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Geological Society of America Bulletin 2000;112:564-578

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Late Archean structural and metamorphic history of the Wind River Range: Evidence for a long-lived active margin on the Archean Wyoming craton

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ABSTRACT

The Archean rocks of the Wind River Range in western Wyoming record a Late Archean history of plutonism that extends for more than 250 m.y. The range is dominated by granitic plutons, including the 2.8 Ga Native Lake gneiss, the 2.67 Ga Bridger batholith, the 2.63 Ga Louis Lake batholith, and late 2.54 Ga granites. These plutons provide a means of distinguishing the complex metamorphism and deformation that affected the range in the Late Archean. Five deformation events are recorded. D1 is a penetrative deformation that occurred during the earliest granulite-facies metamorphism; D2 is a folding event, probably in amphibolite facies, that deforms porphyritic dikes that cut the D1 fabrics. Both D1 and D2 predate the intrusion of the ca. 2.8 Ga Native Lake gneiss. D3 is a folding event, accompanied by upper amphibolite to granulite metamorphism, that deformed the Medina Mountain sequence, a sequence of rocks that was either deposited or thrust upon the Native Lake gneiss. D4 is a fabric-forming event associated with the Mount Helen structural belt (MHSB). It is represented by mylonites in the MHSB, a penetrative fabric in the Bridger batholith, and folding of the D1 fabrics in the Medina Mountain sequence. We consider D1 and D2 to be coeval with the emplacement of the Bridger batholith, and hence to date at ca. 2.67 Ga. The latest structures (D3) are fabrics associated with the folding and thrusting of the 2.65 Ga South Pass sequence.

We recognize at least four metamorphic events. M1 is associated with the D1 fabrics and occurred at high T (>750 °C) and high P (7–8 kilobars). M2 (650–750 °C and 4-5.5 kilobars) is associated with the intrusion of the Bridger batholith and formation of the D2 and D3 structures. The D2 structures of the South Pass sequence record M3, which is low P (~2-3 kilobars) and low T (~500 °C). The final metamorphism, M4, is a contact metamorphism around the Louis Lake batholith. In the south against the South Pass sequence, the metamorphism occurred at ~3 kilobars and at temperatures <700 °C. In contrast, in the north where the Louis Lake batholith is charnockitic, the metamorphism occurred at 6 kilobars and 800 °C. This pressure gradient is probably a reflection of tilting of the Wind River block during the Laramide orogeny.

The composition of the plutons and the structural and metamorphic history of the Wind River Range indicate that during the Late Archean this area occupied the active margin of the Wyoming province. This tectonic environment is similar to the long-lived Phanerozoic margins of North America. The Wind River Range represents the best-documented active margin of Archean age.

Keywords: active margin, Archean, metamorphic petrology, structure, tectonics, Wyoming province.

INTRODUCTION

The Wyoming province is the most southwestern of the Archean provinces of North America. It has a geologic history that is distinctive from that of the Superior province, the largest Precambrian craton within North America. The majority of the Superior province formed rapidly between 2.7 to 2.8 Ga (Card, 1990; Percival et al., 1994), whereas isotopic evidence indicates that the Wyoming province was cratonized before 3.2 Ga (Wooden and Mueller, 1988; Frost, 1993) and that this early crust was repeatedly reworked by later Archean events (Frost et al., 1998). Its history of early cratonization is shared by the Slave, Nain, and Minnesota River Valley provinces, small cratons that ring the Superior and Churchill provinces (Hoffman, 1988).

SUMMARY OF THE GEOLOGY OF THE WIND RIVER UPLIFT

The Wind River Range, a northwest-trending uplift in western Wyoming, exposes Archean rocks over an area of more than 10 000 km2. The range is composed almost entirely of high-grade gneisses and granites that were thrust to the west over Paleozoic and Mesozoic sedimentary rocks during the Laramide Orogeny 40–80 Ma. The thrust, which is sparsely exposed, has been shown seismically to extend to depths of more than 15 km (Smithson et al., 1978). On the east, the Precambrian rocks are covered by Paleozoic and Mesozoic sedimentary rocks, dipping on average 15° NE into the Wind River Basin. The uplift plunges gently to the north in the north and to the south in the south, which suggests that the deepest levels are exposed in the central-west portion of the range (Mitra and Frost, 1981).

Most of the basement in the range consists of granites and granite gneisses (Fig. 1). Frost et al. (1998) recognized four distinct ages of plutonism in the range. In this paper, we use the relation between these plutons and various metamorphic and structural elements in the surrounding gneisses to unravel Late Archean plutonic, sedimentary, metamorphic, and deformation events that occurred in the Wind River Range. These plutons include:

1. The Native Lake gneiss, a locally deformed calc-alkalic pluton in the Washakie terrane with a preliminary zircon U-Pb age of ca. 2.8 Ga (Frost and Frost, 1993).
2. The foliated Bridger batholith, which has been dated at 2670 ± 13 Ma (Aleinikoff et al., 1989), and which makes up a major portion of the northern part of the range.

3. The undeformed Louis Lake batholith, which was intruded at 2630 ± 2 Ma (Frost et al., 1998).

4. A series of latest Archean plutons that have not been dated precisely, but for which Stuckless et al. (1985) suggested an age of 2545 ± 30 Ma. Included in this group are the Middle Mountain batholith and the granite of New Fork Lakes in the northern part of the range, the Bears Ears pluton in the center of the range, and isolated granites at South Pass (Frost et al., 1998).

Three areas within the Wind River uplift are particularly critical for unraveling the geologic history of the range. They include:

1. The Washakie block, a sequence of gneisses that occupies the northeastern portion of the Wind River Range (Fig. 1, area 1) (Frost et al., 1998). This coherent package of tonalitic and calc-alkaline gneisses lacks the intense migmatization associated with the Bridger batholith and provides an important window into the earlier history of the range.

2. A complex migmatite that surrounds the Bridger batholith in which isolated fragments of older gneisses survive within an area of felsic neosomes (Koesterer et al., 1987; Marshall, 1987). Some of the neosomes in the migmatite are of Bridger age, but some of them are clearly older and are presumed to be associated with Native Lake–aged plutonism (Frost et al., 1998). We studied two areas where these migmatites contained infolded belts of supracrustal rocks, one in the Medina Mountain (Fig. 1, area 2), and another in the Crescent Lake area (Fig. 1, area 3). These areas within the Wind River uplift are particularly critical for unraveling the geologic history of the range. They include:

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3. The South Pass area, on the southern margin of the uplift (Fig. 1), which is dominated by a sequence of weakly metamorphosed supracrustal rocks. These rocks, which we call the South Pass sequence, have been the subject of many studies (Bayley et al., 1973; Harper, 1985; Hull, 1988; Hausel, 1991).

**DETAILED GEOLOGIC HISTORY**

We have identified a record of geologic events in the Wind River Range that spans more than 1 billion years (Table 1). To characterize the relative age of events within the range, we have determined the relation between deformations displayed in the rocks and the plutonic events outlined in Frost et al. (1998). Because most structural elements are present in some areas and absent in others, we use the following convention for naming structural elements. We assign a symbol to a given deformation event (i.e., D1, D2,

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**TABLE 1. GEOLOGIC EVENTS IN THE WIND RIVER RANGE**

<table>
<thead>
<tr>
<th>Event</th>
<th>Age (Ga)</th>
<th>Where found</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deposition of early supracrustal rocks, granulite metamorphism, D0</td>
<td>&gt;2.8</td>
<td>CL, MHSB, MM, WB</td>
</tr>
<tr>
<td>Intrusion of porphyritic dikes</td>
<td>&gt;2.8</td>
<td>CL, MM, WB</td>
</tr>
<tr>
<td>D2: folding of D1 fabric and porphyritic dikes</td>
<td>&gt;2.8</td>
<td>WB</td>
</tr>
<tr>
<td>Intrusion of Native Lake gneiss</td>
<td>ca. 2.8</td>
<td>CL, MM, WB</td>
</tr>
<tr>
<td>Deposition of Medina Mountain sequence, metamorphism, and D0: folding of these rocks.</td>
<td>2.7–2.8</td>
<td>MM</td>
</tr>
<tr>
<td>Emplacement of the Bridger batholith, and D3: formation of Mount Helen structural belt and folding of D3 structures</td>
<td>2.67</td>
<td>BB,CL, MHSB, MM, WB</td>
</tr>
<tr>
<td>Deposition of South Pass sequence, metamorphism, D3: folding of SPS and thrusting onto Wyoming craton</td>
<td>2.65</td>
<td>SP</td>
</tr>
<tr>
<td>Intrusion of Louis Lake batholith, contact metamorphism</td>
<td>2.63</td>
<td>LL, SP</td>
</tr>
<tr>
<td>Intrusion of Late Archean plutons</td>
<td>2.54</td>
<td>All of the range</td>
</tr>
<tr>
<td>Intrusion of Proterozoic diabase dikes</td>
<td>1.47</td>
<td>All of the range</td>
</tr>
</tbody>
</table>

**Abbreviations:** BB—Bridger batholith, BE—Bears Ears batholith, CL—Crescent Lake area, LL—Louis Lake batholith, MHSB—Mount Helen structural belt, MM—Medina Mountain area, SP—South Pass, WB—Washakie block.
Early Supracrustal Rocks and the D1 Event

The oldest deformation event identified in the Wind River Range, a penetrative deformation associated with granulite metamorphism, is best displayed in supracrustal rocks from the Washakie block. Rock types in this association include metaperidotite, metabasalt, sulfidic metaquartzite, semipelitic gneiss (with lesser amounts of calc-silicate gneiss), pelitic gneiss, and iron-formation. Most of the rocks carry up-porphyritic-facies assemblages; however, they locally preserve granulite assemblages. This rock association, along with its local granulite metamorphism, is widespread throughout the range and occurs in the Medina Mountain area (Fig. 1, area 2), and D1 and possibly D2 fabric elements may occur in the migmatites in the southern and northern margin of the Bridger batholith. A weak S1 fabric is found throughout the Bridger batholith and older fabrics occur locally in this domain. For example, D1 fabrics are recognized in the Medina Mountain area (Fig. 1, area 2), and D1 and possibly D2 fabric elements may occur in the migmatites in the south and northern margin of the Bridger batholith. The youngest structural domain in the range is defined by D2 deformation of the South Pass sequence. Most of the central portion of the range is overlain by directionless fabrics of the Louis Lake and Late Archean batholiths, which were intruded after D2.

Porphyritic Dikes

Porphyritic, weakly foliated amphibolitized dikes cut the D1 fabrics in the Washakie block. These rocks are invariably boudinaged so that individual dikes typically can be traced for less than a hundred meters. Trains of boudinaged dikes however outline the D2 folds that also deform the D1 fabrics. This is clear evidence that the dikes were intruded between the D1 and D2 events. Porphyritic dikes that cut D1 fabrics are also present both in the Medina Mountain area, where they are called the Victor dikes (Koesterer et al., 1987), and in Crescent Lake area (Marshall, 1987). Because the porphyritic amphibolite dikes show similar textures in all three areas, and because the cross-cutting relations are the same, we consider them to represent the same magmatic event, and the term Victor dikes is applied in all three areas.

D2 Event

The S1.2 schistosities in the Washakie block are folded into broad folds that have a weak axial-planar schistosity that can be recognized only in the fold hinges. On the basis of the shape of these folds, we have divided the Washakie block into two domains, which are separated by the Dry Creek Ridge Structural Zone (Fig. 3). In the Dry Creek domain, these folds are relatively open and plunge at ~45°E (Fig. 4A), whereas in the Bob Lakes domain the folds are nearly isoclinal and plunge ~30°E (Fig. 4B). Because the Victor dikes, which cut the S1.2 fabric, are folded by these later folds, we conclude that this folding event is a distinct deformation event, which we call D2.

Foliations and lineations from the Dry Creek and Bob Lakes domains are plotted on stereo diagrams on Figure 4. The foliations measured are dominantly S1.2–D2 produced weak mineral-preferred orientations in the fold hinges that are easily distinguished from the compositional layering that is distinctive of S1.2. Distinction between lineations is ambiguous. The F2.1 event

Figure 2. Structural domains in the Wind River Range.
produced linear features (L_{2,1}) that are seen mainly as parasitic fold axes and crenulations. There are also L_{1,2} lineations associated with the F_{1,2} folds. The L_{1,2} lineations are easily recognized in the supracrustal rocks, where they form as crenulations and minor folds parallel to the F_{1,2} fold axes, but they are not so easily distinguished in the surrounding gray gneisses. Those lineations that we could clearly identify as L_{1,2} were not plotted on Figure 4. Nonetheless, there is a rather large scatter of lineations from the Dry Creek domain, and we conclude that a fair number of the lineations in Figure 4 formed before D_{2}. This is indicated by a larger scatter of lineations in the Dry Creek domain, where the D_{2} folds are relatively open, than in the Bob Lakes domain, where the folds are nearly isoclinal. We interpret this difference to result from the transposition of the L_{1,2} into parallelism with the L_{2,1} within the Bob Lakes domain (Fig. 4).

**Early Migmatization**

The early migmatites in the Washakie block are represented by the Native Lake gneiss, a weakly foliated to unfoliated calc-alkaline pluton that was emplaced late during or post-D_{2}. This is the earliest body for which we have a direct age; it gives U-Pb zircon dates of ca. 2.8 Ga (Frost and Frost, 1993). There are a few localities where the Native Lake gneiss is inclusion-free. Over most of its occurrence, it is migmatitic, hosting inclusions of all the older rock types discussed above. In a few areas, such as north of Crater Lake (Fig. 3), massive enderbites grade into typical amphibole-bearing Native Lake gneiss. In most of the rest of its exposures, the Native Lake gneiss has a texture similar to retrograded charnockites—the hornblende and biotite form clots rather than distinct grains. The hornblende and biotite in the clots are poikilitic in thin section, with numerous inclusions of quartz. These textures are similar to those produced by secondary biotite and hornblende in charnockites in which the pyroxene has been partially hydrated. We therefore infer that at least locally the Native Lake gneiss was originally pyroxene-bearing.
Weakly foliated felsic orthogneisses that we consider to be correlative with the Native Lake gneiss are found in strain-free zones of the Mount Helen structural belt and in the southwest portion of the Medina Mountain area mapped by Koesterer (1986) (Fig. 5). Like the Native Lake gneiss, the felsic gneiss in both areas contains clotted amphiboles, suggesting an earlier pyroxene-bearing history. Both the Medina Mountain and Crescent Lake areas expose a migmatite that predates the Bridger batholith and that contains inclusions of granulite-grade supracrustal rocks, which are cut by Victor dikes (Koesterer et al., 1987; Marshall, 1987). Our interpretation is that these early migmatites formed at the same time as the Native Lake gneiss. It is possible that these migmatites represent an intrusive event that is distinct from that Native Lake gneiss. With our present knowledge, however, we have no evidence that it is different.

Medina Mountain Sequence and D₃

A belt of supracrustal rocks (the Medina Mountain sequence), is in-folded with the early migmatites in the Medina Mountain area (Fig. 5) (Koesterer et al., 1987). The metasedimentary rocks within this sequence include pelitic and semipelitic gneisses and minor amounts of fine-grained amphibolite (probably metabasalt), quartzite, calc-silicate granofels, and iron formation. In places, such as on the northern limb of the synform that runs through Medina Mountain, a distinct stratigraphy is developed that (from base to top) consists of amphibolite, metapelitic gneiss, amphibolite, and psammitic gneiss. The migmatite at the base of the Medina Mountain sequence is not strongly sheared, which leads us to conclude that the Medina Mountain sequence probably rests in depositional contact on the migmatites. It is possible that the Medina Mountain sequence was thrust onto the migmatites, but, if that were true, then the thrusting must have happened at low temperatures where the strain would have been concentrated in narrow faults rather than in a wide mylonite zone (Sibson, 1977).

The Medina Mountain sequence has been folded into tight synforms during a deformation event (D₃) that clearly postdates the injection of the neosomes in the migmatite. The synforms lack an axial planar schistosity and clearly fold an earlier schistosity that is parallel to the compositional layering. We call the schistosity S₃,1 and the major synforms F₃,1, because these folds are caused by the earliest deformation event that we can identify. The S₃,1 is parallel to the northwest-southeast–trending schistosity in the earlier migmatites and the Bridger batholith. The rocks are poorly lineated; the major linear features are axes of minor folds and crenulations. The L₃,1 lineations have been folded into a broad fold, which is expressed in the field by the way that the F₃,1 folds form closed ovoid structures (Fig. 6). We consider this later fold to be a D₄ structure because it has the same trend as the F₄,2 folds in the Mount Helen structural belt.

Mount Helen Structural Belt and D₄

The Mount Helen structural belt is a major high-temperature shear zone that forms the southwestern margin of the Washakie Block (Figs. 1 and 3). It was named by Granger et al. (1971) who correlated it with the Wilson Creek orthogneiss of Perry (1965) and Barrus (1970). Our mapping shows that the zone is lithologically heterogeneous and measures up to 3 km across strike, much wider than is shown in the map of Granger et al. (1971). The shear zone has a gentle northeast dip and a top-to-the-southwest sense of shear (Hulsebosch, 1993). Where exposed, the footwall contact of the Mount Helen structural belt is sharp, with the footwall rock being a weakly foliated gneiss that resembles the Native Lake gneiss. We have not located the...
hanging-wall contact because it occurs in the steep terrane bordering the North Fork Bull Lake Creek. The extent of the Mount Helen shear zone shown on Figure 3 is the minimum width possible for the zone; it could be considerably wider.

The Mount Helen structural belt consists of intensely foliated, mylonitic, granitic, and tonalitic gneiss. The mylonitic foliation has been folded into tight isoclinal folds that locally preserve rootless hinges of an earlier folding event (Fig. 7). We refer to the earliest folds as $F_{4,1}$ and the isoclinal folds as $F_{4,2}$. Much of the compositional layering in the Mount Helen structural belt was produced by shear early in the deformation and may even predate the formation of $F_{4,1}$. The axial planes of $F_{4,1}$ folds have been folded into broad open $F_{4,3}$ folds (not shown on Fig. 7). The folding of $F_{4,2}$ around these open $F_{4,3}$ folds accounts for the dispersion of lineations from the Mount Helen structural belt (Fig. 7). The major schistosity in the Mount Helen structural belt is axial planar to $F_{4,2}$, hence is designated $S_{4,2}$. The $S_{4,2}$ schistosity anastomoses around giant boudins of weakly sheared rock. It is the anastomosing nature of this foliation that produces the dispersion of the foliations evident on the stereo plot (Fig. 7).

The boudins within the Mount Helen structural belt include weakly deformed granitic gneiss as well as isolated blocks of supracrustal rocks that may be as large as 30 m in any dimension. These supracrustal rocks include metaperidotites, fine-grained metabasites that may have been metabasalts, sulfidic quartzites, and rare metapelitic gneiss. The supracrustal inclusions commonly contain granulite assemblages and have a fabric that is oblique to that of the foliation of the Mount Helen structural belt. We consider these to be equivalent to the $S_{1,2}$ foliations seen in the Washakie block. On the margins of the inclusions are amphibolite-facies assemblages with foliations that are parallel to $S_{4,2}$ (Fig. 7).

About 30% of the area of the Mount Helen structural belt in the North Fork Bull Lake Creek area (Fig. 3) consists of weakly foliated leucogranite typical of that of the Bridger batholith (Hulsebosch, 1993). Some leucogranite dikes have been folded by $F_{4,2}$, whereas others completely cross-cut the fabric of the shear zone. From this situation, as well from the map pattern which shows the Mount Helen structural belt to be truncated by the Bridger batholith near the center of the range (Fig. 1), we conclude that the Mount Helen structural belt was active during the emplacement of the earliest stages of the Bridger batholith.

The northeast-trending fabric of the Mount Helen structural belt is common throughout the Bridger batholith and the migmatites that lie to the south. As noted above, the $S_{3,2}$ foliations lie parallel to the $S_{4,2}$ foliations, and the $D_3$ synform in the Medina Mountain area has been folded around an axis that is coincident with the general orientation of the $F_{4,2}$ folds. These relations suggest that $D_3$ and $D_2$ deformations may have been caused by the same event. Because the deformation style in the Medina Mountain sequence is distinctly different from that of the Mount Helen...
structural belt, and because no evidence indicates that the deformations in the two areas are indeed coeval, we have designated the deformation in each area as a distinct event.

A series of sheared gneisses is found in the Crescent Lake area (Fig. 8). Although the gneisses are locally highly migmatized by the Bridger batholith, enough areas of coherent sheared gneiss survive for structural analysis. The shear zone here has a northeast trend with a top-to-the-northwest sense of shear (Marshall, 1987). Despite the difference in trend, we suggest that it represents the same $D_4$ event as that in the Mount Helen structural belt, because in both areas the shearing was nearly coeval with the intrusion of the Bridger batholith. Furthermore, the country rock affected by the shearing in both areas shows a similar history (cf. Marshall, 1987; Hulsebosch, 1993). In the Crescent lake area the country rock was the early migmatite, whereas for the Mount Helen structural belt it was the Native Lake gneiss. In both areas, the gneiss hosted large inclusions of older supracrustal rocks. By correlating the sheared gneiss in the Crescent Lake area with the Mount Helen structural belt, we infer that the Mount Helen structural belt originally curved from northwest trending in the central portion of the range to northeast trending in the north (Fig. 2).

Bridger Batholith

The Bridger batholith, a weakly foliated orthogneiss that crops out over an area of approximately 650 km$^2$ in the north-central portion of the Wind River Range, is key to establishing the ages of structural events in the northern portion of the range. The batholith was emplaced at 2.67 Ga (Aleinikoff et al., 1989) and places the younger limit on the age of the $D_3$ and $D_4$ fabrics. The batholith ranges in composition from diorite to granite, with granodiorite being the most voluminous phase (Frost et al., 1998), and forms gradational contacts on all margins, which are shown as migmatite on Figure 1. As noted above, the batholith was emplaced late in the $D_4$ event and has a weak foliation. In the southern margin, the fabric has a northwest trend (Hulsebosch, 1993), whereas, in the north, it trends toward the northeast (Marshall, 1987). The foliation is present throughout the body but increases in intensity to the east, toward the Mount Helen structural belt.

South Pass Sequence

Unlike other supracrustal rocks in the Wind River Range, the supracrustal rocks at South Pass are weakly metamorphosed, preserve primary sedimentary structures, and present a coherent stratigraphy. The structurally lowest unit in the South Pass sequence (SPS) is the Diamond Springs formation, which consists of serpentinites and talc-chlorite schists, considered to be komatiitic flows and subvolcanic intrusions (Hausel, 1991). Resting above the Diamond Springs formation is the Goldman Meadows formation, a sequence of mature sediments and iron formation (Bayley et al., 1973) and the Roundtop Mountain greenstone, a series of metavolcanic rocks that range from tholeiitic to komatiitic in composition (Hausel, 1991). Most of the rocks exposed at South Pass are part of the Miners Delight formation and are thrust against the lower units of the SPS (Hull, 1988). The Miners Delight formation consists of two packages. The lower portion is a proximal graywacke that contains minor interbedded calc-alkalic lava flows. Thrust against this is an upper distal sequence that contains only deep-water turbidites (Hull, 1988). Preliminary U-Pb ages of zircons from dacites within the Miners Delight formation indicate that these rocks were deposited at 2.64 Ga (K. R. Chamberlain, unpublished data), only a few million years before the emplacement of the 2.63 Ga Louis Lake batholith.

The relationship between the SPS and the older rocks in the Wind River Range is not clear because the lower contact of the SPS has been
obliterated by the later emplacement of the Louis Lake batholith. Because the SPS contains several north-directed thrust faults, we believe that the SPS was thrust northward upon the craton. A primary sedimentary contact, however, cannot be discounted.

**Louis Lake Batholith**

The Louis Lake batholith, the largest post-tectonic batholith in the Wind River Range, was emplaced at 2.63 Ga (Frost et al., 1998). Although minor enclaves of gabbroic and dioritic rocks are present in the pluton, the major rock types are granodiorite and porphyritic granite. On the northeastern limits of outcrop, in the structurally deepest level of exposure of the Wind River uplift, both rock types contain pyroxene (Frost et al., 1998). East and south from this area, the pyroxene-bearing assemblages are replaced by hornblendic- and biotite-bearing assemblages. Along the southern margin, the Louis Lake batholith intrudes and metamorphoses the South Pass sequence, whereas in the north, the charnockitic portions of the batholith produce granulite-grade metamorphism in the earlier gneisses.

**Late Archean Batholiths**

The last Archean event in the Wind River Mountains was the emplacement of four plutons of porphyritic granite. These are the Bears Ears batholith in the east central, the Middle Mountain batholith in the north central, the Granite of New York Lakes in the northeast, and small plutons in the South Pass area. Whole-rock Pb-isotopic data from these rocks indicate that they are roughly coeval and were emplaced around 2.54 Ga (Stuckless et al., 1985).

**Post-Archean Events**

The Archean rocks of the range are cut by Proterozoic diabase dikes that locally preserve primary igneous textures. These dikes may be up to 50 m wide and can be followed for up to 30 km. Paleomagnetic and preliminary U-Pb baddeleyite ages from these rocks indicate that they are roughly coeval and were emplaced around 2.54 Ga (Stuckless et al., 1985).

The crystalline rocks of the range are also cut by numerous shear zones. Some of the zones are mylonitic, but most are brittle. These zones are commonly associated with chlorite- epidote- or hematite-rich alteration halos that may be as much as tens of meters wide. Although these zones locally have Laramide motion, the green-schist-grade assemblages that they contain are too high-temperature to be Laramide in age. They are considered, therefore, to have formed in the Proterozoic, although their exact age is still unknown. Some large fault zones can be traced from the overlying Phanerozoic rocks into the Archean basement. These are clearly Laramide in age, but they differ from the Proterozoic zones in that they are associated with clay alteration rather than greenschist-grade minerals (Mitra and Frost, 1981).

**METAMORPHIC HISTORY**

We recognize four distinct metamorphic events in the Wind River Range (Table 2), although we are aware that there were other thermal events that we cannot characterize because they were not associated with distinctive structural events or mineral assemblages. The earliest metamorphism, M1, is a granulite-facies event that is associated with D1 deformation. A thermal event certainly was associated with the emplacement of the 2.8 Ga Native Lake gneiss; however, we cannot distinguish assemblages formed from those formed during later metamorphism. The M2 is a regional event that accompanied the emplacement of the 2.67 Ga Bridger batholith. The M3 is a regional greenschist-amphibolite-grade event that accompanied deformation of the 2.65 Ga South Pass sequence. This metamorphism is not seen elsewhere in the range, either because the granitic gneisses of the range are relatively unreactive or because the metamorphism occurred while the South Pass rocks were being tectonically thrust onto the Wyomning craton. The M4 contact metamorphism is associated with the intrusion of the 2.63 Ga Louis Lake batholith and is probably limited to a zone within several kilometers of the batholith.

In addition to using published results of Bayley et al. (1973), Marshall (1987), and Sharp and Essene (1991), we determined the P-T conditions for the various metamorphic events using analyses of minerals from the samples listed in Table 3. Minerals from samples 95Bob25 and 85Fl-2A (see Fig. 1) were analyzed on the JEOL SUPER-PROBE 8900 at the University of Wyoming, whereas H-2 and H-150A (Table 4) come from unpublished analyses of Koesterer (1986) and were performed on a CAMECA™ CAMEBEX microprobe. In the thermobarometric calculations below we have used the TWQ program of Berman (1991) along with the data set of Berman (1988) and the solution models of Berman and Aranovich (1996).

Throughout the range, ion-exchange thermometers were extensively reset to low temperatures (Koesterer, 1986; Sharp and Essene, 1991; Hulsebosch, 1993). During cooling, garnet became more iron-rich whereas cordierite and orthopyroxene become more magnesian-rich.
Thus, in an effort to “see through” the effects of cooling, we have chosen the core compositions, which provide the most magnesian-rich garnet and the most iron-rich cordierite and orthopyroxene compositions (see Table 4).

**M₄ Event**

Before we discuss the metamorphic history of the Wind River Range, we must characterize the effect of Laramide uplift, because different levels of crust are exposed across the range and this will affect how one interprets the results of the thermo-barometry. To do this we use the thermobarometry of the M₄ event, which is the contact metamorphism related to the intrusion of the Louis Lake batholith. In the South Pass region it is characterized by the presence of Kfs in rocks containing And (Bayley et al., 1973) (mineral abbreviations after Kretz, 1983). The And-Sil reaction and the reaction Ms + Qtz = And + Kfs + H₂O intersect at H₂O pressures of 2.5 kilobars. At lower H₂O pressure this intersection moves to higher P. We therefore show the contact aureole to have formed at pressures around 2.5–3.0 kilobars and temperatures around 670–700 °C (Fig. 9). The charnockites along the north contact locally contain pelitic inclusions. One of these inclusions (FL85-2A) contains the assemblage Grt-Opx-Crd-Qtz. Using the displaced reactions written for the Fe and Mg end members from this assemblage, the TWQ program generates 11 equilibria that intersect between 740 °C and 840 bars. The dispersion in T and P is probably caused by Fe-Mg exchange that continued after the peak of metamorphism, as indicated by the discordancy of the cordierite-opx, cordierite-garnet, and opx-garnet Fe-Mg exchange thermometers (Fig. 9). Despite the scatter, it is clear that FL85-2A equilibrated at conditions that were consider-

<table>
<thead>
<tr>
<th>Event</th>
<th>Age (Ga)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>M₄</td>
<td>&gt;2.8</td>
<td>Granulite metamorphism associated with D₁</td>
</tr>
<tr>
<td>M₂</td>
<td>2.67</td>
<td>Amphibolite-granulite metamorphism associated with the emplacement of the Bridger batholith</td>
</tr>
<tr>
<td>M₃</td>
<td>2.65–2.63</td>
<td>Regional metamorphism of the South Pass sequence</td>
</tr>
<tr>
<td>M₃ₘ₉</td>
<td>2.63</td>
<td>Contact metamorphism of South Pass sequence on south side of Louis Lake batholith</td>
</tr>
<tr>
<td>M₃ₚ₉</td>
<td>2.63</td>
<td>Granulite-grade contact metamorphism on north side of Louis Lake batholith</td>
</tr>
</tbody>
</table>
ably hotter and at higher pressure than the rocks along the southern contact.

The barometry indicates that, at the time the Louis Lake batholith was intruded, the present northern contact was about 10 km deeper than the southern contact (Fig. 9). These localities are 65 km apart, and this uplift can be accommodated by a southward plunge in the Laramide structures of less than 10°. The attitude of Phanerozoic sedimentary rocks in the northern portion of the Wind River Range indicate that the Laramide structure is a broad fold that plunges 15°N (Mitra and Frost, 1981). If the southern portion of the structure plunges south at a similar angle, then the pressure difference recorded in the contact aureole of the Louis Lake batholith can easily be accommodated by Laramide deformation, although tilting by Archean or Proterozoic deformation events cannot be disproved.

**M₂ Event**

Although granulite assemblages are not uncommon in the older supracrustal rocks, it is unusual to find assemblages in which ion-exchange thermometers preserve the conditions of this metamorphism (Koesterer et al., 1987; Sharp and Essene, 1991). Because the ion-exchange thermometers in these rocks are strongly reset, distinctive mineral assemblages must be relied upon to constrain these metamorphic conditions. Sharp and Essene (1991) used the assemblage Spl-Sil-Rt-Qtz that occurs as inclusions within garnet to constrain the oldest metamorphic conditions in the Crescent Lake area. We find similar assemblages in the Medina Mountain area and in the Washakie terrane. The assemblage occurs as part of the matrix of the rock in these areas, however, with Sil oriented parallel to S₁, indicating that it formed as part of the M₁ event.

The best-preserved sample with this assemblage is 95Bob25, which comes from the Washakie terrane (Fig. 3). In this rock, Spl is associated with Sil and both are included in Crd or Grt. The Grt contains inclusions of Rt but not Ilm, whereas Ilm occurs in the matrix instead of Rt. As with Spl and Sil, Gar is rimmed by Crd. We infer that the primary assemblage was Qtz-Bt-Ksp-Pl-Grt-Sil-Spl-Rt, whereas the matrix assemblage, which we conclude was M₂, was Qtz-Bt-Ksp-Pl-Grt-Crd-Ilm. Garnet is weakly zoned with moderate enrichment (0.05 mole %) in aln on the margins; likewise XFe in Crd varies by less than 2 mole %. Plagioclase averages a Na₂ and varies by only 0.02 mole % an, with the rims being slightly more sodic than the cores.

Because M₃ was much hotter than the common closure temperature for diffusion in garnet, we consider that the compositions listed in Table 2 formed during M₃. It is likely that Crd, the most magnesian mineral in the rock, grew entirely during M₃, and we believe therefore that the Grt was probably more magnesian during M₃ than is shown by the analyses. It is impossible to back-calculate Grt composition because all Fe-Mg bearing phases (Grt, Bt, and Spl) changed composition as Crd grew. Thus, we would have to know the P-T path followed by the rock to back-calculate. We can use the M₂ compositions, however, to limit the conditions of M₃. Figure 10 shows the location of four limiting curves calculated from the given composition of Grt, Crd, and Spl. The assemblage Spl + Qtz is limited by two reactions:

\[ 2 \text{MgAl}_2\text{O}_4 + 5 \text{SiO}_2 = \text{Mg}_2\text{Al}_5\text{Si}_3\text{O}_{18} \]  

\[ 5 \text{SiO}_2 + 3 \text{MgAl}_2\text{O}_4 = 2 \text{Al}_2\text{Si}_3\text{O}_9 + \text{Mg}_2\text{Al}_3\text{Si}_3\text{O}_{12} \]

**TABLE 3. MINERAL ASSEMBLAGES OF ROCKS FROM THE WIND RIVER RANGE**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Grt</th>
<th>Crd</th>
<th>Opx</th>
<th>Sil</th>
<th>an</th>
<th>Spl</th>
<th>Rt</th>
</tr>
</thead>
<tbody>
<tr>
<td>85FL-2A</td>
<td>X</td>
<td>X</td>
<td>X</td>
<td>N.P.</td>
<td>0.34</td>
<td>N.P.</td>
<td>N.P.</td>
</tr>
<tr>
<td>95Bob25</td>
<td>X</td>
<td>X</td>
<td>N.P.</td>
<td>X</td>
<td>0.33</td>
<td>N.P.</td>
<td>N.P.</td>
</tr>
<tr>
<td>H-2</td>
<td>X</td>
<td>X</td>
<td>N.P.</td>
<td>*</td>
<td>0.29</td>
<td>N.P.</td>
<td>N.P.</td>
</tr>
<tr>
<td>H-150</td>
<td>X</td>
<td>X</td>
<td>N.P.</td>
<td>*</td>
<td>0.33</td>
<td>N.P.</td>
<td>N.P.</td>
</tr>
</tbody>
</table>

Notes: an—anothrite content of plagioclase; X—present in main assemblage, N.P.—absent; *—retrogressive. All rocks contain quartz, K-feldspar, and biotite.

**TABLE 4. MINERAL ANALYSES FROM THE WIND RIVER RANGE**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Garnet</th>
<th>Cordierite</th>
<th>Opx</th>
<th>Spinel</th>
</tr>
</thead>
<tbody>
<tr>
<td>85FL2A</td>
<td>95Bob25</td>
<td>H-2</td>
<td>H-150A</td>
<td>85FL2A</td>
</tr>
<tr>
<td>SiO₂</td>
<td>38.68</td>
<td>38.59</td>
<td>37.01</td>
<td>36.75</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.06</td>
<td>N.A.</td>
<td>0.00</td>
<td>0.01</td>
</tr>
<tr>
<td>Cr₂O₃</td>
<td>N.A.</td>
<td>N.A.</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>21.72</td>
<td>20.81</td>
<td>21.63</td>
<td>21.67</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>27.19</td>
<td>31.52</td>
<td>32.09</td>
<td>32.76</td>
</tr>
<tr>
<td>ZnO</td>
<td>N.A.</td>
<td>N.A.</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>MnO</td>
<td>0.50</td>
<td>0.63</td>
<td>0.97</td>
<td>0.99</td>
</tr>
<tr>
<td>CaO</td>
<td>1.21</td>
<td>2.28</td>
<td>1.27</td>
<td>1.51</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.00</td>
<td>N.A.</td>
<td>N.A.</td>
<td>N.A.</td>
</tr>
<tr>
<td>Total</td>
<td>99.05</td>
<td>100.80</td>
<td>99.62</td>
<td>99.69</td>
</tr>
</tbody>
</table>

Notes: Cation proportions on a basis of 12 oxygens for garnet, 18 for cordierite, 6 for Opx, and 12 for spinel.

\*Includes 0.02 wt% K₂O.

\†Includes 0.90 V₂O₃.
The reactions shown on Figure 10 are calculated for the displaced Mg-end members because these are the extreme limits of the assemblage. The locations displacing the Fe-end-member reactions lie within the field outlined by these reactions; if the minerals in the rock had retained their high-T, Fe-Mg compositions, the locations of the Fe end member and Mg end member for reactions (1) and (2) would have coincided.

The assemblage Rt + Grt is limited by the reaction:

$$3\text{TiO}_2 + \text{Fe}_2\text{Al}_5\text{Si}_3\text{O}_{12} = 3\text{FeTiO}_3 + \text{Al}_2\text{Si}_2\text{O}_5 + 2\text{SiO}_2$$

(1991) also had late Crd. Before the Crd grew in 95Bob25, we assumed that the Crd and Grt equilibrated during M2 and used the compositions of those estimated for the assemblage Grt-Spl-Sil-Qtz from the Crescent Lake area by Sharp and Essene (1991). It must be noted, however, that the matrix contains no Rt, the Crd must have grown at conditions below those indicated by GRAIL, or at pressures below around 5 kilobars (Fig. 10).

We also calculated M2 from two samples from the Medina Mountain area (H-2 and H-150A). Both of these samples have the assemblage Qtz-Crh-Ksp-Pl-Grt-Crd. These rocks differ from 95Bob25 in that they have not undergone M1 and that the whole assemblage was produced during M2 (although the mineral compositions were subsequently modified during cooling). In both rocks, the Grt is moderately zoned, with $X_{\text{Fe}}$ increasing by as much as 8 mol% from core to rim. Crd is not zoned and has $X_{\text{Fe}}$ that varies by about 2% around 0.29 in H2 and 0.32 in H-150A.

In addition to the Grt-Crd thermometer, there are two equilibria involving Grt, Pl, Crd, and Qtz:
Figure 10. P-T diagram showing conditions of the M₁ and M₂ metamorphic events in the Wind River Range. Reaction curves listed are calculated for the average spinel and the composition of core garnet and cordierite from sample 95Bob25. Ruled area is the stability field for garnet-cordierite-quartz-ilmenite. Samples H2 and H150A come from the Medina Mountain area. M, 87—Marshall (1987); S&E, 91—Sharp and Essene (1991). See text for discussion.

\[
\begin{align*}
2 \text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3 \text{SiO}_2 &= 2 \text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3 \text{Fe}_2\text{Al}_2\text{Si}_3\text{O}_{18} \quad (3) \\
2 \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3 \text{SiO}_2 &= 2 \text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 3 \text{Mg}_2\text{Al}_2\text{Si}_3\text{O}_{18} \quad (4)
\end{align*}
\]

By using the most Mg-rich Grt and the most Fe-rich Crd in these rocks, we obtained an intersection of the Grt-Crd thermometer with reactions (3) and (4) at temperatures of ~700 °C and pressures of 4.5–5.5 kilobars (Fig. 10). These conditions are at slightly higher temperature than that obtained from 95Bob25. The differences in T and possibly P may reflect Laramide tilting, since the Medina Mountain rocks, which are structurally deeper in the Laramide uplift, record higher grade conditions.

We conclude that M₂ formed at moderate pressures (4–5 kilobars) and at temperatures around or slightly below 700 °C. This conclusion is consistent with the estimate of metamorphism in the Mount Helen structural belt (Hulsebosch, 1993) and with the core and rim compositions of “matrix” mineral assemblages described by Sharp and Essene (1991), although the temperatures in the Crescent Lake area may have been somewhat higher (~750 °C) than those in the core of the range.

M₃ Event

Evidence of the M₃ event is seen in the regional metamorphism of the South Pass sequence. Although some of the units at South Pass record greenschist conditions, this metamorphism was within the low amphibolite facies over most of the area. We have no quantitative thermobarometry from the area. The coexistence of Crd and And reported by Bayley et al. (1973), however, allow us to limit the conditions to 450–550 °C and 2–3 kilobars (Fig. 11).

Summary

A summary of the metamorphic conditions in the Wind River range is shown in Figure 11. Of particular note is the wide swath in P-T space occupied by the M₃ event. As noted above, this probably reflects Laramide tilting of the range and suggests that, if the other metamorphic events could have been sampled over a suitably wide area, they would record a similar range of conditions. The rocks recording the M₃ event come from locations that are structurally high in the range. Were we to find rocks that equilibrated during this event on the western side of the range, we might expect that they would record even higher T and P. Likewise, had we more locations where M₃ could be estimated and had we better precision, we might be able to see the effect of Laramide tilting on M₃ as well. Finally, because we recognize a nearly 3 kilobar difference between the northern contact of the Louis Lake batholith and the southern contact, and because M₃ followed shortly after M₂, it is likely that the rocks farther north were much deeper and presumably hotter at the time M₃ was being developed on the southern portion of the range.

\section*{P-T TIME PATH FOR THE WIND RIVER RANGE}

We can use our estimation of P-T conditions, the age constraints, and the geological relations outlined above to constrain the temperature path followed by the rocks in the Wind River Range during the Late Archean (Fig. 12). The path is poorly constrained because of the vast amount of time involved, because of the few points (both in space and time) where we could constrain conditions, and by the certainty that each of the metamorphic conditions that occurred over the huge area of the Wind River Range probably was associated with a range of temperature conditions. Even so, it is clear that during this 300 m.y. period of time the rocks in the Wind River Range experienced periodic burial and exhumation. We show the path beginning at surface conditions to account for the deposition of the earliest supracrustal rocks, although the path from this point to the peak conditions at M₁ may have been very long and probably involved thermal events, the evidence for which has long since been lost. The first thermal pulse that we can identify is M₁ at ~800 °C and 6 kilobars (see Fig. 12). As noted above, this event predates emplacement of the ca. 2.83 Ga Native Lake gneiss and the Victor dikes, the age of which is unknown. The X-ordinate in Figure 12 is dashed in this range to emphasize this age uncertainty. The metamorphism could be as old as 3.2 or 3.3 Ga, as indicated by U-Pb SHRIMP (super high-resolution ion microprobe) ages from zircons in migmaitic (Aleinikoff et al., 1989), or it may have predated the Native Lake gneiss by only a few tens of millions of years.

The Victor dikes are porphyritic and locally chloritized, and, therefore, they must have been emplaced into country rocks that were at temperatures lower than M₁. This temperature is unconstrained; although we show it to have been ~200 °C, it could have been much lower or higher. We also have no direct constraints on the emplacement temperature of the Native Lake gneiss. We infer that it must have been emplaced at temperatures >700 °C because it was at least locally chloronormitic.
The Medina Mountain Sequence was either deposited directly on the Native Lake gneiss or was thrust onto it at low temperatures. Thus, we contend that some portions of the Wind River Range were exposed or near the surface after the cooling of the Native Lake gneiss. After deposition of the Medina Mountain sequence, the range was buried again. Some of this may have been accompanied by tectonic thickening along the Mount Helen structural belt and some may have been produced by surface accumulation of effusive components of the Bridger batholith. By the time the Bridger batholith crystallized, the conditions were around 700 °C with pressures of 4.0–5.5 kilobars.

After the cooling of the Bridger batholith at least the southern portion of the range was exhumed to low pressure because the South Pass Sequence was either deposited upon the basement or thrust upon it under low temperature conditions. Shortly thereafter, the Louis Lake batholith was emplaced, bringing temperature again into the 800 °C range. We do not know the thermal history after the cooling of the Louis Lake plutons; we show the area to have cooled slightly before the latest Archean plutons were emplaced at 2.55 Ga and to have cooled slowly thereafter. We postulate this slow cooling to account for the extensive resetting of the ion-exchange thermometers from the M₄ metamorphism.

CONCLUSIONS

Metamorphic and Isotopic Resetting

The Wind River Range was periodically intruded by magmas from ca. 2.8 Ga to 2.55 Ga; this fact accounts for many of the complexities in determining ages and P-T conditions for various rocks from the area. In most places, the original M₁ granulite assemblages were retrograded to amphibolite facies. Distinct hydration halos are seen surrounding dikes of both Bridger and Louis Lake granitic rocks, indicating that during the Late Archean these rocks were periodically flooded with fluids as well as magmas. In those few areas where the original granulite assemblages are preserved, the ion-exchange thermometers have been extensively reset to temperatures that are far too low for granulite assemblages (below 500 °C for two-pyroxene and olivine-spinel thermometers; Koesterer et al., 1987). Even the granulites formed during the latest metamorphism show evidence of significant ion-exchange on cooling (Fig. 10), suggesting that the rocks stayed hot after the intrusion of the Louis Lake batholith.

The prolonged magmatic history and its associated influx of melts and fluids into the crust may explain the complex U-Pb systematics in the older rocks of the range. For example, Aleinikoff et al. (1989) found that a migmatite yielded ²⁰⁷Pb/²⁰⁶Pb SHRIMP ages from zircon that clustered at 2.65, 2.72, 2.85, 3.2, 3.3, and 3.8 Ga. Whereas the older ages reported by Aleinikoff et al. (1989) may be ages of individual detrital grains, the younger three ages probably reflect recrystallization of zircon during the Late Archean magmatism that affected the area. Similarly, DeWolf et al. (1993) found a complex age distribution in monazite from the Crescent Lake area, wherein a single grain yielded ²⁰⁷Pb/²⁰⁶Pb ages that clustered around 2.78, 2.66, and 2.54 Ga. They interpreted these ages to represent distinct periods of monazite growth. It is likely that the two younger ages represent monazite growth due to the influx of magmatic fluids, because the samples studied by DeWolf et al. (1993) occur as a complex roof pen-
taneous with the 2.64 Ga gold mineralization, led to the development of the Wind River Range. The Wind River Range represents a long-lived active continental margin during the Late Archean.

ACKNOWLEDGMENTS

Much of the mapping upon which this paper was based was conducted by Tom Hulsebosch as a seasonal employee of the U.S. Geological Survey and later as part of his Ph.D. dissertation. Tom was killed in a traffic accident in September 1996, and we dedicate this paper to his memory. This manuscript benefited greatly by reviews by Peter Dahl, Dave Mogk, and John Percival.

REFERENCES CITED


