Stratigraphy and sedimentation of the Proterozoic Garnet Range formation Belt Supergroup western Montana

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STRATIGRAPHY AND SEDIMENTATION OF THE PROTEROZOIC
GARNET RANGE FORMATION, BELT SUPERGROUP,
WESTERN MONTANA

By

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The Middle Proterozoic Garnet Range Formation of western Montana comprises the third and uppermost first-order transgressive sequence in the Belt Supergroup. It was deposited as interbedded sandstone and mudstone in marine marginal, nearshore, and offshore environments on a broad, shallow marine platform. Sedimentary structures and textures distinguish four lithofacies. In ascending order, these are: 1) a thinly interlayered sandstone-mudstone facies, deposited in intertidal and subtidal settings; 2) a planar bedded sandstone facies with thin mud partings, probably formed on and between upper shoreface shoals; 3) a rhythmically bedded sandstone-mudstone facies with hummocky bedding, deposited on the lower shoreface during and between storms; and 4) a rhythmically bedded facies similar to facies 3, but with thinner mudstone interbeds and polymodally oriented megaripples, formed on the upper shoreface, in tidal channels, and on beaches.

Commonly, the micaceous, fine grained sandstone portion of facies 3 is tripartite—grading from plane bedded at the base to hummocky bedded at the top, and capped by wave ripples. This sequence suggests that storm currents began as largely unidirectional, gradually changed to largely oscillatory, and then waned.

The Garnet Range represents the only documented storm dominated siliciclastic shelf deposits in approximately 15km of Belt rocks. Perhaps this signals the first wholesale marine inundation of the Belt Basin late in its history. Submergence or tectonic translation of a western landmass may have caused this inundation.
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INTRODUCTION

The Proterozoic Belt Supergroup of northwest Montana, northern Idaho, northeast Washington, and southeast British Columbia contains in its upper part the Garnet Range Formation, a green siliciclastic unit (Figs. 1, 2). The Garnet Range, although included in the Missoula Group, bears little resemblance to those rocks. Whereas Missoula Group Formations from the Snowslip through MacNamara are mostly light green to red, mudcracked, tabular bedded, and deficient in detrital mica, the Garnet Range is green to black, tabular to lenticular bedded, and contains abundant detrital mica and few mudcracks. Smith and Barnes (1966) noted similarities between the Garnet Range and Prichard and Wallace Formations and recognized it as the third and uppermost first order transgressive sequence in the Belt. Bleiwas (1977) believed the detrital mica within the Garnet Range signalled the arrival of sediment from a new source terrain. He and Winston (pers. comm., 1983) proposed that an unconformity may separate the Garnet Range from underlying Missoula Group strata.

The aims of this study are to: a) delineate the internal stratigraphy of the Garnet Range at its type locality for use in intrabasinal correlation, b) interpret the processes and environments of deposition for the Garnet Range based on an examination and analysis of primary sedimentary structures and textures, and c) infer the roles that large scale climatic, oceanographic, and tectonic processes played late in the evolution of the Belt Basin.
Figure 1. Location map of Belt Basin and Garnet Range type locality. After Harrison (1972).
Figure 2. Stratigraphy of the Belt Supergroup near Missoula, Montana. Curve shows first order transgressions (T) and regressions (R). After Harrison (1972).
Stratigraphic studies within the Missoula Group have heretofore concentrated mostly in the red fluval/alluvial deposits of the Mount Sheilds and Bonner Formations (e.g., Winston, 1977, 1978; Quattlebaum 1980; Slover, 1982, Barlow, 1983). The study described herein represents the first detailed stratigraphic work undertaken in the Garnet Range. Although earlier workers agreed that the Garnet Range was deposited seaward from a strandline, they variously interpreted the depositional environment as submergent fan delta (Bleiwas, 1977) and lagoon (Schmidt et al., 1983).

I have restricted this study to the type locality for several reasons. Despite structural complexity and generally sparse outcrops, there are some excellent exposures of the upper part of the Garnet Range at its type locality along the Blackfoot River. Also, the type locality is one of the few areas where the top of the Garnet Range and overlying Pilcher Quartzite have not been removed by pre-Middle Cambrian erosion. Finally, correlation with far removed sections awaits recognition and documentation of an internal stratigraphy within the unit; the type locality is the logical place to describe this internal stratigraphy.

This thesis is divided into three parts. Chapter One describes the stratigraphy and sedimentation of the Garnet Range. In Chapter Two, I synthesize Garnet Range stratigraphy and discuss possible implications for the late history of the Belt Basin. Chapter Three is a detailed, process oriented treatment of storms and their role in forming sandstone-mudstone cycles and hummocky cross stratification on
the lower shoreface during Garnet Range deposition. This chapter is an extension of the Facies Three interpretation presented in Chapter One. It has been separated to allow those readers interested only in Garnet Range stratigraphy and depositional environments to assimilate that information in a single chapter.
Regional distribution, correlation, and thickness

Clapp and Deiss (1931) named the Garnet Range Formation for the extensive exposures along the Blackfoot River at the northwest end of the Garnet Range, several kilometers northeast of Bonner, Montana (Fig. 3). In eastern and southern parts of the Belt Basin, the Garnet Range is underlain by red and green argillite, siltite, and quartzite of the MacNamara Formation. North of Thompson Falls, red beds of the MacNamara pass to green argillite, making it difficult to distinguish from the Garnet Range. The two formations have been grouped as the Libby Formation in that area (Winston, 1984). Purple, red, and white cross bedded and plane laminated quartzite of the Pilcher Quartzite conformably overlies the Garnet Range in and around its type locality near Missoula. The Pilcher is absent in eastern, northern, and western sections of the basin, where the Middle Cambrian Flathead Quartzite rests disconformably atop the Garnet Range and other Belt strata (McGill and Sommers, 1967). On the eastern margin of the basin, this unconformity cuts downsection into rocks of the lower Belt, making it, in effect, a regional angular unconformity. Where the Pilcher is preserved above the Garnet Range, the nature of its contact with the overlying Flathead is disputed (compare Winston, 1977; Elston and Bressler, 1980; Elston, 1984; Illich, 1966; Obradovich and
Figure 3. Geologic map (generalized) with Garnet Range in stippled pattern. Letters show locations of sections measured in this study. From west to east: EC=Eddy Creek, EM=Ellis Mountain, NM=Ninemile, SP=Stuart Peak, PC=Pilcher Creek, JG=Johnson Gulch, LC=LaFray Creek, WB=Wisherd Bridge. See appendix for precise locations. After Watson (1984), Wallace and Lidke (1980), Wells (1974).
Reported thicknesses of the Garnet Range vary widely. Clapp and Deiss (1931) estimated 7600 feet (2316 m) in the type locality along the Blackfoot River, but measured across several faults and folds that repeated the section. Nelson and Dobell (1961) later mapped the Bonner Quadrangle, which encompasses the type locality, and calculated a thickness of about 1800 feet (549 m). Wells (1974), who later mapped the Alberton Quadrangle, reported 8200 feet (2499 m) of Garnet Range in that area, yet presented no data to support his claim. Complicating the picture more, Wallace and Lidke (1980) calculated 5900 to 6900 feet (1798 m to 2103 m) from wilderness mapping that overlaps the Bonner Quadrangle map, yet they also failed to cite the basis of their calculation. Watson (1984), who remapped part of this problematic area in detail, calculated a thickness of about 3800 feet (1158 m) from balanced cross sections. Reliable estimates can be attained for thicknesses in areas west and northeast of the type locality, where structural complications are minimal and permit accurate computations from maps. In these areas, at Trout Creek, near Superior, and in the southern Lewis and Clark Range, thicknesses are approximately 3000 feet (914 m) and 1000 feet (305 m), respectively (Campbell, 1960; McGill and Sommers, 1967). The Flathead Quartzite unconformably overlies the Garnet Range in both areas. For purposes of this report, 1100 m (3500 feet) appears to be a reasonable approximation for the Garnet Range in the type area. This value was derived from recalculation on the maps of Wallace and Lidke, 1980; Wells, 1974; Nelson and Dobell, 1961; and
Watson, 1984; after field checking and minor reconnaissance mapping. There is currently no evidence for significant thickness change across any of the faults bisecting the study area.

In this study, I measured and analyzed eight partial stratigraphic sections in two principal areas (Fig. 3). Five closely spaced sections are located within the Wisherd Syncline in the Jocko Mountains north of Missoula, Montana. Three closely spaced sections are located near Alberton, Montana, 50km west of Missoula. In addition, reconnaissance studies were made northeast and northwest of the area, in the Lewis and Clark Range and near Superior, Montana, respectively. The two principal areas lie on opposite sides of the northwest trending Clark Fork fault. Several thrust faults also separate the areas. Most of these die into folds locally and are therefore not considered to have had large tectonic displacement.

Due to poor exposure and structural complications, the thickest section measured in this study is 273 meters. This thickness probably represents only a fourth or less of the Garnet Range section, which is nowhere completely exposed. Consequently, I constructed a single composite stratigraphic column for the area as a whole from partial measured sections (Fig. 4). Stratigraphic control employed in this method is based on published maps and minor reconnaissance mapping. (See appendix for estimations of stratigraphic position for individual sections.) The composite section illustrates the chief lithofacies of the Garnet Range and their vertical sequence. Facies are described in the next section.
Figure 4. Correlation diagram showing sections measured for this study. See figure 3 for names and locations of sections. COMP is composite, Ym=MacNamara Formation, Yp=Pilcher Quartzite, Cf=Flathead Sandstone.
Facies Descriptions

The Garnet Range Formation is composed throughout of interbedded sandstone (quartzite), siltstone (siltite), and mudstone (argillite), with minor chert and siliceous siltstone. Although the Garnet Range has undergone lower greenschist facies metamorphism, original sedimentary rock fabrics have not been severely altered and normal sedimentary rock nomenclature can be applied to the rocks. Use of this terminology will allow processes and environments to be interpreted more straightforwardly from descriptions. Also, more precise comparisons between facies will be possible using this terminology.

Sedimentary structures and textures characterize four distinct lithofacies. In ascending order, I refer to these as: 1) the thinly interlayered sandstone-mudstone facies, 2) the planar bedded sandstone facies, 3) the cyclic hummocky bedded facies, and 4) the hummocky and cross bedded facies. The vertical distribution of facies has been compiled from relations within the sections themselves, from published geologic maps (referenced above) which show locations of sections within the formation, and from my own unpublished reconnaissance mapping in the vicinity of some sections. As figure 4 indicates, six sections contain two facies, while two sections are entirely within one facies. These are described below from bottom to top of the formation.
Facies One: Thinly interlayered sandstone-mudstone facies

The lowermost facies of the Garnet Range is exposed at Eddy Creek, Ellis Mountain, and at the base of the Johnson Gulch section (see figs. 3,4). The base of the section at Eddy Creek lies approximately 200 feet (61 m) above the uppermost red beds of the MacNamara Formation. This facies therefore forms the lowermost facies of the formation. It attains a maximum measured thickness of about 60 meters at the Ellis Mountain section. Field observations at Trout Creek, near Superior, suggest that this facies might be significantly thicker there. At Johnson Gulch, along the Blackfoot River, this facies is overlain by dark green sandstone of the planar bedded sandstone facies. Near Alberton, the lowermost facies appears to grade into the cyclic hummocky bedded facies. At Superior, the thinly interlayered sandstone-mudstone facies may be overlain by a red equivalent of the planar bedded sandstone facies.

Two principal components characterize this facies—two meter thick thinning and fining upward cycles, and thin non-cyclical sandstone mudstone interbeds (Fig. 5). In addition, a thin, relatively massive black siltstone interval lies at the top of the facies at Johnson Gulch. Thinning and fining upward cycles are best displayed in the railroad cuts at the Eddy Creek section while non-cyclical interbeds occur at both Alberton and Trout Creek, where they are best developed.

The two meter thick thinning and fining upward cycles typically begin with a sharp based, 10-30cm thick, planar laminated, fine grained
THINLY INTERLAYED SANDSTONE MUDSTONE FACIES

Figure 5. Characteristic stratigraphic sequences of Facies One, Garnet Range Formation. Left column shows non-cyclic flaser sandstone-lenticular mudstone interbeds with local beds of reactivated megaripples. Right column shows cyclic thinning and fining upward cycles with thick basal channel sandstone overlain by flaser sandstone and lenticular mudstone.
sandstone bed. Many of these beds pinch out in one or both directions, while some extend continuously across outcrops. Within the lenticular (channel) sandstone beds, laminae are commonly gently inclined. At Johnson Gulch, some of these basal sandstones contain small megaripples and current ripples. The bases of these beds are locally erosive, and mudchips commonly make up a small fraction of the detritus in the otherwise moderately sorted sandstone. Some sandstone beds contain calcium carbonate concretions up to 50cm in diameter.

Basal sandstone beds are typically flat topped and overlain by black siltstone and interbedded white very fine grained sandstone beds that thin upward to the base of the next cycle. Individual interbeds range from one to ten centimeters thick. Some alternating black and white beds are graded while others display varying combinations of flaser, lenticular, and wavy bedding. Small ball and pillow structures lie within the tops of some black siltstone beds, just below thin overlying sandstone beds. Rare bedding plane exposures display runzelmarken (wrinkle marks) and interference ripples. Some interference ripples comprise symmetrical oscillation ripples occupying the troughs between adjacent straight crested asymmetric current ripple crests.

The other component of this facies, thin non-cyclical sandstone-mudstone interbeds, resembles the upper portions of the thinning and fining upward cycles. The absence of thick, commonly lenticular basal sandstones imparts a conspicuously more tabular appearance to the
bedding at Trout Creek. Flaser bedding is especially well developed in many of the sandier beds and lenticular bedding characterizes the muddier beds. In one locality at Trout Creek, the non-cyclic tabular sandstone-mudstone interbeds are interrupted by a bed of almost wholly preserved half meter high sinuous crested dunes (3-D megaripples of Costello and Southard, 1981). The surface below these dunes undulates broadly (centimeters in amplitude over several meters) and is slightly scoured, suggesting these forms migrated in broad, shallow channels. Some foresets within the dunes are draped with mud blankets up to 1.5 cm thick. Dune tops are not appreciably planed off; their three dimensional form remains essentially intact. Other beds within the non-cyclic part of this facies contain isolated stromatolite heads and silicified pelloidal(?) limestone lenses.

As mentioned above, this facies also contains a distinctive fine grained sequence exemplified only in the Johnson Gulch section. There, the thinning and fining upward cycles pass upsection into an approximately 20m thick package of uniform black siltstone with faint planar laminations. This 20m thick black siltstone is in turn sharply overlain by dark green, very fine grained sandstone with thin mud partings that forms the overlying facies. Numerous coarse white silt and very fine grained sandstone based graded beds lie within this 20m interval. Individual graded beds range from two to 20cm thick. One small slump structure and several interference ripple marked bedding surfaces also lie within this interval.
Facies Two: Planar bedded sandstone facies

The second facies from the base of the Garnet Range, also observed only at the Johnson Gulch section (Figs 3,4), is characterized by a high (9/1) sandstone to mudstone ratio, thin to thick, tabular to wedge shaped bedding, and planar lamination with some low angle internal truncation surfaces (Fig.6). The sandstone beds are typically sharp based and variably flaggy. Flaginess is imparted to the sandstones by very thin (in some cases paper thin) layers of light green mudstone or mica. Variable thicknesses of individual sandstone beds are a function of the frequency of fine grained partings within a particular interval. The more massive sandstone beds range up to three meters thick, with only sparse parting surfaces, mostly concentrated near the bottoms and tops of beds. On the other hand, where parting surfaces are plentiful, a three meter sandy interval may be better described as a stack of two centimeter thick flaggy slabs rather than as one continuous bed. Oscillatory ripple marks mantle some sandstone beds. Some sandstones have scoured tops with up to 5cm relief, draped with mud. Other parting surfaces contain dimple shaped craters interpreted to be raindrop imprints.

Occasional mudstone interbeds range up to 10cm thick. These mudstones are typically structureless to fissile and light brown to pale green in color. Mudstone bases and tops are most commonly sharp but both may grade into sandstone. Rare gashed shaped (synaeresis?) cracks penetrate the tops of mudstone layers.
Figure 6. Characteristic stratigraphic sequence of Facies Two, Garnet Range Formation. Facies contains about 90% sandstone, mostly flat laminated. Flaggy sandstone locally interrupted by 1-10cm thick light green mudstone layers. Wave ripples are common.
Facies Three: Cyclic hummocky bedded facies

The cyclic hummocky bedded facies is exposed in measured sections at Wisherd Bridge, LaFray Creek, Stuart Peak, Pilcher Creek, and Ninemile (Figs. 3,4). Nowhere is the base of the facies exposed. In steeply dipping beds at Ninemile, the planar bedded sandstone facies underlies this facies but a large stream gully with no exposure separates the two. At Wisherd Bridge, LaFray Creek, Stuart Peak, and Pilcher Creek this facies grades up into facies four—the hummocky and cross bedded facies. The maximum measured thickness of the facies is about 200m, but it is undoubtedly thicker.

This facies is similar to most of the planar bedded sandstone facies in that it is composed largely of indurated to flaggy sandstone separated by thin mud layers (Fig. 7). However, whereas the green mudstones that formed thin featureless interbeds in facies two appeared to be nothing more than extra thick mud partings, the mud intervals of facies three comprise a separate heterogeneous and variable component of the facies unto themselves. These muddy interbeds are pervasive and thick enough to yield a characteristic sandstone-mudstone cyclicity to the facies. The mudstone beds commonly grade up from underlying sandstones but never grade into overlying ones. The other main difference between facies two and three is that the character of laminations is considerably more undulatory or "hummocky" in the latter sandstones. Although the two do not occur in any of the same sections, vertical relations within the section suggest that facies two grades
Figure 7. Characteristic stratigraphic sequence of Facies Three, Garnet Range Formation. Sandstone-mudstone couplets average one meter thick. Sandstone commonly grades from plane bedded to hummocky bedded to wave rippled. Some sandstone bed tops are scoured, others graded.
upsection into facies three (Fig. 4).

Sandstone-mudstone cycles of this facies are typically one meter thick and commence with an approximately 70 cm thick, sharp based, fine to very fine grained sandstone. The "ideal" sandstone half of the cycle, present in probably less than a third of the sandstone beds, is tripartite, grading from plane bedded at the base to hummocky bedded at the top and capped by a thin veneer of symmetric ripples (Fig. 7).

Textural characteristics as well as sedimentary structures distinguish the plane bedded from hummocky sandstone. Except for the lowermost few centimeters, the plane bedded interval is often more faintly laminated and lighter colored. Mica is sparse in the plane beds and this interval rarely weathers into flaggy slabs. Where slabs do form, parting lineations commonly etch their surface. These features reflect better sorting in the lower sandstone. Only at the base of the plane bedded sandstone has mud scoured from the underlying layer been incorporated into the sand, giving the rock a dark, mottled appearance.

Laminations in the hummocky beds, in contrast, are well defined by concentrations of relatively coarse detrital mica along evenly spaced bedding surfaces. The micaceous layers weather on the outcrop to give the hummocky sandstones their characteristic flaggy parting. Some flaggy sandstones contain enigmatic, anastomosing depressions that would resemble animal trails were it not for their high degree of parallelism. Surfaces of flaggy slabs also typically have a noticeable micaceous glitter or sheen. Although some mica flakes have diameters
of several millimeters, thin section analysis indicates mica layers are only a flake or two thick and form the tops of thin graded laminae. Graded laminae range from less than a millimeter to a centimeter or more and commonly decrease in thickness upward in the sandstone halves of the cycles.

Hummocky bedding, or hummocky cross stratification, was first named by Harms et al. (1975) for the low angle undulatory cross laminations observed in numerous ancient nearshore sandstones. These workers defined four criteria by which to identify hummocky cross stratification: 1) erosional lower bounding surfaces, 2) internal laminae that more or less parallel that bounding surface, 3) laminae often systematically thicken or fan laterally, yielding progressively flatter laminae upwards, and 4) dip directions of the "cross laminae" have no preferred orientation. All of these features are present in the hummocky beds of the Garnet Range.

Swale (a trough between two hummocks) to swale or hummock to hummock wavelength of the structure ranges from about one to more than ten meters. Swale to hummock amplitudes are ordinarily an order of magnitude less, ranging from a few to about 30 cm. Bedding plane exposures of hummocks, although extremely rare, generally display a distribution pattern that appears roughly orthogonal. Individual hummocks are nearly circular in plan view. Several exposures of hummocky bed soles indicate swales are also nearly circular in plan. Except for those laminae in the upper few centimeters of a sandstone bed, which may have been deposited completely from
suspension, all laminae terminate somewhere against erosional surfaces. Where "cross bed" toesets lap onto a hummock, both ends may end at erosion surfaces. Characteristically, a laminae that discordantly overlies another set of laminae can be traced laterally into an area where it concordantly overlies those same laminae.

Some laminae thicken laterally in a set, typically as they fill a swale but in some cases as they mantle a hummock. In this latter sense, domes or hummocks may be thought of as upward aggrading structures. However, the process of upward growth by spatially preferential laminae thickening is from all indications subordinate to the process of hummock formation by chance bed preservation between adjacent scour pits. Conversely, laminae thickening in troughs is significant and has the effect of "filling in the holes" so that most sandstone bed tops are remarkably planar.

Dips of hummocky "cross beds" rarely exceed 15 degrees and commonly are 10 degrees or less, with dip angles often decreasing upward through a bed. Although cross beds may exhibit preferred orientation in part of any one bed, no such preferred orientation exists on the scale of a whole bed or outcrop. Angle of repose bedding occurs in a few significant places within the hummocky sandstones. In isolated cases, swales are partly filled from one side by small lens shaped sets of angle of repose foresets up to 10cm high. These features are an order of magnitude smaller than one coset of hummocky cross laminae and commonly occupy a slot between an underlying concave up laminae and an overlying concave down laminae. These surrounding
laminae are laterally continuous. Orientations of these foresets are scattered.

The tops of tripartite sandstones are commonly capped by a thin veneer of nearly straight crested symmetric ripples. This veneer rarely, if ever, exceeds one ripple amplitude in thickness, and ripples have negligible angles of climb. Internally, the ripples commonly display only one direction of cross stratification, so they may be best classified as combined flow ripples in the terminology of Harms (1969). Both sharp crested and round crested varieties are present, ripple wavelengths range from 6 to 12cm, ripples indexes from 4 to 10. A significant proportion of the ripples are interference ripples. Ripple crest orientations from this facies lack preferred orientation.

The mudstone portion of sandstone-mudstone cycles is commonly about half as thick as the sandstone portion. Many mudstones are heterogeneous and contain combinations of very micaceous very fine grained sandstone, sandy brown siltstone, black siltstone and brown claystone. Parts or all of this interval are locally graded. Except as described below, laminations in this interval are planar. They are defined by various textural features including mica rich laminae in the micaceous sandstone and siltstone, planar fissility in some brown claystones, and faint black/gray color contrasts in the black siltstone.

One or more thin, fine grained sandstone beds commonly punctuate the muddy interval. These are mostly about 10cm thick but some thicken to become the base of an overlying cycle. Cycle boundaries in such
cases are necessarily defined arbitrarily. These thin sandstones are
typically light colored and plane laminated. The plane laminations are
commonly undulatory and truncated, much the same as the thicker
hummocky bedded sandstones lower in the cycle. Symmetric rippled tops
also serve to identify these beds as smaller versions of the middle
tripartite sandstone.

Gash shaped cracks, formed by either synaeresis (see Donovan,
1972) or post-depositional compaction, occur on the tops of many
mudstone intervals. These are most easily identified on the outcrop
where the uppermost part of a mudstone layer sticks to the base of the
overlying resistant sandstone bed. If they do indeed represent
synaeresis cracks formed at the sediment water interface, they are
significant in that they indicate deposition of the overlying sandstone
occurred without scouring.

Variations of the ideal cycle are commonplace. The sandstone parts
of some cycles contain only plane bedded or only hummocky bedded
sandstone. In addition, some sandstones have internal symmetric
ripples as well as those mantling the bed top. The sandstone-mudstone
transition is especially variable. Figure 7 shows the common ways in
which the sandstone interval passes to mudstone. The variations are
remarkably similar to those recognized by Goldring and Bridges (1973)
in a group of sublittoral sheet sandstones they described.

Composite beds, or amalgamations of more than one cycle, are
locally identifiable. They are most easily recognized where an
internal erosion surface can be traced laterally for a distance of
several meters or more. In Phanerozoic deposits formed by similar means, these amalgamation surfaces are often armored with shell lag (e.g., Kumar and Sanders, 1976; Hobday and Morton, 1980). The lack of shells in the Garnet Range makes recognition of composite beds more difficult.

Facies Four: Hummocky and cross bedded facies

This facies is exposed above facies three in sections at Stuart Peak, Pilcher Creek, LaFray Creek and Wisherd Bridge (Fig. 4). Only at LaFray Creek does the section continue uninterrupted into the Pilcher Quartzite. The onset of the gradational transition from facies three to facies four is marked by the lowest occurrence of fairly large but isolated straight crested and sinuous crested megaripples, defined by wedge sets of planar cross beds and lensoidal sets of trough cross beds, respectively (Fig. 8). They are typically about 30cm high with steeply dipping foreset bedding defined by grain size and grain color differences. Foreset tops have been planed off so that the 30cm preserved height represents only a minimum dune height. The erosional tops of some foreset beds are mantled by symmetric ripples. Hummocky sandstone overlies other cross bedded intervals. Mica flakes are less abundant in megaripple beds than in hummocky beds, so that many dunes crop out in resistant, clean white sandstone ledges. Some beds contain several superimposed dunes, but megaripple beds comprise less than 10%
Figure 8. Characteristic stratigraphic sequence, Facies Four of the Garnet Range Formation. Sequence is similar to Facies Three, but grain size is a little coarser, and megaripples, herringbone cross beds, swaley cross-stratification and low angle planar cross-stratification are also present. Facies Four grades into Pilcher Quartzite, which is predominantly cross bedded sandstone.
of facies four, which, like facies three, is predominantly hummocky and plane bedded sandstone.

Also common to both facies three and four are fine grained interbeds. However, facies four interbeds become thinner and a bit coarser upsection, comprising less claystone and black siltstone, and more sandy brown to maroon siltstone. Grain size in the cross bedded sandstones also increases to medium grained in the uppermost facies of the Garnet Range.

Other volumetrically small, but nonetheless important, components of facies four are herringbone cross beds, so-called swaley cross stratification (SCS), and low angle planar laminated sandstone. Half-meter thick herringbone sets of planar-tabular crossbeds occur in several localities; paleocurrent azimuths within individual sets diverge by greater than 160 degrees. Leckie and Walker (1982) interpreted swaley cross stratification, also termed "flat trough cross stratification" (Ricci-Lucci, 1982), as a variety of hummocky cross stratification in which fine grained interbeds and domal structures are absent. In the Garnet Range, swaley cross stratification occurs only at the top of the Wisherd Bridge section, a short distance below the base of the Pilcher Quartzite. The low angle planar cross stratification occurs in an isolated outcrop at Bear Creek, probably within 20 meters of the base of the Pilcher. The red sandstone at Bear Creek is better sorted and rounded than typical micaceous Garnet Range sandstone. Moreover, the outcrop contains about four meters of sandstone with no fine grained interbeds.
Orientations of large bedforms in this facies are polymodal, probably owing to genesis by complex multiple processes (Fig. 9). Depositional strike approximates ENE-WSW, based on current orientations from probable fluvial deposits of the overlying Pilcher Quartzite.

In summary, facies four is both a transition between and a composite of the underlying hummocky bedded Garnet Range Formation and the overlying cross bedded Pilcher Quartzite. Fine grained, micaceous, hummocky sandstone beds and finer grained interbeds decrease in abundance upwards, giving way to coarser grained, better sorted, cross bedded sandstones with minor swaley cross stratification and low angle planar lamination.

The Pilcher Quartzite

The Pilcher Quartzite conformably overlies the Garnet Range Formation in the vicinity of their type localities. As described above, the transition is gradational over 30 to 60 meters. Mappers in the area have placed the contact at the lowest occurrence of abundant red cross bedded sandstone (Nelson and Dobell, 1961). Given the present lack of stratigraphic and sedimentologic knowledge about the Pilcher, I see no reason to redefine the contact.

The Pilcher Quartzite is characteristically pure milky white with purple or red color banding defining cross bedding—by far the most prominent sedimentary structure in the formation. Grain size ranges
from medium to coarse grained sand and commonly alternates in foreset laminae. The style and geometry of cross bedding vary greatly, but wedge shaped sets of tangential cross strata, probably formed by climbing straight to sinuous crested megaripples appear most common. Cosets average 10 to 30 cm in height and are locally separated by plane laminated sandstone. Characteristically, individual beds contain complex composite mosaics of cosets, probably formed by superimposed bedforms climbing and/or migrating simultaneously in more than one direction. These beds range from about one to more than 10 meters thick. At some localities, light colored, cross bedded sandstone is rhythmically interstratified with meter thick or less, dark red, fine grained sandstone, siltstone and mudstone beds. Nelson and Dobell (1961) describe the Pilcher Quartzite more completely.

Paleocurrent data were collected from crossbeds in the Pilcher in order to attain some preliminary information on depositional environments. Approximately forty measurements were taken on megaripple foreset cross laminae at each of six localities around the study area. The suite of rose diagrams (fig. 9) suggests two current regimes: polymodally directed currents at the base of the Pilcher and unimodally directed (NNW) currents in the middle of the Pilcher. These two current regimes will be discussed in the following sections.
### PALEOCURRENT DATA
- MEGARIPPLES -

<table>
<thead>
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<th>COMPOSITE</th>
<th>PROCESSES</th>
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<tr>
<td>MIDDLE PILCHER QUARTZITE</td>
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<td>Fluvial</td>
</tr>
<tr>
<td>LOWER PILCHER QUARTZITE</td>
<td>LC n=52, BG n=40, WR n=32, WR2 n=40, n=164</td>
<td>Transitional</td>
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<tr>
<td>UPPER GARNET RANGE FM.</td>
<td>WB n=13, SP n=10, LC n=24, PC n=25, n=72</td>
<td>Nearsore</td>
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Figure 9. Paleocurrent data for three stratigraphic units, Missoula Group. See appendix and figure 3 for locations of measurements. Depositional strike interpreted as ENE-WSW from probable fluvial deposits of middle Pilcher Quartzite. See text for further interpretations.
Depositional Processes and Environments

Facies One: Thinly interlayered sandstone-mudstone facies

Thin flaser, lenticular, and wavy sandstone-mudstone interbeds of facies one show clearly that current velocity and water depth fluctuated during deposition. Ladderback ripples capped by mud indicate that current rippled beds were occasionally reworked by waves before being draped with mud in standing water. Rare runzelmarken (or wrinkle marks) formed in ripple troughs by wind blowing across a thin film of water. These structures indicate that the bedding surface emerged intermittently, but not long enough to completely dessicate the exposed surfaces. Stromatolites offer further evidence of shallow water, although they do not constrain depth as much as runzelmarken. These structures and processes are very characteristic of tidal flat environments (Klein, 1977; Reineck, 1972).

The two facies components, 1) thinning and fining upward sequences, and 2) tabular non-cyclic sandstone-mudstone interbeds, probably reflect deposition by somewhat different, albeit spatially and temporally overlapping processes. Thinning and fining upward cycles with lenticular, low angle cross bedded sandstone bases likely formed in large part by lateral accretion during tidal channel migration low on the flats. Planar laminae formed in channel bottoms while low angle planar cross laminae formed low on point bars during either upper flow regime flow or non-aggrading lower flow regime flow (terminology of Simons, Richardson, and Nordin; 1965). This second alternative, where
planar laminae form by ripples and dunes migrating at very low angles of climb (see Hunter, 1977), is attractive since it explains the paucity of preserved foreset laminae in deposits of (presumably) low gradient channels. The upper portions of thinning and fining upward cycles may have formed higher on point bars and on mud/sand flats by a combination of sand deposition during lateral channel migration and mud deposition during periods of low current velocity and standing water. Preservation of abundant vertically accreted sediment (terminology of Walker and Cant, 1979) in migrating channel systems indicates slow channel migration rates, low channel quantity, and/or high sediment aggradation rates. These processes characterize the upper, protected reaches of tidal flat systems (Klein, 1977, Reineck, 1972).

Non-cyclic tabular interbeds of flaser sandstone and wavy to lenticular mudstone, like those at Trout Creek, resemble upper point bar and mixed flat deposits at Eddy Creek. The absence of thick, sandy cycle bases at Trout Creek indicates that channels may have been absent or very sparse. In contrast, vertical accretion deposits of the upper tidal flats predominate. Alternatively, the tabular character bedding and the presence of flaser beds may signify that channels were shallow and migrated slowly, such that channel bottom deposits were completely reworked during point bar progradation. Stromatolite beds offer further evidence that vigorously migrating channels weren't present in this paleogeographic setting. Silicified limestone intraclasts or pelloids in this facies may represent reworked pieces of algal mat.

At the top of this facies, at Johnson Gulch, relatively uniform
black siltstone first appears interbedded with clean, white, cross bedded sandstone, interpreted as lower flat channel deposits. Faint planar lamination and graded bedding, both diagnostic of deposition from suspension, characterize most of the siltstone. This siltstone may have accumulated in a restricted tidal lagoon just seaward of the intertidal flats. Locally, coarse silt bases contain truncated planar laminae, indicative of reworking. Large saucer shaped interference ripples in the middle of this interval indicate waves occasionally penetrated to the bottom. One slump may have formed by collapse of low intertidal or subtidal channel banks.

In summary, I interpret this facies to have been deposited in a mixed tidal flat setting with at least two subenvironments. Channels and channel migration were important components and processes, respectively, in lower or more seaward areas of the system. The uppermost black siltstone probably formed in subtidal reaches of this lower flat area. In the upper, landward reaches of the system, flats were either flat (literally), or perhaps dissected by a few shallow channels. These upper restricted regions were sites of gentle sediment reworking, but were stable enough to allow upward growth of stromatolitic algal heads.

Sediment reworking, rather than episodic sediment input, was the predominant process in both lower and upper flat areas. This process, manifested in abundant flaser, lenticular, and wavy bedding, is indicative of persistent sediment agitation by tides (Reineck and Singh, 1980).
There are two hypotheses on the origin of large megaripples with internal mud laminae. The obvious interpretation is that they represent a rare preservation of tidal channel or tidal creek bedforms. Numerous reactivation surfaces indicate that dune migration stopped and started at least several times, undoubtedly due to fluctuations in tidal current velocity. However, the facts that only one megaripple bed exists in a continuous, well exposed, hundred foot section and that the underlying scour surface is very broad and flat suggests that the bedforms may have been migrating well up on the flats themselves. In this case, current pulses may have been generated by wind driven storm currents during a high wind tide or storm. The apparent lack of subsequent modification, as might occur upon emergence of the flats after the storm tide ebbed, leads me to believe these are channel deposits rather than storm driven bedforms.

Facies Two: Planar bedded sandstone facies

Dark green, very fine grained, plane laminated sandstone sharply overlies the uppermost black siltstone. Although muddy interbeds lie between these sandstones, this sharp boundary represents the highest black siltstone for hundreds of meters in the composite Garnet Range section. I interpret this to mark the sharp change from deposition in a semi-restricted, intertidal to subtidal setting to deposition in a more open, aerated shoreface environment. Planar laminations, very low
angle planar truncation surfaces, and symmetric ripples indicate that sandstone was probably deposited by waves, possibly wholly within the zone of breaking waves. One sandstone bed is capped by interference ripples with raindrop imprints, indicating that sand bodies occasionally emerged as shoals. Thin green shale interbeds indicate that upper regime flow was interrupted sporadically by periods of quiescence and suspension sedimentation, perhaps in the lee of shoal complexes. The abruptness with which plane laminated sandstone passes to suspension deposited mud requires there to have been very shallow water (centimeters?) during deposition. Variation in the thicknesses of mud layers from less than one millimeter to several centimeters suggests that the duration of quiet water conditions was non-rhythmic.

This facies was deposited in very shallow water on the upper shoreface of a very low gradient inner shelf. The low gradient nearshore was inherited from the underlying coastal plain over which the nearshore system transgressed (Fig. 10). The low gradient yielded shallow water depths and consequently wave reworking and agitation of sediment was vigorous. The fine grain size, dark color, and presence of mud in these sandstones probably signifies their derivation from underlying coastal plain deposits. Swift (1968, 1983) noted that beach erosion and coastal plain reworking are common processes occurring during transgression.
Figure 10. Depositional model for the Garnet Range Formation. Numbers refer to environments where lithofacies 1-4 were deposited. Note different upper shoreface geometries during transgression and regression. See text for additional descriptions of lithofacies. Depths at left end of cross sections probably on the order of tens of meters.
Facies Three: Cyclic hummocky bedded facies

Rhythmic sandstone-mudstone cycles of the hummocky bedded facies comprise perhaps the most interesting deposits in the Garnet Range Formation. Many workers have speculated on the origin of similar sheet sandstones in deposits from ancient shallow marine environments (see Golding and Bridges, 1973; Brenchley et al., 1979; for excellent summaries). Most attribute transport and reworking of these deposits to storms, sometimes coupled with tidal ebb (Anderton, 1976) and/or turbidity currents (Walker, 1979). More recently, attention has been focussed on the modes and mechanisms of hummocky cross stratification formation in these settings (Dott and Bourgeois, 1982; Hunter and Clifton, 1982; Swift, 1983). Generation of this structure has also been attributed to storms, but, as with the sheetlike sand bodies enclosing it, the identity and precise role of controlling hydrodynamic mechanisms are not well known. Unfortunately, engineering constraints on large bedform generation in flumes and the hazards involved with offshore studies during storms have precluded direct observation of formative processes in modern environments and flumes.

In light of the aforementioned problems, it would be impossible to treat justifiably the cyclic hummocky bedded facies in just a few pages. Therefore, I have included a separate chapter (three) in this thesis which deals with the processes and problems of sediment transport during storms in greater detail. The interpretations presented here, in Chapter One, are a brief summary of those in Chapter Three.
Sharp bases, textural size grading and an upward decrease in the scale of sedimentary structures in typical tripartite sandstone beds imply that sand was deposited during the waning stages of episodic events, probably storms. Planar bedded fine grained sand with mud chips and pebbles was deposited within the upper flow regime early in the depositional event. Large scour and fill structures (hummocky beds) capped by small symmetric ripples record waning flow velocity, along with decreased sediment transport and increased sediment reworking on the depositional interface. Increased concentrations of detrital mica in the upper halves of sandstone beds further reflect decreased flow velocity during the waning stages of deposition.

Rare preserved megaripple foresets in the plane bedded bases of several tripartite sandstone beds suggest early storm currents were unidirectional. Most workers believe hummocky beds form by scour and fill during combined unidirectional and oscillatory flow (Swift, 1983; Hunter and Clifton, 1982; Dott and Bourgeois, 1982). Wave rippled tops on many sandstone beds indicate that currents interacting with the bed were often purely oscillatory at the ends of storms.

A scenario in which early unidirectional storm currents later give way to oscillatory flow is best accounted for in models recently outlined by Brenchley et al (1979), Morton (1981), and Swift (1983). These workers interpreted early sediment entraining storm currents to have been alongshore or offshore directed bottom responses to wind drift surface currents. As storms approach coastlines, surface water is often blown shoreward or alongshore by hurricane force winds. Much
of this water circulates back offshore as bottom return currents. Murray (1970) measured bottom return current velocities of up to 160 cm/sec contemporaneous with onshore blowing winds during Hurricane Camille in the Gulf of Mexico in 1969. Offshore directed bottom currents have also been documented by Creager and Sternberg (1974), and Smith and Hopkins (1974). Later, as storms move landward, deeply penetrating storm waves rework in place sediment previously deposited by bottom currents. Lateral sediment transport during this late stage may be negligible.

Most of the mudstone in sandstone-mudstone cycles was probably deposited from suspension between episodic storm events. However, mud deposited at the ends of storms is often impossible to distinguish from that deposited between storms. Massive black siltstone, which comprises most of many "interstorm" beds, probably signifies a stratified inner shelf. Bottom water may have been oxygenated only during storms. Thin sandstone beds in this interval most likely represent small storm deposits.

In summary, I interpret this facies to have formed on the lower shoreface of a storm dominated inner shelf. Significant lateral transport of sand characterized early storm processes while sediment reworking was most prevalent late in the storm. Mud was deposited at the ends of storms and between storms. These deposits resemble shallow marine deposits described by Brenchley et al. (1979), Cant (1980), and Hobday and Morton (1981).
Facies Four: Hummocky and cross bedded facies

This facies represents a transition between the lower shoreface deposits of facies three and cross bedded deposits of probable fluvial origin in the Pilcher Quartzite above. Many features of this facies are virtually identical to those of the underlying facies and need no further comment. In addition, though, this facies has an array of features not observed below.

The most obvious of these are large megaripples that occur both as scattered isolated bedforms and locally as grouped composite bedforms. As a group, these bedforms owe their genesis to a complex interaction between several nearshore processes. Based on orientations of unimodal rose diagrams from fluvial Pilcher sandstones, the shoreline was probably oriented roughly ENE-WSW (Fig. 9). Polymodal roses from the upper Garnet Range probably reflect a combination of longshore bedform migration, probably during storms, and offshore bedform migration (NNE), possibly by rip currents, tidal currents, or wind-drift bottom return currents. Flood and ebb tidal currents are represented in several sets of large scale planar tabular herringbone cross beds near the top of this facies. South directed current indicators, subordinate to alongshore and offshore directed features, may represent flood tidal deposits or perhaps landward directed shoreface bedforms similar to those described by Clifton et. al. (1971), and Hunter et. al. (1979).

A two meter thick structure resembling swaley cross stratification (SCS) lies near the top of the Wisherd Bridge section. Leckie and Walker (1982) interpreted this type of structure as a shallower, more
proximal variety of hummocky cross stratification in which bed top reworking may have been continuous so that hummock tops were planed off. The lack of fine grained interbeds also characterizes this structure.

Low angle, parallel laminated, medium grained, vitreous sandstone along Bear Creek may be beach deposits. The general absence of beach deposits in measured sections suggests the nearshore may have had a complex configuration, perhaps due to significant tidal influence.

In summary, the hummocky and cross bedded facies of the Garnet Range formed largely on the upper shoreface of the Belt Sea. Upper parts of the facies were deposited in tidal channels and perhaps on beaches. The transition from the wave dominated shoreface deposits of Garnet Range to the fluvial deposits of the Pilcher Quartzite spans several hundred feet of section, and probably represents progradation of a sandy deltaic system out over the inner shelf on the southeast margin of the Belt Sea. More work is needed, specifically within the Pilcher Quartzite, in order to refine further or revise this model. Figure 10, diagramatically illustrating the relationships between facies and depositional environments, undoubtedly depicts this delta in oversimplified form.
Transgression-Regression

The succession of facies documented in the Garnet Range near its type locality indicates a vertical transition from red delta plain sediments of the upper MacNamara Formation through marine margin deposits to lower shoreface or inner shelf deposits in the middle of the Garnet Range Formation. The lack of well developed beaches or a transgressive disconformity in this interval suggests the coastal plain and nearshore were relatively muddy, perhaps similar to the modern Louisiana chenier plain, the Colorado River delta (Thompson, 1968), or the Devonian Catskill Delta (Walker and Harms, 1971). At some point this transgressive trend slowed and was reversed by progradation of the Pilcher fluvial-deltaic system.

Whether the delta existed during the initial transgression is unknown. If it did, the nearshore marine slope probably remained relatively flat, as delta front outbuilding was suppressed by relative sea level rise. The effect of this transgressive planation is seen in the marked absence of current generated bedforms in facies two. Facies two and facies four were both deposited in upper shoreface settings, but contain drastically different amounts of large, current generated bedforms. The relatively steep shoreface shoreface imparted to the inner shelf by a prograding delta front during deposition of facies four yielded sufficient depth for currents and bedforms to move below
the limit of daily wave destruction (Fig. 10). Abundant hummocky bedding, deposited during combined unidirectional and oscillatory flow, also formed in this setting. On the shallower, low gradient, transgressive shelf associated with facies two, on the other hand, waves were better able to dominate sediment movement on the shoreface. The preponderance of flat to low angle wave accretion surfaces with very thin mud partings reflects the lower gradient present during transgressive deposition of facies two. The slightly larger grain size of facies four shoreface deposits with respect to facies two deposits is further evidence that transgression of facies two was marked by coastal plain erosion and sediment reworking while progradation of facies four was marked by first cycle delivery of sand to the basin.

First order transgressions within the Belt include the Prichard Formation over crystalline basement, the Middle Belt Carbonate over red beds of the St. Regis and Spokane Formations, and Libby/Garnet Range green beds over MacNamara red beds (Fig. 2). The Garnet Range transgression is unique in that it records the first appearance of a wide, turbulent nearshore environment in the Belt. The tremendous quantity of sand transported out onto the shelf during this transgression far surpassed that in preceding transgressions. Even during deposition of the dark, siliciclastic, turbidite fan deposits of the Prichard Formation, to which the Garnet Range has been likened (Smith and Barnes, 1966), shallow water shelves were reportedly sites of carbonate (Finch and Baldwin, 1984) and oil shale (Winston, 1984) deposition. Whether the storm dominated siliciclastic depositional
system introduced during the Garnet Range transgression resulted simply from a complex (coincidental?) interplay between climatic deterioration, increased terrigenous clastic input, and coastal transgression or whether it reflects a fundamental change in Belt Basin tectonic configuration and style of sedimentation is unknown. Perhaps this change signals the first wholesale marine inundation of the Belt Basin. This problem, although difficult to answer, poses many interesting questions concerning the late evolution of the Belt Basin.

**Tides**

Given the absence of paleontological data, tidal deposits represent one of the few criteria available by which to distinguish marine from non-marine deposits in Precambrian rocks. Among the evidence cited by other workers for Precambrian marine deposition are glauconite, long symmetric ripple wavelengths, non-channeled sheet sand flow, and widespread flat erosion surfaces (Anderton, 1976; Level 1, 1980; Hobday and Reading, 1972). Winston, whose studies within the Missoula Group have lead him to regard the Belt basin as intracratonic, believes nearshore water level fluctuations were responses to wind tides rather than diurnal and semi-diurnal ocean tides (Winston, 1984). Mauk (1983), who compiled a bimodal-bipolar current distribution from cross bed measurements in the Revett Formation, interpreted the Belt Sea as marine. Harrison (1972) and
Harrison et al. (1974) are also proponents of a marine interpretation. Stewart (1976) questioned evidence for the existence of an adjacent ocean basin during Belt deposition. He believed the Belt and other Middle Proterozoic basins of the western U.S. were deposited in epicratonic troughs.

In the Garnet Range, reactivated megaripples, herringbone cross beds, flaser bedded muddy sandstones, ladderback ripples, polymodal to bimodal-bipolar current distributions and thinning, fining upward sequences comprise a suite of features certainly formed by diurnal tides. Whether tidal deposits are restricted solely to the top of the Belt awaits further study in other nearshore Belt sediments. If they are, they signify a fundamental change in the pattern of sedimentation and perhaps configuration, orientation, and paleoceanographic affinity of the Belt Basin late in its history. Namely, the Garnet Range may record the first time in the history of Belt sedimentation that the basin was an intimate part of an open marine realm. This will be discussed in more detail in a later section.

**Storm Deposits**

Recognition of offshore storm deposits has proceeded vigorously in recent years (Goldring and Bridges, 1973; Brenchley et al, 1979; Dott and Bourgeois, 1982). Most of these are from Phanerozoic deposits with probably the greatest reported from the Cretaceous System.
Precambrian storm deposits have been reported from Norway (Tucker, 1982; Levell, 1980; Johnson, 1977; Hobday and Reading, 1972) and Scotland (Anderton, 1976). Anderton's study indicates that tidal processes accompanied storm processes during deposition of the Jura Quartzite of the Scottish Dalradian. Although the details of those deposits differ markedly from the Garnet Range, they help dispel the myths that some sedimentary processes (e.g., wind, storms, tides) were different or inactive in the Proterozoic.

The Garnet Range Formation contains the only storm dominated siliciclastic shelf deposits in more than 15km of Belt rocks. Acceptance of the concept of event deposition, such as during storms, carries important implications for the stratigraphic record of shallow marine deposits. In these settings, as in, for example, turbidite basins, most sediment is deposited during geologically instantaneous (catastrophic) events (Seilacher, 1982). The stratigraphic record of these environments is far from the idealized ledger of slow, persistent, steady marine deposition. Rather, it records the alternation of punctuated catastrophism and long periods of quiescence. Recognition of this process has lead to the concept of catastrophic uniformitarianism (see Kumar and Sanders, 1976, Brenner and Davies, 1973), which says, in effect, that the stratigraphic record disproportionately preserves exceptional, large magnitude events. This process is nicely illustrated in the shallow marine deposits of the Garnet Range.
Some previous workers (Smith and Barnes, 1966; Winston, 1977; Bleiwas, 1977) have recognized differences between the Garnet Range and other Belt units. Smith and Barnes proposed that the Garnet Range was deposited in deeper water conditions similar to those in which the Prichard and Wallace formations were deposited. They felt the Garnet Range should be separated as a "fifth Belt group" since it bore little resemblance to the Missoula Group as a whole. I concur with their proposal. Winston (1983, pers. comm.) proposed that an unconformity may separate the Garnet Range from the Missoula Group below, and that the Garnet Range represents a much different style of sedimentation than that recorded below. Reconnaissance study of this transition at Alberton, Bear Creek, Trout Creek, and in the Jocko Mountains (Watson, pers. comm., 1983) has produced no evidence of an unconformity. A very sharp contact between red cross bedded sandstone of the MacNamara Formation and rusty brown fine grained Garnet Range sandstone near Phillipsburg may be part of a local transgressive ravinement surface (see Swift, 1968). Neither Smith and Barnes nor Bleiwas, who studied the MacNamara-Garnet Range transition in detail, recognized an unconformity below the Garnet Range. Bleiwas did argue, however, for a change in source terrain as evidenced by the dramatic increase in detrital mica concentration in the Garnet Range. Several lines of reasoning force me to disagree with Bleiwas' interpretation of a new source. Since the contact with the MacNamara is conformable, the sharp transition from mica poor to mica rich strata coincides with,
and most likely reflects the transition to marine or marine marginal facies. Micaceous nearshore and deltaic sands are common on the Mississippi Delta (e.g., Coleman and Prior 1980, p. 58), yet are rare on river floodplains higher in the drainage basin. Furthermore, mica is present in other Missoula Group rocks, especially in sandier intervals. Mica was probably retained in channels in the Missoula Group depositional system(s), which is why it is so sparse in the delta plain sediments of the upper MacNamara. Also, mica was likely to have been preferentially flushed out of the channel systems into nearshore marine environments.

Since the Garnet Range conformably overlies the rest of the Missoula Group, the question becomes: what tectonic controls changed to allow marine transgression into the Belt Basin for the first time? Price (1984) proposed that the Belt was deposited on the rifted edge of the Proterozoic North American craton. Winston (1984) agreed with the rift interpretation, but noted that sediment was derived from a western landmass at least up through deposition of the Middle Belt Carbonate. The Garnet Range, containing the lowest open marine deposits in the Belt, may record the complete disappearence of the western landmass as a tectonic highland. Whether the western landmass was tectonically translated or simply denuded and submerged is unknown. The regional unconformity at the top of the Belt (discussed below), which cuts through mildly warped, metamorphosed and faulted Belt rocks, indicates that a gentle folding event, termed the East Kootenay orogeny (McMechan and Price, 1982), occurred at some later time. This all occurred prior
to Late Proterozoic rifting that initiated formation of the Cordilleran miogeocline (McMechan and Price, 1982; Stewart, 1976).

Whether the Belt records an older, unrelated rifting event or the onset of long-lived rifting that ultimately culminated in Windermere deposition remains one of the unanswered questions surrounding the Proterozoic of the western U.S. Some workers (e.g., Ballard et al., 1983) have contended that the top of the Belt in Montana is time-correlative with the Windermere and may even conformably underlie the Cambrian. Data from the Garnet Range constrain these contentions. O'bradovich and Petermen (1968) dated a sill cutting the Garnet Range near Alberton at 750m.y. Elston and Bressler (1980) used paleomagnetic data to proclaim an age of 1300m.y. for the Garnet Range and Pilcher. These workers and McMechan and Price (1982) agree that a significant hiatus exists between the basal Cambrian and the top of the Belt. Their evidence, coupled with strong negative evidence in the form of metazoan absence in marine deposits of the Garnet Range make the likelihood of a conformable Belt-Cambrian contact, as reported by Ballard et al. (1983), Winston (1973), Illich and Winston (19??), and Nelson and Dobell (1961), very remote, in spite of the fact that the contact is difficult to identify in many areas (Nelson and Dobell, 1961; Illich, 1966; Winston, 1977).
Summary

The above interpretations lead to the important conclusion that the Garnet Range is marine. If the Garnet Range contains the lowermost marine deposits in the Belt, it carries important tectonic implications. The Garnet Range appears to record the change in tectonic configuration of the Belt Basin from intracratonic or restricted marine to open marine. This change may have resulted from removal of a highland terrain from the west side of the basin by either tectonic translation or submergence. The marine interpretation also constrains previous theories on the age of the top of the Belt. In light of data from the Garnet Range, the top of the Belt could conceivably be as young as Upper Proterozoic, but all indications suggest that a large (200-700 m.y.) unconformity separates the Belt from the Lower Cambrian. It is not yet possible to fully answer questions such as the age and paleogeographic and tectonic significance of the top of the Belt. This study lays groundwork for future studies in this important, but relatively poorly understood interval of the stratigraphic record in the northwest United States.
CHAPTER THREE: DYNAMICS OF STORM SEDIMENT TRANSPORT AND HUMMOCKY BEDDING FORMATION, GARNET RANGE FORMATION

As mentioned in Chapter One, many questions surround the formation of offshore storm deposited sandstones. One question centers on the timing and mode of sediment transport. Three ideas are currently popular (Fig. 11). Hayes (1967) first recognized a relatively thick (30cm) sand layer offshore in the Gulf of Mexico as the product of Hurricane Carla in 1961. His study and interpretation gave rise to the concept of storm surge ebb. In Hayes' scenario, water piled up in coastal lagoons by hurricane force winds later discharges back over and between barrier islands after storm winds subside. These storm ebb currents reportedly transport great quantities of nearshore derived sediment offshore. Since Hayes' study, many workers have called upon some form of this mechanism to explain storm deposited sandstone in ancient shallow marine settings (Brenner and Davies, 1973; Goldring and Bridges, 1973; Reineck and Singh, 1972).

An alternative, albeit not mutually exclusive process, has been advocated by Walker (1979) and Hamblin and Walker (1981), who proposed that most sediment is transported in storm dominated shallow marine environments by storm generated turbidity currents. Offshore oriented directional sole marks on hummocky beds along with subwave base turbidites in adjacent paleoenvironments form the basis for the turbidity current flow model.

A third model, proposed recently by Morton (1981), incorporates the
3 Models of Sediment Transport During Storms

Figure 11. Diagram showing three commonly cited models of sediment transport during storms. The third model is most compatible with sedimentary sequences of Facies Three, Garnet Range Formation. F.W.W.B. is fair weather wave base, S.W.B. is storm wave base.
process of sediment entrainment and transport by wind driven currents spatially and temporally ahead of storm tracts. Morton, citing abundant oceanographic data as evidence (e.g., Murray, 1970), reinterpreted the Hurricane Carla "deposit" as the product of cellular wind driven coastal currents. Swift (1983) called upon similar coastal jets to explain large domal bedforms off the east coast of North America, sighted on side scan sonar. Swift claimed the bedforms formed in place in response to combined unidirectional and oscillatory flow during storms.

In this chapter I will review the problems of storm sediment transport and hummocky bedding formation as they apply to the Garnet Range Formation. My specific aims are fourfold: 1) To show that the sandstone halves of cyclic deposits in facies three of the Garnet Range were deposited episodically by high velocity storm driven currents with both unidirectional and oscillatory flow components, 2) To show that unidirectional and oscillatory flow overlapped during storms to form hummocky bedding; 3) To explain the origin of the mud portion of sandstone-mudstone cycles; and 4) To show that nearshore sediment was transported before and during storms, rather than after them, as implied in many popular but perhaps erroneous "storm surge ebb" interpretations.
Evidence of episodic deposition

Workers in Phanerozoic deep water storm deposits attain the wealth of their evidence for event or episodic deposition from biogenic criteria (e.g., Kumar and Sanders, 1976; Hobday and Morton, 1979; Aigner, 1982; Kreisa and Bambach, 1982; Seilacher, 1982). Escape traces, non-bioturbated sand intervals, tool marks, and shell debris accumulations are just a few commonly cited examples of evidence for episodic deposition. Recognition of Precambrian storm deposits is understandably more difficult. Physical sedimentary structures and textures constitute the only sources of evidence on sedimentation rates and event frequency.

In general, sharp bases and the overall size graded texture of individual cycles is the most direct evidence for episodic deposition. Sedimentary structures and textures, detailed below, indicate that sand commenced depositing under high flow velocities coeval with high but subsequently decreasing sedimentation rates. These structures and textures also suggest that the ratio of suspended load to bed load sedimentation increased from the beginning to the end of each depositional event. The sum of the features described below, while not diagnostic, is strong supporting evidence of event deposition.

Plane laminated bases of ideal tripartite sandstones were deposited in the upper flow regime, as evidenced by scoured bases, parting lineation, and matrix supported pebbles. Some isolated black siltstone pebbles apparently travelled many meters during the storm since they do not resemble subjacent lithologies. Moreover, other
clasts are scattered ten or more centimeters above the base of the storm bed suggesting they took longer to arrive at the deposition site than much of the sand.

In one bed, angle of repose cross beds formed by lower regime flow are preserved at the base of these plane beds. Low angle cross beds, which may represent preserved toesets of subsequently eroded megaripples, occur at the bases of several other beds. Preservation of these isolated lower flow regime crossbeds is certainly the exception rather than the rule, probably because most periods of increasing current velocities were marked by net erosion rather than net deposition. Although rare, these cross beds are significant since they record unidirectional flow early in the depositional event.

Interpretation of this current as unidirectional also follows from several other lines of reasoning. If gash shaped mud cracks formed by synaeresis at the sediment water interface during periods of quiescence between storms, then their preservation on the soles of overlying sandstone beds indicates sand was deposited without scouring. Storm deposited sediment must therefore have been transported laterally (by net unidirectional flow) to the deposition site. Furthermore, accumulation of this facies must necessarily have been accompanied by net unidirectional sediment translation into the depositional environment. Simple winnowing, as might happen during a short lived impingement of purely oscillatory flow on the seafloor, cannot account for the hundreds of feet of strata in this facies. The greatest transport rates and entrained sediment volumes likely occurred when
current velocities were highest, namely during upper stage plane bed flow early in the event. Early unidirectional currents were probably more successful delivering sediment to the area than they were in modifying the shape of the bed. As described below, scours are most pronounced in the middle of typical sandstone beds.

It is possible to identify variations in deposition rates and traction to suspended load ratios in the sandstone beds. These variables are difficult to access in the lower plane bedded parts of sandstones where internal scours are rare and laminations are subtle. Assessment is easier in the hummocky intervals. There, "toesets" of hummocky cross laminae commonly abut tangentially onto the sides of scour troughs as if sediment had been swept in from an adjacent synchronously scoured swale. Toeset truncation is characteristic of neither laminae deposited from suspension nor upper stage plane bed laminae. Rather, these laminae likely onlapped or downlapped during low velocity bedload deposition, or perhaps during some period of combined bedload and suspension deposition. Toeset abutment onto the sides of swales, accompanied by slight laminae thickening in swale axes, filled the swales and eventually lead to a progressive flattening of laminae upwards in a sandstone bed. The passage of swale filling toesets in the middle of sandstone beds upward to laminae that drape one or more hummocks and swales signals a transition from predominantly scour and fill sedimentation to suspension drape sedimentation at the end of sand deposition. Erosional surfaces decrease in abundance upwards while individual laminae become more continuous, planar, and
In one common variation on the cycle theme (Fig. 7), plane bedded sand/mica couplets near the top of the sandstone grade upward from 5mm thick nearly pure bedload deposits with very little mica and clay to 1mm thick pure suspension deposits of alternating silt and mica or clay laminae. In these cases, it seems clear that delivery of suspended sediment to the bed continued long after the bed was being reworked by currents.

In another, perhaps more common variety of the cycle, sandstone tops are sharp and rippled. Internally, symmetric ripples often contain only one direction of foresets and rarely if ever climb. These ripples suggest that currents at the ends of storms were oscillatory or combined unidirectional and oscillatory. Moreover, and in contrast to the sandstone tops described above, these ripples indicate that bed top reworking was able to "outlast" the deposition from suspension of sediment entrained during a storm. So, it appears that while some storms climaxed in voluminous suspension deposition from sediment charged water during waning current flow, others were marked by bottom reworking that persisted longer than most entrained sediment could be held in suspension. The latter case has important implications for the timing of sediment transport. I'll comment further on this as well as the significance of the apparent independance of sediment transport and sediment reworking mechanisms in a later section.

The origin of graded laminae can perhaps be best explained in this
scenario as records of individual wave oscillations. Laminae grading is best developed toward the tops of sandstone beds, where, as discussed above, bedload transport of sediment had diminished. Hunter and Clifton (1982) proposed individual wave oscillations to account for similar graded laminae in hummocky deposits in the Cretaceous Cape Sebastien Sandstone of Oregon. This mechanism explains how consistently alternating sizes of sediment were delivered to the bed, either accompanying or succeeding current flow at the sediment water interface. Waves at the heights of storms penetrated deeply and reworked the bed into hummocks. Wave oscillations at the ends of storms often deposited thin graded laminae and occasionally rippled the bed.

In summary, I've argued that the sandstone halves of sandstone-mudstone cycles were deposited by high, but subsequently decreasing velocity currents during periods of high sedimentation rates. Early in storms, sediment was transported and deposited in upper plane bed flow by unidirectional currents. Later, sediment was deposited from suspension while the water column oscillated and the bed was reworked. These processes provide compelling evidence for storm generated episodic sand deposition. Next, I will focus on the formation of hummocky cross stratification in the context of the processes outlined above.
Hummocky cross stratification

Hummocky cross stratification, the structure so prevalent in the cores of many Garnet Range sandstone beds, is believed by many workers to form below fairweather wave base when large storm waves impinge on the sea floor (Walker, 1979). Hummocky beds contain components of two magnitudes: large hummocks and swales, best developed in the middle parts of tripartite sandstone beds, and thin laminae, which are typically most visible at the tops of tripartite sandstone beds. As explained above, the laminae are commonly graded and basically fill and then mantle the larger scale hummocks and swales, who's formation is controversial.

Most workers agree that hummocky bedding results from large scale scour and fill. Their internal form discordant structure indicates that hummocks initially form as erosional remnants between adjacent scour pits, rather than as accretionary knolls. Two questions about their genesis are: what kind of flow created the first order scours and hummocks?, and, do scours form as parts of some larger scale hydrodynamic scheme, i.e., are hummocks and scours attempting to become equilibrium bedforms, or are the scours simply randomly migrating pits on the bed that have little to do with equilibrium flow? Given the lack of flume data, the second question is unanswered. Various workers have proposed that hummocky beds are oscillatory equivalents to unidirectional upper stage plane beds (Dott and Bourgeois, 1982), antidunes, ordinary megaripples modified by wave orbital motion (Swift, 1983), and large wave ripples (Campbell, 1966). Of these, the latter
two seem most compatible with the scour and fill geometry of laminae within hummocky beds. The possibility that scour pits migrate "randomly" rather than in response to some equilibrium bedform propagation is also plausible.

It may be a little easier, as well as more important, to deal with the first question. As outlined above, flow at the base of sandstone-mudstone cycles was almost certainly within the upper flow regime and it was probably in large part unidirectional. As indicated by the absence of any consistent orientation of "cross laminae", hummocky beds record a transition to non-unidirectional flow. A logical first reaction would be to interpret this flow as oscillatory, especially since symmetric ripples mantle many beds and parting surfaces. Graded laminae, if interpreted correctly, offer further evidence of oscillatory flow late in the depositional episode. Shapes of hummocks and swales, both roughly circular in plan, also suggest that no preferred component or direction of flow existed. Carstens et al. (1969) created bedforms in flumes similar to those in the Garnet Range by high velocity oscillatory flow in fine sand. The only apparently significant difference between Carstens' flume generated forms and hummocks observed in the Garnet Range is scale—those formed in the flume being at least an order of magnitude smaller.

One could argue that the lack of oriented cross beds is not conclusive evidence of oscillatory flow. Indeed one could envision similar bedding forming from migration of scour pits (or certain bedforms) in a highly disorganized or modified unidirectional flow.
In fact, local small lenticular beds of angle of repose cross strata in lower parts of hummocky beds provide good evidence for at least temporary unidirectional flow. Moreover, it is hard to equilibrate a process that yields up to 20cm of erosion over broad bowl shaped areas with one that almost simultaneously deposits thin, laterally extensive laminae. In other words, it is unlikely that individual wave orbital oscillations produced deep broad scours while concurrently depositing thin, even, graded drapes from suspension. More likely, wave orbital motion acted: 1) independently to oscillate the top of the sediment laden water column, and 2) in conjunction with a sediment entraining bottom current to rework the bed into a complex hummocky configuration.

Hummocky cross stratification is herein interpreted to have formed by complex superposition of largely unidirectional bottom currents and wave generated oscillatory water column motion. In some rare cases, hummocks may have migrated laterally in response to a net unidirectional component of current. Whether they simply trailed along passively behind randomly migrating scour pits or whether they themselves were acting as migrating bedforms in an "instantaneously" unidirectional current is unknown.

In summary, I have argued that storm currents at any one site on the seafloor began as largely unidirectional and attained greater proportions of oscillatory flow through time before waning. Hummocky bedding formed by scour and fill in response to the overlap between the two current regimes. Other studies that reached similar
conclusions include those of Cant (1980), Brenchley et al (1979), and Hobday and Morton (1979).

The mudstone portion of sandstone-mudstone cycles

The top halves of sandstone-mudstone cycles are similar in many respects to the DE portion of classical turbidites. They are considered here to be a single unit because it is virtually impossible to separate the mud that fell at the end of a relatively short lived depositional event such as a storm from pelagic mud deposited slowly between events, over the course of perhaps tens or hundreds of years. Whereas geologists working with fossiliferous rocks can separate D from E mud on the basis of allochthonous versus autochthonous fossils, Belt geologists are forced to lump the two. Also, bioturbation can’t be used as a clock to time event frequency in Precambrian rocks. Consequently, Garnet Range storm/non-storm cycles encompass two layers—sandstone, mudstone, in contrast to the three layer sandstone-mudstone-bioturbated zone cycles so common in Phanerozoic rocks (e.g., Hunter and Clifton, 1982).

As mentioned above, some sand beds grade continuously up into mud, while others are overlain sharply by mud. It is probably unreasonable to interpret such sharp sand to mud contacts as boundaries between storm and non-storm or everyday deposits. These sharp boundaries probably formed by gentle wave reworking of the bed after storm.
entrained sand was deposited but before storm entrained mud had settled from suspension.

Despite the difficulties described above, it is still possible to decipher some of the between-storm or everyday sedimentary record. Brown siltstone and claystone are locally present at the base of the mud layer. Their color implies that oxidized clay fell from suspension following storms. In contrast, black siltstone, which comprises the upper parts of many mud beds, was deposited and reduced by the incorporation of organic material reigning out of the water column. Thick black siltstone interbeds indicate that reducing conditions sometimes persisted over the inner shelf for significant periods of time between storms.

Two kinds of sandstone beds are commonly intercalated with the mudstones. Some are thin wavy beds of undulatory to plane laminated fine sand, commonly capped by wave ripples. These resemble the thicker hummocky bedded sandstones and are likewise interpreted as small storm deposits. Locally, thin packages of fine white sand with climbing ripples are interbedded in the black siltstone. Because these are fairly rare and aren't reactivated, they are interpreted to record small unidirectional storm generated currents.
Depositional model—hummocky bedded facies

In light of the preceding discussion, we are faced with two sedimentological problems in the interpretation of this facies. First, when, how, and from where was sediment transported to the depositional environment? Second, how do early, sediment-delivering unidirectional currents and later, bed reworking oscillatory currents become superimposed on one another?

It may be most logical to progress toward some reasonable model through a process of elimination. The process of storm surge ebb can be dismissed as a reasonable model because it demands wave impingement and oscillatory flow before unidirectional flow, the reverse of the sequence in the Garnet Range.

Walker's (1979) model of turbidity current flow during storms has gained acceptance among some workers in Cretaceous deposits of the western interior, particularly in cases where hummocky beds are interstratified with turbidites (e.g., Hamblin and Walker, 1981). As yet, the Garnet Range has not been shown to contain or pass basinward to a distal turbidite facies. Hunter and Clifton (1982) challenged the premise that sediment concentrations can become great enough to induce density current flow on marine shelves, even during large storms. Also, organized sole marks are absent on the bases of storm beds in the Garnet Range. But perhaps the best reason to dismiss the turbidite model is that, like the storm surge ebb model, it requires wave impingement and oscillatory flow before unidirectional sediment transport.
The most reasonable explanation for this facies is that sediment was entrained by wind forced unidirectional bottom currents ahead of hurricane scale storm tracts. As sediment was later deposited, it was reworked by deeply penetrating storm waves. Morton (1981) described two kinds of wind driven nearshore current regimes observed during storms. When storms approached coasts from offshore, two-layer cellular water motion occurred as surface waters were blown onshore and return bottom currents flowed offshore. Bottom current velocities of 160 cm/sec, sufficiently fast to move fine sand in the upper flow regime or suspension, were recorded at the sediment water interface in 6.3 m of water during Hurricane Camille in the Gulf of Mexico (Murray, 1970). Similar measurements have been made for other storms (Smith and Hopkins, 1974; Creager and Sternberg, 1974). In cases where storms approached at a low angle (subparallel) to the coastline, surface and bottom water flowed together, alongshore and away from the storm cell. Figure 9 shows orientations of large megaripples from facies four of the Garnet Range. The prominent NE-SW modes probably reflect bedforms generated by such alongshore storm currents. In both the one- and two-layer cases, unidirectional bottom currents preceded impingement of waves on the bottom. Moreover, maximum current velocities preceded maximum wind velocities, suggesting that most sediment was transported before landfall of the storm. The significance of this data becomes apparent when viewed in light of Hayes' original storm surge hypothesis!

To summarize, several lines of evidence suggest that sediment was
transported by wind forced bottom currents during storms on the inner shelf of the Belt Sea during Garnet Range deposition. Hummocky bedding, an enigmatic sedimentary structure, probably formed by scour and fill as deeply penetrating storm waves interfered with the bottom currents. Analysis of storm dominated shallow marine processes in the Garnet Range has just begun. Further study in this part of the Belt will undoubtedly contribute to our understanding of these complicated processes, and help refine or revise the model outlined herein.
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APPENDIX ONE: METHODS

All sections were measured with a Jacobs staff. Sections were first marked off and labelled at every meter, with either silver paint or duct tape. After labelling, sections were described at the scale of one centimeter to the meter. Graphic columns in Appendix Three have been copied directly from field notes. They have been reduced to 75% of original size.

Paleocurrent data were collected with a Brunton and replotted after corrections were made for structural plunge and structural dip. The assumption was made that no tectonic rotation had occurred. Even if this assumption proves to be inaccurate, the data are still internally consistent since all paleocurrent localities lie within one structural plate (Watson, 1984).
APPENDIX TWO: SECTION LOCATIONS

Quadrangles

Measured Sections:
- Eddy Creek (EC) .............. Petty Mountain 7.5 minute quadrangle, Montana
- Ellis Mountain (EM) ............. Alberton 15 minute quadrangle, Montana
- Ninemile (NM) .................. Alberton 15 minute quadrangle, Montana
- Stuart Peak (SP) .............. Stuart Peak 7.5 minute quadrangle, Montana
- Pilcher Creek (PC) ........... Northeast Missoula 7.5 minute quadrangle, Montana
- Johnson Gulch (JG) .......... Blue Point 7.5 minute quadrangle, Montana
- LaFray Creek (LC) ............ Blue Point 7.5 minute quadrangle, Montana
- Wisherd Bridge (WB) ........ Blue Point 7.5 minute quadrangle, Montana

Paleocurrent Localities:
- LaFray Creek (LC) ............ Blue Point 7.5 minute quadrangle, Montana
- Wisherd Gulch (WG) .......... Blue Point 7.5 minute quadrangle, Montana
- Wisherd Ridge (WR1, WR2) ...... Blue Point 7.5 minute quadrangle, Montana
- Blackfoot River (BR) .......... Blue Point 7.5 minute quadrangle, Montana
- Bear Creek (BC) ............. Sunflower Mountain 7.5 minute quadrangle, Montana

Reconnaissance Localities:
- Trout Creek ....................... Superior 15 minute quadrangle, Montana
- Apex Mountain ................. Danaher Mountain 7.5 minute quadrangle, Montana
Eddy Creek (EC): Section begins near center of SE 1/4 of sec 6, T14N, R22W. Section is exposed in railroad excavation below old Highway 10 overpass. The uppermost red beds of the MacNamara Formation crop out in section 5, along the Clark Fork River, approximately 200 feet stratigraphically below the base of this section (see Bleiwas, 1977).

Stratigraphy: 0m-33m: Facies 1.
Ellis Mountain (EM): Section begins in SE 1/4 of SW 1/4 of sec 32, T15N, R22W. Base of section lies below old railroad tunnel, where outcrop protrudes to within several feet of the road. Section continues up over top of tunnel. The concretionary/cherty horizon at 3dm of this section correlates with a similar interval at 14m of the EC section. The base of this section therefore lies about 178m above the base of the Garnet Range.

Stratigraphy: 0m-62m: Facies 1, 62m-97m: Facies 3. Top of this section is problematic; it contains structures characteristic of both Facies 1 and 3. The two may be interbedded. There is evidence of neither Facies 2 nor an unconformity in the section. The F1/F3 contact is placed somewhat arbitrarily at the lowest occurrence of abundant thick bedded sandstone.

Ninemile (NM): Section begins in SE 1/4 of SE 1/4 of sec 28, T15n, R22W. Section is exposed in vertical slabs along abandoned railroad tracks, just below Missoula lakebed outcrops on Highway 10. Base of section lies just west of large gully. Stratigraphic position of this section is very poorly constrained. It lies significantly above the nearest MacNamara Formation, but excessive deformation in the area prohibits precise determination of vertical stratigraphic relations.

Stratigraphy: 0m-52m: Facies 3.
Stuart Peak (SP): Section begins in the NE 1/4 of NE 1/4 of sec 6, T14N, R18W. Access by descending along ridge southeast of Stuart Peak towards unnamed lake. Section begins at first well exposed outcrops about 100 feet above lake. Based on correlation with LC section, base of section lies approximately 150m below base of Pilcher Quartzite.

Stratigraphy: 0m-31m: Facies 3; 31m-141m: Facies 4. F3/F4 contact defined as per LC section.
Pilcher Creek (PC): Section begins in N 1/4 of NW 1/4 of sec 21, T14N, R18W. Access by ascending talus slope above Rattlesnake Creek toward prominent cliffs. Section begins at lowest well exposed outcrops on west side of prominent gully. Section terminates at west directed thrust fault of unknown displacement. Based on correlation with LC section and map relations (Wallace and Lidke, 1980), base of section lies about 195m below base of Pilcher Quartzite.

Stratigraphy: 0m-127m: Facies 3; 127m-195m: Facies 4. F3/F4 contact placed as per LC section.
Johnson Gulch (JG): Section begins in the E 1/4 of NW 1/4 of sec 14, T13N, R18W. Access by walking across Marco Flat foot bridge and then walking west along abandoned railroad tracks. Section begins at first exposures northeast of the old wooden railroad bridge over Johnson Gulch. Section ends where it becomes excessively disrupted by faulting. Based on calculations on map of Nelson and Dobell (1961), section begins approximately 800 feet above base of Garnet Range Formation.

Stratigraphy: 0m-20m: Facies 1; 20m-280m: Facies 2. F1/F2 contact fairly sharp but probably conformable; it is gradational over several meters.
LaFray Creek (LC): Section not located at LaFray Creek, but rather, across the Blackfoot River from LaFray Creek. Section begins in NE 1/4 of NW 1/4 of sec 7, T13N, R17W. Access by walking one mile east from Wisherd Bridge along abandoned railroad railroad tracks. Near point 3485 in section 7, walk north up talus strewn canyon. Ascend slopes past first main "tributary" to the lowest well exposed outcrops. This section crosses the Garnet Range-Pilcher contact; it begins approximately 229m below the contact.

Stratigraphy: 0m-187m: Facies 3; 187m-229m: Facies 4; 229m-273m: Pilcher Quartzite. F3/F4 contact placed somewhat arbitrarily at lowest occurrence of "fairly abundant" megaripples. F4/Pilcher contact placed less arbitrarily at lowest occurrence of abundant megaripples, maroon fine grained interbeds, and white (vs. green) quartzites.

Wisherd Bridge (WB): Section begins in W 1/4 of SE 1/4 of sec 8, T13N, R17W. Access by ascending talus slope above Wisherd Bridge past plastic water pipe to large fir tree. Based on correlation of F3/F4 contact with LC section, this section begins about 130m below the base of the Pilcher.

Stratigraphy: 0m-87m: Facies 3; 87m-105m: Facies 4. F3/F4 as per LC section.
LaFray Creek paleocurrent locality (LC): At top of LC measured section, see preceding page. This section lies at the base of the Pilcher.

Wisher Gulch paleocurrent locality (WG): Located in NE 1/4 of NW 1/4 of sec 8, T13N, R17W. Access by bushwacking up Wisherd Gulch for about one mile. Measurements taken on prominent outcrop approximately 100 feet above the bottom of Wisherd Gulch. Section probably lies more than 200 feet above the base of the Pilcher.

Blackfoot River paleocurrent locality (BR): Located in NW 1/4 of NW 1/4 of sec 16, T13N, R17W. Access by climbing steep hills and cliffs above Blackfoot River about 1/2 mile east of the west end of Wisherd Bridge. Locality is in core of Wisherd syncline and deformation prohibits precise identification of stratigraphic position. It probably lies at least several hundred feet above the base of the Pilcher.
Wisherd Ridge paleocurrent localities (WR1, WR2): WR1 located near center of NE 1/4 of sec 31, T14N, R17W. WR2 located near center of SE 1/4 of same section. Both can be accessed by walking from the hook in the Lockwood Point road west toward the ridge. WR2 is located on ledges in first cirque, WR1 is located on ledges above second cirque. Both localities lie within several hundred feet of the base of the Pilcher.
Bear Creek paleocurrent locality (BC): Locality situated in SW 1/4 of SE 1/4 of sec 13, T13N, R17W. Access by hiking short distance up talus on east side of Bear Creek, just above sharp bend in road. Garnet Range-Pilcher transition is visible just across the road; the paleocurrent locality lies in the lower 200 feet of the Pilcher.

Trout Creek reconnaissance locality: Secs 23, 24, T16N, R26W. Access via Trout Creek Road, 4 mi. southeast of Superior. Exposures are in roadcuts on NW side of road and on slopes above (to SE of) Trout Creek.

Apex Mountain reconnaissance locality: Lower half of sec 26, T18N, R12W. Access by hiking up Monture Creek (just north of Ovando) and crossing Limestone Pass into Bob Marshall Wilderness. Exposures are on the ridge just east of Apex Mountain.
APPENDIX THREE: MEASURED SECTIONS

Legend

Rock types:

- Plane bedded to low angle cross bedded sandstone, very fine to fine grained, locally medium grained in Facies Four.
- Hummocky bedded sandstone, very fine to fine grained.
- Muddy very fine grained sandstone to silty claystone; dot/dash ratio approximates sand/mud ratio.
- Distinctive black siltstone at Johnson Gulch. Slightly coarser than black siltstone of Eddy Creek and lower Johnson Gulch sections.
- Black siltstone, commonly contains lenticular or wavy sandstone too thin to show.
- Interbedded black siltstone and white fine grained sandstone. Siltstone commonly massive to lenticular bedded, sandstone graded to flaser bedded.
- Waxy, light colored (green-brown) mudstone to claystone. Resembles chert in hardness and luster when pure.
- Covered.

Sedimentary structures:

- Current ripples
- Megaripples
- Climbing ripples
- Symmetric ripples
- Slumps
- Scoured top of sandstone bed; erosional remnants form protruding mounds of various shapes
- Mudchips
- Concretions
- Gash-shaped (syneresis?) mudcracks
- Anastomosing grooves on bedding planes (water escape structures?)
- Mottled bedding (water escape structures?)
- Carbonate cement
- Load casts/ball and pillow structures