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Evaporite tectonism in the lower Roaring Fork River valley, west-central Colorado

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ABSTRACT

Evaporite tectonism in the lower Roaring Fork River valley in west-central Colorado has caused regional subsidence of a differentially downdropped area in the southern part of the Carbondale collapse center during the late Cenozoic. A prominent topographic depression coincides with this collapse area, and drainage patterns within the collapse area contrast sharply with those outside of it. Miocene volcanic rocks are downdropped as much as 1220 m in the collapse area. Much of the structural lowering occurred along the margins of the collapse area. Major Laramide-age structures bound the east and west sides of the collapse area, but movement on these structures during late Cenozoic collapse was in an opposite direction to their Laramide movement. Within the interior part of the collapse area faults and folds have as much as ~300 m of structural relief. Large blocks of rock may be rafting into the Roaring Fork River valley as underlying evaporite flows toward the valley. Sinkholes are common in the collapse area, as are closed, or nearly closed, structurally controlled topographic depressions that are formed in both surficial deposits and bedrock. Upper Cenozoic deltaic and lacustrine deposits preserved on ridgelines and mesas document the positions of former structural depressions that were initially filled with sediments and later breached by erosion. At least 450 m of syn-collapse sediments accumulated in a collapse depression on the north side of Mount Sopris. Complexly deformed and brecciated deposits in the interior parts of the collapse center are interpreted as collapse debris. Evaporite flow is an important element in the collapse process, and during early stages of collapse it was perhaps the primary means of deformation. Flow by itself, does not remove evaporite from the collapse area. Dissolution and accompanying transport of dissolved constituents by groundwater and surface water are the ultimate means by which evaporite exits the collapse area. Collapse continues today, as evidenced by historic sinkholes and modern high-salinity loads in rivers and thermal springs. Thick evaporite deposits still underlie much of the collapse area, so collapse will likely continue in the future.

Kirkham, R.M., Streufert, R.K., Kunk, M.J., Budahn, J.R., Hudson, M.R., and Perry, W.J., Jr., 2002, Evaporite tectonism in the lower Roaring Fork River valley, west-central Colorado, *in* Kirkham, R.M., Scott, R.B., and Judkins, T.W., eds., Late Cenozoic evaporite tectonism and volcanism in west-central Colorado: Boulder, Colorado, Geological Society of America Special Paper 366, p. 73–99.

INTRODUCTION

Dissolution and flow of evaporitic rocks in the Pennsylvanian Eagle Valley Evaporite during late Cenozoic time have caused as much as 1220 m of differential subsidence in rocks that overlie the evaporite. This structurally downdropped block forms the southern part of the 1200 km^2 Carbondale collapse center in the lower Roaring Fork River valley between the towns of Glenwood Springs and Basalt (Fig. 1). Kirkham and Scott (this volume) describe the conceptual model of evaporite collapse and its relationships with surface water and groundwater. Evaporite flow plays a major role in the collapse, and it is responsible for lateral movement of large masses of evaporite and causes subsidence in upland areas. However, dissolution is the ultimate mechanism that removes evaporite from the collapse area. Modern high salinity loads in rivers and springs indicate that dissolution is active (Chafin and Butler, this volume; Kirkham and Scott, this volume).

Evaporitic rocks were first recognized in the region in the late 1800s (A.R. Marvine in Peale, 1876). Subsequent geologic investigations in the region reported on various aspects of evaporite stratigraphy, depositional setting, and evidence of localized evaporite tectonism and deformed upper Cenozoic rocks and deposits. Kirkham and Scott (this volume) summarize these previous studies.

Evaporite tectonism, which includes the effects of both flow and large-scale dissolution, has affected depositional patterns and drainage systems in this area since shortly after the evaporitic rocks were deposited (Freeman, 1971, 1972; Tweto, 1977). Kirkham and Streufert (1996) first reported regional late Cenozoic collapse due to evaporite tectonism in the lower Roaring Fork River valley downstream of the town of Basalt (Fig. 1). The discovery of this regional collapse resulted from a 1:24 000-scale geologic mapping program initiated by the Colorado Geological Survey (CGS) in 1993 and supported by geochronologic, geochemical, and paleomagnetic studies by the U.S. Geological Survey.

Geologic mapping provided the structural and stratigraphic background needed to recognize and understand the major elements of the collapse. Accurate correlation of upper Cenozoic volcanic rocks in west-central Colorado was essential to characterize many details of the collapse process and to constrain the timing of the deformation. ^{40}Ar/^{39}Ar geochronologic dating of >80 volcanic flows by Kunk and Snee (1998), Kunk et al. (2001), and Kunk et al. (this volume) and geochemical and isotopic studies of >220 samples by Unruh et al. (2001), facilitated the correlation of these rocks by Budahn et al. (this volume). Average La/Yb and Hf/Ta ratios and preferred ages for the compositional geochemical groups discussed in this paper are listed in Table 1.

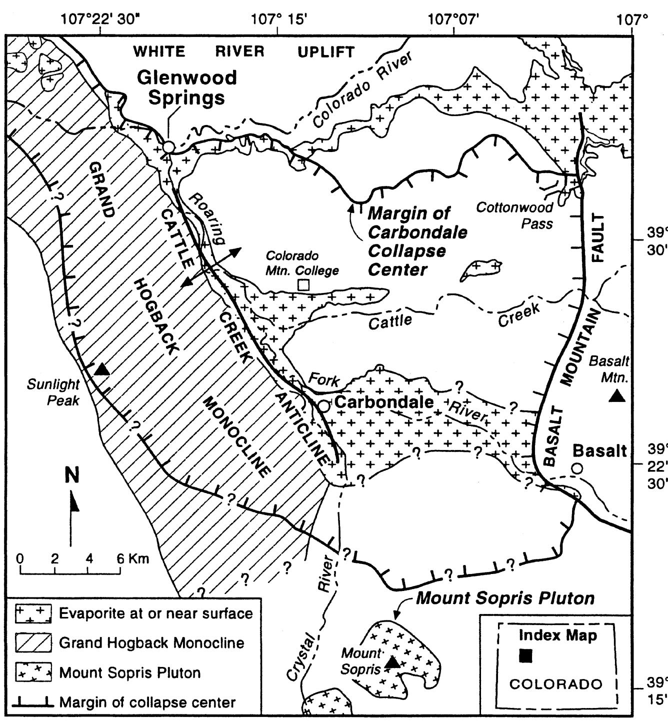

Figure 1. Location map of the southern part of the Carbondale collapse center, lower Roaring Fork River valley (modified from Kirkham et al., 2001b). Queries indicate locations where the collapse area boundary is poorly constrained.

GEOLOGIC SETTING

The southern part of the Carbondale collapse center is defined on at least three sides by major Laramide-age (Late Cretaceous to early Eocene) structures. The northern margin of the collapse area is on the southern flank of the White River uplift, the Basalt Mountain fault marks the eastern edge of the collapse area, and the Grand Hogback monocline forms the western side. A long and narrow extension of the Carbondale collapse center follows the monocline northwestward from near Glenwood Springs (Scott et al., this volume). The southern margin of the collapse center is poorly understood (Kirkham and Scott, this volume), but is placed north of Mount Sopris.

The Middle and Upper Pennsylvanian Eagle Valley Evaporite crops out or underlies surficial deposits in the lower Roaring Fork River valley downstream of Basalt, in the lower Crystal River valley, in lower Cattle Creek, and on the south wall of Glenwood Canyon upstream of Glenwood Springs (Fig. 1). This formation, which consists of halite, gypsum, anhydrite, clastic rocks, and carbonate rocks, was deposited in the Eagle basin part of the Central Colorado trough. Thick sequences of evaporite within the Roaring Fork diapir underlie the lower Roaring Fork River valley. Mallory (1971) and De Voto et al. (1986) reported the evaporite was as much as 2.7 km thick. Perry et al. (this volume) estimate a maximum evaporite thickness of ~1.5 km based on interpretations of seismic reflection data from lines that cross only the ends, but not the center part, of the diapir. Most outcrops of the Eagle Valley Evaporite within the southern Carbondale collapse center include thick

TABLE 1. AVERAGE CHONDRITE-NORMALIZED TRACE-ELEMENT RATIOS AND PREFERRED $^{40}Ar/^{39}Ar$ AND LASER-FUSION AGES OF SELECTED COMPOSITIONAL GEOCHEMICAL GROUPS OF BASALTIC ROCKS IN THE LOWER ROARING FORK RIVER VALLEY

Geochemical group	Average chondrite-normalized La/Yb	Average chondrite-normalized Hf/Ta	Preferred age or age range (Ma)
1b	5.490	0.759	9.75 ± 0.06 to 10.84 ± 0.06
1c	8.005	0.754	10.60 ± 0.07
2b	8.270	0.615	9.68 ± 0.03 to 10.70 ± 0.15
3a	7.926	0.474	No dates
4a	11.000	0.530	No dates
4b	11.659	0.541	9.68 ± 0.03 to 10.49 ± 0.07
5a	10.474	0.388	7.75 ± 0.03
5b	10.312	0.440	7.75 ± 0.04
6b	12.829	0.329	3.97 ± 0.08
6b′	12.118	0.285	2.90 ± 0.01
6b″	11.681	0.364	3.17 ± 0.02
6c	12.389	0.377	3.05 ± 0.04
10a	21.483	0.444	22.56 ± 0.13
12a	11.959	0.765	13.29 ± 0.28
12b	12.349	0.787	13.38 ± 0.06
13a	13.593	0.928	13.57 ± 0.05

Note: From Budahn et al. (this volume); Kunk and Snee (1998); Kunk et al. (2001, this volume).

beds of gypsum that are often highly deformed, but halite and anhydrite are present only in the subsurface. A lithologic log of the 933-m-deep Shannon Oil Rose #1 well, provides the best subsurface information in the southern part of the Carbondale collapse center (unpublished lithologic log by American Stratigraphic Company, now held by the Denver Earth Resources Library). This well, drilled near the mouth of Cattle Creek (Fig. 2), penetrated 18 m of alluvial gravel, then 915 m of gypsum, anhydrite, halite, and clastic and carbonate rocks. The lower 285 m of the well was predominantly halite, and the well did not reach the base of the evaporite sequence.

The Champlin Oil and Refining Blue #1 well was drilled in the center of the collapse area on the north side of the Roaring Fork River between Carbondale and El Jebel (Fig. 2). The well encountered evaporitic rocks to a depth of 707 m, but had to be abandoned. The Mobil Oil Elk Camp Federal #F23X-22P, located in the Grand Hogback monocline on the southwest flank of the White River uplift ~22 km west of Glenwood Springs, penetrated 4810 m of steeply dipping post-evaporite sedimentary rocks before reaching the Eagle Valley Evaporite (unpublished lithologic log by G.E.O., Inc.). Halite was encountered at a depth of 5311 m and immediately began flowing into the borehole (J.L. White, 1998, personal commun.). None of these wells penetrated the entire evaporite sequence, because halite caused drilling problems.

The unusually thick section of evaporitic rocks underlying the lower Roaring Fork River valley may be related to tectonic thickening or flow of evaporite beneath and adjacent to the Grand Hogback monocline. When the monocline formed during the Laramide orogeny, evaporite may have flowed toward the point of greatest curvature of the monocline (Perry et al., this volume). Late Cenozoic flow from beneath adjacent uplands to the valley also thickened the evaporite beneath the valley (Kirkham and Scott, this volume).

CHARACTERISTICS OF THE COLLAPSE AREA

Topography

The southern part of the Carbondale collapse center is a roughly square-shaped topographic depression within an otherwise mountainous region. The topographic depression is readily apparent in an oblique shaded relief image of the region (Fig. 3). The land surface within the collapse area is as much as 1220 m lower than surrounding upland areas.

Conspicuous escarpments that descend rapidly into the collapse area mark the northern and eastern margins of the collapse center (Fig. 3). Prominent, parallel, northwest-trending valleys and ridges west (left) of the confluence of the Roaring Fork and Crystal Rivers are related to differential erosion of Mesozoic and Paleozoic sedimentary rocks within the Grand Hogback monocline. Gently east-sloping, basalt-capped Sunlight Mesa is north-northwest of these prominent valleys and ridges. A broad, bowl-like depression superposed on the mesa is probably a result of differential collapse of the monocline. Northwest-trending, parallel, relatively shallow valleys and low ridges cross Sunlight Mesa. These landforms are associated with bedding-plane faults caused by flexural slip of the collapsing Grand Hogback monocline. Broad flat valleys, such as Spring Valley and Cottonwood bowl, gently sloping basalt-capped mesas like Los Amigos Mesa, which is the landform between Spring Valley and the Roaring Fork River, and the rolling hills of Missouri Heights typify the interior parts of the collapse area (Fig. 3). Los Amigos Mesa and the block of rock northeast of the confluence of the Crystal and Roaring Fork Rivers jut far out into the valley floor. These blocks may be rafting into the valley as the underlying evaporite flows toward the river.

The land surface, particularly in the northeastern and central parts of the collapse area, is hummocky. Drainages of many

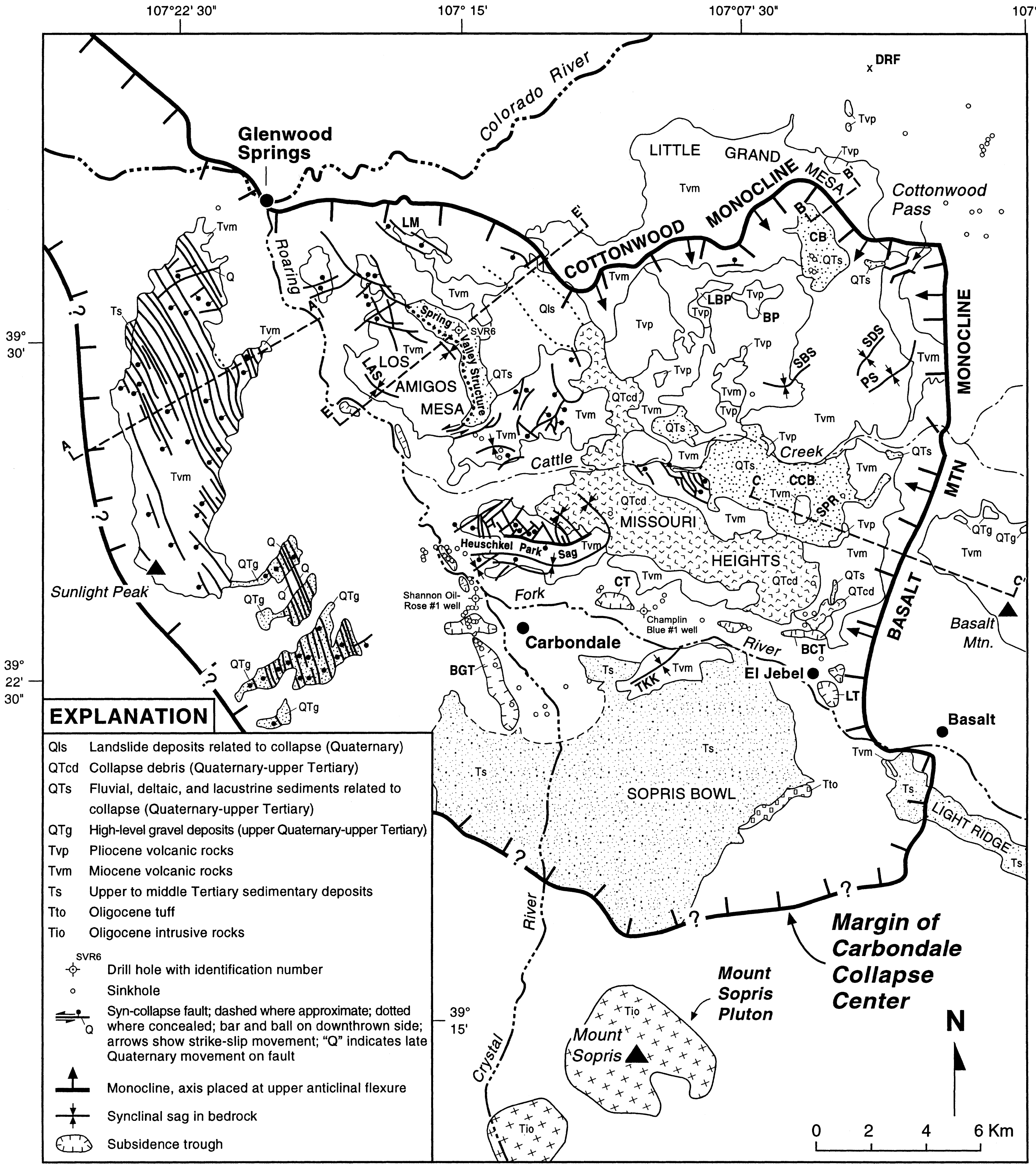

Figure 2. Geologic features of the southern part of the Carbondale collapse center (modified from Kirkham et al., 2001b). BCT—Blue Creek trough; BGT—Barbers Gulch trough; BP—Buck Point; CCB—Cattle Creek bowl; CB—Cottonwood bowl; CT—Crystal terrace trough; DRF—doubly recumbent fold; LAS—Los Amigos sag; LBP—Little Buck Point; LM—Lookout Mountain; LT—Leon trough; PS—Polaris sag; SBS—Shippes Bowl sag; SDS—Shippes Draw sag; SPR—Spring Park Reservoir; TKK—Ti-Ke-Ki sag.

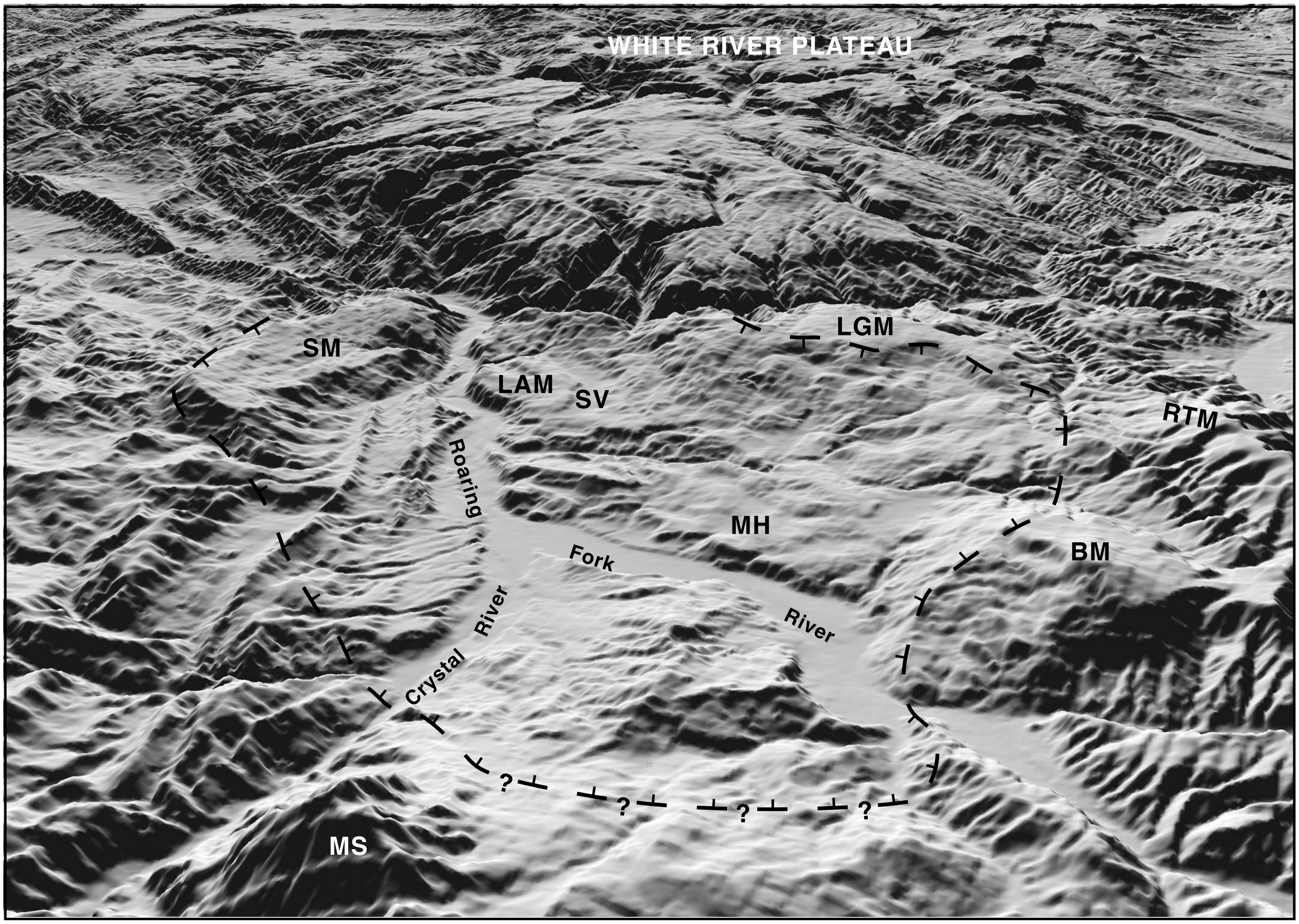

Figure 3. Oblique shaded-relief digital elevation model of the lower Roaring Fork River valley, showing major geographic features (100 m DEM provided by David Catts, 2001, U.S. Geological Survey). View is to north. Dashed line with hachures marks boundary of collapse area. BM—Basalt Mountain; LAM—Los Amigos Mesa; LGM—Little Grand Mesa; MH—Missouri Heights; MS—Mount Sopris; RTM—Red Table Mountain; SM—Sunlight Mesa; SV—Spring Valley. At the west edge of Little Grand Mesa, the collapse margin drops down into Glenwood Canyon and is not visible in this perspective.

of the tributary streams within the collapse area are irregular, poorly integrated, and sometimes interrupted by swallow holes. The tributary streams are incised little and have relatively gentle valley walls compared to the well-integrated and deeply and sharply incised streams outside of the collapse area (Fig. 3). Sinkholes are common in the area (Fig. 2) (Mock, this volume). Several closed or nearly closed depressions lie within the collapse area; they include broad, downwarped river terraces (subsidence troughs) and shallow synclines formed in volcanic rocks (synclinal sags) (Fig. 2).

Margins of the collapse area

We depict our preferred positions of the collapse area margins in Figures 1, 2, and 3. These boundaries are best defined where upper Tertiary volcanic rocks are preserved. Upper Tertiary basaltic flows immediately outside of the collapse center are generally flat lying, whereas in the structural zones along the margins of the collapse areas, these same flows dip as much as 52° into the collapse center.

Western margin. The Grand Hogback monocline, a major down-to-the-west fold created near the end of the Laramide orogeny (Tweto, 1977), forms the western margin of the collapse center in the lower Roaring Fork River valley (Fig. 2). Lower Tertiary and older rocks within the monocline generally dip 40°–60° westward (Murray, 1966; Kirkham et al., 1996, 1997). A subhorizontal erosion surface was cut across the sharply folded lower Tertiary and older rocks during or prior to the middle Miocene. A sequence of basaltic lava flows erupted onto this erosion surface ca. 10.6–9.9 Ma (Kirkham et al., 2001b; Kunk et al., this volume). Individual flows are as much as 10 to 15 m thick. Many flows maintain fairly constant thicknesses across large areas and show no evidence of thickening in the collapse area; some have pahoehoe flow structure. Such characteristics indicate that the lava flows were originally erupted onto the subhorizontal erosion surface (Ken Hon, 1995, written commun.).

Geologic relationships along the western margin of the collapse center demonstrate that the Grand Hogback monocline underwent down-to-the-east relaxation or unfolding during late Cenozoic time. This movement of the monocline is opposite in direction to its Laramide movement and is interpreted to result from removal of evaporite from beneath the monocline, either by flow and/or dissolution (Unruh et al., 1993; Kirkham et al., 2001b). This relaxation or unfolding caused the overlying, subhorizontal basaltic caprock to tilt eastward into the Carbondale collapse center, while simultaneously reducing the dips of the lower Tertiary and older rocks within the monocline. Prominent bends are apparent in the hogback ridge along the monocline west of the confluence of the Crystal and Roaring Fork Rivers where the ridge crosses the margin of the collapse area (Fig. 3). The flattening of dip that was caused by late Cenozoic relaxation or unfolding of the monocline is responsible for the prominent bends in the hogback ridge.

As the monocline relaxed, flexural slip occurred along bedding planes in the sedimentary rocks within the monocline, much like the movement of a deck of cards as it topples over. This slip formed a series of subparallel, down-to-the-west flexural-slip faults (Murray, 1966, 1969; Stover, 1986) that ruptured the basalt flows and some overlying surficial deposits (Figs. 2 and 4). The flexural-slip faults have normal displacement and, because they follow bedding planes within the monocline, probably become listric in the lower or synclinal limb of the monocline.

A large remnant of ca. 10 Ma volcanic flows is preserved on Sunlight Mesa southwest of Glenwood Springs. The flows are broken by numerous northwest-striking, parallel, flexural-slip faults and are tilted eastward into the collapse area (Figs. 2 and 4). Within the blocks bounded by the faults, the flows typically dip 10°–20° east and locally have dips >40°. The flexural-slip faults are recognizable only where lava flows and certain surficial deposits unconformably overlie the folded older sedimentary rocks. A prominent series of northwest-striking, fault-controlled valleys and ridges that trend parallel to the monocline are developed in the basalt cap (Figs. 5 and 6). These parallel ridges and valleys form a fault-controlled trellis drainage pattern.

Faults that cut the volcanic cap on Sunlight Mesa are downthrown to the west, opposite in direction to the overall east tilt of the basalt cap and to the east dip of flows within the fault blocks (Kirkham et al., 1996, 1997). Flexural-slip faults are widely spaced in areas where the basalt cap is underlain by the Upper Cretaceous Mancos Shale and have greater displacements than faults in areas underlain by the Upper Cretaceous Mesaverde Group or lower Tertiary Wasatch Formation. A few northeast-trending cross faults cut the basalt cap near the north end of Sunlight Mesa (Figs. 2 and 5). A fault scarp in upper Quaternary deposits along one of these faults (marked by "A" in Figure 5) suggests latest Pleistocene or Holocene movement (Kirkham et al., 1997).

A structure contour map on the top of the ca. 10 Ma basaltic rocks on Sunlight Mesa (Fig. 7) indicates this formerly flat-lying basaltic cap now has an overall eastward slope of 6°–10° into the collapse center. This map confirms that the bowl-like

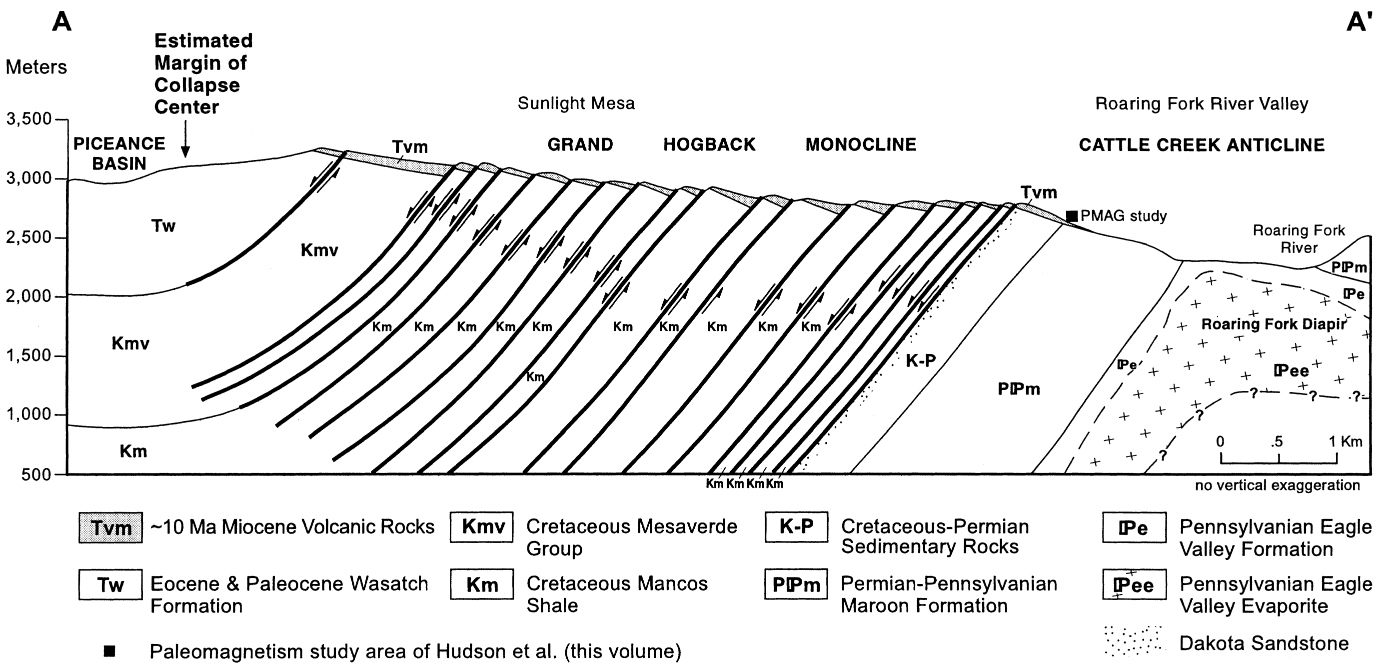

Figure 4. Cross section A–A′ through Sunlight Mesa. Location of cross section shown on Figure 2.

Figure 5. Vertical aerial photograph of the Sunlight Mesa area (National Aerial Photography Program [NAPP] photo 6716–41), where flexural-slip faults cut Miocene volcanic rocks that overlie an angular unconformity eroded across west-dipping sedimentary rocks in the Grand Hogback monocline. Note the prominent, parallel, northwest-striking, light-colored lineaments that are linear grassy valleys (some are marked by arrows). Each valley is underlain by a west-dipping normal fault related to relaxation of the monocline. The linear dark areas between the valleys are ridges formed by fault blocks capped by east-dipping volcanic rocks. Arrows labeled "A" indicate ends of a cross fault that cuts upper Quaternary deposits. Bershenyi terrace has been folded by Pleistocene diapirism. Town of Glenwood Springs in upper right. Scale and north arrow are approximate.

Figure 6. Oblique aerial photograph of the northern end of Sunlight Mesa looking northwest. Prominent fault-controlled ridges and valleys are formed in the ca. 10 Ma basalt flows that cap the mesa. Arrows denote some of the valleys. Note steep westward dip of Dakota Sandstone (Kd) in Grand Hogback monocline.

depression on the mesa has a structural origin and is likely related to differential collapse of the monocline. Late Miocene volcanic flows drop ~1040 m in elevation from the top of Sunlight Peak to the lowest exposure of correlative flows on the west side of the Roaring Fork River. This elevation difference is the minimum amount of post–10 Ma collapse on the western margin of the collapse center. Equivalent-age volcanic rocks are ~180 m lower in elevation on the east side of the river, which suggests ~1220 m of collapse since eruption of the ca. 10 Ma flows.

The average slope of the ground surface varies across Sunlight Mesa (Fig. 3). On the western side of the mesa, near and east of Sunlight Peak, the surface slopes as much as 40% to the east. The surface of the mesa flattens east of the zone of steep slope and remains fairly flat until reaching the subcrop of the Dakota Sandstone beneath the volcanic flows (Fig. 4). Remnants of the basalt cap extend eastward beyond the Dakota subcrop in three areas (Kirkham et al., 1996, 1997), and in all three locations the dip of the flows sharply increases east of the Dakota subcrop. Paleomagnetic measurements at three sites in the basaltic flows in the wedge-shaped remnant east of the Dakota subcrop along cross section A–A′ (Fig. 4) are consistent with an interpretation of eastward tectonic tilting (Hudson et al., this volume).

Flexural-slip faults associated with the Grand Hogback monocline also offset upper Tertiary to lower Quaternary basalt-rich gravel deposits west of Carbondale (Figs. 2 and 8). Light-colored linear features in Figure 8 mark grassy swales underlain by the faults, and the darker areas are the upthrown blocks that are vegetated with sagebrush and/or gambel oak. Prominent uphill-facing scarps mark the down-to-the-west flexural-slip faults in the basalt-rich gravel deposits (Fig. 9). Three of these faults offset upper Quaternary deposits that fill one of the valleys between the mesas (Stover, 1986; Kirkham et al., 1996).

Individual flexural-slip faults offset the Miocene basalt

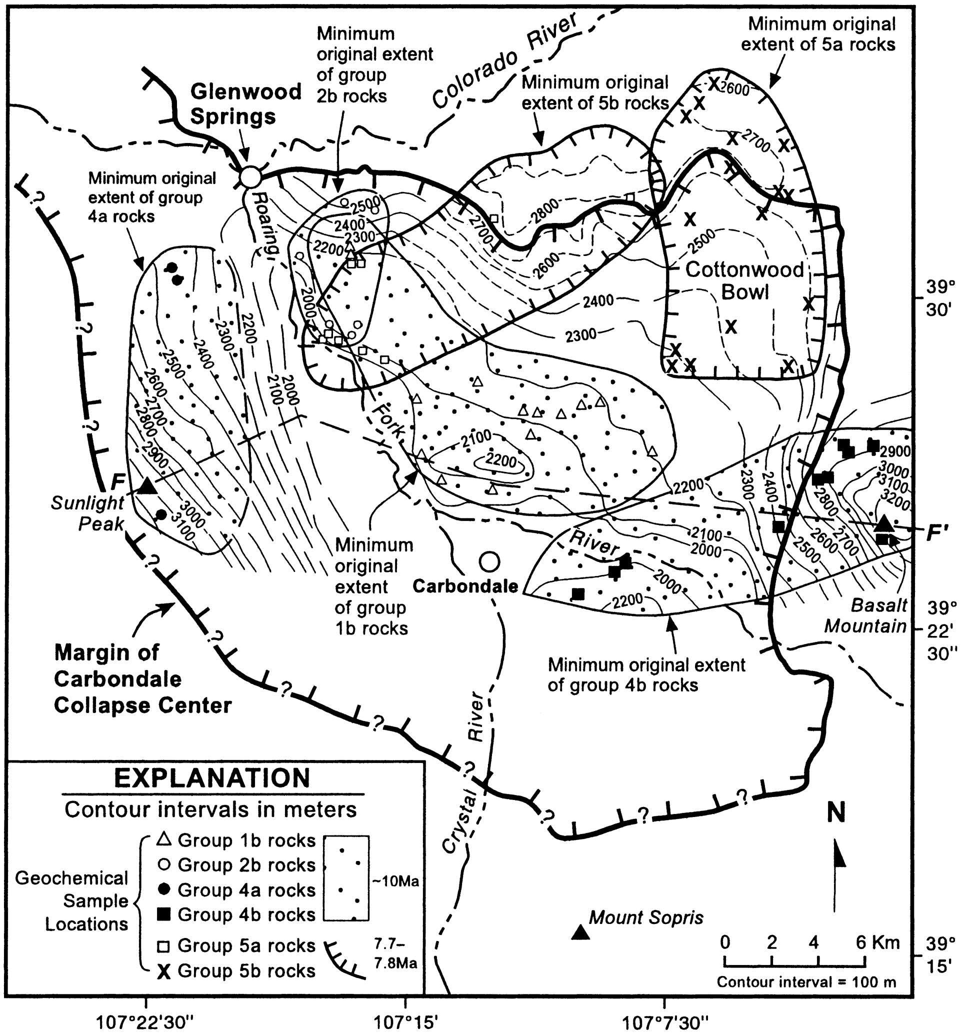

Figure 7. Structure contour maps on the top of the 7.75 Ma volcanic flows in geochemical groups 5a and 5b and on the top of the ca. 10 Ma rocks of groups 1b, 2b, 4a, and 4b (Table 1). Adapted from Kirkham et al. (2001b). Geochemical correlations are by Budahn et al. (this volume). Dashed line indicates approximate location of generalized cross section F–F′ (Fig. 22).

flows as much as 92 m. The upper Tertiary to lower Quaternary gravel deposits are displaced a maximum of ~30 m, whereas the fault throw in the upper Quaternary deposits is only an estimated 3 m. The increasing displacement in successively older deposits indicates the flexural-slip faults associated with the Grand Hogback monocline have experienced recurrent or continuous movement during the late Cenozoic.

Northern margin. East of Glenwood Springs, near Lookout Mountain (Fig. 2), the northern margin of the collapse center is poorly constrained. The margin is arbitrarily placed at the northern limit of evaporite on the south wall of Glenwood Canyon. Evaporitic rocks were probably once present north of their modern outcrop, and late Cenozoic or older evaporite deformation may have occurred there prior to removal of the evaporite by erosion.

Faulted remnants of ca. 10 Ma group 2b volcanic rocks (Table 1) crop out on the south side of Lookout Mountain at substantially lower elevations than other age-equivalent flows that are widely distributed across the White River Plateau (Larson et al., 1975; Kunk et al., this volume). The upper Miocene flows on Lookout Mountain and those to the south and east (Fig. 2) are within the collapse area. Deformation along the collapse margin in the Lookout Mountain area appears to be spread across a broad belt, with the collapse being accommodated by several small faults and by tilting. Elevation differences between the group 2b rocks preserved on the White River Plateau and those on the northern margin of the collapse area indicate a minimum vertical displacement of 680 m (Fig. 7). The maximum amount of post–10 Ma collapse along the northern margin is 1060 m, based on the elevation of ca. 10 Ma flows on the east valley wall of the Roaring Fork River.

East of Lookout Mountain is a 13-km-long basalt-capped mesa called Little Grand Mesa or Dock Flats. A stacked sequence of thick flows caps the mesa, which is up to 5 km wide. Group 5b flows (Table 1) crop out in the western part of Little Grand Mesa, whereas group 5a rocks are found in the eastern end. Both groups are ca. 7.75 Ma (Table 1). The volcanic flows that cap Little Grand Mesa are not noticeably folded or faulted

Figure 8. Vertical aerial photograph of an area west of Carbondale (NAPP photograph 6717-8). Flexural-slip faults related to relaxation of the Grand Hogback monocline underlie the light-colored, grassy meadows in upper Tertiary to lower Quaternary basalt-rich gravel deposits that cap the mesas (some of the faults are marked with arrows). Three faintly visible faults (in box labeled “A”) cut upper Quaternary valley-fill deposits. Faults shown in Figure 9 are labeled “B.” Scale and north arrow are approximate.

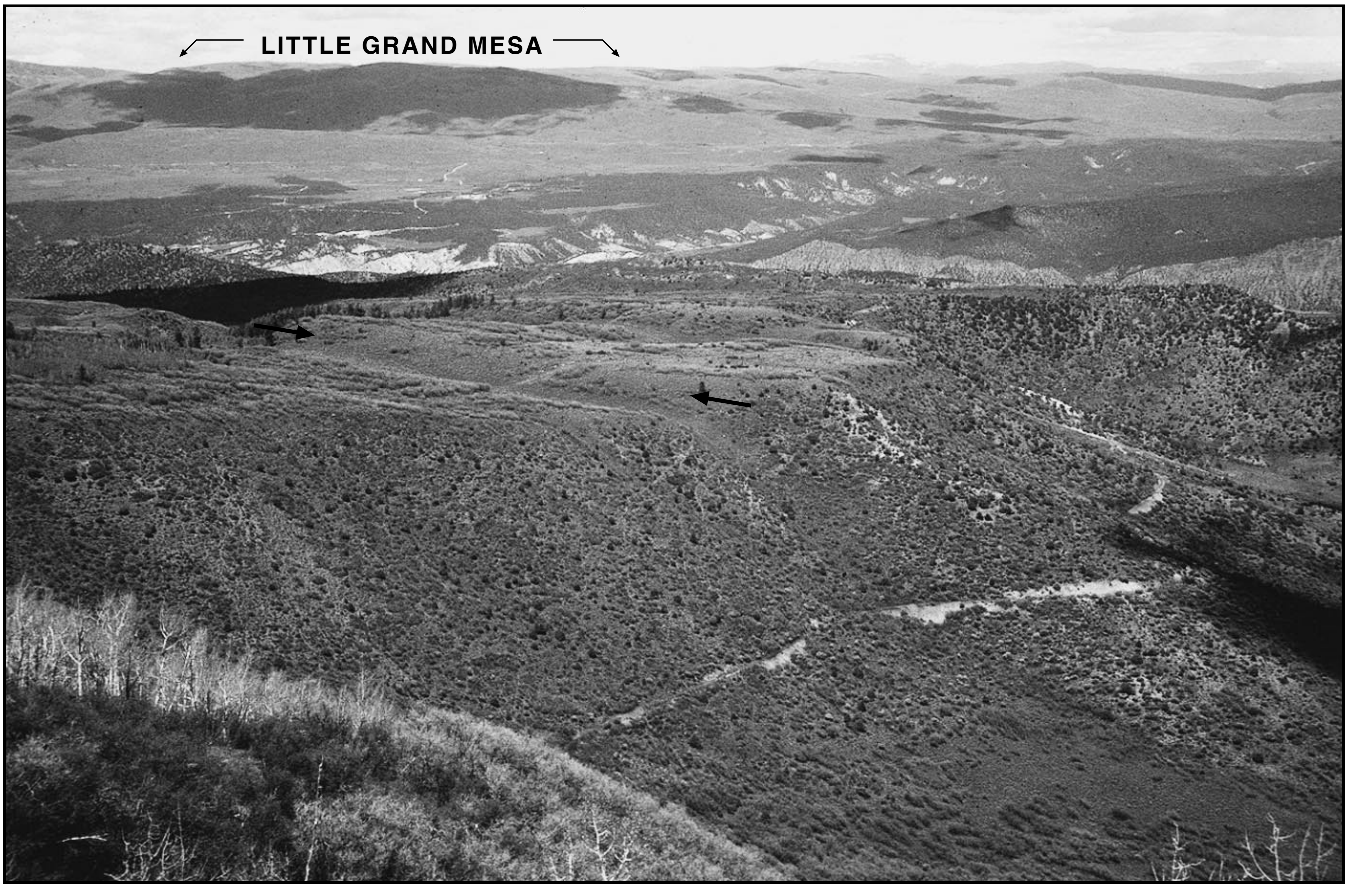

Figure 9. Photograph of flexural-slip fault with prominent uphill-facing scarp in upper Tertiary-lower Quaternary gravel deposits that cap a mesa west of Carbondale. Arrows denote the scarp, which faces to the left. Other more subtle scarps are visible in front of and behind the prominent scarp. View is to northeast, with Little Grand Mesa forming part of the skyline. See Figure 8 for location of photograph.

even though evaporite may occur in the subsurface beneath much of the mesa. In several good exposures, stacked sequences of flows can be traced for hundreds of meters with no apparent deformation. The flows on Little Grand Mesa occur at the highest elevation of any 7.75 Ma flows in the region.

We conclude that differential collapse has not occurred beneath Little Grand Mesa during the past 7.75 m.y., but recognize that the entire basalt-capped mesa may be an intact block that uniformly lowered as underlying evaporite was removed from beneath it. The northeastern margin of the collapse center is placed at the top of a sharp monoclinal drape fold called the Cottonwood monocline, which forms the southern edge of Little Grand Mesa. If the intact block of Little Grand Mesa was lowered by evaporite collapse, then the margin of the collapse center probably follows the evaporite outcrop on the southern wall of Glenwood Canyon, north of Little Grand Mesa (Fig. 1). Evidence of evaporite tectonism, such as complex recumbent folds (Fig. 10), unusual fault patterns in rocks overlying the evaporite, and sinkholes, is present northeast of Little Grand Mesa (Streufert et al., 1997). These features suggest evaporite tectonism has affected this area, but this deformation may predate the late Cenozoic.

Cottonwood monocline (Fig. 2) is a narrow and sinuous fold with several nearly right-angle bends (Streufert et al., 1997; Kirkham et al., 1995). The flat-lying, 7.75 Ma group 5a volcanic rocks (Table 1) capping Little Grand Mesa are abruptly deformed by the monocline (Fig. 11). Within the monocline, these volcanic rocks are moderately well exposed in a small valley cut across the monocline (valley is left of symbol “CM” in Figure 11). Cross section B–B′ (Fig. 12) is drawn perpendicular to the monocline at the location of symbol “CM” on Figure 11. Both the top and bottom limbs of the monocline are sharp. Group 5a basalts dip as much as 44° into the collapse center. They are lowered ~180 m by Cottonwood monocline (Figs. 6 and 12). A few very small displacement faults cut the flows within the monocline, but folding accommodates most of the monocline’s structural relief.

Cottonwood bowl lies at the bottom of the monocline. A veneer of upper Tertiary to lower Quaternary sediments as much as 35 m thick (unit QTs in Fig. 2; unit QTc in Fig. 12)

Figure 10. Photograph of a complex recumbent fold formed in the Eagle Valley Evaporite on the northeast side of Cottonwood Creek. View is to northeast. Fold located northeast of Cottonwood monocline ("DRF" in Figure 2). Photograph by Beth L. Widmann.

blankets much of the floor of Cottonwood bowl. These locally derived alluvial and colluvial sediments were mostly eroded from the face of the monocline as it formed, accumulated in the topographic depression of Cottonwood bowl at the toe of the monocline, and have since been slightly dissected by erosion. The low hills visible on the floor of Cottonwood bowl (Fig. 11) are capped by these sediments, which overlie the downdropped 7.75 Ma group 5a volcanic rocks. Although poorly exposed, the volcanic rocks beneath the floor of Cottonwood bowl appear to be subhorizontal with an overall gentle southward dip. This structural bench continues southward until reaching a west-northwest-striking, broad monoclinal structure on the north side of Cattle Creek.

Eastern Margin. The Basalt Mountain monocline forms the eastern margin of the Carbondale collapse center (Fig. 2). It merges with the Cottonwood monocline at the northeast corner of the collapse center near Cottonwood Pass. The Basalt Mountain monocline coincides with the Basalt Mountain fault, a high-angle, Laramide-age fault with >3000 m of down-to-the-east throw (Streufert et al., 1997; Kirkham et al., 1998). Where the Eagle Valley Evaporite is at or near the ground surface on the west side of the fault there is widespread evidence of major collapse. On the east side of the fault, where the evaporite is overlain by as much as 2500 m of sedimentary and volcanic rocks, there is no known evidence of collapse (Fig. 13).

During the late Miocene, multiple group 4b flows (Table 1) erupted from the Basalt Mountain shield volcano on the east side of the Basalt Mountain fault (Fig. 2) ca. 10.5–9.7 Ma. Flows from the Basalt Mountain volcano extend nearly to Carbondale, ~14 km west of the crater (Fig. 7). As on the western side of the collapse area, these flows were erupted onto a landscape with low topographic relief prior to any significant late Cenozoic collapse (Kirkham et al., 1998). These late Miocene, formerly flat-lying flows are now sharply tilted westward within the Basalt Mountain monocline, which is a broad hinge zone with average width of ~1.6 km. Although the group 4b flows within the monocline are very broken and jumbled, they dip on average ~14° to the west. In the Missouri Heights area west of the monocline, group 4b volcanic flows are broken by faults and fractures, but generally are subhorizontal with only gentle west and southwest dip. The minimum amount of post–10 Ma

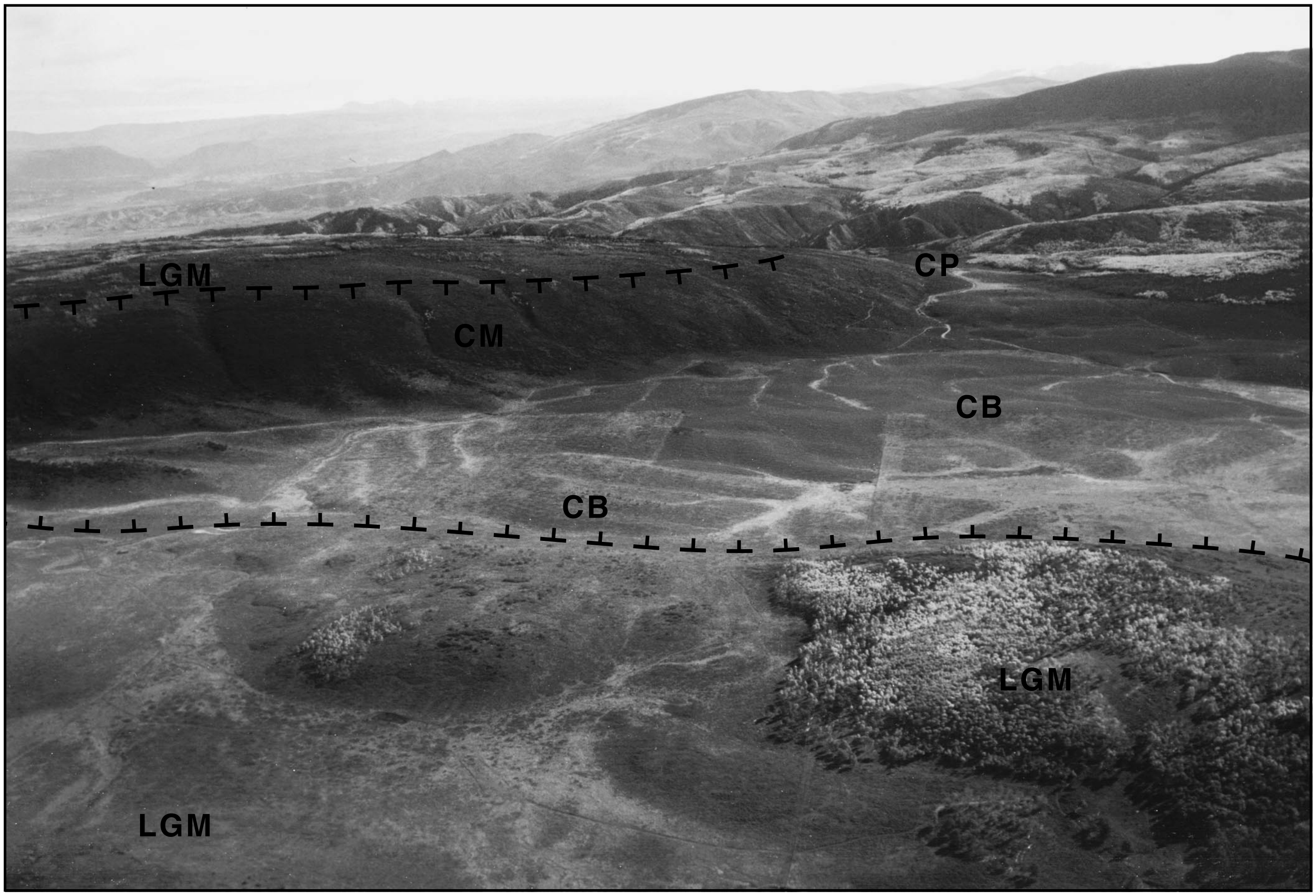

Figure 11. Oblique aerial photograph looking east across Cottonwood bowl (CB) and Cottonwood monocline (CM). Syn-collapse sediments underlie the rounded low hills in Cottonwood bowl. The dashed line with hachures marks top of Cottonwood monocline, which here forms the margin of the Carbondale collapse center. The steep hillside that coincides with the monocline is a dip slope formed by tilted 7.75 Ma basaltic flows. Little Grand Mesa (LGM), the flat surface in the upper left and in the foreground, is capped by nearly horizontal 7.75 Ma basaltic flows. Cottonwood Pass (CP) is in the upper right. Cross section B–B′ (Figs. 2 and 12) crosses the monocline approximately at the location indicated by CM.

collapse on the eastern margin of the collapse center is ~580 m, based on elevation differences between group 4b rocks at the base of Basalt Mountain shield volcano and those exposed in Missouri Heights. If Basalt Mountain was lowered by collapse as a large intact block, a hypothesis discussed by Kirkham and Scott (this volume), then the margin of the collapse center is east of Basalt Mountain and the amount of structural relief reported above represents only a portion of the total vertical subsidence along the eastern margin.

A small cinder deposit and related flow near the southeast end of Spring Park Reservoir (geochemical group 6b′) are preserved within the Basalt Mountain monocline and immediately west of it (unit Tvp on Fig. 2). The cinders and flow have a preferred age of 2.90 ± 0.01 (Table 1). The lava from this volcano originally flowed downhill away from the eruptive center. However, the northeastern and most distal end of the preserved flow is now as much as 73 m higher in elevation than the cinder cone from which it was probably erupted. The flow is tilted westward into the collapse center at an estimated 13°, which is similar to the tilt in the ca. 10 Ma Basalt Mountain flows folded by the Basalt Mountain monocline (Kirkham et al., 1998, 2001b). This suggests that much of the collapse accommodated by the Basalt Mountain monocline occurred during the past 3 m.y.

On the east side of the Basalt Mountain fault, where evaporitic rocks are deep in the subsurface, no evidence of evaporite tectonism is recognized. Basalt Mountain shield volcano and the flows from it appear to be intact and are not noticeably deformed on the east side of the fault. However, with the greater depth to the evaporite, overlying strata may have absorbed strain, so that collapse could have occurred on the east side of the fault without forming obvious structural features visible at the ground surface. In such a scenario, the shield volcano would collapse as an intact block of rock. The ca. 10 Ma flows on Basalt Mountain are at somewhat lower positions in the landscape than are other age-equivalent rocks at other locations in

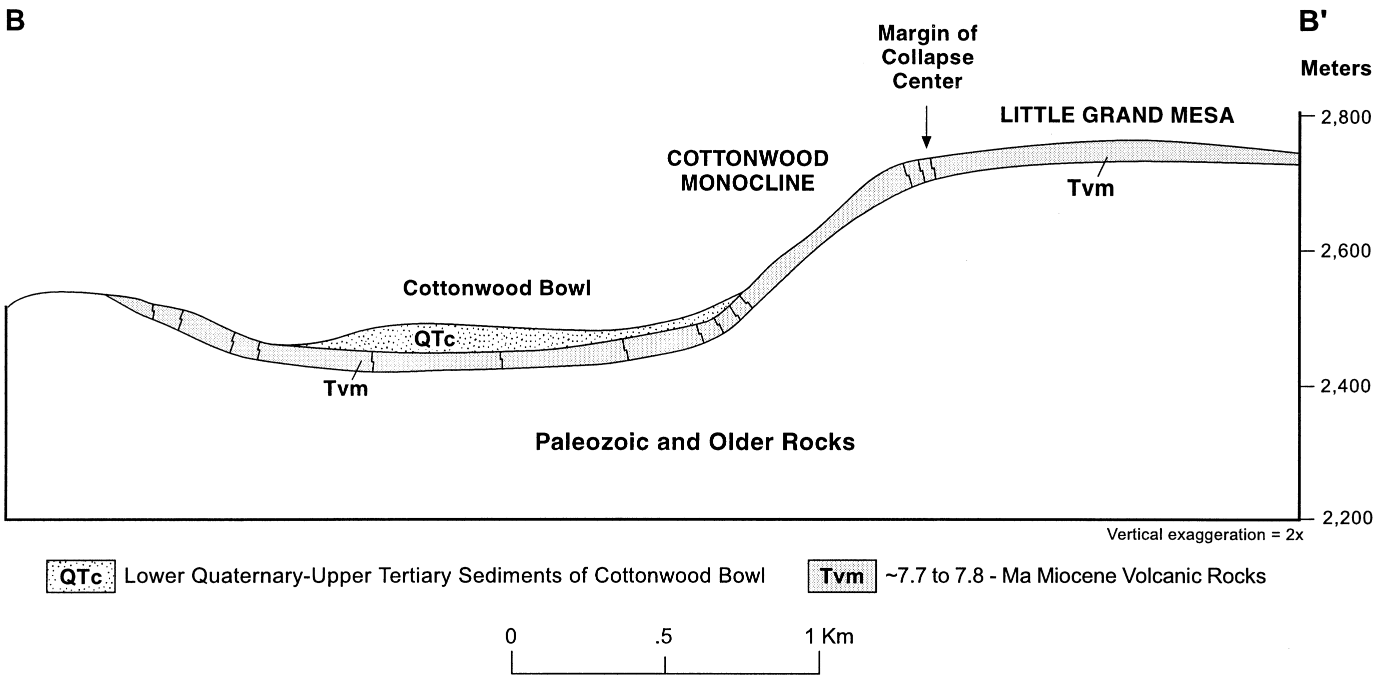

Figure 12. Cross section B–B′ through the Cottonwood monocline on the northeast side of Cottonwood bowl, northern margin of collapse center. Location of cross section shown in Figures 2 and 11.

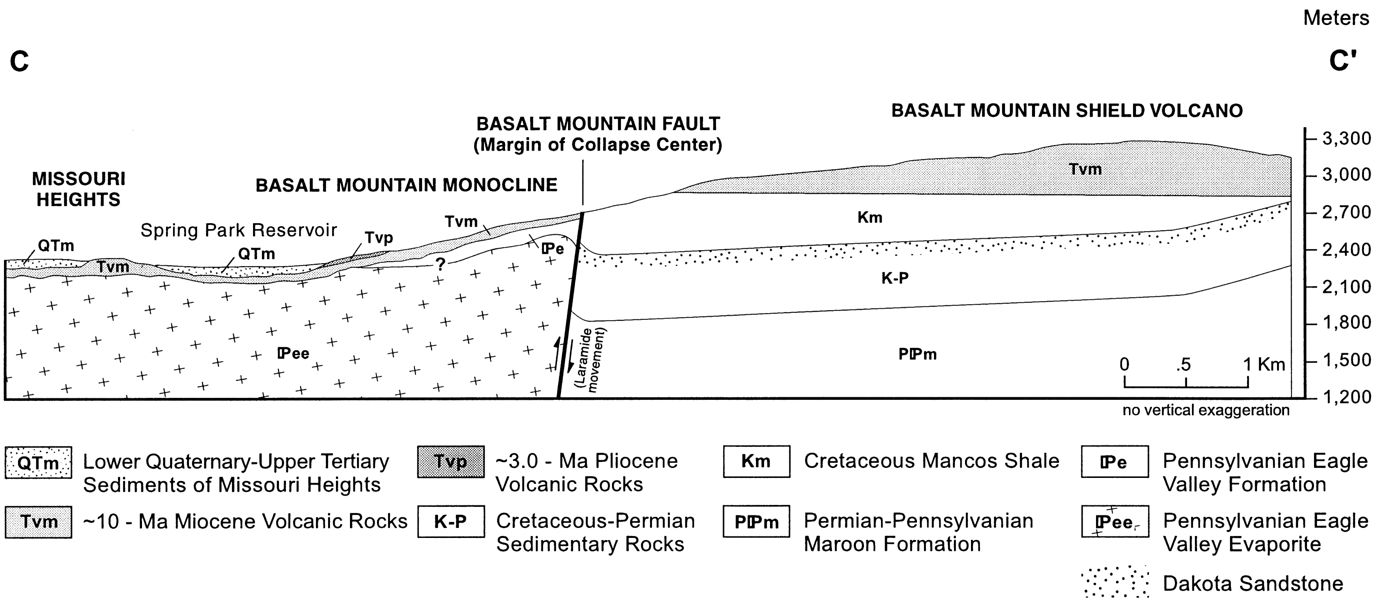

Figure 13. Schematic cross section C–C′ through the Basalt Mountain monocline on the eastern margin of the collapse center. Approximate location of cross section shown in Figure 2.

west-central Colorado (Kirkham et al., 2001a). Thus, the Basalt Mountain volcano either was lowered by collapse as an intact block, or the late Miocene landscape had several hundred meters of relief. Flows from Basalt Mountain volcano are the only ca. 10 Ma eruptions in the study area known to have flowed into and blocked a late Miocene paleovalley, so perhaps there was appreciable topographic relief during the late Miocene.

Southern margin. The southern margin of the collapse center is more poorly defined than the other sides, largely because upper Cenozoic volcanic rocks are scarce (Fig. 2). Thick deposits of middle(?) and late Tertiary fluvial gravel are preserved on the drainage divide between the Roaring Fork and Crystal Rivers and on the west side of the Crystal River. We conclude that these gravels accumulated in a gradually subsiding basin related to evaporite tectonism. Therefore the margin of the collapse center lies south of these deposits.

Fluvial gravels on Light Ridge south of Basalt abruptly thicken where they cross the postulated margin of the late Cenozoic collapse area (Fig. 2). A group 13a basalt flow dated at 13.57 ± 0.05 Ma (Kunk et al., 2001) is interbedded with the fluvial gravels at the north end of Light Ridge. This flow is lowered by collapse at least 730 m, based on its elevation relative to age-equivalent flows outside the collapse area (Larson et al., 1975). To accommodate the basalt flow and align with the abrupt thickening of gravel on Light Ridge, the margin of the collapse center, which trends north-northeast on the west side of Basalt Mountain, must first bend ~110° to an east strike, then swing sharply southward (Fig. 2). The late Cenozoic collapse margin probably turns to the west just south of Light Ridge, and then runs generally west or northwest along the north side of the Mount Sopris pluton until rejoining the Grand Hogback monocline southwest of Carbondale.

A 35.21 ± 0.03 Ma ash-flow tuff (Kunk et al., 2001) on the northeast side of Mount Sopris (Fig. 2) also provides evidence of the amount of subsidence on the southern margin of the collapse center. The top of an age-equivalent intrusive stock crops out at the summit of the nearby Mount Sopris (Streufert, 1999), therefore the land surface at the time of intrusion must have been equal to or higher than the top of the stock. It is unlikely that a deep paleovalley existed in close proximity to the stock at the time of its emplacement. We conclude that the ash-flow tuff was originally erupted onto a ground surface that was probably equal to or perhaps even higher than the top of the Mount Sopris stock and was subsequently lowered by collapse. The tuff now crops out 1290–1460 m lower than the top of the stock. Where last observed, the tuff disappears beneath the Tertiary sediments that fill Sopris bowl and probably continues into the subsurface for some unknown depth beneath its lowest outcrop. If the tuff was erupted onto a surface equal to or higher than the top of the Mount Sopris stock, then the total amount of post–35 Ma collapse probably exceeds 1460 m.

Structures within the interior of the collapse area

Structural deformation within the interior of the collapse center involves both major and minor structures that affect rocks and surficial deposits that overlie the evaporite. The style and amount of deformation in the volcanic rocks, which were erupted onto a low-relief surface prior to collapse, vary greatly. Locally the volcanic rocks are nearly flat lying, even though they were lowered a thousand meters or more in elevation. Elsewhere, well-defined major structures, some with >300 m of local structural relief, disrupt the volcanic rocks. The flows are near vertical and perhaps overturned in at least one structure (Hudson et al., this volume). We call elongate structural depressions formed in volcanic rocks overlying the evaporite synclinal sags. Basin-like structural depressions in bedrock are referred to as bowls. Locally within the collapse center the basaltic rocks are highly broken, fractured, and jostled; in these areas individual structures are difficult to map at a scale of 1:24 000. These intensely deformed deposits are mapped as collapse debris. Minor structural depressions formed in Pleistocene terraces are called troughs. Sinkholes are prevalent throughout the collapse area (Mock, this volume) and common in some troughs.

Heuschkel Park sag. Heuschkel Park sag is a generally east-west-trending, major synclinal structure in volcanic rocks north of Carbondale (Fig. 2). This structure, along with the faulted and tilted basalt-capped mesa north of it, form prominent landforms readily apparent on aerial photographs (Fig. 14). The axis of the Heuschkel Park sag underlies the broad and flat floor of Heuschkel Park. East of the park, the axis of the sag gradually bends northward and eventually turns back to west, creating a fishhook type of map pattern for the fold axis. To the west, the sag axis appears to transition into a faulted structure.

Group 1b flows are widely distributed across the area (Figs. 6 and 15). These 9.7–10.9 Ma flows (Kunk et al., this volume) characterize the structural deformation of the Heuschkel Park sag. Paleomagnetic studies at seven sites in and adjacent to the Heuschkel Park sag indicate the flows on the north limb of the sag may be strongly draped over the limb (Hudson et al., this volume). These data demonstrate south tilts as steep as 53°, which exceeds the slope of the north valley wall. Beneath the floor of Heuschkel Park, surficial deposits conceal the flows, but they are essentially flat lying where they crop out on the east end of the park, an interpretation supported by paleomagnetic studies. The low ridge south of Heuschkel Park (Fig. 15) is capped by group 1b flows in the southern limb of the sag. These flows dip north, into the axis of the sag. The only other volcanic rocks preserved on the south side of Heuschkel Park are on Red Hill, where three samples of the 10.6 Ma group 1c rocks (Table 1) were collected. Although these flows cannot be used to precisely characterize the structure of the south limb of the sag, their age and position in the landscape indicate the structural relief on the south limb of the sag is much less than that on the north limb.

North of the north valley wall is a gently to moderately northward-sloping mesa that is capped by a stacked sequence of basalt flows. The upper three flows in an outcrop on the rim of the mesa north of the east end of the park are group 1b rocks. A small-displacement, east-west-striking fault probably separates the north-dipping flows beneath the mesa from the south-dipping flows on the valley wall of the park. Total structural relief in the group 1b volcanic rocks between the axis of the sag and the north-sloping basalt-capped mesa is ~300 m. The north-sloping basalt-capped mesa is broken by >20 faults that trend either about N35°–55°E or N10°–45°W. Many of the faults abut or intersect at nearly right angles. Grassy meadows are

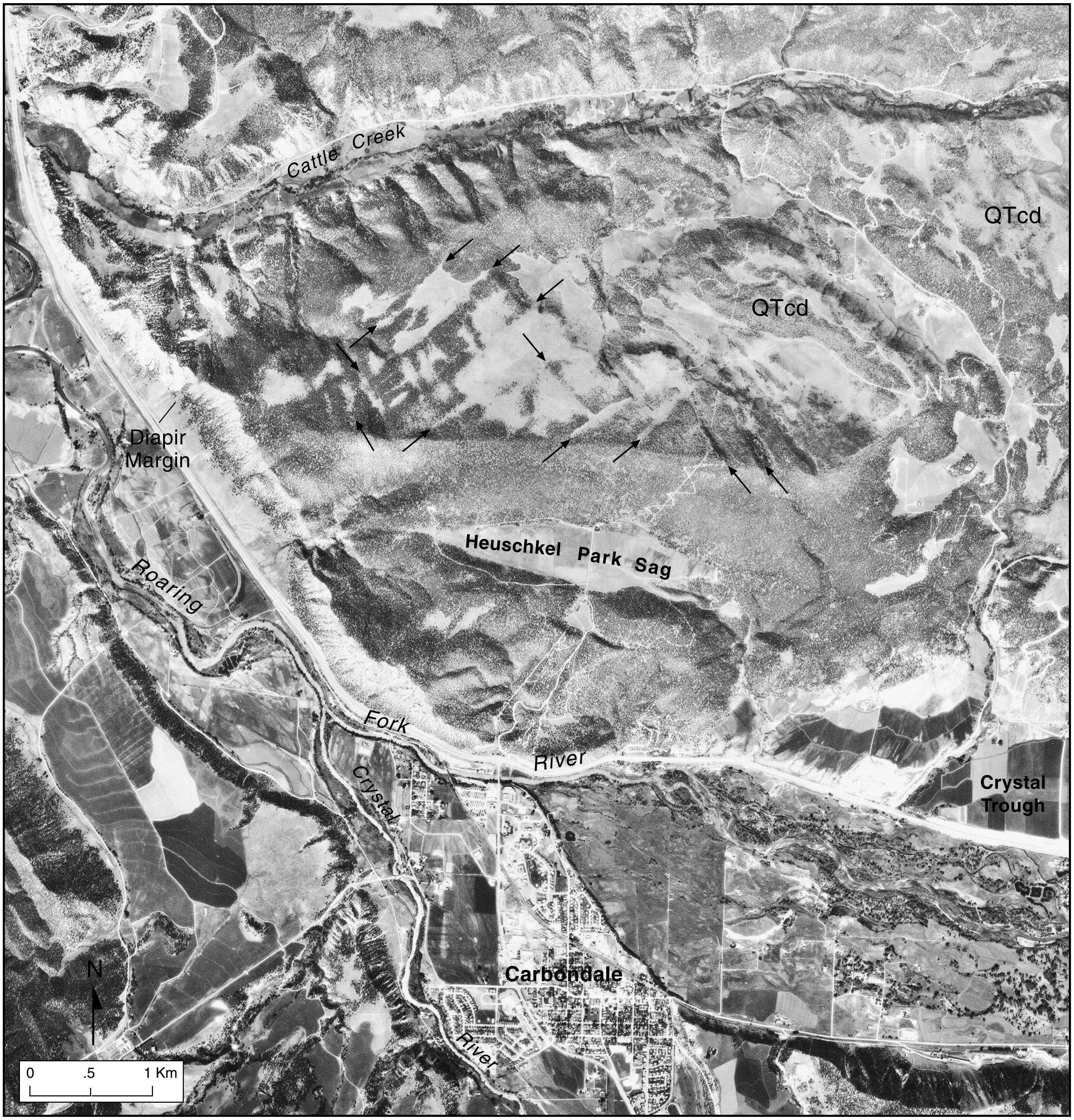

Figure 14. Vertical aerial photograph of the Carbondale to Cattle Creek area (NAPP photograph 6717-236). Axis of Heuschkel Park sag trends approximately east-west through the grassy meadow of Heuschkel Park, which is underlain by nearly horizontal ca. 10 Ma group 1b basaltic flows (Table 1). Steeply south-dipping ca. 10 Ma flows are preserved on the tree-covered slope north of the park. Farther north is the north-tilted mesa capped by ca. 10 Ma basalt, which is cut by several prominent faults (some marked by arrows) that intersect to form a rectilinear pattern. This faulted basalt cap grades northeastward into collapse debris (QTcd) that underlies the hummocky topography in the right side of the photograph. Scale and north arrow are approximate.

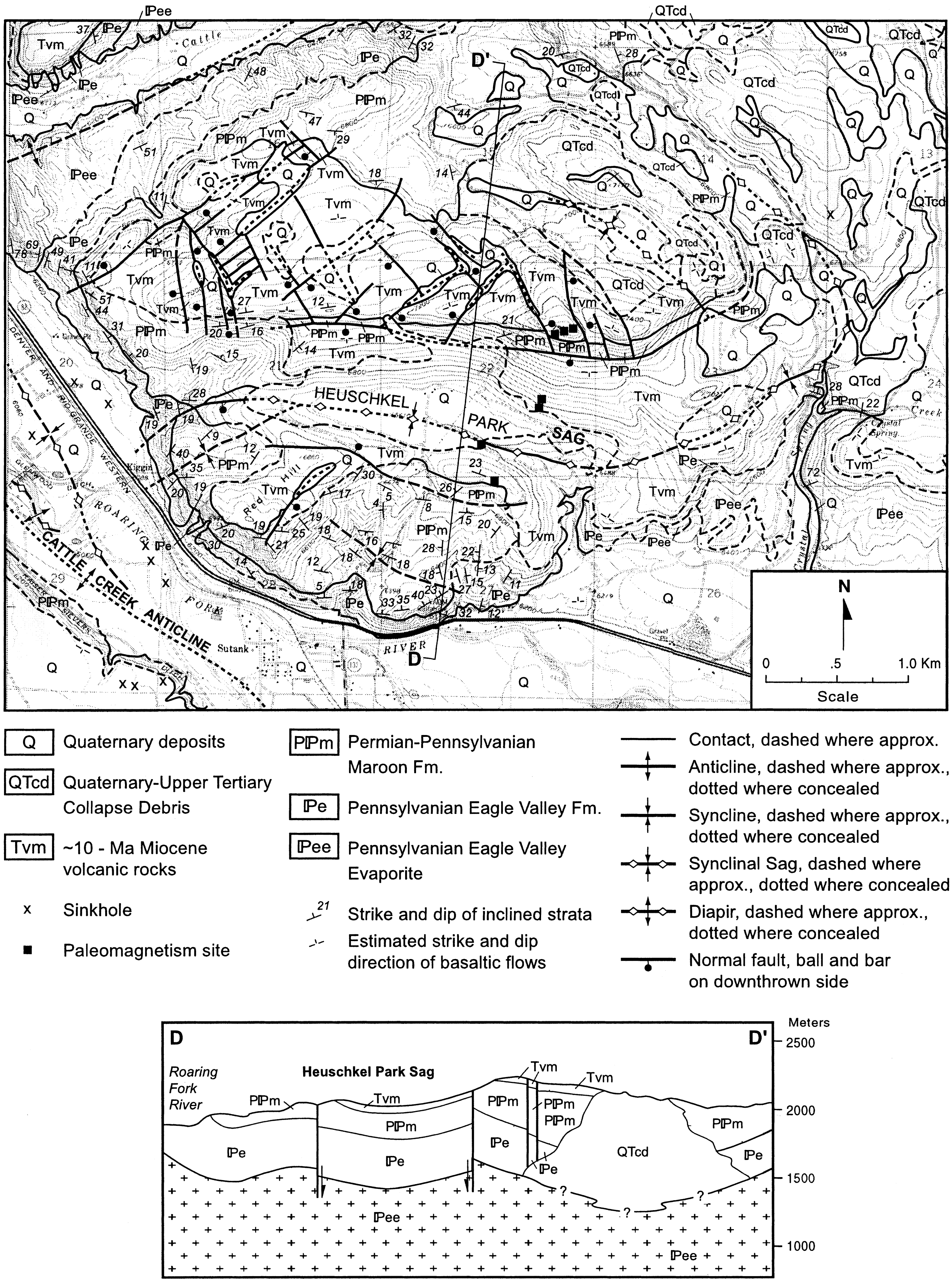

Figure 15. Geologic map of Heuschkel Park sag and cross section D–D′ (adapted from Kirkham and Widmann, 1997).

present along many of the faults, and they contrast sharply with the adjacent pine-covered hills between the faults, creating quasi-rectilinear vegetation patterns on the mesa (Fig. 14).

The Heuschkel Park sag may simply be an asymmetrical syncline due to removal of evaporite from beneath it. However, the block of rock that includes the sag protrudes into and somewhat restricts the valley floor of the Roaring Fork River (Fig. 3). This block of rock may be rafting southward on underlying evaporite, as the evaporite flows toward the river.

Spring Valley structure. Spring Valley is a nearly 6-km-long, broad, flat, arcuate valley between Glenwood Springs and Carbondale on the east side of the Roaring Fork River (Figs. 2 and 3). It is ~300 m higher than the floor of the river valley, and its orientation and position suggested Spring Valley was a former paleovalley of the Roaring Fork River. However, fluvial gravel deposits related to an ancestral Roaring Fork River have not been found in or near Spring Valley. When early settlers first homesteaded this region, Spring Valley was a closed depression and a lake occupied much of the valley floor (Calvin Cox, 1994, personal commun.). The presence of lacustrine sediments beneath much of the valley supports this conclusion. The settlers hand dug a drain ditch at the north end of the valley shortly before the end of the 19th century, drained the lake, and turned the valley floor into agricultural land.

The oldest upper Tertiary volcanic rocks within the collapse area crop out at the northern end of Spring Valley. These group 10a rocks (Table 1), which are 22.56 ± 0.13 Ma, are only ~300 m above the floor of the Roaring Fork River valley and are >1200 m lower in elevation than age-equivalent flows in the White River Plateau (Larson et al., 1975). Some of the best outcrops of these early Miocene flows are within a narrow graben at the northwest end of the Spring Valley structure (Fig. 2). Paleomagnetic studies of the early Miocene flows at three sites within the graben indicate these rocks are strongly tilted (Hudson et al., this volume). A model tilt of 100° ± 15° southwest was calculated for the flow on the east side of the graben, suggesting it is overturned or nearly vertical.

The Spring Valley structure is concealed beneath the surficial deposits that fill the valley floor, and subsurface data are limited, so the structure is poorly understood. Los Amigos Mesa, which is capped by 7.75 Ma flows (group 5b) and ca. 10 Ma flows (groups 1b, 2b, and 3a), lies between Spring Valley and the Roaring Fork River valley. Volcanic flows in the northern end of this mesa are deformed by several small-displacement faults, and at least one synclinal sag occurs on the edges of the mesa (Fig. 2), but overall the mesa is little deformed. Along the western edge of the mesa, the upper Miocene volcanic rocks dip away from the river valley as much as 13°. The underlying Maroon Formation dips more steeply away from the river valley than do the overlying flows, indicating pre–middle Miocene movement on the Cattle Creek anticline and/or Roaring Fork diapir.

Much of the mountain side east of Spring Valley is underlain by broken, jumbled blocks and boulders of basalt in a large landslide complex. This landslide complex developed in response to the collapse of the Spring Valley structural depression, which removed the rock that supported the hillslope east of the valley, effectively oversteepening the east valley wall. This triggered movement of the hillslope toward the valley as a large landslide complex. Large blocks of relatively intact basalt are preserved within the landslide, but much of the basalt was broken into bouldery rubble. Large scarps, which were mapped as down-to-the-west faults by Kirkham et al. (1995) and Kirkham and Widmann (1997), cut both the landslide deposits and adjacent bedrock. These structures may be major slip planes related to massive failure of the entire rock mass and overlying surficial deposits on the east side of Spring Valley.

A 174-m-deep groundwater exploration test hole ("SVR6" in Figure 2) in the middle of the valley failed to reach bedrock (Robin VerSchneider, 2001, personal commun.). Most material penetrated by the well was fine-grained silt and clay, and no basalt flows were encountered. The abundant clay and silt found in the well suggests a lacustrine environment persisted in Spring Valley for much of the time interval represented by the beds penetrated in the drill hole. A thick sequence of reworked volcanic ash and clastic sediments eroded from adjacent hillslopes was present at depths of ~75–91 m in the drill hole. A.M. Sarna-Wojcicki (2002, written commun.) geochemically correlated the ash with the Lava Creek B ash, which was recently dated at 639 ka by Lanphere et al. (2002). The depth of burial of the ash indicates the average sedimentation rate in the valley since the middle Pleistocene was ~140 mm/k.y. This rapid sedimentation rate suggests the Spring Valley structure had a high rate of deformation during the past 639 ka.

Kirkham et al. (1995, 1997) and Kirkham and Widmann (1997) considered the Spring Valley structure to be a half graben whose floor was tilted to the west (Fig. 16, upper diagram). In this model, a large normal fault forms the western margin of the structure, and Spring Valley is a structural low on the downdropped eastern side of the fault. Basaltic flows would underlie the valley-fill deposits and overlie red beds of the Maroon Formation, if the structure is a half graben. As the valley tilted westward, the hills east of the valley gradually became oversteepened and eventually slid toward the valley.

An alternative model for the Spring Valley structure is presented in the lower diagram in Figure 16. In this model, the large block of relatively intact basalt and underlying post-evaporite Paleozoic sedimentary rocks between Spring Valley and the Roaring Fork River valley is being rafted into the river valley along an average travel path of S60°–70°W, as the underlying evaporite flows toward areas with lower lithostatic loads. The release zone for the structure coincides with the topographic outline of the valley, which is a roughly north-south-oriented valley with ends that flare back to the west. The western side of the rafted block would have slowly advanced into the Roaring Fork River valley. As shown in Figure 3, this block of rock does protrude into the river valley. In the rafted-block

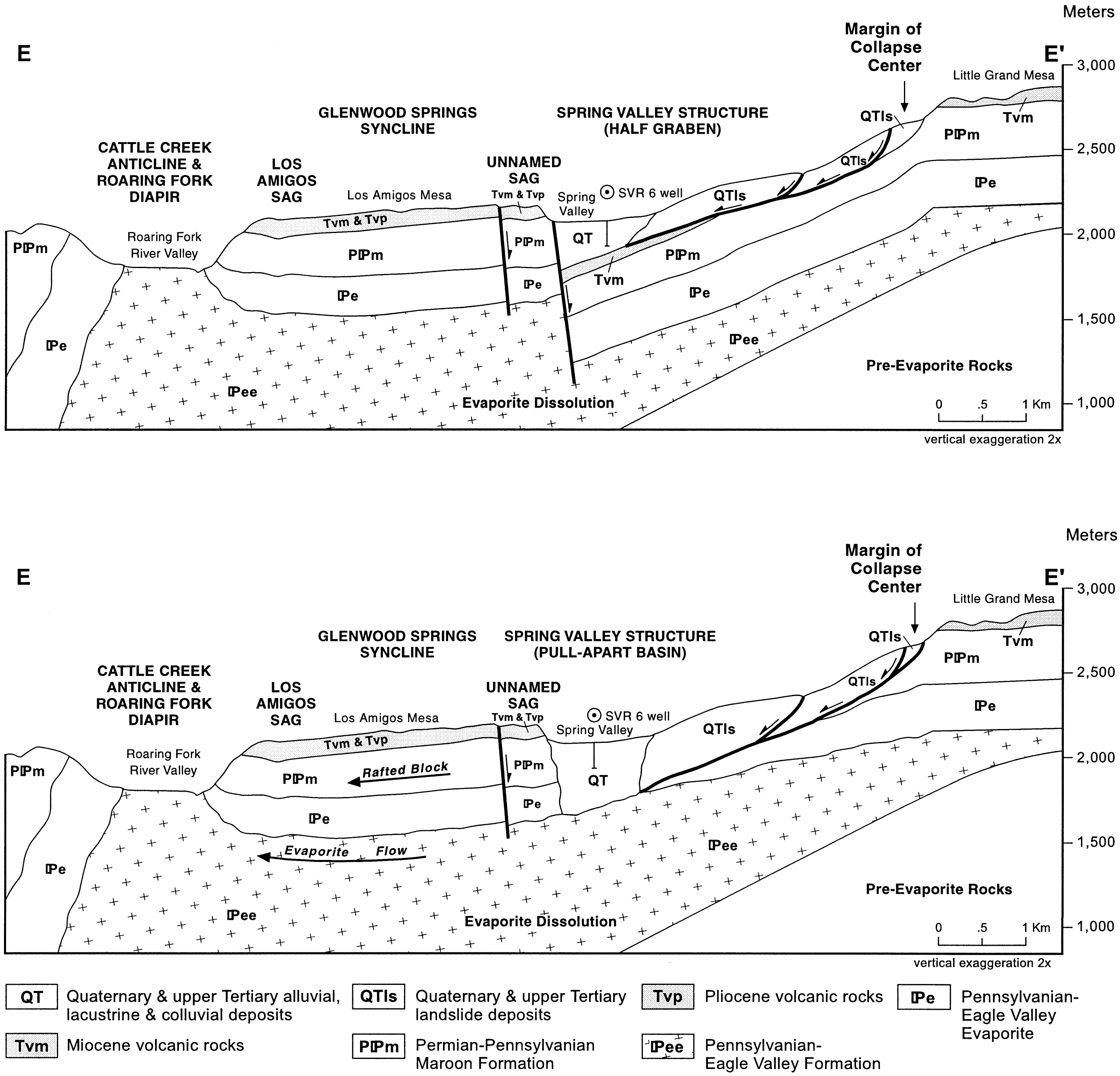

Figure 16. Schematic cross sections (E–E′) showing two possible interpretations of the subsurface geology through Spring Valley. Upper cross section depicts the Spring Valley structure as a half graben chiefly due to removal of evaporite from beneath the valley by dissolution. In the lower cross section, the Spring Valley structure is shown as a pull-apart feature caused by evaporite flow toward the Roaring Fork River valley. Location of cross section shown on Figure 2.

model, the sediment deposited in the Spring Valley structure would directly overlie evaporite.

Strain along the southern margin of the hypothesized rafted block may be largely lateral shearing. The strain would occur as sinistral-strike slip or oblique slip along a narrow, arcuate fault zone that starts in the southeast corner of the structure, bends to a nearly west trend, and disappears into the evaporitic rocks that crop out beneath the volcanic rocks on the east wall of the river valley (Fig. 2). Lateral strain on the northern margin of a rafted Spring Valley structure would in part be accommodated by the narrow graben that extends northwestward from Spring Valley and in part by the complex fault swarm between the graben and the river valley.

Minor structures. Several small synclinal sags deform the volcanic bedrock in the Carbondale collapse center (Fig. 2). They include the Los Amigos sag, Polaris sag, Shippes Bowl sag, Shippes Draw sag, and Ti-Ke-Ki sag (Kirkham et al., 1996, 1998, 2001b). The previously described Heuschkel Park sag has sharp walls and considerable structural relief, but others have broad, relatively flat limbs with only meters or tens of meters of structural amplitude.

Los Amigos sag is a 1500-m-long, slightly arcuate, northwest-trending feature adjacent and parallel to the Roaring Fork River valley. Polaris sag, Shippes Bowl sag, and Shippes Draw sag are 1.6- to 2.6-km-long, northeast- or east-trending structures in the northeastern part of the collapse area (Fig. 2). They lie on a nearly horizontal structural bench underlain by Miocene volcanic rocks between Cottonwood monocline and Cattle Creek (Fig. 7). Topographically, the Polaris sag and Shippes Draw sag are broad, shallow depressions, but a prominent, 240-m-deep, nearly circular topographic depression is associated with Shippes Bowl sag (Fig. 17). Basalt flows are draped across the eastern, northeastern, and perhaps western flanks of Shippes Bowl sag, but are absent on other flanks. The 2.4-km-long Ti-Ke-Ki sag formed in basalt flows on the interfluve between the valleys of the Roaring Fork and Crystal Rivers.

Within the east-central part of the collapse center the basaltic flows are highly broken, fractured, and jostled. Kirkham and Widmann (1997) described these deposits as collapse debris. Individual structures within the collapse debris are typically too complex to map at a scale of 1:24 000. Collapse debris locally includes varying amounts of unconsolidated surficial deposits that have collected over or between the blocks of broken, locally rubbly bedrock. Much of the Missouri Heights area is underlain by collapse debris (Fig. 2).

Ten closed, or nearly closed, linear topographic depressions were identified in Pleistocene terraces along the Roaring Fork and Crystal Rivers (Fig. 2). These depressions, which parallel the subjacent river channels, are interpreted as subsidence troughs resulting from bending due to roof collapse into dissolution cavities formed in evaporite beneath the terraces. Sinkholes are associated with many of the troughs.

The Barbers Gulch trough southwest of Carbondale (Fig. 2), which formed in a middle Pleistocene terrace west of the Crystal River, is ~3.5 km long, 0.6 km wide, and 12 m deep; it is the largest subsidence trough in the area. In addition to locally derived alluvium and colluvium, the trough contains a bed of Lava Creek B ash that lies ~73 m above the Crystal River (Kirkham and Widmann, 1997). A subsidence trough likely existed at this location prior to deposition of the 0.64 Ma ash (Lanphere et al., 2002), as topographic depressions are good environments in which to preserve tephra deposits. The ash bed appears to be slightly tilted (Kirkham and Widmann, 1997), which suggests trough subsidence also postdates tephra deposition.

Other notable subsidence troughs include the Crystal ter-

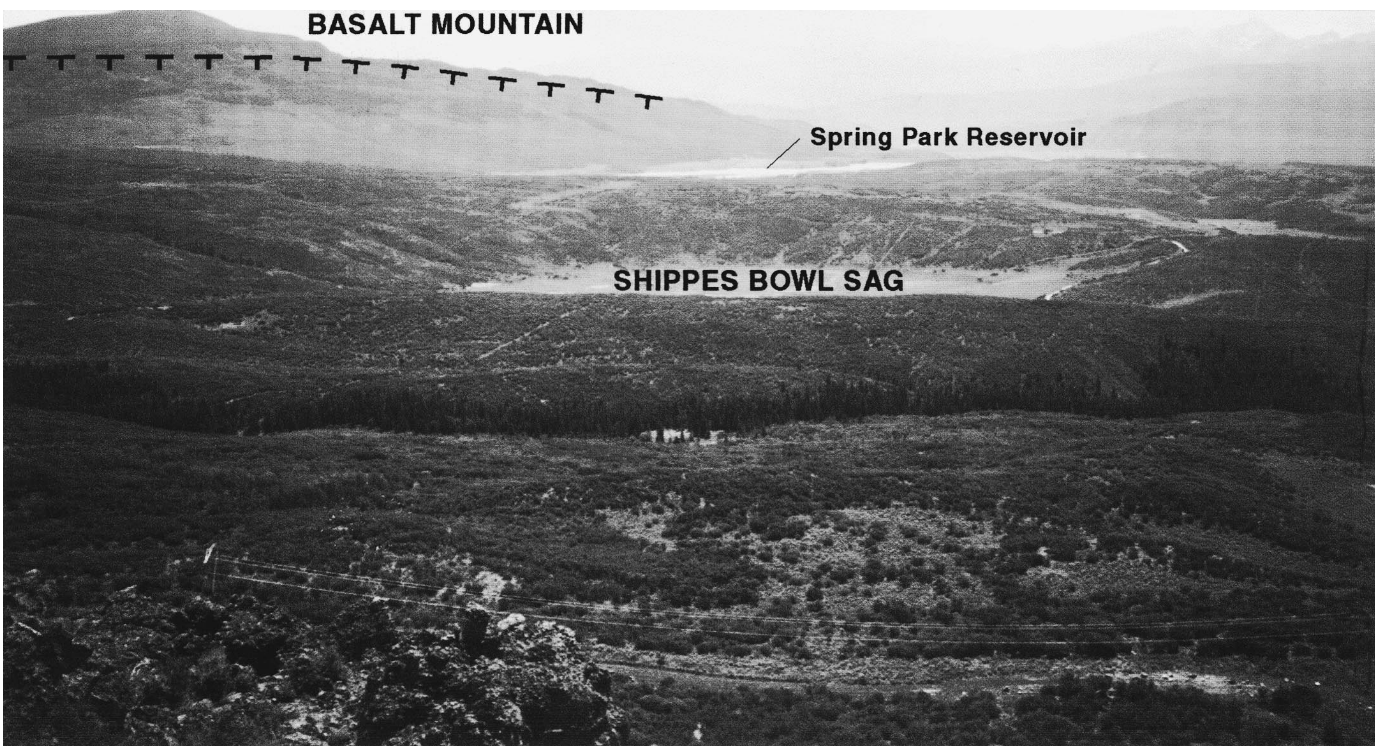

Figure 17. Oblique aerial photograph of Shippes Bowl sag (SBS), looking south. Basalt Mountain forms the skyline on the left side of the photograph, and Spring Park Reservoir is in the middle distance between the sag and the base of Basalt Mountain. Margin of collapse center, immediately below the top of Basalt Mountain, shown by dashed line with tick marks.

race trough, Blue Creek trough, and Leon trough along the Roaring Fork River (Fig. 2). Crystal terrace trough is a well-preserved, symmetrical depression on a late Pleistocene terrace east of Carbondale. The axis of the trough is ~10 m lower than the edges of the terrace. Blue Creek trough is a narrow depression formed on the outer or valley-wall side of a late Pleistocene terrace north of El Jebel. Blue Creek makes a sharp, nearly right-angle bend as it enters the trough and flows roughly parallel to the Roaring Fork River for nearly 1700 m. Leon trough is south of El Jebel. Collapse of this nearly circular depression involves a latest Pleistocene terrace that was dropped to the level of the Holocene valley-fill deposits. Holocene fluvial sediments may partially fill Leon trough. A slightly smaller trough, adjacent to and northeast of Leon trough, lies on the outer edge of the valley floor. This trough, in which a lake for water skiing was recently built, is separated from the Leon trough by a topographic high that likely is a narrow remnant of the original terrace tread.

Sinkholes are prevalent throughout the collapse area (Mock, this volume). They occur in outcrops of evaporite and in bedrock and surficial deposits that overlie evaporite. Sinkholes are most numerous along the Roaring Fork diapir, where on average there is one known sinkhole every 3.9 km^2. The sinkholes provide direct evidence of evaporite dissolution. Some sinkholes may be thousands of years old, but dozens have disrupted the ground surface during the past few years or decades.

Most known sinkholes that formed during historic time are associated with irrigated fields, irrigation ditches, ponds, and lakes. This coincidence may be due to the relatively abundant supply of fresh water found in these areas, which can cause or enhance sinkholes by evaporite dissolution and can induce roof collapse of underground voids by piping. Also, the close attention paid to agricultural fields and water-supply systems could account for more complete reporting of sinkholes in those areas. Spring Park Reservoir, an irrigation reservoir between Carbondale and Basalt Mountain (Fig. 2), was rapidly drained twice during the twentieth century when sinkholes developed in the floor of the reservoir (Steve Callicotte, 1996, personal commun.). Open, sinuous voids are present in many exposures of evaporite. Sinkholes and open voids may be interconnected, forming groundwater flow paths with very high transmissivity. A spring with high discharge flows from an open void ~1.3 km north of El Jebel. Individual sinkholes may be very large; one in a basalt flow that overlies evaporite near Colorado Mountain College (Fig. 1) is ~70 m wide.

COLLAPSE-RELATED SEDIMENTARY DEPOSITS

Syn-collapse sedimentary deposits accumulated in four major structural depressions formed in bedrock: The previously described Spring Valley structure and Cottonwood bowl, the Sopris bowl on the north side of Mount Sopris, and the Cattle Creek bowl along the east-central part of the collapse area (Fig. 2). Both the Spring Valley structure and Cottonwood bowl are topographically well expressed, but Sopris bowl and Cattle Creek bowl were recognized chiefly by the preserved remnants of sediments originally deposited within them.

The thickest and laterally most extensive syn-collapse sedimentary deposits fill Sopris bowl (Kirkham and Widmann, 1997; Kirkham et al., 1998; Streufert et al., 1998; Streufert, 1999). Large remnants of these deposits are found on the interfluve between the Roaring Fork and Crystal Rivers and in the hills on the west side of the Crystal River (Fig. 18). The sediments include gravelly clast-supported fluvial deposits and matrix-supported debris-flow deposits. Clast lithologies indicate an ancestral Crystal River provenance for deposits in the western part of the basin and an ancestral Roaring Fork River provenance in the eastern part. The paleovalley exposed on Light Ridge (Figs. 2 and 18) probably accommodated the ancestral Roaring Fork River.

The exposed thickness of syn-collapse deposits in Sopris bowl exceeds 450 m in the central part of the bowl (Fig. 18).

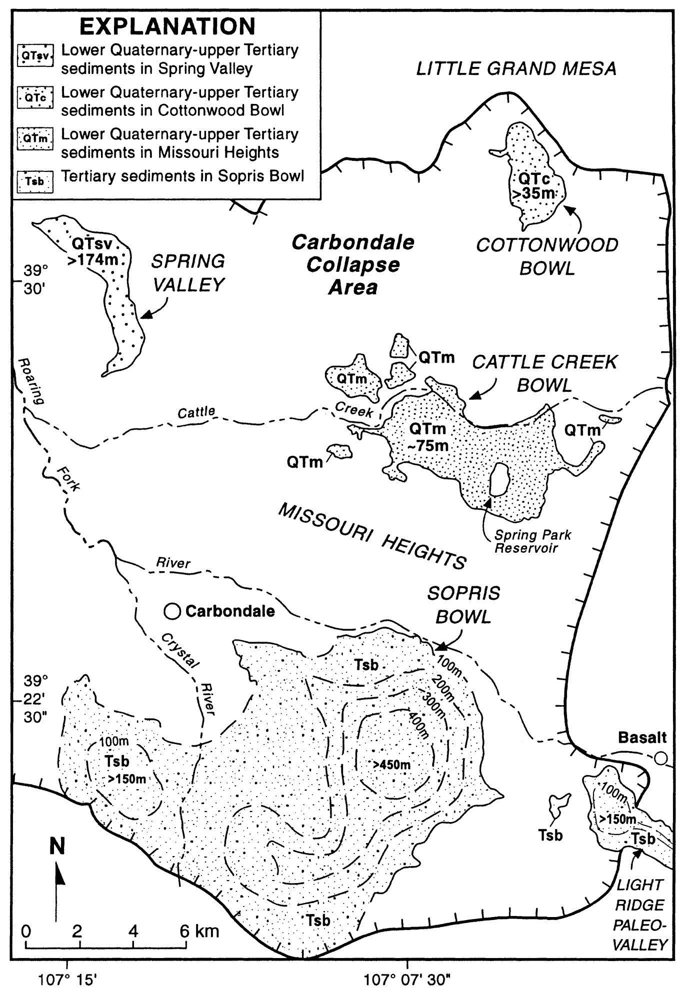

Figure 18. Distribution and minimum preserved thickness of late Cenozoic syn-collapse sediments. Contour interval 100 m.

The base of the syn-collapse deposits is concealed beneath the ground surface in most of the bowl. Where exposed west of Basalt, the basal contact of the syn-collapse sediments slopes into Sopris bowl with an apparent dip of ~9° to the northwest (Streufert et al., 1998). This suggests the maximum preserved thickness of these syn-collapse deposits greatly exceeds 450 m. Seismic reflection data collected from the southwestern part of Sopris bowl suggest the syn-collapse deposits attain a preserved thickness of ~1100 m (Perry et al., this volume).

An ash-flow tuff underlies the syn-collapse sediments in the southeast corner of Sopris bowl (Fig. 2). The tuff dips ~20° to the northeast (Streufert et al., 1998) and contains sanidine dated at 35.21 ± 0.03 Ma (Kunk et al., 2001). Deformation associated with Sopris bowl must have begun sometime after eruption of the ash-flow tuff. Basalt flows west of El Jebel (groups 12a and 12b) with an average age of 13.3 Ma (Table 1) are stratigraphically at or near the top of the syn-collapse sedimentary sequence, which indicates Sopris bowl continued to fill with sediments until near the end of the middle Miocene. Since the 13.3 Ma flows are tilted and are at elevations substantially lower than other middle Miocene flows found outside the collapse area, major subsidence continued in Sopris bowl after 13.3 Ma. The proximity of Sopris bowl to Mount Sopris stock, and age relationships between the formation of the bowl and emplacement of the stock, suggest increased geothermal gradients and other hydrologic changes associated with the stock may have caused or enhanced evaporite dissolution.

A laterally extensive but relatively thin remnant of syn-collapse sediments deposited in Cattle Creek bowl is preserved on the drainage divide between Cattle Creek, Missouri Heights, and Spring Park Reservoir (Fig. 18). A west-flowing stream deposited these sediments in fluvial, deltaic, and lacustrine environments. Coarse-grained, sandy, silty, cobble and pebble gravel eroded from the mountains east of the collapse area was deposited in the eastern and central parts of Cattle Creek bowl. Fine-grained sediments found farther west in the bowl suggest that collapse created closed topographic depressions favorable for lacustrine and deltaic deposition.

The syn-collapse sediments in Cattle Creek bowl unconformably overlie ca. 10 Ma group 1b volcanic rocks, although in places the group 1b volcanic rocks crop out in hills that are topographically higher than adjacent syn-collapse sediments. As the syn-collapse sediments were first deposited, these hills of volcanic rock formed islands that rose above the streams and lakes in which the sediments accumulated. The sediments in Cattle Creek bowl also locally overlie 7.75 Ma volcanic rocks, but stratigraphic relationships with the Pliocene (3.05 ± 0.04 Ma) group 6c rocks are equivocal.

EVAPORITE FLOW

Evidence of evaporite flow in the lower Roaring Fork River valley includes valley-centered anticlines, diapiric contacts between evaporite and overlying rocks, and folded Quaternary deposits. The Cattle Creek anticline, first described by Mallory (1966), underlies the Roaring Fork River valley from Glenwood Springs to near Carbondale (Fig. 1). Evaporite in the core of this anticline is at least locally diapiric (Figs. 4 and 19). This structure coincides with the upper limb of the Laramide Grand Hogback monocline in much of the lower Roaring Fork River valley. Evaporite flowed toward the upper limb of the monocline during the Laramide orogeny (Perry et al., this volume). The valley-centered anticline was enhanced by subsequent late Cenozoic evaporite flow and diapirism related to the Roaring Fork diapir. Over 100 m of diapiric piercement by the Eagle Valley Evaporite into overlying beds in the Middle and Upper Pennsylvanian Eagle Valley Formation is apparent in the prominent exposure on the east valley wall of the Roaring Fork River immediately south of Cattle Creek (Fig. 20).

Quaternary outwash terraces are deformed by evaporite flow at four locations in the lower Roaring Fork River valley between Glenwood Springs and Carbondale (Fig. 19). These tilted deposits dip away from the axis of the valley, even where the deposits are adjacent to the valley wall. This folding prob-

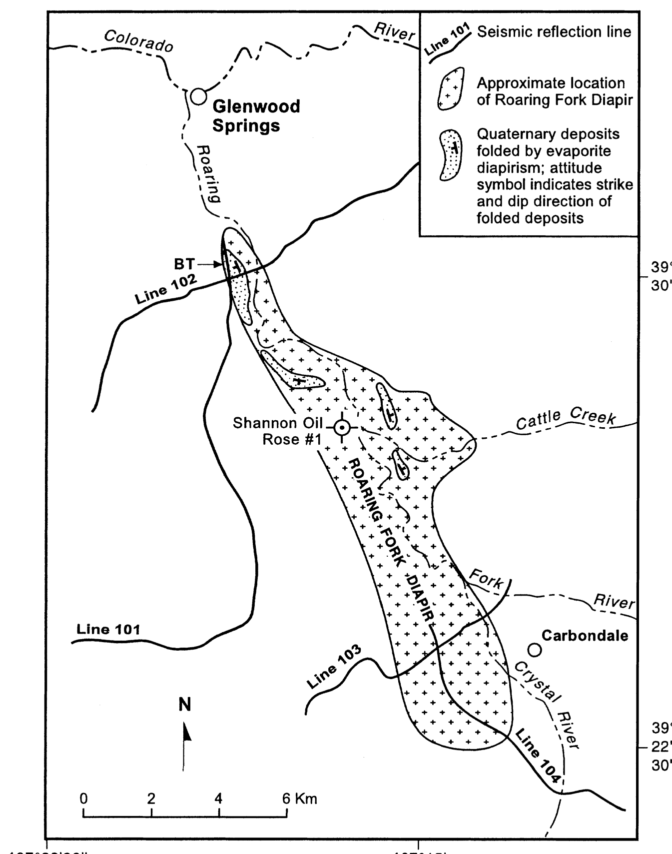

Figure 19. Map of the Roaring Fork diapir, Quaternary deposits folded by evaporite diapirism, and seismic reflection lines used to interpret the diapir and collapse center (modified from Kirkham et al., 2001b). Refer to Perry et al. (this volume) for interpretation of the seismic lines. BT—Bershenyi terrace.

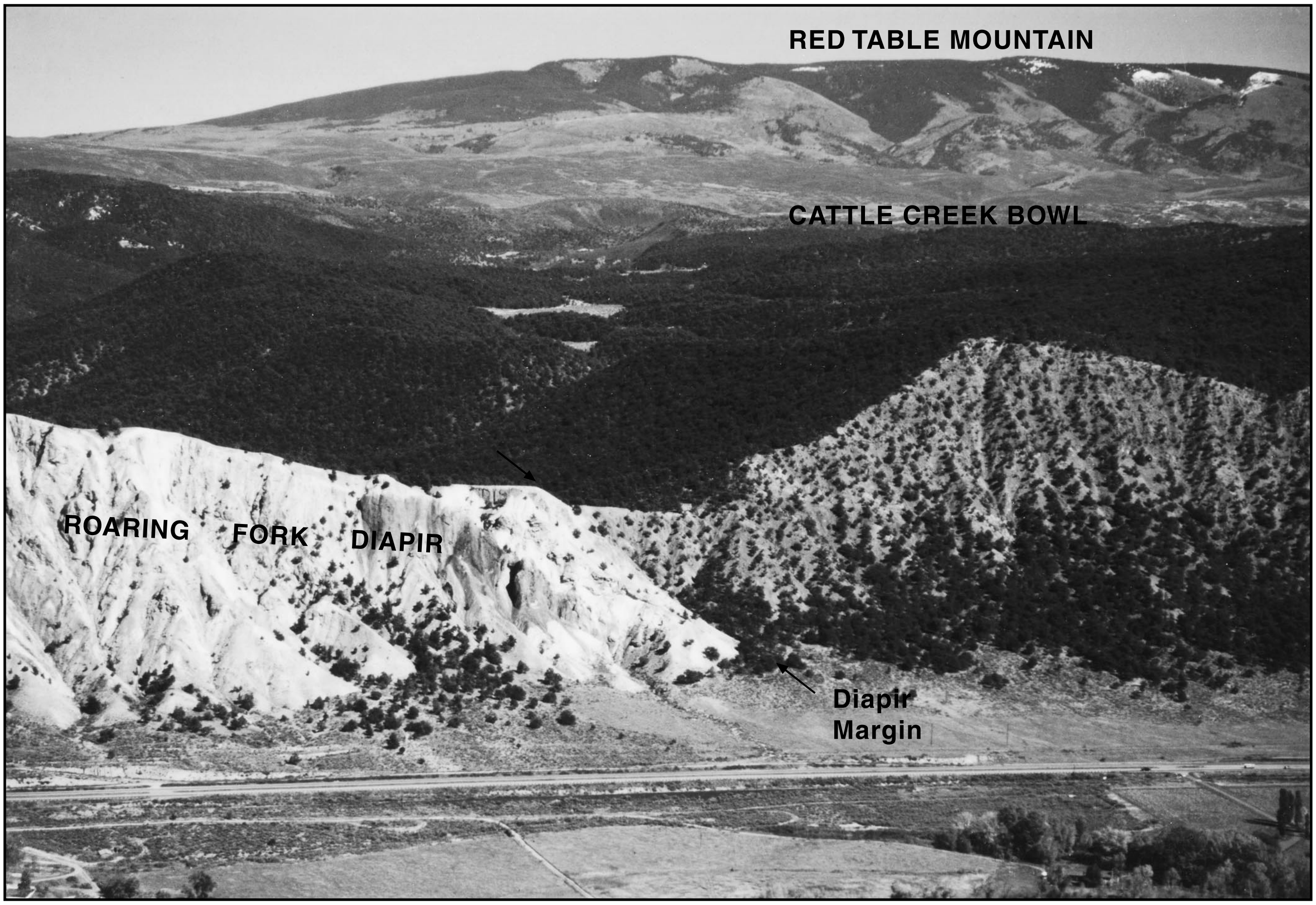

Figure 20. Photograph of Roaring Fork diapir. View is to east. Note diapiric contact between the Eagle Valley Evaporite and overlying Eagle Valley Formation (shown by arrows). Outcrop is on east side of the Roaring Fork River valley between Cattle Creek and the town of Carbondale.

ably resulted from upward evaporite flow in the Roaring Fork diapir during or after the middle Pleistocene, because Quaternary deposits that sag or collapse into subsidence troughs always dip into the subsided area. Unruh et al. (1993) described terraces deformed by evaporite flow along the west wall of the Roaring Fork River valley southwest of Basalt. These deposits appear to be uparched in an anticlinal structure that is oblique to the valley axis, which also is suggestive of deformation by upwelling evaporite.

At Bershenyi terrace, near the northern end of the Roaring Fork diapir (Figs. 5, 19, and 21), a Quaternary terrace and overlying deposits are deformed by evaporite flow. The middle Pleistocene Bershenyi terrace (Piety, 1981) is overlain by a wedge-shaped layer of debris-flow deposits that is thickest at the fan head (Kirkham et al., 1996, 1997). Where closest to the river, these deposits are folded upward and dip away from the river. The downstream (northern) end of the terrace is uparched such that debris-flow deposits at the distal end of the fan are now ~30 m higher in elevation than correlative deposits at the original fan head. The adjacent late Pleistocene airport terrace ("AT" on Fig. 21), is not visibly affected by evaporite tectonism. Measurable evaporite flow last occurred at this location after deposition of the middle to upper Pleistocene debris-flow deposits and prior to deposition of the undeformed, adjacent upper Pleistocene terrace deposits.

Seismic reflection data interpreted by Perry et al. (this volume) support our conclusion that evaporite flow and diapirism locally created evaporite-cored valley anticlines and uparched Quaternary terraces and associated deposits along the valley axis. The interpreted seismic lines suggest the evaporitic strata are thin beneath the Grand Hogback monocline and in the area east of the Roaring Fork River valley but up to 1.5 km thick in the ends of the intrusive diapir under the valley. The evaporite apparently withdrew from the adjacent upland areas and flowed into the valley-centered diapir. The interpretation of Perry et al. (this volume) suggests that the Roaring Fork diapir may locally pierce as much as a kilometer of overlying strata.

Evaporite flow may also be responsible for the Spring Valley structure and Heuschkel Park synclinal sag, which were previously described. Large, relatively intact blocks of rock that

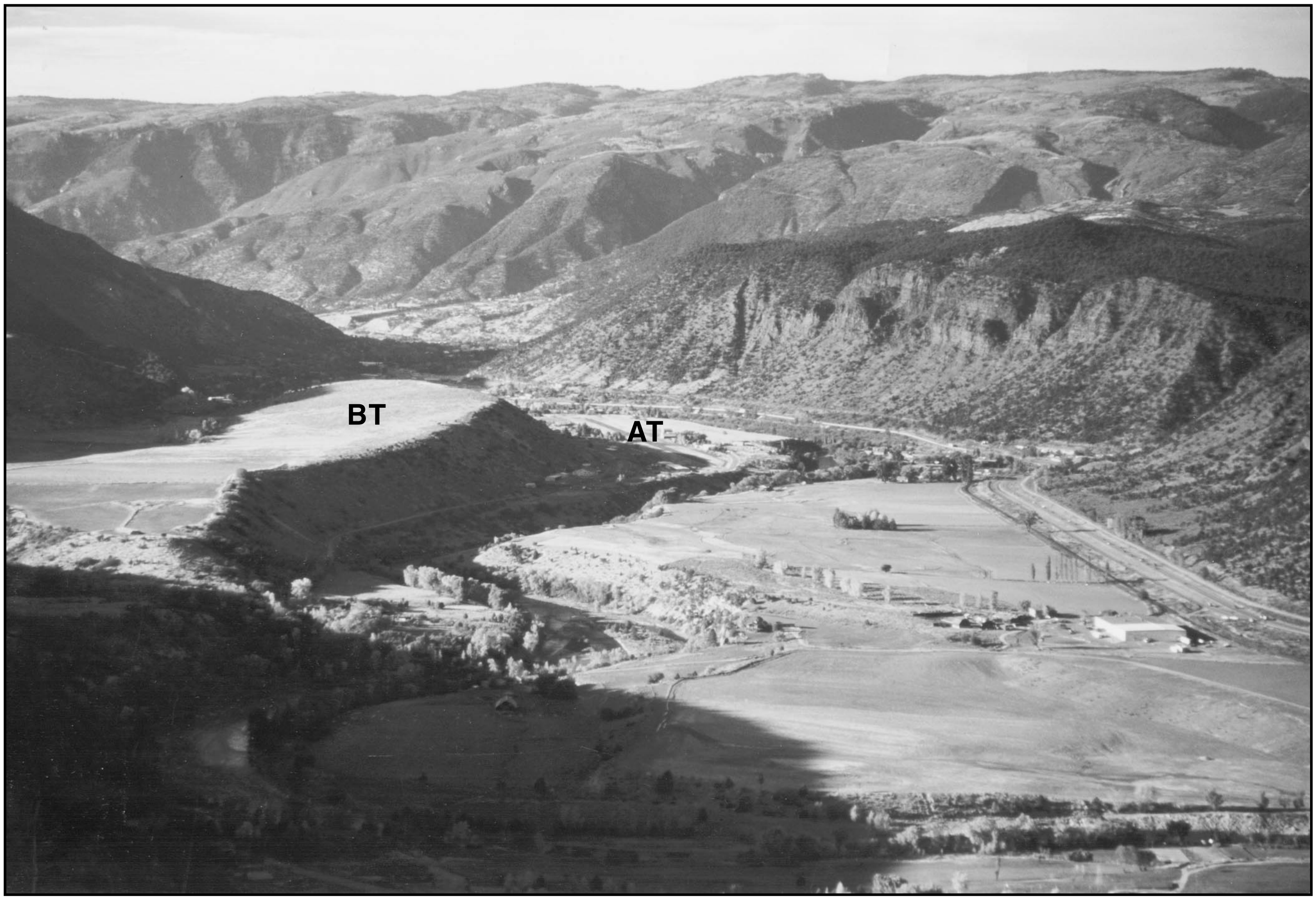

Figure 21. Photograph of Bershenyi terrace (BT, in the left center of photograph) looking north. The downstream end of this middle Pleistocene outwash terrace and the distal (northern) end of an overlying debris-flow fan are upwarped by evaporite flow. The adjacent late Pleistocene airport terrace (AT) is not deformed by evaporite flow. Location of terraces shown in Figures 5 and 19.

overlie the evaporitic strata between these structures and the river valley may be rafting into the valley as the underlying evaporite flows toward the valley.

TIMING OF COLLAPSE

Available data indicate that the rate of evaporite-related collapse has varied temporally and spatially. The earliest recognized evidence of post-Laramide evaporite collapse occurred between 35.21 Ma and 13.3 Ma when a deep, but localized dissolution basin formed immediately north of Mount Sopris. Collapse during this time period apparently was restricted to the area north of Mount Sopris.

Major subsidence across the entire collapse center began sometime after 10 Ma (Fig. 22), but its precise initiation is poorly constrained. In some areas, ca. 10 Ma volcanic rocks are downdropped more than the 7.7 Ma flows, suggesting some of the regional collapse occurred between ca. 10 Ma and 7.7 Ma. Definitive evidence of collapse during this time, such as greatly deformed older flows overlain by less-deformed flows, however, was not observed. Rates of river incision probably increased ca. 10-8 Ma, and greatly accelerated during the past 3 m.y. (Kirkham et al., 2001a). This incision in turn decreased lithostatic pressures on the evaporitic rocks beneath the valleys and triggered or enhanced evaporite flow. As the rivers cut deeper, fresh groundwater circulated to greater depths in the evaporite and promoted dissolution and collapse.

Along the eastern margin of the collapse area, near Spring Park Reservoir, a 2.90 Ma cinder cone and associated volcanic flow (group 6b′) are tilted approximately the same amount by the Basalt Mountain monocline as are nearby ca. 10 Ma group 4b basaltic rocks. Therefore most of the collapse of the Basalt Mountain monocline occurred after ca. 3 Ma (Kirkham et al., 2001b). Major Pliocene and younger collapse can also be inferred from data in the northeastern part of the collapse area. Preserved remnants of the 3.17 ± 0.02 Ma eruptive center at Buck Point (group 6b″) are at an altitude of ~2680 m (Streufert et al., 1997), and the 3.97 ± 0.08 Ma eruptive center at Little Buck Point (group 6b) is at 2620 m (Kirkham et al., 1995). Both of these Pliocene eruptive vents are within the collapse

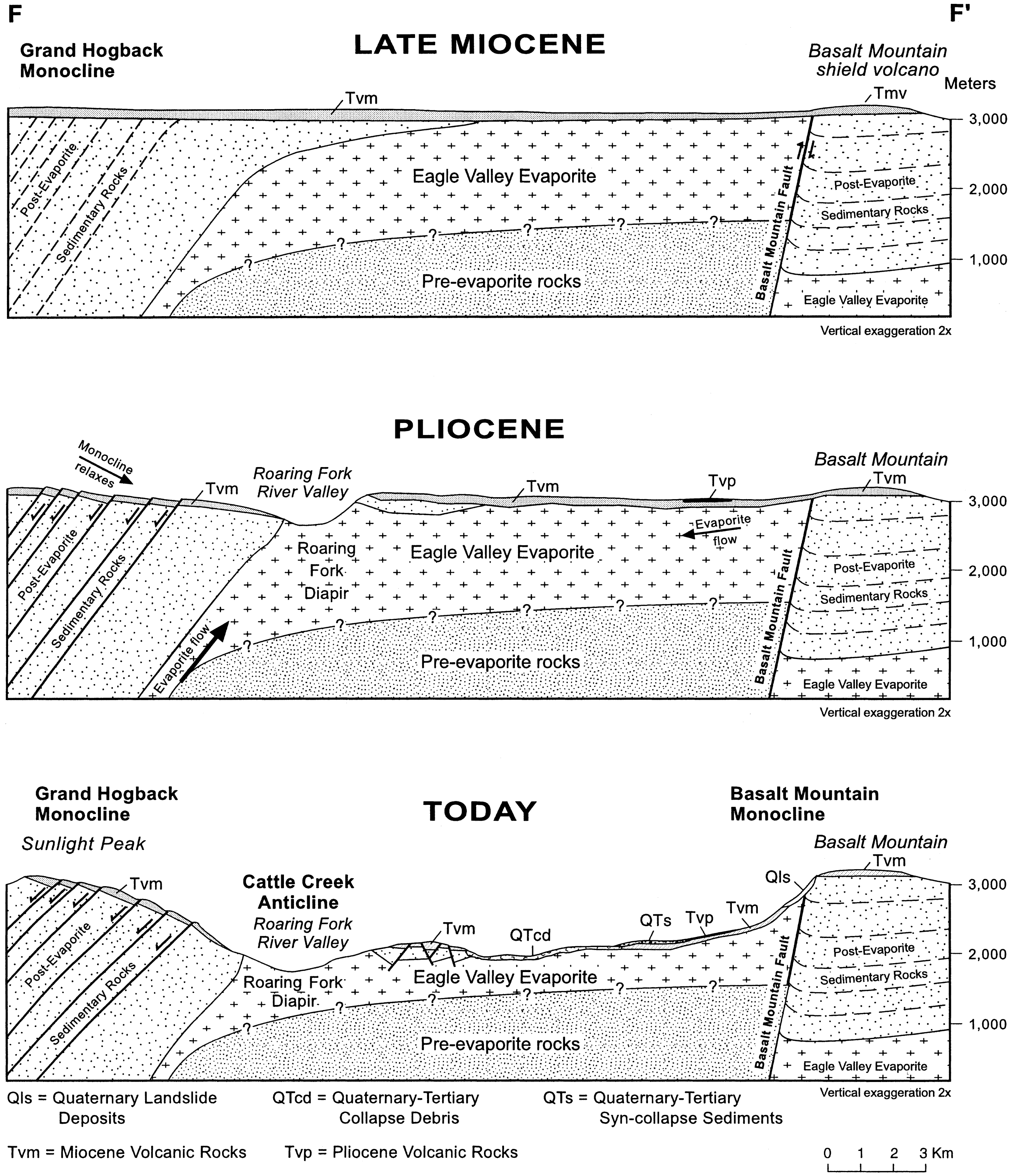

Figure 22. Schematic east-west cross sections (F–F′) showing the generalized evolution of the Carbondale collapse center in the lower Roaring Fork River valley (adapted from Kirkham et al., 2001b). Approximate line of section (shown by dashed line in Figure 7) follows an irregular path from Sunlight Peak to Basalt Mountain.

center, yet they are only slightly lower than or equal in elevation to the 7.75 Ma flows on Little Grand Mesa, which are at an altitude of 2680–2745 m and are outside of the Carbondale collapse center. If significant collapse occurred between 7.75 and ca. 3 Ma, then these younger eruptive vents would have been constructed on a ground surface that was lower in the landscape than Little Grand Mesa. These data suggest that much of the evaporite collapse in the lower Roaring Fork River valley occurred during the past 3 million years, which closely mirrors incision rates (Kirkham et al., 2001a; Kunk et al., this volume). Historic sinkholes and high salinity loads in rivers and thermal springs document active dissolution and deformation. The presence of thick evaporite beneath much of the collapse area portends continuing collapse. Future collapse rates likely will fluctuate as climate and rates of incision and uplift vary.

SUMMARY

Dissolution and flow of Pennsylvanian evaporitic rocks during the late Cenozoic have created a regional collapse area in the lower Roaring Fork River valley in west-central Colorado. Abundant and well-preserved evidence of this evaporite tectonism is widespread in the collapse area. The lateral extent and amount of vertical collapse, as well as the timing and style of deformation, are well constrained by upper Cenozoic volcanic rocks that were correlated using field mapping, $^{40}Ar/^{39}Ar$ geochronology, geochemistry, and paleomagnetism. These volcanic rocks are downdropped as much as 1220 m in the collapse area. Syn-collapse sedimentary deposits accumulated in structurally controlled topographic depressions along the margins of the collapse center and within its interior.

Local collapse initiated during the Oligocene or early Miocene. Collapse became widespread between ca. 10 Ma and 7.7 Ma, and it greatly accelerated during the past 3 million years, largely in response to river incision. When the Roaring Fork River began to downcut through a broad, low-relief erosion surface during the late Miocene, rocks overlying the evaporite beds were eroded from the valley. This created differential lithostatic pressures, which triggered flow of the evaporite from beneath adjacent upland areas where pressures remained high, toward the Roaring Fork River valley where the pressures were reduced. River incision also allowed relatively fresh groundwater to circulate to progressively greater depths, a process that increased dissolution rates. Since thick evaporite still underlies much of the collapse area, continued subsidence is likely.

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