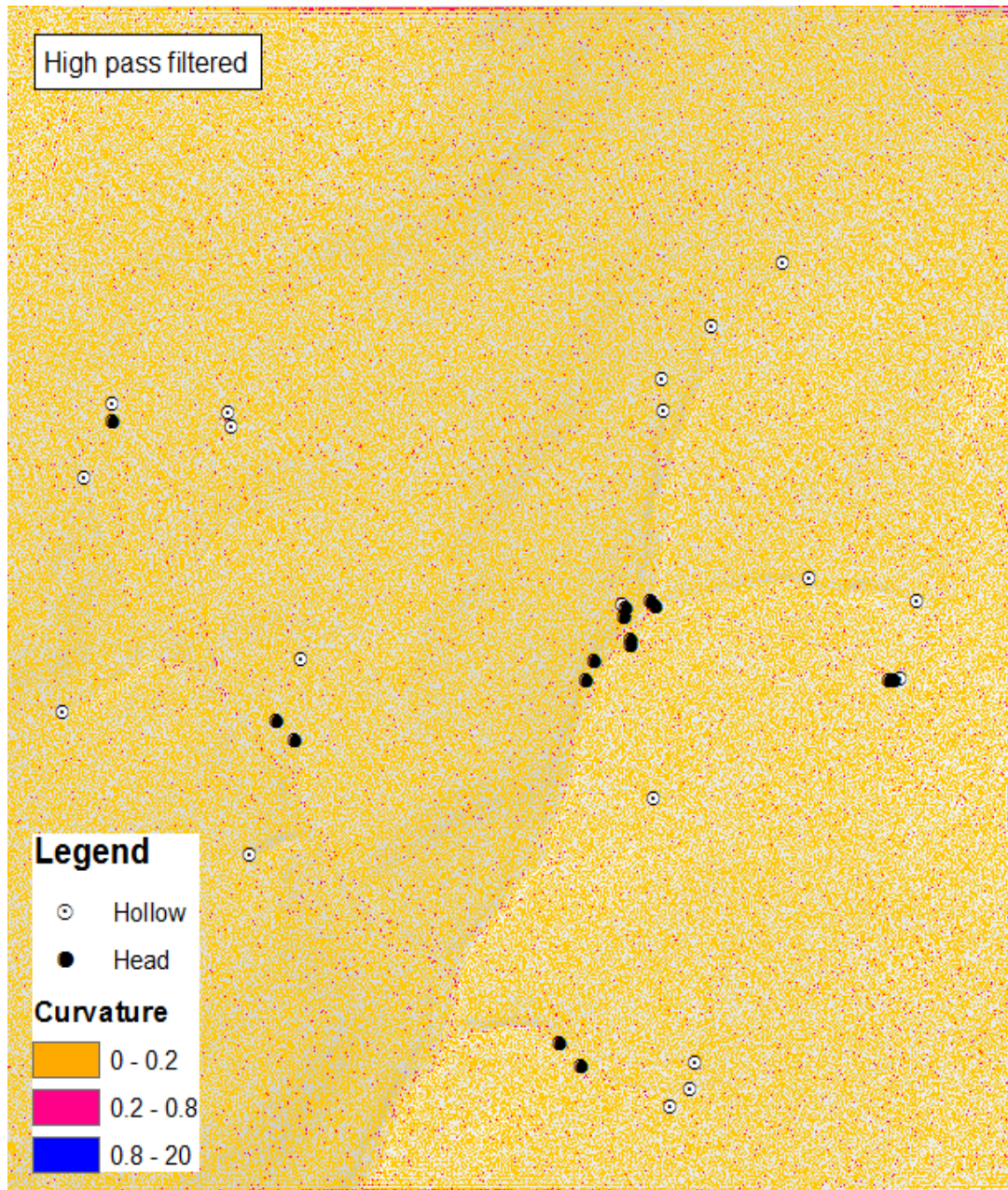


Figure 3.16: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, STR catchment.

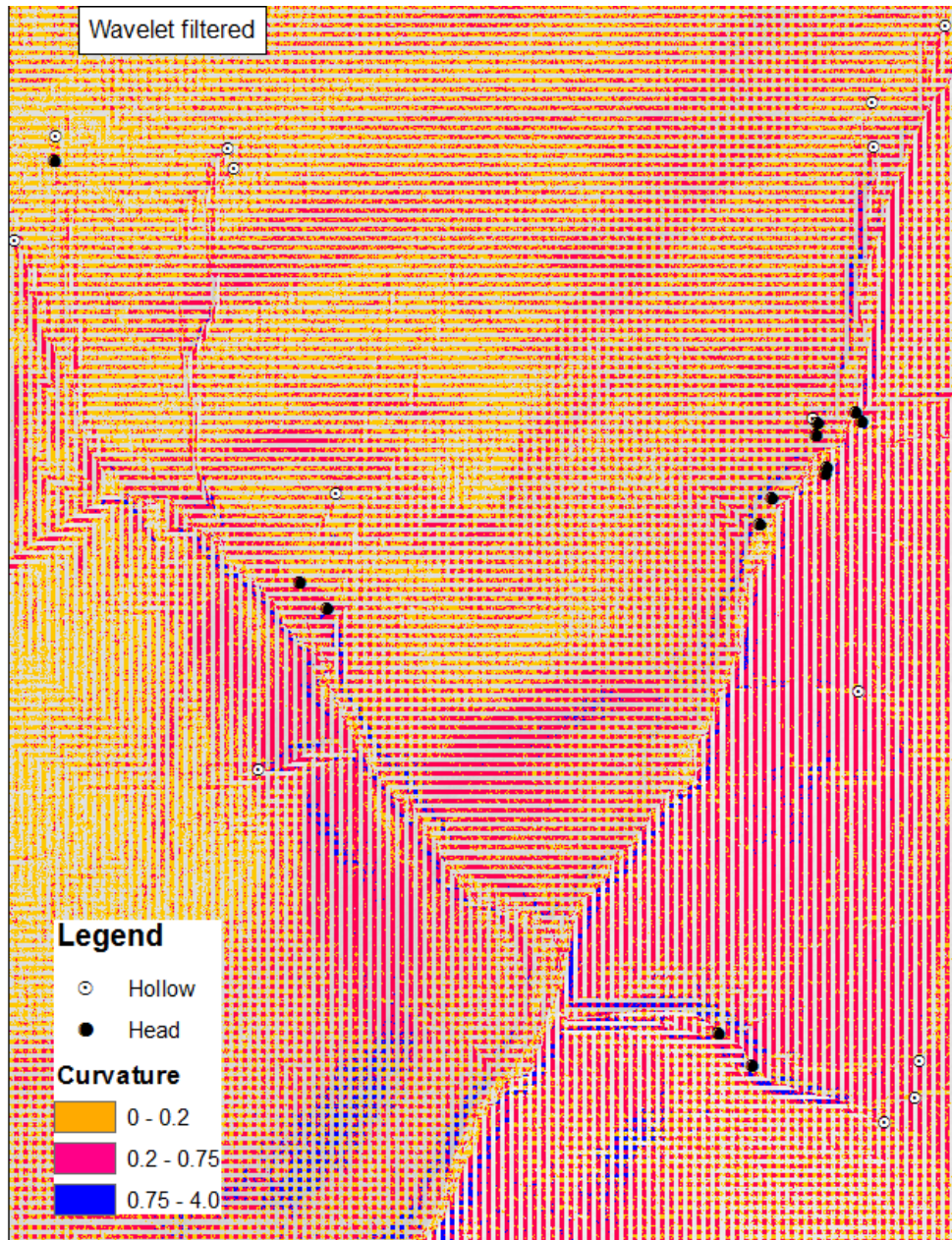


Figure 3.17: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, STR catchment.

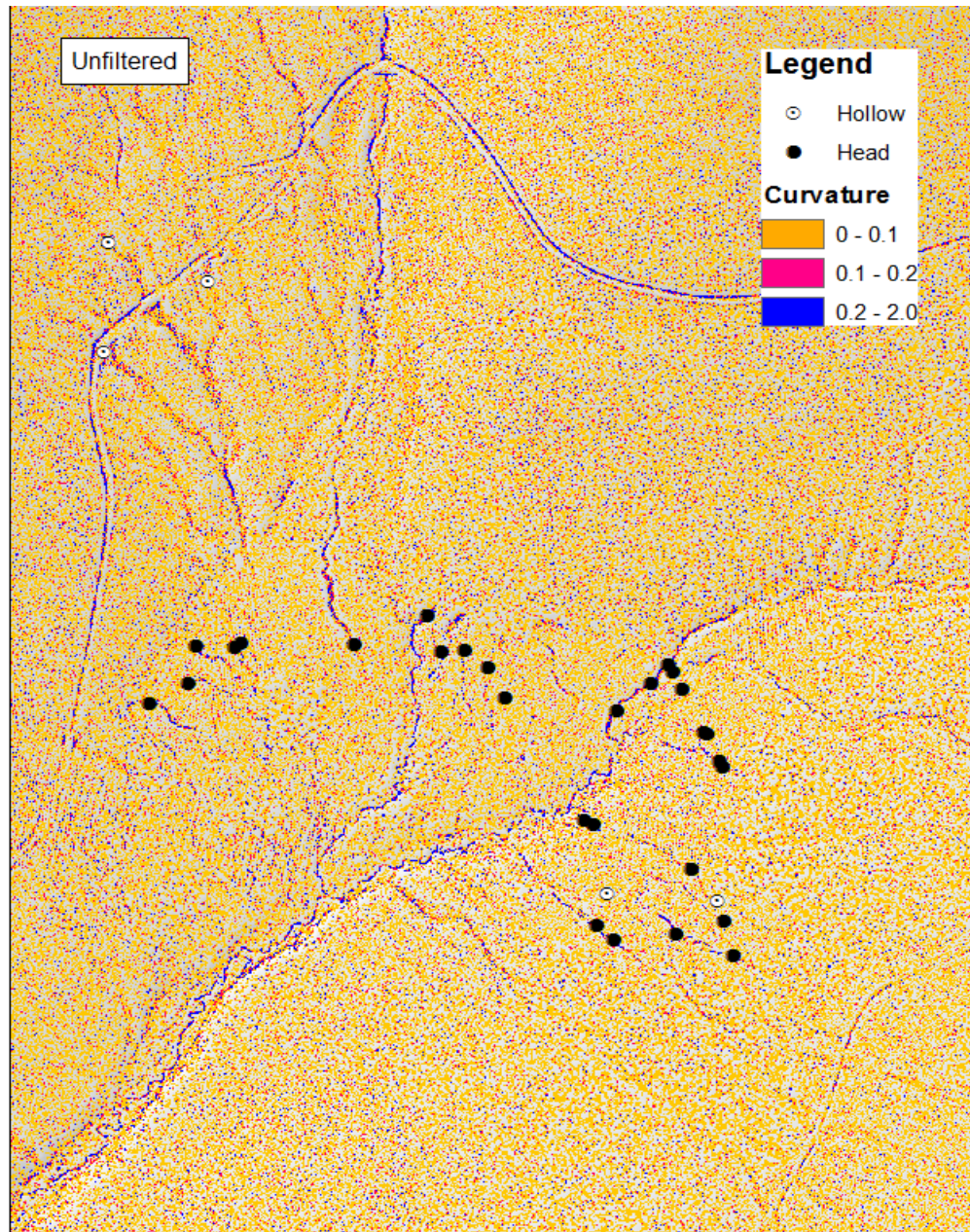


Figure 3.18: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, SPC catchment.

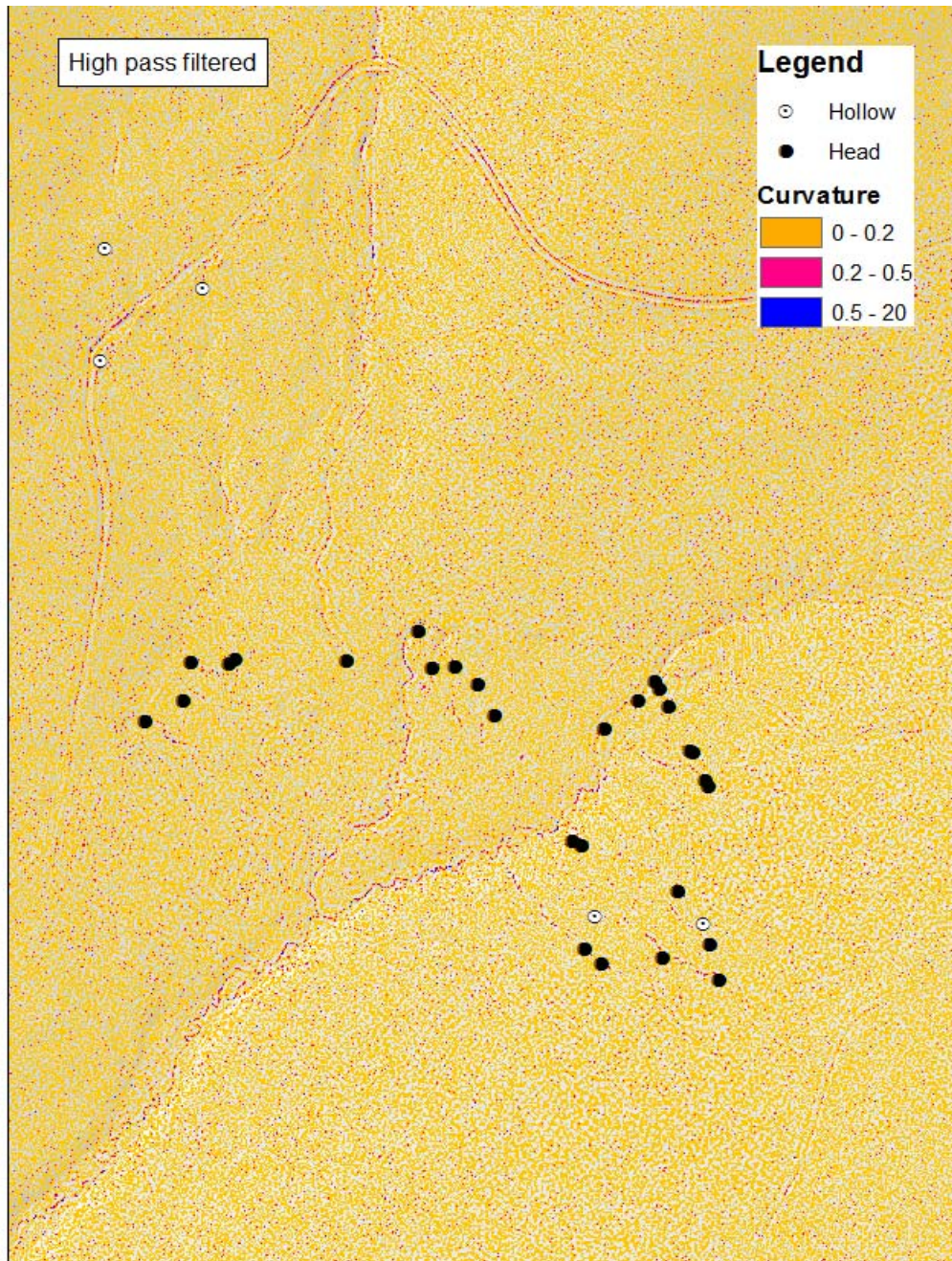


Figure 3.19: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, SPC catchment.

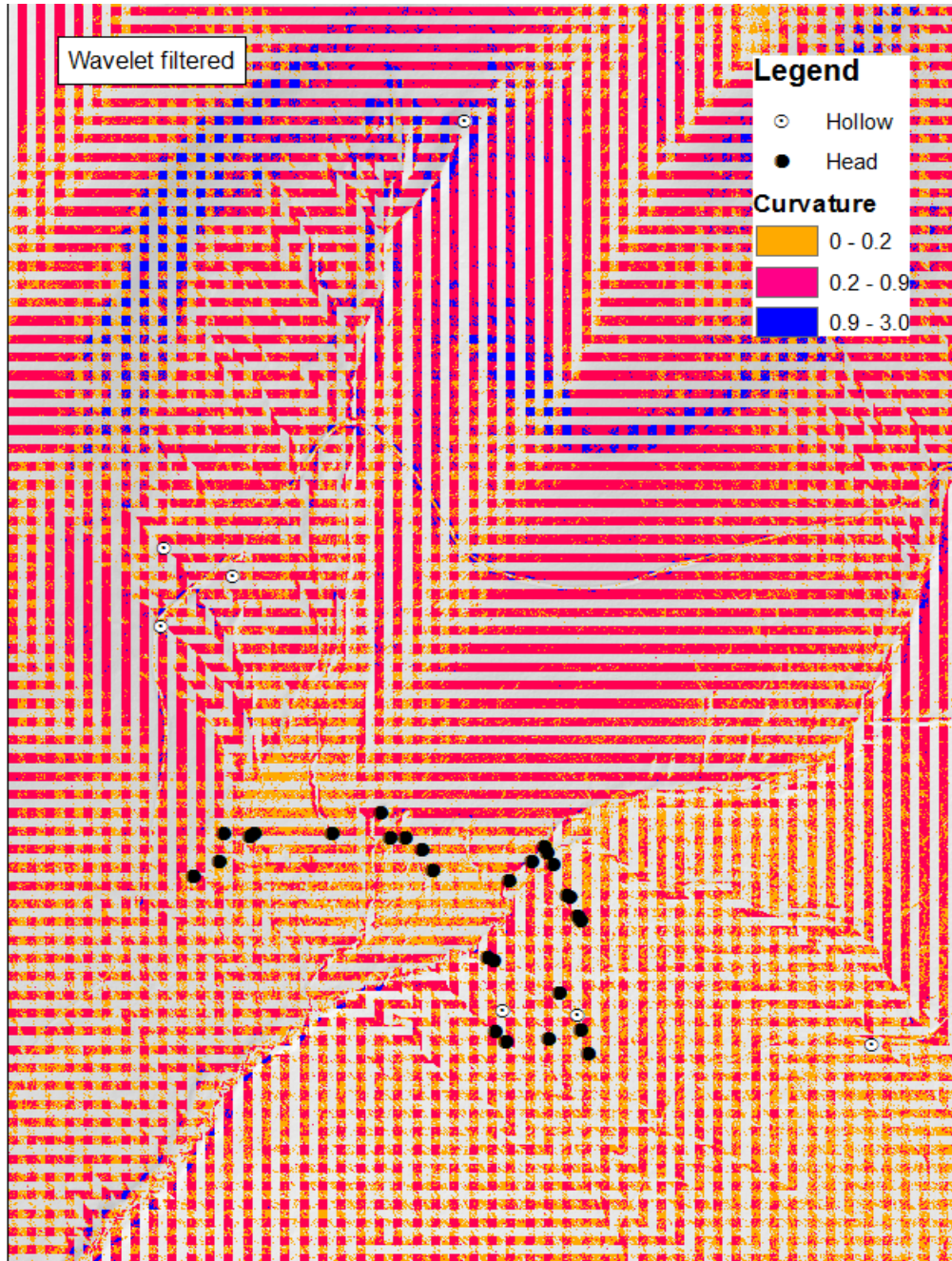


Figure 3.20: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, SPC catchment.

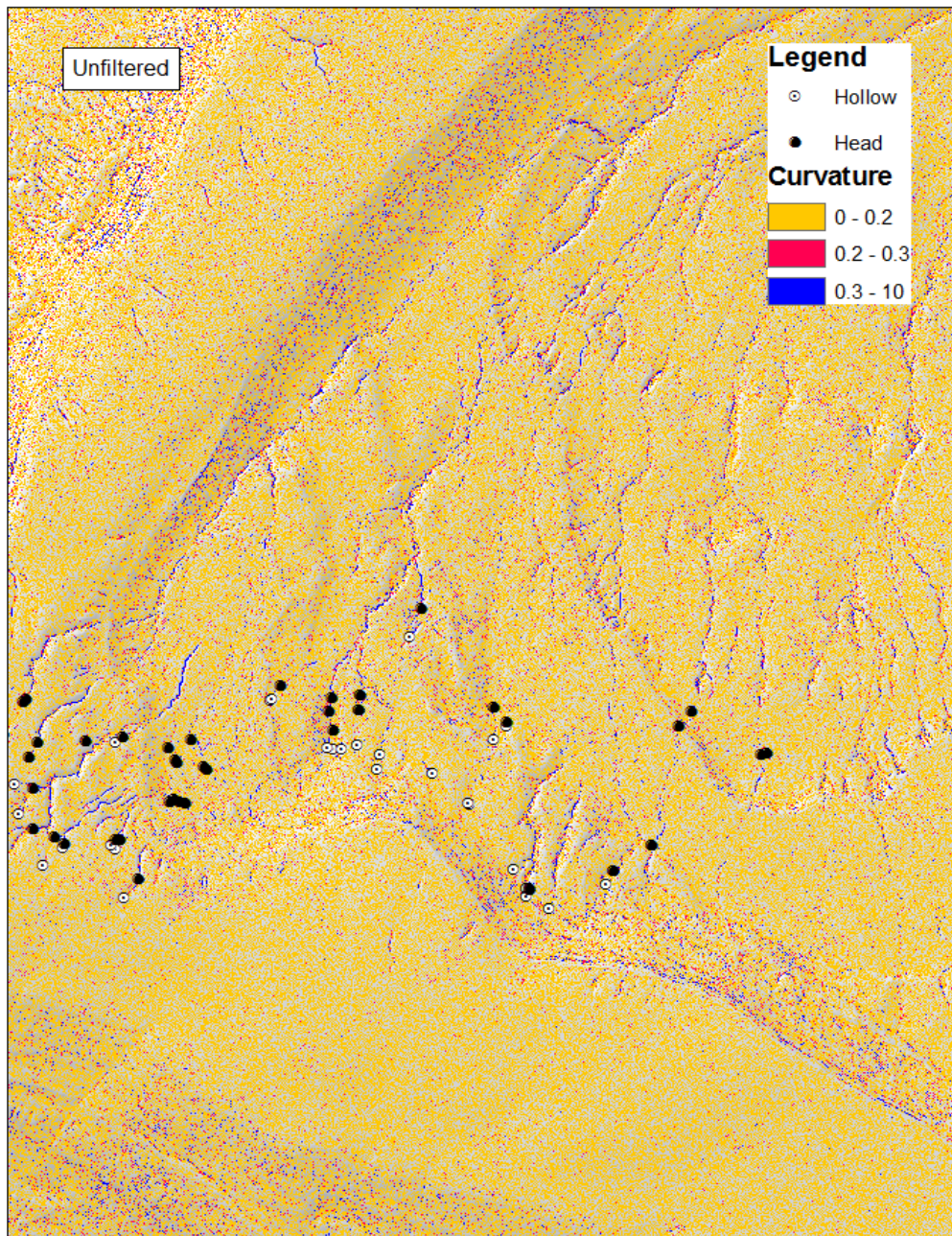


Figure 3.21: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, YM2 catchment.

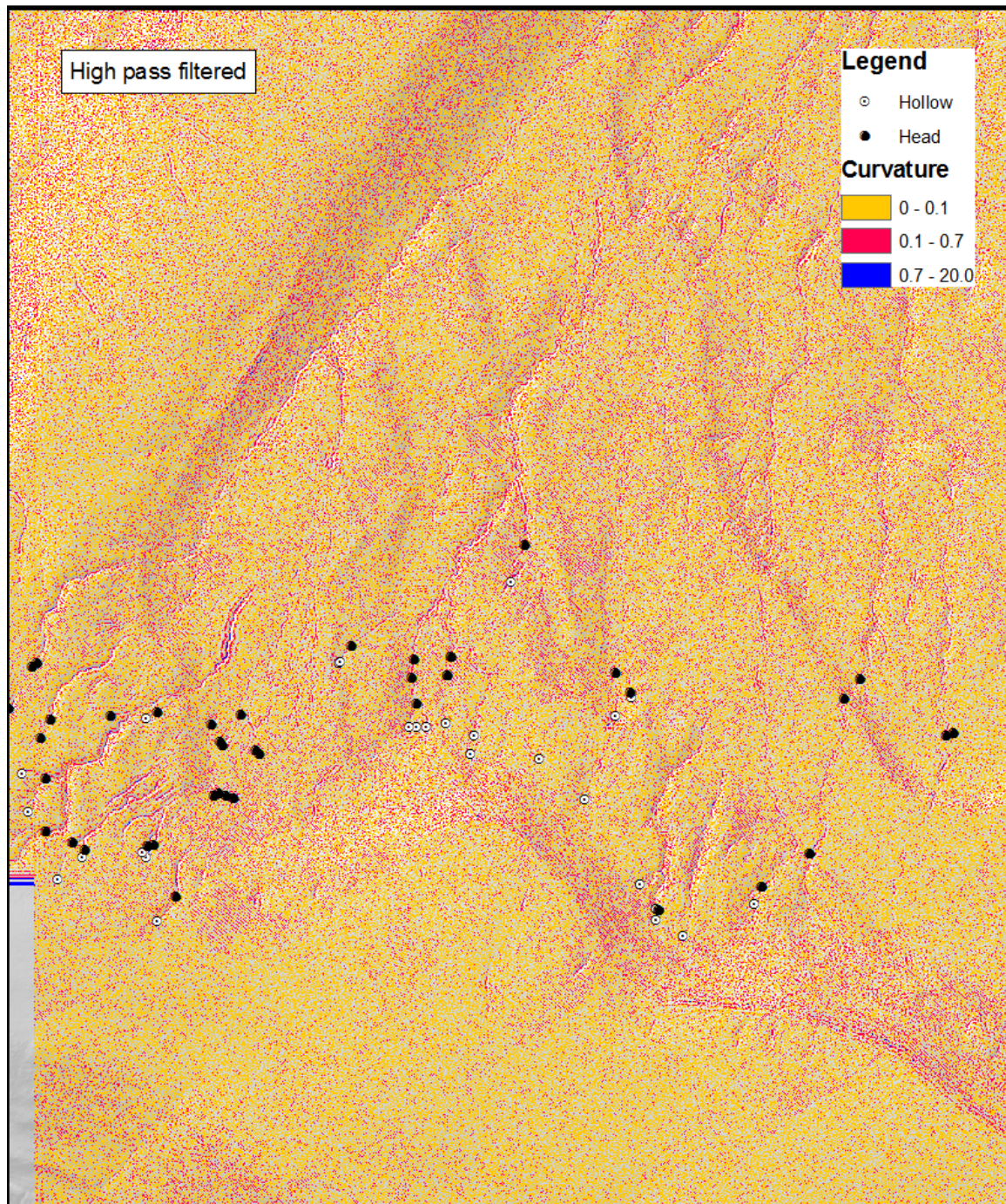


Figure 3.22: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, YM2 catchment.

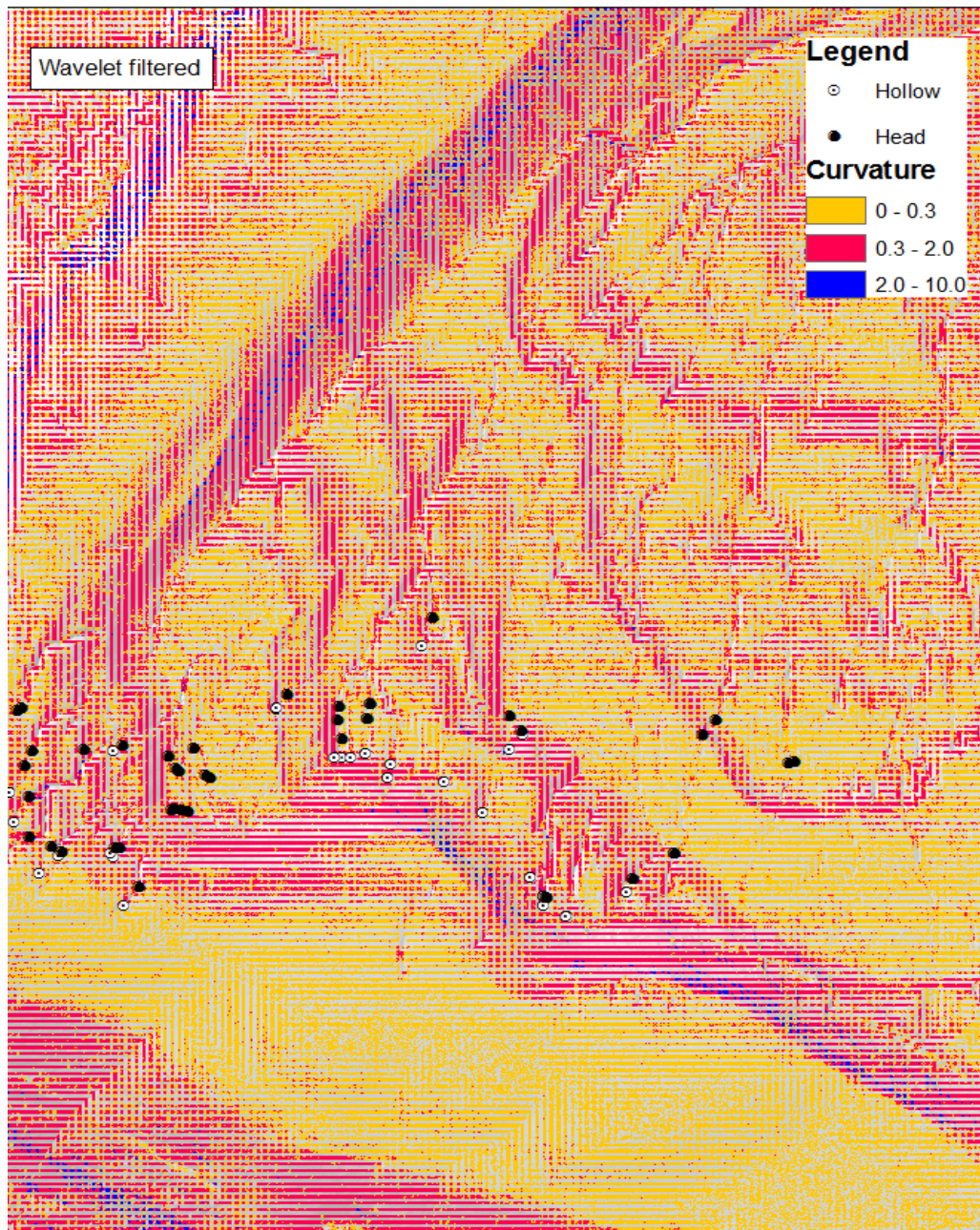


Figure 3.23: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, YM2 catchment.

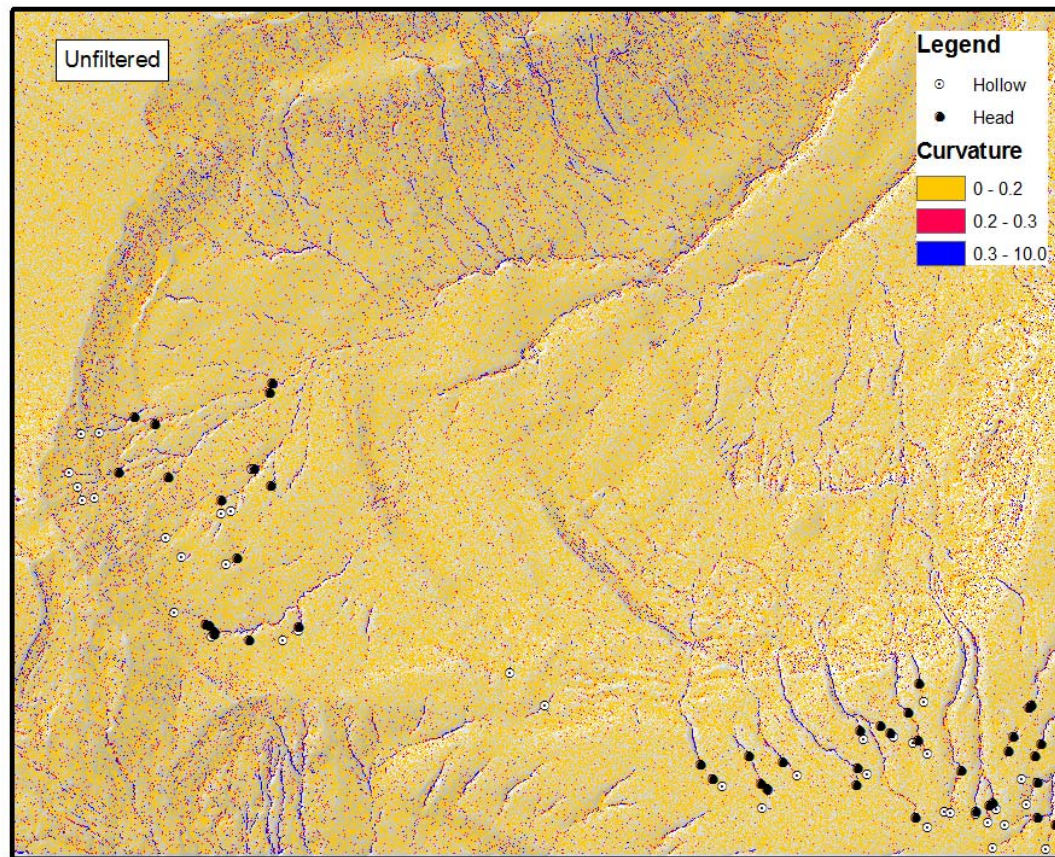


Figure 3.24: Process zones as defined by magnitude-frequency distributions of curvature for unfiltered DEM, YM3 catchment.

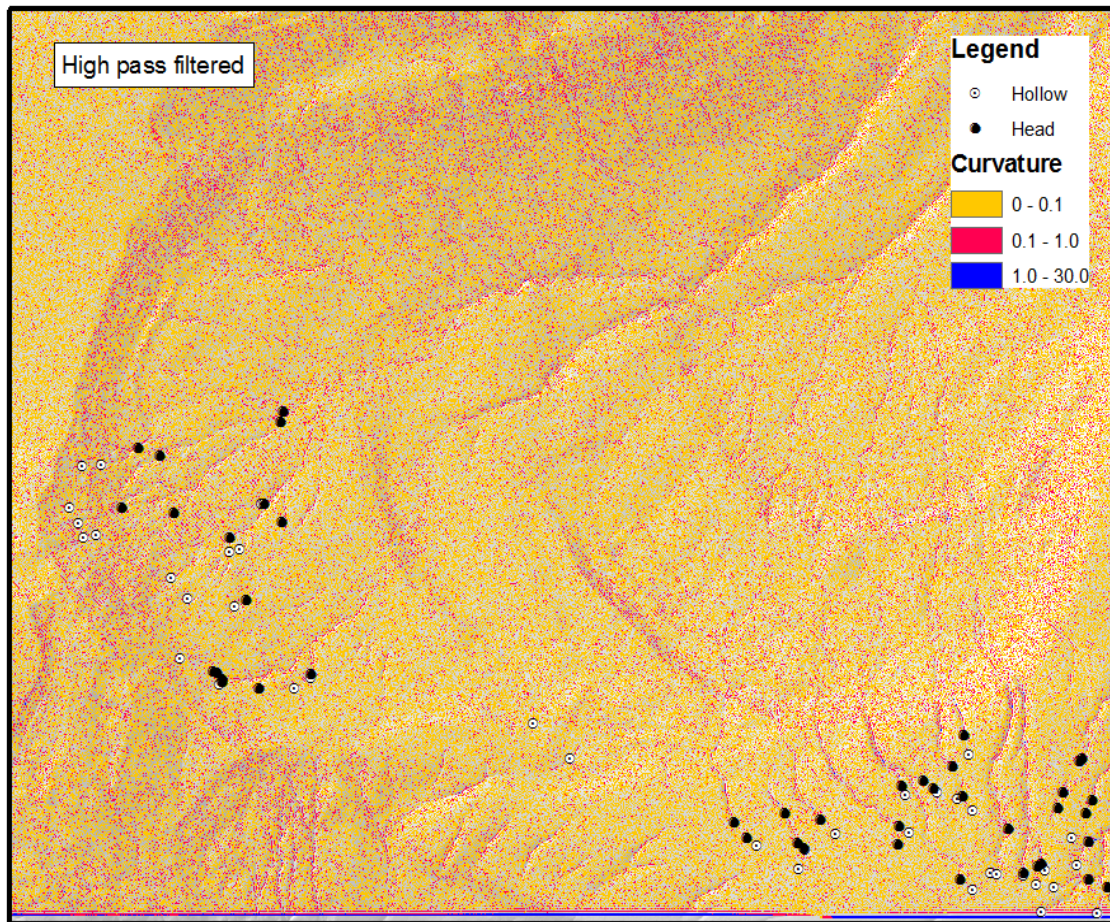


Figure 3.25: Process zones as defined by magnitude-frequency distributions of curvature for high pass filtered DEM, YM3 catchment.

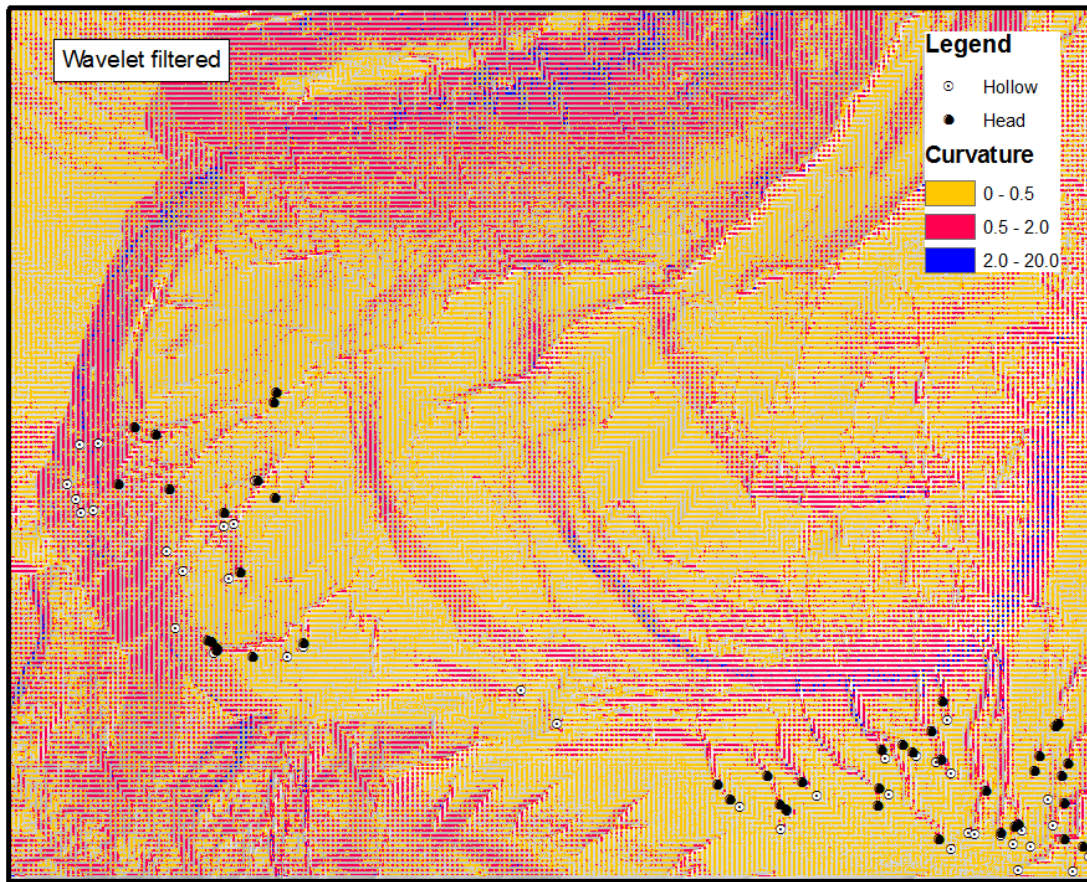


Figure 3.26: Process zones as defined by magnitude-frequency distributions of curvature for wavelet filtered DEM, YM3 catchment.

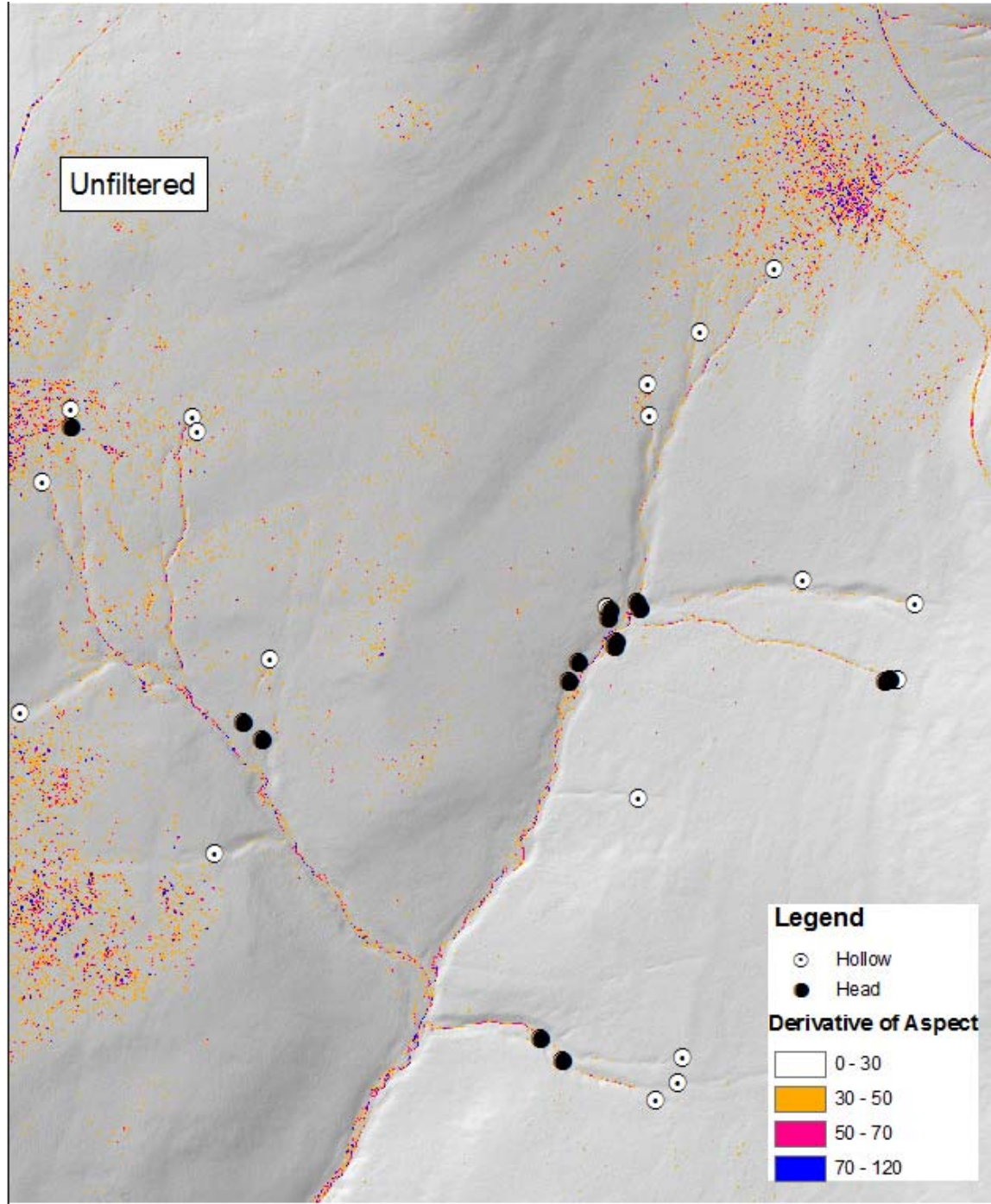


Figure 3.27: Process zones as defined by magnitude-frequency distributions of $d\theta$ for unfiltered DEM, STR catchment.

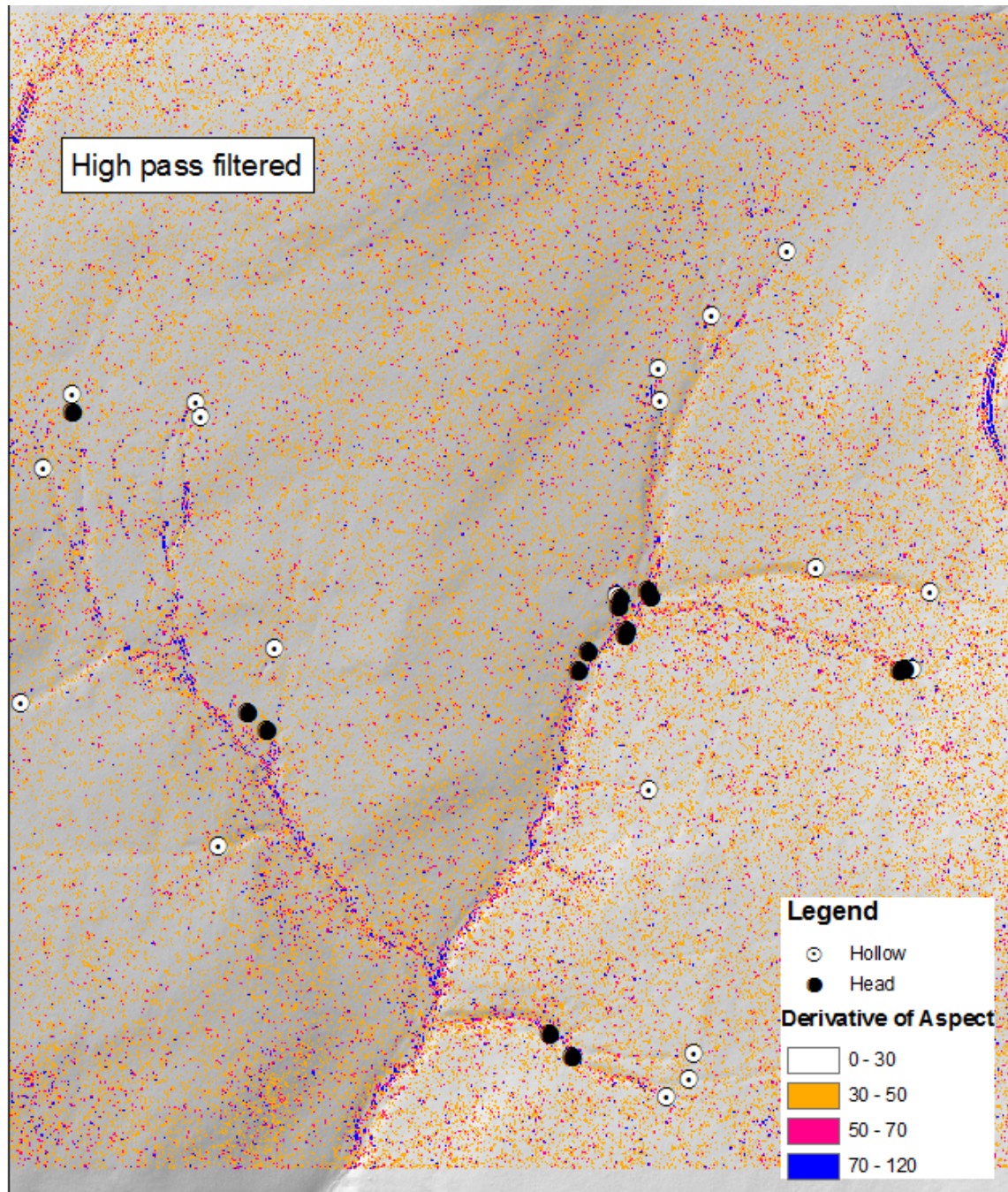


Figure 3.28: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, STR catchment.

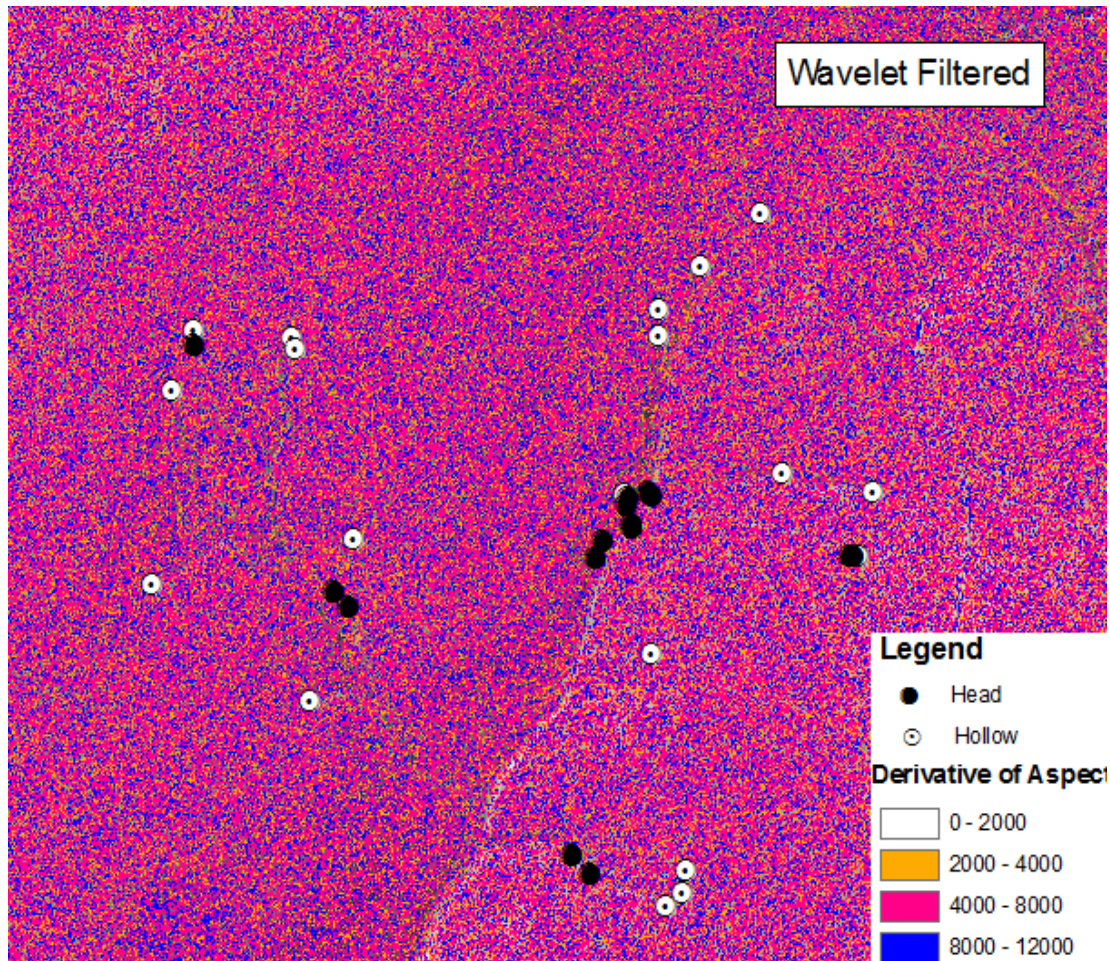


Figure 3.29: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, STR catchment.

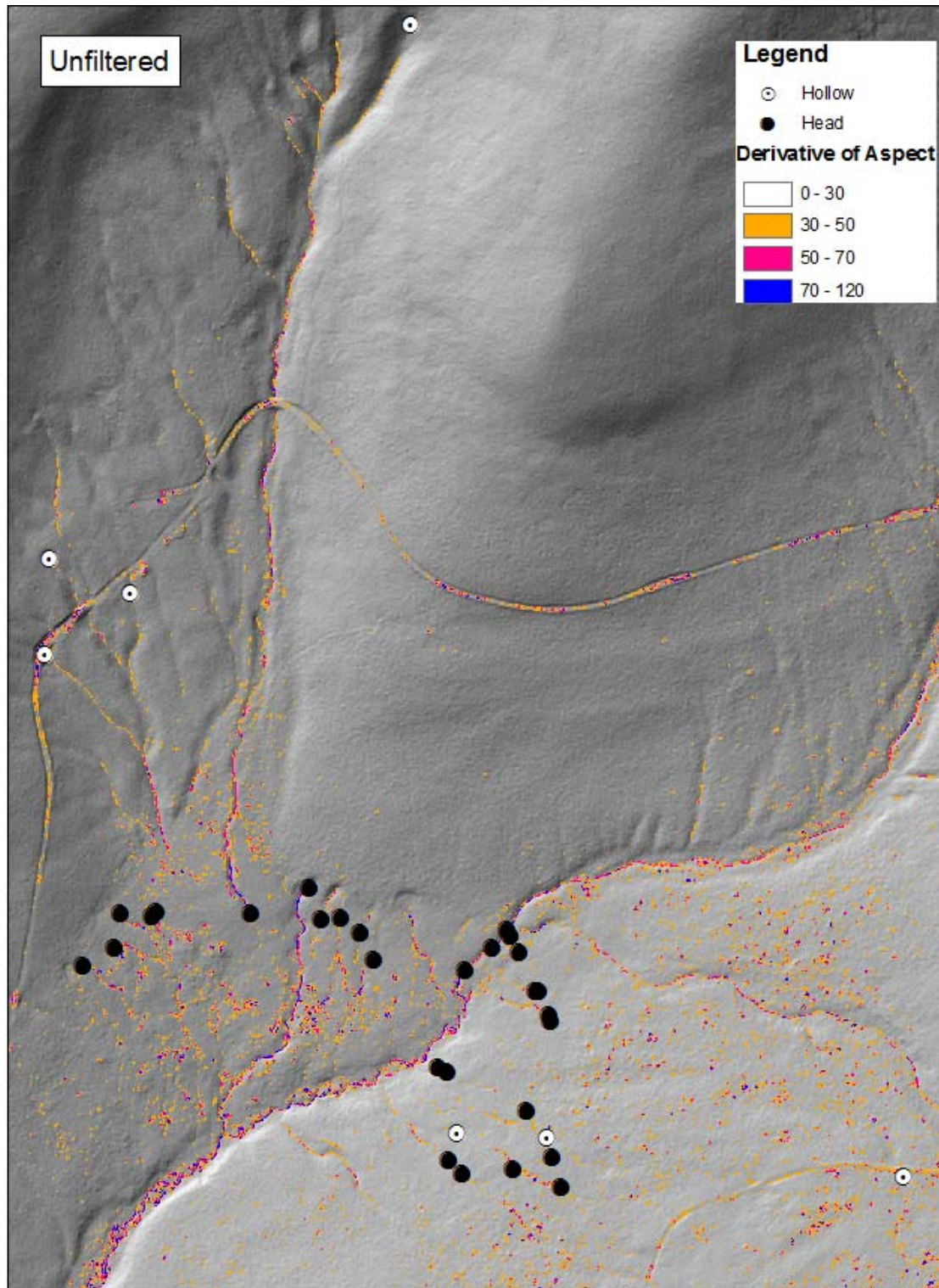


Figure 3.30: Process zones as defined by magnitude-frequency distributions of $d\theta$ for unfiltered DEM, SPC catchment.

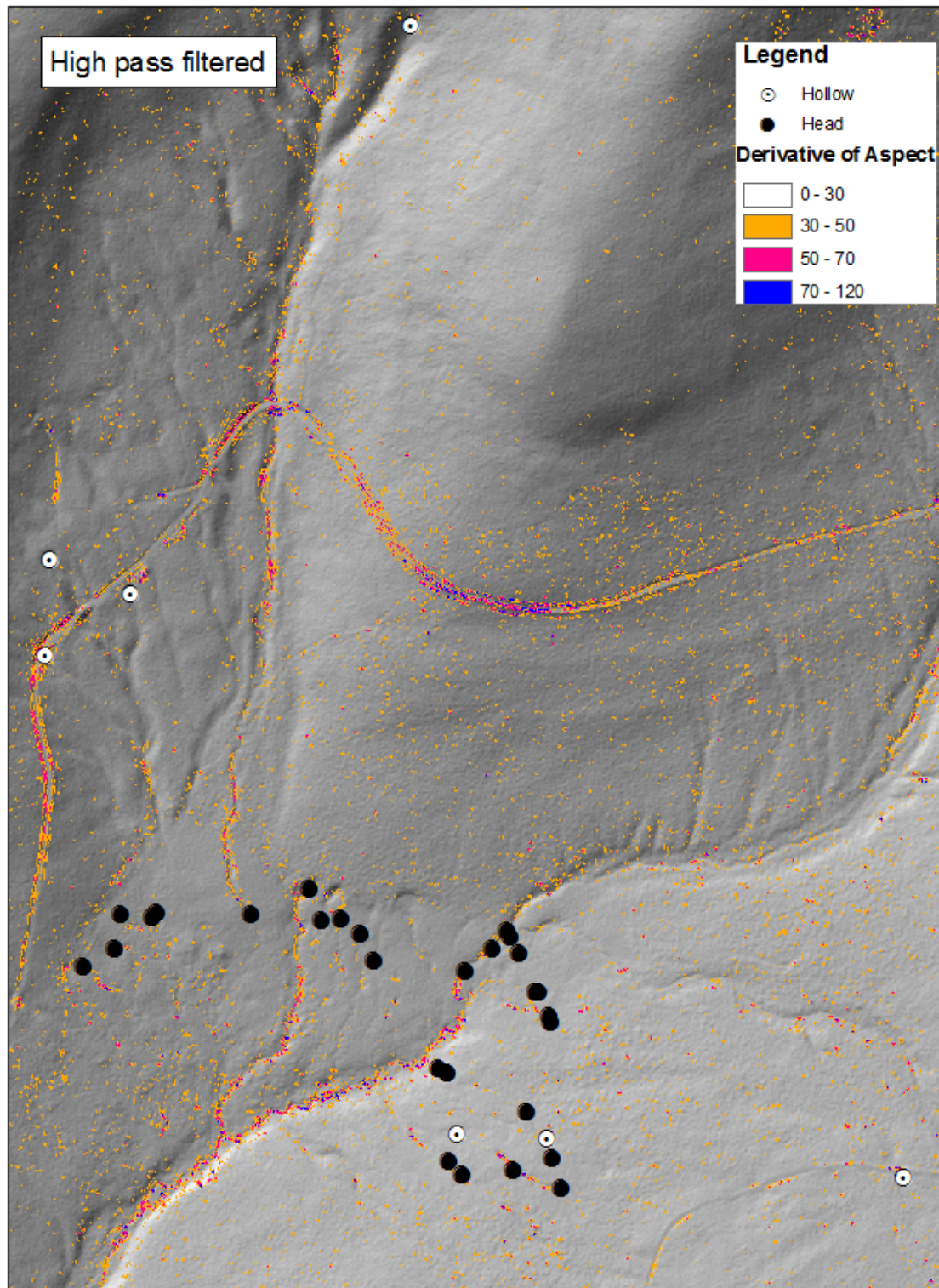


Figure 3.31: Process zones as defined by magnitude-frequency distributions of $d\theta$ for high pass filtered DEM, SPC catchment.

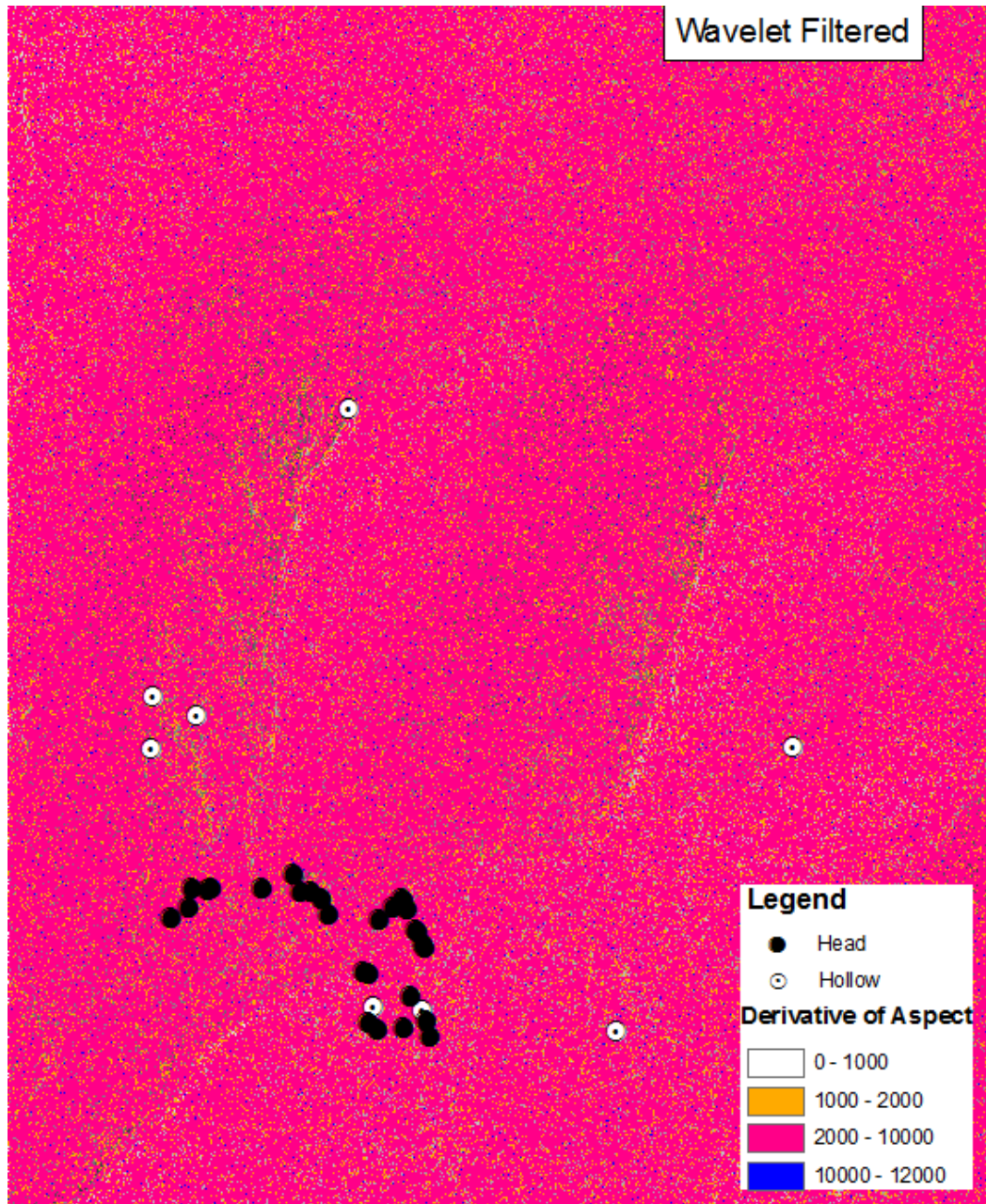


Figure 3.32: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, SPC catchment.

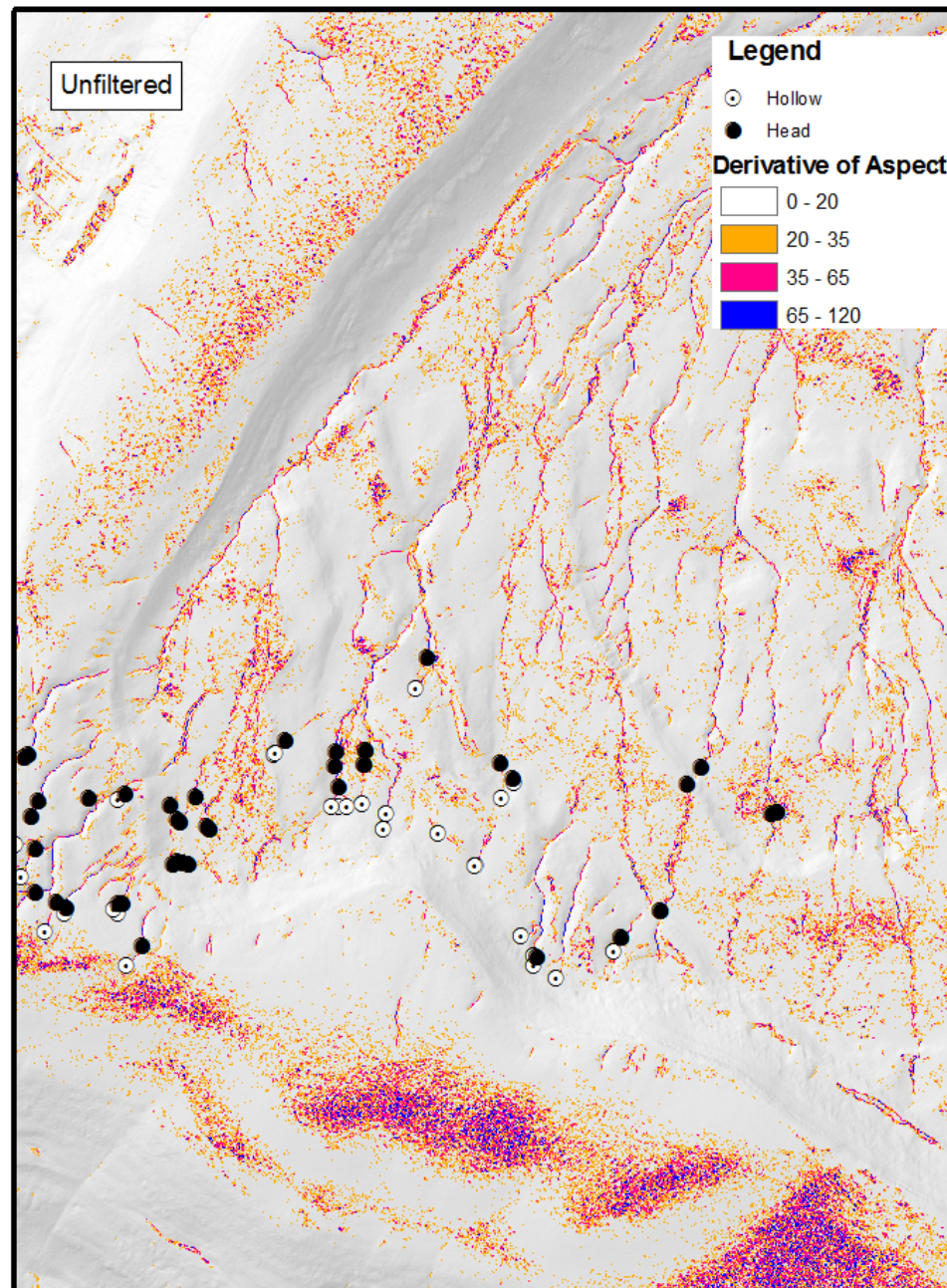


Figure 3.33: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, YM2 catchment.

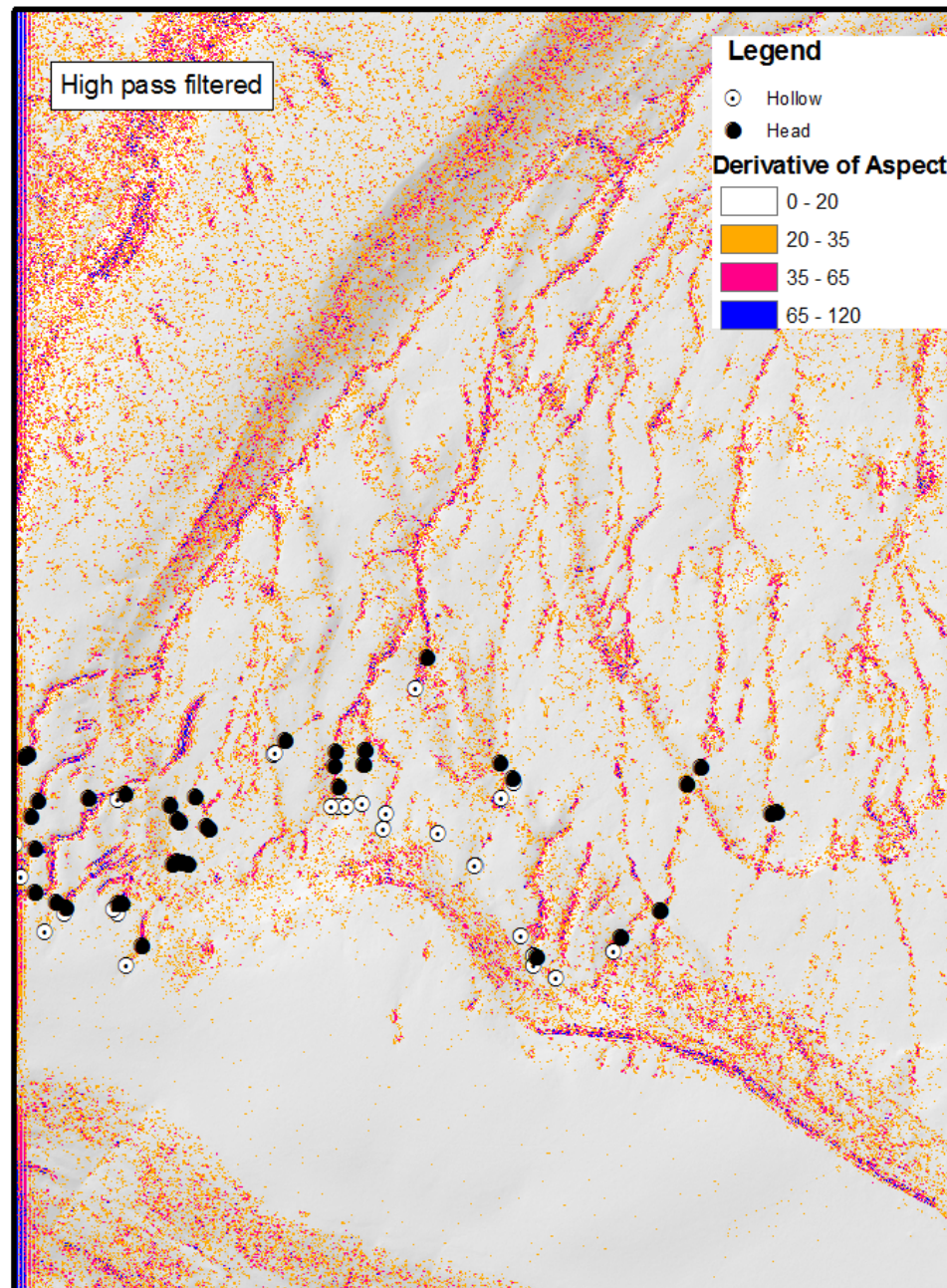


Figure 3.34: Process zones as defined by magnitude-frequency distributions of d_{thet} for high pass filtered DEM, YM2 catchment.

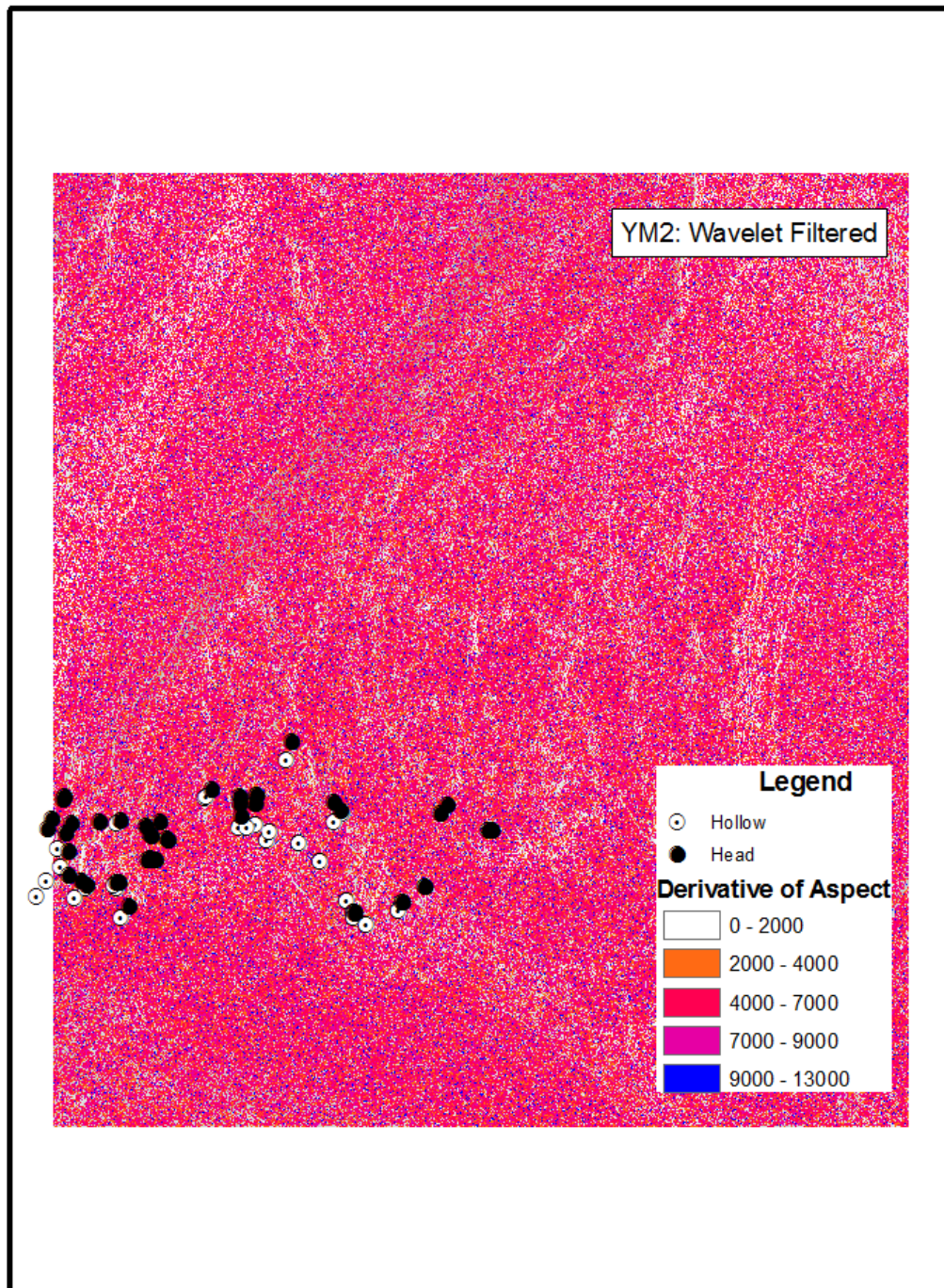


Figure 3.35: Process zones as defined by magnitude-frequency distributions of $d\theta$ for wavelet filtered DEM, YM2 catchment.

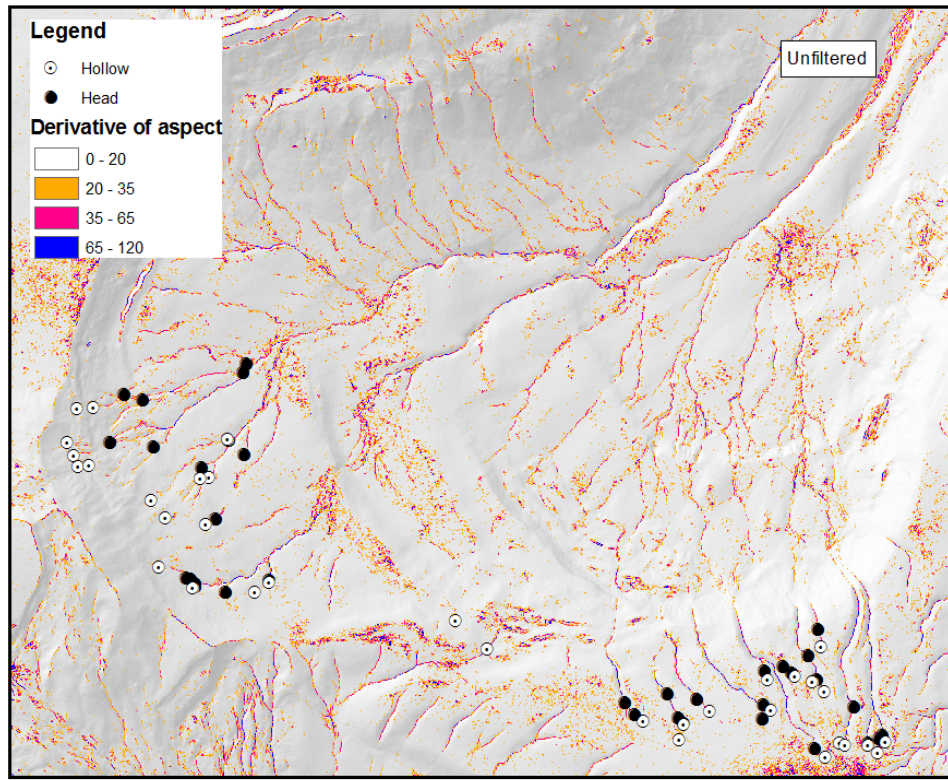


Figure 3.36: Process zones as defined by magnitude-frequency distributions of dthet for unfiltered DEM, YM3 catchment.

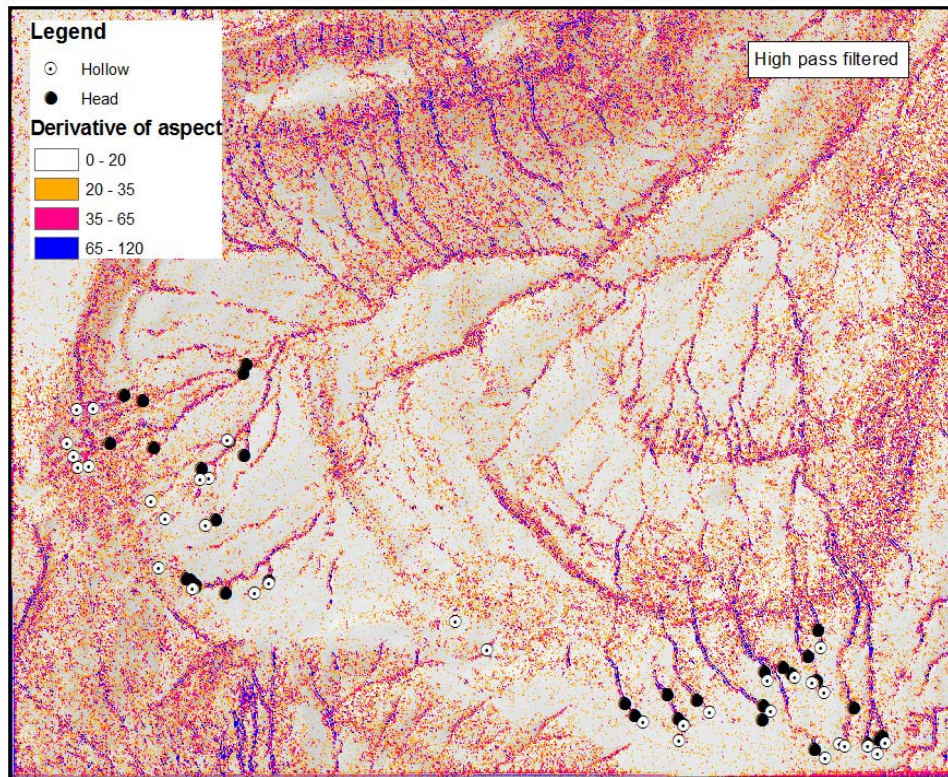


Figure 3.37: Process zones as defined by magnitude-frequency distributions of dthet for high pass filtered DEM, YM3 catchment.

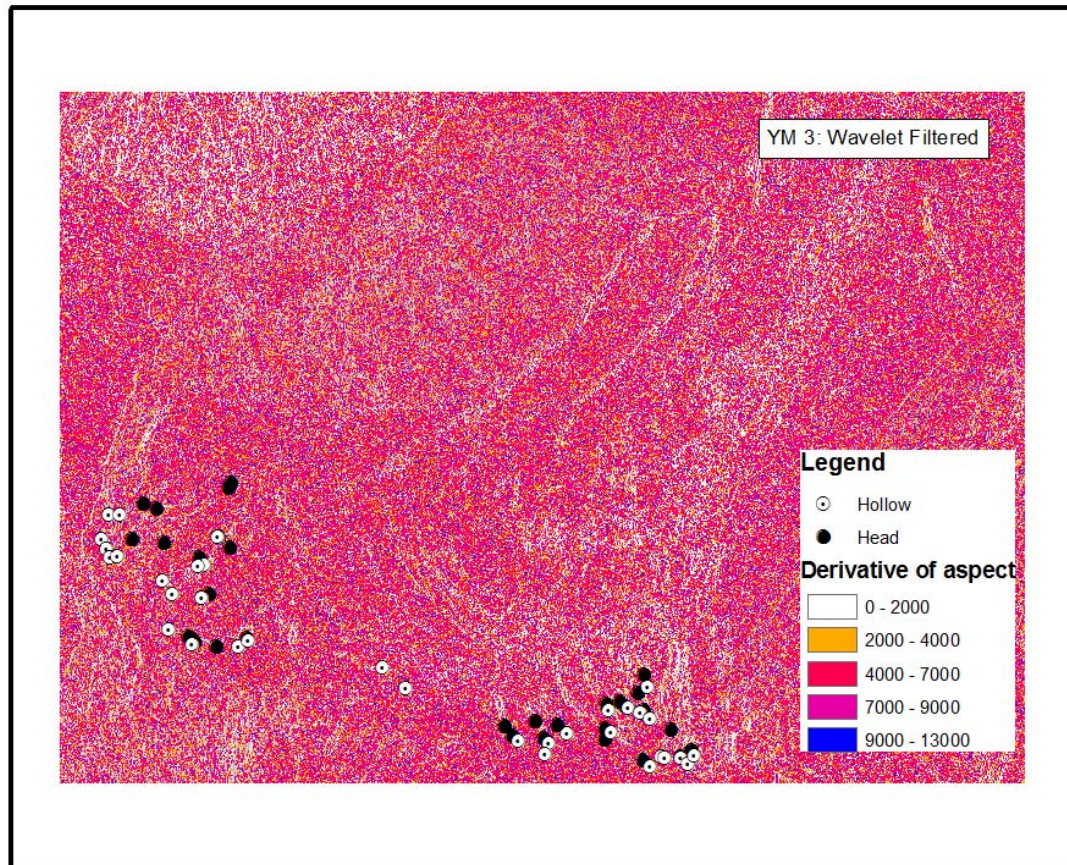


Figure 3.38: Process zones as defined by magnitude-frequency distributions of dthet for wavelet filtered DEM, YM3 catchment.

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CHAPTER 4

IN-CHANNEL PROCESS DOMAINS: IDENTIFICATION OF COLLUVIAL
TO ALLUVIAL TRANSITIONS USING WAVELET TRANSFORMS

Contribution of Authors and Co-Authors

Manuscript in Chapter 4

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Abstract

Multiple channel types are found within the fluvial process domain. Well known channel classification schemes (Schumm, 1977; Rosgen, 1994; Montgomery and Buffington, 1997b) organize channels according to their function, form, and sediment transport capacity. In these schemes, a fundamental distinction is made between colluvial and alluvial channels. The form of colluvial channels is determined by upslope and adjacent hillslopes, rather than by the available fluvial energy. By contrast, alluvial channels create their own form with their own deposits. They also differ in their spatial organization. Alluvial channels tend to produce repeating patterns in form and architecture whereas colluvial channels tend to be random in their arrangement of bedforms and planform. In headwater catchments, the distribution of channel types is likely to consist of both colluvial and alluvial (primarily cascades and step-pools), channels. The upstream extent of dominantly alluvial channel conditions is one measure of the downstream-most extent of a headwater catchment. However, delineating this approximate boundary is non-trivial from field observation much less remote sensing data. An integrated dataset consisting of LIDAR last return points and total station survey data was used to create high resolution DEMs of four channel reaches in three headwater catchments in southwest and central Montana to test for repeating patterns in longitudinal distributions of channel morphology and fluvial energy expenditure. The presence of these patterns is hypothesized to correspond to alluvial subreaches and an absence of patterns is hypothesized to correspond to colluvial subreaches. Using the DEMs, a hydraulic model was constructed of each channel reach to generate longitudinal

distributions of unit stream power, wetted channel width, and width-to-depth ratio. These data were detrended and analyzed by wavelet transforms to identify the location and wavelength of high strength signals, indicating a repeating pattern. Wavelet analysis is a means of decomposing a signal into oscillations (wavelets) highly localized in either time or space. The significance of these peaks in energy was assessed by removing the estimated background spectrum, which was modeled as red noise. This analysis revealed energy peaks in unit stream power with a wavelength ranging from 3 - 10 meters. Wavelengths of the cycles typically increased with downstream distance. Similar patterns were observed in distributions of wetted width. The location of the shorter wavelength peaks appears to correspond to the location of cascade and step-pool channel types.

Introduction

Headwater catchments exhibit significant spatial and temporal variation in geomorphic process domains. At the uppermost part of the catchment, hillslopes meet valleys and channels at the limit of landscape dissection. Further down the incipient channel network, the influence of mass wasting events such as landslides and debris flows on in-channel processes gives way to fluvial processes of local erosion and deposition. This transition can be qualitatively assigned at the interface of the production and transfer spatial zones of the fluvial system (Figure 4.1). This zonation follows downstream trends in discharge, gradient, and grain size, as well as derived parameters such as stream power and shear stress. However, these trends can be interrupted,

particularly in the production and transfer zones, by local variation in controlling variables of lithology, structure, and disturbance regimes. It is these discontinuities that create a mosaic of process domains in headwater catchments, and consequently a mosaic of channel types.

Watersheds have a wide range of stream channel types (Figure 4.2) that tend to follow the generalized fluvial system zones. In the headwaters of the basin, where streams lie in rock and woody debris delivered by gravity from the hillslopes, they are termed “colluvial.” Colluvial channels are characterized by randomly deposited rock and wood debris delivered from hillslopes by mass wasting. They lack the fluvial energy to create their own bedforms and architecture. Bedrock channels are a special case of colluvial channels in the sense that the stream cannot form an equilibrium channel. With increasing drainage area and increasing fluvial energy, channels lie in their own deposits and are then termed “alluvial”. Alluvial channels dissipate energy transporting and rearranging their own deposits to create distinctive bedform architecture such as staircase-like step-pools and repeating meanders of pool-riffle sequences.

Channel type is determined by transport capacity, sediment supply, and grain size of the incoming load (Montgomery and Buffington, 1997b). At the foundation of these concepts is a balance of the driving force of flowing water and the resisting force of the stream channel sediments and woody debris. The structure and function of stream channels varies with the interplay between these driving and resisting forces (Lecce, 1997). The driving force of flowing water has the ability to take disorganized inputs of sediment and woody debris and create repeating bedforms in the vertical plane and

rhythmic, repeating meander patterns in the horizontal plane. When the morphology of a channel reflects only hydraulic and sediment transport processes occurring within the channel, it can display self-organizing behavior. Self-organization in geomorphology includes "periodic structures that form spontaneously on the substrate as water or air transports, erodes, and deposits sediments - ripples and dunes, meanders and bar/pool topography of rivers", as well as "scale invariant structures revealed in drainage networks and topography" (Hallet, 1990).

The classification of channel types is based upon their structure and organization. Madej (2001) described channel structure as the individual architecture of a building and organization as the arrangement of a collection of buildings. This spatial distribution may be random or exhibit a pattern, defined as widespread, repeating signatures in topography (Perron, 2011). Step-pool and pool-riffle channel morphologies are known to produce distinct channel structures and rhythmic patterns of channel organization (Keller and Melhorn, 1978; Grant et al., 1990; Montgomery et al. 1995a; Chin, 1999, 2002). This organization has been linked to how energy is expended in the channel (Leopold, 1966; Yang, 1971; Chin and Phillips, 2007) whether it is minimization of energy expenditure, uniform energy expenditure, or optimization of dissipated energy.

Even recently disturbed channels may exhibit organization. Madej (2001) found that gravel bed streams disturbed by sediment pulses developed topographic regularity in bedform development with time, indicating self-adjustment. Channels subject to external controls such as bedrock and large woody debris also developed topographic regularity but on shorter spatial scales. The development of patterns depends on the frequency and

magnitude of floods able to perform geomorphic work within the channel. It is the distribution of fluvial energy through the channel network that creates a distribution of channel morphologic types and takes disorganized hillslope inputs of wood and sediment and creates channel structure, organization, and regularly spaced bedforms (Madej, 2001).

Force and resistance interactions are fueled by energy. The potential energy of flowing water is converted to kinetic energy, heat of friction, and movement of sediment. In colluvial channels, energy is generally dissipated in flow turbulence and the friction of water flowing over large rocks and debris. In more alluvial channels, energy transports sediment and creates distinctive, repeating forms by preferential erosion and deposition. Sediment transport and deposition mediate the transformation of river energy into organized channel forms. Process and energy are not directly measurable but unit stream power (ω) quantifies energy expenditure, using the properties of mechanical power (Rhoads, 1987).

$$\omega = \frac{QS}{w} = \tau v \quad (4.1)$$

where Q is discharge, S is slope, w is channel width, τ is shear stress, and v is velocity.

Power is the rate at which work is done, and unit stream power is the power dissipated per unit area of the channel bed and a measure of the capacity of the fluvial system to do work. As such, it is a powerful metric in examining the distribution of energy through the channel network. Spatial distribution of stream power in a network has been used to determine extent and magnitude of variations from optimal energy dissipation (Molnar and Ramirez, 1998), calculate deviations from hydraulic geometry

estimates of stream power (Molnar and Ramirez, 2002; Fonstad, 2003), and explain trends in substrate size (Brummer and Montgomery, 2003). These works note the dynamic adjustment between channel form and energy expenditure, as well as the obvious implications for controls on sediment transport capacity and supply. The nonlinear variation of stream power in the downstream direction (Knighton, 1999; Fonstad, 2003; Legleiter et al., 2003) has shown the effect of geologic control on stream power distributions, and natural limitations on attaining the theoretical “optimal state” (Molnar and Ramirez, 1998). In spite of these external controls, energy expenditure has been linked to periodicities in certain channel morphologic types (Chin, 2002) and found to have “a determinant effect on channel morphology” (Marston, 1982).

If the stream expends energy to create repeating patterns in the horizontal and vertical planes, then repeating patterns in unit stream power should also be detectable in alluvial channels. The spatial distribution of energy expended, or unit stream power, should relate to the spatial distribution of channel types (Fonstad, 2003). Spatial distributions of unit stream power were found to correlate with width/depth ratio, a measure of channel shape, in post-fire landscapes (Legleiter et al., 2003). This analysis will test the hypothesis that the transition from colluvial (disorganized) to alluvial (organized) channels in headwater catchments can be detected from spatial patterns of energy expenditure and channel morphology.

Organization of channel morphology is defined as regular patterns in oscillation of the channel in the horizontal and vertical dimensions. Organization in energy expenditure is defined as the presence of predictable, repeating patterns in the spatial

distributions of unit stream power. The presence/absence of spatial patterns should be most evident at the transition from colluvial to alluvial channels. The bedform architecture and topography of colluvial channels, driven by hillslope processes, should be random and poorly organized in terms of repeating patterns. Alluvial channels, as authors of their own geometry (Leopold, 1994), should have distinctive spatial patterns of bedforms. This presence/absence of detectable spatial pattern should be reflected in spatial patterns of energy dissipation.

The transition from colluvial to alluvial process can be suggested by the presence of patterns in alluvial reaches and randomness in colluvial reaches. Similar to the association of power law behavior with fractal structure and scale-invariance, organization, as defined as the presence of patterns, is associated with alluvial process.

Methods

Testing for patterns in energy expenditure and channel form across multiple spatial scales requires high resolution topographic analysis at the scale of headwater channels (Fonstad and Marcus, 2010). Energy expenditure, or stream power, has been used as a proxy to understand process in multiple studies, (Molnar and Ramirez, 1998, 2002; Fonstad, 2003;) but the spatial scales of these earlier studies were limited by sparse and low resolution (30 m.) data coarse compared to the scales at which fluvial process operates. The approach used here integrates remote sensing (here Lidar) with field data to enable spatially comprehensive, quantitative analysis; an approach advocated by Carbonneau et al. (2012).

This analysis uses LIDAR last return point data and total station survey data to create 1.0 m DEMs of two reaches of headwater channels in the TCEF study area and two reaches in the Big Sky study areas. The analysis sequence relies on the development of high resolution hydraulic models of the reaches to generate submeter longitudinal distributions of energy expenditure (unit stream power) of reaches of four headwater channels in the catchments.

Unit stream power was calculated for the four reaches using HEC-GeoRAS modeling and the generated DEMs. Wavelet analyses were applied to each spatial series to detect spatial patterns and identify period, location, and type of spatial patterns in unit stream power and channel form. The wavelet analysis takes the fast Fourier transform of the time (here, spatial) series to produce the power spectra, identifying cycles of highest power or significance, as well as their location. The spatial structure of energy expenditure and channel form was quantified using a one-dimensional continuous wavelet transform. This technique has been used to investigate the spatial relationship between channel structure and aquatic habitat (McKean et al., 2008). McKean et al. (2008) performed wavelet analysis to describe the spatial structure and patterns in stream longitudinal profiles and mapped salmon nests over a 10 km channel segment. They detected patterns in profile that closely followed patterns in salmon nests, with 50 - 100 m. offsets.

Study Area

Four first order channel reaches were surveyed in the U.S. Forest Service Tenderfoot Creek Experimental Forest (TCEF), located in the Little Belt Mountains, and

in the Madison Range, south of Big Sky, Montana. The study areas are comprised of three catchments: Stringer (STR) and Spring Park Creeks (SPC) in TCEF (Figure 2.3) and Second Yellowmule Creeks near Big Sky (YM2), (Figure 2.4).

The channels STR and SPC are north-south trending with a south aspect, sharing a common watershed divide. Bedrock consists of gently dipping Cambrian sedimentary rock, largely Flathead Formation and Woolsey Shale, with some outcrops of Eocene quartz monzonite capping the upper parts of the catchments (Reynolds and Brandt, 2007). The Flathead Formation in this area is described as having a thick middle shale layer beneath a thin and above a massive basal sandstone layer (Weed and Pirsson, 1899). Most of the headwater channels begin in shale (Woolsey or middle Flathead) but the main stem valley floor is mantled with debris flow lag of sandstone flagstones rafted by failing shale layers. The lowest part of the study area is underlain by the lower unit of the Flathead Formation, a resistant sandstone unit. At this change in lithologic type, the channel transitions from a pool-riffle morphology to cascade and step-pool morphology.

The YM2 catchment is also north-south trending, but with a north aspect. The catchments is underlain by Morrison Formation, Cretaceous Muddy Sandstone and Thermopolis Formation and Kootenai Formation, thin sandstone layers separated by thicker shale layers (Kellogg and Williams, 2000). The catchment forms on the gently north-dipping north limb of an east-west-trending asymmetrical anticline, the axial crest of which underlies the southern drainage divide of the West Fork.

The four surveyed reaches are 382 - 203 m. in length. Although the density of surveyed points is high, spaced approximately every 0.5 m., the surveyed reach lengths

are relatively short. High data density ensures channel geometry is accurately modeled but short reaches limit the number of cycles that may be detected, dependent on period. Larger scale effects such as changes in valley type are also unobserved over short reach lengths.

Reaches approximately 281 and 365 meters in length were surveyed on Stringer (STR) and Spring Park (SPC) Creeks, respectively. The STR and SPC reaches were selected to include portions of the perennial channel exhibiting repeating meanders and the presence of pool-riffle sequences. The location of the STR reach was also selected to include portions of channel above and below the confluence with the West Fork of Stringer Creek, a major tributary. Two reaches, YME and YMW, (203 and 382 meters in length) were surveyed on tributaries of Second Yellowmule Creek (YM2). The YM2 tributary surveyed on the east side of the catchment (YME) includes sub-reaches where the channel flows through debris flow lag. This reach is ephemeral, flows over hillslope-derived material, and does not display obvious repeating patterns in either planform or bedform organization. The YM2 tributary surveyed on the west side of the catchment (YMW) is also ephemeral but deeply incised into a thick shale/colluvium unit. It displays a well-developed meander pattern in planform but few organized bedforms.

DEM Generation

The DEM of each channel reach was generated from a triangulated irregular network (TIN) in ArcGIS 9.3. The TIN is constructed from mass points, elevation points used as nodes in the TIN, and breaklines, linear features that define the behavior of the TIN in terms of defining hard edges. Filtered LIDAR laser return points containing the

elevation of ground points reflected by airborne laser were used as mass points. Channel thalweg and banklines were surveyed using a Nikon total station instrument for each of the four reaches (Figures 4.4 - 4.6). Surveyed thalweg and banklines were used as breaklines.

Due to the remote nature of the study areas, benchmarks were established using a mapping grade Trimble GPS receiver. These benchmarks were occupied by the survey instrument to provide known locations to initiate the survey. These benchmarks were later corrected by post-processing to improve the accuracy of the benchmarks. The coordinates of the post-processed benchmarks were used in TransIt, the Nikon total station data reduction software, to update the surveyed data. The accuracy of the total station survey data is a function of the post-processed benchmark accuracy and the total station errors. The accuracy of the post-processed benchmarks was calculated at 0.1 - 0.5 m. by Trimble Pathfinder Office. Closure accuracy of the total station surveys was less than 0.1 m. The surveyed banklines and thalweg elevation points were used to construct breaklines to define the spatial extent of the channel. The resulting TIN was gridded to produce a 1 m. DEM for each channel reach.

Hydraulic Modeling

The U.S. Army Corps of Engineers HEC-RAS (Hydrologic Engineering Center-River Analysis System) software package was used to build a hydraulic model of each of the four reaches. The model solves the one-dimensional energy equation to generate a variety of hydraulic flow parameters at user-defined cross-sections. The results of this model were used to construct longitudinal distributions of unit stream power, or unit

stream power versus downstream distance. Input data to the model include stream channel cross-sections, Manning's roughness values, discharge values, and flow boundary conditions. However, the HEC-GeoRAS plug-in for ArcGIS 9.3 was used to construct the input files. The HEC-GeoRAS plug-in uses a terrain model (either DEM or TIN) and a user-defined cross-section layer to generate the HEC-RAS geometry input files. These files are output in RAS format and read into HEC-RAS. Given the RAS geometry files and user-defined flow files, the HEC-RAS hydraulic model was run for each of the four channel reaches.

The modeled flows in TCEF were selected to roughly match the surveyed bankline elevations. Typical bankfull indicators such as an incipient floodplain were used where available to define the bankline. In the TCEF reaches, the modeled flow (0.3 cms) roughly corresponds to a two-year flood (Parrett and Johnson, 2004). At the downstream termination of the SPC and STR surveyed reaches, contributing area is roughly 3.2 km² in each catchment. In the Big Sky reaches, the modeled flow (0.1 cms) roughly corresponds to a ten- to twenty-five year flood, estimated using USGS regression methods (Parrett and Johnson, 2004). However, the small catchment areas of YME and YMW (less than 0.1 km² each) render suspect any estimates of flood recurrence intervals derived from regression methods commonly developed for channels draining tens of square kilometers. Spatial series of unit stream power versus downstream distance, wetted width, and wetted width to depth ratio, a measure of channel shape, versus downstream distance were saved in an Excel spreadsheet. Wetted width was used in addition to width/depth ratio because it is a primary (not derived) metric.

Wavelet Analysis

A wavelet is a family of functions of irregular and asymmetric waveform, with limited duration and zero mean. The wavelet transform decomposes a spatial signal into energy or amplitude over given spatial frequencies, at each position in the data. Wavelet analysis adjusts its scale to the frequency of interest such that fine scale spatial resolution is used at high frequencies and coarse scale resolution at low frequencies.

The AutoSignal software package was used to perform wavelet analyses on the unit stream power, width to depth (w/d) ratio, and wetted width spatial series. Each series is detrended before wavelet analysis. The Continuous Wavelet Transform (CWT) function is used to decompose the series into Morlet wavelets (small oscillations that are highly localized in time) and construct a space-frequency representation of the series. The results of the wavelet analyses are plots of normalized power spectral density versus downstream distance, showing the strength, location, and spatial extent of cyclic behavior of each parameter. To assess the significance of the generated peaks, the background spectrum was estimated and removed. "Noise" is a random addition to a signal. The power spectral density of the signal is used to assign noise a "color." "Red noise" is a signal whose spectrum plot is characterized by increasing power at lower frequencies. Red noise is commonly associated with natural systems (Roe, 2009) and usually represented as an autocorrelation model with lag 1, AR(1).

$$x(t+1) = x(t)a_1 + \varepsilon(t) \quad (4.2)$$

where $x(t)$ is the value of a variable at time t , a_1 is the autocorrelation coefficient with lag 1, and $\varepsilon(t)$ is a white noise function. The coefficient a_1 reflects the degree of persistence

or memory in the modeled system. Here, the background spectrum is modeled as an AR(1) system with an autocorrelation coefficient of 0.8, reflecting a likely high degree of memory.

Planform Analysis

Regular patterns in the horizontal dimension as a marker of organization was also tested. For each surveyed reach, a valley centerline was digitized in ArcGIS 9.3. The perpendicular distance from the valley centerline to the surveyed stream centerline was determined as a spatial series of valley centerline deviations versus downvalley distance. (Downstream valley distance increases with increasing distance.) A wavelet analysis was performed in AutoSignal on each detrended series to test for repeating patterns. Each series was detrended by removing the mean and a least squares linear fit model. However, this approach retains the long wavelength attributes of the series. To enhance pattern detection at the scale of the channel, long wavelength features were effectively filtered by also detrending each series using a cubic fit. The CWT function was used to decompose both sets of detrended series using the Morlet wavelet. A Fourier analysis was also performed on the spatial series to generate a Lomb-Scargle periodogram for data with unevenly spaced abscissa. This analysis was performed as a secondary test of the significance of generated peaks.

Results

Two different HEC-RAS hydraulic models were run for the SPC reach to account for issues with benchmark coordinate stability. SPC model 1 was constructed using total

station survey data only, while SPC model 2 was constructed using only the 1 m. NCALM DEM. SPC model 2 begins roughly 150 meters upstream of the survey model. The two models share the same stationing with the second model beginning upstream at approximately River Station 450. SPC model 1 also ends roughly 30 meters upstream of the second model due to a lack of survey points. (Note that in all RAS models, direction of flow is from higher river stations to lower.)

Results of the wavelet analysis of unit stream power generated by SPC model 1 (Figure 4.7) show a repeating signal with a period of 3-5 meters from roughly River Station 275 - 225. Another localized signal with a 5-10 meter period is observed from River Station 200 - 175. Red, yellow, green, and blue colors denote significance levels of 99.9%, 99%, 95%, and 90%, respectively. Power spectra of top width and w/d ratio show significant peaks at similar periods and similar locations, as well as additional wavelengths and locations. SPC model 2 produces a power spectrum (Figure 4.8) which shows peaks more localized in location but at similar wavelengths.

Similar issues were encountered in the STR model such that two hydraulic models were constructed. STR model 1 was constructed from last return points and total station survey data but only covers the downstream-most 100 meters of the study reach. STR model 2 was constructed from the 1 m. NCALM DEM for the entire 350 m. reach. Power spectra of both models (Figures 4.9 - 4.10) show a repeating signal in both unit stream power, top width, and w/d ratio at periods of 5 - 10 m. at River Stations 80 - 40. STR model 2 (Figure 4.10) also shows peaks of somewhat higher frequency (5-7 m.) at River Stations 240 - 225.

Integration of last return points and total station survey data was successful in the two surveyed reaches in the YM2 catchment. Consequently, the hydraulic models constructed for YME and YMW were both derived from a DEM generated by both sources of data. Power spectra of YME (Figure 4.11) show high energy signals in unit stream power with period of roughly 3.5 - 8 m sporadically from River Station 350 -225. Spectrum of top width showed peaks of similar period but only from River Station 250 - 225. Spectra of w/d ratio also show peaks at those River Stations as well as River Stations 140-125.

Power spectra of unit stream power in YMW (Figure 4.12) show energy peaks at periods of 2 -5 m. at River Stations 175 -150 and 125- 110. Lower energy peaks are exhibited at periods of 5-7 m. downstream at River Stations 125-25. Between the power spectra of top width and w/d ratio, peaks of similar period and location are observed.

Power spectra of valley centerline deviation series (linear detrending fit) show the presence of two long wavelength features in SPC and STR (Figures 4.13 - 4.14), ranging from 128-155 m. and 71-85 m. Power spectra of YME and YMW (Figures 4.15 - 4.16) share a feature of common wavelength (114-115 m.).

Pattern detection at smaller spatial scales was enhanced by removal of a cubic trend in each series (Figure 4.17). Patterns at the scale of the channel are more likely to be the result of fluvial process. Most of the peaks exhibited by the SPC and STR series detrended by cubic fit have a period of 6 -10 m. (Figures 4.18 - 4.19). The YME series (Figure 20) exhibits peaks with periods of 7 and 12.5 m. Power spectra of YMW (Figure 21) has multiple peaks with periods of 3.5 - 5 m. and 25 m.

Discussion

Power spectra of both SPC models exhibit high energy signals at similar periods (3 - 7 m., 5 - 10 m.) but often different locations. SPC model 2 is limited by a lack of in-channel data. The LIDAR equipment used in TCEF is unable to penetrate water compromising the accuracy of derived channel geometry. However, the total station survey data contains only in-channel information, but of limited density. In addition to differences in location energy peaks, power spectra of the SPC model 2 are extremely localized in location. This behavior may result from discrepancies in the DEM extrapolation across the channel with no intervening last return points. However, power spectra of SPC model 1 exhibits high energy signals over longer subreaches. High energy signals in unit stream power and top wetted width occur at River Stations 350 - 325, 275 - 225, 210 - 150, and 105 - 50. Field observations of SPC from approximately River Station 210 downstream show well-developed meanders and pool-riffle sequences, except where the channel approaches the hillslope toe. At these interfaces, channel morphology converts from pool-riffle to step-pool form.

The planform analysis reveals different results using two detrending models. The linear detrending analysis of the SPC planform shows the presence of a pattern, albeit a limited one. The wavelength of the feature (85 m.) limits its repetition to one or two cycles. The cubic detrending analysis shows a feature with a period of 6 - 8 m. occurring at two locations. Each location corresponds to one full meander wavelength. The upstream location of the first cycle roughly coincides with high energy signals in both unit stream power, top wetted width, and w/d ratio at River Station 150 (Figure 4.7).

This commonality indicates that the channel is able to adjust its boundary in three dimensions. Evidence of statistically significant pattern in variables of energy, morphology, and planform are suggestive of dominantly alluvial conditions. Observed pool-riffle sequences are functional equivalents to patterns in energy, supporting this contention.

Power spectra of the STR models was much more consistent, possibly owing to the morphology of the channel itself. This reach of STR is relatively steep, shallow, and rocky. The LIDAR survey was performed in September, during low flow conditions, potentially enabling the collection of more last return points within the channel. High energy signals are localized at River Stations 250-225, 190-160, and 90-40. Periods range from 5-10 m. One possible explanation for energy at River Stations 190-160 and 90-40 is the confluence of the STR reach with the West Fork of Stringer at approximately River Station 195 and another tributary entering at River Station 83. Downstream of these confluences, the channel exhibits a series of well-developed meanders and pool-riffle sequences shortly before encountering the sandstone unit of the Flathead Formation. Of interest is a notable peak in unit stream power at River Stations 250-225 but a lack of corresponding peaks in either top width or w/d ratio. Possible explanation include modeled flow values (~10.5 cfs) that either wash out changes in morphology or are entirely within the channel margin, or inaccurate channel geometry in the hydraulic model.

The power spectra of the STR planform shows two long wavelength features, similar in period to the SPC planform. The shorter feature (71 m.) occurs near the

confluence with the West Fork of Stringer. The longer wavelength feature (128 m.) extends over much of the reach. Power spectra of the cubic detrended series show features with a period of 7- 13 m. A similar period and location as peaks exhibited in unit stream power and channel form (Figure 4.9). This collocation of statistically significant features of similar periods suggests that alluvial process is creating organization in these subreaches of the channel, in both the vertical and horizontal dimensions.

The power spectra of the SPC and STR planform show remarkable similarity in period of both long and short wavelength features. The period of long wavelength features is many times the width of the channel in this reach, and unlikely to be the result of processes occurring within the channel. More likely, these long wavelength features result from controls of larger spatial and temporal scales such as lithology, structure, and regional tectonic activity. Shorter wavelength features, on the scale of the channel itself, are more likely the result of fluvial process. Channel planform is the result of complex interactions between flood magnitude and frequency; cohesion, strength, and size of bank material, and size and mobility of incoming load. These variables are generally uniform between the two channels, except for the stochastic presence of relict debris flow lag in each catchment.

Power spectra of YME differ from YMW in the number of high energy signals. The YME reach differs from the YMW reach in two important respects: YME is subject to multiple inflows from smaller channels that confluence near approximate River Stations 275, 245, 195, and 130 (Figure 4.11) and is able to access a floodplain surface.

The YMW reach is deeply incised into a thick shale unit and does not confluence with other channels over the reach length. Multiple inflows to YME may explain sporadic spikes in spectral power. Power spectra of top width and w/d ratio exhibit similar peaks of relatively short spatial persistence. The short period of the spectral peaks indicates high frequency features such as the cascades and step-pools observed in the reach.

A point discharge inflow may bring incoming load, and this load is used to form short stretches of alluvial channel. Once this sediment supply is exhausted, the channel resumes its supply-limited condition and is unable to shape channel form. Multiple transitions between alluvial and colluvial conditions may produce closely spaced but discrete high energy peaks.

By contrast, the period of the peaks of the YMW power spectra increases over the reach length, from roughly 2-4 m. to 5-7 m. Locations of the highest energy peaks correspond to the portions of the reach exhibiting the greatest sinuosity.

Power spectra of the YME and YMW planforms also show important differences. The linear detrended YMW planform series shows a repeating cycle of 114 m., centered in roughly the middle of the survey reach, near the same location as high energy peaks in unit stream power and top wetted width (River Stations 125 - 50). However, the short nature of the surveyed reach limits the pattern to one cycle. Although the YME reach also exhibits a feature with a period of 115 m., the normalized power is much less than in YMW. The cubic detrended series of YME planform shows few extended locations of repeating cycles despite the presence of multiple tributary inflows. High energy signals in YME are present in unit stream power and channel morphology but a lack of pattern in

planform suggest that intermittent, in-channel, alluvial conditions may exist during floods of sufficient magnitude to deliver and rework incoming load. However, the channel boundary is largely immobile, except at the largest floods.

In contrast, YMW exhibits sustained repeating pattern in planform, unit stream power, and channel morphology. There is considerable spatial overlap between high energy peaks in all four spectra and very similar periods as well. Similar to SPC, these results suggest dominantly alluvial conditions are achieved in YMW.

Conclusions

Power spectra of unit stream power and two metrics of channel morphology are suggestive of the presence of repeating patterns. However, extrinsic influences of colluvial forcing agents such as immobile lag and hillslope interaction limit the degree of organization possible within these settings. Grant et al. (1990) found the distribution of channel units in high gradient streams influenced by bedrock and hillslope-derived sediment sources. The distribution of channel bed features (bedforms longer than one channel width) exhibited a repeating pattern except in the presence of bedrock outcrop and colluvial boulders. The study reaches may flow over bedrock for short distances but they are considered equivalent to colluvial in that the resisting forces far exceed driving forces. Comparisons between channels are confounded by differing upstream contributing area and consequently different peak discharge values, but comparisons may be made of the period of repeating patterns relative to channel width. For example, Chin (2002) found significant periodicities in the occurrence of step-pools, that varied from

roughly 0.5 - 2.4 channel widths. These ratios of pattern period to channel width are similar to those exhibited here.

In TCEF, STR model 1 exhibited the clearest and most explainable patterns in both unit stream power and channel form. High energy signals in unit stream power and channel form appear subsequent to confluence with tributaries. Spectral analysis of the planform of the SPC reach indicates pattern with a period of 5- 10 m. in the horizontal dimension corresponding to two separate meander wavelengths. The presence of high energy signals of similar period and location in unit power, channel form, and planform of SPC model 1 dominantly alluvial conditions. The similarity in period of the long wavelength planform features of the STR and SPC reaches indicates common controls on planform such as bank cohesion and strength, bedload size, and a similar hydrologic regime.

In YM2, striking difference in power spectra are observed between YME and YMW likely due to multiple channel inflows within the YME reach and none in the YMW reach. The development of weak pool-riffle sequences in the downstream 30 meters of YMW may reflect the change in period of high energy signals from 2-4 m. to 5-7 m. The locations of closely spaced high energy peaks in YME are proximal to tributary inflows and suggest multiple transitions between alluvial and colluvial conditions. Issues associated with fidelity of the hydraulic model to in-channel geometry must also be considered in the assessment of power spectra.

Power spectra of the YME and YMW planforms have similar periods but different signal strength. The YMW planform exhibits pattern roughly coincident with

high energy signals in unit stream power and channel morphology, suggesting that intermittent alluvial conditions exist.

The differences between the TCEF and Big Sky catchments in planform patterns points to underlying differences in the balance of driving and resisting forces. The Big Sky channels are supply-limited, such that channel morphology and planform is dominated by structure and lithology, unbuffered by a more malleable regolith mantle. Consequently, repeating pattern is difficult to create given low excess energy. The YMW reach is an exception to this condition, as this reach is developed in a thick but malleable shale. Both upstream and downstream of this reach, the channel intersects and is dominated by sandstone bedrock. In TCEF, hillslopes are soil and regolith-mantled, and valleys have well-developed soils, such that channels can more readily build and adjust their architecture and planform to create organization.

Channel form in the vertical and horizontal dimensions is the product of fluvial processes. These forms are well-organized when alluvial process is dominant. The common periods exhibited within the YMW, STR, and SPC spectra suggests that the organization of channel form, in vertical and horizontal dimensions, is related to and produced by, the organization of unit stream power. In summary, the pattern of how fluvial work is done may drive the pattern of the resulting channel form.

Although these results are suggestive, further work is needed to calibrate the estimate of background spectra and produce high resolution models of in-channel geometry by remote sensing. The spatial scale of this analysis is limited by the density of Lidar last returns (~ 1 point/m²) and the requirement for detailed channel geometry.

However, detailed local studies provide the opportunity to test our hypothesis in reaches where field observations can confirm analytical results. Next generation Lidar may provide both the required high resolution and the spatial extensiveness required to derive statistically significant data populations.

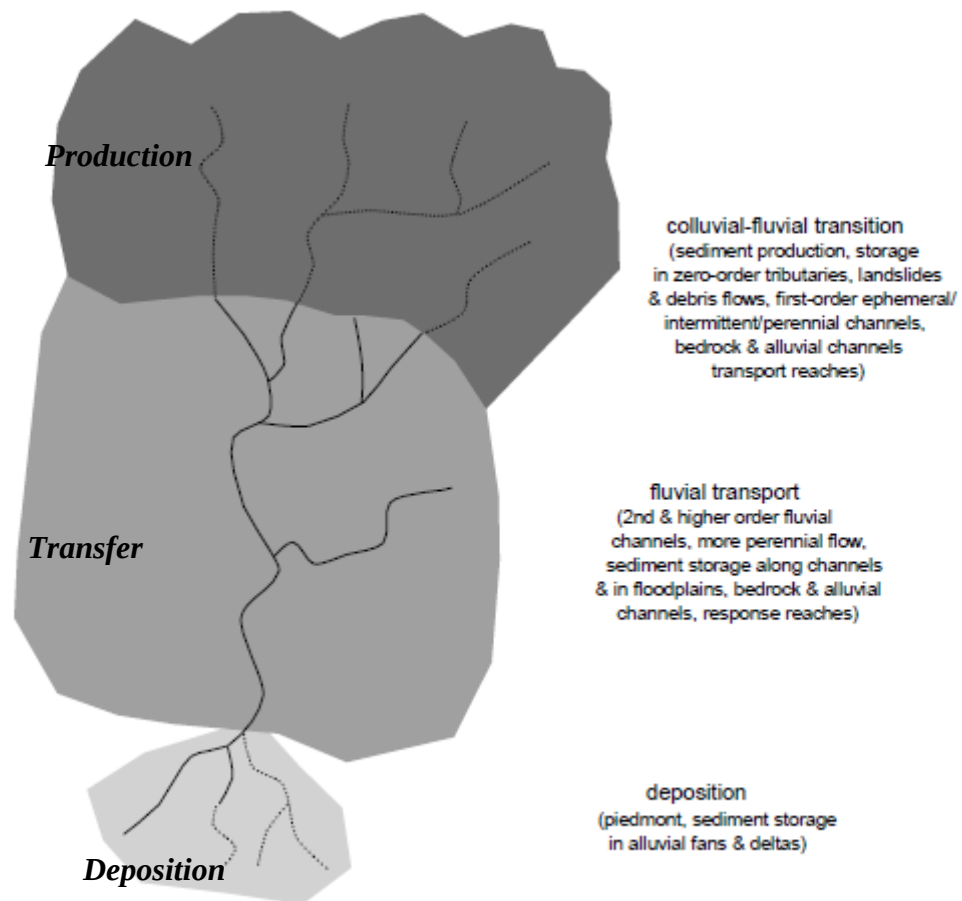


Figure 4.1: Fluvial system spatial zones (Wohl, 2010).

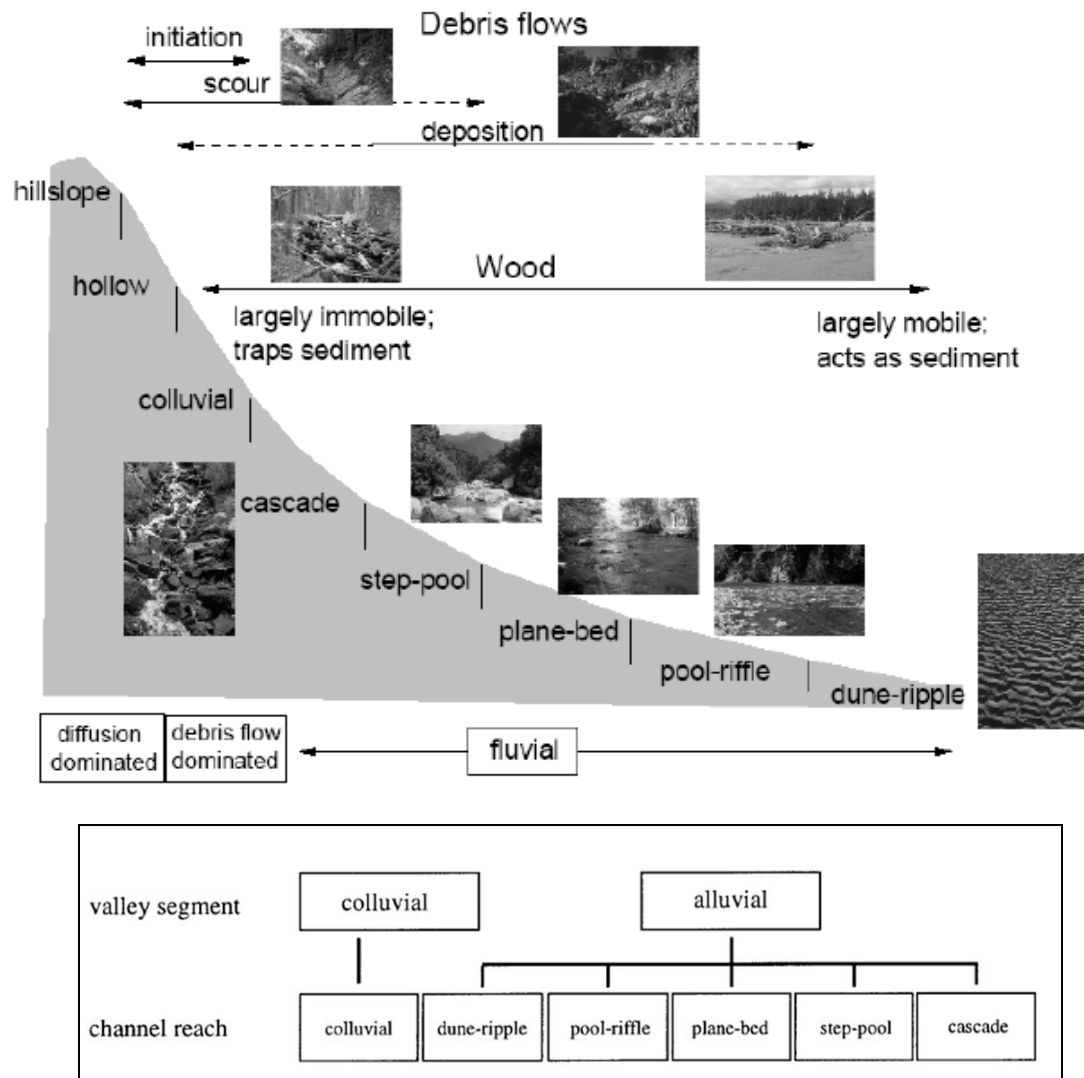


Figure 4.2: Channel types in mountain watersheds (Wohl, 2010; Montgomery and Buffington, 1997b).

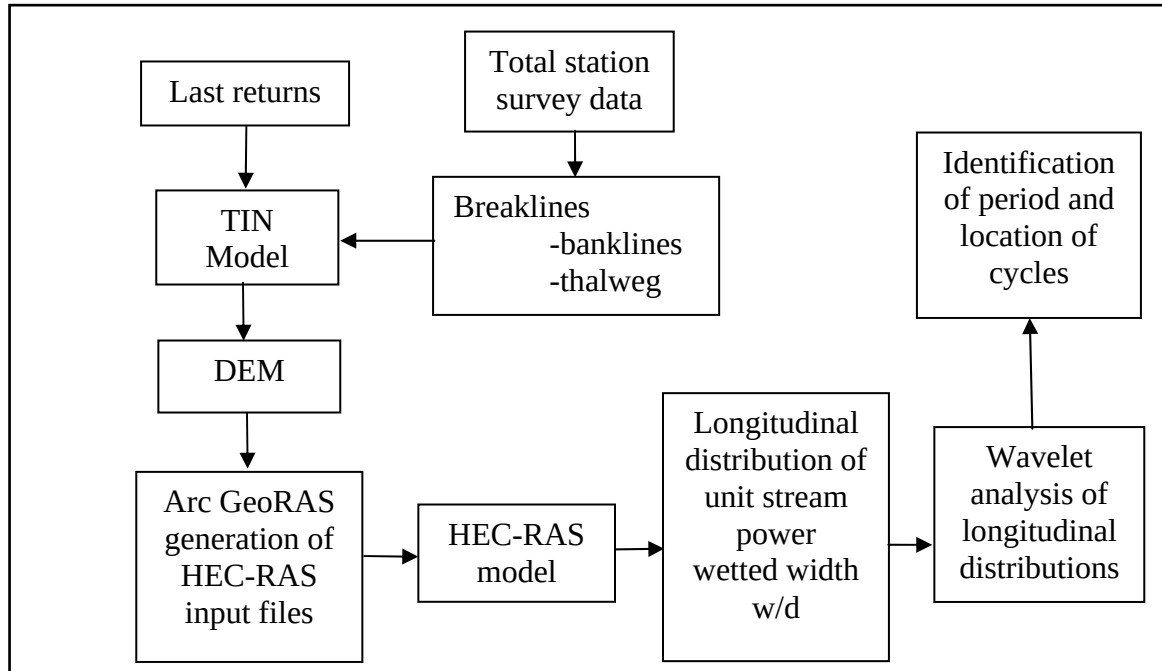


Figure 4.3: Analysis sequence for generation of longitudinal distributions of unit stream power and wavelet analysis.

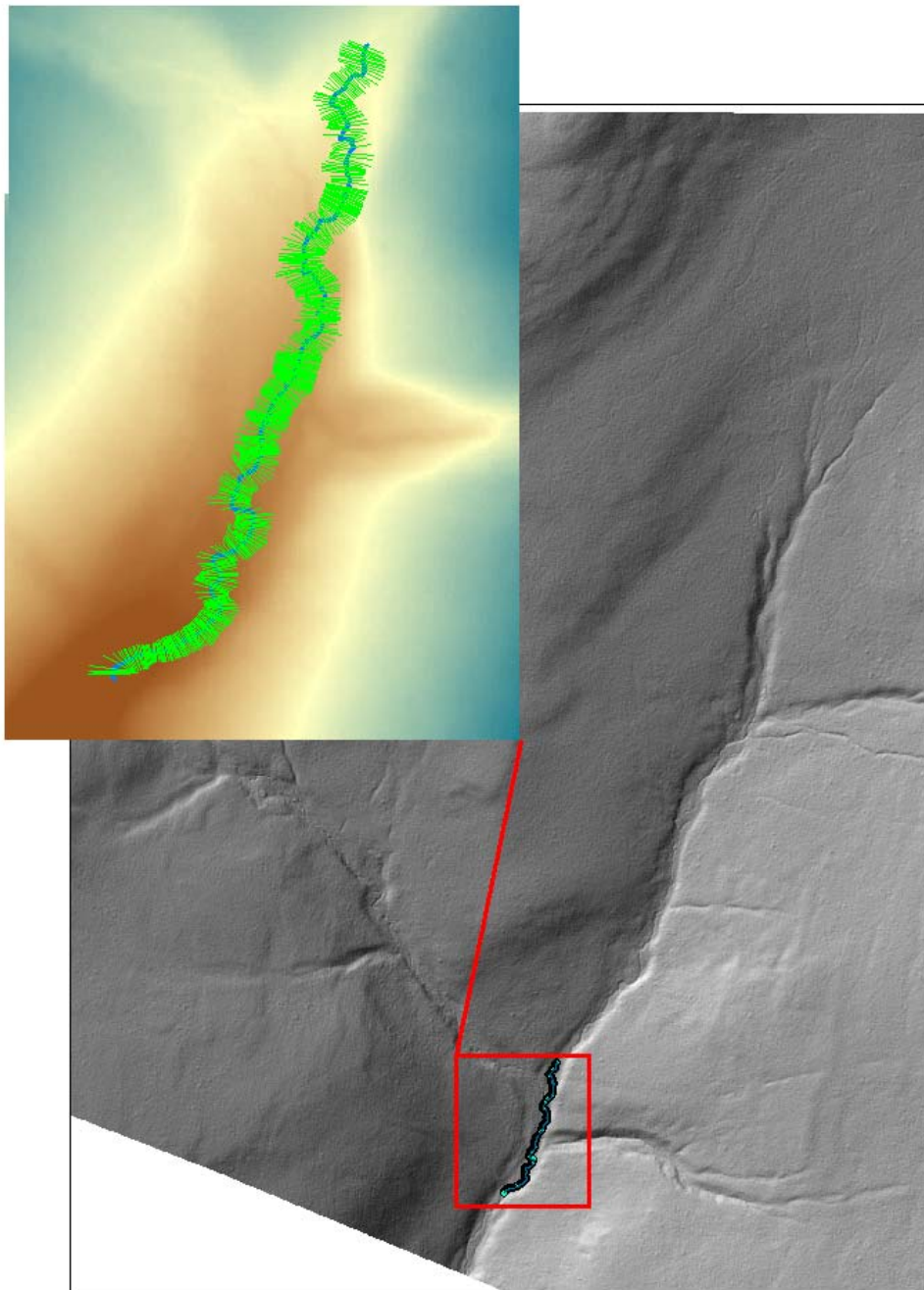


Figure 4.4: Location of Stringer (STR) Creek surveyed reach and inset map showing HEC-GeoRAS cross-sections.

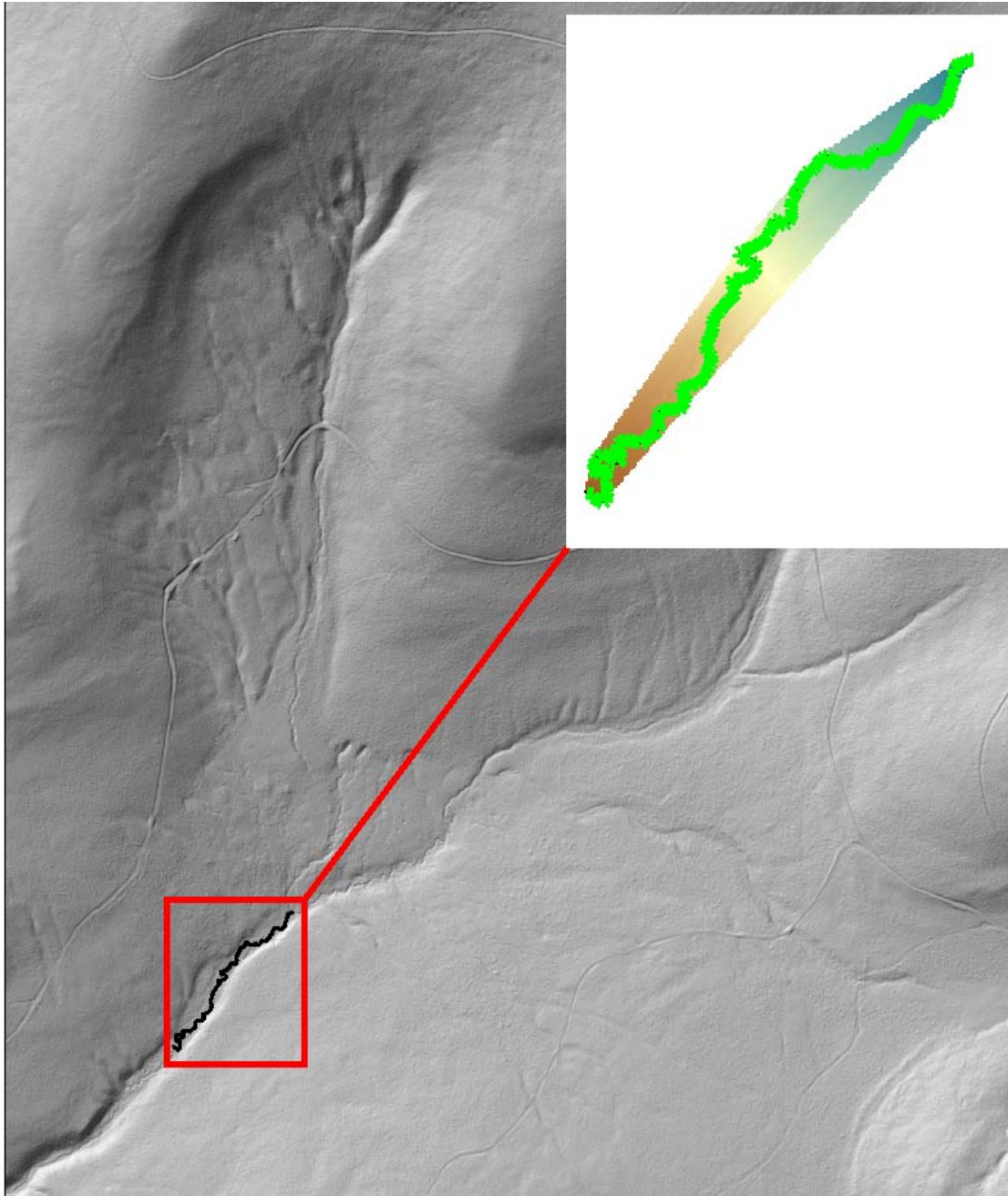


Figure 4.5: Location of Spring Park (SPC) Creek surveyed reach and inset map showing HEC-GeoRAS cross-sections.

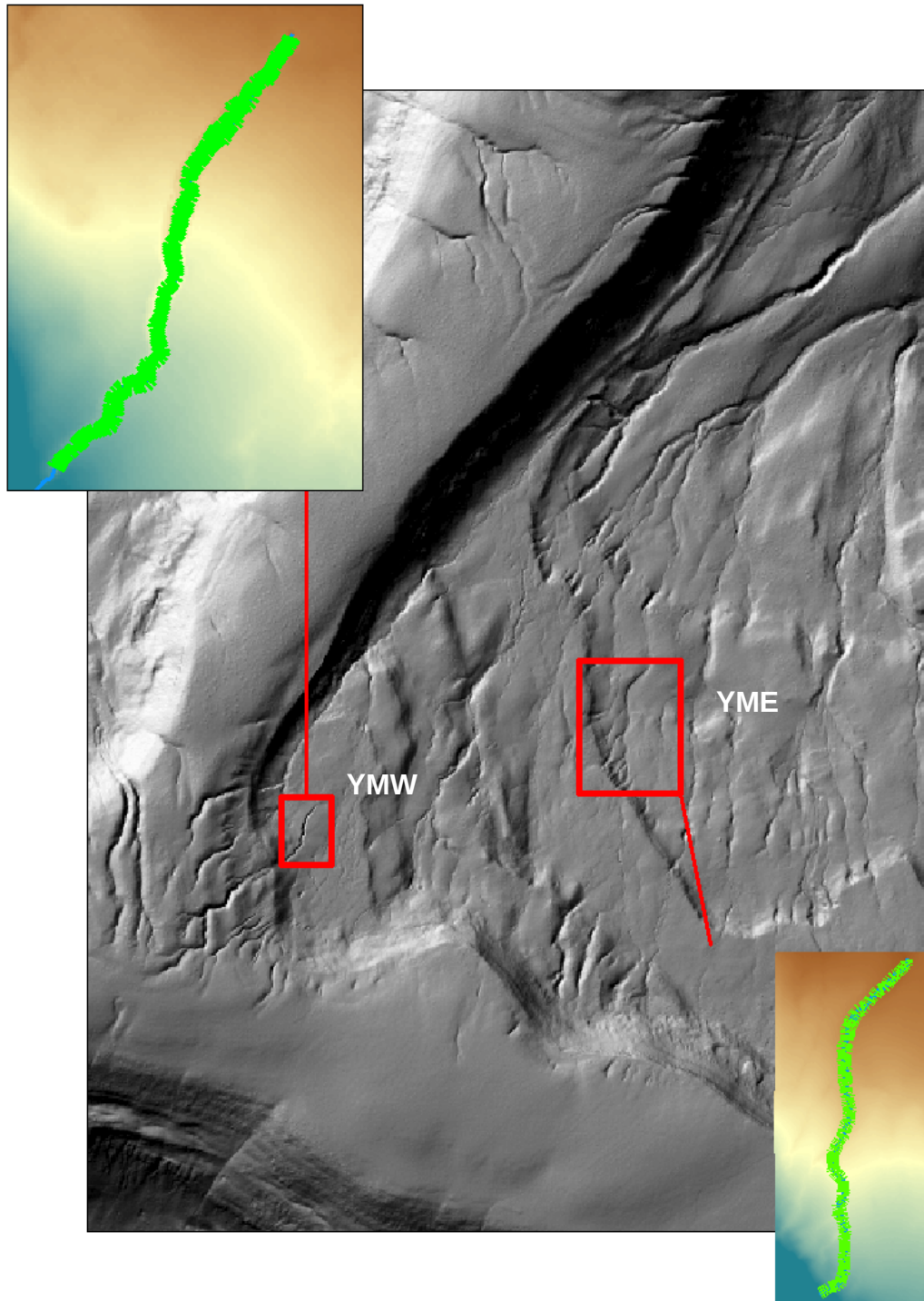


Figure 4.6: Locations of YME and YMW surveyed reaches and inset map showing HEC-GeoRAS cross-sections.

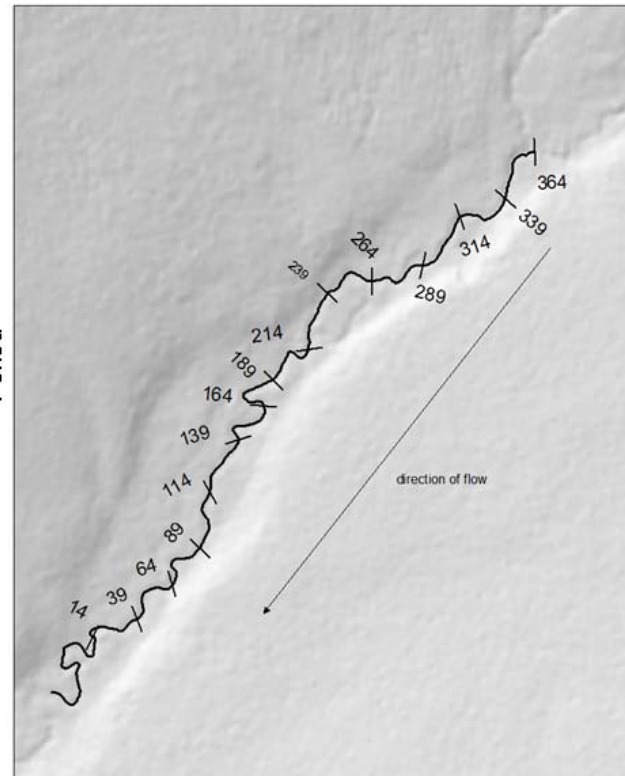
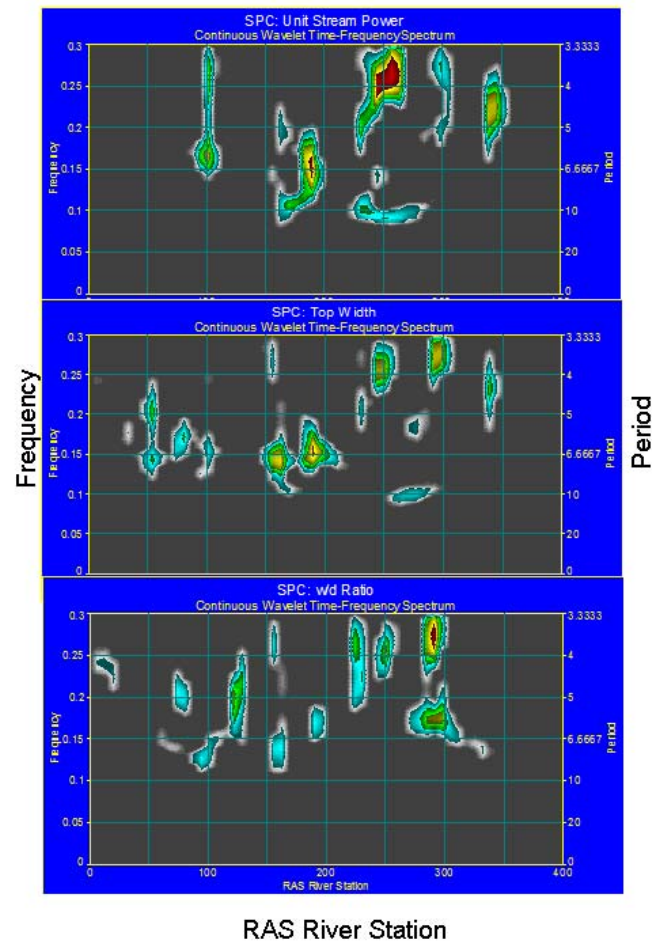


Figure 4.7: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, SPC. RAS model derived from total station survey data only. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

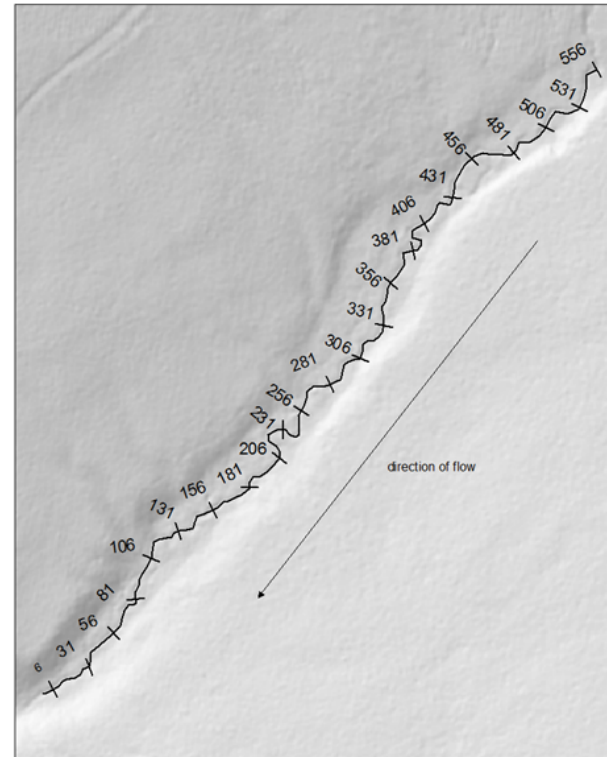
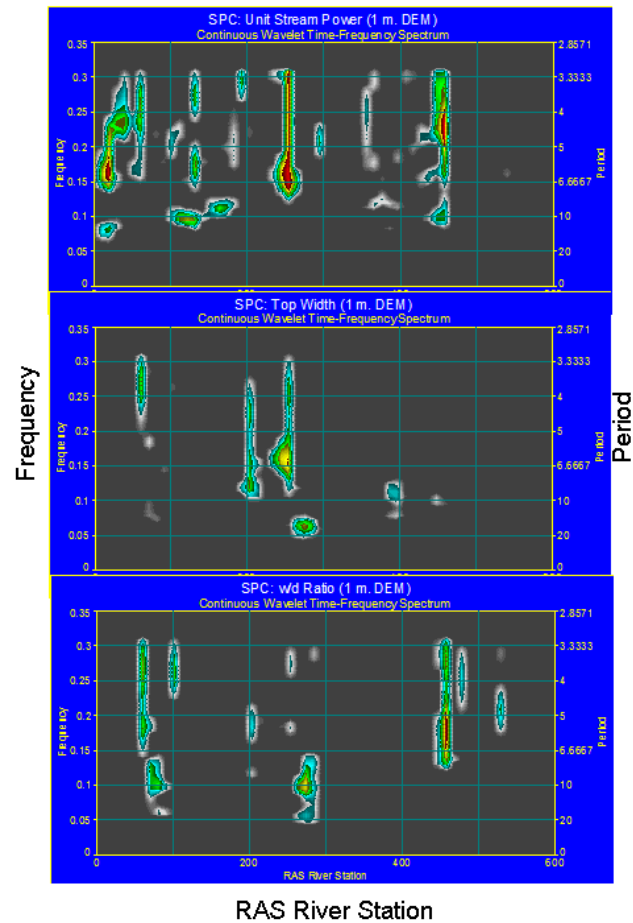


Figure 4.8: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, SPC. RAS model derived from 1 m. DEM only. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

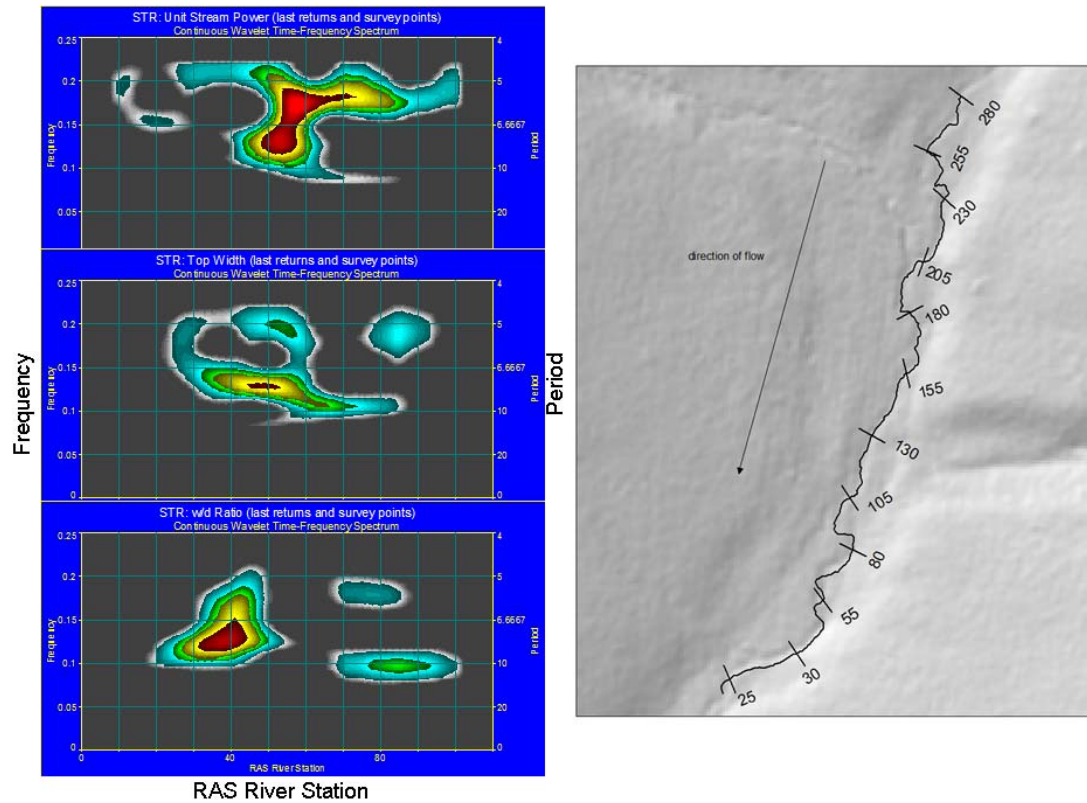


Figure 4.9: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, STR RAS model derived from last returns and total station survey data, downstream-most 100 meters only. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

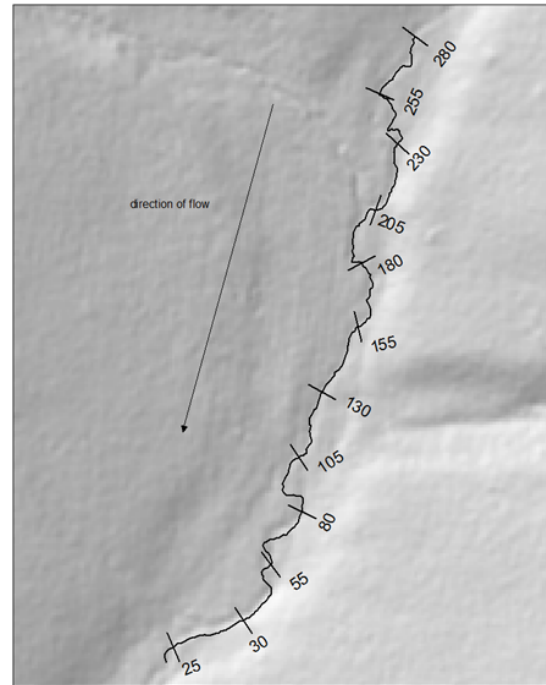
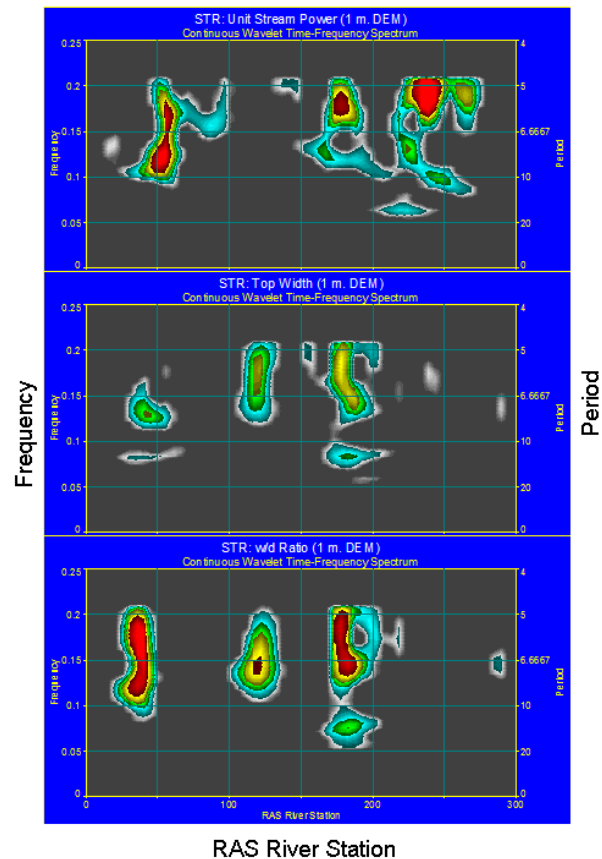


Figure 4.10: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, STR RAS model derived from 1 m. DEM only. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

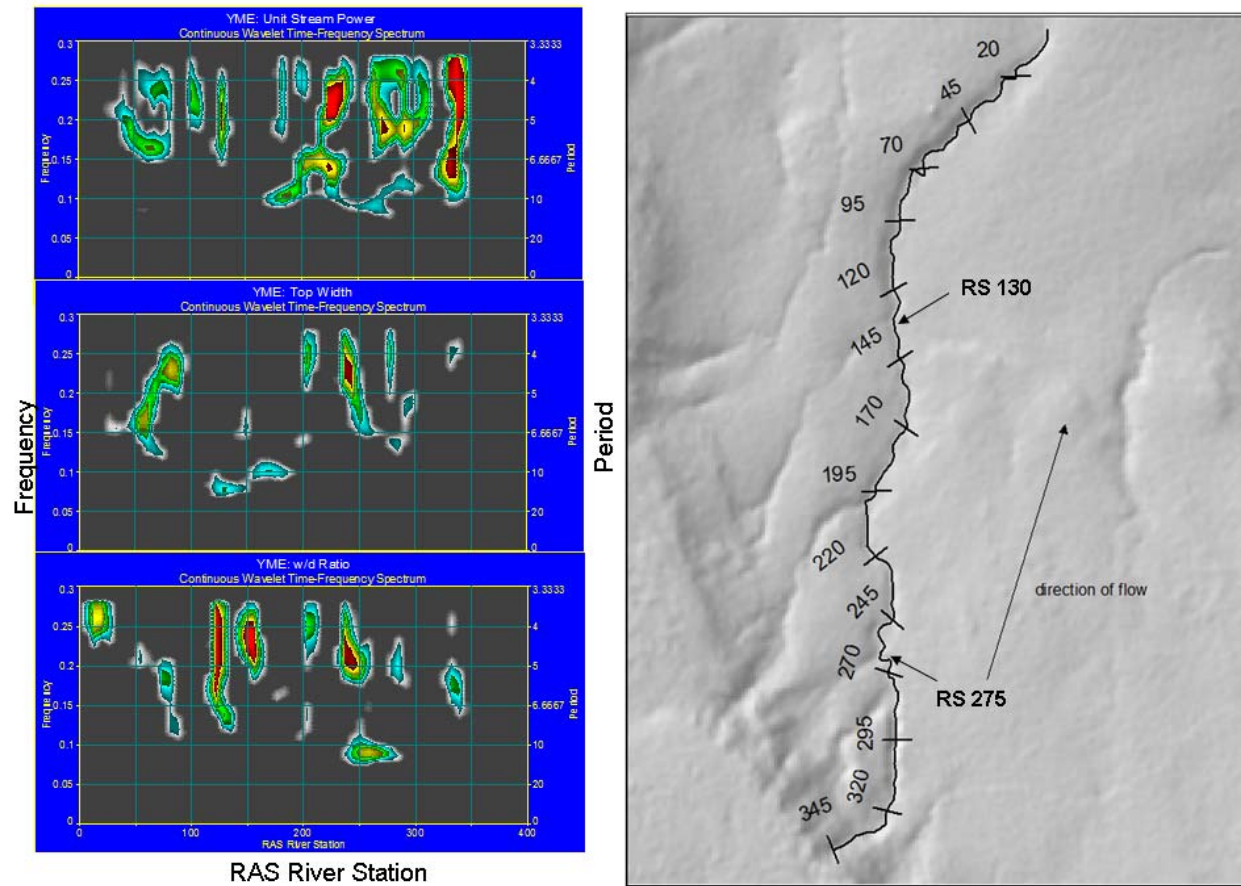


Figure 4.11: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, YME RAS model derived from last returns and total station survey data. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

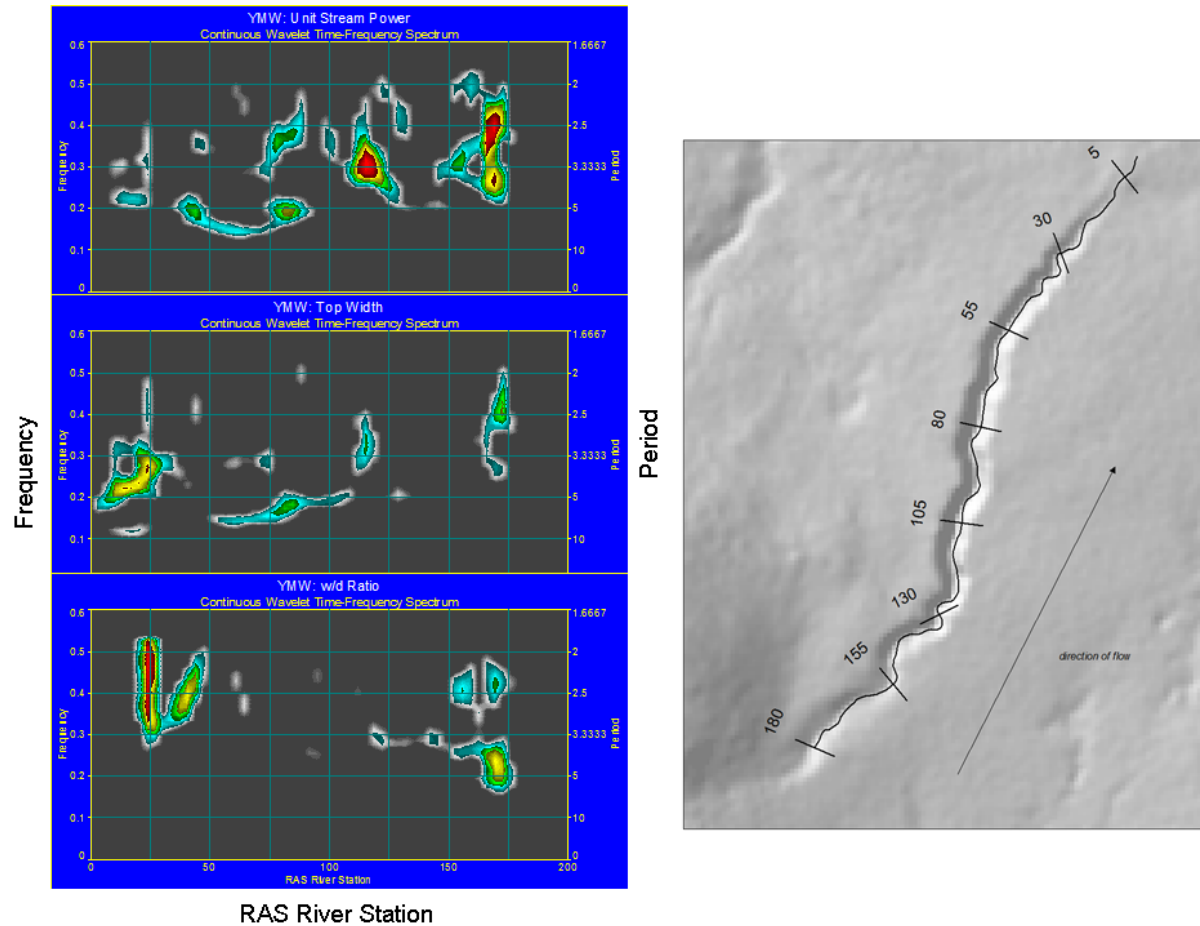


Figure 4.12: Power spectra of unit stream power, top wetted width, and width/depth ratio, and RAS stationed stream centerline, YMW RAS model derived from last returns and total station survey data. RAS river stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

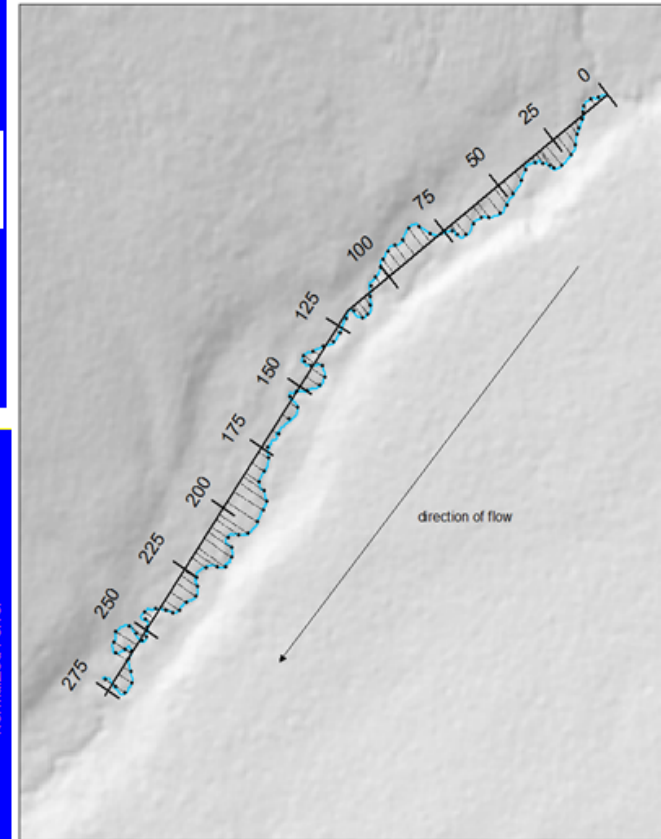
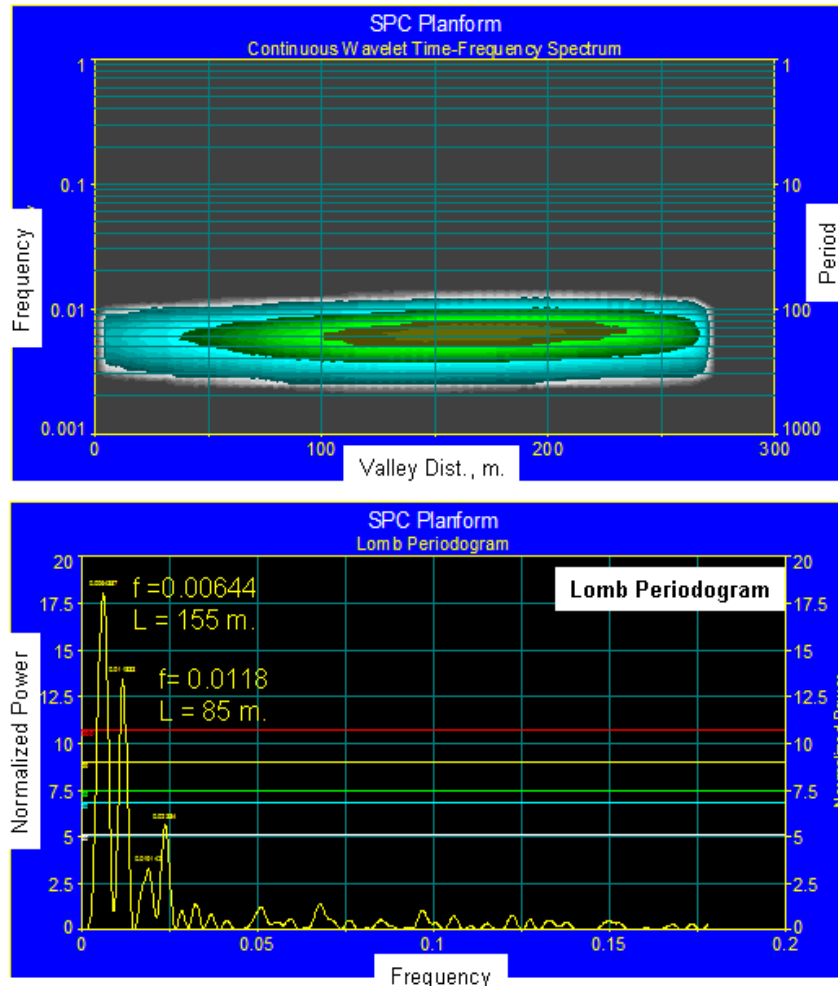


Figure 4.13: Power spectra of planform (deviations from valley centerline), SPC. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

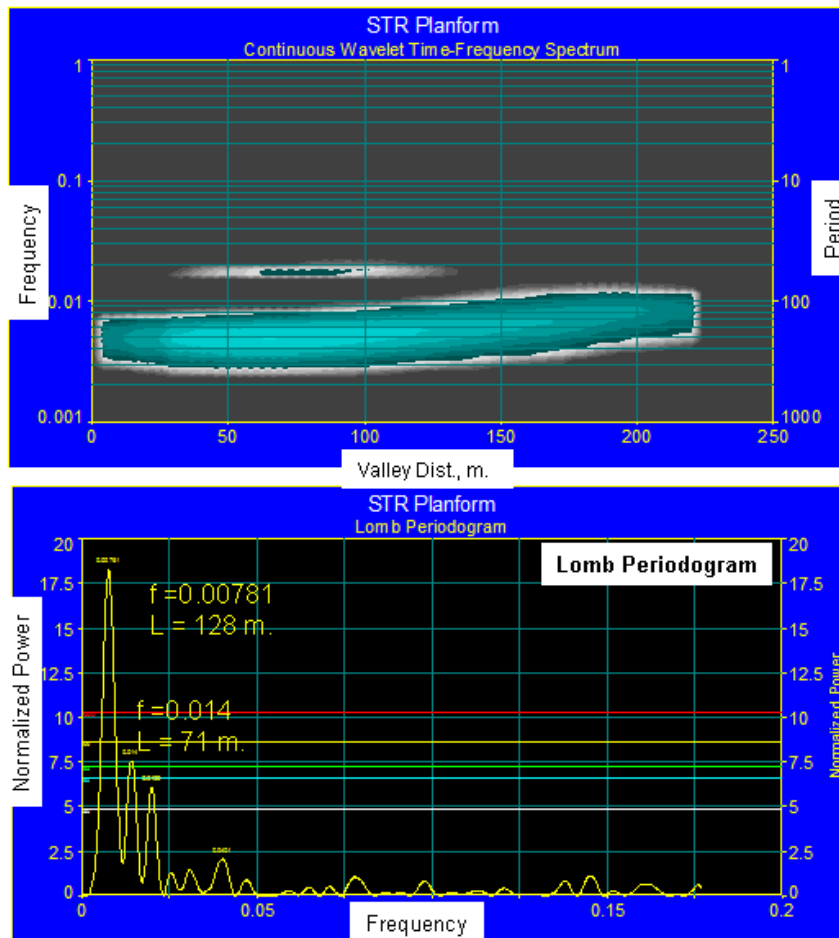


Figure 4.14: Power spectra of planform (deviations from valley centerline), STR. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

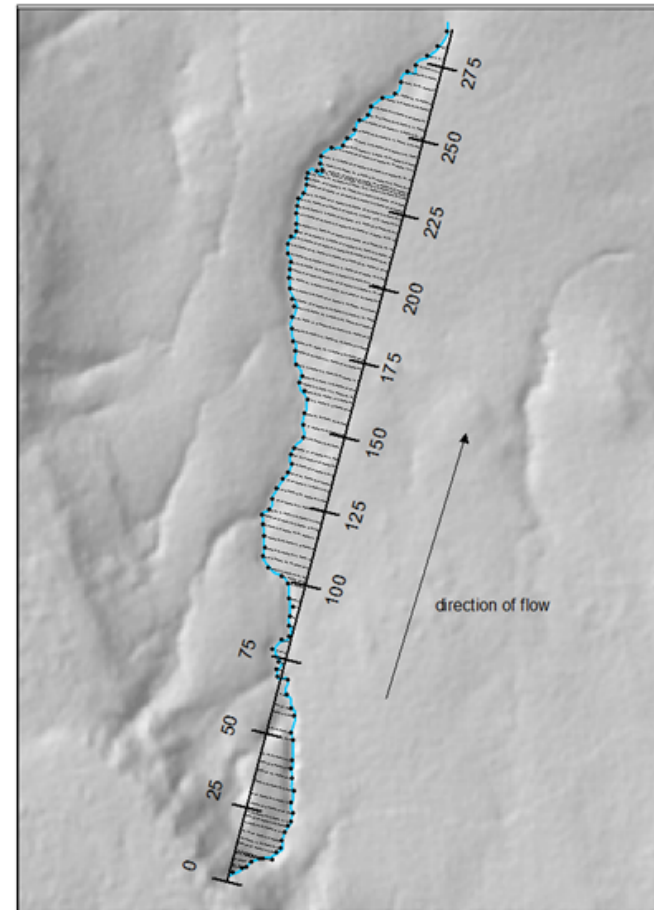
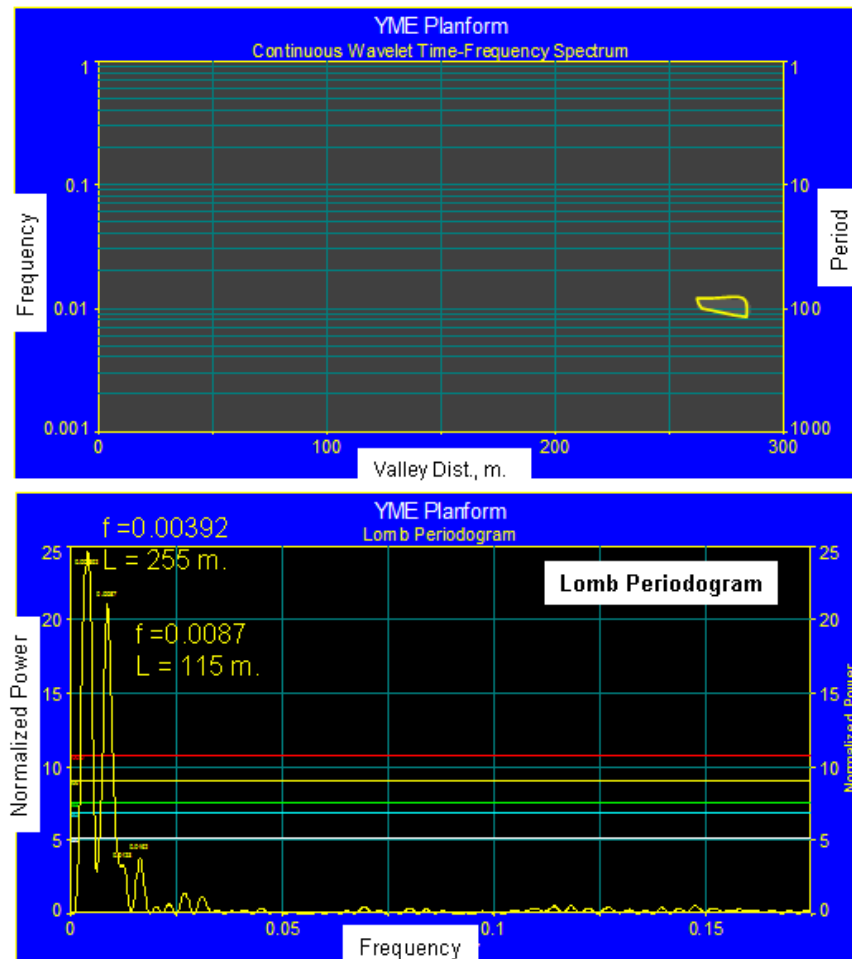


Figure 4.15: Power spectra of planform (deviations from valley centerline), YME. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

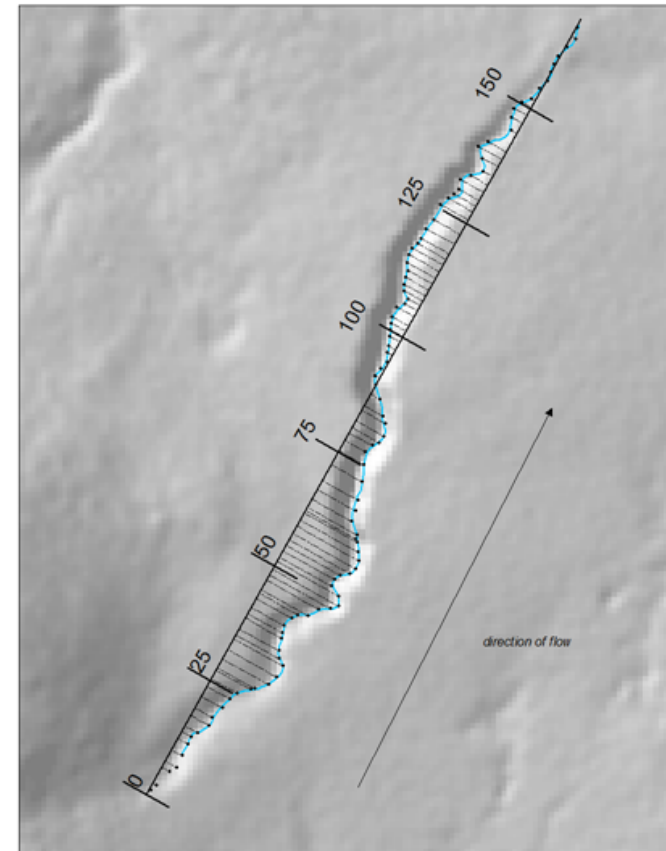
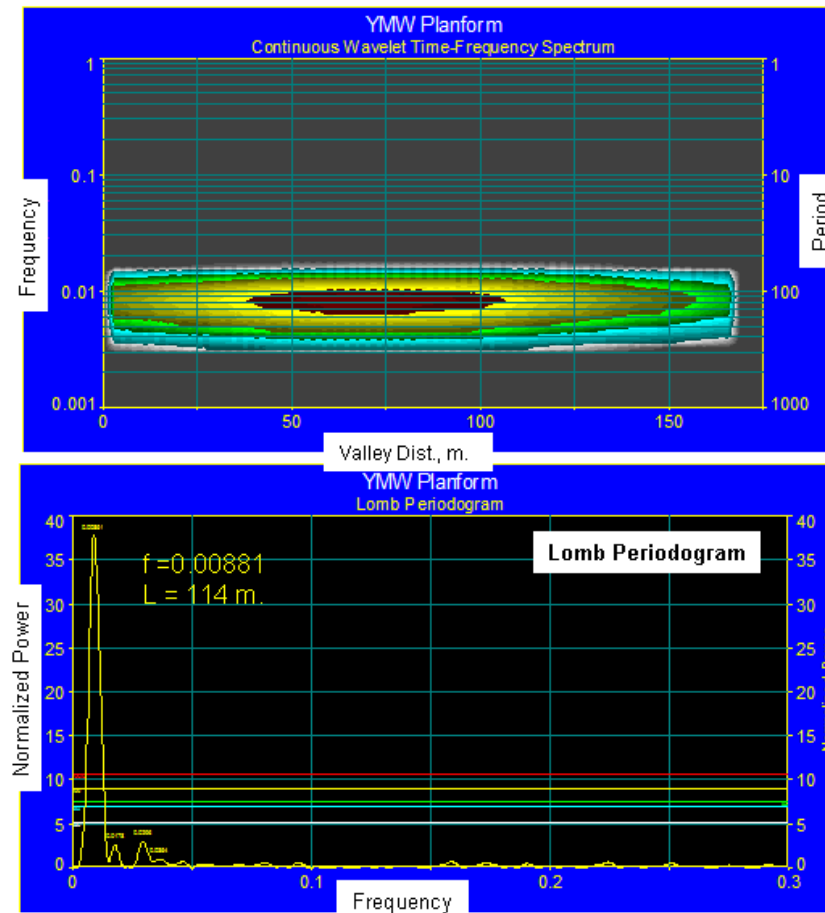


Figure 4.16: Power spectra of planform (deviations from valley centerline), YMW. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

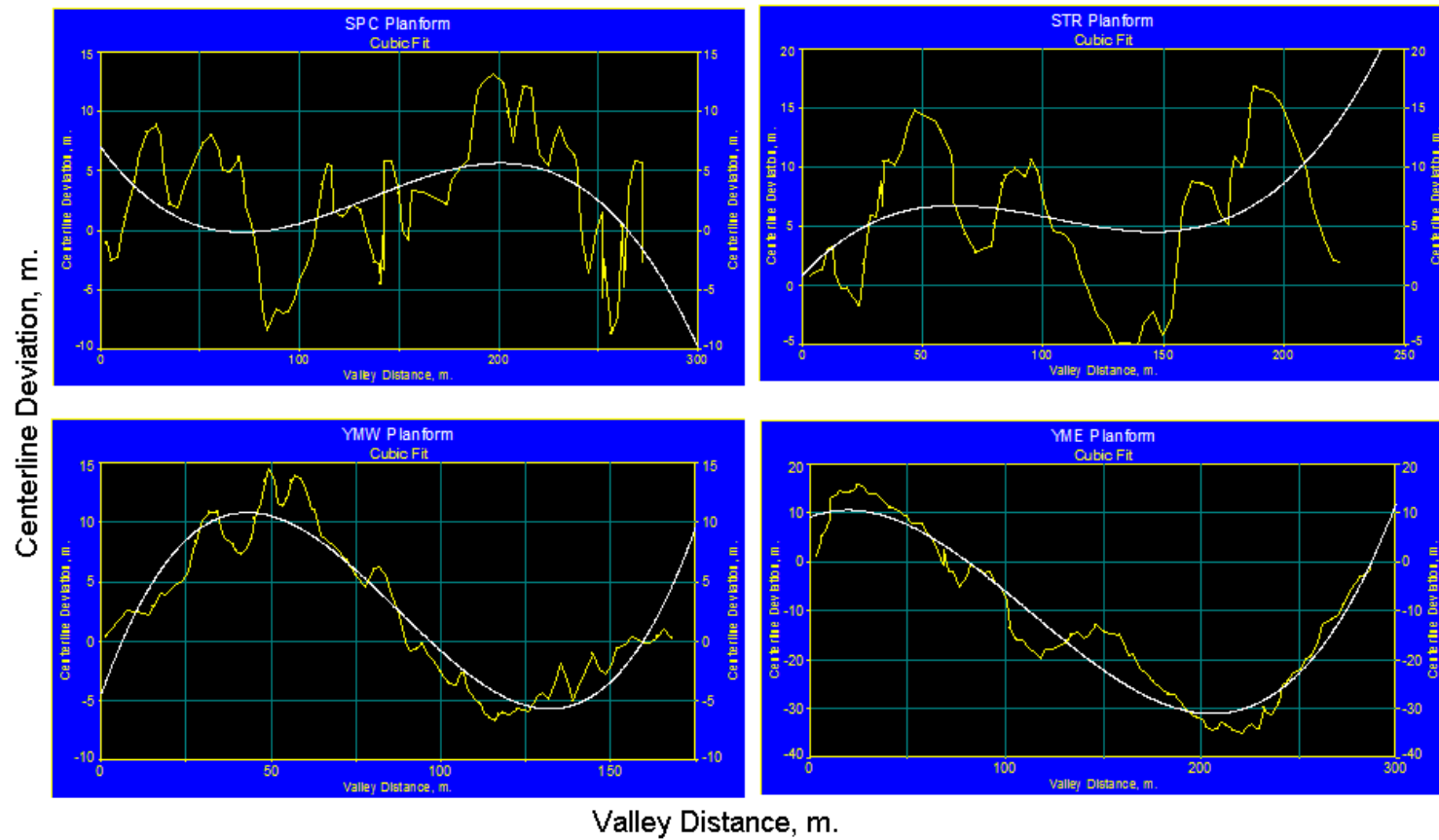


Figure 4.17: Planform series: centerline deviation versus valley distance, cubic fit overlaid.

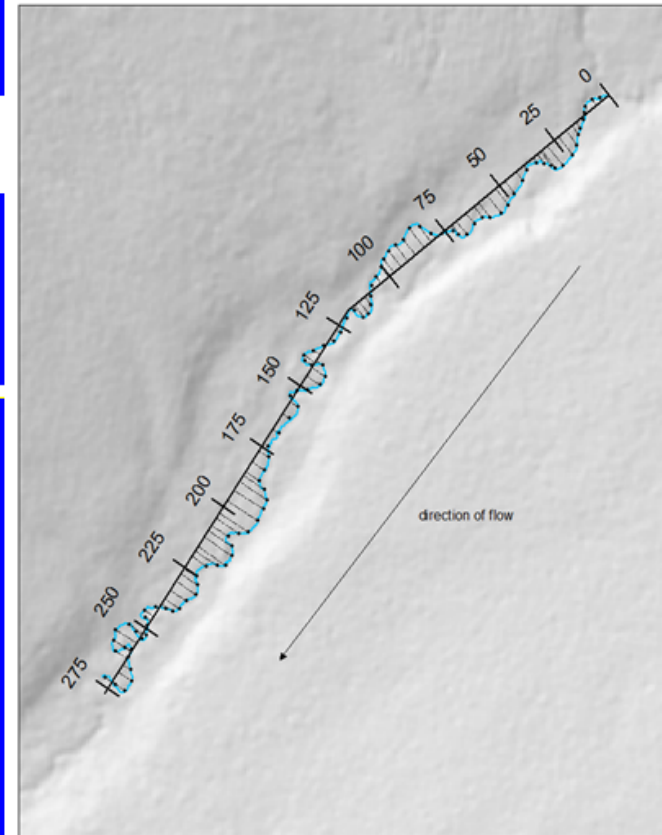
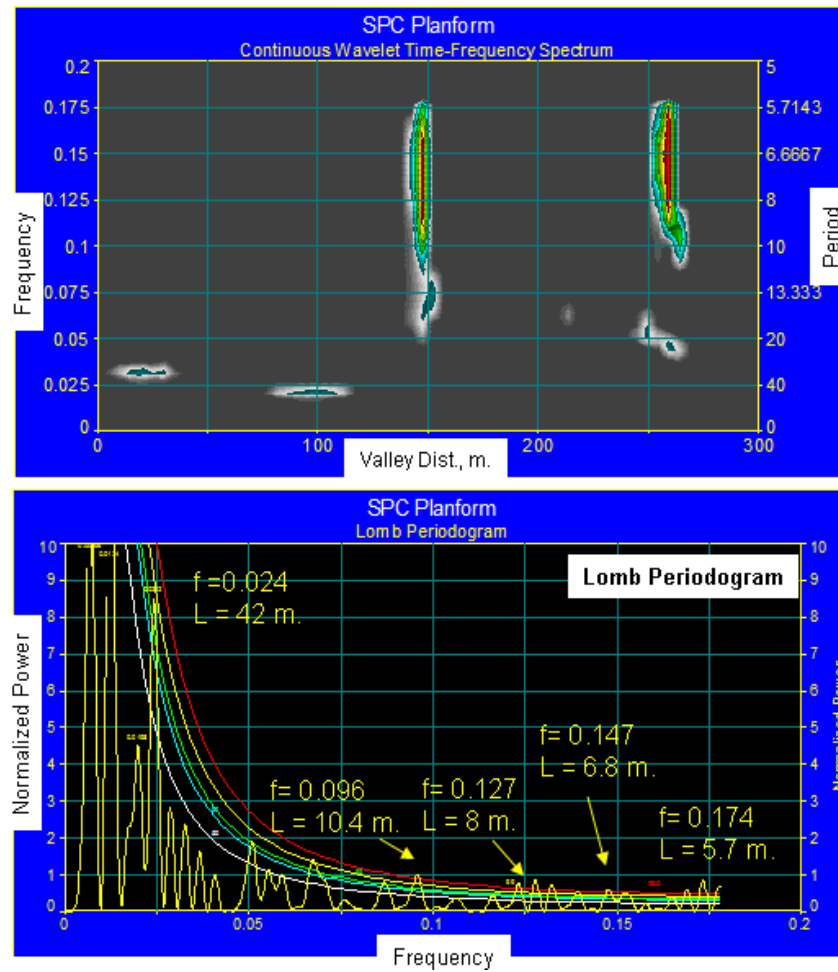


Figure 4.18: Power spectra of planform (deviations from valley centerline) after cubic detrending, SPC. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

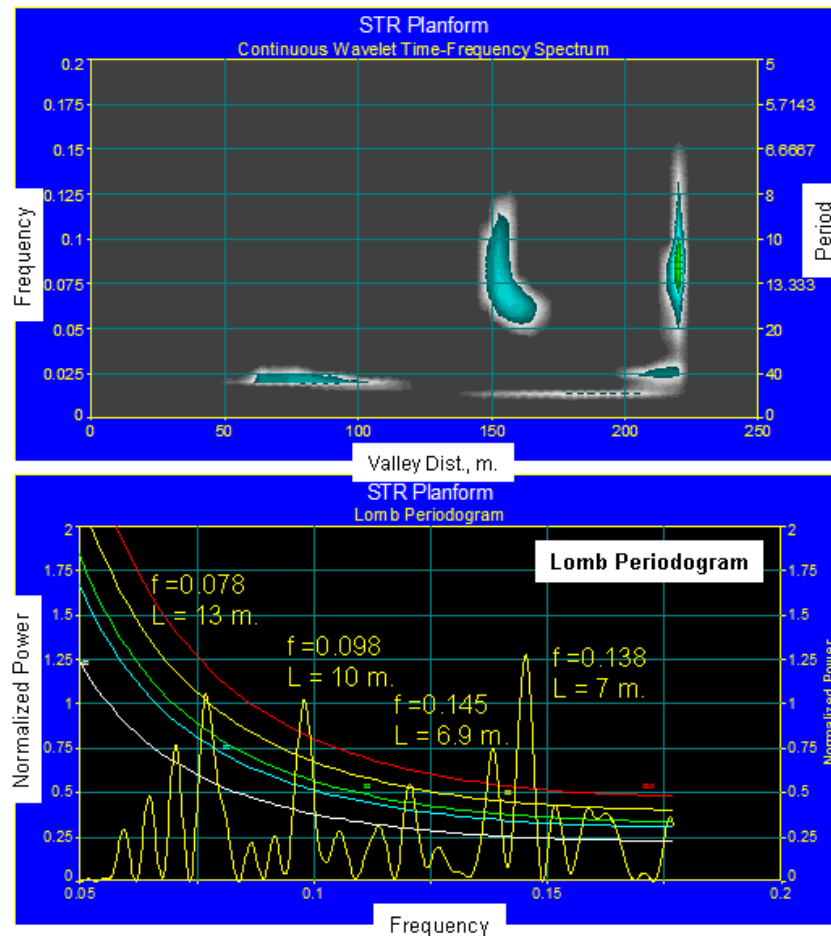


Figure 4.19: Power spectra of planform (deviations from valley centerline) after cubic detrending, STR. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

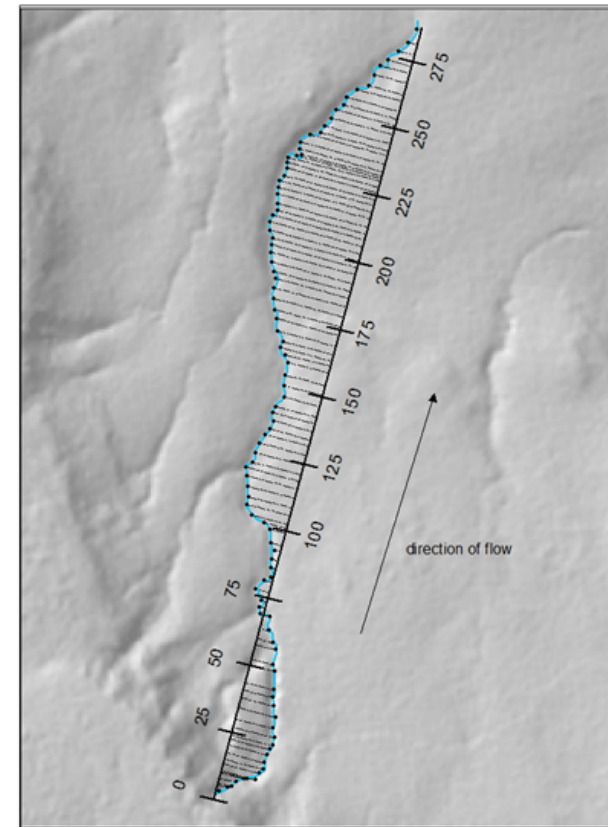
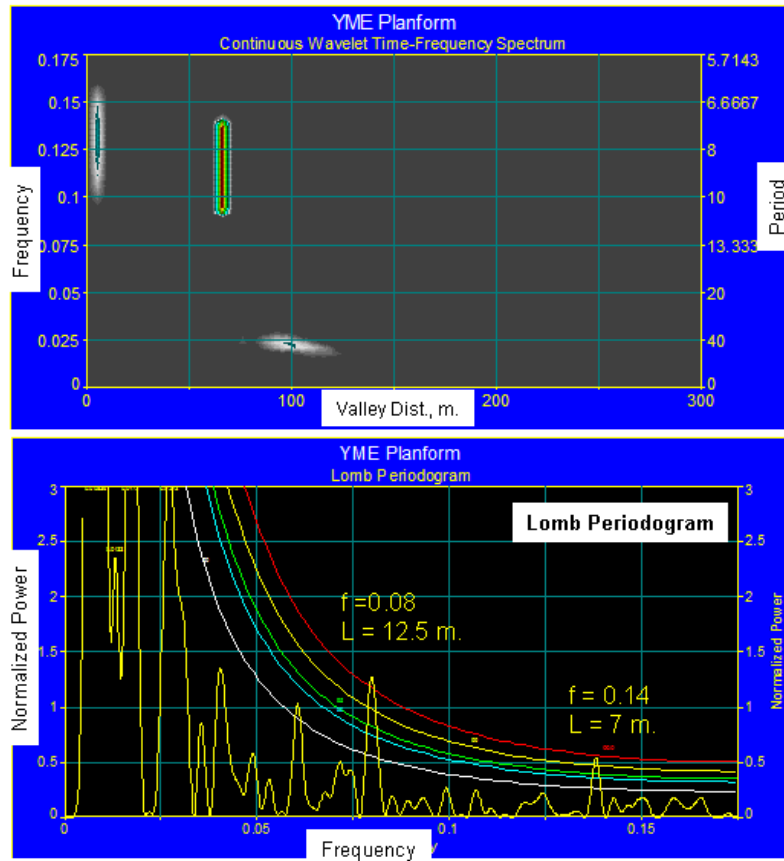


Figure 4.20: Power spectra of planform (deviations from valley centerline) after cubic detrending, YME. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

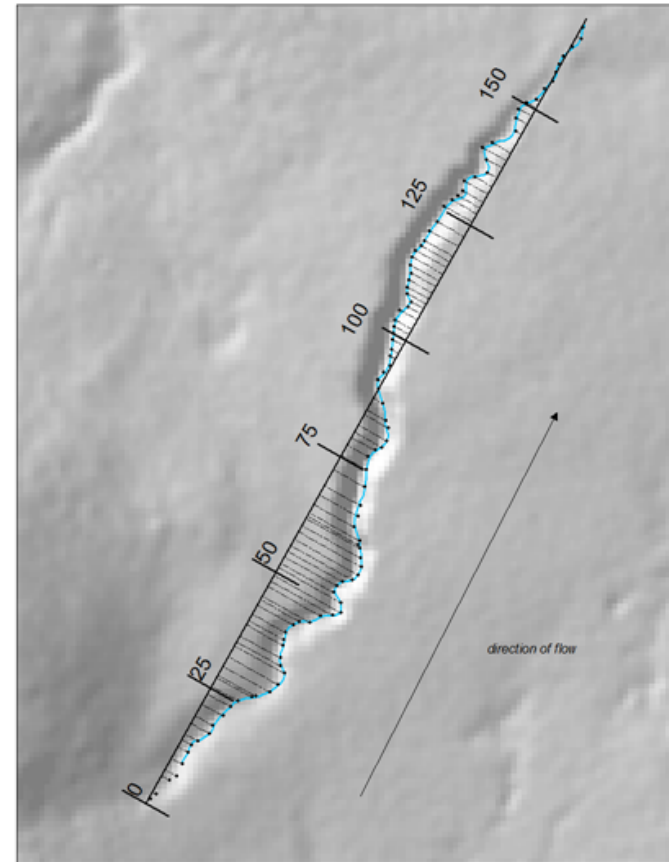
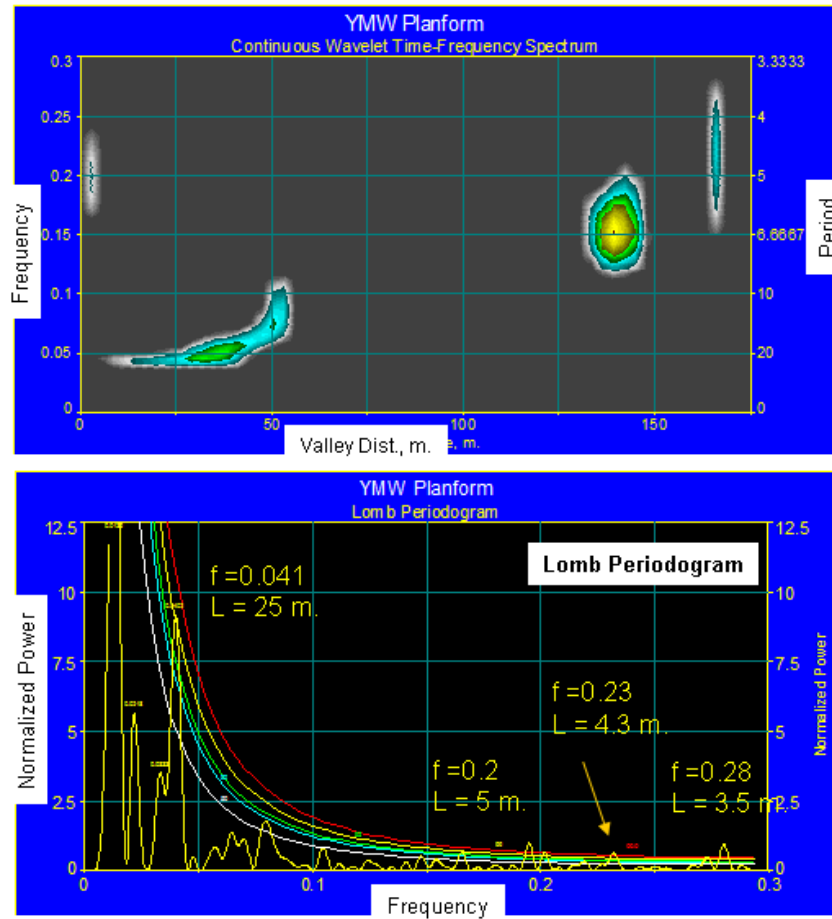


Figure 4.21: Power spectra of planform (deviations from valley centerline) after cubic detrending, YMW. Stationing in meters. Significance indicated by color. Red is significant to 99.9%, yellow is significant to 99%, green is significant to 95%, blue is significant to 90%.

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CHAPTER 5

SUMMARY

The objective of this dissertation was to investigate the transitions between process domains in headwater catchments, using topographic and fluvial signatures, in two pairs of headwater catchments in central and southwest Montana. Headwater catchments are described by multiple physiographic, hydrologic, and ecologic attributes which makes it difficult to delineate their spatial extent by either field observation or remote sensing. From a geomorphic standpoint, headwater catchments consist of an assemblage of landscape units including hillslopes, zero-order basins, discontinuous channels, and perennial first- and second-order channels. The form and arrangement of these landforms is subject to the locally dominant process domain, defined as the degree of interaction between hillslopes and channels, and the mode of sediment transport.

Headwater catchments have been extensively studied in the Pacific Northwest, where process domains are driven by debris flows and landslides, generated by the current geomorphic disturbance regime. In this dissertation, I have discovered important differences in controls between headwater catchments in the Pacific Northwest and two pairs of headwater catchments in the intermountain West. Drier climatic regimes of the intermountain West do not produce either the degree of weathering or vegetation communities found in the Pacific Northwest, or regions of maritime climate such as north Idaho and northwest Montana. Consequently, the studied headwater catchments are much more subject to lithologic and structural controls, unbuffered by thick regolith/soils

and stabilizing vegetation. Ongoing geomorphic processes are dominated by creep and snowmelt-driven fluvial erosion. Portions of the studied headwater catchments exhibit topographic signatures of debris flows and landslides but these are relict of prior disturbance regimes, though they provide the canvas upon which current processes operate. Here, channel initiation is more likely due to structure and the juxtaposition of lithologic units of differing permeability, rather than active mass wasting. Relative to the Pacific Northwest, the catchments are generally supply-limited due to the slow production of regolith in the high elevation, relatively dry climate of mountainous southwest and central Montana. However, between the catchment pairs, the Big Sky catchments can be characterized as supply-limited and the TCEF catchments as transport-limited. Although bedrock outcrops are found throughout TCEF, colluvium and regolith mantles many of the hollows. However, the current geomorphic disturbance regime does not commonly transport weathered material, and geologic contacts drive channel initiation. Some channel initiation is by threshold exceedence of the critical shear stress of the underlying substrate, but the thick colluvium-mantled hollows found in the Pacific Northwest and the Tennessee Valley of Marin County were not commonly observed in the study settings. Consequently, sediment produced by ongoing processes are stored within the channels and sediment produced by past process is either slowly reworked and transported by occasional competent flows or forms an immobile channel boundary and valley floor. Most of the colluvium has likely been produced in post-glacial time and largely remains in the headwaters flooring the valleys. Current colluvium production consists primarily in mechanical attrition of the accumulated lag.

In the Big Sky catchments, the ready visibility of sandstone lenses both in the field and in bare earth imagery highlights the lack of weathered material. In the present geomorphic regime, weathering products are slowly produced. The dominant lithology (sandstone and shale) dictates the size of the weathered products, which are readily transported by slope wash and channel flow. In this supply-limited environment, larger weathered clasts are relict from mass wasting of a more active, prior geomorphic regime. A conceptual model of the formation of the studied headwater catchments (Figure 5.1) emphasizes storage and reworking of colluvium produced in a past geomorphic regime as the dominant influence on present valley and channel morphology.

In this dissertation, I have integrated remote sensing data, statistical analyses, and scale-invariant properties of fluvial systems to characterize the process domains of two pairs of headwater catchments. In Chapter 2, "Hillslope and fluvial process domain transitions and scaling relations," I found that there is spatially significant transition zone between hillslope and fluvial process domains. This transition is best seen in scaling relations of flowpath length and contributing area (a modified Hack's law) and unit stream power versus contributing area. The hypothesized inflections in the modified Hack's law were observed in both catchment pairs though they differ in the length and steepness of the different process regions. This difference reflects hillslope length and how incipient channels confluence. How flowpath length changes with contributing area reflects how the hillslopes interact with the valley in terms of the degree of dissection of the catchment and number and length of flowpaths. The process zones predicted by the

modified Hack's law well predict the initiation of heads and hollows but do not distinguish discontinuous channels.

The hypothesized relation between unit stream power and contributing area also exhibited a well-defined transition zone between hillslope and fluvial domains. Mapped heads and hollows generally fall just before the onset of the hypothesized transitional zone. Process regions show most hollows within the transitional zone and most heads within the fluvial zone. The discontinuous nature of the Big Sky channels was also detected. This relation differs from the modified Hack's law in that it uses a proxy for geomorphic work, an intrinsic fluvial property, rather than a hillslope-driven length property. The flattening and then decline in unit stream power reflects a shift in process from slope-driven to fluvial-driven, as channels establish and increase width with increasing discharge. However, as the channels encounter greater resistance in lithologic type, unit stream power both increases and decreases, as shown in the statistical relations. This chapter shows that hollows and heads reflect different formative processes, but their initiation is nonetheless detectable by scaling relations of parameters that are intrinsic functions of the hillslope and fluvial process domains. Furthermore, the extent of their spatial regions can be mapped and point to the nature of local structural and lithologic controls.

In Chapter 3, "Detection of hollows and channel heads using spectral filtering of DEMs," I first tested the magnitude-frequency distributions of the topographic parameters: curvature and derivative of aspect for power law behavior, indicative of scale-invariant properties, and used values within the rollover region to detect the

hillslope to fluvial process domain transition zone. Additionally, I filtered DEMs of each catchment by high pass and wavelet filtering to remove non-channelized portions of the landscape to sharpen the rollover inflection. Values of curvature and derivative of aspect at the mapped heads and hollows are within an order of magnitude, regardless of the catchment. This may indicate a universal threshold exceedence needed to initiate fluvial features. Values of the derivative of aspect at mapped heads and hollows tend to fall within the power law portion of the magnitude-frequency relation whereas values of curvature tend to fall within the rollover region. The initiation of channels and valleys where statistical distributions of fluvial-related parameters exhibit power law behavior points to a scale-invariance commonly associated with channel network development. However, filtering of the DEMs was not necessarily an asset. Filtering tended to distort power law behavior though it did remove highly textured terrain from classification as part of the channel network. High pass filtering detected more hollows and heads than when an unfiltered DEM was used. Using inflections in the magnitude-frequency distributions to delineate process zones, channelized portions of the DEM were correctly identified though no distinction could be made between hollows and channels. These results indicate the spectral signature of channelized and convergent landforms can be detected and isolated so as to create a more accurate depiction of the network extent and topology. Though this method cannot be used as a watershed network delineator, it gives a coarse-scale estimate of the extent of the fluvial process domain. The characteristic spectra of hillslope, valley, and channeled landforms are used to delineate their spatial

distribution and suggest that transitions between landforms occur where spectral signal turns into random noise.

Chapters 2 and 3 examine different means of identifying process domains and their transitions in headwater catchments. The use of scaling relations between hillslope and fluvial parameters provides a better tessellation of the landscape by the type of flow on, and geomorphic work done by, each part of the landscape. The power law behavior of topographic parameters points to where scale-invariance assume dominance, but the statistical nature of the magnitude-frequency distributions does not permit deterministic prediction of process domains, but only likely locations of transitions.

In Chapter 4, " In-channel process domains: identification of colluvial to alluvial transitions using wavelet transforms", I extend the previous two chapters by examining the transition from colluvial to alluvial process in headwater channel reaches. I hypothesize that the transition from colluvial to alluvial process can be suggested by the presence of repeating signatures in the expenditure of energy and channel morphology in alluvial reaches and randomness in colluvial reaches. Repeating patterns, as well as power law behavior, is associated with the fractal structure and scale-invariance of the fluvial system. Using wavelet analysis of spatial series of unit stream power, width-to-depth ratio, and channel width, I found statistically significant spectral signatures of the form parameters either coincide or are proximal to statistically significant spectral signatures of unit stream power. Additionally, wavelet analysis of unit stream power tended to produce patterns with periods between 3 - 10 meters. Wavelet and periodogram analyses of channel planform shows patterns with similar periods.

Significant repeating cycles in unit stream power, channel morphology, and planform tended to be colocated, suggestive of dominant alluvial conditions in these subreaches. Signatures disappeared in the presence of known colluvial lag, intersection of hillslope toe, or crossing of bands of resistant lithology. Strength of the signatures was variable, generally hinting at the mixture of alluvial and colluvial conditions within the surveyed reaches. Generally, patterns were only detected for one to several wavelengths. Shorter wavelength patterns were detected at the upstream-most end of the surveyed reaches, perhaps associated with step-pool sequences, transitioning to longer wavelength features, perhaps associated with pool-riffle sequences. Patterns in channel form were less clear and may reflect dominant control by an immobile channel boundary or even the presence of random colluvial blocks. It is unlikely that the transition is clean from colluvial to alluvial, rather a noisy series of back-and-forth transitions before fully alluvial conditions are attained. Alluvial conditions are temporally, as well as spatially, variant depending on the probability distribution of disturbances. Clarity of results may also be limited the difficulty of producing an accurate hydraulic model from the integration of LIDAR last return points and total station survey data.

This dissertation adds to our understanding of how hillslope and fluvial process domains interact with each other within headwater catchments, what a mosaic of process domains looks like, and how their transitions and spatial extents can be detected by remote sensing. Finally, I examine the downstream end of headwater catchments: evaluating how colluvial and colluvial-influenced channels transition to alluvial conditions and test a means of detecting those transitions. Each chapter uses a different

approach to investigate process domain transitions. The subsequent tessellation of the headwater catchment varies from approach to approach, each revealing a different characteristic of the ongoing, land-forming process.

Given the results of this study, I suggest that headwater channels are the result of competition between hillslope and fluvial processes. Consequently, their topographic and geomorphic signature is a blurred composite of both processes. The physical signature of headwater channels is also dynamic in time, as they are subject to disturbances variable in frequency and magnitude. This signature may be indistinct such that it is best detected between end members of hillslope signature and fluvial signature. Collectively, these dissertation chapters suggest that studies of headwater channels should consider both hillslope and fluvial characterizations of the landscape to accurately define their spatial extent, and highlights that controls on channels of hillslope runoff generation, hillslope process, fluvial process, and past and present landscape disturbances are both amplified and tightly coupled in the headwater catchment setting.

Based upon the knowledge gained from my dissertation, I make the following recommendations for future research:

1. Consider past geomorphic regimes, lithology, and structure to be first-order controls on channel initiation and channel type in high elevation environments of the intermountain West unless empirically shown otherwise.
2. Use different spatial scales to characterize fluvial and hillslope process domains. Studies of fluvial process requires resolution at the scale of the channel, but hillslope landforms such as hollows may require a coarser scale

to define their spatial extent. Headwater channels should exhibit both scalings from their inception to their downstream end. The intrinsic ability of wavelets to apply multiple spatial scales may provide an efficient means of extracting significant spatial scales of landforms.

3. Current generation Lidar may require additional field topographic data to accurately model headwater channels. Until accurate submeter DEMs are available, alternatives to consider include re-classification of first through last return data using intensity to define the lowest in-channel surface.

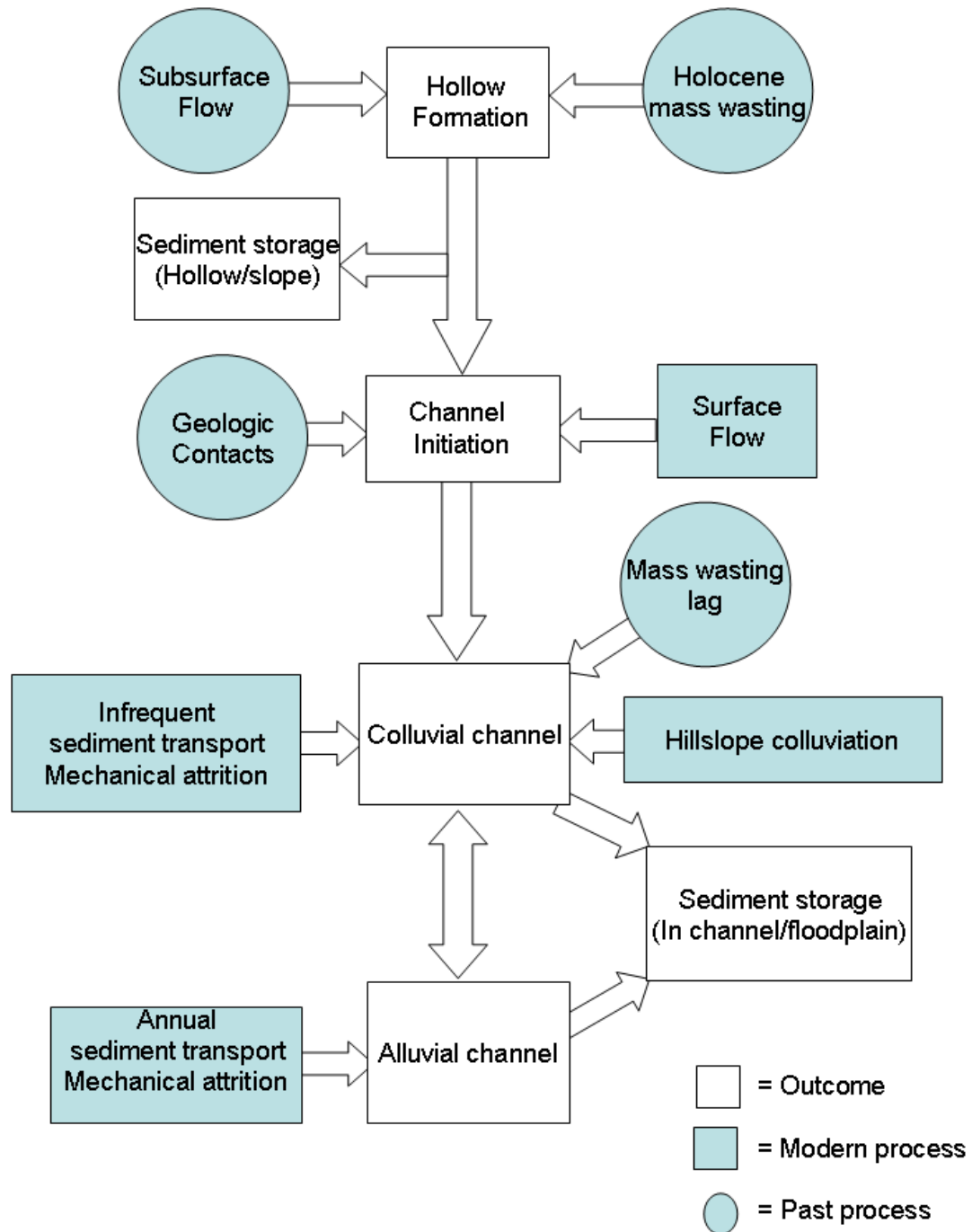


Figure 5.1: Conceptual model of the formation of the study headwater catchments, Big Sky and TCEF, Montana.

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