Laramie Peak shear system, central Laramie Mountains, Wyoming, USA: Regeneration of the Archean Wyoming province during Palaeoproterozoic accretion

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Laramie Peak shear system, central Laramie Mountains, Wyoming, USA: regeneration of the Archean Wyoming province during Palaeoproterozoic accretion

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Abstract: The Laramie Peak shear system (LPSS) is a 10 km-thick zone of heterogeneous general shear (non-coaxial) that records significant tectonic regeneration of middle–lower crustal rocks of the Archean Wyoming province. The shear system is related to the 1.78–1.74 Ga Medicine Bow orogeny that involved the collision of an oceanic-arc terrane (Colorado province or Green Mountain block or arc) with the rifted, southern margin of the Wyoming province. The style and character of deformation associated with the LPSS is distinctive: a strong, penetrative (mylonitic) foliation commonly containing a moderately steep, SW-plunging elongation lineation. In mylonitic quartz-feldspathic gneisses of the Fletcher Park shear zone, shear-sense indicators indicate southside-up, and this interpretation is supported by metamorphic and geochronological studies across the LPSS. We argue that distributed general shear (non-coaxial) involving high-strain zones and multiple folding events yielded a broad, en-masse uplift (Palmer Canyon block) during the late stages of the Medicine Bow orogeny. The LPSS is thus an excellent example of how crystal-plastic strain is distributed in sialic crust during an oceanic arc–continental margin collision. As magmatism (and attendant thermal softening) did not occur in the Wyoming province during its partial subduction beneath the oceanic-arc terranes of the Colorado province, the crystal-plastic strain manifested within the Wyoming province is mechanical in nature and was concurrent with crustal thickening. Strain is localized into discrete shear zones separated by weakly deformed rocks. These high-strain zones are commonly located along contacts between differing rock types and we propose that mechanical and chemical weakening processes may have contributed to strain localization.

The rheological behaviour of the middle and lower crust is commonly cited as playing an important role in the mechanics and geometry of continental plate tectonics (e.g. Molnar 1988; Northrup 1996; Axen et al. 1998; Beaumont \textit{et al.} 2001; Jackson 2002; McKenzie \& Jackson 2002). In many cases, lower crustal deformation is inferred to involve broad-scale, viscous flowage; e.g. Basin and Range province of western North America (Gans 1987; Block \& Royden 1990; Wernicke 1990; Kruse \textit{et al.} 1991). This lower crustal flow appears to compensate for large variations in upper crustal deformation (extension in the case of the Basin and Range province) leading to relatively constant crustal thickness and a flat-lying Moho (Klemperer \textit{et al.} 1986; Hauser \textit{et al.} 1987). However, in other cases, the deep crust does not appear to flow over large areas and Moho topography is maintained over long periods (e.g. Alps of western Europe: Bois \& ECORS Scientific Party 1990). It has been suggested that this variation in the behaviour of the deep crust is associated with the presence or absence of tectonic thickening and/or syntectonic magmatism – processes that raise the temperature of the lower crust and facilitate its viscous flowage (e.g. McKenzie \& Jackson 2002).

This general model of lower crustal deformation is based in a large part on experimental studies, mechanical modelling and geophysical observations. Surface exposures of rocks that previously occupied lower crustal depths are relatively scarce. In this paper we present the results of detailed field mapping of one such exposure of deep crustal rocks from the
Laramie Mountains of SE Wyoming, USA. The Laramie Mountains expose Archean and Proterozoic rocks that are inferred to have occupied middle–lower crustal depths (>25 km) during an inferred Paleoarctozoic arc–continent collision (i.e. the 1.78–1.74 Ga Medicine Bow orogeny). In contrast to many previous studies of deep crustal exposures (e.g. MacCready et al. 1997; Klepeis et al. 2004; Karlstrom & Williams 2005) these rocks did not experience syntectonic magmatism, and thus the strain developed in a purely mechanical fashion without the role of magmatic softening as a deformational agent.

We argue that field observations demonstrate that deformation in the lower crust was characterized by strain localization with discrete shear zones separated by undeformed–weakly deformed rocks. Furthermore, we suggest that this style of deformation is typical of lower crustal deformation in the absence of syntectonic magmatism and thus presents an opportunity to directly observe lower crustal behaviour where broad-scale flowage is absent. Finally, we discuss the implications of these observations to tectonic models of arc–continent collisions. These results have implications for inferred Phanerozoic arc–continent collisions where a similar geological situation has been proposed (e.g. Dewey & Bird 1970; Speed & Sleep 1982; Suppe 1987; Snyder et al. 1996) as well as for general models of collisional orogenesis (e.g. Beaumont & Quinlan 1994; Snyder 2002).

Geological setting and structural subdivisions

The Precambrian rocks of SE Wyoming are exposed in late Cretaceous–early Eocene (Laramide) basement-involved uplifts (Fig. 1). These uplifts typically are bound on one or more of their flanks by a deep-rooted, reverse and/or thrust fault system that developed during the contractional Laramide orogeny. Three ranges (E–W) expose much of the Precambrian rocks of the area: Laramie Mountains, Medicine Bow Mountains and Sierra Madre. Other less extensive exposures of Precambrian rocks in the region include the Elk, Coad and Pennock mountains, Casper Mountain, Richeau and Cooney hills, and Hartville uplift (Fig. 1). The basement rocks exposed in these ranges record a long and complex Precambrian history that probably began in the Middle Archean (?) and continued into the Mesoproterozoic. Numerous papers summarize parts of this history, but especially useful review articles include Karlstrom & Houston (1984), Duebendorfer & Houston (1987), Houston et al. (1989, 1993) Houston (1993) and Chamberlain (1998).

A key tectonic event in the Precambrian geological evolution of SE Wyoming, USA, is the Medicine Bow orogeny (1.78–1.74 Ga) interpreted as a collisional orogeny involving the Colorado province (Palaeoproterozoic oceanic arc and associated rocks; also called the Green Mountain block or arc of Karlstrom & the CD-ROM Working Group 2002; Tyson et al. 2002, respectively) and the Archean Wyoming province (rifted continental margin) (see Chamberlain 1998 for a recent summary). This collisional orogeny is commonly interpreted as the initial accretion event along the southern margin of the Wyoming province (however, see Sims 1995; Bauer & Zeman 1997; Hill & Bickford 2001; Chamberlain et al. 2002; Sims & Stein 2003 for comments on a possible earlier collisional history) and is thus considered the harbinger of the eventual addition of approximately 1200 km of juvenile Palaeoproterozoic crust to the Wyoming craton (Condie 1982; Nelson & DePaolo 1985; Reed et al. 1987; Karlstrom & Bowring 1988; Bowring & Karlstrom 1990; Van Schmus et al. 1993).

Based on petrological, structural and geochronological studies (e.g. Condie & Shadel 1984; Karlstrom & Houston 1984; Reed et al. 1987), previous workers have suggested a tectonic model in which an inferred early S-dipping subduction zone accommodated northward migration (in present geographic co-ordinates) of an oceanic arc (i.e. Colorado province rocks). The Wyoming province was thus drawn into the subduction zone during collision (Houston 1993, fig. 14F). Subsequent uplift and erosion has exposed rocks that occupied deep crustal levels in the Palaeoproterozoic (Chamberlain et al. 1993).

The Laramie Peak shear system is located in the Laramie Mountains approximately 50 km north of the inferred trace of the Cheyenne belt, i.e. the suture zone along which the Proterozoic Colorado province and Archean Wyoming province are juxtaposed (Duebendorfer & Houston 1987). Chamberlain et al. (1993) originally coined the name ‘Laramie Peak shear zone’, although portions of the shear zone (system) had been recognized by previous workers (Condie 1969; Segerstrom et al. 1977; Langstaff 1984; Snyder et al. 1995). Chamberlain et al. (1993) delineated the Laramie Peak shear zone as consisting chiefly of the Garrett–Fletcher Park shear zone but also noted a splay shear zone that they called the Cottonwood Park shear zone (Fig. 1) (Chamberlain et al. 1993, fig. 2). These zones of concentrated ductile
Fig. 1. Geological index map of Precambrian rocks and selected Phanerozoic structural features in SE Wyoming (base derived from Love & Christiansen 1985; Snyder et al. 1995). Inset map was derived from Hoffman (1989, fig. 13). Abbreviations: fragmentary supracrustal sequences (metasedimentary and metavolcanic rocks) – ER, Elmers Rock; BC, Brandel Creek; IM, Johnson Mountain; E, Esterbrook; SM, Sellers Mountain; topographic features – RH, Richleau Hills; CH, Cooney Hills; H, Hartville uplift; CM, Casper Mountain; NMB, Northern Medicine Bow Mountains; MB, Medicine Bow Mountains; SM, Sierra Madre; EM, Elk Mountain; CM, Coad Mountain; PM, Pennock Mountain; R, Rawlins uplift.
Fig. 2. Structural geological map of a part of the Laramie Peak shear system, central Laramie Mountains, Wyoming. See Figure 3 for an explanation of rock units and symbols. NLR, North Laramie River. The map area of Figure 3 is delineated. The area described by Resor et al. (1996) is delineated, which includes the mafic dyke (Xad) from which syntectonic titanite was extracted and radiometrically dated using U–Pb techniques. Structural domains (subareas 1–4) are also delineated. Base derived from the South Mountain and Fletcher Park, Wyoming, US Geological Survey 7.5-minute quadrangles. Geology mapped by P.G. Resor and A.W. Snoke (1993–1998, non-inclusive). Some geological data have been adapted from Snyder et al. (1995).
deformation are inferred to be of Palaeoproterozoic age, although only the crystal-plastic deformation associated with the Fletcher Park shear zone, as exposed in Fletcher Park (Fig. 2), has been directly dated with radiometric age techniques (c. 1.76 Ga, Resor et al. 1996).

In this paper we expand the overall width of the belt of ductile deformation associated with the Laramie Peak shear zone by including another NE–SW-striking zone of concentrated shear strain, the North Laramie River high-strain zone, that occurs SE of the Fletcher Park shear zone (Fig. 2). Recently, Allard (2003) has recognized additional evidence of heterogeneous, high strain deformation to the south and east of our map area, which he correlates with the same deformational phase as manifested in the Laramie Peak shear zone of Chamberlain et al. (1993). Furthermore, fabric elements (NE–SW-striking penetrative foliation and SW-trending elongation lineation) characteristic of the shear zone deformation also occur between the delineated zones of concentrated shear strain as well as sporadically SE of these zones, suggesting a broad, albeit heterogeneous, zone of ductile deformation associated with the LPSS. Given this significant expansion in distribution of high strain rocks, as well as to avoid future confusion in nomenclature, we have designated all these zones of high strain as the ‘Laramie Peak shear system (LPSS)’ in this paper. Within this shear system individual shear zones are given local names as delineated in Figures 1 and 2.

The LPSS is a significant structural, metamorphic and geochronological boundary. The rocks within this broadly defined zone of high strain have experienced upper amphibolite facies dynamic metamorphism and have been locally converted to well-foliated and lineated mylonitic gneisses (typically an L–S tectonite but commonly L > S). Deformation was at a relatively high temperature (>450 °C), and the mylonitic rocks exhibit widespread evidence of recovery and recrystallization.

Rocks to the north of the Garrett–Fletcher Park shear zone are largely undeformed. These rocks have been referred to as the Laramie batholith by Condie (1969) or Laramie Peak block by Patel et al. (1999) (Fig. 1). Narrow zones (typically <1 m wide) of concentrated strain are locally present and perhaps are also related to the shear strain deformation associated with the LPSS. Metamorphic grade is low-pressure amphibolite facies with pressure–temperature (P–T) estimated at c. 2.5 kb and c. 575 °C (Patel et al. 1999). The metamorphism north of the Garrett–Fletcher Park shear zone is interpreted as Archean, based on U–Pb apatite cooling ages that are >2.1 Ga (Chamberlain et al. 1993; Patel et al. 1999; Chamberlain & Bowring 2000).

Rocks to the south of the Garrett–Fletcher Park shear zone are variably deformed and have been referred to as the ‘central metamorphic complex’ (Condie 1969) or the Palmer Canyon block (Patel et al. 1999). Rocks in this block include granitic gneiss–gneissic granite (chiefly Archean in crystallization age but also include subordinate, pre- to syn-tectonic Palaeoproterozoic gneissic rocks: Snyder et al. 1995), Archean migmatitic banded gneiss, Archean supracrustal rocks, ultramafic rocks, and Archean(?) and Palaeoproterozoic mafic dykes. Locally these rocks are strongly deformed and partitioned by anastomosing zones of high strain. Metamorphic grade is upper amphibolite with P–T estimated at c. 7.5 kb and c. 625 °C (Patel et al. 1999). Apatite cooling ages within and south of the Garrett–Fletcher Park shear zone are less than 1.78 Ga (Chamberlain et al. 1993; Patel et al. 1999; Chamberlain & Bowring 2000). In this light, the LPSS is a zone of Palaeoproterozoic high strain that defines the northern margin of the Palmer Canyon block. Consequently, the LPSS is the ‘deformation front’ for Palaeoproterozoic strain in the Archean Wyoming province.

**Rock units**

Although Archean granitic rocks constitute the bulk of the rock types exposed in the central Laramie Mountains, a diverse variety of other rock types provide an unusual spectrum of lithological variation including: an Archean supracrustal sequence, ultramafic rocks, Palaeoproterozoic (pretectonic) granitic rocks, and Archean(?) and Palaeoproterozoic mafic dykes (Fig. 3). Geochronometric data indicate that these rocks were deposited or intruded from the Middle Archean(?) through to the Palaeoproterozoic.

**Granitic rocks**

*Migmatitic banded gneiss.* The migmatitic banded gneiss is considered the oldest rock unit in the area (Johnson & Hills 1976; Snyder et al. 1995) and is most extensively exposed in our study area south of the Fletcher Park shear zone (Figs 2 and 3). Johnson & Hills (1976) reported Rb–Sr whole-rock isochrons of 2759 ± 152 Ma for grey granitic gneiss and 2776 ± 35 Ma for leucogranite from a similar unit in the northern Laramie Mountains. The migmatitic banded gneiss is characterized by compositional banding defined chiefly by variations in biotite content. Foliation is parallel to
composition banding, and mineral lineation is uncommon in these rocks. The compositional banding is commonly folded and cut by leucogranite layers. These layers are interpreted as melt and/or magma derived from in situ anatexis of the enclosing gneiss during high-temperature, Late Archean metamorphism (Allard 2003). The modal composition of the leucocratic layers indicates a minimum melt composition consisting of about equal parts plagioclase feldspar, alkali feldspar and quartz; other phases only account for <1% of the modal mineral content. The gneiss, exclusive of the leucocratic layers, is tonalitic with plagioclase feldspar typically accounting for approximately 60% of the mineral content of the rock. The remaining mineral phases are approximately 30% quartz, 2% alkali feldspar and 8% biotite. The migmatitic banded gneiss is intruded by variably deformed gneissic granite (Late Archean and/or Palaeoproterozoic: Allard 2003), and the migmatitic banded gneiss commonly occurs as enclaves (<1 to >100 m in longest dimension) within gneissic granite.

Undeformed–weakly deformed granite. North of the Garret–Fletcher Park shear zone (i.e. Laramie Peak block of Patel et al. 1999) the predominant granitic rock is an undeformed–weakly deformed granite. The most common rock type is a megacrystic, two-feldspar biotite monzogranite previously referred to as the ‘Laramie granite’ by Johnson & Hills (1976). These authors reported a Rb–Sr whole-rock isochron age of 2567 ± 25 Ma for the granite (Johnson & Hills 1976). Condie (1969) mapped a vast area as the ‘Laramie batholith’ that included the granitic rocks north of the Garret–Fletcher Park shear zone as exposed in our map area. These granitic rocks are commonly intruded by mafic dykes (either Archean?) and/or Palaeoproterozoic in age) and scarce ultramafic bodies (e.g. Elk Park peridotite of Snyder et al. 1995), and locally contain supracrustal enclaves (wallrock xenoliths).

Varially deformed gneissic granite. The variably deformed gneissic granite is the most heterogeneous unit in the map area. The gneissic granite occurs south of the Fletcher Park shear zone but north of the migmatitic banded gneiss (Figs 2 and 3). Thus, it is part of the ‘central metamorphic complex’ of Condie (1969) or ‘igneous and metamorphic complex of Laramie River’ of Snyder et al. (1995, 1998) or Palmer Canyon block of Patel et al. (1999). However, at least parts of this unit bear a marked affinity to granitic rocks north of the Fletcher Park shear zone, and thus these variably deformed granitic rocks may be deeper level (and deformed) equivalents of the Late Archean ‘Laramie granite’. On the other hand, some of the granitic rocks that comprise this unit are distinctly different from any granitic rock type in the ‘Laramie granite’ and may not even be Archean in age. Snyder et al. (1995) reported a U–Pb zircon age of 2051 ± 9 Ma for a distinctive granodiorite pluton in the east-central Laramie Mountains (SE of our map area). Thus, we view the ‘variably deformed gneissic granite’ as a composite unit consisting of deformed Archean granitic rocks but probably also including younger granitic rocks of possible Palaeoproterozoic crystallization age.

The granites of this unit appear to have an intrusive contact relationship with the migmatitic banded gneiss. The contact is highly irregular (e.g. Fig. 3) and enclaves of migmatitic banded gneiss occur in the variably deformed gneissic granite. The gneissic granite can be seen locally cutting across the fabrics of the migmatitic banded gneiss. Compositionally, the granitic rocks are chiefly monzogranitic containing approximately equal proportions of plagioclase feldspar and alkali feldspar in addition to 25–40% quartz; biotite is the typical varietal mineral and garnet is an occasional accessory phase. Virtually all the granitic rocks south of the Fletcher Park shear zone exhibit some evidence of solid-state deformation. The fabric in these granitic rocks is predominantly defined by elongated quartz and feldspar grains, and in this light it is not uncommon for the deformed granitic rocks to be L > S tectonites.

Mylonitic granitic gneiss. Mylonitic granitic gneiss is found in shear zones throughout the map area, but it was only mapped as a discrete unit in the two largest shear zones – Fletcher Park shear zone and North Laramie River high-strain zone (described in the section ‘Structural features and rock fabrics’). The mylonitic granitic gneiss was derived from the other granitic units during high-temperature (>450 °C) crystal-plastic strain. Some outcrops of mylonitic granitic gneiss are characterized by a distinctive pinstripe layering defined by mineralogical and grain-size variations (Fig. 4a). The mylonitic gneisses are typically L–S tectonites with strong mylonitic foliation (fluxion structure) that invariably contains a moderately steep-plunging elongation lineation. Recovery and recrystallization outpaced or kept up with deformation, thus, at the grain scale, evidence of crystal-plastic strain commonly is not well developed. However, at the outcrop scale there is abundant evidence that these rocks experienced substantial strain during their deformational
Geologic map of the South Mountain area, Albany County, Wyoming, U.S.A.

Geology mapped by P.G. Resor (1993-94)
EXPLANATION

SYMBOLS

- Lithological boundary, dashed where approximate
- Gradational boundary, dashed where approximate
- Fault, dotted where concealed; D on downthrown side and U on upthrown side, arrows indicate relative displacement along fault
- Strike and dip of foliation (left), vertical foliation (right)
- Trend and plunge of elongation lineation; combined with foliation
- Trend and plunge of hinge line of minor fold
- Mylonitic foliation associated with the Laramie Peak shear system

ROCK UNITS

Quaternary deposits
- Qal: Quaternary alluvium: Sand and gravel deposits along modern stream valleys

Precambrian rocks
- Xad: Diabasic amphibolite dyke (Palaeoproterozoic?). Black- to brown-weathering dykes from <1m to 30-m-thick. Dykes are variably metamorphosed and deformed. A few dykes contain plagioclase megacrysts
- Xap: Altered peridotite (Palaeoproterozoic?). Green actinolite–chlorite–magnetite rock
- Agr: Granite (Archean to Palaeoproterozoic?). Pink to grey, medium- to coarse-grained granite with subordinate aplite, alaskite, and pegmatite. Commonly contains alkali-feldspar megacrysts; locally contains garnet. Unfoliated to weakly foliated with the exception of small-scale ductile shear zones
- Agn: Gneiss (Archean). Similar to Agr but generally foliated, sometimes mylonitic. Contains enclaves of Abgn ranging from cm to 100 m scale
- Amgn: Mylonitic gneiss (Archean protolith but Palaeoproterozoic crystal-plastic strain). Strongly foliated and linedate pink to grey mylonitic gneiss commonly with feldspar porphyroclasts. Pinstripe layering is distinctive but not ubiquitous
- Aq: Quartzite (Archean). Brown, massive quartzite
- Aa: Amphibolite (Archean). Black, massive or compositionally layered fine- to medium-grained amphibolite
- Abgn: Migmatitic banded gneiss (Archean). Grey and white to pink banded gneiss with leucosome segregations and dykelets. Compositional layering with a biotite foliation; lineation is generally absent. Compositional layering is commonly folded

Fig. 3. Geological map of the South Mountain area, Albany County, Wyoming, USA. NLR, North Laramie River. Base derived from the South Mountain, Wyoming, US Geological Survey 7.5-minute quadrangle. Adapted from Resor (1996).
history. Such evidence includes: (1) shear-sense indicators (especially feldspar porphyroclasts); (2) strong linear fabric, as well as pervasive foliation development (Fig. 4a); (3) relative grain size reduction compared to inferred protolith rock types; (4) highly elongate grains and grain aggregates; and (5) isoclinal folding of compositional layering and mylonitic foliation (Fig. 4b).

**Supracrustal rocks**

The largest, continuous exposure of supracrustal rocks in the map area occurs in the Brandel–Owen Creeks area (Figs 1 and 2; part of the Brandel Creek greenstone belt of Snyder et al. 1995, fig. 2) and is part of the Bluegrass Creek metamorphic suite (Snyder et al. 1995, 1998). This supracrustal sequence is dominated by amphibolite. Other rock types that are part of this supracrustal sequence include porphyroblastic pelitic schist, metamorphosed ultramafic rocks, quartzofeldspathic gneiss (probable felsic metavolcanic and/or wacke protolith), banded iron formation, gedrite schist, marble (scarce) and quartzite (scarce) (Snyder et al. 1995). These supracrustal rocks must have been deposited prior to c. 2.6 Ga as required by the granitic rocks that intrude them (Snyder et al. 1998). This conclusion is supported by U–Pb zircon ages of 2637 ± 10 and 2729 ± 62 Ma from felsic metavolcanic rocks of the metamorphic suite of Bluegrass Creek (Snyder et al. 1998).

Although porphyroblastic pelitic schist is not a common component of the supracrustal sequence, these rocks contain mineral assemblages that provide the best P–T estimates from the area. Patel et al. (1999) demonstrated the following key conclusions derived from P–T data collected on metapelitic schists of the Bluegrass Creek metamorphic suite (of Snyder et al. 1998): (1) the metapelitic rocks exposed in the Owen Creek area reached \( P \approx 7.5 \text{ kb} \) (above the GRAIL reaction; i.e. subassemblage: garnet–rutile–aluminosilicate–ilmenite–quartz); (2) the metapelitic rocks are not migmatitic, so that the peak metamorphic temperature must have been below the water-saturated melting curve; and (3) textural evidence of decompression (from >7 to <3 kb) and retrograde re-equilibration is widespread in these schists.

**Ultramafic rocks**

Scattered bodies of ultramafic rocks occur throughout the central Laramie Mountains...
Diabasic amphibolite dykes are widespread throughout the map area (Figs 2 and 3) and constitute the second most abundant rock type beyond the granitic rocks of the area. These mafic dykes are interpreted to be chiefly members of the c. 2.01 Ga Kennedy dyke swarm (Graff et al. 1982; Cox et al. 2000). The diabasic amphibolite dykes form distinctive low, dark-coloured ridges that can be commonly tracked across the landscape. The dykes vary from <1 m to c. 30 m thick. Dyke lengths are also quite variable from less than 100 m to more than 2 km. In Figure 3, we have shown mafic dykes that were actually mapped out and not inferred from scattered outcrops. Thus, our distribution of mafic dykes represents a minimum of mafic dykes exposed in the map area. We estimate that mafic dykes probably form at least 15% of the bedrock geology in our map area. Most of the mafic dykes strike roughly NE-ward and dip steeply to the south. Two notable exceptions are a thick dyke found south of Menter Draw (Fig. 3; sections 11 and 14, T. 25N R. 72W) and the dyke exposed in Fletcher Park (see Fig. 2 for exact locality), which was the subject of a geochronological study (Resor et al. 1996). These dykes strike northerly.

The diabasic amphibolite dykes were originally pyroxene-bearing diabase. This conclusion is demonstrable in the field as well as under the microscope in that relict pyroxene-bearing diabase occurs sporadically in several of the thicker dykes, including the mafic dyke exposed in Fletcher Park (Fig. 2) (Resor et al. 1996). With progressive metamorphism and deformation, the mafic dykes transform from (1) igneous-textured, pyroxene-bearing diabase to (2) massive, relict igneous-textured amphibolite to (3) lineated and foliated amphibolite (commonly an L > S tectonite) and finally to (4) medium-grained, sometimes banded, amphibolite. This mineralogical and textural transformation resembles the classic metamorphic transformation of dolerite to hornblende schist (amphibolite) originally described by Teall (1885) for Scourie dykes affected by Laxfordian shear zones in NW Scotland.

Diabasic amphibolite dykes that are penetratively deformed and are now either amphibolite or hornblende schist and range from virtually pure L tectonite to L–S tectonite with gradations between these end members (Fig. 5a–c). Thus, the strain regime appears to vary from plane strain (k = 1) to pure shear of the constrictional type (k ≫ 1; Flinn 1994). An interesting aspect of some of the deformed diabasic amphibolite dykes is that the dyke contact with wallrock has locally acted as a strain guide with the strongest deformation localized along dyke walls. Thus, the deformation in the dykes is clearly solid state rather than igneous flow.

Several diabasic amphibolite dykes in the study area contain granitic wallrock breccia (Fig. 5d). These breccias can serve as relative strain indicators. In relatively undeformed dykes, the breccia consists of roughly equidimensional, sub-rounded, granitic clasts within a matrix of mafic dyke rock. In more highly deformed dykes, the clasts are stretched into prolate–triaxial ellipsoids with their long axes parallel to the mineral lineation in the surrounding schistose amphibolitic matrix (e.g. see Bauer et al. 1996, figs 28–30).

Structural features and rock fabrics

The structural chronology of the Precambrian rocks in our map area is most easily understood if the structural features and fabrics are ordered relative to the development of the LPSS. Thus, we view all deformation that predates the development of the LPSS as ‘pre-shear-zone’ deformation. This regional deformation is probably Late Archean in age in our map area, although R.L. Bauer and coworkers (Bauer & Zeman 1997; Pratt et al. 1999; Curtis & Bauer 1999, 2000; Tomlin & Bauer 2000) recognized an important Palaeoproterozoic deformation along the eastern margin of the central Laramie Mountains that they related to the Trans-Hudson orogeny. This deformational phase is poorly dated. However, the relative chronology as
Fig. 5. Montage of deformational features and characteristics of members of the c. 2.01 Ga Kennedy dyke swarm. (a) Moderately steep-plunging hornblende lineation in diabasic amphibolite dyke. (b) Transition from strong linear fabric in diabasic amphibolite dyke rock to relatively undeformed pyroxene-bearing diabase. (c) Narrow diabasic amphibolite dyke distended in foliation of variably deformed granitic gneiss. (d) Xenoliths of granitic rock in diabasic amphibolite dyke. See the text for discussion of these features.
deciphered by Bauer and coworkers indicates that this Palaeoproterozoic deformation predated shear-zone deformation in the LPSS that developed within the time interval of the Medicine Bow orogeny as demonstrated U–Pb radiometric dating (Resor et al. 1996). Thus, in a regional sense the shear-zone deformation (and Medicine Bow orogeny) is D3 (chiefly manifested as S2 and L2 of the LPSS), the earlier Palaeoproterozoic deformation is D2 (Tomlin & Bauer 2000) and the Archean deformation is D1. Recently, the pre-LPSS regional Palaeoproterozoic deformation has been referred to as the ‘Black Hills orogeny’ by Chamberlain et al. (2002) and Allard (2003), and correlated with Palaeoproterozoic deformation in the Black Hills of South Dakota (Dahl et al. 1999).

Superposed on the penetrative fabric formed during the shear-zone deformation are several fold phases broadly referred to as ‘post-shear-system deformation’. All of these deformational phases occurred between the late Palaeoproterozoic and early Mesoproterozoic (c. 1.4 Ga). Subsequently, much younger brittle deformation associated with the late Cretaceous–early Tertiary Laramide orogeny affected parts of our map area (Figs 2 and 3). The Archean–Mesoproterozoic events in the central Laramie Mountains, Wyoming are summarized in Table 1 and discussed in more detail in the following sections.

### Pre-shear-zone structural features and rock fabrics

The key to understanding the pre-shear-zone deformational history is a thorough understanding of the structural development of the migmatitic banded gneiss and its relationship to the younger, Palaeoproterozoic (c. 2.01 Ga) diabase amphibolite dyke swarm. Because intrusion of the dyke swarm predates the c. 1.76 Ga penetrative deformation characteristic of the LPSS (Resor et al. 1996), the structural style of the pre-LPSS deformation can be deduced in outcrops of deformed migmatitic banded gneiss intruded by non-deformed or weakly deformed diabasic amphibolite dykes. Such a situation is exhibited in Figure 6, which shows a moderately dipping, diabasic amphibolite dyke with a chilled margin cutting across both foliation and folds in the migmatitic banded gneiss. Although not apparent in the photograph, but noted on the sketch, a younger shear-zone deformational fabric developed along the lower contacts of the dyke and overprinted both dyke rock and its gneissic wallrocks. This younger, superposed fabric completely transposes the original gneissic fabric within about 0.5 m of the dyke–wallrock contact.

The rock fabrics that predated the intrusion of the amphibolitic mafic dyke swarm and

### Table 1. Summary of Archean–Mesoproterozoic events in the central Laramie Mountains, Wyoming, USA

<table>
<thead>
<tr>
<th>Event Description</th>
<th>Dating</th>
</tr>
</thead>
<tbody>
<tr>
<td>Development of tonalitic orthogneisses (protolith of the migmatitic banded gneiss) – Middle Archean (?)</td>
<td>c. 2.01 Ga</td>
</tr>
<tr>
<td>Intrusion of early mafic dykes (now amphibolite enclaves in migmatitic banded gneiss) – Archean</td>
<td>c. 1.76 Ga</td>
</tr>
<tr>
<td>Pre-LPSS deformation (D1), metamorphism, anatexis and intrusion of leucogranite dykes – Late Archean (?)</td>
<td>c. 1.76 Ga</td>
</tr>
<tr>
<td>Tectonic juxtaposition of migmatitic banded gneiss and Archean supracrustal rocks of the Brandel Creek greenstone belt – Late Archean (?)</td>
<td>c. 1.43 Ga</td>
</tr>
<tr>
<td>Emplacement of granitic rocks into the Laramie Peak and Palmer Canyon blocks – Late Archean – Palaeoproterozoic (including the Boy Scout Camp Granodiorite (2051 ± 9 Ma, Snyder et al. 1995))</td>
<td>c. 1.76 Ga</td>
</tr>
<tr>
<td>Intrusion of members of the c. 2.01 Ga Kennedy dyke swarm along the rifted margin of the Archean Wyoming province (Cox et al. 2000)</td>
<td>c. 1.8 Ga</td>
</tr>
<tr>
<td>Black Hills (Trans-Hudson) orogeny (D2, Palaeoproterozoic (c. 1.8 Ga) as used by Chamberlain et al. 2002; also see Tomlin &amp; Bauer 2000 and Allard 2003 for relative structural chronology)</td>
<td>c. 1.8 Ga</td>
</tr>
<tr>
<td>Deformation and metamorphism related to early phase of the Medicine Bow orogeny (D3, 1.78–1.76 Ga)</td>
<td>c. 1.76 Ga</td>
</tr>
<tr>
<td>LPSS deformation (c. 1.76 Ga (Resor et al. 1996), late phase of the Medicine Bow orogeny (D3)) – broadly transpressive in the regional geological setting but local transtensional effects (Chamberlain 1998); e.g. emplacement of the c. 1.76 Ga Horse Creek Anorthosite Complex in the southern Laramie Mountains – see Scoates &amp; Chamberlain (1997), high-temperature mylonitization (S1) and uplift and decompression (Patel et al. 1999) of the Palmer Canyon block.</td>
<td>c. 1.76 Ga</td>
</tr>
<tr>
<td>Post-LPSS fold events. These deformational events are part of D4 and/or D3 regional deformations related to either a late Palaeoproterozoic regional deformation at c. 1.722 Ma (Allard 2003) or intrusive emplacement of the Mesoproterozoic (c. 1.43 Ga) Laramie Anorthosite Complex (Tomlin 2001)</td>
<td>c. 1.76 Ga</td>
</tr>
</tbody>
</table>

LPSS, Laramie Peak shear system.
development of the shear-zone deformation are only found in the migmatitic banded gneiss. These pre-LPSS deformation rock fabrics include the development of compositional layering with a parallel biotite foliation and abundant outcrop-scale folding of this layering and the foliation. Compositional layering is defined by alternating biotite-rich and biotite-poor layers. Leucogranite layers may lie subparallel to compositional layering or cross-cut it in intrusive dykelet style. The ubiquitous presence of leucogranite layers of variable orientation gives this rock unit its overall migmatitic appearance. Figure 7a-1 is an equal-area, lower-hemisphere projection that summarizes S-surface data (compositional layering and foliation) from the migmatitic banded gneiss as exposed in our map area. The data broadly scatter across the stereogram, and there is no simple pattern to the distribution of the data, suggesting a polyphase deformational history. In Figure 7a-2 a subset of the orientation data in Figure 7a-1 is plotted. These data, exclusively from the Owen Creek area, yield a definitive girdle with a fold axis determined as 21°/N48E. These data suggest NW–SE shortening, roughly perpendicular to the shear-zone foliation and presumably a manifestation of the Medicine Bow orogeny.

Shear-zone structural features and rock fabrics

Distribution and Geometry. The shear-zone deformation is heterogeneously distributed across our map area. As previously noted, the LPSS consists of several distinct segments: Garrett–Fletcher Park shear zone, Cottonwood Park shear zone, North Laramie River high-strain zone, as well as several high-strain zones recognized by Allard (2003). The Fletcher Park shear zone (Figs 1–3 and 7b) exhibits the most concentrated shear-strain deformation in the map area and forms a distinct, Palaeoproterozoic deformation front that separates generally non-deformed–weakly deformed granitic rocks (part of the Laramie batholith of Condie 1969, or Laramie Peak block of Patel et al. 1999) from the variably deformed igneous and metapelitic rocks of the ‘central metamorphic complex’ of Condie (1969) or Palmer Canyon block of Patel et al. (1999). Scattered evidence of the shear-zone deformation does occur north of the Fletcher Park shear zone (Fig. 7c), and the Cottonwood Park shear zone (Figs 1 and 7d) is the most obvious example of concentrated, penetrative shear strain within the Laramie batholith or Laramie Peak block. The Fletcher Park shear zone is a steeply dipping, NE-striking, 300–500 m-wide zone of mylonitic gneisses. These pre-LPSS deformation rock fabrics include the development of compositional layering with a parallel biotite foliation and abundant outcrop-scale folding of this layering and the foliation. Compositional layering is defined by alternating biotite-rich and biotite-poor layers. Leucogranite layers may lie subparallel to compositional layering or cross-cut it in intrusive dykelet style. The ubiquitous presence of leucogranite layers of variable orientation gives this rock unit its overall migmatitic appearance. Figure 7a-1 is an equal-area, lower-hemisphere projection that summarizes S-surface data (compositional layering and foliation) from the migmatitic banded gneiss as exposed in our map area. The data broadly scatter across the stereogram, and there is no simple pattern to the distribution of the data, suggesting a polyphase deformational history. In Figure 7a-2 a subset of the orientation data in Figure 7a-1 is plotted. These data, exclusively from the Owen Creek area, yield a definitive girdle with a fold axis determined as 21°/N48E. These data suggest NW–SE shortening, roughly perpendicular to the shear-zone foliation and presumably a manifestation of the Medicine Bow orogeny.
Fig. 7. Montage of equal-area, lower-hemisphere projections of structural data from the Laramie Peak shear system and environs. (a) Foliation in Archaean migmatitic banded gneiss (1) and (2). (b) Foliation and lineation in Fletcher Park shear zone. (c) Foliation and lineation in granitic rocks north of the Fletcher Park shear zone (but exclusive of the Cottonwood Park shear zone, see (d) below). (d) Foliation and lineation in the Cottonwood Park shear zone. (e) Foliation and lineation in variably deformed gneissic granite (includes rocks of North Laramie River high strain zone). (f) Foliation and lineation in Brandel Creek supracrustal sequence and associated granitic rocks. (g) Foliation and lineation in diabasic amphibolite dykes (throughout map area).
range in the late Cretaceous–early Tertiary. Interestingly, all three mapped brittle faults in Figure 2 exhibit a prominent strike-slip component during their movement history.

The Fletcher Park shear zone has gradational boundaries on both its north and south margins. These gradations can be rather sharp, especially along the northern margin, where the transition

Fig. 7. Continued.
from weakly deformed granitic rocks (immediately north of the shear zone) to mylonitic gneisses occurs over about 10 m. North of the Fletcher Park shear zone, narrow localized shear zones parallel the Fletcher Park zone, but decrease in size and abundance to the north. The largest shear zone, the Cottonwood Park shear zone (Chamberlain et al. 1993), is delimited by an approximately 50 m-wide anastomosing zone of mylonitic granitic gneiss, located about 0.5 km north of the Fletcher Park shear zone (Fig. 1). North of the Cottonwood Park shear zone, localized shear zones are less common and narrower (<5 m wide) and beyond approximately 1.5 km north of the Fletcher Park shear zone the granitic rocks of the Laramie batholith of Condie (1969) are virtually undeformed.

The LPSS deformation is prevalent south of the Fletcher Park shear zone. In this region (the Palmer Canyon block of Patel et al. 1999) shear-zone-parallel foliation and lineation are variably developed with localized high-strain zones separated by weakly–moderately deformed blocks (Fig. 7e). Most of these high-strain zones are relatively narrow (<10 m); however, the North Laramie River high-strain zone is a 200–400 m-wide, anastomosing zone of well-foliated to mylonitic gneisses that parallels the Fletcher Park shear zone, approximately 1.5 km to the south. The North Laramie River high-strain zone is in part localized along the contact between granitic rocks of the Laramie River complex of Snyder et al. (1995) and supracrustal rocks of the Brandel Creek greenstone belt. This strain localization is apparently controlled by the rheology of the rock types on either side of the contact. The Brandel Creek greenstone belt is dominated by amphibolitic rocks, whereas the granitic rocks are considerably more quartz-rich and thus weaker during amphibolite facies crystal-plastic deformation.

**Penetrative fabric elements.** The orientation of penetrative fabrics within individual structural domains and rock units is illustrated in a series of stereographic projections (Fig. 7). The map area was subdivided into four structural domains (subareas 1–4) (Fig. 2): (1) map area north of the Fletcher Park shear zone (Fig. 7c); (2) Fletcher Park shear zone (Fig. 7b); (3) variably deformed granitic rocks including the North Laramie River high-strain zone (Fig. 7e); and (4) supracrustal and associated granitic rocks of the Brandel Creek greenstone belt (Fig. 7f). Furthermore, the following units or structural features were also analysed individually: migmatitic banded gneiss (Fig. 7a-1 & a-2), diabasic amphibolite dykes (Fig. 7g), Cottonwood Park shear zone (Fig. 7d) and late folds in the Fletcher Park shear zone (Murphy Canyon locality, Fig. 8). Also, a synoptic summary of all linear data from the LPSS is presented in Figure 9.

LPSS fabrics can be defined based on the foliation and lineation in the Fletcher Park shear zone (subarea 2, Fig. 2 and Fig. 7b). The shear-zone fabric consists of a foliation striking 55° ± 20° to the NE, dipping 75° ± 15° to the south (Fig. 8). TheNorthLaramieRiverhigh-strainzoneisdelineatedbyanapproximately50m-wideanastomosingzoneofmyloniticgraniticgneiss,locatedabout0.5kmnorthoftheFletcherParkshearzone(Fig.1).NorthoftheCottonwoodParkshearzone,localizedshearzonesarelesscommonandnarrower(<5mwide)andbeyondapproximately1.5kmnorthoftheFletcherParkshearzonethegraniticrocksoftheLaramiebatholithofCondie(1969)arevirtuallyundeformed.

TheLPSSdeformationisprevalentsouthoftheFletcherParkshearzone.Inthisregion(thes PalmerCanyonblockofPateletal.1999)shear-zone-parallelfoliationandlineationarevariablydevelopedwithlocalizedhigh-strainzonestparatedbyweakly–moderatelydeformedblocks(Fig.7e).Mostofthesehigh-strainzonesarerelativelynarrow(<10m);however, the North Laramie River high-strain zone is a 200–400m-wide, anastomosing zone of well-foliated to mylonitic gneisses that parallels the Fletcher Park shear zone, approximately 1.5 km to the south. The North Laramie River high-strain zone is in part localized along the contact between granitic rocks of the Laramie River complex of Snyder et al. (1995) and supracrustal rocks of the Brandel Creek greenstone belt. This strain localization is apparently controlled by the rheology of the rock types on either side of the contact. The Brandel Creek greenstone belt is dominated by amphibolitic rocks, whereas the granitic rocks are considerably more quartz-rich and thus weaker during amphibolite facies crystal-plastic deformation.

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either the SE or NW. This foliation typically contains a lineation trending $220^\circ \pm 25^\circ$ and plunging $60^\circ \pm 15^\circ$ to the SW. Scatter in the stereogram is interpreted as random variation about the mean but may be due in part to Laramide deformation and/or minor late folding.

Foliations and lineations in granitic rocks north of the Fletcher Park shear zone (subarea 1, Fig. 2 and Fig. 7c) show similar fabric orientations, foliation striking $65^\circ \pm 15^\circ$, dipping $70^\circ \pm 20^\circ$ SE or NW, and a lineation trending c. $210^\circ$ and plunging to the SW at approximately $55^\circ$. We interpret these fabrics as the northward-decreasing effects of the shear-zone deformation.

The Cottonwood Park shear zone (Fig. 7d) also shows the same basic fabric elements, and the orientation of these fabrics fall within the distribution of measurements for the Fletcher Park shear zone.

South of the Fletcher Park shear zone (subareas 3 and 4, Fig. 2) there is evidence for both shear-zone fabrics and presumably Archean fabrics. Gneissic granites (Fig. 7e) primarily show evidence of a shear-zone-parallel foliation (strike $55^\circ \pm 30^\circ$, dip $70^\circ \pm 20^\circ$) and lineation (trend $230^\circ \pm 30^\circ$, plunge $50^\circ \pm 20^\circ$); however, there is greater scatter. Migmatic banded gneiss (Fig. 7a-1) shows a much broader scatter suggestive of large-scale folding. In fact, when data from the vicinity of Owen Creek are isolated they exhibit a well-defined girdle consistent with a gently NE-plunging fold hinge (Fig. 7a-2). A biotite lineation was measured in this vicinity, parallel to the interpreted hinge line. The Brandel Creek supracrustals themselves (Fig. 7f) describe a broad girdle consistent with a SW-ward plunging hinge line parallel to mineral lineations in both supracrustal and granitic rocks, and to several late small-scale folds.

The diabasic amphibolite dykes are particularly illustrative of the predominance of the shear-zone fabric (i.e. Palaeoproterozoic). Figure 7g, an equal-area, lower-hemisphere plot of all data collected from the diabasic amphibolite dykes, shows a foliation and lineation distribution that is once again indistinguishable from the Fletcher Park shear zone (Fig. 7b). In fact, a synoptic plot of all the lineation data collected in the study area (Fig. 9) shows a single cluster of data with the one outlier from near Owen Creek that was previously discussed. These data lead to the interpretation that there has been only one penetrative Palaeoproterozoic deformation in the study area (Fig. 2). This stands in contrast to the work of Bauer and coworkers (Bauer & Zeman 1997; Pratt et al. 1999; Curtis & Bauer 1999, 2000; Tomlin & Bauer 2000) to the south and east where they have found evidence for two penetrative Palaeoproterozoic deformation events.

Microstructures. Microstructures in the rocks of the LPSS provide abundant evidence of deformation under high-temperature conditions and indicate that the mylonitic rocks associated with the shear system are best referred to as ‘mylonitic gneisses’ (Snake & Tullis 1998). Feldspar grains commonly exhibit core–mantle structure with large (5–30 mm) grains surrounded by small (0.25 mm) grains, suggesting the importance of subgrain rotation during dynamic recrystallization of these rocks. The importance of grain-boundary migration during deformation of the mylonitic gneisses is suggested by the irregular grain boundaries of the feldspar porphyroclasts. Myrmekitic intergrowths (quartz–sodic plagioclase symplectite) occur in the rims of some of the large alkalifeldspar grains. Simpson & Wintsch (1989) argue that deformation-driven replacement of alkali feldspar by an intergrowth of oligoclase and quartz (myrmekite) is a product of high-temperature ($>450^\circ$ C) deformation in the presence of an aqueous fluid. Mantled feldspar porphyroclasts (Passchier & Trouw 1996) are common in the mylonitic granitic gneisses, although asymmetric tails (e.g. $\alpha$- or $\beta$-type porphyroclasts) are scarce compared to $\phi$- or $\theta$-type porphyroclasts. The abundance of symmetrical winged porphyroclasts (i.e. $\phi$-type) compared to asymmetric porphyroclasts argues for a general shear (non-coaxial) condition during the high-temperature deformation manifested in the LPSS.

In mylonitic granitic gneisses of the LPSS, quartz commonly occurs in elongated aggregates of irregularly shaped grains with interlocking grain boundaries suggesting widespread grain-boundary migration. However, in weakly deformed granitic rocks of the LPSS, quartz grains are recrystallized but contain subgrains and exhibit undulose extinction. Grain size is highly variable from small (c. 0.25 mm) neoblasts with relatively little internal strain to large grains (c. 2 mm) that contain abundant subgrains. Grain boundaries of the large quartz grains are highly irregular.

In diabasic amphibolite within high-strain zones of the LPSS, hornblende grains are commonly subidioblastic with a grain-shape orientation; grain boundaries between hornblende grains, as well as other phases, are straight and indicate textural equilibrium. There is no optical evidence of subgrain development within the hornblende grains. Plagioclase becomes increasingly recrystallized with
increased deformation in the amphibolitic mafic dykes. In highly strained amphibolitic schists, the original tabular, igneous plagioclase grains have been totally recrystallized to an aggregate of intermixed coarse and much finer grains. These petrographic characteristics indicate recrystallization at high temperatures. The reaction of titanomagnetite to titanite (sphene) (Resor et al. 1996) during dynamic recrystallization of the diabasic amphibolite dykes within the LPSS also indicates high-temperature deformation and recrystallization (Frost et al. 2000).

Shear-sense indicators throughout the map area yield a consistent south-side-up, sinistral sense of shear (i.e. reverse sense shear). Shear-sense indicators are most prevalent in the Fletcher Park shear zone, but can also be found in the Cottonwood Park shear zone, North Laramie River high-strain zone, and many smaller shear zones distributed throughout the map area. The most common type of shear-sense indicator is sigmoidal feldspar porphyroclasts. Asymmetric feldspar porphyroclasts are best developed in the marginal zones of the Fletcher Park shear zone where deformation-driven recrystallization is moderate. The asymmetry of recrystallized tails on feldspar porphyroclasts, as well as scarce asymmetric quartzose lenses, consistently indicate south-side-up sense of shear. Composite fabrics (S–C–C') are less common, but also indicate an up-lineation shear sense. Local asymmetric folds at a high angle to the lineation are also consistent with south-side-up shearing.

Evidence for a pure shear (flattening) component of the deformation includes the prevalence of shear-zone–post–directed folds and foliation even in apparently low strain areas (Fig. 2), the predominance of symmetric feldspar porphyroclasts (d-type porphyroclasts), L > S tectonites and the abundance of lineation–parallel folds. Although lineation–parallel folds are commonly interpreted to form at a high angle to lineation and then rotate into parallelism, such a process would probably form shear folds (Cobbold & Quinquis 1980; however, see Alsop & Holdsworth 1999 for an alternative interpretation). The lack of shear folds and abundance of lineation–parallel folds in the LPSS suggest that these folds may have originally formed subparallel to lineation (L3 in the regional sense). Grujic & Mancktelow (1995) demonstrated that viscosity contrasts between layers (and matrix) are fundamental in the generation of folds subparallel to the principal elongation direction X. However, their analogue experiments also indicated that a flattening component during deformation favoured the development of lineation–parallel folds.

**Timing.** The broadest constraints that can be placed on the timing of shear-zone deformation come from observations of fabric development in the diabasic amphibolite dykes. Shear-zone deformation clearly post-dates emplacement of the dykes at c. 2.01 Ga (Cox et al. 2000) and is coincident with high-grade, amphibolite facies metamorphism. Amphibolite facies conditions necessarily predate the uplift and cooling of the Palmer Canyon block to temperatures below approximately 450 °C at c. 1745 Ma, as recorded by U–Pb apatite cooling ages (Patel et al. 1999). Resor et al. (1996) argued that an U–Pb titanite age of c. 1763 Ma from a diabasic amphibolite dyke deformed in the Fletcher Park shear zone provides a more direct constraint on the shear-zone deformation based on structural arguments for syntectonic growth at or below the diffusion-based closure temperature. Cox et al. (2000) subsequently found a similar age for metamorphic titanite from a deformed dyke within the Palmer Canyon block (their sample DC-6). The earliest evidence for Palaeoproterozoic high-grade metamorphism is an U–Pb age for metamorphic zircon from the southern edge of the Palmer Canyon block of c. 1778 Ma (Harper 1997).

**Post-shear-system deformation in the Precambrian**

Bauer and coworkers (Pratt et al. 1999; Curtis & Bauer 1999, 2000; Tomlin & Bauer 2000) and Allard (2003) have recognized several fold phases that post-date the development of the LPSS. These fold phases are best developed south or east of our map area (Figs 2 and 3). These various fold phases are either related to a south or east of our map area (Figs 2 and 3). These various fold phases are either related to a poorly understood event at c. 1722 Ma that Allard (2003) correlated with the Dakotan orogeny of Chamberlain et al. (2002) or the emplacement of the Laramie Anorthosite Complex (Tomlin 2001). Pratt et al. (1999) recognized a series of upright, NE-trending folds that they referred to as the ‘George Creek fold event’. Tomlin (2001) concluded that the ‘George Creek fold event’ was related to the emplacement of the Mesoproterozoic (c. 1.43 Ga) Laramie Anorthosite Complex, and thus is unrelated to the LPSS or Medicine Bow orogeny. Tomlin & Bauer (2000) referred to this deformational event as regional D4 and concluded that deformational features related to Palaeoproterozoic D2 and D3 were reoriented by this younger Proterozoic deformation. Another post-LPSS fold event was called the ‘Open Fold Event’ by Pratt et al. (1999), but the age of this
deformational event is uncertain. Pratt et al. (1999) described this event as late oroclinal bending and considered it associated with the Medicine Bow orogeny. Allard (2003), however, argued that the ‘Open Fold Event’ is younger than c. 1722 Ma based on new geochronometric data and overprinting relationships, and thus post-dates the Medicine Bow orogeny.

**Evaluation of some tectonic models**

The LPSS is a 10 km-thick zone of heterogeneous general shear (non-coaxial) strain that defines the northern margin of the Palmer Canyon block of the central Laramie Mountains. This uplifted tectonic block, chiefly comprised of rocks of the Archean Wyoming province but extensively intruded by various Palaeoproterozoic rocks ranging in composition from ultramafic to granitic, was regenerated during the Palaeoproterozoic Medicine Bow orogeny (Chamberlain et al. 1993; Patel et al. 1999). The Palaeoproterozoic age (c. 1.76 Ga) of the Fletcher Park shear zone (Resor et al. 1996) suggests that the whole LPSS developed within the time interval of the Medicine Bow orogeny (1.78–1.74 Ga).

Several tectonic models have been devised to explain the uplift of basement rocks in a foreland setting during crustal contraction. Possible end-member models for the uplift of a basement block include: (1) tilted block uplift with a frontal listric fault zone that soles into a deep-crustal décollement; (2) en-masse ‘pop-up’ uplift bounded by oppositely vergent thrust (reverse) faults; and (3) crustal-scale duplex consisting of a stack of crystalline thrust sheets each bounded by a ductile fault (shear) zone (Fig. 10).

To date only models 1 and 2 have been evaluated in regard to the uplift of the Palmer Canyon block (Chamberlain et al. 1993; Patel et al. 1999), and we further evaluate these hypotheses in light of our structural analysis of the LPSS. In addition, we also comment on the implications of the crustal-scale duplex model and suggest that additional structural studies of the Palmer Canyon block are necessary to fully test this model.

Chamberlain et al. (1993) originally suggested a block-uplift model to explain the development of the LPSS in conjunction with an arc–continental margin collision along the Cheyenne belt during the Palaeoproterozoic (see Chamberlain et al. 1993, fig. 5). This model involved the uplift of the Palmer Canyon block along a crustal-scale reverse fault that recorded at least 10 km of differential, vertical displacement across the LPSS. Patel et al. (1999) further refined the block-uplift model by emphasizing that their P–T data indicated no significant variation across the Palmer Canyon block, and thus they argued against a rotational component for the hanging-wall block (i.e. model 1 above and see Fig. 10a). If the displacement is chiefly along a deep-rooted, frontal thrust fault, substantial rigid-body rotation of the hanging-wall block is required to achieve structural balance and to avoid ‘space problems’, and is well documented in late Cretaceous–Eocene, Laramide basement-involved uplifts of the Rocky Mountain foreland (e.g. Erslev 1986). To minimize potential rotation of the hanging-wall block, Patel et al. (1999) postulated a complementary back-thrust along the southern margin of the Palmer Canyon block (i.e. en-masse ‘pop-up’ uplift (model 2 and Fig. 10b) similar to a common thrust-belt structural feature discussed by Butler 1982, fig. 16b). Tangible evidence for the back-thrust interpretation is not available, because the back-thrust is inferred to occur in an area now intruded by the Mesoproterozoic Laramie Anorthosite Complex and Sherman batholith. Patel et al. (1999) also suggested that the uplift of the Palmer Canyon block occurred late in the Medicine Bow orogeny during a transpressive tectonic regime. The en-masse ‘pop-up’ uplift model is compatible with a crustal-scale shear zone (i.e. LPSS), a prominent metamorphic discontinuity between the hanging-wall and footwall blocks, the lack of apparent metamorphic variation across the uplifted high-grade hanging-wall block and the age of shear-zone deformation determined from the Fletcher Park shear zone (Resor et al. 1996). One ad hoc element of the ‘pop-up’ uplift model is the back-thrust inferred to bound the uplifted block along its southern margin (Patel et al. 1999).

A key observation that supports an en-masse uplift of the Palmer Canyon block is the apparent lack of metamorphic-grade variation across the block suggesting that significant hanging-wall rotation (i.e. a fundamental characteristic of model 1, Fig. 10a) did not occur despite significant displacement based on contrasting metamorphic grade across the frontal (northern) shear zone (i.e. Garrett–Fletcher Park shear zone). Unfortunately, detailed P–T determinations do not exist across the entire Palmer Canyon block. Furthermore, the Palaeoproterozoic deformational and P–T history of the Palmer Canyon block is complicated by the proximity of the Trans-Hudson Orogen as manifested in the eastern Laramie Mountains, as well as Hartville uplift (Sims 1995; Bauer & Zeman 1997; Pratt et al. 1999; Curtis & Bauer 2000; Tomlin & Bauer 2000). There is considerable controversy concerning the age and magnitude...
Fig. 10. Some tectonic models for the evolution of the Laramie Peak shear system. (a) Tilted basement block bounded by frontal listric thrust fault that soles into a deep-crustal décollement. (b) En-masse ‘pop-up’ uplift bounded by oppositely vergent thrust (reverse) faults. (c) Crustal-scale duplex consisting of a stack of crystalline thrust sheets each bounded by a ductile fault (shear) zone.
of the Trans-Hudson orogenic effects in the Laramie Mountains and environs (Edson 1995; Tomlin & Bauer 2000; Hill & Bickford 2001; Chamberlain et al. 2002; Allard 2003; Sims & Stein 2003). In the light of only scattered P–T determinations and potential overprinting of orogenic events in the Palmer Canyon block, the role of large-scale rotation of the Palmer Canyon block during the development of the LPSS has not been fully evaluated.

Although the en-masse ‘pop-up’ uplift model is supported by a diverse data set, another structural model that may also explain the geological relationships is a crustal-scale duplex consisting of imbricate slices of basement rocks bounded by anastomosing ductile shear zones rooted into a deep-crustal décollement zone. The heterogeneous nature of the shear strain associated with the LPSS raises the possibility that such heterogeneity could exist throughout the Palmer Canyon block. A ductile thrust stack model could explain the distribution of deformation across the LPSS and the high-grade block to the south; however, it fails to explain the consistent uniform grade of metamorphism to the south (although our previous comments about this observation also apply here). In the crustal-scale duplex model, the Fletcher Park shear zone marks the frontal thrust of a thick-skinned imbricate thrust belt (Fig. 10c). Thus, the Fletcher Park shear zone would be a major basement ramp rooted into a deep-crustal décollement zone. In the Laramie Mountains, the frontal reverse fault does mark a significant metamorphic discontinuity; however, metamorphic grade should also increase across the high-grade block in the hanging wall as additional ductile thrusts increase the cumulative displacement to the south. If the crustal-scale duplex model is applied to the uplift of whole Palmer Canyon, the distribution of the shear-zone deformation could be much more widespread than just the presently delineated LPSS. Although structural analysis is not available throughout the entire Palmer Canyon block, the geological mapping of the late George L. Snyder (e.g. Snyder et al. 1995, 1998) and structural studies by Allard (2003) suggest that the shear-zone deformation is restricted to the northern margin of the Palmer Canyon block (i.e. basically the LPSS as defined in this paper).

Given apparent problems with all simple uplift models, we favour a tectonic model for the uplift of the Palmer Canyon block and development of the LPSS by a combination of en-masse block uplift coupled with internal, heterogeneous general shear (non-coaxial) manifested by the development of high-strain zones and complex folding. The Garrett–Fletcher Park shear zone is interpreted as the frontal fault of the uplift (i.e. a major basement ramp) that rooted into a deep-crustal décollement zone that must have underlain the entire Palmer Canyon block during Palaeoproterozoic orogenesis. This décollement zone in turn would have rooted southward (in present geographic co-ordinates) into the evolving collisional zone that must have separated the Wyoming and Colorado provinces (now manifested by the Cheyenne belt, see Duebendorfer & Houston 1987). Internal deformation within the Palmer Canyon block would be manifested by heterogeneous general shear strain and folding. In such a model the ad hoc, S-directed back-thrust suggested by Patel et al. (1999) would not be needed to achieve overall uplift of the Palmer Canyon block, whereas widespread (albeit heterogeneous) Palaeoproterozoic general shear and folding is required. The detailed structural studies by Allard (2003) immediately to the south and east of our map area support such a complex, heterogeneous Palaeoproterozoic deformational history.

On the deformation of sialic crust during an arc–continental margin collision

The central Laramie Mountains provide an unusual opportunity to study how sialic crust responded during an arc–continental margin collision. A modern example of such a tectonic phenomena is the convergence of Australian continental lithosphere with the Banda arc (Snyder et al. 1996). In regard to the Laramie Mountains example, it is inferred, based on regional geological relationships and geochronometric studies, that the Archean Wyoming province was partially subducted beneath a N-facing (in present geographic co-ordinates) oceanic arc (i.e. rocks of the Colorado province). Because the dip of subduction was directed southward (again in present geographic co-ordinates), there was no synorogenic magmatism associated with the underthrusted continental margin (i.e. Archean Wyoming province). Deformation in the Wyoming province was then purely mechanical in nature and not associated with magmatic softening. The chief manifestation of this situation is heterogeneous Palaeoproterozoic strain distributed throughout the partially subducted Wyoming province (Palmer Canyon block). Locally this strain was intense, forming high-temperature, mylonitic gneisses that involved crystal-plastic deformation mechanisms (e.g. dislocation and diffusion creep). The requisite
thermal regime for these chiefly thermally activated deformation mechanisms was facilitated by tectonic thickening related to the emplacement of a Paleoproterozoic oceanic arc and related rocks structurally above the Wyoming province during the Medicine Bow orogeny (Chamberlain 1998).

A fundamental characteristic of the deformation of the Archean Wyoming province during the Medicine Bow orogeny is strain partitioning and localization (Snoke & Resor 1999). Strain is commonly localized along original compositional (i.e. rheological) boundaries. This strain localization varies from outcrop-scale, e.g. along the intrusive contact between an amphibolitic mafic dyke and its quartzofeldspathic host rock, to map-scale examples such as the NW margin of the Brandel Creek greenstone belt and the localization of the North Laramie River high-strain zone. We propose that these contacts may have acted as strength defects (Poirier 1980) affecting geometrical stress concentration and/or chemical and reaction softening (White et al. 1980) associated with localization of aqueous fluid flow. Progressive deformation and amphibolite facies metamorphism of original diabase dykes and the replacement of alkali feldspar by myrmekite demonstrate the importance of aqueous fluids in the deformation process. Developing high-strain zones may have further localized strain through grain-size reduction associated with dynamic recrystallization (Handy 1989) and structural softening (White et al. 1980) associated with the development of compositional layering in pinstripe mylonites and amphibolitic schists.

Strain localization in the Garrett–Fletcher Park shear zone is not clearly influenced by rheological contrasts, although this shear zone may have grown from smaller localized zones in a process similar to the one outlined by Christiansen & Pollard (1997). In a map-scale example, the high strain of the Garrett–Fletcher Park shear zone is the structural front for Paleoproterozoic regeneration of the Archean Wyoming province. Although strain associated with the LPSS occurs north of the Garrett–Fletcher shear zone, this shear zone is clearly a significant structural, metamorphic and geochronological boundary. This important boundary therefore possesses all the essential characteristics of a true orogenic front. Furthermore, this basic conclusion suggests that the Garrett–Fletcher Park shear zone may be an example of a pro-shear zone (Quinlan et al. 1993; Beaumont & Quinlan 1994; Beaumont et al. 1994) rooted into a deep-crustal detachment that developed during the Paleoproterozoic oceanic arc–continental margin collision inferred as the tectonic setting of the Medicine Bow orogeny (Chamberlain 1998).

In summary, we propose that strain partitioning and localization rather than broad-scale flowage at lower crustal depth are common processes in the absence of synorogenic magmatism. Deformation of continental crust in such cases is characterized by the development of distributed zones of general shear (non-coaxial), and some of these shear zones may transect the entire crust. A well-documented Phanerozoic example of a crustal-scale fault/shear zone that developed in absence of synorogenic magmatism is the Laramide-age Wind River thrust of the Wyoming Rocky Mountain foreland (Smithson et al. 1979; Sharry et al. 1986). McKenzie et al. (2000) proposed that one manifestation of the absence of lower-crustal flow is the preservation of crustal thickness variations. Interestingly, a recent seismic study across the Cheyenne belt suture (developed during the 1.78–1.74 Ga Medicine Bow orogeny) indicates that locally thickened crust is associated with this ancient suture (Crosswhite & Humphreys 2003). Thus, crustal thickening associated with the Medicine Bow orogeny, and in part manifested by the LPSS, is apparently still manifested in the overall crustal structure (Moho topography) of this part of the Rocky Mountains.

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