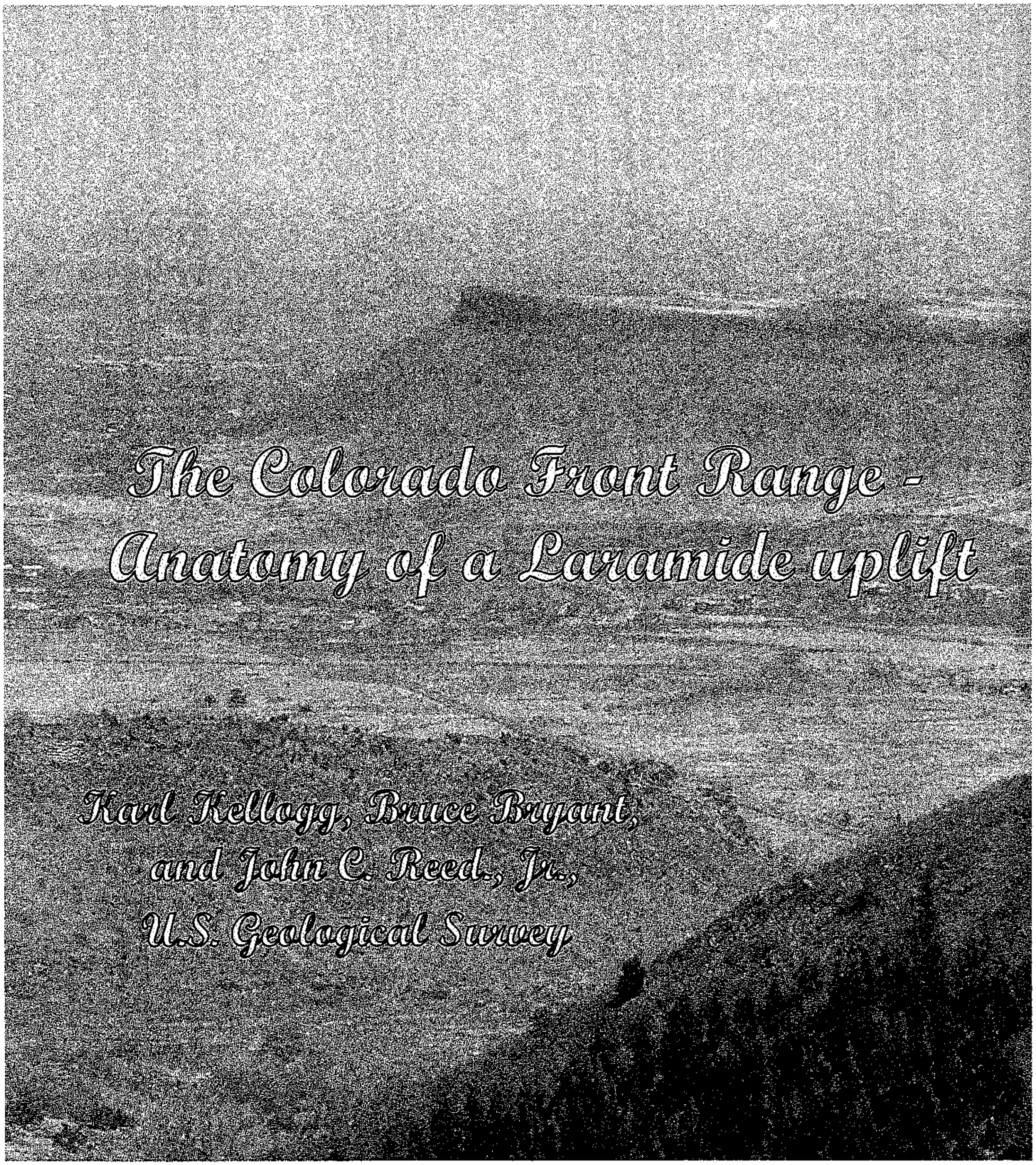

The Colorado Front Range - Anatomy of a Laramide uplift

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Cover photo: View to the southeast toward downtown Denver from Van Bibber Creek, about 2 miles (3 km) north of Golden. Tree-covered slope is underlain by Proterozoic metasediments; linear ridge is underlain by Dakota Group sandstone; isolated outcrop on right side of photo is vertical sandstone of Laramie Formation, which is just east of Golden fault; large mesa is topped by alkalic basalt (shonkinite) of Denver Formation.

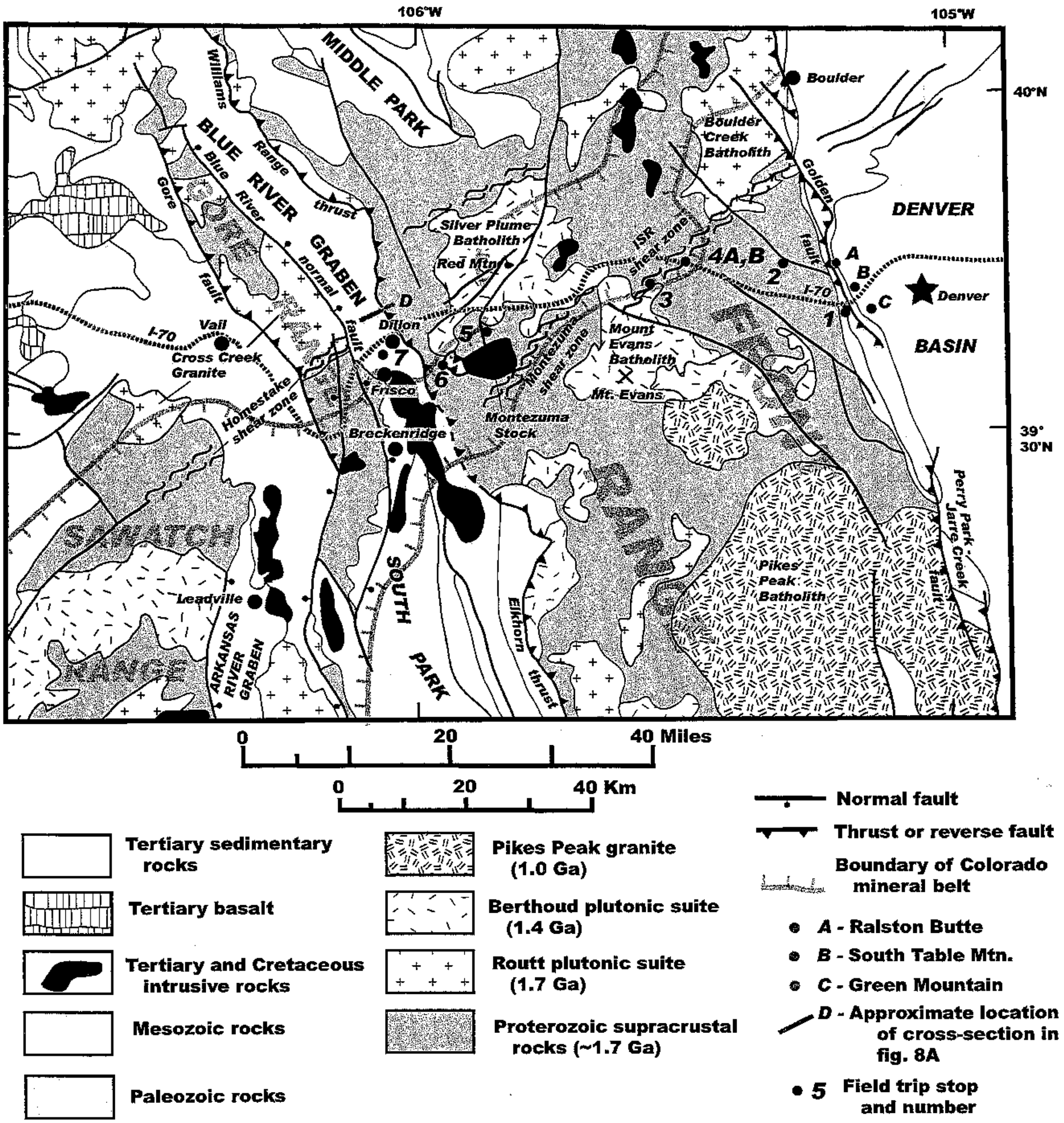

Figure 1. Sketch geologic map of the central Front Range showing field trip stops. "ISR shear zone" is the Idaho Springs-Ralston shear zone.

Introduction

This trip provides an overview of the geology of the Colorado Front Range, from the creation of continental crust during the Early Proterozoic to Neogene deformation and erosion, although the emphasis will be on the Laramide history. As we leave the downtown convention center and drive west toward the Front Range, the mountain front you see represents exhumed Laramide topography, stripped of a thick sequence of Upper Cretaceous to Miocene rocks during late Neogene erosion. As used here, the Laramide orogeny describes the tectonic events that occurred between about 70 Ma and 45 Ma (Late Cretaceous to middle Eocene), although most of the present topography is due to post-Laramide regional uplift and erosion. At the first stop (fig. 1), we will review the Laramide and subsequent history. As a head start, however, perhaps no clearer or more succinct overview of the Laramide orogeny has been given than that by Ogden Tweto (1975):

> "At the beginning of the Laramide orogeny, a blanket of undisturbed Cretaceous and minor older Mesozoic sedimentary rocks 1,500 to 3,000 m thick covered the Southern Rocky Mountains province, and the last of a series of Cretaceous seas was starting to withdraw northeastward across the region. Beneath the blanket was an older and inhomogeneous terrane that in some places consisted of eroded stumps of Late Paleozoic mountain ranges made up of Precambrian rocks, and in other places piles of sedimentary rocks thousands of meters thick on the sites of late Paleozoic basins. At the onset of the Laramide orogeny in Late Cretaceous time, most of the buried mountain ranges were re-elevated, and adjoining Laramide basins, in part inherited from late Paleozoic basins, began to subside and receive orogenic sediments...
>
> ....Once started, uplift of mountain units continued through Paleocene and into Eocene time, as indicated by nearly continuous Upper Cretaceous to Eocene sedimentary sequences in the interiors of the bordering basins. Uplifts grew laterally as they rose vertically. Consequently, the major uplifts today are really larger than those that supplied the first orogenic sediments and their border structures are younger than those sediments.
>
> ...After the Laramide orogeny, the Southern Rocky Mountain region stood somewhat higher than at the beginning of orogeny—when it was at sea level—but at a much lower level than it does today..."

Uplift of the Front Range began about 70 Ma, accommodated on the east side along a series of high-angle, mostly east-directed reverse faults, such as the Golden fault, and on the west side along a low-angle, west-directed structures, including the Williams Range and Elkhorn thrusts (fig. 1). Synorogenic conglomerates shed into the Denver basin (Raynolds, 2002) and South Park basin (Bryant and others, 1981a) record this uplift history. One of the goals of this trip is to compare the deformational styles of the east and west structural margins and consider why they are different.

Much of the following discussion is condensed and modified from Tweto (1975, 1987), Reed and others (1987, 1993), Erslev and others (1999), and Naeser and others (2002). Additional useful references are the Colorado state geologic map (Tweto 1979, the Denver 1°X 2° quadrangle (Bryant and others, 1981b) and the Leadville 1°x 2° quadrangle (Tweto and others, 1978).

Proterozoic history of the Front Range

The Front Range is one of numerous uplifts in the Rocky Mountain region in which Precambrian rocks are exposed. Metamorphic and plutonic rocks, including widespread migmatites, form the core of the Front Range and are part of a Proterozoic terrane called the Colorado province (Bickford and others, 1986) that is interpreted to have formed over an interval of about 130 Ma, beginning about 1,790 Ma. This long orogenic episode is thought to accompany early Proterozoic accretion of island arcs and back-arc basins to the southern margin of an Archean continent (Reed and others, 1987). The suture between the Proterozoic Colorado province and the Archean craton (Wyoming province) to the north is a deformed zone in southern Wyoming called the Cheyenne belt.

Proterozoic rocks of the Front Range include complexly folded and interlayered quartz-feldspar gneiss, amphibolite, biotite schist, and migmatite. The biotite-rich rocks, including migmatite, locally contain layers and lenses of marble, quartzite, and conglomerate, indicating sedimentary protoliths. Amphibolite and quartz-feldspar gneiss are abundant generally in separate areas from the metasedimentary rocks and probably originated as volcanic complexes. Both the metasedimentary and metavolcanic packages are complexly interlayered on a regional scale. The rocks are commonly metamorphosed to high-T, low-P, upper amphibolite assemblages sillimanite-K-feldspar, $\pm$ garnet, $\pm$ cordierite migmatites). Evidence for earlier high T, high P conditions locally have been reported (Selverstone and others, 1997; Munn and Tracy, 1992; Munn and others, 1993).

Details of the history of the Proterozoic rocks in Colorado are only locally well known. Small regions near Gunnison, west of the Sawatch Range, and Salida at the south end of the Mosquito Range, where the metavolcanic rocks are somewhat lower metamorphic grade, are better understood than those elsewhere (Bickford and others, 1989). On the route of our transect across the Front Range, no metavolcanic rocks have been dated despite considerable 1:24,000 scale mapping, and few convincing facing directions or regional marker horizons have been found in the supracrustal rocks. One exception is in Big Thompson Canyon, about 75 km north of our route, where relatively low-grade, low-strain metasediments are preserved and metamorphic isograds have been successfully mapped (Braddock and Cole, 1979). At most places in the Front Range, however, the rocks have been severely deformed in a ductile fashion, recrystallized, and partially melted. In Big Thompson Canyon, peak metamorphism, presumably during crustal accretion, occurred ~1,750 Ma. Compositions of cores and inclusions in garnet and plagioclase grains suggest pressures as great as 10 kb, followed by crystal growth under decreasing pressures down to 4-6 kb (Selverstone and others, 1997). The evidence for the early, extremely high pressures has been questioned (J.C. Cole, written commun., 2004; he asks: How do rocks descent to over 30 km depth and then ascend to mid-crustal levels?) and has not been reported elsewhere in the Front Range.

Coeval with or closely following the ~1,700 Ma metamorphic and deformational event, extensive batholiths and smaller bodies of mostly granodiorite and monzogranite, referred to as the Routt Plutonic Suite (Tweto, 1987), intruded the layered rocks of the Front Range. About 25 km north of our transect, the Boulder Creek Granodiorite, dated at 1,715 Ma (Premo and Fanning, 2000), is synchronous with late folding of the supracrustal rocks (Gable, 2000). Numerous bodies of plutonic rock have been megascopically correlated with this batholith, but many of the larger plutons and most of the smaller plutons have not been dated.

The Colorado province was extensively modified by widespread ~1,400 Ma intrusions, regional heating, and local deformation. Large and small plutons of granite and monzogranite of this age, called the Berthoud Plutonic Suite (Tweto, 1987), are widespread in the Proterozoic rocks of the Front Range. Near our transect, the Mt. Evans batholith of metaluminous granodiorite and monzogranite superficially resembles the Boulder Creek batholith but has a U-Pb zircon age of 1,442 Ma. (Aleinikoff and others, 1993). Part of the Mount Evans batholith has been mylonitized along the northeast-trending Idaho Springs-Ralston shear zone, but that mylonite zone does not extend through the batholith.

The Silver Plume Granite, a peraluminous biotite-muscovite granite and monzogranite forms a batholith that extends from the town of Silver Plume to west of the continental divide. The Silver Plume has a U-Pb zircon age of 1,422 Ma (Graubard and Mattison, 1990) and was derived by limited partial melting of lower crustal material and emplaced possibly as shallow as 8 or 9 km. (Anderson and Thomas, 1985). This batholith and similar ones contain many inclusions of country rock and have complex contacts composed of numerous dikes and irregular bodies intruded into the country rock.

The Middle Proterozoic (~1,400 Ma) magmatism reset the rubidium-strontium and potassium-argon isotopic systems (Peterman and others, 1968; Shaw and others, 1999). $^{40}Ar/^{39}Ar$ dates on muscovite and biotite are all 1,400-1,340 Ma, reflecting cooling through closure temperature after 1,400 Ma. $^{40}Ar/^{39}Ar$ dates on hornblende range from 1,600 to 1,390 and represent variable retention of radiogenic argon (Shaw and others, 1999).

Diabase dikes intrude the metamorphic and granitic rocks of the Front Range. None of them has been dated in the region of our transect, but in the northeastern Front Range similar dikes that both intrude and are intruded by plutons of the Berthoud Plutonic suite with indistinguishable ages of ~1,415 Ma. This age, therefore, must also be the age of the dikes. East-trending lamprophyre dikes are well exposed in Clear Creek Canyon along our traverse, but their age is not known.

Emplacement of the anorogenic Pikes Peak Granite batholith at ~1080 Ma (Unruh and others, 1995) in the southern Front Range marks the final major Proterozoic rock-forming event.

The Proterozoic rocks of Colorado are transected by a number of northeast- to east-trending, discontinuous, en echelon shear zones consisting of steeply dipping mylonitic and non-mylonitic rocks. Our transect (fig. 1) takes us across one of the major shear zones, the Idaho Springs-Ralston shear zone. Northeast of the Mt. Evans batholith in the Front Range, the Idaho Springs-Ralston shear zone marks a discontinuity in the trends of the major folds in the metamorphic rocks, although there is no major lithologic contrast across the zone. North of the zone, folds trend north-northeast, whereas south of the zone they generally trend northwest (Bryant and others, 1981b).

Recent study of another of the major shear zones, the Homestake shear zone west of the Front Range, shows that a zone of steeply dipping, highly strained but non-mylonitic rock was formed during the main metamorphism (~1,750 Ma), and mylonites and ultramylonites formed locally along the shear zone during or slightly after the ~1,400 Ma plutonic event. (Shaw and others, 2001). Similar evidence was obtained on the Idaho Springs-Ralston shear zone (McCoy, 2001). Latest movement on mylonites within the shear zones suggests southeast-up reverse movement (Braddock and Cole, 1979; Selverstone and others, 2000). The shear zones are interpreted to have first developed as a system of diffuse high-strain zones related to continental assembly (Shaw and others, 2001) of terranes to form the Early Proterozoic crust of the region.

Karlstrom and others (2002) have interpreted the rocks on the southeast side of the Idaho Springs-Ralston shear zone to be part of a 1,780-1,730 Ma Gunnison-Salida block and those to the northwest the 1,750-1,700, Ma Rawah block. According to this interpretation, a hypothetical belt of steeply dipping rocks of high metamorphic grade might connect the Idaho Springs-Ralston shear zone with the Homestake Creek shear zone, forming the northwest margin of the Gunnison-Salida block. However, no such belt is yet known. To prove the interpretation of Karlstrom and others, we need to know the details of the age of the volcanic rocks and the metamorphic and structural history of all the rocks on either side of the Idaho Springs-Ralston shear zone.

Paleozoic and Mesozoic history of the Front Range region

During the early Paleozoic, thin continental-shelf sequences of quartz-rich sands and carbonates were deposited in shallow seas over the region. In the late Paleozoic, northwest and north-northwest-trending mountain ranges and basins formed during the basement-involved Ancestral Rocky Mountain orogeny. Over the area traversed by this field trip, erosion during uplift of the Ancestral Front Range removed the earlier Paleozoic sedimentary cover, but these strata are preserved in adjacent basins, where they are overlain by as much as 5 km of Pennsylvanian and Permian, mostly clastic sediments eroded from the Ancestral Rocky Mountain uplifts. Along the east flank of the Front Range at our first stop (fig. 1), pre-Pennsylvanian rocks were eroded before deposition of about 500 m of arkosic sandstone and conglomerate of the Fountain Formation (fig. 2), which forms the "flatirons" at various places, such as west of Boulder and west of Denver near stop 1. The sediment forming the Fountain Formation was shed from the east flank of the Ancestral Front Range. Just south of our Front Range traverse, near Frisco (fig. 1), about 180 m of reddish Triassic?, Permian, and Pennsylvanian sandstone and conglomerate (tentatively correlated with the Chinle and Maroon Formations) overlie basement rock (Kellogg, 2000) and were shed off the west flank of the Ancestral Front Range.

Permian, Triassic, and Jurassic fluvial, eolian, and near-shore deposits overlie the thick clastic sequences derived from the late Paleozoic uplifts. By the middle Jurassic, the Ancestral Rockies had been eroded to low relief and were covered by fluvial and lacustrine deposits of the Morrison Formation, which are 60-100 m thick on both sides of the Front Range. Near the end of the Early Cretaceous, major subsidence coeval with a rise in sea level commenced with deposition of shoreline deposits (Dakota Group) of the western interior seaway over the entire Front Range, followed by over 2 km of marine shale and minor amounts of sandstone and limestone ("Benton Group," Niobrara Formation, and Pierre Shale).

The Laramide orogeny

The early stirrings of the Laramide orogeny, a 20 Ma period of crustal contraction, deformation, and igneous activity during which the present Rocky Mountains were built, were marked by renewed uplift of the Front Range region. The western interior seaway began to withdraw from the region after 69 Ma, the age of the youngest ammonite zone in the Pierre Shale (Scott and Cobban, 1965; Cobban, 1993). This age is based on $^{40}Ar/^{39}Ar$ dating of tuffs elsewhere in that ammonite zone (Obradovich, 1993). The Late Cretaceous-early Tertiary rocks overlying the Pierre Shale record the uplift history, starting with the regressive shoreline deposits of the Fox Hills Sandstone, followed by the coastal plain sandstones and coal beds of the Laramie Formation (from which the Laramide orogeny is named), in turn overlain by the fluvial conglomerates, sandstones, and claystones of the Arapaho Formation and Denver Formations (Raynolds, 1997, 2002). Uplift in this area was geologically rapid; only a few million years separate the age of the Late Cretaceous marine deposits of the Pierre Shale and the earliest conglomerates of the terrestrial Late Cretaceous Arapaho Formation, which contains clasts derived from Proterozoic basement rocks. During this short period, the newborn Rocky Mountains rose from the sea and over 2 km of sedimentary rocks were eroded.

West of Denver, andesitic (field term) debris derived from volcanoes somewhere to the west of the present mountain front forms a major part of the latest Cretaceous-earliest Paleocene Denver Formation. Calc-alkaline and alkalic dikes and stocks intruded into this part of the Front Range, beginning about 68 Ma and continuing until about 27 Ma, although no Late Cretaceous or early Tertiary extrusive equivalents of any of the intrusions are preserved in this region. (Volcanic rocks related to some of the younger intrusions are preserved in the northwestern part of the Front Range (O'Neill, 1981; Braddock and Cole, 1990).

Along our transect in the Golden area, the upper part of the Denver Formation contains ~65 Ma potassic basalt flows that probably erupted from a source a few km to the north (Ralston Buttes). On South Table Mountain, about 5 km northeast of stop 1 (fig. 1), the K-T boundary (65.4 Ma; Obradovich, 1993) is 71 m below these basalts. Paleomagnetic directions from the Ralston Buttes intrusive (fig. 1) are rotated, indicating that the body was emplaced before major movement on the Golden fault (Hoblitt and Larson, 1975). Near the summit of Green Mountain, about 2 km east of stop 1 and 240 m above the basalts, the Green Mountain Conglomerate, which overlies the Denver Formation, contains a 64 Ma tuff (Obradovich, 2002). The similarity of all these ages within a relatively thick sedimentary sequence attests to rapid sedimentation during the close of the Cretaceous and the opening of the Tertiary.

South of Denver, an hiatus in deposition from early Paleocene (~63-64 Ma) to early Eocene (54 Ma) separates two depositional sequences in the Denver basin (Raynolds, 2002; Obradovich, 2002), whereas in the South Park basin, on the west side of the

Front Range, data suggest deposition occurred throughout the Paleocene followed by late Laramide deformation in early Eocene time (Bryant and others, 1981a). The ~80-km-wide Denver basin is highly asymmetric; the structurally deepest part, where the basement rocks are about 2,000 m below sea level (Haun, 1968), is only 8 km east of the mountain front. The elevation difference between the highest basement rocks of the Front Range (4,300 m above sea level) and the surface of the buried basement beneath the Denver basin indicates more than 6,300 m of structural relief.

By the close of the Laramide orogeny, during the early Eocene (about 50 Ma), erosional debris derived from the Laramide Front Range uplift formed a sedimentary apron, now largely eroded, that lapped onto the flanks of the range. At only one place along the east side of the Front Range-Laramie uplifts (the Laramie Range in Wyoming forms a northern extension of the Front Range), at the Gangplank near Cheyenne, is a remnant of the complete apron of sedimentary rocks preserved.

On the east side of the Front Range at the latitude of Denver, the principal Laramide structure is the Golden fault. Seismic sections and a few well data indicate that the Golden fault dips about 50°-70° to the west and has 2-3 km of eastward thrust displacement (Weimer and Ray, 1997) (fig. 3). About 30 km north of I-70, just south of Boulder, there is a transition in the structural style. North of the transition, ENE dipping backthrusts bring the basin side up and the range side down and may sole into a blind extension of the Golden fault in the basement (Erslev and Selvig, 1997).

The Williams Range thrust, the western structural boundary of the Front Range, is a low angle thrust with a minimum lateral displacement of 9 km, as shown in a window through the thrust (stop 6) related to uplift by the 38 Ma Montezuma stock (fig. 1) (Erslev and others, 1999). This thrust is traceable for 100 km along the west margin of the Front Range, and in South Park a similar structure (Elkhorn thrust) is mapped along strike for 35 km. Tertiary intrusive rocks and younger deposits conceal the probable connection between the Williams Range and Elkhorn thrusts. In the western Front Range, many faults and fractures cut the Proterozoic rocks and little is known about their history, although Kellogg (2001) suggests that the fractures may be related to the large overhang along the Williams Range thrust (to be discussed at stop 5).

Thrusts bounding both margins of the Front Range have led to many speculative models for the subsurface architecture (fig. 4). We interpret evidence in this area to indicate that the thrust geometry is significantly different on the east and west margins of the uplift; a viable model for uplift must explain this asymmetry.

The Colorado Mineral Belt

Our transect passes through the northeastern part of the Colorado Mineral Belt (CMB), a northeast-trending zone of Late Cretaceous and Tertiary (68-27 Ma) calc-alkalic and alkalic stocks and dikes, some of which are associated with several world-class ore deposits. The northeast-trending zone of Proterozoic shears discussed previously are interpreted to provide a zone of crustal weakness along which the intrusive rocks were emplaced and ore deposits formed (Tweto and Sims, 1963).

The Central City district near Idaho Springs (stop 3) consists of a ~59 Ma zoned hydrothermal system (fig. 5) having a core of gold-bearing pyrite-quartz veins, an intermediate zone of pyrite veins carrying copper, lead, and zinc sulfide minerals, and a peripheral zone of galena-sphalerite-quartz-carbonate veins (Sims and others, 1963; Sims, 1988; Rice and others, 1982). The ~37 Ma Silver Plume-Georgetown district, west of the Central City district (our route traverses this district), has silver-lead-zinc-bearing quartz-carbonate veins and some gold-silver veins trending east to northeast and controlled by several sets of steeply dipping fractures (Bookstrom, 1988; Bookstrom and others, 1987).

Thermal and uplift history as revealed by apatite fission tracks

Apatite fission-track studies (Bryant and Naeser, 1980; Naeser and others, 2002; Kelley and Chapin, in press) have shown that southeast of the Mineral Belt the top of the Laramide apatite annealing zone (a "fossil isotherm," where T>100-110° C during the Laramide, so that ages in the annealing zone are 70-50 Ma, while those above the zone are older) can be located. Laramide or post-Laramide movements on faults can be determined by displacements of that zone. North of the Mineral Belt most of the Front Range is in the Laramide annealing zone. This may be due to a combination of heat introduced by magma into the crust along the mineral belt and by a northward increase in the thickness of the Pierre Shale (Kelley and Chapin, in press). East and south of Mt. Evans, south of the Mineral Belt, the top of the Laramide annealing zone is at about 3.5 km altitude (Bryant and Naeser, 1980; Kelley and Chapin, in press). Assuming a normal geothermal gradient, rough calculations project the top of the Precambrian basement above Mt. Evans (the top is 4.3 km above sea level) at about 5 km, which would place the Late Cretaceous seafloor at 7.5 km altitude (Bryant and Naeser, 1980). This is about 6.5 km above the deepest part of the Denver basin. South of Mt. Evans, the top of the Laramide annealing zone dips gently east, suggesting eastward tilting. In the western Front Range north of the Mineral Belt, no Laramide annealing zone was found (ages are younger than Laramide). Anomalously young (as young as 6 Ma) Oligocene and Miocene annealing ages in the western Front Range and the Gore Range are interpreted to be due to heating, uplift, and cooling along the northern extension of the Rio Grande rift (Blue River graben) (Naeser and others, 2002).

Post-Laramide modification of the Front Range

A widespread erosional surface formed across the Front Range during the late Eocene (Epis and Chapin, 1975). The surface was not a peneplain, but consisted of a hilly topography, presently between about 2,200 and 2,600 m, that forms a bench-like surface visible from Denver. High peaks similar to those along the Continental Divide protruded above this erosional surface, as they do now.

Renewed uplift of the western Front Range and probable eastward tilting during the Miocene (Naeser and others, 2002) led to erosion of the range and deposition on the plains of the Ogallala Formation, an important aquifer in the region. High-level gravel deposits found on numerous ridges in the Front Range east of the continental divide are probably coarse-grained proximal equivalents of the Ogallala Formation. Between the late Miocene and present, widespread erosion stripped away much of the sedimentary cover from the west side of the Denver basin exposing the more resistant early and middle Proterozoic rocks.

One of the current controversies concerns the Neogene uplift history of the Front Range. One school of thought holds that the range has been at essentially its present elevation since at least Eocene time and that the profound erosion and canyon cutting along the eastern flank is an outcome of Pliocene to Recent climate change (Gregory and Chase, 1992; Molnar and England, 1990). Evidence for this model is from paleobotanical observations on mid-Tertiary leaf morphology, which suggests paleotemperatures and, therefore, lapse rates (Gregory and Chase, 1992). Another school, favored by us, advocates that a broad region, centered on the Rio Grande rift, has been uplifted by as much as a thousand meters during late Miocene to Recent time (Eaton, 1986, 1987), leading to eastward tilting of the Front Range (Steven and others, 1997; McMillan and others, 2002), and that canyon cutting along the east side of the Front range is an outcome of this uplift.

How much of the Front Range has eroded since Laramide uplift? The mining industry has provided important data that indicate that the western Front Range either rose significantly during the late Tertiary or contained mid-Tertiary peaks significantly higher than they are today (Geraghty and others, 1988). A plug of porphyritic rhyolite forms the 3,700 m summit of Red Mountain (fig. 1) and is underlain at depth by the 27-30 Ma Red Mountain intrusive system, now being exploited by the Henderson molybdenum mine. Minimum homogenation pressures were obtained from vapor-rich fluid inclusions collected underground at 2,500 m altitude in distal parts of extensive open-space veins. These pressures indicate 2,900 m of overburden at the time the inclusions formed, which suggests that at the time of intrusion the ground surface was about 2,700 m above the present top of Red Mountain, or at 5,400 m. Either the Front Range was significantly higher during the Oligocene or, more likely, significant uplift has occurred since the Oligocene. In either case, about 1.5 km has been removed from above the summit Red Mountain during Neogene erosion. Detailed paleomagnetic study also shows that the intrusions and associated ore system in the Henderson mine have been tilted east 25° between two faults (Geissman and others, 1992).

Directly west of the Front Range, the Blue River half graben represents the northernmost portion of the Rio Grande rift (fig. 1); extensional faults coeval with the rift extend to the Wyoming border. The Gore Range bounds the half graben on the west. Remnants of valley fill in tilted fault blocks within the graben are as old as late Oligocene. Late Miocene apatite fission-track dates from the eastern part of the Gore Range show that significant displacement occurred along that side of the graben in late Neogene time (Naeser and others, 2002), although the young ages may be due, in part, to an increased geothermal gradient.

The higher parts of the Front Range were carved by glaciers during numerous during the Pleistocene and perhaps the late Pliocene. The last two major glaciations are the Pinedale (16-23 Ka) and Bull Lake glaciations (95->130 Ka) (Chadwick and others, 1997). Detailed knowledge of the glacial history is sketchy for events older than the Bull Lake glaciation.

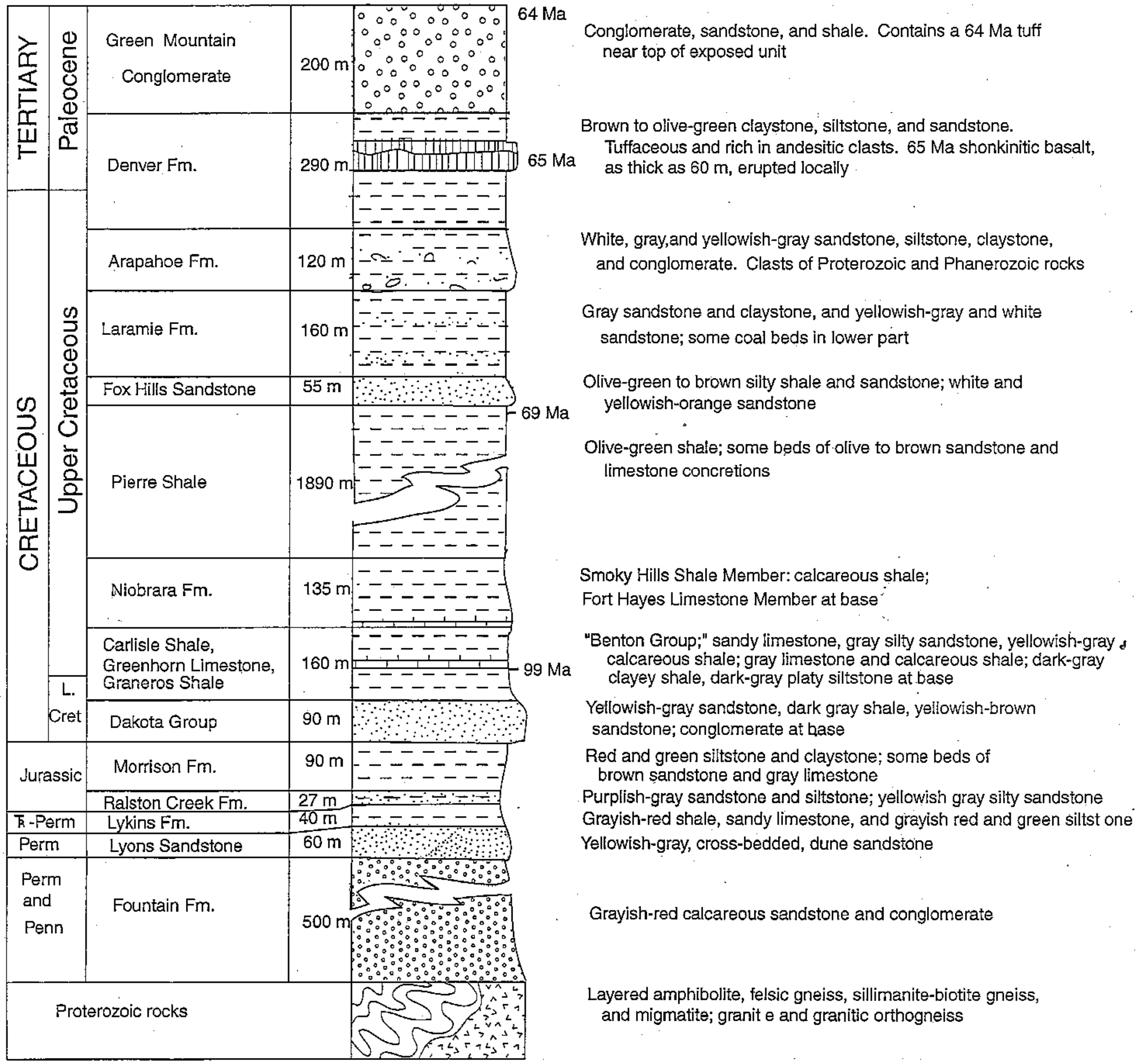

Figure 2. Stratigraphic column of Phanerozoic units along the east flank of the Front Range near Denver (Scott, 1972). West of the Front Range, the Morrison locally rests directly on Proterozoic basement.

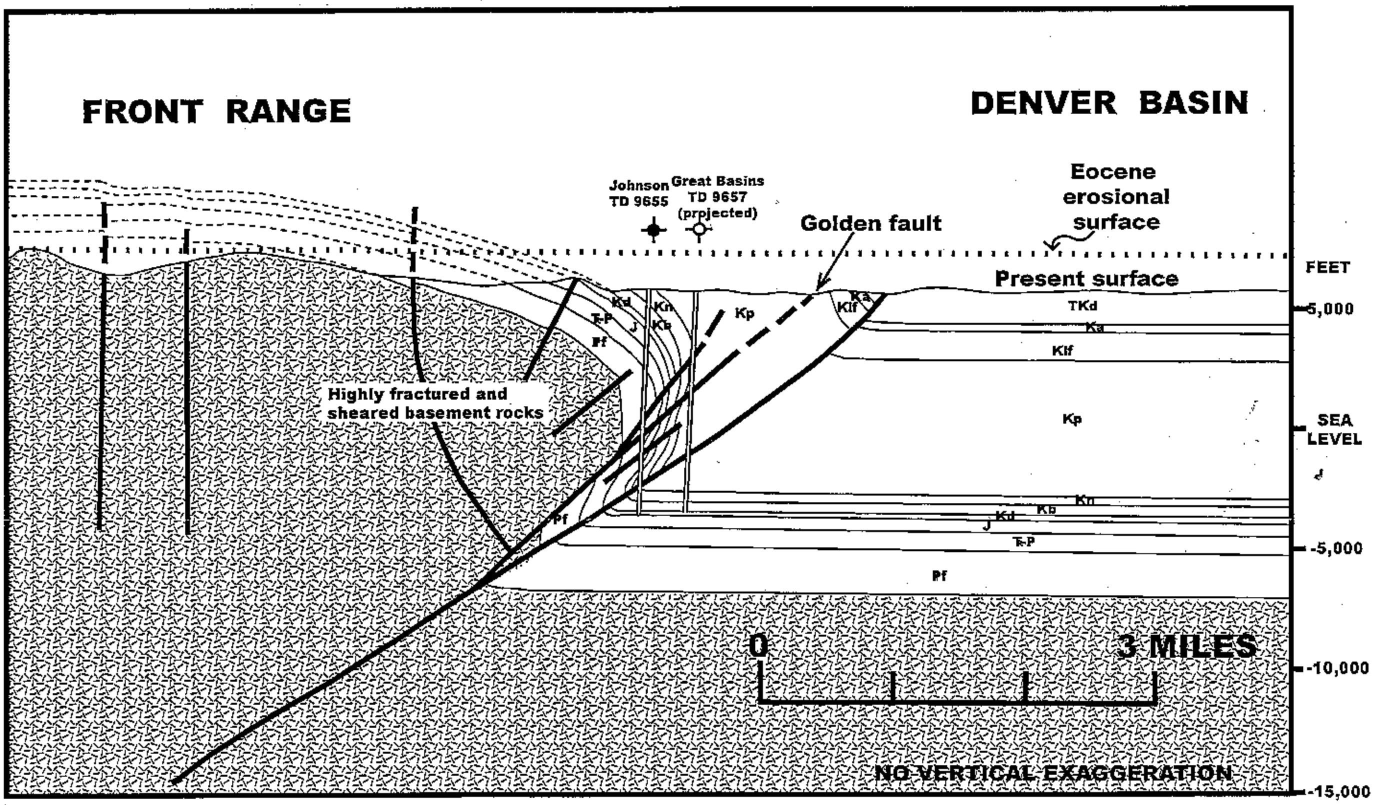

Figure 3. Structural cross section across east flank of the Front Range about 2 km south of stop 1. Adapted from Weimer and Ray (1997).

FIELD TRIP STOPS
All stops are indicated on figure 1

Stop 1. Dakota hogback north of Morrison--the eastern margin of the Front Range uplift

Driving instructions: Follow 6th Avenue west out of Denver to Interstate 70 (I-70). Continue west to exit 259 (for Morrison). After exiting, turn south onto Route 26, go under the I-70 overpass, and take an immediate left turn into the parking lot. Climb the ridge east of the parking lot for a panoramic view of the eastern Front Range.

This ridge and spectacular road cut just below this point exposes Jurassic Morrison Formation and Cretaceous Dakota Group rocks (fig. 2) upturned above the Golden fault, an east-directed reverse fault that lies mostly buried by surficial deposits at the eastern base of the ridge. To the west, the Pennsylvanian and Permian Fountain Formation (arkosic sandstone and conglomerate) forms the dramatic "flatirons" along portions of the eastern foothills of the Front Range. The Fountain Formation unconformably lies on early Proterozoic supracrustal and intrusive rocks that form the core of the Front Range. To the east, Cretaceous and Tertiary rocks comprise the near-surface rocks of the Denver Basin

The Golden fault is a west-dipping reverse fault (Berg, 1962; Weimer and Ray, 1997) that carried basement rocks over the Cretaceous rocks of the interior seaway. Figure 3 shows a cross section across the Golden fault at Turkey Creek, about 6 km south of here. Just east stop 1, the Golden fault places Upper Cretaceous Benton Group rocks (fig. 2) over the upper part of the Upper Cretaceous Pierre Shale, cutting out much of the Benton Group, all of the Niobrara Formation, and about 1 km of the Pierre Shale (Scott, 1972). The Golden fault is one of a series of en echelon west-dipping reverse faults that mark much of the eastern flank of the Front Range. It may be connected, via a blind (subsurface) fault, with the Perry Park – Jarre Creek thrust to the south (fig. 1). The highly asymmetrical shape of the southern and central Denver Basin can be attributed to thrust loading along these and similar faults that mark the west side of the basin.

A profound change in the style of Laramide faulting, visible from stop 1, occurs about 15 miles (24 km) to the north, near Boulder, where the eastern margin of the Front Range steps eastward, giving the margin a more northerly average orientation. A series of these steps is created by northwest-striking, southwest-directed back thrusts that expose basement in the core of fault-propagation folds (Erslev and Rogers, 1993). Reflecting this change in structural style, the northern Denver Basin is both shallower and more symmetric than the southern Denver Basin, and the basin axis is farther from the range.

It is important to remember that the eastern margin of the Front Range is an erosional relic of Laramide uplift and thrust or reverse faulting that placed resistant Proterozoic rocks against more easily eroded sedimentary rocks. Despite the linearity of the mountain front, extension has not played a significant role in forming this topography.

A wide spectrum of models has been proposed to explain Laramide deformation (fig. 4). Earlier vertical-tectonic models (e.g., Prucha and others, 1965; Sterns, 1978; Matthews and Work, 1978; Robinson and others, 1974; Jacobs, 1983) have largely been superceded, with some reservations, by models that invoke horizontal crustal shortening (e.g., Kluth and Nelson, 1988; Gries, 1983; Erslev, 1986, 1993; numerous papers in Schmidt and Perry, 1988; Raynolds, 1997; Erslev and Selvig, 1997).

Stop 2. Ductile (and brittle) structures in Clear Creek

Proceed north for 1.1 miles on Route26. Turn left just past the Heritage Square turnoff, proceed north for about one mile, and turn left at 6th Avenue. Proceed about 2.5 miles to Clear Creek (junction of U.S. Highway 6). Turn left up the canyon for 6.6 miles to Tunnel No. 2. Make a left turn into a parking area immediately upon exiting the west end of the tunnel (be very careful of oncoming and following traffic!).

This part of Clear Creek Canyon lies in a complex, two-mile wide transition between predominantly biotitic rocks mapped by Sheridan and others, (1967) in the Ralston Buttes quadrangle to the north and predominantly felsic gneiss and amphibolites mapped by Sheridan and others (1972) in the Evergreen quadrangle to the south. The units mapped in this zone are based largely on visual estimates of relative proportions of interlayered quartz-feldspar gneiss, hornblende gneiss, amphibolite biotite gneiss, and calc-silicate rocks, which underscores the difficulty of defining map units in this country.

Rocks exposed at this stop are chiefly fine- to medium-grained biotite gneiss, medium-grained strongly foliated migmatitic gneiss and amphibolite, all cut by dikes of pegmatite. The general relations are well exposed on the cliff south of Clear Creek. The various lithologies can be seen in outcrops and talus blocks at the east end of the parking area.

The biotite gneisses consist of various proportions of quartz, oligoclase, biotite, minor amounts of potassium feldspar, and accessory garnet, apatite, and magnetite. Some also contain appreciable amounts of hornblende. Sillimanite is found in similar rocks nearby. Migmatitic phases are characterized by wavy, 1- to 3-cm-thick folia of quartz, plagioclase, and pink potassium feldspar. These folia are parallel to foliation in the enclosing gneisses, and generally display biotite-rich selvages a few millimeters thick. Hedge (1972) showed that the migmatites bear no direct relation to plutons of the ~1,700 Ma Routt Plutonic Suite, and they pre-date the plutons. He concluded that they formed by local partial melting during high-grade regional metamorphism. Olsen (1982) found that some of the migmatites in Clear Creek Canyon formed by partial melting in a closed system, while others required some introduction of granitic or tonalitic material from external, but perhaps nearby, sources. She estimated that the migmatites formed at about 4 kb at temperatures of 650-700°C. Less than half a mile south of here, biotite-hornblende gneiss, similar to some of the rocks exposed here, contains large porphyroblasts of orthopyroxene, suggesting conditions transitional between upper amphibolite and granulite facies (unpublished data). About 11 km to the northeast, pelitic schist contains retrograde muscovite and large undeformed porphyroblasts of andalusite resulting from a low-pressure regional metamorphism at

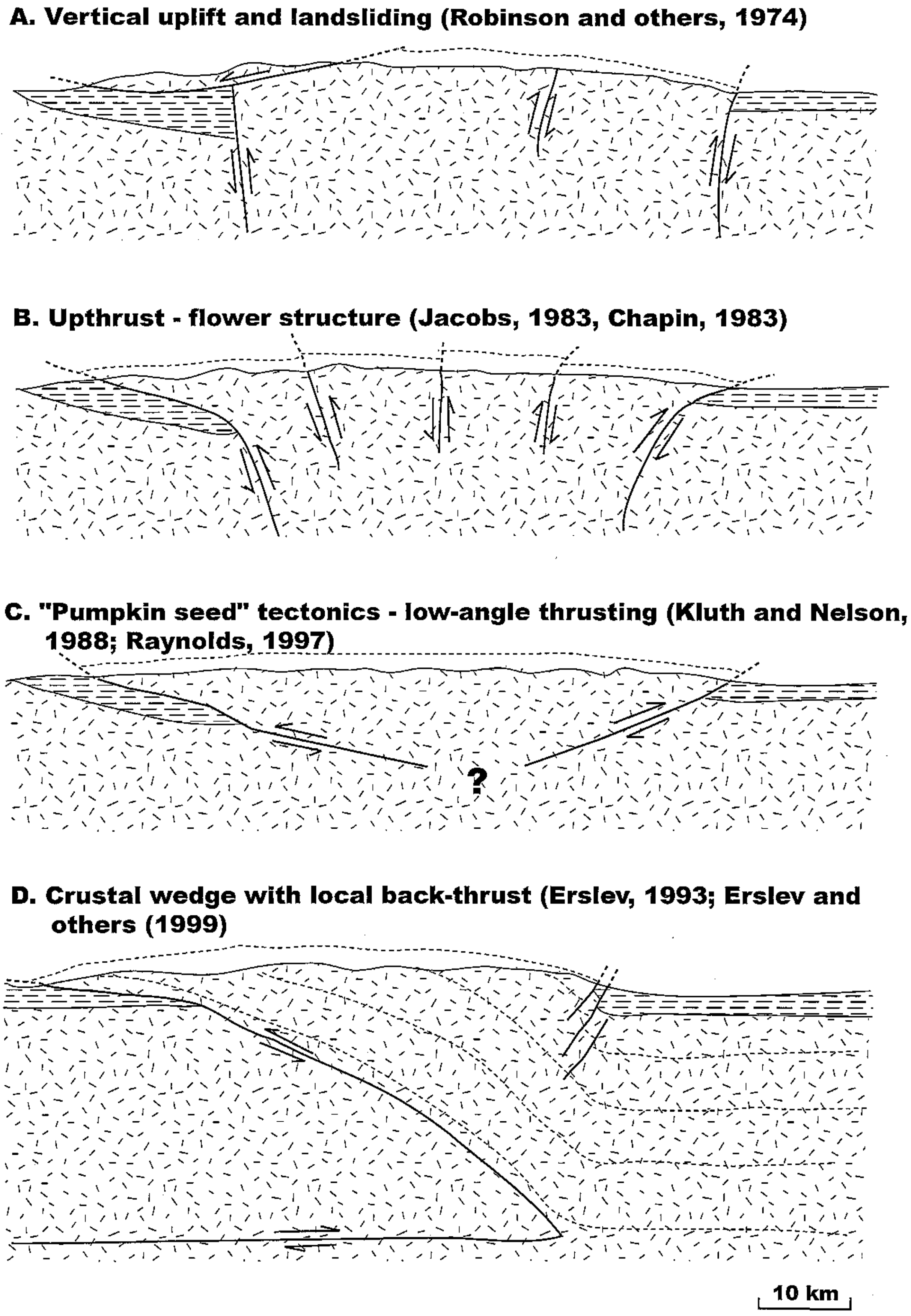

Figure 4. Models for Front Range uplift, adapted from Erslev and others (1999)

about 525°C (Shaw and others, 1999). The effects of this event have not yet been found in Clear Creek Canyon.

In the vicinity of this stop, layering and layer-parallel foliation strike N70°W and dip 40 to 60° southwest. A strong mineral lineation defined by biotite streaks plunges about 10°W. Axes of intrafolial isoclines with amplitudes of meters to tens of meters lie approximately parallel to this lineation; one of them is exposed in the cliff at the eastern end of the parking area. Note that the sense of the small folds in the migmatite folia on the limbs of this isocline is consistent with their having formed as drag folds during formation of the isocline. The complicated pattern of amphibolite pods and layers in the cliff south of the creek is largely due to the fact that the outcrop face is subparallel to the isoclinal axes. The pattern is quite similar to the general pattern of units shown on the geologic maps of the area (Sheridan and others, 1967; Sheridan and others, 1972) suggesting that much of the complex outcrop pattern is due to early layer-parallel isoclinal folding, and that lithologic sequences in these rocks may have little stratigraphic significance. A second generation of larger, more open folds of various trends commonly depicted on 1:24,000 maps affect layering and layer-parallel foliation on the limbs of isoclines of this type.

The layered gneisses are cut by dikes of pegmatite that range from sharply discordant to semi-concordant. Some are folded and boudined, but others seem to post-date ductile deformation of the wall rocks. The pegmatites consist of quartz, potassium feldspar, and plagioclase; some contain biotite and conspicuous knots of magnetite. There are no obvious differences in mineralogy and texture between those that are deformed and those that are not. The lack of muscovite and tourmaline in the pegmatites here (these minerals are found in pegmatites of some Silver Plume granites), and the fact that at least some are clearly deformed suggests that they are related to the ca 1.7 Ga Routt Plutonic Suite, rather than to the ca 1.4 Ga Berthoud Plutonic Suite.

Discordant dikes of dark hornblende lamprophyre also cut the Precambrian rocks here. An offshoot of one of these is conspicuously exposed on the north side of the highway about 15 m west of the tunnel portal. The main dike, about a meter thick, is sub-parallel to the face of the cut and just above it. It is exposed at road level about 50 m west of the portal. The same dike makes conspicuous notches in the ridge above tunnel #2 and in the ridge above tunnel #3, about 0.7 km to the west. It is not clear whether these dikes are Precambrian or Tertiary.

Stop 3. The Idaho Springs-Ralston shear zone in Chicago Creek valley

Proceed 5.0 miles west on Route 6, turn left at the stop light (junction of highway 119), and continue 3.1 miles onto Interstate 70. Drive 4.5 miles on I-70 to the Mt. Evans exit (exit 240) for Idaho Springs exit, turn left (southwest) and continue about 3.2 miles up the Chicago Creek valley to a wide pullout on the left (near Big Spruce Cabins). For this area, see the Idaho Springs quadrangle geologic map (Widmann and others, 2000).

These exposures are near the southeastern side of the Idaho Springs-Ralston shear zone. The Idaho Springs-Ralston shear zone is 1-2 km wide and extends 35 km southwest from the mountain front south of Boulder to where it dies out in the Mount Evans batholith, about 7 km southwest of here (fig. 1). This zone is typical of the numerous northeast-trending shear zones that cut Proterozoic rocks of the Front Range and adjacent ranges. These shear zones have a long and complex history. Recent studies, including monazite dating, on the Homestake shear zone in the Sawatch Range (Shaw and others, 2001) and on the eastern exposures of this shear zone (McCoy, 2001) show that the zones were established during the Early Proterozoic metamorphism (about 1,650 to 1,700 Ma) and that they were reactivated in the Middle Proterozoic (about 1,400 Ma) to form ductile mylonites and later brittle-deformational zones.

Planer, closely spaced, steeply dipping gneiss and schist that strike consistently northeast characterizes the shear zone. The younger generation of mylonitic and brittle-deformed rocks is not exposed at this stop. The rocks are layered biotite-quartz-feldspar gneiss containing some thin layers and partings of biotite schist and numerous stringers, lenses, and pods of pegmatite. The rocks strike N60° E, parallel the Chicago Creek valley, and dip steeply northwest. Locally the layers are in folds ranging from isoclinal to open and steep northwest-plunging mineral lineations are widespread. West along the road is a zone where dips are gentler, and the foliation is folded. This may be a "lump" (phacoid) in the zone where rock escaped much of the deformation. Alternatively, it may represent the hinge zone of a large fold.

A 3-m-thick sill of Tertiary porphyry containing altered alkali feldspar phenocrysts in an aphanitic matrix is well exposed. Sills of Tertiary igneous rock are numerous in the Idaho Springs-Ralston shear zone where it crosses the Central City-Idaho Springs mining area. On the west side of the sill is a 4-5 m-thick layer of mica schist that has been sheared and retrogressed to form a biotite-sericite phyllonite containing porphyroclasts of biotite and muscovite.

At the west end of this stop (where the road curves left), the pegmatites are thicker. At first glance they appear to be undeformed, but close examination shows that they are foliated parallel to the shear zone, so they did not entirely escape shear-zone deformation. Farther west (around the left curve in the road) is more biotite-quartz-feldspar gneiss with steeply dipping lineations, and pegmatite that appears to form tectonic lenses in the sheared gneiss.

Stop 4A. The Colorado Mineral Belt and the Central City and Idaho Springs mining districts

Return to Idaho Springs. Continue over I-70 and two blocks through town. Turn right on Colorado Blvd. and 0.6 miles to a grassy park. The impressive mill and dumps for the Argo Tunnel are visible across Clear Creek.

The Colorado Mineral Belt is an irregular belt that extends northeastward across the mountainous part of Colorado from the western San Juan Mountains to the east flank of the Front Range north of Boulder. The belt contains most of the major metallic-mining districts in the state (major exceptions are the gold-silver districts at Cripple Creek and Silver Cliff). The locus of intrusion of the igneous rocks of the mineral belt seems to be related to a zone of crustal weakness marked by the northeast-trending ductile shear zones in the Precambrian rocks (Tweto and Sims, 1963). The belt of intrusives has been interpreted as the expression of a large subjacent batholith or series of batholiths, a suggestion that is consistent with the fact that the mineral belt is nearly coincident with

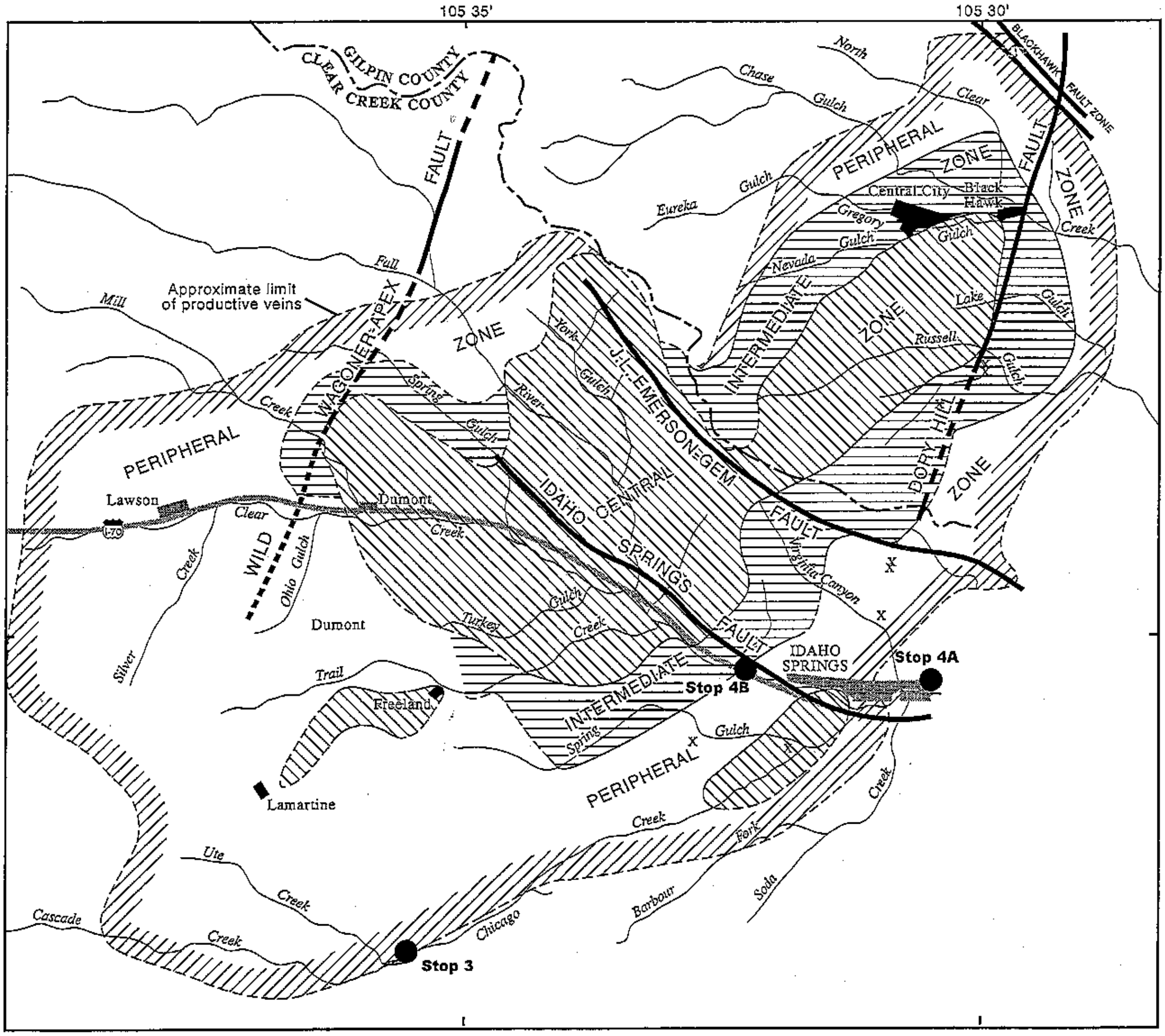

Figure 5. Map showing the zonal arrangement of ores in the Idaho Springs-Central City area. Adapted from Sims and others, (1963).

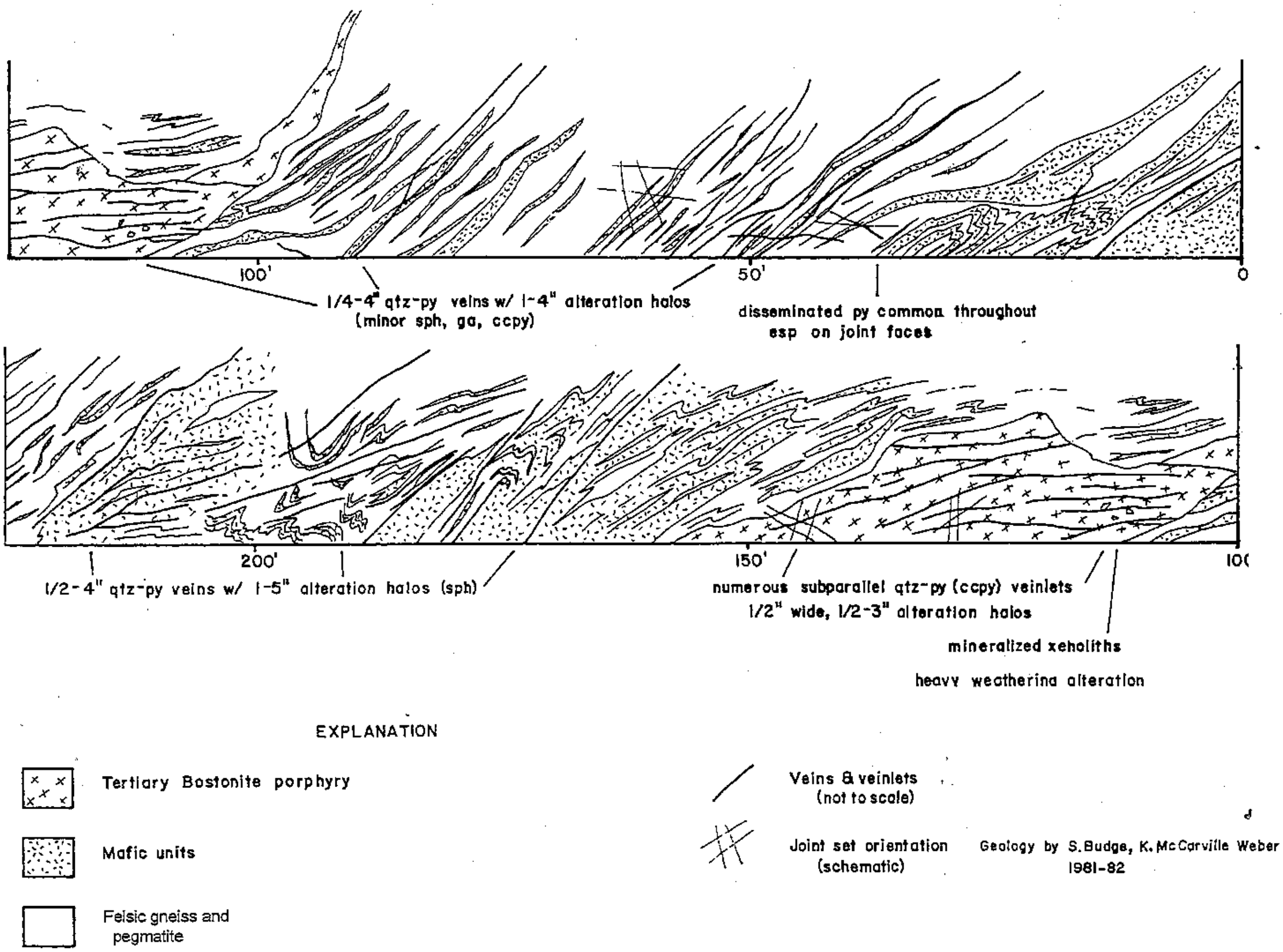

Figure 6. Detailed section across the eastern 230 ft (70 m) of Interstate-70 road cut at the west end of Idaho Springs (Budge and others, 1987).

a major gravity low, one of the lowest Bouguer gravity minimums in the United States (Tweto, 1975).

The Front Range portion of the mineral belt lies in the area shown on figure 1 and is characterized by abundant stocks and dikes of Laramide and post-Laramide age emplaced into Proterozoic rocks. These intrusive bodies have a wide range of composition and a long and complicated history of emplacement. The earliest intrusions in the Front Range region are chiefly monzonites and granodiorites, followed by more alkalic intrusives, all emplaced during the interval from 68 to about 54 Ma (Rice and others, 1982). Ore deposits at Central City and Idaho Springs may have developed from a short-lived (~1 Ma), complex hydrothermal system that developed 59 Ma at the end of the monzonitic and granodioritic sequence of intrusive activity. Farther to the southwest, much larger granitic plutons emplaced at about 40±5 Ma (Late Eocene), include the Montezuma stock, intrusions in the Breckenridge district, and a batholith in the Sawatch Range. Hydrothermal systems that formed the base and precious metal deposits at the Georgetown, Silver Plume, Montezuma, and Breckenridge districts are associated with these younger granitic intrusives.

Some of the largest ore bodies in the mineral belt are associated with the youngest intrusive (30-25 Ma) in the Front Range region, many of which are high-silica alkali granite or rhyolite porphyries (rhyolite-A suite) (White and others, 1981). These include the world-class molybdenum deposits at Climax and Henderson.

The following discussion of the Central City and Idaho Springs mining districts (fig. 5) is slightly modified from Sims (1988) which, in turn, was condensed from Sims (1983) and Sims and others (1963). Both districts are parts of the same ore system. The Central City district produced about 70,000 metric tons of lead, 4,300 metric tons of copper, 8,700 metric tons of zinc, 28,350 kg of gold, 153, 630 kg of silver, and small amounts of uranium. Gold accounted for about 85 % and silver 10% of the value of the ore. Production for the Idaho Springs district is about two thirds of that of Central City district, and silver was more important than gold. Production has been small since 1914.

Mining in the area played an important role in the development of the Rocky Mountain region. Placer gold was discovered near the present site of Idaho Springs in January, 1859, and in May of the same year the first lode discovery was made in Gregory Gulch, between the present towns of Black Hawk and Central City (about 6 km north of here). During the early years of the camp, mining was confined to placer gravels and surface gossans that developed on weathered veins. Later, with the development of better milling and smelting processes, mining took place below the oxidized zone. In 1900 it reached a maximum depth of 685 m in Central City. The Argo tunnel was started at Idaho Springs in 1893 to intersect the veins at depth and provide drainage and easy haulage to the mill at Idaho Springs. The 7.36-km-long tunnel was essentially completed by 1907, but for many reasons it did not stimulate significant new ore production. In 1942, the tunnel was closed because of accidental breakage into old water-filled stopes on a major vein. This that resulted in a disastrous flood and the tunnel was largely blocked by stope fill. It has never been reopened.

The host rocks in the area consist of Early Proterozoic biotite-quartz gneiss, microcline-quartz gneiss, amphibolite, migmatite, and intrusive rocks. Granodiorite and quartz diorite (~1,700 Ma) and two-mica granite (~1,400 Ma) intruded the layered rocks. A slightly younger, approximately 2-km-wide, northeast-trending zone of ductile deformation (Idaho Springs-Ralston shear zone) traverses the southeastern edge of the district. The country rocks in the area are cut by abundant faults, some of which exploited Early Proterozoic structures that were reactivated during the Laramide orogeny. Most of the faults, therefore, are Laramide in age and formed shortly before the mineralization.

The ore deposits in the Central City and Idaho Springs area are sulfide-bearing quartz veins that contain precious metals, base metals, and sparse uranium. They were formed about 59 Ma ago (Rice and others, 1982), and are related to dikes, sills, and small stocks of porphyritic igneous rocks, although they postdate intrusion. The veins are hydrothermal fillings in faults, and are similar in mineralogy, texture, and structure to the deposits classified by Lindgren (1933) as mesothermal. The principal ore minerals are pyrite, sphalerite, galena, chalcopyrite, and tennantite. Less abundant are enargite, telluride minerals, and molybdenite (Rice and others, 1985). The gangue minerals include quartz, barite, fluorite, and rhodochrosite. A portion of the area may represent the upper part of a younger, alkaline porphyry molybdenum system (Rice and others, 1985).

The ores of the area have a well-defined concentric hypogene mineral zoning (fig. 5). A large, central zone containing pyrite-quartz veins with gold is surrounded by an area (intermediate zone) of pyrite-type veins that carry copper, lead, and zinc sulfide minerals. The peripheral zone contains predominantly galena and sphalerite in quartz-carbonate veins.

Stop 4B. Hands on the mineral belt – west end of Idaho Springs

Driving instructions: Drive west on Colorado Blvd. 1.2 miles back through Idaho Springs to a large parking area on the left (south) side of the road just before the entrance to I-70. Do not enter I-70. We will visit the large road cut on the north side of the road.

The rocks along the road cut are mainly migmatitic quartz-microcline-plagioclase-biotite gneiss ("felsic gneiss"), amphibolite, and pegmatite. The layered rocks may represent a metamorphosed bimodal (rhyolite-basalt) volcanic sequence and are folded about northeast-trending axes, generally parallel to the trend of the Idaho Springs-Ralston shear zone. Several irregular dikes of bostonite porphyry (leucocratic alkalic felsite porphyry; contains K-feldspar or sodic plagioclase phenocrysts and very few mafic minerals) and numerous mineralized veins cut the Proterozoic rocks. Oxidation of the widespread pyrite has coated much of the outcrop with an iron-oxide stain, obscuring some of the features. This locality is in the intermediate zone of the district that contains both quartz-pyrite veins and galena-sphalerite-chalcopyrite-bearing veins. Precious metals are apparently associated with both vein types.

Figure 6 outlines a traverse across the eastern 230 feet (70 m) of the outcrop (Budge and others, 1987; refer to this reference for a longer, 880 foot (267 m) traverse). The Proterozoic rocks consist of lenses of amphibolite interlayered with felsic gneiss and are cut by numerous pegmatites. The layered rocks are migmatitic to varying degrees. One irregular, gently dipping bostonite dike cuts the sequence. Several west-dipping quartz-pyrite veins are visible, and one location displays a vein as wide as 2 cm of galena-sphalerite-chalcopyrite. Tennantite ($(Cu,Fe)_{12}As_4S_{13}$, a blackish, lead-gray, isometric mineral) is reported in many veins and may be present in small amounts. Quartz is the dominant gangue mineral at this locality.

Stop 5. Loveland Pass

Driving instructions: Drive 22.8 miles west on I-70 to the Loveland Pass exit (Exit 216) and proceed 4.2 miles U.S. Route 6 to the top of Loveland Pass.

This view from the Continental Divide provides a spectacular outlook on a large part of the western Front Range. The rocks at the pass are mostly highly fractured migmatitic biotite gneiss, and immediately north is a contact with the 1.4 Ga Silver Plume Granite. The shattering is certainly due in part to frost action, but several NNE-trending brittle faults have also been mapped through here (Bryant and others, 1981b). This widespread fracturing and fault gouge caused major engineering problems during the construction of the nearby Eisenhower Tunnel, which takes Interstate 70 under the Continental Divide.

One speculative theory (Kellogg, 1999) attributes the pervasive fracturing to Laramide movement along the low-angle Williams Range thrust, which forms the western structural boundary of the Front Range a few km to the west. The thrust has a demonstrable minimum overhang of 9 km (evidence for this is at the next stop—the Keystone window in the Williams Range thrust). In fact, if a vertical hole were drilled on Loveland Pass, it is reasonable to suspect that it would encounter Cretaceous shale.

The theory suggests that there is a flexure in the thrust plane, from relatively steep at depth to gentle nearer the surface, located approximately where Proterozoic rocks in the hanging wall overlies the eastern extent of Cretaceous rocks in the footwall. The hanging wall was well above the brittle-ductile transition zone (which is at about 15 km) at the time of thrust movement, so the brittle hanging-wall rocks became pervasively fractured as they ramped over the inflection in the thrust. At many places along the west side of the range, gravitational-spreading features such as "sackungen" (deep, sometimes open trenches, commonly with uphill-facing scarps; singular is "sackung"), extensive landsliding, and smooth, rounded mountain tops characterize many of the ridges and their flanks (Varnes and others, 1989; Kellogg, 1999). In contrast to the central and eastern parts of the Front Range, which contain large rocky outcrops with relatively widely spaced fractures, the western side contains only relatively small, strongly fractured outcrops surrounded by gravelly residuum, suggesting that the underlying rocks are relatively weak.

Stop 6. Keystone Window into the Williams Range thrust

Driving instructions: Continue west on Rte. 6 for 8.6 miles to Gondola Rd. (left-turn entrance to Keystone ski area). Take an immediate left turn (still on Gondola Rd.). Cross Montezuma Rd, jogging slightly right onto North Fork Drive (note the black hornfelsed Pierre Shale in the road cuts); park on east side of circle at the end of drive.

Follow a faint trail through the woods toward the prominent cliffs about 200 m to the east. The access crosses private land, so the owner should be notified first.

The Williams Range thrust forms part of the west-central structural boundary of the Front Range and has been mapped from Middle Park, just north of Kremmling, to South Park, where it is probably continuous with the Elkhorn thrust to the south (Bryant and others, 1981b) (fig. 1). North of Middle Park, the low-angle, en echelon Never Summer thrust steps east from the Williams Range thrust and defines the eastern side of North Park.

The age of thrusting on the west side of the Front Range is not precisely known, although the onset of Laramide deformation in this area has been inferred as the age of the 70 Ma Pando Porphyry near Minturn, about 40 km to the west (Tweto and Lovering, 1977). This is approximately the age (69 Ma) of the marine upper Pierre Shale (69 Ma), indicating that the surface was near sea level and the significant uplift had not yet begun. Synorogenic lower Tertiary beds in South Park (South Park Formation) are as young as upper Paleocene and are overridden by the Elkhorn thrust. If the Elkhorn and Williams Range thrusts are synchronous, movement along the Williams Range thrust probably continued into the early Eocene.

Unlike the higher-angle faults of the eastern margin of the Front Range, the Williams Range thrust is low angle to nearly horizontal in most places. At this locality, the Montezuma stock, a quartz monzonite porphyry with an age of 38-39 Ma (Marvin and others, 1989), domed the Williams Range thrust largely along normal faults (fig. 7), forming a thrust window with hornfelsed Pierre Shale in the footwall and Proterozoic gneiss in the hanging wall (Ulrich, 1963). A basement overhang of at least 9 km is indicated by the distance from the thrust window to the frontal exposures of the thrust to the west, combined with the additional sedimentary section that the thrust must cut through with a maximum ramp angle. The ramp angle is unknown, so the amount of overhang is conceivably much greater than 9 km. Contact metamorphism of both the cataclastic basement rocks in the hanging wall and the Pierre Shale in the footwall, forming a resistant hornfels, allow what may be the best exposure of a Laramide thrust in the entire Front Range!

On the walk up to the thrust, stop on the small ridge near an old log cabin where erratic boulders of quartz monzonite porphyry from the Montezuma lie on deformed hornfels. Continue up to the cliff and examine the hornfelsed Pierre Shale. Interlayered sandstone indicates that these exposures are more than 400 m above the base of the formation (Kellogg, 2000). Traverse right around the base of the cliff to a small fault-controlled gully (offset across the gully is less than a meter). The nearly horizontal Williams Range thrust is exposed on both sides of the gully. A lower, 0.3-1.0-m-thick, strongly sheared and silicified hornfels zone marks the base of the thrust zone and is structurally overlain by a 2-m-thick zone of strongly silicified, and oxidized gneiss-breccia (fig. 8). Migmatitic biotite gneiss with a gentle foliation (strike and dip of about N20E, 25E) overlies the breccia. The overall fault orientation is about N90E, 15N and several slickensided surfaces contain lineations that bear due east, consistent with east-directed thrusting. However, the slickensided surfaces may be related to late-stage stock emplacement or Neogene extension, as they contain fault-polished chlorite-

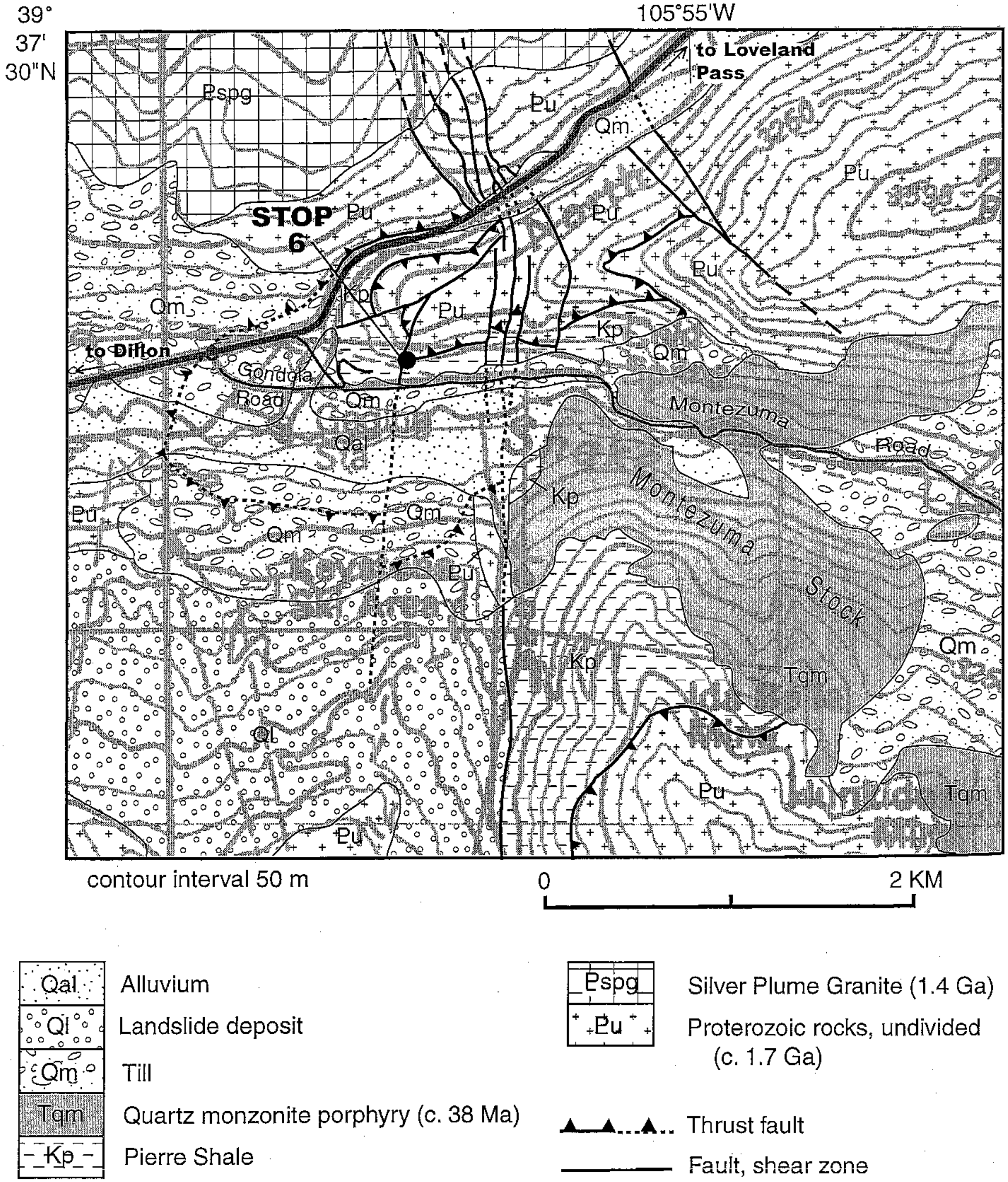

Figure 7. Geologic map of the Keystone window through the Williams Range thrust. Modified from Ulrich (1963).

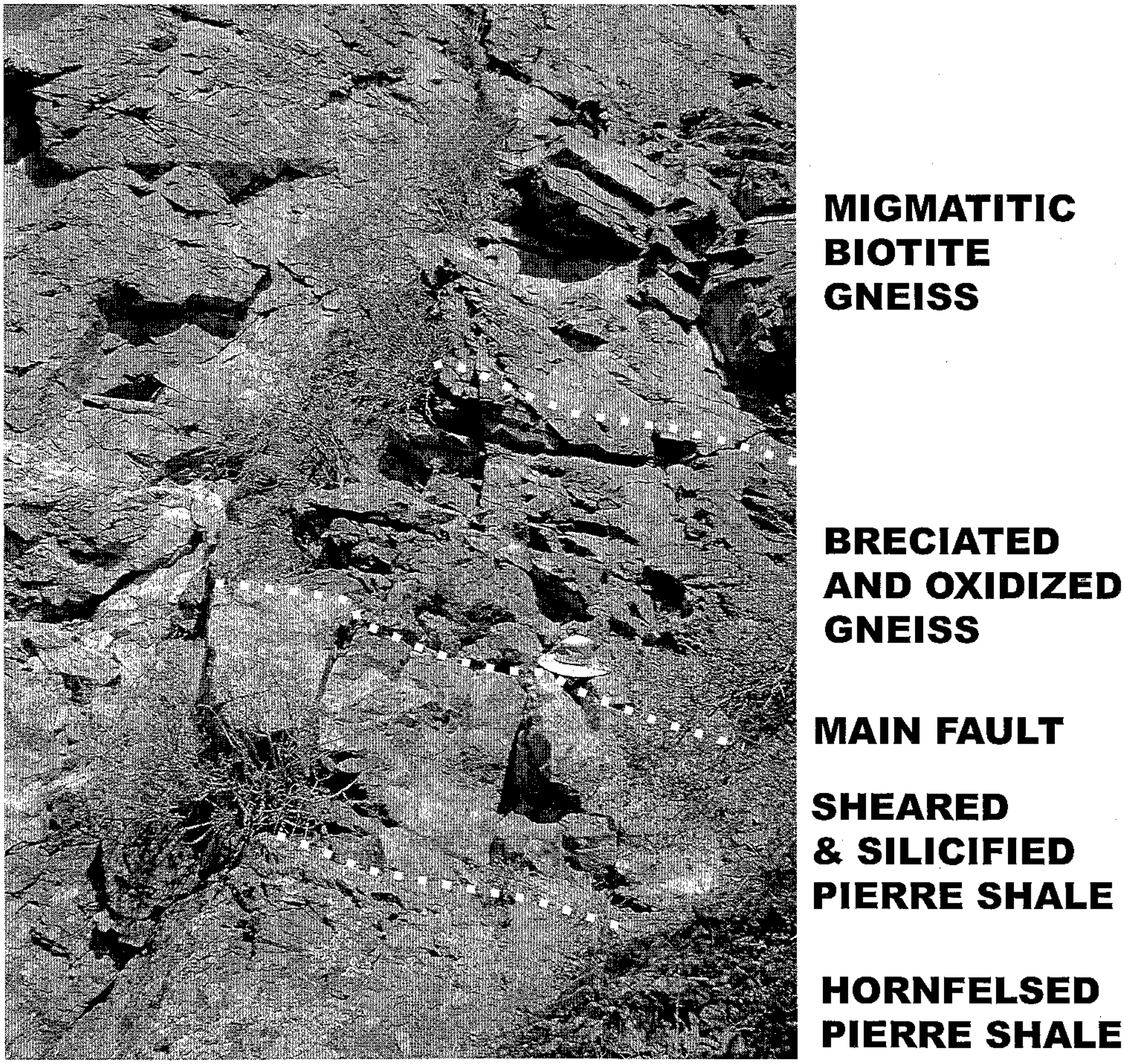

Figure 8. View, facing northwest, showing structural zones of the Williams Range thrust fault near Keystone. Note hat for scale.

epidote alteration. Pervasive chlorite-epidote alteration is probably related to contact metamorphism and hydrothermal alteration by the Montezuma stock.

Stop 7. Western flank of the Front Range and the Blue River half graben

Driving instructions: Rejoin Rte. 6, drive 7.3 miles west to E. Anemone Rd. (just past Dillon Dam Rd.), just west of the town of Dillon. Turn left, proceed 0.4 miles to the cul-de-sac overlooking the Blue River Valley.

The Blue River valley follows part of a belt of mostly Cretaceous sedimentary rocks that extends along the west flank of the Front Range. These rocks lie in a 5-to-9-km-wide half graben, bounded on the west by the Blue River normal fault, a complex Neogene structure that defines the abrupt east margin of the Gore Range (fig. 1). The Blue River half graben is the northernmost major structure of the Rio Grande rift, although a network of Neogene normal faults extends as far north as the Wyoming border (Tweto, 1979; Kellogg, 1999).

Rio Grande rifting began shortly after 29 Ma (Tweto, 1979) and was marked by a change in volcanism from the intermediate compositions that characterized the widespread Oligocene volcanic fields to bimodal basalt-rhyolite (mostly basalt) volcanism that characterizes the Rio Grande rift (Lipman and Mehnert, 1975). Consistent with this style, the main part of the Gore Range contains a few dikes of basaltic to felsic composition. These have not been dated but they may be related to the basaltic rocks of the Yarmony Mountain area (22-24 Ma) west of the Gore Range, part of the bimodal igneous suite related to regional extension. In the Blue River valley a 31 Ma trachytic lacolithic complex (Green Mountain intrusive) intrudes Cretaceous rocks, and a 27 Ma rhyolite welded tuff and 24 Ma trachyandesite flows are preserved in tilted fault blocks within the northern Blue River half graben (Naeser and others, 2002). The tuff is near the base of the sequence, so 27 Ma (late Oligocene) represents the minimum age for initial rifting.

The Blue River normal fault along the west side of the valley has a minimum displacement of 1.2 km, based on the topographic relief between the Phanerozoic sedimentary rocks in the valley and the Proterozoic rocks that form the highest peaks in the Gore Range west of the fault. The latest movement along the fault was probably no later than Pliocene or early Pleistocene (West, 1978), although subtle scarps cutting Bull Lake till (95 ka to >130 ka; Chadwick and others, 1997) were mapped near Frisco, about 5 km south of here (Kellogg, 2000). An extensive apron of glacial deposits emanating from side valleys of the Blue River now covers most of the fault trace and the valley floor west of the Blue River. (The terminus of the major valley glacier that flowed north down the Blue River Valley lies under Dillon Reservoir just to the south of us.)

The half graben is cut by numerous north-striking normal faults that are almost entirely east dipping and bound west-tilted fault blocks (fig. 9A), suggesting that the Blue River graben is a west-tilted structure above a listric fault at depth. Kellogg (1999) suggested that the west-directed Gore fault, a reverse fault along the west side of the Gore Range with significant movement in both Paleozoic and Laramide times, is listric and provided the surface along which Neogene extension beneath the Blue river half graben was accommodated (fig. 9B).

To the northeast, on the west side of the Williams Fork Mountains (part of the western Front Range), the Williams Range thrust defines the structural margin of the Front Range. The thrust is buried beneath an extensive landslide complex except along Interstate 70 about 2.5 km northeast of here (Kellogg, 2001). At that location, the thrust dips about 35° east and is marked by a 3-meter-thick zone of brecciated Precambrian gneiss overlying Pierre Shale. The buried trace of the thrust climbs along the west side of the range to the north and tops the range at Ute Pass, the low point in the ridge visible from here on the east side of the valley.

Landslide deposits mantle most of the slope from the crest of the Williams Fork Mountains to the Blue River and conceal the fault. The landslide deposits may be as thick as several hundred meters and contain blocks of Proterozoic rock tens of meters long. The deposits are deeply eroded and in most places no longer retain hummocky topography, suggesting a late Tertiary or early Pleistocene age (Kellogg, 2001, 2002). A notable contrast in topography is apparent from this location: the crest of the Williams Fork Mountains on the east is rounded and not particularly steep, whereas the Gore Range on the west is rugged and steep. The rocks underlying both ranges are similar—Proterozoic gneiss and granitoid rocks—so the contrast is due to fracture density and gross rock strength underlying each range. At stop 5 (Loveland Pass) a theory for thrust-induced shattering of the hanging-wall rocks was presented and is outlined in more detail in Kellogg (2001).

In late Neogene time, incision of the Blue River undercut the shattered Proterozoic rocks underlying the Williams Fork Mountains, which Kellogg (2001) inferred caused gravitational spreading of the entire mountain ridge. This spreading led to the formation of numerous sackungen ("spreading cracks") along the crest and flanks of the range (Varnes and others, 1989) and caused much of the west side of the Williams Fork Mountains to slide.

Apatite fission-track (AFT) ages from both sides of the Blue River valley reveal important information about the uplift and heating history of the region. The apatite fission-track ages are all younger than Laramide and range from 5 to 37 Ma in the Gore Range and 19 to 48 Ma in the western Front Range (Naeser and others, 2002).

A diagram showing apatite dates at 3000 m altitude from the central Front Range to the White River uplift illustrates the markedly younger dates adjacent to the Blue River valley as compared to areas several tens of km away from the valley, which yield essentially Laramide ages (fig. 10). Similar young AFT ages along major structures related to the Rio Grande rift have been documented by Bryant and Naeser (1980), Lindsey and others (1986), Shannon (1988), and Kelley and others (1992). The youngest ages (5-10 Ma) are at the base of the east flank of the Gore Range west of the Blue River fault. On the tops of the high ridges near the east front of the range, AFT ages are 10-20 Ma. On the west side of the Gore Range ages are 16-25 Ma for the lower altitude samples and 26-37 Ma for the higher altitude samples. Thus, the AFT "thermochrons" (surfaces of equal age) dip west away from the Blue River valley. This indicates higher heat flow along the east side of the Gore Range, in addition to possible westward tilting of the range.

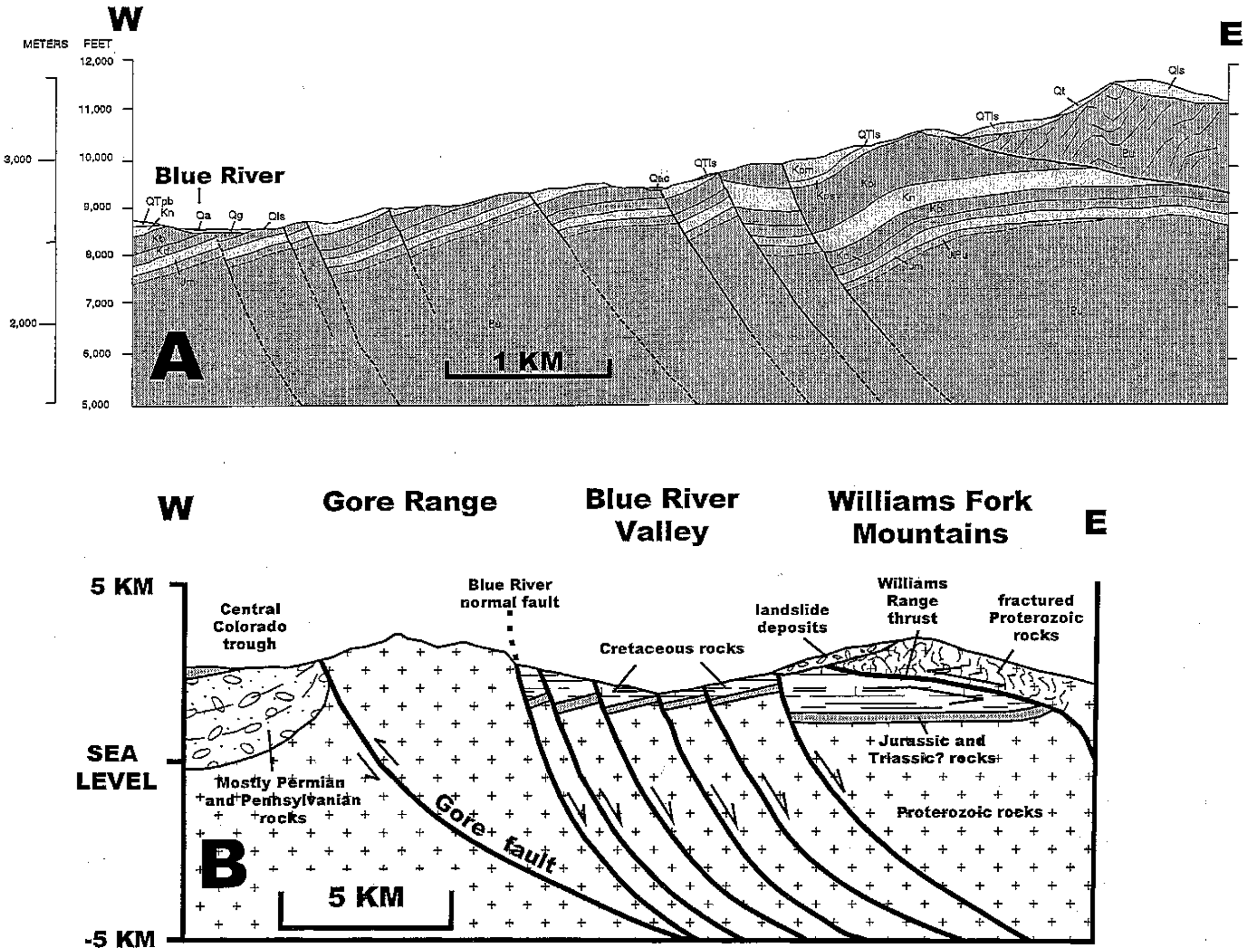

Figure 9. (A) Blue River valley cross section (modified from Kellogg, 1999) showing normal-fault movement, which suggests a model (B) for the western margin of the Front Range in which normal faults merge into a common detachment, exploiting an older fault surface. In this case, the suggested surface is the G ore fault, which has a history at least as old as the late Paleozoic. Symbols in figure 7A are: Qa, alluvium, Qg, terrace gravel; Qls, Holocene landslide deposits; QTpb, pre-Bull Lake diamicton, QTls, old landslide deposits; Kpm, Pierre Shale, sandstone and shale; Kps, Pierre Shale, sandstone; Kpl, Pierre Shale, lower shale; Kn, Niobrara formation; Kb, Benton Shale; Kd, Dakota Group; Ꝑu, Proterozoic gneiss and granite.

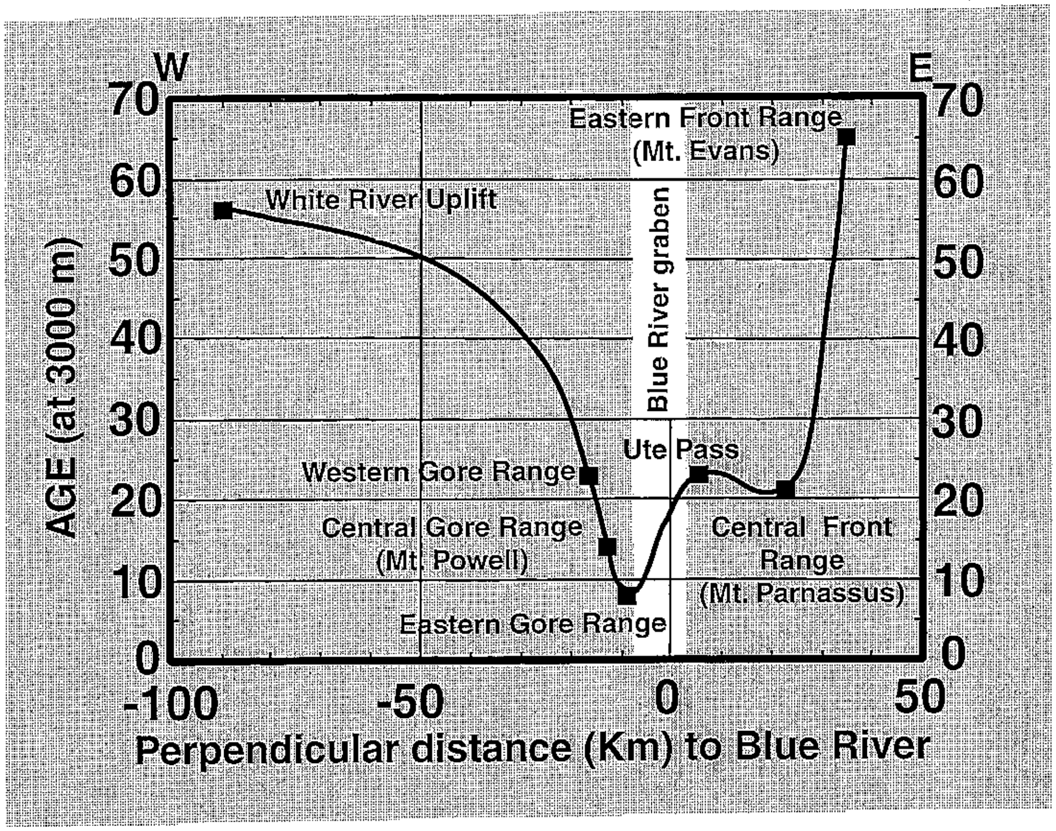

Figure 10. Regional east-west distribution of apatite fission-track ages at 3,000 m altitude from the White Rive uplift in western Colorado to the Front Range, showing the relation of the younger ages near the Blue River half graben (from Naeser and others, 2002).

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