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Stratigraphy and depositional setting of the Proterozoic Snowy Pass Supergroup, southeastern Wyoming: Record of an early Proterozoic Atlantic-type cratonic margin

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ABSTRACT

Metasedimentary rocks in the Medicine Bow Mountains and Sierra Madre are divided into four groups. The >3-km-thick Phantom Lake Metamorphic Suite contains strongly deformed metavolcanic and metasedimentary rocks that are crosscut by late Archean granites. The >2.5-km-thick Deep Lake Group unconformably overlies the Phantom Lake Suite and late Archean granites and contains fluvial sediments, including radioactive quartz-pebble conglomerates, and glaciomarine deposits. Both successions are intruded by large sills of tholeiitic gabbro. The 4.5-km-thick lower Libby Creek Group is inferred to be in thrust-fault contact with older units and contains sediments recording transgressions and regressions across a macrotidal delta. This succession is intruded by the 2,000-m.y.-old Gaps Intrusion and comagmatic tholeiitic to weakly alkalic dikes. This 3-km-thick lower Libby Creek Group is bounded by a thrust fault below and by the Cheyenne Belt above and contains carbonates and marine slates. The early Proterozoic Deep Lake, lower Libby Creek, and upper Libby Creek Groups collectively are named the Snowy Pass Supergroup.

Lithologies and stratification sequence in the well-preserved Medicine Bow Mountain section suggest transgressive, miogeoclinal sedimentation during the early Proterozoic. Paleocurrent data indicate that fluvial, then deltaic, sedimentation of the Deep Lake and lower Libby Creek Groups took place on a southwest-dipping paleoslope, parallel to the inferred south cratonic boundary of the Wyoming Province. This and a few west-directed paleocurrents suggest a continental or microcontinental block to the south, bounding sedimentation in a northeast-elongated basin. A rift setting for deposition of these units explains the transgressive character of the sediments, the deltaic sedimentation with paleocurrents parallel to the cratonic boundary, and the 120° bend in the Cheyenne Belt between the Medicine Bow Mountains and the Sierra Madre. The upper Libby Creek Group is interpreted to represent open marine conditions following separation of the two continental blocks. Tholeiitic sills in the Deep Lake Group and tholeiitic to weakly alkalic dikes in the Libby Creek Group are thought to be related to basaltic igneous activity associated with compound early Proterozoic rifting between 2,300 and 2,000 m.y. ago.

INTRODUCTION

Precambrian rocks in the cores of the Medicine Bow Mountains and the Sierra Madre range in age from more than 2,500 m.y. to about 1,400 m.y. and preserve the most complete and best-exposed record in the western United States of early Proterozoic geologic history. This paper has two main objectives: (1) to present a synthesis of new and published data on the stratigraphy and sedimentology of the early Proterozoic (2,500–1,600 m.y. B.P.) Snowy Pass Supergroup in the Medicine Bow Mountains and Sierra Madre; and (2) to present a plate-tectonic model for deposition of the metasedimentary rocks. The Precambrian time scale used here is that of Harrison and Peterman (1980).

As summarized by Hills and Houston (1979) and Karlstrom and others (1981), several aspects of the sedimentary and tectonic record in southern Wyoming suggest cratonic margin sedimentation in the early Proterozoic followed by continent–island arc collisional orogenesis culminating about 1,700 m.y. ago. The best lines of evidence are that early Proterozoic metasediments of the Snowy Pass Supergroup have many similarities to miogeoclinal, passive, plate-margin successions and that these low-grade miogeoclinal metasediments are juxtaposed against upper amphibolite-facies, late early Proterozoic, eugeoclinal rocks and synorogenic intrusives across a major shear zone. This shear zone, called the Mullen Creek–Nash Fork shear zone in the Medicine Bow Mountains (Houston and McCallum, 1961; Houston and others, 1968), is of regional proportions, having been traced west into the Sierra Madre (Graff, 1978, 1979) and east into the Laramie Mountains (Graff and others, 1981). The shear zone was referred to as the Cheyenne Belt by Houston and others (1979b) and that terminology is retained here. Profound geochronologic discontinuity across the Cheyenne Belt, as shown by the absence of Archean rocks south of the boundary, strengthens the argument that the shear zone marks a major crustal boundary separating the Wyoming Archean Province to the north (Engel, 1963; Houston and others, 1968; Condie, 1976) from younger, accreted Proterozoic island-arc terranes to the south (Hills and Houston, 1979). This paper examines the Proterozoic stratigraphic and sedimentary record in southeastern Wyoming in terms of this plate-tectonic model, with emphasis on the pre-collisional history of the southern margin of the Wyoming Province.

GEOLOGIC SETTING

The Medicine Bow Mountains and Sierra Madre are north-trending anticlines which...
were uplifted during the Laramide orogeny (Houston and others, 1968). As mentioned above, Precambrian rocks in the core of the ranges, shown in Figure 1, are divided into two geologic and geochronologic provinces by the northeast-trending Cheyenne Belt, a zone consisting of mylonites and augen gneisses which reaches a width of 1 km in the western Medicine Bow Mountains. Additional zones of mylonitic rocks are common immediately south of the main shear zone in the eastern Medicine Bows in a belt some 6 km wide.

Rocks north of the Cheyenne Belt consist of an Archean gneissic terrane which is overlain nonconformably by (and in fault contact with) more than 13 km of quartz-rich metasedimentary rocks. Although the ages of these metasedimentary rocks are not precisely known, they range in age from late Archean (more than 2,500 m.y.) to about 2,000 m.y. We have divided the metasedimentary rocks into three successions: the Phantom Lake Metamorphic Suite, which is intruded by tonalites and granodiorites of Archean age; the early Proterozoic Deep Lake Group, which unconformably overlies Phantom Lake Suite metasediments and Archean granitic rocks (Houston and others, 1968; Karlstrom and Houston, 1979a, 1979b); and the early Proterozoic Libby Creek Group (Blackwelder, 1926; Houston and others, 1968), which is crosscut by the 2,000-m.y.-old Gaps Intrusion (Carl Hedge, 1979, personal commun.; Houston and others, in press). The Libby Creek Group is now interpreted to be in thrust-fault contact with older units (Lanthier, 1979; Karlstrom and others, 1981).

Metasedimentary rocks south of the shear zone are mainly hornblende and quartz-feldspathic gneisses with minor sillimanite gneiss and calc-silicate (McCallum, 1964; Hills and Houston, 1979). These gneisses are believed to be partly paragneisses (Houston and others, 1968) that are now complexly intruded by or intercalated with a variety of intrusive rocks, most notably layered gabbroic complexes, quartz-diorite, and synorogenic granodiorite. Details of a possible sedimentary origin of the gneisses are uncertain because of the high degree of deformation and amphibolite-facies metamorphism that has obliterated primary features. However, the gneisses are compositionally similar to basic to intermediate volcanics and volcaniclastic sediments, in marked contrast to the mature siliciclastic metasediments north of the shear zone. Available geologic and geochronologic data are consistent with the interpretation that paragneisses south of the shear zone are island-arc-derived sediments and volcanics that were complexly intruded by synorogenic plutonic rocks, strongly deformed, and metamorphosed to amphibolite grade during continent-island-arc collision about 1,700 m.y. ago (Hills and Houston, 1979). The age of gneisses south of the shear zone is reasonably well-constrained by geochronologic data; gneisses are known to be older than the 1,700- to 1,800-m.y.-old intrusives that cut them (Hills and others, 1968) and are believed to be younger than about 1,800 m.y.

Figure 1. Generalized geologic map of Precambrian rocks of the Medicine Bow Mountains and Sierra Madre, Wyoming.
Paleocurrent analysis in deformed rocks is fraught with difficulties. For simplicity, we have assumed that folds developed by the flexural-slip mechanism, so we have "restored" bedding to inferred prefolding orientations by unfolding about the plunge of local fold axes (Ramsay, 1961). These local fold axes, for individual subareas, were determined statistically on contoured equal-area projections of poles to bedding. This assumption of flexural-slip mechanism seems reasonable for the Deep Lake Group, which consists of well-bedded quartzites that are folded into shallow-plunging, upright, concentric folds and which is believed to be autochthonous.

The assumption of flexural-slip folding in the Libby Creek Group is more questionable. We have postulated thrusting within the Libby Creek Group but have chosen to neglect the possibility of large internal strains associated with this thrusting that could have rotated the current indicators toward the thrusting direction. Our reasons for neglecting this possibility were several. First, the abundance of primary features in the quartzite units and the fact that the majority of cross-beds have normal sedimentary inclinations of 20° to 30° (see Fig. 3 below) suggest that shear folding was not the dominant folding mechanism in these units. According to Ramsay (1961, p. 98), cross-bedding would be obliterated during shear folding in situations like that seen in the Libby Creek Group, where the limbs of the folds are subparallel to the axial plane. Furthermore, the spherical shapes of pebbles in conglomerates and the apparent lack of deformation of stromatolites in the carbonate unit indicate that internal finite strains in these units are small. We believe that the large strains associated with thrusting may have been concentrated in the less competent rock types and in narrow zones within the quartzites, so that reorienting paleocurrents by simply unfolding about the plunge of folds may provide a useful approximation of original paleocurrents within the competent units.

The plunge of folds in the Libby Creek Group varies from horizontal to vertical within the northeast-striking axial plane, a common distribution in thrust terranes. We chose to reorient cross-beds about a horizontal fold axis in the Libby Creek Group because all of the paleocurrent data were taken from the central Medicine Bow Mountains, where the structure is that of a steeply southeast-dipping homocline (some 25 km in length and 6–7 km in stratigraphic thickness) and where shallow fold axes predominate. In areas where subvertical fold axes predominate (for example, the French Creek Syncline in the southwestern Medicine Bow Mountains), sedimentary features have been for the most part obliterated. Paleocurrent directions from the Phantom Lake Suite should be viewed with extreme caution. The rocks are folded into tight to isoclinal folds and have been multiply deformed; there is not yet enough structural data available to accurately restore beds to prefolding positions. However, although directions are suspect, the paleocurrent data still provide useful information with regard to distribution of currents, whether unimodal or polymodal, within individual outcrop areas.

**ARCHEAN BASEMENT ROCKS IN SOUTHEASTERN WYOMING**

**Quartz-Feldspathic Gneisses**

The oldest rocks in southeastern Wyoming are found within a heterogeneous assemblage of quartzo-feldspathic gneisses, hornblende gneisses, biotite gneisses, and quartzites in the western Medicine Bow Mountains and the northeastern Sierra Madre (labeled Agn in Fig. 1). Combined evidence from the Medicine Bow Mountains (Hills and Houston, 1979), the Sierra Madre (Divis, 1976, 1977), the Granite Mountains to the north (Peterman and Hills, 1976) suggests that southern Wyoming's gneissic terranes record a complex history, involving deposition of mafic to intermediate volcanic and volcanoclastic protoliths, perhaps as early as 2,950 m.y. ago, metamorphism of the protoliths accompanying intrusion of tonalitic magmas about 2,700 m.y. ago, and metamorphism and anatexis of gneisses about 2,500 m.y. ago. We consider at least part of the gneissic terrane to have been basement that supplied detritus during deposition of the Proterozoic Snowy Pass Supergroup.

**Phantom Lake Metamorphic Suite**

The Phantom Lake Suite is a volcano-sedimentary succession, more than 3 km thick, that contains about 60% volcanic and volcanoclastic rocks and 40% mature siliciclastic rocks. It is believed to be younger than most of the gneisses, although the two terranes are in direct contact only in the western Sierra Madre, and contact relationships are generally obscured by later intru-
LATE ARCHEAN DEFORMATION AND EARLY PROTEROZOIC UNCONFORMITY

Archean rocks in southeastern Wyoming were deformed and metamorphosed near the end of the Archean. Quartzo-feldspathic gneisses in the Medicine Bow Mountains yield Rb-Sr whole-rock ages of 2,500 to 2,600 m.y., believed by Hills and Houston (1979) to be metamorphic ages. The synorogenic Baggot Rocks Granite yields primary Rb-Sr ages of about 2,400 to 2,500 m.y. (Hills and Houston, 1979; Divis, 1977). F$_1$ folding in Archean metasedimentary rocks of the Phantom Lake Metamorphic Suite appears to have been penecontemporaneous with intrusion of late Archean granites in the northern Medicine Bow Mountains (Karlstrom and others, 1981).

This late Archean deformation episode was followed by a period of relative tectonic quiescence during which Archean rocks were uplifted and eroded. Evidence for this is a regional unconformity between Archean "basement" rocks and the Deep Lake Group. This unconformity is well-exposed in the cirque walls of Crater Lake in the north-central Medicine Bow Mountains, where tightly folded, east-west-trending volcanics of the Phantom Lake Suite are overlain by north-trending, gently west-dipping quartzites of the Magnolia Formation. The unconformity is also evident from map relationships in the northern Medicine Bow Mountains, where basal conglomerates of the Magnolia Formation alternately overlie late Archean granites and various units of the Phantom Lake Suite. Also, Magnolia conglomerates locally contain clasts of each of these underlying units. Together, these data show that the Deep Lake Group is an early Proterozoic metasedimentary succession, younger than 2,500 m.y., that was deposited with angular unconformity on an uplifted and eroded Archean basement.

STRATIGRAPHY AND SEDIMENTARY FEATURES OF EARLY PROTEROZOIC ROCKS IN THE MEDICINE BOW MOUNTAINS

The early Proterozoic section in the Medicine Bow Mountains is more complete and

Figure 2. Stratigraphic column of metasedimentary rocks in the Medicine Bow Mountains.
better-exposed than the section in the Sierra Madre and forms the basis for our stratigraphic interpretations. We first discuss the stratigraphy of the Snowy Pass Supergroup in the Medicine Bow Mountains and then, briefly, correlative units in the Sierra Madre.

As in most complex geologic terranes, stratigraphic nomenclature for metasedimentary rocks in the Medicine Bow Mountains has been modified many times over the years in response to increasingly detailed mapping studies. Table I compares nomenclatures used by previous workers to nomenclature used in this paper. The following section describes each stratigraphic unit of the Snowy Pass Supergroup in the northern Medicine Bow Mountains from oldest to youngest. Figure 2 summarizes our stratigraphy.

**Deep Lake Group**

The Deep Lake Group was defined by Karlstrom and Houston (1979a, 1979b) to include six formations (Table 1). However, we now consider the uppermost unit, the Rock Knoll Formation, to be part of the overlying Libby Creek Group because it is in depositional contact with the overlying Headquarters Formation and is now interpreted to be entirely in fault contact with the underlying Vagner Formation. The basal unit of the Deep Lake Group, the Magnolia Formation, contains important occurrences of uranium-bearing quartz-pebble conglomerate, which is similar to uranium ores of the Huronian Supergroup near Elliot Lake, Ontario. These occurrences are discussed in detail by Houston and others (1977, 1979b), Houston and Karlstrom (1980), and Karlstrom and others (1981).

**Magnolia Formation.** The Magnolia Formation is composed of two members: a basal, radioactive Conglomerate Member, containing muscovitic quartz-pebble conglomerates and arkosic-matrix paraconglomerates, and a Quartzite Member, containing more mature subarkose and arkosic quartzites and quartz-granule conglomerates (see Fig. 4 below).

The Conglomerate Member crops out discontinuously in the cores of anticlines and the limbs of synclines throughout the 35-km-long outcrop area of the Deep Lake Group. Most commonly, the unit occurs as lenticular zones dominated by polymictic paraconglomerate with interbeds of quartz-pebble conglomerate and coarse-grained arkosic quartzite. The matrix of the paraconglomerates is micaceous arkose and subarkose. The clasts are poorly sorted, range as high as tens of centimetres in diameter, are rounded, and consist of quartz, quartzite, schist, mafic volcanic rocks, and granite.

In the Onemile Creek area, at the northern limit of the Magnolia outcrop, the Conglomerate Member is strongly radioactive (with thin zones containing as much as 1,600 ppm U) and is dominated by muscovitic quartz-pebble conglomerates and coarse-grained muscovitic arkoses. The quartz-pebble conglomerates are pyritic and occur in lenses ranging from a single pebble thick to compound zones 20 m thick.

The Conglomerate Member grades laterally and up-section into the Quartzite Member. This unit consists of trough-cross-bedded, coarse-grained to granular

**TABLE I. COMPARISON OF STRATIGRAPHIC NOMENCLATURES FOR METASEDIMENTARY ROCKS IN THE MEDICINE BOW MOUNTAINS**

<table>
<thead>
<tr>
<th>This Paper</th>
<th>Karlstrom and Houston (1979a, b), Lantier (1979)</th>
<th>Houston and Others (1968)</th>
<th>Blackwelder (1926)</th>
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<tr>
<td>French Slate</td>
<td>Towner Greenstone</td>
<td>NASH FORK FORMATION</td>
<td>NASH FORK FORMATION</td>
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<td>SUGARLOAF QUARTZITE</td>
<td>Lockout Schist</td>
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<td>Lookout Schist</td>
<td>Medicine Peak Quartzite</td>
<td>Lookout Schist</td>
</tr>
<tr>
<td>Lookout Schist</td>
<td>Headquarters Formation</td>
<td>Sendek Quartzite</td>
<td>Headquarters Schist (Included Units Now Mapped As Vagner)</td>
</tr>
<tr>
<td>Headquarters Formation</td>
<td>ROCK KNOll FORMATION</td>
<td>Vagner Formation</td>
<td>Marble</td>
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<td>Rock Knoll Formation</td>
<td>Cascade Quartzite</td>
<td>Cascade Quartzite</td>
<td>Quartzite</td>
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<td>Campbell Lake Formation</td>
<td>Campbell Lake Formation</td>
<td>Metaconglomerate</td>
</tr>
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<td>Lindsey Quartzite</td>
<td>Lindsey Quartzite</td>
<td>Metaconglomerate Suite</td>
</tr>
<tr>
<td>Lindsey Quartzite</td>
<td>Magnolia Formation</td>
<td>Magnolia Formation</td>
<td>Metaconglomerate Suite</td>
</tr>
<tr>
<td>Magnolia Formation</td>
<td>Blackwelder (1926)</td>
<td>Blackwelder (1926)</td>
<td>Deep Lake Metamorphic Suite</td>
</tr>
</tbody>
</table>

| Archean | | | |
| Phantom Lake Metamorphic Suite | | | |
| Conical Peak Quartzite | | | |
| Colberg metavolcanics | | | |
| Bow Quartzite | | | |
| Rock Mountain conglomerate | | | |
| Stud Creek volcanoclastics | | | |
| Overland Creek gneiss | | | |
quartzites that vary in composition from micaceous subarkose to micaceous arkose. The Quartzite Member is generally less micaceous, less arkosic, finer grained, and less radioactive than the Conglomerate Member.

We interpret the Magnolia Formation to be a fluvial succession for a variety of reasons. The unit overlies rocks of the Phantom Lake Metamorphic Suite with angular unconformity and contains detritus from the Phantom Lake Suite and from late Archean granites. Both indicate a period of uplift and erosion prior to Magnolia deposition and imply that subaerial conditions dominated at least in the early stages of deposition. Lithologic characteristics, the lenticular distribution, and the internally variable character of the Conglomerate Member also suggest fluvial deposition. The unit is poorly sorted, is relatively immature, and contains complexly intercalated sand, gravel, and boulder lithofacies (including matrix-supported paraconglomerates that may be debris flows), as well as radioactive minerals that are associated with the base of pebble horizons and are believed to represent stream placer accumulations. All suggest high-energy fluvial deposition (Rust, 1979). We envision a paleogeography characterized by fairly high relief and fault-bounded highlands, where paraconglomerates were deposited in alluvial-fan debris flows, and radioactive quartz-pebble conglomerates were deposited in associated high-energy braided streams.

The Quartzite Member is also interpreted to be fluvial in origin because of relatively poor sorting, abundant trough cross-bedding, and gradational (fining-upward) relationships to the underlying conglomerates. We interpret the quartzites to represent deposition in a well-developed and laterally extensive braided river system because of the lateral persistence of the unit over an outcrop length of about 35 km and the change to less micaceous and arkosic quartzites from northeast to southwest, down the paleoslope.

Paleocurrent measurements (Fig. 3, which appears as a foldout with this article), although almost entirely from the Quartzite Member, show a somewhat dispersed polymodal pattern dominated by southwesterly-directed currents. This type of dispersed polymodal pattern is often cited as evidence for intertidal deposition (Herdendorf, 1977; Klein, 1970) but may also represent distributions in braided-stream environments (Ore, 1964; Williams, 1971) or fluvial deposition combining alluvial-fan and braided-river deposition. For individual areas (Fig. 3), some paleocurrent roses show unimodal patterns; others show distinctly polymodal patterns. If the polymodal patterns are real (that is, if they do not reflect sampling problems or measurement problems associated with distinguishing trough from planar cross-beds in small, discontinuous outcrops), they may reflect a southwesterly-directed major river system, bounded on the northwest by areas of high relief, in which alluvial fans were developed that transported sediment in southeasterly directions into the trunk river system. By this explanation, the conglomerates of the lower Magnolia Formation represent the basal portions of these alluvial-fan deposits. The wide distribution of conglomerate lenses and consistently coarse grain sizes of these conglomerates in a downcurrent direction (southwest) in the Medicine Bow Mountains are explained in terms of multiple southeastward-directed fan systems built on fault-bounded highlands to the northwest (see Fig. 9B below). This interpretation remains tentative because cross-bedding is rare in the conglomerates themselves and their paleocurrent distributions are not well known.

Lindsey Quartzite. The Lindsey Quartzite includes trough–cross-bedded quartzarenite and subarkose (Fig. 4) that gradationally overlie the Quartzite Member of the Magnolia Formation. Lindsey quartzites are generally medium grained but also contain scattered pebbles (as much as 1 cm in diameter) along planar foreset beds or in small scours, as well as thin phyllitic layers and partings. The thickness of the Lindsey Quartzite is about 410 m in the central Medicine Bow Mountains, and the unit either pinches out or is unconformably onlapped by the Cascade Quartzite to the north. The Lindsey Quartzite is locally anomalously radioactive, especially near its northern outcrop limit.

Trough cross-bedding is well-preserved in the Lindsey Quartzite, and ripple marks are present locally. The paleocurrent distribution shown in Figure 3 is strongly unimodal, with currents directed southwest and a mean direction (229°) very similar to that of the Magnolia Formation. This type of unimodal pattern suggests fluvial deposition (Allen, 1967; Potter and Pettijohn, 1977). We interpret the Lindsey Quartzite to be a fluvial unit on this basis, as well as on the basis of its gradational relationship to the Magnolia Formation, its fining-upward sets, its abundance of trough cross-beds, and its scattered pebbles and phyllite parts that may represent lag gravels and thin overbank sediments, respectively.

Cascade Quartzite. The Cascade Quartzite is a thin (as much as 85 m) and contains paraconglomerate-phyllite sequence that may represent a debris flow of some type and has been interpreted to be glacial in origin (Karlstrom and Houston, 1979a, 1979b), although we have no conclusive evidence of a glacial origin and the unit might also represent alluvial-fan debris flows (Bull, 1972).

Paleocurrent measurements from abundant planar and trough cross-bedding in the Cascade Quartzite show a unimodal distribution about a west-southwest–directed mean paleocurrent (248°). The dispersion of paleocurrent vectors is appreciably smaller (L is larger, and variance is smaller) than for either the Lindsey Quartzite or the Magnolia Formation. This type of small-variance, unimodal paleocurrent pattern is considered to be one characteristic of fluvial paleocurrent patterns (Potter and Pettijohn, 1977).

We have interpreted the Cascade Quartzite in previous papers (Karlstrom and Houston, 1979a, 1979b) to be a fluvial unit on the basis of the low-variance, unimodal paleocurrent distribution; the presence of both planar and trough cross-bedding (interpreted to represent bars and migrating
SUMMARY OF STATISTICAL PARAMETERS

<table>
<thead>
<tr>
<th>Formation</th>
<th>n</th>
<th>x</th>
<th>s</th>
<th>s²</th>
<th>R</th>
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<td>51</td>
<td>236</td>
<td>573</td>
<td>3282</td>
<td>57.7x10⁻³</td>
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<td>287</td>
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<td>455</td>
<td>2084</td>
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<tr>
<td>Headquarters and Heart Fms</td>
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<td>225</td>
<td>85</td>
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<td>123</td>
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<td>64</td>
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<td>489</td>
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<td>Total Deep Lake Group</td>
<td>466</td>
<td>229</td>
<td>50</td>
<td>670</td>
<td>494</td>
</tr>
</tbody>
</table>

- **n**: number of measurements
- **x**: vector mean azimuth of paleocurrent, in degrees
- **s**: standard deviation, in degrees (Mardia, 1972)
- **R**: Rayleigh test of significance (Curry, 1956); values smaller than 1x10⁻⁸ indicate that the distribution is non-random at the 99% confidence level.

**Percentage of measurements in each 30° segment for for synoptic rose diagrams (left), showing vector mean.**

**Percentage of measurements in each 30° segment for rose diagrams on map (where n>7), showing vector mean, number of measurements and geologic unit (keyed to Explanation).**

**Figure 3.**
Paleocurrent data from Precambrian metasedimentary rocks, Medicine Bow Mountains, Wyoming.

**Explanation**
- Gap, Transhormite: 2000 m.y.
- Mafic intrusive rocks
- Upper Libby Creek Group, [Noosh Fork Fm., Tower Greensite, French Site]
- Sugarloaf Quartzite
- Lookout Schist
- Medicine Peak Quartzite
- Heart Formation
- Lookout Schist
- Medicine Peak Formation
- Rock Knoll Formation
- Medicine Peak Quartzite
- Rock Knoll Formation
- Medicine Peak Quartzite
- Rock Knoll Formation

**Archean granitic gneisses - 2500 m.y.**
- Intrusive contact
- Conical Peak Quartzite
- Medicine Peak Formation
- medicine Peak Formation
dunes in a river system); the layers of well-rounded, well-sorted pebbles considered to be lag gravels; and the unconformities at the base and top of the Cascade Quartzite that suggest subaerial conditions before and after Cascade deposition. However, none of these features is restricted to fluvial environments, and the lateral continuity of the unit (a strike length of more than 50 km in the Medicine Bow Mountains and about 30 km in the Sierra Madre to the west), the mature composition, the relative consistency of maximum pebble sizes in a downcurrent direction, and the presence of one bimodal (north-south) paleocurrent distribution in the northeastern Medicine Bow Mountains (Fig. 3) might also be cited as evidence supporting marine or deltaic deposition. The unimodal paleocurrent distribution for the entire unit and the mature composition are both similar to the Medicine Peak Quartzite (Fig. 3), which we consider to contain both shallow-marine and deltaic deposits (discussed below). At present, our understanding of the Cascade Quartzite permits either an alluvial-plain or deltaic interpretation; or, both types of sediments may be present.

**Vagner Formation.** The Vagner Formation is a paraconglomerate-marble-phyllite-quartzite succession (as much as 800 m thick) that unconformably overlies the Cascade Quartzite. This unit originally was included as part of the Headquarters Schist (Table 1) because of similarities between paraconglomerates in the Vagner and Headquarters Formations and the proximity of these units to one another in the central Medicine Bow Mountains. However, detailed mapping has shown that the paraconglomerate-marble succession of the Vagner Formation is present in the northeastern Medicine Bow Mountains and is part of the Deep Lake Group. The proximity of Vagner and Headquarters paraconglomerates is now interpreted to be related to movement on the Reservoir Lake Fault, the thrust that separates the Deep Lake and Libby Creek Groups (Fig. 1).

The Vagner Formation is an important unit in the Medicine Bow Mountains, from several perspectives. Stratigraphically, it represents the early stages of a major transition in sedimentation: from dominantly fluvial deposition in the Deep Lake Group to dominantly marine deposition in the Libby Creek Group. Structurally, the marbles and the phyllites in the unit were some of the most ductile rocks in the sequence, localizing strain during major Proterozoic thrust faulting and during Laramide uplift so that the unit is intimately associated with major faults. In the central Medicine Bow Mountains, the Vagner Formation forms the footwall of the Reservoir Lake Fault over most of its length; in the northeastern Medicine Bow Mountains, the unit is in the hanging wall of the Arlington Thrust Fault of Laramide age.

The basal unit of the Vagner Formation is a polymictic paraconglomerate (or diamictite) with angular and subangular clasts of granite, quartzite, and mafic schist in a subarkosic matrix (Fig. 4). The unit is thin (average thickness about 300 m), but laterally persistent, and is interpreted to be a glaciomarine conglomerate on the basis of the presence of dropstone clasts, poor sorting, subangular clasts, faint stratification in some conglomerates, interlayered varved phyllites, and the geochemistry of the sand-sized matrix of the conglomerates, which is similar to that of the Gowganda tillites of the Huronian Supergroup and of other early Proterozoic glacial (?) deposits in North America (Sylvester, 1973; Young, 1970, 1973; Houston and others, 1981). Marble and varved phyllite of the Vagner
Formation provide support for a marine depositional setting.

Limited paleocurrent data from the Vagner Formation (Fig. 3) show a change to more westerly directed transport (mean = 256°). One possible explanation for this change involves a highland to the east that could supply detritus. This explanation fits well with our concept of sedimentation in a rift basin during the early Proterozoic (discussed in detail below).

**Lower Libby Creek Group**

The name “lower Libby Creek Group” is used here to refer to the dominantly marine siliciclastic units south of the Reservoir Lake Fault and stratigraphically below the Nash Fork Formation (Table 1). This subdivision, upper versus lower Libby Creek Groups, is introduced for several reasons. First, the lithologies in these two successions are quite different. Whereas the lower succession is predominantly siliciclastic, the upper Libby Creek Group contains dolomite, volcanogenic rocks, and black slate. This major change in lithology indicates very different conditions for sedimentation for the two successions. Second, new map interpretations place a major fault, the Lewis Lake Fault, between the Sugarloaf Quartzite and the Nash Fork Formation on the basis of the disappearance of the Sugarloaf Quartzite in the central part of its outcrop belt, the abrupt thinning of the Medicine Peak Quartzite farther northeast, and various breccia zones along the contact (Lanthier, 1978, p. 26). As discussed by Karlstrom and others (1981), we interpret the fault to be a rotated thrust fault, which implies that the upper Libby Creek Group is allochthonous or para-autochthonous with respect to the lower Libby Creek Group and that the two units may have been deposited in very different tectonic and sedimentary environments. Third, although the exact age relationship between the two successions is not yet clear, there is one indication that the upper Libby Creek Group may be appreciably younger than the lower Libby Creek Group. The 2,000-m.y.-old Gaps Intrusion penetrates the lower succession but not the upper, and regional correlations to other metasedimentary successions in North America (Houston and Karlstrom, 1980) suggest that the upper Libby Creek Group, with its thick stromatolitic dolomites, may correlate better with the 1,900- to 2,100-m.y.-old Marquette Range Supergroup of Minnesota, whereas the lower Libby Creek Group, with its glacial (?) units and aluminous quartzites, may correlate better with the 2,150 (+)-m.y.-old Huronian Supergroup of Ontario (Young, 1973; Houston and Karlstrom, 1980).

We consider the lower Libby Creek Group to represent deltaic deposition in a deepening fault-bounded basin near the south margin of the Wyoming Archean craton. As mentioned above, the two lowest units of the lower Libby Creek Group, as well as the uppermost unit of the Deep Lake Group, are interpreted to be glaciomarine sediments that record a change from mainly continental sedimentation in the Deep Lake Group to mainly marine sedimentation in the Libby Creek Group.

**Rock Knoll Formation.** The Rock Knoll Formation is a quartzite unit that crops out in two places on the south side of the Reser-
voir Lake Fault. Its maximum exposed stratigraphic thickness is 380 m, but the base of the unit in both outcrops is truncated against the Reservoir Lake Fault.

The predominant lithology is a gray, medium-grained, plagioclase-rich arkose (Fig. 5). This unit also contains phyllitic layers and partings up to 30 cm thick and conglomerates as much as 1 m thick. The conglomerates contain quartz, quartzite, and granite clasts. The lithology of the Rock Knoll is very different from underlying Deep Lake Group quartzites, in that plagioclase is overwhelmingly the dominant feldspar (compare Figs. 4 and 5). Plagioclase is also the dominant feldspar in part of the Vagner Formation and throughout the lower Libby Creek Group.

Paleocurrent data from outcrop areas of the Rock Knoll Formation indicate westerly directed paleocurrents (mean = 280°). These westerly current directions are unique to the Rock Knoll and Vagner Formations and may be evidence for an eastern source area that apparently was underlain by plagioclase-rich metavolcanic and plutonic rocks, quartzites, and some granites.

Sedimentary structures in the Rock Knoll Formation include ripple marks, planar cross-bedding, and clay galls. We postulate glacial-related fluvial and deltaic deposition for the Rock Knoll Formation because of its stratigraphic position and its similarity in composition with the Vagner and Headquarters Formations. The quartzites of the Rock Knoll Formation appear to represent a period of glacial retreat between glaciomarine sedimentation episodes represented by the underlying Vagner and overlying Headquarters Formations.

**Headquarters Formation.** The Headquarters Formation, as redefined by Karlstrom and Houston (1979a) and Lanthier (1979), includes a lower member (350 m thick) composed of lenticular paraconglomerates, quartzites, and schists and an upper member (300 m thick) containing laminated schists and phyllites. Since Blackwelder in 1926, the Headquarters Formation has been interpreted to be glacial or glaciomarine in origin (Houston and others, 1968; Sylvestre, 1973; Karlstrom, 1977; Lanthier, 1979; Kurtz and Anderson, 1979; Houston and others, 1981). The unit has been correlated with other early Proterozoic glacial (?) paraconglomerates in North America, most notably the Gowganda Formation of the Huronian Supergroup (Young, 1970, 1973; Houston and others, 1979b; Houston and Karlstrom, 1980).

Petrographically, the unit is quite heterogeneous. The lower member contains several lenses of paraconglomerate (or diamicite) composed of granite, quartzite, and schist clasts (average size 4–6 cm but ranging as high as about 1 m in diameter) in a poorly sorted matrix of sand- and silt-sized quartz, plagioclase, K-feldspar, rock fragments, and micas. At the locality described by Blackwelder (1926), there are three stacked paraconglomerate units separated by quartzites. In other areas, one or multiple paraconglomerates in the lower member appear to be complexly intercalated with quartzites and schists. Quartzites are plagioclase-rich arkoses, very similar in composition to the Rock Knoll quartzites (Fig. 5). The upper member of the Headquarters Formation is a biotite, chlorite, quartz phyllite with laminations formed by alternating quartz-rich and mica-rich layers.

Sedimentary structures in the Headquarters Formation include small-scale planar and trough cross-bedding, laminations in both paraconglomerate and phyllite units, dropstone structures in paraconglomerates, and climbing ripples. In one outcrop, Bouma sequences were observed associated with paraconglomerates. The paraconglomerates themselves are some of the most interesting sedimentary features of the unit. They are massive to slightly stratified, generally nongraded, and poorly sorted; they contain isolated to poorly packed subangular to rounded clasts, a few of which appear to depress underlying strata but to be covered by overlying strata, suggesting ice-rafted dropstones (Sylvestre, 1973). These characteristics of the paraconglomerates are similar to descriptions of other glaciomarine paraconglomerates (Young, 1970) and the overall geochemistry of the Headquarters paraconglomerates is very similar to that of the glacial Gowganda Formation of the Huronian Supergroup (Young, 1973). These, in combination with the association of the paraconglomerates with laminated phyllites (varved metasiltstones), the rapid lateral facies changes in the lower member, and the appearance of multiple conglomerate lenses, are all consistent with a glaciomarine depositional setting. Prodeltaic mudflows and turbidites might have most of the same features but would not be expected to contain dropstones. Kurtz and Anderson (1979) suggested, by analogy to the Antarctic continental-margin sediments, that deposition of the Headquarters Formation took place at some distance from the ice sheet and on at least a moderate slope, as suggested by evidence for turbidity flow in the paraconglomerates, the presence of laminated pebbly and nonpebbly argillites that both occur on the continental slope off Antarctica, the scarcity of the paraconglomerates deposited directly from glacial ice (that is, tillites), and the abundance of finer-grained rocks in the Headquarters Formation. We prefer prodelta and delta-front depositional settings because of the stratigraphic position of the Headquarters Formation, overlying presumed fluvial and deltaic deposits of the upper Deep Lake Group and conformably underlying quartzites of the Heart Formation that we also consider to be prodelta and delta-front sands.

**Heart Formation.** The Heart Formation (Blackwelder, 1926; Houston and others, 1968; Lanthier, 1979) lies conformably between the Headquarters Formation and the Medicine Peak Quartzite throughout the map area. The formation is 670 m thick and is predominantly quartzite. A phyllite unit as much as 90 m thick is locally present about 400 m above the base of the Heart Formation (Fig. 2).

The dominant lithology of the Heart Formation is a highly sericitic, very fine- to medium-grained quartzite (Fig. 5). The unit becomes more quartzose near the top. Plagioclase, chlorite, biotite, and opaque minerals may be present in addition to quartz and sericite. Heavy minerals include zircon, tourmaline, spinel, rutile, and apatite. Lanthier (1978) noted a micaceous quartz-pebble conglomerate at one locality.

A variety of sedimentary structures are present in the quartzites of the Heart Formation. These include small-scale planar and trough cross-bedding, climbing ripples, interference ripples, symmetrical ripples, ball and pillow structures, and graded bedding. The quartzites at the top of the formation are coarser grained and generally massive to plane-bedded. The phyllites are well laminated (Lanthier, 1979).

The Heart Formation is interpreted as prodelta and delta-front sediments associated with a prograding macrotidal (tidal-dominated) delta (Fig. 6). The laminated phyllites and very fine-grained, argillaceous, feldspathic quartzites of the lower two-thirds of the Heart Formation may represent prodelta bottom-set deposits (Reineck and Singh, 1975, p. 273). The presence of graded bedding, climbing ripples, and ball and pillow structures suggests 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; Reineck and Singh, 1972, p. 77; Elliott, 1978). The coarser-grained, subargillaceous, massive quartzites at the top of the Heart Formation represent delta-front sediments (Eriksson, 1979; Elliott, 1978), where deposition can occur by grain flow (Blatt and others, 1972, p. 163). The occasional cross-beds and ripple marks in the upper part of the formation indicate shallowing of the water so that waves and tidal currents could affect the sediments (Elliott, 1978). A limited paleocurrent analysis (Fig. 3) indicates that sediment transport was primarily to the southwest.

**Medicine Peak Quartzite.** 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 metasediments. The quartzite is about 1,700 m thick (Fig. 2) and forms an impressive cliff in the Snowy Range.

The Medicine Peak Quartzite is predominantly a medium- to very coarse-grained quartzite with pebbly zones and layers of quartz-pebble conglomerate. The quartzite ranges from quartzarenite to argillaceous arenite (Fig. 5). Sericite is common; kyanite, pyrophyllite, kaolinite, zircon, tourmaline, and iron oxide are minor constituents. Rarely, plagioclase and biotite may be present. The aluminosilicate minerals (sericite, kaolinite, pyrophyllite, and kyanite) may be in part an in situ alteration product (diagenetic and/or metamorphic) of feldspar, suggesting that part of the unit once was more arkosic than at present (Pettijohn and others, 1972, p. 408; Miyashiro, 1973, p. 199; Young, 1973; Flurkey, 1983). A hematitic quartz-pebble conglomerate, here named the Klondike Lake Conglomerate Member, lies ~125 m below the top of the Medicine Peak Quartzite (Fig. 2). The conglomerate is nearly 17 m thick and can be traced for about 9 km.

The quartzite is generally medium- to thick-bedded and massive to plane-bedded. Planar, tabular cross-bedding (average inclination = 23.5°) is the principal cross-stratification type in the Medicine Peak Quartzite. Trough cross-beds, asymmetrical ripples, large-scale low-angle cross-beds, and graded bedding are infrequent. The graded beds are conglomeratic and may show inverse as well as normal grading. Paleocurrent analysis (Fig. 3) indicates a dominant southwest, largely unimodal directed current system. A few outcrops are polymodal, and one shows only north-directed currents.

Available evidence suggests that the Medicine Peak Quartzite was deposited on the fluvial to subtidal portions of a tidally dominated delta plain (Fig. 6). It appears to reflect an estuarine environment, with features of both tidal and fluvial currents. This is in general agreement with interpretations presented by Kauffman and Steidtmann (1981), who interpreted the upper Medicine Peak to be intertidal to shallow subtidal deposits. Stream-transported arkosic, argillaceous sands were deposited on the delta plain and reworked by the macrotidal currents in the intertidal zone. The general paucity of feldspar in this unit suggests that feldspar may have been selectively abraded and reduced in grain size and thus transported with the finer-grained sediments to lower-energy environments, that is, tidal flats of the delta front (Field and Pilkey, 1969; Balaz and Klein, 1972).

The bulk of the Medicine Peak Quartzite appears to represent the coarser-grained and quartz-rich sediments that remained in the high-energy environments of the delta (Tankard and Hobday, 1977; Button and Vos, 1977; Eriksson, 1979). Large-scale low-angle cross-beds are interpreted to represent deposition of sand on the subtidal delta in large linear shoals elongate in the direction of tidal currents (Wright and others, 1975; Klein, 1970). Planar and tabular cross-beds presumably represent sand waves formed in braided fluvial channels or on the shoals and in the intershoal channels in response to tidal currents. The dominantly unimodal paleocurrent distribution in the Medicine Peak Quartzite suggests that most of the marine-formed sand waves were a response to ebb-tidal currents, in keeping with the time-velocity asymmetry of tidal currents (Klein, 1970). North-directed paleocurrents indicate the dominance of flood tides locally. The local bimodal (south- and west-directed) paleocurrents suggest bed forms moving in channels (south-directed) and on shoals (west-directed). The pebbly quartzite and quartz-pebble conglomerates appear to represent channel fill sequences (Eriksson, 1979).
current analysis (Fig. 3), primarily from fine-grained quartzites at the top of the unit, shows a bimodal-biopolar, northeast-southwest distribution.

The depositional environment of the Lookout Schist appears to be similar to that of the Heart Formation (Fig. 6). The graded and contorted beds, classic dikes, and laminated phyllites suggest prodelta and delta-front deposition (Blatt and others, 1972; Reineck and Singh, 1975; Elliott, 1978). The small-scale planar and hummocky cross-beds, flat-pebble conglomerates, and climbing ripples in fine-grained arkosic quartzites at the top and bottom of the unit imply storm and reversing tidal-current deposits on the shallow-water portion of the delta front (Bourgeois, 1980). The lens of “Medicine Peak-like” quartzite appears to be a delta-front distributary channel fill deposit (Elliott, 1978). Carbonates and possible iron-formation are interpreted to represent sporadic chemical sedimentation in the prodelta basin. Some of the detrital iron-oxide grains may have the same source as similar grains in the Klondike Lake Conglomerate.

Sugarloaf Quartzite. The Sugarloaf Quartzite conformably overlies the Lookout Schist but is in fault contact with the overlying Nash Fork Formation. The quartzite is at least 580 m thick but is commonly thinner or absent due to faulting (Lanthier, 1979). The abrupt disappearance of this thick unit along strike in two places is one of the main justifications for new map interpretations that show major thrust faults in the Libby Creek Group.

The Sugarloaf Quartzite is a monotonously white, medium-grained quartzarenite (Fig. 5). Sericite, tourmaline, and opaque minerals are present in minor amounts. We observed thin beds of hematitic quartzite near the upper end of the unit, and Blackwelder (1926) reported a quartz-pebble conglomerate in a similar stratigraphic position.

The dominant sedimentary structure is medium- to thin-bedded plane beds. Other common sedimentary structures include small-scale planar and trough cross-beds, and symmetric, interference, and climbing ripples. A few larger-scale planar cross-beds were observed. Paleocurrent analysis of cross-bedding (Fig. 3) indicates that sediment transport was dominantly to the west-southwest. Oscillation ripples record north-south and east-west current sense.

The maturity of the sediment and the abundance of small-scale sedimentary structures, especially oscillation ripples, suggest that the Sugarloaf Quartzite was deposited in a shallow-marine environment. The shallow water was probably in response to progradation of a delta system, as in the Medicine Peak Quartzite (Fig. 6). The abundance of oscillation ripples, however, suggests that wave activity was becoming more important as a sediment mover at the expense of ebb-tidal currents (Reineck and Singh, 1975).

Upper Libby Creek Group

The upper Libby Creek Group is here defined to include the three uppermost formations of the metasedimentary succession in the Medicine Bow Mountains: Nash Fork Formation, Towner Greenstone, and French Slate. A major change occurred in the sedimentation pattern between the siliciclastic (deltaic) succession of the lower Libby Creek Group and the more varied lithologies of the (open-marine) upper Libby Creek Group. The change in sedimentation is probably a reflection of evolving tectonic conditions (discussed below).

Nash Fork Formation. As discussed previously, the Nash Fork Formation is in fault contact with the lower Libby Group along the Lewis Lake Fault. The upper contact is not exposed (Houston and others, 1968), but the Nash is structurally conformable with the overlying Towner Greenstone. The unit is at least 1,980 m thick (Fig. 2) but is fault-bounded in part and originally may have been thicker.

The Nash Fork Formation consists predominantly of tan metadolomite with thick lenses of black phyllite. Rarely, thin beds of quartzite, metachert, flat-pebble conglomerate, and iron-formation may be present. The most striking sedimentary feature of the metadolomite is stromatolitic bioherms that occur in a variety of shapes and sizes (Knight and Keefer, 1966; Knight, 1968). The metadolomite consists primarily of dolomite and quartz with variable amounts of plagioclase, muscovite, phlogopite, tremolite, tephroite, apatite, and opaque minerals. The phyllite consists of muscovite, chlorite, biotite (and/or phlogopite), zoisite, sphene, tremolite, quartz, dolomite, and pyrite and other opaque minerals. Pyrite, graphite, and hematite have been observed in a small lens of iron-formation within a larger, black phyllite lens (Houston and others, 1968).

Knight and Keefer (1966) and Knight (1968) mapped about 150 bioherms and 3 reefs in the metadolomite. The reefs are about 60 m thick and perhaps 1 km in length and consist of many bioherms, which
range from 1 to about 30 m in thickness and from 3 to about 30 m in length (Knight, 1968). Knight (1966) subdivided the bioherms into several types and forms. No specific organisms have been identified from these stromatolites. Some of the bioherms show evidence of erosion. Cross-bedding and flat-pebble conglomerates are additional sedimentary structures found in the dolomites (Houston and others, 1968; Knight, 1968).

Fenton and Fenton (1939) suggested that the bioherms grew in clear, shallow water. The abundant siliciclastic particles indicate that at least part of the time there was a considerable influx of clastic debris. Current activity is also shown by cross-bedding, flat-pebble conglomerates, and erosional scour on the bioherms. The presence of pyrite and graphite in the phyllites implies restricted circulation for at least part of the time (James, 1966), suggesting variable currents. The influx of clastic sediments, erosional features, and possible variable currents indicates to us that the Nash Fork Formation was deposited on intertidal flats probably associated with a shallow-marine platform. Similar features have been described from lower Proterozoic carbonate platform. Similar features have been described from lower Proterozoic carbonates in South Africa (Button and Vos, 1977; Beukes, 1977).

**Towner Greenstone.** The Towner Greenstone is conformable with the overlying French Slate (Childers, 1957). This formation is 180 to 490 m thick (Houston and others, 1968) but appears to pinch out to the southwest, perhaps due to faulting. The Towner Greenstone consists of massive to schistose amphibolite with several small lenses of coarse-grained sandstone and fine-grained quartzites that may be metacherts. The greenstones consist primarily of actinolite with variable amounts of chlorite, albite, epidote, carbonates, opaque minerals, and sphene. No diagnostic genetic features have been observed in the greenstones, although the interbedded sandstone lenses suggest an extrusive origin (Houston and others, 1968), and the chert(s?) suggest marine deposition. An intrusive origin cannot be ruled out.

**French Slate.** The French Slate overlies the Towner Greenstone and is truncated by the Cheyenne Belt along its entire upper contact. Its present map thickness is about 610 m. The formation consists primarily of laminated black ferruginous slate and phyllite. The laminae appear primary and consist of layers containing approximately equal amounts of muscovite, chlorite, quartz, and opaque minerals alternating with quartz-rich layers with minor muscovite and chlorite. Metacryystals of biotite and pyrite are common (Houston and others, 1968). We have observed two thick lenses of hematitic iron-formation in the upper part of the unit. The hematite is in a matrix of very fine-grained quartzite (metachert). The French Slate has been complexly folded and crenulated by movement along the shear zone.

The lack of coarse clastics and the presence of iron-formation imply that the French Slate was deposited in a low-energy environment. The thickness of the unit indicates that deposition was in a deep-marine or prodelta basin (Reineck and Singh, 1975).

**STRATIGRAPHIC AND SEDIMENTARY FEATURES OF EARLY PROTERozoIC ROCKS IN THE SIERRA MADRE**

Late Archean and early Proterozoic supracrystal sequences in the Sierr Madre (Fig. 1) are more strongly deformed and metamorphosed than are correlative rocks in the Medicine Bow Mountains, so that stratigraphy and paleogeographic interpretations are based partly on correlation with the Medicine Bow Mountain section. The late Archean Phantom Lake Metamorphic Suite in the Sierra Madre was intruded by the 2,700-m.y.-old Spring Lake Granodiorite (Carl Hedge, 1982, personal commun.) and deformed and metamorphosed in the regional orogenesis of 2,700 to 2,500 m.y. ago. This sequence is discussed in detail by Graff (1979) and Karlstrom and others (1981). Rocks equivalent to the Deep Lake Group (lower Snowy Pass Group in Fig. 1) are believed to unconformably overlie Archean metasediments and gneisses and incomplete sections containing correlative of the upper and lower Libby Creek Groups (middle and upper Snowy Pass Groups in Fig. 1) are preserved in fault slices. These faults are believed to be major reverse faults (Graff, 1978) or steepened thrust faults (Karlstrom and others, 1981).

**Lower Snowy Pass Group**

Correlatives of the Deep Lake Group in the Sierra Madre consist of the Magnolia Formation, Singer Peak Formation, Cascade Quartzite, and parts of the Bottle Creek Formation (see Houston and others, in press). Graff (1978) placed several thrust faults in the section but the thickness variations may also be explained in part by erosional unconformities.

**Magnolia Formation.** This formation consists primarily of quartzite with slightly radioactive conglomerates at the base and phyllites near the top. The thickness of the unit ranges from 460 m on the west to zero on the east. This eastward thinning may reflect on the depositional surface, erosion prior to deposition of the Cascade Quartzite, and/or faulting.

Over much of the Sierra Madre, the Magnolia Formation is difficult to distinguish both structurally and lithologically from the metasediments of the Phantom Lake Metamorphic Suite. There are no basal conglomerates above an obvious regional-scale unconformity in the Sierra Madre like those we have mapped in the Medicine Bow Mountains, although generally coarser grain sizes and the presence of trough cross-bedding serve to distinguish the Magnolia from underlying units. However, structural data from the western Sierra Madre suggest an angular discordance between the Magnolia and the Phantom Lake Suite, either an angular unconformity (Karlstrom and others, 1981) or fault (Graff, 1978).

The quartzites of the Magnolia Formation are primarily arkoses and subarkoses (Fig. 7). Their similarity in lithology to the clastic sediments of the Phantom Lake Metamorphic Suite suggests that the Magnolia sands were derived from the nearby Archean metasediments. The main sedimentary structures are small-scale trough cross-beds. Limited paleocurrent data indicate that the sediment was deposited by south-directed currents (Karlstrom and others, 1981). We interpret the Magnolia Formation to have been deposited by braided streams originating on fault-bound highlands to the north.

**Singer Peak Formation.** This formation (Graff, 1979) consists predominantly of phyllites with minor quartzites and conglomerates. Paraconglomerates are present near the top. The Singer Peak Formation is correlated with the Lindsey Quartzite and Campbell Lake Formation of the Medicine Bow Mountains. This unit is about 831 m thick in the west and thins (due to erosion, facies change, or faulting) to zero in the east.

The Singer Peak Formation is made up
primarily of quartz, feldspar, muscovite, and chlorite (or biotite). We presume that the unit originated as a clay-rich sediment. No sedimentary structures have been observed in this formation. We interpret the metasediment to be an offshore marine facies of fluvial-glacial units in the Medicine Bow Mountains.

**Cascade Quartzite.** This formation is dominated by quartzite with layers of quartz-pebble and black chert-pebble conglomerate. The quartzite is about 560 m thick in the center of its outcrop area and thickens to the east and west. Maximum thickness is at least 1,500 m, but structural complexities and the lack of marker beds make it difficult to determine the exact thickness. In the eastern Sierra Madre, it is the basal unit of the Deep Lake Group. The quartzite is primarily a subarkose (Fig. 7).

Planar and trough cross-bedding and graded bedding have been observed in this unit (Graff, 1978), but detailed sedimentologic analysis has not been made. Its similarity to the Cascade Quartzite of the Medicine Bow Mountains suggests a similar fluvial or deltaic origin.

**Bottle Creek Formation.** The Bottle Creek Formation is a heterogeneous unit consisting of quartzites, phyllites, angular-clast paraconglomerates (diamictites), and metacarbonates. This formation is at least 400 m thick, but its top is always in fault contact with rocks of the Libby Creek Group. No sedimentary structures have been observed in this unit, but it is very similar to the glaciogenic metasediments of the Vagner Formation in the Medicine Bow Mountains. It may also contain rocks correlative with the Headquart Formation.

**Correlatives of the Libby Creek Group**

Divis (1976, 1977) correlated nearly all of the metasedimentary rocks of the Sierra Madre with the Libby Creek Group of Houston and others (1968), whereas Graff (1978) placed only one formation (the Slaughterhouse) into the Libby Creek Group. Here, we correlate Graff's (1978, 1979) Copperton Quartzite with the lower Libby Creek Group and the Slaughterhouse Formation with the upper Libby Creek Group of the Medicine Bow Mountains (see Houston and others, in press, for details of this correlation).

**Copperton Formation.** The Copperton Quartzite of Graff (1979) is here modified to Copperton Formation because of the thick section of phyllite present within it. The formation is at least 1,070 m thick, but as it is bounded by major faults, its maximum thickness is unknown. The unit consists of a lower, coarse-grained, highly sheared, kyanite-bearing quartzite, a few hundred metres of ferruginous phyllite, and an upper sheared quartzite. On the basis of similar lithology, stratigraphic position, and structural setting, the Copperton Formation is correlated with the Medicine Peak Quartzite, Lookout Schist, and Sugarloaf Quartzite of the lower Libby Creek Group in the Medicine Bow Mountains. The Copperton Formation is strongly sheared and its sedimentology has not been studied in detail, but the similarities in lithology and thickness between it and the deltaic sediments of the lower Libby Creek Group in the Medicine Bow Mountains suggest a similar origin.

**Slaughterhouse Formation.** This formation is bounded by a major fault below and the Cheyenne Belt above. This unit is composed of metacarbonate, quartzite, phyllite, and metachert. It is at least 1,160 m thick. This formation is correlated with the Nash Fork Formation of the Medicine Bow Mountains, although none of the stromatolites, characteristic of the Nash Fork, have been observed in the Slaughterhouse Formation. It is presumed to be of marine origin.
Proterozoic Gabbroic Intrusives

All Precambrian rocks in the Medicine Bow Mountains and the Sierra Madre, with the exception of the 1,400-m.y.-old Sherman Granite, are crosscut by mafic bodies of gabbroic composition that have been entirely or partly converted to amphibolite (see Houston and others, 1968, for a discussion of the petrology of these rocks). Structural evidence suggests that some of these may be Archean in age (Karlstrom and Houston, 1979b); others occur as large layered complexes south of the Cheyenne Belt (Fig. 1) and may represent remnants of obducted oceanic crust (Hills and Houston, 1979) or mafic intrusions at intermediate levels of an island-arc massif; still others crosscut the Cheyenne Belt and represent late-stage intrusions following the main orogenic pulse. Two other groups, which are of direct interest here, include large gabbroic sills that crosscut the Phantom Lake Suite and Deep Lake Group and smaller dikes that crosscut the Libby Creek Group. These intrusions are early Proterozoic in age.

Intrusions in the Phantom Lake Suite and the Deep Lake Group are large phacolithic bodies that we loosely refer to as "sills" but which, in detail, crosscut both the tight folds of the Phantom Lake Suite and the open folds of the Deep Lake Group (Karlstrom and Houston, 1979b). These sills also appear to be truncated against the Reservoir Lake Fault (Fig. 1). The faulted contact of the sills suggests that at least some of them were emplaced before movement on the thrust faults. Thus, the sills are early Proterozoic in age but appear to predate the major 1,700- to 1,800-m.y.-old orogenesis. Intrusions in the Libby Creek Group are of direct interest here, include large late-stage intrusions following the main orogenic pulse. Some of these are spatially related to small bodies of the Gaps Intrusion (Karlstrom and others, 1981), which also crosscuts the lower Libby Creek Group and has yielded a 2,000 m.y. age (Carl Hedge, 1982, personal commun.; Houston and others, in press) that we take as the approximate age of the dikes.

Geochemical data from the various gabbroic intrusions are summarized in Figure 8 (see Karlstrom and others, 1981, for analyses). Sills in the Phantom Lake Suite and the Deep Lake Group are quartz-normative tholeiitic gabbros (Cox and others, 1979), with an iron-enrichment trend (Fig. 8A) comparable to that of mid-Atlantic basalts (Irvine and Baragar, 1971). Dikes cutting the Libby Creek Group are also gabbros, but they show a higher average alkali content and lower average silica than do the sills. In both the alkali-silica and AFM plots (Fig. 8), the two groups are located in separate fields, the sills being clearly tholeiitic and the dikes being subalkalic, suggesting that the two groups may represent different intrusive events.

From regional considerations, it seems reasonable to equate the early Proterozoic sills in the Medicine Bow Mountains with the 2,150-m.y.-old Nipissing Diabase of the southern Canadian Shield (Fairbairn and others, 1969; Card and Pattison, 1973), with the 2,100-m.y.-old Blue Draw Metagabbro of the Black Hills of South Dakota (Redden, 1980), and perhaps with 2,100- to 2,200-m.y.-old mafic dikes in the Bighorn Mountains of Wyoming (Stueber and others, 1976). 2,000-m.y.-old dikes in the Libby Creek Group also may be correlatable with these intrusions, or they may be slightly younger. The similarities in age and chemistry between these gabbroic intrusives indicate a continental-scale episode of mafic magmatism that might well be explained in terms of a major rifting event in North America at this time (discussed by Karlstrom and others, 1981).

PLATE-TECTONIC MODEL FOR EARLY PROTERozoic SEDIMENTATION

Geologic and geochronologic data from southern Wyoming suggest orogenesis about 1,700 m.y. ago that involved cratonic margin–island arc collision (Hills and Houston, 1979; Karlstrom and others, 1981). This section attempts to explain the early Proterozoic sedimentary record in the Medicine Bow Mountains in terms of deposition on a rift-formed, Atlantic-type continental margin prior to this collision.

Late Archean sedimentation in southern Wyoming, represented by the Phantom Lake Metamorphic Suite in the Medicine Bow Mountains and the Sierra Madre (Karlstrom and others, 1981), the Elmers Rock Greenstone Belt in the Laramie Mountains (Graff and others, 1981), and the Whalen Group of the Hartville Uplift (Snyder, 1980), was followed by an orogenic episode that culminated about 2,500 to 2,600 m.y. ago with intrusion of synorogenic granitic plutons and amphibolite-facies regional metamorphism (Hills and
This orogenic episode, contemporaneous with the Kenoran orogeny of the Superior Province, marked the end of Archean tectonics in Wyoming, a period characterized by widespread deformation and plutonism and a dominance of volcanic and volcanioclastic rocks in the sedimentary record. Shortly following the 2,500-m.y.-old orogenic period, the tectonic regime in southern Wyoming changed and became characterized by a stable craton capable of being deeply eroded and of supporting thick siliciclastic sediments along its margins.

The Deep Lake Group is the record of this change in tectonic regimes. Conglomerates of the basal Deep Lake Group rest with angular unconformity on late Archean granites and on folded Phantom Lake Suite rocks. The regional extent of the unconformity, the presence of rock fragments from a variety of Archean rock types in the conglomerates, along with the presence of placer accumulations of heavy minerals from both granitic terranes (cordilleran, monazite, zircon, apatite) and supracrustal successions (pyrite, chalcopyrite, bornite, garnet), suggest that the unconformity represents deep weathering of the basement terrane and significant hiatus. Sedimentary evidence indicates that the basal conglomerates are fluvial deposits consisting of both alluvial-fan facies, represented by debris-flow deposits, and braided river sands and gravels, represented by radioactive quartz-pebble conglomerates and quartzites. As shown in Figure 9B, we postulate an intracratonic, fault-bounded basin or graben for Magnolia and Lindsey deposition. This type of paleogeographic reconstruction is supported by the presence of debris-flow deposits in the Magnolia, which suggest syndepositional faulting, and palynological and fossil evidence in the Magnolia that might arise from alluvial fans building out into a southwest-flowing braided-river system. West-directed conglomerates in the Magnolia in one area of the northern Medicine Bow Mountains (Fig. 3) also support this model.

Evidence that this graben was related to incipient rifting comes mainly from the parallelism of fluvial conglomerates in the Magnolia and Lindsey to conglomerates in marine rocks of the overlying lower Libby Creek Group and the parallelism of all paleocurrents to the Cheyenne Belt. Similar current directions in both fluvial and marine sediments would be expected in an elongated intracratonic basin but generally not on an open continental margin. Also, the type section of the Magnolia Formation contains thin beds of mafic volcanics (Karlstrom and Houston, 1979a), implying local syndepositional volcanic activity, typical of rift environments. Evidence from the Sierra Madre shows that fluvial deposition gave way to marine deposition in the western part of the basin, so that the Lindsey Quartzite appears to be absent in the Sierra Madre and the equivalent stratigraphic position is occupied by the marine Singer Peak Formation. This suggests that seas transgressed into the basin from the west (Fig. 9B). The 120° bend in the north margin of the basin shown in Figure 9B mimics the present-day distribution of metasedimentary rocks in the Medicine Bow Mountains and Sierra Madre and the trend of the Cheyenne Belt (Fig. 1) and argues for a triple junction, providing some justification for the aulacogen shown in the hypothetical southern block in Figure 9B.

The rifting story during upper Deep Lake Group and lowest Libby Creek Group deposition is complicated by advance of continental-scale glaciers down the Medicine Bow sedimentary basin and perhaps across the newly forming continental edge (Fig. 9C). Vagner through Headquarters deposition appears to be related to advance and retreat of this ice sheet. One interesting aspect of this succession is the west-directed paleocurrents of the Vagner and Rock Knoll Formations (Fig. 9C), which are our only direct evidence for the existence of the southeastern rift block. These paleocurrents and the increase of plagioclase in the sediments and decrease in K-spar relative to underlyng rocks of the Deep Lake Group indicate that a mafic to intermediate volcanic or plutonic highland to the east was a principal source of detritus during glacial deposition.

Retreat of glacial ice in upper Headquarters time re-established a major delta system similar to one that may have existed earlier, in Cascade time. This delta was fed from the northeast and persisted throughout lower Libby Creek Group deposition (Fig. 9D). It is these rocks that lend most support to the rifting model. This thick, mature, deltaic succession was deposited by paleocurrents directed southeast (and northeast) parallel to older fluvial paleocurrents and parallel to the continental margin as inferred from the major shear zone bounding the Wyoming Province. It is believed that deltaic sedimentation in a trough that was oriented parallel to continental margins also took place in the early stages of Mesozoic rifting of South America and Africa (Burke, 1976, mentioned deltas formed in the Gabon and Cuanza grabens, which became "successful" rift arms, leaving the Benue Trough as a failed arm). Deltaic sedimentation parallel to a rifted continental margin is also taking place in the Gulf of California today (Meckel, 1975). Either of these appears to us to be a reasonable analogue for lower Libby Creek Group sedimentation, and we stress that we have no idea whether the southern block was of continental or microcontinental proportions.

As shown in Figure 9E, by upper Libby Creek Group time, rifting had progressed to the open-ocean stage, and deltaic sedimentation had ceased in southeastern Wyoming. Nash Fork clastic carbonates and shales are interpreted to be intertidal sediments deposited on lower Libby Creek Group deltaic sands. The appearance of volcanic rocks of the Towner Greenstone and deposition of open-marine black shales of the French Slate in the upper Libby Creek Group may be related to approach of an island arc from the south (Hills and Houston, 1979), and they may be broadly contemporaneous with the oldest volcanicogenic rocks of the eugeoclinal terrane.

The gabbroic intrusives that crosscut the metasedimentary successions also appear to fit into a rift model for early Proterozoic sedimentation and tectonics. Similar tholeiitic to slightly alkalic intrusions are interpreted to be related to intracontinental rifting in other areas. Strong (1975) reported continental tholeites in western Newfoundland that he interpreted to be related to early rifting of the Paleoaoicia lapetus ocean, incipient intracontinental rifting in the midcontinent region of North America produced the Keweenawan basaltic flows and dikes of tholeiitic to weakly alkalic affinity (Green, 1975; Baragar, 1975), and tholeiitic sills of Mesozoic age were emplaced in fault-bounded Triassic basins in both North America and Africa during Mesozoic rifting of the Atlantic Ocean (May, 1971). The timing of emplacement of tholeiitic intrusions in southern Wyoming also appears to fit a rift model. Large sills were emplaced after deposition of the Deep Lake Group but before the 1,700- to 1,800-m.y.-old orogenesis. These sills were earlier or contemporaneous with dikes in the Libby Creek Group that were emplaced about 2,000 m.y. ago. Thus, broadly speaking, the gabbros were intruded late in the development of the sedimentary succession. Although the rela-
Figure 9. Plate-tectonic model for deposition of early Proterozoic metasedimentary rocks of southern Wyoming.
tionship between the sills in the Deep Lake Group and dikes in the Libby Creek Group is obscured by structural telescoping of the succession by thrust faults, we suspect that the large sills may be slightly older than the dikes and that their emplacement was related to development of the narrow ocean basin in which the lower Libby Creek Group deltaic sediments were deposited. The 2,000-m.y.-old Gaps Intrusion and the dikes that crosscut the lower Libby Creek Group appear to represent intrusions related to the formation of an open ocean prior to deposition of the upper Libby Creek Group.

Further geochemical and geochronologic work on the gabbros would help to elucidate this hypothesis of compound early Proterozoic rifting in southern Wyoming.

In summary, we believe that the combined stratigraphic, sedimentological, and petrologic data from both early Proterozoic metasedimentary rocks and gabbroic intrusives are best explained by a model involving complex intracratonic rifting in southern Wyoming 2,300 to 2,000 m.y. ago. The sedimentary record indicates a progression of tectonic regimes from intracratonic faulted basins in which fluvial sedimentation predominated, to development of a narrow marine basin that became the site of deposition of a large delta, to development of a true continental shelf on which carbonates and open-marine shales were deposited. We note that, although the details of sedimentation differ, both the thickness (13 km) and duration (200-300 m.y.) of sedimentation in the Snowy Pass Supergroup are similar to those documented from rift-formed basins along the Atlantic margin of North America (for example, Poag, 1979, reported at least 14 km of marine and nonmarine sedimentary rocks deposited in post-Triassic time in the type continental margin in southeastern Wyoming), suggesting complex intracratonic rifting in southern Wyoming.

FIELD ENGINEERING CORPORATION SUBCONTRACT DE-AC13-76GJ01664, the U.S. Geological Survey, the National Science Foundation (Grant no. 43-8409), and the Research Society of North America (Sigma XI). This report has been substantially improved by suggestions of George L. Snyder and John C. Harms, but neither is to be held responsible for the concepts presented.

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