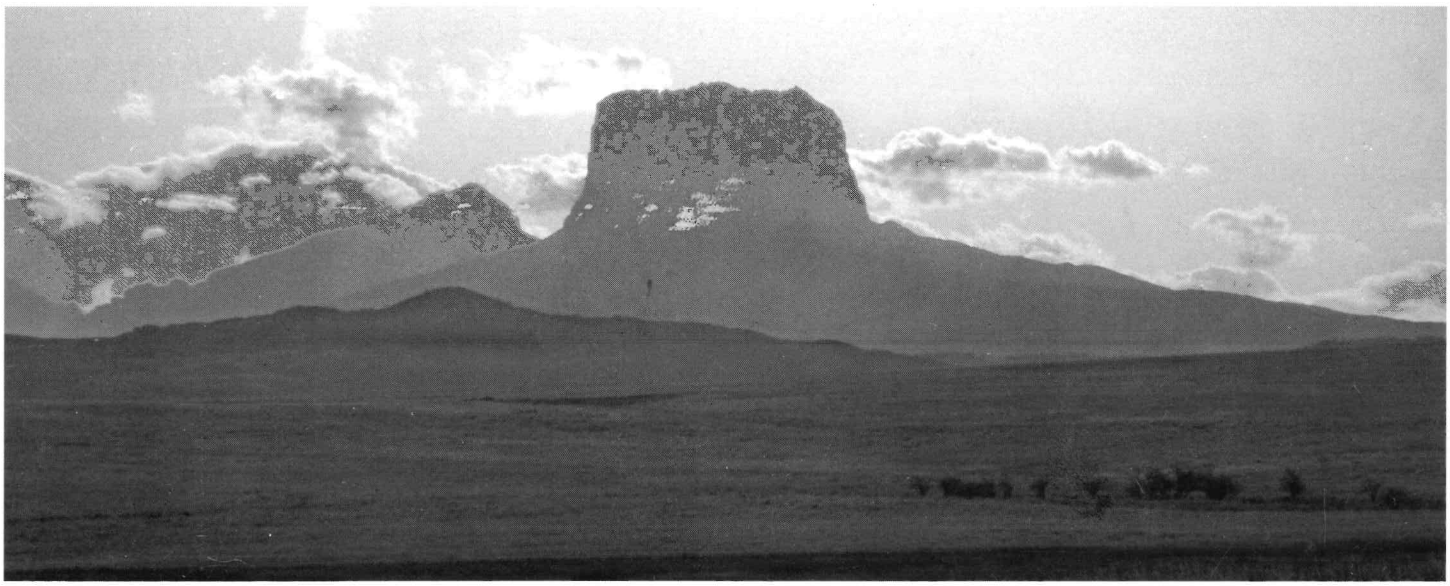
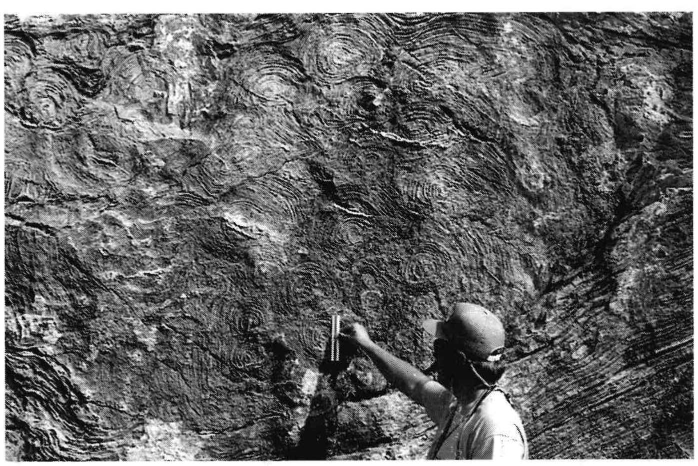
Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and Vicinity, Montana and Adjacent Canada

Paul Karl Link, Editor





Field Trip Guidebook for Belt Symposium III, Whitefish, Montana, August, 1993, Revised 2nd edition, June 1997

Includes Descriptions of Geologic Day Hikes in Glacier National Park along the Highline Trail and the Baring Creek-Siyeh Bend Trail

Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada

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Second Edition, 1997

Field Trip Guidebook for Belt Symposium III,
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Cover illustrations:
Top: *Chief Mountain, northwestern Montana; view looks west from the U.S. border station at Piegan. The Lewis Thrust is located at the break in slope at the base of the vertical cliff. It places Altyn Formation of the Middle Proterozoic Belt Supergroup over Cretaceous shale. The boundary between Glacier National Park and the Blackfeet Indian Reservation is at the summit of the mountain.*
Bottom: *The "Beautiful Bioherm" in the Baicalia-Conophyton stromatolite zone of the Helena (Siyeh) Formation, Going-to-the-Sun Road.*
Photographs by P.K. Link

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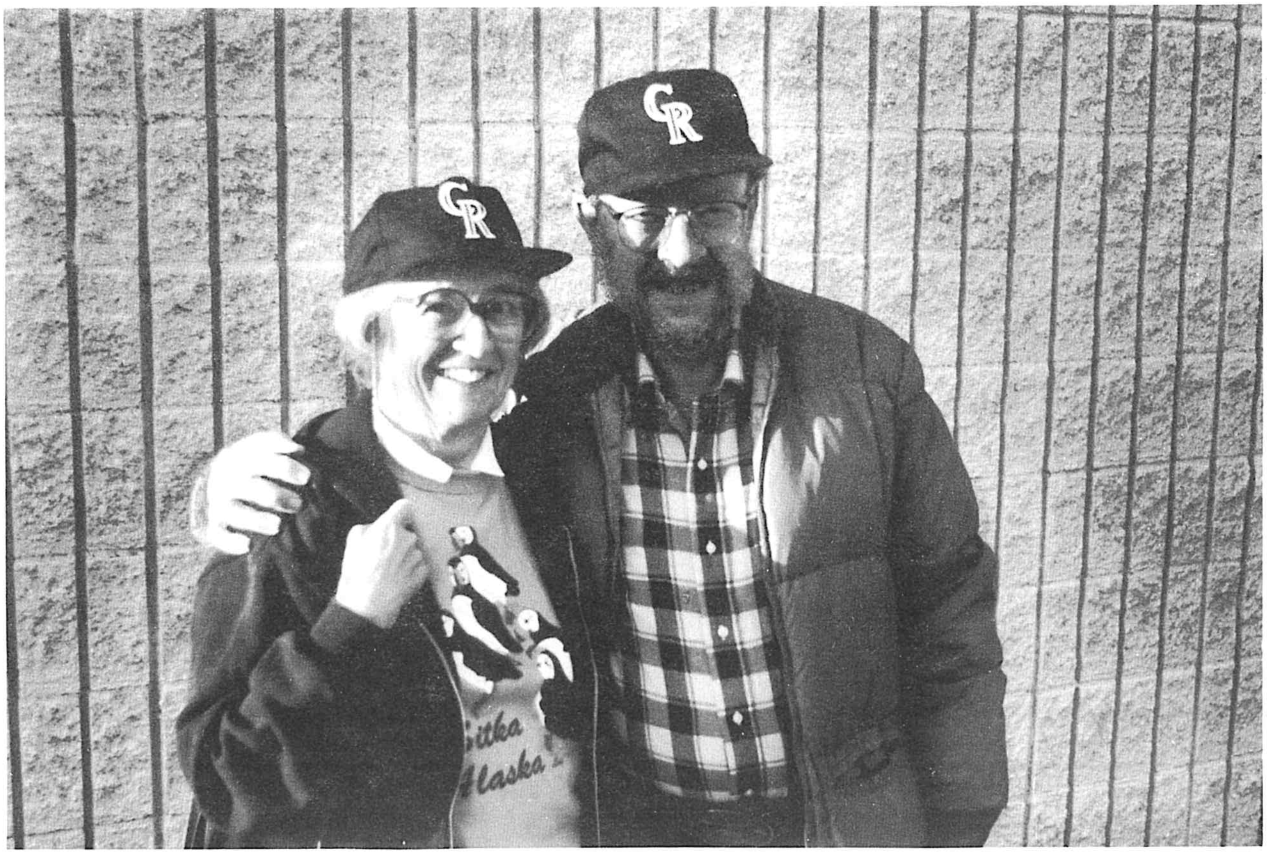
The Belt Association, Inc.

Contents



Jack and Jo Harrison, with Colorado Rockies baseball caps.
Libby Montana, September 1992.

Dedication to Jack and Jo Harrison

The Belt Supergroup has nurtured a series of tough, independent and sometimes world-class geologists. Jack Harrison is one of these.

A member of the U.S. Geological Survey since 1951, Jack has been working in rocks of the Belt Supergroup since 1957 and publishing on them since 1961. His exquisitely detailed geologic maps and geometrically balanced cross sections stand as tributes to field geology as it should be done. Jack also has had the drive, wisdom and luck to see almost all his mapping projects to completion and to availability as colored maps. He has also been active in the general area of Precambrian geology, and has been a long-time participant in the IUGS Subcommission on Precambrian Stratigraphy and Chairman of the IUGS Working Group on the Precambrian for the U.S. and Mexico.

Jack has been a mentor to many Belt geologists, an intransigent protagonist to some, and a friend to all. He was one of the main players in Belt Symposium I (Moscow, Idaho, 1973), an organizer of Belt Symposium II (Missoula, Montana, 1983), and a behind-the-scenes advisor for Belt Symposium III. In many ways much of what we know about the Belt basin we know because of the efforts of Jack Harrison.

Jack and Jo Harrison have come to Montana and Idaho for summer field mapping for over 30 years. They are warm and friendly people who have brightened the summer evenings of many visiting geologists. To them the geological community is much in debt.

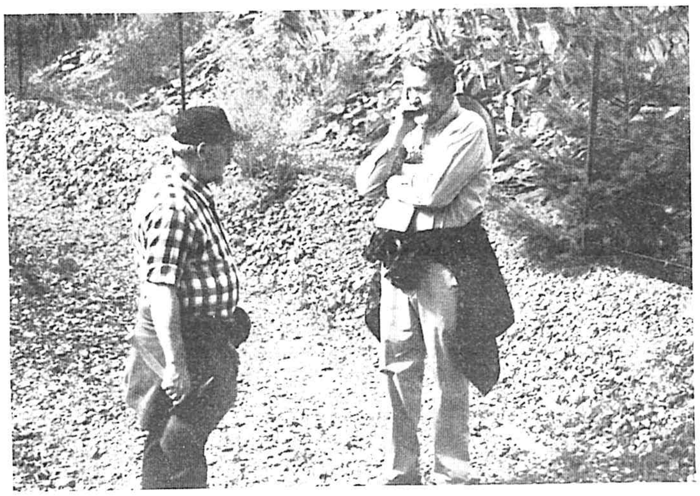
To them Belt Symposium III and this guidebook are dedicated.

Added for Second Edition, May 1997

After participating in Belt Symposium III, in August 1993, Jack Harrison died at his home in Lakewood Colorado on June 2, 1995, at the age of 71, after a two-year battle with lung cancer. Later that year the U.S.G.S. Branch of Central Regional Geology, which Jack had led and shaped, was terminated, in response to the reorganization of the U.S.G.S. and a Reduction in Force within the agency. It is a blessing that Jack did not have to live to see this. As of 1997, U.S. Geological Survey research in the Belt Supergroup has almost totally stopped. The future of Belt research is now in the hands mainly of academic workers.

This 2nd edition includes two new papers, by Whipple and others, and slight revisions to the Winston and Lyons paper. The editor apologizes for the tardy revision of this volume. That tardiness is much related to the events discussed in the preceding paragraph.

Belt Symposium III Organizers at Work (and Play)



Jack Harrison (left) and Don Winston disagree once again about the placement of the contact between the Helena and Wallace formations, north of Troy, Montana. The irresistable force meets the immovable object.

Left to right: Jim Whipple, Jack Harrison, Pier Binda in South Drywood Canyon, Alberta.

Don Winston studies the Grinnell Formation from underneath, South Drywood Canyon, Alberta.

Jim Whipple contemplates the complexities of the Belt basin.

BELT SUPERGROUP STRATIGRAPHY AND STRUCTURE, NORTH-CENTRAL BELT BASIN, NORTHWESTERN MONTANA

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INTRODUCTION

The purpose of this field trip is to see some rock facies of the Belt Supergroup in north-central Belt terrane in the Kalispell, Mont., 1° x 2° quadrangle (fig. 1). The trip emphasizes facies similarities and differences in some of the readily accessible Belt strata and touches on the structural complexities of Late Proterozoic broad open folds that were cut by Cretaceous thrusts and deformed by associated folds, then were sliced by abundant Tertiary listric normal (extension) faults. The first day of the field trip starts in Whitefish and includes trips northward up the Rocky Mountain Trench along the eastern side of the Purcell anticlinorium to Eureka, west across the nose of the anticlinorium, south along the anticlinorium to the Libby thrust belt, and then west into the thrust belt, ending in the town of Libby. The second day includes visits to extensive new road cuts along U.S. Highway 2 west of Libby in the Libby thrust belt and on the nose of the Sylvanite anticline, a return trip to Libby to go south along the trough of the old and now highly faulted Libby syncline, then a traverse east across the Libby thrust belt and onto the Purcell anticlinorium, and eventually a return to the Rocky Mountain Trench at Kalispell and north to Whitefish.

A useful adjunct to the field trip log and illustrations is U.S. Geological Survey Miscellaneous Investigations Series Map I-2267 (Harrison, Cressman, and Whipple, 1992), which includes geologic and structure maps of the Kalispell 1° x 2° quadrangle as well as geologic cross sections and lithologic and stratigraphic columns for various areas of the quadrangle.

Stops on the field trip are shown on the index map (fig. 1), and the formations or members of the Belt Supergroup to be seen are indicated on the historical correlation chart (fig. 2). The chart also shows the increased stratigraphic knowledge through time as geologic mapping connected former isolated areas, and the regional extent of formations and informal members became evident.

Some stops on this field trip are described in various other road logs written, at least in part, by Don Winston (for example, Winston and Woods, 1986). Winston uses "sediment types" (Winston, 1986, p. 89-104) to define and describe Belt rocks. These "sediment types" commonly differ from standard formations and members and result at places in nomenclature, correlations, and basin analysis at variance with that derived from lithostratigraphic mapping and correlation, and strict application of the North American Code of Stratigraphic Nomenclature as required by the U.S. Geological Survey. The geologic map of the Kalispell 1° x 2° quadrangle is based on U.S. Geological Survey practice.

We also follow U.S. Geological Survey practice in restricting areally the use of the term Grinnell to a limited area in and adjacent to Glacier National Park. The Grinnell Formation is primarily defined by locally abundant quartzite beds that contain an influx of well rounded and well sorted white quartz grains and red mud chips that form distinct beds and lenses in a red-bed argillitic sequence that extends all along the eastern Belt terrane from north of Waterton Lakes in Canada to White Sulfur Springs in the Helena embayment in the United States. We use Spokane Formation for that extensive red-bed formation.

We do not consider a road log designed to show rock types and structures in one part of Belt terrane to be a proper forum for extensive scientific debate about a marine or non-marine genesis of the Belt basin. Our preference is for a basin dominated by marine processes, or at least a world-ocean connected basin, as expressed by Cressman (1989) for the lower half (20,000+ feet of the Prichard Formation) of the total Belt stratigraphic column although one of us (DLK) supports a non-marine origin for certain Belt rocks in the

Harrison, J.E., Whipple, J.W., and Kidder, D.L., 1993, Belt Supergroup stratigraphy and structure, north-central Belt basin, northwestern Montana, *in* Link, P.K., ed., **Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada:** Belt Symposium III Field Trip Guidebook, Belt Association, Spokane, Wa., p. 1-19.

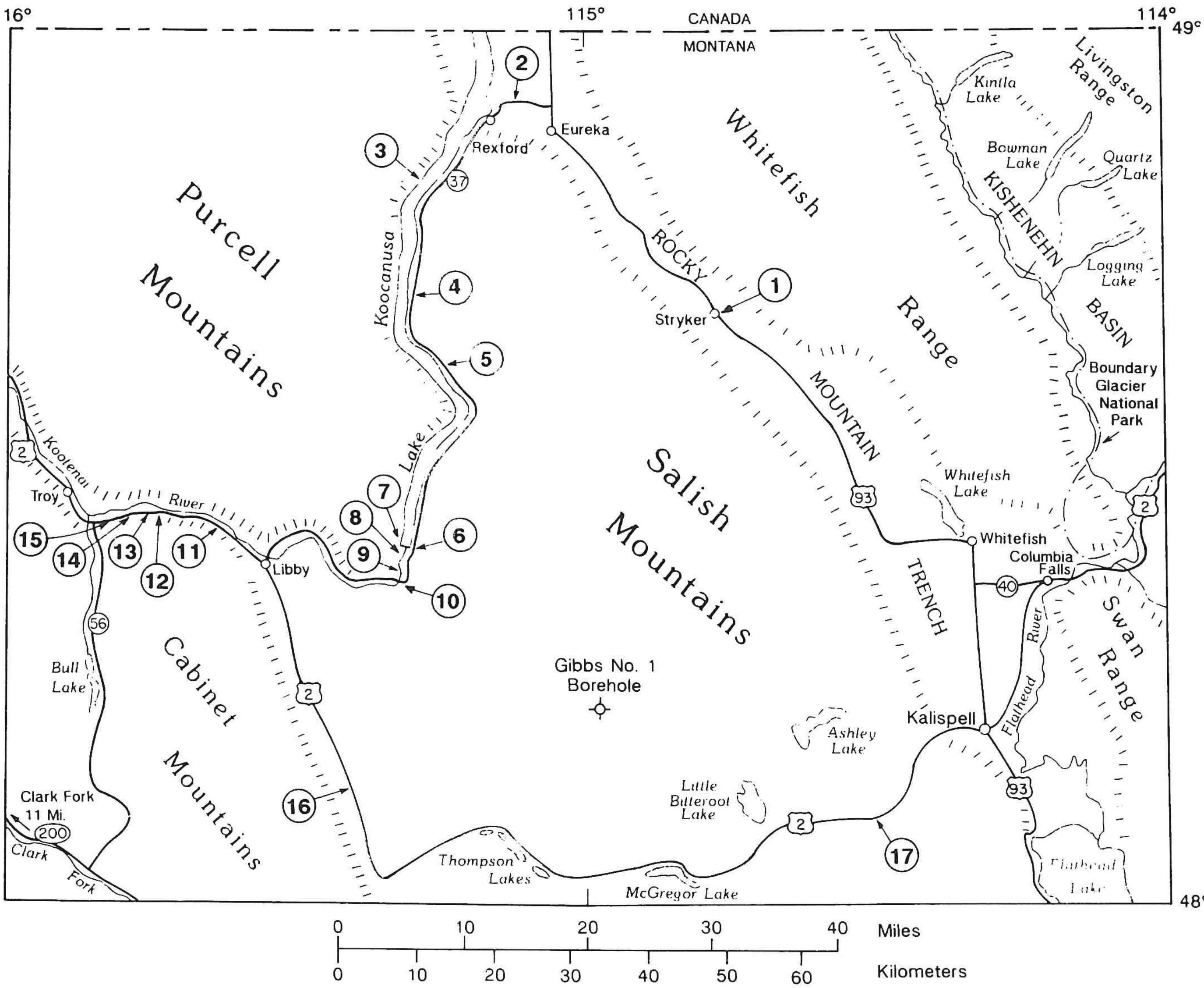

Figure 1. Index map of Kalispell 1° x 2° quadrangle, Montana. Circled numbers indicate locations of stops on two-day field trip.

eastern part of the basin. A contrasting view of the upper Belt as deposits in a lacustrine environment is championed by Winston (1991). We have given a few selected interpretations of depositional environments for some formations or members.

FIELD TRIP LOG

First Day

Whitefish to Libby

Mileage

0.0 Entrance to Grouse Mountain Lodge (about 1 mi west of downtown Whitefish).

U.S. Highway 93 to Eureka heads northwest up the Rocky Mountain Trench, which is defined topographically to the east by the Whitefish Range and less clearly to the west by various parts of the Salish Mountains (fig. 1). Geologically, the trench consists of a series of high-angle Tertiary extension faults loosely related to the topographic boundaries and dividing a stack of thrust plates of the Whitefish Range on the east from the Purcell anticlinorium on the west (Harrison and others, 1992, sheets 1 and 2). The high-angle faults cut through or bound many of the hills within the trench that have been glacially carved into elongate drumlins and drumlinoids. Surrounding these outcrops are glacial deposits as thick as a few thousand feet that consist predominantly of till, outwash, and lesser amounts of lake silts and dune sands. Tertiary sedimentary rocks are known in the southeast part of the trench near Columbia Falls but appear to be absent in the northwest near Eureka. Seismic lines across

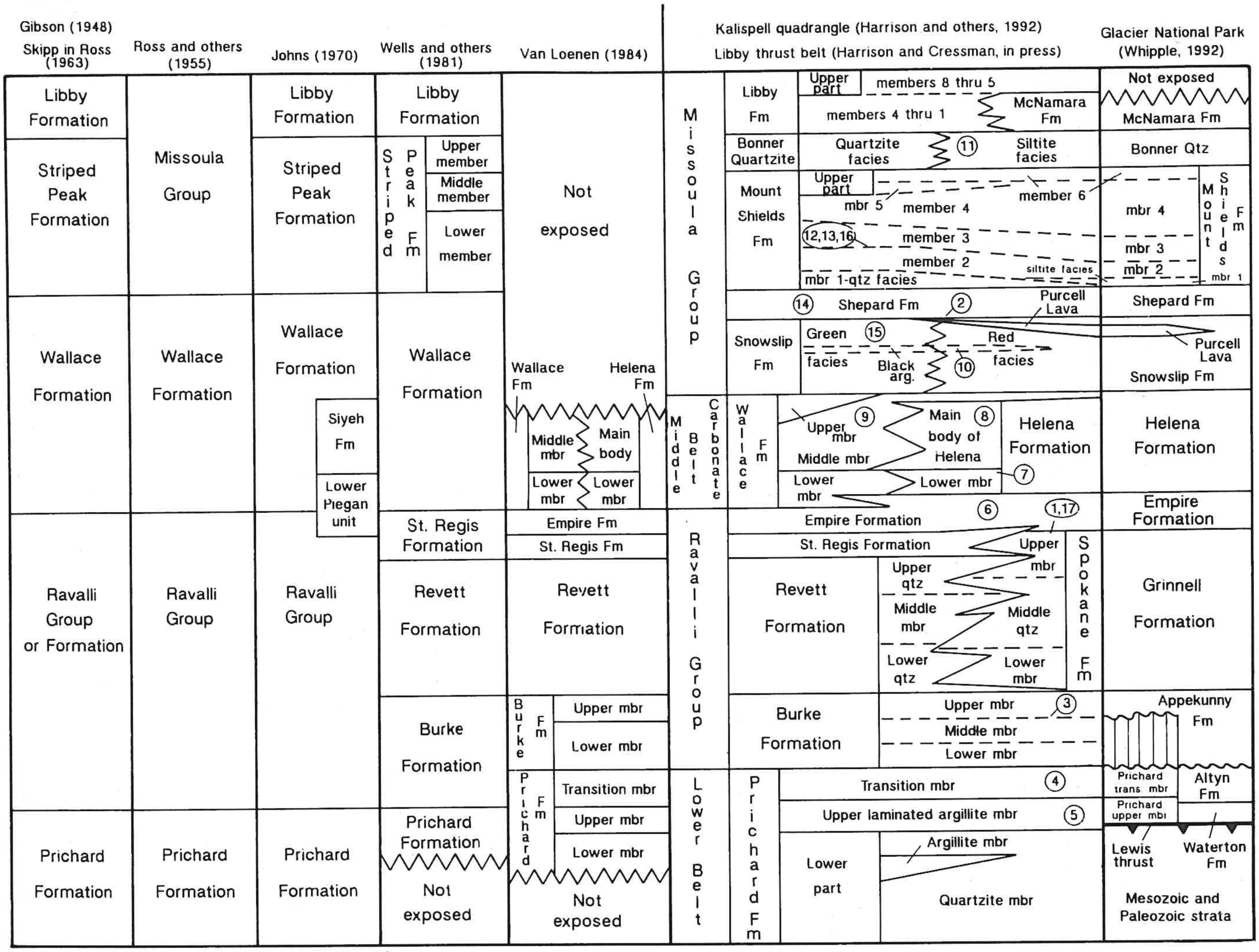

Figure 2. Correlation chart of stratigraphic units in the Middle Proterozoic Belt Supergroup used on various maps in northwestern Montana since 1948. Circled numbers are stops on the field trip. Dashed lines indicate informal members or beds recognized but not mapped.

the trench show a loss of continuity of reflectors within the trench because of energy absorption by the thick valley fill, indecipherable ricochetting of reflections among the buried bedrock hills, or complex and deep structure.

1.0 Traversing west across faulted main body of the Helena Formation.

3.6 Tally Lake road. Highway 93 turns northward along the Rocky Mountain Trench.

5.4 View of Whitefish Range straight ahead.

9.8 Empire Formation in railroad cuts on right.

15.7 Good Creek road.

16.4 Headquarters of Stillwater State Forest. At the request of the State Forest, a special detailed geologic map was prepared for their use (Whipple and Harrison, 1987).

17.2 Road to Olney.

23.1 Road to Radnor.

28.5 Glacial grooves in red beds of the Spokane Formation.

28.8 Lincoln County line.

30.0 Stillwater River.

31.3 **STOP 1.** Turn left for about 150 ft on road to Stryker, then right at Y intersection. Park in wide area on right and climb up to road cut in Spokane Formation. Beware of traffic, particularly logging trucks, rounding curve.

The upper part of the Spokane Formation, largely purple laminated argillite and siltite, is exposed in these road cuts and contains thin white quartzite beds that are characteristic of the formation in this area. The well-sorted quartzite is interpreted to have had a source area of well sorted strandline sands to the east. These white quartzite beds increase in thickness, abundance, and grain size eastward until they compose 20% of the upper part of the Grinnell Formation in Glacier National Park (Whipple, 1992).

The dominant lithology exposed here consists of poorly sorted feldspathic arenite interlaminated with fining-upward couplets of purplish siltite and argillite. These lithologic units and mudchip breccias, accreted mudballs, shrinkage cracks, fluid-escape structures, and small-scale bidirectional cross-beds with mud-draped foresets are interpreted as representing sedimentation on peritidal mudflats largely by storm events (Collins and Smith, 1977).

In the Whitefish Range, a complete section of the Spokane Formation is exposed only in a small area north of Whitefish Lake, where the section is about 4,200 ft thick (Whipple and Harrison, 1987). The lower part of the Spokane in that area does not contain appreciable quartzite. It consists mostly of pale-purple argillite and dark-green siltite commonly as fining-upward couplets, and it is locally spotted with iron carbonate. Those lithologies closely resemble equivalent strata in the Creston Formation in nearby British Columbia.

Return to highway and proceed north toward Eureka.

35.3 North Dickey Lake turnoff and scenic overlook plus road to the community of Trego. Some local citizens insist, with a twinkle in their eye, that the town name stems from European immigrants who worked early on in the logging industry and, rather than yell the traditional warning of "Timber," preferred to yell "Tree go"...

37.1 Murphy Lake Ranger Station.

37.7 Fortine Creek Road.

39.6 Meadow Creek and Fortine road.

40.4 Crystal Lake road.

42.4 Grave Creek road.

44.0 Good view of Whitefish Range to the east.

45.4 Therriault Creek road.

46.4 Glen Lake road.

49.6 Southern end of Tobacco Plains. Named from experiments in the late 1800's by missionary priests and Indians to grow tobacco in the area.

51.0 Snowslip Formation in roadcut. Significant amounts of quartzite in this exposure.

51.3 Town of Eureka.

52.0 Rocky Mountain Trench widens.

53.2 Junction with Highway 37. Turn left (west) onto Highway 37 to Lake Koocanusa, whose name was manufactured from KOOtenai + CANada + USA. International border is 6 mi north.

Crossing Rocky Mountain Trench and approaching crest of Purcell anticlinorium. The anticlinorium is formed from broad, open, double-plunging, north-trending folds that are probably Late Proterozoic in age. In the Eureka area, the plunge is northward, and the Missoula Group is displayed in rotated blocks between east- and west-dipping listric normal extension faults that cut the anticlinorium. A small outcrop of Devonian sedimentary rocks due north at the Canadian border (Harrison and others, 1992, sheet 1) at first seems anomalous; however, the northern plunge of the anticlinorium brings up exposures of Cambrian and younger rocks that lie unconformably above the top of the Belt Supergroup in a normal stratigraphic sequence. The Whitefish Range and its stack of Cretaceous thrusts is separated from the Proterozoic Purcell anticlinorium by a series of Tertiary down-to-the-west extension faults that define the east edge of the Rocky Mountain Trench.

56.2 Tobacco River.

56.6 Road to Black Lake.

57.8 **STOP 2.** Park in pullout on right at Forest Service map of Lake Koocanusa area.

Exposed at this stop is the upper part of the Purcell Lava and overlying lower part of the Shepard Formation in road cuts to the east.

The Purcell Lava in this area is at least 225 ft thick and is mostly pahoehoe flows of vesicular and amygdaloidal basalt. The basalt is typically chloritized, and in some areas of the northern Whitefish Range it is associated with copper mineralization (Whipple and Hamilton, 1984). In an excellent study by McGimsey (1985), the Purcell in Glacier National Park is shown to be alkaline, and initial flows were subaqueous, forming pillow lavas in less than 50 ft of water. The lower pillow facies is not present in these road cuts, possibly due to faulting.

At this stop and locally throughout its exposure, the Purcell Lava is overlain unconformably by the Shepard Formation. Here, we can see a thin dark-green volcaniclastic interval overlying the basalt. This green unit is succeeded by the typical carbonate lithologies of the Shepard. Elsewhere, in the northern Whitefish Range, the lava is overlain by an intraformational conglomerate (Whipple and others, 1984), and in Glacier National Park the lava is interstratified with the upper part of the Snowslip Formation (Whipple and Johnson, 1988).

The carbonate beds of the Shepard consist of a mixture of tan-weathering calcareous siltite and argillite couplets, calcareous arenite, dolomite, and stromatolitic and oolitic limestone. A few pinkish-gray-to-maroon strata punctuate the carbonate lithologies.

The Shepard Formation in this north-central part of the basin appears to represent a barrier bar and shoal deposit. In the northern Whitefish Range, beds of low-angle cross-laminated arenite are common in the lower part of the formation. The Shepard is about 500 ft thick in the Rexford area.

58.8 Rexford Bench Recreation Area. For a rest stop, follow signs to beach. The tan-colored cliffs across the lake expose silts deposited in Glacial Lake Kootenai (Alden, 1953).

At this point, the road turns southward, crosses the Pinkham thrust that places the Prichard Formation over the Helena Formation, and traverses the Pinkham plate from gentle double-plunging folds of the eastward-displaced Purcell anticlinorium to the Libby thrust belt (Harrison and others, 1992, sheet 1).

60.2 Roadcuts in main body of the Helena Formation.

60.9 Pinkham Creek road.

61.5 Crossing Pinkham thrust. Here the thrust dips about 15° W., has a stratigraphic throw of about 8,000 ft, has a horizontal displacement of a few miles, and has been eroded back about 1.5 mi westward from an original overturned fold at the leading edge of a more steeply dipping thrust plane, as exposed to the southeast along the trace of the fault (Harrison and others, 1992, sheet 1). The older (Proterozoic) north-trending broad open folds, which can be identified in several parts of the Belt basin beneath the Cambrian unconformity (Harrison and Cressman, in press), are transected at about 25° by the Cretaceous thrust and its overturned upper plate tight fold. The Precambrian open folds in the Spokane and younger Belt formations to the east are cut off by the overriding Pinkham thrust, and other old folds of the anticlinorium are piggybacked and slightly deformed on the Pinkham plate to the west (Harrison and others, 1992, sheets 1 and 2).

Road cuts for the next 44 miles are predominantly in the Prichard and Burke Formations, which clearly reflect the biotite zone of regional burial metamorphism (Maxwell and Hower, 1967).

61.9 Road cuts in transition member of the Prichard.

65.0 Road to Camp.

65.6 Pinkham Creek. Upper Burke outcrops.

66.6 Rexford bridge. Turn right and cross Lake Koocanusa. Turn right onto west side road and park at first pullout at grizzly bear habitat sign. Beware of logging trucks.

67.2 **STOP 3.** Contact between middle and upper members of the Burke Formation.

The Burke Formation in this general area is about 3,000 ft thick and consists of three informal members (Harrison and others, 1992, sheet 1, fig. 2). The lower member contains 1- to 2-ft-thick alternating beds of green to gray siltite and green argillite that have a distinct light-gray weathering rind and give a blocky to flaggy appearance in outcrop. The middle member is predominantly gray to purple-gray siltite that contains ripple cross-laminations at places. The upper member consists of alternating beds a few tens of feet thick of laminated purple argillite and siltite, and green argillite and siltite that at places contains stratabound copper-sulfide minerals. All members are characterized by secondary biotite and are speckled by tiny euhedral magnetite crystals.

At this exposure, the lower part of the road cut is the middle siltite of the Burke exhibiting a few sedimentary structures such as small-scale cut-and-fill and flame structures, but the rock is generally massive in appearance. Better defined and thinner beds are down section. Purple and green argillitic beds of the upper member increase in number and thickness above road level through an interlayered transition zone about 100 ft thick--a typical Belt contact between formations or members. These argillitic beds have an increasing abundance of mud chips, mud cracks, and ripple marks towards the top of the member.

Both the middle and upper members of the Burke Formation are exposed extensively in road cuts for the next several miles towards Libby. Blocky gray exposures are the middle member; purple and green argillitic rocks are the upper member. The lower member is not exposed and may be missing (due to nondeposition?) at places above the top transition member of the Prichard Formation.

Correlation of the Burke Formation and its members to the north, west, and south within the Kalispell quadrangle is reasonably straightforward; correlation to the east across the Rocky Mountain Trench and into the Whitefish Range is not (Harrison and others, 1992, sheet 1, fig. 2). The authors have agreed to disagree at this time as to whether a mappable olive to green argillitic siltite unit to the east, above the Prichard and below the Spokane, is a facies of Burke or a facies of the Appekunny Formation of Glacier National Park (Whipple and Harrison, 1987).

67.8 Recross bridge and turn right on Highway 37 toward Libby.

70.8 Cliffs in middle member of Burke--a rock climbers favorite!

73.0 Sutton Creek road.

73.4 Sutton Creek.

76.2 Peck Gulch Recreation Site. A narrow, winding road goes down to a lake-level rest area.

78.1 **STOP 4.** Pullout on left in old logging-road junction just after 44 mi road marker. Transition member of the Prichard Formation.

This top transition member of the Prichard Formation is about 2,200 ft thick in this area and consists of a variety of lithologies that in part resemble Burke rocks above and in part Prichard rocks below. Within the transition member three different units--a basal unit of blocky silty argillite, a middle unit of irregularly interlaminated siltite and argillite, and an upper unit similar to the middle but containing quartzite and carbonate--extend over the entire Kalispell quadrangle (Cressman, 1989, p. 34-38; Harrison and others, 1992, sheet 1, fig. 2).

The road cut exposes rocks a few hundred feet down in the upper unit of the transition member of the Prichard Formation. Carbonate has weathered out from a blocky gray siltite, which is one of several such carbonate-bearing beds in this stratigraphic interval. Conspicuous beds of pyritic black argillite are typical of many Prichard strata. Gray biotitic siltite beds have a few small-scale sedimentary features such as mud cracks or cut-and-fill structures.

Carbonate in this stratigraphic interval and in the unit below suggests a distal relation to the carbonate-bearing Altyn Formation of Glacier National Park.

78.2 McGuire Creek. Roadcuts for the next several miles are in the transition member of the Prichard.

79.7 Tweed Creek. White layers a few inches thick are quartzite in the transition member of the Prichard.

81.6 Rocky Gorge Recreation Site.

84.0 Crossing contact into upper laminated argillite member of the Prichard Formation.

84.8 Sheep Creek.

85.5 **STOP 5.** Paved pullout on right. Upper laminated member of Prichard Formation.

This distinctive planar-laminated, biotitic, pyritic or pyrrhotitic argillitic rock consisting of couplets of light-gray argillitic siltite and dark-gray argillite a few tenths of an inch thick can be recognized throughout Belt terrane. It is commonly called the "lined rock." In this area, the exposed part of the laminated member is about 1,500 ft thick, but the base is not seen.

Conspicuous white layers a few inches thick in this rusty-weathering member are quartzite that commonly contains calcite, garnet, and hornblende (fig. 3). This occurrence of amphibolite facies minerals in the biotite zone of regional burial metamorphism probably reflects a closed system where excess CO_2 in metamorphosed dolomitic quartzite caused the unusual mineral assemblage. These Ca-rich quartzite beds are limited to the northeastern part of Belt terrane, although minor carbonate lenses are present in the member on the west side of Glacier National Park, and a few

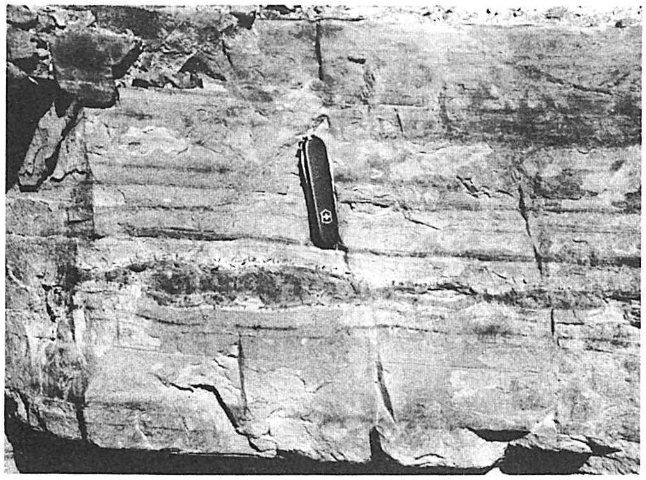

Figure 3. Calc-silicate lens along bedding in the upper laminated Argillite member of the Prichard Formation. Photo by Paul K. Link.

stromatolites occur at places as far west as Clark Fork, Idaho.

Cressman (1989, p. 33) interprets the depositional environment for this member as sedimentation partly from low-density turbidity currents and partly from suspension. The reducing environment required for the abundant authigenic iron sulfides and the varve-like laminae suggests a quiet anoxic basin that had a few local shallow parts where stromatolites could develop.

For the geophysically inclined, we should note that the iron sulfide is sufficiently abundant in this member to cause a distinct conductive zone that can be detected by various electrical and magnetotelluric methods at many places (Wynn and others, 1977; Long and Hoover, 1984; Van Blaricom, 1984).

86.8 Slumps in upper laminated member of the Prichard.

88.2 Ten Mile Creek.

88.5 Osprey nest in top of dead tree on right.

92.5 Stenerson Mountain road.

93.0 Five Mile Creek.

93.5 South side of Five Mile Creek road.

96.2 Warland Creek stock. Small body of syenite.

97.0 Warland Creek.

98.7 Koocanusa Marina.

99.2 Cripple Horse Creek.

100.1 Boundary Mountain road.

102.6 Dip slope in middle member of Burke Formation.

103.7 Canyon Creek road.

104.9 Upper quartzite member of Revett Formation.

105.2 St. Regis Formation

105.6 **STOP 6.** Turn right into parking lot at Libby dam. Rest rooms and observation platform. Empire Formation.

The heavily instrumented face in the road cut is in the Empire Formation, which during construction slid on a bedding plane overnight and left a huge pile of rock on the highway (fig. 4). As noted by geologic colleague and Star Wars fan Dave Kidder, "Aha! The Empire Strikes Back. "Please do not disturb this instrumentation or pound on the rocks near it. Look at the south end of the road cut--the formation is fairly uniform except for more abundant white quartzite beds in the lower part and more abundant calcite segregations in the upper part.

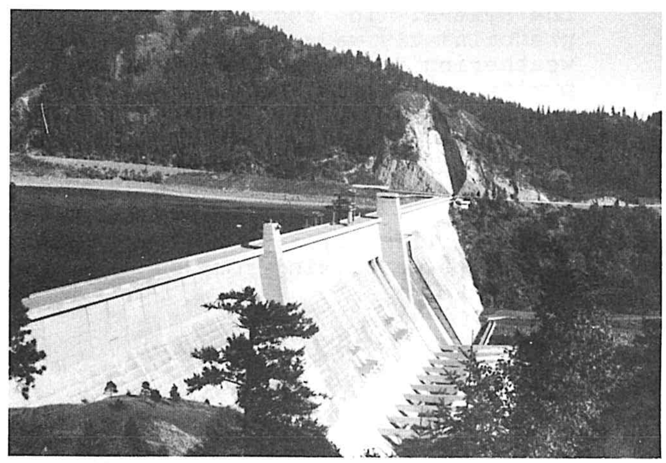

Figure 4. View looking east across Libby dam to road cut in Empire Formation. Missing wedge slid on bedding plane onto highway during construction. Photo by Paul K. Link.

The Empire Formation is about 700 ft thick and is almost completely exposed in the road cut. The formation consists predominantly of wavy laminae of light-green argillite and dark-green siltite in graded couplets. Pyrite, carbonate cement, and carbonate pods are common at places. Fluid-escape structures and mud cracks are displayed at places, as are inch- to foot-thick white quartzite beds. About 250 ft south of the turnoff to the dam is a dense, conchoidal-fracturing, white cherty bed with minor chalcopyrite that was used locally by Indians for tools and arrowheads.

Across the lake, natural exposures and road cuts display a tongue of the Empire about 100 ft thick in the upper part of the St. Regis Formation.

According to Eby (1977), and confirmed by Whipple, the Empire Formation represents the initial stage of marine transgression over red beds of the Spokane and Grinnell Formations culminated by deposition of platform carbonate deposits of the overlying Helena Formation. Thus, a major change in depositional history starts with the Empire.

106.0 Cross dam, turn left through parking lot to west side logging access road.

106.4 Turn right on west side road.

106.6 **STOP 7.** Turn right into viewpoint parking area. Lower member of Helena Formation.

Outcrops beside the parking area and in the road cuts adjacent to it are typical of the lower member of the Helena. The rocks are predominantly wavy laminated orange-weathering dolomite that displays pyrite and calcite pods. Also characteristic of the member are beds of thinly laminated apple-green argillite interbedded with thin layers of brown-weathering quartzite and brown to green silty pyritic dolomite (fig. 5). Gray-green quartzite containing rounded carbonate pebbles and molar-tooth structure (irregular ribbons of calcite in dolomitic rocks) are sparsely present. This lower member is about 950 ft thick in this area, but it contains many layers several tens of feet thick of cyclic-bedded rocks typical of the overlying main body of the Helena Formation.

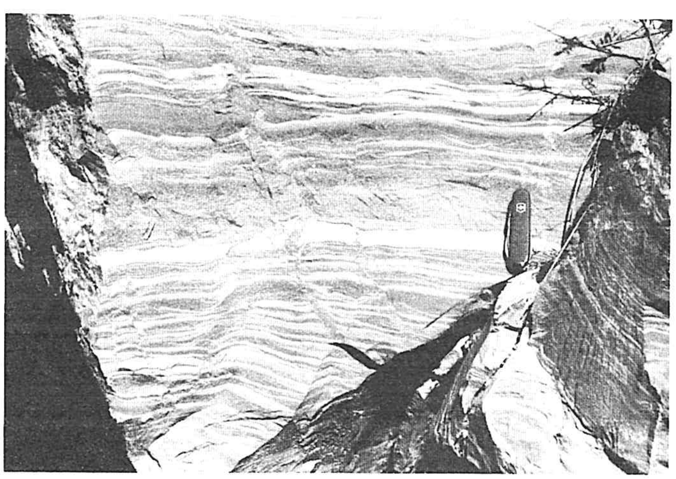

Figure 5. Lenticular couplets and fluid-escape structures in interlaminated green argillite (light colored) and brown quartzite (dark colored) of lower part of Helena Formation. Photo by Paul K. Link.

In the northern part of the Kalispell quadrangle, the lower member of the Helena is intertongued with beds of the lower member of the Wallace Formation (Harrison and others, 1992, sheet 1, fig. 2).

Follow arrows and exit north side of parking area. Turn left on west side road.

106.9 Visitor Center road. Continue on west side road.

107.2 **STOP 8.** Park in paved pullout on left. Main body of the Helena Formation.

The spectacular road cuts display the typical cyclic-bedded rocks characteristic of the main body of the Helena Formation, which is about 2,400 ft thick in this area. A complete tripartite cycle (fig. 6), typically about 8 ft thick, begins with a lower clastic bed of quartzite, or quartzite and black argillite, resting on an erosional surface. The middle bed is laminated dolomitic siltite that contains calcite segregations that change from horizontal pods to vertical pods to vertical ribbons (molar-tooth structure) from base to top of bed. The upper bed is dense, conchoidal-fracturing, orange-weathering dolomite that has an erosional surface at its top. The cycles are not all complete, and where a bed is missing, it most commonly is the lower clastic unit.

Variations in the Helena cycles in other parts of Belt terrane have been described by Eby (1977) and O'Connor (1967). We do not choose to engage in the controversy as to whether the rock cycles represent restricted marine or nonmarine deposition. The pod to molar-tooth upward progression in the middle bed has been interpreted by various observers (for example, O'Connor, 1967) as some sort of diagenetic event, either a filling of voids, or remnants of fluid escape. Regardless of genesis, both the bedding cycle and the internal order of calcite segregations are consistent and are particularly useful in determining tops of beds.

The contact with the lower member of the Helena Formation is exposed

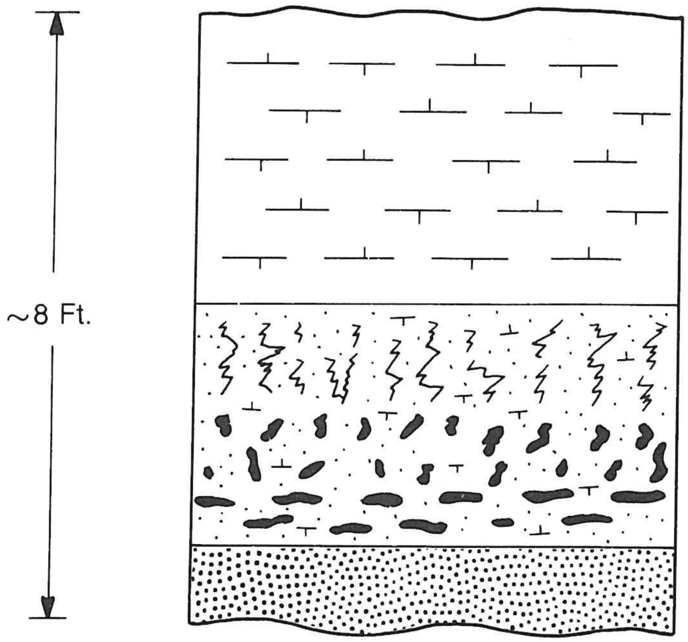

Figure 6. A complete typical depositional cycle in the Helena Formation near Libby dam. The lower bed is quartzite, the middle bed is dolomitic siltite, and the upper bed is dense dolomite. Within the middle bed, black ovoids and irregular lines represent calcite segregations. The cycle is bounded on top and bottom by erosional surfaces.

about 400 ft south of the green sign pointing to the Visitor Center. The contact is marked by the appearance of brown-weathering quartzite and apple-green argillite below the cyclic-bedded main body of the Helena.

A note for the geophysically inclined: the Helena Formation provides faint but discernable seismic reflections in many areas. Reflectors are presumably the dense dolomite beds.

Continue south on west side road.

107.9 Alexander Creek road.

108.2 Powerhouse road.

109.3 **STOP 9.** Pull into road on right. Middle member of Wallace Formation.

In this area, the middle member of the Wallace Formation, described below, is about 3,000 ft thick, but about 20 percent of the unit is intertongued rocks of the main body of the Helena Formation, described above. We coined the field term "Wallena" for these intertongued mixtures.

Rocks in the side-road cut at the pullout are Helena, but the purpose of the stop is to view the outcrops of the middle member of the Wallace Formation in the low road cuts a few hundred feet to the north. The middle Wallace is another distinctive basin-wide unit of the Belt Supergroup. Displayed in the road cuts are the typical rhythmically paired beds of black argillite in irregular, uneven, wavy beds that drape over lenticular beds of tan-weathering, white to gray dolomitic siltite to very fine grained quartzite (the "black and tan" in field jargon). Pyrite is common, as are sedimentary structures such as small-scale cross-bedding, channels, and load and flame structures.

Continue south on west side road.

110.4 Junction with Highway 37. Turn left and cross bridge at confluence of Kootenai and Fisher Rivers.

110.8 Just past bridge, turn right onto Fisher River road.

111.1 Fisher River and railroad bridge.

111.3 **STOP 10.** Go left at Y and park on gravel at junction. This junction is with paved private logging road to mill at Libby. Heavy logging-truck traffic. Road cut is in a black argillite marker interval in the Snowslip Formation.

The road cut exposes black argillite and green to gray siltite in wavy, graded couplets a fraction of an inch to a few inches thick. Beds are commonly pyritic. Small-scale sedimentary structures include slump folds, disrupted beds, cross laminae, cut-and-fill structures, and channels. Scattered stromatolites and carbonate-bearing beds or laminae are visible. This black argillite marker interval is about 700 ft thick here, with the Snowslip Formation being about 3,000 ft thick (fig. 7, Buck Creek section).

The Snowslip Formation has been studied extensively in Glacier National Park, which contains its type section, by Whipple and Johnson (1988). In the park, the Snowslip contains six informal members consisting largely of alternating beds of red and green clastic rocks that are locally calcareous and include the Purcell Lava near the top of the formation. Regional correlation shows a general fining in grain size westward, as well as gradual loss of red-bed sequences and the appearance of black and green argillite that increases in abundance westward. Variations in lithology across the Libby thrust belt are shown in figure 7, and a green facies of the Snowslip that contains few or no red beds is mapped west of Libby

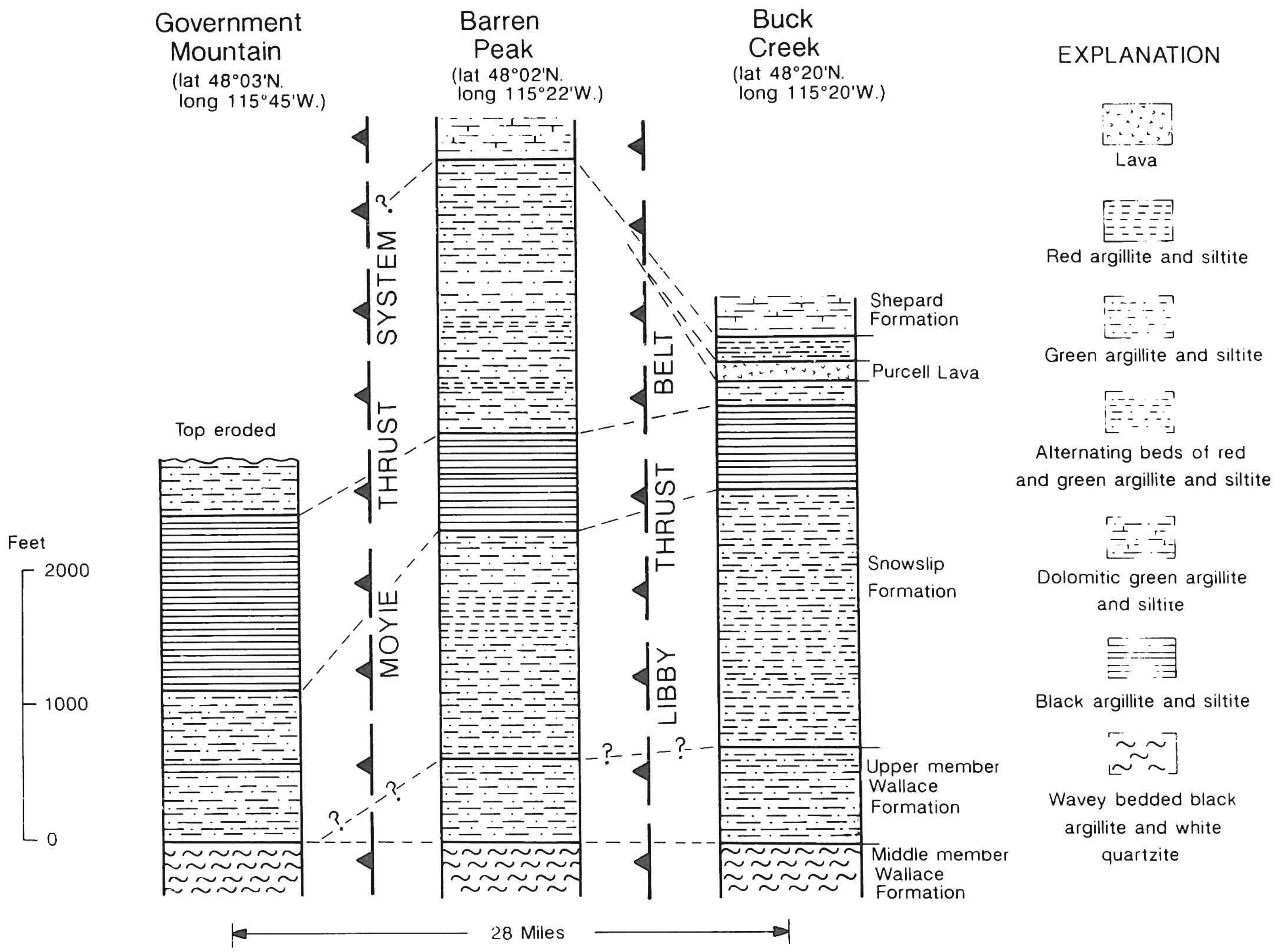

Figure 7. Variations in lithology of the Snowslip Formation across the Libby thrust belt and the Moyie thrust system of northwestern Montana.

(Harrison and others, 1992, sheet 1). This green facies will be seen at **STOP 15.**

Whipple and Johnson (1988) suggest, on the basis of changes in grain size, mineralogic composition, and facies of both the Snowslip and its enclosed Purcell Lava, that the basin slope and sediment transport direction during Snowslip deposition was generally northwestward.

111.7 Go around Y junction and return to Highway 37. Turn left toward Libby.

111.9 Confluence of Kootenai and Fisher Rivers.

112.1 West side road.

113.1 Canoe Gulch Ranger Station. Crossing easternmost thrust in Libby thrust belt.

116.7 Empire Formation outcrops in road cuts.

120.5 Rainy Creek road to Zonalite Mine.

Vermiculite deposits about 0.5 mi northeast, on Vermiculite Mountain, were the main source of vermiculite in the United States until 1991. The deposits occur in an altered laccolithic pyroxenite-syenite complex that was intruded at about 104 Ma into the Belt formations. The complex cuts through the lead thrust fault in the Libby thrust belt (Harrison and others, 1992, sheet 1) and thus establishes a minimum age for the thrusting in this part of the thrust zone that extends from eastern Washington to the Montana Disturbed Belt and ranges in age from about 200 Ma in the west to about 60 Ma in the east (Monger and Price, 1979).

121.0 Steep westward dips of right-side-up beds represent an original west-facing limb of the Purcell anticlinorium that was oversteepened as the Libby syncline was jammed against it during Cretaceous thrusting (Harrison and others, 1992, sheet 1, section A-A').

121.1 View of Cabinet Mountains straight ahead.

123.0 Purcell Lava in road cut.

124.7 Libby Ranger Station.

124.9 Pipe Creek road.

125.0 River road.

125.3 Kootenai River and town of Libby.

126.0 Junction with Highway 2.

End of first day's log.

Second Day

West of Libby, then back to Libby

The first part of this day's trip goes westward on U.S. Highway 2 through the Libby thrust belt, across the south-plunging nose of the Late Proterozoic Sylvanite anticline, and into the upper plate of the Moyie thrust system, which overrides the anticline (Harrison and others, 1992, sheet 1). Extensive deep road cuts display rocks of the Missoula Group that have been thrust eastward and are cut by many high-angle extension faults. The Sylvanite anticline is cored by the Prichard Formation to the north and exposes in its south-plunging nose a section of Belt strata from 20,000 ft down in the Prichard up through the eroded top of the Libby Formation. The section measures about 49,000 ft, including several mafic sills that have an aggregate thickness of 2,000-3,000 ft (Harrison and others, 1992, sheet 1, fig. 2). The base of the Belt is not exposed. This section is the thickest known continuous section of Belt strata and also represents the thickest Middle Proterozoic sedimentary basin in the world.

Mileage

0.0 Take Highway 2 west toward Troy.

Flashing caution light at cross-street adjacent to Venture Inn.

2.0 West edge of Libby.

4.2 Cedar Creek.

4.8 **STOP 11.** Pullout on left. Bonner Quartzite and Libby Formation.

Road cuts to the west are in the west-dipping flank of a fold beneath the Snowshoe thrust.

Exposures at the east end of the road cut are in the siltite facies of the Bonner Quartzite. Here, only a few beds remain of blocky, pink cross-bedded quartzite typical of the Bonner in the Missoula area. The bulk of the Bonner, which is about 1,000 ft thick in this area, consists predominantly of generally even, parallel beds showing parallel to wavy laminae of purple and minor green siltite, argillitic siltite, and argillite. Sparse quartzite channels can be found.

The Bonner is interpreted as an alluvial apron that prograded north and northwest across the Belt basin and then retreated (Winston and others, 1986). These outcrops represent the more distal facies of that apron.

Contact between the Bonner siltite facies and the overlying green to black argillite of the Libby Formation is one of the sharpest in the Belt stratigraphic sequence and is fully exposed and obvious in these road cuts. Here in the Libby thrust belt, the contact is a sheared zone containing obvious folds and faults, but little or no rock appears to be missing.

The Libby Formation is about 5,600 ft thick in this area. A typical section of the Libby (Gibson, 1948) is exposed to the north across the Kootenai River on Flagstaff Mountain, where it has been measured and studied by Kidder (1988, 1992). Kidder has divided the Libby into eight informal members of which all of members 1 and 2 (1 being the bottom member) and part of member 3 are exposed in these road cuts.

The basal member is dark-gray to black argillite thinly interlaminated with dark-green siltite layers that have sharp bases and graded tops. Laminae are predominantly parallel but are wavy in places. Concretions and pyrite occur sparsely. Member 1 is about 180 ft thick.

Member 2 is generally parallel-laminated alternating light-green and dark-green siltite. Sedimentary structures include fluid-escape structures, shrinkage cracks, small-scale cut-and-fill, and flat rip-up clasts of siltite or rare chert. Member 2 is about 280 ft thick.

Member 3, part of which is exposed at the west end of the road cut, is similar to the lower two members but is distinguished by its carbonate content, which occurs mainly as stromatolites, oolites, and brown-weathering carbonate cement. Soft-sediment folds and disrupted beds occur sparsely. Member 3 is about 950 ft thick on Flagstaff Mountain.

These lower members of the Libby Formation plus member 4 are the stratigraphic correlatives of the red and green cherty McNamara Formation to the south and east. Member 4 (not exposed at **STOP 1**) is a greenish-gray silty arsillite member that also correlates to the NcNamara Formation.

The Libby Formation records a shoreline advance over shallow-water to possibly subaerially exposed facies in the Bonner Quartzite (Kidder, 1992). The basal laminated dark-gray argillite records the maximum expression of this event before an influx of silt initiated a shallowing process through members 2 and 3. Stratigraphically higher members on Flagstaff Mountain, across the Kootenai River, record a deep-water environment and a waning of detrital influx. Thick, hummocky, cross-stratified siltstone and very fine-grained sandstones mark a major interval of storm deposition, the deposits of which can be traced as far southeast as Bonner, Mont., where similar deposits occur in the Garnet Range Formation near the top of the Missoula Group.

5.8 Crossing Snowshoe thrust onto Sylvanite anticline.

6.0 Libby Formation in road cuts.

7.0 More Libby Formation.

7.9 Bonner siltite facies in road cuts.

11.1 **STOP 12.** Paved parking lot on right side. Suspension footbridge over Kootenai River below west end of parking lot. Tan beds near stream level are stromatolites in a bioherm in member 4 of the Mount Shields Formation. Several hundred yards upstream on the far side is domal part of bioherm projecting into river.

*****************OPTIONAL****************

Optional side trip to see bioherm and member 4 of the Mount Shields (about 1.5 mi and 1.5 hours).

Take asphalt path eastward through campground to Kootenai Falls viewpoint. Gravel trail to left leads to pedestrian bridge over railroad tracks. About 200 ft past the bridge, the trail forks with the left fork leading to footbridge over Kootenai River. Stromatolite bed is well exposed to left of far bridge abutment. The stromatolites are stacked ellipsoidal heads 1-2 ft across and 3-5 ft high. Measurements of the elongation of the stromatolites and directions of asymmetry on linear ripple marks in beds above the stromatolites have been made and interpreted by C.A. Wallace, U.S. Geological Survey. Stromatolites on the south side of the footbridge are elongated along axes that trend N. 25° W. ±10° and those north of the footbridge along a N. 32° W. ±17° direction (statistically essentially the same). Asymmetric linear ripple marks north of the bridge show a transport direction of N. 30° W. ±70°. Both the elongation of the stromatolites and the ripple marks suggest that northwest currents strongly influenced the growth directions of the stromatolites.

<u>Please</u> do not hammer on these classic exposures!

Return to fork in gravel path and take right fork upstream towards Kootenai Falls. Gravel ends in a few hundred feet at "fisherman trails." Keep walking eastward to Kootenai Falls, where outcrops below falls display abundant ripple marks and salt casts in lower part of member 4 of the Mount Shields. This member, about 400 ft thick in this area, consists predominantly of green dolomitic silty argillite.

C.A. Wallace has also measured the asymmetry of linear ripple marks in these beds. The current direction indicated is N. 80° E. ±43° for the most abundant ripple type.

****

The main purpose of **STOP 12** is to view the road cuts to the west. The Mount Shields Formation on the Sylvanite anticline consists of 6 informal members (fig. 8). Stratigraphic correlations across the region and brief lithologic descriptions are given in figure 8 and by Harrison and others (1992, sheet 1, fig 2). At this locality, member 2 of the Mount Shields is thrust over member 4 of the Mount Shields, probably on a splay from the Moyie thrust. The thrust is below road level but is exposed in the outcrops just above river level on the north wall of the canyon. High-angle faults bound the thrust and create a local window that is too small to show at the scale of the Kalispell quadrangle (Harrison and others, 1992, sheet 1). The folds, cleavage, and small faults in the road cut are upper-plate reflections of the thrust.

At the east end of the road cut are spectacular stromatolite beds that persist over thousands of square miles at the top of member 2 of the Mount Shields, which is predominantly a planar-laminated pink (or rarely

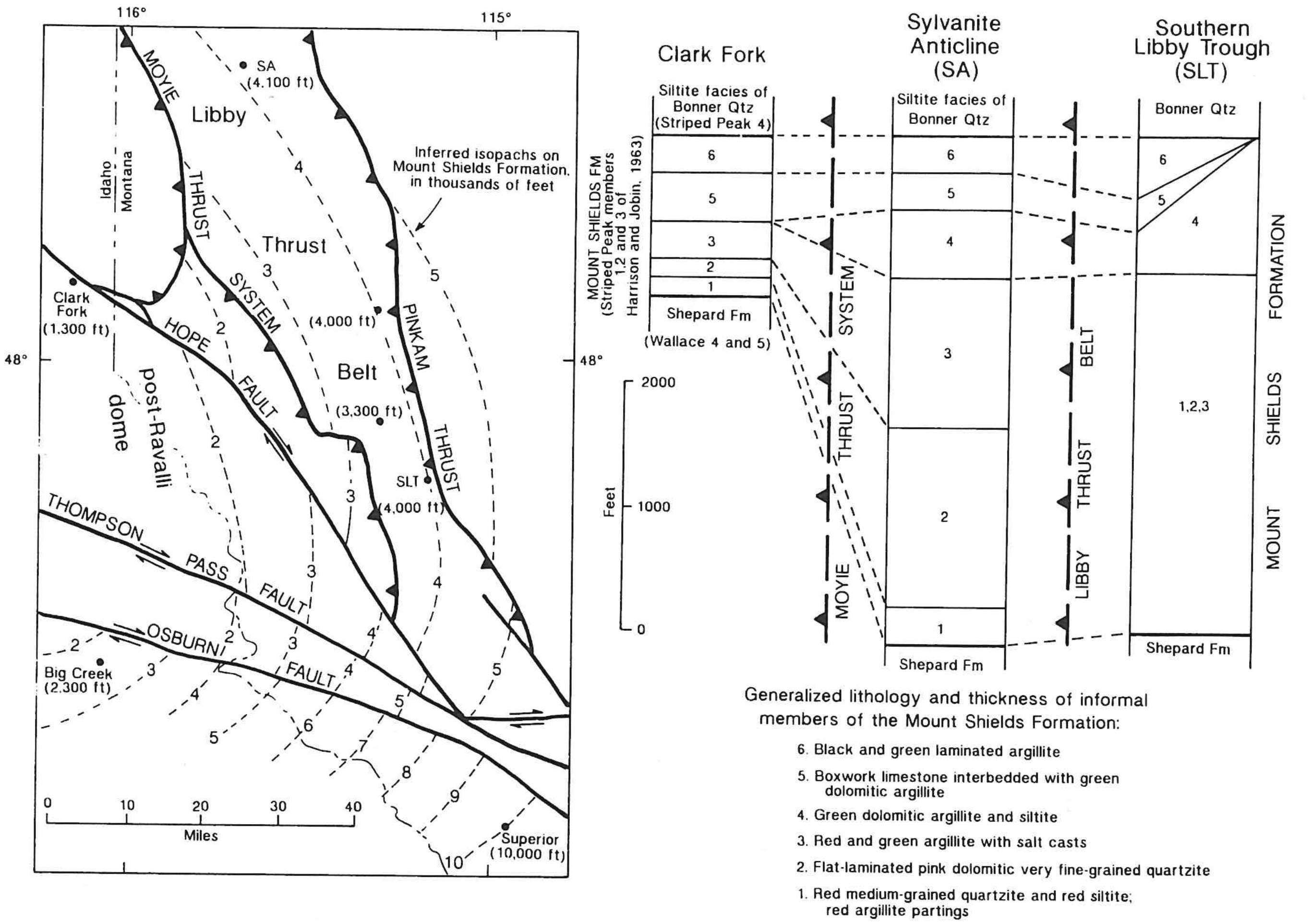

Figure 8. Thickness variations within members of the Mount Shields Formation, and inferred isopach map of the entire formation in northwestern Montana and northeastern Idaho. Numbers in parentheses indicate local thicknesses of the formation.

green), dolomitic, very fine grained quartzite and coarse-grained siltite that is exposed at **STOP 13.** The stromatolite beds are interbedded with quartzite of member 2 and with red argillite characteristic of the overlying member 3 of the Mount Shields. Where the rock steepens at the west end of the exposure, four stromatolite beds in a typical interbedded transition zone between two Belt units are clearly visible.

The covered interval at the west end of the exposure conceals a normal fault that is down to the west (a typical Tertiary extension fault), repeats some of the stromatolite beds to the west, and drops the thrust below the topography.

11.9 **STOP 13.** Paved pullout on left. Member 2 of the Mount Shields.

The east end of the road cut exposes a repeat of the stromatolite transition zone at the top of member 2 of the Mount Shields (fig. 9), and the remainder of the exposure is typical of member 2, which is about 1,400 ft thick in this area. The rock is predominantly very fine grained quartzite that contains brown-weathering carbonate streaks along bedding and sparse climbing ripples (fig. 10). Sparse red to purple argillite beds are present.

Combined, the six members of the Mount Shields Formation form a definitive facies pattern, from coarse-grained red beds at the base to fine-grained black beds at the top, all members containing minor to dominant carbonate. The lower members thin and disappear to the east, west, and north. The entire formation is an excellent display of coarse-grained oxidized facies that fan out from a generally southern source terrane and show distal fine-grained reduced facies northward (Harrison and others, 1992, sheet 1, fig. 2). This simple pattern is disrupted somewhat by thinning of some units where the formation passes over a high in the basin floor near Clark Fork, Idaho, called the "post-Ravalli dome" (Harrison, 1972, p. 1227-1229; fig. 8). A detailed

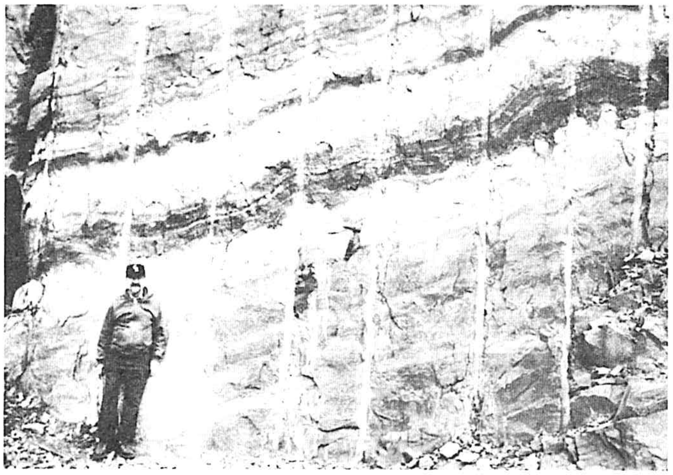

Figure 9. Lowest three of nine stromatolite beds in upper part of member 2 of Mount Shields Formation. Photo by Paul K. Link.

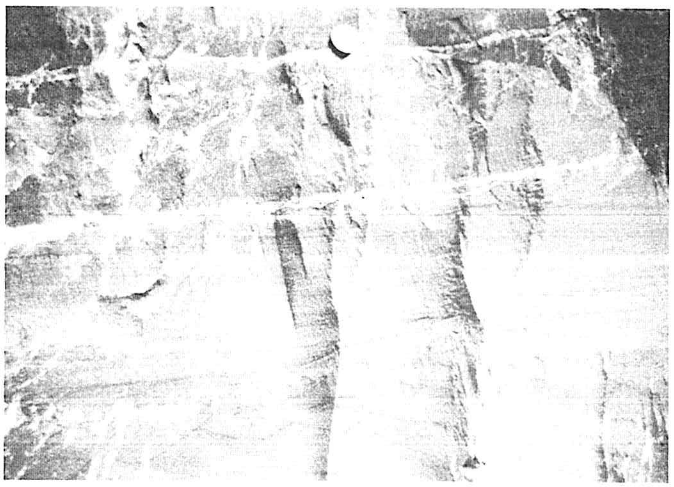

Figure 10. Flat-laminated quartzite displaying climbing ripple marks (lower center of photo) in member 2 of Mount Shields Formation. Photo by Paul K. Link.

analysis of the lower two members by Slover and Winston (1986) concludes by noting that this lower part of the Mount Shields probably represents a fluvial system responding to tectonic uplift in the south and southwest borders of the Belt basin.

12.7 **STOP 14.** Gravel pullout on right. Shepard Formation.

The Shepard Formation consists of a variety of lithologies, generally carbonate-bearing, that vary significantly in abundance from place to place. Thickness also varies by several hundred feet in the area; here the formation is about 2,800 ft thick, but in much of the surrounding area it is only about 2,000 ft thick. Part of this variation may be due to unrecognized faulting in this heterogeneous formation that contains few marker beds.

The predominant lithology in the road cut is green to gray, dolomitic, argillitic siltite in even-parallel to wavy laminae that show graded couplets. This lithology is characteristic of the Shepard; it weathers into orange-brown platy slabs that are displayed toward the east end of the exposure where a low cut is topped by Shepard talus. Pyrite cubes an inch or more across are common in some beds. Less abundant are white quartzite beds, less well-laminated gray argillitic siltite, and laminated green argillite. In places, the beds rich in carbonate also contain small molar-tooth structures, pinkish carbonate laminae, and horizontal gray carbonate pods. Also typical of the Shepard, but not shown at this exposure, are beds of stromatolites and oolites, and red laminated argillite and siltite in beds as thick as 30 ft.

The Shepard Formation has received scant attention to date. The formation has increasing amounts of quartzite to the north and east, loses its red beds to the west where increasing abundance of black argillite occurs, and thins to zero south of Missoula. These facies suggest sediment transport downslope to the west.

13.1 **STOP 15.** Gravel pullout on right near access road across railroad tracks. Green facies of the Snowslip Formation.

Walk west along the railroad tracks while being aware that a metal strap holding bundles of lumber together on railroad flatcars may have broken and will act as a scythe to the unwary as a train passes!

Low cuts along the railroad tracks expose typical beds of the green facies of the Snowslip Formation, which is about 2,500 ft thick in this area. The formation here consists predominantly of two lithofacies: (1) beds of dark-green and light-green laminated argillite and siltite in graded couplets interspersed with layers of green siltite, and (2) pyritic black argillite interlaminated with white or green siltite. The green lithofacies is most abundant and is commonly slightly dolomitic or pyritic and also commonly displays chlorite on bedding surfaces. The black beds are normally tens of feet thick, alternate with green lithofacies, and are similar to the marker interval in the Snowslip seen at **STOP 10**. One slightly pink carbonate-bearing quartzite bed on the north side of the tracks just before the curve may be the last remnant westward of an eastern red-bed facies in the Snowslip.

Laminated black argillite beds of similar lithology and sedimentary features are found in the Wallace, Helena, Snowslip, Shepard, Mount Shields, and Libby Formations in the Kalispell 1° x 2° quadrangle and adjacent areas. No certain lithologic or chemical qualities have been found as yet to distinguish which formation a small outcrop of these beds belongs to except for those with chert, which is only found in some beds of the Libby.

Turn around and return to Libby (or go to rest area 2.2 mi west at junction of Highways 2 and 56 before heading back east).

<u>Libby to Whitefish</u>

Mileage

0.0 Stoplight at junction of Highways 2 and 37. Take Highway 2 east to Kalispell.

The next 30 mi traverses southward along the center of the Libby thrust belt. The structure is complex and represents a Proterozoic syncline that has been crushed during Cretaceous thrusting between the Sylvanite anticline to the west and the Purcell anticlinorium to the east (Harrison and others, 1992, sheet 1, sections A-A' and B-B'). The belt is characterized by multiple Cretaceous thrust faults and numerous Tertiary high-angle extension faults that have left a jumble of slivers and pieces of Belt strata (Harrison and Cressman, 1993, sheets 1 and 2). The trough of the old syncline preserves isolated blocks of the overlying unconformable Middle Cambrian strata, which are exposed in a few places in the Kalispell 1° x 2° quadrangle and more extensively south of the quadrangle in the Fishtrap Creek area of the Wallace 1° x 2° quadrangle (Harrison and others, 1986, sheet 1).

Within the Libby thrust belt, which is defined loosely as the zone of thrusts and high-angle faults within and bounding the Proterozoic Libby syncline, tectonic shortening on folds and thrust faults is about 10 mi, and extension on Tertiary high-angle faults is about 3 mi (Harrison and Cressman, in press). Thus, rocks of the area, excluding those on the Moyie plate, are near their relative depositional positions in the original Belt basin.

0.8 Buff-colored cliffs to east are lake beds of Glacial Lake Kootenai (Alden, 1953).

1.4 Heritage Museum.

3.0 Granite Creek.

3.5 Junction with Highway 482--airport road.

4.4 Leaving greater Libby. Road climbs up on top of lake beds.

4.6 View of Cabinet Mountains to west.

8.2 Bear Creek road.

12.0 Libby Creek road.

12.7 Libby Creek.

13.0 Hills on east and west contain Cambrian strata.

13.8 Road cuts in Cambrian dolomite.

15.4 More Cambrian dolomite.

16.6 Crossing complex high-angle fault and thrust zone from Cambrian dolomite into Mount Shields Formation.

17.0 Narrow valley of Swamp Creek follows thrust fault.

20.7 **STOP 16.** Road to Schreiber Creek on right, almost across from Tepee Creek road. Turn right. Bus turn-around immediately past junction at old logging road; cars can proceed up Schreiber Creek road for about 0.4 mi to car turn-around about 200 ft short of the first bedrock outcrop.

A stop for the skeptical who may be wondering what kind of structural evidence exists for thrust faults in the area. These outcrops expose the stromatolite zone at the top of member 2 of the Mount Shields. Dip of the beds is about 50° to the west and, despite somewhat confusing evidence from crossbeds, the stromatolites are clearly upside down. The nearest outcrops to the east in the narrow valley are right-side-up, east-dipping Bonner and McNamara (Harrison and others, 1992, sheets 1 and 2). Geometric relations require a steep thrust fault in the narrow valley containing Highway 2.

Return to Highway 2.

22.2 Bonner Quartzite.

22.9 Road drops down through lake beds of Glacial Lake Kootenai.

24.8 Miller Creek road.

25.0 West Fisher Creek road.

25.5 Fisher River.

29.1 Silver Butte road. Sweeping left turn through lake beds.

29.6 Houghton Creek road. The highway now heads east crossing the Libby thrust belt toward Kalispell. Wapiti Mountain, directly ahead, is largely formed from the chevron-folded Shepard Formation at the base of a thrust plate.

For about the next 7 mi you are driving past the remains of a devastating forest fire started by lightning in the late summer of 1984. Most of the tree stumps are from timber salvaging operations that began as soon as the fire was out. Surviving structures are a result of valiant efforts by firefighters and luck.

35.1 Crystal Creek road.

36.3 Crossing lead thrust fault in Libby thrust belt. Purcell anticlinorium to east.

37.0 Libby Formation in road cuts.

37.6 Road to McKillop Creek.

38.0 Loon Lake.

39.8 Pleasant Valley Fisher River.

40.9 Happy's Inn.

41.0 Pleasant Valley road.

44.3 Private logging road to Thompson Falls.

44.8 Glacial lake beds.

45.2 Thompson Lake.

46.0 Logan State Park.

48.1 Flat-lying Helena Formation.

49.0 Meadow Peak road.

49.3 Flathead-Lincoln County line.

50.2 Thompson River road.

52.0 About 14 mi due north, on the crest of the Purcell anticlinorium (Harrison and others, 1992, sheet 1), is the Arco-Marathon Gibbs No. 1 borehole (fig. 1). The hole is collared in the Prichard about 4,000 ft below the formation's top, and total depth of the nearly vertical hole is 17,774 ft (Harrison and Cressman, in press, fig. 11). The hole was drilled to explore a high density, nonmagnetic, seismic-reflective, anticlinal zone interpreted to be potential hydrocarbon-bearing, dense carbonate beds of Paleozoic and Mesozoic age, possibly in a structural duplex (a typical thrust-belt oil or gas trap). The hole did not completely penetrate the Prichard Formation, which included about 4,500 ft of altered mafic sills (Harrison and others, 1985; Harrison and others, 1992, sheet 1, cross section B-B'; Harrison and Cressman, in press, fig. 11). The anticlinal structure is part of the broad open Proterozoic fold system that is outlined by surface geology and mimicked by the seismically reflective sills at depth. The gravity (density) high is in part caused by the sills instead of the hoped-for carbonate beds but is also probably due in part to a basement core in the old folds. A lack of magnetic properties is the result of a known widespread autoalteration of mafic sills that destroyed most of their original magnetite, as well as a generally nonmagnetic basement and the Prichard Formation in this area (Harrison and others, 1980, p. 410). Small amounts of methane encountered at a few places in the borehole are more of a novelty than a resource and may represent survival of gas derived from carbon-bearing sediment (black shale) of the Prichard Formation.

Figure 11. Mud cracks and fluid-escape structures in red to purple laminated argillite and siltite of upper part of Spokane Formation. Photo by Paul K. Link.

53.2 Bar Z Mountain road.

54.5 McGregor Lake road.

56.0 Burke outcrop.

56.4 Lost Prairie road.

58.5 Middle member of Burke Formation a few hundred feet below contact with upper member.

60.0 North Fork Murr Creek road.

65.5 Little Bitterroot River.

70.0 Road to Marion and Little Bitterroot Lake.

73.6 Mountain to right has dip slope in the Spokane Formation.

74.7 Road to Ashley Lake.

75.6 Road to Lake Rogers.

77.4 **STOP 17.** Gravel pullout on right. Spokane Formation.

This exposure is on the crest of an anticline of the Purcell anticlinorium and is about 40 mi south of **STOP 2** at Stryker. The rock is the upper member of the Spokane as seen at Stryker, but here the laminated purple carbonate-bearing argillite and siltite (fig. 11) has no beds of the distinctive well-rounded, mud-chip bearing, crossbedded white quartzite seen at Stryker. The white quartzite beds, which occur in the upper and lower members, gradually thin and disappear to the south and west of Glacier National Park. The white quartzite, and perhaps the entire Spokane, has a Canadian Shield source.

77.7 Contact between the Spokane and Empire Formations.

79.7 Road cuts in Empire Formation.

80.2 Highly fractured main body of the Helena Formation.

81.6 Road to Kila.

85.1 Smith Valley Primary School.

87.1 Looking eastward across Rocky Mountain Trench to east wall formed by Swan Range.

89.6 Welcome to Kalispell.

90.8 Junction with Highway 93. Turn left onto Highway 93 to Whitefish. Heading north up Rocky Mountain Trench, which is filled by Pleistocene glacial deposits.

92.7 Flathead Community College.

93.8 Junction with Highway 548.

94.1 Stillwater River.

95.7 Gap in the mountains to the east is Badrock Canyon leading to Glacier National Park. Directly ahead is southern end of Whitefish Range. Ski runs are on Big Mountain, which has Spokane at the bottom and Helena at top.

103.6 Junction with Highway 10.

104.7 Entering Whitefish.

End of road log.

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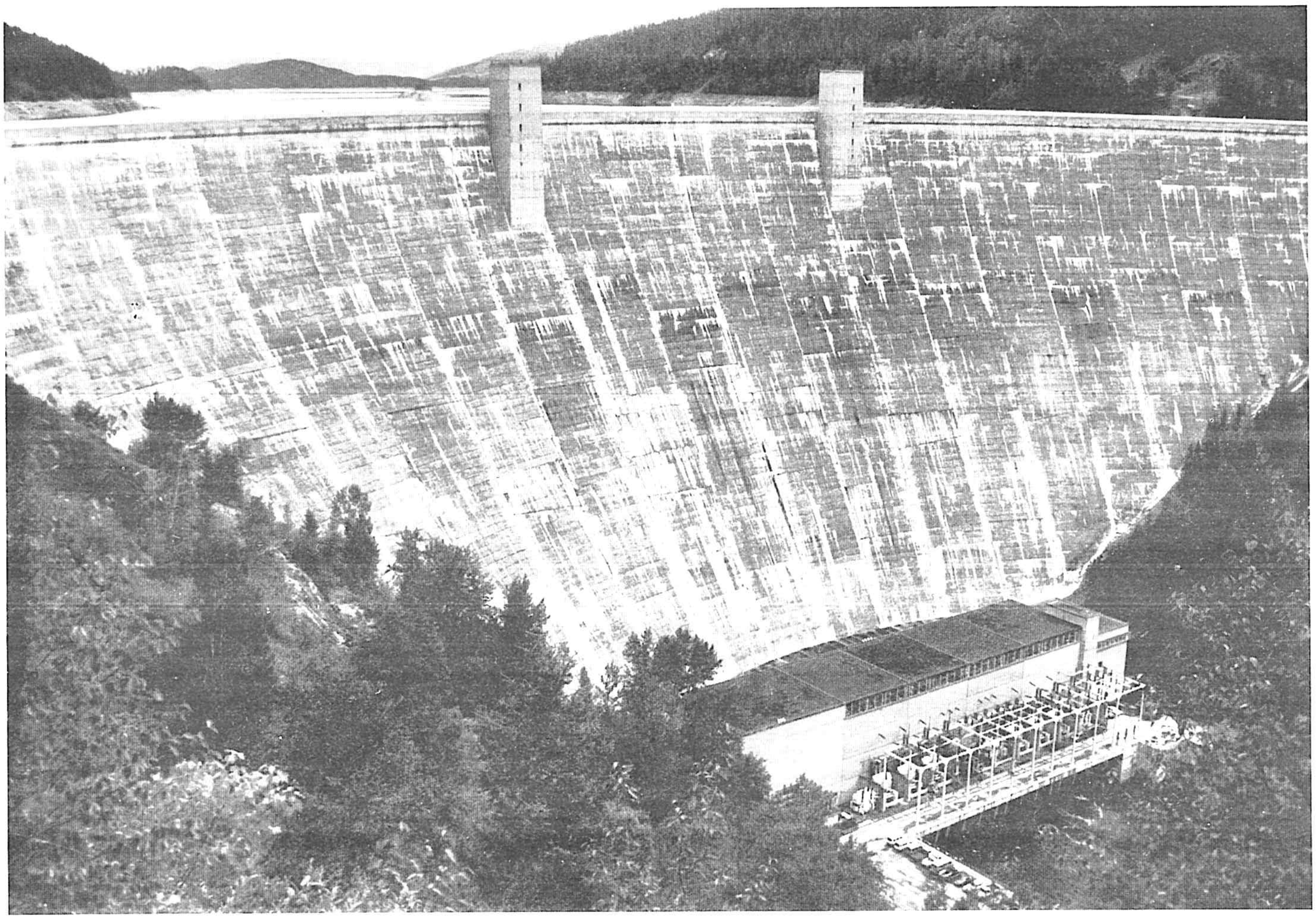

Hungry Horse Dam, on the South Fork of the Flathead River. *Dam is 564 feet high and was completed in 1952.*

SEDIMENTARY CYCLES IN THE ST. REGIS, EMPIRE AND HELENA FORMATIONS OF THE MIDDLE PROTEROZOIC BELT SUPERGROUP, NORTHWESTERN MONTANA

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INTRODUCTION

One of the most striking attributes of rocks deposited in the Belt "sea" is their cyclical character. In the lower Belt, White (1977) described cycles from the Altyn Formation and Zieg (1981) described cycles from the upper Newland Formation. The middle Belt carbonate is also impressively cyclic. Gray weathering siliciclastic layers, a few feet to tens of feet thick, alternate with equally thick tan weathering dolomitic layers through thousands of feet of section and characterize the Helena Formation across the eastern part of the Belt basin. O'Connor (1967, 1972) and Eby (1977) analyzed cycles from the Helena and Grotzinger (1981, 1986) studied cycles in the Wallace Formation. New studies of cycles in the middle Belt carbonate are reported in Belt Symposium III (Johnson, 1993, Winston, 1993).

Stratigraphic, sedimentologic and geochemical analyses of these cycles offer perhaps the best opportunity for answering the important question of whether the Belt "sea" was connected to the open Proterozoic ocean or whether it was landlocked like the

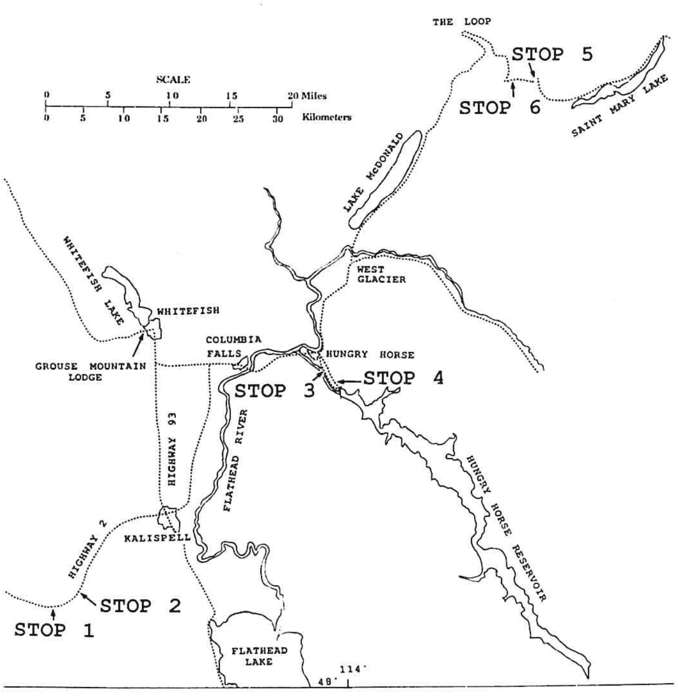


Figure 1. Map showing field trip route and stops.

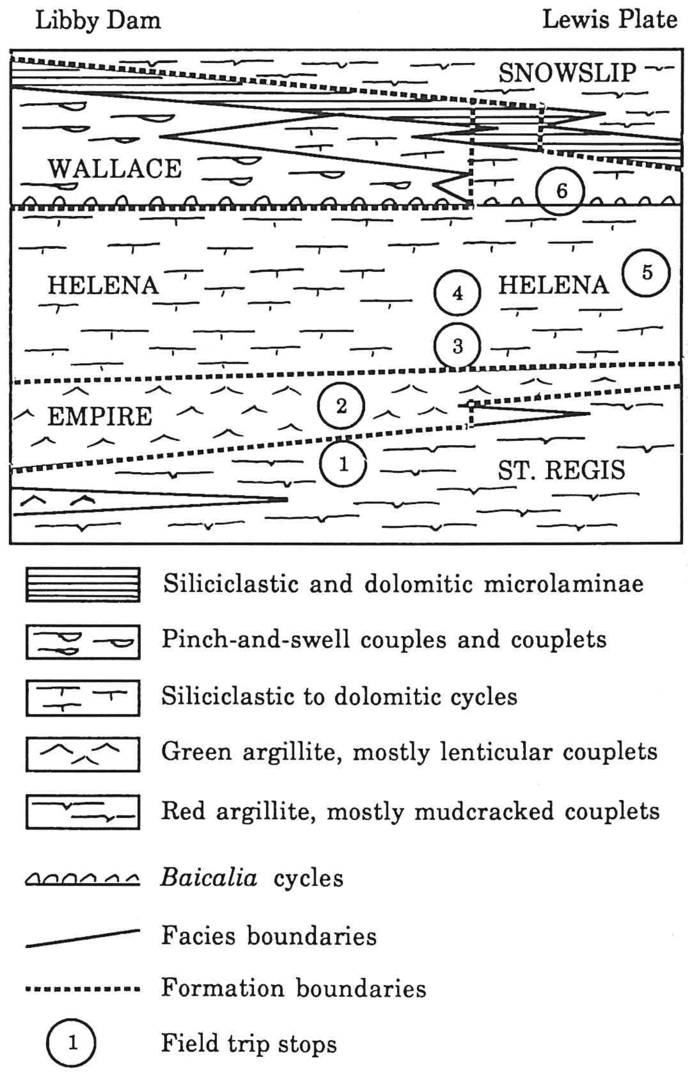


Figure 2. Schematic east-west cross section from Libby Dam to Glacier National Park, showing facies relationships and formational boundaries of the upper St. Regis, Empire, Wallace, Helena and the lowermost Snowslip formations. Figure not to scale. Circled numbers show approximate location and stratigraphic position of field trip stops.

Winston, Don, and Lyons, Timothy, 1993, Sedimentary cycles in the St. Regis, Empire and Helena Formations of the Middle Proterozoic Belt Supergroup, northwestern Montana, *in* Link, P.K., ed., **Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada**: Belt Symposium III Field Trip Guidebook, Belt Association, Spokane, Wa., p. 21-51.

present Caspean Sea. The thesis presented here supports the proposition that the Belt "sea" was landlocked and could more properly be called the Great Belt Lake.

Cycles on this field trip (Fig. 1) will be studied through a vertical succession of six stops beginning in the upper part of the St. Regis Formation, continuing into the Empire Formation, lower and middle parts of the Helena Formation, and ending in the upper part of the Helena Formation (Fig. 2).

SEDIMENT TYPES

Sedimentologic study by Winston and his co-workers has been based on the recognition and analysis of sedimentary lithofacies, termed sediment types, that occur in repeated associations through the Belt. Sediment types are descriptively defined on the basis of sedimentary structures, grain size and inferred original composition. Ongoing study shows that most Belt strata can be classified into one of about 20 sediment types (Winston, and Link, 1993). Some, such as the bouldery mud of the LaHood Formation are limited to a single formation, whereas others, such as the mudcracked even couplet type, occur in association with other sediment types from the lower Belt through the upper Belt. Sediment types are the basic building blocks with which Belt rocks have been analyzed. For a fuller discussion of sediment types see Winston and others (1984), Winston (1986, 1989, 1990), and Winston and Link (1993). Ten sediment types are germane to this field trip. They are: the mudcracked even couplet, mudcracked lenticular couplet, uncracked even couple, uncracked even couplet, uncracked lenticular couplet, microlamina, pinch-and-swell couple, pinch-and-swell couplet, carbonate mud and the coarse sand and intraclast types. These sediment types are schematically sketched and summarized in Figure 3. Note that couples are graded layers 3-10 cm thick, whereas couplets are 0.3 to 3 cm thick. Symbols of sediment types and other lithic constituents in subsequent figures are illustrated in Figure 4

Mudcracked Even Couplet Sediment Type

This sediment type contains flat, continuous layers of very fine-grained sand or silt that grade up to clay (now sericite). The capping clay layers are mudcracked, and transported mud polygons commonly occur as mudchips in the lower sandy half-couplets. Graded layers range from 0.3 cm to 3 cm thick. Some intervals are wholly siliciclastic, whereas others contain carbonate mud (now microcrystalline dolomite and calcite) in the graded siliciclastic sand-to-clay layers.

These graded layers record episodic flood transport and deposition of silt followed by decelerating flow or cessation of flow and accumulation of clay or clay and carbonate mud. After deposition the sediments were exposed and desiccated. Stratigraphic relations in the Ravalli Group show that this sediment type lies lakeward of the mudcracked even couple sediment type, attributed to sheetflood deposition on alluvial sandflats at the toes of alluvial aprons that sloped eastward into the basin. (Winston, 1986, 1989, 1990; Winston and Link 1993). From the alluvial apron and sandflat prospective, the mudcracked even couplet sediment type represents deposition from sheetfloods that flowed from the alluvial aprons and sandflats eastward across playa mudflats and fringing lacustrine mudflats of the perennial Great Belt Lake. As sheetfloods flowed across the exposed mudflats, they deposited fine sand and silt of the lower half-couplets and probably some of the clay in the upper half-couplets. Flood waters trapped in the enclosed lake spread across the flats, permitting suspension settleout of the rest of the clay in the upper half-couplets. As lake water evaporated, the playa or lake mudflat margins were once more exposed and awaited the next flood. Mudcracked and uncracked even couplets characterize modern playas and are nearly identical in form and inferred origin to deposits of the sheetwashed mudflat environment of Hardie and others (1978).

Mudcracked Lenticular Couplet Sediment Type

This sediment type is also characterized by mudcracked, couplet-scale (0.3-3.0 cm) commonly graded silt-to-clay layers with sharp, flat bases. Unlike the even couplet type, silt layers of the lower half-couplets thicken into ripple crests and thin into ripple troughs. They commonly display symmetrical ripple cross-laminae. In some places the upper clay half-couplets rest sharply on the silt and are not graded.

This sediment type records wave reworking of silt and clay into ripples followed by clay suspension settleout. Water then evaporated and the sediments were desiccated. Mudcracked even and lenticular couplets are commonly interlayered: the silt and clay may have been originally transported into the basin by sheetfloods and deposited as even couplets before reworking. This sediment type, like the mudcracked even couplet type, was probably deposited on lacustrine mudflats, but closer to the perennial lake, where the mudflats were submerged for longer periods of time and were reworked by waves.

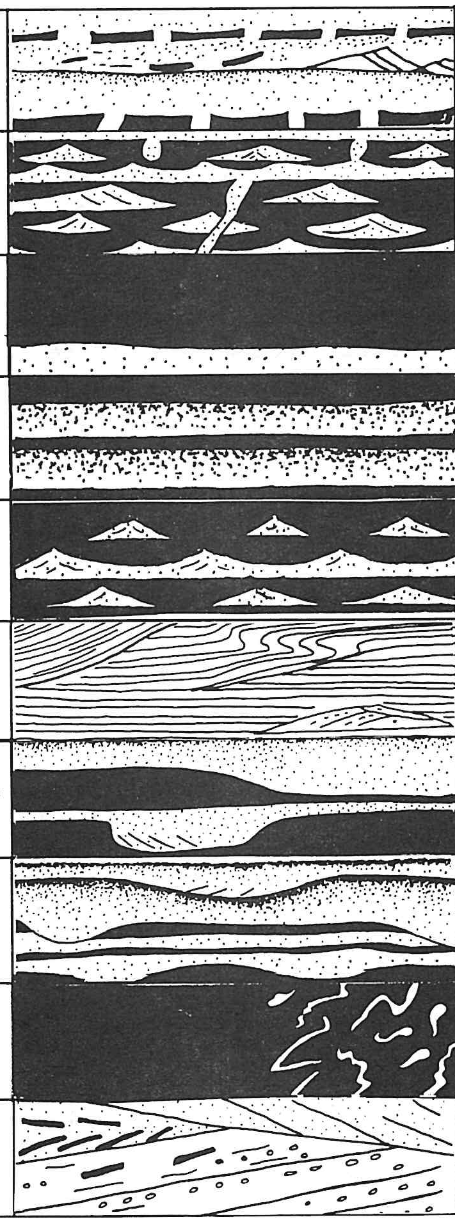
Uncracked Even Couple Sediment Type

In the rocks discussed here, the even couple sediment type contains continuous, uncracked graded silt-to-clay (sericite) or silt-to-clayey dolomite layers from 3 cm to 10 cm thick in which the clayey layers are thicker than the silty layers. They are most common in the upper parts of some Helena cycles and probably record episodic storm deposition followed by long periods of suspension settleout (even couples in other Belt formations have other origins).

Uncracked Even Couplet Sediment Type

Even uncracked couplets are graded, silt-to-clay couplets, similar to the mudcracked even couplets, but tend to be finer grained and lack desiccation cracks. The sediment type has both wholly siliciclastic and carbonate mud varieties.

Sediment of the wholly siliciclastic uncracked even couplets probably came from sheetfloods that crossed the exposed mudflats and entered the perennial Great Belt Lake, forming suspension clouds that settled out into graded layers. Silt-to-clay couplets containing carbonate mud, now in the form of microcrystalline calcite and dolomite, probably represent carbonate-bearing lake floor sediment that was suspended by storms into muddy clouds that settled out into graded layers.



MUDCRACKED EVEN COUPLET	Even, mudcracked, graded, fine sand and stlt-to-mud layers 0.3 to 3 cm thick.	Sheetflood flow across exposed mudflats followed by deceleration, suspension settleout and desiccation.
MUDCRACKED LENTICULAR COUPLET	Oscillation-rippled fine sand and silt lenses, capped by clay laminae, cut by mudcracks.	Wave transport of fine sand and silt, followed by clay settleout and desiccation.
UNCRACKED EVEN COUPLE	Graded silt-to-clay or silt-to-dolomitic clay layers 3 to 10 cm thick	Episodic storm deposition followed by suspension settleout.
UNCRACKED EVEN COUPLET	Even, uncracked graded silt-to-clay couplets 0.3 to 3 cm thick.	Episodic suspension transport and settleout.
UNCRACKED LENTICULAR COUPLET	Oscillation-rippled fine sand and silt lenses, capped by clay laminae, cracked and uncracked.	Wave transport of fine sand and silt, followed by suspension settleout.
MICROLAMINA	Interlayered or graded silt and clay laminations less than 0.3 cm thick.	Alternating silt and clay suspension settleout.
PINCH-AND-SWELL COUPLE	Graded fine sand to dark mud layers with undulating scoured and loaded bases >3 cm thick.	Episodic transport of fine sand and mud by turbidity flows or storms and deposition on scoured or loaded mud surfaces.
PINCH-AND-SWELL COUPLET	Graded fine sand to dark mud layers with undulating scoured and loaded bases, 0.3 to 3 cm thick.	Episodic transport of fine sand or mud by turbidity flows or storms and depositionon scoured or loaded mud surfaces.
CARBONATE MUD	Micrite and dolomicrite without detectable siliciclastic laminations.	Aragonite or calcite precipitation, in places followed by dolomitization.
COARSE SAND AND INTRACLAST	Coarse- to fine-grained, sand and flat clasts, crossbedded and imbricated at various angles.	Transport of coarse sand grains and scoured clasts by breaking waves.

Figure 3. Schematic sketches at varying scales of St. Regis, Empire and Helena sediment types and summaries of their descriptions and interpretations.

Uncracked Lenticular Couplet Sediment Type

This, like the mudcracked lenticular couplet sediment type, is characterized by flat-bottomed symmetrical straight-crested rippled silt layers capped by mud. However, they lack desiccation cracks.

Like the mudcracked lenticular couplet type, this type is interpreted to record wave reworking of flood-derived sediment by fair weather waves in shallow water. Unlike the mudcracked lenticular couplet type, this type remained submerged. It compares closely to wave-reworked, deposits of the ponded mudflat environment of Hardie and others (1978).

Microlamina Sediment Type

This sediment type contains alternating extremely thin silt and clay laminae, sandwiched into layers up to 3 mm thick. Details of this sediment type reveal a variety of forms which may reflect several processes and environments. Some microlaminae are extremely fine and even. Others are mudcracked or grade into thin, lenticular couplets. Still others are crenulate.

Interpretation of this sediment type is less straightforward. It is most widespread along the eastern margin of the basin, farthest from sheetflood influx from the south and west. Therefore, some even microlamina intervals could be the most distal deposits of flood-derived suspension clouds. Others could also record storm resuspension and deposition in very restricted or protected shallow water environments along the tectonically stable east side of the basin. The crenulate, carbonaceous variety may have been bound by mucilaginous films of cyanobacteria in environments of very low sedimentation rate.

Pinch-and-Swell Couple Sediment Type

This sediment type is characterized by graded fine sand-to-mud beds more than 3 cm thick in which the lower sandy and silty layers thicken and thin as a result of loading or cutting into the muddy layers beneath them. Laminae of these sandy layers are commonly hummocky cross stratified at a low angle. The sand to mud transitions are more evenly layered than the lower sand bounding surfaces, and the even, continuous capping clay layers are loaded or cut on their upper surfaces beneath the overlying silty layers. Silt-filled ptygmatically folded cracks commonly cut the muddy layers. The pinch-and-swell couple sediment type is striking in outcrop because the light to medium gray sand and silt layers thicken and thin below the darker, more even mud or tan-weathering dolomitic mud layers.

Although Winston (1986) originally interpreted this sediment type to record turbidite deposition, Johnson (1990) correctly pointed out the dominance of hummocky cross stratification and the amount of scour and fill deposition. He has interpreted the sediment type to record storm deposition, and we now concur.

The silt- and sand-filled ptygmatically folded cracks in the muddy layers of the pinch-and-swell couple and pinch-and-swell couplet (see below) sediment types are especially important. We interpret their curved, discontinuous configuration in plan view and their folding by compaction in cross section to indicate that they formed in wet sediments, and not by desiccation as some have proposed (e.g. Demicco and Gierlowski Kordesch, 1986; Rogers and Astin, 1991). Instead we appeal to the suggestion by Martel and Gibling (1991) that sand dikes in hummocky cross-stratified siltstone of the Lower Carboniferous Horton Bluff Formation of Nova Scotia parallel wave ripple crests and were formed by cyclic loading of passing storm waves.

We envision the following genesis of the ptygmatically folded cracks in the pinch-and-swell sediment of the Belt. First, the capping mud layers of the pinch-and-swell couples and couplets accumulated from suspension and formed water-saturated thin layers on the lake floor. Second, as waves of a next storm passed over the muddy layers they lifted and rumpled them, forming tension cracks that were filled by silt entrained by the storm. Finally, hummocky cross-stratified silt and fine sand layers deposited by storms loaded and compacted the muddy layers, folding the sand- and silt-filled cracks.

Ptygmatically folded cracks are associated with rippled and hummocky cross-stratified beds of the Caithness Flagstone Group of the Devonian Old Red Sandstone, Scotland, (e.g. Rogers and Astin, 1991) and the Cretaceous Dakota Sandstone of North Dakota (Shelton, 1962) This association appears to form a storm-generated lithofacies with potentially widespread recognition and importance.

Pinch-and-Swell Couplet Sediment Type

This is simply a thinner, finer-grained counterpart of the pinch-and-swell couple sediment type. Graded layers range from 0.3 to 3 cm thick. Like the pinch-and-swell couples these couplets have wavy lower bounding surfaces but more even graded silt-to-clay layers. Silt characteristically forms a smaller percentage of the rock than in the pinch-and-swell couples.

Since the pinch-and-swell couplet sediment type contains a similar sequence of sedimentary structures to those of the pinch-and-swell couple type, it too is interpreted to record storm deposition below the level of effective fair weather wave base. It is most common along the eastern to central parts of the basin, mostly eastward and distal from the pinch-and-swell couple type. Therefore, it probably records weaker storm erosion, possibly in shallower water where drag on the bottom dampened wave orbital velocities.

Carbonate Mud Sediment Type

The carbonate mud sediment type is composed of beds of massive calcitic micrite or very fine-grained dolomite. Whereas most Belt sediments record episodic siliciclastic deposition, the carbonate mud sediment type records continuous calcitic, aragonitic or protodolomite mud deposition mixed with such small quantities of silt that graded siliciclastic layers are not discernable. The carbonate mud sediment type is most common in the upper parts of siliciclastic-to-dolomitic cycles in the Helena Formation. It probably reflects decrease in siliciclastic influx concurrent with increasing protodolomite precipitation from saline waters during the arid shrinking phases of the Great Belt Lake.

Coarse Sand and Intraclast Sediment Type

Crossbedded coarse- to medium-grained quartz and oolitic sand grains or thin layers of molar-tooth and micrite intraclasts set in a sparry matrix characterize this sediment type. In Belt cycles this sediment most commonly forms thin beds less than a decimeter thick on scoured surfaces at cycle boundaries. In some places it forms lenses within cycles. Oolitic occurrences tend to be thicker and more strongly crossbedded.

As explained below, scoured surfaces at the cycle boundaries are interpreted to record episodic contraction of the lacustrine mudflats and exposure of the lake bottom. As the lake refilled and expanded over the exposed lake bottom and mudflats, waves worked the strand and deposited thin layers of quartz sand and intraclasts. The coarse- to medium-grained sand, carried into the basin from the east, (Winston, 1989) passes westward to molar-tooth fragments reworked from the periodically exposed lake bed. Thicker beds of oolitic quartz sand are interpreted to have formed along the eastern edge of the lake during prolonged lake highstand.

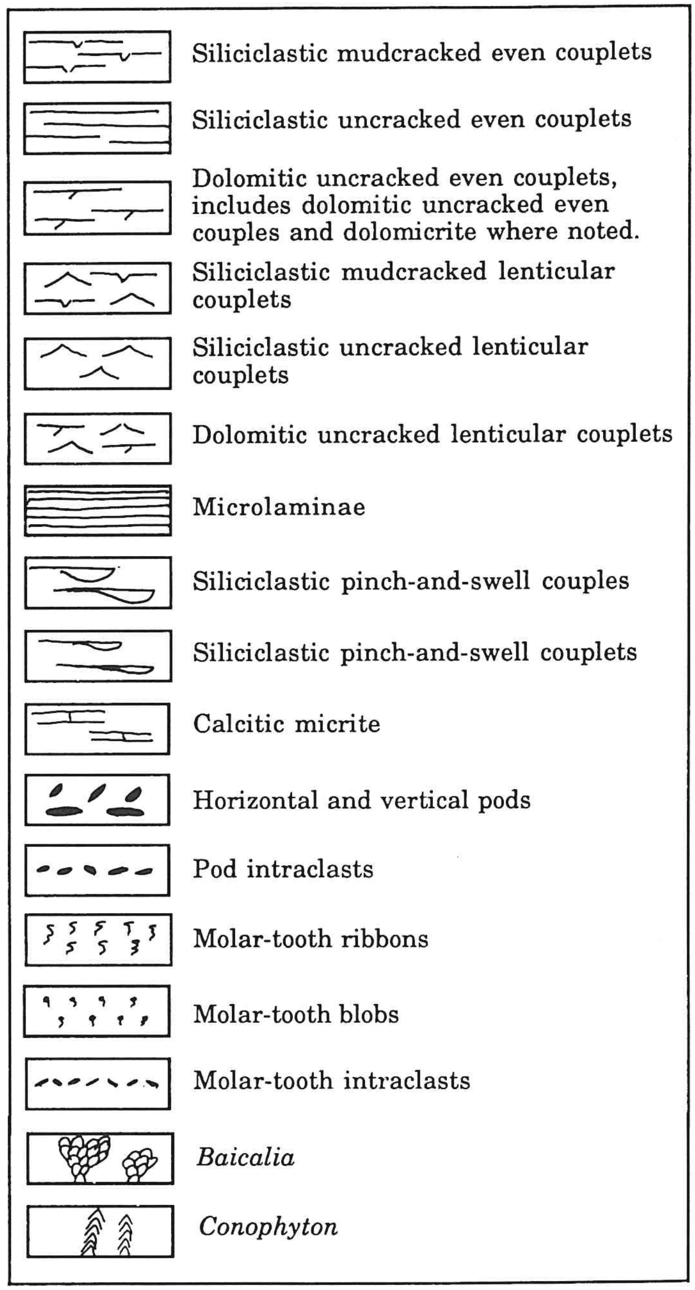


Figure 4. Symbols of sediment types and lithic constituents in subsequent figures.

MOLAR-TOOTH STRUCTURE

Origin of the strange calcitic "molar-tooth" structure has been queried ever since Bauerman (1885) first likened the crenulated ribbons to "markings in the molar tooth of an elephant". More recent interpretations have ranged from biotic (e.g.O'Connor,1967, 1972) to replacement of evaporite minerals (Eby, 1977) to filled open-space structures such as desiccation cracks, syneresis cracks, gas bubbles and fluid voids (Horodyski, 1976, 1983, 1989) to earthquake dewatering structures (Pratt, 1992).

O'Connor (1967, 1972) and Eby described four categories of molar-tooth structure: vertical ribbons, horizontal ribbons, blobs and pods. Vertical ribbons are blade-like structures, filled with uniform, fine-grained blocky calcite crystals 2 to 20 μ across. Vertical ribbons cut through siliciclastic laminae, and laminae above and below commonly drape around the ribbons. Horizontal ribbons are identical to vertical ribbons in composition, but lie parallel to lamination and do not significantly disrupt it. Blobs are ellipsoidal forms less than a centimeter in diameter filled with fine-grained blocky calcite similar to that of the ribbons. They commonly have thin, wiggly cracks projecting from them. Walls of ribbons and blobs are sharp, and we agree with Horodyski (1976, 1983, 1989) that ribbons and blobs represent open-space structures filled with sparry calcite. These are the structures originally noted by Bauerman, and we restrict the term "molar-tooth structure" to them.

Pods, on the other hand, are mostly sheet-like calcitic nodules that commonly contain silt laminae. Therefore, they do not fill open space. Sediment is compacted around pods, and in places pods have been reworked into intraclast beds, indicating that pods probably represent early calcite cement and diagenetic replacement of siliciclastic minerals within centimeters of the depositional interface.

Inferred origin of the ribbons and blobs differs from that of the pods. George Furniss and John Rittle, students at the University of Montana, closely simulated molar-tooth voids by mixing clay, water, sugar and yeast in an aquarium (Furniss, 1990). Fermentation in very soft, water-saturated clay formed gas bubbles that closely mimicked molar-tooth blobs. The gas bubbles rose through the soupy clay and pulled water up behind them, pumping water out of the clay. As the clay lost water, it became too stiff to form bubbles, and gas expansion produced vertical and horizontal cracks that are nearly identical in form to molar-tooth structure Thin cracks projecting from gas bubbles closely simulated ribbons projecting from blobs. Vertical cracks have the same shape and breadth as molar-tooth vertical ribbons and disrupted the surrounding clay. Horizontal cracks simply lifted the clay above them and closely resemble horizontal molar-tooth ribbons.

Based on the exceedingly close replication of molar-tooth forms by gas expansion bubbles and cracks, we interpret molar-tooth blobs and ribbons to have been formed by gas generated in the sediments below the depositional interface. The gas may have been H_2S, generated by sulfate reducing bacteria or methane formed by organic decay, or probably both.

In many places vertical ribbons are broken and offset by compaction of the sediment around them (Smith, 1968), showing that the cracks were filled by calcite before the surrounding sediment was lithified. In addition, molar-tooth fragments form lag concentrations on scour surfaces, indicating that some cracks and bubbles were filled with calcite within centimeters of the depositional interface. Methane and H_2S generation drives the carbonate solution-precipitation equilibrium in the direction of calcite precipitation, and may have promoted rapid calcite filling of the voids.

St. REGIS, EMPIRE AND HELENA CYCLES

The sequence of stops on this field trip illustrates the second great submergence of the Belt basin that began in the upper part of the St. Regis Formation and continued though deposition of the Empire Formation and most of the Helena Formation, through a stratigraphic interval thousands of feet thick in the center of the basin. Superimposed on this large scale submergence are smaller scale cycles from one to 100 feet thick that are separated from each other by scoured surfaces. They have deeper water facies in their lower parts and shallower water facies in their upper parts. And they are inferred to record climactically induced, small scale expansion and contraction of the Great Belt Lake superimposed on the large scale submergence. Sedimentologic style of the cycles evolved through the large scale submergence, so that six types of cycles can be identified (Fig. 5). They began as green-to-purple argillite cycles in the St. Regis Formation and progressed to uncracked to mudcracked green argillite cycles in the Empire Formation. Siliciclastic-to-dolomitic cycles characterize the lower and middle parts of the Helena Formation and culminate in the calcitic *Baicalia-Conophyton* cycles at the base of the upper Helena Formation. The six cycle types are described below and illustrated in Figure 5.

Type 1 Cycles

Description.- Type 1 cycles are characterized by sharp basal contacts overlain by thin layers of medium-grained, ripple-crossbedded quartz arenite, belonging to the coarse sand and intraclast sediment type. These arenite layers are overlain by a meter or two of green, siliciclastic, uncracked lenticular couplets, which pass sharply upward to comparably thick intervals of purple, siliciclastic, mudcracked even couplets, capped by the sharp base of the next cycle. Type 1 cycles are illustrated at

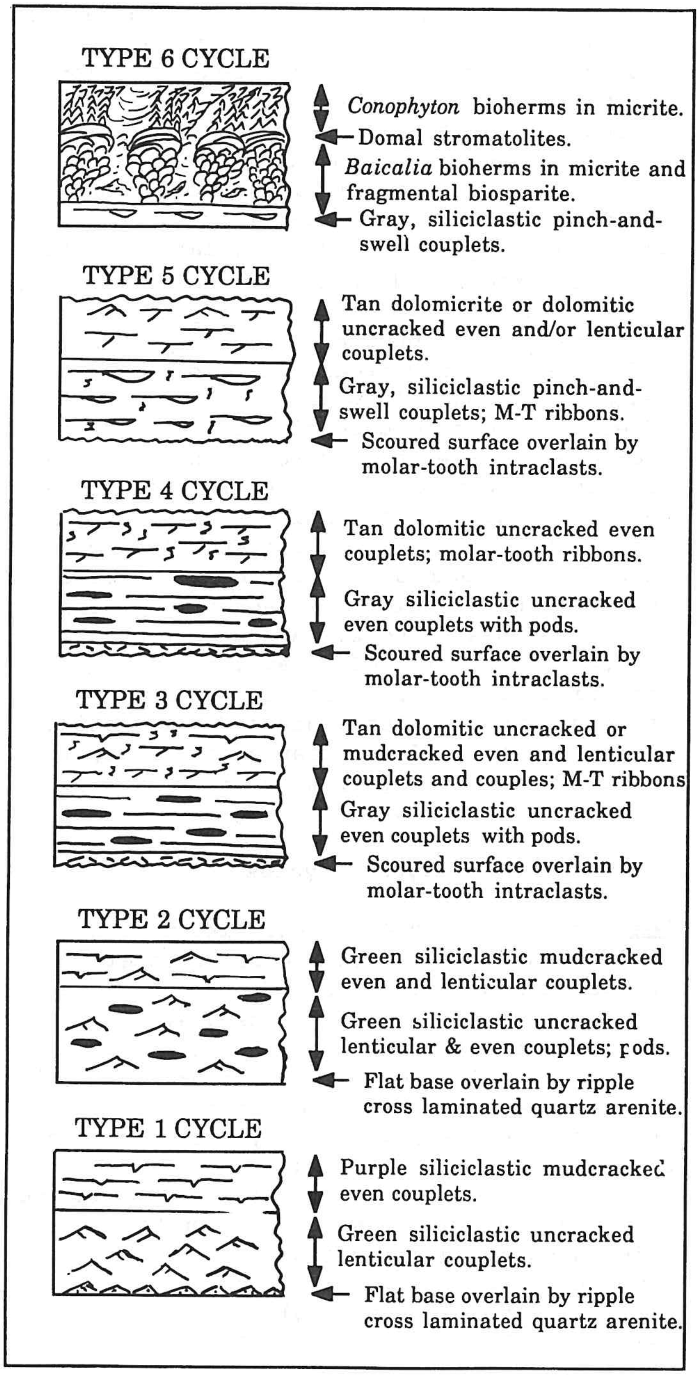


Figure 5. Sketches of cycle types 1, through 6 of the St. Regis, Empire and Helena formations, showing vertical sequences of sediment types.

stop 1 in the upper St. Regis Formation.
Interpretation.- Type 1 cycles are superimposed on the initial phase of the large scale lake filling. The purple, oxidized, mudcracked even couplets of the upper half-cycles are interpreted to record deposition from sheetfloods on exposed playa mudflats (see discussion of the mudcracked even couplet sediment type). Ripple-crossbedded quartz arenite of the coarse sand and intraclast sediment type at the bases of the cycles was deposited in lake strands as they migrated across the playa mudflats during humid climatic periods (see discussion of the coarse sand and intraclast sediment type). The overlying uncracked rippled lenticular couplets of the lower half-cycles record establishment of shallow perennial lakes with flat bottoms within the reach of fair weather waves (see discussion of the uncracked lenticular couplet sediment type). Lenticular couplets are overlain by mudcracked even couplets, signaling the return of arid conditions and exposed mudflats of ephemeral playa lakes.

Although transgressive strands are marked by medium-grained arenite derived from the eastern edge of the basin, exposure of the lake floor was not accompanied by regressive sandy strand deposits. Instead the flat lake floors simply appear to have emerged as mudflats. Exposure may have been so abrupt across the flat, shallow lake floor that sand from the eastern lake margin did not have time to sweep across the rapidly emerging flat. This theme of rapid exposure of flat lake floors carries through to other cycle sequences.

Purple hematite in the mudcracked even couplets of the upper half-cycles reflects oxidation of mud on the exposed flats. Mud carried into the perennial lake was also probably oxidized, but was not preserved in the oxidized state in green argillite of the lower half-cycles. Instead, the ferric oxide was probably reduced to soluble ferrous oxide by small concentrations of organic material in the lake sediments and migrated elsewhere. Consequently, chlorite, in the absence of hematite, imparts the green color to the lower half-cycles.

Type 2 Cycles

Description.- Type 2 cycles are dominantly green and siliciclastic. They, like the type 1 cycles, have sharp bases mantled by thin layers of ripple-crossbedded quartz arenite. Above the basal layers are comparatively thick intervals of uncracked lenticular and uncracked even couplets overlain by thin intervals of mudcracked even and lenticular couplets. Type 2 cycles are illustrated at stop 2 in the Empire Formation.
Interpretation. - Type 2 cycles are quite similar to type 1 cycles. Their thin rippled medium-grained arenite bases are interpreted to record spreading of small strands over exposed, playa mudflats. The uncracked lenticular couplets in the lower half-cycles reflect submergence and reworking by fair weather waves across a shallow, flat lake floor, and the mudcracked even couplets in the upper half-cycles record exposure of sheetflood mudflats. However, the uncracked

even couplets interstratified with the lenticular couplets in the lower half-cycles are interpreted to record episodic influx of suspended silt and clay deposited below the reach of fair weather waves. Thus, the lower half-cycles represent deposition both above and below fair weather wave base, deeper than the lower parts of type 1 cycles.

The lenticular couplets interstratified with mudcracked even couplets in the upper type 2 half-cycles record more standing water and wave reworking than in the upper parts of type 1 cycles. In addition, upper half-cycles are thinner in the type 2 cycles. Thus, although the type 2 cycles record the same basic history of lake expansion and contraction as the type 1 cycles, they reflect deeper, more prolonged submergence accompanied by diagenetic reduction of all the ferric iron. The trace of tan-weathering dolomite in the lower half-cycles at stop 2 may indicate carbonate precipitation from slightly saline waters in a hydraulically closed lake.

Type 3 Cycles

Description.- Type 3 cycles typically have sharp, scoured bases mantled by intraclasts of molar-tooth fragments (coarse sand and intraclast sediment type) instead of quartz arenite as in cycle types 1 and 2. Above the intraclasts are intervals of siliciclastic uncracked even couplets with diagenetic horizontal calcitic pods. They, together with the basal intraclasts, form the gray-weathering lower half-cycles. The upper half-cycles have tan-weathering, dolomitic, siliciclastic uncracked even and lenticular couplets and mudcracked even couplets with diagenetic molar-tooth ribbons and blobs. Type 3 cycles are illustrated at stop 3.

Interpretation.- The sharp, scoured bases of the type 3 cycles are inferred to record exposure and erosion of the uppermost muds of the underlying cycles. Broken molar-tooth ribbons were concentrated in layers of the coarse sand and intraclast sediment type on the erosion surfaces. The intraclasts were probably reworked by waves as the lake rapidly expanded across the nearly level lake floor during the onset of humid climatic cycles. Type 3 cycles formed too far out in the basin to be reached by quartz sand from the eastern basin rim. The overlying siliciclastic uncracked even couplets of the lower half-cycles record episodic deposition below fair weather wave base from either turbid suspension clouds produced by floods entering the western edge of the basin or from lake sediment resuspended by storms and transported by wind-driven currents. The siliciclastic and calcitic composition of the lower half-cycles may reflect nearly fresh waters in an open lake system with an outlet.

Dolomite in the upper half-cycles probably records evaporative concentration and precipitation of carbonate minerals in a closed lake that may have shrunk below its outlet during arid climatic intervals. Gypsum may also have precipitated in the sediments and may have been reduced by bacteria to H_2S, forming the molar-tooth gas expansion blobs and cracks in the upper parts of the cycles. The progressively muddier couplets in the upper half-cycles probably reflect either a decrease in flood-derived sediment during arid periods or a decrease in effective storm resuspension in a shrinking, shallower lake with less fetch and more wave drag on the bottom. Lenticular couplets in the upper half cycles indicate lowering of fair weather wave base across the nearly flat lake floor. Mudcracked even couplets at the tops of type 3 cycles show abrupt exposure as in cycle types 1 and 2. Absence of cracks in the tops of other cycles may indicate formation of an efflorescent salt crust at the cycle tops, which inhibited desiccation and cracking. Efflorescent salt crusts, like those of modern playas, were probably dissolved during lake flooding.

Type 4 Cycles

Description.- Type 4 cycles, like type 3 cycles, have sharp, scoured bases covered by intraclasts of molar-tooth fragments. They also have lower half-cycles of gray siliciclastic uncracked even couplets with horizontal calcitic pods. However, the uncracked even couplets continue up into the upper half-cycles, where they become muddier and dolomitic. Even uncracked couplets become muddier by thinning of the silt in the lower half-couplets and proportionate thickening of the clay and dolomicrite in the upper half-couplets. The upper half-cycles also contain diagenetic molar-tooth blobs and vertical ribbons. Type 4 cycles are interstratified with type 3 cycles at stop 3. They occur in the lower part of stop 4 and in parts of stop 5.

Interpretation. - The lower parts of type 4 cycles, like those of type 3, record rapidly spreading lake waters across exposed mudflats and establishment of a perennial lake. Sediments accumulated from episodic suspended sediment influx on a flat lake floor below fair weather wave base. Diagenetic calcitic pods may reflect nearly fresh waters in an open lake system.

The abrupt introduction of dolomite in the upper half cycles is interpreted to indicate rapid carbonate saturation in the lake waters, probably as a result of evaporation, shrinking, and formation of a closed lake system as proposed for type 3 cycles. However, the continuation of uncracked even couplets through the upper type 4 half-cycles indicates that this chemical change in lake water was not accompanied by a shift in sedimentary structures and sediment types. Instead, even uncracked couplets deposited below fair weather wave base continued up to the tops of the cycles. Decreasing silt in the couplet bases may reflect decreasing sediment influx during arid climatic intervals of the upper half-cycles. Scour at the tops of the cycles indicates either abrupt exposure without time to establish regressive wave rippled surfaces, or possibly deposition of only thin veneers of rippled sediments which were removed by scour at cycle tops.

Type 5 Cycles

Description.- Type 5 cycles are characterized by scoured surfaces mantled by thin layers of molar-tooth intraclasts. Above the intraclasts are siliciclastic lower half-cycles containing pinch-and-swell couplets and couples. The tan-weathering upper half cycles are mostly dolomitic uncracked even and lenticular couplets. Type 5 cycles occur in the upper part of stop 4, in parts of stop 5, and in the lower part of stop 6.
Interpretation.- The scoured surfaces mantled by molar-tooth intraclasts indicate rapid lake expansion and establishment of a perennial lake as in cycle types 3 and 4. The pinch-and-swell couplets and couples in the lower half-cycles record deposition above storm wave base, but below fair weather wave base. (See discussion of the pinch-and-swell couplet and couple sediment types). Introduction of storm deposits in the bases of the cycles probably reflects greater expansion and fetch in the Great Belt Lake, resulting in larger storm waves and deeper, more effective storm reworking (Johnson, 1993). Accordingly, the lower parts of type 5 and type 6 cycles (described below) are interpreted to have the deepest water sediments of the most expanded lake phases.

The vertical succession from pinch-and-swell couplets to uncracked even couplets and uncracked lenticular couplets in the dolomitic upper half-cycles records decrease in storm intensity as a result of a shrinking lake with reduced fetch. Dolomite indicates restriction and precipitation in a closed lake. Cycles with lenticular couplets reflect lowering of wave base to the depositional interface as the lake contracted.

Type 6 Cycles

Description.- Type 6 cycles are the two *Baicalia-Conophyton* cycles seen at stop 6. They have sharp bases locally marked by intraclasts, overlain by intervals of very dark, siliciclastic pinch-and-swell couplets. These are sharply overlain by *Baicalia* bioherms surrounded by micrite and loosely packed *Baicalia* debris in a sparry matrix. They grade up to domal stromatolites, which in turn are overlain by *Conophyton* columns beneath capping micritic even couplets with domal stromatolites at the tops of the cycles. Type 6 cycles are more fully described in the discussion of stop 6.
Interpretation.- Type 6 cycles are essentially type 5 cycles with a superimposed sequence of stromatolitic growth forms. Regional study shows that intraclasts only locally mark the base of the lower *Baicalia-Conophyton* cycle and are absent at the base of the upper *Baicalia-Conophyton* cycle. Therefore, boundaries of type 6 cycles may have been shallow, but not exposed. Pinch-and-swell couplets at the bases of the cycles record expansion of siliciclastic cycle phases and deposition within storm base. *Baicalia* colonies became established and may have induced micrite precipitation. Lenses of loosely packed *Baicalia* debris in a matrix of open work sparry calcite may reflect storm scour of the *Baicalia* bioherms. The upward progression through uncracked evenly laminated micrite with domal stromatolites to *Conophyton* columns in laminated micrite reflects the same decrease in turbulence seen in the type 5 cycles, possibly reflecting a shrinking lake. Continuity of calcitic micrite to the tops of the cycles suggests that Belt Lake waters remained fresh throughout the *Baicalia-Conophyton* cycles and that the lake retained an outlet.

Analysis of the Cycles

Figure 6 shows that the six cycle types form a progressive series from rippled and mudcracked cycles in the St. Regis and Empire formations to cycles with pinch-and-swell and uncracked even couplets in the middle and upper Helena Formation. Sediment types that occur in the lower half-cycles of stratigraphically lower cycles tend to occur in the upper half-cycles of stratigraphically higher cycles, reflecting deeper and deeper water cycles of a larger and larger lake. Thus, all cycles are variants of the same progressively evolving, fundamental cyclic processes. The sedimentary processes of the sediment types and their progressive arrangement in the cycles allow us to place some important constraints on the "Belt sea".

First of all, regardless of whether the cycles are lacustrine or marine, the outcrop extent of the Helena and Wallace formations

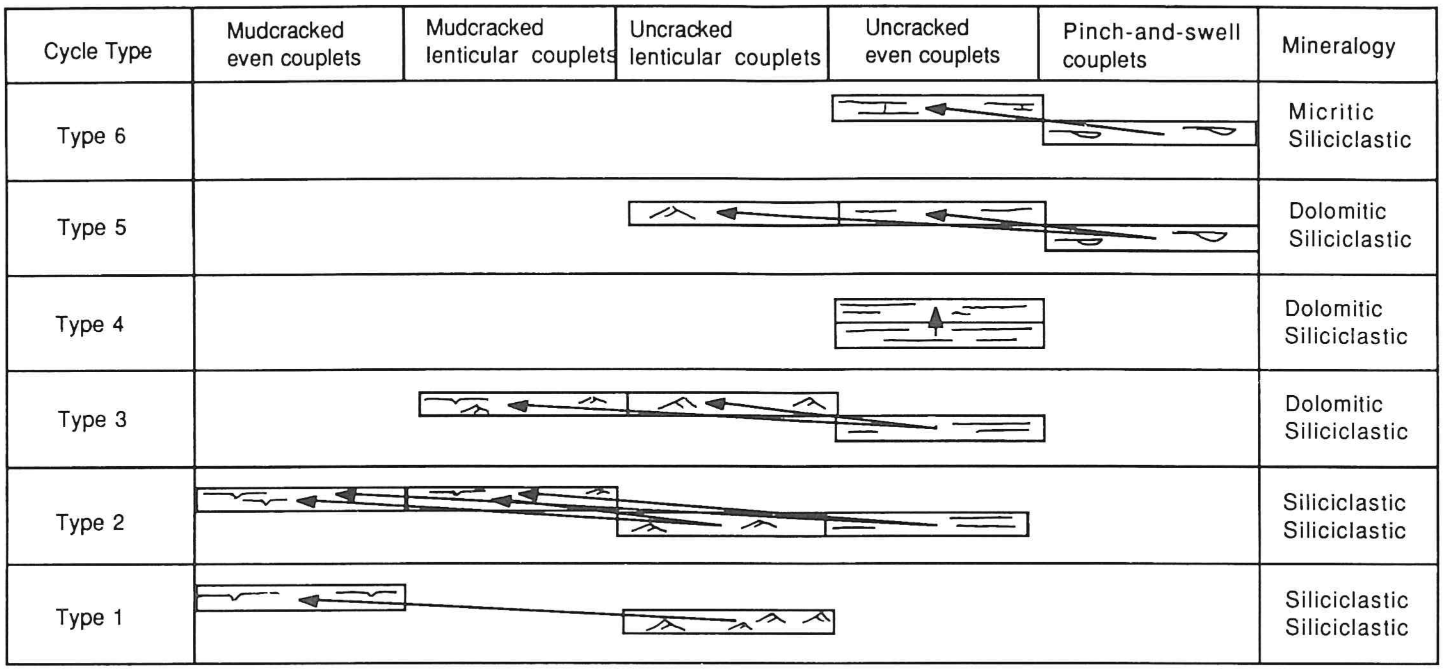


Figure 6. Comparison of sediment types in cycle types 1 through 6, showing the upward progression to deeper, more open water facies.

from the Salmon River of east-central Idaho (Winston and Link, 1993) to southern British Columbia and from the Big Belt Mountains into eastern Washington shows that, during its maximum spread, the "Belt Sea" had a fetch of more than 300 km, comparable in scale to the northern part of the modern Caspean Sea.

Second, hummocky cross stratification in the pinch-and-swell couplet and couple sediment types, inferred to be the deepest water facies, indicates that, even during the greatest lake expansion, the bottom was within reach of storms. Therefore, it was a shallow, exceedingly broad body of water during deposition of the Helena, again, like the northern part of the Caspean Sea.

Third, broad, shallow lakes tend to have mixed water columns, and the dominance of pinch-and-swell couplets and couples in the deepest water facies is supportive evidence of frequent mixing of the water column during deposition of the lower half-cycles.

Forth, the lake must have had a flat, nearly level floor. Its slope along the eastern (present coordinates) margin of the basin must have been exceedingly gentle, without any indication of a carbonate shelf fringed by a clinoform. Evidence for this inference comes from the abruptness with which the scoured cycle boundaries rest upon permanently submerged lake sediments, particularly in type 4 cycles. As pointed out above, regression must have been too rapid to develop a regressive strand deposit. Strands sweep rapidly across modern playas without leaving a beach record, and this explanation is most reasonable for the Helena cycles.

Fifth, regressions of this style, with lowstand deposits sharply sandwiched between highstand deposits are recognized as evidence of forced drops in water level (Plint, 1988; Dam and Surlyk, 1992). Therefore, Helena cycles were not produced by shoaling upward sequences in which sediment accumulated up to a stable marine base level as proposed by Eby (1977), but were induced by periodic rises and falls of water level.

Sixth, the siliciclastic to dolomitic repetition in the cycles probably reflects abrupt basinwide changes in lake water chemistry from relatively fresh and calcite-precipitating to more saline and protodolomite-precipitating. Evidence for this abruptness is found in the sharpness of the calcitic and siliciclastic-to-dolomitic boundaries without the the calcite and dolomite intertonguing as they would if they were contemporaneous facies. Basinwide cyclic changes in water chemistry are more characteristic of enclosed lakes with alternating humid and arid climatic episodes than they are of more stable marine systems. In addition, the calcite-dolomite change in some cycles occurred independently from siliciclastic facies shifts, particularly in type 4 cycles where uncracked even couplets continue to the tops of the cycles.

Finally, the Belt cycles were not deposited in a marine intertidal zone. The uncracked lenticular couplets of the lower half cycles and mudcracked even couplets of the upper half cycles contrast sharply with tidal sequences with which they sometimes have been compared (Price, 1964; McMechan, 1981). Tidal sequences are characterized by tidal channel deposits with herringbone and large accretionary crossbeds and sandflats with climbing ripple flaser bedding (e.g. Reineck and Singh, 1980). All modern tideflats, including those at the head of the Gulf of California (Thompson, 1975; Meckel, 1975), have an arterial system of tidal channels which deliver and receive the flood and ebb tides. Certainly if mudflats, the size of those in the Belt, were flooded and drained by astronomically driven tides, someone, somewhere, would have found one bonafide tidal channel and tidal flaser-bedded sandflat deposit, which they have not. Furthermore, if the red and green argillite sequences were tidal deposits, then one would expect to find associated large beach and barrier island successions, which are lacking in the Belt. Strand deposits in the Belt cycles are much more like small playa beaches (Renaut and Owen, 1991). Oscillation ripples like those that dominate the Belt are largely limited to the more protected parts of tideflats. Mudcracked surfaces are relegated to the supratidal zone and form only the landward fringe of the intertidal complex. Widespread rippled and mudcracked surfaces like those in the Belt are the mark of broad playas, like modern Lake Eyre, Australia. Thus, if a marine interpretation of the Helena cycles is to be argued, they could not have been deposited in a tidally influenced basin.

All of the above seven constraints are compatible with deposition in a great, shallow inland lake. The first five are also compatible with a marine environment, but number six speaks strongly for a chemically isolated lake and number seven repudiates a marine intertidal or peritidal setting.

Comparison with Phanerozoic Lacustrine Cycles

Fair-weather and storm-dominated lacustrine cycles in partly open lakes have not been widely reported. Most lacustrine studies have focused on deeper, narrower lakes than the Great Belt Lake and have thick laminated sub-storm base black shale successions, which the Helena cycles lack. However, two studies of storm-dominated lake cycles have been reported from the Lower Carboniferous Horton Bluff Formation , Nova Scotia (Martel and Gibling, 1991), and the Lower Jurassic East Berlin Formation of the Hartford basin, New England (Demmico and Gierlowski Kordisch, 1986). Belt cycle types 1 through 5 compare well with parts of these cycles (Fig. 7).

The Horton Bluff Formation is a wave-dominated cyclical lacustrine succession (Martel and Gibling, 1991), deposited during the assembly of Pangea. It has three types of cycles: sediment starved, shoreline dominated, and marsh dominated. Belt cycles most closely resemble the shoreline dominated cycles that average 6.5 m thick (Fig. 7), They have the following vertical sequence: 1) gray clay shale and claystone - suspension settleout below storm wave base; 2) hummocky cross-stratified siltstone - storm? generated current settleout; 3) wave rippled sandstone - moderate to high energy wave generated bedload deposition; 4) planar bedded siltstone - shoaling wave deposition in very shallow water; 5) green mudcracked mudstone - lake fringing marsh, and 6) dolomite - mostly pedogenic precipitation. The Horton Bluff Formation cycles have no basal emergent or strand deposits and begin with sub-storm wave base thinly laminated shale and claystone, not represented in Belt cycles. However, the overlying sequence of hummocky cross-stratified, wave-rippled and mudcracked planar bedded mudstone and siltstone facies (Fig. 7) closely resemble the Belt cyclic succession of pinch-and-swell, uncracked or mudcracked lenticular couplet and mudcracked even couplet sediment types, with the Belt uncracked even couplet type missing. Both record cycles of lake expansion followed by contraction and exposure, although Martel and Gibling (1991) argue for tectonically driven cyclicity.

The East Berlin Formation is part of the Newark Supergroup of continental redbeds and lacustrine deposits that filled the rift valleys of eastern North America during the breakup of Pangea. The East Berlin Formation has two styles of muddy cyclic sequences: one containing laminated black mudstone in its lower part (type A) and one containing interbedded sandstone and mudstone in its lower part (type B) (Demmico and Gierlowski Kordisch, 1986). Belt cycles most closely resemble the 2-4 m thick type B cycles (Fig. 7), with the following fining-upward sequence: 1) planar and large-scale cross-stratified sandstone, 2) interbedded sandstone and mudstone, 3) small-scale cross-stratified silty sandstone, and 4) disrupted mudstone. The planar and large-scale cross-stratified facies records sheetflooding across the exposed lake floor at the cycle boundaries and, aside from recording exposure, has no counterpart in the Helena cycles. Above this is interbedded sandstone and mudstone characterized by loaded, continuous and discontinuous rippled and laminated sand overlain by dark shale with ptygmatic folded cracks. It compares closely to the pinch-and-swell couple and couplet

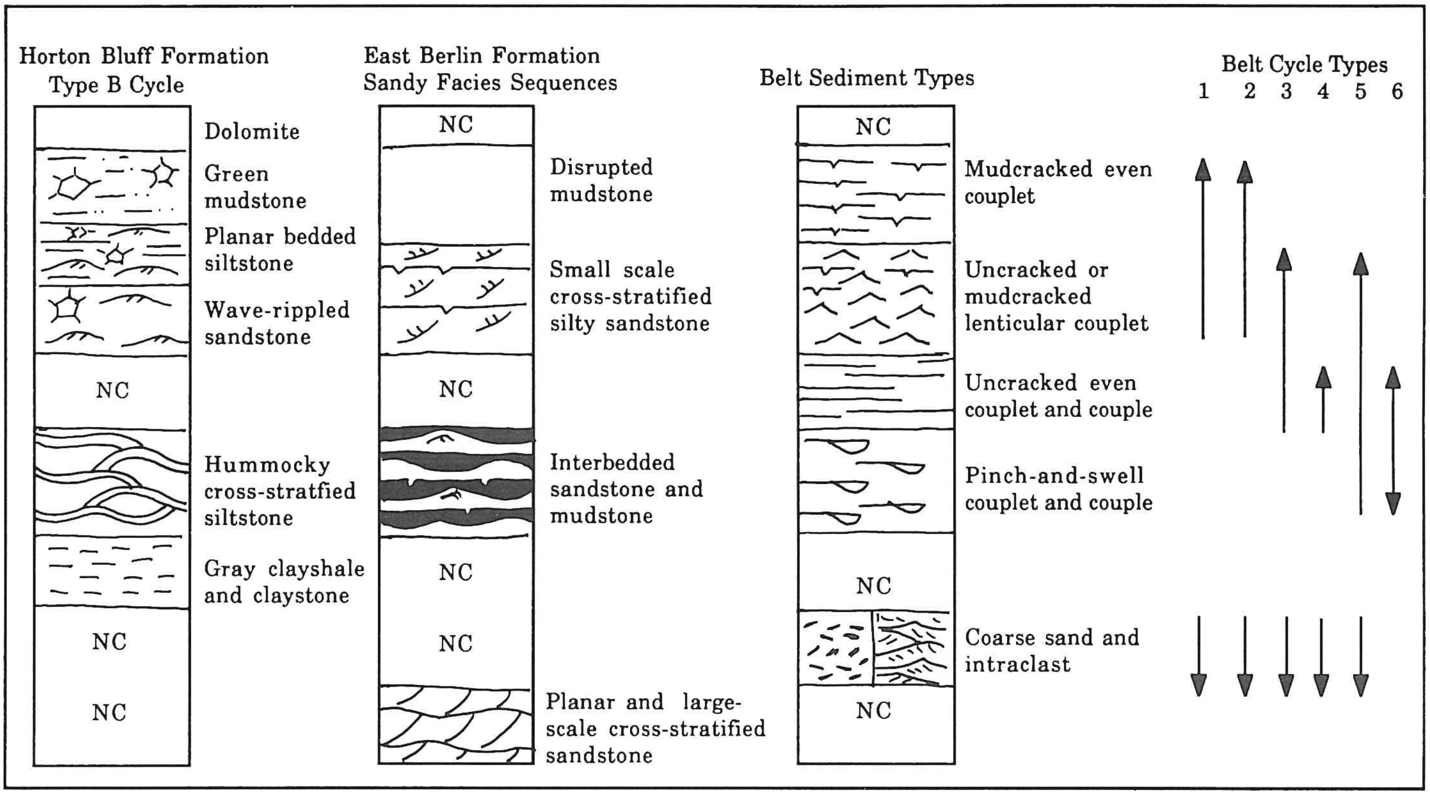


Figure 7. Comparison of Belt cycles with lacustrine cycles from the Carboniferous Horton Bluff Formation after Martel and Gibling, (1991) and from the Jurassic East Berlin Formation after Demmico and Gierlowski Kordisch, (1986). NC means no corresponding facies. Arrows on right show sequences of sediment types for Belt cycle types 1-6. Shafts are discontinuous where sediment types and facies are not represented.

sediment types, and we interpret it to represent subaqueous storm deposits of expanded lake phases. The overlying small-scale cross-stratified silty sandstone is analogous to the fair-weather wave deposits of the uncracked and mudcracked lenticular couplet sediment type and represents shallowing lake phases. Finally, the disrupted mudstone is analogous to the exposed flats of mudcracked even couplet type (Fig. 7). Interpreted in this way, the Horton Bluff, East Berlin and Helena formations record closely comparable cyclic sequences of lake filling and contraction.

ROAD LOG

0.0 Leave Grouse Mountain Lodge and turn right onto U.S.Highway 93 toward Whitefish.

0.3

0.3 Enter Whitefish stay on Highway 93 through town.

0.5

0.8 Cross Stillwater River.

0.3

1.1 Third stop light, turn right and follow U.S. Highway 93 to Kalispell.

2.3

3.4 Montana Highway 40 to Columbia Falls on left, stay straight on U.S. Highway 93 to Kalispell.

9.3

12.7 Cross Stillwater River.

2.1

14.8 Enter Kalispell.

1.2

16.0 Intersection of U.S. Highways 93 and 2. Turn right on U.S. Highway 2 toward Libby.

9.1

25.1 Road to Kila on left. Stay on U.S. Highway 2.

0.2

25.3 Road cuts in tan-weathering carbonate of the lower Helena Formation on both sides of road.

1.0

26.3 Road cuts in green argillite of the Empire Formation on right.

0.7

26.9 Roadcuts of Empire green argillite on right will be seen at Stop 2 on the return leg. Highway continues to cut down section toward Stop 1.

0.1

27.0 The turnout on the right at bottom of hill is where we will park at Stop 2.
1.6
28.6 Purple and green argillite in roadcuts and cliffs on the right for the next mile are in the uppermost St. Regis Formation.
1.1
29.7 **STOP 1** Purple and green argillite roadcuts on right are in the upper part of the St. Regis Formation. Slow down on hill and pull off on small turnout immediately beyond roadcuts. Walk up along highway to examine cycles in the purple and green argillite of the upper part of the St. Regis Formation of the Ravalli Group (Fig. 8).

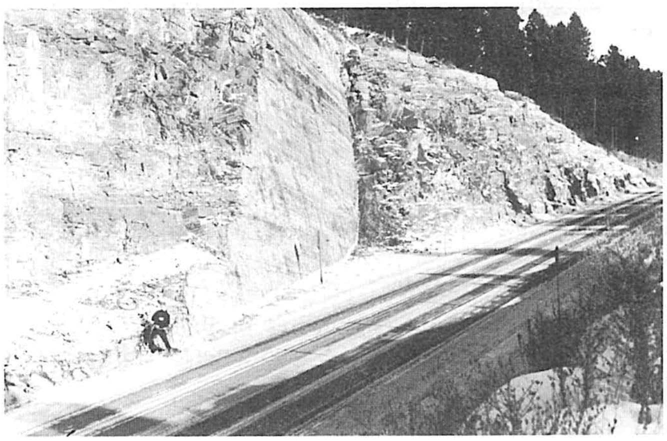


Figure 8. Road cut at stop 1. Nate Hathaway and his dog, Lucy for scale.

Assignment of these rocks to the St. Regis Formation should be explained. On the Kalispell 1° x 2° sheet and in this guidebook Harrison and others (1993) map this unit as the Spokane Formation. As pointed out by Buckley and Ryan (1993), tongues of the Revett Formation were included in the Burke, St. Regis and "Spokane" formations by Harrison and others (1986). A quartzite unit stratigraphically underlies this purple and green argillite and crops out to the west (Harrison, oral communication, 1992). Harrison and others (1993) include this unit in the "Spokane" Formation, although we believe it belongs to the upper member of the Revett Formation. Therefore, we believe this purple and green argillite unit above the upper Revett should be assigned to the St. Regis Formation. We further believe that the redbed package of the Ravalli Group can most clearly be expressed in northwestern Montana by four formations: the Burke, Revett, St. Regis and Grinnell formations. Where the Revett can be mapped, the Burke, Revett and St. Regis formations should be recognized. Where the Revett fines and thins to the east and can no longer be mapped, the redbed package should be mapped as Grinnell. Introduction of the term Spokane from the Helena embayment into northwestern Montana is superfluous and merely confuses stratigraphic analysis.

We begin the field trip with this stop because it displays some of the clearest evidence of cyclic submergence of subaerially exposed mudflats, followed by aqueous retreat and re-exposure (Fig. 9). Both sedimentary structures and diagenetic colors reflect the cyclic pattern. These are type 1 cycles. They are characterized by sharp bases capped by very fine- or medium-grained ripple crosslaminated arenite of the coarse sand and intraclast sediment type a centimeter or two thick. Above the basal arenite are mostly uncracked green lenticular couplets (uncracked lenticular couplet sediment type) in the lower half-cycles. They pass upward to purple mudcracked even couplets of the upper half-cycles. These are in turn sharply overlain by the green bases of the next cycle.

The sharply defined bases of the cycles above mudcracked couplets are interpreted to record scour of an exposed mudflat surface by small breaking waves during cyclic expansion of the Great Belt Lake. The thin ripple crosslaminated, medium-grained quartz arenite layers overlying the basal contact probably represent bedload transport and longshore drift by breaking waves. These thin, rippled sandy beds reflect how small the advancing strands were. Drag on the waves as they crossed the shallow flats probably kept the breaking waves and resulting beaches small.

The ripple crosslaminated, uncracked lenticular silt-to-clay couplets record silt transport and deposition by waves followed by suspension clay settleout. The depositional interface was clearly within fair weather wave base. Chlorite imparts the green color to these rocks. As suggested above, the mud probably contained ferric oxide, which may have been reduced to the ferrous state by traces of organic material in the subaqueous sediments and removed in solution.

The purple mudcracked silt-to-clay couplets (mudcracked even couplet sediment type) are interpreted to record episodic sheetflood transport and deposition across exposed mudflats, followed by desiccation. Ferric iron was probably deposited as brown goethite or limonite and altered to hematite by burial metamorphism.

Notice the absence of tidal channels, flaser bedding with climbing current ripples, and a thick barrier island deposit. These deposits are not ancient tidal flats. They more closely compare to modern and ancient playa lake successions because of the dominance of couplets and the diminutive strand deposits. As in playa lakes, the even silt-to-clay couplets are interpreted to

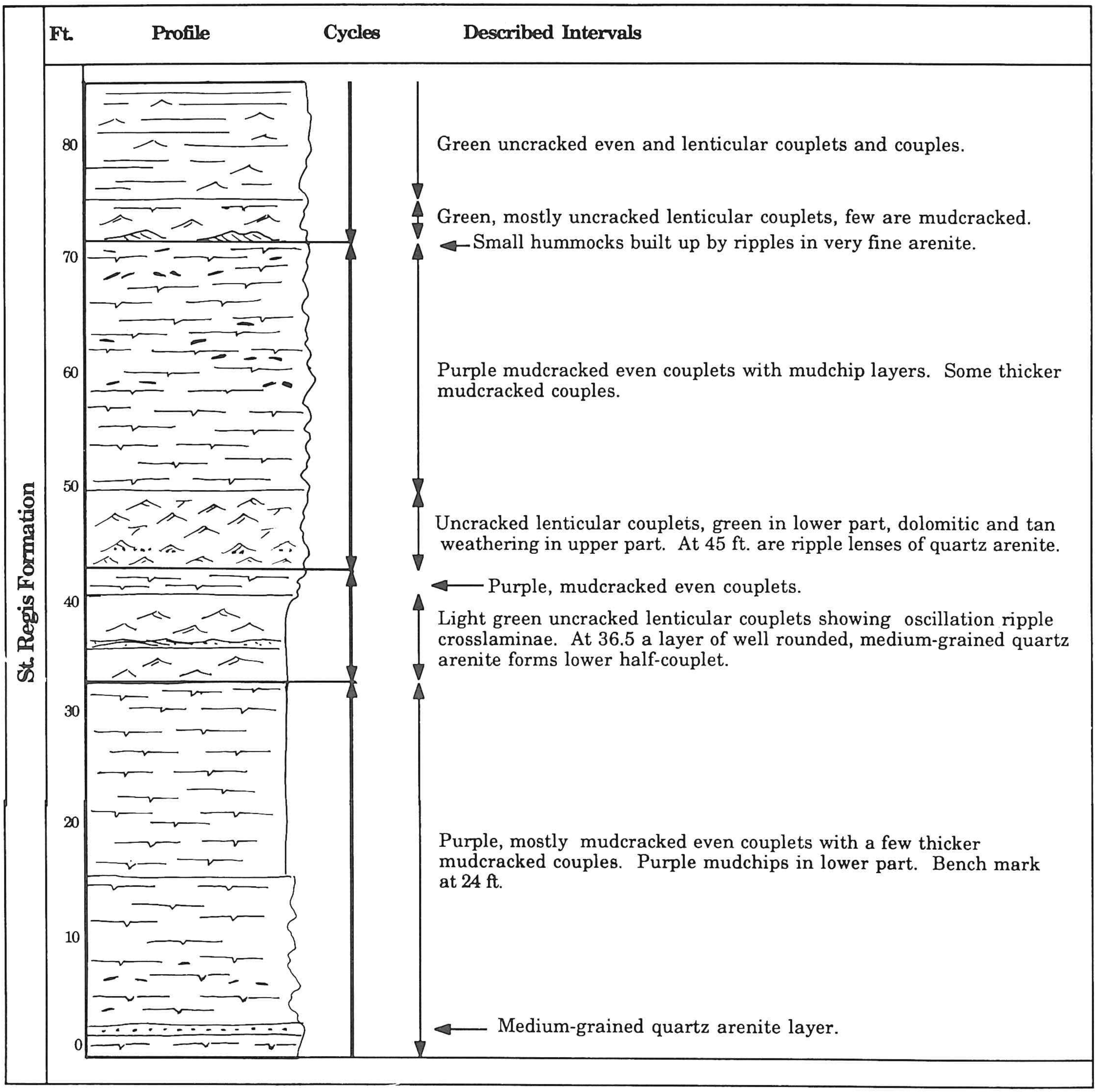


Figure 9. Measured stratigraphic section of stop 1.

record flooding of an exposed surface followed by standing water in an enclosed basin.

This outcrop spans the upper part of one cycle followed by two more cycles and the base of a fourth (Fig. 9). One variation on the above general theme should be noted. In the third cycle medium-grained arenite occurs as a layer of ripple crosslaminated lenses a decimeter above the basal ripple crossbedded very fine grained arenite. Perhaps this records the migration of a small shoal, outboard from the strand.

Return to vehicles and retrace route east on U.S. Highway 2 back toward Kalispell.

2.9

32.6 **STOP 2** Pull off on wide school bus turn around on left side of the highway. Walk about 400 ft. up highway to roadcuts of green argillite in the Empire Formation (Fig. 10). Three full cycles are exposed in this outcrop plus parts of two more cycles (Fig. 11).

These cycles, like those of Stop 1, have sharp bases and lower half-cycles of the uncracked lenticular couplet sediment type overlain by upper half-cycles containing mudcracked even couplets. They are also interpreted to record expansion and contraction of the Great Belt Lake.

However, there are some important differences from Stop 1. The lower uncracked parts of the cycles are proportionately thicker, and the upper mudcracked parts are proportionately thinner. The mudcracked upper half-cycles are green along with the lower half-cycles, and in addition to mudcracked even couplets, the upper half-cycles contain mixtures of mudcracked lenticular couplets and uncracked even and lenticular couplets (Fig. 12). Therefore, they are type 2 cycles. Medium-grained

Figure 10. Road cut at stop 2.

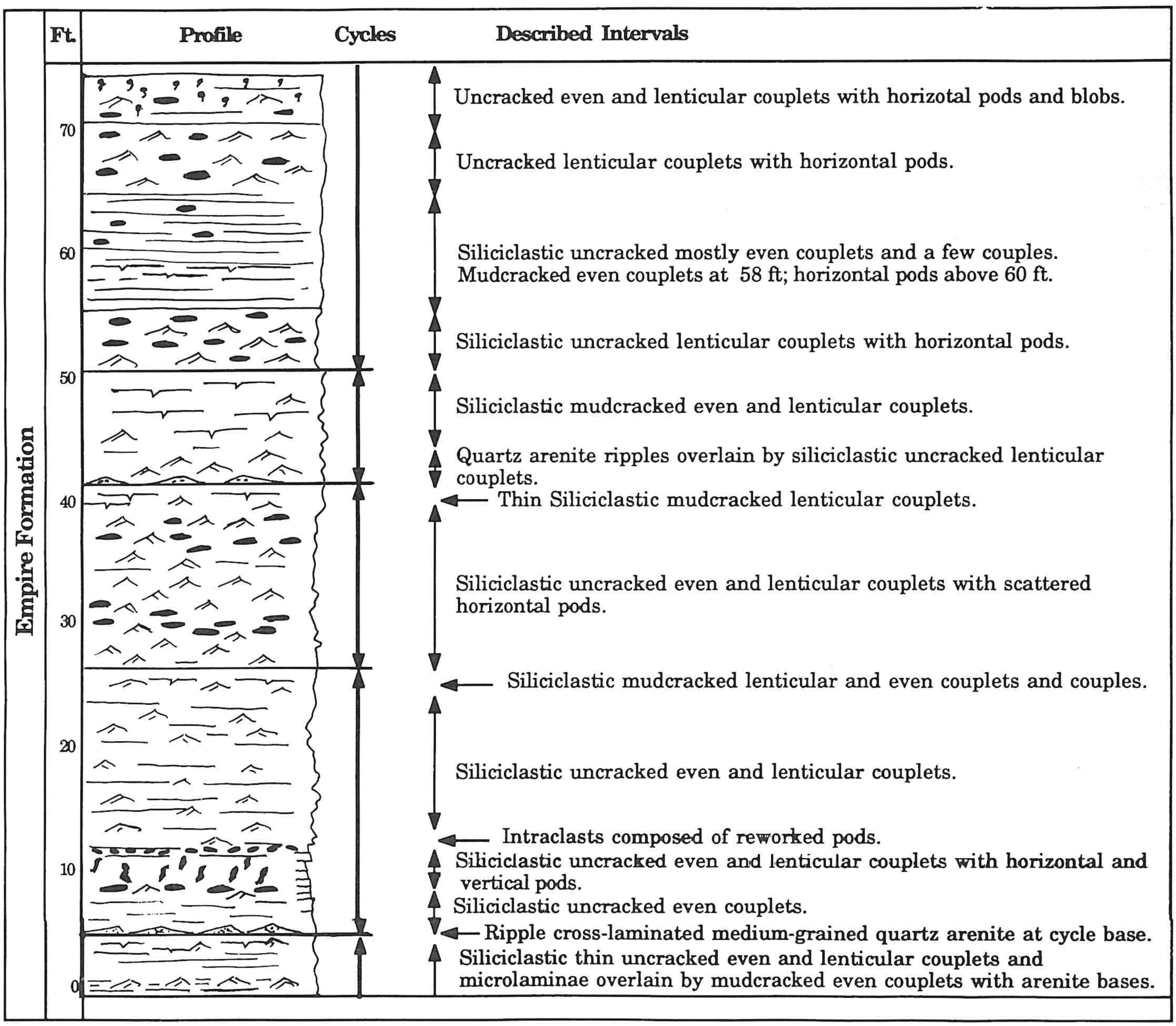


Figure 11. Measured stratigraphic section of stop 2.

quartz arenite layers that mark the bases of the cycles in Stop 1 form the base of only the lowest cycle here. Some intervals of uncracked lenticular couplets in the lower half-cycles contain horizontally and vertically elongate calcitic pods (Fig. 13). At 13 ft.in the measured section (Fig. 11), some pods are reworked into an intraclast conglomerate bed. In addition, molar-tooth blobs, representing gas bubbles, occur above 70 ft.(Fig. 11). The combination of even and lenticular couplets, green color and calcitic pods characterize the Empire Formation.

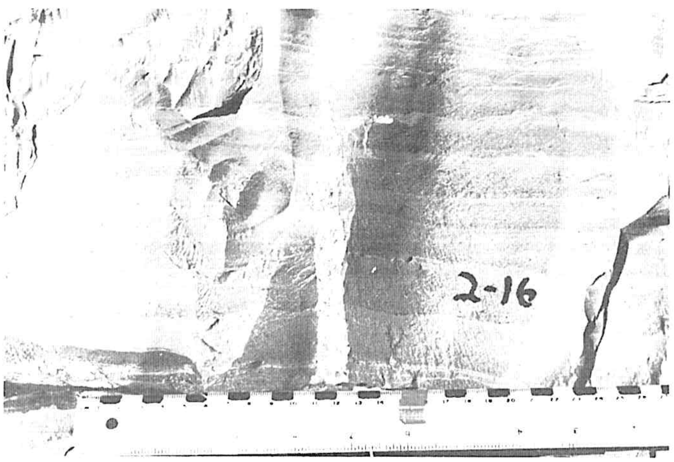


Figure 12. Siliciclastic uncracked even couplets at 16 ft. in measured section of stop 2 (Figure 11).

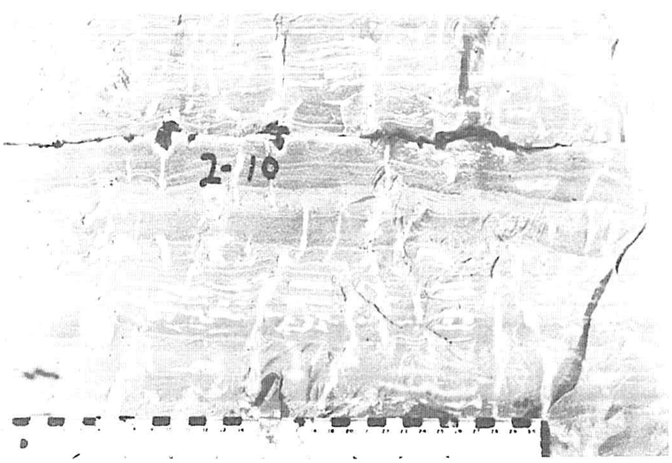


Figure 13. Fresh road cut of horizontal and vertical calcitic pods at 10 ft. in measured section of stop 2 (Figure 11).

The expansion of uncracked even and lenticular couplets at the expense of mudcracked even couplets indicates more prolonged submergence and more abbreviated periods of emergence in the Empire cycles. Silt and clay brought in by sheetfloods was not immediately desiccated and remained submerged longer to be reworked by waves into lenticular couplets. Fluids reduced by organic constituents permeated the sediments, altering all iron to the ferrous state. More carbonate mud may have been generated in the lacustrine waters and is recorded by small percentages of tan weathering dolomite in the siliciclastic mudstone and as early calcite cement of the pods.

The progression from the St. Regis cycles below to the Empire cycles records basinwide large scale expansion of the Great Belt Lake with more subaqueous sediment types. Cycles in the lower part of the Helena Formation seen at stop 3 continue this trend.

Return to vehicles and continue east on U.S. Highway 2.

0.2

32.8 Kila road on right immediately above stop 2. Stay on highway.

10.8

43.6 Intersection of U.S. Highways 2 and 93 again. Continue east on Highway 2 through Kalispell toward Glacier National Park.

1.5

45.1 Cross Stillwater River move to left lane.

0.6

45.7 Bear left at stop light to Columbia Falls and West Glacier, following U.S. Highway 2.

2.4

48.1 Cross railroad tracks. Mountains to east at 2 o'clock are the north end of the Swan Range. Rocks of the Ravalli Group are exposed on range face. Notch to the north is Bad Rock Canyon cut by the Flathead River. Bare mountain with the vegetated bands north of Bad Rock Canyon is Teakettle Mountain. Fumes from the Columbia Falls Aluminum Plant have killed off most of the vegetation,exposing the upper St. Regis, Empire and lower Helena formations.

4.2

52.3 Kalispell airport on left.

4.5

56.8 Stop light and intersection, follow Highway 2 to right (east).

2.0

58.8 Enter Columbia Falls, stay on Highway 2 through town.

1.4

60.2 Cross Flathead River.

1.0

61.2 Intersection with Montana Highway 206. Stop and turn left following U.S. Highway 2 toward Glacier National Park.

2.0

63.2 Enter Badrock Canyon cut by the Flathead River. Red and green argillite on both sides of canyon are in the St. Regis Formation. The prominent arenite beds are thicker and coarser-grained than correlative beds west of the Rocky Mountain Trench, reflecting the eastern source of the coarse quartz sand.

0.7

63.9 Bernie Memorial Park on right with a backdrop of red and green argillite of the uppermost St. Regis Formation. Historical plaque vividly describes Indian ambushes and massacres which gave the canyon its name. We will not stop here because of the bad rocks.

1.3

65.2 Cross South Fork of the Flathead River and enter Hungry Horse.

1.1

66.3 Highway 2 climbs hill as it leaves Hungry Horse on the east side of town. Prepare to turn right beyond sign to Hungry Horse Ranger Station.

0.2

66.5 Turn right onto road leading to Hungry Horse Dam. There are actually two roads: a gravel road on the extreme right that leads to the base of Hungry Horse Dam, and a main paved road that meets the highway perpendicularly and leads to the upper dam viewing facilities. Take the hard right on the gravel road and follow it down the hill.

1.2

67.7 Gravel road is joined on the right by a paved road along the South Fork of the Flathead River. Bear left, on the paved road.

0.2

67.9 **STOP 3** Cliffs on left are in the lower part of the Helena Formation and display well developed cycles. Pull off on right shoulder and cross road to examine cycles (Fig. 14). Twelve full cycles were measured and described in this roadcut (Fig. 15). The cycles are immediately recognizable by the alternating siliciclastic gray-weathering outcrop bands of the lower half-cycles and the upper tan-weathering dolomitic and siliciclastic bands of the upper half-cycles. Each cycle fines upward by the siliciclastic component becoming less silty and more muddy upward and by the upper half-cycles incorporating carbonate (dolomite) mud. Bases of the cycles are sharp. Most are scoured and are veneered by thin continuous and discontinuous layers, up to a few centimeters thick, of molar-tooth ribbon fragments, belonging to the coarse sand and intraclast sediment type. The lower half-cycles above the intraclast beds here are characterized by siliciclastic uncracked even couplets mixed with occasional couples and microlamina intervals. The tan-weathering upper half-cycles are characterized by dolomitic mud and clay with barely enough silt to form muddy uncracked even couples and couplets and lenticular couplets. Cycles with mudcracked and uncracked lenticular couplet upper half-cycles, such as the mudcracked lenticular couplets from 96 to 104 ft. in the measured section are type 3 cycles. Cycles with uncracked even couplets and couples in their upper half-cycles are type 4 cycles. Many of the lower half-cycles contain horizontal pods and the upper half-cycles contain molar-tooth ribbons and blobs.

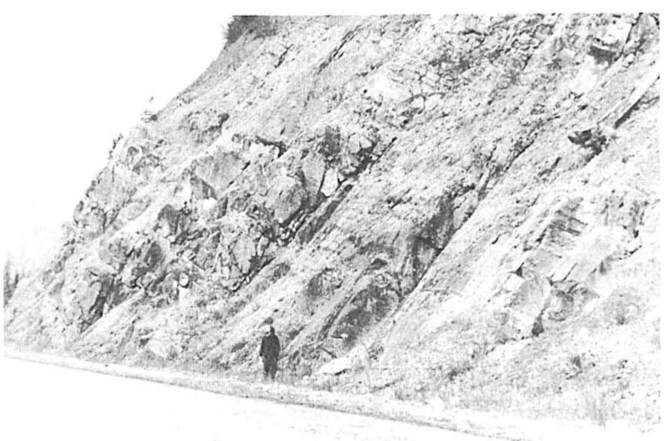


Figure 14. Outcrop at the base of stop 3.

The lower half-cycles are interpreted to record events of suspended load transport and deposition below effective fair-weather wave base. Some uncracked even couplets may record sediment influx from floods that crossed the mudflats and entered the lake waters as suspension clouds. On the other hand, by association with the pinch-and-swell couplets at Stop 4, many of the uncracked even couplets may record resuspension, transport and deposition by storms. These, like floods, were episodic events. The dolomitic upper half-cycles record decrease of silt influx accompanied by carbonate mud deposition now in the form of microcrystalline dolomite.

Like the cycles in the St. Regis and Empire formations, these type 3 and type 4 cycles in the lower Helena Formation are interpreted to record periodic expansion of the Great Belt Lake across exposed and scoured mudflats followed by slow retreat. Evidence for exposure is not so clear as in the St. Regis and Empire formations and only in one cycle are mudcracks well developed. The absence of mudcracks may be due to saline lake waters which kept the sediments moist and inhibited the formation of desiccation cracks. In general the cycles are interpreted to record: 1) subaerial exposure and scour, concentrating the only coarse-grained fragments, namely broken molar-tooth ribbons and blobs and occasional pods, 2) expansion of the lake during humid climatic periods and deposition of the uncracked even couplets by suspended transport of flood derived sediment or sediment resuspended by storms, 3) contraction of the lake during arid periods when siliciclastic sediment

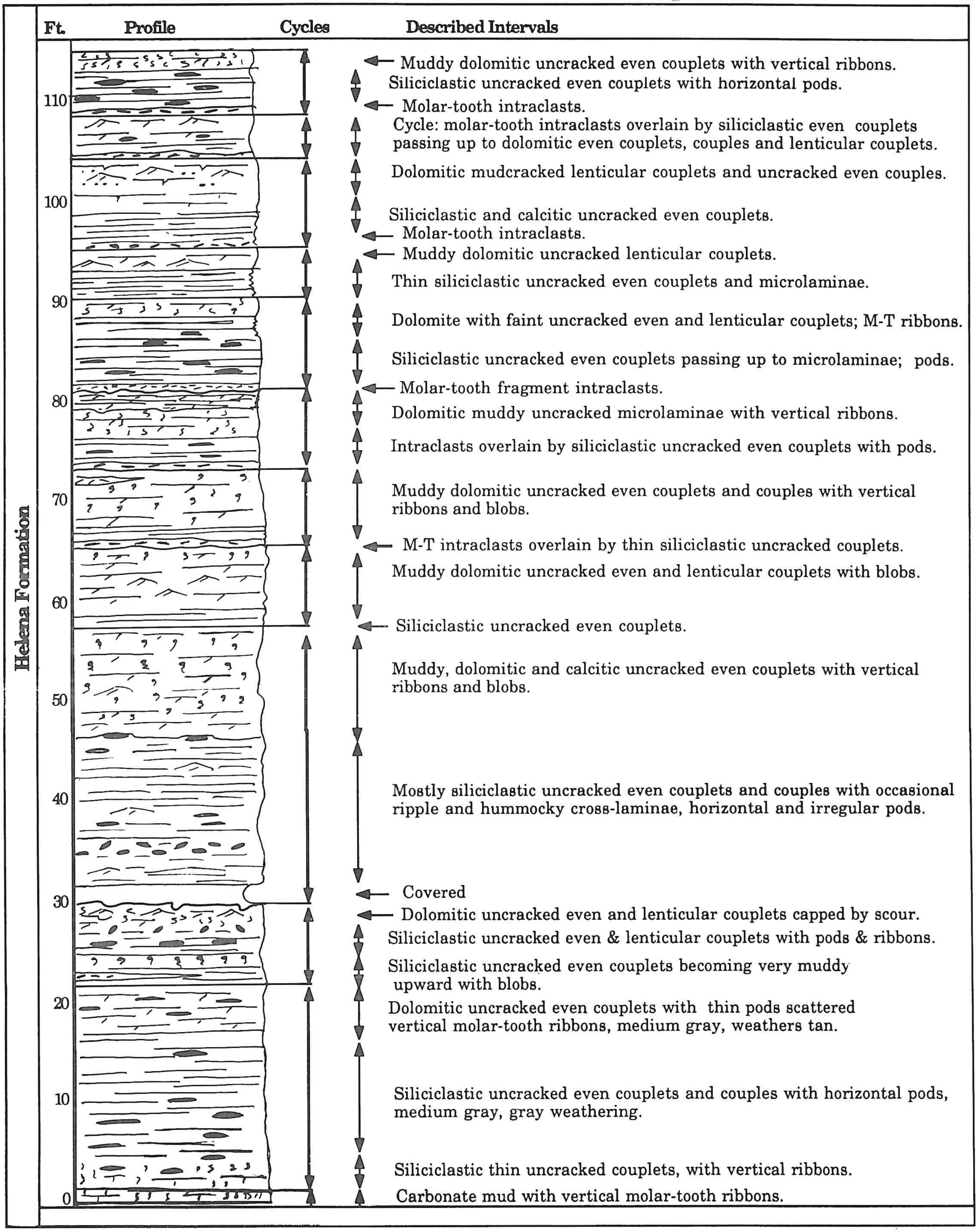


Figure 15. Measured stratigraphic section of stop 3.

influx was reduced and lake waters became supersaturated with respect to $Ca Mg (CO_3)_2$, precipitating protodolomite in clay-rich sediments of the upper half-cycles, and 4) exposure and scour beneath the next cycle.

Diagenesis is evidenced by pods, molar-tooth ribbons and blobs. The vertical succession of horizontal pods to vertical pods, to vertical pods surrounding vertical molar-tooth ribbons to ribbons without pods was recognized by O'Connor (1967, 1972) and commonly overprints Helena depositional cycles.

Return to vehicles and backtrack to intersection of Hungry Horse Dam road and U.S. Highway 2.

1.4

69.3 Intersection of U.S. Highway 2 and road leading to the upper part of Hungry Horse Dam. Take a hard right and follow paved road to Hungry Horse Dam visitor facilities. Road cuts on left expose the Helena Formation.

1.0

70.3 Road to North Lion Lake Picnic Area on left.

0.5

70.8 South Lion Lake Picnic area on left. Road cuts continue in the Helena.

0.7

71.5 Road curves to left above the canyon of the South Fork of the Flathead River. Helena cycles stratigraphically above those seen at Stop 3 are exposed in road cuts on left.

1.1

72.6 Hungry Horse Dam Recreation Area sign on right. More cycles on left.

0.5

73.1 **STOP 4** Pull right into parking area with covered picnic tables. Cross road to examine Helena cycles (Fig. 16) about in the middle third of the Helena Formation, above those seen at stop 3, but below the *Baicalia-Conophyton* cycles. Watch out for the falling rocks.

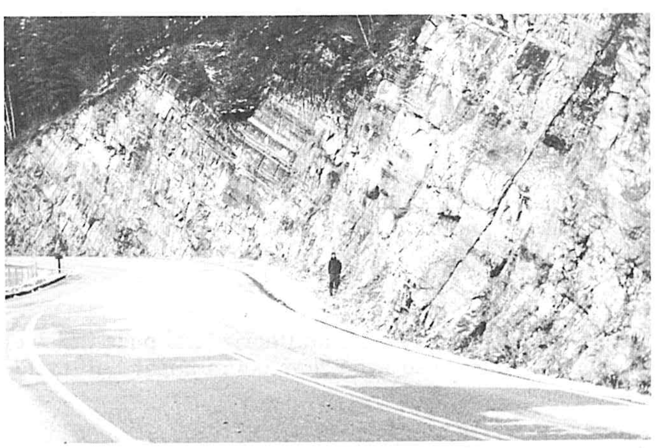


Figure 16. Roadcuts in the middle part of stop 4.

These cycles are similar to those seen at stop 3 in that they have clearly recognizable gray-weathering lower siliciclastic half-cycles and tan weathering dolomitic upper half-cycles (Fig. 17). They also fine upward from muddy couplets and couples to extremely muddy and dolomitic upper half-cycles. Bases of the cycles are sharp and marked by molar-tooth intraclast beds and a few domal stromatolite layers. Cycles with siliciclastic uncracked even couplets in their lower half-cycles and dolomitic uncracked even couplets in their upper half-cycles are type 4 cycles. Cycles high in the section have fine sandy and silty pinch-and-swell couplets and couples in the lower half-cycles and represent type 5 cycles. Diagenetic features include pods in the lower half-cycles that grade up into vertical molar-tooth ribbons and blobs in the upper half-cycles. In the upper part of the section molar-tooth ribbons also occur in the lower half-cycles.

As in the previous stops, these cycles are interpreted to record expansion and contraction of the Great Belt Lake. The scoured surfaces mantled by molar-tooth fragments are interpreted to record subaerial exposure and scour, although evidence for desiccation is missing. Stromatolites coated the higher knobs of the erosion surfaces, probably as the lake expanded across those surfaces during humid climatic intervals. As at stop 3 the uncracked even couplets in the lower type 4 half-cycles possibly record suspension transport by floods or more probably by storms. The introduction of very fine-grained, hummocky crosslaminated pinch-and-swell couples and couplets of type 5 cycles record storm erosion and fine sand transport. These stronger storms may reflect increased fetch and depth of the lake in the upper part of the section. The fining upward in the cycles by reduction in silt percent is interpreted to record smaller floods during more arid climatic intervals and reduction of storm intensity, probably as a result of diminished depth and fetch in a shrinking lake. Lake waters became supersaturated with respect to carbonate ions, and high magnesium calcite or protodolomite mud was precipitated.

To see whether lake expansion and contraction were reflected in the carbon and oxygen isotopes, Tim Lyons sampled through two cycles at one foot intervals from 5 to 25 feet in the measured section. The results of his analysis are presented in the Belt Symposium III Abstracts Volume (Lyons and others, 1993). Lyons and others report that $\partial^{18}O$ values ranged from -12 to $-13^{o}/oo$, too light, they conclude, to have been precipitated from marine water. The $\partial^{13}C$ values ranged from -1 to $+0.5^{o}/oo$. No systematic shift in isotope values was recorded from bottoms to tops of the two cycles. Clearly, the isotopic study at this point is preliminary, but the consistency of the results is very encouraging, and isotopic study of the Belt will continue.

Return to vehicles and backtrack to U.S. Highway 2.

3.8

76.9 Intersection with U.S. Highway 2 turn left toward West Glacier.

0.6

77.5 Martin City on right.

1.7

79.2 Enter Coram.

7.2

86.4 West Glacier. Turn left under railroad overpass, continue through village and cross Middle Fork of the Flathead River.

0.9

87.3 Glacier National Park entrance.

1.1

88.4 Apgar Village intersection. Turn right.

0.9

89.3 Lake McDonald on left.

8.1

97.4 Lake McDonald Lodge on left.

2.1

99.5 Turnout on left to Prichard and bridge.

0.5

100.0 Prichard in roadcut on right.

1.0

101.0 Appekunny member 5 and falls.

0.9

101.9 Cross Avalanche Creek and pass Avalanche Creek Campground.

0.9

102.8 Red argillite on the right marks the base of the Grinnell Formation above the Appekunny.

4.5

107.3 Green argillite of the Empire Formation crops out in cliffs on right. Above the Empire are roadcuts in dark gray muddy carbonate of the Helena Formation. Cycles in this interval are better displayed on the east side of the Lewis plate at Stop 5.

1.3

108.6 Small outcrop of *Conophyton* on right. The *Baicalia-Conophyton* cycles will be seen at Stop 6.

0.3

108.9 Pass through tunnel in the upper Helena Formation.

0.9

109.8 Redbeds on right mark the base of the Snowslip Formation of the Missoula Group. This outcrop is discussed in a stop on Field Trip A1 and B1 (Whipple and others, this volume)..

0.2

110.0 Hairpin turn of the loop cut in red and green argillite of the lower Snowslip. From here Going-to-the-Sun Road climbs to Logan pass through the east limb of the Akamina syncline. Because the dip of the beds is steeper than the road in the

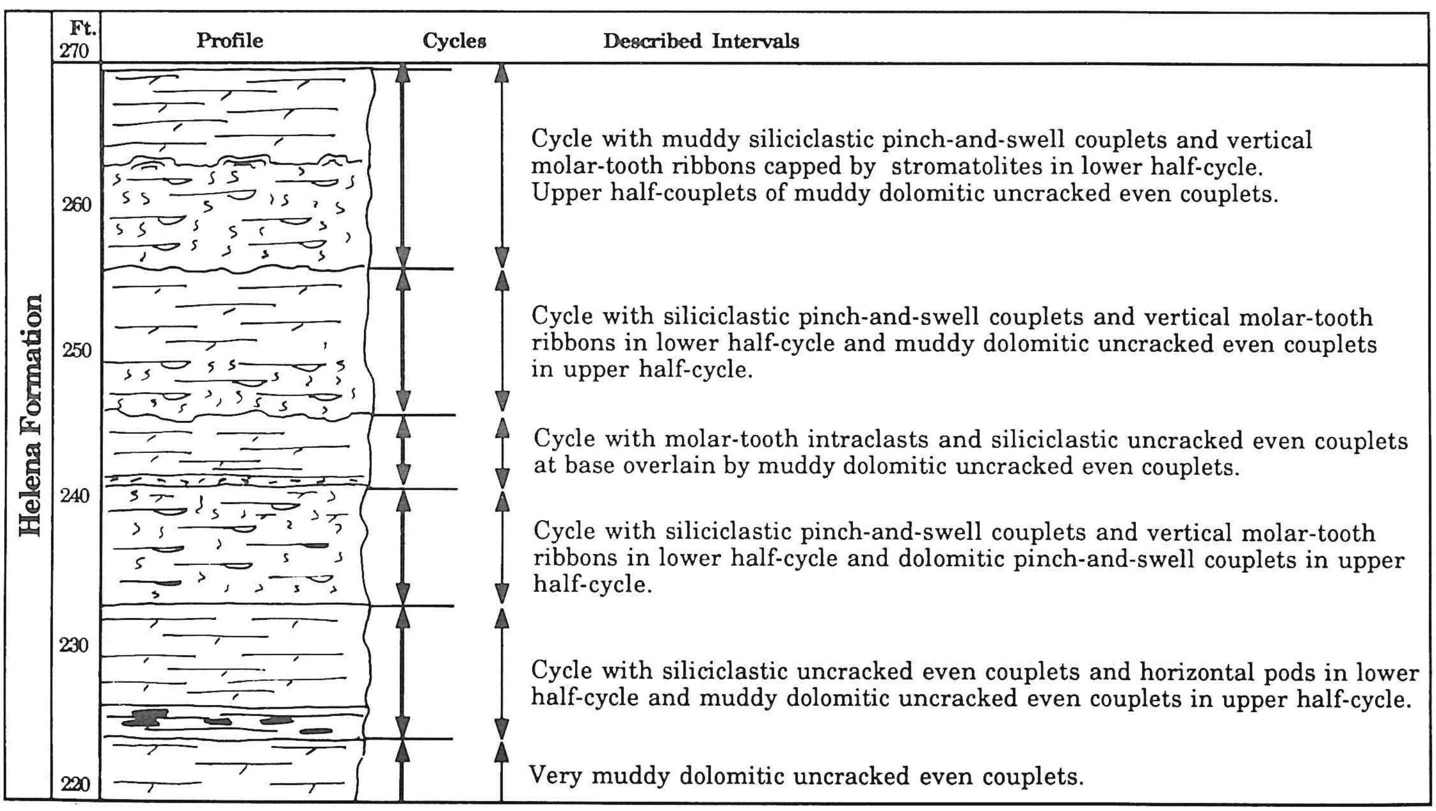


Figure 17c. Measured stratigraphic section of the upper part of stop 4.

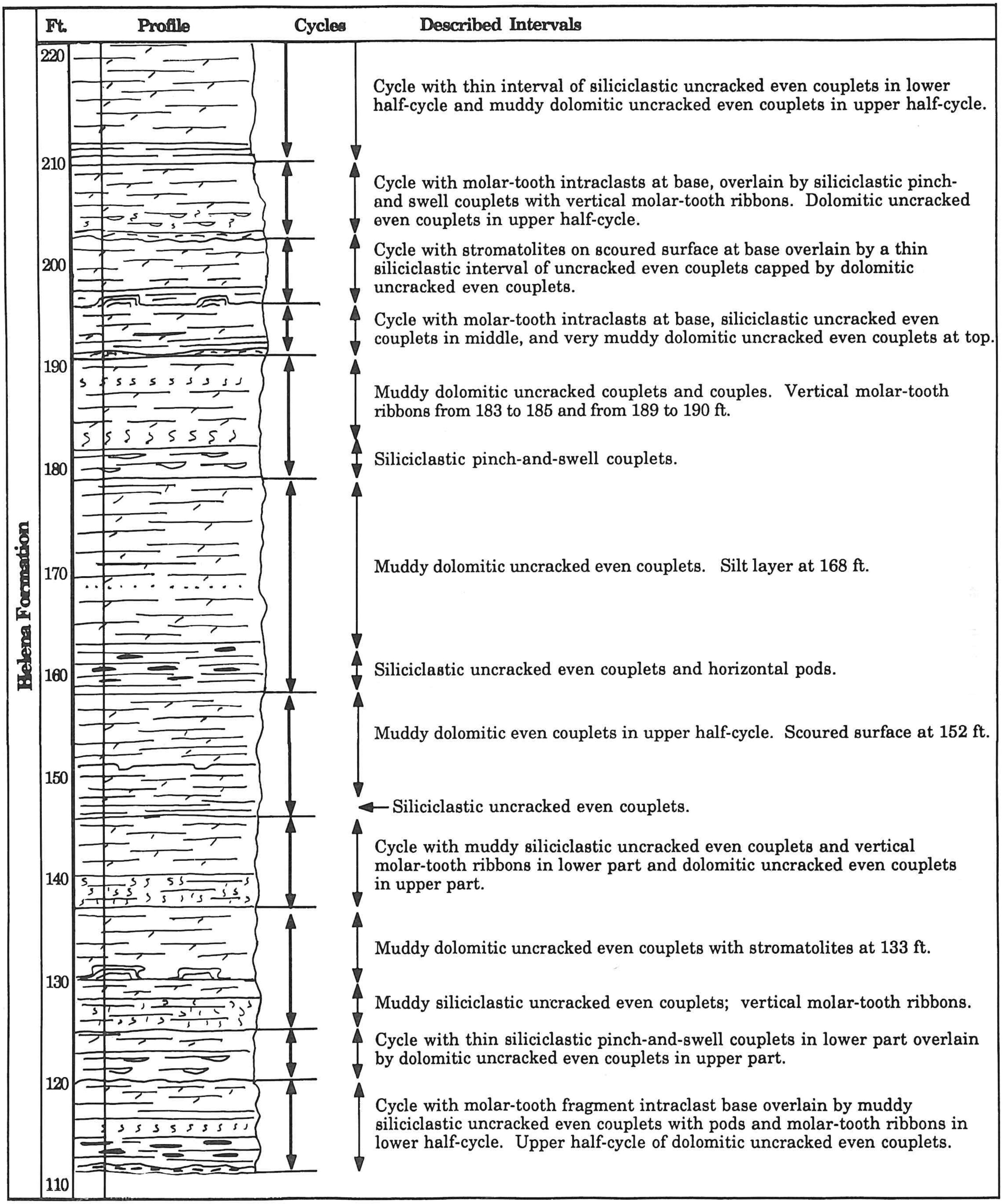


Figure 17b. Measured stratigraphic section of the middle part of stop 4.

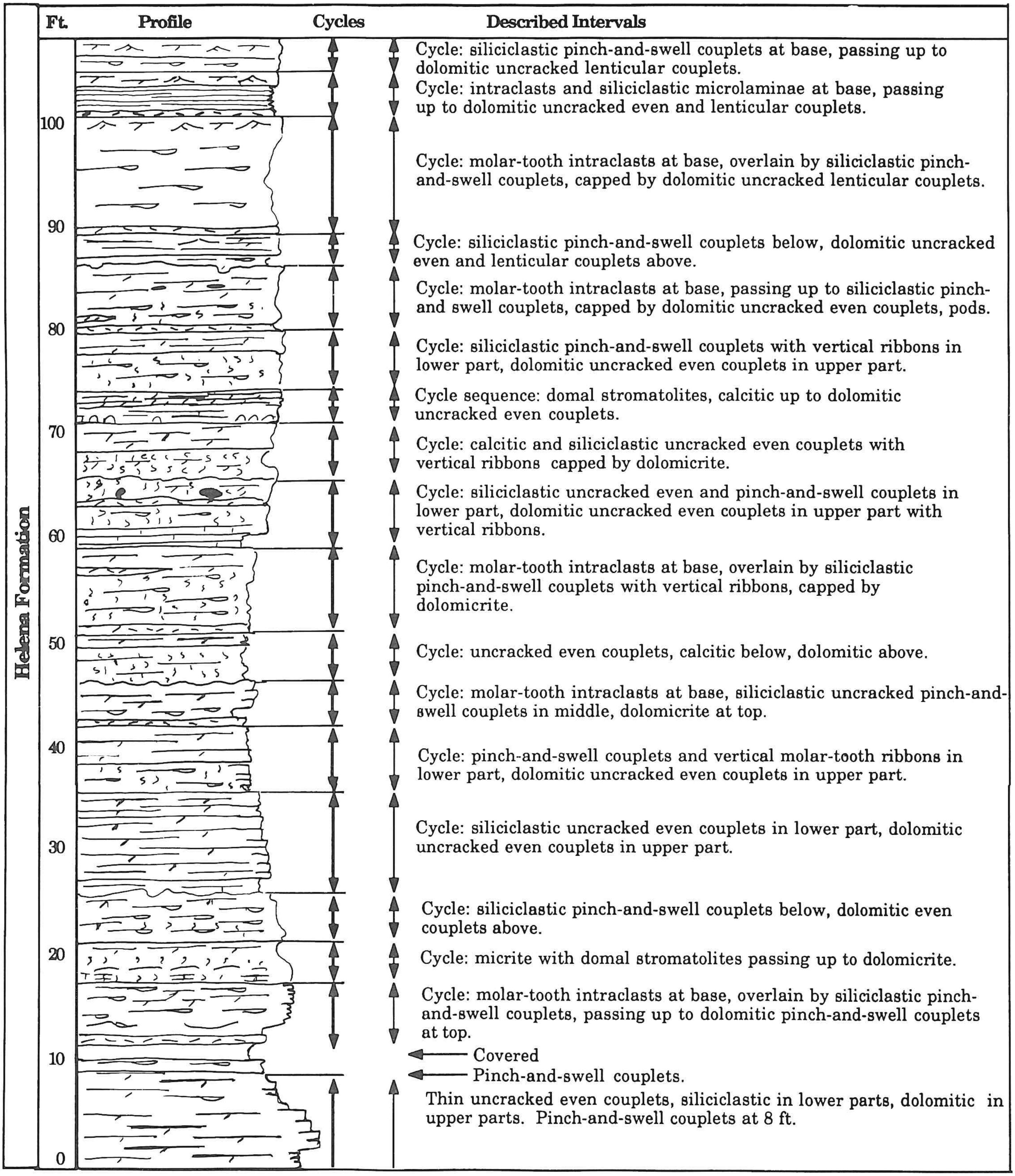


Figure 17a. Measured stratigraphic section of the lower part of stop 4.

lower part of the climb, the road cuts down section into the Helena. It then cuts back up section through the *Baicalia-Conophyton* cycles at Logan Pass.

0.6

110.6 Lowest redbeds in roadcuts on left mark the base of the Snowslip, roadcuts beyond display well developed cycles in the Helena Formation, but traffic prohibits our viewing them here.

0.8

111.4 Thick massive beds on left are the *Baicalia-Conophyton* cycles again. These beds form distinctive ledges that can be easily mapped in Glacier National Park (Rezak, 1957; Horodyski, 1989b). Road continues to cut down into the Helena Formation, and then begins to cut up section.

5.7

117.1 Thick ledges in curve on the left are *Baicalia* Conophyton cycles once again. Here they are contact metamorphosed by the Logan sill. Road curves to left ahead where mountain goats get their kicks by licking up antifreeze from boiled over car radiators.

0.8

117.9 Logan Pass summit and visitors center. Field trip B 5 (Raup and others, this volume) begins here and follows the Highline trail north to Granite Park Chalet.

1.0

118.9 Roadcuts on the left above the curve at Lunch Creek are once again in the *Baicalia-Conophyton* cycles. This outcrop will be seen at stop 6.

0.8

119.7 **STOP 5** Pull out on the upper of two turnouts on right. The dark gray and tan banded cliffs above the road display cycles in the lower to middle part of the Helena Formation (Fig. 18). Walk down to gully

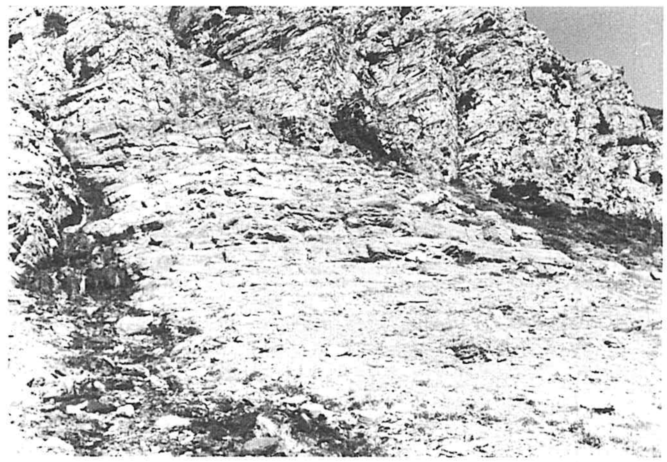


Figure 18. View of stop 5.

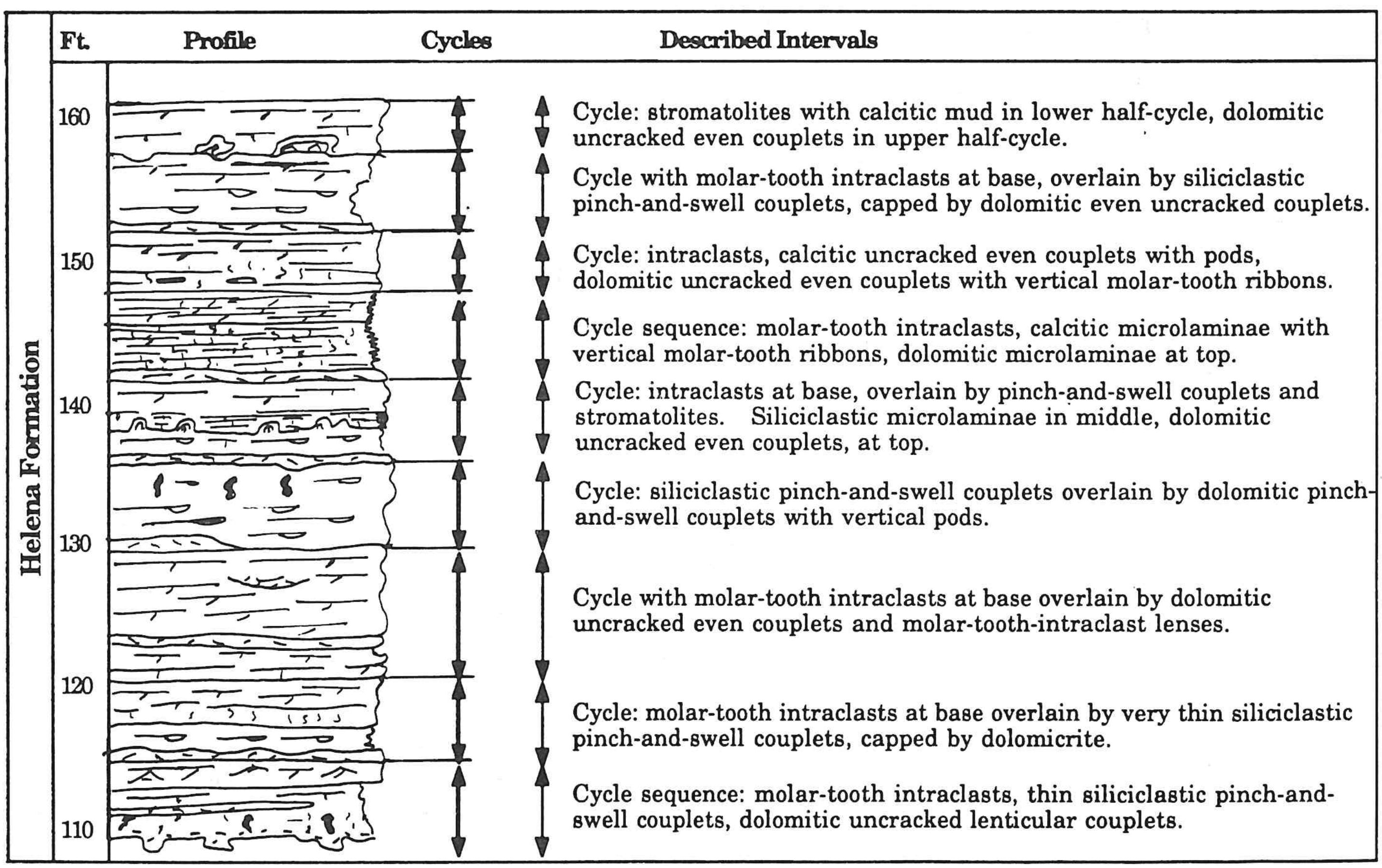


Figure 19b. Measured stratigraphic section of upper part of stop 5.

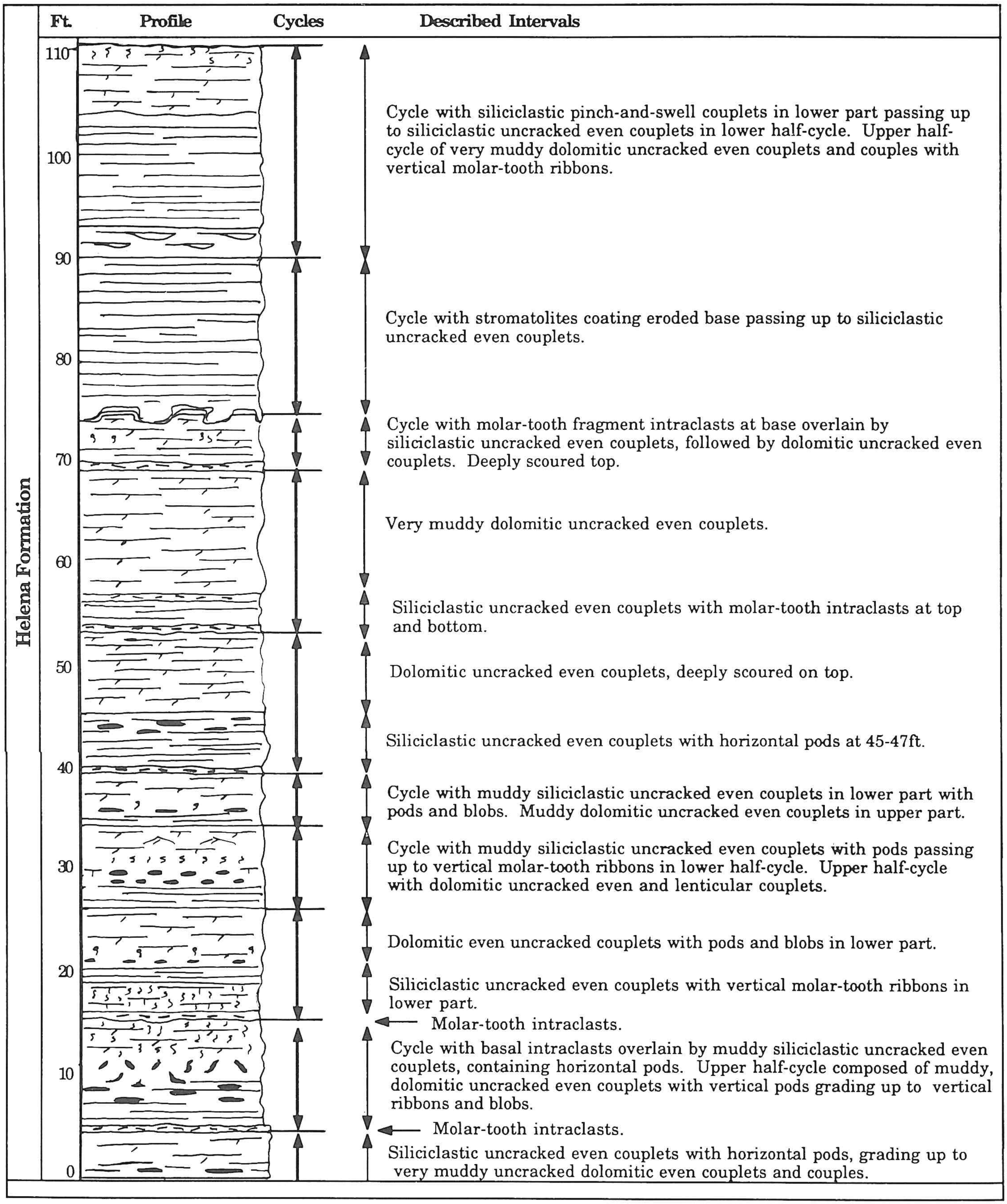


Figure 19a. Measured stratigraphic section of lower part of stop 5.

opposite turnoff and follow path up gully to open scree slope. Cross to north side of drainage to lowest outcrops where measured section begins (Fig. 19).

As in the other Helena cycles, the bases of the cycles are marked by scoured surfaces mantled by molar-tooth fragment intraclasts (Fig. 19). Most of these are type 5 cycles. The lower half-cycles have gray-weathering, silt-to-clay, pinch-and-swell couplets with calcite cement and are overlain by upper half-cycles of dolomitic uncracked even and lenticular couplets and dolomite mud. Other cycles are type 4 cycles with siliciclastic uncracked even couplets in their lower half-cycles and dolomitic uncracked even couplets and dolomite mud in their upper half-cycles. Couplets are mostly thin, and in some intervals grade to microlaminae. Vertical molar-tooth ribbons are common in the upper, dolomitic half-cycles.

These cycles too are interpreted to record expansion and contraction of the Great Belt Lake, but illustrate some important variations on that theme when compared to the other stops. These cycles have thinner couplets nested in thinner cycles than at the other stops. This reduction of the siliciclastic constituents reflects the eastern location of this outcrop, farther from the western and southwestern source of fine grained siliciclastics. At the same time, calcite and dolomite content is proportionately greater, reflecting more carbonate precipitation where siliciclastic influx was diminished. The abrupt change from calcitic micrite in the lower half-cycles to dolomicrite in the upper half-cycles may reflect a change in lake water chemistry from fresher waters during lake highstand to more saline waters during lake contraction.

After returning to vehicles, turn around at Siyeh bend and retrace route back up to Lunch Creek.

1.7

121.4 **STOP 6** Cross Lunch Creek where road curves to the left. Continue 100 meters up the hill and pull off on broad shoulder on left side of road at Going-to-the-Sun Road Geology Stop #8. Cross road to cuts on north side, which expose the two famous *Baicalia-Conophyton* (type 6) cycles (Fig. 20), first analyzed by Horodyski (1983, 1989b), and later by Stickney (1991).

Below the lowest *Baicalia-Conophyton* cycle are three cycles and the upper part of a fourth (Fig. 21), containing lower half-cycles of siliciclastic pinch-and-swell couples and upper half-cycles composed of thinner bedded, dolomitic pinch-and-swell couplets cut by vertical molar-tooth ribbons. The lowest cycle is thickest, and its base is not exposed. A stromatolite head covers a pinnacle on the eroded surface of the highest

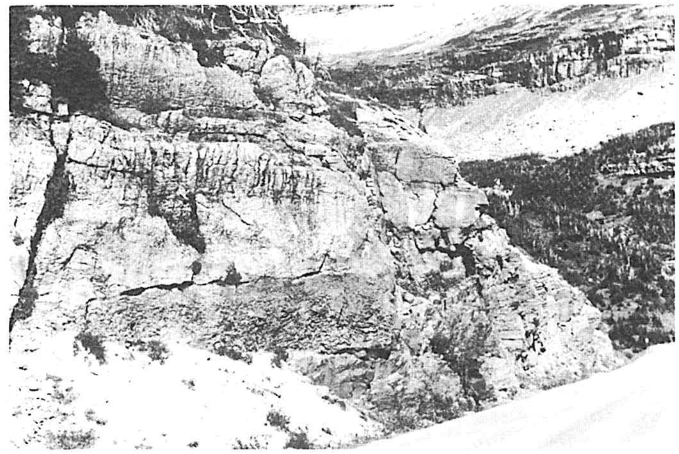


Figure 20. Lower part of stop 6 showing lower *Baicalia-Conophyton* cycle.

cycle. The base of the lower *Baicalia-Conophyton* cycle is marked by an intraclast bed and pinch-and-swell couplets with pods, overlain by thick, dark, siliciclastic pinch-and-swell couples. Near the top of the pinch-and-swell couples is a four-foot-thick bed of dark massive muddy silt containing pyrite crystals. The base of the lower *Baicalia-Conophyton* interval is sharply marked by bioherms of the branching form *Baicalia* separated by nodular masses of light gray micrite and lenses composed of thin rinds of broken *Baicalia* micrite loosely packed in a sparry calcite matrix. The rounded branches of *Baicalia* pass up into vertically stacked, laterally linked columns of the sharply pinnacle shaped stromatolite form *Conophyton* with diameters of 10 cm. These in turn pass upward into larger *Conophyton* heads inclined toward the northeast (Horodyski, 1983, 1989b), at the top of the lowest *Baicalia-Conophyton* cycle, partly covered here.

The base of the upper *Baicalia-Conophyton* cycle is covered in this section, but is marked by thin beds of dark gray siliciclastic pinch-and-swell couplets at the loop. Lowest beds of the upper cycle here consist of small clusters of *Baicalia* that splay upward into larger bioherms. As in the lower *Baicalia-Conophyton* cycle, *Baicalia* branches are surrounded by lenses of *Baicalia* debris set in a sparry calcite matrix and nodular masses of micrite. Above the *Baicalia* bioherms and debris are domal stromatolites (Fig. 22) overlain by carbonate mud in which the original sedimentary structures are largely obliterated by vertical molar-tooth ribbons. Above the molar-tooth ribbons is thinly laminated micrite overlain by vertically stacked, laterally linked *Conophyton* heads that are also linked to the layered micrite beds between the *Conophyton* bioherms. The bioherms spread out over the bedded micrite and terminate sharply beneath thinly laminated lime mudstone with domal stromatolites at the top of the upper *Baicalia-Conophyton* cycle. Pinch-and-swell couples mark the base of the next cycle, bleached white by the overlying Logan sill.

Stromatolites of the *Baicalia-Conophyton* cycles were first recognized by Fenton and Fenton (1933) and named the *Collenia frequens* zone (Fenton and Fenton, 1937). Rezak (1957) redescribed the stromatolites, finding that the cone shaped forms belonged to the genus *Conophyton*. He renamed the unit the *Conophyton* zone 1 and Ross (1959) mapped it extensively across Glacier National Park. Horodyski (1983, 1989b) pointed out the cyclical repetition of *Baicalia* and *Conophyton* and subdivided the cycles into the following lithic units, which he correlated widely throughout the Park: 1) lower *Baicalia* unit, 2) lower small-diameter *Conophyton* unit, 3) lower large-diameter *Conophyton* unit, 4) Middle *Baicalia* unit, 5) middle sedimentary unit, and 6) upper mixed stromatolite unit. Horodyski emphasized the stromatolite growth forms in his study and showed (1989b) that *Baicalia* extended beyond the Park as far west as Sommers, Montana. Stickney (1991) followed Horodyski's work, emphasizing the lithologies in which the stromatolites occur. She found that the *Baicalia-Conophyton* cycles extend as far southwest as Holland Lake, Montana, and that the beds of the *Baicalia-Conophyton* cycles fine and thin upward like those of other Helena cycles. *Baicalia* bioherms surrounded by *Baicalia* intrasparite debris and by micrite characterize the lower half-cycles, and represent the turbulent expansive phase of the cycles. *Conophyton* bioherms in fine-grained pure micrite occupy the upper half-cycles along the eastern, protected side of the basin and reflect a decrease in turbulence. Stickney (1991) interpreted both the sediments and stromatolite growth forms to record turbulent transgressive phases followed by calmer regressive phases.

Ongoing study of the interval from the *Baicalia-Conophyton* cycles up to the base of the Snowslip Formation (Winston, 1993) extends the reach of the two *Baicalia-Conophyton* cycles south to Red Mountain, Montana, west to Greywolf Peak and Libby Dam, and north to South Drywood Canyon, Alberta. Across this area, the lower part of the lower *Baicalia-Conophyton* cycle contains dark gray, siliciclastic pinch-and-swell couples. Whereas in Glacier National Park the pinch-

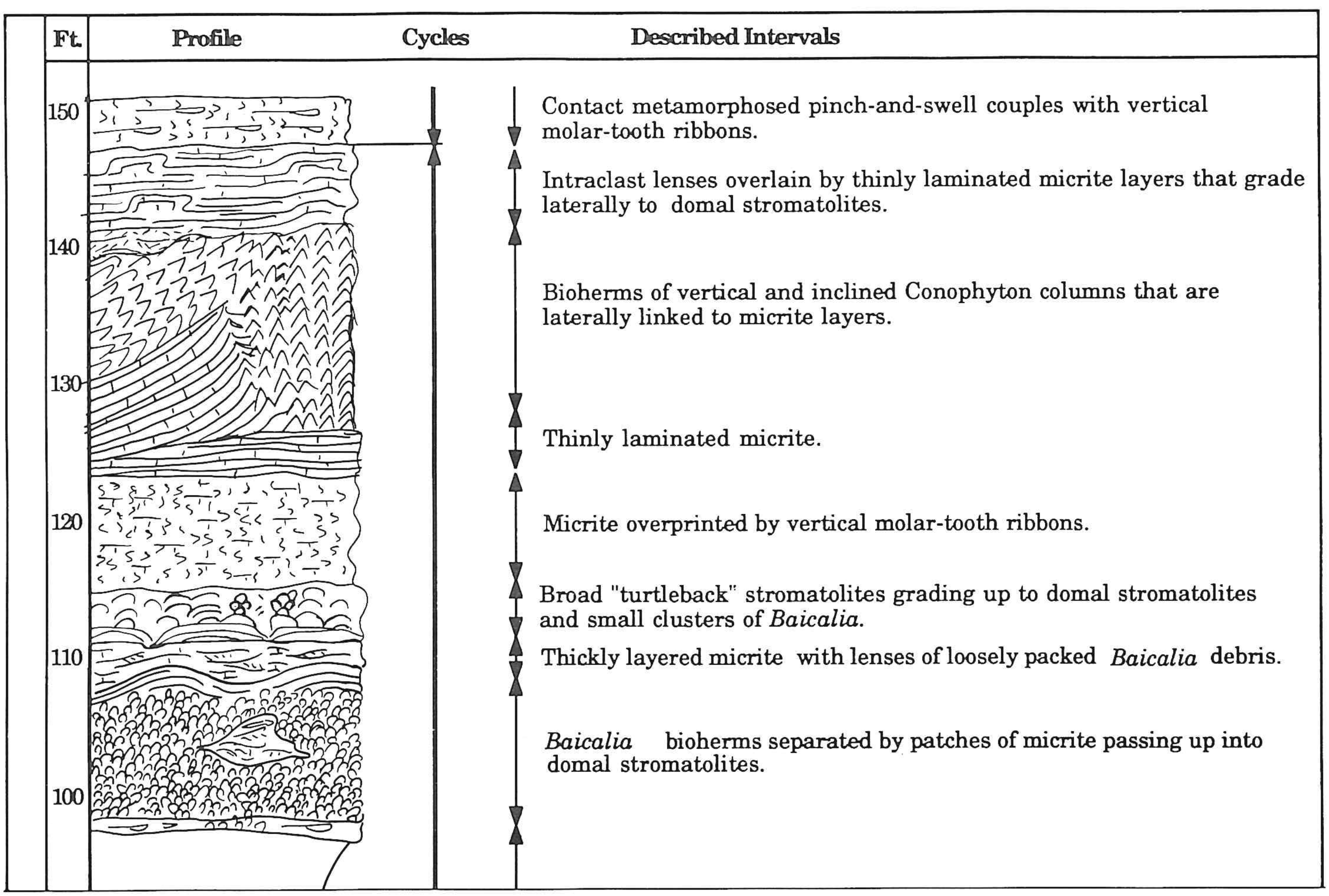


Figure 21b. Measured stratigraphic section of the upper *Baicalia-Conophyton* cycle of stop 6.

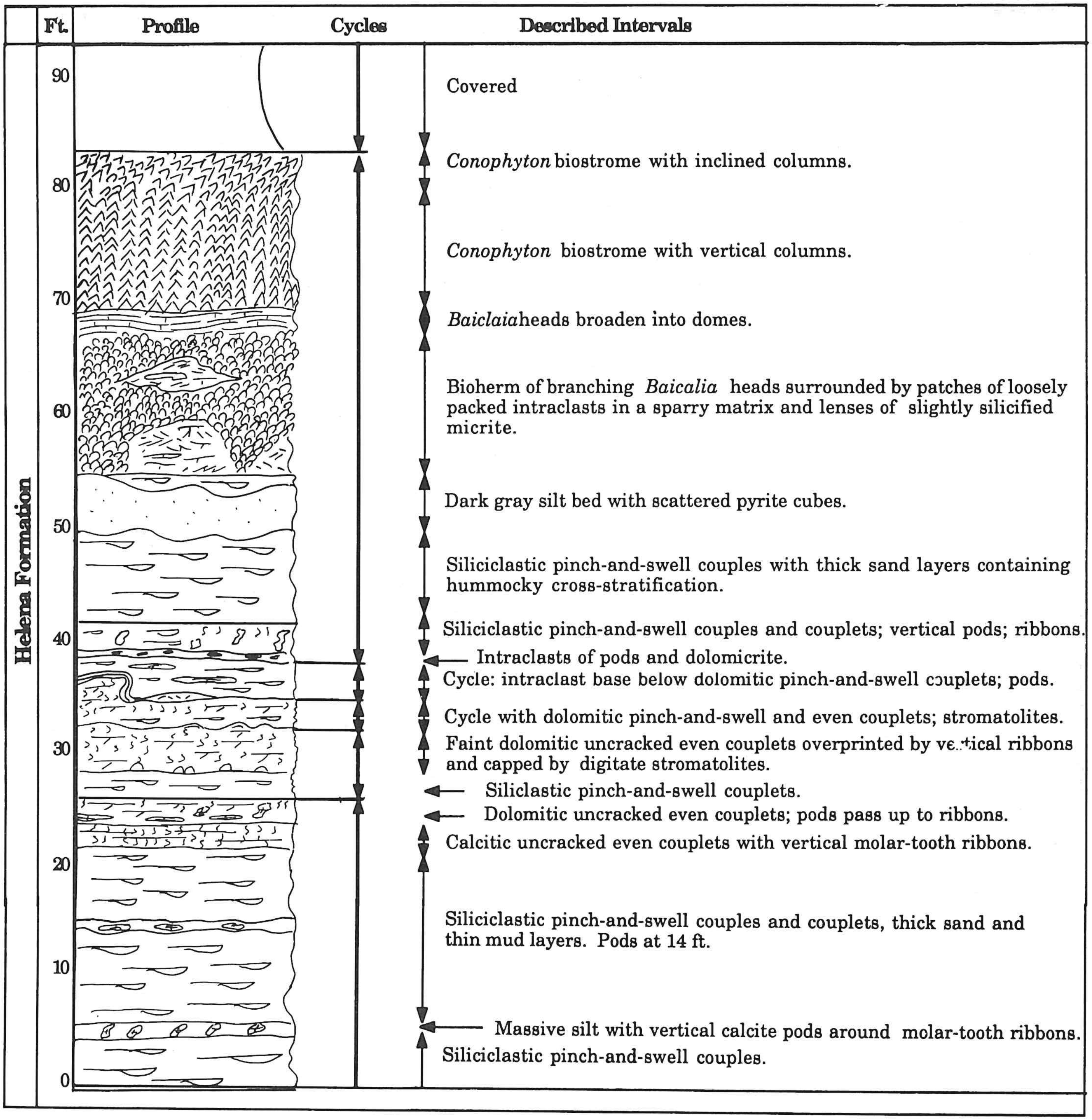


Figure 21a. Measured stratigraphic section of the lower *Baicalia-Conophyton* cycle of stop 6.

and-swell couples are sharply overlain by small *Baicalia* bioherms, west of the Park they are overlain by *Baicalia* debris intrasparite lenses and micrite nodules in turn overlain by *Baicalia* bioherms. *Baicalia* is overlain by *Conophyton* along the eastern margin of the basin from South Drywood Canyon through Glacier National Park, south to Red Mountain. To the west in the Flathead Range *Conophyton* passes to *Baicalia* bioherms and *Baicalia* intraclast and micrite beds. Still farther west in the Swan and Mission ranges and at Libby Dam the *Conophyton* level becomes molar-tooth bearing calcitic pinch-and-swell couples capped by microlaminae at the top of the lower cycle.

The base of the upper *Baicalia-Conophyton* cycle is marked across most of the basin by a thin, sharply defined, siliciclastic interval that fines from silty pinch-and-swell couples in the west to microlaminated black shale in the Swan, Flathead and Livingston ranges, south to Red Mountain. Farther east at South Drywood

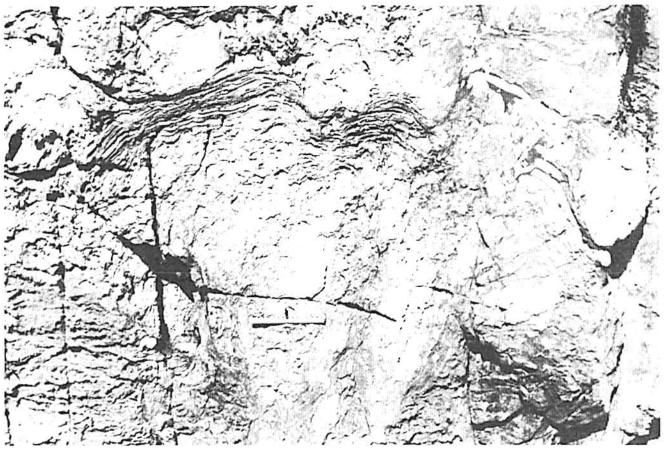


Figure 22. View of *Baicalia* bioherm passing up to domed stromatolites in the upper *Baicalia-Conophyton* cycle.

Canyon the black shale pinches out and *Baicalia* bioherms of the upper cycle rest on *Conophyton* bioherms of the lower cycle. Still farther east at Bald Mountain *Baicalia* is replaced by micritic domal stromatolite heads surrounded by intraclasts. *Baicalia* bioherms, *Baicalia* intraclast debris beds and nodular micrite are spread as widely in the second cycle as in the first, reaching west to Libby, southwest to Greywolf Peak, and south to Red Mountain. In Glacier National Park the *Baicalia* bioherms pass upward through lenses of *Baicalia* intrasparite and micrite, uncracked micritic even couplets and microlaminae (middle sedimentary unit of Horodyski, 1989b) to *Conophyton* bioherms with sharply inclined heads. *Conophyton* heads are laterally linked to the pure micrite that surrounds them. They are overlain by carbonate mud cut by vertical molar-tooth ribbons and capped by thinly laminated micrite, marking the top of the upper cycle. To the west in the Flathead and Swan ranges and to the south at Red Mountain, the upper *Conophyton* interval passes to carbonate mud cut by molar-tooth ribbons. Still farther west in the Mission Range and at Sommers and Libby Dam micrite is mixed with silty lenses of pinch-and-swell couplets capped by a thick microlamina interval.

The regional perspective of each of the two *Baicalia-Conophyton* cycles demonstrates how broadly *Baicalia* became established when conditions were favorable. *Baicalia* bioherms surrounded by intraclastic debris and calcitic mud occupy the position of the lower turbulent half-cycles and record turbulence in the hummocky intraclastic lenses. Because they occupy a single thick stratum, the *Baicalia* bioherms may have grown simultaneously across the northeastern part of the basin during the lower expanded phases of the two lacustrine cycles. Calcite mud precipitation may reflect nearly fresh lake waters. As pointed out by Horodyski (1989b) and Stickney (1991), *Conophyton* growth coincided with calmer conditions accompanied by calcite mud precipitation in Glacier National Park. More siliciclastic uncracked even couplets followed by microlaminae accumulated farther west, completing the fining-upward cycles.

The calcitic *Baicalia-Conophyton* cycles may record the greatest first order expansion of the Great Belt Lake to a point where the lake may have been open with an outlet, and may have been nearly fresh. Eastward shift of the Wallace facies over the Helena facies at the *Baicalia-Conophyton* level in the Whitefish Range (J. Whipple, oral communication, 1992)) supports this expansion. The interval from the base of the *Baicalia-Conophyton* cycles to the top of the Helena (base of the Snowslip Formation) has been subdivided into 5 informal units (Winston, 1993). Cycles in the lower units are dominantly calcareous, but upper half-cycles become progressively more dolomitic, until dolomite dominates the uppermost part of the Helena Formation. This increase in dolomite coupled with the increasing evidence of subaerial exposure are interpreted to indicate that above the *Baicalia-Conophyton* cycles that the Great Belt Lake progressively shrank and became more restricted until continental redbeds of the basal Snowslip Formation spread across the eastern part of the basin.

This is the last planned stop. Return to Grouse Mountain Lodge.

35

154.4 West Glacier and intersection with U.S. Highway 2. Turn right and go back through Coram, Hungry Horse toward Columbia Falls.

15.7

170.1 Intersection of U.S. Highway 2 and Montana Highway 206. Turn right and stay on U.S. Highway 2 through Columbia Falls.

5.4

Intersection of U.S. Highway 2 and Montana Highway 40. Continue straight on Montana 40.

4.6

180.1 Intersection of Montana Highway 40 and U.S. Highway 93. Turn right and continue on Highway 93 to Grouse Mountain Lodge.

3.4

183.5 Grouse Mountain Lodge.

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Alluvial cone below Sexton Glacier. *View is to the north from the middle reaches of Baring Creek, Glacier National Park.*

GUIDE TO THE GEOLOGICAL SETTING OF THE MIDDLE PROTEROZOIC SULLIVAN SEDIMENT-HOSTED Pb-Zn DEPOSIT, SOUTHEASTERN BRITISH COLUMBIA

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Contribution No. 17, Sullivan-Aldridge Project

INTRODUCTION

This field guide focuses on the nature of the Middle Proterozoic Aldridge Formation in the vicinity of the Sullivan mine, southeastern British Columbia. The field guide pays special attention to the nature of the world class Sullivan stratiform Zn-Pb deposit, and to sedimentation, basinal environment, basin structural controls, and the effects of gabbro intrusions and hydrothermal activity during deposition of the Aldridge Formation. This broad view is an attempt to establish a basinal perspective for the origin of the Sullivan deposit.

The character and setting of the Sullivan deposit is the focus of a joint research program led by the Geological Survey of Canada in collaboration with the British Columbia Geological Survey Branch, U.S. Geological Survey, industry and various universities. This field guide represents a part of this research project. A preliminary version (Turner and others, 1992) was published by the British Columbia Geological Survey Branch.

The objectives of this field guide are to demonstrate:
(1) the general structure of the Aldridge Formation (Stops 1-1, 1-5);
(2) the sedimentology of the lower Aldridge (Stops 1-2, 2-2) and middle Aldridge (Stops 1-3, 1-4) Formation;
(3) the distribution and character of the gabbro sill complex within the lower and middle Aldridge Formation, and alteration of sedimentary rocks adjacent to sill contacts (Stops 1-3, 1-7, 1-9, 2-1, 2-3, 2-6);
(4) evidence for minor volcanism during deposition of Aldridge Formation (Stop 1-9);
(5) the nature of hydrothermal alteration and breccia-conglomerate within the Aldridge Formation (Stops 1-6, 1-7, 1-8, 1-9, 1-10, 2-4, 3-2, 3-3);
(6) the nature of the Sullivan stratiform zinc-lead deposit (Stop 3-1) and its relation to the district-scale corridor of altered, mineralized and fragmented strata (Stops 3-2, 3-3); and
(7) the nature of local metamorphism, deformation, and magmatism (Stops 1-1, 2-5, 2-7 and 2-8).

To accomplish these objectives, three field tours are described. The route of **Tour 1** starts in Kimberley and follows highways 95A, 95 and 3 to the vicinity of Creston (Fig. 1). Tour 1 focuses on stratigraphy, gabbro sills and related sediment alteration (Table 1). **Tour 2** focuses on field relationships in the St. Mary valley southwest of Kimberley that illustrate the gabbro sill complex within the lower Aldridge Formation and sill-sediment relationships as well as regional

Turner, R.J.W., Hoy, T, Leitch, C.H.B., Anderson, Doug, and Ransom, Paul, 1993, Guide to the geological setting of the Middle Proterozoic Sullivan sediment-hosted Pb-Zn deposit, southeastern British Columbia, *in* Link, P.K., ed., **Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada:** Belt Symposium III Field Trip Guidebook, Belt Association, Spokane, Wa., p. 53-94.

metamorphism, structure and younger Proterozoic intrusions. **Tour 3** includes the Sullivan mine and adjacent district-scale alteration in the vicinity of Kimberley.

The guidebook uses material from earlier mapping (*e.g.* Höy, 1984b, Höy and Diakow, 1982; Höy and Carter, 1988) and field trips (*e.g.* Höy and others, 1985; Turner and others, 1992). New data and ideas come from recent reports of the B.C. Geological Survey Branch (*e.g.* Höy, 1989; 1993) and Geological Survey of Canada (*e.g.* Leitch, 1991, 1992a, 1992b; Leitch and others, 1991; Leitch and Turner, 1991, 1992; Turner and Leitch, 1992).

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TOUR 1	KIMBERLEY TO CRESTON: REGIONAL SETTING, ALDRIDGE FORMATION
STOP	
1-1	View, setting of Sullivan mine, North Star Ski area
1-2	Lower Aldridge Formation, Mark Creek, Marysville
1-3	Middle Aldridge Formation and Moyie gabbro-sediment contact, Lumberton
1-4	Marker horizon, Moyie River
1-5	Moyie Fault zone, Kitchener Formation
1-6	Pebble wacke pipe, middle Aldridge Formation, Moyie Lake
1-7	Gabbro sill, granophyre and fragmental, Kitchener
1-8	Albitite alteration, middle Aldridge Formation, N. Goat River Road
1-9	St. Joe prospect, volcanic tuff and channel deposit, Middle Aldridge Fm.
1-10	St. Joe prospect, discordant breccias, tourmalinite in Middle Aldridge Fm.
TOUR 2	ST. MARY RIVER VALLEY: GABBRO SILL COMPLEX AND PURCELL AGE MAGMATISM AND TECTONISM
STOP	
2-1	Gabbro-sediment contact, St. Mary River Rd.
2-2	Lower Aldridge Formation, Meachan Cr. Road
2-3	Gabbro-sediment contacts, lower Aldridge Formation, Meachan Cr. Rd
2-4	Fragmental, Lower Aldridge Formation, Meachan Cr. Rd
2-5	Hellroaring Cr. stock, Hellroaring Cr. Rd
2-6	View of gabbro sill-sediment stratigraphy, Bootleg Mtn., St. Mary Valley
2-7	Large fold structure, St. Mary Valley
2-8	Lower Aldridge Formation schist, Matthew Creek
TOUR 3	SULLIVAN MINE AND DISTRICT SETTING
STOP	
3-1	Sullivan mine tour
3-2	Sullivan-North Star corridor, view stop, Sullivan mine road
3-3	Alteration and fragmentals, North Star Hill

Table 1. Summary of field trip stops.

GENERAL GEOLOGY OF THE ALDRIDGE FORMATION

Structure

The Purcell Supergroup rocks in southeastern British Columbia and Alberta, and correlative Belt Supergroup rocks in adjacent United States comprise a thick sequence of clastic and carbonate rocks of Middle Proterozoic age (Price, 1962; Harrison, 1972). In Canada, the Purcell Supergroup lies within the Foreland Thrust and Fold belt, characterized by shallow, easterly verging thrust faults and broad open folds.

The structure of the western part of the Foreland Thrust and Fold belt is dominated by a north-plunging regional anticlinorium structure cored by the Middle Proterozoic Purcell Supergroup and the Late Proterozoic Windermere Supergroup and flanked by Paleozoic miogeoclinal rocks (Fig. 2). The anticlinorium is allochthonous, carried eastward onto underlying cratonic basement by generally north-striking thrust faults during the Laramide orogeny in Jurassic to Paleocene time (Price, 1981). The anticlinorium is cut by prominent northeast-trending faults, including the Moyie and St. Mary faults (Fig. 2) that merge to the east with north-striking thrust faults. These faults are offset by north trending, west-side down normal faults, such as the Gold Creek and Rocky Mountain Trench faults of Eocene age (Fig. 3).

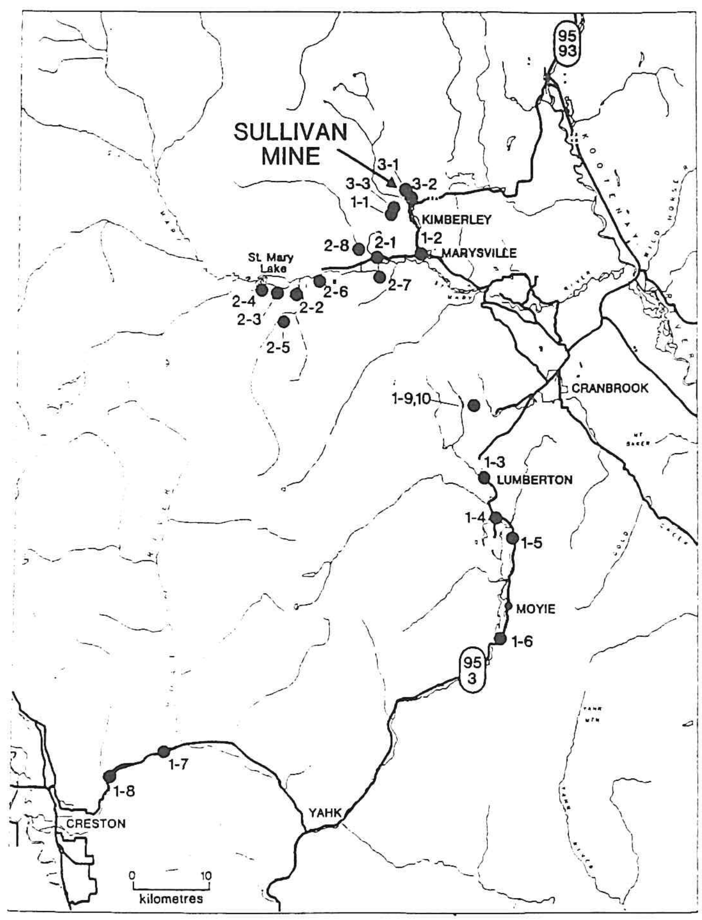

Figure 1. Route map showing locations of field stops for three field tours.

The northeast-trending structures are within or parallel to a broad structural zone that cuts the anticlinorium, crosses the Rocky Mountain Trench and extends northeastward across the Foreland Thrust belt (Fig. 2). A conspicuous change in regional Mesozoic structural grain from northward to northwestward occurs from north to south across this zone. As well, fundamental changes in thickness and facies of sedimentary rocks that range in age from Middle Proterozoic to early Paleozoic are localized here. These changes appear to reflect old Middle Proterozoic structures formed during filling of the Purcell basin that have been reactivated repeatedly during younger periods of tectonism and deformation (Stop 1-5). This structural zone coincides with the western extension of a proposed northeast-trending Proterozoic rift that lies below Phanerozoic cover rocks to the northeast (Kanasewich, 1968; Kanasewich and others, 1969).

Stratigraphy

The Aldridge Formation (Schofield, 1915) is at the base of the Purcell Supergroup in the Purcell Mountains and Lizard Range and overlies Fort Steele Formation in the Hughes Range (Figs. 2 and 4; Hoy, 1993). Reesor (1958) has divided the Aldridge Formation in the Purcell Mountains into three informal units: rusty weathering siltstone, quartz wacke and argillite of the lower Aldridge Formation; grey weathering quartz wacke and siltstone of the middle Aldridge Formation; and laminated argillite of the upper Aldridge Formation. In the Hughes Range, the lower and middle Aldridge Formation much thinner and overlie a distinctive carbonate-bearing unit transitional with underlying fluvial quartzites of the Fort Steele Formation (Hoy, 1993). Elsewhere, the base of the Aldridge Formation is not exposed.

The Aldridge Formation is correlative with the upper part of the Fort Steele Formation in the Hughes Range, the Prichard Formation in Montana, Idaho and Washington (Harrison, 1972) and the Haig Brook, Tombstone Mountain, Waterton and Altyn formations further east (Price, 1964; Fermor and Price, 1983). The thickness of the Aldridge Formation varies from about 2000 m in the northern Hughes Range (Hoy, 1982) to in excess of 4200 m in the southern Purcell Mountains (Edmunds, 1977). To the south, the correlative Prichard Formation is as much as 6000 m thick (Cressman, 1989).

The maximum age of the Aldridge and Prichard formations is constrained by the age of underlying basement and the minimum age by the age of intrusions and deformation. A zircon U-Pb age from the pre-Belt gniess of the Priest River Group in Idaho is 1576 Ma (Evans and Fischer, 1986). The Moyie sills, dated at 1433 Ma (Zartman and others, 1982) and 1445 Ma (Hoy, 1993) by zircon U-Pb method, are interpreted to be penecontemporaneous with deposition of Aldridge sediments (Hoy, 1989). The Hellroaring Creek stock intrudes deformed Aldridge Formation (Ryan and Blenkinsop, 1971) and has recently been dated at 1365 Ma (J. Mortenson, personal communication, 1992).

The Aldridge and Prichard formations form the lower one-forth to one-third of the Purcell and Belt supergroups and differ from overlying Belt Purcell strata by dominance of deep water turbidite deposits, abundant and widespread gabbro sills, interbedded laminated organic-rich siltite, and abundant iron sulfide (Cressman, 1989). Aldridge Formation rocks are composed predominantly of quartz, plagioclase, muscovite, biotite and chlorite. They are regionally metamorphosed to biotite and garnet zone greenschist facies.

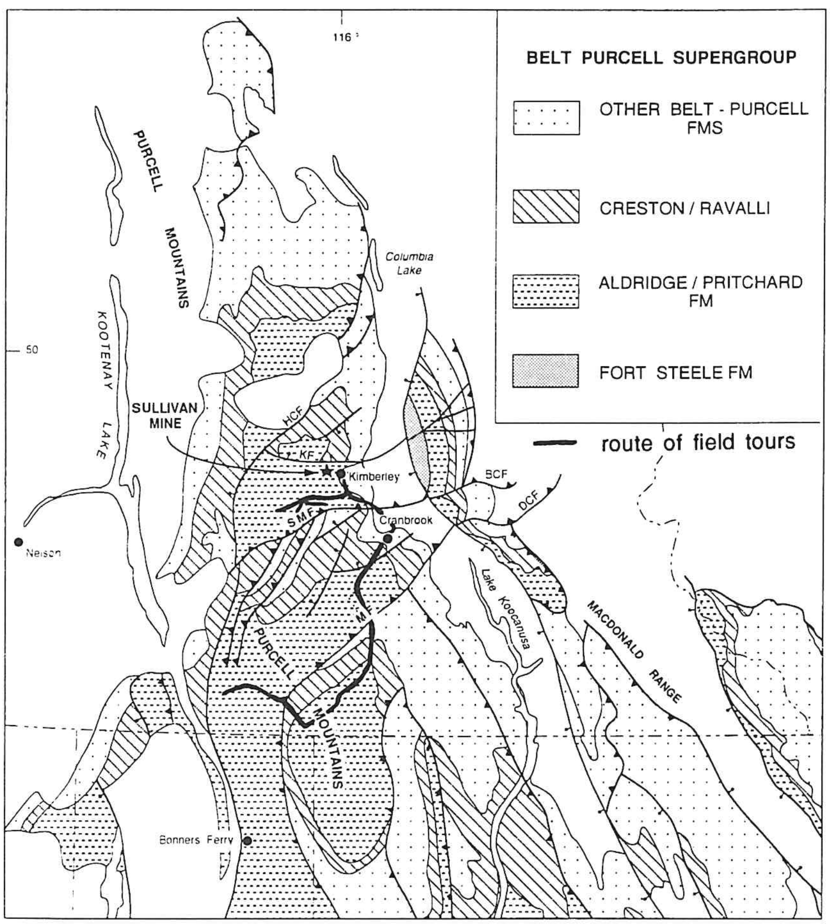

Figure 2. Geological map of Belt-Purcell Supergroup in southeastern British Columbia and adjacent parts of Montana and Idaho (modified from Winston, 1986a; Welbon, 1993). General route of field trip is indicated with heavy black line. HCF = Hall Lake fault; KF = Kimberley fault; SMF = St. Mary fault; MF = Moyie fault; DCF = Dibble Creek fault.

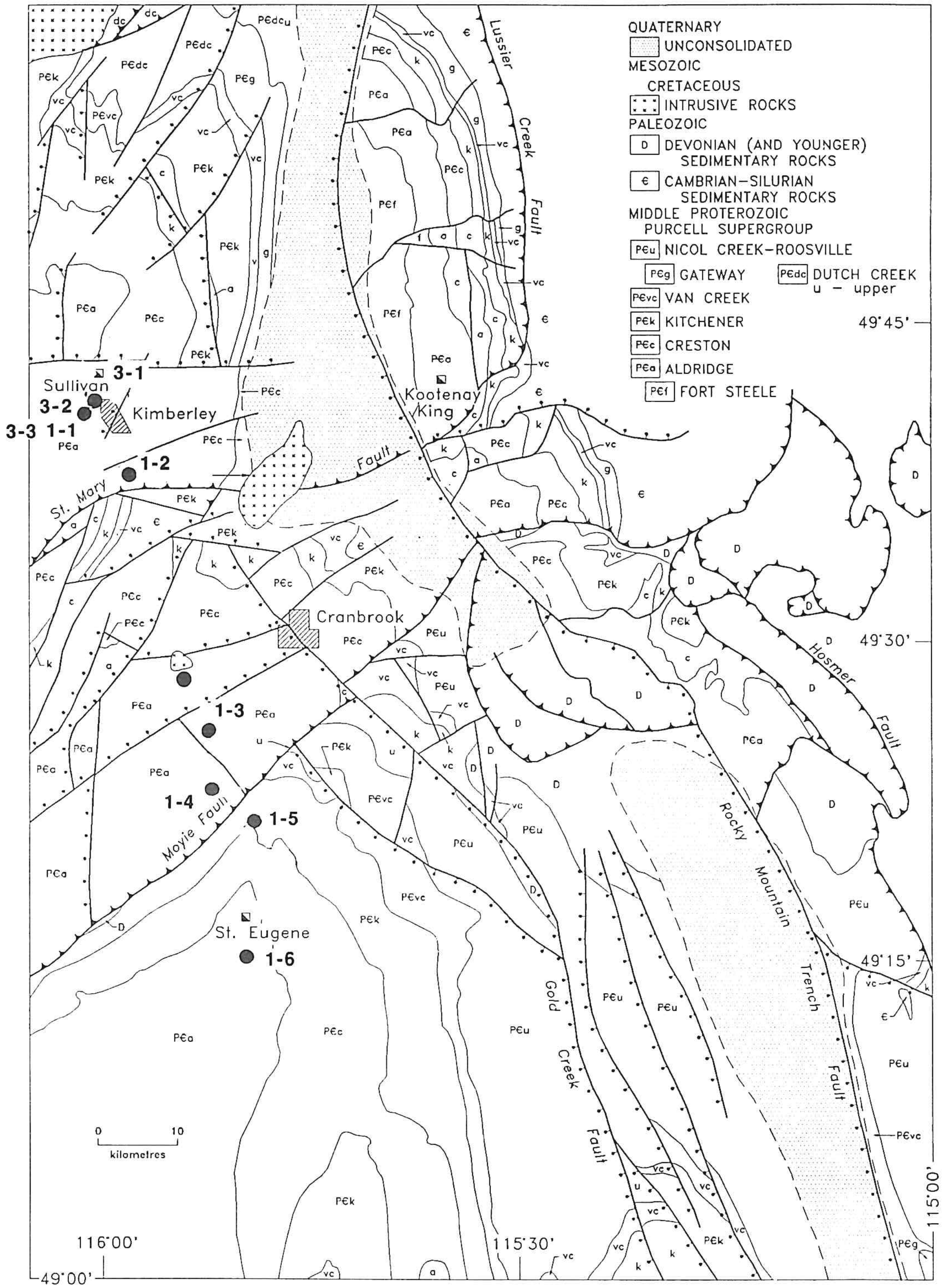

Figure 3. Geological map of the northeastern part of the field trip area (from Höy and Carter, 1988; Höy, 1993). Location of field stops indicated.

Paleogeography of Aldridge-Prichard basin

The Aldridge-Pritchard basin was an elongate trough, presently aligned from southeast to northwest (Finch and Baldwin, 1984; Cressman, 1989; Fig. 5). The basin is bound to the northeast and southwest with shallower-water facies (Price, 1964; Hoy, 1982; Finch and Baldwin, 1984; Cressman, 1989). Basinal strata correlative with Aldridge-Prichard formation in the Helena Embayment such as the Chamberlain and Greyson formations (e.g. Winston, 1986) form the southeastern end of the basin and are bound to the north and south by fault-controlled basin margins (Schmidt and Garihan, 1986; Godlewski and Zieg, 1984). To the northwest along the axis of the basin, argillaceous facies of the

CORRELATION CHART

		NELSON E1/2	FERNIE W1/2	IDAHO, MONTANA	
PALEOZOIC		HAMILL	CRANBROOK		
LATE PROT.	WINDERMERE	HORSETHIEF CREEK			
		TOBY			
MIDDLE PROTEROZOIC	PURCELL	MT. NELSON		GARNET RANGE	MISSOULA
			ROOSVILLE	McNAMARA	
		DUTCH CREEK	PHILLIPS	BONNER	
			GATEWAY	MT. SHIELDS	
			SHEPARD	SHEPARD	
		SIYEH	NICOL CREEK	PURCELL LAVA	
			VAN CREEK	SNOWSLIP	
		KITCHENER	KITCHENER	HELENA EMPIRE	WALLACE
		CRESTON	CRESTON	ST. REGIS REVETT BURKE	RAVALLI
		ALDRIDGE MOYIE SILLS	ALDRIDGE FORT STEELE	PRICHARD	

Figure 4. Table of formations, Purcell Supergroup, and correlation with Belt Supergroup.

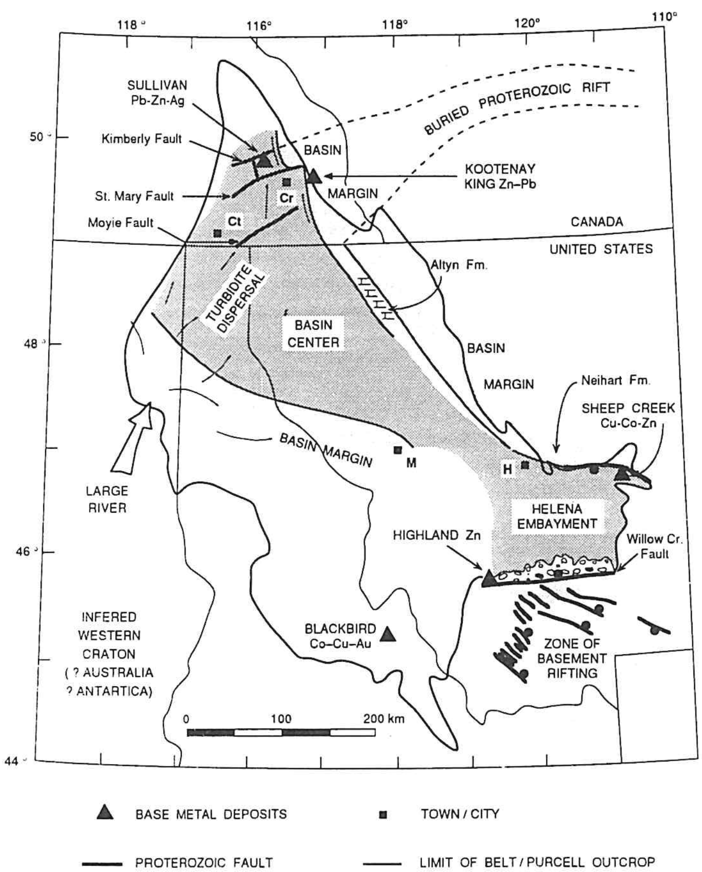

Figure 5. Interpretive paleogeographic map of Aldridge basin. No palinspastic restoration of Mesozoic and Cenozoic faults has been attempted. Based on information in Cressman (1989), Finch and Baldwin (1984), Schmidt and Garihan (1986), and Godlewski and Zieg (1984). Ct = Creston; Cr = Cranbrook; M = Missoula; H = Helena

Prichard Formation are transitional into a thick turbidite sequence interpreted as a turbidite fan derived from a large river entering the basin to the southwest (Cressman, 1989). Progradation of this turbidite fan northward is reflected by the transition from thin-bedded siltite of the lower Aldridge Formation to thick quartz wacke beds of the middle Aldridge Formation in southeastern British Columbia. This Canadian part of the basin is characterized by large northeast-trending transverse structures that offset the margin of the basin and appear related to a northeast-trending rift presently buried under Phanerozoic cover (Kanasewich, 1968; Kanasewich and others, 1969).

Stratiform base metal deposits (e.g. Highland, Sheep Creek) of Middle Proterzoic age are associated with the west-trending synsedimentary structural zones that form the margins of the Helena Embayment (Thorson, 1984; Himes and Petersen, 1990). The Sullivan deposit is associated with the northwestern zone of transverse structures.

Paleotectonic reconstructions of the Proterozoic supercontinent suggest an intracratonic position of the Belt-Purcell basin (Fig. 6). The basin may have been a lake (Winston and others, 1984; Winston, 1986b), possibly analogous to the rift setting of Lake Baikal, Russia or the East African Rift system, or an elongate, narrow seaway, perhaps analogous to parts of the Gulf of California or the Red Sea. In either case, the Belt Basin lay within a large landmass. Paleomagnetic data indicate formation at a low latitude (e.g. Collison and Runcorn, 1960). As the Proterozoic continents were unvegetated, the lake or marine basin was likely set within a hot desert landscape. The southwestern side of the basin could have been the present continental masses of Antarctica or southern Australia (Hoffman, 1991; Ross and others, 1992).

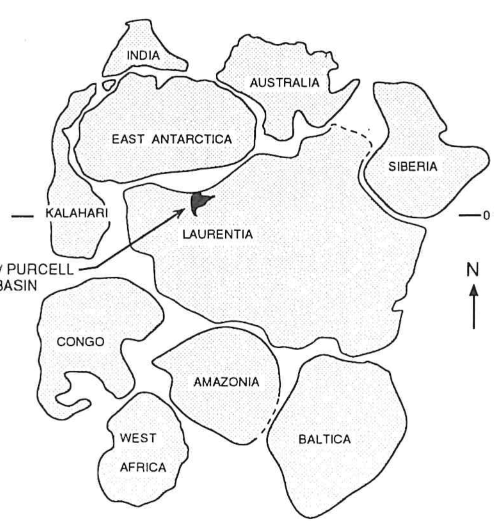

Figure 6. Paleotectonic reconstruction of Proterozoic supercontinent (modified after Hoffman, 1991). Note intracratonic position of Belt-Purcell basin.

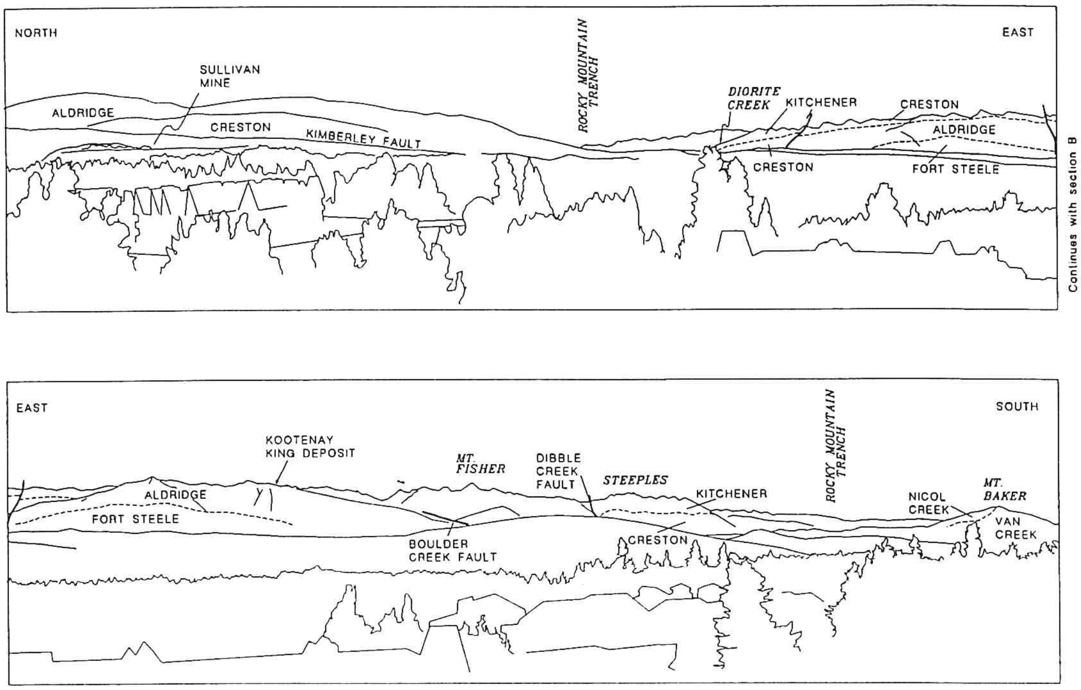

Figure 7. Panoramic view from base of Kimberley Ski area (Stop 1-1) looking from northeast to southeast. Visible are the Sullivan mine, the Rocky Mountain Trench, and the mountain front of the Hughes Range and Steeples east of the Rocky Mountain Trench. Approximate locations of the Kimberley, St.Mary, Boulder Creek and Dibble Creek faults, as well as distribution of rock units are shown. Geological map of area shown on Figure 3.

FIELD TOUR 1

KIMBERLEY TO CRESTON: REGIONAL SETTING, ALDRIDGE FORMATION

All distances are given from the traffic lights at Ross and Wallinger Streets, the main intersection in downtown Kimberley.

0.0 km Ross St. and Wallinger St., Kimberley, B.C. Drive west on Ross St. (which becomes Gerry Sorenson Way) following signs to Kimberley Ski Area. Ascend switchbacks passing Happy Hans Campground. Stay left at fork in road to Purcell, Rocky Mountain and Silver Birch Condominium/chalets. Turn right at top of hill onto road to ski lodge complex; pull off on right side.

STOP 1-1
4.0 km VIEW OF SULLIVAN MINE AND ENVIRONS, NORTH STAR SKI AREA

This stop provides a panoramic view of the major geographic features around the Sullivan mine. The view (Fig. 7) looks across the Rocky Mountain Trench to the Hughes Range in the Western Rocky Mountains (Fig. 3). In the foreground, along the eastern edge of the Purcell Mountains, is the Sullivan mine. The Sullivan mine occurs at the contact of the lower and middle Aldridge Formation. North Star Hill and the town of Kimberley is underlain by lower Aldridge Formation; middle Aldridge Formation underlies the area east of Kimberley.The distal laminated pyrrhotite fringe of the Sullivan deposit subcrops on the low hill just east of Kimberley. In the distance, Fort Steele Formation rocks are exposed on the lower slopes of the Hughes Range, overlain by the Aldridge Formation and younger Purcell Supergroup rocks.

Comment

The Rocky Mountain Trench coincides, at this latitude, with a west-side down Tertiary normal fault, the Rocky Mountain Trench fault. It follows approximately the locus of prominent facies and thickness changes in Middle Proterozoic Purcell rocks. Fluvial quartzites of the Fort Steele Formation at the base of the succession in the Northern Hughes Range (Fig. 8A) are overlain by a carbonate-bearing shallow water facies (Fig. 8B) and a condensed lower and middle Aldridge sequence (Fig. 8C). These strata are correlative with deep-water turbidite facies of the lower and middle Aldridge Formation in the Purcell Mountains to the west.

The Boulder Creek fault, which occurs in the valley just north of Mount Fisher (Fig. 7), also follows the locus of a Middle Proterozoic structure. The shallow water facies of the northern Hughes Range thicken southwards and are transitional to deeper water turbidites south of the fault (Figs. 8, 9). To the south of the fault, turbidites similar to those in the Purcell Mountains are exposed (Figs. 8B, C). Thus the Boulder Creek fault and a segment of the Rocky Mountain Trench fault outline a block of

Purcell rocks in the northern Hughes Range that marks the northern and eastern edge of the Purcell basin. In late middle Aldridge time, turbidite deposition extended over the basin margin (Figs. 8C, 9); evidence of growth faulting is not evident in late Aldridge time as upper Aldridge rocks extend across the basin margin.

Extensional tectonism continued along the eastern Purcell basin margin during deposition of younger Purcell Supergroup rocks. The Nicol Creek Formation reflects extrusion of basaltic lava along the eastern margin of the basin (Höy, 1993). In the Skookumchuck area 30 km northeast of Sullivan, dramatic changes in thickness and facies of the Shepard and Gateway formations were caused by extensional growth faults and differential subsidence. The overlying Phillips Formation, a regional mauve marker unit in both Belt and Purcell Supergroup rocks, also dies out here, supplanted to the north by green facies in the Dutch Creek Formation.

Tectonic instability along the Purcell basin margin is also evident in the Moyie Lake area southeast of Cranbrook. A prominent fluviatile conglomerate, up to 9 m thick, has locally cut through and removed up to several hundred meters of the Nicol Creek Formation (Höy, 1993). Clasts within the conglomerate include some medium-grained intrusive boulders that have been dated at *ca*. 1514 Ma. The size of these boulders and their inferred exotic provenance suggests that basement must have been exposed locally during deposition of late Purcell Supergroup rocks. This implies development of growth faulting with considerable movement as the boulder conglomerate is underlain by at least 9 km of Purcell Supergroup strata.

Return to Kimberley.

0.0 km Ross St. and Wallinger St., Kimberley, B.C. Drive south on Wallinger (Highway 95A).
5.8 km Intersection, St. Marys Lake Road; continue on 95A.

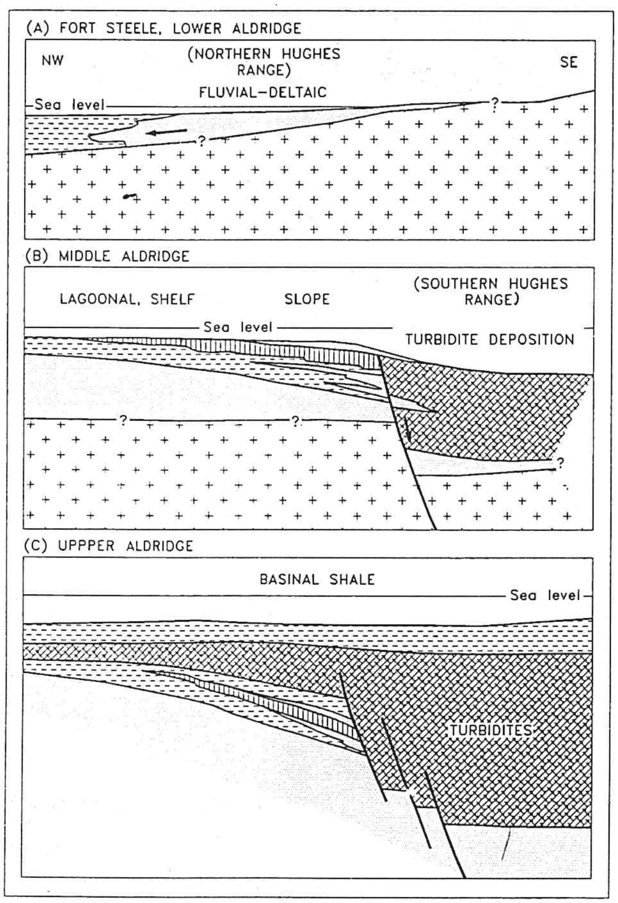

Figure 8. Tectonic evolution of the northeastern margin of the Purcell basin in Fort Steele and Aldridge times based on facies within the Hughes Range to the north and south of the Boulder Creek fault (from Höy, 1993) (Stop 1-1). The Aldridge-age fault zone is currently occupied by the Boulder Creek fault. Legend for patterns: ++ = pre-Purcell crust; -- = argillite; III = dolomitic siltstone; cross-hatch = turbidite.

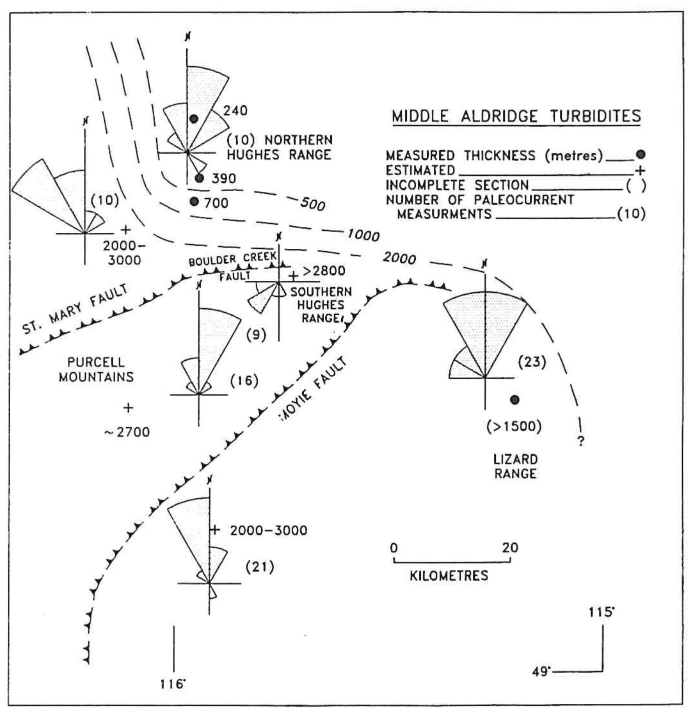

Figure 9. An isopach map of middle Aldridge turbidites, restored for movements on the St. Mary and Moyie faults (Stop 1-1). Rose diagrams are paleocurrent directions from basal scours and cross laminations (from Höy, 1982).

STOP 1-2
6.6 km LOWER ALDRIDGE FORMATION, MARK CREEK, MARYSVILLE

As you enter Marysville, park on north side of road immediately east of bridge across Mark Creek (Fig. 1). Cross road and footbridge to boardwalk trail along west side of creek.
The lower Aldridge Formation, comprising thin-bedded, well-laminated argillite and wacke is well exposed in the walls of the creek, especially downstream from the falls.

Comment

The lower Aldridge comprises dominantly rusty-weathering, thin to medium-bedded, fine-grained siltstone, quartz wacke and argillite (Fig. 10). Quartz wacke beds commonly have graded tops indicative of turbidite deposition. In the Kimberley and Cranbrook areas, a succession of grey-weathering quartz wacke and arenite 150 m thick, the Footwall Quartzite, occurs approximately 150 m below the Sullivan horizon and the base of the middle Aldridge. The contact with the middle Aldridge is placed at the base of the first prominent blocky, grey-weathering quartz wacke bed. To the southwest, east of Creston (*see* Fig. 1), the distinction between the upper part of the lower and the middle Aldridge is less pronounced and it is difficult to separate them.

The average modal composition of "unaltered" lower Aldridge siltstones based on six samples is: quartz 40%, biotite 30%, muscovite 10%, K-feldspar 10%, plagioclase 7%, sphene 3%, tourmaline 2%, pyrite 2%, organic carbon 1%, pyrrhotite 1%, and trace amounts of chlorite, carbonate, epidote, monazite, apatite, rutile, and ?zircon. These modal data compare reasonably well with data of Edmunds (1977) and Cressman (1989, Table 2), though these authors found a very high ratio of plagoclase to K-feldspar. Of significance is the feldspar-rich (~20%) nature of the Aldridge Formation, making it a sensitive protolith to the effects of hydrothermal alteration.

Based on their common occurrence as anhedral grains with a clast-like habit, the quartz, plagioclase, apatite, monazite and possibly some of the coarse muscovite appear to represent original detrital components. Albitization of K-feldspar and plagioclase, common during sediment diagenesis (Boles, 1982) and metamorphism account for dominance of albite; biotite and muscovite is interpreted as metamorphosed detrital and authigenic clays. Tourmaline, characterized by coarse (0.1 - 0.2 mm) euhedral schorl (iron tourmaline), may be of either detrital or diagenetic origin. Epidote replaces some plagioclase and may be related to either diagenesis or metamorphism of feldspars. Sphene appears detrital in unaltered sediments but can also occur as part of a hydrothermal assemblage. Apatite may be detrital, diagenetic or metamorphic after an authigenic phosphatic mineral. Pyrrhotite and lesser pyrite are common minerals, presumably related to diagenetic sulfate reduction in shallowly buried sediments.

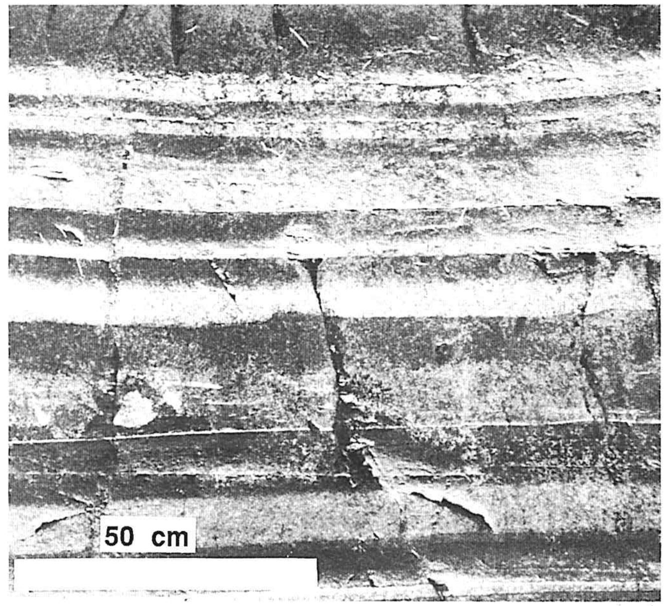

Figure 10. Thin-bedded siltstone (pale grey) and silty argillite (dark grey) exposed

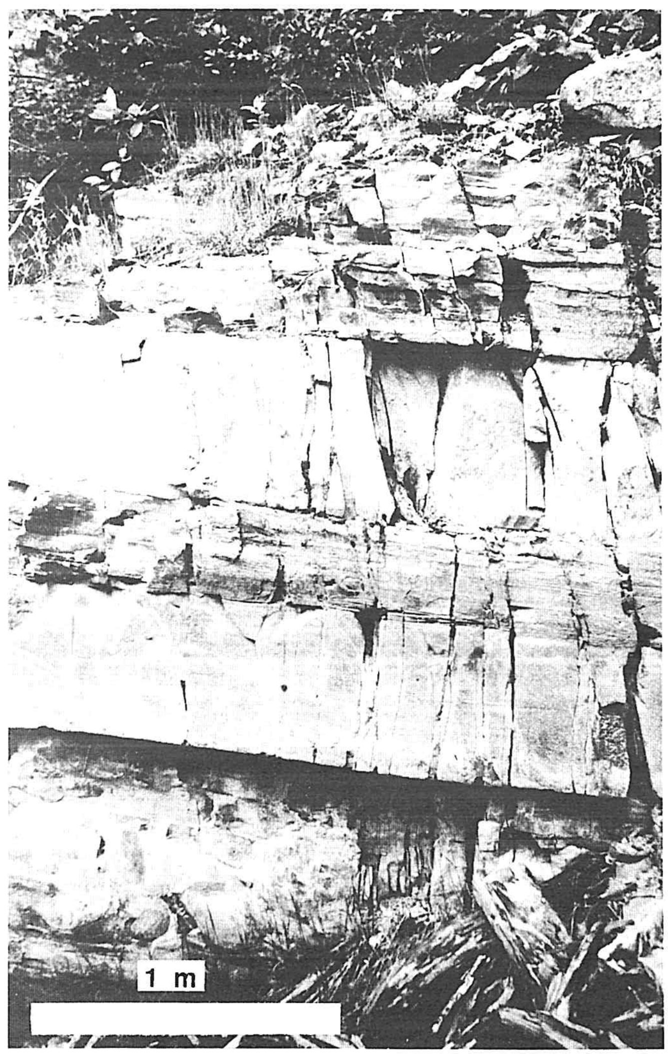

Figure 11. Medium-bedded sandy turbidite interbedded with thinbedded to laminated siltstone and argillite, middle Aldridge Formation (Stop 1-3).

6.7 km Marysville
12.9 km Hill to west (Lone Pine Hill) is underlain by irregular bodies of gabbro (Höy and Carter, 1988).
15.3 km Junction, Wycliffe Road
15.9 km View across St. Mary River. Bluffs on south side expose Pleistocene gravels and sands.
21.0 km Bridge over St. Mary River
26.2 km Railway overpass
28.3 km Junction of Highways 95A and 3/95. Take onramp onto Highway 3/95 heading south. Drive through Cranbrook following signs for Highway 3/95 (at second lights turn right). Continue south on Highway 3/95.
37.4 km Travel Information Center (east side) and Jim Smith Road (west side).
43.3 km Junction, unmarked gravel road (Fassifern Road sign on tree 50 m beyond railway crossing). Access road to St. Joe prospect (Stops 1-9, 1-10).

STOP 1-3
48.4 km: MIDDLE ALDRIDGE FORMATION AND MOYIE SILL-ALDRIDGE CONTACT, JUNCTION: LUMBERTON ROAD-HIGHWAY 95/3

Pull off on west side of Highway 95/3 at junction with Lumberton Road (Fig. 1). Low cliffs on east side of highway expose middle Aldridge strata (Fig. 11). Sole markings and internal textures include poor flutes and longitudinal ridges, and composite graded units containing cross-bedding and convolute laminae. Calcite-filled casts of crystals up to 5 mm length with a swallowtail form weather in negative relief and are found in some of the more argillaceous beds; these casts which may have originally been a sulfate mineral, appear to occur in several restricted stratigraphic intervals in the middle Aldridge. In the trees above the main outcrop are thick composite turbidite beds containing concretions.

A trail at the south end of the roadcut heads east from the road and upslope to the base of the cliffs that are visible from the highway. At the base of the cliff is a contact between gabbro and underlying Middle Aldridge strata.

Comment: Middle Aldridge Formation

The middle Aldridge comprises 2000 to 3000 m of dominantly well-bedded, medium to locally coarse-grained quartz arenite, wacke and siltstone. The basal part is dominated by interbedded quartz wacke and arenite; in the upper part beds are thinner and the proportion of siltstone and argillite increases. The upper middle Aldridge comprises a number of cycles of massive arenite beds, overlying wacke, siltstone and argillite, and uppermost siltstone and argillite.

Arenite and wacke beds of the middle Aldridge have many structures typical of classical turbidite deposits. Beds are laterally extensive and commonly parallel sided, with graded, laminated and cross laminated divisions, and less commonly dish structures. Thick beds are often amalgamated units. Within the finer grained turbidites

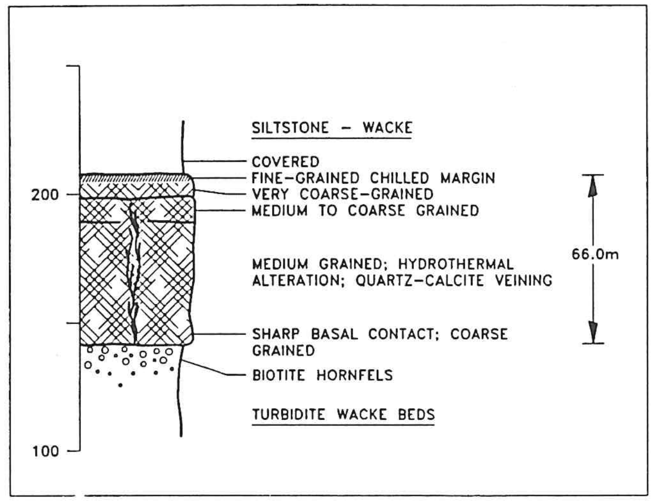

Figure 12. A generalized section through the Lumberton sill showing grain-size variation and cross-cutting veins (Stop 1-3).

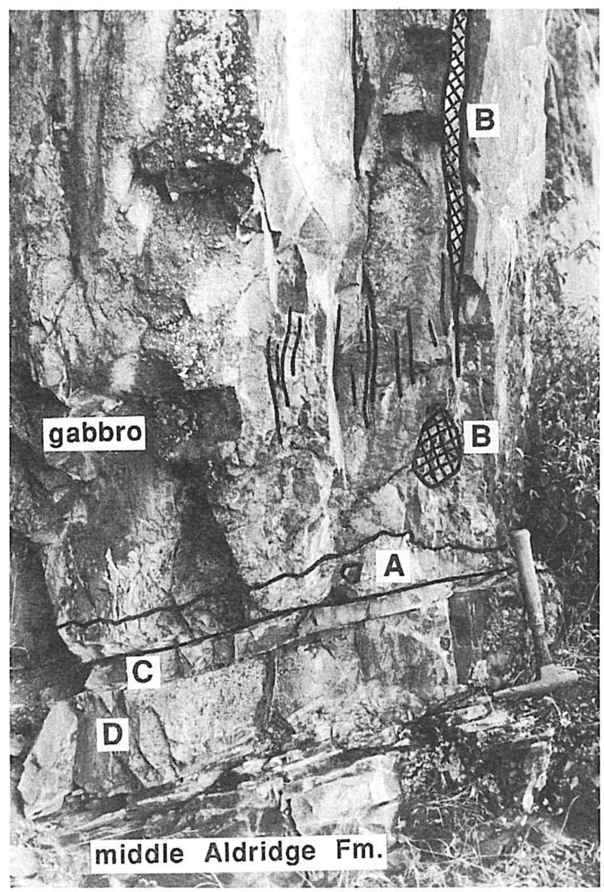

Figure 13. Basal contact of Lumberton sill with underlying Middle Aldridge sedimentary rock (Stop 1-3). A narrow breccia (A) at the contact is overlain by fine to medium gabbro cut by pods and veins of coarse calcite (B). Underlying beds (C, D) are variably altered (see text). Hammer for scale.

are convolute laminations and ripple bedding. Concretions and rip-up clasts occur in some beds. Sole markings, including scours and tool marks, indicate a generally northward current transport direction (Fig. 9).

Contact between Moyie Sill and Aldridge Formation

On the cliffs above the middle Aldridge turbidite exposure is a gabbro sill 66 m thick (Hoy, 1993, pg. 126). It is approximately 1000 m above the base of the middle Aldridge and is one of a number of similar sills within the lower Aldridge and lower part of the middle Aldridge. This stop allows a detailed look at the base of one of these sills, the Lumberton sill, and its contact with the Aldridge Formation.

A schematic section of the Lumberton sill is shown in Fig. 12. The sill is cut by vertical quartz-carbonate-sulfide veins that are visible from the scree slope. The contact between the sill and the Aldridge Formation is well exposed at the base of the cliff and is sharp with alteration restricted to a narrow zone less than one meter thick (Fig. 13). The contact phase of the Lumberton sill (A) is a breccia composed of angular fragments of fine-grained intrusive in a fine-grained matrix of plagioclase, quartz, epidote, biotite and calcite, with coarser poikiloblastic grains of hornblende, chlorite and biotite. The overlying gabbro contains abundant veins and irregular pods of coarse carbonate; the texture of the carbonate suggests formation during early cooling of the sill. An argillite bed immediately adjacent to the sill (C), is altered to fine-grained granular albite, minor dispersed epidote and ragged grains of biotite. A coarser grained, granular quartz wacke (D) contains abundant biotite, dispersed epidote and calcite with large clots of amphibole altered to chlorite. Amphibole clots increase in size and abundance toward the top of the bed.

Underlying bed D is a meter-thick zone of highly attenuated isoclinal folds (Fig. 14). These folds are limited to this narrow stratigraphic interval and are not recognized elsewhere in the section. This deformation style may be related to bedding plane shear during emplacement of the sill. If so this ductile style of deformation would suggest emplacement of the sill prior to complete lithification of the enclosing sediment.

Comment

The Lumberton sill is chemically similar to most Moyie gabbro bodies, which are dominantly subalkaline, low-potassium, high-iron tholeiites (Table 2). Locally, some Moyie sills are alkali basalt in composition (Höy, 1989). Principal early-formed minerals are amphibole (hastingsite group), plagioclase (An_{62}-An_{12}), quartz and ilmenite; amphibole, andesine, scapolite, sphene, ilmenite and biotite formed during interaction of cooling gabbro with heated fluids (i.e. autometamorphism) (Bishop, 1974). Primary pyroxene is rare. Actinolite, clinozoisite, biotite and chlorite formed during later metamorphic events.

Most sill-sediment contacts are sharp and intense alteration of sedimentary rock is limited to a narrow zone. Some Moyie sills have unusual contact relationships with

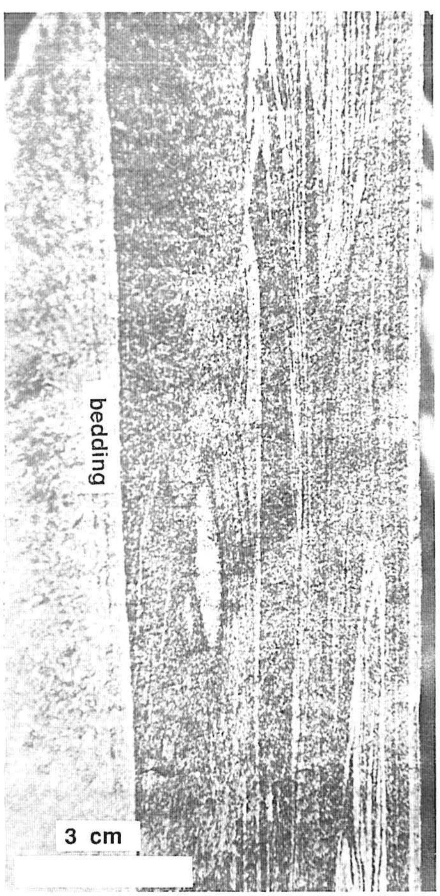

Figure 14. Highly attenuated isoclinal folds in sedimentary rock immediately below base of Lumberton sill (Stop 1-3). Photograph of sample has been rotated to vertical orientation; fold axial planes parallel bedding and contact with sill.

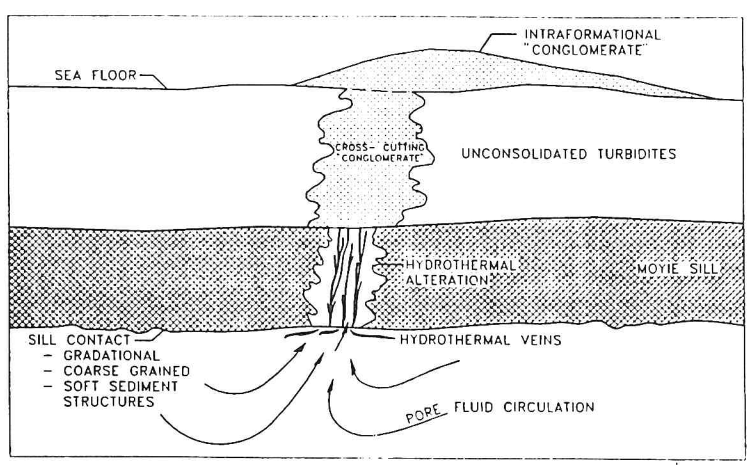

Figure 15. Model for injection of a Moyie sill into unconsolidated middle Aldridge turbidites (from Höy, 1989) (Stop 1-3).

sedimentary rocks that indicate intrusion of unconsolidated wet sediments (Hoy, 1989, 1993). These include irregular contacts with deformed sedimentary rock, anomalous tight to isoclinal folds in sedimentary rocks adjacent to sill contacts (Fig. 15), reconstitution of sediment to a granoblastic or granophyric rock ("granofels") that obliterates original textures (Stop 1-7), widespread autometamorphic alteration of primary igneous minerals in gabbro, and fragmentation of adjacent sediments (Stop 1-7). Sill emplacement may be the cause of large-scale dewatering structures (Fig. 15) (*See* Stop 1-6).

52.2 km Gabbro, road cut
54.9 km Bridge, Moyie River

STOP 1-4
54.9 km MARKER LAMINAE, MOYIE RIVER BRIDGE

Pull off on west side of Highway 95/3 immediately south of bridge. Cross to east side of highway. Climb fence (be careful not to damage fence) and walk to base of cliff exposing middle Aldridge strata. A 3.5 m thick marker unit of laminites are interbedded in the turbidite sequence (Fig. 16). The marker unit is composed of olive grey and black laminae with a generally planar bedded appearance. Olive grey silt laminae 0.5-10 mm thick are massive to faintly banded on a scale of 0.5-1 mm. Lower contacts of silt laminae are generally sharp with local load cast features. Upper contacts may be sharp or gradational. Interbedded black laminae are of similar thickness and faintly microlaminated. Disseminated pyrrhotite grains occur throughout both laminae types. Both laminae types are composed of quartz-biotite-feldspar-epidote-tourmaline-sphene-allanite-carbon silt- and clay-sized grains, and mineralogy typical of middle Aldridge strata (Heubschman, 1973:; Leitch and others, 1991). Dark laminae differ from lighter laminae only by a greater abundance of epidote, sphene, allanite and organic carbon yielding a distinctive Ca-Ti-C-REE geochemical signature.

Comment

Fourteen laminated marker units are recognized within the middle Aldridge Formation. Recognition and correlation of marker units has resulted from the work of Cominco Ltd. over the last 25 years. In particular, detailed knowledge of the characteristics of these markers has resulted from the work of Art Hagen. Each marker unit has a unique sequence of alternating light and dark laminae that have been correlated over distances of several hundred kilometers (Heubschman, 1973). Turbidite beds are variably interbedded within a laminite package, locally thickening the marker unit.

What component of the laminite *(i.e.* dark laminae or light laminae) reflects ambient sedimentation and what reflects episodic sedimentation is an unresolved issue of significant importance to the interpretation of the origin of the laminites. Reconnaissance inspection of marker laminae units throughout the middle Aldridge Formation reveal that the boundaries between pale and dark laminae are typically diffuse, but occasionally the bases of some pale laminae are sharp. These asymmetric beds with sharp bases and gradational tops suggest that rapid sedimentation of pale detrital silt interrupted the ambient deposition of more organic-rich pelagic rain.

Heubschman (1973) interpreted the dark laminae as the fallout of organic blooms in surface waters above a stagnant watermass. Recent work in the Black Sea recognizes laminites that correlate over distances in excess of 1000 km; laminites are composed of alternating organic-rich cocolithophorid and terrigenous laminae (Hay and Honjo, 1989). Spring planktonic blooms in surface waters form organic laminae that interupt ambient terrigenous sedimentation. This model suggests organic laminae are episodic; it is unclear how this model might explain the asymmetric nature of organic-poor laminae in marker units.

An alternate hypothesis for the origin of the marker laminites in the Aldridge basin involves episodic sedimentation of terrigenous material from dust storms. Such aeolian deposits could be very widespread if caused by major seasonal storms and would interrupt ambient, more organic sedimentation. Such dust particulate may have provided an important nutrient source for and caused enhanced organic productivity in the surface layer, resulting in the generally orgainic-rich nature of the markers. Laminites in the Gulf of California that have been correlated throughout a 20 kilometer study area are

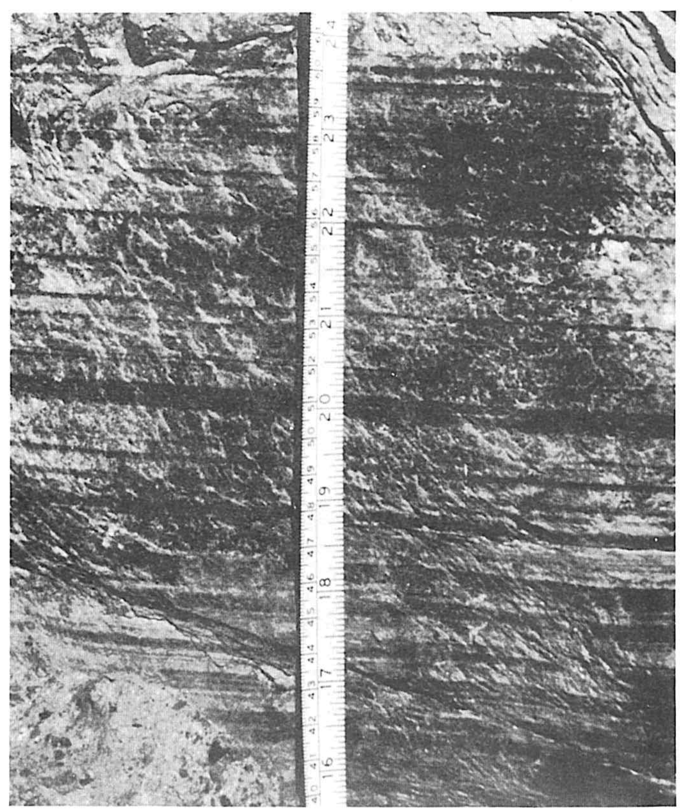

Figure 16. Marker bed laminites, middle Aldridge Formation (Stop 1-4).

composed of alternating diatomaceous and terrigeneous laminae (Baumgartner and others, 1991). Terrigenous laminae are interpreted as aeolian rather than river-derived sediments as the lamina sequence does not correspond to the discharge data for nearby rivers (Baumgartner and others, 1991). Material transport is associated largely with summer convective thunderstorms. Such aeolian processes may have been important during deposition of the Aldridge Formation given the setting as an intracontinental waterbody at equatorial latitudes surrounded by vast desert continents.

56.8 km: Moyie Fault zone. The Moyie fault zone underlies this valley bottom. (*see* Stop 1-5).
57.3 km: Roadcuts of Kitchener Formation.

STOP 1-5
60.0 km: KITCHENER FORMATION, MOYIE FAULT SYSTEM, MOYIE LAKE

Park in pullout on west side of highway at curve to east (Fig. 1). The pullout has an excellent view south and west across Moyie Lake. Roadcuts on the east side of the highway expose Kitchener Formation, part of the shallow water sequence overlying the Aldridge Formation. This stop provides an overview of the Moyie fault system, one of the northeast-trending faults with a history of movement that can be traced back to the Proterozoic.

Kitchener Formation

The Kitchener Formation is dominantly a carbonate unit between the Creston Formation and overlying siltites of the Van Creek Formation (Fig. 4). It is divisible into two members, a lower green dolomitic siltstone and an upper dark grey, carbonaceous, silty dolomite and limestone. The base of the upper member is exposed at this roadcut. It includes pale green silty dolomite and dolomitic siltstone, overlain by argillaceous or silty limestone and dolomite, and a succession of calcareous or dolomitic siltstones. Graded beds, with thin dolomite layers capped by either siltstone or dark grey argillite, are common. Carbonate layers are commonly finely or irregularly laminated, massive or locally crossbedded. Molar-tooth structures are abundant in silty dolomite layers. Siltstone layers are generally graded with argillite caps, are locally cross rippled and may have rippled surfaces. Syneresis cracks and rare mud cracks occur locally, particularly in the upper, more silty section.

Moyie Fault, Proterozoic and Paleozoic Tectonics

Across Moyie Lake is a thick succession of northeast dipping Creston Formation rocks on the west limb of the Moyie anticline (Fig. 17). The Moyie fault is exposed in the valley to the west. In the far distance on the east side of the lake, workings of the St. Eugene deposit are visible. The deposit occurs in middle and upper Aldridge rocks in the core of the Moyie anticline.

Comment

The Moyie fault and its extension east of the Rocky Mountain Trench, the Dibble Creek fault (Figs. 2 and 3), is a right-lateral reverse fault with an estimated displacement of 12 km. Most recent movement was during eastward thrusting in the late Mesozoic (Benvenuto and Price, 1979). It follows the locus of an earlier structure, recognized by prominent facies changes in lower Paleozoic rocks. A tectonic high, referred to as Montania (Deiss, 1941), occurred south of the fault in pre-Devonian time. North of the Moyie-Dibble faults, a thick sequence of Cambrian through Silurian beds is exposed; to the south, Devonian rocks unconformably overlie upper Purcell rocks. A parallel fault, the St. Mary-Boulder Creek fault system, was active during deposition of the Aldridge Formation (see Stop 1; Höy, 1982; 1984b). During the late Proterozoic Goat River Orogeny (McMechan and Price, 1982), the St. Mary fault was a locus of extensional block faulting (Lis and Price, 1976). Along the St. Mary fault near the south end of

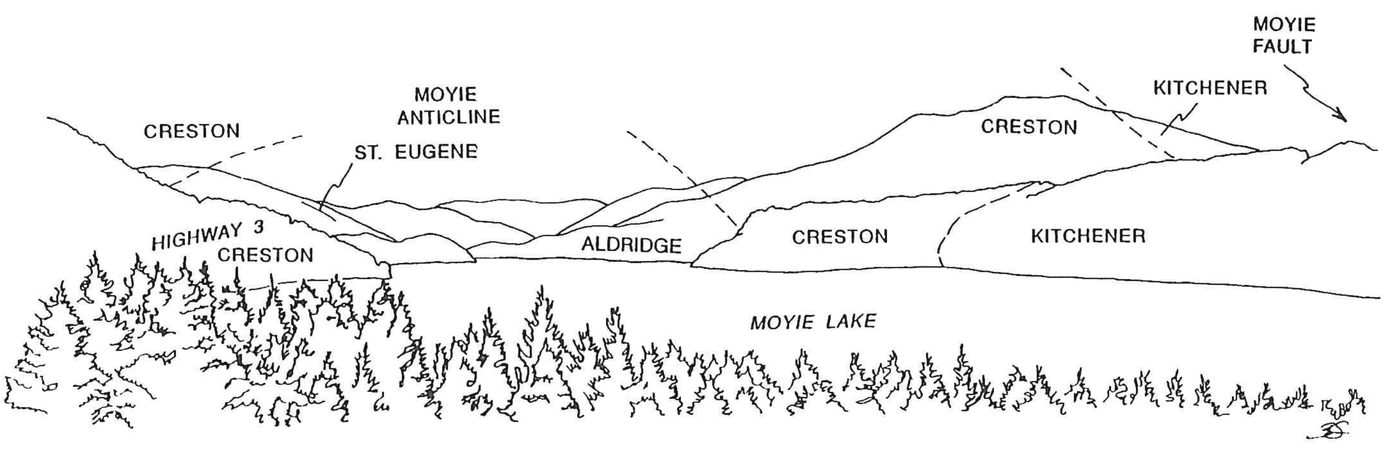

Figure 17. View south along Moyie Lake into the core of the Moyie anticline; note the St. Eugene vein deposit on the slopes above the lake (Stop 1-5).

Kootenay Lake (Fig. 2), the Windermere Supergroup includes a number of conglomerate units that contain clasts derived locally from the underlying Purcell Supergroup.

These northeast-trending structures have important implications for base metal deposits. They parallel crustal structures, recognized by prominent magnetic lineations and a gravity low (Kanasewich, 1968; Kanasewich and others, 1969), which locally controlled the configuration of the Purcell basin margin. These major northeast-trending structures (Moyie-Dibble Creek, St. Mary-Boulder Creek, Kimberley faults) may have acted as conduits for the discharge of hydrothermal fluids that formed stratiform sulfide deposits in the Aldridge Formation such as the Sullivan deposit (see Stop 3-2).

61.8 km: Creston Formation

Roadcuts on the east side of the highway expose rocks of the Creston Formation. The Creston Formation is the basal unit of a succession of shallow water and subaerial clastic, carbonate and volcanic rocks in excess of 5 km thickness that overlie the Aldridge Formation. The Creston Formation represents shoaling of the Aldridge lake/sea to a shallow-water/subaerial environment. It comprises three main subdivisions: a basal silty succession of thin-bedded grey to green siltstone and argillite, a middle quartzite division comprising coarser grained siltstone and quartz arenite, and an upper succession of intermixed green argillaceous siltstone and minor quartz arenite. This stop, near the top of the middle member of the Creston Formation, shows many of the sedimentary structures indicative of shallow-water deposition: mudcracks, syneresis cracks, mud-chip-breccias, scour-and-fill structures, ripple marks, crossbedding and graded bedding. The characteristic bed-form is thin bedded, wavy, commonly laminated argillite-siltite couplets.

67.1 km: Village of Moyie
67.5 km: Dumps, St. Eugene Mine

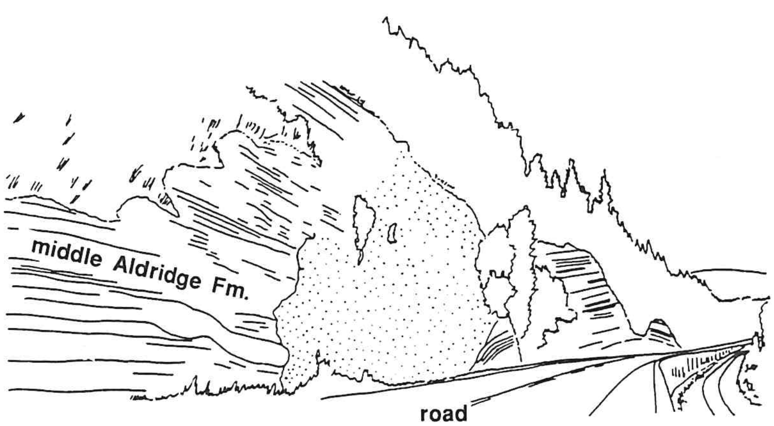

Figure 18. Sketch from photograph of clastic dike exposed in roadcut near Moyie Lake (Stop 1-6). View looking south from west side of road just north of dike. Extent of clastic dike shown in stipple pattern.

STOP 1-6
69.6 km: CLASTIC DIKE, MIDDLE ALDRIDGE FORMATION, MOYIE LAKE

Pull off on west side of Highway 3 just south of guard rail at 70.1 km (Fig. 1). Walk north 0.5 km to bold road cuts on east side of highway. A clastic dike of massive unbedded pebble wacke cuts a bedded sequence of turbidite siltstone and sandstone in the upper part of the middle Aldridge Formation (Fig. 18). The northern contact cuts bedding at a high angle, and is sharp but irregular; a thick wacke bed protrudes into the pebble wacke. This geometry suggests formation prior to lithification of the sediment. The southern contact cuts bedding at a shallow angle near road level. Within the dike the silty sand is very homogeneous, lacks bedding, and locally contains lithic fragments of siltstone and mudstone to 1 cm in length. The sediments within the dike appear to have been disaggregated, mixed and transported some distance.

An anomalously thick bed of quartzitic wacke interbedded with middle Aldridge strata is exposed in outcrop above the roadcut exposure. This thick bed may represent a sand outflow unit erupted from the pipe onto the paleo-basin floor.

Comment

A wide range of coarse clastic rocks including bedded conglomerate, discordant and concordant bodies of massive pebble wacke, discordant conglomerate and breccia dikes, and disrupted strata occur within the lower and middle Aldridge Formation. Collectively these are referred to as fragmental rocks by Cominco geologists. Because of the close association of fragmental rocks with the Sullivan deposit (Stop 3-1) and other sulfide occurrences in the Aldridge Formation (e.g. Stop 3-2), some of the processes that formed fragmental rocks may be linked to ore-forming processes.

The dike and thick sand bed at this stop are interpreted as the upper portion of a sand volcano that erupted on the floor of the Aldridge basin. The sand volcano likely represents liquifaction of shallowly buried sands and buoyant rise of the slurry along a fracture zone to the basin floor. The energy source that induced liquifaction may have been a seismic event, or emplacement of a shallow level intrusion that resulted in fluid boiling and fluid escape (Fig. 15).

100.3 km: Bridge over Moyie River, north end of village of Yahk.
104.7 km: Junction with Highway 3. Turn right (east) on Highway 3.
111.0 km: Goatfell tourmalinite occurence. Prominent knob that rises above railway track several hundred meters south of highway is a tourmalinite body in the middle Aldridge Formation.
123.8 km: Village of Kitchener (McConnel on some maps).

STOP 1-7

125.3 km: MOYIE SILL, ALTERED AND FRAGMENTED ALDRIDGE FORMATION, VILLAGE OF KITCHENER

Pull off on north side of Highway 3, opposite 5-10 m high roadcuts on south side of the highway that extend from mileage 125.2 to 125.7 km (Fig. 1). Two gabbro sills with an intervening section of altered and brecciated middle Aldridge quartzite is exposed near the northeast end of the roadcuts. The sills and associated breccia has been traced intermittently at least 3 km. It contrasts with the basal sill contact at Stop 1-3 and the soft-sediment features described by Höy (1989) in a lower Aldridge sill at Lamb Creek west of Moyie Lake.

The contact of the Aldridge metasediments with the upper sill is sharp. The metasediment-lower sill contact is not exposed; the lower sill crops out approximately 130 m to the southwest along Highway 3.

The exposed metasediment is altered to a granoblastic medium-grained quartz-biotite-feldspar rock refered to here as a granofels. Towards the contact with gabbro, the granofels includes rounded boulders in the granofels matrix. The boulders are similar in composition to the matrix and have somewhat diffuse boundaries. The absence of primary sedimentary structures suggests intense recrystallization or physical mixing of the original sediment. The granofels is composed of recrystallized quartz and minor saussuritized feldspar grains overgrown with abundant biotite and minor plagioclase (albite?) and minor disseminated epidote, chlorite and calcite. Intense alteration may be due to the physical confinement of heated fluids trapped during intrusion of the two adjacent sills.

Chemical analyses of the granofels metasediments and sill samples are shown in Table 2. The composition of the metasediment is fairly typical of Aldridge turbidites (compare Edmunds, 1977) indicating a largely isochemical alteration process. The K_2O values are slightly elevated reflecting increased abundance of biotite (Aldridge turbidites average 1.5 percent K_2O; Edmunds, 1977).

The composition of the sill, listed in Table 2, is comparable to most Moyie sills. It has a subalkaline, tholeiitic basalt composition. High potassium content in sample 92K-13, located approximately 160 m west on Highway 3 reflects the presence of biotite alteration in that gabbro.

129.7 km: Bridge, Goat River
133.0 km: Junction with Arrow Creek East Road. Turn north onto Arrow Creek East Road.
133.4 km: Junction with North Goat River Road. Take right fork (main paved route) on North Goat River Road.

STOP 1-8

133.7 km: ALBITITE, NORTH GOAT RIVER ROAD

Park on south side of road across from prominent road cuts. The roadcut exposes a discordant white to pale grey massive albite ("albitite") alteration zone within a sandstone turbidite sequence within the middle Aldridge Formation (Fig. 19). The albite alteration is most intense along a series of northeasterly-trending steeply-dipping structures in the southern part of the road cut. Laterally, the massive discordant albitite interfingers to the west and east with less altered dark green to grey chlorite-muscovite-albite altered rock. These concordant albitite zones preferentially alter thicker sandstone beds while less intense chlorite-sericite-albite alteration occurs in more thin-bedded sandstone, siltstone and argillite sequences. This relationship suggests fluid flow was focused along the steep structure and spread laterally along permeable sandstone beds. Towards the north end of the road cut, the bedded turbidite sequence is biotitic and is only weakly altered. Note that these least altered sediments are distinctly iron stained (typical of middle Aldridge strata) relative to the albite suggesting that sulfide predates the albitite alteration and that sulfide does not appear to have been introduced during alteration.

The albitite is interpreted as a fault-

Sample Number	Sample Name	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	SUM
(Aa): Moyie Sills - Kitchener Area												
45120	92K-2	54.47	0.63	13.99	11.10	0.18	5.92	7.80	1.03	1.81	0.05	99.65
45122	92K-3	52.28	0.68	13.95	10.82	0.20	7.05	10.27	2.07	0.78	0.05	99.94
45125	92K-13	50.96	0.47	14.72	9.37	0.15	8.79	11.15	1.26	6.38	0.04	99.46
(Ab): Metasediments - Kitchener Area												
45123	92K-5	74.96	0.49	12.40	3.90	0.04	0.82	0.73	1.68	3.32	0.04	99.76
45124	92K-10	67.02	0.79	13.89	7.62	0.11	1.77	2.80	2.38	2.42	0.08	99.77
45126	92K-16	67.44	0.88	11.86	10.15	0.14	1.37	2.46	1.76	2.50	0.11	99.60
(Ac): Moyie Sills - Lumberton and Sullivan Areas												
45130	LUMBD	49.85	0.87	13.14	11.31	0.17	6.13	10.43	1.28	1.74	0.08	99.65
42964	SUL-SILL	53.07	1.30	13.29	13.47	0.29	6.03	7.07	1.80	0.83	0.10	99.71

Lab Number	Sample Number	Sr	Rb	Zr	Y	Nb	Ta	Ce	Cs	La	Sc	V
(Aa): Moyie Sills - Kitchener Area												
45120	92K-2	—	—	—	—	—	—	44	6	17	38	236
45122	92K-3	—	—	—	—	—	—	2	—	—	—	—
45125	92K-13	—	—	—	—	—	—	27	8	—	36	206
(Ab): Metasediments - Kitchener Area												
45123	92K-5	—	—	—	—	—	—	84	--	30	6	48
45124	92K-10	—	—	—	—	—	—	74	7	28	18	147
45126	92K-16	—	—	—	—	—	—	84	10	40	19	129
(Ac): Moyie Sills - Lumberton and Sullivan Areas												
45130	LUMBD	—	—	—	—	—	—	28	13	—	42	289
42964	SUL-SILL	159	35	94	27	6	15	28	5	15	44	364

Table 2. Analyses of Moyie gabbro samples (Aa) and contact metasediments (Ab), stop 1-7 near Kitchener; also included are samples from the Lumberton sill and a Sullivan Mine sill (Ac).

controlled alteration zone. The closest gabbro sill exposed on surface within underlying strata is several kilometers to the northeast in the Iron Range. However, a gabbro sill occurs upslope, approximately 125 m higher in the section. This area lies within an anomalous belt of lower and middle Aldridge strata that extends to the northeast (Iron Range) and southwest (Creston "ramparts") that appear to have been an axis of sill emplacement and hydrothermal alteration. Within this belt is an unusual abundance of gabbro sills and dikes, as well as albite-chlorite altered rock and abundant fragmental rocks. Some large albitite bodies occur along the margins of gabbro sills.

Comment

The origin of albite-chlorite-pyrite alteration appears to be related to fluids circulating during emplacement of the gabbro intrusions. Most occurrences of albitite (+/- chlorite, pyrite) within the Aldridge Formation are spatially associated with Moyie gabbro intrusives. The large volume of albite-chlorite-pyrite alteration within and overlying the vent complex of the Sullivan deposit (Stop 3-1) has been interpreted by some workers to suggest that the albite alteration is related to late stage ore fluids (*e.g.* Hamilton and others, 1983). Turner and Leitch (1992) recognize zoned envelopes of albite-chlorite-pyrite alteration around some gabbro dikes underlying the Sullivan deposit and albitic alteration of some dikes and sills suggesting that alteration post-dated the local intrusion of gabbro. Turner and Leitch (1992) suggest that alteration accompanied intrusion of the gabbro sill complex and that fluids were guided at least in part along sill and dike contacts, as well as along faults. As a result, albitic alteration occurs along some gabbro contacts as well as elsewhere along faults.

Retrace route to junction of Fassifern Road and Highways 3/95 south of Cranbrook.

134.4 km Junction of Arrow Creek East Road and Highway 3. Turn east (left) on highway.
163.7 km Junction of Highway 3 and Highway 95. Turn north (left).

220.0 km Junction, Lumberton Road with Highway 95.

228.1 km Junction, unmarked gravel road (marked Fassifern Road on tree 50 m beyond railway crossing). Turn west onto Fassifern Road. Cross railway tracks. Road turns south, then north and enters a large clearcut area. Look for dirt track that joins road obliquely from the southwest.
231.1 km Junction of dirt track from southwest with gravel road. Four-wheel drive advised from here on. Turn west onto dirt track. Track ascends slope, at times across shelf rock.
233.0 km Intersection with north-south track. Continue ahead on track (westwards).
233.6 km Branch track to left (just past outhouse on left). Take left branch and park. Continue walking down track 100 m to cleared area outcrop.

STOP 1-9
233.7 km: VOLCANOCLASTIC CHANNEL COMPLEX, ST. JOE PROPERTY

Exposed in outcrop in the cleared area is an erosional channel sequence that cuts turbidite strata of the upper part of the middle Aldridge Formation (Fig. 20) (Pighin, 1983). Channel deposits are not commonly observed in the Aldridge Formation. Even more exceptional is the presence of abundant volcanic clasts within this channel deposit.

The channel sequence has been stripped of overburden and cross-cutting veins have been explored by a short adit. It comprises a number of fining upward units, each with a coarse base that scours the underlying unit, and each generally capped by a finer grained laminated or massive unit (Figs. 21, 22). The channel sequence pinches out to the south; its maximum thickness on the north side exceeds 16 m. Though lithic fragments from the Aldridge Formation are dominant, a significant component of the clasts are of volcanic origin.

Measured sections through the channel sequence (Figs. 21, 22) clearly show grading of individual units and truncation of beds by

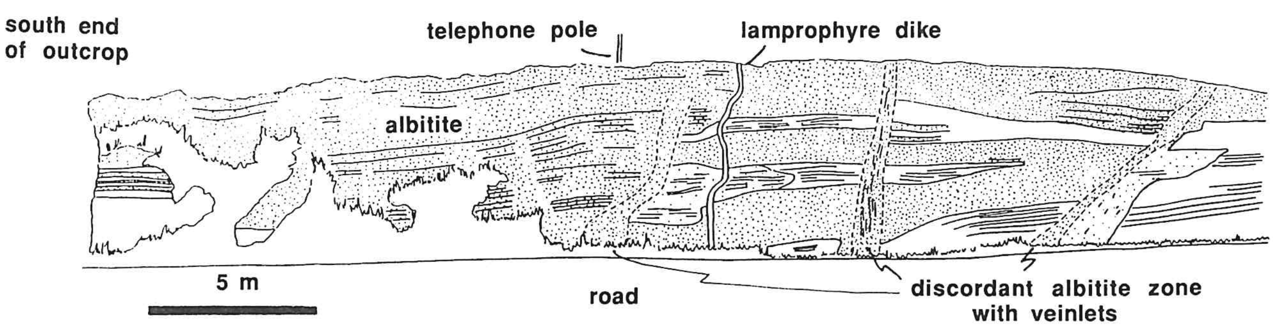

Figure 19. Line sketch of albitite zone, middle Aldridge Formation in roadcut at Stop 1-8. View looks west at southern part of roadcut. Dense stippled areas are massive albitite, open stipple denotes patchy albite with chlorite-muscovite alteration. The boundaries of discordant zones of massive albitite with abundant veinlets are shown in dashed lines. A planar lamprophyre dike appears curviplanar due to irregular surface of roadcut.

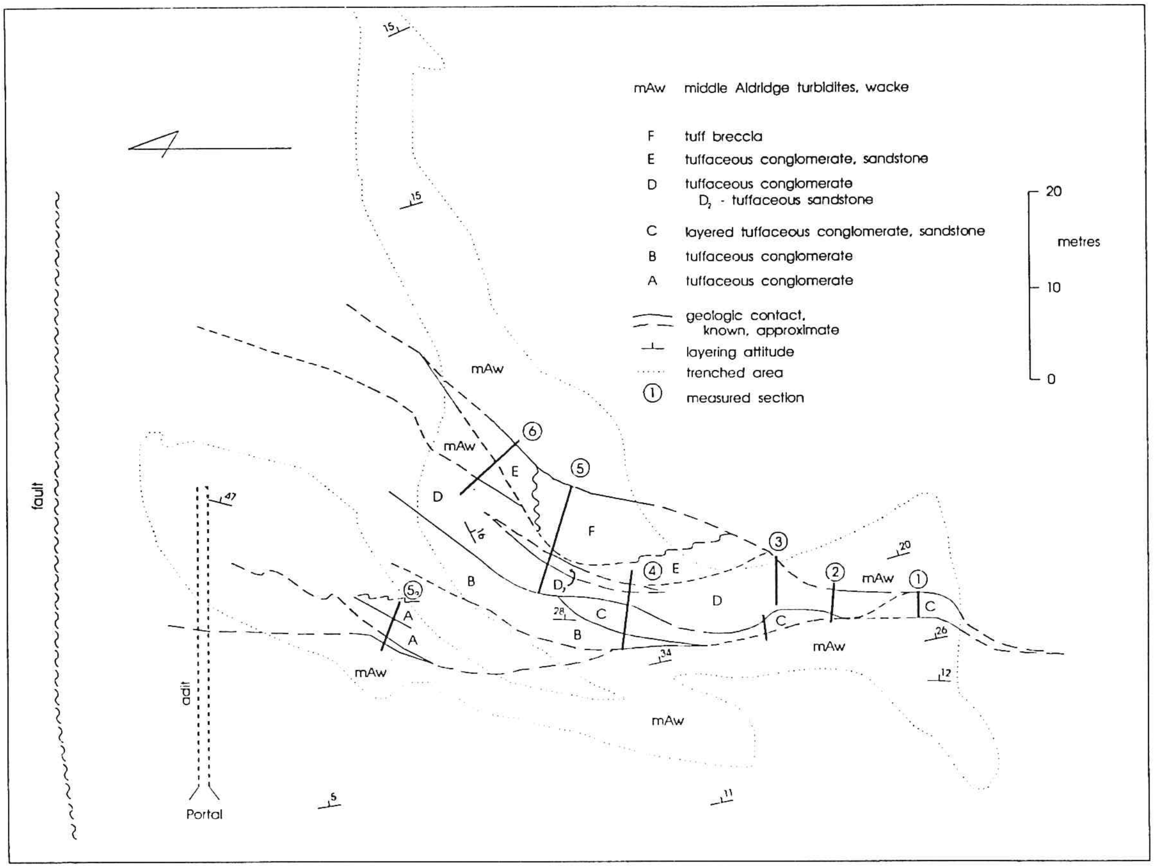

Figure 20. Geological map of the upper fragmental, St. Joe prospect,modified from Pighin, 1983) (Stop 1-9).

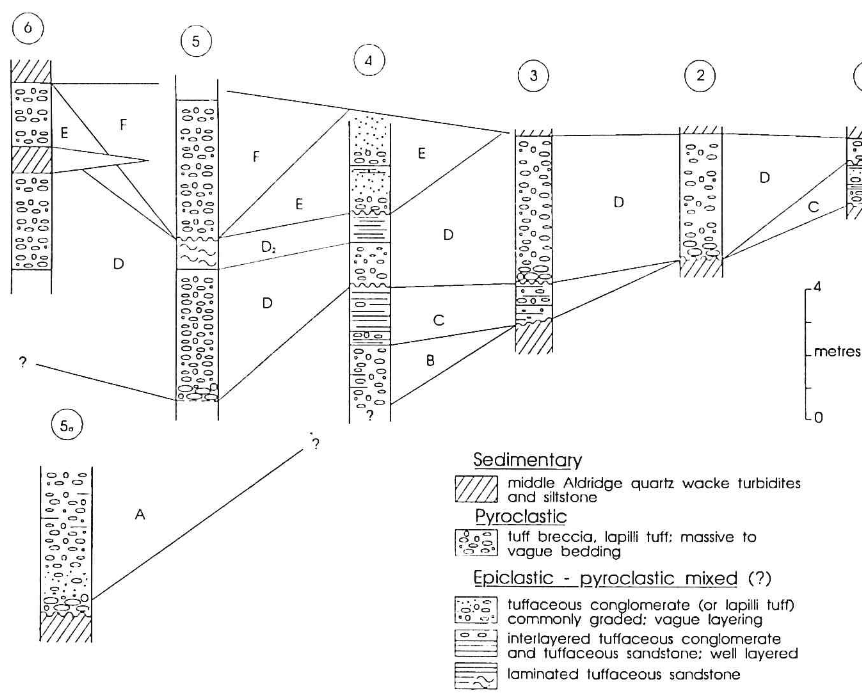

Figure 21. Measured sections through the upper fragmental, St. Joe prospect (Stop 1-9).

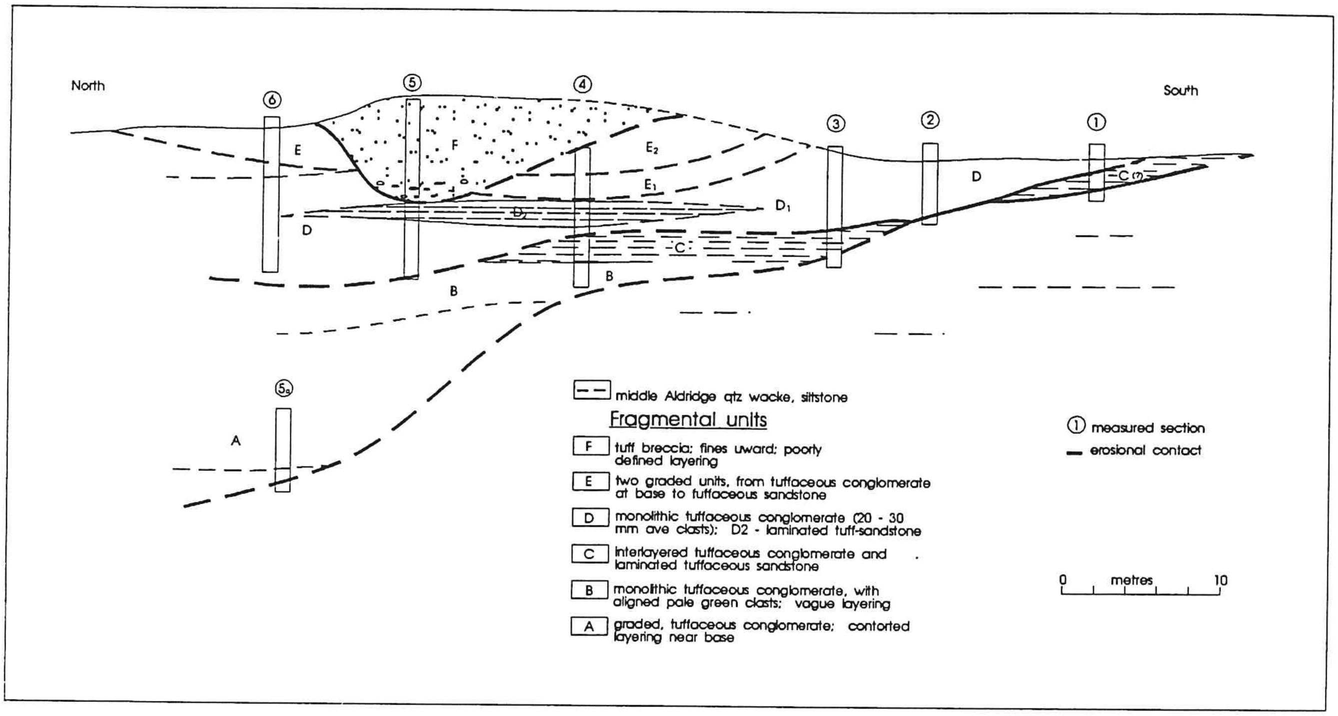

Figure 22. An interpretive section through the upper fragmental, St. Joe prospect (Stop 1-9).

Figure 23(a). Volcaniclastic channel complex, St. Joe prospect (Stop 1-9). Section 4, unit C; well layered tuffaceous conglomerate and laminated tuffaceous sandstone; note general fining-upward nature of unit, occasional large isolated clasts, and local low angle crossbeds.

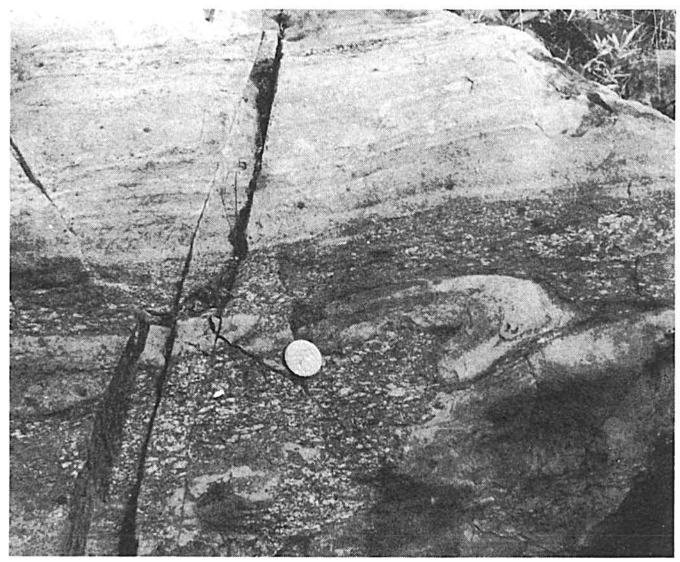

Figure 23(b). Volcaniclastic channel complex, St. Joe prospect (Stop 1-9). Section 4, unit E,?; contorted tuffaceous sandstone layer near the base of a graded tuffaceous conglomerate - laminated tuffaceous sandstone sequence. Quarter for scale.

overlying units (Fig. 23a). The coarsest unit (Unit F) is a reworked tuff breccia to lapilli tuff and occurs near the top and center of the channel deposit. Subrounded blocks up to 10 cm in long dimension occur within a finer-grained granular matrix. Some clasts have well-defined chloritic rims, possibly armored lapilli or pillow fragments; other clasts are comprised dominantly of chlorite, possibly altered mafic glass shards. Clasts that contain feldspar and ?quartz phenocrysts or a fine granular mixture of dominantly feldspar and chlorite are noted. The groundmass is primarily broken feldspar and quartz crystals with abundant biotite and chlorite.

Other clastic units are finer grained but also graded. They scour and locally deform underlying units (Figs. 23b). They are mineralogically similar to the groundmass of Unit F, with abundant broken feldspar and quartz crystals and numerous small angular, aligned chlorite-rich clasts.

A projected section (Fig. 22) illustrates the cross-cutting nature of the channel complex. Successive mass flow events within the channel system deposited the reworked lapilli and crystal tuff, tuffaceous conglomerate and sandstone. The source of this volcanic debris is unknown, presumably it is a small eruptive center within the Aldridge strata. Similar coarse clastic deposits that may have a volcanic component (i.e. Wildhorse River; northwest of the Sullivan mine along Mark Creek) and as well as a possible layered tuff and flow (i.e. Estella mine area, northern Hughes Range) have been recognized elsewhere in the upper part of the middle Aldridge Formation at a similar or slightly higher stratigraphic level. These possible volcanic horizons may be comagmatic with the high-level sills of the Moyie intrusions and record phreatic explosions in a shallower water environment. Alternatively, the presence of significant quartz, some possibly as phenocrysts, within the St. Joe volcanic fragments could suggest a more fractionated felsic phase of the Moyie sills.

STOP 1-10

234.4 km: CLASTIC DIKE AND TOURMALINITE ALTERATION, ST. JOE PROPERTY

From the parking site for Stop 1-9 return to main dirt track. Turn west and continue down hill approximately 0.7 km to junction with track on left (south). Turn onto side track and park. The track is covered with several fallen trees. Follow track across the ravine, continue for about 300 m to a large cleared area of low outcrop.

Exposed within the outcrop is a clastic dike that cuts northwest-striking, shallowly east-dipping strata of the middle Aldridge Formation (Fig. 24). Associated with the clastic dike are small condordant bodies of tourmalinite and a small lens of massive sulfide. This outcrop is 700 m southwest of the volcanic channel complex at Stop 1-9 and about 250 m northeast of a gabbro sill exposed in the valley of Kiakho Creek. These strata are about 300 m below the volcanic and 150 m

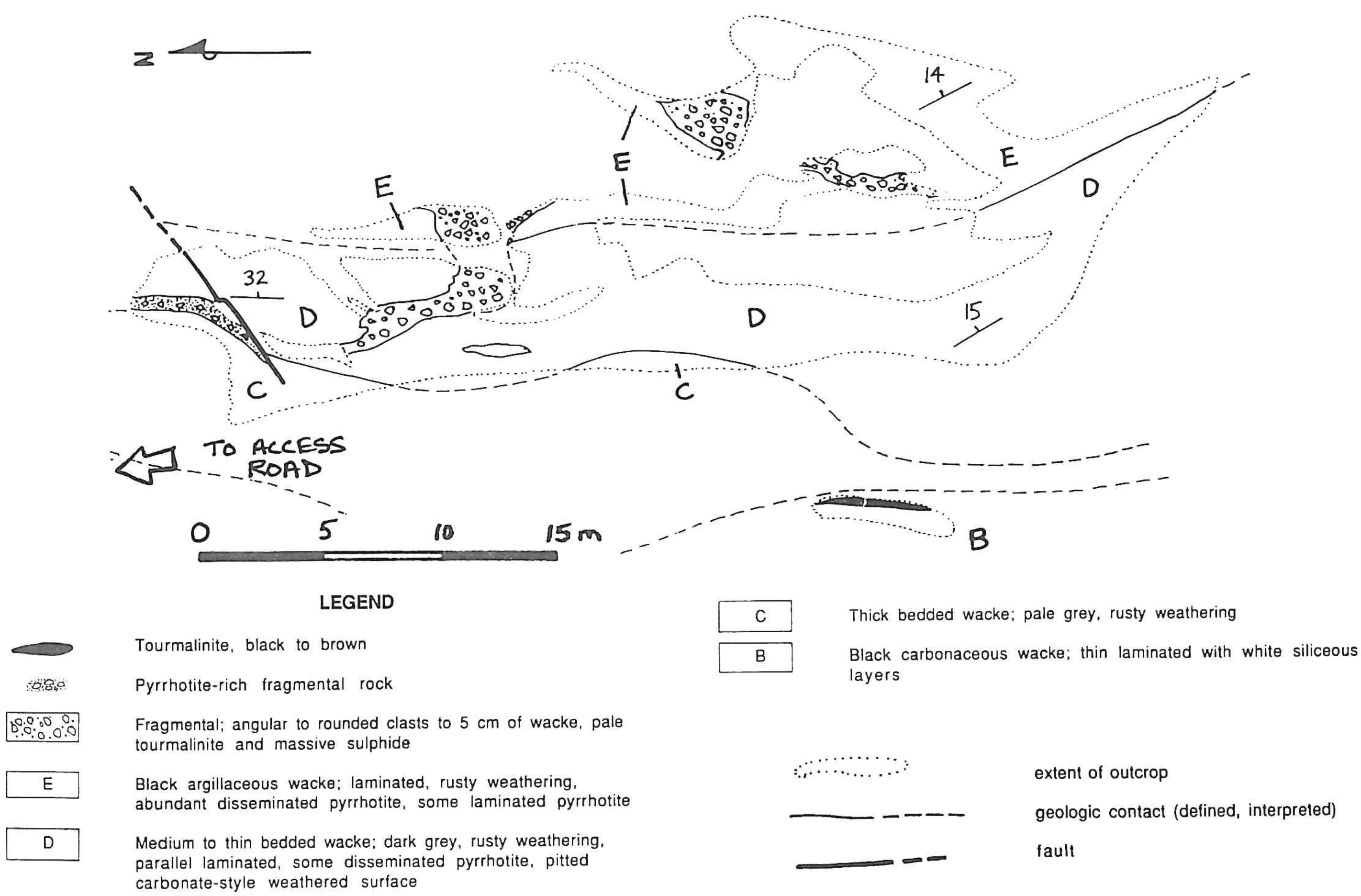

Figure 24. Geological map of the area of clastic dike, St. Joe prospect (modified from Pighin, 1983; Stop 1-10). Map shows large. low outcrop within cleared cleared area. Access road enters cleared area at northwest corner.

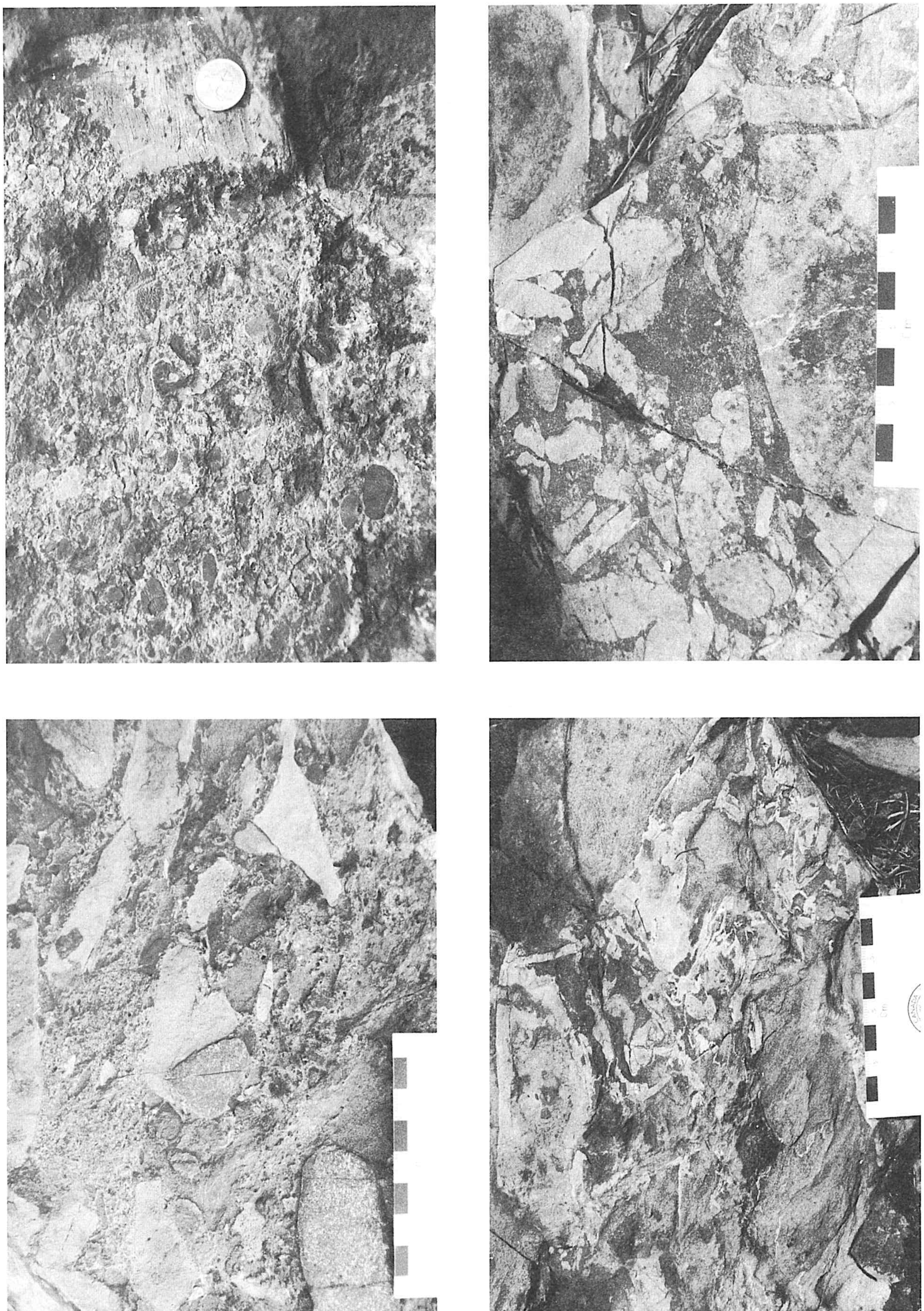

Figure 25. Fragmental textures in and adjacent to clastic dike, St. Joe prospect (Stop 1-10).(a) Pebble conglomerate with abundant angular lithic clasts. (b) Discordant contact between pebble conglomerate and bedded sediment. Dark pebble (T) is black tourmalinite. (c) Irregular breccia in strata adjacent to clastic dike. Wacke clasts have been deformed in a plastic state and lie within a pale argillaceous matrix. (d) Angular breccia adjacent to clastic dike.

above the gabbro (Pighin, 1983).

The clastic dike cuts a dark grey wacke (Unit D, Fig. 24) and overlying black argillaceous wacke with abundant disseminated pyrrhotite (Unit E, Fig. 24) that are exposed in low outcrop within the cleared area. These units are underlain by a highly carbonaceous wacke exposed in cliff exposures and road cuts in the forest immediately west and downslope of the cleared area (Unit B, Fig. 24). This unit is unusually carbonaceous for the Middle Aldridge Formation, and contains small subconcordant lenses of black and pale tourmalinite. Tourmalinite pods are best exposed along road cuts downslope from the main outcrop area. Association of tourmalinite with carbonaceous strata is noted elsewhere in the Aldridge Formation.

The clastic dike is a steeply dipping, northwest-trending body 1 to 2 m wide with several irregular subconcordant lenses that extend along bedding to the north and south. The dike is composed of a variable mix of angular to rounded clasts typically 0.5 to 6 cm in length within a matrix of sand (Figs. 25 a,b). Of particular importance is the well sorted and rounded nature of fragments (Fig. 25b) that suggest winnowing and milling of fragments during upward transport. Clasts appear to be derived from middle Aldridge strata. The presence of rounded clasts of black tourmalinite (Fig. 25b), likely derived from underlying tourmalinite bodies, support upward transport of clasts. Larger angular clasts appear to be derived from immediate wall rocks. Adjacent to the fragmental body, strata are locally veined, brecciated or plastically deformed (Fig. 25c,d) reflecting *in situ* brecciation. Just to the north of the fragmental is a massive pyrrhotite lens 4 m x 0.3 m in dimension that lies concordant to bedding. The pyrrhotite lens contains angular lithic fragments and its origin, whether exhalative or subsurface replacement, is unclear.

Retrace route to Highway 3/95. At highway, turn north and return to Kimberley via Cranbrook.

FIELD TOUR 2

ST. MARY RIVER VALLEY: LOWER ALDRIDGE GABBRO INTRUSIVE COMPLEX, AND PURCELL AGE MAGMATISM AND TECTONISM

Tour 2 visits exposures within the St. Mary valley southwest of Kimberley (Figs. 1 and 26). Field stops focus on lower Aldridge Formation, the gabbro sill complex underlying the stratigraphic level of the Sullivan deposit, altered sedimentary rocks associated with sill contacts, Proterozoic and younger regional metamorphism and structures, and the Hellroaring Creek stock of Middle Proterozoic age.

0.0 km Traffic lights at intersection of Ross St. and Wallinger St., Kimberley, B.C. Drive south on Wallinger St. (Highway 95A), which becomes Warren St. Continue out of town.
5.8 km Turn right (west) on St. Mary Lake Road. Road climbs thick glacial deposits. Pit at top of hill exposes east-dipping foreset beds in Pleistocene gravels.

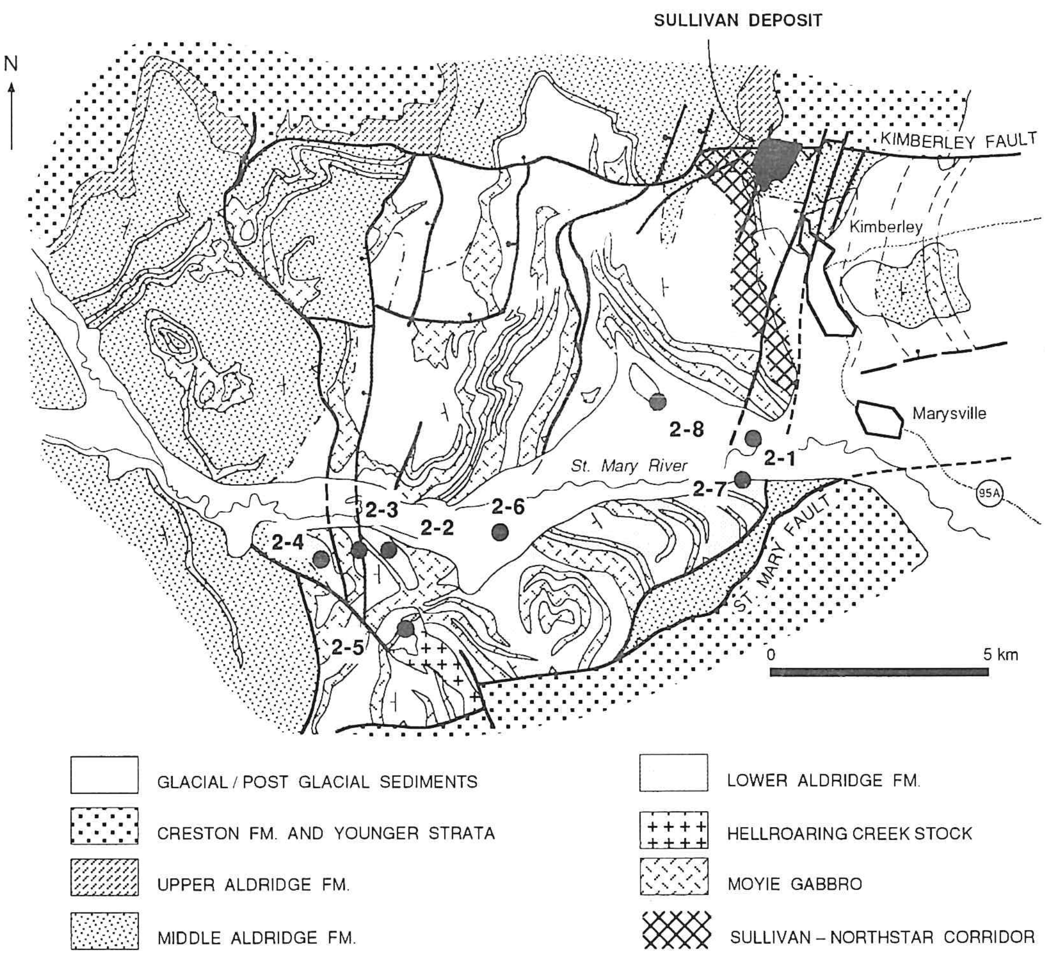

Figure 26. Geologic map of the Kimberley - St Mary valley area showing location of field trip stops. Geology based on regional mapping by Cominco Ltd and Leech (1957).

STOP 2-1
10.3 km ALTERED GABBRO-SEDIMENT CONTACT

Low outcrops lie on the north (right) side of the road opposite a farm to south on the floodplain of St. Mary River (Fig. 26). Outcrops expose lower Aldridge phyllitic metasedimentary rocks, a thin gabbro sill, and altered metasediments (granofels, spotted biotite) adjacent to the gabbro contact.

At the westernmost outcrop, lower Aldridge strata are thin-bedded phyllites and fine-grained quartz-muscovite schist. These phyllitic to schistose rocks are transitional in grain-size and fabric development between lower Aldridge strata exposed in Mark Creek (Stop 1-2) and coarser lower Aldridge schists exposed in Matthew Creek (Stop 2-8). The east-dipping foliation, defined by oriented muscovite grains to 1 mm in diameter, is shallowly oblique to bedding.

A thin zone of pale albitic? alteration occurs along the lower contact with medium-grained gabbro. The contact of the gabbro with overlying metasedimentary rock is obscured by coarse biotite-quartz-feldspar alteration of sediment ("granofels") and chlorite-amphibole-biotite alteration of gabbro. The granofels is similar to sill contact alteration at Stop 1-7, and is transitional away from the sill contact into "spotted biotite" metasediments. This granofels, commonly referred to as "granophyre", is transitional into bedded metasedimentary rock and appears to have formed by alteration of sediments.

12.4 km Junction, St. Mary-Matthew Creek Forest Service Road (access road to Stop 2-8).

Figure 27. Quartz-albite-chlorite vein with albitized gabbro envelope in gabbro, 21.2 km.

21.2 km Gabbro outcrops along north side of road at the base of Bootleg Mountain. Gabbro is cut by a northwest-trending quartz-plagioclase-chlorite vein 5 cm across with white-colored plagioclase (albite?)-chlorite alteration envelope (Fig. 27). This gabbro is part of the lowest thick sill exposed on the slopes of Bootleg Mountain (Stop 2-6).

21.8 km Junction, paved road to left. Turn left.
22.1 km Bridge, St. Mary River
22.4 km Junction. Turn north (right) onto St. Mary-Hellroaring Creek Forestry Road.
23.3 km Junction. Turn west (right) onto St. Mary-Meacham Forest Service Road.

Comment

As you drive west along St. Mary-Meacham Forest Service Road you are climbing upsection through a gabbro sill complex intruded into the lower Aldridge Formation. This section is equivalent to the section exposed to the north across the St. Mary valley on Bootleg Mountain (see Stop 2-6).

24.3 km Low outcrop, gabbro

25.7 km Roadcut cliffs on south side expose medium to coarse-grained gabbro. Some very coarse plumose-textured gabbro crops out on upper cliff and in talus to west. The gabbro is cut by some plagioclase-amphibole veins with chloritic envelopes and albite? veins with albitic envelopes. This gabbro is correlated with the gabbro sill exposed in a quarry along the gravel road immediately north of the eastern portion of St. Mary Lake.

STOP 2-2
26.2 km LOWER ALDRIDGE FORMATION

Low outcrops that extend about 25 m along south side of St. Mary-Meacham Forest Service Road expose a shallowly west-dipping (015, 30W) sequence of lower Aldridge Formation metasediments (Fig. 26). These strata have been intruded by the gabbro sill complex and correlate with strata exposed midway up Bootleg Mountain to the north (Stop 2-6). Strata are a rusty weathering, sandstone-rich sequence of thin- to medium-bedded sandstone and interbedded siltstone/argillite. Sandstone beds 10-70 cm thick compose about 70% of the section. Plane lamination is the dominant internal texture in sandstone; however cross-bedding is present in beds exposed near the upper east end of the outcrop. Conspicuous concretions up to 30 cm diameter occur locally within sandstone beds and are composed of a recessive biotite-rich core and more resistant pale-colored rim (albitic?). These strata are some of the deepest surface exposures of the lower Aldridge Formation.

Medium-grained biotite is abundant in strata, possibly suggesting proximity to a gabbro intrusion. An increase in abundance and grain size of biotite adjacent to sill contacts has been noted elsewhere in the Aldridge Formation. A steep north-trending small fault zone with quartz-pyrite veinlets has an envelope of pale greenish-grey muscovite alteration; this alteration is mineralogically similar to muscovite alteration widespread in the Sullivan-North Star Corridor (Stop 3-2, 3-3).

26.7 km outcrop, lower Aldridge Formation metasediments. Strata are dark in color but not iron stained. Paler sandstone beds (albitic?-biotite) are interbedded with biotitic laminated siltstone. Dark chloritic veins with albite envelopes cut biotitic strata. These strata have a distinctly lower sulfide content than typical lower Aldridge strata, possibly due to weak albitic alteration associated with sills. Decrease in sulfide content of lower Aldridge strata associated with albitic alteration is noted at Stop 1-8.

STOP 2-3
27.2 km GABBRO AND ALTERED LOWER ALDRIDGE FORMATION

Park at small pull-out on north side of St. Mary-Meacham Forest Service Road at the east end of 150 m long roadcut. The road cut exposes a faulted section of gabbro and adjacent albite-chlorite altered lower Aldridge strata (Fig. 26).

Gabbro exposed in the eastern portion of the roadcut is juxtaposed against metasedimentary rocks along a fault zone of chloritic schist. Lower Aldridge strata occupy the central portion of the roadcut and are in contact to the west with gabbro. Strata are variably deformed and altered. The orientation of cross-laminations and scour features indicate bedding is overturned. Small-scale tight folds are associated with a steeply dipping north-tending spaced cleavage. Folding may be related to movement on the chloritic fault zone. Adjacent to the western gabbro contact is an albite-chlorite alteration zone. A transect from west to east from the contact portion of the western gabbro displays the following: (1) fine-grained chloritic gabbro is transitional over 1-2 m to chlorite-albite altered sediment in which bedding is largely obliterated; (2) bedded quartz-biotite-chlorite-albite metasediments; (3) pods of massive albitite with "spotted" biotite within albite-chlorite altered sediments; and (4) bedded quartz-biotite-chlorite-albite metasediments. Boulders of massive albitite occur just above the roadcut and indicate a more intensely albitized zone upslope. The western gabbro is massive, fine- to medium-grained and appears to be pervasively chloritized. Gabbro is cut by very coarse grained tremolite-calcite-limonite (sulfide) veinlets and chlorite-quartz-calcite sheared zones.

Sediment alteration appears to have been localized near the margin of the gabbro, presumably by hydrothermal fluids guided along the intrusion contact.

28.1 km Outcrop of west-dipping lower Aldridge Formation
28.2 km Outcrop of west-dipping lower Aldridge Formation

STOP 2-4
28.7 km PEBBLE WACKE AND CONGLOMERATE, LOWER ALDRIDGE FORMATION

A 50 m long road cut has been blasted where the St. Mary-Meacham Forest Service Road rounds a spur-ridge. The road cut exposes a broad zone of massive unbedded pebble wacke that contains a conspicuous steeply dipping zone of conglomeratic wacke. The pebble wacke is massive and fine-grained wacke with minor dispersed angular to rounded lithic fragments (<1%) up to 1 cm diameter. Contact relationships with bedded lower Aldridge strata are not exposed. The conglomerate is a steeply dipping body 4-5 m wide that lies within the pebble wacke (Fig. 26). Contacts between conglomerate and pebble wacke are transitional. The conglomerate is matrix supported and the wacke matrix is similar to the adjacent pebble wacke. Clasts are distinctly rounded, up to 10 cm diameter, (Fig. 28) and generally aligned parallel with the contacts of the zone. Fragments appear more muscovite-rich than the biotitic matrix. A steeply dipping fault with associated muscovite alteration envelope cuts the pebble wacke adjacent to the conglomerate zone.

The pebble wacke body is texturally similar to the clastic dike near Moyie Lake (Stop 1-6). Whether the pebble wacke is discordant or concordant with lower Aldridge strata is unclear. The diffuse contacts between conglomerate and pebble wacke, and unbedded yet very thick nature of the wacke are unlikely to be of sedimentary origin and a discordant intrusive origin appears more likely.

Regional mapping by Cominco Ltd. suggests the pebble wacke and conglomerate occurs close to the contact between lower and middle Aldridge Formations (Fig. 26).

Figure 28. Conglomerate (Stop 2-4). Field of view is 0.75 m.

Return to St. Mary-Hellroaring Creek Forestry Road.

34.1 km Junction. Turn north (right) on St-Mary Hellroaring Creek Forestry Road.
37.9 km Contact of gabbro and lower Aldridge Formation.
38.6 km Lower Aldridge Formation.
39.4 km Bridge, Hellroaring Creek.
39.5 km Strongly-foliated steeply-dipping lower Aldridge Formation.
40.9 km Shallowly-dipping lower Aldridge Formation intruded by concordant tourmaline-bearing granite pegmatite sills. Along the east side of the creek is strongly foliated amphibolite, presumably a sheared gabbro. The foliation in the gabbro is cut by quartz-feldspar veins.

STOP 2-7
41.0 km HELLROARING CREEK STOCK

Small outcrops on the southeast (upslope) side of the St. Mary-Hellroaring Creek Forestry Road (Fig. 1) expose a coarse pegmatitic phase of the Hellroaring Creek stock (Fig. 26) as well as banded rock (metasediment?). Very coarse grained (7-15 cm) pegmatite composed of quartz, muscovite, feldspar and pale green beryl cuts coarse granite with accessory tourmaline and garnet. Some of the granite displays tourmaline-rich bands.

Elsewhere the stock varies between a coarse-grained granodiorite and a pegmatite, with the southern exposures of the stocks generally finer grained than in the north (Wasylyshyn, 1984). Aplite is locally common, particularly near the intrusion margins. The coarser phases comprise dominantly feldspar (sodic plagioclase and microcline), quartz, muscovite and tourmaline, with minor garnet, some large euhedral beryl crystals and pyrite. The Hellroaring Creek stock has been explored as a possible beryllium resource.

Down the road to the northeast and across the creek, lower Aldridge strata are intruded by concordant tourmaline-bearing granite pegmatite sills. Adjacent to the creek a strongly foliated gabbro is cut by quartz-feldspar veins likely related to Hellroaring Creek stock. Analyses of the foliated gabbro (Sample S12AS, Table 3) suggests that it is part of the Moyie sill suite. The foliation in the gabbro is cut by quartz feldspar veins that are likely related to the adjacent Hellroaring Creek stock. These relationships support a pre-intrusive age of deformation.

Comment

The Moyie sills, dated at 1445 Ma, are interpreted to be penecontemporaneous with deposition of Aldridge sediments (Hoy, 1989). They are deformed by a compressive tectonic event, the East Kootenay Orogeny, that locally produced folds and a penetrative cleavage. The Middle Proterozoic Hellroaring Creek stock (Ryan and Blenkinsop, 1971), with a recent zircon UPb age of 1365 Ma (J. Mortenson, personal communication, 1992), cuts deformed lower Aldridge strata and Moyie sills (Leech, 1962). Hence, it is argued that Moyie sill emplacement is related to extensional tectonics during early development of the Purcell basin (Hoy, 1989), whereas the East Kootenay Orogeny, and possibly associated intrusion of the Hellroaring Creek stock, may be related to tectonics associated with termination of Belt-Purcell sedimentation (McMechan and Price, 1982).

Turn vehicle and return along St. Mary-Hellroaring Creek Forest Service Road to main trunk gravel road that runs along the south side of St. Mary River valley.

47.3 km Junction. Turn east (right) on main trunk gravel road.
48.0 km Bridge, Hellroaring Creek.

STOP 2-6
50.2 km VIEW OF GABBRO SILL COMPLEX, BOOTLEG MOUNTAIN

Stop beside the large clearcut area on the north side of the road. The cut area allows an unobstructed view of Bootleg Mountain, the dominant peak to the north. The slopes of Bootleg Mountain expose about a 2000 meter section of gabbro sill complex and lower Aldridge strata (Figs. 26, 29). The stratigraphic level of the Sullivan horizon is within a few hundred meters of the uppermost gabbro sill exposed. The aggregate thickness of gabbro in this section is estimated as 1200 m; the thickest sill, exposed on the lower slope, is about 700 m thick.

Schofield (1914) notes that the uppermost sill on Bootleg Mountain is composed of 25 m of micropegmatite granite in gradational contact with underlying hornblende gabbro 25 m thick. This sill may correlate with the gabbro-granophyre complex of the highest sill below the Sullivan mine (Stop 3-1). Within the underlying hornblende gabbro sill, Schofield (1914) noted replacement of augite and hypersthene by hornblende, as well as schlieren of "acid material" in the center of the sill.

Comment

The large volume of gabbro sills in the stratigraphic interval underlying the Sullivan deposit (>50%) raises the question , "What was the role of sill emplacement in ore formation and rock alteration?" Leech (1957) recognized two sets of gabbro sills in the St. Mary

Table 3. Analyses of samples from the Hellroaring Creek stock area.

Sample Number	Sample Name	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	SUM	Ce	Cs	La	Sc	V	F
45114	S12AS	58.81	2.08	13.88	12.00	0.23	3.61	8.75	1.11	1.20	0.12	99.50	52	5	22	34	403	640
45116	S13	74.13	0.01	14.72	0.54	0.00	0.02	0.38	2.97	6.16	0.11	99.57	—	9	—	—	240	—

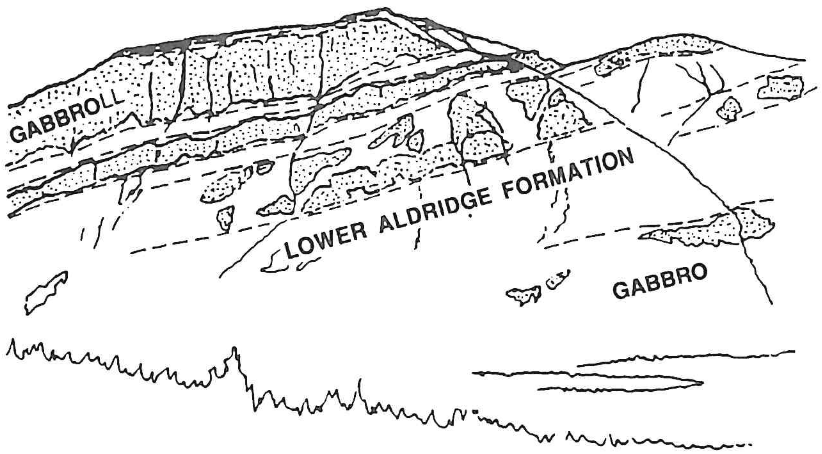

Figure 29. Sketch of Bootleg Mountain gabbro sill complex in lower Aldridge Formation (Stop 2-6).

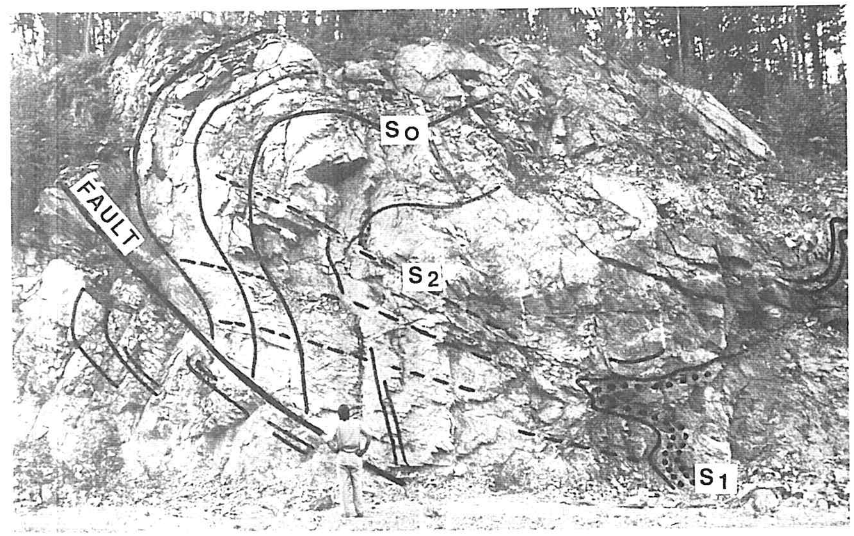

Figure 30. Large fold structure (F_2) folding bedding (S_0) and foliation (S_1) (Stop 2-7). Trygve Hoy for scale.

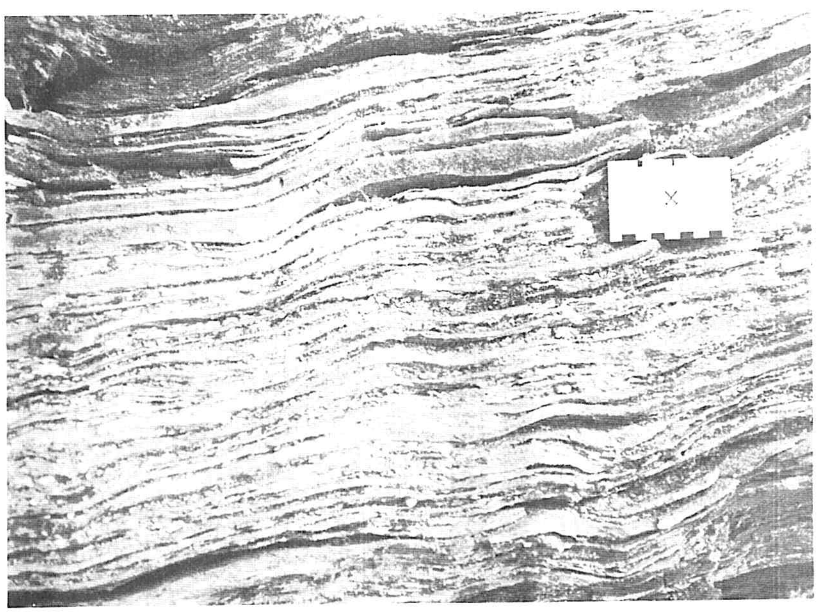

Figure 31. Quartz-muscovite schist at Stop 2-8. Note that foliation (S_1) is parallel to bedding (S_0).

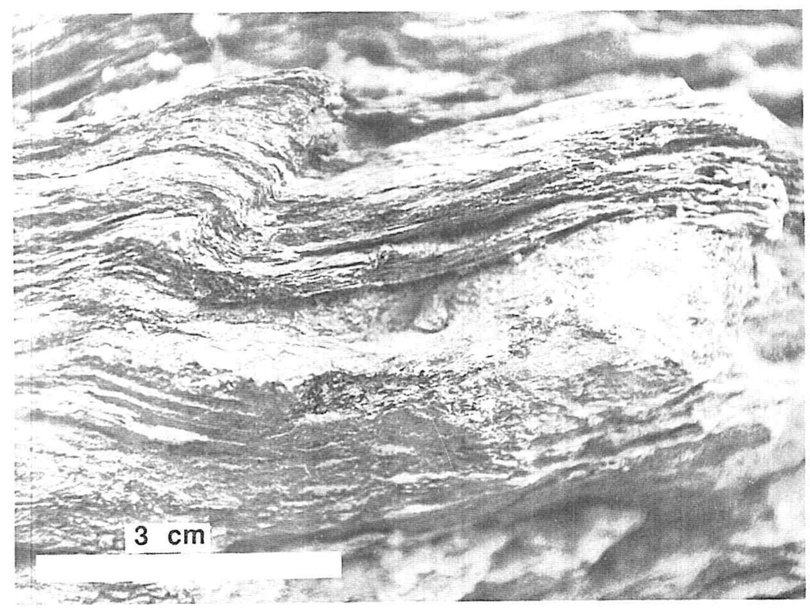

Figure 32. Close up of small-scale folding (F_2) of S_1 foliation, quartz-muscovite schist (Stop 2-8).

valley: a lower and thicker set within the lower Aldridge Formation and a higher and thinner set of sills within the middle of the middle Aldridge Formation. The stratigraphic distribution of Moyie sills throughout the Aldridge basin (Canada) generally displays these two groupings. At the Sullivan mine, the highest sill in the lower sill complex has intruded up to the stratigraphic level of the Sullivan orebody.

Höy (1989) suggested that the Moyie sills were intruded into unconsolidated Aldridge sediment and proposed the Guaymas sedimented rift basin as a modern analogy. Basalt sill complexes are known to intrude modern sediments to within tens of meters of the present seafloor (Einsele, 1982; Gieskes and others, 1982). A model for intrusion by Einsele (1982) assumes upward-younging sills as each subsequent magma injection rises to the top of previously altered sediment, achieves neutral buoyancy and spreads laterally. Such sills are only slightly younger than the sediments they intrude. One might speculate therefore, that the minimum age of the two gabbro sill complexes exposed southwest of the Sullivan mine (Leech, 1957) can be established by the stratigraphic level of the uppermost sill of each complex: an older intrusive event up to the time of formation of the Sullivan deposit, and a younger event during middle Aldridge time. The source magma chamber may have heated basinal fluids that formed the Sullivan orebody, as suggested by Hamilton (1984).

50.9 km Junction, Angus Creek Road. Continue east on main road.

STOP 2-7

58.5 km LARGE FOLD STRUCTURE NEAR ST. MARY FAULT

A prominent road cut cliff exposes a large, east-vergent fold in lower Aldridge Formation (Fig. 30). Massive siliceous sandstone beds are interbedded with minor beds of chloritic schist. A well developed foliation (S_1) in the chloritic schist is subparallel to bed contacts. Both bedding and a penetrative foliation (S_1) are deformed within the large fold structure (F_2). In the hinge zone of the fold structure, the S_1 foliation in the chloritic schist is deformed into a series of crenulation folds that define a S_2 foliation plane. The lower limb of the fold structure is cut by a planar, west dipping fault. The St. Mary Fault lies about 2 km to the southwest.

Comment

Such tight folds are uncommon in the Aldridge Formation except in the vicinity of large faults such as the St. Mary fault zone. This fold is east-vergent and compatible with Mesozoic right-lateral reverse displacement on the St. Mary Fault as has been interpreted on the similar Moyie Fault to the south (Benvenuto and Price, 1979; Stop 1-5). This younger folding, probably of Mesozoic age, clearly deforms the early penetrative cleavage, S1. The early cleavage is likely equivalent to the penetrative foliation in gabbro just east of the Hellroaring Creek Stock (Stop 2-5) and foliation in quartz-muscovite schists and phyllites at stops 2-8 and 2-1 respectively. This early foliation predates intrusion of the Middle Proterozoic Hellroaring Creek stock and formed during the East Kootenay Orogeny.

59.5 km Chlorite schist of St. Mary fault zone. Low outcrops on south side.

Turn vehicle and return west along main gravel road, retracing route to bridge over St. Mary River.

72.0 km Bridge, St. Mary River
72.3 km Junction, St. Mary Lake Road (paved). Turn east (right).
81.6 km Turnoff, St. Mary-Matthew Creek Forest Service Road. Turn north onto Forest Service Road.
82.2 km Junction at pump station, stay left on lower gravel track. Drive several hundred meters to pullout just before gate where road comes to Matthew Creek. Walk just past gate to prominent cliffs along east side of road.

STOP 2-8

82.4 km QUARTZ-MUSCOVITE SCHISTS, LOWER ALDRIDGE FM., MATTHEW CREEK

The cliffs expose gently east-dipping quartz-muscovite schists interpreted as of the lower Aldridge Formation. Schists are conspicuously thin bedded and composed of 1-3 cm quartzitic beds (metasandstone/siltstone) and more recessive 0.5-1 cm muscovite-rich beds that contain tourmaline and garnet (meta-argillites) (Fig. 31). Sillimanite has been reported by Leech, 1962). Foliation (S_1), defined by orientation of muscovite grains, is parallel to or very shallowly oblique to bedding. Small, open southwest-vergent folds (F_2) deform the S1 foliation (Fig. 32). The shape of small sigmoidal lenses of muscovite also suggests top-to-west shear.

Comment

The schists of Matthew Creek lie in a structural and metamorphic culmination centered on Matthew Creek. This structural culmination extends to the southwest and includes Hellroaring Creek stock. To the west across Matthew Creek and upslope is a mylonitic fault zone, the Matthew Creek thrust (Ransom, 1986). If the schists of Matthew Creek are in stratigraphic continuity with lower Aldridge strata to the east on North Star Mountain, then the schists appear to be at a higher stratigraphic level than the base of the Bootleg Mountain section above the Matthew Creek thrust. These relationships support contractional movement on the Matthew Creek thrust.

The intense and penetrative character of the foliation (S_1) is similar to foliation in metagabbro cut by veins related to the Hellroaring Creek stock (Stop 2-5) and presumably is related to the East Kootenay Orogeny of Middle Proterozoic age. Parallelism of foliation and bedding, and absence of evidence of isoclinal folding suggests "static" burial-type metamorphic environment. Small folds (F_2) are likely related to Mesozoic deformation (e.g. Stop 2-7).

Return to St. Mary Lake Road. Turn west (left). Return to Kimberley.

TOUR 3

SULLIVAN DEPOSIT AND ASSOCIATED DISTRICT-SCALE ALTERATION ZONE

Tour 3 will focus on the Sullivan mine and the corridor of altered rock and fragmentals that extends 6 km south of the Sullivan mine.

0.0 km From the traffic lights at intersection of Ross St. and Wallinger St., Kimberley, drive north on Wallinger St., turning left and then right at first stop sign. Follow street past switchback up hill. Continue past hospital and turn left at sign to Sullivan mine. Continue through gates at entrance to Sullivan mine property up hill to Sullivan mine.

Stop 3-1
4.5 km SULLIVAN MINE

Park in parking lot on left; the mine office is the first building past the parking lot. MINE TOURS OF THE SULLIVAN MINE MUST BE ARRANGED IN ADVANCE WITH THE MINE STAFF.

Introduction

The Sullivan deposit, one of the largest massive sulfide base metal deposits in the world, is a stratiform sediment-hosted exhalative (SEDEX) deposit characterised by bedded iron, zinc and lead sulfides formed as hydrothermal sediments on the basin floor. It has been well described in a number of papers including those of Freeze (1966); Ethier and others (1976) and Hamilton and others (1982, 1983) and guidebooks (Ransom and others, 1985).

The Sullivan deposit has produced 134 million tonnes of ore (to Oct, 1991; Höy, 1993) from an original deposit estimated to have originally contained more than 160 million tonnes of 6.5 % lead, 5.6 % zinc, 25.9 % iron and 67 grams per tonne silver. The Sullivan orebody comprises a broadly stratiform convex lens and lesser bands covering an area of 1.6 x 2.0 km and composed mainly of pyrrhotite, sphalerite, galena and lesser pyrite. Contacts with enclosing sediments are sharp and conformable; the orebody is truncated on the north by the Kimberley Fault. The deposit is divided by an irregular transition zone into two parts; a western massive sulfide body and an eastern zone of interbedded sulfides and silicate sedimentary rocks (Hamilton and others, 1981; Fig. 33).

The western massive sulfide body is a part of an extensive vent complex, which includes: (a) the massive pyrrhotite replacement body; (b) an underlying tourmalinite hydrothermal alteration pipe consisting of breccia, altered and fragmented strata and disseminated or veinlet sulfides; and (c) and albite-chlorite altered sediments in both footwall and hangingwall (Figs. 33, 34). The vent complex represents the zone of hydrothermal upflow and discharge at the seafloor.

The eastern bedded ores includes five distinct conformable layers of generally well-laminated sulfides separated by clastic sediments (Figs. 34, 35). The sulfide layers thin to the east away from the transition zone. Sub-ore grade sulfide layers of pyrite and pyrrhotite with subordinate sphalerite and galena persist beyond the eastern limits of the ore-grade sulfides. Bedded sulfides are interpreted as sulphidic sediments deposited on the basin floor adjacent to the vent complex of the western orebody.

Fragmental rocks

A wide range of coarse clastic rocks including breccias, pebble dikes and conglomerates are associated with the Sullivan mine. Collectively these are referred to as fragmentals by mine geologists. On the basis of texture, there are two distinct varieties of fragmentals (Delaney and Hauser, 1983) : (1) pebble fragmental composed of subangular to subrounded, granule to pebble-size fragments; and (2) breccia that consists of angular, granule to boulder-size fragments forming an intact to locally disrupted framework.

Underlying the western orebody is a fragmental body of breccia and pebble fragmental. This discordant body is almost 1 km in diameter and cuts at least 100 to 200 m of lower Aldridge strata (Figs. 33 and 34; Delaney and Hauser, 1983). Within the complex, Jardine (1966) mapped north-trending zones up to 150 m wide and 1000 m long of breccia (Fig. 34). Where unaltered, the clasts and matrix are indistinguishable in composition from enclosing lower Aldridge strata. The top of the fragmental body is a pebble fragmental ("footwall conglomerate") that extends to the east as an eastward-thinning concordant sheet. Below the western orebody, the footwall conglomerate merges downward into underlying breccias and pebble fragmental dikes. The footwall conglomerate is up to 50 m thick below the western orebody, and it is overlain by up to 12 m of bedded wacke that also thin and pinch out to the east (Ransom and others, 1985). The hanging wall conglomerate, a conformable pebble fragmental up to 4 m thick and containing some ore grade clasts, underlies the H sulfide laminations and has been traced for 1000 m along the southern margin of the orebody. A discordant, north-trending pebble dike extends up to the hanging wall conglomerate and probably represents the feeder.

Discordant fragmentals are interpreted to have resulted from violent release of pore overpressure in the sedimentary pile (Delaney

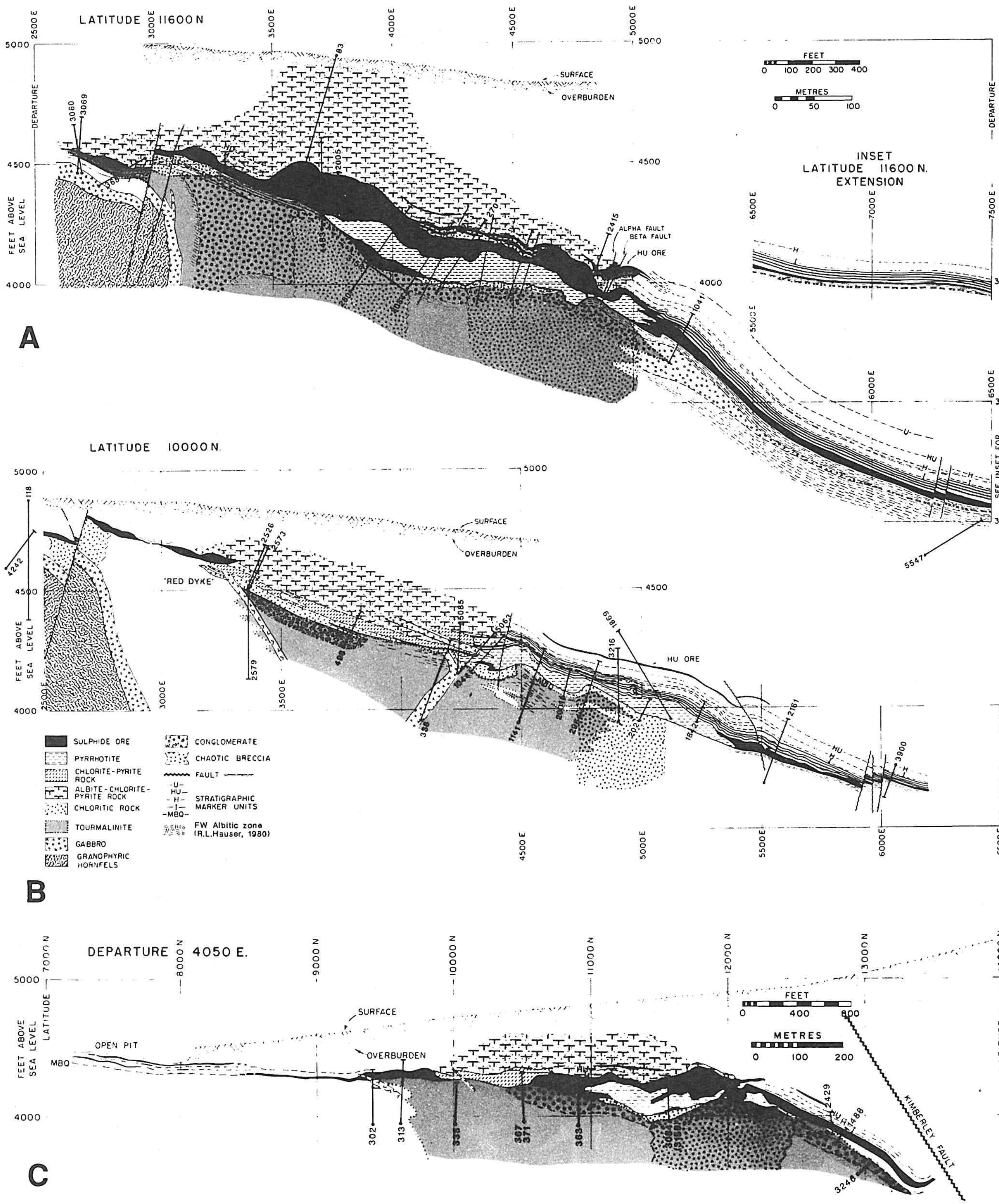

Figure 33. Geologic cross-sections through the Sullivan deposit (modified after Hamilton and others, 1982). (A) East-west cross-section at 11600 N (looking north). (B) East-west cross-section at 10000 N (looking north). (C) North-south cross-section at 4050 E (looking west).

and Hauser, 1983), possibly caused by heat from magma bodies at depth. Rounding of clasts may be explained by abrasion during ascent in a turbulent slurry. Erupted conglomerate may have accumulated as a mound that slumped eastward; accumulation may also have occurred in structural overlying related chaotic breccia bodies and related to withdrawal of material at depth. Bedded wacke between the conglomerate and sulfide body represent resedimented fine sediments related to sediment eruptive processes. Fragmental eruption immediately preceded sulfide deposition.

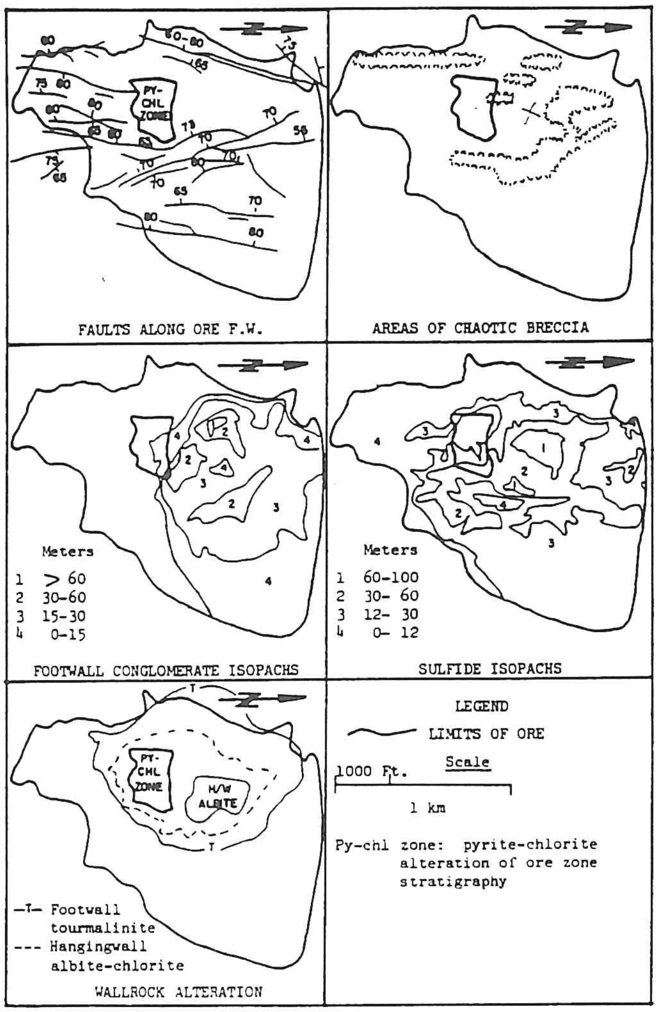

Figure 34. Plan view of Sullivan orebody showing distribution of faults, breccia, conglomerate and sulfide thickness, and wallrock alteration (from Ransom and others, 1985).

Massive sulfide body

The western orebody ranges from 10 to 100 m thick, averages 50 m thick, and is about twice as thick as the eastern bedded ores (Fig. 33). The lower portion consists of: (a) massive pyrrhotite lens up to 35 m thick and 350 m long with wispy laminae of galena; (b) overlain by massive banded galena-sphalerite-pyrrhotite; and (c) uppermost laminated sphalerite, galena and pyrrhotite intercalated with silicate laminae. The massive pyrrhotite passes laterally through an abrupt contact into coarse laminated galena-sphalerite-pyrrhotite of the transition zone. Siliciclastic laminae intercalated with the laminated sulfide are scarce in the area above the massive pyrrhotite and may reflect formation of the western orebody as a constructional sulfide mound feature. Ongoing hydrothermal activity produced the massive pyrrhotite lens over the vent zone by replacement of previously bedded or slumped sulfides and mobilization of Pb, Zn and other minor metals.

The eastern bedded ores average about 30 m in thickness and consist of five laterally persistent sulfide layers called "bands" in sharp conformable contact with interbedded wacke (Figs. 33, 35). From base to top the sulfide bands are named Main, A, B, C and D bands. The basal Main band consists of a series of fine-grained pyrrhotite, sphalerite and galena beds generally less than 3 cm but up to 30 cm thick. The upper part of the Main band and the overlying bands have up to 50% interbedded wacke; sulfide layers are almost monominerallic laminae up to 10 mm thick that persist laterally over 100 m. Eastern bedded ore was probably deposited by one or more of three processes: distal accumulation of sulfides by fallout from a

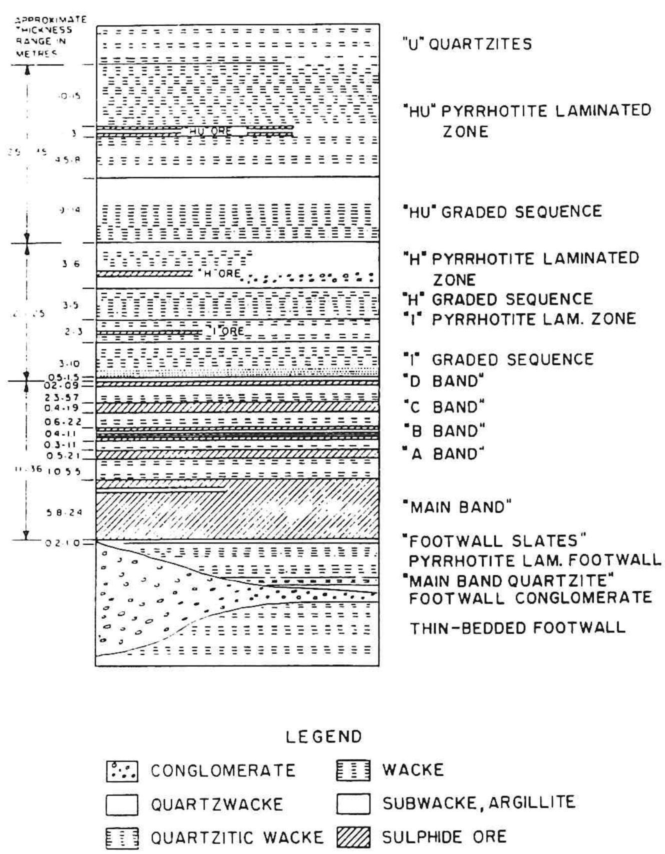

Figure 35. Ideal geological column, eastern part of Sullivan orebody (from Ransom and others, 1985).

hydrothermal plume within a brine pool or open basin water column, or resedimentation of sulfide material from the vent mound area.

The ore bands are interbedded with massive wacke beds ("waste beds") up to 5.7 m thick that compose as much as 40% of the sequence of the eastern bedded ores (Fig. 35). Unlike turbidite beds in the overlying middle Aldridge Formation, the waste beds or not graded. These wacke beds terminate abruptly to the west at the transition zone and generally taper to the east. The wacke beds include a variety of sulfide and non-sulfide clasts, typically angular or irregular in shape. The most significant clast type are sheets of well-laminated pyrrhotite and wacke up to several meters long and up to one meter thick that lie subparallel to the contacts of the wacke bed (Ransom, 1989). Pyrrhotite laminations terminate abruptly at clast margins. Small adjacent clasts suggest disintergration of large clasts during transport within the wacke bed; other clasts are folded suggesting plastic deformation during transport. The laminated pyrrhotite in the clasts is similar to the pyrrhotite-rich fringe of the Sullivan orebody. Waste beds are interpreted as debris flows derived from the marginal parts of the bedded ores (Ransom, 1989). A possible slump scar area 200 by 800 m in area where laminated sulfides are missing has been identified by drillling in the southern marginal area beyond the mining limit.

The eastern bedded ores and parts of the western orebody are overlain by three distinct stratigraphic intervals, the I, H and Hu (Fig. 35). Each interval consists of a lower 1 to 5 m thick graded quartz wacke bed and an upper 2 to 6 m thick argillite characterized by disseminated and laminated pyrrhotite ± sphalerite and rare galena. Distinctive laminae can be matched over 1000 m (Freeze, 1966); zinc and lead content increases up dip and southwards into ore grade material above the western part of the orebody. The texture and graded nature of quartz wacke beds in the I, H and Hu intervals are similar to regional turbidites; however these beds are anomalously thick (3-15 m thick) in the vicinity of the Sullivan mine relative to the thickness of equivalent turbidites east of the mine (0.2-0.3 m thick). This thickening of turbidites in the mine area may reflect the ponding of passing turbidity flows by the bathymetric depression that the Sullivan deposit.

Overlying the Hu interval is the U quartzite which is regarded as the base of the Middle Aldridge (Figs. 33, 35). Middle Aldridge sediments are typically thicker, coarser turbidites (silt to sand-sized wackes) than most found in the lower Aldridge.

Trace metal and gangue distribution

Within the Sullivan orebody, there are minor amounts of arsenopyrite, chalcopyrite, argentian tetrahedrite (freibergite), magnetite, ilmenite, rutile, sphene, hypogene goethite and in places cassiterite, stannite, bismuthian or antimonian boulangerite and/or jamesonite, native Bi-Sb alloy, and bournonite. Ruby silver (pyrargyrite) and scheelite are also reported (Hamilton and others, 1982).

The deposit is zoned, with lead + silver values decreasing toward the margin in the eastern ores. Tin and arsenic are concentrated in the western ores; antimony appears to halo the arsenic (Fig. 36). Higher absolute values of lead and silver and higher Pb/Zn and Ag/Pb ratios overlie the breccia zones (Freeze, 1966; Hamilton and others,

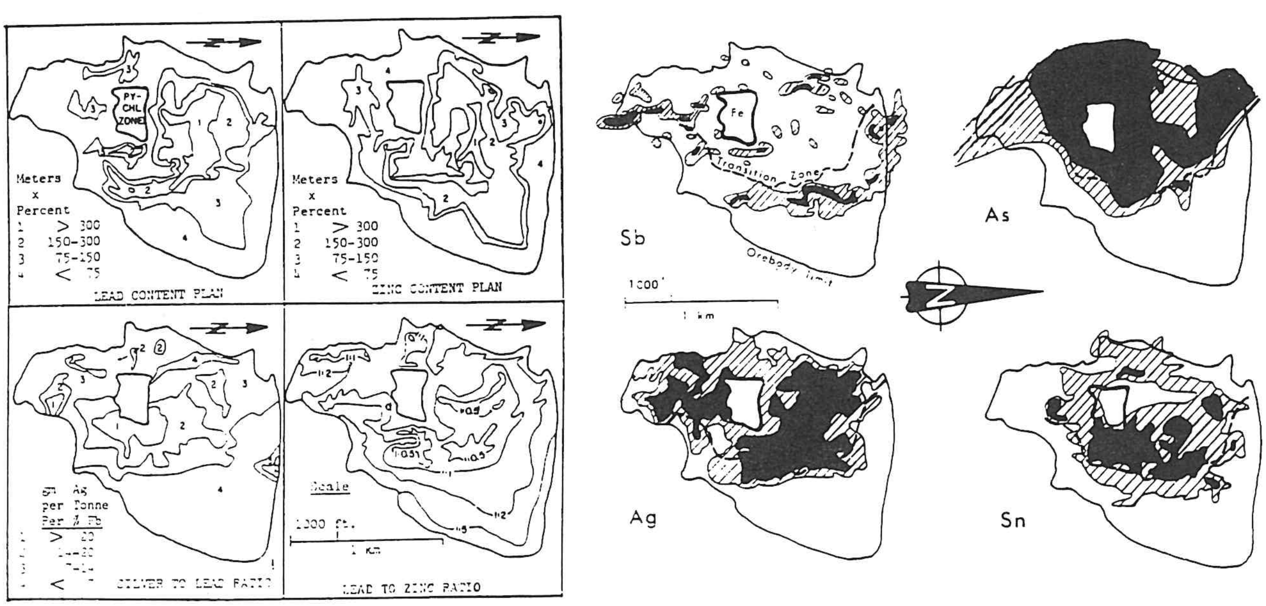

Figure 36. Plan view showing distribution of metals within Sullivan orebody (after Ransom and others, 1985 and Freeze, 1966). For Sb, As, Ag and Sn the white, hachured, and dark areas indicate increasing levels of each element. Fe = central pyrite-chlorite -calcite zone.

1982; Ransom and others, 1985). Vertical metal distribution patterns in the western orebody show increasing Pb and Zn and decreasing Fe upwards; in contrast, patterns in the eastern ore sequence show a general decrease upsection in Pb, Zn and Ag and increase in Fe (Hamilton and others, 1983).

The transition zone, the boundary between the western massive orebody and the eastern bedded ores, is an arcuate zone up to 75 m across that overlies the eastern margin of the footwall tourmalinite-pyrrhotite vent zone. It is characterized by thinning and complex folding of bedded ore. Several minor minerals (cassiterite, stannite, freibergite, arsenopyrite, bismuthian boulangerite and jamesonite, gudmundite, bournonite, ?bismuthinite, Bi-Sb alloy) rich in Sn, Ag and the semi-metals As, Sb and Bi appear concentrated near the transition zone in veins and disseminations associated with muscovite alteration (Leitch, 1992b). Silver is concentrated over the footwall vent zone inside the transition zone. The principal locus of silver may be in freibergite or in solid solution in galena (Freeze, 1966; Hamilton and others,1982). Scheelite probably controls the distribution of W.

Tin, recovered during mining of the western orebody, is primarily in cassiterite and is concentrated in an annular zone inside the transition zone coincident with higher arsenic concentrations (Freeze, 1966; Fig. 36). The association of tin-bearing minerals with muscovite or arsenopyrite-bearing cross-cutting structures suggests that Sn, with As, Bi, and possibly Sb, is related to late upflow of hydrothermal fluids in the transition zone (Fig. 37; Leitch and Turner, 1992).

Microprobe analyses indicate two varieties of chlorite in the deposit: 1) a widely distributed Mg-rich variety and 2) a Fe-rich variety present mainly in the eastern bedded ores. Carbonates vary from calcite to manganoan calcite, ankerite or siderite (Leitch, unpub. data). Garnets are spessartine-rich (Barnett, 1982). Limited microprobe data suggests a zonation in carbonate, garnet and chlorite to higher Mn toward the fringes of the Sullivan deposit.

Footwall sulfide network

The western massive sulfide portion of the Sullivan deposit is underlain by an extensive, in places intensely developed, network of irregular pyrrhotite-quartz-Fe carbonate veinlets in tourmalinite (Fig. 38; Leitch and Turner, 1992). These sulfide veinlets, as well as sulfide disseminations and sulfide matrix in pebble fragmentals are abundant within 50 m of the base of the deposit. The veinlets may also contain minor sphalerite, galena and rare arsenopyrite, chalcopyrite, cassiterite, tourmaline and scheelite. The sulfide network ranges from wispy, irregular veinlets that likely formed within weakly indurated sediments, to planar veins with increasingly abundant quartz and carbonate that appear to have formed within in more indurated parts of the feeder zone. Massive pyrrhotite veins up to 1 m thick occur locally.

A crescent-shaped zone around the margins of the tourmalinite pipe is characterized by the presence of sphalerite and galena in the veinlets (Fig. 38), with associated tourmaline-destructive muscovite or chlorite alteration. This may represent the site of late-stage fluid flow after sealing of the main central conduit (Fig. 37, Stage 2).

Fluid inclusion studies

Within the network veinlets, quartz, and to a lesser extent sphalerite and carbonate, contain abundant secondary or pseudosecondary fluid inclusions, many of which contain halite at room temperature (Leitch, 1992a; Leitch and Turner, 1991). Inclusions are not seen in wallrock detrital quartz grains. Fluids trapped in the

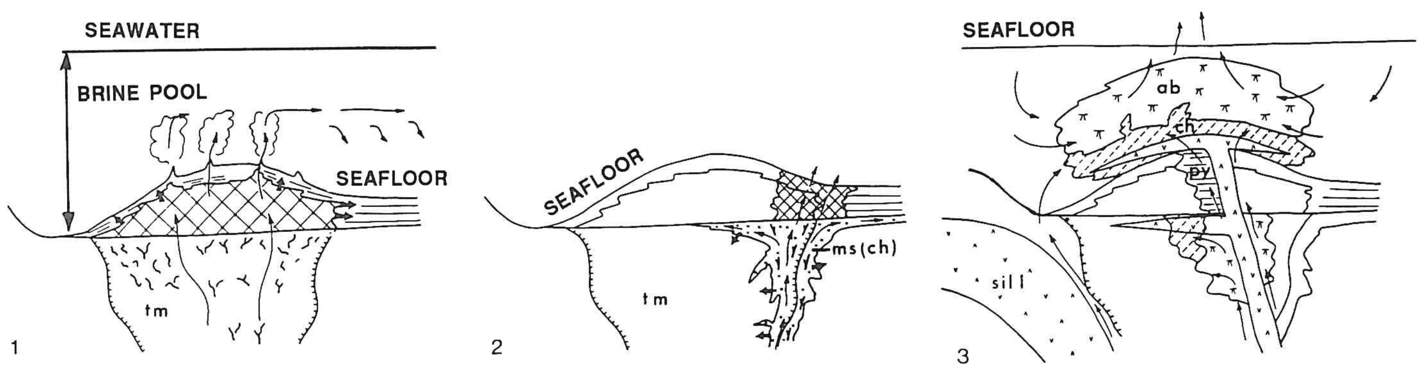

Figure 37. Postulated evolution of the hydrothermal system, Sullivan deposit (Leitch and Turner, 1992). (1) Main stage hydrothermal flow results in sulfide sedimentation in a brine pool, with progressive replacement of early-formed sulfides by massive pyrrhotite and underlying pyrrhotite network and tourmalinite alteration (tm). (2) Late-stage hydrothermal flow is concentrated at the periphery of the main vent zone giving rise to pyrrhotite-sphalerite-galena-sulphosalt veinlets in the transition zone, accompanied by muscovite and ?later chlorite (ms, ch). (3) Post-mineral fluid flow set up by magma body feeding Moyie sills and focused by vertical structures results in albite-chlorite-pyrite alteration (ab = albitite alteration; ch = chlorite-pyrite-pyrrhotite alteration; pyrite = pyrite-carbonate-chlorite alteration).

inclusions are probably samples of the mineralizing fluids, in places diluted and/or reset by metamorphism. Mineralizing fluids are characterized as saline 15-27 wt% ($NaCl+CaCl_2+?MgCl_2$) brines; homogenization temperatures (T_h) range from 200-300°C, but may reflect metamorphism as a pressure correction for 2 kb of 150-200°C would give trapping temperatures of 350-500°C (Roedder, 1984). Primary fluids trapped in 0.5 mm tourmaline recrystallized by chlorite-pyrrhotite alteration (8-18 wt% NaCl±?KCl, T_h=225-300°C) are similar to secondary fluids trapped in quartz associated with albite-chlorite-pyrite alteration (14-20 wt% NaCl±?KCl, T_h=165-250°C), suggesting late fluids were less saline and contained no significant Ca or Mg. Dilution of the mineralizing brines to <5 wt% NaCl, with mainly low but variable CO_2+CH_4, by metamorphic fluids is suspected in several generations of secondary inclusions with T_h = 200-350°C (Leitch, 1991). Lower temperature secondary inclusions in quartz, albite and sphalerite range from 3 to 20 wt% NaCl, T_h 90-150°C. Saline (halite-saturated) fluids are found at several other prospects (North Star, Quantrell, St. Joe and Kidstar) in Aldridge rocks. Limited fluid inclusion data from quartz veins not associated with mineralization and in veins cutting Mesozoic deformation features indicate the presence of lower salinity fluids (Leitch, unpub. data).

Alteration

Alteration at Sullivan is very well developed compared to most SEDEX deposits due to the feldspar-rich nature of the Aldridge Formation. Four main types of alteration are recognized at Sullivan: tourmalinite, muscovite, chlorite-pyrrhotite, and albite-chlorite-pyrite.

Tourmalinite is a very hard, dark brown to black rock with conchoidal fracture that extends at least 450 m below the deposit; within this zone, bedded and fragmental rocks are extensively tourmalinized (Hamilton and others, 1982; Shaw and Hodgson, 1980; 1986). Rare tourmalinite clasts are found in the distal part of the conformable footwall conglomerate outside the tourmalinite pipe. Tourmalinite is found as lenses up to 20 m into the hangingwall below parts of the I, H and Hu sulfide layers, and albitized tourmalinite remnants are found up to 100 m into the hangingwall. These features, combined with the coincidence of the tourmalinite with the footwall sulfide network (Fig. 37), suggest that tourmalinite and sulfide mineralization were contemporaneous (Leitch and Turner, 1992).

Tourmalinite consists of detrital quartz grains in a matrix of fine felted pale greenish-brown tourmaline needles 1-2 x 5-15 µm in size. The texture of the original rock is preserved in fine detail. Pyrrhotite is common as disseminations, laminations, wispy layers, veins, or blebs along microfractures. Tourmalinization has added significant amounts of boron and possibly some MgO and Fe_2O_3 to the sediments, while depleting them in K_2O, Na_2O, and CaO (Shaw and others, 1993a). This suggests preferential replacement of the original feldspar-clay matrix of the wackes by tourmaline.

Muscovite alteration is widespread (Fig. 38) and is recognized by the disappearance of biotite in hand specimen and

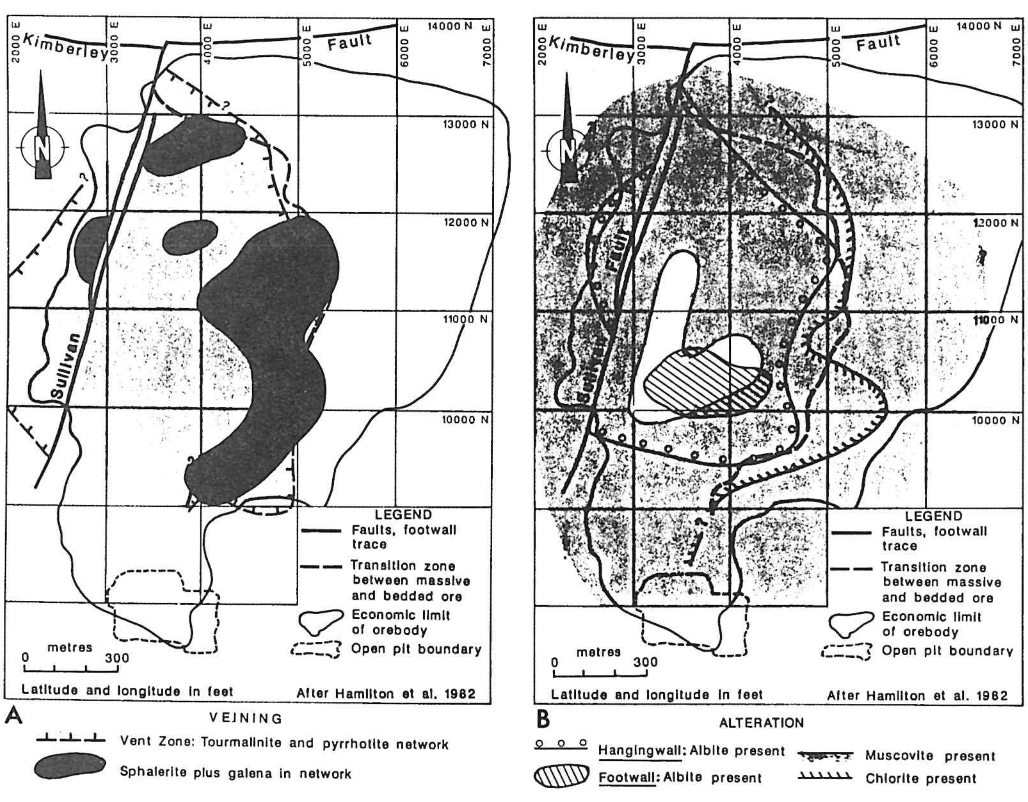

Figure 38. Plan views of the Sullivan deposit showing A) distribution of pyrrhotite-rich and sphalerite-galena-rich footwall mineralization, and B) distribution of albite and muscovite alteration in the hangingwall, albite and chlorite in the footwall (from Leitch and Turner, 1992).

feldspar in thin section, forming a grey to greenish rock in which original texture is blurred as alteration intensity increases. Two types of muscovite alteration are recognized; the first is grey, possibly early, that ranges from most intense near coarse-grained muscovite envelopes around quartz-sulfide-sulphosalt veins underlying the transition zone, to pervasive as an envelope to the entire deposit extending as far as the Concentrator Hill horizon 5 km to the southeast. The second is greenish and possibly later (distinctly fracture controlled), forming envelopes to planar carbonate-bearing veinlets. The envelopes coalesce in places to form semi-pervasive zones; garnet within the envelopes is replaced by carbonate, sericite and chlorite.

Chlorite alteration is mainly found inboard of the transition zone (Fig. 38) and may be divided into chlorite-pyrrhotite and chlorite-pyrite assemblages; the former is confined to the footwall of the deposit whereas the latter is found both below and above. Chlorite-pyrrhotite occurs mainly immediately below the massive pyrrhotite zone at the centre of the orebody in a conformable zone up to 10 m thick. Accessory minerals in this zone are quartz, muscovite, sphene, carbonate and tremolite (Leitch, 1991) with minor but significant allanite (Campbell and Ethier, 1984; Schandl and Gorton, 1992). Chlorite-pyrite occurs at the top of the sulfide body in a semi-conformable zone separating ore from albitite, as breccia matrix to albitite, and as crosscutting zones in the footwall, some of which are cored by albite alteration (see below). A cylindrical pipe of pyrite-chlorite-calcite 350 x 250 m across replaces the central portion of the western orebody (Fig. 34); in the core of this zone pyrrhotite is replaced by pyrite, and sphalerite and galena are absent. In the hangingwall, a 300 m diameter core of albitite, composed of albite-chlorite-sphene and minor quartz and pyrite and extending at least 120 m above the deposit, is surrounded by variably albite-chlorite altered rocks over a 1000 x 600 m area (Shaw and others, 1993b; Fig. 34). This hangingwall core of albite is displaced 240 to 300 m north relative to the center of the pyrite-carbonate alteration of the massive ore zone (McClay, 1983).

The Sullivan deposit is underlain by a thick set of Moyie gabbro sills. The uppermost sill has a central zone of "granophyre", a granular quartz-plagioclase-biotite rock. This sill-granophyre complex lies 450 m below the ore horizon except on the western side where it abruptly cuts up-section to form a 1 km wide north-trending, gently north-plunging arch, the crest of which is in contact with the western orebody (Fig. 33). The root zone of chlorite-albite-pyrite±calcite alteration coincides with a set of gabbro dikes, apophyses of this larger footwall gabbro sill (Turner and Leitch, 1992). Altered tourmalinite and sedimentary rocks adjacent to gabbro are zoned from proximal albite-chlorite to chlorite-pyrite; as some dikes and sills are locally altered this alteration in part postdates intrusion. The position of the dike set appears to have localized the plume of rising hydrothermal fluids that caused albite-chlorite alteration; zoned alteration envelopes on individual dikes indicate fluid flow was channeled along intrusion margins (Fig. 37; Turner and Leitch, 1992).

Albite-chlorite alteration is distinguished by elevated Na_2O (~7%); chlorite-pyrite by elevated MgO (~13%) and total Fe (~29%) (Shaw and others, 1993b). Both alteration types are significantly depleted in SiO_2, K_2O and CaO relative to unaltered Aldridge strata (Hamilton and others, 1982). The ascent of Na-rich, Mg-depleted hydrothermal fluids followed dikes in the Sullivan orebody localizing albitic alteration (Fig. 37). These fluids may be related to deep gabbro intrusions below the Moyie sill complex. Mixing with Mg-rich seawater caused peripheral chloritic alteration, and late collapse of the hydrothermal system caused chloritic alteration to overprint earlier-formed albitite (Fig. 37).

Structural and metamorphic history

Regional metamorphic grade is interpreted as biotite-Mn garnet middle greenschist facies formed at temperatures of about 400°C (McMechan and Price, 1982) and pressures of about 2.0 kb (5.8-7.6 km depth of burial; Edmunds, 1977) and related to the East Kootenay Orogeny of Middle Proterozoic age.

The Sullivan deposit also shows evidence of moderate recrystallization and deformation during Mesozoic thrusting to the east. The ductility contrast between massive pyrrhotite-galena rich sulfides, which have flowed, and the enclosing brittle clastic rocks, has caused strain to be localized in the orebody. Offset of the hangingwall albitite from its postulated roots in the pyrite-chlorite-calcite pipe in the orebody, segmentation of gabbro dikes in the orebody, and recumbent isoclinal folds in the sulfides, suggest overriding of the hangingwall toward the east (McClay, 1983). Continued deformation caused asymmetric folds, commonly overprinting isoclinal folds, and thrust faults at the boundaries of the orebody. This phase of deformation probably included the latest displacement on the Kimberley Fault; subsequently, lamprophyre dikes were intruded. West-verging thrust faults offset the lamprophyres and other earlier structures. Later Eocene crustal extension (Price and others, 1981) resulted in steep west-dipping normal Sullivan-type faults that offset the ore by up to 30 m (Fig. 33), along an important north trending ancestral structure subparallel to the axis of the Sullivan-North Star Corridor and the Belt basin. Early development of the Kimberley Fault is also suggested by albite alteration along it west of the deposit where the fault bends southward, and pyritic alteration similar to that in the center of the deposit, localized along the fault where it cuts bedded eastern ore.

Return down mine access road towards Kimberley

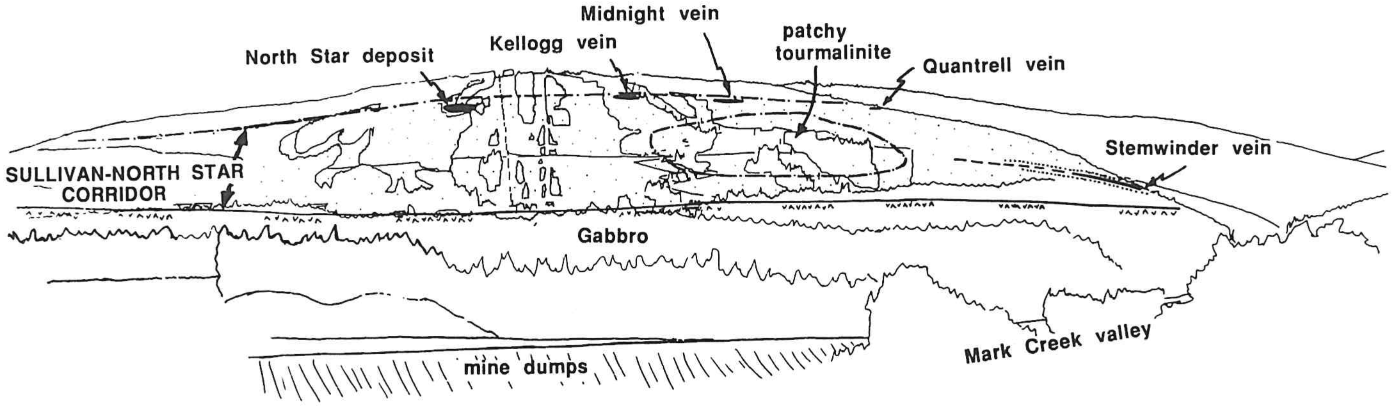

Figure 39. View looking west to North Star Hill and Sullivan - North Star Corridor from Sullivan mine road (Stop 2-4).

STOP 3-2

6.8 km VIEW OF SULLIVAN-NORTH STAR CORRIDOR

Pull off on the wide gravelled area on right (west) just before the mine gate at the sharp turn to the east. Walk to the edge of the bluff overlooking the Mark Creek valley.

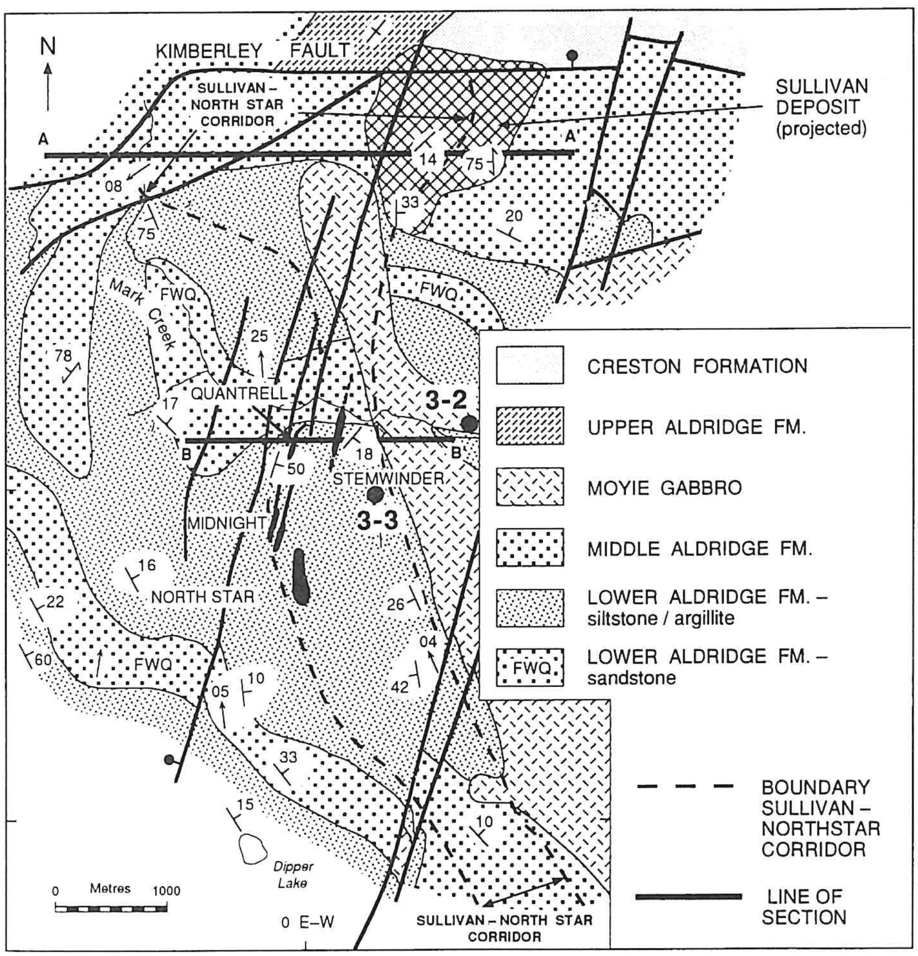

Figure 40. Geologic map of Sullivan - North Star Corridor (modified after Hagen 1985; Höy, 1993). Field stop locations are noted.

The view looks west across the slopes of North Star Mountain and the Kimberley Ski Resort (Fig. 39). Altered and fragmented rocks of the Sullivan-North Star Corridor are exposed on the lower part of the mountain and their distribution is marked by the brown iron-stained soils visible on the cleared ski runs. At the base of the slope, altered rocks of the Corridor are covered by terraces of Pleistocene sediments, now site of the lodge and condominium complexes. Gabbro underlies the Pliestocene sediments below the elevation of the ski lodge (Figs. 39, 40).

The uppermost slopes of North Star Hill are underlain by unaltered east-dipping lower Aldridge strata cut by a variably developed steep, west-dipping cleavage (Figs. 39, 40, 41). The western (upper) boundary of the alteration corridor occurs near north-trending mineralized structures (Midnight, Kellogg and Quantrell veins) and the stratiform and discordant orebodies at the North Star mine (Figs. 39, 40, 41, 42). East (downslope) of this line, fragmental or massive unbedded units are common and rock variably altered and mineralized. The mine dumps of the North Star mine have largely been recontoured but are still visible south of the chairlifts. A 400 m^2 area of patchy tourmalinite and elevated base metal content underlies the area above the Ski lodge. This tourmalinite zone is on strike and semi-continuous with tourmalinite that forms an envelope on the north-trending Stemwinder vein (Figs. 40, 42). The Stemwinder mine area is just out of view on the south side of Mark Creek (Fig. 39).

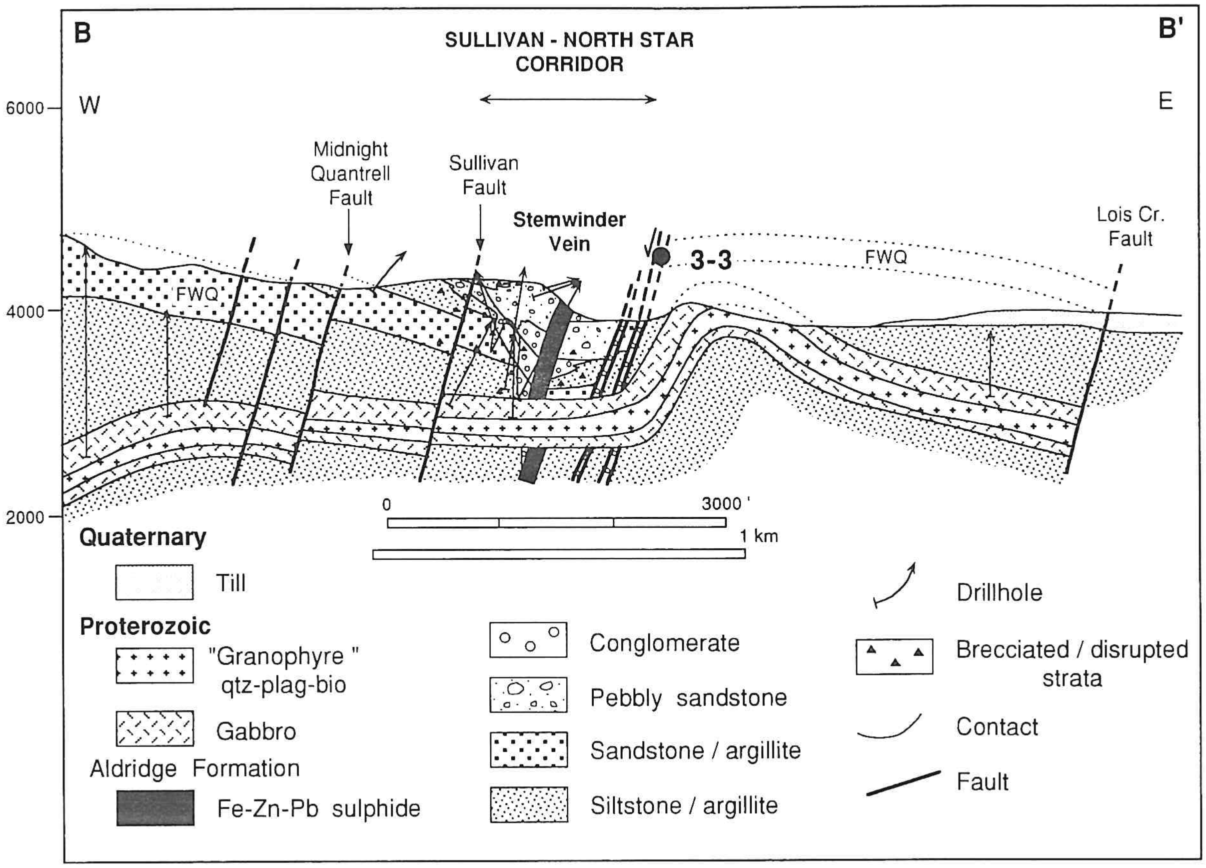

Figure 41. East-west cross-section of central Sullivan - North Star Corridor illustrating the distribution of stratigraphic units, intrusions and the Stemwinder vein. Location of section is shown on Figure 40 and lies along the south side of Mark Creek, south of Stop 3-2. Projected locations of Stops 3-2 and 3-3 are noted.

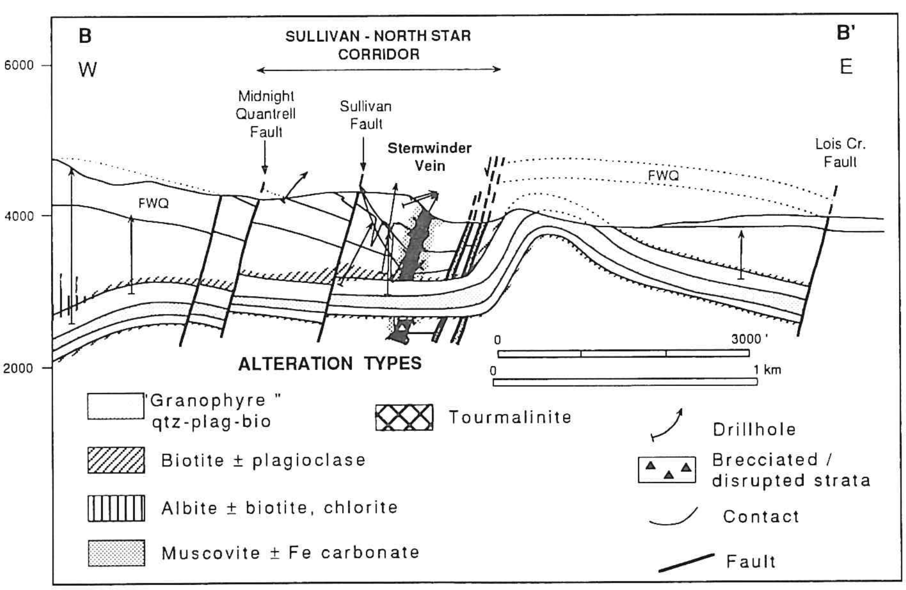

Figure 42. East-west cross-section of central Sullivan - North Star Corridor illustrating the distribution of rock alteration. Same section as Figure 41. Location of section is shown on Figure 40.

Comment

The Sullivan-North Star Corridor is an elongate, north-south-trending zone of fragmental and disrupted sedimentary rocks that are variably altered and mineralized. The Corridor extends 6 km in length, 1-3 km in width, and over 500 m in stratigraphic thickness (Figs. 26, 40, 41, 42, 43, 44). The Kimberley Fault truncates the Corridor to the north while at the southern end the Corridor terminates in outcrops north of the Quaternary deposits of the St. Mary Valley (Fig. 26). The Corridor is widest adjacent to the Kimberley fault.

Three gabbro sills with an aggregate thickness of 650 m have been intersected in drilling below the Sullivan deposit (Fig. 41). These sills are the upper part of a widespread sill complex intruded into the lower Aldridge Formation. The uppermost gabbro sill ("Footwall Sill") forms a discordant arch-like feature roughly coincident with the Corridor and contains a core of plagioclase-quartz-biotite "granophyre". This biotite granophyre may represent injection of a differentiated silicic melt produced by assimilation and melting of Aldridge sediments, or the *in situ* reconstitution of sediment trapped between two gabbro sills by magmatic fluids. The top of the arch coincides with the stratigraphic level of the Sullivan orebody.

<u>Fragmental rocks</u>

Fragmental rocks include bedded conglomerate, semi-concordant bodies of massive pebble wacke, clastic dikes of conglomerate and breccia, and disrupted strata. Bedded conglomerate immediately underlie the Sullivan deposit, extend across the width of the Corridor, and is composed of beds up to 5 m thick and with graded tops. The original distribution of conglomerates is unclear as this stratigraphic level is only preserved in the northern part of the Corridor near the Kimberley fault. The bedded conglomerate unit is typically 30 to 50 m thick, but exceeds 300 m thickness west of the Sullivan deposit (Fig. 41).

Bedded conglomerate is commonly underlain by disrupted strata and clastic dikes that extend at least to the base of the footwall quartzite (Figs 41, 43). Clastic dikes centimeters to meters in width are highly varied in texture and include angular breccia, conglomerate and pebble wacke. Major clastic dikes are commonly associated with sulfide veins. Disrupted strata extend throughout most of the Corridor and are identified by abundantof small displacement faults, variable bedding attitude, contorted beds, small clastic dikes, and elevated sulfide as veinlets and disseminations.

Pebble wacke is composed of small rounded lithic fragments up to 10 mm scattered in a wacke matrix. Pebble wacke occurs as massive units that are conformable or semi-conformable, 10's to 100's of meters thick, and transitional laterally into disrupted strata. The Stemwinder vein cuts a semi-

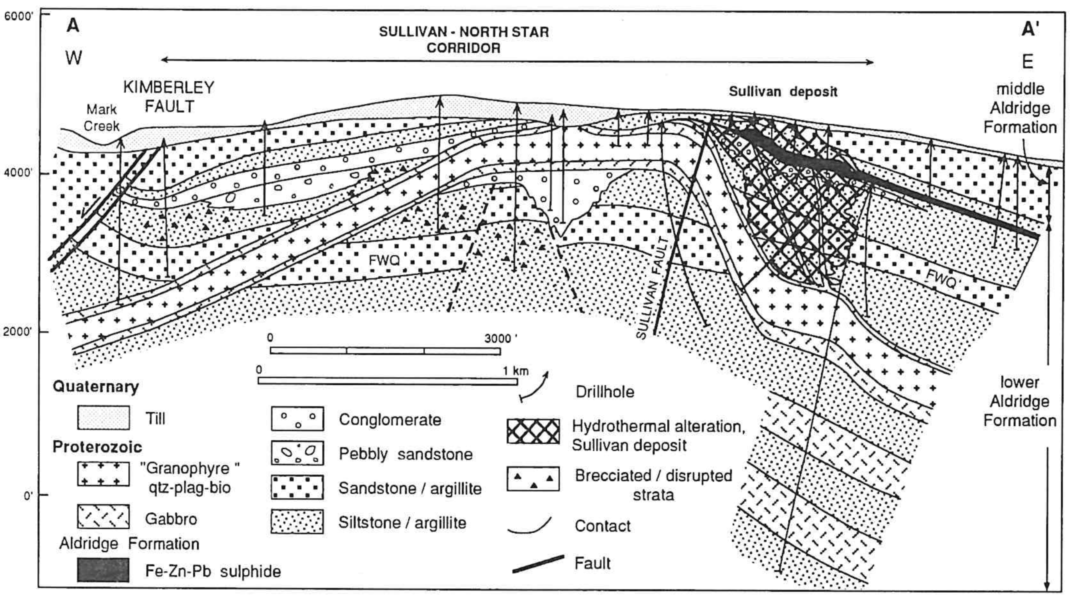

Figure 43. East-west cross-section of northern Sullivan - North Star Corridor illustrating the distribution of stratigraphic units, intrusions and the Sullivan deposit. Location of section is shown on Figure 40.

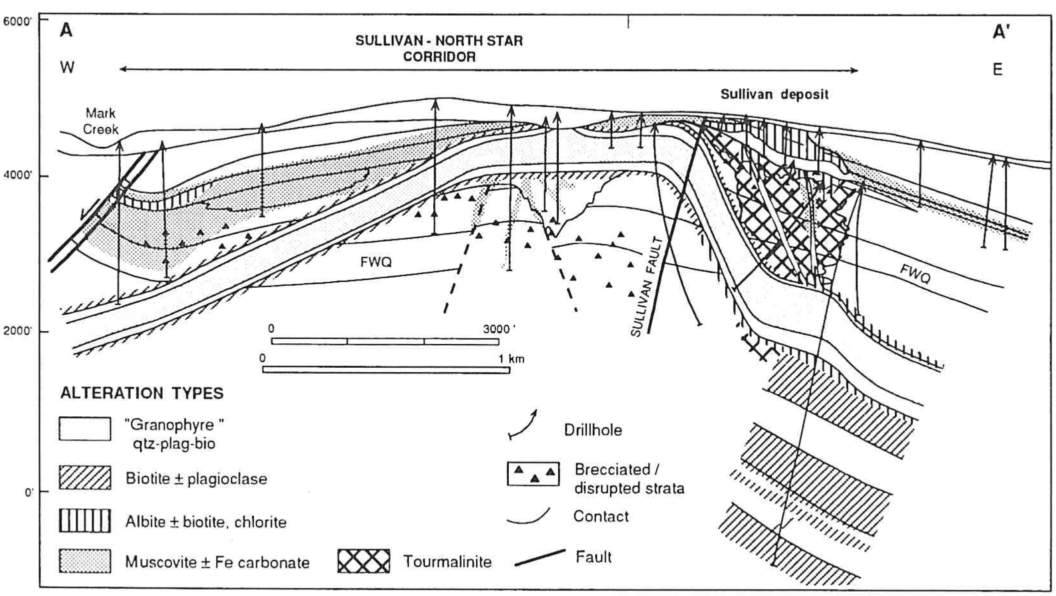

Figure 44. East-west cross-section of northern Sullivan - North Star Corridor illustrating the distribution of rock alteration. Same section as Figure 43. Location of section is shown on Figure 40.

conformable body of pebble wacke that is transitional laterally and down into disrupted strata (Fig. 41).

<u>Rock alteration</u>

Sandstone and siltstone of the Aldridge Formation is typically composed of 15-20% detrital plagioclase. Altered rocks in the Sullivan-North Star Corridor are distinguished from unaltered equivalents by the near absence of feldspars and the combined abundance of muscovite, sulfides, chlorite, epidote and garnet (Fig. 45; Leitch and others, 1991). Sulfides include abundant pyrite and pyrrhotite, lesser sphalerite and galena, and trace chalcopyrite and arsenopyrite (Leitch and others, 1991). Major alteration types include tourmalinite, muscovite-pyrite, garnet, albite-biotite-chlorite, biotite-plagioclase hornfels, and quartz-plagioclase-biotite granophyre (Figs. 42, 44). There is no evidence for addition of hydrothermal silica to altered rocks. Intensity of alteration is highly varied within the Corridor and related to a number of discrete hydrothermal centers, the major ones being the Sullivan deposit (Fig. 44), and the Stemwinder vein system (Fig. 42).

Tourmalinite alteration bodies occur as a pipe underlying the western part of the Sullivan deposit, as an envelope on the Stemwinder vein, as a cluster of small bodies extending south from the Stemwinder vein to the North Star deposit, and as a small body (Myrtle Mountain) at the south end of the Corridor (Shaw and Hodgson, 1980; Hamilton and others, 1982; Hagen, 1983, 1985).

Muscovite alteration results in destruction of detrital feldspar and the lack of biotite formation during later regional metamorphism, resulting in a soft grey to greenish-grey rock (Leitch and others,1991). Muscovite is most abundant as fine (50 micron) flakes. Both pyrite and pyrrhotite can be abundant (distinct from tourmalinite in which pyrrhotite is common and pyrite absent). Muscovite alteration is widespread throughout much of the Sullivan-North Star Corridor and occurs as widespread zones often associated with disseminated and veinlet pyrrhotite and pyrite and roughly coincident with thick fragmental units (Figs. 42, 44). Muscovite alteration also occurs as envelopes on pyrrhotite +/- sphalerite-galena veinlets that cut tourmalinite below the transition zone of the Sullivan deposit (Leitch and Turner, 1992) and as alteration of rock fringing footwall tourmalinite and hangingwall albite-chlorite at the Sullivan deposit and tourmalinite at the Stemwinder deposit.

Disseminated porphyroblasts of garnet occur throughout the Corridor (Leitch and others, 1991). Beds rich in pale pink garnet ("coticule rock") occur associated with bedded sulfides at the Sullivan and North Star deposits; disseminated garnet occurs throughout the Corridor.

Three large albitite bodies occur within the Corridor: (1) overlying the western Sullivan orebody (Shaw and Hodgson, 1980; Hamilton and others, 1982) (2) a kilometer west of the Stemwinder vein immediately overlying the footwall gabbro (Turner and Leitch, 1992); and (3) 3 km west of the Sullivan deposit within the conglomerate

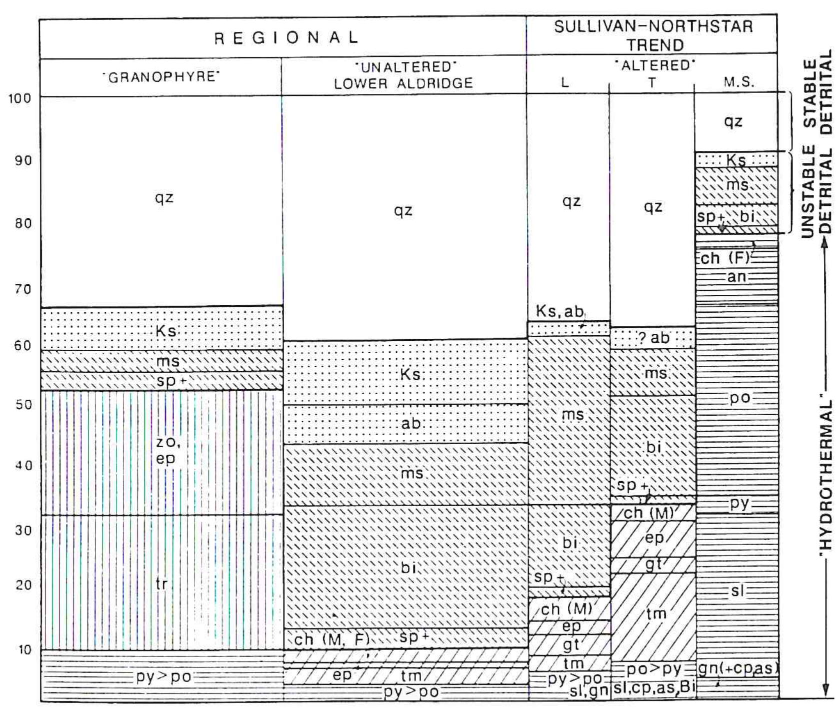

Figure 45. Schematic diagram of averaged modal mineralogy (abundances in per cent) for altered and unaltered rocks in the vicinity of Sullivan-North Star Corridor (from Leitch and others, 1991). L = altered lower Aldridge, T = tourmalinite, M.S. = massive sulfide.

Abbreviations are: Qz=quartz; Ms=muscovite; Bi=biotite (bracketted if relict); Ch=chlorite (*=metacrysts, M=magnesian, F= iron-rich, Hb=hydrobiotite); Ca=calcite (Do=dolomite, ankerite, siderite); Ks=microcline; Ab=albite (Pl=plagioclase); Tm=tourmaline (*=2 types, see text; Dr=dravite, otherwise schorl); Ep=epidote, clinozoisite (Zo=zoisite); Sp=sphene; Mz=monazite; Po=pyrrhotite (± secondary marcasite, pyrite); Py=pyrite (Lm=limonite); Ap=apatite; C?=?carbonaceous matter; Zr=zircon; Ru=rutile; Sl=sphalerite; Cp=chalcopyrite; Gn=galena; Il=ilmenite (Mt=magnetite).

horizon (Figs. 42, 44). Smaller occurrences of albitic rock are associated with altered sediment adjacent to gabbro contacts below the Sullivan deposit. Typically, albite is associated with biotite, with various degrees of chlorite alteration of biotite. Above the Sullivan deposit, massive albitite is associated with chlorite.

Massive fine to medium-grained granular quartz-plagioclase-biotite rock or "granophyre" occurs as the central portion of the Footwall Sill (Figs. 42, 44). Contact between granophyre and gabbro is transitional; elsewhere granophyre is noted to be gradational with biotitic alteration of sediment.

Origin of Sullivan-North Star Corridor

Bedded pebble fragmental appears to have been erupted to the seafloor via clastic dikes (Delaney and Hauser, 1983; Hagen, 1983,1985). Grading at the tops of bedded conglomerates suggests transport on the seafloor as mass flows, possibly due to collapse of an eruptive pebble fragmental mound. The well sorted and rounded nature of the conglomerate yet local derivation precludes simple collapse of strata on a fault scarp; instead conglomerate dikes indicate rounding and winnowing occurred below the seafloor during upward movement. The lack of bedding in pebbly wackes suggests subsurface disaggregation and freezing of silt and sand-rich sediments due to liquifaction rather than seafloor transport and deposition. Disturbed sediments could represent collapse due to evacuation of underlying sediment during eruption (Hagen, 1983, 1985).

The energy required to disrupt and erupt sediment may come from rapid release of overpressured fluids within permeable horizons (e.g. footwall quartzite), possibly due to heating by gabbro intrusion at depth, or liquifaction induced by seismic energy from fault movement. The coincidence of some clastic dikes and large sulfide veins in the Corridor suggest common structural controls for sediment eruption and later hydrothermal upflow. Conglomerate eruption was closely followed by mainstage hydrothermal activity at the Sullivan deposit.

Of the alteration types, tourmalinite appears to be most closely linked to ore formation. Tourmalinite is spatially associated with the two major sulfide systems at Sullivan and Stemwinder, and based on stratigraphic evidence at Sullivan is also temporally associated with ore formation (Leitch and Turner, 1992). The widespread distribution of muscovite alteration around but not directly associated with the major sulfide occurrences suggests it reflects lower temperature fluid-rock interaction away from focused discharge areas. At the Sullivan deposit, muscovite alteration postdates tourmalinite formation and is associated with the waning stages of hydrothermal activity (Leitch and Turner, 1992). Garnet-rich beds associated with bedded sulfides appears to reflect Mn-rich sediments exhaled on the seafloor. Disseminated garnet reflects Mn enrichment of wallrock during alteration of the Corridor (Leitch and others, 1991). The close association of some albite-chlorite and albite-biotite-chlorite alteration with gabbro dike and sill contacts below Sullivan deposit and elsewhere in the Corridor indicate this type of alteration was synchronous with or postdated gabbro emplacement and is later than ore formation (Turner and Leitch, 1992).

Structural controls

The north-south alignment of fragmental and altered rocks, sulfide veins and discordant zone of the Footwall Sill indicate a northerly-trending fault system controlled the position of the Corridor. The widening of the Corridor along the Kimberley fault is in part an artifact of preservation of higher stratigraphic levels such as widespread bedded conglomerates and associated conformable alteration adjacent to the fault. However, the distribution of pebble wacke and disrupted sediments between the footwall quartzite and Sullivan horizon is also significantly wider adjacent to the Kimberley fault than it is further south (this stratigraphic interval is exposed in drilling throughout the Corridor). This distribution suggests an ancestral Kimberley fault was active during formation of the Sullivan deposit. The location of the Sullivan deposit may well reflect a high permeability zone along the intersection of these two fault zones. Early fault movement resulted in fragmentation and eruption of sediments, followed by widespread hydrothermal activity within the Corridor. Focused discharge resulted in formation of a large seafloor sulfide body (Sullivan), as well as a smaller sulfide vein (Stemwinder deposit) and overlying seafloor deposit, now largely eroded (North Star deposit). Emplacement of gabbro intrusions was locally controlled by faults within the Corridor; hydrothermal alteration associated with gabbro emplacement was focusec along discordant limbs of intrusions.

Return to Kimberley.

0.0 km Ross St. and Wallinger St., Kimberley, B.C. Drive west on Ross St. following signs to Kimberley Ski area. Drive up switch backs, past Happy Hans Campground. Stay left at fork to Purcell, Rocky Mountain and Silver Birch Condominium/chalets.
4.0 km Turn right onto road to ski lodge. Drive north on road to large parking area on north side of ski lodge complex and adjacent to base of chairlift. Park vehicle. Walk up main access road to the upper ski area that leaves northwest corner of parking lot at ga and climbs the slope to the north. Permissi to drive the road may be granted by ski area management.

STOP 3-3
4.3 km ALTERATION AND FRAGMENTALS ALONG WESTERN MARGIN OF SULLIVAN - NORTH STAR CORRIDOR

This stop is a walking tour of the lower portion of the ski road that provides a transect across the eastern margin of the central part of the Sullivan-North Star Corridor (Figs. 40, 41) and provides a look at representative alteration types and fragmental rocks of the Corridor. This field stop is several hundred meters south of the section B-B' illustrated in Figures 41and 42. In general, this stop traverses the down-to-west faults immediately west of the gabbro arch (Fig. 42). Based on the presence of clastic dikes that occur within this fault zone, and alteration and mineralization that was focused along some clastic dikes, it is likely that the fault zone was active during Middle Proterozoic hydrothermal activity within the Corridor. The discordant western side of the gabbro arch subcrops just down slope from the parking lot.

The least altered strata occur at the base of the hill. Upslope are tourmalinite bodies, fragmental rocks and muscovite alteration of lower Aldridge strata. Steeply dipping conglomeratic zones are interpreted as important fault structures (Figs. 41, 42) that localized hydrothermal alteration.

0 m Gate at base of access road. Outcrop on the north side is a biotitic siltstone, similar to regional Aldridge strata. The abundance of biotite suggests these rocks are only weakly altered. Within the Sullivan-North Star Corridor, as the degree of alteration increases, the abundance of muscovite increases while the abundance of biotite decreases. This rock may represent less altered rock marginal to the Corridor. A north-trending spaced cleavage is conspicuous.

30 m A pile of rock on west side of road provides a view of the range of rock types within the Corridor. Rocks include: weakly altered conglomerate, intensely muscovite-pyrite altered conglomerate, gossanous breccia, tourmalinite matrix fragmental and gabbro.

100 m Tourmalinite. Road traverses cleared area. Low outcrops are on the west side of the road at edge of the forest. The cutcrop closest to road is a weakly biotitic (i.e. brownish color) muscovite-pyrite altered rock. This rock more altered than biotitic siltstone at gate (i.e. more sulfide, muscovite).

At the edge of the trees is a small tourmalinite body, distinguished by its hardness, black color and prominent NW-trending quartz veins. Quartz veins may reflect brittle nature of tourmalinite during deformation. Just upslope are shallowly dipping muscovite-pyrite altered siltstones. Pyrite is distinctly more abundant than in adjacent tourmalinite.

Ten meters upslope along the tree line is another small tourmalinite body surrounded by muscovite-pyrite altered strata. Some tourmalinite is distinctly soft due to the presence of some muscovite. Timing relationships of tourmalinite and muscovite are unclear but at the Sullivan deposit, tourmalinite is cut by later muscovite-sulfide alteration (Leitch and Turner, 1992). As many tourmalinite bodies within the Corridor are surrounded by muscovite-pyrite altered rock, this detourmalinization process may be widespread.

130 m Muscovite-altered tourmalinite. Road continues through forest. There is a low rib of outcrop 30 m north of the edge of the clearingt on west side of road. Dark, shallowly northwest dipping strata are altered to a tourmalinite-muscovite assemblage. Early tourmalinite may be partially altered by later muscovite alteration (i.e. detourmalinized).

160 m Clastic dike. Continue to low rib of outcrops 60 m north of tree line on west side of road. A shallowly north-dipping sequence of fine-grained sandstone and thin argillite beds is cut along a steep north-trending contact by a matrix-rich pebble wacke to conglomerate. The discordant contact is exposed on the north side of the largest outcrop.

Map distribution of units within lower Aldridge Formation suggest a down to the west fault in this vicinity (Fig. 40); this clastic dike along with others exposed further up the road may represent an early fault zone, active prior to lithification of the rock.

180 m Clastic dike and alteration. Continue to low rib and bench of outcrops 80 m north of tree line on west side of road. A shallowly north-dipping bedded sequence is altered to tourmalinite and muscovite (tourmalinite later altered to muscovite?). Bedded sequence is cut by pebble clastic dike exposed on the flat bench on the north side.

230 m Intensely-altered clastic dike. Continue to the bold north-trending rib with fresh roadcut exposures on west side of road. Rock has been intensely altered to muscovite with abundant disseminated pyrrhotite and minor sphalerite and galena. A steep north-trending foliation is well developed. A weak fragmental texture is preserved in some fresh surfaces; alteration has largely destroyed the original textures.

This clastic dike was a locus of hydrothermal fluid flow and is likely a north-trending fault. This fault set is a part of the structural control of alteration within the Sullivan-North Star Corridor. These clastic dikes are similar to the north-trending fragmental bodies described by Hamilton and others (1982) underlying the Sullivan orebody.

Return to vehicle. Retrace route to Kimberley.

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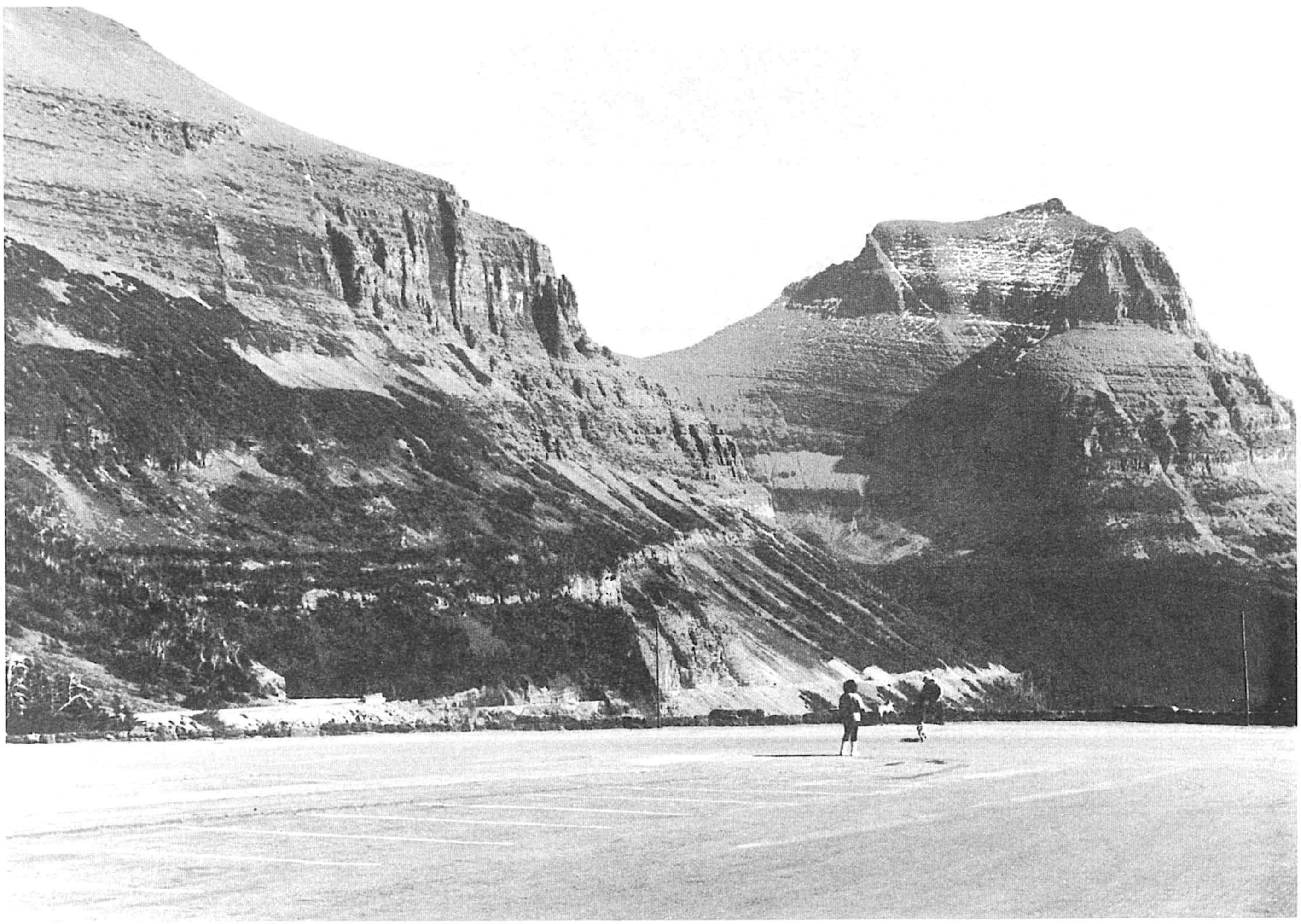
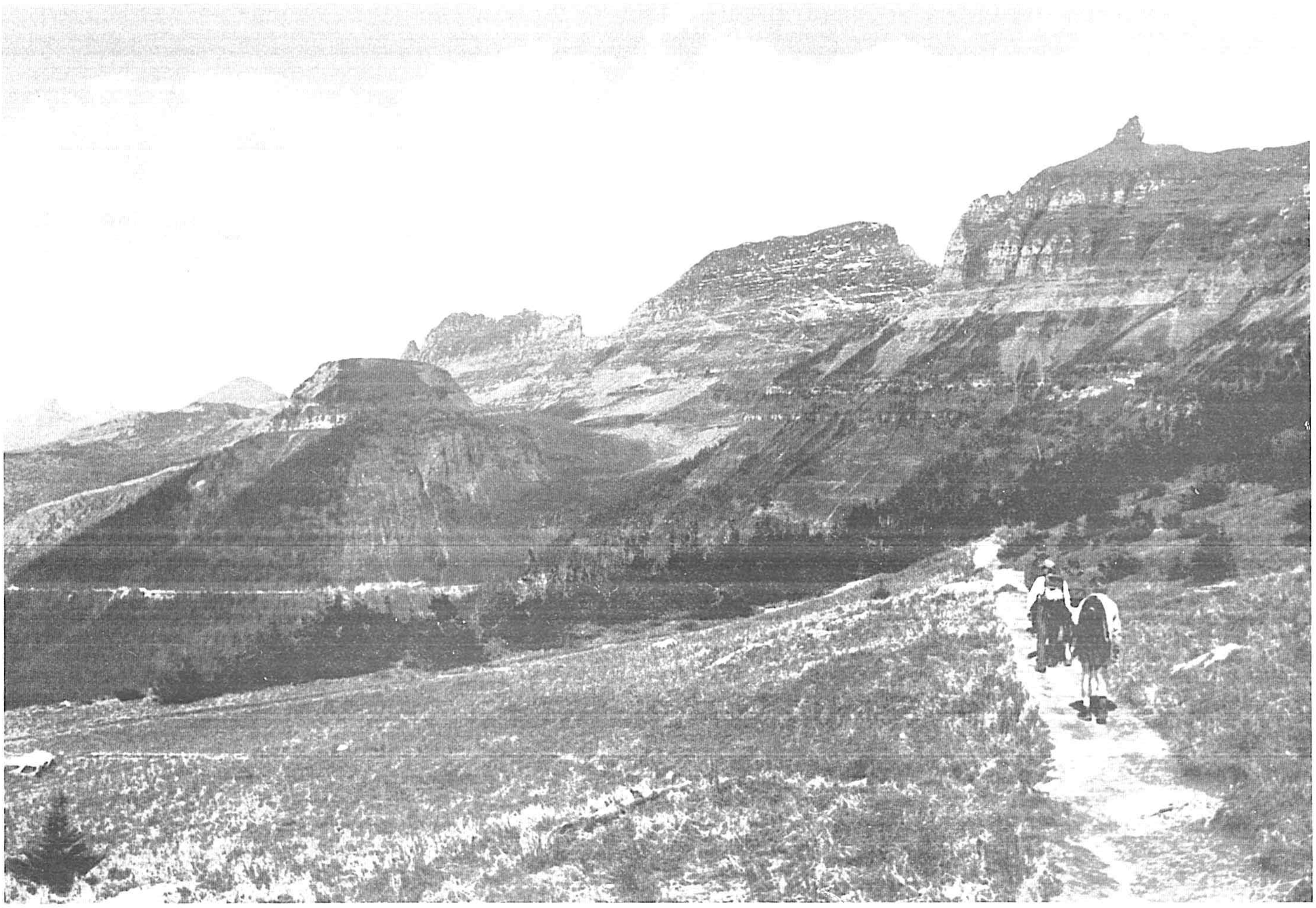
Geologic Tours: Glacier National Park and the Whitefish Range



Relaxing with dogs at Logan Pass. *View is to Piegan Mountain to the east-northeast. The lower part of the mountain is underlain by Helena Formation, with Snowslip Formation starting above the lower shadowed cliff. That limestone cliff is held up by a Proterozoic diabase sill.*



Starting a hike along the Highline Trail just north of Logan Pass.
Going-to-the-Sun-Road is in lower left, below Haystack Butte. Most of the exposed rocks are Helena Formation. The Snowslip Formation forms the highest ridge on the right, with peak called the Bishop's Cap. (Who's luckier than us?)

GEOLOGIC GUIDE FOR THE AREA OF LOGAN PASS, ALONG THE HIGHLINE TRAIL TO GRANITE PARK CHALET, AND TO THE LOOP ON GOING-TO-THE-SUN ROAD, GLACIER NATIONAL PARK, MONTANA

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INTRODUCTION

A walking trip along the Highline Trail, Glacier National Park, Montana, begins at the parking lot of the Visitors Center at Logan Pass, proceeds north across Going-to-the-Sun Road (with a small side trip along the road), contours north along the Garden Wall for 7.1 miles (11.4 km) to Granite Park Chalet, and then decends to the Loop, a hairpin turn on Going-to-the-Sun Road for another 3.8 miles (6.2 km). Total field trip distance is 10.9 miles (17.6 km) (fig. 1). The trail gradually traverses up section through parts of the Middle Proterozoic middle Belt carbonate Helena Formation and diorite sill, the argillaceous Snowslip Formation, and the Purcell Lava (fig. 2). The USGS 1:24,000-scale topographical maps covering the area along the fieldtrip are Logan Pass, Many Glacier, and Ahern Pass quadrangles. The general geology along this route is shown on the Geologic Map of Glacier National Park (Whipple, compiler, 1992).

Because this area is in a National Park, sample collecting, and the use of rock hammers is prohibited.

This park is habitat for a large population of bears. Please observe all of the Park's cautionary advice concerning bears and other wildlife.

GEOLOGIC FEATURES VIEWED FROM THE LOGAN PASS PARKING LOT

Toward the end of the Pleistocene Epoch, about 20,000 years ago, Logan Pass was the site of huge ice fields at the heads of alpine glaciers that flowed into eastern and western drainage basins. These two large glaciers carved the U-shaped valleys that descend both east and west from the pass.

To the south and west of Logan Pass are the prominent mountain peaks of Reynolds Mountain and Clements Mountain. These are both matterhorn-type peaks that were carved to their present shape by the headward erosion of alpine glaciers. At the base of Clements Mountain is an arcuate protalus rampart, formed during the late-Neoglacial expansion. All of the glaciers in the Park reached their late-Neoglacial maximum positions in the mid 1850s, based on tree-ring analysis (Carrara and McGimsey, 1981). The glaciers in Glacier National Park are Neoglacial remnants that occupy cirques eroded during Pinedale glaciation. The Garden Wall, along which the Highline Trail runs, is a sharp-crested arete that separated glaciers flowing east from those that flowed west.

To the northeast of Logan Pass is the rounded profile of Piegan Mountain. During most of the summer months a stream of water emerges from a position high on the west side of the mountain and flows fairly straight down the mountain side. This stream creates a waterfall just west of the tunnel on Going-to-the-Sun Road east of Logan Pass. It is melt-water from Piegan glacier situated in a cirque on the east side of Piegan Mountain. The water flows through the mountain along bedding planes near the top of the Helena Formation that dip west toward Logan Pass.

Rocks exposed around the parking lot, and used in rock walls, are the Helena Formation. They are primarily dolomite, with some limestone, and contain variable amounts of fine-grained quartz arenite and clay minerals. The medium to dark gray rock weathers to various shades of gray and tan. Abundant stromatolites (fossil cyanobacteria) of various forms are also present and will be described in greater detail later in this field guide.

The Snowslip Formation, which contains numerous layers of pinkish argillite and dolomite, overlies the Helena Formation. The contact between the two formations is just a few feet above Logan Pass. Pale red and pinkish rocks in the slopes west and northeast of the pass are part of the Snowslip Formation and are well exposed along the Highline Trail.

METABENTONITE BED ALONG GOING-TO-THE-SUN ROAD

A metabentonite bed in the Helena Formation is exposed in the first road cut 295 ft (90 m) west of Logan Pass. Stratigraphically, it is about 215 ft (65 m) below the contact between the Helena and Snowslip Formations and appears as a conspicuous pale-green layer 4 to 8 inches (10 to 20 cm) thick. The bed was first collected by James L. Wilson and analyzed by Goldich and others (1959) who suggested it is altered volcanic ash. More recently, the bed was studied by Obradovich, Zartman, and Peterman of the U. S. Geological Survey, who were able

Raup, O.B., Whipple, J.W., McGimsey, R.G., 1993, Geologic guide for the area of Logan Pass, along the Highline Trail to Granite Park Chalet, and to The Loop on Going-to-the-Sun Road, Glacier National Park, Montana: *in* Link, P.K., ed., **Geologic Guidebook to the Belt-Purcell Supergroup, Glacier National Park and vicinity, Montana and adjacent Canada:** Belt Symposium III Field Trip Guidebook, Belt Association, Spokane, Wa., p. 97-112.

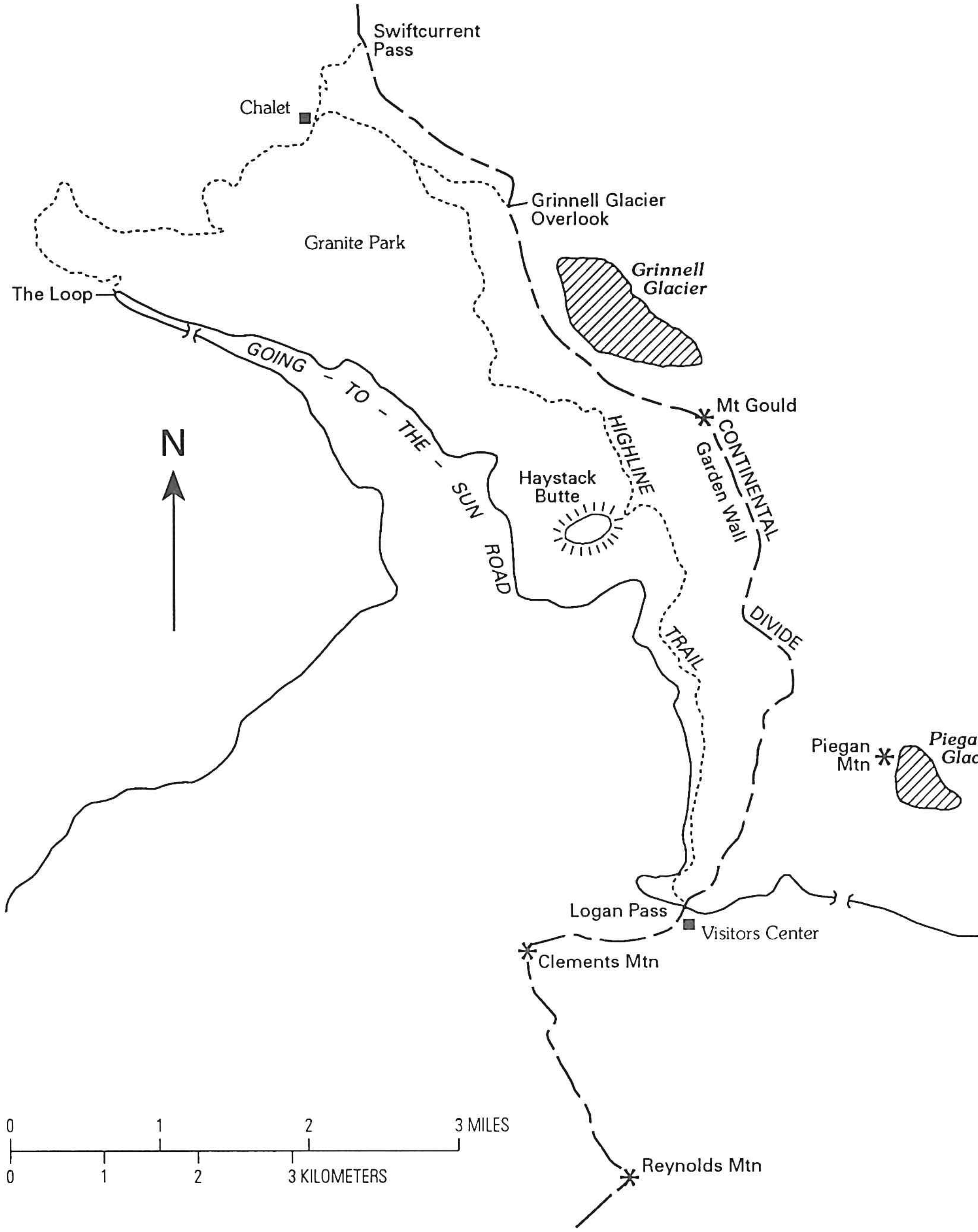
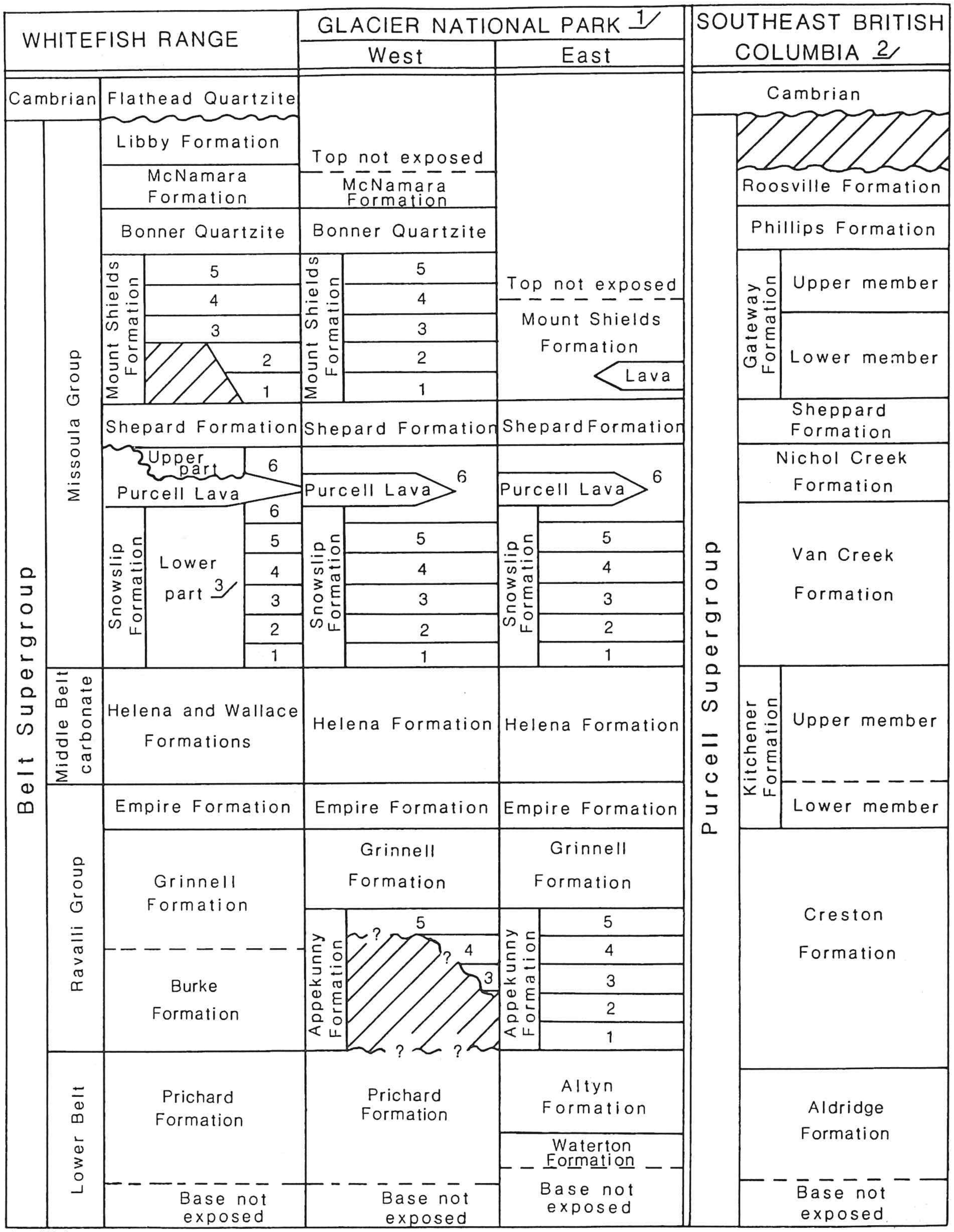


Figure 1. Index map showing the route of the Highline Trail, Glacier National Park, Montana, between Logan Pass and Granite Park Chalet, and the Loop (hairpin turn) on Going-to-the-Sun Road.

to separate small authigenic zircon crystals and extract a U/Pb date of about 1350 Ma (Harrison, 1984). Unfortunately, the zircons were observed to have detrital cores, which introduces some uncertainty as to just what the date reflects. Nevertheless, the reported age appears to correlate with regional magmatic events recorded in similar Proterozoic sequences in the northern Cordillera (Whipple, 1989; Doughty and Chamberlain, 1992). If we can accept the age of the metabentonite as around 1350 Ma, then younger Missoula Group strata could very likely be as young as 900 Ma, as suggested by Obradovich and others (1984).



1/ This report.
2/ Modified from McMechan, 1981.
3/ Upper and lower parts tentatively assigned to Snowslip Formation.

Figure 2. Correlation of Belt and Purcell Supergroups, Glacier National Park, Whitefish Range, Montana, and adjacent part of Canada.

GEOLOGIC FEATURES VIEWED FROM THE HIGHLINE TRAIL

Numerous stops will be made along the trail to point out various geologic features. Because specific stops are difficult to identify, the geologic features in this guide will be described in their approximate order along the trail. Photographs and descriptions will be used to help locate them.

Helena Formation

The first outcrops along the Highline Trail are of the Helena Formation. The Helena Formation in Glacier National Park, where measured along Going-to-the-Sun Road between The Loop and Logan Pass, and along Haystack Creek, the next drainage north of Haystack Butte, is 2,540 ft (774 m) thick and consists primarily of dolomite, limestone, and minor quartz arenite. In this section, it can be divided into three distinct parts.

The lower part is 590 ft (180 m) thick and consists of interbedded quartz arenite and thin-bedded dolomite near the base; thin beds of horizontally laminated and "molar-tooth" dolomite (fig. 7) in the middle; and thick smoky-gray limestone beds near the top (Whipple and others, 1984; 1985).

The middle part of the Helena Formation which is 1,978 ft (603 m) thick, is dominated by dolomitic molar-tooth beds, some as much as 100 ft (30 m) thick. A few thin beds of quartz arenite and stromatolitic limestone are also present.

The upper part consists primarily of interbedded stromatolitic limestone, dolomite, oolitic limestone, and quartz arenite. At the base is an interval of stromatolitic limestone about 100 ft (30 m) thick, the Conophyton zone (Rezak, 1957), composed of Baicalia-Conophyton stromatolite cycles (Horodyski, 1983) (fig. 5). The massive character of the Conophyton zone causes it to stand in relief in most sections of the Helena Formation in the Park.

Diorite sill

A prominent sill of diorite, 130 ft (40 m) thick at this location, is crossed by the Highline trail just beyond the handrail along the narrow part of the trail that traverses the cliff face above Going-to-the-Sun Road (fig. 3). This sill intrudes the Helena Formation throughout much of the Park. It ranges widely in stratigraphic position, from near the base of the Helena in the southeast part of the Park, to the lower part of the Snowslip Formation in the northernmost part of the Park. Most of the sill is medium- to fairly coarse-grained, but near its chilled borders the rock is very fine-grained. It is composed predominantly of pyroxene and plagioclase feldspar, with minor amounts of alkali feldspar, quartz, hornblende and Fe-Ti oxides (Mejstrick, 1975). Low grade metamorphism produced chlorite throughout the rock, and epidote along joints and minor faults.

Fingers of igneous rock, that split away from the main body of the sill, occur in the outcrop adjacent to the trail. The tapered ends of the fingers probably point in the approximate direction of magma injection (fig. 4). It is probable that maximum hydraulic wedging, and therefore maximum separation of the host rock, would be toward the source of the fluid. The least wedging (toward the ends of the fingers) would be away from the fluid source and in the direction of flow. The southwesterly direction of injection in this area is inferred from this and another exposure just east of Logan Pass.

The limestone of the Helena Formation adjacent to the sill has been bleached and altered by contact metamorphism. In areas where heating from the sill was moderate, organic matter was removed and the limestone recrystallized to marble. Where heating was more intense, the carbonate and other minerals in the limestone were recombined into calcsilicate hornfels.

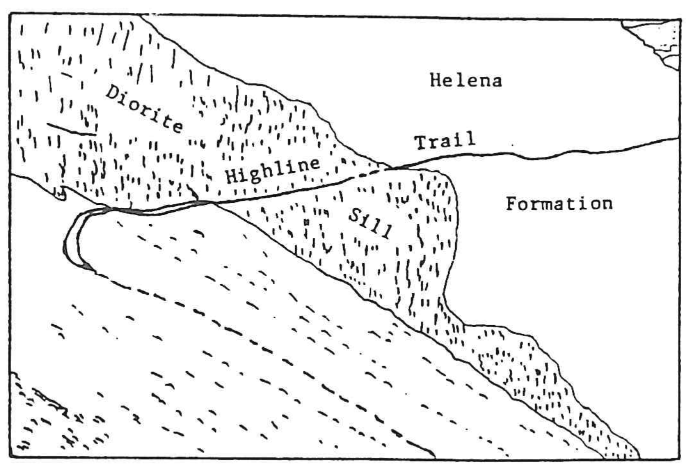


Figure 3. Photograph and sketch of diorite sill viewed toward the south from the Highline Trail north of Logan Pass. The sill is approxomately 40 m thick at this location.

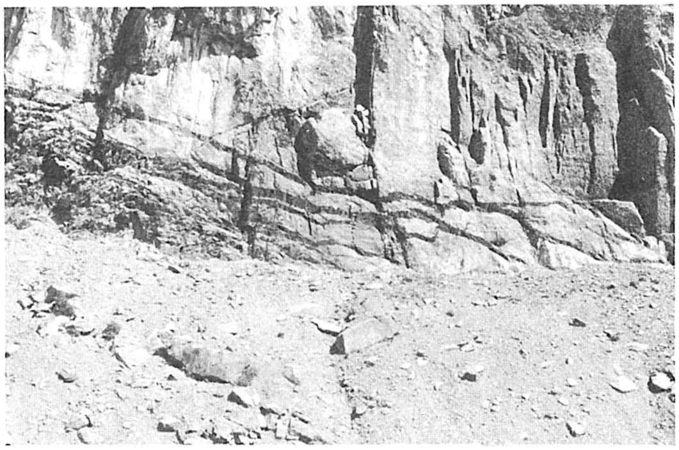


Figure 4. Fingers of diorite sill in the Helena Formation near Logan Pass. This photograph was taken from Going-to-the-Sun Road to the right of (west) and below the area in Figure 3.

The Highline Trail passes the diorite sill again between Haystack Butte and Granite Park Chalet. At this locality, the sill is very coarse-grained, and the large pink alkali feldspar and green epidote crystals stand out in contrast to the typical finer-grained, and darker rock. This is part of a granophyric zone that extends from the Fifty Mountain area in the north part of the Park to Dawson Pass in the south (Mejstrick, 1975).

The age of the sill remains enigmatic. Hunt (1961; 1962) obtained an age of 1,110 Ma from amphibole using the K-Ar method.

More recently, a two-point Rb-Sr isochron yielded ages of 605 and 717 Ma using whole rock and alkali feldspar from samples collected from the granophyric zone (Z. Peterman, unpublished data). Sills of similar composition and thickness in the upper part of the Helena Formation in the neighboring Bob Marshall Wilderness, south of the Park, have been dated at 725 to 775 Ma by the K-Ar method (Mudge and others, 1968).

A good view of the sill and its relationship to the Helena Formation can be obtained by looking back toward Logan Pass from farther along the trail (fig. 3).

Stromatolites

The Conophyton zone, a layer of stromatolites about 100 ft (30 m) thick, is crossed by the trail several hundred feet north of the sill. The stromatolites take many forms; some individual heads resemble cabbages; others are stacked in single or branching columns (fig. 5). Each form was a colony of cyanobacteria (blue-green "algae") that grew in a shallow sea, probably many miles from shore but in water depths ranging from the tidal zone to as much as 60 m. The water had to be fairly shallow and nearly free of suspended sediment to admit enough light to support photosynthesis. The environment probably was much like that at the present time around the Florida Keys and the Bahama Islands. Most of the rock consists of very pure calcium carbonate precipitated by the cyanobacteria.

A distinctive form, Conophyton - so named because it resembles columns of stacked cones - is unique to this zone. The cyanobacteria generally grew as groups of nested cones and they occur near the middle of these stromatolite beds.

This Conophyton zone occurs in the same stratigraphic position in the Helena Formation throughout the Park. As such, it is a distinctive "marker bed" and is a great aid in mapping the geology of the Park.

Figure 5. Outcrop of Baicalia-Conophyton zone (structures formed by carbonate deposition by cyanobacteria, a type of blue-green algae) along the Highline Trail north of Logan Pass. Hammer indicates scale.

Akamina syncline

Both limbs of the Akamina syncline can be seen as soon as Heavens Peak comes into view in the distant mountain range to the northwest. The view is similar to that seen from Going-to-the-Sun Road (fig. 6). The eastward dipping beds in Heavens Peak (across

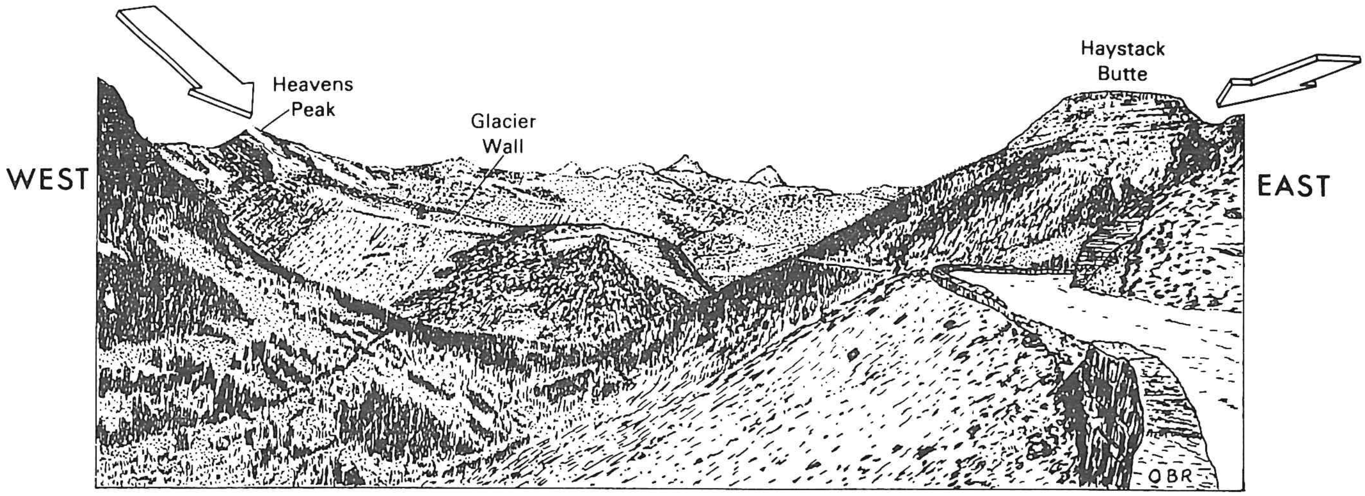


Figure 6. Sketch of the Akamina syncline as viewed toward the northwest from Going-to-the-Sun Road just below the Highline Trail. Arrows show dip angles of beds on either limb of the syncline.

the valley) are on the west limb of the syncline and the westward dipping beds in Haystack Butte (straight ahead along the trail) are part of the east limb. The northward plunging axis passes through the Glacier Wall.

This north-trending syncline is a prominent structural feature in most of the central part of the Park and extends into Canada where it derives its name from the Akamina River drainage just north of the border.

Sedimentation cycles

Sedimentation cycles, and their distinctive rock types, structures, and textures are common features in the Helena Formation. A good examples can been seen in the upper part of the Helena Formation along the Highline Trail in the first outcrop just past a lengthy covered area. This outcrop is next to a stream with a steep gradient that crosses the trail.

The rock textures in complete tripartite cycles, ranging in thickness from 0.5 to 3 m, have the following sequence:

UP Stromatolite zone

↑ Dolomite - molar-tooth structure

Terrigenous zone - clastic material

The cycles are bounded top and bottom by a sharp break in sedimentation. Each cycle begins with a layer of clastic material that can contain quartz grains, oolites, rip-up clasts, and other particles scoured in part from the top of the underlying cycle and indicate a high energy environment and current transport.

The middle part of the cycle contains fine-grained dolomite that usually displays molar-tooth structure. The fine-grained dolomite is interpreted as the product of chemical precipitation in very quiet-water.

The top of the cycle, if present, is generally a layer of stromatolites, stromatolitic detritus, or cryptalgal laminates. This indicates gentle currents carrying very little clastic material derived from outside the site of deposition.

Molar-tooth structures

An unusual structural feature of many dolomite beds in the Helena Formation is called molar-tooth structure. The structure results from many small vertical and horizontal segregations as well as various other shapes composed of microcrystalline calcite. The calcite appears white to light gray in a darker matrix of dolomite. On weathered surfaces the calcite-filled features dissolve back from the face of the rock giving the surface strong relief (fig. 7).

Molar-tooth structures were defined by Bauerman (1885) as vertical to subhorizontal, wrinkled segregations of massive calcite that weather more readily than the host rock. On weathered surfaces they resemble the molar teeth of elephants (Eby, 1977).

In the more than 100 years of study of molar-tooth structures in these rocks, researchers have proposed several theories of origin:

1) They are late-stage structures due to jointing and cleavage planes in the host rocks, with subsequent precipitation of sparry calcite.

2) The structures were the result of some sort of growth pattern of organic material, most likely algal or cryptalgal in origin, in which the different morphologies were partly to largely controlled by wave and current energy, water depth, sediment type and rate of

deposition. The organic materials were later replaced by calcite.

3) The structures resulted from shrinkage of the sediments, possibly by subaerial desiccation, subaqueous syneresis processes, or from voids associated with the decomposition of organic mats; they could also have formed by early post-depositional compaction. The cracks were later filled with a uniform, granular calcite cement.

4) The molar-tooth structures were formed initially as pure evaporite mineral precipitates, both void-filling and displacive within the sediments, underwent rapid and complete replacement by calcite, and finally were deformed during compaction.

Much of the information about molar-tooth structure, given above, is from an excellent summary and discussion in the Ph.D. dissertation by Eby (1977).

Recent work by Furniss (1990), a graduate student of Don Winston at the University of Montana, has yielded new ideas on the origin of molar-tooth structure. Experiments that closely replicate molar-tooth suggest that molar-tooth formed as gas-generated voids that filled with blocky calcite. Experimental models using mud, yeast and sugar in glass aquaria produced bubbles and gas expansion cracks which closely mimic shapes of molar-tooth structures. Mixing $CaCl_2$ and Na_2CO_3 solutions precipitated finely crystalline blocky calcite, similar to molar-tooth calcite. Comparisons of the Belt with the modern Dead Sea, where bacterial decomposition of gypsum is inferred to produce H_2S and CO_2 gases suggests a similar origin for molar-tooth structures and calcite.

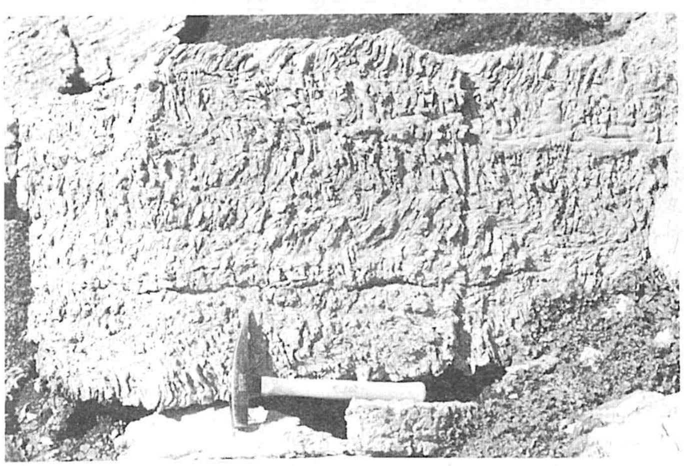


Figure 7. Molar-tooth structure in the Helena Formation along the Highline Trail.

Diorite dike

North of Haystack Butte and about 0.75 mi (1.2 km) south of the Granite Park Chalet, a steep path leads up from the Highline Trail to a prominent saddle on the Garden Wall called the Grinnell Glacier Overlook. Exposed at the overlook, and extending down the slope toward the Highline Trail, a dark, nearly vertical diorite dike cuts across the sedimentary strata of the Helena Formation (fig. 8). The dike is 90.5 ft (27.6 m) thick here and has a contact aureole up to 6.6 ft (2 m) wide on either side. Although vegetation and talus obscure much of the outcrop on the west side of the Garden Wall, excellent exposures exist at the overlook, and in a couloir on the east side directly below the overlook.

The dike, which strikes N 75° W, also crops out in Snowslip Formation strata about 1,970 ft (600 m) west of the Chalet. It is visible on air photographs of the southern end of Flattop Mountain (fig. 11). The most accessible outcrop of the dike is along the Loop Trail about 1,640 ft (500 m) below the Chalet. Walk 70 ft (20 m) northwest along strike from where it crosses the trail.

In mineral and chemical composition the dike is nearly identical to the diorite sill in the Helena Formation. Except for the quenched borders, the rock is medium grained and consists primarily of plagioclase feldspar and clinopyroxene, with minor quartz, hornblende, and iron and titanium oxides. Low-grade contact metamorphic alteration products include pervasive chlorite, epidote, and iron sulfides. In exposures at the overlook, numerous inclusions of Helena Formation country rock occur within the chilled margins of the dike.

Snowslip Formation

The Snowslip Formation overlies the Helena Formation. It varies in thickness from 1,606 ft (489.5 m), at the type section on Snowslip Mountain in the southern part of the Park (Childers, 1963, p. 146), to 1,172 ft (357.2 m) at Hole-in-the-Wall, near the International Boundary. At Hole-in-the-Wall and as far south and west as Granite Park and the Apgar Mountains (fig. 11), it incorporates the Purcell Lava in its upper part, so thicknesses given for the Snowslip in the northern part of Glacier National Park include the Purcell Lava (Whipple and Johnson, 1988). Where the Highline Trail crosses it, the formation is approximately 1,200 ft (365.8 m) thick.

In Glacier National Park, the Snowslip Formation is divided into six informal members, labelled 1 through 6 from bottom to top (Whipple and Johnson, 1988). All six members correlate throughout the Park.

Member 1 rests with apparent unconformity on the Helena Formation. It contains alternating pale-maroon and grayish green successions of calcareous siltite, argillite, and oolitic arenite. It ranges in thickness from 80 to 300 ft (24.4 to 91.4 m). Arenite beds are thin, fine- to coarse-grained, moderatly to poorly sorted, and commonly cross-laminated. Siltite and argillite laminae are commonly arranged as wavy, nonparallel, fining-upward couplets that contain abundant mudchip intraclasts, subaqueous shrinkage cracks, and fluid escape structures.

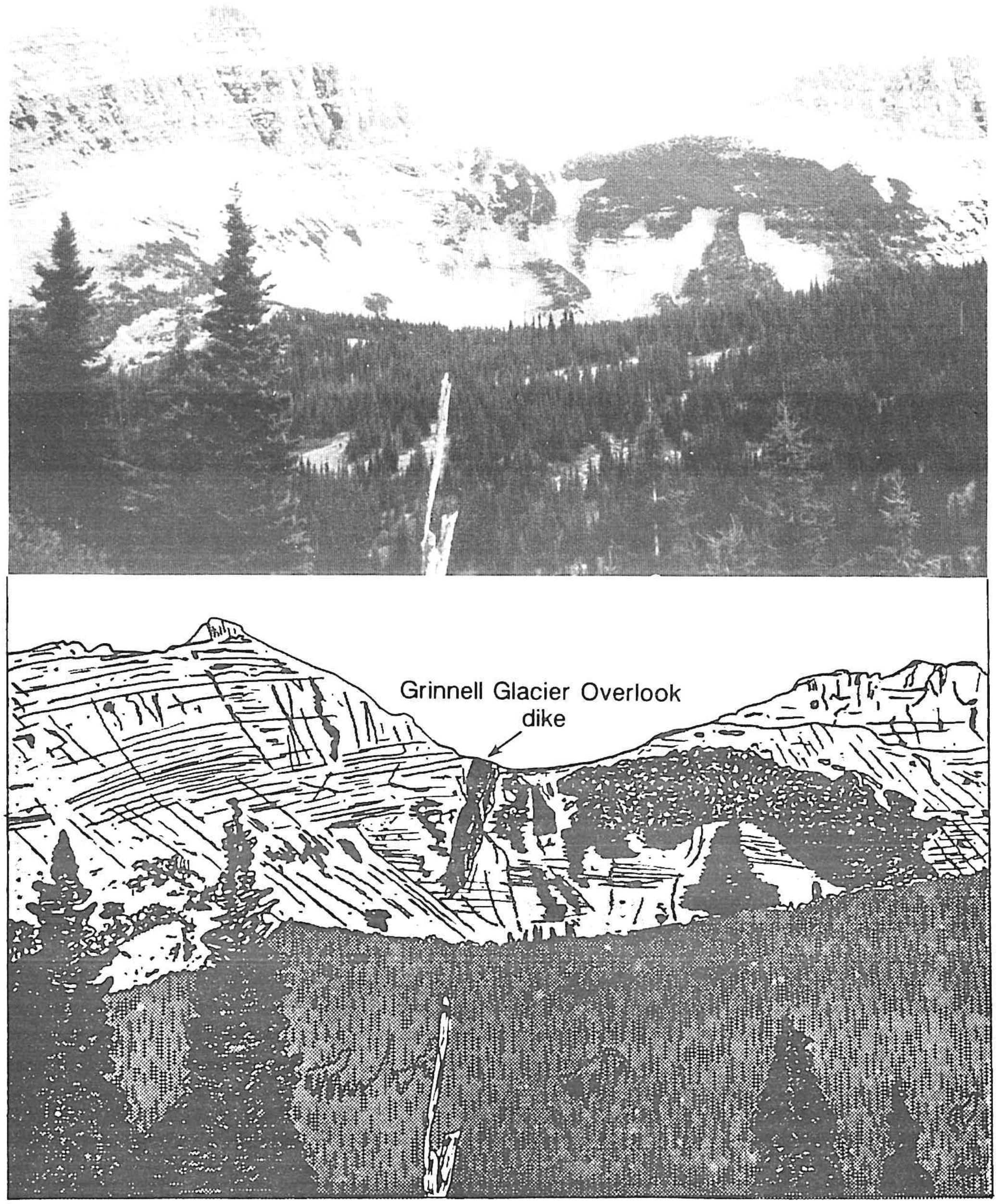


Figure 8. Photograph and sketch of diorite dike viewed along strike toward the southeast from southwest of Granite Park Chalet.

Member 2 is about 400 ft (121.9 m) thick. It consists mostly of wavy, nonparallel-laminated, grayish-green and yellowish-gray calcareous siltite and argillite. It has few interbeds of arenite. Several thin beds of pink stromatolitic limestone (fig. 9), are conspicuous in the lower part.

Member 3 is about 200 ft (61 m) thick. It consists of rhythmic, fining-upward successions as much as 10 ft (3 m) thick. They consist of white to pink quartz and subfeldspathic arenite at the base, and grade upward to dark-red argillite at the top. At the type locality, these rhythmic successions show more complete grading and are thicker than elsewhere in the Park. Sedimentary structures include abundant ripple marks, desiccation cracks, mud-chip breccias, and fluid-escape structures.

Member 4 is about 425 ft (129.5 m) thick. It closely resembles the grayish-green strata of member 2, and also has beds of pink stromatolitic limestone in the lower part.

Member 5 is similar to member 3 but generally thicker. It ranges in thickness from 115 to 450 ft (35 to 137.2 m). It shows better developed and more complete grading in fining-upward successions than does member 3. The base of each succession is erosional, in sharp contact with red argillite at the top of the underlying succession.

The uppermost member 6, is only about 26 ft (8 m) thick in the type section. It consists of interbedded grayish-green and pale-maroon arenite, siltite, and minor argillite. In the northern part of Glacier National Park, member 6 conformably encloses the Purcell Lava. It consists largely of grayish-green siltite and argillite beneath the lava, and alternating beds of pale-maroon and grayish-green arenite, siltite, and argillite above. Where the Purcell Lava is present, thin, discontinuous beds of pink and gray stromatolitic limestone occur in the lower part of member 6.

Outcrops along the Highline Trail, particularly north of the trail junction with Grinnell Glacier overlook and toward the Chalet, expose parts of members 1 through 5. Member 6 is exposed adjacent to the chalet, along with the Purcell Lava. Thin, fining-upward cycles of member 5 crop out locally in low ledges next to the trail just below the chalet. Note the presence of thin white quartz arenite beds that form the base of fining-upward cycles. The clean, well-sorted quartz grains are thought to have originated from strandline sands to the northeast.

The Snowslip Formation is interpreted to have been deposited in a paralic environment. Members 2, 4, and 6 are thought to represent a shallow subtidal environment, which for the most part was reducing. Red bed members 3 and 5 represent fluvial sedimentation probably across the distal apron of an alluvial plain. Interpretation of sedimentary structures of the mixed siliciclastic-carbonate member 1 suggests sedimentation on intertidal to shallow subtidal flats (Whipple and Johnson, 1988).

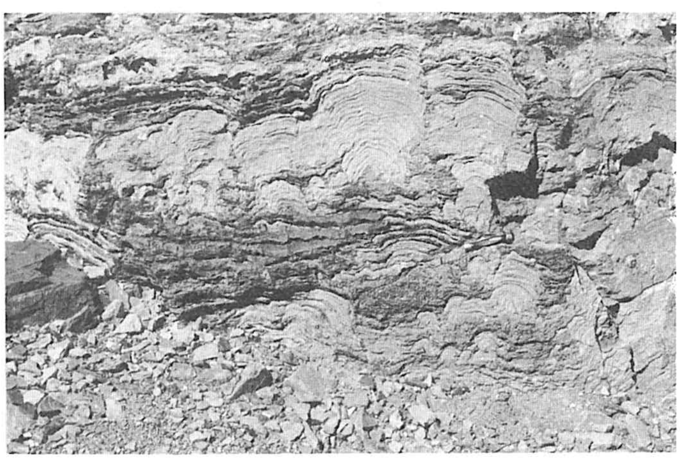
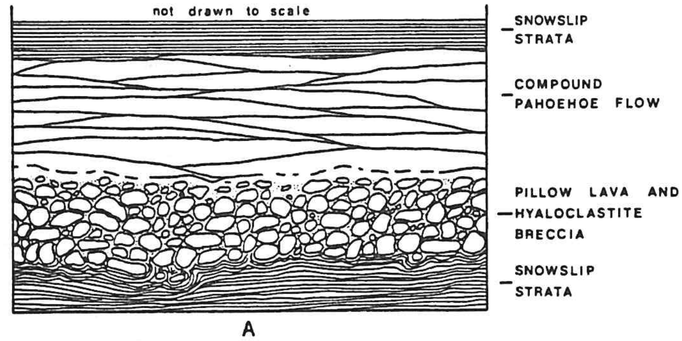


Figure 9. Stromatolitic limestone in member 2 of the Snowslip Formation. Knife at right center is 7.5 cm long.

Purcell Lava

The Purcell Lava is a sequence of amygdaloidal mafic lava flows that forms an important marker in the lower Missoula Group of the Belt Supergroup in both the United States and Canada. In Glacier National Park, it is within member six of the Snowslip Formation. In southeastern British Columbia,



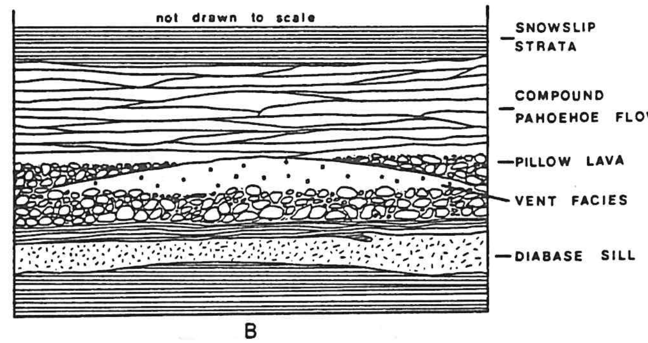


Figure 10. Subaqueous and subaerial phases of the Purcell Lava. A. Schematic diagram depicting a typical section of Purcell Lava in Glacier National Park. The lowermost pillowed section is 9-15 m thick and overlain by massive, intertongued pahoehoe flow units collectively up to 54 m thick. B. Section of Purcell Lava at Hole-in-the-Wall near the Canadian border (fig. 11). Interval includes a zone of 18-m-thick chaotic blocks, both cognate and accidental, that is interpreted to be a near-vent deposit. The entire sequence, including the subaerial flow units, is 77 m thick.

the flows are the Nichol Creek Formation and the Snowslip equivalent is the Van Creek Formation (McMechan and others, 1980; Hoy, 1985).

The flow sequence comprises both subaqueous and subaerial phases (fig. 10). It is well exposed in Glacier National Park, particularly at Granite Park, the Fifty Mountain area, and Hole-in-the-Wall (McGimsey, 1985)(fig. 11). The thickness in Glacier Park ranges from 253 ft (77 m) near the Canadian border to 50 ft (15 m) at the most southern locality in the Apgar Mountains (fig. 11). In general, the flow thickens to the north and west. The exposures in Glacier Park are near the margin of the once extensive volcanic field. Many of the features described can be seen in the area close to Granite Park Chalet.

The basal 30 to 50 ft (9 to 15 m) of the succession ubiquitously contains well-developed pillows where the lava flowed into a shallow marine environment. The total thickness of the pillowed section probably represents the water depth at the time of

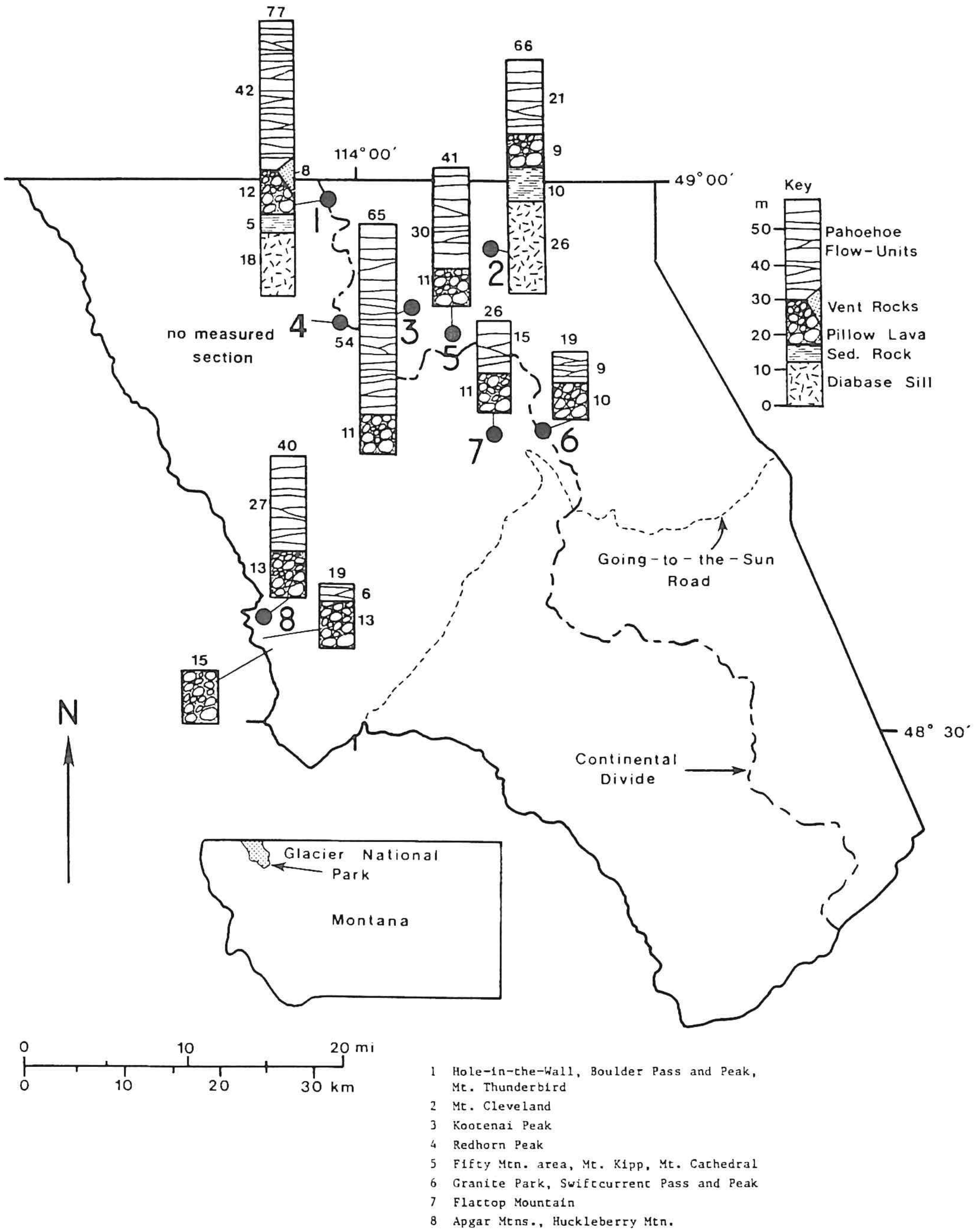


Figure 11. Lithology and thickness of Purcell Lava at various localities in Glacier National Park. Number on top of column is aggregate thickness.

emplacement. Above the pillowed section lies a series of compound subaerial pahoehoe flow units 0 to 177 ft (0 to 54 m) thick. Individual flow units range in thickness from about 4 inches to 20 ft (10 cm to 6 m), and typically cannot be traced laterally for more than a few hundred meters. Two minor facies---near-vent breccia and a hypabyssal diabase sill---occur only in northern exposures in the Park (McGimsey, 1985).

Primary flow structures and features are remarkably well preserved, however, the mineralogy and chemical composition of the Purcell Lava has been extensively altered by diagenesis related to burial. The original composition was that of continental alkaline basalt (McGimsey, 1985).

The age of the Purcell has not been directly determined, primarily because we have been unsuccessful in finding suitable unaltered minerals for dating. However, Hunt (1962) dated the hornfelsed strata beneath the lowest flow at 1075 Ma by the K-Ar method using biotite. Thus, the Purcell Lava may be part of the widespread 1100 Ma anorogenic magmatic event that affected much of the North American craton (Van Schmus and Hinze, 1985; Larson and others, (in press). Alternatively, Link and others (1993) give an age range for the Belt Supergroup, which includes the

Figure 12. A near dip-slope exposure (foreground) of Purcell Lava in a view to the southeast toward Granite Park Chalet and the Garden Wall.

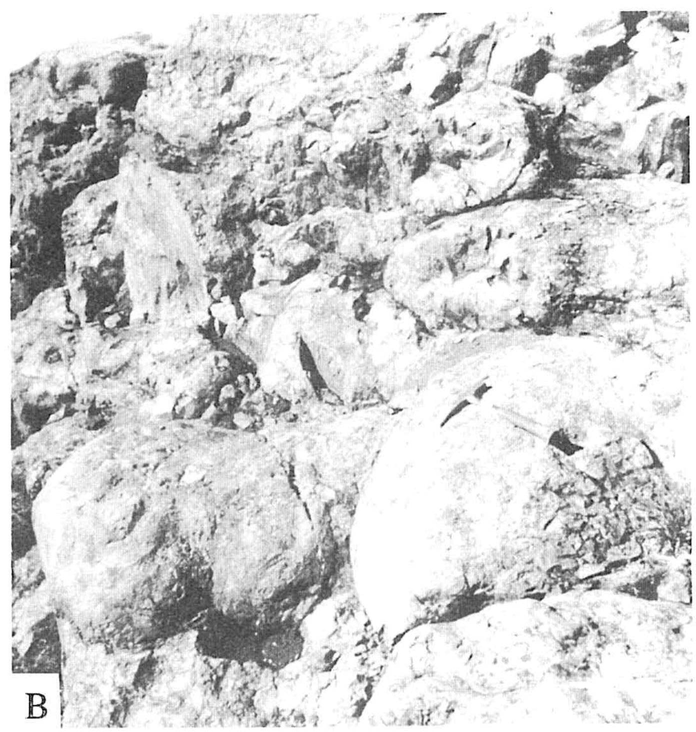


Figure 13. Two examples of subaqueous pillow structures in the Purcell Lava at Granite Park. A. Pillows and interpillow hyaloclastite breccia at Granite Park. B. Bulbous, interconnected pillows near Boulder Pass in the northern part of the Park (fig. 11).

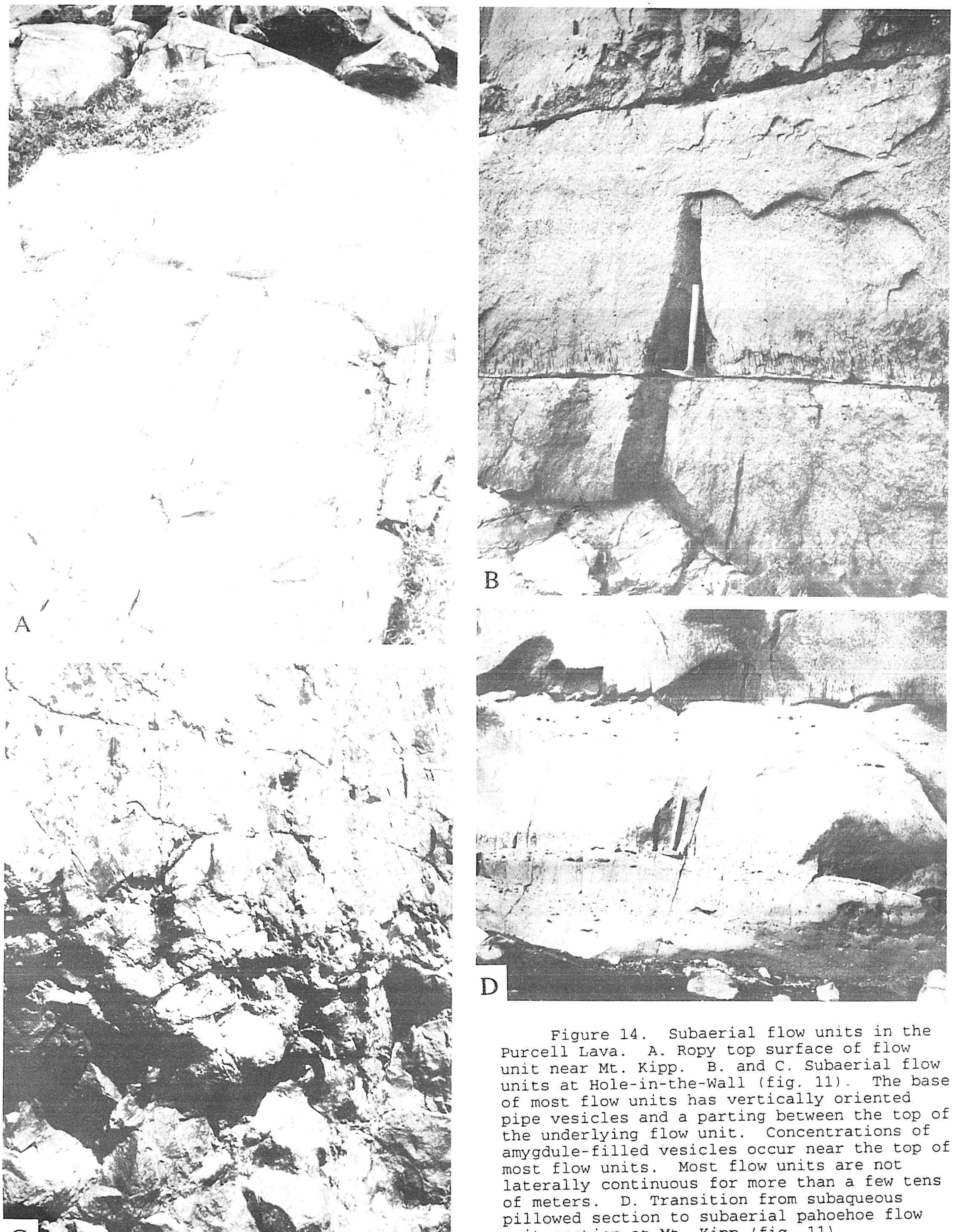


Figure 14. Subaerial flow units in the Purcell Lava. A. Ropy top surface of flow unit near Mt. Kipp. B. and C. Subaerial flow units at Hole-in-the-Wall (fig. 11). The base of most flow units has vertically oriented pipe vesicles and a parting between the top of the underlying flow unit. Concentrations of amygdule-filled vesicles occur near the top of most flow units. Most flow units are not laterally continuous for more than a few tens of meters. D. Transition from subaqueous pillowed section to subaerial pahoehoe flow unit section at Mt. Kipp (fig. 11).

Purcell Lava, of 1,450 to 1.250 Ma.

The Purcell Lava crops out primarily along the limbs of the Akamina syncline in the northern half of Glacier Park. One of the best outcrops is adjacent to, and underlies the Granite Park Chalet (fig. 12). Here, the succession is 62 ft (19 m) thick and crops out along a series of stair-stepped, sparsely vegetated, dip slopes that provide excellent cross sectional and three dimensional exposures. The lava flow sequence dips more steeply than the slope of the ground at Granite Park. Thus, a good way to examine the section is to walk on the trail from the Chalet toward Swiftcurrent Pass to a cliff exposure of pillow lava on the north side of the trail. Then, take a winding traverse down the dipslope back toward, and north of the Chalet, working your way through the pillow lava section and then up into the pahoehoe flow units, which are exposed for a hundred meters below the Chalet.

Many features and structures common to modern subaqueous and subaerial basalt flows are preserved at Granite Park. Soft-sediment deformation occurs locally at the base of the section where the first pillows emplaced bulldozed sediments on the sea floor. The pillows are up to 3.3 ft (1 m) in diameter, are typically elongate and interconnected, and have concentrically oriented but poorly developed vesicles. Although some pillows are tightly packed and conformed to surrounding pillows, interpillow hyaloclastite breccia is common, especially in the upper part of the section (fig. 13). Broken, partially drained pillows, some with diagenetic intergrowths of large dog-tooth quartz, occur in some exposures. The pillowed section at Granite Park is 33 ft (10 m) thick.

Above the pillowed section is a series of subaerial pahoehoe flow units, individually ranging in thickness from 5 inches to 5.6 ft (13 cm to 1.7 m); as many as eight occur in several of the exposures at Granite Park. Most of the flow units have ropy upper surfaces and a zone of filled vesicles near the top. The basal contacts form distinct partings, commonly accentuated by vertical pipe vesicles that extend up from the contact (fig. 14).

Relationship of the Purcell Lava, Diorite dike, and Diorite sill

The Purcell Lava and the diorite sill in the Helena Formation have been inferred to be related (Daly, 1912; Ross, 1959). However, field relations and geochemical differences indicate that the lava flows are neither coeval nor cogenetic with the diorite sill (McGimsey, 1985). The dike exposed at Grinnell Glacier Overlook extends across Granite Park and cuts the Purcell Lava and overlying Snowslip strata. Furthermore, the dike apparently is fed from the sill.

The relationship of the dike to the diorite sill, which are about the same thickness, is unclear on the west side of the overlook because of extensive cover and lack of good outcrops. However, on the east side, the dike and sill are well exposed on the headwall of the Grinnell Glacier cirque. The sill can be traced along the east side of Mt. Gould and the dike extends down from the overlook to the sill where a perennial snowfield obscures the contact. The dike is not present in strata below the snowfield at the base of the cirque, or in the bedrock exposed in front of Grinnell Glacier. The sill thins dramatically north of the projected intersection with the dike and skips erratically upsection for several hundred meters on the south flank of the west ridge of Mt. Grinnell. From the east side of Swiftcurrent Pass (fig. 1), Finlay (1902) discovered exposures of the sill abruptly cutting upsection for 100 m as a dike on the north side of the same ridge. The dike apparently rises from the sill (i. e. the sill becomes a dike as it cuts upsection), intruding the overlying strata including the Purcell Lava. Thus, intrusion of the sill postdates emplacement of the lava flows.

Trace element compositions of the sill and dike are virtually identical but significantly different from that of the Purcell Lava, which further supports cogenesis of the sill and dike (McGimsey, 1985).

The name "Granite Park" is a misnomer since no granite is there. The name apparently originated from early prospectors who, upon discovering nearby occurrences of coarse-grained granophyre from the thick diorite sill that intrudes the Helena Formation, called the bowl-shaped area immediately south of the Chalet, Granite Park (Holterman, 1985).

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Going-to-the-Sun Mountain, with Sexton Glacier below. *View is to the west from near Siyeh Pass on Baring Creek Trail, Glacier National Park.*

THE GRINNELL, EMPIRE AND HELENA FORMATIONS ALONG BARING CREEK AND AT SIYEH PASS, GLACIER NATIONAL PARK

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INTRODUCTION

This field trip consists of a full day's hike up the Baring Creek trail in Glacier National Park, starting from Sunrift Gorge (elev. 4600 ft, 1400 m) on the east side of Logan Pass on Going-to-the-Sun Road. The trail affords access to excellent exposures of the upper part of the Middle Proterozoic Grinnell Formation, and especially the transition to the overlying Empire Formation. The entire Empire Formation is exposed, as is the gradational contact to the Helena Formation above. The field trip log continues up to Siyeh Pass and down the trail to Siyeh Bend (elev. 5900 ft, 1800 m), on Going-to-the-Sun Road 4.5 miles (7.2 km) west of the Sunrift Gorge trailhead. The field trip is best run in an uphill direction because the exposures of the Belt Supergroup are better on the Baring Creek side than on the Siyeh Bend side.

Stratigraphic Setting

The Baring Creek section includes the upper Ravalli Group (Grinnell and Empire formations) and the lower part of the middle Belt carbonate (Helena Formation) (Fig. 1). The term Piegan Group was used for the Helena and Empire formations by Smith and Barnes (1966) and Horodyski (1983) (Fig. 1, this paper).

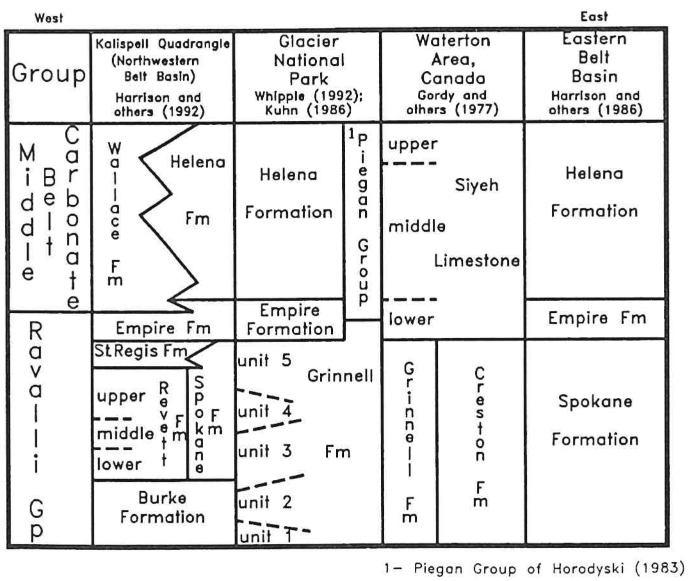

Figure 1. Stratigraphic correlation chart for the Ravalli Group and middle Belt carbonate in Glacier Park and vicinity (see also Harrison and others, 1993, this volume).

The Baring Creek section represents the second major subaerial clastic pulse in the Belt Supergroup (the Neihart Formation contains the first) and the succeeding transition to subaqueous carbonate deposition. The Grinnell Formation is a subaerial red bed succession above the deep-water lower Belt Appekunny Formation and below the second major transgression of the Belt "sea" or "lake" represented by the Piegan Group.

The fully-exposed transition from Grinnell to Empire formations along Baring Creek affords observations of

1) transgressive shoreline deposits containing flood-transported coarse sand grains reworked in standing water; and
2) interfingering between strata deposited on distal portions of southwest- and east-derived sandy alluvial aprons. These aprons represented two distinct source areas on the margins of the intracontinental Belt basin.

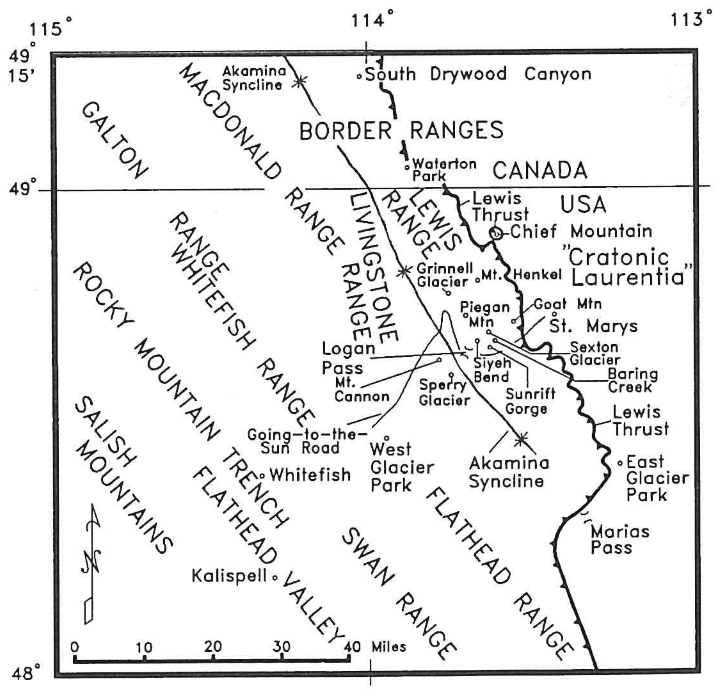

Figure 2. Location map of the Baring Creek trail and localities mentioned in text in Glacier Park and vicinity.

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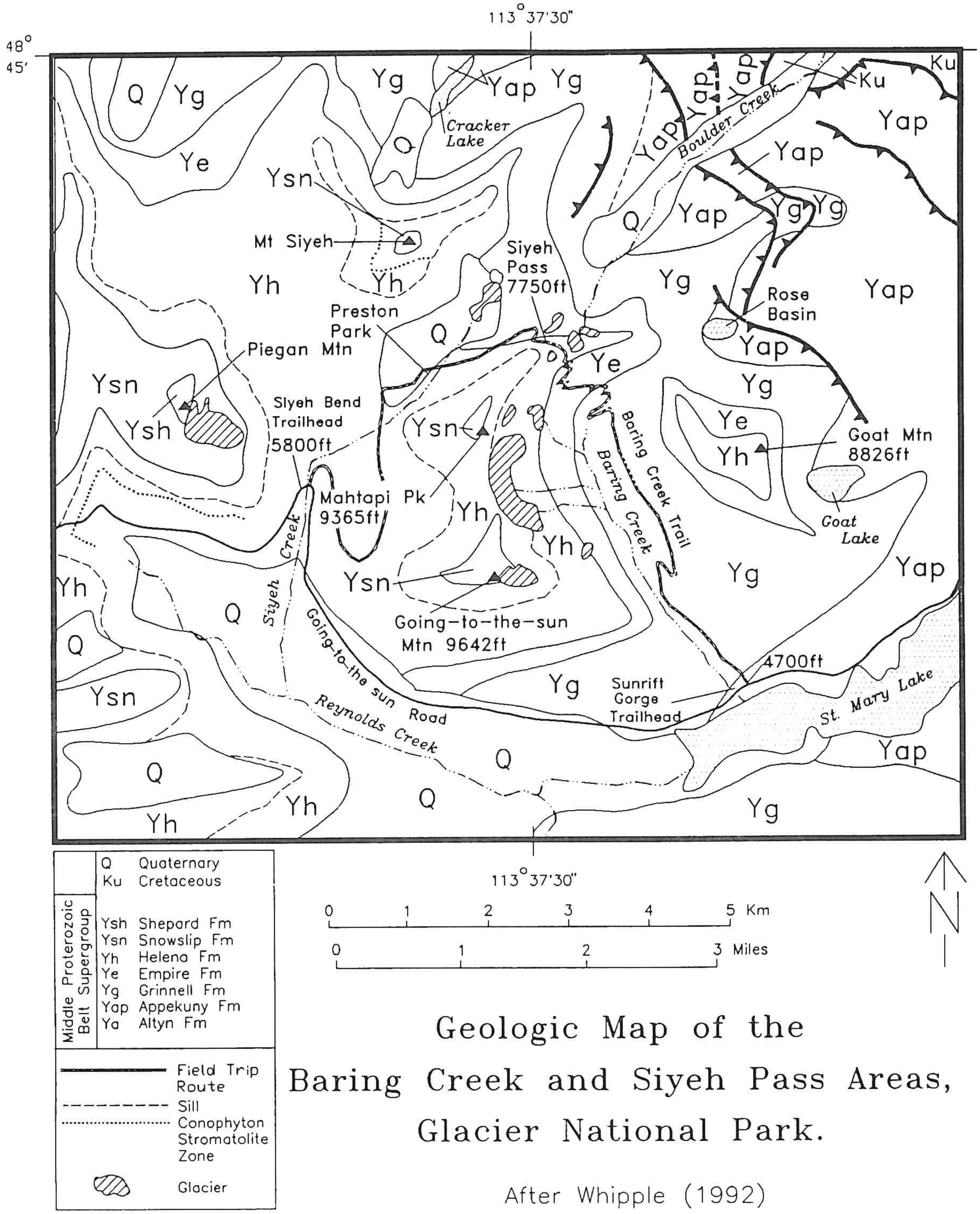

Figure 3. Geologic map of the Going to the Sun Highway (after Whipple, 1992).

Structural Setting

The strata exposed along Baring Creek (Fig. 2, 3) are on the gently west-dipping east limb of the Akamina Syncline that trends about N40°W through Glacier Park and into Canada (Fermor and Price, 1984; Whipple, 1992). The area is about 5 mi west of and 3,000 to 4,000 ft (~900--1,200 m) above the Lewis thrust system (Yin and others, 1989; Hudec and Davis, 1989) (Fig. 3). The Baring Creek area contains several beautifully-exposed small west-vergent kink folds and reverse faults, antithetic to the Lewis thrust (Fig. 4, 5). These antithetic folds are probably related to pre-Lewis thrust movement of the "frontal zone" of the Lewis system (Yin, 1991).

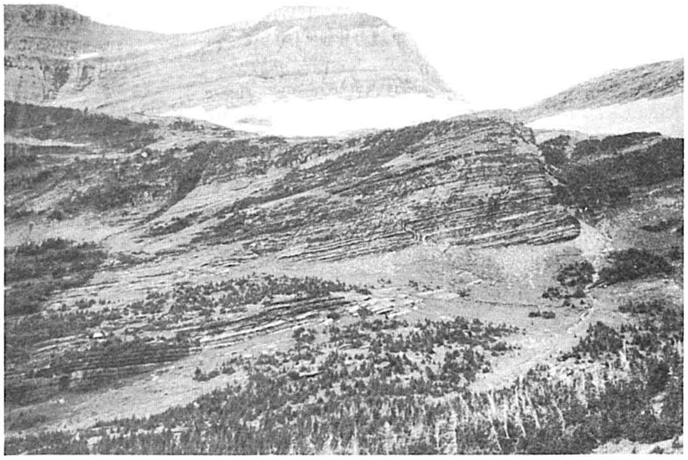

Figure 4. Distant view of head of Baring Creek looking northwest, showing stratigraphic section to be examined on this field trip. Formations are labelled. Yg= Grinnell Formation (1,2,3,4,5, are informal members of Kuhn (1986); Ye=Empire Formation; Yh=Helena Formation. West-vergent kink fold shown in Fig. 5 is in middle center of view.

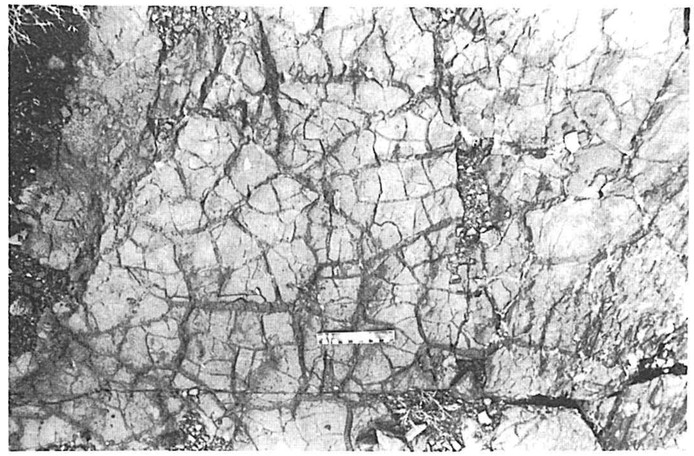

Figure 6. Mudcracked mud sediment type in Grinnell Formation, unit 3, along Baring Creek Trail, elevation 5,200 ft (1,590 m).

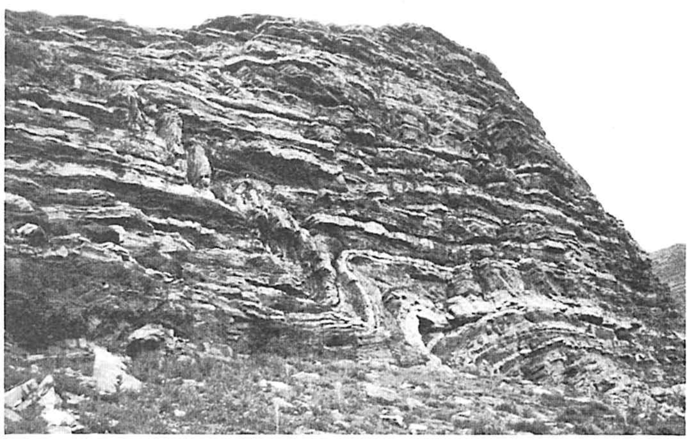

Figure 5. Kink fold in Grinnell Formation unit 5, near top of formation along Baring Creek trail, looking northwest, elevation 6,800 ft (2,070 m). Fold verges to the southwest (to the upper left of the photo), and is thus antithetic to the Lewis thrust system which is several thousand feet (over a thousand m) below.

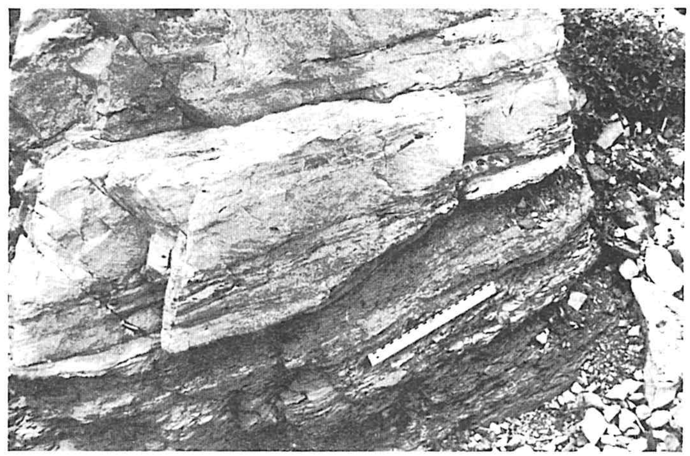

Figure 7. Grinnell Formation, unit 5, Baring Creek Trail, elevation 6,900 ft (2,100 m). Exposure shows coarse-sand and intraclast sediment type in multi-story trough cross-bedded co-set. Base of sand bed is erosional into underlying mudcracked even couplets of argillite and sand. This sand bed is interpreted to have been deposited in a shallow stream channel of a braided system on a sandflat, by successive floods carrying coarse-grained sand grains from the eastern cratonal source area.

FIELD TRIP LOG

Grinnell Formation

History of Nomenclature. The Grinnell Formation was named and described by Willis (1902, p. 316-321), with its type area near Grinnell Glacier in the northeastern part of what is now Glacier National Park (Fig. 2). Outside of Glacier National Park and the Whitefish Range, the Grinnell is mapped as the Spokane Formation, with its type area in the Helena Valley (Fig. 1). The name Spokane Formation was used for a time in the Park (Mudge, 1977; Mudge and Earhart, 1983; Earhart and others, 1984). Whipple (1992), following Connor and others (1984), reinstated the name Grinnell Formation in Glacier National Park and vicinity. General stratigraphic studies which describe the Grinnell include Fenton and Fenton (1937), Smith and Barnes (1966), Horodyski (1983), Connor and others (1984); Whipple and others (1984), Earhart and others (1984), Kuhn (1986), Whipple (1992), and Harrison and others (1992).

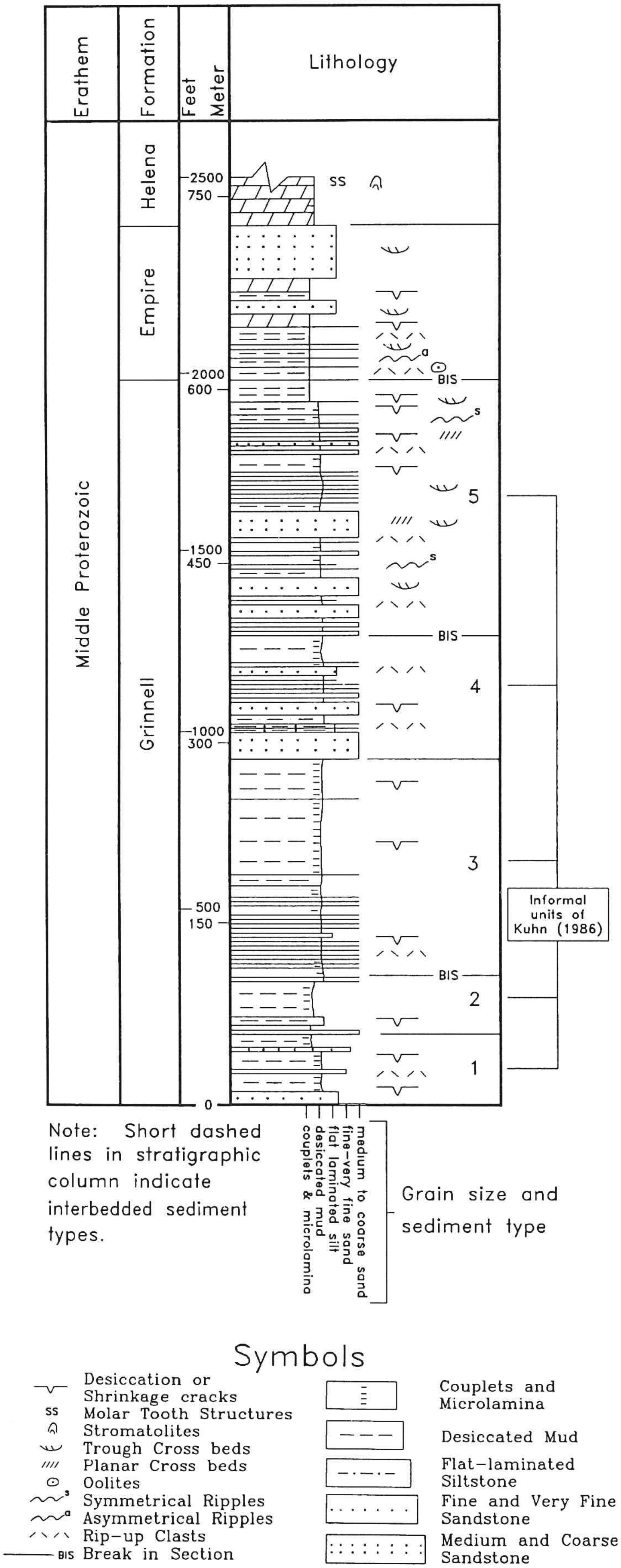

Figure 8. Stratigraphic column of Baring Creek area compiled from Kuhn (1986), Ross (1959), and observations by the author and students.

Description and Thickness. The Grinnell Formation (Figures 5,6,7) consists of bright red to purple (locally green) laminated argillite and blocky siltite, with up to 60 percent white to brown, medium- to coarse-grained quartz arenite and quartz conglomerate. The Grinnell is 1,740-2,590 ft (530 to 790 m) thick in Glacier Park (Whipple, 1992) (Fig. 8). It thickens to over 4,000 ft (1,219 m) to the west in the Whitefish Range, and also to 4,000-5,000 ft (1,200-1,500 m) to the southwest in the Swan Range (Ross, 1959; Kuhn, 1986). The Grinnell thins drastically to the north in Waterton Park, where the equivalent Creston Formation is only 984 ft (300 m) thick (McMechan, 1981; Koopman, 1985). In Glacier National Park, the Grinnell Formation overlies the Appekunny Formation (informal member 5), with the contact placed where green argillite of the Appekunny changes to red argillite and siltite of the Grinnell (Whipple, 1992).

The conglomerate and quartz arenite portion of the Grinnell Formation is thicker and coarser-grained in the eastern part of Glacier National Park and near the top of the formation (unit 5 of Kuhn, 1986) (Whipple, 1992). In the southeasternmost part of the park, the upper Grinnell is nearly 100 percent quartz arenite and conglomerate.

Sediment Types. Strata of the Belt Supergroup, typically cyclic and fine-grained, have proven difficult to describe in meaningful sedimentological terms. Don Winston and his co-workers have developed a set of sediment types by which Belt stratigraphic units may be described and compared (Winston, 1986a; 1986b; 1989; 1991; Winston and Link, 1993; Winston and Lyons, 1993, this volume). Sediment types embody grain size, inferred original mineralogy, and details of sedimentary structure. In this paper the sediment types are used in descriptive fashion.

Sediment types present in the Grinnell Formation include microlamina, flat-laminated sand, mudcracked mud, even couple, mudcracked even and lenticular couplet, and coarse sand and intraclast (Winston, 1991; Winston and Lyons, 1993, this volume; Winston and Link, 1993). Mudrocks in the Grinnell Formation contain both subaerial desiccation cracks and fluid-escape structures (subaqueous shrinkage cracks or "syneresis" cracks). Sandstones generally are medium- to thin-bedded, crossbedded, ripple marked, and contain rip-up clasts and armored mud balls (up to 10 cm in diameter) (Fig. 7). Bi-polar cross-beds occur locally. Some sandstone beds have casts of mud cracks on their undersides, while others have scoured and loaded bases. Sands of the Grinnell contain up to 10% feldspar and alteration products (Kuhn, 1986).

Stratigraphic divisions. Whipple (1992) divided the Grinnell Formation in the northwestern part of Glacier National Park into two members (Fig. 8), a lower unit (365 m, 1,200 ft) of predominantly interbedded blocky siltite and evenly laminated argillite, and an upper unit (425 m, 1,390 ft), containing beds of white quartz arenite (up to 20% of the section) plus siltite and argillite similar to the lower unit. Whipple (1992) did not divide the formation in other parts of the park.

Detailed stratigraphic study by Kuhn (1986) of the Grinnell Formation in three stratigraphic sections (Mt. Henkel, Goat Mountain, and Mt. Cannon, Fig. 2) in Glacier National Park and one in the Whitefish Range, suggests that the unit contains five stratigraphic units (Fig. 1; 8). Horodyski (1983) recognized the same five units, with a sixth unit in the lower part of what is now considered Empire Formation. Kuhn's lower four units make up the lower division of Whipple (1992) and Kuhn's unit 5 is the same as Whipple's upper division. Although the Grinnell regionally thickens to the west and southwest, units 2 and 5 thicken to the east and southeast within Glacier National Park, suggesting an eastern provenance.

Figure 8 shows a stratigraphic column for the Baring Creek area, with the Grinnell 599 m (1950 ft) thick (after Ross, 1959; Kuhn, 1986; and reconnaissance by the author and students).

Unit 1 (55.1 m, 181 ft at Goat Mountain, immediately east of Baring Creek) contains green and minor red argillite, with thin beds of coarse sandstone near the base of the unit (Fig. 6). Using the sediment type terminology, unit 1 contains mainly uncracked couplets and microlamina.

Unit 2 contains a thin (51.4 m, 169 ft at Goat Mountain) west-thinning wedge of medium- and coarse sandstone interbedded with siltite and argillite. The characteristic lithology is stacked beds of the cross bedded and rippled coarse sand and intraclast sediment type (Fig. 7) interbedded with mudcracked mud and mudcracked couplets (Fig. 6). The top of unit 2 is placed at the highest medium to coarse-grained sandstone bed. Unit 2 pinches out westward and is missing in the central western part of the Park at Mt. Cannon.

Unit 3 is 172.6 m (566 ft) at Goat Mountain, thinning to 61 m (200 ft) to the northwest in the Whitefish Range. It contains bright red argillite characterized by the mudcracked mud sediment type, and is generally a fine-grained unit containing couplets and microlamina. Sandstone beds are sparse.

Unit 4 is 99.1 m (325 ft) thick at Goat Mountain and contains mainly green siltite of the flat laminated silt sediment type, with occasional coarse sand and intraclast beds. Unit 4 passes to dolomitic silt to clay couplet sediment type to the north and east.

Unit 5 is 216.6 m (711 ft) at Mt. Cannon. Kuhn (1986) did not measure this part of the section at Goat Mountain. Unit 5 thickens to the east and north (Winston, 1991), and is 425 m (1395 ft) thick in the Whitefish Range to the northwest (Kuhn, 1986). It contains mainly white sandstone and red and green argillite and siltite (Fig. 7). The sandstones contain multi-story thin- and medium-beds of coarse sand and intraclast sediment type. The sandstone beds contain abundant tabular planar, tangential and trough cross bedded sets. Most ripples are symmetrical.

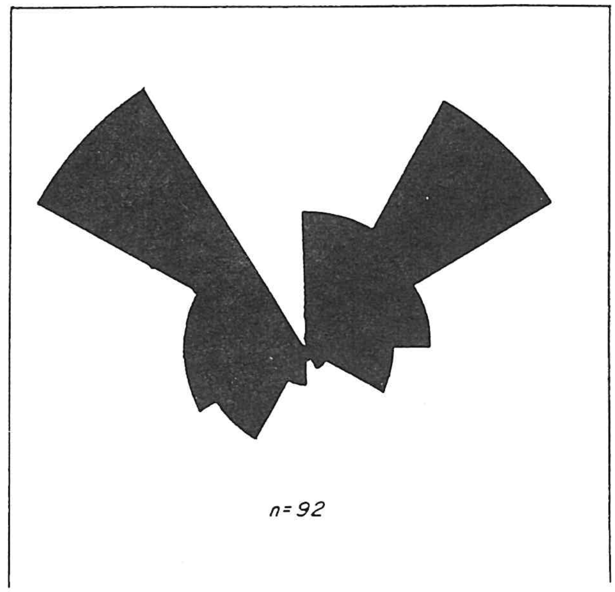

Figure 9. Paleocurrent rose from the base of Unit 5, Grinnell Formation, near Mt. Henkel from Kuhn (1986, Fig. 4-2). Data supports hypothesis of two source areas (southwestern and southeastern) for braided fluvial sandstones.

Interpretation of Depositional Environment and Provenance. Kuhn (1986, p. 3) concluded that "the Grinnell Formation reflects episodic sheet floods which carried medium- to coarse-grained sand across a distal alluvial sand and mud flat bordering a shallow, enclosed to restricted sea". Horodyski (1983) interpreted the Grinnell to have been deposited on an alluvial plain and in an adjacent (marine) body of water. The sand beds were interpreted as deposited by braided or anastomosing streams. Extensive mudcracking seems to indicate that diurnal tides did not exist (Kuhn, 1986), although McMechan (1981) interpreted the white, fine- to-coarse grained quartz arenite of the correlative Creston Formation in Canada to represent fill of migrating tidal channels.

Koopman (1985) noted that trough cross bedding in the upper Grinnell sands appears bidirectional, but the bedsets are only up to 20 cm thick. This limits the possible tidal range if the sands represent tidal channels, and was interpreted to support a braid-delta setting.

Kuhn (1986) and Cronin (1989) correlated sections of the Grinnell Formation from the Whitefish Range into the Spokane Formation in the Mission, Swan, and Flathead Range. Winston (1991) correlated the Grinnell and Spokane Formations of the eastern Belt basin with the entire Ravalli Group (Burke, Revett, and St. Regis Formations) in the Coeur d'Alene area.

These correlations confirm suggestions of others (Smith and Barnes, 1966; McMechan, 1981; Horodyski, 1983; Whipple and others, 1984; Earhart and others, 1984) that the Grinnell represents parts of two alluvial aprons, one derived from cratonic Laurentia to the southeast and one from the southwestern side of the Belt basin. These two alluvial aprons are thought to have interfingered along a desiccated and sometimes submerged mudflat.

Paleocurrent studies of Kuhn (1986) suggest that clastic wedges of coarse-grained white quartzose sand present in units 2 and 5 prograded northwestward from cratonic Laurentia (Fig. 9). These wedges may contain supermature quartz grains eroded from the pre-Belt(?) Neihart Quartzite to the east (Winston and Link, 1993).

These sand wedges interfinger westward and southward in Glacier National Park with flat-laminated siltstone (best represented in Grinnell unit 4) that represents the distal portions of a much larger southwest-derived alluvial apron represented by the Revett and St. Regis Formations (Kuhn, 1986; Cronin, 1989).

Grinnell deposition began in a subaqueous environment, but became largely subaerial in the middle part of the formation. Much of the formation represents mud flats periodically flooded by braided fluvial sand sheets. Generally the Grinnell records westward deepening of the basin. East-derived coarse sand is present near the base of unit 1 and in unit 2. Flat-laminated silt in the top of unit 2 may represent tongues of the southwestern (Burke Formation) sand source (Kuhn, 1986; Cronin, 1989). Unit 3 records extensively exposed desiccated mud flats. Unit 4 records sheetlike deposition of flat-laminated siltstone, with a western (Revett Formation) source (Winston and Link, 1993, Fig. 16). Unit 5 contains a southeastward-coarsening and thickening sand wedge deposited on alluvial flats.

<u>Field stops.</u> The road level at the Sunrift Gorge parking lot is near the base of the Grinnell Formation (Fig. 3). The narrow, fault- or joint-controlled canyon of Sunrift Gorge (Fig. 10) is a short distance off the trail. Unit 1 is exposed a hundred feet (30 m) to the east along the Going-to-the-Sun Road. In the roadcut immediately east of the Baring Creek trailhead, green mudcracked lenticular couplets are interbedded with the coarse sand and intraclast sediment type.

The climb through the trees into the canyon of Baring Creek does not afford much exposure, though there are a few bedding plane outcrops of the mudcracked mud sediment type of unit 3. These outcrops are especially good near the waterfalls along Baring Creek (elevation 5,200 ft, 1,585 m) (Fig. 11).

Above these waterfalls and about an hour from the trailhead, the trail cuts up to the right, switchbacking across a scrub-covered avalanche slope with little exposure (Fig. 3). The climb through the big switchbacks takes about an hour more, before continuous exposure of the upper part of unit 5 and the overlying Empire Formation begins (Fig. 3,4).

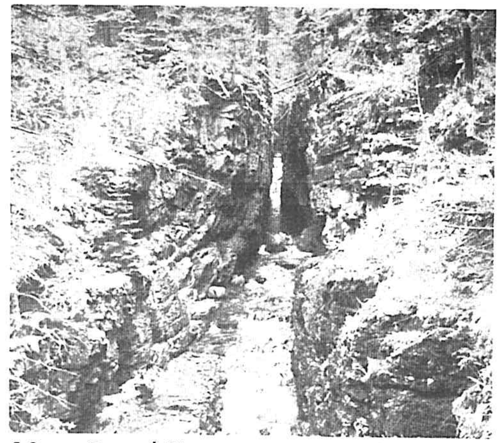

Figure 10. Sunrift Gorge, looking northwest from near the trailhead; straight path of Baring Creek follows fault or joint plane in the lower Grinnell Formation (unit 1 of Kuhn, 1986).

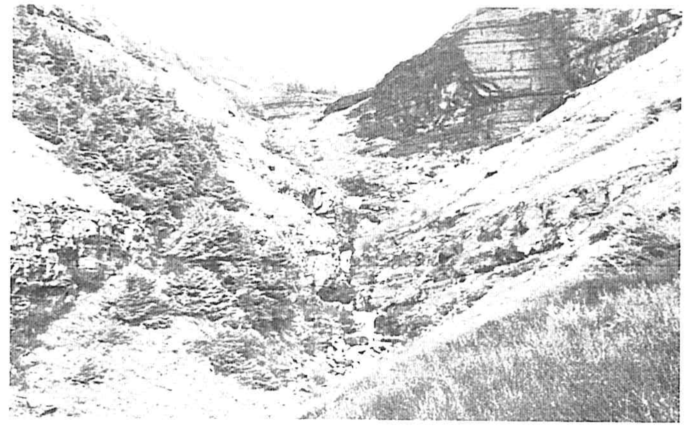

Figure 11. Lower reaches of Baring Creek, flowing through lower Grinnell Formation (unit 3); view looks northwest at elevation 5,000 ft (1,500 m).

The trail cuts across ledgy outcrops and makes several small switchbacks in the upper part of the Grinnell Formation.

One to two hours can be spent working slowly up this section. Unit 5 of the Grinnell Formation contains a sandstone and argillite package, interbedded on a scale of 15-30 ft (5 to 10 m) (Fig. 5, 7). The argillite contains maroon, laminated and mudcracked even and lenticular couplets. The sandstone beds are medium to coarse grained, white weathering, light-gray, submature quartz arenite and subarkosic arenite. Sedimentary structures include erosionally based trough cross-bedded sandstones 10 to 20 cm thick, overlain by parallel laminated fine-grained sandstone beds, with oscillation ripples on the tops of beds. Armored mudballs and abundant argillite rip-up clasts are found in the coarse sand beds. Generally, the sand beds get thinner and display less erosion at their bases toward the top of the section.

The top of the Grinnell is a thick maroon argillite bed followed along strike by a switchback of the trail for about 200 feet (60 m) vertically. Immediately above this argillite, crossed by the trail in two places, is the sulfide ooid sandstone marker bed at the base of the Empire Formation (Fig. 13).

Figure 12. Sexton Glacier in hanging valley at the head of Baring Creek with spectacular alluvial cone below; view looks northwest. Bedrock formations are labelled Yg=Grinnell Formation; Ye=Empire Formation, Yh=Helena Formation.

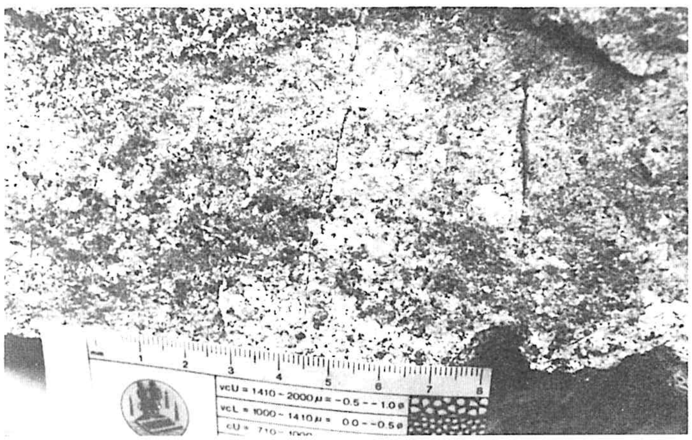

Figure 13. Basal sand bed of Empire Formation, containing medium- and coarse-grained quartz sand grains and ooids composed of pyrite.

Empire Formation

History of Nomenclature. The Empire Shale was defined by Walcott (1899, p. 207), from exposures north of Helena, Montana. The Empire Formation is the lower Siyeh Formation of Mudge (1977), Ross (1959) and previous workers (Fig. 1). The Siyeh usage was also followed by Koopman (1985) and Binda and others (1985; 1989; 1991). The upper Siyeh Formation of Mudge (1977) is now recognized as the Helena Formation (Whipple, 1992). Horodyski (1983) included the lower part of the Empire Formation as unit 6 of his Grinnell Argillite.

The Empire Formation contains mainly grayish green argillite, with subordinate maroon argillite, buff and green, locally dolomitic siltite, and light-gray quartz arenite (Figs. 13, 14, 15, 16). The formation is 400-518 ft (122 to 158 m) thick in Glacier National Park (Whipple, 1992). In South Drywood Canyon, Alberta, the Empire is 377 ft (115 m) thick (Koopman, 1985). The section shown in Figure 8 is generalized from Koopman's descriptions.

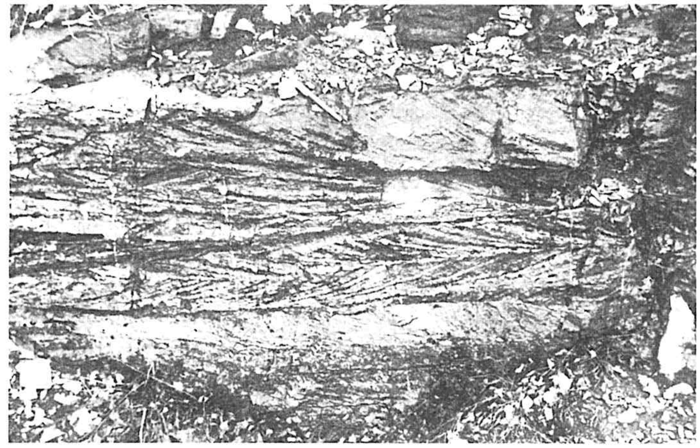

Figure 14. Trough-cross beds in Empire Formation, Baring Creek trail. Horodyski (1983, Fig. 9c) shows a similar photo from the upper Siyeh (Helena) Formation, and interprets it as herringbone cross stratification indicative of tidal currents. An alternate interpretation is washover sands on a beach in a shallow body of water. Note that the sand coset does not have an erosional base.

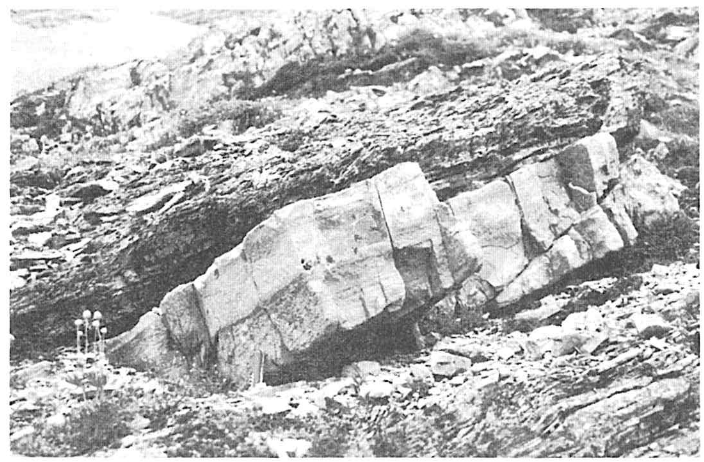

Figure 15. Sandstone bed coset in Empire Formation consisting of lower bed of planar cross-bedded medium-grained sandstone overlain by parallel laminated sandstone containing argillite rip-up clasts at its base. Coset is enclosed in green laminated argillite. Base of coset is not scoured.

In much of northern Glacier National Park, the base of the Empire is placed at a 20 to 50 cm orange-weathering medium-grained quartz arenite bed above red argillite of the Grinnell Formation (Whipple and others, 1984) (Fig. 8, 13). This arenite bed contains distinctive sulfide ooids, and has been described by Binda and others (1985; 1989; 1991). The ooids contain concentric layers of pyrite and chalcopyrite, and are thought to have grown during bacterially-mediated early diagenesis, with a source of metals in black argillaceous sediments above and below the sandstone bed.

The lower 60 feet (18 m) of the Empire Formation contains beds of white to buff quartz arenite that range in thickness from 5 inches to 11 ft (13 cm to 3.5 m) and contain minor carbonate cement and pyrite (Whipple and others, 1984; Whipple, 1992) (Fig. 14, 15). These are arranged in crudely fining-upward siliciclastic cycles (Winston and Lyons, 1993, this volume). Some arenite beds at the bases of cycles have well-developed cross-bedding, load structures and asymmetrical ripple marks (Fig. 16).

The upper two-thirds of the formation contains primarily olive-green and purplish-red beds of argillite one inch to 5 ft (a few cm to 1.5 m) thick. Thin interbeds of dolomite are present near the middle of the formation and increase in number and thickness upward. This upper part of the Empire contains mixed siliciclastic-to-carbonate-cycles 3 to 6 ft (1 to 2 m) thick, that start with a coarse sand bed (Fig. 15), and become finer grained and more carbonate-rich upward, to laminated argillite, and finally molar-tooth and intraclastic beds of dolomite.

<u>Sediment Types Present.</u> The base of the Empire Formation contains cross-bedded and oolitic coarse sand and intraclast sediment type interbedded with plane laminated silt and clay ("argillite" of most workers). The main part of the Empire Formation contains siliciclastic to carbonate cycles, containing uncracked cracked even couplet and microlamina associations at the base (Winston, 1991; Winston and Link, 1993). The upper half of each cycle contains dolomitic uncracked even couplets, microlamina, and carbonate mud sediment types.

<u>Interpretation of Depositional Environments.</u> Generally the Empire Formation represents a transition from the subaerial sand and mudflats of the Grinnell to the subaqueous carbonate-dominated system of the Helena. The first change recorded in the Empire is a deepening and shift from oxidizing to reducing conditions; the second is the waning of the supply of terrigenous clastic material, from both the eastern and the southwestern sources, and its replacement by carbonate mud.

In the Empire the sandstone beds are thicker than in the Grinnell, to 30 cm (8 in), and the bedforms have a longer wavelength, to 1.5 m (5 ft). Also, the bases of the sand beds are not scoured. These sandstones are interpreted by Koopman (1985) as shoreline or foreshore deposits.

The reduction in size to silt and clay of the siliciclastic supply and development of molar-tooth dolomite beds up to a meter (3 ft) thick marks the transitional contact between the Empire and the Helena Formation.

<u>Siliciclastic to Carbonate Cycles.</u> Siliciclastic- to carbonate cycles typical of the Helena Formation appear in the middle of the Empire and become thicker upward. These cycles are described and interpreted by Winston (1991; Winston and Lyons, 1993, this volume). The lower parts are thought to been deposited by episodic siliciclastic sheet floods across a playa mudflat into standing water of an expanding lake, 3 to 6 ft (1 to 2 m) deep. In the upper parts, mixed

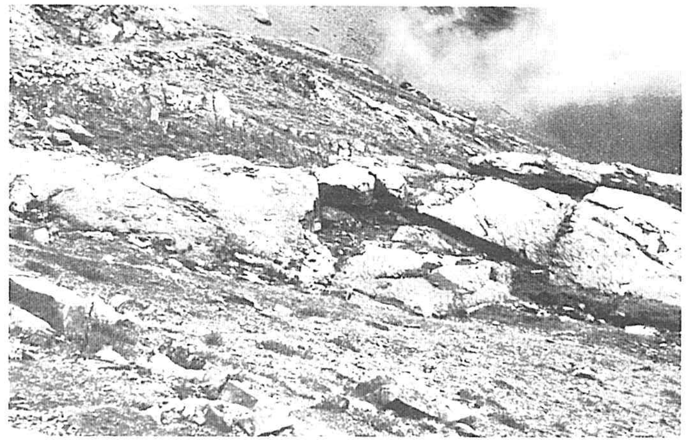

Figure 16. Bedding plane exposure in the clouds, Empire Formation, Baring Creek trail. Upper surface of white sandstone bed has been modified to undulatory asymmetrical ripples.

carbonate-siliciclastic sedimentation of a drying lake bed records shrinking of the lake and supersaturation of $CaCO_3$, culminating in deposition of pure carbonate mud. Fully developed carbonate-only cycles in the Helena Formation are developed to the west of Glacier National Park near Hungry Horse Dam (Winston and Lyons, 1993, this volume; Harrison and others, 1993, this volume). The sediment types above a basal coarse sand and intraclast bed change from pinch and swell couples with loaded bases, to lenticular couplets with rippled tops, to even mudcracked couplets, to molar-tooth beds.

<u>Molar-Tooth Structure.</u> Carbonate mud deposited in the Empire and Helena Formations underwent cementation and deformation by desiccation and water- and gas-escape to form "molar-tooth structure", one of the distinctive facets of Belt sedimentology (O'Connor, 1972). Molar-tooth structure has several genetic and descriptive subdivisions ("blobs, ribbons, and pods").

The process of forming molar-tooth structure may have been initiated by the formation of gas bubbles in carbonate mud. Gas may have been produced by sulfate-reducing bacteria which consumed sea (lake) water sulfate and possibly early gypsum deposits (Binda and others, 1991). The bubbles were filled by fine-grained blocky calcite producing the subdivision known as "molar-tooth blobs" (Winston and Lyons, 1993, this volume). Expansion cracks filled with escaping gas formed in stiffer, more dewatered mud. When cemented these become "molar-tooth ribbons". Dense gray "pods" probably were formed by early calcitic cement and subsequent replacement by spar. Any of these cemented cavities could be ripped up and redeposited as intraclasts during storms.

<u>Field Stops.</u> The Empire Formation is entirely exposed in the upper part of the switchbacks on the Baring Creek trail. The oolite bed (Fig. 13) parallels with the trail for nearly 300 ft (100 m). The upper contact with the Helena Formation is about 300 ft (100 m) west of the junction with the Sexton Glacier side trail.

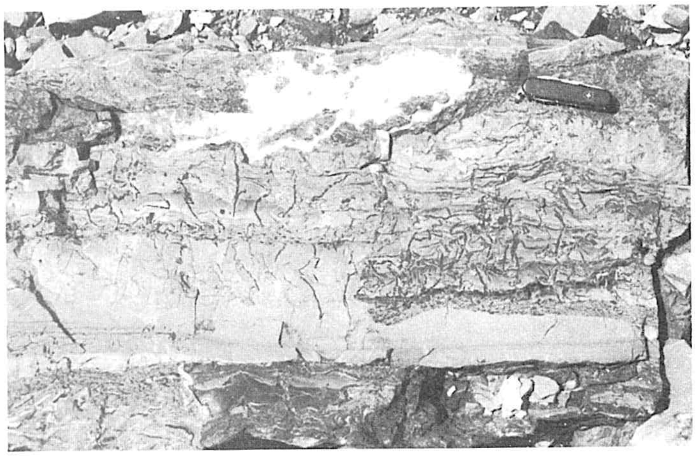

Figure 17. Molar-tooth structure, Helena Formation. The "ribbon" and "blob" variants are shown, as is a channel, itself filled with dolomitic mudstone containing a variety of molar-tooth structures.

Helena Formation

History of Nomenclature. The Helena Formation was named by Walcott (1899), for the area near Helena and correlated with the Siyeh Limestone of Glacier Park. Ross (1959; 1963) included the Helena in the Missoula Group. The original correlation of the type Helena with the Siyeh Limestone of the Piegan Group in Glacier National Park was again suggested by Smith and Barnes (1966) and adopted by Harrison (1972), Mudge (1977) and subsequent workers.

In Glacier National Park, the Helena Formation makes up the middle Belt carbonate. Formerly, the strata of the Empire and Helena Formations in Glacier Park were referred to as Siyeh Limestone (Ross, 1959; 1963) (Fig. 1). The name Siyeh was used by Koopman (1985) and Binda and others (1985; 1989) for the South Drywood Canyon area in Alberta, north of Waterton Park. The Helena Formation consists of the middle and upper Siyeh Limestones of Horodyski (1983).

The name Piegan Group was first proposed for Piegan Mountain in Glacier National Park by Fenton and Fenton (1937, p. 1890-1900), to include what are now recognized as Empire, Helena, Snowslip, and Shepard Formations, that is the carbonate-bearing beds between carbonate-free terrigenous argillite and sandstone above and below. After confused correlations summarized by Ross (1959; 1963), Smith and Barnes (1966) correctly established the lithostratigraphic entity of a middle Belt carbonate and applied the name Piegan Group to it. However, Mudge (1977) officially abandoned the name because of its blemished history. Horodyski (1983), however, retained the name Piegan Group for the upper Empire and Helena Formations in Glacier National Park, and Winston and Link (1993) predict that the Group will be resurrected and redefined, since the awkward and informal term "middle Belt carbonate" is generally used, at group level, for what could easily be called Piegan Group.

The Helena Formation in Glacier National Park is about 2,460-3,380 ft (750 to 1,030 m) thick and contains dolomite, limestone and minor quartz arenite (Whipple, 1992) (Fig. 8). The base of the Helena Formation is placed at the first bed of dolomite with molar-tooth structure above a 6 ft (1.8 m) interval of green argillite in the Empire Formation (Whipple and others, 1984).

The lower part of the Helena, 590 ft (180 m) thick, consists of interbedded quartz arenite and thick-bedded dolomite near the base, thin beds of horizontally laminated and molar-tooth dolomite (Fig. 17) in the middle, and thick smoky-gray limestone beds near the top (Whipple and others, 1984; Whipple, 1992). The middle part is 1,180 ft (360 m) thick and dominated by dolomitic molar-tooth beds, some as thick as 98 ft (30 m). A few thin beds of quartz arenite and stromatolitic limestone are present in the middle part. The upper part (235 m, 771 ft) consists of interbedded stromatolitic limestone, dolomite, oolitic limestone and quartz arenite. At the base of the upper part is a 100 ft (30 m) interval of stromatolitic limestone known as the Conophyton zone (Rezak, 1957), that is composed of Baicalia-Conophyton stromatolite cycles (Horodyski, 1983). Field stop 6 of Winston and Lyons (1993, this volume) examines these stromatolite cycles.

Tucker (1984) studied carbonate microfabrics in the Helena (Siyeh) and basal Snowslip formations. He interpreted calcite ooids to have been originally composed of a mixed mineralogy, consisting of calcite, aragonite, and calcite-aragonite.

Sediment types present. The Helena Formation in Baring Creek contains cycles of pinch-and-swell couplet, uncracked even couplet and microlamina associations at base overlain by dolomitic uncracked even couplet, microlamina, and carbonate mud at top (Winston, 1991; Winston and Link, 1993, Fig. 20). These cycles are interpreted to have formed from periodic expansions and contractions of the Belt lake on the shallow perennial lake floor. The Helena Formation also contains thin intervals of cross-bedded, oolite-bearing coarse sand and intraclast sediment type, indicative of beaches (Tucker, 1984; Winston and Lyons, 1993, this volume).

Field Stops. The Helena Formation underlies the upper portion of the Baring Creek trail (above 7,600 ft, 2,300 m)) (Fig. 12), but much of it is not well exposed (Fig. 12). The best exposures are at the base of the unit immediately above the Empire Formation, and west of the main trail toward Sexton Glacier. Laminated dolomite and limestone, exposed at Siyeh Pass, is the stratigraphically highest rock seen on this field trip. Most of the lithologies of the Helena Formation can be examined in blocks of float along the upper part of the Baring Creek trail.

Figure 18. Valley of Baring Creek looking southeast from Siyeh Pass. Saint Mary Lake is in the distance. Sexton Glacier is out of the view to the right, but cut this U-shaped valley during the Pinedale glaciation.

Sexton Glacier

Sexton Glacier is one of many glaciers and permanent snowfields remaining in Glacier National Park. It occupies a hanging valley at the head of Baring Creek, and calves onto a steep waterfall and alluvial fan with a 1,600 ft (490 m) drop in elevation. Like many glaciers in the Park, Sexton Glacier is fronted by a bare-rock moraine that formed in the mid 19th-century (Carrara, 1989). For Proterozoic sedimentologists, one of the most interesting things about a side-trip to Sexton Glacier is the magnificent array of clean float blocks that lie in front of the ice. These blocks are derived from the Helena and Snowslip Formations and the Late(?) Proterozoic diorite and diabase sill that intrudes the Helena high on the cliff above.

Continuation over Siyeh Pass

The walk over Siyeh Pass (elev 7,750 ft, 2,360 m) is scenic and aerobic, but it does not afford much opportunity to examine well-exposed rocks. East of the Pass the views to the southeast down Baring Creek (Fig. 18) and toward Sexton Glacier are imposing.

At the top the view to Boulder Creek to the northeast demonstrates the abrupt drop from alpine heights to aspen- and sage-covered glaciated plains of eastern Montana (Fig. 19). The topographic escarpment produced by the Lewis thrust and the Belt Supergroup above it underlies the continental divide and the change from the semi-arid Great Plains of eastern Montana to the forested Rocky Mountains of the western part of the state. The rocks exposed on Boulder Creek include, from southwest to northeast (top to bottom), the Empire, Grinnell, Appekunny, and immediately above the Lewis thrust, the Altyn Formation. The Lewis thrust crosses Boulder Creek at elevation 6,000 ft (1,830 m), 5 miles (8 km) northeast of Siyeh Pass (Fig. 3).

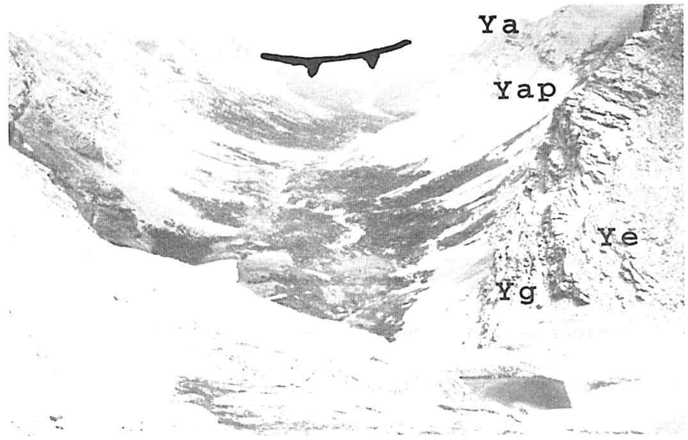

Figure 19. View looking northeast down Boulder Creek to the Great Plains of eastern Montana. Exposed strata consist of west-dipping lower part of Belt Supergroup in Glacier National Park. They are labelled Ya=Altyn Formation; Yap=Appekunny Formation; Yg=Grinnell Formation; Ye=Empire Formation; Trace of Lewis thrust is shown.

Figure 20. View looking northwest to Siyeh Peak, showing west-vergent anticline in Helena Formation. Diabase sill with bleached limestone on its margins is in upper part of view. Baicalia-Conophyton zone is above sill.

North and west of the Pass the trail descends onto a talus field containing mainly float of the Helena Formation above poorly exposed rocks of the Empire Formation. Glacial deposits form much of the low ground in Preston Park. The lower Helena Formation underlies the tree-covered country as the trail drops toward Siyeh Bend. The best exposures of the Helena Formation are in the bed of Siyeh Creek immediately northwest of the Siyeh Bend trailhead.

The most geologically interesting parts of the descent below Siyeh Pass are the views of the Helena, Snowslip, and Shepard Formations, to the north toward Mt. Siyeh (Fig. 20), to the east to Matahpi Peak, and to the west toward Piegan Mountain. On Mt. Siyeh and Matahpi Peak, the light-colored cliff marking the Baicalia-Conophyton stromatolite cycles is above the dark cliff of the diorite sill. To the west, on Piegan Mountain and in the Logan Pass area, however, the sill is below the Baicalia-Conophyton zone.

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GEOLOGIC GUIDE TO GLACIER NATIONAL PARK, MONTANA AND AREAS ADJACENT TO WATERTON, ALBERTA

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INTRODUCTION

This article guides a field trip to localities that examine Belt/Purcell Supergroup rocks and the structural features that disrupt them in Glacier National Park and adjacent areas near Waterton, Alberta. Parts of this road log come from other published sources that include Gordy and others (1977; 1982), Bryant and others (1984), Tucker and French (1984), Tucker and others (1984), Whipple and others, (1985), Winston and Woods (1986), and Winston and others (1989). The two-day trip focuses on the stratigraphy and facies relations among units from the Prichard Formation to the McNamara Formation.

The guide for this trip begins and ends at the Grouse Mountain Lodge in Whitefish, Montana. On Day 1, you travel east to Glacier National Park, thence over Going-to-the-Sun Road, with several stops in the Park (See Fig. 1a, Stops 1-5). The latter part of the day is spent travelling north to Waterton Townsite in Waterton National Park. Day 2 begins at Waterton and proceeds north along the Rocky Mountain front to the Drywood Canyon area near the Waterton Gas Plant to examine the Grinnell Formation and overlying units (Figs. 1 and 19); it then returns to Whitefish by way of Marias Pass with stops along the way. All roads are passable by passenger car, although some are gravel.

The Middle Proterozoic strata of Glacier National Park (Fig. 2) represent perhaps the easternmost facies of the Belt/Purcell Supergroup in the basin. Several units such as the Grinnell Formation contain an abundance of sand indicating its proximity to the strandline. Offshore, deep-water deposits such as the Prichard Formation locally contain carbonate fragments and cement indicating a possible foreslope environment of deposition. Lithofacies of thick carbonate units such as the Helena and Altyn Formations indicate deposition in a platform environment with shoals and beach sands. Sharp breaks in sedimentation and facies patterns suggest local disconformities, consistant with a shallow shelf enviroment. The stratigraphic section in Glacier National Park has a maximum thickness of only 17,100 ft (Whipple and others, 1984) as compared to 49,000 ft thick near Libby, Montana, just 90 mi west (Harrison and others, 1992).

Some stratigraphic units such as the Waterton, Altyn, Appekunny, and Grinnell Formations are unique lithofacies to this part of the Belt/Purcell basin and represent separate and distinct environments and source areas from other parts of the basin. Glacier National Park is the type locality for the Snowslip, Shepard, and the Mount Shields Formations of the Missoula Group (Fig. 2) which are recognized nearly basinwide.

The principal geologic structures of Glacier National Park trend about N. 40° W (Fig. 1). From west to east, they consist of: 1) a Tertiary-filled graben in the Middle and North Fork Flathead River drainages flanking the west side of the park (middle left of Fig. 1b); 2) a fold and thrust belt on the western edge of the Livingston and Lewis Ranges (east of the Flathead fault on Fig. 1b); 3) a broad, open syncline in the central part of the park called the Akamina syncline; and 4) a spectacular thrust belt on the eastern edge of the Livingston and Lewis Ranges, featuring the Lewis thrust fault (east side Fig. 1b and 1c).

The Lewis thrust fault is considered to be the easternmost and structurally lowermost thrust fault that transported Middle Proterozoic strata of the Belt Supergroup over sedimentary rocks of Cretaceous age. Hudec and Davis (1989) suggested that the Lewis "...may be a composite of multiple thrust surfaces that were not all active at the same time." The Lewis is estimated to have had at least 40-60 km of lateral displacement (Dahlstrom, 1970; Gordy and others, 1977) on a gently southwest-dipping detachment surface (less than 20°) that is interpreted to be broadly folded and locally displaced by normal faults (Fig. 1b; 1c). The transport direction of the Lewis allochthon is generally to the northeast, ranging from N. 40°±10° E. to N. 70°±10° E. (Davis and Jardine, 1984; Hudec and Davis, 1989). Locally at Marias Pass (Fig. 1a), kinematic indicators suggest southward-directed thrust movement (Kelty, 1985).

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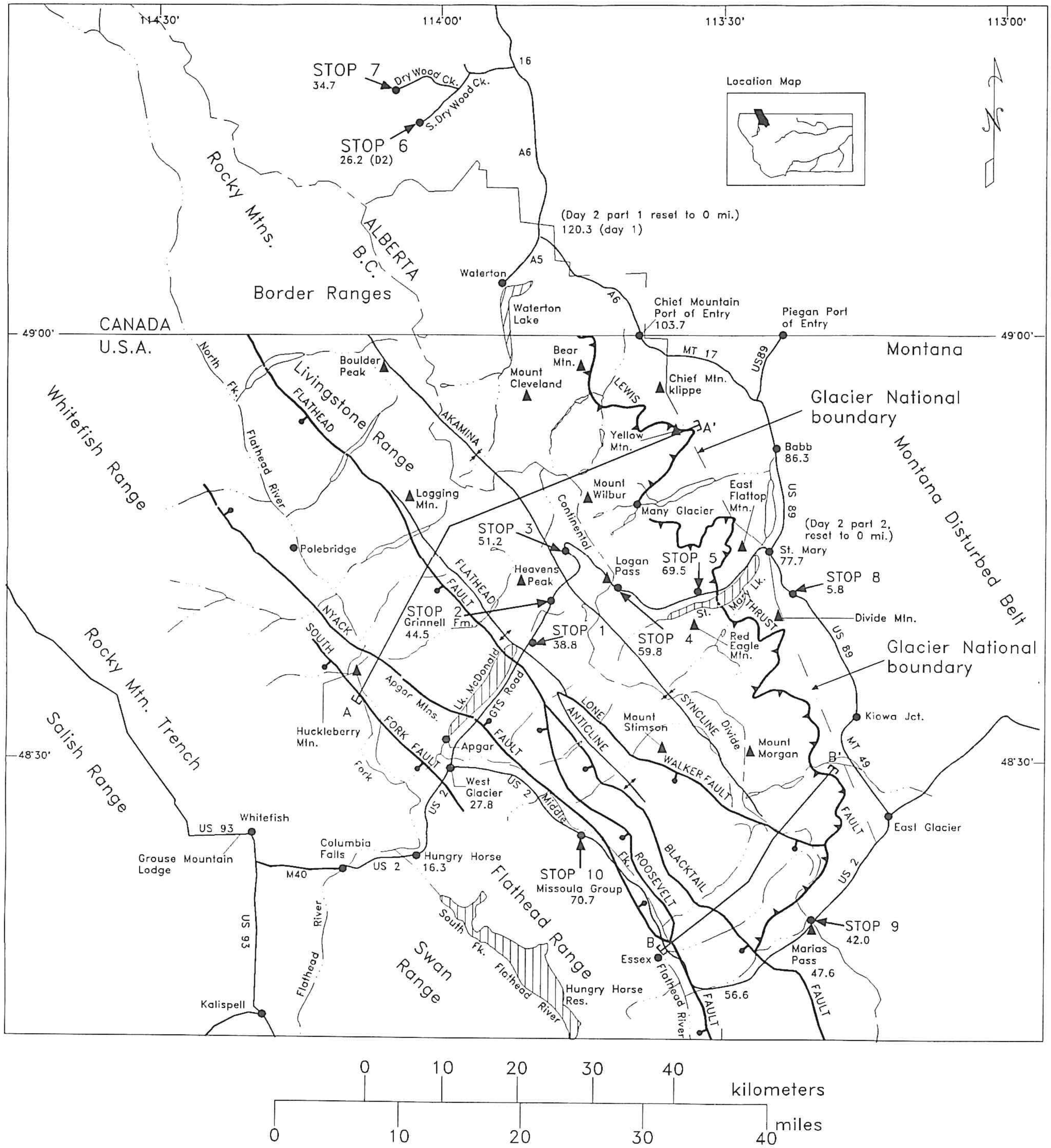


Figure 1. Field trip route and stops, Glacier-Waterton National Park and adjacent area.

Within the Lewis allochthon, imbricate thrust belts, on the western and eastern edges of the Livingston and Lewis Ranges, appear to have formed over ramps in the Lewis detachment surface and are the site of several duplex structures, some of which are as much as 300 m thick (Davis and others, 1989). In general, the duplexes contain panels of intensely faulted and deformed Belt strata that are floored by the Lewis thrust and roofed by local thrust faults.

The northwest-southeast-trending Tertiary-filled graben on the west side of Glacier National Park (Fig. 1b, 1c) is bounded on the northeast by the Flathead fault, a large-displacement, listric normal fault, and on the southwest by the Nyack fault, a steep east-dipping to overturned normal fault that is antithetic to the Flathead fault (Constenius, 1988). Tertiary sediments were deposited in an asymmetrical graben synchronously with movements on the Flathead

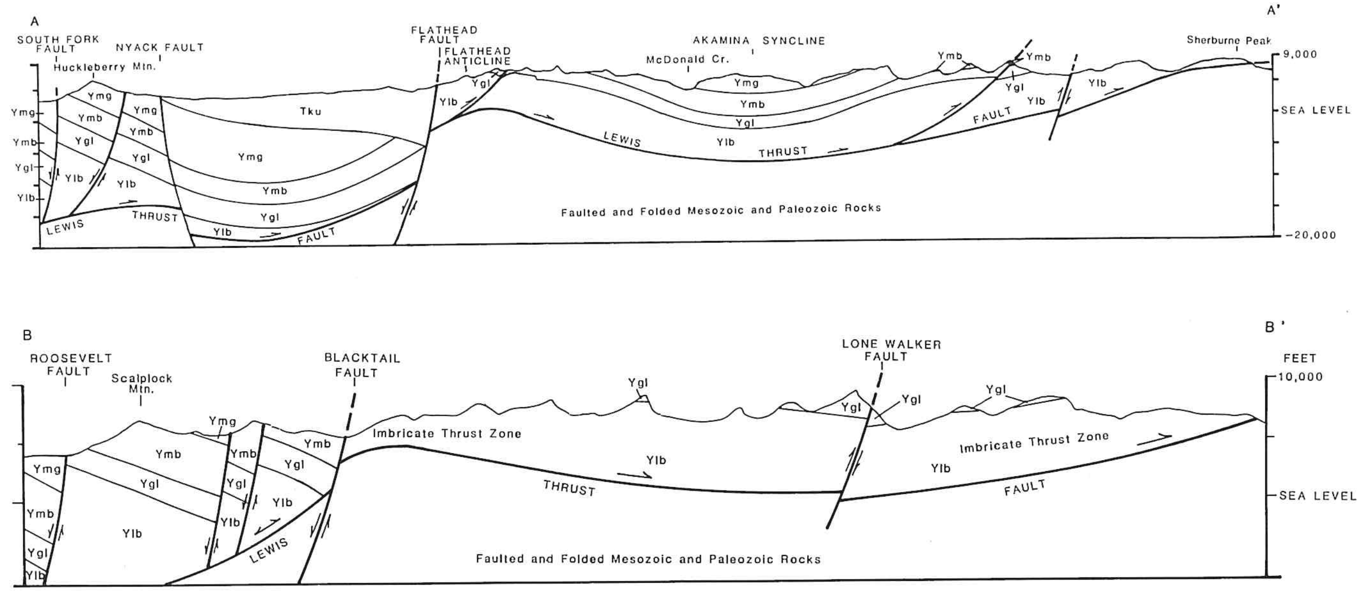


Figure 1a. Simplified geologic cross sections of Glacier National Park. Locations shown on Figure 1

fault. This event resulted in a wedge of Tertiary strata that dips and thickens east toward the Flathead fault (Fig. 1b). The Flathead fault has at least 8,000 ft of stratigraphic displacement. The listric Flathead fault is interpreted to offset the Lewis thrust at depth, as proposed by Childers (1963), and to sole into a detachment below the Lewis. Southeast of Lake McDonald, the Flathead splits into the Roosevelt and Blacktail faults (Fig. 1c). The Roosevelt fault becomes the northeast-bounding fault of the Tertiary graben and the Blacktail is considered to be the southern extension of the Flathead because of its similarity in dip (about 40° southwest) and greater displacement compared to the Roosevelt.

Broad regional folds warp the central part of the park and the Lewis allochthon into a northwest-plunging syncline (which is the southern extension of the Akamina syncline in the Clark Range of Canada) and an anticline flanking the Flathead fault called the Flathead anticline (Fig. 1b). Belt rocks are gently folded about the axis of the Akamina syncline, but they are tightly folded and faulted within the Flathead anticline. Both folds are interpreted to have warped the Lewis thrust fault, however the anticlinal fold might be associated with ramping of the Lewis thrust.

DAY 1

Miles

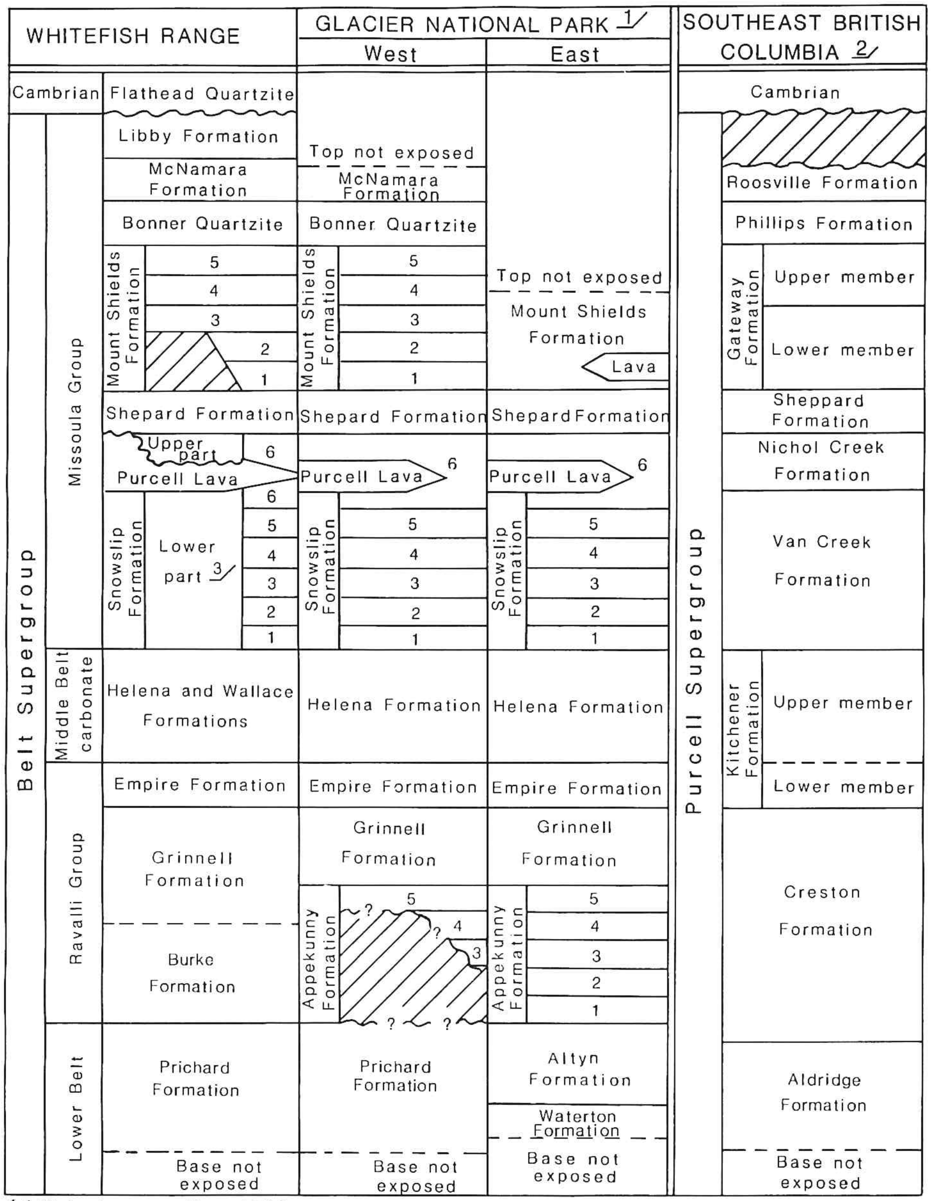
0.0 Grouse Mountain Lodge. Begin field trip by turning right (east) onto U.S. Highway 93. Proceed into the town of Whitefish and follow road signs to stay on U.S. 93 (Fig. 1).

1.2 Turn right (south) at second traffic light in Whitefish.

3.6 Junction: Turn left (east) on Montana 40 and proceed to Columbia Falls, Montana.

8.0 Junction: Montana 40 and U.S. Highway 2. Proceed straight ahead (east) on four-lane highway (U.S. 2).

Entering broad Flathead Valley of the Rocky Mountain trench, a prominent physiographic feature that extends from far north in Canada into Montana and terminates at the Saint Mary fault along the Proterozoic Jocko line (Winston and Woods, 1986). Along its length the trench is topographically expressed by a range of geologic structures; in southern British Columbia it is bordered by Tertiary listric normal faults. Northwest of Columbia Falls the Rocky Mountain trench follows the trace of a series of closely spaced normal faults separating the Whitefish Range on the east from the Salish Mountains on the West. South of the Flathead River the normal faults of the trench appear to split into the Mission and Swan Valley faults which extend south, forming the steep west faces of the Mission and Swan ranges.



1/ Whipple, compiler, 1992.
2/ Modified from McMechan, 1981.
3/ Upper and lower parts of Whipple, 1984.

Figure 2. Correlation of Belt and Purcell Supergroups, Glacier National Park, Whitefish Range, and adjacent parts of Canada.

9.9 Cross railroad tracks. Entering Columbia Falls. Postal records (1891) indicate that the town was formerly known as Monaco.

11.4 Cross Flathead River.

12.4 Junction: Montana Highway 206 and U.S. Highway 2; turn left (northeast) toward Glacier National Park.

13.6 Columbia Mountain Road on right.

14.8 Entering Badrock Canyon.

The intermittent exposures on the right are the Empire Formation (Fig. 2), which passes down into red and green argillite interbeds of the uppermost part of the Grinnell Formation. Red and green argillite of the Empire and Grinnell is mostly of the even couplet sediment type (Winston, 1986), laminated on a centimeter scale. Interstratified with the argillite intervals are beds, centimeters to decimeters thick, of coarse cross-bedded quartz sand of the coarse sand and intraclast sediment type (Winston and Woods, 1986). Tectonic compression that accompanied thrusting is reflected in these outcrops by locally intense penetrative cleavage that cross-cuts the couplets. Small folds and thrusts are outlined by the quartzite beds. These open folds, tighter folds and thrusts illustrate at a small scale the style of folding and thrusting of the much larger scale characteristic of the Belt terrain west of the Rocky Mountain trench ahead. There Belt rocks are compressed into broad open folds and thrusts of probably short displacement.

Across the Flathead River to the north are interbedded green and tan beds of the lower part of the Helena Formation that reflect the transition from tan-weathering Helena dolomite downward into green argillite of the Empire Formation.

16.3 Cross South Fork of the Flathead River. The South Fork joins the combined flows of the North and Middle Forks to form the master river. The town of Hungry Horse is just ahead.

From Hungry Horse to West Glacier, the highway crosses a broad valley floored mostly by glacial till. Large erratic blocks of Belt rocks lie scattered on the rolling, glacially sculptured hills. Belt bedrock exposed in the mountains to the north and south projects under the glacial deposits, and faults mapped between West Glacier and Hungry Horse are mostly projected from faults mapped in bedrock to the north and south. These high-angle faults, like those to the east, are mostly listric normal faults of Tertiary age. In portions of the valley ahead, Tertiary rocks are exposed beneath the glacial till.

17.5 Roadcuts in the Helena Formation, junction: Paved road on right leads to Hungry Horse Dam.

19.9 Town of Coram.

22.7 Cross projection of the South Fork fault. The South Fork fault is a major, down-to-the-west listric normal fault that extends far to the southeast along the northeast shore of Hungry Horse Reservoir and deep into the Bob Marshall Wilderness (Fig. 1a). The South Fork of the Flathead River follows the trace of the South Fork fault along most of its course and cuts Tertiary valley fill deposits in much the same way as the Middle Fork does along the Nyack fault. Here, Tertiary sediments deposited on the western down-thrown block are faulted against covered Belt rocks on the east.

25.5 Glacier Campground on right.

26.9 Junction: U.S. Highway 2 and Going-to-the-Sun Road; turn left on Going-to-the-Sun Road.

27.2 Cross the Middle Fork of the Flathead River. The boundary of Glacier National Park is centered on the river.

27.8 Entrance to the Park.

29.0 Junction: Turn right at stop sign. Apgar Village is to the left.

Rocks of the Belt Supergroup that are exposed along Going-to-the-Sun Road consist of clastic and carbonate units that are interpreted to have been deposited in shallow-shelf and nearshore environments. In Glacier National Park the succession is in the upper plate of the Lewis thrust fault and has a maximum thickness of 17,100 ft (Whipple and others, 1984). About 90 mi west of here, Belt rocks are estimated to have a maximum thickness of about 49,000 ft (Harrison and others, 1992).

29.8 Lake McDonald is on the left. Road cuts on the right are in the middle part of the Helena Formation. The Helena in these exposures consists of mostly thin-bedded, molartooth dolomite, stromatolitic and oolitic limestone, and thin beds of white, calcareous arenite. Fining-upward couplets of calcareous siltite are common and increase in abundance upward in the section.

30.4 The contact between the Helena and the overlying Snowslip Formation is exposed at the far end of this road cut. The base of the Snowslip is marked by lenticular beds of gray, medium- to very coarse-grained arenite with mudchips and interlaminated dark-red argillite. Upsection is a sequence of gray-green siltite containing stromatolite beds as thick as 30 cm.

37.8 Junction: Lake McDonald Lodge to the left. Proceed straight ahead.

38.8 **STOP 1 - Prichard Formation**
Small turnout on left - be careful of on-coming traffic around blind corner.

The road cuts at this stop (Fig. 3) are in the upper part of the Prichard Formation and expose very thin, even parallel laminae of light-gray to black siltite and argillite on the west limb of the Flathead anticline (Fig. 4). Pyrite and pyrrhotite are abundant in some laminae and upon oxidation, stain outcrops with limonite. Well developed, small-scale slump structures are common in these exposures (Fig. 5). The strata and slumps are interpreted as turbidite deposits formed as part of a large submarine fan complex (Cressman, 1989).

The Prichard Formation until recently had not previously been recognized this far east in exposures of the Belt Supergroup. It is recognized only on the west side of the Park where it is the lowermost stratigraphic unit and is overlain disconformably by only parts of members 3, 4, and 5 of the Appekunny Formation (see Figs. 2 and 13). (Whipple, 1992).

Figure 3. Roadcuts at Stop 1 exposing the Prichard Formation.

Figure 4. Looking north from Stanton Mountain down the axis of the Flathead anticline.

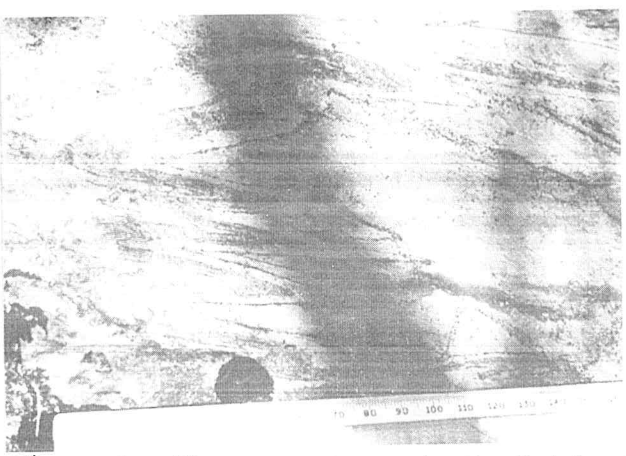


Figure 5. Slump structures in the Prichard Formation at Stop 6.

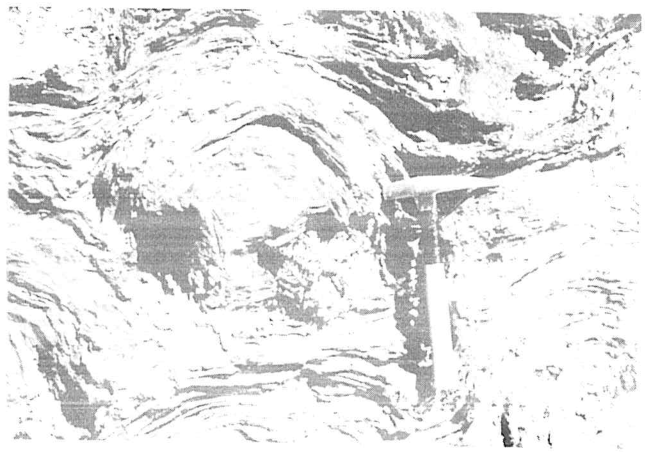


Figure 6. Close-up of stromatolites in the *Conophyton* Zone near the tunnel on the west side of Going-to-the-Sun Road.

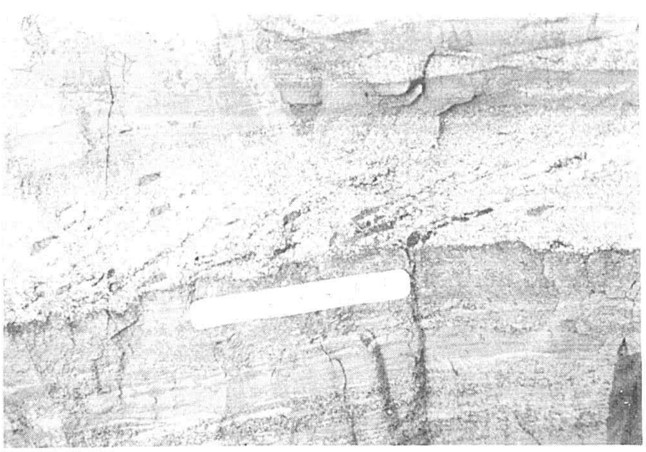


Figure 7. Mudchips aligned on foresets of lithic arenite bed in the Snowslip Formation at Stop 3.

Figure 8. Photomicrograph of broken hematite-coated ooliths in the lowermost part of Snowslip Formation at Stop 3.

In the Park, the Prichard is subdivided into lower and upper parts. The lower part is about 4,300 ft thick and characterized by thin, even, parallel laminae of rusty-weathering, blackish-gray argillite and light-gray siltite that contain disseminated pyrite and pyrrhotite (rocks of this stop). Some small-scale cross-lamination is present in siltite laminae. Carbonate occurs locally near the top of the lower part as cement in thin siltite laminae and as pods and nodules of black manganiferous limestone. Slump structures are common locally. The upper part of the Prichard consists of wavy, nonparallel laminae of greenish-gray to medium-gray calcareous siltite. Locally, it contains thin lenticular beds of white quartz arenite and discontinuous beds of fragmental limestone or breccia and stromatolitic limestone. This upper part of the Prichard is equivalent to the transition zone of the Prichard as described by Cressman (1989). Its thickness ranges from 800 to 1,200 ft.

Within a short distance from Stop 1, the route crosses the axis of the Flathead anticline and into the Akamina syncline.

Small road signs marked 'GEOLOGY' designate stops along Going-to-the-Sun Road described in a road guide prepared by Raup and others (1983). Some stops made by this field trip will visit the same locations.

43.4 Avalanche Creek

44.5 **STOP 2 - Grinnell Formation**

Pullout is on bend to the left - use caution when crossing traffic.

Redbed exposures in roadcuts across from the pullout and outcrops in McDonald Creek are the Grinnell Formation. The Grinnell at this location is mostly a sequence of red siltite and argillite that contain conspicuous beds of white to gray quartz arenite. Sedimentary structures abound in these beds and include sand-filled subaerial shrinkage cracks, ripple marks, fluid-escape structures, mudchip breccias, and mudchips aligned along foresets of cross-laminated arenite beds. The association of sedimentary structures suggests deposition on extensive mudflats. The white arenite beds are interpreted to have a source from the northeast.

In Glacier National Park, the Grinnell consists mostly of quartz arenite on the east side of Park and interlaminated siltite and argillite on the west side (the field trip by Link, this volume, examines the Grinnell Formation in Baring Creek). The contact between the Grinnell and the underlying Appekunny Formation is placed where red argillite and siltite of the Grinnell change to green argillite of the Appekunny. The thickness of the Grinnell ranges from 1,700 to 2,600 ft.

In southeasternmost part of park, the Grinnell averages 60 percent quartz arenite, and its upper part is nearly 100 percent quartz arenite or quartz conglomerate. Quartz arenite beds in eastern exposures are typically white, medium- to coarse-grained, lenticular, ripple marked, and prominently cross-bedded, and, in general, become more common upward in the formation. Basal scours and red argillite chips, pellets, and cobbles are common. Red or purplish-red laminated siltite, silty argillite, and argillite are commonly interbedded. Lamination ranges from even parallel to wavy nonparallel and locally includes ripple cross-lamination; mud cracks and fluid-escape structures commonly disrupt bedding in these red beds.

Interbedded quartz arenite and red argillite in the eastern exposures change northwestward to a lithofacies that contains less quartz arenite and instead is composed mainly of pale, grayish-green and grayish-purple siltite and argillite that closely resembles the Creston Formation of the Purcell Supergroup. In the northwestern part of the Park, the Grinnell can be subdivided into two parts. The lower part is 1,200 ft. thick and is predominantly interbedded blocky siltite and evenly laminated argillite that contains a few thin lenses of ripple-marked, white quartz arenite. Bedding is disrupted by abundant shrinkage cracks, fluid-escape structures, and interlayers of mud-chip breccia. The upper part is 1,400 ft thick and is similar to the lower part except that it contains more lenses of rippled, white quartz arenite (locally as much as 20 percent of the section). In sequences on both sides of the Park, greenish-gray siltite and argillite are locally present in the upper and lower transition zones with adjacent formations.

The Grinnell Formation was subdivided into members by Fenton and Fenton (1937), but that stratigraphy has not stood the test of more recent structural mapping. The fully exposed Grinnell section on Mount Henkel 5 km to the west (Horodyski, 1983; Kuhn, 1986) can be subdivided two different ways by using two different sets of criteria. By focusing on the silt and clay size populations interpreted to have been derived from the west (Cronin, 1989), one can subdivide the Mount Henkel section into six informal units and correlate them to intervals in the Burke, Revett and St. Regis formations of the Ravalli Group (Winston, 1989). In essence these units in the Grinnell are rather subtly distinguished distal tongues of thicker, more clearly defined units to the west. These units are: 1) flat-laminated silt and even couples that represent the distal tongue of the lower Burke fine quartzite and siltite; 2) red mudcracked mud and microlaminae that correlate with the upper part of the Burke; 3) even siltite beds and couples that represents the distal tongue of the lower part of the Revett; 4) green and red mudcracked mud, microlaminae, and even and lenticular couplets that correlate with the middle part of the Revett; 5) an upper red unit of even siltite beds and couples that represent the distal tongue of the upper part of the Revett; and finally, 6) red mudcracked mud, and even and lenticular couplets that correlate with the St. Regis Formation to the west.

The other subdivision of the Grinnell focuses on the more obvious interval of white quartz sand interpreted to have been derived from the east. This subdivision permits the recognition of a lower sandy interval that generally coincides with the lower Burke sand, a middle red muddy interval that includes the upper Burke and lower Revett correlatives, and an upper interval of quartzite beds that occur with the distal upper Revett and St. Regis facies. These two criteria have produced a rather schizophrenic approach for subdividing and mapping the Grinnell and equivalent rocks in the Ravalli Group to the west.

The coarse sand population of the lower Grinnell and upper Grinnell was probably derived from the east and flowed as sheetfloods more or less northwest, down the depositional axis of the basin. The coarse grains of this population were transported as bedload and build large scale bedforms that coalesced into tabular flat beds of the tabular variety of the cross-bedded sand sediment type. They closely resemble in form crossbedded sand layers produced by the great Lake Eyre flood of 1968 (Williams, 1971). These sand beds in the Creston Formation of the Purcell Group of Canada have a broad scatter of paleocurrent directions, which McMechan (1981) interpreted to indicate tidal deposition.

The red Grinnell passes northward into the mostly green Creston Formation of the Purcell Supergroup, which may record a more distal setting along a northwest-trending topographic basin that, during Ravalli deposition, was situated over the eastern part of the Belt structural basin. Sediments, principally from the west, simply filled the Belt basin, and waters could have flowed northward out of the basin either as subareal or groundwater flow. Microlaminae which form thick intervals in the Spokane Formation in Prickly Pear Canyon may have been deposited in a large lake that occupied the central part of the basin and was impounded by a delta of coarse sand that entered the eastern side of the basin near Two Medicine Lake, where much of the Grinnell is composed of coarse quartz arenite (Winston, 1989).

47.4 Restrooms on right.

48.9 Road cuts on the right expose the gray-green dolomitic siltite of the upper part of the Empire Formation. Limonite staining in exposures is from oxidation of large pyrite cubes. The contact with the overlying Helena Formation is present about at the upper end of these exposures.

The Empire in Glacier National Park consists primarily of argillite, siltite, and subordinate amounts of arenite and dolomite (see also Link, this volume). The lower part of the Empire is composed largely of white to buff quartz arenite beds that range in thickness from 5 in to 11.5 ft and contain minor carbonate cement and pyrite. Most arenite is well sorted; however, within some beds grain size ranges from fine to coarse. Locally, arenite beds have well-developed cross-bedding, load structures, and asymmetrical ripple marks. Arenite beds decrease in number and thickness from bottom to top of Empire. The lower contact is placed at the base of the lowermost bed of white quartz arenite that overlies the uppermost sequence of red argillite of the Grinnell Formation. The upper part is composed primarily of olive-green and a few purplish-red argillite and siltite beds that range in thickness from a few inches to 5 ft. Thin dolomite beds are present near the middle of the Empire and increase in number and thickness upward. The Empire ranges in thickness from 520 ft on Scalplock Mountain at the south end of Glacier National Park to 400 ft near Grinnell Glacier.

Road cuts ahead are all in the Helena Formation.

The Helena Formation in Glacier National Park, where measured along Going-to-the-Sun Road between The Loop and Logan Pass and along Haystack Creek, is 2,540 ft thick and consists primarily of dolomite, limestone, and minor quartz arenite (Whipple and others, 1984). In this measured section, the Helena can be divided into three distinct parts. The lower part is 590 ft thick and consists of interbedded quartz arenite and thin-bedded dolomite near the base; thin beds of horizontally laminated and "molartooth" dolomite in the middle; and thick smoky-gray limestone beds near the top. The middle part of the Helena, which is 1,180 ft thick, is dominated by dolomitic molartooth beds, some as much as 100 ft thick. A few thin beds of quartz arenite and stromatolitic limestone are present in the middle part of the Helena. The upper part consists primarily of interbedded stromatolitic limestone, dolomite, oolitic limestone, and quartz arenite. At the base of this upper part is an interval of stromatolitic limestone about 100 ft thick, known as the *Conophyton* zone (Rezak, 1957), that is composed of *Baicalia-Conophyton* stromatolite cycles (Horodyski, 1983). A diorite sill 130 ft thick intrudes the Helena in this part of the measured section just above the *Conophyton* zone. This sill, which is present throughout much of the Park, varies widely in stratigraphic position, from near the base of the Helena in the southeast part of the Park, to the lower part of the Snowslip in the northernmost part of the Park.

50.3 *Conophyton* stromatolite zone is exposed in the low outcrops to the right (Fig. 6). Stop 4 will examine this zone in more detail. The mountain that rises above all others looking west across the McDonald Creek drainage is Heavens Peak, elevation 8,987 ft.

50.5 On the right just before the tunnel is an exposure of a metadiorite sill and tannish hornfels. This sill is exposed throughout the Park and is mostly dark gray to greenish-black metadiorite that is fine- to medium-grained, and equigranular. The metadiorite is commonly pyritic and contains abundant chlorite that replaces amphibole and pyroxene. The laterally continuous sill in the Helena Formation cuts up and down section locally, ranges in thickness from 53 to 295 ft and is commonly flanked by bleached zones of hornfels. Sills (1.5-215 ft thick) also occur locally in the Grinnell, Empire, and Snowslip Formations. Dikes intrude the Appekunny and Altyn Formations on east side of park and intrude the Purcell Lava and overlying strata near Granite Park (McGimsey, 1985); dikes are typically 10-20 ft thick (Ross, 1959). The sills and dikes closely resemble Late Proterozoic intrusive rocks in west-central Montana reported to be 750±25 Ma (potassium-argon age method, Mudge and others, 1968).

50.9 At about this point, roadcuts expose the uppermost part of the Helena Formation. The uppermost part consists of a shoaling-upward sequence about 75 ft. thick of mostly oolitic and stromatolitic limestone and gray to greenish-gray calcareous siltite. A short walk back from Stop 3 just ahead will bring you to these exposures.

51.2 **STOP 3 - Lower part of the Snowslip Formation and upper part of the Helena Formation.**

Park on right below switchback called The Loop and walk back downhill through the section. The field trip guide walking guide by Raup and others (this volume) ends here.

Interbedded red and green bed sequences typical of the lowermost member of the Snowslip in this region are exposed along here. Green beds consist of wavy, non parallel couplets of calcareous siltite and argillite interbedded with white to gray, cross-laminated arenite. Arenite is mostly composed of medium to coarse quartz and lithic fragments (Fig. 7). Pinkish stromatolites occur locally in some beds. Fluid-escape structures, mud-chip breccias, and subaqueous shrinkage cracks are common. Red beds contain much the same lithologies and structures as the green beds except arenite units are not as laterally continuous, stromatolites are absent, and shrinkage cracks form large polygons typical of types formed in subaerial environments.

Pink to gray oolitic arenite is common in the lowermost beds of this member. Unlike the underlying Helena Formation, ooliths in the lower Snowslip are broken, deformed, and reformed, and generally dusted with hematite (Fig. 8). In combination with the large number of lithic fragments, the broken ooliths suggest a source from eroded parts of the Helena (Whipple and Johnson, 1988).

The contact with the Helena is not exposed here, but elsewhere it is represented by a sharp break from the carbonate shoals of the Helena to the siliciclastic mudflats of the Snowslip. The Helena-Snowslip contact marks the boundary between the middle Belt carbonate and the Missoula Group of the Belt Supergroup (Fig. 2).

Dolomitic strata of the uppermost part of the Helena, downsection of the Snowslip, is punctuated by conspicuous beds of gray oolitic limestone. Oolite beds are commonly cross-laminated, pyritic, and contain stromatolitic debris and elongated limestone clasts. Herringbone cross-lamination is obvious in some of the more weathered beds.

The Snowslip Formation can be divided into 6 informal members at the type section at Snowslip Mountain in southern part of Glacier National Park, designated 1 through 6 in ascending order (Whipple and Johnson, 1988). The members can be recognized and correlated in exposures throughout the Park. The contact with the underlying Helena Formation is sharp, apparently disconformable, and placed at base of the first occurrence of red lithic arenite that overlies gray limestone or dolomite of the Helena. The Snowslip ranges from about 1,180 ft thick (including 311 ft of Purcell Lava) at a reference section on the west wall of Hole-in-the-Wall cirque (Whipple and Johnson, 1988) to 2,080 ft thick in Apgar Mountains including 106 ft of Purcell Lava; it is about 1,606 ft at the type section (Childers, 1963).

Member 1 is characterized by alternating pale-maroon and grayish-green sequences of calcareous siltite, argillite, and oolitic and lithic arenite. Arenite grains are fine to very coarse, moderate to poorly sorted, subrounded to rounded; arenite beds are thin and commonly cross-laminated. Siltite and argillite laminae are commonly arranged as wavy, nonparallel, fining-upward couplets that contain abundant mud-chip intraclasts, shrinkage cracks, and fluid-escape structures. Thickness ranges from 80 to 295 ft.

Member 2 is predominantly wavy, nonparallel-laminated, grayish-green and yellowish-gray calcareous siltite and argillite. It has a few interbeds of very fine-grained arenite and several thin, conspicuous beds of pink stromatolitic limestone, particularly in the lower part of the member. The lower contact is placed at the base of the lowermost sequence of calcareous grayish-green strata. Member 2 ranges in thickness from about 230 to 495 ft.

Member 3 consists of rhythmic, fining-upward successions from arenite to argillite. Successions are as much as 10 ft thick but are mostly incomplete. At the type section, the rhythmic successions in member 3 show a more regular change in grain size, grading from arenite to argillite and are thicker than elsewhere in the Park. For example, a succession could have arenite at the base and argillite at the top, but no siltite in the middle, or a succession could be all siltite, relatively coarse grained at the base and very fine grained at the top. Thickness ranges from about 50 to 215 ft.

Member 4 is nearly identical to member 2. It includes a few interbeds of arenite and several thin beds of pink stromatolitic limestone, particularly in the lower part of the member. Member 4 ranges in thickness from about 280 to 4605 ft.

Member 5 is similar to member 3, but rhythmic, fining-upward successions are more complete and thicker. They are as much as 25 ft thick, but typically 6.5-10 ft thick, and composed of very fine- to medium- grained, white to pink quartz arenite and subfeldspathic arenite, fining upward to siltite in middle of succession and dark-red argillite at top (Fig. 9). The base of each succession is erosional and forms a sharp contact with dark-red argillite at the top of the underlying succession. Sedimentary structures include abundant ripple marks, subaerial shrinkage cracks, mud-chip breccias, and fluid-escape structures. The lower contact is placed at the base of lowermost fining-upward succession that rests on calcareous strata of member 4. Thickness ranges from 115 to 475 ft.

Member 6 consists of interbedded noncalcareous, grayish-green and pale-maroon, fine-grained arenite, siltite, and minor argillite at the type section. In the northern part of the Park, member 6 conformably encloses the Purcell Lava and consists mostly of grayish-green siltite and argillite beneath the lava and alternating beds of pale-maroon and grayish-green, very fine-grained arenite, siltite, and argillite above the lava. Where the Purcell is present, thin discontinuous beds of pink and gray stromatolitic limestone occur in lower part of member 6. The base of the member is placed on top of the uppermost red, fining-upward succession of arenite and argillite of member 5. Thickness ranges from 25 to 425 ft.

Proceeding uphill and around The Loop, one goes slowly downsection back into the Helena in the east limb of the Akamina syncline.

52.0 Approximate contact between the Snowslip and Helena Formations.

52.6 Immediately to the left about at this location you may catch a glimpse of the "Beautiful Bioherm", a particularly nice exposure of a stromatolite bioherm in the *Conophyton* zone (Fig. 10 and lower photograph on the cover of this guidebook). Unless there is noone else on the road, please glimpse quickly!

54.1 Scenic pullout on your right to view Birdwoman Falls in distance and Haystack Creek in the foreground.

54.8 Nicely sculptured U-shaped valley looking to the west.

56.0 Weeping Wall - tears over Helena.

58.2 Above the bend in the road to your left, outcrops of metadiorite sill along the Highline Trail cut across the Helena Formation (see description from Raup and others, this guidebook).

59.1 Logan Pass. You may want to stop here to see the visitors center. Restrooms are available. Short hikes from the visitors center are very scenic. The Raup and others field trip hike (this volume) starts here. Strata viewed are mostly Helena Formation and lower Missoula Group rocks. Continue down east side of the Pass. Descending from the Pass, the road cuts slowly downsection in the east limb of the Akamina syncline. The Shepard and Snowslip Formations cap many of the peaks along the Continental Divide with the Helena making up most of their relief. As the road drops to the east, red beds of the Grinnell Formation and green beds of the Appekunny Formation become progressively exposed. Where the road exits the east side of the Rocky Mountains, faulted and sheared parts of the Altyn Formation are exposed.

59.8 STOP 4 - *Conophyton* Stromatolite Zone

Gray limestone exposed in road cuts on to the left is a thick stromatolite zone composed mostly of *Conophyton* and *Baicalia* stromatolites (Fig. 11). This zone is present throughout Glacier National Park and forms a distinct biostratigraphic marker. In the Whitefish Range west of here, the zone is composed mostly of *Baicalia* stromatolites; *Conophyton* types are absent.

Figure 9. Fining-upward successions in member 5 of the Snowslip Formation at Hole-in-the-Wall in the north part of Glacier National Park.

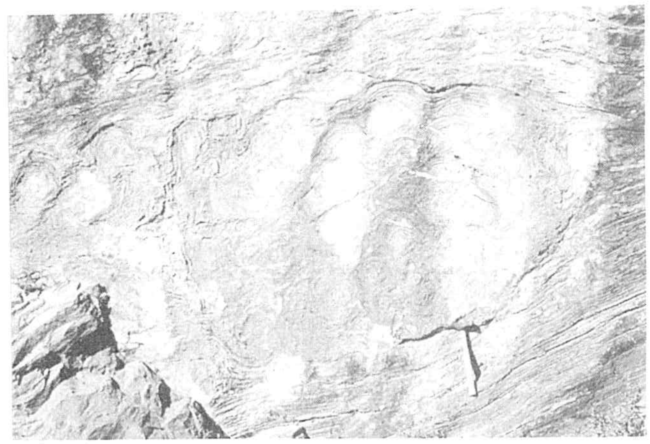


Figure 10. "Beautiful Bioherm" of stromatolites in the *Conophyton* Zone exposed along Going-to-the-Sun Road.

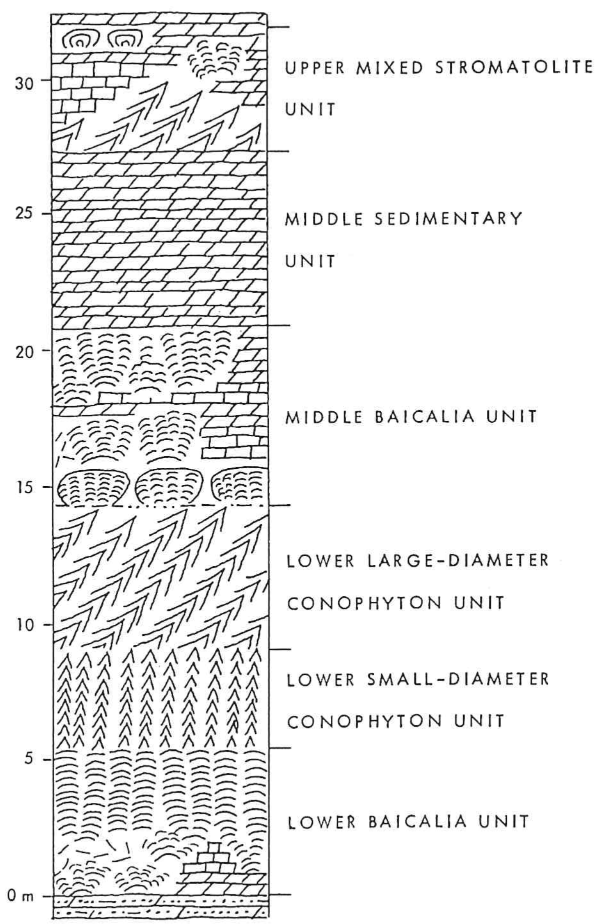


Figure 11. Diagrammatic representation of the *Baicalia-Conophyton* Stromatolite Cycles on Little Chief Mtn., central Glacier National Park, Montana, by R.J. Horodyski in Whipple and others, 1985.

Figure 12. Road cuts at Stop 5 that expose upper part of member 2 of the Appekunny Formation.

At this stop we will examine the prominent stromatolitic unit that occurs in the upper part of the Helena (Siyeh) Formation in Glacier National Park and vicinity. This unit, the *Baicalia-Conophyton* stromatolite cycles, is a massive, cliff-forming unit that can be recognized on mountains many miles away. such as on Matahpi Peak, north of Going-to-the-Sun Road, and on Little Chief Mountain and Citadel Mountain, south of Going-to-the-Sun Road. This stromatolitic unit is unique in the Belt Supergroup, but it is similar to parts of stromatolite cycles that are much more extensive and repetitious in the Siberian platform.

The following discussion of the stromatolites at this stop is from Winston and others (1989, p. 87-88). At the first roadcut to the west of Lunch Creek, one can see: 1) the dark-colored mudstone and muddy sandstone that characteristically underlies this unit (note the soft-sediment deformation); 2) relatively small-diameter branched columnar stromatolites that form small bioherms (note the eroded stromatolite debris between bioherms) in the lower part of the Lower *Baicalia* Unit; and 3) broader, unbranched columnar stromatolites near the top of the Lower *Baicalia* Unit. Proceed to the west end of the outcrop and climb up grass-covered slopes to the old carriage road to Logan Pass (along the way are some large-diameter *Conophyton* in an outcrop near the road).

The conspicuous interval, 3-6 ft thick, of pyritic, blackish-green mudstone at the base of the *Conophyton* Zone at this stop occurs nearly everywhere the stromatolite zone is exposed. The mudstone is not noticeably laminated and breaks along conchoidal fractures. Preliminary geochemical interpretations imply that the interval has a volcanic component and may be mostly a metatuff. Similar beds occur in the Whitefish Range near the base and within the *Baicalia* Zone. A metabentonite bed has been identified upsection near Logan Pass that has similar lithologic character, and zircons from that bed have been dated at about 1350 Ma (see description from Raup and others, this guidebook). One could speculate that the possible volcanic unit had a cause and effect relationship with a prolific algal bloom that resulted in the *Conophyton/Baicalia* Zone.

60.3 Tunnel - burrow through the Helena Formation.

61.9 Sieyh Bend - trailhead for spectacular scenic hikes on the east side of the Park. The hiking field trip by Link (this volume) ends here.

63.8 View of Jackson Glacier. The following discussion of its historical record is paraphrased from Carrara and McGimsey (1981).

The Jackson Glacier, in the Lewis Range, lies along the eastern side of the Continental Divide and presently sits on a steep (15 degree) dip slope of the Helena Formation. Air photo inspection reveals a forest trimline indicating that the Jackson Glacier had flowed north down its bedrock dip slope for 2.2 km before advancing 0.6 km across an area of ground moraine to reach its maximum late-Neoglacial position.

The Jackson Glacier, and the Blackfoot Glacier to the east, were until shortly before the present century one continuous ice body which at its maximum late-Neoglacial extent had a surface area of about 8 square km. Detailed mapping by Dyson (1941) revealed that by August 1939, the Jackson Glacier had separated from the Blackfoot Glacier by 0.5 mile (.85 km). At this time, the Jackson Glacier had a surface area of 0.29 square miles (.7 sq km) and a length of 1 mile (1.7 km). From 1932 to 1944, measurements of the Jackson Glacier's recession (410 ft, 129 m) were carried out by the National Park Service. Once the Jackson Glacier became confined to its bedrock dip slope, retreat was rapid and was probably similar to that of the Agassiz Glacier in that sections of the glacier would break off and slide down the dip slope. The Jackson Glacier presently has a surface area of about 0.29 sq mi (0.71 sq km) representing approximately 24% of its late-Neoglacial maximum size. The Jackson Glacier is evidently still thick enough to be capable of flowage.

65.6 Grinnell Formation exposed in roadcuts for a mile or so. Saint Mary Lake on the right.

67.1 Approximate contact between the Grinnell and the Appekunny Formations. The walking field trip by Link (this volume) starts here at the Sunrift Gorge trailhead.

67.2 Junction - road to right goes to Rising Sun Point. Picnic tables, restrooms and scenic views of Saint Mary Lake and surrounding peaks.

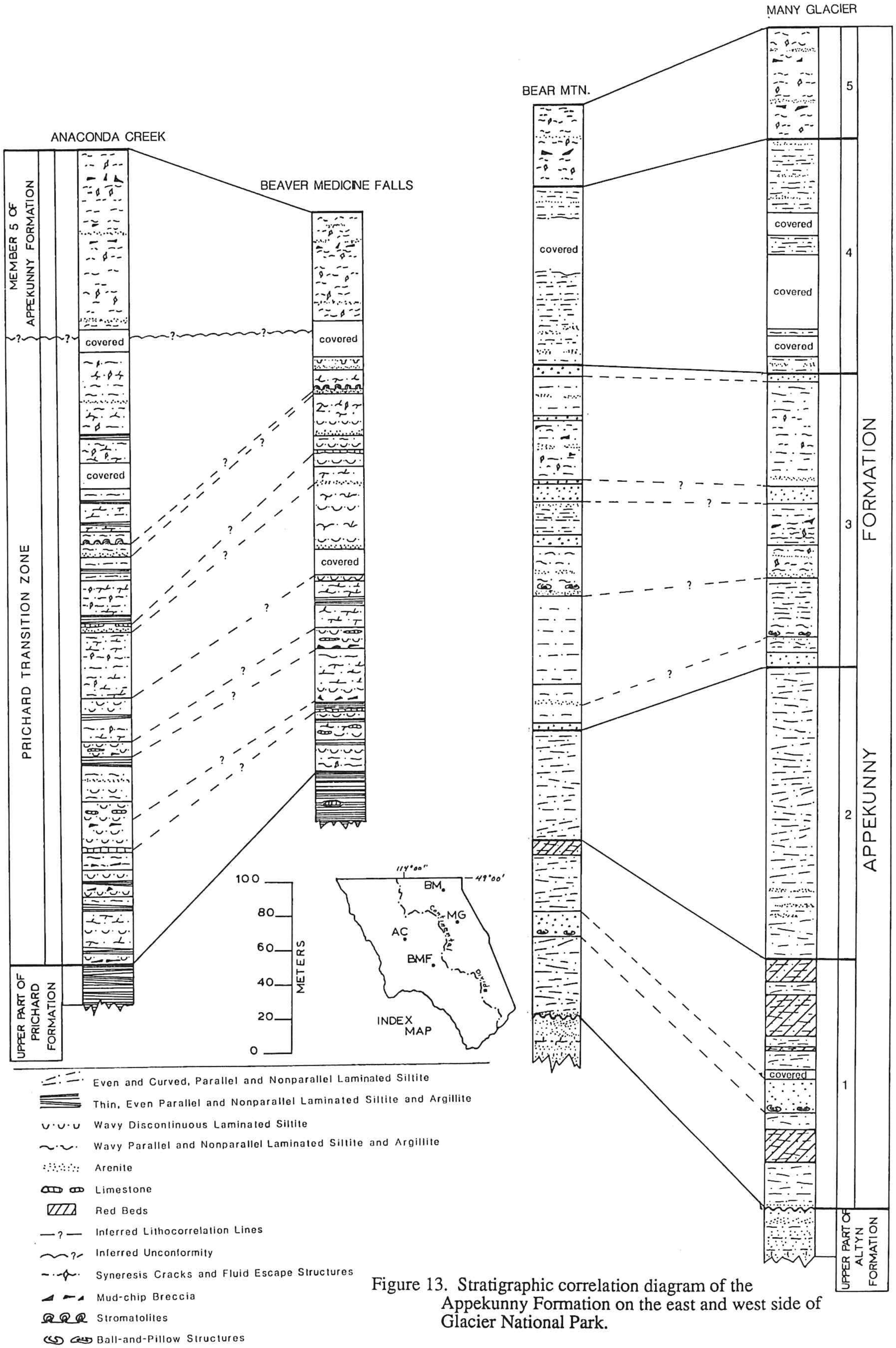


Figure 13. Stratigraphic correlation diagram of the Appekunny Formation on the east and west side of Glacier National Park.

69.1 **STOP 5 - Appekunny Formation**

The Appekunny Formation is exposed in roadcuts on the left (Fig. 12). Parking on right. At the type area near Apikuni Mountain in the northeastern part of the Park, the Appekunny is informally subdivided into five members designated 1 through 5 in ascending order (Fig. 13).

Member 1 consists of alternating successions of interlaminated pale-maroon siltite and minor argillite with grayish-green siltite and argillite Bed lamination is even parallel to nonparallel and curved nonparallel; some beds show broad, low-angle hummocky cross-lamination and small-scale, scour-and-fill structures but lack shallow water sedimentary structures. A quartz arenite interval forms a key marker about 180 ft above the base of member 1. This interval thins gradually northward from about 80 ft at Elk Mountain (at south end of park) to 50 ft at Bear Mountain (near U.S.-Canada boundary). Member is about 440 ft thick.

Member 2 closely resembles member 1 except maroon beds are absent and laminae are generally thinner in member 1 than in member 2. Thin arenite beds, 2.5-7.5 cm thick, are common in the lower part. The lower contact is placed on top of the uppermost maroon sequence of member 1 and generally coincides with an increase in thickness of siltite laminae in member 2. In areas where maroon beds are absent, the contact between members 1 and 2 may be indistinguishable. Member 2 is about 540 ft thick.

Member 3 is about 540 ft thick and characterized by interlaminated and interbedded grayish-green siltite, yellowish-brown pyritic arenite, and lesser amounts of grayish-green argillite. Subaqueous shrinkage cracks, load structures, and mud-chip breccia are common and lamination is wavy nonparallel. The lower contact is placed at the base of the lowermost bed of pyritic arenite where the pyritic arenite and overlying beds are wavy laminated and contain numerous shallow-water sedimentary structures.

Member 4 is poorly exposed because outcrops are mostly cleaved and easily weathered. The member consists of thin to very thin laminae of olive siltite and thin lenticular beds of rusty-brown arenite that are commonly stained by iron and manganese oxides. Member 4 is about 440 ft thick and is commonly cleaved, folded, and sheared by thrust faults.

Member 5 contrasts sharply with member 4 and consists of bright-green argillite and lesser amounts of siltite. Lamination is wavy, within nonparallel, fining-upward silt-clay couplets. Mud-chip breccias, fluid-escape structures, and dolomite-filled subaqueous shrinkage cracks are common; it is about 195 ft thick.

The upper part of member 2 of the Appekunny Formation is exposed at this stop. The low-angle, hummocky cross-lamination is particularly unique to this part of the Appekunny Formation (Fig. 14). Note the broad, low hummocks looking along the exposures. Small clots and thin discontinuous laminae of pink ferroan calcite are common in some beds.

At the top of these roadcuts is a thick, light-green siltite that appears to have been rapidly deposited in a saturated state, based on abundant fluid-escape features that totally disrupted internal lamination, and the many asymmetrical synsedimentary faults, folds and slumps (Figs. 15, 16 and 17).

On east side of the Park, the Appekunny overlies disconformably the Altyn Formation (Fig. 13). The contact between the two is placed on top of the uppermost dolomite bed of the Altyn and locally shows as much as 2 m of erosional relief (Fig. 18). The thickness on the east side of the Park ranges from 1,740 to 2,265 ft.

70.3 Exposures on the left for the next 0.5 mi are the uppermost part of the Altyn Formation.

The Altyn Formation is exposed on the east side of the park and its easternmost exposures contain a unique eastern facies. The formation is completely exposed at Yellow Mountain (Fig. 1) and northward in the northeast part of the Park. In exposures south of Yellow Mountain, the base is not exposed and the formation is truncated by the Lewis thrust fault. In the Yellow Mountain area, the Altyn can be informally subdivided into 3 members designated 1 through 3 in ascending order (Jardine, 1985). At Yellow Mountain, the Altyn ranges in thickness from 780 to 840 ft.

Member 1 is mostly thin to thick (6 ft) beds of yellow- to orange-weathering, dark-gray to black dolomite and thin, lenticular interbeds of fine-grained arenite. Stromatolites are common in the lower part. The member is about 400 ft thick.

Member 2 is massive, medium- to thick-bedded, white to gray dolomite. Some medium- to coarse-grained, poorly sorted arenite beds are in the upper part. Stromatolites and dark-orange dolomite blebs occur locally in the lower part. The member contains contains black asphaltic veinlets near the Lewis thrust fault. The thickness of member 2 ranges from 190 to 225 ft.

Member 3 ranges from 180 to 200 ft thick and consists of interbedded and interlaminated light-gray to brownish-yellow dolomite, dolarenite, and arenite. Dolomite beds are 1-6 in thick; arenite

Figure 14. Low angle truncation of laminae by hummocky and swaley beds in member 2 of the Appekunny Formation near Bear Mountain in the north end of Glacier National Park.

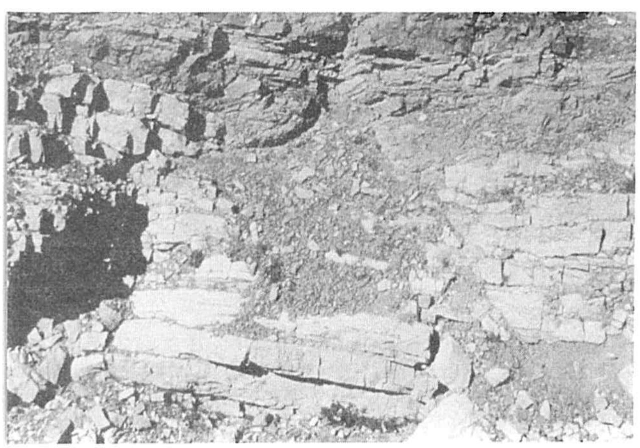
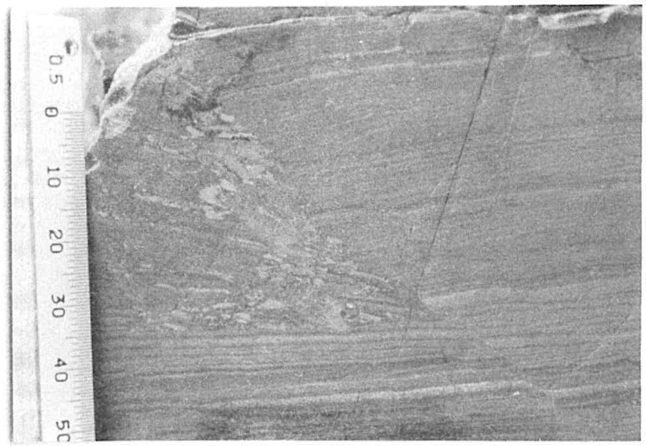
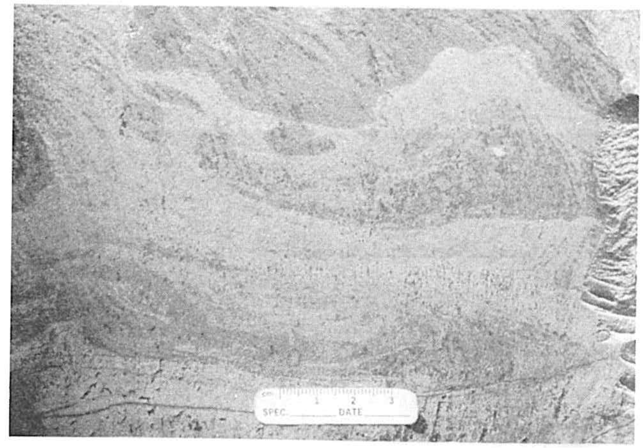


Figure 18. Contact of the light-colored Altyn Formation below and the Appekunny Formation above showing erosional unconformity between formations. Outcrops on east side of Spot Mountain.





Figures 15, 16 and 17. Soft sediment deformation associated with slumping of units in the upper part of member 2 at Stop 5.

beds are medium to coarse grained and commonly cross-bedded with some herringbone lamination. Stromatolites and styolites are common.

An eastern facies of the Altyn is partially exposed in thrust-fault plates in the Divide Mountain area (Stop 8). This facies is similar to the main body of the formation except the middle member (member 2) contains mostly thick beds of brownish-weathering, coarse-grained quartzite and minor interbedded argillaceous gray dolomite (Hill and Mountjoy, 1984). Low-angle cross-lamination is common.

71.6 Rising Sun Lodge and Campground is on the left - Saint Mary Lake on the right.

77.2 East entrance of Glacier National Park and visitor center.

77.7 Junction: U.S. Highway 89 and Going-to-the-Sun Road in Saint Mary. Turn left and proceed north on U.S. 89. The second part of the Day 2 roadlog starts here and heads south on U.S. 89.

79.0 Lower Saint Mary Lake is on the left.

85.0 Cross Saint Mary River.

86.3 Junction at the town of Babb. Road to left (west) goes to Many Glacier Hotel in the Park. Continue north on U.S. 89.

87.0 Babb School on left. From here to the next stop the route will cross and recross several thrust faults of minor displacement involving the Cretaceous Two Medicine, Virgelle, and Telegraph Creek Formations.

91.4 Cross Kennedy Creek.

91.6 Junction U.S. 89 and State Highway 17, turn left (northwest) onto Hwy 17. To the right U.S. 89 continues six miles to the International Border at Piegan Port of Entry.

93.2 Good exposures to the left of massive Cretaceous Virgelle Sandstone overlying Telegraph Creek shale at crest of the Babb anticline. The Babb anticline is in the hanging wall of a thrust fault that here overrides Cretaceous Two Medicine Formation.

93.8 Good view of Chief Mountain, a klippe of Belt strata, to the west. Massive, cross-bedded Virgelle Sandstone in roadcut.

95.7 Cross Roberts Creek.

96.6 Roadside rest area on the right. Chief Mountain overlook. Chief Mountain is in the foreground to the west, with Sherburne Peak in the background. At Chief Mountain, numerous west-dipping inbricate thrust faults of the Lewis thrust repeat strata of the Altyn Formation and part of the Appekunny Formation. Similar imbrications repeat strata of the Altyn at Yellow Mountain, to the south of Chief Mountain.

The Lewis thrust, although only locally exposed, can be inferred beneath talus and landslide deposits by the position of the Altyn Formation. Along the east side of the Park, the trace of the Lewis generally trends northwest and dips between 3 and 8 degrees southwest. The amount of easterly translation of the Lewis here was initially estimated to be at least 15 miles. At the south edge of the Park it is estimated that the Lewis moved eastward at least 40 miles, and in Canada, a minimum eastward movement of 32 miles is calculated.

Upper Cretaceous rocks underlie the Lewis in the eastern part of the Park. They are mostly thrust faulted and folded strata of the Marias River Shale and the Two Medicine Formation.

Much of the bedrock east of the Rocky Mountains is mantled by deposits related to multiple periods of alpine glaciation that advanced from cirques in the Park during early and middle Pleistocene time.

102.3 Leave Blackfeet Indian Reservation, enter Glacier National Park.

103.7 Parking area to left. Canada-United States International Boundary; Chief Mountain Port of Entry. For the next 2 miles Alberta Provincial Highway 6 follows the formation contact of Wapiabi (Marias River) and Belly River (Two Medicine) Formations. The Belly River forms the highlands to the east end the Wapiabi shale is in the valley to the west.

108.5 Belly River Campground to the left. Belly River sandstone crops out along the road.

108.8 Point of interest. Blood Indian Timber Reserve. Across the river lies a 4050 acre tract of land set aside in 1883 by Treaty No. 7 Northwest Territories. Bordered on three sides by Waterton Park, this land owned and administered by the Blood Indians is used by them for cattle grazing, logging and natural gas production.

109.7 Belly River Bridge. Enter Waterton National Park.

119.8 Junction with Highway 5. Bear left toward Waterton Park.

120.2 Cross Waterton River.

120.3 Junction Highways 5 and 6, entrance gate to Waterton Lakes National Park, turn left (southwest). The Park was set aside in 1895 and covers an area of 204 square miles along the eastern slope of the Rocky Mountains immediately north of the International Boundary.

121.0 Lower Waterton Lake on the left. Elevation 4,187 feet.

122.1 Point of Interest: Iceblock depression, the Lower Waterton Lakes occupy part of a huge trough which extends from the Waterton River Bridge to the Waterton townsite. This depression was formed during the last ice age when a block of ice broke away from the Waterton Glacier and became stationary. The surrounding active glaciers continued depositing glacial till against the ice block until the Ice Age ended and the block melted some 12,000 years ago, leaving this depression.

123.3 Junction. Red Rock Canyon road to the right. Continue straight ahead.

123.4 Cross Blakiston Brook

125.0 Driveway at the left to Prince of Wales Hotel built on Altyn Formation.

125.2 Waterton Park townsite. Altyn Formation in road cuts. End of Day 1.

DAY 2

Day 2 begins at the entrance gate to Waterton Lakes National Park (junction of Highways 5 and 6), travels north to the Pecten gas refinery and nearby Drywood Canyon to examine the Grinnell Formation and overlying units, and then returns to Whitefish, Montana, via U.S. Highway 2.

Reset odometer to 0.0 at gate.

The discussion and log for the Drywood Canyon area and Waterton gas field on this day was prepared by Pier L. Binda, Geology Department, University of Regina, Regina, SK, Canada, S4S 0A2 and Henry T. Koopman, 2205 22nd Street Southwest, Calgary, AB, Canada T2T

This excursion is intended to give an overview of the stratigraphy of the Grinnell and Siyeh Formations of the Purcell Supergroup. The lower part of the Siyeh Formation (Douglas, 1952; Price, 1966) is being equated with the Empire Formation (Whipple and others, 1984; Whipple, 1992; Link, this volume) and referred to as such in the discussion. Stop 6 (Fig. 19) will include examination of a reference section of a local facies of the Empire. At Stop 7, an outcrop of the Purcell Lava, which overlies the Siyeh Formation (Table 1), will be examined.

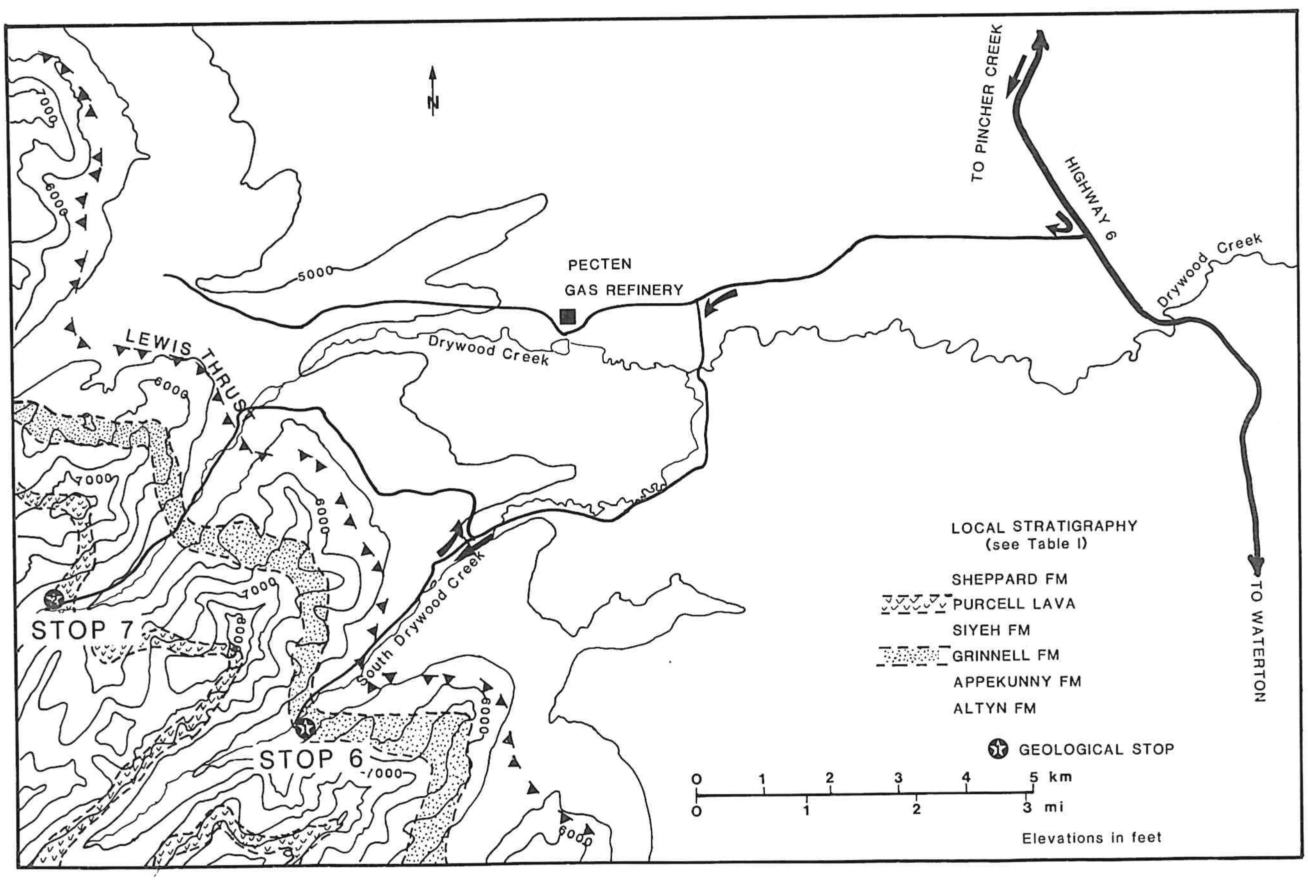


Figure 19. Sketch map of field trip area (Geology simplified after Price, 1966).

Other noteworthy geological features of the area include the Waterton Gas Field and the Lewis thrust.

Considerable parts of the following field guide are derived from field guide books by Gordy and others,(1982) and by Koopman (1985).

PHYSIOGRAPHY AND STRUCTURES

Regional

The Rocky Mountains of southern Canada can be divided into four physiographic and structural provinces termed from east to west: Foothills, Front Ranges, Main Ranges, and West Ranges. To the west, they are bounded by the Rocky Mountain Trench. To the east, the fold-and-thrust belt gives way to relatively undisturbed Phanerozoic sedimentary rocks of the Western Canadian Basin. The Rocky Mountains rest on a basement of Precambrian crystalline rocks which are the western continuation of the Canadian Shield.

The Foothills have a gently rolling topography and consist of multiple imbricates of Mesozoic clastics resting on Paleozoic carbonates cut by a few, flat to slightly westward dipping thrusts. The Front Ranges are formed by major thrust sheets stacked in imbricate fashion and involving mainly Paleozoic carbonates north of approximately latitude 51° N and Proterozoic Purcell-Belt rocks to the south. The Main Ranges consist of lower Paleozoic and Proterozoic rocks thrust eastward over the adjoining Front Ranges. The Western Ranges are characterized by tightly folded lower Paleozoic rocks.

Excursion Area

Neither the Main Ranges nor the Western Ranges of the Rocky Mountains extend south of the Crowsnest Pass (approximate lattitude, 49° 30' N). Therefore, at the latitude of the present field trip, only Foothills and Front Ranges occur between the Interior Plains and the Rocky Mountain Trench.

The sharp break between the gently rolling topography of the Foothills and the rocky high mountains to the west marks the surface trace of the Lewis thrust, the major structural feature of the region. The Lewis thrust extends some 225 km north and south from the International Boundary. In the excursion area above the thrust lies a slab of Proterozoic Purcell rocks which have been displaced some 80 km northeast relative to the underlying Upper Cretaceous molasse. The complex tectonics of the region are discussed by, among others, Price (1962), Mudge and Earhart (1980), and Fermor and Price (1987).

0.0 Junction, Waterton Lakes National Park gate, turn left (north) onto Alberta Highway 6.

1.8 Cross Galway Creek.

6.3 Crossing Indian Springs Ridge, good view of Waterton area at 8:00.

8.4 Dungarvan Creek bridge.

9.5 Spread Eagle Road to right.

11.6 Town of Twin Butte.

12.9 Yarrow Creek bridge.

16.0 Junction, turn left (west) onto paved road to Pecten gas refinery.

The Waterton Gas Field

The Waterton Field is the largest gas field of the southern Foothills of Alberta, and is also one of Canada's largest, with estimated ultimate reserves of about 57.7 x 10^9 m^3 (2.05 TCF).

The field was discovered as a result of regional seismic reconnaissance programs conducted by Shell in 1953 and 1954, which led to the recognition of a complex stack of at least three major thrust sheets involving Paleozoic carbonates. Subsequent drilling revealed reservoirs in the Mississippian Mount Head and Livingstone Formations and in the Devonian Palliser Formation.

The Waterton Sheet III Pool, whose axis lies west of this mountain front, is one of Alberta's most prolific reservoirs. Most of the wells in the southern part of the pool initially had open flow potentials in excess of 2.8 x 10^6 m^3/day, with some exceeding 28.2 x 10^6 m^3/day.

The gas plant, owned and operated by Shell Canada Resources, Ltd., was originally built in 1961 and expanded in 1967. A second plant was added in 1971. Combined product capacities quoted by Koopman (1985) are:

Raw Gas Intake (497 MMCFPD)	14.0 x $10^6 m^3$
Sales Gas (311 MMCFPD)	8.8 x $10^6 m^3$
Ethane (5,900 barrels/day)	940.0 m^3/d
Propane (2,770 barrels/day)	440.0 m^3/d
Butane (2,515 barrels/day)	400.0 m^3/d
Pentanes-Plus (12, 600 barrels/day)	2,000.0 m^3/d
Sulphur (3,197 tons/day)	2,900.0 t/d

Sales gas is contracted to Alberta and Southern Co. Ltd. and shipped by the Alberta Gas Truck Line (Nova) and other pipelines to markets in eastern Canada and the western United States. The propane is sold to local distributors. Butane is sold to refineries as feed stock for jet fuel and gasoline. Sulphur, an important by-product of the gas plant, is made into slates and shipped by railroad to Vancouver for international markets.

17.7 Valleys of South Drywood Creek and Drywood Creek to the southwest (Fig. 19).

Rocks of the Purcell Supergroup and its correlative, the Belt Supergroup in the United States, occur over an area of approximately 77,200 mi^2 in southwestern Canada and adjacent northwestern United States. Belt-Purcell rocks consist of 6-12 mi of alternating clastics and carbonates (Harrison, 1972). Regional stratigraphic correlation of these rocks is complicated by the intertonguing of sedimentary prisms from different source areas around the Belt-Purcell Basin. The Purcell Supergroup is Middle Proterozoic in age. The Purcell Lava, located in the upper third of the sequence, has been dated at 1,075 $\pm$ 65 Ma (Hunt, 1962), but thie date is almost certainly too young (see Raup and others, this volume). The degree of metamorphism of Belt-Purcell rocks increases from northeast to southwest: rocks along the eastern margin are virtually unmetamorphosed, those in the southwest are of greenschist facies to amphibolite facies (Harrison, 1972).

In the Clarke Range of southwestern Alberta, the Purcell Supergroup is represented by approximately 11,500 ft of exposed strata (Table 1 and Fig. 20) occurring as allochthonous sheets which have been thrust northeastward over the Cretaceous Belly River Formation.

19.8 Junction: turn left on dirt road leading to South Drywood Creek. As you follow this log to Stop 6 and encounter new roads or junctions, follow the signs that direct you to Shell well site #30.

23.8 Junction: Stay left on road up South Fork of Drywood Creek. You will return to this junction and take the road to the right which goes to Stop 7 in Drywood Creek.

As we enter the South Drywood Creek Valley (Fig. 19) there is a near-complete exposure of the section from the Appekunny Formation to the Siyeh Formation on both sides of the valley. Green and gray argillite and thinly-bedded green arenite of the Appekunny Formation contrast markedly with the overlying red argillite and white arenite of the Grinnell Formation. Above the Grinnell Formation, interbedded black and green argillite, thin arenite and buff-weathering dolostone of the Empire Formation (lower Siyeh) form a recessive slope. The middle Siyeh Formation (Helena Formation equivalent) consists of cliff-forming buff dolomite, gray algal limestone and gray argillite, which in turn are overlain by a recessive sequence of interbedded dolomite, arenite, and green and red argillite of the upper Siyeh Formation (Snowslip Formation equivalent). Above this sequence, the Purcell Lava forms a dark resistant cap.

26.0 Junction with dirt road to well site, turn left downhill.

TABLE I

Stratigraphy of the Purcell Supergroup

In southwestern Alberta

	thickness		lithology
gabbro			sill intrusions
KINTLA	3000 feet (900 m)		red, green and grey argillite and quartzite
SHEPPARD	600 feet (180 m)		grey dolomite and dolomitic argillite; recessive, brown weathering
PURCELL LAVA	200 feet (60 m)		dark purplish green, amygdaloidal; resistant, dark grey weathering
SIYEH	2000 to 3000 feet (600 to 900 m)	Upper (600 feet, 180 m)	interbedded dolomite, quartzite, algal limestone, green argillite; three bands of red argillite
		Middle (900 feet, 275 m)	massive to thickly bedded dolomite, molar tooth and algal limestone, grey argillite; resistant, cliff forming, grey weathering
		Lower (500 feet, 150 m)	interbedded grey dolomite, quartzite, green and black argillite; recessive, buff weathering
GRINNELL	750 to 1000 feet (230 to 300 m)		recessive, bright red argillite at the base, inter-bedded with green argillite and white, green and red quartzite and conglomerate in resistant upper part
APPEKUNNY	1100 to 1600 feet (330 to 500 m)	Upper (600 to 1000 feet, 180 to 300 m)	massive, green laminated argillite and thinly bedded green quartzite; with maroon and red argillite in the northeast
		Lower (500 to 600 feet, 150 to 180 m)	interbedded green quartzite, dolomite and quartz pebble conglomerate, green and red argillite; grading southwestward into massive, green, laminated argillite
ALTYN	500 to 1400 feet (150 to 425 m)	Upper (100 to 200 feet, 30 to 60 m)	thinly bedded, sandy, gritty dolomite, algal dolomite, black argillite
		Middle (200 to 400 feet, 60 to 120 m)	massive, cliff forming, sandy dolomite, algal dolomite, dolomite and quartz pebble conglomerate; light grey weathering; in the northeast brown weathering gritty dolomite at the base
		Lower (200 to 800 feet, 60 to 250 m)	thinly bedded, laminated, grey dolomite; light buff weathering, recessive
WATERTON	600 feet (180 m)		red, green grey dolomite and limestone

26.2 **STOP 6 - Grinnell and Empire Formations** (including the Drywood argillite member of the Empire Formation (Shell No. 30 Waterton LSD 5-34-4-1 W5). Park at well site.

Approximately 160 ft to the southwest is a small tributary of South Drywood Creek. Excellent exposures of the upper Grinnell and of the Empire (lower Siyeh) Formation can be examined along the tributary. A lithostratigraphic and geophysical log of the two formations is shown in Figure 21. Near the wellsite bright red argillite of the lower part of the Grinnell Formation can be seen cropping out at the junction of the tributary and South Drywood Creek. The lower part of the Grinnell Formation consists of 740 ft of predominantly bright red argillite, sparsely interbedded with thin quartz-arenite.

The upper part of the Grinnell Formation is exposed along the tributary upstream from the trail crossing. In this 250 ft-thick unit, beds of white quartz-arenite and green argillite become thicker towards the top. The upper part of the Grinnell Formation consists of repeated sequences of red argillite overlain by medium grained white quartz-arenite. Some of the thinner arenite beds pinch out over short distances, but thicker beds can be correlated over distances of a few miles. Sedimentary structures visible within these arenite beds include: small scale tabular cross-bedding, symmetrical (Fig. 22) and asymmetrical ripple marks with straight, sinuous and bifurcating crests (Fig. 23) which frequently have mud drapes with desiccation polygons preserved in the ripple troughs.

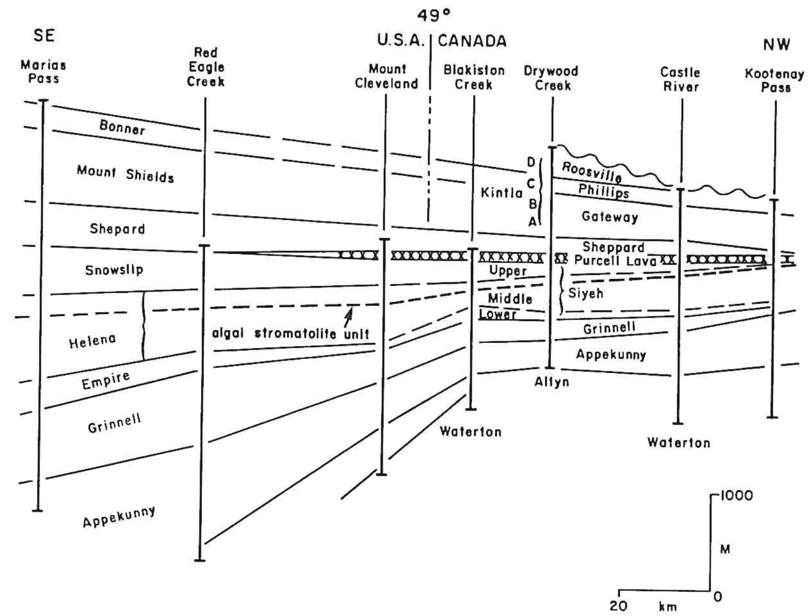


Figure 20. Correlation diagram of the Proterozoic formations between North Kootenay Pass, Canada, and Marias Pass, Montana (from Binda and others, 1989).

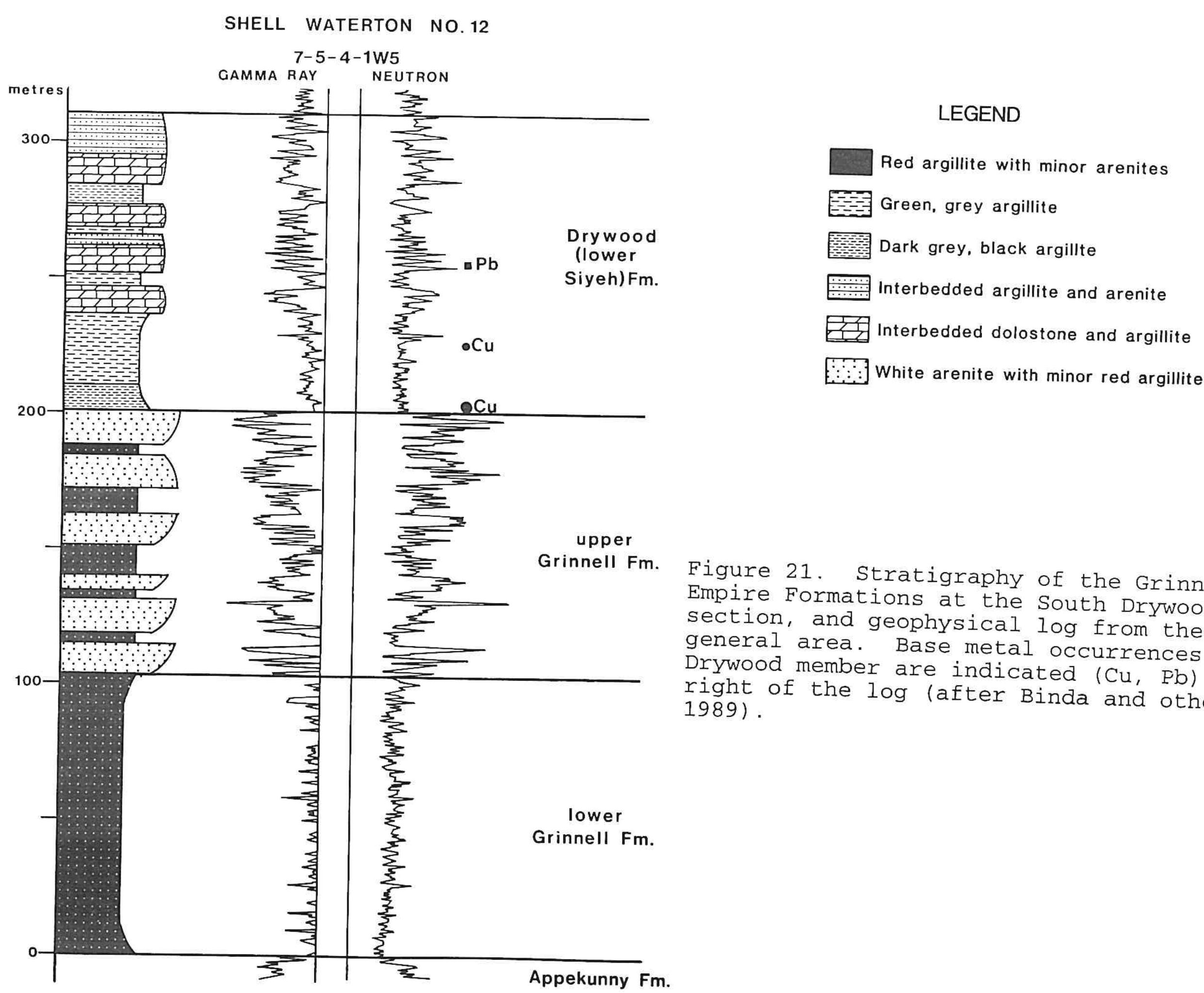


Figure 21. Stratigraphy of the Grinnell and Empire Formations at the South Drywood Creek section, and geophysical log from the same general area. Base metal occurrences in the Drywood member are indicated (Cu, Pb) to the right of the log (after Binda and others, 1989).

Cross-bedding measurements indicate multidirectional paleocurrents. Argillite beds display abundant desiccation cracks that have been infilled by sand (Fig. 24). Flakes of argillite, sometimes concave up and club ended, are lifted from desiccated surfaces and incorporated in arenite beds to form intraformational conglomerate.

The sedimentary sequences and associated structures suggest that deposition occurred on a low gradient mud flat, with possible tidal influence, which was cyclically desiccated and flooded by storms.

In the upper 100 ft of the Grinnell Formation, arenite beds become thicker, and together with green argillite, more abundant. The top of the formation is capped by a 13 ft-thick clean quartz-arenite, with few visible sedimentary structures other than some rippled surfaces. The increase in arenite content of this unit suggests that marine incursions became more frequent until a stable strandline was established.

The unit overlying the Grinnell Formation is herein called the Empire Formation; it was previously known as the lower part of the Siyeh Formation. The Empire Formation conformably overlies the Grinnell and consists of 395 ft of black and green argillite interbedded with thin beds of arenite and with buff-weathering dolomite higher up in the section. The lower part of this unit is lithologically distinct from the lower part of the Empire Formation in Glacier National Park (Whipple, 1980) and is herein designated as the Drywood argillite member of the Empire Formation. The lowermost 130 ft consists of black argillite passing upwards to green argillite. Rippled sand lenses up to 6 in thick are intercalated within these argillite beds. Birdfoot and linear syneresis cracks formed by subaqueous dehydration of mud are abundant throughout this lower 130 ft. Similar sedimentary structures have been reported by Whipple (1980) from the basal Empire Formation in Montana. Approximately 5 ft above the top of the Grinnell is a 1-2 ft limonite-stained, laterally continuous arenite bed which contains diagenetic sulfide ooids. This marker bed is also present on the Baring Creek Trail north of Going to the Sun Road in Glacier National Park (Link, this volume).

Overlying this unit is a 115 ft-thick carbonate-rich sequence composed of interbedded buff-weathering dolomite, green to gray argillite, carbonate tempestites, stromatolites and minor arenite lenses. Carbonate tempestites (Fig. s 25 and 26) are beds of intraformational conglomerate formed by storm waves which rip up partially consolidated shelf muds and deposit the mud flakes in mound-shaped bodies. These mounds pinch and swell to a maximum thickness of 2 ft and pinch out laterally within a few feet to laminated argillaceous dolomite. The stromatolites (Fig. 27) are bulbous (hemi-spherical) and up to 50 cm (20 in) high. They are predominantly dolomitic, although some are silicified.

A 33 ft-thick interval of interbedded green argillite and calcareous arenites separates this lower carbonate-rich unit from an overlying 80 ft-thick carbonate unit similar to that just described. The Drywood argillite member is capped by 50 ft of an interbedded sequence of green argillite and calcareous arenite.

Marine transgression over the broad mudflats of the Grinnell Formation resulted in deposition on a broad shallow shelf under reducing conditions, as indicated by the black argillite of the Empire Formation. Continued transgression of the sea across the shelf resulted in less restricted conditions and the deposition of the interbedded carbonates and argillite. The two arenite-rich intervals within the carbonate sequence may represent minor progradation of near shore sands. Above the last arenite-rich interval, carbonate sedimentation predominated, producing the cliff-forming carbonates of the Siyeh Formation (typical middle part of the Siyeh Formation of Price, 1966, and the Helena Formation of Glacier National Park).

The stratigraphic transition from the upper part of the Grinnell Formation to the Drywood argillite member of the Empire Formation is analogous to sequences that host some of the largest stratiform copper deposits of the world, e.g. Zambian Copperbelt and German Kupferschiefer. Geochemical analyses have revealed the presence of a 6.5-13 ft-thick, laterally continuous copper-rich zone of black argillite and gray arenite at the base of the Drywood member. Copper values tend to be higher in arenite than in associated argillite. The highest copper values do not exceed 4,000 ppm. Chalcopyrite is the main copper mineral, however malachite and azurite are visible on weathered surfaces.

Located within this copper-rich zone is a laterally continuous arenite bed approximately 5 ft above the base of the Empire Formation which contains diagenetic sulfide ooids composed of chalcopyrite and pyrite (Binda and others, 1985). The ooid-bearing bed can be traced 12 mi to the northwest and has been reported approximately 75mi to the south in Glacier National Park (Link, this volume). The presence of these sulfide ooids and of rims of chalcopyrite around argillite clasts in clean quartz arenite at the top of the Grinnell Formation indicate movement of copper-rich solutions in the porous sands and precipitation of sulfides at permeability barriers or favorable nucleation sites (Binda and others, 1989).

Return to road junction with Drywood Creek.

28.6 Junction: Turn left, north, on road to Drywood Creek.

29.5 Bathing Lake on left.

30.1 Bear right of road junction.

31.5 Cross Drywood Creek.

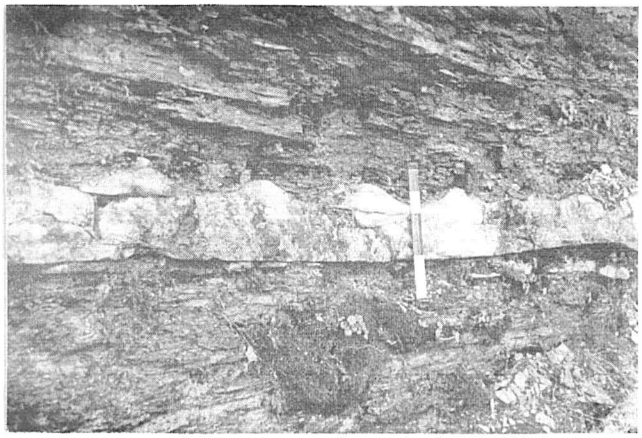


Figure 22. Symmetrically rippled quartz-arenite of the upper Grinnell Formation.

Figure 23. Rippled upper surface of quartz-arenite in the upper Grinnell Formation.

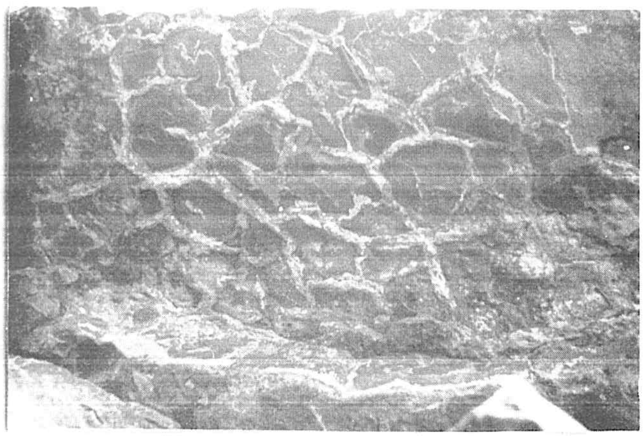


Figure 24. Casts of desiccation cracks on the underside of an arenite bed, upper Grinnell Formation.

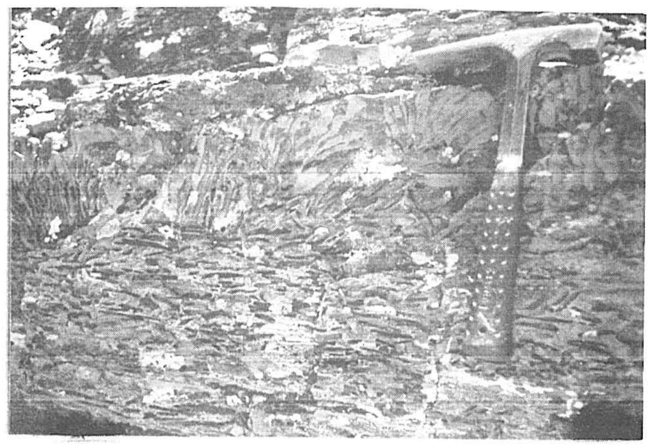


Figure 25. Carbonate intraformational conglomerate (tempestite) of the Drywood member, Empire Formation. Note change in clast orientation from base to top.

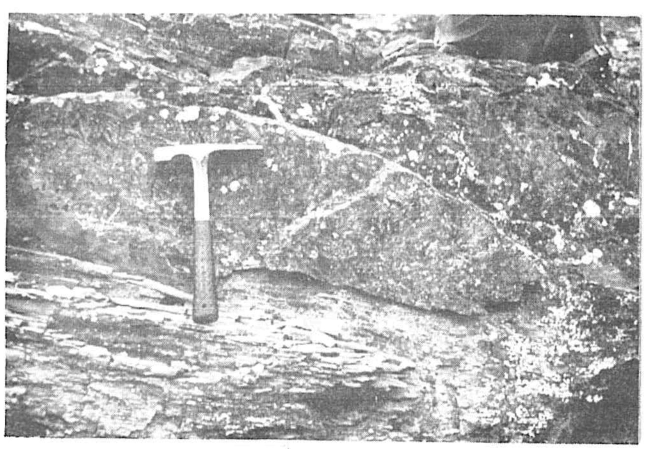


Figure 26. Carbonate tempestite mount within dolomitic argillite of the Drywood member, Empire Formation.

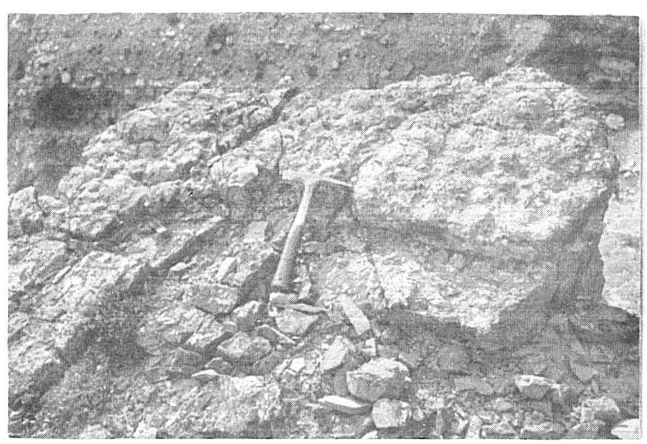


Figure 27. Small bulbous stromatolites in argillaceous dolomite of the Drywood member, Empire Formation.

31.7 Junction: Turn left up Drywood Creek road (you will return to this point and take the road to the right to return to the main highway) and proceed about 3 mi to Stop 7 at the end of the road at Shell No. 12 Waterton well site.

32.1 Cross Drywood Creek.

33.4 Cross back over Drywood Creek to the north side.

Entering the Drywood Creek Valley we again see a near complete exposure of the Appekunny to Siyeh Formations on both sides of the valley.

34.6 **STOP 7 - Purcell Lava.**

The Purcell Lava forms the ledge at the waterfalls 1,300 ft west of the Shell No. 12 Waterton well site. It consists of 200 ft of dark green and reddish or purplish green amygdaloidal andesite. Pillow structures are visible in the basal part of the flow, while the upper surface displays a ropey (pahoehoe) texture, suggesting a change from subaqueous to sub-aerial flow.

Above and below the Purcell Lava are red argillite beds of the Sheppard (Shepard) and upper Siyeh (Snowslip) Formations respectively.

From Stop 7, return to the junction at mileage 31.7.

37.5 Junction: Proceed straight ahead (road to the right returns to Stop 6).

38.5 Junction: Turn right on gravel road and head east toward the gas plant.

41.0 Junction with paved road at the Waterton gas plant - turn right and return to main highway, Alberta Highway 6.

Retrace field trip route to the junction of U.S. 89 and Going-to-the-Sun Road at Saint Mary (see Day 1 log). At this point reset odometer to 0.0 for the last leg of the field trip.

0.0 Junction: U.S. Highway 89 and Going-to-the-Sun Highway to Saint Mary. Road to the right goes 51 miles to West Glacier via Logan Pass. Continue south on U.S. 89.

The discussion below and at Stop 8 is authored by G.A. Davis in Whipple and others, 1985.

Looking west into the Park from Saint Mary, we can see the Lewis allochthon in perhaps its most simple geometric development on the front of East Flattop Mountain (Fig. 1). A horizontal sequence of Appekunny and Altyn strata (greenish and yellowish colors respectively) overlies Cretaceous Marias River Shale which is concealed by forested slopes, talus and landslide deposits. The widespread, but misleading, popular notion of the Lewis allochthon as an essentially undeformed slab lying above Cretaceous strata probably was derived from the atypical East-Flattop-Mountain segment of the thrust fault as seen here. Two, steep southwest-dipping reverse faults can be seen offsetting white-weathering quartz arenite marker beds in the Appekunny Formation and its contact with underlying Altyn strata. Stratigraphic displacement on the most important of the two faults is approximately 200 ft. Geologic mapping elsewhere in the allochthon, for example, on Curly Bear Mountain south of Saint Mary Lake and west of Divide Mountain, indicates that these reverse faults formed prior to emplacement of the Lewis allochthon in this area. Hence, they are regarded as transported structures which are truncated downwards by the Lewis thrust fault. The thrust is not exposed beneath East Flattop Mountain, but its location is closely approximated by the base of cliffs developed in Belt strata.

5.8 **STOP 8 - Divide Mountain viewpoint.**

Divide Mountain (elev. 8,665 ft) (Fig. 1) is a klippe of Belt strata that overlies the Upper Cretaceous Two Medicine Formation on the boundary between Glacier National Park and the Blackfeet Indian Reservation. Like the much better known Chief Mountain klippe farther north, Divide Mountain consists of two major upper-plate structural assemblages: (1) an upper, essentially undeformed sequence of Belt strata, here Appekunny and Altyn Formations, overlying (2) a complexly deformed assemblage of lower Belt strata of still uncertain stratigraphic assignment. Figure 28 is a line-drawing of the southeastern flank of Divide Mountain. Dark-green argillite of the Appekunny Formation caps the peak and overlies cream-to orange-colored carbonate strata of the Altyn Formation. The Divide Creek thrust fault separates this undeformed sequence from an intensely imbricated and folded assemblage of carbonate and clastic strata below, which appears to be much more variegated in color. This lower sequence is interpreted to be an easterly facies of the Altyn Formation. The Divide Creek thrust may represent the roof thrust of a duplex fault zone, the floor thrust being the Lewis fault. The two faults merge to the southwest beneath Curly Bear Mountain, but their possible merger northeastward in the direction of thrust transport--requisite for duplex core geometry--is not demonstrable because of the present erosional configuration of upper plate rocks. Geologic mapping (G.A. Davis, unpub.) indicates that the Divide Creek thrust fault formed very late in the emplacement of the Lewis allochthon, and postdates, or at least outlasted, movements along imbricate faults below it and above the Lewis thrust. Between Divide Mountain and Curly Bear Mountain, the Lewis thrust has a synformal configuration and appears to have been broadly folded along a northwest-southeast trend prior to hindward development of the higher, very planar Divide Creek thrust fault.

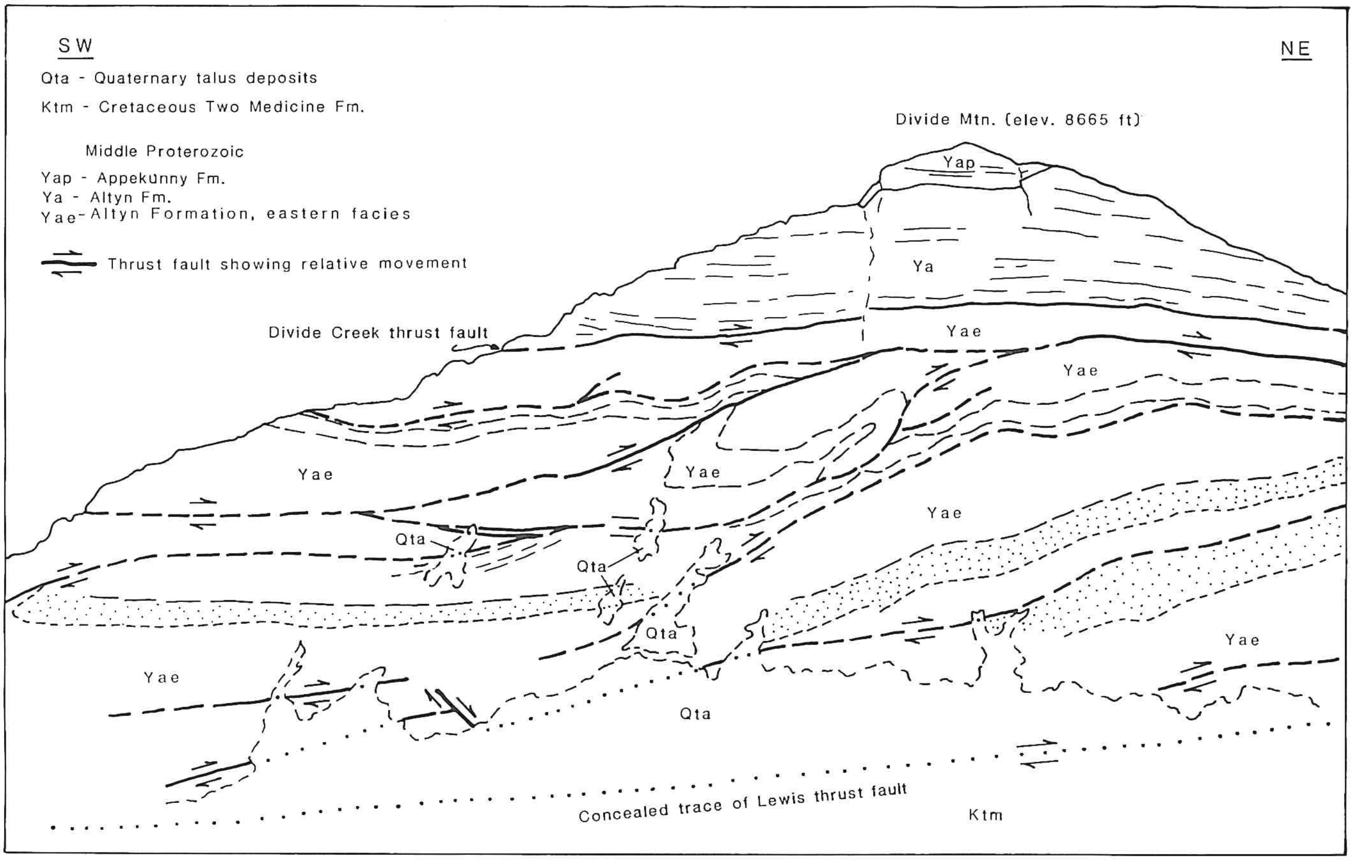


Figure 28. Line drawing of the southeast side of Divide Mtn. showing geology of the Divide Mtn. duplex Modified from drawing by G.A. Davis, in Whipple and others, 1985

19.0 Kiowa Junction: Turn right (southwest) onto Montana Highway 49.

20.4 Deformed beds of Upper Cretaceous Telegraph Creek Formation and Virgille Sandstone are exposed in roadcuts.

22.3 On the left side of the road are deformed glacial deposits of interbedded conglomerate and silt.

28.0 Cross Two Medicine River.

30.7 Junction: U.S. Highway 2 and Montana Highway 49 in East Glacier, turn right (southwest) on U.S. 2.

Above the railroad tracks to the north stand cliffs of the Lewis thrust plate above gentler, tree-covered slopes of Mesozoic shale and sandstone. The Lewis thrust plate is one of the most famous structural features in North America. It was first described by Willis (1902), who pointed out that the resistant rocks forming the spectacular scenery of Glacier National Park are Precambrian sedimentary Belt rocks that have ridden eastward over much younger Paleozoic and Mesozoic rocks. On the basis of geophysical studies, geologists believe that Paleozoic rocks below the exposed Mesozoic rocks project beneath the Lewis plate as far west as the Rocky Mountain trench. Therefore the rocks of the Lewis plate were deposited west of the Rocky Mountain trench, and have been thrust perhaps as much as 40 miles eastward to their present position (Mudge and Earhart, 1980). The Lewis thrust plate carries, in its lower part (Fig. 1b, 1c), buff-weathering, cliff forming limestone of the Altyn Formation, overlain by the darker, greener beds of the Appekunny Formation. From here to the summit of Marias Pass, the Lewis plate is spectacularly exposed. The highway crosses Mesozoic rocks, which underlie the projected plane of the Lewis plate and have been folded and thrust at a much smaller scale. The town of East Glacier Park rests on marine shale of the Upper Cretaceous Marias River Formation.

42.0 **STOP 9 - Lewis thrust fault and geologic features immediately north of Marias Pass**

(From Kelty, 1984).

"The precipitous slopes of Little Dog Mountain rise directly to the north of Marias Pass. Upon these south facing slopes is one of the most outstanding continuous exposures of the Lewis thrust fault (Fig. 29). The Lewis thrust can be traced from Little Dog Mountain eastward approximately 5.5 mi to Calf Robe Mountain by locating the subhorizontal orangish-white Altyn Formation (100-130 ft thick) which almost always lies directly above or immediately adjacent to the Lewis thrust fault. The approximate attitude of the Lewis thrust along this segment of the fault is N 68° E, 12° NW.

Note that even from a distance, the Altyn Formation changes thickness. These changes in thickness usually coincide with imbrication or extensional faulting that either thickens or thins the Altyn Formation, respectively. Imbrication has been observed to triple the normal stratigraphic section of the Altyn Formation in a few localities. Thinning of the Altyn Formation along the Lewis thrust fault usually corresponds to movement on listric extensional faults that formed at the time of movement along the Lewis.

Above the Lewis thrust fault, on the steep south-facing slopes of Summit Mountain and Little Dog Mountain, is a system of relatively steeply dipping, imbricate faults that have kinematic indicators that suggest a direction of movement to the south. The major lines of evidence for southward directed movement comes from large scale folds which are related to thrust faulting along the Lewis. These large scale folds have axes that trend east-west. Geologic mapping of this system of thrust faults has not resolved whether the folds are related to south-directed thrust faulting or are "mega-drag" of the once more north-south oriented structures.

Above the east-west trending system of structures, lies the highest structure exposed on the slopes of Summit Mountain and Little Dog Mountain: The Ole Creek detachment. The Ole Creek detachment is a shallow-dipping, east-directed fault in the upper exposures of the greenish Appekunny Formation which overlie the Altyn Formation. Above this fault are preserved some of the least disrupted rocks along this segment of the Lewis.

Exposed below the Lewis thrust fault are intensely deformed Cretaceous rocks. These rocks are best exposed between talus slopes below Little Dog Mountain. The most obvious structural features within the Cretaceous rocks are: (1) a large recumbent fold and (2) structural discontinuities between rock packages. The recumbent fold has a fold axis that trends 15° N, 10° W (approximately perpendicular to the direction of transport along the Lewis thrust fault: N 65° to 70° E). The structural discontinuities within the Cretaceous rocks, upon closer inspection, are a system of compressional faults that are truncated by the Lewis."

47.6 Approximate location of the Blacktail fault. The Blacktail fault (Fig. 1c) is a listric normal fault that merges along strike with the Flathead-Roosevelt fault system about 30 mi north of here and has been interpreted to either truncate the Lewis thrust (Childers, 1963; Whipple, compiler, 1992), or merge with the Lewis at depth (Constenius, 1982). The fault has about 12,800 ft of stratigraphic displacement at this point and places rocks of the McNamara Formation against rocks of the Appekunny Formation.

48.3 Snowslip, Montana on the right. Mountains on the north side of the drainage are the type locality of the Snowslip and Mount Shields Formations.

50.2 Tranquil Falls on left cascades over dolomitic outcrops of the Helena Formation.

53.9 About here, the road crosses the Roosevelt fault (see Fig. 1) which is another listric normal fault similar to the Blacktail fault. East-dipping strata of the Shepard Formation are down dropped to the west against the upper part of the Grinnell Formation. Much of the fault is covered by glacial till in this area.

54.4 Confluence of Bear Creek and Middle Fork of the Flathead River. At this location, the highway turns northwestward and proceeds north down the Middle Fork of the Flathead River.

54.7 Railroad overpass. Entering Glacier National Park boundary. The first white man to penetrate the spectacular region of present-day Glacier National Park was probably the Hudson Bay Company trapper, Hugh Monroe, in 1815. The Blackfeet say that Father DeSmet visited the region in 1846 and named the Saint Mary lakes. Later, explorer John Stevens discovered Marias Pass in 1889, thus linking east-west access across the Continental Divide. The Great Northern Railroad, realizing the lucrative tourist potential of the scenery, completed its rail line in 1892. Ultimately, on May 11, 1910, Glacier National Park was created by an Act of Congress.

55.2 The road crosses the Roosevelt fault again and proceeds back onto the up thrown block of Grinnell, east of the fault.

56.1 Turnoff to Goat Lick overlook. Exposures in roadcut are brecciated Grinnell Formation.

56.6 At this point, the road again crosses to the western, hanging wall side of the Roosevelt fault system, but here the fault consists of two segments. The easternmost segment separates the

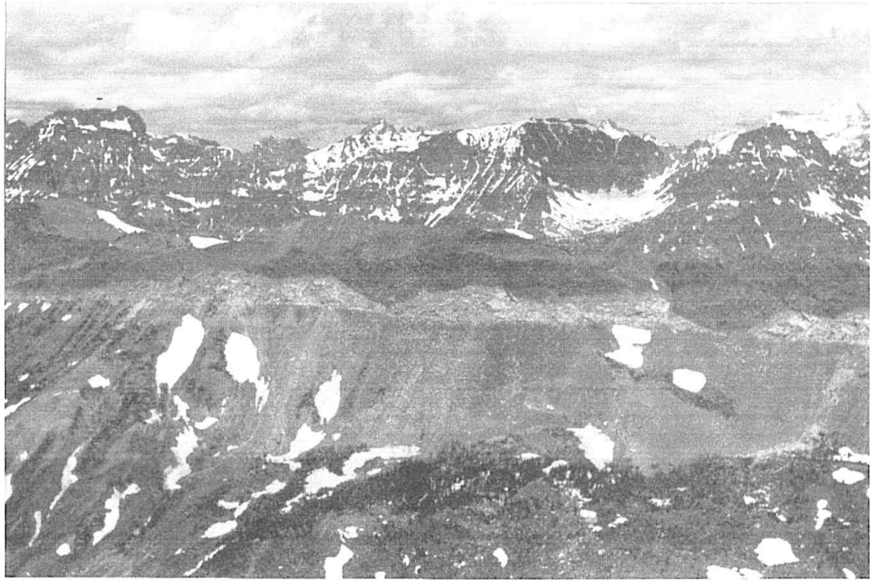


Figure 29. Aerial view of the Lewis thrust fault looking north near Marias Pass. The trace of the Lewis is just below light-colored rock of the Altyn Formation in the center of the photo.

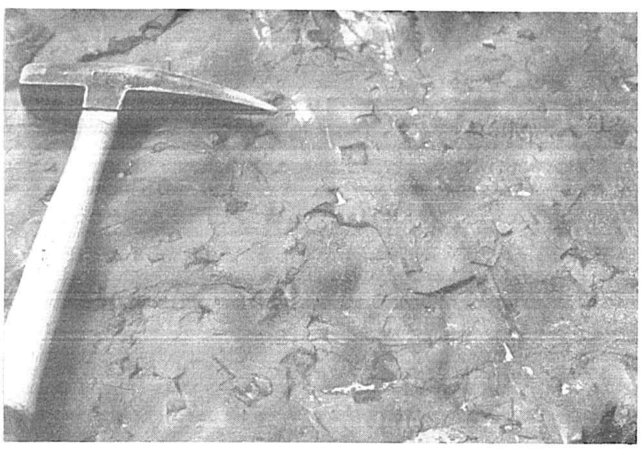


Figure 31. Salt casts in red argillite of member 3 of the Mount Shields Formation.

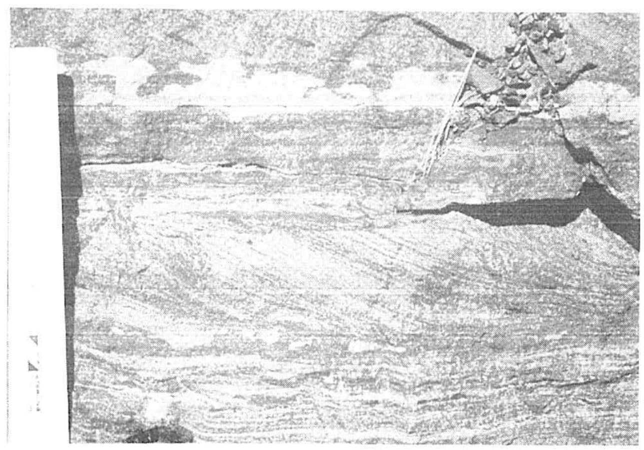


Figure 32. Cross-bedded arenite in lithofacies of the Bonner Quartzite at Stop 10.

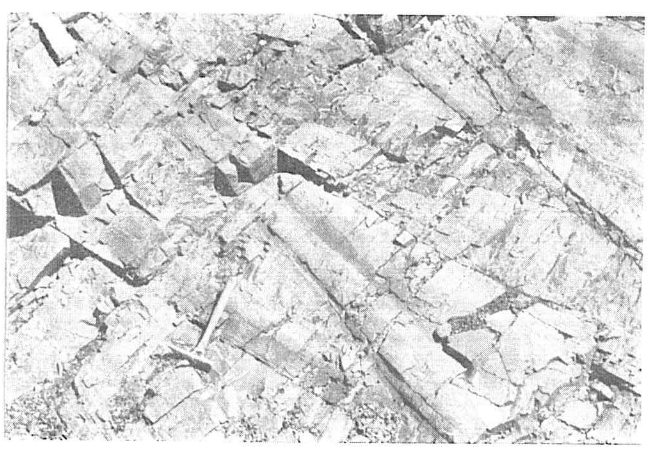


Figure 33. Thick beds of channeled quartzite in the Bonner at Stop 10.

lowermost part of the Empire Formation from a down dropped block of Helena Formation, which in turn is faulted by the remaining segment against the down dropped middle part of the Snowslip Formation 0.4 mi further along the highway. The stratigraphic displacement on the fault in this area is about 7,200 ft. Natural salt licks on the river bank below at the Goat Lick viewpoint are located near the traces of these faults and attract mountain goats to this area. Analysis of material at the lick sites indicates that the principal salt is calcium sulfate which has probably been remobilized from evaporites in the fault-bounded Kishenehn Formation to the north.

57.2 Lower part of the Snowslip Formation (Members 1 and 2).

58.2 Leave Glacier National Park boundary; cross Middle Fork of the Flathead River. Highway climbs to the top of a high river terrace underlain by rocks of the Mount Shields and Shepard formations.

58.9 Paved road to Essex, Montana. Just north of here, the Nyack and Roosevelt faults merge and form the southernmost juncture of the Tertiary basin (see Fig. 1).

62.2 Cross over the Nyack fault into Tertiary sedimentary rocks.

66.1 Exposures of the Tertiary Kishenehn Formation in roadcut on the left.

68.5 Approximate location of the Nyack fault. This fault is the structural boundary for the west side of the Kishenehn basin, and so at this point, we pass from Tertiary rocks to Proterozoic rocks . The Nyack fault has been interpreted to be an antithetic normal fault to the Flathead-Roosevelt fault system (Constenius, 1982). At the surface, the fault dips steeply to the northeast, but is inferred to be progressively overturned with increasing depth (Constenius, 1982).

70.7 **STOP 10 - Missoula Group strata**

The upper part of the Mount Shields Formation, all of the Bonner Quartzite, and the lower part of the McNamara Formation are exposed in a series of roadcuts at this stop. A wide turnout and parking area is on the left at the far end of the roadcuts.

At type section near Mount Shields, the formation is informally subdivided into five members designated 1 through 5 in ascending order; it has a maximum thickness of about 2,800 ft. (Whipple and others, 1984; Whipple, 1992.)

Member 1 consists of thinly laminated, maroon to pale-purple argillite, brick-red siltite, and some interbedded arenaceous siltite and thin intervals of greenish-gray siltite and argillite. The lower contact is placed on top of uppermost sequence of dolomitic siltite of the Shepard Formation. Near the northern boundary of the Park, member 1 encloses an interval of basaltic lava about 35 ft thick. Member 1 is about 100 ft thick.

Member 2 consists mostly of thin, fining-upward successions of brick-red, very fine-grained arenite and coarse-grained siltite capped locally by dark-red argillite; member 2 contains more arenite than the other members and is about 885 ft thick. Ripple cross-lamination and some even, parallel lamination is common in the lower part of successions. Pink to cream limestone beds at the top of the member contain oolites and small stromatolite heads; this zone is recognized throughout northern part of Belt basin.

Member 3 is mostly couplets of siltite and argillite. Siltite laminae in couplets successively change color upward from brick red in the lower part of the member to purplish gray in the middle part to dark grayish green near the top; argillite laminae in couplets remain dark red to pale purple throughout the member. Salt casts (Fig. 31) and ripple marks are common, but salt casts become less abundant downward as arenite beds increase and argillite beds decrease. The lower part contains pale-purple to brick-red, very fine-grained arenite similar to that in member 2 but in equal proportion to siltite and argillite. Member 3 is the thickest (1,475 ft) member of the Mt. Shields Formation throughout Glacier National Park.

Member 4 consists mostly of grayish-green, fining-upward couplets of siltite and argillite and contains carbonate mostly as cement in siltite. Salt casts are common, particularly in the lower part. The member is about 55 ft thick at the type section but appears to thicken and contain more carbonate beds northward in the park.

Member 5 is about 30 ft thick and is characterized by very thinly laminated, blackish-green argillite and some thin, lenticular beds of arenaceous siltite that are more abundant near top of member. This distinctive succession of blackish-green argillite is locally calcareous and sharply overlain by pale-green, coarse-grained, poorly sorted feldspathic arenite of the Bonner Quartzite.

At this stop, couplets of pale-purple argillite and dark-green siltite, which compose most of member 3, are in contact with member 4 at the extreme west end of the roadcut. Salt casts are abundant in member 3 and less so in member 4. The contact between 3 and 4 is transitional and is placed at the base of the first grayish-green siltite and argillite sequence that contains carbonate. Here, member 4 is thin, but to the northwest, it thickens to about 800 ft and contains several beds of dolomite (Whipple, 1984).

Small-scale fluid-escape structures and subaqueous shrinkage cracks are common. The contact between members 4 and 5 is placed at the base of a sequence of thinly-laminated, olive to blackish-gray argillite and siltite that is commonly stained by limonite. Cryptalgal lamination, carbonate cement, and thin lenticular beds of arenite are present locally in member 5.

The Bonner Quartzite in this area is about 800 ft thick. The base of the formation is sharp and placed at the at the base of the first bed of feldspathic arenite that overlies the blackish strata of Mount Shields 5. The Bonner consists dominantly of pinkish-gray to pale-red, very fine- to medium-grained feldspathic arenite and lesser amounts of interbedded siltite and dark-red argillite that commonly occur as rhythmic, fining-upward sequences as much as 10 ft thick. Five lithofacies are recognizable in the exposures at this stop, several of which are shown in Figure 32:

1) Cross-bedded arenite -- pink to maroon subfeldspathic and feldspathic arenite (middle bed, Fig. 32); fine to medium grained; planar cross-bedding; mudchips common on foresets and at base of lithofacies.
2) Even parallel-laminated arenite -- pink to red, fine-grained feldspathic arenite; very thin mudchips along laminae.
3) Faint, wavy parallel-laminated arenite -- dark red, very fine-grained arenite to coarse siltite; commonly contains buff-colored bleach marks (upper part, Fig. 32).
4) Ripple cross-laminated arenite -- pink to dark red, fine to medium-grained feldspathic arenite; ripple marks and mudchips common (lower part, Fig. 32).
5) Thinly-laminated argillite -- dark-red argillite, discontinuous lamination common; top of lithofacies eroded.

Fining-upward sequences begin generally with lithofacies 1 and become finer-grained upward in numerical order, but commonly do not contain all 5 lithofacies. Generally, the cross-bedded arenite lithofacies is the most abundant, particularly in the middle part of the Bonner, where it appears to form broad shallow channels (Fig. 33). The sequences are interpreted to have formed in a distal fluvial (alluvial apron) environment during sheet-flood events and wanning-current flows (Winston, 1984).

The McNamara Formation is exposed in roadcuts at the far east end of this stop. The contact between the Bonner and the McNamara is not exposed. For the most part, the McNamara in this area consists of grayish-green siltite and argillite and locally interbedded oolitic and stromatolitic limestone, calcareous arenite, and mud chip breccia. Silicified mudchips and discontinuous laminae are characteristic of the McNamara.

77.5 Steeply dipping exposures on the left are the Shepard Formation.

The Shepard Formation in Glacier National Park typically consists of yellowish-gray to greenish-gray dolomitic and pyritic siltite and argillite and minor thin beds of coarse-grained calcarenite, quartz arenite, limestone, and dolomite. Thin beds of stromatolitic limestone are common in the southern exposures but rare in the northern part of the Park. Lamination is generally wavy, non parallel, and composed of fining-upward couplets. Fluid-escape structures, shrinkage cracks, ripple marks, miniature molar-tooth structures, and mud-chip breccias are common in the Shepard. Because of the carbonate and pyrite content of the strata, most exposures weather tan to dusky orange.

The Shepard in Glacier National Park ranges considerably in thickness; it thins from 1,550 ft at the southern edge to 550 ft at the northern edge.

79.0 Roadcuts expose the upper part of member 5 and most of member 6 of the Snowslip Formation. The contact between member 6 and the Shepard Formation is inferred to be near the east end of the north roadcut.

80.1 Highway curves to the west. Contact between the Helena Formation and the Snowslip Formation (Missoula Group) is exposed in the roadcut. The contact is placed on top of the uppermost oolitic limestone bed which is overlain by beds of calcareous siltite and lithic arenite of the Snowslip.

80.2 Cross Ousel Creek.

83.6 Roadcuts in the broad curve expose the uppermost part of the Empire Formation. The Empire in this area is about 1,200 ft thick and consists primarily of interlaminated argillite and siltite arranged commonly as fining-upward couplets less than 1 in thick. Carbonate cement is present in the uppermost part, and thin quartz arenite beds are common in the lower part of the formation. At the east end of the roadcut, the Empire is in contact with the overlying Helena Formation. The contact is placed at the lowest bed (about 8 in thick) of calcareous quartz arenite above which dolomitic siltite beds of the Helena Formation contain numerous small molartooth structures or crinkly segregations of gray, sparry calcite.

84.7 Steeply dipping to slightly overturned beds in roadcut on the left are the lowermost part of the Empire Formation that pass downsection into the uppermost part of the Grinnell Formation. The Grinnell is the oldest red-bed assemblage in the Belt succession.

85.3 Junction of U.S. 2 and road to Glacier National Park. This is the end of the field trip log. Proceed straight ahead on U.S. 2 and return to Whitefish and Grouse Mountain Lodge.

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TABLE I

Stratigraphy of the Purcell Supergroup

in southwestern Alberta

	thickness	lithology	
gabbro		sill intrusions	
KINTLA	3000 feet (900 m)	red, green and grey argillite and quartzite	
SHEPPARD	600 feet (180 m)	grey dolomite and dolomitic argillite; recessive, brown weathering	
PURCELL LAVA	200 feet (60 m)	dark purplish green, amygdaloidal; resistant, dark grey weathering	
SIYEH	2000 to 3000 feet (600 to 900 m)	Upper (600 feet, 180 m)	interbedded dolomite, quartzite, algal limestone, green argillite; three bands of red argillite
		Middle (900 feet, 275 m)	massive to thickly bedded dolomite, molar tooth and algal limestone, grey argillite; resistant, cliff forming, grey weathering
		Lower (500 feet, 150 m)	interbedded grey dolomite, quartzite, green and black argillite; recessive, buff weathering
GRINNELL	750 to 1000 feet (230 to 300 m)	recessive, bright red argillite at the base, interbedded with green argillite and white, green and red quartzite and conglomerate in resistant upper part	
APPEKUNNY	1100 to 1600 feet (330 to 500 m)	Upper (600 to 1000 feet, 180 to 300 m)	massive, green laminated argillite and thinly bedded green quartzite; with maroon and red argillite in the northeast
		Lower (500 to 600 feet, 150 to 180 m)	interbedded green quartzite, dolomite and quartz pebble conglomerate, green and red argillite; grading southwestward into massive, green, laminated argillite
ALTYN	500 to 1400 feet (150 to 425 m)	Upper (100 to 200 feet, 30 to 60 m)	thinly bedded, sandy, gritty dolomite, algal dolomite, black argillite
		Middle (200 to 400 feet, 60 to 120 m)	massive, cliff forming, sandy dolomite, algal dolomite, dolomite and quartz pebble conglomerate; light grey weathering; in the northeast brown weathering gritty dolomite at the base
		Lower (200 to 800 feet, 60 to 250 m)	thinly bedded, laminated, grey dolomite; light buff weathering, recessive
WATERTON	600 feet (180 m)	red, green grey dolomite and limestone	

GUIDE TO THE GEOLOGY OF THE NORTHERN WHITEFISH RANGE, MONTANA

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INTRODUCTION

The purpose of this road log is to discuss the geology of the northern Whitefish Range at selected stops with reference to regional structures and stratigraphy. Stops have been selected to illustrate typical exposures of Middle Proterozoic and Paleozoic sedimentary rock units in this part of the Rocky Mountains. These rock units and the various thrust faults that shuffle them are part of the overthrust belt known to contain hydrocarbons just north of the 49th parallel in Canada.

This field trip guide begins and ends at Grouse Mountain Lodge in Whitefish, Montana and takes a full day to complete. The roads traveled are mostly gravel and in some places quite rough, but for the most part, they are suitable for passenger cars. The first part of the route travels up the North Fork of the Flathead River past a few exposures of Belt rocks in the southern part of the range (Fig. 1). No stops are made until you turn west up Trail Creek approximately 50 mi up the North Fork.

During the last decade, exploration in this region for oil and gas resulted in several seismic lines being run across the Whitefish Range. Much of the road log follows one of the more frequently shot seismic lines that generally follows a narrow, one-lane mountain road up Trail Creek, over a pass south of Bald Mountain, and down to Grave Creek (Fig. 1). You should be very cautious of on-coming vehicles. The latter part of the field trip deviates some from the seismic route by traveling up Grave Creek and into the Weasel Creek drainage to examine stratigraphic units in the upper part of the Belt Supergroup. After the stops in the Weasel Creek drainage, the field trip route proceeds down Grave Creek to the junction of U.S. Highway 93. The return to Grouse Mountain Lodge is south down the Tobacco Valley along U.S. 93. No stops are made along this part of the route; however, you can refer to the field trip log by Harrison and others in this volume for additional discussion of the geology.

The Whitefish Range extends from the 49th parallel southeast about 60 mi to Columbia Falls in the Flathead Valley and is flanked on the west by the Rocky Mountain Trench and on the east by the Kishenehn basin, a Tertiary graben (Fig. 1). The geology of the northern Whitefish Range consists generally of a sedimentary and igneous rock sequence at least 34,000 ft thick, ranging from Middle Proterozoic to Early Tertiary in age (see stratigraphic table, Fig. 2), that is structurally rearranged by thrust faults and normal faults. The sedimentary rocks are mostly Middle Proterozoic Belt Supergroup overlain locally by Paleozoic shallow marine strata (figs. 2 and 3), Mesozoic, or Lower Tertiary strata. Igneous rocks consist of the Purcell Lava in the upper part of the Belt Supergroup and several sills and dikes of probable Proterozoic age.

Stratigraphic terminology used in this log combines nomenclature used in the United States for Belt Supergroup strata and a mixture of U.S. and Canadian terms for Paleozoic and Mesozoic units. Descriptive rock terminology comes mostly from field observations made during geologic mapping. Detailed descriptions, stratigraphic correlations, and discussion of Belt Supergroup rocks in the northern Whitefish Range are presented by Whipple (1984) in a report on the Ten Lakes Wilderness Study Area. For your reference, the general geology of the complete Whitefish Range is published on the Kalispell, Montana 1° x 2° geologic map (Harrison and others, 1992).

All of the rock units in the northern Whitefish Range are part of an allochthonous structural block between the Rocky Mountain Trench and the Kishenehn basin. The allochthon consists, in many places, of a series of stacked thrust faults that have been folded and oversteepened by what are interpreted to be younger movements northeastward along a gentle, west-dipping basal detachment, see Figure 4. Locally, this style of faulting is thought to form duplexes at depth. Some of the oversteepened, piggybacked compressional structures were subject to backsliding during younger extensional events and formed listric normal faults such as the Wigwam fault (see Fig. 15). The structural block has a regional, north-plunging fold geometry.

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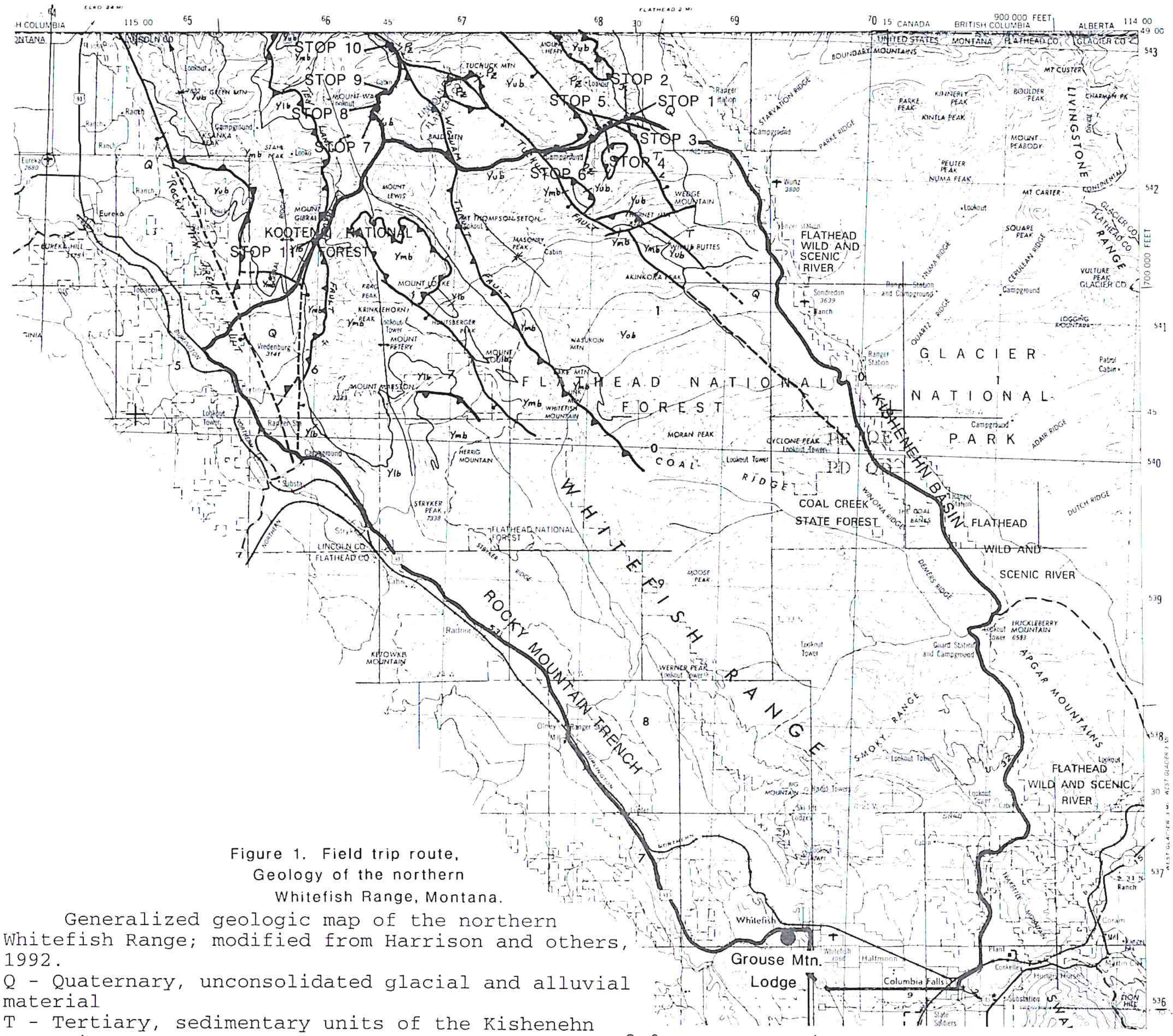


Figure 1. Field trip route, Geology of the northern Whitefish Range, Montana.

Generalized geologic map of the northern Whitefish Range; modified from Harrison and others, 1992.

Q - Quaternary, unconsolidated glacial and alluvial material
T - Tertiary, sedimentary units of the Kishenehn Formation
JR - Jurassic, sedimentary units of the Fernie Formation
PZ - Paleozoic sedimentary units, see Fig. 3
Yub - Middle Proterozoic Belt Supergroup, sedimentary units of the Missoula Group
Ymb - Middle Proterozoic Belt Supergroup, Helena and Empire Formations
Ylb - Middle Proterozoic Belt Supergroup, St. Regis/Grinnell Formation

FIELD TRIP LOG

0.0 Begin Field Trip from Grouse Mountain Lodge, west side of Whitefish Mountain-turn right out of entrance of Lodge and proceed east into the town of Whitefish.

1.2 Turn right at second street light and continue south on U.S. 93.

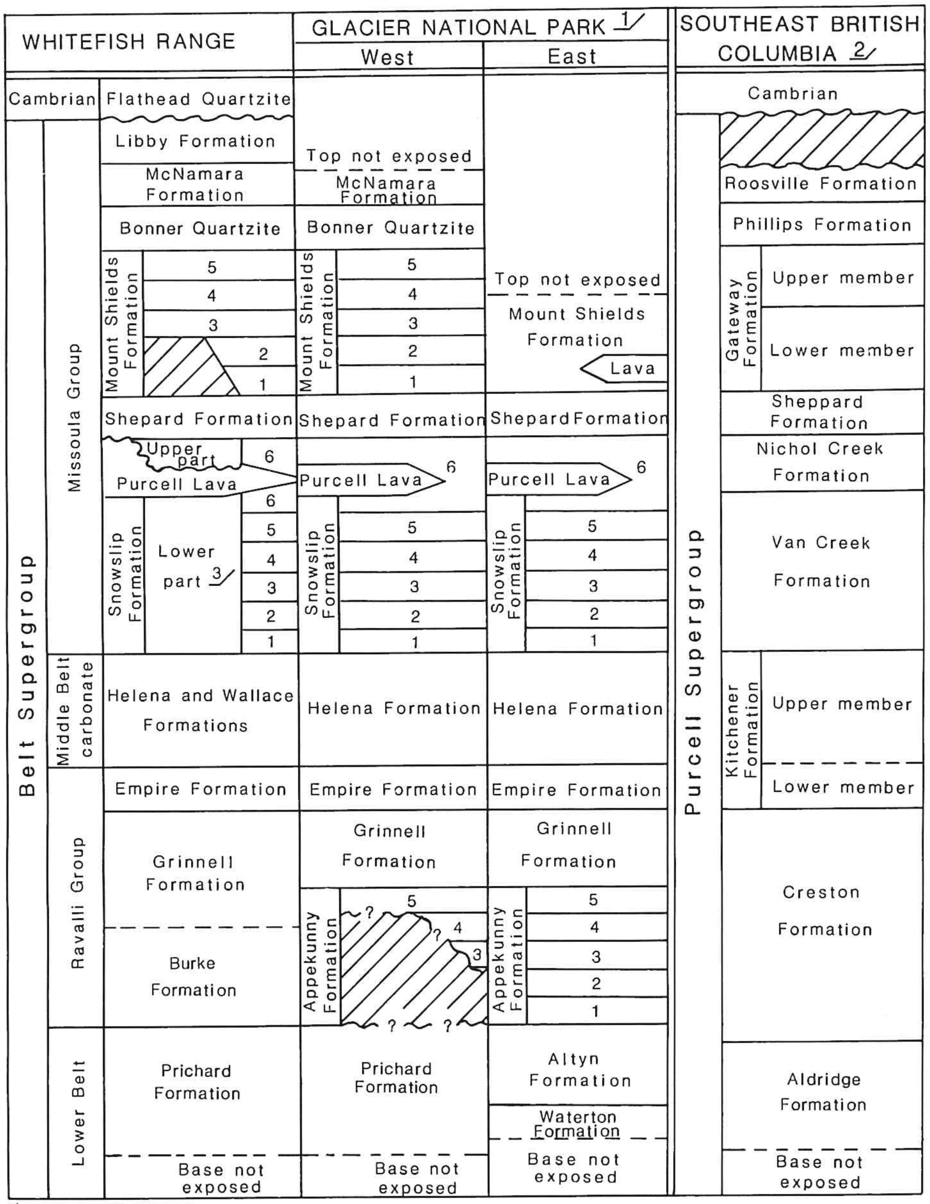
3.6 Turn left on Montana State Highway 40 and proceed west to the town of Columbia Falls, Montana.

8.0 Junction: U.S. Highway 2 and MT 40. Proceed straight ahead to Columbia Falls, MT.

11.1 Turn left on Nucleus Ave. (second stop light), go through the main center of town (0.7 mi) and turn right at the T-junction onto the North Fork Road (Highway 486). You are now traveling north on a narrow paved road through some isolated residential areas. Stay on the main road. Rocks exposed along the North Fork road for the next 25 mi are middle Belt carbonate and Missoula Group strata of the Middle Proterozoic Belt Supergroup.

17.1 Road cuts on both sides of the road for the next few miles expose gray dolomitic of the Helena Formation that weather locally to a tan or orangish-brown.

21.2 Junction with the Blankenship Bridge cutoff road. Continue north on paved road.



1/ Whipple, compiler, 1992.
2/ Modified from McMechan, 1981.
3/ Upper and lower parts of Whipple, 1984.

Figure 2. Correlation of Belt and Purcell Supergroups, Glacier National Park, Whitefish Range, and adjacent parts of Canada.

24.0 Pavement ends. Exposures in road cuts and outcrops along river are member 3 of the Mount Shields Formation. (see Fig. 2).

25.2 McGinnis Creek Road junction.

26.1 Fool Hen Hill. As you proceed along the curvy stretch of road, for the next few miles you pass by exposures of Missoula Group rocks beginning with the Snowslip Formation and working upsection to the Mount Shields Formation. The Purcell Lava, which is present in the northern Whitefish Range, pinches out southward at about this location, and although not exposed in road cuts, its talus is present locally. The exposures in these road cuts are intensely sheared and broken by northwest trending regional normal faults that cut through this area (see Harrison and others, 1992).

The Apgar Mountains across the North Fork of the Flathead River to the east are composed of Missoula Group rocks for the most part. In some fault blocks and low along the river level, outcrops of Helena Formation are present (see Whipple, compiler, 1992). The Apgars are particularly well known for grizzly bears that congregate there in the late summer to feed on huckleberries.

Rocky Mtn. Fm.		Pennsylvanian/ Permian	PALEOZOIC
Etherington Fm.		Mississippian	
Mount Head Fm.	Carnarvon Mbr.		
	Marston Mbr.		
	Loomis Mbr.		
	Salter Mbr.		
Livingstone Fm.			
Banff Fm.			
Exshaw Sh. (?)		? ?	
Palliser Fm.		Devonian	
Erosional Unconformity			
Unnamed Unit		Middle Cambrian	
Elko Fm.			
Gordon Sh.			
Flathead Qtz.			

Figure 3. Paleozoic nomenclature and stratigraphy, northern Whitefish Range, Montana.

30.0 Dip slope exposures of Shepard and Mount Shields Formation.

32.1 Big Creek campground and work center.

32.6 Big Creek road junction - Stay right on main road.

33.9 Long exposure of the Snowslip Formation. The Snowslip Formation in and around Glacier National Park consists of 6 informal members that consist mostly of red and green clastic units with 2 members being largely dolomitic siltite (Whipple and Johnson, 1988). These exposures are in the upper part of member 3.

39.8 Coal Creek. Tertiary beds of the Kishenehn Formation contain low-grade coal deposits that were mined locally near the confluence of Coal Creek and the North Fork of the Flathead River around the turn of the century. The Kishenehn Formation of possible Eocene to Miocene age (Constenius and others, 1989) underlies all of the North Fork valley and is interpreted to have been deposited in a graben contemporaneous with movements on the east-bounding Flathead Fault (Constenius, 1982).

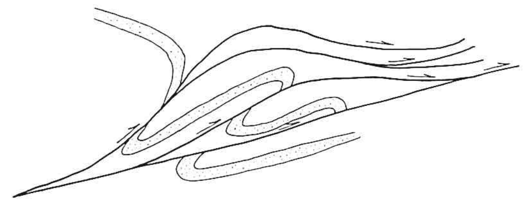


Figure 4. Diagrammatic sketch of a typical series of stacked and folded thrust faults over a basal detachment. Sedimentary unit shown to indicate sense of movements.

46.0 Junction Hay Creek Road--stay right.

47.4 Road junction to the metropolis of Polebridge, Montana a quarter of a mile east. Last stop for gas, groceries, and grog. This berg remains one of the few places where you can experience frontier living. Stay left of the junction and continue up the North Fork about to Trail Creek.

53.0 Red Meadow Creek road junction--stay right at this junction and right of other side roads that merge with the North Fork road. The Red Meadow road goes west across the Whitefish Range and comes out at Upper Whitefish Lake.

57.7 Whale Creek road junction--stay right, Ford Work Center on your right just ahead.

61.9 Cross Trail Creek bridge. Montana's first oil well was drilled by the Butte Oil Company due east of here in 1901 near the head of Kintla Lake where oil was first reported seeping from glacial debris in 1892 (DeSanto, 1985). Attempts to complete the well failed in 1903 when escaping, flammable gas ignited and burned down the derrick. The well reached a depth of about 1,450 ft and was reported to be in black limestone and iron. Based on recent mapping by the U.S. Geological Survey in Glacier National Park, the rock unit was probably the Prichard Formation. Hydrocarbons are thought to have seeped upward along faults that extend from Paleozoic source rocks at depth into the overthrust Precambrian rocks, similar to the setting at Sage Creek just to the north in British Columbia (Fermor and Price, 1983).

Other early wells on both sides of the 49th parallel have been drilled into the Tertiary Kishenehn Formation in search of oil. Small shows of oil in these wells are probably associated with oil shales in the Kishenehn (Boberg, 1984; Constenius and Dyni, 1983).

62.9 Trail Creek road junction--Turn left here off main road. From here west, the road is a single lane road with turnouts--be careful of on-coming vehicles.

The first few stops examine outcrops of Paleozoic strata, beginning in Pennsylvanian/Permian rocks and generally going down section to the west; see Figure 3.

65.9 Small Jeep trail to right leads to Thoma Lookout and Hefty Ridge trail.

Over the next 2 miles, stops 1 through 6 examine successively older Paleozoic stratigraphic units on the east limb of a north plunging anticline . Stops begin at exposures of the Pennsylvanian Rocky Mountain Formation and end downsection in the Mississippian Mount Head Formation, see Figure 3. The Livingstone Formation is exposed in the creek bottom along the axis of the anticline.

66.6 **STOP 1: Rocky Mountain Formation:** At this stop, the Pennsylvanian/Permian Rocky Mountain Formation is exposed in a small road cut on the right. The buff-colored, subfeldspathic sandstone forms cross-bedded planar parallel sets 15-30 in thick (Fig. 5). Internal lamination is even parallel on a millimeter scale. Iron banding and staining are common. Carbonate occurs locally in these exposures as clots of cement 0.2-0.4 inches in diameter and along some laminae (Fig. 6). Where the carbonate has weathered, exposures display dimpled surfaces . Framework grains in the sandstone are well sorted, fine to medium grain size. Burrows are common.

The Rocky Mountain Formation is the youngest, Paleozoic unit present in the northern Whitefish Range (Fig. 3) and has a maximum exposed thickness of about 1,000 ft. It forms bold outcrops (Fig. 7) and where mapped in this area, the Rocky Mountain Formation is structurally thickened or truncated by thrust faults. The relationship with the overlying Mesozoic units is not clear due to lack of exposure. Strata of possible Jurassic age are poorly exposed in a thrust plate along the northeast flank of Mount Hefty. On Cleft Rock Mountain and Mount Hefty, the Rocky Mountain Formation underlies klippen of Belt Supergroup and Middle Cambrian rocks.

Proceed west and downsection to the next stop at the underlying Etherington Formation.

66.8 **STOP 2: Etherington Formation.** The Mississippian Etherington Formation is exposed at this stop in a series of low road cuts on the right (the north side of the road). Here, the Etherington is composed of very fine- to fine-grained, calcareous sandstone that shows a variety of small-scale cross-lamination styles generally in sets less than 1.2 in thick (Fig. 8). Sparry calcite and limey mud segregations along laminae impart a banded character to exposures. Dark-gray to black chert nodules and irregular patches 2-8 in diameter are present locally.

In the northern Whitefish Range, the Etherington is composed mostly of interbedded light-brown to gray dolomite, arenaceous dolomite, micritic limestone, bioclastic limestone, and calcareous sandstone. Discontinuous interbeds, bedding-parallel segregations of chert, and weathering patterns impart a banded appearance to outcrops. Small solitary corals and crinoid columnals are common locally in limestone and dolomite beds. The thickness of the Etherington ranges from 250-400 ft. The differences in thickness appear to be related to the differences between thrust plates.

Continue west, stopping four times to look at the 4 members of the underlying Mount Head Formation

67.5 **STOP 3: Mount Head Formation, Carnarvon Member** Stop at large pile of gray limestone blocks on your right. These blocks are from the Carnarvon Member of the Mount Head Formation. Four members of the Mississippian Mount Head Formation are recognized in this part of the northern Rocky Mountains; they are in ascending order: Salter, Loomis, Marston, and Carnarvon Members (Price, 1961; Barnes, 1963), see Fig. 3. The Mount Head Formation is about 800 ft thick.

Across the road to the south and flanking Trail Creek are gray massive outcrops of micritic limestone typical of the Carnarvon Member. Fragments of large horn corals and brachiopods are present in some beds. Note the presence of dead oil as fracture fillings.

The Carnarvon Member in the Whitefish Range is a distinct, massive cliff former consisting of mostly gray, vuggy micritic limestone, laced with styolites and calcite veinlets (Fig. 9). A few feet of the lowermost beds are crinoidal grainstone. Large horn corals and scattered pods, patches, and discontinuous laminae of gray to black chert as thick as 1.2 in are present particularly in the uppermost part of the member (Fig. 10). Thin beds containing abundant brachiopod fragments are present locally. The Carnarvon Member is about 250 ft thick.

Figure 5. Cross-bedded subfeldspathic sandstone of the Rocky Mountain Formation at Stop 1.

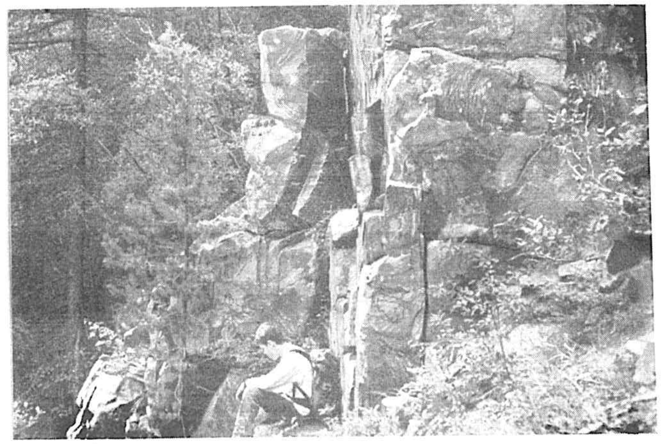


Figure 7. Cliff-forming outcrops of Rocky Mountain Formation on Cleft Rock Mountain.

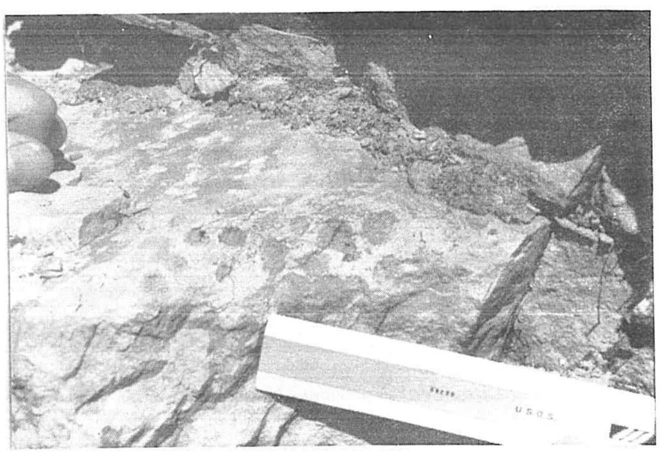


Figure 6. Carbonate clots of early cement in sandstone beds of the Ruby Mountain Formation at Stop 1.

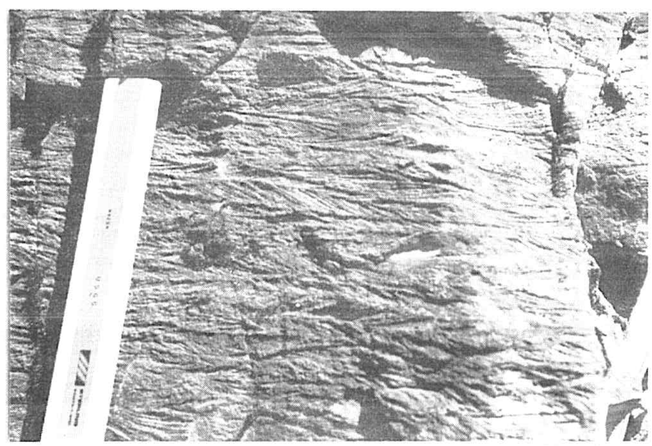


Figure 8. Small-scale ripple cross-lamination in sandstone beds of the Etherington Formation at Stop 2.

67.8 Junction with Forest Service road 114A; stay left on main roadway.

68.0 **STOP 4: Marston Member of the Mount Head Formation:**
The Marston Member of the Mount Head Formation is exposed in low road cuts on your right (Fig. 11). The Marston underlies conformably the Carnarvon Member and weathers to form distinct orangish-brown outcrops and soils. At this stop, the Marston consists of gray dolomite beds 4-12 in thick containing thin, wavy cryptalgal laminae. The dolomite is interbedded with light- to medium-gray shaley intervals generally less than 20 in thick and thin beds of gray micritic and fragmental limestone (Fig. 12). Small calcareous mudstone concretions and segregations are present locally.

The Marston Member is poorly exposed in the Whitefish Range because of its weathering characteristics. The member is recognized most often by orangish soils developed over dolomite beds. In the uppermost part of the member, a few interbeds of gray micritic limestone and bioclastic limestone form a transition interval with the overlying Carnarvon. The Marston is about 150 ft thick.

68.2 **STOP 5: Loomis Member of the Mount Head Formation:**
Stop at a small pullout at left. Cliff forming gray outcrops to the north across road are the Loomis Member of the Mount Head Formation which underlies the Marston Member. In this area, the Loomis is mostly a gray, crinoidal grainstone composed of massive beds that are largely a framework of disarticulated crinoid columnals. Thin, gray micritic limestone strata are interbedded locally. Solitary and colonial corals are commonly scattered throughout the member.

Within the Whitefish Range, the Loomis has much the same appearance and lithology as the Livingstone Formation which underlies the Mount Head Formation; however, the Loomis is much thinner and generally about 150 ft thick, whereas the Livingstone is about 450 ft thick. The Loomis in the Whitefish Range is much thinner than that reported from the Flathead and MacDonald Ranges to the north, where it has a reported thickness of 300-400 ft (Price, 1961).

68.5 Thoma Creek bridge.

68.6 **STOP 6: Salter Member of the Mount Head Formation:**
Pull out on left side of road just up from prominent road cuts. Road cuts at this stop expose the upper part of the Salter Member of the Mount Head Formation. The gray crinoidal grainstone of the overlying Loomis Member is exposed in the uppermost part of the road cut. These exposures consist of thick (20 in) beds of brownish-gray to tan dolomite and thin interbeds (<8 in) of light-green to pale olive shaley mudstone (Fig. 13). Some arenaceous dolomite beds display small-scale cross-lamination similar to the Etherington Formation at Stop 2. Obvious white to light-gray chert stringers and nodules as thick as 2.4 in punctuate the strata. Exposures below the road in Trail Creek contain abundant small solitary corals.

The Salter Member in the Whitefish Range is the lowest member of the Mount Head Formation and rests with apparent conformity on the Livingstone Formation. Like the Marston Member, the Salter weathers recessively which limits exposure of strata. Light-brown dolomite is typically banded by discontinuous layers and segregations of chert. Thin beds of dolomitic limestone and arenaceous dolomite are commonly present. The Salter in the Whitefish Range is about 250 ft thick.

68.8 At this point the road crosses the toe of a large landslide that originated from Cleft Rock Mountain across the creek to the south. The slide mass is composed mostly of debris from the Belt Supergroup McNamara Formation exposed in a klippe on Cleft Rock Mountain. Stratabound copper occurrences in rock debris from the slide prompted a prospector to drive an adit into the slide--his hopes were short-lived when he struck Paleozoic limestone within a few feet!

69.9 The light-gray cliffs in the creek bottom are the Mississippian Livingstone Formation that are exposed in a broad anticline. From this point the road travels upsection again until it crosses the Hefty thrust which places Belt strata over the Paleozoic section.

Older parts of the Paleozoic section are exposed in various structural blocks or thrust plates in the northern Whitefish Range, but access is difficult and requires bushwhacking.

70.3 Cliffs of the Carnarvon Member of the Mount Head Formation exposed to right in the west limb of anticline.

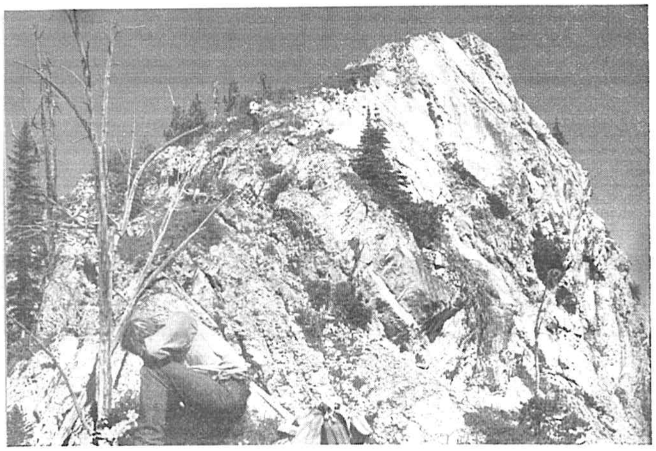


Figure 9. Cliff-forming massive limestone of the Carnarvon member of the Mount Head Formation on Cleft Rock Mountain; Steve Roof in foreground.

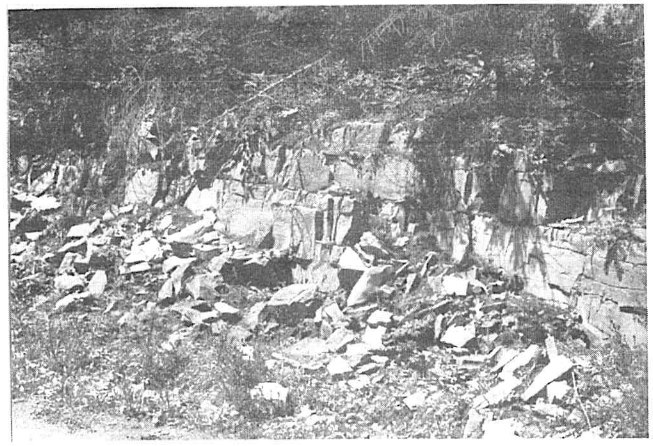


Figure 11. Dolomite beds of the Marston member of the Mount Head Formation in roadcut at Stop 4.

Figure 10. Horn corals on bedding plane of the Carnarvon member on Cleft Rock Mountain.

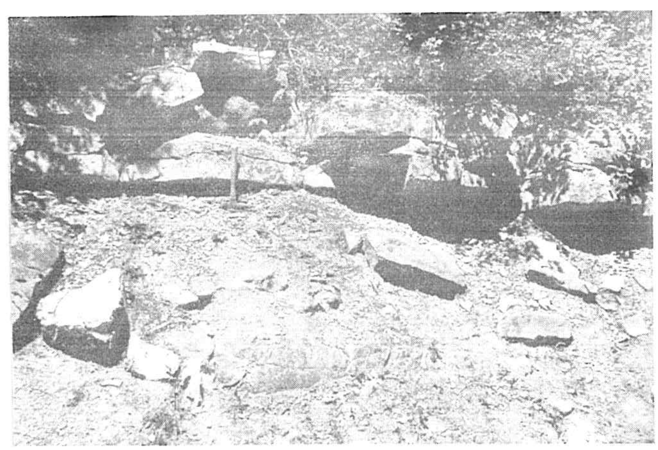


Figure 12. Dolomite and interbedded shale of the Marston member at Stop 4.

70.6 About at this point, the route crosses the Hefty thrust fault. From here east, mostly Belt rocks are exposed along the road until you cross the Whitefish Range Divide and drop into the Grave Creek-Wigwam River drainage. The remaining stops on this field trip examine Belt strata in those drainages. Stay on the main traveled roadway.

71.4 Bridge over Tuchuck Creek; Tuchuck campground to the left--stay right on main road. A prominent west-dipping normal fault trends northwest-southeast along the Tuchuck drainage and displaces Missoula Group rocks about 1,700 feet stratigraphically. The Mount Shields Formation is in the downthrown block.

Figure 13. Interbedded dolomite and shaley mudstone of the Salter member of the Mount Head Formation at Stop 6. Note discontinuous bed parallel chert stringers.

Figure 16. Salt casts in dolomitic siltite and argillite of Mount Shields Formation at Stop 7.

Figure 14. Landslide-dammed pond on Yakinikak Creek above Tuchuck campground. Landslide moved from right to left and is composed mostly of Belt debris.

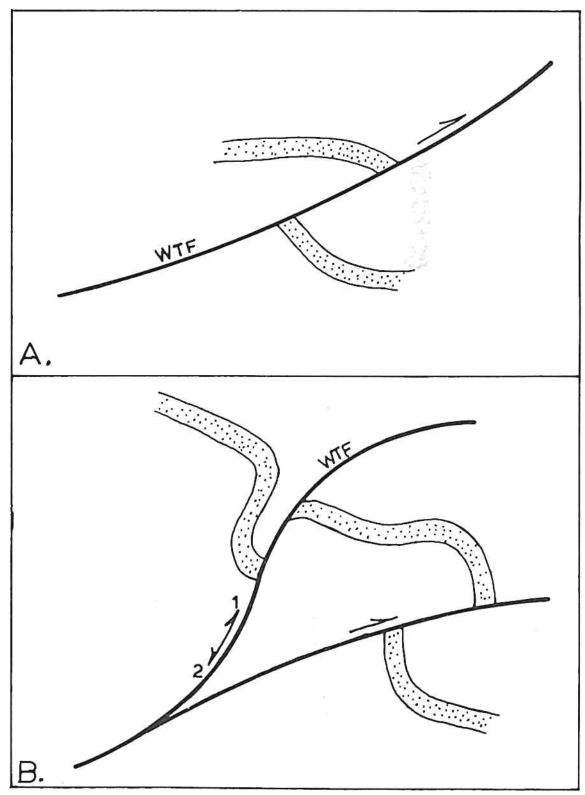


Figure 15. Diagrammatic sketch of the geometry and kinematics of the Wigwam thrust fault (WTF) in the northern Whitefish Range. A. - During regional compressional stress. B. - After piggybacking and oversteepening above younger thrust then backsliding during post-thrusting extensional stress. Numbered double arrow shows sequencing. Sedimentary unit shown to indicate sense of movements.

72.8 At about this point, the road proceeds around the toe of a large landslide that once dammed Yakinikak Creek. The resultant lake has drained but left excellent habitat for moose and cutthroat trout (Fig. 14). Landslide debris consists mostly of upper units in the Mount Shields Formation and covers both sides of the creek.

74.4 Exposures in road cuts of member 4 of the Mount Shields Formation (see Fig. 2). Stop 7 in Grave Creek will examine this unit more closely.

75.3 Outcrops of the Bonner Quartzite. Stop 8 will discuss the Bonner in detail.

From here the road begins an ascent to a low pass below Bald Mountain. Small west-dipping normal faults offset the section locally. Outcrops high on south-facing slopes are strata of the Mount Shields, Bonner, and McNamara Formations.

76.1 Cross the Wigwam thrust fault (Price, 1961). The Wigwam thrust fault at this location has experienced back sliding during post-thrusting extension (Fig. 15) and presently is a listric normal fault that exposes the upper part of the McNamara in the downthrown block.

77.4 Whitefish Range Divide and low pass. Proceed west downgrade with caution--narrow road with few turnouts.

77.9 Maroon argillite/siltite exposures to right are member 3 of the Mount Shields Formation.

79.5 Cross Grave Creek bridge and join Forest Service road 319--upper Grave Creek Road. Turn right and head updrainage .

At stops 7 through 10 over the next 5.3 miles, you will examine successively younger formations of the Missoula Group, starting with the upper part of the Mount Shields Formation and ending with a look at the Libby/Garnet Range Formation, see Fig. 2.

Note: The field trip route returns to this point and continues south.

80.9 **STOP 7: Mount Shields Formation-member 4.**
Road cuts at this stop expose a dip slope of member 4 of the Mount Shields Formation The Mount Shields is informally divided into 6 members in northwest Montana (Harrison and others, 1992). In the northern Whitefish Range, only members 3, 4, and 6 are recognized (see Fig. 2).

Member 4 exposed here consists mostly of interbedded couplets of gray-green siltite and argillite and couplets of brown-weathering dolomitic siltite and argillite. Thin beds of very fine-grained calcareous arenite and dense, gray dolomite locally punctuate the succession. Couplets are mostly wavy, non-parallel laminated and fine upward. Ripple marks and salt casts are abundant (Fig. 16). In this area the member is about 600 ft thick.

This member of the Mount Shields Formation closely resembles the Shepard Formation which underlies the Mount Shields. The carbonate-bearing lithologies are the principal criteria for recognition of this member. West, in the vicinity of Libby, Montana, member 4 is overlain by a thick unit (maximum thickness 500 ft) of stromatolitic limestone and dolomite designated member 5 by Harrison and others (1992). Differentiation between isolated outcrops of member 4 and the Shepard Formation can be realized by recognition of salt casts which are common to most of member 4, particularly in the lower part, and rare in the Shepard. In addition, the Shepard in the northern Whitefish Range contains abundant beds of pyritic arenite and oolites (Whipple, 1984).

Member 4 in this area is overlain by member 6 and underlain by member 3 of the Mount Shields Formation. Member 6 is about 180 ft thick and consists of thinly laminated pale-olive to grayish-green siltite and blackish-green argillite all stained by limonite. Conspicuous, white to light-gray, discontinuous beds 1.2-4.0 in thick of very fine-grained quartz arenite are scattered throughout the member. Member 3, which underlies the carbonate member 4, is mostly a sequence of gray-green siltite and pale-purple argillite arranged as thin, fining upward couplets. Outcrops and talus slopes are commonly tinged with reddish hues due to parting along argillite laminae, but the principal lithology is greenish siltite. Member 3 is informally called the "salt-cast" member of the Mount Shields because of abundant salt casts throughout its exposure in the Belt Basin. Total thickness of the Mount Shields Formation in the northern Whitefish Range is about 1,780 ft.

Changes in lithofacies of the Mount Shields in the Whitefish Range and Glacier National Park suggest a gradual shift from shallow water in the south and east of this region to deeper water in the north and west (Whipple and others, 1984). The lower members 1 and 2 of Mount Shields Formation are present in the southern Whitefish Range but pinch out northward toward this stop. Farther west, near Libby, the deeper-water lithofacies of member 6 have nearly doubled in thickness to 300 ft (Harrison and others, 1992).

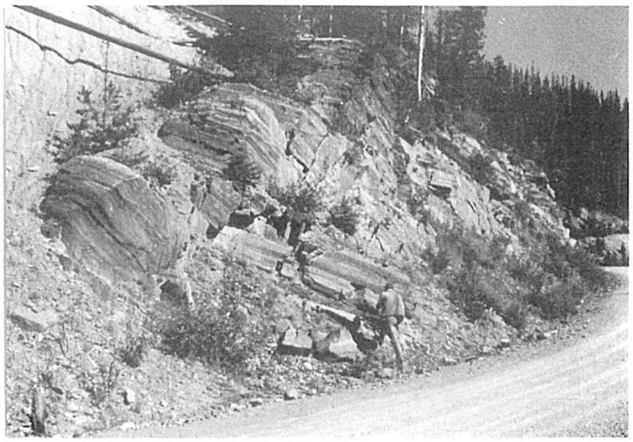


Figure 17. Roadcuts in Bonner Quartzite at Stop 8.

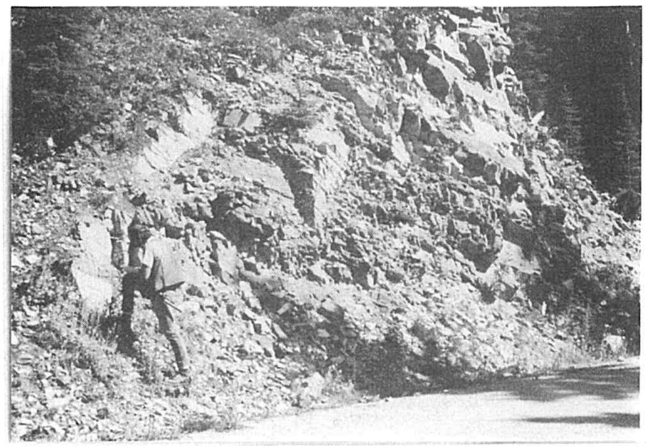


Figure 19. First roadcut at Stop 9 exposing green beds of the McNamara Formation.

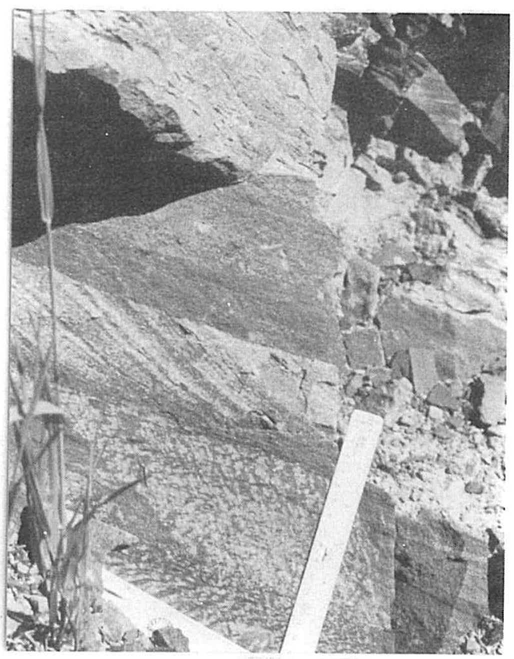


Figure 18. Cross-laminated arenite beds in the Bonner Quartzite at Stop 8.

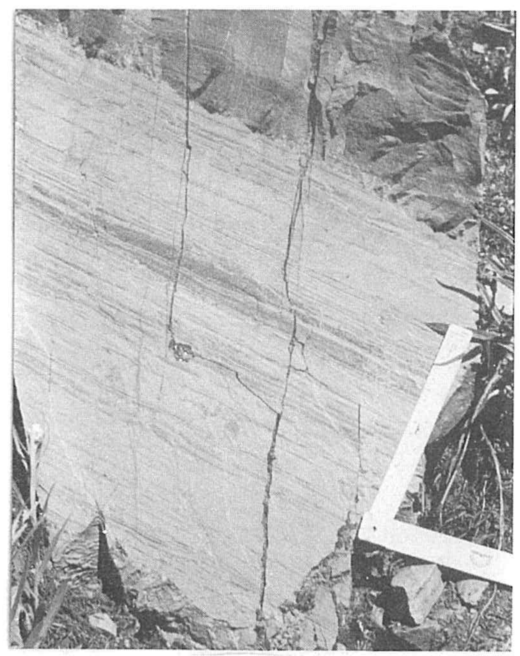


Figure 20. Wavy non parallel couplets of grayish-green siltite and argillite of the McNamara Formation at Stop 9. Note fluid-escape structures in center of photo.

81.3 **STOP 8: Bonner Quartzite.** This stop is at the point of the ridge where the road cuts maroon beds of the Bonner Quartzite (Fig. 17). Pull out on right and walk upsection through road cuts. These exposures are some of the northernmost in the Bonner sedimentation system and represent a fine-grained distal lithofacies. The sequence here consists of several moderately to poorly sorted arenite units that are mostly feldspathic to subfeldspathic in composition and separated by intervals of pink to maroon siltite and argillite (Fig. 18). White quartz arenite beds that contain maroon mud-chips are interbedded throughout the formation. Arenite beds are commonly cross-bedded in curved, tangential sets as much as 8 ft thick but typically less than 3 ft thick. Note the apparent lack of scouring and channeling at the base of these sets. Small-scale, ripple cross-lamination is common. The apparent lack of regular, fining-upward sequencing of these interbedded lithofacies is interpreted as deposition by sporadic flood events in the distal part of a fluvial system on mostly exposed mud flats.

Where exposed in the Whitefish Range, the Bonner Quartzite is uniformly about 600 ft thick. Changes in lithofacies are subtle and generally reflect a change from fining-upward, fluvial sequences near the south end of Glacier National Park (Whipple and others, 1985) to the irregular sequences observed at this stop. The Bonner rests sharply on the Mount Shields Formation and represents a rapid change in environments probably caused by tectonic adjustments in the Belt basin.

82.1 Weasel Lake road junction, stay left and stop to examine road cuts on the left.

STOP 9: McNamara Formation. A series of three road cuts at this stop expose gray-green strata of the middle to upper part of the McNamara Formation (Fig. 19). Begin at the first cut and proceed upsection. The McNamara is mostly a green-bed lithofacies consisting of fining-upward couplets of gray-green siltite and pale-green argillite. Couplets are typically wavy, non-parallel laminated and contain abundant fluid-escape structures, mud-chips, and shrinkage cracks (Fig. 20). Interlaminated yellowish-brown weathering calcareous siltite is present locally. Several couplets have ripple cross-laminated siltite at their base.

Distinct red-bed sequences are present in the second road cut. Note that most of the red-bed lithofacies resemble the green-bed type which is interpreted to mean that red beds here formed from the oxidation of green beds. The uppermost red-bed sequence in the road cut is composed of distinct units of maroon argillite and pinkish-gray, ripple cross-laminated siltite (<2 in thick) in the upper part. Scattered, small circular depressions on the bedding planes of some argillite units resemble rain-drop impressions. The last red-bed sequence is overlain by a 6 in thick oolitic limestone bed before passing upward into the green-bed lithofacies (Fig. 21).

The strata in these sequences are interpreted to have been deposited in a subtidal environment (green-bed lithofacies) that became emergent during deposition of the uppermost red-bed assemblage. The formation of oolites on top of emergent red beds and the succeeding deposition of green beds suggests a transgressive event and resubmergence.

As you move upsection, note the increase in carbonate content of the coarse size fraction of the couplets which is expressed by pale-yellowish weathering. Close inspection of the last road cut at this stop may reveal copper oxide stains from disseminated stratabound chalcopyrite which occurs locally in the McNamara Formation.

In the Whitefish Range, the McNamara Formation is about 4,000 ft thick. In addition to the lithologic units exposed at this stop, the McNamara contains several stromatolitic limestone beds 4-12 in thick in the green-bed lithofacies and at one locality on Mount Hefty, in red-bed lithofacies. Thin, white, quartz arenite beds are scattered throughout the formation. One of the basin-wide distinguishing characteristics of the McNamara is the presence of silicified or cherty mud-chips and discontinuous laminae. Although not obvious at this stop, the cherty features are present in exposures elsewhere in the Whitefish Range.

84.3 Junction with Weasel cabin/Frozen Lake road. Stay left on main road.

84.8 **STOP 10: Libby/Garnet Range Formation.** Stop 10 examines low road cuts on your right that expose vertical to overturned strata of the Libby/Garnet Range Formation (Fig. 22). The exposures consist of interbedded brownish-green, very fine-grained lithic arenite and siltite and thinly laminated black, micaceous, silty argillite. Arenite beds are commonly hummocky cross-stratified or even-parallel laminated and contain black mud-chips and small-scale load structures (Fig. 23). Black argillite units that drape arenite beds

Figure 21. Author perched at the top of red-bed sequence in the McNamara Formation where it is overlain by a 6 in thick oolite limestone bed on roadcut at Stop 9.

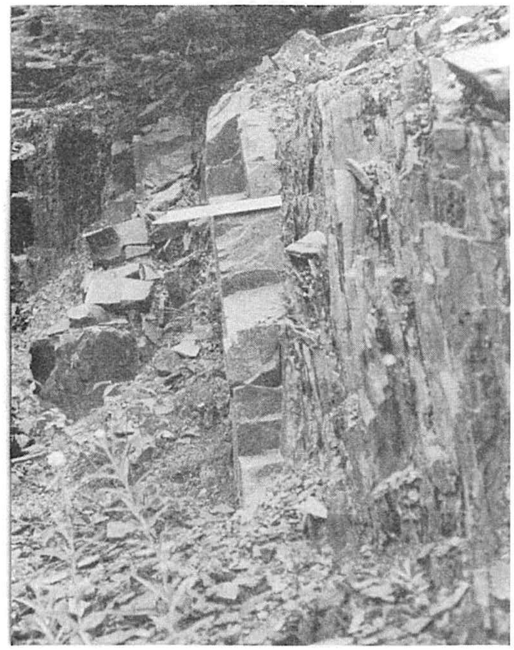


Figure 23. Looking along a hummocky bed of very fine-grained arenite at Stop 10.

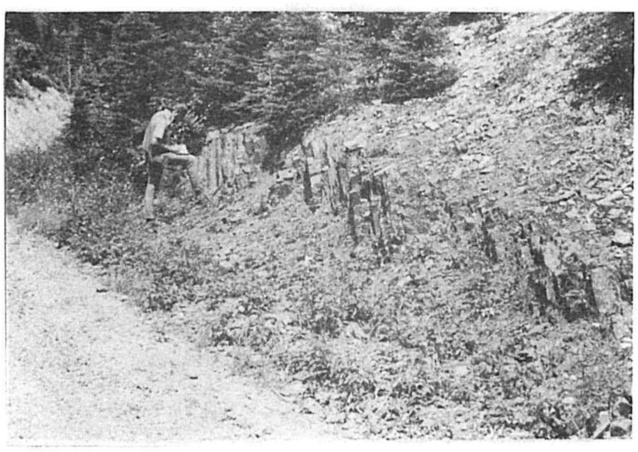


Figure 22. Roadcut exposure of near vertical beds in the Libby/Garnet Range Formation at Stop 10.

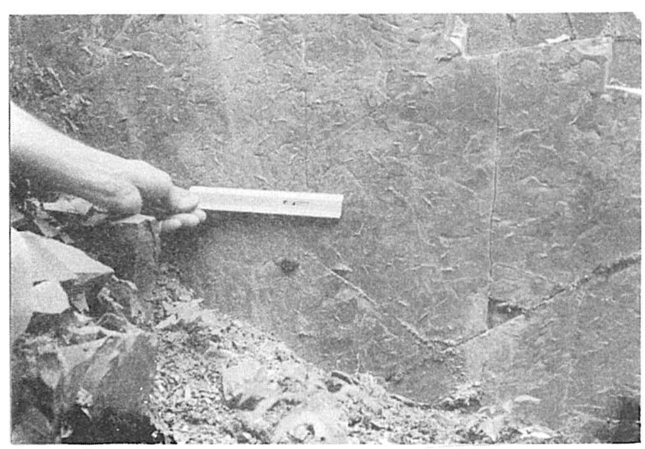


Figure 24. Subaqueous shrinkage cracks in black argillite of the Libby/Garnet Range Formation at Stop 10.

contain abundant shrinkage cracks interpreted to be subaqueous in origin (Fig. 24). The bedding style closely resembles lenticular couplets of the Wallace Formation.

The strata here were mapped as the Libby Formation by Whipple (1984), but perhaps the proper nomenclature for these strata should be Garnet Range Formation (Winston, oral communication, 1992). This road log, however, is not the appropriate forum for that determination.

Within the Whitefish Range, the Libby/Garnet Range Formation is only exposed in the northern part. It forms the uppermost stratigraphic unit in the Belt Supergroup and is overlain unconformably by the Middle Cambrian Flathead Quartzite. Maximum thickness is about 3,200 ft. The Libby/Garnet Range Formation in this area can be subdivided into two parts (Whipple, 1984) that generally correlate with the upper two parts of the Libby Formation in its type area (Kidder, 1985).

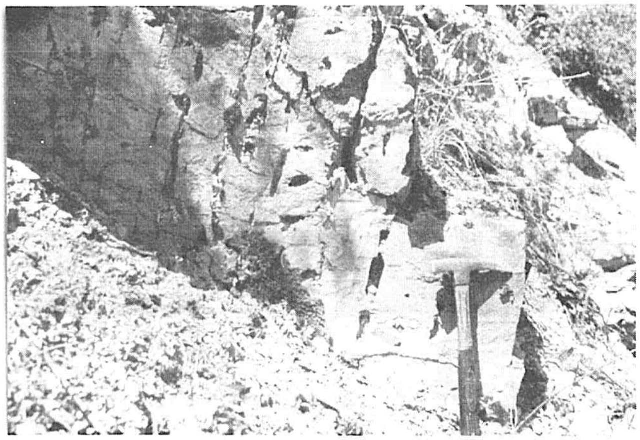


Figure 25. Weathered pods of carbonate cement in near vertical beds of the Empire Formation near its contact with the Helena Formation at Stop 11.

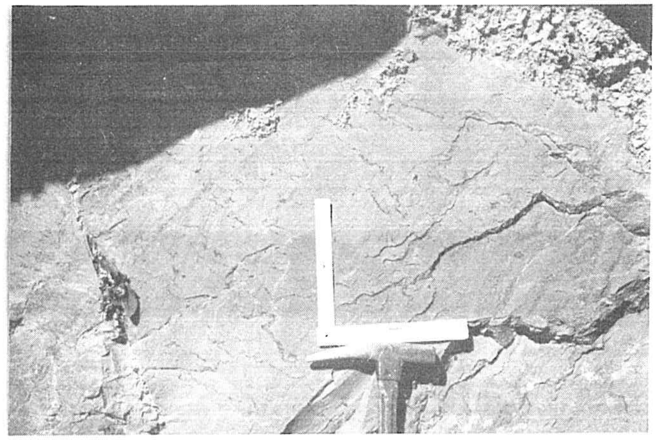


Figure 26. Ribbon-like segregations of sparry calcite, called molartooth structure, in basal dolomite beds of the Helena Formation at Stop 11.

At this point, the field trip makes an about face and returns to the road junction with Lewis Creek (5.3 mi) where you came over the Whitefish Range divide. From there the route travels down Grave Creek where the last stop will examine the Empire and Helena Formations.

RESET ODOMETER TO 0.0

0.0 Junction Forest Service road 319 (Grave Creek) with road 114 (Lewis Creek). **Reset odometer** and continue south on main Grave Creek Road (#114, see Fig. 1).

3.5 Clarence Creek bridge.

3.7 Junction of Grave Creek road with Stahl Creek road--stay left on main road.

Rocks exposed along this part of the route are in the Wallace Formation which overlies the Helena Formation in the northern Whitefish Range. The Wallace is characterized by wavy, lenticular beds a few centimeters thick that form paired, fining-upward sets. The lowermost part of the set is mostly gray, cross-laminated dolomitic siltite. The dolomitic siltite is draped by black argillite which forms the top of the set. Sets are separated by scoured surfaces generally accompanied by load structures.

4.3 Poorly exposed part of *Baicalia* stromatolite zone that approximately separates the Helena from the overlying Wallace Formation in this area. The *Baicalia* zone is as much as 36 ft thick (Whipple, 1984).

6.3 **STOP 11: Upper Empire and lower Helena Formations.**
Park on the left at the far end of the road cuts and work your way back up the road. At this last stop, you will have an opportunity to pick the contact between the Empire Formation and the overlying Helena Formation. The section is vertical to slightly overturned.

The contact between the two formations is transitional, but in these exposures, sedimentary structures and lithologies used to differentiate the two are nicely exposed and illustrate the position of the contact as mapped in this region.

The first exposures are in the Empire Formation and consist of gray-green dolomitic siltite and argillite couplets similar in appearance and sedimentary structures to the green-bed lithofacies of the McNamara Formation at Stop 9. Note the abundance of large pyrite cubes and attendant limonite staining. Also present are scattered pods and segregations of early carbonate cement typical of the uppermost part of the Empire and lowermost part of the Helena Formation (Fig. 25). The contact is placed within this interval of pods and mapped at the base of the first gray dolomite bed (weathers orangish brown) which contains molartooth structure. Molartooth structure is defined as small segregations of sparry calcite that weather to form wrinkled, ribbon-like structure in outcrop (Fig. 26).

The lower Helena in these exposures shows some well-developed cycles ranging in thickness from 5 to 25 feet (Fig. 27). These cycles are typical of much of the formation elsewhere in the Belt basin (see log for field trip by Winston and Lyons, this volume). When complete, the cycles begin with a intraclastic limestone unit, 2 to 30 inches thick, composed mostly of stromatolitic debris. More commonly, the cycle simply consists of the two, thicker dolomitic lithofacies, as shown in Fig. 27. Each ranges from 3 to 12 feet thick.

In the northern Whitefish Range, the Helena Formation is about 1,300 ft thick and is overlain by about 600 ft of Wallace Formation (Whipple, 1984). The *Baicalia* stromatolite zone approximately separates the two formations. South in the Whitefish Range and west near Lake Koocanusa, the Wallace and Helena are intimately intertongued on a scale not easily mapped and consequently are commonly lumped as one unit. This intertongued succession suggests deposition at the juncture of the eastern carbonate shelf (Helena Formation) with the axial trough (Wallace Formation) of the Belt basin.

This concludes the field trip. The route back to Grouse Mountain Lodge proceeds down Grave Creek about 7.5 mi where it connects with U.S.. Highway 93.

Turn left on the main highway and continue south toward Whitefish. This part of the return route is described in the log by Harrison and others, this volume. You may find it useful to consult the log from that trip--you will be traveling in reverse order to that log.

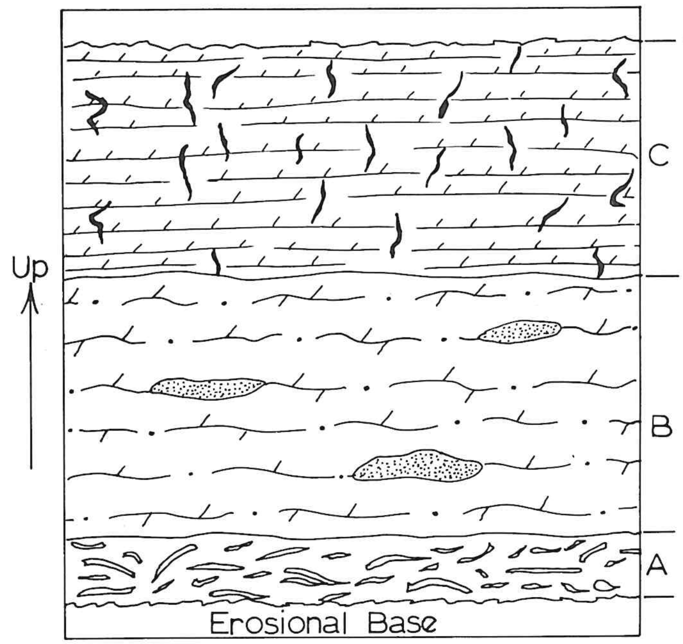


Figure 27. Diagrammatic sketch of typical cycle in lower part of the Helena Formation, Stop 11. A - Gray intraclastic limestone, clasts of stomatolite common; 2-30 in thick. B - Gray-green to tan weathering couplets of dolomitic siltite, rounded voids and pods of early cement; 3-12 ft thick. C - Gray to orangish weathering, thinly laminated dolomite, molartooth structure common; 3-12 ft thick.

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Idaho State University Geology Students Discover the Belt Basin.

Marias Pass, looking north, October, 1992. *Brennan Jordan meets John Frank Stevens. Eastbound Burlington Northern freight train in left rear has just mounted the continental divide. The Lewis thrust forms ledge on cliff in right distance. An early winter storm is lifting.*

Burlington Northern Quarry, Essex, Montana. *Quarry is in Snowslip Formation with exquisite sedimentary structures. Rachel Lawrence dances.*

Routes of Geologic Field Trips Described in this Volume

Authors' names are shown by geographic area. See Table of Contents.