

OPEN-FILE REPORT 02-5

Geologic Map of the Cheyenne Mountain Quadrangle, El Paso County, Colorado

By

Peter D. Rowley, John W. Himmelreich, Jr., Donald H. Kupfer,
and Christine S. Siddoway



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Peter D. Rowley of the Mineral Resources and Geologic Mapping Section of the Colorado Geological Survey (currently with Geologic Mapping, Inc., of New Harmony, Utah) completed field work on this project from May 2001 to October 2001. John W. Himmelreich, Jr., of John Himmelreich & Associates, Colorado Springs, Colorado; Donald H. Kupfer, former Professor, now of Canon City, Colorado; and Christine S. Siddoway, Professor, Colorado College, Colorado Springs, Colorado supplied data and interpretations based on their earlier mapping in the area. Field Assistant N. Nicole Jones, a student at Adams State College, Alamosa, Colorado, assisted with the field work in August 2001.

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Matt Morgan of the CGS instructed the senior author on the use of the plotter/digitizer that transferred contacts drawn on aerial photos in

the field to the base map. Karen Morgan turned the crude field map into the final digital geologic map. Cheryl Brchan provided the final layout.

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INTRODUCTION

Geologic mapping of the Cheyenne Mountain 7.5-minute quadrangle was undertaken by the Colorado Geological Survey (CGS) as part of the STATEMAP component of the National Cooperative Geologic Mapping Program. Geologic maps produced by the CGS through the STATEMAP program are intended as multi-purpose maps useful for land-use planning, geotechnical engineering, geologic-hazards assessment, mineral-resource development, and ground-water exploration. Mapping of surficial units was emphasized because these deposits underlie most homes and businesses and because their study assists the analysis of climate change in this drought-prone region (Madole, 1994).

Figure 1 shows the status of geologic mapping in the Colorado Springs area. The Cheyenne Mountain quadrangle is the fifth quadrangle in the Colorado Springs area to be mapped by the CGS; we plan to map additional quadrangles in the future. The map area includes the southern suburbs of Colorado Springs. Fort Carson Army Base occupies most of the area east of State Highway 115. Both the Cheyenne Mountain NORAD facility and Cheyenne Mountain State Park are in the northwestern part of the map area. This geologic map was based on (1) prior published and unpublished geologic maps and reports; (2) interpretation of black and white, 1:24,000-scale aerial photography flown in 1992 and color infra-red, 1:40,000-scale aerial photography flown in 1988; (3) unpublished data from field work by Himmelreich, Kupfer, and Siddoway before 2001; and (4) field investigations by Rowley in 2001. The field data and interpretations supplied by Himmelreich are based on

his long-term familiarity with the surficial geology, especially landslides and other geologic hazards, in the Colorado Springs area, where he lives and has a geologic consulting business. The field data and interpretations by Kupfer are based on a long-term geologic mapping project of the Ute Pass fault zone during the period 1960–1975. Kupfer did most of this work while Director of the Louisiana State University (LSU) Geology Camp along Little Fountain Creek and Professor at LSU. His detailed mapping of the camp area considerably aided the interpretation of this geologically complex part of the fault zone and adjacent features. The field data and interpretations by Siddoway are based on mapping with her students, starting in the 1990s as Professor, Geology Department, Colorado College, Colorado Springs, Colorado. The field work by Rowley was in all parts of the map area except restricted military areas, where the reconnaissance geologic mapping of Scott and Wobus (1973) was used.

The mountains in the western part of the map area are part of the southern Front Range. This range continues south to Canon City and north to the Wyoming border. The area of Colorado east of the Front Range and extending from south of Pueblo north to nearly the Wyoming border is the Colorado Piedmont. The Colorado Piedmont is significantly lower topographically than the High Plains east of it because the South Platte and Arkansas Rivers and their tributaries have stripped the area of Miocene sedimentary rocks that now underlie the High Plains. These Miocene sedimentary rocks formerly extended to, and were derived from erosion of, the Front Range (Madole, 1995).

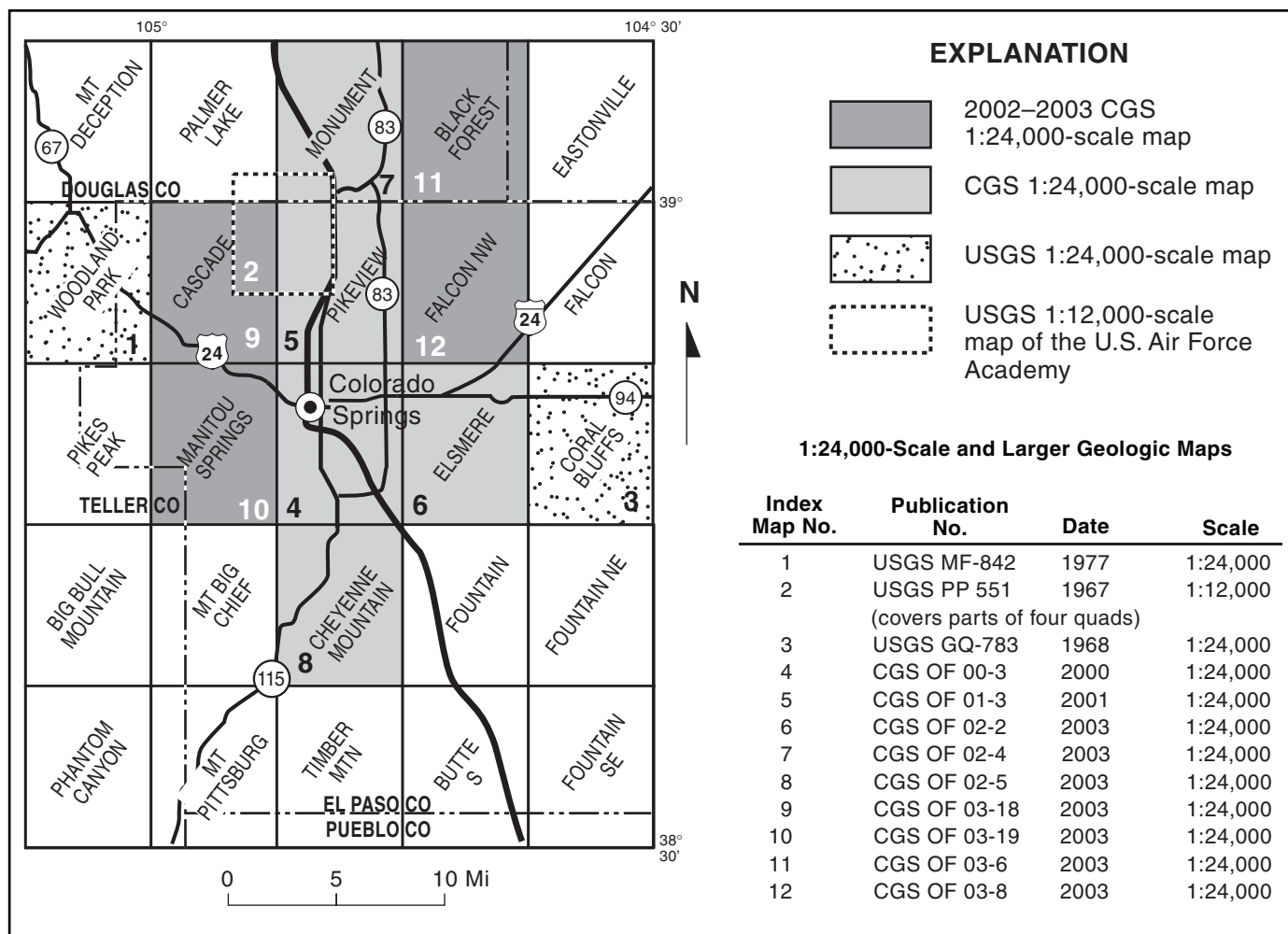


Figure 1. Index of published geologic maps, Front Range area near the Cheyenne Mountain quadrangle area.

DESCRIPTION OF MAP UNITS

SURFICIAL DEPOSITS

Surficial deposits shown on the map are generally more than 5 ft thick but may be thinner locally. Because it is common on most slopes, colluvium is not mapped unless it obscures bedrock relationships. Residuum and artificial fills of limited extent as well as the significant cover of concrete and asphalt in urbanized areas were not mapped. Contacts between surficial units are generally gradational and approximately located, and mapped units locally may include deposits of adjacent types. Age of Quaternary units is given on the chart on the map plate, following Madole (2003); the reader should see Madole (2003) for a detailed discussion of the exact ages of the Quaternary and how they were determined. Relative age assignments for surficial deposits are estimated chiefly on the basis of their height above present streams, degree of post-depositional modification of original surface morphology, and degree of soil development—especially the morphology and thickness of calcium-carbonate-enriched horizons. Older deposits thus are typically higher in the landscape, more eroded, and contain more secondary calcium carbonate than younger deposits. Scott and Wobus (1973) established the Quaternary stratigraphy for the Cheyenne Mountain and adjacent areas, but they did not use formal names; nonetheless, they correlated many informal names with units in the Denver/Boulder area or other parts of the Colorado plains. We prefer to use instead the terminology of Carroll and Crawford (2000) and Madole (2003), in which such correlations are considered tentative.

HUMAN-MADE DEPOSITS

af Artificial fill (latest Holocene)—Consists of waste rock and fill placed during construction of roads, buildings, dams, landfills, and military installations. Map unit is composed mostly of unsorted silt, sand, rock fragments, and construction materials. The maximum unit thickness is about 60 ft. If not adequately compacted, artificial fill may be sub-

ject to settlement when loaded with buildings, roads, etc.

ALLUVIAL DEPOSITS—Clay, silt, sand, and gravel deposited in stream channels, on flood plains, on pediments, on alluvial fans, and as sheetwash along Fountain Creek and its tributary drainages and along the mountain front. Cut or strath terraces and pediments represent the same processes, whereby meandering streams bevel older deposits and leave a thin lag of alluvium, but terraces are younger and can be identified with the stream that caused them. Fill terraces, however, represent valley aggradation. Terrace alluvium two and three, perhaps representing fill terraces, were deposited mostly during late or late-middle Pleistocene glacial stages, whereas pediment gravels one, two, three, and four were deposited during middle Pleistocene and older glacial and interglacial stages (Scott and Wobus, 1973). The approximate terrace and pediment heights reported for each unit are the elevation differences measured between the creek bed and the top of the original or remnant alluvial surface near the creek edge of the terraces. Depositional processes in the alluvial fans include debris flows, and the fans themselves may also include material from rockfall, talus, sheetwash, and other colluvial processes. Deposition of alluvial units may represent a cessation of downcutting because sediment loads temporarily clogged the drainages (R.F. Madole, Madole and Associates, written commun., 2002). The downcutting recorded between the various alluvial units represents mostly regional uplift of the Front Range and Great Plains in late Miocene through present time, estimated by Gable and Hatton (1983) to be 5,000 to 6,600 ft for the eastern Front Range and 300 to 1,600 ft for the eastern Great Plains. Holocene deposits mark potential flood and debris-flow hazards to persons, buildings, and transportation infrastructure.

Qt₁ Terrace alluvium one (Holocene)—Silt, sand, and minor pebble to boulder gravel deposited in modern channels and flood plains of perennial and intermittent streams.

Clasts are as much as 3 ft in diameter and well rounded to subangular, and they consist of diverse lithologies from a wide variety of bedrock formations in the drainages, especially Precambrian igneous and metamorphic rocks and Paleozoic and Mesozoic sedimentary rocks. Unit includes low stream-terrace alluvium that is as much as 10 ft above modern stream level and some local fan and sheetwash alluvium and hillslope colluvium. Madole (1989) studied three Holocene alluvial units at an archeological site along Turkey Creek, about 7 mi south of the mapped area. The well dated (^{14}C) units studied by Madole, which collectively make up what is herein mapped as terrace alluvium one, demonstrate a complex history of Holocene aggradation, the two oldest of which were followed in turn by the incision of Turkey Creek to a depth of 13 to 20 ft during the well-known episode of arroyo cutting of the last 100 to 150 years (Graf, 1983) in the western United States. Terrace alluvium one correlates with flood-plain alluvium and the Husted Alluvium of Varnes and Scott (1967), as mapped at the U.S. Air Force Academy 15 mi north of the Cheyenne Mountain quadrangle. Scott and Wobus (1973) considered the deposits to correlate with Piney Creek Alluvium of the Denver area (Hunt, 1954) but the Piney Creek appears to be a collection of different deposits, most of which appear to have been deposited between 3,000 and 1,500 years BP (before present) according to three ^{14}C ages in the Elsmere area (R.F. Madole, written commun., 2002). Maximum thickness is about 30 ft. Low-lying areas that are underlain by this deposit are subject to flooding.

Qsw

Sheetwash deposits (Holocene and late Pleistocene)—Pebbly and locally cobbly sand, silt, and clay. The sand, silt, and clay consist partly of clasts of Pierre Shale (Kp) transported by sheetflow and deposited on valley sides. Unit is locally well bedded and is deposited on hillslopes below landslides and alluvial fans, and in basins. Map unit locally grades into and interfingers with colluvium (Qc) on steeper hillslopes. Maximum thickness is about 15 ft. Where not compacted, unit may be prone to soil collapse (hydrocompaction), settlement, and piping.

Qt₂

Terrace alluvium two (late Pleistocene)—Sand and pebbly to bouldery gravel that

underlie dissected surfaces about 10 to 20 ft above the level of modern streams. Clasts are subrounded to rounded and reflect the varied lithologies of their drainage basins, notably Precambrian rocks. Unit is correlated with the Monument Creek Alluvium on the Air Force Academy (Varnes and Scott, 1967) and the Broadway Alluvium in the Denver area (Scott and Wobus, 1973). The Broadway was considered by Scott and Wobus (1973) and Madole (1991) to be of Pinedale glaciation age (which began about 30,000 years ago and ended about 12,000 years ago) on the basis of stratigraphic position, weathering, soil development, fossil mammals, and archeological evidence. Maximum thickness is about 15 ft. The unit is a source of sand and gravel in several pits in the mapped area.

Qt₃

Terrace alluvium three (late middle Pleistocene)—Sand and pebbly to cobbly gravel that underlie dissected surfaces about 25 to 30 ft above the level of modern streams. Distinguished from terrace alluvium two (Qt₂) by being more weathered and locally containing more films and nodules of secondary calcium carbonate. Unit was correlated with the Kettle Creek Alluvium on the Air Force Academy (Varnes and Scott, 1967) and the Louviers Alluvium in the Denver area (Scott and Wobus, 1973). The Louviers was considered by Scott and Wobus (1973) and Madole (1991) to be of Bull Lake glaciation age on the basis of weathering, soil development, and isotopic dates. Madole and others (1998) place the age of the Bull Lake at between about 300,000 years ago and 130,000 (note that the 130,000 has been updated to 127,000 years in the Quaternary time chart published here). Maximum thickness is about 10 ft. Unit is a potential source of sand and gravel.

Qg₁,

Qg_{1l},

Qg_{1u}

Pediment gravel one (middle Pleistocene)—Light-reddish-brown clay, sand, and pebbly to bouldery gravel that overlie dissected, gently sloping pediment surfaces about 30 to 100 ft above the level of modern streams. Uppermost 6 ft is not as coarse grained as the rest of the deposit. Two levels of erosion/deposition were noted along Little Fountain Creek: unit Qg_{1l} is younger (lower in elevation) than unit Qg_{1u} (upper in elevation); unit Qg₁ is used where the surface could not be assigned to Qg_{1l} or Qg_{1u}. Clasts

of Precambrian rocks, as large as 3 ft in diameter near the mountain front, are moderately weathered and locally contain minor coatings of calcium carbonate. Unit locally includes sheetwash, colluvium, and eolian sand. Correlated with the Pine Valley Gravel on the Air Force Academy (Varnes and Scott, 1967) and the Slocum Alluvium in the Denver area (Scott and Wobus, 1973). The Slocum was considered by Madole (1991) to be about 240,000 years old based on his interpretation of a uranium-series date, and thus an early Bull Lake age is used by Madole and others (1998) for this unit. As mapped here, the lower part of the map unit may predate Slocum Alluvium and thus be as old as 640,000 years (Madole, 2003). Maximum thickness is about 30 ft. The unit is a potential source of sand and gravel.

Qg ₂ ,	Pediment gravel two (middle Pleistocene) —Locally consolidated, medium-reddish-brown, and brown, well stratified pebble to cobble gravel and sand that overlie dissected, gently sloping pediment surfaces from 150 to 250 ft above the level of modern streams. Deposit includes occasional boulders. Two levels were noted by Scott and Wobus (1973): unit Qg _{2l} is younger (lower in elevation) than unit Qg _{2u} (upper in elevation); unit Qg ₂ is used where the surface could not be assigned to Qg _{2l} or Qg _{2u} . Clasts are commonly weathered to some degree and locally cemented with calcium carbonate. Correlated with the Douglass Mesa Gravel of the Air Force Academy (Varnes and Scott, 1967) and the Verdos Alluvium in the Denver area (Scott, 1963; Scott and Wobus, 1973). The upper part of the Verdos is interbedded with the Lava Creek B ash (Scott, 1960) in the Denver area; this tuff, erupted from the Yellowstone area, was recently dated at 640,000 years (Lanphere and others, 2002). Thickness generally is about 30 ft or less. The unit is a source of sand and gravel in the mapped area.
Qg _{2l} ,	
Qg _{2u}	

Qg ₃	Pediment gravel three (early Pleistocene) —Slightly consolidated, dark-reddish-brown, and brown, stratified cobble and boulder gravel that overlies dissected, gently sloping pediment surfaces from 200 to 250 ft above modern streams. Lithologically similar to pediment gravel two (Qg ₂) but has a higher percentage of cobbles and boulders than pediment gravel two, and weathered upper
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surfaces characterized by scattered to widely scattered boulders; boulders are as large as 3 ft in diameter. Clasts of Precambrian rocks are weathered and coated by calcium carbonate. Correlated with the Lehman Ridge Gravel of the Air Force Academy (Varnes and Scott, 1967) and the Rocky Flats Alluvium in the Denver area (Scott and Wobus, 1973). Based on stratigraphic position, the Rocky Flats is considered by Scott (1960) to be pre-Illinoian and by Madole (1991) to be probably early Pleistocene. Thickness is generally about 20 ft or less. Unit is a source of sand and gravel in the mapped area.

QTg ₄	Pediment gravel four (early Pleistocene? to Pliocene) —Moderately consolidated, reddish-brown and brown, stratified pebble to boulder gravel and sand that overlie a dissected, gently sloping pediment surface from 550 to 700 ft above the level of modern streams. Uppermost 30 ft is less coarse. Contains boulders as large as 4 ft in diameter along the western edge of the map. Clasts, mostly of Precambrian rocks, are weathered and punky. Locally derived cobbles and boulders of the Dakota Sandstone (Kdp) and Keeton Porphyry (Yk) are particularly diagnostic of the deposit. The consolidation of the unit is due to a coating on the pebbles of calcium carbonate, which in some places results in the incipient development of a horizontal calcrete layer 3 to 6 ft thick and several feet below the reddish-brown B horizon of the pediment surface. This calcrete, however, is incompletely developed (discontinuous laterally and vertically) and locally is confined to subvertical joint fillings. Present only in the southern part of the mapped area. Correlated with the Nussbaum Alluvium (Scott and Wobus, 1973), which they considered to be largely if not entirely Pliocene, although the upper parts may be early Pleistocene, based on fossil mammals (Scott, 1982). The Nussbaum forms surfaces that project to altitudes slightly lower than those surfaces overlain by the Ogallala Formation (Madole, 1991). Maximum thickness is about 100 ft. Unit is a source of sand and gravel.
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Qfy	Younger fan deposits (Holocene) —Poorly stratified, poorly sorted deposits of sand, gravel, silt, and clay that are coarser near the heads of fans. May include small landslide deposits. Includes tongues of debris-flow deposits (storm deposits composed of rocks,
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vegetative material, sand, and gravel) that formed in a watery slurry that cut their own channels and left levees on their edges and a lobe of material on their downslope side. Such debris-flow deposits were studied just north of the mapped area (Himmelreich, 1996a; Hoffmann, 1998; Himmelreich, 2002a) and in the northern part of the mapped area (Terry and Hoffmann, 2002a,b). Maximum thickness of the unit is about 40 ft. Areas mapped as young fan deposits are subject to future flooding, debris flows, and sheetwash; of these events, debris flows are potentially the most serious threats to human life in the map area. Terry and Hoffmann (2002a) cited eyewitnesses to a storm event in 1965, during which “highly erosive flows” of water were observed just north of the NORAD facility. Construction in developed areas may exacerbate debris-flow and flooding problems by removing vegetation or by altering drainage patterns. Where fine-grained and uncompacted, deposits may be subject to settlement, hydrocompaction, piping, and landsliding.

Qfo

Older fan deposits (Holocene and late and middle Pleistocene)—Poorly stratified, poorly sorted deposits of sand, gravel, silt, and clay that are coarser near the heads of fans. Unit may include small landslide deposits. Deposit is dissected, so that future sediment deposition on the fan surface is unlikely, but the channels cut into it may still be active and subject to flooding and plugging by debris flows, and thus floods may spill out onto the fan surface. Some units have a soil profile developed in them. Maximum thickness is about 200 ft. Where the deposit is thin and overlies the Pierre Shale, foundation construction in or on it may be subject to expansive soil and landsliding problems. Unit may be a source of sand and gravel.

MASS-WASTING DEPOSITS—Mass-wasting deposits are earth materials that have been transported downslope primarily by gravity. Most of the material moves as a mass rather than as individual fragments or particles borne by water or wind. Water is an important constituent of most mass movements and commonly triggers movement, but generally the water is part of the mass. Consists in the map area of landslide, talus, and colluvium deposits. Of these, colluvium is sepa-

rately defined because the term has been used in many ways. As used here, it consists of clasts of any size transported short distances by gravity and deposited on and below hillslopes; it has no structures indicative of sedimentation or stratification by water in channels and does not include significant sheetwash alluvium. These deposits are of particular interest because the processes that form them are potentially hazardous to persons, buildings, and transportation infrastructure.

Qlsr

Recent landslide deposits (late Holocene)—

Mostly angular, unsorted, and unstratified rock debris moved by gravity from nearby bedrock slopes or cliffs. The map unit is restricted to recently active landslides that have been observed, through aerial photos and consultants reports, or inferred to have moved within the last 20 years. The map unit has fresh morphological features (lateral shear zones, hummocky terrain, headscarps, “step-and-bench” features, and toes). The map unit is present in the quadrangle due to extensive areas underlain by incompetent shale of the Cretaceous Pierre Shale (Kp) and surficial deposits derived from it. Hillslopes of less than even 10 degrees may slide if triggered by earthquakes, rainstorms, etc. Furthermore, the map area is undergoing rapid population growth, and construction activity is commonplace. New roads and other excavations have removed the toes of slopes formed in Pierre Shale and other landslide deposits (Qls, Qlso), allowing new slides to form or reactivating older landslide deposits that occur in the area. Loading the heads of currently stable deposits can trigger slope failure. Most small landslides (Qlsr, Qls, Qlso) in the map area are rotational landslides, in which shearing takes place on a well defined concave-upward curved surface, resulting in backward rotation of the displaced mass. Many other slides (Qlsr, Qls, Qlso) in the map area, especially large ones, are translational landslides, in which shearing takes place on a planar surface that is parallel to the general ground surface, resulting in displacement of the entire mass down the slope. Maximum thickness is about 30 ft. Recent landslides are likely to experience future movement and, like older landslides, exhibit expansive soil and settlement problems. Where of significant size, recent landslide deposits should either be avoided or be properly engineered.

Landslide deposits (Holocene and late

Pleistocene)—Mostly angular, unsorted, and unstratified rock debris moved by gravity from nearby bedrock slopes or cliffs. Mapped where there is no evidence that the slide has been active in the last 20 years. The deposits contain fresh to subdued (older) morphological features. The map unit is common in the quadrangle because extensive areas are underlain by incompetent shale of the Pierre Shale (Kp) and surficial material derived from it. The Pierre Shale has generally failed where present on steep to moderate slopes and commonly where present on slopes of less than 10°. Landslide deposits have been encountered frequently in housing developments in the northwestern part of the map area and just north of the map area, and thus have been studied by geological and engineering firms (for example, Hill, 1974; Himmelreich and others, 1982a, b; Morris and Essigmann, 1983a, b; Mock, 1987; Himmelreich, 1996a; Hoffmann, 1998; CTL/Thompson, 2001; Pardi and Allen, 1996; West, 1997; Attwooll, 1998; Higgins, 1998; Himmelreich, 1998a, b; Morris and Morris, 1998a, b; Himmelreich, 1999a; Terry and Hoffmann, 2002a,b). Here many steep slopes expose dissected edges of older pediment surfaces; the unconsolidated and permeable gravels that underlie these surfaces allow rainwater to pass through as ground water to the underlying impermeable Pierre Shale, where it creates springs at the base of the pediment gravel deposit (Essigmann and others, 1997). The ground water moving into the Pierre Shale destabilizes the Pierre. In fact, water is a particular problem in landslide terranes. Whether from springs, ground water, storms, rainwater in undrained depressions, lawn watering, broken water mains, or septic systems, it increases pore-water pressure, adds weight, weakens rocks by saturation, decreases shear strength by lubricating slip planes, lubricates clay minerals, and causes expanding soils. Unit includes rotational landslides, translational landslides, earthflows, and extensive slope-failure complexes. The largest landslides are translational landslides. Translational slides have moved along various planar slip surfaces (shear planes) that dip gently in the direction of the general slope. Such shear zones are commonly only 0/5 to 2 in. thick; the shale above and below them locally may not appear to be deformed. Although slip surfaces of many translational landslides com-

monly follow bedding planes (Brooker and Peck, 1993), in the map area this is not necessary because the Pierre Shale is so thick that the attitudes of slip surfaces are controlled more by topographic slopes. Such translational slides moving along planar shear planes have been described in several engineering reports of landslides that have been excavated and studied in housing developments just north of the mapped area (Carroll and Crawford, 2000). One basal shear surface just north of the mapped area underlies a large (about 140 acres, intended for housing developments), older (post-Qg₂), presumably now stable, translational landslide whose toe contains a 13-acre multi-family development; the shear surface within the Pierre Shale has a dip of about 6° and a depth that averages about 40 ft below the ground surface (Mock, 1987). Other translational slides were documented just north of the mapped area (Himmelreich and Essigmann, 1982; Himmelreich and others, 1982; Morris and Essigmann, 1983a, b; West, 1997; Himmelreich, 1998b; Morris and Morris, 1998a, b; Himmelreich, 1999a). While accompanied by J.L. White and T.C. Wait (both of the CGS), the senior author visited another shear surface in September 2001 just north of the mapped area. This 2-in.-thick planar shear zone was well exposed in an excavation just west of the intersection of Academy Boulevard and Cheyenne Meadows Road in Pierre Shale overlain by Quaternary sheetwash (Qsw) and terrace alluvium one (Qt₁) (Carroll and Crawford, 2000). Translational landslides also have been studied in the northern part of the Colorado Springs suburbs (Hamacher, 1999; Himmelreich, 1998b, 2001a,b, 2002b). A likely basal shear plane of a translational landslide was described in the northern part of the mapped area (Terry and Hoffmann, 2002a), although the authors of the report did not interpret it as a shear plane. The maximum thickness of the map unit may exceed 100 ft. Landslide deposits range from active to inactive, but regardless of this they may be subject to future movement. Foundation excavations within landslide deposits should be individually evaluated for stability. Deposits may be prone to settlement when loaded and may have low internal structural stability. Expansive soils may be associated with landslide deposits, and shallow (perched) ground water may be present within the deposits.

Qfro

Older landslide, fan, and rockfall deposits, undivided (late to middle? Pleistocene)—

Unstratified, unsorted, clast- and matrix-supported boulders, cobbles, pebbles, sand, and clay at the base of Cheyenne Mountain. Characterized by boulders as much as 20 ft in diameter that are almost as common and large at the eastern margin of the deposit as they are at the mountain front. In the upper part of the deposit, the matrix that supported these boulders has been removed by erosion, leaving the boulders on the surface. Scott and Wobus (1973) and Trimble and Machette (1979) interpreted the deposit to have formed as a large, catastrophic rock avalanche, perhaps triggered by faulting (seismicity) in the Quaternary and perhaps emplaced on a cushion of air. Carroll and Crawford (2000) interpreted the deposit to result from a series of “multiple rockfall events and other types of colluvial processes,” including alluvial fans, that rest on landslide deposits. We likewise consider the deposit to result from multiple processes along this, the steepest part of the mountain front in the quadrangle. However, the widespread nature of the large boulders suggests to us that the primary process involved was a catastrophic landslide or a debris avalanche, which may have transformed itself into a rapid earthflow (earth slide) or debris flow at its eastern toe. Examination of basement excavations for new homes being constructed in the deposit in and just north of the mapped area shows that most of the large boulders are enclosed with sand and gravel deposits; all boulders, sand, and gravel are derived from Precambrian rocks exposed in the mountains. But at one basement site, located near the intersection of Ellsworth Street and Broadmoor Bluffs Drive just north of the quadrangle boundary, this deposit is underlain by a loose breccia made up of angular pebble- to boulder-sized clasts of Pierre Shale in a finer-grained matrix of the same material—clear evidence of a landslide origin. Terry and Hoffmann (2002a,b) examined a proposed development of 125 acres in the mapped unit in the northern part of the map area (Sec. 18, T. 15 S., R. 66 W.) and concurred with the assessment by Scott and Wobus (1973) that the map unit is a landslide. Their 15 drill holes found that the mapped unit is at least 90 ft thick and consists of “deposits of erratic combinations of silty and clayey sand, gravel, cobbles and boulders overlying weathered claystone, claystone and/or shale bedrock.”

Probably this “claystone” is the same lower part of the landslide deposit seen in the basement near the intersection of Ellsworth Street and Broadmoor Bluffs Drive, and indeed Terry and Hoffmann (2002a) also noted that “expansive claystone was encountered near the ground surface in about 10 percent of the site.” The deposit overlies the Pierre Shale and in places the pediment gravel two (Qg₂) that in turn was dissected on the Pierre; thus the map unit postdates pediment gravel two and is middle or late Pleistocene in age. Springs locally occur along the Ute Pass fault zone (Osborne, 1963) and these springs may cause failure in the Pierre Shale or other shales east of the fault zone. Spring water may reduce the shear strength of the slopes east of the fault, which eventually leads to failure that takes part of the adjacent mountain front with it (Himmelreich, unpub. data, 2001). Maximum thickness is at least 100 ft thick. Where the upper sand and boulder gravel is thin, foundation construction may be subject to expansive soil and landsliding problems from the underlying breccia that is rich in clasts of the Pierre Shale. More recent failure of the map unit has resulted in recent landslide deposits (Qlsr). Channels incised into the deposit are prone to flooding and debris flows, and debris-flow deposits locally overlie the map unit near the mountain front. Large boulders that have eroded out and now rest on top of the surface are subject to movement when undermined by erosion or recent flooding.

Qlso

Older landslide deposits (late?, middle, and early Pleistocene)—

Mostly angular, unsorted, and unstratified rock debris that moved by gravity from nearby bedrock slopes or cliffs. Mapped where morphological features are subdued and dissected, and where dated as old. But such deposits may be still active. Older landslide deposits are common in the quadrangle because of the widespread Pierre Shale (Kp) and surficial material derived from it. Unit includes both rotational and translational landslides. The unit contains the largest landslides in the mapped area, which are in the southwestern part of the quadrangle. Here large areas have failed along the flanks of mesas formed where underlain by pediment gravel four (QTg₄). These older landslides postdate deposition of pediment gravel four and locally predate deposition of pediment gravel one (Qg₁). Most of these landslides are translational slides that have moved along

various planar slide surfaces (shear planes) that dip gently away from the mesa top. For such landslides, bedding symbols reflect the attitude of bedrock (Pierre Shale) or intact landslide masses (mostly pediment gravel four) that have slid. Slide surfaces, which are abundantly exposed along and south of Little Fountain Creek, are commonly only 1 to 3 in. thick, made up of sheared shale commonly discolored by ground water that presumably was able to migrate along this zone of higher permeability. Locally the shear planes are anastomosing, or several may be seen within a 10-ft stream exposure. The shale above and below the zones, however, commonly does not appear to be deformed. Maximum thickness of map unit may exceed 100 ft.

Qt

Talus deposits (Holocene and late Pleistocene)—Angular cobbly and bouldery rubble on steep slopes developed on Precambrian igneous and metamorphic rocks on the western side of the map area. This debris was transported downslope by gravity, as rock-falls, rockslides, or rock topples. Unit commonly lacks matrix material. Maximum thickness is about 30 ft. Because of the steepness of their landforms and drainage channels, areas mapped as talus deposits may be subject to flooding, debris-flow, or instability hazards from storms or seismic shaking. Talus deposits may be a source of riprap and decorative stone.

Qc

Colluvium (Holocene and late Pleistocene)—Unsorted, matrix-supported clayey sand and sandy clay and less common pebble to cobble gravel or boulders in a sandy matrix. Derived largely from weathered bedrock and other surficial deposits that are transported downslope primarily by gravity aided locally by sheetwash. Colluvium is gradational into sheetwash deposits, which are found on slopes of generally less than 10 percent. Includes alluvium in steep channels and mountain valleys. Colluvium is not mapped unless it conceals underlying bedrock units and structures. Maximum thickness is about 30 ft. Areas mapped as colluvium may be subject to future deposition as well as sheetwash, rockfall, debris flows, mudflows, and landslides that may accompany colluvium on the relatively steep slopes where it is found. Where fine grained and not compacted, may be prone to collapse.

EOLIAN DEPOSITS—Sand, silt, and clay deposited by wind on level to gently sloping surfaces. Because deposition of eolian sand is correlated with episodes of prolonged drought (loss of plant cover), distribution of different ages of Holocene sand provides information on past climate changes in eastern Colorado (Madole, 1994).

Qes

Eolian sand (Holocene and late Pleistocene)—Fine- to coarse-grained, tan-colored frosted sand and silt deposited by wind. Mostly vegetated (stable), generally unstratified and poorly exposed as only a thin veneer on other surficial deposits. No dune forms were observed. Mapped unit follows the distribution mapped by Scott and Wobus (1973) because the unit is almost entirely within a military restricted area. Maximum thickness is about 10 ft, although it is thicker east (R.F. Madole, oral commun., 2001) and north (Carroll and Crawford, 2000) of the mapped area, where it may be a sand resource. Southeast of the mapped area, eolian sand covers an extensive area and was subdivided into three units ranging in age from late Pleistocene to late Holocene (Madole, 1994, 1995).

BEDROCK

Kp

Pierre Shale (Upper Cretaceous)—Soft, medium- and dark-gray and greenish-gray, shale and local sparse thin beds of tan siltstone and fine-grained sandstone. The Pierre Shale weathers tan to light- and medium-gray to olive-green and commonly into chunks (“popcorn”) and shale chips. Unit locally contains tan spheroidal septarian concretions that are flattened in the plane of bedding. The concretions are as much as 3 ft in longest diameter and may contain invertebrate fossils. Fossils, notably ammonites, cephalopods, clams, and oysters, are abundant in the unit. Scott and Cobban (1986) subdivided the formation into seven units, largely on the basis of ammonites. W.A. Cobban of the U.S. Geological Survey (oral commun., 2001) identified *Inoceramus mortoni* and *Baculites* fossils collected from an outcrop in the northeastern part of the mapped area (NE¹/₄ NE¹/₄ Sec. 16, T. 15 S., R. 66 W.) and *Inoceramus mortoni* and *Baculites scotti* from a streamcut (SW¹/₄ SW¹/₄ Sec. 8, T. 16 S., R. 66 W.) on the northern side of Little Fountain Creek and 1.6 mi west-northwest of Haymes Reservoir; these fossils are character-

istic of the middle and upper parts of the formation. The Pierre Shale has relatively weak strength properties and has several problematic characteristics for development construction on or in it. The Pierre Shale contains bentonite, derived from wind-blown volcanic ash that can be thin discontinuous layers and beds. The bentonitic clay has high shrink-swell potential and can cause heaving bedrock near the surface, such as those noted by Pardi and Allen (1996) and many other geological engineering firms in developments just north of the mapped area. Where exposed on natural or excavated slopes greater than about 10 degrees, the unit is unstable and prone to landsliding. Shallow water tables are common in the Pierre, a result of water perched on impermeable shale beds. White salts are commonly found at the surface where perched water exists. Perched water tables are highly irregular and discontinuous, and may enhance slope instability and other engineering problems (Hoffmann, 1998). Of marine origin. Regionally, the Pierre ranges from about 3,600 to 5,300 ft thick (Scott and Cobban, 1986); in the mapped area, it probably is about 4,000 to 5,000 ft thick.

Kpt

Teepee zone of Gilbert (1896) of the Pierre Shale—Tan, moderately resistant, siltstone and fine-grained sandstone containing abundant fossils. It underlies two small hills in the map area. The hills are located about 0.7 mi south-southwest of Haymes Reservoir (SE¹/₄ SW¹/₄ Sec. 16, T. 16 S., R. 66 W.). The hills, which are about 250 ft apart in a north-trending line and have a relief of 10 to 20 ft, represent bioherms in the Pierre Shale. The bioherms consist almost entirely of clams identified by W.H. Cobban (oral commun., 2001) of the U.S. Geological Survey as *Nymphalucina occidentalis*. The bioherms were interpreted to form along submarine methane seeps (W.H. Cobban, oral communication, 2001). Such cone-shaped hills are called tepee buttes, and the overall fossil zone was mapped by Scott and Cobban (1986) as the tepee zone of Gilbert (1897). Thickness is about 20 ft thick.

Kn

Niobrara Formation (Upper Cretaceous)—Consists of two members, the poorly exposed Smoky Hill Shale Member and the underlying distinctive Fort Hayes Limestone Member. Berman and others (1980) noted that in Colorado the contact of the Niobrara

with the Pierre is generally conformable and transitional from slightly calcareous silty shale of the Pierre to shaly chalk of the Smoky Hill. The Smoky Hill Member is made up of soft, light- to dark-gray, yellowish-orange, and brown, thin-bedded and laminated, limey shale and local thin beds of slightly resistant, gray and white chalk and limestone; the unit commonly weathers to small light-gray and buff shale chips. The Fort Hayes Member consists of resistant, light- to moderate-gray (light-gray weathering), well-bedded, medium-bedded, fine-grained limestone interbedded with thin beds of light-gray shale. It contains solution stylolites. The Niobrara Formation is of marine origin. The Smoky Hill Member is about 400 ft thick, and the Fort Hayes Member is about 100 ft thick.

Kcgg

Carlile Shale, Greenhorn Limestone, and Graneros Shale, undivided (Upper Cretaceous)—The Carlile Shale, Greenhorn Limestone, and Graneros Shale are poorly exposed marine units in the mapped area. They are best exposed in Deadman Canyon, in the southwestern part of the area. These formations were formerly called the Benton Group (Grose, 1960) and more recently have been referred to as the Colorado Group (Carroll and Crawford, 2000). The Carlile Shale, the upper formation, consists of several members that are not exposed in the mapped area, except for the uppermost ones, the Juana Lopez Member and the underlying Codell Sandstone Member. In road cuts in Deadman Canyon, the basal several inches of the ledges mapped as the Fort Hayes Member of the Niobrara, consist of a distinctive buff, marine limestone coquina of small flat oyster shell fragments. In southern Colorado, western Kansas, and northern New Mexico, the bed is much thicker and is broken out and mapped as the Juana Lopez Member of the Carlile Shale (Berman and others, 1980). The Codell Member of the Carlile Shale is a moderately resistant light-gray (light-tan when weathered because of brown stringers of iron oxide), mottled, burrowed, fine- to medium-grained sandstone. It is well bedded in detail because it is interbedded with dark-gray shale, although the member weathers to rounded form. The rest of the Carlile Shale in the mapped area consists of soft, yellow, blue-gray, and black limey shale. The Greenhorn Limestone con-

sists of thin (less than 1 in. to 2 ft) beds of moderately resistant, medium-gray, mottled fine grained limestone beds that weather to light-gray and that occur separated by soft dark-gray marine shale and bentonite. The Graneros Shale is a soft dark-gray and black shale with local 0.5 in.-thick, tan and brown silt beds. The unit weathers to platy chips about 0.25 in. on a side; its lower part locally contains bentonite. The Codell Member is about 20 ft thick, whereas the rest of the Carlile is about 220 ft thick. The Greenhorn Limestone is about 50 ft thick. The Graneros is about 250 ft thick.

Kdp

Dakota Sandstone and Purgatoire Formation, undivided (Lower Cretaceous)—Forms a prominent hogback, or commonly a dual hogback, in the southern part of the mapped area, and it is exposed elsewhere in the western part of the mapped area in faulted sections. The Dakota is particularly well exposed on the northern side of Little Fountain Creek, where it consists, from top to base, of yellow, medium-bedded, ripple-marked, cross bedded, fine- to medium-grained sandstone about 10 ft thick; a covered interval of shale about 30 ft thick; a gray and light-yellow, fine-grained sandstone about 10 ft thick; a 50-ft-thick sequence of medium-gray carbonaceous shale and yellow and gray mudstone containing two 2-ft-thick lenticular sandstone beds; a 10-ft-thick gray, greenish-gray, and locally red-stained, well-bedded, fine-grained sandstone and subordinate light-green siltstone and shale that overall contains burrows and flute casts; a dark-gray to gray carbonaceous shale and green shale about 6 ft thick; a massive gray and light-yellow, fine to medium-grained, red-stained sandstone about 15 ft thick; a 7-ft-thick interval of fine-grained sandstone containing three thin green siltstone and shale beds; and a tan, gray, and pink, thin- to medium-bedded, locally cross bedded, fine-grained sandstone about 25 ft thick. Facies change significantly within short distances laterally so the unit looks different within short distances, although the general characteristics of resistant, gray and light-yellow, mostly planar bedded, fine-grained sandstone and interbedded carbonaceous shale and low-grade coal remain. Where folded or faulted, the unit may exhibit deformation bands, which are thin semi-

brittle shear zones that are noticeable in porous sandstone (Davis, 1999). The lower part of the mapped unit is the Purgatoire Formation, which consists of the Glencairn Shale Member and underlying Lytle Sandstone Member. The Glencairn is a poorly to moderately resistant yellow and gray, poorly bedded sandstone and tan shale, whereas the Lytle is a moderately resistant, light-gray, pebbly, medium- to coarse-grained sandstone, of which a maximum of about 15 ft is exposed. Weimer (1970) interpreted the Dakota Sandstone to represent shoreline regression that resulted in coastal-plain sediments spreading over the underlying marine rocks; he interpreted the Glencairn Member to represent marine deposition and the underlying Lytle Member to represent fluvial channel and overbank deposition. At Little Fountain Creek, Weimer (1970) measured thicknesses of 175 ft for the Dakota Sandstone, 90 ft for the Glencairn Member, and 50 ft for the Lytle Member; they become thinner southwest, west, and northwest of the mapped area.

Jmr

Morrison Formation and Ralston Creek Formation, undivided (Upper Jurassic)—

The Morrison Formation is generally not well exposed, but where exposed in Deadman Canyon, it consists of soft, reddish-brown, light-gray, and light-green, well bedded, thin-bedded mudstone, shale, and limestone. The underlying Ralston Creek Formation consists of soft, gray, medium-grained sandstone underlain by light-red and gray laminated shale and in turn by gray and red sandstone, siltstone, and shale; the shale and siltstone beds weather to small plates. Both formations are continental in origin. The Morrison is about 220 ft thick, whereas the Ralston Creek is about 70 ft thick.

TRPI

Lykins Formation (Lower Triassic? and Upper Permian)—Soft to moderately resistant, reddish-brown and light-gray shale and siltstone interbedded with thin beds of light-gray and tan, mottled, fine- to coarse-grained limestone. On the northern side of Little Fountain Creek and in Deadman Canyon, only the gray and tan limestone in the lower 5 ft of the unit is exposed. About 0.5 mi south of the Hitch Rack Ranch in the southwestern corner of the mapped area, the lower 33 ft of the unit consists of 7 ft of lami-

nated and ripple-marked limestone and interbedded red shale and several white gypsum beds, underlain by 20 ft of pink siltstone, shale, and sandstone, then in turn by 6 ft of purple shale, siltstone, and limestone. Overall, the Lykins in the mapped area is about 150 ft thick.

Ply

Lyons Sandstone (Upper and Middle? Permian)—Made up of upper and lower ridge-forming units of moderately consolidated, pink to red, medium-bedded, commonly cross bedded, well sorted, fine- to medium-grained sandstone. Basal parts include thin red shale. The middle unit is made up of soft red siltstone and conglomerate. Where folded and faulted, the unit commonly also contains abundant deformation bands (Davis, 1999). The unit is of dune and stream origin (Noblett and others, 1987). The Lyons Sandstone is as much as 700 ft thick in the Colorado Springs area (Grose, 1960), but in the mapped area it is about 500 ft thick.

PIf

Fountain Formation (Lower Permian and Pennsylvanian)—Moderately consolidated, orange-red, light-red, light-reddish-brown, pink, and light- to medium-gray, well bedded, locally cross bedded, fine- to coarse-grained arkosic sandstone and pebble to boulder conglomerate. Includes subordinate thin, medium- to dark-red, reddish-brown, and dark-purplish-red beds of shale and mudstone. The most clearly identifiable clasts are subangular to well-rounded vein quartz and subordinate weathered fine- to coarse-grained intrusive rocks. Considered by Suttner (1989) to have been deposited eastward in an intercalated marine delta and nonmarine (subaerial) alluvial fan draining the “Ancestral Rockies,” which here had been uplifted along the ancestral Ute Pass fault zone. Unit is more than 4,000 ft thick in the Colorado Springs area (Grose, 1960), but thickness is highly variable in the mapped area; it attains a maximum thickness of 1,500 ft in the southwestern part of the map area.

Om

Manitou Limestone (Lower Ordovician)—The Manitou Limestone crops out only in several box canyons south of Little Fountain Creek and northwest of Deadman Canyon. The Manitou Limestone is a moderately resistant, pink, light-gray, and pinkish-tan, medium-bedded, fine-grained dolomite. The Fremont Formation (Middle Ordovician) and

underlying Harding Sandstone (Middle Ordovician) have been reported above the Manitou in this area by workers at the LSU Geology Camp (Daniel Holm, written commun., 2001). Scott and Wobus (1973) also suggested that the Harding is present in that area. However, only one doubtful small outcrop of Fremont Formation seems possible. At the northern point that the Manitou is exposed in the map area, it is shown on the map with an attitude of N.55° E., 86° E.; here it is overlain in angular unconformity by the Fountain Formation with an attitude of N. 20° E., 42° E. Near the western margin of the map area, the Manitou appears to rest unconformably on the Precambrian sequence. Where Berg and Ross (1959) studied the Manitou Limestone in its type area near the town of Manitou Springs, about 9 mi north-northwest of the mapped area, it is mostly fossiliferous marine limestone about 195 ft thick. In the map area, the Manitou is at least 160 ft thick.

€s

Sawatch Sandstone (Upper Cambrian)—Resistant medium-purple, light-red, light-gray, and khaki-green, poorly bedded, locally cross bedded, locally bioturbated, pebbly medium-grained quartzite and sandstone. The Sawatch Sandstone is sparsely exposed in the western parts of the map area as thin faulted such as at 0.5 mi north of Rock Creek, and at two places 0.5 to 0.6 mi north of Little Fountain Creek. Where exposed just south of the northern boundary of the mapped area, the unit partly occurs as thin slices bounded by high-angle faults within Precambrian rocks along the mountain front. In some places, the boundary faults are not obvious and these subvertical pieces of sandstone have been interpreted as clastic dikes of Sawatch Sandstone (Harms, 1965; Kupfer and others, 1968). The sedimentary unit is well exposed in the Manitou Springs area, 9 mi north-northwest of the mapped area, where it consists of 14 ft of sandstone deposited along a shoreline and near shore, overlain by 43 ft of glauconite-bearing subaqueous dune sandstone, in turn overlain in angular unconformity by the Manitou Limestone (Myrow, 1998; Myrow and others, 1999). Although derived by erosion of, and deposited directly on, the Precambrian basement, the Sawatch is almost pure quartz sandstone, indicative of a

long episode of chemical weathering before deposition (Myrow, 1998). In three thin sections cut from samples collected in the mapped area, the grains are well-rounded, with a bimodal size distribution of fine- and medium-grained sand consisting mostly of quartz with 1.0 to 1.6 percent altered plutonic rock fragments, 0.4 to 3 percent microcline, and a trace to 0.2 percent plagioclase. Because of its high cementation and resistance, the Sawatch appears to have behaved mechanically like basement rocks during deformation. Maximum thickness in the mapped area is about 20 ft.

Yk

Keeton Porphyry of Murray (1975) (Middle Proterozoic)—Generally resistant red dikes and sills of porphyritic granodiorite or quartz monzonite. Murray (1975) suggested that in one place in the area, the rock unit was extruded onto a conglomerate surface, but these exposures were not seen during the present study. The Keeton Porphyry is generally poorly exposed where present in the southwestern part of the map area. Zoned plagioclase phenocrysts as much as 0.5 in. long make up about 20 percent of the rock and occur in a coarse-grained aphanitic matrix; smaller phenocrysts of potassium feldspar and beta quartz are also present. Sparse ferromagnesian minerals that originally consisted of biotite and hornblende now are altered or weathered to chlorite (Murray, 1975). The presence of beta (high temperature) quartz is consistent with a shallow intrusive origin or even an extrusive origin. The unit makes a distinctive, common clast in the Fountain Formation (PIPf) and pediment gravel four (QTg₄). Murray (1975) noted that Keeton clasts in the Fountain increase in abundance northward, suggesting that the porphyry formerly was more extensive to the north but has now been removed by erosion. Murray (1975) mapped a small outlier of the unit as far north as north of Rock Creek, but such outcrops were not seen in the present study, although red rocks of the quartz monzonite (Yqm) are present there. Murray suggested a Late Proterozoic age for the Keeton Porphyry, but this designation is not well constrained and the unit more likely is Middle Proterozoic.

Ypp

Pikes Peak Granite (Middle Proterozoic)—Resistant, red, medium- to coarse-grained granite intrusions. The Pikes Peak Granite is

characterized by equigranular to porphyritic texture and by abundant red microcline, abundant quartz, moderate plagioclase, and low (generally at or less than 1 percent) amounts of biotite. It is classified as a granite according to the IUGS classification (Streckeisen, 1976). The rock unit belongs to the Pikes Peak batholith, a huge anorogenic plutonic mass that outside the map area has many late-stage alkalic intrusive centers, several of which display all or parts of large oval- to circular-shaped ring faults and dikes, perhaps representing calderas. The Pikes Peak Granite has sharp intrusive contacts, which is in contrast to the generally conformable contacts of the quartz monzonite (Yqm) and the granodiorite (Xgd), as noted in the southern Tarryall region about 30 mi to the northwest (Hawley and Wobus, 1977). Unit includes aplite and pegmatite veins. It commonly weathers to grus on north-facing slopes, but can form cliffs on south-facing slopes. The age of the Pikes Peak Granite is about 1.08 to 1.02 Ga (Hedge and others, 1967; Noblett and others, 1993; Bickford and others, 1989).

Yqm

Quartz monzonite (Middle Proterozoic)—Resistant, pink, locally foliated, medium- to coarse-grained, locally porphyritic, quartz monzonite intrusions. The unit contains high amounts of microcline, moderate plagioclase, low to moderate quartz, and relatively large crystals of abundant (0.5 percent) accessory apatite. It is classified as quartz monzonite and low-quartz granite according to the IUGS classification (Streckeisen, 1976). Quartz monzonite has significantly higher biotite (3.5 to 5 percent) and hornblende (about 1 percent) and lower quartz than the Pikes Peak Granite (Ypp), but significantly less biotite and hornblende than the granodiorite (Xgd). It includes granite pegmatite and aplite dikes. Lithologically it is similar to the Silver Plume Quartz Monzonite of the central Front Range, which has an age of about 1.4 Ga (Hedge, 1969; Nyman and others, 1994). Because of this similarity, the quartz monzonite was correlated with the Silver Plume by Scott and Wobus (1973) and Trimble and Machette (1979). Other silicic intrusions of Silver Plume composition and age in the Wet Mountains (20 mi southwest of the mapped area), Front Range, and other areas go by different names (Bickford and

others, 1989; Nyman and others, 1994; Noblett and others, 1998). We prefer, therefore, to follow Carroll and Crawford (2000) and apply a lithologic name to the rock unit.

Xgd

Granodiorite (Early Proterozoic)—Resistant, gray and pink, medium- to coarse-grained, generally porphyritic granodiorite and quartz diorite intrusions and locally intervening country rocks, now metamorphosed to biotite schist, biotite-feldspar gneiss, and migmatite. The composition of these plutonic rocks is variable, ranging from low-quartz granite to quartz diorite (IUGS composition, Streckeisen, 1976) that is characterized by a gneissic foliation. Phenocrysts are mostly microcline, generally pink, that are as long as 1 in. Depending on the rock composition, the minerals include at least several percent biotite and much subordinate hornblende; sphene and apatite are locally abundant accessory minerals. The unit includes gray and pink, granite to quartz monzonite aplitic and pegmatite dikes as wide as 100 ft,

dark-gray and dark-greenish-gray fine-grained hornblende diorite dikes as wide as 50 ft, and quartz veins as wide as 1 ft. The unit is hydrothermally altered and metamorphosed to schist within several hundred yards of where intruded by the younger Pikes Peak Granite in the northwestern part of the mapped area. Lithologically the unit is similar to the Boulder Creek Granodiorite of the northern Front Range, which has an age of about 1.7 Ga (Hedge and others, 1967; Reed and others, 1987; Noblett and others, 1998). Because of this similarity, the granodiorite was correlated with the Boulder Creek by Scott and Wobus (1973) and Trimble and Machette (1979). Other foliated silicic intrusions of Boulder Creek composition and age in the Wet Mountains (20 mi southwest of the mapped area) and Front Range go by different names (Reed and others, 1987; Bickford and others, 1989). We prefer, therefore, to follow Carroll and Crawford (2000) and apply a lithologic name to the rock unit.

ECONOMIC GEOLOGY

Known mineral resources in the Cheyenne Mountain quadrangle consist mainly of sand and gravel, mostly from pediment gravel two (Qp₂), pediment gravel three (Qp₃), and pediment gravel four (Qtp₄). These resources have been

exploited for local use, especially on Fort Carson Army Base, but unlike in the Colorado Springs area (Carroll and Crawford, 2000), there are no major mines in the map area.

GEOLOGIC HAZARDS AND ENGINEERING CONSTRAINTS

Landslides and related features such as earthflows and soil creep are common in the Colorado Springs area because the Pierre Shale underlies most of the urbanized area east of the mountain front (Hill, 1974; Himmelreich, 1996b, 1998a; Attwooll, 1998; Himmelreich and Noe, 1999; White and Wait, in preparation). Shales in the Niobrara Formation, Carlile Shale, Graneros Shale, and Morrison Formation also are prone to the same types of earth failure but are less signif-

icant because they underlie a much smaller area. Within the Cheyenne Mountain quadrangle, virtually all significant slopes underlain by the Pierre Shale have flowed or slid, particularly along the eroded margins of pediments. When slopes eroded into the Pierre Shale are adversely modified during construction, they commonly become more susceptible to slope failure. In addition, when the hydrologic regime is modified by clearing slopes of vegetation, altering existing

drainage, applying irrigation, watering lawns, or installing septic systems, the slopes become more susceptible to movement. The slopes at the edges of dissected pediments are amenable to earth movement because they have the steepest slopes outside of the mountains and because bedrock beneath the pediment deposits was subaerially exposed to weathering. The granular pediment deposits are permeable and allow rain water and ground water to move vertically through them and into the underlying shale (Essigmann and others, 1997). Shaking of incompetent shales during earthquakes also can trigger landslides (Essigmann and Himmelreich, 1986). Such shaking may have triggered many of the larger slides we have mapped, and thus the study of young faults and the causes of seismic events that take place in and near the mapped area is of practical interest. An older landslide in the Pierre or other units may appear stable but such deposits are more likely than intact, undisturbed Pierre Shale to renew movement during construction or shaking. Because the quadrangle underlies not only the southern side of the rapidly growing suburbs of Colorado Springs but also two military bases, landslides have the potential to create severe economic hardships on public facilities and private homeowners if not correctly identified, analyzed, and engineered. In those places in the mapped area where a significant cover of Quaternary deposits overlie the Pierre, construction has little influence on the inherent stability of the slope.

The largest landslides in the map area are also the oldest (Qlso), coming off the slopes eroded into the highest pediment surface (QTg₄). Although their forms are subdued, they also are the most dissected and have been cut down to their basal shear planes. These landslides are translational slides, in other words they moved as a block on a planar shear plane. Many of these planes are exposed in the southwestern part of the map area. Each shear plane slopes away from the pediment cap and dips at as much as 12 degrees. The ones that are exposed are only from about 0.5 to 3 in. thick and composed of sheared rock, but little of the rock above and below them appears to be brecciated. Booker and Peck (1993) reported that such near-horizontal shear planes resemble bedding planes in Cretaceous overcon-

solidated clay for low-angle shales, and they noted the difficulty of their identification and interpretation. Translational slip planes might be mistaken for low-angle joints or even for bedding. Nonetheless, translational landslides are recognized in some parts of the Colorado Springs area (Himmelreich and Essigmann, 1982; Himmelreich and others, 1982; Morris and Essigmann, 1983a, b; Mock, 1987; West, 1997; Himmelreich, 1998b; Morris and Morris, 1998a, b; Hamacher, 1999; Himmelreich, 1999a, 2001a, b, 2002b). In certain other places in the Colorado Springs area in or north of the mapped area, developments are being undertaken where such shear planes were likely misidentified when exposed in excavations or were not even seen in standard auger geotechnical drilling. In earlier practice, engineering firms who have prepared site investigations in the Colorado Springs area did not have the trained staff and either misidentified or denied that such shear planes exist. If recognized, they were not given the importance needed in determining the stability of the site. Such shear planes need to be recognized for what they are and should be more thoroughly investigated prior to construction for their possible future role in slope instability. The city of Colorado Springs has recently recognized the potential hazards and in requiring additional studies and subsurface investigations such as trenches and continuous sampling core in more recent development proposals where sites are susceptible to landslides (Jonathan White, CGS, written commun., 2002). In addition, special study-area maps are being published to show where landslide-susceptible areas lie in Colorado Springs (White and Wait, in preparation).

Other types of geologic hazards and engineering constraints in the map area should be evaluated. The Pierre Shale and some other shale beds in other formations have high shrink-swell potential. Foundations and roads built on them may be broken, especially if water is allowed to accumulate on fresh exposures. In addition, perched ground water may be close to the surface in the largely impermeable Pierre Shale. Such water may lead to wet basements and instability of pipelines and shallow foundations and roads, as is the case in developed parts of Fort Carson

Army Base (L.P. Lakin, Directorate of Public Works, Fort Carson, Colo.). Rockfall is a hazard in and near areas mapped as talus and colluvium. Large boulders in the older landslide, fan, and rockfall deposits, undivided (Qfro) have the potential to roll or move during floods or earthquakes, or if their natural setting is modified by construction. Debris flows during violent storms, especially in deep, narrow stream channels, are life-threatening hazards.

Do safe areas exist for construction? Generally the best areas are on surficial deposits that are relatively thick (20 feet or more) and away from slopes. Such thick deposits keep

foundations from being placed directly on Pierre Shale. Broad pediment surfaces are particularly good, especially older ones that are well above those large active streams that are the most likely areas to receive significant floods and debris flows. Areas of Precambrian rocks generally weather to relatively stable, permeable sand and gravel, but many of these rocks are in areas of high relief so are expensive to build on and they may be subject to rockfall and sheetwash flooding. Ideal lower areas would include areas on Paleozoic sandstones, such as on the Fountain Formation, where they do not erode to steep hogbacks.

STRUCTURAL GEOLOGY



UTE PASS FAULT ZONE

The Ute Pass fault zone is a broad zone of high-angle reverse faults. It strikes generally north to northeast in the mapped area and separates the resistant Precambrian igneous and metamorphic rocks of the mountain front from the mostly soft, Paleozoic and Mesozoic sedimentary rocks underlying the Colorado Piedmont to the east. The zone passes along nearly the entire western side of the mapped area, then continues north and then north-northwest of the mapped area for 20 mi to the town of Woodland Park, before striking north for another 10 mi as it forms the western side of the Manitou Springs graben, then continues north for another 18 mi before veering west as the Willow Creek fault (Scott and others, 1978; Epis and others, 1980, Figure 14; Dickson and others, 1986, Figure 2). Southward, the fault zone is shown by Scott and Wobus (1973), Scott and others (1978), Trimble and Machette (1979), and us as dying out within our southwestern part of the mapped area. Scott and others (1978) showed it reappearing again 7 mi south-southwest of where they show it dying out. The Ute Pass fault zone has traditionally been considered to be largely of Laramide age (Late Cretaceous to early Cenozoic) and to dip west and be down-thrown to the east. It was also active during late Paleozoic Ancestral Rockies events, resulting in

great thicknesses of orogenic sedimentary rocks of the Fountain Formation (Kluth, 1997). It may have been active in early Paleozoic or Late Proterozoic, as well as in middle to late Cenozoic.

Enough of the Ute Pass fault zone has been mapped to know that its west dip is generally steep. Furthermore, the linear nature of the fault zone over most of the mapped area argues that, where exposed at the surface, it is a high-angle feature (Kupfer and others, 1968). Information is lacking, however, on whether it gets steeper or gentler with depth. In other words, is the fault a high-angle reverse fault or a low-angle thrust fault? Furthermore, is the last phase of movement entirely of Laramide age and, if so, why does the Front Range now exhibit some of the greatest topographic relief in the West? Significant interpretation has revolved around both the fault geometry and age because of its implications to the origin of the Front Range and of other foreland ranges in the Rocky Mountains. In addition, the nature of this fault is of great human interest because of its potential seismicity (Kirkham and Rogers, 1981) and of great scientific interest because it is a fundamental structural feature that has been the subject of controversy.

During 2001, the Ute Pass fault zone was found well exposed in only one place within the mapped area, just south of Rock Creek canyon

(NE $\frac{1}{4}$ SW $\frac{1}{4}$ Sec. 36, T. 15 S., R. 67 W.). Here the shear plane is only several inches wide and dips 78° east. Its hanging wall to the west consists of deeply weathered and sheared, crumbly rocks of granodiorite (unit Xgd), whereas its footwall to the east consists of deeply weathered and sheared, crumbly rocks of the Fountain Formation (PIPf) dipping 57° to the east. But this exposure are only about a square yard in extent, along an old dirt road on the property of John May, several hundred yards south of the May Museum. This old road, at the base of the mountain front, continues up the mountain side through many switchbacks, exposing dozens of faults of similar attitude, all interpreted to be part of the Ute Pass fault zone but within rocks of the hanging wall of granodiorite. Few of these faults can be shown on the map, but clearly in this area the fault zone locally is at least 1,000 feet wide.

Similarly wide zones of steeply to gently west-dipping shears in hanging-wall rocks are exposed locally in other parts of the map area. Probably the most prominent of these broad fault zones is at the northern edge of the map area (N $\frac{1}{2}$ NW $\frac{1}{4}$ Sec. 13, T 15 S, R 67 W); this structural zone is shown on the geologic map but likewise without the detail. Here many faults of the fault zone are exposed in the hanging wall of the granodiorite (Xgd), some containing fault slivers as much as 20 ft wide of pebbly quartzite that is here interpreted as the Sawatch Sandstone (Cs). Others persons have interpreted the sandstone slivers at these outcrops as clastic dikes (U.S. Army Corps of Engineers, 1966), in keeping with an interpretation of such sandstone dikes elsewhere along the Ute Pass fault zone in and near the Colorado Springs area (Harms, 1965; Kupfer and others, 1968). The main strand of the fault zone that separates granodiorite from sedimentary rocks is not exposed at this location near the northern edge of the mapped area, but instead is concealed beneath a covered interval of only about 20 feet that separates granodiorite on the west from Dakota Sandstone to the east. East of the Dakota Sandstone, the Graneros Shale, Greenhorn Limestone, Carlile Shale, and the Fort Hayes Member of the Niobrara Formation are exposed in order from west to east. Osborne (1963) also noted these footwall relationships.

During the 1950s and 1960s, Professor D.H. Kupfer managed the Louisiana State University (LSU) Geology Field Camp (central part of W $\frac{1}{2}$ Sec. 2, T. 16 S., R. 67 W.), at the entrance to the Precambrian part of the canyon of Little Fountain Creek, in the southwestern part of the mapped area. When not teaching, he and his students used these occasions to conduct a regional study of the Ute Pass fault zone (they called it the Cheyenne fault zone) in and north of the mapped area. They found that in almost all places the fault zone dipped west at between 55° and 80° and in places was as wide a zone as 6,000 ft (Kupfer and others, 1968). Low-angle dips in the rare places where they were found were best explained by the hanging wall slumping eastward over the footwall. In places they found horizontal slickensides and concluded that the fault, especially where exposed north of the mapped area, had considerable left-lateral slip.

As part of the teaching of the field camp, Kupfer and his students, and other professors and students since then, mapped the geology of the area of and near the field camp, including the Ute Pass fault zone. Here field relations are particularly complicated and, unfortunately, the exposures are sufficient only to torment all who might interpret the rocks, whether the greenest of students or the most veteran of professionals. Major complications near the field camp involve alternating west-northwest- to east-northeast-trending ridges of granodiorite (Xgd) and embayments of Paleozoic rocks. The most problematic of these is just south of the field camp itself. At the eastern end of this ridge, the Ute Pass fault zone is not exposed but was interpreted by three-point geometry to dip west at either a low (Bucher, 1933; C.S. Siddoway, unpub. mapping, about 1995) or a moderate (46°) angle (D.H. Kupfer, unpub. mapping, about 1964) between the granodiorite in the hanging wall and the Fountain Formation in the footwall. Relationships are not exposed between the Precambrian and Paleozoic units on the northern and southern sides of the granodiorite ridge, but one can walk on outcrops of the Fountain Formation all the way around the ridge. South of the western end of the ridge, the Ute Pass fault zone can be inferred striking about N. 15° E. at a high angle

between Precambrian and Ordovician rocks. To the north of the western end of the ridge, the Ute Pass fault zone can be inferred to strike also at about N. 15° E. between granodiorite and Fountain Formation. About a mile north of Little Fountain Creek, where the prominent hogbacks die out to the north, the Ute Pass fault zone can again be interpreted as striking north-northeast, then almost due north.

So how does one interpret these relationships, which provide both a delight for professors and a nightmare for students? The simplest explanation, and probably the most common for field camp students in their mapping projects, was to show the fault zone bending east around the ridge, before continuing both south of the ridge and north around the meadow ("Keaton Meadow") north of camp. This certainly is a reasonable interpretation, based on the Fountain outcrops and if one concludes that the low- or moderate-angle dip on the fault plane at the eastern end of the ridge is representative of the dip of the fault plane elsewhere. An alternative interpretation is that of Kupfer and others (1968), who concluded that high-angle, west-northwest-striking (transverse) faults displaced the Ute Pass fault zone on the northern side of the ridge and perhaps on the southern side of the ridge, but they interpreted that most of the ridge represents a landslide or slumped mass of granodiorite that moved eastward a quarter mile from the main fault zone. This also is a reasonable interpretation, inasmuch as west- to west-northwest faults are common in both the granodiorite and the Mesozoic rocks in this part of the map area. Yet another interpretation, suggested by C.S. Siddoway (unpub. mapping, 1995), is that a steep west-northwest striking fault on the southern side of the granodiorite ridge is a right-lateral transfer fault that allowed eastward translation of granodiorite relative to Fountain Formation upon a gently west-dipping thrust segment. This interpretation explains the overturned bedding of the Fountain Formation east of the ridge and the prominent deflection of the trend of the north-striking hogback ridge of Lyons Formation south of Little Fountain Creek. With these interpretations in mind, the senior author mapped the area and also concluded that west-northwest-striking transverse faults displace

the Ute Pass fault zone. But he found the granodiorite on the ridge too coherent to support the interpretation of a slump or landslide origin except perhaps at its eastern end. Lacking exposures on the northern and southern sides of the ridge, he inferred these sides to be controlled by the transverse faults. The low- to moderate- dip of the Ute Pass fault zone at the eastern end of the granodiorite ridge is shown as a thrust segment, as suggested by Siddoway, although this segment could have attained its low angle by slumping to the east, as seen by Kupfer in other places along the Ute Pass fault zone. Also, Rowley and Kupfer interpreted another transverse fault north of the LSU Camp and "Keaton Meadow" and continuing eastward past where the prominent hogbacks end about a mile north of Little Fountain Creek.

Probably the best known cultural facility in the mapped area is that of the NORAD (North American Air Defense Command) Combat Operations Center along the eastern side of Cheyenne Mountain. This command and control communications center is housed in large chambers and tunnels (an area of about 4.5 acres) in intrusive and metamorphic rocks of the granodiorite (Xgd). The chambers and tunnels were designed both to withstand a nuclear detonation and to operate for a sustained period without outside contact (Chapman, 1996). The center was excavated during 1961 and 1962 (it became operational in 1966) out of the granodiorite on the eastern slope of Cheyenne Mountain (Underwood and Distefano, 1967; Chapman, 1996). Apparently the best exposures of the Ute Pass fault zone were created by construction of the NORAD facility. In particular, the north-portal tunnel into the NORAD underground facility was excavated through the fault zone (see cross section A-A'). Unfortunately, the senior author did not see the fault zone during his 2001 field work, although he examined the entrances to the northern and southern portals of the facility. Precambrian rocks in the hanging wall of the fault zone are exposed, as well as several subsidiary faults in the hanging wall. Parking lots, buildings, and grout conceal other structural relationships, and no footwall rocks are now visible. Prior to construction, a geologic map of the entire site was

put together in about 1960 by the Omaha District of the U.S. Army Corps of Engineers, who did the design and construction of the tunnel and underground cavity. Unfortunately, this geologic map has been lost. Several reports based on this work, most of which is unpublished, have survived, but these provided no detailed descriptions of the Ute Pass fault zone. Instead, the reports focused on engineering aspects of the facility, with the primary geologic focus on fractures. Underwood and Distefano (1967) and Samuelson (1970) reported that joints in the quadrangle into which the NORAD facility was excavated consist of an open tensional joint set that strikes from N. 10° W. to N. 30° W. (it dips an average of 74° NW), and a closed compressional joint set that strikes from N. 40° E. to N. 70° E. (it dips an average of 84° NE). Ground water locally follows the open joint set, whereas granite dikes and shear zones parallel the closed set. Samuelson (1970) noted that most of the basalt dikes at the NORAD facility had the same strike as the closed joints and a thickness of as much as 30 ft. Moore and others (1993) reported two N. 20° W. faults (dipping steeply eastward) that cut the roof and floor of the NORAD facility and parallel the strike of the open joint set described by Underwood and Distefano (1967) and Samuelson (1970). These two faults localize ground water in the main cavity. We mapped similarly striking faults in and west of Cheyenne Mountain.

Kupfer and others (1968) studied the Ute Pass fault zone at the north portal prior to construction and found it, as well as parallel faults in the hanging wall, to be dipping west at, or greater than, about 55°. The granodiorite is altered along the fault zone, with alteration intensity diminishing to the west over about 200 feet (D.H. Kupfer, unpub. data, 1961). Osborne (1963), in an unpublished report prepared shortly after construction of the NORAD facility, noted that the Dakota Sandstone and, just to its east, the Graneros Shale were exposed about 100 ft northeast of the entrance to the north portal, on the northern side of a small east-trending gully coming out of the mountain front. In unspecified locations nearby, Osborne (1963) noted the Fort Hayes Limestone Member of the Niobrara Formation in the footwall of the fault zone, and between the north and

south portal, Pierre Shale in roadcuts east of the fault zone. He observed that all of these footwall rocks were overturned and dipped west, with those closest to the fault generally dipping at about 45° W. Based on drill holes that passed through the fault zone into Cretaceous rocks in the footwall, Osborne (1963) calculated dips of the main fault of 37.5° and 48°. Samuelson (1970) reported that, at the entrance to the north portal, the fault strikes between N. 30–35° W. and dips between 45° and 50° SW; the rock is hydrothermally altered along the fault, and basalt dikes were injected along the zone. Hembre and TerBest (1997, Figure 3E) drew a cross section through the NORAD facility, based on one of the diamond-drill holes drilled from the floor of the cavity. The section was accompanied by no supporting text and no supporting data other than portraying a slant hole drilled down and eastward through the fault zone, to a total depth of 776 ft. This section showed the Ute Pass fault zone to have a westward dip of about 45° between the surface and the intersection with the slant hole, below which Hembre and TerBest (1997) interpreted the fault to flatten to about 30°. Based on the early reports and their cross section, we interpret the geometry of the fault in our cross section A–A'. Our portrayals of the Ute Pass fault zone in this section and section B–B' are partly influenced by the portrayal of similar frontal faults given by Berg (1962). We suggest that the Laramide part of the fault zone consists of many faults that feed into a single high-angle reverse structure at depth. Between these faults are panels of sedimentary strata that show variable dips in the overturned western limb of a footwall syncline. We have made no attempt to portray what the Paleozoic (Ancestral Rockies) or earlier geometry of this fault zone might have looked like, although we are convinced that a fault zone of one or more Paleozoic or Late Proterozoic ages existed, at approximately the same location. A speculative middle to late Tertiary normal fault zone is also shown in section B–B'.

WESTWARD REENTRANTS

The area of the mountain front south of the valley of Little Fountain Creek appears to be a

structural reentrant (embayment) that projects westward into the Front Range. In other words, the relatively straight north-trending mountain scarp veers westward. The thickest and best exposed sequences of Upper Cretaceous (as young as Niobrara Formation) to Paleozoic rocks in the quadrangle are exposed in this reentrant. Specifically, at a point in the NW¹/₄ of Sec. 1, T. 16 S., R. 67 W., the character of the mountain front and much of its rock section changes. To the south of this point, strong hogbacks of Lyons Formation through lower Niobrara Formation are exposed, whereas these same rock units to the north are thin, or relatively soft and poorly exposed. The Fountain Formation, which underlies the rocks making up these hogbacks, is much thicker south of the NW¹/₄ Sec. 1. The thick Fountain in turn is underlain in angular unconformity by the only exposures of Ordovician rocks in the map area. The next embayment to the north, the Manitou Springs area, contains Cambrian and Ordovician rocks (Noblett and others, 1997; Myrow, 1998; Myrow and others, 1999).

The differences in mountain-front geology of the reentrant (south of the NW¹/₄ Sec. 1) versus that of the area to the north may be correlated in part with the differences in the size of the Ute Pass fault zone. To the north of Sec. 1, the fault is a major structure, thousands of feet across. Nearly the entire scarp between Sec. 1 and Rock Creek canyon to the north contains steeply west-dipping reverse faults, as seen along an old road noted previously. Thus to the north the Paleozoic and Mesozoic rocks were absent, thinned (attenuated), sheared, and so badly deformed that they do not form significant hogbacks. In contrast, it is not certain where the Ute Pass fault zone is south of the NW¹/₄ Sec. 1. One coauthor (Kupfer) interprets that the fault zone dies out in the northern part of Sec. 1 by splaying into two or more bedding-plane faults in the near-vertical hogbacks and losing all displacement southward before crossing Little Fountain Creek. Another coauthor (Rowley) interprets that the Ute Pass fault zone is displaced westward by transverse faults and continues southward, defining the contact between Precambrian and Paleozoic rocks before it dies out or passes westward into Precambrian rocks

before it reaches the western map boundary. At the map boundary, the southern part of the Manitou Limestone appears to rest unconformably on the Precambrian rocks.

Another difference north and south of Sec. 1 is that the northern boundary of the reentrant seems to be defined by west-northwest- to east-northeast-striking (transverse) faults. The northern of the transverse faults (in Sec. 1) continues east almost to State Highway 115, where it appears to cut off the Ute Pass fault zone. Are these easterly faults the northern structural margin of an embayment, which subsided to allow accumulation of sediments that are not present or are thinner to the north? If so, one is tempted to interpret the faults as recurrently active, like the Ute Pass fault zone that defines the eastern side of the Front Range. The oldest of these recurrent ages of deformation in the reentrant and the Ute Pass fault zone may be Late Proterozoic to Ordovician in order to explain the presence of Ordovician rocks. Another significant age of deformation on the Ute Pass fault zone must be late Paleozoic in order to explain the uplift, exposure, and erosion of the Precambrian rocks of the Ancestral Rockies, resulting in the deposition of the Fountain Formation. Post-Manitou, pre-Fountain deformation next is required, whether by movement on the Ute Pass fault zone or other structures in the area of the reentrant, so as to explain the angular unconformity. Laramide movement on the Ute Pass fault zone explains the uplift, exposure, and erosion of the Precambrian rocks and deposition of orogenic sediments of Late Cretaceous to Paleocene age. Laramide displacement of the transverse faults also is required where they cut rocks as young as Niobrara Formation south of Little Fountain Creek.

It could be argued that Laramide displacement along the northern of the easterly-striking faults could not have been significant because the Lyons Sandstone, Dakota Sandstone, and Niobrara Formation, which form strong hogbacks south of the fault, are also present north of this northern fault. Furthermore, it could be argued that these three units and the other Paleozoic and Mesozoic rocks do not appear to have been displaced significant amounts. The explanation, however,

would seem to be that fault displacement, as with the Ute Pass fault zone itself, is large in the brittle Precambrian rocks but dies out upwards into the cover rocks, which express the displacement by folding.

The northern of the easterly-striking faults appears to mark a point to the north of which the Ute Pass fault zone is a major structure, and to the south of which the Ute Pass fault zone is dying out. What kind of structures are these transverse faults? The apparent offset of the Ute Pass fault zone, our preferred interpretation as shown on the geologic map, by easterly-striking faults north and south of Little Fountain Creek could be explained by either strike-slip motion or vertical motion along the transverse faults, or a combination of both. If they define the northern margin of a marine embayment, the motion would seem to be largely vertical, at least in Paleozoic time. Their motion during Laramide deformation, when the Ute Pass fault zone was a reverse fault, would more likely be strike slip so as to agree with the stress field of Laramide time. It is also possible that these structures are the same type as east-striking "transverse zones" of Rowley (1998) and Rowley and Dixon (2001), which are interpreted to act like transform faults in ocean basins in that they define long-lived east-west boundaries that allowed the crust north and south of them to deform at different amounts, styles, and rates of strain.

ORIGIN OF THE SOUTHERN FRONT RANGE

The geometry of the Laramide faults and related folds defining the eastern flank of the Front Range appears to be largely similar to those of the margins of most other basement-cored Rocky Mountain foreland ranges in Colorado and Wyoming. This structural style has been interpreted by most geologists in terms of two end members, as summarized by Berg (1962), Harms (1965), Stone (1984), and Mitra and Mount (1998). These are: (1) the fold-thrust model (Berg, 1962), in which most movement of the ranges is vertical and involves sharp monoclinical folding in the basement rocks and overlying sedimentary cover rocks, followed by high-angle or low-angle (thrusting) reverse faults that may not penetrate

upward to the surface; and (2) the thrust-fold model (Stone, 1984), in which most movement of the ranges is horizontal and involves initial thrusting in basement rocks, which causes monoclinical folding of the cover rocks. In addition to these models, a third model involving movement along strike-slip faults, in which transpression and thrusting resulted (Sales, 1968; Stone, 1969) or the ranges or parts of them may dome upward due to flower structure (Hembre and TerBest, 1997; Kelley and Chapin, 1997). However, such faults appear to be present more commonly across or in the interior of ranges and do not seem to characterize range margins. The fold-thrust model dominated thinking in the 1960s and 1970s (e.g., Boos and Boos, 1957; Berg, 1962; Harms, 1965; Tweto, 1975; Matthews, 1978), but in the 1990s Erslev (1993) claimed that the fold-thrust model is "erroneous" and Erslev and Rogers (1993) suggested that "current consensus" favors the thrust-fold model. Under the thrust-fold model (e.g., Gries, 1983; Brown, 1984; Stone, 1984; Erslev, 1993; Erslev and Rogers, 1993; Erslev and Selvig, 1997), the range-front faults are interpreted to flatten with depth; the local presence of back-thrusts (that is, thrusts with opposite vergence to the primary eastward to southward vergence direction) is better explained. Under the thrust-fold model, "triangle zones" (Erslev and Rogers, 1993; Erslev and Selvig, 1997) are now widely interpreted as a zone of penetrative shear that initiates in the middle limb of the monocline and with progressive fault displacement is translated to a footwall position, above an underlying, downward-steepening blind thrust. Erslev and Rogers (1993) attributed the present popularity of the thrust-fold model to new drilling and seismic data obtained in the 1980s (e.g., Gries, 1983; Jacob, 1983; Gries and Dyer, 1985). These data reveal low-angle thrusts along range flanks, and balanced cross sections based on seismic interpretation that support the geometries of the thrust-fold model in many areas. Yet low-angle thrusts are also well substantiated in the fold-thrust model, as shown by many examples given by Berg (1962), although such thrusts are interpreted as steepening with depth. According to Berg (1962), the thrust-fold model explains the extensive overturned section beneath the thrust by

drag beneath a thrust plane, but explaining it by plastic folding by a fold-thrust model makes more sense. Jacob (1983) proposed an origin in which the core of the Front Range uplifted vertically, whereas the flanks spread laterally over the adjacent sedimentary basins; such a model is similar to the fold-thrust model. Tweto (1975) also argued a fold-thrust model, pointing out that the growth of foreland ranges took place with uplift of a central axis, which expanded outward as it expanded upward; under this concept, it was his idea that mountain-border faults formed later in the process. Whatever the model, it is likely that the Precambrian basement will break when deformed, and faults will propagate upward so that the cover rocks can fault if displacement is large but more likely will fold.

Fold-thrust structures are also called "forced folds," in which the monocline-with or without thrusts or reverse faults-results directly from the differential uplift and rotation of a basement block (Matthews and Work, 1978; Matthews, 1987). Laramide monoclines on the Colorado Plateau are generally interpreted to originate by this mechanism (Davis, 1999). Monoclines on the Colorado Plateau, however, do not commonly have overturned middle limbs, nor are they accompanied by thrusts. In other words, such monoclines could be attributed to a block uplift, as this term is used by Berg (1962). In contrast, fold thrusts as applied by Berg (1962) differ from block uplifts in that they represent the results of "uplift by folding which progressed into overturning that produced great horizontal displacement." Nonetheless, block uplifts and fold thrusts differ only by degree, that is, different amounts of vertical versus horizontal movements. Clearly foreland ranges have significant amounts of vertical throw, and adjacent basins go down by corresponding amounts, so great vertical structural relief characterizes these ranges and basins (Berg, 1962; Tweto, 1975). Tweto (1975), in fact, noted the long history of "buoyancy" of the Rocky Mountains, in which vertical movements in the late Precambrian, late Paleozoic, late Mesozoic, and late Cenozoic led to massive stripping of younger cover rocks and deep dissection of Precambrian basement rocks. Kupfer and others (1968) noted early and late Paleozoic movement

along the Ute Pass fault zone and suggested that late Paleozoic movement was greater than Laramide movement. We likewise are impressed by long-lived vertical movements, and therefore we have interpreted the Ute Pass fault zone as steepening downward (cross sections A-A', B-B'). The underlying cause of vertical movements is well beyond the scope of this geologic map.

The age of the faults in the southern Front Range, including the Rampart Range, is of major significance because of the potential seismicity of the faults. As Kupfer and others (1968) and Tweto (1975) noted, faulting has taken place during the Precambrian, early and late Paleozoic, and Laramide. Trimble (1980) and Bilodeau (1986), like many others, concluded that the latest fault movement in the area is Laramide. In fact, Laramide deformation of the Front Range took place in several pulses, resulting in deposition of orogenic sediments in the Denver Basin (Tweto, 1975; Kelley and Chapin, 1997; Gregory and Chase, 1994; Raynolds, 1997; Thorson and others, 2001; Thorson and Madole, 2003). If, however, the latest movement on the Ute Pass fault zone were Laramide, then why do the present mountains show so much relief? Certainly the present Front Range coincides with the Laramide Front Range, which can be looked upon as a faulted anticline (Steven and others, 1997; T.A. Steven, oral commun., 2001).

How much of the relief we see between the Front Range and Colorado Piedmont is due to late Cenozoic faulting, and how much is due to broad-scale (epeirogenic) uplift? Does some of this relief represent extensional block faulting? Late Cenozoic extension is certainly the case in some ranges farther west, which represent the northern part of the Rio Grande rift, as best exposed in New Mexico. The late Cenozoic faults, which began movement in the late Oligocene and continue to be active into the late Cenozoic, resulted in erosion of upthrown areas (horsts) and deposition of basin-fill clastic material in downthrown areas (grabens). Epis and others (1980) proposed that most of the northwest- and north-northwest-striking faults in the mountains west of Colorado Springs are reactivated older (Precambrian to Laramide) faults that in early Miocene became extensional block faults. But if

the ranges that bound the Colorado Piedmont also significantly represent late Cenozoic block faults, why do we not see normal faults bounding the eastern side of the Front Range? There are many faults that cut Precambrian rocks of the Front Range, but the Phanerozoic cover has been eroded away and thus the age of these faults can not be determined more precisely than as Proterozoic or post-Proterozoic. Faults also cut Paleozoic and Mesozoic rocks and some rocks as young as Paleocene east of the mountain front, but there are few middle or late Cenozoic rocks in contact with the faults, so that their ages can only be considered as Laramide or younger.

What can we use to date the uplift we see today? After the Laramide, a low-relief erosion surface was formed across the entire area. Gregory and Chase (1994) cited evidence that the surface formed at about 2 to 3 km elevation and attributed its development to climatic factors, but such an explanation would not seem to be necessary inasmuch as streams certainly can cut erosion surfaces at elevations that are well above sea level (e.g., Mackin, 1937). Remnants of this erosion surface are now widespread across the mountains of central Colorado, and the surface is particularly apparent at the crest of the Rampart Range (Epis and others, 1980, Figures 7, 8). Furthermore, this surface is faulted, as particularly well exposed in the Woodland Park area, where the Ute Pass fault zone defines the Manitou Springs graben (Epis and others, 1980, Figure 17) with at least 1,100 ft of post-late Eocene offset that displaces the erosion surface (Steven and others, 1997). The age of this surface has been interpreted to have formed in late Eocene or early Oligocene (Epis and others, 1980; Trimble, 1980; Bilodeau, 1986), on the basis that the erosion surface was covered by Tertiary volcanic rocks, generally the oldest of which is the widespread 36.6-Ma Wall Mountain Tuff (Epis and others, 1980; Gregory and Chase, 1994). Steven and others (1997), however, noted that cutting of erosion surfaces on the mountains of central Colorado has taken place throughout the Cenozoic and, in fact, such erosion has removed most of the volcanic rocks that provided the age control for the oldest of these surfaces and locally cuts down into source Tertiary plutons. Such sur-

faces are not necessarily entirely of low relief, for they grade into subdued mountains. Steven and others (1997) suggested that the culmination of cutting of such surfaces was in middle to late Miocene and that, in fact, the cutting of most of the Rampart surface dates to this age. Their evidence is that the Rampart surface, as well as the Sherman surface in the northern Front Range, is locally capped by conglomerates that are the lateral equivalent to the latest Miocene Ogallala Formation of the Great Plains.

As is well known by looking at the present topography, the low-relief erosion surfaces have been cut by deep canyons, with incised meanders. Such canyon cutting, the "canyon cycle" of erosion of 5- to 4-Ma summarized by Steven and others (1997), clearly records a younger episode of uplift. Additional confirmation of the age of this erosion has been provided by K-Ar dates of 5 to 1 Ma from supergene alunite and jarosite in the San Juan Mountains (Rye and others, 2000), which underwent similar uplift and canyon cutting. Trimble (1980) and Bilodeau (1986) attributed the relative relief we see today between the Front Range and Great Plains to stream erosion during rapid broad (epeirogenic) uplift of both the Colorado Piedmont and the Front Range over the last several million years, following the end of deposition of the Ogallala Formation at about 5 Ma. This is the traditional view of Colorado geology: the mountains we see are the result of erosion of Laramide structures. In our view, however, the steep, youthful terrain of the mountain front cannot be explained entirely by epeirogenic uplift and consequent river incision and erosion. Steven and others (1997) proposed that the uplift was not uniform, instead many local structures formed as it was taking place. Among these are faults, as in the Woodland Park area mentioned above, which post-date the Ogallala Formation or equivalent formations (Steven and others, 1997). A "mosaic of fault blocks offset remnants" of the erosion surface north of Woodland Park along the Ute Pass fault zone (Steven and others, 1997). Uplift, whether by faulting or warping, was greater at the sites of the present mountains, although it also involved adjacent parts of the Colorado Piedmont, and tilting was a common effect, as

recorded by the inclination of the erosion surfaces (e.g., Steven and others, 1997, Fig. 3). The southern Front Range and adjacent Colorado Piedmont, for example, were tilted eastward, and the hingeline to such tilting is indicated by the fact that stream excavation becomes progressively less to the east and is nonexistent near North Platte, Nebraska (Steven and others, 1997). Only a tilt of a degree or two would be sufficient to create the mountains we see today (T.A. Steven, oral commun., 2001).

Late Cenozoic faulting has been proposed at one location in the mapped area. Following Scott and Wobus (1973), we mapped a Quaternary fault north of Rock Creek and west of State Highway 115 (Sec. 25, T. 15 S., R. 67 W. and Sec. 30, T. 15 S., R. 66 W.). Scott and Wobus (1973) even speculated that seismic shaking along this fault or parts of the Ute Pass fault zone could have brought down the older landslide, fan, and rockfall deposits (Q_{fro}) from Cheyenne Mountain. As also discussed by Kirkham and Rogers (1981), this fault appears to displace pediment gravel two (Q_{g2}). Unfortunately, this fault is poorly exposed and the two levels of pediment gravel could alternately be explained by two episodes of pediment cutting.

Does evidence exist nearby for late Tertiary or Quaternary block faulting? Dickson and others (1986) found no evidence for Quaternary movement along the northern end of the Ute Pass fault zone, but they did not evaluate Quaternary movement in or near the mapped area. Quaternary extensional faults, however, have been mapped by Scott (1970) and Dickson (1986) as close as the Rampart Range fault zone at the U.S. Air Force Academy less than 20 mi north of the mapped area. Kirkham and Rogers (1981) noted several subtle anomalous lineaments in

Quaternary rockfall deposits along the Ute Pass fault zone within 2 mi north of the mapped area, as well as deformed (faulted?) rocks along and east of the Ute Pass fault zone as close as 4 mi north of the mapped area.

Occasional small to large earthquakes take place along the Front Range, supporting a concept of local tectonism, including late Cenozoic faulting. Certainly contemporary tectonic stresses exist in the area, and one explanation for modern earthquakes is reactivation of Laramide or older faults (Wong, 1986; Warner, 1986). Warner (1986) concluded that the modern stress regime in the eastern Front Range is compressional and thus he suggested that Cenozoic movement along faults represents the decay of remnant stresses from Laramide tectonism. However, restudying of microearthquakes suggests that most Front Range earthquakes were extensional (R.M. Kirkham, written commun., 2002). Regardless of this interpretation, some faults in Colorado have been active clearly in the late Cenozoic; most of these strike northwest or north-northwest and they include the Ute Pass fault zone in the vicinity of Woodland Park (Dickson and others, 1986; Warner, 1986). Kirkham and Rogers (1981) concluded that, based on "circumstantial evidence," that "it is likely" that the Ute Pass fault zone moved significantly at least once during the Quaternary, and that it is potentially active. Certainly small earthquakes are not uncommon near it (e.g., Scott, 1970; Himmelreich, 1996a) so the potential for larger earthquakes exists. Seismicity is dangerous to human beings in itself, of course, but a more important hazard in the Colorado Springs area would be if an earthquake triggered landslides or lead to the failure of buildings and other structures.

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