

FAULT SEGMENTATION CONTROL ON
ALLUVIAL FAN AND FAN DRAINAGE BASIN MORPHOMETRY,
LEMHI RANGE, EAST-CENTRAL IDAHO

by

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ABSTRACT

Alluvial fan and fan drainage basin morphometric parameters are proposed to vary as a function of distance from closest fault segment boundary along the Lemhi fault in east-central Idaho. Large normal faults are broken into segments along strike that have unique rupture histories. Since potential earthquake magnitude is related to rupture area and therefore segment rupture length, accurate delineation of segments can have implications for earthquake hazard assessment.

Alluvial fans were mapped on airphotos, and drainage basins were extracted from digital elevation models (DEMs). Morphometric parameters calculated for each drainage basin/fan pair based on DEM-extracted data include: elongation ratio (measures basin roundness in map view), Melton ratio (relates absolute relief and basin area), fan area to drainage basin area ratio, and hypsometric integral. Hypsometric integral is the area under the hypsometric curve, which displays relative elevation as a function of relative area. Standard regressions compared variations in these parameters with distance to fault segment boundaries mapped by previous workers.

Drainage basins closer to fault segment boundaries tend to have lower elongation ratios (more elongate shape), higher Melton ratios, higher fan area to drainage basin area ratios, higher hypsometric integrals, and straighter (less sinusoidal) hypsometric curves. No parameter is strongly correlated with distance to closest segment boundary, but the strongest correlation occurs with the Melton ratio. High Melton ratios have been associated with basins dominated by flows with higher sediment-to-water concentrations compared to basins with low Melton ratios. The observed along-strike morphometric variations can influence conceptual models of extensional footwall drainage development and hangingwall basin stratigraphic evolution. However, the relationships are not strong enough to inform seismic hazard or similar studies requiring a high degree of confidence and strong correlations.

In agreement with previous studies, fan area and drainage basin area are directly correlated, fan slope and drainage basin area are inversely correlated, and drainage basin slope and area are also inversely correlated. This study recognizes differences between slope calculated conventionally and slope calculated using standard grid-based methodology. This observation requires future studies to consider which slope definition is most representative of gravity-driven hydrogeomorphic processes.

INTRODUCTION

Morphometric variations in alluvial fans and their drainage basins along the extensional range front of the Lemhi Range in eastern Idaho (Figure 1) are proposed to be controlled by normal fault segmentation. Evolving geologic features such as alluvial fans and their catchments can be characterized by morphometric measurements, which in turn can help elucidate processes shaping the landforms. Active and ancient extensional basins are important because the sediments they contain provide information about paleogeographic, climatic, tectonic, base (lake/sea) level, ecologic, and earth surface process changes that took place during the period of deposition (Leeder and Jackson, 1993; Janecke, 1994; Bull, 2000; Allen and Allen, 2005; Leleu et al., 2005). In addition, basin-fills can contain hydrocarbon reservoirs, mineral and gravel deposits, or serve as aquifers or areas for waste disposal, while processes on active fans can pose a threat to human infrastructure and development (Warne, 1984; Cooke et al., 1993; Blair and McPherson, 1998; Blair, 1999; Gawthorpe and Leeder, 2000; Leeder and Mack, 2001; Gawthorpe and Hardy, 2002; Allen and Allen, 2005).

Weathering of bedrock to produce sediment that is subsequently removed and redistributed shapes a mountain range and surrounding valleys. Drainage basins develop as pathways for sediment evacuation and gravitationally-driven sediment transport from higher to lower elevations. Mountain-valley systems can thus be classified into regions defined by the dominant sedimentary regime – the erosional and depositional sectors of

the sediment routing system (Allen and Allen, 2005; Allen, 2008). Topography is the net result of tectonic processes that generate relief, erosional processes that destroy relief,

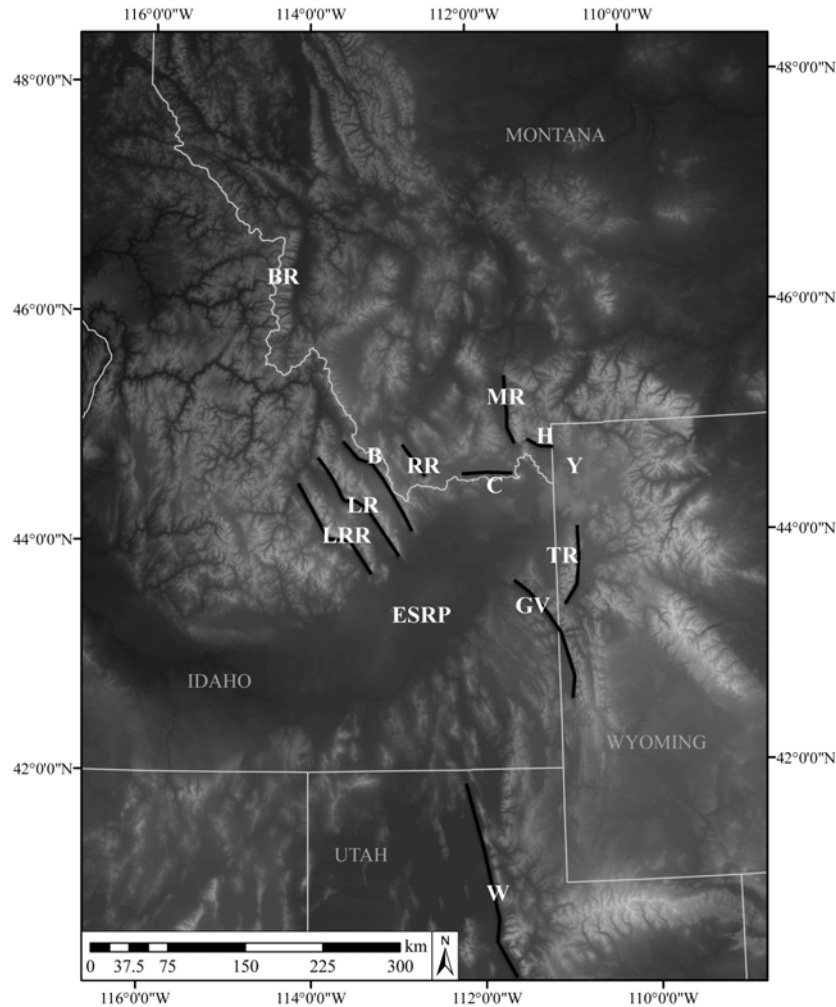


Figure 1. Elevation map of the western US with selected ranges, faults, and volcanic provinces labeled. Lighter colored areas are higher elevation and darker areas are lower elevation. Faults are generalized. B: Beaverhead Mountains, BR: Bitterroot Range, C: Centennial Range, ESRP: Eastern Snake River Plain; GV: Grand Valley fault, H: Hebgen fault, LR: Lemhi Range, LRR Lost River Range, MR: Madison Range, RR: Red Rock fault, TR: Teton Range, W: Wasatch Range, Y: Yellowstone. Elevation data from the US Geological Survey's GTOPO30 dataset (see Appendix A for data source information). All other elevation data are sourced from the US Geological Survey's National Elevation Dataset.

and depositional processes that reposition erosional products (Friedmann and Burbank, 1995; Hanks, 2000; Allen and Allen, 2005; Willett et al., 2006).

Numerical modeling and analog studies are two main methods of investigation of tectonic systems (Gawthorpe and Leeder, 2000). The case study presented in this research is an analog study that addresses quantification of variation of topographic parameters in an active extensional system.

Study of a modern, active sediment routing system provides input to conceptual models useful in interpreting the evolution of ancient mountain ranges, whose former presence exists now only as sedimentary packages deposited elsewhere (Leeder and Jackson, 1993; Leleu et al., 2005; Allen, 2008). The Lemhi Range in Idaho (Figure 1) presents an excellent opportunity to study a modern sedimentary routing system due to active uplift of the range and well-exposed alluvial fans with minimal vegetation or human modification.

Alluvial Fans and Tectonic Signals

Alluvial fans are located in proximal settings and are more sensitive repositories of sediment source dynamics and faulting activity than distal depositional environments (Blair and McPherson, 1994; Harvey et al., 2005; Pope and Wilkinson, 2005). In an active extensional basin, the uplifted footwall serves as the sediment source to fans of the developing hangingwall basin. The type of alluvial fans (debris flow or sheet flood) that develop is controlled by fan drainage basin size, relief, and bedrock geology (Blair and McPherson, 1998; Blair, 1999). These three sediment production variables are structurally controlled by segmentation of the active normal fault system and any

preexisting structural features in the extensional footwall (Leeder and Jackson, 1993; Janecke et al., 1999). Sediment supply to the basin is also controlled by climate, due to influences on weathering rates, erosion rates, vegetation biomass, and water balance (Pierce and Scott, 1982; Friedmann and Burbank, 1995; Blair and McPherson, 1998; Gawthorpe and Leeder, 2000). Within a set of climatic and bedrock lithology boundary conditions, annual water and sediment discharge is controlled by drainage basin area (Moore et al., 1991; Milana and Ruzicky, 1999; Gawthorpe and Leeder, 2000; Harvey et al., 2005).

Stream length is controlled by tectonic slope, which is created by extensional faulting (Leeder and Gawthorpe, 1987; Leeder and Jackson, 1993; Gawthorpe and Leeder, 2000). Portions of fault segments with higher throw will have steeper, larger drainage basins (Gawthorpe and Leeder, 2000). Catchments located at breached segment boundaries can also be anomalously large (Gawthorpe and Leeder, 2000). Streams incise into uplifting mountain blocks and erode headward during periods of lesser uplift (Turko, 1988). Thus, in a more tectonically active area, the drainage divide is closer to the fault front and the lengths of streams draining the range will be shorter. In a less active area, the drainage divide migrates away from the fault as the streams lengthen and cut headward (Turko, 1988; Leeder and Jackson, 1993). The change in slope from the steep footwall to the subsiding hangingwall results in the deposition of sediment – alluvial fans, talus cones, and alluvial fan deltas are deposited in terrestrial environments (Gawthorpe and Leeder, 2000). The shape and size of accommodation space available to these depositional systems in the hangingwall is influenced by along-strike changes in normal fault geometry (Shaw et al., 1997).

Alluvial fans along active mountain ranges may reflect fault slip rate – steeper surface slopes with rapid down-fan decrease in grain size indicate more rapid fault slip rate (Allen, 2008). Size, shape, drainage network, bedrock lithology, volumes and type of stored colluvium/sediment, vegetation, and other drainage basin characteristics are major factors controlling alluvial fan morphology and processes (Blair and McPherson, 1994; Blair and McPherson, 1998; Keefer, 1999). Along a single range front, changes to the climatic boundary conditions are relatively uniform, but fault-slip varies along strike. Climate-driven changes to sediment or water yield usually occur on all hillslopes in a watershed; aggradation and entrenchment events are commonly closely spaced in time for adjacent watersheds or even within a single region (Pierce and Scott, 1982; Ritter et al., 1995; Bull, 2000). This study focuses on a group of watersheds with similar aspect and glaciation histories, so along-strike climatic variations are considered minimal for any given time span. While precipitation data are minimal for the study area, vegetation patterns indicate relatively consistent high-elevation precipitation along strike (Johnson et al., 2007).

Models of Extensional Basin Evolution and Resulting Stratigraphy

Continental extension occurs in multiple settings, including: high-elevated extensional collapse regions of contractional orogens, back-arc regions where the trench is backstepping, long-lived intracratonic rifts, and areas of very high extension commonly associated with metamorphic core complex development (Horton and Schmitt, 1998). Several models have been proposed to describe the structural and stratigraphic evolution of different types of extensional basins.

Wernicke and Burchfiel (1982) interpreted uniformly tilted fault blocks to be bound by planar faults; listric fault geometry at depth was interpreted for imbricate faults with different tilt between blocks and a high degree of rotation of beds and faults. Differential tilt between strata in the hangingwall and footwall of a normal fault was interpreted as evidence of listric geometry of the fault (Wernicke and Burchfiel, 1982). Planar faults were described as capable of rotating beds and faults, but mainly were associated with no rotation. Rotational faulting produces growth strata that display an increase in dip with an increase in age (Wernicke and Burchfiel, 1982).

Leeder and Gawthorpe (1987) proposed four facies models for extensional basin sedimentary and stratigraphy development covering environments ranging from continental basins with interior drainage or axial through-drainage to coastal environments along marine gulfs or shelf basins with carbonate facies. The footwall may be more the more dominant source volumetrically, but hangingwall-sourced sediment will be more spatially extensive. Asymmetric subsidence related to the hangingwall tilting toward the fault controls facies migration perpendicular to fault strike (Leeder and Gawthorpe, 1987). In the continental facies models, axial or basin center (playa or lake) depocenters will tend to be located proximal to the fault, where the maximum subsidence occurs (Leeder and Gawthorpe, 1987).

In another model of intracontinental extensional basin development, two end-member types are proposed: rift basins and supradetachment basins (Friedmann and Burbank, 1995). Rift basins have slower slip rates, lower total extension, steeper faults and different composition of magmatic products compared to supradetachment basins (Table 1). Additionally, development of drainage networks and resulting stratigraphy

varies between the end-members. While supradetachment basins tend to have long, transverse, footwall-sourced drainages yielding thin basin fills with rapid sedimentation rates and distal depocenters, rift basins tend to be dominated by axial or hangingwall-sourced drainages creating thick basin fills proximal to the bounding fault (Friedmann and Burbank, 1995).

Table 1. Rift basin and supradetachment basin characteristics, two end-members of intracratonic extensional basins. Based on data in Friedmann and Burbank (1995).

Intracontinental Extensional Basin End-Member Characteristics		
	Rift basins	Supradetachment basins
Tectonic setting	Crust of normal thickness; thermally equilibrated	Recently deformed and overthickened crust; back-arc setting
Magmatism	Alkali to tholeiitic	Calc-alkalic
Dip of bounding faults	45-60°	15-35°
Slip rates	<1 mm yr ⁻¹	>2 mm yr ⁻¹
Extension magnitude	10-25%	Often >100%
Duration of extension	Typically >25 Myr	<10 Myr

A detailed stratigraphic and structural study of a supradetachment basin in southwest Montana, the Eocene-Oligocene Muddy Creek half-graben (Figure 2), yielded facies patterns and basin architecture results inconsistent with the predictions of Friedmann and Burbank (1995) for a supradetachment basin (Janecke et al., 1999). Instead, the deposits more closely fit the half graben/rift basin model of Leeder and Gawthorpe (1987), despite the low angle of the bounding fault system of the Muddy Creek basin (Janecke et al., 1999). Further refinement of models of the stratigraphic and

structural evolution of extensional basins is needed to resolve which models best describe reality (Janecke et al., 1999).

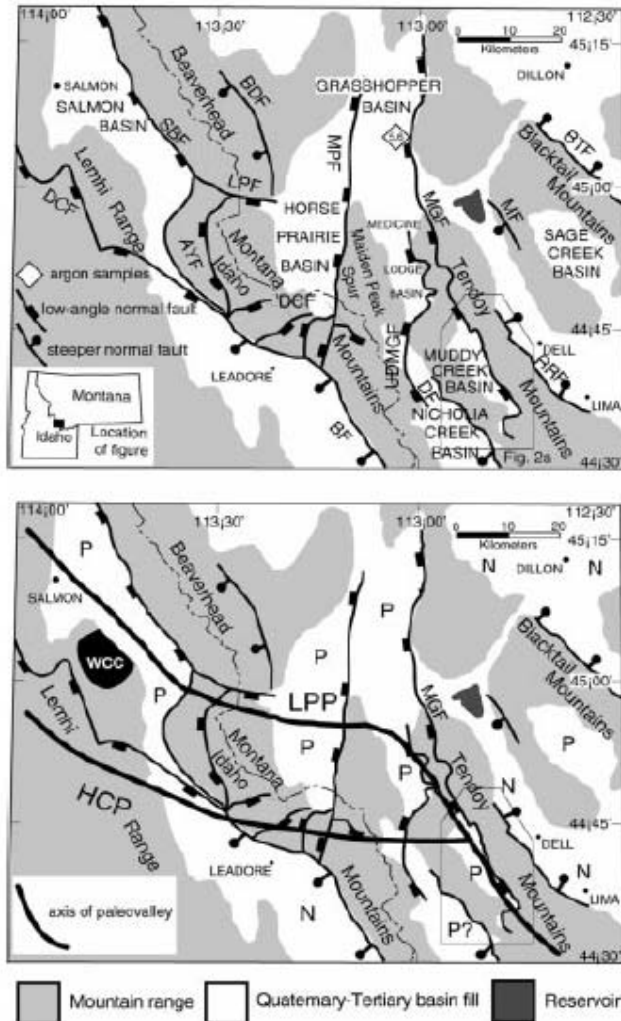


Figure 2. Above, large normal faults and Paleogene basins close to the modern Lemhi Range. AYF: Agency-Yearian fault, BF: Beaverhead fault, BTF: Blacktail fault, DCF: Divide Creek fault, DF: Deadman fault, LPF: Lemhi Pass fault, MGF: Muddy-Grasshopper detachment fault, MPF: Maiden Peak fault, RRF: Red Rock fault, UDMGF: upper detachment fault of the Muddy-Grasshopper detachment fault system. Paleogene basins in this region include the Salmon basin, Horse Prairie basin, Nicholia Creek basin, Sage Creek basin, Grasshopper basin, Medicine Lodge basin and Muddy Creek basin. Below, Middle Eocene paleovalleys. HCP: Hawley Creek paleovalley, LPP: Lemhi Pass paleovalley, MGF: Muddy-Grasshopper detachment fault, N: predominately Neogene-Quaternary basin fill, P: predominately Paleogene basin fill, WCC: Withington Creek caldera. This figure from Janecke (1999, p. 145).

Extensional tilt-block ranges of homogeneous lithology typically are asymmetric, with shorter, steeper drainages on the footwall side of the main drainage divide, and longer, more gently sloping drainages on the hangingwall side (Leeder and Jackson, 1993). Uplift of an extensional footwall is isostatically driven, thus local uplift can be at an angle to the regional extension direction because negative loading of the footwall by removal of the hangingwall load is controlled by the strike of the individual fault. Individual faults may or may not be aligned in exact agreement with the regional extension direction (Wernicke and Axen, 1988).

Segmentation of Normal Faults

All major (>40 km long; Crone and Haller, 1991) active normal faults are segmented along strike. Fault zones that break the seismogenic crust typically consist of 20-25 km long segments (Gawthorpe and Leeder, 2000; Keller and Pinter, 2002). Workers describe segments as short as 10-20 km (Leeder and Jackson, 1993) or as long as 40-70 km (Machette et al., 1991); longer segments tend to be identified on longer faults. Fault zones with earthquakes nucleating at depths of 10-15 km or more are considered to be those that break the seismogenic crust (Leeder and Jackson, 1993). Fault length and displacement are thought to be related in that $D = cL^n$ where D is maximum displacement, c is a constant reflecting rock properties, L is maximum fault trace length, and n ranges from 1-2 (Gawthorpe and Leeder, 2000). In a fault zone, the D - L scaling is consistent between the entire zone and single segments within the zone. Neighboring faults can affect each other's displacement profiles, creating asymmetric profiles (Gawthorpe and Leeder, 2000).

Topographically, a segment boundary can sometimes be recognized by hangingwall highs and footwall lows (Crone et al., 1987; Gawthorpe and Leeder, 2000); however, relative relief may be insufficient to identify which segments have the highest likelihood of future surface-faulting seismicity (Smith and Arabasz, 1991). Areas within segment boundary zones may experience less frequent rupture and less net slip than the areas located in the interior of adjacent segments (Crone et al., 1987). If segment boundaries are persistent along a fault it may cause morphologic differences between mountain front segments (Crone et al., 1987; Turko, 1988; Hemphill-Haley et al., 2000). Along-strike variability in the basin bounding fault system affects three-dimensional basin geometry (Janecke et al., 1999), as well as the drainage basins draining the footwall (Leeder and Jackson, 1993). Higher uplift rates generate sharper, straighter mountain/piedmont junctions with steep, deep, narrow stream valleys. When denudation rates are higher than uplift rates, stream valley floors widen and the mountain/piedmont junction becomes more sinuous (Rockwell et al., 1985; Piety et al., 1992; Keller and Pinter, 2002).

Structurally, a segment boundary is not a point on the earth's surface, but rather a complicated zone of intersecting faults compared to the more simple linear trace of the interior of a segment (Janecke, 1993), or a lateral ramp between *en echelon* neighboring segments (Crone and Haller, 1991; Leeder and Jackson, 1993). Before a normal fault breaks the surface, fault-propagation folding will occur above the fault, manifested as a triangular zone of deformation extending upward and outward from the subsurface fault tip (Gawthorpe and Hardy, 2002). Thus, where the normal fault has not broken the surface, a monoclinial fold will develop (Gawthorpe and Hardy, 2002); relay ramps

between *en echelon* fault segments may be blind transfer faults expressed as monoclines at the surface (Leeder and Gawthorpe, 1987). Segment boundaries are often, but not always definitively, controlled by transverse geologic structures that allow segments on either side to behave somewhat independently (Haller, 1988; Turko, 1988; Crone and Haller, 1991; Turko and Knuepfer, 1991; Piety et al., 1992; Janecke, 1993; Gorton, 1995). These structures can be bedrock highs or range-front bends where cumulative displacement, geometry, complexity or roughness of a fault changes; or cross faults (inherited from earlier deformation events) at a high angle to the main range-bounding fault that may accommodate some of the offset during a faulting event on the younger range-bounding normal fault (Machette et al., 1991; Janecke, 1993; Gorton, 1995; Petrik, 2008). Sometimes gravity saddles are associated with segment boundary zones, indicating locations with subsurface hangingwall highs or bedrock ridges (Crone and Haller, 1991; Machette et al., 1991).

Seismically, segment boundaries are typically barriers to fault rupture, and often have an increased density of faults with small displacement (Crone et al., 1987; Janecke, 1993; Gawthorpe and Leeder, 2000). Surface displacement (as measured by scarp offsets) increases towards the midpoint of rupture for a given event, and decreases towards the edges of rupture (Gorton, 1995). “Leaky” segment boundaries allow some rupture to occur on the neighboring segment during a seismic event (Crone and Haller, 1991). A segment boundary may stop rupture during the main seismic event, while aftershocks may affect other segments to a lesser degree (Crone et al., 1987). Neighboring segments have differing style and earthquake histories, although very large earthquakes can rupture multiple segments during a single event (Crone et al., 1987; Crone and Haller, 1991;

Piety et al., 1992; Keller and Pinter, 2002; Harkins et al., 2005). Segment boundaries may persist through several seismogenic periods in a fault's history (Haller, 1988).

Detailed segmentation studies in Idaho, Montana, Wyoming and Utah have focused on several large normal faults (Figure 1) in the region including the Lost River fault (Scott et al., 1985; Crone et al., 1987; Crone and Haller, 1991), Lemhi fault (Haller, 1988; Turko, 1988; Baltzer, 1990; Crone and Haller, 1991; Turko and Knuepfer, 1991; Janecke, 1993; Gorton, 1995; Hemphill-Haley, 2000), Beaverhead fault (Haller, 1988; Crone and Haller, 1991), Red Rock and Monument Hill faults (Haller, 1988; Crone and Haller, 1991; Harkins et al., 2005; Regalla et al., 2007), Centennial fault (Petrik, 2008), Grand Valley fault (Piety et al., 1992), and Wasatch fault (Machette et al., 1991).

Tectonically Driven Evolution of the Extensional Footwall as a Sediment Source

Tectonic activity affects the ratio of fan area to catchment area (Leeder and Jackson, 1993; Allen and Densmore, 2000; Allen and Allen, 2005). Sediment transport events affect only portions of a fan – thus the fan surface is an amalgamation of many events over the course of time (Blair and McPherson, 1994; Blair and McPherson, 1998). The ratio of fan area to catchment area, termed ϕ (Allen and Hovius, 1998; Allen and Allen, 2005), reflects the sum of these events, and thus scales on tectonic rather than sediment-transport-event parameters (Allen and Densmore, 2000). The ratio of fan area to catchment area is inversely related to fault slip rate in extensional settings (Ferrill et al., 1996; Allen and Densmore, 2000). An increase in fault slip rate causes increased uplift of the footwall, encouraging headward erosion of the drainage network and

potential piracy of adjacent drainage areas (Leeder and Jackson, 1993; Allen and Densmore, 2000). Meanwhile, the increase in accommodation space (due to downdropping of the hangingwall) ensures that even as fan volume increases, fan plan area remains small. Thus, all other factors being held constant, for a given catchment area, fan area will be smaller when fault slip rates are higher (Allen and Densmore, 2000). A higher fault angle results in greater vertical offset (more accommodation space is created) for a given amount of horizontal extension when compared to a lower fault angle (Ferrill et al., 1996).

At shorter time scales (10^2 – 10^4 yr), climatic elements (i.e. increased sediment and water discharge from the catchment resulting from a change in climate) determine fan geometry (Allen and Densmore, 2000; Harvey et al., 2005). At longer time scales ($>10^4$ yr), fault slip rate and fault geometry are the determining factors (Leeder and Jackson, 1993; Allen and Densmore, 2000). Tectonic activity and drainage basin evolution act on similar time scales to affect changes in fan depositional facies and stacking patterns (Harvey et al., 2005; Leleu et al., 2005).

Large transverse drainages can have a variety of origins in orogenic zones (Oberlander, 1965; Leeder and Jackson, 1993; Mather and Hartley, 2006). The traditional conceptual model of extensional footwall drainage development is that as an extensional footwall system develops from a series of *en echelon* normal fault segments, large transverse drainages develop sourced in the footwall relay zone or fractured segment boundary zones while smaller drainages (and smaller fans) develop along the main border faults (Gawthorpe and Leeder, 1987; Leeder and Jackson, 1993; Horton and Schmitt, 1998; Allen and Allen, 2005). However, numerical landscape evolution modeling has not

supported this hypothesis (Densmore et al., 2003; Allen and Allen, 2005); instead, large antecedent drainages coincidentally located at relay zones may be responsible rather than local fault geometries (Allen and Allen, 2005).

Alluvial Fan and Drainage Basin Area and Slope Relationships

Power-law functions have been applied to describe the relationships between alluvial fan area (A_f), fan drainage basin area (A_b), fan slope (S_f), and drainage basin slope (S_b). These can be expressed as:

$$A_f = uA_b^v \quad (1)$$

$$S_f = wA_b^x \quad (2)$$

$$S_b = yA_b^z \quad (3)$$

where u , v , w , and y are positive constants; and x and z are usually negative constants (Hooke, 1968; Funk, 1977; Rockwell et al., 1985; Cooke et al., 1993; Allen and Hovius, 1998; Mack and Leeder, 1999; Milana and Ruzycki, 1999; Al-Farraj and Harvey, 2005; Harvey, 2005; Leleu et al., 2005; Lin et al., 2009). In arid regions, v is often about 0.7-1.1, while u ranges more widely from about 0.1 to 2.1, reflecting catchment lithology (Al-Farraj and Harvey, 2005). Equation 2 is often a weaker relationship, but x usually ranges between about -0.15 and -0.35, and w ranges from about 0.03 to 0.17 (Al-Farraj and Harvey, 2005). Climate may influence w , with greater values of w associated with more arid regions (Milana and Ruzycki, 1999).

Catchment lithology (relative erodability of exposed units), tectonic tilting, precipitation, and available accommodation space all influence fan area (Rockwell et al., 1985; Cooke et al., 1993; Mack and Leeder, 1999); but the primary factors are often

simply described by sediment and water discharge from the catchment (Leeder and Jackson, 1993; Milana and Ruzyski, 1999; Al-Farraj and Harvey, 2005). Fan slope is related to water discharge, sediment supply, grain size, and base-level influences (Al-Farraj and Harvey, 2005). As the ratio of sediment supply to water discharge increases, slope increases. Fan slope also increases with increasing grain size (Rockwell et al., 1985; Blair and McPherson, 1994; Blair and McPherson, 1998).

Variations from the usual inverse slope-area correlations (positive x and z instead of negative) have been attributed to conditions where large basins with high uplift and erosion rates yield large sediment and water discharge with little long-term sediment storage in the basins (e.g. case study: uplifting Taiwan Central Range in the monsoon zone, Lin et al., 2009). Variations in these relationships can also occur due to reworking of fan deposits. Alluvial fans are considered “space-sharing systems” (Cooke et al., 1993) in that deposition on one fan affects neighboring depositional systems. If one fan increases in volume, whether by vertical aggradation, lateral avulsion and deposition, or increased deposition basinward, neighboring depositional systems (alluvial fan, basin-floor playa, and/or axial fluvial systems) will be affected (Leeder and Gawthorpe, 1987). Neighboring depositional systems can also limit the available space for alluvial fan deposition by actively eroding fan deposits (Mack and Leeder, 1999; Milana and Ruzyski, 1999; Leeder and Mack, 2001; Harvey, 2005).

Basin Elongation Ratio, Hypsometry, and Melton Ratio

Elongation ratio is a measure of basin shape and roundness (Singh and Jain, 2009); it can be calculated as the ratio between the diameter of a circle with identical area

as the basin and the length of the basin. Higher elongation ratios indicate rounder basins, while lower elongation ratios indicate more elongate basins (Figure 3). In a young extensional range, elongate basins will be associated with segment interiors and areas of higher fault activity due to headward erosion of the drainage system (Regalla et al., 2007). Once range half-width is established, however, catchments can only increase in size by lateral capture (Densmore et al., 2005). The Lemhi Range, ID (Figure 1), the focus of this study, has been an active extensional footwall for potentially as long as the last 7 Myr (see “Structural History” section in Regional Setting chapter), and has an established range half-width (Densmore et al., 2005).

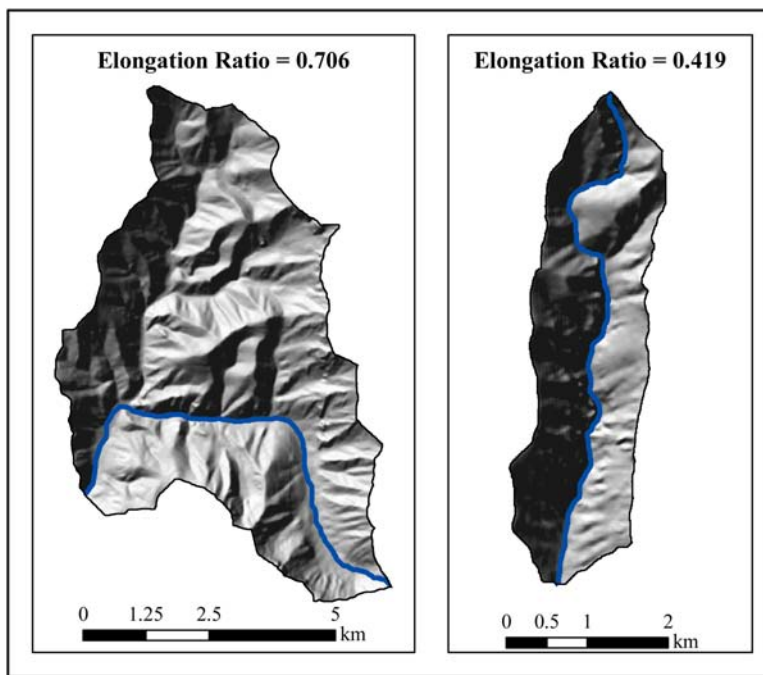


Figure 3. Elongation ratio comparison of two basins. A basin with the shape of a perfect circle would have an elongation ratio of one. Basin on the left is more round, so it has a higher elongation ratio. Basin on the right is more elongate, so it has a lower elongation ratio. Elongation ratio is dimensionless. Blue lines are flow lines of the longest drainage path within each basin (equivalent to basin length). For increased figure readability, the two basins are depicted at different scales.

Hypsometry is the distribution of elevations within a given area (Keller and Pinter, 2002; Rasemann et al., 2004). Hypsometric curves are constructed with relative elevation (percent of total basin relief) on the vertical axis and relative area (percent of total basin area) on the horizontal axis (Keller and Pinter, 2002). The hypsometric integral is defined as the area under the hypsometric curve, and can be approximated by dividing the difference between mean elevation and minimum elevation by the total relief (Keller and Pinter, 2002; Rasemann et al., 2004). Higher hypsometric integrals indicate that more area in the basin is located at higher elevations, and indicates more tectonic activity (Keller and Pinter, 2002; Harkins et al., 2005). Intermediate values of the hypsometric integral and sigmoidal hypsometric curve shapes are associated with decreased tectonic activity (Keller and Pinter, 2002; Harkins et al., 2005). Median basin elevation and mass are higher in areas of more tectonic activity within a given mountain range (Ritter et al., 1995).

The Melton ratio is another morphometric parameter used to describe fan drainage basins and is defined as the ratio between watershed (basin) relief and square root of watershed (basin) area (Wilford et al., 2005). In regions with basins dominated by different hydrogeomorphic processes (stream floods, hyperconcentrated flows, and debris flows), Melton ratios have been used in conjunction with basin length to characterize the different basin process classes (Wilford et al., 2005). Floods were the dominant process in basins with lower Melton ratios, while hyperconcentrated flows were the dominant process in basins with higher Melton ratios and longer basin lengths, and debris flows were the dominant process in basins with higher Melton ratios and shorter basin lengths (Wilford et al., 2005).

Alluvial Fan Processes

Individual depositional events on an alluvial fan may activate only portions of the fan (Blair and McPherson, 1998). An alluvial fan represents a collection of depositional events, erosional events, and other secondary modifications (Table 2, Blair and McPherson, 1994). Drainage basins factors control the primary depositional events on alluvial fans.

Table 2. Deposition, erosion, and modification of alluvial fans.

Alluvial Fan Processes (modified from Blair and McPherson, 1994)	
Primary Processes <i>Generated in fan drainage basin</i>	Secondary Processes <i>Modifications to the fan surface</i>
Generated by Bedrock Cliff Failure	• Overland flow winnowing
<u>Sediment-gravity flows</u>	• Rilling and gullying • Rainsplash erosion • Wind erosion and/or deposition • Plant rooting • Animal burrows
• Rockfalls • Rock slides • Rock avalanches	
Generated by Colluvial Slope Failure	• Sediment weathering
<u>Sediment-gravity flows</u>	• Soil development • Groundwater flow • Case hardening • Subsurface cementation
• Colluvial slides • Cohesive debris flows • Noncohesive debris flows	
<u>Fluid-gravity flows</u>	• Lateral erosion ('toe cutting'; Leeder and Mack, 2001) • Tectonic faulting or folding • Development of desert varnish (Blair and McPherson, 1998)
• Sheetfloods • Incised-channel floods	

Flows reaching alluvial fans are directly influenced by the character of the drainage basins that source them. Steep slopes with stored sediment and numerous first-order streams can rapidly transfer overland flow from across the drainage basin into the main channel sourcing the alluvial fan (Blair and McPherson, 1994). Two main types of alluvial fans, sheetflood and debris flow, have been identified based on the type of dominant flow (Blair and McPherson, 1994). Sheetfloods are unconfined, supercritical (Froude number > 1) fluid gravity flows with high attenuation and deposition rates as they progress downfan. Debris flows are cohesive sediment gravity flows of laminar character with sediment concentrations of 70-90% by weight (Pierson and Costa, 1987; Costa, 1988; Blair and McPherson, 1994). Flooding (fluid gravity streamflow) can occur in incised fan channels and will have similar characteristics to sheetflooding (turbulent, supercritical) without the rapid lateral expansion until reaching the intersection point downfan (Blair and McPherson, 1994).

Earthquake-induced landslides can also affect alluvial fans in several ways, including direct deposition on fan surfaces, removal or mobilization of previously deposited fan material, and by changing the drainage basins sourcing the fans (vegetation cover, sediment availability for future transport, connectivity of channel networks) (Keefer, 1999). Landslides are also capable of constructing entire landslide fans along active extensional range fronts.

Axial River Influences on Alluvial Fans

Due to increased subsidence near the fault in a half-graben system, axial river systems tend to be displaced toward the fault (hangingwall–footwall boundary), creating

an asymmetry in the axial sedimentation in the graben (Schumm et al., 2000). In slight contrast, a two-stage model of basin-filling argues that this is true during the synorogenic phase, but during the postorogenic phase, sediment shed from the uplifted footwall will displace the axial river back away from the fault (Mack and Leeder, 1999).

The bajada on the west side of the neighboring Lost River Range (Figure 1) has been noted for ‘toe-cutting’ by the axial river (Leeder and Mack, 2001). This process takes place when a non-incising axial river avulses to occupy the same space as the distal alluvial fan or bajada, eroding sediment and leaving steep fan-margin scarps (Leeder and Mack, 2001; Al-Farraj and Harvey, 2005). Once the river avulses again and abandons the fan edge, the fan can ‘heal’ the eroded areas by renewed deposition of fan material (Leeder and Mack, 2001).

In contrast, lobate fan margins grade into asymptotic contact with the adjacent valley floor where fans are not affected by toe-cutting. The Lost River bajada may have undergone up to 300 m of retreat during the Holocene due to toe-cutting by the axial river. Tectonic tilting or climate-change induced variations in discharge may affect the efficacy of toe-cutting, depending on basin context and time period in question (Mack and Leeder, 1999; Leeder and Mack, 2001). When climate change-induced toe-cutting occurs, it should be similar in occurrence and magnitude along neighboring fault segments (Mack and Leeder, 1999; Leeder and Mack, 2001). In contrast, occurrence and magnitude of toe-cutting will vary from segment to segment when the process is more dominantly tectonic in origin (Leeder and Mack, 2001). The effects of lateral movement of axial rivers on alluvial fans (toe-cutting) has been recognized in both modern settings

(Milana and Ruzyski, 1999; Leeder and Mack, 2001) as well as in the rock record (Mack and Leeder, 1999).

Study Area and Previous Regional Morphometric Work

This study focuses on alluvial fans and fan drainage basins of the southwestern flank of the Lemhi Range, east-central Idaho (Figure 4). Observations of the Lost River, Lemhi, and Beaverhead Ranges (Figure 1) at the range-bounding fault scale (as opposed to the segment scale) indicate that catchment relief and across-strike range width increase from the fault tips for about 15 kilometers before reaching relatively uniform values for the range interiors (Densmore et al., 2005). The range width of the Lemhi Range is particularly constrained due to limiting geometry of domino-style tilted fault blocks (Lost River Range and Beaverhead Mountains) on either side of the range (Densmore et al., 2005).

Early work (Funk, 1977) characterizing the sedimentology, stratigraphy, and geomorphology of fans in Birch Creek Valley to the northeast of the Lemhi Range (Figure 5) showed fan area and fan drainage basin area to be directly correlated. Funk (1977) stated that tectonics was a major control on few fans – the primary difference between fans was observed in the gradient and area of older fans (steeper and larger older fans), which was interpreted to be the result of increased runoff during glacial times.

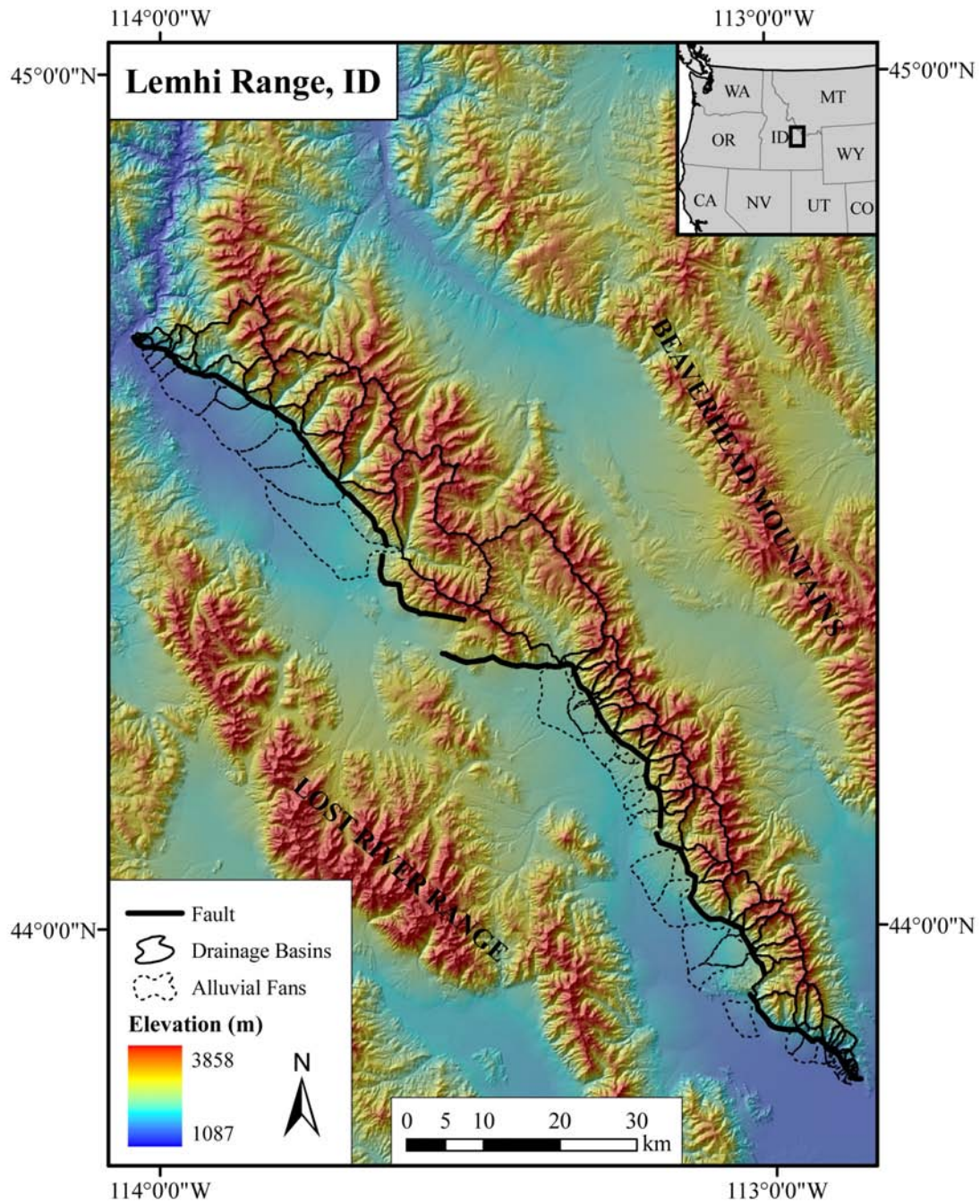


Figure 4. Location map of the study area showing the Lemhi Range with alluvial fans, fan drainage basins, and range-bounding Lemhi fault. Only basins that reach the main range drainage divide and their associated fans are mapped. The Lost River Range is located to the southwest of the Lemhi Range and the Beaverhead Mountains are to the northeast. For clarity, segments of the Lemhi fault are not depicted on this map; Figure 11 shows segment locations used in this study.

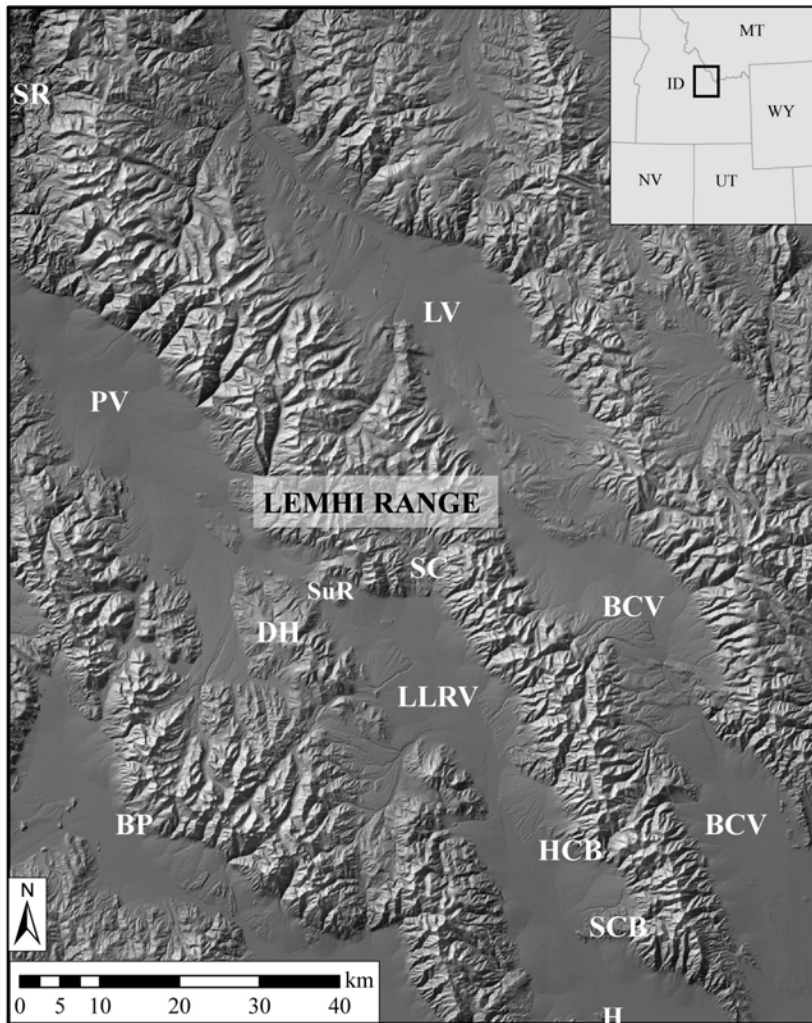


Figure 5. Detail map of Lemhi Range and nearby features. BCV: Birch Creek Valley, BP: epicenter of 1983 Borah Peak earthquake, DH: Donkey Hills, H: Howe, HCB: Horse Creek block, LV: Lemhi Valley, LLRV: Little Lost River Valley, PV: Pahsimeroi Valley, SC: Sawmill Canyon, SCB: South Creek block, SR: Salmon River, SuR: Summit Reservoir.

Hypothesis Questions

Since fault slip necessarily increases toward the center of segments, along-strike morphometric variations in drainage basin and alluvial fan parameters should result

(parameters described in Table 3). This hypothesis will be investigated by the following three questions that relate to morphometric variation along strike:

1. Does the ratio of fan area to drainage basin area (ϕ) decrease toward the interior of fault segments?

2. Are drainage basins more elongate (lower elongation ratio) closer to fault segment boundaries and more round (higher elongation ratio) in the interior of fault segments?

3. Does hypsometry change along strike with contrasting hypsometric curve shapes in the interiors of segments compared to near segment boundaries? Specifically, are drainage basins in segment interiors characterized by higher hypsometric integrals and are drainage basins closer to segment boundaries characterized by lower hypsometric integrals and a more sigmoidal hypsometric curve shape?

A fourth question will allow the drainage basins and alluvial fans to be placed in context with other studies of Quaternary alluvial fans:

4. Do the fans along the western Lemhi Range follow the traditionally accepted relationships described in equations 1-3 (“Alluvial Fan and Drainage Basin Area and Slope Relationships” section above)? In other words, are fan area and fan drainage basin area directly related; fan slope and drainage basin area inversely related; and drainage basin slope and drainage basin area inversely related?

Finally, the Melton ratio will be used to compare drainage basins as a function location along strike relative to segment boundaries. However, no hypothesis is proposed for how basin location along a segment boundary will affect Melton ratio. This ratio was not included in the original *a priori* questions and is therefore used in an exploratory

sense. Stated as a question: Is Melton ratio correlated to drainage basin distance from nearest fault segment boundary?

Table 3. Morphometric parameters used in this study to characterize alluvial fans and fan drainage basins.

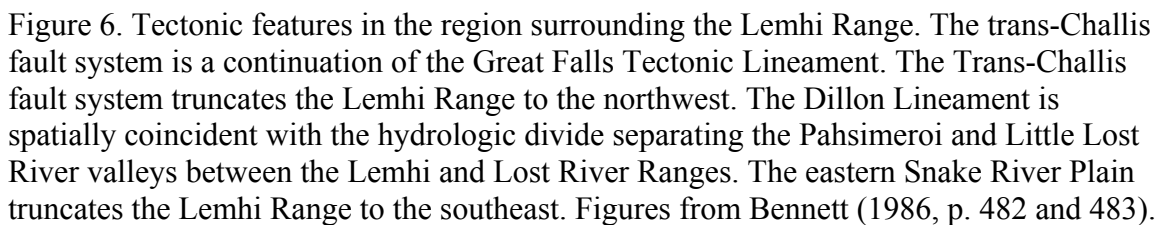
Parameter	Description	Acquisition Method	Format
Basin Area	Planimetric area of watershed (km ²)	Extracted from DEM	Polygon feature with area as a number
Fan Area	Planimetric area of fan (km ²)	Mapped on airphoto	Polygon feature with area as a number
Basin Slope	Min, max, range, mean slope (degrees, percent)	Extracted from DEM (zonal statistics)	Numbers
Fan Slope	Min, max, range, mean slope (degrees, percent)	Extracted from DEM (zonal statistics)	Numbers
Basin elevation	Min, max, range, mean elevation (m)	Extracted from DEM (zonal statistics)	Numbers
Longest Stream Path	Path of longest flow route for a basin (measured in m)	Extracted from DEM with digitization of calculated seed point	Line feature
Elongation Ratio	Measure of basin shape (roundness = width to length ratio)	Computed from DEM extracted data	Number (dimensionless)
Hypsometric Curve	Area-elevation distribution	Computed from DEM extracted data	Smooth curve of x-y point data
Hypsometric Integral	Area under the hypsometric curve	Calculated from hypsometric curve points	Number
Melton ratio	Ratio of watershed relief to square root of watershed area	Extracted from DEM	Number

REGIONAL SETTING (GEOLOGY, TECTONICS, AND CLIMATE)

This study focuses on the southwestern flank of the Lemhi Range in east-central Idaho (Figures 1, 4). The study area covers about 2,500 km². The range is part of the late Mesozoic Cordilleran fold and thrust belt as well as the Cenozoic Basin and Range structural province (Rodgers and Janecke, 1992; Janecke, 1993). The Lemhi Range is within the region classified as the Cordilleran extensional province, and in east-central Idaho the direction of maximum horizontal compressive stress (S_{hmax}) is oriented WNW (Zoback and Zoback, 1991). Maximum horizontal compressive stress is the intermediate principal stress in an extensional system; the maximum principal stress is vertical or $S_v > S_{hmax} > S_{hmin}$ (Zoback and Zoback, 1991).

The north-trending Salmon River fault truncates the Lemhi Range to the northwest (Figure 6). The Salmon River fault is part of the trans-Challis fault system (Bennett, 1986), or is a possibly a continuation of the Great Falls Tectonic Zone (O'Neill and Lopez, 1985). This fault has also been interpreted as a tear fault related to the Mesozoic Poison Creek thrust system (Figure 7) but with more recent normal movement (Tysdal, 2002). The eastern Snake River Plain truncates the Lemhi Range to the southeast (Figure 1).

The Lemhi Range is the central of three structurally similar northwest-trending ranges (including the Lost River Range to the southwest and the Beaverhead Range to the northeast, Figure 1) bound by active normal faults on at least the southwest flanks (Crone and Haller, 1991; Gorton, 1995). The ranges are east-tipping (westward dipping) fault



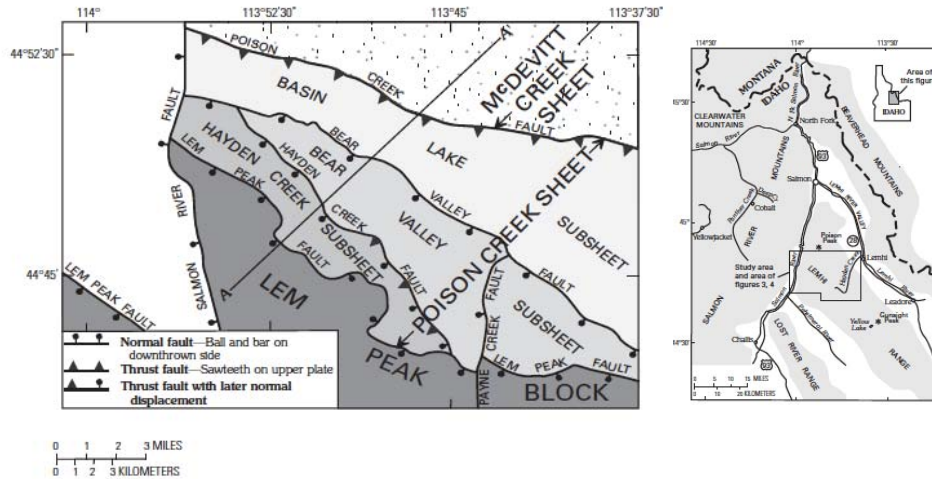


Figure 7: Left, Mesozoic Poison Creek thrust system in the northern Lemhi Range (location map at right). The Salmon River fault has been interpreted as a tear fault originating during Mesozoic compression but with more recent normal movement (Tysdal, 2002). Figures from Tysdal (2002, p. 2 and 5).

An arch trending SW-NE breaks up the axial valleys of the three ranges. Between the Lemhi and Lost River Ranges, the arch in the Donkey Hills - Summit Reservoir area (Figure 5) serves as a hydrologic divide (2,088 m) between the Pahsimeroi River Valley draining to the northwest and the Little Lost River Valley draining to the southeast (Knoll, 1977). Similarly, the axial valleys between the Lemhi and Beaverhead Ranges are hydrographically divided by a topographic high (Gilmore Summit at 2,190 m) toward the center of the main valley, separating the drainages for the northwest-flowing Lemhi River and the southeast-flowing Birch Creek (Figure 5). The highs dividing the north-flowing and south-flowing axial rivers are aligned along a N70E trend (Haller, 1988). The topographic highs are coincident with a lineament informally termed the Dillon line (Bennett, 1986). A left-stepping lateral ramp in the Hawley Creek thrust sheet may coincide with the structural high (Rodgers and Janecke, 1992). On the western side of the Lemhi Range, the northwest-flowing Pahsimeroi River flows out of the valley and

becomes a tributary of the Salmon River; and the southeast-flowing Little Lost River disappears into alluvium south of Howe, ID (Figure 5).

Structural History

Before delving into a discussion of mid-late Cenozoic deformation, a review of prior events is critical, as crustal rheology, flexural rigidity and fault geometry all are influenced by the inherited factors of previous tectonism, including heat flow, crustal thickening, and preexisting structural features (Friedmann and Burbank, 1995; Janecke, 2007). East-central Idaho and adjacent western Montana have a complex history of tectonic development extending from Holocene back to at least middle Proterozoic time (O'Neill and Lopez, 1985). The folding and faulting history of the range is one of alternating compressional and extensional deformation. Mesoproterozoic compression was followed by Neoproterozoic to early Paleozoic extension, Late Cretaceous to early Cenozoic compression, and onset of extension by the Eocene, with some extensional deformation continuing to present (Skipp and Link, 1992; Tysdal, 2002). NE-striking faults developed in the Eocene, followed by Eocene-Oligocene volcanism, and regional uplift, volcanism and NW striking faults later in the Cenozoic (Crone and Haller, 1991).

Some work indicates that if a Proterozoic compressional deformation event occurred, any cleavage or other compressional structures are coincident and indistinguishable from those resulting from Cretaceous compression (Tysdal, 2002). Regardless, in the southern Lemhi and Beaverhead Ranges, thin lower Paleozoic sedimentary rocks unconformably overlie Precambrian rocks, indicating possible sporadic early Paleozoic uplift (Skipp and Link, 1992).

In the western U.S. Cordillera, fold-and-thrust deformation and arc magmatism began around 120 Ma, intensifying by 105-100 Ma with major deformation occurring 105-80 Ma (Livaccari, 1991). Deformation prograded into the foreland beginning around 80-75 Ma, and in the northern Rockies, compressive deformation ceased around 56 Ma (Livaccari, 1991).

In east-central Idaho, shortening began in the Cretaceous and ended by at least the middle Eocene and onset of Challis volcanism (Rodgers and Janecke, 1992). At least two middle Eocene paleovalleys, the Lemhi Pass paleovalley and the Hawley Creek paleovalley (Figure 2), have been identified in the Lemhi Range-Beaverhead Mountains area (Janecke et al., 2000). These were likely among the youngest components of a major late Cretaceous – early Eocene paleoriver system draining the Sevier uplands toward the foreland and may be the source for the Pinyon and Harebell conglomerates of northwest Wyoming (Janecke et al., 2000; Janecke and Blankenau, 2003). While these paleovalleys currently contain middle Eocene rocks, the paleovalleys were likely formed during the Cretaceous (Janecke et al., 2000). The Cordilleran thrust belt hinterland has subsequently undergone Cenozoic extensional collapse following the cessation of predominately Late Cretaceous compressional deformation (Tysdal, 2002). East-central Idaho and neighboring western Montana have been considered similar to the Basin and Range structural province to the south by many workers for several decades, despite separation from the main Basin and Range region by the Snake River Plain (see Reynolds, 1979).

Extension of the Cordilleran thrust belt has been attributed to several plate tectonic factors including gravitational collapse of overthickened crust resulting from plate convergence, changes in slab orientation and subduction angle (due to changes in

composition or age of subducting slab or rate of subduction), triggering by Tertiary magmatism, and, in the late Cenozoic, extension and shear associated with development of the San Andreas fault zone (Atwater, 1970; Atwater, 1989; Livaccari, 1991; Janecke, 1994). Gravitational collapse is inferred from multiple lines of evidence including location of metamorphic core complexes to the west of the fold and thrust belt front and minimum horizontal compressive stress (S_{hmin}) generally oriented perpendicular to the topographic bulge defining the eastern boundary of the Cordilleran extension province – this spatial coincidence suggests a causal relationship (Coney and Harms, 1984; Livaccari, 1991; Zoback and Zoback, 1991). Flattening of the subducting slab in the late Cretaceous (~80 Ma) may have caused the eastward sweep of the Cordilleran magmatic arc (Atwater, 1989). The inboard magmatism occurred in the northern U.S. Cordillera (Idaho and Montana) starting around 80 Ma, and reached inboard in more southern regions (western Texas and the Mexican Sierra Madre Occidental) by about 35 Ma (Atwater, 1989). Onset of extension (early Basin and Range) is interpreted to have occurred due to removal of the shallowly subducting slab and transition back to steeper subduction along the western coast (Atwater, 1989). The broad, diffuse, uplifted area of extension has been linked to shallow subduction of buoyant (young, hot) lithosphere under western North America (Allen and Allen, 2005).

The late Tertiary (Neogene) deformation of the western U.S. (Basin and Range province and parts of Oregon, Washington, and Idaho) has been interpreted as a megashear zone related to the San Andreas fault following the widespread acceptance of plate tectonic theory (Atwater, 1970; Smith and Sbar, 1974; Smith, 1977). San Andreas shear does not explain early Cenozoic extension in the western U.S. because the

transcurrent plate boundary postdates some of the extension. Notably, onset of mid-Tertiary (Paleogene) core complex activity preceded San Andreas development. In the intermountain region, metamorphic core complex formation in the Sevier thrust belt occurred within 50-70 Myr of thrusting, and in some cases began within 1-2 Myr (Friedmann and Burbank, 1995). The Pacific and North American Plates did not come into contact until around 27-25 Ma (Atwater, 1970; Coney and Harms, 1984; Atwater, 1989). In the early Cenozoic (43 Ma), subduction of the Farallon and its derivative plates dominated along the western margin of the North American plate; the San Andreas and associated strike-slip and extensional deformation became progressively dominant starting in the middle to late Cenozoic (Atwater, 1989). The right-lateral movement along the San Andreas causes tension to the northwest and “we might consider western North America to be a very wide, soft boundary between 2 rigid moving plates” (p. 3525, Atwater, 1970; Smith, 1977).

A recent two-dimensional finite element analysis of the modern North American plate considered boundary (plate margin), internal (gravitational collapse) and basal (craton root drag) loads (Humphreys and Coblenz, 2007). The main factors contributing to ongoing deformation in the western U.S. are shear associated with movement along the San Andreas, extensional collapse of the Cordillera, and southern Cascadia pull (Humphreys and Coblenz, 2007). Gravitational potential energy can create intraplate horizontal stress; the stresses are a function of the horizontal gradient of the vertical mass of overlying rock. Gravitational collapse occurs when the horizontal deviatoric stresses cause extension (Coney and Harms, 1984; Humphreys and Coblenz, 2007). At present, the crust in the western U.S. is not anomalously thick, so the high gravitational potential

energy is no longer simply due to overthickened crust (Humphreys and Coblenz, 2007). Instead, Laramide and post-Laramide changes to the lithosphere as well as the sinking Farallon slab and Yellowstone hotspot all contribute to the modern high elevation and thus high gravitational potential energy of the western U.S. (Humphreys and Coblenz, 2007).

Modern deformation in the northern Basin and Range is characterized by three different styles: steeply (45-60°) dipping planar normal faults bounding simple basins; low angle listric or planar normal faults bounding asymmetric tilt basins; or planar and listric normal faults connected to a low-angle detachment defining complex basins and subbasins (Smith and Arabasz, 1991).

Multiple Phases of Cenozoic Extension in east-central Idaho

The northern Basin and Range Region (east-central Idaho and southwest Montana) underwent at least four, and in some areas up to six phases, of kinematically distinct extension during the Cenozoic (Janecke 1993; Janecke, 1994; Janecke, 2007; Petrik, 2008), with more recent deformation events overprinting older ones (Bennett, 1986). Earliest extension may have occurred during the late Cretaceous to early Paleogene with SW dipping normal faults, but widespread extension did not occur until sometime in the middle Eocene (Janecke, 2007). The emplacement of the Challis Volcanic Group in the Eocene (~50-45 Ma, Janecke, 2007) occurred after the compressional deformation had ceased (Rodgers and Janecke, 1992, Tysdal, 2002) and also postdates the onset of widespread extension (Tysdal, 2002; Janecke and Blankenau, 2003). Several faults with most recent normal-sense movement have been identified as

reactivations of older (Cordilleran) thrust faults (including the Bear Valley and Hayden Creek faults in the northern Lemhi Range as mapped by Tysdal, 2002, Figure 7). The current topography of the range is primarily an expression of Cenozoic basin-and-range style extension (Tysdal, 2002).

The main phases of extension discussed below are: 1) Eocene metamorphic core complex development in the north with lesser extension on high-angle NE-striking normal faults in the south; 2) Paleogene supradetachment basin development in a N-S trending belt; 3) tilt-block (basin and range) style N to NW trending normal fault development; modified by: 4) NE to E trending normal faults associated with the passage of the Yellowstone hotspot.

Eocene Metamorphic Core Complex Development: In Idaho, the area of Eocene extensional deformation is bound by the Lewis and Clark line to the north and by the Snake River Plain to the south (Bennett, 1986). Extreme extension and metamorphic core complex development occurred from the Bitterroot Range (Figure 1) northward, while in central Idaho lesser extension occurred along NE-striking high-angle normal faults (Janecke and Blankenau, 2003; Janecke, 2007). During the Eocene, metamorphic core complexes, including the Anaconda and Bitterroot, formed as a result of partial middle crustal melt and magmatic emplacement. The crust would have been thermally and mechanically weakened due to thickening and magmatism (both subduction-related magmatism as well as magmatism related to radiogenic heat flow). Emplacement of the batholiths occurred as regional deformation switched from Late Cretaceous – early Paleogene contractional (Sevier-Laramide) deformation to extensional deformation

(Bennett, 1986; Janecke, 2007). During a time frame similar to the development of the core complexes, small NE-striking normal faults developed south of the Bitterroot Range. These faults included the Lem Peak and Divide Creek faults just north of the Lemhi Range (Janecke and Blankenau, 2003; Janecke, 2007). These small NE-striking faults first developed in the early-mid Eocene concurrent with Challis volcanism; large-magnitude extension in central Idaho began after the cessation of Challis volcanism (Janecke and Blankenau, 2003; Janecke, 2007).

Paleogene Supradetachment Basins: During the Late-Middle Eocene to Oligocene (or possibly early Miocene, Janecke, 2007) a narrow (100 ± 25 km) dominantly N-S trending rift zone north of the eastern Snake River Plain developed and contained multiple supradetachment basins (Janecke, 1994; Janecke and Blankenau, 2003). This period started during the waning phases of Challis magmatism: Challis magmatism terminated early during this phase of extension with a rhyolitic phase (Janecke, 1994). Continued gravitational collapse of the fold and thrust belt led to formation of many mostly east-tilted half graben that filled with up to 2.6 km of Eocene to Oligocene conglomeratic, sandstone and tuffaceous basin fill (Janecke, 1994). Eocene-early Miocene supradetachment basins of east-central Idaho and western Montana include the Salmon, Horse Prairie, Grasshopper, Medicine Lodge, Muddy Creek, and Nicholia Creek basins (Figure 2, Janecke and Blankenau, 2003). Complex extensional faults and folds (Xiao and Suppe, 1992; Shaw et al., 1997) complicate the present geometries of the basins (Janecke, 2007).

One of the Eocene-early Miocene supradetachment basins, the Muddy Creek half-graben of southwest Montana, has been used to test proposed models of extensional basin development (see “Models of Extensional Basin Evolution and Resulting Stratigraphy” in Introduction chapter). While this basin had the structure of a supradetachment basin, the stratigraphy does not match the predictions of the supradetachment model of Friedmann and Burbank (1995); instead, the basin seems to fit the models of Leeder and Gawthorpe (1987) more closely (Janecke et al., 1999). The Muddy Creek basin is characterized by: rare mass-wasting deposits, low volume of hangingwall-sourced coarse debris, dominant basin-center deposystems, rare angular unconformities, syntectonic basin fill up to 2.5 km thick, dominantly footwall-sourced sedimentation and depocenters proximal to the basin-bounding fault (Janecke et al., 1999).

In the Salmon basin just north of the Lemhi Range (located between the northern Lemhi Range and Beaverhead Mountains), the mid-Eocene – early Miocene phase has been further split into three smaller phases that slightly overlap (Janecke and Blankenau, 2003). These are termed the Tendoy phase (~ 45-30 Ma, mid-Eocene– Oligocene); the Lemhi Pass phase (~35 -27 Ma, Oligocene); and the Sacagawea phase (~31 – 20 Ma, mid-Oligocene – early Miocene). In addition to differing activity on local faults, each of these smaller phases is characterized by a slightly different maximum extension direction, from NE in the Tendoy phase to NNE in the Lemhi Pass phase to slightly ENE in the Sacagawea phase (Janecke and Blankenau, 2003).

Basin and Range Extension: The ongoing phase of northeast-southwest extension is termed the Basin and Range phase (Janecke and Blankenau, 2003) and began in the mid-Miocene (~12 Ma in east-central Idaho, Janecke and Blankenau, 2003). This ongoing phase has a broader spatial distribution than the more confined Paleogene extension (Janecke, 2007). This extensional phase is characterized in east-central Idaho by a shift in extension direction from the previous phase with the onset of predominately northwest-trending fault systems, namely the Lost River, Lemhi, and Beaverhead faults (Figure 1, Janecke, 2007). These faults mimic the geometry of the Cretaceous fold and thrust belt (Rodgers and Janecke, 1992; Janecke, 2007). The faults have large amounts of Quaternary displacement generating tilted range blocks via footwall uplift and asymmetric hangingwall subsidence (Smith and Arabasz, 1991). For normal faults dipping 45-60°, hangingwall subsidence is 2-4 times the amount of footwall uplift (Smith and Arabasz, 1991), sometimes more specifically reported as 0.85:0.15 hangingwall subsidence to footwall uplift (Leeder and Jackson, 1993).

Northern Rockies basin-and-range style extension began around the time emplacement of the Columbia River and Oregon Plateau flood basalts (~ 16 Ma) and has continued concurrently with Yellowstone hotspot migration through eastern Snake River Plain (Pierce and Morgan, 1992; Janecke, 2007). The southern axial valley between the Lemhi and Beaverhead Ranges is the Birch Creek Valley (Figure 5). Valley fill dates to at least part of the Miocene and Pliocene based on Miocene vertebrate fauna and tuff deposits, indicating the basin was subsiding by this time (Funk, 1977).

The northern Basin and Range province is bound by the eastern Snake River Plain to the south, the Lewis and Clark line in Montana to the north, the Idaho batholith to the

west, and the Rocky Mountain front to the east (Petrik, 2008). Older research placed the northern boundary of northwest trending structures associated with basin and range extension further south at the trans-Challis fault system (an extension of the Great Falls lineament in Montana, Bennett, 1986). However, more recent studies indicate active extension in the northern Basin and Range occurs north to, and includes seismicity along, the Lewis and Clark line (Stickney and Bartholomew, 1987; Smith and Arabasz, 1991).

The basins associated with this ongoing deformation are characterized by Friedmann and Burbank (1995) as closer to the rift basin end-member in their proposed supradetachment-rift basin spectrum (see “Models of Extensional Basin Evolution and Resulting Stratigraphy” in Introduction chapter). Individual basins and their bounding faults tend to be characterized by relatively low slip rates with small amounts of extension, steep (45-70°) bounding faults, and major hangingwall sediment sourcing (Friedmann and Burbank, 1995). However, many of the younger high-angle faults bounding the modern rift-style basins cut older low-angle faults that may have bound supradetachment-style basins (Friedmann and Burbank, 1995). The transition from a supradetachment basin with high rates of extension along low-angle faults to a rift/half-graben basin may be due to strengthening of the lithosphere as it returns to a thickness of 30-40 km. The crust of the Idaho-Wyoming thrust belt was likely 30-40 km thick pre-compression but reached thicknesses of 50-60 km during Middle Triassic to early Tertiary thrusting (Coney and Harms, 1984). This overthickened crustal welt became the site of later Eocene core complex development (Coney and Harms, 1984). Examples of systems where supradetachment basins have been overprinted by younger rift-style deformation include the late Miocene Greater Amargosa Chaos-Buckwheat-Sperry Hills

Basin/Death Valley; pre-Rio Grande basin/Rio Grande rift; and the Northern Aegean Strymon Basin/active Aegean extensional basins (Friedmann and Burbank, 1995). However, as described above (“Paleogene Supradetachment Basins” section), some workers disagree that the stratigraphy of Paleogene supradetachment basins in the northern Basin and Range region fit the model of Friedmann and Burbank (1995).

Modification by Passage of the Yellowstone Hotspot: As the Yellowstone hotspot (see “Yellowstone Hotspot and the Snake River Plain” below) moved northeast relative to the North American plate towards its present location in NW Wyoming, it affected the stress regime of surrounding areas (Janecke, 2007). The current location of the Yellowstone-influenced strain field includes the southern end of the Red Rock fault and the Centennial fault (Figure 1, Janecke, 2007). As the hotspot migrated northeastward, E-striking cross faults were generated that cut the NW-striking basin-bounding faults (Janecke, 2007). These E-W faults have small offsets, and some dip north while others dip south (Janecke, 2007).

Regional Stratigraphy

The oldest rocks in the Lemhi Range are Meso- and Neoproterozoic (Rodgers and Janecke, 1992; Skipp and Link, 1992; Tysdal, 2002). Proterozoic rocks are almost without exception metamorphosed to lower greenschist facies (Tysdal, 2002). Paleozoic rocks in the Lemhi Range are progressively younger to the southeast (Rodgers and Janecke, 1992; Tysdal, 2002). Rocks in the northern part of the range are chiefly Mesoproterozoic; while the central and southern extent of the range also include parts of the

Neoproterozoic and Phanerozoic sections (Tysdal, 2002). The northern Lemhi Range contains the Gunsight, Swauger, and equivalent formations of the Belt Supergroup as well as the Apple Creek Formation (Janecke and Blankenau, 2003), which are quartzites and siltstones with minor conglomeratic and dolomitic lenses (Turko, 1988). The southern Lemhi Range contains dominantly Paleozoic and Mesozoic sedimentary rocks (Janecke and Blankenau, 2003). The Eocene Challis Volcanic Group is exposed in places (Tysdal, 2002). The south-central and southern Lemhi Range has less variable lithology and contains mostly Paleozoic carbonates (limestone and dolomite) with quartzite and some siliciclastic layers (Turko, 1988; Johnson et al., 2007).

Lemhi Alluvial Fans

The alluvial fans on the west side of the Lemhi Range are not typified by the classical cone morphology with distributary drainage. Rather, the fans are in large part coalesced to form a bajada. The surface of much of the bajada displays contributory drainage and many of the fans display linear troughs parallel to the fan radius (Figure 8). These are known as “channel ways” and represent erosion of fan debris (Patterson, 2006). Higher precipitation, lower angle faults, and lower slip rates all favor the development of prograding, coalesced alluvial fans (bajada-form) due to increased sediment production (higher precipitation) or decreased accommodation (fault geometry). Lower precipitation, higher fault plane angle, and higher slip rates favor nonprograding, isolated fans due to decreased sediment production (lower precipitation) or increased accommodation (fault geometry) (Allen and Densmore, 2000). In the study area, alluvial fan deposits resemble the first case: coalesced bajada-form fans. Of the three factors identified by Allen and

Densmore (2000), the least likely factor contributing to the coalesced form of the Lemhi Fans is low fault angle, as geophysical evidence does not indicate a low-angle fault (see “Quaternary Faulting and the Intermountain Seismic Belt” below). A study of alluvial fans on the western side of the neighboring Lost River Range (Figure 1) showed that three neighboring drainage basins of similar bedrock lithology (carbonate within the same formation) produced sheetflood-dominant and debris flow fans. Drainage basin size was interpreted to be the controlling factor rather than bedrock lithology (Patterson, 2006).

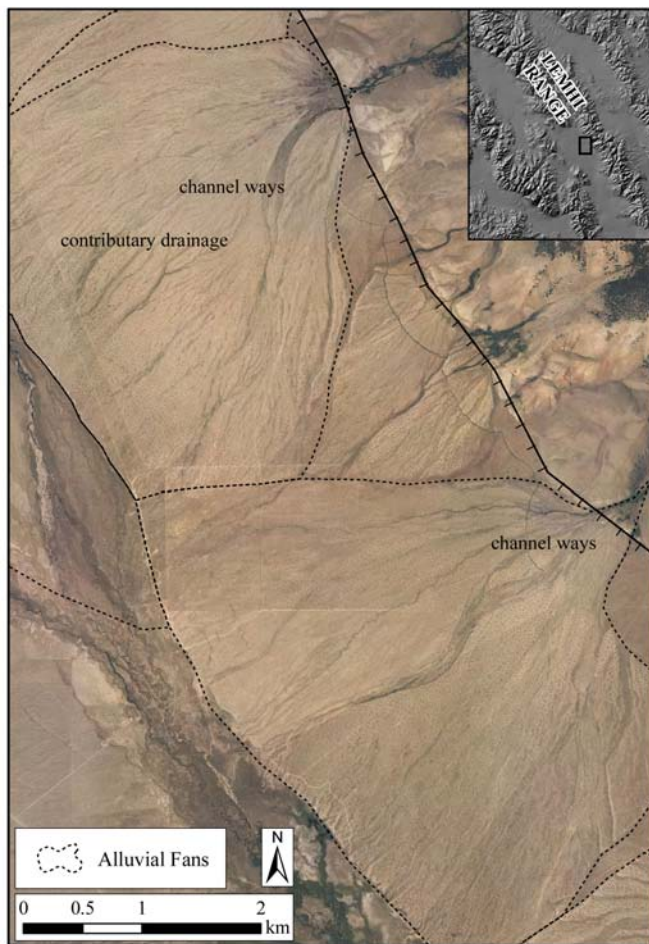


Figure 8. Channel ways (incised channels) and contributory drainage on older surfaces of coalesced alluvial fans on the western flank of the Lemhi Range. Black line is fault trace, bars on downthrown side.

Modification of fan surfaces, whether by humans or by adjacent geomorphic systems can affect the ease of mapping fan units accurately. Human modification of the Lemhi fans involves cattle grazing, drainage ditch construction and water diversion, gravel/dirt road construction and some agricultural activity. However, most fan surfaces are relatively undisturbed by human modification. Distal portions of some of the alluvial fans have been eroded by the axial river to varying degrees (Figure 9). This is the ‘toe-cutting’ process described by Leeder and Mack (2001). Thus, the mapped alluvial fans represent a minimum surface area of deposition sourced from each drainage basin.

Alluvial depositional forms on the western side of the Lost River Range are characterized by large, dissected, inactive Pleistocene sheetflood alluvial fans that are proximally buried by smaller, young (Holocene) debris flow alluvial fans (Patterson, 2006). Increasing steepness with age of many of the Lost River Range fan surfaces indicates progressive tilting of the fan surfaces through time, perhaps by transfer of hangingwall fan surfaces to the uplifting footwall block as the bounding fault steps basinward (Patterson, 2006). While not a focus of this study, similar features to those described by Patterson (2006) were observed on the Lemhi fans including apparently steeper older surfaces and proximal debris flow deposition on fans that appeared to have been previously sheetflood-dominated. Given the similar structural, stratigraphic, and climatic histories of the two ranges, it is not unreasonable to suspect that the fans sourced from the footwalls of both ranges show similar depositional patterns through time.

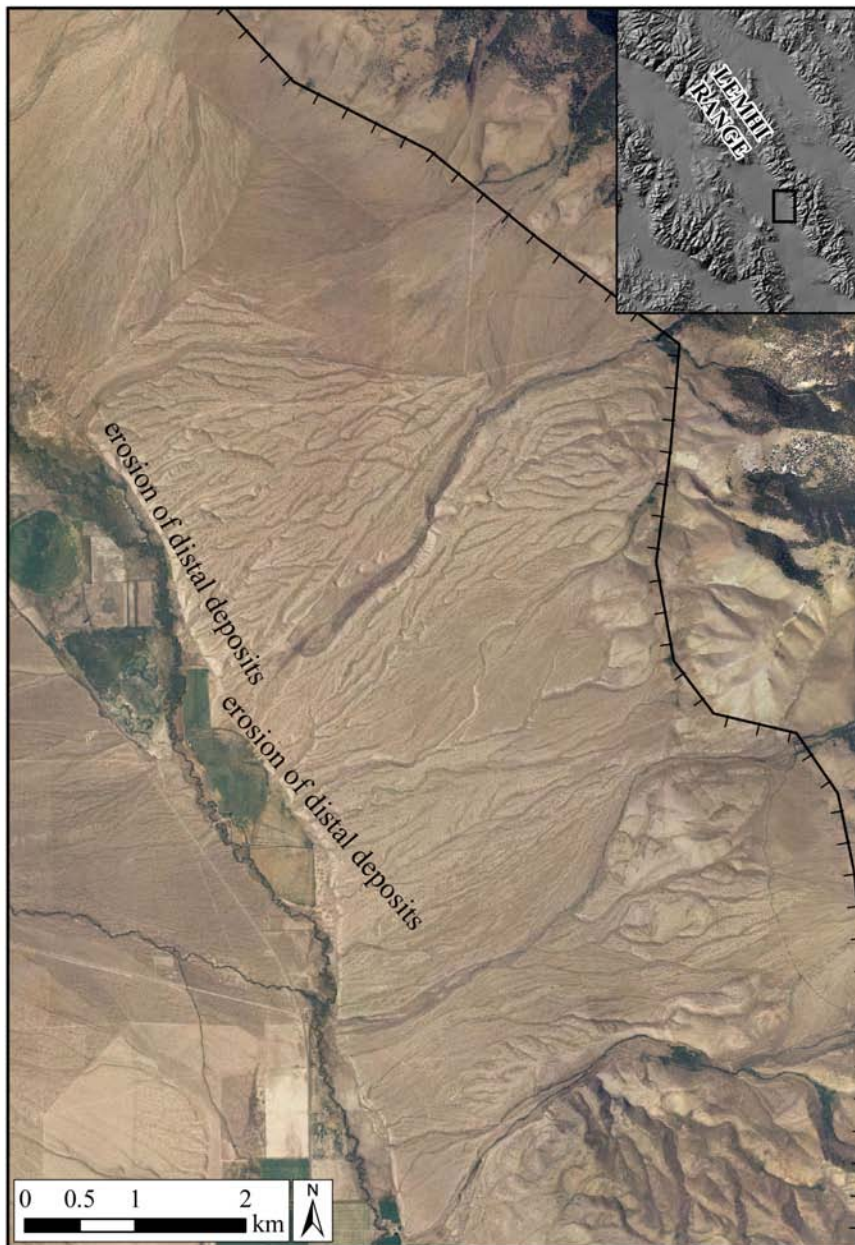


Figure 9. Toe-cutting by axial river on the western flank of the Lemhi Range. Black line is fault trace, bars on downthrown side. Note steep scarps formed near agricultural activity. These scarps formed when the axial river migrated or avulsed closer to the fans and eroded distal fan deposits.

Quaternary Faulting and the Intermountain Seismic Belt

Quaternary Seismicity of the Lemhi and Surrounding Ranges

The Lost River, Lemhi, and Beaverhead Ranges of Idaho and the Tendoy Mountains of Montana (Figure 1) are strikingly similar from a structural perspective, with fault segment boundaries aligning along an ENE line subparallel to the trend of the Snake River Plain. Each range is bound on at least one side by a normal fault and at least 21 large (surface rupturing) earthquakes have occurred in the last 30,000 years on the four range-bounding faults. (Haller, 1988). Quaternary fault scarps occur dominantly on the western side of the Lost River, Lemhi, and Beaverhead Ranges (Turko, 1988). All range-bounding faults (faults of the same name as the range except the Red Rock fault bounding the Tendoy Mountains) have had a seismic event resulting in surface offset along at least one segment during the Holocene (Haller, 1988; Crone and Haller, 1991). Extensional deformation in the Lost River and Lemhi Ranges may have been concentrated during the last 4-7 million years based on stratigraphic relationships (Scott et al., 1985). Specifically, uplifted ignimbrites in the southern Lemhi Range are older than 4.3 Ma but mostly younger than 6.5 Ma (Pierce and Morgan, 1992).

The mean slip rate in the central section of the Lemhi and Lost River Ranges is estimated at 0.3 m/1000 yrs, whereas the maximum mean slip rate at the southern ends of the ranges is estimated at 0.1 m/1000 yrs (Scott et al., 1985). This 3-fold increase in slip rate along strike is consistent with other normal fault systems: the Wasatch (Figure 1) fault zone in Utah displays an increase in Holocene displacement rates by a factor of up

to five between distal and central segments (Machette et al., 1991). The timing of the most recent fault movement shows an along-strike trend for all three ranges. The southern segments show Late Pleistocene movement at the latest, while the central and north-central segments display latest Quaternary movement and the northernmost segments may only have had most recent movement earlier than the late Pleistocene (Scott et al., 1985). Scott and others (1985) cited offset of Pinedale (as young as 12,000-15,000, see “Pleistocene Glaciations” below) deposits and Quaternary gravels (unspecified age) located along segments in the central parts of the ranges as evidence for latest Quaternary movement (Scott et al., 1985). The central portions of the ranges have been active more recently, and have had more activity than the northern and southern portions of the ranges (Scott et al., 1985). Based on scarp morphology in the Lemhi Range, the Sawmill Gulch and Patterson segments (Figure 10, terminology of Haller, 1988, see “Previous Work Delineating Fault Segment Along the Lemhi Range” below) both have Holocene displacement, and the segment between them (Goldburg) shows evidence of rupture around 15,000 years ago (Haller, 1988). The longest segments are located in the central or south-central regions of the ranges (Crone and Haller, 1991). This along-strike footwall pattern, with longer segments, higher slip rates, and highest footwall topography toward the center of the ranges is typical for a large active normal fault zone (Gawthorpe and Leeder, 2000). Alternate segmentation models, including more recent models and the one used in this study (Janecke, 1993) do not indicate longer segments in the central Lemhi Range.

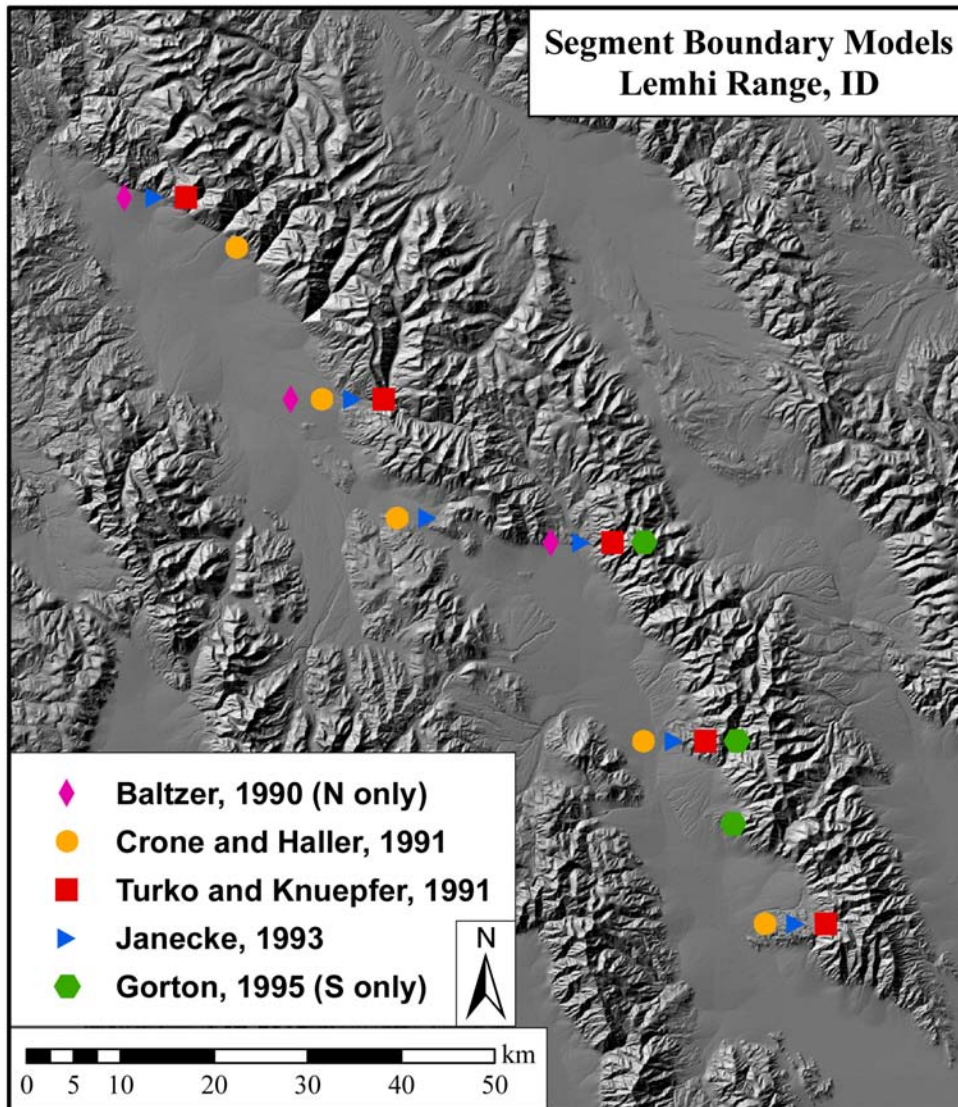


Figure 10. Segment boundaries of the Lemhi fault as defined by different workers. Boundaries at the extreme northern and southern tip of the range are not depicted for clarity. For segment boundaries identified at the same location by multiple workers, the symbols are offset horizontally. Segment names (Baltzer, 1990) from north to south are: Ellis, May, Big Gulch, and Warm Creek (did not map further south than the Big Gulch-Warm Creek boundary). Segment names (Crone and Haller, 1991) from south to north are: Howe, Fallert Springs, Sawmill Creek, Goldburg, Patterson, and May (same as Haller, 1988). Segment names (Turko and Knuepfer, 1991) from south to north are: Howe, Fallert Springs, Warm Creek, Big Gulch, Falls Creek, and Ellis (refined from model by Turko, 1988). For Janecke (1993) segment names, see Figure 11. Gorton (1995) did not change fault segment names, but suggested the Howe-Fallert Springs boundary be moved north of where previous workers placed it. Gorton (1995) did not map further north than northern Warm Creek boundary.

The 1983 Borah Peak Earthquake

The October 28, 1983 M_S 7.3 ($M_W = 7.0$, Keefer, 1999; $M_M = 6.8$, Gorton, 1995; for a discussion of earthquake magnitude scales, see Appendix B) Borah Peak earthquake resulted in a 36-km long zone of surface faulting along the western front of the north-central Lost River Range (Crone, 1987; Crone et al., 1987). This earthquake resulted in two deaths in Challis, ID, and an estimated \$15 million in damage to the surrounding area (Crone and Machette, 1984). Over 40 sediment boils and sinkholes formed near Chilly Buttes about 23 miles from the epicenter (Waaq, 1984). During the earthquake, a debris slump with surface area of 40,000 m² (volume approximately 100,000 m³) and 5-10 m of displacement with headward-rotated blocks occurred on the Birch Springs alluvial fan (Keefer, 1999). Movement was dominantly dip-slip, with some component of sinistral-slip; an average 0.17 m of sinistral slip occurred for every 1.00 m of dip slip (Crone and Machette, 1984; Scott et al., 1985; Crone et al., 1987). Average vertical displacement along the fault scarp was 0.8 m, with a maximum of 2.7 m (Smith and Arabasz, 1991). No foreshocks were recorded in the area before the earthquake (Richins et al., 1987). Much of the 1983 faulting followed preexisting (pre-1983 Holocene) scarps and shear zones, supporting the theory that the Lost River fault, or at least the fault segment ruptured in 1983, is subject to “characteristic earthquakes” or large earthquakes of similar magnitude and similar surface displacement (Crone, 1987; Keller and Pinter, 2002). The earthquake nucleated at a depth of about 16 km, and aftershocks dominantly occurred on adjacent secondary structures instead of on the fault plane of the initial rupture (Smith and Arabasz, 1991). The ruptured area of the fault occurred along the Thousand Springs

segment during the primary seismic event, though aftershocks occurred on the Warm Spring and Challis segments as well (Crone et al., 1987). The fault is planar and dips about 45° to 53° southwest – a listric geometry is not supported by geodetic data (Smith and Arabasz, 1991).

Normal faults in the Cordillera of southwestern Montana and east-central Idaho have been interpreted as listric at depth based on seismic data, including the faults in the northern section of the Lemhi Range (Tysdal, 2002). However, there is no historical evidence for nucleation or slip on low-angle or listric faults in the entire Intermountain Seismic Belt (Smith and Arabasz, 1991). The segment of the Lost River fault that ruptured during the 1983 Borah Peak earthquake is interpreted to have an average dip of 40° to 50° to the southwest from the surface fault trace based on aftershock hypocenters (Richins et al., 1987).

Intermountain Seismic Belt

The Intermountain Seismic Belt (ISB) is a roughly north-trending 100-200 km zone that spans at least 1500 km from northern Arizona to northwestern Montana; the ISB is characterized by shallow (usually <20 km) spatially diffuse seismicity and is a classic area for the study of intraplate extensional tectonics (Smith and Sbar, 1974; Smith and Arabasz, 1991; Humphreys and Coblenz, 2007). The ISB is thought to be a region that has undergone continuing deformation since the late Precambrian (Smith and Arabasz, 1991). The Lemhi Range is located within the central ISB, which extends from 42.25° to 45.25° N (encompassing the region from southern Idaho and western Wyoming to southern Montana), and includes the parabolic deformation zone associated with the

Yellowstone-Snake River Plain hotspot (Smith and Arabasz, 1991, see “Yellowstone Hotspot and the Snake River Plain” below). The overall stress field in the western part of the central ISB is defined by ENE-to-ESE-trending S_{hmin} (minimum horizontal compressive stress) orientations (Smith and Arabasz, 1991). Surface rupture in the ISB is thought to occur only for earthquakes with M_L above 6.0-6.5 (Smith and Arabasz, 1991). The central ISB differs from the southern and northern ISB in that it contains the two largest historical surface-faulting events (1959 Hebgen Lake and 1983 Borah Peak earthquakes) and is dramatically influenced by the Yellowstone-Snake River Plain system (Smith and Arabasz, 1991).

From 1900-1985, the 1988 Seismicity Map of North America (see Engdahl and Rinehart, 1991; also known as the DNAG catalog) records 49 earthquakes of magnitude 5.5 or greater in the Intermountain region (Smith and Arabasz, 1991). Other than the 1983 Borah Peak, ID earthquake, historic earthquakes in the ISB include: the M_W 6.6 1934 Hansel Valley, Utah earthquake (Smith and Arabasz, 1991); an earthquake swarm of 2,281 felt earthquakes during 14 months starting in 1935 near Helena, Montana, (including two $M > 6$) which damaged more than half the city’s buildings and killed four people (Smith and Sbar, 1974; Qamar and Stickney, 1983; Smith and Arabasz, 1991); and the M_S 7.5 (previously assigned magnitude = M 7.1 [Smith and Sbar, 1974; Qamar and Stickney, 1983], see Smith and Arabasz, 1991, p. 200) Hebgen Lake, Montana earthquake, which created vertical displacement of up to 6 meters on the Red Canyon fault as well as displacement on the Hebgen and Madison faults, caused the Madison Canyon Landslide and killed 28 people (Qamar and Stickney, 1983; Smith and Arabasz, 1991). In the Lost River, Lemhi, and Beaverhead areas, future faulting is most likely

along the high-relief central segments of the Lemhi and Beaverhead faults that have ruptured in latest Quaternary time and along the Mackay segment of the Lost River fault (Scott et al., 1985). Analysis of historic earthquakes in Montana and the surrounding regions of Idaho and Wyoming indicated that a $M \geq 5$ event is expected every 1.5 years, a $M \geq 6$ event every 10 years, and a $M \geq 7$ event every 77 years in western Montana and central Idaho, with the highest recurrence rates occurring around southwest Montana (Qamar and Stickney, 1983). An analysis covering the entire ISB estimated inter-event times of 6 years for magnitude 6.0 earthquakes and 17 years for magnitude 6.5 earthquakes (Smith and Arabasz, 1991).

In early studies of the ISB, the belt was described as a boundary between North American subplates, separating the Basin and Range province from the Colorado Plateau-Middle Rocky Mountains and the Northern Rocky Mountains (Smith and Sbar, 1974; Smith, 1977). Unlike other lithospheric plates, “subplates” are not completely surrounded by seismic activity, and the seismic activity that does occur is characterized by a much lower return rate of large earthquakes than along plate boundaries (Smith and Sbar, 1974; Smith, 1977). While fault plane solutions in the northern ISB (in the terminology of Smith and Sbar, 1974, the northern ISB includes the northern Utah – Yellowstone area) indicate predominately normal faulting on moderately dipping faults, faults to the south (Colorado Plateau – Great Basin Area) tend to show higher-angle reverse or normal sense movement (Smith and Sbar, 1974). Perhaps a decrease in extension away from Yellowstone is the cause (Smith and Sbar, 1974). North of the Yellowstone caldera, the trend of the entire Intermountain Seismic Belt swings to the northwest – south of the caldera the belt is oriented more north-south (Smith and Sbar, 1974). The Snake River

Plain is aseismic in contrast to the ISB (Piety et al., 1992), though earlier work suggested ongoing deformation due to aseismic creep (Smith and Sbar, 1974). The two major zones of shear and extension in the western U.S. (southern Great Basin and northern Interseismic Belt) are thought to accommodate convergence between the Pacific and North American plates in a manner similar to areas of lateral extrusion/tectonic escape associated with the convergence of the Indian and Eurasian plates (Smith, 1977).

The area of the Intermountain Seismic Belt located between the Hebgen Lake and Borah Peak earthquakes has been termed the “Centennial Tectonic Belt” (Stickney and Bartholomew, 1987; Gorton, 1995; Petrik, 2008). Alternatively, the area from Hebgen Lake to the Stanley Basin in central Idaho (or even to eastern Oregon; Smith and Sbar, 1974) is known as the Idaho Seismic Zone and separates the Northern Rocky Mountains from the Great Basin (Smith and Sbar, 1974; Qamar and Stickney, 1983). More recently, Smith and Arabasz (1991) have grouped the Idaho Seismic Zone back with the ISB, noting that the correlative (and likely causal, see “Yellowstone Hotspot and the Snake River Plain” below) relationship with the Yellowstone-Snake River Plain hotspot ties the zone to the main ISB (Smith and Arabasz, 1991). The recent division of the ISB splits the region into three parts: the northern, central and southern ISB (Smith and Arabasz, 1991). Historical seismicity in central Idaho indicates a NNE-SSW extensional strain with horizontal strain rate on the order of $3.3 \times 10^{-16} \text{sec}^{-1}$ and deformation rates of 2.0 mm/yr (Smith and Arabasz, 1991).

Evidence for recent faulting in the ISB in nearby ranges includes offset of surfaces from the most recent glacial period in the Centennial and Teton Ranges of Montana and Wyoming (Figure 1). In the Centennial Range, average surface offset for

Pinedale deposits (12,000 – 14,000 years old, see “Pleistocene Glaciations” below) is 6.7 m (estimated to have been generated by 2-3 large earthquakes), while the average surface offset for Bull Lake deposits is 11.3 m (estimated to have been generated by 3-4 large earthquakes). The estimated number of earthquakes using surface offset is based on the historically observed surface offset of 1.5 – 3 m resulting from large ($>7.0 M_w$) magnitude earthquakes in the Basin and Range Province (Petrik, 2008). Pinedale glacial deposits in the Teton Range of Wyoming contain evidence for up to 30 m of vertical surface offset since the late Pleistocene (Pierce and Morgan, 1992; Hampel et al., 2007).

Seismicity and the Idaho National Engineering and Environmental Laboratory

An additional concern regarding seismicity in the Lemhi region is the presence of the Idaho National Engineering and Environmental Laboratory (INEEL) site (Gorton, 1995; Hemphill-Haley et al., 2000). The INEEL is located immediately to the southeast of the southern extent of the Lost River and Lemhi Ranges. The 2,300 km² area was designated the National Reactor Testing Station in 1949, established by the U.S. Atomic Energy Commission (later the U.S. Department of Energy) for peace-time atomic energy research as well as defense programs (Nace et al., 1972; Bartholomay et al., 2002). The name of the facility changed to the Idaho National Engineering Laboratory in 1974, and again in 1997 to the current name (Bartholomay et al., 2002). Historic waste-disposal practices included chemical- and radiochemical-bearing wastewater discharge until 1983 to ponds and wells located at the facility, as well as burial of solid and liquid chemical and radiochemical wastes (Nace et al., 1972; Bartholomay et al., 2002). Hydrogeological studies have focused on known and potential migration of the contaminants in the Snake

River Plain aquifer and possible mitigation efforts (Nace et al., 1972; Bartholomay et al., 2002). Long-term concerns of safe disposal of hazardous materials at the Radioactive Waste Management Complex also hinge on proper information regarding land surface erosion, including the aggradational and erosional history of sediments within the complex (Dechert et al., 1993).

Yellowstone Hotspot and the Snake River Plain

The eastern Snake River Plain (SRP, Figure 1) is a linear 90-km wide trench filled with calderas subsequently covered by basalt flows (Pierce and Morgan, 1992). The axis of the SRP trends northeast (N52E). Silicic volcanism has been ongoing for about 16 million years and has progressively moved northeastward at an average rate of about 3 cm/yr to its present location in the Yellowstone area (Scott et al., 1985; Pierce and Morgan, 1992). The modern extension regime and Yellowstone-Snake River Plain volcanism began between 10-15 million years ago (Smith and Arabas, 1991).

While some workers have argued diverse origins of the Yellowstone thermal anomaly, including an eastward-propagating rift zone, localized volcanism along a crustal flaw, and a meteorite impact, the most likely origin is a deep-seated mantle plume (Pierce and Morgan, 1992). In the model of Pierce and Morgan (1992), around 16-17 Ma, a mantle plume head with a diameter of about 300 km encountered the base of the lithosphere near the Idaho/Oregon/Nevada border. Around 10 Ma, the plume chimney of smaller diameter (10-20 km) broke through the stagnating plume head near American Falls, Idaho. As the North American plate has moved southwest over the plume, associated volcanism has progressed northeastward (Pierce and Morgan, 1992).

The active seismic region located in close proximity to the Yellowstone volcanic province has inspired numerous studies and tectono-magmatic models of the regional deformation patterns (Smith and Sbar, 1974; Smith, 1977; Rodgers et al., 1990; Smith and Arabasz, 1991; Pierce and Morgan, 1992; Fritz and Sears, 1993). Yellowstone area seismicity is greatly influenced by magmas, partial melts, and hydrothermal activity (Smith and Arabasz, 1991). Upper-crustal blocks in Yellowstone are deformed due to these influences as well as the regional tectonic stress field (Smith and Arabasz, 1991). Further from the present location of the hotspot, in areas where the hotspot was previously located, seismicity patterns have been linked with passage of the hotspot. For instance, changes in recurrence intervals of large paleoearthquakes along the Swan Valley segment of the Grand Valley fault (Figure 1) have been attributed to passage of the Yellowstone hotspot and associated caldera formation on the Snake River Plain (Piety et al., 1992).

In early studies, thermal uplift on the shoulders of the migrating volcanic front was been compared to the “bow wave” or wake of a moving boat (Scott et al., 1985). According to Scott and others (1985), if the thermal upwelling is related to tectonic activity in the region, a period of higher faulting activity should be spatially and temporally associated with the thermal front. As the thermal anomaly moves on and the area cools, tectonic activity should decrease (Scott et al., 1985). In detail, as the thermal wave approaches, fault segments closest to the front will first undergo brittle failure, then a period of relative quiescence once the crust is heated. Brittle failure will increase just in advance of the thermal front. As the hotspot passes and heat dissipates, brittle failure will resume again (Haller, 1988).

The ranges adjacent to the SRP to the northeast display the most Quaternary activity within a belt on the order several 10's to 100 km from the SRP margin, termed the "Subsidence Shoulder" (Smith and Arabasz, 1991). In contrast, the zones of high Quaternary activity in the Intermountain Seismic Belt ranges to the south of the Teton Range are 100-200 kilometers from the SRP margin (Scott et al., 1985). The two zones of activity converge in the Yellowstone area and define a double parabola (Scott et al., 1985; Smith and Arabasz, 1991). Near the plume, deformation and associated seismicity and volcanism have been predicted to be highest, with predicted migration of this seismic region to the east as the North American plate continues to move west (Smith and Sbar, 1974). Thus, the Intermountain Seismic Belt can be considered a weak zone of lithosphere resulting from the Yellowstone plume passage where east-west extension is accommodated (Smith and Sbar, 1974). The exact mechanism of influence of the Yellowstone hotspot has been debated – but potential thermomechanical changes to the lithosphere could be due to lateral variations in deviatoric stresses, lithospheric subsidence, high temperatures, or loss of strength of upper-crustal material (Scott, 1985; Smith and Arabasz, 1991). Passage of the Yellowstone hotspot resulted in lateral changes of deviatoric stress, influencing seismogenic potential at distances of more than 100 km away from the Snake River Plain (Smith and Arabasz, 1991).

A slightly more recent model of faulting in the wake of the Yellowstone hotspot defined three belts of seismic activity, with a fourth belt containing inactive faults (Pierce and Morgan, 1992). The three seismically active belts subparallel the eastern SRP, with Belt I furthest from the SRP, Belt III closest to the SRP, and Belt II located between Belts I and III (Figure 11, Pierce and Morgan, 1992). Differences between the model of Pierce

and Morgan (1992) and earlier models are likely due to earlier models using historical seismicity while Pierce and Morgan used geologic evidence in addition to historical seismicity to define their zones (Pierce and Morgan, 1992). The belts form parabola or V-shape that converges on Yellowstone. The Lemhi Range has segments in each of the three belts of seismic activity (Figure 11).

Belt II is the most active belt with faults active since 15 ka, and range front relief > 700 m and is interpreted to be culminating in activity. The 1959 Hebgen Lake and 1983 Borah Peak earthquakes were located in Belt II (Pierce and Morgan, 1992). Belt I contains new (or recently reactivated faults) that have only generated less than 200 m of range front relief and are in early stages of activity. Belt III contains faults active during 15-120 ka that have range front relief of more than 500 m and are waning in activity. Belt IV only exists on the southern side of the SRP; range-bounding faults are currently inactive but were active during the late Tertiary, this belt is interpreted as having completed its cycle of tectonic activity (Pierce and Morgan, 1992).

In addition to seismic activity variations between the belts, Pierce and Morgan (1992) note elevation and uplift differences as well. A crescent-shaped area is elevated 0.5-1 km higher than adjacent areas – the crescent is about 350 km across and extends more than 400 km from the apex of the crescent, located just NE of Yellowstone (Pierce and Morgan, 1992). Pierce and Morgan suggest that the axial drainage divide separating the Pahsimeroi and Little Lost River valleys (Figure 5) may have origins related to the Yellowstone hotspot. Variations in the relative length of Bull Lake and Pinedale advances (see “Pleistocene Glaciations” below) reflect uplift occurring between the glaciations, including in the central Lemhi Range. Typically, Pinedale advances were

shorter in length than Bull Lake advances; where the opposite relationship is observed locally, uplift between the two glaciations is inferred (Pierce and Morgan, 1992).

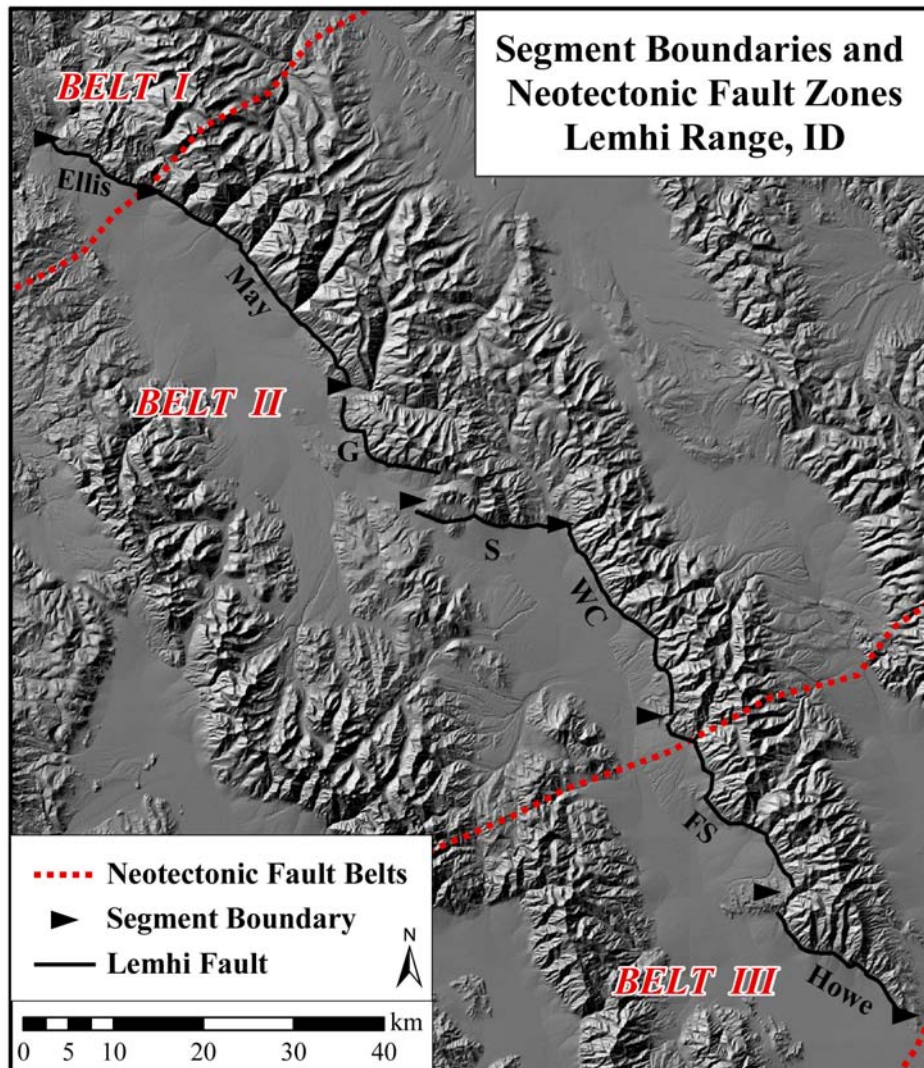


Figure 11. Segment boundaries used in this study and belts of seismic activity related to the passage of the Yellowstone hotspot (neotectonic fault belts generalized from Pierce and Morgan, 1992; segment boundaries from Janecke, 1993). Fault segments from south to north: Howe, Fallert Springs (FS), Warm Creek (WC), Summerhouse (S), Goldberg (G), May, and Ellis. Belt I is characterized by recently (Holocene offset) active faults with little (<200 m) relief, Belt II is characterized by recently (Holocene offset) active faults with large (>700 m) relief, Belt III is characterized by older (late Pleistocene offset) faulting with intermediate (>500 m) relief (Pierce and Morgan, 1992). Note that Belt IV is located only south of the eastern Snake River Plain, and is thus south of the area mapped above. North of the eastern Snake River Plain, only Belts I-III are recognized (Pierce and Morgan, 1992).

Climate

Modern Climate

The modern climate in the study area is continental arid. For the period of August 1948 – December 2005, average annual precipitation in May (Pahsimeroi Valley, Figure 5) was 7.77 inches (19.7 cm), including 19.2 (48.8 cm) inches of annual average snowfall. Mean January minimum temperature was 6.4°F (-14.2°C), and mean July maximum temperature was 86.0°F (30°C) (Western Regional Climate Center, original values reported in inches and °F). Regional snowpack telemetry (SNOTEL) data show precipitation to be approximately 75 cm/yr with annual mean temperature of ~0.5°C at 2800 m elevation (Johnson et al., 2007).

Pleistocene Glaciations

Pinedale and Bull Lake are names for the two most recent glacial episodes in the Rocky Mountain region (Crone and Haller, 1991). Most segmentation studies of the Lemhi and surrounding areas have used similar dates for glaciation-related events as the following: regional deglaciation around 14,000 years ago, Pinedale glaciation 15,000-30,000 years ago (with “younger Pinedale” around 15,000 years ago and “older Pinedale” 25,000-30,000 years ago), and Bull Lake around 140,000 years ago (Crone and Haller, 1991; Pierce and Morgan, 1992). These dates were based on early regional glacial studies including Pierce and others (1976); for a more recent discussion of Pleistocene glaciations and the ages of associated deposits in the Rocky Mountains, see Pierce (2004).

Periods of Fan Deposition and Glacial/Interglacial Cycles

Major periods of fan deposition and relative inactivity during the late Quaternary in southeastern Idaho have been noted, with particular study of the Lost River Range fans (Pierce and Scott, 1982; Baltzer, 1990; Gorton, 1995). Fan aggradational periods have been linked with glacial climates and fan incision or entrenchment periods with interglacial climates (Pierce and Scott, 1982; Ritter et al., 1995; Leeder and Mack, 2001). During glacial periods, peak meltout occurred later in the season. This resulted in longer melt days with higher solar angle than an earlier meltout and thus yielded higher runoff, with an estimated increase in peak stream discharge of at least an order of magnitude (Pierce and Scott, 1982). Glaciated and unglaciated valleys were both affected by this phenomenon (Pierce and Scott, 1982). However, this increase was not the result of increased precipitation, because precipitation was likely lower in the region during the last glacial period (Pierce and Scott, 1982). Rather, lower temperatures resulted in more precipitation occurring as snow (thicker spring snowpack) as well as decreased evaporation, transpiration, and sublimation (Pierce and Scott, 1982). Increased periglacial activity likely produced more coarse-grained material in mid-high elevation drainage basins (including those in the Lemhi Range) as evidenced by patterned ground, stone stripes and rock glaciers that are largely inactive in the modern climate (Pierce and Scott, 1982; Johnson et al., 2007). During interglacial periods, lower peak discharge led to lower sediment load and thus erosion of previous fan deposits (Pierce and Scott, 1982; Baltzer, 1990). The same pattern of more expansive Pinedale deposits being incised during the Holocene with late Pinedale – Holocene deposition of smaller fans near the

mountain front was described for the western Lemhi fans (Gorton, 1995). Some workers argue that the transition period of deglaciation may represent the time interval with greatest water discharge available for geomorphic work (Patterson, 2006; Koppes and Montgomery, 2009). Furthermore, the deglacial transition can also represent the highest sediment yields from glacierized catchments, due to thinner ice and warmer basal temperatures and increased flow at the bed of glaciers (Koppes and Montgomery, 2009).

Cycles of aggradation and erosion on an alluvial fan in the San Joaquin Valley, California have been attributed to changes in accommodation space on the fan driven by glacial-interglacial fluctuations in the ratio of sediment supply to water discharge (Weissmann et al., 2002; Weissmann et al., 2005). Koppes and Montgomery (2009) concluded that the end of a glacial period/start of an interglacial period corresponded to a period of fan incision. Incision was interpreted to have occurred as a result of decrease in the ratio of sediment supply to discharge or as a result of an increase in stream power (Weissmann et al., 2002; Weissmann et al., 2005). Periods of widespread fan aggradation have been correlated with glacial periods and a higher ratio of sediment to discharge (Weissmann et al., 2002; Weissmann et al., 2005). The fan intersection point and deposition shifted downfan (more distal) during periods of incision, and shifted up fan (more proximal) during periods of aggradation, when the sediment to discharge ratio increased (Weissmann et al., 2002; Weissmann et al., 2005). Periods of incision were marked by widespread eolian reworking and soil development on the perched proximal fan surfaces (Weissmann et al., 2002; Weissmann et al., 2005). The fan intersection point was noted as the transition between negative accumulation space up fan and positive accumulation space downfan (Weissmann et al., 2002; Weissmann et al., 2005).

Other workers argue for an erosion-deposition threshold on alluvial fans that can have different responses to climatic changes in drainage basins with differing boundary conditions (Bull, 2000). In some cases (hot desert environments), aggradation occurs with the onset of interglacial climate, while in other environments (high-altitude or high-latitude), aggradational cycles occur during full glacial times (Bull, 2000). Climatic changes influence vegetation and thus sediment production from drainage basins; in some cases this may supersede the importance of changes in water discharge from the basin (Pope and Wilkinson, 2005). In the Plio-Pleistocene Palomas half graben in the southern Rio Grande rift, New Mexico, sediment yield from drainage basins was lowest during wetter climates due to increased vegetation density and associated sediment storage. During drier climates, the stored sediment was not protected by vegetation and thus was flushed out of the catchments (Mack and Leeder, 1999).

On a time scale longer than glacial/interglacial transitions, some workers invoke the late Cenozoic trend of increased snowcover in the Rocky Mountains to explain erosion of upwards of 1.5 km from intramontane basins since the middle Miocene (Pelletier, 2009). For areas with elevation between 1.5 and 3 kilometers, snowmelt became an increasingly large contribution to total river discharge during the climate cooling of the late Miocene and Plio-Quaternary (Pelletier, 2009). Increased snowcover may have had similar impacts for mid- to high-elevation mountains worldwide, providing a greater flood magnitude and hence bedload sediment flux during seasonal snowmelt periods (Pelletier, 2009). Before the late Miocene, temperatures were warmer and seasonal snowmelt may have only been a large contributor to runoff in areas where elevation is greater than 3 km above sea level (Pelletier, 2009). During glacial periods,

the timing and magnitude of meltout was altered, with meltout of elevations greater than 3 km in July or August, and mid-elevations melting out in June (Pierce and Scott, 1982). The highest runoff periods during glacial periods of basins of 2 km average elevation are thus likely comparable to modern runoff from basins of 3 km average elevation (Pierce and Scott, 1982). Both size of the snowpack and warm-season temperature affect the meltwater flux; a larger snowpack provides a greater volume of water, while warmer temperatures ensure faster meltout and more intense flooding (Pelletier, 2009).

Janecke (1994) suggests the more humid Paleogene climate might explain the contrast in observed facies between the Eocene-Oligocene basins of the east-Idaho rift zone and those of late Cenozoic half-graben forming in arid regions. The Paleogene basins were characterized by coarse dominantly footwall-sourced deposits far into the basin with little modification by axial systems and minor preservation of fine-grained deposits (Janecke, 1994).

Climate-related Changes in Seismicity

Changes in climate have also been correlated with changes in seismic activity along normal faults (Hetzel and Hampel, 2005; Hampel et al., 2007). In a normal fault system, the maximum principal stress, σ_1 , is perpendicular to the Earth's surface and the minimum principal stress, σ_3 , is horizontal and is the direction of extension (Hetzel and Hampel, 2005; Hampel et al., 2007). Glacial loading alone cannot explain the observed tendency of several normal faults (including the Teton fault, Wyoming and the Wasatch, West Valley, Oquirrh and Stansbury faults, Utah) to undergo increased seismic activity following deglaciation, as glacial loading would increase σ_1 and fault slip rate (Hetzel

and Hampel, 2005). However, glacial loading also causes lithospheric flexure which increases σ_3 , decreasing differential stress ($\sigma_1 - \sigma_3$) and slip rate (Hampel et al., 2007). Isostatic rebound due to glacial unloading then causes increased slip rates on normal faults following deglaciation. The effect of isostasy explains the seemingly contradictory correlation of glacial unloading with increased fault slip rates (Hetzel and Hampel, 2005). In the case of the Utah faults, the drainage of Lake Bonneville was an additional factor influencing postglacial isostatic rebound, and thus, fault slip (Hetzel and Hampel, 2005). These studies imply that earthquake recurrence rates based on a simply dividing the elapsed time by the number of seismic events may not accurately capture earthquake clustering behavior (Hampel et al., 2007).

Climate, Tectonics, and Erosion: Coupled Systems with Complex Responses

Changes in climate and tectonics both control alluvial fan morphology, and much of the alluvial fan literature centers on debate over which is the dominant factor over different spatial and temporal scales (Friedmann and Burbank, 1995; Ritter et al., 1995; Mack and Leeder, 1999; Hartley et al., 2005; Harvey, 2005; Harvey et al., 2005; Pope and Wilkinson, 2005). Fan progradation and aggradation are often associated with glacial periods, and decreased deposition attributed to interglacial times (Pierce and Scott, 1982; Patterson, 2006). Deposition during interglacials may occur downfan of deposition that occurred during the previous glacial period, as sediment bypasses the upper fan reaches and flows occur in entrenched channels (Ritter et al., 1995); or deposition during interglacials may be concentrated proximal to the range front, sometimes associated with

a change in depositional processes (Patterson, 2006). Other workers correlate changes in fan deposits with deformation and uplift. The split in opinion is often between geomorphologists (different surfaces correlated with climate changes) and sedimentologists (gravel packages associated with different periods of tectonic activity) (Mack and Leeder, 1999). The split may reflect the order of magnitude difference in time scale ($>10^4$ yr for tectonic influences, 10^2 – 10^4 yr for climatic influences) the studies consider (Leeder and Jackson, 1993; Allen and Densmore, 2000; Harvey et al., 2005; Patterson, 2006). One implication for alluvial fan interpretation in the rock record is that in a case where drainage basin bedrock lithology is not distinctive, the transition between intervals dominated by debris flow deposits and intervals dominated by sheetflood deposits could be attributed to a climatic change (affecting water and sediment discharge from a single source basin) or attributed to the overlapping of deposits sourced from multiple drainage basins with contrasting dominant processes (Patterson, 2006).

A study focusing on a specific process that operates on one time scale may completely neglect other pertinent influences that operate on other timescales – this is likely to happen frequently as climatic, tectonic, and geomorphic processes operate over several different temporal orders of magnitude (Bishop and Shroder, 2004a; Allen and Allen, 2005). However, since spatial and temporal scales in many systems are similar, the time span over which the most dominant processes shape morphology increases with system size (Shroder and Bishop, 2004). Tectonics may simply create the relief conditions (uplifted block and accommodation space) required for alluvial fan development, while climatic changes influence alluvial fan sequence deposition (Ritter et al., 1995; Weissmann et al., 2002; Pope and Wilkinson, 2005). Alternatively, tectonics

can be considered to play a more important role in determining fan setting, accommodation space, and catchment lithology-related variations in sediment supply rates (Harvey, 2005).

Despite the ongoing arguments of the relative influence of climatic, tectonic and erosional factors (Beaty, 1961; Funk, 1977; Ritter et al., 1995; Finlayson et al., 2002), some workers envision less of a cause-and-effect relationship between these factors, and instead suggest a quasi-equilibrium state of balance with complicated feedbacks (Molnar and England, 1990; Keller and Pinter, 2002; Allen and Allen, 2005; Roe et al., 2006; Willett et al., 2006; Molnar, 2009). For instance, Molnar uses the following mnemonic inspired by the perfect gas law ($PV = nRT$): “**P**recipitation rate x **V**ertical component of velocity = **n**umber of floods per annual precipitation x **R**ock removal rate x **T**ectonic movement” (p.44, Molnar, 2009) as a conceptual model of how the processes interact. Evidence of uplift can often be confused with evidence of mean elevation change or even evidence of climate change, making cause-and-effect relationships difficult or impossible to assign conclusively (Molnar and England, 1990).

Relief is the energy source for flowing water, while lithology is a control on weathering, sediment production, and fluvial erosion rates (Bull, 2000). Crustal thickening and the associated isostatic response, as well as faulting, folding, and relative elevation increases resulting from base level lowering all increase channel gradients and rates of fluvial erosion and sediment transport (Willett et al., 2006). Uplifted areas cause changes in patterns of precipitation, in turn influencing fluvial processes. As topography reaches high elevations, glaciation ensues and affects erosion regimes (Willett et al.,

2006). Erosion redistributes mass and thus can alter tectonic processes, and erosion can alter climate via carbon cycle dynamics (Willett et al., 2006).

The dataset in this study is estimated to incorporate influences exerted on the tectonic time scale ($>10^4$ yr, Leeder and Jackson, 1993; Allen and Densmore, 2000), since tectonic activity and drainage basin evolution have been interpreted to act on similar time scales to affect changes in fan depositional facies and stacking patterns (Harvey et al., 2005; Leleu et al., 2005). A more detailed examination of estimated time scale of analysis is located in the discussion section.

Previous Work Delineating Fault Segments Along the Lemhi Range

Following the 1983 Borah Peak earthquake, the USGS Earthquake Hazards Reduction Program funded a segmentation study of the Lemhi fault (Baltzer, 1990). Multiple workers mapped fault segments along the Lemhi Range using a variety of methods (Scott et al., 1985; Haller, 1988; Turko, 1998; Baltzer, 1990; Crone and Haller, 1991; Turko and Knuepfer, 1991; Janecke, 1993; Gorton, 1995). While each study yielded a slightly different segmentation model for the fault, some boundaries were more commonly agreed upon (Figure 10). The difference in identified boundaries is likely due to the different time scales of rupture elucidated by the various methods, as well as increasing availability of data as more studies were completed. When differentiating fault segments based on rupture history, dating resolution of offset surfaces inherently limits what can be resolved – ruptures of adjacent segments must occur at times sufficiently

greater than the uncertainty in dating offset deposits in order to be resolved as separate events (Turko and Knuepfer, 1991).

Many of the segment boundaries coincide with gravity highs and many of the interiors of segments coincide with gravity lows (Haller, 1988; Crone and Haller, 1991), indicating that the location of fault segment boundaries along the Lemhi Range may have persisted through several seismogenic cycles (Haller, 1988). Gravity lows indicate thick basin fill (Crone and Haller, 1991).

This study will use the segment boundaries as defined by Janecke (1993). Janecke used a combination of segment boundaries from multiple workers (Haller, 1988; Turko, 1988; but primarily Baltzer, 1990; Crone and Haller, 1991; and Turko and Knuepfer, 1991), as well as her own mapping (Janecke 1992a; Janecke 1992b) to define segment boundaries along the Lemhi fault (Janecke, 1993).

Individual Study Descriptions

Haller (1998) based segment definitions on changes in along-strike fault-scarp morphology, without detailed structural or bedrock geology mapping. The main factors under consideration were variations in: 1) offset of contemporaneous deposits; 2) scarp-height versus maximum-slope-angle relation; and 3) degree of scarp preservation (Haller, 1988). The study notes that while some boundaries are located near footwall structural features, others are not associated with any known feature or change in range-front geometry (Haller, 1988). One boundary associated with a structural element is the Howe-Fallert Springs boundary, which is located near a thrust fault in the footwall block. All six of the segments mapped by Haller (Figure 10 shows segments, named from south to

north: Howe, Fallert Springs, Sawmill Gulch, Goldburg, Patterson, and May) have scarps on alluvium and likely ruptured within the past 30,000 years. Haller stated that a segment boundary could possibly exist near the Sawmill Canyon embayment, but there was not sufficient evidence based solely on scarp morphology to indicate a segment boundary. One possible explanation for the similar scarp morphology across the potential segment boundary is that the adjacent segments ruptured separately, but within 1000 years of each other. This case is indistinguishable from the case where there is no segment boundary and the entire length ruptured simultaneously (Haller, 1988). The Sawmill Gulch segment has evidence of two faulting events in the last 15,000 – 20,000 years (the most recent in the middle Holocene), with as many as six in the last 100,000 – 150,000 years (Haller, 1988). The Goldburg segment has evidence of three faulting events in the last 50,000 years, the most recent of which occurred around 15,000 years ago (Haller, 1988). Haller (1988) also mapped segment boundaries in the Lost River, Beaverhead, and Tendoy ranges. However, Haller's segment boundaries for the Lemhi Range stand in contrast to those identified by the more localized studies of Turko (1988) and Baltzer (1990). Notably, Haller's (1988) segment delineation does not include the May-Ellis division included by Baltzer (1990) and Turko and Knuepfer (1991).

Turko (1988) used fault scarp diffusion modeling (scarp profile degradation) and cluster analysis to determine the location of segment boundaries, as well as an analysis of mountain-front morphology. This approach allowed multiple time scales to be considered. Using the youngest event scarps ($10^3 - 10^4$ year time scale), six or possibly eight segments were identified (Turko, 1988). Multiple-event scarps ($10^4 - 10^5$ year time scale) yielded as many as four segments, while mountain-front morphology ($10^5 - 10^6$

year time scale) indicated four segments (Turko, 1988). The longer time scale segment determinations are the same as some of the youngest-event determinations, perhaps indicating that only a few of the segment boundaries have persisted through multiple earthquake rupture cycles (Turko, 1988). Turko's (1988) unnamed segments include a three-fold division of Baltzer's (1990) May segment under Turko's eight-segment model.

Baltzer (1990) delineated segments along the northern 80 km of the Lemhi Range based on surficial mapping of late Quaternary surfaces and exploratory trenching across fault scarps offsetting key surfaces. Four segments were identified, from north to south: Ellis, May, Big Gulch and Warm Creek (Figure 10, Baltzer, 1990). The segments were initially based on previously identified segments (Turko, 1988), which were field checked and critical localities were trenched. Four age groups were assigned using a combination of relative dating techniques, dating of similar deposits in nearby areas by other workers, and a single radiocarbon date; these age groups were designated: Holocene (0-12 ka), late Pinedale (12-30 ka), early Wisconsin (30-60 ka), and pre-Wisconsin (>80 ka) (Baltzer, 1990). The older three surfaces correspond to depositional periods of activity related to glacial climates as identified by Pierce and Scott (1982). Trenching revealed the following rupture history: late Pinedale rupture at Warm Creek (Warm Creek segment); early Holocene rupture at Summerhouse Creek (Big Gulch segment); and Holocene rupture at Falls Creek (May segment) and at Warm Creek (Baltzer, 1990). Haller (1988) indicated that a segment boundary might exist at Sawmill Canyon, but the rupture history was considered indistinguishable between two cases: 1) Sawmill Canyon is not a segment boundary; and 2) Sawmill Canyon is a segment boundary and the segments on either side of Sawmill Canyon ruptured within 1000 years

of each other. In agreement, Baltzer (1990) found evidence for either one large event, or two smaller clustered events at Summerhouse Canyon. However, trenches at Warm Creek, on the other side of the possible Sawmill Canyon boundary from Summerhouse Canyon, display two distinct events separated by a soil-forming interval – a different history than that of Summerhouse Canyon, and thus an indication that Sawmill Canyon is indeed a segment boundary (Baltzer, 1990). A combined interpretation of offset of late Quaternary surfaces (Baltzer, 1990) with scarp profile data (Turko, 1988) is that the northern four segments of the Lemhi Range (May, Ellis, Big Gulch, and Warm Creek) have been persistent for at least the last 60 ka (Baltzer, 1990).

Crone and Haller (1991) defined segments along the Lost River and Lemhi ranges, and Beaverhead and Tendoy Mountains based on fault scarp morphology, age of offset deposits, and range front morphology with largely the same interpretations as Haller (1988). The Lemhi segments mapped in their study from south to north are: Howe, Fallert Springs, Sawmill Gulch, Goldburg, Patterson, and May (same as Haller, 1988). Most recent movement on individual segments ranges from late Pleistocene to middle Holocene (Crone and Haller, 1991). The Sawmill Gulch and Patterson segments have evidence of middle Holocene rupture, while the other segments most recently ruptured in the late Pleistocene. Crone and Haller (1991) stated that the Goldburg segment boundaries might be revised following further study.

Turko and Knuepfer (1991) recognized six segments of the Lemhi fault, from south to north: Howe, Fallert Springs, Warm Creek, Big Gulch, Falls Creek, and Ellis (Figure 10). Most of their proposed segment boundaries are defined to within 1 km (Turko and Knuepfer, 1991). In a refinement of Turko's (1988) segmentation model, the

boundary between Ellis and Falls Creek is located at the contact between Eocene volcanic rocks and Proterozoic quartzite and siltstones just north of Tater Creek (Turko and Knuepfer, 1991).

Janecke (1993) investigated cross faults located at potential segment boundaries in the Lost River and Lemhi Ranges. Seven segments were recognized, from south to north: Howe, Fallert Springs, Warm Creek, Summerhouse, Goldburg, May and Ellis (Figures 10, 11). Using three-dimensional structural models, some preexisting (crosscutting) normal faults were demonstrated to intersect range front faults at hypocentral depths. For the northern part of the Lemhi Range, Janecke's (1993) model follows the model of Baltzer (1990) closely, except Baltzer's Big Gulch segment is split into the Goldburg and Summerhouse segments. Field mapping and observations of the bedrock high and zone of distributed faulting located at the axial drainage divide (splitting the Pahsimeroi Valley and Little Lost River Valley) indicate that the zone may be a potential segment boundary zone, though detailed paleoearthquake history reconstruction may not support the independence of these two segments (Janecke, 1993). Haller (1988) identified this zone as a segment boundary; Baltzer (1990) and Turko and Knuepfer (1991) did not. The Warm Creek-Summerhouse boundary coincides with the Paleogene basin-bounding north-striking Sawmill Canyon normal fault. While Sawmill Canyon fault activity was dominant in the Paleogene, some Quaternary reactivation has occurred, and may represent accommodation of some seismic events along central segments of the modern range-bounding Lemhi fault (Janecke, 1993). Haller (1988) indicated a segment boundary might exist at Sawmill Canyon; while Baltzer (1990) and Turko and Knuepfer (1991) mapped a segment boundary at Sawmill Canyon. The Ellis-

May boundary is likely influenced by the Allison Creek normal fault, which near the Lemhi fault strikes NE, is nearly vertical and has had down to the NW movement (Janecke, 1993). The fault brings Eocene volcanic rocks into contact with Proterozoic metasedimentary rocks and was likely formed in Eocene-Oligocene time (Janecke, 1993). Three-dimensional structural modeling of cross faults proposed to control segment boundary location of major range-bounding faults is necessary because the two-dimensional map view of nonvertical faults does not reflect the location of the fault plane at depth, and thus does not reflect how the cross faults may interact with range-bounding faults at seismogenic depths (Janecke, 1993).

Gorton (1995) mapped segments of the southern half of the Lemhi fault (south of Sawmill Canyon). Similar to the methodology of Baltzer (1990), late Quaternary surfaces were mapped and faults scarps offsetting those surfaces were profiled and trenched to delineate segments (Gorton, 1995). A number of relative dating techniques, as well as thermoluminescence dating were used to classify surfaces into age groups (Gorton, 1995). Seven age groups of surfaces were identified: Holocene, late and early Pinedale, early Wisconsin, and three groups of pre-Wisconsin age (Gorton, 1995). Gorton (1995) identified three segments – a northern segment corresponding to the Warm Creek segment of Turko and Knuepfer (1991); a central segment corresponding to the northern portion of the previously defined Fallert Springs segment (Haller, 1988; Crone and Haller, 1991; Turko and Knuepfer, 1991); and a southern segment corresponding to the southern portion of the previously defined Fallert Springs segment and the entire Howe segment (Haller, 1988; Crone and Haller, 1991; Turko and Knuepfer, 1991). The segmentation model proposed by Gorton (1995) moves the segment boundary between

the southern two segments from the South Creek Block north to around Coyote Springs. The main reason for moving the segment boundary was an observed 15-17 km gap along the range front lacking scarps on late to middle Pinedale deposits – this gap defined the central segment equivalent to the northern half of the previously defined Fallert Springs segment (Gorton, 1995). The discordance between segmentation models regarding the potential segment boundary at the South Creek block can possibly be explained as follows: 1) the block serves as a “leaky” segment boundary, or 2) the block has not been a persistent boundary – acting as a rupture barrier during some earthquake cycles and not acting as a rupture during others (Gorton, 1995). A leaky segment boundary does not completely stop rupture, but allows some rupture on the neighboring segment during a seismic event (Crone and Haller, 1991; Piety et al., 1992).

A more recent seismic hazard study of the southern two geomorphically defined segments (Howe and Fallert Springs; Crone and Haller, 1991; Turko and Knuepfer, 1991) used thermoluminescence dating, radiocarbon dating, and soil development observations at four locations to determine the timing of the most recent event and two or three previous events at each site (Hemphill-Haley et al., 2000). The four sites were located along the two southern segments defined by previous workers. Results indicated that some seismic events may have ruptured at least the northern portion of the Howe segment (and potentially the entire Howe segment) together with at least the southern half of the Fallert Springs segment (Hemphill-Haley et al., 2000). Late Pleistocene events may have ruptured the entire Fallert Springs segment, while during the Holocene, only the northern section of the segment ruptured (Hemphill-Haley et al., 2000). With the exception of this possible Holocene rupture of the northern portion of the Fallert Spring segment,

segmentation models of the earlier geomorphic studies are permissible given the newer data bracketing timing of seismic events (Hemphill-Haley et al., 2000).

Preferred Segmentation Model

This study will use the model of Janecke (1993) with segment boundaries for the seven segments (north to south: Ellis, May, Goldberg, Summerhouse, Warm Creek, Fallert Springs, and Howe) located as follows (Figure 11). The Ellis-May boundary is located at the Proterozoic/Eocene lithologic contact near Tater Creek. The May-Goldburg boundary is near Big Creek. The Goldberg-Summerhouse boundary is located at the axial drainage divide separating the Pahsimeroi Valley from the Little Lost River Valley. The Summerhouse-Warm Creek boundary is located at the mouth of Sawmill Canyon. The Warm Creek-Fallert Springs boundary is located at the Horse Creek block. The Fallert Spring-Howe boundary is located at the South Creek block (Janecke, 1993). Two segment boundaries are agreed upon by all studies – the May-Goldburg boundary near Big Creek and the Warm Creek-Fallert Springs boundary at the Horse Creek block (segment terminology of Janecke, 1993).

From the Warm Creek-Fallert Springs segment boundary northward, Janecke's (1993) model represents incorporation of the most field mapping, trending data, gravity data, and structural analysis of subsurface locations of faults and interactions. Additional work conducted by Gorton (1995) and Hemphill-Haley and others (2000) suggested possibly moving the Fallert Springs-Howe boundary north from the South Creek block to near Coyote Springs. This was based on thermoluminescence dating of material in offset surfaces as well as along-strike changes in fault scarp morphology. However, the original

location of the Fallert Springs Howe boundary (Janecke, 1993) was considered “permissible” given the new data (Hemphill-Haley et al., 2000). This study will use the boundaries of Janecke (1993) without the alternate modification suggested by Gorton (1995) and Hemphill-Haley and others (2000), but this has an acknowledged potential impact on the calculation of drainage basin morphology as a function of along-strike position relative to closest segment boundary.

METHODOLOGY

This study is an application of geographic information science (GIS) to test geomorphologic hypotheses. In “Geographic Information Science and Mountain Geomorphology”, written in part as a response to the United Nations International Year of Mountains 2002, Bishop and Shroder make a call for researchers to link GIS concepts with theoretical geomorphology (Bishop and Shroder, 2004a). Studies integrating these two fields will provide increased understanding of space-time interactions of surface processes. New geomorphic hypotheses can be tested using GIS data analysis techniques (Bishop and Shroder, 2004a). Conversely, exploratory spatial analysis can inform new geomorphic hypotheses based on observed patterns (Bishop and Shroder, 2004a). “GIScience not only serves as a foundation for development of new technology and technology transfer, it also potentially enables scientific formalizations through computer-based research to produce new geomorphic knowledge and better understandings of the geodynamics of mountains” (p. xiii, Bishop and Shroder, 2004b).

A comparison of along-strike variation in footwall drainage basins necessitates a delineation of drainage basin boundaries, extraction of appropriate morphometric parameters from each basin, and quantification of basin locations relative to fault segment boundaries. Alluvial fans must be mapped using airphotos. Fan morphometric parameters are extracted from elevation data, and then the fan morphometric parameters can be associated in a relational database with the mapped alluvial fans.

Digital elevation data can be stored in several different data structures, including triangulated irregular networks and regular rectangular grids (Mark and Smith, 2004). The term *digital elevation model* (DEM) can be used to refer to any digital elevation dataset, regardless of the data structure (Mark and Smith, 2004). The U.S. Geological Survey (USGS) produces different elevation products, including “digital elevation models” and “digital terrain models”, both of which are stored in regular grid format but which differ in compilation (Mark and Smith, 2004). Regardless, elevation data are fundamentally stored, processed, and displayed using a field (rather than object) data structure. In this paper, “DEM” will refer to regular rectangular gridded elevation data in general, which in this project is sourced from USGS “digital elevation model” products.

Data Sources and Accuracy

The elevation data used in this study are a subset of the USGS National Elevation Dataset (NED). The DEM data are supplied as $1^{\circ} \times 1^{\circ}$ tiled grids of 1/3-arc-second (approximately 10-meter) resolution. The source data for the DEMs in the study area are 10-meter resolution. Error is reported as a probability that the given spatial coordinate is within a certain distance of the true location (Rasemann et al., 2004). The most recent vertical accuracy assessment available is from the 2003 version of the NED; vertical accuracy was reported as root mean square error 2.44 m or 3.99 m for the 90% confidence interval of National Map Accuracy Standards (Gesch, 2007).

Uncertainty in landform evolution analysis can be epistemic – due to lack of understanding of geologic processes and relationships; or, aleatory – due to random variations in the data (Hanks, 2000). Using high-resolution high-accuracy DEM data

reduces aleatory uncertainty, since the effect of any given pixel in a high-resolution dataset is very small compared to a low-resolution dataset. Systematic errors are common in DEMs as a result of methods used to generate the DEM (Rasemann et al., 2004). For instance, when contour maps are used as base data, computer interpolation of contours can result in step-like features and other artifacts (Rasemann et al., 2004).

All data used in this project are available online for free. Data processing was conducted primarily using the free open-source System for Automated Geoscientific Analysis (SAGA), although ArcMap was used for initial data preparation and final data analysis. SAGA was chosen primarily due to the flexibility of selecting specific flow algorithms and transparency of processes used by watershed modules. Statistical analysis was conducted using the free open-source statistical package R. See Appendix A for data and software sources.

Data Preprocessing

NED tiles were downloaded, reprojected into Idaho Transverse Mercator (projection parameters, Appendix C, mosaiced and clipped to the extent of the area of interest in the ArcMap environment. The grid data were then imported into SAGA.

The source DEM was observed to have a striping error on the scale of 10 by 10 pixels in the N-S and E-W directions. This striping is a frequent artifact of elevation data compilation, and is a vertical error that leads to parallel linear ridges and depressions (Perego, 2009). Since striping can alter flow paths, flow routes, and thus calculated drainage patterns, the striping was smoothed using the Directional Average 1 module for SAGA by Perego (2009). This module calculates the average elevation along the length

of each stripe, and then compares the difference in elevation between adjacent stripes (depressions and ridges). The difference between the average elevations of adjacent stripes is the striping error and is removed from the DEM, effectively “smoothing” the data (Perego, 2009).

In order to properly define watersheds, the DEM needs to be hydrologically correct and water outflow from each pixel of the DEM needs to eventually route to an outlet on the border of the DEM. Sinks (pits or surface depressions) in the DEM are localized, small features that “catch” flow, prevent it from exiting the DEM, and thus lead to the identification of spurious drainage subbasins (Wang and Liu, 2006).

Depressions in DEMs can be data artifacts or can be actual landscape features (Moore et al., 1991). Data artifacts can result from rounding or truncating elevation data values to lower precision, interpolation of values, or simply data input errors. Depressions representing landscape features can be related to karst, recent glacial activity, mass wasting, and landscape alteration by humans (quarries, artificial ponds, etc.). Regardless of origin, sinks must be filled so that all cells will have downslope flow that eventually exits the DEM boundaries (DeMers, 2002; Wang and Liu, 2006). Sinks were filled using the SAGA Fill Sinks XXL module created by Wang and Liu (2006). This module uses a least-cost search algorithm to assign each cell a spill elevation value, which is defined as the minimum elevation necessary for flow to leave the cell and drain to the edge of the DEM. For all pixels that are not sinks, the spill elevation is the same as the actual elevation. Pixels that are in sinks will have higher spill elevations than true elevations, so that all flow leaves the cell. The resulting destriped and “filled” DEM is the input for the terrain analysis.

The smoothing and filling of the DEM necessarily alter some of the pixel values. To ensure that the alterations were not extreme, changes in elevation value between the original DEM and terrain analysis input (smoothed and filled) DEM were inspected. The change in elevation values can be split into standard deviations from the mean change. Values are reported for the entire clipped elevation area, which includes area adjacent to, but not included within, the formal study boundaries. The formal study boundary includes alluvial fans and their drainage basins on the western flank of the Lemhi Range (Figure 4). The mean change in elevation from the original to the modified DEM was 0.33 m, with a standard deviation of 3.4 m. Thus, 95% of elevation points were changed by no more than 7.1 meters. The range in elevation change was -91 m to +82 m, but pixels with values changed so drastically can be considered outliers. Visual inspection of the spatial distribution of changes to elevation revealed no areas of concern within the study boundary. As expected, elevation changes are particularly aligned along horizontal and vertical lines due to destriping. The majority of clumped elevation changes occurred to the east of the main drainage divide (eastern flank of the Lemhi Range), out of the study boundary. The altered DEM was thus considered suitable for further use.

The Lemhi fault was digitized based on previous mapping (Haller, 1988; Turko, 1988; Baltzer, 1990; Crone and Haller, 1991; Turko and Knuepfer, 1991; Janecke, 1993; Gorton, 1995; Hemphill-Haley, 2000) and minor adjustments were made at a mapping scale of 1:24,000. These adjustments were based on the author's interpretation of airphoto and DEM data. Janecke's (1993) segment boundary definitions were used, but other workers provided more detailed mapping of segment interiors. The most detailed map was used as a guide for digitizing the fault in any given area. Additionally, different

workers mapped segment boundary locations at different scales. Several segment boundary locations were identified by multiple workers to be within a few kilometer wide zone. These discrepancies are minor compared to overall difference between contrasting models (e.g. Crone and Haller's (1991) Patterson-May boundary is far removed from the May-Ellis boundary identified by the other workers, Figure 10). Segment boundaries of one model that were within a few kilometers of a boundary from another segmentation model were considered to be at the same location. Although the boundaries were mapped at a larger scale (higher resolution) than depicted in this study and were thus placed at slightly differing localities by different workers, for the sake of comparison of overall segment models, these small variations were removed. For instance, Janecke (1993) places the Ellis-May segment boundary at the lithology change between Eocene volcanics and Proterozoic quartzite/siltstone just south of Tater Creek. Turko (1988) explicitly states that the May-Ellis segment boundary is not located at this lithology change, but rather approximately 1.5 miles northwest (along the range front) of the lithologic change. Later, Turko and Knuepfer (1991) revised the boundary to be located at the lithologic contact (same as Janecke, 1993). Given slight adjustments such as these of effectively the same segment boundary, for the small-scale (coarse resolution) comparison of varying segmentation models in the current study, these were considered to be the same boundary location. Additionally, several workers mapped a segment boundary zone rather than a discrete point. If this zone encompassed a boundary mapped by another worker as a discrete point, the zone is depicted where the second worker mapped a point boundary. One example is the Big Gulch/Warm Creek boundary zone of

Baltzer (1990), which is mapped in this study and cited by Janecke (1993) as the point boundary at Sawmill Creek.

Terrain Analysis and Morphometry

Drainage basins are delineated in part using overland flow simulation. Catchment areas (used here synonymously with drainage basins) are a measure of surface and shallow subsurface runoff from contributing area upslope (Moore et al., 1991). The area measured in this study is planimetric area.

The Standard Terrain Analysis module in SAGA (Conrad, 2006) was executed on the destriped (smoothed) and filled DEM. This module outputs numerous grids: analytical hillshade, slope, aspect, plan curvature, profile curvature, convergence index, curvature classification, catchment area, wetness index, stream power, LS-factor, channel network (in grid and shapefile formats), altitude above channel network, channel network base level, and watershed subbasins. Grids used in following steps include: catchment area, L-S factor, and channel network.

Basin Delineation

Alluvial fans form where confined, channelized flow leaves the drainage basin catchment and becomes unconfined after reaching the break in slope at the mountain front. This transition point is referred to as the drainage basin outlet, and corresponds to the transition from the erosional and depositional sectors of the sediment routing system (O'Callaghan and Mark, 1984). Outlet points were digitized based on the following factors. First, outlet points were digitized based on orthophoto imagery. Points were

located where the sediment routing system transitioned from the erosional sector (drainage basin) to depositional sector (alluvial fan). Then, outlet points were adjusted slightly (by no more than a few pixels) to correspond to a location along a calculated drainage path (calculated during the Standard Terrain Analysis). Since all calculated flow must exit the drainage basin through this point, the outlet must be located along the calculated drainage path; this ensured the entire upstream area would be included in the next steps. The L-S factor served as a final check on correctly identifying this transition. The dimensionless L-S factor is the product of slope length and slope steepness parameters (Van Remortel et al., 2001). Longer slope length and higher slope steepness result in higher overland flow velocities and thus higher erosion rates. Thus, the L-S factor is a suitable proxy for identifying the location where flow regime transitions from erosional in the drainage basin to depositional in the alluvial fan – this occurs where the L-S factor drops to less than one. Visual inspection of the L-S grid confirmed proper location of outlet points.

Drainage basins were created using the interactive Upslope Area module of SAGA (Conrad, 2006, see Moore et al., 1991) with identified outlet points as input. Drainage basins were numbered 1-43 from south to north (Figures 12-14, Table 4). During later vector-based table joins, basin number was used as the key to associate drainage basins with fan data, basin morphometric data and distances from segment boundaries.

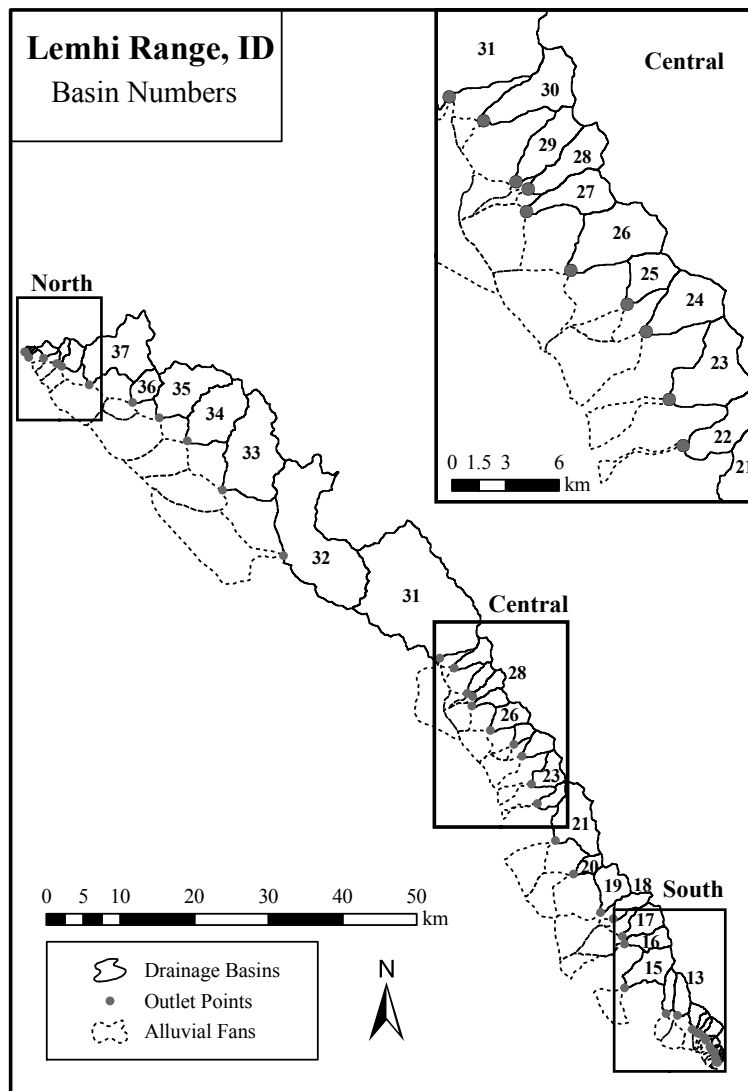


Figure 12. Assigned drainage basin numbers for the Lemhi Range. Basins were numbered 1-43 from the south to the north. Detail of area labeled “Central”, upper right; detail of area labeled “North”, figure 13; detail of area labeled “South”, figure 14.

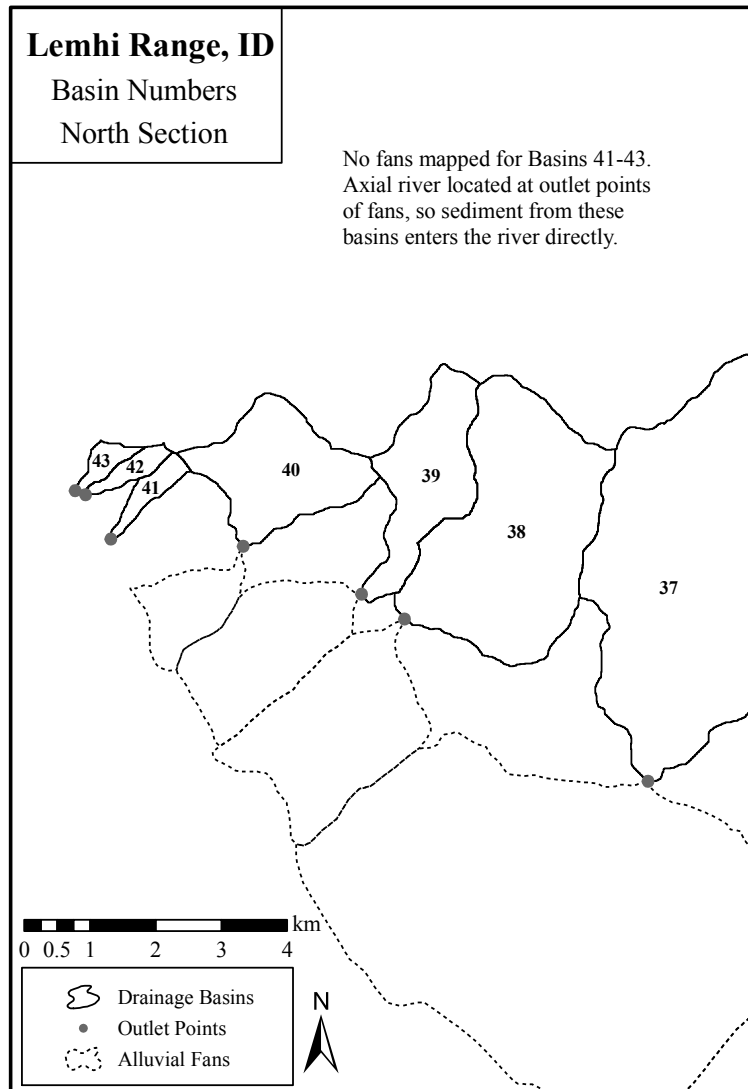


Figure 13. Assigned drainage basin numbers for the northern part of the Lemhi Range. Basins were numbered 1-43 from the south to the north. See Figure 12 for location of this figure relative to the entire study area. Gaps between mapped drainage basins (such as between basins 37 and 38) are due to the fact that only drainage basins that reach the main range drainage divide were considered in this study. Thus, the drainage basin that is located between basins 37 and 38 and its associated alluvial deposits were not considered in this study and are not depicted in this map.

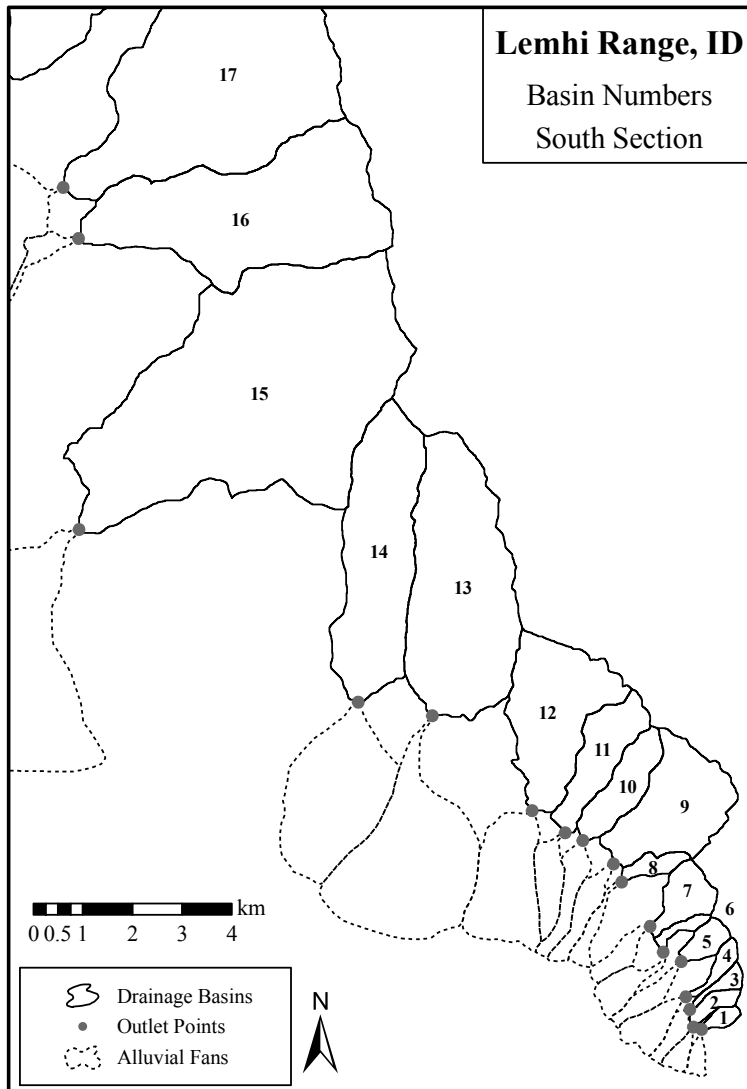


Figure 14. Assigned drainage basin numbers for the southern part of the Lemhi Range. Basins were numbered 1-43 from the south to the north. See Figure 12 for location of this figure relative to the entire study area.

Table 4. Basin names and basin numbers. These are the same numbers as the numbers in Figures 12-14. Only drainages that have names on 1:100,000 scale topographic maps are listed. Basins 1-12 and 39-43 are smaller than the basins listed below.

Basin Number	Basin Name	Basin Number	Basin Name
13	East Canyon	26	Basinger Canyon/Bell Mtn Creek
14	Middle Canyon	27	Mahogany Creek
15	South Creek	28	Meadow Creek
16	Camp Creek	29	(north of Meadow Creek)
17	North Creek	30	Warm Creek
18	Mormon Gulch	31	Sawmill Canyon
19	Uncle Ike Creek	32	Big Creek
20	Boulder Creek	33	Patterson Creek
21	Badger Creek	34	Falls Creek
22	Williams Creek	35	Morse Creek
23	Cedar Run Canyon	36	Tater Creek
24	Deep Creek	37	Morgan Creek
25	Black Creek	38	Ennis Gulch

Basin Area, Flow Path Length,
Elongation Ratio, and Melton Ratio

The Deterministic-8 (D8) flow algorithm was preferred instead of Deterministic- ∞ ($D\infty$, Tarboten, 1997) or Multiple Flow Directions (MFD, Freeman, 1991; Quinn et al., 1991) algorithms. Drainage is contributory in the fan drainage basins. The D8 algorithm works well for contributory drainage, and only one outflow direction per pixel is assigned, unlike other algorithms. Thus, for all flow calculations in the erosional sector (drainage basins) of the sediment routing system, the D8 algorithm was chosen over alternate flow algorithms.

Basins were exported as ASCII files and then imported into ArcMap as gridfiles and converted to polygons. Basin polygons were combined into a single shapefile. Planimetric area was calculated using built-in functionality.

Longest flow path length within a basin is the maximum transport distance for a particle to exit the drainage basin (i.e. the distance from the furthest ridgeline on the drainage divide with another basin or the main range divide to the outlet point). Flow path was determined by 1) calculating overland flow distance using the Overland Flow Distance to Channel Network SAGA module (Conrad, 2006); 2) finding the maximum overland flow distance for a given basin around its upper elevation margin; 3) digitizing that point as a seed; and 4) running the Interactive Downslope Area module for the input seed (Conrad, 2006) for each basin using the D8 flow algorithm (Figure 15). For each basin, the raster of longest flow path was exported as an ASCII file, imported into ArcMap, converted to polyline format, intersected with the corresponding basin, and planimetric length of the resulting line calculated using built-in functionality.

Elongation ratio was calculated for each basin as the ratio between the diameter of a circle with the same area as the basin and the length of the longest flow path (see “Basin Elongation Ratio, Hypsometry, and Melton Ratio” in Introduction chapter). Melton ratio was calculated as the ratio between basin relief and square root of basin area (see “Basin Elongation Ratio, Hypsometry, and Melton Ratio” in Introduction chapter).

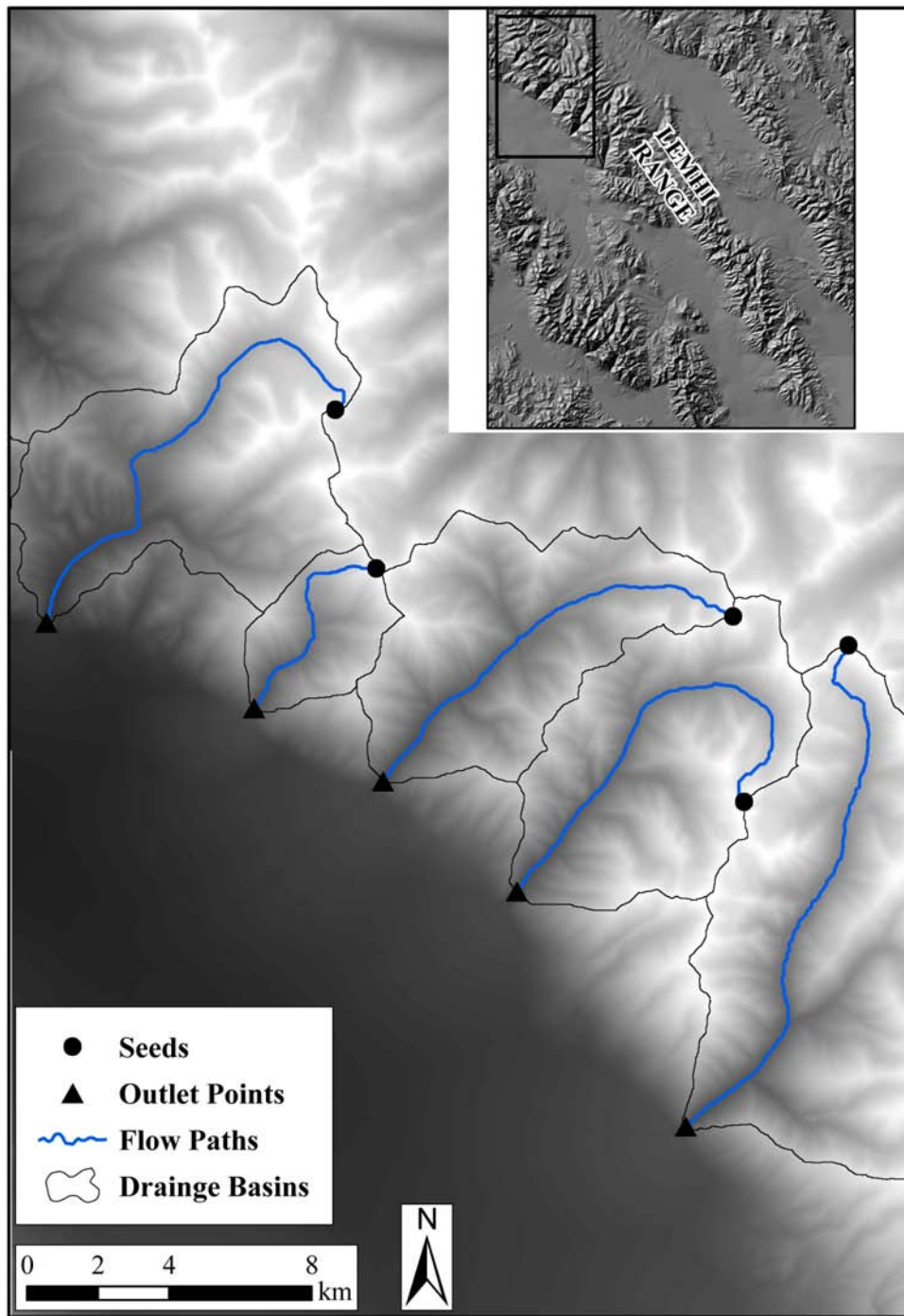


Figure 15. Flow paths for five basins. Outlet points (triangles) are located at the transition from erosion in the drainage basin to deposition on the alluvial fan. Seed (circles) are located where a particle leaving the basin would have the longest travel distance along calculated flow paths to reach the outlet. Background is elevation, with darker values indicating lower elevation areas and lighter values indicating higher elevation areas.

Fan Area

Fans were mapped on 1-m digital orthophotos in ArcMap using field mapping conducted by Baltzer (1990) and Gorton (1995) as a reference. In this study, all fans were mapped onto the orthophotos at resolutions of 1:15,000 or better. One fan was mapped per drainage basin – multiple lobes originating from the same drainage were considered the same fan. Thus, for an individual basin, the mapped fan represents the surficial area covered by deposits originating from that basin over several periods of aggradation and incision (see Appendix D for fans mapped on airphotos). Due to the coalesced nature of the fans and toe-cutting by axial streams, the mapped fans necessarily represent a minimum depositional area for each fan. Fan planimetric area was then calculated using the built-in functionality of ArcMap.

Drainage Basin Slope and Fan Slope

Drainage basin slope can be calculated in a few ways. In one method, drainage basin slope is calculated by simply dividing basin relief by basin length (Al-Farraj and Harvey, 2005; Harvey, 2005). This method is used primarily when mapping drainage basins on topographic maps. Fan slope is conventionally calculated using topographic maps as well (Milana and Ruzycki, 1999; Harvey, 2005). Slope is measured along a transect where contour lines are semicircular, from the fan apex to the downfan limit of deposition (Milana and Ruzycki, 1999), or along a transect on the proximal half of each fan (Harvey, 2005).

Another method is available when using DEMs instead of topographic maps. A slope grid can be generated and zonal statistics used to extract an average slope value for each drainage basin (DeMers, 2002). The latter method was initially used in this study – slope was calculated using zonal statistics with inputs of drainage basins (zones) and the smoothed, filled DEM. Fan slope was calculated in the same manner. Minimum, maximum, range, and mean slope were calculated for each drainage basin and fan, and slope was calculated as both percent slope and in degrees.

Since the two slope methods yield different results (see “Two Different Methods of Calculating Drainage Basin Slope” in Discussion chapter), drainage basin slope was also calculated using the conventional method – by dividing drainage basin relief by length. Relief was calculated using zonal statistics with inputs of the drainage basins and the corrected DEM. Basin length was previously calculated for flow path length and elongation ratio (see “Basin Area, Flow Path Length, Elongation Ratio, and Melton Ratio” above).

Basin Hypsometry

Hypsometry was calculated using the Hypsometry module in SAGA (Conrad, 2006) using constant elevation intervals (Harlin, 1978; Luo, 2000). Basin 35 was used as a test case to determine an appropriate number of elevation intervals. The hypsometric curve shape and hypsometric integral value stabilized above 500 classes (Figure 16). Hypsometric curves were generated for all basins using 1000 classes. Hypsometric curves were plotted in R, and the hypsometric integral was calculated as the area under the curve.

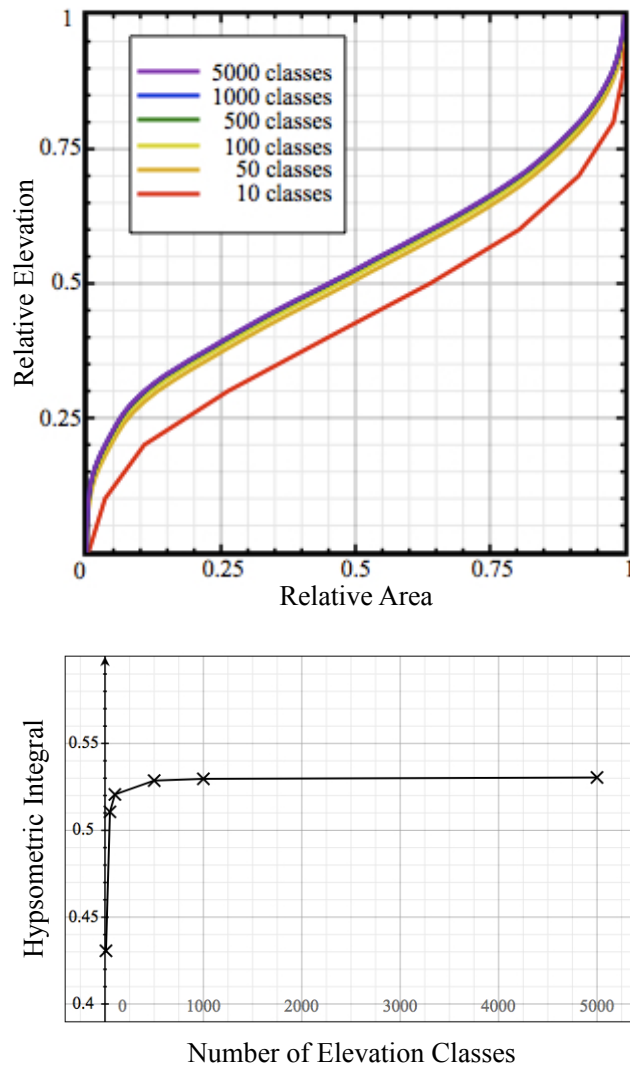


Figure 16. Determination of how many classes to use for hypsometric analysis. Above: hypsometric curve is very similar for $n > 100$ classes. Below: hypsometric integral stabilizes for $n > 500$ classes. Based on these results, 1000 classes were used to generate hypsometric curves for all basins.

Other Zonal Statistics: Elevation

Maximum elevation, minimum elevation, mean elevation, and relief (difference between maximum and minimum elevation) were calculated for each drainage basin

using zonal statistics with the mapped basins and the smoothed, filled DEM as input. The same methodology was used for extracting elevation statistics for each alluvial fan.

Distance from Segment Boundary

For each basin, outlet points needed to have a quantified location with reference to nearest fault segment boundary. Fault segment boundaries are not points on the Earth's surface, but rather diffuse zones of complex structure (Janecke, 1993). However, for the purpose of quantifying distance from segment boundary, segment boundaries had to be treated as discrete locations. Digitized fault traces based on prior work (Haller, 1988; Turko, 1988; Baltzer, 1990; Crone and Haller, 1991; Turko and Knuepfer, 1991; Janecke, 1993; Gorton, 1995; Hemphill-Haley, 2000) were edited in order to connect different traces of the same segment as one continuous line feature. These edited segments ended exactly at segment boundary points.

Each basin's outlet point location was then determined along the edited segment line that crossed through the area of the associated alluvial fan-drainage basin pair. Since outlet points did not fall exactly on edited segment lines, a 'projected' outlet point was determined by constructing a perpendicular line from the edited segment to the outlet point (Figure 17). The intersection of the perpendicular line with the edited segment was defined as the 'projected' outlet point, and the edited segment was cut at this point. Thus a segment was cut into two pieces for each basin outlet. The length of the shorter segment piece was assigned as the distance from segment boundary for each basin. Two basins, basin 31 and basin 32 (basin 32, Figure 18), had outlet points located within segment

boundary zones rather than along individual segments and thus could not accurately be assigned an along-segment distance from nearest segment boundary.

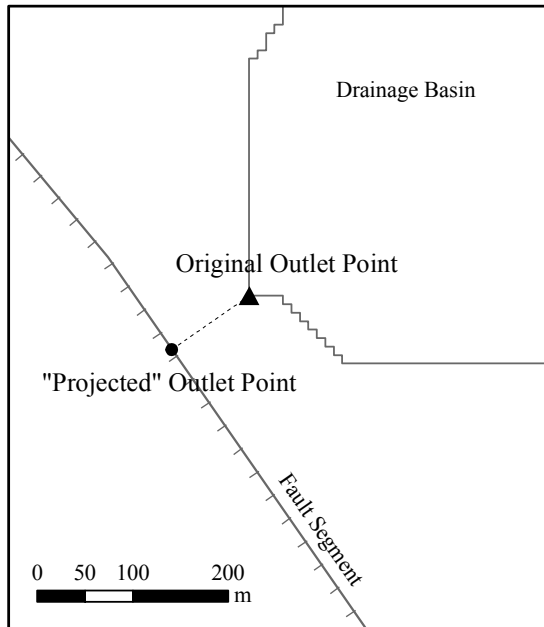


Figure 17. Each outlet point was “projected” to the nearest fault segment by constructing a perpendicular line (dashed) from the segment to the original outlet point. The “projected” point was then used to cut the fault segment into two pieces. The length of the shorter segment piece was assigned to the outlet point as distance from segment boundary. Note very different scale of the figure compared to maps in previous figures.

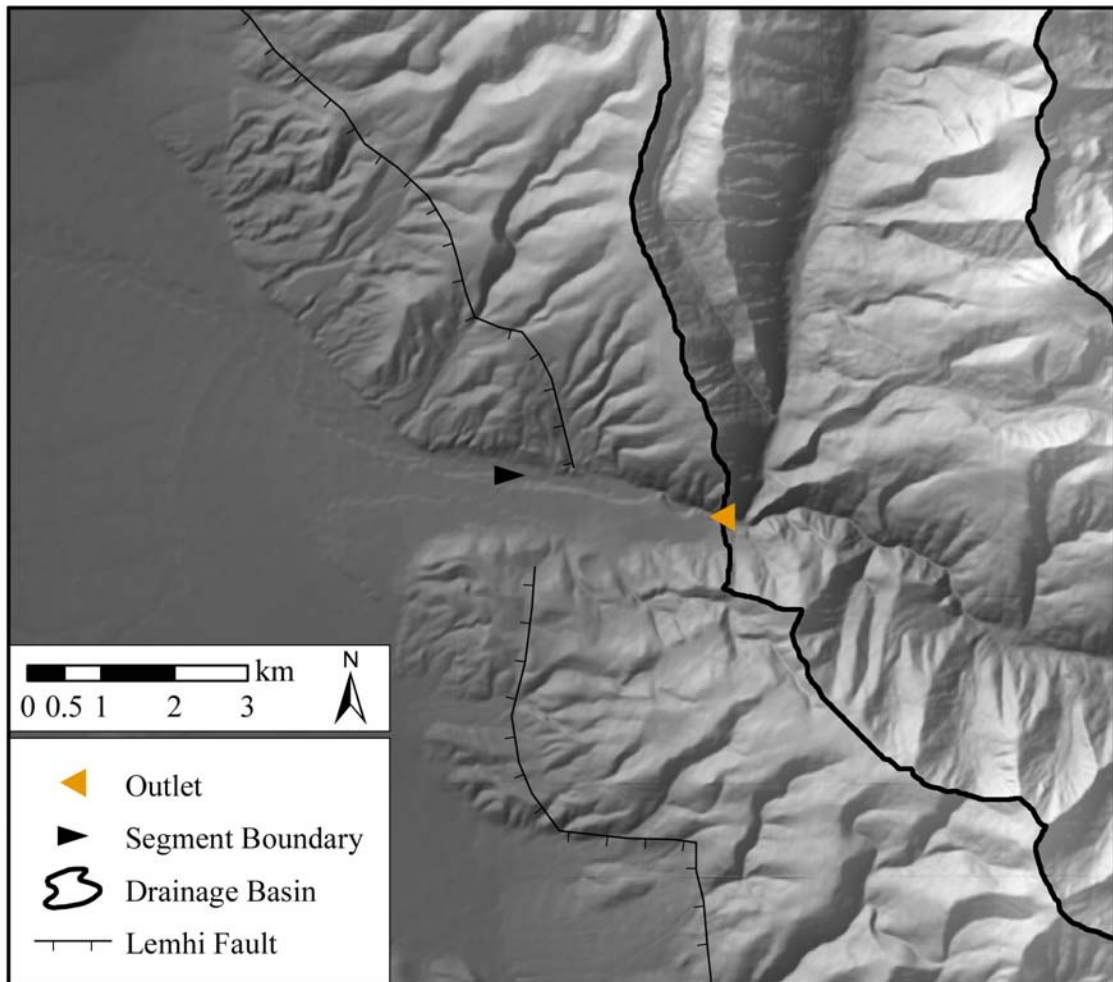


Figure 18. Basin 32. Outlet is located within segment boundary zone. Since the outlet cannot be assigned to either of the two segments of the Lemhi Fault that are closest, the drainage basin has no calculated distance from nearest segment boundary.

Standard Regressions

Standard regression was carried out using the statistical package R. Simple linear regressions include: elongation ratio, hypsometric integral, Melton ratio and fan-area to drainage-area ratio individually considered with respect to distance from nearest segment boundary. The morphometric parameters were considered as response variables and distance from segment boundary as the causal variable. For the entire range, fan area was

considered with respect to drainage area, fan slope with respect to drainage area, and drainage slope with respect to drainage area. Following previous work, standard linear regressions were performed on log-transformed drainage area, fan area, drainage slope, and fan slope (Milana and Ruzycki, 1999; Al-Farraj and Harvey, 2005; Harvey, 2005; Leleu et al., 2005; Lin et al., 2009).

RESULTS

General Statistics for Drainage Basin and Fan Areas and Slopes

Footwall topography is defined by the 43 drainage basins that reach the main drainage divide. Basins are numbered 1-43, south to north along strike of the range-bounding fault system (Figures 12-14, Table 4). Alluvial fans were mapped for basins 1-40. Fans were not mapped for basins 41-43 because the outlet points are very close (within a few pixels) to the axial river and alluvial fans are not well developed. The axial river transports sediment produced by drainage basins 41-43 away from the drainage basin outlets, precluding the development of alluvial fans. Measurements of drainage basin and fan area and slope are reported in detail in Appendix E.

Elevation values are reported to three significant digits. Maximum drainage basin elevation for the 43 individual basins ranges from 1,720-3,710 m, minimum drainage basin elevation ranges from 1,430-2,320 m, and mean drainage basin elevation ranges from 1,540-2,890 m. Basin area ranges from 0.231 to 190 km², with a mean area of 19.5 km². Mean basin slope (calculated on a pixel-by-pixel basis) ranges from 13.6 to 32.3 degrees (24.9 – 65.7 percent) for individual basins. Slope of individual drainage basins (calculated conventionally) ranges from 5.48 to 26.6 percent (discussion of different slope calculations is presented in “Two Different Methods of Calculating Drainage Basin Slope” in Discussion chapter).

The 40 mapped alluvial fans have maximum elevations ranging from 1,500 to 2,320 m, minimum elevations ranging from 1,430 to 1,970 m, and mean elevation

ranging from 1,470 to 2,050 m for individual fans. Fan area for individual fans ranges from 0.230 to 82.0 km², with a mean area of 11.4 km². Mean fan slope (calculated on a pixel-by-pixel basis) ranges from 0.886 to 7.26 degrees (1.55 to 12.9 percent).

Morphometric Parameters and Distance from Fault Segment Boundaries

Drainage basin parameters were calculated for all 43 basins. Alluvial fan parameters were calculated for the 40 mapped alluvial fans (associated with basins 1-40). Therefore, ratios involving only drainage basin parameters were calculated for all 43 basins, while ratios involving fan parameters were calculated for 40 basins (basins 1-40). The outlet points of basins 31 and 32 (basin 31 is Sawmill Canyon and basin 32 is Big Creek) were located within segment boundary zones (basin 32). Rather than assign these two basins an arbitrary distance from the nearest segment boundary (representing distance along the interior of a segment), they are excluded from regressions in which a morphometric variable is considered as a function of along-strike distance from segment boundary. Thus, the following regressions use 41 basins (basins 1-43 excluding 31 and 32) for morphometric parameters specific to drainage basins and 38 basins (basins 1-40 excluding basins 31 and 32) for morphometric parameters that also rely on fan data.

Fan Area to Drainage Basin Area Ratio

The ratio of fan area to drainage basin area as a function of distance to segment boundary was considered for 38 basins. Fan area to drainage area (ϕ) is inversely related to drainage basin outlet distance from the closest segment boundary (R-squared = 0.0294,

residual standard error = 0.592, p-value=0.304). The regression equation describing the relationship is:

$$\phi = 0.988 - 0.0312D$$

where ϕ = ratio of fan area to drainage area and D = distance from the nearest fault segment boundary (Table 5, Figure 19).

Table 5. Fan drainage basin and alluvial fan area in square kilometers. Fans were not mapped for basins 41-43 (see “General Statistics for Drainage Basin and Fan Areas and Slopes” above). Basin areas larger than 50 square kilometers are bolded.

Basin	Basin Area (km ²)	Fan Area (km ²)	Ratio Fan: Basin Area	Basin	Basin Area (km ²)	Fan Area (km ²)	Ratio Fan: Basin Area
1	0.231	0.230	0.999	21	39.6	24.2	0.612
2	0.354	0.279	0.788	22	7.00	1.44	0.205
3	0.275	0.352	1.28	23	13.6	5.30	0.388
4	0.350	0.746	2.13	24	10.2	13.0	1.27
5	0.525	0.520	0.992	25	5.59	2.81	0.502
6	0.259	0.679	2.62	26	14.1	12.8	0.907
7	1.09	0.502	0.459	27	7.25	10.2	1.40
8	0.450	0.399	0.886	28	7.52	1.12	0.149
9	4.01	0.768	0.191	29	7.31	3.77	0.516
10	1.68	0.817	0.487	30	11.0	1.65	0.150
11	2.20	0.912	0.414	31	190.	33.5	0.176
12	5.05	3.18	0.629	32	140.	82.0	0.586
13	10.1	7.18	0.709	33	80.7	37.8	0.468
14	7.69	5.37	0.698	34	49.0	28.7	0.587
15	23.2	13.5	0.581	35	47.3	44.7	0.945
16	12.0	9.19	0.765	36	13.4	9.25	0.690
17	15.9	15.6	0.983	37	57.3	28.5	0.497
18	10.2	1.49	0.146	38	8.48	5.44	0.642
19	19.5	23.2	1.19	39	3.19	3.84	1.21
20	7.04	18.6	2.64	40	3.50	1.58	0.452

Fan-Area to Drainage-Area vs. Distance from Segment Boundary

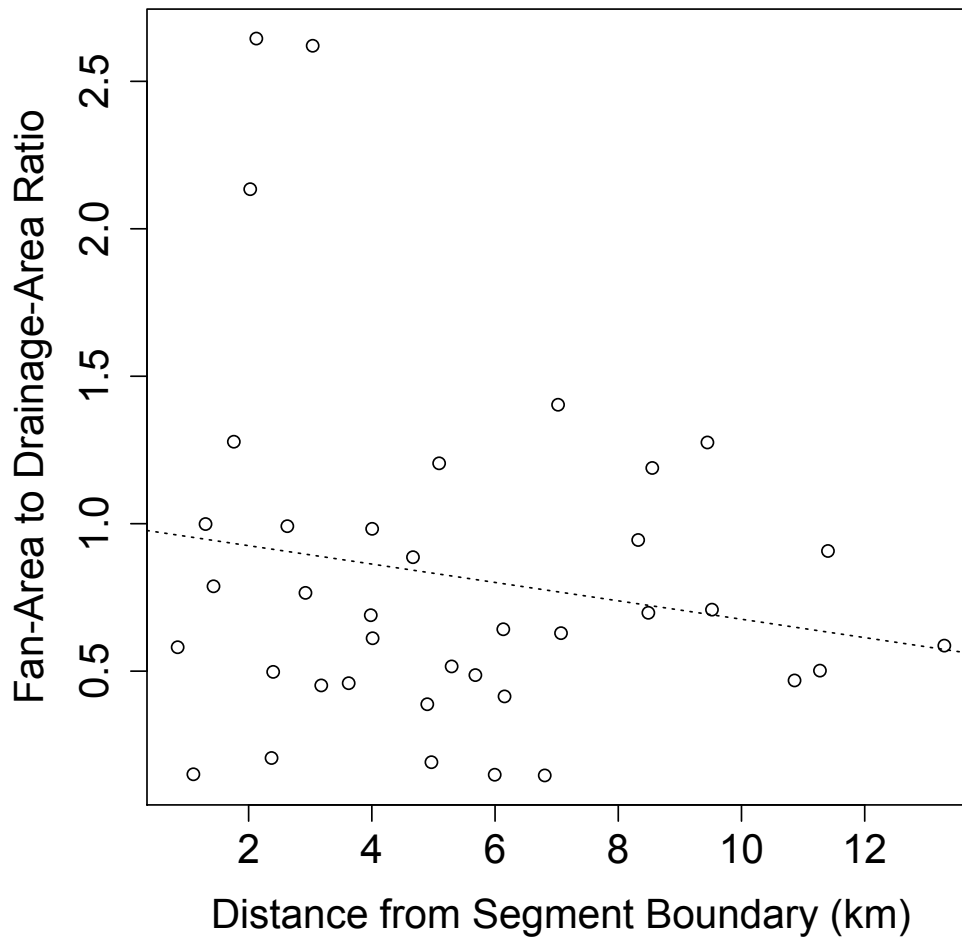


Figure 19. Fan area to drainage basin area ratio (ϕ) plotted against distance from segment boundary (D). Dotted line is regression line for standard linear regression: $\phi = 0.988 - 0.0312D$, R-squared = 0.0294.

Drainage Basin Elongation Ratio and Melton Ratio

Drainage basin elongation ratio as a function of distance to segment boundary was considered for 41 basins. Drainage basin elongation ratio is directly related to drainage basin outlet distance from the nearest segment boundary (R-squared = 0.0604, residual

standard error = 0.0830, p-value=0.121). The regression equation describing the relationship is:

$$ER=0.489+0.00615D$$

where ER = elongation ratio and D = distance from the nearest fault segment boundary (Table 6, Figure 20).

Table 6. Elongation ratio, Melton ratio and hypsometric integral for all drainage basins.

Basin	Elongation Ratio	Melton Ratio	Hypsometric Integral	Basin	Elongation Ratio	Melton Ratio	Hypsometric Integral
1	0.536	0.454	0.631	23	0.560	0.319	0.453
2	0.462	0.395	0.665	24	0.560	0.344	0.460
3	0.406	0.501	0.556	25	0.592	0.508	0.531
4	0.384	0.485	0.502	26	0.523	0.389	0.549
5	0.612	0.404	0.545	27	0.520	0.411	0.440
6	0.371	0.574	0.568	28	0.430	0.392	0.446
7	0.660	0.408	0.588	29	0.477	0.416	0.502
8	0.439	0.612	0.581	30	0.507	0.342	0.439
9	0.460	0.261	0.534	31	0.658	0.094	0.578
10	0.457	0.390	0.456	32	0.814	0.124	0.521
11	0.438	0.381	0.429	33	0.598	0.163	0.444
12	0.539	0.460	0.580	34	0.560	0.206	0.471
13	0.563	0.432	0.529	35	0.593	0.212	0.530
14	0.419	0.538	0.537	36	0.613	0.363	0.514
15	0.560	0.316	0.541	37	0.557	0.221	0.542
16	0.527	0.363	0.495	38	0.668	0.307	0.572
17	0.591	0.314	0.497	39	0.403	0.491	0.550
18	0.510	0.375	0.366	40	0.648	0.319	0.549
19	0.536	0.307	0.476	41	0.421	0.645	0.562
20	0.552	0.443	0.454	42	0.429	0.693	0.567
21	0.706	0.260	0.590	43	0.405	0.625	0.675
22	0.514	0.432	0.514				

Elongation Ratio vs. Distance from Segment Boundary

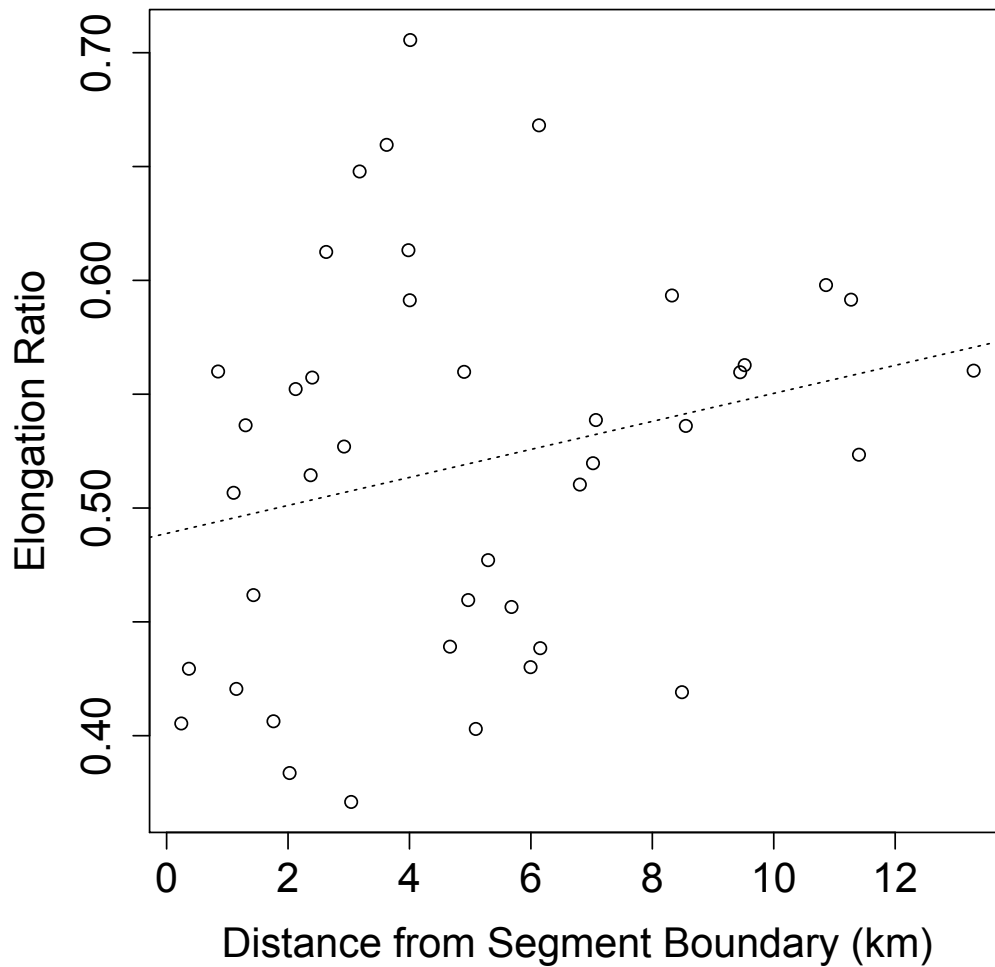


Figure 20. Elongation ratio (ER) plotted against distance from segment boundary (D). Dotted line is regression line for standard linear regression: $ER = 0.489 + 0.00615D$, R -squared = 0.0604.

Melton ratio as a function of distance to segment boundary was considered for 41 basins. Drainage basin Melton ratio is inversely related to drainage basin outlet distance from the nearest segment boundary (R -squared = 0.143, residual standard error = 0.114, p -value = 0.0149). The regression equation describing the relationship is:

$$MR = 0.470 - 0.0136D$$

where MR = Melton ratio and D = distance from the nearest fault segment boundary (Table 6, Figure 21).

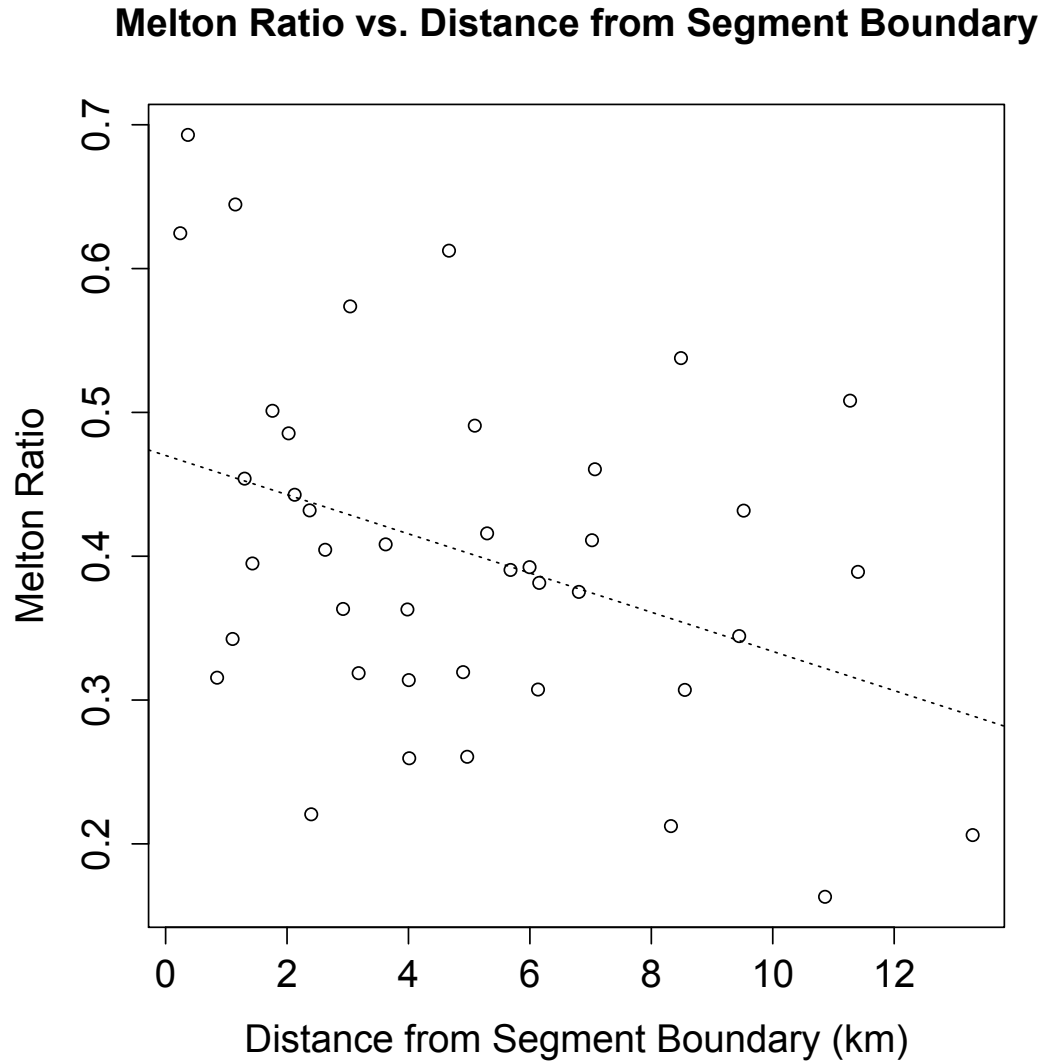


Figure 21. Melton ratio (MR) plotted against distance from segment boundary (D). Dotted line is regression line for standard linear regression: $MR = 0.470 - 0.0136D$, $R^2 = 0.143$.

Drainage Basin Hypsometric Integral and Hypsometric Curves

Drainage basin hypsometric integral as a function of distance to segment boundary was considered for 41 basins. Drainage basin hypsometric integral is inversely

related to drainage basin outlet distance from the nearest fault segment boundary ($R^2 = 0.150$, residual standard error = 0.0603, p -value=0.0122). The regression equation describing the relationship is:

$$HI = 0.560 - 0.00742D$$

where HI = hypsometric integral and D = distance from the nearest fault segment boundary (Table 6, Figure 22). Hypsometric curves for individual basins can be found in Appendix F. Drainage basins with outlets further from segment boundaries tend to have more sigmoidal-shaped hypsometric curves, while basins closer to segment boundaries have straighter or slightly convex-up hypsometric curves (Figure 23).

Hypsometric Integral vs. Distance from Segment Boundary

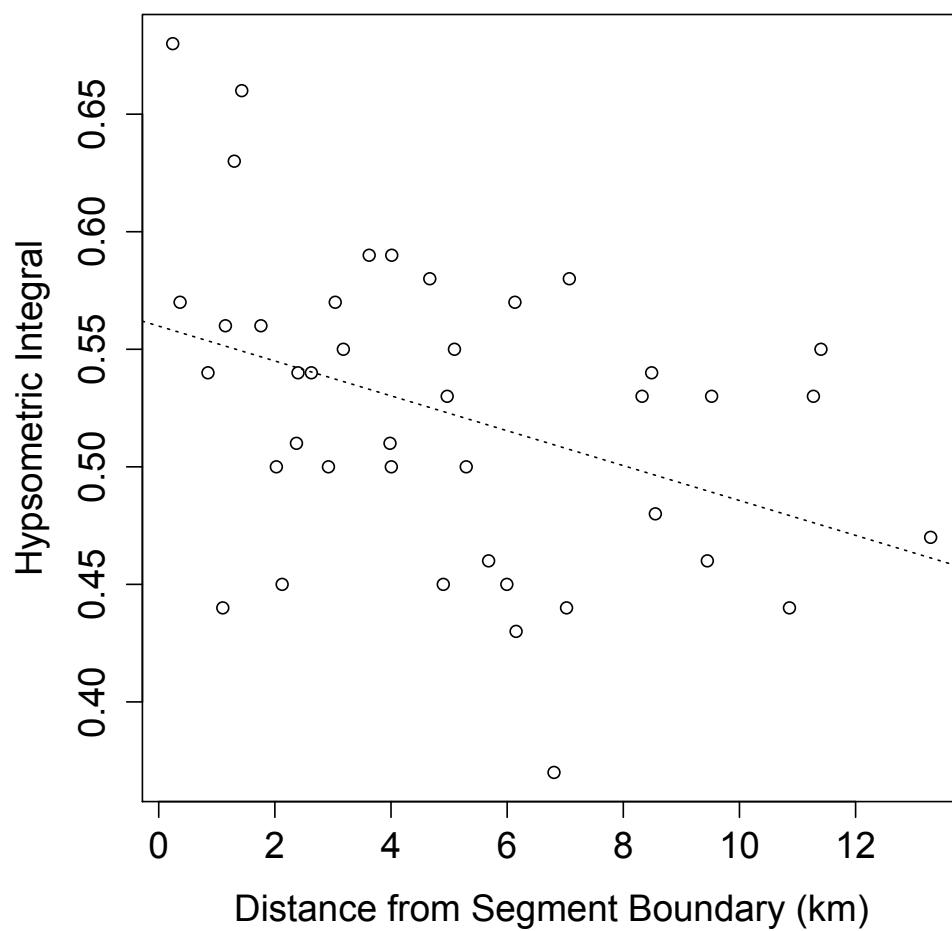


Figure 22. Hypsometric integral (HI) plotted against distance from segment boundary (D). Dotted line is regression line for standard linear regression: $HI = 0.560 - 0.00742D$, $R\text{-squared} = 0.150$.

Hypsometric Curves Vary with Distance from Segment Boundary

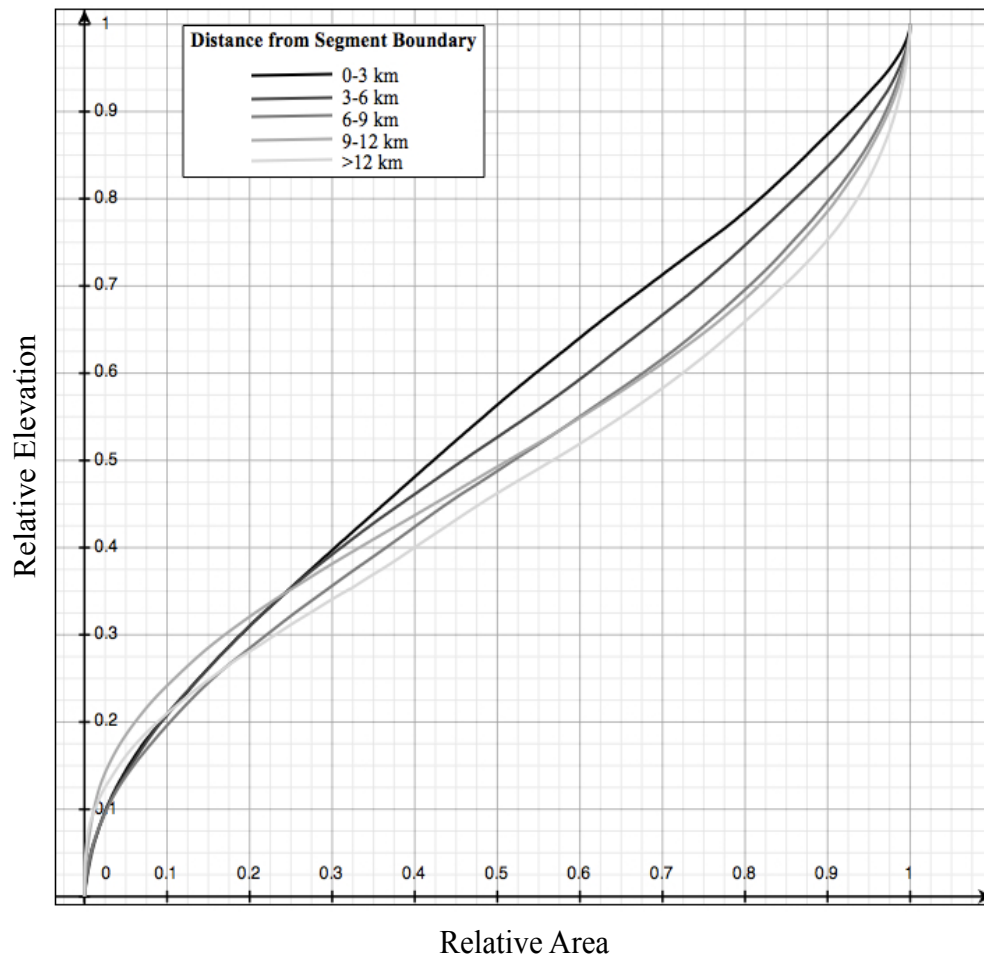


Figure 23. Hypsometric curve variations with drainage basin distance to segment boundary. Each curve was created by averaging the hypsometric curves of drainage basins within a 3-km distance to segment boundary class (0-3 km: 14 basins; 3-6 km: 13 basins; 6-9 km: 8 basins, 9-12 km: 5 basins, >12 km: 1 basin). Curves of basins closer to segment boundaries are straighter; while curves for basins further from segment boundaries are more sigmoidal. Basins further from segment boundaries have more area at relatively lower elevations, and basins closer to segment boundaries have more area at relatively high elevations.

Drainage Basin and Fan Area and Slope Relationships

Visual inspection of fan area vs. drainage basin area, fan slope vs. drainage basin area, and drainage basin slope vs. drainage basin area (Figures 24-27) prompted the decision to plot relationships between all four variables (fan slope, fan area, drainage basin slope, and drainage basin area) on log-log plots. Linear regression was performed on log-transformed data. Log-log plots are standard constructs used in previous studies (Hooke, 1968; Funk, 1977; Rockwell et al., 1985; Cooke et al., 1993; Allen and Hovius, 1998; Mack and Leeder, 1999; Milana and Ruzyski, 1999; Al-Farraj and Harvey, 2005; Harvey, 2005; Leleu et al., 2005; Lin et al., 2009). Since these parameters were not considered as a function of segment distance, the two basins excluded in the analysis described in the above section (“Morphometric Parameters and Distance from Fault Segment Boundaries”) are included in these plots. Thus, relationships involving only drainage basin parameters included all 43 basins; relationships involving drainage basin and fan parameters included 40 basins (basins 1-40).

Fan Area and Drainage Basin Area

Fan area as a function of drainage basin area was considered for 40 basins. Fan area is directly related to drainage basin area (R-squared = 0.827, residual standard error = 0.299 log units, p-value=4.74x10⁻¹⁶). The regression equation describing the relationship is:

$$A_f = 0.830A_b^{0.848}$$

where A_f = fan area and A_b = drainage basin area (Figure 28).

Fan Area vs. Drainage Area

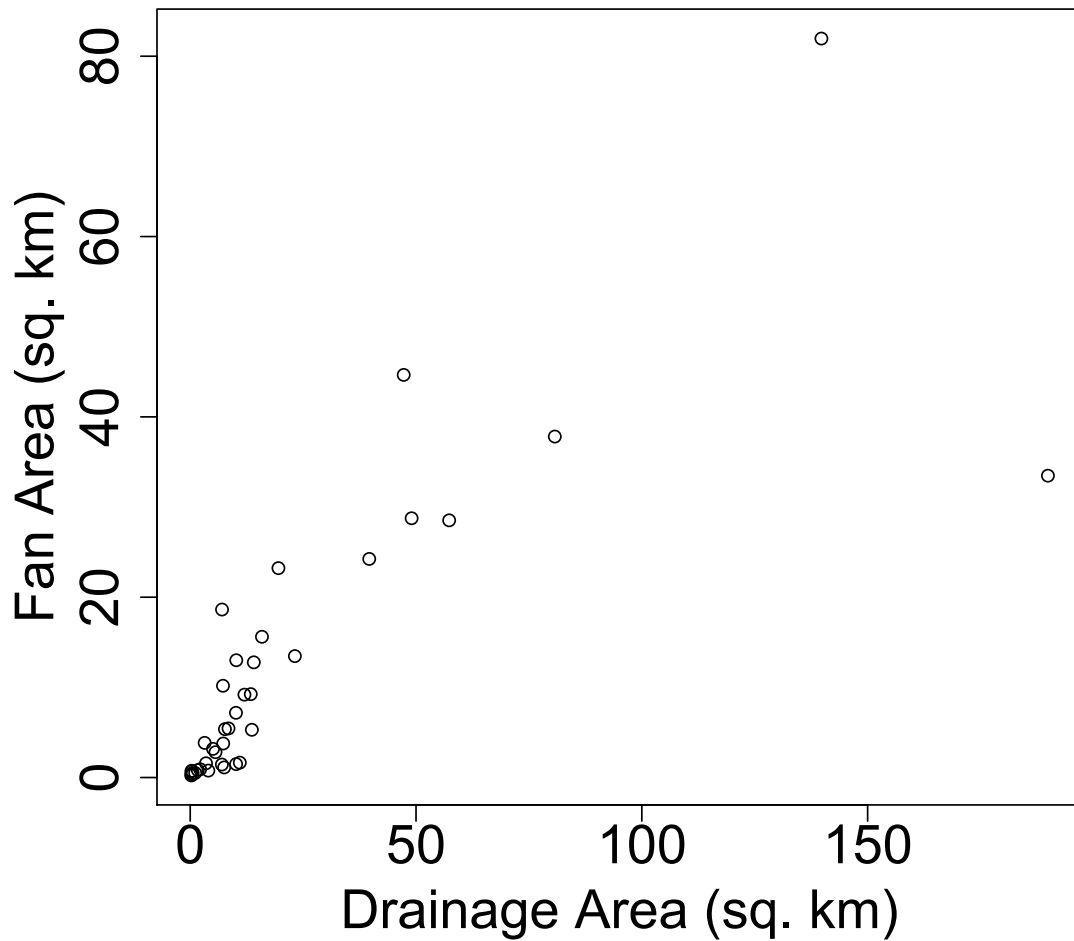


Figure 24. Linear plot of fan area against drainage basin area for 40 basins (basins 1-40). Note that fan area tends to increase as drainage area increases. Compare with the log-transformed graph of these data (Figure 28) and note the need to log-transform the data.

Fan Slope vs. Drainage Area

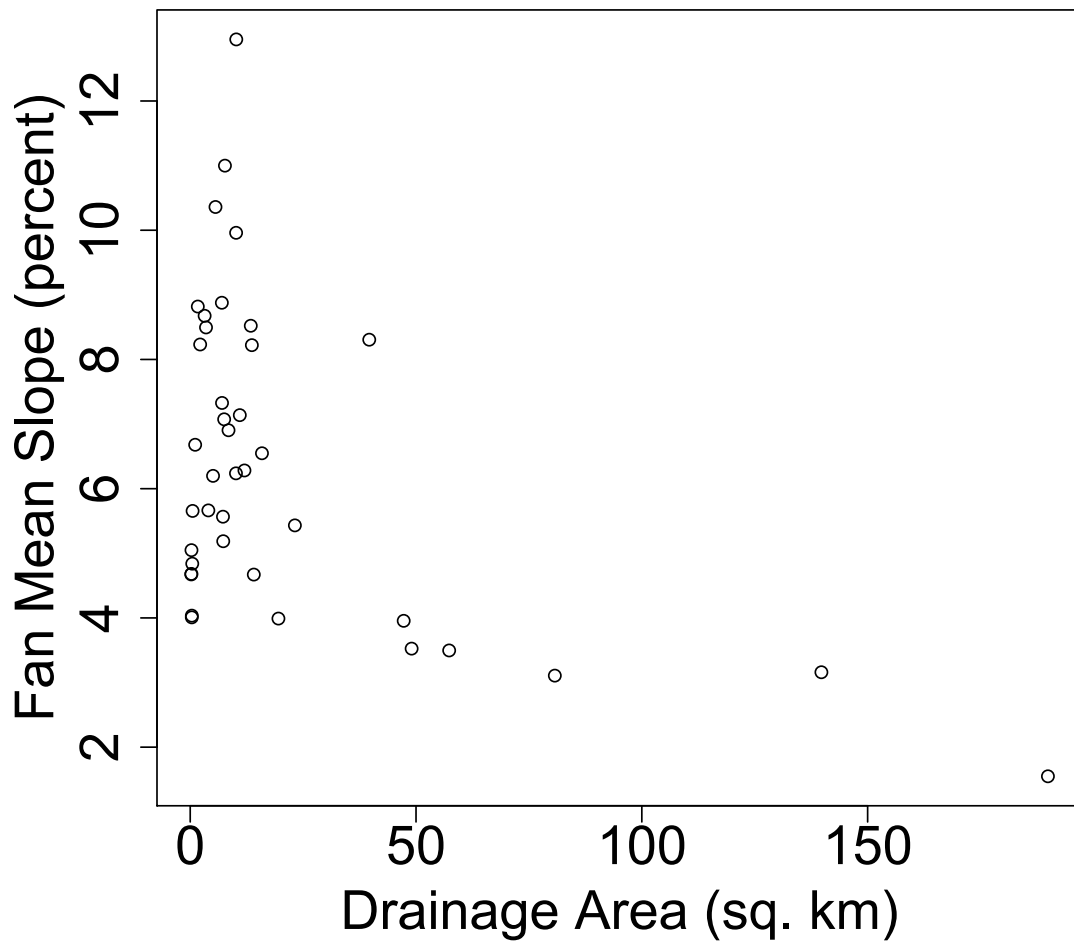
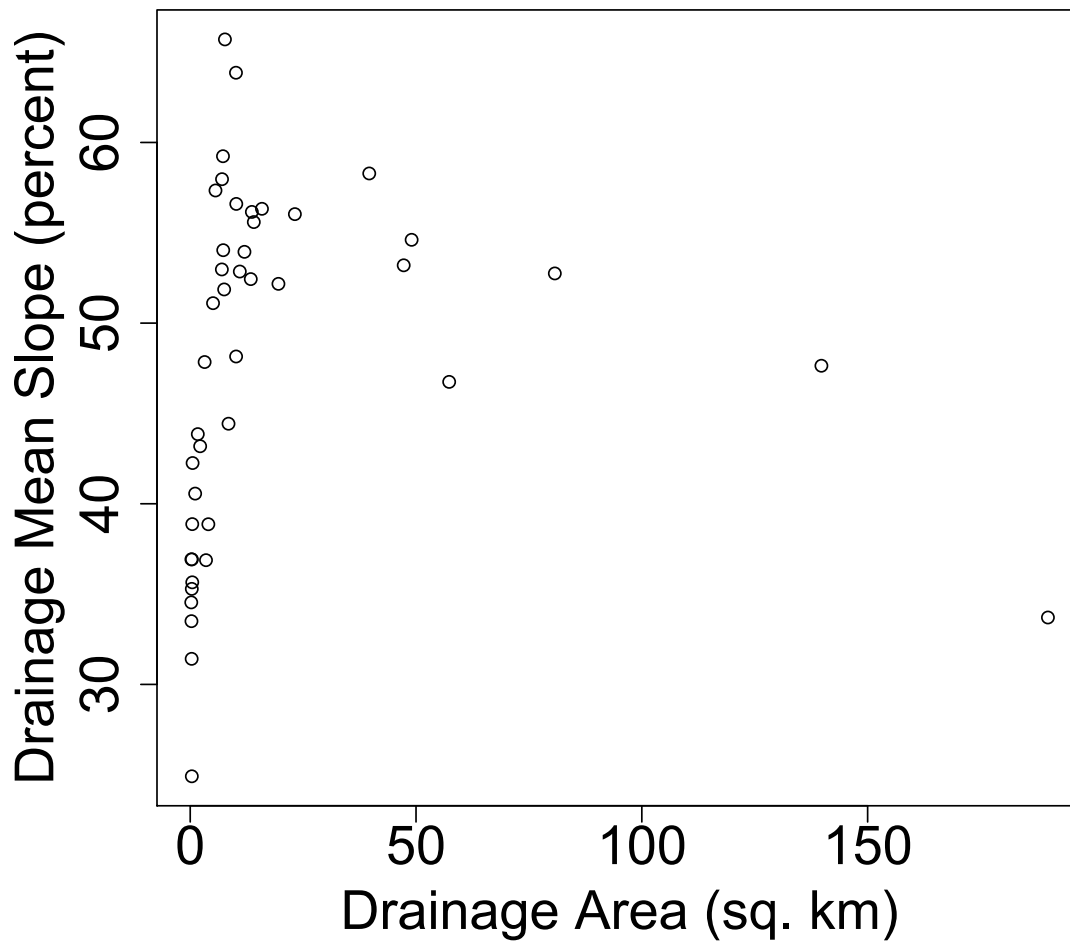


Figure 25. Linear plot of fan mean slope (in percent slope) against drainage basin area for 40 basins (basins 1-40). Slope was calculated using the “pixel” method – fan mean slope is the average slope of all pixels within the fan as obtained using zonal statistics.

Drainage Slope vs. Drainage Area



Drainage Slope vs. Drainage Area

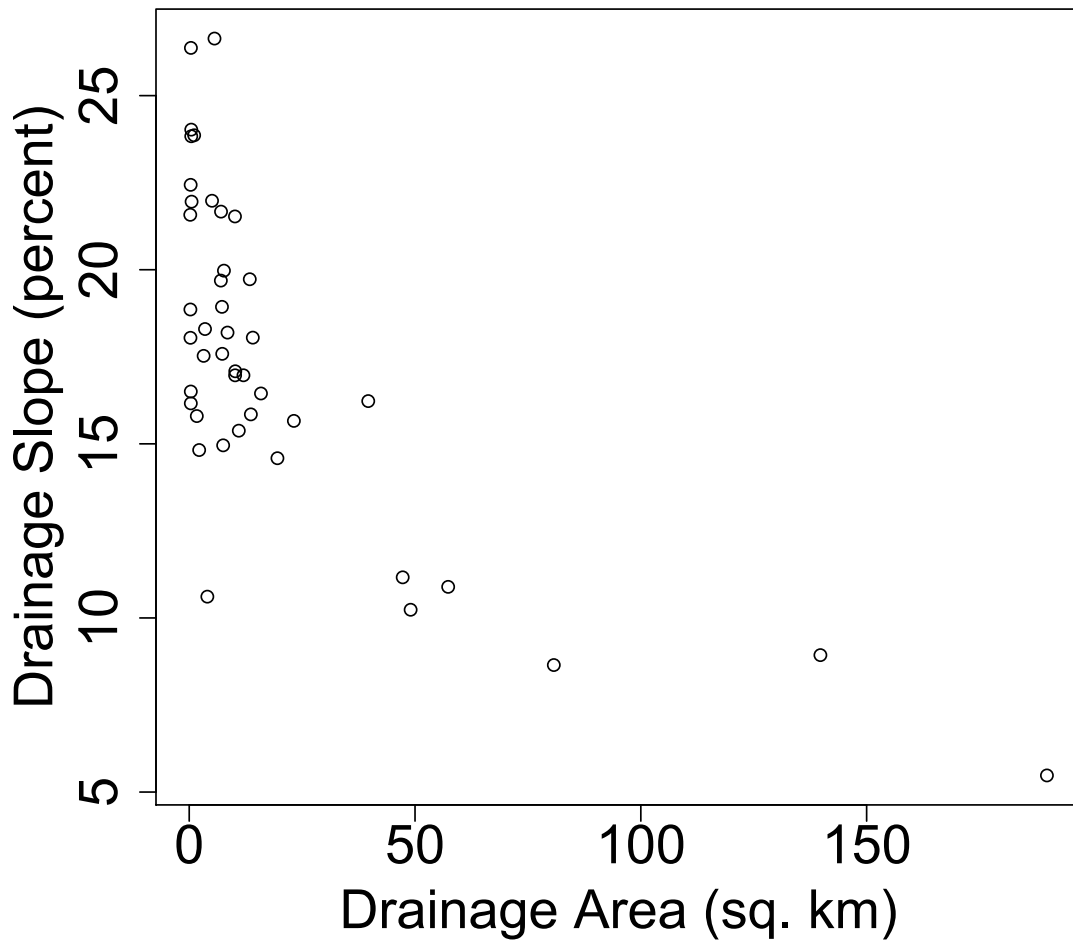


Figure 27. Linear plot of drainage basin slope against drainage basin area for 43 basins. Note the need to log-transform the data and tendency for smaller drainage basins to have greater slopes. Slope was calculated using the “conventional” method – dividing basin relief by basin length. Note wide differences between slopes calculated using this method and slopes calculated using zonal statistics (previous graph).

Fan Area vs. Drainage Area

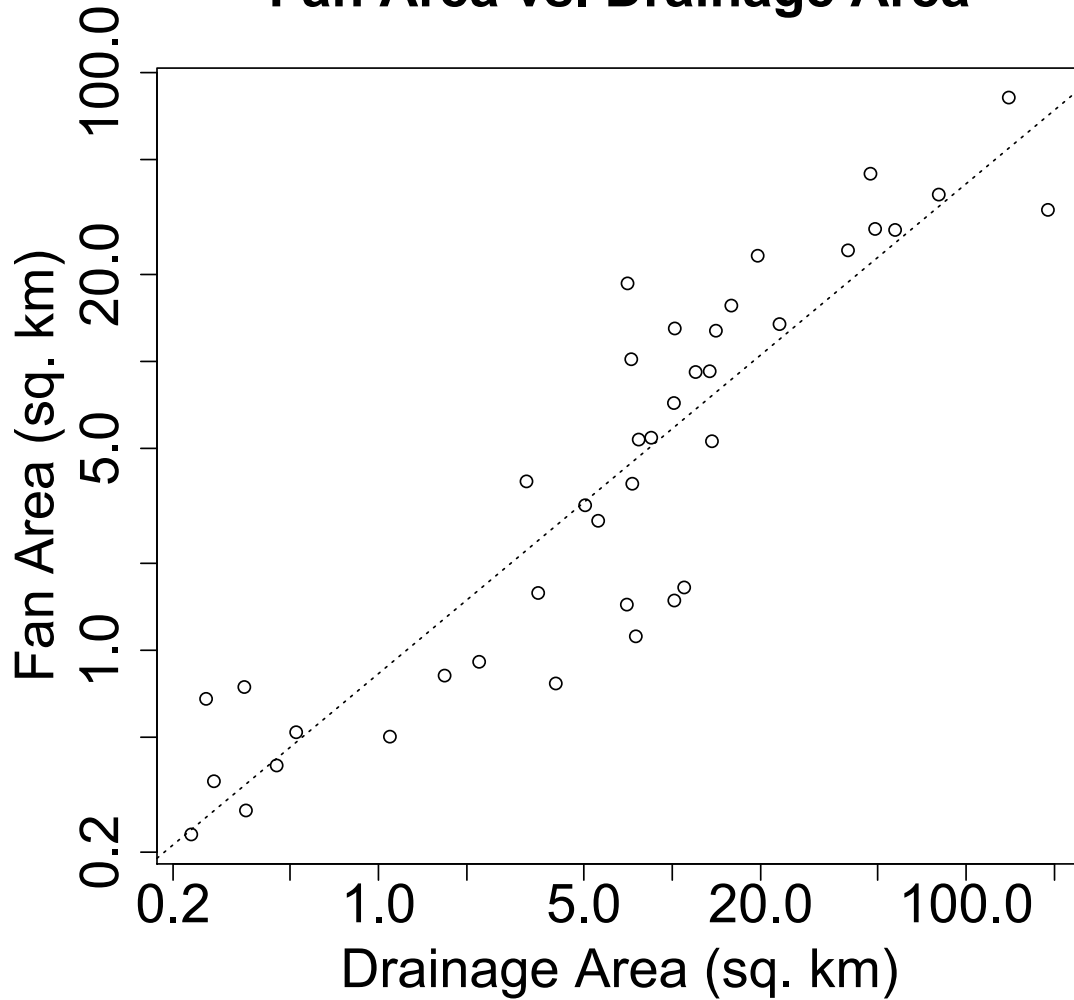


Figure 28. Log-log plot of fan area against drainage basin area. Dotted line is regression line for standard linear regression on log-transformed data: $A_f = 0.830A_b^{0.848}$, R-squared = 0.827.

Fan Slope and Drainage Basin Area

Fan mean slope (in percent slope) as a function of drainage basin area was considered for 40 basins. Fan mean slope is inversely related to drainage basin area (R-squared = 0.0498, residual standard error = 0.180 log units, p-value=0.166). The regression equation describing the relationship is:

$$S_f = 6.51A_b^{-0.0535}$$

where S_f = fan mean slope and A_b = drainage basin area (Figure 29).

Fan Slope vs. Drainage Area

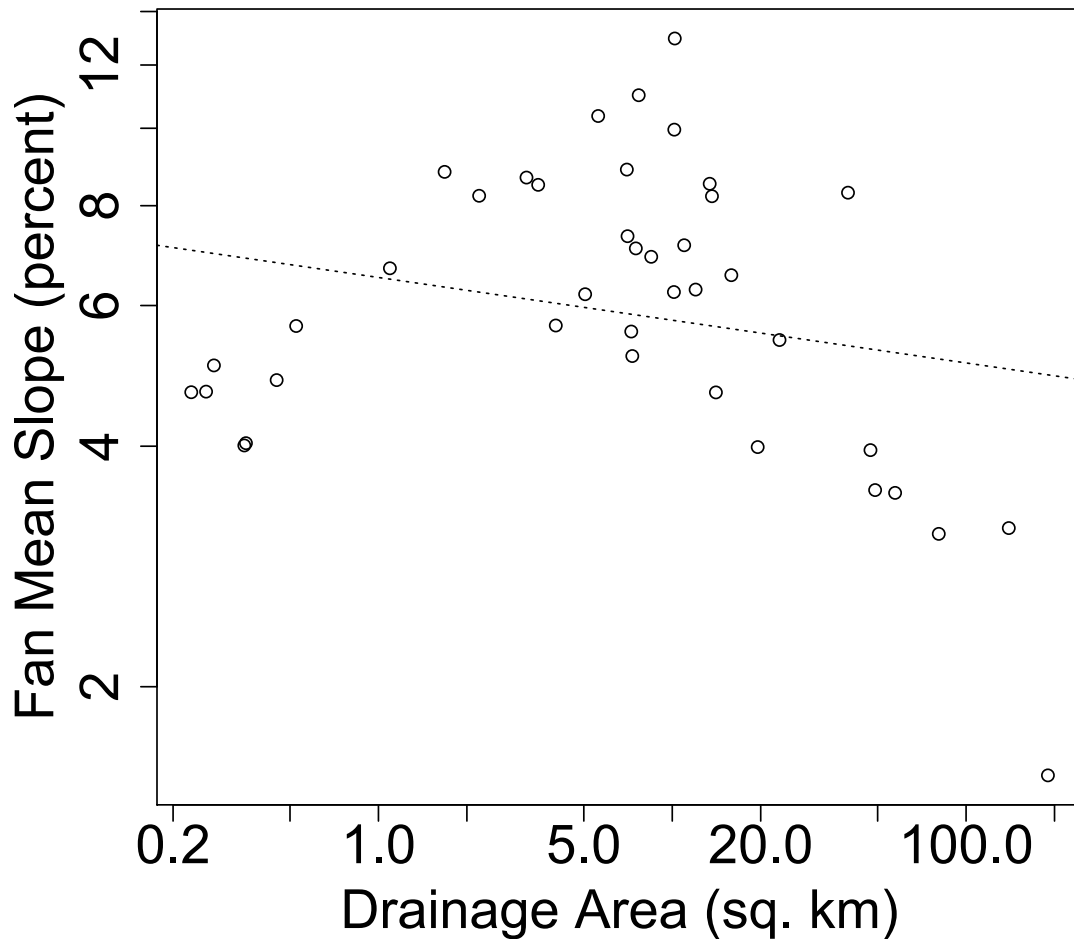


Figure 29. Log-log plot of fan mean slope (in percent slope) against drainage basin area. Fan mean slope was calculated as the average slope of all pixels within the fan using zonal statistics. Dotted line is regression line for standard linear regression on log-transformed data: $S_f = 6.51A_b^{-0.0535}$, R-squared = 0.0498. Note that a linear regression of log-transformed data is probably not the most appropriate regression for this data distribution, but is the standard regression used to compare variations in fan slope with variations in drainage basin area. Since one question being tested if the data distribution of the current study fits well with previous studies, the standard regression relationship was preformed.

Drainage Basin Slope and Drainage Basin Area

Drainage basin mean slope (in percent slope) as a function of drainage basin area was considered for 43 basins. Drainage basin mean slope is directly related to drainage basin area (R-squared = 0.418, residual standard error = 0.0739 log units, p-value = 2.81×10^{-6}). The regression equation describing the relationship is:

$$S_b = 41.0A_b^{0.0777}$$

where S_b = drainage basin mean slope and A_b = drainage basin area (Figure 30).

Drainage basin slope was also calculated conventionally, by dividing basin relief by basin length. Calculated conventionally, drainage basin slope (in percent slope) as a function of drainage basin area was considered for 43 basins. Drainage basin slope is inversely related to drainage basin area (R-squared = 0.453, residual standard error = 0.104 log units, p-value = 7.58×10^{-7}). The regression equation describing the relationship is:

$$S_b = 20.2A_b^{-0.118}$$

where S_b = drainage basin slope and A_b = drainage basin area (Figure 31).

Drainage Slope vs. Drainage Area

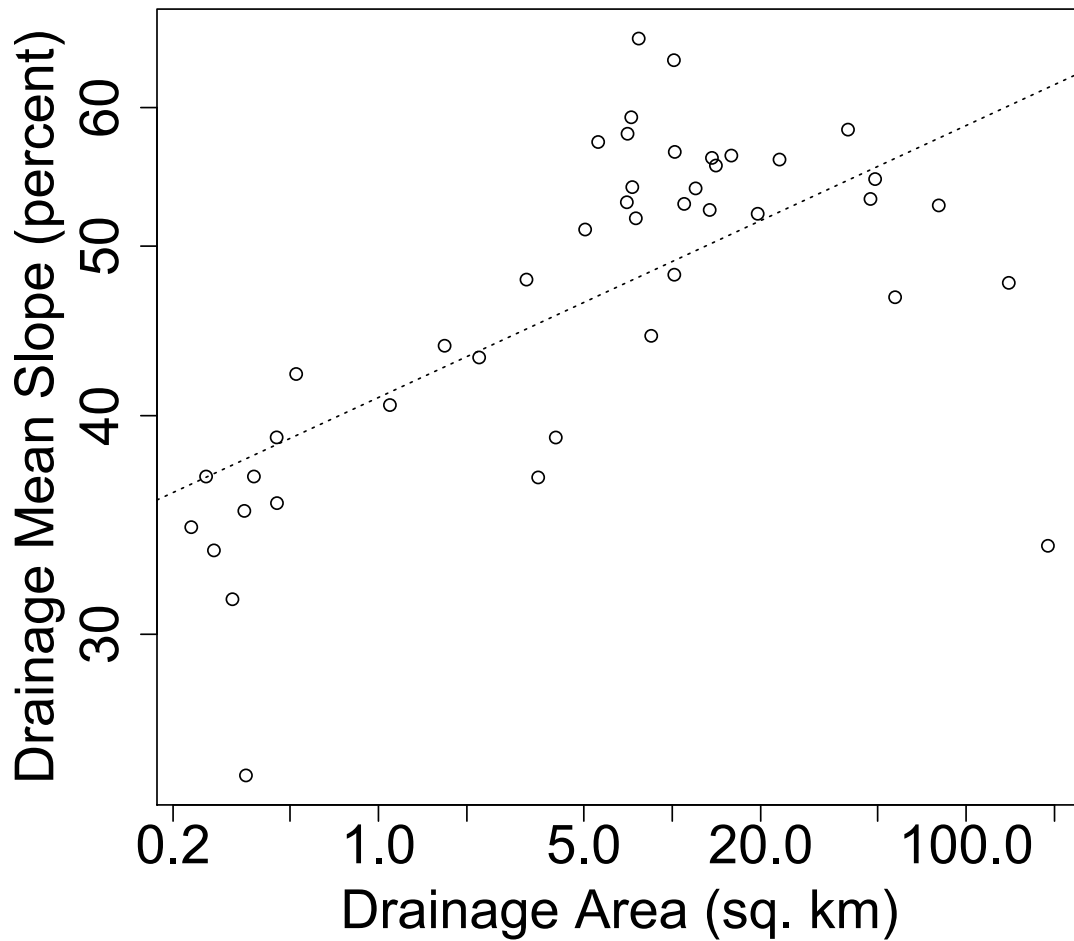


Figure 30. Log-log plot of drainage basin mean slope (in percent slope) against drainage basin area. Drainage basin mean slope was calculated as the average slope of all pixels within the drainage basin using zonal statistics. Dotted line is regression line for standard linear regression on log-transformed data: $S_b = 41.0A_b^{0.0777}$, R-squared = 0.418.

Drainage Slope vs. Drainage Area

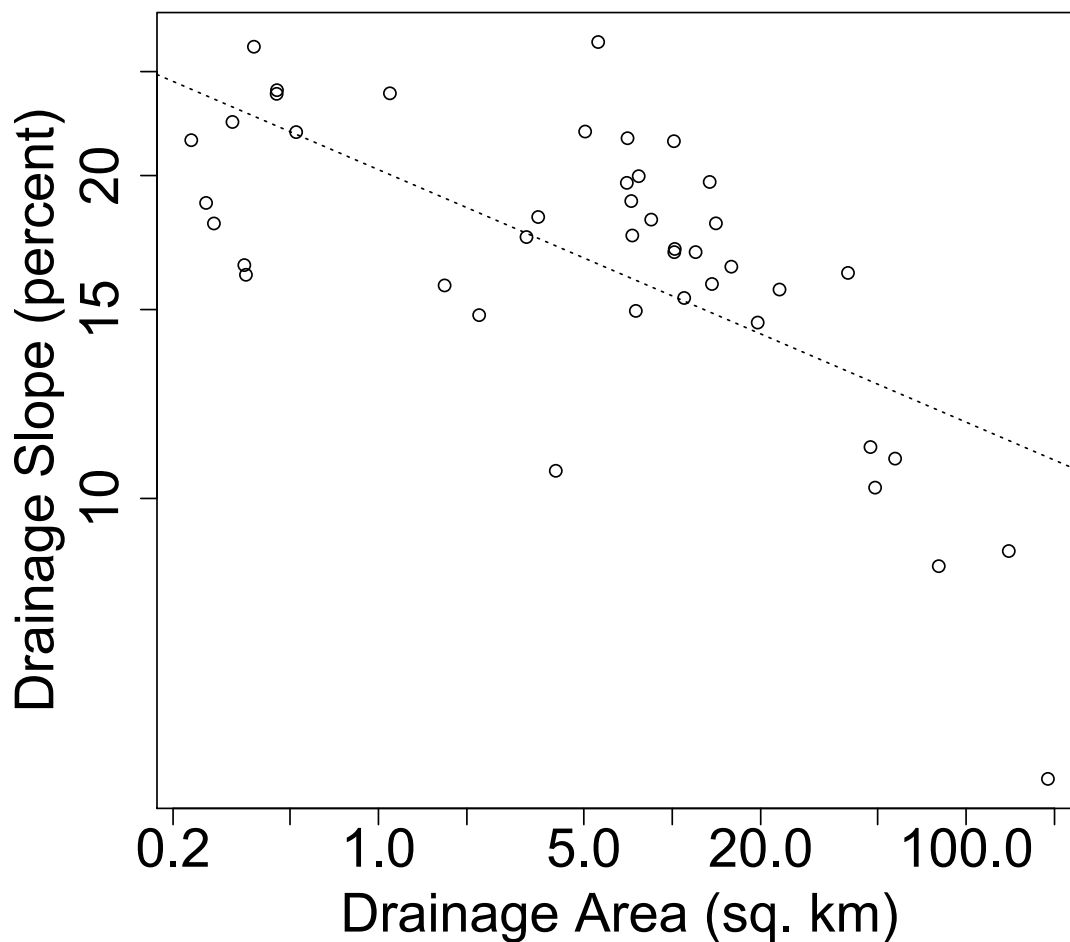


Figure 31. Log-log plot of drainage basin slope against drainage basin area. In contrast to the previous figure, drainage basin slope was calculated conventionally – by dividing basin relief by basin length. Dotted line is regression line for standard linear regression on log-transformed data: $S_b = 20.2A_b^{-0.118}$, R-squared = 0.453.

DISCUSSION

Drainage Basin Area

Only three of the 43 drainage basins that reach the main drainage divide have areas larger than 50 km² (Table 5). The largest drainage basin (basin 31, Sawmill Canyon drainage) has an area of 190 km². The other two drainage basins larger than 50 km² are basin 32 and basin 33 (see Figure 12 and Table 4 for basin location and names). One potential explanation for the large area of at least the Sawmill Canyon drainage, as well as possibly the Big Creek drainage (basin 32) is the spatial coincidence of Eocene-Oligocene basin fill and faults at high angle to the modern range-bounding fault (Janecke, 1994).

Eocene-Oligocene basins in the Lemhi vicinity included the Donkey basin in what is now the area of Donkey Hills, Goldburg Hills, and Stinking Creek near Big Creek, and the Sawmill Canyon basin, which was part of a single N-trending basin that included the Pass Creek/Wet Creek basin in the central Lost River Range (Janecke, 1994). These basins are interpreted to be part of a narrow N-S trending Neogene rift (see “Paleogene Supradetachment Basins” in Regional Setting chapter). Particularly in Sawmill Canyon (basin 31), Paleogene basin fill is exposed in the footwall of the Lemhi fault (Janecke, 1992a). This contrast in lithology, as well as the presence of Paleogene normal faults at high angle to the Miocene-Recent Lemhi fault, may have increased the relative erodability of these footwall drainages.

The two largest drainage basins (basins 31 and 32) are located within segment boundary zones (Figure 18). Thus, they were excluded from regressions relating morphometric parameters to drainage basin distance from nearest fault segment boundary. Previous studies have also noted that anomalously large drainage basins are often located in segment boundary zones (Gawthorpe and Leeder, 1987; Leeder and Jackson, 1993; Horton and Schmitt, 1998; Allen and Allen, 2005).

Morphometric Parameters and Distance from Fault Segment Boundaries

Three hypothesis questions were raised regarding variation in drainage basin morphometric parameters as a function of distance from fault segment boundary. These three questions are answered briefly below and then addressed individually in greater detail in the following sections.

1. Does the ratio of fan area to drainage basin area (ϕ) decrease toward the interior of fault segments? Yes, ϕ decreases towards segment interiors and increases towards segment boundaries (Figure 19). However, the relationship is very weak.

2. Are drainage basins more elongate (lower elongation ratio) closer to fault segment boundaries and more round (higher elongation ratio) in the interior of fault segments? Yes, basins are more round in fault segment interiors and more elongate closer to fault segment boundaries (Figure 20). However, the relationship is very weak.

3. Does hypsometry change along strike with contrasting hypsometric curve shapes in the interiors of segments compared to near segment boundaries? Specifically, are drainage basins in segment interiors characterized by higher hypsometric integrals

and are drainage basins closer to segment boundaries characterized by lower hypsometric integrals and a more sigmoidal hypsometric curve shape? Hypsometry does change along strike, but in the opposite direction from this hypothesis, and the relationship is very weak. Drainage basins closer to segment boundaries have higher hypsometric integrals and straighter hypsometric curve shapes, while drainage basins located in segment interiors have lower hypsometric integrals and a more sigmoidal hypsometric curve shape (Figures 22-23).

The Melton ratio was also used to compare drainage basins. No hypothesis was proposed for how basin location along a segment boundary would affect Melton ratio. The Melton ratio of drainage basins decreases towards the interior of segments, but the relationship is very weak (Figure 21).

Fan Area to Drainage Basin Area Ratio

The ratio of fan area to drainage basin area (ϕ) decreases with increasing drainage basin distance from the closest fault segment boundary (D). The equation describing this relationship is: $\phi = 0.988 - 0.312D$. However, R-squared is very low for the linear regression (R-squared = 0.0294), indicating that very little of the variance in ϕ can be explained by drainage basin position along a fault segment. While the correlation is weak, the scatter plot (Figure 19) does not clearly indicate that a non-linear correlation would be preferred. Plotting ϕ against drainage basin outlet distance from the southern range tip (Figure 32) reveals that for any given segment, ϕ varies widely and with no strong pattern for individual segments.

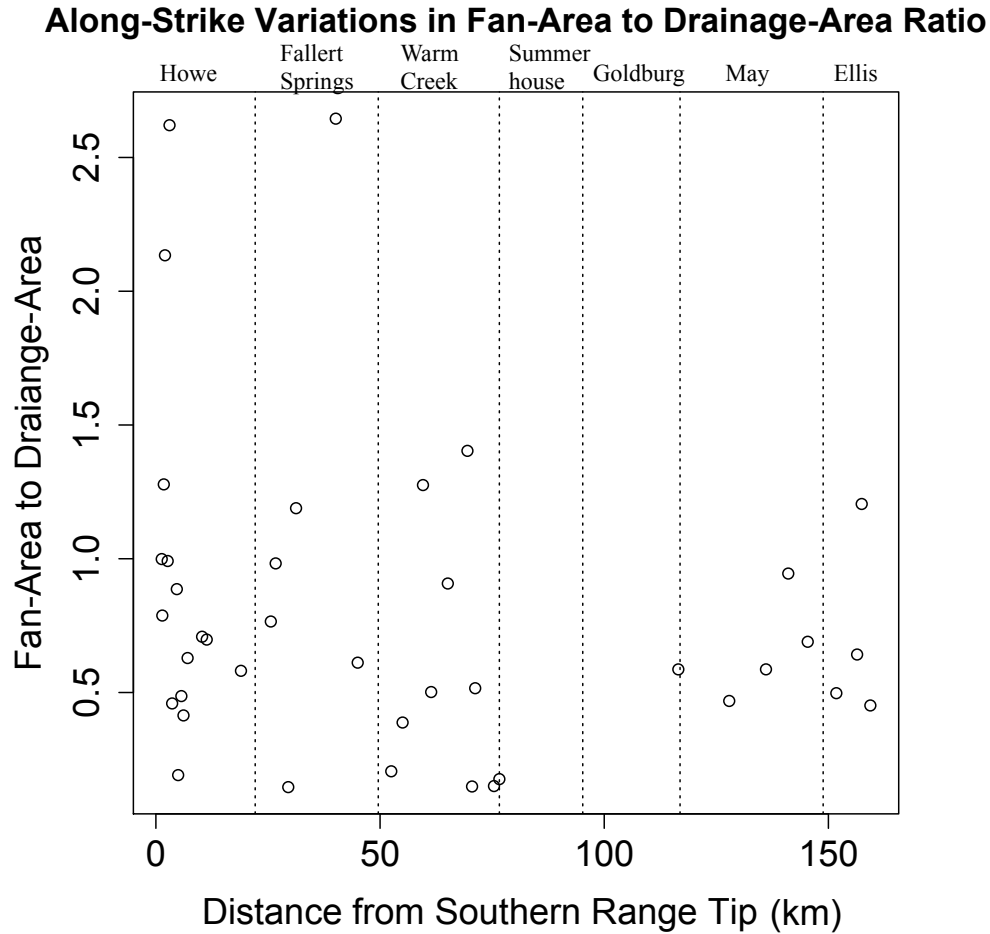


Figure 32. Fan area to drainage basin area ratio (ϕ) plotted against drainage basin outlet distance along strike from the southern range tip. For any given segment, ϕ varies widely – there is no strong pattern for individual segments. Dashed lines denote segment boundary locations used in this study. From south to north, segments are: Howe, Fallert Springs, Warm Creek, Summerhouse, Goldburg, May, and Ellis. For map of segments, see Figure 11.

Despite the weakness of the relationship, the direction of the observed relationship (inverse correlation between ϕ and distance to segment boundary) supports the conceptual model that increased fault slip rate and thus increased accommodation space towards the interior of fault segments encourages alluvial fan packages to stack more vertically. In contrast, fans located closer to segment boundaries are closer to

hangingwall highs and experience lower slip rates. Thus, fan packages near segment boundaries will tend to stack laterally, leading to higher plan-view fan area for a basin of a given size if the basin outlet is located closer to a fault segment boundary than if it is located in the interior of a segment.

Drainage Basin Elongation Ratio and Melton Ratio

Drainage basin elongation ratio (ER) increases with increasing drainage basin distance from the closest fault segment boundary (D). The equation describing this relationship is: $ER = 0.489 + 0.00615D$. However, R-squared is very low for the linear regression (R-squared = 0.0604), indicating that very little of the variance in ER can be explained by drainage basin position along a fault segment. While the correlation is very weak, the scatter plot (Figure 20) does not clearly indicate that a non-linear correlation would be preferred. Plotting elongation ratio against drainage basin outlet distance from the southern range tip (Figure 33) reveals that for any given segment, elongation ratio varies widely and with no strong pattern for individual segments. The drainage basins with the lowest elongation ratios (most elongate basins) are located along either the northernmost or southernmost fault segment.

Elongation ratio is related to the efficacy of sediment removal from the drainage basin (Densmore et al., 2005). Footwall drainage patterns evolve via weathering, erosion (including mass wasting) and knickpoint incision (Friedmann and Burbank, 1995), as well as piracy or capture of neighboring drainages (Densmore et al., 2005). However, once range half-width is established (as is the case with the Lemhi Range), drainage basins on extensional footwalls become more round (lower elongation ratio) by widening

along strike via piracy or drainage capture of neighboring catchments rather than expanding towards the main range drainage divide (Densmore et al., 2005). Increasing elongation ratio (increasing basin roundness) toward segment interiors implies that more effective drainage competition is occurring in segment interiors. This allows the drainage divide to be drained by fewer, but larger, rounder basins in segment interiors.

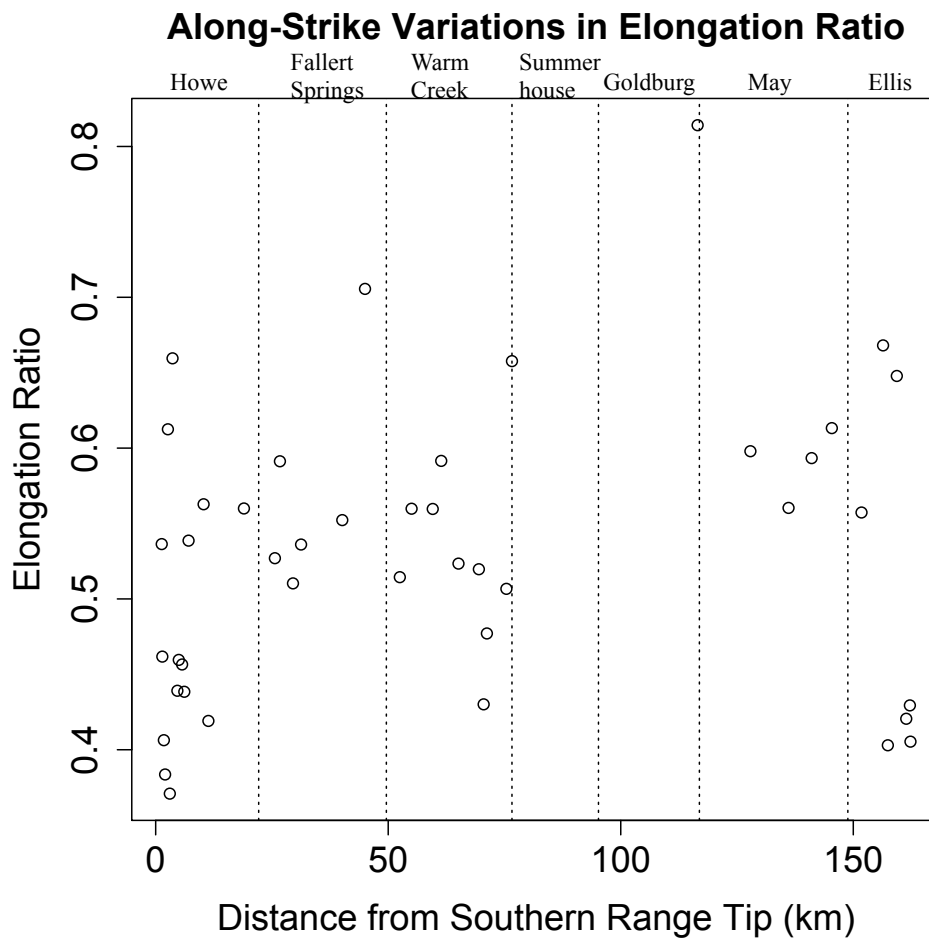


Figure 33. Elongation ratio plotted against drainage basin outlet distance along strike from the southern range tip. For any given segment, elongation ratio can vary widely. The lowest elongation ratios (most elongate basins) are located primarily along the segments on either end of the range. Dashed lines denote segment boundary locations used in this study. From south to north, segments are: Howe, Fallert Springs, Warm Creek, Summerhouse, Goldburg, May, and Ellis. For map of segments, see Figure 11.

Drainage basin Melton ratio (MR) decreases with increasing drainage basin distance from the closest fault segment boundary (D). The equation describing the relationship is: $MR = 0.470 - 0.0136D$. However, R-squared is low for the linear regression (R-squared = 0.143), indicating that little of the variance in MR can be explained by drainage basin position along a fault segment. While the correlation is not strong, the scatter plot (Figure 21) does not clearly indicate that a non-linear correlation would be preferred. This was the strongest (highest R-squared and lowest p-value) linear correlation of any morphometric parameter with distance to segment boundary. Plotting Melton ratio against drainage basin outlet distance from the southern range tip (Figure 34) reveals a relatively strong pattern at the range scale: Melton ratio trends downward along strike toward the range interior, with the highest Melton ratios mostly towards either edge of the range. This pattern is still weak, but is strong relative to the lack of range-scale patterns displayed by other parameters. Segment-scale patterns in Melton ratio variation are not as apparent as the range-scale pattern.

Variations in drainage basin Melton ratio have been associated with variations in dominant drainage basin hydrogeomorphic process (Wilford et al., 2005). Fan catchments were differentiated based on variations in Melton ratio and length for a study area in west central British Columbia (Wilford et al., 2005). For their specific study area, drainage basins of any length with Melton ratios lower than 0.3 were associated with floods as the dominant hydrogeomorphic process. Hyperconcentrated flows dominated in basins with any basin length and Melton ratios of 0.3-0.6 or in basins with length ≥ 2.7 km and Melton ratios > 0.6 . Debris flows dominated in basins with lengths < 2.7 km and Melton ratios > 0.6 (Wilford et al., 2005). While these numbers are likely not directly applicable

to the Lemhi Range, the important relationship between Melton ratio and dominant hydrogeomorphic process is that basins with higher Melton ratios are dominated by flows with a higher sediment-to-water concentration (Wilford et al., 2005).

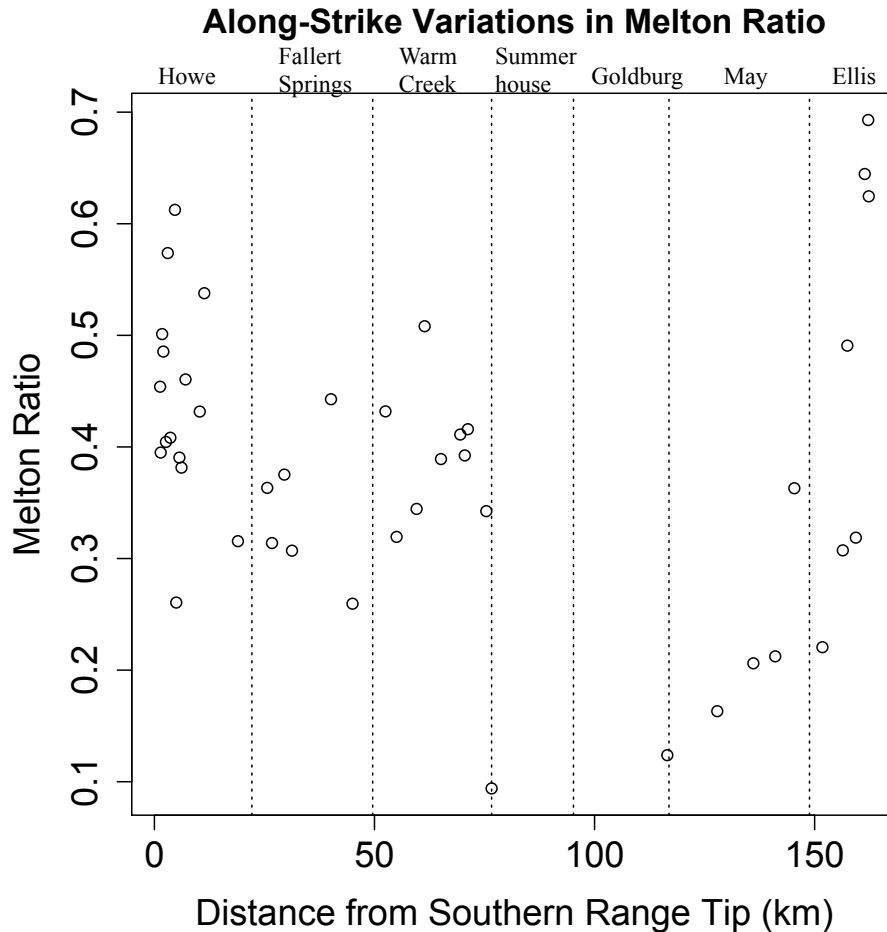


Figure 34. Melton ratio plotted against drainage basin distance along strike from the southern range tip. Drainage basin Melton ratio trends downward toward the range interior, with the highest Melton ratios mostly toward either end of the range. Segment-scale patterns are not as apparent. Dashed lines denote segment boundary locations used in this study. From south to north, segments are: Howe, Fallert Springs, Warm Creek, Summerhouse, Goldberg, May, and Ellis. For map of segments, see Figure 11.

While no flow process data were collected in this study for the drainage basins of the Lemhi Range, the Melton ratio was calculated for the drainage basins reaching the main drainage divide. If the relationship observed in the Lemhi Range of higher drainage

basin Melton ratios closer to segment boundaries holds for other normal-fault bounded ranges, conceptual models of extensional basin stratigraphic evolution could be refined. Higher Melton ratio values indicate dominant flows with higher sediment-to-water concentration. This may mean that drainage basins (particularly short drainage basins; Wilford et al., 2005) close to segment boundaries will have more flows with higher sediment-to-water concentration than basins located in fault segment interiors. Since flow character strongly influences primary porosity and permeability of the resulting deposits, there should be along-strike facies changes due to higher sediment-to-water concentrations closer to fault segment boundaries. The present study does not supply enough evidence to strongly support these stratigraphic implications; rather, it points to drainage basin Melton ratio as the morphometric parameter (of those investigated in this study) of most interest for future studies of along-strike morphometric variation.

Drainage Basin Hypsometric Integral and Hypsometric Curves

Drainage basin hypsometric integral (HI) decreases with increasing drainage basin distance from the closest fault segment boundary (D). The equation describing the relationship is $HI = 0.560 - 0.00742D$. However, R-squared is very low for the linear regression ($R\text{-squared} = 0.0603$), indicating that very little of the variance in HI can be explained by drainage basin position along a fault segment. While the correlation is weak, the scatter plot (Figure 22) does not clearly indicate that a non-linear correlation would be preferred. Plotting hypsometric integral against drainage basin outlet distance from the southern range tip (Figure 35) reveals that for any given segment, the hypsometric integral varies widely and with no strong pattern for individual segments.

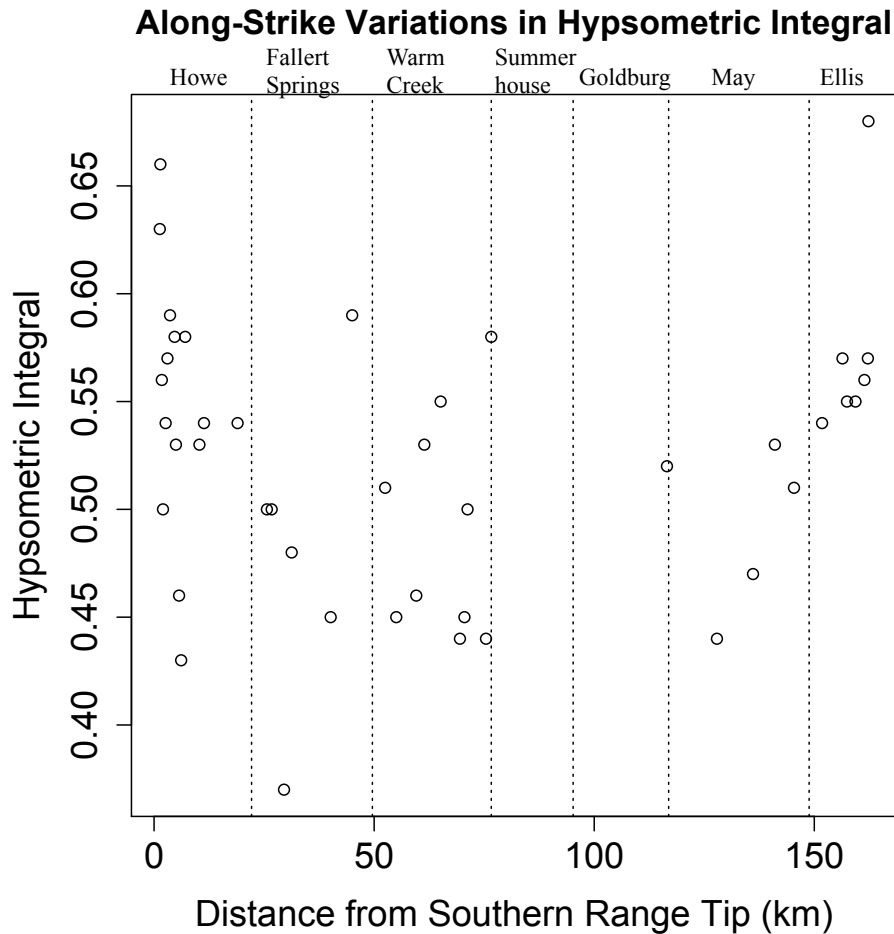


Figure 35. Hypsometric integral plotted against drainage basin distance along strike from the southern range tip. For any given segment, drainage basin hypsometric integral values vary widely with no strong pattern for individual segments. Dashed lines denote segment boundary locations used in this study. From south to north, segments are: Howe, Fallert Springs, Warm Creek, Summerhouse, Goldberg, May, and Ellis. For map of segments, see Figure 11.

Lower drainage basin hypsometric integrals towards segment interiors indicates these basins have less area at relatively high elevation compared to basins closer to segment boundaries. Drainage basins closer to segment boundaries have straighter, less sigmoidal hypsometric curves (Figure 23). Straighter (less sigmoidal) hypsometric curves indicate increased tectonic activity (Rasemann et al., 2004). Relative values of the other morphometric indices (elongation ratio and fan area to drainage basin area ratio) support

the idea that drainage basins closer to segment boundaries experience a smaller fault slip rate than drainages towards segment interiors. The apparent contradiction of greater tectonic activity close to segment boundaries indicated by higher values of drainage basin hypsometric integrals and straighter hypsometric curves is unresolved. However, due to the lack of a strong relationship between hypsometric integral and distance from nearest fault segment boundary, the apparent contradiction likely does not warrant resolution or further study in the Lemhi Range.

Drainage Basin and Fan Area and Slope Relationships

A fourth question placed the drainage basins and alluvial fans of the Lemhi Range in context with other studies of Quaternary alluvial fans. Do the fans along the western Lemhi Range follow the traditionally accepted relationships described in equations 1-3 (see “Alluvial Fan and Drainage Basin Area and Slope Relationships” in Introduction chapter)? In other words, are fan area and fan drainage basin area directly related; fan slope and drainage basin area inversely related; and drainage basin slope and drainage basin area inversely related?

The Lemhi fans and drainage basins do follow the traditionally accepted relationships. Fan area and fan drainage basin area are directly related, while fan slope is inversely related to drainage basin area, and drainage basin slope (when calculated conventionally) is inversely related to drainage basin area.

Fan Area and Drainage Basin Area

Fan area increases with increasing drainage basin area. The equation for the linear regression of the log-transformed data is $A_f = 0.830A_b^{0.848}$. The R-squared for the relationship is 0.827, indicating that a large amount of variation in fan area can be explained by variation in drainage basin area. The exponent (0.848) of the regression fits well with previous studies, which found the exponent to usually range between about 0.7 and 1.1 (Al-Farraj and Harvey, 2005). The value of the constant (0.830) also fits within the previously reported range of about 0.1 to 2.1 (Al-Farraj and Harvey, 2005).

Fan Slope and Drainage Basin Area

Fan slope decreases with increasing drainage basin area. The equation for the linear regression of the log-transformed data is $S_f = 6.51A_b^{-0.0535}$. The R-squared value is very low (0.0498) for this correlation, indicating that very little of the variance in fan slope can be explained by drainage basin area. The log-log scatter plot of the data indicates that a non-linear regression might be more appropriate to describe the relationship between fan area and drainage basin area (Figure 29). However, since linear regression is used in other studies to describe the fan slope to drainage basin area relationship, it was used in this study as well to place the results in context with other studies. The direction of the relationship is the same as in previous studies (inverse correlation). The poor fit of the regression in this study may be due to different methodology in calculating fan slope (see “Drainage Basin Slope and Fan Slope” in Methodology chapter).

Drainage Basin Slope and Drainage Basin Area

Drainage basin slope was calculated using two different methods. Slope was calculated conventionally by dividing basin relief by basin length, and slope was calculated using the “pixel” method or as the mean slope of all DEM pixels within the drainage basin. When calculated conventionally, drainage basin slope decreases with increasing drainage basin area. The equation for the regression of the log-transformed data is $S_b = 20.2A_b^{-0.118}$. The R-squared value for this correlation (0.453) indicates that almost half the variance in drainage basin slope can be explained by variance in drainage basin area. The direction of this relationship (inverse correlation) agrees with previous studies (see “Alluvial Fan and Drainage Basin Area and Slope Relationships” in Introduction chapter).

However, when the “pixel” method (see “Drainage Basin Slope and Fan Slope” in Methodology chapter and “Two Different Methods of Calculating Drainage Basin Slope” below) was used to calculate drainage basin slope, a direct correlation between drainage basin slope and drainage basin area was observed. The equation for the regression of the log-transformed data using “pixel” slope is $S_b = 41.0A_b^{0.0777}$ and the R-squared is 0.418. Log-log scatter plots of drainage basin slope (using either calculation method) against drainage basin area do not indicate that a non-linear regression would be preferred (Figures 30-31).

Two Different Methods of Calculating Drainage Basin Slope

In studies of alluvial fans and fan drainage basins, drainage basin slope is conventionally calculated by dividing the relief of the basin by the length of the basin, typically derived from topographic maps (Al-Farraj and Harvey, 2005; Harvey, 2005). When a DEM is used instead of a topographic map, an additional method is available for slope calculation. Instead of determining a single slope value based on basin relief and length, a slope value is calculated for each pixel in the drainage basin. These values can then be averaged to give what is perhaps a more representative mean slope for the basin. In practice, the DEM covering the study area is used to calculate a slope grid and zonal statistics are used to extract minimum, maximum, and mean slopes within each identified drainage basin. The method of calculating slope as the mean slope of all pixels within the drainage basin will herein be referred to as the “pixel” method of slope calculation, and the basin relief/length method as the “conventional” method.

Fan studies have often cited an inverse relationship between basin area and basin slope – smaller basins tend to be steeper than larger basins (Lin et al., 2009). In this study, basin slope was initially calculated using the pixel method. Instead of the expected inverse relationship, a direct relationship was observed (Figure 30). Smaller basins were less steep than larger basins. Slope was then calculated for each basin using the conventional method. With slope calculated by dividing basin relief by basin length, the expected inverse relationship between slope and drainage basin area was observed (Figure 31).

Using the two methods to calculate basin slope yields very different results (Figure 36). However, the two methods are assessing different processes driving sediment transport in the basin. Conventional slope is a proxy for channel slope of the main channel of the drainage basin. In contrast, pixel slope incorporates hillslope scale information in addition to channel slope; slope of the hillsides above the main channel affect mass wasting processes and sediment delivery to the channel itself.

Different Slope Methods

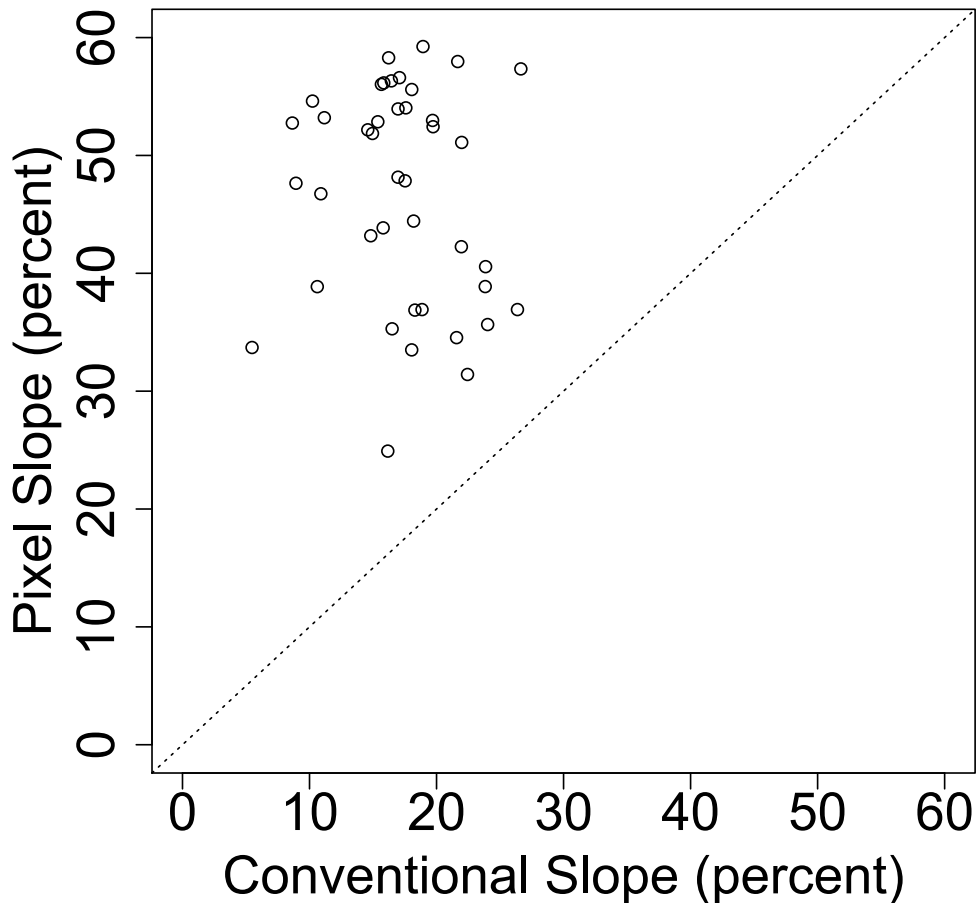


Figure 36. Two different methods of calculating drainage basin slope yield very different results. “Pixel” slope is the mean slope of all pixels in the basin, while “conventional” slope is the basin relief divided by basin length. Dotted line is $y=x$, and denotes the location of where the points would fall if the two methods yielded the same results.

In order to put the results of this study in context of previous studies, the regression using slope from the conventional method is used. However, this may not be the most representative slope for the basins and may not be the most accurate reflection of sediment production and removal, as conventional slope represents only the main channel slope. An accurate quantification of slope is important because slope affects sediment transport processes, gravitational potential energy and slope stability. An example of a case in which the two slope methods would yield very different results is a large basin compartmentalized into smaller steep subbasins that branch off the main drainage path. Slope calculated using the conventional method would simply take the longest flow path and divide it by the basin length, and not account for slope values within the various compartments. In the study area, this is exemplified by basin 33 (Figure 37). The different slope methods yield similar results for basins that are not as compartmentalized and are straighter like basin 5 (Figure 37). Future studies of drainage basins will need to consider which slope definition is more representative of the sediment production and sediment transport processes of interest. Regardless of which definition is chosen for a particular study, researchers should ensure that similar methods are used when comparing their results with previous research.

The discrepancy between results obtained using different slope methods has been noted (Wobus et al., 2006), but there does not seem to be consensus in the literature about which method should be used to calculate slope. However, most studies of alluvial fans and their drainage basins do not seem to recognize that different methods are available or the discrepancies that result from comparing results obtained using different methods. While Wobus and others (2006) describe a method to extract slope from a DEM that

approximates the conventional method, there seems to be no discussion as to whether this is the proper objective to pursue in the first place. Wobus et al. (2006) note that using values from a 3x3 moving window slope grid (equivalent to “zonal statistics” as described in methods section of this paper) will yield higher values, particularly in large drainage areas with steep bedrock canyons. Their recommended calculation method for channel slope involves resampling raw elevation data at equal vertical intervals (preferably equivalent to the contour interval of the original elevation data source), which yields a smoother profile (Wobus et al., 2006).

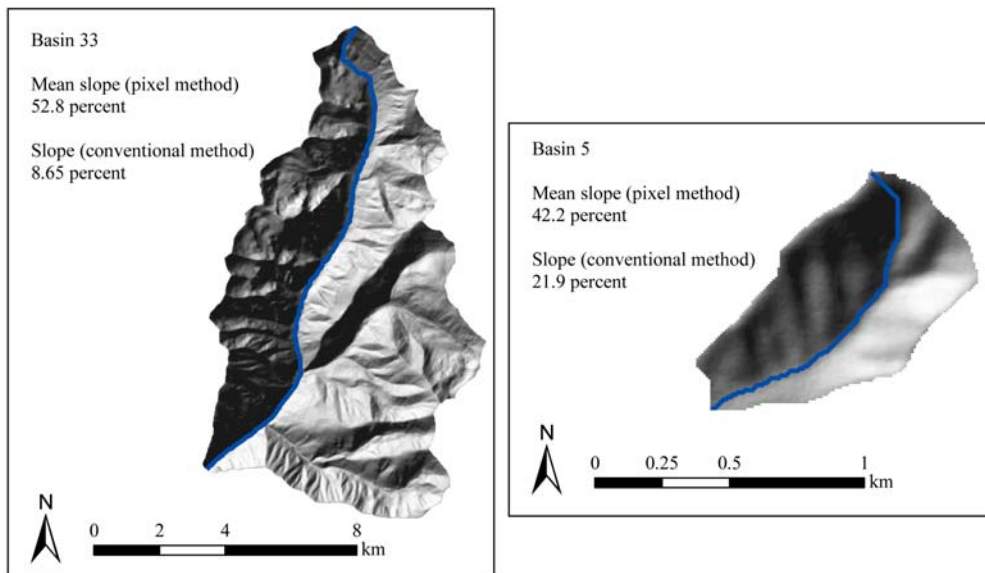


Figure 37. Above, basin 33 displays a wide difference between slope calculated using the pixel method and slope calculated using the conventional method. Basin 33 has several different catchment compartments. Below, basin 5 displays a smaller difference between slope calculated using the pixel method and slope calculated using the conventional method. Basin 5 is a simple basin, without any major subbasins. Note the very different scale bars for the two basins.

Time Scale of Analysis

Landforms may be in equilibrium when one time scale is considered, but in disequilibrium when other time scales are considered (Bishop and Shroder, 2004a; Allen and Allen, 2005). Timescale of measurement affects rates that are measured in erosional systems or tectonically uplifting areas (Gardner et al., 1987; Keller and Pinter, 2002; Allen and Allen, 2005; Koppes and Montgomery, 2009). For instance, apparent erosional rates for glaciated catchments are at least an order of magnitude larger if shorter (1-100 yr) timescales are considered compared to lower rates when timescales of $>10^3$ yr are considered (Koppes and Montgomery, 2009). In other words, measured process rates are partly dependent on measured time scale, and a simple comparison of rates derived over different scales of measurement may not be valid without adding a corrective scaling factor (Gardner et al., 1987; Allen and Allen, 2005). Observed rates tend to decrease with increasing measured time interval (Gardner et al., 1987; Allen and Allen, 2005). Mountainous topography results from a combination of climatic, tectonic, and surface processes operating over a variety of spatial and temporal scales (Shroder and Bishop, 2004; Allen and Allen, 2005). Therefore, an attempt is made to quantify the time scale on which the landforms (alluvial fans and their drainage basins) in this study respond to changes in factors (particularly fault slip rate) that are measured by proxy.

Previous studies have indicated that $\sim 10^6$ years are required for evolution from a flat landscape to the stable relief of a normal fault block range (Ellis et al., 1999). Once this state is obtained, topographic adjustments to changes in fault slip rate occur on the

order of $\sim 10^5$ yr (Ellis et al., 1999; Allen and Densmore, 2000). Climate-driven changes have been cited as driving factors for evolving fan geometry on time scales of 10^2 - 10^4 yr (Allen and Densmore, 2000; Harvey et al., 2005). Fault slip rate and fault geometry drive changes in the associated systems on time scales of $>10^4$ yr (Leeder and Jackson, 1993; Allen and Densmore, 2000). Therefore, by quantifying morphometric variation of drainage basins as a proxy for fault slip rate, this study may be assessing changes that occur on time scales of at least 10^4 yr, and perhaps around 10^5 yr.

Applications to Seismic Hazard Assessment

In addition to providing information on the evolution of footwall drainages and the applicability to basin stratigraphy, results from studies of evolving normal faults can be useful for future earthquake hazard assessment studies. Segment boundary delineation is critical to earthquake hazard assessment as the length of rupture is related to the maximum expected earthquake magnitude (Wyss, 1979; Crone and Haller, 1991; Smith and Arabasz, 1991; Piety et al., 1992; Gorton, 1995; Keller and Pinter, 2002). Specifically, maximum potential earthquake magnitude is proportional to rupture area, with area of rupture being approximately the product of depth to earthquake focus and rupture (segment) length (Wyss, 1979; Crone and Haller, 1991). For example, based on historical seismicity, mapped fault segment lengths, and thickness of seismogenic crust, maximum proposed magnitudes for future earthquakes in eastern Idaho are usually around 7.5 (Smith and Arabasz, 1991). In areas lacking segment boundary models, there may be a high degree of uncertainty in the maximum segment length, and thus, uncertainty in the maximum expected earthquake magnitude.

Earthquake prediction is often based on tectonic activity that is, in the geologic context, rather recent. However, historical seismic activity may not accurately reflect the earthquake potential of an area due to the short time period of observation. The realm of neotectonics is concerned with post-Miocene deformation (Schumm et al., 2000). Some workers have argued that accumulation of displacement is episodic along fault zones, with alternating periods of earthquake clusters and relative tectonic quiescence, on time scales ranging from several thousands to millions of years (Gawthorpe and Leeder, 2000). On an even shorter time span, “active tectonics” has been defined by some to be “those tectonic processes that produce deformation of the Earth’s crust on a time scale of significance to human society” (Keller and Pinter, 2002, p. 2).

The challenge, therefore, is to incorporate information from a neotectonic time scale to accurately assess future seismic potential. Syntectonic sedimentation and landform evolution can be used to provide insight into tectonic activity across a range of timescales (Schumm et al., 2000). Traditional fault segmentation studies involve large amounts of (sometimes costly) fieldwork and (often costly) dating of material in offset deposits. Detailed mapping and analysis of fault scarp morphology, offset surfaces, deposits exposed in trenches, and range-front morphology as well as relative and numerical dating of material in offset surfaces are typical methods employed (Haller, 1988; Turko, 1988; Baltzer, 1990; Janecke, 1993; Gorton, 1995; Hemphill-Haley, 2000). These studies allow the time frame under consideration to be extended past the instrumental seismic record.

This study proposes that drainage evolution in the extensional footwall is tied to segmentation and associated variations in fault slip rate and offset. If appropriate

drainage basin metrics can be identified that vary systematically with distance from segment boundary, morphometric analysis of the extensional footwall could be added as a method in identifying segments along a large normal fault. Since three-dimensional drainage basin geometry reflects processes operating on tectonic time scales (10^4 - 10^5 yr), morphometric analysis of drainage basins can serve as a proxy for tectonic activity over that time period (Harkins et al., 2005; Singh and Jain, 2009).

Geomorphometry is defined as the quantitative characterization of topography (Keller and Pinter, 2002), or more specifically, as “the quantitative characterization of geometric and topological relationships of the landscape” (p. 22, Bishop and Shroder, 2004a). The strength of geomorphometry in informing studies of active tectonics is the capability to determine a large amount of information over a broad geographic area of interest from elevation data, aerial photos, and/or topographic maps (Keller and Pinter, 2002; Wobus et al., 2006). With the increasing availability of digital elevation data worldwide and at increasingly higher resolutions, footwall morphometry could be a less costly and time consuming way to identify fault segments and segment boundaries in new areas, and thus, inform estimates of maximum expected earthquake magnitudes. However, the parameters investigated in this study show such weak (if any) segment-scale patterns that no recommendation for use of these morphometric parameters to delineate segment boundaries or determine seismic hazard is warranted.

Future Work

At the scale of the Lemhi Range, future work could include further data analysis of the existing dataset as well as collection of auxiliary data. Future work with the

existing dataset could include an analysis of residuals from fan area and fan slope regressions similar to work done by Al-Farraj and Harvey (2005) and Harvey (2005). Additional data could be collected – in particular, fan slope could be measured in the field using GPS technology along the most recent active lobe of each individual fan. Other morphometric parameters, including k-values (the stream gradient index, a measure of stream power), bifurcation ratios, number of first order streams, and range front curvature could be measured and compared against along-strike distance to fault segment boundary (with similar methodology to Mather and Hartley, 2006). Perhaps these or other morphometric parameters change along strike more strongly as a function of distance from fault segment boundary than those addressed in this study. Any morphometric parameters that are found to systematically vary with distance from fault segment boundaries in the Lemhi Range might spur investigation in other large, active, normal-fault bounded ranges. However, the lack of strong trends in the data collected and analyzed in this study indicate that further study may not be strongly warranted.

More generally, this study represents an application of geographic information technology to test geomorphic hypotheses. Geographic information technology (GIT) can be used in studies covering topics as diverse as hazard assessment, water resource potential, and mountain agriculture (Shroder and Bishop, 2004). GIT represents an improvement to traditional geomorphic studies by providing a framework for storage, manipulation, and analysis of large amounts of data, including spatial reference of data sets, visualization, rapid quantitative topographic analysis, and coupling with dynamic landscape evolution models (Shroder and Bishop, 2004). As stated by Goodchild (2004, p. xviii), “The rapid advances that have been made in terrain representation now underpin

a host of GIS applications in rugged environments, particularly mountains. Researchers are asking much more sophisticated and interesting questions: what are the fundamental spatial components of landscape, and can we develop automated methods for extracting them...what is the relationship between a representation and the processes that modify terrain, and what representations are most useful for understanding process?"

Recent studies have used GIS technology to test geomorphic hypotheses. A GIS erosion model based on elevation and precipitation data was created to test the "tectonic aneurysm" hypothesis in the Himalaya (Finlayson et al., 2002). Stream gradient index values have been related to uplift rate, and extrapolation from areas of known uplift rates to areas of unknown uplift rates has been accomplished using stream gradient index as a proxy (Mather and Hartley, 2006). Since it is easier to derive stream gradient values than obtain direct uplift estimates, this method may provide a first approximation of uplift rates for wide areas (Mather and Hartley, 2006).

Future structural and geomorphic studies will be assisted not only by new geographic information systems, but by new data products as well, including remotely sensed imagery and elevation data. Novel approaches with manipulating the new information can yield first-pass answers to tectonic, geomorphic, and natural hazard questions. For instance, the Space-Shuttle Radar Topography Mission provided a 90m resolution digital elevation model for land areas between 60°N and 56°S, representing coverage of 80% of Earth's land (Spencer, 2010). Investigation of these elevation data provided structural information on footwall curvature, fault curvature, and extension direction for three active or recently active extensional detachment faults; results from the study will be used to guide future field-based structural and thermochronologic

investigations (Spencer, 2010). Geodetic laser scanning offers another new possibility for increasing data coverage (Carter et al., 2007). A laser pulse is sent onto the surface to be mapped, and the reflected pulse is used to obtain the location of the reflector relative to the laser source (Carter et al., 2007). The instrument can be ground-based, airborne, or even space-based; for an airborne instrumentation, the obtained coordinates can be accurate to 5-10 cm vertically and 20-30 cm horizontally on a few-decimeter-scale grid (Carter et al., 2007). Filters can be used to extract a “bare-Earth” elevation dataset from the data obtained, allowing high-resolution elevation data coverage of heavily vegetated areas (Carter et al., 2007). The Puget Sound Lidar Consortium used this technology to more finely delineate scarps, terraces, and other fault-related features and even identified some previously unrecognized faults (Carter et al., 2007). These studies (Carter et al., 2007; Spencer, 2010) used GIS technology to analyze elevation data in order to obtain structural and geomorphic information.

In contrast, this study of the Lemhi Range did not result in interpretations strong enough for practical applications such as seismic hazard assessment. The relationships between the tested morphometric parameters and location along fault segment were very weak. However, the datasets, software, and methods used in this study represent an example of how GIS and high-resolution data can be used to test geomorphic and tectonic hypothesis. The case studies described in the previous paragraph represent outcomes where observed relationships were strong enough to have practical applications (Carter et al., 2007; Spencer, 2010). The combination of high-resolution data coverage of large areas with evolving software products and methods will allow more thorough testing of tectonic and geomorphic hypotheses than was possible during prior decades.

CONCLUSIONS

1) Drainage basins closer to fault segment boundaries tend to have higher Melton ratios, lower elongation ratios (more elongate shape), higher fan area to drainage basin area ratios, higher hypsometric integrals, and straighter (less sinusoidal) hypsometric curves than drainage basins located near fault segment interiors. Of the morphometric parameters examined, drainage basin Melton ratio varies most strongly as a function of drainage basin distance to nearest fault segment boundary. However, none of the morphometric parameters are strongly correlated with distance to fault segment boundary. Thus, while the observed relationships can influence conceptual models of extensional footwall development and hangingwall basin stratigraphic evolution, the relationships are not considered strong enough for purposes such as delineation of segment boundaries for seismic hazard assessment. Other morphometric parameters (including stream gradient index, bifurcation ratios, and range front curvature) might prove more useful for purposes requiring stronger relationships between predictor and response variables at a high level of confidence. Future studies should determine which, if any, morphometric parameters are appropriate.

2) Slope and area relationships for alluvial fans and fan drainage basins on the western side of the Lemhi Range agree with previous studies. Fan area is directly correlated to drainage basin area, and the relationship is strong. The relationships between fan slope and drainage basin area and between drainage basin slope and drainage basin area are much weaker, but the correlation direction of the relationships mirrors

those of studies of fan-basin relationships in other areas. Fan slope and drainage basin area are inversely correlated. Drainage basin slope and area are inversely correlated when drainage basin slope is calculated conventionally.

3) Conventional slope calculation methods do not yield results equivalent to standard grid-based zonal slope calculations. Slope is conventionally calculated by dividing drainage basin relief by basin length. Grid-based zonal slope calculations determine the mean slope of all DEM pixels in the drainage basin. The conventional methods were developed with topographic maps as the available data product. With development of newer, higher resolution data products (digital elevation models derived from various data sources), the appropriate definition of “slope” in various contexts should be reexamined. While conventional methodology can be used with newer data products, the availability of additional methods raises a question. Which definition of slope is most related to gravitationally-driven hydrogeomorphic processes? The most process-related definition of slope within the context of a given study is the value that should be calculated, though alternate slope values may need to be calculated as well to compare results with studies using other slope definitions. Conventional slope reflects main channel slope processes, while grid-based zonal slope incorporates hillslope-scale processes as well as channel slope processes.

4) GIS technology, combined with the availability of high-resolution elevation data, allows rapid morphometric characterization of large areas at relatively low cost. These results can be especially useful in the initial planning stages of a project to highlight zones of particular focus for fieldwork within the area of interest. The remote

nature of the data collection and analysis is also useful for areas where fieldwork is prohibitively costly for financial, logistical, and/or political reasons.

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APPENDICES

APPENDIX A

DATA AND SOFTWARE SOURCES

Data

DEM source information:

Title: National Elevation Dataset (NED), edition 2, 1/3 arc-second

Originator: U.S. Geological Survey

Publication date: 2009

Original Projection: (unprojected) geographic coordinates, NAD83

Data used in this project obtained: May 30, 2010 and September 8, 2010

Available (3/24/11) at: <http://seamless.usgs.gov/>

See also:

<http://ned.usgs.gov/>

Gesch, D., Oimoen, M., Greenlee, S., Nelson, C., Steuck, M., and Tyler, D., 2002, The National Elevation Dataset: Photogrammetric Engineering and Remote Sensing, v. 68, no. 1, p. 5-11.

Orthophoto source information:

Title: Digital Orthoimagery Series of Idaho (2004, 1-meter, Natural Color)

Originators: USDA-FSA-APFO Aerial Photography Field Office

Publication date: Feb 10, 2005

Original Projection: UTM NAD83

Data used in this project obtained: April 2, 2010 and October 7, 2010

Available (3/24/11) at: http://inside.uidaho.edu/webapps/search/iso_browse.aspx

Elevation source information for Figure 1 (Western US) – this dataset not used in analysis, only used for reference figure

Title: GTOPO30, 30 arc-second (1 km) resolution

Originator: U.S. Geological Survey's Center for Earth Resources Observation and Science

Publication date: late 1996

Original Projection: North American Lambert Azimuthal Equal Area

Data used in this project obtained: April 15, 2011

Available (4/25/11) at:

http://eros.usgs.gov/#/Find_Data/Products_and_Data_Available/gtopo30_info

Open-Source Software

R: The R Project for Statistical Computing. Version 2.12.2. Available at: <http://www.r-project.org/>

SAGA: System for Automated Geoscientific Analysis. Versions 2.0.4 and 2.0.6

Available at: <http://www.saga-gis.org/en/index.html>

APPENDIX B

EARTHQUAKE CLASIFICATION SCALES

A variety of scales are used to report earthquake magnitudes. Listed below are several different magnitude definitions (Engdahl and Rindhard, 1991).

M_W = estimate of seismic moment

M_S = surface-wave magnitude (maximum amplitude)

M_L = local or Richter magnitude (maximum amplitude 100 km from epicenter, see Keller and Pinter, 2002)

M_D = signal duration

m_b = body-wave magnitude (maximum amplitude)

MM = Maximum Modified Mercalli intensity

Richter local magnitudes (M_L) are determined from seismographs located closer to the epicenter than epicenters used to determine surface-wave magnitudes (M_S).

However, since M_S and Richter M_L can be saturated for larger earthquakes, the moment magnitude (M_W) is preferred for larger earthquakes, particularly those greater than about 7.5 (Keefer, 1999). M_W is also preferred due to its wider applicability and physical basis (Keller and Pinter, 2002). Moment magnitude is the product of average fault offset, fault area, and rigidity of material along the fault (Keefer, 1999). Specifically,

$$M_W = 2/3 \log (M_O) - 0.6$$

$$\text{where } M_O = S A \mu$$

where S = the average slip on the fault (in meters) and A = the area (in square meters) of the fault plane that ruptured and μ = shear modulus of rock that failed .

The Modified Mercalli intensity scale has allowed estimation of historic earthquakes for which there are not seismographic records. The following scale is an abridged version of the Modified Mercalli intensity scale and is quoted directly from Qamar and Stickney (1983, p. 18).

Modified Mercalli Intensity Scale of 1931 (Abridged)

- I. Not felt except by a very few under especially favorable circumstances.
- II. Felt only by a few persons at rest, especially on upper floors of building. Delicately suspended objects may swing.
- III. Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize the motion as an earthquake. Standing motor cars may rock slightly. Vibration similar to that of a passing truck.
- IV. During the day, felt indoors by many; felt outdoors by few. At night, some awakened. Dishes, windows, doors disturbed; walls make creaking sound. Sensation similar to that of a heavy truck striking building. Standing motor cars rock noticeably.
- V. Felt by nearly everyone; many awakened. Some dishes and windows broken; a few instances of cracked plaster; unstable objects overturned. Disturbance of trees, poles and other tall objects sometimes noticed. Pendulum clocks may stop.
- VI. Felt by all; many frightened and run outdoors. Some heavy furniture moved; a few instances of falling plaster or damaged chimneys. Damage slight.
- VII. Everyone runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built, ordinary structures; considerable in poorly built or badly designed structures. Some chimneys broken. Noticed by persons driving motor cars.
- VIII. Damage slight in specially designed structures; considerable in ordinary, substantial buildings, with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Destruction of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. Sand and mud ejected from ground in small amounts. Changes in well water. Persons driving motor cars disturbed.
- IX. Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; damage great in substantial buildings, with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken.
- X. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations. Ground badly cracked. Rails bent. Considerable amount of landslides from river banks and steep slopes. Shifted sand and mud. Water splashed over banks.
- XI. Few, if any, masonry structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.
- XII. Damage total. Waves seen on ground surface. Lines of sight and level distorted. Objects thrown upward into the air.

APPENDIX C

IDAHO TRANSVERSE MERCATOR PROJECTION

All spatial data were reprojected to the Idaho Transverse Mercator Projection, and all new datasets created were created in this projection.

Table C.1. Projection parameters for the Idaho Transverse Mercator Projection

Idaho Transverse Mercator (IDTM83)	
Units	Meters
Datum	NAD83
Scale factor at central meridian	0.99960
Central meridian	-114° 00' 00"
Latitude of origin	42° 00' 00"
False easting	2,500,000
False northing	1,200,000
Vertical datum	NAVD88

Parameters from: Idaho Department of Water Resources, accessed Feb 28, 2011, http://www.idwr.idaho.gov/GeographicInfo/gisdata/IDTM/projection_parameters.htm

In this thesis, Idaho Transverse Mercator Projection is referred to as “ITM”, not “IDTM83”.

APPENDIX D

MAPPED DRAINAGE BASINS AND FANS

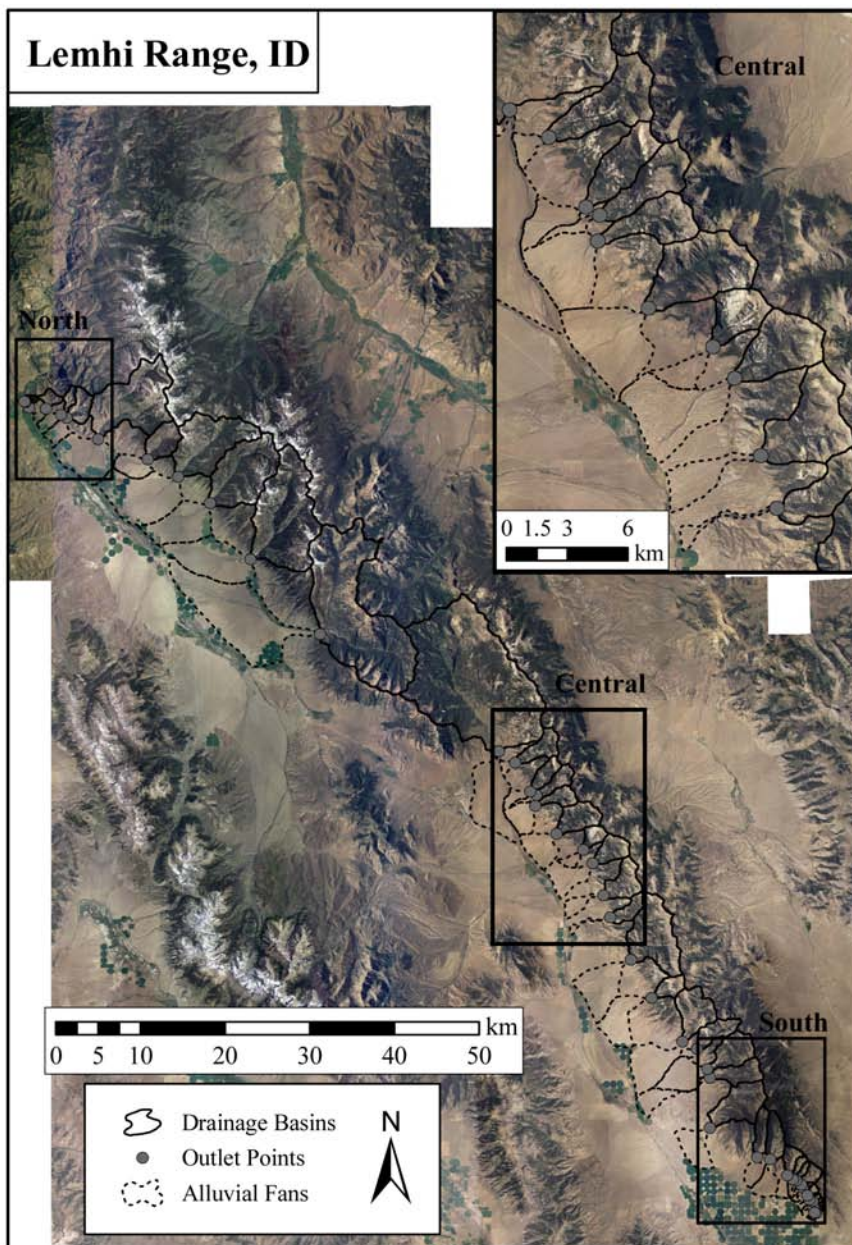


Figure D.1. Mapped alluvial fans, drainage basins, and outlet points with orthophoto as reference. Upper right inset is the central portion of the range, areas labeled “North” and “South” are shown in Figure D.2 and Figure D.3.

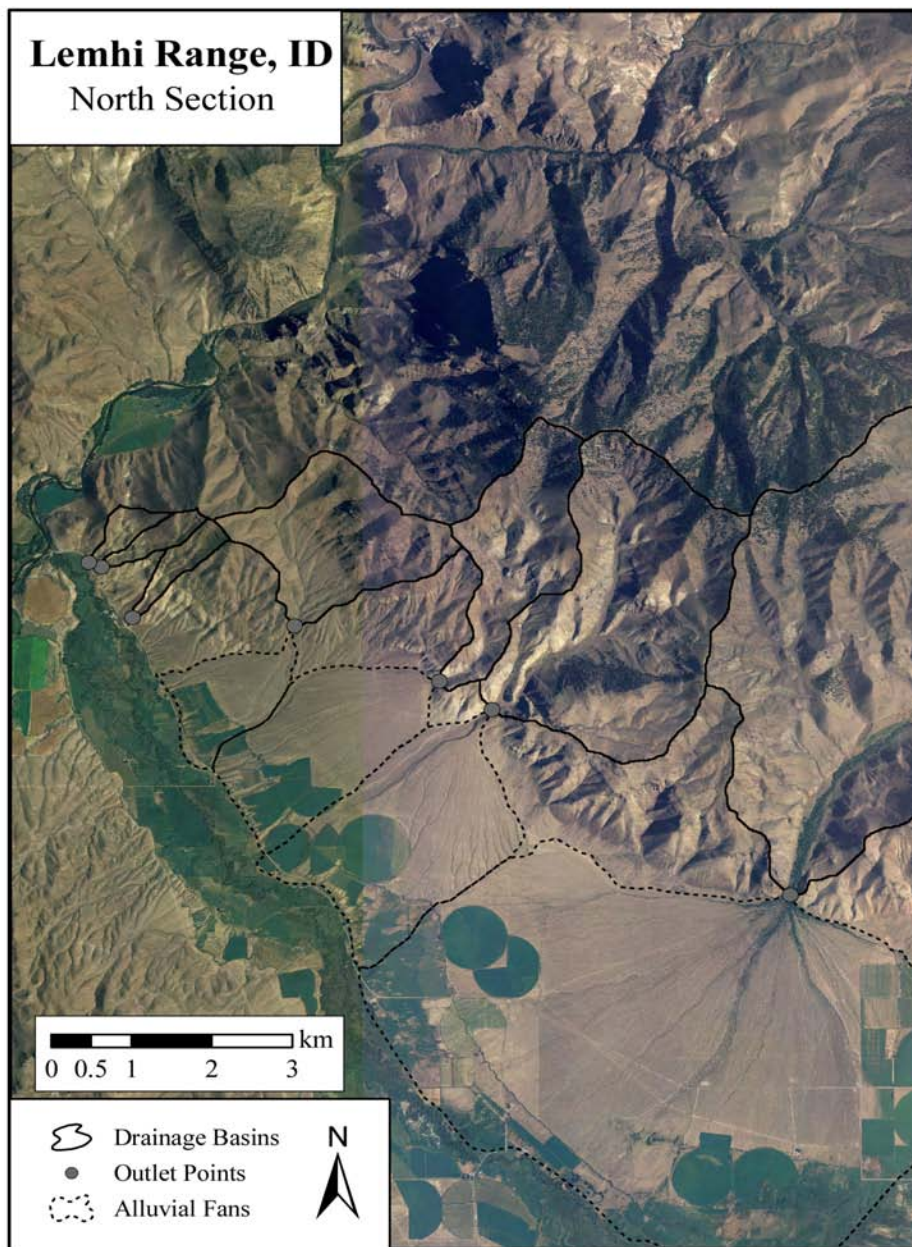


Figure D.2. Mapped alluvial fans, drainage basins, and outlet points with orthophoto as reference for the northern portion of the Lemhi Range. For location of this figure, see Figure D.1.

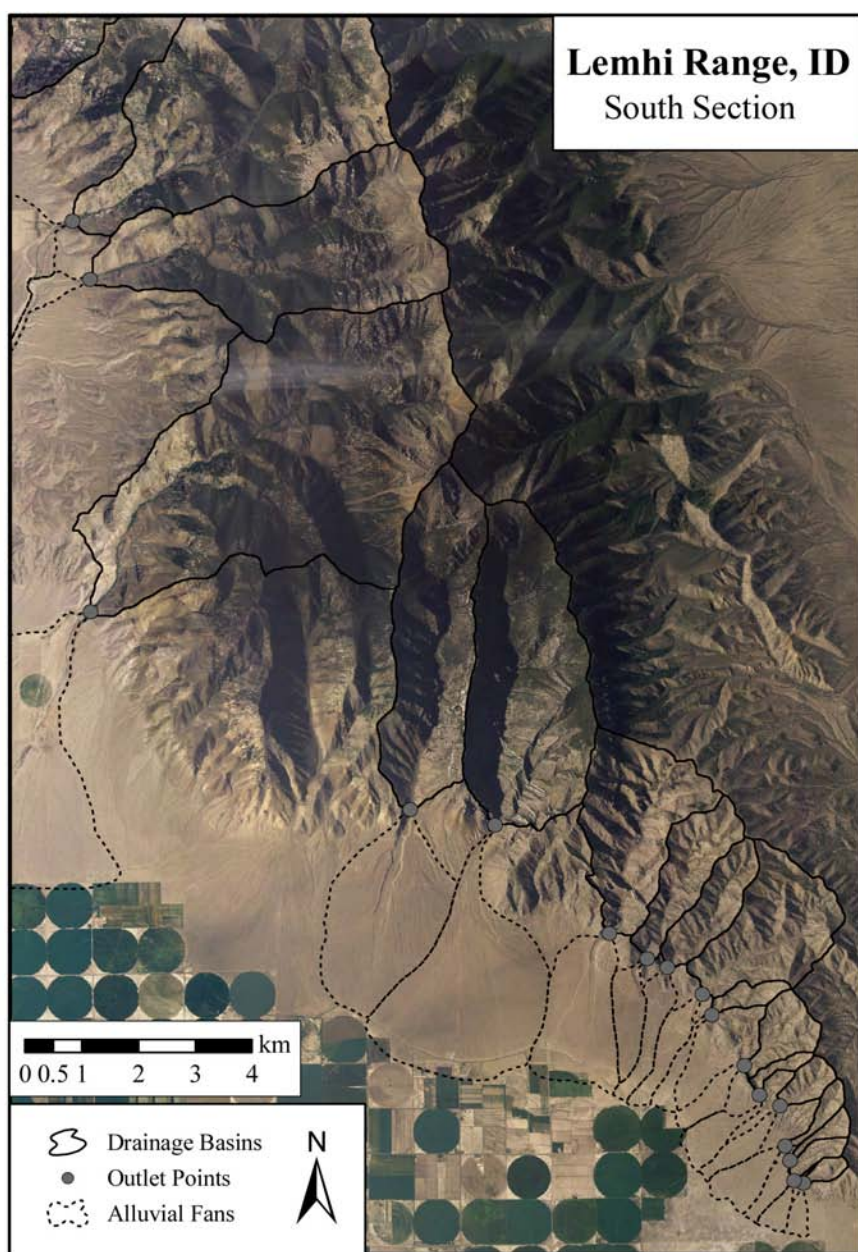


Figure D.3. Mapped alluvial fans, drainage basins, and outlet points with orthophoto as reference for the southern portion of the Lemhi Range. For location of this figure, see Figure D.1

APPENDIX E

DRAINAGE BASIN AND FAN MORPHOMETRICS

Table E.1. Summary statistics for elevation, slope and area for all 43 drainage basins. “Pixel” and “conventional” denote the two different slope calculation methods used, see “Drainage Basin Slope and Fan Slope” section in Methodology chapter and “Two Different Methods of Calculating Drainage Basin Slope” section in Discussion chapter for discussion of the two methods. All values are reported to three significant figures.

	Minimum	Mean	Maximum
Minimum elevation (m)	1430	1810	2320
Mean elevation (m)	1540	2280	2890
Maximum elevation (m)	1720	2780	3710
Relief (m)	218	966	1670
Minimum slope (pixel, degrees)	(0)	0.235	2.28
Mean slope (pixel, degrees)	13.6	24.7	32.3
Maximum slope (pixel, degrees)	32.2	52.5	72.2
Minimum slope (pixel, percent)	(0)	0.438	3.99
Mean slope (pixel, percent)	24.9	47.6	65.7
Maximum slope (pixel, percent)	63.0	145	312
Slope (conventional, percent)	5.48	17.4	26.6
Area (km ²)	0.231	19.5	190

Table E.2. Summary statistics for elevation, slope and area for the 40 alluvial fans sourced by drainage basins 1-40. “Pixel” denotes the slope calculation method used, see “Drainage Basin Slope and Fan Slope” section in Methodology chapter and “Two Different Methods of Calculating Drainage Basin Slope” section in Discussion chapter for discussion of other methods. All values are reported to three significant figures.

	Minimum	Mean	Maximum
Minimum elevation (m)	1430	1610	1970
Mean elevation (m)	1470	1690	2050
Maximum elevation (m)	1500	1850	2320
Relief (m)	38.0	237	599
Minimum slope (pixel, degrees)	(0)	(0)	(0)
Mean slope (pixel, degrees)	0.886	3.62	7.26
Maximum slope (pixel, degrees)	5.70	31.2	63.4
Minimum slope (pixel, percent)	(0)	(0)	(0)
Mean slope (pixel, percent)	1.55	6.38	12.9
Maximum slope (pixel, percent)	9.98	66.35	200.
Area (km ²)	0.230	11.4	82.0

Table E.3. Elevation and location statistics for the 43 drainage basins. Basin outlet distance to closest segment boundary is not reported for basins 31 and 32, as these basins are located within segment boundary zones. Outlet distance along-strike to the southern range tip is reported for all basins. All values are reported to three significant figures.

Basin	Min. Elevation (m)	Mean Elevation (m)	Max. Elevation (m)	Outlet Distance to Closest Segment Boundary (km)	Outlet Distance to Southern Range Tip (km)
1	1510	1590	1720	1.30	1.30
2	1500	1580	1740	1.43	1.43
3	1520	1640	1780	1.76	1.76
4	1530	1670	1810	2.03	2.03
5	1570	1700	1860	2.63	2.63
6	1570	1690	1860	3.04	3.04
7	1570	1740	1990	3.62	3.62
8	1580	1750	1990	4.67	4.67
9	1580	1820	2100	4.97	4.97
10	1620	1890	2120	5.68	5.68
11	1620	1940	2190	6.15	6.15
12	1610	2040	2640	7.07	7.07
13	1740	2390	3120	9.52	10.3
14	1790	2480	3280	8.49	11.3
15	1760	2460	3280	0.848	19.0
16	2000	2640	3260	2.92	25.6
17	2010	2640	3260	4.00	26.7
18	2060	2820	3250	6.81	29.5
19	1930	2630	3280	8.55	31.3
20	2250	2890	3420	2.13	40.1
21	2070	2740	3710	4.01	45.0
22	2170	2730	3320	2.37	52.5
23	2140	2790	3330	4.90	55.0
24	2160	2750	3260	9.45	59.6
25	2320	2880	3520	11.3	61.4
26	2060	2720	3520	11.4	65.1
27	2100	2720	3210	7.02	69.5
28	2180	2780	3260	5.99	70.5
29	2130	2690	3260	5.29	71.2
30	2120	2750	3250	1.10	75.4
31	2010	2560	3310	--	76.6
32	1990	2690	3450	--	117
33	1850	2660	3310	10.9	128
34	1820	2580	3260	13.3	136
35	1800	2480	3260	8.32	141
36	1820	2470	3150	3.98	145
37	1660	2430	3340	2.40	152
38	1660	2050	2560	6.13	156
39	1690	2080	2560	5.09	157
40	1600	1870	2190	3.18	159
41	1460	1640	1890	1.15	161
42	1450	1630	1880	0.368	162
43	1430	1540	1780	0.241	162

Table E.4. Drainage basin slope statistics. All slopes from the “pixel” method except the last column (alternate slope), which is “conventional” method. See “Drainage Basin Slope and Fan Slope” section in Methodology chapter and “Two Different Methods of Calculating Drainage Basin Slope” section in Discussion chapter for discussion of different slope calculation methods. All values are reported to three significant figures.

Basin	Min Slope (degrees)	Mean Slope (degrees)	Max Slope (degrees)	Min Slope (percent)	Mean Slope (percent)	Max Slope (percent)	Alternate Slope (percent)
1	2.29	18.7	34.9	3.99	34.5	69.8	21.6
2	(0)	13.6	35.6	(0)	24.9	71.5	16.2
3	1.14	18.1	34.5	2.00	33.5	68.7	18.0
4	1.14	18.8	35.3	2.00	35.3	70.8	16.5
5	1.14	22.5	40.6	2.00	42.3	85.8	22.0
6	(0)	20.0	35.3	(0)	36.9	70.7	18.9
7	1.14	21.7	36.2	2.00	40.6	73.2	23.9
8	(0)	21.0	34.5	(0)	38.9	68.7	23.8
9	(0)	20.9	43.2	(0)	38.9	93.8	10.6
10	(0)	23.4	45.3	(0)	43.9	101	15.8
11	(0)	22.9	49.6	(0)	43.2	118	14.8
12	(0)	26.4	60.3	(0)	51.1	176	22.0
13	(0)	31.2	72.0	(0)	63.9	308	21.5
14	(0)	32.3	72.2	(0)	65.7	312	20.0
15	(0)	28.6	70.0	(0)	56.0	275	15.7
16	(0)	27.6	59.1	(0)	53.9	167	17.0
17	(0)	28.8	63.9	(0)	56.3	204	16.4
18	(0)	25.0	55.4	(0)	48.2	145	17.0
19	(0)	26.9	60.5	(0)	52.2	177	14.6
20	(0)	29.4	56.0	(0)	58.0	148	21.7
21	(0)	29.5	64.2	(0)	58.3	207	16.2
22	(0)	27.5	53.1	(0)	53.0	133	19.7
23	(0)	28.8	59.8	(0)	56.2	172	15.8
24	(0)	29.1	54.7	(0)	56.6	141	17.1
25	(0)	29.1	56.1	(0)	57.3	149	26.6
26	(0)	28.3	60.9	(0)	55.6	180	18.0
27	0.339	30.0	57.3	0.592	59.2	156	18.9
28	0.0577	26.7	58.3	0.101	51.9	162	15.0
29	0.0512	27.9	62.5	0.0893	54.0	192	17.6
30	0.327	27.2	61.9	0.571	52.9	187	15.4
31	0.00586	18.2	57.4	0.0102	33.7	156	5.48
32	0.00592	24.8	58.9	0.0103	47.6	166	8.93
33	0.00617	27.1	62.9	0.0108	52.8	195	8.65
34	0.00586	28.0	61.9	0.0102	54.6	187	10.2
35	0.00695	27.4	57.6	0.0121	53.2	157	11.2
36	0.00687	27.2	50.7	0.0120	52.4	122	19.7
37	0.00620	24.3	62.5	0.0108	46.7	192	10.9
38	0.259	23.4	54.9	0.388	44.4	142	18.2
39	0.201	25.1	44.1	0.352	47.8	96.8	17.5
40	0.478	19.8	45.9	0.834	36.9	103	18.3
41	0.179	19.2	44.9	1.00	35.7	99.7	24.0
42	0.774	20.0	42.2	1.92	36.9	90.8	26.4
43	0.539	17.3	32.2	0.941	31.4	63.0	22.4

Table E.5. Elevation statistics for the 40 alluvial fans (sourced from drainage basins 1-40). All values are reported to three significant figures.

Basin	Min. Elevation (m)	Mean Elevation (m)	Max. Elevation (m)
1	1460	1480	1510
2	1460	1470	1500
3	1460	1490	1520
4	1460	1480	1530
5	1460	1500	1570
6	1460	1490	1570
7	1470	1510	1570
8	1480	1500	1580
9	1480	1520	1580
10	1480	1530	1620
11	1480	1530	1620
12	1470	1520	1610
13	1470	1540	1740
14	1490	1620	1790
15	1510	1600	1760
16	1590	1690	2000
17	1670	1850	2010
18	1830	1890	2060
19	1610	1710	1930
20	1640	1790	2250
21	1690	1860	2070
22	1780	1880	2170
23	1790	1900	2140
24	1820	1950	2160
25	1910	2050	2320
26	1840	1910	2060
27	1880	1960	2100
28	1950	2030	2180
29	1910	1980	2130
30	1970	2020	2120
31	1870	1930	2010
32	1600	1770	1990
33	1580	1710	1850
34	1550	1670	1820
35	1500	1630	1800
36	1540	1670	1820
37	1460	1540	1660
38	1450	1530	1660
39	1450	1540	1690
40	1430	1490	1600

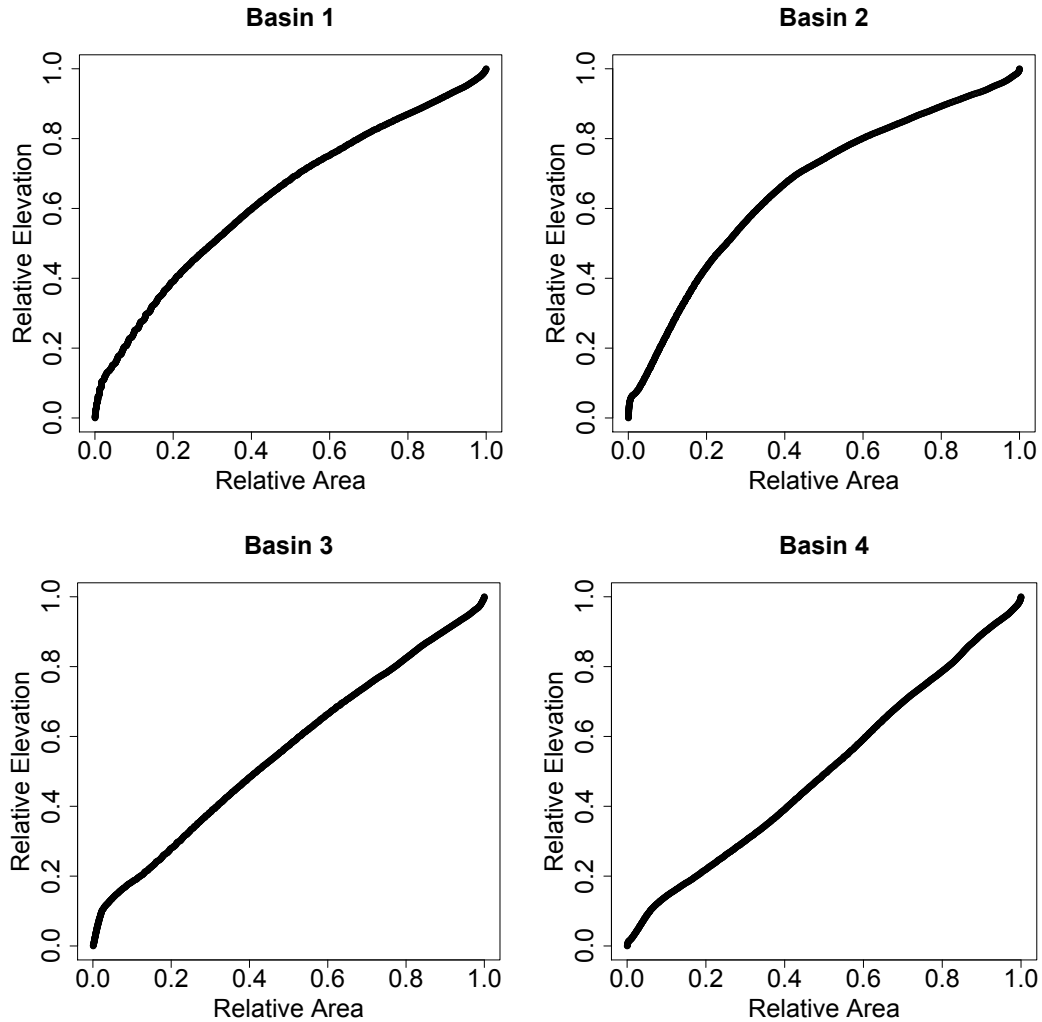
Table E.6. Alluvial fan slope statistics. All calculated with the “pixel” method, see “Drainage Basin Slope and Fan Slope” section in Methodology chapter and “Two Different Methods of Calculating Drainage Basin Slope” section in Discussion chapter for discussion of different slope calculation methods. All values are reported to three significant figures. Minimum slope (percent and degrees) is 0 for all fans.

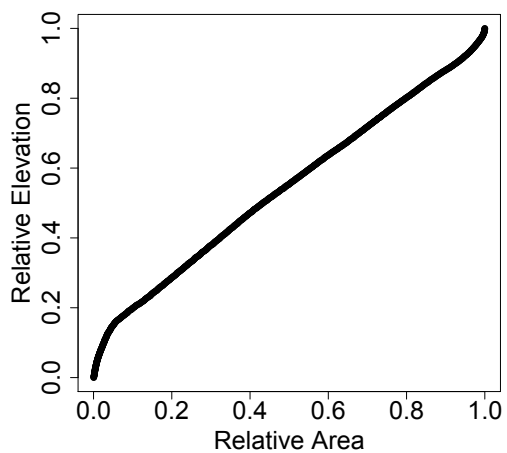
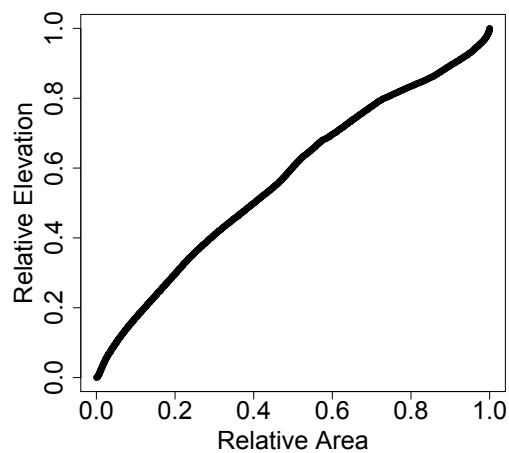
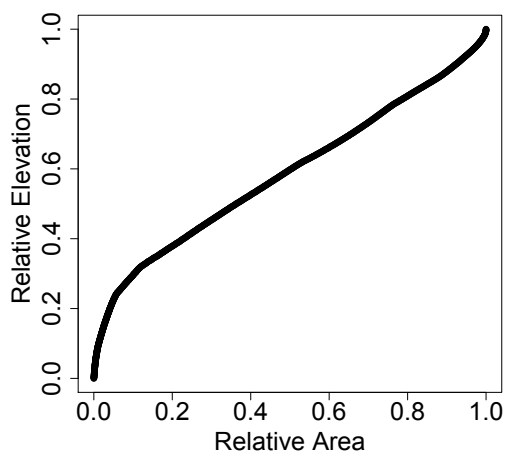
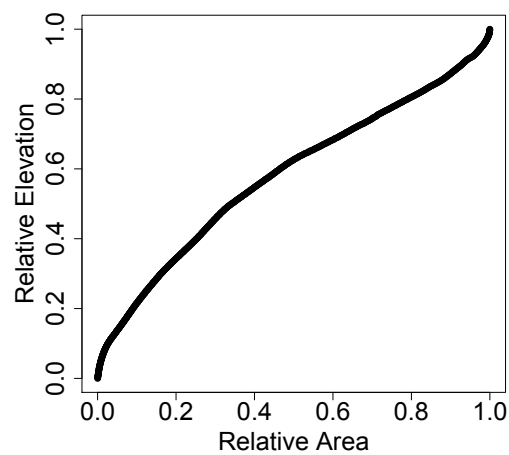
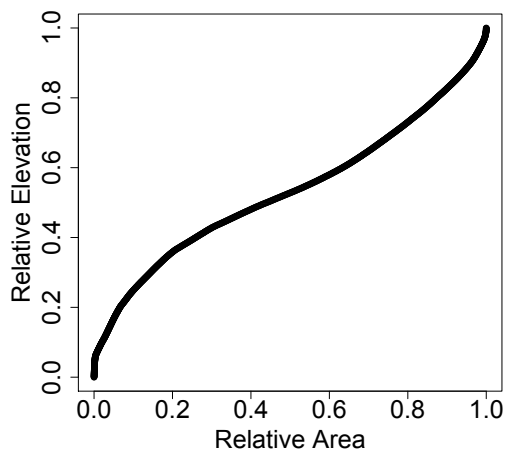
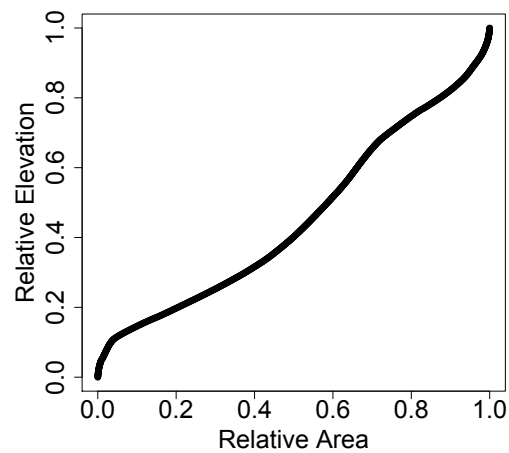
Basin	Mean Slope (degrees)	Max Slope (degrees)	Mean Slope (percent)	Max Slope (percent)
1	2.67	8.26	4.67	14.5
2	2.31	5.70	4.04	9.98
3	2.89	6.13	5.05	10.7
4	2.29	7.19	4.01	12.6
5	3.23	15.8	5.66	28.2
6	2.67	28.9	4.68	55.2
7	3.81	21.8	6.68	39.9
8	2.76	21.0	4.84	38.4
9	3.23	20.4	5.66	37.3
10	4.93	49.6	8.82	117
11	4.63	47.8	8.23	110
12	3.51	38.0	6.20	78.0
13	3.57	24.7	6.24	45.9
14	6.19	59.5	11.0	170
15	3.10	33.0	5.43	64.9
16	3.55	43.4	6.28	94.5
17	3.72	42.1	6.55	90.3
18	5.67	30.8	9.96	59.5
19	2.28	25.0	3.99	46.6
20	4.16	38.1	7.33	78.5
21	4.70	40.5	8.31	85.3
22	5.03	42.6	8.88	91.9
23	4.67	47.4	8.22	109
24	7.26	46.3	13.0	105
25	5.90	33.1	10.4	65.1
26	2.67	24.8	4.67	46.3
27	3.18	21.2	5.57	38.7
28	4.04	18.9	7.07	34.3
29	2.97	18.0	5.19	32.5
30	4.07	31.0	7.14	60.0
31	0.886	32.4	1.55	63.5
32	1.79	63.4	3.16	200
33	1.78	31.6	3.11	61.5
34	2.01	42.4	3.52	91.4
35	2.26	37.2	3.96	75.9
36	4.86	27.4	8.52	51.9
37	2.00	21.8	3.50	39.9
38	3.94	35.9	6.90	72.3
39	4.94	36.4	8.68	73.6
40	4.84	28.5	8.50	54.2

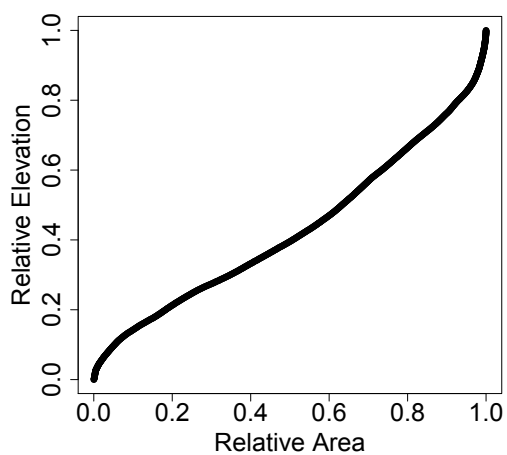
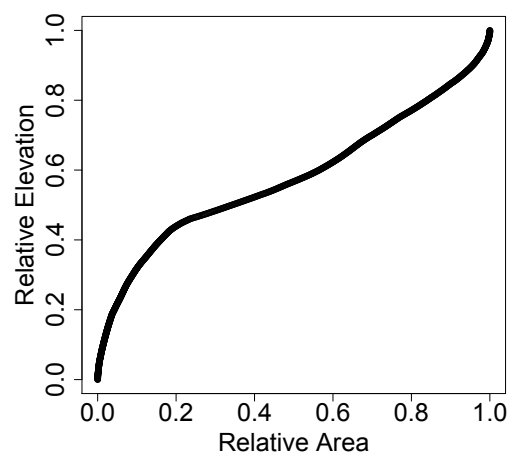
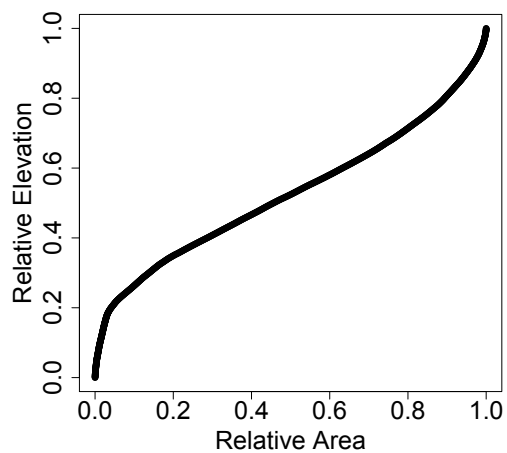
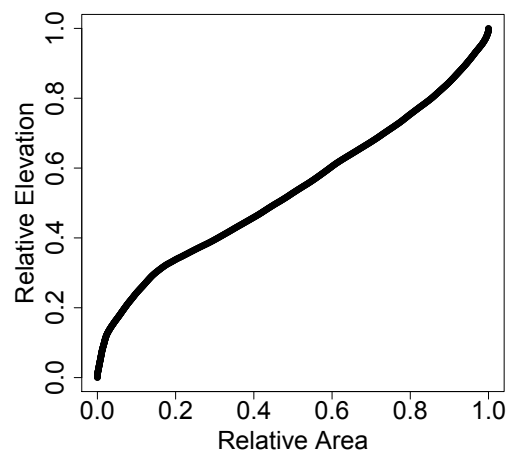
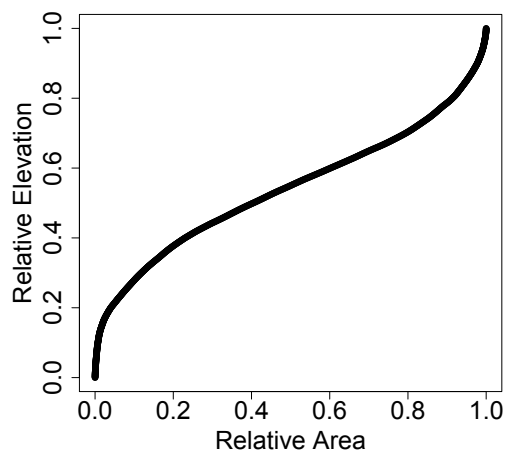
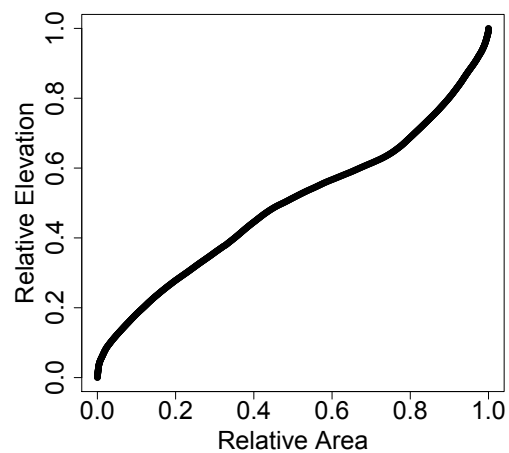
APPENDIX F

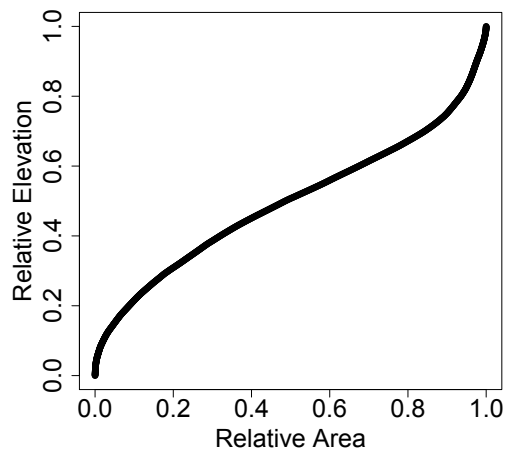
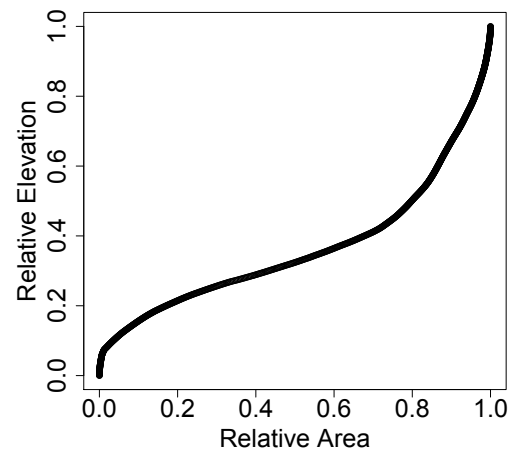
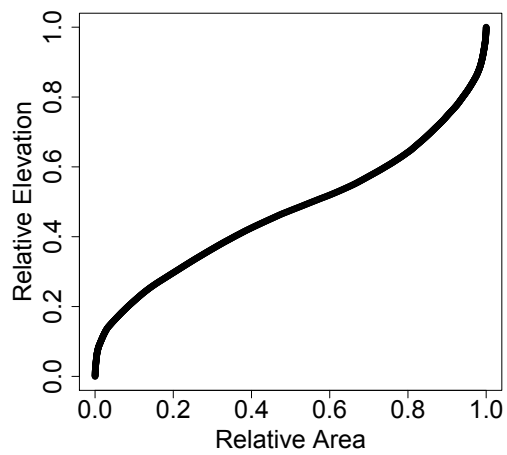
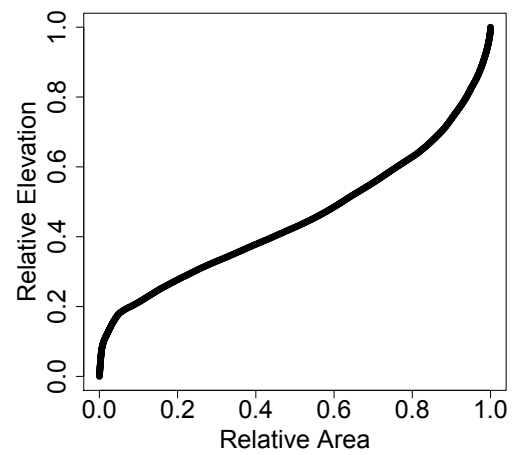
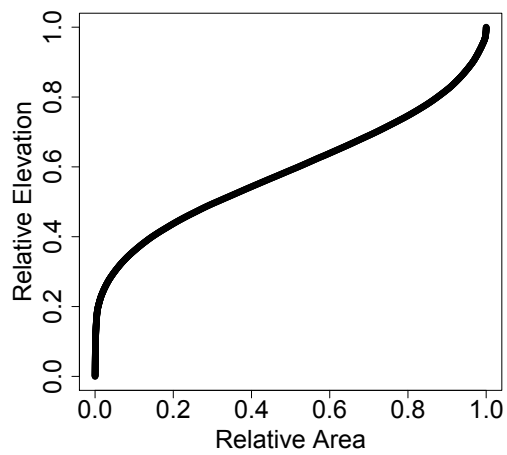
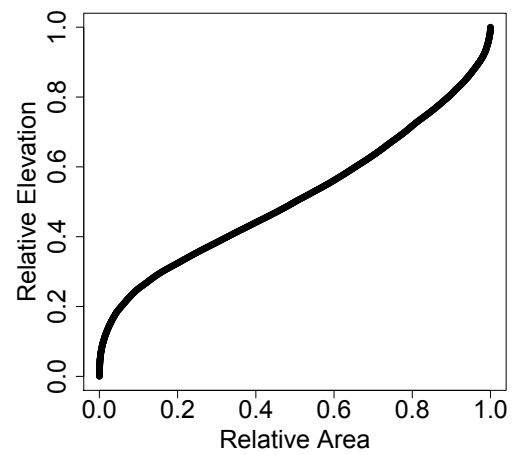
DRAINAGE BASIN HYPSONETRIC CURVES

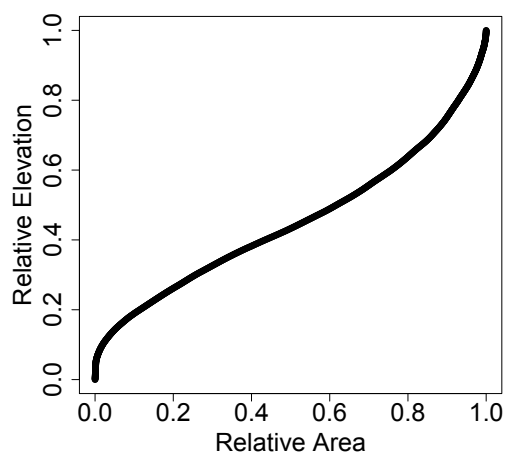
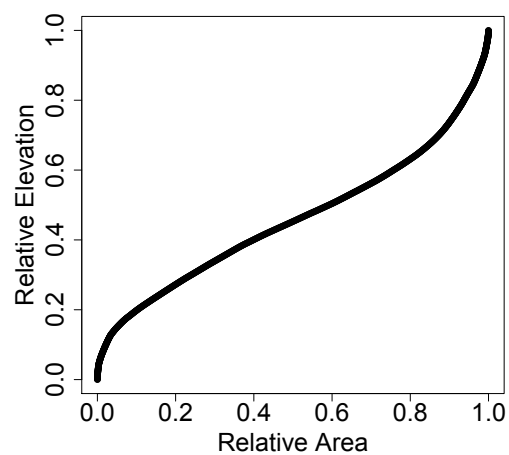
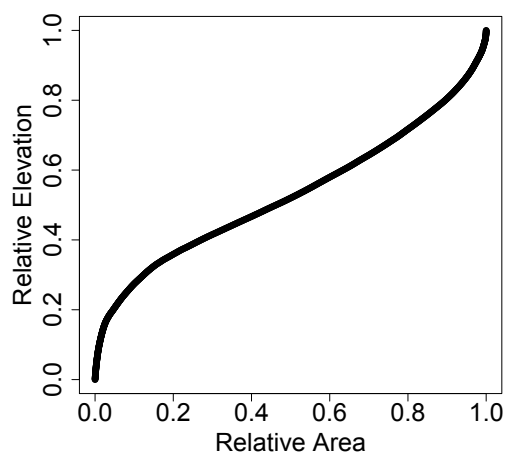
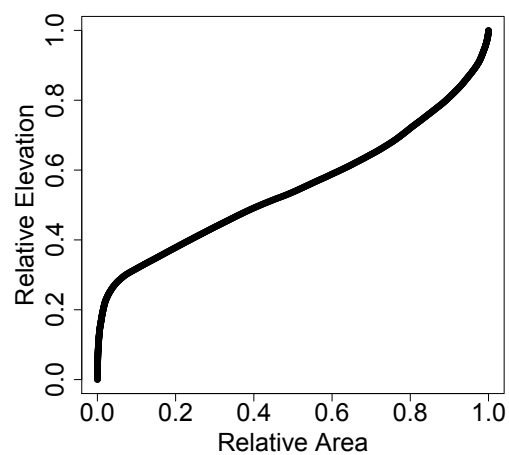
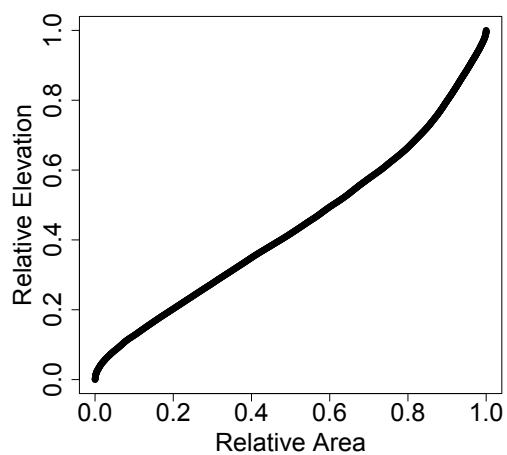
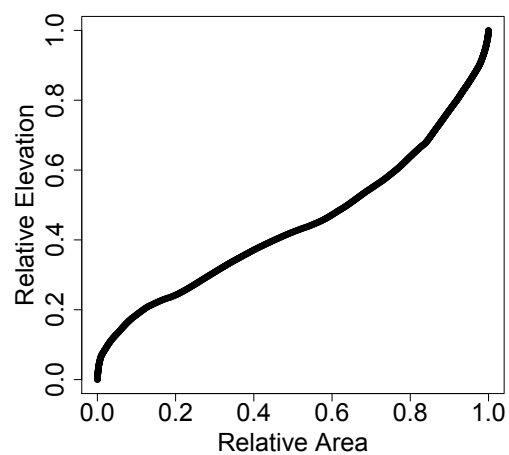
The following graphs are hypsometric curves for individual basins. Individual captions are omitted for clarity. On each graph, the vertical (y) axis is relative elevation (percent) and the horizontal (x) axis is relative area (percent). The area under the curve is defined as the hypsometric integral for that basin.

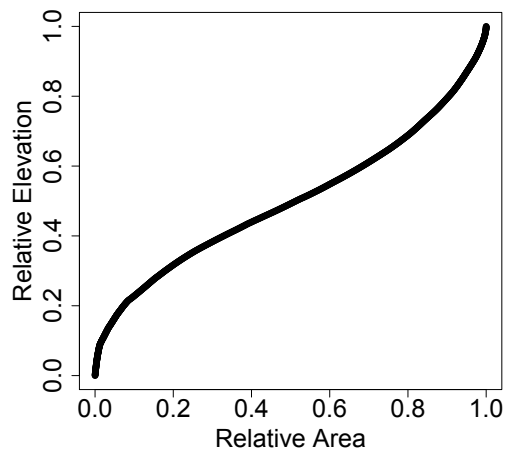
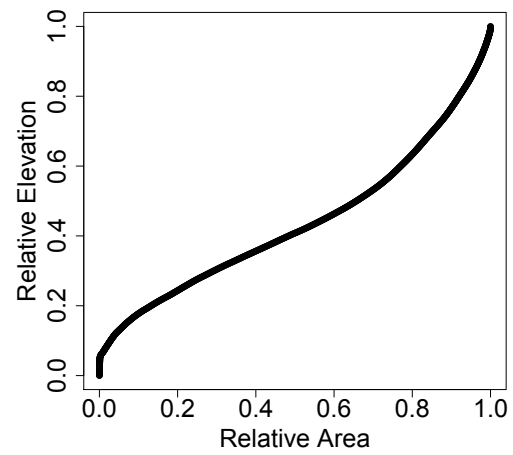
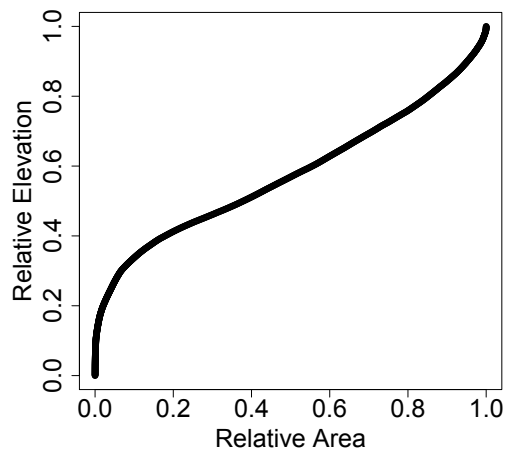
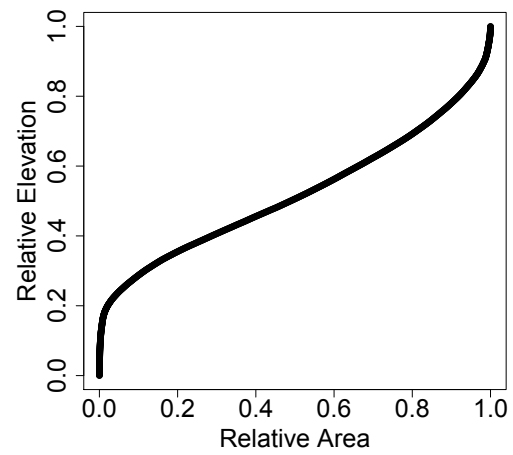
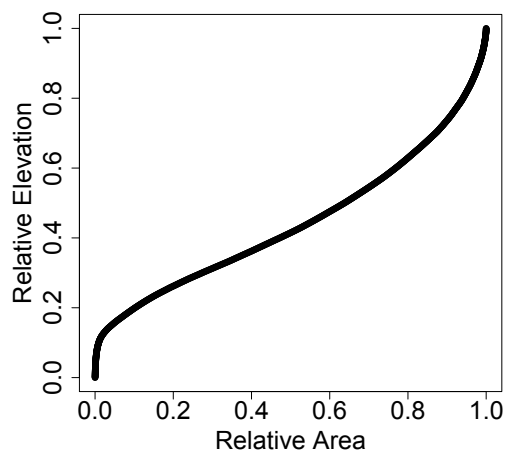
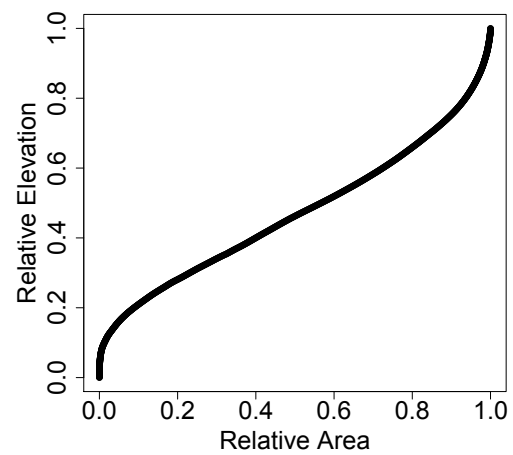


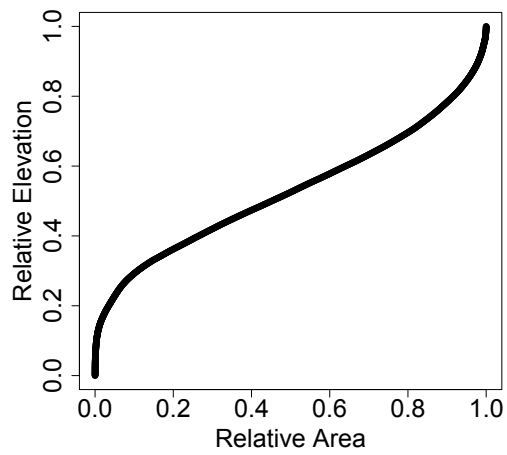
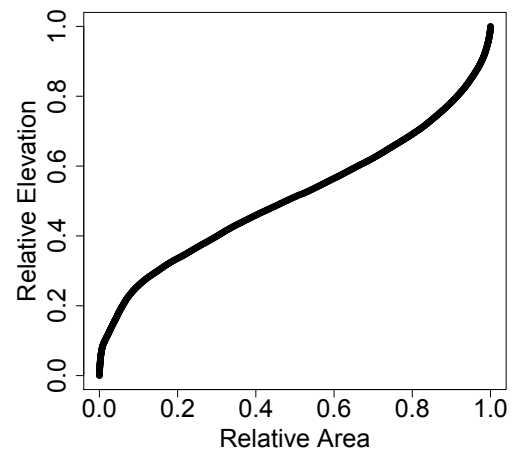
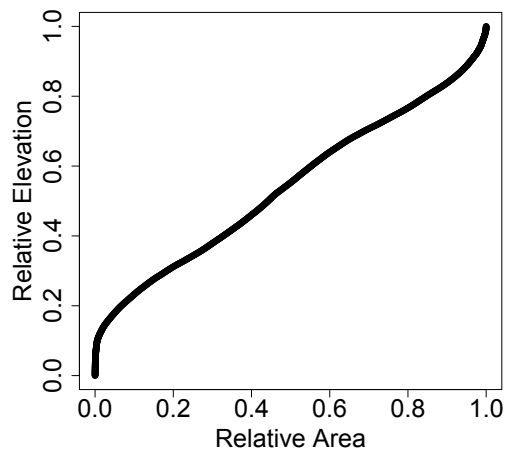
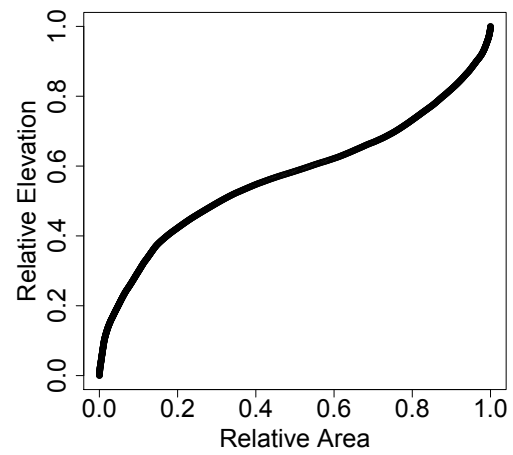
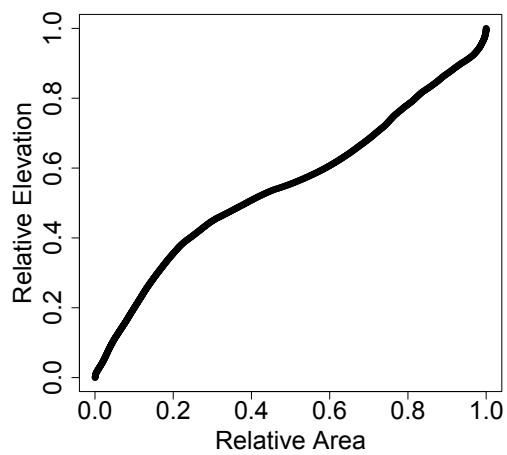
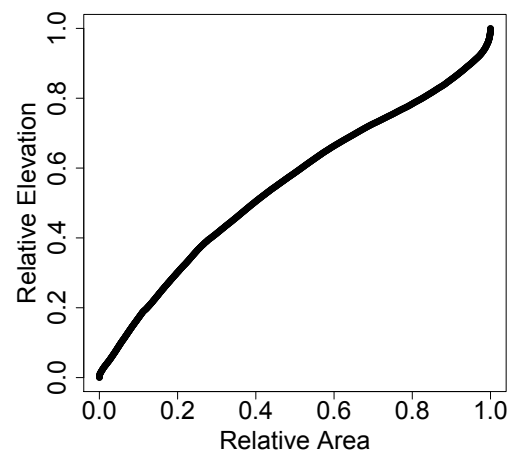
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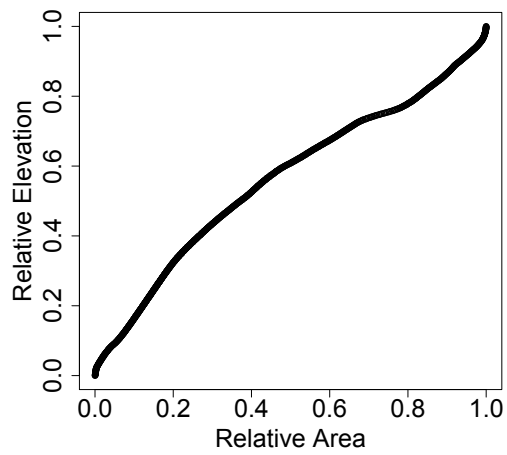
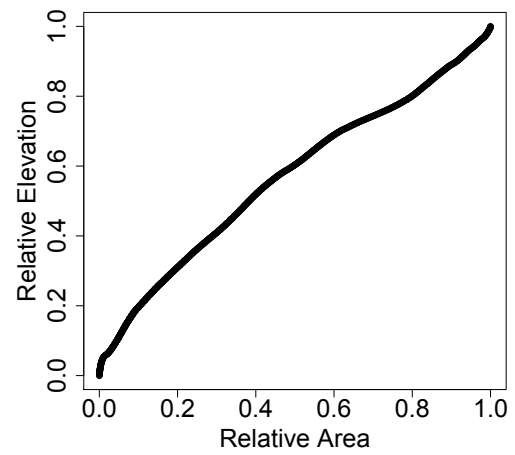
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Basin 17**Basin 18****Basin 19****Basin 20****Basin 21****Basin 22**

Basin 23**Basin 24****Basin 25****Basin 26****Basin 27****Basin 28**

Basin 29**Basin 30****Basin 31****Basin 32****Basin 33****Basin 34**

Basin 35**Basin 36****Basin 37****Basin 38****Basin 39****Basin 40**

Basin 41**Basin 42****Basin 43**