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Geologic Map of Precambrian Metasedimentary Rocks of the Medicine Bow Mountains, Albany and Carbon Counties, Wyoming

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INTRODUCTION

The Medicine Bow Mountains of Wyoming are 30 mi west of Laramie in southeastern Wyoming. The mountain area is a complex of Precambrian rocks that is the core of a large asymmetric anticline bounded by west-dipping thrusts on the east flank. It is bordered on the east by the north-trending Laramie basin, which contains sedimentary rocks ranging in age from Paleozoic to Holocene, and on the west by the northwest-trending Saratoga Valley, a graben that contains sedimentary rocks of Tertiary age.

Precambrian rocks of the Medicine Bow Mountains are separated into two segments by a major northeast- to east-striking fault system, the Cheyenne Belt of Houston and others (1979). The northernmost fault of the Cheyenne Belt is the Mullen Creek-Nash Fork shear zone; north of this shear zone a thick succession (52,000 ft thick) of Archean and Proterozoic metasedimentary and metavolcanic rocks overlies Archean basement rocks. South of the shear zone no rocks of Archean age have been recognized, and the area consists of metasedimentary and metavolcanic rocks cut by numerous mafic and felsic intrusions, all of Proterozoic age.

The Sierra Madre is the northern extension into Wyoming of the Park Range of Colorado, and its eastern border in Wyoming is only about 600 ft west of the westernmost exposure of Precambrian rocks of the Medicine Bow Mountains. The Precambrian geology of the Sierra Madre is very similar to the Medicine Bow Mountains and rock relationships and geochronology established in the Sierra Madre were used to aid in interpreting similar units of the Medicine Bow Mountains and vice versa.

This pamphlet has been prepared for the reader unfamiliar with the geology of this area and is designed to update the stratigraphy and clarify problems of age relationships of this succession. The following discussion relies heavily on previous reports (Houston and others, 1968; Hills and Houston, 1979; Karlstrom and others, 1981a, 1983; Karlstrom and Houston, 1984).

The reader will have a better understanding of the contents if he keeps in mind the plate tectonic models for this area as proposed by Hills and Houston (1979) and Karlstrom and Houston (1984). In essence, the Early Proterozoic succession is regarded as a rifted Atlantic-type trailing margin that collided with island arcs in a manner similar to the Taconic history of the Appalachians.

HISTORY OF INVESTIGATIONS

Geologic investigations began with the early territorial surveys of the United States government; a summary of this early work is in Van Hise and Leith (1909, p. 833–849). Investigations between 1900 and 1968 are summarized in Houston and others (1968) and include a stratigraphic study by Blackwelder (1926) and regional mapping by students and staff of the University of Wyoming, which was sponsored by the Geological Survey of Wyoming. The U.S. Geological Survey and U.S. Bureau of Mines conducted field studies in the Snowy Range Wilderness of the central Medicine Bow Mountains in 1976–78. The wilderness and parts of the Medicine Bow Mountains to the west and north were recognized as possible areas for uranium- and gold-bearing conglomerate of the Blind River type (Miller and others, 1977; Houston and others, 1977). For this reason, it was determined that additional geological and geochemical studies both within and outside of the wilderness area were needed in order to evaluate the mineral resource potential. Additional work was sponsored by the National Uranium Resource Evaluation Program of the U.S. Geological Survey (Houston and others, 1978) and by the U.S. Department of Energy (Houston and Karlstrom, 1980; Karlstrom and others, 1981a, b; Borgman and others, 1981).

STRATIGRAPHY

Metasedimentary and metavolcanic rocks are present in three successions ranging in age from Archean (about 2,900–2,700 Ma) to Early Proterozoic (about 2,500–1,700 Ma). These three successions include, from oldest, the Overland Creek Gneiss, the Phantom Lake Metamorphic Suite, and the Snowy Pass Supergroup.

OVERLAND CREEK GNEISS

The Overland Creek Gneiss, which is exposed in the northeasternmost Medicine Bow Mountains near Arlington, is interpreted as a remnant of an Archean greenstone belt. Major lithologic rock types of the isoclinally folded succession are hornblende gneiss and amphibolite, interpreted as mafic volcanic rocks, and biotite-feldspar-quartz gneiss, interpreted as graywacke. Units having fragmental texture, interpreted as lithic tuff, have been identified in the hornblende gneiss, and graded bedding is preserved locally in the biotite-feldspar-quartz gneiss. Thin layers of quartzite (2–3 ft), marble, and paraconglomerate are present in the biotite-feldspar-quartz gneiss, but these rocks make up a minor part of the succession. The rocks of the Overland Creek Gneiss are amphibolite facies metamorphic grade. The Overland Creek Gneiss of the Medicine Bow Mountains is not dated but is considered Archean by analogy with similar rocks exposed in the northwestern Medicine Bow Mountains and the Sierra Madre.

In the Medicine Bow Mountains, the Overland Creek Gneiss is invaded by a pink gneissic granite that is undated. Pink gneissic granite of at least two ages is present in the Medicine Bow Mountains and Sierra Madre. It includes a pink quartzo-feldspathic orthogneiss exposed in the northeastern Sierra Madre, which has been dated as 2,665±28 Ma and 2,683±5 Ma by the uranium-lead-zircon method (Premo, 1983), and a pink gneissic granite (Baggot Rocks Granite) exposed in the western Medicine Bow Mountains, which is dated as 2,430±5 Ma by the same method (Premo, 1983; W.R. Premo, oral commun., 1984). These gneissic granites do not allow us to distinguish between a Late Archean or earliest Proterozoic age for the Overland Creek Gneiss.

There is, however, additional evidence to support an Archean age for the Overland Creek Gneiss if we correlate it with similar successions in the western Medicine Bow Mountains and Sierra Madre. The Archean basement gneiss of the western Medicine Bow Mountains and eastern Sierra Madre is mostly quartzo-feldspathic gneiss and biotite-feldspar-quartz gneiss of uncertain parentage that has been subject to more than one episode of deformation and has been invaded by at least two and possibly three sets of mafic dikes of tholeiitic composition, which are also deformed (Houston and others, 1968; in press). This basement gneiss has masses of hornblende gneiss, amphibolite, quartzite, paraconglomerate, and marble infolded as keels and irregular-shaped bodies both on a small scale and large scale. These mafic gneisses and metasedimentary rocks are similar to rocks of the Overland Creek Gneiss, and the similarity is particularly striking for the largest masses of these rocks, the Vulcan Mountain Metavolcanics, exposed in the north-central Sierra Madre. The Vulcan Mountain Metavolcanics are intruded by a syntectonic granodiorite, the Spring Lake Granodiorite, dated as 2,700 Ma by the whole-rock Rb-Sr method (Hedge in Houston and others, in press) and as 2,710±12 Ma by the uranium- lead-zircon method (Premo, 1983). In addition, smaller masses of mafic gneiss and metasedimentary rocks of the eastern Sierra Madre are intruded by pink orthogneiss dated as 2,683±5 Ma by Premo (1983). Therefore, by analogy, the Overland Creek Gneiss is considered to be Late Archean.

PHANTOM LAKE METAMORPHIC SUITE

The Phantom Lake Metamorphic Suite is a succession of metasedimentary and metavolcanic rocks having much better preserved textures and structures than the Overland Creek Gneiss, but still so severely deformed and metamorphosed that critical evidence for tops of beds is often missing. Although a tentative stratigraphic succession has been established for the Phantom Lake Metamorphic Suite, the term "suite" has been used to describe the succession rather than the term "group" to emphasize the uncertainty in the stratigraphy. From oldest to youngest, the Phantom Lake Metamorphic Suite consists of the Stud Creek Metavolcaniclastics, the Rock Mountain Conglomerate, the Bow Quartzite, the Colberg Metavolcanics, and the Conical Peak Quartzite.

In the northeastern Medicine Bow Mountains near Arlington, divisions of the Phantom Lake Metamorphic Suite are either in fault contact with the Overland Creek Gneiss or are separated from it by mafic sills that are introduced at the contact. Tops of basal beds of the Phantom Lake Metamorphic Suite indicate that it overlies the Overland Creek Gneiss in this area, and similar structural relationships are noted between the basal beds of the Phantom Lake Metamorphic Suite of the Sierra Madre and the Vulcan Mountain Metavolcanics. In the northwest Sierra Madre, the basal beds of the Phantom Lake Metamorphic Suite are radioactive conglomerate that overlies both Archean basement gneiss and rocks that are probably equivalent to the Vulcan Mountain Metavolcanics and Overland Creek Gneiss. Rocks of the Phantom Lake Metamorphic Suite are therefore considered younger than the Overland Creek Gneiss and Vulcan Mountain Metavolcanics.

STUD CREEK METAVOLCANICLASTICS

The Stud Creek Metavolcaniclastics are in the cores of anticlines exposed about 1 mi southwest of Arlington. Rocks in the cores of the anticlines are tightly appressed and folded so that evidence for tops of beds is lost. The thickness of the Stud Creek is unknown, but a maximum exposed thickness is 1,650 ft. The rocks of the Stud Creek Metavolcaniclastics can be divided into four groups: Pelitic schist (50 percent); amphibole schist, gneiss, and amphibolite (30 percent); quartzite and conglomerate (20 percent); and calcareous rocks (less than 1 percent). Pelitic rocks are predominantly biotite- and muscovite-quartz schist but there is also schist containing garnet, staurolite, chloritoid, or kyanite. The pelitic schist has fragmental textures locally, which suggests that some schist was tuff. Amphibolite and amphibole schist include rock types having the chemical composition of tholeiitic basalt, and some quartz-rich amphibolite and amphibole schist may have been andesite. Locally the amphibolite has well-preserved amygdules. Quartzite includes fine-grained micaceous quartzite, fuchsitic quartzite, granule conglomerate, and quartz-rich schistose paraconglomerate. The conglomerate is radioactive and presumably contains heavy minerals containing thorium and (or) uranium; the quartzite locally has recognizable planar bedding and crossbedding. Calcareous rocks range from impure marble to calcareous pelitic schist. In this and subsequent discussions of the various rock units, we will make a best guess on depositional environment, but, especially in the Phantom Lake Metamorphic Suite, poor preservation of textures and structures limits the reliability of these interpretations. Rocks of the Stud Creek Metavolcaniclastics probably represent depositional environments ranging from fluvial (as suggested by radioactive conglomerate) to marine (carbonate), but the variability of rock types, rapid facies changes, the absence of thick and continuous graywacke, and the absence of pillow basalt suggest subaerial deposition during periods of volcanism.

ROCK MOUNTAIN CONGLOMERATE

The Rock Mountain Conglomerate crops out on Rock Mountain about 3 mi southwest of Arlington where it has a gradational contact with the Stud Creek Metavolcaniclastics. The maximum thickness of the Rock Mountain is 1,300 ft. The unit is predominantly granular to pebbly muscovite quartzite containing paraconglomerate beds ranging from less than 3 ft to 500 ft thick. The paraconglomerate has a matrix that is chiefly muscovite and quartz, with lesser amounts of feldspar (mostly potash feldspar) and, locally, garnet. It contains stretched clasts of quartz, quartzite, amphibolite, amphibole schist, and a bright-green fuchsitic schist. The clasts of the paraconglomerate are probably locally derived from the underlying Stud Creek Metavolcaniclastics because the distinctive fuchsitic schist is confined to the Stud Creek. The Rock Mountain Conglomerate is radioactive, having radiation counts as much as 5 times background and containing as much as 270 ppm uranium and 95 ppm thorium. A drill hole that penetrated the Rock Mountain Conglomerate encountered 400 ft of a coarsening upward succession ranging from poorly sorted quartzite at the base to paraconglomerate in the upper half (Karlstrom and others, 1981b). The Rock Mountain Conglomerate is interpreted as a prograding alluvial fan deposit (Rust, 1979), on the basis of poor sorting, coarse grain sizes, coarsening-upward successions, the presence of anomalously high radioactivity which may reflect fossil placer accumulations of uranium- and thorium-bearing heavy minerals, and the limited lateral extent of the unit. The source of the Rock Mountain Conglomerate may have been a fault-bounded nearby terrane underlain, in part, by the Stud Creek Metavolcaniclastics.

BOW QUARTZITE

The Bow Quartzite is exposed throughout the northern part of the map area. It lies conformably on the Rock Mountain Conglomerate or on the Stud Creek Metavolcaniclastics where the Rock Mountain is missing. The Bow Quartzite also crops out in the core of an anticline near Arrastre Lake (sec. 9, T. 16 N., R. 80 W.) in the west-central part of the map area. The Bow Quartzite ranges in thickness from 650 ft to 1,950 ft. The unit consists chiefly (95 percent) of fine-grained (0.1–2 mm) foliated quartzite, with local beds of conglomerate, biotite and hornblende schist, phyllite, and quartz-rich carbonate. The quartzite is muscovitic (averaging about 10 percent muscovite) in the Arrastre Lake area but becomes more feldspathic to the north where it averages about 9 percent muscovite and biotite and 21 percent feldspar (potash feldspar>plagioclase). This increase in feldspar in the north suggests a northern source for the Bow Quartzite.

The most prevalent sedimentary structures in the Bow Quartzite are medium- to large-scale planar crossbeds (amplitude about 1.5–3 ft; mean inclination 23°). When combined for the entire unit, these crossbeds yield a bimodal paleocurrent distribution having a prominent mode directed southwest and a secondary mode directed northeast. Several oscillation ripple marks in the unit confirm a bimodal current pattern but record east-west-directed currents. Multiple ripple sets in one outcrop show east-west- and north-south-directed current directions.

The fine-grained clast size of the Bow Quartzite suggests low-energy deposition; large-scale planar crossbeds may be sand waves; and the bimodal paleocurrent distribution may indicate ebb and flood tides. Perhaps fluvial deposition of the Rock Mountain Conglomerate to the northeast, close to tectonically active highlands, gave way to marine deposition to the south and up-section.

COLBERG METAVOLCANICS

The Colberg Metavolcanics include some of the most interesting and distinctive rocks in the Medicine Bow Mountains. Metavolcanic rocks predominate and were initially amygdaloidal basalt; pillow basalt; rhyolite; fragmental metavolcanic rocks showing graded bedding and ranging in composition from basaltic to rhyolitic; and metasedimentary rocks including paraconglomerate, quartz-pebble and granule conglomerate, fine-grained quartzite, and metagraywacke that has local preservation of graded beds and contains thin beds of marble. The entire Colberg Metavolcanics succession varies greatly in thickness; in the south and east, thickness is 8,200 ft, whereas to the north and west, the thickness decreases to 300 ft.

Numerous faults and minor folds and the general absence of evidence for tops of beds makes it difficult to establish a reliable stratigraphic succession in the Colberg Metavolcanics. Although not well exposed, the contact between the Bow Quartzite and lower units of the Colberg Metavolcanics is conformable. The lower part of the Colberg Metavolcanics is mostly metavolcanic rocks with some interbedded metagraywacke, and, locally, calcareous schist, marble, and paraconglomerate. The volcanic rocks include amygdular basalt, andesitic basalt, rhyodacite, rhyolite, basalt tuff, andesitic-basalt tuff, and rhyolite tuff. Although mafic volcanic rocks exceed felsic volcanic rocks by a ratio of 4:1 or greater in most areas, felsic volcanic rocks are abundant in an area southeast of the old mining village of Colberg (sec. 10, T. 18 N., R. 79 W.) where a mafic to felsic cycle is exposed. There, basalt grades upward through rhyodacite to rhyolite, which is, in turn, overlain by chert containing disseminated sulfides. The presence of stratiform sulfides in chert suggests a marine cycle, but pillows have not been recognized in the mafic volcanic rocks at the base of the cycle, as might be expected in a marine succession.

Pillow basalt is present near the middle of the Colberg succession and is well exposed in an area east of Crater Lake (SE¼ T. 18 N., R. 79 W.). The most distinctive rock type in the Colberg section is paraconglomerate, which is interbedded with basalt and quartzite above the pillow basalt succession. Thin beds of paraconglomerate are present lower in the succession, but near the top they are well developed, ranging to as much as 400 ft thick. The paraconglomerate contains varying proportions of rounded boulders (diameters as much as 20 in.) of granite, porphyritic felsic igneous rocks, and quartzite. Clasts of mafic volcanic rocks are also abundant in the paraconglomerate but these clasts tend to be stretched and flattened. The matrix of the paraconglomerate is mafic, consisting of amphibole, biotite, and quartz.

Field studies determined that the Colberg succession exhibits facies changes on all scales. One of the best examples of this is in the upper part of the succession northeast of Colberg, where interbedded mafic volcanic rocks and paraconglomerate can be traced northeast around the limb of a large syncline. The thick paraconglomerate thins to the north and in its place metagraywacke, fine-grained quartzite, thin calc-schist, marble, and thin beds of quartz-pebble conglomerate are interbedded with mafic volcanic rocks. The graywacke and volcanic rocks (tuff) exhibit graded bedding locally, suggesting subaerial deposition, but interbedded basalt is amygdular and does not exhibit pillow structure. On the north limb of the syncline the succession gradually gives way to granule-bearing quartzite containing local beds of radioactive quartz-pebble conglomerate.

Chemical analyses of the Colberg Metavolcanics are limited to five samples taken primarily from the area south of Colberg. Two samples were classed as basalt in the field and proved to be andesitic basalt containing high amounts of iron and low amounts of aluminum. Three samples were tuff, one having the composition of andesite, one rhyodacite, and another rhyolite. An alkali, iron, magnesium plot shows an iron enrichment trend characteristic of tholeiitic magmas (Irvine and Baragar, 1971), and all samples are in the tholeiitic field in alkali-silica plots (Karlstrom and others, 1981a, p. 221–222).

The Colberg Metavolcanics are considered to be interbedded subaerial and marine rocks. Most of the basalt is amygdular and does not have pillow structure, and this, along with the wide variety in lithology and complex facies changes, suggest subaerial deposition. Some metatuff and metagraywacke show graded bedding, and pillow basalt has been recognized in part of the succession. The volcanic rocks were probably deposited near a shoreline where there was periodic encroachment of the sea or where deposition of volcanic material occasionally prograded seaward. The Colberg paraconglomerate may have been deposited in alluvial fans or possibly submarine channels and fans which developed adjacent to fault scarps bounding volcanic highlands.

CONICAL PEAK QUARTZITE

The Conical Peak Quartzite conformably overlies the Colberg Metavolcanics, occupies the core of a major syncline southeast of Arlington, and unconformably underlies the Magnolia Formation of the Deep Lake Group in the northeastern and central parts of the map area. The dominant lithology is a white, fine-grained micaceous and feldspathic quartzite. Metabasalt is considered to be part of the Conical Peak Quartzite that crops out above the quartzite near the north fork of Rock Creek (sec. 22, T. 17 N., R. 79 W.), near the head of Deep Creek (sec. 3, N. 17 N., R. 79 W.), south of Onemile Creek (sec. 7, T. 18 N., R. 78 W.), and on the upper main fork of the Medicine Bow River. Large-scale and planar crossbeds (amplitude about 3 ft; mean inclination 22.6°) are common in the Conical Peak, but trough crossbeds have been recognized locally. The paleocurrent distribution shows a bimodal, bipolar distribution with currents directed northeast and southwest.

The quartzite of the Conical Peak is similar in texture and composition to that of the Bow Quartzite except that the Conical Peak has more plagioclase as indicated in the statistical mode. Perhaps the Conical Peak was derived by reworking of the sediment of Bow Quartzite with addition of plagioclase from Colberg Metavolcanics. Some beds in the upper Conical Peak are rich in heavy minerals and are slightly radioactive. These beds are in quartzite that has low-angle cross-stratification that resembles beach deposits. The resemblance to beach deposits plus fine grain size, large-scale planar crossbeds, and bimodal paleocurrents suggest a marine origin for the Conical Peak. Additional evidence for a marine origin comes from metabasalt, which has textures suggestive of pillows, but these textures are not as well developed as in basalt of the underlying Colberg Metavolcanics.

AGE OF THE PHANTOM LAKE METAMORPHIC SUITE

Evidence for age of the Phantom Lake Metamorphic Suite is similar to that of the Overland Creek Gneiss. Beds of the Colberg Metavolcanics are cut by pink gneissic granite in outcrops along Foote Creek about 3.5 mi west of Arlington. Inasmuch as the gneissic granite could be either Late Archean or earliest Proterozoic, the relationship between the gneissic granite and the Colberg does not establish age except to show that part of the Phantom Lake (that is equivalent to the Colberg) is older than about 2,430 Ma. In the Sierra Madre, rocks considered equivalent to the lower units of the Phantom Lake Metamorphic Suite of the Medicine Bow Mountains are the Jack Creek Quartzite (equivalent to Bow Quartzite) and Silver Lake Metavolcanics (equivalent to Colberg Metavolcanics). Both the Jack Creek Quartzite and Silver Lake Metavolcanics are cut by the 2,700 Ma Spring Lake Granodiorite of this area, indicating a Late Archean age for at least the lower part of the Phantom Lake Metamorphic Suite. The Phantom Lake Metamorphic Suite is therefore considered to be Late Archean, but we emphasize that it is only the lower part of the succession that is dated.

SNOWY PASS SUPERGROUP

The Snowy Pass Supergroup is a succession of Early Proterozoic clastic rocks that is about 32,000 ft thick. It is divided into two groups, a lower group, the Deep Lake Group, which is mostly fluvial, and an upper group, the Libby Creek Group, which is mostly marine.

DEEP LAKE GROUP

The Deep Lake Group is a Proterozoic succession consisting of five formations, the Magnolia Formation (oldest), Lindsey Quartzite, Campbell Lake Formation, Cascade Quartzite, and Vagner Formation (youngest). Beds of the Deep Lake Group are exposed throughout the northern and central part of the map area from Onemile Creek in the northeast to South Brush Creek in the central Medicine Bow Mountains. Basal conglomerate of the Magnolia Formation unconformably overlies rocks of the Phantom Lake Metamorphic Suite from Onemile Creek in the northeast to localities on the Medicine Bow River about 2 mi south of Stillwater Park. The basal Magnolia overlies the Conical Peak Quartzite but cuts down section into the Colberg Metavolcanics in several areas. In the vicinity of Arrastre Creek in the western part of the map area the Magnolia lies with angular unconformably on the Colberg Metavolcanics.

MAGNOLIA FORMATION

The Magnolia Formation is composed of two members: (1) a basal conglomerate member, which consists of micaceous quartz-pebble conglomerate and paraconglomerate interbedded with pebbly, coarse-grained muscovitic-feldspar quartzite, overlain by (2) an upper quartzite member containing more mature subarkosic and arkosic quartzite interbedded with quartz-granule conglomerate. The conglomerate member is not always present in its full thickness. It reaches its maximum thickness of 1,300 ft in the northeast along Onemile Creek and thins to the southwest where it is less than 300 ft thick in outcrops along the Medicine Bow River. The conglomerate member thickens again in the vicinity of South Brush Creek where it exceeds 750 ft.

Paraconglomerate of the conglomerate member is thickest near the base and in areas where the member is thickest. The matrix of the paraconglomerate consists of quartz, feldspar, and muscovite with lesser amounts of sand-size rock fragments. Clasts, which are as much as tens of centimeters in diameter, are quartzite, schist, mafic volcanic rocks, and granite. Paraconglomerate in the Magnolia is distinctly different from that of the Colberg Metavolcanics in that it has an arkosic matrix with little or no amphibole and biotite and there are relatively few clasts of mafic volcanic rocks. The proportions of the various rock types, which are clasts in the conglomerate, change in different localities. For example, granite clasts are more abundant in the northeast, and quartzite and metavolcanic clasts are more abundant in the southwest. Another distinction between paraconglomerate of the Colberg Metavolcanics and the conglomerate member of the Magnolia is radioactivity; the paraconglomerate of the conglomerate member is distinctly radioactive and may contain as much as 50 ppm uranium and 200 ppm thorium.

In the northeast, from Onemile Creek to localities along Rock Creek, the conglomerate member has abundant beds of muscovitic quartz-pebble conglomerate in coarse-grained muscovite-feldspar quartzite. The quartz-pebble conglomerate layers range in thickness from beds 1 pebble thick to composite conglomerate zones as much as 20 ft thick. Some quartz-pebble conglomerate layers might be best classed as quartz-granule conglomerate inasmuch as clasts are mostly in the granule range, but most beds are coarser grained and contain pebble-size clasts. The clasts are quartz, quartzite, and granite in a matrix of quartz, feldspar, and mica (chiefly muscovite). The matrix of the quartz- pebble conglomerate is mostly mica and makes up about 30 percent of the conglomerate, and mica content ranges to as much as 75 percent in some samples. These quartz-pebble conglomerate layers are radioactive and, although uranium is leached from surface outcrops, conglomerate layers have 20 times the background radioactivity and contain as much as 177 ppm uranium and 915 ppm thorium on the surface and as much as 1,620 ppm uranium and 1,143 ppm thorium in the subsurface. The quartz-pebble conglomerate beds are not uniformly mineralized and individual beds may have as little as 10 ppm uranium. There are two principal zones of radioactive quartz-pebble conglomerate in the Onemile Creek area, which are as much as 60 ft thick. Typical uranium assays for quartz-pebble conglomerate of the upper zone are 100 ppm, but uranium values as much as 400 ppm over a thickness of 10 ft are present. Resource estimates from geostatistical analyses of core and surface samples of the Onemile Creek area (Borgman and others, 1981) estimate uranium resources at about 1,801 tons U_3O_8, at an average grade of 310 ppm U_3O_8, and thorium resources at 1,106 tons ThO_2, at an average grade of 290 ppm ThO_2.

Heavy minerals in the quartz-pebble conglomerate are pyrite, zircon, coffinite, thorite, thorogummite, monazite, huttonite(?), ilmenorutile, apatite, galena, chalcopyrite, bornite, marcasite, sphene, ilmenite, columbite, magnetite, anatase, rutile, and spessartine garnet. Some of these heavy minerals are clearly detrital, including ilmenorutile, rutile, zircon, garnet, apatite, magnetite, ilmenite, and columbite. Unfortunately, the sulfide minerals and the principal uranium-thorium minerals are recrystallized to the extent that they cannot be shown to be detrital. Although not part of the heavy mineral suite, graphite is sparsely present in round grains. The graphite may be of organic origin, because microprobe studies identified sulfur in the graphite (Desborough and Sharp, 1979).

The conglomerate member of the Magnolia Formation has characteristics of an alluvial fan system. For example, at Onemile Creek the conglomerate member can be divided into 5 units that progress towards finer grained successions up-section. Unit one is an arkosic paraconglomerate containing abundant large granite clasts interlayered with subarkose. Unit one grades up-section into unit two, which is trough-crossbed subarkose containing thin lenticular beds of radioactive quartz-pebble conglomerate. Unit two grades upward into unit three, which is a granular subarkose containing thin, lenticular beds of radioactive quartz-pebble conglomerate. Unit three is overlain by unit four, which is biotite-chlorite schist containing paraconglomerate lenses, and unit four is overlain by unit five, which is muscovite-rich subarkose containing thick and continuous beds of radioactive quartz-pebble conglomerate. The paraconglomerate at the base of this section is thought to represent mostly proximal mudflow deposits of an alluvial fan system; the mixed paraconglomerate and quartz-pebble conglomerate represent mid-fan deposits; and the coarse-grained quartzite containing beds of quartz-pebble conglomerate represent(s) parts of the distal fan.

The quartzite member of the Magnolia Formation is considered to be a distal facies of the conglomerate member on the basis of somewhat better sorting, the presence of a generally finer grain size, the presence of fining- upward sets having well-developed trough cross-bedding, and the presence of a polymodal paleocurrent pattern dominated by southwesterly directed currents. The quartzite member is thought to have been deposited in a well-developed and laterally extensive braided river system because of the lateral persistence of the unit (outcrop length of about 22 mi), average thickness of about 1,300 ft, and the change to less micaceous and arkosic quartzite from northeast to southwest, down the paleoslope.

LINDSEY QUARTZITE

The Lindsey Quartzite conformably overlies the Magnolia Formation and has many characteristics in common with the quartzite member of the Magnolia Formation, except that it is generally finer grained, better sorted, and is characterized by thin phyllitic layers and partings which are not present in the Magnolia. The Lindsey Quartzite is about 1,350 ft thick in the central Medicine Bow Mountains, and the Lindsey either pinches out or is conformably onlapped by the Cascade Quartzite to the north. The

Lindsey Quartzite exhibits some of the best preservation of trough crossbedding in the sedimentary successions; ripple marks are present locally, and pebble beds are along planar foreset beds and in small scours. The paleocurrent distribution is better defined than in the Magnolia because of the well-preserved directional indicators, and the distribution is strongly unimodal with currents directed southwest, similar to the Magnolia Formation. The unimodal paleocurrent distribution, abundance of trough crossbeds, and thin pebble beds and phyllite partings, which may represent lag gravels and thin overbank deposits, respectively, suggest a fluvial origin for the Lindsey. The general succession from conglomerate member to quartzite member of the Magnolia to Lindsey Quartzite suggests a transgressional sequence bringing more distal parts of a braided river system (Lindsey) over another braided river system (quartzite member of Magnolia) which encroached upon an alluvial fan system (conglomerate member of Magnolia).

CAMPBELL LAKE FORMATION

The Campbell Lake Formation is a thin (as much as 215 ft), discontinuous paraconglomerate-quartz phyllite succession which serves as a useful stratigraphic marker in the central Medicine Bow Mountains where it is thickest. The paraconglomerate contains poorly sorted subangular clasts (as much as 30 in. in diameter) of granite, quartzite, and phyllite in a poorly sorted micaceous arkose to subarkose matrix. The overlying quartz-phyllite is laminated, with laminae formed by alteration of quartz and mica-rich layers. The origin of the Campbell Lake Formation has not been determined, although the presence of a laminated phyllite similar to that in glacial deposits of the Vagner and Heart Formations suggests a glacial origin.

CASCADE QUARTZITE

The Cascade Quartzite is the thickest (as much as 2,800 ft) and most laterally extensive formation of the Deep Lake Group. It unconformably overlies units ranging in age from Archean Overland Creek Gneiss and an unnamed granite, in the northeast, to the Campbell Lake Formation, in the central part of the map area, to Archean quartzofeldspathic gneiss, in the southwest. The Cascade is composed of well-sorted, pebbly quartzite with some arkosic quartzite. Quartz and black chert pebbles are present in layers 5–10 cm thick.

Paleocurrent measurements from planar and trough crossbedding in the Cascade show a unimodal distribution about a west-southwest-directed mean paleocurrent (azimuth 248°). The small variance of the unimodal paleocurrent pattern has been interpreted as characteristic of fluvial systems (Potter and Pettijohn, 1977), and this plus the presence of both planar and trough crossbeds (interpreted to represent bars and migrating dunes in a river system), layers of well-rounded, well-sorted pebbles interpreted as lag gravels, and the unconformity at both base and top of the Cascade, which may indicate subaerial conditions before and after deposition, all suggest deposition in braided streams and rivers (Karlstrom and Houston, 1979a, b). However, none of these features is restricted to fluvial environments, and lateral continuity, mature composition, and relative consistency of pebble sizes in a down-current direction can be cited as evidence of a marine or deltaic origin.

VAGNER FORMATION

The Vagner Formation, which unconformably overlies the Cascade Quartzite, consists of interbedded paraconglomerate, marble, phyllite, and quartzite. Beds of the Vagner Formation are interpreted as glaciomarine; in fact, a glacial interpretation has been in favor for this succession since Blackwelder (1926). The basal unit of the Vagner Formation is a paraconglomerate (diamictite) containing angular and subangular clasts of granite, quartzite, and mafic schist in a subarkosic matrix. The paraconglomerate is a laterally persistent unit averaging 1,000 ft thick, which has features suggestive of glacial origin, such as dropstone clasts, poor sorting, subangular clasts, locally faint stratification, and a sand- size matrix having a chemical composition similar to that of other Early Proterozoic glacial deposits of North America (Sylvester, 1973; Young, 1970, 1973; Karlstrom and others, 1981a). The interbedded marble and phyllite of the Vagner Formation provide supporting evidence for a glaciomarine origin (Sylvester, 1973). Limited paleocurrent data from the Vagner show a change to more westerly directed transport (mean azimuth = 256°).

LIBBY CREEK GROUP

The Libby Creek Group is a marine succession that is in fault contact with rocks of the Deep Lake Group in the map area. The fault is a reverse fault that we postulate to have been a thrust fault later rotated to steep attitudes. This thrust fault brought younger beds of the Libby Creek Group over older beds of the Deep Lake Group. The fault extends from the northeast limit of outcrop to the southwest limit of outcrop of the Libby Creek and Deep Lake, and it is folded about a steeply plunging synformal axis (Houston and Parker, 1963).

The Libby Creek Group is divided into two successions for several reasons. The lower part of the Libby Creek Group is predominantly siliciclastic, whereas the upper part of the Libby Creek Group contains stromatolitic dolomite, volcanogenic rocks, and black shale, which were probably deposited at some distance seaward of the siliciclastic succession. The lower part of the Libby Creek is cut by a felsic intrusion, the Gaps Intrusion, dated as approximately 1,900–2,150 Ma by the Rb-Sr whole-rock method (Hedge in Houston and others, in press), whereas the upper part of the Libby Creek Group rocks are not invaded by felsic intrusions. The Deep Lake Group and lower part of the Libby Creek Group rocks can be correlated remarkably well with the Huronian Supergroup of Canada (Young, 1973; Houston and others, 1979; Karlstrom and others, 1981a), which is believed to have been deposited between 2,400–2,100 Ma (Young, 1983). However, the upper part of the Libby Creek Group can be correlated with rocks of the Marquette Range Supergroup of Michigan (Houston and Karlstrom, 1980), which may be composed of rocks less than 2,150 Ma, although the age of the lower part of the Marquette Range Supergroup is still debated (Young, 1983).

LOWER PART OF LIBBY CREEK GROUP

Rock Knoll Formation

The Rock Knoll Formation is a quartzite succession, which is in fault contact with formations of the Deep Lake Group and is unconformably overlain by the Headquarters Formation. The fault that brings Rock Knoll against the Deep Lake Group is interpreted as a thrust fault which has been subsequently rotated to a near vertical dip. The displacement on the fault is unknown. The maximum exposed thickness of the Rock Knoll Formation is 1,250 ft, but the base of the formation has been removed by faulting.

The Rock Knoll Formation consists of arkosic quartzite and interbedded phyllitic layers as much as 1 ft thick and diamictite layers as much as 3 ft thick. The diamictite contains quartz, quartzite, and granite clasts. The arkosic quartzite is similar in composition to the Vagner Formation of the Deep Lake Group and the Headquarters Formation of the Libby Creek Group in that plagioclase is the dominant feldspar (ranging from 2:1 to 5:1 plagioclase over potash feldspar). This is in contrast to most older quartzite of the Deep Lake Group where potash feldspar dominates.

Sedimentary structures in the Rock Knoll Formation include ripple marks, planar crossbedding, and clay galls. Paleocurrent data from the Rock Knoll indicate westerly directed paleocurrents (mean azimuth = 280°). The Vagner Formation, Rock Knoll Formation, and Headquarters Formation may all be glacial deposits as suggested by sedimentary features of the Vagner and Headquarters, but no diagnostic evidence of glacial origin has been found in the Rock Knoll. The abundance of plagioclase in all three formations may be an indication of climatic deterioration rather than a change of source, and might indicate glacial origin for these formations.

Headquarters Formation

The Headquarters Schist of Blackwelder (1926), which originally included rocks of the Vagner Formation and Rock Knoll Formation, was redefined by Karlstrom and Houston (1979b) and Lanthier (1979) as the Headquarters Formation and was restricted to two principal rock types. It includes a lower part (1,140 ft thick) composed of lenticular beds of paraconglomerate, quartzite, and schist and an upper part (980 ft thick) containing laminated schist and phyllite. Since the time of Blackwelder (1926), the Headquarters Formation has been interpreted to be glacial or glaciomarine in origin (Houston and others, 1968; Kurtz and Anderson, 1979; and Houston and others, 1981). The Headquarters Formation has been correlated with other Early Proterozoic glacial(?) deposits in North America, most notably the Gowganda Formation of the Huronian Supergroup of Canada (Young, 1970, 1973; Houston and others, 1979; Houston and Karlstrom, 1980).

The lower part of the Headquarters Formation contains several lenses of paraconglomerate (or diamictite) composed of granite, quartzite, and schist clasts (average size 4–6 cm, but ranging to as much as about 1 m in diameter) in a poorly sorted matrix of sand- and silt-size quartz, plagioclase, K- feldspar, rock fragments, and mica. At one locality, west of Twin Lakes, there are three stacked paraconglomerate units separated by quartzite. In other areas, paraconglomerate is interlayered with quartzite and schist. The quartzite is plagioclase arkose, similar in composition to the quartzite in the Rock Knoll. The upper part of the Headquarters Formation is a biotite- chlorite-quartz phyllite containing laminations formed by alternating quartz-rich and mica-rich layers.

Sedimentary structures in the Headquarters Formation include small-scale planar and trough crossbedding, laminations in both paraconglomerate and phyllite beds, dropstone structures in paraconglomerate, and climbing ripples. In one outcrop, Bouma A–B–C turbidite sequences are associated with paraconglomerate. The paraconglomerate beds are massive to slightly stratified, generally nongraded, poorly sorted, and contain isolated to poorly packed subangular to rounded clasts, some of which depress underlying strata but are covered by overlying strata, suggesting ice-rafted dropstones (Sylvester, 1973). The association of the paraconglomerate with laminated, presumably marine phyllite (metasiltstone), the abrupt lateral facies changes in the lower part, and the appearance of multiple conglomerate lenses are all consistent with a glaciomarine depositional setting. Prodeltaic mud flows and turbidites might have similar features but would not contain dropstones.

Kurtz and Anderson (1979) suggested, by analogy to Antarctic continental margin sediment, that deposition of the Headquarters Formation took place at some distance from the ice sheet on a moderate slope, as suggested by evidence for turbidite flow in paraconglomerate, the presence of laminated pebbly and nonpebbly argillite, which both occur on the continental slope off Antarctica, the scarcity of paraconglomerate deposited directly from glacial ice (tillite), and the abundance of finer grained rocks in the Headquarters Formation.

Heart Formation

The Heart Formation (Blackwelder, 1926; Houston and others, 1968; Lanthier, 1979) lies conformably between the Headquarters Formation and the Medicine Peak Quartzite. It is 2,200 ft thick and is predominantly quartzite. A phyllite unit as much as 300 ft thick is present locally about 1,300 ft above the base.

Quartzite of the Heart Formation is composed of quartz (52 percent) and sericite (31 percent) with lesser percentages of plagioclase, chlorite, biotite, and opaque minerals. A variety of sedimentary structures are present in the quartzite of the Heart Formation. These include small-scale planar and trough crossbedding, climbing ripples, interference ripples, symmetric ripples, ball and pillow structures, and graded bedding. The quartzite at the top of the formation is generally massive to plane bedded. The phyllite is well laminated (Lanthier, 1979).

The Heart Formation is interpreted as prodelta and delta-front sediment associated with a prograding macrotidal (tide-dominated) delta (Karlstrom and others, 1983). The laminated phyllite and very fine grained, argillaceous, feldspathic quartzite of the lower two-thirds of the Heart Formation may represent prodelta bottom-set deposits (Reinick and Singh, 1972, p. 273). The presence of graded bedding, climbing ripples, and ball and pillow structures suggest rapid deposition of sediment temporarily thrown into suspension, perhaps as a result of slumping on the delta-front slope (Blatt and others, 1972, p. 131; Elliot, 1986).

Medicine Peak Quartzite

The Medicine Peak Quartzite was initially defined by Blackwelder (1926) as a two-fold unit consisting of a lower bluish-green, kyanitic quartzite having well-developed crossbedding and an upper, white, medium- to coarse- grained quartzite marked by thin beds of conglomerate. Detailed mapping of the upper part of the Medicine Peak Quartzite by Flurkey (1983) demonstrated that the quartzite is a complex succession consisting of five facies. Only one of these facies, the Klondike Lake Conglomerate Member, is a mappable unit that can be traced for any distance along strike (5.5 mi).

The Medicine Peak Quartzite conformably overlies the Heart Formation and is conformably overlain by the Lookout Schist. In several localities it is in fault contact with younger metasedimentary rocks. The quartzite is folded and truncated by the Mullen Creek-Nash Fork shear zone at its southwest limit of outcrop and extends to the northeastern margin of the Medicine Bow Mountains where it is thrust over Phanerozoic successions of the Laramie Basin. The quartzite is about 5,600 ft thick and forms an impressive cliff in the Snowy Range. The following discussion is mostly from Flurkey (1983).

Although the Medicine Peak Quartzite is predominantly a medium- to very coarse grained quartzite containing pebbly zones and layers of quartz-pebble conglomerate, it is quite variable in composition with compositional changes related to the facies as defined by Flurkey (1983). Thirty samples from all facies studied by Flurkey (1983) contained 84 percent quartz, 0.5 percent total feldspar, 11 percent sericite, 3.1 percent aluminosilicates (kaolinite, pyrophyllite, and kyanite), 1 percent opaque minerals, and the remainder carbonate, zircon, tourmaline, and biotite. There is an inverse relationship between feldspar and the aluminosilicate minerals kaolinite, pyrophyllite, sericite, and kyanite, suggesting that the aluminosilicates may be an in situ alteration product (diagenetic and (or) metamorphic) of feldspar and that some part of the unit was more arkosic than indicated by modal analyses (Flurkey, 1983). Flurkey's (1983, p. 10–63) detailed study of the mineralogy of the Medicine Peak Quartzite led him to conclude that most of the argillaceous quartzite (quartzite containing 10–15 percent sericite, kyanite, kaolinite, and pyrophyllite) was subarkosic sandstone, initially, and that the very argillaceous quartzite (>25 percent aluminosilicates) may have been derived from diaspore-rich sandstone or represent kaolinite-rich sediment.

The best exposures and best developed structures and textures in the Medicine Peak Quartzite are in the central area of outcrop from Lake Marie (sec. 24, T. 16 N., R. 80 W.) to Telephone Lakes (sec. 9, T. 16 N., R. 79 W.). Inasmuch as the lower part of the Medicine Peak is covered or poorly exposed in this area, information critical to establishing depositional environment had to come from the upper part of the quartzite. Flurkey (1983) measured four partial sections in the central area and was able to define seven facies, two at the base of the Lookout Schist and five in the Medicine Peak Quartzite. Facies three of Flurkey represents the top of the Medicine Peak Quartzite and is composed of fine- to medium-grained quartzite, which is thin bedded and contains some phyllite beds and partings. The quartzite is sericitic and contains abundant small-scale sedimentary structures. Feldspar is rare.

Facies four of Flurkey is present locally in cliffs above Lake Marie. It is a medium- to coarse-grained quartzite characterized by an abundance of large-scale crossbedding. Facies five and six of Flurkey constitute the bulk of the upper part of the Medicine Peak Quartzite and interfinger on all scales. Facies five is a medium- to very coarse grained, massive to plane- bedded quartzite. Some plane beds are slightly discordant to bedding and may be large-scale, low-angle crossbeds. More distinctive crossbedding may be abundant locally and consists of solitary small- to large-scale planar crossbeds. The quartzite of this facies is sericitic and is considered to be altered subarkose by Flurkey (1983, p. 82). Facies six of Flurkey is coarse- to very coarse grained, pebbly quartzite and quartz-pebble conglomerate. Clasts of the conglomerate are as much as 75 mm in diameter and are quartz, black chert, and rare red jasper. Conglomerate has both graded and reverse graded bedding. The graded bedding sets have sharp or scoured bases, and the conglomerate grades upward into a pebbly, coarse-grained quartzite. The pebbly quartzite may be overlain by nonpebbly quartzite having planar crossbeds. Facies seven of Flurkey is the Klondike Lake Conglomerate Member, which is a zone of bluish-black hematite-bearing conglomerate and coarse- grained quartzite as much as 55 ft thick that is about 410 ft below the top of the Medicine Peak Quartzite. As noted above, this conglomerate layer is the only mappable unit in the Medicine Peak Quartzite and can be traced about 5.5 mi from near Silver Lake (sec. 35, T. 16 N., R. 80 W.) to a locality just east of the Gaps (sec. 8, T. 16 N., R. 79 W.).

Flurkey considered the Medicine Peak Quartzite to be primarily marine. The upper part of the quartzite was thought to represent shallow marine and subtidal delta-plain sediment. For example, facies three, which is finer grained and marked by small-scale sedimentary structures, was considered to be sand deposited on the margins of a subtidal delta plain where wave and (or) tidal currents created a variety of small-scale structures. Facies six, with its fining upward sets and basal conglomerate, was interpreted as having been deposited in the fluvial-dominated part of the delta plain where marine reworking was limited. Details of the sedimentology and additional evidence to support the delta model are in Flurkey (1983) and Karlstrom and others (1983).

We must emphasize that the above discussion refers to the upper part of the Medicine Peak Quartzite and that less is known about the lower three quarters of the unit. The best exposures of the lower part of the Medicine Peak are in the southwest part of the map. Houston and others (1968) describe a section from sec. 35, T. 15 N, R. 81 W., as having a 350-ft-thick unit of dark-blue, dark-green, and dark-gray, medium- to coarse-grained quartzite at its base, which is overlain by 730 ft of blue to blue-gray, coarse-grained, kyanitic quartzite having well-developed crossbedding. The thickness of the crossbedded sets range from 3 in. to 2 ft. The crossbedded unit is overlain by 620 ft of light-blue, light-green, and white, coarse-grained quartzite, which has local kyanite-rich areas. Kyanite and rose quartz veinlets are common in the lower part of the Medicine Peak Quartzite. The middle part of the Medicine Peak is similar to facies five and six of Flurkey.

The lower 2,000 ft of the Medicine Peak is richer in kyanite; 11 samples described by Houston and others (1968) and Young (1973) contained approximately 10 percent kyanite, 8 percent sericite, and 80 percent quartz. Only one of these samples contained feldspar.

The Medicine Peak Quartzite has many characteristics in common with Early Proterozoic aluminous quartzite from North America. As pointed out by Young (1973, p. 101), Early Proterozoic quartzite is unusually thick, commonly overlies deposits interpreted as glacial, commonly underlies iron-formation and (or) stromatolite-bearing carbonate rocks, has unusual aluminous mineralogy, such as kaolinite, diaspore, pyrophyllite, kyanite, and andalusite, and may have special textural attributes, such as an association of clay minerals and rounded quartz grains and a bimodal distribution of the matrix grains. The abundance of aluminous minerals has attracted the attention of most workers (Robertson and Card, 1972; Young, 1973; Nesbitt and Young, 1982) and has been interpreted as a climatic factor. Tropical climate may have been present prior to deposition (if all aluminous minerals are detrital), during deposition (if some aluminous minerals are detrital and some are diagenetic), or after deposition (if all aluminous minerals are diagenetic). Flurkey's (1983) description of the textures and mineralogy of the Medicine Peak Quartzite suggest that the aluminous minerals are mostly diagenetic with later metamorphic recrystallization. This implies that a subtropical or tropical climate existed during deposition of the Medicine Peak and was confined to Medicine Peak time, since neither the underlying Heart Formation or overlying Lookout Schist contain aluminous minerals other than sericite.

Lookout Schist

The Lookout Schist is a laminated quartz-muscovite schist of great lithologic and structural variability that conformably overlies the Medicine Peak Quartzite and conformably underlies the Sugarloaf Quartzite. It is found throughout the central Medicine Bow Mountains and extends to the eastern limit of Precambrian outcrop, but it thins to a wedge-edge to the southwest in sec. 4, T. 15 N., R. 80 W. The Lookout Schist is an incompetent unit between the two massive bodies of quartzite and is complexly folded, but, despite the severe deformation, primary structures are preserved in most outcrops. The Lookout is variable in thickness and has a maximum thickness of about 1,300 ft near Lake Marie.

Houston and others (1968, p. 30) measured a section of Lookout Schist west of Telephone Lake that had a basal gray, laminated quartz-muscovite schist (730 ft thick), which has well-developed convolute structure, and is overlain by white, crossbedded quartzite (31 ft thick) overlain by a unit composed of beds of white quartzite (5 ft thick) that alternate with gray schist (71 ft thick), an amphibolite sill (43 ft thick), gray, laminated quartz-muscovite schist with convolute laminations (151 ft thick), a diabase sill (77 ft thick), blue, fine-grained, crossbedded quartzite (22 ft thick), greenish-gray chlorite schist (53 ft thick), and, at the top of the section, blue, medium- to fine-grained, crossbedded quartzite (8 ft thick). The amphibolite and diabase sills are not part of the Lookout. This section may not be complete because the upper part of the Lookout Schist is in fault contact with Nash Fork Formation at this locality, but the lithology is typical of the Lookout except that beds of brown dolomite 12 in. thick are present locally in the thick, laminated quartz-muscovite schist units, and magnetite-rich quartzite has been identified in the Lewis Lake area (sec. 17, T. 16 N., R. 79 W.).

The principal sedimentary structures in the Lookout Schist are laminations that result from alternations of quartz-rich and mica-rich layers in the quartz-muscovite schist. These laminations are crosscut by one or more metamorphic foliations, including probably slaty cleavage (Wilson, 1975). These thick schist beds have zones of thin, layered graded beds sandwiched between more homogeneous sequences, and convolutions are present in all but the thinnest beds, but, according to Wilson (1975), other structures are uncommon.

In the Lake Marie-Mirror Lake area, graded units that are much larger scale (some 6–7 ft thick) are present that consist of sets of quartzite and mica schist layers, with the quartzite layers several times thicker than the schist. Small-scale planar crossbeds are common in the quartzite, along with hummocky cross-stratification and climbing ripples. Sole marks have been recognized at the base of some graded sets. Large-scale graded beds in the Lewis Lake area are coarse grained at the base and contain pebble-size clasts. According to Wilson (1975), quartzite layers containing over 40 percent magnetite are in the large-scale graded sets. The magnetite may be disseminated throughout the quartzite but may also be in laminae, including crossbedded laminae, suggesting that the magnetite is detrital rather than chemical in origin.

Injection structures, such as clastic dikes and flame structures, are common in the Lookout Schist. Clastic dikes are present in both the graded sets and in the more homogeneous quartz-mica schist. According to Wilson (1975), the source of the dikes is the more quartzitic bed. Material in the dike can be readily traced to the source bed, and boundaries of the dikes are always sharp.

The composition of the Lookout Schist is dependent on rock type. The quartz-mica schist has 30–70 percent muscovite, and quartzite has as much as 90 percent quartz but averages 70–80 percent quartz. All rocks tend to be feldspathic and may contain as much as 33 percent feldspar, which is almost entirely plagioclase. Chlorite and biotite are in a few samples but are uncommon. Heavy minerals are abundant in some samples and include zircon, tourmaline, sphene, apatite, epidote, garnet, and magnetite. Tremolite and siderite have been reported in the Lookout by Houston and others (1968).

Karlstrom and others (1983) have suggested that the graded and contorted beds, clastic dikes, and laminated schist were deposited in a prodelta and delta-front environment. They also suggest that the small-scale planar and hummocky crossbeds and climbing ripples in the fine-grained arkosic quartzite indicate storm and reversing tidal currents of the shallow-water part of the delta front.

Sugarloaf Quartzite

The Sugarloaf Quartzite conformably overlies the Lookout Schist and is in fault contact with the overlying Nash Formation. The Sugarloaf is 1,900 ft thick but is significantly reduced in thickness in the east-central Medicine Bow Mountains due to faulting.

The Sugarloaf Quartzite is a white, fine- to medium-grained quartzite that is composed almost entirely of quartz. A small amount of sericite, tourmaline, and opaque minerals can be identified in thin section, but samples studied by Houston and others (1968) and Karlstrom and others (1981a) contained 97–99 percent quartz. Thin beds of hematitic quartzite are near the top of the Sugarloaf, and Blackwelder (1926) reported thin beds of quartz- pebble conglomerate near the top of the unit.

Sedimentary structures are rare or difficult to identify in the Sugarloaf because of its purity, but medium to thin plane beds, small-scale planar and trough crossbeds, and symmetric, interference, and climbing ripples have been recognized in several areas. In addition, a few large-scale planar crossbeds were observed.

The maturity of the quartzite and the presence of small-scale sedimentary structures, such as oscillation ripples, suggests that the Sugarloaf Quartzite was deposited in a shallow marine environment. Karlstrom and others (1983), in keeping with their deltaic model for the lower part of the Libby Creek Group, interpret the lowering of water level during Sugarloaf time as a response to progradation of a delta system.

Paleocurrent Directions in Lower Part of Libby Creek Group

Paleocurrent directions in rocks of the lower part of the Libby Creek Group were measured by Lanthier (1979), Karlstrom and others (1981a), and Flurkey (1983). A summary of these measurements, as well as those made in all other metasedimentary rocks of the northern Medicine Bow Mountains, is in Karlstrom and others (1981a, pl. 2).

With the exception of the Lookout Schist, all formations of the lower part of the Libby Creek Group have southwest-directed paleocurrent directions. The mean vector azimuths of these paleocurrent measurements is relatively small: Rock Knoll 75, Headquarters and Heart 83, Medicine Peak 65, and Sugarloaf 61. This suggests that during deposition of the lower part of the Libby Creek Group the paleoslope was directed southwest. The fluvial paleocurrent directions measured in the Deep Lake Group are also oriented southwest, and this orientation is parallel to the main faults of the Cheyenne Belt.

Parallelism of paleocurrent directions measured in marine and fluvial successions might be expected in an elongate intracratonic basin or in elongate rifted basins (Larue, 1981, 1983) but not on an open continental margin where the paleoslope should be perpendicular to the shoreline. Karlstrom and others (1983) recognized this problem and suggested that the Deep Lake Group and lower part of the Libby Creek Group were indeed deposited in a rifted basin. These writers (Karlstrom and others, 1983, p. 1271–1273, fig. 9) developed a triple junction model with the Medicine Bow segment oriented northeast, the Sierra Madre east-west, and a third segment south-southeast. Details of this model will not be discussed here, but it does offer an explanation for the interesting paleocurrent orientations on a trailing margin.

In this model, the upper part of the Libby Creek Group is not confined to the rifted basin but is laid down on a marine shelf that faces open ocean.

UPPER PART OF LIBBY CREEK GROUP

The upper part of the Libby Creek Group includes the three uppermost formations of the metasedimentary succession of the Medicine Bow Mountains: Nash Fork Formation, Towner Greenstone, and French Slate. The basal formation of the upper part of the Libby Creek Group, the Nash Fork Formation, is in fault contact with rocks of the lower part of the Libby Creek Group. This fault is interpreted as a thrust fault that has been steepened during later deformation and is believed to have major displacement that caused the upper part of the Libby Creek off-shore marine facies to be juxtaposed upon the deltaic facies of the lower part of the Libby Creek Group.

Nash Fork Formation

Blackwelder (1926) mapped a three-fold succession, which he named the Nash Marble, Anderson Phyllite, and Ranger Marble, because, in his type locality, a black graphic and pyritic phyllite was present between two metadolomite successions. Houston and others (1968) determined that the phyllite is at different stratigraphic horizons in the succession and renamed the entire succession the Nash Fork Formation.

As noted above, the Nash Fork Formation is in fault contact with the lower part of the Libby Creek Group, but it appears to be structurally conformable with the overlying Towner Greenstone. It is nearly 6,500 ft thick and may have been thicker because of lost stratigraphic section due to faulting.

The Nash Fork Formation is predominantly tan, siliceous metadolomite containing thick lenses of black, graphitic phyllite and, locally, thin beds of quartzite, metachert, flat-pebble conglomerate, and sulfide- facies iron-formation. The most striking sedimentary feature of the metadolomite is stromatolitic bioherms which are in a variety of shapes and sizes (Knight and Keefer, 1966; Knight, 1968).

The metadolomite consists primarily of dolomite and quartz and variable amounts of plagioclase, muscovite, phlogopite, tremolite, talc, apatite, and opaque minerals. Phyllite is variable in composition but consists chiefly of quartz and mica. The different types of mica are muscovite, chlorite, biotite and (or) phlogopite, and other constituents are graphite, tremolite, dolomite, zoisite, sphene, pyrite, and other opaque minerals.

Knight and Keefer (1966) and Knight (1968) mapped about 150 bioherms and three reefs in the metadolomite of the central Medicine Bow Mountains. The reefs are about 180 ft thick and as much as 0.6 mi in length and consist of many bioherms. The bioherms range from about 3 ft to about 100 ft thick and 10 ft to 100 ft in length (Knight, 1968). Knight (1968) subdivided the bioherms into several structural types and forms which are discussed later in the text. No specific organisms have been identified from these stromatolites. Some of the bioherms show evidence of erosion. Crossbedding and flat-pebble conglomerate are present in the dolomite (Houston and others, 1968; Knight, 1968).

Fenton and Fenton (1939) suggested that the bioherms grew in clear, shallow water. The abundance of siliciclastic particles indicates that at least part of the time there was a considerable influx of clastic debris. Current activity is also shown by crossbedding, flat-pebble conglomerate, and erosional scours on the bioherms. Pyrite and graphite in the phyllite implies restricted circulation and reducing conditions for at least part of the time. As suggested by Houston and others (1968), the algal mats may have been deposited in intertidal flats, and the reducing environments developed in adjacent restricted bays; however, an alternative is that a part of the Nash Fork Formation was deposited on subtidal mud flats some distance from shore. In this model, the fine clastic material, represented by shale, may have settled from suspension, and the algal "reefs" developed on the mud surface. Knight (1968, p. 113) points out that some of the bioherms he studied are completely enclosed in slate and he infers that they formed on a mud-silt facies. This mud-flat subtidal model does not explain the evidence of current activity or erosion, but some of these features may not be primary.

The extraordinary size of the Nash Fork stromatolites as compared to modern stromatolites may be due to a lack of grazing and burrowing organisms in the Precambrian (Garrett, 1970) but would also seem to require greater water depth than modern localities such as Shark Bay in Australia (Awramik, 1971; Hoffman, 1976). We thus have some evidence supporting intertidal environments and some evidence supporting subtidal environments, and we suggest that both facies are probably present in the Nash Fork Formation.

It is instructive to compare stromatolites of the Nash Fork Formation with those of similar age described by Hoffman (1976) in the Great Slave Supergroup, which is east of Great Slave Lake in Canada. Hoffman interpreted the Pethei Group of the Great Slave Supergroup as a carbonate platform consisting of five facies. Hoffman's platform-shoaled facies consists of heterogeneous alternations of laminar stromatolites and columnar stromatolites. The laminar stromatolites are flat, corrugate, or mammillate in morphology; the mammillate form corresponding to the laterally linked hemispheroids of Logan and others (1964). Hoffman also described tabular oncolites as being typical of his platform-shoaled facies. Hoffman's platform-shoaled facies has edgewise conglomerate and ooid grainstone in ripple laminated units, which are additional support for his concept of deposition in shallow water. Knight and Keefer (1966) described a type II stromatolite in the Nash Fork Formation which has a morphology remarkably similar to stromatolites of Hoffman's platform-shoaled facies, including the laterally linked hemispheroids and tabular oncolites. This is also the facies of the Nash Fork in which flat-pebble conglomerate and crossbedding is recognized. We suspect these stromatolites of the Nash Fork were deposited in an environment similar to Hoffman's platform-shoaled facies. Knight (1968, pl. 1) shows the platform-shoaled-type stromatolites in two principal horizons of the upper one-half of the Nash Fork Formation.

Hoffman's platform-submerged facies contains stromatolitic beds that he describes as having great monotony. These stromatolites are in narrow, closely spaced columns and may exhibit discontinuous vertical partitions between the stromatolite columns. These stromatolites are similar to Knight's (1968) type I stromatolite, except that Knight considered some of these stromatolites to be large spherical oncolites formed around a nucleus of foreign material. This correlation of Knight's type I stromatolite with those of Hoffman's platform-submerged facies is less convincing than the correlation of Knight's type II and Hoffman's platform-shoaled facies, but the overall similarity of the Nash Fork stromatolites with those of Hoffman's platform facies suggests that this part of the Nash Fork was laid down on a carbonate platform.

The platform-submerged facies of Hoffman grades into a basin-floor facies, which has no analogue in the Nash Fork, and this platform-submerged facies, in turn, grades into a basin-slope facies that consists of "even- bedded, very fine grained, light-gray limestone with partings of dark-gray shale" (Hoffman, 1976, p. 562). This basin-slope facies of Hoffman has no analogue in the Nash Fork Formation of the Medicine Bow Mountains but is very similar to the Slaughterhouse Formation of the Sierra Madre, which occupies the same stratigraphic position as the Nash Fork Formation and is correlated with it (Houston and others, in press). Hoffman's basin-slope facies grades into a graywacke-turbidite facies which is the last of Hoffman's facies and helps to establish the overall correlation from shallow to deep water. No graywacke-turbidite facies has been identified in either the Medicine Bow Mountains or Sierra Madre.

We suggest that the Nash Fork Formation represents a relatively shallow carbonate platform which has a basin slope facies to the south in the Slaughterhouse Formation of the Sierra Madre.

Towner Greenstone

The Towner Greenstone is a dark chlorite amphibolite consisting of both massive and schistose amphibolite. No contacts have been observed with the underlying Nash Fork Formation, but Childers (1957, p. 27) noted a conformable, sharp contact with the overlying French Slate in the northeast wall of Silver Creek Canyon. The greenstone has a maximum thickness of 1,600 ft near Towner Lake (sec. 15, T. 16 N., R. 79 W.) but thins to the southwest, is missing in sec. 15 and 16, T. 15 N., R. 80 W., and reappears in the French Creek syncline as a thin amphibolite bed 165 ft thick. According to Childers (1957), the basal part of the Towner Greenstone contains many small lenses of poorly cemented quartzite about 0.4 m in length and 1 cm thick. The uppermost part of the Towner Greenstone also contains lenses of coarse-grained quartzite as well as very fine grained quartzite that may be chert.

The Towner Greenstone was regarded as a marine volcanic succession by Blackwelder (1926) and Houston and others (1968) because of the presence of lenses of quartzite and chert(?), but these authors did not recognize primary structure in the thick amphibolite bodies. Blackwelder (1926, p. 645) stated that the Towner's weathered surfaces resembled those of the volcanics of the Hemlock Formation of northern Michigan. In 1985, P.K. Sims of the U.S. Geological Survey visited this area with one of the authors (Houston) and examined outcrops of the Towner and recognized structures similar to those in pillow lavas of the Lake Superior area. Subsequently, poorly developed pillow structures have been recognized in most of the amphibolite of the Towner Lake area. The Towner can now be classed as marine volcanic rocks with some certainty.

French Slate

The French Slate overlies the Towner Greenstone and is truncated by the Mullen Creek-Nash Fork shear zone along its entire upper contact. It has a minimum thickness of 2,000 ft. This formation consists primarily of laminated, black ferruginous slate and phyllite. The laminae consist of layers containing muscovite, chlorite, quartz, and opaque minerals that alternate with quartz-rich layers containing minor muscovite and chlorite. Metacrysts of biotite and pyrite are common. We have observed two thick layers of hematitic iron formation in the middle and upper part of the unit. The hematite is in a matrix of very fine grained quartzite (metachert?).

In sec. 11, T. 15 N., R. 80 W., a sulfide-rich zone is poorly exposed at the base of the French Slate near its contact with amphibolite of the Towner Greenstone. The French Slate is graphitic and has disseminated pyrite. The area has been prospected, and beds of massive sulfide (pyrite) a few cm thick can be observed in pieces of core left from diamond drilling. The sulfide- rich zone must be extensive because a large gossan can be observed for a distance of approximately 1 mi along a timber road in the S sec. 11, T. 15 N., R. 80 W. The Towner Greenstone of this area is massive amphibolite without primary structure. Its actual contact with French Slate is not exposed. The proximity of the sulfide-rich French Slate and the volcanogenic Towner Greenstone suggest a genetic relationship and that the sulfide may be volcanogenic.

The stratigraphy of the French Slate has not been established primarily because of poor outcrop. Blackwelder (1926) suggested that the lower part was more graphitic and contained more iron (thin layers of hematite), and the upper part was a more distinctly laminated slate or phyllite. This general transition has also been recognized by the authors.

The stratigraphic position and composition of the French Slate are typical of Early Proterozoic foredeeps as discussed by Hoffman (1987). This depositional setting was suggested by Hills and Houston (1979) who compared the Proterozoic continental margin of this area with the Phanerozoic Appalachian trailing margin and suggested that the Towner Greenstone and French Slate were developed as an island arc approached the rifted continental margin from the south. Recent models of Phanerozoic foredeep basins propose that they develop as a result of attempted subduction of a stable shelf of a trailing margin beneath approaching volcanic arcs or microplates (Shanmugan and Lash, 1982). The foredeep basin develops between a fold and thrust belt that progrades towards the continental margin as the arc or microplate approaches (Shanmugan and Lash, 1982, p. 564, fig. 4). Crustal thinning beneath the basin, accompanied by normal faulting (Hoffman, 1987), promotes basaltic magmatism and explains the volcanics of the Towner Greenstone as well as gabbroic sills that are common in the French Slate and underlying platform succession. Foredeep deposits are laid down conformably on continental shelf deposits (typically, deposits of a carbonate bank) and show an overall coarsening-upward succession beginning with starved-basin deposits followed by thick turbite successions and conglomerate that are derived from the approaching arc or microplate. Hoffman (1987) suggests that the Proterozoic foredeeps are the proper setting for iron-formation which he states was deposited on the outer ramp and would appear in the basal part of a foredeep succession. Hoffman (1987) describes hematitic iron-formation and euxinic shale as characteristic of basal foredeep deposits of Early Proterozoic Coronation Supergroup in the northwest territory of Canada that are very similar to the succession of the lower part of the French Slate.

We interpret the lower exposed French Slate as the lower part of a foredeep succession and the upper exposed French Slate as possible distal turbidites that are more typical of a lower middle foredeep succession.

METAMORPHISM

Metasedimentary and metavolcanic rocks of the Overland Creek Gneiss and Phantom Lake Metamorphic Suite are amphibolite facies. Rocks of the Deep Lake Group increase in rank from greenschist facies (biotite zone) in the north- central area of outcrop, where they are less deformed, to amphibolite facies in the northeast, where they show evidence of increasing deformation. Rocks of the Libby Creek Group are greenschist facies except where they are in contact with the Mullen Creek-Nash Fork shear zone. Directly north of the Mullen Creek-Nash Fork shear zone, the French Slate shows an inverted metamorphic gradient (Duebendorfer, 1988). The metamorphic grade increases from chlorite facies to staurolite facies where the French Slate is in contact with the shear zone. Duebendorfer (1988) interprets this increase in grade to be the result of conductive heating associated with the emplacement of a hot upper plate over a cooler lower plate.

STRUCTURE

ARCHEAN

The Archean rocks of the Medicine Bow Mountains have undergone multiple episodes of deformation as shown by successive emplacement and deformation of basalt dikes and as shown by structural analyses (Houston and others, 1968). The best developed structure in the gneiss and schist trends northwest and includes the structure of the Archean Overland Creek Gneiss in the map area. This northwest-striking structure is not present in the metasedimentary rocks and probably developed prior to deposition of rocks of the Phantom Lake Metamorphic Suite.

The Archean(?) Phantom Lake Suite records several deformational events as shown by the presence of folded folds and by the observation of more than one s surface in individual outcrops. These metasedimentary and metavolcanic rocks are now folded into tight folds and are often overturned (axial planes dipping northwest). These folds probably formed during intrusion of granitic magma, because granite intrudes both the Overland Creek Gneiss and Phantom Lake Metamorphic Suite, has a well-developed foliation parallel to that of the metamorphic rocks it intrudes, and granite and pegmatite dikes are folded with the foliation in the Overland Creek Gneiss and Phantom Lake Metamorphic Suite (Karlstrom and others, 1981a, p. 315–317). Unfortunately, we do not know the exact age of the granite, but it probably is between about 2,700–2,400 Ma, Late Archean to earliest Proterozoic.

Faults in the Phantom Lake Metamorphic Suite parallel the dominant foliation and show both vertical and horizontal movement. Vertical movement on these faults may have been in response to the same stress that developed the major folds. Horizontal movement (left-lateral) may be related to a later fold system that developed about an east-west axis.

PROTEROZOIC

DEEP LAKE GROUP

The earliest Proterozoic deformation is recorded in rocks of the Deep Lake Group and produced upright, northeast-to-east-trending folds which are approximately coaxial with the more tightly appressed folds of the Phantom Lake Suite. A single anticline-syncline system in the Deep Lake Group transverses the entire northern Medicine Bow Mountains. The fold axes vary in strike from east-west to north-northeast, and plunges are shallow, less than 30 degrees to the southwest and northeast.

The relationship between folds of the Deep Lake Group and Phantom Lake Suite can be seen in the Crater Lake area (sec. 35, T. 18 N., R. 79 W.), where open folds in the Deep Lake Group are coaxial with tight to isoclinal folds in the unconformably underlying Phantom Lake Suite.

The timing of the development of the open folds of the Deep Lake Group can be constrained by uranium-lead dating of zircon in the Magnolia Formation and by ages determined by Rb-Sr whole-rock method and by uranium-lead zircon on gabbroic sills and dikes that cut the Deep Lake Group. Zircons of the Magnolia Formation of the Sierra Madre have been dated by Premo (1983) as 2,451±12 Ma, gabbroic sills that cut the Cascade Quartzite in the Sierra Madre have been dated as 2,092±4 Ma (Premo, 1983, U-Pb zircon), and a differentiate of a gabbroic dike that cuts the Lookout Schist of the Medicine Bow Mountains has been dated as about 2,100 Ma (Hedge in Houston and others, in press) by Rb-Sr whole-rock method. The deformation must have taken place between about 2,500 and about 2,100 Ma.

Two additional fold systems can be observed in the Deep Lake Group and underlying Phantom Lake Metamorphic Suite. In the northeastern Medicine Bow Mountains, the generally east- to northeast-striking folds of the Deep Lake Group and Phantom Lake Metamorphic Suite are rotated, and the deformation was intense enough locally to obscure earlier structures of the metasedimentary-volcanic successions and to develop a strong foliation in gabbroic sills. Plots to poles of bedding and foliation in both the Phantom Lake Suite and Deep Lake Group form a great circle girdle defining a west-plunging fold axis (Karlstrom and others, 1981a, p. 321). This same east-west fold system is well developed in the Phantom Lake Suite of the Sierra Madre, but we are uncertain of the involvement of the Deep Lake Group in this area. If the gabbroic sills of this northeast area are the same age as those noted above (2,100 Ma), this deformation took place after 2,100 Ma and may have involved the lower part of the Libby Creek Group. The proximity of this structure to Laramide thrust faults suggests that it may have developed as late as Laramide, but the plastic deformation, seen in these structures, suggests a Precambrian age.

In several areas, mesoscopic folds in the northern Medicine Bow Mountains exhibit a northwest-trending foliation that crenulates folds associated with the west-plunging fold axis. Similar northwest-trending folds and foliation are present in incompetent beds in local areas of the central Medicine Bow Mountains, especially near Twin Lakes (secs. 14 and 23, T. 16 N., R. 80 W.). In addition, folds and gabbroic intrusions are gently warped about the northwest axes. This is probably a Proterozoic event but we have no way to constrain it in time except to say it postdated intrusion of gabbroic sills.

LIBBY CREEK GROUP

The structural style of the Libby Creek Group differs from older successions and resembles that of rocks south of the Mullen Creek-Nash Fork shear zone more than it does those to the north. Rocks of the Libby Creek Group strike northeast and dip steeply southeast. They have characteristics of the steep limb of a south-facing monoclinal flexure in which cleavage is generally parallel to bedding. Sedimentary structures are common in the Libby Creek Group and they show a consistent top-to-the-southeast orientation. Several major reverse faults have been recognized in the Libby Creek Group. One reverse fault follows the contact between the Libby Creek and Deep Lake Groups and another, the Lewis Lake fault of Houston and others (1968), roughly follows the contact between lower and upper parts of the Libby Creek Group successions. The evidence for such faults, in addition to breccia and fault scarps along their trace, is that massive quartzite sections of both the Medicine Peak and Sugarloaf abruptly pinch to zero thickness, then reappear in normal thickness along strike. In addition, major transverse faults in the lower part of the Libby Creek Group terminate at the inferred thrusts and are not present either in footwall rocks of the Deep Lake Group or younger hanging-wall rocks of the upper part of the Deep Lake Group. Tectonic attenuation of this magnitude, as well as differing structural styles between hanging wall and footwall suggest major movement, and we postulate that these reverse faults were originally thrust faults that have been subsequently rotated to steep attitudes.

Rocks of the Libby Creek Group and major faults are folded at the southwest limit of outcrop into a steeply plunging syncline, the French Creek fold of Houston and Parker (1963). This fold was interpreted (Houston and Parker, 1963) as a late flexural slip fold related to left-lateral strike-slip movement on the Mullen Creek-

Nash Fork shear zone. This interpretation is questioned on the basis of new evidence which will be discussed next.

CHEYENNE BELT

The Cheyenne belt is a series of fault blocks separated by mylonite zones in which the northernmost mylonite zone is the Mullen Creek–Nash Fork shear zone. The Cheyenne belt separates rocks of the Archean Wyoming craton, to the north, from Proterozoic rocks of the Colorado Front Range, to the south. It has been variously interpreted as a transform fault (Warner, 1978) and a Proterozoic suture associated with island arc-continent collision (Hills and Houston, 1979; Karlstrom and Houston, 1984).

The Cheyenne belt proper does not include the Libby Creek Group, but we will consider it a component of the belt because of structural similarities. The Cheyenne belt contains four major mylonite zones, which are, from north to south, the Mullen Creek–Nash Fork shear zone, the central mylonite zone, the southern mylonite zone, and the Rambler shear zone. All of these mylonite zones converge to the southwest into one major fault.

Duebendorfer and Houston (1987) have distinguished two blocks within the Cheyenne belt, the Bear Lake Block, bordered on the north by the Mullen Creek–Nash Fork shear zone and on the south by the central mylonite zone, and the Barber Lake Block, bordered on the north by the central mylonite zone and on the south by the southern mylonite zone. An additional unnamed block mapped by McCallum (1964) is bordered on the north by the southern mylonite zone and on the south by the east-northeast-trending Rambler shear zone.

On its southwestern extremity, the Bear Lake Block is a bimodal sequence of amphibolite and granite gneiss, and, in the northeast, it is composed of granodiorite-tonalite laminated gneiss, biotite granodiorite augen gneiss, and amphibolite. Texture, modal composition, and major-element chemistry suggest to Duebendorfer (1986) that the gneiss was intermediate plutonic rock of a calc-alkaline suite. The Bear Lake Block is amphibolite facies. The Barber Lake Block is composed of metasedimentary and metavolcanic rocks, metaperidotite, migmatitic augen gneiss, and garnet-biotite granite. Duebendorfer (1986) interprets the metasedimentary-volcanic sequence as representing graywacke, siltstone, shale, and intermediate to mafic volcanic rocks based on his analyses of mineralogy, major-element chemistry, and texture. The mineral assemblages are characteristic of the sillimanite-K-feldspar zone of the upper amphibolite facies (Duebendorfer and Houston, 1987). The garnet-biotite granite is peraluminous ($Al_2O_3/CaO+K_2O+Na_2O=1.2–1.4$) and corundum-normative and is viewed as having formed by partial melting of the associated rocks. Duebendorfer (1986) suggests that thermal conditions of metamorphism and anatexis were about 650 °C.

If we include the Libby Creek Group, the various blocks that compose the Cheyenne belt show a distinct increase in metamorphic grade from north to south, that is, from greenschist to amphibolite to upper amphibolite.

The earliest structure recognized in the Cheyenne belt is preserved in the Bear Lake and Barber Lake blocks in zones that were not strongly overprinted by later events. Subhorizontal isoclinal folds that deform compositional layering of unknown origin are present in the Bear Lake and Barber Lake blocks. Subhorizontal lineations defined by elongate microcline megacrysts are present locally in the Bear Lake Block. Duebendorfer and Houston (1987) have suggested that these early structures may have resulted from low-angle thrusting of unknown transport direction and dextral strike slip.

The most pervasive and widely developed structures of the Cheyenne belt are folds and lineations that define the northeast-striking, subvertical fabric of the belt and that overprint the structures described above. In the Libby Creek Group (Lookout Schist and French Slate primarily), bedding is transposed to develop a set of intrafolial, isoclinal folds and to develop an axial-planar, northeast-striking, subvertical schistosity. Fold hinges plunge steeply northeast. In the east-central Bear Lake Block, the transition from folds that plunge gently northeast to southwest and verge northwest to folds that steeply plunge can be seen in outcrop; in some outcrops, steeply plunging sheath folds have been observed (Duebendorfer and Houston, 1987, p. 561). Penetrative, down-dip, mineral lineations, which parallel the more steeply plunging fold axes, are developed by elongation of hornblende, biotite, and microboudinaged staurolite. Sense-of-shear indicators measured in rocks showing the down-dip lineation have a consistent south-side-up shear sense (Duebendorfer, 1986).

The mylonite zones (including the Mullen Creek-Nash Fork shear zone and Rambler shear zone) consist of mylonite, ultramylonite, and protomylonite. They have steep dips (60° to vertical) and range in width from 160 to 1,300 ft. They may have sharp or gradational boundaries. These mylonite zones appear to localize strain; the margins of blocks where they are in proximity to the mylonite zones usually exhibit a tightening of folds, dismemberment due to limb alteration, hinge arcuation, and steepening of fold hinge lines (Duebendorfer and Houston, 1987).

The French Creek fold of the Libby Creek Group has a steeply plunging axis that is approximately parallel to the fold axes and lineations discussed in the two preceding paragraphs. Karlstrom and Houston (1984, p. 433–435) have suggested that the French Creek fold developed in the same manner as the folds and lineations, that is, by "kinematic conditions characterized by subvertical extension (present orientation) related to dip-slip movements on the shear zone."

The last unequivocal Precambrian event documented in the Cheyenne belt is the development of semi-penetrative, subhorizontal lineations that are defined by chlorite and epidote fibers on mylonitic foliation (Duebendorfer and Houston, 1987). Kinematic indicators on both macroscopic and microscopic scale indicate right-lateral (dextral) shear. Related to the lineations are steeply plunging kink folds, which deform the northeast foliation of slate of the Libby Creek Group, rocks of the Bear Lake Block, and the Mullen Creek-Nash Fork shear zone, and show a consistent Z-shaped down-plunge profile and support the dextral sense of movement.

The proposed thrusting of Libby Creek Group rocks and blocks of the Cheyenne belt over autochthonous Deep Lake Group and older successions probably took place during the development of a fold and thrust belt as island arcs collided with a rifted margin. The rocks of the Libby Creek Group were probably metamorphosed at this time and an attempt was made to date this metamorphic event by Hills and others (1968). Whole-rock Rb-Sr methods were used to date the Lookout Schist (1,710±60 Ma) and French Slate (1,620±425 Ma). These dates have large margins of error but suggest an age of metamorphism between 1,700–1,800 Ma. A better approach has been used in the Sierra Madre where a granite (the Sierra Madre Granite of Divis, 1976), which intrudes the shear zone and is itself sheared, was dated by Premo (1983) using the U-Pb zircon method as 1,749±8 my and 1,763±6 Ma. Inasmuch as this granite must have been emplaced towards the end of the deformational events, 1,750 Ma must represent a minimum age.

In the Medicine Bow Mountains, quartz monzonite cuts the rocks of the Cheyenne belt and is affected by the later dextral event, suggesting that this event took place after emplacement of the quartz monzonite. The quartz monzonite has not been dated.

TECTONIC SIGNIFICANCE OF MAFIC INTRUSIONS

The Archean gneissic basement of the Medicine Bow Mountains and Sierra Madre is best shown on maps of Houston and others (1968) and Houston and Ebbett (1977) but includes the Overland Creek Gneiss and the unnamed granite of this map area. The basement gneiss is cut by numerous small dikes and sills and larger intrusive bodies of mafic and ultramafic composition. Some of the dikes and sills were folded at the time the northwest-trending structure developed in the gneissic terrain. If this northwest-trending structure is Archean as previously suggested, the folded dikes and sills may document at least two structural events in the Archean.

Dikes and larger mafic and ultramafic intrusions that strike northeast to east cut the foliation of the gneiss and the earlier dikes and sills. Because these mafic bodies parallel the structure of the Proterozoic successions, they may be Proterozoic. Only one of these mafic bodies has been dated, the Spring Lake mafic body of Shaw and others (1986). This intrusion cuts rocks of the Vulcan Mountain Metavolcanics, which are part of the Archean basement of the Sierra Madre. It is dated as 1,990±30 Ma by the Sm-Nd method (Shaw and others, 1986).

Dikes are uncommon in the Phantom Lake Metamorphic Suite, but large sills, which generally conform to but may crosscut the bedding, are common. None of the sills in the Phantom Lake have been dated. Similar sills are present in the Deep Lake Group, and one of these that intrudes the Cascade Quartzite of the Sierra Madre is dated as 2,092±4 Ma by the uranium-lead zircon method (Premo, 1983).

Large sills are less common in the Libby Creek Group except in the vicinity of the French Creek fold where a number of sills and one large intrusive body is present. Dikes and small sills as much as 100 ft thick are common in the Libby Creek Group, especially in the Medicine Peak Quartzite and Nash Fork Formation. A felsic differentiate of a mafic dike (the Gaps Intrusion) that intrudes the Lookout Schist has been dated as 1,900–2,150 Ma by the Rb-Sr whole-rock method (Hedge in Houston and others, in press).

The geochemistry of the mafic intrusives is limited to seventeen analyses of samples from the Archean Phantom Lake Suite and the Proterozoic Deep Lake Group and Libby Creek Group. There is a wide variation in the normative compositions of these rocks, but, in a general way, compositions of mafic intrusions of the Deep Lake and Phantom Lake fall in tholeiitic fields, and at least some of the analyses of Libby Creek rocks are more alkalic (Karlstrom and others, 1981a). Karlstrom and others (1981a) interpret the tholeiitic intrusions as rift-related. They suggest that they were emplaced during a protracted rifting event that occurred between 2,000–2,150 my. This is certainly an oversimplification; for example, the Proterozoic Spring Lake mafic body and other similar intrusions studied by Tracey and others (1986) have an olivine-bronzite andesite affinity, and they suggest that these rocks originated as "mantle-derived magmas representing high degrees of partial melting of somewhat depleted mantle beneath an active Early Proterozoic Island arc" (Tracey and others, 1986, p. 1265). The more alkalic intrusions have not been dated but may have been emplaced during the development of the foredeep, as exemplified by the French Slate.

In summary, we believe that the geological and geochemical data suggest the presence of several different tectonic settings for the Proterozoic mafic intrusives and probably more than one episode of intrusion. More detailed petrological, geochemical, and geochronological studies may help solve this problem.

PLATE TECTONIC MODEL

This model has been developed over the last decade and is based on information obtained from mapping, remapping, studies of depositional structures, and geochronological and geochemical investigations in both the Medicine Bow Mountains and Sierra Madre of southern Wyoming. The interested reader will find details of the models as they evolved in the following references: Hills and Houston, 1979; Karlstrom and others, 1981a; Karlstrom, Flurkey, and Houston, 1983; Karlstrom and Houston, 1984; Duebendorfer and Houston, 1987; and Houston and others, 1989.

LARAMIDE STRUCTURE

Laramide uplift, accompanied by the development of a complex set of west-dipping thrust faults on the east margin of the Medicine Bow Mountains, began in Late Cretaceous and continued into the early Eocene (Knight, 1953). Some of these thrust faults, which strike generally north to north-northwest, paralleling the mountain front, have associated tear faults which are reactivated Precambrian faults that strike northeast (Houston and others, 1968).

The main Laramide frontal fault of the northern Medicine Bow Mountains is the Arlington thrust fault that hinges about 12 mi north-northwest of Arlington. Its displacement increases from Arlington to a point a few miles south of Cooper Hill, which is east of the map area (Blackstone, 1987), and then decreases in displacement, passes beneath Corner Mountain, and becomes part of a splay of the Rambler fault system (see Blackstone, 1987, figs. 1–11, for details of this structure). Where the Arlington fault intersects the Precambrian, breccia and gouge zones can be recognized in well-exposed areas of the mylonitic and cataclastic fault zone.

The Corner Mountain fault that is west of and overrides the Arlington fault to the south (Blackstone, 1987) begins as a normal fault about 8 mi south of Arlington, where it is confined to the Precambrian. It becomes a thrust fault to the south-southeast where the hanging wall is offset by a series of northeast-striking tear faults that can be related to faults of Precambrian age. The principal tear fault is south of Corner Mountain where the Laramide fault intersects the southern mylonite zone (Blackstone, 1987, fig. 1).

The Corner Mountain fault is controlled, in part, by Precambrian structure. It parallels bedding of the Deep Lake Group but is perpendicular to bedding of the Libby Creek Group and foliation of the Cheyenne belt. All tear faults of the Corner Mountain thrust intersect Precambrian shear zones to the southwest. It has not been possible to identify the Laramide faults in the Precambrian shear zones, but all Precambrian faults and shear zones are overprinted by cataclasis and have local breccia and gouge that may represent the Laramide faults.

Another example of a fault that has characteristics similar to the Corner Mountain fault is the Middle Fork thrust fault, which is south of Centennial in T. 15 N., R. 78 W. This fault system has a tear fault that is related to the main Rambler shear zone, and the tear fault has modest displacement where the block is thrust to the east on the north-striking fault trace.

The French Creek fault (Houston and others, 1968) in T. 15 N., R. 80 and 81 W., is a major fault which is similar to the Middle Fork fault except that it is confined to Precambrian rocks. The northeast-striking part of the fault is interpreted as a tear fault, having dextral movement, that becomes a northwest-striking thrust fault. The dextral movement is based on the geometry of folded dikes north of the fault (west of the map area), as shown on the map in Houston and others (1968, pl. 1). The fault trace of the northeast-striking segment can be observed in the field. It is a zone of cataclasis and brecciation in gneiss, and it is a zone of brecciation where it cuts through quartzite. The brecciated quartzite is cemented by hematite. The character of the northwest-striking part of the fault has not been determined, but we suspect it is a thrust fault.

The French Creek fault is interpreted as a Laramide structure because of its similarity to Laramide structures on the east flank of the Medicine Bow Mountains and to Laramide faults of the northwest Medicine Bow Mountains outside of the map area (Houston and others, 1968, pl. 1). If we are correct in this interpretation, the brecciation and cataclasis, which is common in many of the major shear zones and mylonite zones, may be, at least in part, Laramide in age.

The Mill Creek fault, in the east ½ T. 16 N., R. 81 W., has a geometry similar to the French Creek fault. The northeast segment of this fault is Precambrian in age because the fault contains amphibolite of Precambrian age. Breccia and gouge along the amphibolite-granite contact indicates Laramide movement. The northwest-trending segment of the fault is a thrust fault that has thrust Precambrian rocks in contact with the Triassic Chugwater Formation (Houston and others, 1968, pl. 1). We interpret the Mill Creek fault as a reactivated Precambrian structure.

FAULTS OF TERTIARY AGE

Northwest-striking normal faults that offset the North Park Formation of Miocene and Pliocene age are common in the western Medicine Bow Mountains and Sierra Madre (Houston and others, 1968; Houston and Ebbett, 1977; Montagne, 1955). These faults, which may be the most northern manifestation of the Rio Grande Rift, are all outside the map area, with one exception, the Rock Creek Fault of Knight (1953). This fault strikes north to north-northwest from the vicinity of Centennial for a distance of about 12 mi, where it turns northeast and offsets the Arlington thrust. The Rock Creek Fault is a normal fault that is down-dropped on the west. A remnant of North Park Formation is preserved in the down-dropped block of the fault exposed in Rock Creek Canyon (Knight, 1953).

The strike of the Rock Creek Fault is parallel to a set of north- northwest-trending faults exposed on Corner Mountain. These faults are Precambrian in age, because one of these faults is intruded by a rhyolite dike that is probably a satellite dike of the Sherman Granite (Duebendorfer, 1986).

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