



Quaternary tectonic activity within the northern arm of the Yellowstone tectonic parabola and associated seismic hazards, southwest Montana
by Chester Allan Ruleman, III

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in
Earth Sciences
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Abstract:

The area of southwest Montana encompassing the Madison, Absaroka-Beartooth, and Bridger Ranges lies within the northern arm of the Yellowstone tectonic parabola (YTP). Detailed mapping along the Madison Range fault (MRF) and morpho-metric analyses were compared to the morphometry of two other faults in southwest Montana (Emmigrant/Deep Creek/Luccock Park fault - EDLF, and Bridger Range fault - BRF) in order to determine long-term tectonic activity rates and their correlation to the YTP. Morpho-metric analyses, including Ms, Vf mountain-front morphology, and the RTAC, quantify long-term tectonic activity rates of > 0.1 mm/yr for all three faults. However, the basal facet reconnaissance technique yields late Pleistocene slip rates for the BRF of 0.01 mm/yr. The basal facet analyses provide a more accurate late Pleistocene slip rate for the BRF. This possibly suggests that the BRF experienced pre-late Pleistocene slip rates an order of magnitude greater (> 0.1 mm/yr) than the late Pleistocene (< 0.01 mm/yr), which is not modeled by the YTP. Mid- to late-Pleistocene tectonic activity between these three faults has been greater southward towards the Yellowstone volcanic field. In addition to late Pleistocene heterogeneous spatial and temporal earthquake activity, pre-existing structural controls on Quaternary normal faulting suggest that a simplistic parabolic deformational model does not adequately explain the neotectonics between the faults under investigation and southwest Montana.

Along with the neotectonic implications, the calibration and application of morpho-metric analyses to the northern Basin and Range province, coupled with surficial geology and contemporary seismicity, derive a seismic hazard assessment for the MRF, EDLF, and BRF. The MRF and EDLF have late-Pleistocene to Holocene slip rates ranging from 0.1-0.5 mm/yr. Based on basal facet analyses, the BRF has late-Pleistocene to Holocene slip rates of 0.01 mm/yr, but morpho-metric analyses (i.e., RTAC) quantify long-term average Quaternary slip rates of > 0.1 mm/yr. All three faults possibly have long-term tectonic activity rates > 0.1 mm/yr, with bifurcations in slip rates over periods > 100 ky. Maximum earthquakes for the three faults range from Mw 6.5-7.2, in accordance with historic seismicity of the Basin and Range province.

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MONTANA STATE UNIVERSITY
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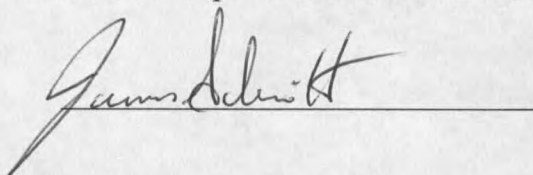
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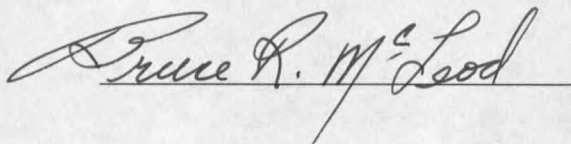
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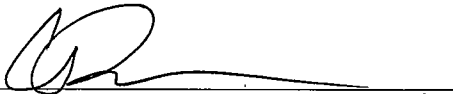
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ABSTRACT

The area of southwest Montana encompassing the Madison, Absaroka-Beartooth, and Bridger Ranges lies within the northern arm of the Yellowstone tectonic parabola (YTP). Detailed mapping along the Madison Range fault (MRF) and morpho-metric analyses were compared to the morphometry of two other faults in southwest Montana (Emmigrant/Deep Creek/Luccock Park fault - EDLF, and Bridger Range fault - BRF) in order to determine long-term tectonic activity rates and their correlation to the YTP. Morpho-metric analyses, including M_s , V_f , mountain-front morphology, and the RTAC, quantify long-term tectonic activity rates of > 0.1 mm/yr for all three faults. However, the basal facet reconnaissance technique yields late Pleistocene slip rates for the BRF of 0.01 mm/yr. The basal facet analyses provide a more accurate late Pleistocene slip rate for the BRF. This possibly suggests that the BRF experienced pre-late Pleistocene slip rates an order of magnitude greater (> 0.1 mm/yr) than the late Pleistocene (< 0.01 mm/yr), which is not modeled by the YTP. Mid- to late-Pleistocene tectonic activity between these three faults has been greater southward towards the Yellowstone volcanic field. In addition to late Pleistocene heterogeneous spatial and temporal earthquake activity, pre-existing structural controls on Quaternary normal faulting suggest that a simplistic parabolic deformational model does not adequately explain the neotectonics between the faults under investigation and southwest Montana.

Along with the neotectonic implications, the calibration and application of morpho-metric analyses to the northern Basin and Range province, coupled with surficial geology and contemporary seismicity, derive a seismic hazard assessment for the MRF, EDLF, and BRF. The MRF and EDLF have late-Pleistocene to Holocene slip rates ranging from 0.1-0.5 mm/yr. Based on basal facet analyses, the BRF has late-Pleistocene to Holocene slip rates of 0.01 mm/yr, but morpho-metric analyses (i.e., RTAC) quantify long-term average Quaternary slip rates of > 0.1 mm/yr. All three faults possibly have long-term tectonic activity rates > 0.1 mm/yr, with bifurcations in slip rates over periods > 100 ky. Maximum earthquakes for the three faults range from M_w 6.5-7.2, in accordance with historic seismicity of the Basin and Range province.

CHAPTER 1

INTRODUCTION

The study of neotectonics and paleoseismology in the northern Basin and Range province (Figure 1) has attracted the attention of researchers in the past due to the large-magnitude seismic history of the region (e.g., 1959 Hebgen Lake, MT, $M_s 7.5$, and 1983 Borah Peak, ID, $M_s 7.3$). Several models have been developed to explain deformational patterns and causative stress fields associated with the interaction of Basin and Range extension and the Yellowstone volcanic field (e.g., Anders et al., 1989; Pierce and Morgan, 1992; Smith and Braile, 1993). Fairly extensive paleoseismologic studies have determined slip rates along multiple faults proximal to the southern boundary of the Snake River Plain (SRP) (Piety et al., 1986; McCalpin, 1993; McCalpin and Warren, 1993), all of which show that westward from the Yellowstone volcanic field, greatest tectonic activity is located increasingly farther southward from the Snake River Plain (SRP). This southward displacement of tectonic activity is modeled to be reflected as a mirror image across the SRP to form the north arm of the Yellowstone tectonic parabola (YTP) (Figure 2) (Anders et al., 1989; Pierce and Morgan, 1992).

Extensive paleoseismologic studies in southwest Montana have involved the Hebgen fault system (Haller et al., 2000; Pierce et al., 2000; Schwartz et al., 2000; Van der Woerd et al., 2000) and the Red Rock and Blacktail faults (Stickney et al., 1987); however, other neotectonic studies have predominantly involved the mapping of fault

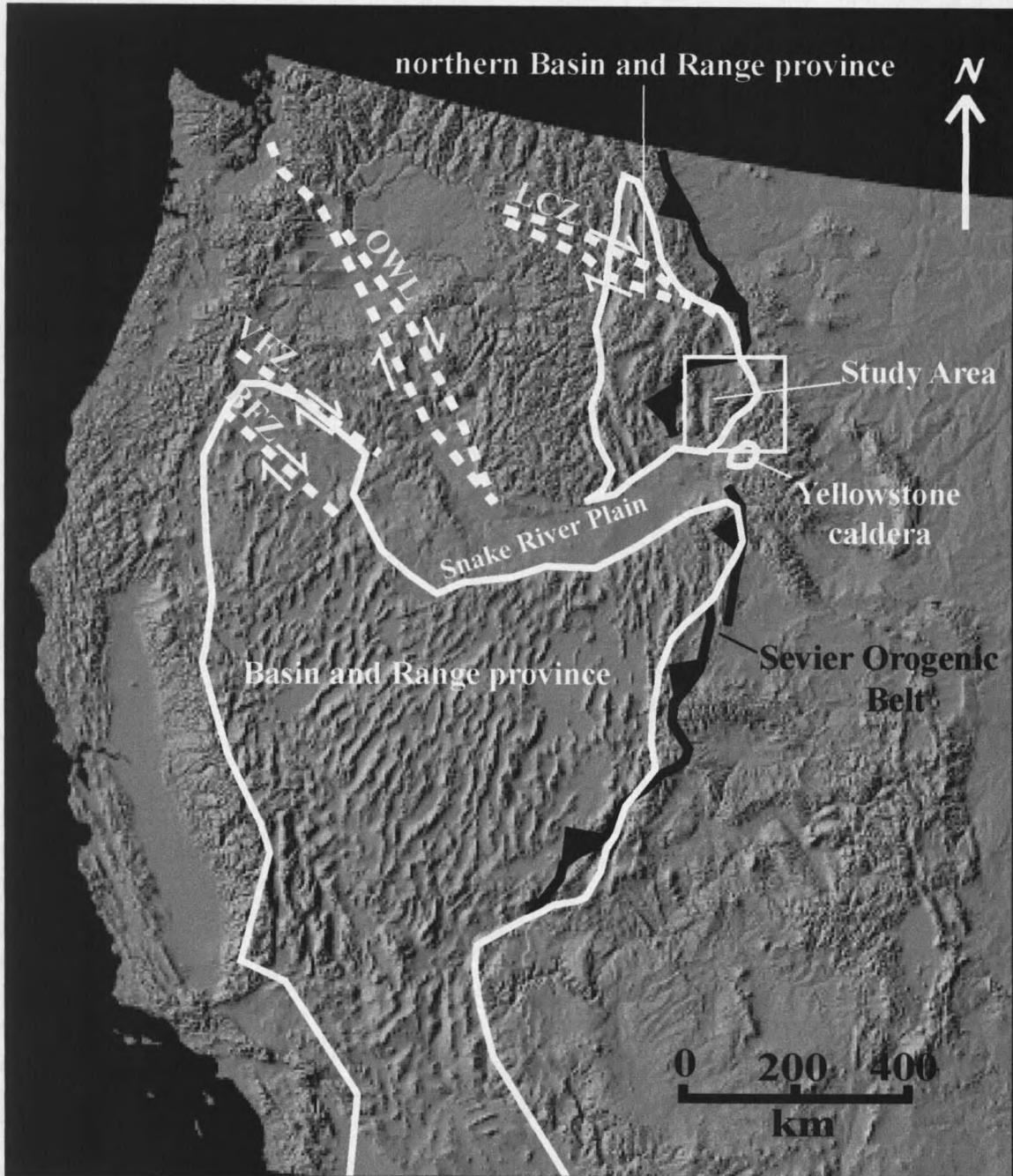


Figure 1. Western North America and location of study area. Regional zones of lateral shear in relation to the Basin and Range province include BFZ - Brothers fault zone, VFZ - Vale fault zone, OWL - Olympic-Wallowa lineament (Mann and Meyer, 1993), LCZ - Lewis and Clark zone (Hamilton and Meyers, 1966). Basin and Range province boundaries modified from dePolo (1994) and Lageson and Stickney (2001). Sevier Orogenic Belt boundary from Snoke (1993).

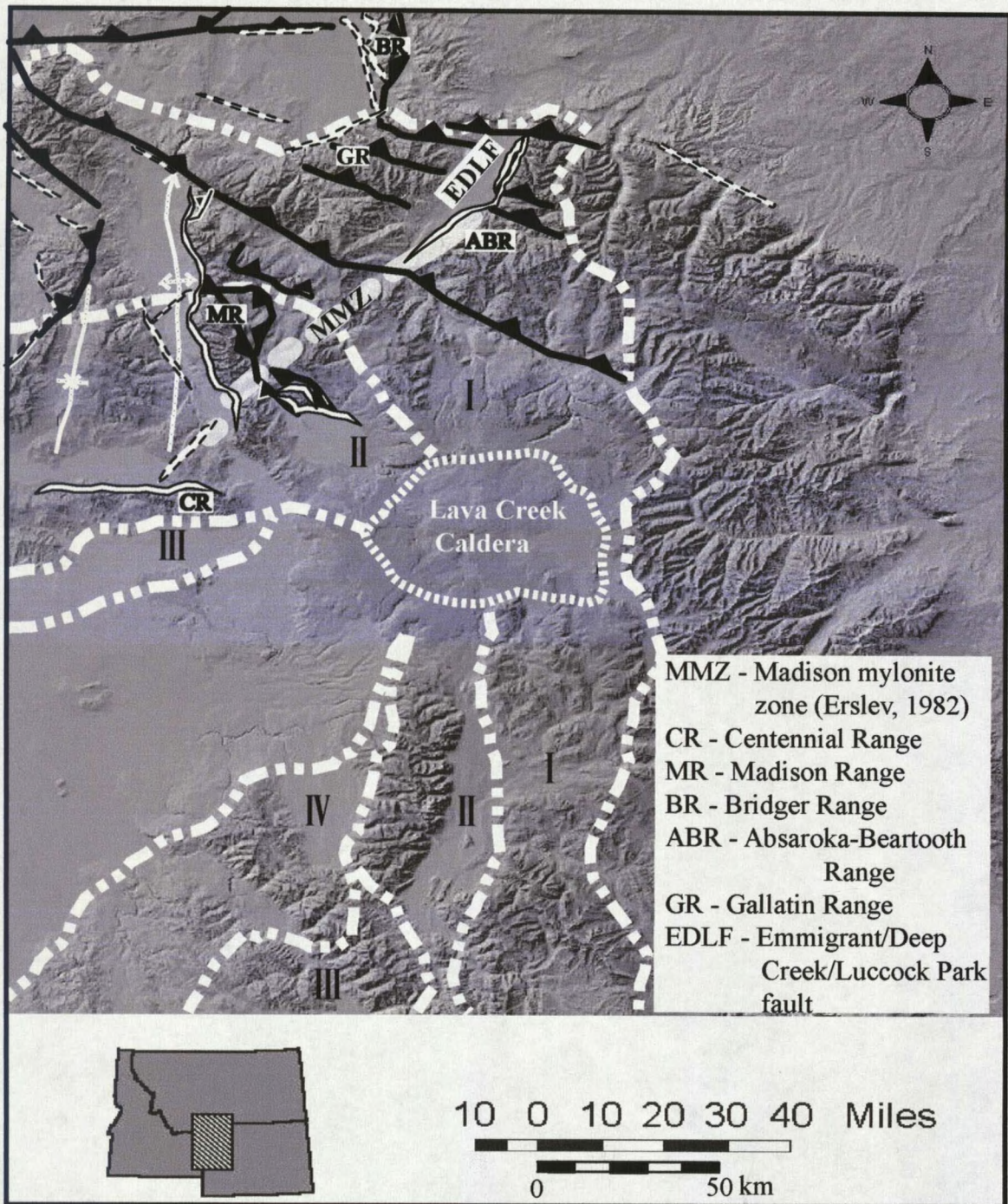


Figure 2. Physiography of the Yellowstone tectonic parabola (YTP). Dashed white lines, and roman numerals show tectonic belts (Pierce and Morgan, 1992). Laramide structures shown as black lines with sawteeth on hangingwall blocks, axes of large folds by white lines with anticlinal and synclinal symbols. Quaternary faults shown as white lines with black outlines, solid lines indicate faults with late Quaternary (< 130 ka) ruptures and dashed lines indicate faults with Quaternary (< 1.6 Ma) offset, modified from Stickney and others (2000).

scarps on Quaternary deposits and developing inferred long-term average slip rates from the amount of offset of late Pleistocene deposits (Madison Range fault - Mathieson, 1983; Lundstrom, 1986; Schneider, 1985; Emmigrant/Deep Creek/Luccock Park fault - Personius, 1982). Through morpho-metric analyses, this study attempts to quantify slip rates within the north arm of the YTP, southwest Montana, in order to determine long-term (> 130 ka) tectonic activity rates and aid in the understanding of northern Basin and Range province deformational rates.

Tectonic geomorphic analyses have been extensively used in the Great Basin to determine slip rates and resulting seismic hazards (Bull, 1984, 1987b; dePolo, 1998). These tectono-morphic reconnaissance techniques provide an inexpensive and relatively fast means to estimate slip rates, refine extensional models, and hopefully better constrain seismic hazards for the region. By calibrating and utilizing the geomorphic indices developed in the Basin and Range province of Nevada and Utah, this study documents fault activity rates along two major range-front faults within the YTP and one late Cenozoic fault just outside the YTP in southwest Montana (Madison Range fault - MRF, Emmigrant/Deep Creek/Luccock Park fault - EDLF, and Bridger Range fault - BRF). This paper treats in detail the MRF (Figure 2 and Plates 1, 2, and 3).

This research addresses the following questions: 1) Can the tectono-morphic methods, used to quantify Quaternary slip rates on large range-bounding normal faults in the Great Basin, be applied to the northern Basin and Range province north of the Snake River Plain, where climate and the direct influence of the Yellowstone hotspot are variables? 2) Can slip rates established by tectono-morphic analyses, in addition to

historical/contemporary seismicity and the paleoseismologic record, be used to identify areas of greater seismic potential? and 3) Do Quaternary slip rates measured for the MRF, EDL, and BF agree with previously published neotectonic models for the tectonic parabola that flanks the Yellowstone hotspot track?

CHAPTER 2

DISCIPLINARY CONTEXT

Normal fault slip rates and less precise relative rates of offset along active range fronts can be reasonably approximated through the study of tectonic geomorphology, or tectono-morphology (Bull, 1984). By solving for the variables that create landforms along fault-generated mountain fronts, models have been developed to describe long-term tectonic histories (Mayer, 1986). However, tectono-morphic studies have only been performed within a few regions, such as the central and southern Basin and Range province of North America. The practicality of these tectono-morphic models depends on the calibration and usage from region to region.

Perceiving the morphology of mountain fronts as a temporal indicator of fault activity, and therefore a direct tool for evaluating range-front evolution rates and seismic hazards, began as early as G.K. Gilbert's studies of the Basin and Range province (Gilbert, 1883, 1909). W.M. Davis followed with geometric models for the evolution of range fronts and faceted spurs (Davis, 1903, 1913). These first ideas were primarily observational and highly qualitative. A more quantitative relationship between fault offset, basin spacing, and facet development followed (Wallace, 1978). This classic study showed the relationship between fault-generated landforms and erosional processes, and the time periods that these landforms could persist. Further investigations led to a greater understanding of measured time intervals and coincident landforms, such as relationships between differing-aged alluvial fans along faults of differing tectonic

activity rates and basal facet heights along range fronts (e.g., Keller, 1986; Bull, 1987b; Gardner et al., 1987; dePolo, 1998). Through these neotectonic models, one can potentially estimate the seismogenic potential for regions with a particular landform development (Michetti and Hancock, 1997).

Over the past two decades, advances have been made in understanding the morphologic evolution of mountain fronts. A catalog of Quaternary geochronologic "tools" that can be utilized to determine age of events and inferred offset rates was compiled by Pierce (1986). Numerical dating techniques are generally reliable (e.g., ^{14}C), but samples are often difficult to obtain and expensive to analyze. However, calibrated-age and relative-age methods are inexpensive and broadly applicable, if properly calibrated to a region. Understanding the relationships between stream-channel downcutting, piedmont aggradation, piedmont degradation, and relative slip rates along a range front have led to the development of a *Relative Tectonic Activity Class Designation Chart* (RTAC) (Bull, 1987b; McCalpin, 1996) (Table 1).

The *Relative Tectonic Activity Classes* have been determined from a variety of morpho-metric analyses. *Valley-floor width to valley height ratios* (as discussed in the Methods chapter) yield values that quantify stream channel vertical incision versus lateral erosion within the footwall of normal-faulted range fronts (Keller and Pinter, 1996). Tectonic activity rates directly affect these values along normal-faulted range fronts. Higher tectonic rates create narrower, deeper canyons orthogonal to the range front as footwall erosion is accelerated by the increased gradient caused by dip-slip displacement along the range front. Conversely, slower tectonic rates allow for more lateral erosion to

Table 1. Relative Tectonic Activity Class Designation Chart. Morpho-metric indices of Ms - mountain front sinuosity and Vf - valley-floor width to valley height are used along with geomorphic characteristics in order to determine an inferred slip rate to produce the index values and accompanying landforms. McCalpin (1996) states that these are only for semi-arid climates and that subhumid to humid regions with the same indice values and landforms may have inferred slip rates several times larger than the ones listed above for each class. Modified from Bull and McFadden (1977), Bull (1984, 1987), and McCalpin (1996). Inferred slip rates can span the entire order of magnitude and should be listed for reconnaissance purposes only.

Classes of Relative Tectonic Activity	Piedmont Landforms	Footwall Landforms	Range-front Sinuosity	Valley Floor-Width to Valley Height Ratio	Inferred Slip Rate (m/ka)
1-Maximal	Unentrenched alluvial fan	V-shaped valley in bedrock, U-shaped valley in alluvium and/or soft bedrock			1.0-5.0
2-Rapid	Entrenched alluvial fan	V-shaped valley	1.1-1.3	0.06-0.53 (mean 0.15)	0.5
3-Slow	Entrenched alluvial fan	U-shaped valley	1.6-2.3	0.2-3.5 (mean 1.5)	0.05
4-Minimal	Entrenched alluvial fan	Embayed mountain front	> 2.5	0.4-3.8 (mean 1.7-2.5)	0.005
5-Inactive	Dissected pediment	Pediment embayment	2.6-4.0	0.9-39.4 (mean 7.4)	< 0.005

occur within the footwall, resulting in higher ratio values. Bedrock lithology and local structural heterogeneity must be considered when performing these analyses.

Mountain front sinuosity (as discussed in the Methods chapter) is the sinuous length of the mountain-piedmont junction divided by the straight-line length (Kellar and Pinter, 1996). This measurement reflects the balance between fault offset and stream embayment of mountain fronts (see Methods chapter). It is critical to determine the exact fault location (e.g., fault scarp) in order to accurately measure the embayment of the mountain front from the fault. Another geomorphic indicator of offset rate includes the

analysis of *alluvial fans* located at the mountain-piedmont junction (Keller, 1986).

Relationships between older and younger fans and the degree of stream entrenchment relate directly to fault offset rates. Higher tectonic activity rates along normal-faulted range fronts produce greater accommodation space proximal to the range front, thus allowing deposits to aggrade. This results in the stacking of deposits of differing age. Low tectonic activity commonly results in entrenchment of older deposits proximal to the range front and progradation of younger deposits further out into the valley.

These morpho-metric analyses provide a means to compare inferred rates of fault offset from range to range and between tectonic provinces. Bull (1987b) comments that the offset rates developed from the Relative Tectonic Activity Class values are for semi-arid to arid regions, and that humid to sub-humid regions with similar values may have offset rates up to several times larger for each class. Southwest Montana lies in a semi-arid to arid climatic zone (25-66 cm/yr) (National Water and Climate Center-USDA, 2000), similar to a large portion of the Basin and Range province. However, the southern and central Basin and Range province have a somewhat drier climate (25-51 cm /yr) (National Water and Climate Center-USDA, 2000). Therefore, the use of tectono-morphic methods developed in semi-arid to arid regions south of the Snake River Plain will potentially yield minimum tectonic activity rates for southwest Montana.

Faceted-spur analyses (as discussed in the Methods chapter) provide another means to estimate rates of offset for reconnaissance purposes. Triangular facets represent the eroded fault plane along a range front and indicate that sufficient time to erode the facets between uplift events has not elapsed (Gilbert, 1883, p. 50; Mayer, 1986). The

height of the basal set of facets has been shown to reveal relative rates of offset, whereas the average dip of the facet surface is largely a function of bedrock lithology (McCalpin, 1992). The most significant process controlling facet height is tectonic activity or slip rate, with bedrock lithology and piedmont aggradation or dissection having minimal effects (Menges, 1988). An estimated slip rate reconnaissance technique has been developed for the Basin and Range province south of the Snake River Plain (dePolo, 1998). The calibration of this slip rate estimation technique (dePolo, 1998) to the Yellowstone tectonic parabola provides a useful and inexpensive tool to understand rates of extension and associated seismic hazards in the region.

Quantifying the paleoseismology and reconstructing the neotectonic history of a region are the basis for practical information for seismic hazard analyses (McCalpin, 1996). The quantification of earthquake size and distribution in space and time is a key outcome of these studies. These neotectonic and paleoseismic investigations provide the means to determine maximum potential earthquakes (MPE) for regions, to calibrate rates of continuing tectonic activity, and to understand the effects of recurrent earthquakes on a particular geomorphic setting (Michetti and Hancock, 1997). By determining fault geometry and rupture segmentation, mean recurrence intervals, displacement per event, and seismic moment/moment magnitude for paleoevents, a quantification of seismic hazards can be developed (Schwartz and Coppersmith, 1986). This is one goal of this project and all earthquake research in general.

CHAPTER 3

PREVIOUS WORK

Regional

South of the Snake River Plain, paleoseismologic studies have constrained temporal and spatial patterns of Quaternary faulting. Trenching on range-bounding normal faults along the south arm of the tectonic parabola (e.g., Grey's River, Star Valley, and Rock Creek faults of Wyoming) has quantified rates of offset (Jones and McCalpin, 1993; McCalpin and Warren, 1993; Warren and McCalpin, 1992; Piety et al., 1992; Anders et al., 1989). Southward and eastward-migration of tectonic activity indicates that temporally-clustered events probably initiated nearer the hotspot and moved outward (Anders et al., 1989; Pierce and Morgan, 1992; McCalpin, 1993). The migrational patterns of late Cenozoic faulting along the southern arm of the YTP are also present on the northern arm with inferred temporal and spatial patterns of a parabolic, quasi-symmetrical YTP (Pierce and Morgan, 1992).

Most of the previous work entails descriptions of fault scarps and relative age of offset deposits, with slip rates on three faults constrained by paleoseismologic studies (Stickney et al., 1987; Bartholomew et al., 1999; Schwartz et al., 2000). Pardee (1950) gives a comprehensive description of Late Cenozoic faulting in western Montana based upon structural relationships, such as tilting of basin deposits containing volcanic ash beds and the presence of fault scarps. Descriptions of the geomorphic expression and relative-timing of most recent movement for the MRF, EDLF, and BRF are mentioned in

his study. Witkind (1975) compiled the first topographic-based Quaternary fault map for Montana. Ruppel (1993) gives an overview of the sequence of structural events from Laramide contractional tectonics to the present extensional regime, focusing on the influence of pre-existing structures on Quaternary deformational patterns in southwest Montana.

The northern arm of the YTP in southwest Montana consists of five main range-bounding normal fault zones, the Emmigrant/Deep Creek/ Luccock Park faults (EDLF), the Hebgen/Red Canyon faults, the Madison Range fault (MRF), the Centennial fault, and the Red Rock Valley fault (Pierce and Morgan, 1992). Pierce and Morgan (1992) interpreted the EDLF and the northern half of the MRF to be in *belt I, lesser and reactivated Holocene faults* and *lesser late Pleistocene faults* that are in Miocene basins and reactivated Miocene range-front faults. *Belt I* is characterized by offset younger than 15 ka, but with the absence of a high, steep, active, faceted range fronts. This belt contains faults that are considered to be early in their present cycle of seismic activity (Pierce and Morgan, 1992). The Hebgen/Red Canyon faults, southern half of the MRF, Centennial, and Red Rock faults are interpreted to be within *belt II* of the YTP, *major Holocene faults*. These faults are characterized by displacement since 15 ka, with high, steep, active facets and >700 m of relief (Pierce and Morgan, 1992), and are considered to have maintained high rates of activity over the last few million years. The Bridger fault (BRF) lies just outside the YTP.

Paleoseismologic studies have shown that slip rates vary an order of magnitude 0.01 to 0.1 mm/yr within the vicinity of the MRF, EDLF, and BRF. Stickney and others

(1987) excavated trenches across the Blacktail and Red Rock faults, with slip rates of 0.054 m/ka over the last 120,000 years and 0.11 m/ka for the last 23,000 years, respectively (Bartholomew et al., 1999). Trenches along the Hebgen fault system (Pierce et al., 2000, Haller et al., 2000, and Hecker, et al., 2000) show late Pleistocene to Holocene vertical slip rates of 0.3-0.5 mm/yr (Schwartz et al., 2000), with at least two events in the last 11-15 ka (Van der Woerd et al., 2000). Other neotectonic and tectonomorphic studies describe the scarp morphologies of the MRF (Gary, 1980; Matthieson, 1983; Schneider, 1985, 1994; Lundstrom, 1986) and EDLF (Personius, 1982).

Madison Range Fault

Shelden (1960) gives a general overview of the location and characteristics of the MRF. Work by Paul and Lyons (1960) and Sloane (1960) contributed to the identification of different-aged terraces and benches within the Madison Valley and showed that older mid-Pleistocene deposits are tilted into the range front. Weinheimer (1982) mapped and discussed intra-basin faults between Wolf and Curlew Creeks demonstrating the complexity of the MRF. For the southern MRF, the most recent offset is estimated to be 500-6,000 yr B.P. based on offset relative-age deposits and soil geomorphology (Schneider, 1985), and less than 4,000 yr B.P. by diffusion-equation modeling of fault scarps (Lundstrom, 1986). Mathieson (1983) gives a post-Pinedale slip rate of 0.4 to 0.7 m/kyr. This section of the southernmost Madison Range fault had co-seismic slip during the 1959 Hebgen Lake earthquake. Structure contours on top of the Huckleberry Ridge Tuff (2.0 Ma) (HRT) within the southern Madison Valley show more

than 1100 m of relief with at least 700 m of relief related to Quaternary faulting (O'Neill et al., 1996). This results in a minimum long-term average Quaternary slip rate of 0.35 m/ka, a slip rate in accordance with those determined from previous investigations. However, there is uncertainty in the relief on the pre-existing topography that the HRT mantled.

Emmigrant-Deep Creek-Luccock Park Fault

The EDLF was first described by Gail (1937), who recognized the fault scarp at the base of Dome Mountain at the southern end of the fault. Horberg (1940) calculated at least 5000 ft (> 1500 m) of throw on the fault with intermittent activity since the Miocene. Personius (1982) profiled and interpreted the fault to have an age of most recent offset of 10,000-12,000 yr B.P. with a late Quaternary slip rate of approximately 0.25 mm/yr. Pierce and Morgan (1992) identified two pulses of tectonic activity expressed by tilted Miocene volcanic and sedimentary rocks. They found a late Quaternary slip rate of 0.25 mm/yr for the last 0.5 Ma, preceded by fault activity quiescence between 0.5-8.4 Ma, and preceded by activity tilting Miocene sediments 8-10° from 8.4 Ma to at least 15 Ma. New dating of nearly horizontal basalts of about 2.0 Ma rather than 5-9 Ma removes the argument for the 0.5-8.4 Ma interval of quiescence (Kenneth L. Pierce, written communication, 2002).

Bridger Range Fault

The Bridger Range received early attention due to its precipitous western range front, stream morphology, and presence of faceted spurs. Fix (1941) describes offset Recent and Pleistocene deposits at the mouth of Bostwick Canyon. A scarp 1.5 m high was described and interpreted to have been formed within the last few hundred years, based on his belief that seasonal flood waters would rapidly erode the scarp across drainages at the mountain-piedmont junction (Fix, 1941). However, this scarp has not been recognized by subsequent workers. Pardee (1950) suggests that the presence of sharp faceted spurs between Cottonwood and Ross Creeks, along with a general interpretation of the Bridger Range summits as dislocated parts of the Tertiary peneplain, indicates that most of the displacement is Pleistocene and younger (i.e., ≥ 3000 ft of Quaternary displacement along the Bridger Range fault). Pardee (1950) also described a double fault scarp near the fish hatchery at the southern end of the fault. Investigations by the author found a subdued escarpment several meters high, but heavy anthropogenic alterations preclude accurate measurements of displaced surfaces.

Seismicity

Advancements in the seismic network in Montana over the past twenty years have greatly expanded our understanding of contemporary seismicity and its relationship to regional tectonics (Stickney and Bartholomew, 1987; Stickney et al., 2000; Lageson and Stickney, 2001). Stickney and Bartholomew (1987) suggest an overall T-axis orientation of S45°W for late Quaternary faulting in southwest Montana, with localized effects of the

Yellowstone caldera resulting in a T-axis orientation of S4°W. The thermal influence of the hotspot can restrict earthquake foci to much shallower depths, 4 to 6 km, than in the more typical Basin and Range province (Smith, 1988), resulting in the high frequency/low magnitude seismicity of the Hebgen Lake and West Yellowstone regions (Stickney, 1995). Doser (1985) and Barrientos and others (1987) used historic seismicity to determine focal mechanisms to ≈ 15 km depth in order to constrain fault plane dips ($\approx 55^\circ$) and characteristics of faults associated with the 1959 Hebgen Lake, MT and the 1984 Borah Peak, ID earthquakes.

CHAPTER 4

REGIONAL TECTONIC ACTIVITY RECONNAISSANCE TECHNIQUE UTILIZING
BASAL FACET HEIGHTS AND ITS CALIBRATION TO THE NORTHEASTERN
BASIN AND RANGE PROVINCE

Basal, triangular faceted spurs at the mountain-piedmont junction are the result of co-seismic offset along a range front (Davis, 1903). Once bedrock is exposed, erosional processes begin to dissect the bedrock, which counter-balances the effect of continuous offset along a range front (i.e., basal facets are in a steady-state balance between erosion and fault-generated offset, where offset rates equalize or exceed erosion rates) (Hamblin, 1976). dePolo (1998) used basal faceted spurs to distinguish range-bounding faults of differing slip rates for the Basin and Range province in Nevada. He found that dip-slip faults of differing vertical slip rates have dramatically different geomorphic expressions, of which the basal facet height is a factor. By mapping basal facets and acknowledging the presence or absence of alluvial/colluvial fault scarps with previously established methods (e.g., Hamblin, 1976; Anderson, 1977; McCalpin, 1982, 1992) along faults of known slip rates, the maximum basal facet height can be correlated with slip rates of known orders of magnitude (dePolo, 1998).

dePolo (1998) equates the change in height of basal facets over time (dh/dt) to the slip rate(s), minus the erosion rate (w), minus the burial rate (b) at the range front:

$$dh/dt = s - w - b .$$

The erosion rate (w) is equated to a local erosion constant (E) multiplied by the height of basal facets (h) :

$$w = E h .$$

The local erosion constant is similar to the constant of diffusivity for semi-arid to arid climatic regimes (Hanks et al., 1984). dePolo (1998) states that the complete system modifying the height of basal facets is much more complex than what is described by the above equation; however, the above equation incorporates enough of the physical processes to show that a steady-state solution might exist in the real physical world.

Thus, the basal facet height as a function of time is defined as (dePolo, 1998):

$$h(t) = [(s-b) / E] + A_0 e^{-Et}$$

This relationship shows that basal facet height increases as slip rates increase.

In order for slip rates to be compared to maximum basal facet height, three considerations must be evaluated: 1) Does a fault have scarps in alluvium/colluvium? 2) Does a fault have active or relict facets (i.e., sharp, well-defined vs. incised and deeply rilled facets)? 3) What is the height, in meters, of the largest basal facet? From these observations, dePolo (1998) assigned orders of magnitude to estimated slip rates based on the following range-front conditions:

Type 3 - Faults without alluvial scarps and active facets that have a slip rate of 0.001 m/kyr, with a range of uncertainty from 0.0005 to 0.01 m/ka. This slip rate magnitude has the potential to produce an earthquake with a ground rupture of 1-2 m every 1-2 million years. The slip rate is slow compared to erosion rates. Fault scarps would only be present 10% of the time.

Type 2 - Faults with alluvial/colluvial scarps and relict facets that have a slip rate 0.01 m/kyr. This slip rate has the potential to produce 1 m of offset every

100,000 yrs. Hanks and others (1984) showed that scarps become indistinct after 100,000 yrs, so faults in this category could have scarps the majority of the time. However, slip rates are too slow to maintain active facets along the range front. Any facets that exist are relict from earlier periods of higher rates of fault activity, and are primarily preserved in rocks with relatively higher resistance to erosion. Characteristics of **Type 2** faults are highly eroded facets and range fronts, retreating range fronts, and small remnants of offset alluvium or pediments between the range front and fault, but with the presence alluvial/colluvial fault scarps.

Type 1 - Faults with alluvial/colluvial scarps and active facets (i.e., minimal dissection and relatively smooth, sharp facets) have an estimated slip rate coincident with the maximum basal facet. At a slip rate of 0.01 m/kyr, 3 million years is needed to produce facets 30 m high. With the inclusion of any erosion rate, active facets are unlikely to develop at this order of magnitude of slip rate. Thus, 0.1 m/kyr is the lower bounding order of magnitude for slip rates to produce faults with active basal facets and the presence of range-front fault scarps.

Following the methods of dePolo (1998) for **Type 1** faults, basal faceted spurs must be mappable on 1:24,000-scale topographic maps. A minimum basal facet height of 30 m is used. The base of the facet is mapped along the fault scarp and crests are located at the top of the basal facet and the base of the spur ridge. It is best to align the topographic map with the range-front fault dipping towards the viewer and the range

front trending orthogonal to the viewer. After all basal facets along a range front are mapped, the largest height, measured in meters vertically from the elevation of the main fault trace to the facet crest, is used as the scaling parameter. dePolo (1998) derived a linear regression from known slip rates versus maximum basal facets in the Great Basin:

$$\text{Log } S = 0.00267 H - 0.963,$$

where S is slip rate and H is the height of the maximum basal facet for Type 1 faults.

The calibration of this technique to the Yellowstone tectonic parabola was performed by applying dePolo's (1998) methodology to quantify relative slip rates along faults with known slip rates from the Swan-Grand-Star Valley fault in southeastern Idaho and western Wyoming (Piety et al., 1986). Once the utility of this technique was determined for this fault system, it was then applied to faults north of the Snake River Plain where climatic regimes are largely comparable. The fault scarps located near the bases of both of the Snake River and Salt River Ranges lie in semi-arid to arid regions with < 50 cm/yr precipitation (National Water and Climate Center – USDA, 2000), making the application of this morpho-metric analysis comparable between portions of the Great Basin and similar dry climatic regions in the northern Basin and Range province. Climatic data for southwest Montana for the period 1931 to 1961 (Department of Commerce, 1980) and from 1961 to 1990 (National Water and Climate Center - USDA, 2000) show an average of 25-66 cm/yr for valleys and range fronts. These averages are comparable with average annual precipitation of ranges in the central and southern Basin and Range province, 10-50 cm depending on locality, and Idaho, 10-70 cm for the Swan-Grand-Star Valley (National Water and Climate Center - USDA, 2000).

Schneider (1985) states that exogenic processes between these regions should be comparable if the materials are the same. Bull (1987b) notes that humid to sub-humid environments with the same quantified morpho-metric indices will have higher slip rates to maintain the same values as semi-arid to arid environments. Thus, indices (e.g., V_f , M_s , and *basal facet heights*) for regions with slightly greater amounts of annual precipitation quantified by techniques used in the Basin and Range province of Nevada (e.g., RTAC and *basal facet analyses*) yield minimum slip rate values for regions with slightly greater precipitation and erosion rates.

Geomorphic processes working to weather and erode facets are a function of climate and annual precipitation. Figure 3 shows that denudation rates (cm/yr) and sediment yield (tons/mi²/yr) for the southern and central Basin and Range province are comparable with that of the Star and Swan Valleys of Wyoming and Idaho as well as the valleys and range fronts in southwest Montana under investigation. Regions with less precipitation have less vegetation preventing erosion, yet low annual precipitation levels

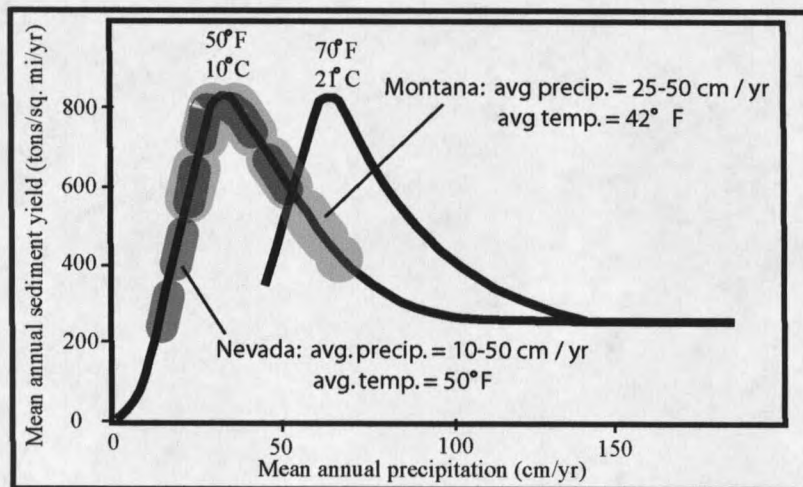


Figure 3. Precipitation versus mean annual sediment yield. Modified from (Chorley et al., 1984). For more arid environments, the sediment yield is constrained by amount of annual precipitation. As the annual precipitation increases to semi-arid and subhumid, the annual sediment yield is constrained by an increase in vegetation.

have restricted sediment yields (Chorley et al., 1984). As annual precipitation increases above a certain value, vegetation increases, which decreases the erosive potential of precipitation. The Star and Swan Valley and southwest Montana (in particular, the Bridger, Madison and Absaroka-Beartooth Range fronts) have slightly greater annual precipitation rates (≈ 10 -20 cm) than the main south-central Basin and Range province of Nevada (National Water and Climate Center – USDA, 2000). The decrease in annual precipitation and subsequent decrease in vegetation to the south produces similar sediment yields to the valleys of southwest Montana and eastern Idaho, where greater amounts of vegetation and annual precipitation occur (Chorley et al., 1984). Therefore, I propose that exogenic processes working to degrade and bury basal faceted spurs are comparable between these two regions.

The Star Valley has documented late Quaternary slip rates of 0.6 to 1.1 m/kyr (Piety, et al., 1992). The faceted spur estimation scheme yields a *type 1* fault with a slip rate of 0.53 m/kyr, using the maximum basal height of 268 m. This basal faceted spur is between Graveyard Canyon and Hale Creek, just south of Swift Creek in Star Valley on the Afton, WY 7.5' USGS topographic map (Figure 4). The Swan Valley segment has a documented slip rate of 0.019 m/kyr (Anders et al. 1989; Piety et al., 1992). Faceted spurs along this reach of the fault are moderately dissected and resemble relict facets, but fault scarps are present. Thus, the faceted-spur technique assigns this segment to be a *Type 2* fault with an estimated slip rate of 0.01 m/kyr.

Morpho-metric analyses, including *valley-floor width to valley height ratios* (V_f), *mountain front sinuosity* (M_s) and dissection of range front deposits, are factors in the

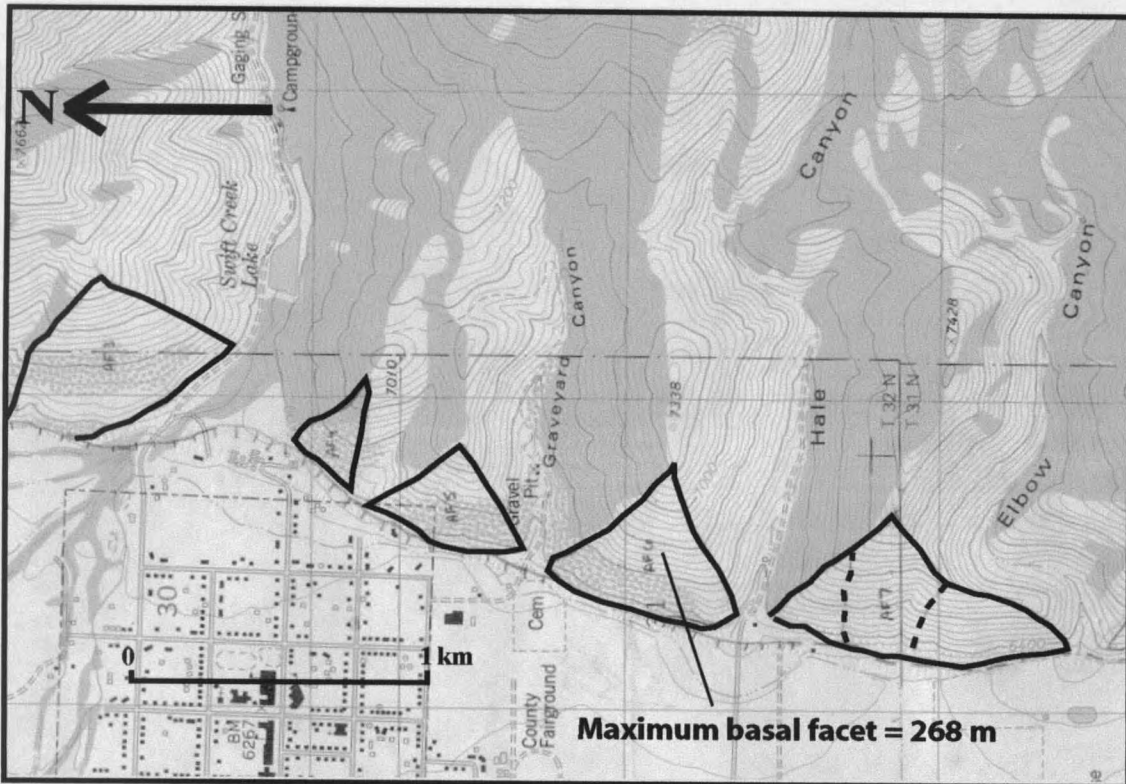


Figure 4. Maximum basal facet for the Star Valley segment of the Grand-Swan-Star Valley fault system, southeast Idaho and western Wyoming. Examples of other mapped basal faceted spurs are shown. Fault scarp shown by hachured lines at the base of the range front.

Relative Tectonic Activity Class Chart (RTAC) (McCalpin, 1996; Bull, 1987, 1984), yield inferred tectonic activity rates of 0.5 m/ka for both the Star Valley and the Swan Valley. Average V_f ratio values are 0.06 for the older and Holocene Star Valley segment and 0.21 for the Swan Valley segment (Piety et al., 1986), indicative of a tectonic activity rate > 0.1 m/ka for both segments. I hypothesize that the morpho-metric index similarities between the Star Valley and Swan Valley segments represent remnant morphology along the Swan Valley segment indicative of a previous cycle of greater activity rates documented by Anders and others (1989). This also leads to the conclusion

that these values (e.g., M_s and V_f) are affected on a time period greater than 10^5 years in semi-arid to arid climates. The faceted spur technique provides greater accuracy for the late Pleistocene slip rate along the Star and Swan Valleys than the RTAC. A comparison of documented slip rates gathered from trench data, offset basalts and ash-flow tuffs, the Relative Tectonic Activity Classification Chart, and the faceted-spur analyses show a correlation in order of magnitude for slip rates quantified by the faceted spur analysis and paleoseismology, but the Relative Tectonic Activity Chart only yields consistent slip rates for the Star Valley segment (Table 2).

Table 2. Comparison of slip rates determined through several analyses for the Star and Swan Valley faults, southeast Idaho and western Wyoming

<u>Slip Rates Quantified by Several Techniques for the Star-Swan Valley faults, Idaho</u>			
	<u>Paleoseismology</u> <u>(Pietry et al, 1986)</u>	<u>Faceted Spur</u> <u>technique</u>	<u>Relative Tectonic</u> <u>Activity Classification</u>
Swan Valley	0.019 m/ka	0.01 m/ka	0.5 m/ka
Star Valley	0.6-1.1 m/ka	0.53 m/ka	0.5 m/ka

CHAPTER 5

TECTONIC ACTIVITY AND SLIP RATES: HOW THEY RELATE

Slip rate is defined as the net tectonic displacement on a fault during a measurable period of time (Wallace, 1968; Schwartz and Coppersmith, 1986). For fault slip rates to be assigned to a specific fault, two criteria must be met: 1) known amount of fault displacement, and 2) time between successive ruptures (Machette, 1987). This poses a problem for reported slip rates determined from amount of offset of a datum that has experienced only one offset. Since short-term slip rates can vary greatly from long-term slip rates, average long-term slip rates (over time periods > 10 ka) must be distinguished from short-term slip rates (< 10 ka). For example, a datum of assigned age 17 ka that is offset 3.0 m with one offset event does not quantify a contemporary slip rate of 0.18 m/ka. Instead, this datum only yields a summation of the tectonic activity along the fault in the last 17 ka, with the uncertainty of no known slip rate per unit time during the previous 17 ka interval, nor number of coseismic ruptures. The recurrence interval for this particular fault could be 100 ka with one event preserved within the fault scarp, making the seismic hazard minimal for the next 80,000 years (M.N. Machette, personal communication, 2002). Conversely, if a 17 ka datum has experienced 5 offsets, then the slip rate and recurrence intervals are relatively well constrained. However, temporal clustering greatly complicates the assumption of relative slip constancy over time.

Within plate boundary fault systems, where slip rates along individual faults are generally higher than intraplate regions, calculated Holocene slip rates, primarily through

trench explorations, can yield pertinent information on stress concentration and strain partitioning over long time intervals (>100 ka). Within intraplate regions, such as the Basin and Range province, Holocene rupture along range fronts may not have occurred, but this does not necessarily equate to a lower seismic hazard (dePolo and Slemmons, 1998). Stress may be distributed across many faults and a greater volume of crust, making it very difficult to predict where and when the next ground rupture will occur. If the uncertainty is great in determining slip rates over short time intervals (< 10 ka) along intra-plate faults, then a longer record of tectonic activity along a specific fault should be utilized for determining seismic potential and hazards for specific faults. The grouping and migration of historic and paleoseismic faulting events within intra-plate regions clearly show that tectonic activity shifts between sub-provinces and belts over time intervals greater than 10^3 - 10^4 years (Wallace, 1987). Thus, the tectonic activity preserved in datum older than 10 ka (e.g., glacial moraines, alluvial fans, terraces) can aid in quantifying long-term average slip rates for fault segments and associated hazards, along with patterns of migration of tectonic activity.

Models of fault behavior show that displacement rates and recurrence intervals can vary along strike of a fault. Data gathered through paleoseismic studies leads to the development of models to characterize fault behavior over time, such as, characteristic earthquake model of Schwartz and Coppersmith (1984), and variable and uniform slip models of Berryman and Beanland (1991). These models predict varying slip rates and displacement per event along a fault system. If accumulated slip over short time intervals can vary, such as perfectly periodic, time-predictable, and slip-predictable models of

earthquake recurrence (Scholz, 1990), long-term displacement from an older datum may yield a better estimate of the seismic potential and relative tectonic activity for a specific fault than offset of a young datum. Therefore, these long-term average slip rates acquired from older datums and the RTAC provide key data for the assessment of fault activity and hazard to society.

In order to better quantify slip rates and the relationship with tectonic activity rates between sub-provinces, morpho-metric analyses of the range front and fault zone provide a useful comparison to trench data. If geomorphic rates are similar across a region over a specific time interval, then the tectono-morphic signature should vary along strike on segments of differing slip rates and long-term behavior. Furthermore, long-term averages of fault displacement rates help quantify average recurrence intervals for earthquakes along individual fault segments (C.M. dePolo, 2001, personal communication). From the average recurrence interval, a probability of an event to occur along a specific segment can be determined in order to assign seismic hazard to regions.

Wallace (1986) states that active faults should be defined as "those that may have displacement within a future period of concern to humans." If seismic hazard is in question, these crude slip rates can be helpful. These long-term (> 50 ka) slip rates may be significantly lower or higher than short-term (< 50 ka) rates, but nevertheless provide an estimated slip rate for the fault and provide reconnaissance data for further investigation. Uncertainty and error are due to the following: datum-age calculations, displacement or height of scarp, preservation of evidence of number of events preserved in large fault scarps, and time interval between events. Trench data is the most accurate

means of determining slip rates. Slip rates obtained by trenching allow correlation between geomorphic indices and calculated slip rates.

Lastly, these morpho-metric analyses provide a means to compare regional shifts in tectonic activity across broad intraplate regions. Since the parameters measured in the calculations are affected on time periods involving several seismic cycles, a comparison of faults having equivalent M_s and V_f ratios and different range-front expressions (e.g., presence of fault scarps and fresh, sharp facets vs. deeply rilled, "skeletal" facets and lack of fault scarps) shows where more recent Quaternary tectonic activity has been concentrated. The morpho-metric indice values assist in distinguishing faults that are active and faults that once were active, but have experienced more recent tectonic activity quiescence. Therefore, morpho-metric analyses used to quantify tectonic activity provide estimated minimum long-term slip rates. However, they provide another analysis to determine Quaternary migration and grouping patterns of surface faulting between sub-provinces and belts to compare with trench data for a more accurate seismic hazard assessment.

CHAPTER 6

METHODS

Slip/Tectonic Activity Rate

Inferred slip rates derived from tectonic activity rates for range-bounding normal faults along the Bridger Range, Absaroka-Beartooth Range (the Paradise Valley), and the Madison Range are quantified by the following tectono-morphologic methods: 1) valley-floor width to valley height ratios, 2) mountain-front sinuosity, and 3) degree of alluvial fan development and stream incision. *Valley-floor width to valley height ratios* (V_f) yield values that can be compared between ranges of differing uplift rates (Figure 5). Normal-faulted range fronts undergoing slower tectonic activity rates will have larger ratio values because lateral fluvial erosion is dominant over vertical incision of the footwall. Values are calculated by measuring the width of the valley floor that contains the active channels. A constant distance of between 0.5 to 1.5 km into the footwall is desired for the measurements. This distance usually occurs within a reach of the drainages that has not been heavily affected by glacial processes and is a good distance downstream from the catchment basin. Also, this reach of a drainage correlates with the area most recently affected by tectonically-induced footwall incision. The set distance into the footwall must be held constant for all measurements along the range. Therefore, it is necessary to analyze the entire drainage network of the range front in order to determine the set distance into the footwall where the measurements are made. Valleys that have been

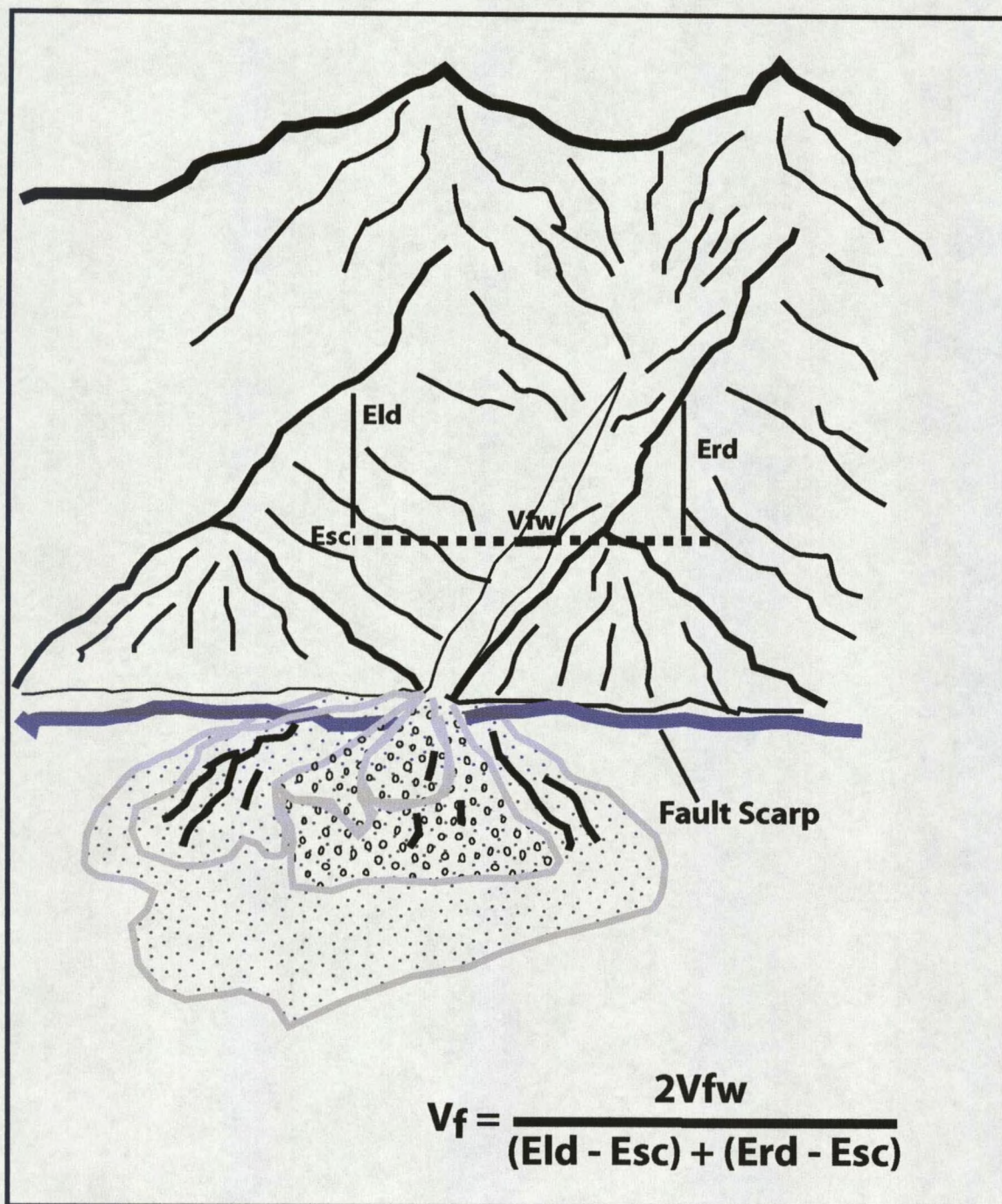


Figure 5. *Valley-Floor Width to Valley Height Ratio* (V_f) measurements and calculations. Eld = elevation of left divide, Erd = elevation of right divide, Esc = elevation of stream channel, Vfw = valley-floor width.

heavily glaciated require finding a location where fluvial incision is easily distinguished from glacial scour on a transect perpendicular to the drainage that the measurements are made. The elevations of the adjacent divides of the drainage are recorded, along with the elevations of the valley floor. These data are then applied to the equation:

$$V_f = 2V_{fw} / [(E_{ld} - E_{sc}) + (E_{rd} - E_{sc})]$$

where V_{fw} is the width of the valley floor, E_{ld} and E_{rd} are the respective elevations of the left and right divides, and E_{sc} is the elevation of the valley floor (Keller, 1986).

Mountain front sinuosity is a measure of the embayment of the mountain front from the trace of the bounding fault zone, and is the sinuous length of the mountain-piedmont junction divided by the straight-line length (Figure 6). The equation used is:

$$S_{mf} = L_{mf} / L_s$$

where L_{mf} is the length of mountain front along the mountain-piedmont junction and L_s is the straight-line length of the front (Keller, 1986). Changes in fault trends (salients and embayments) are accounted for when measuring the straight-line length of the fault.

Slower tectonic rates tend to produce more diffuse and highly embayed range-fronts.

Conversely, more rapid slip rates, with shorter recurrence intervals, will tend to keep the fault more proximal to the mountain-piedmont junction and bedrock/colluvium contact.

Alluvial fan development and stream morphology show a relationship between rates of uplift and erosion, thus requiring more thorough field observations. Spatial relationships between older and younger fans (Figure 7), as well as degree of stream entrenchment, can relate directly to offset rates (Keller and Pinter, 1996; McCalpin, 1996).

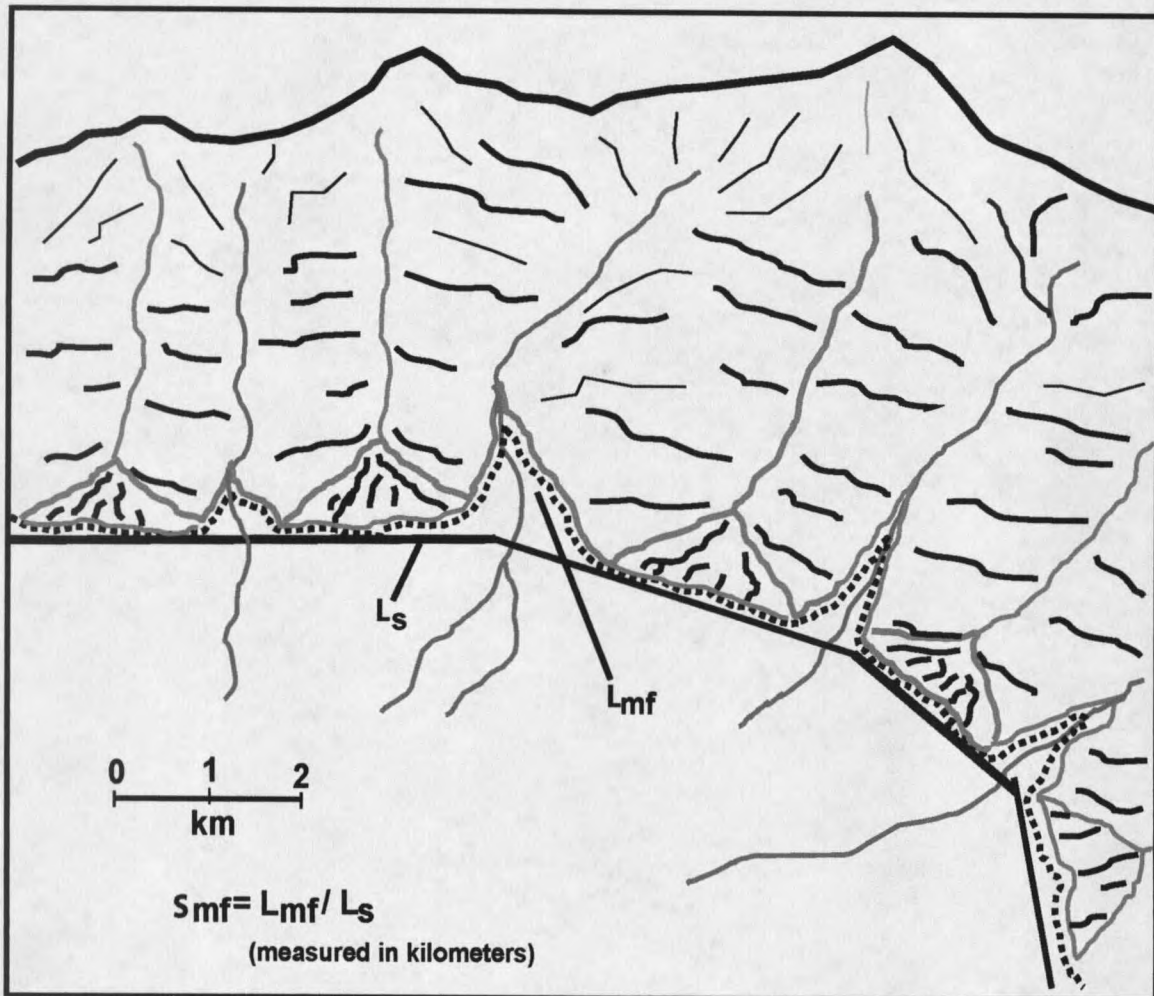


Figure 6. Mountain front sinuosity (S_{mf}) measurements and calculations. L_s = straight-line length of mountain front determined from fault trace and L_{mf} = length of the mountain front at the mountain-piedmont interface. Note that the trace of the fault is used as the straight line length of the mountain front. Some faults may have a more irregular, fault trace pattern and this must be incorporated into the measurement (Kellar, 1986) by adding multiple straight segments for the straight-line length.

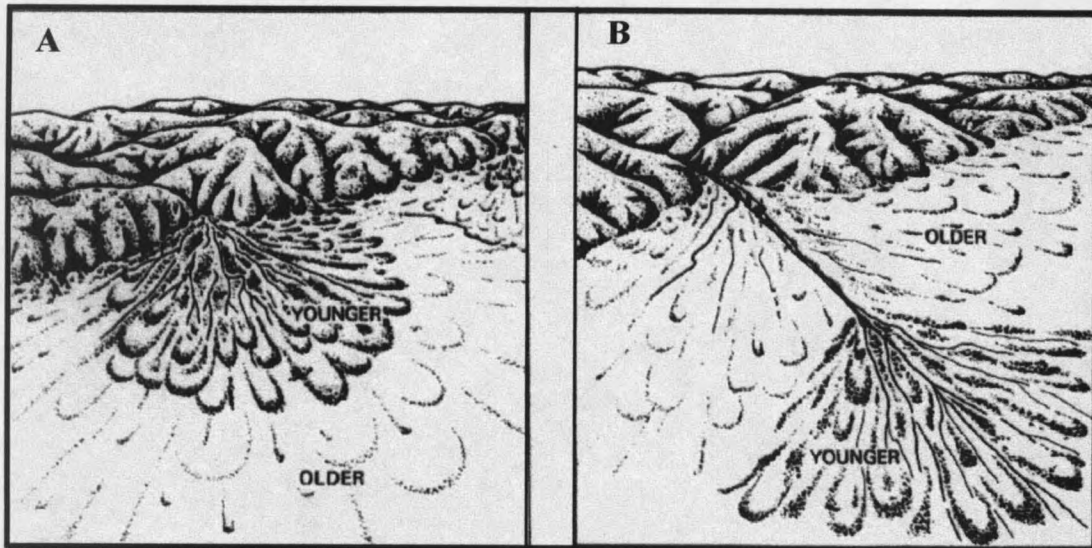


Figure 7. Relationships between tectonic activity and alluvial fan deposition. Modified from Bull (1977). A) With higher slip rates, accommodation space is more constantly being produced. Subsidence rates exceed erosion rates and deposition occurs near the range front. B) With slower slip rates, fan-head entrenchment occurs and deposition occurs further away from the mountain-valley junction.

Decreasing slip rates will lower rates of accommodation space production. This produces progradation of transverse fans, fan-head entrenchment, and basin-ward shift of axial rivers (Burbank and Anderson, 2001). These three analyses are assembled into the RTAC to determine an inferred slip rate for the range-bounding fault under investigation (Table 1) (Bull, 1984, 1987b; McCalpin, 1996).

Fault Scarp Analyses

Fault-scarp profiles were measured across the Madison Range fault by standard USGS methods utilizing an abney or digital level and stadia rod following the procedure

of Bucknam and Anderson (1979) and Machette (1982). The stadia rod is placed at meter increments on relatively smooth surfaces and the digital level is placed on the rod to measure the slope. Where vegetation (sagebrush) and surficial topography interfered with slope measurements, a tape measure was placed at the hip between two surveyors for distance measurement and the abney level was sighted to the same height from the viewer to the other surveyor. Measurements of scarp profiles were then entered into a USGS "scarp" Excel program by Barton Kiburz, provided by M.N. Machette, to determine total surface offset. Measurements of offsets from fault scarps, in addition to rupture length, were then applied to determine moment magnitude of individual paleo-events for seismic hazard analyses.

Basal Facet Height Reconnaissance Analyses

A reconnaissance technique utilized by the Nevada Bureau of Mines and Geology that correlates the height of basal facets at the mountain-piedmont junction to slip rates (dePolo, 1998) was calibrated using faults within the south arm of the YTP, where slip rates have been calculated by trenching and radiocarbon dating, as described earlier (Warren and McCalpin, 1992; Piety et al., 1992). Application of this basal facet reconnaissance tool to range fronts with documented slip rates will reveal if calibration needs to be modified to use this technique in southwest Montana along the Bridger, Absaroka-Beartooth, and Madison Ranges (see Chapter 4). Slip rates (based on mean, minimum, and maximum facet height for this study, but dePolo's (1998) technique was solely developed for maximum basal facet heights) calculated from faceted-spur analyses

are compared and contrasted with the quantified slip rates from the Relative Tectonic Activity Chart.

Field Mapping

The Madison Range fault and adjacent bedrock and surficial geology was mapped in the field at 1:12,000 scale. The use of aerial photography and previously published maps (Montagne, 1960; Paul and Lyons, 1960; Sheldon, 1960; Sonderegger et al., 1982; Weinheimer, 1982; Matthieson, 1983; Lundstrom, 1986; Bearzi, 1987; Ritter et al., 1990; Ritter et al., 1995; O'Neill, in press) aided the identification and mapping of Quaternary landforms that extend into the valley. The previously published maps of Kellogg and Williams (2000), Wilson and Elliott (1997), and Erslev (1981, 1982) assisted in mapping the range-front bedrock geology.

Relative-ages of deposits were mapped primarily by standard geomorphic techniques, including degree of original surface morphology preservation, degree of dissection, and cross-cutting relationships by incision and faulting (Bucknam and Anderson, 1979). Soil development and the presence of a thick loess unit distinguishes Bull Lake equivalent (≥ 130 ka) from Pinedale equivalent (15-20 ka) and younger deposits (Lundstrom, 1986; Bearzi, 1987; Ritter et al., 1990; Schneider, 1994; K.L. Pierce, 2002, oral communication). The measuring and correlation of weathering rinds on stones from differing deposits was attempted to determine relative ages using the methods of Coleman and Pierce (1981). Approximately 30 stones were gathered from the upper B horizon at each site. The stones are then cracked open and measured with a

7x pocket comparator with a 10 mm reticle with 0.1 mm increments. Stones with very thin rinds (0.01 to 0.1 mm) were assigned thicknesses based upon whether the rind obscured the texture of the rock; rinds were assigned 0.0 mm where the texture of the rock could be seen through the rind and 0.1 mm where the rind obscured the texture of the rock. However, footwall rocks and their weathering susceptibilities differ greatly along the Madison Range front making correlations from the north to the south impossible; therefore, this technique proved to be of little practical use.

CHAPTER 7

THE MADISON FAULT ZONE

The NNW-striking Madison Range fault is approximately 95 km in length, lying in both the northeast Basin and Range province and the Yellowstone tectonic parabola. Detailed mapping (plates 1-3) and fault scarp profiling (plates 1-3 and Appendix A) reveal greater paleoseismic activity towards the south, with at least three post-Pinedale paleoevents along the southern portion of the fault zone. For the Madison Canyon/Missouri Flats and the Taylor-Hilgard segments and the section between Mill and Bear Creeks, early- to mid-Holocene fans have vertical surface offsets that average between 2.0-3.0 m, which defines the characteristic single-event surface offset. Pinedale lateral moraines have vertical surface offsets as great as 12.0 m. As suggested by Personius (1982) for the EDLF, Late Pleistocene to Holocene multiple-event fault scarps show little evidence of beveling, suggesting short recurrence intervals and potential late Pleistocene/Holocene temporal clustering. Long-term average tectonic activity rates indicate slip rates ranging from 0.18-0.6 mm/yr. Based on a comparison of fault scarp height versus maximum slope angle of known regression lines developed from other paleoseismic investigations, the most recent event ranges from 1-5 ka.

The northern section of the fault zone is defined by multiple normal faults detaching the hanging wall of Laramide thrust faults within the Paleozoic and Mesozoic strata. This results in the partitioning of extension along multiple pre-existing structures and less displacement along individual normal faults. Structural controls on lateral

propagation of individual paleoevents involve the position of lateral ramps along pre-existing Laramide contractional faults. This results in greater displacement within the larger basement-cored structures along the southern sections south of Indian Creek where extension is accommodated by one inferred principal basement-involved normal fault.

Structural Setting

The Madison Range and Valley overlap the location of four distinct tectonic episodes and related structures: 1) an Early Proterozoic shear zone, the Madison mylonite zone (Erslev, 1982); 2) Laramide foreland province (Snoke, 1993); 3) the Basin and Range province and associated Early to Middle Miocene, northeast-trending grabens (Sears and Fritz, 1998), followed by late Cenozoic northwest-trending grabens (Ruppel 1993); and 4) the Yellowstone tectonic parabola (Pierce and Morgan, 1992). The Madison fault zone exposes structures related to all of these provinces culminating in the present extensional regime (Figure 8).

Beginning in Early Proterozoic, pre-Belt sedimentary rocks were deformed and metamorphosed to produce the northeast-striking, northwest-dipping Madison mylonite zone (MMZ) (Erslev, 1982). Seismic tomography data supports Erslev's (1993) idea that the MMZ is a major crustal shear zone (Dueker et al., 2001). North of the MMZ to Indian Creek, the Madison Range front exposes Archean granitic gneiss along with migmatite and mixed suites of quartz-feldspathic gneiss. Following the Early Proterozoic continental suturing event (O'Neill, 1993), Middle Proterozoic northwest-

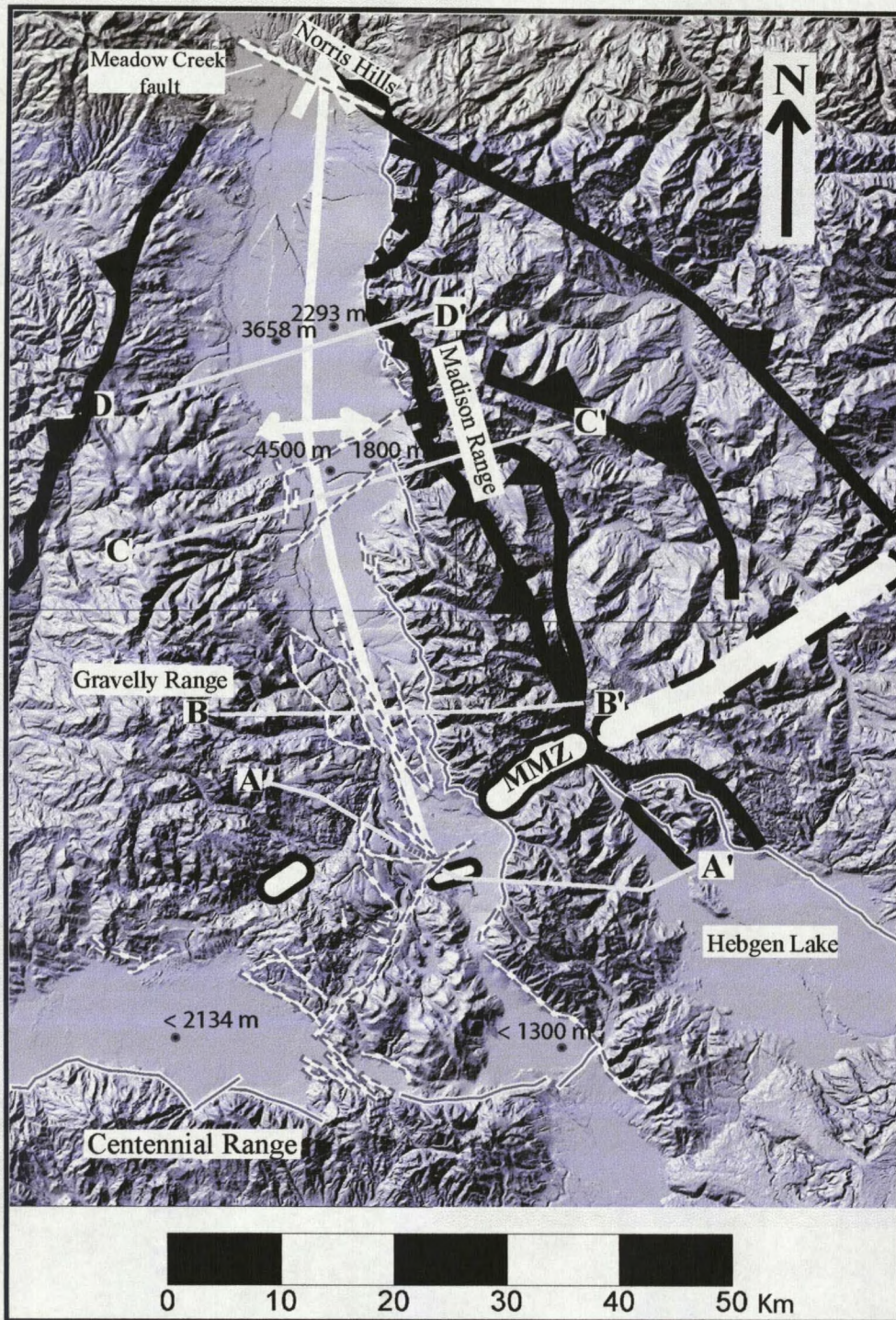


Figure 8. Madison Range structural setting. Gray lines show recognized Quaternary faults modified from Stickney and others (2000). Black lines show Laramide structures with white line showing fold axis of Madison-Gravelly arch. Early Proterozoic shear zone (Madison mylonite zone - MMZ) shown by white zone with black outline from Erslev (1993). Cross sections described in text shown in white. Basin depths from Schofield (1981), Peterson and Witkind (1975), and Kellogg and others (1995).

trending extensional faults developed within the Archean rocks, resulting in the opening of the Belt Basin (Schmidt and Garihan, 1986).

During the late Cretaceous and early Tertiary, contraction within the Laramide foreland province produced the following basement-involved structures: 1) NNW-striking, WSW-dipping thrusts subtending the Madison-Gravelly Arch, a large Archean basement-cored, fault-bend fold (Scholten, 1967; Kellogg et al., 1995), and 2) the NW-striking, NE-dipping Spanish Peaks fault (Garihan et al. 1982) (Figure 8). The eastern margin of the Madison-Gravelly Arch is defined by the Hilgard thrust system, which juxtaposes Archean rocks of the southern Madison Range with Cretaceous rocks to the east within the Madison synclinorium (Scholten, 1967; Tysdale, 1986; Ruppel, 1993). Timing of movement along the Hilgard thrust system is constrained by thrusts stitched by dacitic sills dated at 68-69 Ma (Tysdale et al., 1986). Unfaulted Eocene volcanic rocks overlap the southeastern end of the Spanish Peaks fault, showing no displacement along the fault since at least the Eocene (Kellogg et al., 1995). From north of Bear Creek to Cedar Creek, the Hilgard thrust system consists of numerous imbricate thrust faults within the Phanerozoic section, striking subparallel to the Madison Range front fault (Kellogg et al., 2000). Dip-separation decreases to the north along the Hilgard thrust system, with a maximum dip separation of 5000 m (Kellogg et al., 1995).

Initial extension of the Madison-Gravelly Arch began in the middle Eocene (Hanneman and Wideman, 1991). During the middle Miocene, rapid extension began to form the present topography of the Madison Range and valley (Fields et al., 1985). As much as 4500 m of Cenozoic basin fill exists south of Ennis, MT (Schofield, 1981).

These basin fill deposits are divided into the Oligocene Renova Formation, Neogene Six Mile Creek Formation, and approximately 300 to 500 m of Quaternary deposits (Rasmussen and Fields, 1983). Within the field area, drill data from Texaco and Helis Estate show the Renova Formation overlying Archean basement rock, which is the same relationship seen along the Madison River north of the Norris Hills (Figure 8). These basin deposits thicken to the south, with the thickest section beginning near the confluence of Indian Creek and continuing south for approximately 10 km to Deadman Creek and the Missouri Flats section of the fault zone (Schofield, 1981) (Figure 9). The Madison Valley is considered to be a half-graben bound by faults on the eastern side along the Madison Range front. However, the southern portion of the valley is characterized by a full graben with apparent intra-basin normal faults. One focal mechanism is consistent with a steep ($> 50^\circ$), east-dipping normal fault along the western side of the Madison Valley (Stickney, 1995) indicating that the Madison Valley is a graben bound by steep inward-dipping faults.

The northern end of the Madison Valley is bound by the Meadow Creek fault (Figure 8). This normal fault strikes northwest parallel to the Spanish Peaks fault, with down-to-the-southwest displacement. Schmidt and Garihan (1986) state that Tertiary extension along the Meadow Creek fault was controlled by the position of the Spanish Peaks fault. Most movement along the Meadow Creek fault had ceased by mid-Miocene (Schmidt and Garihan, 1986), with only 40 m of post-Miocene displacement (Young, 1985).

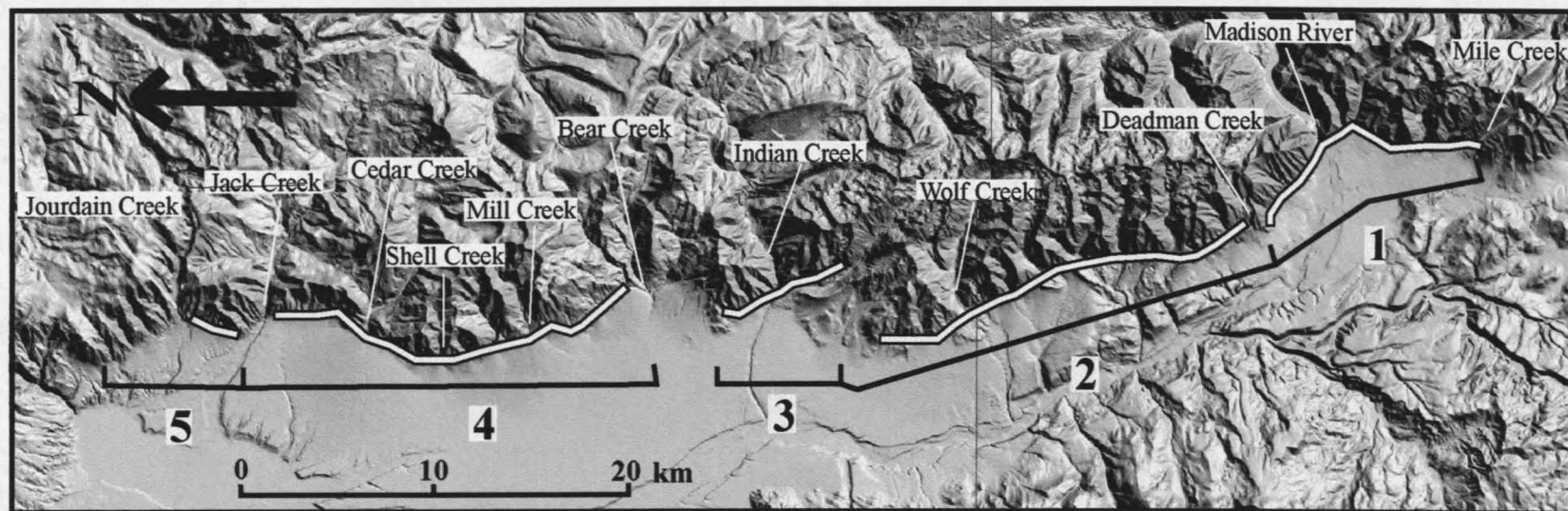


Figure 9. Madison Range fault (MRF) physiography. White line showing late Pleistocene (< 130 ka) surface ruptures. Sections listed as follows: 1 - Madison Canyon/Missouri Flats, 2 - Taylor-Hilgard, 3 - Indian Creek, 4 - Cameron, 5 - Spanish Peaks.

Five main sections of Quaternary faulting have differing structural control and morphologic expression: 1) Missouri Flats, 2) Taylor-Hilgard-Indian Creek section, 3) central Madison Valley from Bear to Jack Creeks, and 4) Jack Creek to Watkins Creek (Schneider, 1990) (Figure 9). The southern extent of the Madison Range at Mile Creek and Reynolds Pass, within the Missouri Flats region, displays a complex pattern of faulting. Basin depths are less than 1000 m (Peterson and Witkind, 1975) and Archean basement within the hangingwall of the Madison Range fault is present at the surface (Horn Mountains). The Missouri Flats region is a transition zone between the E-W trending Centennial fault and the N-S trending MRF displaying complex fault orientations. A narrow trough capped by the Huckleberry Ridge Tuff (2.1 Ma) connects the two basins, through which the Centennial Valley drained northeast into the Madison Valley until mid-Pleistocene time (Gary, 1980).

Historical and contemporary seismicity show a wide array of focal mechanisms for events in the Madison Canyon/Missouri Flats segment (e.g., normal, thrust, and strike-slip; Stickney, 1995). Focal mechanisms show that this region has T-axis orientations of NNW, sub-parallel to the Madison Range fault (Stickney, 1995). Evidence for structural control of Quaternary extension exists in lineaments along-trend of pre-existing shear zones (e.g., Early Proterozoic Madison mylonite zone, Erslev, 1982), minor embayments and re-entrants marked by coincident lateral thrust ramps preserved within the footwall, and historic fault plane solutions showing a connection between pre-existing structures and Neogene high-angle normal faults (Doser, 1985)

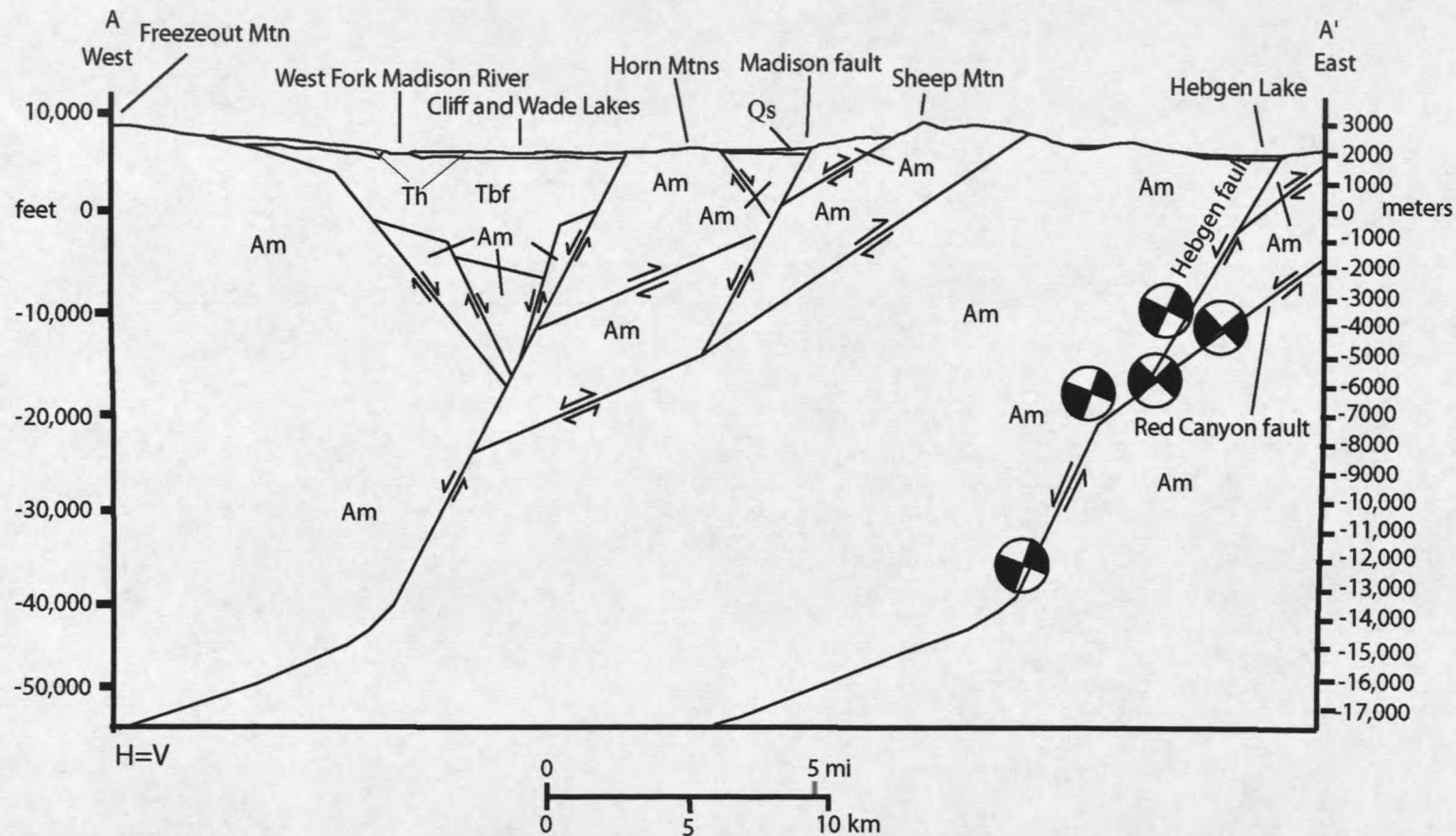


Figure 10. Simplified cross section line A-A' southern Madison Range fault. Focal mechanisms and Hebgen/Red Canyon fault system from Doser (1985). Map units listed as follows: Am - Precambrian metamorphic rocks, Tbf - Tertiary basin fill, Th - Huckleberry Ridge Tuff (2.0 Ma), and Qs - Quaternary sediments. Cliff and Wade Lakes are the location of a paleodrainage flowing north from the Centennial Valley to the Madison Valley.

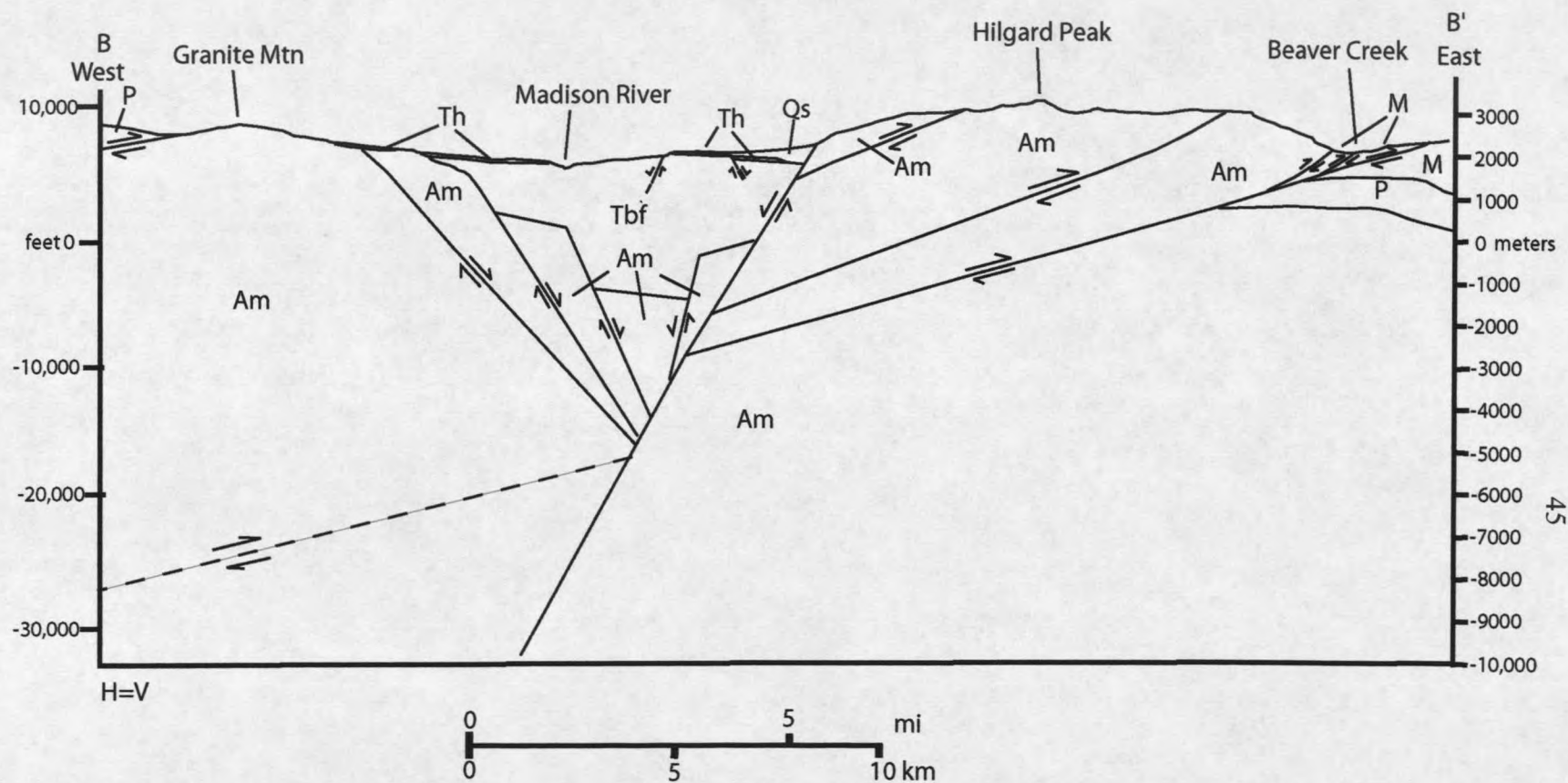


Figure 11. Simplified cross section B-B' Taylor-Hilgard section of the Madison Range fault. Maximum basin fill estimated as 3300 m (Gary, 1980) and 4500 m (Rasmussen and Fields, 1985). Units listed as follows: Am - Precambrian metamorphic rocks, P-Paleozoic rocks, M - Mesozoic rocks, Tbf - Tertiary basin fill, Th - Huckleberry Ridge Tuff (2.0 Ma), and Qs - Quaternary sediments.

(Figure 10). Therefore, the wide array of historical and contemporary seismicity may be a product of regional Basin and Range province extension interacting with pre-existing structures, rather than multiple stress fields creating the seemingly chaotic seismicity.

The Taylor-Hilgard segment (Figure 11) is characterized by Quaternary deposits juxtaposed with Archean basement rock. If modeled after the fault plane solutions for the 1959 Hebgen Lake earthquake and more recent seismicity, the main normal fault extends to the base of the seismogenic crust, approximately 15 km, at a 60°-70° dip (Doser, 1985) (MBMG, 2001, unpublished data) (Figure 12). This section of the range also contains the greatest thickness of basin deposits. Holocene fault scarps are nearly continuous along this section.

A bench held up by the Huckleberry Ridge Tuff lies 200 to 300 m above the Madison River between Moose and Curlew Creeks and is tilted into the range front, but erosion and poor exposure make an accurate measurement of the degree of tilt difficult. Intra-basin faults with small total slip appear to have rotated basin sediments into the MF. If gravity data analyses are accurate, this section of the fault zone is approximately 10 km wide and as great as 3000 to 4000 m in depth to Archean basement rocks (Rasmussen and Fields, 1983; Schofield, 1981). This contrasts with sections to the north from Indian Creek to Watkins Creek, where basin depths consisting of Cenozoic basin fill decrease to approximately 2100 m (Rasmussen and Fields, 1983). Thrust faults within the Hilgard thrust system are not exposed within the footwall of the Madison fault north of Indian Creek. Figure 13 shows a cross section of the Indian Creek segment and structural relationships between Laramide thrust faults and Neogene normal faults.

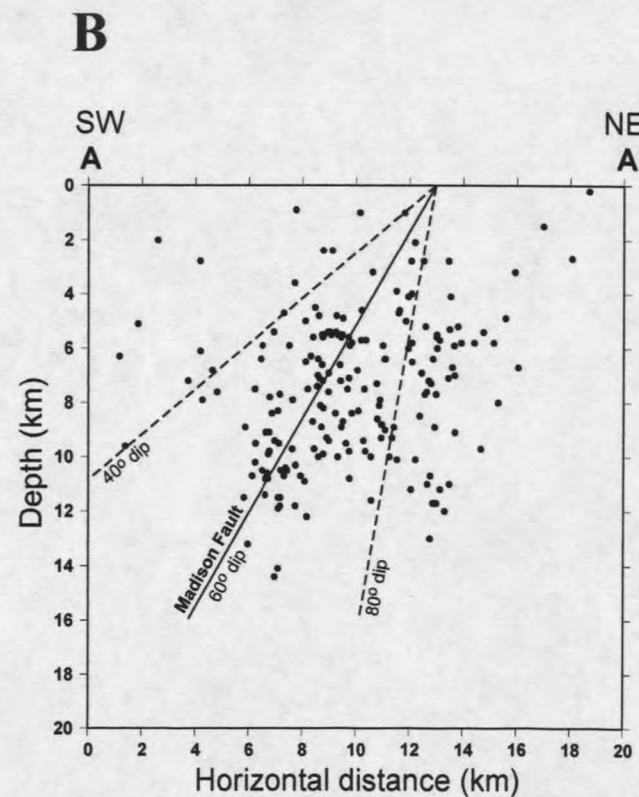
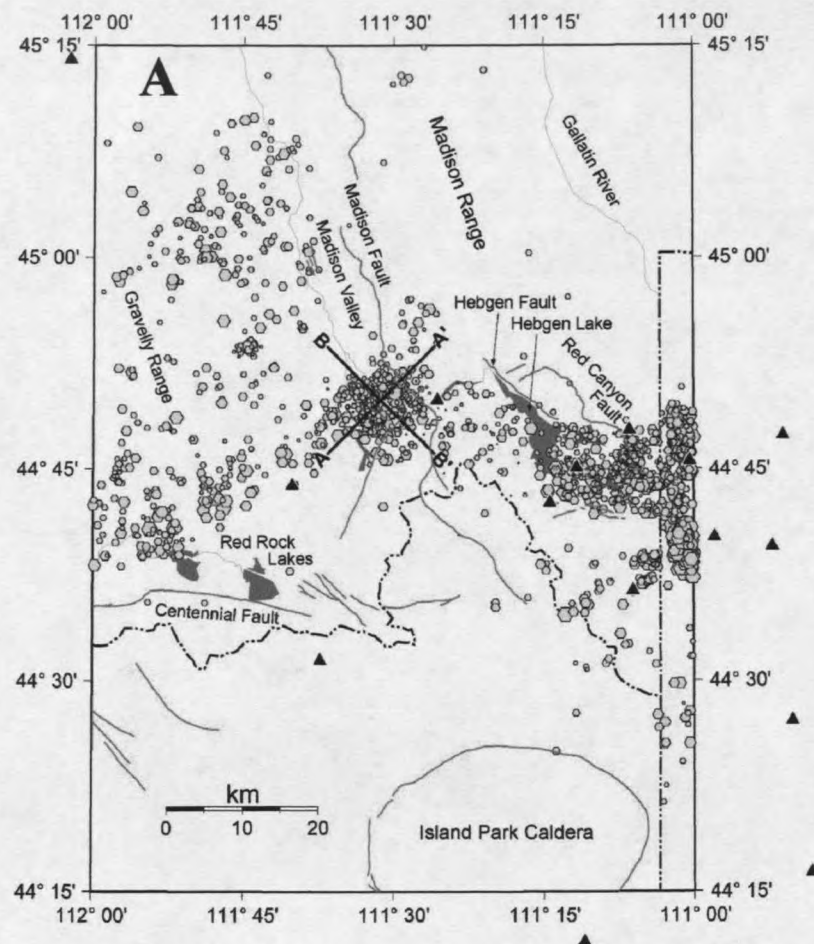


Figure 12. Seismicity of the Hebgen Lake/Southern Madison Range region 1982-2000. A) Epicenters and location of major faults within the southern Madison Range/Hebgen Lake/ Centennial Valley region. B) Foci across line A-A' and inferred Madison Range fault projection (M.C. Stickney, 2001, personal communication).

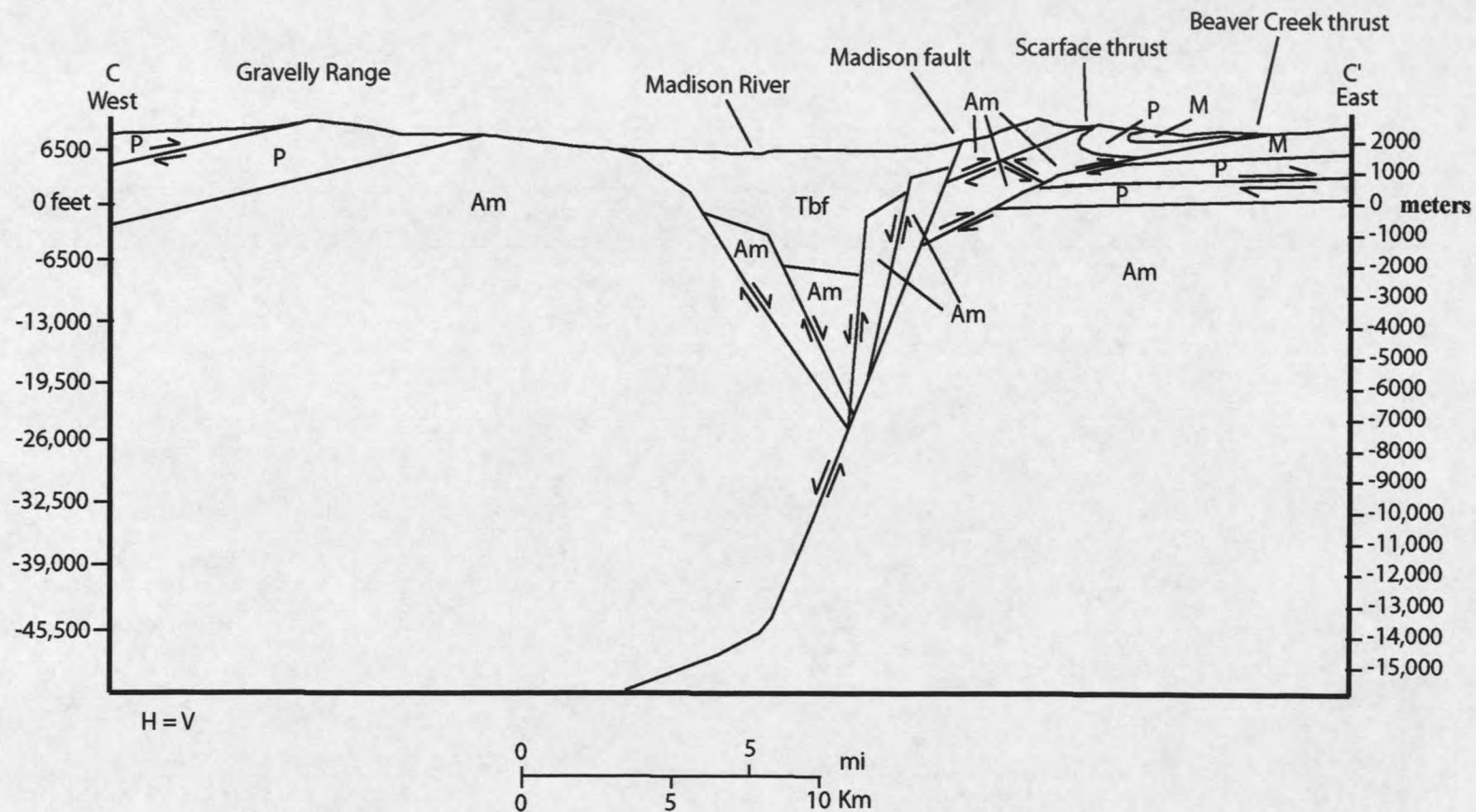


Figure 13. Simplified cross section C-C', Indian Creek segment of the Madison Range fault. Rock units are listed as follows: pCm - pre-Cambrian metamorphic rocks, P - Paleozoic sedimentary rocks, Tr-K - Triassic to Cretaceous sedimentary rocks, T - Tertiary basin fill. Modified from Schmidt and others (1993).

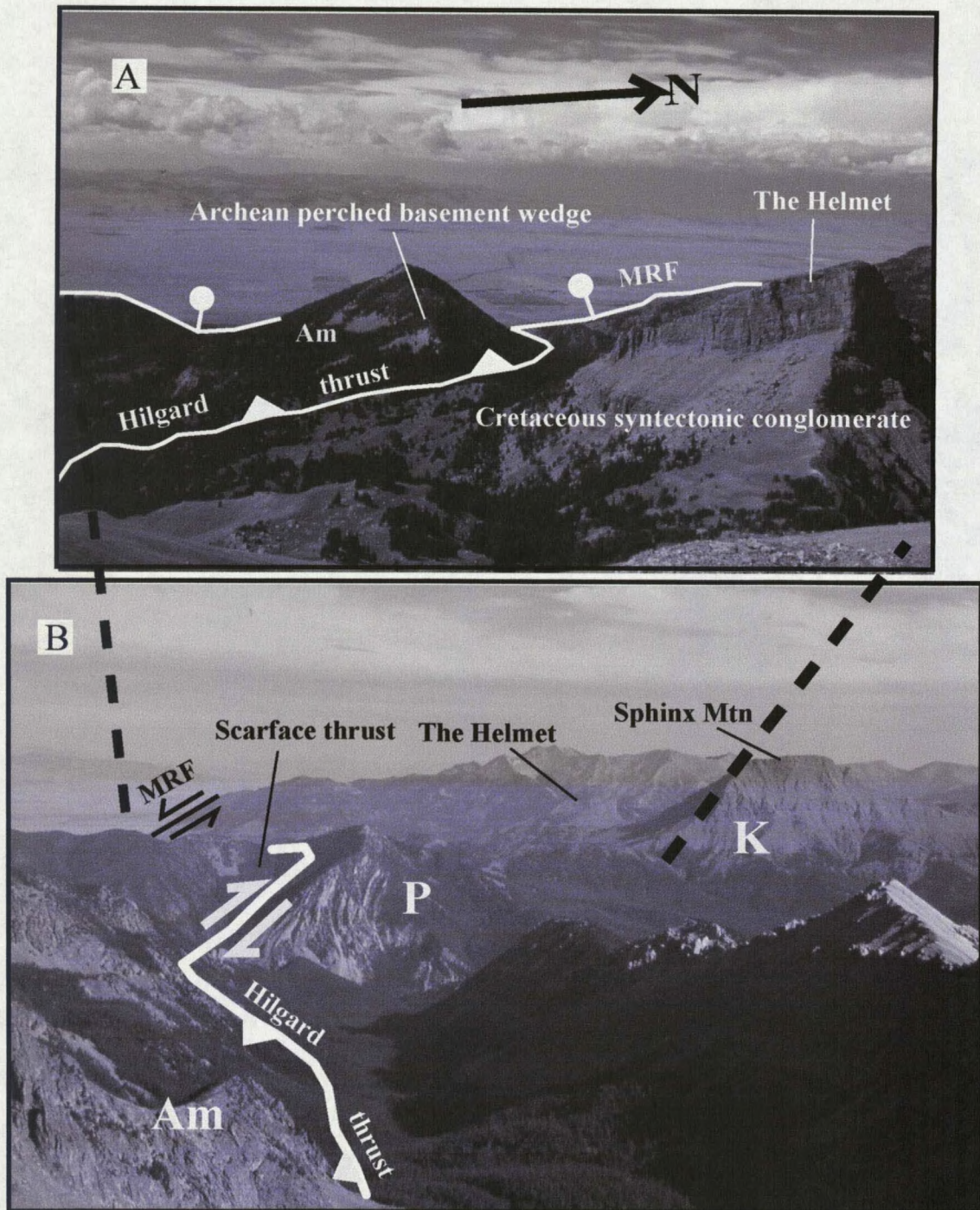


Figure 14. Structural relationships between the Hilgard thrust system and the Madison Range fault. A) Hilgard thrust cut by MF west of Sphinx Mountain, resulting in a "perched basement wedge" (Lageson and Zim, 1985). B) View from Koch Peak looking north at overturned Paleozoic rocks and the Scarface thrust (Schmidt et al., 1993) of the Hilgard thrust system (Kellogg et al., 1995). Am - Archean metamorphic rocks, P - Paleozoic rocks, K - Cretaceous rocks, MRF - Madison Range fault.

Figure 14 shows field relationships between the Neogene range-bounding normal fault of the Madison fault, the Hilgard thrust, and syntectonic conglomerate rocks, resulting in a "perched basement wedge" (terminology from Lageson and Zim, 1985).

For the Cameron section (Figure 9), evidence for normal dip-slip displacement along multiple synthetic faults is observed between Bear and Mill Creek. Along this reach of the Madison fault zone, imbricate thrusts within the Cretaceous Everts and Virgelle Formations have late Pleistocene normal dip-slip fault scarps displacing the hangingwall of these pre-existing thrust faults. The section of the range front from Bear Creek to Cedar Creek coincides with thrusts and folds predominantly within the Phanerozoic section. Along this section of the Madison fault zone, the Hilgard thrust system (Kellogg et al., 1995) is only exposed within the footwall between Tolman and Shell Creeks where a "perched basement wedge" consisting of Archean marble crops out. From Tolman Creek to Cedar Creek, the range front is characterized by nearly vertical beds of the Mississippian Madison Formation and the Pennsylvanian Quadrant Formation with late Pleistocene bedrock fault-line scarps, or *nastri di faglia* (Yeats et al., 1997, p. 293), at the bedrock-colluvium junction detaching thrusts and folds within the Phanerozoic section.

A left-step in the Madison range front occurs approximately 1.5 km south of Boulder Creek (Figure 15) along the Cameron section (Figure 9), where Mississippian limestone has been thrust over Pennsylvanian sandstone. The left-stepping re-entrant occurs at the junction of a lateral ramp, trending sub-parallel to the mountain-piedmont

junction. Late Pleistocene offsets terminate in this minor embayment and steps to the west around the salient. No ruptures are found along the range front coincident with the re-entrant, indicating that the westward step in the range front is controlled by the

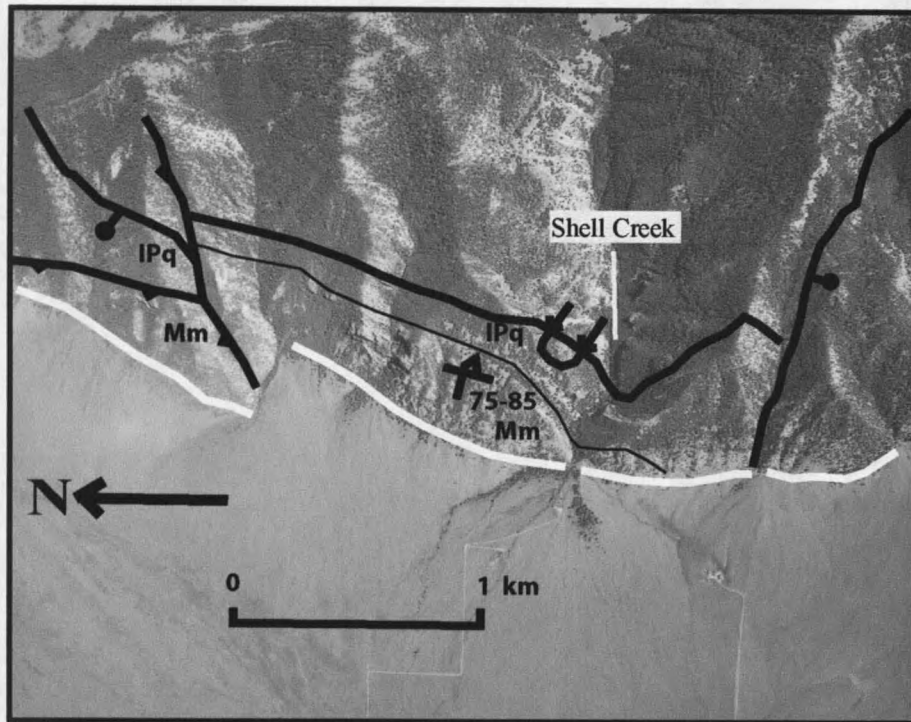


Figure 15. Thrusts and overturned syncline with normal faults at Shell Creek. White line shows trace of late Pleistocene surface rupture. Mm is Mississippian Madison Formation and IPq is Pennsylvanian Quadrant Formation.

position of the thrust fault and lateral ramp. This suggests that these imbricate thrusts are reactivated as synthetic normal faults with lateral propagation of surficial ruptures controlled by the position of the lateral ramps on individual thrust sheets. If these synthetic normal faults extend into the main detachment fault, as modeled in Figure 16, a large earthquake, $> M 7.0$, could potentially produce a co-seismic rupture of several or all of these connected synthetic faults, resulting in a complex pattern of surface rupture.

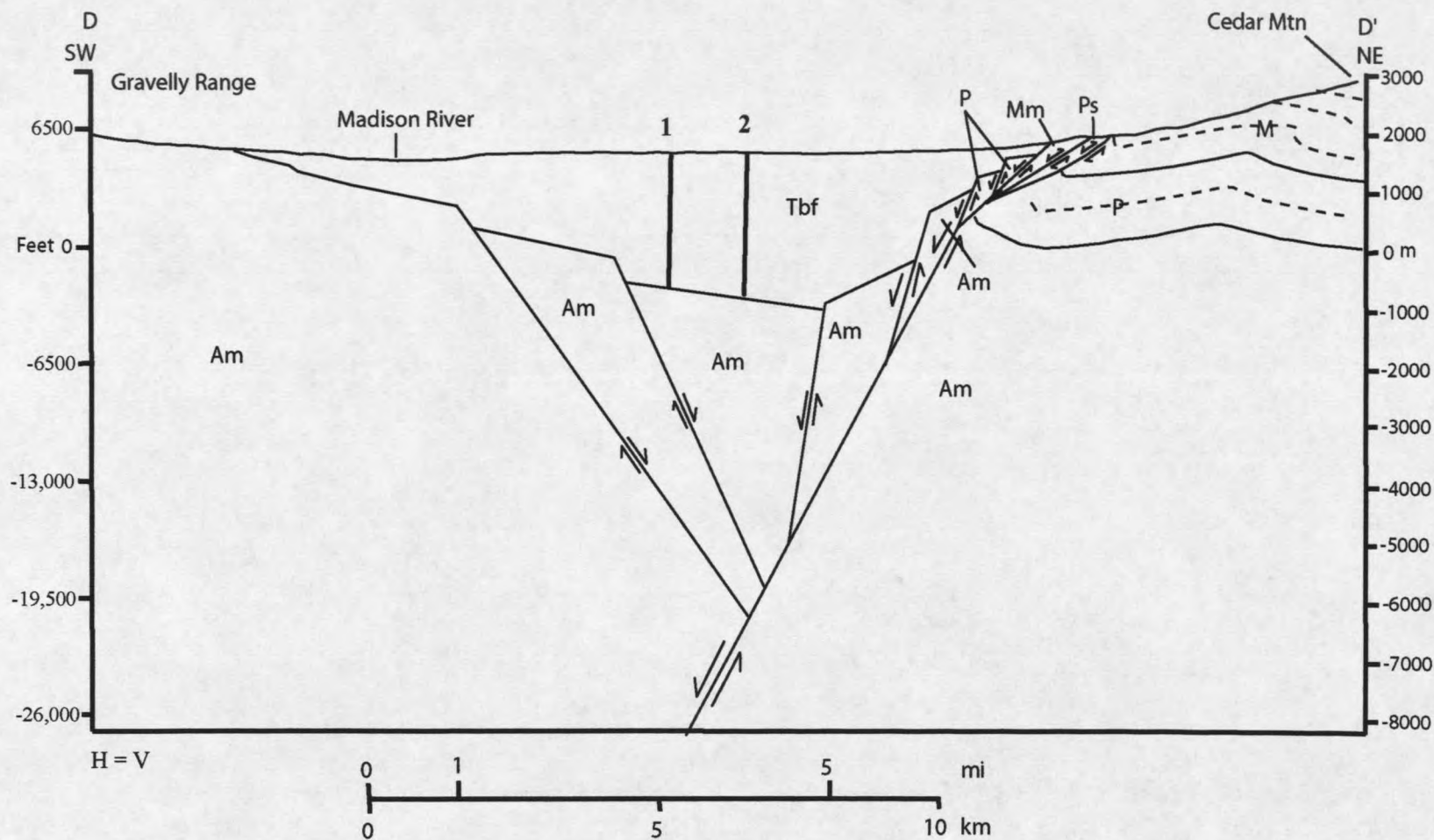


Figure 16. Madison Range cross section line D-D', Cameron section. Depth to Precambrian metamorphic rocks constrained by 1) approximately 2200 m (Gary, 1980) and 2) 1984 Texaco Granger 1, NE SE sect. 18, T.25, R.1E (pC at ~ 2293 m, Tertiary at or near surface, and Tertiary tufa at pC contact). Units listed as Am - Archean metamorphic rocks, P - Paleozoic rocks, M - Mesozoic rocks, Tbf - Tertiary basin fill, Ps - Lower Permian Shedhorn Sandstone, and Mm - Mississippian Madison Limestone.

Along the Cameron and Spanish Peaks sections from Cedar Creek north to Watkins Creek, the Hilgard thrust system is exposed along the range-front within the footwall as a "perched basement wedge". The range front consists of Archean quartzofeldspathic gneisses and amphibolite. Between Cedar and Jack Creek, a complex pattern of normal faulting is the result of extension along pre-existing thrusts where the Hilgard thrust converges with other minor thrusts within the Phanerozoic section. Normal-slip reactivation of multiple thrust splays coincides with lateral and frontal ramps of the Hilgard thrust system exposed in the footwall, again showing pre-existing structural control on patterns and geometry of extensional deformation. From Jack Creek north to Watkins Creek, the range front consists of Archean metamorphic rocks thrust over the Phanerozoic section. The range front is relatively linear and trends sub-parallel to the Hilgard thrust system. Between the North and South Forks of Jourdain Creek the trend of the range front changes from approximately 025° to 000° . This change is coincident with two high-angle reverse faults, one dipping southwest and the other northeast, associated with the Hilgard thrust system and Spanish Peaks fault, respectively.

In summary, extensional patterns of deformation displayed along the Madison Range fault relate to pre-existing contractional structures formed during prior tectonic episodes. Normal faults truncate at the intersection of lateral ramps, resulting in range-front steps. Greater basin depths to the south coincide with the region of greatest thrusting of the Hilgard thrust system, with a maximum dip-separation of 5000 m (Kellogg et al., 1995), where one main detachment appears to be accommodating all of the displacement in the core of the Madison-Gravelly arch. The greatest amount of Neogene

extension occurs where the greatest amount of dip separation occurred within the Hilgard thrust system.

Quaternary Deposits and Depositional History

The surficial deposits adjacent to the MRF consist almost entirely of the penultimate and the last glacial periods, Bull Lake and Pinedale, respectively. These probably correlate with marine oxygen isotope stages 6 and 2. Embayments located at the South Fork of Indian Creek and Bear Creek have perched older surfaces that were present before Bull Lake glaciation. These surfaces are relay ramps located between fault zone segments at Indian and Bear Creeks. Deposition of Bull Lake and Pinedale moraines on top of a bedrock surface in between Indian and Corral Creeks suggests that these older perched surfaces were erosional until the Bull Lake event. Displacement along the MRF has resulted in accommodation space for the deposition of Bull Lake and Pinedale-aged deposits on the down-thrown side.

The southern portion of the Madison Valley is capped by the Huckleberry Ridge Tuff (HRT) (2.0 Ma) (Christiansen and Blank, 1972). Coarse gravels beneath the Huckleberry Ridge Tuff contain Proterozoic Belt Supergroup argillite, quartzite, and sandstone clasts. Schneider (1994) concludes that the coarse, well-rounded characteristics of these deposits indicate a relatively high-energy fluvial system. The sources for these Proterozoic clasts lie to the NNW, suggesting an ancient paleovalley draining southward prior to the emplacement of the Huckleberry Ridge Tuff (Fritz and Sears, 1993).

Although not within the mapped area, freshwater limestones exist west and south of Ennis, to the central Madison Valley. Sandstone and gravel facies of this limestone deposit contain Huckleberry Ridge Tuff clasts according to Schneider (1994). The limestones have been interpreted as perennial, freshwater lacustrine deposits (Hadley, 1980) and are restricted to the northern half of the valley, north of Indian Creek. Schneider (1994) proposed that the southern Madison Valley was built up and dammed by the HRT. The Cameron Surface was down-dropped to the north and tilted into the northeast-striking, northwest-dipping Indian Creek fault (Plate 2) prior to deposition of Bull Lake and Pinedale equivalent deposits. This faulting, perhaps coupled with increased surface runoff induced by the welded tuff cap, promoted the development of perennial "Lake Hadley" (Schneider, 1994). Therefore, the maximum age of this lake may be 2.0 Ma.

The maximum basin fill level is recorded by the Cameron Surface (Paul and Lyons, 1960). Within the Madison valley between Bear and Mill Creeks, the Cameron Surface is tilted 4°-6° into the MRF causing Pinedale and Bull Lake fan deposits to grade into one another (Paul and Lyons, 1960). Schneider (1994) demonstrated that Bull Lake and Pinedale terraces are cut-and-fill features and that post-Bull Lake incision by the Madison River exceeds 50 m. Based upon regional correlations between timing of glacial maximums, the ages of these glacio-fluvial fans correlate with the ages of the Bull Lake (~140 ka) and Pinedale (perhaps 15-20 ka) glacial moraines and outwash (Pierce, 1979).

Glacial deposits consisting of lateral and terminal moraines extend onto the piedmont at Mile, Deadman, Papoose, Squaw, Moose, Wolf, and Indian Creeks, as first noted by Sheldon (1960). Pierce (1979) showed that glaciers extending from the Yellowstone icecap occupied the Paradise Valley, approximately 50 km to the east, until about 20 ka. Within the Paradise Valley, cosmogenic nuclide dating of surface boulders on Pinedale moraine sequences yield ages for outer Pinedale moraines of 16.5 ± 0.4 ^3He ka, with deglaciation of the Yellowstone Plateau by ~ 14 ^{10}Be ka (Licciardi et al., 2001). Since the Madison Range glaciers were separate alpine glaciers and not part of a much larger ice cap, the 20-25 ka age for Pinedale maximum (Pierce, 1979) provide a rough estimate for the Pinedale terminal moraines and associated outwash within the Madison Valley. These glacial deposits predominantly emanate from the Taylor-Hilgard section of the range. At Wolf, Squaw, and Papoose Creeks, older Bull Lake moraines lie beneath Pinedale moraines (See Plate 1).

No Bull Lake-equivalent deposits could be found within the footwall adjacent to the MRF to determine amount of offset. At Wolf Creek along the Taylor-Hilgard section, a crude approximation of total offset of Bull Lake deposits yields between 70-80 m, based on projections of older Bull Lake moraine surfaces into the footwall. At Indian Creek a Pinedale moraine is perched on top of Bull Lake age moraines, and both lie above of a bedrock surface consisting of Archean gneiss and Eocene or Oligocene volcanic and volcanoclastic rocks (Kellogg and Williams, 2000).

Within the southern Madison Valley along the Missouri Flats section, Pinedale and Bull Lake glacio-fluvial fans have interacted with fluvial deposition and trimming by

the Madison River system. Lundstrom (1986) shows how the series of terraces at the mouth of Madison Canyon at Quake Lake are directly related to glacial damming of the Madison River upstream at Beaver Creek. The Cliff and Wade Lake vicinity is characterized by a deeply incised, rhyolitic benchland held up by the HRT (Gary, 1980).

Holocene deposition within the Madison Valley has been limited to small alluvial cones and debris flows proximal to the range front, along with alluvium in active drainage channels. Holocene deposits are predominantly fine-grained, limited in extent, and proximal to large drainages, indicating discharge flows much smaller than those occurring with Pinedale glaciation, as demonstrated by Pierce and Scott (1982). Dissection of Pinedale glacio-fluvial fans has been minimal, 2 to 5 m. The current flow regime and energy of the Madison Valley fluvial system is assumed to be indicative of the Holocene.

Morpho-Metric Analyses

Fault Scarp Profiles and Offset Deposits

A total of 103 profiles of the MRF were measured along the Madison fault zone (Plates 1-3 and Appendix A). A Bucknam-Anderson "log scarp height versus maximum slope angle" graph was derived from fault scarp profiles (Figure 17). Regression lines determined for known-age events and fault scarps from Drum Mountain, Bonneville (Bucknam and Anderson, 1979), and Lost River, Idaho region (Pierce, 1985; Pierce and

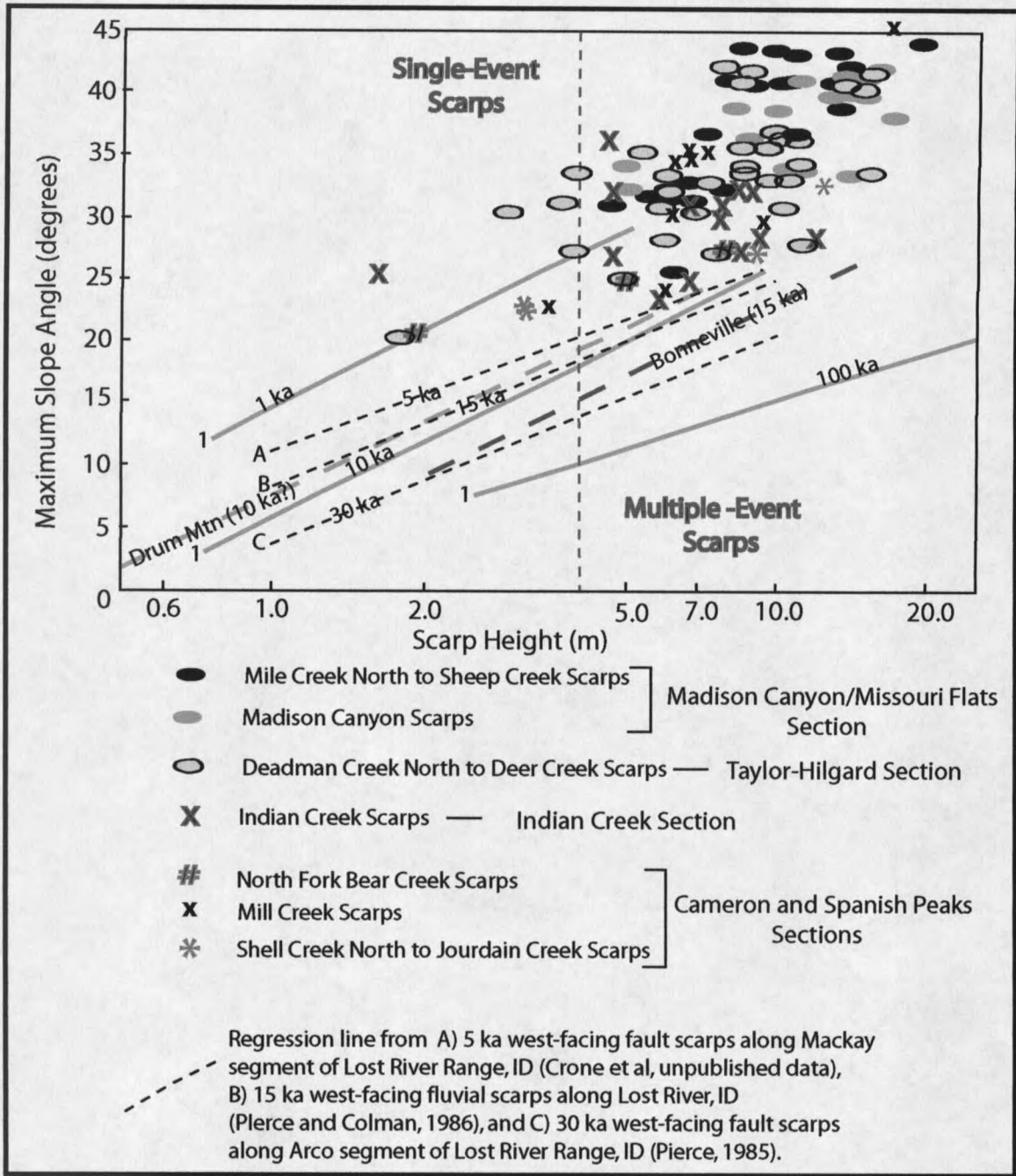


Figure 17. Scarp height vs. maximum slope angle plot for fault scarp profiles along the Madison Range fault. Scarp height (SH) and maximum slope angle for segment location of fault scarp profiles indicated by symbols listed above. Other fault scarp regression lines are placed for comparison. Solid gray lines, labeled 1, from Bucknam and Anderson (1979). Based on regression lines from previous studies within comparable climatic regimes, most recent event is between 1 ka and 15 ka for the entire Madison fault zone. Sections of fault for each symbol listed to the left of symbols.

Colman, 1986; Crone et al., unpublished data) are plotted to show relationships with Madison fault scarp profile data. Mid-Pleistocene to Holocene surface ruptures characterize the majority of the Madison Range front from Mile Creek north to Jourdain Creek. The two centrally-located embayments show no evidence of late Pleistocene offset.

The southern two segments, Taylor-Hilgard and Madison Canyon/Missouri Flats, both have fault scarps on late Pleistocene moraines and fans, as well as early- to mid-Holocene fan and debris flow deposits (Figure 18). Surface offsets on early- to mid-Holocene deposits range from 2.2-4.0 m, with a mean around 3.2 m. These are predominantly single-event fault scarps with average maximum slope angles of 32.8° . From these fault scarps, a characteristic earthquake (Schwartz and Coppersmith, 1984; Hecker and Schwartz, 1994) with surface offsets ranging from 2.0-3.0 m is reasonable for maximum earthquake analyses. Proposed single-event fault scarps within the Missouri Flats area show slightly higher surface offsets than the Taylor-Hilgard segment, possibly created as secondary effects from large earthquakes in the region over-printing single-event fault scarps (Wallace, 1979, 1980). However, it is still unclear if the 1959 slip along the MRF was coseismic or secondary "shake-down" deformation (Myers and Hamilton, 1964).

The Madison Canyon/Missouri Flats segments have fault scarp surface offsets ranging from 2.2-14.4 m. Late Pleistocene fan deposits are offset 6.8-14.4 m (Figure 18). Small alluvial cones and debris flows of early- to mid-Holocene age are offset 2.2-6.5 m. A large landslide at the mouth of an unnamed canyon north of Little Mile Creek shows a

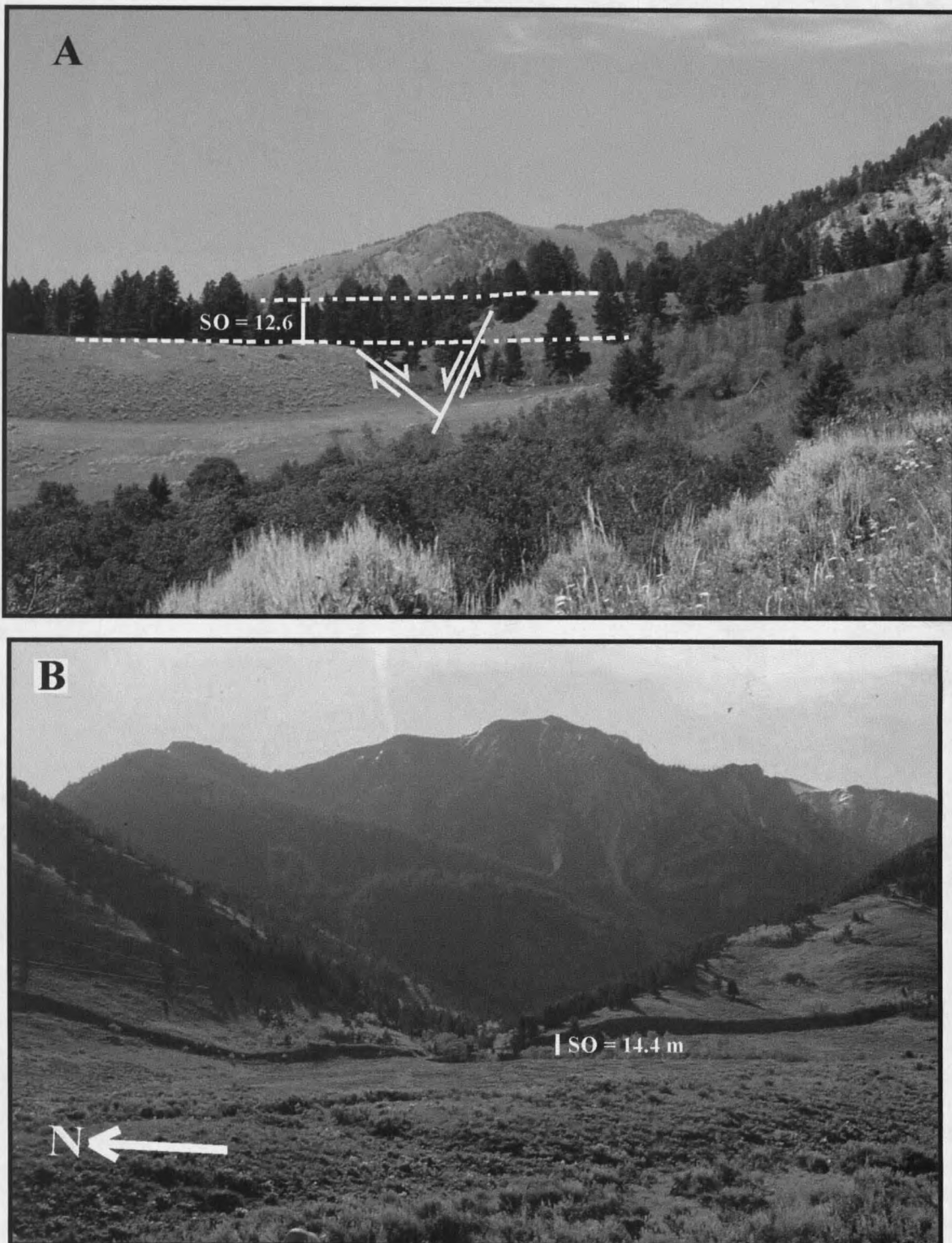


Figure 18. Taylor-Hilgard and Madison Canyon/Missouri Flats sections fault scarps. (A) Offset Pinedale lateral moraine at South Fork Squaw Creek, SO = 12.6 m, looking north. (B) Offset Pinedale fan alluvium at mouth of Little Mile Creek, SO = 14.4 m.

surface offset of 8.5 m. Between Sheep Creek and this unnamed canyon, the fault is right at the bedrock/alluvium-colluvium contact. The fault tapers into Mile Creek canyon and offsets a Pinedale moraine 7.0 m.

Late Pleistocene and Holocene deposits are offset from Deer Creek south to Deadman Creek. Late Pleistocene recessional outwash deposits have surface offsets of 5.9 m to 9.4 m, with younger early- to mid-Holocene fan deposits offset 2.3 m to 3.7 m (Figure 19B). Pinedale moraines are offset the greatest with surface offsets ranging from 7.6 m to 12.6 m (Figure 18).

Early- to mid-Holocene alluvial cones and debris flows are offset along the Indian Creek and Cameron segments. Two profiles show surface offsets of 2.4 m and 3.0 m on these deposits at Indian Creek and Mill Creek, respectively. This gives further evidence for a characteristic earthquake with surface offset in the 2.0-3.0 m range. North of Mill Creek, early- to mid-Holocene deposits are not offset.

The Indian Creek segment shows a maximum surface offset of 7.1 m on late Pleistocene deposits. This segment is bound by the east-west trending salient on its northern boundary. Displacement of this bedrock salient is influenced by the northeast-trending intrabasin Indian Creek fault, which does appear to have late Pleistocene displacement. Late Pleistocene surface ruptures do not propagate past this salient and surface offsets decrease to 3.1 m on late Pleistocene deposits. Early- to mid-Holocene deposits have a surface offset of 2.4 m.

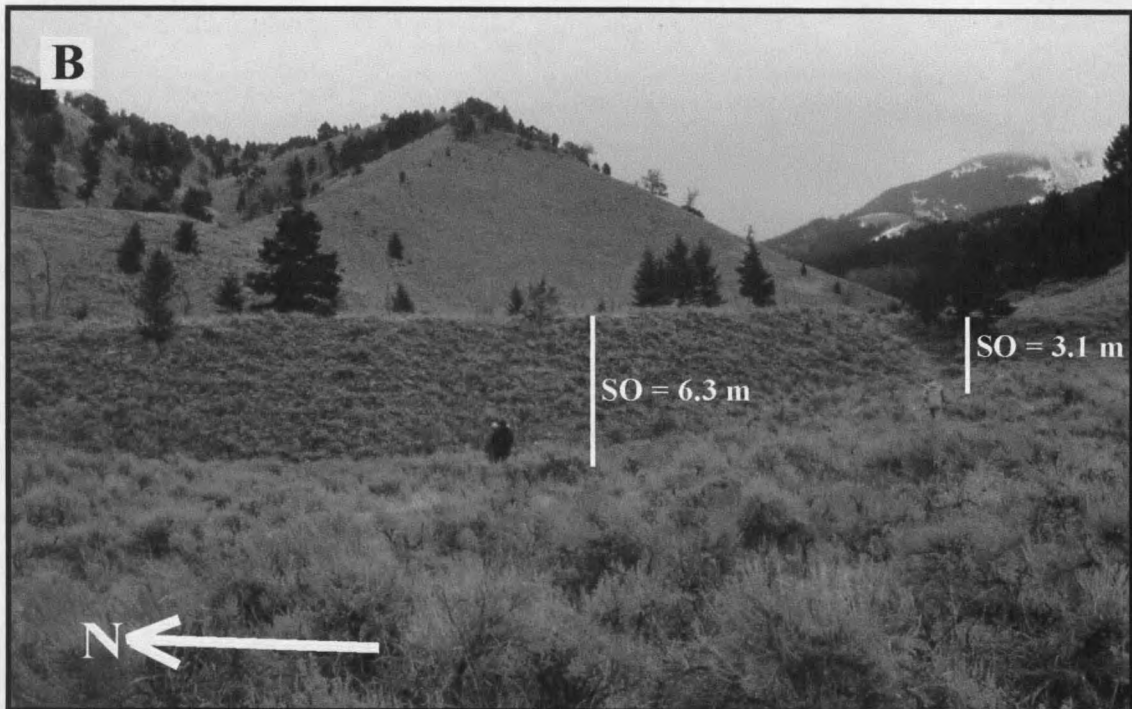
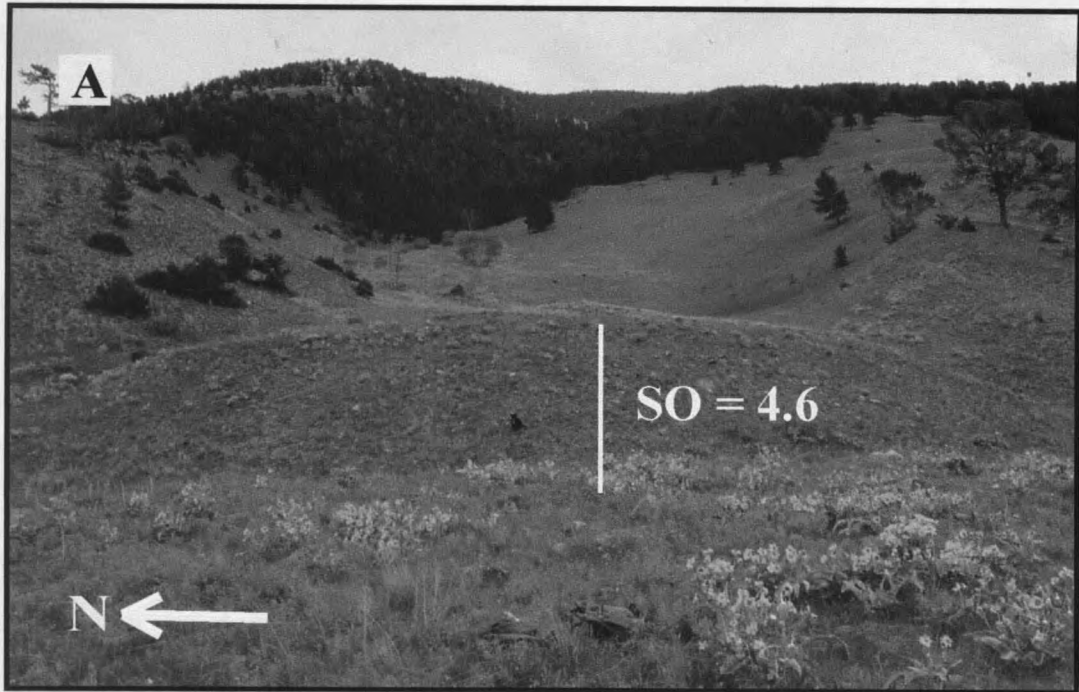


Figure 19. Cameron and Taylor-Hilgard section fault scarps. (A) Fault scarp on late Pleistocene debris flow at Mill Creek, SO = 4.6 m. Dog for scale in center of fault scarp. (B) Fault scarp at Stock Creek offsetting late Pleistocene and mid to early Holocene fan alluvium and debris flows. Note person for scale.

Figure 19A shows a fault scarp on a late Pleistocene debris flow one drainage south of Mill Creek. Between Mill and Bear Creeks, the fault zone is expressed as discontinuous fault scarps trending along-strike of pre-existing thrust faults mapped in the footwall, and appear to be detaching the hangingwall of these thrust faults as previously discussed. Late Pleistocene deposits are offset 4.0-6.4 m. Based on moment magnitude parameters (Hanks and Kanamori, 1979), with an inferred characteristic earthquake displacement of 2.0-3.0 m, at least two post-Pinedale events have occurred along this segment of the MRF.

Between Cedar and Jack Creeks, where a complex pattern of normal faulting appears to be related to pre-existing contractional structures, no late Pleistocene deposits are offset. Just south of Jack Creek the fault is expressed as a fault-line scarp at the bedrock-colluvium contact. Late Pleistocene fans are deposited on the downthrown side of a larger mid-Pleistocene fault scarp. Along this section fault scarps appear to have been fluvially trimmed and steepened by Holocene erosion. Late Pleistocene deposits are inset approximately 1.0 m into older colluvium-alluvium. Therefore, some faulting had to occur at this location prior to deposition of the late Pleistocene fans. This suggests mid-Pleistocene faulting and late Pleistocene to Holocene quiescence along this reach of the fault.

From Jourdain Creek south to Mill Creek, fault scarps on late Pleistocene fan alluvium and undivided colluvium/alluvium have surface offsets ranging from 1.3 m to 8.8 m just south of Shell Creek. A small single-event fault scarp (SO = 1.3 m) exists between Jack and Jourdain Creeks (Figure 20A). At McDeed Creek, just south of Cedar

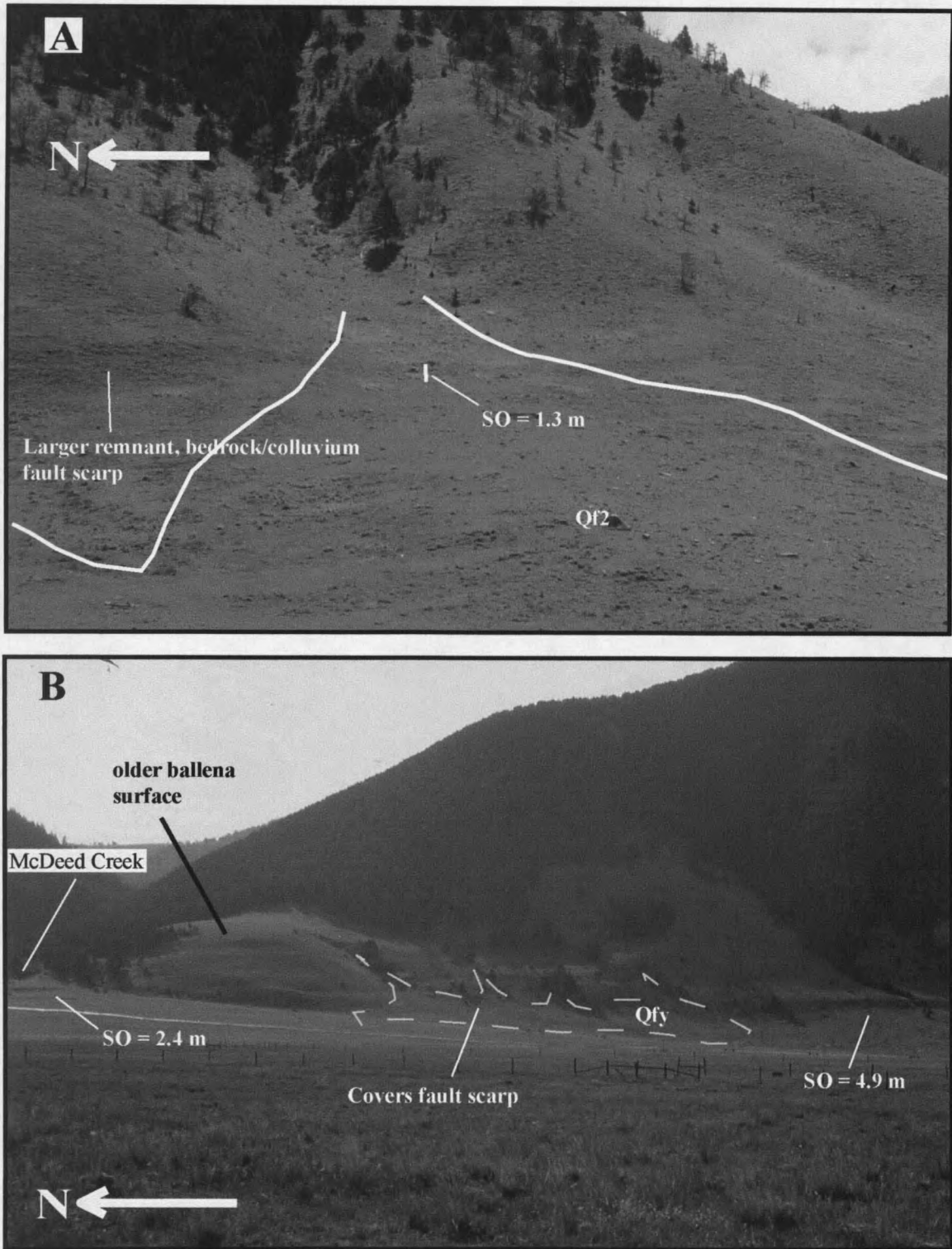


Figure 20. Fault scarps along northern MRF. (A) Fault scarp between Jack and Jourdain Creeks, Qf2 - Late Pleistocene alluvial fan. (B) Fault scarp south of McDeed Creek, early- to mid-Holocene debris flow (Qfy) covers the late Pleistocene fault scarp.

Creek, a late Pleistocene fan is offset 2.4 m. To the south older range-front colluvium/alluvium is offset 4.9 m (Figure 20B). From McDeed Creek south to Tolman Creek, the fault scarp is right at the bedrock/colluvium junction and is expressed as *nastri di faglia*, or bedrock fault scarps with freshly exposed surfaces. South of Shell Creek a late Pleistocene fan deposit is offset 8.8 m.

The log scarp height versus maximum slope angle plot (Figure 17) shows that all single-event fault scarps along the Madison Range fault zone would plot above the 10^3 year and 10^4 year regression lines from Bucknam and Anderson (1979). This indicates that all profiled fault scarps along the Madison Range fault zone have been formed or regenerated within the last 10,000 years. Most of the fault scarps plot above the 5 ka regression (Crone, unpublished data) from the Lost River fault, ID, leading to an even younger age for the most recent event. Fault scarps along the Madison Canyon-Missouri Flats and Taylor-Hilgard segments plot proximal to the 10^3 year regression line, suggesting a most recent event within the last 1-2 ka. The Indian Creek, Cameron, and Spanish Peaks segments fault scarps plot between the 10^3 year regression and the 10^4 year regression, but plot above the 5 ka regression from the Lost River Valley, ID.

Based on the Bucknam-Anderson plot (Figure 17), the most recent event for the entire MF ranges between 1-12 ka, with the most recent activity occurring along the southern half of the fault system, the Taylor-Hilgard and Madison Canyon/Missouri Flats segments. With the given fault scarp data, the southern half of the MF appears to have slip rates double that of the northern half (0.6 mm/yr versus 0.3 mm/yr, respectively) and recurrence intervals half of those to the north. This may suggest that activity to the south

is concentrated along the southern sections of the MRF. However, the other morphometric analyses show that tectonic activity and the resulting slip rates average out to suggest similar slip rates along the entire range front. This is also supported by the high topography (ranging between approximately 3200-3400 m) along the entire range crest. If the overall tectono-morphic signal suggests the entire fault zone has a slip rate on the order of 0.1 mm/yr for 10^5 - 10^6 year periods, the mid to late Pleistocene bifurcations in slip rate along the fault could possibly show evidence for late Pleistocene-Holocene temporal clustering of events. Late Pleistocene grouping and migration of events induced by stresses associated with the Yellowstone volcanic center could be a cause for the slip rate fluctuations on the order of 10^3 - 10^5 years.

Relative Tectonic Activity Class Designation Chart

Valley-floor width to valley height ratios for the Madison Range were measured approximately 0.75-1.5 kilometers into the footwall, depending on which segment was measured. All ratio values are less than 0.1 except at Mile Creek, where the Southern Madison/Missouri Flats segment boundary occurs and glacial erosion have induced higher V_f ratio values. V_f ratio for the majority of the range front is from 0.01-0.08, with a mean around 0.04 (Figure 21 and Appendix B). All drainages orthogonal to the range front show deep, narrow incision. Drainages just south of Madison Canyon are characterized by steep gradients and narrow channels almost completely dominated by landslides and debris flows. This same characteristic is found between Mill and Cedar

Creeks. Along this section of the fault, all of the drainages are deeply incised into Paleozoic carbonate and siliciclastic rocks and V_f values range from 0.03-0.06.

Analysis of V_f values (Figure 21) for each segment of the Madison Range fault leads to several observations. Between Bear Creek and Cedar Creek (V_f measurements 7 to 15), V_f values range from 0.06 at the north and south ends to 0.02-0.04 between the two drainages. An increase in V_f ratio value from the center of this segment towards the north and south suggests a tapering of tectonic activity towards the ends of the Cameron segment, with the greatest elevation difference occurring in the center coincident with lower V_f ratios. The Indian Creek segment has values of 0.05-0.06, which yields a slightly higher average than to the north and south, suggesting less tectonic activity.

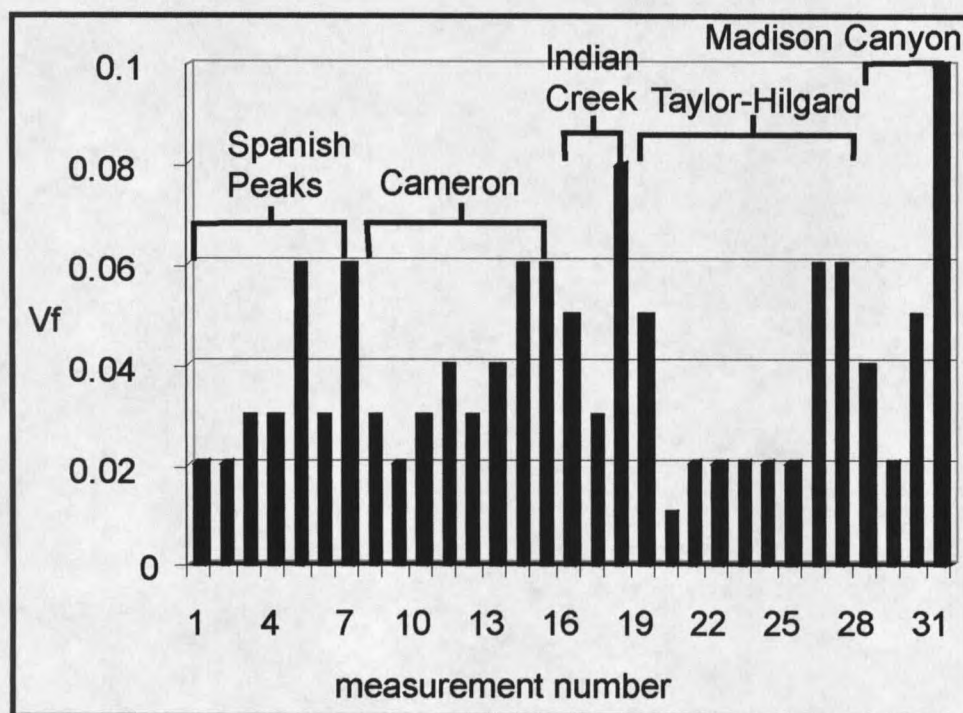


Figure 21. V_f values for the Madison Range fault. Sections of the Madison Range fault listed above graph. See Appendix B for data.

Along the Taylor-Hilgard segment, five of the eleven drainages have a value of 0.02 (V_f measurements 19 to 26, see Appendix B), increasing to the north and south of these five drainages, again suggesting a tapering of tectonic activity towards the ends of distinct segments of the MF.

Although some perturbations in V_f ratio values exist along the entire length of the Madison fault system, all of the values are within and below the range considered to be correlated to a Relative Tectonic Activity of 0.5 m/ka (Bull, 1984, 1987). As previously discussed, the tectonic activity rate assigned to differing values of V_f is a long-term average slip rate required to sustain these index values. Since morpho-metric indices are affected on a scale of greater than 10^5 years and recurrence intervals along individual segments can vary greatly on the order of 10^3 and 10^4 years, V_f ratio values for the Madison Range fault yield a long term tectonic activity rate of 0.5 m/ka.

Mountain front sinuosity values of 1.05 for the MRF infer a relative tectonic activity of 0.5 m/ka (Bull, 1984, 1987). The bedrock range front almost directly coincides with mid- to late-Pleistocene rupture patterns with little to no embayment of the range front at drainages. Therefore, the morphology of the Madison Range front suggests continuous tectonic activity maintained over periods on the order of $>10^5$ years. A long-term average tectonic activity rate of 0.5 m/ka for the last 250,000 years seems reasonable to maintain the range-front morphology. Based upon a 6° southeast tilt of the HRT on the west side of Missouri Flats, Weinheimer (1979) calculated at least 1000 m of offset along the MF, resulting in a long-term average slip rate of 0.5 mm/yr. This is in

accordance with the rates determined by mountain front sinuosity and valley-floor width to valley height ratios.

Late Pleistocene and mid Pleistocene fan deposits are entrenched; however, the entrenchment of these deposits may be more related to climatic fluctuations over the last 150,000 years than to low tectonic activity. Tectonic activity has been rapid enough to allow the basin to subside and provide the accommodation space to stack Pinedale stream and possibly glacial deposits above Bull Lake deposits, rather than cut and fill with entrenched deposits. Without these rapid tectonic activity rates, Pinedale glaciers would have simply entrenched and reworked Bull Lake moraines rather than depositing on top of the older deposits proximal to the range front.

Basal Facet Reconnaissance Technique

The Madison Range is marked by pronounced triangular faceted spurs along the entire length of the range front. Most facets are steep with little to no dissection and lack well-developed erosional rills along their faces. Basal facet heights along the range front range from 61 m to 305 m (Figure 22). Sections of the range front that have apparent greater mid- to late-Pleistocene surface offset also correlate to greater basal facet heights.

The southern Madison Range from Mile Creek to Corral Creek (Madison Canyon and Taylor-Hilgard sections) has basal facet heights ranging from 85 m to 280 m, with a mean of 194 m. Applying these basal facet heights to the basal facet regression equation developed through fault analyses in the southern Basin and Range province (dePolo, 1998) yields a minimum slip rate of 0.18 mm/yr for the minimum height, 0.61 m/ka for

the maximum height, and 0.36 m/ka for the average facet height. The presence of two anomalous basal facets of 85 m proximal to segment boundaries along the MRF suggests apparent differential offset. Because late Pleistocene and Holocene offset is similar along this entire segment of the fault system, this could be the result of either segment linkage at the two locations with the 85 m facet heights, prior to the more recently generated 85 m basal facet, or previous segment boundaries where offset tapered dramatically.

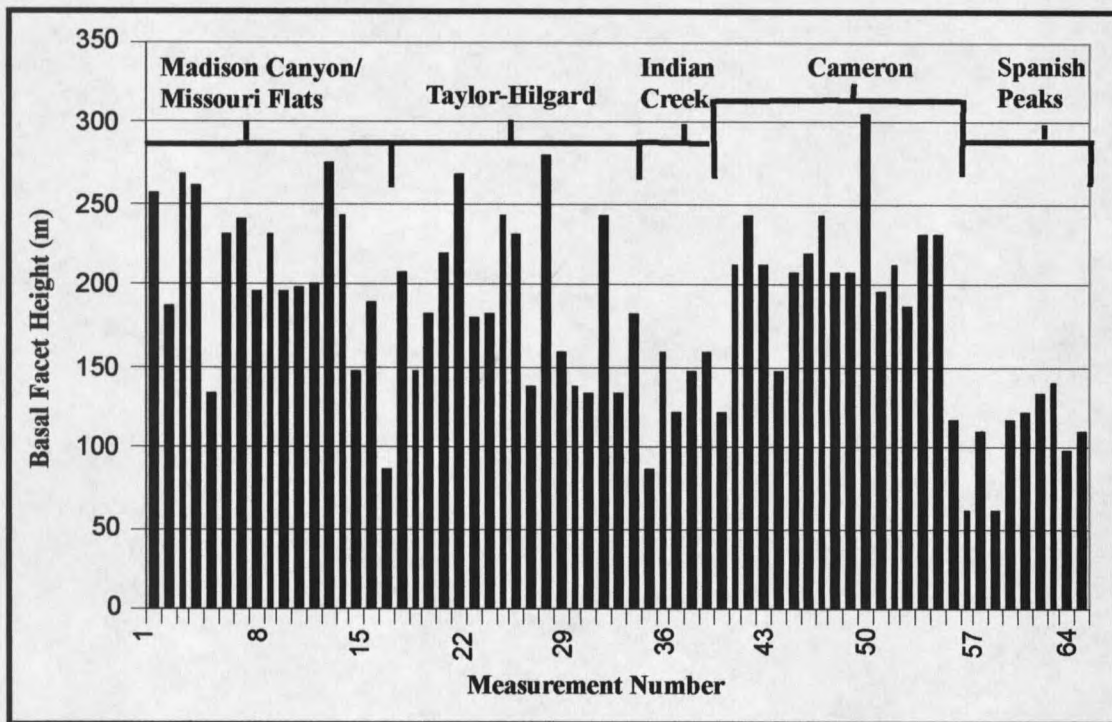


Figure 22. Basal facet heights for the Madison Range fault (MRF). Measurements were recorded south (1) to north (66). Sections for the MRF listed above facet heights. Anomalous low basal facet heights occur at segment boundaries suggesting a tapering of net displacement towards the ends of segments. See Appendix C for data.

The Indian Creek segment has basal facet heights ranging from 122 m to 244 m with a mean basal facet height of 183. Applied to the facet equation (dePolo, 1998), these heights yield a minimum slip rate of 0.23 m/ka, a maximum slip rate of 0.49 m/ka, and a mean slip rate of 0.34 m/ka. The Cameron section from Bear Creek to Jack Creek has basal facet heights ranging from 61 m to 305 m, with a mean of 166 m. Bedrock lithologies are variable from Cretaceous volcanic and volcanoclastic rocks to Archean quartzo-feldspathic gneiss. This section of the range front contains back-sliding on Laramide thrust faults with apparent differing amounts of mid- to late-Pleistocene surface offset along individual segments. Slip rates determined by the faceted spur reconnaissance technique yield a minimum of 0.16 m/ka, maximum of 0.71 m/ka, and a mean rate of 0.30 m/ka. The Spanish Peaks section from Jack Creek to Jourdain Creek has basal facet heights ranging from 98 m to 140 m, with a mean of 121 m. These heights yield slip rate values with a maximum rate of 0.26 m/ka, a minimum rate of 0.2 m/ka, and a mean slip rate of 0.23 m/ka.

In summary, a comparison of the slip rate values determined by analysis of basal facets shows a northward decrease in estimated mean slip rate from 0.36 m/ka to 0.34 m/ka to 0.3 m/ka to 0.23 m/ka. It must be noted that the Holocene to late Pleistocene slip rate between Bear and Cedar Creeks is variable with some sections having offset early- to mid-Holocene deposits and others only having late Pleistocene deposits offset. I hypothesize that slip is being partitioned between several reactivated Laramide fault surfaces with differing amounts of slip occurring on each fault during large events initiated at depths greater than 10.0 km).

CHAPTER 8

EMMIGRANT/DEEP CREEK/LUCCOCK PARK FAULT

Paradise Valley is bounded on the east by the Emmigrant-Deep Creek-Luccock Park fault system (EDLF). The EDLF lies at the northern apex of the belt I within the Yellowstone tectonic parabola as well as the northeastern extent of the Basin and Range province. Structural controls on segmentation involve the interaction of Neogene-Quaternary faulting patterns with Laramide and Precambrian structures. The eastern limit of the Emmigrant segment coincides with an Early Proterozoic shear zone revealed by bedrock geology (Erslev, 1993) and tomographic studies demonstrating the presence of the Early Proterozoic continental suture zone (O'Neill, 1993; Dueker et al., 2001). The southern boundary of the EDLF terminates into the NW-striking, NE-dipping, Laramide Bismark-Spanish Peaks-Monarch fault and the centrally-located abrupt bend in the fault occurs where the WNW-striking, NNE-dipping, Mill Creek fault intersects the range-bounding normal fault (Wu, 1995). In addition to the presence of 12-m-high Late Pleistocene- to Holocene fault scarps, tectono-geomorphic analyses of faceted spurs, valley height-to-valley floor width ratios, and mountain front sinuosity, reveal rapid Pleistocene-Holocene slip rates of 0.5-1.0 m/ka. However, the distribution of historic seismicity in this part of the northern Intermountain Seismic Belt does not correspond to the tectono-geomorphic and paleoseismological records. Despite the evidence for substantial Holocene surface rupture(s) along the Emigrant fault system and measured

rates of long-term displacement, the area within and proximal to the Paradise Valley has been markedly seismically quite in historic time (Ruleman et al., 2000).

Structural Setting

The Paradise Valley lies at the interface of four major tectonic provinces within southwest Montana: 1) the northeast-trending Early Proterozoic Madison Mylonite Zone (Erslev, 1982); 2) the Laramide foreland province represented by the Beartooth uplift and several southwest-verging reverse faults (Reid et al., 1975; Garihan et al., 1983; Schmidt and Garihan, 1986), ; 3) the northeastern limit of the Basin and Range province (Personius, 1982); and 4) the current upwelling and extensional regime associated with the Yellowstone volcanic field (Pierce and Morgan, 1992; Smith and Braile, 1993) (Figure 23). Cenozoic extensional deformation has apparently utilized, or at least exposed, the structures, particularly faults, related to these prior tectonic episodes. Being within the Laramide foreland province, Neogene extensional faulting is inferred to behave structurally similar to the Madison Range fault system and other large normal faults within the foreland province, with range-front normal faults dipping approximately 60°-70° and projecting to the base of the seismogenic crust (Doser, 1985; Barrientos et al., 1987).

Paradise Valley is bound by the Deep Creek-Luccock Park normal fault system on the east side of the valley. Archean metamorphic rocks and Tertiary intrusive rock are present within the footwall of this fault system. Estimated maximum throw along the Deep Creek fault ranges from 1200 m (Wu, 1995) to 2400 m (Bonini et al., 1972).

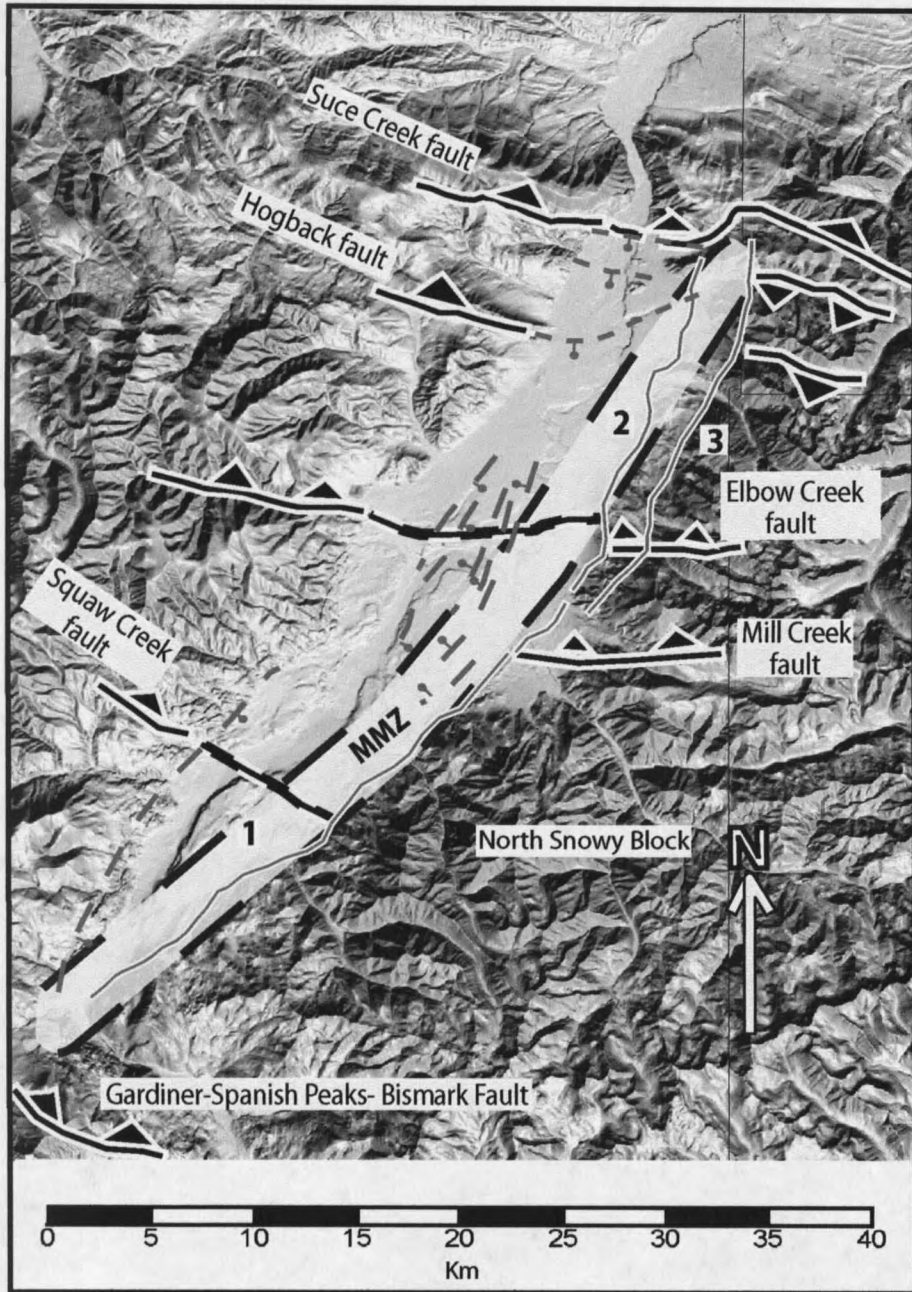


Figure 23. Paradise Valley structural framework. Segments of the EDLF shown as gray and white line and are listed as follows: 1 - Emmigrant, 2 - Deep Creek, and 3 - Luccock Park. Laramide structures shown as black line with teeth on the hanging wall (Wu, 1995). Inferred location of the Early proterozoic shear zone MMZ (Madison mylonite zone) (Erslev, 1993) indicated by white line. Transparent-dashed-Gray lines show Neogene normal faults (Wu, 1995). Gray-white outlined line shows location of mid- to late-Pleistocene surface ruptures (Personius, 1982).

Personius (1982) estimated a maximum throw of 1500 m on the Deep Creek fault. Three-point calculations along the surface trace of the Deep Creek fault suggest dip angles of 70°-90° (Robbins and Erslev, 1986). Interpretation of seismic profiling and gravity modeling suggest fault dip angles around 65° (Wu, 1995).

West-northwest trending Laramide reverse faults have strongly influenced basin depths and geometry of normal faulting (Wu, 1995). The Paradise Valley is bound on the northeast by the Suce Creek fault and on the southwest by the Gardiner-Spanish Peak-Bismark fault. Both of these are steep ($>45^\circ$), northeast-dipping reverse faults. Other Laramide structures that flank the valley are the Squaw Creek fault, Mill Creek fault, Elbow Creek fault, and the Hogback fault, all of which are west-northwest striking, north-northeast dipping reverse faults (Figure 23). The EDLF truncates these Laramide structures. However, patterns of normal faulting and the resulting basin depths are at least partly controlled by these cross-cutting Laramide features. The change in trend of the range-front fault from N40E to N20E and change in basin depths at Mill and Elbow Creek faults are interpreted as first-order structural control by these pre-existing Laramide faults (Wu, 1995).

Quaternary Deposits and Depositional History

Quaternary deposits from the penultimate and most recent glacial periods (Bull Lake and Pinedale), dominate the surficial geology of the Paradise Valley. Clarke (1991) estimates a maximum thickness of Quaternary deposits to be approximately 250 m. During the penultimate and most recent glacial stades, ice extending from the

Yellowstone ice cap flowed north into Paradise Valley (Pierce, 1979). Glacial deposits related to Pinedale glacial maximum are preserved within the valley floor, the Eightmile and Chico moraines (Licciardi et al., 2001). Pierce (1979) constrained the minimum age of 20-25 ka for Pinedale glacial maximum deposits. Sturchio and others (1994) determined Pinedale glacial maximum age ranging between 22.5-30 ka. Cosmogenic ^3He and ^{10}Be ages indicate that the Paradise Valley outlet glacier from the Yellowstone icecap reached its maximum at 16.5 ± 0.4 ^3He ka and $16.2 \pm .3$ ^{10}Be ka (Licciardi et al., 2001). Ice thicknesses near Pray and Emmigrant during the Eightmile advance reached approximately 300 m (Pierce, 1979).

Glacial deposits of mid-Pleistocene Bull Lake advances from the Yellowstone ice sheet were over ridden and largely obliterated by the Pinedale advance. Old deposits, perhaps from glaciers from the Beartooth uplift are preserved within the bench between the Deep Creek and Luccock Park faults and as higher remnant surfaces several kilometers into the footwall of the Deep Creek fault at Mill and Elbow Creeks. Alluvial deposits and terraces within the Paradise Valley almost entirely consist of Pinedale-equivalent deposits. The highest terrace level within the Paradise Valley is correlated with the Eightmile moraines of Pinedale age. This surface is well developed as an outwash surface northward from the Eightmile moraines north. Younger, inset terraces are correlated with Pinedale recession ending around 12-13 years B.P. (Personius, 1982). Therefore, the Paradise Valley has a similar chronology of Quaternary system to that of the Madison Valley, although the southern half is dominated by a large outlet glacier from Yellowstone: 1) glacial aggradation and progradation, 2) subsequent entrenchment

of glacial deposits during interglacials, and 3) aggradation within the dissected channels during the following glacial event.

Morpho-Metric Analyses

Fault Scarps and Offset Deposits

Personius (1982) found three distinct periods of movement along the Deep Creek fault, based upon equivalent maximum scarp height for different sections of the fault with different slope angles indicating age of displacement. He concluded that the most recent surface rupturing activity along the fault occurred 10,000-12,000 years B.P. This was preceded by two prior periods of activity, the Pool Creek scarps of potential 30,000-50,000 years B.P., and the Strong's Ranch scarps of 100,000-200,000 years B.P. This results in an average slip rate since the early Pleistocene of 200-300 m/ Ma, or a 0.2-0.3 mm/yr rate. Personius (1982) also notes that this slip rate is consistent with slip rates determined from uplift of the Short Hills block, 200-300 m/Ma for the mid- to late-Pleistocene. He considered these to be single-event fault scarps, based on the interpretation that the Hebgen Lake fault scarp was a single-event rupture. However, it is likely that single-event surface ruptures tend to have a maximum range from 2-3 m of surface offset within this region (i.e., based on moment magnitude parameters, a fault scarp with surface offset of 6.0 m will likely have more than one event preserved within the scarp, especially for a segment only 20 km in length) (Haller et al., 2000; Pierce et al, 2000; Schwartz et al, 2000; and Van der Woerd, 2000).

The Deep Creek fault offsets Pinedale and Holocene deposits south of Pool Creek beginning with the Barney Creek scarp (Personius, 1982). Based on an alluvial fan surface offset of approximately 5.1 m at the base of Dome Mountain with a maximum slope angle of 36.5° , two events have occurred on this post-Pinedale fan (Personius, 1982). Entrenched remnant fan deposits proximal to the bedrock-colluvium junction at the mouth of this drainage be could be latest Pleistocene, related to the first phase of deposition following the retreat of the Yellowstone Valley glacier approximately 12,000 years B.P. (Personius, 1982), while the fan surface that is offset 5.1 m could be an early to mid Holocene debris flow fan.

Based on the fault scarps at Pool Creek, Personius (1982) found another period of faulting preserved on Bull Lake moraines, but not found on Pinedale moraines. Surface offsets on Bull Lake deposits (>130 ka) range from 3.8-6.0 m, with maximum slope angles ranging from 27° - 30° . Personius (1982) found highly-dissected range-front colluvium offset 3.5-7.9 m with maximum slope angles of 23° - 26° at the far northern end of the fault zone along the Strong's Ranch section, which he considers to be an even older event. However, slope angles are not indicative of fault scarps older than 100,000 years. Pierce (1979) found that Bull Lake moraines were offset roughly five times more than Pinedale moraines in the Pine Creek area.

Relative Tectonic Activity Class Designation Chart

V_f ratio values for the Deep Creek-Luccock Park fault system range from 0.02-0.27 (Figure 24 and Appendix B). From Mill Creek south to Yankee Jim Canyon, the ratio values range from 0.04-0.23, of which nine out of twelve of the drainages have values between 0.04-0.09, with four drainages having values of 0.04. North of Mill Creek along the Deep Creek fault to Pine Creek, V_f ratio values remain between 0.03-0.09 with a mean of 0.06. The Luccock Park fault has V_f ratio values ranging from 0.02-0.07 with a mean of 0.05.

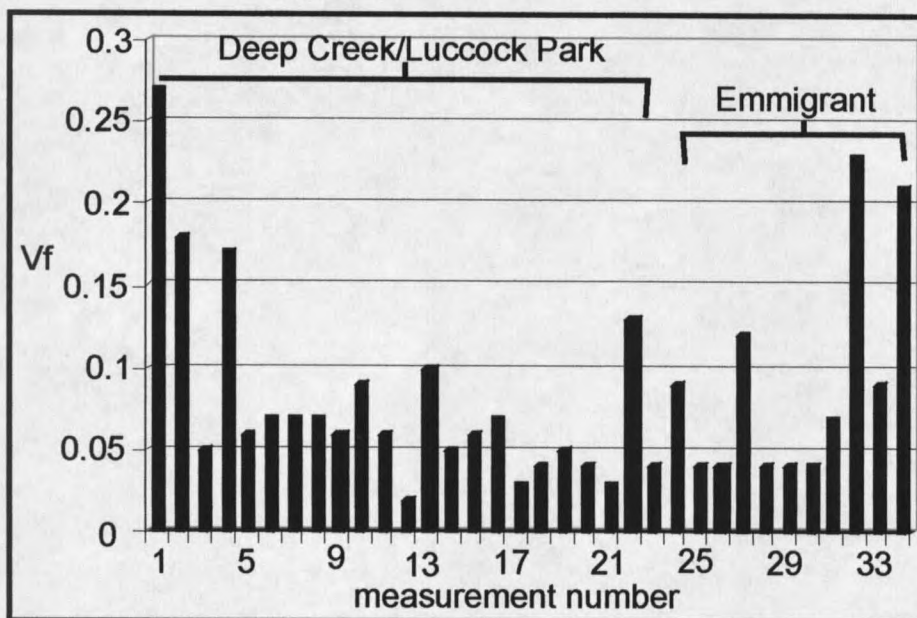


Figure 24. V_f values for the Emmigrant/Deep Creek/Luccock Park fault. See Appendix B for data.

Anomalous high V_f ratio values exist north of Pine Creek, ranging from 0.17-0.27, and to the south as follows: at South Fork MacDonald Creek along the Deep Creek

fault (0.1); South Fork Elbow Creek along the Deep Creek fault (0.13); at Emmigrant Gulch (0.12); at Sheep Creek (0.23); and at Yankee Jim Canyon (0.21). The increase in V_f values north of Pine Creek correlates with decreased Quaternary tectonic activity at the northern end of the fault zone. The increase in V_f at South Fork MacDonald Creek corresponds to the southern end of the Short Hills Block and is the result of tectonic activity tapering towards the southern end of Short Hills Block. At South Fork Elbow Creek, the increase in V_f values corresponds to change in fault and fault scarp trends and potential segment boundaries. Laramide structures preserved within the North Snowy block could be the cause of segmentation here. At Emmigrant Gulch and Sheep Creek, the higher V_f values are a result of glacial scouring and erosion, broadening the floor of the valley and creating anomalous values. The Yankee Jim Canyon value corresponds to tapering of tectonic activity at the south end of the EDLF.

M_s values for the Deep Creek-Luccock Park fault system are low with a sinuosity of 1.11. The mapped fault traces and fault scarps occur proximal to the mountain-piedmont junction with a very sharp junction. Little embayment has occurred along both the Deep Creek and Luccock Park faults. Using the RTAC (Table 1), both fault systems show V_f and M_s values indicative of long-term average slip rates of > 0.1 mm/yr. However, knowing that these indices are affected over periods greater than 100 ka, these rates may be considered consistent with the 0.2-0.3 mm/yr determined by Personius (1982).

Basal Facet Reconnaissance Technique

Basal triangular, facets along the Deep Creek-Luccock Park fault system have heights ranging from 122-256 m (Figure 25 and Appendix C). The average basal facet height is 147 m. Most of the range front exhibits a steep bedrock-colluvium contact. However, the facets along the Emmigrant section of the fault are highly irregular in shape and resemble uplifted and scoured bedrock glacial features more than “triangular” faceted spurs.

Based on the basal facet slip rate estimation equation of dePolo (1998), the maximum basal facet height of 256 m yields a slip rate of 0.53 mm/yr. Based on the lowest basal facet height, the slip rate would be 0.23 mm/yr. A mean basal facet height of 147 m yields an estimated slip rate of 0.27 mm/yr. All of these estimated slip rates fall within the range of slip rates determined by the uplifted Short Hills Block, as well as fault scarp morphometry (Personius, 1982) and the RTAC.

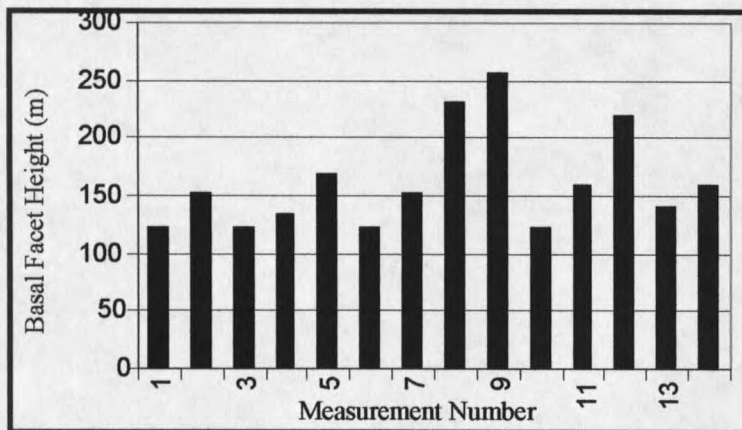


Figure 25. Basal facet heights for the Emmigrant/Deep Creek/Luccock Park fault (EDLF). Measurements recorded south (1) to north (14). Change in facet height occurs proximal to Mill Creek (between measurements 7 and 8), where the northern segment boundary for the Emmigrant segment and the southern boundary for the Deep Creek/Luccock Park segments intersect. See Appendix C for data.

CHAPTER 9

THE BRIDGER RANGE FAULT SYSTEM

The Bridger Range fault is characterized by a lack of identifiable late Pleistocene to Holocene fault scarps, suggesting low slip rates (< 0.1 mm/yr) and long recurrence intervals (> 10 ky). Morpho-metric analyses infer slip rates an order of magnitude greater than those inferred from the surficial geology and absence of fault scarps (> 0.1 mm/yr), suggesting a prior period of greater tectonic activity rates older than 130 ka. Basal facet reconnaissance infers slip rates consistent with low late Pleistocene (< 130 ka) slip rates derived from the absence of fault scarps.

Although no fault scarps have been identified, six segments are identified and analyzed for seismic hazard analyses. Structural controls on segmentation are consistent with those observed along the MRF and EDLF, and are defined by major bends in the range front occurring at the intersection of Laramide structures exposed in the footwall along with embayments and salients.

Structural Setting

The Bridger Range is located within a region that overlaps four major tectonic provinces: 1) the Middle Proterozoic Belt Basin; 2) the Sevier fold and thrust belt; 3) the Laramide foreland province of basement-involved deformation; and 4) the Basin and Range province (Lageson, 1989). Figure 26 depicts these structural relationships. The

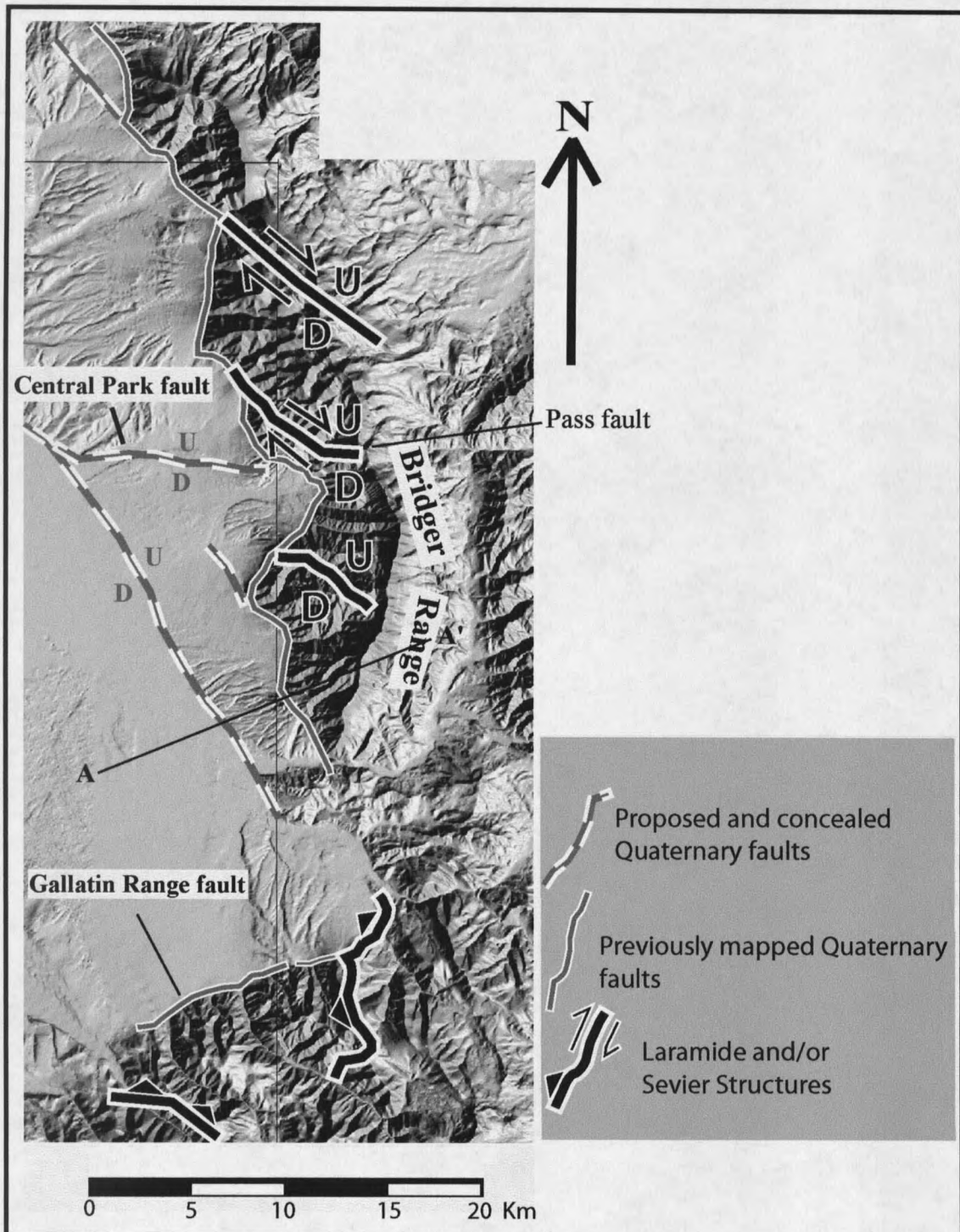


Figure 26. Bridger Range/Gallatin Valley structural framework. Quaternary faults are modified from Stickney and others (2000). Laramide structures compiled from McMannis (1952), Miller and Lageson (1993), Wilson and Elliott (1997), Skipp and others (1999).

Pass fault coincides with the southern structural margin of the Belt Basin. Between 1450 MA-850 Ma, this fault was active as a normal fault with down-to-the-north displacement (McMannis, 1952). Evidence for activity during the Paleozoic is supported by sedimentary facies changes proximal to the fault (Lageson, 1989). During the Paleocene, coincident with the Sevier fold and thrust belt, the Pass fault and other east-west trending faults were reactivated as high-angle, oblique-slip, lateral thrust ramps along the transverse fault zone of the Helena salient (Lageson, 1989). Contractional deformation culminated with the basement-cored Bridger anticline during the Laramide orogeny in latest Paleocene to earliest Eocene.

The Bridger Range marks the eastern extent of the Basin and Range province at the latitude 46°N. Quaternary faulting appears to be discontinuous and irregular along the range front. The range front is characterized by NW-trending sections with well expressed basal facets and accumulation of Quaternary sediments proximal to the range front. Northeast-trending sections of the range front have less facet development and have Tertiary strata juxtaposed with the range front, suggesting minimal mid- to late-Quaternary displacement. This is supported by the presence of diffuse, eroded facets, dissection of older Tertiary basin fill remnants, and the progradation of younger Quaternary deposits away from the range front into the valley.

Northwest-trending sections of the range front show more obvious evidence of Quaternary tectonic activity. Small alluvial cones and fans emanating from the range front have range-front proximal deposition and the footwall is marked by well-expressed, but rilled, basal facets. The northwest-trending sections of the range front parallel

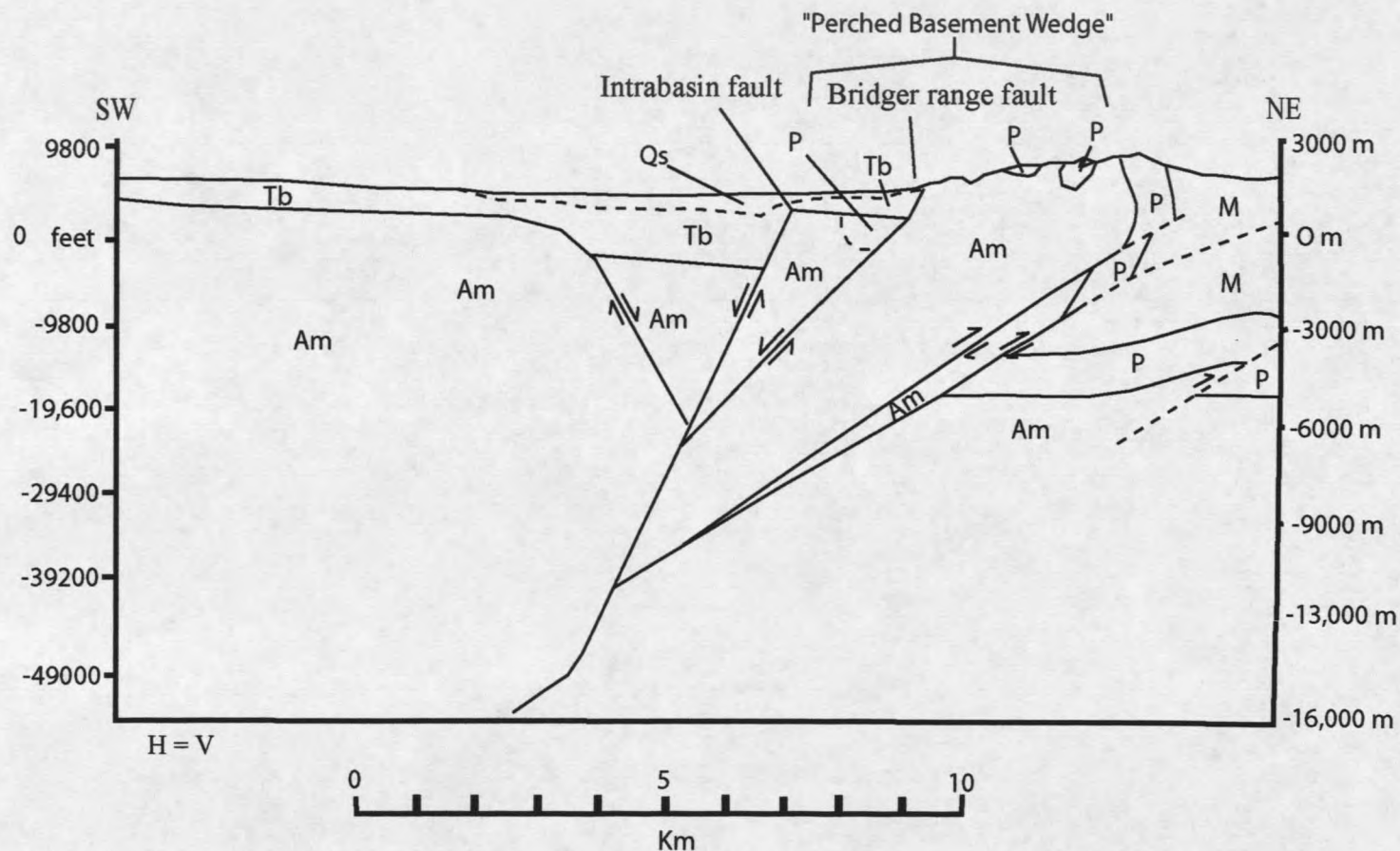


Figure 27. Cross Section of the southern Bridger Range showing relationships between Laramide compressional structures and Neogene normal faults. Cross section line A-A' from Figure 26. Rock units are as follows: Am - Archean metamorphic rocks, P - Paleozoic sedimentary rocks, M - Mesozoic sedimentary rocks, Tb - Tertiary basin fill deposits, Qs - Quaternary sediment. Modified from Lageson (1989).

WNW-trending oblique thrust ramps exposed in the footwall (Skipp et al., 1999). This suggests a strong structural control on the faulting patterns within the Bridger Range fault system.

A topographic lineament that trends north-northwest and separates the Bridger Range piedmont and the Gallatin Valley, may suggest the presence of a piedmont fault approximately 3 km west of the range front. This could represent the main Bridger detachment (Figure 27). This escarpment is at the boundary between tributary alluvial deposits from the Bridger Range and the mainstream alluvial deposits of the East Gallatin and Gallatin Rivers. Whether this scarp represents faulting in addition to undercutting by the Gallatin stream system is uncertain. If this intra-basin fault is the main Bridger detachment that the range-front fault(s) merge into at depth, mid- to late-Pleistocene displacement could be expressed by the Gallatin Valley fluvial network. The East Gallatin River flows parallel to this lineament and the Gallatin River flows north directly into it before turning west at the Horseshoe Hills. Mid- to late-Pleistocene displacement along the intra-basin fault would allow the piedmont between the range-front faults and the basin fault to be a pediment surface, and explain the dissected Tertiary strata and deposition of Quaternary deposits onto the downthrown side, the Gallatin Valley.

Quaternary Deposits and Depositional History

Quaternary deposition proximal to the Bridger Range front has been limited by low subsidence rates based on calculated tectonic activity rates. A detailed study of Quaternary deposits and their correlation to the previous glacial maximums, Bull Lake

and Pinedale, was not attempted for this study. Sections of the range front with north-northeast trends have older dissected Tertiary deposits proximal to the range front. Apparently, Quaternary deposition has prograded further out into the valley to reach the base level of the Gallatin and East Gallatin Rivers, leaving older remnant surfaces dissected. Along these reaches of the range front, mid- to late-Pleistocene deposition was apparently confined to incised channels emanating from larger drainages, along with smaller localized debris flows, landslides, and colluvial slopes. This relationship can be seen along the southern end of the range between Walton and Lyman Creeks. Displacement along this section of the fault system has not been enough to allow deposition of Quaternary deposits on top of Tertiary deposits. Instead, Quaternary deposits have become entrenched into the Tertiary Bozeman Group and prograde further out into the valley.

Quaternary deposits exist along northwest-trending sections of the range front in the form of alluvial fans that display range-front proximal deposition. This suggests that activity along these sections of the fault system have had more recent activity than the north-northeast trending sections. Tectonic activity has downdropped older deposits and created space for younger deposits to aggrade above. These relationships exist between Limestone and Smith Creeks and Bostwick and Middle Cottonwood Creeks. Detailed surficial geologic maps for the entire range front have not been completed, so further investigation is required to fully understand these depositional relationships and the tectonic and climatic signals.

Morpho-metric Analyses

V_f and M_s values for the Bridger Range show similarities to those calculated for the Madison and Absaroka-Beartooth Range. Three V_f measurements were made per drainage along the Bridger Range in order to distinguish any anomalies in the data set caused by lithologic and erosional differences down channel. All values measured for the Bridger Range are between 0.093 and 0.019 (Figure 28), which is correlative to an inferred long-term slip rate of > 0.1 mm/yr based on the Relative Tectonic Activity Chart (Bull, 1987b). V_f values do show a pattern of increasing towards potential segment boundaries, or sections that once were boundaries.

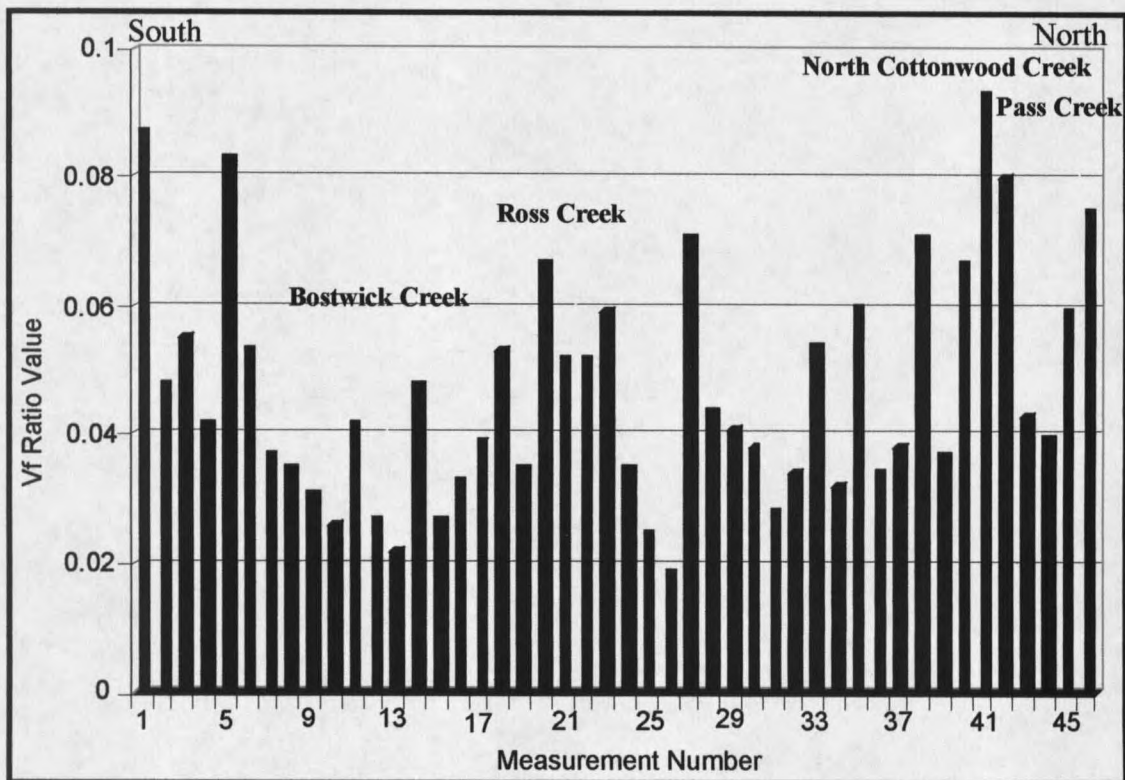


Figure 28. V_f ratio values for the Bridger Range fault. Measurements listed south (1) to North (46) and creeks shown at approximate centers and ends of proposed section boundaries. See Appendix B for data.

From Walton Creek north, values jump down from 0.063-0.059 to 0.026-0.042, then increase back up to 0.057 north to Ross Creek the values taper. North from Ross Creek the values range from 0.026-0.08, and are in concordance with the Relative Tectonic Activity Chart (Table 1) values for an inferred slip rate of magnitude > 0.1 mm/yr.

M_s values for the Bridger Range fault range from 1.11-1.13, also correlating to an inferred slip rate of 0.5 mm/yr from the Relative Tectonic Activity Class Designation Chart (Table 1). The Bridger Range front is characterized by both sharp and diffuse bedrock-colluvial junctions, with deeply incised drainages trending generally orthogonal to the range front. However, the lack of evidence for fault scarps and actual location of the fault impairs the M_s measurement.

In one scenario, the inferred intra-basin fault approximately 3-10 km west of the range front would be the main fault. Erosion and embayment would be utilizing pre-existing weaknesses within the pre-existing footwall structures in order to make a highly sinuous range front. This seems improbable because the range front is marked by sharp colluvial/bedrock junctions, but with rilled, older facets, probably generated during previous periods of greater tectonic activity. The range front morphology quantified with the Relative Tectonic Activity Chart (Table 1) appears to be a signal for long-term average slip rates on the order of > 200 ka. Fault scarps are absent along the length of the range front, but remnant facets are present with cataclasites preserved in hoodoos north of Bostwick Creek and elsewhere, suggesting a range front that is relatively freshly exposed.

Based upon the regional faceted spur reconnaissance technique (dePolo, 1998), the Bridger Range is assigned a type 2 range front, with an inferred slip rate of 0.01 mm/yr. This is based on the fact that remnant, older, rilled facets are present along the majority of the range front, but fault scarps are not present. This infers that slip rates are high enough (0.01 mm/yr) to develop a range front with facets, but recurrence intervals are long enough that fault scarps are eroded and facets are not as fresh in appearance. No evidence of late Pleistocene or younger (< 130 ka) offset is present.

CHAPTER 10

SEISMIC HAZARD ANALYSES

The region within and proximal to Montana is associated with many Quaternary faults and high frequency of historic seismicity. Although this study does attempt to quantify seismic hazards along the Madison Range fault (MRF), Emmigrant/Deep Creek/Luccock Park fault (EDLF), and the Bridger Range fault (BRF), the inherent seismic risk to society is not determined. Through future studies, hopefully utilizing the techniques used herein, risk to communities proximal to large faults can be determined. Furthermore, seismic hazards delineated through these analyses are conservative measurements of maximum earthquakes, recurrence intervals, and slip rates, which does not discredit the potential for much larger earthquakes ($> M_w 7.2$) to occur. Investigations of historic earthquakes (e.g., 1992 Landers, CA and 1959 Hebgen Lake, MT earthquakes) show that what is expressed by the surficial geology prior to a large earthquake may not predict or forecast what will occur during the next earthquake. For example, the chaotic rupture patterns and segmentation of the 1992 Landers, CA earthquake (Lazarte et al., 1994) and greater basin subsidence than the fault scarps revealed during the 1959 Hebgen Lake, MT earthquake (Myers and Hamilton, 1964) would probably not have been predicted by paleoseismologic and fault scarp investigations.

Madison Range Fault

Seismicity

The historic seismic activity on and near the Madison fault system indicates seismic quiescence within the central and northern segments of the fault and intense microseismicity along the southern reaches proximal to the Yellowstone volcanic field (Stickney, 1995). The few fault-plane solutions determined for the region are generally consistent with NE-SW extension (Stickney, 1995). The southern Madison is characterized by a wide array of fault-plane solutions. Few fault-plane solutions directly lie on the Madison fault and seismicity is concentrated along smaller faults (Stickney, 1995).

Seismicity within the Madison Range and Madison Valley is marked by intense microseismicity within the Madison Canyon/Missouri Flats segment showing a wide array of focal mechanisms, with relative seismic quiescence north of latitude 45.0° (Stickney, 1995; MBMG, 2001, unpublished data). It has been noted that major faults with Holocene offset and contemporary seismic quiescence within the Intermountain Seismic Belt are common (Smith and Sbar, 1974; Smith and Arabasz, 1991). Stickney (1995) suggests that this could be the result of tensional stresses being completely or almost completely alleviated along large faults during large earthquakes ($M_w > 7.0$), but regions without large faults detaching the seismogenic crust must maintain lesser magnitude and more frequent seismicity in order to release tectonic strain. However, the southern MRF is the location of concentrated low magnitude seismicity and large normal

faults, suggesting a concentration of low magnitude seismicity proximal to large normal faults, especially following the 1959 Hebgen Lake earthquake which was also accompanied by movement on the southern MRF.

The Centennial Tectonic Belt trends west from the Yellowstone caldera through the Hebgen Lake basin and the Centennial Valley. Savage and others (1993) found a total of 8.0 mm/yr of extension in a N15°E direction within the Hebgen Lake basin over the period of 1973-1987. This direction is coincident with regional Basin and Range extension (Stickney, 1995). The greatest concentration of seismicity west of the Hebgen Lake basin occurs within the Missouri Flats area. The West Fork of the Madison River is along trend of a NE-trending zone of epicenters. Focal mechanisms show a steeply-dipping, right-lateral fault that parallels the West Fork of the Madison River (Stickney, 1995). North of Wade Lake, Stickney (1995) found seven events with nodal planes trending perpendicular to the Madison fault. These events suggest oblique slip along intrabasin faults that strike orthogonal to the Madison fault. Focal mechanisms for two events (20 October 1990 and 2 February 1995) show left-lateral slip along the Madison fault (Stickney, 1995). Fault-plane solutions also show faults antithetic to the Madison fault within the Missouri Flats region. The inferred Quaternary faults on Plate 1 are well established by surficial geology and consistent with fault-plane solutions.

Oblique-reverse and reverse dip-slip focal mechanisms on faults in the southern Madison region have also been determined (Stickney, 1995). On 4 January 1992, an oblique-reverse event occurred within the footwall of the Madison fault (Stickney, 1995). This event was along trend of the Madison fault, but had a P-axis trending the direction

that most of the regions T-axes trend, NNE (Stickney, 1995). This event may be related to shifting of intrabasin fault-bound blocks, creating a wide array of possible focal mechanisms. Fault-plane solutions have also been located on a fault antithetic to the Madison fault along the west side of the valley as far north as Ennis, MT. Fault-plane solutions near Pony, MT show strike-slip motion consistent with WNW-ESE directed extension (Stickney, 1995). Low magnitude and infrequent historic seismicity defines the region east of the Madison Valley and north of the Centennial Tectonic Belt (MBMG, 2001, unpublished data).

Segmentation and Maximum Earthquakes

Crone and Haller (1991) give four characteristics of segment boundaries along Basin and Range province normal faults: 1) major en echelon offsets or pronounced gaps in the continuity of fault scarps; 2) distinct, persistent, along-strike changes in fault-scarp morphology that indicate different ages of surface faulting events; 3) major salients in the range front; and 4) transverse bedrock ridges with less cumulative throw than other places along the fault zone. Based upon these criteria, the Madison Range fault can be broken into five distinct structural segments: 1) the Madison Canyon/Missouri Flats segment, 2) the Taylor-Hilgard segment, 3) the Indian Creek section, 4) the Cameron segment, 5) the Spanish Peaks segment (Figure 29), modified from Schneider (1990).

The Madison Canyon/Missouri Flats segment is considered a separate segment from the Taylor-Hilgard segment because it may have experienced displacement during the 1959 Hebgen Lake event (Myers and Hamilton, 1964), where as the Taylor- Hilgard

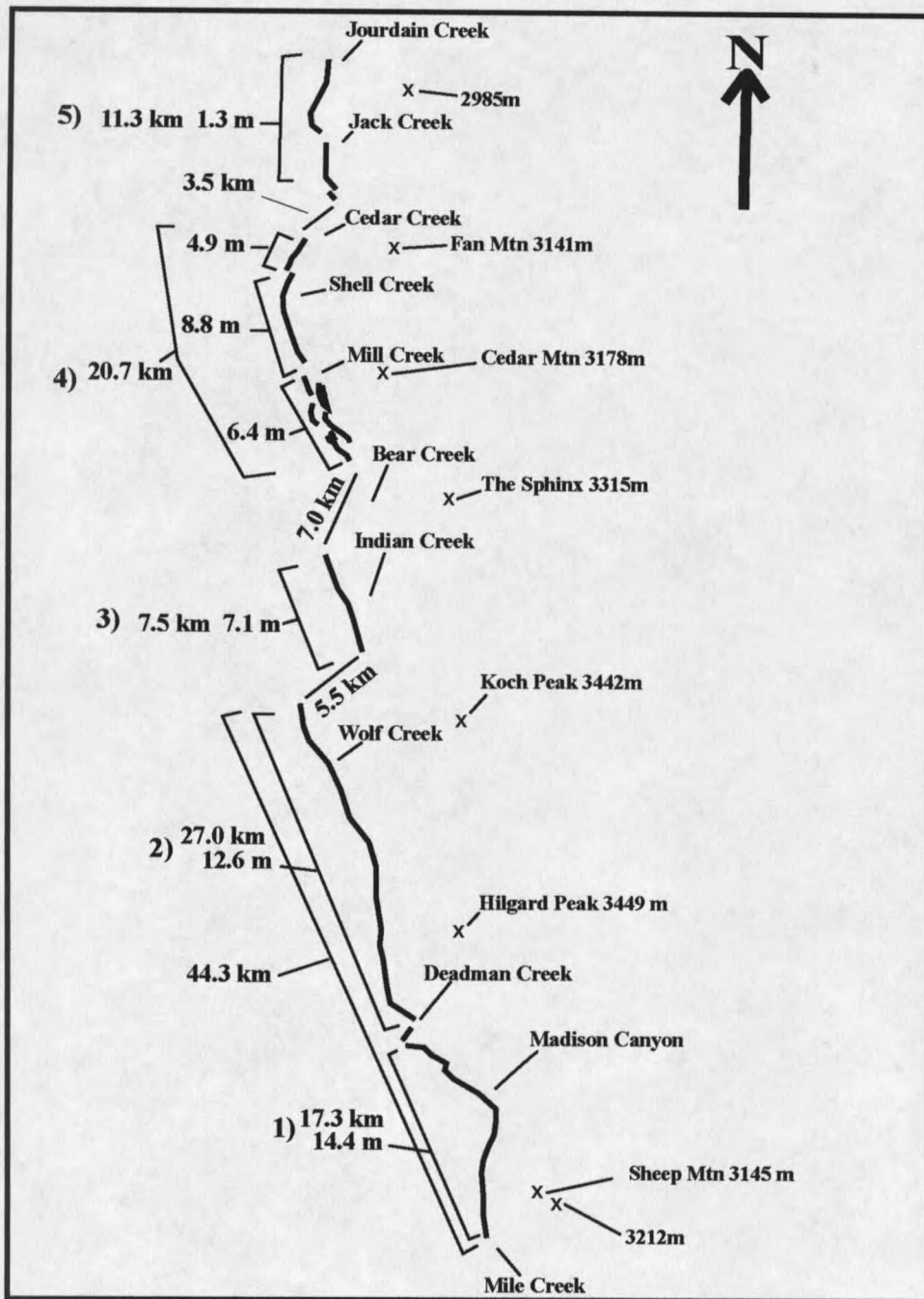


Figure 29. Madison Range fault late Pleistocene to mid Holocene (< 130 ka) surface ruptures and sections. Maximum surface offset for sections of the fault given in meters and length of segments in km. Structural discontinuities between sections shown. Sections shown as 1) Madison Canyon-Missouri Flats, 2) Taylor-Hilgard, 3) Indian Creek, 4) Cameron, and 5) Spanish Peaks.

segment was not displaced. However, the 1959 displacement could have been coseismic subsidence ("shakedown") within the basin (Wallace, 1979, 1980), as previously discussed. Both the Taylor-Hilgard and Madison Canyon/Missouri Flats segments have offset early- to mid-Holocene and late Pleistocene deposits. In this seismic hazard analysis, these segments were considered to behave independently and the Taylor-Hilgard and Madison Canyon-Missouri Flats sections were considered to rupture coseismically. Refer to Table 3 for MRF moment magnitude parameters and the maximum credible earthquake (MCE).

By using a maximum focal depth of 14 km, based on the 1959 Hebgen Lake, MT earthquake and the 1983 Borah Peak, ID earthquake (Doser, 1985; Barrientos et al., 1987), moment magnitude for individual segments along the MRF were calculated following Hanks and Kanamori (1979) (Table 3). Because both the Taylor-Hilgard and Madison Canyon/Missouri Flats segments appear to have at least three to four post-Pinedale events that also have offset early- to mid-Holocene deposits, it is necessary to consider the seismic potential if these segments ruptured together. The small salient south of Deadman Creek that could possibly split the Taylor-Hilgard and Madison Canyon-Missouri Flats sections has late Pleistocene rupture and does not appear to have a different late Pleistocene displacement history than the rest of the segment from Dear Creek to Deadman Creek. This justifies the possible coseismic offset of both segments during a large earthquake ($M_w > 7.0$).

Table 3. Madison Range fault (MRF) moment magnitude (M_w) parameters and maximum earthquake. Calculations made from equations for moment magnitude Hanks and Kanamori (1979).

Segment	Surface Rupture Width W (km)	Surface Offset (m)	Down-Dip Rupture L (km)	M_w
Madison Canyon/Missouri Flats	17.3	2.0	16.0	6.8
Madison Canyon/Missouri Flats	17.3	3.0	16.0	6.9
Taylor-Hilgard	27.0	2.0	16.0	6.9
Taylor-Hilgard	27.0	3.0	16.0	7.0
Madison Canyon/Taylor-Hilgard	44.3	2.0	16.0	7.1
Madison Canyon/Taylor-Hilgard	44.3	3.0	16.0	7.2
Indian Creek	7.5	2.0	16.0	6.5
Indian Creek	7.5	3.0	16.0	6.6
Cameron	20.7	2.0	16.0	6.8
Cameron	20.7	3.0	16.0	6.9
Spanish Peaks	11.3	2.0	16.0	6.7
Spanish Peaks	11.3	3.0	16.0	6.8

The Indian Creek segment is situated north of a structural discontinuity approximately 5.5 km wide in map view. No surficial faulting has occurred along the northeast-trending range front connecting the Taylor-Hilgard and Indian Creek segments. The Indian Creek and Cameron segments are separated by a structural discontinuity approximately 7.0 km wide in map view (Figure 29, Plate 2). This structural discontinuity is marked by a bedrock salient and lack of fault scarps. Large lateral ramps are mapped within the footwall proximal to this structural discontinuity (Plate 2). To the north, the Cameron segment is approximately 20.7 km in length, with a footwall exposing

imbricate thrusts and folds in the Phanerozoic section. Late Pleistocene rupture has occurred along this segment and late Pleistocene deposits are offset as much as 8.8 m. To the south of Mill Creek, fault scarps are discontinuous and fill gaps between mapped bedrock normal faults. These scarps are thought to rupture coseismically with a large (M_w 6.5) event as synthetic, shallow (< 6 km depth) normal faults that merge at depth.

The Spanish Peak segment is defined by fault scarps on late Pleistocene deposits with surface offsets of 1.3 m, indicating longer recurrence intervals and slower slip rates than the rest of the fault. The footwall geology of this segment consists of Archean metamorphic rocks thrust over the Phanerozoic section and is marked by irregular steps in the range front controlled by the position of pre-existing thrust faults.

Based upon these segment boundaries, the maximum earthquake for the Madison Range fault ranges from M_w 6.5 to 7.1 (Table 3). However, if the right-stepping pattern between the Taylor-Hilgard and Indian Creek segments is controlled by a lateral ramp in the Hilgard thrust system, these two segments might merge at depth (> 5 km) and rupture coseismically during a large event. In this case, M_w 7.0 and 7.2 earthquakes would be necessary to create a surface offset of 2.0 m and 3.0 m, respectively, with a total rupture length of 51.8 km (Hanks and Kanamori, 1979). This would lead to a maximum earthquake for the Madison fault of M_w 7.2.

Slip Rate and Recurrence Intervals

Late Pleistocene deposits are offset along nearly all the Madison Range fault (Figure 29). Along the Madison Canyon-Missouri Flats segment, surface offsets on

Pinedale-aged deposits (12-30 ka) are as great as 14.4 m. Based on offset Bull lake moraines, Matthieson (1983) assigns a late Pleistocene slip rate of 0.3-0.6 mm/yr, but a mid Pleistocene slip rate of only 0.01-0.02 mm/yr, showing a definite increase in slip rate during the late Pleistocene. Weinheimer (1979) estimated a Quaternary slip rate of 0.5 mm/yr based on displacement and tilting of the Huckleberry Ridge Tuff (HRT) (2.1 Ma). Shaw estimated a displacement of 460 m on the HRT, suggesting a long-term average slip rate of 0.23 mm/yr. Seismic moment rates and return periods determined for the southern Madison are 2.2 N-m/yr and $4,500 \pm 800$ years (Doser, 1984). Morpho-metric analyses (i.e., RTAC and basal facet analyses) yield slip rates ranging from 0.18 mm/yr to 0.61 mm/yr for the MRF. Based upon morpho-metric analyses, all of the Madison fault segments have a slip rate of > 0.1 mm/yr.

Recurrence intervals for the southern Madison fault range from $4,500 \pm 800$ years based on moment rates (Doser, 1984). Using diffusion modeling, Lundstrom (1986) determined a late Holocene most recent event of 4,000 years. Single-event fault scarp profiles plotted on Bucknam-Anderson scarp height versus maximum slope angle show that all of the fault scarps are within 1-10 ka for the most recent event. The amount of surface offset of late Pleistocene deposits decreases to the north, from 12.6-14.4 m along the Taylor-Hilgard and Madison Canyon/Missouri Flats segments, to 7.1 m along the Indian Creek segment, to 8.8 m along the Cameron segment, to 1.3 m along the Spanish Peaks segment. The overall tectono-morphic signature of the range front shows a slip rate of > 0.1 mm/yr. However, the grouping and migration of surface faulting events

since the late Pleistocene along this 100 km long fault zone is evident in the differing amounts of surface offset along segments.

Because single-event displacements appear similar, recurrence intervals must decrease from north-to-south in order to explain the greater amount of displacement on late Pleistocene deposits to the south. More offset events have occurred along the Taylor-Hilgard and Madison Canyon/Missouri Flats segments than to the north to explain the differing late Pleistocene offsets of 12.6-14.4 m for the southern half of the fault, compared to the 1.3-8.8 m of surface offset north of Bear Creek to Jourdain Creek. If recurrence intervals are on the order of 1-10 ka and the most recent event was 1-10 ka ago, then the southern segments are in the later part of a seismic cycle. However, some have concluded that slip rates along this fault have varied an order of magnitude from the mid- to late-Pleistocene (0.01 to 0.6 mm/yr, respectively) (Matthieson, 1983), then this must be considered for the seismic hazard analyses. Based upon mid- to late-Pleistocene variations in slip rate, the long-term average slip rates determined through the RTAC and basal facet analysis provide a more generalized, but likely accurate slip rate.

Emmigrant/Deep Creek/Luccock Park fault

Seismicity

Ruleman and Lageson (2000) refer to the Paradise Valley as a potential "seismic gap" within south-central Montana. Historic seismicity is absent within the Paradise

Valley, but continues to the north, south and west of the valley. Tectono-geomorphic analyses of faceted spurs, valley height-to-valley floor width ratios, and mountain front sinuosity, reveal rapid Pleistocene-Holocene slip rates of 0.5-1.0 m/ka. However, the tectono-geomorphic and paleoseismological records do not correlate to the lack of historic seismicity in this part of the northern Intermountain Seismic Belt.

Despite the evidence for substantial Holocene surface rupture(s) along the Emigrant fault system and measured rates of long-term displacement, the area within and proximal to the Paradise Valley has been markedly seismically quiet in historic time (MBMG, 2000, unpublished data). The accurate location of over 16,000 earthquakes in western Montana (Stickney, 1995) clearly shows a lack of seismicity in the Paradise Valley area, forming a mirror image to the seismic gap along the Teton normal fault south of Yellowstone (Ruleman and Lageson, 2000). Extensional strain along the Emigrant fault system may be transferred to smaller normal faults within the seismically active Norris-Mammoth corridor to the south, just as the Teton fault appears to splay into several seismically active, but smaller normal faults to the north (Pierce and Morgan, 1992).

It is possible that large-magnitude earthquakes on the Emigrant and Teton faults may occur as a "catch-up" process with respect to these smaller faults. Therefore, a regional seismo-tectonic model should incorporate major locked normal faults north and south of Yellowstone, with transitional zones of distributed seismicity between them and the Yellowstone (Lava Creek) caldera. This pattern may be induced by the Yellowstone caldera, where earthquake foci are restricted to the upper seismogenic crust (Smith and

Braile, 1993). Proximal to and within the Yellowstone caldera, stress is alleviated more rapidly, but outside the thermal effects of the caldera, stress accumulates on major locked normal faults where the brittle-ductile transition is deeper (e.g., Teton fault, EDLF, MRF).

Segmentation and Maximum Earthquakes

In order to provide a seismic hazard assessment for the Paradise Valley, segmentation is considered along the three reaches of the fault: 1) the Emmigrant fault from Mill Creek to Yankee Jim Canyon, 2) the range-bounding Luccock Park fault from Elbow Creek to Pool Creek, and 3) the Deep Creek fault from Mill Creek to Pool Creek (Table 4). These three potential segments are considered to rupture independently. For a maximum earthquake, the EDLF is considered to rupture from Yankee Jim Canyon north to Pool Creek. Although the EDLF only shows late Pleistocene-Holocene rupture from Yankee Jim Canyon to Barney Creek, the potential of linkage with older displacements to the north between Strong's Ranch and Pool Creek is not improbable. However, mid- to late-Pleistocene rupture patterns (Personius, 1982) and structural controls suggest that the fault zone is segmented at Mill Creek, arguing against a maximum earthquake rupturing the entire fault zone from Yankee Jim Canyon to Pool Creek. Based on these constraints (e.g., segment lengths, rupture depths of 14 km, and displacement per event), the maximum earthquake is M_w 6.9.

Table 4. Emmigrant/Deep Creek/Luccock Park fault (EDLF) moment magnitude (M_w) parameters and maximum earthquake. Equations for moment magnitude from Hanks and Kanamori (1979).

Segment	Surface Rupture Length (km)	Surface Offset (m)	Down-Dip Rupture W(km)	M_w
Yankee Jim - Mill Creek	20.6	2.0	16.0	6.8
Yankee Jim - Mill Creek	20.6	3.0	16.0	6.9
Deep Creek fault	19.0	2.0	16.0	6.8
Deep Creek fault	19.0	3.0	16.0	6.9
Luccock Park fault	23.0	2.0	16.0	6.9
Luccock Park fault	23.0	3.0	16.0	7.0

Tectonic Activity Rates and Inferred Slip Rates

Based on Personius' (1982) fault scarp morphology data and uplift along the Short Hills Block, a slip rate of 200-300 mm/yr was assigned to the EDLF system. As determined by the RTAC, a long-term average slip rate is assigned to be > 0.1 m/ka for the entire fault zone. The faceted spur slip rate estimation analyses led to a slip rate of 0.53 mm/yr.

Recurrence Intervals

Personius (1982) suggests three periods of tectonic activity along the Deep Creek-Luccock Park fault system. He defines three periods of seismic activity post-Pinedale (<12 ka), pre-Pinedale (30-50 ka), and a possible grouping of events between 100-200 ka.

This suggests clustering of events within a probable 10 ka period followed by quiescence of up to 50 ky for the late Pleistocene, with recurrence intervals between these clustered events decreasing over the late Pleistocene from 50 ka for the early-mid late Pleistocene to approximately 20 ka for the latest late Pleistocene (30,000-10,000 years B.P.). In addition to the fault scarp regression analyses (Personius, 1982), the morphology of the fault scarps along the EDLF of all ages show little to no sign of multiple events and are characterized by consistent slopes with surface offsets indicative of two to three events, further suggesting this clustering followed by quiescence model. However, trench data must be attained in order to constrain timing of all faulting events along the Deep Creek/Luccock Park fault system.

Bridger Range Fault

Seismicity

Stickney (1995) reports that seismicity from 1982-1995 has been concentrated along two seismic source zones southwest and northwest of Bozeman. Other than these zones, the Bozeman area experiences less seismicity than to the north and south. Stickney (1995) found a zone of seismicity 9 km south of Gallatin Gateway. Fault-plane solutions reveal a NNW-trending, steeply SW-dipping fault plane with primarily normal dip-slip motion and a small component of right-lateral slip. The other source zone is near Churchill, an inferred intra-basin left-lateral strike-slip fault.

Segmentation and Maximum Earthquakes

The Bridger Range is characterized by a highly irregular, west-dipping range-front normal fault. Rupture segmentation may be controlled by pre-existing Laramide structures exposed within the footwall of the fault system. As noted in Zhang et al. (1999), an analyses of seven well-documented historical Basin and Range province earthquakes showed that the sizes of cross-cutting structural discontinuities vary the ability to terminate rupture. However, since no evidence of surface rupture patterns exists and in order to make a complete analysis with the current data, the pre-existing structures are modeled as either (1) preventing lateral propagation of rupture or (2) creating a segment boundary and a rupture propagating through these pre-existing

Table 5. Bridger Range fault (BRF) moment magnitude (M_w) parameters and maximum earthquake. Equations used from Hanks and Kanamori (1979).

Section	Surface Rupture Length (km)	Surface Offset (m)	Down-Dip Rupture W(km)	M_w
Walton Creek - Bostwick Creek	8.1	2.0	16.0	6.6
Walton Creek - Bostwick Creek	8.1	3.0	16.0	6.7
Bostwick Creek - Ross Creek	8.2	2.0	16.0	6.6
Bostwick Creek - Ross Creek	8.2	3.0	16.0	6.7
Walton Creek - Ross Creek	16.3	2.0	16.0	6.8
Walton Creek - Ross Creek	16.3	3.0	16.0	6.9
Ross Creek - Cottonwood Salient	12.0	2.0	16.0	6.7
Ross Creek - Cottonwood Salient	12.0	3.0	16.0	6.8
Cottonwood Salient - Blacktail Mtn	16.8	2.0	16.0	6.8
Cottonwood Salient - Blacktail Mtn	16.8	3.0	16.0	6.9
Intrabasin fault	23.0	2.0	16.0	6.9
Intrabasin fault	23.0	3.0	16.0	7.0

structures (Table 5). All of the moment magnitudes quantified (Table 5) are equal to or exceed above the maximum background earthquake value for the Basin and Range province, $M 6.5$ (dePolo, 1994).

Tectonic Activity and Inferred Slip Rates

Morpho-metric analyses incorporating the RTAC and the faceted spur regional reconnaissance technique, determine a mid to late Pleistocene slip rate of 0.01 m/ka . V_f and M_s values suggest a more rapid slip rate ($> 0.1 \text{ mm/yr}$). These values may represent a long-term ($> 200 \text{ Ka}$) average slip rate of 0.1 m/ka for the Bridger Range, based upon the RTAC. However, the V_f and M_s values can be influenced by changing local base levels and broad regional uplift.

It has been shown in the Grand Valley fault system to the south (Piety et al., 1986; Anders et al., 1989) that slip rates can vary orders of magnitude over different time intervals, apparently associated with the passage of the Yellowstone hotspot. Based upon the RTAC, the Bridger Range fault appears to have undergone more rapid slip rates ($> 0.1 \text{ mm/yr}$) during previous tectonic intervals ($> 200 \text{ ka}$) in order to produce the V_f and M_s values. If V_f and M_s values are not influenced by changing base levels within an open basin, then subsequent late Pleistocene slip rates appear to be at least an order of magnitude less than the previous tectonic interval, of which little is known besides the formation of basal facets and morphometry indicative of slip rates on the order of 0.1 m/ka . With the absence of fault scarps and deeply rilled basal faceted spurs, the Bridger Range is assigned a mid to late Pleistocene type 2 fault class with the faceted spur

technique (dePolo, 1998). Type 2 faults have a slip rate of 0.01 mm/yr. This is an order of magnitude less than that quantified by the RTAC.

Recurrence Intervals

Evidence for a late Pleistocene (< 130 ka) surface rupturing event is generally absent along the Bridger Range front. This means that either there has not been a surface rupture within the past 130 ka or it has been eroded and concealed by rapid erosional and burial processes. Based upon the lack of well-defined fault scarps on late Pleistocene deposits associated with the last glacial maximum (≈ 20 ka), the recurrence interval is at least > 20 ka. With the Bridger Range fault being assigned a slip rate of 0.01 mm/yr, a maximum recurrence interval would be 100 ka for a 1-2 m surface rupture. Recurrence intervals for the Bridger Range fault may range between a minimum of 60 ka and a maximum of 100 ka for the mid to late Pleistocene.

CHAPTER 11

DISCUSSION

Methodology

The morpho-metric analyses used herein to determine regions of differing tectonic activity provide a method to quantify long-term tectonic activity within normal-faulted provinces and to correlate between sub-provinces and zones of differing extensional deformation rates. By solving for variables involved in tectonic deformation and the exogenic processes working to dampen the tectonic signal, one can quantify a more long-term average tectonic activity rate to be compared with slip rates quantified over shorter time intervals (< 130 ka). Tectonic activity rates derived from these morpho-metric analyses for the Grand Valley fault system, MRF, and EDLF all show a direct comparison to slip rates determined through other analyses (i.e., trench data, offset and tilted volcanic rocks, and fault scarp morphometry with age of offset landforms). This confirms the utility of the RTAC for regional reconnaissance studies within semi-arid to arid southwest Montana.

The faceted spur regional reconnaissance technique provides yet another method to estimate tectonic activity. A calibration of this technique to the northeast Basin and Range province shows good agreement with tectonic activity rates determined through other analyses (e.g., fault scarp morphometry). When the surficial geology and regional tectonic framework are considered together, this method provides a tool to quantify

tectonic activity rates to produce the resulting range-front morphology. When the RTAC, faceted spur reconnaissance technique, and surficial geology are all in accordance with the tectonic activity determined for the specific fault, a reliable reconnaissance database may be compiled to compare between sub-provinces and zones for future more detailed paleoseismologic investigations.

Clearly, the variables quantified by these morpho-metric analyses are affected on time intervals greater than 100,000 years. They provide a means to quantify and compare long-term tectonic activity to shorter-term deformation preserved within the late Pleistocene and Holocene (time scales $< 10^5$ years). These analyses provide methods to determine regional grouping and migration patterns of faulting and concentrations of tectonic activity along individual fault systems over periods greater than 100 ka.

All three faults under investigation show similar V_f and M_s ratio values; however, the MRF and EDLF show late Pleistocene-Holocene surface ruptures and the BRF has limited, uncertain expression. This suggests that all three fault systems have undergone similar tectonic activity rates during previous episodes of deformation, but the late Pleistocene-Holocene activity has been concentrated on the MRF and EDLF. As previously mentioned, the BRF RTAC analyses, indicating more rapid tectonic activity rates, could be a result of local base level changes induced by regional uplift. This would cause greater incision of the footwall deposits and bedrock independent of range-front fault activity rates. An implication for regional seismic hazard assessments from these morpho-metric analyses are that faults that appear to have no late Pleistocene to Holocene surface rupture, but have remnant morphology from prior tectonic episodes of

greater rates, may be towards the end of the present seismic cycle and associated seismic quiescence.

Regional Neotectonics

Late Quaternary slip rates derived from the RTAC, surficial geology, and basal facet reconnaissance technique show accordance with Pierce and Morgan's (1992) modeled belts of differing tectonic activity rates around the Yellowstone volcanic field. The Madison Range fault does have a definite increase in late Pleistocene slip rate towards its southern end (*belt II* of the YTP) and towards the Yellowstone volcanic field. The EDLF shows similar late Pleistocene slip rates as the northern MRF, in accordance with *belt I* of the YTP, and the BRF lies outside of the YTP (Pierce and Morgan, 1992). Based upon basal facet reconnaissance and surficial geology, the Bridger Range fault is assigned a slower late Pleistocene (< 130 ka) slip rate (< 0.01 mm/yr), than the EDLF and the MRF, which are both > 0.1 mm/yr. However, long-term Quaternary tectonic activity rates between the MRF, EDLF, and BRF quantified by the RTAC in this study have minimum values of > 0.1 mm/yr with a decrease in late Pleistocene slip rates along the BRF shown by the basal facet reconnaissance technique. The high tectonic activity rates quantified by the RTAC for the BRF may be more of a product of regional incision rather than faulting. Even so, Late Pleistocene slip rates increase and large historic earthquakes (1925 Clarkston Valley M 6.75) have occurred north of the BRF towards the Lewis and Clark fault zone (LCZ), a major accommodation zone for northern Basin and Range extension (Figure 1) (Hamilton and Meyers, 1966; Stickney, 1995; Stickney et al., 2000).

Late Pleistocene faulting and seismicity north and south of the BRF show heterogeneity in temporal and spatial earthquake activity that cannot wholly be explained by a simplified model of parabolic deformation around the Yellowstone volcanic field.

Within southwest Montana, a southward increase in late Pleistocene tectonic activity does correlate to a reflection over the SRP of northward increasing tectonic activity along the south arm of the YTP proximal to the Yellowstone volcanic field. Further west from the Yellowstone volcanic field and south of the SRP, the Grand Valley fault system has slip rates ranging from 0.6-1.1 mm/yr along the Star Valley fault to 0.019 mm/yr along the Swan Valley, showing a southward increase in order of magnitude for late Quaternary (< 130 ka) slip rates (Anders et al., 1989; Piety et al., 1992). However, if V_f and M_s have not been a product of local base level changes, the RTAC analyses suggest that the long-term Quaternary tectonic activity has been comparable between these two segments of the Grand Valley fault system. North of the SRP west of the region proximal to the Yellowstone volcanic field, late Pleistocene slip rates increase towards the Yellowstone caldera (e.g., MRF and EDLF) and south of the SRP slip rates increase towards the SRP.

RTAC analyses suggest that mid-Pleistocene slip rates along the Bridger Range fault were greater than those predicted by the parabolic model of Pierce and Morgan (1992). North of the Bridger Range in the Helena area, late Pleistocene surface ruptures are present (Stickney et al., 2000) and seismicity increases (Stickney, 1995). The tectono-morphic signature for the MRF, EDLF, and BRF suggests long-term tectonic activity rates > 0.1 mm/yr, with late Pleistocene heterogeneity in spatial and temporal

earthquake activity. Wallace (1987) suggests that nonuniformity in spatial and temporal occurrences of surface rupturing events characterizes seismogenic faulting in the Basin and Range province. This could be a possible explanation for the heterogeneity in late Pleistocene slip rates on faults between the Yellowstone volcanic field and the Lewis and Clark zone (LCZ). Further paleoseismologic studies are needed to constrain migrational patterns of late Pleistocene tectonic activity.

In addition to temporal and spatial heterogeneity of surface rupturing events, there is clear evidence of pre-existing structural control on fault orientation (Schmidt et al., 1986; O'Neill, 1993; Ruppel, 1995). Southwest Montana is characterized by complex geologic structure and a wide array of Quaternary fault orientations, resulting in heterogeneity in net slip vectors and reduced vertical slip rates along faults not oriented orthogonal to compressive stress fields. Nonuniform fault slip rates along large seismogenic faults may be a product of structurally-controlled fault orientation within a complex stress field.

CHAPTER 12

CONCLUSIONS

The results of this study obtained through morpho-metric analyses, surficial geology, and seismicity permit the following conclusions: 1) morpho-metric techniques used in the Great Basin are applicable to the northeast Basin and Range province, where climate and the Yellowstone volcanic field are additional variables; 2) Areas of greater seismic potential can be identified with the utilization and collaboration of the morpho-metric analyses used herein; and 3) Late Quaternary slip rates derived for the MRF, EDLF, and BRF do show accordance with the northern arm of the YTP (Pierce and Morgan, 1992); however, the mid-Pleistocene to Holocene temporal and spatial heterogeneity of earthquake activity within the study area shows that tectonic activity is more complex than the parabola associated with the Yellowstone volcanic field.

Comparable climatic regimes and related geomorphic rates allow the use of morpho-metric analyses, developed in the Great Basin, in the northeast Basin and Range province and southwest Montana. After calibrating and utilizing the RTAC and basal facet height reconnaissance technique along the Swan and Star Valleys of Idaho and Wyoming, where slightly greater annual precipitation and the presence of the YTP are variables, these techniques can be applied to similar climatic and tectonic regimes north of the Snake River Plain (SRP) in southwest Montana. It must be noted that when utilizing these techniques in more humid regions (e.g., west-northwest Montana), the quantified slip rates provide minimum rates. Slip rates will have to be more rapid in a

humid environment to maintain values equivalent to those quantified in arid to semi-arid environments.

Areas of greater seismic potential and resulting hazards can be identified with the collaboration of the analyses used in this study. Through the RTAC, all of the three ranges under investigation yield Quaternary slip rates of the same order of magnitude (> 0.1 mm/yr). By comparing and constraining these rates with surficial geology and the basal facet reconnaissance technique, a more accurate assessment of the seismic potential for a specific fault can be made. When used with detailed surficial mapping, the basal facet reconnaissance technique (dePolo, 1998) can constrain late Pleistocene (< 130 ka) slip rates (e.g., Bridger Range fault). The RTAC parameters and measurements are affected on periods longer than 100 ka, yielding an assessment of fault activity over the last 0.5 Ma. Other geomorphic process must be considered for the when there is discordance in slip rates quantified by differing techniques (e.g., Bridger Range fault, MT and the Swan Valley segment of the Grand Valley fault, ID and WY). Slip rates can vary on the scale of 10-100 ka intervals; however, the long-term slip rate derived from the RTAC yields a more thorough assessment of seismic potential. If the RTAC is more indicative of the long-term rate, then the quiescence could be a minor bifurcation in the long-term average slip rate (Wallace, 1987), making it erroneous to discredit the seismic potential and associated hazards of faults with morpho-metric values indicative of more rapid slip rates, but lack of late Pleistocene (< 130 ka) surface rupture. In addition to these considerations, the basal facet reconnaissance technique yields a more approximate late Pleistocene slip rate among the faults under investigation.

The Bridger Range fault may possibly have experienced > 100 ka tectonic activity rates greater than those predicted by the YTP of Pierce and Morgan (1992). Late Pleistocene tectonic activity increases both to the north and south of the BF, which is not predicted by the YTP. Lastly, there is clear evidence that the diverse array of Quaternary faults in southwest Montana is controlled by pre-existing structure, indicating that fault orientation cannot be solely related to a simplistic parabolic stress field model.

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APPENDICES

APPENDIX A

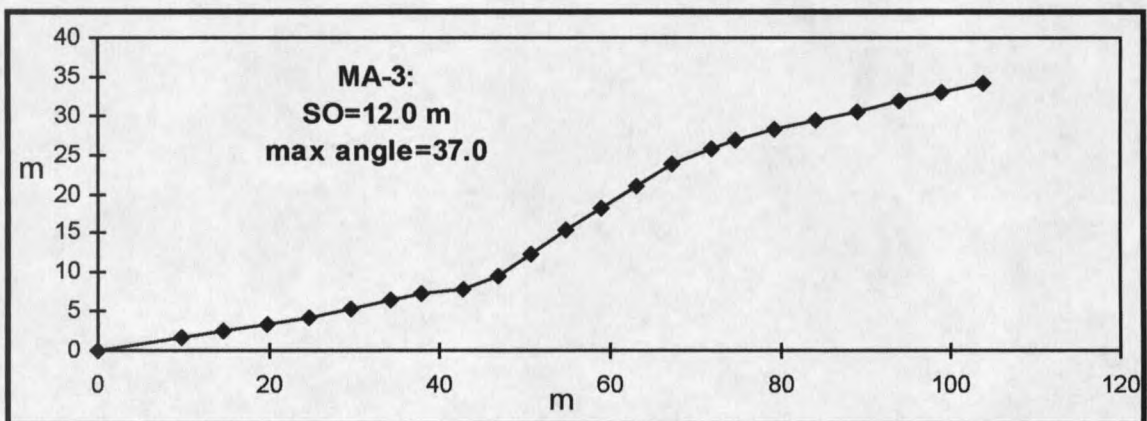
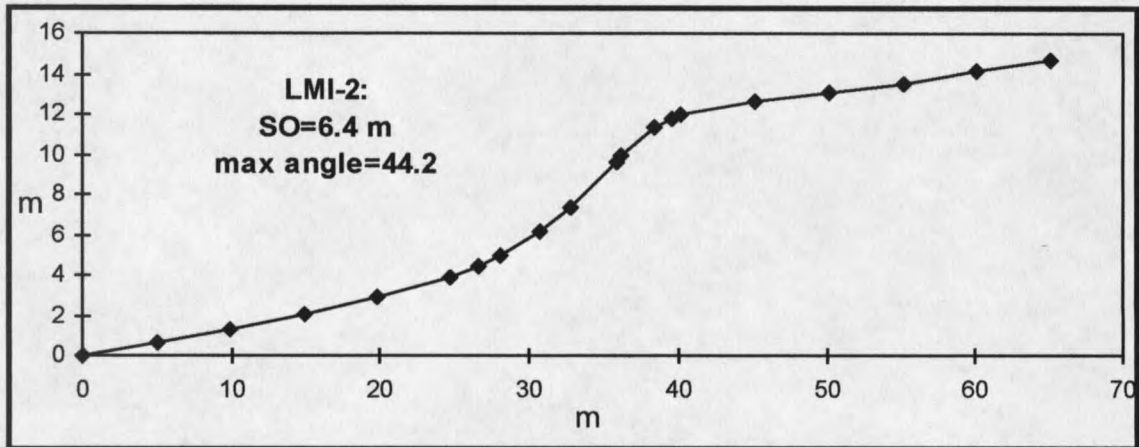
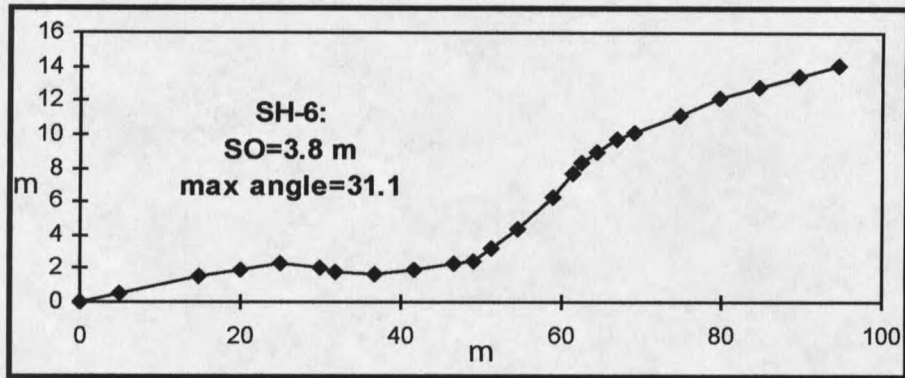
MADISON RANGE FAULT SCARP PROFILE DATA

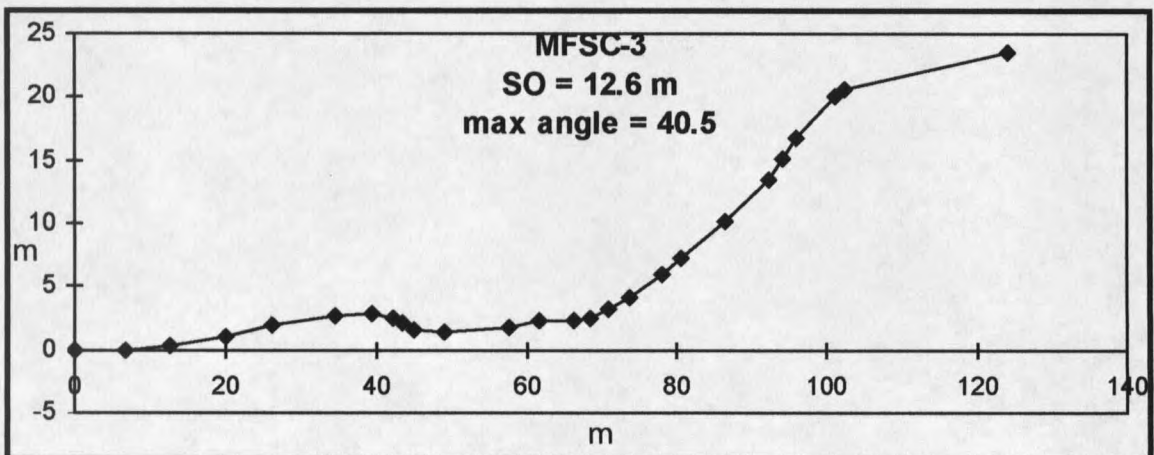
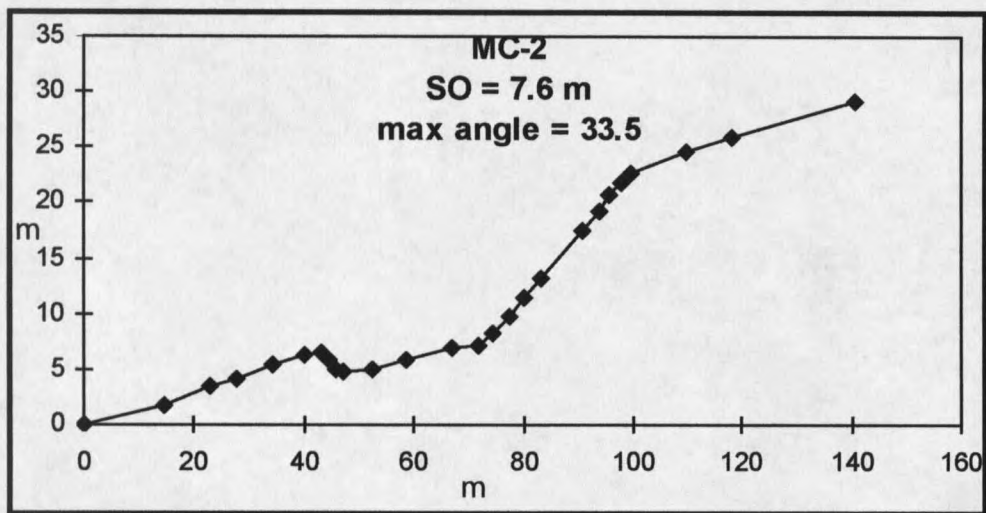
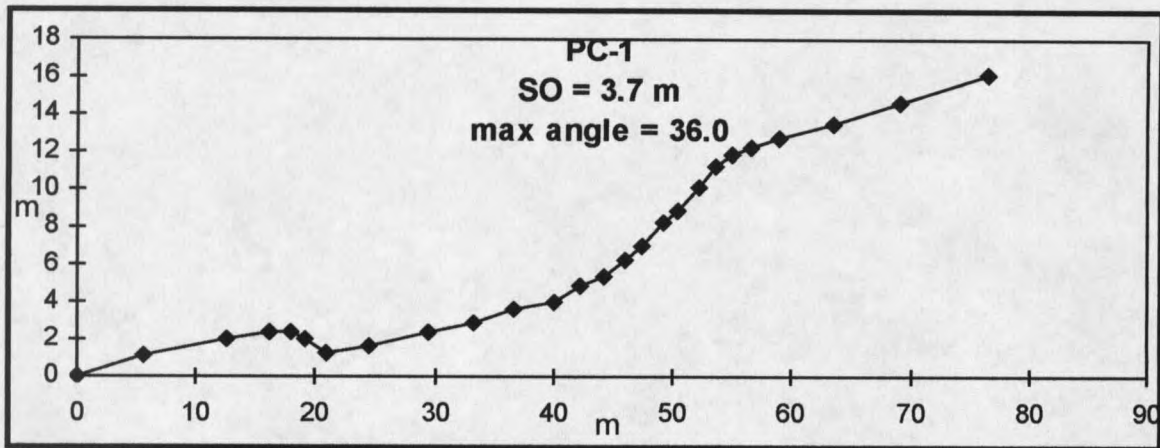
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MI-2	5.8	7.1	33.1	Qfp	
MI-3	3.6	6	35.5	Qfy	
MI-4	5.1	5.1	31.9	Qfp	
MI-5	7	6.4	37.2	Qmp	
MI-6	3.3	6.2	25	Qfy	
LMI-1	14.4	18.6	41.5	Qfp	
LMI-2	6.4	7.3	44.2	Qfy	
LMI-3	7.3	14	38.3	Qca	
LMI-4	6.5	6.9	31.3	Qls	
LMI-5	11	18.7	43.8	Qca	
LMI-6	11.1	14.7	42.1	Qca	
LMI-7	3.3	5.8	31.5	Qfy	
LMI-8	8	13.8	50.5	Qca	
SH-1	10.3	11.3	39.8	Qfp	
SH-2	8.7	9	43.7	Qfp	
SH-3	5.1	7.1	41.4	Qfy	
SH-4	9.3	11.2	40.4	Qfp	
SH-5	6.8	8.3	40.1	Qfp	
SH-6	3.8	4.7	31.1	Qfy	
SH-7	4	6.1	32.8	Qfy	
SH-8	11.5	15.1	39	Qca	
MA-1	9.4	17.7	42	Qfo	
MA-2	12.4	15.2	41	Qfp	
MA-3	12	16.8	37	Qfo	
MA-4	2.4	4.9	33.9	Qfy	
MA-5	7.3	11.2	34.7	Qca	
MA-6	9.8	13.7	38.7	Qca	
MA-7	5.8	8	39.3	Qca	
MA-8	11.1	15.5	40.3	Qca	
MA-9	10	18.1	38	Qfo	
MA-10	9.9	15.1	34	Qfp	
MA-11	8.5	15.6	39.5	Qfp	
MA-12	2.2	4.1	35	Qfy	
MA-13	6.3	12.5	40.5	Qfy	
MA-14	6.5	12.5	33.5	Qfy	
MA-15	3.9	8.5	36	Qfy	
MA-16	5	10.9	38	Qca	
DC-1	5.1	8.5	41	Qca	
DC-2	7.2	9.8	35	Qfp	

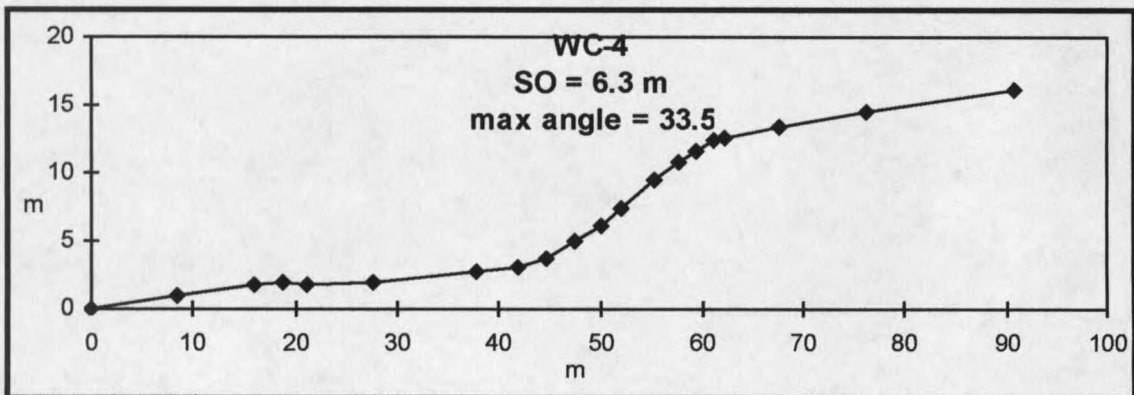
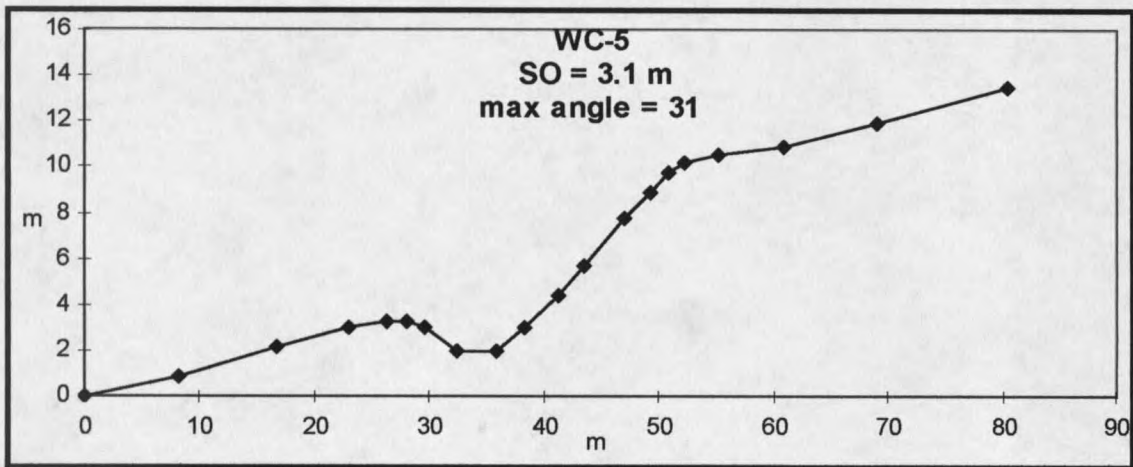
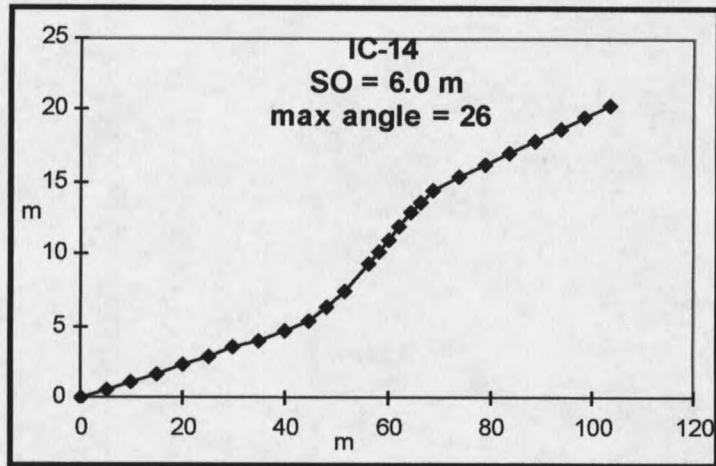
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DC-6	11.2	16.5	33	Qca
DC-7	3.2	4	27	Qfp
DC-8	5.9	7.3	34	Qfp
PC-1	3.7	4.8	36	Qfy
PC-2	6.1	6.7	30	Qfp
SFSC-1	5.9	7.1	43.5	Qfp
MFSC-1	7.2	9.3	36.5	Qca
MFSC-2	2.3	3.2	31	Qfy
MFSC-3	12.6	14.4	40.5	Qmp
NFSC-1	8.4	8.8	34	Qmp
NFSC-2	8.2	8.7	34.5	Qfp
NFSC-3	5.2	7.7	42.5	Qca
MC-1	10.5	11.7	36.5	Qmp
MC-2	7.6	9.9	33.5	Qmp
MC-3	6.9	10.8	36	Qfp
BL-1	5.9	11.4	34.5	Qfp
BL-2	8.2	13.1	41	Qfp
BL-3	9.4	13.1	38	Qfp
WC-1	10	11.1	33.5	Qmp
WC-2	3.1	5	31.5	Qfy
WC-3	8.8	10.8	30.5	Qca
WC-4	6.3	7.5	33.5	Qfp
WC-5	3.1	4	31	Qfy
WC-6	8.1	13.2	28	Qfb
WC-7	4.2	6.2	28.5	Qfb
WC-8	2.9	4.2	34.5	Qfy
DR-1	5.4	9.4	27.1	Qca
DR-2	7	13.3	32.6	Qca
DR-3	3.6	6.4	36.3	Qca
IC-1	4	6.5	31.5	Qfp
IC-2	5.5	8.8	32.6	Qfp
IC-3	5.6	7.2	31.4	Qfp
IC-4	3.7	4.5	36.3	Qfp
IC-5	6.3	9.4	28.1	Qfp
IC-6	1	1.6	25.4	Qca
IC-7	4.1	7.9	30.1	Qfp
IC-8	2.4	4.5	31.9	Qfy
IC-9	3.3	6.2	23.5	Qca
IC-10	6.3	9	31.8	Qfp
IC-11	4.7	6.7	24.8	Qca

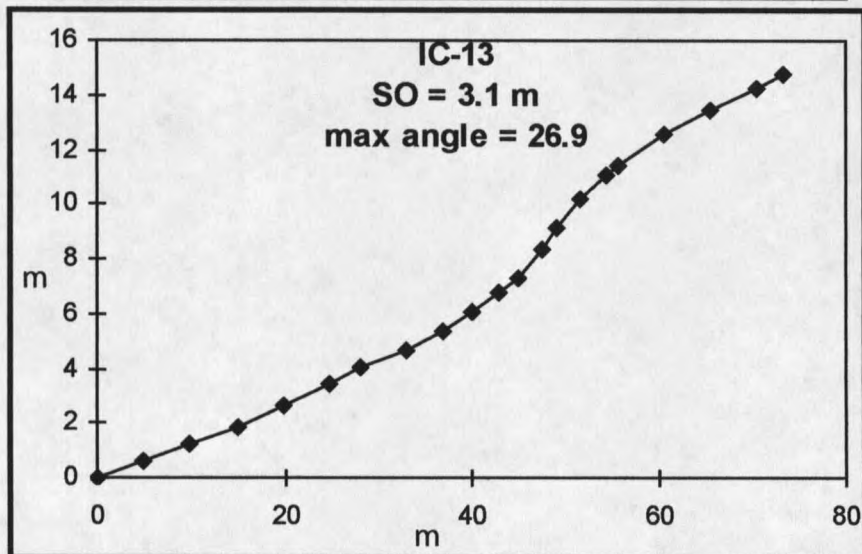
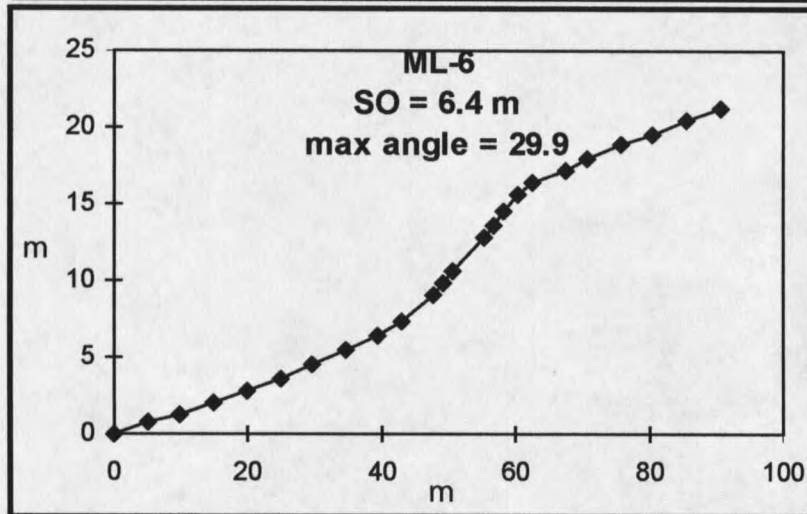
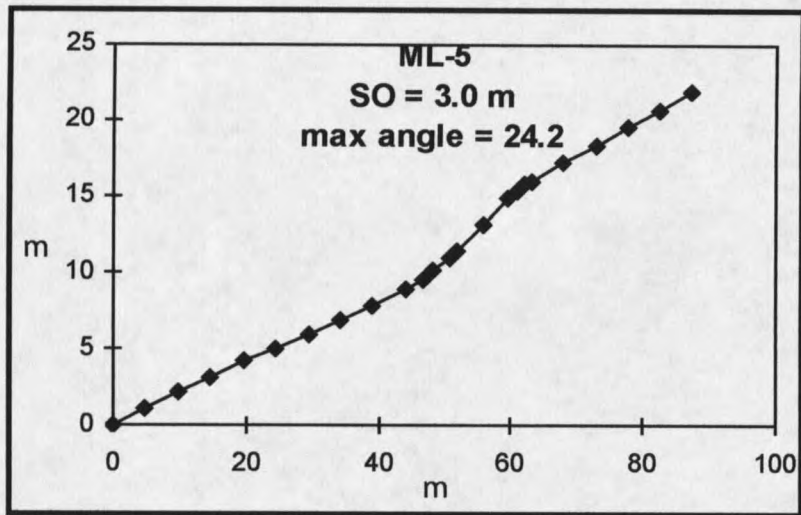
IC-12	7.1	12.2	27.9	Qca
IC-13	3.1	4.7	26.1	Qfp
IC-14	6	8.4	26.3	Qfp
NFBC-1	4	7.4	27	Qca
NFBC-2	3.6	5.1	25	Qca
NFBC-3	1	1.8	20.5	Qca
ML-1	5.5	7.3	36.5	Qfp
ML-2	3.5	5.6	35.5	Qca
ML-3	2.6	3.8	23	Qfp
ML-4	4	6.5	31	Qfp
ML-5	6.4	9	29.9	Qfp
ML-6	3	5.9	24.2	Qfy
ML-7	4.7	6.5	34.6	Qfp
ML-8	4.6	6.3	36.9	Qfp
ML-9	8	16	46.3	Qca
SL-1	8.8	12	33.3	Qfp
CC-1	2.4	3.1	22.2	Qfp
CC-2	4.9	9.2	28.9	Qca
JC-1	1.3	3.1	22.6	Qfp

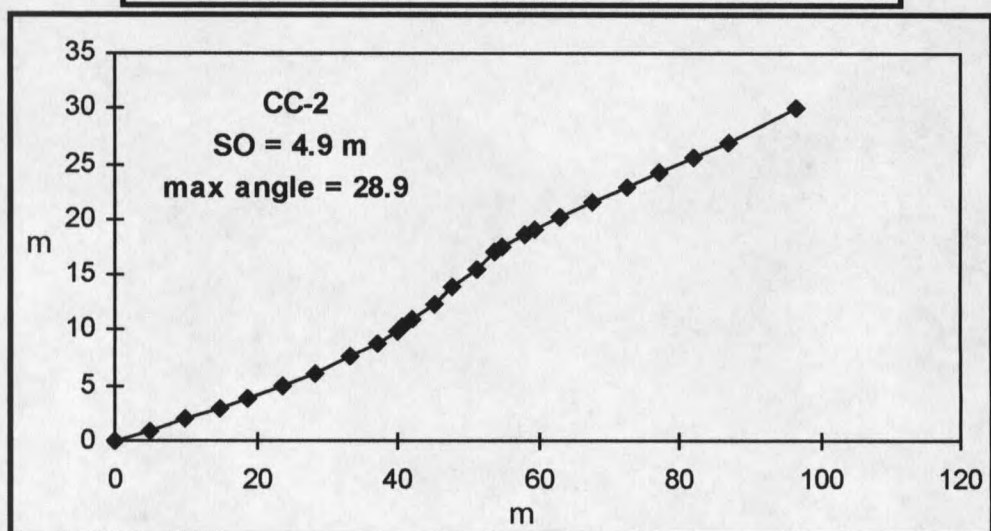
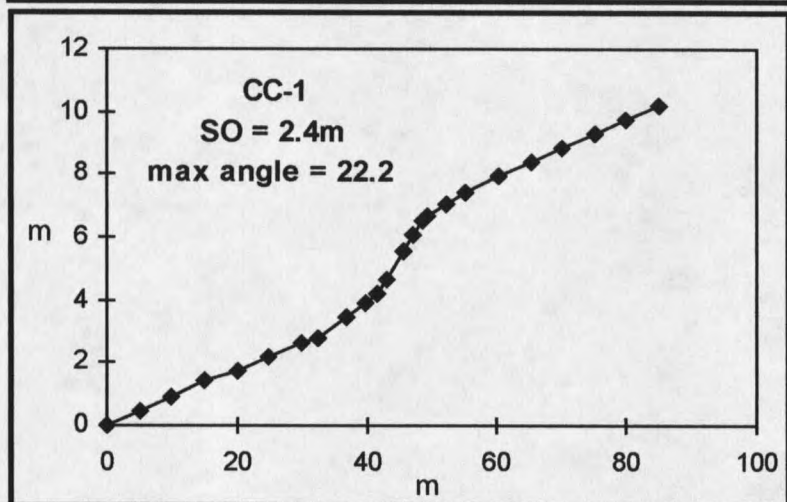
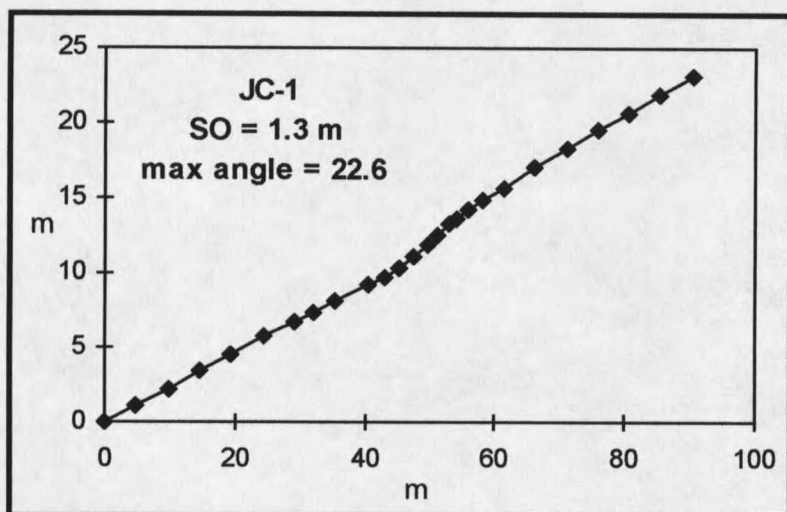
SELECTED FAULT SCARP PROFILES











APPENDIX B
VfRATIO DATA

MADISON RANGE V_F RATIO DATA

Location	E_{ld}	E_{rd}	E_{sc}	V_{fw}	V_f
St. Joe Creek	7080'	7640'	6400'	20'	0.02
Short Creek	7320'	7240'	6400'	20'	0.02
Crooked Creek	7228'	7126'	6390'	20'	0.03
Watkins Creek	7001'	7228'	6400'	20'	0.03
Jourdain Creek	7080'	7001'	6570'	30'	0.06
Jack Creek	7161'	8087'	5580'	60'	0.03
Cedar Creek	7382'	8154'	6260'	90'	0.06
Shell Creek	8120'	8423'	6880'	40'	0.03
Tolman Creek	7640'	8419'	6740'	30'	0.02
Mill Creek	8908'	8920'	7200'	50'	0.03
Cameron Creek	7400'	7712'	7100'	20'	0.04
North Fork Bear Creek	8242'	8472'	7190'	30'	0.03
Middle Fork Bear Creek	8688'	8242'	6700'	70'	0.04
Trail Fork Bear Creek	8650'	8688'	7010'	100'	0.06
Indian Creek	7440'	7951'	6260'	80'	0.06
South Fork Indian Creek	8452'	8680'	7100'	70'	0.05
Corral Creek	9037'	9040'	7880'	30'	0.03
Deer Creek	7770'	7640'	7170'	40'	0.08
Wolf Creek	8140'	9162'	7000'	80'	0.05
Bad Luck Creek	8942'	8721'	7560'	15'	0.01
Moose Creek	9390'	8942'	7840'	30'	0.02
North Fork Squaw Creek	9706'	9765'	8450'	20'	0.02
Squaw Creek	9414'	9706'	7830'	40'	0.02
Middle Fork Squaw Creek	9745'	9414'	7590'	30'	0.02
Papoose Creek	8551'	8958'	7480'	25'	0.02
Curlew Creek	8000'	8000'	7640'	20'	0.06
Deadman Creek	8200'	8600'	7780'	45'	0.06
Sheep Creek	7860'	7600'	6740'	40'	0.04
South of Sheep Creek	8500'	8800'	7400'	30'	0.02
Little Mile Creek	8380'	8500'	7320'	50'	0.05
Mile Creek	8680'	9120'	7460'	80'	0.1

average V_f : 0.04

(measured approx. 1.5 km into footwall)

ABSAROKA-BEARTOOTH RANGE/PARADISE VALLEY V_f RATIO DATA

Location	E_{ld}	E_{rd}	E_{sc}	V_{fw}	V_f
Suce Creek	6360'	6350'	5160'	320'	0.27
North Fork Deep Creek	6598'	6360'	5600'	160'	0.18
South Fork Deep Creek					
Pool Creek	5930'	5600'	5420'	60'	0.17
Pine Creek	9041'		5960'	180'	
Barney Creek a (Deep Creek Fault)	6120'	6120'	5520'	40'	0.07
Barney Creek b (Luccock Park Flt)	7330'	7650'	7000'	35'	0.07
Cascade Creek a (Luccock Park Flt)	9321'	9748'	7720'	120'	0.07
Cascade Creek b (Deep Creek Fault)	5720'	6120'	5650'	15'	0.06
George Creek (Deep Creek Fault)	5520'	5600'	5400'	15'	0.09
N.F. Macdonald Creek a (Deep Crk)	5752'	5480'	5180'	25'	0.06
N.F. Macdonald Creek b (Luccock Park Fault)	8655'	8160'	7600'	15'	0.02
S.F. Macdonald Creek a (Deep Creek Fault)	5762'	5752'	5560'	20'	0.1
S.F. Macdonald Creek b (Luccock Park Fault)	9658'	9125'	7520'	100'	0.05
Strawberry Creek a (Deep Creek Fault)	5960'	5762'	5680'	10'	0.06
Strawberry Creek b (Luccock Park Fault)	9760'	9480'	7160'	160'	0.07
Johnson Creek (Deep Creek Fault)	6296'	5982'	5800'	10'	0.03
Kauffman Creek (Deep Creek Fault)	6331'	6296'	5800'	20'	0.04
South of Kauffman Creek (Deep Creek Fault)	6128'	6331'	5900'	15'	0.05
Magee Creek (Luccock Park Fault)	6662'	6536'	6240'	15'	0.04
North Fork Elbow Creek (Luccock Park Fault)	7490'	9760'	7160'	40'	0.03
South Fork Elbow Creek (Deep Creek Fault)	8720'	7490'	6760'	180'	0.13
Mill Creek	7800'	8440'	5380'	120'	0.04
Davis Creek	7005'	6720'	5990'	80'	0.09
Between Davis and Conlin Creeks	7120'	7080'	6400'	30'	0.04
Conlin Gulch	7780'	7280'	6600'	40'	0.04
Emigrant Gulch	7615'	7780'	5740'	240'	0.12
Creek North of Gold Run Creek	6648'	6736'	6240'	20'	0.04
Gold Run Creek	7561'	7287'	7000'	15'	0.04
Gold Prize Creek	7746'	7135'	6400'	40'	0.04
Sixmile Creek	7336'	6860'	5670'	100'	0.07
Sheep Creek	7020'	6912'	6580'	90'	0.23
Dailey Basin Drainage	6851'	6852'	6200'	60'	0.09
Yellowstone River - Yankee Jim Canyon	6640'	7280'	6040'	190'	0.21
Average V_f = 0.08					

BRIDGER RANGE V_F RATIO DATA

Location	E_{fd}	E_{rd}	E_{sc}	V_{fw}	V_f	avg V_f
Walton Creek a	6160'	6200'	5720' 40'	0.087		
Walton Creek b	6620'	6640'	6000' 30'	0.048	0.06	
Walton Creek c	6682'	6806'	6200' 30'	0.055		
Snypes Canyon a	6220'	6400'	5600' 30'	0.042		
Snypes Canyon b	6064'	6120'	5490' 50'	0.083	0.06	
Snypes Canyon c	6520'	6600'	5800' 40'	0.053		
Middle Cottonwood a	6800'	6400'	5520' 40'	0.037		
Middle Cottonwood b	6800'	6720'	5600' 40'	0.035	0.03	
Middle Cottonwood c	6842'	6362'	5800' 25'	0.031		
Schaefer Creek	6240'	7180'	5760' 25'	0.026	0.03	
Bostwick Creek a	6680'	6760'	5760' 40'	0.042		
Bostwick Creek b	7120'	7520'	6200' 30'	0.027	0.03	
Bostwick Creek c	7560'	7172'	6000' 30'	0.022		
Truman Gulch a	7160'	6350'	5500' 60'	0.048		
Truman Gulch b	7240'	6760'	5880' 30'	0.027	0.04	
Truman Gulch c	7160'	6680'	5720' 40'	0.033		
Jones Creek a	6760'	6790'	5760' 40'	0.039		
Jones Creek b	6320'	6320'	5560' 40'	0.053	0.04	
Jones Creek c	6720'	6760'	5880' 30'	0.035		
Ross Creek a	6080'	6120'	5800' 20'	0.067		
Ross Creek b	6050'	5920'	5600' 20'	0.052	0.06	
Ross Creek c	6460'	7080'	6000' 40'	0.052		
Potters Gulch	6520'	6640'	6240' 20'	0.059	0.06	
Dry Canyon a	6640'	7200'	5760' 40'	0.035		
Dry Canyon b	7400'	7200'	6120' 30'	0.025	0.03	
Dry Canyon c	6690'	7200'	5880' 20'	0.019		
Limestone Canyon a	6760'	6920'	6280' 40'	0.071		
Limestone Canyon b	6440'	6300'	5920' 20'	0.044	0.05	
Limestone Canyon c	7240'	7441'	6600' 30'	0.041		
Evan Reese Creek a	6440'	6560'	5720' 30'	0.038		
Evan Reese Creek b	6360'	6660'	5800' 20'	0.028	0.03	
Evan Reese Creek c	6760'	7020'	6000' 30'	0.034		
Tom Reese Creek	6560'	6640'	6040' 30'	0.054	0.05	
Smith Creek a	6720'	6600'	5880' 25'	0.032		
Smith Creek b	6920'	7640'	6280' 60'	0.06	0.04	
Smith Creek c	6840'	6640'	6000' 25'	0.034		
N. Cottonwood Creek a	7820'	7320'	6520' 40'	0.038		
N. Cottonwood Creek b	8080'	7520'	6680' 80'	0.071	0.05	
N. Cottonwood Creek c	8380'	8480'	6800' 60'	0.037		
Mill Creek a	8310'	7667'	6790' 80'	0.067		
Mill Creek b	8360'	7520'	7080' 80'	0.093	0.08	
Mill Creek c	8400'	7852'	7000' 90'	0.08		
Johnson Canyon (N. fork)	7600'	7240'	6480' 40'	0.043	0.04	
Pass Creek a	7200'	7340'	6010' 50'	0.04		
Pass Creek b	7120'	7240'	6160' 60'	0.059	0.06	
Pass Creek c	6880'	6520'	5900' 60'	0.075		

APPENDIX C
BASAL FACET HEIGHTS

MADISON RANGE BASAL FACET HEIGHTS (SOUTH TO NORTH) (m)

256
186
268
262
134
232
241
195
232
195
198
201
274
244
146
189
85
208
146
183
220
268
180
183
244
232
137
280
159
137
134
244
134
183
85
159
122

Indian Creek Segment

146
159
122
213

244

213

Bear Creek to Jack Creek Section

146

207

220

244

207

207

305

195

213

186

232

232

116

61

110

61

116

Jack Creek to Jourdain Creek Section

122

134

140

98

110

PARADISE VALLEY BASAL FACET HEIGHTS (LISTED SOUTH TO NORTH) (m)

122

152

122

134

168

122

152

232

256

122

159

220

140

159

APPENDIX D
STAR VALLEY, WY DATA

Star Valley Vf Ratio Data

Location	Eld	Erd	Esc	Vfw	Vf	Avg. Vf
Spring Creek a	7840'	7392'	6810'	80'	0.099	
Spring Creek b	7371'	7360'	6930'	30'	0.069	0.078
Spring Creek c	7720'	7360'	7090'	30'	0.067	
Dry Creek a	7489'	7715'	6465'	90'	0.079	
Dry Creek b	7080'	7400'	6520'	30'	0.042	0.057
Dry Creek c	8461'	8836'	6830'	90'	0.05	
Swift Creek a	7330'	7662'	6390'	50'	0.045	
Swift Creek b	8555'	8160'	6440'	50'	0.026	0.03
Swift Creek c	8606'	8836'	6560'	40'	0.019	
Phillips Creek a	7080'	7378'	6390'	100'	0.12	0.114
Phillips Creek b	7080'	6930'	6540'	50'	0.11	
Willow Creek a	8160'	8280'	6640'	90'	0.06	0.06
Dry Canyon a	9000'	8920'	7080'	94'	0.05	0.05
Strawberry Creek a	8920'	8200'	6640'	141'	0.07	0.07
Lost Creek a	7995'	8127'	6700'	50'	0.037	
Lost Creek b	8600'	8720'	7000'	40'	0.024	0.028
Lost Creek c	8970'	9170'	7280'	40'	0.022	
Cedar Creek a	8160'	8060'	6560'	66'	0.043	
Cedar Creek b	8575'	8649'	6760'	40'	0.022	0.027
Cedar Creek c	9200'	8730'	7000'	30'	0.015	
Prater Canyon a	8400'	7820'	7160'	60'	0.063	
Prater Canyon b	8952'	9273'	7620'	40'	0.027	0.038
Prater Canyon c	9080'	9000'	7860'	30'	0.025	
Lee Creek south a	8230'	7400'	6640'	55'	0.047	
Lee Creek south b	8690'	8280'	7380'	40'	0.036	0.035
Lee Creek south c	9273'	8890'	7860'	35'	0.029	
Lee Creek north a	7400'	8000'	6720'	45'	0.05	
Lee Creek north b	8280'	8200'	7080'	35'	0.03	0.033
Lee Creek north c	8890'	8910'	7640'	25'	0.02	
Call Creek a	7840'	8080'	6900'	55'	0.052	
Call Creek b	8560'	8555'	7240'	35'	0.027	0.032
Call Creek c	8840'	8820'	7600'	20'	0.016	
Corral Creek a	8200'	8160'	7040'	35'	0.031	
Corral Creek b	8695'	8355'	7560'	25'	0.026	0.026
Corral Creek c	9080'	8860'	8000'	20'	0.021	
Stewart Creek	8200'	8036'	6888'	80'	0.065	0.065
Dry Creek	7380'	8200'	6560'	180'	0.146	0.146

STAR VALLEY BASAL FACETED SPUR HEIGHTS (NORTH TO SOUTH) (m)

171
85
134
122
134
183
146
110
98
98
195
183
61
122
110
110
158
220
232
134
122
183
73
189
61
232
183
183
268
165
152
195
195
159
209
183
244
171
226
122
256
183
256
220
256
268
98
110
146
171
137

APPENDIX E
MADISON RANGE MAP UNIT DESCRIPTIONS

- Qal** **Stream alluvium, unit 4 (mid Holocene to Present)** – Stream alluvium within active channels and flood plains. This includes reworked older coarse-grained deposits within active channels and over-bank fine-grained sediments. Contains localized colluvium from oversteepened banks.
- Qaly** **Stream alluvium, (early to mid-Holocene)** – Silt-to boulder-size, moderately sorted to well sorted, unconsolidated sediments; sub-angular to rounded clasts; contains bars of sand and gravel; contains localized poorly-sorted colluvium near steep embankments. Surfaces characterized by bar-and-swale topography of modern drainage channels and floodplains. Surfaces and terraces consist of channel fill along entrenched Q_{f2} , Q_{f1} , Q_{m2} , and Q_{m1} deposits.
- Qalp** **Stream alluvium, (late Pleistocene, Pinedale equivalent)** – Silt- to boulder-size, moderately sorted to well sorted, unconsolidated sediments; sub-angular to rounded clasts; occur at elevations of 3 to 45 m above present river channel at Madison River Canyon (Lundstrom, 1986); contains carbonate coatings of 1-3 cm (Lundstrom, 1986); loess cap of 30-50cm (Lundstrom, 1986). Obsidian hydration analyses yielded an age of 30 ka for the Pinedale terminal moraine at West Yellowstone with younger recessional deposits of 10 to 15 ka (Pierce et al., 1976). Pinedale equivalent terraces at Madison Canyon 15-30 ka (Lundstrom, 1986).
- Qalb** **Stream alluvium, (mid Pleistocene, Bull Lake equivalent)** – Silt- to boulder, size, stream alluvium; 60-70 m above active river channel at Madison River Canyon (Lundstrom, 1986); contains a loess cap of 1m (Lundstrom, 1986). Surface is characterized by a lack of large boulders and subdued bar and swale micro-topography. This alluvium was deposited at the maximum fill level within the basin and correlates with the Cameron Surface.
- Qfy** **Fan alluvium, (early to mid-Holocene)** – Subangular to rounded clasts; poorly sorted; pebble and cobble gravel in a matrix of sand and silt. Also includes small debris flows consisting of angular to subangular, poorly sorted, pebble- to boulder-sized clasts in a clay-rich matrix. Restricted to the mountain-piedmont junction along ephemeral drainage channels. Form fresh, steep alluvial cones along the mountain front. Deposited by intermittent streams and debris flows onto older alluvial surfaces (units Q_{f2} , Q_{al2} , Q_{al1} , and Q_{f1}). Lack of substantial soil development, and includes fine-grained alluvial deposits at toe of cone.
- Qfp** **Fan alluvium, (late Pleistocene, Pinedale equivalent)** – moderately sorted to well sorted, subangular to round clasts; deposition in high-energy, braided fluvial system with fluctuating water and sediment discharge allowing rapid aggradation to occur due to conditions of greater sediment yield and discharge (Funk, 1976). Surfaces are inset into Q_{f1} as great as 11 m; juxtaposed and graded to Q_{m2} and Q_{al2} .
- Qfb** **Fan alluvium, (late to upper middle Pleistocene?, Bull lake equivalent and older)** – Moderately to well-sorted, round to well-rounded clasts in a sand and silt matrix. Surfaces are characterized by smooth, subdued, extensive surfaces grading from Q_{m1} to Q_{qal1} and the Cameron Surface. Contains a loess cap ranging from 0.5 m to 2.0 m in swales.
- Qmp** **Moraine, (late Pleistocene, Pinedale equivalent)** – unconsolidated, poorly sorted, silt- to boulder-size, angular to subangular deposits adjacent to the range front at Deadman, Papoose, Squaw, Moose, Wolf, and Indian Creeks. Surface characterized by partially buried boulders surrounding fine-grained depressions and fully exposed smaller sub-angular cobbles abundant on the surface. Sharp crests show little modification. 60-9 cm thick Bt (Ritter et al., 1995). Soils support more conifers (Matthieson, 1983).
- Qmb** **Moraine, (late Pleistocene, Bull Lake equivalent)** – Characterized by a smooth surface mantled by sparse, partially buried and weathered boulders. Well developed rinds on clasts of all lithologies along the range front (> 1.0 mm); 120 cm thick Bt (Ritter et al., 1995; Reheis, 1987); less pronounced, subdued, hummocky morphology with low angle slopes; absence of undrained, conical depressions (i.e., kettles); overlain with variable thickness loess (> 1.0 m); supports more vegetation due to the moisture holding capacity of loess.

- Qls** **Landslide deposits (late Pleistocene to Holocene)** – clay- to boulder-size, unsorted deposits with little to no sorting. Surfaces characterized by large boulders and hummocky topography. Relative-age of different landslide deposits is not determined within this study. Within the Cliff and Wade Lake benchland and the West Fork of the Madison River, these deposits include landslides, debris flows, and earthflows. Spatial relationships and surficial morphology suggest that age of deposits span from mid-Pleistocene to presently active landforms. Along and within the Madison Range these deposits are characterized by catastrophic mass-wasting events of mid-Pleistocene to late Pleistocene age, with the exception Deadman Creek and Madison Canyon, where seismically-induced landslides occurred with the 1959 Hebgen Lake M_s 7.5 earthquake occurred.
- Qca** **Colluvium and alluvium, undivided (mid Pleistocene to Holocene)** – Consists of poor to moderately sorted, slopewash colluvium along with stream and fan alluvium. Includes localized slumps and rockfall. Predominantly mantles surfaces where direct distinction of landforms is ambiguous and where relative-age of deposit cannot directly be determined.
- QTbf** **Quaternary-Tertiary basin fill (Pliocene-mid Pleistocene)** – coarse-grained to fine-grained, inter-bedded, silt, sand, and well-rounded cobbles that are poorly consolidated to unconsolidated. These deposits are overlain by mid Pleistocene to Holocene deposits and are considered to be younger than Pliocene (Kellogg and Williams, 2000).
- Thrt** **Tertiary Huckleberry Ridge Tuff (Pliocene, 2.0 Ma)** – grey to light-brown, massive welded tuff; contains sparse phenocrysts of sanidine, quartz, and plagioclase; matrix primarily consists of devitrified glass shards, opaque minerals, and aphanitic minerals; contains sparse pumice fragments. Thickness increases to the south with maximum thickness of 50 m (Kellogg and Williams, 2000).
- Tsmc** **Tertiary gravel deposits, Six Mile Creek Formation (Pliocene, ≥ 2.0 Ma)** – poorly consolidated to unconsolidated, massive to well-bedded silt, sand, and well-rounded cobbles (Kellogg and Williams, 2000). Contains clasts of Proterozoic Belt Supergroup quartzite, argillite, and sandstone. Underlies the Huckleberry Ridge tuff from Moose to Papoose Creeks, exposed along the Madison River. Thickness unknown, but overlies 4,500 m of Cenozoic basin-fill deposits (Rasmussen and Fields, 1983).
- Tls** **Tertiary landslide deposits (Pliocene and Miocene)** – older landslide deposits consisting of poor to unsorted, pebble to boulder size clasts with a clay, silt, and sand matrix; significant erosion and modification of the deposit suggests a Tertiary age.
- Tg** **Tertiary volcanoclastic sandstone and gravel deposits (Eocene)** – coarse-grained, well-bedded, cross-bedded volcanoclastic sandstone, pebbly sandstone, and pebble and cobble conglomerate that underlie the Oligocene-Eocene Virginia City volcanic field (Kellogg and Williams, 2000).
- Tv** **Tertiary volcanic rocks, undivided (Oligocene-Eocene)** – andesite and basalt lava flows; coarse-grained, well-bedded, cross-bedded volcanoclastic sandstone; pebble sandstone and pebble-cobble conglomerate (Kellogg and Williams, 2000).
- Kd** **Dacite porphyry of Fan and Lone Mountains (Late Cretaceous)** – greenish-grey, porphyritic dacite, $^{40}\text{Ar}/^{39}\text{Ar}$ and K/Ar ages from hornblende are 68-69 Ma (Tysdale et al., 1986).
- KI** **Livingston Formation (late Cretaceous)** – composed of three members, upper member is a cobble-boulder, volcanoclastic conglomerate; middle member consists of brown, maroon, and volcanic flows ranging from dacite to basalt in composition with welded tuffs and volcanoclastic sandstone; lower member lies conformably on Everts formation and consists of intertonguing volcanoclastic breccias and sandstone with basalt flows. Maximum thickness is approximately 700 m (Kellogg and Williams, 2000).
- Kev** **Everts Formation and Virgelle Sandstone undivided (Upper Cretaceous)** – thin- to medium-bedded, gray to dark gray, poorly sorted to well-sorted sandstone. Contains a few thin beds of limestone, porcellanite, and coal (Kellogg et al., 2000). Everts formation conformably overlies

Virgelle, which is a thin- to thick-bedded, trough-crossbedded sandstone that forms prominent white ledges. Total thickness is as much as 475 m (Kellogg and Williams, 2000).

- Kc** **Cody Shale (Upper Cretaceous)** – thin-bedded black fissile shale interbedded with bioturbated, calcareous, fine-grained sandstone. Lower section consists of black fissile shale. Thickness approximately 300 m.
- Kf** **Frontier Formation (Upper Cretaceous)** – interbedded black shale and buff-colored, thin- to thick-bedded sandstone, approximate thickness is 140-180 m (Kellogg and Williams, 2000).
- Kmo** **Mowry Shale (Upper Cretaceous)** – brownish-gray to greenish-gray tuffaceous mudstone and shale with thin sandstone beds. Approximate thickness 90-180 m (Kellogg and Williams, 2000).
- Kmt** **Muddy Sandstone and Thermopolis Shale undivided (Lower Cretaceous)** – thin- to medium-bedded, brown to brownish-gray salt-and-pepper sandstone above black fissile shale. Lowest member consists of thin- to medium-bedded buff-colored quartz arenite. Maximum thickness approximately 180 m.
- Kk** **Kootenai Formation (Lower Cretaceous)** – upper 10-15 m consists of gastropod, oolitic biomicrite. Middle section consists of purple, red, yellow, and gray shale, mudstone, sandstone, and siltstone. Lower 40 m consists of medium- to coarse-grained, gray, salt-and-pepper sandstone with a basal chert-pebble conglomerate. Maximum thickness ranges from 90-130 m (Kellogg and Williams, 2000).
- JTru** **Morrison Formation (Upper Jurassic), Ellis Group (Upper and Middle Jurassic), Woodside Siltstone (Lower Triassic), and Dinwoody Formation undivided (Lower Triassic)**– 20-30 m of predominantly black shale. Lower section consists of thin-bedded, gray, yellow, red, orange, and green shale interbedded with grey quartz arenite and limestone. Ellis Group consists of the Swift Sandstone, a brown, medium- to coarse-grained glauconitic sandstone, thickness <12 m (Kellogg et al., 2000), Rierdon Formation, a yellowish-gray to brownish-gray, fine-grained oolitic limestone, thickness 8-30 m (Kellogg et al., 2000), Sawtooth Formation is yellowish-brown interbedded limestone and shale, thickness 25-55 m (Kellogg et al., 2000). Woodside Formation consists of brick-red to orangish-red siltstone and mudstone, interbedded with gypsum and thin beds of gray limestone, forms prominent red soil, total thickness 0-220 m (Tysdale, 1990). Dinwoody Formation is a brown silty sandstone interbedded with brown limestone layers, thickness between 20-80 m (Kellogg and Williams, 2000).
- Ps** **Shedhorn Sandstone (Lower Permian)** – medium- to massive-bedded, gray, fine-grained to coarse-grained, quartz-rich sandstone. Dolomitic near the base, maximum thickness 70 m (Kellogg and Williams, 2000).
- IPMqa** **Quadrant Sandstone (Pennsylvanian), Amsden Group (Lower Pennsylvanian and Upper Mississippian), and Snowcrest Range Group (Upper Mississippian)** – Quadrant Sandstone is a medium- to thick-bedded, white to yellowish-tan, well-sorted, fine- to medium-grained dolomite-cemented quartz arenite, approximately thickness 75 m. Amsden Group varies from brick-red, reddish-brown, and pink calcareous siltstone, silty shale, shale, sandstone, and light-gray limestone and yellowish-gray dolomite to medium- to thin-bedded dolostone that grades into gray limestone. Maximum thickness is 88 m. Snowcrest Range Group contains 57 m of thin- to medium-bedded gray limestone that becomes more dolomitic towards the base (Lombard Formation). Below the Lombard Formation, the Kibby Formation consists of 75 m of thin- to medium-bedded, yellowish-gray to maroon, friable dolomitic sandstone, sandy dolostone, and siltstone. Descriptions after Kellogg and Williams (2000).
- Mm** **Madison Group undivided (Upper and Lower Mississippian)** – contains the Mission Canyon Limestone consisting of approximately 240 m of medium- to massive-bedded, light gray to brownish gray medium-crystalline limestone with small amounts of dolostone. Lodgepole Limestone is approximately 180 m thick and contains thin- to medium-bedded, gray to brownish-gray fine-grained limestone that grades into yellowish- to buff-colored silty limestone beds several centimeters thick. Total thickness 480 m (Kellogg and Williams, 2000).

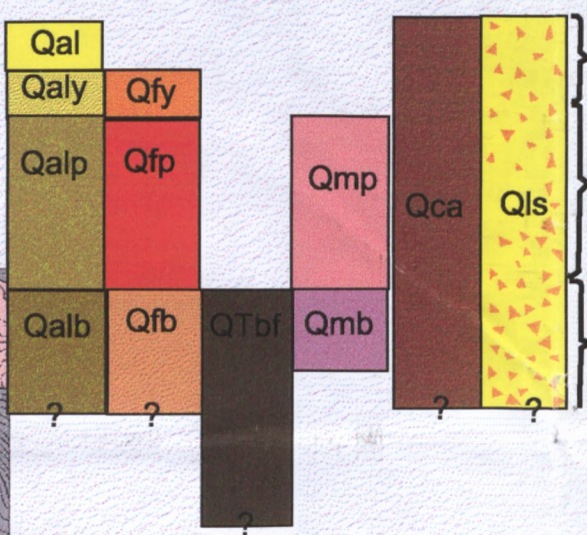
- MDtj** **Three Forks Formation (Lower Mississippian and Upper Devonian) and Jefferson Formation (Upper Devonian)** – Jefferson Formation consists of 40-60 m of mostly thin-bedded, yellowish-orange to yellowish-tan silty limestone and siltstone with minor amounts of thin interbeds of red-weathering siltstone. Jefferson Formation consists of approximately 110 m of thin- to thick-bedded, black, brown, dark-gray, and light-gray, petroliferous, medium-crystalline to coarsely crystalline dolostone (Kellogg and Williams, 2000).
- OCrp** **Bighorn Dolostone (Ordovician), Red Lion Formation (Upper Cambrian), Pilgrim Dolostone (Upper Cambrian), and Park Shale (Middle Cambrian)** – Bighorn Dolostone's presence within the map area is questionable. Red Lion Formation is 40-60 m thick and consists of thin-bedded, medium-gray to tan, siliceous dolostone with orange-tan to reddish-tan cherty stringers up to 2 cm thick. Pilgrim Dolostone is approximately 60 m thick and is a medium- to massive-bedded, locally oolitic, medium-crystalline dolostone varying in color from gray, light-gray, and brownish-gray, contains dark-gray mottles. The Park Shale is a greenish-gray to tan, fissile shale with a thickness of 30-50 m (Kellogg and Williams, 2000).
- Cm** **Meagher Limestone (Middle Cambrian)** – thin- to massive-bedded, light-gray to brownish-gray, fine-grained limestone. Upper 30 m contains orange mottles. Upper few meters consists of greenish-gray shale. Middle 50 m contains medium- to massive-bedded limestone with prominent tan mottling. Lower 30 m contains a few intercalated micaceous shale beds. Total thickness is 100-150 m thick (Kellogg and Williams, 2000).
- Cwf** **Wolsey Shale and Flathead Sandstone undivided (Middle Cambrian)** – Wolsey Shale consists of 30-65 m of mostly thin-bedded, greenish-gray to brownish gray, micaceous sandstone, siltstone, and shale. Sandstone beds are commonly bioturbated with obvious tracks and trails. Middle part contains 10-15 m of thin-bedded, dark-gray, argillaceous limestone. Flathead Sandstone consists of thin- to medium-bedded, medium to coarse-grained, reddish-brown, tan, and purplish-tan, quartz-rich, feldspathic sandstone. Two thin zones of fine-grained, micaceous, greenish-gray argillaceous sandstone are near the top of the formation. Basal unit contains well-rounded clasts of Archean basement rocks (Kellogg and Williams, 2000).
- Xm** **Proterozoic metasedimentary rocks, undifferentiated** – mapped in southern Gravelly Range from O'Neill (unpublished mapping).
- Xmm** **Proterozoic mylonitic gneiss (Madison Mylonite) (early Proterozoic)** – consists predominantly of equigranular quartz, plagioclase, magnetite, and epidote with thin flakes of folded muscovite and biotite. A sharp contact between the quartzofeldspathic gneiss (pL_{qf}) and mylonite (pL_{mm}) has interlayered pink, granular gneiss and dark grey mylonite (Erslev, 1981).
- PCph** **Phyllite (Proterozoic or Archean)** – greenish-gray, greenish-red, gray-red, and dark-gray, contains quartz, biotite, chlorite, muscovite, hematite, magnetite, and sparse garnet (Kellogg and Williams, 2000).
- Aqf** **Archean quartzofeldspathic gneiss** – banded, light- to medium-grey, microcline-plagioclase-quartz-biotite gneiss of unknown origin, contains porphyroblasts of feldspar in a mylonitic fabric. Includes dioritic gneiss, granite gneiss, granitic augen gneiss, foliated granite, granitic migmatite-gneiss, leucotonalite gneiss, tonalitic migmatite-gneiss, tonalitic biotite gneiss.
- Aam** **Archean hornblende-plagioclase gneiss and amphibolite** – gray to black, medium-grained, equigranular, moderately foliated to well-foliated hornblende-plagioclase-garnet gneiss and amphibolite (Kellogg and Williams, 2000). Xenoliths of amphibolite locally present in gneiss
- Ama** **Archean marble** – white, coarse-grained, massive to moderately well foliated dolomitic marble, weathers orange-brown.
- Ado** **Archean dolomitic marble** – dark to light grey dolomite-quartz marble; contains widespread internal deformation of quartzite and dolomite layers.

- Abs** **Archean biotite schist** – medium- to coarse-grained biotite schist; localized chloritic retrograde overprinting.
- Acs** **Archean chlorite-hornblende schist** – contains well-banded green phyllonite, chloritic schist with quartz augen, and granulated medium-grained amphibolite (Erslev, 1981). Crenulation, boudinage, and smeared mica fabrics exist throughout the unit.
- Aq** **Archean quartzite** – white, medium- to coarse-grained, foliated to massive, fuchsite-bearing giving a green hue.
- Aqms** **Archean quartz-mica schist** – contains mainly fine-grained quartz beds that weather yellow and shredded muscovite in tightly folded patterns.
- Aqc** **Archean chloritic quartzite** – light-green quartzite mapped by Erslev (1983) in the Madison Canyon/Missouri Flats region.
- At** **Mylonite of the Crooked Creek Shear Zone (Archean)** – composition ranges from quartzofeldspathic gneiss to amphibolite. Light- to medium-gray mottled mylonite with granoblastic texture (Kellogg and Williams, 2000)

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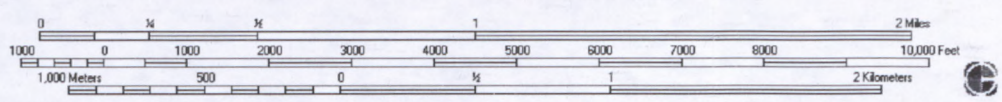
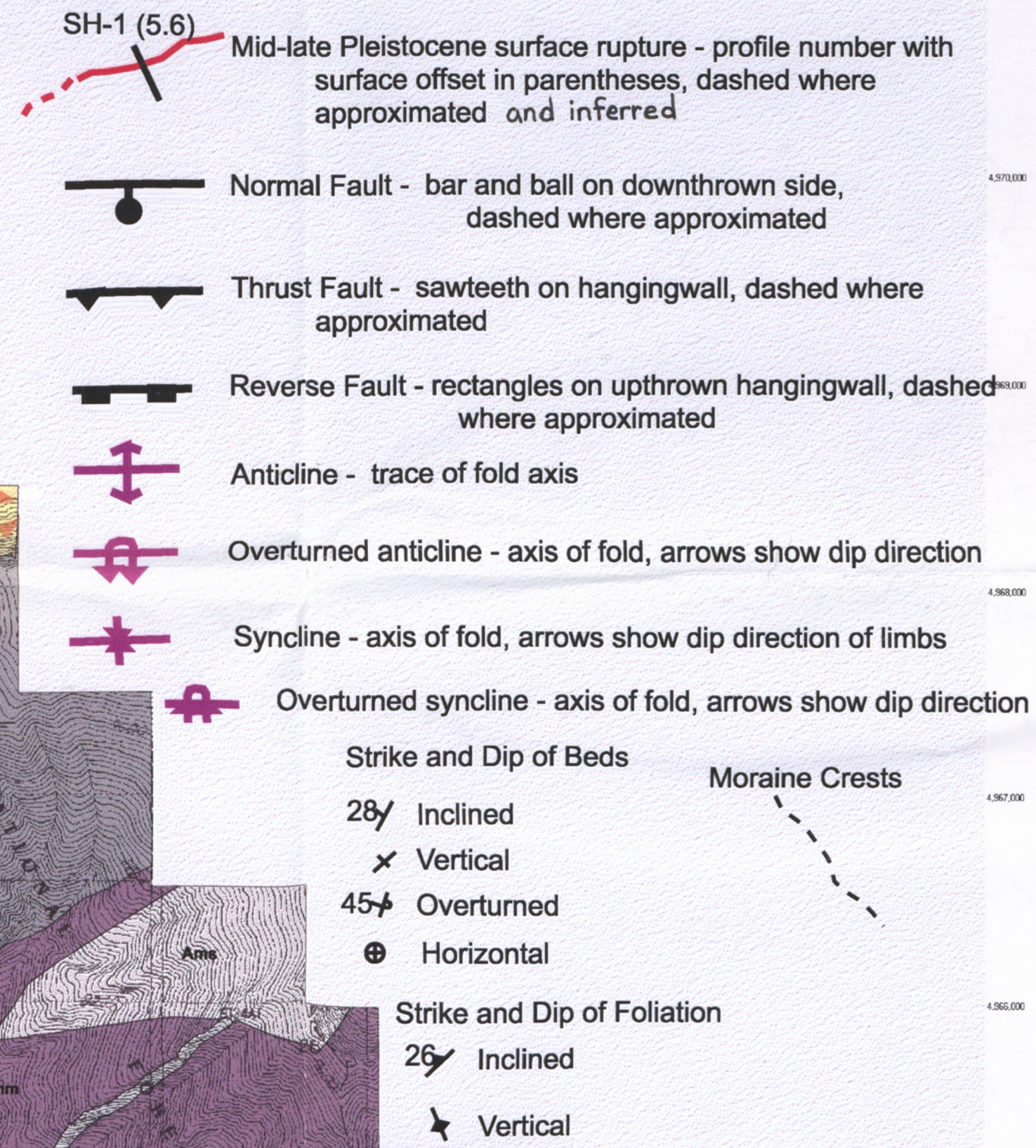


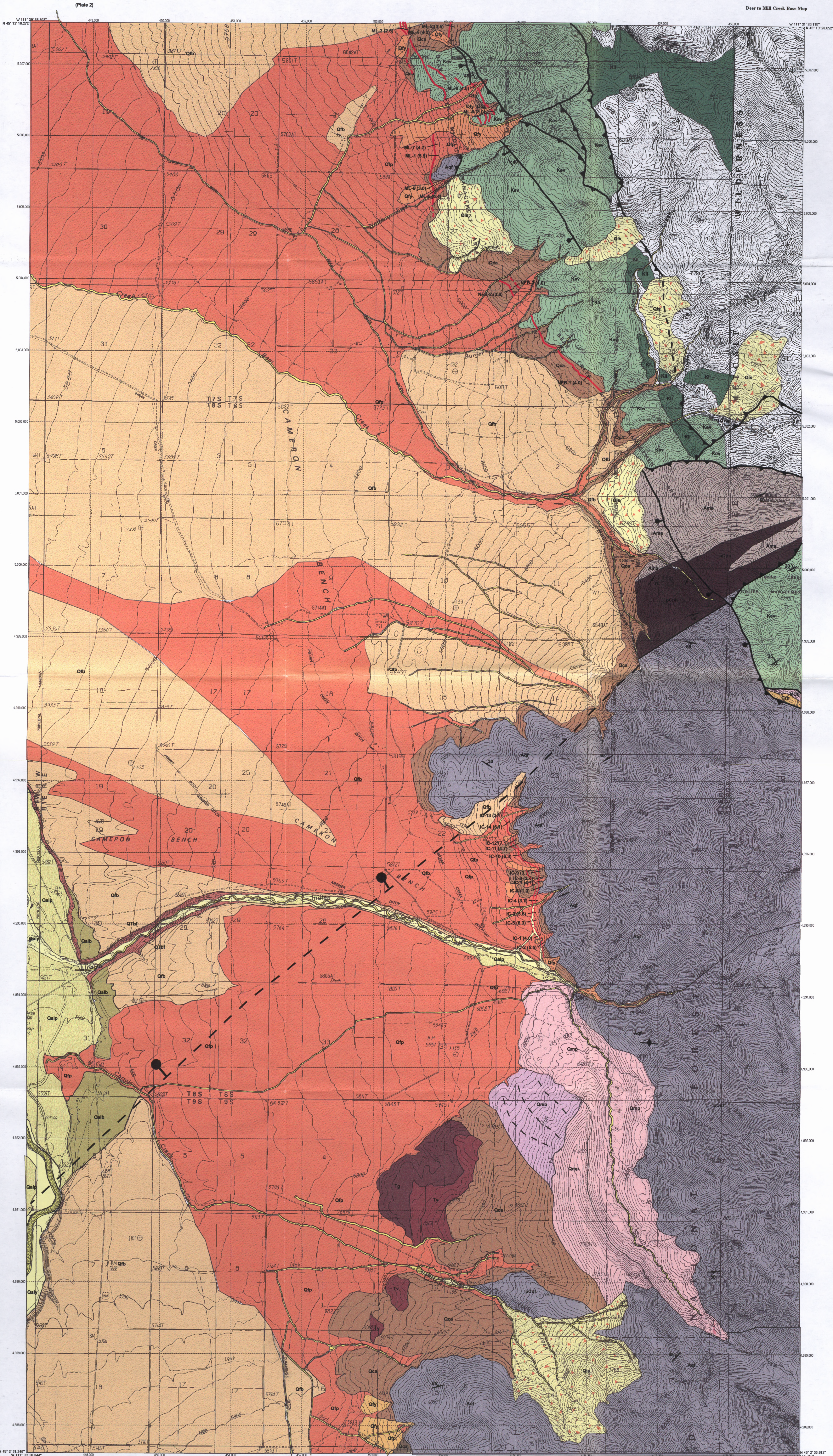
Surficial Geology of the Madison Range fault from Mile Creek to Watkins Creek
(Plate 1)



List of Map Units

- Qal - Mid to Late Holocene Alluvium
- Qaly - Early to Mid Holocene Alluvium
- Qalp - Late Pleistocene Alluvium (Pinedale equivalent)
- Qalb - Mid Pleistocene Alluvium (Bull Lake equivalent)
- Qly - Early to Mid Holocene Fan Alluvium
- Qlp - Late Pleistocene Fan Alluvium (Pinedale equivalent)
- Qlb - Mid Pleistocene Fan Alluvium
- Qls - Mid-Late Pleistocene to Holocene Landslide (Quake Lake slide 1959)
- Qmp - Pinedale Equivalent Moraine
- Qmb - Bull Lake Equivalent Moraine
- Qca - Alluvium/Colluvium undivided
- QTbf - Pliocene to mid Pleistocene coarse- to fine-grained alluvium
- Thrt - Quaternary-Tertiary Huckleberry Ridge Tuff (2.0 MA)
- Tsmc - Tertiary Six Mile Creek Formation (Miocene-Pliocene)
- Tg - Volcaniclastic sandstone and gravel (Eocene)
- Tv - Andesite and basalt flows (Eocene?)
- Kd - Dacite porphyry of Fan and Lone Mountains (Late Cretaceous)
- Kl - Livingston Formation (Upper Cretaceous), undivided
- Kev - Everts Formation and Virgelle Sandstone (Upper Cretaceous), undivided
- Kc - Cody Shale (Upper Cretaceous)
- Kf - Frontier Formation (Upper Cretaceous)
- Kmo - Mowry Shale (Upper Cretaceous)
- Kmt - Muddy Sandstone and Thermopolis Shale (Lower Cretaceous), undivided
- Kk - Kootenai Formation (Lower Cretaceous)
- JTru - Morrison Formation (Upper Jurassic), Ellis Group (Upper and Middle Jurassic), and Woodside Siltstone and Dinwoody Formation (Lower Triassic), undivided
- Ps - Shedhorn Sandstone (Lower Permian)
- IPMqa - Quadrant Sandstone (Pennsylvanian), Amsden Group (Lower Pennsylvanian and Upper Mississippian), and Snowcrest Range Group (Upper Mississippian), undivided
- Mm - Madison Group, undivided (Upper and Lower Mississippian)
- MDlj - Three Forks Formation (Lower Mississippian and Upper Devonian) and Jefferson Formation (Upper Devonian), undivided
- OCrm - Bighorn Dolostone (?) (Ordovician), Red Lion Formation (Upper Cambrian), Pilgrim Dolostone (Upper Cambrian), and Park Shale (Middle Cambrian), undivided
- Crp - Red Lion Formation (Upper Cambrian), Pilgrim Dolostone (Upper Cambrian), Park Shale (Middle Cambrian), undivided
- Cm - Meagher Limestone (Middle Cambrian)
- Cwf - Woolsey Shale and Flathead Sandstone (Middle Cambrian)
- Xm - Proterozoic metasedimentary rocks
- pCph - Proterozoic or Archean phyllite
- Xmm - Proterozoic Madison Mylonite Zone
- Aam - amphibolite (Archean)
- Abs - biotite schist (Archean)
- Acs - chloritic schist (Archean)
- Adm - dolomitic marble (Archean)
- Ama - marble (Archean)
- Amqs - muscovite-quartz schist (Archean)
- Aq - Quartzite
- Aqc - Chloritic Quartzite
- Aqf - Quartz-Feldspathic Gneiss
- At - Mylonite of the Crooked Creek Shear Zone





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