NORTHWEST GEOLOGY

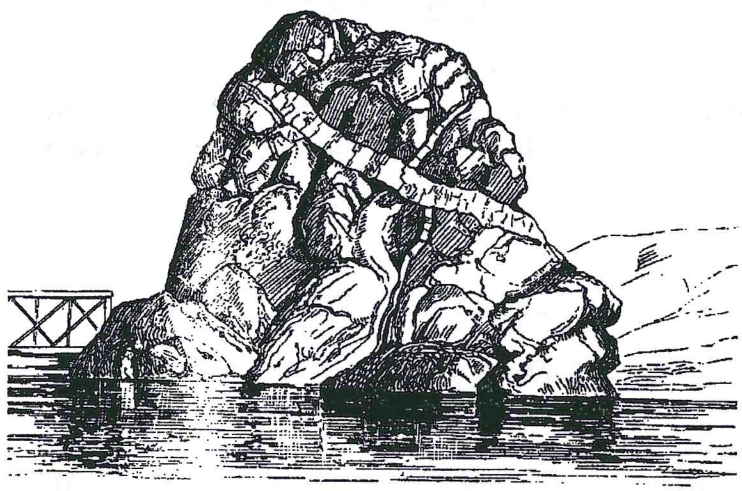
The Journal of The Tobacco Root Geological Society

Volume 33, 2004

29th Annual Field Conference

White Sulphur Springs, Montana

August 5-8, 2004



Published by The Tobacco Root Geological Society, Inc.
P.O. Box 2734
Missoula, Montana 59806
http://trgs.org

Edited by: Richard I. Gibson and Robert C. Thomas

Cover: Galt Mine, Neihart Mining District, Little Belt Mountains, 1895.
From W.H. Weed, Geology of the Little Belt Mountains, U.S.G.S. 20th Annual Report, Part III, 1898-99

The Tobacco Root Geological Society, Inc.

P.O. Box 2734
Missoula, Montana 59806

Officers, 2004:

President: Larry Smith, Montana Bureau of Mines and Geology, Butte
Vice-President: James Sears, Dept. of Geology, Univ. of Montana, Missoula
Scretary-Treasurer: George Furniss, MT Dept. of Environmental Quality, Helena
Corresponding Secretary: Emily Geraghty, Dept. of Geology, Univ. of Montana, Missoula
Web Mistress: Ann Marie Gooden

Board of Directors, 2004:

Richard B. Berg, Montana Bureau of Mines and Geology, Butte, MT
Bruce E. Cox, Stillwater Mining Co., Nye, MT
Marie Marshall Garsjo, Natural Resources Conservation Service, Ft. Worth, TX
Richard I. Gibson, Gibson Consulting, Butte, MT
Robert C. Thomas, Dept. of Environmental Sciences, U. of Montana-Western, Dillon, MT

Conference Organizers, White Sulphur Springs Field Conference:

Mitchell Reynolds, U.S. Geological Survey
Gerald Zeig, Teck Cominco American, Inc.

ISSN: 0096-7769

© 2004 The Tobacco Root Geological Society, Inc.

http://trgs.org

NORTHWEST GEOLOGY

The Journal of The Tobacco Root Geological Society

Volume 33, 2004

White Sulphur Springs Field Conference

Table of Contents

Field work and fossils in southwestern Montana: the 1898 field season of C. D. Walcott

Ellis L. Yochelson
Research Associate, Department of Paleobiology, National Museum of Natural History, Washington, DC 20013-7012
G. Zieg
Senior Geologist, Teck Cominco American Inc., East 15918 Euclid, Spokane, WA 99216

ABSTRACT

The diary of Charles Doolittle Walcott provides a brief daily account of his investigations of Cambrian and Precambrian rocks, mainly in the Belt Mountains during one field season. These entries also give some notion of the trials of field work before the development of motels and internal combustion engines. A key result of that season's work was the discovery of organic remains in Belt strata. Although by present day standards, these forms were misinterpreted, Walcott's publication on Precambrian fossils laid a solid basis for further investigations and made the search for fossils in such ancient rocks a respectable scientific pursuit.

INTRODUCTION

In 1879, Charles Doolittle Walcott (1850-1927) (Yochelson, 1998) joined the new U. S. Geological Survey (USGS) and July 1, 1894, became the third director of the agency. Shortly before that time the USGS had several field parties starting to investigate mining districts in Montana and Idaho. There was no overall stratigraphic succession, nor clear correlation from one mining district to another. In 1895, Walcott took a first quick trip through the Belt Mountains. In the vicinity of Neihart, Montana, he collected Middle Cambrian fossils (Weed, 1900). These fossils established that Lower Cambrian rocks were absent from the area and thus the Belt strata (or Algonkian, as Walcott called them) were pre-Cambrian in age The unhyphenated usage and the lack of capitalization of "formation" are relatively late developments in stratigraphic nomenclature.

For more than fifty years, Walcott used a small pocket diary and with his comments one can trace his route and gain some notion of how field work was conducted before the days of the rapid automobile transportation. Sundays are marked with an asterisk (*), but in the field "Sunday" was controlled by the weather and what had to be accomplished, not by the calendar. Inconsistencies in capitalization, spelling and punctuation are rendered as accurately as they can be interpreted from the diary.

THE FIELD SEASON OF 1898

Walcott left Washington, DC. on Saturday, July 30, accompanied by his wife Helena, and their oldest child, Charles, Junior; the younger children were left at home. The next day they were in Chicago and headed west to see his brother in La Crosse, Wisconsin. After a quick visit, Wednesday night they were in Livingston, Montana, spending the night on the train.

> *August 4 - Mr. F. B Weeks met us at Livingston & after attending to errands we drove out to camp 8 mi south. Out on the hills P. M. with Mr. Weeks looking for Cambrian section. 1) Source Creek Camp.*

To begin annotations of the diary entries, F. B. Weeks was a nearly constant field companion during the time when Walcott was director. When he was not in the field, Weeks was a librarian-bibliographer and was responsible for several of the early USGS compilations of American geological literature. The camp was run by Arthur Brown, officially a "messenger" with the USGS. This season, as with many, Walcott numbered the camp sites.

The Walcotts drove to camp in either a buckboard or a buggy, and probably the latter for Helena to use during the trip. Walcott spent the next day searching for Cambrian fossils and then the party moved south. On Sunday August 7 they were at Mammoth Hot Springs in Yellowstone National Park. They continued south to the Jackson, Wyoming, before returning. Walcott's official duties were to examine the new forest reserves - though he also examined Cambrian outcrops - and to make an official inspection of the National Park for the Secretary of the Interior. Camp 21, on September 3, was named by Walcott as "Terminal camp."

> **September 4 - Broke camp & packed up early + going to Livingston. Stopped at the Albemarle Hotel. A cl[ou]dy - cold day. Snow fell at night.-*

After a month of tent living, one can appreciate how the group, and especially Helena, must have appreciated being under a roof.

> *September 5 - Rainy morning. Packed up fossils & rocks and at 145 P.M. left on the freight train for Bozeman. Stopped at Bozeman House 530 P. M. Helena & Chas are well & strong. Mr. F. B. Weeks & Arthur Brown busy with camp outfit for trip down Gallatin. [River].*

Throughout his diary entries, Walcott showed great concern for times and distances. Insofar as one can check entries more than a century old, he was always accurate on distances and railroad schedules, so presumably his times in connection with field work are also accurate.
The distinction in usage between Mr. Weeks and Arthur Brown probably is attributable to Arthur Brown being a mulatto, as classified by the U. S. Census. On the other hand, Arthur was very nearly a member of the family (Yochelson, 1998a) and when Walcott referred to him by only his first name, it was never in a paternalistic manner. So far as Mr. Weeks is concerned, except in rare times of haste in writing when only a last name was used, with the exception of one person - "Joe," that is Joseph Paxton Iddings - Walcott's diary entries almost always use a formal title with a person, no matter how long the relationship.

> *September 6 - Called on President Rich, Dr. Lafhagen a.m. - Writing letters etc. P.M.*

The gentlemen mentioned must have been connected with the state college in Bozeman. No works by Lafhagan are listed in the bibliography of North American geology. In 1897, the Federal forest reserves had been established. Although Walcott was in the field, official mail always followed him. With both the USGS and the forest reserves to administer, Walcott never lacked for letters which required a reply.

> *September 7 - Left Bozeman 830 A. M. & drove buckboard to forks of E &*

W Gallatin rivers. Camped at ranch of Alex Proffutter [?]. Dr Lafhagen accompanied party. 22) Gallatin river camp.

This was a drive of about 30 miles to west northwest, quite a respectable distance with a buckboard. The route must have been near Montana highway 205, to the north of Interstate 90. If they arrived, say by 4:00 PM, in time to set up a tent and begin cooking before dark, they would have averaged about four miles per hour. Although no details are given, it seems reasonable that Helena, Charles Jr. and all the camp gear was in the buckboard, with Arthur Brown driving. Whether the other four were on horseback or there was another vehicle is unknown.

September 8 - Out all day on the Cambrian rocks rock north of the E. Gallatin river. Collected a lot of Cambrian fossils.

This is a classic locality and though no precise age is given, almost certainly the fossils were from Middle Cambrian shales.

September 9 - [Walcott's diary contains ditto marks for the day's entry, a rare practice for him when out in the field]

September 10 - Moved camp to a fine camping ground across the river from Logan. Out P. M. on Cambrian rocks with Mr. Weeks. Collected a lot of fossils from the M.[iddle] - C [ambrian] shales. 23) Logan Camp.

Logan is on highway 205 and the camp move would have put the party on the north side of the Gallatin River about 3-4 miles east of the confluence of the two forks of the river.

**September 11- In camp writing etc. At 11<u>17</u> a.m. Helena & Cha's left for Helena [Montana] on freight train to return at night en route to St. Paul Minn. Miss them greatly after our six weeks on train & in camp.*

The number of Walcott publications help to quantify just how much of a workaholic he was, yet by all measures he was also a devoted and caring husband and father.

September 12 - Broke camp 7<u>45</u> a.m. & moved to near mouth of Deep creek <u>via</u> Logan, Three Forks & Toston. 35 miles over good roads. 24) Road side camp. Lost camp kitten.

The three men would have traveled more or less parallel to Interstate 90 and at Three Forks, turned north north-east, about where US 287 is today. They crossed the Missouri River just before reaching Toston. 35 miles is a good day's travel with a buckboard.

September 13 - Moved camp up Deep Creek to about 11/2 mi below glendening's house at mouth of canyon. After lunch went out on hills south of Deep Creek to see Cambrian rocks. Found them altered & with very few fossils. 25)Brook Camp.

The Cambrian rocks on the western edge of the Big Belt Mountains are somewhat altered and poorly exposed. The area is a broad pediment, covered with gravels shed from the Big Belt Mountains to the east., dissected by Deep Creek and its tributaries. Cambrian exposures are limited to a few places along the edges of Deep Creek.

US highway 12 bisects the Big Belt Mountains and Walcott's party essentially followed its route.

September 14 - Drove up Deep Creek Canyon about ten miles to examine rocks. On return took photographs of conglomerate & near mouth of canyon found traces of fossils in the silicious [sic] Belt <u>shaly</u> slates.

Many years later, in 1914 when he was secretary of the Smithsonian Institution, Walcott (1915) took another trip through the area and published a photograph of a conglomerate in the Belt from this general locality; whether it is the same photograph is not clear. As is well known, a shale is a fine grained rock, originally a mud, but with water squeezed out and some of the minerals reoriented so that it splits into fine layers, parallel to the bedding. Slate is a metamorphic rock in which the minerals are both changed and reoriented. Slate can be cleaved into flattened pieces, but these are controlled by slaty cleavage imposed on the rock and cleavage is almost never parallel to the original bedding, so that searching for fossils in slate is a most difficult task. From his entry, apparently Walcott was trying to convey how relatively little the rock had been altered.

The "traces of fossils" mentioned occur in the lower part of the Greyson above the conglomerate. Despite its induration and relatively high siliceous content, the Greyson is typically quite fissile. Although unpublished , this may be the first notation of the relatively low grade of metamorphism in the eastern Belt Basin.

The conglomerate itself is a diamictite near the base of the Greyson Formation. In most areas, the lower part of the Greyson consists only of shale, but in a few restricted areas quartzite beds occur. The diamictite is thus an unusual rock for this stratigraphic position, but it occurs in the lower part of Deep Canyon and in the Lion Creek area northeast of White Sulphur Springs. It contains basement rocks and older sedimentary rocks. One clast containing stromatolites was found by Zieg along strike to the north, but no specimens of "Newlandia" have been noted in the diamictite.

Some geologists have interpreted the diamictite as debris flow, but there is little agreement as to whether it came from the south margin or the north margin of the Helena salient.

> *September 15 - Collecting in Belt slaty-shales all day. Only found a few bits of a crustacean - suggestive of a limuloid type.*

Walcott had found a few nearly black fragments of what he later named *Beltina*. His diary entry suggests that he immediately thought the pieces might have come from an animal related to *Limulus*, the horseshoe crab. Outcrops on North Fork Road near the mouth of the canyon have yielded *Grypania spiralis* (Walcott), but Walcott may have collected from another part of the Greyson Shale.

> *September 16 - About camp attending to mail, studying fossils collected in the Belt rocks.*

The run on nature of this entry might be interpreted to mean that the fossils were puzzling. Arthropods are complex animals and to find them in ancient rocks was quite unexpected.

> *September 17 - Moved camp up to near head of Deep Creek canyon & examined Belt silicious [sic] shales & slaty shales in P. M. Fine camp under old pine trees. 26) Pine camp.*

Near the head of Deep Creek Canyon both the Newland and Greyson Formations outcrop and from the limited information, it is difficult to determine on which he camped.

Walcott was looking at two slightly different units exposed locally and was attempting to record their distinction. The siliceous shales, as the name implies, had extremely fine grains of silica intermixed with the mud, whereas the slaty shales were closer to pure mud originally and probably formed much thinner layers. Presumably these were different beds in the Greyson Shale.

> **September 18 - Moved camp 33 miles en route to Neihart <u>via</u> White Sulphur Spgs. Camped near Newland creek about 4 <u>mi</u> below Kinney's ranch. Clear beautiful weather. 27) Newland creek camp.*

The move was more or less along the route of US highway 89, and after leaving White Sulphur Springs the distance would have placed the camp on the southern foothills of the Little Belt Mountains along the lower reaches of the creek. The precise spot cannot be determined but a reasonable supposition was that they may have stopped near the mouth of Charcoal Gulch. Newland Creek is now Newlan, but the Newland Formation was named before that change and retains the "d."

The carbonates along Newlan Creek consist of two cycles, shoaling upward, of clean, even-bedded limestone or dolomite, interbedded with silty and shaly carbonates and may have been deposited below wave base. The shale at the top of the second cycle is a siliceous silty shale and forms the base of the Greyson Shale. The carbonates do not show any "molar-tooth" structure and the interbedded siliclastic units also lack small scale, hummocky cross stratification that accompany "molar-tooth" dolomites in other parts of the Belt Supergroup.

"Newlandia" was named by Walcott on specimens from these rocks. The presumed algal structure from the clean carbonates has been interpreted as a product of pressure solution and related diagenetic processes (Zieg, 1981).

Below the carbonates of the upper part of the Newland are about 3,000 feet of platy, calcareous shale. In Deep Creek Canyon, the section is less calcareous. In the Newlan Creek area, the rocks are interpreted as microturbidtites (Feeback, 1997) which may be more proximal than distal, for more of the bases of the subunits of the individual microturbidites are preserved. The inferred direction of transport was from north to south.

> *September 19 - Moved camp 23 miles to a point near the base of Neihart Mt. on Sawmill creek. Collected a lot of Cambrian fossils en route. 28) Neihart camp.*

Almost certainly the campsite was upstream from Neihart in and beautiful spot near the confluence of Chamberlain Creek and Belt Creek. As indicated on a map of the area (Weed, 1900), it is roughly 17-18 miles by road from Wolsey's ranch where Walcott had collected in 1895, and he may have collected additional material from that place. Sawmill Creek does not appear on this map; the name was changed to Belt Creek. Lead-silver ore was the principal product of the Neihart mining district.

As regards the early part of the route, the Cambrian section is well exposed on the low divide between Newland Creek and Sheep Creek. Approaching Sheep Creek, one crosses the Volcano Valley fault, named by Weed. This fault thrusts Newland northward over the Cambrian which is overturned locally. Toward Sheep Creek the Cambrian is upright and then flattened so that he would have crossed the Flathead Sandstone, Wolsey Shale, Meagher Limestone, Park Shale, Pilgrim Limestone and Red Lion Formation of today's terminology. In crossing the divide from Sheep Creek into Sawmill (also known as Belt) Creek, Walcott would have traveled north and down through the Middle Cambrian to where the Flathead Sandstone lies on the Belt strata.

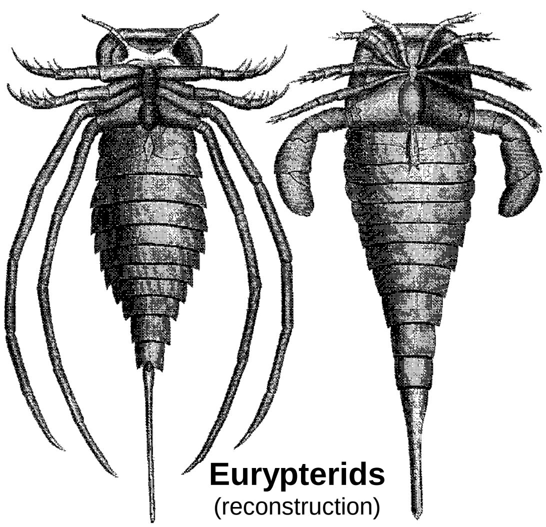
> *September 20 - Drove to Neihart & thence to Belt Park. Collected a few Cambrian fossils & returned to camp 5<u>30</u> P.M.*

This is another locality which Walcott had visited in 1895. "Drove" implies a team and buckboard. The group may not have had saddle horses with them. Although the price of silver had crashed in 1893, Neihart must still have been somewhat active as a mining camp.

In Belt Park, the Cambrian lies on crystalline basement rocks, rather than Belt. This demonstrates a post-Belt, pre-Flathead tectonic event, wherein downdropped blocks in the Little Belt Mountains contain areas of older Belt strata preserved below the Flathead. The downdropped areas are along the eastern extension of the Sheep Creek fault. That fault is parallel to the Volcano Valley fault.

> *September 21 - Studying section of Belt rocks in Sawmill canyon. Found a lot of fragmentary crustacean remains (Eurypteroid) in shale above <u>the</u> limestone.*

As mentioned above, a limestone unit occurs near Newlan Creek, and, not surprisingly, it was later named the Newland Limestone. Walcott was trying to make sense of the stratigraphic sequence, hence his remark on the occurrence of the fossils above the limestone. Eurypterids occur through out most of the Paleozoic, but are best known from the Silurian. The state fossil of New York is a eurypterid.



Eurypterids
(reconstruction)

It is possible that occurrence of this presumed fossil may have led to a complication in correlation. In 1973, Keefer mapped the carbonates exposed on Belt Creek as Newland and the overlying shales as Greyson, though this has since been recognized as a miscorrelation. However, the carbonates at Belt Creek are low in the section, just above the black, silty, Chamberlain Shale, and may have been deposited in shallow water. They show "molar-tooth" structure, are primarily dolomitic and somewhat unevenly and thickly bedded; in general aspect they are much like the middle Belt Helena or Wallace formations. Further, interbedded shales show small scale, hummocky cross stratification again typical of the middle part of the Belt. Feeback (19--) shows that calcareous shale typical of the lower part of the Newland Formation rests *above* these dolomites.

> *September 22 - At work on Belt rocks a.m. At noon started over devide [sic] on return trip. Camped at old saw mill near Wolsey's ranch house. <u>Rainy</u> <u>afternoon</u> 29. Sawmill camp.*

As noted above, during his hasty 1895 trip Walcott had visited Wolsey's Ranch and probably collected from the Middle Cambrian shales now named the Wolsey Shale. One consistency in Walcott's diaries was his misspelling of divide; for an inconsistency, he now omitted the ")" from the camp number.

> *September 23 - Moved camp. Lunched near White Sulphur Springs & camped at night 13 miles from W.[hite] S.[ulphur] on road to Townsend. A high wind all day. 30. Shelter camp.*

A few years later, Collen joined Walcott in the Swan Creek Range, and still later sent his large collections from the White Sulphur Springs area to Walcott, leading to Walcott's 1914 descriptions of "Newlandia" and related forms (Walcott, 1914). There are no comments on the rocks, as Walcott had traveled this road in 1895. The trip must have been at least 35 miles.

This campsite would have been close to where stromatolites were obtained from the Spokane Shale. Mannasseh Collen, for whom *Collenia* was named by Walcott homesteaded on nearby Battle Creek. A few years later Collen joined Walcott in the Swan Creek range, and still later sent his large collections from the White Sulphur Springs area to Walcott, leading to Walcott's 1914 description of "Newlandia" and related forms.

> *September 24 - Left camp at 7<u>30</u> a. m. - crossed over the Big Belt range & camped in glendenning's field in the same spot as Sept. 13-16. Warm & pleasant in the valley of Deep Creek. 31. Brook camp.*

> **September 25 - About camp. Examined & packed the Belt -fossils -*

Took a bath & wrote a few letters. Renal colic at night.

The first part of the diary entry is difficult to interpret. Occasionally Walcott used a dash rather than a period so his entries have "run on" quality. The dash after "fossils" fits this pattern, but the one before does not; one possible interpretation is that was his way of expressing a bit of uncertainty concerning the nature of the material identified as organic.

This would have been a relaxing true Sunday had Walcott not become ill. The illness today is termed constipation and he was severely impacted.

September 26 - In camp. Sick all night with renal colic. Mr. Weeks and Arthur up with me & worked hard. Rested during the day & drove to Townsend in the evening. Freight train to Helena. Went to Hotel Helena 130 a.m.

For Walcott to have abandoned field work, he must have been in severe pain. What one might do as camp doctoring for this malady is better left unexplored.

September 27 - Saw Dr E. S. Kellogg & went to his house to get well & brace up. Quietly resting P.M. & retired early.

So far as known, there was no prior contact with Dr. Kellogg, and for him to have offered hospitality to Walcott and for Walcott to have accepted may be an indication of what poor shape the man was in.

September 28 - Much better. Attended to a large mail rec[eive]'d a.m. Called on Senator Carter, Surgeon General Beatie etc. p.m.

When Walcott was out of Washington and time permitted, he was a master at finding the important people in any area and chatting them up.

September 29 - About town a.m. After lunch went out on Cambrian rocks with Leon S. Griswald. In the evening Supt. Of Forest Reserves J. B. Collins called & talked over Montana reserves.

Griswald left Harvard for a career as a mining geologist. Between 1891 and 1898, he published 11 papers and abstracts, but seemingly nothing further. Only one of these (Griswald, 1898) was germane to Montana.

Probably the two traveled across Missouri valley to reach the Cambrian outcrops.

September 30 - Out all day studying Cambrian & precambrian rocks with L. S. Griswald. Found an unconformity between the Flathead sandstone] & the Belt Terrace.

From the few fossils that the Flathead Sandstone had yielded, it was accepted as being Middle Cambrian in age. The presence of an unconformity demonstrated that the Belt rocks were older. This unconformity was the final piece of field evidence needed to prove that the Belt clastics were undoubtedly Precambrian. It is unfortunate that there is no more detailed information on the locality; the best guess is that this may have been in the Spokane Hills section along the Missouri River. A similar tectonic-stratigraphic gap occurs in the Belt Mountains, but presumably it was more apparent here.

October 1 - Rain, snow, cold. Wrote an article on Forest Reserves & in the evening spoke before the Mining Engineers Club & audience on Survey work & the "Forest Reserves."

It is not generally known but from 1897 until 1905, the USGS was essentially in charge of the forest reserves, now the national forests. Walcott (1898) gives only the general outline of events, but it was he who convinced Republican President McKinley to support the action of Democrat President Cleveland who had designated the reserves just before leaving office.

**October 2 - About house most of the day. Reading & writing. Weather moderating. Snow melted away before night.*

This was a restful, authentic Sunday for a change.

October 3 - Drove from Helena to N. Hilger's ranch with Mr. L. S. Griswald. Cold. Cldy day. After lunch went out on the rocks at Gate of the Mountain where the Missouri cuts thro'. Out until 6 p.m. Stopped at Hilger's.

It was near the Gates of the Mountains that Lewis and Clarke were led to believe that they had only to cross one large ridge to enter into western draining rivers. The Gates themselves exhibit a beautiful section of Devonian and Mississippian carbonates cut by the Missouri River. Probably Walcott's route was more or less parallel to, and perhaps slightly to the east of Interstate 15. It is likely that this excursion was by buggy.

October 4 - Left Hilgers, forded the Missouri & crossed at Beaver Creek - thence to Trout Creek, York & across the Missouri at Trout Creek ferry stopping at ferry house, staples. The section on Beaver Creek is unusually fine. Cold but the sunshine warmed the air at noonday.

The route is not clear but the two geologists may have recrossed the Missouri River near what is now Canyon Ferry on Montana highway 284. An unpaved road from York trends south to that town.

Presumably they saw a small amount of the Paleozoic section, but in particular the Spokane, Greyson, and Newland formations are exposed along these drainages; York is situated in the lower part of the Greyson Shale. The carbonates in the upper part of the Newland are not present in this area, and it is difficult to draw a contact between the Newland and the Greyson. Mining around York was focused on gold vein mineralization developed in a mudstone facies of the Greyson (possibly a deeper water deposit) which shows widespread silicification, potassic alteration and low levels of gold mineralization.

The younger Belt rocks, Empire Formation and Helena Formation are also exposed in this area. It is possible that Walcott recognized them as different from the Precambrian units studied to the east, but no field notes exist to confirm this supposition.

October 5 - Left ferry at 8a.m. & drove to Canyon Ferry & to White's gulch Big Belt Mtns. Ret's to Helena via Canyon Ferry- reaching Dr Kellogg's at 8 P. M. A fifty mile drive. Weather clear & cold. Found much of geological interest.

By any definition, that was a full day's work! Probably they drove though at least the middle part of the Belt, the Ravalli Group, and perhaps the Greyson Formation. White's gulch exposes a section from Mississippian carbonates down to the lower part of the Belt Supergroup.

October 6 - Writing letters and field notes a.m. At 3 45 p.m. left Helena for Marysville on N.[orthern] P.[acific] train. Mr. L. S. Griswald went up on horseback. Met Supt. Alex Burwell of Drumluman [sic] mine.

The railroad to the mine has long disappeared. Today, one travels northeast on Montana highway 279 and then travels west on an unimproved to the remains of Marysville. What precious metals are in the area today come from the skiers. Historically, however, the Drumlummon mine, as it is correctly spelled, produced several hundred thousand ounces of gold and it was the most productive gold mine in the one of the most productive gold districts in Montana.

The veins are emplaced mainly in rocks of the middle part of the Belt. They are peripheral to a molybdenum porphyry system at Bald Butte.

October 7 - Out all day on the section west of Marysville with Mr. Griswald. In the evening spent 21/2 h[our]s in the Drumluman mine with Supt. Alex Burwell.

After the geologists had a full day on the outcrop in October, the mine visit would have provided an opportunity to warm up underground. It is likely that they were looking at the Empire and Helena formations.

In 1906, Walcott named the Empire shale - capitalization of lithology was a later development. Two locations are given: the canyon walls just below Marysville; and on the ridge north of Empire, given as 12 miles west of Marysville. Presumably, Empire Gulch gave its name to the long-gone camp of Empire.

October 8 - Left Marysville 8 00 a.m. At Helena 9 30. Attending to correspondence etc. the remainder of the day & evening.

**October 9 - Left Helena 8 30 a.m. with L. S. Griswold & drove to Clendenin's [sic] ranch via Townsend - 51 miles. Stopped at ranch 530 p.m.*

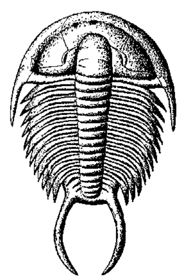
The distance is impressive and indicates a buggy, rather than a buckboard. In other entries Glendening appears and it is likely that Clendenin refers to the same ranch; in places, Walcott's handwriting is subject to various interpretations. North of Neihart, is a small mining camp named Clendenin and that may have added to confusion.

The best guess one can make as a reason for this long excursion so late in the season is that Walcott wanted to show the fossiliferous strata to a local geologist who might see similar material at other localities, as well as a desire to collect a few more specimens.

October 10 - Collecting Belt fossils until 3.p.m. We then drove across Greyson Creek & down the canyon & out to Townsend reaching there at 745 P.M.

This trip would have been a return west more or less along the present US route 12. It would have been south from Deep Creek to Greyson Creek drainage and then west to that creek. At that time of year the last part of this trip would have been in the dark.

October 11 - Left Townsend at 7 30 a.m. Drove to foothills of Boulder range west of Townsend. At 11. A.m. rain stopped work & we drove back to Helena stopping in Prickly Pear canyon. Went to Dr. Kelloggs.



This was the end of the field season in Montana. Presumably Mr. Weeks and Arthur Brown arranged for storing of the camping gear and then returned to Washington, for there is no further mention of them as being in the field. Over the years, Walcott remained in contact with Dr. Kellogg and as late as 1908 was writing him concerning a tonic and a cure for Helena's foot trouble.

Walcott spent the following day writing letters and packing before leaving for Butte. The next day he saw a senator and a local editor before leaving for Collinston, Utah, where he arrived 540 a.m. Walcott hired a team and driver and was off to examine the Cambrian rocks near Malad City, Idaho. Saturday, Sunday, Monday, and part Tuesday he was collecting and measuring sections before returning to Collinston; on Monday Walcott mentioned that he was tired!

After packing up on the 19th, Walcott was on the Union Pacific heading east. Thursday he was in Cheyenne, calling on a senator and discussing water problems with the state engineer. He spent most of the next day visiting state officials before leaving for Omaha, but still was delayed going home, for October 22, he was at the Omaha exposition grounds for 14 hours meeting people. *The Gov't Exhibit & Mining Exhibit are good - the former unusually so &*

the latter as good as possible. Most of Sunday was spent at the exposition until he took the evening train to Chicago. Never one to waste an opportunity, he met with University of Chicago geologists and after lunch visited a few family members.

> *October 25 - On B.[altimore] & O.[hio] train until 455 p.m. Helena & little Stuart met me at the station in Washin[gton]. All well at home & I am happy to be here.*

Anyone who has been in the field for a long season would have the same sentiments. Benjamin Stuart, the third son and the fourth and last of the Walcott children was born July 8, 1895.

AN IMMEDIATE REPORT ON FIELDWORK

A short summary of the season's effort in the Belt Mountains was soon published by the USGS. *Early in September a camp outfit was obtained at Bozeman, Montana, and a study of the Belt Mountain series of rocks was begun. From the time of the Hayden survey of this region there had been differences of opinion as to the correct stratigraphic position of a series of shales, sandstones, and limestones, 10,000 feet or more in thickness that form the Big and Little Belt Mountains. By studying and measuring a number of local sections and one crossing from Helena to Neihart, data were obtained which showed that the Belt Mountain formations were unconformably beneath the Cambrian rocks; also that they contain the oldest traces of highly organized animal life known* (Walcott, 1899, p 60-61).

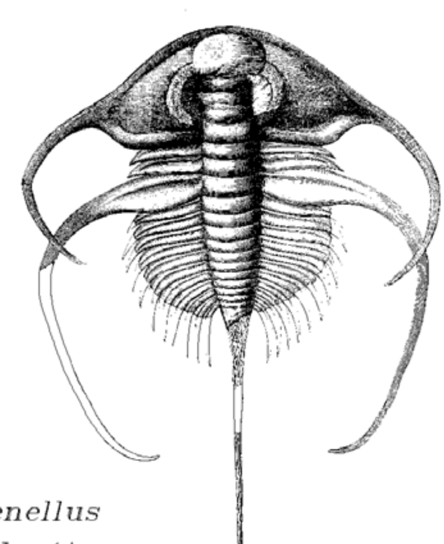
A MAJOR PUBLICATION AND DISCUSSION OF FOSSILS

It may well be that the discovery of organic remains in the Belt was the impetus for what, in theory at least, is a key paper in history of geology (Walcott, 1899a). In it, Walcott provided a fairly extensive summary of the Precambrian of Montana and its relationship to the overlying Cambrian. He then went on to a shorter treatment of the Precambrian in other parts of North America: the Grand Canyon of Arizona; the Llano area of Texas; eastern Newfoundland; and the Lake Superior region. He had worked in the first three of those regions. The final section of the paper is concerned with reports of fossils from these areas. He discounted several reports, but was clear that the Precambrian of Grand Canyon had yielded fossils, as did the Belt Series.

The fossils thus far discovered in the Belt Terrace occur in the Greyson shales, in a belt of calcareous shales about 100 feet above the Newland limestone, at a horizon 7,700 feet beneath the summit of the Belt Terrane at its maximum development. Indications of fossils were first discovered near the mouth of Deep Creek canyon, a short distance above Glenwood postoffice. Subsequently they were found in Sawmill canyon, about 4 miles above Neihart (Walcott, 1899a, p. 234). That the fossils were apparently found at the same stratigraphic level in places 35-40 miles apart gave promise fossils which might be useful for biostratigraphic correlation. As shown, however, the two occurrences are from different stratigraphic units.

Walcott went on to formally name and illustrate two species of *Planolites* Nicholson and three species of *Helminthoidichnites* Fitch, two of which were questionably assigned to the genus. On the caption of plate 24, these are identified as "Annelid Trails on Grayson [sic] Shales." He also named a new genus *Beltina* and the new species *B. danai*, assigning it the Merostomata within the Arthropoda.

In retrospect, the best one can write is that his systematics have fared quite poorly. There were earlier comments discounting some of the forms, as summarized by Horodyski (1993, Appendix 1, p. 18), but the first critical systematic restudy of some of the material was by Walter, Oehler and Oehler (1976). Yochelson had examined Walcott's specimens in preparation for a talk at the Twenty-fourth International Congress of Geology and later he called them to the attention of Malcolm Walter. Walter *in* Walter, Oehler and Oehler (1976), Yochelson (1979) and Horodyski (1993) concur that *Planolites superbus* Walcott and *P. corrugatus* Walcott are almost certainly of inorganic origin; Hofmann (1992, p. 418) illustrates both, and considers that they fall within his category of dubiofossils. In partial defense of Walcott, one should note that interpretations of both trace fossils and sedimentary structures are far advanced, compared to the state of knowledge a century ago.



Olenellus gilberti

The material which Walcott assigned to three species of *Helminthoidichnites* was described in general terms by Yochelson (1979, p. 274-275). In their revision of this material Walter, Oehler and Oehler (1976) retained one of the specimens as *Helminthoidichnites meeki* Walcott, but questioned its assignment to that genus. The genus is based on what was at the time assumed to be a trace fossil, but the authors interpreted it as of algal origin and implied that the type of the genus might also be algal. The original description has one "i" on the trivial name and "ii" on the plate description. If this form is a "plant," the use of *meekii* is correct. Hofmann (1992a, p. 355) informally suggests that it might belong within *Grypania*.

In examining Walcott's material, Walter, Oehler and Oehler (1976) determined that two new genera and species were present but unrecognized by him. They named *Lanceforma striata* and *Proterotainia montana* and assigned *Helmintoidichnites? neihartensis* Walcott to the latter genus. All three species were considered filamentous algal body fossils, rather than trace fossils. Horodyski (1986) considered the first of these as possibly a fragment of a microbial mat, the second as nonbiogenic, and the third as possibly nonbiogenic. (Hofmann, 1992a) reproduced one the 1976 illustrations of *Lanceforma* and suggested it might possibly be a synonym of *Beltina*. In a table, Hofmann, 1992b) interprets *P. montana* as inorganic and *P. neihartensis* as algal; what that interpretation does to the nomenclatural status of the latter species is not clear. Horodyski (1993, p. 19) maintained his earlier views on these three taxa, and emphasized that *P. neihartensis* was probably inorganic.

The third species which Walcott assigned to the genus, *Helminthoidichnites? spiralis*, became the type species of *Grypania* Walter, Oehler and Oehler. This genus seems to be generally accepted as an authentic body fossil rather than as a trace of movement, and one that is widespread. Indeed, it is the hallmark of a Middle Proterozoic assemblage (Hofmann and Bengtson, 1992, p. 503). One specimen of *Grypania spiralis* (Walcott) was reillustrated and the genus was briefly discussed by Hofmann (1992a). Horodyski (1993, p.19) judges it to a eucaryotic alga, and perhaps a senior synonym of a genus described from the Precambrian of China. Walcott's discovery of this form is certain more important than his misclassification. Again in partial defense of the original work, during the 1890s no one was seriously considering that algal filaments could be preserved, let alone preserved in ancient sedimentary rocks.

Finally one arrives at the genus *Beltina* and its type and only species *B. danai* This may have been Walcott's greatest error, at least in terms of assigning this material to the arthropods. Two unrelated points may be helpful in trying to interpret Walcott's view. First, in earlier work, he had extended the record of eurypterids from Silurian downward into the Ordovician (Walcott, 1882), so there was no reason to automatically question that the group might occur in still older rocks. Second, in terms of color, disarticulation, and fragmentation of some specimens, the material of *Beltina* does resemble that of the famous New York Silurian Passage Gulf occurrence of eurypterids. Walcott was clearly puzzled that such an advanced form should occur in older rocks, but the other side of that particular coin is that a variety of arthropods occur in the Early Cambrian and obviously they had ancestors. One may also speculate that Walcott was particularly attuned to megascopic arthropods and had no understanding of algal mats. This may have been another example of the well-known phenomenon that "the eye beholds what the mind perceives."

Hofmann (1992a , p. 354) places this material in an informal group of Beltinoid remains, which are widespread. Horodyski (1993, p. 19) has summarized the literature on this form and his conclusion is that it is based on fragments of microbial mats or megascopic algae. One of Walcott's specimens was illustrated by Yochelson (1979, p. 273). It is twisted in much the same way at the type specimen of *Lanceforma striata*, indicating a flexible form, though it is far larger than an algal filament.

One can argue that in the face of such major revision and dismissal of much of Walcott's interpretation of organic material, the paper is insignificant. Alternatively, one can argue that by emphasizing the general stratigraphic setting and distribution of Precambrian rocks, and by questioning the earlier reports of fossils, Walcott laid a sound basis for future investigations of life in the Precambrian. Biostratigraphic investigation of megafossils in North America had resolved correlation problems in the Paleozoic in general, and the Cambrian in particular.

Near the turn of the 19th century, enough material had been found to suggest that hard work and luck might result in further paleontological discovery in the Precambrian However, if one sets aside study of stromatolites, and the work of Fenton and Fenton during the 1930s, investigations by paleontologists languished for half a century.

When there was renewed paleontological investigation of the Precambrian, it came about in part as a result of improvements in technology for the study of microfossils. Knoll (2003, p.90) has dubbed the late Elso Barghoorn as "the father of Precambrian paleontology" and that is indeed a reasonable label, but, if so, Walcott is surely the grandfather of this field. To repeat a statement previously quoted, "The man who probably expended more time and energy than any other individual in trying to find pre-Cambrian fossils was Walcott" (Raymond, 1935, p.388).

ACKNOWLEDGMENTS

The late R. J. Horodyski kindly showed Yochelson the *Grypania* locality from which he had collected many specimens. During the same trip, in Glacier National Park, he later demonstrated Belt stratigraphy and stromatolites. He also led a group to the Precambrian outcrop showing the megafossil which later was named *Horodyskia* (Yochelson and Fedonkin, 2000; Fedonkin and Yochelson, 2002).

REFERENCES

Fedonkin, M. A., and Yochelson, E. L., 2002. Middle Proterozoic (1.5 Ga) *Horodyskia moniliformis* Yochelson and Fedonkin, the Oldest Known Tissue-Grade Colonial Eucaryote. Smithsonian Contributions to Paleobiology number 94, 29 p.

Feeback, D. L. 1997, Stratigraphy and sedimentation of the Chamberlain Formation -Newland Formation boundary, lower Belt Supergourp (Middle Proterozoic), central Montana, M.S. thesis, University of Montana.

Griswald, L. S., 1898. The geology of Helena, Montana, and vicinity. Association of Engineering Societies, Journal, 20:51-68.

Hofmann, H. J., 1992. Megascopic dubiofossils: 413-419. *in* Schopf, J. W. and Kline, C. (eds.) The Proterozoic Biosphere- A multidisciplinary study. Cambridge University Press, 1348 p.

_____, 1992a. Proterozoic carbonaceous films: 349-357. *in* Schopf, J. W. and Kline, C. (eds.) The Proterozoic Biosphere- A multidisciplinary study. Cambridge University Press, 1348 p.

_____, 1992b. Proterozoic and selected Cambrian megascopic carbonaceous films: 957-979. *in* Schopf, J. W. and Kline C. (eds.) The Proterozoic Biosphere - A multidisciplinary study. Cambridge University Press, 1348 p.

Hofmann, H. J., and Bengtson, S., 1992. Stratigraphic distribution of megafossils: 501-506. *in* Schopf, J. W. and Kline, C. (eds.) The Proterozoic Biosphere- A multidisciplinary study. Cambridge University Press, 1348 p.

Horodyski, R. J., 1986. Walcott's *Helminthoidichnites* assemblage from the Middle Proterozoic Belt Supergroup. Geological Society of America, Abstracts with Program, 18: 640.

______, 1993. Precambrian paleontology of western United States and northwestern Mexico: 558-565, plus two appendices on microfiche. *in* Reed J. C., Jr., Bickford, M. E., Houston, R. S., Link, P. K., Rankin, D. W., Sims, P. H., Van Schmus, W. R. (eds.) Precambrian: Conterminous United States. Geological Society of America, The geology of North America, volume C-2, 657 p.

Knoll, A. H. 2003. Life on a Young Planet; The first three billion years of evolution on Earth. Princeton University Press, Princeton, NJ and Oxford, England, 277 p.

Raymond, P. E., 1935. Pre-Cambrian life. Geological Society of America, Bulletin, 46: 375-392.

Walcott, C. D., 1882. Description of a new genus of the Order Euryiperida from the Utica slate. American Journal of Science, 23: 213-216.

_____, 1898. The United States Forest Reserves. Appleton's Popular Science Monthly, 52: 1-13.

_____, 1899. 20th Annual Report of the Director of the United States Geological Survey to the Secretary of the Interior, 1898-99, part I: 9-209.

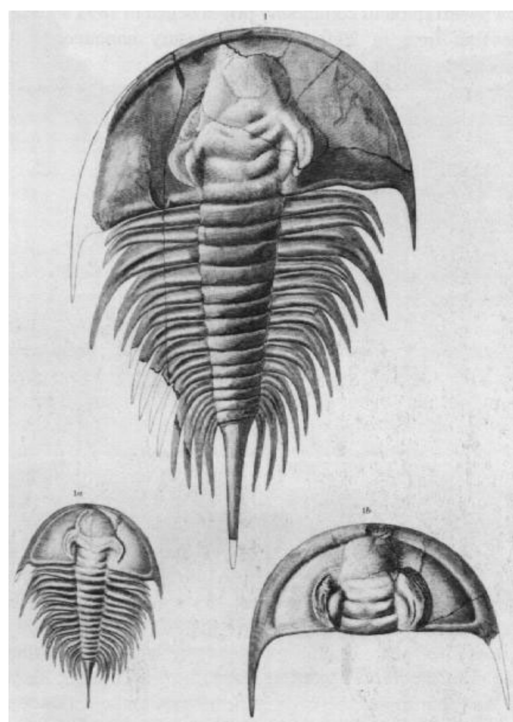
_____, 1899a. Pre-Cambrian fossiliferous formations. Geological Society of America, Bulletin 10: 199-214.

_____, 1914. Precambrian Algonkian algal flora: Smithsonian Miscellaneous Collections, v.64 no. 2, p.77-156.

______, 1915. Geological explorations in the Rocky Mountains. *in* Explorations and field-work of the Smithsonian Institution in 1914. Smithsonian Miscellaneous Collections, 65: 1-10

Walter, M. R., Oehler, J. H., and Oehler, D. Z.. 1976. Megascopic algae 1300 million years old from the Belt Supergroup, Montana: a reinterpretation of Walcott's *Helminthoidichnites*. Journal of Paleontology. 50: 872-881.

Weed, W. H., 1900. Geology of the Little Belt Mountains, Montana, with notes on the mineral deposits of the Neihart, Barker, Yogo, and other districts by W. H. Weed, accompanied by



Some of Walcott's trilobites (USGS image)

a report on the petrography of the igneous rocks of the district by L. V. Pirsson. 20th Annual Report of the Director of the United States Geological Survey to the Secretary of the Interior, 1898-1899, Part III: 257-581. [Only part I of the 20th Annual Report appeared in 1899].

Yochelson, E. L., 1979. Charles D. Walcott - America's pioneer in Precambrian paleontology and stratigraphy: 26l-292. In Kupsch, W. O. and Sarjeant, W. A. S. (eds.) History of concept in Precambrian geology. The Geological Association of Canada, Special Paper 19, 292 p.

_____, 1998. Charles Doolittle Walcott, Paleontologist. Kent State University Press. Kent Ohio, 591 p.

_____, 1998a. Arthur Brown: The forgotten "assistant for all seasons." Marrella 7:14-18.

Yochelson, E. L., and Fedonkin, M. A., 2000. A new tissue-grade organism 1.5 billion years old from Montana. Biological Society of Washington, Proceedings, 113 (3): 843-847.

Zieg, G.A., 1981, Stratigraphy, sedimentology and diagenesis of the Precambrian upper Newland Limestone, central Montana: M.S. thesis, University of Montana, Missoula, 182 p.

Lithology and age of pre-Belt Precambrian basement in the Little Belt Mountains, Montana: implications for the role of the Great Falls Tectonic Zone in the Paleoproterozoic assembly of North America

James J. Vogl, David A. Foster, Paul A. Mueller
Department of Geological Sciences, University of Florida, Gainesville, Florida 32611

Joseph L. Wooden
U.S. Geological Survey, Menlo Park, California 94050

David W. Mogk
Department of Earth Science, Montana State University, Bozeman, Montana 59717

ABSTRACT

Although it appears that much of the Precambrian nucleus of North America was rapidly assembled between ~1.9 and 1.6 Ga, many questions remain regarding accretion along the southwest margin. One of the more enigmatic features in that region is the Great Falls Tectonic Zone (GFTZ), which separates two Archean cratonic blocks, the Wyoming Province (WP) and the Medicine Hat block (MHB). Various models have been proposed to explain the origin and evolution of the GFTZ, with the ambiguities resulting from the fact that substantial pre-Beltian rocks within the GFTZ are limited to the Little Belt Mountains (LBM). Our recent mapping in the LBM combined with new U-Pb and geochemical data shed new light on this issue.

Mapping and U-Pb zircon geochronologic data for the LBM indicate that most of the exposed basement comprises meta-plutonic rocks emplaced between ~1870 and 1790 Ma. These intrusive rocks have geochemical characteristics typical of subduction-generated magmas suggesting closure of an ocean basin and collision during the Paleoproterozoic. Paleoproterozoic plutonic rocks intruded Archean (~2800-2600 Ma) meta-diorite, Paleoproterozoic pelitic gneisses, and metavolcanic rocks.

Amphibolite or higher grade metamorphism occurred after ~1840-1870 Ma magma emplacement in the northern part of the area and prior to ~1800-1770 Ma cooling. The development of steep WNW-striking foliations accompanied metamorphism. The southern part of the area was affected by amphibolite facies metamorphism, and, near the town of Neihart, was also mylonitized at greenschist facies conditions. Mylonitic and adjacent metamorphosed rocks display a shallow WNW-ESE-trending stretching lineation and sinistral shear-sense indicators. Deformation in this region has not yet been dated.

Structural fabrics in the study area are highly oblique to those in high-grade metamorphic rocks to the south that are believed to record collisional underthrusting of the WP beneath the MHB following ocean basin closure and emplacement of arc-related plutons in the Little Belt Mountains. It is not clear if fabrics in the study area are part of a single WP-MHB collisional orogenic event or if the fabrics record reactivation during accretionary events to the west and/or along the southern margin of the WP.

Introduction

Between ~1900 and 1600 Ma numerous Archean and Proterozoic fragments were rapidly

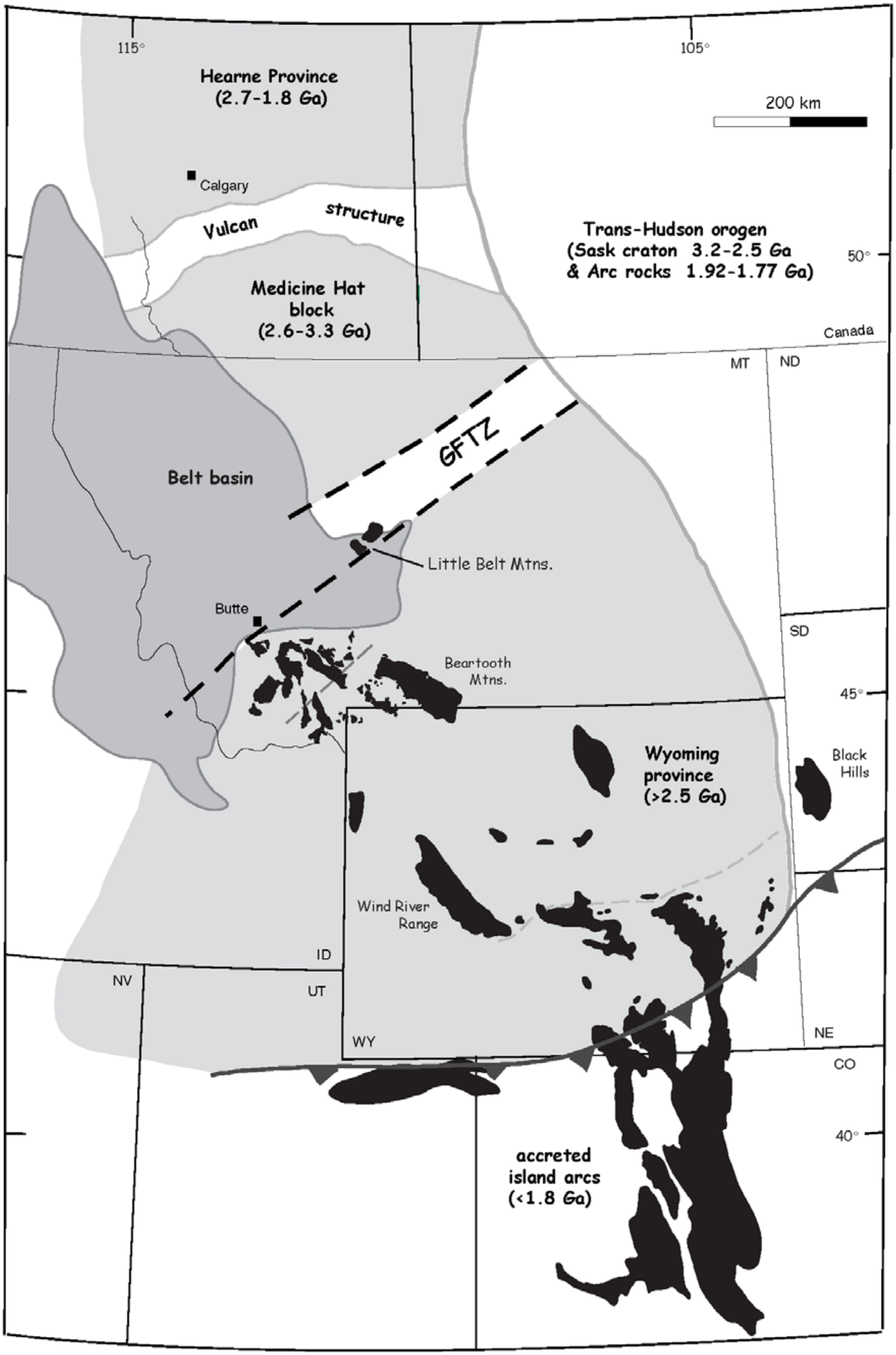


Figure 1. Map showing Precambrian basement provinces of southwestern Laurentian margin and setting of Great Falls Tectonic Zone (GFTZ).

assembled to form Laurentia (e.g., Hoffman, 1988). The nature and timing of accretion along the southwestern margin of Laurentia is poorly understood, largely because the western margin of the craton is buried by Mid-Proterozoic and Phanerozoic strata. In spite of a generally poor understanding of the basement in this region, it has been featured in a number of proposed supercontinent (e.g., Rodinia, Columbia) reconstructions based on geological and geochemical correlations (e.g., Sears and Price, 2000, 2002; Borg and DePaulo, 1994; Karlstrom et al., 1999; Burrett and Berry, 2000). A more complete knowledge of the nature of the crust and timing of assembly is, therefore, critical for meaningful reconstructions of Precambrian continents and Laurentian assembly. Furthermore, the features associated with the Precambrian assembly of North America appear to have had a profound impact on the evolution of Phanerozoic basins, magmatic provinces, thrust and extensional structures, and metallogenesis.

One of the more ambiguous aspects of Laurentian assembly concerns the accretion of the Wyoming Province to the rest of Laurentia. The Wyoming province is a geochemically and geophysically distinct Archean craton (e.g., Wooden and Mueller, 1988; Mueller et al.,1993) that is separated from the Archean Superior Province to the east by the Paleproterozoic Trans-Hudson orogen and is sutured with Proterozoic arcs to the south along the Cheyenne Belt (Fig. 1; Karlstrom and Houston, 1984; Chamberlain, 1998). To the north of the Wyoming Province (WP) lies the Archean Medicine Hat Block (MHB; e.g., Villeneuve et al., 1993). The boundary between the WP and MHB is generally taken to be the Great Falls tectonic zone (GFTZ), which in western Montana is delineated by a zone of NE-trending Proterozoic to Tertiary faults, intrusions, and depositional patterns (O'Neill & Lopez, 1985). These features coincide with a NE-trending zone of basement-related potential field anomalies (Fig. 2) leading to the suggestion that their position is controlled by inherited Precambrian basement structures.

O'Neill & Lopez (1985) initially suggested that the GFTZ originated as a Paleoproterozoic suture between the WP and MHB. On the basis of a lack of geophysical features hypothesized to be characteristic of more extensively studied Paleoproterozoic orogens in Canada, it was subsequently argued by Boerner et al. (1998) that the GFTZ was an Archean structure that was reactivated as an intrcontinental shear zone. Lemieux et al. (2002) argued that filtered potential field anomalies in the MHB and WP are continuous across part of the GFTZ prompting them to suggest that the GFTZ is not a crustal suture. Dipping reflectors in the mantle north of the GFTZ, were imaged by Gorman et al. (2002), who interpreted them as a relic subduction zone that accommodated WP-MHB suturing, which they suggested was Archean based on the arguments of Boerner et al. (1998). Ross (2002), however, interpreted the "relic subduction zone" as Paleoproterozoic based on the age data in Mueller et al. (2002). These widely ranging interpretations highlight the debate of the role of the GFTZ in the assembly of Laurentia. This debate remains unsettled largely because the lack of Precambrian basement exposures along the GFTZ has limited study of the zone to geophysical surveys and xenolith samples from Cenozoic rocks. The only exposures of basement lying directly within the GFTZ occur in the LBM, Little Rocky Mountains of west-central Montana, and scattered thrust-bounded outcrops in SW Montana, which therefore represent the only places to directly assess the role of the GFTZ in Paleoproterozoic assembly of Laurentia.

Despite the importance of these relatively extensive exposures in the LBM, the Pre-Beltian basement has never been completely mapped or studied in any comprehensive way. Previous work has been limited to local mapping, partial lithologic descriptions, geochronology along U.S. Highway 89, and studies aimed at Cretaceous/Tertiary mineralization (Pirsson, 1900; Weed, 1900; Schafer, 1935; Catanzaro and Kulp, 1964; Woodward, 1970; Witkind, 1971; Keefer, 1972; Holm and

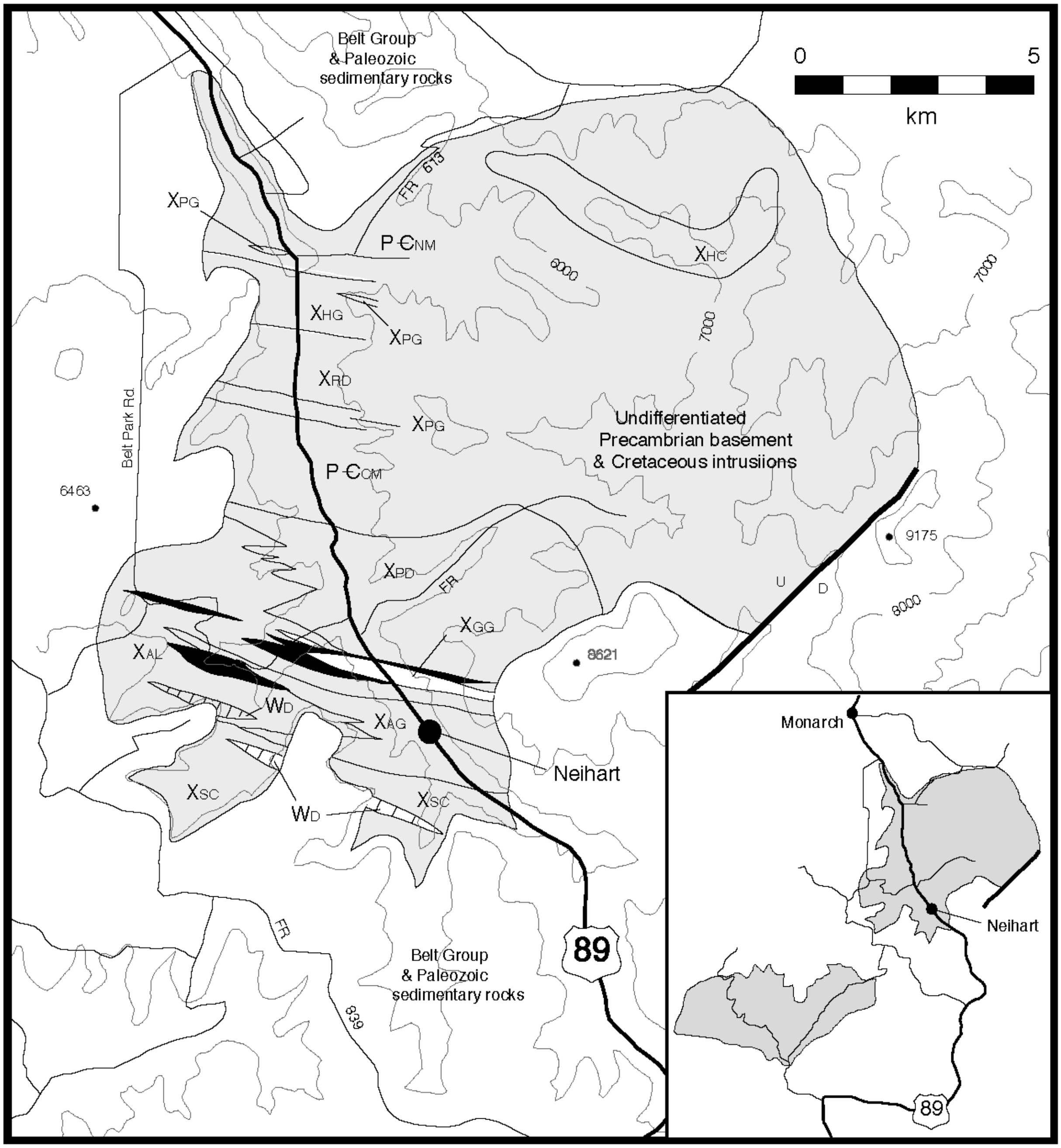


Figure 2. Geological map of the northern area of pre-Beltian basement exposures in the Little Belt Mountains. Inset shows distribution of Pre-Beltian exposures (gray shading) in Little Belt Mountains. White areas are unmetamorphosed Belt and Paleozoic rocks. See Figure 1 for location and Table 1 for map unit symbols, descriptions, and ages.

Schneider, 2002).

Consequently, we have undertaken a study to assess the age, character, structural evolution, and thermal history of basement in the LBM to test models regarding the origin and evolution of the GFTZ and to provide a more complete framework for understanding of the assembly of southwestern Laurentia. Our ongoing study integrates mapping, isotopic dating (U-Pb and $^{40}Ar/^{39}Ar$), geochemical analysis, and structural analysis. In this contribution we present preliminary results of our work, emphasizing the results of mapping and U-Pb ion microprobe zircon dating. These results provide strong evidence supporting models of Paleoproterozoic suturing between the MHB and WP.

Lithology and mapping

Pre-Beltian Precambrian basement in the Little Belt Mountains is exposed in two areas separated by a ridge of Paleozoic and Belt Supergroup sedimentary rocks (Fig. 2). The northern area comprises several mappable, predominantly meta-igneous rock units (Fig. 2), as well as numerous other meta-igneous lithologies that are too small to be mapped. The map units appear to have roughly WNW-ESE map traces, although the eastern extent of the units has not been mapped more than a few kilometers east of U.S. Highway 89. The southern exposures are entirely intrusive rocks. Although numerous lithologies have been identified, the poor exposure and small-scale lithologic variations precludes delineation of the map-scale geometry.

General lithologic descriptions, along with U-Pb ages, are given in Table 1 (at end of paper). On the basis of lithology, field relations, and U-Pb ages we divide the Pre-Beltian rocks into three groups: Paleoproterozoic intrusive rocks, meta-volcanic/sedimentary rocks, and Archean orthogneiss country rocks. In the following sections we summarize the salient features of these groups. All ages reported here are from U-Pb dating of zircons on the ion microprobe at the Stanford-U.S.G.S. facility. Several of the unit names used here are informal terms used here for the first time; others such as the Pinto diorite, Augen gneiss, and Gray gneiss have bee retained from previous work.

Paleoproterozoic intrusive rocks

The northern area is dominated by seven foliated meta-intrusive rock units: Augen gneiss, Gray gneiss, Pinto diorite, Ranger diorite, Helispot granite/migmatite, O'Brien Creek unit, and Hoover Ridge unit (Fig. 2). The compositions range from Bt-granite and Grt-leucogranite to Hbl-Cpx-diorite (all mineral abbreviations from Kretz, 1973). Concordant and upper intercept U-Pb zircon ages from this study combined with two dates from (Mueller et al., 2002) range in age from ~1790 Ma to ~1870 Ma, indicating magma emplacement over a protracted interval.

Major element chemistry for intrusive rocks indicates a calk-alkaline suite (Mueller et al., 2002), and previous and new trace element data show significant enrichments of large-ion lithophiles relative to high-field-strength elements. The geochemical characteristics , including relatively primitive initial Nd isotopic compositions (Mueller et al., 2002), of the Paleoproterozoic intrusive rocks, therefore, suggests magma generation in a subduction zone setting.

The Pinto diorite, because of its distinctive appearance and prominent outcrops, is one of the most notable units in the LBM, appearing on nearly every geologic map of the area. Mineral assemblages in the Pinto diorite include Pl+Hbl+Bt+Qtz+Kfs displaying polygonal recrystallized textures. One of the striking features of the Pinto is the high abundance of gray to mint-green plagioclase megacrysts up to 3 cm across. The megacrysts are locally well aligned producing a foliation or lineation. This foliation, along with dikes (diorite to granodiorite), are locally isoclinally folded where discrete shear zones cut the diorite (Fig. 3b). Twenty-five, low discordance (<10%) ion-probe analyses yield a weighted mean

$^{207}Pb/^{206}Pb$ age of 1864±5 Ma (2) (Mueller et al., 2002).

In the map area south of the Gray gneiss and Pinto diorite exposures, within the northern area, interlayered leucogranite and amphibolite are abundant (Fig. 2). These units are intercalated at a range of scales with other units such as the Augen gneiss and Bt granite of the O'Brien Creek unit. Thus, the map pattern in Figure 2 shows the relative proportions of the different lithologies and is not reflective of the detailed spatial complexities in this region. One of these leucogranites (Red gneiss of Schafer) was dated by Mueller et al. (2002), yielding an upper intercept age of ~1851 Ma.

The Sheep Creek complex (southern area) comprises weakly foliated leucogranite sheets that cut across strongly foliated intrusive units that range from granite to amphibolite. The leucogranite sheets contain only minor amounts of biotite, and locally contain garnet and secondary (?) muscovite. Although the southern area is poorly exposed, it appears that the older (cross-cut) units are limited to outcrop-scale blocks and lenses. A strongly foliated amphibolite yields a concordant age of ~1810 Ma, which is interpreted as an emplacement age. Within the outcrop the amphibolite is cut by leucogranite, thus, 1810 Ma is considered a maximum age for the leucogranites. Direct dating of the leucogranites has proven difficult due to the combination of lack of zircons and post-emplacement alteration and Pb-loss. Overall, the mild deformation and age constraints suggest that the Sheep Creek leucogranites were intruded late in the magmatic and orogenic history of the LBM. The leucogranites are similar in composition to many of the leucogranite veins that are numerous in the southern and western parts of the northern exposure, but it is unknown if they were intruded during the same event.

Overall, the age data collected thus far suggest that units north of the Augen gneiss are the oldest Paleoproterozoic intrusive rocks, yielding ages of ~1870-1840 Ma (Table 1, at end of paper). The only date younger than this is from a melt pod within the Cemetary migmatite (see below; Fig. 2). The Augen gneiss and units to the south (including the Sheep Creek intrusive complex) yield ages between 1790 and 1820 Ma (Table 1).

Meta-sedimentary and meta-volcanic(?) rocks

Although the Paleoproterozoic intrusive rocks make up the majority of the volume of the Pre-Beltian basement exposed in the study area, other lithologic packages have been mapped. The most prominent of these occurs in intrusive contact with the northern margin of the Pinto Diorite. These rocks comprise a compositionally variable layered migmatite sequence. Most compositions contain assemblages with two feldspars, variable amounts of clinopyroxene, biotite, hornblende, and quartz. The rocks display meter-scale lithologic variations. We interpret these features as an indication of a meta-volcanic origin for this unit, which we refer to as the Cemetary migmatite. We have not yet dated the meta-volcanic protoliths. A leuogranite pod from the migmatite yielded an upper intercept age of 1817±17 Ma and probably represents the time of migmatization and melt injection. We are currently attempting to date the protolith, which combined with geochemical analysis, will determine whether this sequence represents a Paleoproterozoic volcanic sequence that pre-dates or is related to the intrusive suite.

Along the northern edge of the Cemetary migmatite, pelitic gneiss occurs as a distinct mappable band between the meta-volcanic rocks and the Ranger diorite. Compositionally similar rocks also occur as narrow discontinuous layers near the boundary between the Helispot granite and Northern migmatite. Notable minerals in the pelitic gneisses include garnet, sillimanite, spinel, K-feldspar, cordierite, and anti-perthite, which indicate very high temperatures of metamorphism. Preliminary U-Pb age data were collected for eleven zircons in order to constrain depositional and metamor-

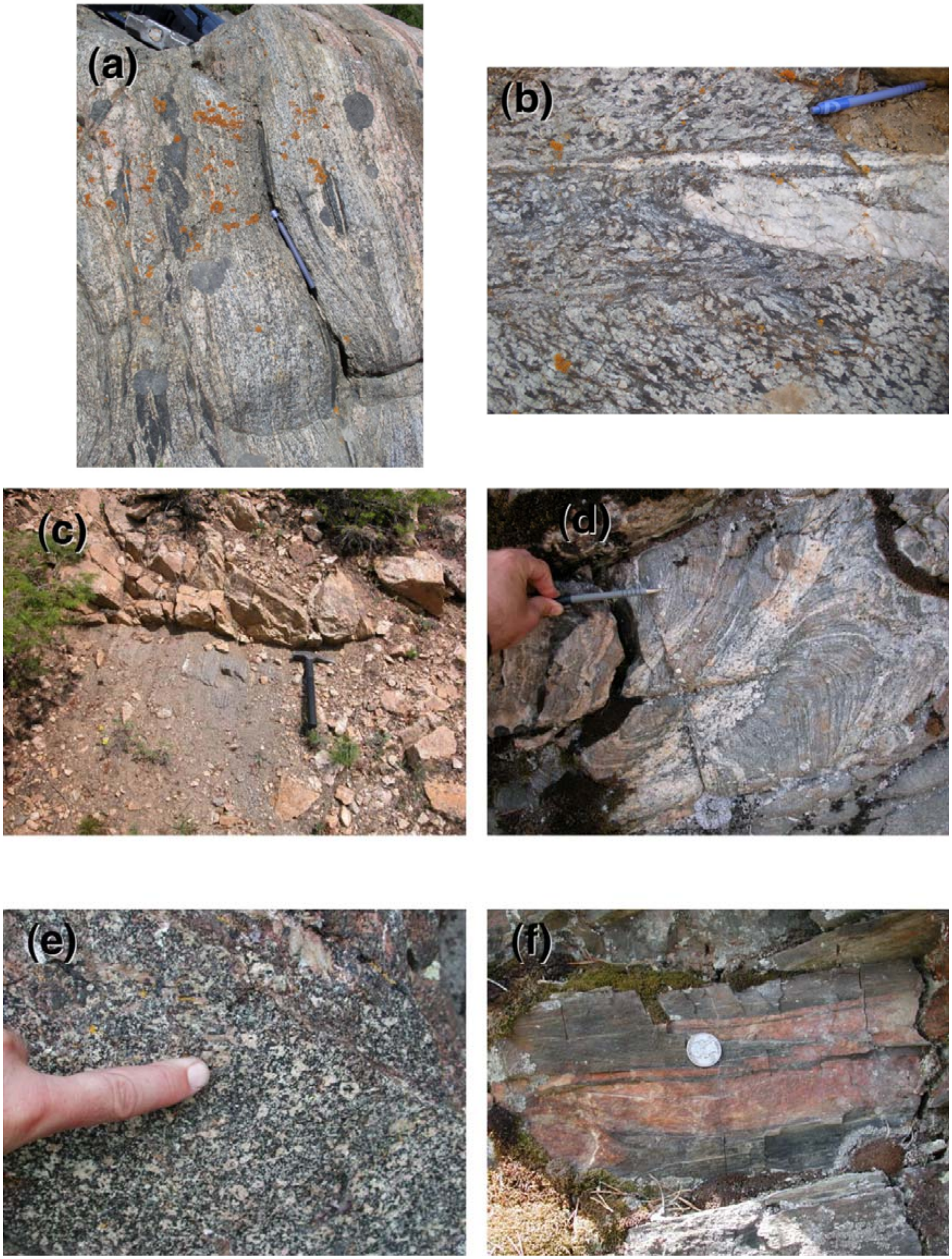


Figure 3. Photographs of basement outcrops. (a) High-grade felsic gneiss with amphibolite boudins; Northern migmatite. (b) Foliated Pinto diorite isoclinally folded, along with vein, in small scale shear zone. (c) Strongly banded dioritic gneiss cut by weakly foliated leuocogranite; Sheep Creep intrusive complex (d) Leucogranite melt pods and layers in folded metavolcanic(?) rocks; Cemetary migmatite. (e) Archean meta-diorite. (f) Isoclinally folded red-stained leucogranite vein in mylonitized intermediate plutonic rocks. Photo looking downdip of a moderately S-dipping foliation. Outcrop on ridge west of the town of Neihart.

phic ages within the pelitic gneiss. The sample yielded concordant ages at ~2600 Ma, ~2360-2380 Ma, and ~1850 Ma. Thus, this unit is clearly Paleoproterozoic, having been deposited after 2360 Ma (detrital zircon age) and prior to metamorphism (~1817 Ma). More analyses are being done to decipher whether the youngest concordant ages reflect deposition of detrital zircons or the age of metamorphism.

Archean meta-diorite

Mappable exposures of variably foliated meta-diorite are found near the southwest edge of the northern area. A sample from this meta-diorite yielded discordant U-Pb ages that indicate emplacement between 2600 and 2800 Ma. Discontinuous pods of similar meta-diorite can be found throughout the southern part of the northern area near, and west of the town of Neihart.

Structure and metamorphism

All units in the northern area contain structural fabrics and variably recrystallized textures indicating post-emplacement deformation and metamorphism. The northernmost units have reached granulite facies conditions as evidenced by the presence of garnet, cordierite, spinel, and sillimanite in pelitic gneisses and by the presence of orthopyroxene in the Hoover Ridge unit to the east. The presence of spinel + cordierite suggests that the metamorphic peak may have been reached at low pressures (<4-5 kbar; e.g., Spear et al., 1999). Low metamorphic pressures may also explain the lack of garnet in the amphibolites.

An important question regards the timing of emplacement relative to the peak of metamorphism. The Helispot granite contains highly recrystallized textures as well as local disruption of fabrics and nebulous quartzo-feldspathic regions in the northernmost outcrops that give the outcrops a chaotic migmatitic appearance. Furthermore, some of the felsic high-grade migmatitic gneisses in the northern migmatite are compositionally similar to the Helispot granite, perhaps suggesting it is a high-grade equivalent. Recrystallized polygonal hornblende and feldspar textures in foliated Pinto diorite samples also indicate post-emplacement amphibolite-facies or higher grade metamorphism. The ~1817 Ma age on a leucogranite pod in the Cemetary migmatite may establish the age of metamorphism. If these interpretations are correct, they suggest that the peak of metamorphism was reached following 1870-1840 Ma emplacement of the northern most intrusive rocks. However, Dahl et al. (2000) report garnet growth at 11.86 Ma in a metapelite from the LMB, so that some metamorphism did accompany the intrusion of the diorites. South of the Pinto diorite, amphibolites do not contain clino- or orthopyroxene, suggesting middle to lower amphibolite conditions. Thus, regional metamorphic grade appears to decrease toward the south.

Foliations in the northern part of the northern area formed during peak metamorphic conditions. On the map scale, gneissic foliations generally strike NW and are subvertical to shallowly NE dipping. At the outcrop scale, the structural fabrics in the northernmost migmatitic units have variable orientations with complicated geometries.

South of the Pinto diorite, rocks show prominent lower grade deformational fabrics compared with those farther north. Many of the amphibolites contain well-developed subhorizontal WNW-trending lineations and interlayered leucogranites locally have strong foliations. Intensely developed mylonitic fabrics are prominent near the town of Neihart, prompting our reference to this area as the Neihart mylonite zone. Mylonitization has affected a range of lithologies including Augen gneiss, amphibolites, pegmatite, and granitic rocks. The mylonite has a strongly banded character due to intense transposition of leucogranite veins within more mafic meta-igneous compositions (Fig. 3f). Outcrops display a flaggy appearance in the zones of highest strain. Mylonitization occurred at amphibolite-facies conditions, but continued to greenschist-facies conditions, as evidenced by the replacement of hornblende and biotite by actinolite, epidote, and chlorite;

feldspars are also partially replaced by white micas. The strike of the mylonitic foliation ranges from WNW to WSW with moderate to steep southward dips. Mylonites are locally folded on the outcrop scale. Lineations are generally shallowly plunging with variable WSW to WNW trends. Shear-sense indicators such as C-S fabrics, porphyroclast tails, and quartz grain-shape fabrics indicate a sinistral shear sense, consistent with shear bands in the Gray gneiss to the north.

Leucogranite sheets of the Sheep Creek complex display steeply dipping, weakly to moderately developed foliations that strike WNW, parallel to strongly foliated units that they cross-cut. Leocogranite veins that have injected the older units are locally folded with the foliation in both units axial-planar to the folds. Microstructural observations are suggestive of lower amphibolite-facies conditions during deformation. We interpret these relations as synkinematic injection of the leucogranite sheets into crust undergoing horizontal NNE-SSW shortening.

In the northern area, the age of the high-grade metamorphism and associated deformation is constrained by the age of the intrusive units and the ~1800-1770 Ma hornblende and biotite $^{40}Ar/^{39}Ar$ obtained by Holm and Schneider (2002). The $^{40}Ar/^{39}Ar$ age constraint only applies to the Gray gneiss and units to the north (sample locations of Holm and Schneider, 2002). We are in the process of dating areas to the south, including the Neihart mylonite zone and Sheep Creek complex, which will allow evaluation of the spatial distribution of deformation and cooling across the area.

Implications for accretion of the Wyoming Province

The most important result of our study to date is the documentation of voluminous Paleoproterozoic magmatism within the LBM segment of the GFTZ. Given the subduction zone geochemical signatures of these rocks, the age data provide strong evidence for closure of an ocean basin between the MHB and WP during the Paleoproterozoic. Thus, the age and geochemical data are in agreement with models depicting the Great Falls tectonic zone as a Paleoproterozoic suture (e.g., O'Neill, 1998) and do not support the Archean suture or intracontinental transform models (e.g., Boerner et al., 1998).

One of the questions critical to understanding the accretion of the Wyoming province is the nature of the crust upon which the Little Belt arc was built. Possibilities include the MHB, the WP, or an intervening crustal fragment. The Medicine Hat block is not exposed so the only data available come from xenoliths and drill cores. Six of seven samples of deformed MHB intrusive rocks (from drill core & xenoliths) yield U-Pb zircon ages between 2612 and 2840 Ma, while one sample yielded an age of 3278 Ma (Davis et al., 1985; Ross et al., 1991, Villenuave et al., 1993). These ages are similar to the ~2610 to 2813 Ma $^{207}Pb/^{206}Pb$ ages from the meta-diorite intruded by Paleoproterozoic magmas in the Little Belts. In contrast, meta-igneous rocks of the northern WP yields ages of 3100-3300 Ma, whereas 2600-2800 Ma rocks are most common in the southernmost part of the province. Thus, our present data are most consistent with a model involving northward subduction beneath the Medicine Hat block prior to terminal collision with the Wyoming province, although we cannot rule out the Little Belt arc having formed on an intervening crustal fragment. This interpretation is also consistent with the north-dipping mantle reflections interpreted by Gorman et al. (2002) as relic subduction zones. Ongoing Pb and Nd isotopic analysis is aimed at fully documenting the geochemical character of the Paleoproterozoic intrusions and testing whether there is any contamination from the geochemically distinct Wyoming province.

Paleoproterozoic thermo-tectonism has long been recognized along the northwest margin of the Wyoming province. This recognition stems largely from the widespread Paleoproterozoic K-Ar and $^{40}Ar/^{39}Ar$ ages from Archean rocks in

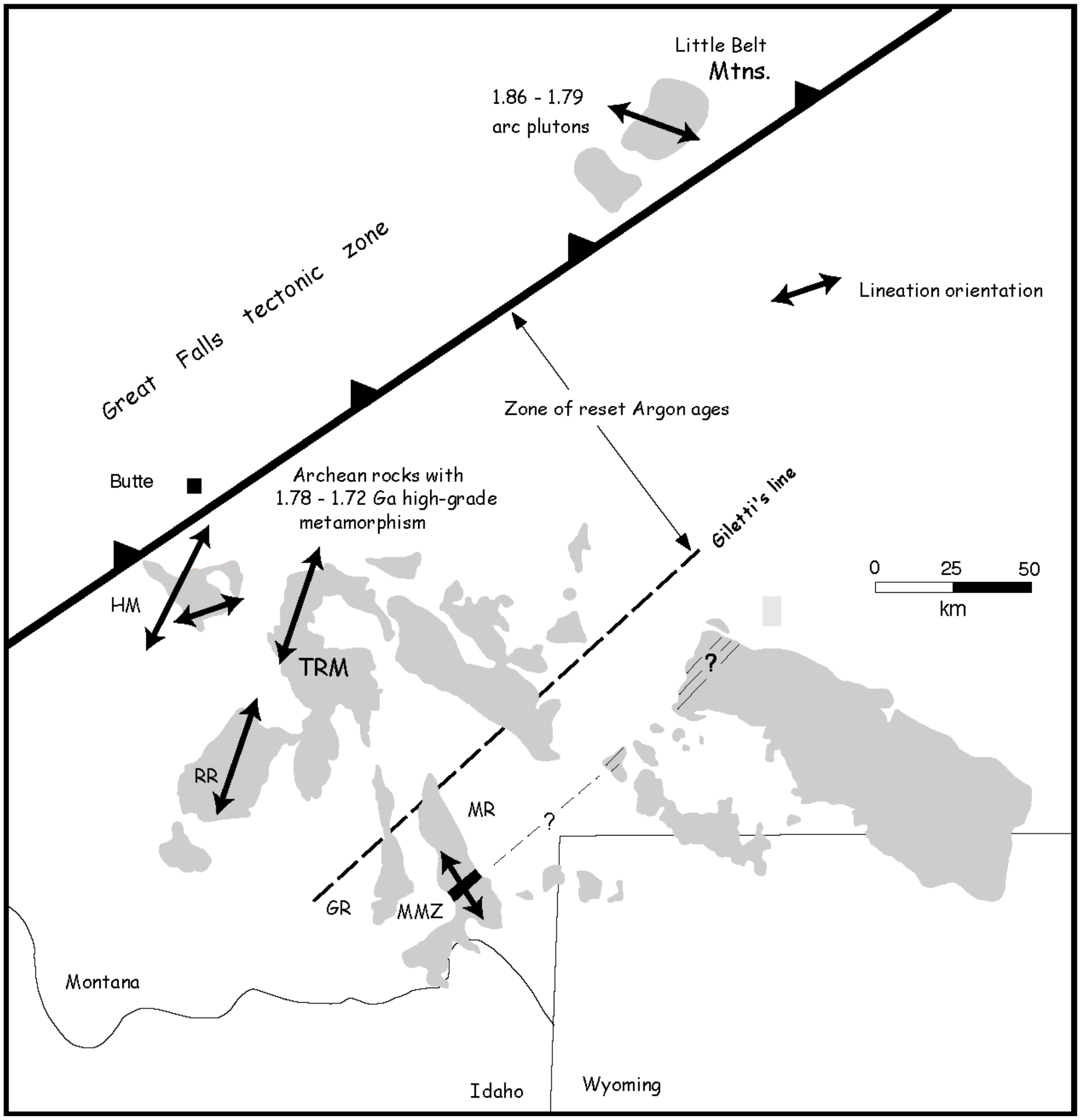


Figure 4. Map outlining orientation of stretching lineations formed during Paleoproterozoic tectonism. GR, Gravelly Range; HM, Highland Mountains; MMZ, Madison mylonite zone; MR, Madison Range; RR, Ruby Range; TRM, Tobacco Root Mountains. Data from Erslev and Sutter (1990), Harms et al. (2004b), Garihan (1979), Duncan (1976), and this study.

southwestern Montana (Fig. 4; Tobacco Root Mtns., Highland Mtns., Ruby Range; e.g., Giletti, 1966). The evidence for thermal/tectonic reworking of Archean rocks and the nature of structures and metamorphism has led to the suggestion, first by O'Neill and Lopez (1985) and O'Neill (1998), and later by Harms et al. (2004b), that the basement uplifts of SW Montana represent different elements of a single orogenic belt.

South of the Little Belt Mountains, the Tobacco Root and Highland Mountains and the Ruby Range, record high temperature metamorphism (Fig. 4) and are interpreted as the metamorphic core of the orogen, where the northwestern margin of the Wyoming province was underthrust beneath a colliding block (perhaps the MHB). The most detailed work has been done in the Tobacco Roots, where recent data has documented penetrative deformation and accompanying high-P/high-T metamorphism around 1780 Ma evolving to lower-P/high-T conditions and cooling by ~1710 Ma (Brady et al., 2004; Cheney et al., 2004a,b; Harms et al., 2004a,b; Mueller et al., 2004). Farther south, mylonite zones in the Gravelly and Madison Ranges have been interpreted as basement shear zones in the foreland region of the orogen (e.g., Erslev and Sutter, 1990; O'-Neill, 1998).

Given the subduction-related magmatism that dominates the LBM, this region is considered to be the continental (?) arc under which the Wyoming province was thrust. Thus, it is considered allochthonous with respect to the Wyoming province. The peak of tectono-thermal activity in the Tobacco Roots post-dates magmatism, deformation, and cooling documented thus far in the northern part of the LBM. Thus, in typical orogenic fashion, deformation/ metamorphism may have progressed from the colliding terrane (LBM) during or shortly after arc magmatism outward (south- or southeastward) increasingly affecting rocks of the underthrust margin (Archean rocks of the northern WP). We note, however, that we currently do not have an age for the youngest leucogranites of the Sheep Creek intrusive complex or the Neihart mylonite zone. Thus the spatial evolution of deformation and metamorphism may prove to be more complicated upon completion of our dating of intrusions and structures of the southern LBM.

One of the problems with this simplified view of the orogen concerns the variation in Paleoproterozoic structural trends within the different basement blocks of southwestern Montana (Fig. 4). The foreland Madison mylonite zone with a thrust sense of displacement contains NNW-trending lineations, which are approximately perpendicular to the northeast trend of the GFTZ suture assumed from the orientation of potential field anomalies. Lineations within the metamorphic core (Highland Mountains, Tobacco Root Mountains, and Ruby Range) generally trend NNE and are assumed to record large shear strains (Tobacco Roots, Harms et al., 2004a). These lineations would indicate transport that was highly oblique to the suture zone and, if formed during convergence, would have imparted a sinistral sense along the suture. In the LBM the most prominent lineations trend WNW, approximately perpendicular to those in the metamorphic core. The variability in structural orientations can be explained by (1) structures forming at different times during a kinematically evolving collision zone, (2) partitioning into oblique, orogen-normal, and orogen-parallel components, (3) local late-orogenic folding (perhaps in the LBM) or post-orogenic rotation, and/or (4) structures recording the effects of distant accretionary events such as in areas west of the study area and/or along the southern margin of the Wyoming Province (Karlstrom and Houston, 1984; Chamberlain, 1998), in addition to MHB-WP collision. Our continued dating and kinematic analysis of structural fabrics in the southern LBM, combined with study of crustal amalgamation to the west will provide important tests of these models.

Concluding remarks

New mapping and geochronologic data have shed light on the origin of the Great Falls tectonic zone and the accretionary history along the SW Laurentian margin. In particular, data and observations reveal that a large fraction of the exposed pre-Beltian basement rocks consist of compositionally variable Paleoproterozoic (~1870-1790 Ma) intrusive rocks with a subduction zone geochemical character. These data are most consistent with a model involving closure of an ocean basin and associated arc construction, followed by collision between the Wyoming Province and Medicine Hat block all during the Paleoproterozoic. Ongoing geochemical studies are aimed at testing whether the Little Belt arc was constructed on the Medicine Hat block, Wyoming Province or a fragment of intervening crust.

Metamorphism and structural fabrics in the Little Belts record WP-MHB convergence, but may also record more distal events that affected the southwest margin of Laurentia. We are continuing to date the Little Belt fabrics to establish the diachroneity of deformation both within the Little Belt exposures and between the Little Belts arc and northern margin of the Wyoming Province. Establishment of the chronology of plutonism, metamorphism, and deformation will greatly enhance our understanding of the accretionary history of the southwest margin of Laurentia. This will in turn provide critical constraints for testing models for the construction of Rhodinia and subsequent dispersal of continental fragments.

Acknowledgements

Funding for our research was provided by NSF grant EAR 0106592. We thank Mike Hartley, Daniel Gorman and Monica Lipinski for assisting with sample preparation.

References Cited

Boerner, D., Craven, J., Kurtz, R., Ross, G., and Jones, F., 1998, The Great Falls Tectonic Zone: Suture or intracontinental shear zone?: Canadian Journal of Earth Science, v. 35, p. 175-183.

Borg, S. and Depaolo, D., 1994, Laurentia, Australia, and Antarctica as a Late Proterozoic supercontinent: Constraints from isotopic mapping: Geology, v. 22, p. 307-310.

Brady, J., Kovaric, d., Cheney, J., .Jacob, L., and King, J., 2004, 40Ar/39Ar ages of metamorphic rocks from the Tobacco Root Mountains, southwest Montana, in Brady, J., Burger, H., Cheney, J., and Harms, T., eds., Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America Special Paper 377, p. 131-149.

Burrett, C. and Berry, R., 2000, Proterozoic Australia-western United States (AUSWUS) fit between Laurentia and Australia. Geology, v. 28, p. 103-106.

Catanzaro, E., and Kulp, J., 1964, Discordant zircons from the Little Belt (Montana), Beartooth (Montana) and Santa Catalina (Arizona) Mountains: Geochimica et Cosmochimica Acta, v. 28, p. 87-124.

Chamberlain, K., 1998, Medicine Bow orogeny: Timing of deformation and model of crustal structure produced during continent-arc collision, ca. 1.78 Ga, southeastern Wyoming: Rocky Mountain Geologist, v. 33, p. 259-277.

Cheney, J., Brady, J., Tierney, K., DeGraff, K., Mohlman, H., Frisch, J., Hatch, C., steiner, M., Carmichael, S., Fisher, R., Steffen, K., Cady, P., Lowell, J., Archuleta, A., Hirst, J., Wegmann, K., and Monteleone, B., 2004a, Proterozoic metamorphism of the Tobacco RootMountains, SW Montana, in Brady, J., Burger, H., Cheney, J., and Harms, T., eds., Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America Special Paper 377, p. 105-129.

Cheney, J.T., Webb, A.A., Coarth, C.C., and McKeegan, K, 2004b, In situ ion microprobe 207Pb/206Pb dating of monazite from Precambrian metamorphic suites, Tobacco Root Mountains, Montana, in Brady, J.B., Burger, H.R., Cheney, J.T., and Harms, T.A., eds., Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America Special Paper 377, p. 151-179.

Davis, W., Berman, R., and Kyarsgaard, B., 1995, U-Pb geochronology and isotopic studies of crustal xenoliths from the Archean Medicine Hat block, northern Montana and southern Alberta: Paleoproterozoic reworking of Archean crust, *in* Ross, G.M., ed., 1995 Alberta Basement Transects Workshop: Lithoprobe Secretariat, University of British Columbia, Lithoprobe Report 47, p. 330-335.

Duncan, M.S., 1976, Structural analysis of the pre-Beltian metamorphic rocks of the southern Highland Mountains, Madison and Silver Bow Counties, Montana. Ph.D. thesis, Indiana University, 222 pp.

Erslev, E. and Sutter, J., 1990, Evidence for Proterozoic mylonitization in the northwestern Wyoming province: Geological Society of

America Bulletin, v. 102, p. 1681-1694.

Garihan, J.M., 1979, Geology and structure of the central Ruby Range, Madison County, Montana. Geological Society of America Bulletin Part II, v. 90, p.695-788.

Giletti, B, 1966, Isotopic ages from southwestern Montana. Journal Geophysical Research, v. 71, p. 4029-4036.

Gorman, A., Clowes, R., Ellis, R., Henstock, T., Spence, G., Keller, G., Levander, A., Snelson, C., Burianyk, M., Kanasewich, E., Asudeh, I., Zoltan, H., and Miller, K., 2002, Deep Probe: imaging the roots of western North America: Canadian Journal of Earth Science, v. 39, p. 375-398.

Harms, T.A., Burger, H.R., Blednick, D.G., Cooper, J.M., King J.T., Owen, D.R., Lowell, J., Sincock, M.J., Kranenburg, S.R., Pufall, A., and Picornell, C.M., 2004a, Character and origin of Precambrian fabrics and structures in the Tobacco Root Mountains, Montana, in Brady, J. B., Burger, H.R., Cheney, J.T., and Harms, T. A., eds., Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America Special Paper 377, p. 203-226.

Harms, T.A., Brady, J.B., Burger, H.R., Cheney, J. T., 2004b, Advances in the geology of the Tobacco Root Mountains, Montana, and their implications for the history of the northern Wyoming province, in Brady, J.B., Burger, H.R., Cheney, J.T., and Harms, T.A., eds., Precambrian Geology of the Tobacco Root Mountains, Montana: Geological Society of America Special Paper 377, p. 227-243.

Hoffman, P., 1988, United plates of America, the birth of a craton: Annual Reviews of Earth and Planetary Sciences, v. 16, p. 543-603.

Holm, D and Schneider, D, 2002, $^{40}Ar/^{39}Ar$ evidence for ca. 1800 Ma tectonothermal activity along the Great Falls tectonic zone, central Montana. Can. Jour. Ea. Sci., v. 39, p. 1719-1728.

Karlstrom, K., and Houston, R., 1984, The Cheyenne Belt: Analysis of a Proterozoic suture in southern Wyoming: Precambrian Research, v. 25, p. 415-446.

Karlstrom, K.E., Harlan, S.S., Williams, M.L., McLelland, J., Geissman, J.W., and Ahall, K.I., 1999, Refining Rodinia: Geologic evidence for the Australia -Western U.S. connection for the Proterozoic: GSA Today, v. 9, no. 10, p. 1-7.

Keefer, W.R., 1972, Geologic map of the west half of the Neihart 15-minute quadrangle, Central Montana.. U.S.G.S. Miscellaneous Geologic Investigations Map I-726, scale 1:62,500.

Kretz, R., 1973, Symbols for rock-forming minerals, American Mineralogist, v. 68, 277-279.

Lemieux, S., Ross, G., and Cook, F., 2000, Crustal geometry and tectonic evolution of the Archean crystalline basement beneath the southern Alberta Plains, from new seismic reflection and potential-field studies: Canadian Journal of Earth Sciences, v. 37, p. 1473-1491.

Mueller, P., Heatherington, A., Kelly, D., Wooden, J., and Mogk, D., 2000, The Great Falls tectonic zone and its role in the Paleoproterozoic assembly of southern Laurentia. Geological Society of America Abstracts with Programs, v. 32, no. 7, p. A318.

Mueller, P.A., Shuster, R., Wooden, J., Erslev, E., and Bowes, D., 1993, Age and composition of Archean crystalline rocks from the southern Madison Range: Implications for crustal evolution in the Wyoming craton: Geological Society of America Bulletin, v. 105, p. 437-446.

O'Neill, J.M., 1998, The Great Falls tectonic zone, Montana-Idaho: An early Proterozoic collisional orogen beneath and south of the Belt Basin, *in* Berg, R.B., ed., Belt Symposium III – 1993: Montana Bureau of Mines and Geology Special Publication 112, p. 222-228.

O'Neill, J.M., and Lopez, D., 1985, Character and regional significance of Great Falls tectonic zone, east-central Idaho and west-central Montana: American Association of Petroleum Geologists Bulletin, v. 69, p. 437-447.

Pirsson, L., 1900, Petrography of the igneous rocks of the Little Belt Mountains. U.S. Geological Survey, 20th Annual Report, Pt. III, p. 463-581.

Sears, J.W., and Price, R.A., 2000, New look at the

Siberian connection: No SWEAT: Geology, 28, 423-426.

Sears, J.W., and Price, R.A., 2002, Thy hypothetical Mesoproterozoic supercontinent Columbia: implications of the Siberian-West Laurentian connection. Gondwana Research, 5, 35-39.

Schafer, P., 1935, Geology and ore deposits of the Niehart mining district, Cascade County, Montana: Montana Bureau of Mines and Geology Memoir 13, 62 p.

Spear, F.S., Kohn, M.J., and Cheney, J.T., 1999, P-T paths from anatectic pelites. Contributions to Mineralogy and Petrology, v. 134, p. 17-32.

Villeneuve, M., Ross, G., Theriault, R., Miles, W., Parrish, R., and Broome, J., 1993, Tectonic subdivision and U-Pb geochronology of the crystalline basement of the Alberta basin, western Canada: Geological Survey of Canada Bulletin 447, 86 p.

Weed, W., 1900, The geology of the Little Belt Mountains, Montana: U.S. Geological Survey 20th Annual Report, Pt. III, p. 257-461.

Wooden, J., and Mueller, P., 1988, Pb, Sr, and Nd isotopic compositions of a suite of Late Archean, igneous rocks, eastern Beartooth Mountains: Implications for crust-mantle evolution: Earth and Planetary Science Letters, v. 87, p. 59-72.

Woodward, L.A., 1970, Time of emplacement of Pinto diorite, Little Belt Mountains, Montana. Wyoming Geological Association Earth Science Bulletin, p. 15-26.

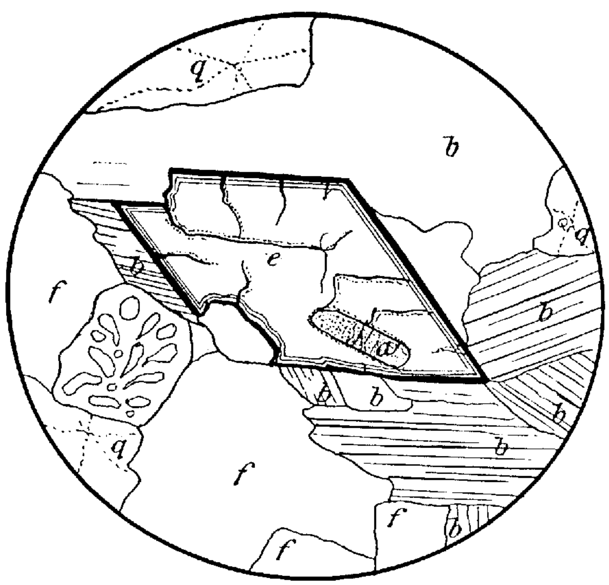
TABLE 1.

Map unit symbol	Unit	Description	U-Pb age & rock type dated
		Paleoproterozoic intrusive rocks	
XHR	Hoover Ridge Unit	Contains several rock types not mapped in detail. Includes Coarse-grained Grt-leocogranite, diorite, Opx-bearing coarse-grained granite, medium-grained variably banded Opx granitic gneiss, and amphibolite.	1846 ± 9 Ma (ui & Pb/Pb) Grt-leucogranite
PCNM	Northern migmatite	Incudes numerous lithologies with abundant granodioritic to dioritic gneiss and amphibolite lenses and boudins. Bt-bearing granitic gneiss that compositionally resembles the Helispot granite is common (Fig. 3a). Contains some granitic and pegmatitic intrusive veins, as well as locally derived melt zones. Amphibolites contain more pyroxene than Cemetary migmatite to the south.	-
XHG	Helispot granite	Weakly to strongly foliated and recrystallized granite to granodiorite. Assemblages contain Kfs+Pl+Qtz in addition to either Bt or Hbl+Bt±Cpx. Locally contains strongly foliated recrystallized porphyroclastic mylonitic textures. In the northerly regions, foliations are disrupted and chaotic and exhibit features suggestive of re-melting	1842 ± 42 Ma (ui)
XRD	Ranger Diorite	Coarse-grained, moderately to strongly foliated diorite. Typical assemblages include Pl+Kfs+Cpx+Bt±Hbl with minor quartz. Spatially associated with fine-graned dioritic rocks.	1856 ± 9 Ma (ui & Pb/Pb)
XPD	Pinto diorite	Coarse-grained unit with abundant gray to mint green megacrysts. Contains primary assemblage of Pl+Hbl+Bt+Qtz+Kfs. Unit is variably foliated with polygonal recrystallized microtexture. See figure 3b and text.	1864 ± 5 Ma (Pb/Pb)[a]
XGG	Gray gneiss	Medium-grained, homogeneously well-foliated and locally well-lineated granodiotic rock containing Hbl, Bt, Pl, Kfs, and Qtz. Finer grained and slighty more mafic rocks, as well as amphibolite, are commonly found near the margins.	1867 ± 6 Ma (Pb/Pb)[a]
XAL	Amphibolite and leucogranite	Alternating bands/lenses of well lineated fine- to medium-grained amphibolite and leucogranite. Leucogranite is fine- to medium-grained and variably red- or pink-stained. Granites contain little or no biotite, but locally contain garnet. Small zones of Hbl-bearing granitic rock also ocur in southwest part of the northern basement exposures.	1851 ± 11 Ma (ui)[a] leucogranite
XAG	Augen gneiss	Moderately well foliated to strongly mylonitic orthogneiss. Contains distinct red/pink K-feldspar megacrysts/ porphyroclasts up to 2 cm across. Feldspars are set in fine grained dark gray grain-size reduced matrix containing Hbl and Bt.	1791 ± 10 Ma (ui & Pb/Pb)
XSC	Sheep Ck. Intrusive complex	Gently dipping leucogranite sheets that cut compositionally variable strongly foliated units. Leucogranites are variably pink/red stained, weakly to moderately foliated, and locally contain Grt and Ms. Older units are more strongly foliated, range from granite to amphibolite, and are cut by leucogranite (Fig. 3c)	1817 ± 11 Ma (Pb/Pb) amphibolite

TABLE 1, CONTINUED.

Map unit symbol	Unit	Description		U-Pb age & rock type dated
		Paleoproterozoic metasedimentary rocks and meta-volcanic rocks		
XAP	Aspen pelitic gneiss	Migmatitic pelitic paragneiss with local melt layers and melt injections. Occurs as narrow lenses near the Helispot granite and northern migmatite boundary as well as distinct mappable band that crosses U.S. 89 immediately south of Ranger diorite. Contains large garnets up to 3 cm across that are commonly surrounded by plagioclase. Other minerals include sillimanite, cordierite, spinel, biotite, and antiperthitic plagioclase.		
PCCM	Cemetary migmatite	Locally banded migmatite displaying leucogranite melt pods and layers (Fig. 3d). Migmatite is derived from igneous protoliths with varying compositions. Metamorphic assemblages range from Hbl+Pl±Cpx±Qtz amphibolite to more felsic Pl+Kfs+Qtz+Bt+Cpx±Hbl. Textures and composition vary on the outrop scale.		1817 ± 16 Ma (ui) leucogranite pod
		Archean rocks		
Wmd	Archean metadiorite	Medium to coarse-grained black and white plutonic rock with a primary assemblage of Pl+Hbl+Bt, with minor quartz and K-feldspar (fig. 3e). Displays distinct white weathering plagioclase phenocrysts/porphyroclasts up to 1 cm across. Commonly not strongly foliated but contains mm-scale shear bands. Metamorphic minerals include blue-green Hbl rims and actinolite and minor calcite. Occurs as prominent lense or pod on the ridge between Graveyard Gulch and upper Harley Creek. Small lenses of compositionally similar metadiorite occuring east and southeast of these exposures may be part of the same unit.		~2600 - 2800 Ma

Notes: all mineral abbreviations from Kretz (1973). “a” indicates date from Mueller et al. (2002). “ui” denotes upper intercept age. Pb/Pb indicates $^{207}Pb/^{206}Pb$ age


q
b
f
e
b
q
b
a
b
b
b
q
f
b
f
b
f

Upper Cretaceous Maudlow and Sedan Formations of the Livingston Group and Their Regional Significance, west-central Montana

Betty Skipp
U.S. Geological Survey (Retired)

Introduction

Many advances have been made in refining the Late Cretaceous timescale in western Montana (Obradovich, 1993; Cobban, 1993; Dyman, and others, 1994; Dyman, and others, 1995) since the publication of the definitions of the Maudlow and Sedan Formations of the Livingston Group (Skipp and McGrew, 1972, 1977; Skipp and Peterson, 1965; Skipp, 1977). This paper updates, extends, and adjusts some of the older correlations based on new and recalculated radiometric ages, and paleontology.

Adjustments in the correlation of volcanic units from the Elkhorn Mountains Volcanics on the west to the lower part of the Livingston Formation in the Madison Range, as well as the Livingston Group on the southern and western margins of the Crazy Mountains Basin (fig.1) show that most primary volcanic units and many volcaniclastic units of the formations of the Livingston Group are correlative with the Elkhorn Mountains Volcanics (EMV). Some of the primary volcanic rocks in the Madison Range, however, may have had local sources (Tysdal and others, 1987).

One andesitic welded tuff (ignimbrite) unit dated at about 80 Ma (Tilling, 1974) ties isolated remnants of the volcanic field in western Montana to the marine sequences of central Montana. This unit forms Member D of the Maudlow Formation, the Welded Tuff Member of the Sedan Formation (Skipp and McGrew, 1977), the "grey welded tuff" of the Madison Range (Tysdal, 1987), and a prominent lower tuffaceous zone in the Cokedale Formation (Roberts, 1972). This lower tuffaceous zone contains a local marine fauna of the *Baculites obtusus* faunal zone in the Livingston area (Roberts, 1972, p. C39; Cobban, 1993), the same faunal zone that is recognized in the lower part of the marine Claggett Shale in central Montana (Dyman and others, 1994).

Both the welded tuff unit, in the Madison Range (Tysdal and Nichols, 1991) and the Welded Tuff Member of the Livingston Group along the western margin of the Crazy Mountains Basin (Skipp and McGrew, 1977) are dated radiometrically at about 79.8 Ma. These welded tuffs most resemble thin andesitic tuff in the lower part of the EMV (Rutland and others, 1989). The voluminous rhyolitic tuffs of the middle part of the EMV are not present as welded tuffs in the Livingston Group, but probably are represented by airfall tuffs in the middle part of the Cokedale Formation and its correlatives.

The abundant tuffs in the Sulphur Flats Sandstone Member at the base of the Miner Creek Formation were correlated by Roberts (1972) with the marine *Baculites compressus* and *Baculites cuneatus* faunal zones of the Lennep Sandstone. Recent examination of palynological assemblages from the Sulphur Flats Member at its type section (Tysdal and Nichols, 1991) corroborates this correlation and indicates that the Lennep Sandstone Member at the top of the Sedan Formation must correlate with the Sulphur Flats Sandstone Member at the base of the Miner Creek Formation. The tuffs in the Lennep Sandstone and Sulphur Flats Sandstone, therefore, are approximately 73.35 + 0.39 Ma (Obradovich, 1993) and are just slightly younger than the Adel Mountain Volcanics dated recently at 75 to 73.7 Ma (Harlan and others, in press). No sources for the tuffs of the upper part of the Livingston Group which includes the upper part of the Miner

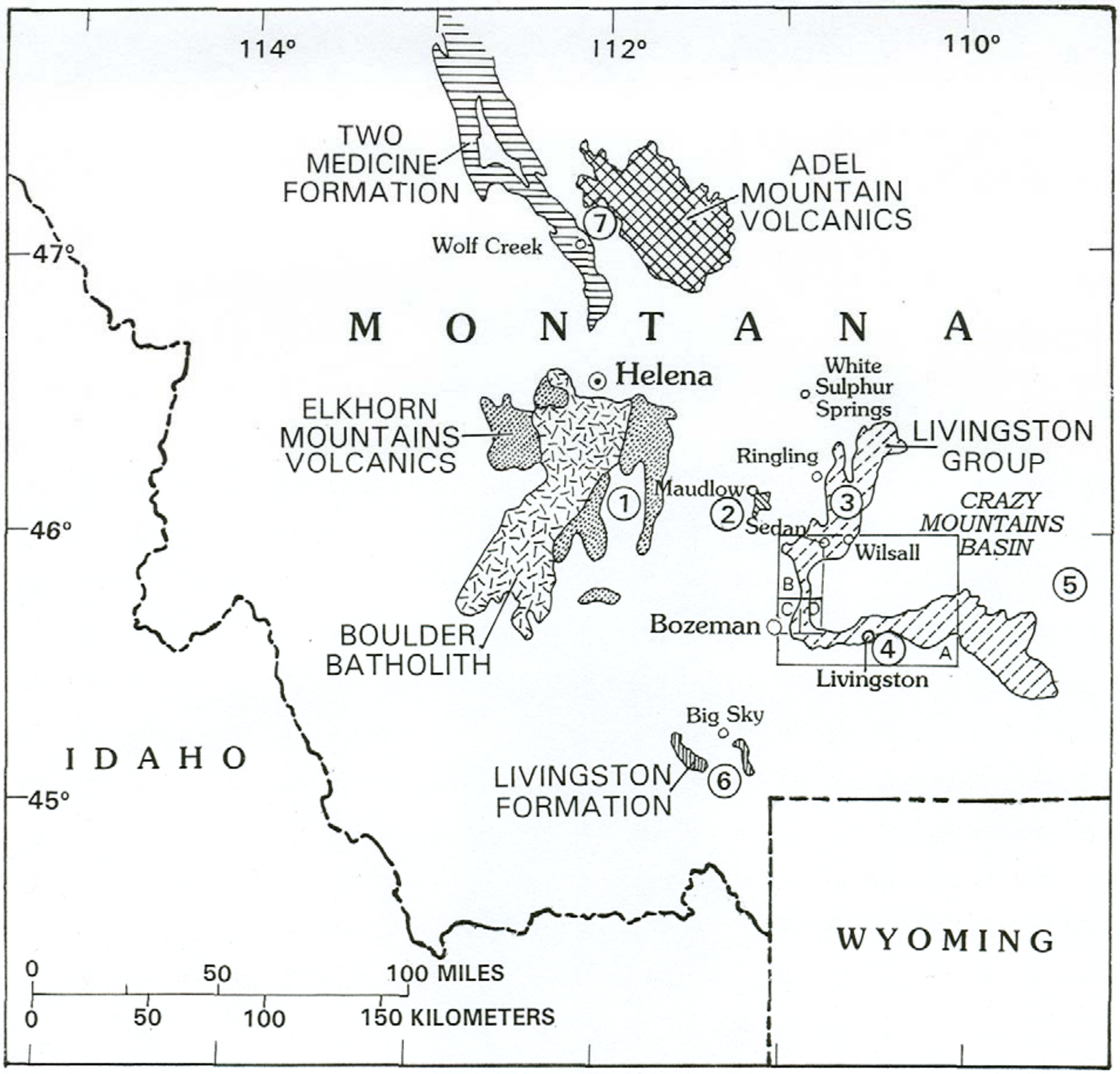


Figure 1. Location map of western Montana showing towns and geographic localities cited in the text and outcrop areas of Upper Cretaceous volcanic and volcaniclastic rocks discussed in the text.

①: Elkhorn Mountains Volcanics. ②: Maudlow Formation of Livingston Group. ③: Sedan Formation of Livingston Group in Sedan-Ringling area. ④: Type area of Cokedale, Miner Creek, and Billman Creek Formations in type area of the Livingston Group. ⑤: Upper Cretaceous marine sequences in central Montana that correlate with the Livingston Group. ⑥: Livingston Formation in the Madison Range. ⑦: The Two Medicine Formation and unconformably overlying Adel Mountain Volcanics in the northern Big Belt Mountains and the Adel Mountains.

A designates the Livingston 30' x 60' quadrangle (Berg and others, 2000). B is the Sedan 15' quadrangle (Skipp, Lageson, and McMannis, 1999). C is the Bozeman Pass 7½' quadrangle (Roberts, 1964b). D is the Fort Ellis (Kelly Creek) 7½' quadrangle (Roberts, 1964a).

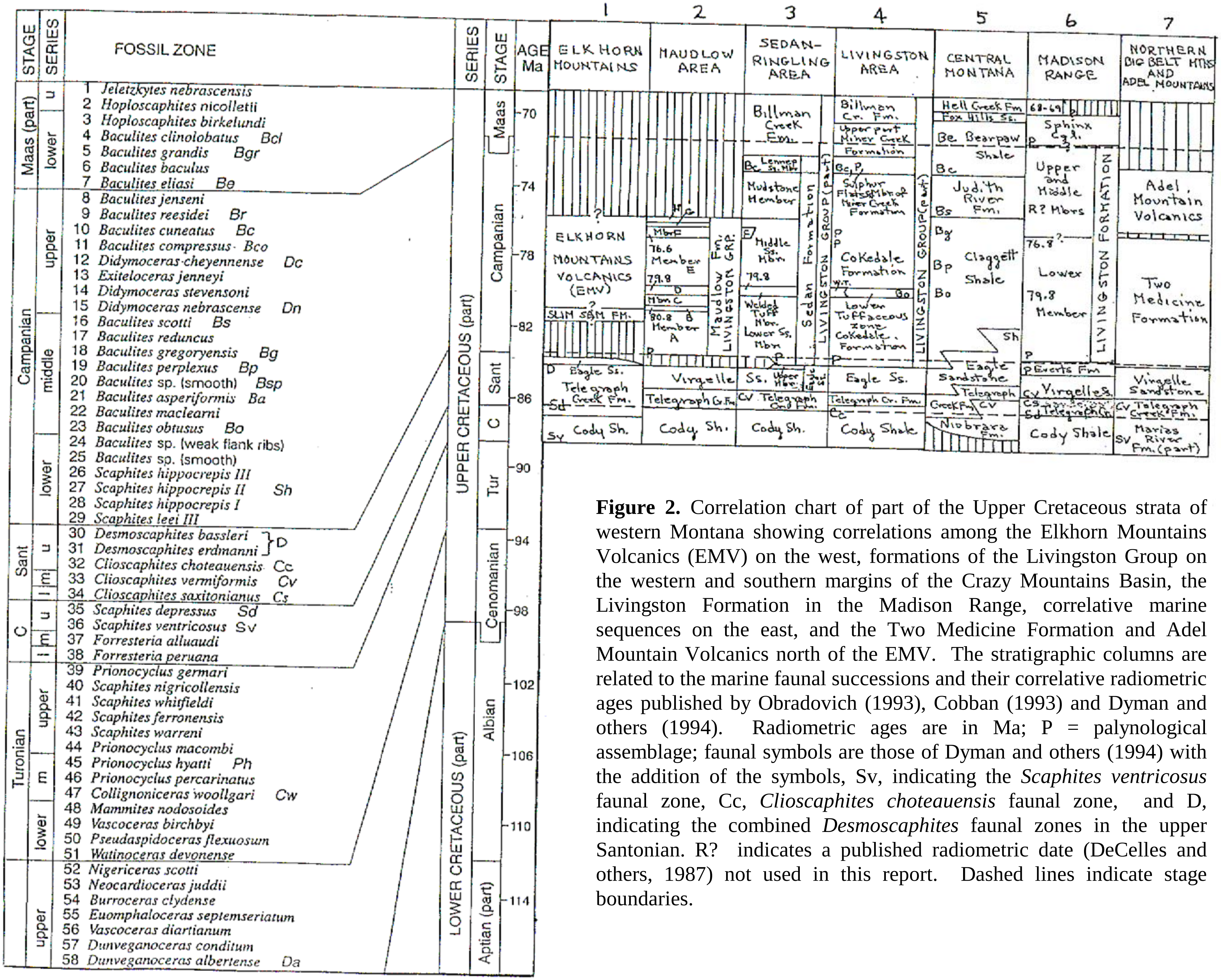


Figure 2. Correlation chart of part of the Upper Cretaceous strata of western Montana showing correlations among the Elkhorn Mountains Volcanics (EMV) on the west, formations of the Livingston Group on the western and southern margins of the Crazy Mountains Basin, the Livingston Formation in the Madison Range, correlative marine sequences on the east, and the Two Medicine Formation and Adel Mountain Volcanics north of the EMV. The stratigraphic columns are related to the marine faunal successions and their correlative radiometric ages published by Obradovich (1993), Cobban (1993) and Dyman and others (1994). Radiometric ages are in Ma; P = palynological assemblage; faunal symbols are those of Dyman and others (1994) with the addition of the symbols, Sv, indicating the *Scaphites ventricosus* faunal zone, Cc, *Clioscaphites choteauensis* faunal zone, and D, indicating the combined *Desmoscaphites* faunal zones in the upper Santonian. R? indicates a published radiometric date (DeCelles and others, 1987) not used in this report. Dashed lines indicate stage boundaries.

Creek Formation, the Billman Creek Formation, and the overlying Hoppers Formation, have been recognized to date.

The stratigraphic columns of figure 2 (columns 1-7) are each discussed below with reference to the available radiometric ages and paleontology.

Discussion

<u>Elkhorn Mountains Volcanics</u>

The Elkhorn Mountains Volcanics (EMV) - figure 2, column 1 – are remnants of an extensive volcanic field that may have covered as much as 26,000 square kilometers, and were likely to have been at least 3.5 km thick, perhaps, as much as 4.6 km thick (Smedes, 1966). Smith (1960) considered tuffs of the EMV to have been one of the earth's largest ash-flow fields. The volcanic field has been dated as being active from about 81 to 76 Ma, and, the Boulder batholith dated at about 80 to 60 Ma intrudes the EMV (Rutland and others 1989).

Klepper and others (1957) divided the EMV into 3 informal volcanogenic members. The lower member consists predominantly of lava flows, autobrecciated lavas, mudflow breccia, water-laid volcaniclastic rocks, and sparse thin sheets of andesitic ash-flow tuff. These ash-flow tuffs may have been mobile enough to spread out as far as the Madison Range (fig. 2, locality 6), and the western margin of the Crazy Mountains Basin (fig. 2, localities 2 and 3), where they have been radiometrically dated at 79.8 Ma (Tilling, 1974). The middle member of the EMV consists of many sheets of rhyolitic welded tuff and interbedded andesitic and basaltic volcaniclastic rocks (Rutland and others, 1989). These rhyolitic welded tuffs and interbedded volcaniclastic rocks have not been identified with certainty in any of the Livingston Group rocks, but probably are contemporaneous with many of the water laid and airfall tuffs found in the volcaniclastic rocks stratigraphically above the welded tuffs in the Sedan-Ringling and Livingston areas. The upper member of the EMV consists entirely of volcaniclastic deposits derived from units below except in the northernmost exposures, which comprise a large field of basalt lava flows (Rutland and others, 1989; Smedes, 1966).

The Slim Sam Formation conformably underlies the EMV (Klepper and others, 1957). Tysdal (2000) has restricted the name Slim Sam Formation to the volcaniclastic rocks that make up the upper part of the Slim Sam Formation as originally defined by Klepper and others (1957).

The lower part of the Slim Sam Formation of Klepper and others (1957) is assigned to the Eagle Sandstone, Telegraph Creek Formation, and the Cody Shale, in descending order by Tysdal (2000). An unconformity marks the top of the Eagle Sandstone. An ammonite identified by W.A. Cobban as representative of the upper Santonian *Desmoscaphites* faunal zones was recovered from near the top of the Telegraph Creek Formation below the Eagle Sandstone (Tysdal, 2000), and *Scaphites depressus* was recovered from near the base of the Telegraph Creek Formation. Radiometric dates from the base of the EMV and the late Santonian age of the Eagle Sandstone indicate that the time span between the base of the Eagle Sandstone and the EMV could be a maximum of 4 million years (Tysdal, 2000). This gap may be the same erosional gap as that between the Virgelle Sandstone and the Maudlow Formation of the Livingston Group in the Maudlow area (fig. 2, locality 2).

<u>The Maudlow Area</u>

The lower part of the Livingston Group, represented by the Maudlow Formation, in the Maudlow area -fig. 2, column 2 - as interpreted by Skipp and McGrew (1972, 1977), is revised only in that recalculated radiometric ages (Marvin and others, 1989) make regional correlation clearer. The base of the Maudlow Formation - Member A - contains a palynological assemblage that was identified by R.A Tschudy

(Skipp and McGrew, 1977) as being latest Santonian to earliest Campanian in age, and possibly older than assemblages from the base of the type Cokedale Formation. Reexamination of those assemblages from the Cokedale Formation, however, indicates that they may be the same age, or, possibly, older than Member A (Tysdal and Nichols, 1991). The restricted Slim Sam Formation (Tysdal, 2000) occupies a stratigraphic position similar to that of Member A.

Member D has a recalculated age of 79.8 Ma (Tilling, 1974), and is similar in age and composition to the "grey welded tuff" of the lower member of the Livingston Formation of the Madison Range (Tysdal and others, 1987). Member D is also in the same Stratigraphic position as the fossiliferous lower tuffaceous zone in the lower part of the type Cokedale Formation (Roberts, 1972), and is the same age, about 80 Ma, as the marine *Baculites obtusus* faunal zone (Obradovich, 1993). Hornblende from a dacite flow in Member B gives a recalculated K/Ar age of 80.8+1 Ma (Marvin and others, 1989), close to the oldest age of the EMV. Hornblende from a dacite flow in Member F yielded a recalculated K/Ar age of 76.6+1 Ma (Marvin and others, 1989). These dacite flows, the youngest primary volcanic rocks in the Maudlow Formation, coincide with the time of cessation of active dacitic volcanism preserved in the EMV. The Maudlow Formation, therefore, is in part equivalent in age to the EMV and to the lower volcaniclastic parts of both the Sedan and Cokedale Formations in their type areas.

<u>The Sedan-Ringling Area</u>

The Upper Cretaceous stratigraphy of the Sedan-Ringling area – fig. 2, column 3 – is compiled from the early publications of Skipp and McGrew (1972, 1977), and Skipp and others (1999) modified by a reevaluation of palynological assemblages from the type Sulphur Flats Member of the Miner Creek Formation in the Livingston area (Tysdal and Nichols, 1991).

The new map of the 30 x 60-minute Livingston quadrangle (Berg and others, 2000) shows a correlation between mapped members of the Sedan Formation (Skipp and others, 1999) and the mapped formations of the Livingston Group in the Bozeman Pass and Fort Ellis quadrangles (Roberts, 1964a,b) that indicates a boundary discordance. The discordance between the two maps indicated to Skipp and McGrew (1977), that the Mudstone Member just below the Lennep Sandstone Member of the Sedan Formation correlated with the lower part of the Miner Creek Formation. Roberts (1972), however, proposed a correlation of the Sulphur Flats Sandstone Member of the Miner Creek Formation (the lower part) with the Lennep Sandstone based on tracing a prominent ridge of sandstone northward from the Livingston area to Wilsall (fig. 1), where a marine fauna of the *Baculites cuneatus* and *Baculites compressus* faunal zones had been identified. In 1991, Tysdal and Nichols corroborated Roberts' correlation by identifying palynomorphs from the type Sulphur Flats sandstone Member that are correlative with the Lennep marine faunas at Wilsall. In the Sedan-Ringling area, the Lennep Sandstone Member of the Sedan Formation had been mapped from Wilsall southward to the join between the Sedan quadrangle (Skipp and others, 1999) on the north, and the Bozeman Pass quadrangle (Roberts, 1964b) on the south (fig.1). Therefore, the Lennep Sandstone as mapped in the Sedan quadrangle must be equivalent to the type Sulphur Flats Sandstone Member of the Miner Creek Formation, and the Billman Creek Formation as mapped in the Sedan Quadrangle must contain both the upper part of the type Miner Creek Formation and the entire Billman Creek Formation. A west-directed thrust mapped along the eastern margin of the Sedan quadrangle (Skipp and others, 1999) southward into the Fort Ellis (Kelly Creek) quadrangle (Roberts, 1964a) is located within this combined unit. The area of these quadrangles' join (fig. 1,) forms a structural triangle zone much like that described further north (Skipp and others, 1999, Berg and others, 2000) and probably involves a complex of faults that haven't been recognized. Mapping in the Miner Creek-Billman Creek interval is difficult because the contact between the two is marked by a thick mudstone unit which is covered in many places. The area needs to be looked at again. All of the Sedan Formation excepting the Lennep Sandstone Member, therefore, correlates with the type Cokedale Formation. The Lennep Sandstone Member correlates with the Sulphur Flats Sandstone Member at the base of the type Miner Creek Formation (fig. 2).

The Welded Tuff Member of the Sedan Formation has been mapped into the northern part of the Bozeman Pass Quadrangle (Roberts, 1964b) near the top of the Cokedale Formation (Berg and others, 2000). However, the age of the welded tuffs, about 79.8 Ma, suggest correlation with a lower stratigraphic position, probably with the lower tuffaceous zone of the Cokedale Formation that contains fossils of the *Baculites obtusus* faunal zone (Roberts, 1972, p. C39). The zone of *Baculites obtusus* is radiometrically dated at 80.54+0.55 Ma (Obradovich, 1993), and seems to correlate with a period of extensive explosive volcanism at about 80 Ma during which widespread ash-flow tuffs spread across western Montana.

Palynomorphs from the basal beds of the Cokedale Formation at its type section have been identified as Upper Santonian (Tysdal and Nichols, 1991), and may correlate with the floras identified by R. H. Tschudy (Skipp and McGrew, 1977) from a lignitic coal at the base of the Sedan Formation (D. J. Nichols, oral communication, 2004).

The Livingston Area

The Livingston area -fig. 2, column 4 - includes the type sections for all of the formations and members of the Livingston Group in the southern part of the Crazy Mountains Basin (Roberts, 1972). These include, in ascending order, the Cokedale, Miner Creek, Billman Creek, and Hoppers Formations. The Miner Creek Formation consists of a basal Sulphur Flats Sandstone Member rich in tuff, quartzose sandstone, volcaniclastic sandstone, and conglomerate. The member is correlated northward along strike with the Lennep Sandstone at Wilsall. The Lennep Sandstone at Wilsall contains a marine fauna that includes the Baculites compressus and Baculites cuneatus faunal zones (Roberts, 1972). The Sulphur Flats Sandstone Member is massive and is overlain by the upper part of the Miner Creek Formation, composed of largely tuffaceous siltstone that is olive gray to grayish yellow green and contains minor thin-bedded sandstone. The overlying Billman Creek Formation consists of interbedded mudstone, claystone, siltstone, and sandstone, and is characteristically grayish red with some grayish green beds in the lower part. The overlying Hoppers Formation is not considered in this report.

Reexamination of palynological assemblages from the type section of the Cokedale Formation and the Sulphur Flats Sandstone Member of the Miner Creek Formation indicate a middle to late Campanian age for the middle part of the Cokedale and a late Campanian age for the type Sulphur Flats Member, and correlation of the Sulphur Flats Member with the marine Lennep Sandstone (Tysdal and Nichols, 1991). The marine *Baculites obtusus* fauna recovered from the lower tuffaceous zone in the Cokedale Formation by Roberts (1972) lies below the palynological collections referenced above (Tysdal and Nichols, 1991). As noted, palynological collections from near the base of the Cokedale have been assigned to the upper Santonian. These age assignments are used here for correlation with members of the Maudlow and Sedan Formations mapped to the west and northwest (figs.1 and 2).

Central Montana

Central Montana - fig. 2, column 5 - is adapted from Dyman and others (1995). The marine sequences are well developed, and the identification of numerous critical faunal zones (Cobban, 1993) provides correlations between the Upper Cretaceous marine sequences of central Montana and the nonmarine volcanic and volcaniclastic sequences to the west. *Baculites obtusus*, *Baculites compressus*, and *Baculites cuneatus* faunal zones are widespread and suggest regional correlations as far west as the Elkhorn Mountains (fig. 2, locality 1). The usefulness of these zones for correlations between the type Livingston Group volcaniclastic rocks and the marine sequences to the east were noted earlier by Roberts (1972).

<u>Northern Madison Range</u>

The Livingston Formation in the Madison Range- fig. 2 - column 6 – is adapted from Tysdal and others (1987). The Livingston Formation is underlain by the Everts Formation, the Virgelle Sandstone, the telegraph Creek Formation and the Cody Shale, in descending stratigraphic order. Marine faunal zones, *Clioscaphites saxitoniatus* and *Scaphites depressus* form overlapping faunal zones at a locality in the Telegraph Creek Formation and indicate these strata lie astride the Coniacian-Santonian boundary (Dyman and others, 1997). A diagnostic palynomorph collection from the upper 300 ft (90 m) of the Everts Formation indicates deposition during the latest Santonian to earliest Campanian time (Tysdal and Nichols, 1991), and another collection from the lower member of the Livingston Formation indicates an earliest Campanian age (fig. 2).

Stratigraphically above these palynomorph collections, radiometric ages of 79.8+2.9 Ma and 76.8+2.5 Ma were obtained from welded tuffs in the lower member of the Livingston Formation (Tysdal and others, 1987). Their "grey welded tuff" bears petrographic and compositional resemblance to the welded tuffs of Member D of the Maudlow Formation in the Maudlow area, and the Welded Tuff Member in the Sedan-Ringling area, and appears to be close in age. Member F of the Maudlow Formation is made up of grayish-red dacite flow and flow breccias lithologically similar to those of the middle member of the Livingston Formation in the Madison Range. A radiometric age of 76.6 +1 Ma was obtained on hornblende from Member F (Marvin and others, 1989). Thus, though much of the Livingston sequence in the Madison Range is quite different overall from the Maudlow sequence, there are some similarities. Possibly, both the EMV and local sources provided the volcanogenic material for the Livingston Formation in the Madison Range.

The Sphinx Conglomerate above the informal middle and upper members of the Livingston Formation is shown by Tysdal and Nichols (1991) to be restricted to the Maastrichtian, although queries at the lower and upper contacts indicate uncertainty due to a lack of reliable fossil and/or radiometric dates. A radiometric date of 75.6+1.8 Ma was published by DeCelles and others (1987) from the lower part of the Sphinx Conglomerate. But, they also noted Maastrichtian palynomorphs from the middle part of the formation, a determination that conflicts with the radiometric date. The position of this radiometric date is indicated by R? on the correlation chart of figure 2, column 6.

The Livingston Formation and the Sphinx Conglomerate are both cut by dacitic dikes that were emplaced about 68-69 Ma. These dates suggest that the youngest Livingston strata of the Madison Range are no younger than mid-Maastrichtian (Tysdal and Nichols, 1991).

<u>Northern Big Belt Mountains and Adel Mountains</u>

The Upper Cretaceous stratigraphic relationships in the northern Big Belt Mountains and Adel Mountains - figure 2, column 7 - are based on recent work by Harlan and others (in press). Radiometric ages reported in that paper show that the Adel Mountain Volcanics were emplaced between 75 and 73.7 Ma. The unconformably underlying Two Medicine Volcanics were emplaced before the Adel Mountains Volcanics and after deposition of the Virgelle Sandstone. A fauna representing the middle Santonian *Clioscaphites vermiformis* faunal zone (Cobban, 1993) was identified from the Telegraph Creek Formation below the Virgelle Sandstone, and an upper Coniacian *Scaphites ventricosus* fauna was found in the upper part of the underlying Marias River Formation below the Telegraph Creek Formation.

Skipp and McGrew (1972) suggested that the volcanic centers in the Adel Mountains may have provided volcanic detritus for the tuffs present in the Billman Creek Formation of the Livingston Group. These new ages, however, show that deposition of the Two Medicine Formation and the Adel Mountain Volcanics coincided closely with EMV volcanism, and a younger age is ruled out.

Conclusions

The Upper Cretaceous volcanic and volcaniclastic sequences – Livingston Group or Formation - found in several isolated areas of central and western Montana can be correlated with one another on the basis of radiometric dating combined with paleontologic studies. Radiometric ages come from several sources, and the paleontologic correlations are based chiefly on the work of Obradovich (1993) and Cobban (1993). Two stratigraphically linked faunal and radiometric zones are particularly useful; the *Baculites obtusus* faunal zone at 80.54 + 0.55 Ma (Obradovich, 1993), Middle Campanian, and the *Baculites compressus* faunal zone at 73.35 + 0.39 Ma, Upper Campanian (Obradovich, 1993). In addition, recent palynological studies of floras from the Livingston Group at the type section (Tysdal and Nichols, 1991) have provided key correlations with marine faunas to the north and east.

The Middle Campanian *Baculites obtusus* faunal zone provides a timeline that relates deposition of the lower part of the EMV to the Maudlow, Sedan, and Cokedale Formations of the Livingston Group in western Montana and correlates these formations eastward to the marine Claggett Shale. Similarly, the Upper Campanian *Baculites compressus* faunal zone provides a key, along with palynological determinations (Tysdal and Nichols, 1991), to the correlation of the Lennep Sandstone Member of the Sedan Formation with the Sulphur Flats Sandstone Member of the type Miner Creek Formation. This correlation indicates that the Billman Creek Formation mapped in the Sedan-Ringling area (Skipp and others, 1999) includes both the upper part of the Miner Creek Formation above the Sulphur Flats Sandstone Member at its type section, and the Billman Creek Formation at its type section. This correlation suggests that new mapping is needed in the area of the join between the Sedan and Fort Ellis (Kelly Creek) quadrangles (fig. 1) northwest of Livingston.

In addition, a radiometric age and description of a "grey welded tuff" in the Livingston Formation of the Madison Range (Tysdal and others, 1987) suggests correlation of that tuff with Member D of the Maudlow Formation and the Welded Tuff Member of the Sedan Formation which can be traced into the Cokedale Formation; this suggests that at least part of the Madison Range sequence is related to the EMV. This zone then correlates with the marine *Baculites obtusus* faunal zone to the east.

New radiometric ages from the Adel Mountain Volcanics (Harlan and others, in press) indicate that these volcanics largely correlate with the deposition of the EMV and, therefore, cannot be the source of volcanic detritus found in the Billman Creek Formation of the Livingston Group as previously suggested by Skipp and McGrew (1972).

References

Berg, R.B., Lopez, D.A., and Lonn, J.D., 2000, Geologic Map of the Livingston 30' x 60' quadrangle, southwest Montana: Montana Bureau of Mines and Geology Open-File Report 406, scale 1:100,000.

Cobban, W.A., 1993, Diversity and Distribution of Late Cretaceous ammonites, Western Interior, United States, in Caldwell, W.G.E., and Kauffman, E.G., eds., Evolution of the Western Interior Basin: Geological Association of Canada Special Paper 39, p.435-452.

DeCelles, P.G., Tolson, R.B., Graham, S.A., Smith, G.A., Ingersoll, R.V., White, J., Schmidt, C.J., Rice, R., Moxon, I., Lemke L., Handschy, J.W., Follo, M.F., Edwards, D.P., Cavazza, W., Caldwell, M., and Bangor, E., 1987, Laramide thrust-generated alluvial-fan sedimentation, Sphinx Conglomerate, southwestern Montana: American Association of Petroleum Geologists Bulletin, V. 71, no. 2, p. 135-155.

Dyman, T.S., Cobban, W.A., Fox, J.E., Hammond, R.H., Nichols, D.J., Perry, W.J., Jr., Porter, K. W., Rice, D.D., Setterholm, D.R., Shurr, G.W., Tysdal, R.G., Haley, J.C., Campen, E.B., 1994, Cretaceous rocks from southwestern Montana to southwestern Minnesota, Northern Rocky Mountains, and Great Plains region: in G.W. Shurr, G.A. Ludvigson, and R.H. Hammond, eds., Perspectives on the eastern margin of the Cretaceous Western Interior Basin: Geological Society of America Special Paper 287, p. 5-26.

Dyman, T.S., Porter, K.W., Tysdal, R.G., Cobban, W.A., Fox, J.E., Hammond, R.H., Nichols, D. J., Perry, W.J., Jr., Rice, D.D., Setterholm, D. R., Shurr, G.W., Haley, J.C., Lane, D.E., Anderson, S.B., and Campen, E.B., 1995, West-East Stratigraphic Transact of Cretaceous rocks in the northern Rocky Mountains and Great Plains regions, southwestern Montana to Minnesota: U. S. Geological Survey Miscellaneous Investigations Series Map I-2474-A.

Dyman, T.S., Tysdal, R.G., Perry, W.J., Jr., Obradovich, J.D., Haley, J.C., and Nichols, D.J., 1997, Correlation of Upper Cretaceous strata from Lima Peaks area to Madison Range, southwestern Montana and southeastern Idaho, U.S.A.: Academic Press Limited; Cretaceous Research, p.751-766.

Harlan, S.M., Snee, L.W., Reynolds, MW. Mehnert, H.H., Schmidt, R.G., Sheriff, S.D., and Irving, A.J., in press, 40Ar/39Ar and K-Ar geochronology and tectonic significance of the Late Cretaceous Adel Mountain Volcanics and spatially associated igneous rocks, northwestern Montana: U.S. Geological Survey Professional Paper 1696, 71p.

Klepper, M.R., Weeks, R.A., and Ruppel, E.T., 1957, Geology of the southern Elkhorn Mountains, Montana: U.S. Geological Survey Professional Paper 292, 82 p., map scale 1:31,250.

Marvin, R.F., Mehnert, H.H., Naeser, C.W., and Zartman, R.E., 1989, U.S. Geological Survey radiometric ages-Compilation "C", Part five-Colorado, Montana, Utah, and Wyoming: Isochron West, no. 53, P.14-19.

Obradovich, John, 1993, A Cretaceous time scale, in Evolution of the Western Interior Basin, Caldwell, W.G.E., and Kauffman, E.G., eds.: Geological Association of Canada Special Paper, p. 379-396.

Roberts, A.E., 1964a, Geology of the Fort Ellis quadrangle, Montana: U.S. Geological Survey Miscellaneous Investigations Map I-397, scale 1:24,000.

Roberts, A.E., 1964b, Geology of the Bozeman Pass quadrangle, Montana: U.S. Geological Survey Miscellaneous Investigations Map I-399, scale 1:24,000.

Roberts, A.E., 1972, Cretaceous and Early Tertiary depositional and tectonic history of the Livingston area, southwestern Montana: U.S. Geological Survey Professional Paper 526-C, p. C1-C120.

Rutland, Carolyn, Smedes, H.W., Tilling, R.I., and Greenwood, W.R., 1989, Volcanism and plutonism at shallow crustal levels; the Elkhorn Mountains Volcanics and Boulder batholith, southwestern Montana, in Hyndman, D.W., ed., Cordilleran volcanism, plutonism, and magma generation at various crustal levels, Montana and Idaho: 28th International Geological Congress Field Trip Guidebook T337, p. 16-31.

Skipp, Betty, 1977, Geologic map and cross section of the Wallrock quadrangle, Gallatin and Park Counties, Montana: U.S. Geological Survey Geologic Quadrangle Map GQ-1402, scale 1:24,000.

Skipp, Betty, Lageson, D.R., and McMannis, W.J., 1999, Geologic map of the Sedan quadrangle, Gallatin and Park Counties, Montana: U.S. Geological Survey Geological Investigations Series Map I-2634, scale 1:48,000.

Skipp, Betty, and McGrew, L.W., 1972, The Upper Cretaceous Livingston Group of the western Crazy Mountains Basin, Montana: Montana Geological Society Guidebook 21st Annual Field Conference, Crazy Mountains Basin, p.99-106.

Skipp, Betty, and McGrew, L.W., 1977, The Maudlow and Sedan Formations of the Upper Cretaceous Group on the west edge of the Crazy Mountains Basin, Montana: U.S. Geological Survey Bulletin 1422-B, p. B1-B68.

Skipp, Betty, and Peterson, A.D., 1965, Geological map of the Maudlow quadrangle, southwestern Montana: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-452, scale 1:24,000.

Smedes, H.W., 1966, Geology and igneous petrology of the northern Elkhorn Mountains, Jefferson and Broadwater counties, Montana: U.S. Geological Survey Professional Paper 510, 116 p.

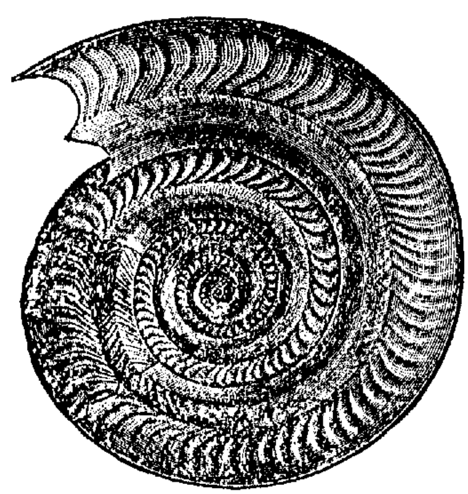
Smith, R.L, 1960, Ashflows: Geological Society of America Bulletin, v.71, p.795-842.

Tilling, R.I., 1974, Composition and Time Relations of Plutonic and Associated Volcanic Rocks, Boulder Batholith Region, Montana: Geological Society of America Bulletin, v.85, 1925-1930.

Tysdal, R.G. 2000, Revision of Cretaceous Slim Sam Formation, Elkhorn Mountains Area, Montana: U.S. Geological Survey Professional Paper 1601-B, p. B1-B8.

Tysdal, R.G., and Nichols, D.J., 1991, Biostratigraphic correlation of Santonian and Campanian formations in the northwestern Part of Yellowstone National Park, and the Madison Range and Livingston area of southwestern Montana: U.S. Geological Survey Bulletin 1962, p. 10-19.

Tysdal, R.G., Nichols, D.J., and Winkler, G.R., 1987, The Livingston Formation in the Madison Range of southwestern Montana: U.S. Geological Survey Bulletin 1665, 15p.



Geology of the Little Bear Creek Copper Skarn, Troy, Idaho

Michael C. Rowe

Geology Department, Washington State University, Pullman, WA 99163
Present address: Oregon State University, Department of Geosciences, Corvallis, OR, 97331
rowem@geo.oregonstate.edu

Lawrence D. Meinert

Geology Department, Washington State University, Pullman, WA 99163
Present address: Smith College, Department of Geology, Northampton, MA 01063
lmeinert@smith.edu

ABSTRACT

The Little Bear Creek copper skarn, located near Troy, Idaho, underlies Miocene basalt flows, which have subsequently been eroded by Little Bear Creek. Three igneous intrusions, a monzonite, pegmatite and quartz monzonite porphyry, in addition to a granitic gneiss, are all found in outcrop and drill core, associated with skarn mineralization. However, geochemical analysis indicates that the pegmatite and monzonite were probably not associated with skarn formation. Relatively homogenous garnet compositions ($Gr_{71\text{-}89}Al_{12\text{-}28}Py_{0.1\text{-}0.3}$) support a single source hypothesis. Pyroxene compositions are diopsidic ($Di_{52\text{-}75}Hd_{21\text{-}45}Jo_{3\text{-}4}$) and match those commonly associated with copper skarns around the world. We hypothesize that fluids derived from the NE-striking dike of quartz monzonite porphyry are responsible for the formation of the skarn deposit, based on correlations from geochemical analysis, drill core, and field observations. A chilled margin at the contact between the quartz monzonite porphyry and the skarn, and jointing exposed in the porphyry suggest a relatively shallow depth of emplacement, consistent with previous models of copper skarn genesis. The Little Bear Creek copper skarn is significant because it implies that more skarn mineralization may be associated with shallow, intrusive dikes/plutons and Belt Group carbonates in northern Idaho and eastern Washington.

INTRODUCTION

Skarn deposits, such as that found near Troy, ID, can be an important economic source for many rare and valuable minerals. Copper skarn deposits, such as the Little Bear Creek copper skarn, often result from the shallow intrusion of magma either in the form of dikes/sills or plutons into calcareous country rock. However in eastern Idaho and western Washington relatively little is known about the country rock due to thick sequences of Miocene Columbia River Basalt (CRB) flows overlying most of the preexisting geology (Potter et al., 1999; Priebe and Bush, 1999). Pre-CRB geology is often only found as isolated buttes or in river/stream drainages where sufficient downcutting has occurred to expose basement rock (see for example: Webster and Nunez, 1980; Hooper and Rosenberg, 1970). The Little Bear Creek copper skarn is therefore geologically important for understanding pre-Columbia River Basalt igneous activity and its relationship to the formation of ore bodies in Eastern Washington and Idaho. The purpose of this study has therefore been to document the extent of skarn mineralization and determine its chemical and spatial relationship to igneous intrusions in the area.

Geologic Setting

The Little Bear Creek copper skarn lies in a complex geologic setting. The skarn deposit is located to the southwest of Moscow Mountain, a large granitic body related to the Idaho Batholith, dated at 67.8 +/- 2.5 Ma (Webster and Nunez, 1982). Also prominent in Latah County are Precambrian Belt Rocks that consist of beds of quartzite, siltite, argillite, marble, schist and gneiss (Potter et al., 1999). Located north of Little Bear Creek, in the Harvard 7.5 minute Quadrangle of Latah County, are Tertiary Potato Hill volcanics, which include stocks of quartz monzonite porphyry and dikes of basaltic andesite and dacite (Potter et al., 1999). Overlying all of the local geology near Little Bear Creek are the Grande Ronde and Priest Rapids basalt flows of the Columbia River Basalt Group (Potter et al., 1999; Reidel et al., 1989). The Columbia River Basalt Group erupted throughout much of the Miocene from 17 million years to 6 million years, although the bulk of the basalts erupted during the first 1.5 million years (Grande Ronde basalt flows), and can reach a total thickness of greater than 3 km (Reidel et al., 1989; Tolan et al., 1989).

The Little Bear Creek skarn is located 5 km east of Troy, Idaho, in Latah County (Fig. 1). In the Pacific Northwest, the Little Bear Creek skarn is the westernmost copper skarn related to cratonic rock, (e.g. east of the 0.704/0.706 $^{87}Sr/^{86}Sr$ line). Based on the size and scarcity of mine dumps, production appears to have been limited. The Little Bear Creek area was reported to have first been mined in the 1940s and 1950s although records of production and grade are not readily available. However, newspaper insulation in the walls of one building (Spokane Spokesman Review, 1913) possibly dates an early period of exploration. Intermittent exploration and development resulted in at least 7 shallow adits and mine shafts. Rock exposure is only visible at ground level in four of these shafts and adits. In 1991 seven diamond drill holes (DDH's) were drilled by Kennecott Northwest Exploration and analyzed for Cu, Au, and Ag (Fig. 1; Wakeman, 1991). No further economic exploration has occurred since the Kennecott drilling.

METHODS

The geologic outcrop map (Fig. 1) is based on a limited number of outcrops in the heavily vegetated terrain and float mapping was often required to discern approximate contacts. The Kennecott drill core, stored at a nearby farm, was logged for contacts, alteration mineralogy, and igneous and metamorphic fabric. The Washington State University geoanalytical laboratory performed the whole rock major and trace element analyses, using x-ray fluorescence (XRF). For a detailed discussion of the XRF analytical procedures see Johnson et al (1999). Electron microprobe analysis of skarn pyroxenes and garnets was also conducted at Washington State University.

RESULTS

Four distinct rock types, excluding the skarn material and overlying basalt flows, crop out in the approximately 0.4 km^2 study area. The four rock types were assigned the following field terms: granitic gneiss, monzonite, pegmatite, and quartz monzonite porphyry. Classification of these rocks was based on mineral composition as observed in outcrop and hand specimen. Crosscutting relations both in drill core and in the field suggest that the granitic gneiss is the oldest rock unit in the area, followed by the monzonite, pegmatite, and quartz monzonite porphyry. These relative ages are corroborated by similarities of the intrusions to geologic units found elsewhere in the region. The monzonite and pegmatite are texturally and chemically similar to other Cretaceous intrusions in the region whereas the quartz monzonite porphyry is similar to dikes and stocks of porphyritic material associated with the Tertiary age Potato Hill volcanics (Potter et al., 1999).

Petrography

Granitic Gneiss

The granitic gneiss, the oldest rock unit in the region, varies compositionally, texturally and structurally throughout the field area and in relation to the monzonite (Fig. 1). Thin section estimates of the granitic gneiss indicate approximately 25% alkali feldspar, 20% pyroxene, 20% quartz, 15% plagioclase, 10% amphibole, 5% biotite, and 5% chlorite as an alteration phase, with a strongly foliated gneissic to schistose texture (Fig. 2a). Trace amounts of sulfides are also present. Sample LBC-1-30 of the granitic gneiss is relatively rich in pyroxene (>20%), possibly due to its close proximity to the skarn. Several 1-10 mm bands of weathered mafic minerals, mostly biotite, separate wide (<0.3 m) sections of more felsic material in the northernmost outcrops and are representative of the banded texture of the granitic gneiss.

Of the eight outcrops of granitic gneiss, five have a steeply dipping foliated texture that strikes in a northeasterly direction. The other three outcrops have foliations that dip steeply to the northeast and strike southeasterly. In vertical drill core, the banded texture of pyroxene, biotite and amphibole dip approximately 30-50°.

Monzonite

In the field area, the monzonite occurs in outcrop just south of the granitic gneiss in the northernmost region of the field site (Fig. 1). The monzonite is characterized in hand specimen by medium grained plagioclase, alkali feldspar, hornblende, and biotite (Fig. 2b). XRF analysis indicates a low SiO_2 content (57.89%), much less then either the porphyry or the pegmatite (Table 1). Several trace minerals were also identified in thin section including sphene, epidote, calcite and several sulfides. The monzonite has a slight foliation, defined by the biotite and amphibole, which is most easily observed in the drill core.

Pegmatite

The pegmatite is identified in the field by its extremely coarse-grained feldspars and micas, and lack of mafic minerals (Fig. 2c). The pegmatite minerals are generally >0.5 cm in diameter. Modal estimates from thin section indicate the pegmatite is composed of approximately 45% alkali feldspar, 25% quartz, 15% plagioclase, and 15% muscovite/biotite. Feldspar alteration to clay is most evident in thin section. The pegmatite is found in some large outcrops but for the most part occurs as thin, irregular sills and dikes intruding into other rock formations. The pegmatite is observed in the field associated with both the granitic gneiss and the monzonite (Fig. 1).

Quartz Monzonite Porphyry

The quartz monzonite porphyry is exposed in two locations along Little Bear Creek (Fig. 1). The quartz monzonite porphyry is characterized by a fine-grained, gray, groundmass (approximately 50%) and a phenocryst assemblage of medium-grained anhedral quartz and feldspar, and fine-grained biotite (Fig. 2d).

An elongated linear contact between the quartz monzonite porphyry and the skarn material on the east bank of the Little Bear Creek suggests that the quartz monzonite porphyry was intruded as a dike striking NNE. Jointing of the porphyry is also observed in the Little Bear Creek streambed. In drill hole # 1, at a depth of 8 m, the porphyry grain size diminishes towards the skarn and has a chilled margin near a contact with skarn (Fig. 3). Approximately 0.5 m from the contact the matrix begins to alter to a green color due to endoskarn alteration of the porphyry.

Skarn alteration and mineralogy

The majority of the Little Bear Creek Cu skarn occurs in a region 150 m x 100 m located northeast of the Camps Canyon road stream-crossing (Fig. 1). The skarn is identifiable in outcrop based on its dark brown color and green staining from the oxidation of Cu-bearing minerals. The Little Bear Creek skarn is dominated by fine to medium-grained pyroxene and garnet, with lesser amounts of amphibole, epidote, and calcite and trace Cu sulfides, including pyrite, chalcopyrite and bornite (Fig.

2a). Garnet and pyroxene proportions are zoned relative to the igneous intrusions with a proximal garnet-rich zone and a distal pyroxene-rich zone, a typical feature in skarn deposits (Meinert, 1997). Garnet compositions are strongly calcic and relatively homogenous, ranging from ($Gr_{71-89}Al_{12-28}Py_{0.1-0.3}$). Some zonation is evident in several garnets with a slight decrease in CaO and increase in FeO from core to rim (Table 2). Pyroxene compositions in the skarn material are strongly diopsidic ($Di_{52-75}Hd_{21-45}Jo_{3-4}$), and fall within the compositional field characteristic of copper skarn deposits (Table 2; Fig. 4; Meinert 1992). The Little Bear Creek copper skarn is smaller but otherwise chemically and similar to copper skarns in Idaho (Mackay district; Chang and Meinert, 2002, 2004) and Montana (Garnet district; Umpleby, 1914, Wilson et al., 1995). A dark black amphibole is commonly observed as a reaction rim approximately 0.5 cm thick along the contacts between the monzonite and pegmatite and the skarn material. Hydrothermal veining, numerous quartz veinlets and pegmatite dikes are also evident in many of the skarn samples.

DISCUSSION

In the field and drill core, the garnet-pyroxene-rich skarn is spatially associated with the quartz monzonite porphyry, pegmatite, and monzonite suggesting that there is at least a spatial relationship to these intrusions and the skarn (Figs. 1, 3). In several of the thicker sequences of skarn material in drill core, garnet-pyroxene zonation is evident relative to the pegmatite intrusion. The pegmatite and monzonite appear to be part of the western margin of the Idaho Batholith as summarized by Hyndman (1983). In general, these rocks are not strongly mineralized but do have minor base and precious metal mineralization (Bennett, 1980). Although the drill core suggests that skarn material is at least spatially associated with all of the igneous intrusions, geochemical evidence suggests that the quartz monzonite porphyry is most similar to igneous intrusions typically associated with copper skarn mineralization. The Rb/Sr ratio for the porphyry (0.12) matches well with the worldwide average for intrusions associated with Cu skarns, whereas the pegmatite has a higher ratio (0.42) and the gneissic granite has a lower ratio (0.07; Fig. 5; Meinert, 1995). The relatively homogeneous garnet compositions help to support the hypothesis that the skarn was formed from a single intrusion (Table 2). If multiple fluid phases had caused the skarn formation then several distinct chemical compositions of the garnets, or distinctive zoning within the garnets, would be observed, associated with the varying compositions of the magmas.

The chilled margin with endoskarn alteration in DDH #1 also suggests that the quartz monzonite porphyry is genetically related to the skarn (Fig. 3). The two-foot zone of skarn mineralization at the quartz monzonite porphyry contact, as well as the close proximity in outcrop, suggests that there is a spatial, in addition to the geochemical, relationship between the skarn and the quartz monzonite porphyry (Fig. 3). The chilled margin and jointing of the quartz monzonite porphyry suggest a shallow depth of intrusion, characteristic of many copper skarns (Meinert, 1992; Dilles and Proffett, 1995). Therefore, although drill core observations are inconclusive on their own, in conjunction with the geochemical analyses and field observations, the evidence suggests that the skarn mineralization resulted from fluids derived from the quartz monzonite porphyry.

The orientation of the quartz monzonite porphyry dike has significant implications for the possible location of other bodies of ore beneath the Columbia River Basalts. The scenario of a NNE striking dike is consistent with the location of the outcropping skarn mineralization, and explains why skarn material occurs in DDH #5. Based on this model, outcrops of skarn and mineralization can be predicted along contacts of the porphyry. With the porphyry dike striking in a northeasterly direction it suggests the occurrence of skarn material just east of the postulated buried dike. However, there is no skarn or porphyritic intrusive material in DDH #6, although localized structural complexities, or variations in the carbonate component of the host rock, might explain the lack of skarn material.

CONCLUSIONS

In the Pacific Northwest, the Little Bear Creek skarn is the westernmost Cu skarn known related to cratonic rock, although it is chemically similar to copper skarns in Idaho (Mackay district; Chang and Meinert, 2002, 2004) and Montana (Garnet district; Umpleby, 1914, Wilson et al., 1995). Observed both in the field and in drill cores 1, 2, 3 and 7, the Little Bear Creek copper skarn is spatially associated with monzonite, pegmatite, and quartz monzonite porphyry intrusions. However, endoskarn alteration in the quartz monzonite porphyry suggests that the porphyry was the major source of fluids for the formation of the skarn. Geochemical data also indicates the quartz monzonite porphyry as a likely source for skarn formation because of its chemical similarities to other Cu skarn plutons. This correlation is especially evident when using the differentiation index of Rb/Sr (Fig. 5). In contrast, both the pegmatite and the monzonite have trace element signatures that are dissimilar to intrusions related to typical Cu skarns.

The Little Bear Creek skarn is a copper skarn deposit composed almost entirely of garnet and pyroxene. The pyroxene compositions are diopsidic with low manganese ($Di_{52-75}Hd_{21-45}Jo_{3-4}$), and thus are similar to those found at copper skarn deposits worldwide (Meinert et al, 1997). Garnet compositions are relatively homogenous ($Gr_{71-89}Al_{12-28}Py_{0.1-0.3}$), although there is a slight zonation of decreasing CaO and increasing FeO from core to rim in several garnets. The Little Bear Creek skarn is most significant because of its presence in an area not known for this type of ore deposit.

This research is instrumental in further understanding mineralization along the western margin of the Idaho Batholith and for its implication that more mineralized zones may be present, buried by younger basalt flows. Based on the abundance of Belt Group rocks present, especially to the north of the Little Bear Creek deposit, and scattered outcrops of quartz monzonite porphyry in Latah County, the existence, and further discovery, of more skarn material can be anticipated upon additional mapping of carbonate-rich lithologies and shallow intrusives.

ACKNOWLEDGEMENTS

This project was funded by the Washington State University Honors College and the College of Sciences. We would also like to thank Dr. Reed Lewis (Idaho Geologic Survey) and Dr. David Gaylord (WSU) for their helpful comments and discussion of the manuscript.

REFERENCES

Bennett, E.H., 1980, Granitic rocks of Tertiary age in the Idaho Batholith and their relation to mineralization: Economic Geology, v. 75, p. 278-288.

Chang, Z., and Meinert, L.D., 2004, Skarn and mineralization at the Empire mine, Idaho: Geological Society of America, Abstracts with Programs, v. 36, p.

Chang, Z., and Meinert, L.D., 2002, Magmatic processes and skarn formation - An example from the Empire Cu-Zn skarn deposit, Custer County, Idaho: Geological Society of America, Abstracts with Programs, v. 34, p. 337.

Dilles, J., and Proffett, J., 1995, Metallogenesis of the Yerington batholith, Nevada, *in* Pierce, F., et al., eds., Porphyry Copper Deposits of the American Cordillera: Arizona Geological Society Digest 20, p. 306-315.

Hooper, P.R., Rosenberg, P.E., 1970, Petrology of granite point, southeastern Washington: Northwest Science, v. 44, p.

131-142.

Hyndman, D.W., 1983, The Idaho Batholith and associated plutons, Idaho and western Montana, *in* Roddick, J.A., ed., Circum-Pacific plutonic terranes: Geological Society of America Memoir 159, p. 213-240.

Johnson, D.M., Hooper, P.R., and Conrey, R. M., 1999, XRF analysis of rocks and minerals for major and trace elements on a single low dilution Li-tetraborate fused bead: Advances in X-ray Analysis, v. 41, p. 843-867.

Meinert, L.D., 1992, Skarns and skarn deposits: Geoscience Canada, v. 19, p. 145-162.

Meinert, L.D., 1995, Compositional variation of igneous rocks associated with skarn deposits- Chemical evidence for a genetic connection between petrogenesis and mineralization, *in* Thompson, J.F. H., ed., Magmas, fluids, and ore deposits, Mineral Association of Canada Short Course Series, v. 23, p.401-418.

Meinert, L.D., 1997, Application of skarn deposit zonation models to mineral exploration: Exploration and Mining Geology, v. 6, p. 185-208.

Meinert, L.D., Hefton, K., Mayes, D., Tasiran, I., 1997, Geology, zonation, and fluid evolution of the Big Gossan Cu-Au skarn deposit, Ertsberg District, Irian Jaya: Economic Geology, v. 92, p. 510-517.

Potter, G., Garwood, D., Bush, J., 1999, Bedrock geology of the northwestern part of the Clearwater Embayment; Part I, Deary and Stanford quadrangles, Latah County, Idaho: Geological Society of America Abstracts with Programs, vol. 31, no. 4, p. 52.

Priebe, J., and Bush, J., 1999, Bedrock geology of the northwestern part of the Clearwater Embayment; Part II, Little Bear Ridge and Texas Ridge quadrangles, Latah County, Idaho: Geological Society of America Abstracts with Programs, vol. 31, no. 4, p. 52.

Reidel, S.P., Tolan, T., Hooper, P., Beeson, M., Fecht, K., Bentley, R., Anderson, J., 1989, The Grande Ronde Basalt, Columbia River Basalt Group; stratigraphic descriptions and correlations in Washington, Oregon, and Idaho, *in* Reidel, S.P, et al., eds., Volcanism and tectonism in the Columbia River flood-basalt province: Boulder, Colorado, Geological Society of America, Special Paper 239, p. 21-53.

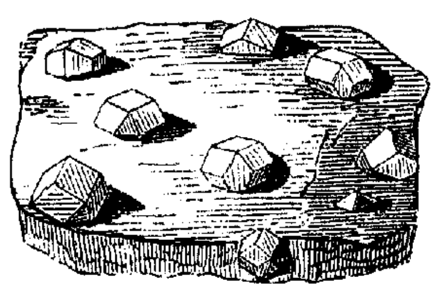
Tolan, T.L., Reidel, S., Beeson, M., Anderson, J., Fecht, K., Swanson, D., 1989, Revisions to the estimates of the areal extent and volume of the Columbia River Basalt Group, *in* Reidel, S., et al., eds., Volcanism and tectonism in the Columbia River flood-basalt province: Boulder, Colorado, Geological Society of America, Special Paper 239, p. 1-20.

Umpleby, J.B., 1914, The genesis of the Mackay copper deposits, Idaho: Economic Geology, v. 9, p.307-357.

Wakeman, B., 1991, Little Bear Creek Project Report: Kennecott Exploration file report, 13 p.

Webster, G.D., and Nunez, L., 1982, Geology of the steptoes and Palouse Hills of eastern Washington, a roadlog of the area south of Spokane, Washington, *in* Roberts, S., et al., eds., Guidebook from the 5th Annual TRGS Field Conference, Spokane, WA., p. 45-57.

Wilson, A.B., Soulliere, S., and Skipp, B., 1995, Mineral deposits of the Mackay



MOLAR TOOTH STRUCTURES IN PROTEROZOIC CARBONATE ROCKS: A PROPOSED BIO-GEOCHEMICAL MODEL FOR GENERATING CALCITE FILLED GAS VOIDS AND DOLOMITE BY SUCCESSIONAL SYNTROPHIC MICROBES MEDIATING GAS FORMATION AND CARBONATE PRECIPITATION

John F. Rittel*, Don Winston and George Furniss
Geology Department, University of Montana, Missoula, Montana 59812, U.S.A.
** P.O. Box 1863 Helena, Montana 59624, U.S.A.*

ABSTRACT

Molar tooth structures (MTS) in the Belt Supergroup, Montana, are well preserved, compacted, macroscopic features of complex, thin, interconnected, vertical ribbons (VR) or sheets, horizontal ribbons (HR), and spheroidal blobs (O'Connor 1967, 1972) all contained in cyclic units of muddy, ferroan dolomite host rock. MTS are composed of tightly packed nearly pure, uniform, microcrystalline calcite crystals 5-15μ in diameter displaying late, Fe^{++} enriched rims and inter-crystalline cements as revealed with potassium ferricyanate staining. Bishop *et. al. (*2003) discovered similar Fe^{++} rich rims in African MTS. Furniss *et. al.* (1994, 1998) created expansion cracks in mud-gas models in the lab with nearly identical morphologies to MTS and precipitated similar calcite crystals concluding that MTS are biogenic, gas induced structures that were preserved by early, calcite precipitation.

Careful observation of MTS indicate that spheroids (bubbles) were formed first, followed by small cracks, that commonly initiate from the spheroid wall and taper outward in a radial fashion. Next, HR and VR form, which in many specimens completely dissect these spherical, bubble-features and form a complex interconnected network through the sediment. HR originate from expansion cracks that trend along bedding planes. VR often express vertical, parallel orientation. In outcrop, VR commonly display bending, shearing or breakage patterns and are commonly re-deposited as angular broken "hash" fragments in storm beds suggesting they formed, filled and solidified prior to lithification of the host sediment. The host sediment and some MTS calcite contains finely disseminated pyrite, indicating anoxic conditions with dissimulatory sulfate reduction and early calcite precipitation.

The host dolomite rocks in thin-section show a finely laminated fabric containing dispersed organic particulate remnants, amorphous kerogen and occasional smooth-walled acritarcs. Dolomitic cycles interpreted in the sediments, containing MTS may very well represent ancient microbial "blooms" or eutrophication events that resulted in rapid accumulations of organic matter (OM) mixed with solution precipitated sulfate (gypsum) and detrital iron oxide silt (hematite/goethite) deposited on the bottom of the pool. Subsequent microbial remineralization of OM with the inorganic components would lead to large volumes of CH_4, CO_2 and bicarbonate generated within the sediments.

The presence of fossil gas bubbles or blobs interpreted in the MTS host rock suggest decomposition of the OM substrate likely began with microbial fermentation. In this initial acetogenic decomposition, CO_2 gas is rapidly expelled from the low pH pore water solution and trapped as gas bubbles within the cohesive

mud sediment. As the redox potential of the substrate decreases, anaerobic sulfate-reducing bacteria become dominate and consume organic acids produced by the fermenters using sulfate as the primary electron acceptor: CHO + SO_4 4 $H_{22}S$ + HCO_3 (Claypool and Kaplan, 1974; Carothers and Kharaka, 1980). The lack of siderite in Belt rocks suggests amorphous iron oxides were limited in the sediments, while sulfate was abundant (Bonneville *et. al.* 2003; Vargas *et. al.*, 1998). H_2S produced by the sulfate reducing bacteria additionally binds iron (Berner, 1981) and inhibits both iron reducing and methanogenic bacteria (Claypool and Kaplan, 1974). Some sulfate-reducing bacteria can reduce Fe oxides, but it is probably not an energy conservation process (Lovley *et. al.*, 1993; Coleman *et. al.*, 1993). With rising pH, and increasing $CO_{2,}$ calcium bicarbonate becomes supersaturated as a solution in the sediment void spaces. The sediment trapped gas expands into bubbles, micro-cracks develop from bubbles and merge forming complex networks and conduits allowing eventual CO_2 degassing of the sediment, contributing to a further rise in pH. With decreasing PCO_2, calcium bicarbonate saturated in the pore fluids rapidly precipitates as vaterite in the void spaces created by the microbial gas $CaH_2(CO_3)_2$ $CaCO_3 + CO_2 + H_2O$ (Gellatly and Winston, 2000).

When the sulfate supply is exhausted, acetigenic and iron-reducing bacteria flourish producing H_2, CO_2, formate and acetate. Crystalline, detrital Fe^{+++} is slowly reduced to Fe^{++} and iron becomes mobile in the pore waters forming a detectable late Fe^{++} component of MTS carbonate cements and the dolomite crystal precipitates $2H_2O + CHO + 8Fe^{+++}$ $8Fe^{++} + 2CO_2 + 8H^+$ (Surdam et. al., 1984). With Fe^{+++} and sulfate inhibitors removed, methanogenic bacteria finally play the dominate role in mineralization of OM. Methanogenic consortia are restricted to H^+, H_2, CO_2, formate or acetate produced by the previous microbial populations, and generate CH_4 and CO_2 from these organic substrates. Host sediment organogenic dolomites were likely formed during methanogenesis at the cellular level, during microbial metabolism and membrane transport of dehydrated Ca^{++} and Mg^{++} ions as enzyme cofactors and by active cellular dialysis and excretion of Ca^{++}, Mg^{++} and CO_2 (Kempe and Kazmierczak, 1994). As Mg^{++} becomes depleted in the system, Fe^{++} substitutes for Mg^{++} in the dolomite crystal (Mazzullo, 2000). Ca^{++} and Mg^{++}sources come from cellular assimilation of SO_4/PO_4 and organic-acid Ca/Mg salt complexes, dissolution of precursor carbonates, cationic exchange from clay particles, overlying pool and pore-water recharge.

Carbonates of methanogenic origin are typically ^{13}C enriched due to the discriminatory nature of methanogens for ^{12}C (Mazzullo, 2000; Bottinga, 1969), and Middle-Belt dolomites in Montana show ^{13}C values of near zero (Frank *et. al.* 1997) indicating that Belt dolomites are indeed methanogenically derived.

Intense studies of microbial gas formation in modern cohesive marine sediments by Van Kessel and Van Kestern, (2002) corroborate our work and show that after carbon dioxide and methane gas generated by bacterial decomposition of organic matter in sediments becomes saturated in pore solution and when the vapor pressure exceeds the liquid pressure, bubbles will be formed. In fine-grained sediments, bubble pressure works directly on the sediment skeleton, whereas in course grained, sandy sediments the gas displaces the pore water between the sediment grains. Viscosity and shear strength of the sediment and temporal hardening of material over a bubble prevent its rise and escape, making this an unlikely mechanism of gas transport. Once bubbles reach a critical size, however, they become unstable and initiate cracks in the sediment by ductile and hydraulic fracture processes. After a fracture plane is formed it becomes saturated with pore water and adjacent bubbles connect by coalescence of fractures. Gas can then be transported through the crack openings when the width exceeds the pore size for gas entry. Once the cracks are established between bubbles, the cavities are filled with both gas and pore water. Channels are created as a result of pore water drainage during consolidation and by merging cracks produced during bubble growth. Consolidation channels have a vertical dimension and merge with bubble cracks thus creating parallel venting channels from deep within the sediment.

References

Berner, R.A.,1981, A new geochemical classification of sedimentary environments: Jour. Sed. Pet., v.51, p.359-365.

Bishop, J. W., and Sumner, D.,2003, Geochemistry of molar tooth structure microspar from the neoarchean Montiville Formation, South Africa: GSA programs with abstracts Seattle annual meeting, paper no. 187-6.

Bonneville, S., Behrends, T., Haese, R., Van Cappellen, P., 2003, Thermodynamic constraints on microbial iron oxide reduction: Geophy. Res. Absts., v5, p. 12591. European geophysical soc.

Bottinga, Y., 1969, Calculated fractionation factors for carbon and hydrogen isotope exchange in the system calcite-carbon dioxide-graphite-methane-hydrogen-water vapor: Geochimica et Cosmochimica Acta v. 33, p. 49-64.

Carothers, W.W., and Kharaka, Y.K., 1980, Stable carbon isotopes in oil-field waters and the origin of CO2 : Geochimica et Cosmochimica Acta, v. 44, p. 323-332.

Claypool, G.E., and Kaplan, I.R., 1974, The origin and distribution of methane in marine sediments, *in* Kaplan, I.R., ed., Natural gases in Marine Sediments: Plenum Press, p. 99-139.

Coleman, M.l., Hedrick, D.B., Lovley, D.R., White, D.C., and Pye, K.,1993, Reduction of Fe(III) in sediments by sulphate-reducing bacteria. Nature v. 361, p.436-438.

Frank, T.D., Lyons, T.W. and Lohmann, K.C, 1997. Isotopic evidence for the paleoenvironmental evolution of the Mesoproterozoic Helena Formation, Belt Supergroup, Montana, USA. Geochimica Cosmochimica Acta, 61, 5023-5041. (5)

Furniss, G., 1990, Gas bubble and expansion crack origin of "molar tooth" calcite structures in the Middle Proterozoic Belt Supergroup, Western Montana [unpubl. M.S. thesis]: University of Montana, Missoula, 45 p.

Furniss, G., Rittel, J.F., and Winston D. 1994, Gas bubble and expansion crack origin of "Molar tooth" calcite structures in the Middle Proterozoic Belt Supergroup, Western Montana. Northwest Geol. V.23, p. 93-96.

Furniss, G., Rittel, J.F., and Winston D. 1998, Gas bubble and expansion crack origin of "molar tooth" calcite structures in the Middle Proterozoic Belt Supergroup, Western Montana, Jour. Sed. Res., vol 68 no.1, p 104-114.

Gellatly, A.M, and Winston, Don, 1998, Is "molar-tooth" calcite in the Helena Formation (Middle Proterozoic Belt Supergroup, Montana) inverted vaterite?: [Abstract] Geological Society of America Program with Abstracts, vol. 30, no. 7, p. 333.

Kempe, S., and Kazmierczak, J., 1994, The role of alkalinity in the evolution of ocean chemistry, organization of living systems and biocalcification processes: Bulletin de l'Institut Oceanographique Manaco, v. 13, p. 61-117.

Kessel van, T and Van Kesteren, W.G.M., 2002, Gas production and transport in artificial sludge depots, Waste Management v. 22/l, p. 53-62.

Kesteren, W. van and Kessel, T. van, 2002, Gas bubble nucleation and growth in cohesive sediments. In: J.C. Winterwerp and C. Kranenburg (Eds.), Fine Sediment dynamics in the Marine environment. Proc. In Marine Sci. 5, Elsevier, Ansterdam. pp. 329-341.

Lovley, D.R., Roden, E.J., Phillips, J.P., and Woodward, J.C. 1993, Enzymatic iron and uranium reduction by sulfate-reducing bacteria. Mar. Geol. V.113, p. 41-53.

Mazzullo, S.J., 2000, Organogenic dolomitization in peritidal to deep-sea sediments: Jour. Sed. Res. V. 70, no. 1, p. 10-23.

O'Connor, M.P., 1967, Stratigraphy and petrology across the Precambrian Piegan Group-Missoula Group boundary, Southern Mission and Swan ranges, Montana [unpubl. Ph.D. thesis]: University of Montana, Missoula, 269 p.

O'Connor M.P., 1972, Stratigraphy and environmental interpretation of the cryptalgal organo sedimentary "Molar-tooth" structure from the Late Precambrian Belt-Purcell Supergroup: J. Geol., v. 80, no.5, p. 592-610.

Surdam, R.C., Crossey, L.S., and Lahann, R., 1984, Mineral oxidants and porosity enhancement (abs.): American Association of Petrol. Geol. Bull., v68, p.532.

Vargas, M., Kashefi, K.., Blunt-Harris, E.L., and Lovey, D.R., 1998, Microbial evidence for Fe (III) reduction on early earth. Nature 395: 65-67.

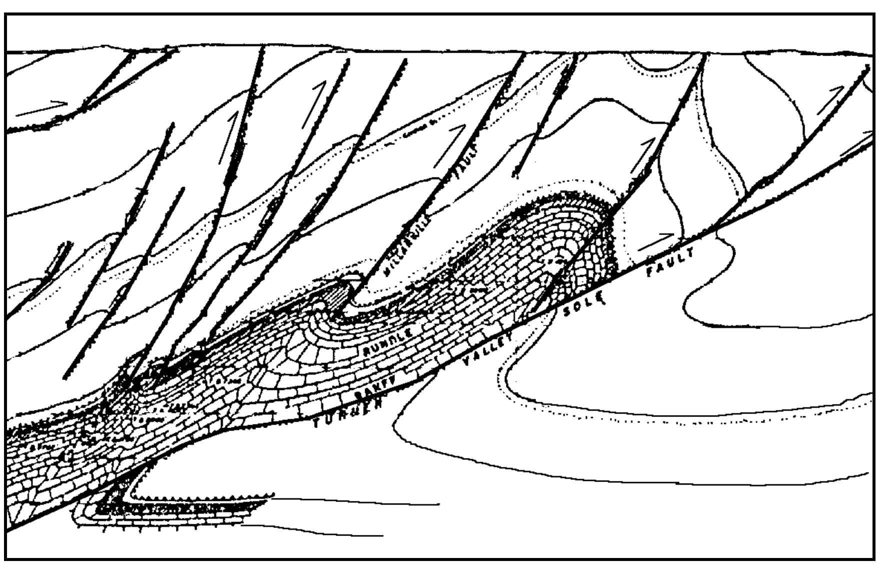
CRETACEOUS SYNCONTRACTIONAL EXTENSION IN THE SEVIER HINTERLAND, SOUTHWESTERN MONTANA

Jeff Lonn and Katie McDonald

Montana Bureau of Mines and Geology, Montana Tech, Butte, MT 59701

New 1:24,000 scale mapping in the Pintler Range of southwestern Montana suggests that extensional features developed synchronously with regional crustal contraction in the Late Cretaceous. The Kelly Lake 7½' quadrangle lies within the hinterland of the Sevier thrust belt and also in the footwall of the Eocene? Anaconda metamorphic core complex. The north-northeast-striking Georgetown thrust bisects the quadrangle, with hanging wall rocks to the northwest and footwall rocks to the southeast. The hanging wall consists of a 10,000-foot thick section of Proterozoic Missoula Group sediments intruded by Late Cretaceous plutons and deformed into open, upright folds. It also contains a major low angle normal fault, the Shadow Lake detachment. This detachment is also folded and intruded by the 73 Ma Sapphire Batholith.

Footwall rocks are more complexly deformed and contain a ductilely strained, tectonically attenuated stratigraphy with a Missoula Group section as thin as 200 feet. These rocks have been thinned by both broad, diffuse zones of ductile shear and by distinct, nearly bedding-parallel faults that omit section. Both types of structures appear to have formed simultaneously and are attributed to extension. Repetition of one of the faults, the Sawed Cabin detachment, by a thrust indicates that a convergent tectonic setting persisted. The structures were tightly folded by at least two subsequent fold events and intruded by granitic plutons that place a minimum age of Late Cretaceous on the postulated extensional features. However, at this latitude, foreland thrusting is known to have continued until the Paleocene. The fault zone previously mapped as the Cretaceous Georgetown thrust is actually comprised of Tertiary high-angle faults that cut the structures described above. Although the zone does juxtapose the hanging wall and footwall of the thrust, the original thrust geometries are not preserved.

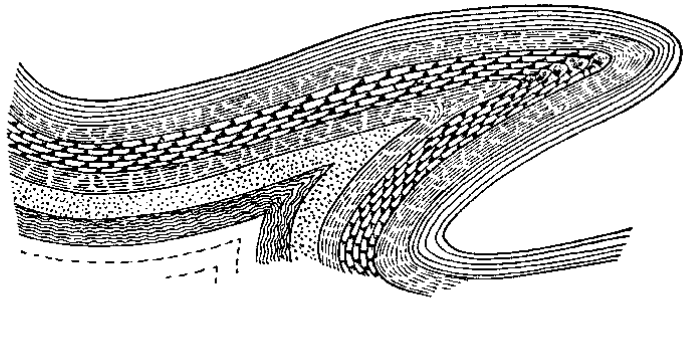


TIGHTLY FOLDED THRUSTS, TECTONICALLY THINNED STRATIGRAPHY, THE INVISIBLE GREAT UNCONFORMITY, AND OTHER MAPPING COMPLICATIONS IN SOUTHWESTERN MONTANA

Catherine McDonald and Jeff Lonn
Montana Bureau of Mines and Geology, Montana Tech, Butte, Montana, 59701

The Montana Bureau of Mines and Geology's STATEMAP program mapped two key 7½' quadrangles in 2003 to help resolve questions on regional Proterozoic stratigraphy and the extent of early Tertiary extension associated with the recently recognized Anaconda metamorphic core complex. The Dickie Hills quadrangle, located south of the core complex in the West Pioneer Mountains, contains plutonic, sedimentary, and high-grade metamorphic rocks similar to hanging wall and footwall rocks in the core complex. New mapping shows that the high-grade, dynamically metamorphosed rocks grade into unmetamorphosed sedimentary and plutonic rocks that are structurally continuous with those of the hanging wall of the core complex. The detachment zone is therefore postulated to swing along the southeast flank of the Pintler Range northwest of the Pioneer Mountains. The major structural features in the quadrangle include the Cretaceous Johnson thrust fault that is deformed into tight, northwest-trending folds, and northwest-striking sinistral (?) faults that cut the thrust and probably developed in a transtensional setting related to the Anaconda core complex. The thick quartzite section in the hanging wall of the Johnson thrust includes a thick Cambrian unit as well as previously mapped Proterozoic Belt units.

The Kelly Lake quadrangle, located in the central Pintler Range along the southwestern extension of the Anaconda core complex, illustrates the complex structural history of this region. At least five episodes of deformation are recorded that produced thrusts and detachments, a bedding-parallel mylonitic foliation, at least two subsequent fold generations, and younger high-angle faults. The high-angle faults separate the hanging wall and footwall of the Cretaceous Georgetown thrust and are of probable Tertiary age. Original thrust-fault relationships are not preserved. Tectonically attenuated Proterozoic Belt Supergroup strata in the footwall appear to be spatially associated with the bedding-parallel foliation. This structural thinning of Belt strata may help resolve



A field tour of the Castle mining district and its geologic setting

G. A. Zieg
Senior Geologist, Teck Cominco American Inc., East 15918 Euclid, Spokane, WA 99216

This trip focuses on the Castle mining district and examines intrusive phases, various mineralization styles, and host rocks for this historic camp. Participants can scrutinize both the silver and lead rich part of the district and the previously undescribed copper porphyry style alteration and mineralization.

Mile 0.0 — White Sulphur Springs. From the main intersection in town, where northbound Highway 12/89 turns east, head east through town on the main highway toward Great Falls and Billings. The view to the southeast from White Sulphur Springs shows the rugged pinnacles on the west facing flank of the Castle Mountains. These erosional remnants are composed mainly of the granite of the Castle stock and inspired the name of the range.

Mile 1.2 — Roadcuts are Proterozoic Greyson Shale. To the north is the valley of the North Fork of the Smith River, and the low hills beyond are the foothills of the Park Hills. The valley is incised in a graben containing Miocene sediments. The Proterozoic sediments lay near the base of a major thrust sheet and are carried over Permian sediments by the Willow Creek Fault.

Mile 3.0 — Here U.S. Highway 12 splits from U.S. Highway 89 at the '3-mile Y'. Continue east on U.S. Highway 12.

Mile 5.2 — These roadcuts expose Amsden Formation, which is capped by Quadrant quartzite exposed along the ridge crest further north. These and nearby exposures of Quadrant quartzite are the most northeasterly in Montana. Upper Paleozoic sediments are eroded away from here across the top of the Little Belt Mountains, and do not appear in the exposures of upper Paleozoic rocks on the north slopes of the Little Belt Mountains.

Mile 6.0 — Ramshorn caverns, developed in Devonian Jefferson dolomite, lay to the northwest (Caves in Montana, MBMG Bulletin ___). The caves are on private land and closed to the public.

Mile 6.85 — This is Four Mile Creek. This road is an alternate access to the interior of the Castle Mountains.

Mile 11.0 — To the north of the highway is Sutherlin Reservoir, built on the North Fork of the Smith River. The reservoir is named after Robert Sutherlin, the editor of the Rocky Mountain Husbandman newspaper, which he started in Diamond City in the 1860's and moved to White Sulphur Springs when the placer gold played out. Sutherlin was well known for criticizing cattlemen for overstocking of the open range in the early 1880's. This resulted in the disaster in the hard winter of 1886-7 when thousands of cattle perished. The event is best memorialized by a C.M. Russell postcard sketch of a skeletal cow staggering through a blizzard which he titled 'Last of the 5000'.

Mile 12.1— Newlan Creek Road turn off. These hills are composed of middle Proterozoic Spokane shale and Greyson shale. The contact is exposed on the hills just north of the highway. Garry Anderson (1986) provides the most complete description of internal Greyson stratigraphy in this area. Continue on highway.

Mile 15.0 – Jamison Trail turn off. This is a freight trail which provided access between Fort Benton, north of the mountains, and the area south of the Little Belt Mountains which included the Copperopolis mining camp (see below); Fort Howe, a military fort constructed along the Carroll trail about eight miles southeast of here at the junction of Spring Creek and the North Fork of the Musselshell River; Brewers Springs (later White Sulphur Springs); and Fort Logan, another military fort constructed along the Carroll trail. The primary purpose of these freight routes and forts was to ensure a supply infrastructure to the placer gold camps at Diamond City and beyond. Southwest Montana gold played a significant role in financing the latter days of the Civil War and the following reconstruction, and during these times these freight trails were heavily used.

In this location, we're on a low divide between the Castle Mountains to our south and the Little Belt Mountains to our north. The south slopes of the Little Belt Mountains, visible in the distance, consist of Paleozoic carbonate. The Belt rocks are carried north over the carbonates by the Volcano Valley thrust fault (Anderson, 1986), also called the Wagner Gulch fault by Blumer (1969). A core hole drilled by Kennecott through the Volcano Valley fault and through the base of the Paleozoic section north of here shows that Greyson shale underlies the middle Cambrian Flathead sandstone. In this location, Spokane shale rests on Greyson and underneath the Cambrian unconformity. Also in this area north of the highway are young basalts extruded from vents along the Jamison Trail and Volcano Butte. South of the highway Blumer (1969) mapped some Tertiary rhyolite and welded tuff in topographically low areas.

Mile 16.5 — On the right is the location of Copperopolis, which produced the first copper in the state of Montana in 1863, according to a Harpers Weekly edition of the day. The site is on private land presently controlled by the Bair Company. This ore was sacked on site, hauled overland along the Jamison Trail north to Fort Benton, transferred to steamboats, transported downstream to New Orleans and then transferred into larger ships for transport across the Atlantic to Swansea, Wales, for smelting. Later shipments went to Utah and to Anaconda. When W. H. Weed visited the district in the mid-1890's the camp was abandoned (Weed and Pirsson, 1896). Calcite-quartz veins with bornite-chalcopyrite-chalcocite-pyrite cut the upper Greyson shale (Blumer, 1969) and are best developed where they cut quartzite beds. The near vertical veins trend both N60W and N60E. The age of the veins is unknown, but they haven't been shown to cut nearby Paleozoic rocks. Five miles northeast in the Spring Creek drainage, the Greyson shale hosts similar veins but many of these consist of sphalerite and galena in quartz and calcite, as well as copper sulfide minerals. Copperopolis was largely abandoned by the turn of the century, though small miners produced sporadic shipments of copper ore until 1961. Kennecott and Rio Algom explored the Copperopolis and Spring Creek areas in the 1990's, including drilling on Spring Creek

South of Copperopolis, Blumer (1969) maps a northwest trending thrust which carries Mississippian Mission Canyon limestone from the southwest over Cambrian Pilgim limestone and older rocks. This younger over older relationship suggests more complex structure in the area than is immediately apparent.

Mile 19.5 — The reservoir northeast of the highway is the Bair Reservoir, built on the North Fork of the Musselshell River. This reservoir is named after Charles Bair, a prominent sheepman of the Musselshell Valley in the late 19^{th} and early 20^{th} centuries.

The exposures in the roadcuts on the right are Greyson formation and further along, the underlying carbonate-rich upper Newland Formation. The hills on the left are lower Newland formation calcareous shale. Rhyodacite dikes and sills intrude these Proterozoic sediments and contribute to the extensive alteration visible in the exposures.

Mile 20.9 — Checkerboard. Turn right at the

Checkerboard Bar on the east edge of town and start up Checkerboard Creek on the U.S.F.S. road. Checkerboard sits in carbonate and calcerous shale of the upper Newland Formation. The route we'll follow passes up section through the Belt rocks into Paleozoic sediments. Exposures are limited until we reach the Paleozoic carbonates.

Mile 21.7 – this is the approximate location of the Greyson-Newland contact.

Mile 23.5 – this is the approximate location of the Greyson-Spokane contact.

Mile 24.8 (530,220E; 5154290N) – This is the approximate location of the unconformity between the middle Cambrian Flathead sandstone, exposed along the brow of this hill, and the underlying Belt sediments (Note that for this and following location, UTM locations are provided and use the North American 1927 datum, Zone 12, projection).

Mile 25.25 (529620E; 5153990N) – There is a fork in the road here, take the left hand fork. Somewhere in the subdued terrain in the next two miles, the southeast trending Eight Mile fault that Blumer (1969) mapped south of Copperopolis passes through and juxtaposes northeast dipping Cretaceous shale against southwest dipping Cambrian carbonate and shale. Only Tanner (1949) has mapped this area, and at a scale of limited use to precisely locate this fault.

Mile 27.75 (527230E; 5151070N) – This is Limestone Ridge, made of Mississippian Madison Group carbonates, Mission Canyon limestone and Lodgepole limestone. Winters (1964) maps this as the northeast limb of a southeast plunging anticline.

STOP 1

Mile 29.1 (526710E; 5149090N) **Stop 1** — This is the Judge Mine and the beginning of the Castle Mountains mining district. The Castle Mountains consist of a complex of Eocene calc-alkaline magmas intruded into middle Proterozoic and Phanerozoic sediments. The main mass of the range consists of unaltered granite. This pluton appears orthorhombic in shape and occupies about 50 square miles. Sediments are altered over a broad area (approximately 25 square miles) off the east margin of the Castle granite. Within this area of altered sediment, and separated from the eastern boundary of the granite by about a mile, is a north-northwest trending body called the Blackhawk diorite. It cores the ridge line between Robinson Creek and the Blackhawk area in a mass about one mile wide and three miles long. The Castle granite and the Blackhawk diorite show little alteration. Other intrusives in and near the altered area are hypabyssal and include diorite, granodirotie, monzonite,quartz monzonite, andesite, dacite, rhyodacite, rhyolite and mafic dikes. Most of these intrude the northwest trending one mile wide band of altered sedimentary rock between the Castle Granite and the Balckhawk diorite. Extrusive rhyolite post-dates all the other igneous rocks and its exposures east of Bonanza Creek have a K-Ar date of 46 m.a. (Chadwick, 1980).

This location is still on the east side of the Blackhawk diorite. One half mile west of here was the small mining community of Blackhawk. The Cambrian section and Belt shales are all exposed within one half mile of us on the ridge north of the Blackhawk cabin. At this location, the Judge mine and the Felix mine were developed on Ag-Pb mineralization concentrated along the contact between a dacite porphyry dike and Jefferson dolomite on the south limb of a southeasterly plunging anticline. The crest of the anticline is upthrown behind us across a northwest trending fault. Galena, cerrusite, and anglesite occur in jasperoidal material. This is typical of most of the carbonate-hosted Ag-rich deposits in the district. Similar deposits occur in prospects deeper in the stratigraphy in Cambrian carbonates. Generally, the best hosts for sizable base metal and Ag deposits in the district are the Cambrian Pilgrim Limestone, the Devonian

STOP 2

Jefferson Dolomite, and the Mississippian Lodgepole Limestone.

Mile 29.3 (526720E; 5148770N) — Turn left at the junction of the Checkerboard road with U.S.F.S. 581 and continue southeast.

Mile 29.6 (526720E; 5148770N) **Stop 2** — This is the location of the Blackhawk mine, developed on Ag-rich MnO in the Mississippian Lodgepole Limestone of the Madison Group. The breccia body that hosts the mineralization is 5000 feet long and strikes about N78W. The mineralized zone reaches 100 feet wide, and according to Weed and Pirsson (1896) grades ran between 5 and 15 ounces Ag per ton. Workings seem limited to several hundred feet of strike length. The nearby Annie Maud and Alice mines were developed on small shoots of similar material localized along a northeast trending dike.

STOP 3

Mile 31.9 (526210E; 5146230N) **Stop 3** — This is an outcrop of Blackhawk diorite. In this area, it is only weakly altered. Thompson (2001) recognizes some internal variation in the composition of this unit but regards it as a quartz monzonite. He describes weak potassic alteration in four samples manifest as secondary orthoclase rimming both earlier plagioclase and orthoclase, and some secondary biotite after biotite. He recognized no sericitization or propylitic alteration of this rock. Cominco American drilled into it near here and also described it as weakly potassically altered quartz monzonite.

STOP 4

Mile 32.4 (525520E; 5146420N) — This was the location of the mining community of Robinson. In 1884 Mr. C. Barnes, the U.S. Postmaster in White Sulphur Springs, located the Blue Bull claim, the first lode mining claim in the district, along this southern boundary of the Blackhawk diorite.

Mile 32.9 (525200E; 5146510N) **Stop 4** — Turn to the right here to leave U.S.F.S. road 581. This location is inside the 'marble line' of the Castle Mountains alteration system. These are marbleized carbonates of the Mississippian Madison Group. To the north about 1000 feet is the Iron Chief mine which is developed on a magnetite rich deposit in the Mississippian Lodgepole Limestone. Winters (1968) described the paragenesis of ore and gangue minerals at the Iron Chief as typical of carbonate hosted deposits in the district. According to his analysis, the paragenesis is, from earliest to youngest, magnetite — pyrite — magnetite+/-hematite — jasperoid — chalcopyrite+/- bornite +/- enargite — pyrite — crystalline quartz – sphalerite (marmatite) +/- chalcopyrite – galena – MnO +/-chalcedony. In spite of the high Ag grades, no Ag minerals have been positively identified. Oxidation of deposits usually ranges from depths of 100 to 500 feet, and results in mineralization consisting of goethite, limonite, cerrusite, anglesite, and smithsonite, or in Cu-rich areas malachite, azurite, tenorite, cuprite, chrysacolla and rare native Cu. Most district production is from oxidized cerrusite and anglesite rich portions of mineral deposits.

STOP 5

Continue up the road onto Yankee Jim Ridge. This route may require 4-wheel drive.

Mile 33.9 (524310E; 5147650N) **Stop 5** — This is the top of Yankee Jim Ridge. Here one can enjoy a view of the top of the Castles and examine calc-silicate alteration in the nearby mine dumps of the Milwaukee mine. Note that mineralization accompanies an epidote-rich alteration assemblage in a fine grained intrusive rock. Since the last stop, most of the ridge is held up by a fine grained multiphase, generally porphyritic rock type called dacite porphyry (Winters, 1968; Thompson, 2001); rhyodacite (Cominco American, 1999) and quartz porphyry by Weed and Pirsson (1896) and by Tanner (1949). We will accept dacite porphyry based on the most recent petrographic by

Thompson (2001). This rock intrudes the mile-wide band of sediments between the Blackhawk diorite and the Castle granite. It shows abundant alteration and is accompanied by a variety of intrusive phases, most strongly altered. According to Thompson (2001) the dacite porphyry shows both plagioclase and orthoclase phenocrysts, quartz eyes, and both biotite and hornblende phenocrysts. Potassic alteration ranges from secondary orthoclase rims on earlier feldspar to complete replacement of earlier feldspar, and secondary biotite grows on both primary biotite and amphibole. Propylitic alteration (chlorite+/-calcite+/-epidote) is ubiquitous in unsericitized rocks. Sericite alteration follows and masks the other alteration types, and is quite strong in some areas. Some of the epidote may be related to the calc-silicate alteration event prevalent in the nearby sedimentary rocks.

Two widely separated 800-foot core holes between here and the last stop show that the dacite porphyry gives way to quartz monzonite within 100 feet of surface. The furthest one from our present location (2000 feet toward Stop 4) shows pervasive weak quartz-sericite-pyrite veinlets throughout the hole. The nearest one (1000 feet toward Stop 4) shows only sparse chlorite-pyrite veinlets and traces of galena overprinting older weak potassic alteration. In a large soil sample gride, this area shows the most Cu relative to Pb and Zn of anywhere else in the area (very low (Cu+Pb+Zn)/Cu ratios).

STOP 6

A core hole just up the road 200 feet encountered, above 400 feet, a complex assemblage of rhyodacite, diorite (modally quartz monzonite); garnet hornfels skarn; garnet pyroxene skarn; and dacite. Much of the skarn alteration overprints Cambrian Meagher limestone. Below this to a total depth of nearly 1000 feet, the hole encountered mainly quartz monzonite with some secondary biotite, areas of K-spar 'flooding' and bits of visible chalcopyrite. of the Cu grades are very low. The quartz monzonite in these holes on Yankee Jim Ridge are probably part of the Blackhawk diorite (quartz monzonite). The contact between the quartz monzonite and the dacite porphyry dips westerly into the Hensley Creek drainage.

Mile 34.4 (523800E; 5148300N) **Stop 6** – This is an outcrop of pyroxene hornfels, and petrographic work shows that much of it is a diopside-scapolite rock. In this location, there is little or no garnet, and this suggests we're somewhat distal from the heat source for the skarn. This hornfels is probably developed in Cambrian Park Shale. Our drilling has shown that this band of calc-silicate altered sediments exposed in upper Hensley Creek consist of Cambrian rocks, not Belt Series Piegan Group as mapped by Winters (1968). One hole 2000 feet to our south-southwest encountered Meagher Limestone, Wolsey shale and a thin Flathead quartzite resting on a 400-foot section of Spokane Formation shale above upper Greyson Formation shale. 4000 feet northwest of here, in the upper Four Mile Creek drainage, we've mapped extensive surface outcrops of east-dipping Flathead quartzite resting on unaltered Spokane shale. These outcrops strike south into this upper Hensley Creek area of hydrothermal alteration.

STOP 7

STOP 8

Mile 35.3 (522460E; 5148190N) **Stop 7** — Off the road to the northwest is rubble crop of Castle granite. The granite shows very little, if any, alteration. Thompson (2001) describes the granite as a leucocratic, largely holocrystalline granite with quartz > K-spar > plagioclase. Biotite and amphibole occur in low quantities. Thompson recognized widespread potassic alteration as secondary orthoclase overgrowths on earlier feldspar and secondary biotite on earlier biotite and amphibole. Thompson (2001) recognized little propylitic and no argillic alteration in the granite.

Mile 36.15 (522370E; 5147250N) **Stop 8** — Along this road is strong quartz-sericite-pyrite alteration overprinting the Castle granite. This is one of the few areas of Castle granite which shows strong alteration. The alteration zone is broad here, measureing over 3000 feet north-south, and extends and narrows east toward the forks of Hensley Creek. Its abrupt southern boundary appears controlled by an east-west fault which manifests topographically as a drainage bottom. Drilling here encountered 815 feet of intensely brecciated and quartz-sericite-pyrite altered quartz monzonite breccia above more weakly altered granodiorite and quartz monzonite. The brecciated area is cut by altered granodiorite and dacite porphyry dikes. The hole contains a thin zone (36 feet grading 0.2 % Cu) of weak supergene copper enrichment manifest as chalcocite coatings on pyrite. Deeper in the hole, Zn and Pb concentrations reach 0.5% over tens of feet within the breccia zone. The unusual mineral vivianite occurs in radial clusters in a few places between 400 and 600 feet deep in the hole. Below 600 feet, alteration diminishes and the hole penetrates nearly fresh granodiorite and quartz monzonite. This quartz monzonite approaches granite in composition and could be a phase of the Castle granite. The sericite in the altered zone gives a K-Ar age date of 50 m.a. This area shows high Pb relative to other metals in an area wide soil sample survey (very low (Cu+Zn+Pb)/Pb ratios)

STOP 9

500 feet northeast another hole intersected 300 feet of sericite-altered quartz monzonite resting on unaltered quartz monzonite and granite cut by weakly altered dacite porphyry containing scattered quartz-sericite-pyrite veins and veinlets with galena and chalcopyrite.

Mile 37.0 (523640E; 5147140N) **Stop 9** — In this location we've re-entered calc-silicate altered sedimentary rocks. This is the upper end of the Belle of the Castle claim. A drillhole 500 feet east of this location passed through a quartz eye tonalite and encountered a marble zone bounded by garnet-pyroxene skarn and hornfels with weak copper mineralization. The hole ends in diorite. On the surface here, you can find copper mineralization in garnet-pyroxene skarn; presumable some of the same material encountered in the drillhole.

Another drillhole about 600 feet to the north-east encountered quartz feldspar porphyry with strong quartz-sericite-pyrite alteration and small zones of magnetite and hornfels and skarn. This hole encountered 130 feet of 0.7% Cu near its top, an interval which included all the calc-silicate altered sediment in the hole. Alteration of the porphyry continued to the base of the hole at 340 feet. This hole is quite close to the discovery shaft for the Belle of the Castle claim.

Continue down the road to lower elevations.

Mile 37.6 (524100E; 5146390N) — This is the Yellowstone Mine, one of the largest producers in the district and typical of the oxidized Zn-Pb-Ag replacement deposits in the carbonate. The mine was discovered in 1886 and the first smelter in the district, located along Castle Creek, was built to process its ore. According to Winters (1968) its high grade Pb-Ag bodies were localized along the contact between a dacite porphyry sill and Cambrian Pilgrim limestone. The ore is galena and cerrusite in jasper. Park shale and intrusive rocks are argillized in this area. Work in this mine continued as recently as the mid-1970's.

Mile 38.15 (524770E; 5145850N) – This is the junction of ridge road with main U.S.F.S. road 581. For a continued tour of the center of the skarn system, turn left and proceed to the Hensley Creek drainage bottom.

Mile 38.6 (524510E; 5146500N) — Turn left and drive up Hensley Creek. This route may require 4-wheel drive. There is little exposure in the creek bottom, but the route passes through Paleozoic marble and dacite porphyry. Park by the locked gate at the forks of the creek. A hole here encountered garnet-rich endoskarn

STOP 10

with 122 feet grading 0.4% Cu. The mine portal across the creek is part of the Belle of the Castle mine, located in 1889 by the Hensley brothers. Little development of these copper prospects took place until after 1900. Weed and Pirsson (1896) make no mention of the copper prospects in this part of the district. Continue walking up the right hand fork of the creek and climb onto the dozer trails up the hill on your right to Stop 10.

Mile 39.4 (523800E; 5147530N) **Stop 10** – Park and walk up the right hand drainage and climb the caterpillar cuts up the hill to approximately 523870E; 5147710N. These cuts are in garnet pyroxene skarn which overprints Cambrian sediments on the Copper Bowl and Copper Kettle claims. Much of this material is gossan, some is also magnetite rich. Cu carbonates are abundant. A variety of sills and dikes cut the hillside and have been variably logged in local drillholes as tonalite, trachyte (K-altered andesite), rhyodacite, diorite, and granodiorite. Some of the material is endoskarn but the majority is banded and is derived from a Cambrian sedimentary protolith. It is impossible to tell which Cambrian units are mineralized but none of the drilling is deep enough to encounter the quartzite which was encountered in a hole 300 feet up the drainage (described at Stop 6). Precious metals values are low, though Winters (1968) reports a surface sample of $400/ton Au in this area. Values up to 1 g/t Au have been encountered in drilling, but usually the concentrations are undetectable.

$\boxed{\text{STOP 11}}$

Return to the road and travel back down the creek.

$\boxed{\text{STOP 12}}$

Mile 40.2 (524510E; 5146500N) — Turn right at the junction with U.S.F.S. road 581.

Mile 40.64 (524770E; 5145850N) – Turn right onto ridge road.

Mile 40.93 (524420E; 5146170N) – Take the road fork to the left which angles down the hill slope.

Mile 41.3 Stop 11 — The Hamilton mine workings (524120E; 5146170N) access the same mineralization as the Yellowstone mine. This is a good place to examine ore types similar to those decribed from the nearby Yellowstone mine. After examining some material on the mine dump, continue down the drainage to the main road and continue on it down drainage to the next stop.

Mile 42.7 Stop 12 — The Cumberland mine (524450E; 5144070N) was the largest producer in the district, and in 1891 was the largest producer of lead ore in Montana. The deposit was discovered in 1886 by Lafe Hensley and his sons. Production began in 1888 with shipments grading 60% Pb and 25 ounce/ton Ag. A smelter was located on site (note the black slag pile) and operations went well until the Ag market collapsed in 1893. The mine operated only intermittently until 1956, after which it had a decade of good production. The mine then operated intermittently until the mid-1970's. The deposit is at the contact between the Madison Group limestone and Castle granite. The ore types are similar to those at the Yellowstone.

$\boxed{\text{STOP 13}}$

The route downstream continues up section through southeast dipping stratigraphy. The road passes through Mississippian Big Snowy Group and Amsden Formation, and then across an unconformity and through a thin Jurassic section into Cretaceous Kootenai Formation and then Colorado Group shale.

Mile 43.75 (525300E; 5142800N) **Stop 13** — This is the site of Castle Town, and as you can see many of the original buildings are left standing. It is on private property, and can only be entered with permission of the landowners. The heyday of Castle Town lasted only a few short years, when it reached a population of 1,500 and was serviced by the Jawbone railroad, a spur from the Milwaukee line. R. A. Harlow (after whom Harlowton is named) was the main promoter for the railroad, and is said to have 'wagged his jawbone' until he managed to get financing to build it. The Jawbone railroad was completed in 1896 but saw appreciable service for only two years. Among other personalities of the day, Calamity Jane is reported to have spent time in Castle Town. The site is in Colorado Group shale. From here on for about one mile Winters (1968) mapped scattered syenite sills in the Colorado Group and overlying Eagle sandstone and Judith River Formation. Further on, the Cretaceous stratigraphy includes the Bearpaw shale, Lennep sandstone, and Hell Creek Formation. Winters (1968) and Weed and Pirsson (1896) mapped glacial morainal material on the low ridge tops north and south of the valleys of Castle Creek and Alabaugh Creek.

The main road route to Lennep from Castle covers about 7.2 miles. At Lennep, you can turn right onto the state highway and follow it west to its junction with Highway 89, then turn north and drive back to White Sulphur Springs. An alternate route is to turn left at Lennep and follow the state highway 12 miles to Martinsdale, and then turn north to the junction of the state highway with U.S. highway 12. There, one can travel back to White Sulphur Springs or east toward Harlowton and Billings.

This concludes our tour of the Castle mining district.

REFERENCES

Anderson, G.E., 1986, Geology of the southern two-thirds of Volcano Butte and Hoover Springs quadrangles, Meagher County, Montana, M.S. thesis, Montana College of Mineral Science and Technology.

Blumer, J.W., 1969, Geology for the Deadman Canyon-Copperopolis area, Meagher County, Montana, M.S. thesis, Montana College of Mineral Science and Technology.

A field tour through the Little Birch Creek section of the Middle Proterozoic upper Newland Formation

G. A. Zieg

Senior Geologist, Teck Cominco American Inc., East 15918 Euclid, Spokane, WA 99216

INTRODUCTION

This half-day field trip examines excellent exposures of the stratigraphy of the upper Newland and highlights C. D. Walcott's Newlandia assemblage. This is the type locality for the spectacular specimens collected for Walcott in 1914 and initially interpreted by him as algal remains. The stops also provide a fine opportunity to discuss the changing depositional conditions in the middle Proterozoic Helena embayment in the transition from Newland to Greyson Formations.

This trip is entirely on private land, and permission must be granted from the landowners prior to accessing these locations.

ROADLOG

Mile 0.0 — White Sulphur Springs. From the main intersection in town where northbound Highway 12/89 turns east, head west on the Fort Logan road. The travel route will cross the northwest axis of Smith River Valley. Gravity work by Gogas (1984) and Gierke (1987) show that the valley bottom is probably less than 2000 feet below surface.

Mile 1.8 — Here the highway curves to the right, but a graveled county road continues straight. Go straight on the graveled county road (Duck Creek road) and continue west.

Mile 4.9 — Smith River bridge.

Mile 5.5 — Greyson shale on River Hill/ Woods Creek. This is section in the middle of the Greyson Formation shale is unusual in that it contains interbeds of buff-weathering silty limestone.

Mile 8.2 — Junction between private access road to left, and Birch Creek road, which curves to the right and crosses Birch Creek just beyond this point. Take the private access road on your left.

The ranch buildings along Birch Creek straight ahead are on the location of the Dave Folsom ranch. Folsom was a pioneer friend of mountain man Jim Bridger who on his advice toured Yellowstone Park prior to the better known 1872 expedition. One half mile further along the Birch Creek road on the right are the former headquarters of the Birch Creek Ranch, once owned by the Ringling Brothers of the Ringling Brothers Circus. The Ringling Brothers had extensive holdings in Meagher County during the early 20th century. They later sold the property to Wellington D. Rankin, a prominent Helena attorney with extensive ranch property in Montana and brother of Jeanette Rankin, the Montana congresswoman. This is also the general location of a successful effort by a group of fur trappers led by Jeremiah "Liver-Eatin" Johnson to liberate some of their stolen horses from a band of Blackfeet warriors (from Vardis Fisher's *The Mountain Man*).

Mile 9.2 — This broad dike is a Tertiary granodiorite. This location will also be the end of the walking tour of the upper Newland and lower Greyson section here. After stop 7, return to here and proceed to stop 8.

Mile 9.5 — Take the right hand turn and proceed down the coulee on your right into the Little Birch Creek drainage. Continue around the switchback and drive into the ranch headquarters. This is the starting point for a walking tour of the upper Newland. Refer to the map in Figure 1 for directions to the stops on the walking tour. Each stop will be identified on the map and with GPS coordinates for precise locations. From the ranch buildings, head upstream on the east side of the creek bottom, through the gate in the fence, and take the trail along the base of the hill. You may have to search for an appropriate deer trail to take you to the first outcrop.

Mile 9.9 — This is the Zieg ranch house. From here you must walk beyond the house up the Little Birch Creek drainage along the base of the steep hill. Beyond the fence about 100 yards, look for a deer trail angling up the hill on your left. The first stop is beyond the steep drainage bottom which is crossed by the deer trail. The stops in this road log include UTM coordinates using a projection of North American datum 1927, Zone 12. They are estimated at about 5 meter accuracy.

STOP 1

Stop 1 (492825E; 5153320N). — The stratigraphic divisions and nomenclature used in these descriptions follows the scheme described by Zieg (1981, 1986) who reported two 'shallowing upward' sequences with a relatively clean carbonate at the base (Unit III and Unit VI), a silty carbonate or calcareous shale with more current transported sediment capping this (Unit IV and Unit VII), and a carbonate deficient rusty weathering silty shale capping each cycle (Unit V and lower Greyson Formation). The lone outcrop below the base of the cliff is an exposure of Unit III of the upper Newland Formation. Here, Unit III consists of a medium gray weathering limestone with poorly defined bedding and black chert stringers and nodules. Further up the drainage, Unit III exposures show forms of Walcott's Newlandia assemblage, which we'll visit and describe at later stops.

Up the hill 20 feet are excellent cliff face exposures of Unit IV of the Upper Newland Formation. These are silty limestone beds with shaly partings between them. Note the apparent dissolution of carbonate on the margins of the beds, and the lateral gradation of some carbonate beds into shaly partings. Proceed further up the hill in the break between the cliffs, and look at the upper part of the unit.

STOP 2

Stop 2 (492855E; 5153345N). — These exposures, near the top of Unit IV, show beautifully rippled silts within the carbonate beds, and 'pseudonodules' formed by silt ripples which sank into unconsolidated carbonate mud below. Silt ripples could indicate storm current activity. Again, note the apparent dissolution of carbonate beds resulting in carbonate nodules encased in shaly partings. A short distance above these in the upper bank of cliffs is a black limestone unit which caps Unit IV. This is an unusual rock type in the Newland. The route to Stop 3 is primarily in silty, noncalcareous shale of Unit V. In the small coulee which cuts down the hillside between stops 2 and 3 is a small sand dike cutting silty shale of Unit V.

STOP 3

Stop 3 (492955E; 5153545N). — These outcrops of Unit V show scattered sandy lenses in noncalcareous, fissile, rusty weathering shale. This lithology is typical not only of this unit, but also of the Greyson Formation.

STOP 4

Stop 4 (493035E; 5153560N). — These outcrops of Unit V silty shale contain arkosic sand lenses and layers, and carbonate concretions. Scattered carbonate concretions are typical in this unit but unusual on lower Greyson shale except near the contact with Newland Formation.

STOP 5

Stop 5 (493110E; 5155590N). — Note that the rusty weathering silty shale has given way to a dark gray, fissile shale with little silt or sand. A few limestone beds similar to those in overlying Unit VI occur at the transition between

STOP 6

the shale types. Between here and the limestone cliffs at the next stop, black limestone nodules become abundant in the gray shale. These contain no sedimentary structures and appear concretionary.

Stop 6 (493160E; 5153610N). — This is Unit VI, a clean, medium gray, thin to medium bedded carbonate. This unit hosts abundant forms of Walcott's Newlandia assemblage. Visible in these outcrops are examples of *Newlandia major*, *Newlandia concentrica*, *Copperia tubiformis*, and *Greysonia basaltica*. Zieg (1981) renamed these based on morphologic appearance and describes them as bedded ropes; plates arranged oblique to bedding; and hemispheroids (both concave and convex upward). Hemispheroids, both upright and inverted, occur stacked and dominate the base of the outcrops. A layer of ovoids occupies the bases of these stacks. Both hemispheroidal layering and concentric banding of ovoids grade from dark to light in toward the center of the form. The color change results from slight changes in the proportion of insoluble residue to undissolved calcite. These bands and layers overprint the primary bedding lamination, and so are a post-depositional feature. Stacks of hemispheroids and ovoids are bounded by solution boundaries with concentrations of insoluble residue. Ropes and plates dominate the upper part of the outcrops. Etching of ropes and plates shows that both types transgress original bedding lamination, and lay in a matrix of insoluble residue. Plates are easily explained as a result of dissolution along a regularly spaced fabric, such as a cleavage, oriented oblique to the plane of bedding. Ropes result from further dissolution of plates along bedding planes. Preserved laminations in ropes show a chaotic variety of orientations, a result of the collapse of laminations during dissolution of the bed.

The Little Birch Creek section is the type location of the Newlandia assemblage first described by C. D. Walcott (1914) as algal forms. These forms have been subsequently described as inorganic products of weathering of carbonate along cleavage (Fenton and Fenton, 1936), pseudofossils and dubiofossils (Gutstadt, 1975), and as products of pressure solution (Zieg, 1981). Eby (1977) has referred to these as a type of molar tooth structure. Verrall (1955), McMannis (1963), Boyce (1975), and Bonnet (1979) retained molar tooth and algal interprations for similar forms in the Newland carbonate of the Horseshoe Hills. Trunk and Smith (1979) have interpreted these forms in the Horseshoe Hills as a result of soft sediment deformation. Proper interpretation of these features has some bearing on understanding the depositional setting of the upper Newland Formation. Zieg's (1981) subwave base interpretation of the upper depositional environment conflicts with an algal interpretation for Newlandia. An inorganic interpretation for the origins of the Newlandia assemblage precludes its use as a biostratigraphic marker.

Walcott's descriptions of his Newlandia assemblage was based on a box of rocks sent him from this location by a local rancher named Manassa Collen. Collen also included specimens of stromatolites from the Spokane formation in the box and Walcott, in appreciation, named these Collenia. Awramik and others (1993) recently relocated these sites. Collen apparently had a strong appreciation of local and general geology. He aided C. D. Walcott, W. H. Weed, and Ransome and Calkins in a number of their field excursion in the 1890's and early 1900's. Collen owned and lived on a ranch on Battle Creek about 15 miles south of White Sulphur Springs between the years of 1896 and 1916. He left no personal information and he apparently had no family. We have copies of his letters to Walcott and to Weed, and in them he discusses the merit of various structural, stratigraphic and sedimentologic interpretations for rocks in the area, as well as the genesis of ores in the local mining districts. Some of the writing refers to the mundane details of his shipping tons of samples back to Washington D.C. for further study. Together with his geologic insights, his references to the geology of other parts of the world suggest he had some formal scientific training. One significant interpretation made by Collen based on his observations of field relationships along the Smith River was that the Volcano Valley Fault showed Precambrian movement. From the tone of Collens writing, Walcott and Weed apparently disputed the idea, and Collen was again attempting to convince them of its merit. We know now that it is indeed a Precambrian fault. For me, Collen remains a mystery geologist who had an underappreciated influence on the first serious geologic investigations in the area by C. D. Walcott and W. H. Weed.

STOP 7

Stop 7 (493105E; 5153830N). — Capping Unit VI is Unit VII, here dominated by calcareous shale. A megaripple of calcarenite suggests deposition within storm wave base. In other areas (such as along Newlan Creek, in the type Newland section) Unit VII exhibits a facies identical to the Unit IV facies along Birch Creek. Here in Unit VII, the facies is quite shaley, similar to both Unit IV and Unit VII along Decker Gulch. Lateral transitions between calcareous shale facies and silty carbonate facies is typical of both Unit IV and Unit VII. Just up the hill crossing the road are outcrops of a Tertiary granodiorite dike which cuts the Proterozoic section at the Newland-Greyson contact.

STOP 8

From this point, one should drive to Stop 8 via the Little Birch Creek road. **Reset odometer to 0.0 miles.**

Mile 1.7 Stop 8 (492170E; 5151940N) — In the bottom of the draw by the road are exposures of Unit I of the Upper Newland Formation.. Here, Unit I is a buff-weathering, even and medium bedded limestone. Though limestone units can occur deeper with the the lower Newland shale, this represents the first regionally mappable carbonate unit in the Newland Formation, and so on that basis Zieg (1981) chose it as the base of upper Newland. In the roadcut are exposures of Unit II, evenly laminated calcareous shale. This lithology is identical to lower Newland calcareous shale. Walk northeast toward the low bank of dark gray outcrops on the hill slope.

STOP 9

Stop 9 (492340E; 5152090N). — These are outcrops of Unit III of the upper Newland Formation. They contain several members of Walcott's 1914 Newlandia assemblage including *Newlandia lamellosa*, *Copperia tubiformis*, *Greysonia basaltica*, and *Kinneyia simulans*. Zieg (1981) renamed these on the basis of their morphology as ropes, plates, thinly laminated structures, hemispheroids and mottling. Here in Unit III, diagenetic black chert forms a matrix to ropes and plates. In thin section, the ropes and plates here are rimmed by radialaxial calcite. As in the other location, we interpret these as forms of diagenetic dissolution, and likely pressure solution, of the carbonate. In both Unit III and in Unit VI, the symmetry of hemispheroidal, ovoid, or circular forms are arranged about a vertical axis. This suggests downward stress, either from gravity or from lithologic load, formed the fabric which subsequently focused the dissolving fluids.

STOP 10

Return to the vehicles and drive back along the Little Birch Creek road toward the Duck Creek road. **Reset odometer to 0.0 miles.**

STOP 11

Mile 0.53 — park on the road and walk up the slope to the band of outcrops which encircle the butte.

Stop 10 (493000E; 5152000N) The this band of outcrops represent Unit VI. It is similar in all aspects to the Unit VI at stop 6. The float and minor outcrop between here and Stop 11 are calcareous shale of Unit VII.

Stop 11 (493350E; 5151650N) The arkosic quartzite on top of this butte forms the base of

STOP 12

the Greyson formation in this area. The quartzite pinches out to the north, but is present in upper Deep Creek Canyon 10 miles south, and in Lion Gulch 20 miles east. In lower Deep Creek Canyon, and to some extent in Lion Gulch, diamictite with clasts of Newland and crystalline basement rock also occurs at this stratigraphic position. Above the basal quartzite is typical Greyson shale. Return to the vehicle and continue toward the main road.

Mile 2.0, Stop 12 (493600E; 5153970N) Non-calcareous, fissile, rusty-weathering silty shale of the Greyson formation is visible here in the road bank . Within the shale, tiny silt lenses occasionally show cross lamination, and suggest a quiet sub-wave base shelf environment with occasional storm currents.

This concludes the field tour through the Little Birch Creek section. Continue back to the Duck Creek road and onward to your next destination.

References

Awramik, S., Kidder, D., and Zieg, G.A., 1993, *Collenia undosa*, Charles Walcott, and lacustrine stromatolites, Geological Society of America, 1993 annual meeting, Abstracts with Programs - Geological Society of America. 25; 6, Pages 357. 1993.

Bonnet, A.T., 1979, Lithostratigraphy and depositional setting of the limestone-rich interval of the Lahood Formation (Belt Supergroup), southwestern Montana.

Boyce, R., 1975, Depositional systems in the LaHood Formation, Belt Supergroup, southwest Montana: Ph.D. dissertation, University of Texas, Austin, 247 p.

Eby, D. E, 1977, Sedimentation and early diagenesis within eastern portions of the 'middle Belt carbonate interval' (Helena Formation), Belt Supergroup (Precambrian Y), western Montana, Ph.D. dissertation: State University of New York at Stoney Brook, N.Y., 702 p.

Fenton, C.R., and Fenton, M.A., 1936, Walcott's "Pre-Cambrian Algonkian algal flora" and associated animals: Geol. Soc. America, Bull., v. 47, p. 609-620.

Gierke, W.G., 1987, Structural geology and geothermal investigation of the White Sulphur Springs Area, Montana, M.S. Thesis, Western Michigan University.

Gogas, J.G., 1984, A gravity investigation of the Smith River basin and White Sulphur Springs area, Meagher County, Montana, Masters thesis, Montana College of Mineral Science and Technology, Butte.

Gutstadt, A.M., 1975, Pseudo- and dubiofossils from the Newland Limestone (Belt Supergroup, Late Precambrain), Montana: Jour. Sed. Petrology, v. 45, no. 2, p. 405-414.

McMannis, W.J., 1963, LaHood Formation 0 A coarse facies of the Belt series in southwestern Montana: Geol. Soc. American Bull., v. 74, p. 407-436.

Trunk, F. J. and Smith, D.L., 1979, Soft sediment deformation of laminated algal carbonates of the LaHood Formation (Precambrian Y), southwestern Montana: Geological Society of America, Abstracts with Programs, v. 11, no. 6 p. 304.

Verrall, P. 1955, Geology of the Horseshoe Hills area, Montana: Ph.D. dissertation, Princeton University.

Walcott, C.D., 1914, Pre-Cambrian Algonkian Algal Flora: Smithsonian Institution Miscellaneous Collections, v. 64, p. 77-156.

Zieg, G.A., 1981, Stratigraphy, sedimentology and diagenesis of the Precambrian upper Newland Limestone, central Montana: M.S. thesis, University of Montana, Missoula, 182 p.

Zieg, G.A., 1986, Stratigraphy and sedimentology of the Middle Proterozoic upper Newland Limestone: in S. Roberts, (ed.), Belt Supergroup, A Guide to the Proterozoic Rocks of Western Montana and Adjacent Areas: Montana Bureau of Mines and Geology Special Publication 94, p. 125-141.

A field tour through the Proterozoic rocks of the southern Little Belt Mountains

G. A. Zieg, D. L. Feeback, J. J. Vogl, D. A. Foster, P. A. Mueller, and D. W. Mogk

Senior Geologist, Teck Cominco American Inc., East 15918 Euclid, Spokane, WA 99216

INTRODUCTION

This trip examines the Belt Supergroup strata including Ravalli group and lower Belt as well as underlying rocks of the southern Little Belt Mountains. The trip includes are review of the Sheep Creek copper-cobalt-zinc-lead-iron prospect and describes the dynamic interplay of Proterozoic tectonics, sedimentation and hydrothermal activity that resulted in these extensive sulfide deposits. The trip also includes a unique opportunity to visit the base of the Belt, and ends in the crystalline basement complex in the vicinity of the Big Ben molybdenite deposit and the Neihart silver camp.

ROADLOG

Mile 0.0 — White Sulphur Springs. From the main intersection in town where northbound Highway 12/89 turns east, head west on the Fort Logan road. Stay on the blacktop. This is the approximate route of the Carroll Trail, the freight road from the mouth of the Musselshell River to Diamond City and beyond. Prior to the advent of the railroad, steamboats ascended the Missouri River with suppliers for the placer miners in the gold rich gulches of southwest Montana. Most of the year, they were able to reach Fort Benton. However, the Carroll Trail received heavy use in the spring, when the Missouri was still icebound between the Musselshell and Fort Benton. In 1868, James Brewer opened a store and roadhouse to service travelers on the trail at Trinity Springs, about a mile west of the present town site of White Sulphur Springs. By 1873, a community had grown up around the hotsprings at the present townsite and the U.S. postal service established a post office here.

To the north of the road and across the north Fork of the Smith River, Newland Formation of the lower Belt Supergroup has been carried northward over Paleozoic sediments as young as Permian Quadrant quartzite along the Willow Creek thrust fault (Phelps, 1969). These are the youngest Paleozoic exposures in the Smith River drainage basin. The Newland Formation here consists of the upper Newland (carbonate and shale) and the upper part of lower Newland (calcareous shale). Greyson shale caps the Newland and is exposed on the north and east margins of the town site. Cominco American core drilling one mile north of town has shown that a normal fault downdrops the Belt section along the north margin of the North Fork valley by at least 800 feet. Southwest of here, normal faulting downdrops the valley floor about 1,100 feet (Gierke, 1987).

Mile 5.63 (500780E; 5158810N) — Turn right onto the Newlan Creek Road and head north. Note that for this road log, UTM coordinates use a North American Datum 1927, Zone 12 projection.

Mile 8.70 (501750E; 5163280N) — Our route up Newlan Creek takes us down through the Paleozoic section which outcrops on either side of the valley.

STOP 1

Mile 10.90 (504790E; 5164590N) **Stop 1** — Resistant outcrops of middle Cambrian Flathead sandstone form a narrow spot in the Newlan Creek drainage, chosen for construction of a reservoir. The Flathead rests in angular unconformity over a section of Spokane shale, which represents the Ravalli group in the Helena embayment. Spokane red and green argillite contains raindrop imprints, salt casts, and mud cracks which record its subaerial exposure. In the Smith River area, the Spokane also contains stromatolites of the genus Collenia (Walcott, 1914). During a snowy day in August about 10 years ago Stan Awramiks, Dave Kidder, and I relocated the original sites for Collenia on Battle Creek, about 15 miles south of White Sulphur Springs, and near Keep Cool Reservoir, about 20 miles west of White Sulphur Springs. The productive horizon lies about 700 feet above the base of the Spokane Formation. No Collenia have been identified in the Spokane section at Newlan Creek Reservoir.

STOP 2

Mile 13.37 (507710E; 5166950N) **Stop 2** — The Greyson Formation consists mainly of dark gray, rusty weathering silty shale. At this stop, arkosic sand interrupts the shale. The contact between Spokane shale and Greyson shale represents the boundary between reduced sediments and oxidized sediments that further west separates the Ravalli Group rocks from lower Belt Rocks. Small amounts of pyrite are preserved in the Greyson shale, and its oxidation upon weathering contributes to the rusty color.

Mile 14.70 (509410E; 5168190N) - This is the junction of Highway 89 and the Newlan Creek road. Turn left (north).

STOP 3

Mile 15.45 (509520E; 5169450N) **Stop 3** — This is an excellent exposure of Unit VII (Zieg, 1981) of the upper Newland. Here, Unit VII consists of medium beds of silty limestone. In places, lenses of cross-laminated silt are preserved in the carbonate beds. Note that some beds pinch out laterally, though the primary silt laminations seem to carry through. This suggests pressure solution of these carbonate beds resulted in only preservation of some of them. Most dark shaley material between beds consists of insoluble residue. Up section, carbonate beds give way to silty calcareous shale, which in turn grades into noncalcareous silty shale of the lower Greyson formation.

Mile 18.0 (510820E; 5173130N) - At this junction, turn left from Highway 89 onto the gravel U.S.F.S. road and proceed up Miller Gulch. This location is called the Miller Gulch cow camp. Here, we are very close to the base of the Newland formation. Across the highway to the northeast and up the Newlan Creek drainage are exposures of Chamberlain formation against the Volcano Valley thrust fault. Northwest of the cow camp in an area called Mumbro Park are exposures of Chamberlain formation. In these areas, the transition from tan-weathering lower Newland shale to black silty Chamberlain shale contains multiple thin silicified debris flows. Some of these are exposed on the hillside behind the cow camp. About two miles northwest of this area, uppermost Chamberlain silty shale stratigraphy contains debris flows containing olistoliths of molar tooth bearing dolomite (Godlewski and Zieg, 1984). Above this, at the Chamberlain-Newland contact, are a few iron oxide gossan zones. One mile further north along Sheep Creek, drilling shows that copper-rich stratabound sulfides in what we locally call the 'lower sulfide zone' (LSZ) occupy this stratigraphy (Zieg and Leitch, 1993). Also at Sheep Creek, capping the sulfide zone is a thin-bedded limestone unit about 100 feet thick which is nowhere exposed on surface. Below the sulfide zone, in the upper Chamberlain, are scattered beds of molar tooth bearing dolomite. We believe that these beds correlate with molar tooth bearing dolomite exposed above the Chamberlain shale on Belt Creek and Chamberlain Creek. There are no carbonate units in the lower Newland or upper Chamberlain exposed in this area.

STOP 4

Mile 19.50 (510030E; 5174650N) **Stop 4** — This is typical calcareous shale of the lower Newland formation. Then entire lower Newland section is preserved in a traverse from the upper reaches of Newlan Creek, about 4 miles east of here, and up our route to the upper part of Miller Gulch. We interpret the lower Newland as basinal shale, deposited in a reducing, subwave base depositional environment. Studies of lower Newland layering by Feeback (1997) and by Cominco American geologists show that they are distal microturbidites. Turn around and head back to the highway. At the junction with Highway 89, turn left (north).

Mile 22.20 (511380E; 5175030N) — On the left is a quarry. If you turn in, be very careful because oncoming traffic has poor visibility on this curve in the highway. This quarry is in lowermost Newland formation noncalcareous shale and is within 200 stratigraphic feet of the Newland-Chamberlain contact. The shale contains abundant small soft sediment folds.

Mile 24.00 (510370E; 5177690N) — This location is our best approximation of the trace of the Black Butte fault (BBF). This is a reverse fault that carries Chamberlain and lower Newland northward over Upper Newland carbonates, which amounts to at least 2500 feet of displacement. From this vantage point, one can look west to Black Butte. Note the notch in the top of Black Butte; this is the approximate intersection of the BBF with the Volcano Valley Fault (VVF), which trends northeast at Black Butte. The BBF then traces straight from the notch to this point, while the VVF swings northeast, turns east, then turns southeast across the highway and again intersects the BBF to our east. At this point, the VVF continues east up the Newlan Creek and on to the upper Musselshell drainage where it dies in an anticline. The combined displacements of the BBF and the central segment of the VVF (where it lies north of the BBF) approximate the amount of displacement on those segments of the VVF east and west of the ends of the BBF. Thus, the Black Butte fault appears to handle the amount of displacement that the northerly arc of the VVF can't accommodate during thrusting.

This stop is on the drainage divide between Newlan Creek, which flows south, and Sheep Creek, which here flows west. In the basin north of us, Sheep Creek enters from the east, flowing on or near bedrock. In the basin center, drilling shows that over 300 feet of gravels capped by lacustrine sediments fill the basin. As Sheep Creek exits the basin on its west margin, it again cuts through bedrock. During its course of flow toward this basin from the east, Sheep Creek cuts through a young basalt flow. On the divide where we're standing, basalt clasts are apparent in the gravels. We hypothesize that Sheep Creek once flowed south from this basin into the present day Newlan Creek drainage. Subsidence of the basinal area in front of us, possibly during Miocene extension, blocked the path of Sheep Creek and it cut its way west through the present day Sheep Creek Canyon. Eventually Sheep Creek abandoned its former drainage to Newlan Creek, formerly a tributary and now an underfit stream.

STOP 5

Mile 25.00 (510300E; 5179250N) **Stop 5** — Turn left here onto the Moose Creek road. At this point the VVF crosses the highway and trends southeast toward upper Newlan Creek. Here, upper Newland Formation is carried over Paleozoic rocks. The exposures on the corner show Flathead sandstone, Wolsey shale, and Meagher limestone overturned beneath the sole of the thrust. To the north, the Paleozoic section rests on unexposed lower Newland shale. If that statement made you uncomfortable, it should have; we've just described a younger over older relationship (upper Newland over lower Newland) across the same fault that shows reverse motion (Belt over Paleozoic). Data collected from drilling in this area shows net normal displacement between Newland sections. The amount of dip slip movement was considerably greater than the amount of reverse motion displacing the Paleozoic section. We believe that the simplest explanation for this is that the dip slip movement must have occurred before Paleozoic sedimentation, and subsequent reverse movement wasn't great enough to cancel out the displacement. One mile north of here along Sheep Creek, unfaulted Flathead sandstone conceals an east-west structure we call the 'buttress' fault. This fault drops Newland on the south against lower Chamberlain shale and underlying Neihart quartzite on the north. We have traced this fault in the subsurface for about five miles east along the Sheep Creek drainage where its throw must reach 2000 feet. Along its trace to the west, the VVF cuts the buttress fault. West of Black Butte, the VVF and the buttress fault merge and are indistinguishable.

Extensive mapping and drilling of the lower Newland during mineral exploration work in the area by Cominco American, BHP, Anaconda, Exxon, and Kennecott provided excellent documentation of the internal stratigraphy of the Newland Formation. This work shows that near the buttress fault there is an abundance of thick debris flow units of black silty shale clasts, typical of the underlying Chamberlain Formation. Microturbidite laminations show an abrupt thickening into debris flows coincides with proximity to the buttress fault and VVF. West and east, some areas show broad calcarenite fans and olistoliths of stromatolitic limestone within the lower Newland, also focused against the Volcano Valley fault (Zieg and Godlewski, 1986). The evidence strongly suggests a shelf to basin margin roughly coincident with the Volcano Valley Fault and buttress fault. Changes in lithology of clasts suggest a deeper water shelf north of the Black Butte area, and a shallower water shelf both east and west. This interpreted shelf margin also coincides with the appearance of molar-tooth bearing dolomite in the upper part of the Chamberlain shale, as shown by drilling the Newland-Chamberlain contact along the Sheep Creek valley.

STOP 6

Mile 26.70 (507700E; 5179940N) **Stop 6** — This stop provides a view of a gossan exposure from weathering of lower Newland formation bedded sulfide. This is the 'upper sulfide horizon' (USZ) whose top occurs about 400 feet below the top of the upper Newland formation. The mineralization in this zone is stratabound and stratiform and usually is less than 100 feet thick, though it can persist over great thicknesses of stratigraphy. In one nearby area, the USZ consisted of 25% sulfide over 1000 stratigraphic feet. Though the zone consists mainly of framboidal pyrite, both chalcopyrite and cobaltiferous pyrite are concentrated toward the base of the zone. In this area, known as the Strawberry Butte area after the small butte northwest of us, Cominco American reported a resource of 5 million tonnes grading 2.5% Cu and 0.12% Co near here. A considerably larger quantity averages about 2% Cu. Also in this area, they reported 4 million tonnes grading 4% Cu in the LSZ at the Newland-Chamberlain contact (Zieg, et.al., 1989). This resource lies at depths greater than 1300 feet below surface. The LSZ is known only between the VVF and the buttress fault, and is truncated by both. The net normal displacement along the VVF shortens the distance between the USZ in its hangingwall and the LSZ in its footwall by about 700 feet. Distal areas of the USZ to the south contain Zn-Pb concentrations.

Much of the USZ gossan shows some silicification, in places pervasive. At this location, the USZ contains abundant barite, which is typically concentrated above the Cu-Co rich zone. It occurs as large blades and masses in fine grained massive pyrite. Here, it also occurs in a peculiar texture in which blebs of barite appear set in an FeOx (formerly sulfide) matrix. In a shaly horizon about five feet above the gossan in this outcrop is a small debris flow horizon with mounds of intergrown, silicic hollow tubes (the outcrop is picked over and specimens are now hard to find). The cross sectional texture between these and the barite-sulfide masses are identical. These textures are interpreted as preservation of a hydrothermal vent fauna, and probably formed as algal filaments were surrounded by a crust of aragonite (?) (J. Farmer, pers. comm.) which was later replaced by sulfide within the sulfide horizon, and silica in the areas above the horizon. Within the sulfide zone, barite, calcite, dolomite or microcrystalline quartz gangue filled the tubes as sulfide replaced the walls. There is a strong correlation between this texture and the areas of the most extensive and richest sulfide development. These could represent one of the oldest and best preserved hydrothermal vent communities documented on the planet. This, plus middle Proterozoic Pb dates (1340 m.y.) on the mineralization, abundant synsedimentary textures involving sulfide, and S isotope values consistent with Proterozoic seawater sulfate, provide strong evidence that the bedded pyrite is synsedimentary in origin, and overprinted with Cu-Co mineralization, apparently diagenetic based on texture and its similar post-depositional history. The VVF cuts the sulfide zones.

The LSZ, USZ, and an additional less well developed sulfide zone in the basal upper Newland (Unit II shale zone) appear to represent discrete and separate events of seafloor hydrothermal venting. Each of these involved formation of syngenetic pyrite framboid mud followed by burial and a diagenetic overprint of barite and pyrite followed by silicification and Cu (LSZ) or Cu-Co (USZ) mineralization. Near vent sites, the persistence of bedded pyrite throughout the lower Newland section show that at least low levels of hydrothermal activity were nearly constant. The USZ and LSZ represent major hydrothermal events. Extensive debris flows lie immediately footwall to the USZ and LSZ, and provide evidence that these hydrothermal events accompanied basin subsidence along synsedimentary Proterozoic faults.

The USZ is downdropped east of us by a northeast trending fault, which forms the west margin to the Tertiary basin described from two stops ago (Newland Creek - Sheep Creek divide). In this area, both the USZ and LSZ are well mineralized, and Drilling in this area defined another northeast structure which, because if its great vertical extent of bedded sulfide, silicification, hydrothermal brecciation, and Cu-Co mineralization, must have acted as a feeder structure for mineralization. Thicknesses of sulfide zones are radically different on either side of this structure as well. Both the USZ and LSZ are generally well mineralized in this area.

The intersection of the Proterozoic buttress fault and a broad zone of northeast trending faults (the ancestral Great Falls tectonic lineament) localized mineralization. This Proterozoic extensional faulting resulted in a shale basin against a 'buttress' of crystalline basement and quartzite. After some additional post-mineral normal faulting, this was capped by Paleozoic sediments, but the juxtaposition of shale faulted against more massive rocks controlled the location of the later Volcano Valley thrust/ reverse fault.

STOP 7

Mile 27.16 (507090E; 5180340N) — Go straight here. The turn off to the left leads to Butte Creek, Iron Butte, Copper Creek, and Horse Prairie which all contain extensive exposures of gossan after bedded sulfide in the lower Newland Formation. These exposures extend over a strike length of nearly 10 miles.

Mile 28.20 (506790E; 5181830N) **Stop 7** — Road cut exposures at this stop along Sheep Creek are Neihart Quartzite. On the hill west of the road, the Volcano Valley fault carries Upper Newland carbonate over middle Cambrian Flathead quartzite. 1000 feet west of here, drilling through the Flathead quartzite found it rests on Chamberlain shale and Neihart quartzite. Neihart quartzite exposures north of the Volcano Valley Fault persist over 10 miles west beyond the Smith River.

STOP 8

Mile 28.96 (507070E; 5182880N) — Take the left hand fork onto the Sheep Creek road.

Mile 29.45 (506330E; 5183150N) **Stop 8** — These are exposures of the granitic basement rock which underlies the Neihart Quartzite. Some slices of this rock are caught along the Volcano Valley fault west of Black Butte, between Newland Formation on the south and Cambrian sediments on the north. A wedge of this rock is also caught between areas of Newland Formation along the west end of the Black Butte fault. At this writing, the granite remains undated.

STOP 9

The granite and amphibolite are part of the Sheep Creek Intrusive Complex (Vogl et al., this volume). The Sheep Creek intrusive complex is variably exposed throughout the southern outcrops of basement in the Little Belt Mountains. This complex comprises weakly foliated, shallow-dipping, leucogranite sheets that cut across strongly foliated intrusive units that range from granite to amphibolite. The leucogranite sheets contain only minor amounts of biotite, and locally contain garnet and secondary (?) muscovite. The older (cross-cut) units are limited to outcrop-scale blocks and lenses. A strongly foliated amphibolite from the complex yields a concordant U-Pb age of ~1810 Ma, which is interpreted as an emplacement age (Vogl et al., this volume). Within the outcrop the amphibolite is cut by leucogranite, thus, 1810 Ma is considered a maximum age for the leucogranites. Overall, the mild deformation and age constraints suggest that the Sheep Creek leucogranites were intruded late in the magmatic and orogenic history of the Little Belt Mountains basement. The leucogranites are similar in composition to many of the leucogranite veins that are numerous in the southern and western parts of the northern exposure, but it is unknown if they were intruded during the same event.

STOP 10

Mile 31.60 (508380E; 5184990N) **Stop 9** — This is an exposure of amphibolitic basement rock, typical of much of the crystalline basement exposed in the Little Belt Mountains. It is intruded by the granite of the last stop.

Mile 36.60 (513560E; 5190330N) — This is the approximate location of the Wet Creek fault, which juxtaposes crystalline basement rocks against Chamberlain shale. This fault continues west through the head of Coyote Creek and forms a southern boundary to exposures of both Neihart Quartzite and Chamberlain shale.

Mile 38.10 (515780E; 5191120N) **Stop 10** — These are exposures of dark silty, micaceous shale typical of the Chamberlain formation, and yellowish or brownish tan- weathering uneven beds of dolomite. Published mapping (Keefer, 1972) has generally mapped this as Newland Formation. Drilling from lower Newland Formation shale through to Chamberlain shale along sheep Creek has encountered this same carbonate type. This carbonate type is nowhere present in the upper Newland Formation. Because it is always interbedded with silty, micaceous, Chamberlain type shale, we have informally called it 'upper Chamberlain carbonate' to distinguish it from the more evenly bedded, limey fine grained mudrocks of the Newland. South of Sheep Creek, this type of carbonate disappears from the upper Chamberlain.

STOP 11

Mile 41.20 (516010E; 5193920N) — Rocking Chair Park; this area is typical of the Moose Creek-Belt Creek divide area where erosional remnants of Middle Cambrian Flathead quartzite, Wolsey shale, and Meagher limestone cap the Proterozoic rocks. Some exposures of mid-Tertiary 'andesite porphyry' sills (McClernan, 1969) intrude the Cambrian section. At the T-junction, turn right.

Mile 48.36 (523810E; 5188400N) — This is the junction with U.S. Highway 89. Turn left (north) and drive down the Belt Creek drainage. The route will pass down section through middle Cambrian Wolsey shale and Flathead quartzite to the unconformity above the Proterozoic. The next stop will be just below the unconformable contact between the Proterozoic and overlying Flathead quartzite.

STOP 12

Mile 50.85 (524700E; 5190980N) **Stop 11** — Here, the roadcuts show exposures of both the 'upper Chamberlain' dolomite examined in the last stop, and overlying calcareous shale identical to lower Newland Formation shale. To my knowledge, this is the only surface exposure of the upper Chamberlain-lower Newland contact in the area north of the Volcano Valley fault. Note that the upper Chamberlain dolomite is somewhat thicker bedded than in Moose Creek. Interbedded shales are silty and micaceous. Some molar tooth structure is present in the dolomite. Similar exposures along Chamberlain Creek north of here contain matrix supported debris flows with very angular, long and skinny laminated limestone clasts. This carbonate has little in common with the upper Newland carbonate south of the Volcano Valley fault.

STOP 13

Mile 51.87 (523980E; 5192430N) **Stop 12** — These silty and sandy shale exposures are typical of Chamberlain Shale. They contain channels with silty and sandy lenses and a quartz pebble conglomerate. Note the carbonaceous films on the shale surfaces; some if these are round and Horodyski (1980) describes flattened organic-walled filamentous and spheroidal envelopes from the Chamberlain shale at this approximate stratigraphic level from near the confluence of Chamberlain and Jefferson Creeks, 1.5 miles northeast of here.

STOP 14

Mile 53.36 (523670E; 5194770N) — Turn right from the highway onto the Jefferson Creek road and proceed upstream.

Mile 53.67 (524110E; 5194760N) **Stop 13** — This quarry exposes the top of the Neihart Quartzite, the basal Belt Supergroup unit in this part of the Belt Basin. Tabular beds of fine-grained Neihart Quartzite become interbedded with the black micaceous Chamberlain Shale as one traverses upstream from the quarry. Low angle cross beds with broad shallow channels are apparent in these exposures. Further up Jefferson Creek are excellent exposures of the carbonate-rich section which rests above the silty and sandy Chamberlain shale. Return to the highway and turn right (downstream).

STOP 15

Mile 54.56 (522970E; 5195310N) **Stop 14** — The massive Neihart quartzite here at the Devils Chair shows little internal texture. The Neihart is approximately 800 feet thick in this area. Toward its base, Winston (1989) reports ventifacts and interprets lower Neihart as eolian sands. Keefer (1972) interprets the Neihart as fluvial sandstone.

Mile 56.00 (521030E; 5196420N) — These granitic gneiss exposures represent the basement rock upon which the Neihart quartzite rests. This stop is at the east edge of Neihart, a late 19th century mining camp grown up around a swarm of high grade silver veins that cut through Belt Creek canyon in this area. Though the silver market crashed in the mid-1890's, Neihart survived as a resort community and intermittent mining town.

Mile 56.90 (520050E; 5197470N) **Stop 15 —** Park behind Bob's Bar near the north end of the town of Neihart. Walk across the bridge over Belt Creek and turn left along the creek. The Neihart mylonite is exposed in outcrops along the edge of Belt Creek. Mylonitization in this shear zone has affected a range of lithologies including Augen gneiss, amphibolites, pegmatite, and granitic rocks. The mylonite has a strongly banded character due to intense transposition of leucogranite veins within more mafic meta-igneous compositions (Vogl et al., this volume). Outcrops display a flaggy appearance in the zones of highest strain. Mylonitization occurred at amphibolite-facies conditions, but continued to greenschist-facies conditions, as evidenced by the replacement of hornblende and biotite by actinolite, epidote, and chlorite;

STOP 16

STOP 17

feldspars are also partially replaced by white micas. The strike of the mylonitic foliation ranges from WNW to WSW with moderate to steep southward dips. Mylonites are locally folded on the outcrop scale. Lineations are generally shallowly plunging with variable WSW to WNW trends. Shear-sense indicators such as C-S fabrics, porphyroclast tails, and quartz grain-shape fabrics indicate a sinistral shear sense, consistent with shear bands in the Gray gneiss to the north.

Mile 57.62 (519520E; 5198510N) **Stop 16** — Medium-grained, homogeneously well- foliated and locally well-lineated granodiotic rock known as the Gray gneiss is exposed around the Florence Mine. Finer grained and slightly more mafic rocks, as well as amphibolite, are commonly found near the margins of this intrusive unit. Mueller et al. (2002) report a U-Pb zircon date of 1867 ± 6 Ma for the Gray gneiss.

Mile 58.45 (518840E; 5199560N) **Stop 17** — Here at the confluence of Carpenter Creek and Belt Creek are exposures of the Pinto diorite. The Pinto diorite intruded into older gneiss around 1.9 Ga (Catanzaro and Kulp, 1964). Mueller et al., (2002) report a ^{207}Pb/^{206}Pb age of 1864±5 Ma (2_) for the Pinto diorite. Olmore (1991) describes the basement complex in this area as, from oldest to youngest, older gneiss, Pinto diorite, amphibolite, mylonite schist, granitic gneiss, granite, pegmatite, and diorite dikes. The older gneiss may be Archean in age. The Pinto diorite is one of the most notable basement units in the LBM, appearing on nearly every geologic map of the area. Mineral assemblages in the Pinto diorite include Pl+Hbl+Bt+Qtz+Kfs displaying polygonal recrystallized textures. One of the striking features of the Pinto is the high abundance of gray to mint-green plagioclase megacrysts up to 3 cm across. The megacrysts are locally well aligned producing a foliation or lineation. This foliation, along with dikes (diorite to granodiorite), are locally isoclinally folded where discrete shear zones cut the diorite (Vogl et al., this volume).

STOP 18

The entire assemblage is intruded by the Eocene age Big Ben molybdenum porphyry complex. The silver veins exploited at Neihart are peripheral to the porphyry center. AMAX drilled out the Big Ben deposit in the 1970's, and according to Olmore (1991) the 'probable reserve' is 120 million tons grading 0.16 MoS2.

STOP 19

Mile 60.00 (518360E; 5201950N) **Stop 18** — The stop is north of Neihart on the east side of US 89 just after a large bridge over Belt Creek. Turn right into Neihart cemetery and park. Although the Paleoproterozoic intrusive rocks make up the majority of the Pre-Belt basement exposed in the northern Little Belt Mountains, other lithologic packages are exposed. The most prominent of these occurs in intrusive contact with the northern margin of the Pinto Diorite. These rocks comprise a compositionally variable layered migmatite sequence. Most compositions contain assemblages with two feldspars, variable amounts of clinopyroxene, biotite, hornblende, and quartz. The rocks display meter-scale lithologic variations. We interpret these features as an indication of a meta-volcanic origin for this unit, which we refer to as the Cemetery migmatite (Vogl et al., this volume). A leuogranite pod from the migmatite yielded an U-Pb zircon upper intercept age of 1817±17 Ma and probably represents the time of migmatization and melt injection.

Mile 61.56 (517650E; 5204270N) **Stop 19** — Pull off into the north entrance of the USFS Aspen campground and park. Walk to north end of campground and 50 feet or so into bushes to an outcrop and talus pile of pelitic gneiss and granite. We refer to this migmatitic pelitic paragneiss unit as the Aspen paragneiss (Vogl et al., this volume). This unit occurs as a distinct mappable band between the meta-volcanic rocks and the Ranger diorite (see map in Vogl et al., this volume). Notable minerals in the pelitic gneisses include garnet (up to 3 cm), sillimanite, spinel, K-feldspar, cordierite, and anti- perthite, which indicate very high temperatures of metamorphism.

From the north end of the Aspen campground, continue walking on west side of highway 89 to exposures of coarse grained diorite, know as the Ranger Diorite. This coarse- grained, moderately to strongly foliated diorite is another of the suite of calc-alkaline, subduction-related igneous intrusions in the northern Little Belt Mountains. Typical mineral assemblages include Pl+Kfs+Cpx+Bt±Hbl with minor quartz. The Ranger diorite is associated with micro-diorite and leucogranite.

This is the end of our trip through the Proterozoic rocks of the Little Belt Mountains. The quickest return route to White Sulphur Springs is to drive back through Neihart and continue south on highway 89. By continuing north on the highway, you can pass through spectacular exposures of the Paleozoic section in the vicinity of Monarch, and continue north into Cretaceous exposures further into the foothills of the north slope of the Little Belt Mountains. Eventually, the this road intersects the Lewistown-Great Falls highway.

References

Catanzaro, E.J. and and Kulp, J.I., 1964, Discordant zircons from the Little Belt (Montana), Beartooth (Montana), and Santa Catalina (Arizona) mountains: Geochimica et Cosmochimica Acta, v. 28, p. 87-124.

Feeback, D. L. 1997, Stratigraphy and sedimentation of the Chamberlain Formation -Newland Formation boundary, lower Belt Supergroup (Middle Proterozoic), central Montana, M.S. thesis, University of Montana.

Gierke, W.G., 1987, Structural geology and geothermal investigation of the White Sulphur Springs Area, Montana, M.S. Thesis, Western Michigan University.

Godlewski, D.W., and Zieg, G.A., 1984, S. Hobbs (ed.), Belt Symposium II, The Belt, Absstracts with Summaries, Montana Bureau of Mines and Geology Special Publication 90, p. 2-4.

Horodyski, 1989, Paleontology of the Middle Proterozoic Belt Supergroup, in Middle Proterozoic Belt Supergroup, Western Montana, Field Trip Guidebook T334, 28th International Geological Congress.

Keefer, W.R., 1972, Geologic map of the west half of the Neihart quadrangle, Montana: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-726.

McClernan, H. G., 1969, Geology of the Sheep Creek area, Meagher County, Montana: M.S. thesis, Montana College of Mineral Science and Technology.

Mueller, P., Heatherington, A., Kelley, D., Wooden, J., and Mogk, D. (2002) Paleoproterozoic crust within the Great Falls Tectonic Zone: Implications for the assembly of southern Laurentia. Geology. v. 30, p. 127-130.

Olmore, S.D., 1991, Geologic summary of the Big Ben molybdenite deposit near Neihart, Montana, in D. Baker and R. Berg (eds), Central Montana Alkalic Province: Montana Bureau of Mines and Geology Special Publication 100, pgs 124-127.

Phelps, G.B., 1969, Geology of the Newlan Creek area, Meagher County, Montana, M.S. thesis, Montana College of Mineral Science and Technology.

Vogl, J. J., Foster, D. A., Mueller, P. A., Wooden, J. L., and Mogk, D. W., 2004, Lithology and age of pre-Belt Precambrian basement in the Little Belt Mountains, Montana: implications for the role of the Great Falls Tectonic Zone in the Paleoproterozoic assembly of North America, this volume.

Walcott, C.D., 1914, Pre-Cambrian Algonkian Algal Flora: Smithsonian Institution Miscellaneous Collections, v. 64, p. 77-156.

Winston, D., Whipple, J.W., and Horodyski, R. J.,1989, Field Trip Road Log, in Middle Proterozoic Belt Supergroup, Western Montana, Field Trip Guidebook T334, 28th International Geological Congress.

Zieg, G.A., 1981, Stratigraphy, sedimentology and diagenesis of the Precambrian upper Newland Limestone, central Montana: M.S. thesis, University of Montana, Missoula, 182 p.

Zieg, G.A., 1986, Stratigraphy and sedimentology of the Middle Proterozoic upper Newland Limestone: in S. Roberts, (ed.), Belt Supergroup, A Guide to the Proterozoic Rocks of Western Montana and Adjacent Areas: Montana Bureau of Mines and Geology Special Publication 94, p. 125-141.

Zieg, G.A. and Godlewski, D.W., 1986, Road Log No. 1, a traverse across the eastern Belt basin from Neihart to Townsend Montana, in S. Roberts, (ed.), Belt Supergroup, A Guide to the Proterozoic Rocks of Western Montana and Adjacent Areas: Montana Bureau of Mines and Geology Special Publication 94, p. 1-16.

Zieg, G.A. and Leitch, C.H.B., 1993, The geology of the Sheep Creek copper deposits, Meagher County, Montana, R. Berg (compiler), Belt Symposium III abstracts, Montana Bureau of Mines and Geology Open-File Report 381, pg. 67-69.

NORTHWEST GEOLOGY

The Journal of The Tobacco Root Geological Society

Volume 33, 2004

White Sulphur Springs Field Conference

Table of Contents