Maps Showing Geology, Structure, and Geophysics of the Central Black Hills, South Dakota

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Introduction

Preparation of the 1:100,000-scale geologic map (map A) began in the early 1980s with the original intent of limiting the map to the Precambrian rocks, where there had been considerable recent mapping by U.S. Geological Survey personnel. Because there was also recent work on the Phanerozoic rocks, the map was extended to cover the area included in the U.S. Geological Survey folio of the central Black Hills (Darton and Paige, 1925). The map in that folio has served as a basis for subsequent studies for more than 70 years. The reader can compare the original folio map with the present map and see that there are relatively few differences in the portrayal of the Phanerozoic rocks. If better base maps had been available, there would have been fewer differences. The work by Darton and Paige was done without many of the present roads and with relatively primitive transportation. This early work is most impressive, considering these difficulties and the general geologic knowledge of that time.

Darton was a self-made geologist whose education was limited to grade school. His powers of observation and keen mind led him to become, quite possibly, the U.S. Geological Survey's greatest field geologist by mapping approximately 1,000,000 mi^2 during his lifetime. We stand in awe of his great accomplishments.

Since 1925, two large-scale maps of the Black Hills updated the regional knowledge of both Phanerozoic and Precambrian rocks (Kleinkopf and Redden, 1975; DeWitt and others, 1989). The first map summarized geophysical data available for the region. The second updated the Precambrian stratigraphy and structure. Initial compilations of selected parts of the region at 1:100,000 scale were contained in an appraisal of the mineral resource potential of the Black Hills (DeWitt and others, 1986).

Compilation of map A involved many sources of data. Major data sources are listed on the map. An arbitrary area approximately equivalent to a 7½-minute quadrangle or the contribution of major basic data was used to credit authorship (fig. 1). The senior author is responsible largely for the compilation, interpretation of correlation, and structure, which may not always be in agreement with the original investigator. Some areas were mapped at different times by different individuals, and the present map is an interpretation of the original data where differences existed. The relatively recent compilations by A.L. Lisenbee and the senior author of areas involving Tertiary igneous rocks in the northern Black Hills greatly improved the quality of this map. A chemical classification of those igneous rocks is used on the map largely because of the difficulty of correlating aphanitic or sparsely porphyritic specimens with chemically analyzed samples. Refer to Lisenbee and DeWitt (1993) for a description of the classification.

It should also be apparent from the sources of data that the detail of the map is not everywhere uniform. Areas mapped from aerial photographs and subjected to only limited field checking may be less accurate than areas compiled from published geologic maps. Also, authors of the different source maps did not necessarily map the same units of Tertiary igneous rocks and Quaternary surficial deposits both in Precambrian and Phanerozoic areas. The authors have attempted to achieve a general uniformity but recognize that errors will exist. Precambrian rock units also may differ somewhat from those in the original source maps. Such deviations—justified or not—are the full responsibility of the authors.

Because the Black Hills has the easternmost exposures of rocks generally referred to as Laramide or part of the Rock Mountain orogeny, and is the only exposed Precambrian part of the Trans-Hudeon orogeny in the northern United States, it is widely visited and sampled by geologists. Hence, samples of the Precambrian rocks are the subject of many scientific papers involving geochemistry, isotopes, and so forth

Precambrian Rock Units

Continuity of Precambrian rock units in the Black Hills and their correlation are locally problematic because of complicated structure resulting from repeated deformation. Also, pronounced facies changes, lack of relict facing structures in some units, lack of exposures in some key areas, effects of younger igneous events, and several episodes of metamorphism complicate interpretations. Added factors include lack of detailed mapping in some areas and the human element of interpretation during mapping. Primary map data come from many sources but chiefly from work done by U.S. Geological Survey personnel, including Richard Bayley, J.J. Norton, James Ratté, R.G. Wayland, and the authors. Published and open-file reports of contiguous and noncontiguous areas have used both formation and lithologic names. Later work has shown that some of the formation names cannot be extended from the original areas. Also, in places, names were used to include subunits that are of different ages, for example Ratté's (1986) revision of the Loues and Vanderlehr Formations originally defined by Redden (1968).

Because the prototypes of most of the rocks can be identified with assurance, and primary structures can generally be readily identified in spite of the metamorphism, the different rock units generally are named first according to protolith and then by common metamorphic rock names. The more detailed "Description of Map Units" on map A indicates the various formation names used in different areas, changes in facies, and the interpreted lithologic or stratigraphic equivalent. In a few examples, an accepted formation name is used for a unit if there is little probability of erroneous correlation. Depositional environments that produced certain rock types, such as graywacke (turbidites), quartzite and conglomerate, or carbonate-facies iron-formation, existed at several times. Thin units, such as the iron-formation, are not shown everywhere, either because of omission during the original mapping or problems of scale. Some units, such as the carbonate-facies iron-formation, are locally shown in detail (although necessarily

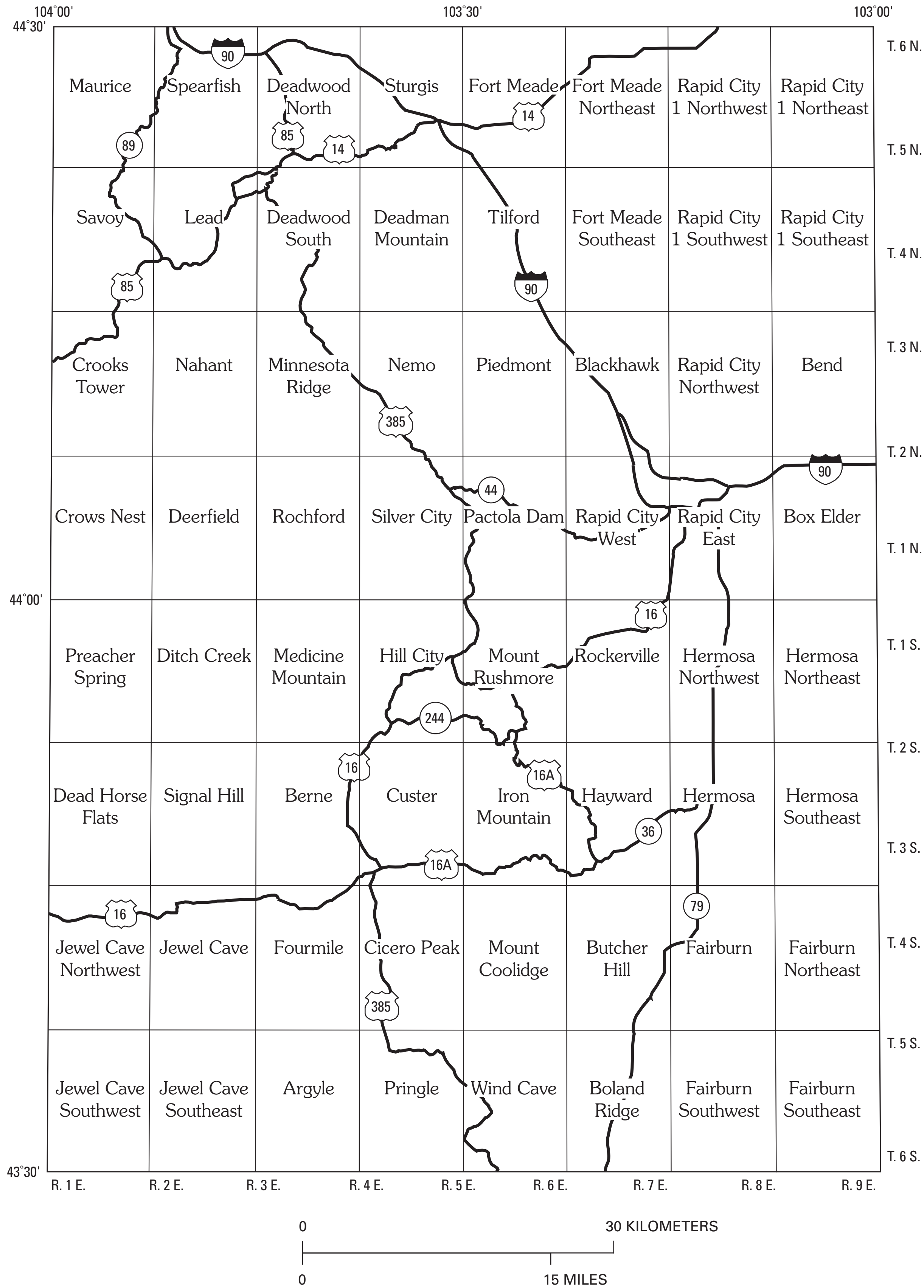


Figure 1. Index map of the central Black Hills showing location and names of 7.5-minute quadrangles.

requiring exaggeration of thickness) where they are critical either in stratigraphic interpretation or other aspect, for example the Homestake Formation, host of gold deposits in the Lead area.

Archean Units

Archean metasedimentary units have limited exposures both in the Little Elk Creek area (T. 3 N., R. 6 E.) and at Bear Mountain (T. 2 S., R. 3 E.). The older metasedimentary rocks (unit Wos) at Little Elk Creek appear to have been intensely metasomatized by the emplacement of the 2.56-Ga I-type Little Elk Granite (Gosselin and others, 1988). Age-equivalent rocks contained in the same map unit (unit Wos) at Bear Mountain are 2.6-Ga S-type pegmatitic granite and trondhjemite that intrude Archean metamorphic rocks (McCombs and others, 2004). The oxide-facies (predominantly hematite) banded iron-formation (unit Wif, Nemo iron-formation) is inferred to be Archean because clasts of similar rock are found in the earliest Proterozoic rocks in the Nemo area (Redden, 1987b) and because units Wif and the Benchmark Iron-formation (unit Xbi) have different concentrations of rare-earth elements (REE) (Frei and others, 2008). Faulting and limited exposures do not permit recognition of exact stratigraphic relationships of the Archean rocks at Little Elk Creek, but an unconformity exists between these units and Early Proterozoic conglomerate of the Boxelder Creek Formation (unit Xbcg). Adjacent, undifferentiated Archean rocks (unit Wu) in the Nemo area are too poorly exposed to be characterized. A major fault separates unit Wu from Proterozoic rocks to the west.

The oxide-facies iron-formation (unit Wif) can be extended, under Phanerozoic cover, from the Little Elk Creek area to along the east side of Precambrian rocks in the Nemo area by interpretation of magnetic anomalies. A major gravity low (map E) centered near the town of Piedmont suggests that the exposures of the Little Elk Granite (unit Wgr) to the west are part of a much larger granite body or block of Archean granitoid rocks concealed by the Phanerozoic cover. In contrast, gravity and magnetic data at Bear Mountain indicate a much smaller mass of Archean rocks surrounded by Proterozoic metamorphic rocks.

Oldest Early Proterozoic Units (Pre-2,480 Ma)

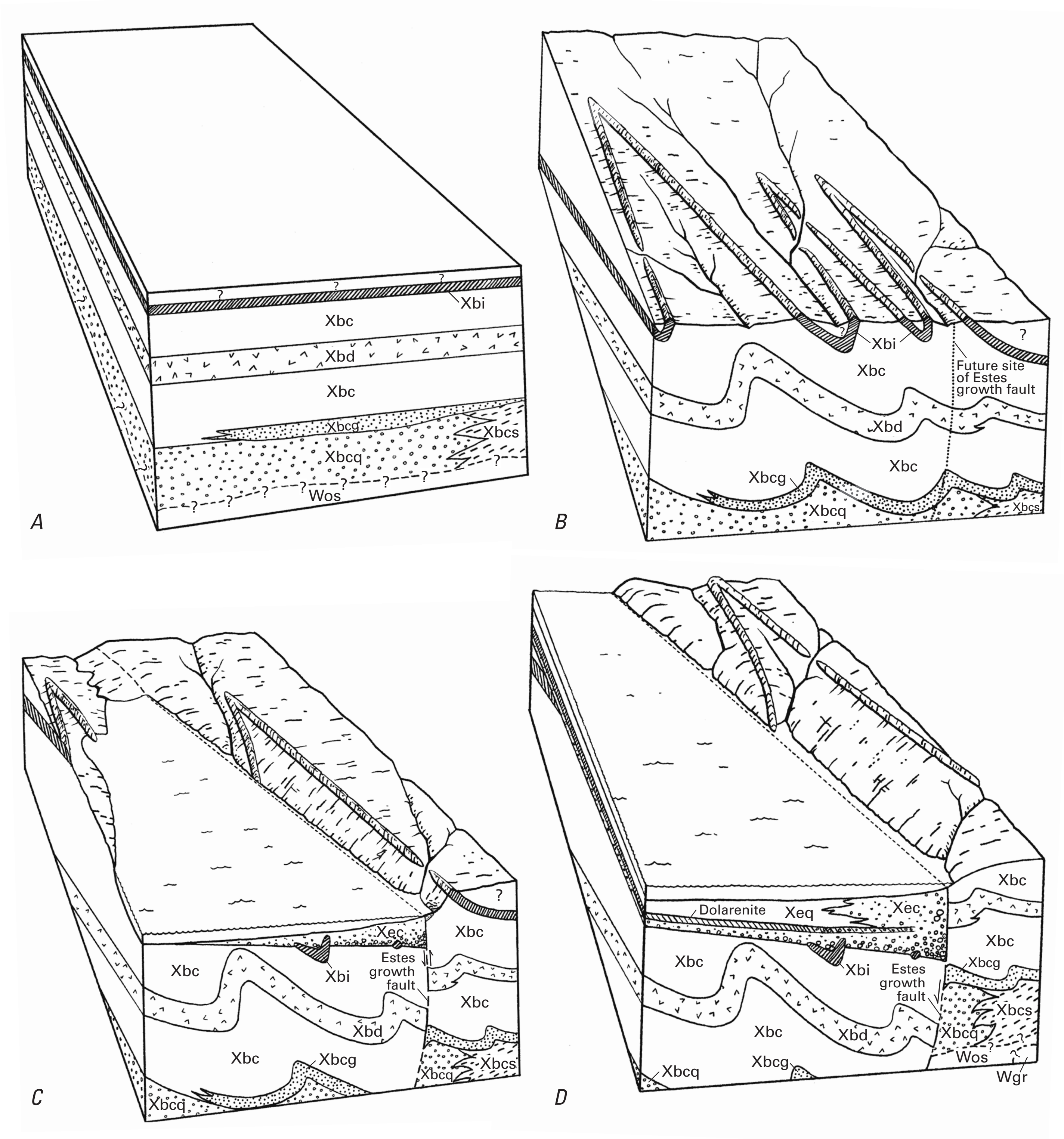
The oldest Early Proterozoic rocks are confined to the Nemo area, where the depositional environment differed markedly from that in younger Early Proterozoic rocks (Redden, 1980). These rocks in general dip steeply to the north and are overturned, the map pattern closely resembling a cross section. The oldest unit is the heterogeneous Boxelder Creek Formation. The lower part of the Boxelder Creek consists of intertonguing rock types including conglomerate and paraconglomerate, containing predominantly quartzite and banded iron-formation clasts, and chloritic quartzite (unit Xbcq). A few clasts of dolomite and vein quartz are present but igneous clasts are lacking. In general, this unit grades laterally to the northeast into distal equivalents consisting of siltstone, shale, and dolomite (unit Xbcs). Locally, abrupt terminations of thick conglomerate subunits indicate the likelihood of growth faults. A thin, somewhat uraniferous and auriferous, pyritic fluvial conglomerate and grit unit (Xbcg) stratigraphically overlies the conglomerate and quartzite (unit Xbcq) in the eastern and southeastern parts of the Nemo area. However, in the area south of Benchmark the conglomerate and grit unit pinches out into thick-bedded, fluvial quartzite (unit Xbc). Detrital chromite is especially abundant in unit Xbcq but is also a common heavy mineral in most of the clastic parts of the Boxelder Creek Formation. The distribution and type of sediment in the Boxelder Creek Formation indicate a rift environment (table 1). Abundant trough cross bedding in the conglomerate and grit (unit Xbcg) indicates a nearly constant up-dip current direction. Restoration of the entire overturned section of rock along an east-west axis suggests a source area to the north and transport along northerly trending active fault scarps (Redden, 1980, 1987b). Sedimentation patterns suggest that the source area for the Boxelder Creek rocks may have been somewhere to the west or northwest in the area now covered by younger Proterozoic rocks of the main part of the central Black Hills basin.

The thick quartzite of the Boxelder Creek is concordantly overlain by the Benchmark Iron-formation (unit Xbi). Due to folding prior to deposition of quartzite and conglomerate of the Estes Formation (unit Xec), the Benchmark is preserved only in synclines below the erosion surface that truncates the Benchmark and all older rocks (fence diagram on map A and fig. 2*B*).

The Blue Draw Metagabbro (unit Xbd) forms a 1,000-m-thick, gravity-differentiated sill that intrudes the thick quartzite (unit Xbcq) of the Boxelder Creek Formation. Two sphene grains, analyzed in 16 places by SHRIMP techniques, have a weighted mean upper intercept of 2484±11 Ma (Dahl and others, 2006). Six zircon grains, analyzed in 10 spots by SHRIMP, have $^{207}Pb/^{206}Pb$ ages of 2284–2458 Ma and do not define a precise crystallization age. We use the 2,484 (rounded to 2,480) Ma date as the intrusive age of unit Xbd. Therefore, both the Benchmark and the Boxelder Creek are older than approximately 2,170 Ma.

Older Early Proterozoic Units (2,480–1,900 Ma)

The oldest Early Proterozoic stratified units, preceding, were intruded by the Blue Draw Metagabbro at 2,480 Ma, folded, and eroded prior to deposition of the Estes Formation. The Estes includes a lower conglomerate and quartzite (unit Xec) and an upper quartzite (unit Xeq). These rocks were deposited along growth faults (fig. 2) in three separate fans (Redden, 1980, 1987b). Although the sources of most


A
Xbi
Xbc
Xbd
Xbc
Xbcg
Xbcs
Xbcq
Wos
B
Xbi
Xbc
Future site of Estes growth fault
Xbd
Xbc
Xbcg
Xbcq
Xbcs
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of the clasts are the oldest Early Proterozoic units, abundant feldspar and blue quartz in the uppermost Estes Formation (unit Xeq) indicate a probable source from some of the Late Archean Little Elk Granite (unit Wgr). The presence of minor lenses of dolarenite, dolomite-cemented conglomerate, and the general absence of primary structures (other than bedding) suggest a moderately shallow marine environment on an intracontinental rift basin margin (fig. 2; table 1). One clast of oxide-facies banded iron-formation of an apparent drop stone in the dolarenite suggests ice rafting. The main part of the basin is interpreted to have been centered in a general way on the present-day central Black Hills, and the source area was to the east.

Along Rapid Creek near the east boundary of the Pactola Dam quadrangle and also northeast and east of Johnson Siding are limited exposures of thick arkose beds, quartzite, and pelite, which are considered to be the equivalent of the upper part of the Estes Formation (unit Xeq) in the Nemo area. Contradictory younging structures suggest that this largely concealed unit of Xeq is in fault contact with adjacent rocks to the west. The stratigraphic sequence is believed to have been derived from Archean basement to the southeast that was uplifted along growth faults similar to those responsible for the deposition of the Estes Formation in the Nemo area. This interpretation is supported by a significant gravity low south of Rapid City that is interpreted to be a basement block of Archean granitoid rocks extending to the south end of the map area (see “Geophysics” section).

Shallow marine conditions persisted during deposition of dolomite and silty pelite (unit Xds) and dolomite (unit Xd) units in the Nemo area. These carbonate-rich rocks include thin lenses of conglomerate and are more distal continuations of the fan deposition (although locally in fault contact) of the younger rocks in the Nemo area. Sparse stromatolitic structures are recognized near the center of sec. 32, T. 3 N., R. 5 E. The only other significant carbonate units in the Black Hills are at Bear Mountain and inliers within the central part of the Harney Peak Granite (unit Xh). Because of similar lithologic sequences in the three areas, the carbonate units are assumed to be correlative. At Bear Mountain the underlying quartzite (unit Xeq), including a very thin basal conglomerate (not shown) and some arkosic beds, unconformably overlie Archean rocks and are inferred to be equivalent to lithologically similar rock types in the Nemo area. Similar quartzite and conglomerate (unit Xeq) are present as screens and inliers within the central part of the Harney Peak Granite.

West of the Nemo area the stratigraphic sequence above the older carbonate rocks consists of tholeiitic pillow basalt (unit Xbo), ferruginous chert and minor carbonate-facies iron-formation (unit Xfc), mafic tuff and shale (unit Xmt), and interbedded quartzite, siltstone, and shale (unit Xqs). Earlier publications (Bayley, 1972b; Redden and others, 1990) used essentially the reverse of this sequence because it was not realized that extensions of the pillow basalt (unit Xbo) are present along Bogus Jim Creek in an area previously mapped as metagabbro. The tholeiitic pillow basalt near Benchmark is apparently truncated by faults to the south but must be equivalent to the basalt along part of Bogus Jim Creek. Unit Xbo is repeated in elongate belts to the west in cores of anticlines. The main unit Xbo belts contain numerous pillowed flows, commonly separated by thin-bedded pelitic interflow intervals that may have been tuffaceous in part. Pillow structures are generally common and are also present in the two areas of unit Xbo in the Lead area.

The pillowed basalt of unit Xbo north of Lead is the so-called Yates member of the Poorman Formation (informal name used by Homestake mine staff; Dodge, 1942; originally considered a sill by Noble and Harder, 1948). Thin amphibole schist units at Bear Mountain and within the Harney Peak Granite dome also have a tholeiitic composition and are also

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Figure 2 (facing page). Generalized block diagrams showing stratigraphic and structural events in the Nemo area and development of Estes growth fault. Surface terrane is hypothetical, and view is to the north. Not to scale. Unit queried where unknown.

A. Original stratigraphic sequence of Boxelder Creek Formation (units Xbc, Xbcg, Xbcq, and Xbcs) and overlying Benchmark Iron-formation (unit Xbi). Rocks overlying unit Xbi may have been present but were eroded before deposition of the younger Estes Formation. Gravity-differentiated, 2,480-Ma Blue Draw Metagabbro (unit Xbd) sill also shown. Contact with Late Archean metasedimentary rocks (unit Wos) shown as unconformable although it is nowhere exposed. Unconformity based on oxide-facies iron-formation clasts found in Boxelder Creek Formation that resemble the Nemo Iron-formation (unit Wif, not shown).

B. Reconstruction of earliest (pre-D_1) deformation and erosion preceding development of Estes fault. Note that orientation and plunge of folds are unknown but that rotation to horizontal of overturned unconformity at the base of the Estes Formation suggests a gentle plunge to the north. Foreshortening of the block diagram due to folding is ignored in the cross-sectional view.

C. Early stages of development of the Estes fault showing deposition of lower conglomerate and quartzite (unit Xec) in alluvial fans. Restriction of estuary to the north is purely conjectural. Although only vertical fault movement is shown, considerable right-lateral horizontal movement is likely, resulting in the left side of the block moving northward, away from the observer. Erosion causes fold noses to migrate northward at the surface.

D. Continued development of Estes growth fault and associated conglomerate and quartzite (unit Xec) and quartzite (unit Xeq). Dolarenite (part of unit Xec), which merges into a dolomite-cemented conglomerate toward the Estes fault, is based on similar rock types exposed in the largest fan of the Estes Formation. Uppermost conglomerate and arkose of the Estes suggest that Little Elk Granite (unit Wgr) east of Nemo is ultimately exposed on the upthrown fault block. Inferred Archean older metasedimentary unit (Wos) shown as basement.

Table 1. Late Archean and Early Proterozoic depositional and tectonic events in the central Black Hills.

[---, not applicable]

Age (Ma)	Tectonic events	Igneous activity	Sediment type, environment, and source (where applicable)
1,700–1,540	Northwest-southeast or north-south compression (D_5). Development of local S_5 foliation. Late faulting of indeterminate age. Possible development of minor east-west "buckle" fold zones.	Isotopic re-equilibration(?)	---
1,715±3	Major diapiric doming (D_4), low-pressure metamorphism, local late S_4 foliation, minor F_4 folds, reorientation of adjacent F_1 and F_2 structures, major thrust faults.	Harney Peak Granite and associated pegmatite. A similar concealed granite east of Deadwood and also in the Crook Mountain area may be slightly younger.	---
About 1,715 but could be slightly older.	Northeast-southwest compression (D_3). Development of local folds (F_3) and foliation (S_3). Possible warping of F_2 and other structures by competent Archean block to the northeast.	None	---
Post-1,880, pre-1,715.	East-northeast–west-southwest compression (D_2). Isoclinal folding produces F_2 folds accompanied by regional metamorphism and development of steeply dipping S_2. Presumed warping of F_1 folds into general conformity with Archean structural highs. Development of right-lateral, north-northwesterly trending faults follows F_2.	None	---
	Northwest-southeast compression (D_1). Isoclinal folding produces F_1 folds locally overturned to northwest and probably accompanied by thrust faults.	None	---
1,884±29 for alkalic tuff in Rochford area. 1,883±5 for gabbroic sill on Prairie Creek in Pactola Dam quadrangle.	Regional uplift, tilting, and erosion followed by major, but irregular, subsidence along north-trending epicratonic basin of central Black Hills.	Submarine alkalic pillow basalt flows and volcaniclastic rocks (unit Xby) and minor ash fall tuff centered at Rochford and pillow basalt west of Rockerville. Localized siliceous and iron-sulfur-rich thermal spring deposits commonly associated with volcanic flows. Associated gabbroic sills and dikes are assumed to be same age as dated gabbroic sill on Prairie Creek, which is inferred to be approximate age of unit Xby. Sparse gabbroic dikes intrude unit Xgw_2 and may be somewhat younger.	Local debris-flow deposits, conglomerate, and quartzite (Xqc), characteristic of a steep slope environment, unconformably overlie older shelf deposits along part of east edge of basin. Unit changes facies to deeper water turbidite deposits (graywacke) to the north and west in central part of basin, and the unconformity below unit Xts cannot be recognized. Turbidite deposits are probably derived mainly from the east and pinch out to the west where a volcanic center existed near Rochford. Volcanism produced extensive volcaniclastic rocks whose laterally equivalent tuff and shale (unit Xts) represents submarine, weathered alkalic tuff that is extensively distributed throughout west-central part of Precambrian rocks. Isolated unit of graywacke (Xgw_3) along east side of Bear Mountain may have had a western source. Sources of graywacke units southwest of Grand Junction fault are unknown. Extensive shale (unit Xs) apparently indicates final deposition in main basin of central Black Hills Precambrian.

No available age for basalt (unit Xbo). Gabbroic sill below basalt along Bogus Jim Creek is 1,964±15 and tuff in unit Xqg above basalt in Lead area is 1,974±8.	Ensialic rifting in Nemo area forms north-striking, right-lateral(?) growth faults along east margin of area, which becomes the main intracratonic basin of the central Black Hills. The rift section is largely overturned to the north during the D_1 event.	Tholeiitic, pillow basalt (unit Xbo) adjacent to Nemo area and repeated in fold belts to the west and in Lead area. Thermal spring deposits common between flows. Gabbroic sills intrude all rock units.	Fanglomerate and quartzite (units Xec and Xeq) in Nemo area unconformably overlie, and are derived from, older Early Proterozoic rocks and Archean rocks to the east. Younger shallow-water dolarenite and pelite (units Xd and Xds) are overlain by pillow basalt (unit Xbo) and thermal spring deposits of ferruginous chert and carbonate facies iron-formation (unit Xfc). The overlying mafic tuff and shale (unit Xmt) are transitional to younger quartzite, rippled siltstone, and shale (unit Xqs) characteristic of a shelf environment that persisted along central eastern margin of Precambrian rocks. A shallow-water environment existed in Lead area where pillow basalt is overlain by chemical sediment, including carbonate-facies banded iron-formation (unit Xif), and quartzite with pelite (unit Xqg). The latter changes facies to the south-southwest into turbidite deposits northwest of Rochford. In central part of main basin, individual flows in pillow basalt (unit Xbo) pinch out north of Pactola Lake into a black shale unit (Xbs_1) that probably represents deeper basin deposits. Source area and facies changes in small areas underlain by rocks believed to be correlative with preceding units are generally unknown. Quartzite (unit Xq) south and east of Custer may be older than quartzite and pelite (unit Xqs) to the northeast.
	Uplift and erosion	None	---
	Moderate deformation in Nemo area, east-west(?) compression; north(?)-trending, probably open folds.	None	---
2,484±11	None	Emplacement of gravity-differentiated Blue Draw Metagabbro.	---
About 2,490	Ensialic rifting; faulting in Nemo area, uplift to west; inferred early intracratonic basin east(?) of Black Hills.	None	Fanglomerate, fluvial conglomerate, sandstone, and distal equivalents in lowermost Boxelder Creek Formation (units Xbc, Xbcg, Xbcq, and Xbcs). Capped by oxide-facies Benchmark Iron-formation (unit Xbi). Probable sources are Archean sedimentary rocks and chromite-bearing anorthosite to the west or northwest. Detrital chromite recycled in part from Archean sedimentary rocks.
About 2,500	Uplift and erosion	None	---
2,549±11	None	Little Elk Granite	---
2,550	None	None	Feldspathic quartzite and minor conglomerate (unit Wos) in fault contact with oxide-facies iron-formation (unit Wif) in Little Elk creek area.
2,596±11	None	Granite at Bear Mountain	---
pre-2,600	Burial and deformation. Few details available.	None	Poorly exposed metasedimentary rocks at Bear Mountain (unit Wos); details poorly known.

believed to be approximately equivalent to unit Xbo although no pillow structures are known.

Many of the interflow intervals are characterized by massive or thin-bedded chert and iron-rich rocks that grade into a carbonate-facies iron-formation (unit Xif). Many prospect pits in these repeated rock types expose gossanlike material, commonly carbon-rich, and the fresh rock is almost certainly sulfide-bearing. The relatively massive chert (streaked quartzite) locally may be tens of meters thick but typically extends only a few hundred meters along strike. Individual flows in the belt of unit Xbo, which passes through Pactola Dam, decrease in abundance to the north and intertongue with black shale (unit Xbs_1) ending near Estes Creek. The next belt to the east ends in a fold nose at Rapid Creek outlined by the underlying ferruginous chert unit (Xfc). To the north this Xbo unit seemingly has a conformable contact with black shale (unit Xbs_1). The latter has a fault contact on its west side separating it from younger graywacke. Near Roubaix the pillow basalt is exposed in cores of refolded F_1 folds and is overlain by a thin unit of black shale believed to be equivalent to unit Xbs_1.

The age of the metamorphosed tholeiitic basalt (unit Xbo) is not known. A quartz-bearing metagabbro sill that intrudes unit Xds on the north side of Bogus Jim Creek just south of Green Mountain (T. 2 N., R. 4 E.) has a U-Pb zircon age of 1,964±15 Ma (Redden and others, 1990) but it cannot be proven to be related to the basalt. However, the sill resembles other sills in the adjacent metabasalt belt, which suggests the 1,964-Ma date would be the approximate age of the basalt (unit Xbo).

A thin ferruginous metachert (unit Xfc) generally overlies the pillow basalt and extends across parts of the Nemo, Piedmont, and Pactola Dam quadrangles. At numerous localities massive ferruginous chert forms resistant wall-like exposures 3–15 m thick and as much as 25 m high. Float from the chert commonly masks an equal or greater thickness of thin-bedded chert and carbonate-facies iron-formation and sulfide-bearing beds. Elsewhere the massive chert is thin or absent and the unit consists largely of thin chert and carbonate-facies iron-formation. Except for its continuity, the ferruginous metachert unit closely resembles some of the interflow rocks in the adjacent metabasalt (unit Xbo) as well as the carbonate-facies iron-formation (unit Xif) in the Lead area. Locally, there may be a few meters thickness of basalt below the unit.

The mafic tuff and shale unit (Xts) includes Bayley's (1972b) Gingrass Draw Slate and the lower part of the unit that he included in the Buck Mountain Quartzite. At the type locality of the Gingrass Draw, in the southwestern part of the Nemo quadrangle, the unit is thin-bedded green chlorite-biotite phyllite and schist. However, to the south in the Pactola Dam quadrangle the unit changes facies and includes tan quartzite beds and siltstone. Chlorite is less abundant, and the unit is difficult to distinguish from the quartzite, siltstone, and shale (unit Xqs). Chlorite also decreases to the north, and a few thin lenses of chert are noted in the northernmost exposures of the unit. The eastern part of the unit in the Pactola Dam quadrangle is more siliceous, and the contact with the overlying quartzite, siltstone, and shale (unit Xqs) is transitional. Much thicker, tan to gray quartzite beds and thin, commonly ripple marked silt beds characterize the overlying Xqs unit. Unit Xqs is apparently considerably thicker in the Pactola Dam quadrangle where abundant relict structures indicate a shallow shelf depositional environment.

Carbonate-facies iron-formation (unit Xif) in the older Early Proterozoic rocks is relatively common as thin lenticular or continuous subunits within many of the thicker units shown on the map. For example, the older basalt (unit Xbo) locally contains lensoid iron-formation subunits between flows as previously noted. These subunits are generally not shown on the map. At low metamorphic grade the iron-formation is largely recrystallized chert and iron-rich carbonate minerals. However, these react approximately at the garnet metamorphic isograd to produce cummingtonite-grunerite. Sulfide- and carbon-rich beds are common. This is evident in gold-bearing iron-formation ore from the Homestake mine (photograph, sheet 2). Carbonate-facies iron-formation associated with volcanic or volcanogenic units may grade into massive, somewhat iron-stained metachert or quartzite lenses. Such quartzite typically forms topographic highs. Some of the iron-formation units have remarkable continuity, such as in the Lead area where the Homestake Formation (Hosted and Wright, 1923), which separates the Poorman and Ellison Formations, is traceable for more than 15 km through intensely folded outcrops. The ferruginous chert (unit Xfc, previously described) has a considerably longer strike length than the Homestake Formation but differs somewhat as it commonly contains massive chert. In general, most of the iron-formation is associated with volcanic and tuffaceous deposits and it is believed to have been deposited by, or be closely associated with, marine thermal spring activity. Hence, iron-formation unit Xif is believed to be restricted to either proximal vent areas or to local basins.

A large area containing quartzite and minor pelite east and southeast of Custer (unit Xq) is interpreted to be in fault contact with all other rock units. Rocks in this area are all above the sillimanite isograd, and sedimentary structures characteristic of the Xqs unit are not recognizable. Tentative correlation of these rocks with those of the eastern belt of quartzite and pelite (unit Xqs) that crosses Rapid Creek is based on general similarity of rock types and proximity of the two areas.

Quartzite, pelite, and graywacke unit (Xqg) west of Nahant is largely graywacke, but north and northeast of Rochford it consists of quartzite and pelite and was named the Moonshine Gulch Quartzite (Bayley, 1972c). Stratigraphic-facing criteria indicate the quartzite and pelite are correlative and equivalent to the Ellison Formation in the Lead area (Hosted and Wright, 1923). At Lead, minor carbonate beds in surface exposures of the Ellison suggest a shallow marine depositional environment. West of Nahant, quartzite (subgraywacke) and graywacke suggest a deeper turbidite environment for rocks of the same age. A tuff bed in the basal part of the Ellison Formation at Lead has a U-Pb zircon age of 1,974±8 Ma (Redden and others, 1990), which makes unit Xqg about 90 million years older than the main graywacke units (units Xgw$_2$ and Xgw$_3$) in the eastern and southeastern parts of the central Black Hills.

Younger Early Proterozoic Units (Post-1,900 Ma)

In the east-central part of the Black Hills, a heterogeneous unit of conglomerate, debris flow sediments, quartzite and pelite (unit Xqc), carbonate-facies banded iron-formation (unit Xif), and younger alkalic basalt (unit Xby) apparently unconformably overlies the older Early Proterozoic rocks. Although the unconformity is not evident in individual exposures, several lines of evidence, including the ages of the rocks in the Lead and Rochford areas, favor its existence (Redden and others, 1990). In the Lead area the Northwestern Formation (unit Xst) is apparently truncated below the unconformity (Noble and Harder, 1948). The presence of quartzite clasts in boulder conglomerate in the easternmost exposures of unit Xqc in the Pactola Dam quadrangle that are lithologically similar to quartzite in the adjacent older quartzite and pelite (unit Xqs) suggests an unconformity. Farther north at Rapid Creek the contact of unit Xqc truncates the west limb of an anticline cored by basalt (unit Xbo), ferruginous metachert (unit Xfc), and mafic tuff and shale (unit Xts). Clasts similar to the ferruginous metachert are noted within unit Xqc near Rapid Creek. Along trend to the north of Rapid Creek, unit Xqc thins considerably and probably changes facies, becoming equivalent to the upper part of the older graywacke (unit Xgw$_1$). This facies change is consistent with changes to the west in the Silver City quadrangle, where unit Xqc cannot be positively identified north of the Silver City fault and where several mappable units become diluted with turbidite deposits (graywacke) and cannot be continued to the north, through the Nemo quadrangle. It is also consistent with other facies changes in the heterogeneous unit Xqc.

The predominant rock types in the heterogeneous unit are conglomerate, debris flows, quartzite, shale, and lenses of carbonate-facies iron-formation (unit Xqc), which are generally indicative of a steep-slope depositional environment. This is in general agreement with the base of the unit, at least locally, marking the boundary between the younger and older Early Proterozoic rocks. The unit is best developed in the Pactola Dam and Mount Rushmore quadrangles. Conglomerate containing a range in clast size from sand to boulder is largely limited to the easternmost exposures in the Pactola Dam quadrangle. Debris flow deposits characterized by isolated clasts (locally exceeding 2 m in diameter) in a fine-grained matrix are best developed in parts of the Mount Rushmore quadrangle and commonly have associated lenticular, thick, structureless quartzite beds. Localized lenses of carbonate-facies iron-formation (unit Xif) also tend to be more abundant in the Mount Rushmore quadrangle and are commonly associated with localized younger alkalic basalt flows (unit Xby). In general, the characteristic rocks of unit Xqc disappear to the north and west with an increase of graywacke representing deeper water deposits. Thus the angular unconformity to the east apparently becomes a disconformity to the west. In the general area north and south of Hill City there is no angularity at the base of unit Xqc. Some clasts in debris flow-like rocks resemble rocks of adjacent graywacke units, and extrapolation of unit Xqc as a single map unit in that area may be incorrect.

Carbonate-facies iron-formation (unit Xif) is found at several stratigraphic intervals in the younger Early Proterozoic rocks. In the Rochford area some of the iron-formation forms relatively continuous thin units. Two such named units (the Rochford and Montana Mine Formations; Bayley, 1972c) may be the same unit repeated by folding. However, several beds of iron-formation are within younger alkali basalt (unit Xby) and the shale, tuff, and volcaniclastic rocks (unit Xtv) in the Rochford area. The relatively wide map units of iron-formation in the Sheridan Lake area (T. 1 S., R. 6 E.) have little outcrop, but extensive gossanlike float suggests sulfide-rich iron-formation. Thin iron-formation also is present locally within graywacke in the Berne quadrangle (Redden, 1968). These carbonate-facies iron-formation units in the younger Early Proterozoic rocks appear to have an origin in common with those units in the older Early Proterozoic rocks. Both appear to have been deposited by marine thermal springs.

The tuff and shale unit (Xts) in much of the west-central part of the Black Hills is lithologically and geophysically distinct because of its anomalous concentration of ilmenite and magnetite, and sparse, but widespread, copper minerals, which can create malachite-stained surfaces. Subunits similar to black shale separate the purer tuffaceous subunits. The latter commonly contain 60 percent or more muscovite and 15 percent biotite, a level indicating a very high aluminum concentration. This strongly suggests that the tuff reacted with seawater prior to lithification. The tuff and shale become diluted with graywacke beds to the east and northeast of the Hill City area. Although unit Xts is shown north of Harney Peak, it cannot be recognized to the northeast. Micaceous schist, included in the unit Xgw$_2$ graywacke but close to its contact with unit Xqc, east and northeast of Mount Rushmore, may be at the equivalent stratigraphic level. Northeast of Mystic, the tuff and shale unit (Xts) contains very limited magnetite but includes a thin alkali pillow basalt (unit Xby).

West and northwest of Mystic, the tuff and shale unit (Xts) is interlayered with alkalic basaltic flows (unit Xby) and diluted with coarse volcaniclastic material derived from a volcanic center near Rochford. There the correlative unit is designated shale, tuff, and volcaniclastic rocks (unit Xtv). The general age of the tuff and shale unit (1.88 Ga) is based on the 1,884±29 Ma U-Pb zircon age of thin alkalic ash-fall beds associated with carbonate-facies iron-formation (Redden and others, 1990). Although Bayley (1972c) separated unit Xtv (his Flag Rock Group) into several formations, it seems best to include all of the tuffaceous rocks having alkalic volcanic affinity in a single unit. This conclusion was made in part because somewhat lower metamorphic grade rather than stratigraphic change probably accounts for the disappearance of Bayley's garnet-bearing Poverty Gulch Slate south of Rochford. Because the southern Rochford area has not been remapped in detail, the upper contact of the shale, tuff, and volcaniclastic rocks in that area is very indefinite.

The eastern limb of the shale, tuff, and volcaniclastic rocks (unit Xtv) at Rochford can be followed nearly continuously from Rochford to the Lead area except for a narrow covered interval of Cambrian rocks. At Lead the name Flag Rock Formation was used for these rocks (Hosted and Wright, 1923). Thin carbonate-facies iron-formation and massive chert locally crop out within the Flag Rock in the Lead area but are not shown on the map. Alkali basalt and volcaniclastic rocks are sparse in the area between Rochford and Lead but are more widespread at Lead, possibly indicating another volcanic center.

The younger alkali basalt (unit Xby) in the Rochford area includes pillowed flows, fragmental volcaniclastic units, and well-bedded, impure tuff. The tuff typically is rich in amphibole and biotite. Fragment size in the volcaniclastic material decreases northward from Rochford, and amphibole-bearing beds are diluted by input of pelitic material. Similar alkali basalt associated with quartzite and conglomerate (unit Xqc) in the Keystone-Rockerville area has calc-alkalic as well as alkalic components, and it is believed to be approximately the same age as alkali basalt in the Rochford area. A differentiated gabbroic sill along Prairie Creek in the Pactola Dam quadrangle has a U-Pb zircon age of 1,883±5 Ma (Redden and others, 1990), which documents igneous activity in that area corresponding to the alkalic tuff in the Rochford area. These mafic volcanic rocks and associated thin sills (unit Xgby) are not found within the quartzite and conglomerate (unit Xqc) to the northwest.

Graywacke dominates the south-central part of the Precambrian area, where it is shown as two units (units Xgw_2 and Xgw_3) separated by tuff and shale (unit Xts) and quartzite and conglomerate (unit Xqc). Graywacke units Xgw_2 and Xgw_3 can be recognized only for about 6 km north of Pactola Lake, where the Xts unit becomes so diluted with graywacke that it can no longer be recognized as a mappable unit. A similar situation is noted with the Xbs_2 unit about 4 km farther north, where older graywacke (units Xgw, Xgw_1, and Xgw_2) cannot be distinguished due to an increase in graywacke within unit Xbs_2. West of the Pactola Lake area graywacke subunits pinch out to the west, and apparently equivalent graywacke was never deposited in the Rochford or Lead areas. The uppermost graywacke (unit Xgw_3) does reappear near Bear Mountain, which may indicate a source from the west side of the original basin. Correlation of graywacke subunits along the west side of the Harney Peak granite dome is uncertain due to possible facies changes in the quartzite and conglomerate unit (Xqc) or possibly a repetition of rock types.

Thick graywacke units predominate west of the Grand Junction fault in the Berne and Fourmile quadrangles, but the graywacke subunits do not fit the subdivisions (Xgw_1, Xgw_2, and Xgw_3) in the south-central part of the Precambrian area. Instead, distal (unit Xgwd) and proximal (unit Xgwp) facies are shown on the map west of the Grand Junction fault. Because the fault displacement is unknown, the units may not correlate with the other graywacke units as shown on map A. The youngest distal unit was originally named the Mayo Formation (Redden 1963); an older unit consisting of both proximal and distal facies was named the Bugtown Formation (Redden, 1963). The proximal facies consists virtually entirely of thick Bouma A beds. The Bugtown and Mayo Formations are separated by a thin unit of alkalic tuff and titaniferous basalt (Crow Formation), which is correlated with unit Xby. Magnetic data and the inferred contacts based on those data (map D) suggest the titaniferous basalt (Crow Formation) extends northwesterly beneath Phanerozoic rocks on the west side of the Bear Mountain dome, where it probably joins the equivalent tuff and shale (unit Xts) exposed near Deerfield. If the general correlation of the alkali tuff and titaniferous basalt is correct, a disconformity may exist with or within the underlying graywacke (Bugtown Formation).

Phanerozoic Rock Units

The units shown on map A are standard in most respects, and specific descriptions of rock units are not repeated here. The following comments emphasize only the differences noted in preparation of the map based on new observations (Redden, 1996) and various theses from students at the South Dakota School of Mines and Technology (Bush, 1982; Daly, 1981; Kleiter, 1988; Rawlins, 1978). Quadrangle names, place names, and section-township-range designations are noted in figure 1 for convenience in locating areas mentioned in the text. Place names that are also quadrangle names are not identified separately.

Paleozoic Sedimentary Units

Although Darton (Darton and Paige, 1925, p. 5) recognized that the Lower Ordovician and Upper Cambrian Deadwood Formation thins to the south along the Black Hills, map A indicates a complete pinchout of the unit for about 3 km along the southeast edge of the Precambrian (T. 4 S., R. 6 E.) where the Englewood Formation overlies, at least locally, low-dipping Precambrian basement quartzite. Another variation, which may not be readily apparent at the scale of map A, is the local relief on the nonconformity, which is reflected in thicknesses and lithologic changes in the Deadwood Formation. Early work suggested that the maximum relief was only about 30 m, but relief of approximately 90 m exists in the Nemo area, where resistant Precambrian quartzite formed hills that were engulfed by the Cambrian seas onlapping from the north (Redden, 1987a). In general, resistant Precambrian rocks throughout the Black Hills form highs underlying thinner sections of Deadwood. Lensoid basal conglomerates are noted locally on the slopes of these highs and are considerably higher stratigraphically than the discontinuous basal conglomerates found in old topographic lows.

The Ordovician rocks of the Winnipeg Formation and Whitewood Dolomite are eroded to a feather edge approximately along an east-northeast-trending line through Pactola Lake. A few meters of Winnipeg probably extend farther south but the Winnipeg is largely concealed. The younger Whitewood persists, although discontinuously, to just south of Little Elk Creek on the east side of the Precambrian core and to about the middle of the Nahant quadrangle on the west.

In the Fourmile, Pringle, and Cicero Peak quadrangles, the Upper Devonian and Lower Mississippian Englewood Limestone is underlain by a few meters of very well rounded, unfossiliferous sandstone, which, although included with the Deadwood Formation, could have formed after the major Silurian erosional break.

Depressions filled with reddish-brown sandstone at the top of the Mississippian Pahasapa Limestone mark local well-developed karst topography along the disconformity below the Pennsylvanian and Lower Permian Minnelusa Formation, a condition common in the Western United States. In contrast, above pre-Minnelusa topographic high areas, such as between Rapid City and Piedmont, one may find limestone beds of the Minnelusa in outcrop-scale conformity with underlying Pahasapa Limestone. About 6 km southwest of Pringle a pinching out of lower Minnelusa subunits against the Pahasapa Limestone indicates as much as 60 m of relief along the disconformity. This is the largest erosional high exposed along this contact in the Black Hills.

The Minnelusa Formation more than doubles in thickness from less than 120 m in the northeastern Black Hills to more than 350 m in drill holes in the extreme southwestern part of the map area. Surface exposures of the upper Minnelusa in the southern part of the map area have extensive breccias produced by solution of as much as 80 m of gypsum and anhydrite (Bowles and Braddock, 1963). The solution results in considerably thinner total thicknesses at the outcrop as compared to thicknesses obtained in drill holes several kilometers downdip. This solution and collapse was initiated in the early Tertiary but is known to be continuing today in areas south of the map area (Tim Hayes, oral commun., 1997). The gypsum and anhydrite beds of the upper part of the Minnelusa are thin or absent in drill holes along the northeastern part of the map area.

Mesozoic Sedimentary Units

Triassic rocks are limited to the predominantly soft red beds of the Spearfish Formation, which underlies most of the well-known “Red Valley” rimming the Black Hills. The Upper Jurassic Unkpapa Sandstone and Morrison Formation are shown as a single unit that is as much as 110 m thick along the eastern side of the Black Hills. In this area the thickness of the eolian Unkpapa generally has an antithetic relation with that of the Morrison. The Unkpapa pinches out by intertonguing with the uppermost Morrison a few kilometers northwest of Sturgis and also pinches out south of the map area at approximately the north-south axis of the central Black Hills uplift. The Morrison is only about 25 m thick in the southwest corner of the map area.

The Lower Cretaceous fluvial Lakota and Fall River Formations of the Inyan Kara Group form the major hogback beyond the Red Valley but are difficult to separate on aerial photographs. The contact between the two generally was placed just below the uppermost massive sandstone. The Fuson Shale of Darton and Paige (1925) is probably included with the Lakota Formation on the map. On the basis of field checking and well logs, it is evident that the Fall River is thinnest between Rapid City and Sturgis, where there is apparently an intertonguing contact between thin beds of iron-stained concretionary sandstone and black shale of the overlying Skull Creek Shale.

The remaining Cretaceous section is characterized, in general, by nonresistant units and poor exposures. Therefore, map contacts are only approximately located. N.H. Darton mapped numerous conical knobs in the Pierre Shale that are underlain by small bulbous lenses of fossiliferous limestone as teepee buttes. Only a few of those shown on the map have been field checked for location. Most locations shown on map A were made by inference from modern topographic maps rather than by direct transfer from the Darton and Paige map. The alignment of some in linear traces supports an initial origin along joint-controlled submarine vents. Such structures are widespread throughout the Cretaceous Western Interior Seaway where they formed from vented methane-rich fluids (Kauffman and others, 1996).

Tertiary Sedimentary Units

Subdivision of the upper Eocene and Oligocene White River Group (unit Tw) into the Chadron and Brule Formations, commonly used in the Badlands area to the east, could not be used here because of facies changes near the Black Hills. Darton and Paige (1925) used a general separation of boulder gravel from finer gravel and clay to distinguish between deposits labeled Quaternary gravel and White River deposits on their geologic map. Unfortunately, this criterion is not reliable in areas at higher elevations. The small area of White River shown about 2 km west of Norris Peak (T. 2 N., R. 6 E.) is covered by boulder gravel, but a road-cut exposure of pinkish clay and clasts also contains fossil material that indicates it is White River (Philip Bjork, oral commun., 1985). Elevation of the outcrop is only 54 m above the present level of Rapid Creek. Just south of this outcrop, at Johnson Siding, rhyolite boulders are noted in gravel about 18 m above Rapid Creek. Because no rhyolite is exposed in the Rapid Creek drainage, early Tertiary erosion could conceivably have extended to this level. At Hisega, a few kilometers to the south, documented White River is about 33 m above the present level of Rapid Creek. The elevation of the documented

White River indicates that the depth of the maximum erosion prior to deposition of the White River was, in many areas, very similar to the present depth of erosion on Precambrian and younger rocks. Darton and Paige (1925) correctly identified White River deposits in the lowest part of the valley in the town of Lead and close to the bottom of Fourmile Creek south of Custer. But elsewhere White River deposits can crop out at considerably lower elevations than those originally shown on their map. Consequently, numerous gravel deposits along Rapid Creek and southwest of Rockerville are interpreted as White River although the exposures lack specific lithologic or fossil evidence of being White River. Such deposits cannot be accurately identified on the basis of surface material alone. Darton and Paige (1925) showed a few small Tertiary deposits above the Minnelusa Formation exposures in the Black Hawk and Nemo quadrangles. These consist largely of angular fragments from the underlying Minnelusa Formation and were not included with the Tertiary deposits on map A. However, elevations of the deposits indicate that it is likely that they are very close to the Tertiary erosion surface.

In the Fairburn area drill holes document an ancient White River “French Creek” drainage that was diverted to the northeast by the more resistant rocks exposed in the Fairburn anticline (Kleiter, 1988). The bottom of this White River valley is nearly 80 m below the elevation of French Creek where the two drainages cross about 3 km west of Fairburn. About 5 km north of Fairburn, the White River valley turns to the east-southeast and continues to the Cheyenne River, where it is equivalent to the Red River valley of Clark (1937).

Along Pleasant Valley (T. 4 S., R. 3 E.), southwest of Custer, Tertiary deposits directly overlie Precambrian rocks and are only 5–10 m above the present drainage incised across the Phanerozoic rocks. Redden (1963) described the evidence for extensive solution in the Pahasapa Limestone along this valley prior to the deposition of auriferous White River gravel deposits. In the Argyle quadrangle to the south, Bowles and Braddock (1963) and unpublished mapping by Redden documented Tertiary gravels in an area that overlies collapse structures in the lower Minnelusa Formation. The latter structures likely result from down-dip solution of the Pahasapa Limestone during the Tertiary. Also, numerous sink holes were noted in other soluble Phanerozoic rocks near known White River deposits. These observations suggest that before and during deposition of the White River deposits there was extensive solution of carbonate rocks and gypsum-anhydrite, at least locally, on the old erosion surface (Redden, 2000). Many of the caves in the Pahasapa Limestone near the White River erosion surface are likely to have developed or began to develop at that time. It is probable that this extensive period of solution correlates in a general way with the widespread development of a red oxidized zone below White River deposits overlying Cretaceous shales that Clark (1967) called the Interior paleosol. This weathered zone is locally nearly 90 m deep in the Cretaceous shales (King and Raymond, 1971). Although most Tertiary deposits within the Inyan Kara hogback are in flat interdivide areas, considerable evidence indicates that deeper channels, possibly as much as 100 m in depth, existed in several of the major water gaps (Redden, 1995a, b). As a final note, during mapping of the Custer quadrangle, distinctive weathered skarns were noted at the 6,200-ft elevation in the Needles area south of Harney Peak. These are suspected to have formed during White River erosion and deposition. Thus, some White River deposits may have existed at higher elevations than those that have been preserved.

Tertiary Igneous Units

The various Tertiary igneous or igneous-related units shown are predominantly alkalic iron-rich rocks, both saturated and undersaturated relative to silica, and have a considerable range in composition (Lisenbee and DeWitt, 1993). The rocks range in texture from aphanitic to phaneritic. Most textures are porphyritic, and flow structures are common. The intrusive forms include dikes, sills, laccoliths, small pipes, and irregular-shaped bodies that were emplaced at relatively shallow depths.

Several diatreme pipes (unit Tbx) contain tuff and other sediment layers plus an abundance of wall rock fragments or small and large foundered blocks of overlying sedimentary rocks (which are shown as the original sedimentary unit on map A where large enough). Two diatremes are localized along the inferred northern extension of the Benchmark fault. The largest known diatreme, immediately north of Brownsville (T. 4 N., R. 4 E.), contains rhyolitic pitchstone, bedded tuff, and also fossiliferous fragments of Cretaceous age, an assemblage that indicates a volcanic pipe or volcano existed there before unroofing of the Phanerozoic rocks in that area (Redden and others, 1983). Some of the other diatremes and breccia bodies are also likely to be sites of former volcanoes.

Most of the rhyolite consists of a cream to white groundmass and sparse phenocrysts of biotite, feldspar, and minor quartz. Phonolite bodies in Precambrian rocks tend to be markedly discordant with the trend of the host rock and form distinctive resistant outcrops in contrast to the more easily weathered, intermediate-composition rocks. Phenocrysts in the intermediate-composition rocks are commonly feldspar and hornblende. Phenocrysts of aegerine-augite are common in the more alkalic rocks.

Two major differentiation trends from olivine gabbro and alkali gabbro to phonolite or rhyolite are suggested (Lisenbee and DeWitt, 1993). Most of the igneous activity occurred between 58 and 50 Ma, shortly after the beginning of the Black Hills uplift (Lisenbee and DeWitt, 1993). A 54-Ma rhyolite in the diatreme near Brownsville that contains inclusions of fossiliferous Cretaceous rocks indicates approximately 1 km of Phanerozoic rock cover at that location during early Eocene time (Redden and others, 1983). However, nearby White River deposits indicate this cover was removed by late Eocene.

Quaternary Units

Only the more extensive Quaternary deposits are shown on map A. Alluvium covers the bottoms of virtually all of the major and minor stream valleys. Terrace deposits of sand and gravel tend to be abundant outside of the Inyan Kara hogback and at different elevations. They dominate the south-facing slopes of the major valleys and indicate southerly shifts in the major streams across the incompetent Cretaceous rocks. Tertiary deposits are essentially all removed from interdivide areas north of Rapid City. This and other data on the distribution of Tertiary deposits in the source areas at higher elevation suggest that the northern part of the central Black Hills has been elevated 80–100 m relative to the southern part. This probably occurred during the Quaternary. North- and northeast-flowing streams, for example, Spearfish Creek, are incised to considerable depths below the old erosion surfaces compared to the southern Black Hills.

Glacial deposits are unknown, although gently dipping small slumps at higher elevations that involve coherent segments of Precambrian rocks likely indicate past periglacial conditions (Norton and Redden, 1960). Some small landslides are currently active during wet years. Most of the larger landslide deposits shown on map A occur on the north-facing scarps in interdivide areas underlain by the Cretaceous shales. These are a result of the southward shift in streams previously mentioned. Minor slides too small to show on the map are in the Jurassic rocks on the outcrop slope of the Inyan Kara hogback along the edge of the Red Valley. Virtually all of these are presently inactive. However, because of the extensive development of housing and other construction in the Red Valley, recognition of the potential for failure due to excavations in the Jurassic rocks is of special importance.

Proterozoic Plutonism and Metamorphism

The peraluminous S-type pegmatitic Harney Peak Granite (unit Xh) and associated pegmatites, which dominate the southern part of the map area, have been described in considerable detail by numerous authors, including Redden and others (1982), Duke and others (1990), Nabelek and others (1992), and Krogstad and Walker (1996). The main granite mass shown on the map is composite and consists of hundreds of separate intrusive bodies and many country rock inliers or screens. Most of the bodies are sills, some of which can be nearly 100 m thick, such as those visible on the aerial view (sheet 2). However, some pluglike masses, as well as dikes, are within the main granite mass. More than 24,000 separate bodies of pegmatite and granite crop out between the main mass of granite and the outer limit of pegmatite and granitic bodies in the surrounding country rock (Norton and Redden, 1990). Attitudes of the inliers, tabular lenses of granite, and textural layering within the granite define a dome that is reflected by the attitude of country rock around the main mass of granite (Duke and others, 1990). Emplacement of major sills creates noticeable protrusion of the granite, adjacent rocks, and earlier faults, especially in areas around Custer. Several hundred of the many outlying bodies of pegmatite are zoned, and a few of the larger bodies may have giant crystals such as those shown at the Etta mine (sheet 2). Very few zoned pegmatites are within the main granite mass. A regional zonation in the mineralogy of the zoned pegmatites also exists; pegmatites with lithium-rich and rarer minerals tend to be near the outer limit of pegmatite and granite bodies shown on map A (Norton and Redden, 1990). Country rock inliers in the center of the dome are the same rock types and sequence as the older Early Proterozoic units found at Bear Mountain. The main body of Harney Peak Granite is undeformed. A few thin sills within country rock marginal to the main granite mass or near the core of the dome are weakly flow foliated as shown by oriented biotite.

The main granite consists largely of plagioclase, perthitic microcline, quartz, and muscovite. Biotite is relatively common in the central part of the main mass but largely lacking in pegmatitic masses. Tourmaline is most abundant near the margin of the granite and in pegmatitic bodies throughout the granite. Monazite from a somewhat foliated biotite-bearing sill near the southern contact of the main mass has a concordant U-Pb and Th-Pb age of 1,715±3 Ma (Redden and others, 1990). Isotope data indicate that the Harney Peak Granite formed by partial melting of both Proterozoic and Archean crust (Krogstad and others, 1993). The central part of the granite may have a greater component of Archean crust than the margin. The inner, biotite-bearing part of the granite has lower $\delta 18O$ concentrations compared to the outer, more tourmaline-rich part (Nabelek and others, 1992).

Metamorphic events include an earlier, generally lower grade regional metamorphism and a later, high-temperature/low-pressure metamorphism associated with emplacement of the Harney Peak Granite and other similar but largely unexposed granites. It is likely that the two events were not widely separated in time, and the earlier may have led to the later. Except for the kyanite isograd around the Bear Mountain dome, the isograds shown on the map do not differentiate between the two events.

The earlier, higher pressure regional metamorphism is indicated by early kyanite at the Bear Mountain dome where data by Terry and Friberg (1990) indicate a clear pressure-temperature-time (PTt) loop caused by superposition of lower pressure Harney Peak-type metamorphism at Bear Mountain on the earlier regional metamorphism (Friberg and others, 1996). Large euhedral cordierite and andalusite porphyroblasts found along the edge of the Bear Mountain dome indicate Harney Peak Abukuma-type metamorphism (lower pressure, higher temperature). However, tectonic deformation of the porphyroblasts in some exposures indicate continued late deformation during or after doming. Muscovite in boudins on the Bear Mountain dome has a Rb-Sr age of 1,680±25 Ma (R.E. Zartman, *in* Ratté, 1986). This age, as well as the deformed porphyroblasts, supports multiple metamorphic events, the latest one coincident with formation of the Bear Mountain dome and emplacement of Harney Peak Granite at depth as shown diagrammatically in cross section *C–C′*.

Earlier, largely low grade regional metamorphism affected most of the central and northern parts of the Precambrian rocks, where a predominant north-northwesterly striking foliation exists. Whether the distribution of the garnet isograd everywhere in the map area is related to earlier regional metamorphism or to the Harney Peak event is uncertain. In the Rochford area the garnet isograd coincides in a general way with a major antiform that may predate the Harney Peak Granite. This coincidence suggests that garnet development there was related to the development of the structure during the regional metamorphic event. However, the garnet isograd and nearby staurolite isograd along the northeastern and eastern edge of the Precambrian exposures indicates increasing metamorphic intensity to the northeast below Phanerozoic rocks. Here this increase is believed to be due to emplacement of granite bodies similar in composition, but possibly slightly younger than, the Harney Peak Granite. Pegmatite similar to bodies around Harney Peak is present as rafts within Tertiary quartz trachyte on Whitewood Peak, about 3 km northeast of Deadwood. Farther northeast, near Crook Mountain (T. 5 N., R. 4 E.), a granite similar in composition to the Harney Peak Granite was drilled (Homestake Mining Company personnel, oral commun., 1984). A model age using the intial $^{87}Sr/^{86}Sr$ ratio of the Harney Peak Granite (0.7189, DeWitt, unpub. data, 2006) of 1.68 Ga is calculated. Model ages as young as 1.60 Ga can be cakulated assuming an initial ratio as high as 0.724. A small body of pegmatite is also present above the staurolite isograd in the Galena area (T. 4 N., R. 4 E.). Similar undeformed pegmatitic granite was found by drilling through Phanerozoic rocks just north of Nemo (Stan Ellingwood, oral commun., 1978). The presence of the garnet isograd near the eastern limit of the Precambrian exposures at Rapid Creek (T. 1 N., R. 6 E.) also suggests additional concealed Proterozoic granite to the east.

Metamorphic isograds in the southern part of the Precambrian area conform in a general way to the distribution of the Harney Peak Granite as indicated by the outer limit of pegmatite bodies shown on map A. The garnet isograd in the central Precambrian area is based on the general first appearance of garnet and also pseudomorphs believed to be altered garnet. The pseudomorphs tend to have a somewhat wider distribution than unaltered garnet, but the isograd conforms in a general way with calc-silicate assemblages of grossularite and amphibole found in metamorphosed concretions in the graywacke units. Below the garnet isograd the concretions retain a carbonate mineralogy. Because the deformation of the concretions is believed to be related to the development of the regional foliation, the garnet isograd in the south-central part of the area is believed to be due to the regional metamorphism.

The staurolite isograd nearer the Harney Peak Granite conforms more closely with the distribution of pegmatite bodies. In addition, porphyroblasts of andalusite and cordierite have a spotty distribution within the staurolite isograd, prompting Helms and Labotka (1991) to draw an andalusite isograd near the staurolite isograd shown on map A. Between the staurolite and sillimanite isograds, but closer to the staurolite boundary, are local areas where the original foliation is considerably modified by late mineral growth, including biotite. In extreme examples the original foliation is nearly destroyed. In other areas, generally nearer the Harney Peak Granite, a younger schistosity is developed that is believed to be related to the doming and emplacement of the hundreds of intrusive bodies that make up the central area of Harney Peak Granite. Some of the intrusive bodies locally have a flow foliation parallel to their contacts. Calculated pressures near the main granite body and over most of the higher grade metamorphic rocks are only 3–4 kb; temperatures exceed 650°C above the high sillimanite zone (Friberg and others, 1996). However, pyroxene-bearing screens from stratigraphically lower country rock exposed in the central part of the Harney Peak dome formed at much higher pressures and apparently were diapirically emplaced, as discussed in the later section on structure. Textural evidence indicates that most of the present porphyroblasts in the rocks surrounding the granite are relatively late and related to emplacement of the granite. Some large sillimanite knots are deformed into flat discs indicating growth prior to completion of the granite emplacement. Local replacement of porphyroblasts by mica in areas within the outer limit of intrusive bodies indicates continued fluid movement through the country rock during emplacement and crystallization of the many bodies of Harney Peak Granite.

Preservation of earlier, regional, higher grade metamorphic minerals is apparently limited to the Bear Mountain area and adjacent areas to the south. Locally, schist around the Harney Peak dome has unidentifiable relict metamorphic minerals indicative of an earlier, presumably regional metamorphic event that took place at higher pressure than the metamorphism related to emplacement of the granite. Although exposure of these earlier higher(?) grade metamorphic minerals is likely due at least in part to later doming, it seems that the earlier regional metamorphism was more intense to the south.

Metamorphism associated with the emplacement of the main mass of Harney Peak Granite was accompanied by extensive metasomatism in the surrounding country rock. The resultant mineralogic and textural data indicate a complex system probably resulting from a prolonged interval of plutonism. Elements added to the wall rock around granitic and pegmatite bodies include Li, K, Rb, Be, and Cs, as well as B. Analyses of approximately 6,000 grid samples of country rock over 80 km^2 along the north and northeast side of the granite indicate lithium was introduced into the country rock at least as far as 5 km from the main granite (Redden and Norton, 1992). Rubidium and cesium are also added in lithium-enriched areas. These alkalis do not increase uniformly toward the granite contact but instead are concentrated in “highs,” or anomalies, whose central area may have as much as or more than 10 times the average lithium concentration of metagraywacke far from the main Harney Peak Granite. These "highs" tend to be relatively small, ranging from about 0.3 to 1.0 km^2. Because extensive alkali-enriched haloes are known around individual zoned pegmatites in the Black Hills (Tuzinski, 1983; Shearer and others, 1986), the anomalies are almost certainly due to concealed (or in some cases exposed) bodies of pegmatite and granite satellitic to the main Harney Peak Granite. Similar, much smaller haloes are known around small quartz-graphite veins (Duke, 1995). Because the individual intrusive bodies tend to increase in number and size toward the main granite contact, there is a general increase in the average lithium concentration of the country rock near the granite. The anomalous lithium-enriched areas also tend to have coarser grained metamorphic minerals, and in some areas higher metamorphic grade minerals than those in the surrounding area (Redden and Duke, 1996). The texture and mineralogy suggest that the anomalies represent areas of enhanced fluid flow. The addition of Li, Rb, Cs, and B indicates that most of the fluid must have been derived from the crystallizing granite and pegmatite (Redden and Norton, 1992; Duke, 1996). However, the average lithium concentration of the main granite is considerably less than that of the surrounding schists according to data of J.J. Norton (oral commun., 1989).

Although local potassium metasomatism adjacent to pegmatite bodies such as pictured on sheet 2 has been well documented (Page and others, 1953; Redden, 1963), no analytic study testing regional potassium metasomatism has been done. However, indirect mineralogic evidence based on the presence of extensive areas of schist containing biotite-rich spots representing earlier garnet and staurolite suggests large-scale introduction of potassium. The spotted schists are especially well developed in the tuff and shale unit (Xts) on the north side of the main granite mass, where the spots are deformed by a late, low-dipping foliation (S_4). This late foliation is restricted to schist relatively close to the main granite contact. Potassium metasomatism is inferred because only a few percent of post-metasomatic garnet and staurolite were able to develop, in contrast to much greater concentrations of earlier garnet and staurolite (fig. 3). The added potassium effectively locked up initial iron and magnesium in biotite and some of the aluminum in newly developed muscovite. A similar addition of potassium is required to explain the complete absence of almandite in schist derived from turbidite

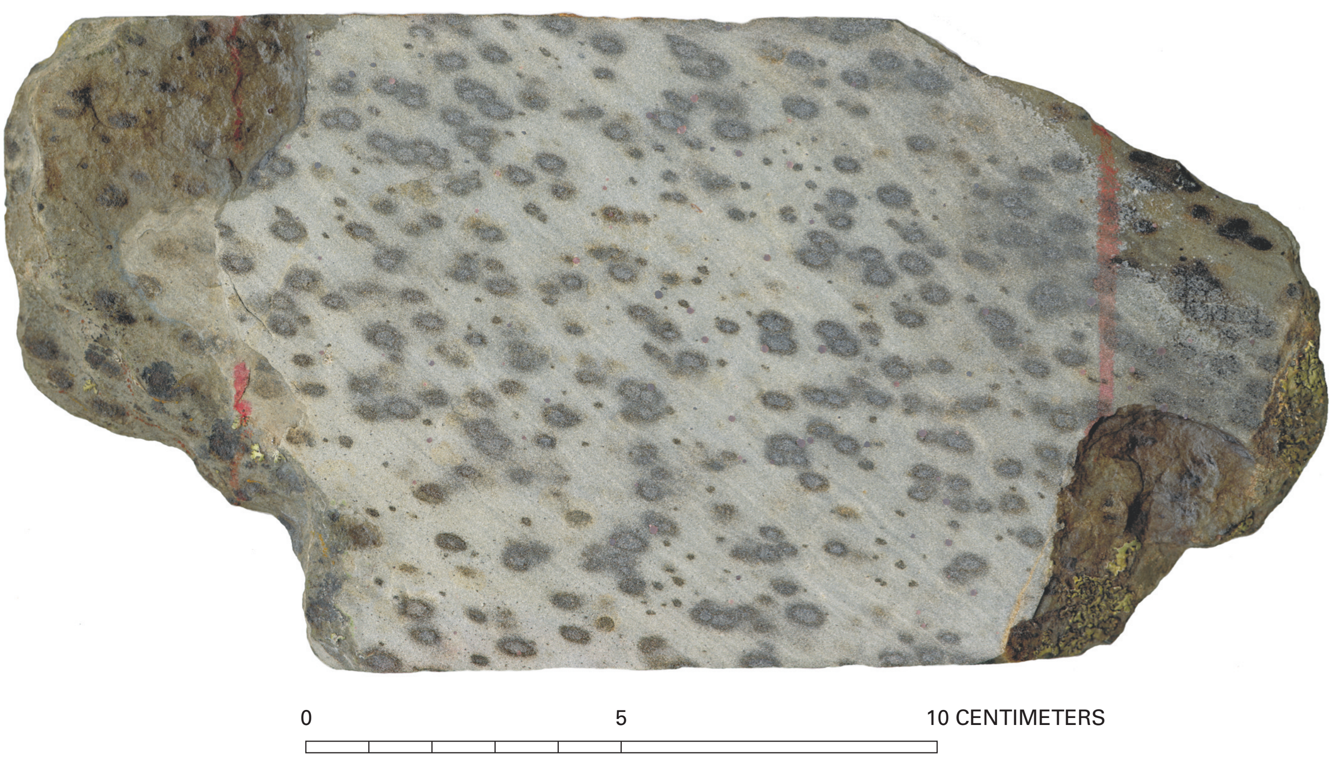

Figure 3. "Spotted" muscovite-biotite schist from the mica schist unit (Xts) about 3 km north of the main Harney Peak Granite mass. The larger dark, zoned oval forms of various cross-sectional shapes that make up about 15 percent of the rock are former staurolite porphyroblasts replaced by biotite rims and nearly pure quartz cores. About 0.2 percent of young garnet crystals have formed after the staurolite was altered. Small dark spots consist of mainly biotite and are probably former garnet. Some samples of spotted schist also contain traces or a few percent of younger staurolite and garnet. The replacement by biotite is approximately contemporary with development of the S_4 foliation. The major reduction in the volume percent of garnet and staurolite results from potassium metasomatism and the formation of biotite utilizing iron and magnesium of the earlier garnet and staurolite. The polished surface is subparallel to the S_4 foliation associated with the Harney Peak and Bear Mountain domes.

Bouma DEF sequences near the main granite contact. Small metamorphosed concretions within the associated Bouma A beds have their typical zoned mineralogy of hornblende, grossularite, diopside, and calcic plagioclase but apparently were essentially impermeable to the introduction of potassium.

In summary, the Harney Peak Granite and associated pegmatite represent a complex system involving repeated magmatic emplacement, development of alkali-rich fluids during crystallization, and metamorphism. Textural data of the adjacent metamorphic rocks indicate repeated metasomatism related to the emplacement of granite and pegmatite. It is likely that these processes extended over a considerable period of time.

Precambrian Structure

Structural detail in the small areas of Archean metasedimentary rocks is relatively limited, as is information about the development of the structure (table 1). In the Little Elk Creek area northeast of Nemo, Gosselin and others (1988) described northeast-striking bedding and an earlier foliation in metasedimentary rocks intruded by the Little Elk Granite and similar strikes within large inclusions in the granite. A relatively well developed, nearly vertical, northwest-striking foliation in the granite is defined by large oriented feldspar augen, but the foliation is absent in northeast-striking inclusions within the granite. The northwest strike is similar to the younger regional foliation strike in nearby Proterozoic rocks. The feldspar augen foliation is therefore inferred to have formed syntectonically with the granite emplacement (flow foliation), but was later overprinted by the regional Proterozoic foliation.

Folds within large clasts in the younger earliest Proterozoic rocks suggest that Archean deformation was probably widespread. At Bear Mountain any minor Archean folds that may have been present seem to have been molded into approximate conformity with the unconformably overlying Proterozoic rocks during doming similar to examples described by Ramsay (1967).

Stratified Early Proterozoic rocks have a complex structural evolution that began about 2,500 Ma with rifting of the Archean crust and ended sometime after 1,700 Ma with uplift and cooling of the Harney Peak Granite. The resulting sedimentation, igneous activity, and tectonic events are summarized in table 1. The structure of some areas is relatively straightforward, but in others there is considerable uncertainty in the interpretation of the rock pattern. Bedding and foliation attitudes are mostly very steep except where younger domal structures have modified earlier structures. Most of the steep bedding attitudes are apparently the result of later deformation so that even where younging structures or stratigraphic sequences indicate a significant fold, the predominant structural attitudes in available outcrops do not necessarily confirm the primary structure. Plunges of minor folds also typically vary considerably and may reverse in a single outcrop. All these features indicate multiple structural events. Accurate projection of contacts and folds to any significant depth is generally questionable. Because of the overall complexities and space limitations, the structures of only a few areas are discussed in some detail.

Folds and Foliation

The earliest Proterozoic folding is known to have involved only the oldest Proterozoic rocks (units Xbc, Xbcg, Xbcq, Xbcs, Xbi, and Xbd) in the Nemo area. These units were folded prior to the development of an unconformity below units Xec and Xeq as shown diagrammatically in figure 2. Both the older, pre-unconformity folded rocks and the post-unconformity rocks were subsequently overturned to the north. Because of the steep dip and plunge of the folds, the map view closely resembles a cross section. The timing and mechanism of the overturning relative to later Proterozoic structure are unknown. These overturned rocks were also deformed later and a nearly vertical, north-northwest-striking foliation developed, which is apparently equivalent to similar, generally penetrative foliation found throughout the Precambrian rocks of the central Black Hills. If we assume that the general surface of the unconformity developed prior to the deposition of the Xec and Xeq units was approximately a level datum, and if the surface is unfolded and restored to a subhorizontal position along an east-west axis, the early folds would have had northerly trends and low plunges (Redden, 1987b). The axial traces of these folds are not shown on the map, but the folds would be approximately coaxial with the post-unconformity folds (F_2), which are shown on map C (sheet 2). No foliation related to these earliest folds has been recognized. Two minor folds having northeast trends are also known, but these may represent early slump structures.

Excluding the restricted folding event just described that involved only older Early Proterozoic rocks, Archean and Early Proterozoic rocks of the main Black Hills basin were also affected by multiple deformations of varying intensity that for clarification are divided into five events (D_1 through D_5). Two major regional deformations (D_1 and D_2) are responsible for the major rock distribution and regional metamorphism (D_2). These were followed by a minor deformation (D_3), which locally produced minor folds and a foliation. A fourth event (D_4) consisted of igneous-related doming that strongly modified earlier structures in the Hill City, Keystone, and Custer area and was accompanied by largely thermal metamorphism. This doming was followed by a minor later event that locally produced a late foliation or spaced cleavage (S_5) and no significant folds. In addition, there are some younger minor outcrop-scale folds whose ages are uncertain. The various events are described in the following sections.

D_1 Event and Related Structures

F_1 folds include broad, regional-scale structures and also intermediate-scale folds indicated by the stratigraphic sequence or younging structures (maps A and C). Although intensely modified by younger events, the major folds are, at least locally, nearly isoclinal. However, the original attitudes of many of the F_1 fold limbs are generally unknown, and it is likely that they have been steepened considerably due to later deformation, especially where the F_1 trend is close to the trend of younger folds. Overall, the D_1 event was characterized by F_1 folds that are believed to have had originally north- to northeast-striking, relatively steeply dipping axial surfaces. The actual plunges of F_1 folds are generally unknown, but it is likely that most had relatively low initial plunges. It is difficult to determine the trends of the axial surfaces of F_1 folds due to lack of younging structures or detailed mapping. In such areas the axial trace is shown ending in a query on maps A and C although the fold likely continues.

At least three major F_1 folds are relatively well documented, although the exact location of their axial traces is poorly known. These include (1) the Slate Prairie syncline south of Mystic (T. 1 N., R. 4 E.), whose axial trace is contained within the shale (unit Xs) in the central Black Hills; (2) the adjacent Union Hill anticline to the southeast, which is cored by black shale (unit Xbs_2); and (3) the Rockerville syncline to the east (maps A and C). The axial trace of the F_1 Slate Prairie syncline is shown dipping southerly beneath the northern part of the younger Bear Mountain dome (D_4) in cross section *C–C′*. However, the trace of the axial surface east of the cross section is only approximately known. A minor overturned anticline cored by the Xts unit is also shown in the cross section, but the trace of its axial surface is not known because of the lack of detailed mapping to the south in an area where the axial surface should be exposed. This anticline is inferred to have a southeast plunge nearly parallel to the outcrop of the tuff and shale unit (Xts) because the overturned section at Deerfield Lake goes through a vertically dipping zone a few kilometers to the southeast of the lake and then into an upright (east-dipping) section equivalent to the upper limb of the anticline. A magnetic high area to the east (map D) suggests the plunge swings easterly below the thick shale (unit Xs) section. An easterly plunge at depth is indicated by the reappearance of the Xts unit at Redfern Mountain, which is interpreted as a refolded F_1 nose on part of the west limb of the Union Hill anticline, west and northwest of Hill City.

The Union Hill anticline is based on repetition of stratigraphic units and younging structures and was originally cited (Redden and others, 1990) as a typical example of a refolded F_1 structure. If the fold is followed from south to north and to the east, the axial surface apparently changes from a steep westerly dip to a moderately steep southerly dip north of Lowden Mountain but is essentially vertical where the fold trace is shown ending on maps A and C. Apparently the fold dies out to the northeast as shown, and the core rocks in a large area extending to the south beyond Hill City are repetitions due to later F_2 cross folding and doming on the upright limb of the anticline rather than a separate anticlinal nose as originally mapped by Ratté and Wayland (1969). The apparent core of older black shale (unit Xbs_2) can be followed to Pringle, where faults bound the unit as a major décollement. The extent of the apparent core strongly suggests that the structure was initially a major nappe-like fold having northward vergence. However, the plunge of the original nose of the structure and the trace of the axial surface are problematic. If the fold is a large nappe and its west (upper?) limb extends beneath the large area of younger rocks along Thompson Draw as is believed, the core should be expected to reappear at the north end of the Bear Mountain dome area. Possibly the fold nose suggested by cross section *D–D´* (sheet 2) could be the equivalent structure, but it would have a much diminished amplitude, which suggests it is unlikely to be the same fold. It is possible to draw subfolds in the nose area of the Union Hill anticline northwest of Hill City that retain the concept of an original, probably northwest trending nappe whose limbs were subsequently greatly steepened. However, additional detailed mapping will be required in key areas before the original structure is fully understood and axial traces better defined. Possibly the earlier interpretation (Redden and others, 1990) that F_1 folds were not strongly overturned before later deformation may ultimately prove to be correct and the fold was never a major nappe.

The Rockerville F_1 syncline differs from most other F_1 folds because a north-northwest trend over much of its length suggests that it has been warped by D_2. Only on its south end does it curve to the west and northwest. The extension of the trace of the syncline in the block between the Empire and Silver City–Blue Lead faults is very questionable as there is no obvious nose of younger rocks to the southwest. The general trend of the trace of the axial surface diagonally to the gross rock pattern may be due to intense later overprinting by north-northwest F_2 folds. Although minor F_1 folds are difficult to recognize, the reversal of fold patterns (Z versus S) on opposite sides of the north end of the small dome a few kilometers southeast of Keystone is consistent with small F_1 folds having approximately northeast trends in that area. This geometry would be in agreement with the general trend of the major syncline to the north. Southwest of Keystone and west of the Empire fault, the generalized trace of the syncline continues to the southwest until it encounters the dome around the Harney Peak Granite. Any continuation south of the dome is uncertain. The trends of intermediate-scale F_1 folds about 1.5 km west-northwest of Keystone are consistent with the major structure, although they also have been refolded. A few kilometers south of Rapid Creek, the trace of the synclinal axis is offset by two faults and apparently warped by younger folds. The map pattern suggests that the plunge of the fold may reverse for a few kilometers, but this apparent reversal could be a result of the faulting.

Just north of Rapid Creek and east of the Rockerville syncline, near an area known locally as Placerville, is the nose of a well-defined anticline cored by the pillow basalt (unit Xbo). The axial trace extends to the north-northwest before being offset by a right-lateral fault. The southeast extension of the fold is less accurately known because of extensive facies changes in the younger rock units. In the nose of the structure at Rapid Creek, thin, nearly vertical metagabbro sills outline the fold nose, and minor folds of presumably younger age have nearly vertical plunges. The distribution of the pillow basalt and overlying ferruginous chert (unit Xfc) indicates the F_1 fold plunges gently to the south. Younger units to the southeast may indicate a local reversal of the low plunge.

A major syncline should exist in the quartzite, siltstone, and shale (unit Xqs) east of the F_1 anticline at Placerville. However, the location of the trace of the fold is questionable as most known younging directions in that unit are to the west or southwest. Possibly an unrecognized early fault removed part of the west limb of the structure. Detailed mapping in the Pactola Dam quadrangle indicates that many closed doubly plunging folds are apparently due to younger folding. The map pattern near Hat Mountain northeast of Johnson Siding is also inconsistent with known F_1 structures and may require an unrecognized fault or faults to explain.

The inferred anticline passing south of Pactola Lake is also probably a major F_1 structure whose east limb was concealed or removed by the Pactola Lake fault. The fold axis is displaced by the right-lateral Silver City fault, and its extension west of the fault is interpreted to be about 7 km to the northwest. The extension of this southwest-trending anticline into the block separating the Silver City and Empire faults seems straightforward, but several F_1 folds to the east are anomalous. Near the Silver City fault the axial traces of these folds turn northeasterly, but when traced to the south their trend changes to the southeast. In addition, the fold closures indicate several of the folds are doubly plunging. The northeast trend near the Silver City fault is probably due to drag produced by movement on the right-lateral fault. However, the reason for the southeast trends roughly parallel to the Empire fault is unknown. Possibly the fold trends there are the result of D_3 modification of F_1 folds, or alternatively the F_1 folds were modified by some unrecognized event.

In the Rochford area the axial traces of several intermediate-size isoclinal F_1 folds form U-shaped patterns due to F_2 refolding (map C). Details of these structures are poorly known because of a general lack of younging structures and rapid facies changes in the volcaniclastic and related rocks of unit Xtv. The simplest structure is an innermost F_1 anticline cored by unit Xqg that is warped into a fishhook pattern by D_2. Other intermediate-scale F_1 folds show somewhat similar patterns but locally may have been modified by D_3. Several kilometers south of Rapid Creek the older rocks are shown as the core of a very irregular, westerly trending, large F_1 anticline that is assumed to be more or less parallel to the Slate Creek syncline farther south. The trace of the axial surface of the anticline is inferred to extend northeast of Mystic across a right-lateral fault, but the older rocks there could be in a separate fold. Isolated exposures of basaltic flows (unit Xby) along Castle Creek are probably surrounded by pelitic parts of unit Xtv, but because of the lack of detailed mapping in some areas only a thin unit of Xtv is shown on map A. Detailed mapping of some of these isolated areas cored by metabasalt (unit Xby) and adjacent phyllite indicate steep, doubly plunging structures resembling sheath folds. It is inferred that they were originally minor F_1 folds that were subsequently deformed. Another refolded F_1 syncline is shown along the east side of the Rochford area and extending to the northeast. This syncline is also shown extending across the south end of the central Rochford structure and westward into the large area of younger shale (unit Xs) west of Rochford. However, the location of the axial trace through this shale is uncertain. The fold may be equivalent to the north-northwest-trending F_1 syncline shown on the west side of the main Rochford structure. However, that structure is parallel to D_2 structures and may not be an original F_1 fold.

In the Roubaix area (T. 4 N., R. 3 E.) several apparently early F_1 folds follow irregular, inverted U-shaped patterns, which suggest both northeast and northwest crossfolding (Krahulec, 1981). The original trend of the F_1 folds is uncertain.

Generally, minor F_1 folds are difficult to recognize, especially where overprinted by similar-trending younger folds. Locally, they may be recognized because of warping and variation in plunge due to younger deformation. About 3 km north of Storm Hill (T. 2 S., R. 6 E.), on the Rapid City–Sheridan Lake Road, a well-exposed minor F_1 syncline shows no megascopic evidence of mineral lineations or penetrative fabric. The plunge of the synclinal fold has been rotated past 90°, and both limbs are cut by a north-northwest-striking, nearly vertical schistosity. Other minor F_1 folds about 1.5 km southwest of Rockerville along U.S. Highway 16 have much shallower plunges to the south-southwest, and the fold limbs are crossed by well-developed bedding or foliation lineations caused by the younger north-northwest schistosity. The plunge of an individual minor F_1 fold can vary more than 90° along the axis, and it seems that no noticeable metamorphism or strongly penetrative fabric accompanied D_1—at least in this part of the map area. The available evidence suggests that the D_1 event resulted from generalized west-northwest–east-southeast compression at a relatively high crustal level, at least for part of the map area.

D_2 Event and Related Structures

The D_2 event (maps A and C) is characterized by generally north-northwest trending, nearly isoclinal folds designated F_2. These folds vary in size and have a nearly vertical, generally axial-plane foliation (S_2). On the limbs of many tight F_2 folds, S_2 is essentially parallel to bedding, especially in beds of contrasting competency or in areas where the general trend of bedding is northerly, dips steeply, and no minor folds are recognized. Where large antiforms and synforms developed during D_2, there are generally recognizable left- and right-handed folds (S verses Z) on opposite limbs of the structures. However, these smaller F_2 folds may have irregular shapes and considerable variation in plunge. Isolated calcareous concretions in the graywacke units are typically deformed into triaxial ellipsoids having steep plunges (except in areas affected by D_4) parallel to F_2 minor fold noses (Redden, 1963, 1968). Larger clasts in the conglomerate and debris flow rocks are also flattened and elongated with generally steep plunges. The plunge associated with D_2 was apparently generally steep. The D_2 event and accompanying regional metamorphism apparently initially affected the entire Precambrian area, although later metamorphism and structural doming associated with the emplacement of the Harney Peak Granite locally modified and reoriented D_2 structures. The largest F_2 folds are the major antiforms and synforms shown in the central part of the Precambrian rocks (maps A and C). The clearest example of the effects of the F_2 folding is along the Lowden Mountain synform, which overprints the early Union Hill anticline and part of the Prairie Creek F_1 syncline north of Hill City. The thick, boomerang-shaped exposure of the tuff and shale (unit Xts) at Redfern Mountain in the southeastern part of the Rochford quadrangle results from the intersection of a minor F_2 antiform with a southwest-trending F_1 anticline on the overturned limb of the Union Hill F_1 anticline as described in the preceding section. The Lowden Mountain synform cannot be recognized to the south where modified by the Harney Peak Granite dome. The adjacent unnamed antiform to the east and associated minor folds are also warped to the southwest near Hill City. This warping is believed to be the result of a buried subsidiary granite dome that is inferred to be centered just east of Hill City. The eastern unnamed antiform may also extend farther north than shown, but data are inadequate to locate the axial trace.

The Rochford F_2 antiform trends northerly toward the Lead area but disappears beneath Phanerozoic cover. Bayley (1972c) referred to the structure as the "Rochford anticlinorium," but F_1 folds wrap around in the nose of the structure as described previously. North of Rapid Creek, dips are steep and the antiform is slightly overturned to the west. To the south, lack of uniformity of dips suggests an antiformal structure and the trace of the antiform is shown ending on maps A and C. Possibly this is due to proximity to the axial trace of the Slate Prairie F_1 syncline, whose axial surface is apparently dipping very steeply in this area. Farther south and essentially on trend with the Rochford antiform is the Thompson Draw synform, which is parallel in part with the Thompson Draw syncline of Ratté (1986). Dips are nearly vertical and locally overturned near the axial trace but are inward a few kilometers on each side. Farther away from the axial trace, inward dips are apparently due to the Harney Peak and Bear Mountain domes. However, it is uncertain how much of the Thompson Draw synform is due to these late domes and how much is due to F_2 folding. It is believed that the basic structure is largely a downwarp on the south limb of the Slate Prairie F_1 syncline due to F_2 folding.

A minor south-plunging F_2 synform is shown intersecting the northern part of the Bear Mountain dome at Deerfield Lake. The structure apparently produces a west-dipping, south-plunging synformal nose in the inverted section of graywacke on the overturned F_1 fold limb. The adjacent relatively small syncline to the west is considered to be an F_2 structure but may have been an earlier intermediate-size F_1 structure.

In structurally complex areas such as south of Rochford, F_2 closed folds cored by basalt (unit Xby) have somewhat modified conical geometry as previously described. Pi diagrams in figure 4*A*–*C* represent bedding, foliation, and fold plunges from a 3-km² area that includes some of these closed folds. The diagram in figure 4*A* shows the abundance of very tight F_2 folds in the schist surrounding the more massive basalt, and two foliations (S_2 and S_3) are recognizable in figure 4*B*. Minor fold plunges (F_2) shown in figure 4*C* merge in a single general maximum elongated to the north-northwest. This pattern contrasts somewhat with plunges of minor F_2 folds from areas including both limbs of major F_1 folds such as the Union Hill F_1 anticline northwest of Hill City. Diagrams of plunges of minor folds from such areas typically show a north-northwest–south-southeast girdle having a major maxima that indicates a moderate plunge to the south-southeast and a submaxima that indicates very steep plunges to the south-southeast. Such patterns confirm the superposition of D_2 deformation on F_1 folds that were originally overturned to the north or northwest.

The major bend to the west and northwest in the southern part of the Rockerville F_1 syncline is interpreted to be caused by a major F_2 structure. Where the structure crosses the Rockerville syncline, prevailing bedding attitudes indicate it is synformal. However, north of Spring Creek, dips are nearly vertical and there is no consistency to indicate a synform.
In the Rockerville area, as well as areas to the north having generally northerly trends, the angle between F_1 and F_2 was typically low and D_2 flattening was intense. As a result, some minor F_2 folds are tightly doubly plunging (fig. 5) and produce complex patterns due to interference with F_1 folds. In general, F_2 minor folds are difficult to recognize by map pattern alone where both F_1 and F_2 have north-northwest trends. However, most outcrop-scale folds tend to be of D_2 age, and the effects of D_2 are generally greatest where there are pronounced differences in competency between adjacent units.

The major structure in the Lead area is the nearly north trending Poorman anticline, which is overturned to the west and is probably an F_2 structure (Caddey and others, 1991). Another, smaller anticline (the Lead anticline) lies immediately to the east. Underground gold mining at the Homestake mine has traced the Lead anticline for about 8 km down plunge from surface exposures at Lead. The fold has an average plunge of about 40° to 35° SE., whereas the plunge of the larger Poorman anticline is almost due east at depth. Many very tight, smaller folds (generally not shown at the scales of these maps) modify the incompetent rocks overlying the older metabasalt (unit Xbo) in the core of the larger structure. Most of these small folds have essentially an axial-plane foliation, which suggests they are of F_2 origin. A few low-plunging folds in the mine area may have been earlier but seemingly do not indicate any major earlier F_1 structure. The larger north-

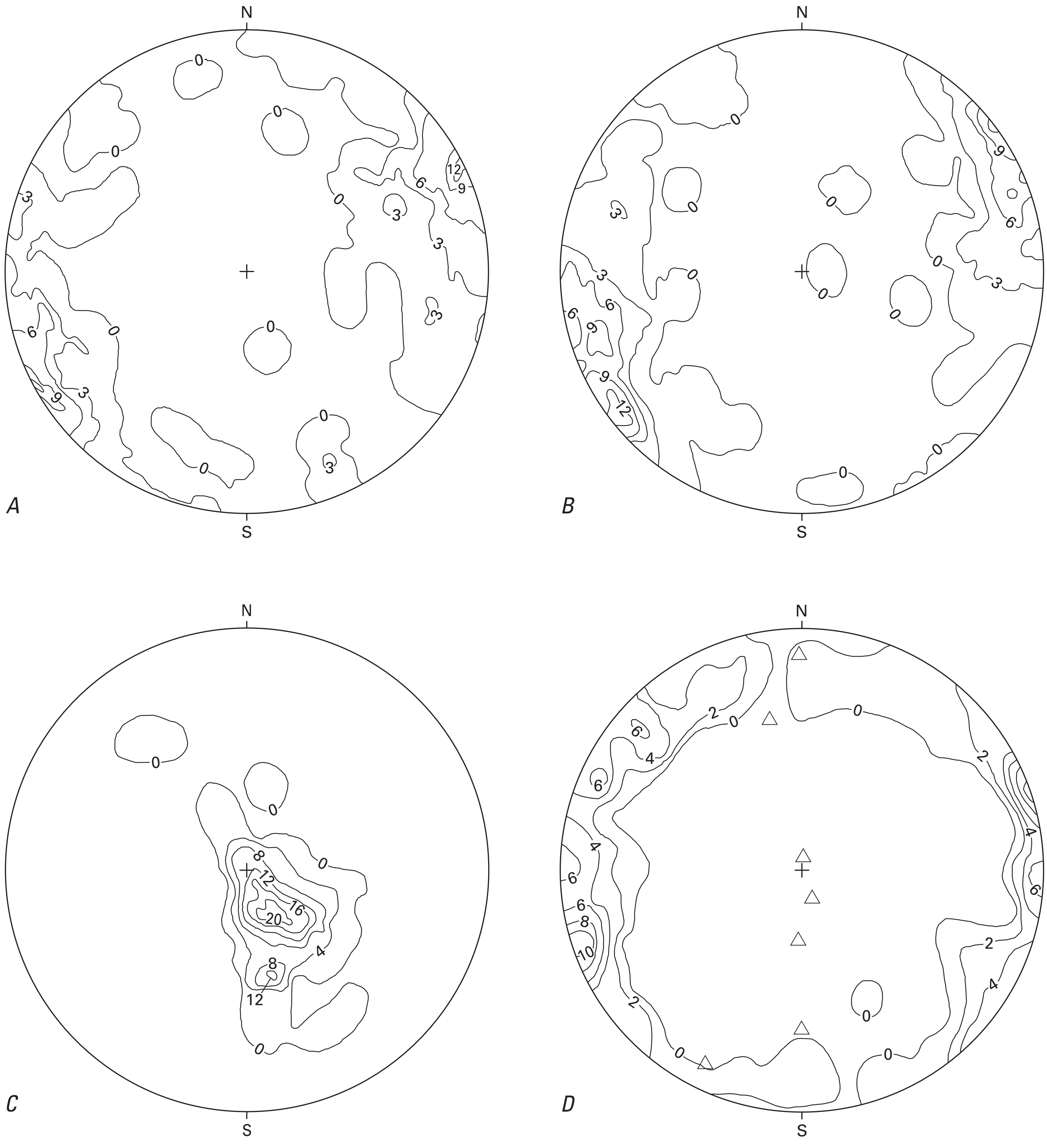

Figure 4. Structural diagrams illustrating bedding, foliation, and plunge of minor folds in selected areas of F_2 cross folds. All plots are lower hemisphere.

A. Plot of 112 poles to bedding. Data from sec. 31, T. 2 N., R. 3 E., and sec. 1, T. 1 N., R. 3 E., Rochford 7.5-minute quadrangle (McMillan, 1977). Contour interval 3 percent.

B. Plot of 150 poles to foliation in same area as diagram *A*. Contour interval 3 percent.

C. Plot of 53 fold plunges of mainly F_2 folds in same area as diagram *A*. Contour interval 4 percent.

D. Plot of 100 poles to bedding over 9 km^2 area in southeast corner of Silver City 7.5-minute quadrangle. Contour interval 2 percent. Triangles indicate plunges of minor F_2 folds.

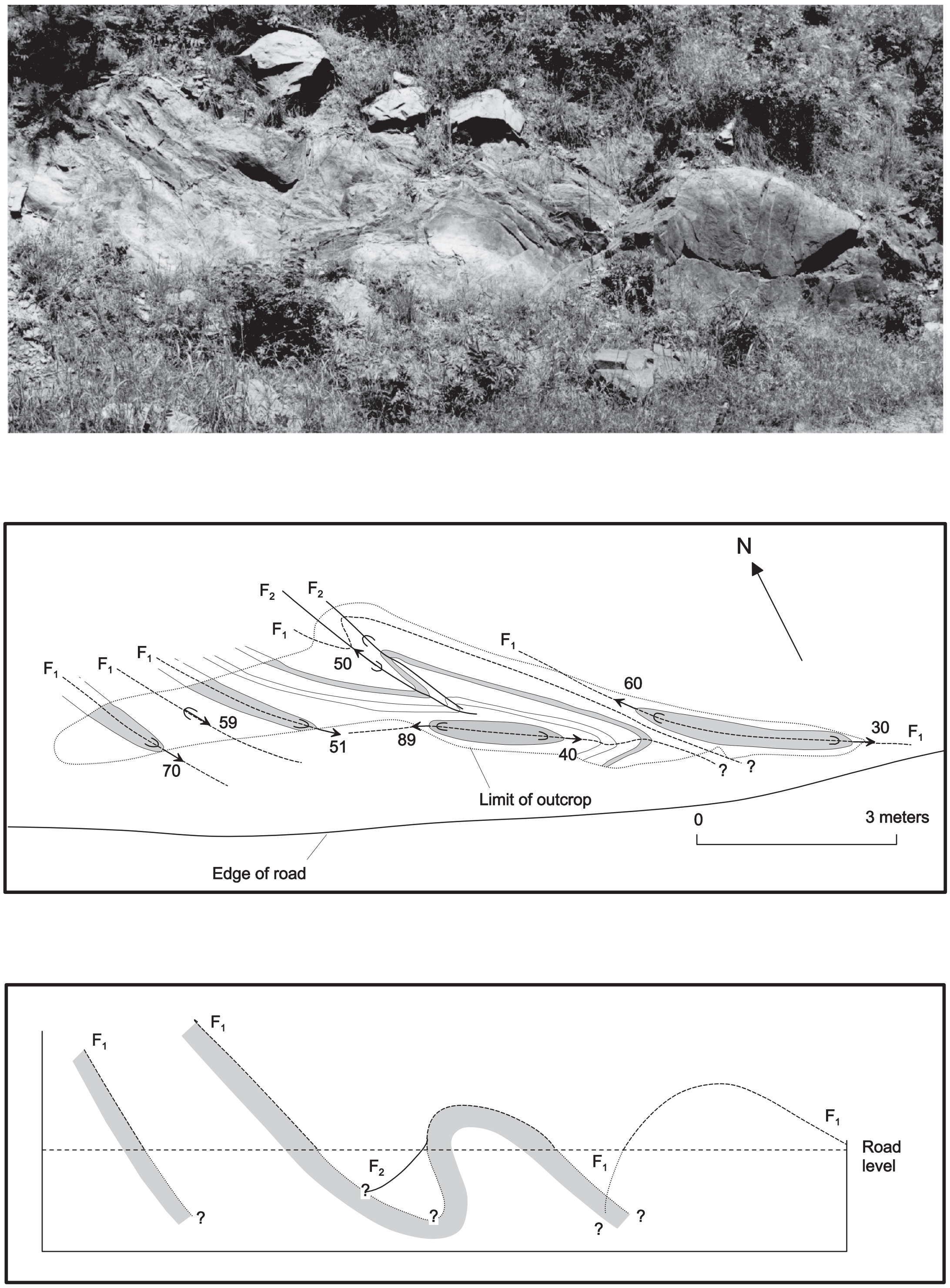

Figure 5. Photograph and reconstructions showing doubly plunging minor folds in unit Xqs along Rapid Creek near McGee (T. 1 N., R. 6 E.). Patterned areas in plan view below are two thin quartzite beds separated by mica schist and a third thin unpatterned quartzite bed. Although the arrows at the fold noses indicate plunge direction and amount, the anticline-syncline symbols may be erroneous because inconclusive data on younging suggest the section may be overturned. Most fold plunges in this area are gentle to moderate and trend southerly, but reversals in plunge may exceed 135°. Although the fold axes are designated F_1 and F_2, they may be essentially contemporaneous and likely of D_2 age. The two sets of folds are nearly coaxial, and extensive flattening of the folded competent-incompetent assemblage may have caused buckling along the axes of early F_2 folds. Plan and profile view modified from Fricke (1982).

trending folds west of the Lead area are interpreted on the basis of stratigraphic repetitions and in general have nearly vertical bedding and foliation that have a northerly strike. This trend is similar to the F_2 trends in the central part of the Rochford antiform to the south. Some of the larger folds in the Lead area may have been earlier F_1 folds having original trends similar to F_2 folds.

Foliation of S_2 ranges from a slaty cleavage to a well-developed schistosity depending on the metamorphic grade. Although it is generally penetrative, especially in the more incompetent rocks, it is absent or virtually absent in the more competent rocks such as large metagabbro bodies, thick basaltic flows, or thick beds of quartzite. The foliation may be present near contacts of the larger metagabbro bodies or in narrow zones within the metagabbro. Although the north-northwest trend of S_2 is relatively constant over many areas, it tends to follow bedding where the latter has somewhat similar trends. This is especially true in bedded rocks of differing competency. Also, the trend of the major foliation (S_2) is modified in several areas due to younger events. For example, in the northwestern part of the Nemo quadrangle and in the nearby Roubaix area, the main foliation basically follows bedding trends and gradually curves to a north-northeast trend. This change is believed to be due to late warping around a largely concealed block of Archean granite to the east, which is described in the “Geophysics” section. Warping of S_2, and other F_2 structures also occurs around the late domes cored by Harney Peak Granite. The foliation near the main Harney Peak dome is probably largely S_2 but it has been modified by the later metamorphism and utilized during the later deformation where appropriately oriented. Also, along some strike-slip faults such as the Silver City fault there has been considerable reorientation of S_2 and minor F_2 folds. The reorientation agrees with the apparent lateral movement along the fault and is apparently related to the deformation that produced the fault.

Several intermediate and larger scale folds in the area around and to the southeast of the Harney Peak dome are shown as F_2, but their age is poorly known. Because of distortion due to D_4, the original folds may have been of D_1 or D_2 age. Some possibly are D_4 structures. The northeast-trending Summit Peak anticline and Humbolt Mountain syncline southeast of Hill City are subparallel to other smaller F_2 folds to the northwest that are influenced by distortion due to D_4 doming. The Humbolt Mountain syncline is believed to have a gentle inward plunge although the continuation of the southernmost part is poorly known. The plunge of the northeast end of the Summit Peak anticline is unknown because of the superposition of probably the youngest F_4 folds such as shown in figure 7. About 5.5 km south of Hill City, an inverted, northeast-closing anticline outlined by graywacke unit Xgw_3 plunges gently to the southwest. This inverted anticline, although bounded on the east by the inferred St. Elmo fault, is believed to close about 8 km to the south where minor folds and elongate calc-silicate ellipsoids indicate a gentle northerly plunge. The inverted anticline is interpreted to be the nose of the Summit Peak anticline, which was overturned by a buried D_4 dome or possibly domes in the general Hill City area.

D_3 Event and Related Structures

The D_3 event apparently had a relatively minor influence on the larger map patterns. It is generally characterized by a poorly developed, nearly vertical cleavage (S_3) that strikes northwest. In many exposures it cannot be readily distinguished from S_2 and is either absent or is parallel to S_2. Detailed maps available for the compilation of map A generally show only a single type of foliation symbol, but Pi diagram plots typically reveal the two foliations (fig. 4*B*). Because of this ambiguity, S_3 is shown only in a few areas on map A. Map-scale D_3 folding (F_3) is sparse and shown largely in the Deerfield and Roubaix areas. In the latter area S_3 is especially well developed. Minor F_3 cross folds are noted in the Homestake mine area at Lead, where their pattern indicates a right-lateral shear across F_2 folds. Several minor folds south of Rochford have northwest trends, but it is uncertain whether they are F_3 or modified F_2 folds. Minor northwest-trending upright folds near the intersection of highways 16 and 385 northeast of Hill City have a nearly vertically dipping late axial-plane schistosity in a 0.5-m-wide zone along the axial plane. Some northwest-trending map-scale folds in that area may be of F_3 age. A similar late schistosity limited to the axial-plane areas of fold noses is known in canoe-shaped refolded folds southeast of Hisega, in the Pactola Dam quadrangle. No large folds of apparent F_3 age are known in the earliest Proterozoic rocks in the Nemo area. However, S_3 is locally present in the more competent rocks and a similarly oriented late foliation is known in the Archean Little Elk Granite. No F_3 structures have been noted in the higher metamorphic grade southeastern part of the Precambrian area where limited mapping has been done.

The D_3 event apparently resulted from general northeast-southwest compression. The apparent warping of F_2 structures around the inferred Archean block to the east of the Nemo area may have occurred during D_3. Although D_3 structures are oriented relatively close to D_2 structures in many areas, it is believed that the two events are distinct because of the local overprinting of a relatively good schistosity (S_3) on the earlier S_2. Possibly the age of D_3 was relatively close in time to the emplacement of Harney Peak Granite-like intrusions to the northeast of Nemo beneath Phanerozic cover.

D_4 Event and Related Structures

Late D_4 structures are apparently all associated with emplacement of Harney Peak Granite or equivalent granite and are restricted to exposed or inferred domal areas. A shallow-dipping schistosity, S_4, wraps around the Harney Peak and Bear Mountain domes. This schistosity is most easily recognized where the adjacent rocks have orientations markedly discordant to the domes, such as on the north end of the Harney

Peak dome and on the southeast edge of the Bear Mountain dome. Where adjacent rocks are subparallel to a D_3 dome, the predominant or only schistosity is subparallel to bedding or is axial planar to very tight asymmetric, presumably F_2 folds. This S_2 schistosity has been both spatially reoriented by the large Harney Peak D_4 dome and also recrystallized as a result of the accompanying metamorphism. On the concordant sides of the somewhat elongate major domes, boudins and necking structures indicate considerable thinning of the rock units due to extension parallel to, or less commonly perpendicular to, the average dip. Outcrops of shallowly dipping country rock along the southwest side of the main Harney Peak Granite have exceptionally abundant boudin structures that document the stretching of country rock over the Harney Peak dome. The metamorphosed graywacke shown in figure 6 demonstrates that this thinning or stretching occurred, at least in part, following the development of coarse sillimanite (fibrolite) knots. Duke and others (1990) cited evidence based on the

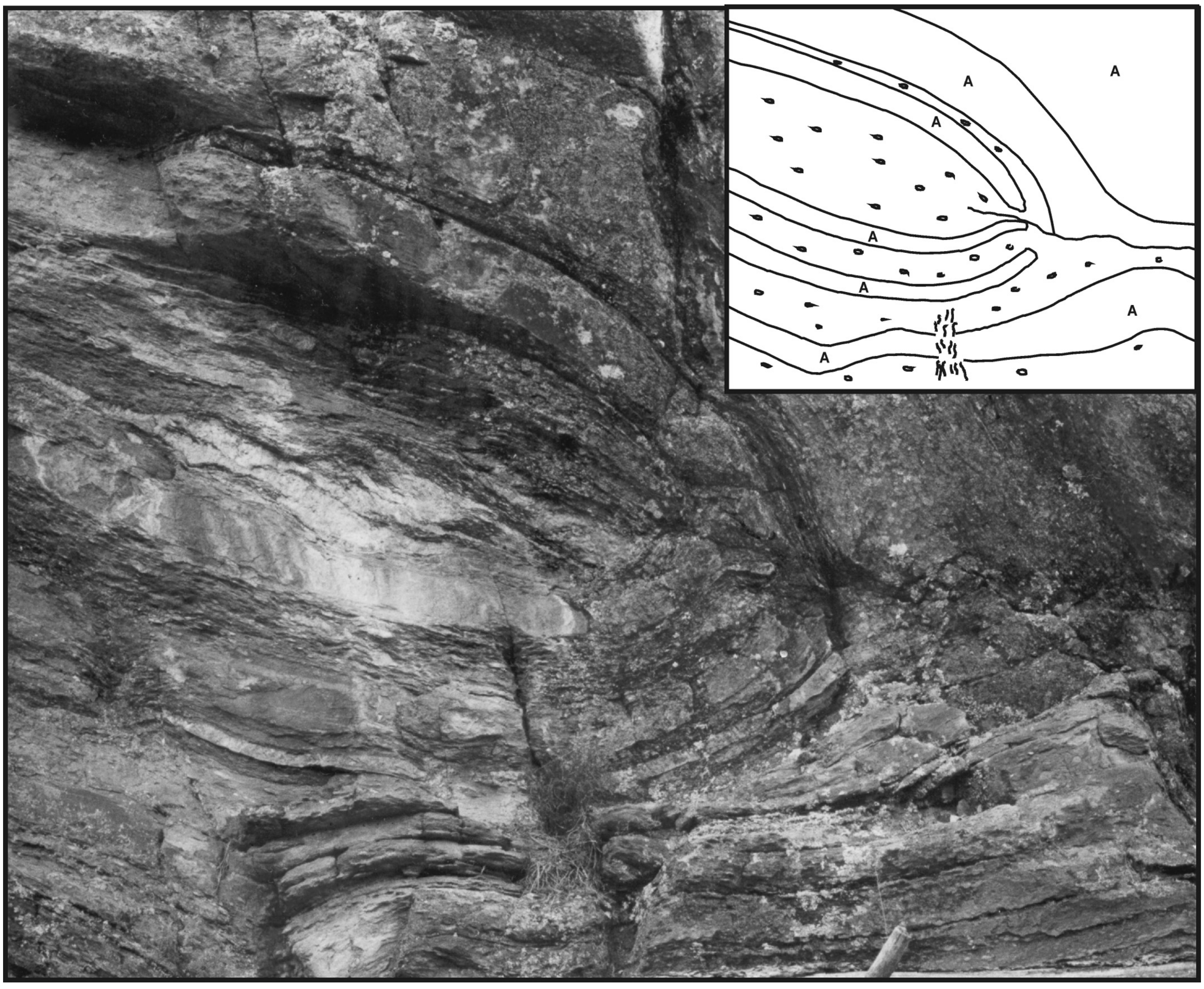

Figure 6. Photograph and drawing showing boudin development in metagraywacke on the shallow-dipping southern part of the Harney Peak dome, 6 km north of Custer along Willow Creek. In this area above the high-sillimanite isograd, the thicker units that boudin are predominantly or entirely biotite-rich schist containing large, somewhat flattened sillimanite (fibrolite) knots as much as 4 cm across that develop in Bouma DEF lithologies. The quartzose, but much finer grained Bouma A (symbol A in drawing) beds tend to flow into the boudin gaps. The rock behavior indicates boudin development was at a relatively late stage in the formation of the Harney Peak dome. Presumably, continued emplacement of granite effectively stretched the cover of the dome and resulted in much thinning of the stratigraphic units. Boudinaged sillimanite-biotite schist beds such as those shown in the figure typically do not reappear along strike at an outcrop scale of several meters.

development of a minor dome suggesting that the main Harney Peak dome developed in part by addition of hundreds of tabular intrusive bodies to the periphery of the central core of granite. This caused extensive inflation of the adjacent schist and development of S_4.

Small-scale F_4 folds associated with the elongate Harney Peak and Bear Mountain domes are asymmetric and consistent with upward movement of the central part of the dome, yielding a "Christmas tree" pattern in a regional cross section. Axial planes are low to moderate dipping away from the dome. Many small F_4 folds typically have a chevron geometry similar to that shown in the lower part of figure 7, but some are flattened isoclinally and have a well-developed S_4.

Another more deeply buried Harney Peak-type dome is inferred to exist in the Hill City area in order to explain the low dips and the bulge in the boundary line showing the outer limit of pegmatite and granitic bodies on map A. The middle parts of the Humbolt Mountain syncline and Summit Peak anticline have low northwesterly dipping axial planes and are shown as being of indeterminate age because they could be reoriented F_1 or F_2 folds that were presumably modified by D_4 doming. The young chevron F_4 recumbent fold (fig. 7) at Mitchell Lake, 2 km north of Hill City, trends at nearly right angles to the adjacent Humbolt Mountain and Summit Peak folds and has a poorly developed axial plane foliation (S_4) restricted to the fold hinge area. Bedding plane foliation is noted on the limbs. The exposures of turbidites shown in figure 8 are nearer the middle of a small dome about 5 km east of Hill City and show boudin structures that plunge at a low angle subparallel to the plunge of the recumbent fold in figure 7. These structures are believed to result from the general upward movement and elongation parallel to an earlier anticlinal fold due to granite emplacement. About 6 km south of Hill City minor late folds indicate upward movement to the north, which supports this concept. The D_4 domes resulted from diapiric intrusion of the 1,715-Ma Harney Peak Granite. Calculated peak metamorphic pressures on pyroxene-bearing metabasalt inclusions (unit Xbo) near the center of the Harney Peak dome are as much as 6.3 kb (Friberg and others, 1996). This pressure is approximately 3 kb greater than the peak pressures at which metamorphic minerals surrounding the Harney Peak Granite formed. The field evidence indicating major thinning of rocks surrounding the dome and the pressure difference (about 3 kb) suggest an upward movement of 8–9 km of the core of the dome at Harney Peak. The absence of any significant gravity anomaly over the Harney Peak dome (map D) also indicates that the granite does not have extensive roots below the dome.

Stratigraphic relationships indicate that the Bear Mountain F_4 dome deforms at least part of the southern limb of the

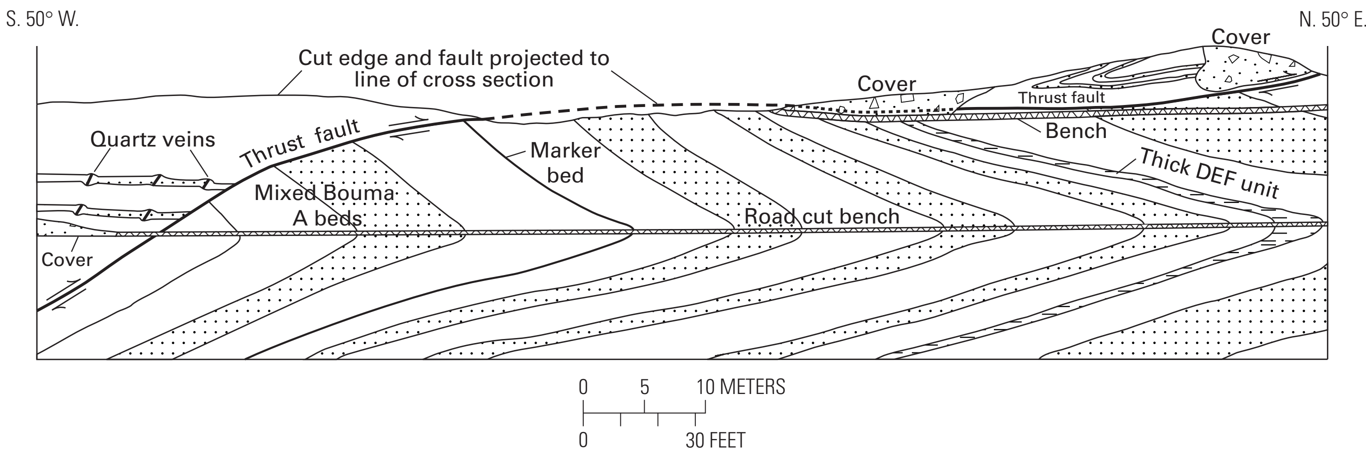

Figure 7. Diagrammatic cross section of late (D_4) recumbent anticline and small thrust fault exposed in road cut along U.S. Highway 16 at Mitchell Lake (D_4), northeast of Hill City. Rock units are all metagraywacke turbidites (unit Xgw_3). Approximately 70 partial or individual Bouma sequences are present. The contacts shown represent boundaries of predominantly Bouma A sequences (patterned) versus mixed or predominantly Bouma DEF sequences that could be readily recognized in photographs. The late, chevron-style fold plunges gently to the northwest and has poorly developed axial-plane foliation in the hinge area along the bench. Above the apparent thrust fault, small crenulations and mineral lineations plunge to the southwest parallel to earlier (F_2?) folds. Although the road cut is somewhat curved, and the cross section was prepared from photographs, the units shown are approximately true thicknesses because of the low plunge of the fold. The greater apparent thicknesses in units above the road-cut bench are due to slight differences in strike and in cut exposure. Massive quartz veins in the two individual Bouma A beds shown on left above fault zone are rotated and indicate right-lateral shear. They also document extensive thinning of host Bouma A beds away from the rigid quartz veins. The degree of thinning is somewhat exaggerated in the drawing.

Slate Prairie F_1 syncline as previously discussed. The presence of early kyanite in the Proterozoic rocks near the center of the dome indicates higher pressures than do the metamorphic minerals that formed around the periphery of both the Bear Mountain and Harney Peak domes. Presumably the kyanite represents a higher pressure regime elevated by the D_4 doming. Late andalusite and cordierite are noted in the tuff and shale unit (Xts) relatively close to the dome and in areas to the east within the staurolite zone. Locally, the andalusite is overprinted by minor folds that are believed to have developed late in the D_4 event. A Rb-Sr age of 1,680 Ma (R.E. Zartman, *in* Ratté, 1986) for coarse muscovite in boudins on the edge of the dome indicates that the dome is contemporaneous with the Harney Peak event even though no known Harney Peak Granite is exposed.

The anticlinal fold at the north end of the Bear Mountain dome shown in cross section *C–C´* is shown to be cored by a flexure in the older rocks of the dome. The anticlinal interpretation is based on general dip reversals along the domal axis at Deerfield Lake. This interpretation, if correct, suggests the possibility that a north- or northwest-trending nappe preceded doming. If the synclinal structure that is shown as an F_2 fold on map C crossing the upper part of Deerfield Lake were a northwest-southeast-trending F_1 fold, it is possible that the core of the dome could have been a major nappe cored by Archean rocks.

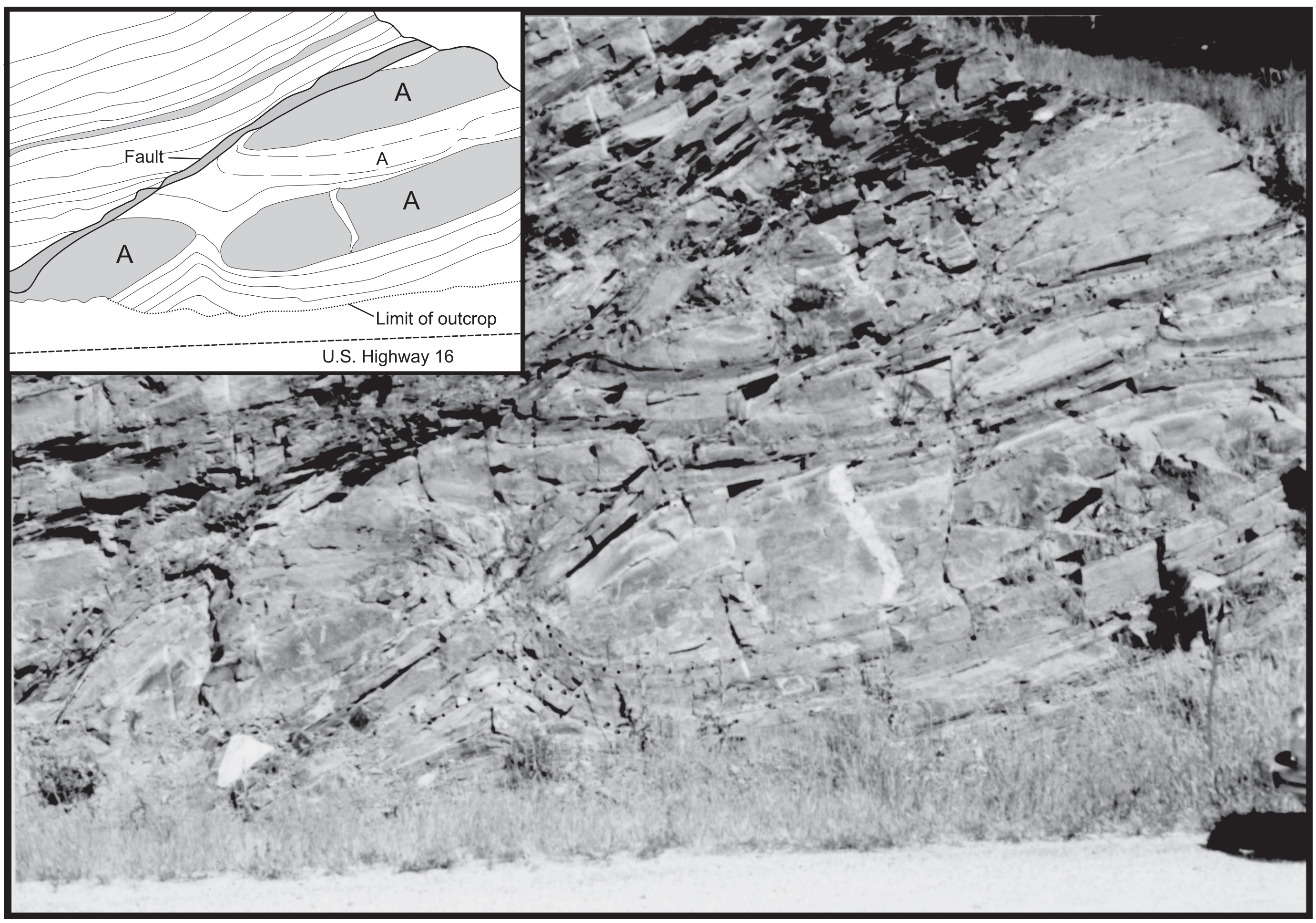

Figure 8. Boudinaged Bouma A units in staurolite-grade metamorphic zone turbidites (unit Xgw_3). The large boudin in the thickest Bouma A bed also shows right-lateral rotation, which is probably related to the development of the small thrust(?) fault. The fold below the fault is inaccessible for detailed examination and could conceivably be one side of a boudin. Incompetent Bouma DEF beds deform into the tensional break of the Bouma A beds. Plunges of the boudin ends are to the northwest, similar to the large late F_4 fold of figure 7. Both fold and boudin development are believed to be related to a buried D_4 domal structure underlying the Hill City area. Compare deformation of the competent and incompetent turbidite beds here with those at high metamorphic grade shown in figure 6. Grain size is the major factor in establishing the relative competency in these bedded rocks. Location on U.S. Highway 16 at 0.6 km east of Hill City and 1.3 km west of Mitchell Lake.

D_5 Event and Related Structures

A late, generally northeast- to east-striking, steeply southeast dipping foliation (S_5) is locally present throughout many parts of the Precambrian core (map A). Flattening of sillimanite aggregates in the plane of the foliation (Redden, 1963) indicates that the deformation postdates emplacement of the Harney Peak Granite and the peak of the lower pressure metamorphism. In areas of lower metamorphic grade, the S_5 foliation is typically a spaced cleavage marked by abundant chlorite or biotite. Generally, minor folds related to this event are absent or limited to very small folds having limbs measured in centimeters. An exception is in the area southwest of Nahant, where this S_5 foliation locally is very well developed and is axial-planar to small-scale folds having limbs a few meters in length (Wynn, 1992). Determination of the age of this D_5 event is difficult. Strontium isotope data on whole-rock samples taken over several meters laterally from two widely separated areas show a re-equilibration age of about 1,540 Ma, which was suggested as the possible time of formation of this post-D_3 event by Redden and others (1990). In the Berne quadrangle Redden (1968) noted oriented amphibole crystals crossing the exterior of elongate calc-silicate ellipsoids (deformed concretions). Structural data indicated the orientation was due to the intersection of the late foliation and the deformed concretions, which were elongated and metamorphosed during D_2. If correctly interpreted, the lack of cataclastic effects on the amphibole suggests the late foliation developed relatively near the time of metamorphism associated with the emplacement of the Harney Peak Granite when temperatures were still relatively high. The absence of any evidence of S_5 within the granite dome supports this approximate age for the development of the foliation. Minor small, late, steeply plunging folds, having general east-west trends, are sparsely present in a general belt across parts of the Rochford, Silver City, and Pactola Dam quadrangles. These folds indent the limbs of the more commonly northwest trending F_2 folds and might be related in some way to the deformation causing the D_5 spaced cleavage (S_5). However, S_5 is not evident in these younger folds. In rocks of contrasting bed competency, these late folds more closely resemble buckle folds, as if the general northwest-trending host units had experienced shortening perpendicular to their trend. The lack of correlation of these somewhat isolated outcrops with the development of S_5 probably indicates they are unrelated. Possibly these late "cross" buckles resulted from differential shear along fault blocks or belts of rocks of different relative competency at some later time.

Folds of Indeterminate Age

Additional intermediate-scale folds of indeterminate age are noted just north of Custer. The trends of the folds vary, and the limbs of some of the larger folds are clearly refolded, as is the St. Elmo fault. The general plunge of linear features is moderately steep to the south, and the original age of the larger folds is unknown. A few indeterminate folds are shown on map C southeast of Custer in the large area of quartzite and schist (unit Xq). Much of this higher metamorphic grade area has not been mapped in detail, and the geometry of the folds is uncertain.

Faults

Numerous faults of diverse types and ages cut the Precambrian rocks in the central Black Hills (table 2). Growth, or rift, faults formed in the Nemo area prior to emplacement of the 2,480-Ma Blue Draw Metagabbro sill (unit Xbd). Similar faults formed after these older rocks were folded, uplifted, and eroded, producing the Estes Formation (units Xec and Xeq). Several faults that are subparallel to contacts are believed to have been thrusts, and possibly more exist along seemingly conformable contacts. In general, these are believed to have been overthrust from the southeast or south during the D_1 event. Other, largely strike-slip faults cut many of the stratified Early Proterozoic rocks prior to emplacement of the 1,715-Ma Harney Peak Granite. Several faults, especially in the Nemo area, probably had repeated movements over time. Postmetamorphism breccias along some faults may indicate local Laramide movement during uplift of the Black Hills.

The character and distribution of the oldest Early Proterozoic sedimentary rocks in the Nemo area indicate two ages of north-northwest-striking faults having opposite senses of vertical movement (tables 1 and 2) prior to later overturning of the section. Older normal faults, responsible for formation of the basin that contains the Boxelder Creek Formation, were downthrown on the east. The younger growth or normal faults, which controlled the geometry of fanglomerate wedges of the Estes Formation, were downthrown on the west. A slight amount of much later post-folding movement also occurred. The development of the relatively large Estes growth fault is shown diagrammatically in figure 2. Although not shown in figure 2, lateral movement on the fault is likely. The southern extent of the Estes fault is well known, but its northern continuation is offset by cross faults and partly concealed by Phanerozoic rocks. Due to a decrease in the angle of dip at depth, the fault probably is subparallel to stratigraphic units to the north.

Later faults in the Nemo area and their movement histories are also complex. The Benchmark fault is clearly right lateral but may have had some normal or even thrust movement early in its history. Geophysical data indicate the north end of the Benchmark fault curves to the east beneath the Phanerozoic rocks. It truncates magnetic anomalies similar to those made by oxide-type banded iron-formation in the Nemo area. The curved trace of the fault parallels an inferred Archean block of granite east of the Nemo fault (see "Geophysics" section). The latter fault is believed to connect in some way beneath the Phanerozoic cover with the fault contact of the Late Archean Little Elk Granite. The Nemo fault probably has an origin and movement similar to that of the Benchmark fault. The curvature of both of these faults, as well as the distribution of adjacent younger rocks to the west, suggests late warping. A possible explanation is that the block of Archean granite to the east acted as a buttress some time after the D_2 event. However, the S_2 foliation has a relatively constant attitude in the older granite and in the younger, more nearly east-west striking rocks between the faults. A possible explanation is that the granite and the rocks between the faults are all competent in contrast to the younger bedded rocks to the west. The younger bedded rocks tend to be foliated parallel to bedding.

The Blue Lead fault in the Sheridan Lake area is probably one of the earliest faults and may well have been a thrust, although its present attitude is apparently steep. The fault is interpreted to form the contact between graywacke (unit Xgw_2) and the older carbonate-facies iron-formation (unit Xif) plus the quartzite and conglomerate (unit Xqc). The contact (fault) is locally discordant as far north as the north side of Sheridan Lake. However, its continuation there to the northeast to the Silver City fault appears to be concordant with folded but southeast-younging graywacke. The extension of the fault across the younger Silver City fault and also across the Pactola Lake fault is not known inasmuch as the displacement of these faults is uncertain in this general area. Possibly the extension of the Blue Lead fault to the east is within part of the thick graywacke (unit Xgw_2), where the fault would not be readily recognizable. The likely trace of the fault suggests that the original displacement may have occurred during D_1 deformation. The fault probably had an original southeasterly dip, but there is inadequate information to confirm this suggestion.

The right-lateral Silver City fault likely developed relatively early, because the axial traces of F_1 folds and D_2 structures are considerably deformed where the fault apparently refracts as it crosses competent graywacke units southwest of Pactola Lake. Also, the fault does not offset the garnet isograd. Many small quartz veins are noted near the fault trace, and minor pyrrhotite-arsenopyrite-gold mineralization associated with the fault in the Mount Rushmore quadrangle is apparently equivalent to lower temperature silver sulfosalt mineralization localized along the fault northwest of Silver City. Fine-grained graphite is common in exposures of the fault, but brecciated schist gouge suggests there was minor later (Laramide?) movement.

The southern extension of the Silver City fault was originally believed to trend southerly west of Silver Mountain and extend along the east side of a late dome about 5 km southeast of Keystone (Redden and others, 1990). However, the apparent displacement of stratigraphic units on this fault reverses to the south and it is possible that the Pactola Lake fault borders the east side of the small dome. If so, the southern extension of the Silver City fault might be the more southeasterly trending inferred fault located about 1 km west of Silver Mountain, which crosses a wide expanse of graywacke (Xgw_2) and is inferred to be located along a zone of relatively abundant quartz veins. To the northwest, the Silver City fault offsets an unnamed small(?), earlier thrust(?) fault that is largely subparallel to lithologic contacts. This fault may have been associated with the F_1 folding. The inferred Humbolt Mountain fault near Hill City is also locally subparallel to lithologic units and may be an extension of the unnamed thrust(?) faults offset by the Silver City fault.

The northwest-striking, left-lateral Empire fault is marked by a pronounced structural contrast in the development of D_2(?) fabrics. Southwest of the fault, in the Hill City area and the area north of the Harney Peak Granite, northeast-striking rock units are typical. Northeast of the Empire fault, F_1 or possibly F_3 structures locally trend largely northwest, parallel to the fault. The inferred buried dome of granite in the Hill City area would favor upward movement on the southwest side of the fault as interpreted in table 2.

In the Keystone area the Empire fault is joined by the smaller, northwest-striking Keystone fault. Farther south, the Empire fault apparently joins either the right-lateral Silver City or the Pactola Lake fault. The presence of galena, pyrite, and calcite along minor faults at the Spokane mine, near the junction of these faults and along the Keystone fault farther north, suggests that these fault systems may have had repeated movement, possibly into the Laramide.

It is likely that the Grand Junction fault had north-to-northeast, east-vergent thrust movement. Northwest of Custer the fault is deflected by a small D_4 dome characterized by D_4 folds, S_4 foliation, and potassium metasomatism, presumably related to an underlying core of Harney Peak Granite. The age of the fault is clearly pre-D_4 doming. Development of the fault subparallel to the overturned limb of the Atlantic Hill anticline west of Custer suggests a relationship to the deformation producing the fold. However, the fault is clearly younger than D_2 folding that involves both hanging wall and footwall rocks. The extension of the Grand Junction fault south of Custer (and an adjacent, unnamed fault to the east) is along a metamorphosed shale (unit Xbs_2), which suggests movement along décollement surfaces of unknown extent. At the Bear Mountain dome the inferred extension of the Grand Junction fault is shown to be domed and to cut out, at least in part, the tuff and shale (unit Xts). Part of the Grand Junction fault may be reactivated by a younger fault striking north-northwest along the east side of the dome. This younger fault, which has post-metamorphic movement, was originally considered to be the continuation of the Grand Junction fault. Extreme thinning of the tuff and shale (unit Xts) due to the D_4 doming, rather than faulting, may explain the disappearance of this unit. Ratté (1986) and Redden (1963) did not recognize the cutting out of the Xts unit on the south end of the dome.

The northwest-striking Pactola Lake fault dips steeply and probably has a relatively large displacement on its north end near Roubaix, where the fault truncates northwesterly trending F_3 fold axes. The north end also curves to the northeast similar to the Benchmark fault. The fault truncates stratigraphic units where it is close to U.S. Highway 385 in the Merritt area, but the lateral displacement is uncertain because the fault marks the contact with the Xbs_1 unit and various

Table 2. Characteristics of major faults in the central Black Hills shown on maps A and C.

[N, north; NW, northwest; NNW, north-northwest; WNW, west-northwest; NE, northeast; NNE, north-northeast]

Name of fault; location[1]	Strike	Apparent movement	Apparent maximum displacement	Type	General relationships	Relative age
Unnamed; Nemo area.	WNW	Left-lateral	Small	Unknown	Offsets north-northwest-striking earlier faults.	Post-Harney Peak metamorphism and D_2 folding. Youngest Precambrian faults in Nemo area.
Keystone (11)	NNW to NW.	Left-lateral	About 3 km	Tear	Offsets F_2 folds	Post-Harney Peak metamorphism and D_2 folding.
Empire (29)	NNE to NW.	Left-lateral, downthrown on east(?).	About 2 km(?)	Tear(?)	Truncates northeast-trending F_2 structures north and northeast of Hill City and area north of Harney Peak dome. Possibly downthrown on northeast.	Post-Harney Peak metamorphism and D_2 folding.
Benchmark (3)	NW to NE.	Right-lateral	About 1 km minimum.	Tear	Truncates oldest Early Proterozoic stratified rocks and cuts unit Xeq. Fault apparently warped around block of Archean granite.	Post-Harney Peak metamorphism(?) and D_2 folding.
Nemo (5)	NNW to N	Right-lateral(?)	Large(?)	Tear	Probably similar to Benchmark fault. Warps F_1 folds	Post-Harney Peak metamorphism(?) and D_2 folding.
Pactola Dam (7).	NNW	Left-lateral(?)	Small	Thrust or tear.	Truncates F_2 folds. Apparently warps F_1 fold axes. Southern extension uncertain.	Post-Harney Peak metamorphism(?) and D_2 folding.
Pactola Lake (4).	NW	Mainly vertical(?); minor left-lateral.	Moderate; increases to the north.	Normal?	Truncates east limb of major fold and F_3 structures farther to the north.	Post D_3.
St. Elmo (18)	NW to NE.	Unknown	Unknown	Normal	Inferred fault that may cut out strata along margin of concealed dome. Bounds contrasting F_2(?) fold patterns northeast of Custer.	Probably pre-D_4 doming.
Burnt Fork (22).	N, NNE	Normal; downthrown on west.	About 1 km(?)	Normal	Displacement may increase near Harney Peak dome and extend farther south than shown on map.	Late D_4 doming(?).
Silver City (8)	NNW to NW.	Right-lateral; downthrown on west.	About 5 km	Tear(?)	Truncates and warps F_1 folds and F_2(?) structures where fault refracts across graywacke units. Southern extension uncertain.	Probably close to time of regional metamorphism with later movement.
Grand Junction (17).	N to NW	Thrust(?); south to north or northwest.	Moderate to possibly large.	Thrust	Cuts F_2 folds; domed by D_4 fold in eastern part of Berne quadrangle and at Bear Mountain. Fault later warped by emplacement of Harney Peak Granite(?).	Pre-D_4; post D_2.
Laughing Water; between (13) and (17).	N to NW	Thrust(?); south to north or northwest.	Moderate to possibly large.	Thrust(?)	Generally concordant with adjacent units. May extend around west side of Harney Peak dome.	Similar age as Grand Junction fault.

Stockade Lake (13).	N to NE to NW.	Thrust(?); southeast to northwest(?).	Unknown	Thrust(?)	Tight, reclined F_1(?) folds parallel general trace of fault near southern margin of Harney Peak Granite.	Pre-D_4; post-D_2.
Humbolt Mountain (23).	NE	Thrust(?); southeast to northwest.	Unknown	Thrust(?)	Locally cuts out unit Xts; probably later reactivation as normal fault.	Pre-D_4(?).
Unnamed; subparallel to contacts near Silver City and southwest of Pactola Dam.	Generally NE.	Moderate to possibly large.	Unknown	Thrusts(?)	Repeats or cuts out some rock units; possibly related to F_1 folds, but some movement may be post-F_2(?). May be equivalent to Humbolt Mountain fault.	D_1(?).
Blue Lead; south of Sheridan Lake.	Variable	Thrust(?); westerly(?).	Unknown	Thrust(?) with later steepening of dip.	Southern part warped; inferred northern part required to repeat stratigraphic section northwest of Blue Lead Mountain. Extensions beyond Silver City and Pactola Lake faults unknown.	Probable initial movement during D_1 but later movement following local D_2 is possible.
Estes (6)	N to NW	Normal; west side down. Right-lateral(?).	About 8 km	Rift or growth.	Offsets oldest Early Proterozoic rocks and provides source for units Xec and Xeq; major movement completed before deposition of unit Xd. Slight post-Xd movement. Diagrammatically shown in figure 2.	Post-2,480 Ma; pre-1,970 Ma.
Unnamed; 2 km ENE of Green Mountain, Nemo area.	NNW	Normal; west side down. Probably considerable strike-slip movement.	Unknown	Rift or growth.	Offsets upper part of Boxelder Creek Formation. Conglomerate and quartzite (Xec, Xeq) wedges on west side. Concealed extension displaces magnetic anomalies related to unit Wif.	Post-2,480 Ma; pre-1.970 Ma.

[1]Number in parentheses is number of fault as shown on map C.

younger units. Nearer Pactola Lake the fault seemingly has no obvious lateral displacement. To the south the fault either coincides with or produces a right-lateral offset of the Silver City fault. The uncertainty of the continuation of the fault relative to the Silver City fault was described previously.

The Stockade Lake thrust fault, southeast of the Harney Peak Granite dome, is nowhere exposed, but its existence is inferred because of the apparent truncation of several mapped subunits. The trace of the fault appears to be subparallel to steeply dipping beds on both sides of the fault in the area southeast and east of the main Harney Peak Granite. Therefore, the fault was probably steepened by the later emplacement of the granite.

Several inferred faults are shown around the west side of the Harney Peak dome. These faults are placed where stratified rock units are seemingly missing. However, other causes, such as facies changes, poor exposures, or extreme tectonic thinning, could explain the apparent omission of strata. The St. Elmo fault, in the area about 1 km northeast of Custer, is locally placed between apparently contrasting fold structures. Also, the fault is interpreted to have been folded. The age of the folding is uncertain, but it is likely that the fault developed slightly before or during the formation of the D_4 Harney Peak dome.

Phanerozoic Structure

The Black Hills consists of two major structural blocks: a higher, north-trending central Black Hills block that underlies much of the map area and a lower, northwest-trending western block that is largely in Wyoming (Lisenbee and DeWitt, 1993). The blocks are essentially separated by the Fanny Peak monocline, which dies out just within the map area (no. 48, map B). The blocks are Laramide in age, and data from sedimentary records in the basins that flank the Black Hills indicate that uplift began at 65–63 Ma and ended about 55–43 Ma (Lisenbee and DeWitt, 1993).

The various faults and folds in Phanerozoic rocks (map B) developed more or less contemporaneously with the uplift, as did the small domes and associated structures shown in the northern one-third of the map. The domes are due to the emplacement of Tertiary igneous rocks, producing laccoliths in the sedimentary section. The larger Lead dome (no. 51, map B) resulted from emplacement of a considerable volume of Tertiary igneous rock (estimated at 10 km^3 by Redden and French, 1989) in the Precambrian basement. This dome is northeast of the main axis of the structurally high central Black Hills block. Recent work (Redden, 2000) suggests some faults, such as the Fourmile and Jewel Cave faults (nos. 38, 39 on map B), were reactivated and displaced earlier White River erosion surfaces.

Many of the various structures shown on map B have long been known, but there are also a number of newly recognized structures. For example, the previously unrecognized structural high near Crook's Tower (no. 43, map B) is interpreted to be related to emplacement of Tertiary igneous rocks in the lower Paleozoic section. Although the anticline at Loring Siding (no. 30, map B) was shown as an anticline (Redden, 1995b), it is actually a relic erosional high on the Pahasapa Limestone, which results in pinching out and thinning of the lower units in the younger Minnelusa Formation. A slight anticlinal closure is presumably due to differential compaction.

The many small domes shown in the Minnekahta Limestone in the extreme southwestern part of the map area were considered to have been caused by solution of several hundred meters of anhydrite in the upper Minnelusa Formation during the uplift and erosion of the Black Hills (Braddock, 1963). Although solution may be the cause, the nearly circular shape of many suggests the possibility that hydration of anhydrite to gypsum with subsequent volume expansion may have contributed to the formation of these minor domes.

In the Argyle quadrangle many small dome and trough structures are noted over an area of approximately 8 km^2 in sections 17, 18, 19, 20, and 29, T. 5 S., R. 4 E. Bedding attitudes are affected over approximately the lower 75 m of the Minnelusa Formation due to solution of the underlying Pahasapa Limestone. A small deposit of White River gravel overlying part of the area indicates that the structures likely developed during early Tertiary time (Redden, 2000). If collapse occurred over true caves, the size of some of the depression areas suggests a former cave system in the underlying Pahasapa Limestone much larger than any now known.

Numerous breccia pipes are localized in strata between the upper Minnelusa Formation and the Lakota Formation and are shown in the Jewel Cave SW and Argyle 7.5-minute quadrangles, respectively (Braddock, 1963; C.G. Bowles, unpub. mapping, 1983). Similar pipes undoubtedly are present in adjacent quadrangles, but the pipes were too small to be identified on aerial photographs. Most of these pipes originated due to solution of 50 m or more of anhydrite in the upper Minnelusa (Braddock, 1963; Gott and others, 1974) and may extend upward more than 400 m into Lower Cretaceous rocks. A few pipes bottom in solution breccias in the underlying Pahasapa Limestone. A pipe cutting the upper Sundance Formation in a road cut a few kilometers west of Hermosa contains brecciated Lakota Formation and is the northernmost pipe recognized. The pipes apparently decrease in abundance or are not present northward along the east side of the Black Hills, where the thickness of anhydrite in the Minnelusa decreases considerably. The thickness of anhydrite increases in the Spearfish area and to the west in Wyoming. No deep breccia pipes are known there, and relatively thick sections of anhydrite are known in the outcrop in Wyoming. The Tertiary erosion surface was at a considerably higher elevation in this general area, and the valley above Triassic rocks must have been deepened relatively recently. Major springs in the Red Valley near the Wyoming border indicate recent solution activity and probably involve removal of both anhydrite and older limestone to form buried breccia pipes. The role of these breccia pipes in transferring ground water between aquifers is briefly described in the “Ground Water” section.

Geophysics

Magnetic and gravity data grids for the central Black Hills are portrayed on 1:300,000-scale maps (maps D and E, respectively). Generalized geologic units from the geologic map (map A) that have recognizable geophysical signatures are included on these maps. Regional magnetic data grids generated from east-west flight lines spaced 4.8 km apart were compiled from the National Uranium Resource Evaluation (NURE) of the Gillette, Newcastle, Rapid City, and Hot Springs 1° × 2° quadrangles (EG&G geoMetrics, 1979a–c, 1980a, b; Bendix Field Engineering Corporation, 1982a–d; Union Carbide Corporation, 1982). In the central Black Hills detailed magnetic data grids generated from east-west flight lines spaced 0.15 and 0.3 km apart were merged with the regional data grids (Meuschke and others, 1962, 1963; U.S. Geological Survey, 1969; Hildenbrand, 1981; Hildenbrand and Kucks, 1981, 1985). Gravity data from the U.S. Department of Defense, National Imagery and Mapping Agency (NIMA) were utilized to construct the Bouguer gravity anomaly map. Gravity stations throughout much of the area are spaced about 1 km apart. However, an area in the northwestern Black Hills, west of long 103°37′30″ N. and between lat 44°7′30″ W. and lat 44°30′ W., has gravity stations spaced as widely as 8–10 km. This area includes the Deadwood North, Deadwood South, Minnesota Ridge, Spearfish, Lead, Nahant, Maurice, Savoy, and Crooks Tower 7.5-minute quadrangles.

Regional magnetic and gravity data grids for the eastern part of the Powder River Basin, the Black Hills uplift, and surrounding region are portrayed on 1:1,500,000-scale maps (maps F, G, and H, respectively). Outlines of the Early Proterozoic core of the Black Hills and selected Laramide structures are shown for reference. Sources of data for these maps are the same as for the core of the Black Hills. All geophysical data were compiled, rasterized, and plotted by Robert Kucks.

Magnetic Anomaly Map

The central Black Hills contains rocks of relatively low magnetic susceptibility, which result in a low total magnetic intensity for the area of exposed Early Proterozoic to Late Archean rocks (map D). Concealed basement rocks to the northeast of the Black Hills uplift have a somewhat stronger magnetic signature.

The highest magnetic high in the Black Hills is in the Nemo area, where hematite-rich oxide-facies banded iron-formations (units Wif, Xbi) are associated with magnetite-rich schist. Magnetite is also abundant in clastic deposits (units Xec, Xeq, Xbcq) that were derived in part from older banded iron-formations. These units are relatively thin, but their high magnetite concentration overwhelms the more extensive adjacent nonmagnetic rocks. A prominent magnetic low situated over Paleozoic rocks to the north of this major high is assumed to be caused by nonmagnetic Archean granite; gravity data support this interpretation. An additional thin, locally magnetite-bearing ferruginous chert (unit Xfc) creates a separate magnetic high along the west side of these older Early Proterozoic rocks west of Nemo. This magnetic high is also partly due to magnetite-bearing contact zones between thin metagabbro sills and dolomitic rocks in the adjacent unit Xds. A magnetic high of less intensity to the north in the Deadman Mountain quadrangle is over an area of Phanerozoic sedimentary and Tertiary igneous rocks. Ground magnetic surveys in the Little Elk Creek area (Bayley, 1972b) suggest this magnetic high is due to a continuation of the Archean oxide-facies iron-formation (unit Wif) exposed along Little Elk Creek.

Other smaller, discontinuous magnetic highs are coincident with outcrops of several rock types (Kleinkopf and Redden, 1975). Metabasalt (units Xbo, Xby) creates local highs above thin interflow units that commonly contain banded or massive chert as well as concentrations of pyrrhotite. Similar small highs associated with metamorphosed shale (unit Xs) are also due to pyrrhotite. Small areas of intermediate magnetic intensity in the Rochford area are associated with metagabbro sills or dikes (unit Xgby) that contain veinlets and segregations of asbestiform minerals. Magnetite is a byproduct of the alteration and metamorphism of these sills. Minor magnetic highs also are associated with some exposures of carbonate-facies banded grunerite-bearing iron-formation, which contains magnetite. In areas of high metamorphic grade, such as near Pringle, pyroxene-bearing banded iron-formation contains magnetite and is responsible for the narrow, north-south-striking, well-defined magnetic high that extends from east of Custer to south of Pringle. Beneath the Phanerozoic sedimentary cover south of Pringle, the high is apparently offset by the concealed east-northeast-striking Pringle fault. Gott and others (1974) considered this fault to have left-lateral offset on the basis of their interpreted offset of the “Black Hills gravity high.” We favor right-lateral movement based on the interpretation that the narrow gravity high in the Wind Cave area is equivalent to the north-trending high east of Pringle.

The most continuous magnetic highs are associated with mica schist of the tuff and shale (unit Xts) in the area between Bear Mountain and Hill City, as shown originally by Kleinkopf and Redden (1975). Accessory magnetite and ilmenite in the relatively thick unit produce linear highs that, with some exceptions, closely follow the surface exposures. On the west side of the Harney Peak dome, destruction of magnetite in the mica schist by metamorphic reactions associated with the emplacement of the Harney Peak Granite reduces the magnetic susceptibility of the unit. Although magnetite is present at the surface in the vertically dipping mica schist, it presumably is absent at depth. On the northeast side of the Bear Mountain dome, near Deerfield Lake, the mica schist unit does not cause magnetic highs. Perhaps the complex structure on the north part of the dome, as previously described, has in some way affected the magnetic susceptibility of the unit in this area, or, alternatively, the unit is flat lying or truncated at a shallow depth by a concealed fault.

Large, linear areas of high magnetic intensity on the northwest side of the Bear Mountain dome indicate extension of unit Xts from exposures west of Deerfield Lake southward beneath Phanerozoic rocks on the west side of the dome as shown on map D. These highs may be offset somewhat if the Grand Junction fault is projected farther to the northwest below Phanerozoic cover, but it is believed that they connect with the correlative stratigraphic unit of alkalic metabasalt (unit Xby) mapped as the Crow Formation (Redden, 1968) that trends northwesterly from the Berne quadrangle on map D. This unit thickens and increases in volcanic content to the north at the boundary with the Phanerozoic rocks and may include magnetic mica schist as it is traced to the northwest.

The overall magnetite concentration of the mica schist (unit Xts) decreases to the north as the schist is diluted with metamorphosed tuffaceous volcaniclastic material of stratigraphically equivalent rocks (unit Xtv) in the Rochford area. Likewise, magnetite in the mica schist decreases to the northeast as graywacke beds become more abundant. Therefore, no obvious magnetic highs are coincident with the unit in those areas.

In the Rochford area and to the north, unit Xtv is not correlated with any continuous magnetic highs. Rather, numerous small highs are apparently associated with pyrrhotite in spatially associated carbonate-facies iron-formation (unit Xif). One such high is along the linear north-south-striking western limb of the Rochford antiform. This high can be traced under Phanerozoic rocks and is associated with age-equivalent rocks in the western part of the Lead dome. In the Lead dome the magnetic signatures of the Early Proterozoic rocks are largely masked due to magnetism associated with Tertiary igneous intrusive bodies. The latter typically have reversely polarized dipoles above laccoliths such as those at Elkhorn Peak, Whitewood Peak, and Crook Mountain.

A small area of low magnetic intensity above the Bear Mountain dome reflects the closure of the inner stratigraphic units of the dome. In contrast, the larger Harney Peak dome lacks any significant magnetic response.

In the southern part of the Deadwood South quadrangle the center of an elongate major low is located about 0.5 km west of Brownsville. This large-intensity low has a relief of approximately 1,500 gammas and is by far the lowest magnetic reading in South Dakota according to the statewide magnetic survey (Petsch, 1967). The low is mainly above Early Proterozoic metamorphosed shale (unit Xs), but a few Tertiary igneous rocks are present. This major low could result from a concealed Tertiary dike or pluton that plunged moderately to the south-southeast (Everson and Roggenthen, 1988). Because existing Tertiary pipes in the area are nearly vertical, it seems more likely that this low is the result of hydrothermal alteration of the Precambrian schist by Tertiary intrusive bodies. The area has not been mapped in detail nor has the schist been studied petrographically. Several lows in the Lead dome area coincide with areas of known Tertiary mineralization and pyritization of the Early Proterozoic basement. These lows are likely to be due, at least in part, to destruction of magnetite in the basement. The prominent low in the northern part of the Deadman Mountain quadrangle is located over an area where Tertiary igneous rocks are rich in magnetite and have been altered. The low is probably caused by reverse polarity of magnetite-bearing igneous rocks.

Gravity Map

In the central Black Hills small-intensity gravity highs (map E) are associated with areas of metabasalt (units Xby and Xbo) and abundant metagabbro (unit Xgb). The areally extensive gravity high about 8 km northwest of Nemo is due to the outcrop of two separate units of metabasalt (unit Xbo), which extend eastward beneath Phanerozoic cover. A similar high, somewhat masked by overlying Phanerozoic sedimentary rocks, is northwest of Bear Mountain. On the basis of the trend of magnetic anomalies (map D), this gravity high is believed to connect with exposed northwest-striking metabasaltic rocks (unit Xby) about 5 km south of Bear Mountain.

Slightly higher than average gravity measurements characterize the exposed Early Proterozoic rocks. Slightly lower than average measurements are representative of areas known or inferred to be underlain by Archean crystalline rocks and granite. A distinct low centered at Piedmont is interpreted to be caused by concealed Archean granite, presumably similar in composition to the Little Elk Granite northeast of Nemo. A small area coinciding with the Bear Mountain dome is a gravity low. The Harney Peak Granite and dome are not coincident with any major gravity features, which suggests that the granite does not have roots beneath its surface exposures not does it extend laterally away from its outcrop pattern.

The gravity data suggest that Early Proterozoic metabasalt and metagabbro in the northern part of the Pactola Dam quadrangle extend eastward below Phanerozoic rocks in an arc interpreted to lie on the south edge of an Archean granite block centered at Piedmont. Similar Early Proterozoic rocks probably rim the Archean rocks on the east, but the reason for an intervening minor gravity low is not apparent. A broad gravity low extends from south of Rapid City southward to the edge of the map area. This low is interpreted to be a north-trending block of largely Archean granite that forms the southeast border of the reconstructed Black Hills Proterozoic depositional basin.

In general, gravity highs associated with the Lead dome and surrounding area are due to the exposed Early Proterozoic rocks and the relatively thin cover of Phanerozoic sedimentary rocks. West of Spearfish is a large area of low gravity that crosses the general axis of the central Black Hills uplift and is believed to be due to Archean granitoid rocks in the subsurface. The gravity signature in the northeastern part of the map area is varied and is believed to indicate Archean basement consisting of diverse rock types.

Because gravity stations are widely spaced on the edge of the Black Hills uplift, minor Laramide structures generally cannot be correlated with known gravity features. One exception is a small gravity high northeast of Sturgis that overlies a Tertiary laccolithic dome just west of Bear Butte (cross section *A–A′*, sheet 2).

Regional Geologic Setting based on Geophysics and Exposed Rocks

Analysis of regional magnetic and gravity features (maps F and G, respectively) in the eastern Powder River Basin and around the Black Hills uplift can assist in interpreting the relationship of the central Black Hills to older Archean terranes to the east and west. Precambrian rocks of the central Black Hills are largely of Early Proterozoic age and have been assumed to belong to the Trans-Hudson orogen, which separates the Archean Wyoming and Superior provinces. These Early Proterozoic rocks were deposited in an epicontinental, long-lived rift basin(s). The midpoint between exposures of Archean rocks in the Wyoming province to the west and the Superior province to the east trends north-south through South Dakota and North Dakota slightly west of the middle of the two States. Approximately 600 km north of the Black Hills, in northern Manitoba, are the nearest outcrops of rocks of the Trans-Hudson orogen. The Trans-Hudson orogen is truncated to the south at approximately the Nebraska State line by the Central Plains orogen. The central Black Hills contains the only exposures of Trans-Hudson rocks in the United States and is important in understanding the extent and chronology of the orogen.

The subdued magnetic response and north-south trends of local narrow magnetic highs (map F) indicate that known Early Proterozoic rocks of the central Black Hills extend south in the subsurface to the Nebraska State line (lat 43° W.). These relationships are confirmed by Proterozoic ages determined from basement rocks in two wells near Edgemont (Klasner and King, 1990). A relatively narrow, east-northeast-trending positive magnetic anomaly in the southeast corner of the central Black Hills area is unlike any other magnetic response that can be correlated with Early Proterozoic rocks. The apparent truncation of this trend to the west is considered to be the west edge of the Archean. The exposed rocks west of this Archean(?) block are predominantly Early Proterozoic shelf-facies quartzose and feldspathic schists. These rocks are believed to have been derived from the Archean(?) block to the east in a manner analogous to those in the Nemo area of the central Black Hills. In the Nemo area rocks slightly older than the quartzose and feldspathic schists formed along the west side of a rift in Archean basement that was controlled by growth faults. The upper part of the deposits at Nemo was derived from an eastern source that probably included the Archean Little Elk Granite. Gravity data south of lat 43°30′ W. (map G) indicate that an inferred Archean granite south of Rapid City extends to the Nebraska State line. This inferred Archean granite is presumed to be similar to the Little Elk Granite at Nemo. A granite basement is confirmed by two drill holes in the southern part of the block (Klasner and King, 1990), but there are no age data available. The gravity and magnetic data are consistent with the extension of Early Proterozoic rocks north of the central Black Hills to lat 45° W. (map H). An east-west gravity low saddle near lat 44°30′ W. suggests the possibility of a structural high in the Archean basement, but no drill-hole data exist to confirm this suggestion.

The location of the probable west edge of Early Proterozoic rocks is uncertain. Line A (map H) marks the west edge of the very distinctive subdued magnetic signature attributed to Early Proterozoic rocks in the central Black Hills. Robbins (1994) recognized this subdued magnetic response in his study of the Powder River Basin and suggested that the southeastern part of the basin could be underlain by Early Proterozoic rocks. Comparison of the magnetic data grid of Robbins (1994) with that of this study suggests that the western margin of Early Proterozoic rocks may coincide with line A (map H). The northern extension of line A crosses a small but prominent gravity high near lat 45° W., and it seems unlikely that this part of the line is the western boundary of Early Proterozoic rocks. Note that Archean rocks of the Wyoming province could extend east of line A as suggested by the inference that metamorphic rocks in the Tinton area are Archean.

A linear feature striking N. 10° E. on both gravity and magnetic maps (line B, map H) is likely a major Precambrian fault because it crosses the main axis of the central Black Hills uplift. Because the extreme northwest corner of the central Black Hills has a gravity response suggesting Archean rocks (map G), that part of line B north of lat 44°15′ W. may represent the northwestern limit of Early Proterozoic supracrustal rocks. The southern extension of line B coincides with the Fanny Peak monocline. Farther south the line could coincide with the Hartville-Rawhide fault (Sims, 1995). In the Hartville area this fault cuts Archean rocks. A possible suture separating Proterozoic rocks on the east from Archean rocks on the west was placed east of the Hartville-Rawhide fault (Sims, 1995). The extension of the suture to the north by Sims (line S, map H) approximately coincides with line B. The suture was extended north of lat 45° W. and curved to the west into Montana at the North Dakota State line by Sims (1995). Sims proposed that the suture represented the eastern boundary of the Wyoming province on the premise that Early Proterozoic stratified rocks in the Black Hills Proterozoic were related to arc processes. However, the detailed geology of the central Black Hills suggests an intracontinental basin for Early Proterozoic time. Also, geochemical and isotope data suggest that the exposed Black Hills Archean rocks have affinities with the Wyoming province rather than the Superior province (Walker and others, 1986; Gosselin and others, 1988). Hence, it seems unlikely that boundary S is a suture. The eastern boundary of the Wyoming province likely lies east of the Black Hills, possibly near long 103° N.

The location of S (map H) could mark the approximate western extent of Early Proterozoic rocks in the southern part of the Black Hills basin inasmuch as line S is approximately parallel to gravity anomalies that trend north-south in the Edgemont area. Unfortunately, details of the structure in the basement near Edgemont are unknown, and the 600–900 m of relief on the Fanny Peak monocline effectively masks gravity features to the west. Magnetic data indicate no obvious changes, although there is a slight magnetic gradient associated with the Fanny Peak monocline. Apparent lack of a unique western boundary for the Early Proterozoic rocks to the south might be due to thinning of the stratigraphic section to the south in response to formation within a half graben. Such a graben would have its greatest displacement on the east side, consistent with data from rocks in the Nemo area, on the east edge of the central Black Hills.

Geochronology

Early Proterozoic metasedimentary and mafic metavolcanic rocks in the Black Hills range in age from greater than 2.48 Ga to less than 1.88 Ga (table 1). This age range is similar to that of rocks associated with the Trans-Hudson orogen in Canada. Regional metamorphism in well-studied parts of the Trans-Hudson orogen took place before 1.85–1.80 Ga. Geophysics and drill-hole data show that metamorphic and other basement rocks strike north-south throughout much of eastern Montana and the western Dakotas. These are presumably foliated in a similar direction, and the fabric has been interpreted as Trans-Hudson in age and related to the east-west collision of the Archean crustal blocks of the Superior and Wyoming provinces. A similarly trending fabric present in the Black Hills is related to the D_2 event, which was accompanied by greenschist- to amphibolite-facies metamorphism.

The age of this regional metamorphism based on new $^{40}Ar/^{39}Ar$ data on various minerals and published Rb-Sr muscovite dates is presented in figure 9 and discussed in the following sections. These data show that the Trans-Hudson orogeny, as defined to the north in Saskatchewan, did not affect the Black Hills, or at least there is no evidence of any old metamorphic event that formed hornblende.

The oldest cooling dates are in the Tinton area, west of Lead (fig. 9), where two amphiboles have plateau and near-plateau ages of 1,730–1,720 Ma, respectively. There are no 1.7-Ga ages in the Lead and Galena areas, but high-temperature gas fractions as old as 1,670 Ma may indicate an age for prograde metamorphism close to 1.7 Ga. Amphiboles from the Bear Mountain and the Tin Mountain areas have plateau and near-plateau ages of 1,686 and 1,680 Ma. Amphiboles from north of Keystone and at Silver Mountain have plateau and near-plateau dates of 1,695–1,686 Ma. Actinolitic amphiboles, such as those at Castle Peak, Roubaix, and, possibly, Bear Mountain, have erratic release spectra and do not aid in defining the age of prograde metamorphism.

The southwestern Black Hills, from Bear Mountain to south of Tin Mountain, and including the western part of the 1,680-Ma Harney Peak Granite, cooled rapidly at 1,690 Ma following prograde metamorphism. Coarse muscovite from the Tin Mountain pegmatite and the Harney Peak Granite have Rb-Sr dates of 1,675–1,665 Ma. Lithium-rich muscovite and cesium-rich biotite have dates of 1,640 and 1,660 Ma, respectively. These data are in agreement with plateau ages of most muscovite in the same area (Holm and others, 1997), which range from 1,648 to 1,611 Ma. All the data suggest that this part of the Black Hills cooled from above 500°C at 1,690 Ma (hornblende dates) to below 300°C at about 1,640 Ma (fine-grained muscovite dates). A cooling rate of about 4° per million years is indicated. No subsequent thermal disturbances are indicated in this part of the Black Hills.

Near-plateau dates of 1,610 Ma in Custer State Park and 1,620 Ma at Deadwood are evidence for an additional event also similar in age to the Central Plains orogeny (Sims, 1995). Rb-Sr muscovite dates as young as 1,613 and 1,576 Ma for the Hugo pegmatite near Keystone substantiate a possible thermal event, but Rb-Sr muscovite dates as old as 1,704 Ma for the nearby Harney Peak Granite suggest that the event could have been local in extent. In the Deadwood area, regional metamorphic grade increases to the northeast, toward Crook Mountain, where a buried muscovite-garnet granite was intersected in drill core (Homestake Mining Company, written commun., 1986). Rb-Sr model ages for the single sample of granite range from 1.68 Ga, for an initial ratio of 0.7189, to 1.6 Ga, for an initial ratio of 0.724. The possibility of a 1.6-Ga thermal event in the Lead area is strengthened by Pb-Pb whole-rock isochrons for the metamorphosed Homestake Formation of 1.64–1.59 Ga (Ed DeWitt, unpub. data, 1999).

A near-plateau date of 1,980 Ma for metagabbro that intrudes rocks younger than the Homestake Formation south of Lead is evidence that the very coarse, splay-oriented amphibole likely is igneous in origin. The 1.98-Ga date is in general agreement with the 1.97-Ga U-Th-Pb zircon age of the Ellison Formation. Apparently, the regional uppermost greenschist-facies metamorphism in that part of the Lead area has not been intense enough to partially reset coarse-grained amphibole in the central part of the metagabbro body.

The $^{40}Ar/^{39}Ar$ hornblende data indicate that the maximum age of D_2 metamorphism in the northwestern Black Hills is about 1,730 Ma. In the Lead area the maximum age is older than 1,670 Ma. Near Harney Peak and to the southwest the age is well defined as 1,695–1,685 Ma. Dated samples in both the Lead area and northeast of the Harney Peak Granite were collected at the first appearance of hornblende in amphibolite-facies rocks. No older metamorphic assemblages were present. The 1,695- to 1,685-Ma dates indicate when the rock cooled below about 500°C following the D_4 event. Our $^{40}Ar/^{39}Ar$ hornblende dates in the Bear Mountain area do not indicate a prograde metamorphic event much older than about 1,685 Ma, which would agree with the D_4 age. Pb-Pb monazite dates as old as 1,760–1,715 Ma from metapelitic rocks in that area (Dahl and Frei, 1998; Dahl and others, 1999) likely

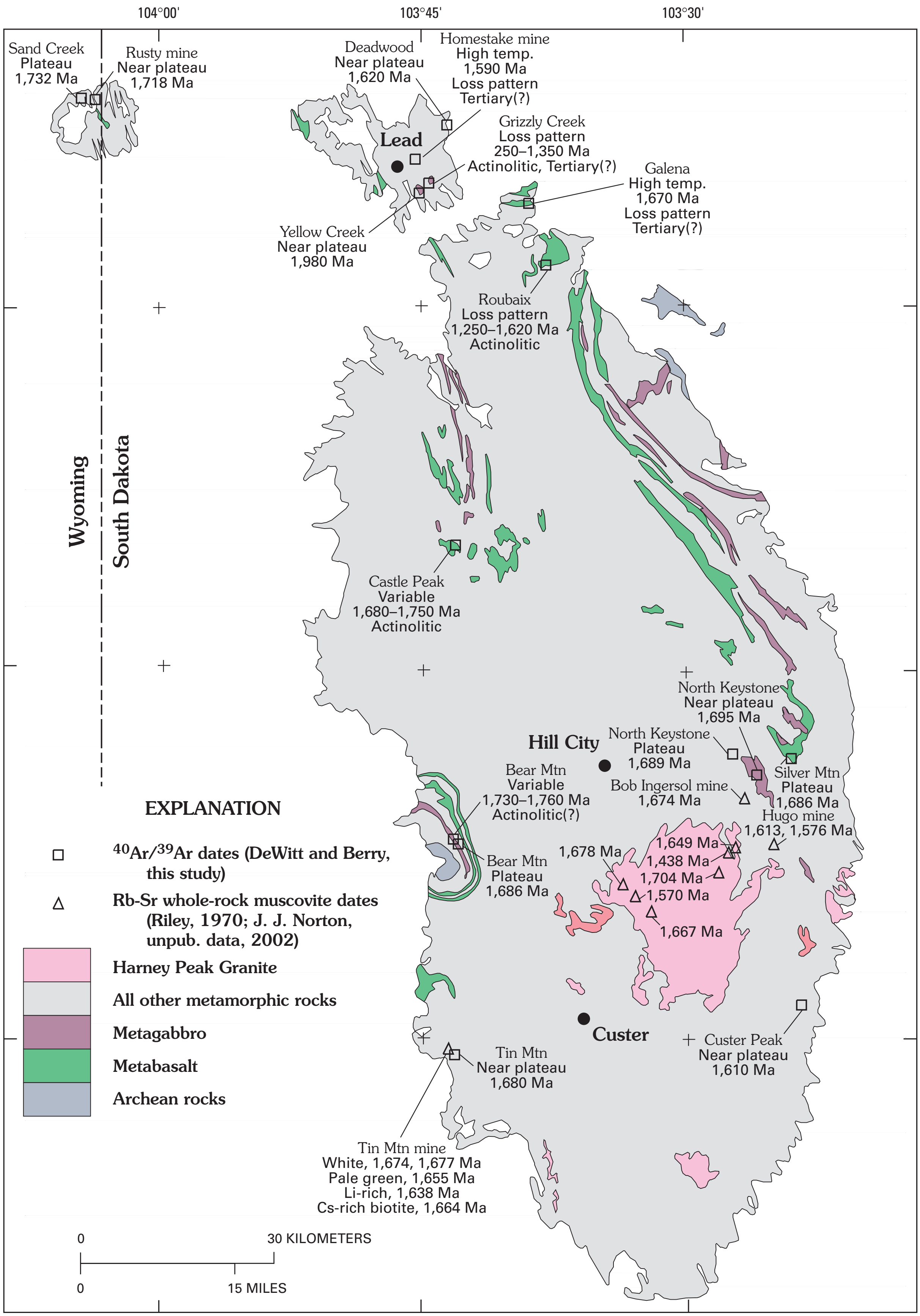

Figure 9. $^{40}Ar/^{39}Ar$ dates on amphiboles and Rb-Sr whole-rock muscovite dates in the central Black Hills.

represent the D_2 age. Step-leach monazite dates may not be able to distinguish old, detrital seed cores from metamorphically formed crystals, as detrital monazite is variably resorbed and thorogenic Pb is incorporated into the metamorphic monazite without the preservation of the detrital core (DeWitt and J.N. Aleinikoff, unpub. data from the Homestake deposit, 2004).

The data on cooling ages in the Black Hills thus indicate an age for the regional metamorphism (D_2) that is not much older than the superposed lower pressure thermal metamorphism (D_4) associated with the emplacement of Harney Peak Granite-type rocks. It seems plausible that the two events were related. It is also evident that these two events were temporally associated with the Central Plains orogeny rather than the Trans-Hudson. How the D_2 event is related to Central Plains deformation is uncertain. Perhaps it indicates a more widespread unrecognized event characteristic of some margins of the Central Plains orogenic belt. Available data also do not define the age and tectonism that produced D_1 folding and no metamorphism. If initial trends of F_1 folds were northeasterly, as seems likely, it is most probable that they were also related to Central Plains deformation.

Mineral Deposits

Exploration and development of mineral deposits played a major role in the early settlement and development of the central Black Hills. Placer gold was discovered near Custer in 1874 by the General Custer Expedition. Mining of various types of deposits has continued to the present and contributes significantly to the economy of the Black Hills. The region is probably most famous for its production of gold, largely because of the world-class Homestake mine at Lead, which has produced more than 40 million ounces of gold between 1878 and 2000, when mining ceased. Mining extended to the 8,000-ft level, and underground workings exist less than 1 km north of the line of cross section *A–A′*.

Gold deposits in the Black Hills are commonly divided into Precambrian and Tertiary types. Details of the different types of mineralization are described by Paterson and others (1988) and Redden and French (1989). The Precambrian gold deposit at the Homestake mine was essentially restricted to the Homestake Formation, a thin carbonate-facies banded iron-formation (unit Xif) separating quartzite, pelite, and graywacke of the Ellison Formation (unit Xqg) from carbonate-rich black shale of the Poorman Formation (unit Xbs_1). Numerous individual ore bodies were concentrated in nine named "ledges." One of these (Main Ledge) has yielded 75 million tons of ore as it was followed down plunge for approximately 5 km (Caddey and others, 1991). The ore bodies constitute only a small part of the Homestake Formation and consist of bedded sulfide ore (pyrrhotite dominant over arsenopyrite) and quartz-vein-related ore. One is present from below to above the nearly vertical garnet isograd. Magnesian chlorite alteration is common in quartz-vein-related ore bodies. Irregular quartz veins associated with ore tend to be more abundant in the higher metamorphic grade eastern part of the mine.

The localization of this world-class gold deposit almost entirely in a single lithologic unit has led to several theories of origin: (1) syngenetic (Paige, 1924); (2) epigenetic during the Tertiary (Noble, 1950, but disproved by subsequent work); (3) syngenetic with modification during metamorphism (Rye and Rye, 1974; Rye and others, 1974); and (4) epigenetic, but of Precambrian age from solutions derived from buried Harney Peak-type granite (Caddey and others, 1991). Recent mineralogic data indicating extreme reducing, sulfur-poor depositional conditions of bedded ore similar to that shown in the photograph (sheet 2) in the Homestake mine suggest original deposition by thermal springs (DeWitt, 1996a, b). The extreme reducing conditions, as well as Pb-isotopic data suggesting mineralization near the age of the host rock, favor initial gold deposition during sedimentation or early diagenesis. If the gold were present in an original sulfide-bearing iron-carbonate chert host rock, mobilization by metamorphic(?) fluids explains the quartz-vein-related ore.

Other small gold deposits in the Precambrian rocks are in quartz veins in graywacke or are spatially associated with carbonate-facies iron-formation (unit Xif) in the Rochford and Keystone areas. Sulfide-poor carbonate-facies iron-formation in those areas is known to have a higher gold concentration than other rock types (DeWitt, 1996a, b), and this may be important in the genesis of these smaller deposits. Faults in the Keystone area appear to be important in localizing vein-related deposits associated with the iron-formation. Throughout the Black Hills, iron-formation (unit Xif) appears to have the highest mineral resource potential of any rock type (DeWitt and others, 1986).

Since 1980, several low-grade gold deposits have been developed in Tertiary breccia pipes, Tertiary intrusive rocks, and Paleozoic strata adjacent to Tertiary intrusive rocks in the northern Black Hills (Paterson and others, 1988). Such low-grade mines are located in areas where early miners selectively mined high-grade gold ores. Open-pit mining of this low-grade gold ore has become economically possible because of cyanide heap-leaching techniques.

High-purity limestone makes up virtually the entire thickness of the Minnekahta Limestone and represents the most important nonmetallic deposit in the region. Most of the limestone is mined at Rapid City for use in making cement, aggregate, and lime. There are very large reserves of this limestone as it forms extensive dip slopes around the entire Black Hills.

Numerous other mineral resources have been locally developed, such as silver, tungsten, lead, copper, light-weight aggregate, building stone, and gypsum. However, to date, no major deposits have been discovered. Extensive iron reserves measured in tens of millions tons, at grades generally between 28 and 30 weight percent iron, are known in the oxide-facies Benchmark Iron-formation (unit Xbi) and Nemo iron-formation (unit Wif) in the Nemo area. The Nemo iron-formation is presently mined as an iron and silica additive for the manufacture of cement at Rapid City. As population centers and demand change, these iron-formations could provide large quantities of iron, but the ore would require beneficiation.

The area around Harney Peak is world famous for pegmatite deposits that have been mined since the late 1800s (sheet 2). Potassium feldspar, sheet and scrap mica, lithium minerals, cassiterite, beryl, columbite-tantalite, and quartz are the major minerals mined. Numerous articles have been written about the zoned pegmatites, their structure, ore deposits, mineralogy, and origin. The most comprehensive description is by Page and others (1953). A complete listing of deposits, commodities, and references is in Wilson and DeWitt (1995). The known near-surface pegmatite deposits have been extensively mined, and their reserves are limited. However, the area will continue to be a source of mineral specimens for avid collectors and mineral dealers.

Ground Water

The Deadwood Formation (unit OЄd), Pahasapa Limestone (unit Mp), Minnelusa Formation (unit Pℙm), and the Inyan Kara Group (units Kl and Kf) are the major bedrock aquifers in the Black Hills area. The recharge area for these aquifers is the central part of the Black Hills uplift (Darton, 1909, 1918; Driscoll and others, 1996). Because these aquifers extend in the subsurface under the adjacent plains, the Black Hills is the source of ground water over a large area in western South Dakota. This was recognized long ago by N.H. Darton (Darton, 1918; Darton and Paige, 1925). Some Precambrian rock units are local sources of ground water for homes and small towns in the central Black Hills. Also, alluvial deposits along valleys provide a moderate amount of water to shallow wells and serve to recharge underlying aquifers.

The Pahasapa Limestone (or "Madison Limestone," as some workers have referred to it with respect to the subsurface) is by far the most prolific aquifer in the area. The uppermost Pahasapa locally has extensive secondary pore space (including caverns) because of karst development before deposition of the overlying Minnelusa Formation. Several perennial streams lose all or much of their surface flow in areas underlain by the Pahasapa Limestone. For example, Boxelder Creek typically loses its entire flow during fall and winter where it crosses the Pahasapa Limestone. In such areas surface waters are diverted to karst-controlled ground-water systems and part of the water re-emerges as springs down gradient (Rahn and Gries, 1973). Along the east and northeast sides of the Black Hills uplift, the hydrostatic head is about the elevation of the lowest stream recharge in the Pahasapa. This relationship produces a significant rise in the water level in wells drilled into the Pahasapa at lower ground elevations around the Black Hills. As a result, many of the early wells in those locations were artesian, or flowing. The water table is generally not much above the elevation of streams and canyon floors that cross the upper contacts of the aquifer units. Consequently, aquifer units that are exposed at higher elevations in interdivide areas contain no water unless the lower contact of an aquifer unit is below the stream or valley.

Extensive outcrops of the Pahasapa Limestone in the central western part of the map area are at a relatively high elevation, and the area is commonly referred to as the "limestone plateau." In the eastern part of this area, the more deeply incised canyons drain the Pahasapa and the water table is approximately at the base of the Pahasapa. Except for local perched water above shale subunits in the Minnelusa Formation, the only available ground water in much of the central western part of the map area is in the Deadwood Formation. Similar conditions exist south of the Jewel Cave fault, and water is encountered only in the deepest levels of Jewel Cave. In the area northwest of Pringle and east of Fourmile Creek, wells encounter little or no water, even in the Deadwood Formation. Variable but generally small amounts of water are encountered in the Precambrian rocks in the area. Apparently, most ground water is moving down slope along the top of weathered Precambrian regolith and along fractures in the Precambrian rocks below the Deadwood Formation. Evidence supporting ground-water movement in the basement rocks includes anomalous high temperatures in water wells in the Pahasapa in the Edgemont area about 20 km south of the map area. Knirsch (1980) and Hildenbrand and Kucks (1985) concluded that the anomalous geothermal gradient indicated by the temperatures was due to circulation into, and transfer of heat from, the Precambrian basement. The transfer becomes relatively easy because the intervening Cambrian rocks are extremely thin or absent in the Edgemont area and the Pahasapa directly overlies Precambrian rocks. Additional supporting evidence for the water movement is the occurrence of anomalous amounts of lithium in these geothermal wells in the Edgemont area, as well as detectable amounts of other rare alkalis and metals (Knirsch, 1980). No known source of lithium exists in the sedimentary section, but schist near the Harney Peak Granite and outlying bodies of granite and pegmatite are extensively enriched in lithium (Redden and Norton, 1992). Norton (1984) noted as much as 700 ppm lithium in schist exposed near pegmatite sills at Pringle, which is near the southern limit of outcrops of Precambrian rocks.

The Minnelusa Formation is a major aquifer due in part to its extensive outcrop area and thickness but also due to its composition. Near the top of the formation is a considerable thickness of anhydrite generally interbedded with massive sandstone. The removal of anhydrite in outcrop areas by solution produces a blanket of highly permeable founder, or collapse, breccias (Braddock, 1963) suitable for rapid recharge. The blanket founder breccias extend downdip into the subcrop in stream divide areas. Localized solution, probably controlled by faults and joints, extends even farther and causes the formation of breccia pipes that can extend as far upward as the Lakota Formation. For the adjacent area south of the map area (sheet 1), Gott and others (1974) presented strong evidence, based on water composition and geothermal gradients, that these pipes acted as recharging conduits to transfer artesian-pressured water from the Minnelusa into Lower Cretaceous sandstone. Solution breccias and fractures also extend downward into the Pahasapa Limestone, which is the major source of artesian recharge to overlying aquifers. Such recharging of overlying aquifers is likely wherever there are extensive soluble rocks in the section or extensive fracturing and faulting.

In the northern Black Hills, especially in areas of higher rainfall to the west, shallow ground water is generally more abundant. Local structures resulting from emplacement of laccoliths and sills in the Paleozoic strata can produce ideal conditions for wells. The igneous rocks generally act either as aquicludes or, where fractured, transport zones. Jones Spring, near Elk Creek in the Tilford 7.5-minute quadrangle, is an excellent example where a tabular Tertiary intrusive body cutting both the Pahasapa Limestone and the lower Minnelusa Formation results in a perched water table that feeds a perennial spring. Because of the structural complexity of areas including Tertiary igneous rocks, individual well sites must be chosen on the basis of the local detailed geology.

In areas underlain by Precambrian rocks, ground water is most easily obtained in the valley bottoms containing alluvial deposits. In areas adjacent to the valleys, domestic wells commonly must be drilled below the elevation of the adjacent valley floor. However, drilling to such elevations will not ensure an adequate water yield in many of the Precambrian areas. The most favorable rock units are the massive graywacke units or rock units of pronounced difference in competence, such as quartzite and schist. Such units are especially favorable well sites along Laramide faults. Potential water yields in wells along Precambrian age faults are variable. The Nemo area, for example, contains many faults, and excellent breccias are found along some faults that cross quartzite and related rocks. However, most of the breccias are silicified and yield no water. Precambrian faults having the best water potential in general are those that are most discordant and that offset competent rock units. In the area around the Harney Peak dome, small granite or pegmatite sills locally act as subsurface dams. Relatively good water yields are possible in such settings. Water at Mount Rushmore National Memorial is obtained from such a site (Rahn, 1990). In alluvial areas the intersection of moderately dipping pegmatite sills and the valley floor are usually good well sites.

Because the Black Hills serves as the recharge area for a number of aquifers that supply the surrounding area and the high plains to the east, avoidance of ground-water contamination is of utmost importance. As an example, the Minnekahta Limestone is a very minor ground-water source in the Triassic Red Valley north of Rapid City. However, recent residential construction on the dip slope of the limestone used improperly designed septic systems and has contaminated the few residential wells in this shallow aquifer that are located downdip. Areas of solution and brecciation are especially sensitive to surface contamination. Recent work suggests that areas of solution are related to the Tertiary erosion surfaces that developed prior to, and during deposition of, the White River Group (Redden, 2000). Contamination can affect not only the exposed aquifer but also another aquifer receiving recharge from the first. Gypsum beds in the Spearfish Formation are prone to solution and can provide rapid movement of contaminants into alluvial deposits. All alluvial deposits along streams in the central Black Hills can be easily contaminated by improper septic siting. Resulting contamination can ultimately affect aquifers recharged from the streams or alluvial water. Thus, good water-management plans are vital for the Black Hills and surrounding area.

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This summary does not include a bibliography of the most recent publications (or even some older publications on some aspects of Black Hills geology) but does include references that provide isotopic age determinations more recent than those in Redden and others (1991).

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Rhonda Driscoll completed the mineral separations and Ross Yeoman completed the isotopic analyses for $^{40}Ar/^{39}Ar$ dating shown in part of figure 9.

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