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Notes

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. D_1 is a penetrative deformation that occurred during the earliest granulite-facies metamorphism; D_2 is a folding event, probably in amphibolite facies, that deforms porphyritic dikes that cut the D_1 fabrics. Both D_1 and D_2 predate the intrusion of the ca. 2.8 Ga Native Lake gneiss. D_3 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. D_4 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 D_3 structures in the Medina Mountain sequence. We consider D_3 and D_4 to be coeval with the emplacement of the Bridger batholith, and hence to date at ca. 2.67 Ga. The latest structures (D_5) are fabrics associated with the folding and thrusting of the 2.65 Ga South Pass sequence.

We recognize at least four metamorphic events. M_1 is associated with the D_1 fabrics and occurred at high *T* (>750 °C) and high *P* (~7–8 kilobars). M_2 (650–750 °C and 4–5.5 kilobars) is associated with the intrusion of the Bridger batholith and formation of the D_3 and D_4 structures. The D_5 structures of the South Pass sequence record M_3, which is low *P* (~2–3 kilobars) and low *T* (~500 °C). The final metamorphism, M_4, 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 km^2. 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).

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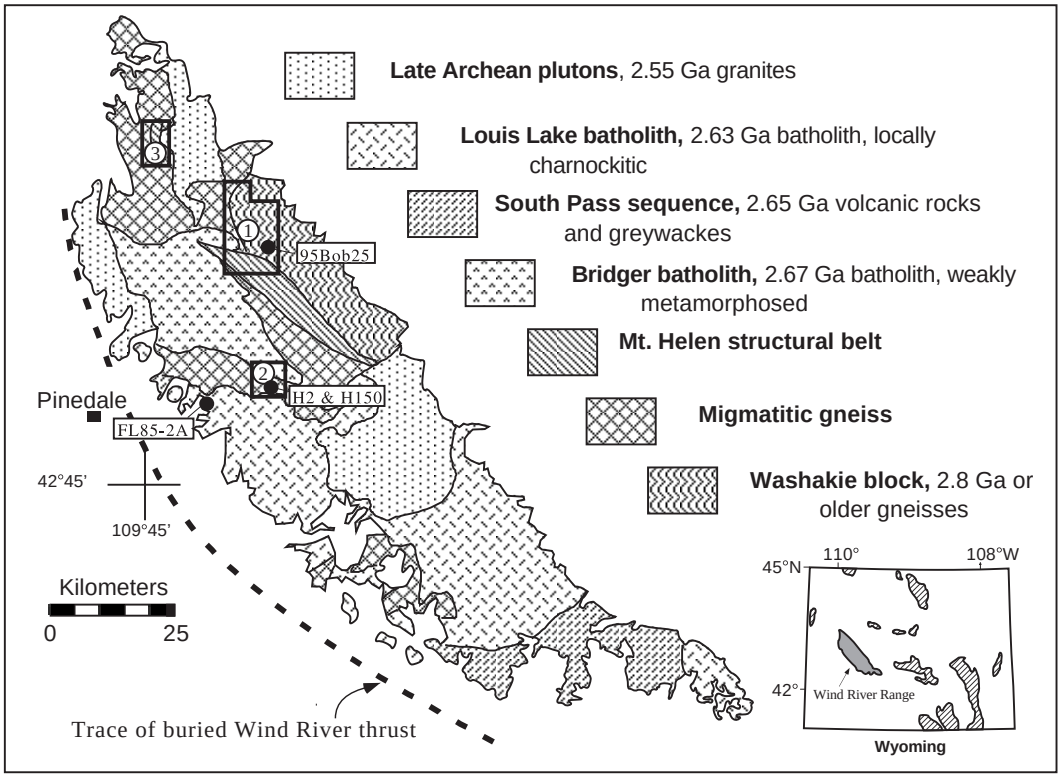

Figure 1. Geologic map of the Wind River Range compiled from Granger et al. (1971), Worl et al., (1986), Koesterer et al. (1987), Marshall (1987), Hulsebosch (1993), and our unpublished data. Boxes outline areas discussed in detail in the text. These are: (1) Dry Creek–Bob Lakes–North Fork Bull Lake Creek region, (2) Medina Mountain area, (3) Crescent Lake area. Dots locate samples used in thermobarometry.

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-alkalic 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).

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., D_1, D_2,

TABLE 1. GEOLOGIC EVENTS IN THE WIND RIVER RANGE

Event	Age (Ga)	Where found
Deposition of early supracrustal rocks, granulite metamorphism, D_1	>2.8	CL, MHSB, MM, WB
Intrusion of porphyritic dikes	>2.8	CL, MM, WB
D_2: folding of D_1 fabric and porphyritic dikes	>2.8	WB
Intrusion of Native Lake gneiss	ca. 2.8	CL, MM, WB
Deposition of Medina Mountain sequence, metamorphism, and D_3: folding of these rocks.	2.7–2.8	MM
Emplacement of the Bridger batholith, and D_4: formation of Mount Helen structural belt and folding of D_3 structures	2.67	BB,CL,MHSB, MM, WB
Deposition of South Pass sequence, metamorphism, D_5: folding of SPS and thrusting onto Wyoming craton	2.65	SP
Intrusion of Louis Lake batholith, contact metamorphism	2.63	LL, SP
Intrusion of Late Archean plutons	2.54	All of the range
Intrusion of Proterozoic diabase dikes	1.47	All of the range

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.

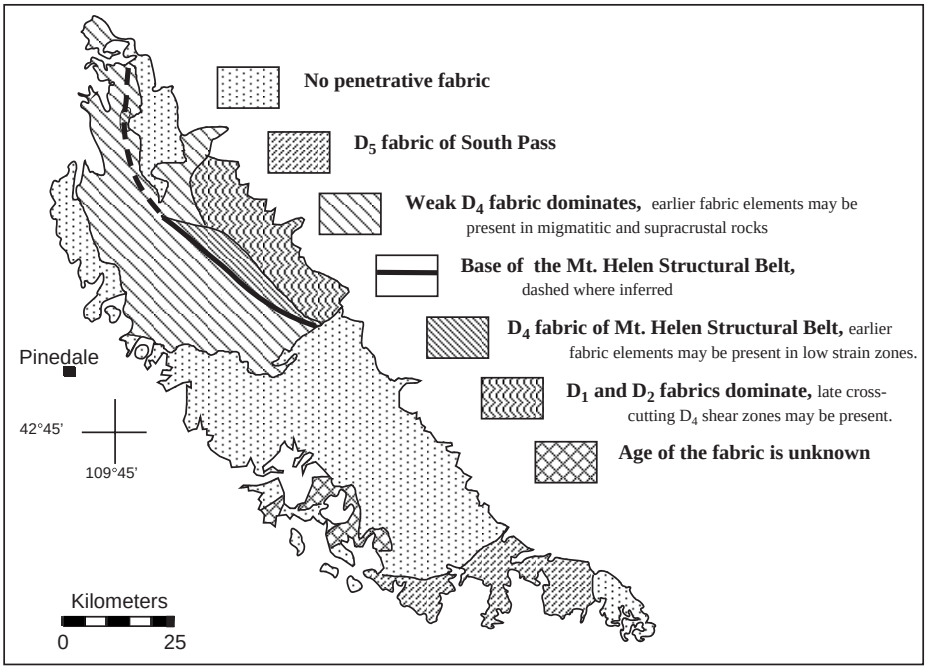

Figure 2. Structural domains in the Wind River Range.

etc.) with an older event having a lower number in the subscript. Each fabric element is labeled with twin subscripts—the first refers to the deformation event and the second to the generation of that particular feature within the event.

We divide the Wind River Range into five structural domains (Fig. 2). The oldest is in the Washakie block, in which younger fabrics are absent or are present only as sparsely occurring cross-cutting mylonite zones. The dominant fabrics in this block were produced by the D_1 and D_2 events. This block is bounded on the south by a major deformation zone, the Mount Helen structural belt (Granger et al., 1971). This D_4 zone has thrust displacement and separates the Washakie block from a domain where the earlier fabrics have been largely obliterated by the Bridger batholith. A weak S_4 fabric is found throughout the Bridger batholith and older fabrics occur locally in this domain. For example, D_3 fabrics are recognized in the Medina Mountain area (Fig. 1, area 2), and D_1 and possibly D_2 fabric elements may occur in the migmatites on the southern and northern margin of the Bridger batholith. The youngest structural domain in the range is defined by D_5 deformation of the South Pass sequence. Most of the central portion of the range is underlain by directionless fabrics of the Louis Lake and Late Archean batholiths, which were intruded after D_5.

Early Supracrustal Rocks and the D_1 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 upper amphibolite-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 (Koesterer et al., 1987), Crescent Lake area (Worl, 1968; Marshall, 1987; Sharp and Essene, 1991), and Mount Helen structural belt (Hulsebosch, 1993) (Fig. 1). In these areas, however, later migmatization has disrupted the supracrustal sequence to the extent that early fabric relations are obliterated.

In the Washakie block, supracrustal rocks occur as lenses that may extend for a kilometer or more (Fig. 3) interfolded with gray gneisses. In several places they clearly occur in the keels of tight, isoclinal $F_{1,2}$ synforms. Locally within these synforms, refolded isoclinal $F_{1,1}$ folds occur. The major schistosity in this area lies parallel to the limbs of the $F_{1,2}$ folds, and we designate it as $S_{1,2}$ (Fig. 4). Where the rocks are in granulite facies, the preferred orientation of granulite minerals lies parallel to $S_{1,2}$. Thus, we conclude that at least the later stages of D_1 occurred during granulite metamorphism.

It is unclear whether the supracrustal rocks were deposited on the gray gneisses, are in tectonic contact with them or are intruded by the gray gneisses. Locally, we have recognized D_1 folds and granulite assemblages in the gray gneisses. This fact means that at least some of the gray gneisses were present during the earliest structural and metamorphic events recognized in the range.

Porphyritic Dikes

Porphyritic, weakly foliated amphibolitized dikes cut the D_1 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 D_2 folds that also deform the D_1 fabrics. This is clear evidence that the dikes were intruded between the D_1 and D_2 events. Porphyritic dikes that cut D_1 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.

D_2 Event

The $S_{1,2}$ 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 $S_{1,2}$ fabric, are folded by these later folds, we conclude that this folding event is a distinct deformation event, which we call D_2.

Foliations and lineations from the Dry Creek and Bob Lakes domains are plotted on stereo diagrams on Figure 4. The foliations measured are dominantly $S_{1,2}$. D_2 produced weak mineral-preferred orientations in the fold hinges that are easily distinguished from the compositional layering that is distinctive of $S_{1,2}$. Distinction between lineations is ambiguous. The $F_{2,1}$ event

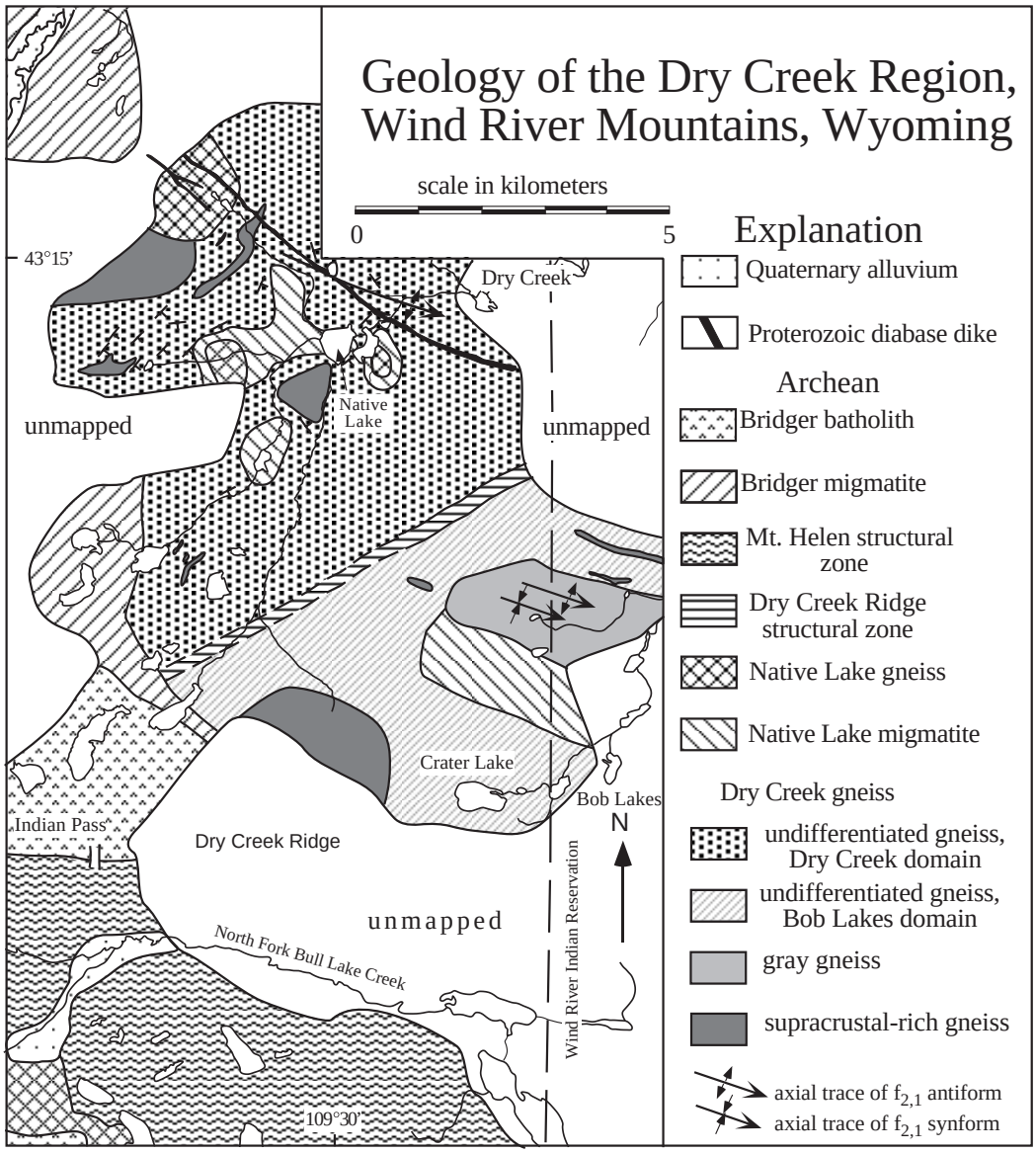

Figure 3. Geologic map of the Dry Fork–Bob Creek–Bull Lake Creek region (area 1, Fig. 1). Map covers portions of U.S. Geological Survey Bob Lakes, Fremont Peak North, Hays Park, and Ink Wells 7.5′ Quadrangles. Geology by B. R. Frost, T. Hulsebosch, and K. Chamberlain, 1982–1995.

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-alkalic 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.

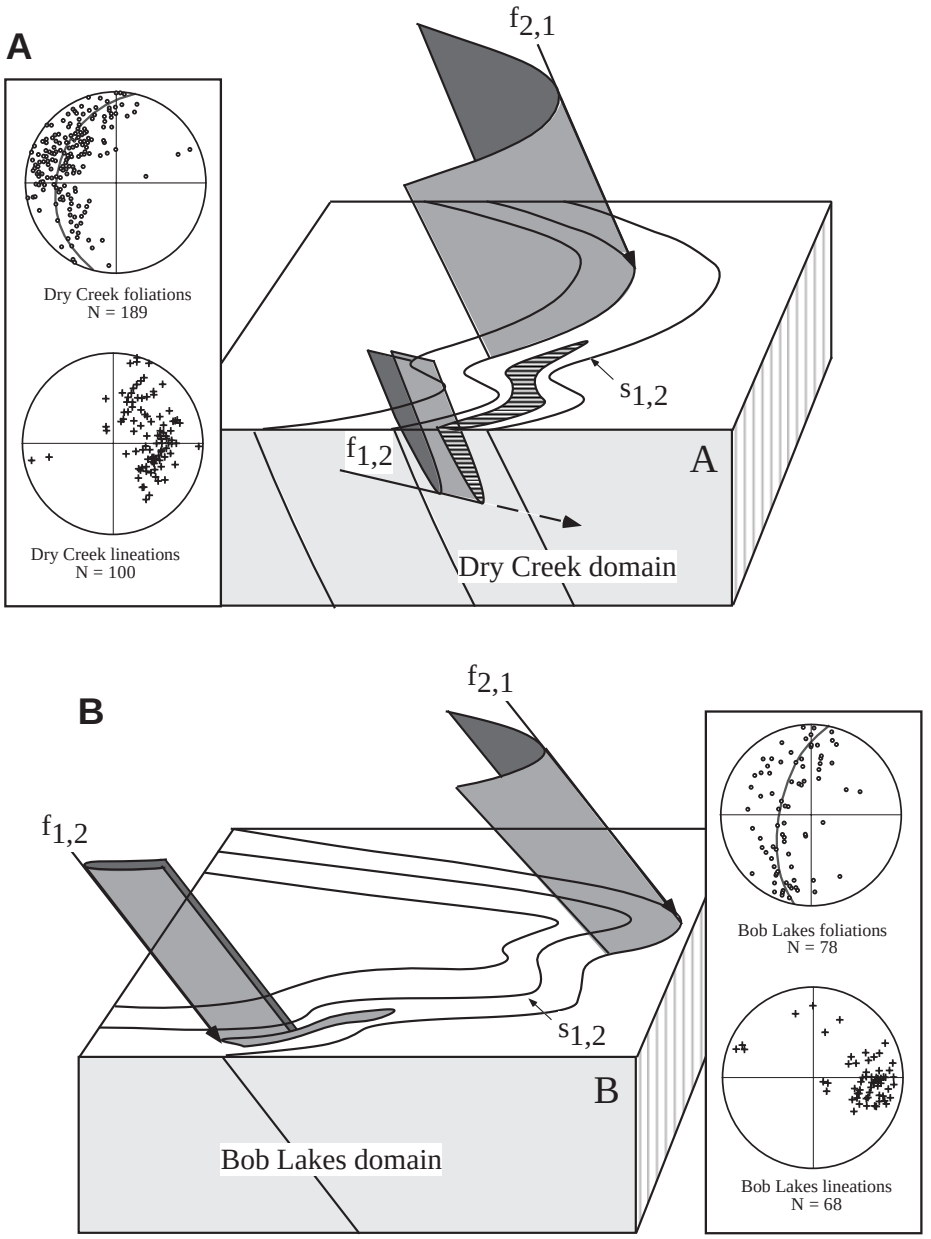

Figure 4. Sketch comparing the structural framework of the Dry Creek and Bob Lakes domains. In both areas, the supracrustal rocks contain rootless isoclinal folds not shown on the figure that we label as $f_{1,1}$.

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 D_3

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 (D_3) 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 $S_{3,1}$ and the major synforms $F_{3,1}$ because these folds are caused by the earliest deformation event that we can identify. The $S_{3,1}$ 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 $L_{3,1}$ lineations have been folded into a broad fold, which is expressed in the field by the way that the $F_{3,1}$ folds form closed ovoid structures (Fig. 6). We consider this later fold to be a D_4 structure because it has the same trend as the $F_{4,2}$ folds in the Mount Helen structural belt.

Mount Helen Structural Belt and D_4

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

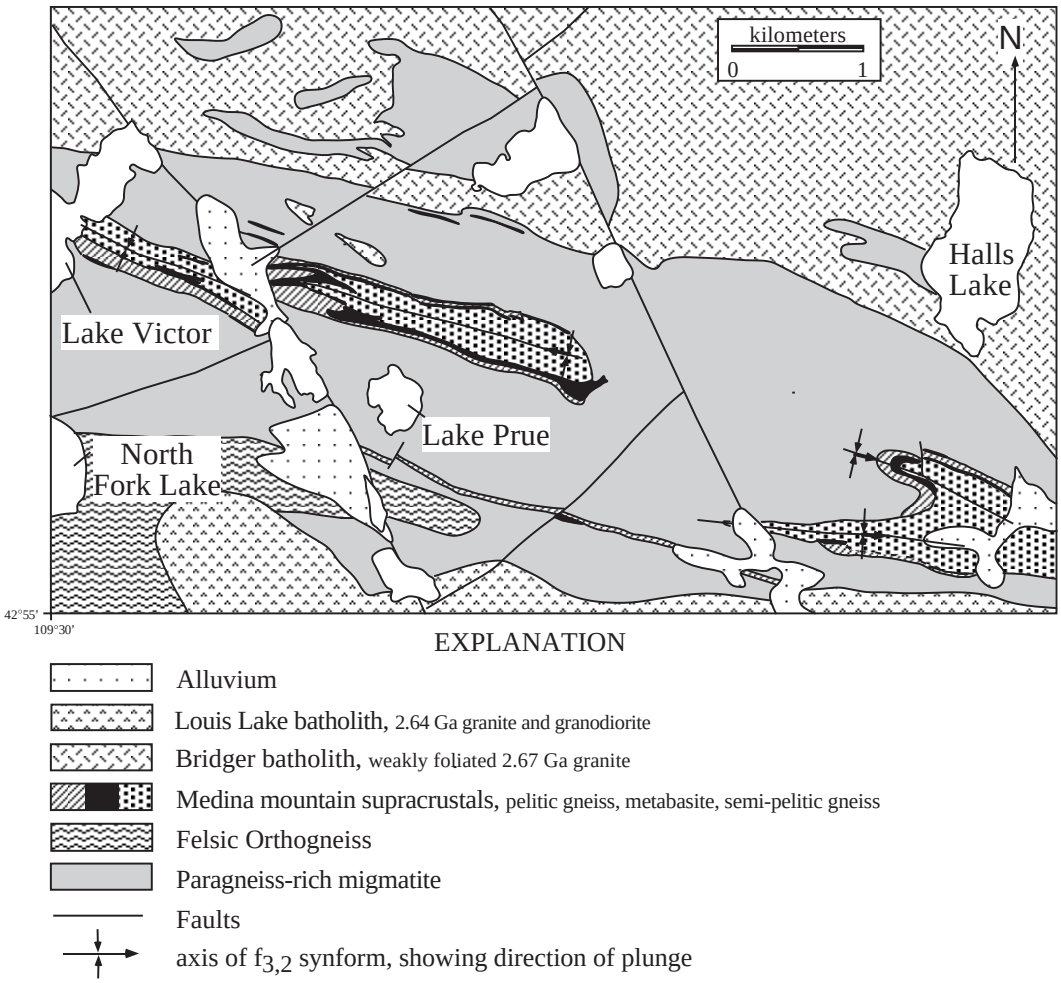

Figure 5. Geologic map of the Medina Mountain area (area 2, Fig. 1), modified from Koesterer (1986). Map covers areas of the U.S. Geological Survey Hall Mountain 7.5′ Quadrangle map.

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 $F_{4,2}$ 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 as 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_4 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

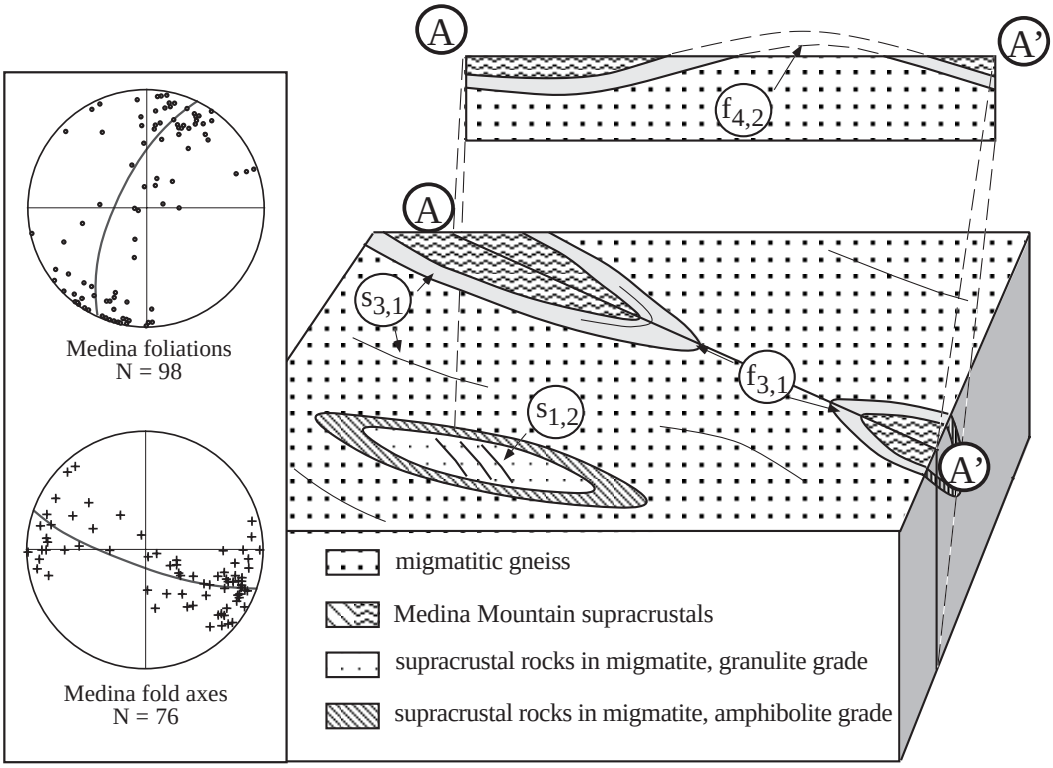

Figure 6. Sketch of the structural relations in the Medina Mountain area looking north. Earliest fabric elements are found in isolated blocks of supracrustal rocks floating in migmatite. We designate the schistosity in these rocks as $S_{1,2}$ on the basis of similarity to fabric elements in the Washakie block. Included within the migmatites are supracrustal rocks that are recorded in the keels of $f_{3,1}$ synforms. The axes of these folds have been folded as shown in cross section A–A′ around open folds that we suggest formed in the D_4 event.

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 2 630 Ma (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 hornblende- 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.

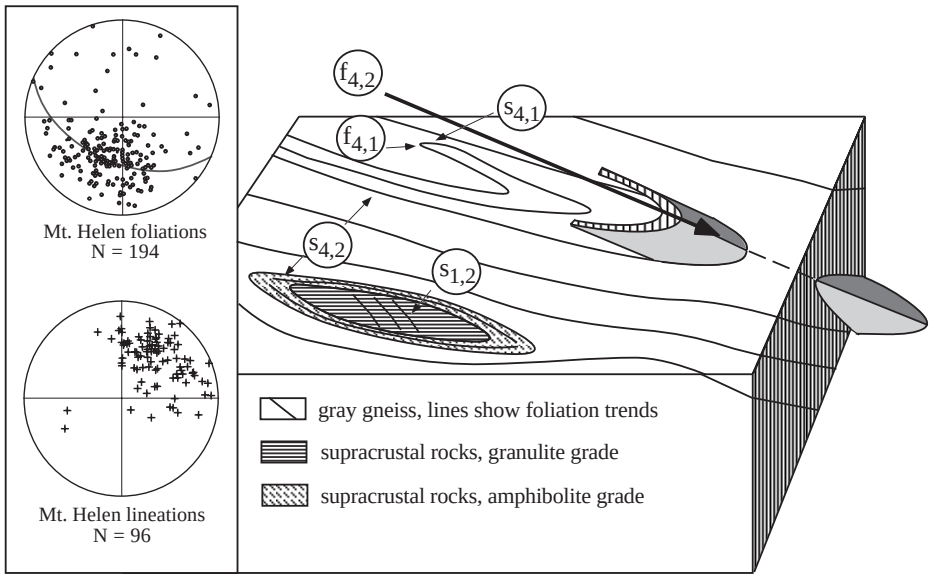

Figure 7. Sketch of structural relations in the Mount Helen structural belt looking roughly north. The earliest structural elements are foliations found in granulite-grade inclusions within the mylonitic gneiss. These we label as $s_{1,2}$. The earliest fabric element in the mylonitic gneiss is an intense mylonitic foliation that has been folded into at least two isoclinal folds, which we designate $f_{4,1}$ and $f_{4,2}$. Not shown are the broad, open $f_{4,3}$ folds that fold the axial planes of the $f_{4,2}$ folds and bodies of the Bridger granite, which may be involved in the $f_{4,2}$ and $f_{4,3}$ folds, but which may also cut the shear zone entirely.

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 Fork 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 of these dikes indicate that there may be two ages of dikes, one at around 2.0 Ga (Harlan, personal commun.) and another at 1.47 Ga (Chamberlain and Frost, 1995).

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 greenschist-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, M_1, is a granulite-facies event that is associated with D_1 deformation. A thermal event certainly was associated with the emplacement of the 2.8 Ga Native Lake gneiss; however, we cannot distinguish assemblages formed then from those formed during later metamorphism. The M_2 is a regional event that accompanied the emplacement of the 2.67 Ga Bridger batholith. The M_3 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 Wyoming craton. The M_4 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 SUPERPROBE 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.

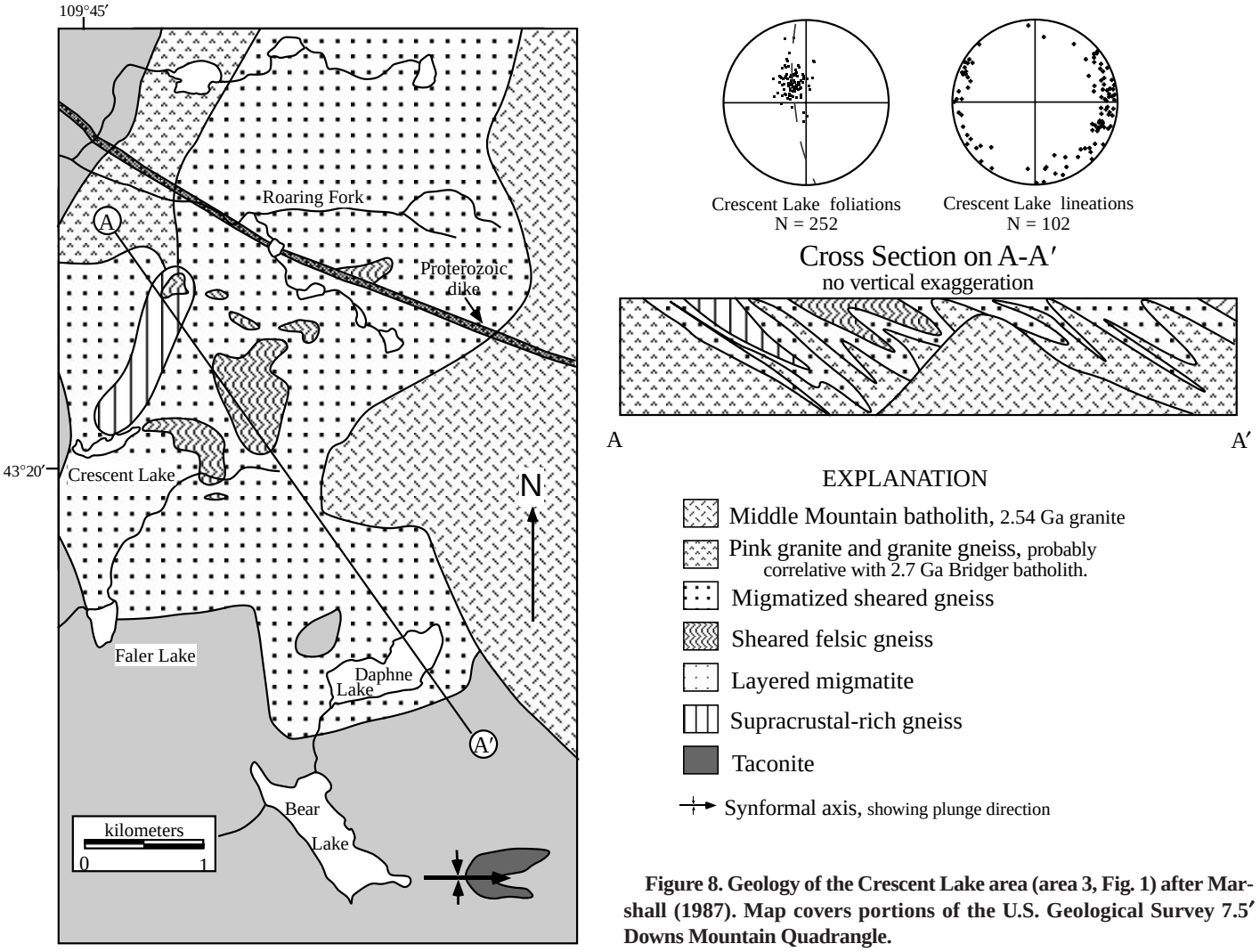

Figure 8. Geology of the Crescent Lake area (area 3, Fig. 1) after Marshall (1987). Map covers portions of the U.S. Geological Survey 7.5′ Downs Mountain Quadrangle.

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).

M_4 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 thermobarometry. To do this we use the thermobarometry of the M_4 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 + H_2O intersect at H_2O pressures of 2.5 kilobars. At lower H_2O 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 °C and 5 400 and 6 800 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 2. METAMORPHIC EVENTS IN THE WIND RIVER RANGE

Event	Age (Ga)	Description
M_1	>2.8	Granulite metamorphism associated with D_1
M_2	2.67	Amphibolite-granulite metamorphism associated with the emplacement of the Bridger batholith
M_3	2.65–2.63	Regional metamorphism of the South Pass sequence
M_{4S}	2.63	Contact metamorphism of South Pass sequence on south side of Louis Lake batholith
M_{4N}	2.63	Granulite-grade contact metamorphism on north side of Louis Lake batholith

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 Laramie deformation, although tilting by Archean or Proterozoic deformation events cannot be disproved.

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

Sample	Grt	Crd	Opx	Sil	an	Spl	Rt
85FL-2A	X	X	X	N.P.	0.34	N.P.	N.P.
95Bob25	X	X	N.P.	X	0.33	X	X
H-2	X	X	N.P.	*	0.29	N.P.	N.P.
H-150	X	X	N.P.	*	0.33	N.P.	N.P.

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

M_1 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 earliest 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 $S_{1,2}$, indicating that it formed as part of the M_1 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 M_2, was Qtz-Bt-Ksp-Pl-Grt-Crd-Ilm. Garnet is weakly zoned with moderate enrichment (0.05 mole %) in alm on the margins; likewise X_{Fe} in Crd varies by less than 2 mole %. Plagioclase averages an_{68} and varies by only 0.02 mole % an, with the rims being slightly more sodic than the cores.

Because M_2 was much hotter than the common closure temperature for diffusion in garnet, we consider that the compositions listed in Table 2 formed during M_2. It is likely that Crd, the most magnesian mineral in the rock, grew entirely during M_2, and we believe therefore that the Grt was probably more magnesian during M_1 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 M_2 compositions, however, to limit the conditions of M_1. 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:

$$2\ MgAl_2O_4 + 5\ SiO_2 = Mg_2Al_4Si_5O_{18} \quad (1)$$

$$5\ SiO_2 + 3\ MgAl_2O_4 = 2\ Al_2SiO_5 + Mg_3Al_2Si_3O_{12} \quad (2)$$

TABLE 4. MINERAL ANALYSES FROM THE WIND RIVER RANGE[1]

	Garnet				Cordierite				Opx	Spinel
Sample	85Fl2A	95Bob25	H-2	H-150A	85Fl2A	95Bob25	H-2	H-150A	85Fl2A	95Bob25
SiO_2	38.68	38.59	37.01	36.75	49.77	50.19	48.46	48.93	47.93	0.00
TiO_2	0.06	N.A.	0.00	0.01	0.00	0.00	0.03	0.00	0.09	0.07
Cr_2O_3	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	4.02
Al_2O_3	21.72	20.81	21.63	21.67	34.29	33.65	33.93	33.95	6.89	56.28
FeO	27.19	31.52	32.09	32.76	5.16	5.75	6.64	6.91	25.23	30.70
ZnO	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	4.07
MnO	0.50	0.63	0.97	0.99	0.01	0.00	0.04	0.07	0.04	0.06
MgO	9.69	6.96	6.65	6.00	10.43	10.09	9.79	9.75	19.40	4.24
CaO	1.21	2.28	1.27	1.51	0.01	0.00	0.03	0.00	0.07	N.A.
Na_2O	0.00	N.A.	N.A.	N.A.	0.03	0.28	0.05	0.11	0.00	N.A.
Total	99.05	100.80	99.62	99.69	99.72*	99.95	99.15	99.72	99.65	100.34†
Note: Cation proportions on a basis of 12 oxygens for garnet, 18 for cordierite, 6 for Opx, and 12 for spinel.										
Si	2.999	3.013	2.938	2.928	4.975	5.019	4.919	4.938	1.817	0.000
Ti	0.004	0.000	0.000	0.001	0.000	0.000	0.032	0.000	0.003	0.003
Cr	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	N.A.	0.136
Al	1.985	1.916	2.023	2.035	4.040	3.967	4.059	4.038	0.308	2.833
Fe	1.763	2.058	2.130	2.183	0.432	0.481	0.564	0.583	0.800	0.731
Mn	0.033	0.042	0.065	0.067	0.001	0.000	0.004	0.006	0.001	0.002
Mg	1.120	0.810	0.787	0.713	1.554	1.504	1.482	1.467	1.096	0.180
Ca	0.100	0.191	0.108	0.129	0.001	0.000	0.003	0.000	0.003	N.A.
Na	0.000	N.A.	N.A.	N.A.	0.005	0.054	0.011	0.026	0.000	N.A.
Almd	0.585	0.664	0.689	0.706						
Py	0.371	0.261	0.255	0.231						
Gros	0.033	0.062	0.035	0.042						
Spes	0.011	0.014	0.021	0.022						
XFe	0.612	0.718	0.730	0.754	0.218	0.242	0.276	0.284	0.422	0.802

Notes: Analyses of minerals from H-2 and H-150A come from Koesterer (1986); opx—orthopyroxene; N.A.—not analyzed.
*Includes 0.02 wt% K_2O.
†Includes 0.90 V_2O_3.

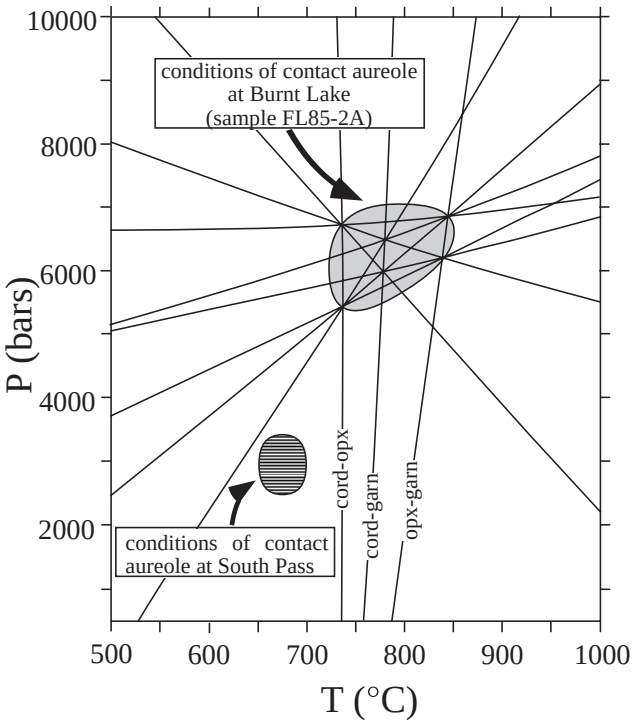

Figure 9. *P-T* conditions for metamorphism along the northern and southern margins of the Louis Lake batholith. Sample M5-2A with the assemblage garnet-cordierite-orthopyroxene-quartz comes from the northern contact, where the pluton is charnockitic. Eleven equilibria generated from the Fe and Mg end-members of these phases using the TWQ program of Berman (1991) intersect in an area around 6 kilobars and temperatures around 750 °C. The southern contact with an assemblage K-feldspar-andalusite-sillimanite-muscovite-quartz (Bailey et al., 1973) records temperatures below 700 °C and pressures of 2.5–3.0 kilobars. We interpret the differences in these conditions to record different degrees of crustal exhumation in response to Laramide deformation.

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:

$$3\,TiO_2 + Fe_3Al_2Si_3O_{12} = 3\,FeTiO_3 + Al_2SiO_5 + 2\,SiO_2 \quad \text{(GRAIL)}$$

Our sample has Rt but not Ilm in the core of the Grt. Therefore, to estimate the location of GRAIL in Figure 10, we used the most magnesian Grt, pure Rt (the Rt in the rock shows only Ti peaks on EDS) and pure Ilm. This establishes the minimum pressure for this reaction, because solution of Mg and Fe_2O_3 into Ilm will stabilize it to higher *P*. The fourth reaction on Figure 10 is the Grt-Crd thermometer, which, as noted above, was calculated using the most Mg-rich Grt and the most Fe-rich Crd.

We consider that the Grt-Rt-Spl-Sil-Qtz assemblage in sample 95Bob25 formed at conditions indicated by the vertically ruled area in Figure 10. This field is limited at high pressures by reaction (2), at low temperatures by reaction (1), and at low pressures by GRAIL. As noted above we contend that Crd growth during M_2 caused all minerals to become more Fe-rich as Mg was sequestered in Crd. If the original Grt and Spl were more Mg–rich, reactions (1) and (2) would both have moved to higher *T*, restricting the stability field for Spl + Qtz.

Although imprecise, the conditions determined for M_1 in 95Bob25 are consistent with those estimated for the assemblage Grt-Spl-Sil-Qtz-Rt-Ilm from the Crescent Lake area by Sharp and Essene (1991). It must be noted, however, that the rocks studied by Sharp and Essene (1991) also had late Crd. Before the Crd grew in these rocks, the Grt and Spl were likely to have been richer in Mg and this change in bulk composition would cause the parallelogram in Figure 10 to move to lower *P* and higher *T*. A small change in Grt bulk composition would have caused the Sharp and Essene (1991) estimate to overlap the *P-T* estimate by Marshall (1987). Marshall (1987) estimated conditions from the occurrence of sapphirine in metaperidotitic blackwall rocks with assemblages that bracketed the reaction Crd + Spl = Opx + Spr. Because of the Laramide tilting, there is no reason to expect that the metamorphic conditions for M_1 were the same in 95Bob25 as at Crescent Lake, which is located 25 km to the northwest. Our results and those of Marshall (1987) and Sharp and Essene (1991) indicate, however, that M_1 occurred at high pressure (6–8 kilobars) and was very hot (T > 800 °C). This is consistent with textural relations suggestive of pigeonite in iron formation from the northernmost margin of the range (Sykes, 1985) and with locally preserved (but strongly reequilibrated) garnet in mafic granulites elsewhere in the range.

M_2 Event

The M_2 event is recorded in amphibolite-facies halos around granulite boudins in the Mount Helen structural belt and in assemblages in the Medina Mountain sequence. In most areas, M_2 was in upper amphibolite facies, however, locally in the Medina Mountain area, where metabasites abut partially melted pelite gneiss, granulite-grade assemblages are formed. This local granulite metamorphism probably reflects areas of low water activity around granite melts (Koesterer, 1986).

We calculated the conditions for the M_2 event from the cordierite-bearing assemblage in 95Bob25. We assumed that the Crd and Grt equilibrated during M_2 and used the compositions of the Grt core and the most Fe-rich Crd. Because 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 M_2 from two samples from the Medina Mountain area (H-2 and H-150A). Both of these samples have the assemblage Qtz-Bt-Ksp-Pl-Grt-Crd. These rocks differ from 95Bob25 in that they have not undergone M_1 and that the whole assemblage was produced during M_2 (although the mineral compositions were subsequently modified during cooling). In both rocks, the Grt is moderately zoned, with X_{alm} increasing by as much as 8 mol% from core to rim. Crd is not zoned and has X_{Fe} that varies by only 3 mol%. Pl is irregularly zoned and varies by about 2% around an_{29} in H2 and an_{32} in H-150A.

In addition to the Grt-Crd thermometer, there are two equilibria involving Grt, Pl, Crd, and Qtz:

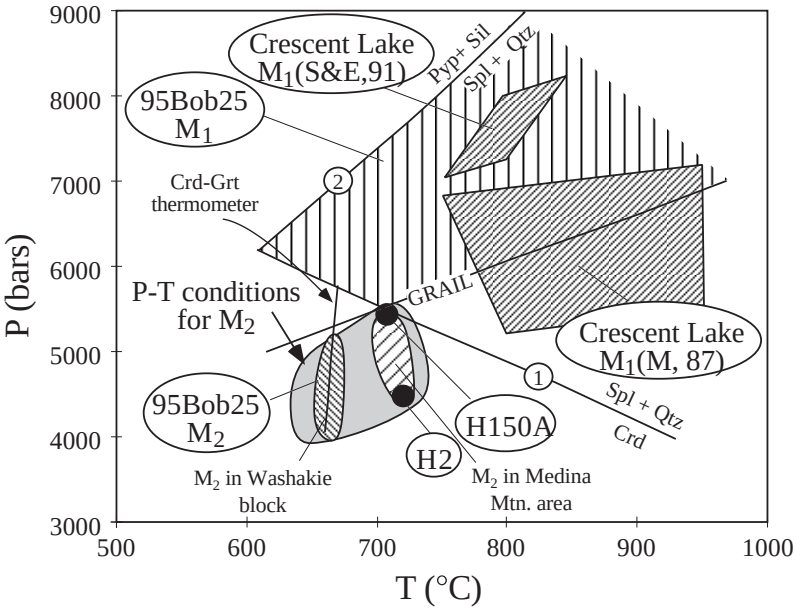

Figure 10. *P-T* diagram showing conditions of the M_1 and M_2 metamorphic events in the Wind River Range. Reaction curves listed are calculated for the average spinel and the composition of core garnet and cordierite from sample 95Bob25. Ruled area is the stability field for the M_1 assemblage garnet-sillimanite-spinel-quartz-rutile. Area indicated as M_2 for this sample 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.

$$2\ Fe_3Al_2Si_3O_{12} + 6\ CaAl_2Si_2O_8 + 3\ SiO_2 = 2\ Ca_3Al_2Si_3O_{12} + 3\ Fe_2Al_4Si_5O_{18} \quad (3)$$

$$2\ Mg_3Al_2Si_3O_{12} + 6\ CaAl_2Si_2O_8 + 3\ SiO_2 = 2\ Ca_3Al_2Si_3O_{12} + 3\ Mg_2Al_4Si_5O_{18} \quad (4)$$

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 M_2 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.

M_3 Event

Evidence of the M_3 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 M_4 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 M_1 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 M_2 could be estimated and had we better precision, we might be able to see the effect of Laramide tilting on M_2 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 M_4 followed shortly after M_3, it is likely that the rocks farther north were much deeper and presumably hotter at the time M_3 was being developed on the southern portion of the range.

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 M_1 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 M_1 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 migmatite (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 chilled, and, therefore, they must have been emplaced into country rocks that were at temperatures lower than M_1. 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 charnockitic.

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 eruptive 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 M_4 metamorphism.

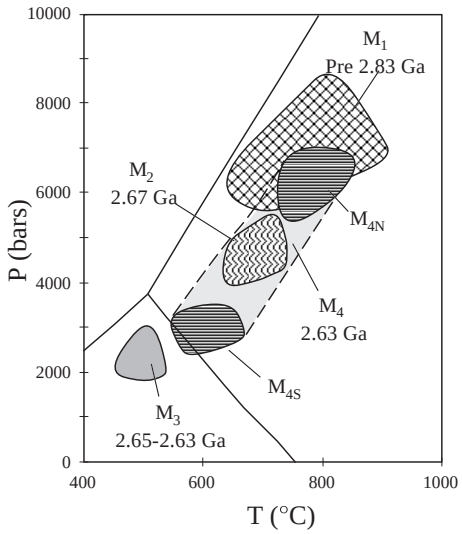

Figure 11. *P-T* diagram comparing conditions of the M_1, M_2, M_3, and M_4 metamorphic events.

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 M_1 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 $^{207}Pb/^{206}Pb$ 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 $^{207}Pb/^{206}Pb$ 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-

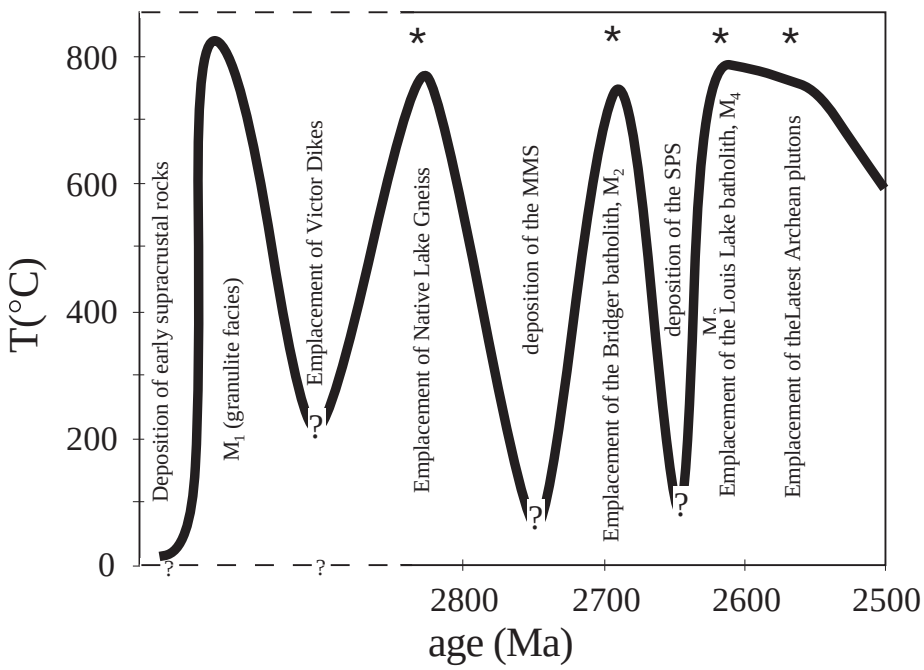

Figure 12. Archean thermal history of the Wind River Range incorporating the thermometry observed from the metamorphic rocks. Asterisks denote age constraints discussed by Frost et al. (1998).

dant that lies on the contact between the 2.7 Ga Bridger batholith and the 2.54 Ga Middle Mountain batholith (Fig. 8).

Tectonic Implications

Although we cannot determine the polarity of the 2.8 Ga magmatic event in the Wyoming province, distribution of the Late Archean granitoids indicates that subduction forming the 2.7, 2.63, and 2.54 Ga plutons was directed from the west and south toward the craton (Frost et al., 1998). During the Phanerozoic, three hundred million years is not an exceptionally long period of time for an active continental margin to remain stationary. The locus of calc-alkalic magmatism on the western margin of North and South America has occupied essentially the same position for nearly that long. For example, the Canadian Cordillera has been the site for magmatism periodically for the past 200 million years (Armstrong and Ward, 1991, 1993). Likewise, the Sierra Nevada was magmatically active from 180 to 70 Ma and is still active in the North today (Armstrong and Ward, 1991). Like the Wind River Range, both the Canadian Cordillera and the Sierra Nevada areas contain fragments of sedimentary sequences that have from time to time been deposited or thrust onto the magmatic arc.

The existence of active continental margins, in which calc-alkalic plutons have been emplaced into an older continental crust, has been postulated in many Archean terranes. In some localities, such as Lac des Iles area of the Superior Province (Brügmann et al., 1997), the Eastern Goldfields superterrane of the Yilgarn province (Swager, 1997), and the northern magmatic belt of Zimbabwe (Kusky, 1998), the arcs were short-lived and were built upon crust that was only slightly older than the plutons themselves. Other areas, such as the Limpopo belt and the Slave and the western Yilgarn provinces, are similar to the Wyoming province, in which calc-alkalic plutons were emplaced into a basement that was considerably older than the plutons.

In the central zone of the Limpopo belt, calc-alkalic plutons dated from 2.6 to 2.55 Ga were emplaced into a basement as old as 3.2 Ga. (Holzer et al., 1998). The older rocks of the Slave province are invaded by voluminous 2.70–2.58 Ga granitoids (van Breemen et al., 1992). The western Yilgarn province, including the Murchinson, Lake Grace, Boddington, and Balingup terranes, consists of fragments of old greenstones and gneisses, including some as old as 3.5 Ga, included in large volumes of Late Archean granites. Granitic magmatism lasted from 2.70 to 2.60 Ga in the Murchison province (Wiedenbeck and Watkins, 1993); in the other terranes in the western Yilgarn, granitoids date from 2.67 to 2.53 Ga. (Wilde et al., 1996).

No consensus exists on the origin of the Late Archean calc-alkalic plutons in the Slave and Yilgarn cratons. The 2625 to 2695 Ma plutons in the Slave craton have been ascribed to subduction, but it is unclear whether the subduction was west-dipping (Kusky, 1989) or east-dipping (Fyson and Helmstaed, 1988). Likewise, although Archean granites in each of the blocks in the Yilgarn province are of different ages, ages do not progress across the province (Wilde et al., 1996). This lack of progression, along with a widespread distribution of 2.64 Ga gold mineralization, led Qiu and Groves (1999) to conclude that the widespread postorogenic plutons in the Yilgarn are not products of subduction but rather were produced by mantle delamination. Thus, the composition, age, and distribution of granites and gneisses in the Wind River Range represent the best documented example of a long-lived active continental margin during the Late Archean.

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