DEPARTMENT OF THE INTERIOR

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Stratigraphy, micropaleontology, petrography, carbonate geochemistry, and depositional history of the Proterozoic Libby Formation, Belt Supergroup, northwestern Montana and northeastern Idaho

by

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ABSTRACT

The Proterozoic Libby Formation of northwestern Montana is subdivided in to eight informal members that total a thickness of 1700 meters at the type locality near the town of Libby, Montana. The lower four of these members can be correlated southwest to Clark Fork and southeast to Fishtrap Creek. Hummocky cross- stratified siltite and very fine grained arkosic arenite in the upper Libby Formation serve as a useful marker that can be correlated to the Garnet Range Formation as far southeast as Bonner, MT and to the Libby Formation in the Whitefish Range.

Compositional and textural differences between the lower and upper Libby Formation suggest a renewed influx of detrital material from Beltian source terrain(s). The lower Libby Formation is characterized by subarkosic wackes with grains that are generally well-sorted and have a high sphericity. The upper Libby Formation is composed of arkosic wackes and arenites with grains that are moderately well sorted and have low sphericity. There is no significant difference in grain roundness between the lower and upper Libby Formation.

A well preserved assemblage of oolites in the carbonate- bearing portion of the Libby Formation contains a variety of fabrics, part of which is interpreted as a mixture of originally calcitic and aragonitic ooids. Microprobe analysis of ooids, cements, and ooid replacement fabrics reveals different trace element composition for some of these calcites. It may be possible to at least partially relate varied trace element compositions to the diagenetic sequence in these rocks allowing a crude tracking of the chemical evolution of the pore waters in these rocks. It may also be possible to approximate the trace element composition of Proterozoic sea water.

Acritarchs have been preserved in some of the dark gray argillite zones of the Libby Formation. The biotas examined thus far are not distinctive enough to be used for age interpretation of the Libby Formation.

Comparison of sedimentology and facies patterns in the lower Libby Formation to Tertiary lacustrine deposits and modern sediments of the tidal flats on the Colorado River delta does not provide the resolution needed to clearly distinguish whether the lower Libby Formation was deposited under marine or non-marine conditions. However, the unique assemblage of ooids in the lower Libby Formation may indicate deposition under restricted marine conditions. An interpreted deepening of water in the upper Libby Formation near Libby combined with the influx of 150+ meters of sand derived from rejuvenated source terrains suggests subsidence in the central Belt basin accompanied by uplift in the source area(s) during Libby time.

INTRODUCTION

The Belt Supergroup is one of the thickest Proterozoic successions in the world. The wide distribution of these rocks makes the Belt basin (Fig. 1) one of the largest preserved Proterozoic basins in the world. Despite the fact that these rocks are well known, controversies abound

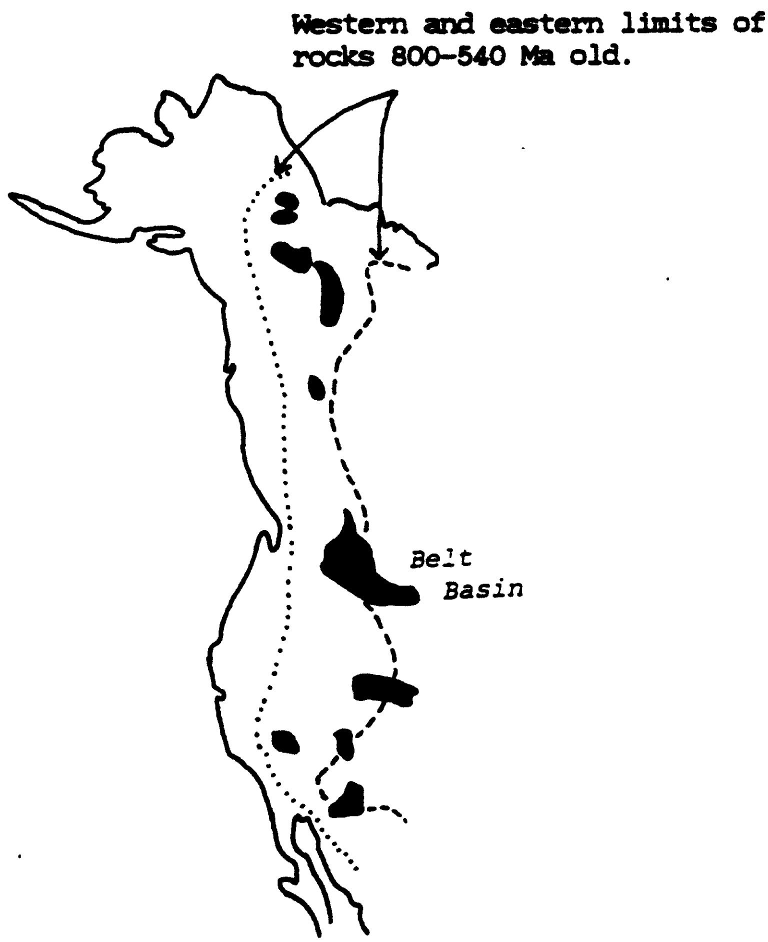


Figure 1. Distribution of Middle and Late Proterozoic sedimentary rocks in western North America after Stewart (1976). Black areas denote Belt-Purcell age rocks. Windermere and lower Paleozoic strata are lumped as shown.

regarding the Belt. Was the supergroup deposited in a marine shelf setting (Price, 1964; Harrison, 1972; McMechan, 1981)? Or was it an epicratonic trough or intracratonic basin (Stewart, 1976; Winston et al., 1984)? Over what length of time were Belt rocks deposited? Five hundred million years (Obradovich et al., 1984)? Two hundred million years (Elston, 1984)? What are the tectonic implications of these differing time spans? Was Belt sedimentation continuous or discontinuous? The answers to these questions are largely unknown and/or controversial.

Superb preservation of sedimentary features is commonly the rule rather than the exception in Belt rocks due to the lack of bioturbation and low grade metamorphism. Locally extensive exposures provide a detailed record of mud deposition. However, determination of facies relations using these features is often hampered by the overall poor exposure and/or regional structural complications, making broad-scale basin analysis difficult.

The top of the Belt Supergroup was eroded prior to deposition of the Windermere Supergroup, which is part of an extensive package of Late Proterozoic and lower Paleozoic miogeoclinal rocks that extend from Alaska to Mexico. The uppermost units of the Belt (Pilcher, Garnet Range, McNamara, and Libby Formations) have only received limited attention compared with the rest of the supergroup. Clearly the best way to understand the relationship between sedimentation that appears to have occurred mainly in isolated basins in the Middle Proterozoic in contrast to the extensive miogeoclinal deposits of the Late Proterozoic is to examine the rocks stratigraphically closest to the unconformity separating these two supergroups. Were there any changes in depositional environment that might have signalled the onset of whatever event(s) that produced this unconformity? Or was there no such signal? Or has it also been eroded away?

This study attempts answer these questions, by detailed examination of some of the youngest rocks in the northwestern part of the Belt basin, the Libby Formation. The results show that not only did sedimentation patterns change significantly near the top of the supergroup, but that facies, petrographic, and geochemical analysis have, in these well-preserved mudrocks, the potential to increase our understanding of mud deposition and possibly even the composition of Proterozoic sea water.

GEOLOGIC SETTING

A striking feature of rocks of the Belt Supergroup is the tremendous thickness of fine-grained sedimentary rocks that are widely distributed and, at first glance, appear similar in many formations within the supergroup. The Belt-Purcell succession (the Purcell Supergroup is the Canadian equivalent to the Belt Supergroup) is commonly greater than 15 km thick and at places reaches a thickness of 20 km (Harrison et al., 1980). These rocks are presently distributed over 128,000 km in northwestern Montana, northern Idaho, northeastern Washington, southeastern British Columbia, and southwestern Alberta (Fig. 1). The original distribution of Belt rocks probably covered an even larger area (Reynolds, 1984).

Although some coarse-grained and conglomeratic lithologies are in the Belt-Purcell Supergroup, most of these weakly metamorphosed sedimentary rocks are composed of particles less than or equal to fine-grained sand in size. Formal units within the supergroup often resemble one another resulting in what has been described as a monotonous sequence of fine-grained rocks.

These rocks have survived a long and complex history of faulting and at places intrusion. Much of the Belt terrane is allochthonous as a result of thrust faulting that translated rocks as much as 180 km to the east (Harrison et al., 1980). Major structural features within the Belt terrane include the Lewis and Clark line and the Rocky Mountain trench. The Lewis and Clark line (Fig. 2) is a broad, complex shear zone that cuts west-northwest across the central part of Belt terrane (Harrison et al., 1974; Reynolds, 1984). The Rocky Mountain trench is a

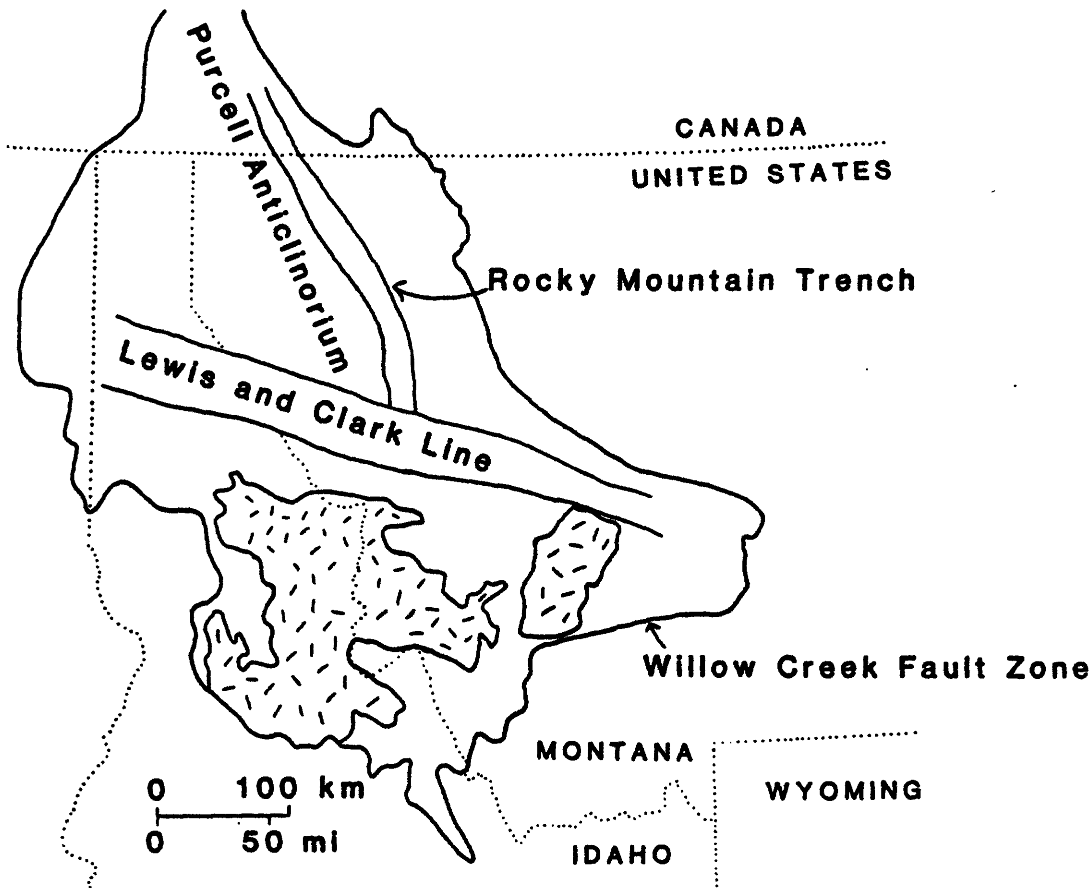


Figure 2. Distribution of Middle Proterozoic rocks in northwestern United states and associated structures (after Harrison, 1972 and Reynolds, 1984). Area within solid line represents limits of Middle Proterozoic rocks in region. Irregularly shaped areas enclosing tick marks are Mesozoic-Cenozoic batholiths.

graben-like structure that extends from the east-central part of the Belt terrane north-northwestward into Canada (Fig. 2). Much of the rest of the Belt terrane is marked by thrust faulting and folding in the northern region (Harrison et al., 1980) and faulting and intrusion by large Cretaceous-Tertiary batholiths in the southern part of the basin (Harrison, 1972). Metamorphism in the Belt Supergroup is generally to lower greenschist facies. Metamorphic grade increases from east to west and from top to bottom of the Belt Supergroup.

The study area occurs within and just west of the Libby thrust belt (Fig. 3). This thrust belt displays a complex areay of faults and folds (Harrison and Cressman, 1985). Backsliding and other extensional movements also characterize the Libby thrust belt, which, overall has experienced a tectonic shortening of about 10 km (Harrison and Cressman, 1985).

Despite the complicated structure of the region, a nearly continuous, 14 km thick section of the Belt Supergroup is preserved in gently dipping exposures of the Sylvanite anticline (Harrison and Cressman, 1985). The upper part of the section preserves the thickest, most complete exposures of the Libby Formation (1700 m) in the Belt basin. Supplementary sections (Fig. 3) are less complete because of post-Libby Formation erosion (Clark Fork, ID) or because of erosion at the tops of those sections combined with loss of strata due to faulting (Fishtrap Creek).

STRATIGRAPHY

Stratigraphic relations in the upper part of the Missoula Group have, to date, been only loosely resolved. The Bonner Quartzite has received focused attention over the last decade (Winston, 1978; Winston et al., 1986), and can be traced to most parts of the Belt basin, excluding the Helena embayment (Harrison, 1983), but with the exception of studies by Bleiwas (1977) and McGroder (1984) on the McNamara and Garnet Range Formations, post-Bonner rocks have received very little attention. On a crude, regional scale, the McNamara and Garnet Range Formations successively overlie the Bonner Quartzite in the southern and eastern parts of the Belt basin, and the Libby Formation overlies the Bonner Quartzite in the northwestern part of the basin. Precise stratigraphic relationships between these units have been, and still are controversial. Winston (1984, 1986) correlated the McNamara Formation to the lower Libby Formation and diagramed equivalence between the Garnet Range Formation and the upper Libby Formation. Wallace et al (1984) equated the Libby and Garnet Range Formations directly and stated that the McNamara Formation is absent where the Libby Formation is present. Whipple (1984) and Whipple et al (1984) correlated the McNamara Formation from the southeast northward to the Whitefish Range and referred to the rocks overlying the McNamara Formation as the Libby Formation.

The detailed stratigraphic work of this study focuses first on the internal stratigraphy of the Libby Formation, and second on correlation of the Libby Formation to other formations in the southern and eastern parts of the basin. Tentative correlations are also made to the Mount Nelson Formation of the Purcell Supergroup in Canada. A clear understanding of the stratigraphic relationships in the upper Belt Supergroup is essential before beginning depositional and basin modeling for these rocks.

The Libby Formation was named by Gibson (1948) for exposures of light-gray and dark gray argillite (the terms argillite, siltite, and quartzite are used to describe the low-grade metamorphic equivalents of shale, siltstone, and quartz sandstone respectively) on the north side of the Kootenay River about 8 miles west of the town of Libby, MT. Informal members recognized in the course of this study are best exposed on Flagstaff Mountain. These members are described in detail and are correlated to less completely exposed sections at Clark Fork, ID

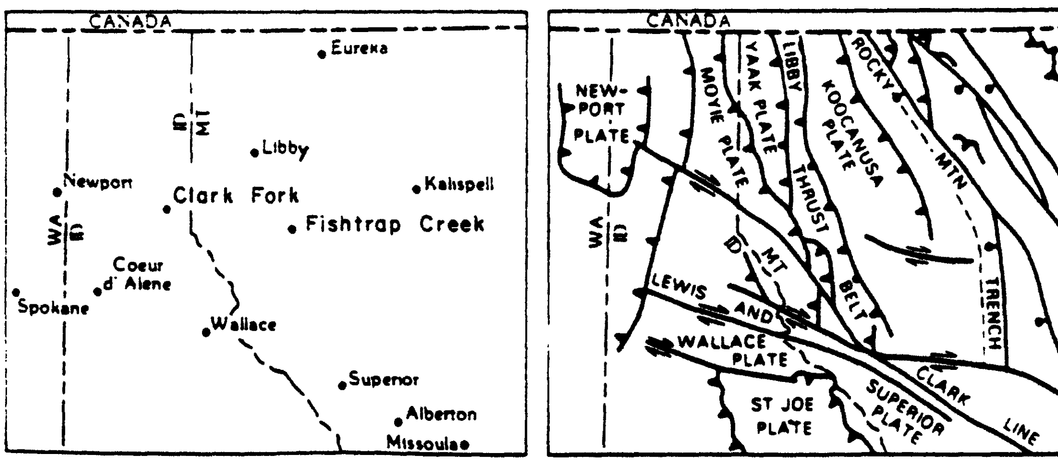


Figure 3. Generalized maps of study area showing major structural features of the Libby thrust belt (modified from Harrison, 1984). Dashed line marks western margin of Rocky Mountain trench.

and Fishtrap Creek, MT (Fig. 4). Subsequent stratigraphic discussion focuses on correlation of the Libby Formation to lithostratigraphically equivalent rocks throughout the Belt-Purcell basin. This basinwide correlation builds on the results of many previous workers, and is strengthened by my reconnaissance of many of the localities described by others.

Bonner Quartzite

The Bonner Quartzite is an important marker unit below the Libby Formation and is therefore briefly treated here before description of the Libby Formation. The Bonner Quartzite in the Libby area consists of approximately 300 meters of fine to medium grained quartzite interbedded with siltite and argillite. The overall color of the Bonner Quartzite is reddish purple. Quartzite beds range from pink to buff colored, and argillites are usually purple.

The quartzites are generally medium to thick bedded and display a variety of sedimentary structures including planar bedding, low angle trough cross-bedding, scour surfaces, and load structures. Interlaminated quartzites and argillites display wavy lamination. Tops of beds are argillitic and are characterized by oscillation ripples with wavelengths averaging 5 cm. The quartzite:argillite ratio ranges from 4:1 to 20:1.

A generalized sequence for individual beds (commonly 1/2 meter thick) within the Bonner Quartzite can be recognized. The base of a typical bed is sharp (sometimes scoured), and approximately the lower two thirds of the bed consists of planar bedded and low angle cross laminated quartzite and coarse siltite. The upper third of the bed is wavy laminated and grades upward into interlaminated siltite and purple argillite toward the top of the bed, which consists entirely of rippled purple argillite. Flat chips of purple argillite (several cm long and generally <1cm thick) are common in the quartzite layers.

Contact with the Libby Formation

The contact between the Bonner Quartzite and Libby Formation is one of the most abrupt contacts in the Belt Supergroup. Many intra-Belt contacts are gradational over many tens of meters or more (Harrison pers. comm., 1985), whereas the Bonner/Libby contact is confined to one meter of section in the type area of the Libby Formation. The best exposure of this contact is along Highway 2 at milepost 27, west of the town of Libby. Steeply dipping beds of Bonner Quartzite undergo a transition from pink quartzites and purple argillites to green quartzite that is 15 m thick. This quartzite abruptly passes upward into the dark gray argillite that Gibson (1948) defined as the base of the Libby Formation. One meter of green siltite separates the green quartzite from the dark gray argillite. This contact is less well exposed on Flagstaff Mountain (across the Kootenay River) from which most of the following informal members were described.

The Libby Formation

The Libby Formation on Flagstaff Mountain has been subdivided into eight informal members (Fig. 5). Contacts between members are gradational over several meters except where otherwise stated. Most of the members change laterally only slightly over as much as 60 miles. Therefore, detailed description is restricted to the type locality on Flagstaff Mountain. Significant lateral changes in each member are briefly discussed immediately following description for each member. Correlation of members to Clark Fork is straightforward. Some members can be traced to Fishtrap Creek with moderate success, but this requires some interpretive facies correlation.

Member A is the basal unit of the Libby Formation. This member is a laminated dark gray

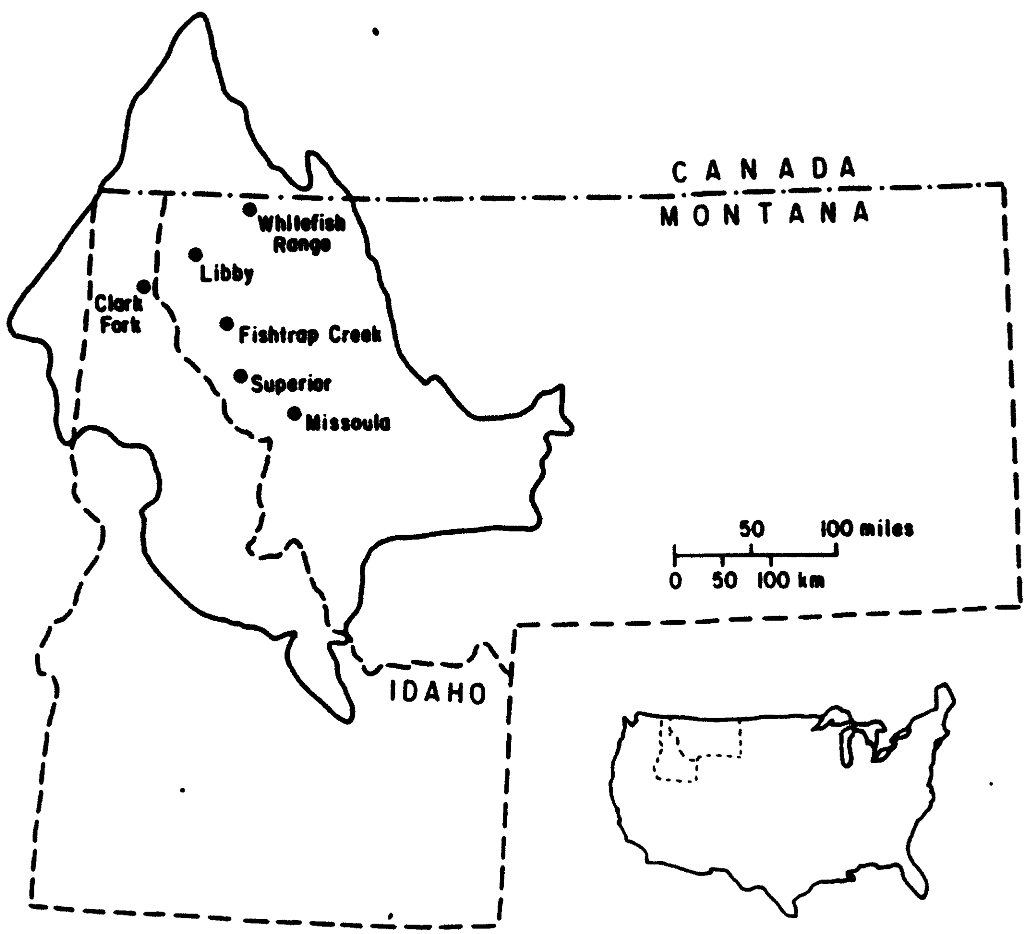


Figure 4. Map showing present distribution of Belt rocks (solid line) and principal localities referred to in text. (Harrison, 1972).

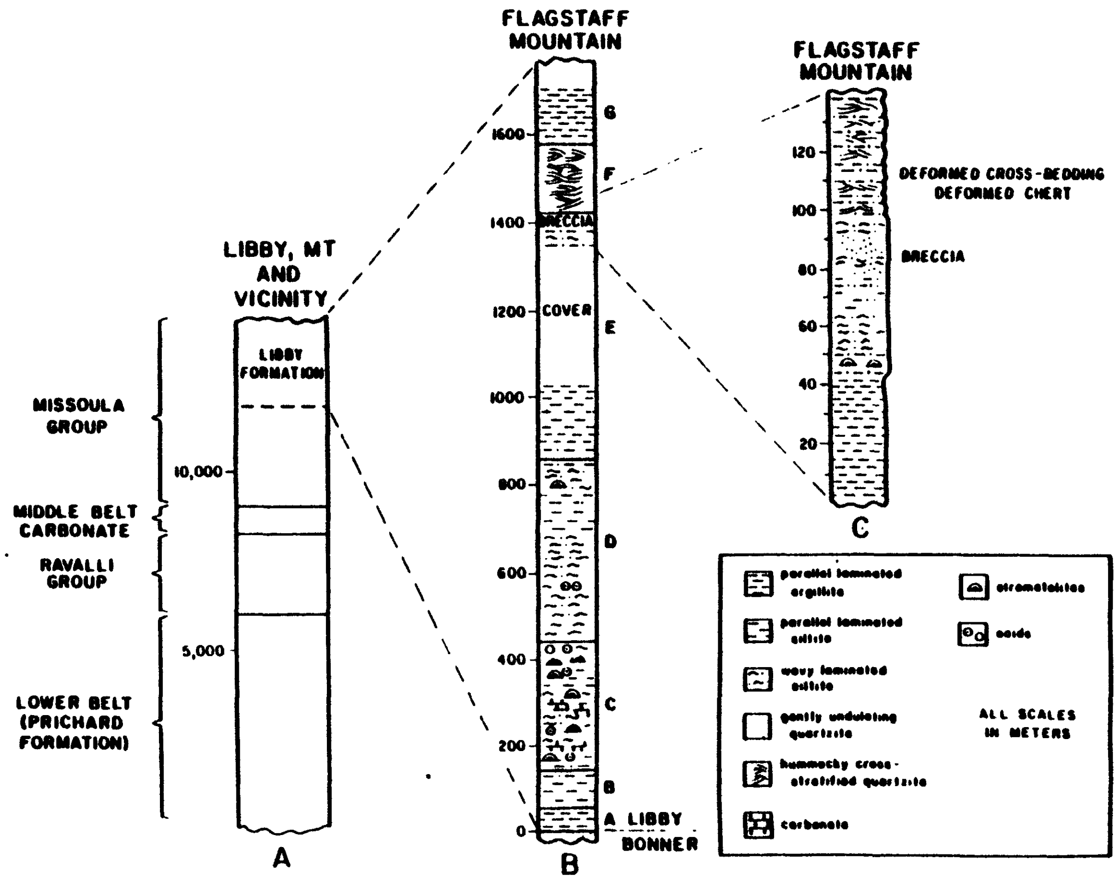


Figure 5. Nested stratigraphic diagram of Proterozoic rocks exposed near Libby, Montana. A. Generalized column of the Belt Supergroup in the Libby region, B. Stratigraphic column of the Libby Formation on Flagstaff Mountain approximately 8 miles west of Libby. C. Close-up of the transition to hummocky cross-stratified facies in the upper Libby Formation on Flagstaff Mountain.

argillite. The laminations are 1mm to 1cm thick and consist of dark green silt layers that have sharp bases and grade upward into dark gray argillitic laminae. Laminae are usually even and parallel, but are sometimes slightly wavy, and occasional zones display lenticular lamination. Concretions occur rarely, and authigenic pyrite is occasionally present.

Member A is 55 m thick at Libby. At Clark Fork, it thins to approximately 30 m (minimum of 25 m) with no noticeable lithofacies changes. Recognition of member A at Fishtrap Creek is complicated by intertonguing relationships with green siltite that not only weaken stratigraphic correlation within the Libby Formation, but also involve the problem of correlation of the lower Libby Formation to the McNamara Formation. The relationship between the Libby and McNamara Formations is treated in detail later. The contact with member B is gradational.

Member B is composed of green siltite. This unit is characterized by predominantly even, parallel, thick lamination produced by alternating layers of dark greenish gray siltite and light greenish-gray, siltite. These dark and light shades of greenish gray usually weather to rusty tan and pale greenish gray respectively. Zones up to a meter thick display wavy lamination. Other sedimentary structures include shrinkage cracks, starved ripples, fluid escape structures, small-scale cut and fill structures, and load structures. Also present are flat rip-up clasts of siltite and rarely, chert.

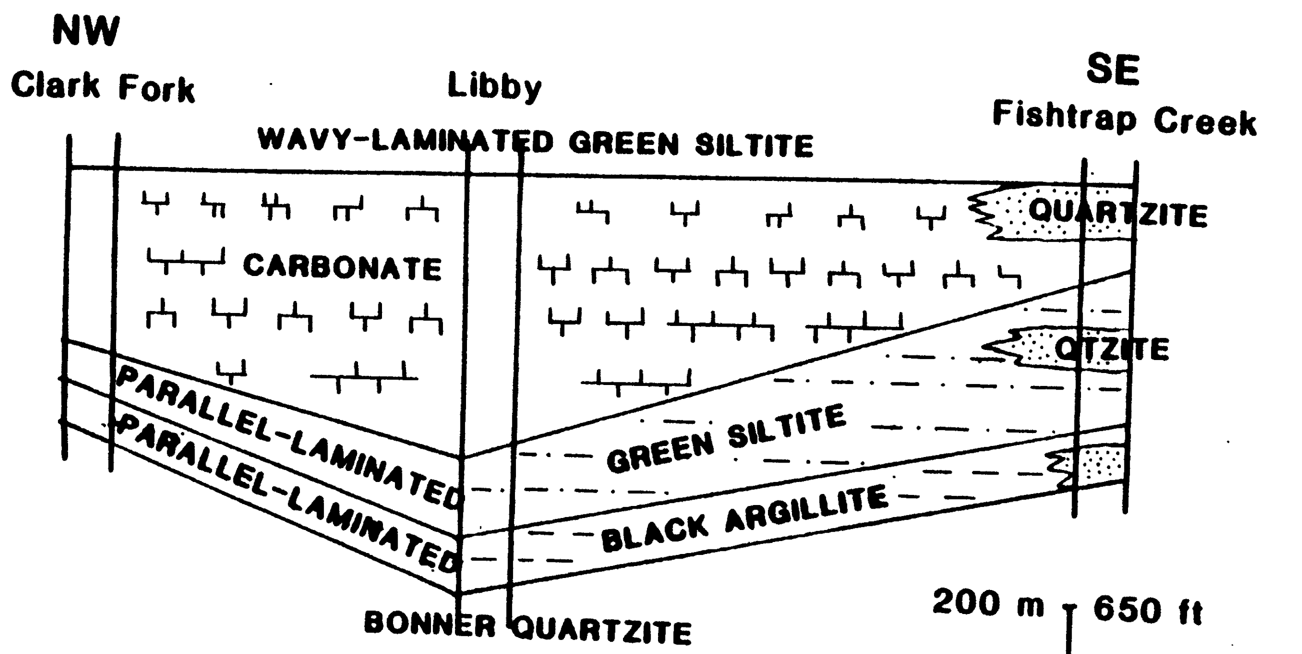
Member B is 85 m thick at the type locality on Flagstaff Mountain. It thins to approximately 30 m at Clark Fork (13 m minimum, 60 m maximum) with no significant facies changes. This member maintains its typical character at Fishtrap Creek except for the presence of occasional quartzite beds with sharp bases that are generally about 1 m thick. At Fishtrap Creek, this sandier facies of member B is 133 m thick. The contact with member C is gradational.

Member C is a carbonate-rich stromatolitic and oolitic siltite and argillite unit. Parallel to wavy laminated green siltite is the principal rock type in this unit, but dark gray parallel laminated argillite is also common. The lithology and sedimentary structures are generally similar to the features of member B with a few exceptions. Flat to mound- shaped stromatolites, usually less than one meter thick, have an uneven vertical distribution occurring every 5 - 30 meters. Some of these stromatolitic zones are associated with oolites at the base or top of the stromatolitic bed. Discontinuous oolite beds up to a half-meter thick also occur within this carbonate-bearing unit. Shrinkage cracks are much more common in this member than any of the others. Preserved cracks are difficult to observe in plan view because of the rarity of bedding plane surfaces, but cross sections of these cracks are abundant, particularly in the dark gray argillite.

Member C is 290 m thick on Flagstaff Mountain. At Clark Fork the unit thins to 180 m. In addition to the stromatolites and oolites, fenestral fabric is present in some of the carbonate strata in a quarry 7 miles east of Clark Fork on Idaho Highway 200.

Correlation of this member from Flagstaff Mountain to Fishtrap Creek lacks resolution because of facies changes. Only a few stratiform stromatolites occur at Fishtrap Creek. Some of the shrinkage cracks in this area are well-defined, polygonal shrinkage cracks formed by desiccation that occur in thin red argillite layers. Bleiwas (1977) observed fenestral fabric in this part of the section at Daisy Creek (one of the localities from which the composite section of this study was constructed). Intercalation of thin quartzite beds obscures direct lithostratigraphic correlation of much of this unit to the type locality. With the exceptions of the mud-cracked red argillite, quartzite beds, fenestral fabric, and overall lesser amount of carbonate, this member resembles member C on Flagstaff Mountain. It seems reasonable, therefore, to correlate the stromatolite-bearing and desiccation-cracked part of the Fishtrap Creek section to Member C at Flagstaff Mountain (Fig. 6) because of similarities in lithology and sedimentary structures and a similar stratigraphic position.

A unique feature of the Fishtrap Creek section above the stromatolitic zone is the < 1m thick beds of fine grained pale green quartzite and greenish gray siltite that fine upwards into a thin cap of grayish red to pale red (5 R 4/2 to 5 R 6/2) argillite with desiccation cracks. Some of the quartzite beds pinch out laterally over a distance of 10-30 meters. These quartzites are interpreted



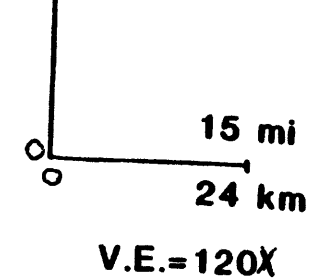


Figure 6. Lithostratigraphic and facies correlation diagram for the lower four members of the Libby Formation extending from Clark Fork, ID to Libby, (Flagstaff Mountain) and Fishtrap Creek, MT. Symbols: dots = quartzite; dot-dash = siltite; dash = argillite; bricks = carbonate. V.E. = vertical exaggeration.

as broad, shallow channels, but it is difficult to discern whether the bases of the beds are flat or concave-up. The zone containing stromatolites and desiccation- cracked red layers is 90 m thick. Once above the red argillite, correlation confidence diminishes, and the rocks are characterized by parallel to wavy laminated green siltites and argillites that are marked by load structures, small channels, and flat rip-up clasts of green siltite. One thin oolitic (?) carbonate bed was noted.

At this point in the Fishtrap Creek section, exposure becomes poor and a normal fault of uncertain displacement occurs. J.E. Harrison (pers. comm. 1985) stated that this fault has at least 120 meters of throw. Stratigraphic correlation between the sections at Fishtrap Creek and Libby suggest that as much as 700 stratigraphic meters may have been removed by post-Belt faulting. The contact with member D is gradational.

Member D consists primarily of wavy and lenticular laminated green siltite. Lithologically, it is characterized by 1mm to 1cm scale dark green siltite and light green silty argillite laminae that, at first glance resemble those in members B and C. However, wavy lamination is the dominant sedimentary structure in this unit. Lenticular and parallel lamination are less common. Stratiform stromatolites occur very rarely in isolated zones. A zone of ovoidal carbonate concretions occurs along bedding in the upper part of this member. The concretions range in long dimension from a few cm to 30 cm, and are 20-30 cm thick.

Member D is 370 m thick on Flagstaff Mountain. It correlates to a very similar lithology at Clark Fork that has a minimum thickness of 300 m, but the top of that section has been eroded. Correlation of this member from Libby to Fishtrap Creek is difficult. Siltite interbedded with quartzite above the stromatolite and red argillite zone resembles the wavy laminated siltite on Flagstaff Mountain. It is presumed that the interbedded siltites and quartzites are a facies equivalent of member D. This facies equivalent is at least 100 m thick, but the normal fault mentioned earlier and exposure problems prohibit precise correlation of this unit. The contact with member D' is gradational.

Member D'. This member is a transition zone between members D and E that is highlighted by resistant exposures of parallel to wavy laminated green siltite similar to that found in members B, C, and D. These siltite zones alternate with recessive zones of poorly exposed dark gray argillite that is parallel laminated where exposed. Thickness of both rock types varies from one to several meters. This transition zone is approximately 200 m thick. The contact with member E is gradational.

Member E is a poorly exposed parallel laminated dark gray argillite that attains an approximate thickness of 360 m on Flagstaff Mountain. Traces of this member can be recognized above the faulted and covered zone at Fishtrap Creek where there are approximately 30 m of wavy laminated dark gray-green silty argillite (5G 5/2) with thin interbeds of dark gray argillite. These limited exposures would not by themselves be confidently correlative to Flagstaff Mountain, but solid correlation of the overlying member facilitates recognition of member E at Fishtrap Creek.

Member F is a hummocky cross-stratified quartzite and coarse-grained siltite unit. It is characterized by blocky weathering 1-5 m thick quartzite beds with silty argillite partings that are up to 3 cm thick. Some of the thicker quartzite beds are amalgamated. The bases of the quartzite beds are sharp and sometimes display cut and fill structures that contain quartzite and rip up clasts of green siltite and/or black chert. The quartzite beds are characterized by low angle cross-stratification and bed tops are commonly hummocky. Hummocks display 10-20 cm of relief and are spaced 1-2 m apart. At places the argillitic partings contain shrinkage cracks.

The contact with the underlying dark gray argillite member is gradational over a variable thickness of 30-50 m. The transition into this member (Fig.5C) consists of an increase in grain size and sedimentary structures that record a progressively higher flow regime. Parallel laminated green siltite overlies the dark gray argillite of the top of member E. A 2 cm thick, poorly developed, stratiform stromatolite occurs within the parallel laminated zone. The siltite is overlain successively by wavy laminated green siltite, planar bedded coarse siltite, undulating coarse siltite, and finally, hummocky cross-stratified fine-grained quartzite and coarse-grained siltite. The transition from argillite to quartzite is accompanied by a zone of soft-sediment deformation. This zone is characterized by contorted layers, recumbently folded cross-strata, soft-sediment breccias with clasts containing deformed laminae, and cherty load structures (Figs. 7,8).

Member F is 155 m thick on Flagstaff Mountain. At Fishtrap Creek, as much as 100 m of the hummocky cross-stratified quartzite are truncated by the pre-Middle Cambrian unconformity. The lower part of the hummocky beds contains soft-sediment breccias that are very similar to those on Flagstaff Mountain. Recumbently folded cross-strata are also present at Fishtrap Creek. The contact with member G is gradational.

Member G is a dark olive-gray silty argillite unit that is massive at most exposures. However, millimeter-scale laminations are observable at places. There are occasional zones of fine- grained quartzite within this unit. This member has a minimum thickness of 125 m, although as much as 180 m may be present on Flagstaff Mountain. The top of the Libby Formation is eroded.

Correlation of the Libby Formation to other Belt units

McNamara Formation

The McNamara Formation resembles the lower members of the Libby Formation in many respects. Each formation was originally defined in a different part of the Belt basin. The McNamara Formation is characterized by green and red laminated siltite and argillite with some interbedded quartzite. Parallel to wavy lamination is common, flat rip-up clasts of siltite (sometimes cherty) and shrinkage cracks are also common. Both formations immediately overlie the Bonner Quartzite.

The McNamara Formation was defined by Clapp and Deiss (1931) for exposures of argillite and quartzite between the now abandoned town of McNamara's Landing and the town of Bonner. The McNamara Formation was divided into two argillitic members separated by a quartzite member (Fig. 9). When Nelson and Dobell (1961) revised the stratigraphy of the Missoula Group, they elevated the three members of the McNamara Formation of Clapp and Deiss (1931) to formation status. The new formations were, in ascending order, the Miller Peak Argillite, the Bonner Quartzite, and the McNamara Argillite (Fig. 9). The Bonner Quartzite and McNamara Formation are still in use as defined by Nelson and Dobell (1961), but the Miller Peak Formation is now subdivided into the Snowslip, Shepard, and Mount Shields Formations (Fig. 9).

The revision of Nelson and Dobell (1961) for the McNamara Formation is not in accord with present principles of stratigraphic nomenclature. Article 19g of the North American Stratigraphic Code (North American Subcommission on Stratigraphic Nomenclature, 1983) states (p. 855): "When a unit is divided into two or more of the same rank as the original, the original name should not be used for any of the divisions." Thus the present use of McNamara is technically incorrect. However, inasmuch as the rocks in question have been referred to as McNamara by workers in the Belt basin for more than twenty-five years, to change it at this point over a technicality would, in my opinion be unnecessary.

The contact between the Bonner Quartzite and the McNamara argillite is gradational in the type area of these formations. There are alternating beds of pink quartzite and green siltite and argillite over a transitional zone of roughly 30 m. Nelson and Dobell (1961) defined the top of the Bonner Quartzite as the highest typical pink quartzite, which places most of the transitional beds in the Bonner Quartzite.

The similar appearance of the McNamara Formation and lower four members of the Libby

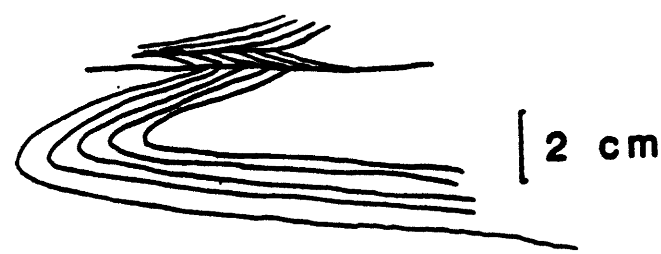


Figure 7. Field sketch of recumbently folded cross strata from Member F on Flagstaff Mountain.

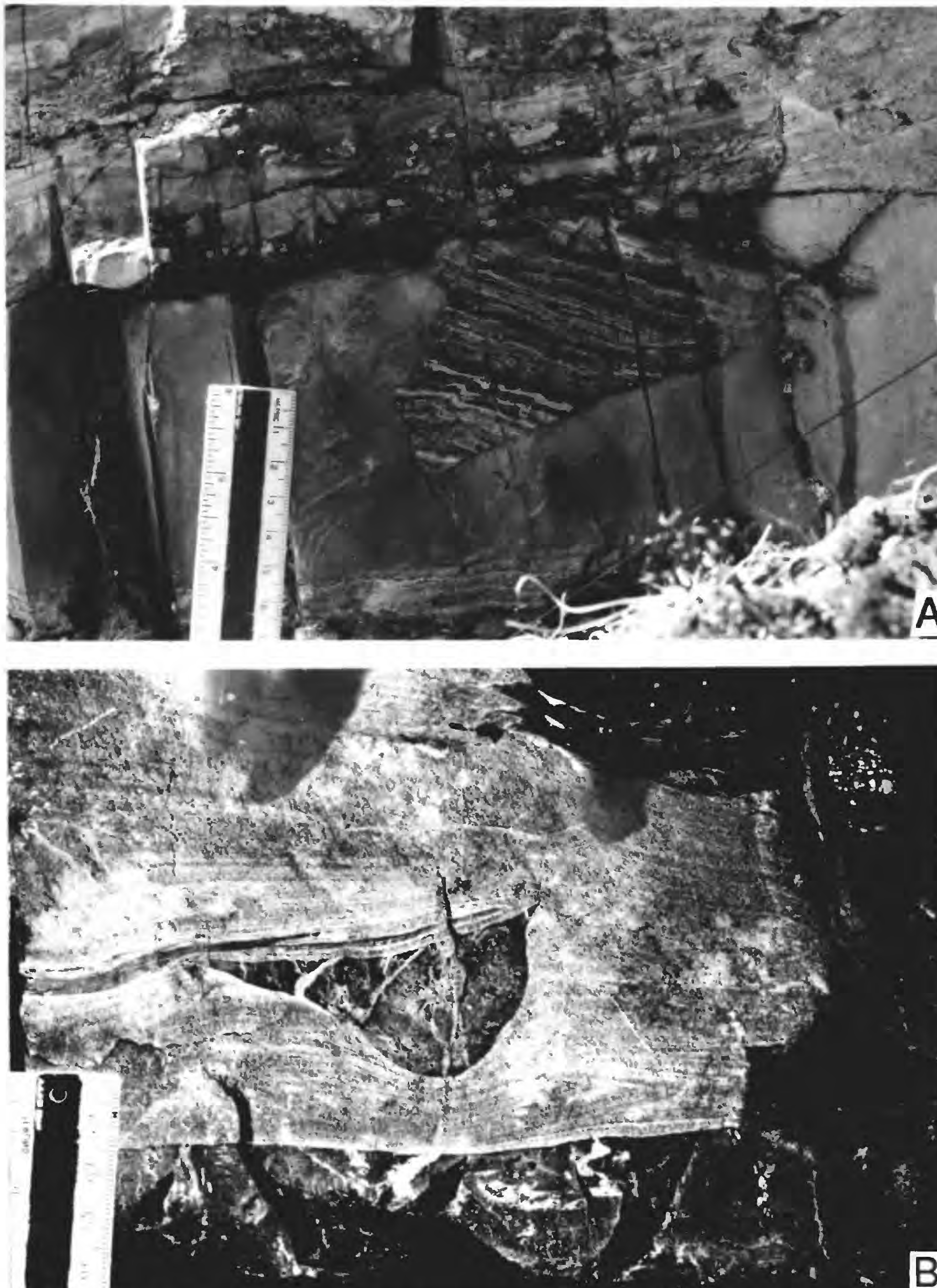


Figure 8. Photographs of soft-sediment deformation features near base of hummocky cross-stratified facies (member F) in upper Libby Formation, north side of Flagstaff Mountain. A. Angular clast of laminated greenish-gray siltite exhibiting deformed laminae. B. Deformed chert (black) in low angle cross- stratified very fine grained arkosic arenite. Right side of ruler scaled in millimeters in both photos.

<table>
<tr><th colspan="2">Clapp and Deiss (1931)</th><th>Nelson and Dobell (1961)</th><th>Harrison (1983)</th></tr>
<tr><td colspan="2">Sheep Mountain formation</td><td>Pilcher quartzite</td><td>Pilcher Quartzite</td></tr>
<tr><td colspan="2">Garnet Range formation</td><td>Garnet Range quartzite</td><td>Garnet Range Formation</td></tr>
<tr><td rowspan="3">McNamara formation</td><td>Upper member</td><td>McNamara argillite</td><td>McNamara Formation</td></tr>
<tr><td>Middle member</td><td>Bonner quartzite</td><td>Bonner Quartzite</td></tr>
<tr><td>Lower member</td><td rowspan="3">Miller Peak argillite

Hellgate quartzite member</td><td rowspan="3">Mount Shields Formation

(4 members)</td></tr>
<tr><td colspan="2">Hellgate formation</td></tr>
<tr><td colspan="2">Miller Peak formation</td></tr>
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Figure 9. Correlation diagram showing history of terminlogy regarding what are now referred to as Mount Shields, McNamara, Garnet Range, and Pilcher Formations and Bonner Quartzite near the town of Bonner, MT.

Formation to each other and their common stratigraphic position above the Bonner Quartzite suggests the stratigraphic equivalence of these units. Their differences in color and grain size warrant retention of their present nomenclature.

Garnet Range Formation

The type locality of the Garnet Range Formation was described by Clapp and Deiss (1931) for exposures of "brown, green-gray to gray thin bedded siliceous, micaceous, coarse- grained quartzite, with argillitic and coarse-grained quartzitic sandstones near the base" along the Blackfoot River just east of the town of Bonner (about 10 miles east of Missoula). Thicknesses reported for the Garnet Range Formation in its type area range from 549 m to 2316 m. McGroder (1984a) summarized thickness estimates and calculations by a number of workers. Based on detailed mapping in the type area by Watson (1984) and comparison with measured thicknesses of the formation in less structurally complex areas, McGroder (1984a) used a thickness of 1100 m as a best approximation for his study of the Garnet Range Formation in its type area.

McGroder (1984a) subdivided the Garnet Range Formation into four lithofacies that he described from different areas in the Missoula region and compiled those facies into a composite stratigraphic column. His lithofacies, in ascending order, are: (1) thinly interlayered sandstone-mudstone facies, (2) planar- bedded sandstone facies, (3) cyclic hummocky-bedded facies, and (4) hummocky- and cross-bedded facies. Correlation of these facies to members in the Libby Formation is shown in Figure 10.

Whitefish Range

Whipple (1984) subdivided Missoula Group rocks in the Whitefish Range (Fig. 4) originally defined as the Rooseville Formation by Daly (1912) into the McNamara and Libby Formations. The 1220 meter thick McNamara Formation closely resembles the rocks at the type locality of the McNamara Formation in both overall appearance and stratigraphic position. The overlying Libby Formation (Fig. 10) is 975 meters thick and has been subdivided into two units (Whipple, 1984; Whipple et al., 1984). The lower unit is approximately 300 meters thick (J. Whipple, pers. comm. 1986) and consists of even, parallel laminated and hummocky cross-stratified, poorly sorted, micaceous arenite. The upper unit is approximately 700 meters thick (J. Whipple, pers. comm. 1986) and consists of interbedded, olive, arenaceous siltite and black, micaceous argillite.

The McNamara Formation in the Whitefish Range is easily correlated to its type locality near the town of Bonner. Correlation of the McNamara Formation to the lower part of the Libby Formation on Flagstaff Mountain is not quite as simple inasmuch as the base of the Libby Formation at Flagstaff Mountain contains the dark gray argillite horizon that the Whitefish Range section lacks (Fig. 10). The red zones present near the base, middle, and top of the McNamara Formation in the Whitefish Range are lacking in the lower Libby Formation. The best way to correlate units in this case is to use the top of the Bonner Quartzite as a marker and to correlate the argillitic and silty beds of the basal McNamara to the basal argillite of the Libby Formation. The overall general similarity in stratigraphic position, lithology, and sedimentary structures between the two formations warrants the correlation, but the differences in facies justify the different formation names.

The upper Libby Formation on Flagstaff Mountain correlates fairly closely with the Libby Formation mapped by Whipple (1984) in the Whitefish Range. The hummocky cross-stratified quartzite (Fig. 10) is an important marker between the two sections. The dark gray argillite below the hummocky cross-stratification on Flagstaff Mountain (Member E) is a facies

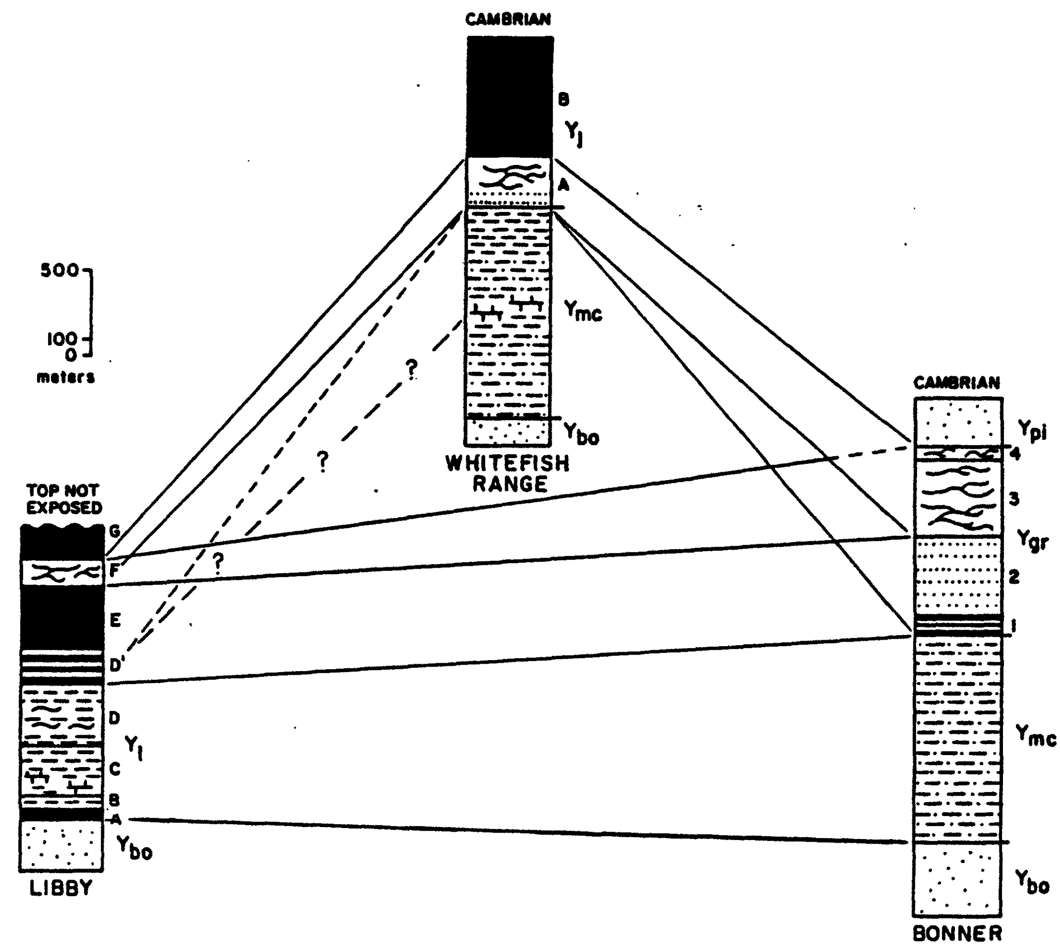


Figure 10. Correlation diagram between most complete and well-studied Libby-equivalent sections. Y_{bo} = Bonner Quartzite, Y_l = Libby Formation, Y_{mc} = McNamara Formation, Y_{gr} = Garnet Range Formation, Y_{pi} = Pilcher Quartzite. Solid black represents dark gray argillite. Otherwise, standard lithologic symbols apply (see Key to Appendix I). Units represented solely by wavy lines arehummocky cross-stratified. See Figure 26 for locations.

equivalent to green and red siltite in the upper McNamara in the Whitefish Range. Dark gray argillite above the hummocky cross-stratification at both sections is similar, but much thicker in the Whitefish Range.

Superior area

The region surrounding the towns of Superior and Alberton (Fig. 4) contains the thickest recorded exposures of Libby- equivalent rocks. Unfortunately, these rocks have received little detailed attention. Near Superior, Campbell (1960) mapped and named the Sloway and Bouchard Formations which now correspond roughly to the more recently mapped McNamara and Garnet Range Formations respectively (Harrison et al., 1986). The McNamara Formation here lithologically resembles the McNamara of the Missoula and Whitefish Range areas. It is a coarser grained lithofacies equivalent of the lower Libby Formation near Libby.

The Garnet Range Formation near Superior can be correlated to other areas by several marker units. The basal part of the formation consists of dark gray argillite interbedded with thin to medium bedded quartzite that corresponds to facies 1 of McGroder (1984) (Fig. 11). This unit may be a coarser facies of the transition zone (member D') on Flagstaff Mountain. Correlation across much of the Belt basin can be made by use of the hummocky cross-stratified quartzite present at Libby, Superior, and Missoula (Fig. 11). This lithology has a minimum thickness of 150 meters at Superior in a faulted section.

Wells (1974-1975) subdivided the McNamara and Garnet Range Formations into four and eleven units respectively based on color variations in argillite and siltite. These subdivisions are shown in Figure 11, but no confident correlations are made here. The dark gray argillite at the base of the Garnet Range Formation near Alberton (Fig. 11) probably correlates to facies 1 of McGroder (1984a) and may be a proximal lithofacies of member D' on Flagstaff Mountain.

Fishtrap Creek

Correlation to the Fishtrap Creek area (Fig. 6, 11) is extremely difficult because the section is disturbed by faulting and exposure is poor. Correlation of informal members within the Libby Formation from Flagstaff Mountain to Fishtrap Creek has already been discussed. Correlation of units at the formation level and associated problems are discussed below.

A transitional zone of alternating pink quartzites and green siltites from the upper Bonner quartzite into the overlying siltites and argillites resembles that at the type localities of both the Libby (Flagstaff Mountain) and McNamara (McNamara's Landing) Formations. The base of the overlying unit as previously discussed, consists of approximately 40 m of green siltite suggesting that this unit could be the McNamara Formation. However, parallel laminated dark gray argillite virtually identical to that at the base of the Libby Formation at its type locality overlies the green siltite. At Daisy Creek, this argillite is 22 m thick. Further south, at Deerhorn Mountain, the dark gray argillite is 35 m thick, but is interbedded with green siltite. So, even within the Fishtrap Creek region, there are facies changes complicating stratigraphic relations. Above the dark gray argillite, the rocks could be easily called Libby or McNamara Formation except for a thin red horizon, which is more characteristic of the McNamara Formation. Hummocky cross-stratified quartzite is present above a poorly exposed interval in which some section has been lost due to normal faulting. The problem of how to correlate the McNamara and lower Libby Formations is thus bracketed by the Bonner Quartzite below and by the hummocky cross-stratified quartzites above. Clearly, the problem is a matter of intertonguing facies. A stratigraphic solution is presented below.

Inasmuch as the problem involves the nomenclature of <u>formations</u> (lithologically mappable units), the solution should be resolved in terms of units that can be mapped rather than

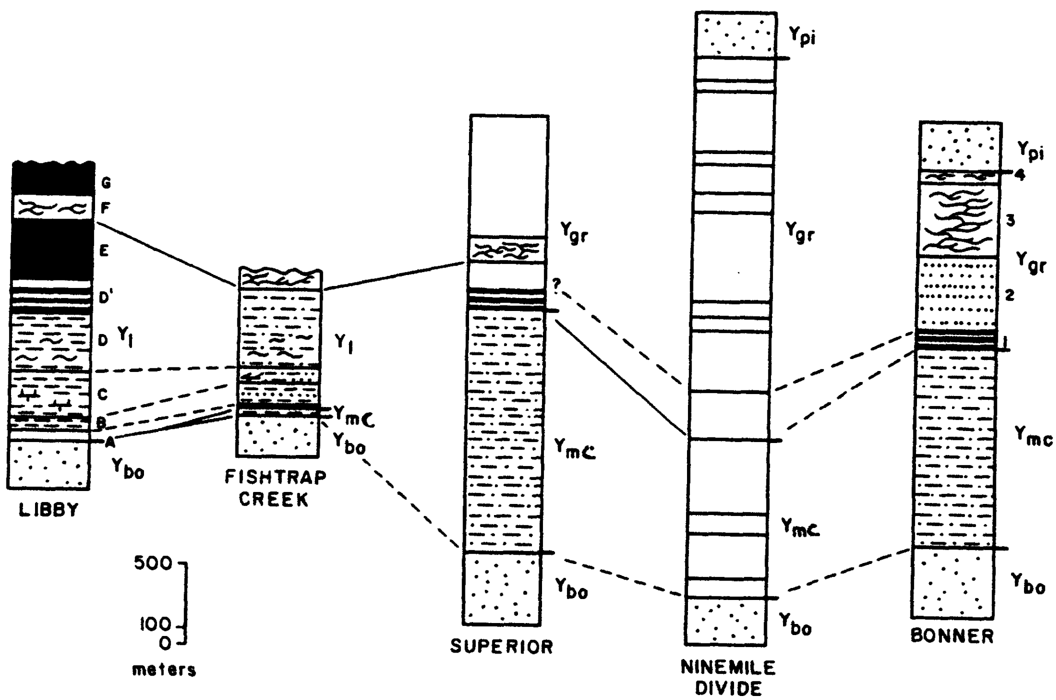


Figure 11. Correlation diagram for central Belt basin in Libby-equivalent rocks. Ybo = Bonner Quartzite, Yl = Libby Formation, Ymc = McNamara Formation, Ygr = Garnet Range Formation, Ypi = Pilcher Quartzite. Standard lithologic symbols apply (see Key to Appendix 1 on p. 158). Units represented solely by wavy lines are hummocky cross-stratified. See Fig. 26 for locations.

intercalating facies. Facies interpretations can always be made regardless of formation names. It is important to have a consistent nomenclature.

Nelson and Dobell (1961) never defined a base for the McNamara Formation at its type locality. They defined the top of the Bonner Formation as the highest pink quartzite in the Bonner/McNamara transitional zone. Gibson (1948) defined the base of the Libby Formation as the dark gray argillite overlying the Striped Peak Formation (Striped Peak member 4 = Bonner Quartzite) near the town of Libby. Based on those definitions, it seems most reasonable to define the top of the Bonner Quartzite and the base of the Libby Formation at Fishtrap Creek. The intermediate 40 m of green siltite can then be referred to and mapped as McNamara Formation as has been done by Harrison et al., 1983). Most of the section at Fishtrap Creek stratigraphically between the Bonner Quartzite and the Cambrian Flathead Quartzite is named Libby Formation. The lower part is a facies equivalent of the McNamara Formation, and at least the hummocky cross-stratified quartzite in the upper part correlates to the Garnet Range Formation. Supportive reasons for the choice of Libby Formation for this interval are:

1) Closer proximity to the type locality of the Libby Formation than to that of the McNamara Formation.

2) The improper use of the name McNamara by Nelson and Dobell (1961).

3) Adherence to the preferred practice in the North American Stratigraphic Code of defining the bottoms rather than tops of units.

Eastern Washington

Very little Libby-equivalent rock is exposed in eastern Washington, but thin exposures of dark gray argillite overlying equivalents to the Bonner Quartzite may represent the Libby Formation in this region. Miller and Clark (1975) correlated Member D of the Striped Peak Formation (Bonner Quartzite) to their Member F of the Striped Peak Formation in the Chewelah region. Glover (1978) and McMechan (1981) correlated the basal quartzite of the Mount Nelson Formation (British Columbia) to both the upper Striped Peak (Bonner) Formation and to the Buffalo Hump Formation of the Magnesite Belt near Chewelah, Washington (Campbell and Loofbourow, 1962). Correlation to the Buffalo Hump Formation is hampered by rapid facies changes in the Magnesite Belt (Campbell and Loofbourow, 1962). However, if the Buffalo Hump is correlative with Bonner-basal Mount Nelson rocks, then dark gray argillite that overlies the Buffalo Hump quartzite may be equivalent to the basal Libby Formation.

British Columbia

Correlation of the Bonner Quartzite to the northwest is critical in relating rocks in the upper Belt Supergroup to comparable units in the upper Purcell Supergroup. McMechan (1981) correlated the basal quartzite of the Mount Nelson Formation (Fig. 12) to member 4 of the Striped Peak Formation (Harrison and Jobin, 1963). Member 4 is correlative to quartzites and siltites mapped as the Bonner Quartzite throughout much of western Montana (Harrison et al., 1983; 1986; Wallace et al., 1983). This correlation is supported by unit by unit correlation between the Dutch Creek Formation and upper Wallace and Striped Peak Formations near Lake Pend Oreille (units now mapped as Snowslip, Shepard, and Mount Shields Formations in Montana, Harrison et al., 1983; 1986; Wallace et al., 1983).

Although the basal quartzite of the Mount Nelson Formation can be traced to many sections in British Columbia (Fig. 12), McMechan (1981) does not correlate this quartzite to the Phillips

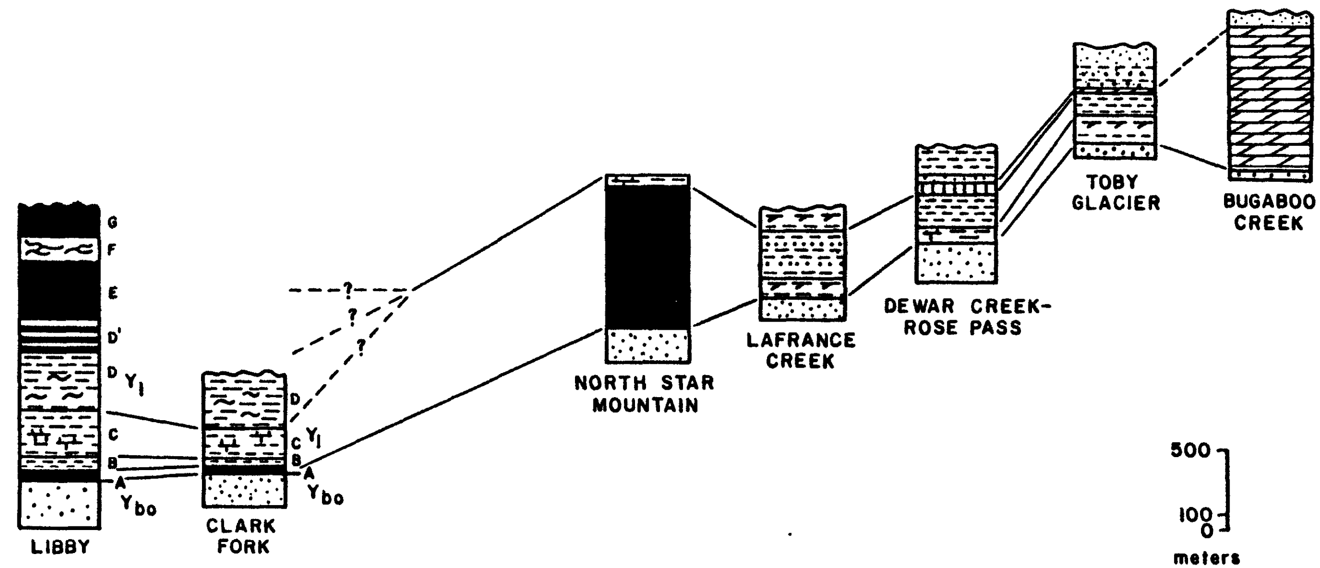


Figure 12. Correlation diagram for northwestern Belt basin in Libby-equivalent rocks. Y_{bo} = Bonner Quartzite, Y_l = Libby Formation. Rocks to the right of Clark Fork section are Mount Nelson Formation. Standard lithologic symbols apply (see Key to Appendix 1 on). Units represented solely by wavy lines are hummocky cross-stratified. See Fig. 26 for locations.

Quartzite. Examination of sections from Clark Fork to the Whitefish Range shows stratigraphic and sedimentologic similarities indicating that all of those quartzites are equivalent to the Bonner Quartzite (they are mapped as Bonner Quartzite and include siltite facies for the Bonner by Harrison et al., 1983). McMechan's reason for not correlating as mentioned is because of an apparent pinch-out of the Phillips (Bonner) Quartzite on the map of Leech (1960). I suggest that that area needs to be re-examined and that it is likely that the Bonner/Phillips/Striped Peak 4 are all correlative with the basal quartzite of the Mount Nelson Formation.

The rocks overlying the Bonner/basal Mount Nelson quartzite are roughly similar, but undergo facies changes over long distances. The dark gray argillite at the base of the Libby Formation on Flagstaff Mountain thickens considerably to a section measured by Glover (1978) in the Summit Creek area (Fig 12). More northerly exposures of the Mount Nelson Formation are argillitic, but more carbonate-rich in their basal parts (Dewar Creek-Rose Pass, LaFrance Creek, Toby Glacier, and Bugaboo Creek). The basal carbonate-bearing argillites of all of these Canadian localities except Bugaboo Creek are overlain successively by argillite and carbonate units. These units vary slightly in thickness (substantially at Summit Creek), but all correspond generally to the lower three units of the Libby Formation at Clark Fork and on Flagstaff Mountain.

It is tempting to speculate that the quartzites in the upper Mount Nelson Formation at Toby Glacier and Bugaboo Creek might be hummocky cross-stratified. Inasmuch as more carbonate is in those sections, sedimentation might have been somewhat slower resulting in thinner deposits in the lower Mount Nelson Formation. A thinner basal quartzite is consistent with suggestion of slower sedimentation possibly as a result of being further from southern source areas. If thinner deposits account for longer amounts of Libby time, then, based on the stratigraphic position of the upper Mount Nelson quartzites (Reesor, 1984, not observed in this study) at Toby Glacier and Bugaboo Creek, it is not unreasonable to expect them to be hummocky cross-stratified.

Summary

Nearly basinwide correlation of the Libby Formation and its equivalents can be accomplished with the aid of two key marker units: (1) The distinctive reddish Bonner Quartzite highlights the base of the Libby and McNamara Formations in Montana. The lithostratigraphically equivalent white quartzite that forms the basal unit of the Mount Nelson Formation facilitates correlation of this British Columbian formation to the Bonner Quartzite and Libby Formation, and (2) a second important marker unit is the hummocky cross-stratified quartzite that strengthens previously established (Winston, 1986a) correlation of the upper part of the Libby Formation to the Garnet Range Formation. It is postulated that the quartzite in the upper exposed parts of the Mount Nelson Formation at Toby Glacier and Bugaboo Creek may be correlative with the hummocky cross-stratified beds.

Correlation of other units within the Libby and equivalent formations is complicated by facies changes. Some of the units can be correlated locally (up to 60 miles). The dark gray argillite (member A) at the base of the Libby Formation can be traced to Clark Fork and Fishtrap Creek. This member may be present at the top of the section in the Magnesite Belt in eastern Washington. Member A is apparently considerably thicker in the Summit Creek area (Glover, 1978). Argillitic carbonate units just above the basal Mount Nelson quartzite at Dewar Creek-Rose Pass, LaFrance Creek, and Toby Glacier may all be facies equivalents of the basal Libby dark gray argillite member.

A carbonate zone several hundred meters above either the Bonner or basal Mount Nelson quartzite serves as a marker that can be traced west to Clark Fork, south to Fishtrap Creek, east to the McNamara Formation in the Whitefish Range, and northwest to sections of Mount Nelson Formation at LaFrance Creek, Toby Glacier, Dewar Creek-Rose Pass, and possibly Summit

Creek.

PETROGRAPHY OF SILTSTONE AND SANDSTONE

Most of the terrigenous clastic rocks in the Libby Formation are composed of silt- and clay-sized particles. Nineteen of the coarsest samples were selected for detailed mineralogical and textural analysis of detrital components. Because many of the samples contain a high percentage of matrix, five hundred points were counted for each sample.

Mineralogy

Point count analysis reveals significant differences in detrital composition between the lower and upper Libby Formation. The results of the point count analyses are summarized in Table 1. These rocks are primarily arkosic and subarkosic wackes and arenites (Pettijohn et al., 1973 modification of Dott, 1964 classification). A suite of samples from the Flagstaff Mountain section is characterized by subarkosic wackes in the lower Libby Formation and arkosic wackes and arenites in the upper Libby Formation. Limited samples from Fishtrap Creek and Clark Fork show mineralogical trends that are consistent with the distribution of clastic material in the Flagstaff Mountain section.

Differing detrital mineralogy between the upper and lower Libby Formation is clearly delineated in quartz/feldspar/lithic fragments (QFL) and quartz/K-feldspar/plagioclase (QKP) ternary plots (Fig. 13). In both cases, the lower Libby Formation has a quartz-rich composition, whereas the upper Libby Formation is richer in feldspar when analyzed for QFL (Fig. 13A), and richer in plagioclase (twinned) in QKP analysis (Fig. 13B). Samples in the lower Libby Formation from members B and C on Flagstaff mountain, member C at Fishtrap Creek, and member C at Clark Fork each cluster into distinct QKP compositions (Fig. 13B). These unique populations for particular members in different areas may signal subtle shifts in source terrain during deposition of the lower Libby Formation. Conversely, shifts in detrital composition in the lower Libby Formation may reflect winnowing of more unstable detrital grains as the samples became more compositionally mature up section in the lower Libby Formation.

Other grain types in the Libby Formation include argillaceous rip-up clasts, zircon, and white mica. The percentage of detrital mica is variable with stratigraphic position (Table 1).

Textures

Semi-quantitative textural analyses of these siltites show differences in grain size, sorting, and sphericity between the upper and lower parts of the Libby Formation (Fig. 14). Roundness of grains does not vary significantly with stratigraphic position.

Size

The maximum, minimum, and average size range of grains were measured for the coarsest samples in the Libby Formation. Average grain size ranges from coarse silt to very fine sand. Several samples have grains ranging into the fine-grained category.

Grain size varies with stratigraphic position in the Libby Formation. At the Flagstaff Mountain section (Fig. 14), there is a general increase in grain size up section from predominantly very coarse silt in the lower Libby Formation to a mixture of very coarse silt and very fine sand-sized grains in the upper Libby Formation. Data are insufficient at this time to recognize grain size trends at other localities.

Sample No.	Meters	Quartz	Undulatory	Polyxtln	Embayed	K-spar	Microcline	Plagioclase	Lithics	White mica	Muscovite	Chlorite
FE-132	1509	18.2	0.0	0.0	0.0	1.8	0.0	20.8	0.8	3.0	0.0	1.6
FE-122	1497	7.2	0.0	0.0	0.0	2.2	0.0	20.8	0.0	6.2	0.6	0.2
FE-84	1459	13.0	0.0	0.0	0.0	0.6	0.0	9.6	0.0	11.8	1.0	0.2
FE-78a	1453	12.2	0.0	0.0	0.0	3.6	0.0	15.8	1.0	8.4	1.4	0.0
FE-78	1453	28.4	0.0	0.0	0.0	7.6	0.0	14.0	0.0	4.6	0.4	0.2
FE-42	1417	18.6			0.0	1.4	0.0	9.0	0.0	6.8	0.6	0.6
FE-34	1409	7.2		0.0	0.0	1.0		9.8	0.0	12.2	0.6	0.2
FE-20	1395	14.4			0.0	2.0		11.2	2.2	3.6	10.2	0.0
FC-227	547	31.0	1.6			8.4		5.2	27.4	4.4	0.0	0.2
FC-206	526	19.2	0.0	0.0	0.0	2.2	0.0	1.2	0.0	7.2	0.0	0.4
FB-177	217	23.6		0.0	0.0	3.4	0.0	4.4	0.0	8.4	0.2	0.0
FB-137	187	21.8		0.0	0.0	0.2	0.0	5.2	0.01	23.2	0.6	0.2
FB-104	144	17.4	0.0	0.0	0.0	0.8	0.0	5.0	0.0	8.0	0.0	0.0
FB-50	90	28.6		0.0		3.2	0.0	5.4	0.6	8.2	2.4	0.2
FB-14	54	33.8			0.0	4.2	0.0	4.6	0.0	13.4	0.0	1.0
DCB-39	—	18.0	0.0	0.0	0.0	2.6	0.0	3.6	0.0	4.4	1.0	
DCB-1	—	38.2		0.0	0.0	8.0		12.4	2.4	9.8	0.0	1.0
DCA-67	—	10.6	0.0	0.0	0.0	1.0	0.0	2.6		16.4	1.8	
CFQ-5	—	9.6	0.0		0.0	2.4	0.0	2.2	1.4	4.6	0.4	0.0

Sample No.	Meters	Heavy Minerals	Altered	Opaques	Hematite	Org. Wisps	Matrix	Chert cement	Carbonate	Rxtl/auth.	Brown crud	Total
FE-132	1509	0.0	0.8	0.0	0.0	0.0	8.4	2.8	7.4	36.4	0.0	100.0
FE-122	1497	0.2	7.4	0.0	0.0	1.2	32.2	3.8	3.6	14.2	0.0	99.8
FE-84	1459	0.0	0.8	0.0	0.0	0.0	34.0	12.6	7.8	8.4	0.0	100.0
FE-78a	1453	0.0	2.4	0.0	0.0	0.0	36.2	12.8	4.8	1.4	0.0	100.0
FE-78	1453	0.4	6.6	0.0	0.0	0.0	12.0	10.8	4.6	12.4	0.0	100.0
FE-42	1417	0.2	1.6	0.0	0.0	0.0	2.2	17.4	0.0	40.4	1.2	100.0
FE-34	1409	0.0	2.8	0.0	0.0	0.0	44.8	6.8	2.6	9.2	2.8	100.0
FE-20	1395	0.2	2.2	0.0	0.0	0.0	51.8	1.8	0.2	0.0	0.0	100.0
FC-227	547		0.0	0.2	0.0	0.0	5.6	0.0	15.0		0.0	92.6
FC-206	526	0.2	0.0	0.2		0.0	54.4	14.6	0.0	0.0	0.0	99.8
FB-177	217		0.0	0.2	0.4	0.0	58.0	0.4	0.0	0.0	0.0	99.0
FB-137	187	0.0	3.8	0.0			40.0	0.4	0.6	1.4	0.0	101.6
FB-104	144	0.2	0.0	0.0	2.8	0.0	25.0	30.4	12.4	0.0	0.0	100.0
FB-50	90	0.4	0.0	1.2	0.0	0.0	39.0	1.2	11.6	0.0	0.0	100.0
FB-14	54	0.2	0.2	0.0	0.0	0.0	29.4	0.4	0.0	0.0	0.0	87.2
DCB-39	—	0.0	1.6	0.0	0.0	0.0	9.2		42.0	17.4	0.0	100.0
DCB-1	—	0.8	13.4	0.0	0.4	0.0	3.8	0.4	0.0	9.4	0.0	100.0
DCA-67	—	0.0	0.0			0.0	62.2	0.6		3.0	0.0	98.2
CFQ-5	—	0.0	0.0	0.0	0.0	0.0	78.4	0.6	0.4	0.0	0.0	100.0

Table 1. Point count data for siltite samples in the Libby Formation. Each analysis based on 500 points. Abbreviations are explained as follows: meters - meters above base of Libby Formation at Flagstaff Mountain section (only applies to FB, FC, and FE); polyxtln -polycrystalline quartz; altered - altered grains (usually feldspar); org. wisps - wisps of organic material; rxtl/auth - recrystallized or authigenic silicates; brown crud - unidentified light brown material.

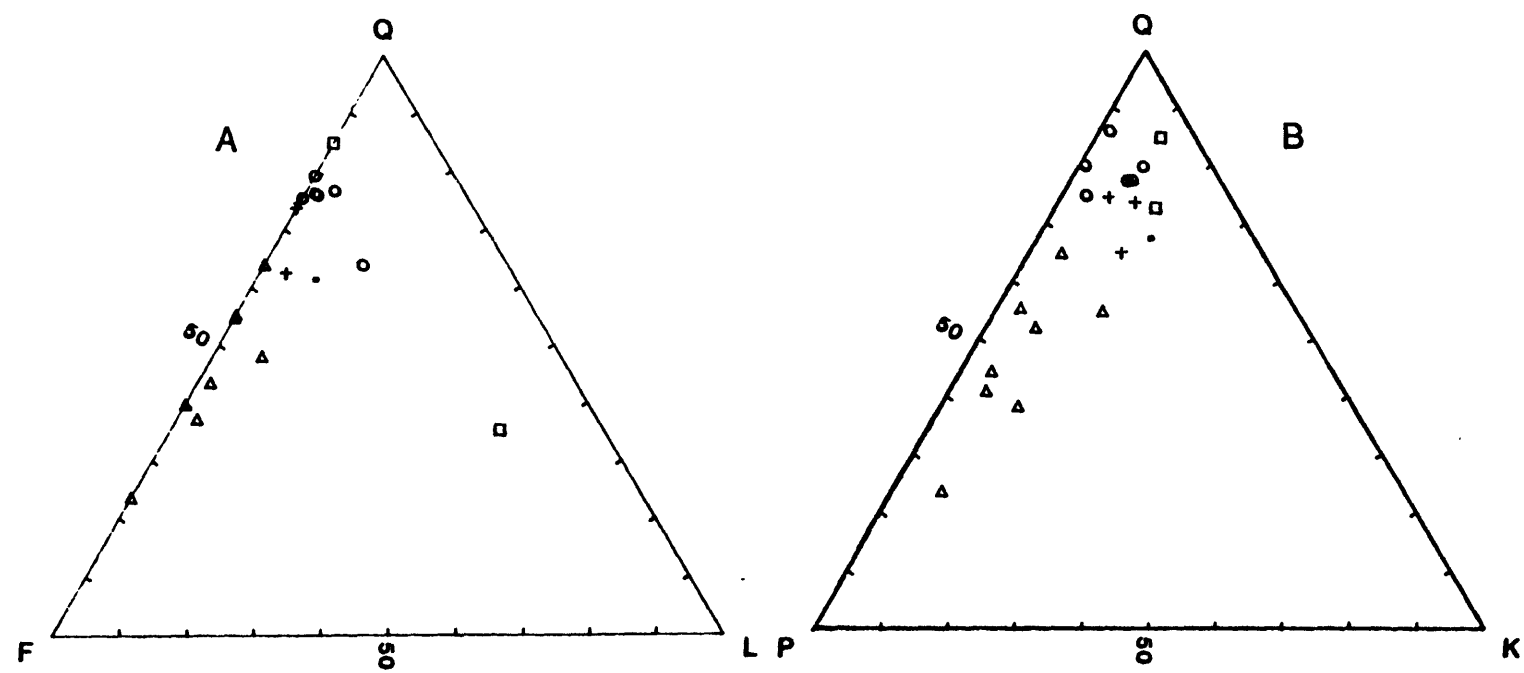


Figure 13. Ternary plots of detrital grain variation in the Libby Formation. Each symbol based on a point count of 500 points. Key to symbols: circle and square - lower Libby Formation at Libby;triangle - upper Libby Formation at Libby; plus (+) - lower Libby Formation at Fishtrap Creek; dot - lower Libby Formation at Clark Fork.

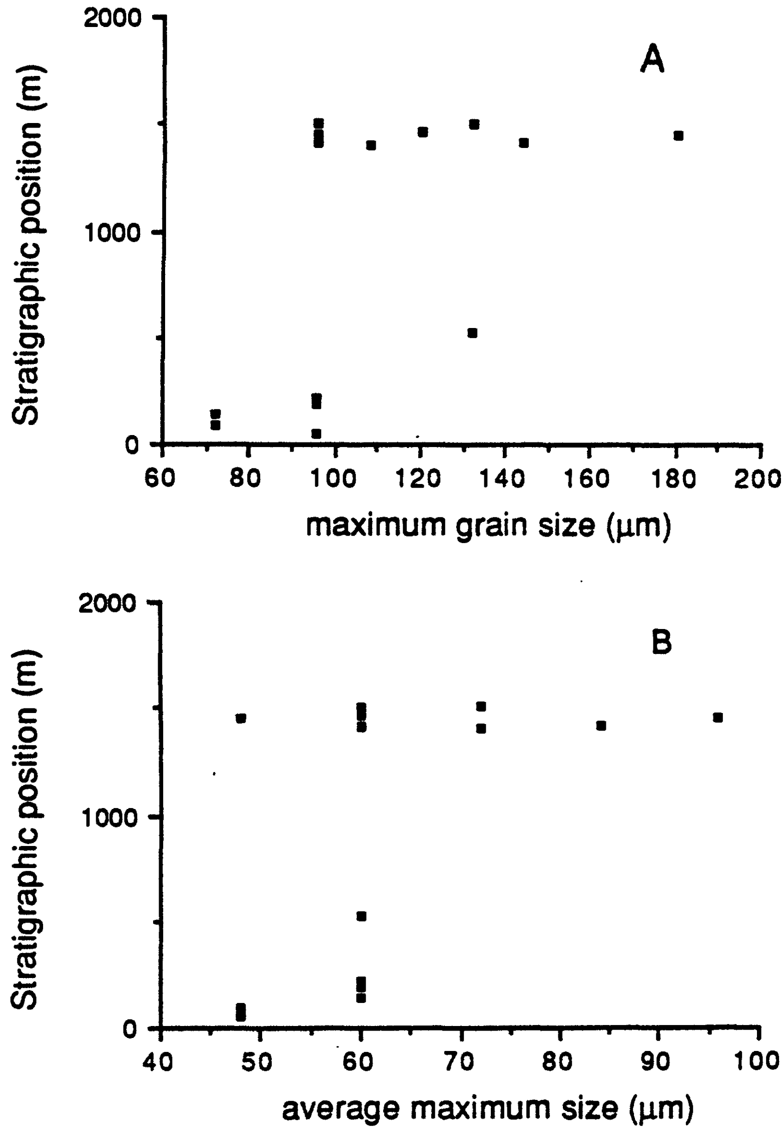


Figure 14. Scatter plots of grain diameter variation vs. stratigraphic position in the Libby Formation on Flagstaff Mountain. A. Absolute maximum grain diameter. B. Coarsest grain sizes in the general size range of a given sample. 0 on vertical scale represents base of the Libby Formation.

Sorting

Sorting also varies with stratigraphic position. Standard sorting categories were assigned numerical values (Fig. 15A). Semi-quantative estimates of sorting indicate that the rocks on Flagstaff Mountain are progressively more poorly sorted up section. Samples in the lower Libby Formation are generally well sorted, whereas upper Libby rocks are mostly moderately sorted, and are commonly poorly sorted.

Roundness

Visual estimates were made of the most common grain roundnesses for each sample. As with sorting, standard roundness/angularity categories were assigned numerical values (Fig. 15C). Roundness within samples is quite variable, ranging from well-rounded or rounded ot angular within a thin section. There is no significant variation in grain roundness with stratigraphic position.

Sphericity

Sphericity of detrital grains varies considerably within samples. Semi-quantitative analysis was performed by visual estimate of the per cent ratio of grains with high and low sphericity. These percentages were then used to plot sphericity variation with stratigraphic position. Sphericity decreases up section. Sphericity in the upper Libby Formation ranges from high sphericity to low sphericity, but grains in the lower Libby Formation generally do not contain more than 50% low sphericity grains (Fig. 15B).

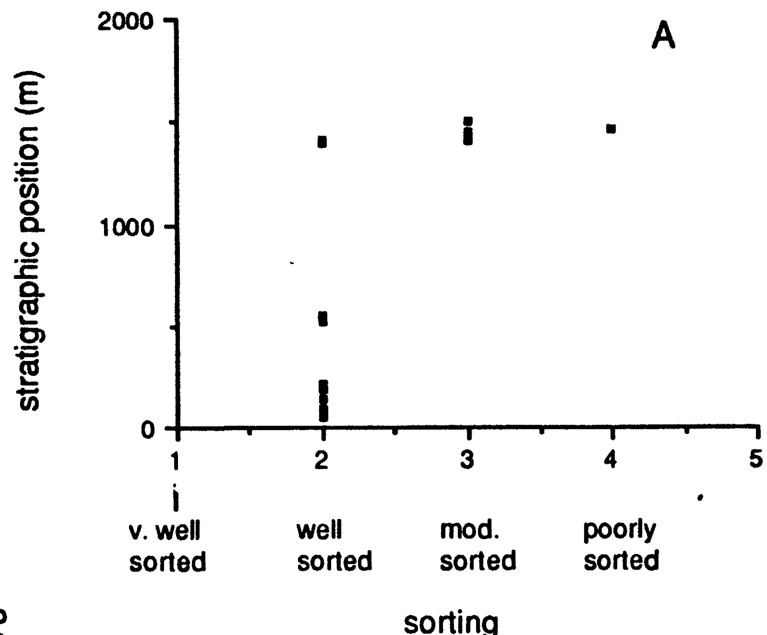
Diagenesis

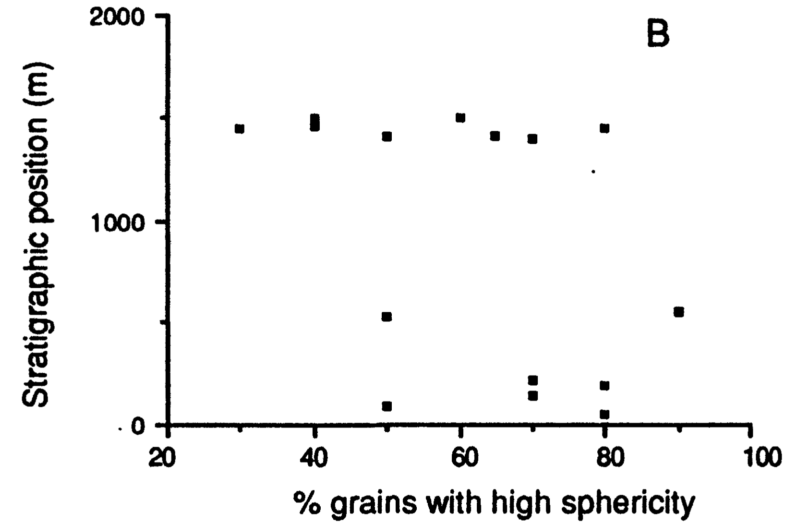
Interstitial material in the Libby Formation is predominantly detrital matrix (Table 1). However, a variety of authigenic minerals are present in variable quantities. These minerals include: pyrite (<1/%), chert (<1%-30%), equant/subequant quartz (<1%-40%), chalcedony (<2% in some oolitic mudstones), and patchy carbonate cement (<1%-42%).

The degree of feldspar alteration is, at present, unknown. Many of the feldspars in the Libby Formation are very fresh and unaltered. In some samples, grains interpreted as altered feldspar reach percentages as high as 13%. The unaltered, twinned feldspars could be albitized or could be well-preserved as a result of low permeabilities in these mudrocks. Given that many feldspars are extremely altered in Beltian "sandstones" in a number of formations to the south and east of this study area (R. Horodyski and K. Liebold, pers. comm., 1987), it is possible that the fresh, unaltered feldspars in the Libby Formation are the result of good preservation. The fine grain size makes interpretation of diagenetic relationships in these rocks difficult to assess.

Clastic Petrofacies Analysis

Recent work by Land and Milliken (1981) and Boles (1982) suggests that petrofacies analysis of sandstones using feldspar may be suspect due to albitization of both plagioclase and potassium feldspar. However, Blatt (1985) recently championed the potential usefulness of mudrocks in petrofacies analysis. Low permeabilities generally prevent alteration of feldspars except for several examples from alluvial fan muds interpreted as deposited in at least somewhat arid environments (Hubert and Reed, 1978; Walker, 1978; Goodbread, 1978; and Blatt and Caprara, 1985). Lack of described examples of altered feldspars in mudrocks from other environments could be a function of lack of study. This analysis provides an example of successful silt petrofacies analysis, and shows that significant information can be gained from such a study even if feldspars have been albitized.





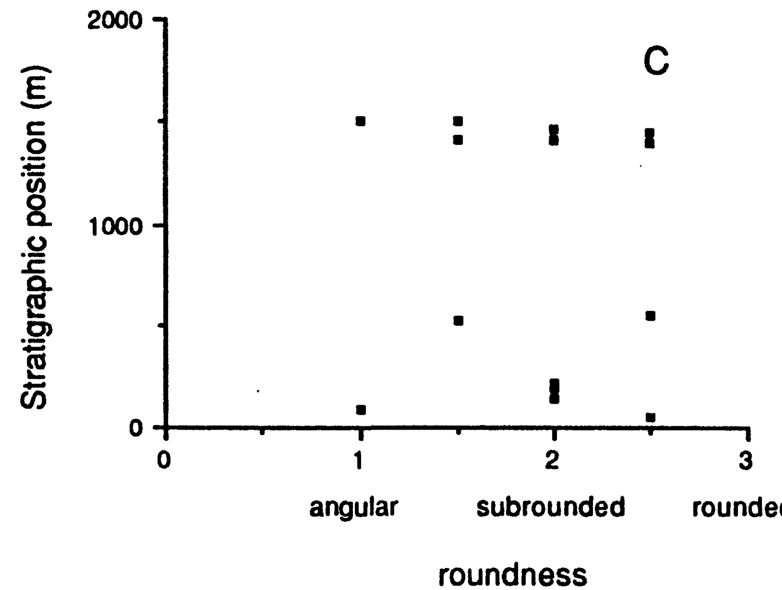


Figure 15. Semiquantitative scatter diagrams of textural variation (other than size) of framework grains plotted against stratigraphic position in the Libby Formation on Flagstaff Mountain. A. Sorting. B. Sphericity. C. Roundness. For A and C, numerical values were assigned to general textural categories to facilitate plotting. 0 on vertical scale represents base of the Libby Formation.

Detrital Mica

Previous petrofacies research on post-Bonner Formation rocks in the Belt Supergroup is restricted to the observation by Bleiwas (1977) that there is a marked upward increase in the percentage of detrital mica across the McNamara-Garnet Range Formation contact. The results of this study do not show any systematic trends in the distribution of detrital mica in the stratigraphically equivalent Libby Formation. The lack of any such trends could be the result of the fact that the Libby Formation represents a more distal setting than the McNamara- Garnet Range transition, or the result of influence by a source terrain different from that which supplied sediment to the sections studied by Bleiwas (1977) in the central Belt basin.

QFL and QKP variation

Examination of the relative percentages of detrital grain types shows distinct compositional fields for the lower and upper Libby Formation. The lower Libby Formation is quartz rich and consists of texturally mature rocks. The upper Libby Formation contains plagioclase that is texturally less mature than the grains in the lower Libby Formation.

Potential source areas

The commonly invoked sources for Belt rocks are near the southern and eastern margins of the Belt basin (Weed, 1899; McMannis, 1963; McClerman, 1969; Keefer, 1972). These rocks are usually gneisses with variable percentages of K-feldspar and plagioclase, although diorite has intruded some of the gneisses (Keefer, 1972). Another possible source for plagioclase-rich rocks in the Libby Formation might be the Purcell lava.

Continued petrofacies analysis of the upper Libby Formation as compared with composition of potential source terrains might reveal the source(s) from which the upper Libby plagioclases were derived. If the plagioclase is not albitized, geochemical comparison of feldspars in the Libby Formation with basement terranes and the Purcell lava might prove fruitful.

Implications of feldspar alteration

Although fresh-looking feldspars are common in the siltites of the Libby Formation, detrital composition may have been influenced by alteration of feldspar. The low percentages of K-feldspar in the Libby Formation are striking. For plagioclase to so dominate the detrital composition, either the source terrain had to be very rich in plagioclase with minimal reworking of the sediment (otherwise unstable plagioclase would have been altered and removed preferentially over more stable K-feldspar), or original K-feldspar has either been albitized or altered to clay minerals. The increased amount of plagioclase in the upper Libby Formation could be a function of enhanced permeability resulting in albitization of plagioclase or potassium feldspar. Many of the upper Libby samples are arenites, whereas all of the samples from the lower Libby Formation on Flagstaff Mountain are wackes. Land and Milliken (1981) noted a progressive loss of K-feldspar with depth in Oligocene Gulf coast sediments. Eslinger and Sellars (1981) noted a similar relationship in the Belt Supergroup near Clark Fork, Idaho. Both studies attribute the decrease in K-feldspar to alteration of feldspar to clay minerals. It first needs to be clearly demonstrated, however, that the decrease in K-feldspar does not reflect changes in source terrain or transport history.

Application of silt petrofacies to tectonic regimes

Petrofacies analysis has not been applied to many Proterozoic rocks, due to lack of confidence regarding the tectonic setting of the source terrain(s) (Dickinson and Suczek, 1979). Comparison of QFL ternary plots for the lower Libby Formation (Fig 13) with ternary plots compiled by Dickinson and Suczek (1979) for Phanerozoic petrofacies trends is consistent with derivation of detritus from a craton interior. QKP data could be interpreted as craton interior, uplifted basement, collision orogen, or foreland uplift.

Comparison of QFL plots for the upper Libby Formation with the diagrams discussed above is consistent with the uplifted basement source inferred for the upper Libby Formation from limited knowledge of Belt source terrains. The best fit obtained by comparing QKP data from the upper Libby Formation with the diagrams of Dickinson and Suczek (1979) is with a magmatic arc, but this result is grossly inconsistent with what is known about Beltian source terrains. However, an uplifted basement source, as compiled by Dickinson and Suczek (1979) is moderately consistent with the upper Libby data.

This preliminary analysis suggests that at least in some cases QFL analysis may be applicable to Proterozoic rocks, but variation in QKP solutions for source terrains indicates that this latter approach does not work in Middle Proterozoic rocks.

Summary

This preliminary study provides a Proterozoic example of mudrock petrofacies analysis. Further work needs to be done to determine the potential of diagenetic alteration of detrital grains in mudrocks. Nevertheless, even if the feldspars in the Libby Formation have been altered, their greater abundance in the upper Libby Formation attests to a provenance shift that is at least consistent with what little is known about Beltian source terranes.

This study also demonstrates that QFL/QKP analysis can, with care, be successfully applied to Middle Proterozoic rocks. QFL plots for the lower Libby Formation strongly suggest derivation of detrital material from a cratonic source, whereas QKP analysis suggests four possible tectonic source terrains, only one of which (cratonic) is consistent with what is known about Beltian source terrains. QFL plots for the upper Libby Formation clearly suggest an uplifted basement source from a continental block province. QKP plots are more ambiguous suggesting either magmatic arc or uplifted basement.

OOLITES

Oolitic beds up to 1 m thick occur sporadically in Member C on Flagstaff mountain, at Clark Fork, and at a section exposed in a quarry along Lake Creek, approximately six miles south of Troy, Montana. Many of these beds pinch out laterally over several to a few tens of meters; however, their maximum lateral continuity is not known, inasmuch as a few oolitic beds on Flagstaff Mountain maintain their thickness laterally until they are covered by talus.

Petrography

Most of the oolitic samples examined petrographically from Flagstaff Mountain and Clark Fork are oolitic mudstones associated with ripped-up mud clasts. Many of these ooids have been silicified, whereas others are recrystallized to calcite. Minor amounts of chalcedony and blocky quartz cement are present in interstitial areas. Oolitic grainstones from the Lake Creek quarry preserve a spectacular array of ooid fabrics and a complex distribution of cements and replacement fabrics. The remaining discussion of ooid fabrics and cements focuses on the Lake Creek quarry samples.

Ooid types

The highly variable internal structures of these ooids are summarized in Figure 16. Ooids can be subdivided into calcitic and cherty types.

Calcitic ooids. Radial/concentric ooids (Fig. 16, Fig. 17A,B) represent the most abundant calcitic ooid type (Table 2). These ooids are composed of fine crystals of needle-like habit arranged in a radial pattern. Distinct concentric rings are also common in these ooids. Ooids composed of coarsely crystalline, twinned calcite are also abundant. These ooids are preserved in several habits including: ooids composed of a single unit crystal of twinned calcite, ooids composed of two or more concentric zones of calcite, each zone with its own twin orientation, and ooids with two or more concentric zones of twinned calcite that are divided, within each zone, into areas of uniquely twinned calcite (Figs. 16, 17C,D). It is possible that some of this variety is due to differing angles of view along the plane of the thin section. Many of the ooids replaced by coarse twinned spar preserve ghosts of concentric structure. Multiphase ooids consisting of both radial and twinned calcite are also present (Fig. 16). Some of these ooids have radial cores surrounded by a twinned cortex (Fig. 17E), and others have a twinned core surrounded by a radial cortex. Less common ooid types include: ooids with an untwinned blocky calcite core and a radial cortex and rim, ooids composed of irregular calcite (Fig. 16, Fig. 17F), and ooids composed of multiple nucleii (small ooids and intraclasts).

Cherty ooids. Patterns of silicification are variable in the ooids from Lake Creek quarry. Macroscopically, silicification occurs in discrete, irregular, discontinuous zones approximately one cm thick. Some ooids in these zones are completely silicified (Figs. 16, 18A), whereas others are only partly silicified (by chert) in a variety of ways including: silicified cores/calcitic rims, silicified rims/calcitic cores, calcitic cores and rims/concentric cherty zones within the ooid, and ooids that are nearly completely silicified, but have several thin concentric zones of calcite preserved within them (Figs. 18B- E). Some of the dominantly silicified ooids contain rhombic and irregularly shaped calcite crystals. Minor amounts of megaquartz have partially replaced some ooids in a selective fashion. At a few places, tiny fingers of quartz have followed the concentric zonation within an ooid (Figs. 17B, 19). Selective silicification by both chert and megaquartz along certain concentric laminae within ooids may have been triggered by clay- rich zones along those laminae.

Cements and replacement fabrics

A complex array of a few cement morphologies characterizes the interstitial areas in these oolitic grainstones. The main types of cements (summarized in Fig. 20) are: bladed crystals (Fig. 17B), blocky spar (Fig. 17A), coarse crystals of twinned calcite, and microspar (Fig. 17A). The bladed crystals apparently nucleated on ooid surfaces. These crystals compose partial isopachous rims around some ooids, but no examples of complete preservation of such a rim have been observed. The bladed crystals now compose approximately one per cent of the rock, but they may have been more ubiquitous earlier in the history of the rock. The blocky interstitial cement fills the central parts of many interstitial areas and comprises 27-30% of the rock (Table 2). In some areas, the blocky spar coarsens toward the centers of interstitial voids. Large crystals of twinned calcite occupy some void spaces and occasionally extend into adjacent ooids. Microspar is also present as both a replacement of grains and a fill in interstitial areas.

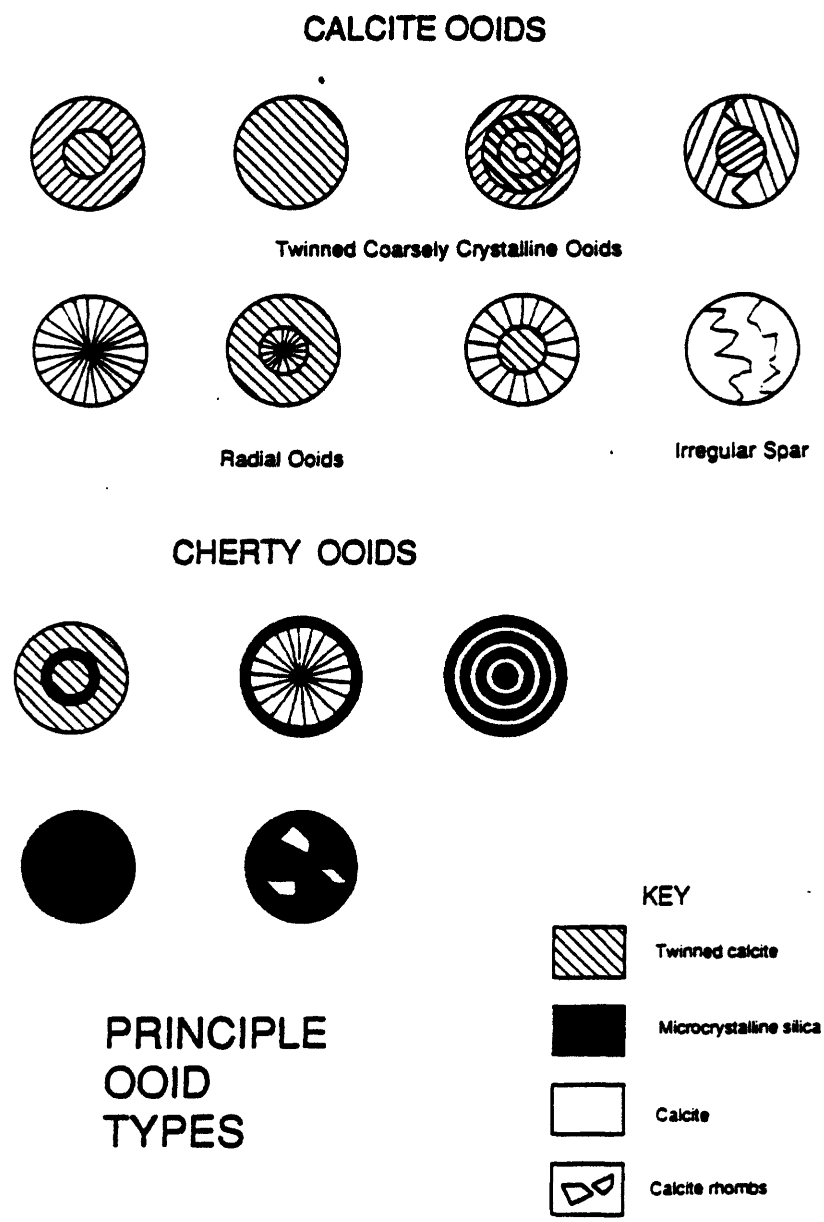


Figure 16. Generalized diagram showing most common ooid types from Member C of the Libby Formation in the Lake Creek Quarry, about 6 miles south of Troy, MT.

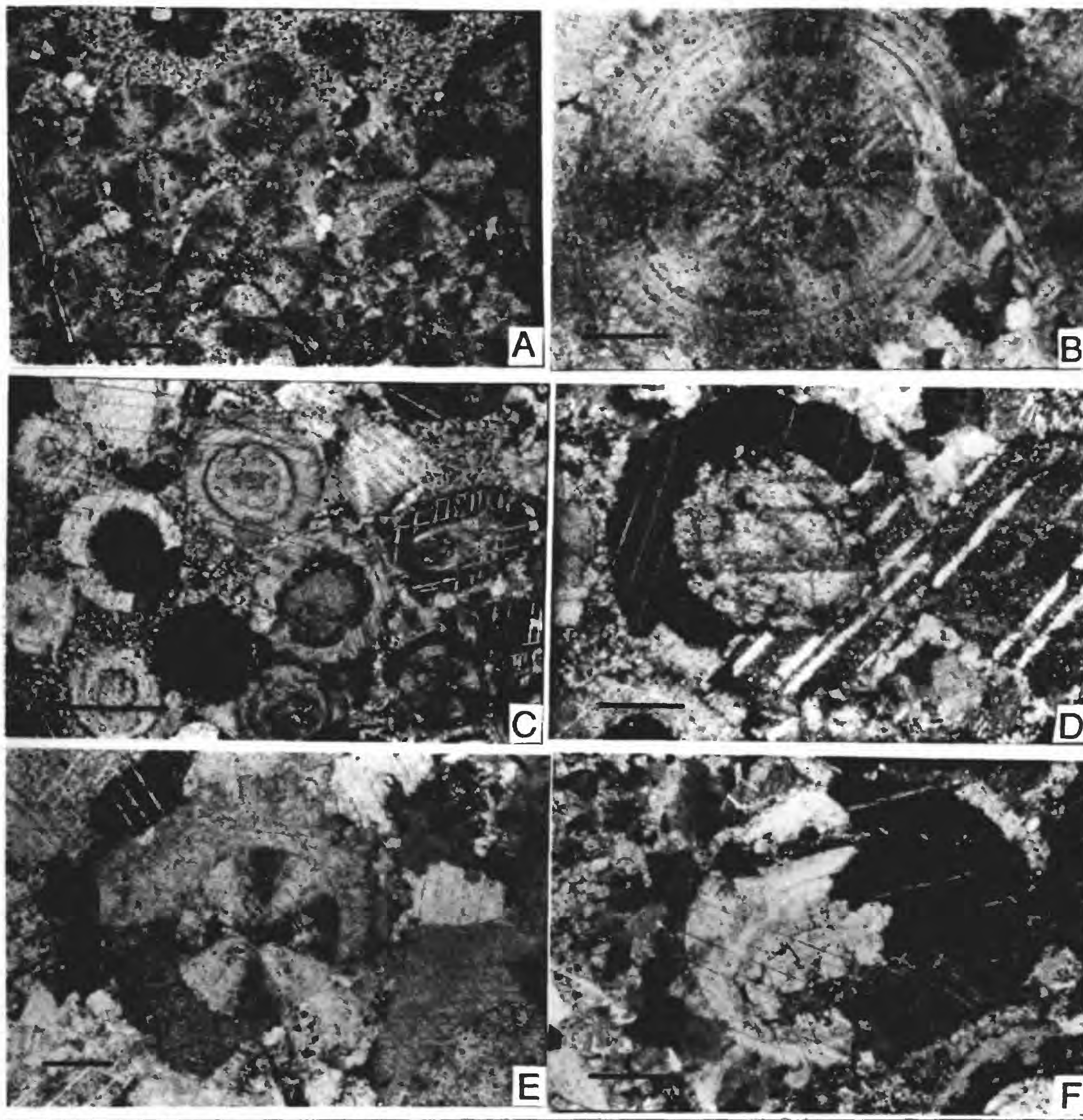


Figure 17. Photomicrographs of ooids from Lake Creek Quarry. A, B. Radial ooids. Right side of ooid in B is selectively replaced by quartz. C, D. Ooids replaced by coarse, twinned calcite. Note relic concentric structure in replaced ooids in C. E. Ooid with radial calcite core; cortex and rim replaced by coarse twinned spar. F. Ooid replaced by irregular spar. All photos taken under crossed polarizers. Bar scales: A, C = 1.0 mm, B, D-F = 0.3 mm.

	Count #1 Vol. %	Count #2 Vol. %
Radial ooids	33.6	29.0
Twinned spar	11.9	15.9
Microspar	14.6	12.0
Irreg. calcite	2.6	2.3
Bladed	1.0	0.6
Blocky calc.	27.0	29.3
auth. chert	5.0	7.3
auth. qtz.	3.6	1.6
detrital gns.	0.3	0.0
auth. feldspar	0.0	0.3
intraclasts	0.0	1.3

Table 2. Point count analyses of oolitic grainstone from Lake Creek Quarry (Member C - Libby Formation) near Troy, Montana. Each count based on 300 points.

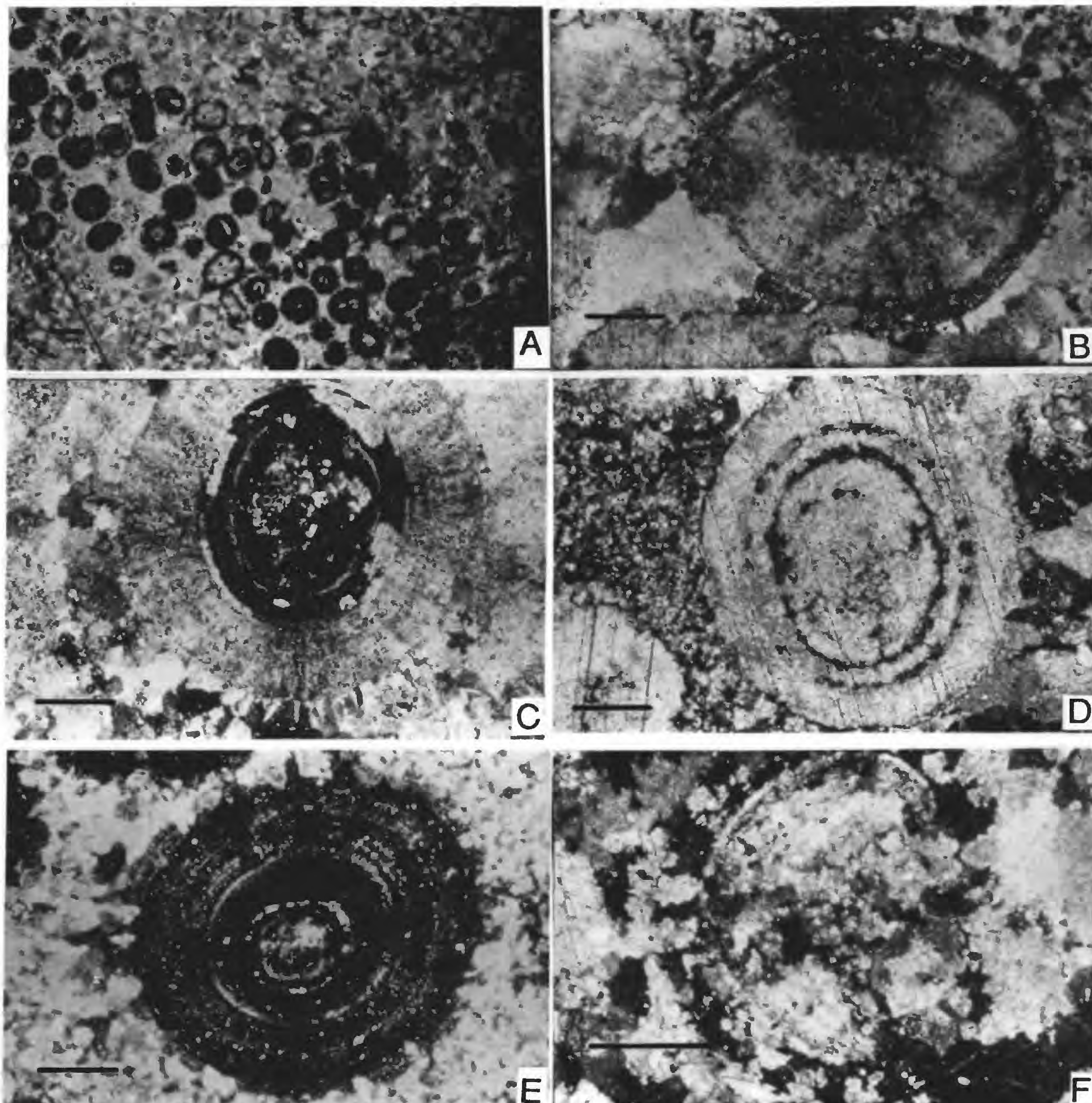


Figure 18. Photomicrographs of Lake Creek Quarry ooids. A. Band of silicified ooids (black ooids trending NW-SE in photo). B. Radial ooid with thin chert rim (speckled black). C. Radial ooid with chert core (black). Note bladed spar at southern margin of ooid. D. Coarsely crystalline calcite ooid with crudely concentric chert laminae in cortex.
E. Silicified ooid with thin concentric relics of calcite (white).
F. Silicified ooid extensively replaced lby irregularly shaped calcite crystals. All photos taken under crossed polarizers.
Bar scales: A = 2.0 mm, B-F = 0.3 mm.

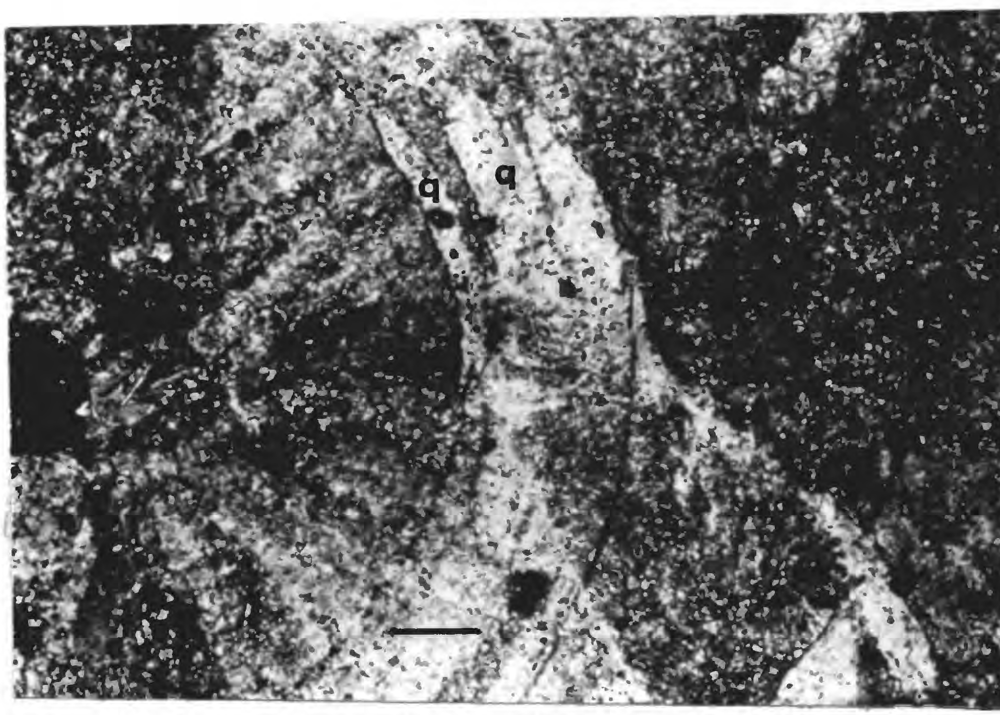


Figure 19. Photomicrograph of ooid from Lake Creek Quarry. Radial calcite ooid selectively replaced by quartz (q), perhaps, in part, along clay-rich laminae. Photo taken under crossed polarizers.
Bar scale = 0.1 mm.

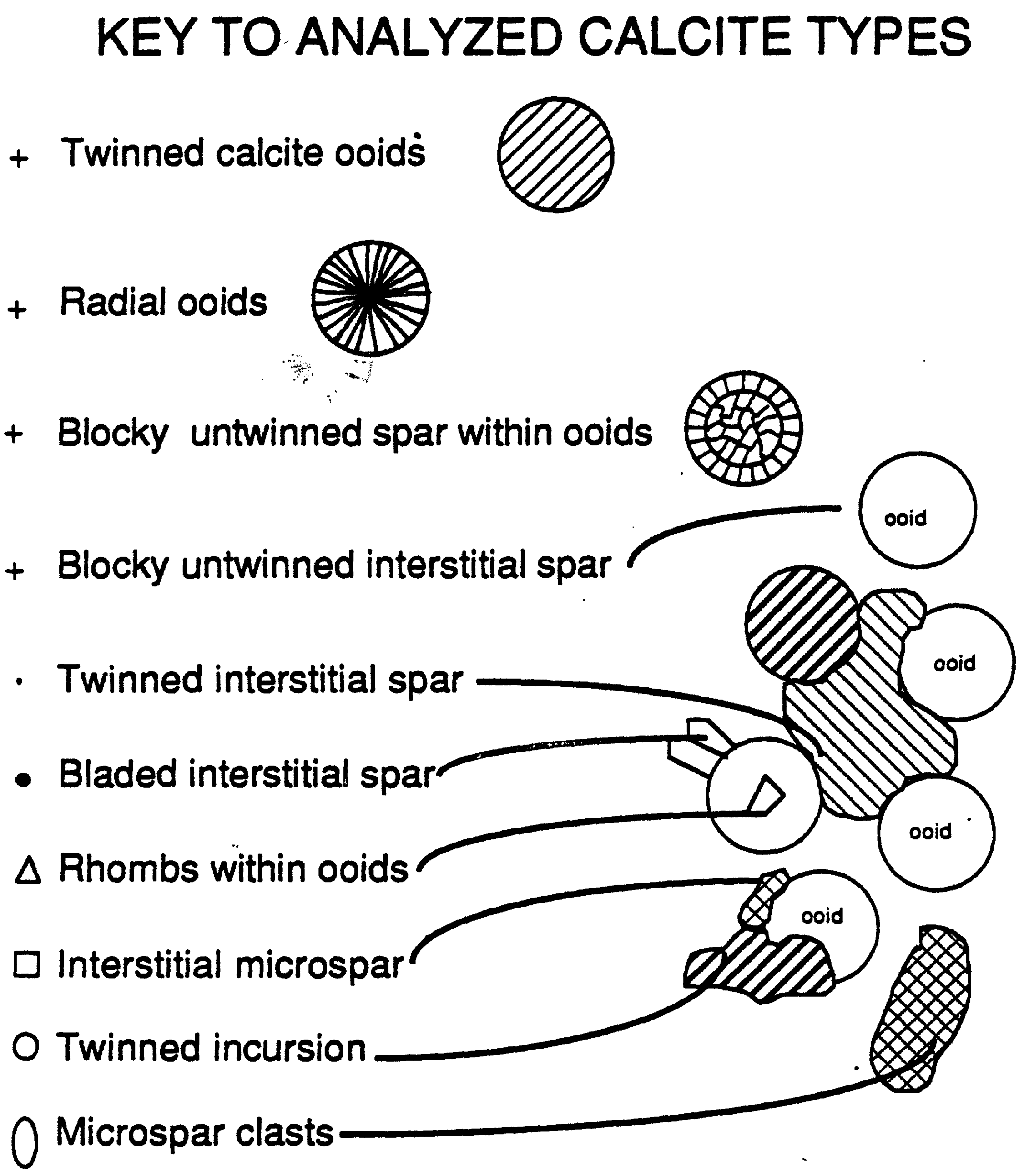


Figure 20. Schematic illustration of calcite types and key to microprobe analyses in figs. 22,23.

Other calcite within ooids consists of irregular to rhombic shaped crystals that vary in abundance from one or two isolated rhombs to pervasive replacement of selected grains. Some blocky spar is present in the centers of isolated ooids with a radial/concentric cortex and rim. Silica cement occurs locally as equant crystals of quartz in interstitial areas. Minor amounts of authigenic feldspar (<1% by volume) cross-cut ooid edges and adjacent bladed calcite cement.

Diagenetic history

Although the diagenetic relationships are complex due to lack of clear cross-cutting relationships between cements and replacement fabrics, some statements can be made about the diagenetic history of these rocks. The inferred paragenetic sequence is shown in Figure 21. Precipitation of the preserved remnants of probably originally isopachous rims of bladed crystals was one of the earliest diagenetic events. Silicification of ooids by chert was early in specific areas as evidenced by cross-cutting later phases. The neomorphic, coarse, twinned calcite spar within ooids followed chert silicification. In some examples early concentric rings of chert within ooid cortices are engulfed within single crystals of coarse neomorphic calcite spar (Fig. 18D). Blocky interstitial spar that at places coarsens toward the centers of interstitial areas followed precipitation of the bladed crystals, but could not be directly related temporally to the cherty ooids. Megaquartz cross-cuts both radial/concentric ooids and adjacent rims of bladed crystals indicating that it post-dates the bladed crystals. This type of quartz has not been observed cutting adjacent blocky calcite. Therefore, either all silicification followed precipitation of the bladed crystals, or there were two phases of early silicification, one, a cherty phase that preceded or formed synchronously with the bladed crystals, and a later, coarser phase that followed formation of the bladed crystals. Authigenic feldspar formed at the same time as the replacive megaquartz as evidenced by similar textural relationships. Blocky quartz crystals and microspar in interstitial areas are difficult to place in the diagenetic sequence because of the lack of cross-cutting relationships. The latest diagenetic phases are small irregular calcite crystals (some of which may be dolomite) and calcite that occurs in vein fillings. The irregular calcite cuts all previous fabrics and are in turn cut by the calcite veins. The veins may have developed during the Cretaceous-Tertiary tectonism that uplifted these rocks.

The minimal amount of compaction in these rocks indicates that cementation occurred fairly early in the diagenetic history of these rocks. The good preservation of ooid fabrics and cements may have resulted from low permeabilities in the mud-dominated section. Such low permeabilities may have restricted influx of diagenetic waters, leaving a record of the early cementation history preserved in these oolitic grainstones. The remnants of the bladed crystals and the blocky interstitial spar that at places coarsens toward the centers of voids is similar to the style of cementation recorded in many recent and inferred ancient freshwater phreatic zones (Longman, 1980). This realization, of course, sheds no light on the depositional environment of the Libby Formation during depositon of member C, but indicates that regardless of whether the environment was marine or non-marine, the local environment was more likely humid rather than arid. The distinct, irregular, bedding parallel zones of silicification may reflect movement of the mixing zone through these rocks early in their diagenetic history.

Geochemistry

A variety of ooid types, cement types and replacement fabrics were analyzed with an ARL Model EMX-SM electron microprobe for their trace element composition. Analyses were taken for 60 second counting periods at 15 kV acceleration potential. These calcite types are summarized in Figure 20. The raw data, expressed in weight per cent are shown in Table 3.

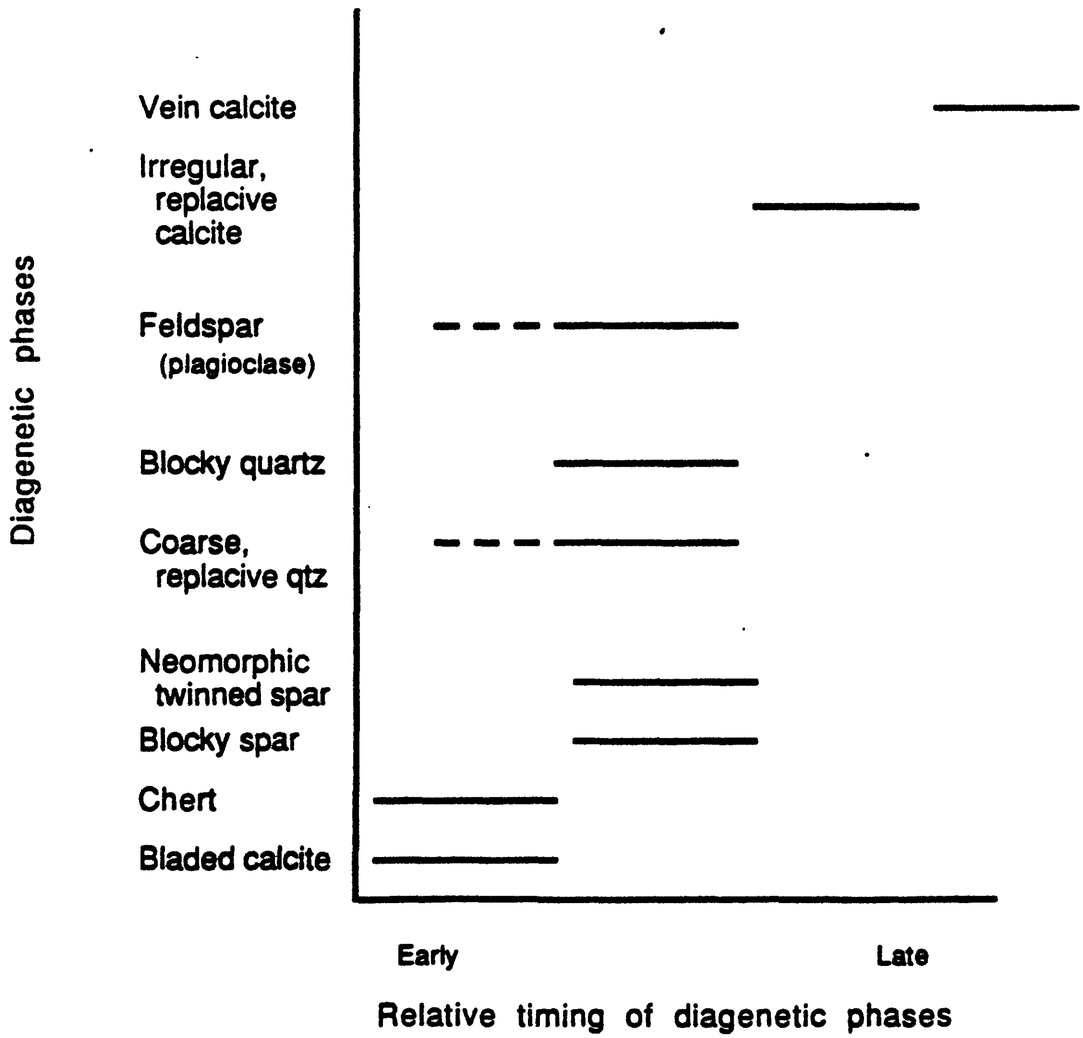


Figure 21. Generalized paragenetic sequence for cements and replacement fabrics in Lake Creek Quarry oolite (member C).

LCQ	FeO	MnO	MgO	CaO	SrO	SiO_2	Al_2O_3	SO_3	CO_2	Total
1	0.19	0.62	1.04	53.75	0.40	0.04	0.18	0.14	44.352	100.492
2	0.04	0.84	1.02	53.50	0.30	0.18	0.19	0.04	44.165	99.905
3	0.10	0.75	0.96	53.65	0.20	0.22	0.08	0.11	44.145	99.915
4	0.06	1.03	0.97	53.92	0.14	0.33	0.19	0.19	44.533	100.843
5	0.14	0.61	1.07	52.63	0.35	0.92	0.27	0.03	43.440	98.270
6	0.05	0.88	0.89	54.06	0.35	0.34	0.07	0.07	44.524	100.824
7	1.30	0.73	2.79	47.16	0.41	0.23	0.19	0.35	41.817	94.557
8	0.06	0.96	0.95	52.58	0.26	0.19	0.15	0.14	43.449	98.399
9	0.22	0.51	1.12	53.10	0.12	0.12	0.25	0.05	43.742	98.862
10	0.01	0.92	1.12	52.85	0.30	0.03	0.14	0.10	43.804	99.104
11	1.31	1.18	1.61	52.19	0.11	0.83	0.08	0.14	44.727	101.267
12	0.79	1.01	1.45	52.38	0.04	1.79	0.21	0.18	44.227	100.077
13	1.22	1.13	1.72	52.42	0.21	0.88	0.06	0.07	44.978	101.748
14	0.26	0.81	0.96	52.76	0.10	0.15	0.04	0.05	43.543	98.483
15	0.15	0.82	1.01	55.49	0.23	0.14	0.18	0.02	45.749	103.469
16	0.01	1.16	1.10	56.03	0.33	0.10	0.12	0.01	46.490	105.130
17	0.82	0.67	1.91	54.98	0.28	0.12	0.22	0.27	46.643	105.573
18	0.77	0.59	1.52	54.69	0.50	0.19	0.12	0.16	45.990	104.220
19	0.32	0.63	1.14	52.78	0.03	1.35	0.61	0.25	43.626	98.776
20	0.27	1.02	1.01	51.30	0.37	0.11	0.16	0.28	42.723	96.973
21	0.09	1.10	0.96	55.38	0.29	0.06	0.01	0.01	45.809	103.639
22	0.07	0.89	0.97	53.52	0.37	0.35	0.08	0.05	44.212	100.082
23	0.13	0.39	0.96	53.04	0.10	0.16	0.32	0.30	43.365	98.285
24	0.31	0.77	1.14	50.94	0.06	4.13	0.81	0.07	42.285	95.575
25	0.15	0.66	0.95	52.40	0.41	0.11	0.11	0.01	43.196	97.776
26	0.31	0.57	1.02	55.36	0.47	0.24	0.01	0.06	45.665	103.455
27	0.13	0.67	0.95	51.47	0.13	0.26	0.07	0.04	42.339	95.729
28	0.46	0.46	1.36	55.20	0.33	0.11	0.12	0.19	45.860	103.860
29	0.06	0.78	0.99	55.97	0.20	0.06	0.10	0.02	46.009	104.029
30	1.20	1.00	3.23	53.56	0.12	0.03	0.07	0.09	47.375	106.575
31	0.22	0.79	0.97	55.11	0.43	0.07	0.04	0.06	45.509	103.089
32	0.18	0.63	0.99	56.96	0.25	0.17	0.11	0.16	46.771	105.941
33	0.06	1.32	1.21	53.76	0.24	0.04	0.08	0.34	44.930	101.860
34	0.11	0.88	1.01	52.89	0.27	0.35	0.10	0.07	43.733	98.963
35	0.19	0.90	0.99	54.77	0.38	0.13	0.14	0.36	45.307	102.897
36	0.31	0.90	1.09	56.44	0.34	0.17	0.08	0.18	46.792	106.052
37	0.15	1.06	1.05	55.70	0.26	0.37	0.10	0.30	46.154	104.674
38	0.20	1.54	1.09	52.90	0.16	0.68	0.13	0.19	44.338	100.418
39	0.35	2.90	1.29	49.72	0.08	2.26	0.25	0.05	43.137	97.527
40	0.55	2.67	1.46	51.52	0.35	1.20	0.15	0.35	44.804	101.704
41	0.66	2.07	1.18	49.72	0.24	1.24	0.10	0.25	42.642	96.762
42	0.43	1.64	1.03	50.81	0.23	0.88	0.40	0.09	42.868	97.098
43	0.55	1.92	1.18	52.45	0.34	0.27	0.02	0.11	44.660	101.210
44	0.48	0.53	1.26	50.85	0.22	1.07	0.07	0.14	42.333	95.813

LCQ	FeO	MnO	MgO	CaO	SrO	SiO_2	Al_2O_3	SO_3	CO_2	Total
45	0.08	0.94	1.34	53.74	0.43	0.43	0.37	0.33	44.860	101.720
46	0.35	0.44	0.94	51.97	0.28	0.47	0.03	0.17	42.747	96.897
47	0.17	0.66	0.83	58.91	0.19	0.42	0.19	0.29	48.130	109.180
48	0.11	0.53	0.79	54.10	0.38	0.29	0.25	0.16	44.230	100.300
49	1.22	0.89	0.74	54.74	0.32	0.01	0.02	0.24	45.606	103.756
50	0.12	0.94	0.53	53.33	0.19	0.21	0.01	0.20	43.576	98.886
51	0.18	0.61	0.48	54.29	0.32	0.05	0.03	0.01	44.120	100.010
52	0.19	0.47	0.49	48.44	0.15	3.49	1.04	0.07	39.338	89.148
53	0.36	0.97	0.56	54.69	0.29	0.21	0.36	0.15	44.894	101.914
54	0.01	0.62	0.46	56.19	0.17	0.97	0.42	0.27	45.439	103.159
55	0.19	1.03	0.52	54.07	0.32	0.11	0.27	0.16	44.315	100.605
56	0.12	0.81	0.51	49.88	0.19	3.31	0.01	0.20	40.730	92.440
57	0.02	0.06	0.43	53.62	0.08	0.22	0.02	0.13	42.921	97.261
58	0.04	0.90	0.47	55.79	0.38	0.15	0.01	0.15	45.454	103.184
59	0.04	1.28	0.50	56.49	0.39	0.39	0.16	0.19	46.333	105.223
60	0.17	0.97	0.54	56.50	0.26	0.17	0.11	0.22	46.174	104.834
61	0.23	1.08	0.52	54.73	0.17	0.11	0.08	0.05	44.836	101.616
62	0.06	0.64	0.50	53.00	0.17	0.18	0.07	0.06	43.009	97.439
63	0.30	0.87	0.50	54.33	0.23	0.05	0.01	0.19	44.407	100.827
64	0.05	0.80	0.51	54.50	0.07	0.05	0.01	0.27	44.279	100.479
65	0.70	0.43	1.06	53.07	0.34	0.01	0.04	0.42	43.977	99.997
66	0.60	0.65	0.90	53.90	0.29	0.06	0.05	0.51	44.544	101.394
67	1.10	0.80	1.28	54.24	0.12	0.23	0.04	0.38	45.574	103.494
68	0.69	1.00	1.11	53.60	0.28	0.09	0.04	0.30	44.853	101.833
69	0.69	0.76	0.90	52.89	0.25	0.37	0.23	0.36	43.868	99.718
70	0.24	0.57	0.64	52.16	0.01	0.09	0.04	0.27	42.487	96.377
71	0.59	0.71	0.66	54.64	0.21	0.07	0.04	0.18	44.873	101.863
72	0.71	0.85	1.12	54.86	0.27	0.08	0.01	0.24	45.752	103.802
73	0.02	0.42	0.44	56.34	0.35	0.03	0.01	0.23	45.467	103.267
74	0.60	0.49	1.07	53.25	0.29	0.23	0.15	0.38	44.094	100.174
75	0.61	0.74	0.82	43.27	0.11	7.42	4.83	0.03	36.057	81.637
76	0.06	0.77	0.50	56.04	0.32	0.37	0.23	0.27	45.573	103.533
77	0.19	0.64	0.53	55.84	0.19	0.13	0.01	0.45	45.374	103.214
78	0.31	0.49	0.62	47.84	0.30	2.20	1.10	0.36	39.157	89.077
79	0.16	0.92	0.66	54.42	0.39	0.22	0.17	0.37	44.672	101.592
80	0.66	2.39	0.66	52.85	0.24	0.31	0.09	0.19	44.791	101.781
81	0.32	0.69	0.63	54.35	0.42	0.27	0.02	0.39	44.519	101.319
82	0.45	0.69	0.64	53.33	0.34	0.80	0.67	0.08	43.771	99.301
83	0.32	0.60	0.61	53.32	0.35	0.27	0.09	0.19	43.586	98.976
84	0.47	0.78	0.65	56.62	0.38	0.12	0.04	0.27	46.478	105.648
85	0.76	0.51	1.00	54.45	0.42	0.03	0.07	0.06	45.134	102.334
86	0.35	0.87	0.61	51.92	0.14	0.27	0.15	0.03	42.616	96.536
87	0.36	0.49	0.79	54.15	0.22	0.41	0.09	0.34	44.323	100.673
88	0.21	0.91	0.70	52.96	0.46	0.01	0.05	0.11	43.614	98.964

LCQ	FeO	MnO	MgO	CaO	SrO	SiO_2	Al_2O_3	SO_3	CO_2	Total
89	0.14	0.60	0.53	54.35	0.12	0.05	0.01	0.37	44.106	100.216
90	0.07	1.07	0.58	54.90	0.26	0.01	0.02	0.19	44.968	102.038
91	0.07	0.87	0.52	54.91	0.59	0.23	0.06	0.25	44.898	102.108
92	0.05	0.95	0.45	53.94	0.32	0.20	0.10	0.01	43.991	99.711
93	0.13	0.99	0.52	52.83	0.25	0.04	0.03	0.01	43.239	97.969
94	0.01	0.65	0.46	56.71	0.28	0.18	0.16	0.06	45.920	104.090
95	0.33	1.26	0.52	56.35	0.51	0.18	0.01	0.05	46.457	105.477
96	0.01	0.81	0.44	47.83	0.20	0.11	0.01	0.24	38.972	88.502
97	0.05	0.82	0.44	43.02	0.63	15.90	0.21	0.24	35.385	80.585
98	0.15	0.88	0.48	55.72	0.26	0.03	0.07	0.19	45.411	103.091
99	0.21	0.88	0.50	55.96	0.33	0.07	0.03	0.12	45.689	103.689
100	0.06	0.76	0.52	56.86	0.41	0.23	0.02	0.06	46.273	104.943
101	0.19	0.62	0.51	53.89	0.40	0.21	0.31	0.01	43.884	99.504
102	0.10	0.67	0.50	55.26	0.14	0.40	0.19	0.17	44.829	101.669
103	0.10	0.86	0.47	56.14	0.39	0.01	0.06	0.17	45.742	103.872
104	0.21	0.46	0.50	56.08	0.30	0.02	0.01	0.13	45.452	103.132

Table 3. Microprobe analyses of ooid types, cements, and replacement fabrics from Lake Creek Quarry oolite (Member C). See Appendix 2 for location. Analyses are in weight per cent oxides. A key to the types of calcites analyzed follows:

LCQ#	Calcite type
1-49	Coarse twinned spar ooids
50-66	Radial ooids
67-72	Blocky untwinned spar within ooids
73-81	Blocky untwinned interstitial spar
82-84	Twinned interstitial spar
85-93	Bladed interstitial spar
94-96	Rhomb within ooid
97-100	Rhombic incursion
101	Interstitial microspar
102	Microspar clast

Trace element data

Iron, magnesium, and manganese substitute in varying amounts for calcium in the different calcite types. (Figs 22, 23). With the exception of only several analyses, the radial/concentric ooids and the twinned calcite ooids have very similar trace element compositions (Fig. 22A,B, 23A,B). Most of the cement and replacement types are characterized by unique compositions (Fig. 22C,D, 23C,D). Bladed interstitial spar has a composition that is fairly similar to that of the dominant ooid types, except for a slight enrichment in iron. Blocky, untwinned interstitial spar has a wider range of values that any of the three calcite types just discussed. Some of the analyses of the blocky spar are very similar to the principle ooid types and bladed cements, but others have slightly richer iron content than any of the other types mentioned thus far. Except for some overlap with twinned ooids, blocky untwinned spar within ooids has a unique, Mn-rich composition (Fig. 23D). Twinned spar between ooids has a composition similar to, but slightly more Mg-rich, than the early bladed calcite crystals (Fig. 22D). Microspar has a highly variable composition; and consequently the limited number of points analyzed prohibits interpretation of any compositional trends. Only one rhombic crystal within an ooid was analyzed. Its composition is similar to the bladed calcite except for a slight depletion with respect to Ca.

The available data permit only limited interpretation of the geochemical variation among calcite types. One crude temporal trend, that of a slight increase in iron content with progressively younger cement age is observed in comparison of both radial/concentric and twinned calcite ooids, bladed calcite, and blocky, untwinned, void-filling spar. The Mn-rich nature of the blocky spar within ooids may be attributable to conversion of aragonite to calcite as has been documented in Pleistocene corals from Barbados (Pingitoire, 1978). The general overlap in composition of twinned calcite ooids with radial/concentric ooids suggests that in a broad sense, the pore waters from which the neomorphic twinned spar crystallized was similar to the water from which the radial ooids formed.

Discussion

Until 1975, most ancient ooids were inferred to have originated as aragonite by analogy to ooids from modern marine settings (Bathurst, 1975). Detailed textural analyses of ancient ooids since then (Sandberg, 1975; Wilkinson and Landing, 1978; Rich, 1982; Tucker, 1984) combined with documentation of radial calcite ooids in modern settings (Land et al., 1979; Popp and Wilkinson, 1983) and in reworked modern settings (Marshall and Davies, 1975; Milliman and Barreto, 1975) indicate that not all modern ooids are aragonitic and that many ancient radial ooids may have been originally calcite.

The radial/concentric ooids in the Libby Formation petrographically resemble the ooids that were examined in more detail under the scanning electron microscope by Sandberg (1975) and in particular radial/concentric ooids from the Siyeh (Helena) and Snowslip Formations of the Belt Supergroup (Tucker, 1984). Based on this textural similarity to other calcitic radial/concentric ooids, and by the close association with (presumably) originally aragonitic ooids that have been replaced with coarse calcite, the radial/concentric ooids are interpreted as primary calcitic ooids.

If the radial/concentric ooids are preserved primary calcitic ooids, their association with originally aragonitic ooids and their trace element composition may have important implications regarding the waters from which they formed. The mixed calcitic/aragonitic ooids may reflect a unique depositional setting. The ooid composition would, then, reflect the composition of the water from which they formed. The composition of the Libby ooids can be compared with the that of carbonate in younger settings where there is a better understanding of the derivation of those carbonates. Questions that can be addressed are: Was Proterozoic "sea water" similar or different to that of today? and Were the waters from which the lower Libby Formation was

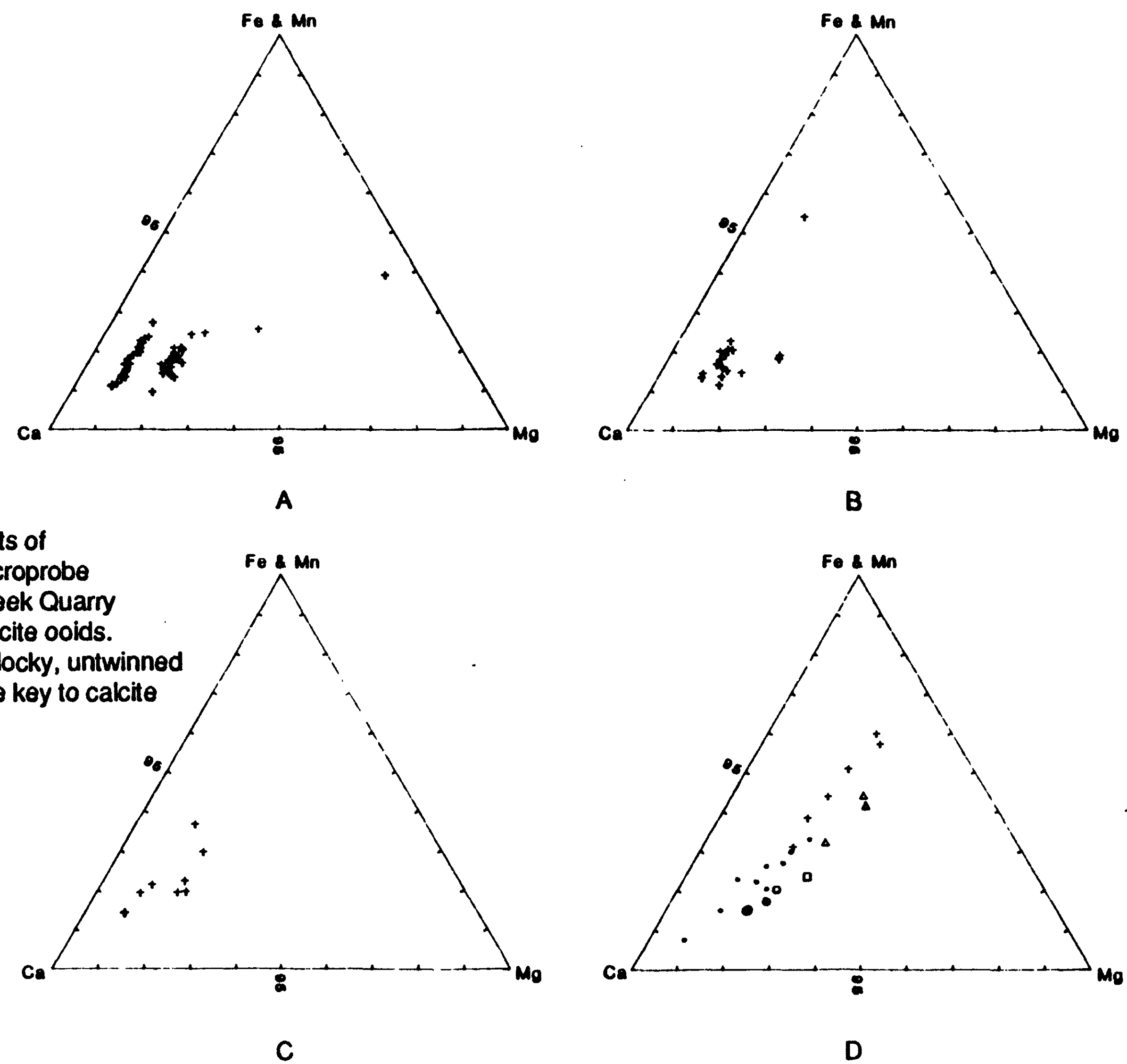


Figure 22. Ternary plots of Ca, Fe + Mn, Mg of microprobe analyses from Lake Creek Quarry oolite. A. Twinned calcite ooids. B. Radial ooids. C. Blocky, untwinned interstitial spar. D. See key to calcite types in Fig. 20.

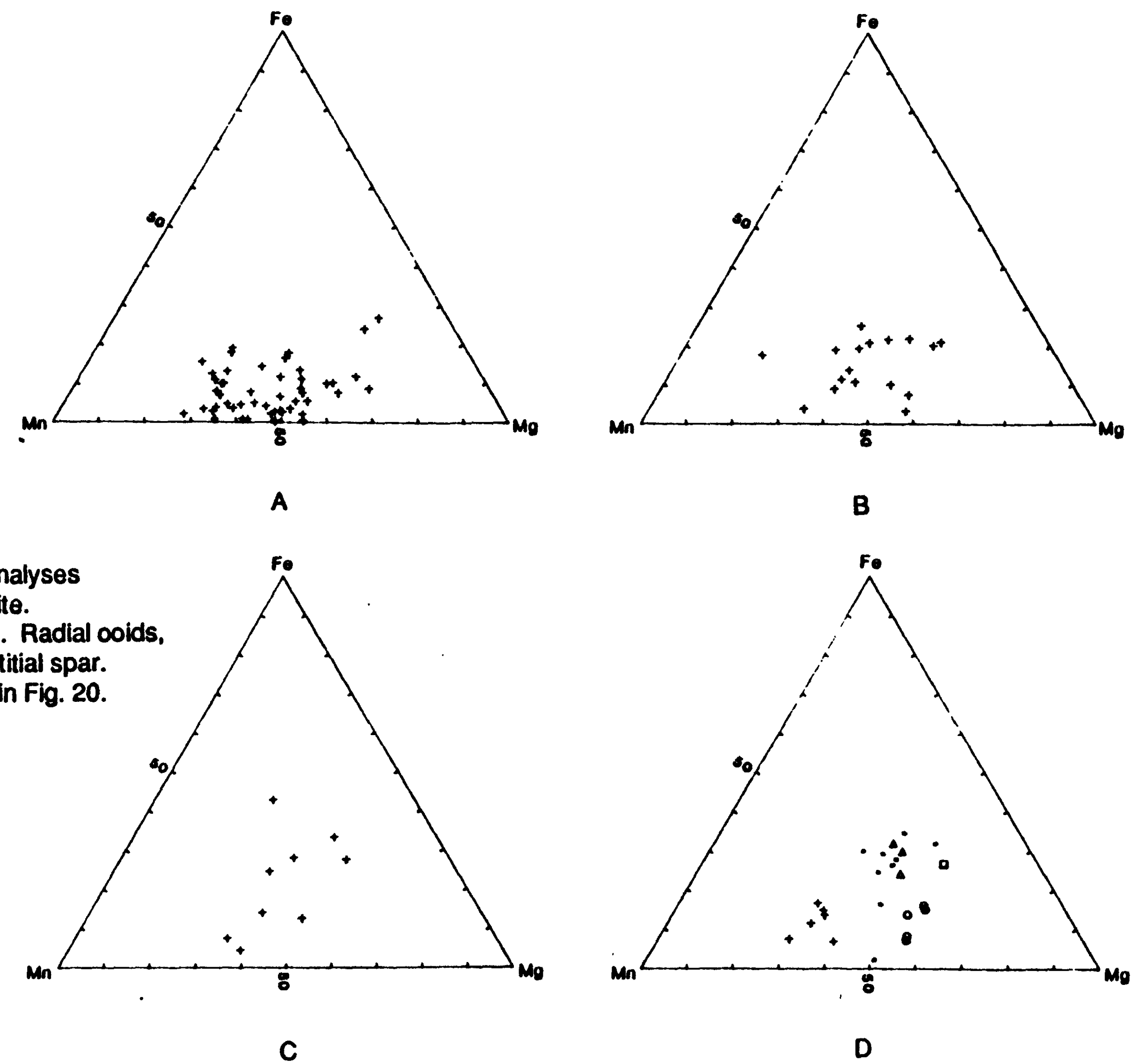


Figure 23. Ternary plots of
Mn, Fe, Mg of microprobe analyses
from Lake Creek Quarry oolite.
A. Twinned calcite ooids. B. Radial ooids,
C. Blocky, untwinned interstitial spar.
D. See key to calcite types in Fig. 20.

deposited marine or non-marine?

Tucker (1984) pointed out that the only documented occurrence of mixed calcitic-aragonitic ooids is the assemblage from the hyper-hypo-saline water of Baffin Bay, Texas (Land et al., 1979). He used this information to interpret similar ooids from the Belt Supergroup as having been deposited in a salinity- stressed environment similar to that of Baffin Bay. The petrographically similar ooids from the Libby Formation lend themselves to a similar interpretation. The only other known modern in situ radial calcite ooids are from Pyramid Lake, Nevada (Popp and Wilkinson, 1983). This strengthens the interpretation that there was a non-marine influence on the formation of the radial/concentric calcite ooids in the Belt Supergroup. However two deep-water examples of reworked radial calcite ooids from off the Great Barrier Reef (Marshall and Davies, 1975) and off the Amazon shelf (Milliman and Barreto, 1975) need to be carefully evaluated before strong support can be given to Tucker's (1984) interpretation that radial calcite ooids form in non-marine or salinity-stressed settings.

Comparison of the trace element composition of the Libby ooids and cements with Miocene-Pleistocene carbonate cements from the San Joaquin valley of California shows some interesting similarities. Nearly all of the analyses from the Lake Creek quarry have similar compositions to carbonate cements from late marine diagenetic environments documented by Boles and Ramsayer (1984). One exception is the blocky untwinned spar from within ooids that bears a composition more like the non-marine cements analyzed by Boles and Ramsayer (1984).

These geochemical comparisons conflict with the origin of the cement fabrics, which is interpreted as meteoric phreatic. The discord between the textural and geochemical interpretations of the calcites discussed here could be a function of different composition of Proterozoic (marine or non-marine) water, or a breakdown in the textural relationships summarized by Longman (1980) that appear to hold up from the Recent back through the Paleozoic.

The results of these preliminary geochemical analyses demonstrate that even Middle Proterozoic rocks have the potential for geochemical tracing of the evolution of diagenetic waters. The recent successful application of a similar study from Upper Proterozoic rocks of southern Australia (Singh, 1987) suggests that geochemical analyses combined with detailed textural analyses of carbonate rocks may allow backtracking of diagenetic history of Proterozoic rocks and approximation of the composition of Proterozoic sea water. Tucker (1982) proposed that dolomite was the principal carbonate mineral precipitated from both sea water and diagenetic water in the Precambrian based on study of the Middle Proterozoic Beck Spring Dolomite of Death Valley, California. The good preservation of calcite fabrics and textures in the Libby Formation differs from the dolomitic composition of many Beltian and other Middle Proterozoic rocks. Examples such as these oolites from the Libby Formation, calcite oolites from older Belt rocks (Tucker, 1984), and calcite oolites from the upper Proterozoic of Australia (Singh, 1987), suggest that dolomitization probably results more from diagenesis than direct precipitation from sea water as has been accepted since the early 1970's (Zenger et al., 1980).

SHRINKAGE CRACKS

The Libby Formation contains three distinct types of shrinkage cracks:

I) Cracks that are polygonal and completely connected in plan view. These cracks are several mm wide and are v-shaped in cross section. These cracks have only been observed in association with purple argillites in member C at Fishtrap Creek.

II) Cracks that display a bird's foot or triple-junction morphology in plan view, although some are in close proximity to slit- like bedding plane cracks. In cross section, these cracks are distinctly v-shaped and occur in green siltite and argillite and dark gray argillite. These cracks occur in members B an C at Flagstaff Mountain, Fishtrap Creek, and Clark Fork. This crack type is extremely common in the carbonate-bearing member C and is moderately common in member B.

III) Cracks that occur as narrow, slit-shaped features 1-2 mm wide in plan view. In cross section, these cracks are v- shaped, but their walls are often nearly parallel. The morphology of these cracks is often highly modified by compaction. This shrinkage crack-type occurs in many of the units, but is most common in the silty partings between the sandy beds in the hummocky cross-stratified unit. These cracks occur in the hummocky facies at Flagstaff Mountain, Fishtrap Creek, Superior, and near Bonner.

Interpretation

Cracks of type I are classic desiccation cracks. Type II cracks are interpreted as incipient desiccation cracks because of their association with the shallow water members B and C. Crack type III is interpreted as having formed by syneresis. This interpretation is based on 1) their resemblance to laboratory generated syneresis cracks (Burst, 1965), 2) their poor development in cross section, which suggests that shrinkage merely broke the surface of the mud and all subsequent development is from compaction of sand from overlying beds, and 3) their preferential association with the hummocky cross-stratified facies, which at Libby is interpreted as a deeper environment than the facies in which crack types I and II occur.

DEPOSITIONAL ENVIRONMENTS

Environments of deposition within the study area have been interpreted by integration of information on sedimentary structures and their distribution, paleocurrent data, grain size distribution, and lithofacies maps. Extrapolation of these interpretations to other parts of the Belt basin have been made in areas where data from other workers are available. In this section, brief environmental interpretations for each member on Flagstaff Mountain will be extended as far as possible throughout the Belt-Purcell basin. Then an overview of depositional environments will be presented for the lower Libby Formation (members A-D) and upper Libby Formation (members D'-G).

Winston (1986d) has developed a sedimentological classification for Belt rocks. These terms are commonly used in this and other sections. Brief descriptions of the Winstonian terms are provided here and in the accompanying figure (Fig. 24):

<u>Even couplet</u>. "...very fine sand and silt layers that grade up to clay, forming parallel-laminated to slightly wavy, continuous couplets 0.3 to 3 cm thick."

<u>Lenticular couplet</u>. "...discrete ripple lenses and continuous rippled laminae or fine sand and silt, rather sharply overlain by mud laminae, forming couplets 0.3 to 3 cm thick."

<u>Pinch and swell couplets</u>. "...fining-upward fine sand- and silt-to-clay couplets in which the sandy and silty layers thicken and thin across outcrops, producing uneven, pinch-and-swell stratification."

<u>Microlamina</u>. "...millimeter-scale laminae of quartz or dolomitic silt sharply overlain by equally thin clay laminae."

<u>Source Area and Paleocurrents</u>

The major source area for the Libby Formation and equivalent rocks was in the southern part of the Belt basin based on paleocurrent data largely taken from the Bonner Quartzite and Garnet Range Formation (Winston et al., 1986; McGroder, 1984a) and on grain size distribution in the Libby, McNamara, and Garnet Range Formations. Several zones within the Libby Formation

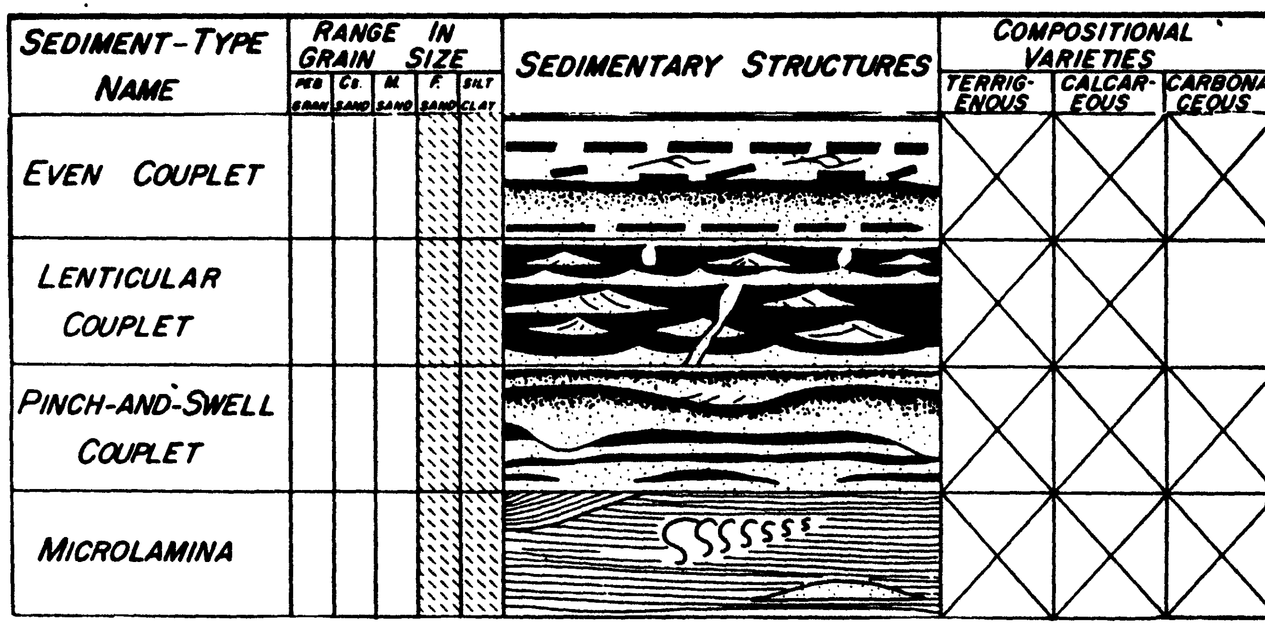


Figure 24. Illustration of Beltian sediment types of Winston (1986d) found in Libby Formation. See text for explanation.

(especially members B,C, and F) coarsen from the type locality, near Libby, to the south and, for member F, to the east. Paleocurrent measurements within the Libby Formation (Fig. 25) are on small-scale structures such as ripples and centimeter- scale channels. These measurements consistently suggest oscillatory transport from north-northwest to south-southeast.

Lithofacies maps

A series of lithofacies maps (Figs. 26A-F) has been constructed for each member at the type locality of the Libby Formation. These maps are preliminary and can be expanded after further field studies in the areas near the geographic limits of Belt exposures. Discussions for each member first outline the distribution of lithofacies followed by an environmental interpretation. Often, on a local scale, a lithofacies will represent the same rocks as a lithostratigraphic unit, but on a larger scale, these units change laterally, and must be described as lithofacies.

Member A. The base of the Libby Formation marks a sudden change in rock type from the quartzite of the Bonner Quartzite to much finer grained deposits. The dark gray, microlaminated argillite of the study area thickens drastically to the northwest where Glover (1978) reported 275 m of dark gray argillite in the Summit Creek area. This part of the section becomes progressively more carbonate-rich to the north. Southward and eastward from Libby, the rocks are generally coarser-grained and the McNamara Formation is predominantly green with some red zones. To the west, member A retains its character as far as Clark Fork, and isolated exposures of dark gray argillite in the Magnesite belt (Campbell and Loofbourow, 1963) may extend this facies considerably further westward.

Presently available data suggest the interpretation of progradation of terrigenous clastics from the south and possibly east across mudflats that were periodically exposed in southern and eastern areas. To the north and west of Libby, the supply of terrigenous material diminished allowing possible development of a carbonate bank to the north (Bugaboo Creek region) and dark gray argillite in the northwest (Summit Creek). The northern region was marked by waters that were clearer and/or shallower than the lithofacies correlative sections in the United States. The repetitive couplets of siltite and argillite indicate frequent agitation of the bottom that would produce turbid waters. Although carbonate deposition prevailed in the north, the sudden thickening of dark gray argillite at Summit Creek suggests the possibility of a local topographic depression in that area. Alternatively, this sudden thickening may reflect proximity to a shelf edge. The widespread distribution of this finer grained facies indicates either a relative sea level rise or a sudden cut-off of source material from the southeast.

Member B. This lithofacies is marked nearly basinwide by deposits that are coarser than member A. In the study area, even couplets of green siltite overlie the microlaminated dark gray argillite of member A. The even couplets are sparsely mudcracked at Libby and Fishtrap Creek. At Fishtrap Creek, sandy zones in parts of the section are suggestive of progradation from the south. The general character of this rock type is similar in the Whitefish Range and near the town of Bonner. Some zones to the south and east of the study area are reddish purple. To the northwest (Summit Creek) the dark gray argillitic facies persists and carbonate rocks are somewhat diluted by clastics to the north (Bugaboo Creek).

A general northwesterly and westerly progradation of detrital silt and fine-grained sand is suggested by lithofacies relationships. Rapid deposition is indicated by abundant fluid escape structures. Occasional mudcracks indicate periodic subaerial exposure as far north and west as Libby and Clark Fork. The "carbonate bank" deposits in the northern Belt-Purcell basin still persisted, but were more restricted to the region of Bugaboo Creek.

Member C. This member displays the widest facies variation of any member in the Libby Formation and equivalent rocks. This unit marks the maximum development of carbonate environments in the study area. Thin, laterally discontinuous stromatolites, oolites and mixed stromatolites and oolites interbedded with siltite and argillite may represent scattered, local

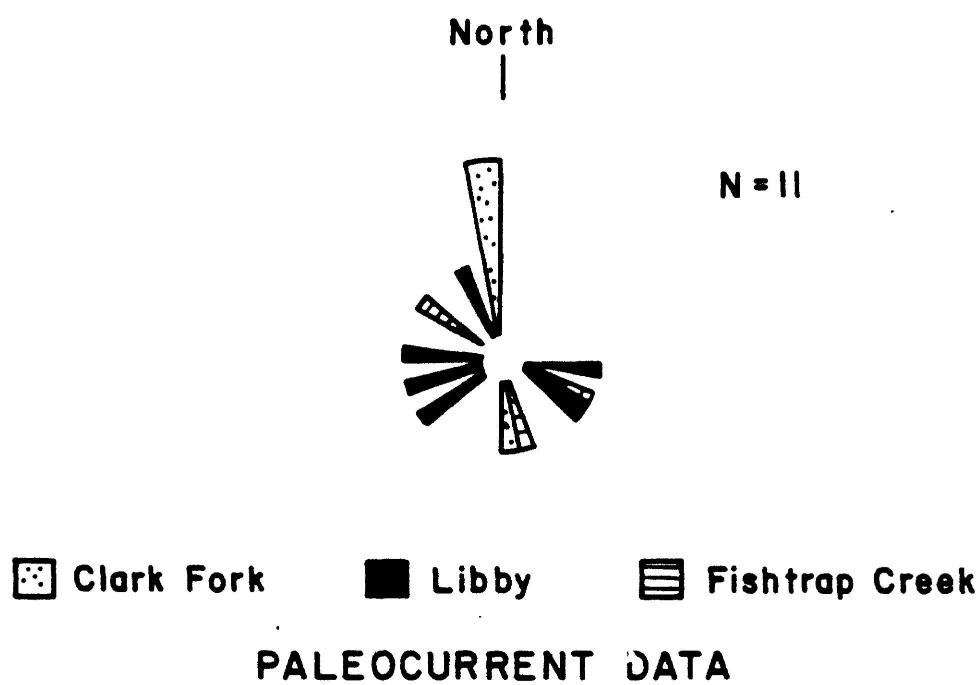
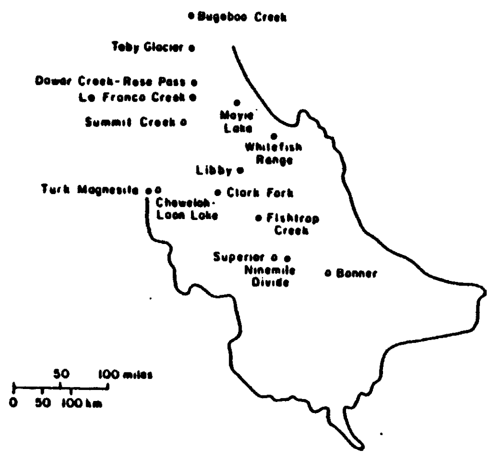
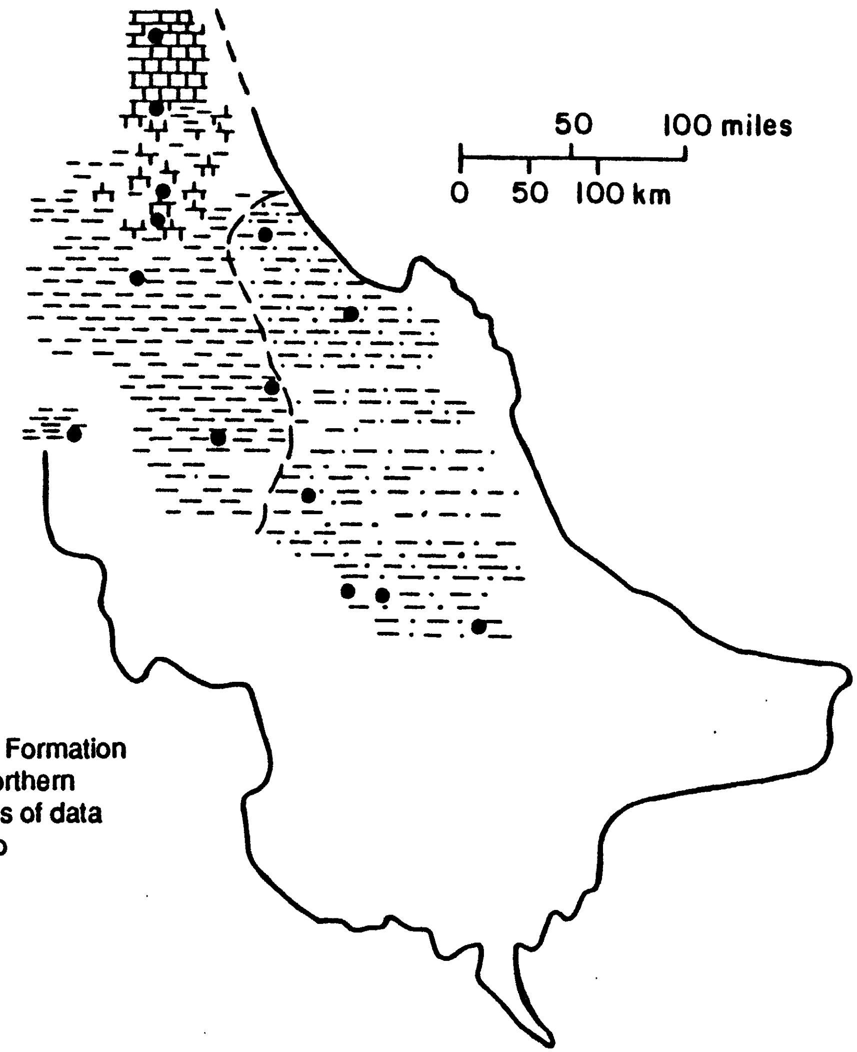


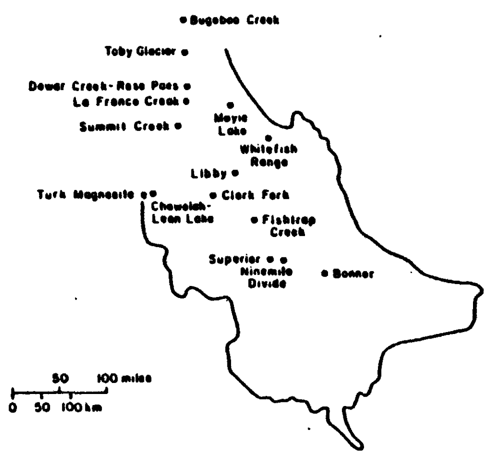
Figure 25. Paleocurrent analyses of small-scale sedimentary structures (e.g. starved ripple foresets, small-scale channels, perpendicular axis to ripple crests) from Libby area, Clark Fork, and Fishtrap Creek.

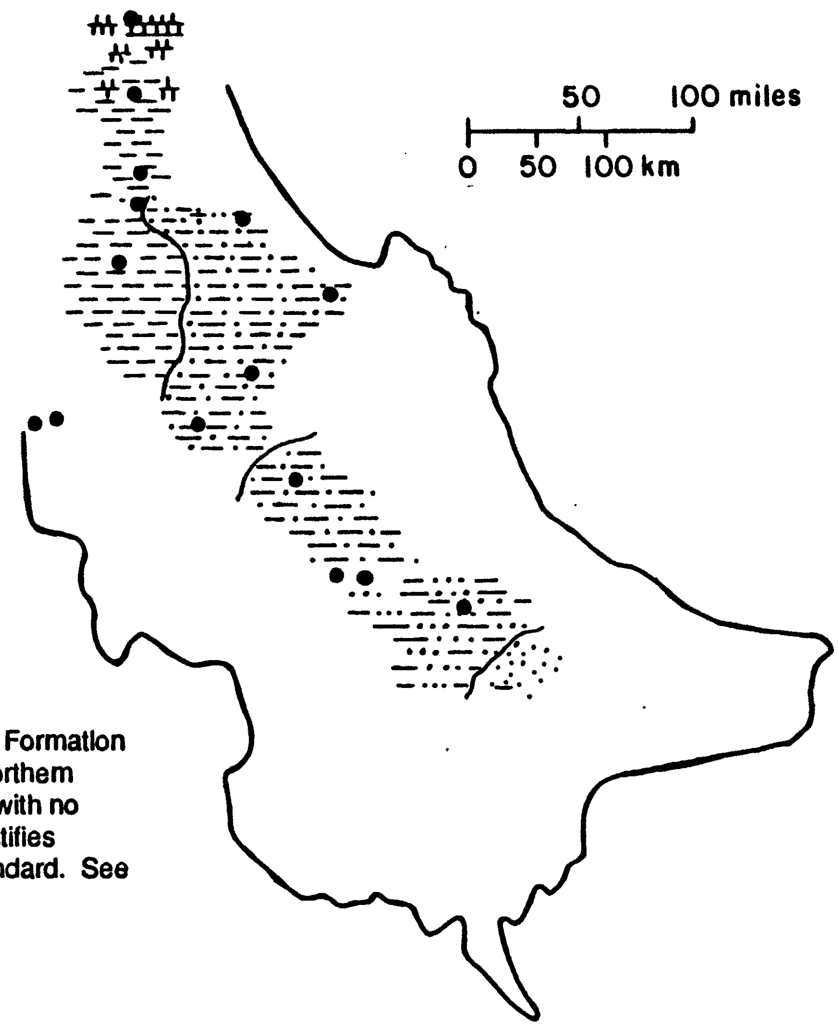




Member A time

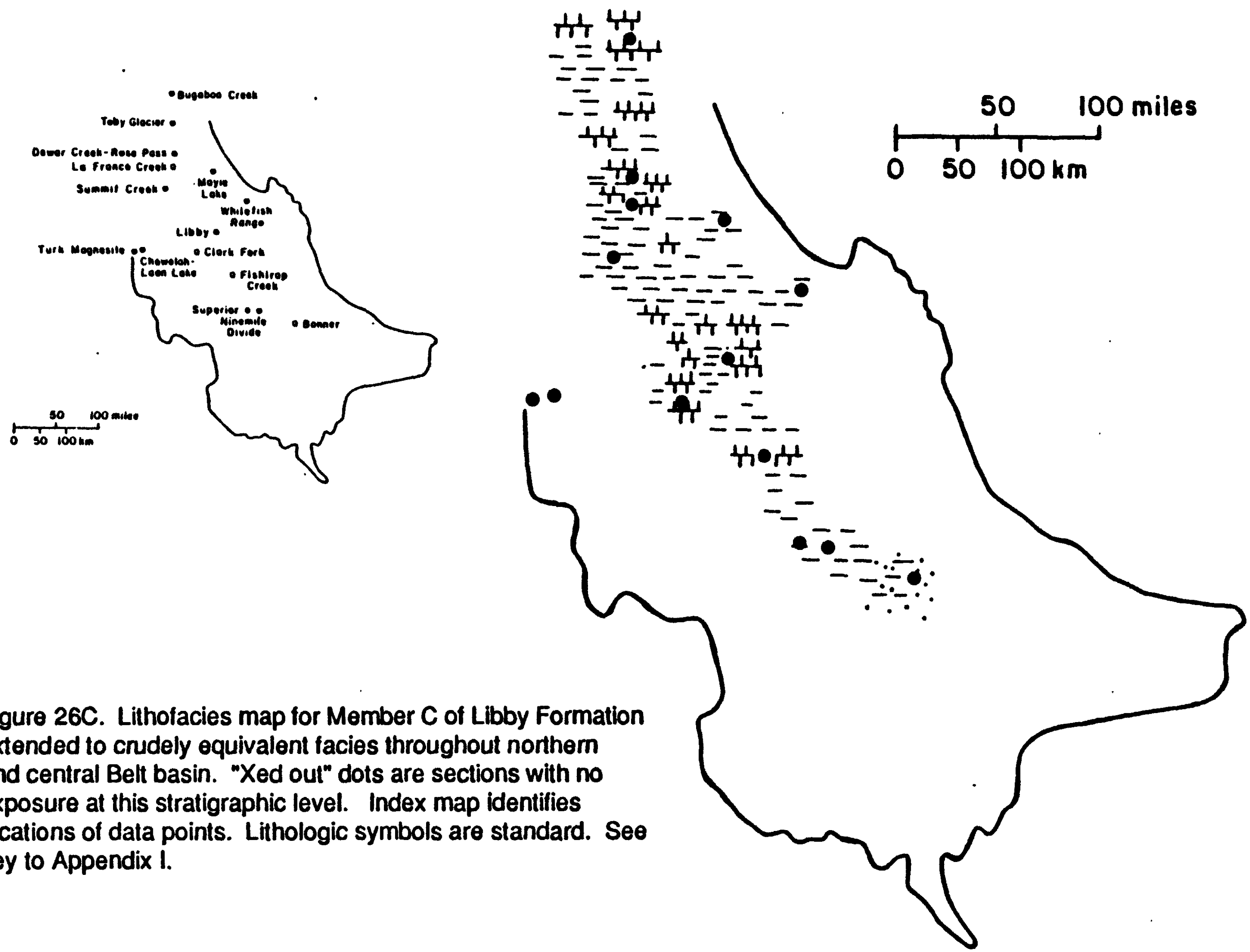
Figure 26A. Lithofacies map for Member A of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.





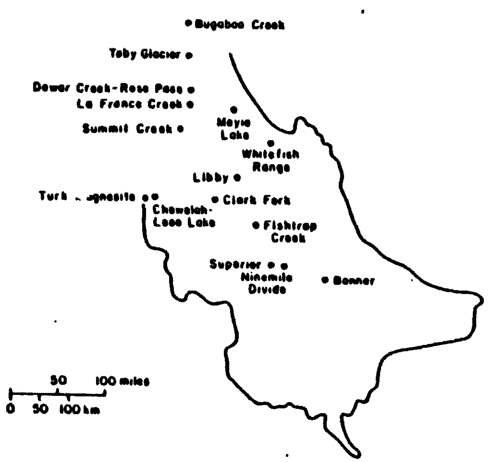
Member B time

Figure 26B. Lithofacies map for Member B of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. "Xed out" dots are sections with no exposure at this stratigraphic level. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.



Member C time

Figure 26C. Lithofacies map for Member C of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. "Xed out" dots are sections with no exposure at this stratigraphic level. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.



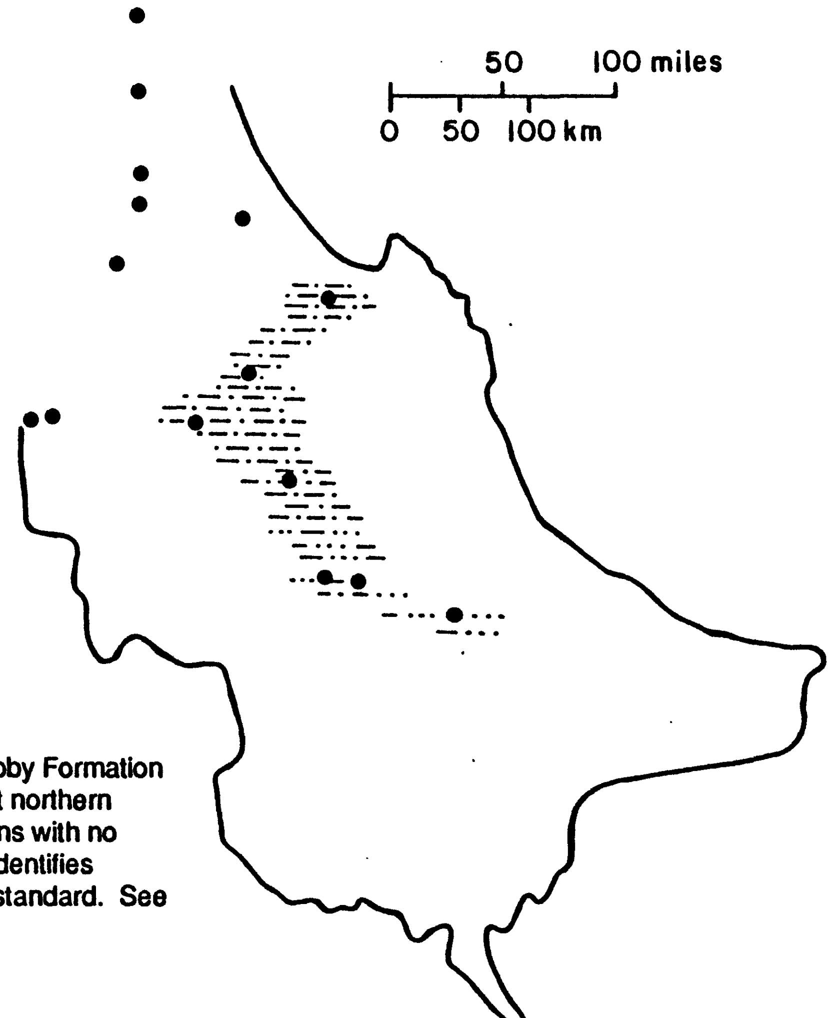
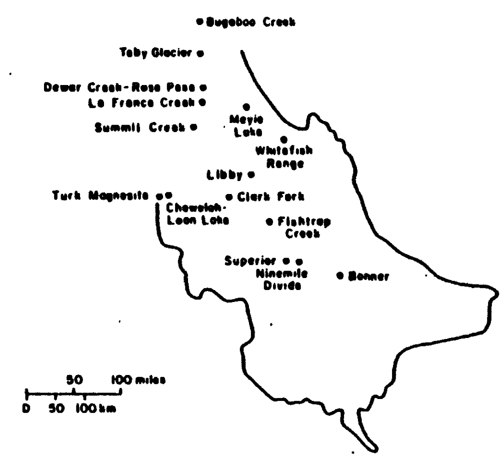


Figure 26D. Lithofacies map for Member D of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. "Xed out" dots are sections with no exposure at this stratigraphic level. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.

Member D time



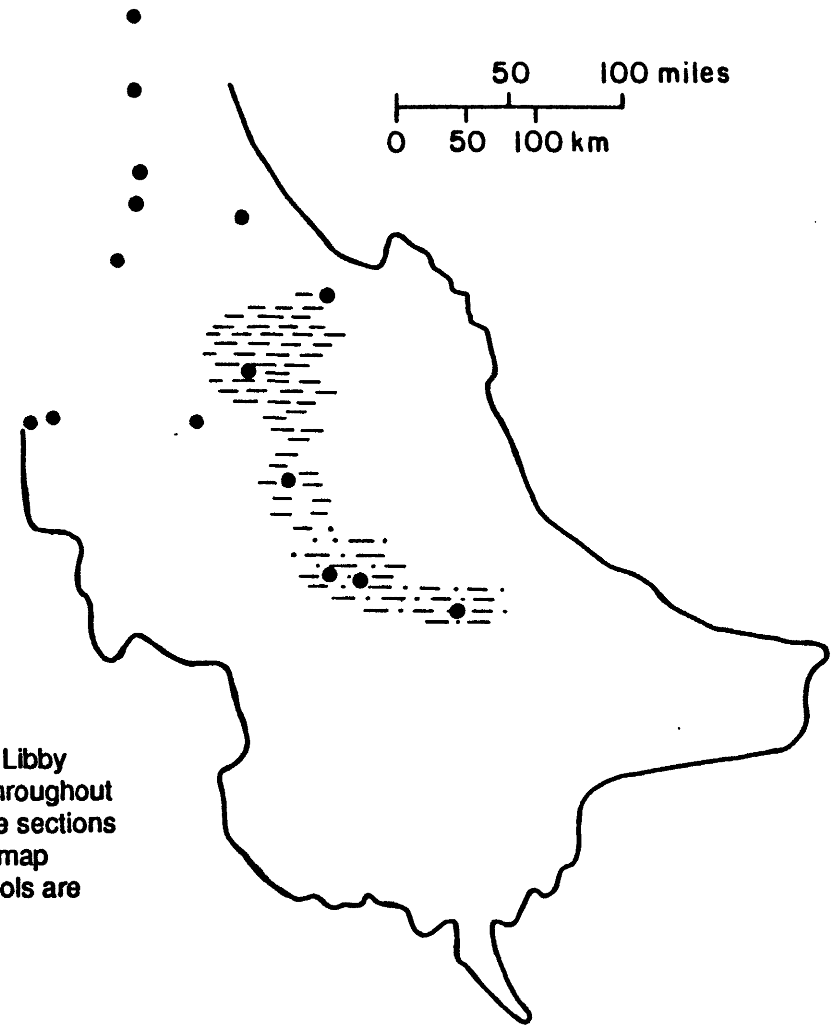
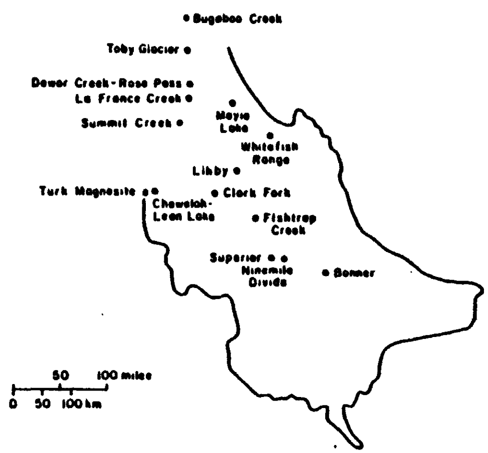
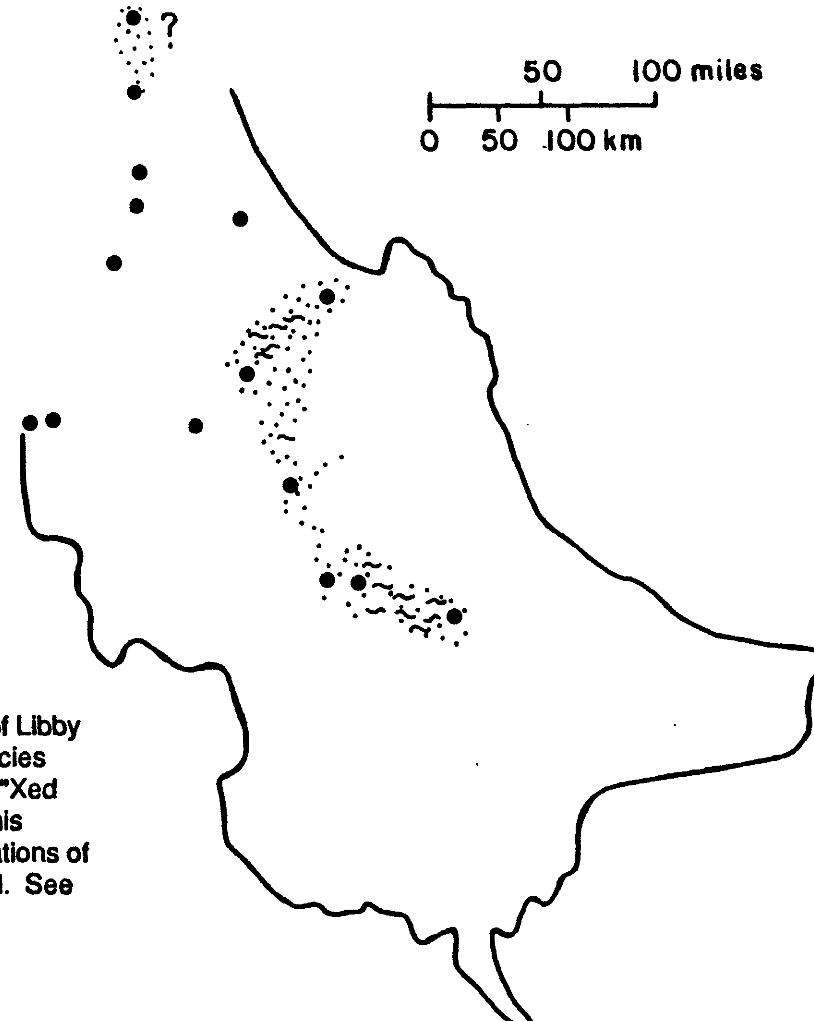


Figure 26E. Lithofacies map for Member D' + E of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. "Xed out" dots are sections with no exposure at this stratigraphic level. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.

Member D' + E time





Member F time

Figure 26F. Lithofacies map for Member F of Libby Formation extended to crudely equivalent facies throughout northern and central Belt basin. "Xed out" dots are sections with no exposure at this stratigraphic level. Index map identifies locations of data points. Lithologic symbols are standard. See key to Appendix I.

development of carbonate shoals. Detrital influx from the south largely quenched the development of these environments at Fishtrap Creek. Overall, the facies of this unit were deposited in shallower water as evidenced by well-developed mud cracks at Fishtrap Creek and the much greater abundance than in other areas of shrinkage cracks interpreted as desiccation cracks at Libby and Clark Fork. Supportive evidence for subaerially exposed environments comes from the purple/red argillitic zones at Fishtrap Creek that are in approximately the same stratigraphic position in the Whitefish Range. Lithofacies equivalent rocks in British Columbia are also characterized by carbonates at this stratigraphic level.

Member D. Basinwide study of Libby equivalents becomes difficult at this horizon because erosion of the top of the formation reaches this level at many places, particularly in northern sections. Additionally, detailed study of the upper McNamara Formation has not been done, which further hampers construction of lithofacies maps because precise correlation needed to relate lithofacies over wide areas cannot be made to southern sections.

The green siltite of this member marks a renewed influx of terrigenous material. This influx is associated with a predominance of wavy and lenticular lamination, indicating higher energy environments than most of those recorded lower in the Libby Formation. This lithofacies can be traced to Clark Fork and Fishtrap Creek without any significant changes other than thickness. Shrinkage cracks are rare in this unit suggesting that periodic subaerial exposure was greatly diminished. This environment probably represents a subaqueous mudflat more-or-less constantly affected by gentle wave activity.

Member E. The transition into this member (the transition has been described in the stratigraphy as member D', but a separate lithofacies map has not been constructed) consists of interbedded siltite and argillite interpreted as several pulses of terrigenous detrital influx punctuated by deposition of dark gray argillite. A coarser lithofacies of this transition zone can be recognized as far away as the town of Bonner, recording the beginning of an incursion of a quiet, reducing environment across much of the Belt basin. This incursion is recorded in member E proper by a wedge of dark gray argillite that thickens from Fishtrap Creek to the northwest.

This depositonal change is significant in terms of the Belt for two reasons. First, it marks the lithostratigraphic contact between the McNamara and Garnet Range Formations which can now be informally extended to separate the upper from the lower Libby Formation. Second, it marks the beginning of a change to a new type of environment that is largely absent from other Belt formations. This new environment is recorded by the hummocky cross-stratified quartzite in member F.

Member F. The thick and widespread hummocky cross- stratified quartzite is a facies unique to Belt deposition. Thin, localized occurrences of this facies occur in several other Belt formations (Mt. Shields Formation, C.A Wallace pers comm., Newland Formation, J. Schieber pers. comm.; Prichard Formation, E.R. Cressman, pers. comm.), but none of these approach the thickness and lateral extent recorded in the Libby and Garnet Range Formations (Figs. 10-12). Deposition at the base of this facies was accompanied by some sort of high energy event recorded over a large area that resulted from either direct tectonic activity (e.g. earthquakes) that cause slumping, disruption of beds, and recumbently folded cross-strata in member F and slumping in the Garnet Range Formation (McGroder pers. comm. 1986), or subsidence that produced a basin that was more open to higher energy waves that upon impinging on a liquified bottom, produced the same features described above.

Member G. This member has only been recognized with confidence at Libby and in the Whitefish Range. It is much thicker in the Whitefish Range than at Libby. (Fig. 10). Because this lithofacies is truncated by erosion and because of its limited geographic distribution, it does not warrant construction of a lithofacies map. The simple lithostratigraphic correlation (Fig. 10) adequately illustrates its distribution. The predominance of structureless muds with thin, sandy interbeds is suggestive of a flysch-like or pre-flysch-like rock.

Environmental Interpretations

Lower Libby Formation

Members A-E of the Libby Formation contain sediment types that can be genetically related to one another laterally and vertically (even though member E is lithostratigraphically part of the upper Libby Formation). Members F and G characterize different environments of deposition, and are treated separately. The sediment types in the lower six members fit the classification developed by Winston (1986d) for Beltian sediment types. Winston interpreted the environments of deposition in some Belt formations by comparison of these sediment types and their lateral and vertical relations with potential modern and ancient analogs. Winston (1986d) did not incorporate the upper Libby/Garnet Range Formations into his hypotheses.

The lower Libby Formation contains distinctive sediment types that can be compared to similar sediment types found in various modern and recent settings. The main features of these sedimentary rocks include:

1) Predominance of silt and clay sized material.

2) Predominance of four principal sediment types, namely (terminology after Winston, 1986d - see Fig. 24): 1) microlaminated sediment type, 2) even couplet, 3) pinch and swell couplet, and 4) lenticular couplets.

Other features recorded in the lower Libby Formation include stromatolites, ooids, and fenestral fabrics.

The depositional setting of the lower Libby Formation can be interpreted with the aid of member and lithofacies correlation from Libby to Clark Fork and Fishtrap Creek. The spatial and vertical relations among members A through D are shown in Figure 27. The basal microlaminated argillite of member A represents an incursion of a deeper, more offshore environment than succeeding members B and C as evidenced by the lack of shrinkage cracks (which B and C contain), and the finer-grained, more reduced lithology. Factors influencing this incursion could have been a transgression and/or loss of supply of terrigenous clastics. Members B through E are interpreted as a series of facies that represent progressively more offshore environments. A spatial reconstruction using Walther's Law is shown in Figure 27C. The even couplets of member B were occasionally exposed producing weakly developed mudcracks. The greater abundance of mud cracks in the carbonate-bearing member C suggests slightly shallower water or, at least more frequent subaerial exposure than in member B. Perhaps the development of patches of small stromatolites and oolite shoals created enough relief to either cause a broad scale shallower environment or to produce isolated pools that could dry up if cut off from the rest of the waters in the basin in which the Libby Formation was deposited. The irregular wavy laminated siltite in member D was not subaerially exposed. The parallel laminated dark gray argillite of member E is interpreted as offshore from member D.

The thick, widespread influx of fine-grained terrigenous clastic material is a distinctive feature recorded in the Belt Supergroup (and hence, the lower Libby Formation). Terrestrial vegetation did not occupy the source regions of these Middle Proterozoic rocks. Three analogs from different sedimentological settings that approximate the above-mentioned conditions of Beltian sedimentation are considered here in an attempt to explain the distribution of sediment types and other features of the lower Libby Formation. These analogs include Eocene lacustrine environments (Green River Formation), modern restricted marine tidal flats (Colorado River Delta region), and modern open marine shoreface muds (Suriname).

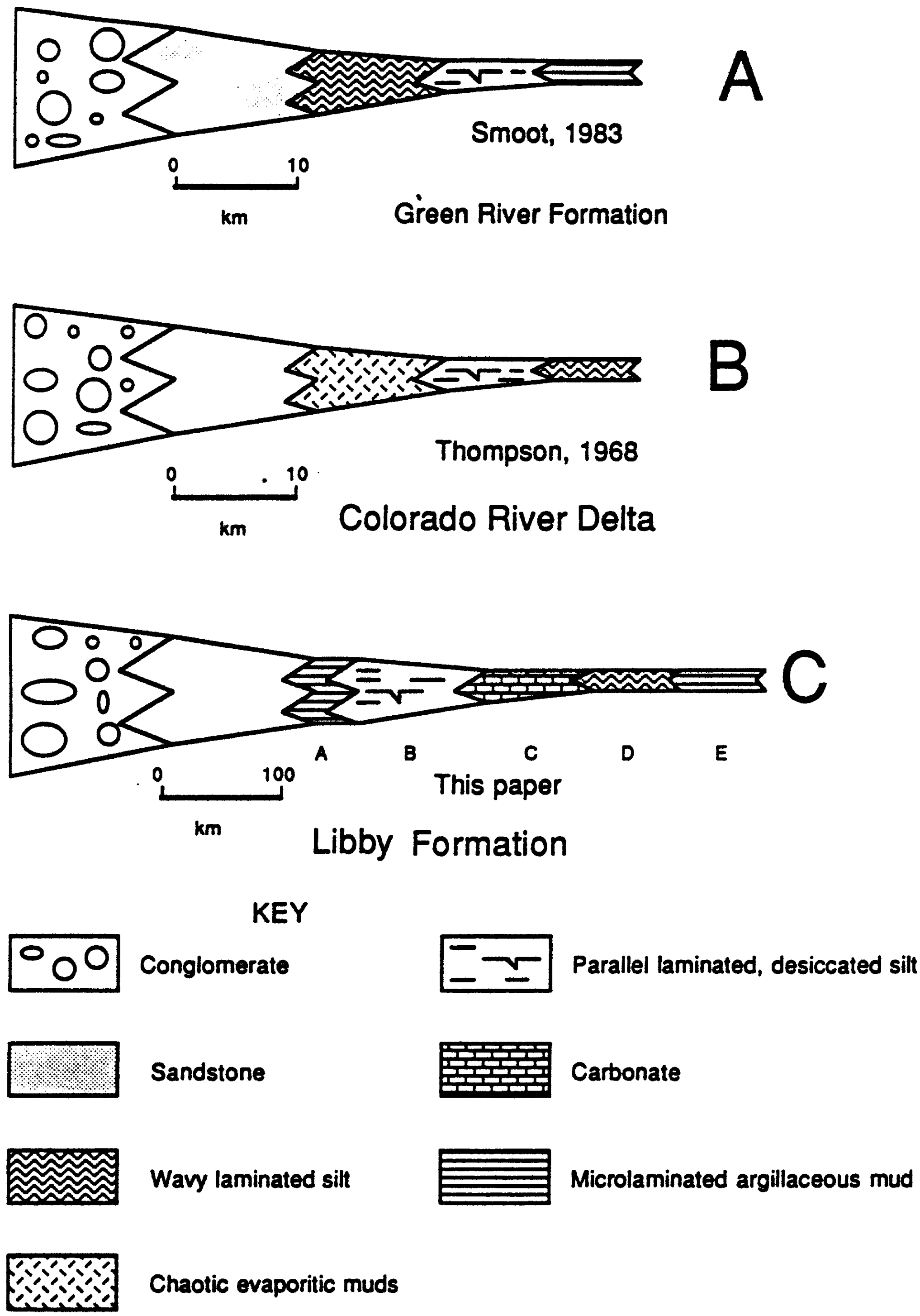


Figure 27. Comparison of facies relationships from Colorado Delta, Green River Formation, and members A-E of Libby Formation. Libby Formation relationships derived through application of Walther's Law. More complete explanation in text.

Restricted Marine Analog

Tidal flats on Colorado River Delta

The tidal flats on the Colorado River delta have been chosen as a potential analog for the Libby Formation because these tidal flats formed in a drainage area characterized by an arid climate. The lack of vegetation makes this a more suitable tidal flat analog than the classic North Sea tidal deposits. A second reason for choosing the Colorado delta is that the sediment on the delta and on the tidal flats is primarily composed of silt and clay-sized material (Thompson, 1968) much like the sediments in the Libby Formation. A significant difference between these modern tidal flats and those in the Libby Formation is the abundance of evaporites that developed on the surface of the tidal flats in the Salton Trough. These evaporites are recorded as laminae and crack-fillings of gypsum. The crystallization of gypsum often deforms the laminae in the upper few centimenters of the sediment column into chaotic muds (Thompson, 1968). The only evidence for evaporite deposition in the Libby Formation or equivalent rocks comes from thin salt cast horizons in the McNamara Formation at Bonner (Smith and Barnes, 1966). The general lack of salt casts at this level in the Belt section suggests that although modern arid conditions may closely approximate Proterozoic physical sedimentation patterns, sediments in the Libby Formation and equivalents may have been deposited under more humid conditions.

Thompson (1968) divided the tidal flats on the Colorado River Delta into three subenvironments, each with representative sediment types. The high tidal flats are dominated by chaotic muds composed of gypsum, halite, and silty clay. The intertidal flats are composed of brown laminated muds with laminae 1-8 mm thick, but commonly 2-3 mm thick. These laminae are regular and even in the upper part of this zone and become more irregular towards the lower part of the intertidal flats. Sand bars are present at places in the lower intertidal zone. The subtidal flats were only scantily studied by coring. They have been observed to 11-12 m deep and the sediments are characterized by gray silty clay. Lamination is irregularly spaced at intervals of 1-10 cm.

Comparison of sediment types in the Libby Formation with photographs from Thompson's (1968) work shows that the even couplets of member B closely resemble those of the brown laminated muds of the intertidal zone. Irregular lamination of the wavy and lenticular couplets of member D resembles the irregularly laminated silty clay of the subtidal zone off the Colorado River Delta.

The distribution of laminae types in the Colorado Delta region is consistent with the spatial reconstruction of sediment types in members A-E of the Libby Formation (Fig. 27B). The major difference between this modern analog and the Libby Formation is that the reconstruction for the Libby Formation includes stromatolitic and oolitic carbonate (member C) that was apparently subjected to more frequent subaerial exposure than members B or D. This may have been due to increased relief produced by the development of stromatolitic zones and oolite shoals.

Although the tidal range on the tidal flats of the Salton Trough is one of the highest in the world (8-10 m at spring tide), there are virtually no tidal channels in the region (Thompson, 1968).

Lacustrine analog

Wilkins Peak Member of Green River Formation

An assemblage of facies similar to many of those in the Belt Supergroup and those from the Colorado River Delta region has been described in detail by Smoot (1983). These sediment types show basinward fining (Fig. 27A) from conglomerates to sheet flood sands to wavy laminated dolomitic mudstones to extensively mudcracked, more evenly laminated dolomitic mudstones, and finally to laminated organic-rich mudstones. The progression of sedimentary facies and their associated structures closely resembles those in the Libby Formation (Fig. 27C). Indeed, Winston (1986d) compared the facies described by Smoot (1983) in detail with sediment types in older Belt rocks and found a close correlation between the types of sediments and their lateral and vertical relations. Data from the present study show that the lower Libby and McNamara Formations (not incorporated into Winston's studies) conform to the facies patterns recognized by Winston in older Belt rocks.

Open Marine Analog

Mud shoreface of Suriname

Several workers have proposed an open physiographic setting for deposition of Belt-Purcell sediments (Price, 1964; Harrison, 1972; McMechan, 1981). Most modern shallow, open-marine environments, however, are characterized by sandy (terrigenous or carbonate) settings. A unique exception of extensive open- marine, shallow-water mud deposition occurs along at least one hundred km of the coast off Suriname (Rine and Ginsburg, 1985). The facies of this muddy shoreface represent another potential modern analog for the lower Libby Formation.

The two main features of the Suriname muddy shoreface are shifting mud banks and interbank zones (Rine and Ginsburg, 1985). Sedimentologically, the mud banks are dominated by massive, homogeneous beds that are as thick as 2.2 m. Supplementary sedimentary structures consist of parallel and subparallel laminations with individual laminae that have sharp bases and graded tops, wavy and lenticular lamination, scour and fill features, micro cross-lamination, and soft-sediment deformation features. All of these subordinate structures are present in the lower Libby Formation, and x-ray radiographs of these modern features resemble those seen in outcrops of the Libby Formation. Muds of the interbank zone have a pudding-like consistency which is more consolidated than the yogurt-like or fluid-mud consistency of the mud-bank muds. The most common sedimentary features of the interbank foreshore are furrow structures, burrows, mud cracks, and armored mud balls (Rine and Ginsburg, 1985). Bioturbation dominates the subtidal interbank zones.

The massive nature of the mud banks was interpreted by Rine and Ginsburg (1985) as the result of fluid flow as the banks migrate. The sedimentary structures resembling those in the Libby Formation are present in only trace amounts, and no characteristic pattern of spatial relationships between these sedimentary structures has been documented. Were it not for the dominance of the shifting mud banks, this environment might compare more closely to those in the Belt Supergroup. However, it appears that a suitable modern or recent analogue for the Libby Formation in an open marine setting has yet to be documented.

Discussion

Deposition of large volumes of mud in shallow water can only be accomplished in several ways. Quiet water deposition is most commonly invoked for such deposition (Pettijohn, 1975) although there are documented cases of high energy muds (Potter et al., 1980) including the fluid-muds deposited on the Suriname shoreface (Rine and Ginsburg, 1985). Dominance of massive muds such as on the Suriname shoreface does not occur in the lower Libby Formation even though many of the subordinate sedimentary structures are similar. It is possible that the Suriname analogue could explain the dark gray massive mudstones in the upper Libby Formation, but the exposures on Flagstaff Mountain do not permit such a comparison. Thicker exposures of this lithology in the Whitefish Range (Whipple, 1984) may provide sufficient information to allow comparison to the Suriname analogue.

The two remaining potential analogues in many ways compare closely with the deposits in the lower Libby Formation. Facies relationships for these three settings are compared in Figure 27. The lateral progression of facies is generally similar in all three cases although the scale is quite different.

Conglomeratic facies in Figure 27 are inferred for the lower Libby Formation (~McNamara Formation). Sandy facies are present at Porters Corner near Phillipsburg, MT, in the McNamara Formation. These sandstone beds may coarsen to the south in the area of the Dillon quadrangle.

Comparison of muds from the Libby Formation, Colorado River Delta tidal flats, and Wilkins Peak Member shows several differences. Relative to the parallel laminated facies, the Colorado River Delta region and the Libby Formation are both characterized by wavy laminated facies in more offshore positions than in the Wilkins Peak Member. A microlaminated facies was not reported by Thompson (1968), but inasmuch as the wavy laminated facies (Fig. 27B) represents the most distal portion of his sampling, it is possible that there are microlaminated muds further offshore from the subtidal deposits studied on the Colorado River Delta. The presence of halite/gypsum-bearing chaotic muds differentiates the Colorado River Delta tidal flats from both of the other two analogs. This suggests that although the modern arid environment simulates the physical processes of transporting and depositing detrital particles in the Proterozoic, environmnents in the Libby Formation were generally not evaporative except for parts of the McNamara Formation discussed earlier.

Winston et al (1984) embraced a lacustrine model in part because tidal channels have generally not been observed in Belt rocks. However, deposits of the modern Colorado River Delta tidal flats are mostly lacking in channels (Thompson, 1968). When present, these channels are broad and shallow. Possible broad and shallow channels in the lower Libby Formation were described earlier from Fishtrap Creek. These may have been tidal channels. The point is, that the tidal flats of the Colorado River Delta provide an analog that generally lacks tidal channels even though the tidal range is 8-10 m.

The facies relationships of both the Colorado River tidal flats and the lacustrine deposits of the Green River Formation resemble those of the Libby Formation. Although the basinward facies progression of the Colorado Tiver tidal flats more closely resembles that of the lower Libby Formation, this resemblance does not prove a lacustrine or restricted marine setting for the lower Libby.

Other considerations

Arid lacustrine environments are characterized by a suite of unusual evaporite minerals that are not present in arid shallow marine settings (Picard and High, 1972). Grotzinger (1986) has compared this assemblage with the carbonate mineral associations in the Wallace Formation (Middle Belt Carbonate) which has been considered to be a "more restricted" environment than the upper Libby/Garnet Range Formations (Winston 1986d). The Wallace Formation lacks the unique mineral assemblage often found in arid lacustrine settings.

The oolitic assemblage described in the petrography section is interpreted as an originally mixed assemblage of calcitic and aragonitic ooids. The only modern examples of calcitic ooids are from the restricted marine Baffin Bay, Pyramid Lake, Nevada, and redeposited ooids from the Great Barrier Reef (Marshall and Davies, 1975) and the Amazon shelf (Milliman and Barreto, 1975). Tucker (1984) suggested that the presence of calcite ooids in the Belt Supergroup (Siyeh (Helena) and Snowslip Formations) might be used as evidence for restricted marine deposition. However, there are enough Phanerozoic examples of calcitic ooids that were deposited in environments interpreted as marine (Sandberg, 1975, Wilkinson and Landing, 1978) to negate definitive environmental interpretation of the Belt based on inferred primary ooid composition.

Upper Libby Formation

Two facies dominate the upper Libby Formation and the Garnet Range Formation. These are: 1) hummocky cross-stratified quartzite and coarse-grained siltite with argillitic partings, and 2) mostly parallel laminated dark gray argillite. On Flagstaff Mountain, the hummocky cross-stratified facies is sandwiched between dark gray argillite packages. Near the town of Bonner, the hummocky cross-stratified facies is underlain by dark gray argillite interbedded with planar bedded quartzite units. At Bonner, the hummocky cross-stratified facies is overlain by the Pilcher Quartzite (McGroder, 1984a). In the Whitefish Range, the hummocky cross-stratified quartzite (HCS) is overlain by a thick dark gray argillite unit and is underlain by reddish purple argillite (Whipple, 1984). All three of these localities are characterized by an interval of quiet water deposition represented by argillitic rocks interrupted by deposition of coarser sediments under higher energy conditions.

The hummocky cross-stratified quartzite is best exposed and coarsest-grained in the type area of the Garnet Range Formation near Bonner (McGroder, 1984a). A number of features are shared between a typical bed of HCS at Libby and the HCS near Bonner. These common features are: 1) scoured bases of beds with occasional rip-up clasts of silty argillite (or sometimes chert at Libby), 2) low angle cross-stratification, and 3) hummocky bed tops of comparable scale. The beds in the Garnet Range Formation are graded at the top and capped by symmetrical ripples which are absent in the upper Libby Formation. The bases of beds in the Garnet Range Formation are commonly planar bedded. This is a much more subtle feature in the finer grained and less well exposed deposits at Libby. Figure 28 shows an generalized view of the key features observed in hummocky cross-stratified beds near Bonner and in less satisfactory detail on Flagstaff Mountain. McGroder (1984a) did a detailed analysis of the hummocky strata in the Garnet Range Formation. I have seen many of the rocks studied by McGroder, I find his arguments sound, and I accept his model for deposition of the HCS in the Garnet Range Formation and apply it to the Libby Formation with the restriction that the energy of deposition was somewhat less in these more distal deposits. McGroder (1984a) evaluated the hummocky cross-stratification in terms of three models of sediment transport:

1) Storm ebb

2) Density current generation

3) Wind driven bottom current

The storm ebb model (Hayes, 1967) used to explain the deposits of Hurricane Carla has largely been rejected in favor of more sophisticated models (Walker, 1984). The density current generation model (Hamblin and Walker 1981) would produce directional sole marks on the bases of sand beds. Such sole marks are absent at good exposures of the hummocky cross- stratified quartzites at Bonner, Superior, Fishtrap Creek, and Libby, thus making the density current model an unlikely mode of transport for these sandstones. Transport by wind generated bottom currents (Morton, 1981) explains all of the features reported by McGroder (1984a) and in this study except the symmetrical ripple marks that occur on the tops of many hummocky cross-stratified beds in the Garnet Range Formation. If these ripples were generated by a waning wind generated bottom current as stated by McGroder, they should be asymmetrical instead of symmetrical. An alternative explanation for these ripples is that once the wind driven bottom current had waned, fair-weather processes were then able to express themselves on the bottom sediments. The absence of these ripples in the upper Libby Formation can be explained by deposition in slightly deeper environments (below fair weather wave base).

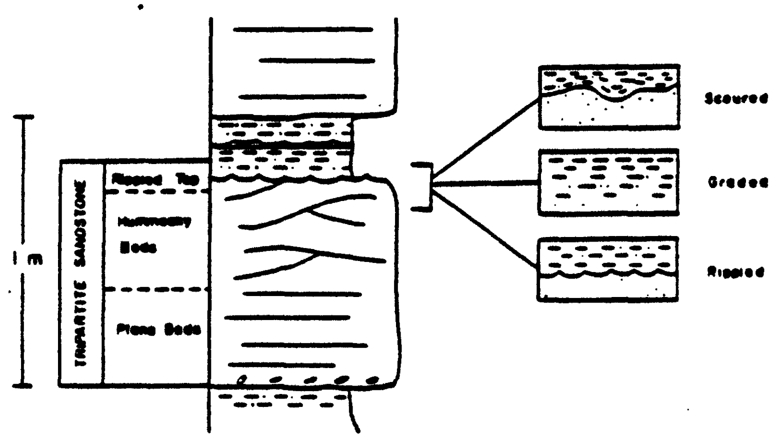


Figure 26. Characteristic stratigraphic sequence of facies three (cyclic hummocky bedded sandstone facies) of McGroder (1984) in the Garnet Range Formation in the Missoula region. Sandstone commonly grades from plane bedded to hummocky bedded to wave rippled. Some sandstone bed tops are scoured, others graded (from McGroder, 1984a).

The dark gray argillite above the hummocky cross-stratified quartzite at Libby and in the Whitefish Range is similar in that it contains thin zones of quartzite with some hummocky cross-stratification. This represents a waning of energy and/or loss of sand supply. Deposits such as these are recorded in the Newland (Limestone) Formation deposited in the lower Belt Supergroup in the eastern part of the basin (Schieber, 1984). These largely structureless mudstones could represent deposition as mud banks like those on the modern Suriname shoreface (Rine and Ginsburg, 1985) or the massive muds interbedded with thin sandy beds might record a pre-flysch type of deposition at the top to the Libby Formation.

Summary

Detailed facies analysis of the lower Libby Formation does not allow resolution as to whether the environment of deposition was marine or non-marine. Unique minerals that form in arid lakes are not present in the lower Libby Formation. The distinctive assemblage of what are interpreted as mixed aragonitic and calcitic ooids only has one modern analog in the hyper-hypo-saline water of Baffin Bay, Texas (Land et al., 1979). The only statement that can be made with certainty about the lower Libby environments is that they were shallow. Environments become shallower from the town of Libby to the south and east as the equivalent McNamara is characterized by textbook examples of desiccation cracks and red beds in addition to the stromatolites, oolites, ripple marks and incipient shrinkage cracks present at Libby. Consideration of all data presently suggests the interpretation that deposition occurred under restricted marine conditions in a sub-humid to humid climate.

The upper Libby Formation was deposited in deeper water than the lower Libby Formation. The upward disappearance of oolites, stromatolites, abundant symmetrical ripple marks, and abundant shrinkage cracks interpreted as incipient desiccation cracks combined with the predominance of shrinkage cracks interpreted as syneresis cracks suggests that the hummocky cross-stratified facies in the Libby region was deposited below fair weather wave base in the Libby region, although this facies was probably deposited above fair weather wave base in the Missoula region.

MICROPALEONTOLOGY

The recent success of many micropaleontological investigations in upper Proterozoic, fine-grained sedimentary rocks (Vidal and Knoll, 1983) has laid the groundwork for similar investigation in Middle Proterozoic rocks. This recent research has focused on acritarchs, organic-walled microfossils of uncertain taxonomic affinity, often found preserved in shale and silt facies like those of the Libby Formation. Inasmuch as the Libby Formation contains the least metamorphosed rocks in the western part of the Belt basin, it was hoped that these rock types that so often yield microfossils in younger Proterozoic rocks might preserve microfossils that could be used to estimate the biostratigraphic age of the Libby Formation.

Both acid macerations and bedding-parallel thin sections of thinly-laminated, dark gray argillite from the stromatolitic and oolitic member C of the Libby Formation on Flagstaff mountain have yielded acritarchs. These microfossils are moderately well to poorly preserved, and by techniques used in this study to date, recovered only in small amounts (approximately 20 - 30 microfossils per microscope slide). One large acritarch was observed in a bedding-parallel thin section of dark gray argillite.

Techniques

The following procedures were followed in maceration of samples:

1) 30 - 50 mg samples of argillite and siltite were scrubbed and soaked in a solution of potassium dichromate and reagent grade sulfuric acid to remove modern organic contaminants from surfaces of samples.

2) Samples were crushed with a mortar and pestle to particles with maximum diameters in the granule size range.

3) Some powdered sample from the crushed material was squirted with 10% HCl to check for a carbonate reaction. If no carbonate was present in the sample, the next two steps were skipped.

4) Carbonate-bearing samples were soaked in 10% HCl for several days to two weeks depending on the duration of the reaction.

5) Samples from step 4 were rinsed by multiple centrifuge treatment.

6) Carbonate-free samples were soaked in concentrated HF for several days. These samples were periodically stirred and supplied with fresh HF until most or all of the chunks were dissolved.

7) Mud from HF procedure was rinsed by placing samples in sealed bags constructed of dialysis tubing and left in running deionized water overnight.

8) Samples were mounted on glass slides in a mixture of Kumar resin and xylene for examination under the microscope.

Systematic Paleontology

Group Acritarcha Evitt, 1963

Genus cf. Kildinosphaera Vidal, 1983

(in Vidal and Siedlecka, 1983)

Figure 29A

cf. Kildinosphaera sp. Light brown, smooth-walled fragment of acritarch with distinctive, overlapping folds. The fragment represents approximately one fourth of a spherical fossil. Its maximum dimension is 27 um, and an extrapolated diameter for the entire fossil is approximately 108 um, which is within the range of several species of Kildinosphaera described from the Upper Riphean of northern Norway (Vidal and Seidlecka, 1983). The limited information available for this fragment is also consistent with the definition for the form genus Leiosphaeridia.

Stratigraphic range: The apparent stratigraphic range for Kildinosphaera is Upper Riphean-Vendian (Vidal and Seidlecka, 1983).

Unnamed spheroid A

Figure 29B

Sphaeromorph characterized by narrow, well-defined thickened dark brown rim. Central area is composed of thin, smooth walls that are greenish gray or brownish gray. Size ranges from 8 - 32 um (x- = 17.5, n = 26, Fig. 30). Some examples display short (3-5 um) wispy folds and/or crudely defined bumps 1-3 um in diameter in the central, otherwise smooth-walled region. Many examples have narrow, slit-shaped cracks in the central, thin-walled region. Two examples are perforated with < 1 um diameter circular holes. One example has perforations evenly distributed throughout the central region, whereas the other only has holes in a localized

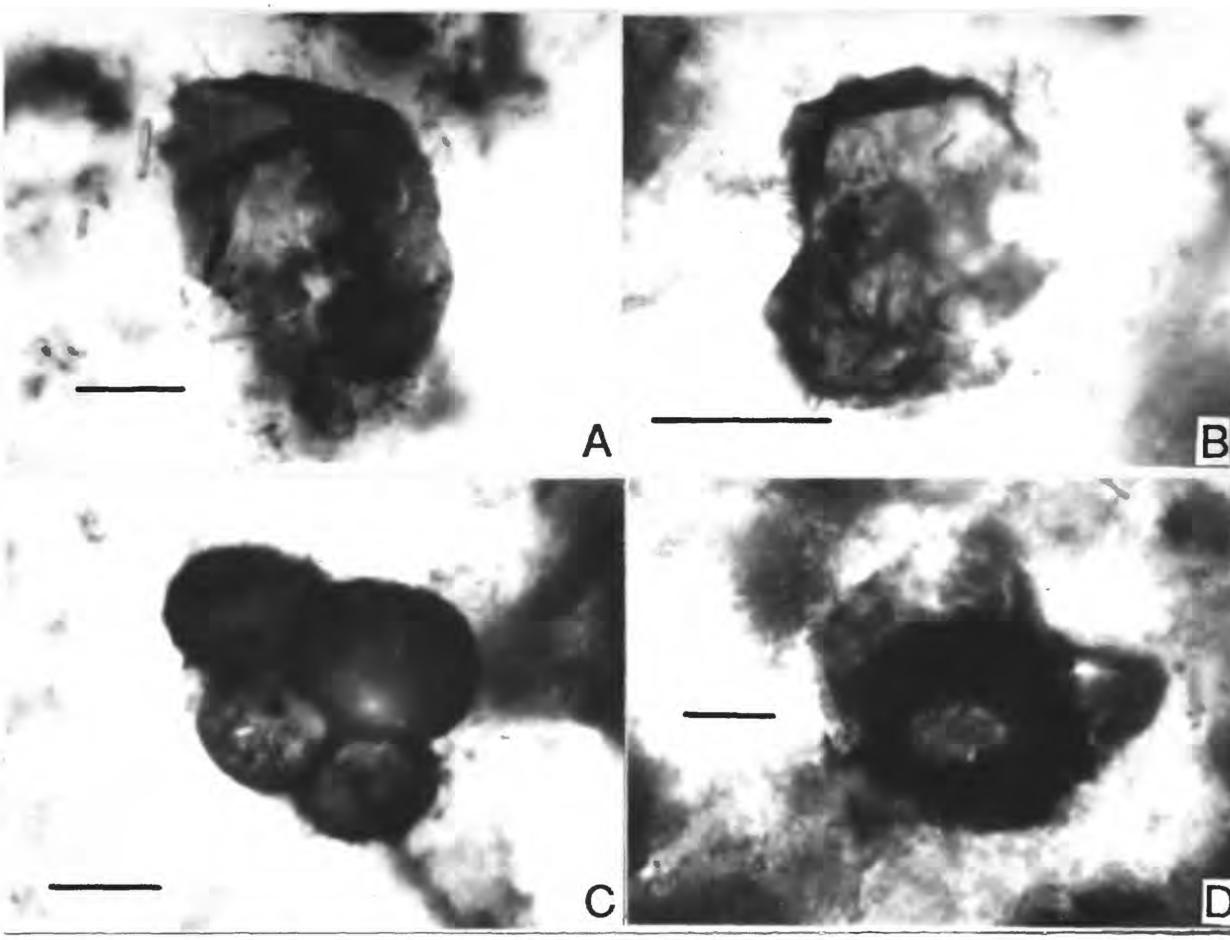


Figure 29. Acritarchs preserved in macerals of dark gray argillite in member C of the Libby Formation on Flagstaff Mountain. A. cf. Kildinosphaera fragment. B. Ovoidal example of unnamed spheroid A with partial thickened rim preserved. C. Cluster of four spheroids, each with an internal spot. D. Torus-shaped microfossil (fragment) with wing-like projections connected to torus. Bar scales for A-C are 10 micrometers long. Scale for D = 5 micrometers.

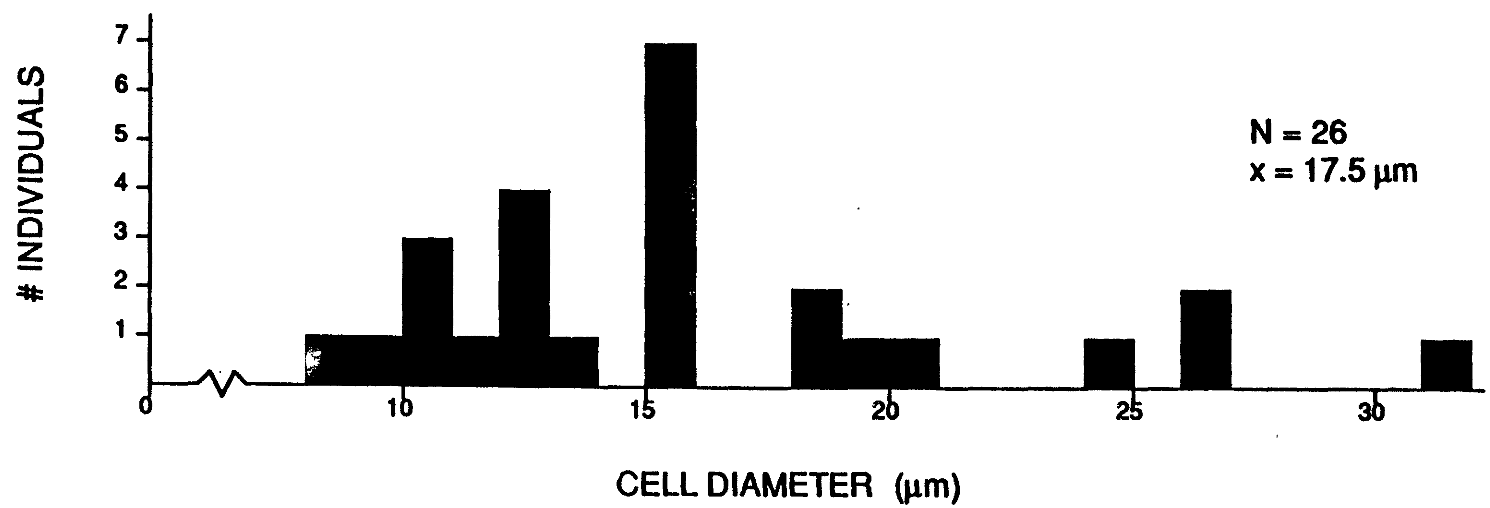


Figure 30. Size-frequency distribution plotted for unnamed spheroidal acritarch A.

area. One example was pear-shaped.

Unnamed spheroid B

Medium-brown sphaeromorph with smooth, featureless walls. Diameters of the four specimens examined are 19 um, 19 um, 27 um, and 30 um. One example is characterized by a feature that can be interpreted as a median split.

Unnamed spheroid C

Medium-brown to greenish-brown sphaeromorph with distinctive, short (3-5 um) linear folds near and parallel with the margin of the fossil. Cell walls are thin and smooth. Diameter (two specimens) ranges from 13-17 um.

Unnamed tetrad

Figure 29C

Four dark greenish-brown spheroids form a cluster in a cubic open packing configuration. The spheroids partially overlap rather than abutting adjacently to one another. Each spheroid is 9 um in diameter and has a 1-2 um diameter dark brown spot near its center.

Unnamed fusiform fossils

The best preserved fusiform morphs are dark brown, have smooth walls, and are characterized by abundant, irregularly oriented and irregularly distributed folds crudely sub-perpendicular to cell margins. These two morphs range in length from 22-27 um and in width from 17-19 um. Ten additional fusiform morphs that lack surface folds display colors of greenish-gray, light, medium, and dark brown. Length of these forms ranges from 13-49 um, and width ranges from 8-38 um. Length plots as a straight line against width with a good fit (Fig. 31, $R = 0.95$) suggesting a possible taxonomic affiliation between all of these fusiform morphs.

Large sphaeromorphs

A small number of large, ovoidal sphaeromorphs were recovered. These are generally not well-preserved. The three best examples are: 1) a dark brown to black ovoidal fossil (54 um x 75 um in size) with a few tiny, slit-like cracks near the cell margin, 2) a medium brown, plain, smooth-walled spheroid, 30 um in diameter, adjacent to an ovoidal fragment of similar composition that is 80 um long and approximately 40 um wide, and 3) a 150 um diameter ovoidal fragment with a thickened dark brown rim and a highly degraded and cracked, thin-walled, greenish-gray central region. The latter example is preserved in a bedding- parallel thin section.

Unnamed torus

Figure 29D

Dark brown, thickened ring-like structure with two light greenish-gray, smooth, scoop-shaped projections (wings) attached to the inside of the ring. The torus is 13 um in diameter. Wing-span is 19 um.

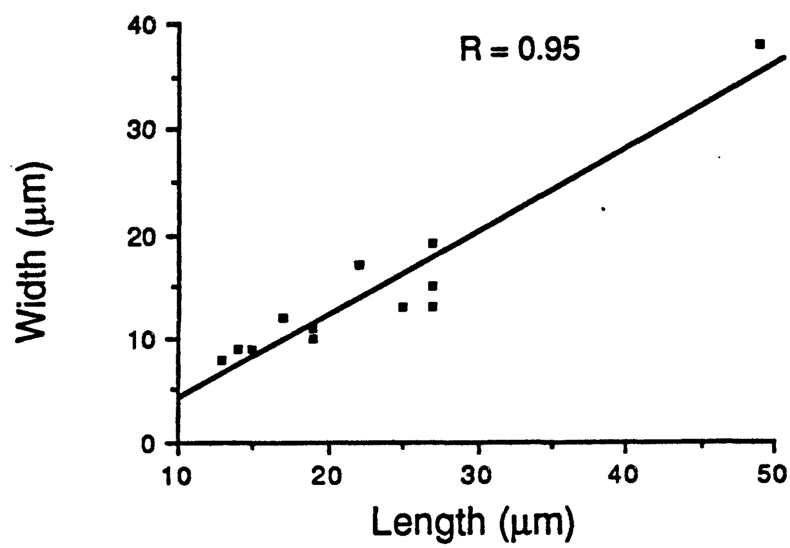


Figure 31. Scatter plot of length vs. width of fusiform acritarchs.

Unnamed club-shaped fragment

Greenish-brown fragment with smooth, thin walls, straight folds, and narrow, slit-like cracks. Preserved length is 45 um, width at tip is 15 um, and width at broken end is 8 um.

Discussion

The low diversity acritarch biota in the Libby Formation is consistent with any of the widely divergent ages of 900 Ma (Obradovich et al., 1984) or 1250 Ma (Elston, 1984) for the upper Missoula Group. It is clear that Late Proterozoic acritarchs in open marine settings are quite diverse (Vidal and Knoll, 1983), but the diversity level of Middle Proterozoic acritarchs is poorly known although some studies show lower diversity characterized by simpler morphs than in the Late Proterozoic (Peat et al., 1978; Horodyski, 1980; Vidal and Knoll, 1983).

Late Proterozoic acritarch diversity can be strongly influenced by depositional environment. Paleoecological studies of Proterozoic acritarchs have shown that restricted sedimentary environments preserve low diversity microbiotas, whereas open marine assemblages record high diversity (Vidal and Knoll, 1983; Knoll, 1984). Micropaleontological success in the Proterozoic to date has been achieved with marine rocks. Therefore, even if the acritarch data seemed more biostratigraphically definitive, it would still be suspect because of the uncertainty of the depositional environment of the Belt Supergroup discussed in other parts of this report.

Acritarch size distribution may provide another approach to the question of environment versus age. A tabulation of the size distribution of all of the microfossils recovered from the Libby Formation shows that 75% of the fossils are smaller than 30 um, 23% and 2% of the acritarchs are greater than 30 um and 100 um respectively. Comparable studies of younger Proterozoic planktonic organisms from restricted environments show size distributions as follows:

Location	x <30um	30um>x<100um	100um<x
Bitter Springs Fm (Vidal & Knoll, 1983)	most	rare	unknown
Hunnberg Fm (lagoonal) (Knoll, 1984)	96%	3%	1%
Hunnberg Fm (open coastal) (Knoll, 1984)	81%	15%	4%

Comparison of the size distribution of the acritarchs of the Libby Formation with the above results supports an open marine setting for the Libby Formation if acritarch size distribution can be used as a barometer of depositional environments. However, much further work must be done before acritarch size can be used as an indicator of any of marine vs. non-marine depositional environments.

TECTONICS

The Proterozoic tectonic setting of the Belt basin figures prominently in several interpretations for the tectonic evolution of the middle and upper Proterozoic margin of western North America (Stewart, 1972; 1976; Monger et al., 1972; Burchfiel and Davis, 1975; Monger and Price, 1979). The Belt basin has been interpreted as being a cratonic reentrant (Harrison et al., 1974), an epicratonic trough (Stewart, 1976), and an intracratonic basin (Winston et al., 1984).

Numerous workers have reported evidence for tectonic activity on a variety of scales during deposition of the Belt Supergroup (McMannis, 1963; Harrison et al., 1974; Winston, 1978, 1986a,b; Wallace et al., 1984, Godlewski and Zieg, 1984; and many others). Most of this work has focused on areas in the southern, eastern, and central parts of the basin in rocks older than the Libby Formation. Patterns of sedimentation in the upper Libby Formation differ from those in underlying Belt rocks. The hummocky cross-stratified facies represents a new style of sedimentation in the upper Belt Supergroup. Changing tectonics may have caused the shift in sedimentation from a pattern that had dominated during deposition of at least seven kilometers of underlying rocks. Although many studies of the interplay between tectonics and sedimentation involve coarse grained rocks, this study shows that lithofacies patterns in fine-grained sedimentary rocks can also be used successfully to interpret the manner in which tectonic movements are expressed in the deposition of sediments.

The hummocky cross-stratified lithofacies is coarsest and thickest at its southernmost known occurrence and becomes thinner and finer grained to the north and west. This lithofacies is widely distributed in the upper Libby Formation and lithostratigraphically equivalent beds in the Garnet Range Formation. The hummocky beds extend southeast from the town of Libby to the Missoula region where McGroder (1984a) reported a minumum thickness of 250 m and estimated as much as 500 m of this lithofacies in the Garnet Range Formation. Near Superior the hummocky lithofacies is at least 150 m thick. In a faulted section at Fishtrap Creek, the hummocky cross-stratified quartzite is fine-grained and approximately 100 m thick. At Libby (Fig. 10), the facies is 160 m thick. In the Whitefish range, the hummocky cross-stratified quartzite is may be as thick as 300 m (J.W. Whipple, pers. comm. 1986), but little is known of the facies in that region. Grain size varies from fine- to medium-grained sand near Missoula (McGroder, 1984a) to fine- grained sand at Superior and Fishtrap Creek. At Libby, the grain size is primarily fine-grained sand and coarse-grained silt.

Detailed analysis of the depositional environments of the Libby Formation suggests a deepening trend from the lower to upper Libby Formation. The upper Libby Formation was deposited below fair weather wave base and above storm wave base. Flysch- like rocks above the hummocky cross-stratified beds may signal a continuation of the deepening trend. Hummocky cross-stratified beds in the Missoula region were probably deposited above fair weather wave base.

Thickness and grain size trends in the hummocky cross- stratified quartzite in the Libby and Garnet Range Formations suggest a source area to the south and east. Paleocurrent data from the Libby Formation (Fig. 25) and Garnet Range Formation (McGroder, 1984a) are consistent with a southeasterly source terrain as are studies of the Bonner Quartzite (Winston, 1978, 1986d; Winston et al., 1986) which underlies the Libby and McNamara Formations. Thicker, coarser deposits in correlative units in the Whitefish Range (Whipple, 1984) indicate an eastern terrigenous source component.

The quartzite in the upper Libby and Garnet Range Formations is the only thick, widespread, sand-rich hummocky cross- stratified quartzite reported to date in Belt rocks. Schieber (1984) reported hummocky cross-stratified sandstone in the Newland Limestone (lower Belt) in the eastern Belt basin. These deposits, however are much more shale-rich and not as widespread as those discussed in this report. C.A. Wallace (pers. comm. 1986) has observed hummocky cross-stratification locally in the Mount Shields Formation (Missoula Group). E.R. Cressman (pers. comm.) reported scattered hummocky cross-stratified beds in one member of the Prichard Formation near Plains, MT, and in the Coeur d'Alene district of Idaho.

Petrofacies analysis shows an influx of detrital feldspar associated with progradation of the hummocky cross-stratified facies from areas near source terrains into the central part of the basin. This feldspar is texturally more immature than the terrigenous detritus in the lower Libby Formation. These data suggest that the hummocky cross-stratified sands were derived from newly rejuvenated source terrains. It is not possible with the available data to match the feldspars in the upper Libby Formation to any particular source terrain.

Discussion

The occurrence of thick, widespread, hummocky cross- stratified quartzite only near the top of the Belt Supergroup indicates that depositional conditions had changed near the close of recorded Belt sedimentation. Comparison of the differences between the lower and upper parts of the Libby Formation in the Libby area suggest that there are three major differences between the lower five members of the Libby Formation and the hummocky cross-stratified member F. These three differences are the upward change into:

1) sandier environments

2) interpreted deeper environments

3) interpreted higher energy environments

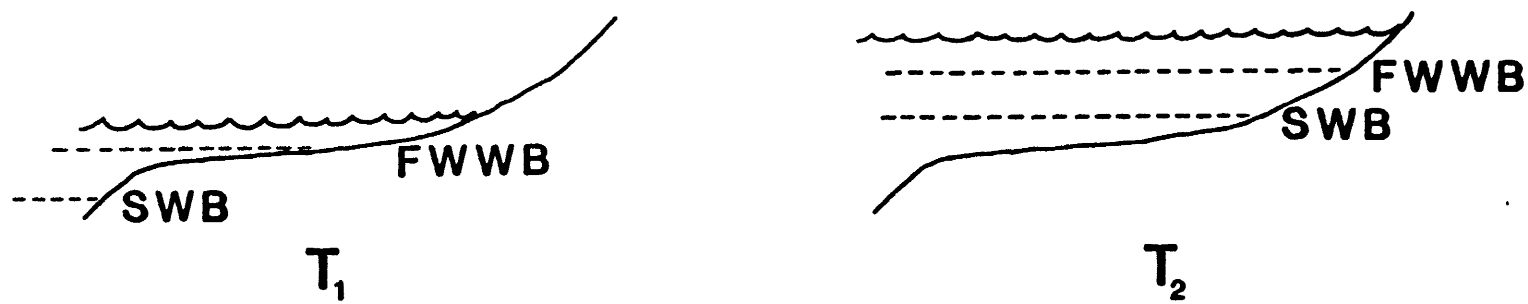
Two possible settings are evaluated here in an attempt to explain these changes. 1) Eustatic rise in sea level. (Fig. 32) A eustatic rise in sea level could result in depositional environments for the upper Libby Formation below fair-weather wave base, which would explain the presence of hummocky cross-stratification and the absence of shallow-water sedimentary structures common in the lower Libby Formation. Although it may be possible to transport a thickness of 100-200 meters of sand across several hundred kilometers of sea floor by storm activity, it seems more likely that a eustatic transgression would have trapped sand near source areas, as was the case along the east coast of North America following the Pleistocene ice age.

2) Offshore subsidence and onshore uplift. (Fig. 33) An increased rate of subsidence would have produced deeper and higher energy environments. Uplift of nearby source terrains would explain deposition of 100-200 m of sand over a large area.

The zone of soft-sediment deformation at the base of the hummocky cross-stratified quartzite can be interpreted as the result of a sudden deepening of the basin that may have been accompanied by tectonic movement. Allen and Banks (1972) concluded from a detailed analysis of deformed cross-bedding, particularly recumbently folded cross-bedding, that the most likely way to produce this structure is by liquification of the sediment with an earthquake shock followed by current drag on the sediment. Liquified sediment would also have facilitated development of the soft-sediment breccias and zones of deformed cherty clasts and thin beds. However, these soft-sediment features do not, by themselves, indicate tectonism. Liquified sediment could also be produced by high-energy waves or loss of support on the slope of the sea bottom (Allen, 1984).

Independent evidence for tectonic activity that may be related to the interpreted record of increased subsidence is available, but imprecise. Harrison (1986) documented broad-scale folding that affected much of Belt terrane, but all that is known of the age of this deformation is

EUSTATIC RISE IN SEA LEVEL



FWWB = fair weather wave base

SWB = storm wave base

Figure 32. Generalized cross-section view of idealized eustatic sea level rise in which water deepens, but terrigenous influx is restricted.

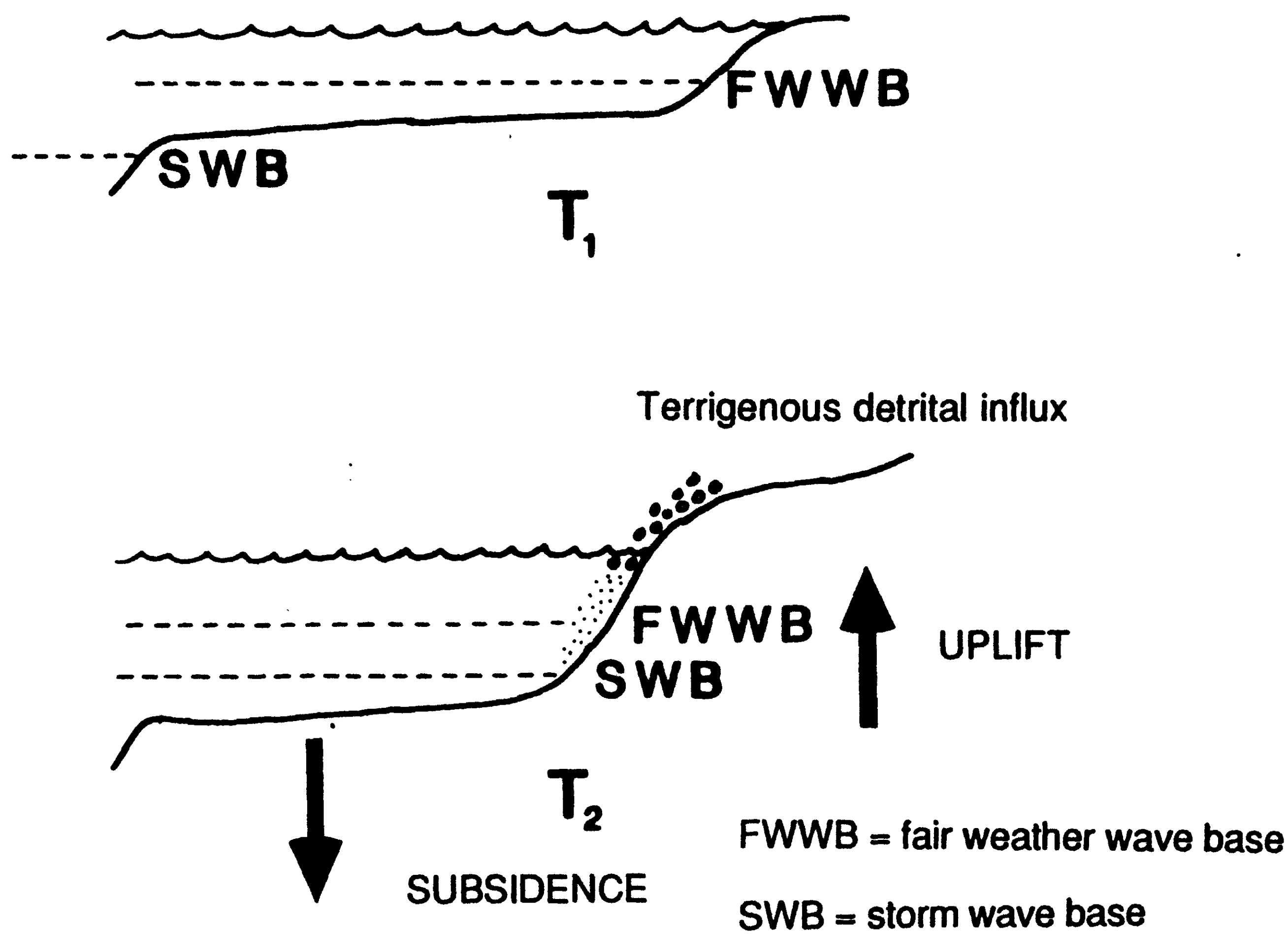


Figure 33. Generalized cross-section view of idealized combination of nearshore uplift combined with offshore subsidence showing deepening effect and terrigenous detrital influx that would result from uplifted source terrains.

that it occurred after deposition of the Belt Supergroup and before deposition of the Middle Cambrian Flathead Sandstone. The angular unconformity that cuts into rocks equivalent to the Libby Formation in Washington and British Columbia is capped by the conglomeratic base of the Windermere Supergroup. Some clasts of this conglomerate in British Columbia were derived from the Mount Nelson Formation (Lis and Price, 1976), which is partially equivalent to rocks of the Libby Formation. This unconformity is evidence for tectonic activity that post-dated the Belt Supergroup and pre-dated the Windermere Supergroup, but the amount of Belt sediment deposited above the Libby Formation and its stratigraphic equivalents will never be known, so the gap in time between Belt-Purcell and Windermere deposition is difficult to evaluate.

The development of this unconformity could have been related to pre-middle Cambrian block faulting and uplift reported by Lis and Price (1976) in British Columbia. Perhaps compression associated with the "East Kootenay orogeny" was related to this folding. This "orogeny" has been restricted in its definition to intrusion of the Hellroaring Creek Stock dated at 1305 $\pm$52 Ma using Rb-Sr whole-rock isochron analysis (Ryan and Blenkinsop, 1971; McMechan and Price, 1982).

The lack of reliable age control on Belt rocks and on tectonic events that affected those rocks makes it difficult to develop accurate tectonic models for the Belt basin. The most confident statement that can be made about the relationship between time of deformation and Belt deposition is that the deformation that produced the angular unconformity between rocks stratigraphically equivalent to the Libby Formation and rocks of the Windermere Supergroup was post-Belt and pre-Windermere. One can speculate that subsidence recorded in the Libby Formation signaled the beginning of terminal Belt tectonism. If, for example, Belt sedimentation was terminated by compressional tectonics, subsidence in the basin and uplift of southern source terranes could be explained by buckling of crust in an early compressional stage. Subsequent compression would have ultimately produced overall uplift resulting in eventual erosion of overlying (post-Belt/pre-Windermere) strata.

An earlier tectonic interpretation based on the hummocky cross-stratified lithofacies in the Garnet Range Formation in the Missoula region (McGroder, 1984; Winston, 1986b) suggested that deposition of the hummocky beds recorded the first incursion of marine waters into the Belt basin. This requires that Belt rocks older than the Garnet Range Formation (and upper Libby Formation) were not deposited in open marine waters (Winston et al., 1984; Winston, 1986d). Sedimentologic and stratigraphic results from this study of the Libby Formation do not as yet provide evidence for demonstrating a definitively marine or non-marine origin for the lower Libby Formation. Based on the available data, tectonic adjustments occurring during deposition of Garnet Range/upper Libby rocks can be explained by either offshore subsidence and onshore uplift in either a marine or non-marine setting or by opening of a restricted basin to open-marine waters.

Summary

Hummocky cross-stratified quartzite is a thick, widespread facies near the top of the Belt Supergroup. Although there are other occurrences of hummocky cross-stratification in the supergroup, none are as thick or extensive as this facies.

Analysis of depositional environments in the Libby Formation suggests deepening within the Libby Formation from environments above fair-weather wave base in the lower Libby Formation to environments below fair-weather wave base and above storm wave base in the upper Libby Formation.

The upper Libby Formation differs from the lower Libby Formation in that it is characterized by rocks deposited in sandier environments that are interpreted as forming under deeper and higher energy conditions. The simplest way to account for these changes is with subsidence in the Libby region and uplift in the Belt source terrain to the southeast, although it is quite possible that tectonic activity was more complex than this.

REGIONAL CORRELATION OF THE MIDDLE PROTEROZOIC OF WESTERN NORTH AMERICA

Proterozoic correlation must commonly be achieved by combination of several methods, and these methods may not produce accurate rock relations. The relative lack of biostratigraphically useful fossils compared to the Phanerozoic and imprecise radiometric age dating proves frustrating to the stratigrapher. The use of radiometric dating techniques has been encouraged over other methods of correlation for Precambrian rocks (Harrison and Peterman, 1980, 1984), but the necessary data are not available everywhere. Despite the often poor age control, correlation is being done successfully in many Proterozoic successions. Traditional lithostragraphy is useful only for intrabasin correlation, although distinctive lithologies such as upper Proterozoic diamictites (Harland, 1964; Hambrey and Harland, 1981) can be used as crude markers for interbasinal correlation. The use of unconformities will, as more detailed studies become available, become an increasingly important tool in Proterozoic correlation. Proterozoic biostratigraphy is coming into its own. However, correlation using columnar, columnar branching, and conical stromatolites often has limited resolution (Preiss, 1976) and acritarchs have only been sufficiently studied in rocks younger than 800 Ma (Vidal, 1976; Timofeev, 1976). Paleomagnetic approaches using magnetostratigraphy and polar wandering curves (Elston and Bressler, 1980; Elston, 1984; Lin et al., 1985; Roy, 1983; Park, 1984; Park and Aitken, 1986; Zhang and Zhang, 1985) have great potential, but the application of these techniques to Proterozoic rocks is still in its infancy. This section compares different methods of correlation for selected Middle Proterozoic sequences in western North America.

Radiometric Ages

Radiometric ages suggest that deposition of Belt rocks may have occurred over a time span of more than 500 million years. Ages that delineate the absolute maximum and minimum ages of the Belt Supergroup consist of a ~1700 Ma Rb-Sr whole rock isochron age for metamorphosed crystalline basement rocks beneath the Belt (Giletti, 1966) and ages ranging from approximately 750-900 Ma for the volcanic assemblage in the Windermere Supergroup (Miller et al., 1973; Armstrong et al., 1982). The most reliable intra- Belt age is 1433 ± 13 Ma determined by U-Pb analysis of zircon from the Crossport C sill in Idaho (Zartman et al., 1982). Ages for stratigraphically younger Belt rocks are less reliable. Rb-Sr whole rock isochron analyses (Obradovich and Peterman, 1968, 1973; Obradovich et al., 1984) yield ages of approximately 1300 Ma (Ravalli Group), 1100 Ma (Missoula Group), and 900 Ma (Garnet Range and Pilcher Formations). K-Ar analyses of glauconite (Obradovich and Peterman, 1968) yield ages of ~1100 Ma for formations ranging from Empire through McNamara (see Fig. 34 for techniques and rock types for Belt ages).

The most direct radiomet ic age estimate for the Libby Formation comes from a stratigraphic correlation (Harrison, 1972) to a 900 Ma old age based on Rb-Sr whole rock analysis of argillite in the Garnet Range Formation in the Alberton region approximately 30 miles west of Missoula (110 miles SE of Libby).

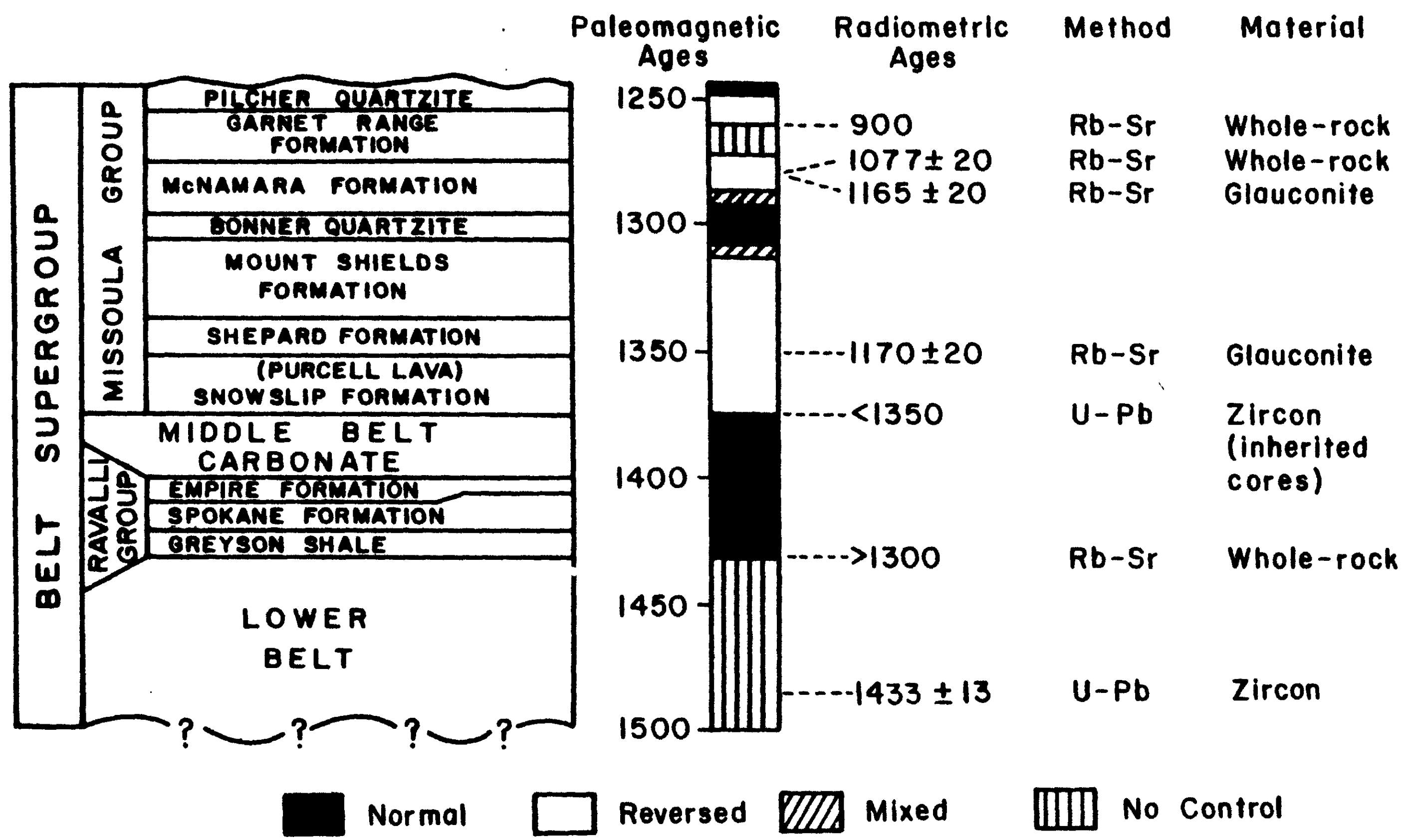


Figure 34 . Comparison of paleomagnetically derived (Elston, 1984) and radiometric (Obradovich et al., 1984) ages for the Belt Supergroup. All ages in Ma.

Lithostratigraphy

Crudely correlative diamictites are common in the Late Proterozoic of western North America (Stewart, 1972). These diamictites are sometimes associated with a volcanic assemblage that has been dated by K-Ar at 827-918 Ma (Miller et al., 1973) and Rb-Sr at 766-769 Ma (Armstrong et al., 1982). The diamictite assemblage that overlies Belt rocks and possible equivalents in Washington, southern Canada, Death Valley, Utah, Wernecke Mts. and the associated volcanics in Washington, southern Canada, Death Valley, and the Wernecke mountains are roughly synchronous given the uncertainties of the dating methods used on these often metamorphosed rocks (Stewart, 1972). In stratigraphic sections of the Belt and Death Valley regions, one or more unconformities are associated with the diamictite/volcanic assemblage (Fig. 35). This gap in the record may correlate to the unconformity between the Chuar and Unkar Groups of the Grand Canyon where diamictites are lacking. Chuaria is present in all sections above, but not below the unconformity(ies).

The generally similar stratigraphic position of major unconformities in many western North American sequences must be viewed with caution. The events that produced the unconformities could be temporally closely related to one another, diachronous, or completely unrelated. Until better age control is established, these unconformities must maintain a supportive role, at best, in regional correlation.

Lithostratigraphic arguments and radiometric age considerations on volcanic events suggest that the Wernecke Supergroup and the Pinguicula Group in northwestern Canada may be equivalent to the Belt Supergroup (Young, 1982). The Unkar Group and the lower parts of the Uinta Mountain Group (Crittendon and Wallace, 1973) and the Death Valley sequence may be Belt correlatives in the United States. Similar radiometric ages for lavas in the Belt Supergroup (Purcell Lava) and the Unkar Group (Cardenas Lavas) may provide an additional line of correlation between the Belt and Grand Canyon sections (Fig. 35). An unconformity separating the Pinguicula Group from the underlying Wernecke Supergroup (Fig. 35) occurs at a horizon may suggest relationship between tectonic events in the northern and southern Cordillera at this stratigraphic horizon.

Biostratigraphy

The difficulty in finding distinctive, rapidly evolving fossils in Proterozoic rocks older than 800 Ma allows only a few approaches for biostratigraphic correlation. This discussion will focus on three methods: 1) acritarch assemblages, 2) stromatolites, and 3) the use of Chuaria in Proterozoic correlation and consideration of why it is absent from the Libby Formation.

Acritarchs. Correlation using acritarchs is one successful biostratigraphic tool in Proterozoic rocks. Acritarchs have been successfully used in biostratigraphic correlation of Late Proterozoic (<800 Ma) sedimentary rocks from many regions of the world (Timofeev et al., 1976; Vidal, 1976), but have met with limited success in the biostratigraphy of older strata. Acritarchs have been recovered from rocks in the lower Belt in the eastern Belt basin (Horodyski, 1980) and from the upper Missoula Group in the northwestern Belt basin (this study). Lower Belt acritarchs from the Chamberlain Shale and Newland Limestone (Horodyski, 1980) and those of this study from the Libby Formation are represented by simple spheroidal forms, some of which have been deformed by shrinkage and compaction. A few of the microfossils from the Libby Formation vaguely resemble structures described by Jankauskas from the 1400 - 1600 Ma sequences in the Ural Mountains. (Vidal, 1984), but these comparisons are too tentative to make any definitive age comparisons. As discussed in the micropaleontology section, the assemblages of acritarchs from this study and from Horodyski's (1980) study could be dominated by simple spheroidal forms for reasons of either age or a restricted environment.

Stromatolites. Most Beltian paleontologic work has focused on stromatolites, which, though

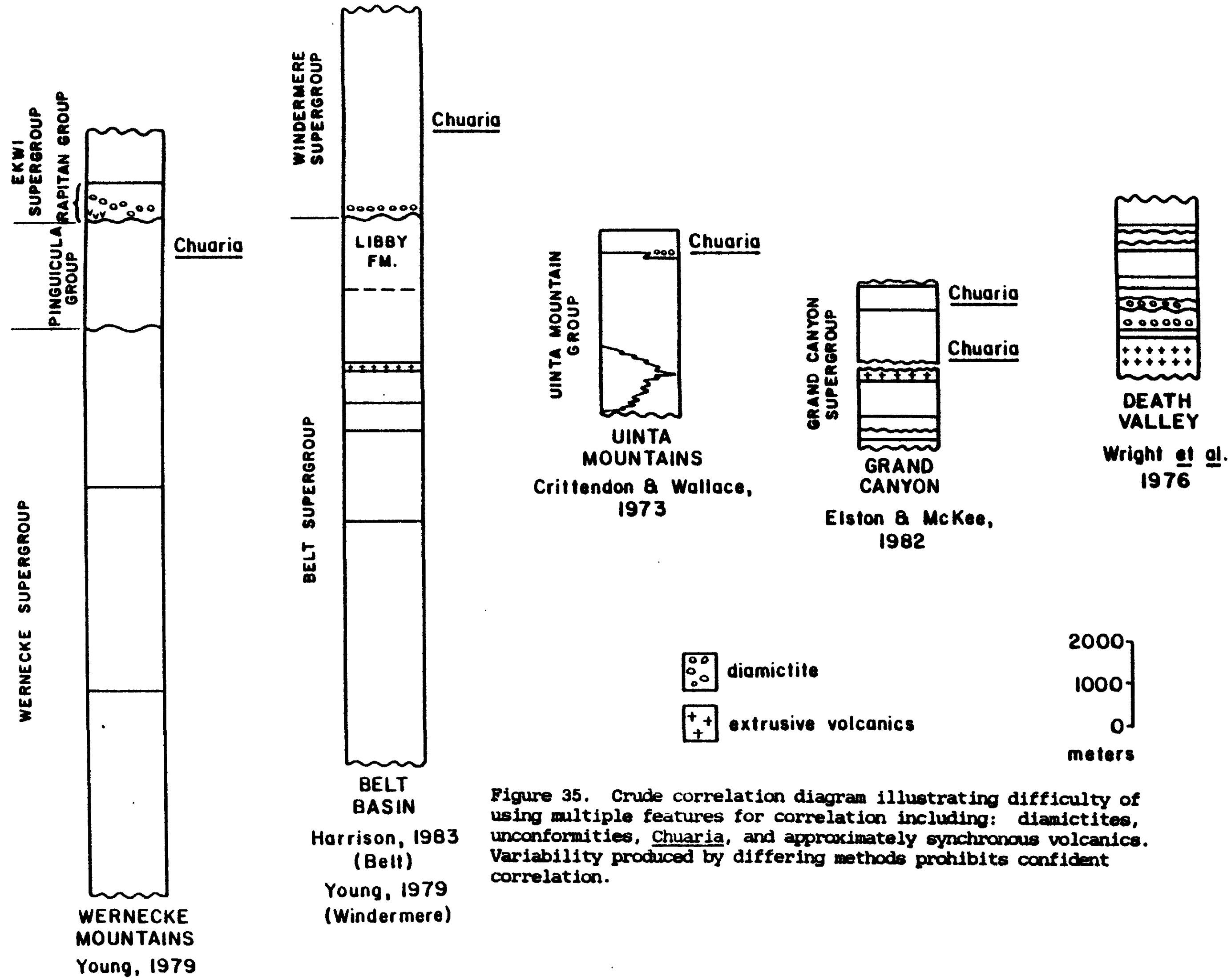


Figure 35. Crude correlation diagram illustrating difficulty of using multiple features for correlation including: diamictites, unconformities, Chuaria, and approximately synchronous volcanics. Variability produced by differing methods prohibits confident correlation.

useful in a broad sense, do not have the biostratigraphic resolution needed for resolving the age problems encountered in the Belt. A number of stromatolite groups ("genera" in a crude sense) have been reported and described from Belt rocks (Fenton and Fenton, 1931, 1937; Rezak, 1957; Horodyski, 1975, 1976a,b White, 1984). Some of the more apparently useful taxa in Proterozoic biostratigraphic correlations such as Conophyton and Baicalia, occur in the Belt Supergroup (Rezak, 1957; Horodyski, 1976a, 1983). These groups suggest at best a Middle to Late Riphean age (1350-680 Ma) for Belt rocks younger than the Altyn Formation (lower Belt) of Glacier National Park. There may be potential for stromatolite biostratigraphy in some parts of the Belt Supergroup; however, Horodyski (1983) suggests that the Baicalia-Conophyton cycles should not be used as a chronostratigraphic unit. Stromatolites in the Libby Formation are flat to mound-shaped structures, as opposed to the more distinctive columnar branching forms normally found useful in biostratigraphic work, thus no contribution to Beltian biostratigraphy appears to be possible from the Libby stromatolites.

Chuaria. Interpretation of the correlation chart for Proterozoic rocks of western North America (Fig. 35) suggests a number of reasons that support the potential for the occurrence of Chuaria in rocks of the Libby Formation. These reasons include: radiometric ages, stratigraphic position, and rock type. As previously mentioned, the most direct age for the Libby Formation is ~900 Ma based on Rb-Sr whole-rock analysis of argillite in the Garnet Range Formation. The Libby Formation occurs stratigraphically just below the unconformity separating the volcanic/diamictite assemblage commonly considered to be about 800-850 Ma (Stewart, 1972). In western North America, Chuaria has been found both above (Hector Formation, Red Pine Shale of the Uinta Mountain Group) and below (Little Dal Group) this diamictite horizon. There are no diamictites in the Grand Canyon sequence, but Chuaria occurs in several zones of the Chuar Group below a series of unconformities in the Sixtymile Formation. Based on available lithostratigraphic, biostratigraphic, and geochronologic data, it is possible that the diamictite assemblage was eroded from the top of the Grand Canyon Supergroup. The unconformities of the Sixtymile Formation may correlate to the unconformity that separates the Belt Supergroup from the Windermere Supergroup.

The reduced dark gray and greenish gray argillites and siltites of the Libby Formation are among the best rock types for preserving Chuaria. Most occurrences of Chuaria from around the world are in reduced, fine-grained sedimentary rocks (Table 4). Detailed tabulated data from Vidal (1976a) for the rock types in which Chuaria was preserved in the Visingso beds show that Chuaria can be preserved in a variety of sedimentary rock types, but the highest fossil yields come from black argillite. Vidal's (1976a) data also show that bedding type has a variable influence on the preservation of Chuaria, and finer grained sediments tend to favor Chuaria preservation, although some sandy samples have been productive. The dark gray and greenish-gray argillites and siltites of the Libby Formation therefore seem well suited for the preservation of Chuaria. The argument that these upper Belt rocks are too metamorphosed to preserve Chuaria is not valid, because reasonably good preservation of smaller acritarchs is found in these rocks.

Possible reasons for the absence of Chuaria could be a function of age of the rocks, depositional environment, or both. It is also possible that Chuaria exists in the Libby Formation, but was not observed. However, careful, bed-by-bed search for megafossils like Chuaria and Tawuia, and trace fossils was conducted during measurement and description of the Libby Formation sections to minimize the chances of Chuaria's absence being due to the lack of detailed examination. If the Libby Formation was deposited before the evolutionary appearance of Chuaria, then the answer is simple: The Libby Formation is older than the oldest Chuaria.

The age of the oldest Chuaria is in itself problematical. Most Chuaria occurrences cluster around 800-900 Ma, however, Hofmann and Jinbiao (1981) reported this fossil from the Jixian section of China in rocks yielding an age of 1800 Ma. The radiometric age interpretation of the Jixian section is suspect. Age dates of 1800 - 1000 Ma come from K-Ar analysis of phlogopite,

UNIT	LOCALITY	AGE	LITHOLOGY	DEPOSITIONAL ENVIRONMENT	REFERENCE
Scandinavia					
Visingso Beds	Sweden	0.75-0.80	sandstone, micacous silty shales	fluvial, shallow marine prodelta, intertidal	Vidal, 1974, 1976a
Nyborg, Lower Tillite, Tanafjord, Vadso Fms.	Finnmark	0.85-0.87	silty shale, sandy tillite matrix	Transgressive shallowing upward sequence--see Banks	Vidal, 1981
Hunnberg Fm.	Svalbard	0.80	micrite--not clear	shallow open coastal facies	Knoll, 1984
Veteranen Gp.	Svalbard	0.90	sandstone, siltstone, shale	shallow tidal	Knoll & Swett, 1985
Murchisonfjorden Supergroup	Svalbard		dark gray siltstones & shales	coastal mudflats	Knoll, 1982
Eleanore Bay Group	Greenland	0.80	black slates, black & gray shales		Vidal, 1976b
Thule Group	NW Greenland	0.65-0.80	shale		Vidal & Dawes, 1979
Hailuoto	Finland	0.58-0.80	siltstone		Tynni & Joakim, 1980
North America					
Chuar Group	Arizona	0.70-0.80	thin bedded dark gray & black shale	shallow water	Vidal & Ford, 1985, Ford & Breed, 1973a,b, Walcott, 1899

Uinta Mountain Group	Utah	0.82	thin bedded gray shale	shallow starved basin	Hofmann, 1977, Vidal & Ford, 1985
Little Dal Group	Northwest Territories	0.80-1.1	black shale siltstone	subtidal to peritadal platform	Hofmann & Aitken, 1979, Aitken, 1981, Hofmann, 1985
Hector Formation	Alberta	0.6-0.7			Gussow, 1973
<u>Africa</u>					
Atar	Mauritania	0.89	black shale		Amard, 1986
Nama Group	Nambia		gray & black shale	braided stream to quiet subtidal	Germs et al., 1986
<u>Antarctica</u>					
Robertson Bay Group	Antarctica	0.65	mudstone, muddy sandstone sandstone limestone		Cooper et al., 1982
<u>Asia</u>					
Bhima "series"	India	0.85	thinly laminated dark to light gray shale		Suresh & Sundara, 1983
Qingbaikou	China	0.65-0.80	green silty shale		Rulin & Lifu, 1985
Jixian	China	1.8			Hofmann & Jinbiao, 1981

Table 4. Comparison of ages, lithologies, and depositional environments for Churia from numerous localities.

Pb-Pb analysis of shale, and U-Pb concordia analysis of shale (Laboratory of Isotope Geology, Institute of Chemistry, Academy of China, 1977). Given the uncertainty of these dates, and questionable validity of the fossils (Awramik, pers. comm. 1987), it is probably safe to say that Chuaria does not occur in rocks older than approximately 1000-1100 Ma. It is equally possible that Chuaria's absence from the Libby Formation is a function of depositional environment although the rock types are well-suited to preserve Chuaria.

The marine vs. non-marine controversy for deposition of the Belt Supergroup has already been pointed out and discussed. Comparison of facies in the Libby Formation cannot as yet resolve this problem, although mixed calcitic/aragonitic ooids suggest a restricted marine setting. Many of the reports on the occurrence of Chuaria either state or imply marine conditions for the rocks in which Chuaria is preserved (Table 4). A more localized study of the microbiota of the 800 Ma Hunnberg Formation of Svalbard (Knoll, 1984) showed that Chuaria is restricted to facies reflecting open coastal environments and does not occur in lagoonal deposits. Similar studies of Paleozoic acritarchs show that restricted environments are characterized by low diversity biotas with small individuals (Staplin, 1961; Smith and Saunders, 1970). If the Belt basin were non-marine or perhaps even restricted marine, current knowledge of the distribution of Chuaria indicates it would not be found in these rocks. If the Belt basin opened to marine waters for the first time during deposition of the Garnet Range/upper Libby Formation, as previously suggested (Winston et al., 1984; McGroder, 1984, Winston, 1986), the high energy deposition that formed hummocky cross-stratified beds might not favor preservation of Chuaria. If, by coincidence the age of the upper Libby Formation were appropriate and the environment was indeed marine, then the best place to find Chuaria in the Belt Supergroup is in the silty partings of the hummocky cross- stratified facies and in the uppermost dark gray argillite unit (member G). These rocks generally do not exhibit good bedding plane exposures, so that if Chuaria does occur in member G, it could have been easily missed. If Chuaria is indeed not present in the Libby Formation, better information on the age and depositional environments are needed to explain the reason(s) for its absence.

Paleomagnetism

Paleomagnetic studies suggest that the entire Belt Supergroup may be older than 1200 Ma (Elston and Bressler, 1980; Elston, 1984; Fig 34). Although the magnetic reversals may provide time lines for correlation within the Belt Basin, the ages of the time lines depend primarily on correlation of the Pilcher Formation normal-polarity pole to a statistically identical normal-polarity pole for the Bass Limestone of the basal Grand Canyon Supergroup. Supportive correlation of the Missoula Group to the Pioneer Shale (Apache Group, central Arizona) and the Sibley Group (Lake Superior region), based on reversed-polarity intervals and similar pole positions, have been determined by Elston (1984). Additionally, these correlations are consistent with a polar wandering curve (Elston, 1984). Paleomagnetic and radiometric age determinations are in relatively close agreement for the lower Belt (~1450 Ma), but the results diverge widely in the younger Missoula Group (Fig. 34).

An unpublished correlation chart by Elston shows some consistency between radiometric, paleontologic and paleomagnetic ages. The Little Dal Group, Uinta Mountain Group, and Chuar Group are indicated as correlative by radiometric ages and the co-occurrence of Chuaria. Elston's data show a series of normal polarity intervals punctuated by intervals of no data for all three of these groups, and the paleomagnetism is consistent with already established correlations. However, the evidence provided by similarity of reversal stratigraphy among these three groups is not strong enough evidence to prove the correlations.

One problem with Elston's paleomagnetic work is that most of the data being used are from red bed sequences in which the magnetic carrier may be diagenetic rather than depositional. Numerous samples have been analyzed from non-red bed units such as the Prichard, Greyson, Helena, and Garnet Range Formations, but these rocks were reported as magnetically unstable (Elston and Bressler, 1980).

Clearly, paleomagnetism is producing reversal sequences and polar wandering curves in Proterozoic sequences. This approach provides great potential for correlation in these old rocks, but more complete coverage of these sequences and a better understanding of magnetic carriers is needed before Proterzoic paleomagnetic correlation can be accepted as reliable.

Discussion and Implications

Although paleomagnetically determined correlation and radiometrically derived ages for the lower Belt Supergroup are in agreement, synthesis of the available information on correlation of the Proterozoic in western North America produces serious conflict for the stratigraphic position and age of the upper Belt Supergroup. The ~900 Ma radiometrically derived age from the Garnet Range Formation (Obradovich and Peterman, 1968) differs drastically from the >1250 Ma age suggested by Elston and Bressler (1980). The incomplete information from lithostratigraphy and imperfect nature of our understanding of the biostratigraphy are, nevertheless, roughly consistent with this younger age. Preliminary attempts to correlate using unconformities separating major groups of rocks and other events such as volcanic extrusions in the Missoula Group and Grand Canyon Supergroup are also consistent with the younger age and such methods may hold future promise, but even if more confident correlations can be established, it is difficult at present to determine the degree of diachroneity in the erosional and volcanic events.

CONCLUSIONS

The results of this study show that a great deal of information can be gleaned from analysis of mudrocks on scales from regional facies analysis to microprobe investigation. Too often, mudrocks are ignored in stratigraphic, sedimentologic, and petrologic study because of their fine grain size. The fact that about 65% of the cratonic stratigraphic record consists of mudrock (Blatt, 1985) suggests that a valuable resource of data in the rock record is being overlooked. This resource may, at times preserve features that have been destroyed in coarser, more permeable rocks.

Eight major conclusions are derived from this study of the mudrocks that comprise the Proterozoic Libby Formation of Montana and Idaho:

1) The Libby Formation can be correlated to lithostratigraphic or lithofacies equivalents throughout the northern and central Belt basin over an area of 50,000 square km. The Libby Formation is subdivided into eight informal members. The lower four members correlate to the McNamara Formation in the southern and eastern Belt basin. The upper four members correlate to the Garnet Range Formation in the southern Belt basin and to the Libby Formation in the Whitefish Range. The basal quartzite of the Mount Nelson Formation (British Columbia) is equivalent to the Bonner Quartzite, and the rest of the Mount Nelson Formation is equivalent to the lower Libby Formation except for some quartzites high in the stratigraphic section at Toby Glacier and Bugaboo Creek, which may correlate to the upper Libby Formation.

2) The lower Libby Formation was deposited in a restricted marine setting above fair-weather wave base. Detailed facies comparisons between the Libby Formation and modern restricted marine, open marine, and Eocene lacustrine facies relationships show that the Libby Formation has no modern open marine sedimentologic analog. Both lacustrine and restricted marine facies relationships in settings that approximate conditions of Proterozoic sedimentation compare closely with facies relationships in the Libby Formation. The only modern example of a mixed calcitic/aragonitic ooid assemblage is in Baffin Bay, Texas, a restricted marine environment that experiences drastic salinity fluctuations. Unique minerals commonly associated

with lacustrine deposits are absent from the Libby Formation.

3) The upper Libby Formation was deposited below fair weather wave base and above storm wave base. The absence of abundant shallow water sedimentary structures in the upper Libby Formation suggests deposition below fair weather wave base. The thick, widespread hummocky cross-stratification throughout the central Belt basin in the Libby and Garnet Range Formations indicates deposition above storm wave base. Hummocky cross- stratified quartzite in the Garnet Range Formation was probably deposited above fair weather wave base.

4) Silt petrofacies analysis is successfully applicable to these Middle Proterozoic mudstones and has the potential to be a useful tool in mudrocks from other ages and areas where alteration of sandstones severely hampers petrofacies analysis. Blatt (1985) recently pointed out the potential of mudrock petrofacies analysis. This study is the first such application to Middle Proterozoic rocks. The distinct difference between the framework grain mineralogy in the lower and upper Libby Formation shows that even Middle Proterozoic mudrocks can adequately preserve detrital mineralogy for petrofacies analysis.

5) Low permeabilities in these Middle Proterozoic mudrocks may have preserved an assemblage of now calcite ooids, interpreted as an original mixture of calcite and aragonite ooids, sufficiently well that geochemical analysis may allow an approximation of Proterozoic sea water composition or at least early diagenetic water composition.

6) The distinct trace element composition of different cement types in the ooids in the Libby Formation may show the chemical evolution of pore water in these Middle Proterozoic carbonates. Preliminary analysis shows an increase in iron as cementation progressed in these rocks. Increased iron content in late diagenetic phases has been documented in both clastic and carbonate rocks (Boles, 1978; Heckel, 1983).

7) Although the acritarchs found in the Libby Formation can not yet be successfully be used in biostratigraphic analysis, it is significant that they can be reasonably well-preserved even in lower greenschist facies rocks.

8) Careful stratigraphic and depositional facies analysis in predominantly mudrock sequences can be successfully used in the interpretation of basinal tectonic shifts.

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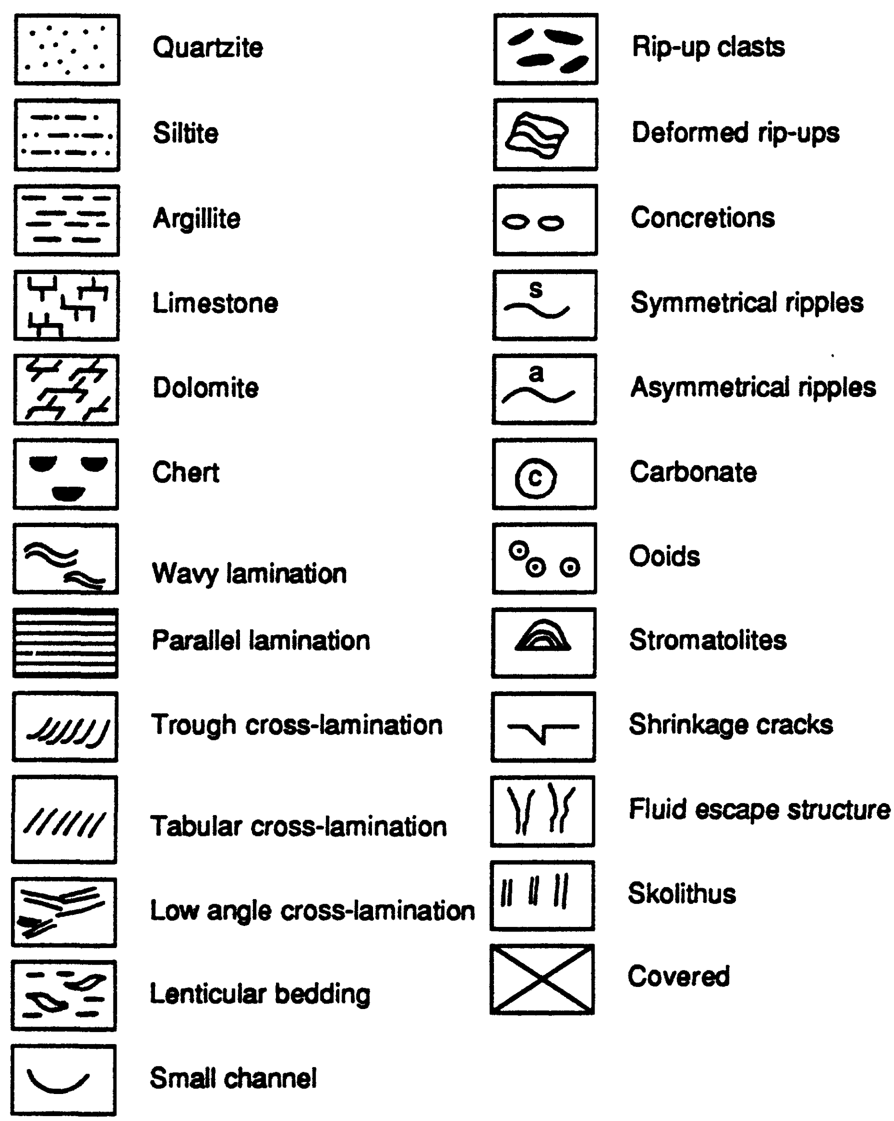
APPENDIX 1

Detailed stratigraphic columns of the Libby Formation. Localities shown in Appendix 2. Key to lithologic symbols and sedimentary structures shown below. Color key: dg = dark gray, gg = greenish gray, pr = purplish red.

KEY TO APPENDIX I



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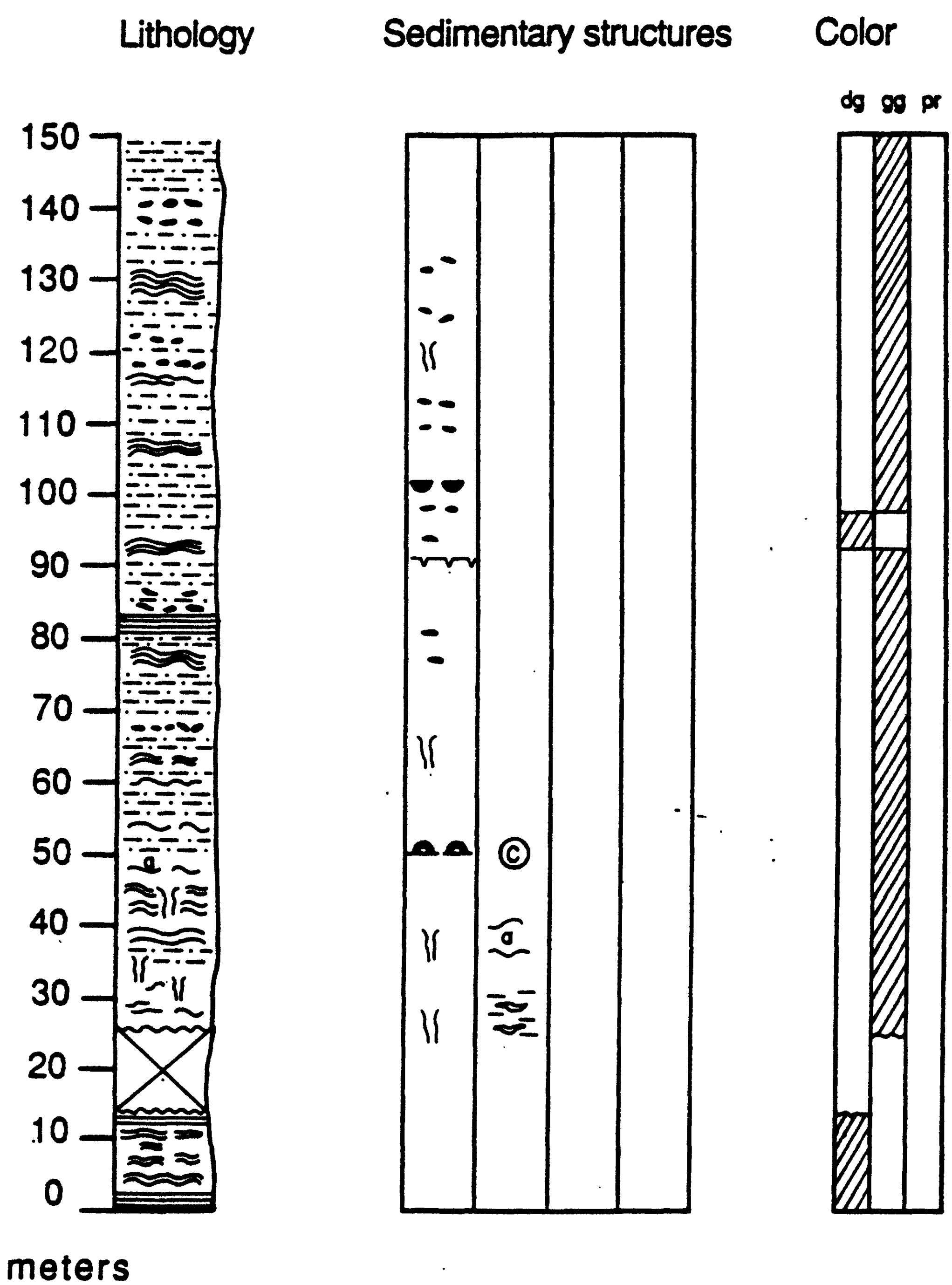
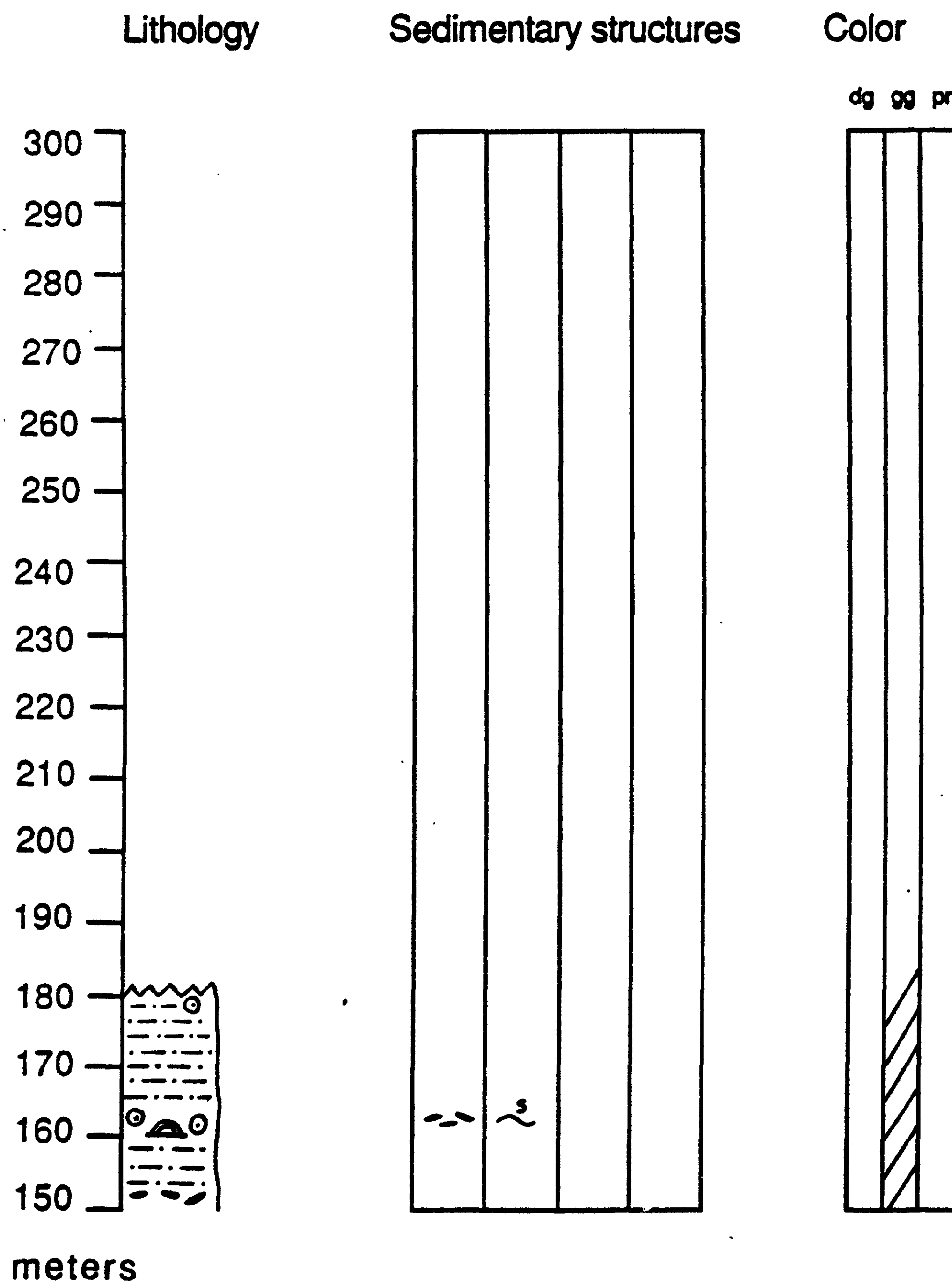
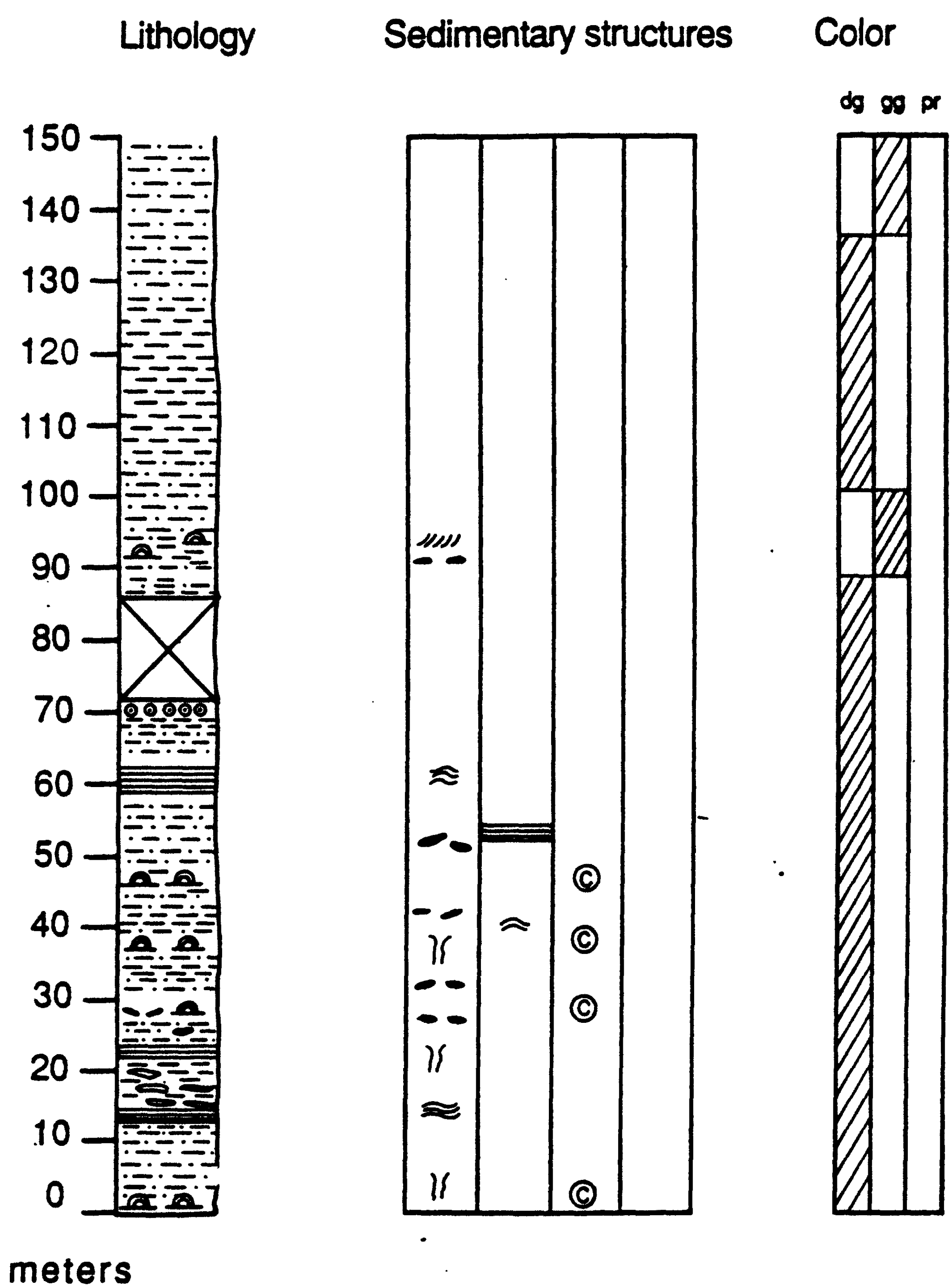

Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

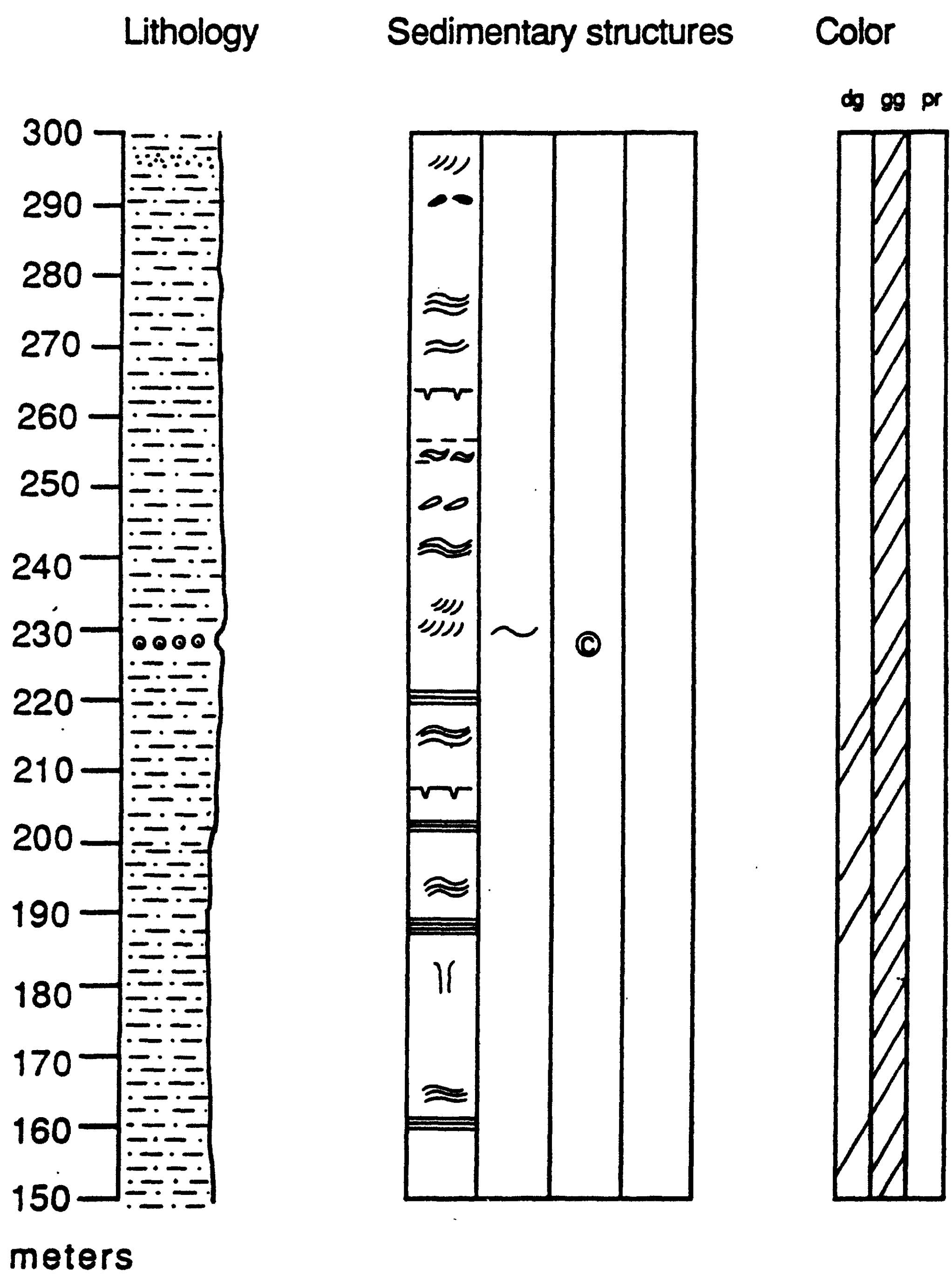
SECTION # FB


Lithology
Sedimentary structures
Color
dg gg pr
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
meters

SECTION # FC

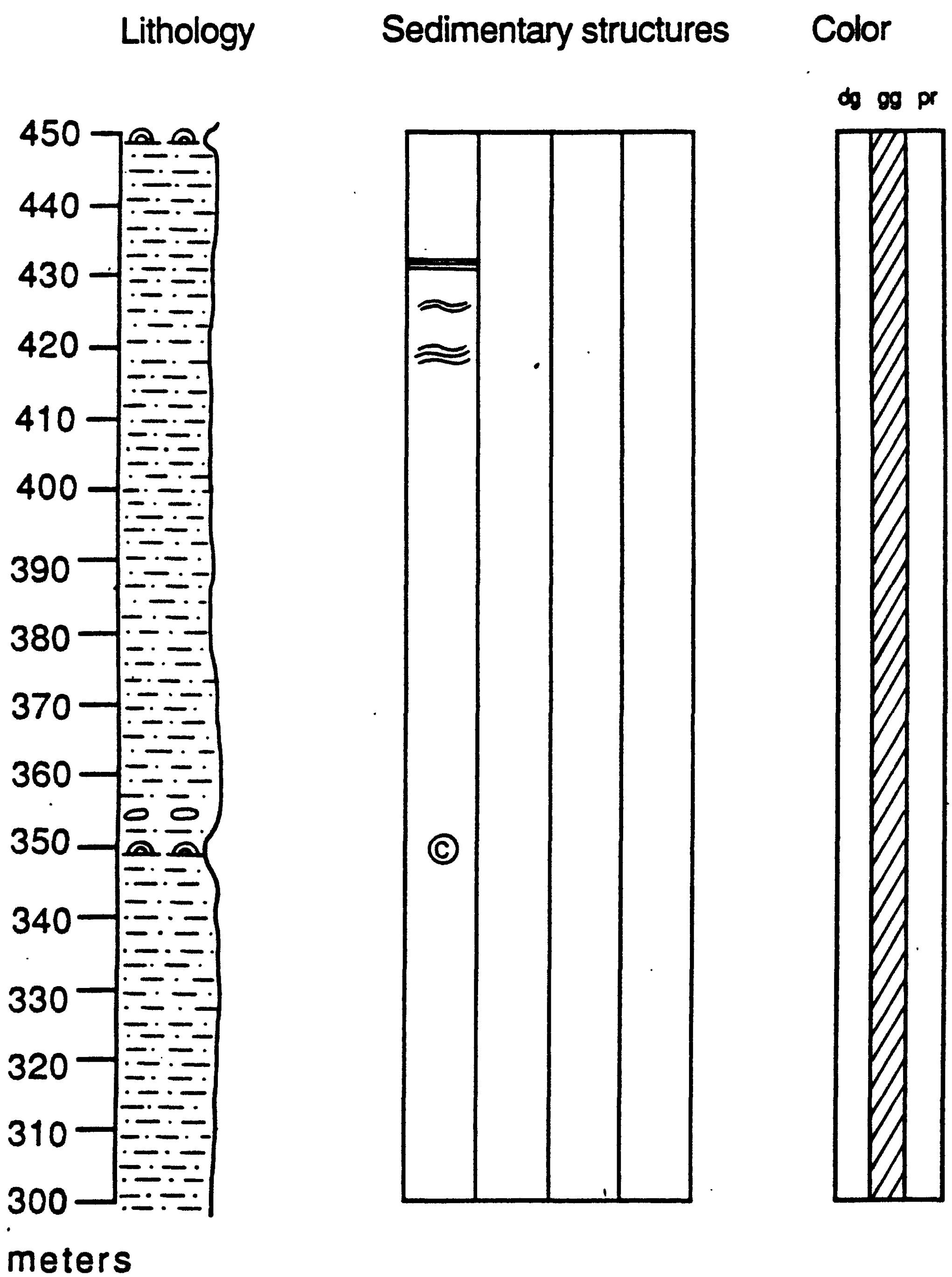

Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

SECTION # FC



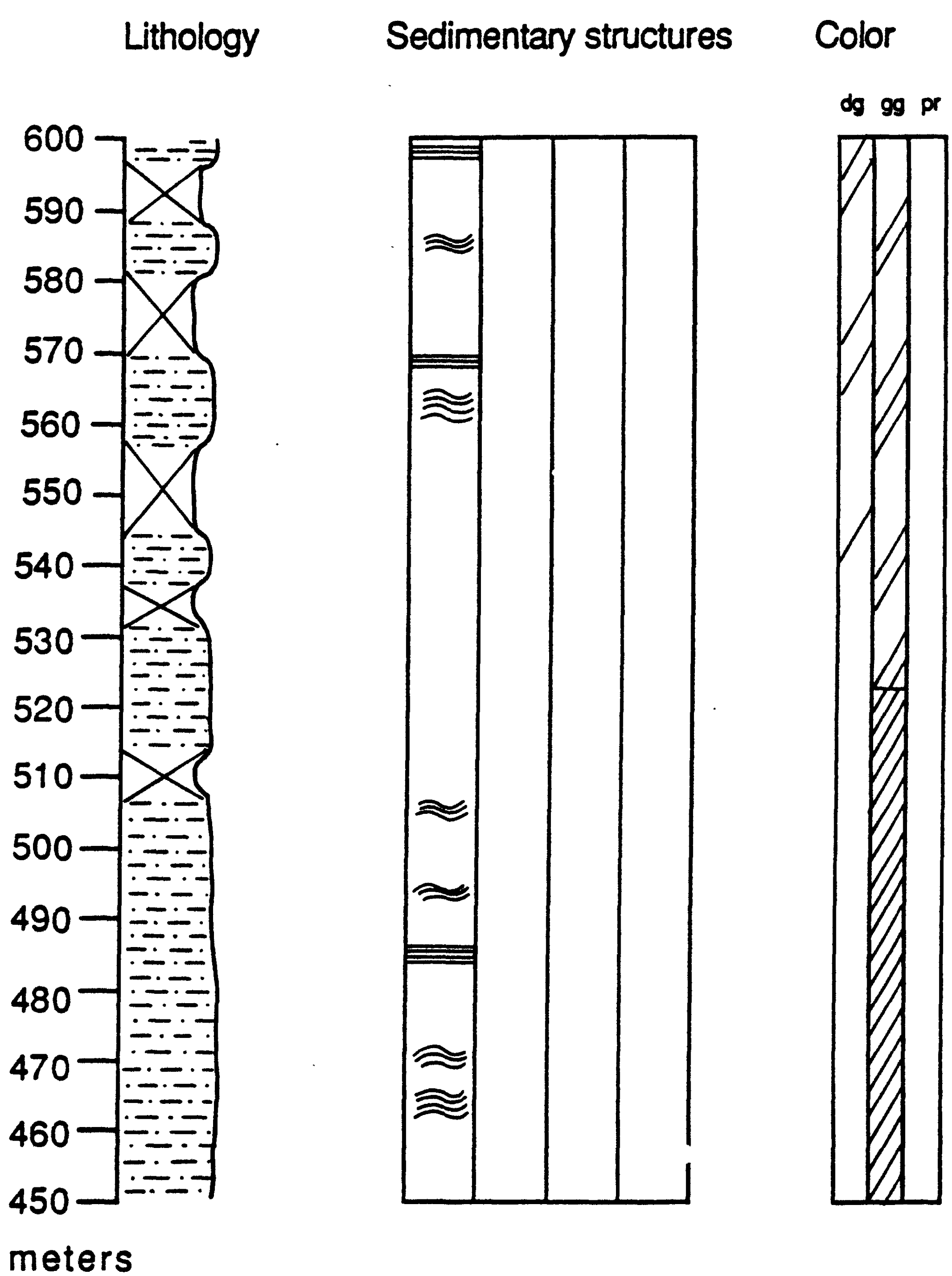
Lithology
Sedimentary structures
Color
dg gg pr
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
meters

SECTION # FC

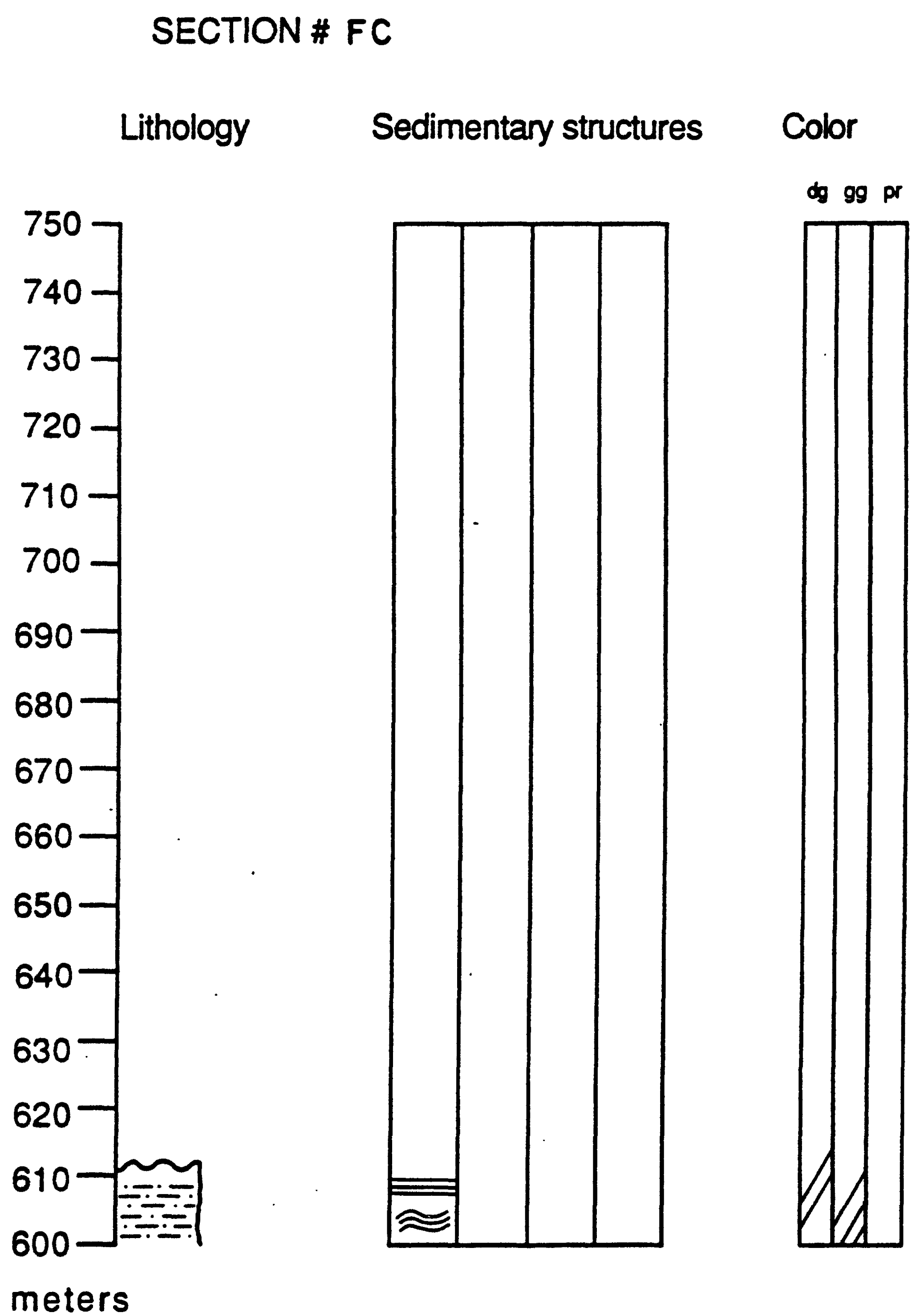


Lithology
Sedimentary structures
Color
dg gg pr
450
440
430
420
410
400
390
380
370
360
350
340
330
320
310
300
meters

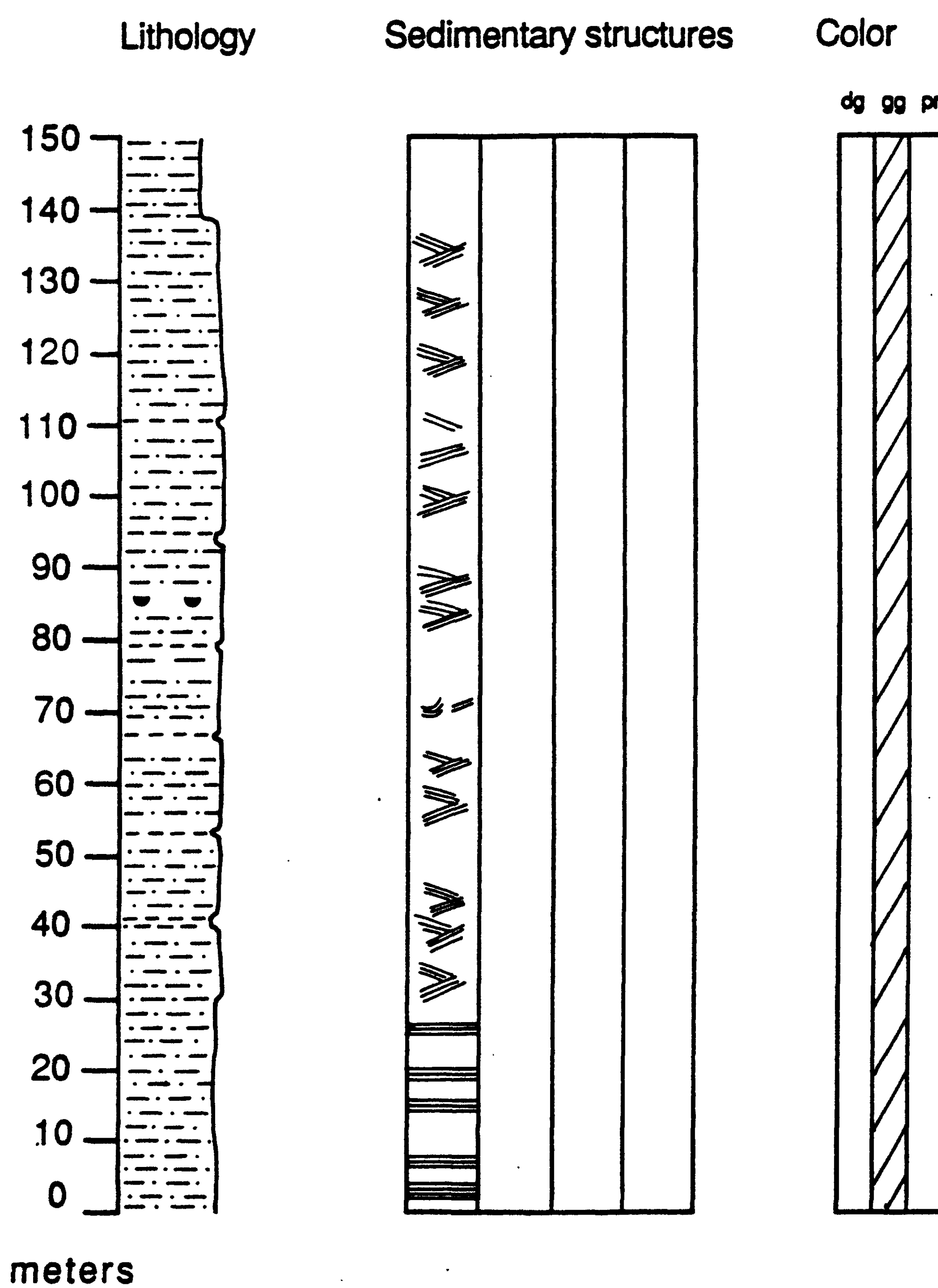
SECTION # FC



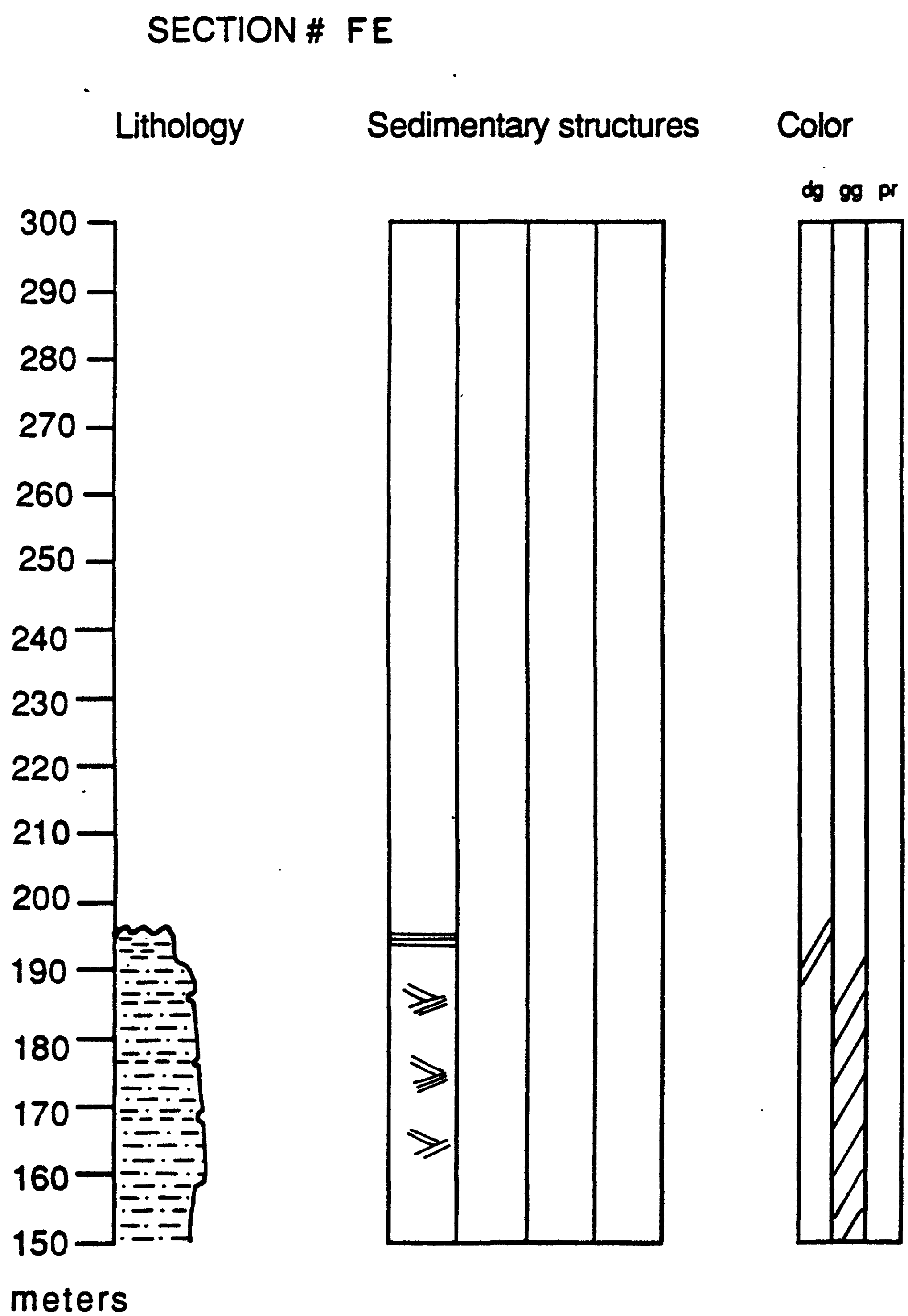
Lithology
Sedimentary structures
Color
dg gg pr
600
590
580
570
560
550
540
530
520
510
500
490
480
470
460
450
meters


SECTION # FC
Lithology
Sedimentary structures
Color
dg gg pr
750
740
730
720
710
700
690
680
670
660
650
640
630
620
610
600
meters

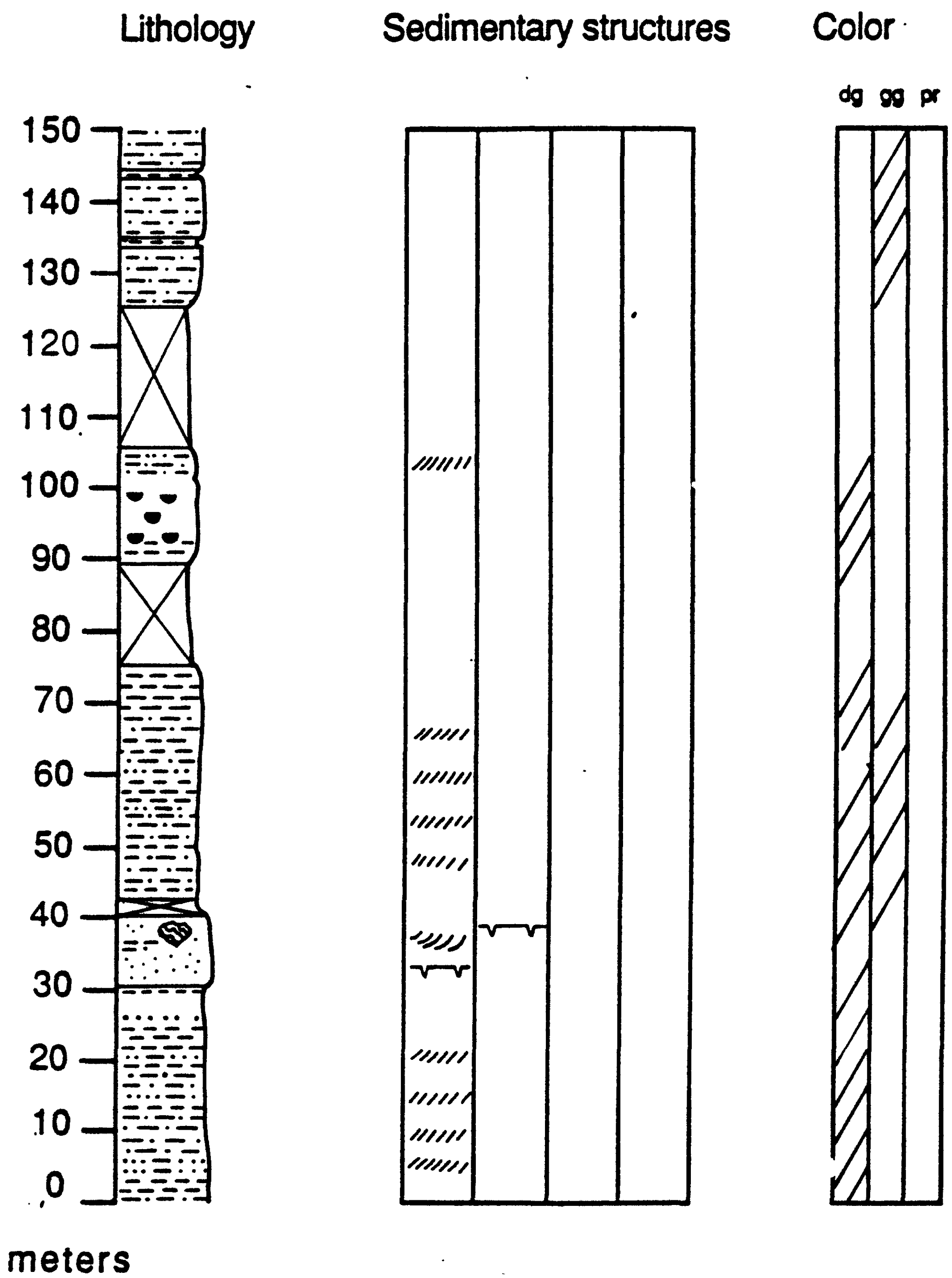
SECTION # FE



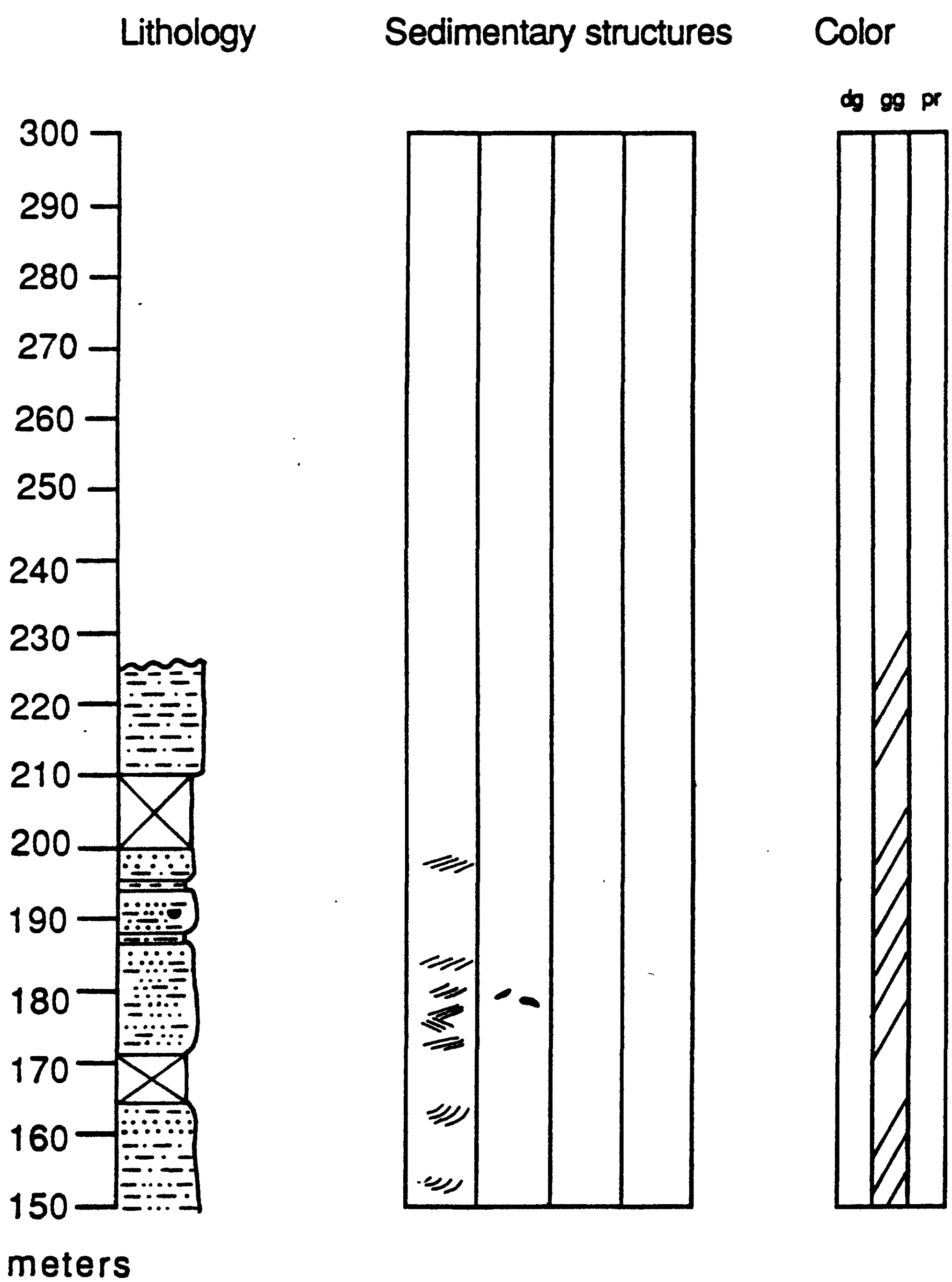
Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters


SECTION # FE
Lithology
Sedimentary structures
Color
dg gg pr
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
meters

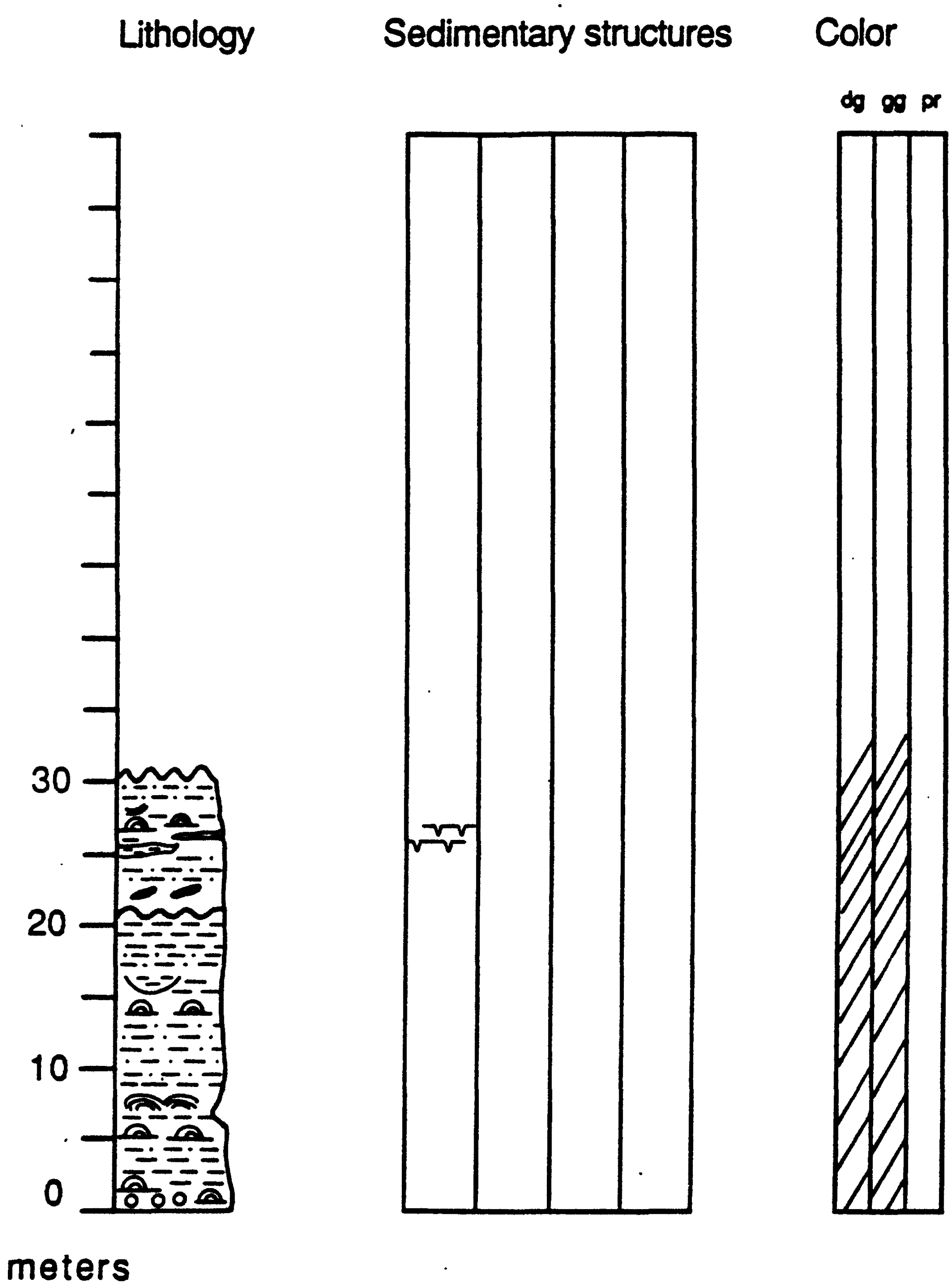
SECTION # FG


Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

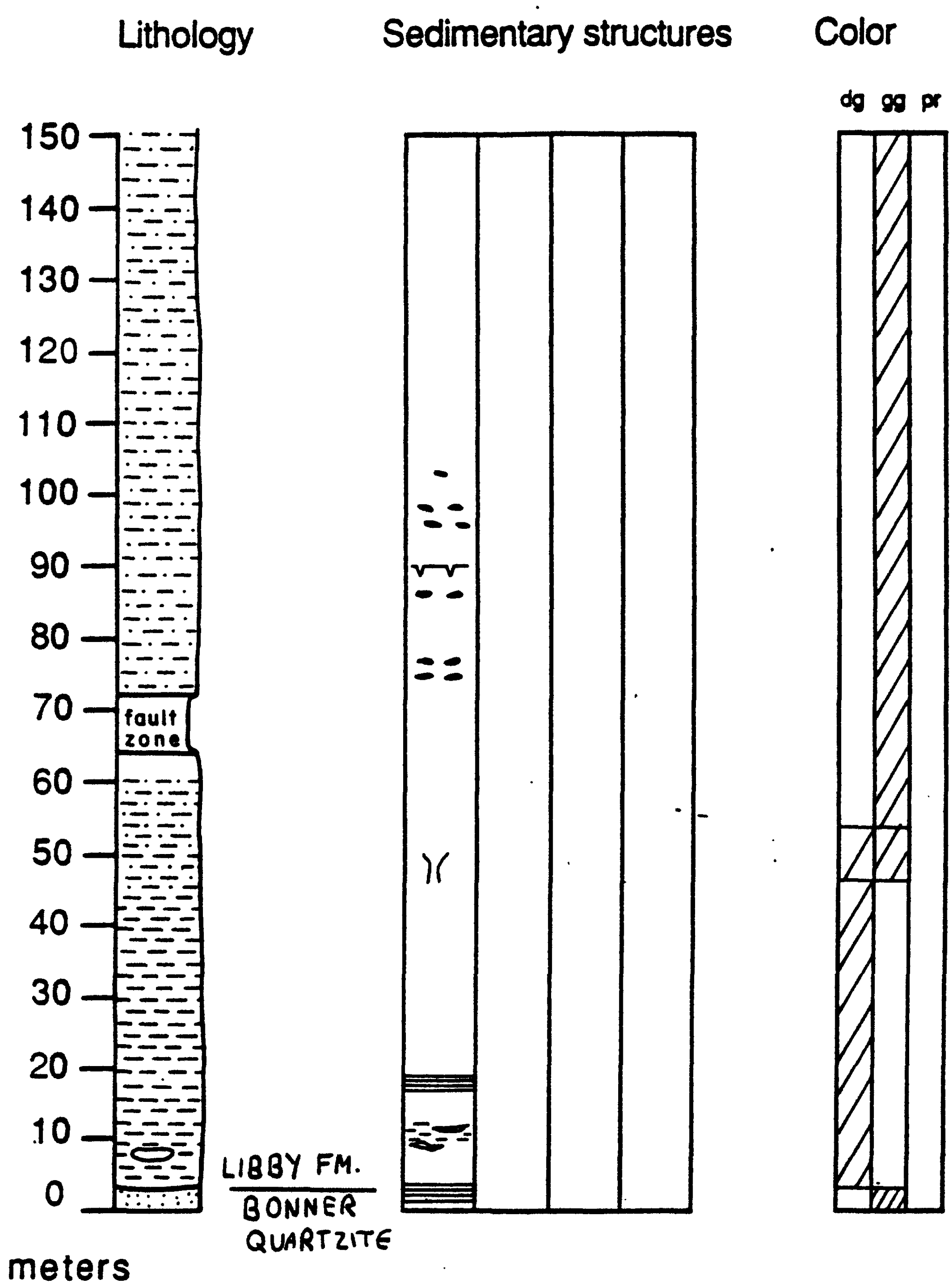
SECTION # FG


Lithology
Sedimentary structures
Color
dg gg pr
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
meters

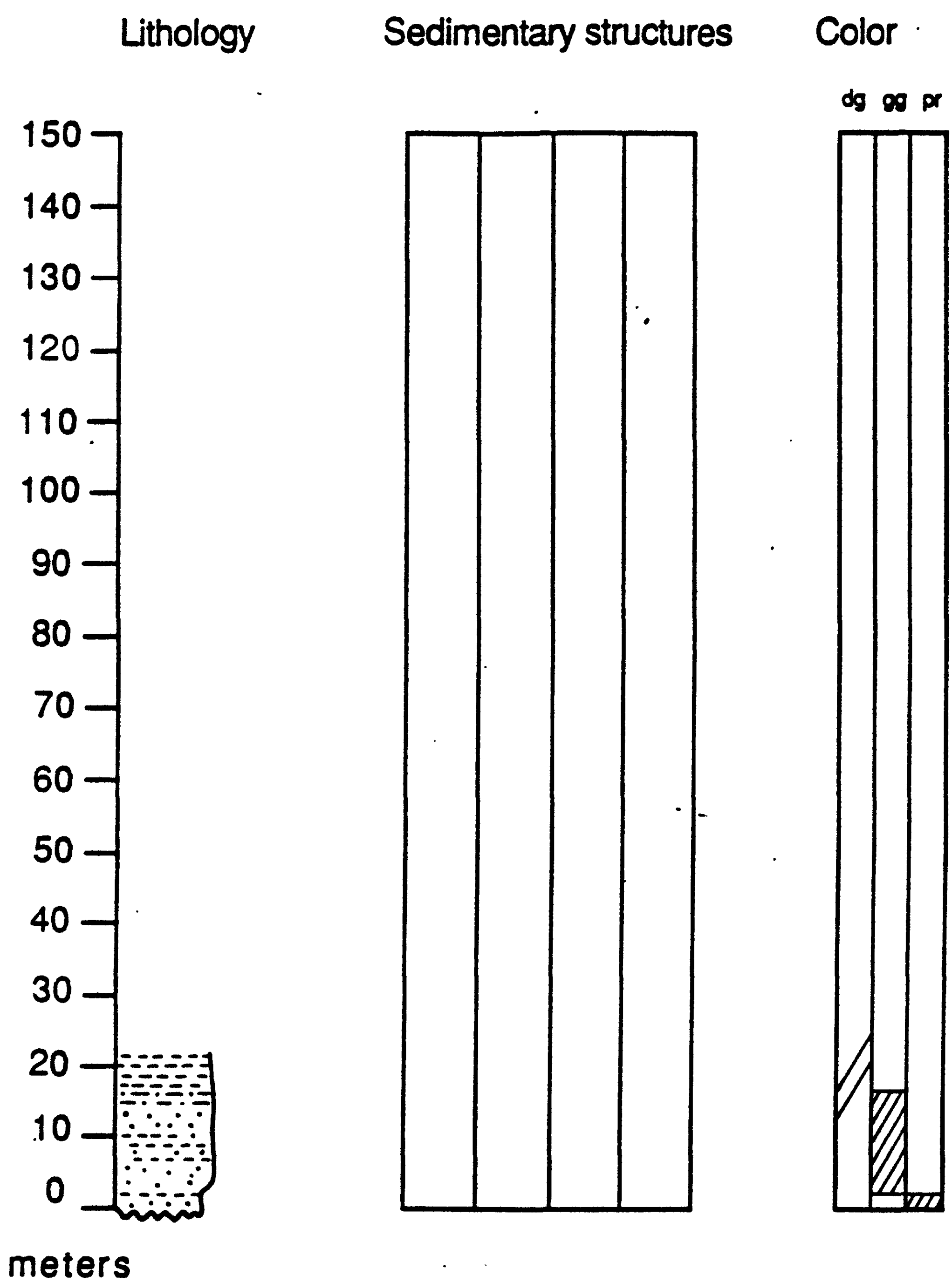
SECTION # 2-MP-25



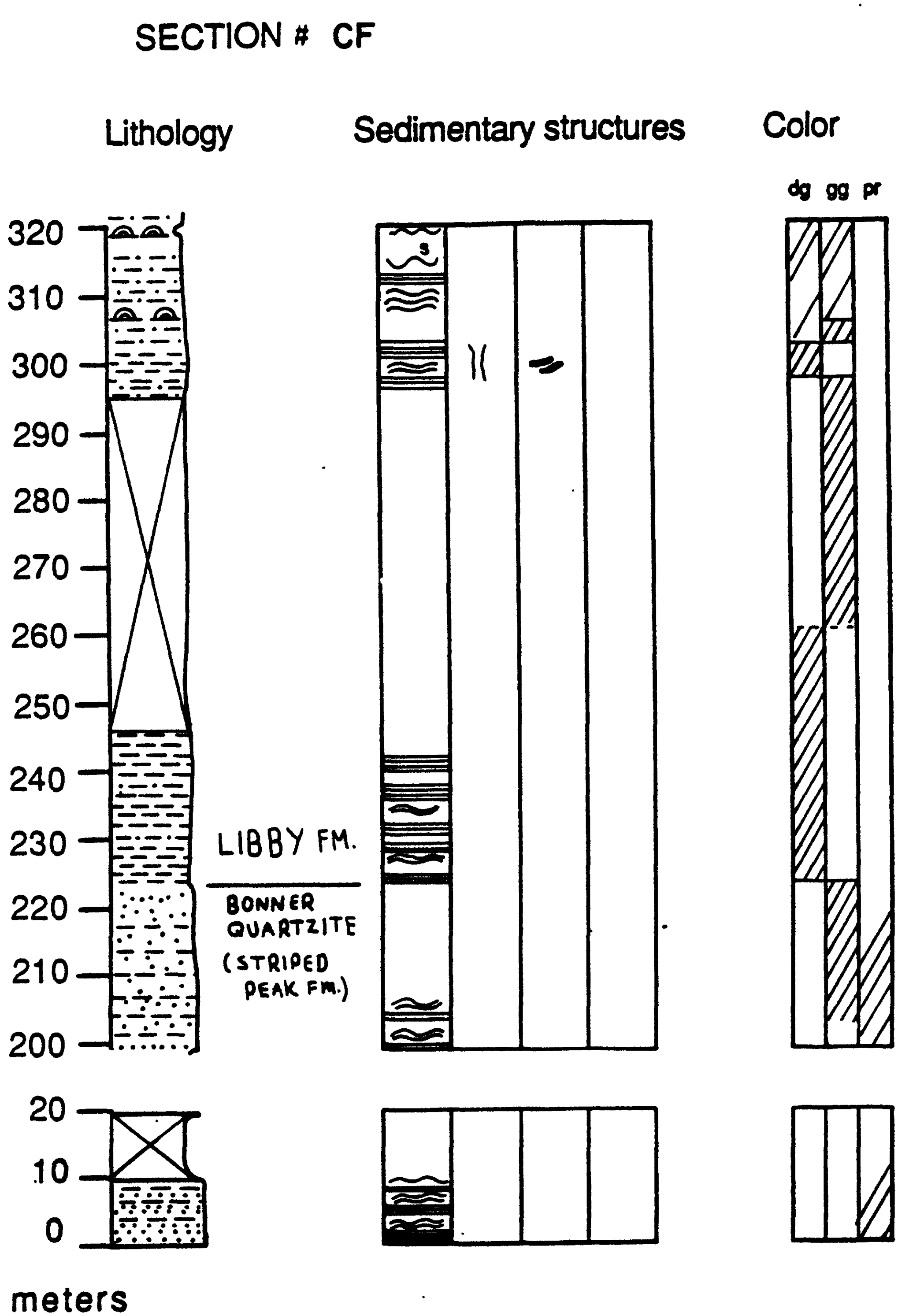
SECTION # 2A

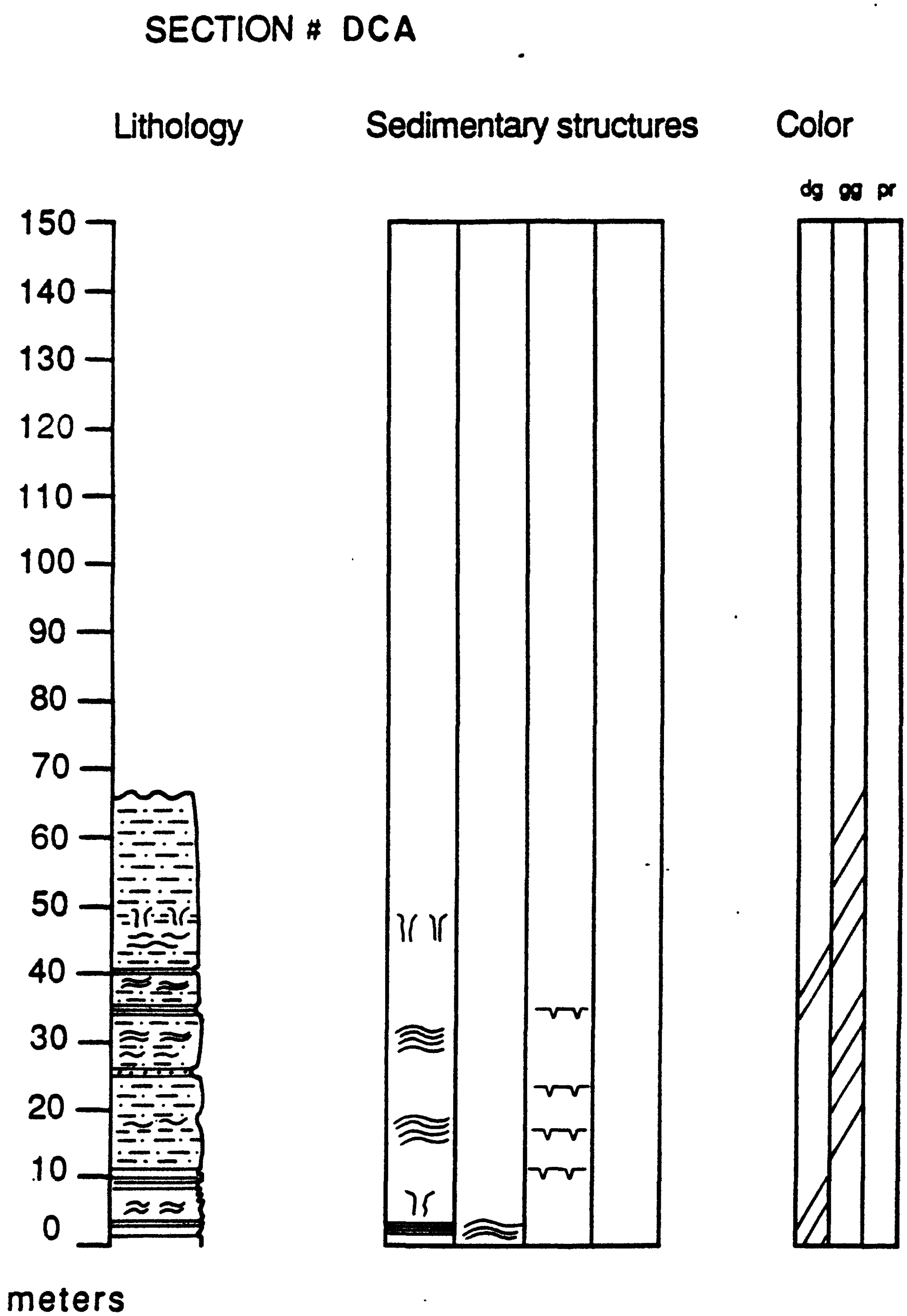

Lithology
Sedimentary structures
Color
dg
gg
pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters
fault zone
LIBBY FM.
BONNER QUARTZITE

SECTION # 2AA

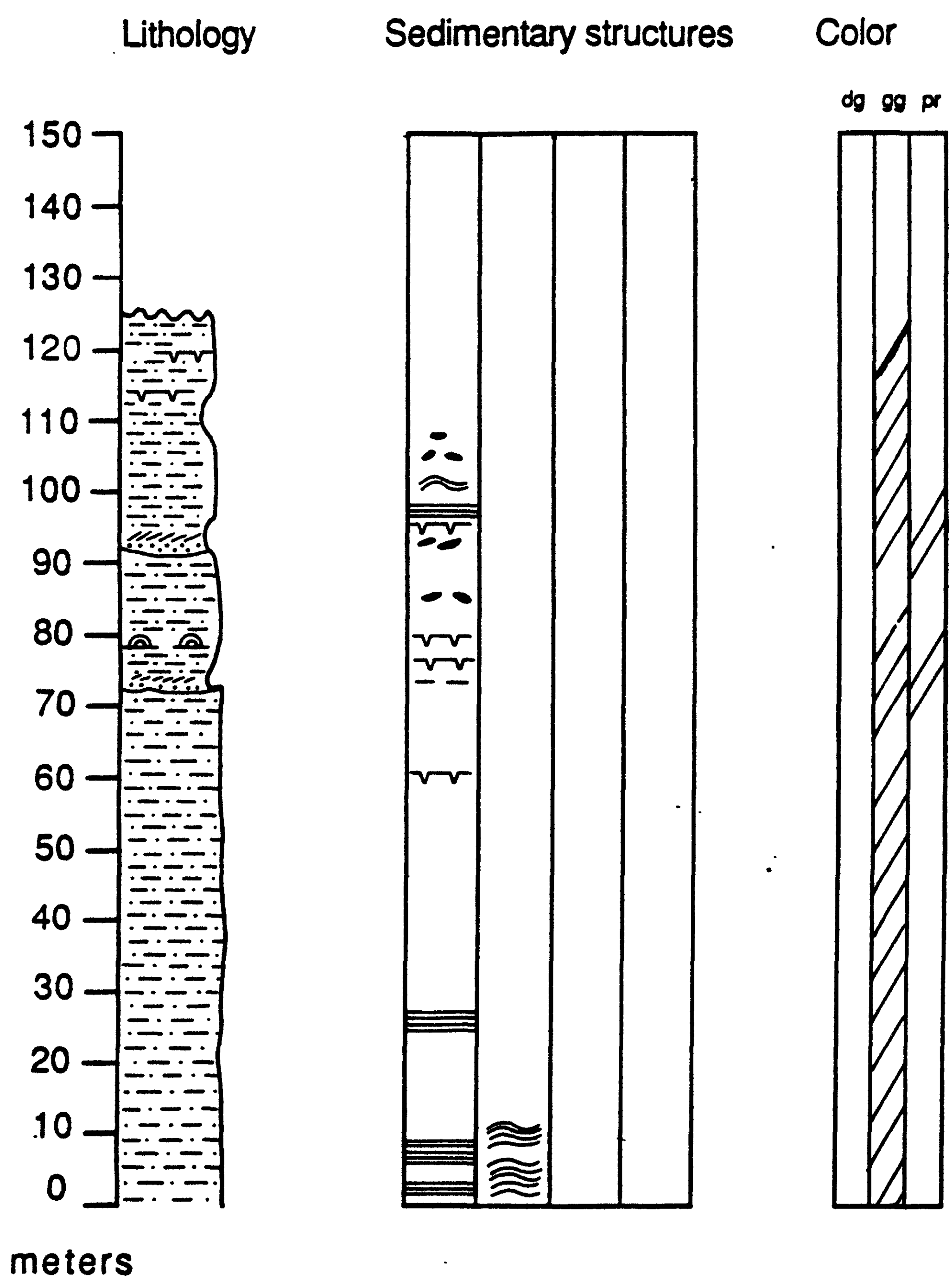


Lithology
Sedimentary structures
Color
dg
gg
pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

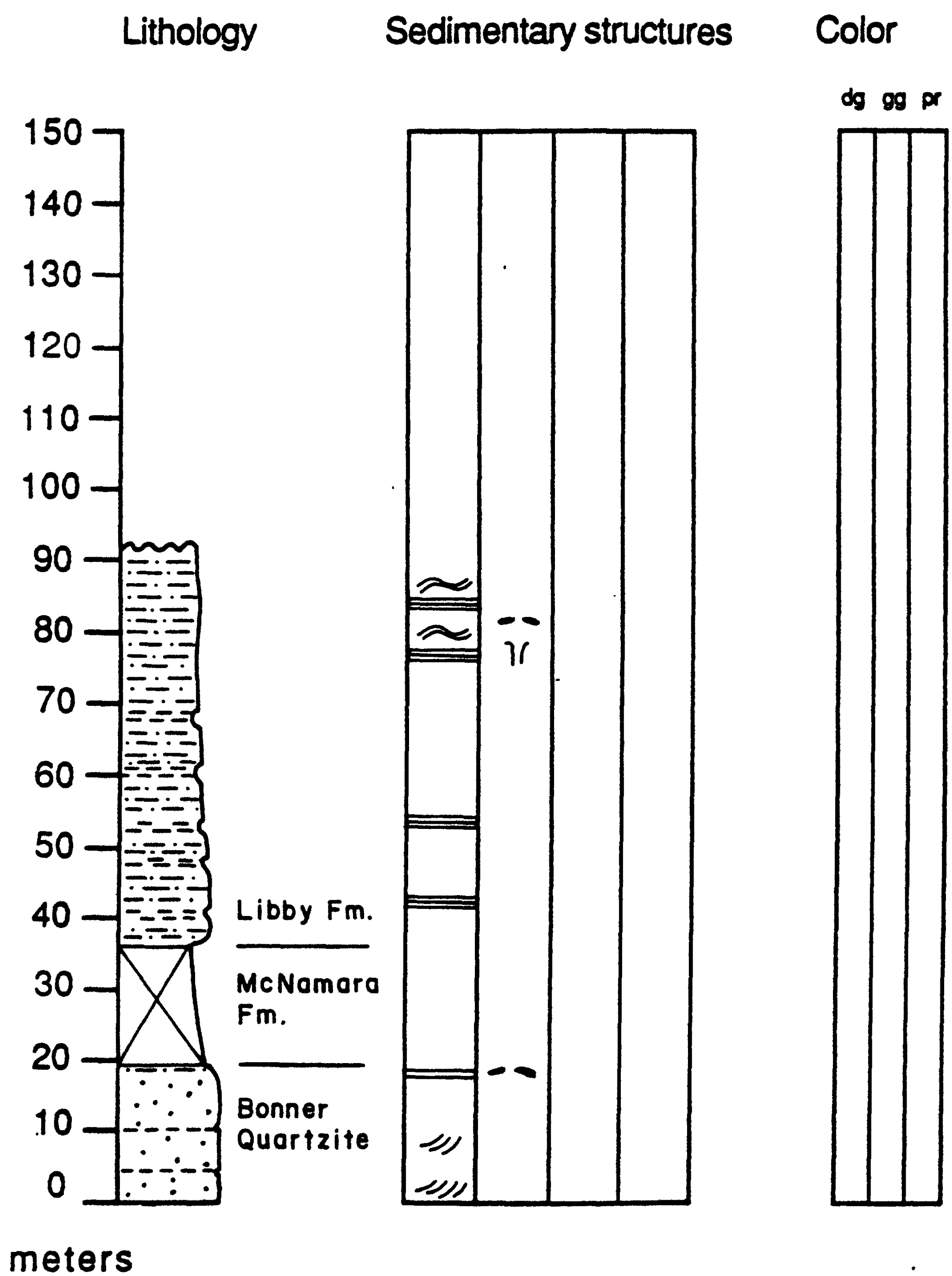

SECTION # CF
Lithology
Sedimentary structures
Color
dg gg pr
320
310
300
290
280
270
260
250
240
230
220
210
200
20
10
0
meters
S
LIBBY FM.
BONNER QUARTZITE (STRIPED PEAK FM.)


SECTION # DCA
Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

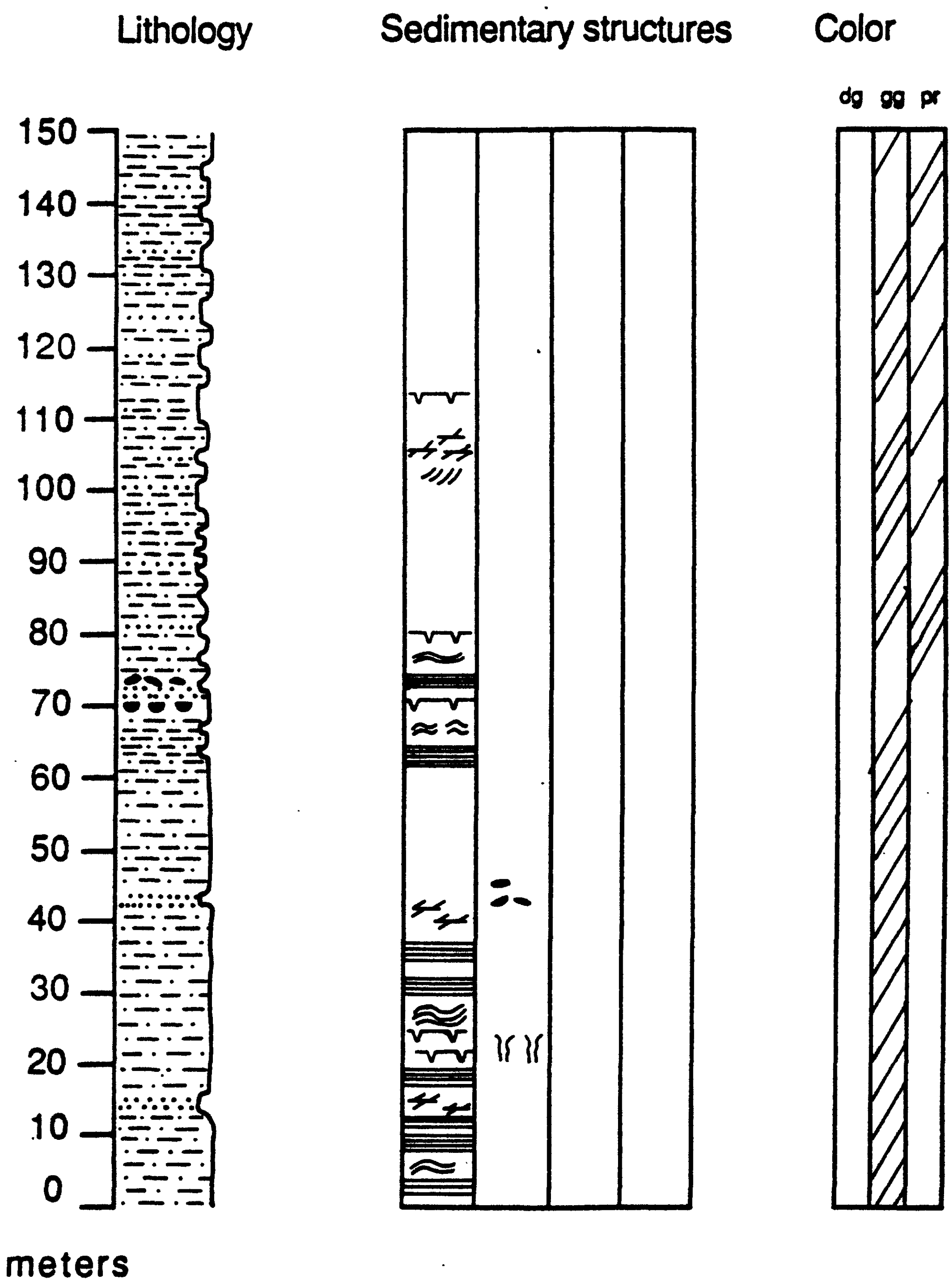
SECTION # DCB


Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

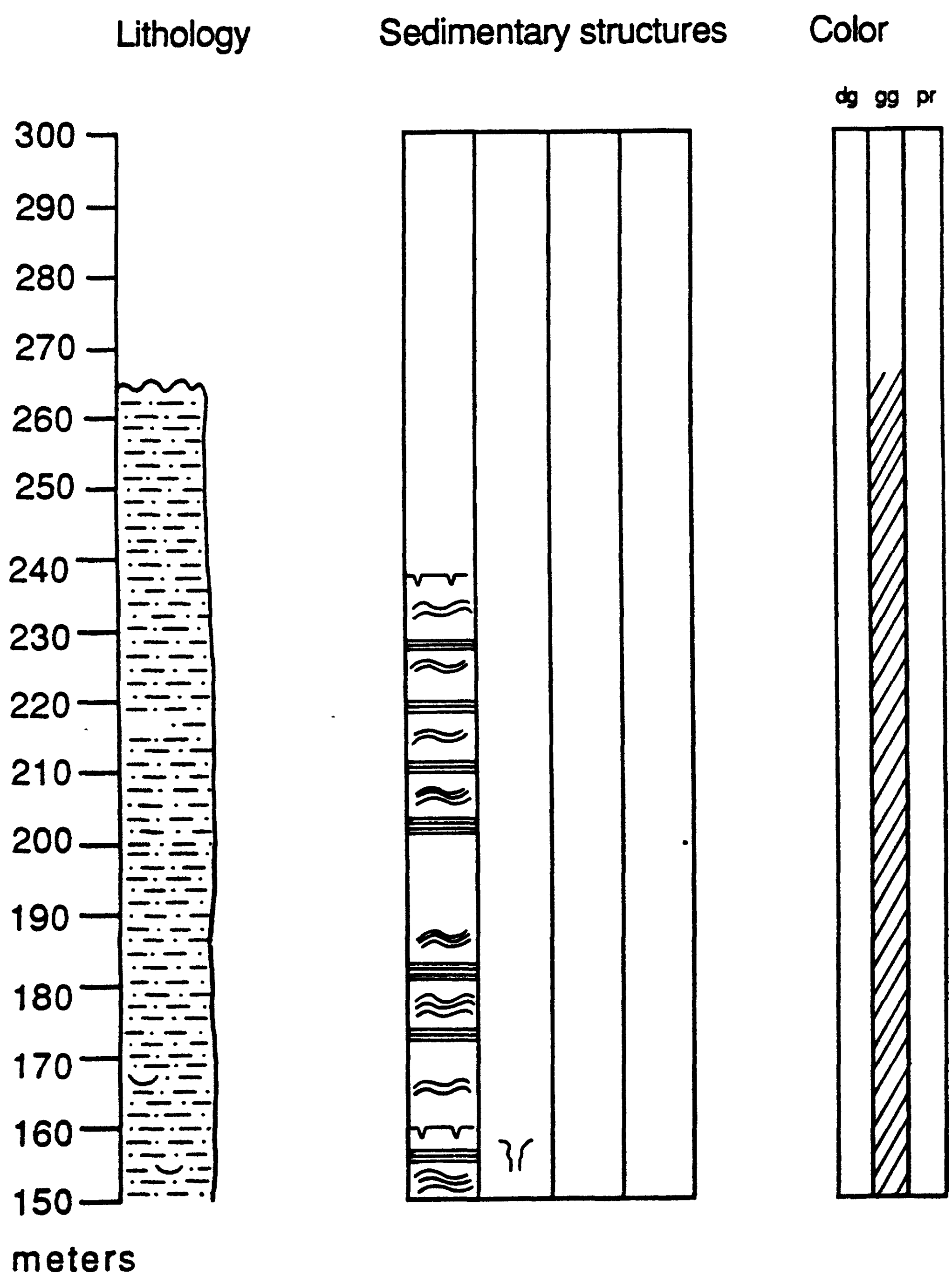
SECTION # DMA


Lithology
Sedimentary structures
Color
dg gg pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters
Libby Fm.
McNamara Fm.
Bonner Quartzite

SECTION # DMB

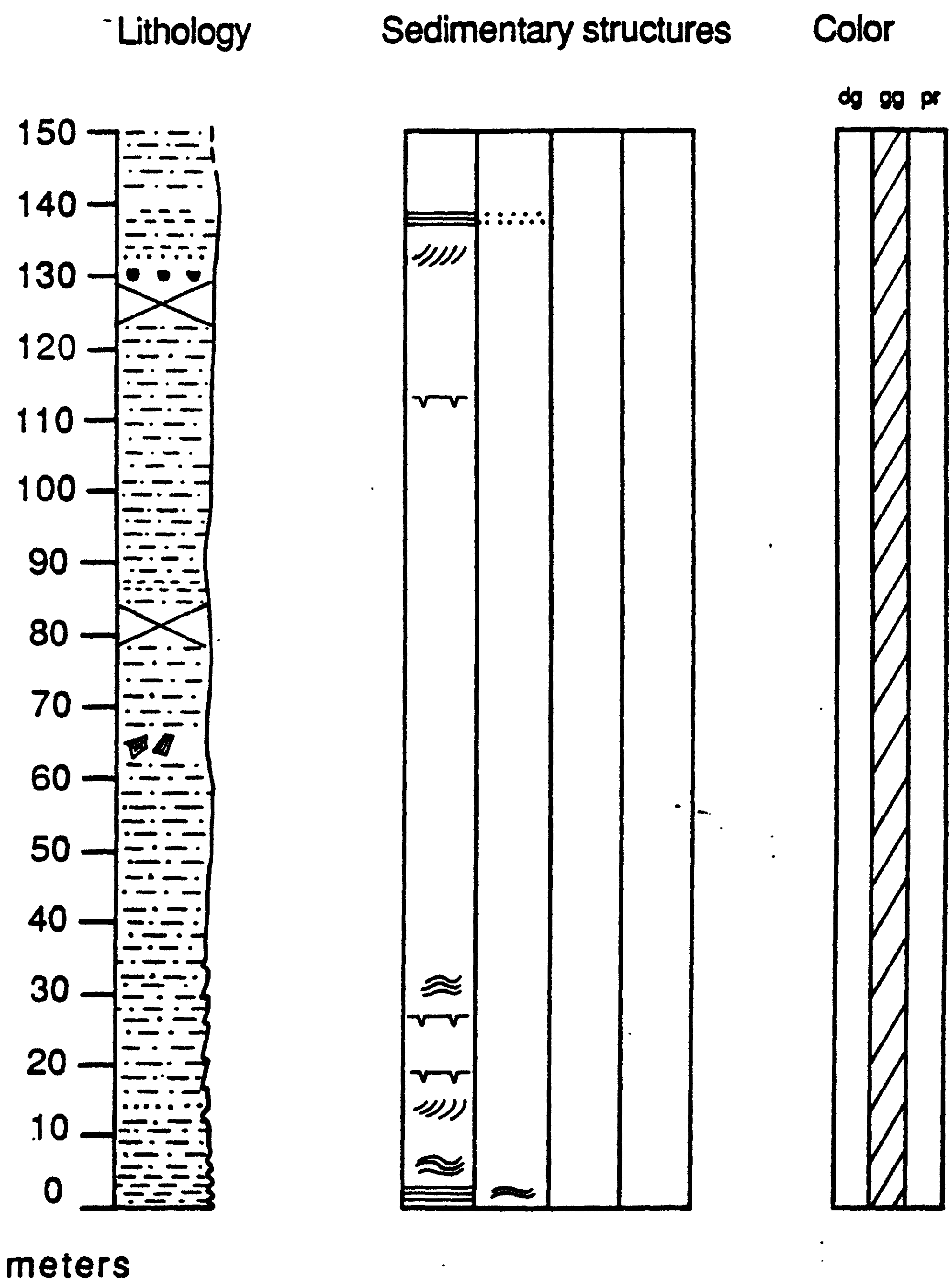

Lithology
Sedimentary structures
Color
dg
gg
pr
150
140
130
120
110
100
90
80
70
60
50
40
30
20
10
0
meters

SECTION # DMB

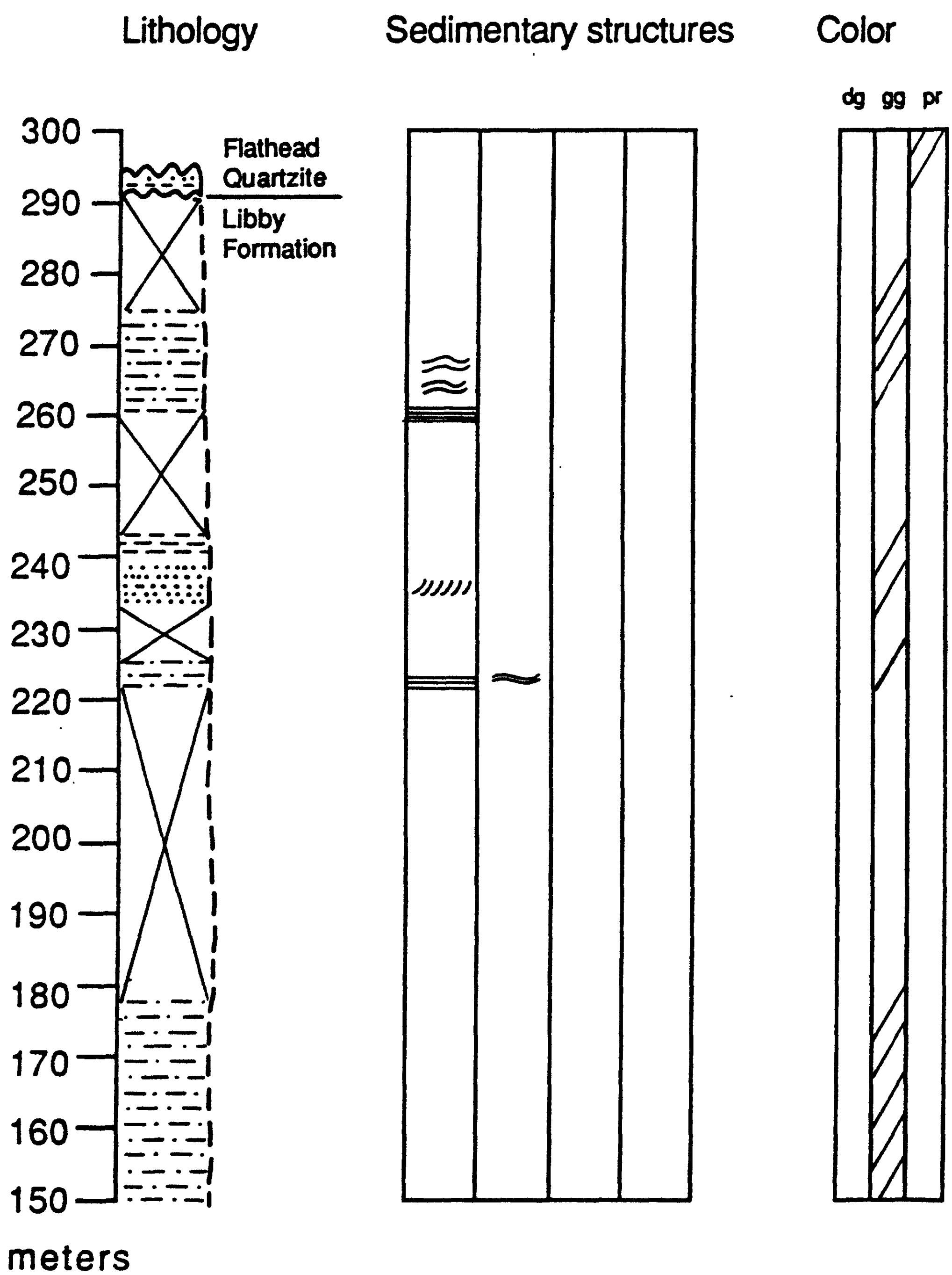


Lithology
Sedimentary structures
Color
dg gg pr
300
290
280
270
260
250
240
230
220
210
200
190
180
170
160
150
meters

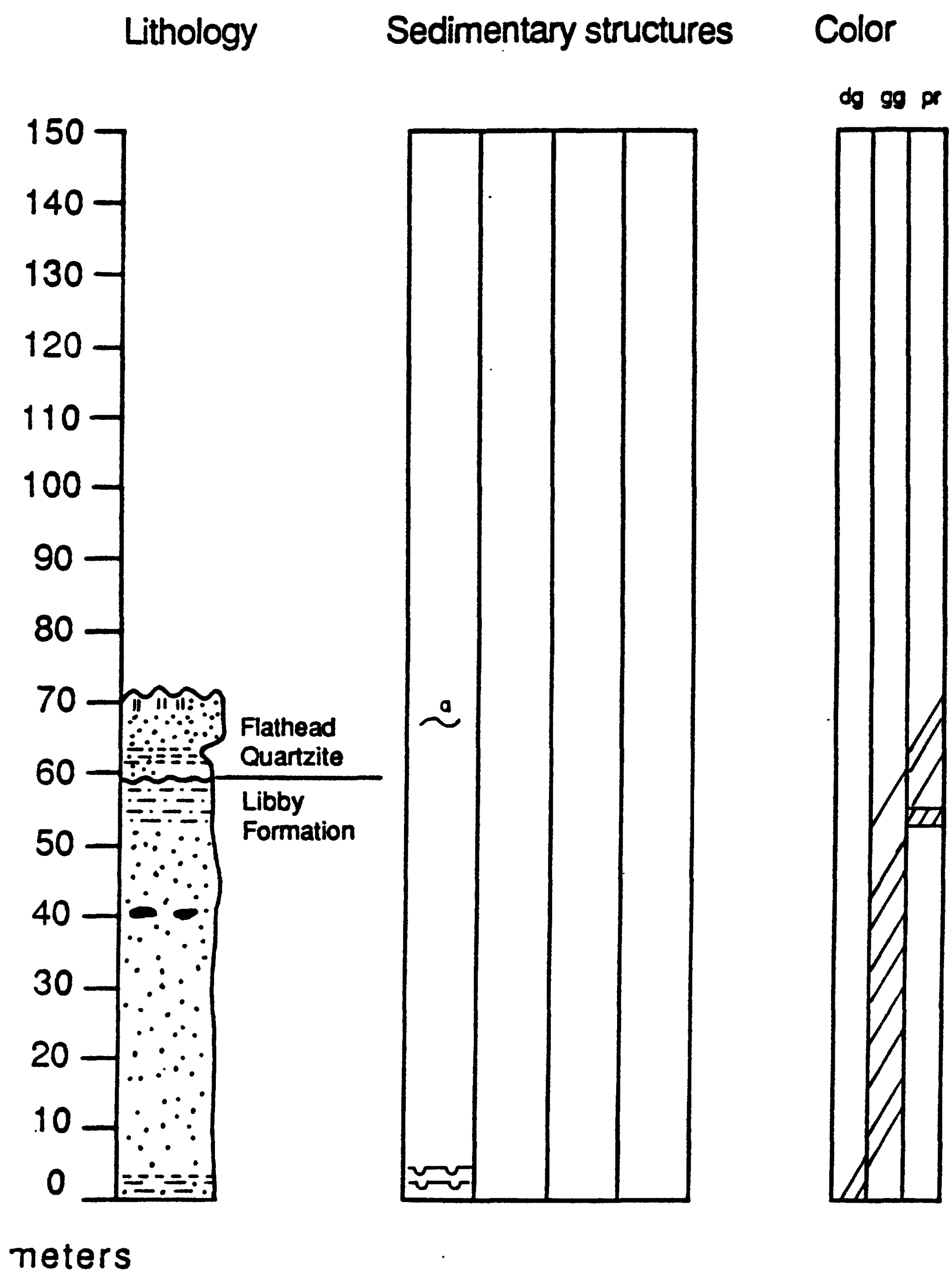
SECTION # SQRA



SECTION # SQRA

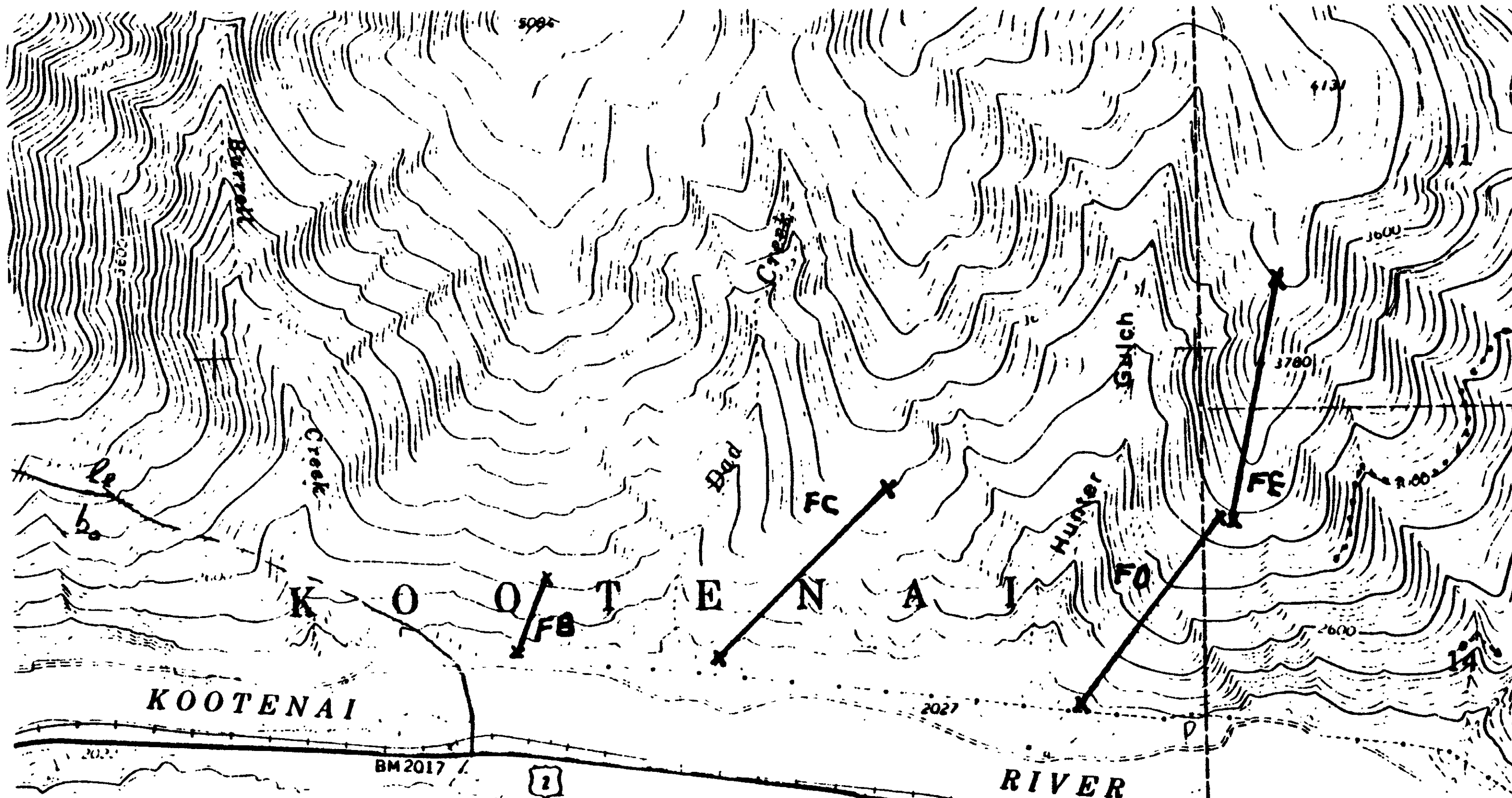


SECTION # SRQB

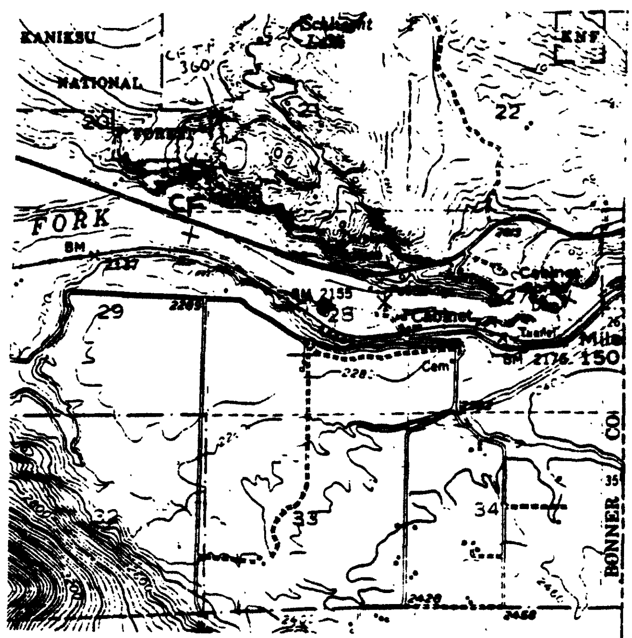


APPENDIX 2

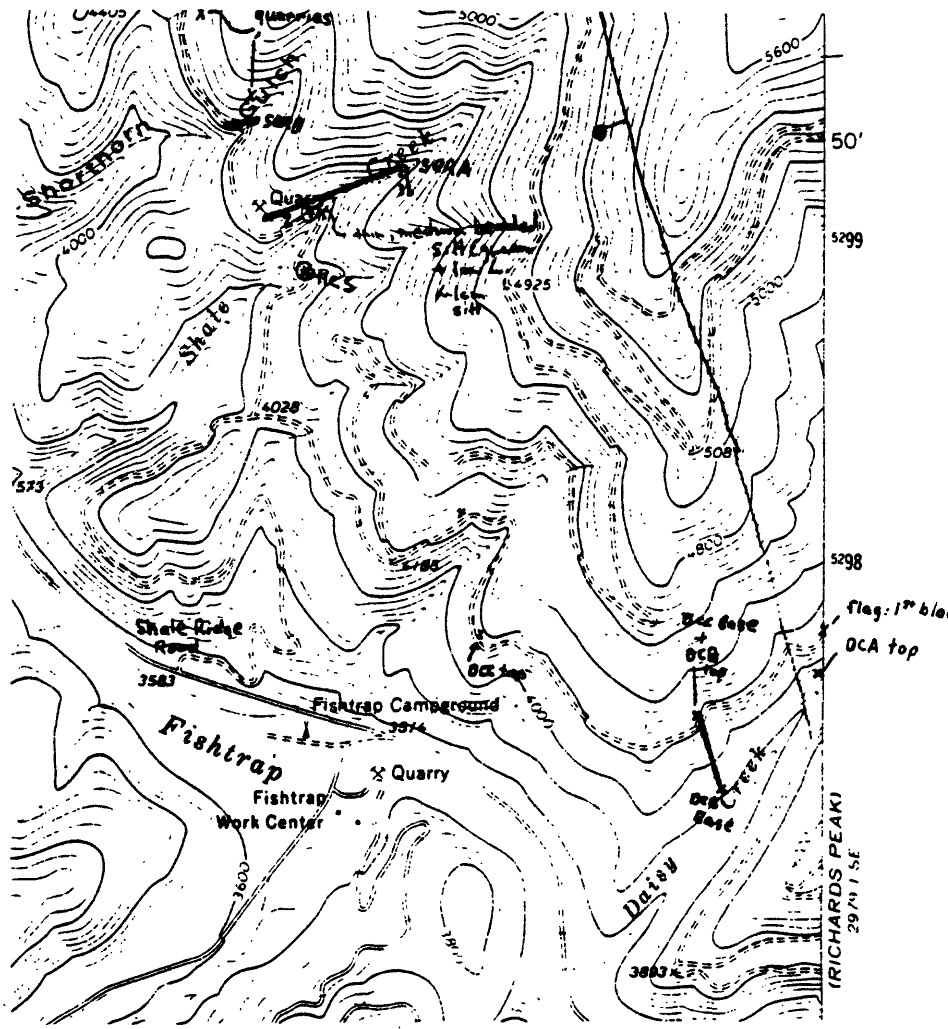
Localities for measured sections and for Lake Creek Quarry ooids. Localities for sections 2A, 2AA, and 2-MP-25 are as follows: 2A, 2AA located at milepost 27 on U.S. Highway 2 west of Libby, Montana. 2-MP-25 located as milepost 25 on U.S. Highway 2 west of Libby, Montana.



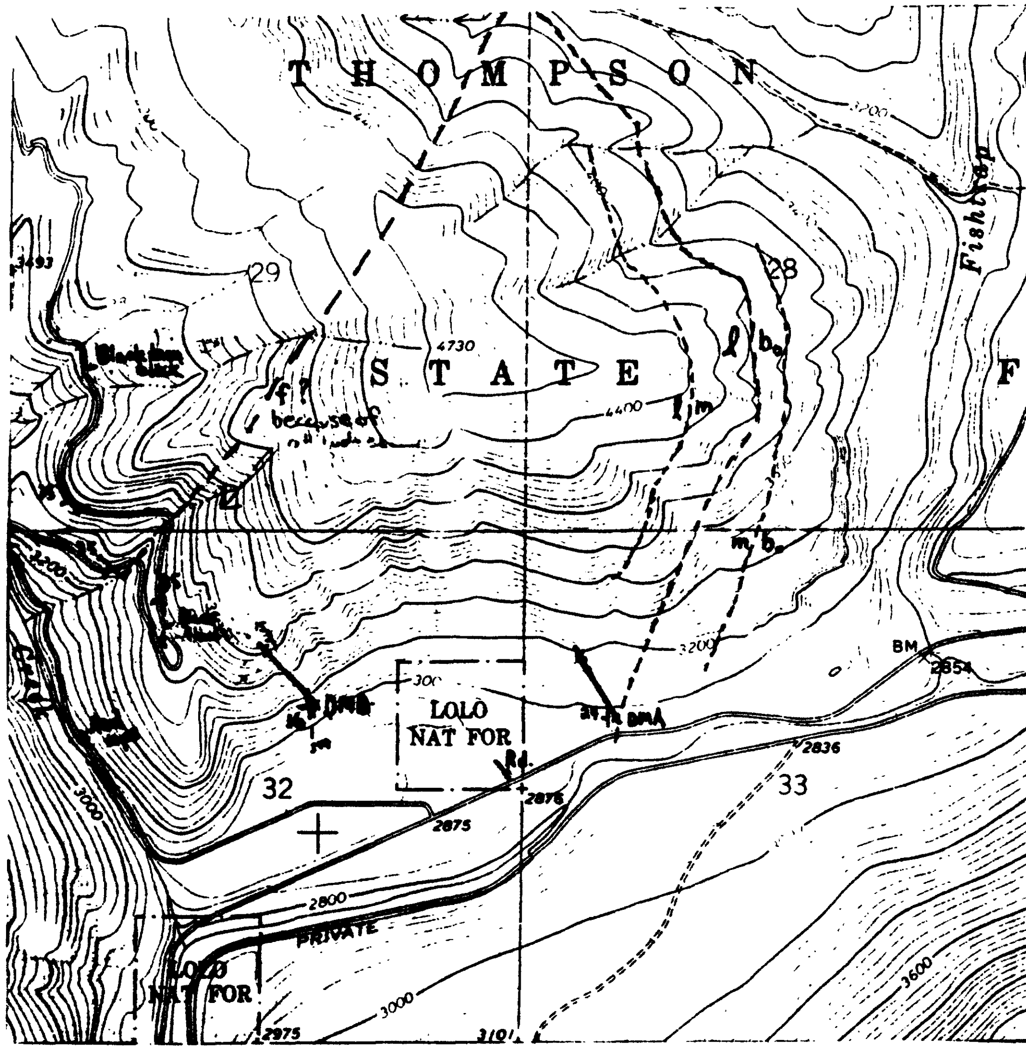
Location of principal F (Flagstaff Mountain) sections on the south flank of Flagstaff Mountain, just north of the Kootenai River about 8 miles west of Libby, MT. Base of Libby Formation was measured at milepost 27 on the south side of the river.



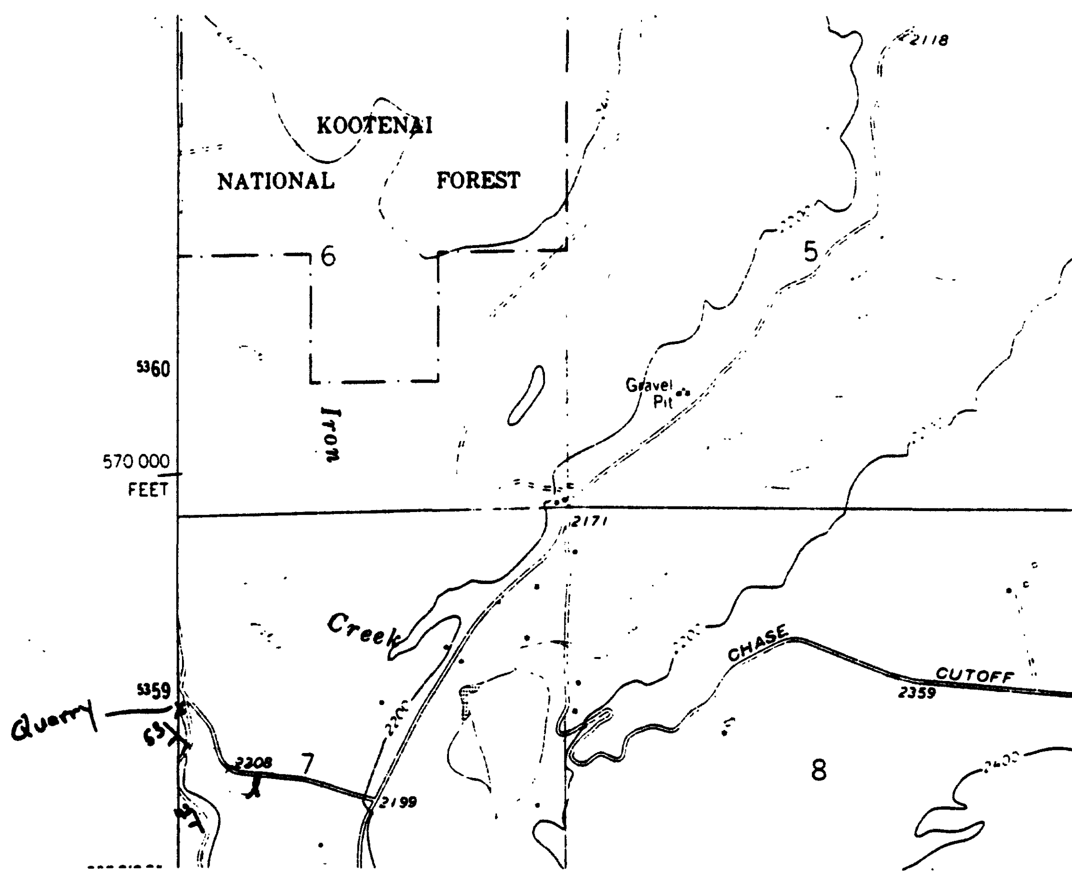
Location of CF (Clark Fork) section on the Clark Fork 15 minute quadrangle. Section shown as solid line about 1 1/2 miles west of Cabinet Gorge Dam.



Location of DC (Daisy Creek) and SQR (Shale Ridge Quarry) sections on Fishtrap Lake 7 1/2 minute quadrangle. DCA starts at lowest exposures in Daisy Creek drainage (top of section shown at "x"). DCB shown with solid line. DCC measured along Daisy Creek road from top of DCB. SQRA and SQRB sections shown as solid lines.



Location of DM (Deerhorn Mountain) sections on the south flank of Deerhorn Mountain just north of the Thompson River Road on the Calico Creek 7 1/2 minute quadrangle. Sections DMA and DMB shown as solid lines. Heavy dashed line is a fault (f). Light dashed lines are contacts between Bonner Quartzite (b_o) and Libby (ℓ) and McNamara (m) Formations.



Location of LCQ quarry shown as "x" in southwest corner of Kootenai Falls 7 1/2 minute quadrangle about 6 miles south of Troy, MT.