Soil development, morphometry, and scrap morphology of fluvial terraces at Jack Creek, Southwestern Montana
by James Paul Bearzi

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
Analysis of exceptionally well-displayed fluvial terraces and pediments along lower Jack Creek, a modest gradient (0.020) tributary of the Madison River in southwestern Montana, reveals a Late Quaternary chronology of terrace development. Soil stratigraphy of the surfaces reveals two distinct populations. "Higher Group" surfaces (2 terraces and 1 pediment, 40 to 60 m above Jack Creek) are mantled by a loess cap, contain stage III carbonate morphology, and were formed during pre-Pinedale time. All lower surfaces (7 terraces and 1 pediment), termed "Lower Group" surfaces, lack a loess cap, have remarkably similar weakly developed soils with stage I to stage II carbonate morphology, and were formed during late Pinedale and early post-glacial time. Deglaciation of the Madison Range, 15-12 ka (ka = thousand years) ago, initiated downcutting from the late Pinedale (highest Lower Group) terrace.

Mountain front tectonism was not responsible for Lower Group terrace formation; it may have contributed to the formation of Higher Group terraces. Post-glacial climate changes may have influenced the development of steep (~0.030 gradient) Lower Group terraces. Uplift of the Norris Hills throughout Quaternary time has resulted in intermittent damming and subsequent aggradation and degradation of the Madison River. Thus, base level fluctuation has been the primary terrace forming factor at Jack Creek. Jack Creek has been aggrading to the modern floodplain for several thousand years.

Morphologic dating of terrace scarps at Jack Creek shows that slope processes that have operated on the scarps at Jack Creek are much more complex than can be modeled with the diffusion equation. Discriminant function scores define Lower Group scarps as Holocene and Late Pleistocene in age. Linear regression of scarp slope versus scarp height for individual scarps and for aspect groups reveals that scarp morphology at Jack Creek is greatly dependent on height' and aspect.

Jack Creek exhibits a detailed post-glacial terrace flight primarily because the Madison Valley is tectonically active and the Madison River still aggrading. Terraced landscapes in less tectonically active basins are usually of Pleistocene age.

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James Paul Bearzi

This thesis has been read by each member of the thesis committee and has been found to be satisfactory regarding content, English usage, format, citations, bibliographic style, and consistency, and is ready for submission to the College of Graduate Studies.

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ABSTRACT

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INTRODUCTION

The Problem

Fluvial terraces and mountain-front pediment surfaces are common along both the Madison River and many of its tributaries in southwestern Montana. The terraces are floodplains abandoned through stream downcutting and were formed during periods of equilibrium or threshold conditions of the parent stream (Bull, 1979). As such, they can be used to decipher a chronology of the evolution of the terraced landscape. An understanding of this kind of chronology as well as the factors controlling downcutting in the Madison Valley will shed considerable light on the Late Cenozoic evolution of the area.

Application of graded stream concepts (e.g., Gilbert, 1914; Mackin, 1948; Leopold and Bull, 1979) and terrace analysis in climatically and tectonically diverse areas have proven useful in reconstructing the Late Cenozoic evolution of several large basins (e.g., Ritter,1967; Palmquist, 1983; Reheis, 1984, 1987; Bull and Knuepfer, 1987). An exceptionally well-displayed flight of terraces and pediments along lower Jack Creek (Figure 1), a major tributary of the Madison River, is amenable to this type of analysis. Since the Jack Creek terraces are the best developed flight in the Madison Valley, a detailed history of landscape evolution can be pieced together at Jack Creek; this is probably not possible elsewhere.

The primary purposes of this study are to determine the timing and causes of major downcutting events at Jack Creek. Establishment of a relative chronosequence (timing) of terrace development at Jack Creek can be obtained by analyzing time-dependent characteristics of terrace development (e.g., soils). Soil development on terrace treads will generally increase with increasing height above the present stream (Birkeland, 1984). The causes of terrace formation can be determined by comparing the timing of terrace formation with that of potential terrace forming activity (i.e., tectonism, climate change, and base level fluctuation) in the Jack Creek area.

The interpretation of pediments is more complex. Pediments are essentially surfaces of transport during periods of basin stability (Mabbutt, 1977) and thus are more subject to reactivation than are fluvial terraces. Morphologically, range-front pediments differ from fluvial terraces in that their gradients decrease dramatically from the mountain zone to the pediment zone to the alluvial zone (Figure 2). Fluvial terraces are subparallel to the master stream throughout their length and have a much less pronounced concave-up profile. Pediments also lack a well-defined drainage basin upslope and are not directly related to nearby ephemeral or perennial streams (Johnson, 1932).

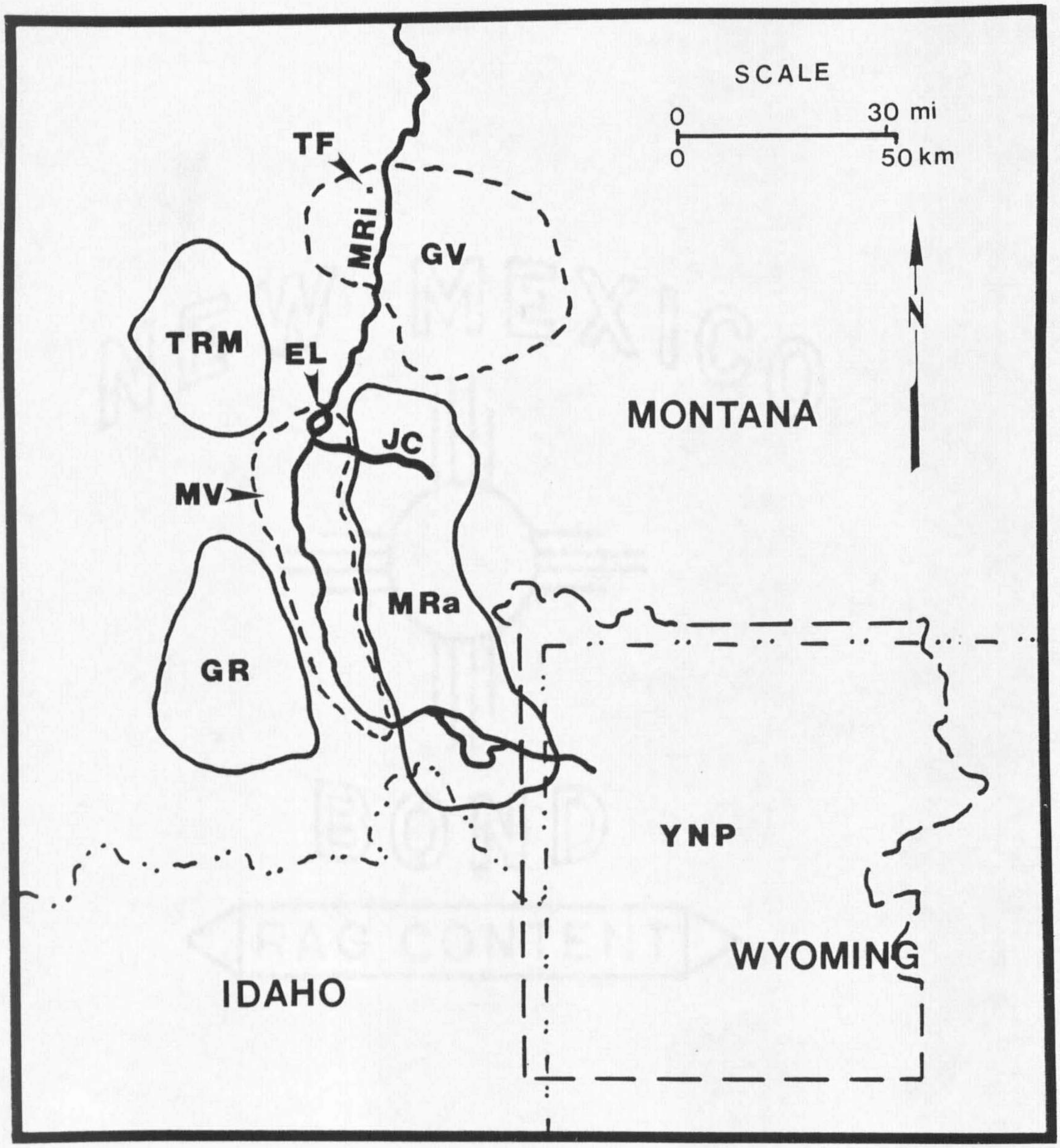

Figure 1. Geographic setting of Jack Creek in southwestern Montana. Solid lines bound mountain ranges; dashed lines bound valleys. JC = Jack Creek; MRi = Madison River; MV = Madison Valley; MRa = Madison Range; EL = Ennis Lake; GR = Gravelly Range; TRM = Tobacco Root Mountains; TF = Three Forks; GV = Gallatin Valley; YNP = Yellowstone National Park.

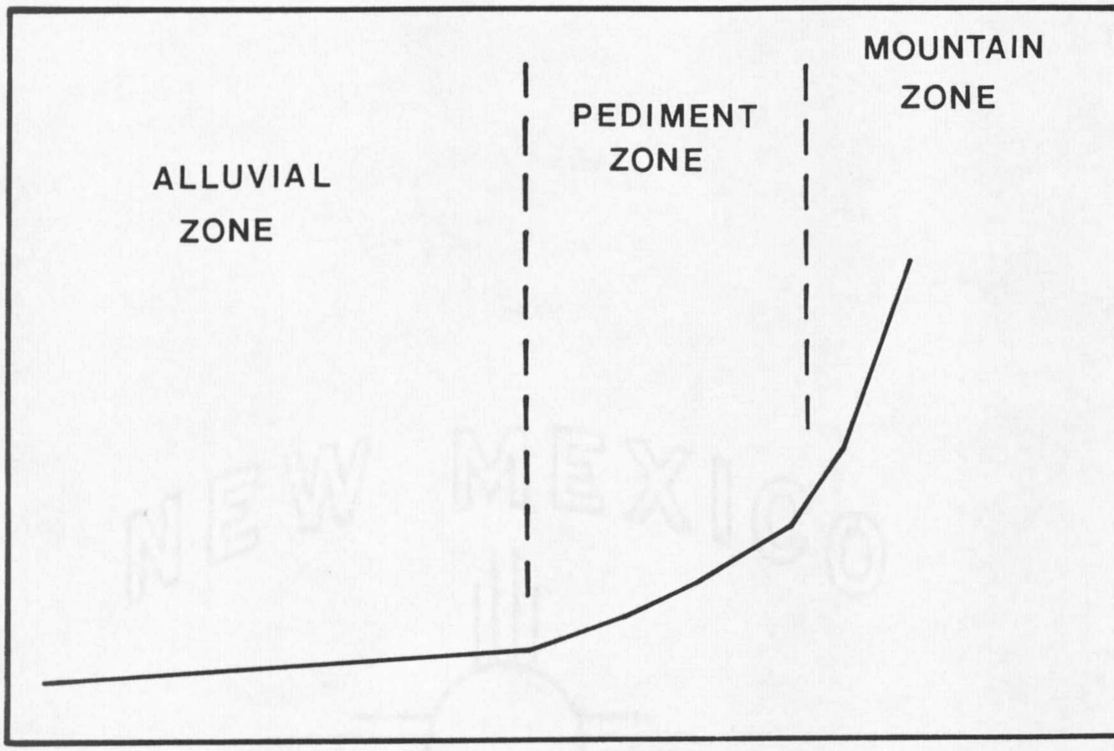

Figure 2. Generalized longitudinal profile of a pediment. Degradation, transport, and aggradation characterize the mountain, pediment, and alluvial zones, respectively. Modified from Mabbutt (1977).

Genetically, pediments differ from terraces in that they are not caused by simple downcutting of a fluvial system. Weathering with subsequent rill work and sheet flow (Bryan, 1925; Davis, 1938) and lateral planation of streams (Bryan, 1922; Howard, 1942) are the primary explanations of pediment formation. Since either hypothesis invokes pedimentation as an episodic process, pedogenesis may be interrupted and soils truncated by periods of sheetflood or planation, transport, and deposition on pediments. Thus, soils developed on pediments and terraces of similar age may be morphologically dissimilar.

Previous Work

Fluvial terraces in the Madison Valley were first noted by Peale (1896) during initial reconnaissance of the Three Forks (Montana) area. Most later work on Madison Valley terraces concerns the age of the most prominent fluvial surface in the Madison Valley, the Cameron Bench. The Cameron Bench has been described as both depositional and erosional (Paul and Lyons, 1960). Alden (1953) suggested that this surface is as old as Early Pleistocene while Montagne (1960) constrained it to post-Bull Lake but pre-Pinedale in age based on its relationship to Bull Lake (?) fan deposits in the valley. Schneider and Ritter (1987) suggested that the Cameron Bench is actually time transgressive, with entrenchment of the Madison River occurring in the northern part of the valley while aggradation or still-stand was still occurring in the southern part of the valley.

A flight of lower terraces in the Madison Valley was grouped by Alden (1953) and termed Intermediate terraces by Paul and Lyons (1960). These terraces represent still-stands of the Madison River and its tributaries after they cut through the Cameron Bench.

Detailed studies of Intermediate terraces have been restricted to the south part of the valley. Lundstrom and Burke (1985) identified three levels of paired glacial outwash terraces 3, 18, and 35 m above the modern floodplain. They are assigned Pinedale, Bull Lake, and pre-Bull Lake ages based on loess cap thicknesses, loess weathering, and carbonate buildup. More detailed work in the same area (Lundstrom, 1986) suggested reinterpretation of paired terraces at 3-15 m, 30 m, and 45 m above the modern river channel as Pinedale correlatives due to gross similarities in soil development. Minor increases in carbonate coating and loess cap thickness with increasing height above the modern stream suggested correlation with three different Pinedale stades. An unpaired terrace 60 m above the modern floodplain is correlated with Bull Lake glaciation due to a significantly thicker loess cap and thicker carbonate clast coatings.

The modern floodplain of the Madison River, called the Lower surface (Paul and Lyons, 1960), is less than 1.5 km wide in the southern part of the valley but widens to over 3 km in the northern part near Jack Creek. This is in part due to increased lateral cutting of the Madison River as it changes from single to braided channel morphology 20 km south of Ennis Lake. Further downstream, in Beartrap Canyon, the river returns to a single channel but flows on 100 feet (30 m) of alluvium, suggesting that the present floodplain is not the lowest level to which the Madison River has cut in the recent past (Paul and Lyons, 1960).

The Study Area

The terraces at Jack Creek range from Cameron Bench equivalents (based on projection from the highest Jack Creek terrace to the Cameron Bench) to the modern floodplain. At least seven intermediate terraces exist between these extremes. The study area includes all these surfaces and is bounded on the north by Ennis Lake and Jourdain Creek, on the west by the Madison River floodplain, on the south by Cedar Creek alluvial fan, and on the east by the Madison Range (Figure 3). Access is reasonably good since parts of all the terraces are one landowner's property; exposure of terrace stratigraphy, however, is poor and limited to manually excavated soil pits, irrigation ditches, and a few road cuts. Like lower Jack Creek, access in the Jack Creek basin is good since it lies almost entirely within the Madison Range and is owned by the federal government and private timbering concerns.

Geologic History

The Madison Valley formed as a result of post-Laramide Basin and Range-style extension, characterized by listric normal faulting with structurally controlled fault blocks and associated depressions. Early stages of basin formation were characterized by high rates of subsidence causing blocked and internal drainage (Fields and others, 1985) similar to that of today's Madison Valley near Ennis Lake (see Figure 3). Contemporaneous subsidence and sedimentation during early development (Kuenzi and Fields, 1971; Rasmussen and Fields, 1983) created broad, shallow basins. The Madison Valley was probably recognizable as a sediment depocenter by mid-Eocene time (Schmidt and others, 1984).

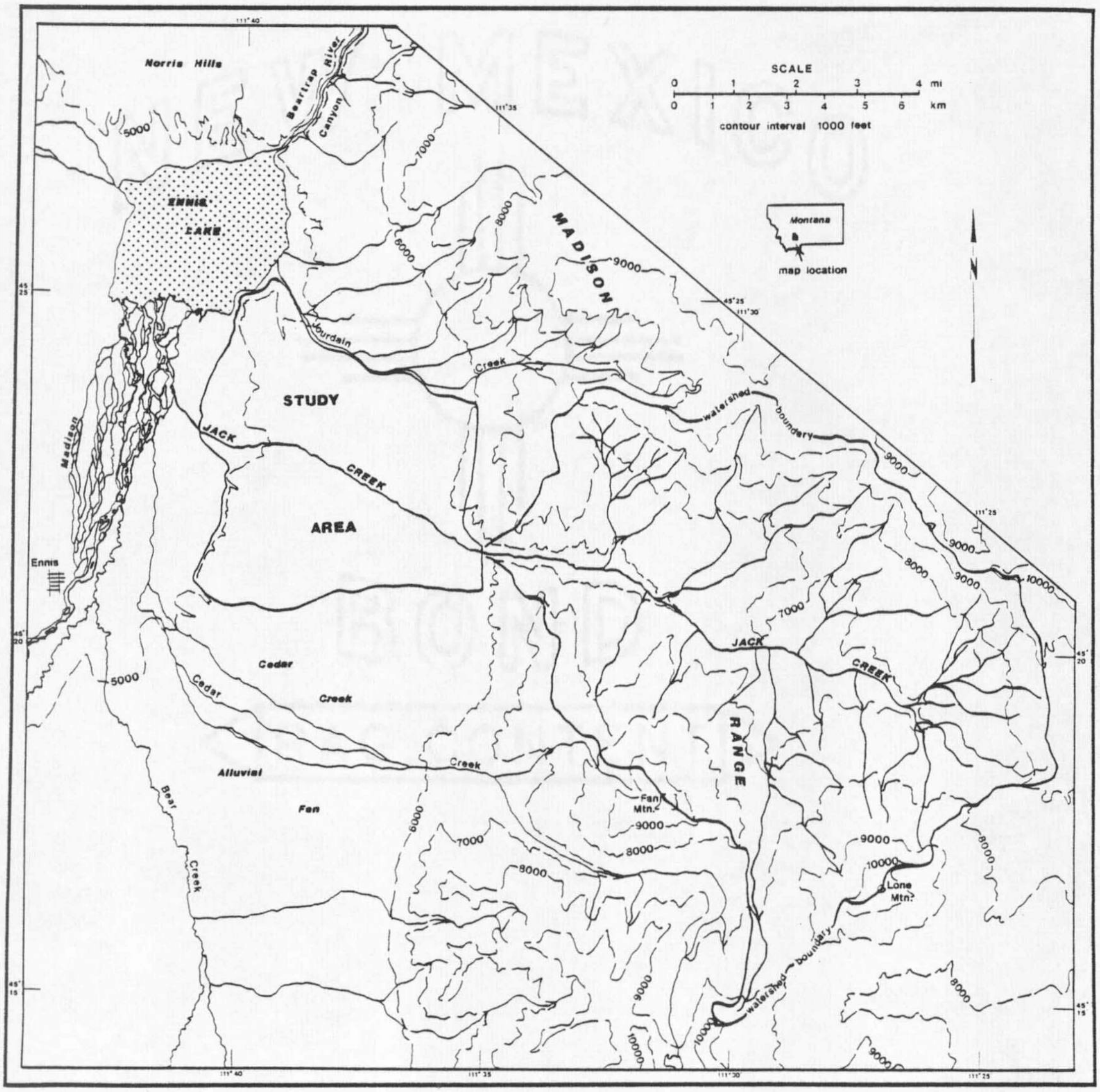

Figure 3. The study area at Jack Creek. Important locations mentioned in the text are labeled.

Tertiary sediments of the Madison Valley probably consist of Bozeman Group equivalents. The upper member of the Bozeman Group, the Sixmile Creek Formation, crops out throughout the Madison Valley and consists of coarse conglomerate deposited by high energy braided streams (Kuenzi and Fields, 1971). Sixmile Creek sediments accumulated until late Pliocene time when a period of exhumation began (Kuenzi and Fields, 1971). Gravity data reported by Rasmussen and Fields (1983) indicate 7,100 ft (2,200 m) of Cenozoic sediment in the Madison Valley north of Beartrap Canyon. From Ennis Lake to the south, the basin becomes progressively deeper (15,000 ft (4,600 m) or more). The Spanish Peaks fault separates these two parts of the basin (Tysdal, 1986) and provides local base level control for the upper Madison River.

Three major structural features in the western part of the Madison Range influence Jack Creek (Figure 4). First, at the junction of the Madison Range and the piedmont in the Madison Valley, Jack Creek crosses the Madison Range Fault System. This fault system is a series of north trending en echelon Cenozoic normal faults (Swanson, 1950; Schneider, 1985; Tysdal, 1986; Tysdal and others, 1986) which controls the topography and position of the western mountain front of the Madison Range. Structural relief provided by this range bounding fault system exceeds 3,000 m. Thus, Precambrian and Phanerozoic rocks and associated structures exposed in the Madison Range probably exist beneath the floor of the Madison Valley covered by hundreds of meters of Tertiary sediment.

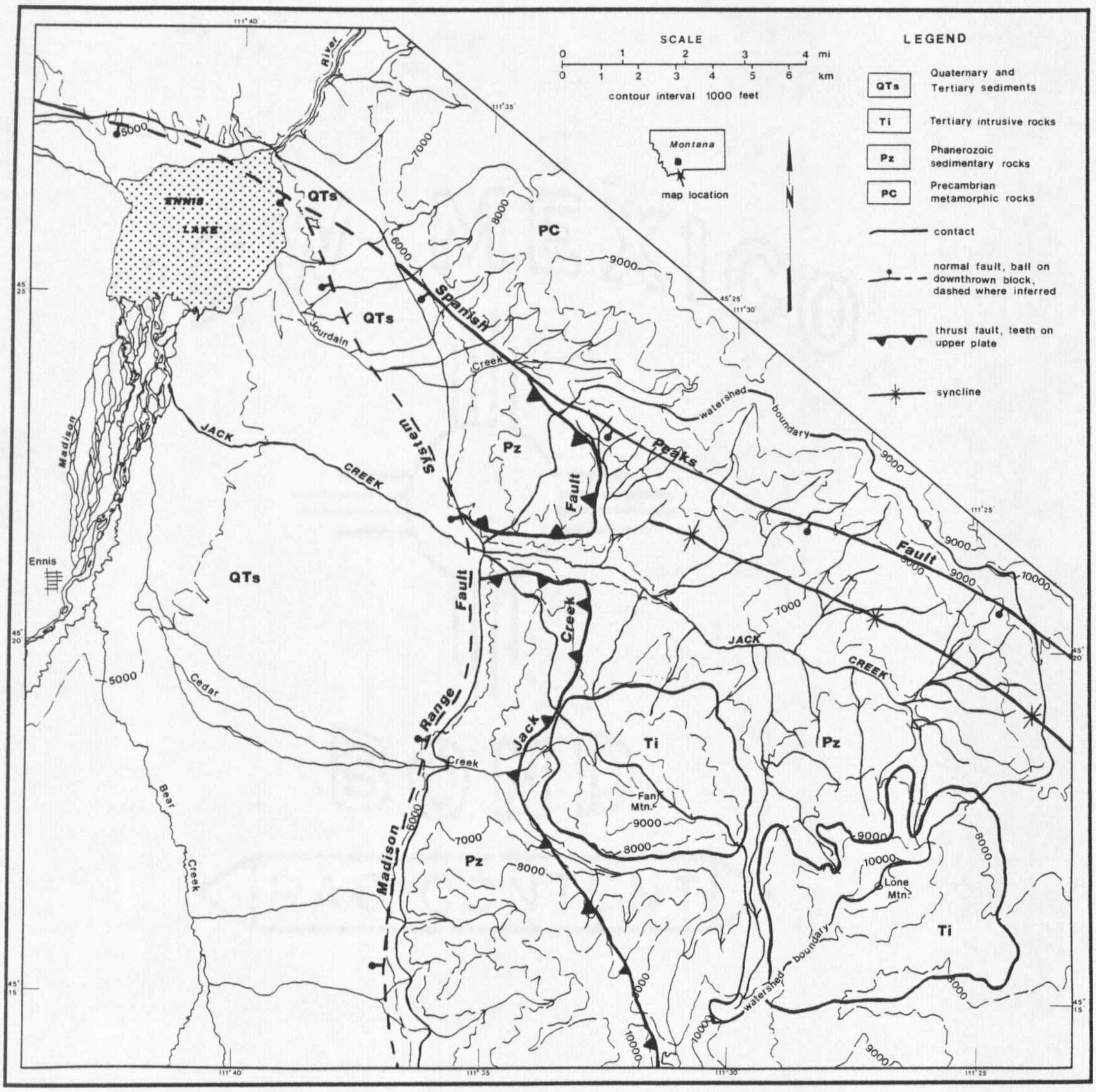

Figure 4. Generalized geology and structure of the Jack Creek area. Sources: Swanson (1950), Grabb (1977), and Tysdal (1986).

The second structural feature influencing Jack Creek is the Spanish Peaks fault (see Figure 4). This prominent down-to-the-south northwest trending structure juxtaposes Precambrian crystalline rocks to the north with a thick Phanerozoic cover to the south (Tysdal, 1986) in the Jack Creek drainage. It also is probably responsible for the relative upwarping of the Norris Hills / Beartrap Canyon area (Montagne, 1960). This area of uplifted topography provides local base level control for the Madison Valley at Ennis Lake. The Madison River in turn is the local base level for Jack Creek. Movement along the Spanish Peaks fault (i.e., uplift of the Norris Hills) changes base level for the Madison River. Adjustment of the Madison River causes concomitant base level changes for Jack Creek.

The Spanish Peaks fault is thought to have a Precambrian ancestry (Schmidt and Garihan, 1983) with later reactivation during the Laramide Orogeny (Tysdal, 1986) and quite possibly the Quaternary. Indeed, on July 21, 1987, an earthquake registering 4.5 on the Richter Scale occurred in southwestern Montana. Its epicenter was $45^{\circ}29.22'$N $111^{\circ}36.19'$W (in the Norris Hills / Beartrap Canyon area) at .39 km depth. Although no surface rupture was reported, the location of the epicenter near the Spanish Peaks fault is eloquent testimony to active tectonism in the Norris Hills and recent fluctuation of the base level for Jack Creek (i.e., the Madison River).

Finally, Laramide-age thrust faults of the Hilgard system (Tysdal, 1986), known as the Jack Creek fault in the

Jack Creek area (Swanson, 1950), thrust Lower and Middle Paleozoic strata eastward over Mesozoic rocks. Stream evolution was influenced by drainage development over structural weaknesses and lithologies of varying resistance. As a result, most tributaries of Jack Creek have developed along strike. Jack Creek appears to have adjusted fully to the different lithologies since no nickpoints exist on the stream within the drainage basin.

The structural features discussed above control the distribution and types of lithologies present in Jack Creek basin (see Figure 4). In the Madison Range, four groups of rocks are exposed: Precambrian crystalline rocks, Phanerozoic sedimentary rocks, Phanerozoic intrusive rocks, and Quaternary and Tertiary sediments.

Precambrian gneiss, schist, and intensely sheared rocks crop out in the northern part of Jack Creek basin along the northern drainage divide. Their presence is the result of relative upward movement of about 4,900 m along the Spanish Peaks fault (Tysdal, 1986; Tysdal and others, 1986).

A large west-trending syncline immediately south of the Spanish Peaks fault is a drag fold genetically related to Laramide movement along the Spanish Peaks fault (Garihan and others, 1983). The syncline consists of a thick section of Paleozoic and Mesozoic sedimentary rocks including sandstone, shale, limestone, and dolomite. Additionally, the Jack Creek fault crops out in Jack Creek basin and is deformed by drag folding along the Spanish Peaks fault. The age of Laramide deformation along the Jack Creek fault is younger than that of the Spanish Peaks fault (Garihan and others, 1983; Tysdal and others, 1986).

The sedimentary rocks in the vicinity of Fan and Lone Mountains were intruded by laccolithic rocks after structures related to the Hilgard system were formed (Becraft and others, 1970). These rocks are mainly andesites and porphyritic dacites (Swanson, 1950; Tysdal and others, 1986) and are largely undeformed.

Quaternary deposits mantle much of upper Jack Creek basin (Figure 5). Glacial, periglacial, alluvial, and mass movement processes all contributed to their deposition. Tills of both Pinedale and Bull Lake glaciations have been identified (Hall, 1960a; Grabb, 1977). Periglacial deposits in the form of rock glaciers and geliflection lobes exist on both Fan and Lone Mountains. Dip slopes of Cretaceous shales have caused mass movement to dominate the post-glacial Quaternary record in Jack Creek basin (Hall, 1960b).

Climatic History

Climate change in latest Pliocene time initiated downcutting of Tertiary basins through Bozeman Group equivalent sediments in southwest Montana. This episode of downcutting is thought to be related to a relatively warm wet climate (Thompson and others, 1982) compared to that of Middle Pliocene time. The onset of Pleistocene glaciation in North America, however, probably ended this period of downcutting by ushering in a cooler, possibly humid climate interspersed with relatively warm, possibly dry interglacial periods.

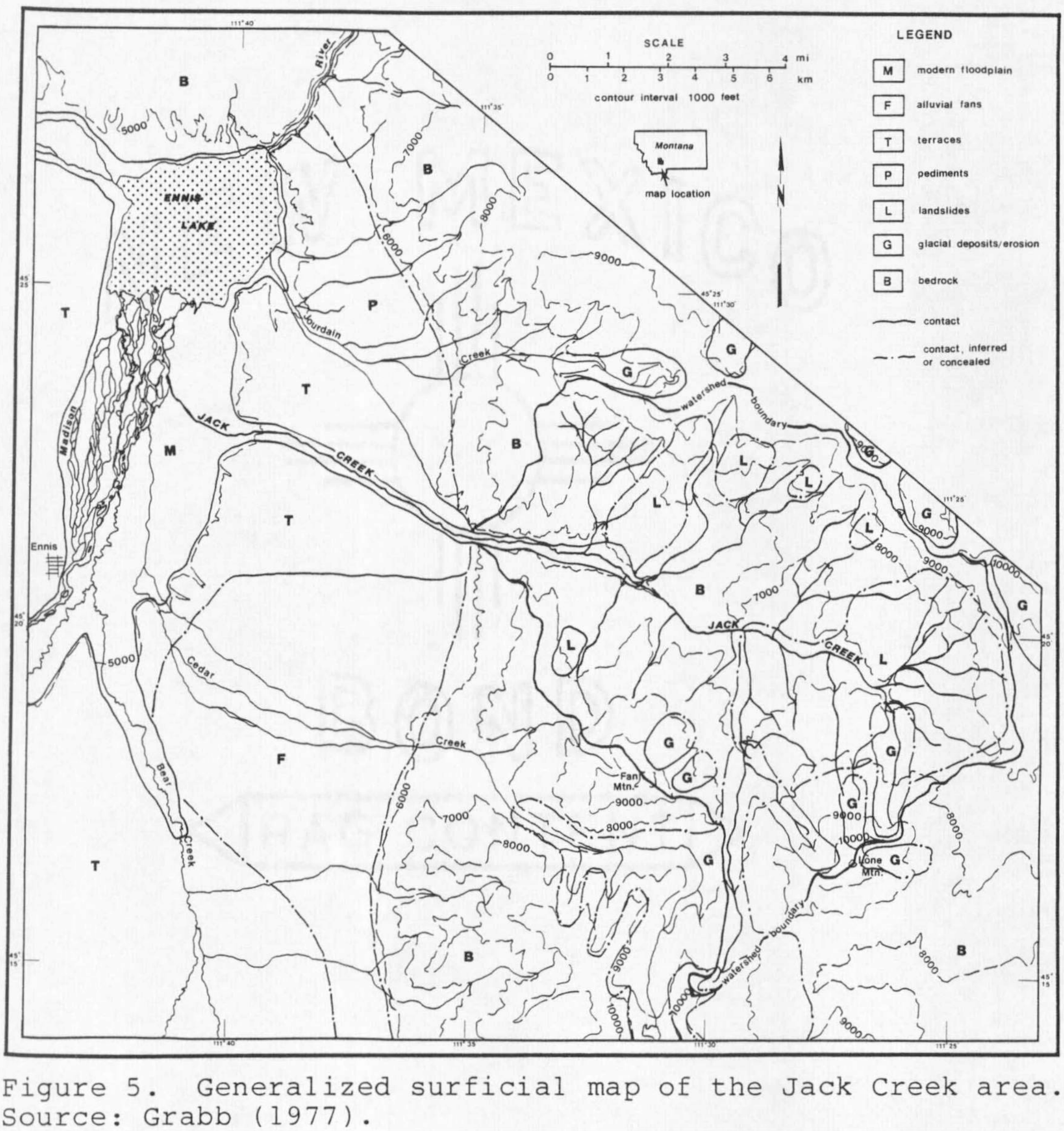

Figure 5. Generalized surficial map of the Jack Creek area. Source: Grabb (1977).

At least three Pleistocene glacial climatic episodes in southwestern Montana have been identified on the basis of moraine and outwash deposits. Bull Lake deposits have been identified in the West Yellowstone basin (Pierce and others, 1976), the Madison Range (Hall, 1960a), and in Jack Creek basin (Grabb, 1977). This major advance occurred approximately 150-140 ka (ka = thousands of years) ago and is correlative with the Illinoian advance of the Laurentide ice mass (Porter and others, 1983).

A post-Bull Lake interglacial period starting 125-118 ka ago (Barry, 1983), correlative with the post-Illinoian Sangamon interglacial stage, was characterized by a warmer, possibly drier climate compared to that of glacial times (Emiliani, 1972). In Yellowstone Park, Richmond (1986) has identified a post-Sangamon but pre-Wisconsin glacial episode, the Eowisconsin, 120-110 ka and 90-80 ka ago.

The most recent Pleistocene glacial episode in southwestern Montana has been identified through Pinedale equivalent (35-30 ka old) deposits (Hall, 1960a; Pierce and others, 1976) generally found nested within Bull Lake deposits (see Pierce (1979) for an exception). Pinedale glaciation represents the final episode of Pleistocene glaciation in North America. Deglaciation of the Madison Range was probably synchronous with much of the Rocky Mountains -- approximately 15-12 ka ago (Colman and Pierce, 1986; Madole, 1986); the range has been ice-free since.

The climate of the Holocene Epoch (the last 10 ka) is thought of as another interglacial period similar to the

Sangamon interglacial (Barry, 1983) -- warmer than that of glacial climates. Descriptions of Holocene climate have centered around the idea of a post-glacial thermal optimum, called the Hypsithermal, Altithermal, or climatic optimum, based on the probable former extent of grassland vegetation communities into areas now containing forest communities (Deevey and Flint, 1957; Hopkins, 1975). Leopold (1951) and Leopold and Miller (1954) envision this period of probable global warming, hereafter called the Altithermal, as having similar mean annual precipitation as today. The frequency of small rains, however, was less during the Altithermal. Thus, the influence of large convective storms in the summer was increased (e.g., increased arroyo cutting) without a change in the total amount of precipitation. Pollen data suggest the Altithermal occurred 7.5-4.5 ka ago in Yellowstone National Park (R. Baker, 1983).

Three small glacial advances occurred during the otherwise warm, dry Holocene. One occurred 9-8 ka ago, prior to the Altithermal (Beget, 1983; Burke and Birkeland, 1983). The two advances which occurred after the Altithermal are grouped together and termed Neoglaciation (Porter and Denton, 1967; Hopkins, 1975). Neoglacial advances in the Rocky Mountains occurred 4.5-3 ka ago and 2-1 ka ago (Porter and Denton, 1967; Burke and Birkeland, 1983).

The terraces at Jack Creek have probably never supported a forest cover during Holocene time (Sindelar, 1971). Thus, the climate changes that caused Holocene glaciation were apparently not extreme enough to severely alter vegetation communities or soil types on the valley side of the mountain front. Today, the dominant vegetation types along lower Jack Creek are grasses. Woody plants dominate the modern floodplain. The soils are of three orders: Aridisols, Mollisols, and floodplain Entisols. Mean annual precipitation is 11 to 14 in (29 to 36 cm) and mean annual temperature is 44°F (7°C) (Soil Conservation Service, 1987). Most precipitation falls as a result of convective storms in the summer.

Fluvial Response to Terrace Forming Factors

Tectonic activity and climate change are two primary influences on stream stability (Bull, 1984). Rivers are sensitive to variations in valley floor slope which are caused not only by offset along faults (Wallace, 1967) but also by gradual aseismic deformation (Ouchi, 1985; Schumm, 1986). Climate change influences channel morphology by affecting the load and discharge parameters of stream equilibrium (Leopold and Maddock, 1953). Streams are also influenced by an indirect manifestation of tectonic or climate change: base level fluctuation. Base level for a stream can be established through extrabasinal as well as intrabasinal tectonic or climatic variation.

The response of the fluvial system to a stimulus provided by tectonic activity, climate change, or base level fluctuation is ultimately controlled by the power of the stream. Stream power is defined as the time rate of potential energy expenditure as water flows downslope in a channel (Rhoads, 1987). Power refers to the capacity of flowing water to do work and is directly related to discharge (Rhoads, 1987). The stream will be able to more effectively carry its load once the power exceeds a critical threshold of transport (Bull, 1977). Any power beyond that threshold will allow the stream to downcut. Similarly, a decrease in power below the critical threshold of transport will cause aggradation.

Variations in stream power will cause concomitant changes in stream gradient (Hack, 1973). Stream competence, as a measure of stream power, is directly related to the product of the channel slope and length along a reach (Hack, 1973). Since terraces are abandoned floodplains and so record the gradient of the river when it flowed at the terrace level, terrace gradient is similarly related to the competence of the river before abandonment.

Tectonic Activity

Relative uplift of fault-generated landscapes by movement along range bounding normal faults is manifested in adjustments in the fluvial systems which cross them (Bull, 1973). These manifestations include stream downcutting within mountains, piedmont aggradation, and piedmont degradation (Bull and McFadden, 1977). Net downcutting of the piedmont implies either quiescence of the range bounding normal fault or increased channel downcutting in the mountains relative to piedmont degradation (Bull, 1984). Longitudinal profiles of piedmont terraces can be deformed by subsequent activity along a range-bounding fault, thus obscuring original relationships between fluvial and tectonic systems.

Active tectonism along mountain fronts is revealed by the presence of all or part of a typical landform assemblage. Highly active mountain fronts exhibit low mountain front sinuosity (Bull and McFadden, 1977; Mayer, 1986), triangular facets, V-shaped cross-valley profiles in bedrock (U-shaped in alluvium), and unentrenched aggradation surfaces at the piedmont. Relative uplift would be greater than combined channel downcutting and piedmont aggradation (Bull, 1984).

Moderately active mountain fronts may show cross-valley profiles similar to that of active mountain fronts. However, embayed mountain fronts, degraded triangular facets

(Mayer, 1986), and entrenched aggradation surfaces on the piedmont distinguish this assemblage from that of more active areas (Bull and McFadden, 1977). Channel downcutting in the mountains would be greater than both relative uplift and piedmont degradation (Bull, 1984).

Inactive mountain fronts generally show dissected embayments and piedmont surfaces. If piedmont degradation is greater than channel downcutting and relative uplift, however, many characteristics of active mountain fronts may be present (Bull, 1984).

Climate Change

While uplift causes local changes in slope and thus stream power, climate change primarily influences the discharge and load of a fluvial system. Discharge fundamentally affects the available stream power while load affects the critical stream power (i.e., threshold of bedload transport). Increases in discharge are generally the result of increased precipitation in the drainage basin or increased size of the drainage basin. The associated increase in power will result in greater competence and capacity of the stream.

Glaciation greatly influences stream power by increasing bedload relative to discharge in the meltwater systems (Chorley and others, 1984). Aggradation will generally occur (V. Baker, 1983). Aggrading outwash surfaces form as plains, valley trains, or fans.

Glaciofluvial systems are further characterized by development of braided channel morphologies and gradients steeper than their non-glacial counterparts to transport the increased bedload (Leopold and Maddock, 1953).

The transition between glacial and non-glacial climates also greatly affects the fluvial system. Deglaciation is thought to occur relatively quickly (Thompson and Jones, 1986). In small fluvial systems, it affects downstream reaches by increasing discharge relative to sediment load. Stream power is effectively increased beyond the point needed to transport the load so the stream experiences degradation. Large fluvial systems are affected by glacial meltwater to a minimal extent (Pierce, 1979).

Non-glacial climate changes similarly influence the fluvial system. In moderately humid climates where vegetation cover is not extreme, discharge is high. Thus, stream power is also greater and stream downcutting is common (Brackenridge, 1980; Knox, 1983). Stream gradients tend to be lower (Hack, 1957, 1973) and channels adjust by attaining a meandering channel morphology (Schumm, 1977) given constant base level. Fluvial systems in arid climates are characterized by lower stream power. Steeper gradients are needed to transport similar loads out of the drainage basin (Knox, 1983). No excess stream power exists so aggradation is common.

Base Level Fluctuation

The third primary factor that influences stream adjustment is that of base level change. The two major types of base level change that affect streams are vertical and horizontal.

Vertical changes in base level occur due to sea level fluctuation (absolute base level) or aggradation and degradation of a master stream (local base level) (Figure 6). Absolute base level is controlled by eustatic changes in sea level and uplift and subsidence on a plate-wide scale. Local base level is controlled by extrabasinal factors such as local damming of a master stream or climatic and tectonic factors that directly influence the master stream or the tributary in question.

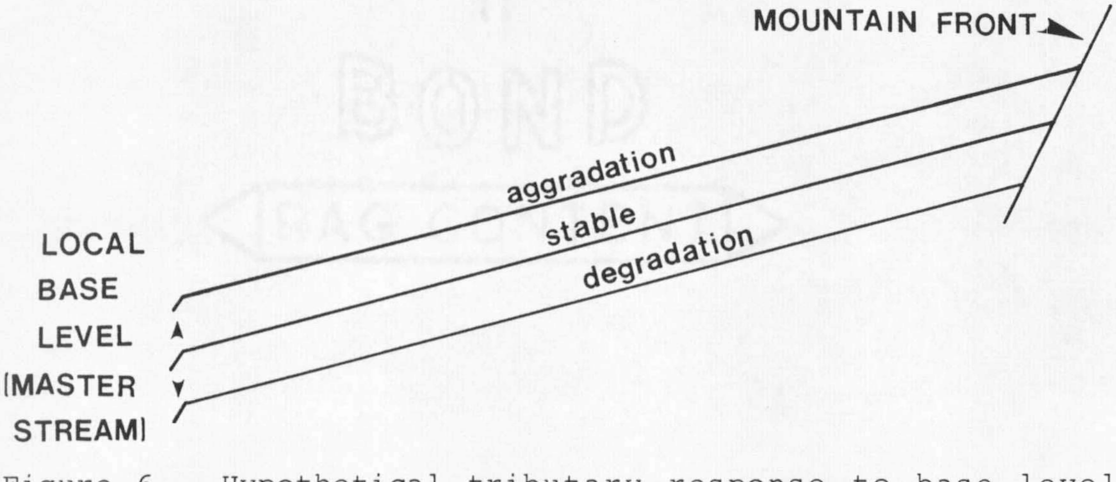

Figure 6. Hypothetical tributary response to base level change of the master stream.

Horizontal changes in base level cause undercutting of the confluence of a tributary stream with the master stream.

Channel migration or avulsion of a master stream can undercut tributary streams. This effectively lowers base level (Figure 7). Channel migration in the opposite direction can effectively raise base level, creating an aggradational response in the fluvial system. Vertical changes may develop paired abandoned surfaces, however, while horizontal changes may not. Additionally, base level changes will not be reflected in appreciably different stream gradients before and after base level change given near complete relaxation time since the stream is not responding to changes in its hydraulic regimen.

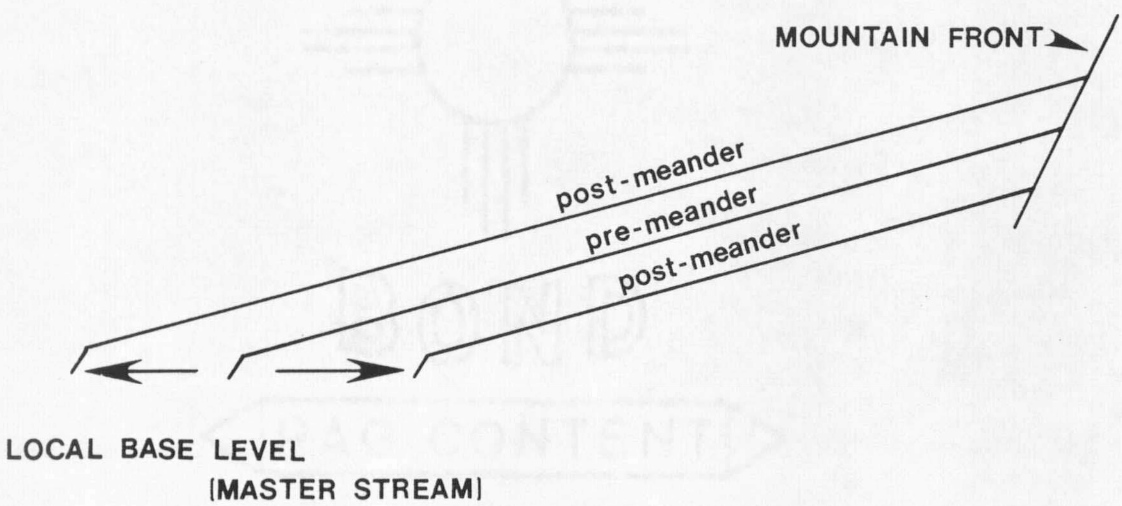

Figure 7. Hypothetical tributary response to lateral migration of the master stream.

The presence of terraces at Jack Creek shows that the stream has responded to external stimuli -- tectonism, climate change, and/or base level fluctuation -- by downcutting. The timing of a stimulus can be established by determining the timing of the stream response, in this case, floodplain abandonment. The methodology of this study reflects this relationship between stimulus and response.

METHODS

The use of terraces in analyzing fluvial landscapes is well established. The presence of terraces implies net downcutting of the parent stream and subsequent exposure of the terrace treads and scarps to pedogenic and slope processes. Methods used in this study were chosen to reveal differences in terrace morphology (height above Jack Creek and longitudinal terrace gradient) and age-dependent characteristics (tread soil development and scarp evolution) of the terraces at Jack Creek.

Terrace Morphology

Much can be learned about terrace development and terrace forming processes by observing morphology (see discussion of Terrace Forming Factors above). The terraces at Jack Creek were first defined by interpreting 1:24,000 topographic maps (Figure 8). Presence of the interpreted surfaces was later field checked and confirmed. Longitudinal profiles and terrace gradients (horizontal error = ± 40 ft; vertical error = ± 20 ft) were also determined using map interpretation. These longitudinal profiles reflect channel slope (Hack, 1957, 1973) when Jack

Creek was at still-stand (i.e., equilibrium) at each respective terrace level (Bull, 1984).

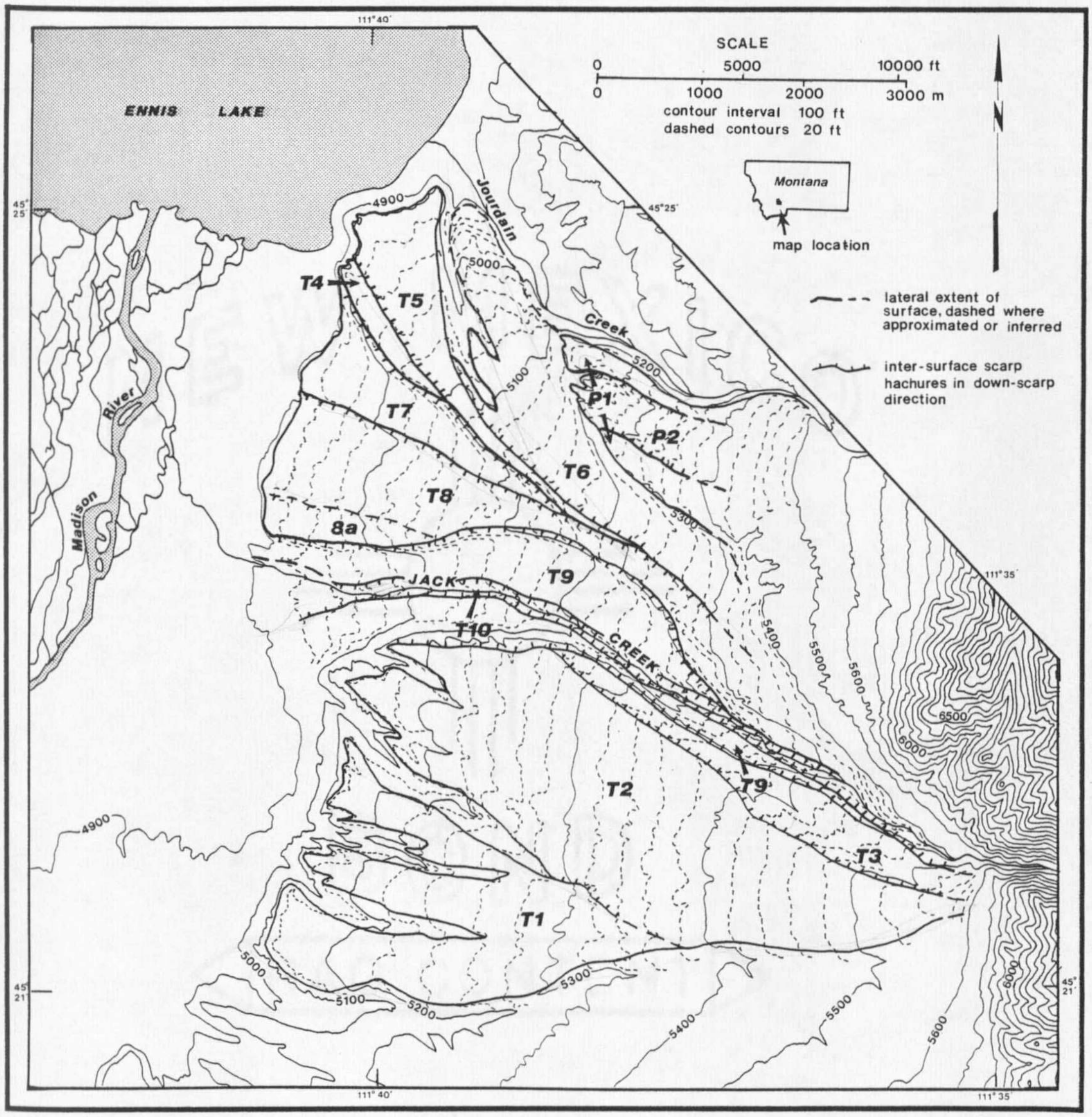

Figure 8. Map of the terraces at Jack Creek. T1 is the highest terrace above Jack Creek, T10 the lowest. P1 and P2 are the high and low pediment surfaces, respectively.

Soils Analysis of Treads

A relative chronosequence of terrace development can often be determined using soil analyses on terrace treads (e.g., Harden, 1982; Reheis, 1984). When a floodplain is abandoned (i.e., a terrace is formed), pedogenic processes begin to act on the floodplain material. Barring subsequent submergence, the degree of soil development is directly related to the time of terrace formation (Birkeland, 1984).

Two soil pits were excavated with pick and shovel on each surface at Jack Creek (Figure 9). Each pit was 1 m deep in order to expose most of the soil profile. The soil profiles were described using the soil taxonomy of Soil Survey Staff (1985). Carbonate stage was described using the criteria of Birkeland (1984). In addition, the lithology of 30 to 40 clasts from the C horizon was recorded to better understand the nature of the parent material. Weathering rind development of andesitic clasts was inhibited due to carbonate clast coatings (Grim, 1968) and so was not recorded.

Bulk density was determined in the field using a sand excavation method modified from McLintock (1959) and Cassidy (1981). Sub-rounded sand was sieved to obtain the medium sand size-range and its bulk density determined. Known weights of sand were then placed in containers for use in the field. Small excavations were made in the field of each major soil textural class found. A plexiglass plate with a 20 cm diameter hole cut in it was placed over the site to be measured and a 20 cm deep hole was excavated through the hole in the plate. This procedure minimized variation in the size and shape of the excavations. The excavated material was placed in a container of known weight and weighed. The hole was filled to the level of the plexiglass plate with the sieved sand. The remaining sand in the container was weighed to determine the volume needed to fill the hole and thus the bulk density of the excavated material. This is summarized in the equation:

$$BD_e = \frac{m_e BD_s}{m_o - m_r} \qquad (1)$$

where BD_e is the bulk density of the excavated soil, m_e is the weight of the excavated soil, BD_s is the bulk density of the sand, m_o is the original weight of the sand in the container, and m_r is the weight of the remaining sand in the container. It is necessary to know bulk density not only for characterization of the profile but also for analysis of soil carbonate content.

Samples of each profile were collected across 15 cm depth intervals for laboratory analysis. Coarse fragment (> 2 mm) weight-per cent was determined by sieving. The fine fraction (< 2 mm) was split to obtain lab samples of 1 to 3 gm, each precisely weighed. Carbonate content of each sample was determined using the Chittick apparatus (AOAC, 1950; Dreimanis, 1962) and then converted to grams per square centimeter column through the profile (Machette, 1978, 1985). Between five and ten splits were analyzed per sample to obtain maximum precision. Accuracy was limited by variability of calcium carbonate content within surfaces.

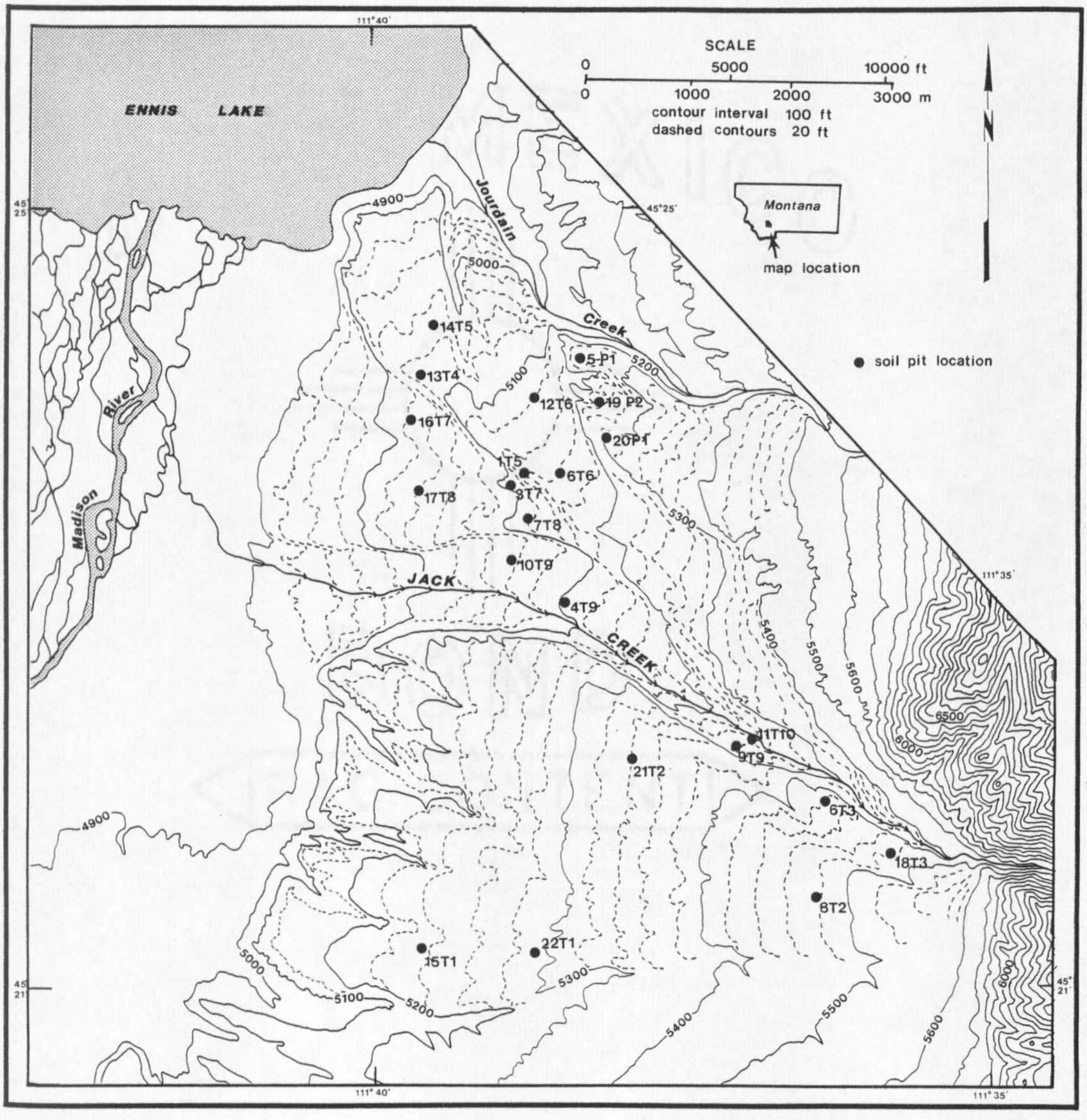

Figure 9. Locations of excavations for soil profile descriptions and analyses.

Profile Analysis of Scarps

The degradation of escarpments with time has long been recognized. Early studies of Lake Bonneville shorelines (Gilbert, 1890) revealed that the present morphology of wave-cut landforms was time-dependent; older slopes are more degraded. Bucknam and Anderson (1979) were the first to show that maximum scarp angle may be directly related to the logarithm of scarp height (scarp height is defined as the vertical distance between two adjacent terraces along a survey line) and that the slope of the regression line may be related to scarp age (Figure 10). Other studies of wave-cut (Hanks and others, 1984; Nash, 1980b), faulted (Wallace, 1977, 1980; Nash, 1980a; Mayer, 1984), and fluvial (Nash, 1984; Pierce and Colman, 1986) scarps have attempted to quantify the relationship between scarp morphology and time. This relationship has been used to relative age-date (RD) many Quaternary scarps and scarp-forming events. This RD technique is known as morphologic dating.

Theory of Morphologic Dating

Morphologic dating of scarps involves making several assumptions regarding both the initial morphology of the scarp and dominant processes acting to modify the slope since its formation. These assumptions severely limit the method, and care must be taken to clearly identify the assumptions.