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Ductile Deformation fo Mid–Crustal Archean Rocks in the Foreland of Early Proterozoic Orogenies, Central Laramie Mountains, Wyoming

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INTRODUCTION

The Laramie Mountains of southeastern Wyoming contain an exposed core of Precambrian rocks that range in age from greater than 2.7 Ga to 1.43 Ga. Rocks in the central part of the range are predominantly Archean granitic gneiss, metasedimentary rocks and greenstone (Snyder, 1984, 1986) that represent the southeasternmost exposures of the Wyoming Archean province. In the southern parts of the range, Archean rocks were intruded ~1.43 Ga by the Laramie Anorthosite Complex and Sherman Granite (**Fig. 1**). However, recent radiometric dating in the central Laramie Mountains has also identified Early Proterozoic ages of ~2.01 Ga for widespread mafic dikes and local granitic plutons that intruded the Archean rocks. Deformation features and fabrics within these Early Proterozoic units and comparable features in their Archean host rocks indicate that the southeastern margin of the Wyoming Archean province sustained a complex history of thick-skinned deformation at mid-crustal levels during at least two periods of Early Proterozoic deformation. This field trip examines evidence for the Early Proterozoic deformation and considers the relationship of deformation features to the Trans-Hudson orogen, Cheyenne belt suture, and Central Plains orogen that mark the boundaries of this southeastern margin of the Wyoming province (**Fig. 2**).

Much of the Archean crustal block north of the Laramie Anorthosite Complex and south of the Laramie Peak shear zone, referred to here as the central Laramie Mountains (**Fig. 1**), has undergone regional, thick-skinned uplift (Chamberlain et al., 1993) and complex ductile reworking (Bauer et al., 1995, 1996) during the Early Proterozoic. Stops on the first day of the trip are along the southern margin of this deformed and uplifted block (**Fig. 3**). In this area, supracrustal rocks in the Elmers Rock

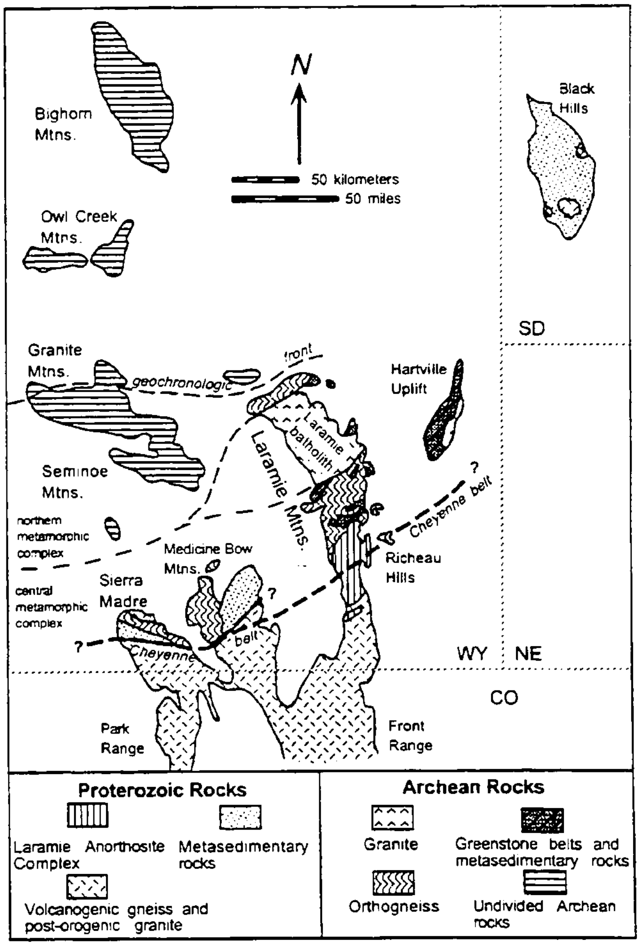

Figure 1. Geologic sketch map of eastern Wyoming and parts of adjacent states showing exposures of Precambrian rocks and subdivisions of the Archean and Proterozoic terranes as discussed in the text.

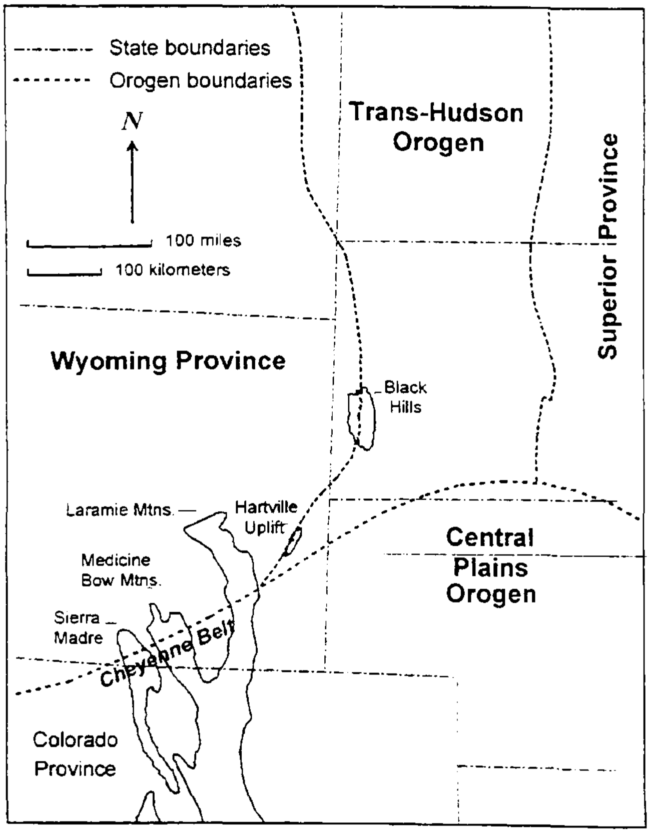

Figure 2. Map illustrating the regional geologic setting of the southeastern Wyoming Province.

greenstone belt, adjacent areas of granitic gneiss, and Early Proterozoic mafic dikes that intruded the Archean units contain fabrics and folds associated with at least two periods of Early Proterozoic deformation. East-west-trending deformation fabrics and features along this southern margin of the Archean terrane, may be related to collision along the Cheyenne belt suture zone, but they have also been locally reoriented by northwest-trending folds. Fabrics in Early Proterozoic mafic dikes constrain the timing of this deformation to younger than 2.01 Ga (Cox et al., 1995; Snyder et al., 1995).

Stops on the morning of the second day include granitic gneiss, Early Proterozoic granites and mafic dikes along the eastern margin of the range. These rocks contain north-south trending folds and fabrics that may be a product of the Trans-Hudson orogeny. However, these deformation features have also been locally deformed and reoriented by younger ductile deformation. Stops during the afternoon of the second day include granitic gneiss and a mafic dike that have been deformed in the Laramie Peak shear zone (Resor et al., 1996) that marks the northern boundary of the reworked and uplifted Archean block. Syndeformational sphene from a mafic dike deformed in the shear zone yielded a U-Pb age of 1,763 ± 7 Ma. This age is interpreted as the time of ductile shearing along the Laramie Peak shear zone (Resor et al., 1996).

REGIONAL TECTONIC SETTING

The southeastern margin of the Wyoming Archean province developed as a rifted passive continental margin during the Early Proterozoic, broadly constrained from ~2.3 to 1.9 Ga (Karlstrom et al., 1983; Karlstrom and Houston, 1984). Igneous products of this event include a well-developed mafic dike swarm (Kennedy Dike swarm of Graff et al., 1982) and local granitic plutons. These rocks intruded Archean gneiss and supracrustal rocks that make up most of the Archean rocks along the rifted margin. Early Proterozoic passive margin sedimentary rocks, ranging from an early fluvial succession through deltaic and local carbonate platform units (Karlstrom et al., 1981), are preserved along the southern margin of the craton in the Medicine Bow Mountains and Sierra Madre (**Fig. 1**).

Following deposition of the passive margin sedimentary sequence, the margin of the Wyoming province sustained a complex period of Early Proterozoic convergent margin tectonism associated with development of the Trans-Hudson orogen, Cheyenne belt, and Central Plains orogen (**Fig.** 2). Among these features, the Cheyenne belt is the best exposed and most widely studied.

The Cheyenne Belt

The Cheyenne belt (**Fig. 1**), marks the southern boundary of the Wyoming province and the northern boundary of the Colorado Proterozoic province. It has been interpreted as an Early Proterozoic collision zone (~1,780 Ma) marking the boundary between the Archean craton of the southern Wyoming province and accreted Early Proterozoic island-arc terranes to the south (Hills and Houston, 1979; Karlstrom and Houston, 1984; Duebendorfer and Houston, 1986). Deformation along the collision zone was initiated by ~1,780 Ma and may have continued to as late as ~1,740 Ma (Premo and Van Schmus, 1989). Well-exposed segments of the Cheyenne belt in both the Medicine Bow Mountains and Sierre Madre display a complex structural history of shortening, thrusting, and late dextral transcurrent shear.

The Early Proterozoic rocks of the passive margin sequence north of the Cheyenne belt include as much as 10 km of quartzite-dominated rocks of the Snowy Pass Supergroup in the Medicine Bow Mountains as well as quartzite and marble in the Sierre Madre that have been correlated with the Snowy Pass Supergroup (Houston et al., 1978; Karlstrom et al., 1983). These rocks record a transgressive depositional sequence ranging upward from dominantly fluvial conditions in the Deep Lake Group through deltaic/shallow marine to deep-water marine conditions in the overlying Libby Creek Group.

In contrast to the Early Proterozoic miogeoclinal sequence (Snowy Pass Supergroup) north of the belt, the rocks south of the Cheyenne belt, in both the Medicine Bow Mountains and the Sierra Madre, include a complex Early Proterozoic eugeoclinal sequence. These rocks include: (1) gneiss and migmatite of probable mixed volcanic-sedimentary origin (Hills and Houston, 1979; Houston et al., 1989); (2) interlayered mafic and felsic metavolcanic rocks and associated metasedimentary rocks of the Green Mountain Formation (Divis, 1976; Condie and Shadel, 1984); and (3) plutonic rocks, ranging from pre - to post-tectonic and from mafic to intermediate to felsic compositions. The maximum ages obtained for rocks of this sequence are 1,792 ± 15 Ma for metavolcanic rocks of the Green Mountain Formation and 1,779 ± 5 Ma for plutonic rocks intruded into the Green Mountain Formation (Premo and Van Schmus, 1989). No trace of an Archean component has been recognized in Nd isotopes south of the Cheyenne belt (DePaolo, 1981; Condie, 1982; Ball and Farmer, 1991).

Structural analysis by Duebendorfer and Houston (1986, 1987) in and along the Cheyenne belt defined three distinct deformational events. From oldest to youngest, these are: (1) north-directed thrusting associated with suturing of Proterozoic island arc-marginal basin rocks to the southern margin of the Archean Wyoming craton; (2) local dextral strike-slip faulting along the southern margin of the belt, under greenschist-facies conditions; and (3) late-stage cataclasis. Duebendorfer (1988) presented evidence for an inverted metamorphic gradient in the Snowy Pass Supergroup that he interpreted to be a result of the early north-directed thrusting of amphibolite-facies rocks over the Snowy Pass Supergroup along the northern mylonite zone.

A Tectonic Model for the Cheyenne Belt

On the basis of the lithologic, sedimentologic, structural, and geochronologic data noted above and geophysical data discussed below, Houston and coworkers (Hills and Houston, 1979; Houston et al., 1979; Karlstrom et al., 1983; Karlstrom and Houston, 1984; Duebendorfer and Houston, 1986; Houston et al., 1989; Houston, 1993) proposed and refined a convergent margin plate tectonic model for the Cheyenne belt and adjacent areas during the Early Proterozoic. General aspects of the model include:

(1) Progressive rifting of the southeastern Wyoming craton between about 2.3 and 2.0 Ga, and development of a passive margin with miogeoclinal sedimentation (Snowy Range Supergroup).
(2) Development of oceanic island arcs south of the Wyoming Archean craton (~1.8 Ga).
(3) Closure of the marine basin along a south-dipping subduction zone by accretion of younger island-arc terranes from the south starting ~1.8 Ga.
(4) Tectonic burial of the southern margin of the Archean craton beneath a fold-and-thrust belt rooted in the Cheyenne belt. Thrust faulting, occurring between 1,780 and 1,750 Ma (Premo and Van Schmus, 1989; Houston et al., 1989), was followed by steepening of the rocks in the thrust root zone and subsequent dextral strike-slip displacement along parts of the Cheyenne belt.
(5) Erosion of the thrust-belt rocks from the Archean Wyoming craton north of the Cheyenne belt.

Regional geophysical data are consistent with this model. Seismic reflection profiles across the Cheyenne belt (Allmendinger et al., 1982; Tempelton and Smithson, 1994) display south-dipping reflectors that are consistent with north-directed thrusting and south-dipping subduction. North-south gravity profiles across the Sierra Madre, Medicine Bow Mountains, and Laramie Mountains (Johnson et al., 1984) are also consistent with this interpretation. In each of the three profiles, models of the long-wavelength gravity data indicate a simple ramp in the Moho about 20-50 km south of the Cheyenne belt, with the base of the crust stepped down to the south about 5-8 km. The position of these steps is consistent with a south-dipping contact between the Wyoming and Colorado provinces.

Although Proterozoic thrust belt rocks are no longer present over a broad extent of the Wyoming craton (point 5 above), geochronologic evidence from the Archean rocks north of the Cheyenne belt suggests significant burial of much of the southeastern margin of Wyoming province during the Early Proterozoic. The Archean basement rocks within a 150 km-wide zone that extends from the Cheyenne belt to the northern Laramie Mountains and southern Granite Mountains (**Fig. 1**) have K-Ar and Rb-Sr biotite ages of 1500-1300 Ma (Hills and Armstrong, 1974; Peterman and Hildreth, 1978). North of this zone, both the whole-rock ages and the mineral ages are Archean and agree within error. The Proterozoic mineral ages have been interpreted to reflect progressive cooling of the Archean rocks during uplift (Hills and Armstrong, 1974; Peterman and Hildreth, 1978) and erosion of the overlying Proterozoic thrust-belt pile (Karlstrom and Houston, 1984).

The Cheyenne Belt and the Central Plains Orogen

The northeastern projection of the Cheyenne belt into the Laramie Mountains (**Fig. 1**) is concealed by the intrusion of the Laramie Anorthosite Complex (~1,434 Ma; Scoates and Chamberlain, 1995) and Sherman Granite at ~1,435 (Aleinikoff, 1983; Chamberlain, unpublished data). The extrapolated projection of the Cheyenne belt passes southeast of the Hartville uplift and possibly near the Richeau Hills (**Fig. 1**). However, a more definitive location for the Cheyenne belt is difficult because most of the Precambrian basement east of the Laramie Mountains is covered by Phanerozoic sedimentary rocks. Metasupracrustal rocks in the Laramie Mountains, well to the south of the projection of Cheyenne belt (**Fig. 1**), include metavolcanic and metasedimentary rocks similar to, and probably correlative with, those in the Medicine Bow Mountains and Sierra Madre south of the Cheyenne belt (Houston et al., 1989).

The projection of the Cheyenne belt to the east (**Fig. 2**) also marks the northern boundary of the Central Plains orogen (Sims and Peterman, 1986), which is delineated based on subsurface drill-hole data from Nebraska, Kansas, and Missouri. Recovered metamorphic and granitoid rocks from the orogen have ages in the range 1.63-1.80 Ga. Based on isotopic ages, rocks types, and regional gravity and magnetic data, Sims and Peterman (1986) concluded that the Central Plains orogen is the eastward extension of the Early Proterozoic orogen south of the Cheyenne belt. However, the relationship of the northern margin of the Central Plains orogen and the Cheyenne belt is not clear due to the extensive Phanerozoic cover east of the Laramie Mountains.

The Trans-Hudson Orogen

In the north-central United States, the Trans-Hudson orogen (THO) (Hoffman, 1981) bounds the eastern margin of the Wyoming province and separates the Wyoming province from the western margin of the Archean Superior province to the east (**Fig. 2**). However, the orogen is best exposed and has been studied most extensively in northern Manitoba and Saskatchewan (e.g., Lewry and Collerson, 1990) where it separates the Superior province and the Rae and Hearne Archean provinces to the northwest. Studies in this region indicate that the THO formed during a complex history of microcontinent and juvenile Early Proterozoic island arc accretion during collision (~1.83-1.80 Ga) of the Superior craton with the combined Wyoming, Hearne, and Rae provinces (cf., Bickford et al., 1990).

From the exposures in northern Canada, the THO can be traced for 100s of kilometers to the south, beneath Phanerozoic sedimentary cover, based on geophysical data and drill core (e.g., Dutch and Nielsen, 1990; Klasner and King, 1986, 1990; Thomas et al., 1987), to its apparent truncation in northwestern Nebraska by the Central Plains orogen (**Fig. 2**). This north-south trending subsurface segment of the orogen, known as the Dakota segment (Dutch and Nielsen, 1990), is partially exposed in the uplifted core of the Black Hills. The western core of the Black Hills contains reworked Archean basement of the Wyoming craton that is overlain to the east by a complex sequence of Early Proterozoic metasedimentary rocks that sustained pronounced east-west shortening during the THO (Redden et al., 1990).

Klasner and King (1990) presented a model for the THO in North and South Dakota, based on gravity and drill hole data, that depicts the Superior craton partially underthrusting the Wyoming craton and juvenile terranes within the THO. They also suggest that large slices of the Superior craton were uplifted toward the east along deep, west-dipping thrust faults. East-west trending COCORP seismic reflection profiles, taken near the U.S. - Canada border (Nelson et al., 1993; Baird et al., 1996) both support and suggest modifications to Klasner and King's model. West-dipping reflectors in the western part of the profile support Klasner and Kings suggestion of west-dipping underthrusting of the Wyoming craton. However, the reflectors change to an easterly dip in the eastern part of the profile, suggesting a crustal-scale anticlinorium and possible east-dipping overthrusting of the Superior province over a continental fragment lodged between the Superior and Wyoming provinces during terminal collision (Baird et al., 1996).

GEOLOGY OF THE CENTRAL LARAMIE MOUNTAINS

General Overview

The Precambrian exposures in the Laramie Mountains can be subdivided into four segments (**Figs.** 1 and 3); from north to south these are: (1) Archean granitic orthogneiss and volcanic and metasedimentary supracrustal rocks along the northern margin of the Laramie batholith. This segment is part of the northern metamorphic complex of Condie (1969). (2) The Laramie batholith (2.5 -2.6 Ga, Rb-Sr whole rock dates of Hills and Armstrong, 1974) and Esterbrook greenstone belt. The batholith is a medium - to coarse-grained, massive to locally gneissic biotite monzogranite. The southern boundary of the batholith with the next segment to the south is marked by a complex system of northeast-striking shear zones, collectively referred to as the Laramie Peak shear zone (Chamberlain et al., 1993; Resor et al., 1996). (3) The central Laramie Mountains , which is the focus of this field trip, is a complex gneiss-migmatite terrane that extends from the Laramie Peak shear zone along the southern margin of the Laramie batholith to the northern and northeastern margin of the Laramie Anorthosite Complex. It is part of the central metamorphic belt of Condie (1969) and has sustained regional high-grade Barrovian metamorphism.

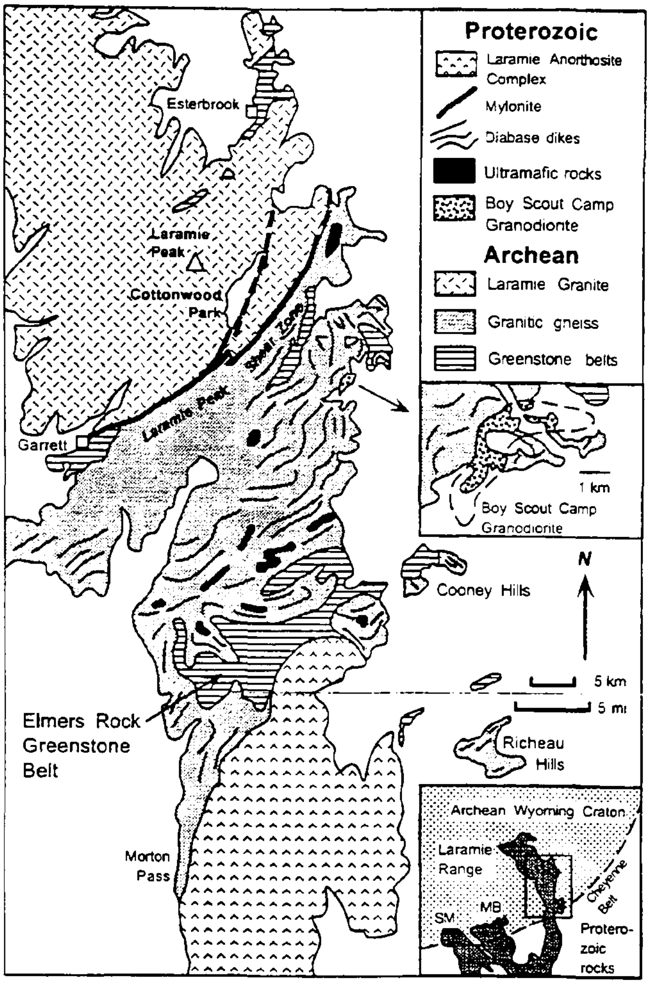

Figure 3. Sketch geologic map of the central Laramie Mountains with an inset map of the area containing the Boy Scout Camp Granodiorite. The lower right inset shows the figure region relative to the adjacent Sierra Madre (SM) and Medicine Bow Mountains (MB). (Generalized from Snyder, 1984, 1986).

The geology in this segment is considered in more detail below. (4) Proterozoic rocks of the Laramie Anorthosite Complex, Sherman Granite, and volcanogenic and pelitic gneiss south of the Cheyenne belt. As noted above, the projection of the Cheyenne belt in the Laramie Mountains is concealed by the Laramie Anorthosite Complex (**Fig. 1**).

The central Laramie Mountains can be further roughly subdivided into a northern, gneiss-dominated region and a southern supracrustal-rich region (**Fig. 3**). The following two sections describe the geology of these two regions of the central Laramie Mountains. The first section, on the northern, gneiss-dominated region, describes relationships between intrusive phases, deformation fabrics, and radiometric ages. The second section describes general features of the supracrustal sequence in the Elmers Rock greenstone belt and the complex metamorphic history recorded in the rocks.

Northern, Gneiss-Dominated Region - Evidence for Early Proterozoic Deformation

The northern part of the central Laramie Mountains contains primarily interlayered Archean granite and granitic gneiss intruded by deformed sills, dikes and lenses of mafic and ultramafic rocks. Granitic rocks include banded granitic gneiss, migmatitic gneiss, and variably foliated granite. The banded granitic gneiss (~2.7 Ga [Rb-Sr], Johnson and Hills, 1976) contains a biotite foliation, compositional layering, and extensive leucogranite veining that at least partially predated the intrusion of foliated Laramie Peak granite. Along the southern part of the gneiss-dominated region, the gneissic banding is cut by a recently dated Archean granite with a preliminary U-Pb age of 2.6 Ga (K. Chamberlain, unpublished data), indicating an Archean age for this gneissic banding.

The foliated granite of the central Laramie Mountains includes a number of individual bodies that are compositionally and texturally heterogeneous. Fabrics in the granites include variably developed foliation and lineation defined by biotite, hornblende, quartz, and/or feldspar. The majority of the foliated granite is assumed to be contemporaneous with the intrusion of the extensive ~2.6 Ga Laramie batholith (Hills and Armstrong, 1974; Condie, 1969); however, local smaller bodies of younger granite also exist. One of these bodies (Boy Scout Camp Granodiorite of Snyder, 1986) is a hornblende-biotite granodiorite with a U-Pb zircon age of 2,051 ± 9 Ma (Snyder et al., 1995). This intrusion occurs along the eastern margin of the range (**Fig. 3**) and will be examined on the morning of the second day of the trip (Day 2, Stop 2). We have recently recognized a second Early Proterozoic granitic unit in this area, informally referred to here as the Sheep Mountain granite, that has a tentative U-Pb zircon date of ~2.0 Ga (K. Chamberlain, unpublished data).

All of the Archean units, as well as the Boy Scout Camp Granodiorite (BSCG) and Sheep Mountain granite (SMG), are intruded by abundant mafic dikes (**Fig. 3**) (Snyder, 1984, 1986; Snyder et al., 1990). Holden and Snyder (1983) recognized two sets of dikes: an early set of high-alumina, plagioclase-phyric basaltic dikes cut by more abundant tholeiitic diabase dikes. Snyder et al. (1988) further recognized that the dikes were emplaced prior to at least one deformation and metamorphic episode because they locally contain metamorphic garnet, and some of the dikes are folded and contain locally strong deformational fabrics. The dikes are commonly conformable with the adjacent banding in the Archean granitic gneiss. However, dike contacts cutting the gneissic banding were observed in several locations.

Our recent work indicates that the mafic dikes and all older units, including the BSCG and SMG were deformed by at least two and locally three periods of deformation that must be younger than the 2.05 Ga granodiorite and mafic dikes. The deformed Sheep Mountain granite and mafic dikes are exposed within the Archean gneiss along the eastern margin of the range and will be examined on the morning of the second day of the trip (Day 2, Stop 3).

Ultramafic lenses and layers, as much as a kilometer wide, also intruded the granitic gneiss (**Fig. 3**). These bodies have mutually cross-cutting relationships with dated mafic dikes, and preliminary U/Pb geochronology suggests that they are contemporaneous (Cox et al., 1995; D. Cox, unpublished data). The ultramafic layers are locally folded about axial plane orientations similar to those of the folded mafic dikes. Both the mafic dikes and the ultramafic bodies contain local amphibolitized ductile shear zones in areas where their folding can be demonstrated. These shear zones do not extend into the adjacent gneiss and are apparently one of the ways in which the dikes and ultramafic bodies accommodated the strains associated with their folding.

Southern, Supracrustal-Rich Region and High-Grade Metamorphism

The Elmers Rock greenstone belt (Graff et al., 1982) and associated metasedimentary rocks (Hodge, 1966; Holden, 1978; Snyder, 1984) comprise the major supracrustal sequence in the southern part of the central Laramie Mountains. These rocks are bound on the southeast by Red Mountain Syenite of the Laramie Anorthosite Complex, and they are surrounded by Archean granitic gneiss to the north, south, and west (**Fig. 3**). A northeast-trending arm of the belt extends to the northeast of the main part of the belt forming a complex synformal pattern between domes of granitic gneiss to the northwest and southeast (**Fig. 3**). The rock sequence consists of layered mafic volcanic rocks and various interlayered metasedimentary rocks. A U-Pb zircon age of 2,637 ± 10 Ma was obtained from a rhyolite layer interlayered with the metasedimentary rocks (K.R. Ludwig, referenced in Snyder et al., 1989). Mafic dikes, similar to those in gneissic rocks discussed above, also intrude the greenstone belt, but they are much less common or more difficult to recognize (Snyder et al., 1990).

The volcanic rocks of the greenstone belt include tholeiitic and komatiitic flows (Holden and Snyder, 1983; Smaglik, 1987). The tholeiitic flows locally contain preserved amygdaloidal pillow lavas but primary igneous textures have typically been destroyed by metamorphic recrystallization and deformation. The metasedimentary rocks include graywacke, pelite, calc-silicate, marble, quartzite, iron-formation, and metaconglomerate (metamorphic suite of Bluegrass Creek of Snyder, 1984). Despite considerable deformation of the supracrustal rocks, remnant pillows are preserved in the basaltic flows, and graywacke layers locally contains cross beds and graded beds allowing determination of stratigraphic tops in some areas. Snyder et al. (1988) interpret the sequence to have been deposited in a shallow marine environment. All of the rocks have been metamorphosed under amphibolite-facies conditions.

Graywacke and interlayered pelite of the belt contain a well-developed regional Barrovian series of mineral assemblages (Hodge, 1966; Holden, 1978; Snyder, 1984). The peak metamorphic assemblage is muscovite + biotite + kyanite + staurolite + garnet ± sillimanite + rutile. Although evidence for this metamorphic series is best represented in the Elmers Rock greenstone belt, similar assemblages occur in local metasedimentary pods throughout the central Laramie Mountains from the Elmers Rock greenstone belt to the Laramie Peak shear zone (Patel, 1992).

The timing of high-grade metamorphism is constrained to the Early Proterozoic on the basis of U-Pb ages of metamorphic sphene (~1.76 Ga) collected from a number of diabase dikes from throughout the central Laramie Mountains (Cox et al., 1993; Cox, unpublished data; Resor et al., 1996). U-Pb sphene ages range from 1.76 to 1.75 Ga and record the time that the central Laramie Mountains cooled through 580-620 °C during uplift and exhumation. Peak metamorphism may have been slightly earlier as thermobarometry indicates peak temperatures of at least 650 °C with rapid decompression and cooling to 550-500 °C (Patel, 1992; Chamberlain and others, 1993). The Early Proterozoic timing of Barrovian metamorphism is also supported by Rb-Sr data from metapelitic rocks throughout the central Laramie Mountains (Patel, 1992), which have been interpreted to indicate isotopic rehomogenization at ~1.8 Ga (Patel et al., 1991).

The mineral assemblages and porphyroblasts associated with this event, nevertheless, display evidence of both a complex deformational and thermal history. Garnet and staurolite porphyroblasts commonly contain inclusion trails of an early foliation that was folded. In some units this early foliation has been completely transposed and is evident only from the inclusion trails. Kyanite and sillimanite are commonly aligned in well-developed planar and linear fabrics but are also locally folded.

Within the aureole of the Laramie Anorthosite Complex, the Barrovian assemblages have been thermally overprinted. The thermal overprinting produced high-temperature, low-pressure recrystallization and local partial melting near the contact aureole (Bochesky and Frost, 1982; Grant and Frost, 1986; Grant and Frost, 1990;

Patel, 1992). The typical Barrovian assemblages in the schist distal to the contact include: quartz, muscovite, aluminosilicate (kyanite or sillimanite), biotite, garnet, and staurolite. In rocks progressively closer to the contact with the Red Mountain Syenite of the Laramie Anorthosite Complex, kyanite is replaced by andalusite and then andalusite is replaced by prismatic sillimanite. Staurolite is pseudomorphically replaced by cordierite + aluminosilicate + spinel, and quartz + muscovite assemblages are replaced by K-feldspar + sillimanite. The highest grade assemblages in the contact aureole are K-feldspar + cordierite + garnet + biotite and K-feldspar + cordierite + orthopyroxene + biotite (Spacuzza, 1990).

Laramie Peak Shear Zone and Evidence for Regional Uplift

The Laramie Peak shear zone is a system of heterogeneously distributed shear strain, including localized mylonitic zones as much as 500 m-wide, that strikes northeastward across the north-central Laramie Mountains from the vicinity of Garrett to Harris Park (**Fig. 3**). Shear zone fabrics include a northeast-striking mylonitic foliation (**Fig.** 4) and a steep southwest-plunging mineral-elongation lineation (**Fig.** 5). Kinematic indicators within the shear zone (asymmetric porphyroclasts, composite foliations and asymmetric folds) all yield a consistent south-side-up sense of shear (see Stop 6, Day 2 figures). The crystal-plastic deformation of feldspars and the stability of hornblende and plagioclase within mafic rocks indicate that deformation occurred under amphibolite-facies conditions.

Evidence from both metamorphic and isotopic studies indicate that the central Laramie Mountains (i.e., south of the Laramie Peak shear zone) sustained regional uplift in response to Early Proterozoic deformation. Patel (1992) and Chamberlain et al. (1993) reported thermobarometric data from locations ranging from the Elmers Rock greenstone belt to the Laramie Peak shear zone that yield steep decompression-cooling trends. Garnet-aluminosilicate-silica-plagioclase (GASP) and garnet-rutile-aluminosilicate-ilmenite (GRAIL) barometers coupled with garnet rim-biotite thermometry indicate P-T ranges for single samples ranging from ~7 kb at 650 ° C to as low as 3.5 kb at 500° C. Textural relationships in the regional metamorphic assemblages are also consistent with decompression. In pelitic schist this includes local examples of garnet, kyanite, or staurolite with overgrowths of cordierite at various locations across the central Laramie Mountains. Decompression is also seen in gedrite-rich schist, wherein kyanite and gedrite never touch; cordierite has formed between them. In contrast, metamorphic assemblages from rocks north of the Laramie Peak shear zone display consistent andalusite-grade assemblages in

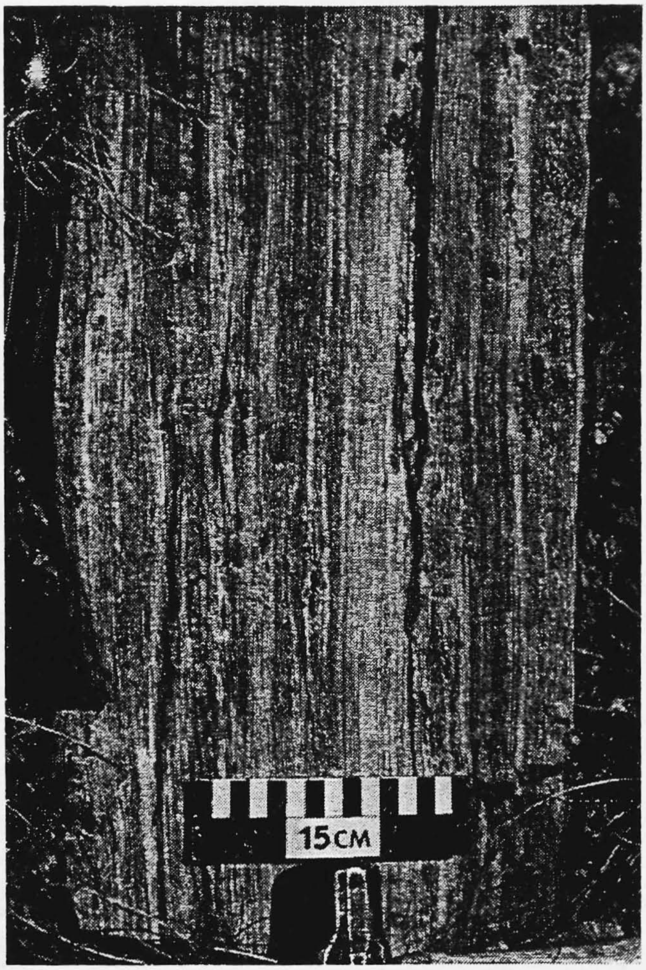

Figure 4. Photograph of "pin-striped" mylonitic foliation in granitic gneiss of the Laramie Peak shear zone.

pelitic assemblages such as quartz-muscovite-chloritoid-andalusite-garnet-plagioclase (Snyder and Bow, 1992). Peak pressures recorded by the rocks are less that 4 kb.

U-Pb apatite ages, interpreted as cooling ages, also show a distinct break across the Laramie Peak shear zone (Chamberlain et al., 1993). Samples from the Laramie batholith, north of the shear zone, yield ages of ~2.1 Ga, and have been interpreted as evidence that rocks of the batholith have not been hotter than about 450 °C since 2.1 Ga (Chamberlain et al., 1993). Chamberlain and coworkers attributed this age to Proterozoic uplift of these rocks during ~2.2-2.1 Ga rifting along the southern margin of the Wyoming craton. U-Pb apatite ages from the Archean gneiss south of the shear zones have consistent apatite closure ages of 1,736 ± 20 Ma (Chamberlain et al., 1993; Chamberlain, unpublished data). Chamberlain et al. (1993) suggested that the displacement on the shear zones occurred at about 1.8 Ga as a thick-skinned thrust in response to the Cheyenne belt collision. Subsequently,

Figure 5. Steep, southward-plunging hornblende lineation on foliation plane in amphibolitic dike rock developed during the shear-zone deformation (D_2). The site of the photograph was in the Fletcher Park mafic dike (see Stop 5, Day 2) but south of the Laramie Peak shear zone.

Resor et al. (1996) obtained a U-Pb age of 1,763 ± 7 Ma Ga for syndeformational sphene from a metamorphosed and deformed mafic dike in the Laramie Peak shear zone (Day 2, Stop 5). This age is consistent with the age range of 1,780 to 1,740 Ma suggested by Premo and Van Schmus for the Cheyenne belt collision. Biotite K-Ar dates on both sides of the shear zones show a similar range (1.56-1.25 north, 1.51-1.36 south) (Hills and Armstrong, 1974), indicating that both sides of the shear zone passed through the biotite closure temperature for Ar (~300-350°C) after displacement on the shear zones.

Early Proterozoic Folding and Regional Structural Trends

The rocks of the central Laramie Mountains have undergone at least two periods of Early Proterozoic deformation, and the Archean rocks also experienced older, Archean deformation. The Early Proterozoic deformation features occur in all of the Early Proterozoic intrusive igneous rock units as well as the Archean host rocks. The granitic gneiss contains a well-developed compositional banding that is cut by the Early Proterozoic intrusive rock units and is also cut by a recently dated Archean granite with a preliminary U-Pb age of 2.6 Ga (K. Chamberlain, unpublished data). We have not attempted to decipher the Archean deformation history of the area; however, it is primarily represented by gneissic banding and an early foliation that is locally preserved in the supracrustal rocks of the Elmers Rock greenstone belt. For the ease of discussion of younger events, we assign all Archean deformation to a D_1 event. The two Early Proterozoic events, D_2 and D_3 respectively, are the principal focus of our analysis.

The dominant structural trends across the region are a product of the Early Proterozoic deformation and are typified by the trends of the mafic dikes shown schematically in Figure 3. Along the southern margin of the gneiss-dominated region, just north of the Elmers Rock greenstone belt, the dikes and compositional banding in granitic gneiss primarily strike to the east or northeast. However, to the north, along the eastern margin of the range, the structural trends change progressively to the north. This pattern appears to be a product of the D_2 deformation in both regions. A well-developed D_2, L to LS fabric occurs in both the Archean and Early Proterozoic rock units and shows this same spatial variation in trend as the dike pattern. Along the eastern margin of the range, the lineations generally plunge moderately to the south and are coaxial with the hingelines of large-scale, z-symmetry, F_2 folds of the Archean gneissic banding. Along the southern margin of the gneiss-dominated region and in the Elmers Rock greenstone belt, the L_2 lineations are parallel to the hingelines of F_2 folds that have eastward-striking axial surfaces that dip variably to the north. The orientation of the F_2 fold hinges and coaxial L_2 lineations vary from westward plunging in the western part of the greenstone belt to northward plunging in reclined folds in the central part of the greenstone belt. The locus of change in the orientation of the F_2 fold axial planes from easterly to northerly is in the northeastern part of the greenstone belt where supracrustal rocks are complexly infolded with granitic gneiss (**Figs.** 3 and 4).

The D_3 deformation produced large-scale open folds that locally refold D_2 fabrics and F_2 folds about upright, northwest-striking axial surfaces. F_3 folds deform L_2

lineations and other D_2 features in all of the rock units. A D_3 crenulation foliation occurs locally in response to tight folding of L_2 lineations, but no penetrative D_3 fabrics have been recognized.

SUMMARY

Early Proterozoic (~2.0 Ga) granitic plutons, mafic dikes, and peridotite intruded Archean granitic gneiss in the central Laramie Mountains during a phase of continental rifting along the southern and eastern margins of the Wyoming Archean craton. These intrusive rocks cut an Archean gneissic banding (D_1), but all of the Early Proterozoic units contain well-developed ductile deformation features that formed during at least two periods of Early Proterozoic deformation (D_2 and D_3). Deformation features associated with these events include large-scale folds and rock fabrics that presumably formed during east-west shortening during the Trans-Hudson orogeny and north to northwest shortening during formation of the Cheyenne belt suture zone.

The northern boundary of this region of ductile reworking is marked by the Laramie Peak shear zone. South-side-up, sinistral ductile shear along this zone occurred ~1, 760 Ma in response to collisions to the south along the Cheyenne belt.

The uplift of the central Laramie Mountains in response to the Cheyenne belt collision is reasonably well constrained (Chamberlain et al., 1993; Resor et al, 1996). However, the relationship of the Early Proterozoic ductile reworking of the central Laramie Mountains to the Trans-Hudson orogeny and the Cheyenne belt collision are not as clear (Bauer et al., 1995, 1996). The purpose of our on-going studies is to further define the geometry, age, and tectonothermal history of the Early Proterozoic deformation and metamorphism associated with these orogenies, with the ultimate goal of determining the relationship between these regional orogenic episodes that intersected along this southeastern margin of the Wyoming craton. The field stops described below contain some of the features that are the basis of our current interpretations of this region of Early Proterozoic reworking.

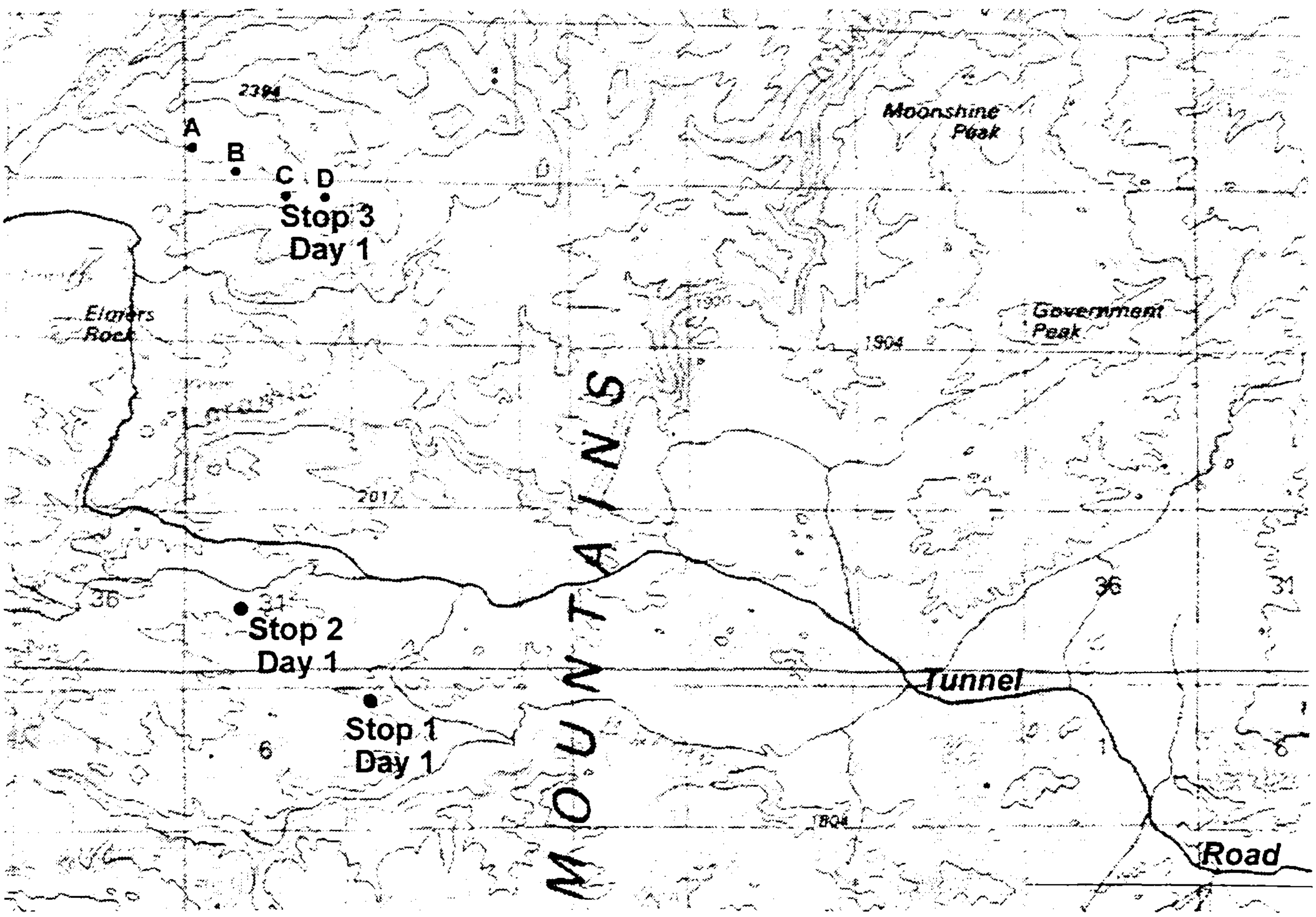

Figure 6. A portion of the U.S. Geological Survey, Rock River, 30x60 minute topographic map showing the location of field trip stops for day 1 (scale: 1:70,000). See Road Travel descriptions in the text for Townships and Ranges.

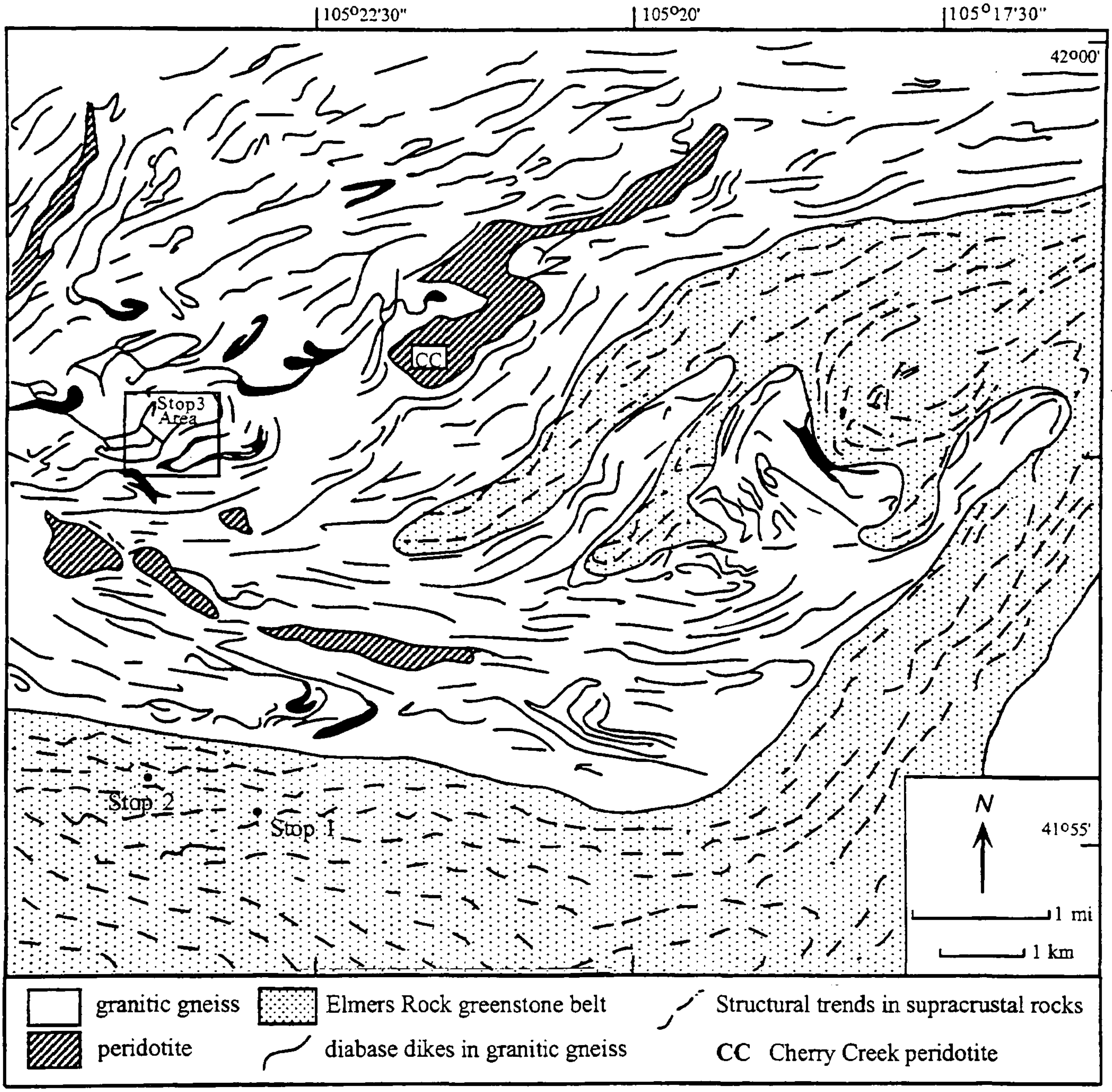

Figure 7. Geologic sketch map showing the location of Day 1 field stops in the northeasternElmers Rock greenstone belt and adjacent areas of deformaed granitic gneiss, mafic dikes and peridotite (simplified from Snyder, 1984).

FIELD TRIP STOPS

All of the stops on this field trip are on private ranch property and are not directly accessible from public roads. As part of our arrangements with the private landowners, we have not provided detailed road logs on private roads. We have, however, provided a road log for the public access to the private property and include a section from a USGS 1:100,000 topographic map (**Fig. 6**) with the locations marked for each of the Day 1 stops. Some of the stops involve hiking from one location to another within the stop area. In these cases, we have provided location descriptions and township, range, section, subsection locations.

Day 1 - Elmers Rock Greenstone Belt and Adjacent Granitic Gneiss

Itinerary

During the first day of the trip we will examine three areas that illustrate the nature of Early Proterozoic deformation along the southern part of the central Laramie Mountains. The first two stops examine deformed supracrustal rocks in the Elmers Rock greenstone belt. These rocks contain well-developed F_2 and F_3 folds and locally intense ductile shear that predates F_2 folding. The third stop examines deformed Archean granitic gneiss, granite and schist, and Early Proterozoic mafic dikes in the gneiss-dominated region north of the Elmers Rock greenstone belt. All of the rock units in this area have been deformed by both the F_2 and F_3 fold events. Each stop will involve short hikes to parts of the stop area to examine specific features included in the stop descriptions.

Road travel to Stop 1

From Wheatland, Wyoming, travel approximately 5.9 miles south on Interstate 25 to the Wyoming Highway 34 exit. Turn right and continue to the west on Highway 34. Travel west and southwest approximately 15.6 miles to Tunnel Road (paved road makes a T-intersection from the right with Highway 34). Turn right (west) on Tunnel Road and proceed approximately 9.1 miles to the entrance to the Bluegrass Creek Ranch and Cattle Company. Turning left enters private ranch property. The outcrop area of interest is in the NW1/4 of section 5, T.22N., R.71W. (**Figs.** 6)

Stop 1 - F_2 folding in marble of the Elmers Rock greenstone belt

The Elmers Rock greenstone belt (**Figs.** 3 and 7) contains an extensive sequence of supracrustal rocks that Snyder (1984) named the metamorphic suite of Bluegrass Creek. The most extensive members of this suite are metabasalt (amphibolite), metagraywacke, and pelitic schist. However, marble and calc-silicate rocks occur locally within the sequence, and this stop contains the most extensive marble exposure in the greenstone belt. Snyder (1984) suggested the marble is possibly the youngest stratigraphic unit in the metamorphic suite, and he noted that its contact with many different rock types in the belt may indicate an unconformity at its base.

The marble at this stop contains well-bedded, white to gray units that are interbedded with abundant, more siliceous, buff-colored beds. The metamorphic mineral assemblages includes: calcite, diopside, tremolite, and quartz with minor plagioclase and sphene. The more siliceous buff-colored beds are more resistant than the white marble and stand out in relief on the weathered surfaces. During deformation, these beds were also more competent than the adjacent calcite-rich marble; as a result, they contain many of the well-developed F_2 folds that pervade this unit (**Fig.** 8).

Figure 8. F2, s-symmetry folding in banded marble at Stop 1 (pencil parallel to fold axial trace for scale).

This outcrop area lies near the axial trace of a major, reclined, F_2 fold. Both regionally and in this outcrop area, the fold closes to the west and plunges moderately to the north about a westerly striking axial plane that dips to the north. This fold geometry is recognizable in the field by looking to the north across the small amphitheater valley that contains the eastern contact of the marble with amphibolite. The marble-amphibolite contact crops out along the north side of the valley in the core of a large-scale F_2 fold plunging to the north. The contact is marked by local areas of probable clastic metasedimentary rock and amphibolite. Sparse fragments of amphibolite in the marble near the contact suggest an unconformable contact with structural facing to the west, toward the marble. From this contact, looking west, well-developed F_2 folds are visible in the marble. Most of the smaller scale F_2 folds have s- to m-symmetry (e.g., **Fig.** 8), consistent with their position on the larger fold structure. The F_2 fold hinges, which plunge to the north, are parallel to a well-developed L_2 linear fabric that is developed locally in the amphibolite. The L_2 extension is also evident in the folded buff-colored marble beds where small, closely spaced extension fractures, normal to the F_2 fold hinges, are filled with more resistant quartz.

The compositional layering in the marble, with alternating white carbonate-rich layers and buff-colored silica-rich layers, is readily interpreted as a primary bedding feature. However, many of the other supracrustal units in the greenstone belt have an earlier foliation that is folded by F_2 folds. In the sedimentary rocks where primary bedding is still preserved, this early foliation is typically parallel to bedding. Such an early bedding-parallel foliation is not apparent in the marble. However, local areas that contain small-scale, type-1 fold interference patterns occur locally in this outcrop (**Fig.** 9) and may be a result of interference between F_2 and "F_1" folds.

Figure 9. Deformed marble containing type-1 interference patterns between F_2 and probable "F_1" folds (knife scale in all photos is 6 cm long).

The marble exposures continue along a ridge to the northeast of the amphitheater valley toward a small area of quarried marble. Highly folded marble along this ridge contains F_2 folds as well as small-scale F_3 folds with axial traces that trend to the north or northwest (**Fig.** 10).

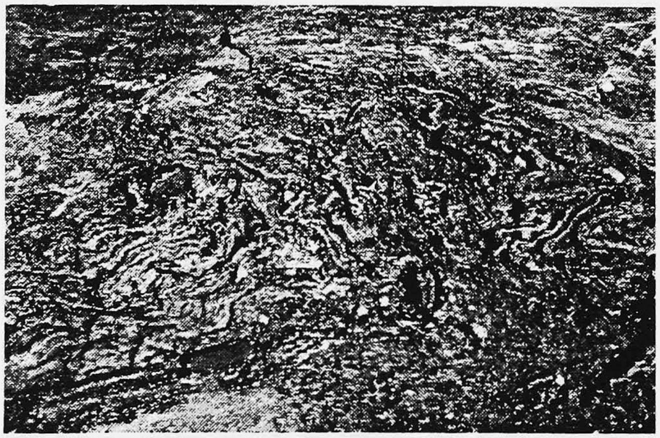

Figure 10. Deformed marble containing F_2 folds on the right side of the photo, closing to the east, and more open F_3 folds above the pocket knife with axial traces trending to the northwest.

Road travel to Stop 2

Return to Tunnel Road and turn left (west). Approximately 0.6 miles to the west, a dirt road to the left continues to the west where Tunnel Road curves to the northwest (**Fig.** 5). A locked gate marks the intersection of the road with the section 31-32 boundary. Stop 2 is in the NE1/4 of the SW1/4 of section 31, T.23N., R.71W., approximately 0.9 miles west of the road intersection, and over the ridge on the south side of the dirt road.

Stop 2 - Sheared and folded amphibolite and metaconglomerate

Banded amphibolite that contains local evidence of intense shearing crops out on the ridge marking the northern part of this stop. Over the ridge, to the south, the amphibolite is in contact with a unit that has been mapped and described by all previous workers as a metaconglomerate (Hodge, 1966; Graff et al., 1982; Snyder, 1984). The metaconglomerate contains matrix-supported, granitic to pegmatitic clasts and layers, ranging from pebble to boulder size, in a biotite schist matrix.

Both the amphibolite and the metaconglomerate show evidence of deformation by locally intensive shear. The conglomerate clasts commonly display asymmetric shapes that indicate sinistral, north-side-up shear (**Figs.** 11 and 12). Granitic layers in the amphibolite are locally folded into zones of high strain that also indicate a significant north-side-up shear component (**Fig.** 13). Despite the evidence for shearing in these rocks, they apparently recrystallized subsequent to the shearing and are not mylonitic.

Figure 11. Asymmetric clasts, indicative of sinistral shear, in the metaconglomerate.

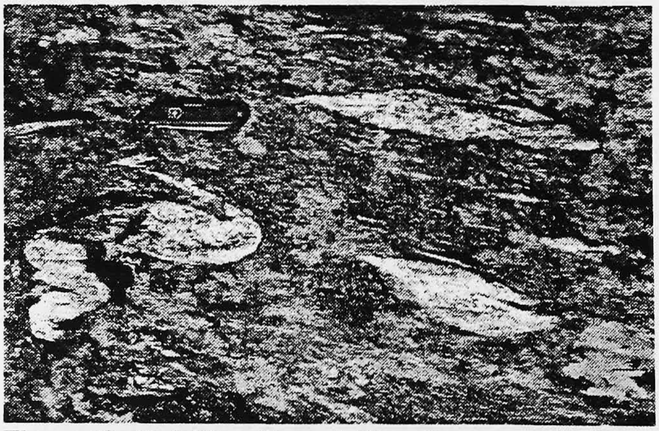

Figure 12. F_2 folding (lower left) and asymmetric clasts in metaconglomerate.

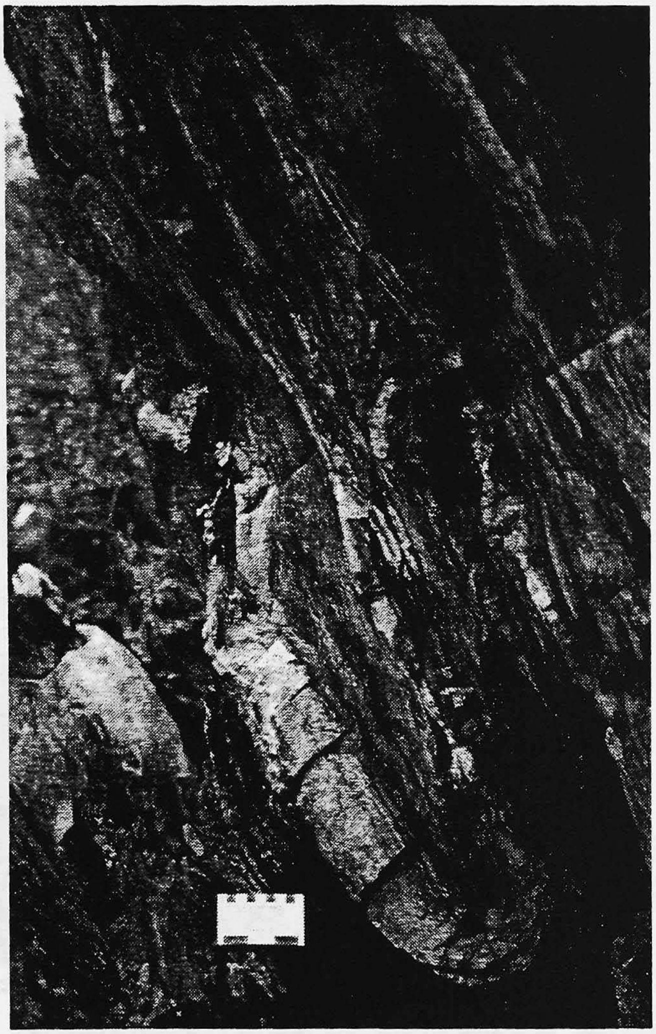

Figure 13. Folded granitic vein in sheared amphibolite contains a strongly attenuated northern limb indicating north-side-up shear.

The shearing and development of clast asymmetry predate the F_2 folding in the area. This is evident along a ridge containing a metaconglomerate that extends to the south. This ridge crosses the axial trace of a large F_2 fold and contains numerous flattened and asymmetric clasts and thin granitic layers that are folded by F_2 folds (**Fig.** 14) In the hinge area of the large F_2 fold, the clasts are also stretched in a strong L_2 fabric that parallels local F_2 fold hinges and plunges steeply to moderately to the north-northeast. The axial planes of local F_2 folds strike westerly and dip steeply to moderately to the north.

Small-scale F_3 folding is locally developed in the amphibolite. These folds vary from open warps of the foliation to folds and crenulations with a typical z-symmetry. These folds plunge northerly, similar to the F_2 folds, but typically have north- to northwest-striking axial planes.

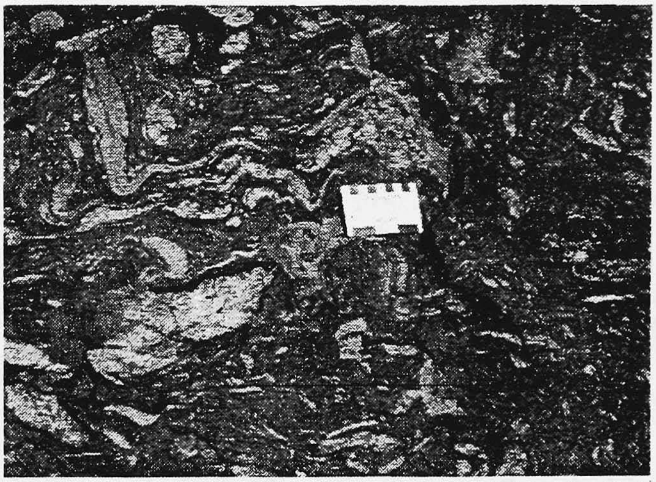

Figure 14. Thin granitic layers and clasts folded by F_2 folding in the metaconglomerate.

The origin of the metaconglomerate is worth further consideration. Although this unit may be a clastic unit, as described above, it could be a product of intense shear deformation of a veined migmatite. With a clast composition generally restricted to granite, granitic gneiss, granite pegmatoid, - and with no volcanic clasts, it is certainly not a lahar, such as might be expected to occur locally in a greenstone terrane. Its matrix-supported character (clasts surrounded by biotite schist) also argues against a channel deposit. There are remnant granitic veins and layers in the unit that show evidence of disruption by the locally concentrated shear (**Fig.** 15)and such disrupted units could have provided fragments that appear to be granitic clasts.

Figure 15. Incipient development of an asymmetric pull-apart in a granitic vein deformed in the sheared conglomerate.

Road travel to Stop 3

Retrace route back to Tunnel Road and turn left (west). Proceeding west, the road crosses the Laramie River in approximately 1.35 miles and curves progressively to the north and crosses the contact between supracrustal rocks of the Elmers Rock greenstone belt and the granitic-gneiss-dominated terrane to the north of the greenstone belt. The road continues up Gibbs Canyon (local name, not printed on maps) with granite outcrops that contain Early Proterozoic mafic dikes on either side of the road. As the road emerges from the tight part of the canyon, approximately 2.75 miles from the road intersection from Stop 2, the prominent peak of granite on the left (west) side of the road is Elmers Rock. Continuing north, the road curves to the left and enters a flat-lying stretch of road at approximately 3.5 miles. A dirt road to the right is the road to Stop 3. Parts of this road are only passable in a 4-wheel-drive vehicle. The peak approximately 1.3 miles to the northeast from this point is Sugar Loaf (elevation: 7,387 ft, 2,394 m). The outcrops to be examined as Stop 3 are in the SW1/4 of section 18, T.23N., R.71W. and the northern half of section 19, T.23N., R.71W (**Figs.** 6 and 7).

Stop 3 - Granitic gneiss, granite, kyanite/sillimanite schist, and mafic dikes deformed by F_2 and F_3 folding. Local metamorphic evidence for regional uplift

This stop includes a series of outcrops with a wide variety of rock types that contain deformation features formed during both D_2 and D_3. The oldest rocks in the area are Archean banded gneiss. The banding in the gneiss is locally cut by a more massive pink granite with a preliminary U-Pb zircon date of 2.6 Ga (K. Chamberlain, unpublished data). This is probably the same granite that crops out along the road up Gibbs Canyon (see road travel description above), for which Snyder et al. (1988) report an unpublished U-Pb zircon date of 2.6 Ga (K. Ludwig, unpublished data). The granitic gneiss contains local, small, isolated pods of coarse-grained pelitic schist containing sillimanite or kyanite. Early Proterozoic mafic dikes that are variably recrystallized to amphibolite and garnet amphibolite cut the gneiss. Some outcrops contain a net-veined, plagioclase-phyric, garnet-bearing amphibolite, with irregular veins of granite.

The mapped distribution of mafic dikes in this area (Snyder, 1984), illustrated in the boxed area of Figure 7, suggests probable multiple folding of the dikes by both F_2 and F_3 folds. One of the principle objectives of this stop is to illustrate the features associated with this deformation in the mafic dikes and other rock units of the area.

Four outcrop areas are examined at this stop (**Fig.** 6):

Area A. Outcrops on the north side of the South Fork of Cherry Creek, in section 18, T.23N., R.71W., just east of the section 18 - section 13 boundary.

One of the characteristics of the deformation in the Early Proterozoic mafic dikes is a strong local partitioning of the strain and associated fabrics that develop in the

Figure 16. Strong linear fabric, parallel pencil, in mafic dike (amphibolite) passes into relatively undeformed, pyroxene-bearing diabase in the upper part of the photograph.

dikes. In this area, the mafic dikes show extreme variation in recrystallization and fabric development. A mafic dike along the base of this outcrop area shows very little recrystallization. However, a mafic dike exposed near the top of the outcrop, to the west, shows a strong linear fabric in amphibolite along its margin. The foliation, nevertheless, dies out within two feet of the dike margin, yielding to diabase with preserved clinopyroxene (**Fig.** 16). Such strain partitioning is not restricted to dike margins. Deformation fabrics and recrystallization in the mafic dikes show a complex distribution in response to folding. This selective recrystallization and strain partitioning have resulted in areas of well-preserved mafic dike that are amenable to radiometric dating.

In the field, these pods of preserved magmatic assemblages can be recognized by the orange-brown weathering of pyroxene and olivine, a slight brownish-green tint to the plagioclase, and the rounded bouldery appearance of the outcrops. In contrast, zones of recrystallized diabase that are dominated by metamorphic hornblende and plagioclase tend to be less weathered and dark black in color, with pure white plagioclase. Magmatic zircon separated from one of these pyroxene-rich magmatic pods yielded a U-Pb age of 2010 ± 10 Ma (Cox et al., 1995). Baddeleyite, which only occurs in the olivine-rich dikes, was rarely preserved, even in demonstrably magmatic pods. In general, the baddeleyite has been partially to completely recrystallized to polycrystalline zircon. U-Pb data from the baddeleyite - zircon composite grains from a second dike define a mixing line from ~2.0 to 1.75 Ga (Cox et al., 1993; Cox, unpublished data), consistent with evidence for crystallization ~2.0 Ga from the magmatic zircon.

The gneiss at this location contains a small pod of coarse-grained muscovite-biotite schist with clots of sillimanite and feldspar. Both the gneiss and the schist are deformed locally by F_2 folds with steeply plunging hingelines that are parallel to a steep L_2 lineation. The fold axial planes are typically steeply dipping and strike to the west-northwest.

Area B. Outcrop ridge on the south side of the South Fork of Cherry Creek in the SE1/4, SW1/4 of section 18, T.23N., R.71W.

Figure 17. Relatively massive pink granite, cutting banding in the granitic gneiss, is folded into an open fold. The photographed exposure is approximately 2 meters wide.

This ridge contains banded granitic gneiss cut by the pink massive granite (**Fig.** 17). Both the gneiss and granite are folded by F_2 folds that plunge steeply about steeply dipping, westward-striking axial planes. A broad, open F_3 warp has locally reoriented one of the limbs of a large F_2 fold in the gneiss (**Fig.** 18)

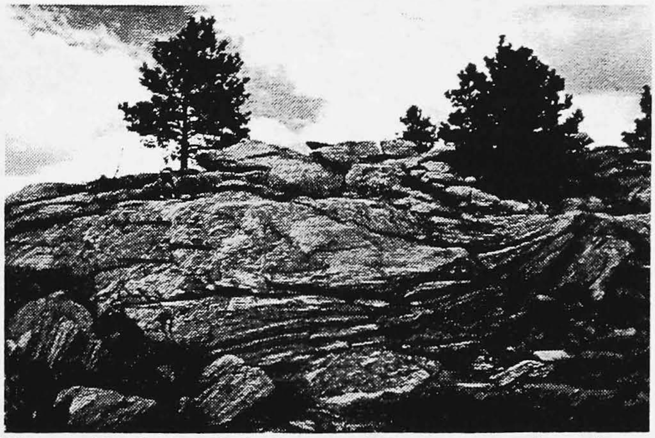

Figure 18. Open F3 fold (especially attenuated on the right side of the photo) folding the southern limb of an F2 fold dipping toward the viewer. The view is to the north. (Note person near left tree for scale.)

Area C. Outcrop area in the NW1/4, NE1/4, of section 19, T.23N., R.71.W.

This area contains part of the multiply folded mafic dikes mapped by Snyder (1984) and shown in the boxed area in Figure 4. This dike, as well as adjacent mafic dikes, contain evidence of F_2 folding and a locally strong L_2 lineation. F_3 folding produced the open warps in the dikes shown in Figure 4 about northwest-trending axial traces. This outcrop area contains an inferred F_2 fold closure shown in the southeastern part of the boxed area in Figure 4. This closure cannot be demonstrated in the field, but dike outcrops near the inferred closure do contain a very strong L_2 fabric. L_2 lineations in this area are locally reoriented by F_3 folding.

Mutually cross-cutting relationships and overlapping preliminary U-Pb zircon ages of the tholeiitic diabase dikes, plagioclase-phyric mafic dikes and metaperidotite bodies, suggest that all of the mafic intrusions were contemporaneous (Cox et al., 1995). Major, minor and trace element compositions can be interpreted to suggest that all of the mafic intrusions are genetically linked (Cox et al., 1995), although multiple sources have also been proposed (e.g., Hall et al., 1987)

Both tholeiitic diabase and plagioclase-phyric mafic dikes occur in this area. Unlike the tholeiitic dikes, the plagioclase-phyric dikes are locally netveined by white granite to pegmatite veins.

A small pod of coarse-grained biotite-kyanite schist in the gneiss contains a foliation that is folded into numerous small-scale folds and crenulations (**Fig.** 19). The schist apparently contained kyanite grains as much as 2-3 inches long (**Fig.** 20). Although large kyanite grains are still present in the schist, much of the kyanite has been replaced by intergrowths of quartz, plagioclase and cordierite. The cordierite commonly occurs directly replacing the kyanite

Figure 19. Crenulated kyanite schist

Figure 20. Coarse-grained kyanite schist. Most of the kyanite has recrystallized to quartz, plagioclase and cordierite.

and is interpreted to be a response to decompression associated with uplift of the central Laramie Mountains during collision along the Cheyenne belt (Chamberlain et al., 1993). Much of the schist in this area does not contain cordierite but contains the assemblage garnet + staurolite + kyanite + sillimanite + rutile. This assemblage records pressures of 7000 bars and temperatures around 650 ° C (Patel, 1992). However, even in rocks that do not contain cordierite, decompression to around 4000 bars is recorded in thin, Ca-poor rims of the garnet (Patel, 1993). Evidence for decompression is also seen in the mafic dikes, where plagioclase rims separate garnet porphyroblasts from surrounding hornblende. This texture is interpreted to have formed as the grossular component of the garnet was destabilized during decompression (cf., Kohn and Spear, 1990).

Area D. Outcrop area in the NE1/4, NE1/4, of section 19, T.23N., R.71.W. (optional)

The ridge northwest of the saddle at this location contains folded mafic dikes and granitic gneiss along the axial trace of a large-scale F_3 fold. The F_3 axial trace trends to the northwest and broad open warps of the gneissic banding with steeply plunging hingelines occur in the granitic gneiss on the inner arc of the fold. Smaller scale F_3 folds do not occur in the dikes or the granitic gneiss at this location, and no new fabric developed during the F_3 folding.

Return to Wheatland via Tunnel Road and Highway 34.

Day 2 - Deformation Along the Eastern Margin of the Central Laramie Mountains and the Laramie Peak Shear Zone.

Itinerary

During the morning of second day of the trip, we will examine three areas that contain varying degrees of Early Proterozoic deformation along the eastern margin of the central Laramie Mountains. The first stop contains granitic gneiss and an Early Proterozoic mafic dike that show relatively little evidence of Early Proterozoic deformation. The second and third stops examine areas with well-developed Early Proterozoic deformation features that affect two Early Proterozoic granitic units and mafic dikes as well as the host Archean gneiss.

Stops during the afternoon include the Laramie Peak shear zone, a mafic dike deformed in the shear zone that constrains the age of ductile shearing, a Laramide fault cutting the basement rocks, and a stop illustrating the relationship of an Early Proterozoic mafic dike to the Archean gneiss away from the reworked eastern margin of the central Laramie Mountains.

Road travel to Stop 1

Leave Wheatland heading west on county road 310 (Oak Street). The road crosses over I-25 on an overpass without entrances or exits. From the I-25 - Highway 310 intersection, travel 4.0 miles west to the intersection with N. Hightower Road (Highway 311). (The road names here are confusing. The stretch of Highway 310 from I-25 west changes from Oak St. to a part of Highway 310 that is called Hightower Road. At the intersection of Highway 310 and Highway 311, the stretch of Highway 311 to the north is N. Hightower Road, the stretch of Highway 311 to the south is Hightower Road). Turn right (north) on Highway 311 (N. Hightower Road). The road curves sharply to the left (west) 3.0 miles north of this intersection and takes another sharp curve to the right (north) in 1.3 miles. At this point, N. Hightower Road becomes the Fletcher Park Road. Continue to the northwest on the Fletcher Park Road for approximately 7.9 miles. The road changes from paved asphalt to gravel at this point and intersects the Van Ortwick Hill Road from the left in a T intersection. Turn left (west) on the Van Ortwick Hill Road (gravel). Continue 3.6 miles west. At this point the road jogs to the left onto private property. Continue on the fork of the road to the west, but the road may be blocked by locked gates beyond this point. Stop 1 is located beyond locked gates on the Van Ortwick Hill Road in the NW1/4, NE1/4, of section 18, T.25N., R.70W.

Stop 1. Banded Archean granitic gneiss and D_1 folds cut by undeformed Early Proterozoic mafic dikes

This stop is west of a region of Early Proterozoic ductile reworking that occurs along the eastern margin of the central Laramie Mountains. The banding in the Archean gneiss at this location is folded by recumbent folds. Steep mafic dikes cut across the banding in the gneiss and are unaffected by the recumbent folding. The gneiss and the dikes lack the linear fabric that characterizes the D_2 deformation that deformed the rocks to the east. The dikes are commonly recrystallized, but the only fabric they contain is a weakly developed foliation that occurs along the margins of the dikes near their contact with the gneiss.

Road travel to Stop 2

Return to the public part of the Van Ortwick Hill Road. Approximately 0.6 miles east of this point, a dirt ranch road intersects the Van Ortwick Hill Road in a T intersection from the north. This road leads to Stop 2, but it is on private property and contains locked gates. Stop 2 is located beyond locked gates in the vicinity of an abandoned Boy Scout Camp on the south bank of the North Laramie River in the SW1/4, SW1/4, Sec 3, T.25N., R.70W.

Stop 2. F_2 and F_3 folding in Early Proterozoic granodiorite and mafic dikes, and Archean gneiss.

Outcrops at this stop contain Archean granitic gneiss intruded by the Boy Scout Camp Granodiorite (BSCG) of Snyder (1986) and Snyder et al. (1995). Both units are intruded by mafic dikes. Steep cliffs on the north side of the North Laramie River, near a bend in the river, northeast of the abandoned cabin on this site, show the BSCG cutting the banding in the Archean granitic gneiss. Snyder et al. (1995) reported a U-Pb zircon date of 2051 ± 9 Ma for the BSCG.

Both F_2 and F_3 folding are well developed at Stops 2 and 3. The regional geometry of this deformation is illustrated in Figures 21 and 22. Figure 21 shows the

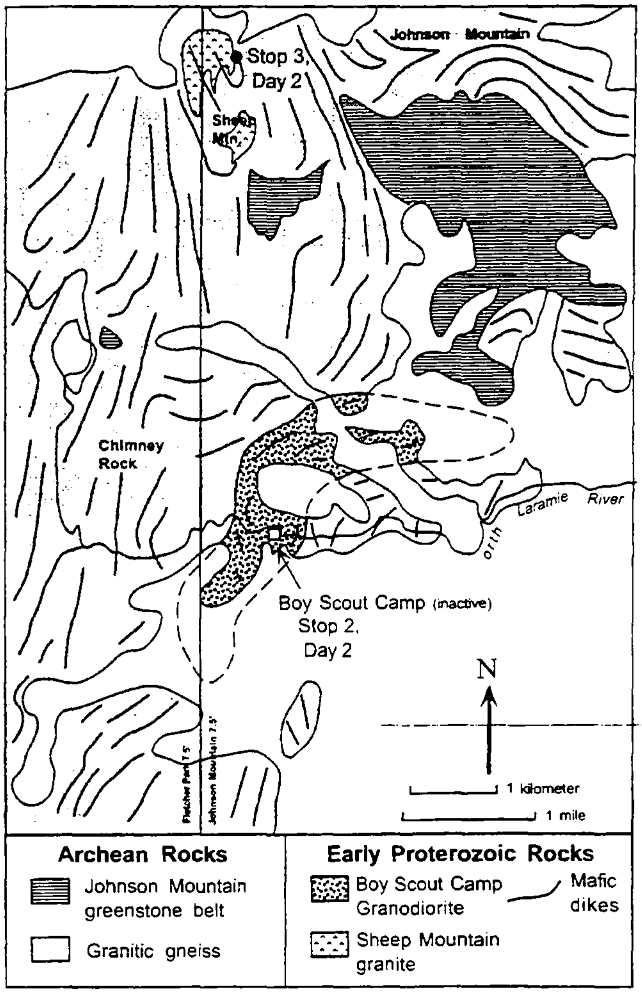

Figure 21. Geologic sketch map of the Boy Scout Camp - Sheep Mountain area showing the location of Stop 2 and Stop 3 for Day 2 of the field trip (Simplified from Snyder, 1986).

distribution of rock units in the region (modified from Snyder, 1986) and a schematic illustration of the distribution of mafic dikes across the region (simplified from Snyder, 1986). Figure 22 shows the distribution of F_2 and F_3 fold axial traces in the area based on work of Gresham (1994). The granodiorite body has a curved lenticular shape (**Figs.** 3 and 21) that mimics the shape of F_3 folds that deformed the gneiss and mafic dikes.

The BSCG at this stop is typical of this hornblende-biotite granodiorite unit. It contains a very strong L_2 linear fabric (**Fig.** 23) that is also well represented in the adjacent gneiss and the mafic dikes. The lineations throughout the entire pluton have a dominant southerly plunge. However, F_3 folding that produced the arcuate shape of the pluton has variably reoriented the lineations on the opposing limbs of the fold. Lineations on the southwestern limb of the fold tend to plunge moderately to the south; whereas those

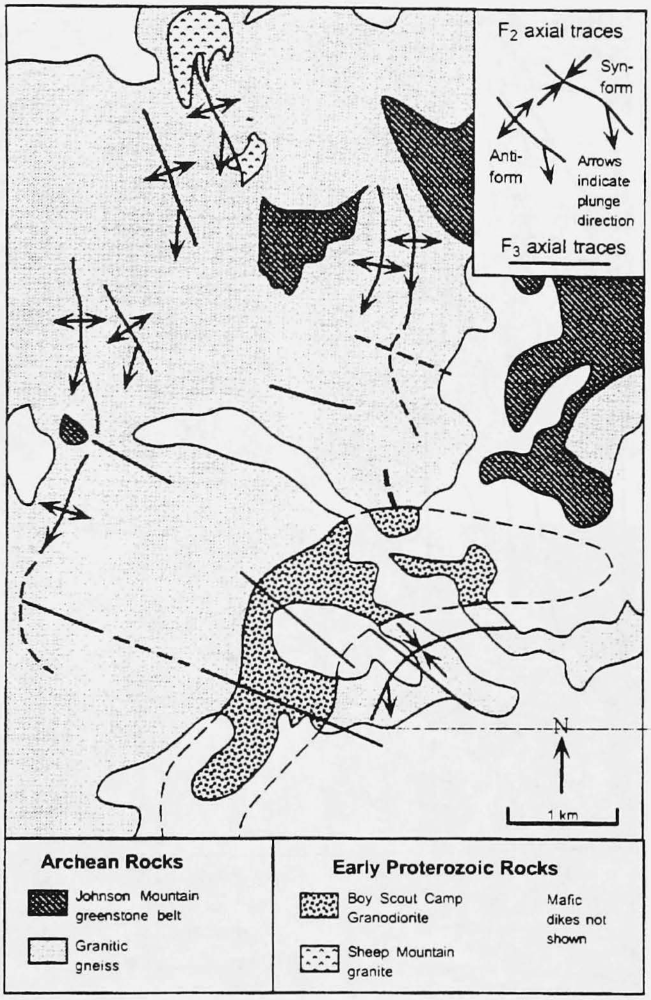

Figure 22. Geologic sketch map of the Boy Scout Camp - sheep Mountain area showing the distribution of F2 and F3 fold axial traces (after Gresham, 1994).

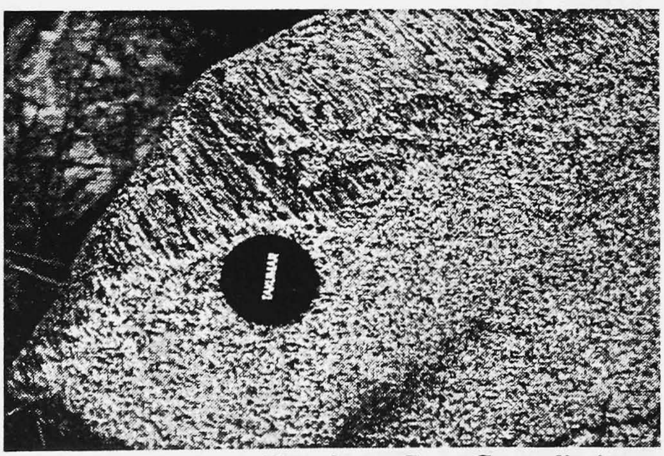

Figure 23. Outcrop of Boy Scout Camp Granodiorite showing the strong linear fabric that is typical in this unit.

on the northeast limb tend to plunge more steeply to the southwest (**Fig.** 24). Similar variations in the orientation of L_2 lineations occur in adjacent areas of the gneiss that have undergone F_3 folding. Reorientation of L_2 lineations in response to small-scale F_3 folding occurs locally within the pluton and such an example occurs at this stop (**Fig.** 25).

A mafic dike that cuts both the BSCG and gneiss at this location is variably recrystallized and deformed. It contains local areas with a strong L_2 lineation and other areas with preserved clinopyroxene.

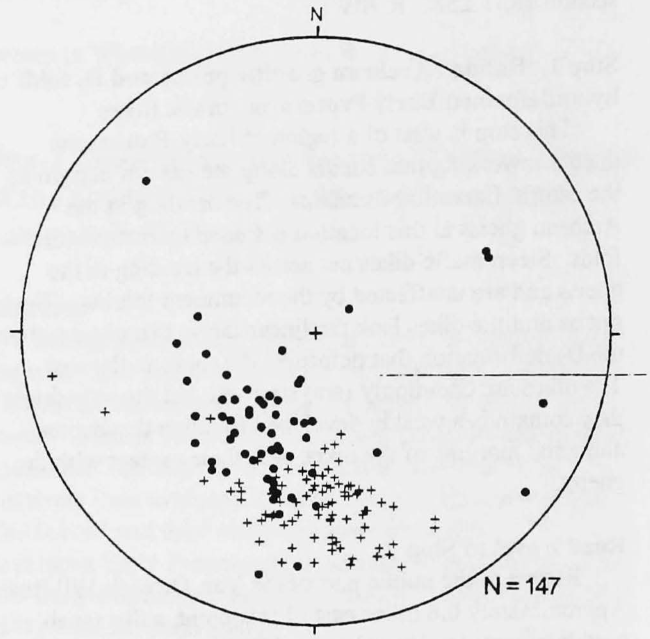

Figure 24. Stereographic projection of lineations measured in the Boy Scout Camp granodiorite. The filled circles are data collected on the northern limb of the pluton and the + symbols are data collected on the southwestern limb of the pluton.

Road travel to Stop 3

Retrace route back to the Van Ortwick Hill Road. Turn left (east) on the Van Ortwick Hill road and return to the intersection with the Fletcher Park Road. Turn left (northwest) onto the Fletcher Park road and travel approximately 7.4 miles to the northwest. The peak of Sheep Mountain is approximately 1 mile south of the road at this point. The northeastern side of Sheep Mountain was burned in a fire over a ten-day period in August of 1979. Stop 3 is on private property along the lower slope on the east side of the ridge that extends to the north from the east side of Sheep Mountain. The site is located in the NE1/4, NE1/4, of section 21, T.26N., R.70W.

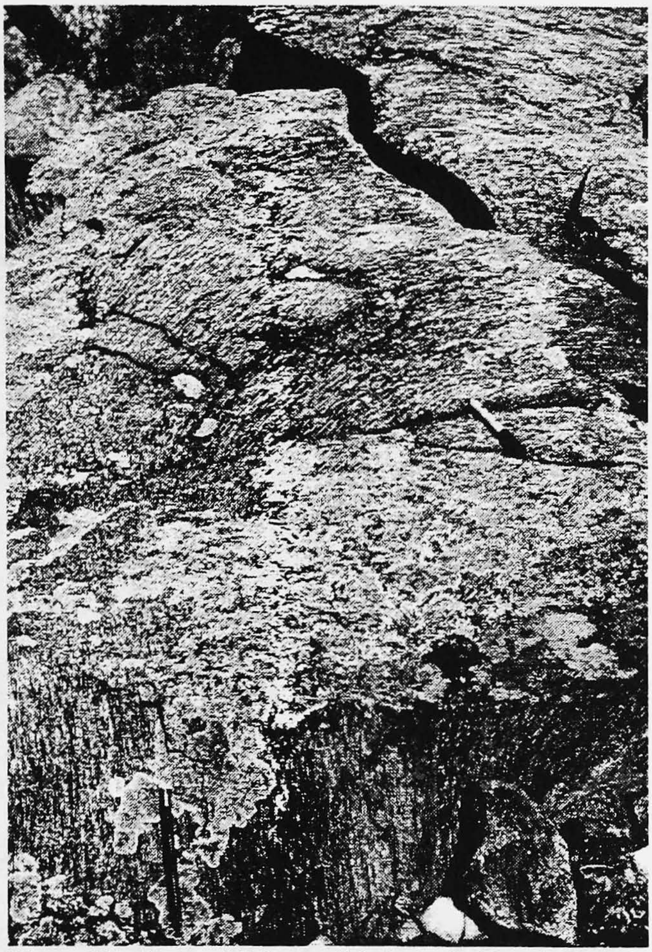

Figure 25. Strong L2 linear fabric in the Boy Scout Camp Granodiroite (parallel to pencil) is folded by a open F_3 fold with an axial trace parallel to the marker on the top of the outcrop.

Stop 3 - D_2 and D_3 Deformation in Archean granitic gneiss intruded by Early Proterozoic granite and mafic dikes

This stop contains a complexly deformed combination of both Archean and Early Proterozoic units, including a recently identified Early Proterozoic granite, referred to here informally as the Sheep Mountain granite (SMG). The SMG occurs intimately mixed with the Archean granitic gneiss on the northwest side of Sheep Mountain and may extend considerably further to the west of the area where it is has currently been mapped (**Fig.** 21). The unit is a gray, biotite-magnetite granite that weathers to a red or pink color, especially in this burned area. It has a preliminary U-Pb zircon date of ~2.0 Ga (K. Chamberlain, unpublished data). At several locations, including this stop, the granite contains locally well-developed segregations of coarser granite (**Fig.** 26). The segregations occur as discontinuous tubular features and probably represent fluid-rich pockets that formed in the granitic magma during its emplacement and crystallization. They tend to contain higher concentrations of magnetite (0.1-1.0 cm in diameter) than the adjacent finer grained granite and commonly have thin biotite-rich rims along their contacts. The SMG can locally be observed cutting the banding in the Archean gneiss, but it appears to be locally banded where it contains the coarser segregations because the segregations are aligned and locally folded by younger deformation (**Fig.** 27).

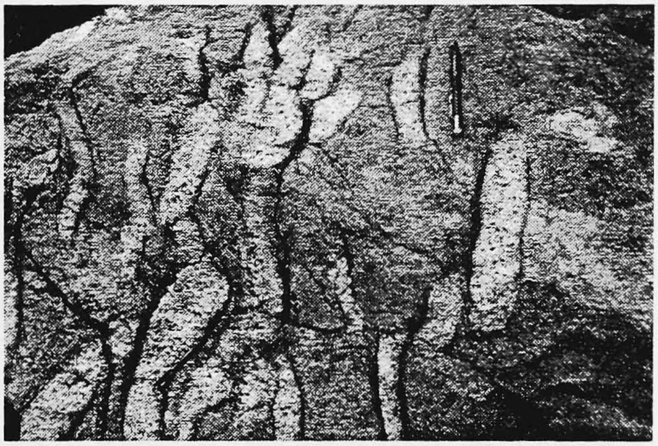

Figure 26. Sheep Mountain granite containing elongate segregations of coarser grained granite.

The rocks of this area have seen a complex sequence of Early Proterozoic deformation. The granitic gneiss between Stop 2 and this stop is folded into z-symmetry F_2 folds that plunge to the south and have a strong L_2 lineation parallel to their hingelines. In the area of this stop, the rocks and the L_2 lineations are folded by s-symmetry folds with steeply dipping, northeast-striking axial planes and by small-scale folds with westerly striking axial planes that dip gently to the north or northeast. The axial plane orientations of these younger fold events is not similar to that of the F_3 folds that occur as open warps with northwest-striking axial planes, but they are distinctly younger than the L_2 lineation. The relationship of these folds to the F_3 folding is not clear at this point.

The relationship of the s-symmetry folding to the L_2 linear fabric is dramatically illustrated in a folded mafic dike at this stop. One of the dikes contains abundant inclusions of granitic fragments. This is probably an intrusion breccia in a dike that carried gneissic fragments to its site of emplacement, but the origin of this unit will undoubtedly be a point of discussion on the outcrop. Where this unit is least deformed, the included fragments vary from less than a centimeter to as much as 4 centimeters across; however, the fragments are strongly deformed and ideally characterize the complex strains and associated fabrics in this unit. The earliest fabric in the rock is a strong linear fabric that is characteristic of the regionally

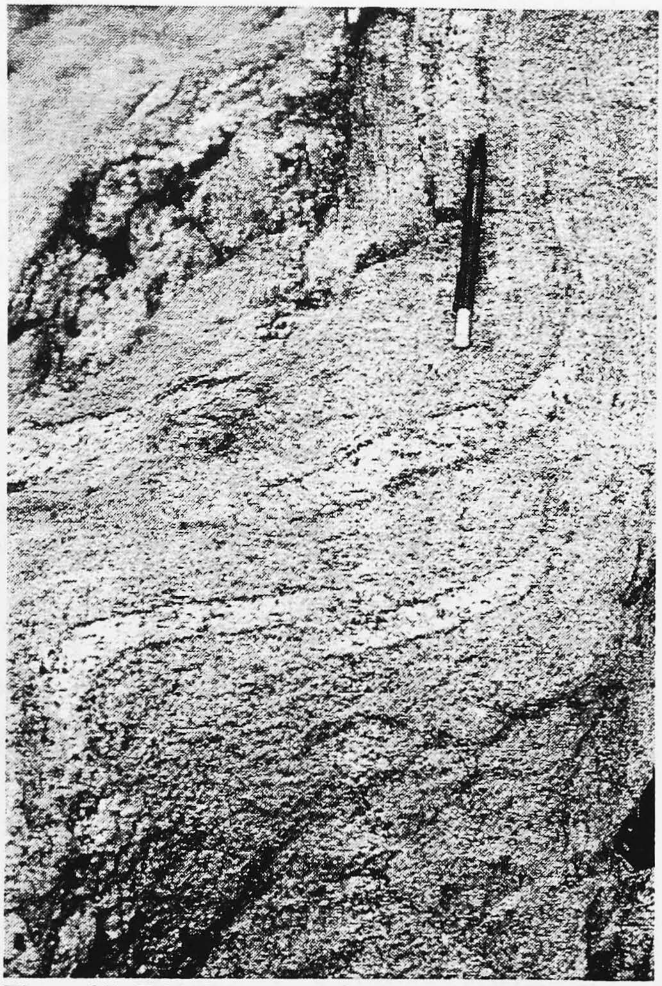

Figure 27. Folded coarse-grained granite segregations in the Sheep Mountain granite.

Figure 28. Strong L2 linear fabric developed in mafic dike intrusion breccia.

developed L_2 fabric (**Fig.** 28). However, this unit, the adjacent granitic gneiss, another mafic dike, and layers of the SMG have all been deformed into large s-symmetry folds that deform the L_2 lineation. Figure 29 shows an example of the strong strain gradients that develop locally within the fragment-bearing dike unit, and Figure 30 shows an example of s-symmetry folding in some of the strongly strained parts of the unit.

Early, probable F_2 folds in some of the adjacent gneiss are strongly flattened or complexly reoriented near the contacts of the folded dikes, and the gneissic banding is also folded into large, s-symmetry folds.

The younger folds with northerly dipping axial planes are not as strongly developed at this specific site, but they are locally well developed in outcrops to the southeast of this stop. Examples of this event are best developed in the SMG where a strong southerly plunging L_2 lineation is deformed about a northerly dipping axial plane.

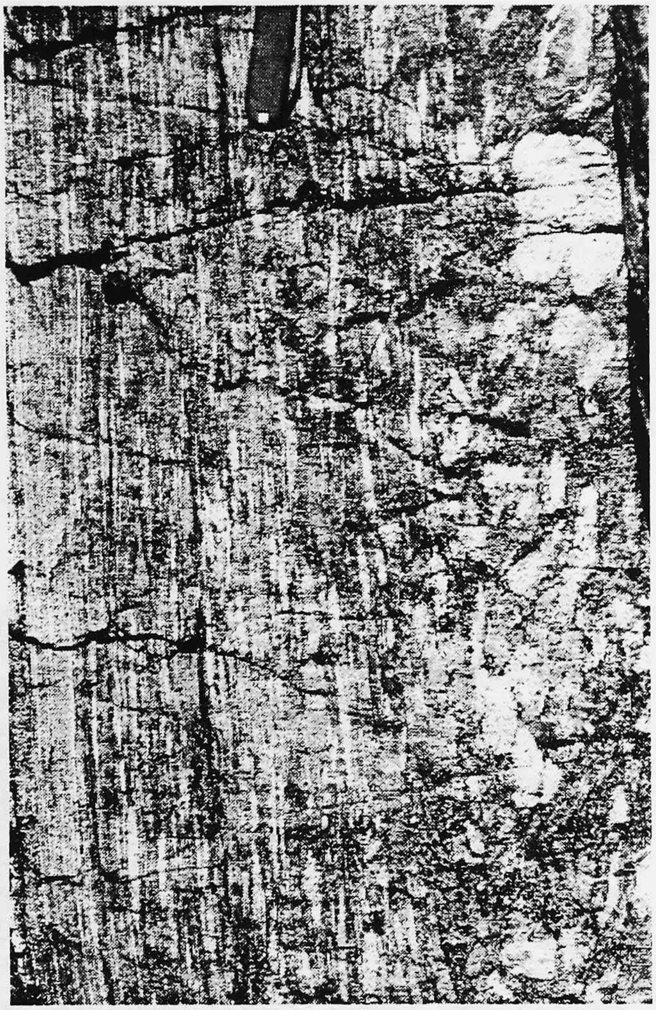

Figure 29. Strong lateral strain gradient in the mafic dike intrusion breccia.

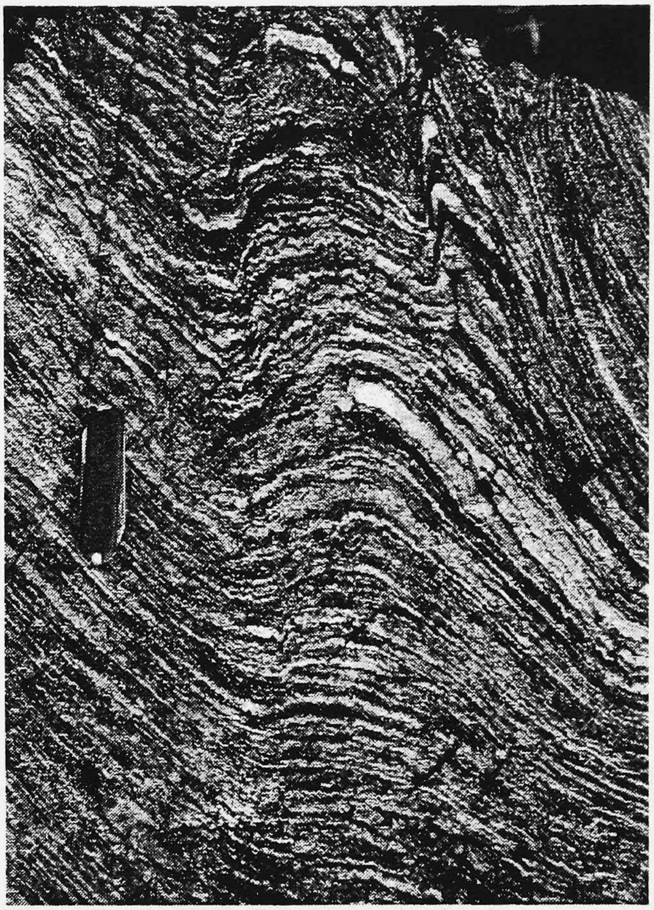

Figure 30. S-folding of highly strained mafic intrusion breccia.

Road travel to Stop 4

Continue westward on the Fletcher Park Road. Approximately 2.0 mi beyond Stop 3, we leave Platte County and cross the Albany County line. Drive another 2.7 mi to reach Stop 4, a roadcut along Fletcher Park Road. The parking is tight at this locality, and the road is narrow. This stop is situated in the NW 1/4, NE 1/4 of section 23, T.26N., R.71W (**Fig.** 31).

Stop 4 - Foliated granite containing folded schlieren layer, Fletcher Park Road

This roadcut exposure provides an example of the variably deformed granite south of the Laramie Peak shear zone but here affected by the D_2 shear zone deformation manifested by a S_2 foliation and a L_2 elongation lineation. A dark layer (schlieren) has been folded into an open fold, and S_2 is subparallel to the axial surface of this fold. Small-scale ductile shear zones occur within this set of exposures and represent local strain localization during the D_2 deformation.

Road travel to Stop 5

Continue northwestward along the Fletcher Park Road. Approximately 1.1 mi beyond Stop 4, we enter Fletcher Park, a flat open area at ~6900 feet elevation. After 0.35 miles, we cross a saddle through a large mafic dike, take the next right and drive 0.1 mi into an open meadow. Park here and continue by foot along the abandoned log road. After ~300 m, the road passes through a prominent saddle in a mafic dike. This is the same dike that the Fletcher Park Road passed through about 0.25 mi to the south along strike. This stop will involve approximately 0.6 mi of walking (round trip). The start of our traverse in the saddle along the log road is in the east-central part of section 15, T.26N., R.71W.

Stop 5 - Mafic dike deformed in the Laramie Peak shear zone.

This dike is one example of an extensive swarm of dikes that intruded the granites, gneisses, and supracrustal

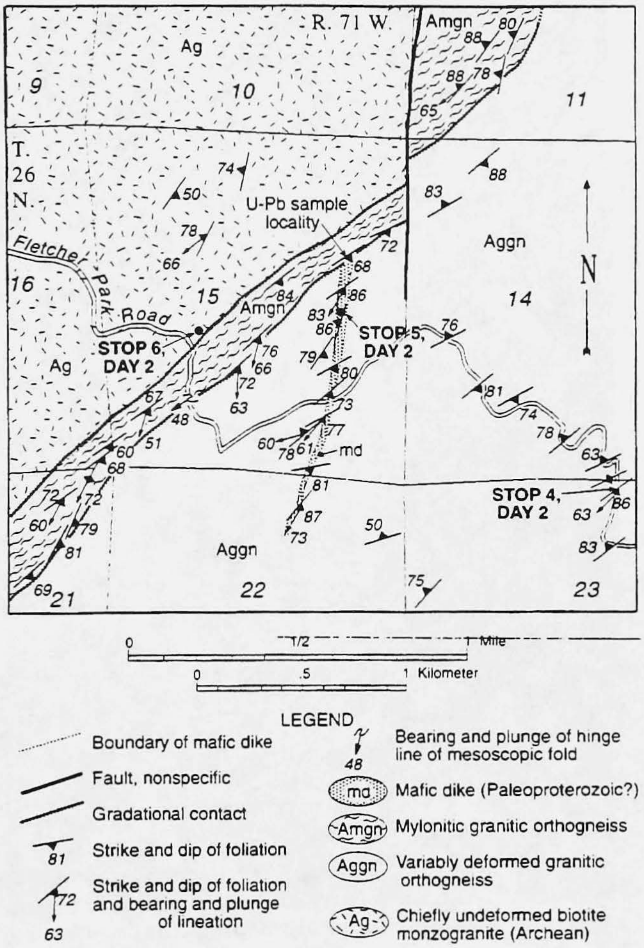

Figure 31. Geologic map of the Laramie Peak shear zone and environs near Fletcher Park, central Laramie Mountains, Wyoming. The base map was derived from the U.S. Geological Survey Fletcher Park 7 1/2 minute quadrangle. Geology by A.W. Snoke (1993 and 1994). Only the mafic dike (referred to as the "Fletcher Park dike") that was dated with U-Pb techniques is shown on the map. The area is widely intruded by Early Proterozoic mafic dikes; the distribution of these dikes is shown in Snyder et al. (1995). The sites of Stops 4, 5, and 6 for Day 2 are located on the map.

rocks of the Archean Wyoming province during the early Proterozoic. Several related dikes have been dated at ~2.0 Ga (Cox et al., 1995). This dike is unique because it has a more northerly strike (~10°) than other dikes near the Laramie Peak shear zone and merges into the shear zone at a high angle ~600 m north of the Fletcher Park Road (**Fig.** 31). Starting at the abandoned log road, walk north along the crest of the ridge formed by the dike. Note the heterogeneity of both metamorphism and deformation as you walk along the dike. Local, anastomosing shear zones cut across the dike. These shear zones have steeply-dipping, northeast-striking foliations and steep, southwest-plunging hornblende lineations subparallel to elongation lineations common in mylonitic granitic gneiss of the Laramie Peak shear zone, several hundred meters to the north. In these shear zones, dike rock contains an amphibolite-facies metamorphic assemblage including: hornblende, plagioclase (An_{57}), quartz, sphene, ilmenite, and magnetite (**Fig.** 32). Just before the crest of the hill, part of the dike retains its original igneous texture and mineral assemblage. This assemblage includes: orthopyroxene, clinopyroxene (including exsolved pigeonite), plagioclase (An_{64}), ilmenite, and magnetite (exsolved titanomagnetite) (**Figs.** 33 and 34). There is a strong spatial relationship between deformation and amphibolite-facies metamorphism in the dike that suggests that the deformation zones may have channelized fluids and therefore localized metamorphism.

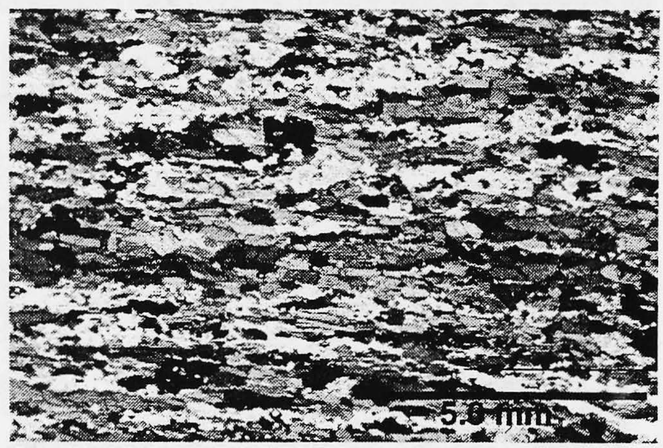

Figure 32. Schistose amphibolite dike rock (Fletcher Park dike) from within the Laramie Peak shear zone. This rock contains an amphibolite-facies assemblage: hornblende (gray)-plagioclase (~An_{57}, colorless)-quartz (colorless)-sphene (colorless, high relief), and an overall texture that indicates pervasive deformation and recrystallization. The rock is compositionally banded with mm-scale plagioclase- and hornblende-rich layers. Hornblende has a well-developed preferred orientation, and plagioclase is recrystallized into subequant, fine-grained (0.1-0.5 mm) crystals.

Just north of the high point on the ridge, the dike intersects the southern margin of the Laramie Peak shear zone. At this point the dike bends to the east and thins as it merges into the shear zone. The dike within the shear zone is a well-foliated, lineated amphibolite. This is the sample locality used for dating the deformation in the Laramie Peak shear zone. Syntectonic sphene growth has been dated at 1,763 ± 7 Ma, coeval with Cheyenne belt deformation in the Sierra Madre (1.74-1.76 Ga) (Resor et al., 1996).

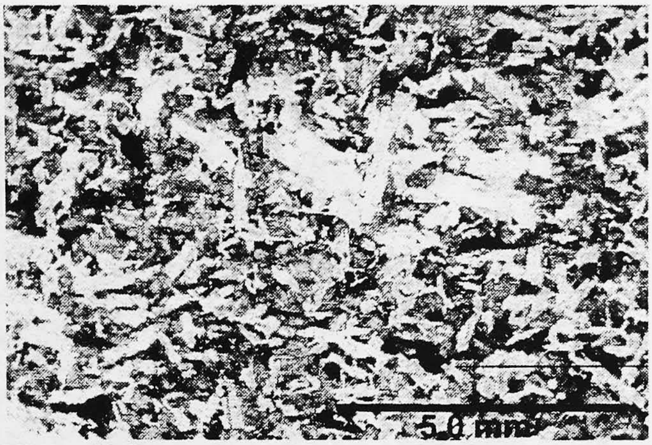

Figure 33. Two-pyroxene diabase exhibiting subophitic to intergranular texture. Large, zoned tabular plagioclase grains (colorless, low relief; average composition ~An_{65}) are separated by anhedral to subhedral hypersthene and pigeonite grains (colorless, high relief). Opaque grains are lamellar intergrowths of magnetite and ilmenite that appear to have formed from exsolution of original titanomagnetite. Hornblende growth (gray), particularly prevalent around opaque grains, indicates minor alteration under amphibolite-facies conditions.

Figure 34. Amphibolitic dike rock with relict igneous texture. The original igneous mineral assemblage has been replaced by the amphibolite-facies assemblage: hornblende (gray)-plagioclase (~An_{55}, colorless)-scapolite (colorless)-quartz (colorless)-sphene (colorless, high relief); however, a relict, original igneous texture is still preserved. Metamorphism was not accompanied by significant deformation in this sample. Plagioclase grains generally retain their igneous tabular habit, although they exhibit irregular grain boundaries with hornblende suggesting partial replacement, and are also partially replaced by scapolite. Unusual clusters of sphene grains in the upper right corner suggest near complete replacement of ilmenite lamellae from an exsolved titanomagnetite grain.

Continue to walk to the northeast along the dike for ~200 m to the last visible outcrop of amphibolitic dike rock. From this point head north-northeast downhill to a prominent outcrop of mylonitic granitic gneiss. Smaller outcrops throughout this area display mylonitic textures and lineation-parallel folds. The large outcrop is made up of mylonitic hornblende-biotite monzogranite with conspicuous feldspar augen. The mylonitic fabric in the granite includes a southeast-dipping, northeast-striking foliation and a southwest-plunging elongation lineation. This outcrop appears to have been rotated to a lower dip due to mass-wasting processes, so that the orientation of structural data differ from surrounding outcrops. Note the abundant feldspar porphyroclasts. The majority of the porphyroclasts do not yield a conclusive sense-of-shear, however, some asymmetric porphyroclasts do yield a consistent south-side-up shear-sense. An asymmetric fold within a mafic enclave in the lower part of this outcrop is also consistent with a south-side-up sense-of-shear.

Road travel to Stop 6

Return to the Fletcher Park Road and turn right. The road takes a sharp bend and crests an open ridge in 0.7 miles. The prominent granitic knob on the ridgeline to the right is the "Lookout." The "Lookout" is located on the southeastern contact of the Laramie Peak shear zone and affords an excellent view of the surrounding country. However, continue northwestward about 0.2 mi along the Fletcher Park Road until you reach a log road on your right; turn onto this road and park immediately on the left side of the road.

Stop 6 - Northwest Margin of the Early Proterozoic Laramie Peak shear zone near the Lookout

The discussion for this stop will begin in the NE 1/4, SW 1/4 of section 15, T.26N., R.71W. Walk northwest to rounded, bouldery appearing exposures of granite. These exposures consist of undeformed to very weakly deformed, locally megacrystic biotite monzogranite (~2.6-Ga Laramie Peak-type granite) with the occasional mafic enclave. This rock type, intruded by variably textured mafic dikes, forms the bulk of the terrane northwest of the Laramie Peak shear zone. The Archean migmatitic banded gneiss unit, common southeast of the Laramie Peak shear zone, has not yet been identified in our map area northwest of the shear zone.

Walk approximately east-southeast toward the small bouldery hillock with prominent exposures. Along the ridgeline behind this set of exposures is a rock body shaped much like a gunsight when viewed from our present vantage point. The Lookout, with its thin wooden cross on top, is toward the southeast. Before you reach the hillock, note small exposures of protomylonitic, foliated and lineated biotite monzogranite; these rocks define the northwest margin of the Laramie Peak shear zone. These protomylonitic granitic rocks exhibit heterogeneous strain containing thin zones of concentrated grain-size reduction; σ-type porphyroclasts that indicate southside-up displacement along the Laramie Peak shear zone are locally present (**Fig.** 35).

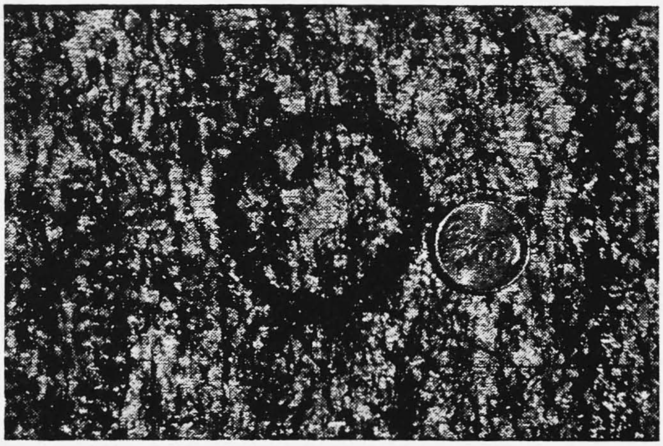

Figure 35. Feldspar porphyroclast in mylonitic granitic gneiss, Laramie Peak shear zone, near Fletcher Park. The view is approximately perpendicular to foliation and subparallel to lineation; southeast is to the right. This sigma-type porphyroclast with conspicuous asymmetric tails indicates a south-side-up sense-of-shear. The site of the photograph is Stop 6, Day 2.

At the bouldery hillock, small-scale folds that fold the mylonitic foliation indicate sinistral vergence in concurrence with the sigma-type porphyroclasts at the previous exposure. Beyond the exposures that form the bouldery hillock are other exposures of mylonitic granitic gneiss that exhibit excellent sense-of-shear indicators. In one exposure, a σ-type quartzose lens (**Fig.** 36) clearly indicates a southside-up displacement history. The mylonitic rocks in this part of the shear zone include color-banded, laminated granitic mylonite (**Fig.** 4).

Road travel to Stop 7

Return to the Fletcher Park Road and turn right and thus continue to the west until you reach the junction with the Cottonwood Park Road. Turn left onto the Cottonwood Park Road (i.e., southward). After 0.9 mi, bear left at a fork and continue another 0.4 mi to a sharp right-hand turn. Park on the left-hand shoulder. Walk north and uphill from the bend in the road to several distinctive, blocky red outcrops. This locality (S top 7) is near the center of section 19, T.26N., R.71W.

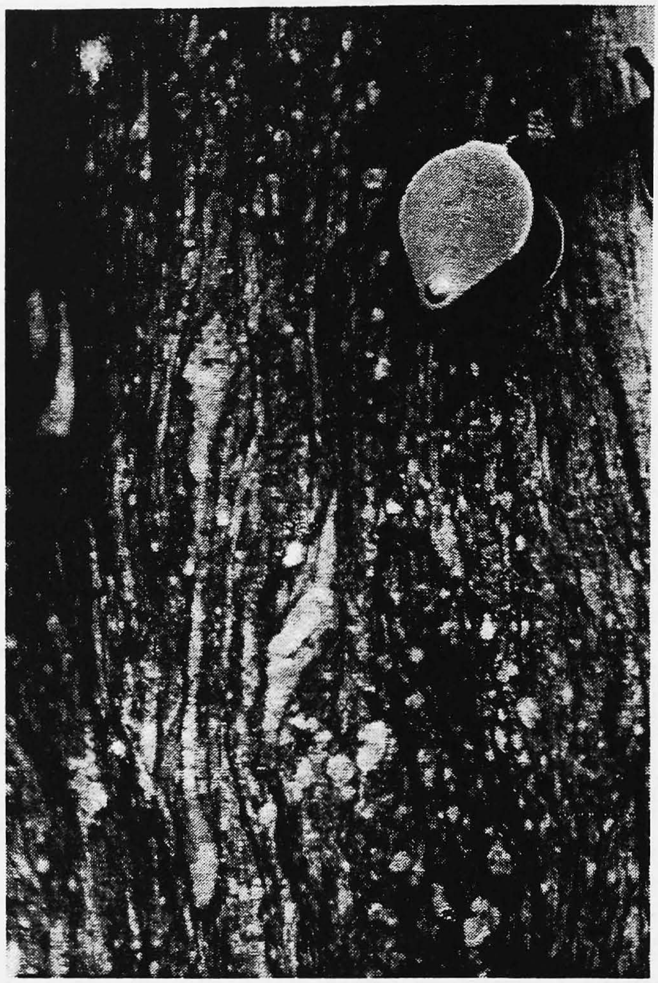

Figure 36. Asymmetric quartzose lens within mylonitic granitic gneiss, Laramie Peak shear zone, near Fletcher Park. The view is approximately perpendicular to foliation and subparallel to lineation. The sigma-type shape of the lens indicates a south-side-up sense-of-shear. Note hand lens for scale. The site of the photograph is Stop 6, Day 2.

Stop 7 - Sturgeon Creek fault zone

These blocky outcrops are part of the Sturgeon Creek fault, a steeply dipping brittle fault that continues to the south for >10 km, generally parallel to Sturgeon Creek. The fault also presumably continues north through Cottonwood Park and may be the continuation of one of the faults mapped by Snyder (1993) that bounds the basement block containing the Esterbrook greenstone belt.

The Sturgeon Creek fault is characterized by a 5-10 m-wide zone of brecciation with significant hydrothermal alteration. Note the oxidation and quartz, calcite, and epidote veining within the fault zone. Slickenside lineations can be found on several of the surfaces at this outcrop and are generally plunging gently to the north (**Fig.** 37). Note that the fault zone cuts across the

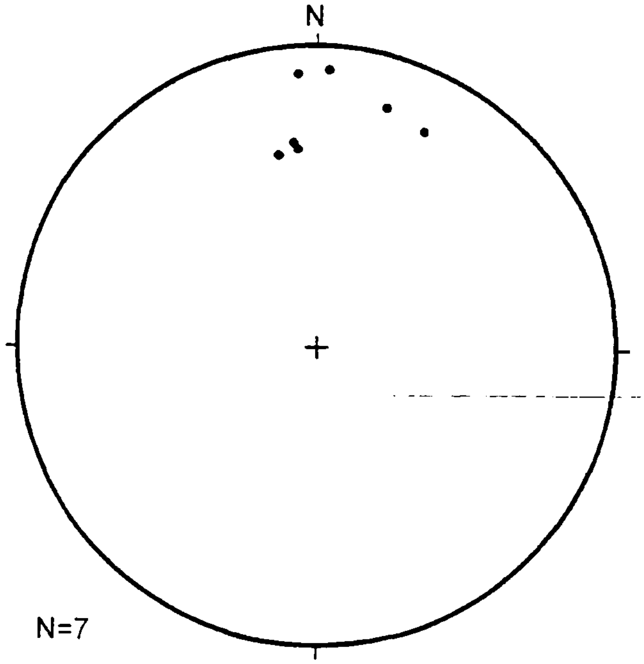

Figure 37. Stereographic projection of the trend and plunge of slickenside lineations in the Sturgeon Creek fault zone.

topography indicating a steep dip.

The absolute displacement across the fault is uncertain, however, the fault is associated with a 1.25-km horizontal separation of the Laramie Peak shear zone. This horizontal separation of the shear zone, a steeply dipping marker, and the gently dipping slickensides at these outcrops suggest a significant strike-slip component for the fault displacement.

The age of the fault is unconstrained, but it is presumed to be Laramide, based on its brittle deformation and associated extensive low-grade (hydrothermal) alteration. The Sturgeon Creek fault is one of a set of north-striking Laramide faults that transect the Precambrian basement in the north-central Laramie Mountains. A second set of faults, including the North Laramie River fault ~3 km south of this outcrop, strikes northeast, subparallel to the Precambrian structural grain.

Road travel to Stop 8

Continue south on the Cottonwood Park Road. Breccias of the Sturgeon Creek fault zone are exposed again in roadcuts 0.7 mi to the south. More breccias of the Sturgeon Creek fault zone are exposed at "Turkey Rock" about 0.9 mi farther south along the Cottonwood Park Road. Still farther south, the road crosses the North Laramie River (location of another brittle fault) and then passes the Double Four ranch on the left. As we drive farther south along the Cottonwood Park Road, we continue to travel subparallel to the Sturgeon Creek fault zone and pass through several prominent saddles that expose altered fault-zone rocks. After driving south about 6.1 mi from Stop 7 and just before the road begins to climb west out of the Sturgeon Creek valley, pull off the road and park. We will then cross a hay meadow and Sturgeon Creek to reach a prominent outcrop above the creek (**Fig.** 38). This large outcrop which is the focus of this stop is situated in the SW 1/4, NW 1/4 of section 19 of T.25N., R.71W.

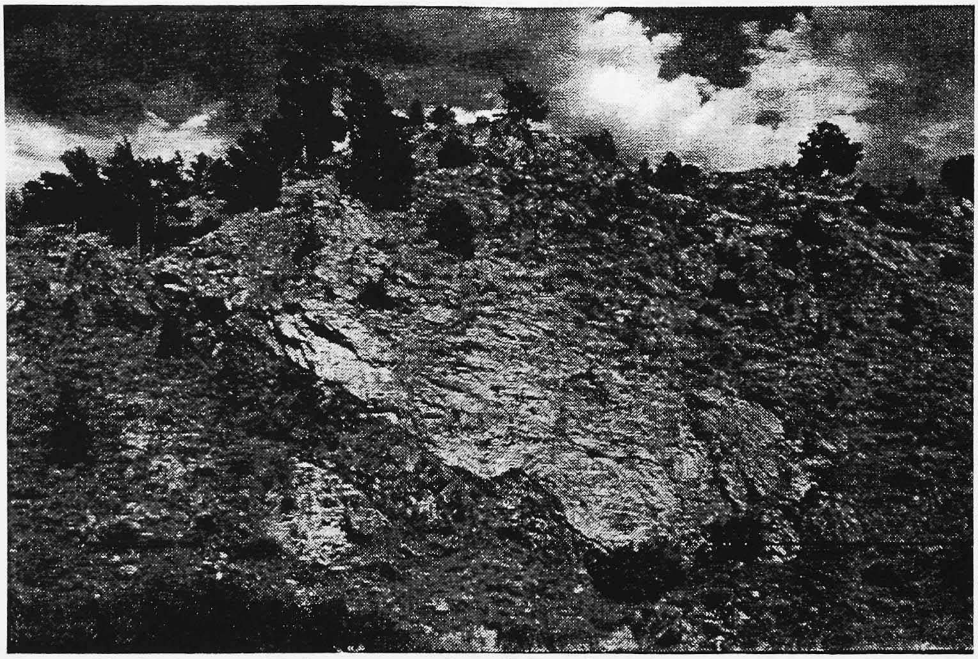

Figure 38. Photograph of outcrop at Stop 8, Day 2 showing a large mafic dike [Early Proterozoic(?)] intruded discordantly across gneissic foliation of the Archean, migmatitic banded gneiss unit. In this same exposure, a shear-zone fabric is developed along the lower margin of the mafic dike and is locally superposed on both the dike and its banded gneiss wall rocks. This locality thus indicates that the gneissic banding (D_1) pre-dates the mafic dike (inferred as a member of ~2.0 Ga suite), whereas the shear-zone deformation (D_2) is post dike swarm (this deformation has been dated as ~1.76 Ga at Stop 5, Day 2).

Stop 8 - Mafic dike in migmatitic banded gneiss unit

This large outcrop exhibits the relative temporal relationships between two high-temperature deformational phases and a widespread mafic magmatic event in the north-central Laramie Mountains. Specifically, these elements are: 1) a late Archean (>2.6 Ga) high-temperature event, 2) a widespread Early Proterozoic (~2.0 Ga) mafic dike swarm , and 3) an Early Proterozoic (~1.76 Ga) shear-zone deformation.

A late Archean to Early Proterozoic high-temperature deformation led to the formation of compositional banding, foliation, folding, and leucogranitic layering in the migmatitic banded gneiss. These elements are all older than the intrusion of the amphibolitic mafic dikes (~2.0 Ga). A prominent dike cuts across fabrics in the banded gneisses. This relationship can clearly be seen along the upper contact of the dike at the upper (northern) end of the cliff face, where gneissic banding and the contact of the mafic dike are nearly orthogonal (89°). In the southern half of the cliff, the dike appears to cut an outcrop-scale, northward vergent antiform within migmatitic banded gneiss. Relative temporal relationships from other outcrops indicate that these fabrics are late Archean as they are also cut by massive late Archean (~2.6 Ga) granites.

Heterogeneous early Proterozoic (~1.76 Ga) shear-zone deformation transposes the late Archean fabric within the migmatitic banded gneiss and the mafic dike. A narrow (~0.5-wide) mylonitic shear zone is localized along the lower contact of the dike. The shear-zone foliation is defined by elongated quartz and feldspar grains and transposes the older gneissic banding. A parallel foliation is also found in localized deformation zones within the amphibolitic mafic dike.

ACKNOWLEDGEMENTS

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REFERENCES CITED

Aleinikoff, J.N., 1983, U-Th-Pb systematics of zircon inclusions in rock-forming minerals: a study of armoring against isotopic loss using the Sherman Granite of Colorado-Wyoming, U.S.A.: Contributions to Mineralogy and Petrology, v. 83, p. 259-269.

Allmendinger, R.W., Brewer, J.A., Brown, L.D., Kaufman, S., Oliver, J.E., and Houston, R.S., 1982, COCORP profiling across the Rocky Mountain Front in southern Wyoming, Part 2: Precambrian basement structure and its influence on Laramide deformation: Geological Society of America Bulletin, v. 93, p. 1253-1263.

Baird, D.J., Nelson, K.D., Knapp, J.H., Walters, J.J., and Brown, L.D., 1996, Crustal structure and evolution of the Trans-Hudson orogen: Results from seismic reflection profiling: Tectonics, v. 15, p. 416-426.

Ball, T.T, and Farmer, G.L., 1991, Identification of 2.0 to 2.4 Ga Nd model age crustal material in the Cheyenne belt, southeastern Wyoming.: Implications for Proterozoic accretionary tectonics at the southern margin of the Wyoming craton. Geology, v. 19, p. 360-363.

Bauer, R.L., Gresham, D.A, and Edson, J.D., 1995, Early Proterozoic ductile reworking of Archean basement in the central Laramie Range: A complex response to the Cheyenne belt, Trans-Hudson and Central Plains orogens: 12th International Conference on Basement Tectonics, v. 12, p. 21-22.

Bauer, R.L., Gresham, D.A, and Edson, J.D., 1996, Multiphase Early Proterozoic ductile reworking of Archean basement in the central Laramie Mountains, SE Wyoming: Geological Society of American, Abstracts with Programs, v. 28, no. 4, p. 1-2.

Bickford, M.E., Van Schmus, W.R., and Zietz, I., 1986, Proterozoic history of the midcontinent region of North America: Geology, v. 14, p. 492-496.

Bickford, M.E., Collerson, K.D., Lewry, J.F., Van Schmus, W.R., and Chiarenzelli, J.R., 1990, Proterozoic collisional tectonism in the Trans-Hudson orogen, Saskatchewan: Geology, v. 18, p. 14-18.

Bochensky, P., and Frost, B.R., 1982, Contact metamorphism produced by the Laramie Anorthosite Complex, Morton Pass, Wyoming: Geological Society of America Abstracts with Programs, v. 14, p. 303.

Chamberlain, K. R., Patel S.C., Frost, B.R., and Snyder, G.L., 1993, Thick-skinned deformation of the Archean Wyoming province during Proterozoic arc-continent collision: Geology, v. 21, p. 995-998.

Condie, K.C., 1969, Petrology and geochemistry of the Laramie batholith and related metamorphic rocks of Precambrian age, eastern Wyoming: Geological Society of America Bulletin, v. 80, p. 57-82.

Condie, K.C., 1982, Plate-tectonics model for Proterozoic continental accretion in the southwestern United States: Geology, v. 10, p. 37-42.

Condie, K.C., and Shadel, C.A., 1984, An early Proterozoic volcanic arc succession in southeastern Wyoming: Canadian Journal of Earth Sciences, v. 21, p. 415-427.

Cox, D.M., Chamberlain, K.R., Heaman, L.M., and Snyder, G.L., 1993, U-Pb ages of mafic intrusions from the southeast margin of the Wyoming province: Implications for timing of rifting and deformation: Geological Society of America, Abstracts with Programs, v. 25, no. 5, p. 24.

Cox, D.M., Chamberlain, K.R., and Snyder, G.L., 1995, Timing and petrogenesis of Early Proterozoic high Mg mafic magmatism in the southern Wyoming province, U.S.A.: Geological Association and Mineralogical Association of Canada Annual Meeting, Program with Abstracts, v. 20, p. A20.

DePaolo, D.J., 1981, Neodymium isotopes in the Colorado Front Range and crust; Mantle evolution in the Proterozoic: Nature, v. 291, p. 193-196.

Divis, A.F., 1976, Geology and geochemistry of the Sierra Madre Range, Wyoming: Colorado School of Mines Quarterly, v. 71, 127 p.

Duebendorfer, E.M., 1988, Evidence for an inverted metamorphic gradient associated with a Precambrian suture, southern Wyoming: Journal of Metamorphic Geology, v. 6, p. 41-63.

Duebendorfer, E.M., and Houston, R.S., 1986, Kinematic history of the Cheyenne belt, Medicine Bow Mountains, southeastern Wyoming: Geology, v. 14, p. 171-174.

Duebendorfer, E.M., and Houston, R.S., 1987, Proterozoic accretionary tectonics at the southern margin of the Archean Wyoming craton: Geological Society of America Bulletin, v. 98, p. 554-568.

Duebendorfer, E.M., and Houston, R.S., 1990, Structural analysis of a ductile-brittle Precambrian shear zone in the Sierra Madre, Wyoming: Western extension of the Cheyenne belt?: Precambrian Research, v. 48, p. 21-39.

Dutch, S.I., and Nielsen, P.A., 1990, The Archean Wyoming Province and its relations with adjacent Proterozoic provinces, *in* Lewry, J.F., and Stauffer, M.R., eds., The Early Proterozoic Trans-Hudson orogen of North America: Geological Association of Canada, Special Paper 37, p. 287-300.

Graff, P.J., Sears, J.W., Holden., G.S., and Hausel, W.D., 1982, Geology of the Elmers Rock greenstone belt, Laramie Range, Wyoming: Geological Survey of Wyoming Report of Investigations 14, 23 p.

Grant, J.A., and Frost, B.R., 1986, Decompression, metamorphism and melting in the aureole of the Laramie Anorthosite Complex: Geological Society of America Abstracts with Programs, v. 18, p. 620.

Grant, J.A., and Frost, B.R., 1990, Contact metamorphism and partial melting of pelitic rocks in the aureole of the Laramie Anorthosite Complex, Morton Pass, Wyoming: American Journal of Science, v. 290, p. 425-472.

Gresham, D.A., 1994, Early Proterozoic multiple deformation in the Archean gneiss of the east-central Laramie Range, Wyoming [M.S. thesis]: Columbia, University of Missouri, 99 p.

Hall, R.P., Hughes, D.J., Friend, C.R.L., and Snyder, G.L., 1987, Proterozoic mantle heterogeneity: Geochemical evidence from contrasting basic dykes. *in* T.C. Pharoah, R.D. Beckinsale and D. Richard, eds., Geochemistry and mineralization of Proterozoic volcanic suites: Special Publication of the Geological Society of London, v. 33, p. 9-21.

Hills, F.A., and Houston, R.S., 1979, Proterozoic tectonics of the central Rocky Mountains, North America: University of Wyoming Contributions to Geology, v. 17, p. 89-109.

Hills, F.A., and Armstrong, R.L., 1974, Geochronology of Precambrian rocks in the Laramie Range and implications for the tectonic framework of Precambrian southern Wyoming: Precambrian Research, v. 1, p. 213-225.

Hodge, D.S., 1966, Petrology and structural geometry of Precambrian rocks in the Bluegrass area, Albany County, Wyoming [Ph.D. dissertation]: Laramie, University of Wyoming, 135 p.

Hoffman, P.F., 1981, Autopsy of Athapuscow aulocogen: a failed arm affected by three collisions, *in* Campbell, F.J.A., ed., Proterozoic basins of Canada: Geological Survey of Canada Paper 81-10, p. 97-102.

Holden, G.S., 1978, The Slate Creek metamorphic terrain, Albany County, Wyoming [Ph.D. dissertation]: Laramie, University of Wyoming, 95 p.

Holden, G.S., and Snyder, G.L., 1983, Compositional variation of mafic rocks from an Archean granite-greenstone terrane, central Laramie Range, Wyoming: Geological Society of America Abstracts with Programs, v. 15, p. 423.

Houston, R.S., 1993, Late Archean and Early Proterozoic geology of southeastern Wyoming, *in* Snoke, A.W., Steidtmann, J.R., and Roberts, S.M., eds., Geology of Wyoming: Geological Survey of Wyoming Memoir No. 5, p. 78-116.

Houston et al., 1978, A regional study of rocks of Precambrian age in that part of the Medicine Bow Mountains lying in southeastern Wyoming - with a chapter on the relationship between Precambrian and Laramide structure (second printing of 1968 edition with updated preface): Geological Survey of Wyoming Memoir, No. 1, 167 p.

Houston, R.S., Karlstrom, K.E., Hills, F.A., and Smithson, S.B., 1979, The Cheyenne belt; A major Precambrian crustal boundary in the western United States: Geological Society of America Abstracts with Programs, v. 11, p. 446.

Houston, R.S., Duebendorfer, E.M., Karlstrom, K.E., and Premo, W.R., 1989, A review of the geology and structure of the Cheyenne belt and Proterozoic rocks of southern Wyoming, *in* Grambling, J.A., and Tewksbury. B.J., eds., Proterozoic geology of the southern Rocky Mountains: Boulder, Colorado, Geological Society of America Special Paper 235, p. 1-12.

Johnson, R.C., and Hills, F.A., 1976, Precambrian geochronology and geology of the Box Elder Canyon area, northern Laramie Range, Wyoming: Geological Society of American Bulletin, v. 87, p. 809-817.

Johnson, R.A., Karlstrom, K.E., Smithson, S.B., and Houston, R.S., 1984, Gravity profiles across the Cheyenne belt, a Precambrian crustal suture in southeastern Wyoming: Journal of Geodynamics, v.1, p. 445-472.

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

Karlstrom, K.E., Houston, R.S., Flurkey, A.J., Collidge, C.N., Kratochvil, A.K., and Sever, D.K., 1981, A summary of the geology and uranium potential of Precambrian conglomerates in southeastern Wyoming: Bendix Field Engineering Corporation Report DJBX-139-81, 541 p.

Karlstrom, K.E., Flurkey, A.J., and Houston, R.S., 1983, Stratigraphy and depositional setting of Proterozoic metasedimentary rocks in southeastern Wyoming; record of an Early Proterozoic Atlantic-type cratonic margin: Geological Society of American Bulletin, v. 94, p. 1257-1294.

Klasner, J.S., and King, E.R., 1986, Precambrian basement geology of North and South Dakota: Canadian Journal of Earth Sciences, v. 23, p. 1083-1102.

Klasner, J.S., and King, E.R.,1990, A model for tectonic evolution of the Trans-Hudson orogen in North and South Dakota, *in* Lewry, J.F. and Stauffer, M.R., eds., The Early Proterozoic Trans-Hudson orogen of North America: Geological Association of Canada, Special Paper 37, p. 271-285.

Kohn, M.J., and Spear, F.S., 1990, Two new geobarometers for garnet amphibolites, with applications to southeastern Vermont. American Mineralogist, v. 75, p. 89-96.

Lewry, J.F., and Collerson, K.D., 1990, The Trans-Hudson orogen: Extent, subdivision, and problems, *in* Lewry, J.F. and Stauffer, M.R., eds., The Early Proterozoic Trans-Hudson Orogen of North America: Geological Association of Canada, Special Paper 37, p. 1-14.

Nelson, K.D., Baird, D.J., Walters, J.J., Hauck, M., Brown, L.D., Oliver, J.E., Ahern, J.L., Hanjal, Z., Jones, A.G., and Sloss, L.L., 1993, Trans-Hudson orogen and Williston Basin in Montana and North Dakota: New COCORP deep-profiling results: Geology, v. 21, p. 447-450.

Patel, S.C., 1992, Rb-Sr Isotope systematics and Proterozoic Barrovian metamorphism, Laramie Mountains, Wyoming [Ph.D. dissertation]: Laramie, University of Wyoming, 184 p.

Patel, S.C., Frost, B.R., and Snyder, G.L., 1991, Extensive Early Proterozoic Barrovian metamorphism in the southeastern Wyoming province, Laramie Range, Wyoming: Geological Society of America Abstracts with Programs, v. 23, p. 59.

Peterman, Z.E., and Hildreth, R.A., 1978, Reconnaissance geology and geochronology of the Precambrian of the Granite Mountains, Wyoming: U.S. Geological Survey Professional Paper 1055, 22 p.

Premo, W.R., and Van Schmus, W.R., 1989, Zircon geochronology of Precambrian rocks in southeastern Wyoming and northern Colorado, *in* Grambling, J.A., and Tewksbury, B.J., eds., Proterozoic geology of the southern Rocky Mountains: Geological Society of American Special Paper 235, p. 1-12.

Resor, P.G., Chamberlain, K.R., Frost, C.D., Snoke, A.W., and Frost, B.R., 1996, Direct dating of deformation: U-Pb age of syndeformational sphene growth in the Proterozoic Laramie Peak shear zone: Geology, v. 24, p. 623-626.

Ridgley, N.H., 1971, Precambrian rocks in the Blackhall Mountain area, Carbon County, Wyoming [M.S. thesis]: Laramie, University of Wyoming, 50 p.

Scoates, J.S., and Chamberlain, K.R., 1995, Baddeleyite (ZrO_2) and zircon ($ZrSiO_4$) from anorthositic rocks of the Laramie Anorthosite Complex, Wyoming: Petrologic consequences and U-Pb ages: American Mineralogist, v. 80, p. 1317-1327.

Sims, P.K., and Peterman, Z.E., 1986, Early Proterozoic Central Plains orogen: A major buried structure in the north-central United States: Geology, v. 14, p. 488-491.

Smaglik, S.M., 1987, Petrogenesis and tectonic implications for Archean mafic and ultramafic magmas in the Elmers Rock greenstone belt, Laramie Range, Wyoming [M.S. thesis]: Golden, Colorado School of Mines , 126 p.

Snyder, G.L., 1984, Preliminary geologic maps of the central Laramie Mountains, Albany and Platte Counties, Wyoming: U.S. Geological Survey Open-Files Report 84-358 (Parts A-M), scale 1:24,000.

Snyder, G.L., 1986, Preliminary geologic maps of the Reese Mountain and part of the Hightower SW 7.5 minute quadrangles (part A) and parts of the Fletcher Park and Johnson Mountain 7.5 minute quadrangles (part B), Albany and Platte Counties, Wyoming: U.S. Geological Survey Open file Report 86-201, scale 1:24,000.

Snyder, G.L., Frost, B.R., Grant, J.A., Lindsley, D.H., and Peterman, Z.E., 1988, The crust of the young Earth - Guide to Precambrian continental core of southeastern Wyoming, *in* G.S. Holden, ed., Geological Society of America Field Trip Guidebook (Trip 1): Colorado School of Mines Professional Contributions no. 12, p. 1-34.

Snyder, G.L., Hall, R.P., Hughes, D.J. and Ludwig, K.R., 1990, Early Precambrian basic rocks of the USA, *in* Hall, R.P. and Hughes, D.J., eds., Early Precambrian basic magmatism, Glasgow, Blackie, p. 191-220.

Snyder, G.L., Hausel, W.D., Klein, T.L., Houston, R.S., and Graff, P.J., 1989, Precambrian rocks and mineralization, southern Wyoming province: Washington D.C., American Geophysical Union, 28th International Geological Congress, Field Trip Guidebook T332, 48 p.

Snyder, G.L., Siems, D.F., Grossman, J.N., Ludwig, K.R., and Nealey, L.D., *with a section on* the Fletcher Park shear zone by Frost, B.R., Chamberlain, K.R., and Snyder, G.L., 1995, Geologic map, petrochemistry, and geochronology of the Precambrian rocks of the Fletcher Park - Johnson Mountain Area, Albany and Platte Counties, Wyoming: U.S. Geological Survey Miscellaneous Investigations Series, Map I-2233, scale 1:24,000.

Spacuzza, M.J., 1990, High grade metamorphism and partial melting in the Bluegrass Creek suite, central Laramie Mountains, Wyoming [M.S. thesis]: Duluth, University of Minnesota, 115 p.

Tempelton, M.E., and Smithson, S.B., 1994, Seismic reflection profiling of the Cheyenne belt Proterozoic suture in the Medicine Bow Mountains, southeastern Wyoming: A tie to geology: Tectonics, v. 13, p. 1231-1241.

Thomas, M.D., Sharpton, V.L., and Grieve, R.A.F., 1987, Gravity patterns and Precambrian structure in the North American central plains: Geology, v. 15, p. 489-492.

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