GEOLOGIC MAP OF THE ST. GEORGE AND EAST PART OF THE CLOVER MOUNTAINS 30' x 60' QUADRANGLES, WASHINGTON AND IRON COUNTIES, UTAH

by

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GEOLOGIC MAP OF THE ST. GEORGE AND EAST PART OF THE CLOVER MOUNTAINS 30' x 60' QUADRANGLES, WASHINGTON AND IRON COUNTIES, UTAH

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ABSTRACT

This map is derived from 29 published 1:24,000-scale geologic maps that encompass the St. George 30' x 60' quadrangle and Utah part of the Clover Mountains 30' x 60' quadrangle. It also incorporates significant new unpublished field mapping of the Smith Mesa quadrangle and of six quadrangles in the Bull Valley and Pine Valley Mountains, late Tertiary and Quaternary deposits of the Mesquite basin, and the Pintura area. The resulting compilation shows the regional geology of the southwest corner of Utah in unprecedented detail.

Southwest Utah straddles the transition zone between the Colorado Plateau and Basin and Range Province and so includes a remarkable diversity of rocks, landforms, and geologic structures. Sedimentary and volcanic rocks that collectively total over 8 miles (12 km) in thickness are exposed in the map area, representing every geologic period except the Ordovician and Silurian. The oldest rocks exposed in the map area, 1.7-billion-year-old metamorphic and igneous rocks, form the core of the enigmatic Beaver Dam Mountains. Triassic and Jurassic "redrock," famously exposed at Zion National Park and Snow Canyon and Quail Creek State Parks, are among the most widespread and visible strata, for they also surround major urban areas in southwest Utah. The past decade has witnessed significant new discoveries of dinosaur tracks and fossils in these local strata, shedding new light on this part of North America's geologic history. These and other rocks, including the enormous Pine Valley laccolith and dozens of basaltic lava flows and cinder cones, are exceptionally well exposed because of widespread erosion that began when the Colorado Plateau and Basin and Range began to split apart about 17 million years ago.

Major geologic structures in the map area include thrust faults and folds associated with the Late Cretaceous to early Tertiary Sevier orogeny (including the Square Top Mountain and Taylor Creek thrust faults and Kanarra, Pintura, and Virgin anticlines), large normal faults that break apart the western margin of the Colorado Plateau (including the Hurricane, Washington, and St. George faults), mostly east-trending oblique-slip faults and rotated blocks in the Bull Valley Mountains (part of the Caliente-Enterprise zone, a broad, left-lateral transfer zone that was accompanied by significant north-south shortening), and perplexing faults and folds at the east margin of the Basin and Range Province (including the Beaver Dam Mountains culmination, Shivwits syncline, Jackson Wash-Pahcoon Flat fault zone, and the Gunlock-Reef Reservoir-Grand Wash fault zone). Though well exposed, these latter faults and folds have engendered conflicting interpretations of their age and origin. We suggest that they were not produced during the Sevier orogeny, but are best explained as a result of late Tertiary and Quaternary displacement on left-lateral oblique-slip faults that are consistent with a regional pattern of significant left-lateral strike- and oblique-slip faulting at the east margin of the Basin and Range, in combination with footwall uplift associated with the Piedmont–Red Hollow range-front fault zone that created the main Beaver Dam culmination.

This map displays relationships among nearly 50 individually mapped lava flows of late Pliocene to late Quaternary age, many of which form classic inverted valleys and several of which cross and are offset by faults associated with Basin and Range extension. These primarily basaltic lava flows document long-term incision rates across the transition zone between the Colorado Plateau and Basin and Range Province, which increase from west to east from about 0.2 feet per thousand years (0.06 m/kyr) around St. George, to 0.35 feet per thousand years (0.11 m/kyr) at Hurricane, to about 1.15 feet per thousand years (0.35 m/kyr) east of the Hurricane fault. Lava flows that cross and are offset by the Hurricane fault zone dramatically demonstrate that this is the most active fault in southwest Utah.

The main geologic resources in the area include sand and gravel, decorative stone, gypsum, and water, and silver, gold, uranium, rare earth metals, and oil were produced in the past. Flooding, expansive soil and rock, landslides, surface faulting, liquefaction, rock fall, collapsible soil and rock, and gypsiferous soil and rock are among the most common geologic hazards in the map area.

INTRODUCTION

Southwest Utah straddles the transition zone between the colorful, flat-lying strata of the Colorado Plateau to the east and the highly faulted Basin and Range Province to the west (figure 1). The transition zone melds parts of each so that southwest Utah contains a remarkable diversity of rocks, structures, and landforms—essentially doubling the geologic diversity in the region. Strata from every geologic period except the Ordovician and Silurian are present in the map area, and Paleoproterozoic (1.7-billion-year-old) metamorphic and igneous rocks are exposed in the core of the enigmatic Beaver Dam Mountains. Collectively, these strata total over 8 miles (12 km) in thickness and record the geologic evolution of southwest Utah.

Southwest Utah also straddles parts of four ecoregions (large areas having geographically distinct plant and animal communities): Mojave Basin and Range, Central Basin and Range, Colorado Plateaus, and Wasatch and Uinta Mountains (figure 2). Most people in the county live in the Virgin River lowlands, part of the dry, sparsely vegetated Mojave Desert. The community of St. George is at an elevation of about 2700 feet (800 m), where the winters are mild and the summers hot, but at just over 10,000 feet (3050 m) in elevation, the Pine Valley Mountains, a dozen miles (20 km) to the north, tower over southwest Utah's urban areas (figure 3). Their high elevation, as well as that of the Kolob Plateau in the northeast part of the quadrangle, traps moisture and ultimately provides an important part of the region's water supply; their cool ponderosa pine forests offer a welcome respite from the warm Virgin River lowlands below. Southwest Utah thus encompasses many climatic zones, and the region's diverse geology is complemented by a diverse flora and fauna. With its characteristically sparse vegetation highlighting but not concealing a diverse geological terrain, the greater St. George area is an ideal outdoor geological and biological classroom.

In this introductory text we discuss the most widely exposed rocks in the greater St. George–Zion National Park area, especially as they relate to changing depositional environments. We also provide a summary of the major geologic structures in the map area, including extraordinarily complicated faults and folds at the east margin of the Basin and Range, whose interpretation has puzzled generations of geologists. Finally, we discuss the evolution of the modern landscape and touch on the geologic resources and geologic hazards of the area. More information about the rocks and structures of the map area can be found in the map unit descriptions and references therein (in the appendix), in published map unit descriptions of the individual 7.5' quadrangle maps from which this map was compiled (figure 4), and in a number of popular accounts of the geology of southwest Utah, including those of Hintze (1993, 2005), Hamilton (1995), Biek (1999), Biek and others (2003, 2004, 2009), Higgins (2003), Hamblin (2004), Hayden (2004c), Eves (2005), Biek and Rohrer (2006), Orndorff and others (2006), Biek and Hayden (2007), and Hintze and Kowallis (2009). Colorful paleogeographic maps and accompanying text of the recently published *Ancient Landscapes of the Colorado Plateau* (Blakey and Ranney, 2008) give readers an excellent summary of the rock formations of the Colorado Plateau and the changing depositional environments they record.

At its simplest, the story in the great stack of rock layers in southwest Utah can be summarized as follows: (1) The Proterozoic Eon, known in southwest Utah for metamorphic and igneous rocks exposed in the Beaver Dam Mountains, marks a time of collision and suturing of island-arc terranes to the supercontinent Rodinia. This was followed by the breakup of Rodinia, which created a broad zone of attenuated continental crust that reached inland as far as central Utah (the north-trending boundary between this thinned crust and the thicker crust to the east is known as the Utah hingeline, which has been a dominant geologic feature throughout geologic time). (2) The Paleozoic Era was dominated by shallow-marine sedimentation on the cratonic margin. (3) The Mesozoic Era saw the development of broad, low-elevation interior basins drained by northwest-flowing river systems, which were subsequently overwhelmed by great sand deserts, then in turn by encroachment of the Sevier orogenic belt. (4) The Cenozoic Era opened with regional uplift and development of early Tertiary intermontane basins, followed by explosive calc-alkaline volcanism in the middle Tertiary, and finally, in the late Tertiary and Quaternary, structural and topographic differentiation of the Colorado Plateau and Basin and Range physiographic provinces and creation of the modern landscape.

Today, erosion is the dominant geologic process in southwest Utah. Since the inception of modern basin-range extension beginning in late Tertiary time about 17 million years ago, erosion has stripped away a tremendous thickness of rock, in effect providing a geologic cross section across southwest Utah revealing the effects of more than 1.7 billion years of Earth history. The timing and history of erosional exhumation is recorded by fault relationships, the stratigraphy of associated basin-fill deposits, and Pliocene and Quaternary basaltic lava flows and their spectacular examples of inverted topography. Erosion has also revealed a wide variety of compressional and extensional geologic structures, whose age is summarized on the stratigraphic column on plate 2, including (1) dramatic frontal folds and thrust faults of the Sevier orogenic belt, (2) large normal faults that tear apart the western margin of the Colorado Plateau, (3) left-lateral oblique-slip faults and associated structures that define the east edge of the Basin and Range Province, (4) low-angle normal faults and gravity-slide blocks in the Beaver Dam Mountains, (5) enormous gravity-slide structures associated with intrusion of the Pine Valley and other laccoliths, and (6) an east-trending transverse zone in the Bull Valley Mountains. Considerable disagreement exists between interpretations of some of these structures, especially those in the west part of the map area. In the structure section of this report, we outline the major structural elements of the map area and reveal the profound effect of late Tertiary extensional overprinting on the leading edge of the Late Cretaceous-early Tertiary fold and thrust belt.

REGIONAL GEOLOGIC SETTING

Parts of four north-trending, regional-scale features dominate the geology of southwest Utah (figure 5): (1) the leading edge of the Sevier orogenic belt, (2) the transition zone between the

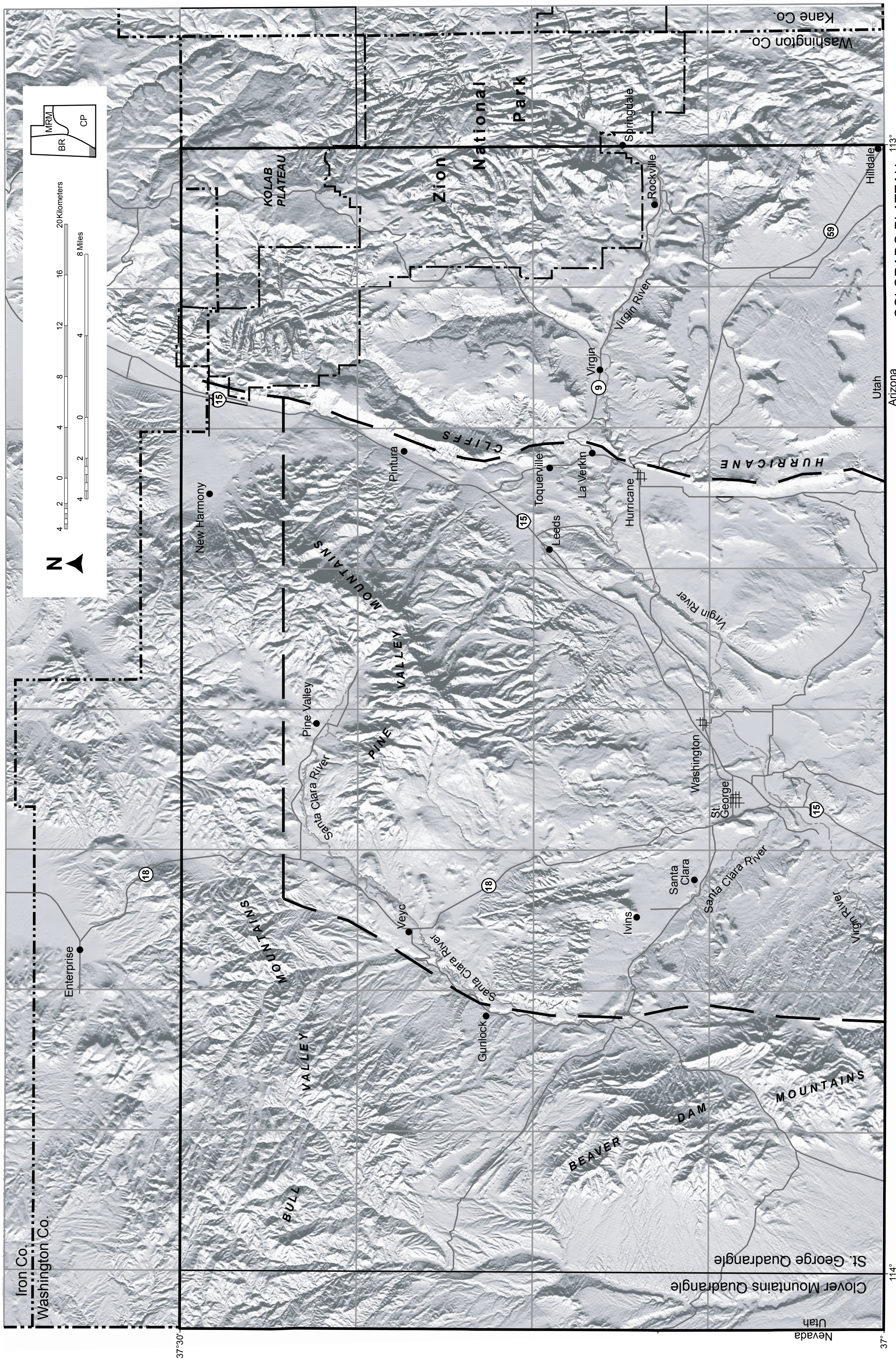

Figure 1. *Shaded-relief image of Washington County showing major fault zones that form the boundaries of the Colorado Plateau and Basin and Range Province, and the transition zone between these two physiographic provinces. Heavy line marks boundary of the map area (St. George 30' x 60' quadrangle and the east part of the Clover Mountains 30' x 60' quadrangle); grid shows boundaries of 7.5' quadrangle maps. Compare with illustration on plate 2 and figure 4, which show major geologic features and names of 7.5' quadrangles. Inset shows major physiographic provinces of Utah: BR, Basin and Range; CP, Colorado Plateau; MRM, Middle Rocky Mountains.*

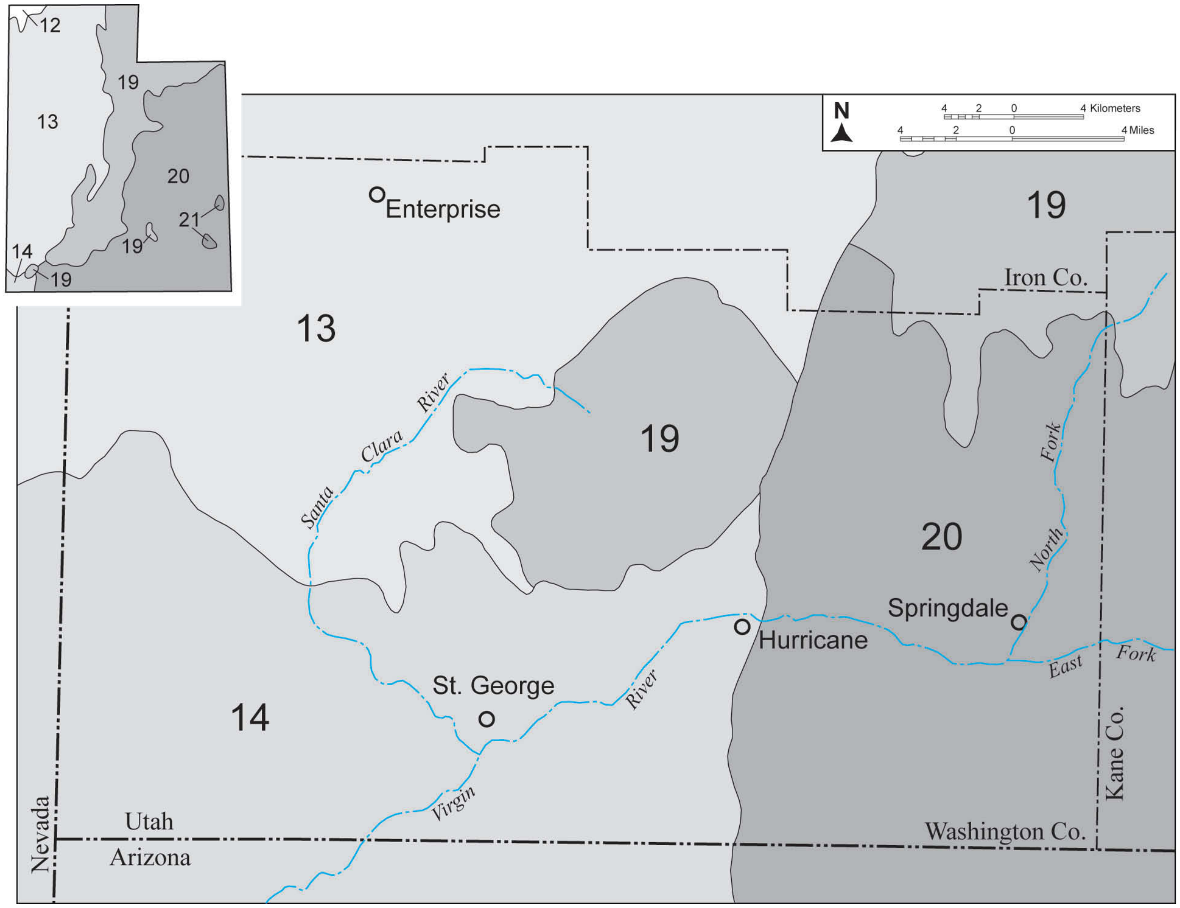

Figure 2. *Large areas having similar physiographic and climatic characteristics, and thus geographically distinct plant and animal communities, are known as ecoregions or bioregions. In southwest Utah, ecoregions (numbered on the map and inset) include the Central Basin and Range (13), Mojave Basin and Range (14), Colorado Plateaus (20), and Wasatch and Uinta Mountains (19), reflecting the great diversity of plant and animal communities found here. Modified from Ecoregions map of Utah, available at http://education.usgs.gov/common/resources/mapcatalog/images/environmental/ECO_UTAH1_11x14.pdf.*

Figure 3. *View north to the Pine Valley Mountains from Washington Black Ridge; Interstate 15 cuts through the ridge at the center of the photo. The Pine Valley Mountains are the eroded remnants of one of the world's largest laccoliths, a mushroom-shaped igneous intrusion that was emplaced 20.5 million years ago. The cinder cone at the left margin of the photo is the source of the 900,000-year-old Washington basaltic lava flow, which now forms a classic inverted valley capping Washington Black Ridge. The lava flowed 5.5 miles (8.8 km) down the ancestral Grapevine Wash to the Virgin River.*

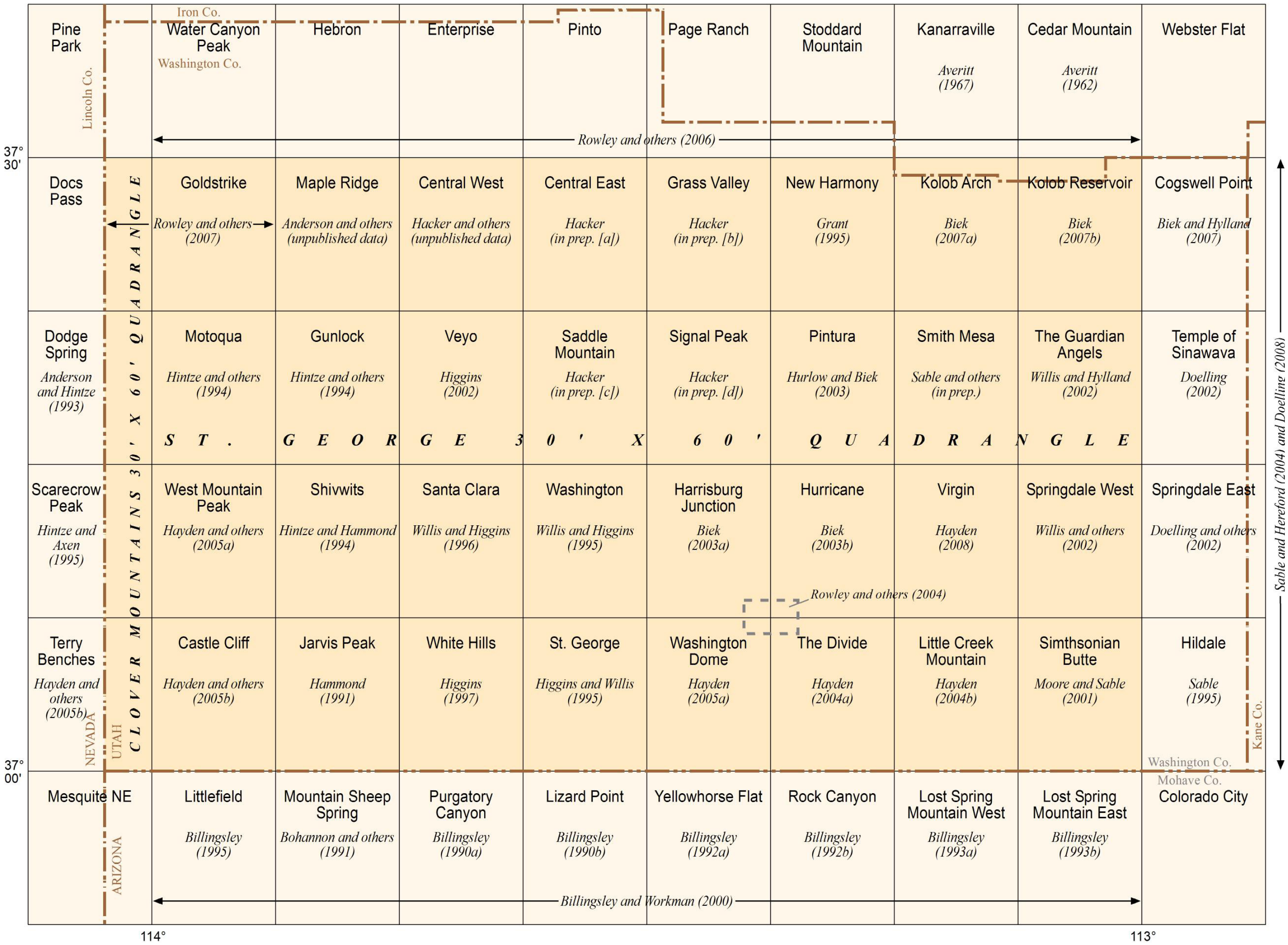

***Figure 4.** Index to U.S. Geological Survey 7.5' quadrangles and principal sources of geologic mapping used in this compilation.*

Colorado Plateau and Basin and Range physiographic provinces, (3) the southern part of the Intermountain seismic belt, and (4) the major controlling feature, the north-trending Utah hingeline, which has influenced Utah geology for more than 600 million years. The Utah hingeline originated as a long-lived boundary between a stable continental shelf to the east and a subsiding marine basin to the west, and is manifest today as the boundary between the Basin and Range and Colorado Plateau–Rocky Mountains Provinces. Collectively, these four regional-scale features overprint Paleoproterzoic metamorphic and igneous rocks exposed in the Beaver Dam Mountains—and the vast thickness of overlying sedimentary and volcanic rocks in the remainder of the map area—and help us to understand the distribution and structure of the rocks that make this corner of the state so interesting.

The Sevier orogeny (part of the larger Cordilleran orogeny, which produced a zone of deformation that stretched from Mexico to Alaska along the western margin of North America) was one of the most important events in Utah's geologic history. The orogeny, or mountain-building event, was a direct result of the subduction of the Farallon tectonic plate (the Farallon plate was part of an earlier Pacific Ocean basin) beneath the North American plate. In Utah, deformation associated with this collision is expressed as long folds and thrust faults that formed during the Cretaceous to Eocene, about 100 to 45 million years ago (see, for example, Willis, 1999, 2000). In southwest Utah, such deformation is dramatically displayed at Square Top Mountain, where a thrust fault by that name places Pennsylvanian and Permian rocks on top of Jurassic and Cretaceous rocks. Late Cretaceous sediments shed from the southern Utah part of the Sevier thrust belt are preserved on the southern edge of the Bull Valley Mountains, in the Pine Valley Mountains, and in the Markagunt Plateau, but late Tertiary erosion stripped away much of this story in the greater St. George area. Still, two of the most prominent geologic features visible today in southwest Utah mark the leading edge of the Sevier compressional deformation—the Virgin and Kanarra anticlines (figure 6).

The area of what are now the Colorado Plateau and Basin and Range physiographic provinces was uplifted several thousand feet during early Tertiary time. But by about 17 million years ago, western Utah began to "collapse" away from eastern Utah due to changing motions of the Earth's tectonic plates, thus dividing the Basin and Range from the Colorado Plateau (Rowley and others, 1978, 1979; Rowley and Dixon, 2001). Today, the Earth's crust is being pulled apart east-to-west across the Basin and Range. Compared to the Colorado Plateau—a wide area underlain by mostly gently warped sedimentary rocks and only locally disrupted by broad uplifts and igneous intrusions—the Basin and Range crust is thinner, is broken into north-trending fault blocks, and has experienced widespread igneous activity.

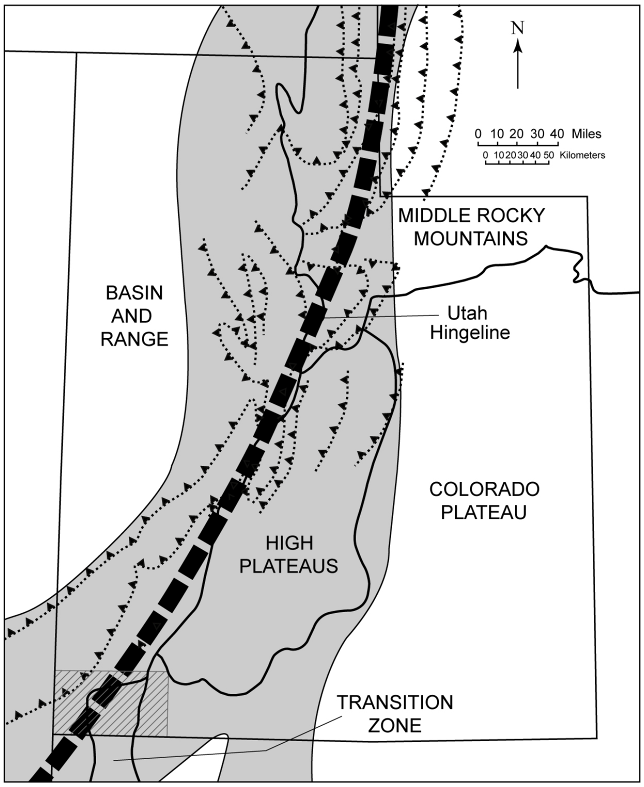

Figure 5. *Regional-scale tectonic features of Utah. The Utah hingeline (also known as the Wasatch hingeline), which originated as a long-lived boundary between a stable continental shelf and a subsiding marine basin, now controls the location of three other regional-scale features: the Sevier orogenic belt, the boundary between the Basin and Range and Colorado Plateau (the High Plateaus are a subprovince of the Colorado Plateau), and the Intermountain seismic belt (shaded). Major thrust faults of the Sevier orogenic belt are shown as dotted lines with barbed teeth on upper-plate rocks. Hatchured area shows location of map area.*

In southwest Utah, the western margin of the Colorado Plateau is broken by several north-trending faults that step down from the Colorado Plateau to the Basin and Range, creating a transition zone between the two physiographic provinces. These faults comprise the southern part of the Intermountain seismic belt, a north-trending zone of pronounced seismicity that extends from northern Arizona to western Montana (Smith and Arabasz, 1991). In southwest Utah, the transition zone is bounded by the Hurricane fault zone on the east and the Gunlock–Reef Reservoir–Grand Wash fault zone on the west. These major faults not only form the boundaries of the transition zone, but also, in conjunction with other faults of the eastern Basin and Range, greatly influenced the late Tertiary-Quaternary landscape evolution of southwest Utah.

STRATIGRAPHY

Southwesternmost Utah contains exposed rocks as old as Paleoproterozoic in the Beaver Dam Mountains, as young as Late Cretaceous on the Kolob Plateau above Zion National Park, and younger still in the Pine Valley and Bull Valley Mountains where Late Cretaceous and early Tertiary sedimentary strata and Oligocene and Miocene volcanic and intrusive rocks are present. But by far the most colorful and widespread strata are those of Triassic and Jurassic age that are spectaculary exposed in the heart of the St. George and Zion areas. What follows is an overview of these rocks, emphasizing Triassic and Jurassic strata. Additional information on these and other rocks in the mapped area are in the map unit descriptions (see appendix).

Proterozoic

(2500 to 542 million years ago)

The oldest rocks exposed in southwest Utah are metamorphic and igneous rocks of the Beaver Dam Mountains (figure 7). These rocks form what geologists call the Precambrian crystalline basement, literally the foundation upon which continents are built. Basement rocks in Utah are typically deeply buried by a vast thickness of sedimentary and volcanic rocks and so are seldom exposed at the Earth's surface. In fact, seeing basement rocks at the surface is indicative of significant crustal uplift and erosion.

Gneiss—a foliated, granite-like metamorphic rock where light-colored (principally feldspar and quartz) and dark-colored (mostly biotite and hornblende) minerals are segregated into distinct bands—is the most common basement rock in the Beaver Dam Mountains. Garnet, indicative of high temperatures and pressures of metamorphism, is common in the gneiss. Schist, a micaceous metamorphic rock, and amphibolite, a metamorphic rock composed dominantly of dark-colored amphibole minerals such as hornblende, are also common in the Beaver Dam Mountains. The mineralogy of these metamorphic rocks shows that they experienced pressure-temperature conditions in the range of 0.7 to 0.8 GPa (billion pascals) and 1300 to 1450°F (700 to 800°C), indicative of mid-crustal levels, perhaps 10 to 15 miles (16–24 km) below the Earth's surface; the rocks also show evidence of an earlier, higher pressure phase of metamorphism, suggesting that the rocks once occupied a lower crustal position (Colberg, 2007; Fitzgerald and Colberg, 2008). Igneous rocks of mostly granitic composition are also present in the Beaver Dam Mountains. Pegmatite, a coarse-grained igneous rock composed of large feldspar and quartz crystals with minor muscovite, is common as dikes and sills that cut across other basement rocks in the range.

The Precambrian crystalline basement of the Beaver Dam Mountains is about 1730 million years old (Paleoproterozoic) (Nelson and others, 2007). The rocks are part of the Mojave Province, and represent part of the initial assembly of the Rodinian supercontinent. Rodinia formed between about 1.7 and 1.0 billion years ago when the Mojave, Yavapai, and other provinces collided with and were sutured to the Wyoming Province, probably along an island-arc

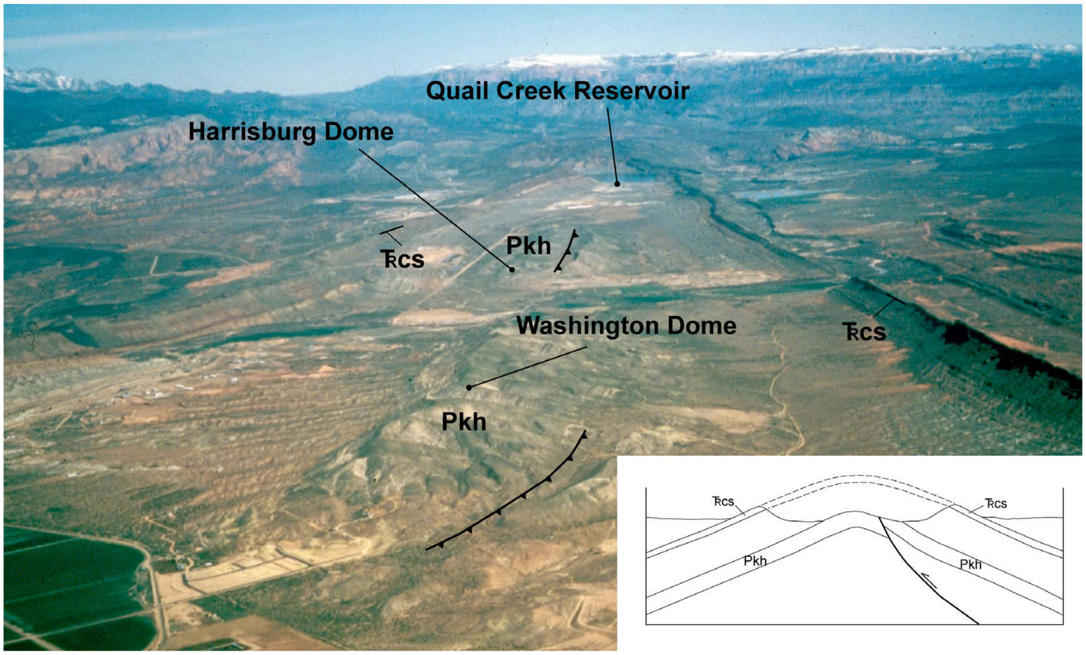

Figure 6. *Oblique aerial view to the northeast of the Virgin anticline, a spectacular frontal fold of the Sevier orogenic belt. The Virgin River cuts across the anticline in the center of the photo. Quail Creek Reservoir is nestled in the north end of the fold and is used for offline storage of Virgin River water. Washington Dome is south of the river, Harrisburg Dome is to the north; both domes expose the Harrisburg Member of the Kaibab Formation (Pkh), and each is cut by a small axial reverse fault. Note the resistant hogbacks of the Shinarump Conglomerate Member of the Chinle Formation (TRcs) along the flanks of the anticline. Schematic cross section shows how these strata are folded into this upwarp. See figure 41 for another view of Harrisburg Dome.*

subduction zone (Nelson and others, 2002; Karlstrom and others, 2005); it included Laurentia (North America), Australia, Antarctica, Greenland, and Baltica (northern Europe) (figure 8).

Figure 7. *Rugged western escarpment of the Beaver Dam Mountains, which are eroded from 1.7 billion-year-old Precambrian crystalline basement. These are the oldest rocks exposed in southwest Utah—they formed at great depths (perhaps 10 to 15 miles [16–24 km]) in the Earth's crust as island-arc terranes collided with and sutured to the North American craton during initial assembly of the Rodinian supercontinent. The fact that they are exposed at the surface today—and that thousands of feet of overlying Paleozoic, Mesozoic, and Cenozoic strata are eroded away—is dramatic evidence of the tremendous amount of uplift and erosion that has occurred at the western margin of the transition zone between the Basin and Range and Colorado Plateau during the past 20 million years. West Mountain Peak, 7680 feet (2341 m) above sea level, forms the highest point of the Beaver Dam Mountains.*

The 1730-million-year-old basement rocks of the Beaver Dam Mountains are unconformably overlain by 540-million-year-old Tapeats Sandstone. This gap in the rock record spans about 1.2 billion years and is known in the Grand Canyon region as the "Great Unconformity." However, in the Grand Canyon region, basement rocks are overlain by the Meso- to Neoproterozoic rocks of the Grand Canyon Supergroup, which consists of more than 12,000 feet (3650 m) of sandstone, siltstone, and shale preserved in tilted fault blocks. As described and illustrated in Blakey and Ranney (2008), the Grand Canyon Supergroup may have accumulated in a basin formed during final assembly of Rodinia, and then was preserved in tilted fault blocks during breakup of the supercontinent. Equivalent strata may have been deposited in southwest Utah, but if so, they were entirely removed by erosion that produced the Great Unconformity.

Early and Middle Paleozoic
(Cambrian to Pennsylvanian, 542 to 299 million years ago)

In late Neoproterozoic time, about 725 million years ago, Rodinia began to break up or rift apart. North America split away from Australia and Antarctica, creating the proto-Pacific Ocean, leaving the western edge of North America, then in central Nevada, as a passive margin (passive margins form on the trailing edge of Earth's tectonic plates) (figure 9). During all of Paleozoic time, Utah lay near the equator along a coastline probably geologically similar to the eastern United States today. The rifting created a broad north-trending zone of attenuated continental crust that reached inland as far as central Utah. The north-trending boundary between weaker, thinned crust to the west and thicker, more stable crust to the east is known as the Utah hingeline, which has affected sedimentation and structure at and near the western margin of North America since its formation.

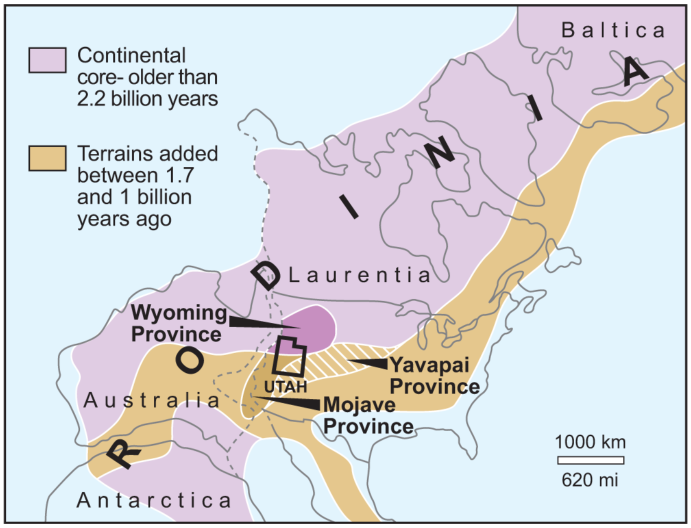

Figure 8. The Rodinian supercontinent formed between 1.7 and 1.0 billion years ago as island-arc terranes were accreted to the southern margin of older, Archean cratonic rocks. Rodinia included Laurentia (North America), Australia, Antarctica, Greenland, and Baltica (northern Europe). About 725 million years ago, Rodinia began to break apart when Australia and Antarctica rifted away from Laurentia. The north-trending rift, shown by dashed lines, ran through central Nevada and created a broad zone of attenuated continental crust that reached inland as far as central Utah. The north-trending boundary between this thinned crust and the thicker crust to the east is known as the Utah hingeline, which has affected sedimentation and structure at and near the western margin of North America ever since. Modified from Karlstrom and others (1999).

Subsidence along the passive western margin of North America created accommodation space in which to preserve a great thickness of shallow-marine strata. West of the Utah hingeline, where basin subsidence was greatest, strata are especially thick; to the east of the hingeline (which includes the entire Colorado Plateau) correlative strata are much thinner since they were deposited on the more stable continental shelf. In the Beaver Dam Mountains of southwest Utah, which straddle the hingeline, sedimentary strata of Cambrian through Pennsylvanian age total over 8000 feet (2400 m) thick; equivalent strata are much thicker to the west in Nevada and west-central Utah and are correspondingly much thinner on the Colorado Plateau (see, for example, Hintze, 1986; Hintze and Kowallis, 2009). One interesting but confusing complication about such thickness (and lithologic) variations is that correlative strata tend to be known by different names on either side of the Utah hingeline. For example, the Cambrian Bonanza King Formation is at least 2800 feet (800 m) thick in the Beaver Dam Mountains (and thicker still in nearby Nevada); equivalent strata on the southwestern Colorado Plateau are known as the Mauv Limestone, and are only 650 feet (200 m) thick in the central Grand Canyon area.

In Early Cambrian time, western North America was partly inundated by the Cambrian sea, which transgressed (flooded) the continent from west to east (figure 10). In southwest Utah, resistant, silica-cemented sandstone (orthoquartzite) and minor quartz-pebble conglomerate, known as the Tapeats Sandstone, represent a coarse-grained, nearshore and intertidal facies of the Cambrian sea that transgressed over eroded crystalline basement rocks. Little relief is present on the underlying unconformity, showing that by Early Cambrian time in this area, Precambrian crystalline basement rocks had been eroded to a low-relief coastal plain. That plain, however, featured only bare rock and shallow, sediment-filled valleys barren of life. The “Cambrian Explosion,” which ushered in the astounding diversity of complex multicelluar organisms (and that actually began near the end of the Neoproterozoic during the Ediacaran Period, about 625 million years ago), filled the seas with life, but terrestrial plants and animals had not yet evolved. The landscape would have appeared as barren as that on Mars or the ice-free valleys of Antarctica.

Thus, as the sea transgressed eastward, coarse-grained sediment of the Tapeats Sandstone was deposited near shore. Finer-grained clay and silt accumulated offshore (represented by the Bright Angel Shale), and limestone with algal stromatolites, brachiopods, and other fossils (represented by the Bonanza King Formation and Nopah Dolomite) was deposited in warm, clear, shallow water far from the land’s edge. This is a classic example of a marine transgression—sand overlain by mud overlain by limestone—and shows that these rock units, like so many others, are time transgressive. The Tapeats Sandstone is older in western exposures and becomes progressively younger to the east, and the same is true of other transgressive strata deposited on the passive western margin of the continent.

No rocks of Ordovician and Silurian age (488 to 416 million years ago) are present in southwesternmost Utah, nor in all of central and eastern Utah, due to widespread erosion in the Early Devonian, likely associated with the Antler orogeny in central Nevada. (The Antler orogeny was the first mountain-building event to affect the western margin of North America following the breakup of Rodinia, and occurred when island-arc terranes collided with the continent in what is now western and central Nevada.) About 1600 feet (500 m) of strata of this age is present, however, in the upper plate of the Tule Springs thrust in the adjacent Scarecrow Peak quadrangle, just 2 to 4 miles (3–6 km) west of the Utah-Nevada state line (Hintze and Axen, 1995). During the Early Ordovician, plants and animals began to colonize the land.

Devonian time in southwest Utah is represented by sandy and cherty dolomite of the Muddy Creek Dolomite (equivalent to the Temple Butte Limestone of the Virgin Mountains and Grand Canyon areas). Scattered coral, crinoid, gastropod, and brachiopod fossils show that this package of rocks was deposited in a warm, shallow sea. The uppermost part of the Muddy Creek Dolomite typically weathers to pinnacles or hoodoos and so contrasts strongly with the overlying massive limestone cliff represented by the Mississippian-age Redwall Limestone.

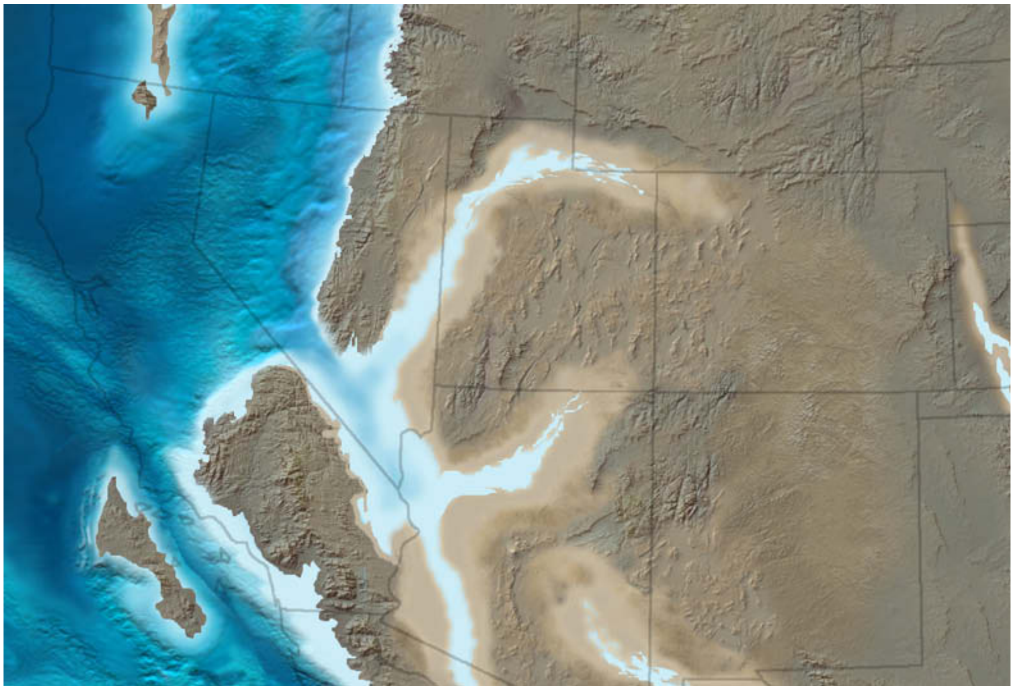

Figure 9. *Schematic reconstruction of Late Neoproterozoic paleogeography of the southwest U.S. (about 650 million years ago). Note the active rift that separates North America from Gondwanaland on the south, marking the breakup of the Rodinian supercontinent. Compare with figure 10 to see the location of the Utah hingeline. Over the following 100 million years, the western margin of North America was eroded down to a low-relief coastal plain, setting the stage for accumulation of great thickness of shallow-marine Paleozoic strata. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

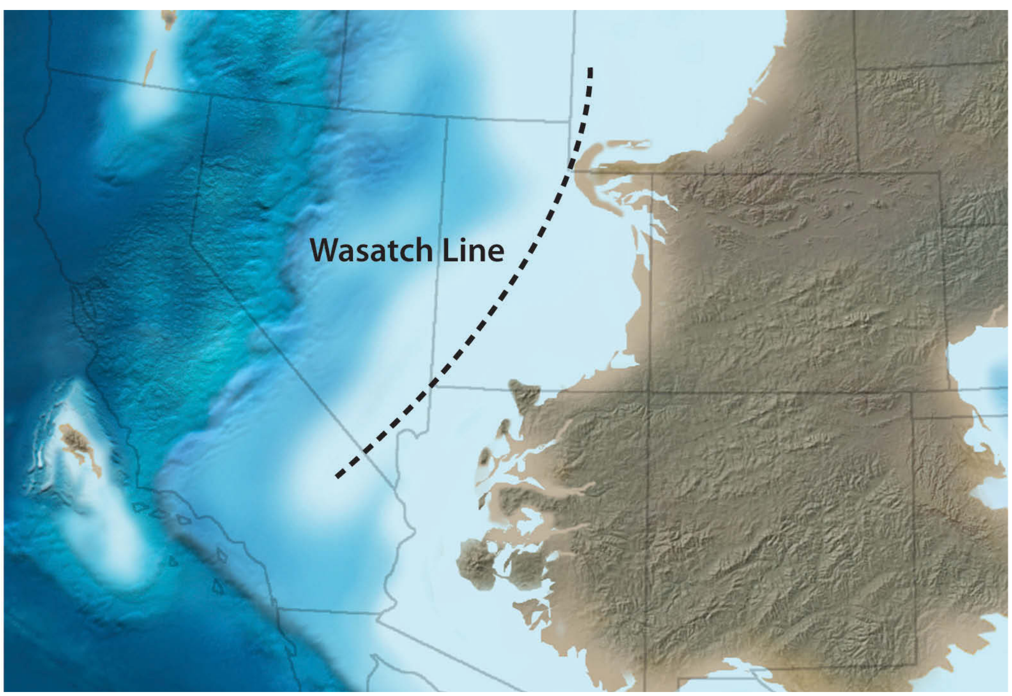

Figure 10. *Schematic reconstruction of Middle Cambrian paleogeography of the southwest U.S. (about 510 million years ago). As the shallow Cambrian sea flooded the continent from west to east, a classic transgressive sequence of sandstone (Tapeats Sandstone), overlain by mudstone (Bright Angel Shale), overlain by limestone (Bonanza King Formation) developed. The Utah hingeline (also known as the Cordilleran or Wasatch line) separates thicker strata to the west (that accumulated in a subsiding basin) from equivalent but thinner strata to the east (that accumulated on a more stable shelf platform). Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

Pioneering American geologist Grove Karl (G.K.) Gilbert named the Redwall Limestone for one of the most recognizable rock units in the Grand Canyon. The Redwall forms the highest single cliff in the canyon, a 500-foot-tall (150 m) massive limestone cliff stained red by mud eroded from the overlying Supai Group rocks. The Redwall Limestone forms a similarly imposing cliff in the Beaver Dam Mountains, but one that is more fractured and ledgy due to the range's position in the highly faulted transition zone (figure 11). Because the red shale and sandstone of the Supai Group is not present in the Beaver Dam Mountains, the true gray color of the Redwall Limestone—unstained by those red rocks—is apparent in southwest Utah. A wide variety of brachiopod, coral, echinoderm, and bryozoan fossils in Mississippian-age limestone beds throughout western North America shows that this was a time of widespread carbonate deposition in warm, shallow, tropical seas. In the map area, limestone deposition took place in a foreland basin associated with the Antler orogeny of central Nevada (figure 12).

Pennsylvanian time witnessed the collision of two great continents, Laurasia (composed of North America, Europe, and Asia) and Gondwanaland (most of the modern southern continents), which created the supercontinent Pangea. The colliding land masses also created the ancestral Appalachian and Rocky Mountains and associated basins, and marked the end of passive-margin deposition at the western edge of North America (figure 13). In southwesternmost Utah, shallow-marine deposition continued with the limestone and cyclically interbedded sandstone and siltstone of the Callville Limestone, although areas to the east experienced more varied terrestrial and marine depositional environments due to proximity to the ancestral Rocky Mountains and adjacent basins. Cyclically bedded strata is common in Pennsylvanian-age sedimentary rocks and resulted from worldwide sea-level fluctuations due to widespread glaciation in the southern hemisphere. During glacial maximums, so much fresh water was locked up in glacial ice that sea level fell several hundred feet (and correspondingly rose several hundred feet when the ice melted), creating constantly changing depositional environments.

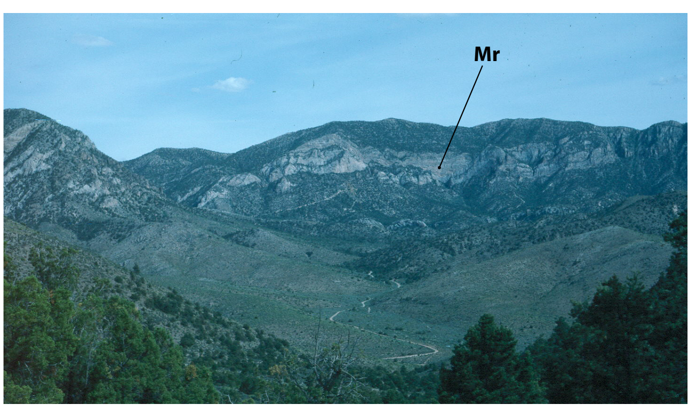

Figure 11. *The Redwall Limestone (Mr) is easily recognizable because it forms an imposing cliff in the Beaver Dam Mountains. This view is to the south across Castle Cliff Wash from just west of the Utah Hill area; the Bright Angel Shale forms a strike valley traversed by the dirt road.*

Permian
(299 to 251 million years ago)

Permian strata in the Virgin River Gorge, Beaver Dam and southern Bull Valley Mountains, Hurricane Cliffs, and in domes along the Virgin anticline are the oldest strata widely exposed in the St. George basin (figure 14).These strata accumulated on the western margin of the great supercontinent Pangea (figure 15). Cherty dolomite of the Pakoon Dolomite, cross-bedded sandstone of the Queantoweap Sandstone, and limestone, cherty limestone, and gypsum of the Toroweap and Kaibab Formations were deposited in the Early Permian when southwest Utah lay just north of the equator, at the Pangean margin. The flat landscape was alternately inundated by warm, shallow seas and exposed as broad coastal beaches and sabkhas (broad, flat surfaces near sea level where high evaporation rates commonly lead to the accumulation of salts in sediments, such as today in some parts of the Arabian Peninsula) (McKee, 1938; Rawson and Turner-Peterson, 1979; Nielson, 1986).

The upper contact of the Kaibab Formation, the youngest Permian rocks in southwest Utah, is an erosional unconformity that spans 10 to 20 million years during Late Permian and Early Triassic time (Nielson, 1981, 1991; Sorauf and Billingsley, 1991). This is the TR-1 unconformity of Pipiringos and O'Sullivan (1978), the first regional unconformity of the Triassic Period in the western U.S. The unconformity represents an episode of dramatic, worldwide sea-level drop and the largest global extinction event in Earth's history (see, for example, Ward, 2004). In the greater St. George area, erosion associated with the Permian-Triassic unconformity typically cuts out several tens of feet of the Harrisburg Member of the Kaibab Formation; locally, along major paleochannels now exposed in the White Hills southwest of St. George, 500 feet (150 m) of Permian strata is missing at this unconformity (Higgins, 1997) (figure 16).

Triassic
(251 to 200 million years ago)

The Moenkopi Formation of southwestern Utah, with its alternating reddish-brown, white, and gray layers, documents renewed shallow-marine sedimentation along the western margin of Pangea. The Moenkopi consists of three transgressive members (the Timpoweap, Virgin Limestone, and Shnabkaib Members that each record an interval of sea-level rise), each of which is overlain by an informally named regressive red-bed member (the lower, middle, and upper red members, respectively, which record sea-level fall) (figure 17); the Rock Canyon Conglomerate Member locally forms the base of the Moenkopi Formation (Reeside and Bassler, 1921; Stewart and others, 1972a; Dubiel, 1994). These members thus record a series of incursions and retreats of a shallow ocean across a gently sloping continental shelf, where sea-level changes of several feet translated into shoreline changes of many miles (Blakey and others, 1993; Dubiel, 1994) (figure 18).

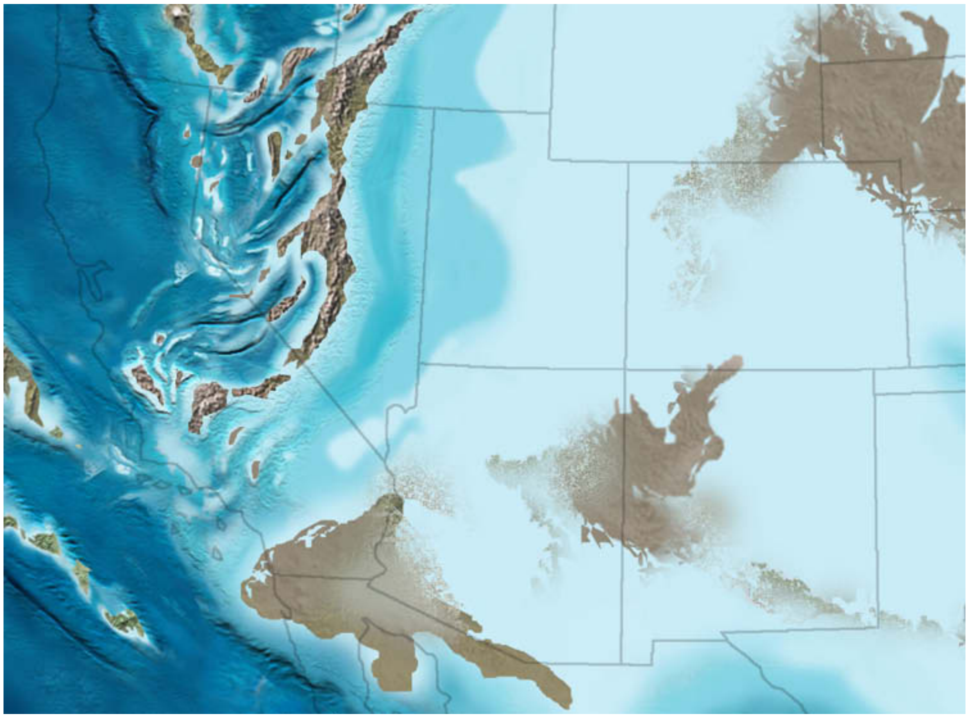

Figure 12. *Schematic reconstruction of Early Mississippian paleogeography of the southwest U.S. (about 340 million years ago). The Misissippian was a time of widespread limestone deposition across most of western North America. In southwest Utah, these strata belong to the Redwall Limestone, which forms a prominent cliff along the east side of the Beaver Dam Mountains. The Redwall contains common coral, brachiopod, and other fossils indicative of warm, shallow seas. Note position of foreland basin that developed east of the Antler orogenic belt. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

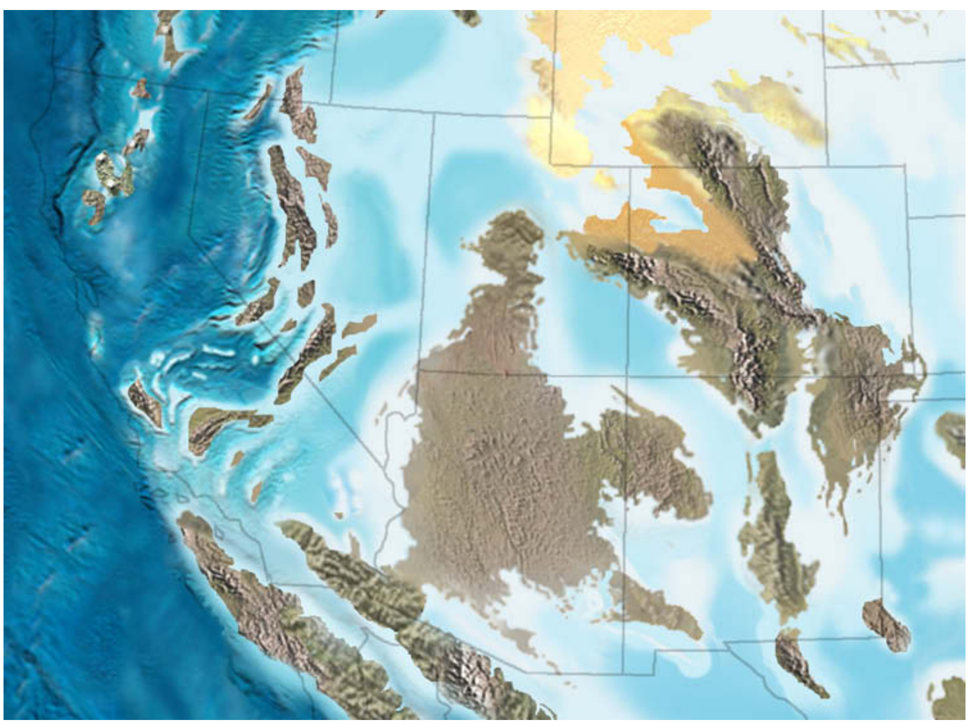

Figure 13. *Schematic reconstruction of Pennsylvanian paleogeography of the southwest U.S. (about 300 million years ago). Note the Ancestral Rocky Mountains and associated basins, which formed during this time as continental landmasses collided to form the Pangean supercontinent. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

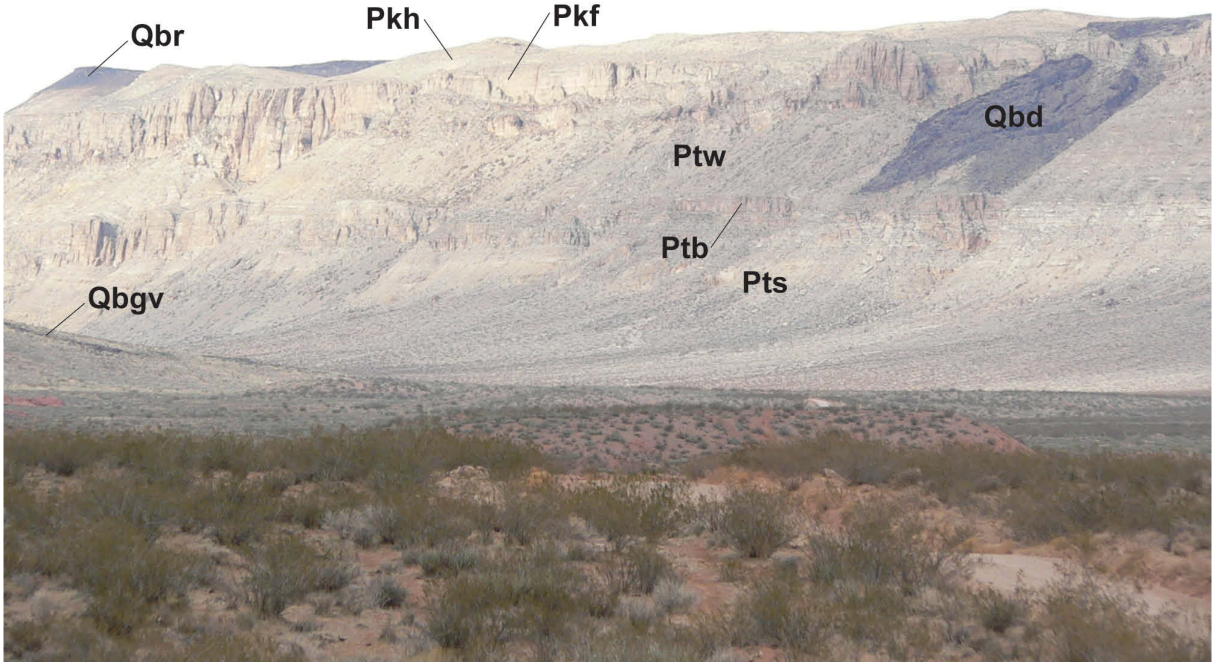

Figure 14. View northeast to the Hurricane Cliffs, about 8 miles (5 km) south of the Virgin River. Here, the Early Permian is represented by the cliff- and slope-forming shallow-marine rocks of the Kaibab and Toroweap Formations. Pkh, Harrisburg Member and Pkf, Fossil Mountain Member of the Kaibab Formation; Ptw, Woods Ranch Member, Ptb, Brady Canyon Member, and Pts, Seligman Member of the Toroweap Formation. The Rock Canyon Conglomerate Member and Timpoweap Member of the Moenkopi Formation cap the rounded hills at the top of the cliffs. The 410,000-year-old Divide lava cascade (Qbd) is visible at the right, the 1.0-million-year-old Remnants lava flow (Qbr) caps lower Moenkopi strata atop the Hurricane Cliffs at the left, and the 1.1-million-year-old Grass Valley lava flow (Qbgv) caps Navajo Sandstone at the base of the cliffs. The Hurricane fault zone is at the base of the cliffs; the cliffs are, in essence, a little-eroded fault scarp that dramatically exhibits several thousand feet of displacement on this part of the fault.

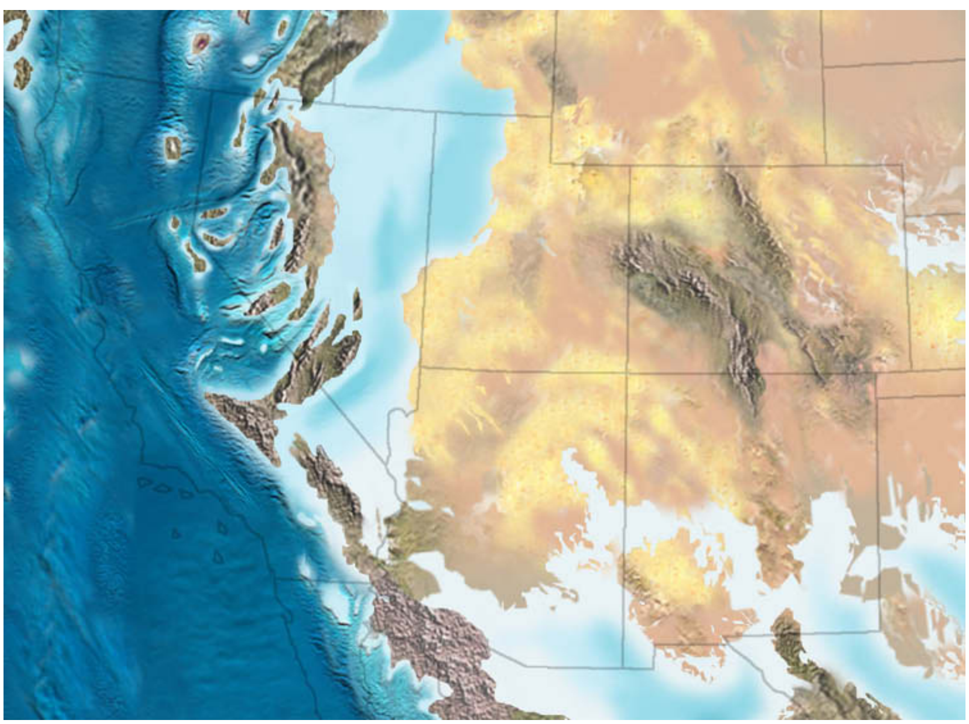

Figure 15. Schematic reconstruction of Early Permian paleogeography of the southwest U.S. (about 280–270 million years ago). What is now southwest Utah was alternately inundated by shallow seas and exposed as broad coastal beaches and sabkhas. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

Perhaps the most interesting part of the Moenkopi Formation is its basal member, the Rock Canyon Conglomerate. It consists of two main rock types: a rounded pebble and cobble conglomerate found in paleovalleys (figure 16), and a widespread but thin regolithic breccia. Clasts in both types are chert and minor limestone derived exclusively from the underlying Kaibab Formation. The conglomerate deposits are restricted to paleochannels as much as 500 feet (150 m) deep in the St. George area (Higgins, 1997), whereas the breccia deposits, which resulted from in-place weathering of Harrisburg strata, are found over paleotopographic high areas between channel-form conglomerates (Nielson, 1991).

Figure 16. *View northeast across Curly Hollow Wash in the White Hills, west of St. George. Here, Late Permian to Early Triassic erosion associated with a worldwide drop in sea level cut out all of the Harrisburg Member and part of the Fossil Mountain Member (Pkf) of the Kaibab Formation, so that river channel deposits of the Rock Canyon Conglomerate Member of the Moenkopi Formation (TRmr) rest on and are deeply inset into the Fossil Mountain strata.*

The Timpoweap Member and Virgin Limestone Member of the Moenkopi Formation are shallow-marine limestone and mudstone that thicken from east to west across the Utah hingeline, across the former continental shelf of southwest Utah. The Shnabkaib Member also thickens westward, but its abundant gypsum, mudstone, and stromatolites (algal mounds) indicate deposition in a restricted-marine and coastal-sabkha environment; the member contains limestone in the Beaver Dam Mountains, indicating more open-marine conditions existed to the west. Conversely, the red-bed members thin to the west, away from the former continental margin; common ripple marks and other features indicate that they were deposited in tidal-flat and coastal-plain environments (Stewart and others, 1972a). Locally, as in the White Hills, the Timpoweap and lower red members are restricted to paleochannels, showing that paleotopography was not completely buried until the Virgin Limestone was deposited.

In southwest Utah, the TR-3 regional unconformity (Pipiringos and O'Sullivan, 1978) separates Lower Triassic (Moenkopi Formation) and Upper Triassic (Chinle Formation) rocks and marks a change from mostly shallow-marine to continental sedimentation. This change occurred as oceanic island arcs continued to collide and accrete to the western Pangean margin in what is now California and southwest Arizona during the Sonoman orogeny. These collisions deformed the western continental margin, forming a large chain of mountains and volcanoes. In Utah, by the Late Triassic, the environment changed from a passive continental margin to a large interior basin drained by north- and northwest-flowing rivers. Volcanic ash from the western volcanoes drifted east to the interior basin where it was deposited as the colorful layers of the Chinle Formation (figure 19). As mapped here, the Chinle Formation of southwest Utah consists of the Shinarump Conglomerate Member and the Petrified Forest Member (Stewart and others, 1972b). The resistant Shinarump Conglomerate forms a prominent hogback along the Virgin anticline (figure 20) and elsewhere typically forms bold mesas (figure 17). The overlying non-resistant Petrified Forest Member is commonly involved in and covered by landslide deposits, but where well exposed, it typically displays bright, variegated mudstone slopes (figure 21).

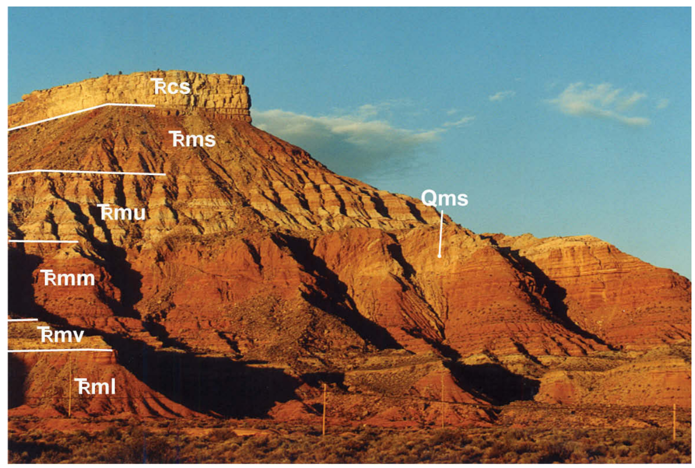

Figure 17. *View north to Hurricane Mesa from State Highway 9. Hurricane Mesa, capped by the resistant Shinarump Conglomerate Member of the Chinle Formation (TRcs), has a World War II-era test track once used by the Air Force for testing ejection seats. The lower red member (TRml), Virgin Limestone Member (TRmv), middle red member (TRmm), Shnabkaib Member (TRms), and upper red member (TRmu) of the Moenkopi Formation are also visible. Note landslide (area of disrupted bedding, Qms) in middle red and Shnabkaib strata at the right center of the photo.*

The sandstone, pebbly sandstone, and pebbly conglomerate of the Shinarump Conglomerate Member were deposited in braided-stream channels, whereas the Petrified Forest Member was deposited in floodplains, lakes, and high-sinuosity stream channels in a low, forested basin (Stewart and others, 1972b; Dubiel, 1994). Amphibians, reptiles (including the crocodile-like phytosaur), freshwater clams, snails, ostracods, and fish made their home on this once vast, coastal lowland, and petrified conifer trees are common in Chinle strata; fossil cycads, ferns, and horsetails are also known, as are dinosaur tracks (Stewart and others, 1972b; Blakey and others, 1993; Dubiel, 1994; DeCourten, 1998). Abundant bentonitic mudstone, which swells when wet and shrinks when dry (giving weathered surfaces a "popcorn" appearance), was derived from volcanic ash carried in from the magmatic arc to the south and west. This mudstone

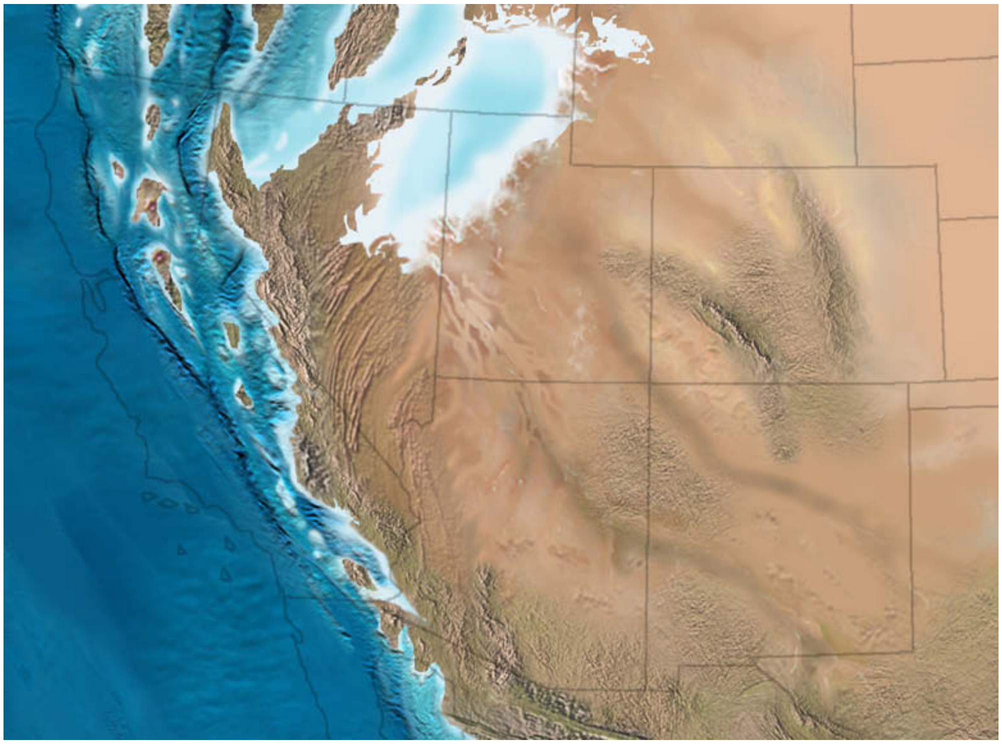

Figure 18. *Schematic reconstruction of Early Triassic paleogeography of the southwest U.S. (about 240 million years ago). What is now southwest Utah was near the gently sloping western margin of Pangea, alternately inundated by shallow seas and exposed as broad tidal flats and coastal plains. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

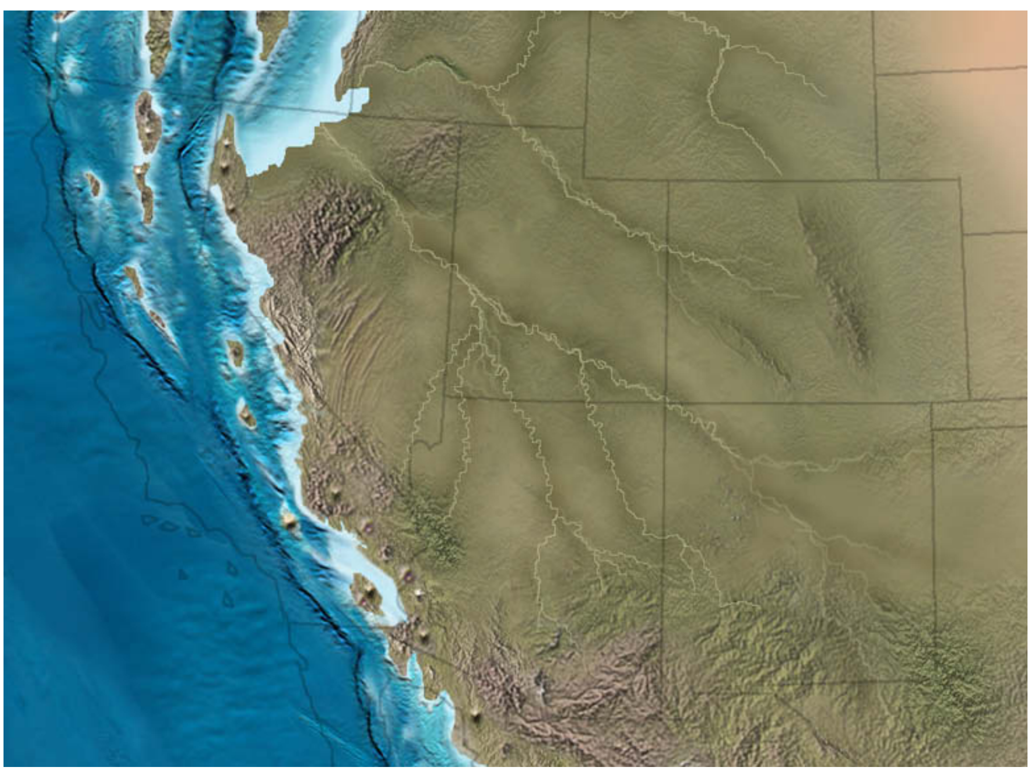

Figure 19. *Schematic reconstruction of Late Triassic paleogeography of the southwest U.S. (about 220 million years ago). Note interior basin setting dominated by northwest-flowing rivers in what is now southwest Utah. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

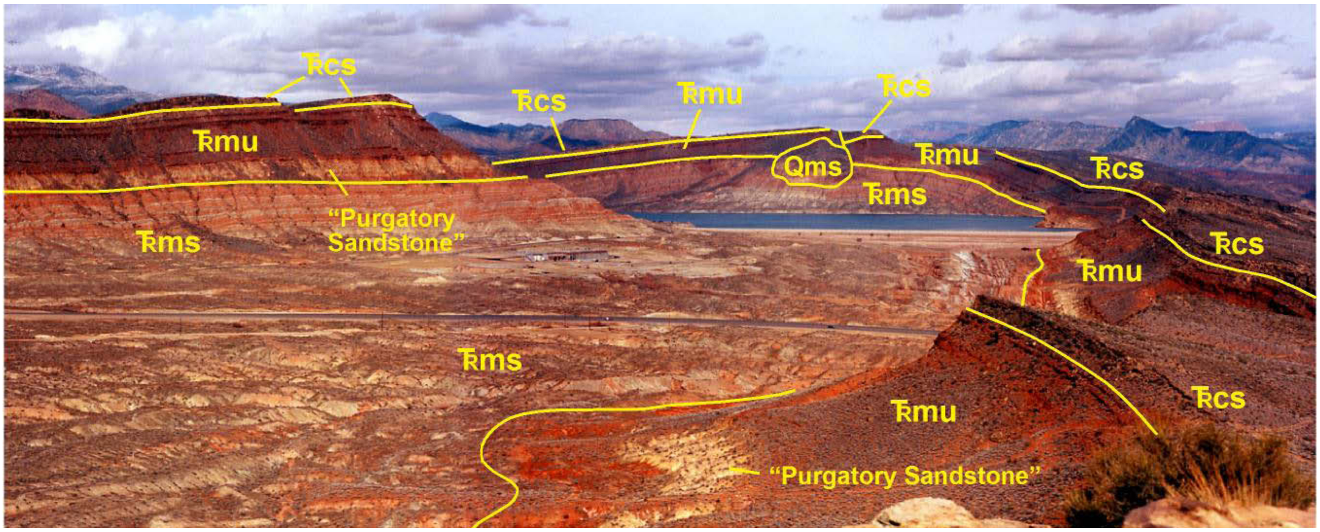

Figure 20. *View northeast to Quail Creek Reservoir, near the north end of the Virgin anticline. Note resistant hogbacks of the Shinarump Conglomerate Member (TRcs) of the Chinle Formation that overlie the upper red member (TRmu) and the "bacon-striped" Shnabkaib Member (TRms) of the Moenkopi Formation. A landslide (Qms) in upper red and Shnabkaib strata is also shown. The "Purgatory sandstone" has been bleached by reducing fluids in the core of the anticline. Dissolution of gypsiferous Shnabkaib strata led to catastrophic collapse of the Quail Creek south dam on January 1, 1989 (see Biek, 2003a, and references therein). The dam was rebuilt in 1990 and continues to be an important part of the Washington County Water Conservancy District water supply.*

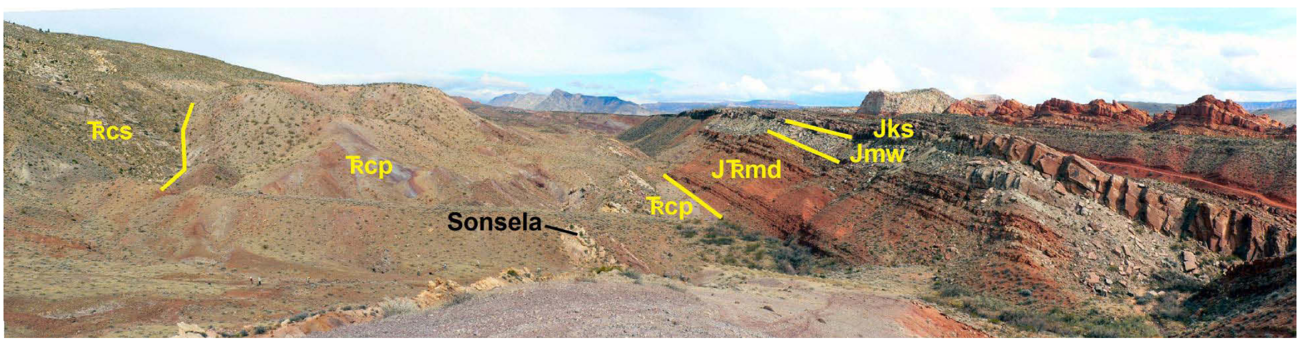

Figure 21. *View north of the strike valley eroded into the Petrified Forest Member of the Chinle Formation at East Reef, just north of where Grapevine Wash passes through the reef. East Reef, so named for the resistant hogback (or "reef") of Springdale Sandstone Member once mined for silver in the late 1800s, provides some of the best exposures of Petrified Forest strata in the greater St. George area. TRcs, Shinarump Conglomerate Member, and TRcp, Petrified Forest Member of the Chinle Formation; JTRmd, Dinosaur Canyon Member and Jmw, Whitmore Point Member of the Moenave Formation; Jks, Springdale Sandstone Member of the Kayenta Formation. The Sonsela is a sandstone bed commonly found in the middle part of the Petrified Forest Member. The Hurricane Cliffs near Toquerville are visible in the distance.*

is notoriously susceptible to landslides and has caused numerous failures of building foundations and roads in southwest Utah (figure 22). Ancient soil horizons (paleosols) are common in the Petrified Forest Member and are noted for their mottled beds and limestone nodules. The upper contact of the Petrified Forest Member, typically marked by a thin chert-pebble conglomerate of the basal Dinosaur Canyon Member of the Moenave Formation (figure 23), is the "J-0" unconformity and represents a gap of about 10 million years during the Late Triassic (Pipiringos and O'Sullivan, 1978). (The J-0 unconformity of Pipiringos and O'Sullivan [1978] was assumed to be at the base of the Jurassic, but new evidence suggests this unconformity is actually within Triassic strata and that the Triassic-Jurassic boundary is within Dinosaur Canyon strata [Molina-Garza and others, 2003; Lucas and others, 2005; Kirkland and Milner, 2006].)

Jurassic
(200 to 145 million years ago)

The interior basin setting that characterized Late Triassic time persisted into the Early Jurassic with deposition of the Moenave Formation in a variety of river, lake, and floodplain environments (Clemmensen and others, 1989; Blakey, 1994; Peterson, 1994) (figure 24). With the discovery of exceptionally well-preserved dinosaur tracks and associated fossils at the St. George Dinosaur Discovery Site at Johnson Farm in 2000 (figure 25), the Moenave Formation has been the focus of numerous stratigraphic and paleontological studies in recent years (for example, see Kirkland and Milner, 2006; Milner and others, 2006). The formation consists of two distinctive members: the Dinosaur Canyon Member below and the Whitmore Point Member above (figure 26).

Figure 22. *First recognized in the early 1980s, the Truman Drive landslide developed in the Petrified Forest Member of the Chinle Formation, and has remained intermittently active to the present; major periods of reactivation occurred in 1992 and again in 2002–03 (Lund and others, 2007b). In an effort to stabilize the main landslide scarp and prevent further damage to homes and utilities, large-diameter boreholes were drilled along the top of the slide, culverts reinforced with rebar were inserted in the boreholes, and the culverts were filled with concrete to create "soldier piles." However, the landslide has continued to enlarge, destroying several of the piles. Large landslides in Petrified Forest strata are common, especially where exposed in steep hillsides protected from erosion by resistant cap rock or terrace gravels. The brightly colored swelling claystone of the member also exhibits a high shrink-swell potential (it expands when wet and shrinks upon drying), commonly leading to road and foundation problems.*

If the Dinosaur Canyon Member is noted for its uniformity—mostly reddish-brown, thin-bedded, fine-grained sandstone and silty sandstone deposited in fluvial and floodplain environments—the Whitmore Point Member is known for its variety. The Whitmore Point contains sandstone and siltstone similar to the Dinosaur Canyon, but it also contains reddish-purple to greenish-gray mudstone and claystone and thin dolomitic limestone beds. The limestone is bioturbated and contains small, reddish-brown chert nodules and blebs, poorly preserved and contorted algal structures, and locally abundant fossil fish scales and bones of semionotid fish, including *Semionotus kanabensis* (Hesse, 1935; Schaeffer and Dunkle, 1950; Milner and others, 2006). The fish fossils were originally thought to be restricted to the Triassic and so conflicted with palynomorphs from the Whitmore Point Member that indicate the unit is Early Jurassic (latest Sinemurian to earliest Pliensbachian) (Peterson and others, 1977; Imlay, 1980). Olsen and Padian (1986) later found that *Semionotus kanabensis* is not age-diagnostic, which resolved the long-standing debate on the Early Jurassic age of the Moenave and Kayenta Formations and the Navajo Sandstone. Dinosaur tracks and fossil plants, fish, and invertebrates at the St. George Dinosaur Discovery Site, preserved on the margin of "Lake Dixie," also confirm an Early Jurassic age for most of the Moenave Formation in southwest Utah (Kirkland and Milner, 2006).

Figure 23. *The "J-0" unconformity between the Late Triassic Chinle Formation and the Upper Triassic to Lower Jurassic Moenave Formation in Warner Valley. This unconformity, at the base of the lowest ledge, represents an erosional gap of about 10 million years in Upper Triassic time. Throughout Washington County, this contact is marked by a chert-pebble conglomerate and abundant anhydrite nodules.*

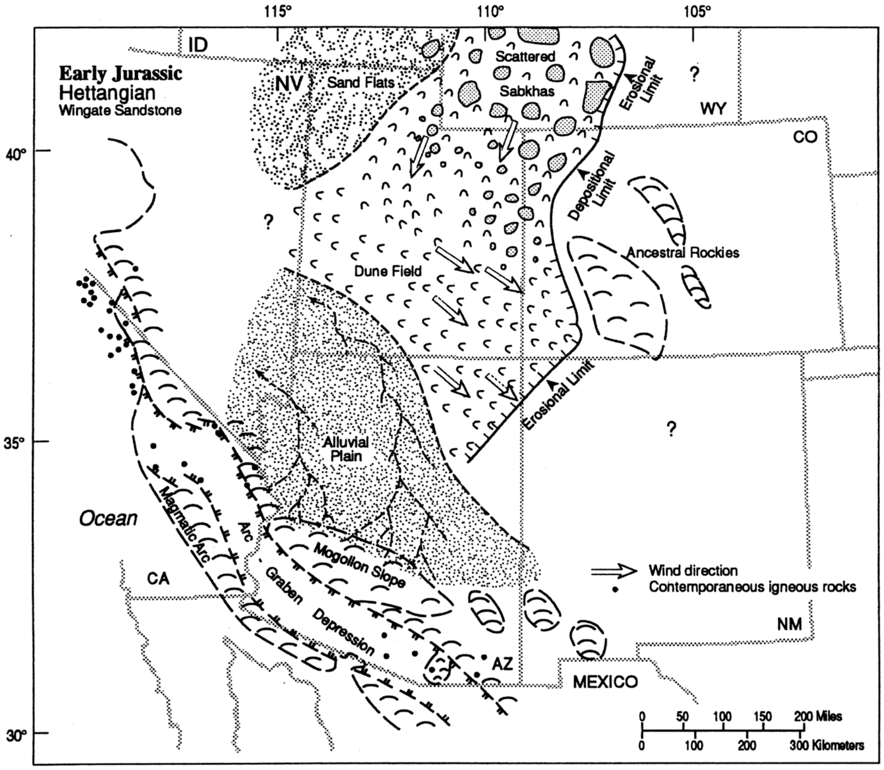

Figure 24. *Schematic reconstruction of earliest Jurassic paleogeography of the southwest U.S. (about 200 million years ago). What is now southwest Utah was part of a low-elevation interior basin flanked on the north and east by enormous dune fields now represented in part by the Wingate Sandstone (from Peterson, 1994).*

Figure 25. *Natural cast of a three-toed* Eubrontes *dinosaur track shortly after its discovery at the Johnson Farm tracksite in February 2000 by Sheldon Johnson, a retired optometrist who was using a backhoe to prepare his land for development; boxwork-like lines are natural casts of mudcracks. Inset shows close-up of a nearby print, showing detail of two toes and claws. Sheldon donated the land to the City of St. George and a museum opened in 2005. The site contains thousands of dinosaur tracks in multiple track layers, including some of the best preserved tracks in the world, with skin and claw impressions, tail drag marks, squatting marks, and the first unequivocal evidence of swimming dinosaurs anywhere in the world. Most tracks are* Eubrontes, *likely made by the crested, meat-eating dinosaur* Dilophosaurus, *which reached 20 feet (6 m) long, 7 feet (2 m) high at the hips, and weighed about 1000 pounds (450 kg). Many other types of tracks and huge, gar-like fish are also found at the site. Pencil is 5.5 inches (14 cm) long.*

Apart from several low sandstone cliffs in the upper Kayenta Formation—including the Lamb Point Tongue of the Navajo Sandstone present only in the east part of the map area at Zion National Park—the Springdale Sandstone Member of the Kayenta Formation forms the first significant cliff or ledge below the Navajo Sandstone. Because the Springdale Sandstone was deposited principally in river channels, it contains thin, discontinuous lenses of intraformational conglomerate, with mudstone and siltstone rip-up clasts and poorly preserved, petrified and carbonized fossil plant remains. Such channel deposits near Leeds, at the northeast-plunging nose of the Virgin anticline, were the primary ore horizons of the Silver Reef mining district (figure 27) (Proctor and Shirts, 1991; Biek and Rohrer, 2006).

The upper part of the Kayenta Formation conformably and gradationally overlies the Springdale Sandstone Member and was deposited in river, distal river/playa, and, especially in the lower part of the formation, minor lake environments (Blakey, 1994; Peterson, 1994). It forms a deep-red ledgy slope between cliffs of the Springdale Sandstone below and the Navajo Sandstone above (figure 28). At Zion National Park, an eolian (wind blown) sandstone known as the Lamb Point Tongue of the Navajo Sandstone forms a ledge about one-third of the way down from the base of the Navajo. Thus, in eastern exposures, the lower two-thirds of the Kayenta above the Springdale Sandstone are designated the main body of the Kayenta Formation. The upper third is known as the Tenney Canyon Tongue. The Lamb Point Tongue pinches out westward so that the upper part of the Kayenta Formation is undivided throughout most of the map area. In southwest Utah, the Kayenta is probably best known for its dinosaur tracks.

As movement of the Earth's tectonic plates continued to carry North America northward, and as mountains in what are now Nevada and California created a rain shadow, Utah entered the arid, low-latitude, trade-winds climatic belt. Vast dune fields similar to the modern Sahara eventually overwhelmed the Kayenta playas, resulting in the Navajo Sandstone, part of the world's largest coastal and inland paleodune field (Blakey and others, 1988; Blakey, 1994; Peterson, 1994) (figure 29). This transition is clearly recorded in strata that enclose the gradational contact between the Kayenta Formation and Navajo Sandstone (figure 30). During this transitional time, the area was a sabkha, a broad, flat evaporation pan with a high water table. As sand was blown onto the sabkha, it adhered to the wet surface. Crinkle bedding and salt casts formed through the growth of evaporite minerals. Eventually, the sabkha was overridden by the wind-blown sand, but the water table remained high. This led to the formation of planar sandstone beds in the lower Navajo as dry sand was blown away and wet sand remained behind. With time, large sand dunes eventually formed. The great, sweeping cross-beds of these sand dunes are preserved in the middle and upper parts of the Navajo Sandstone, where uncommon planar sandstone and limestone beds provide a record of widely scattered oases.

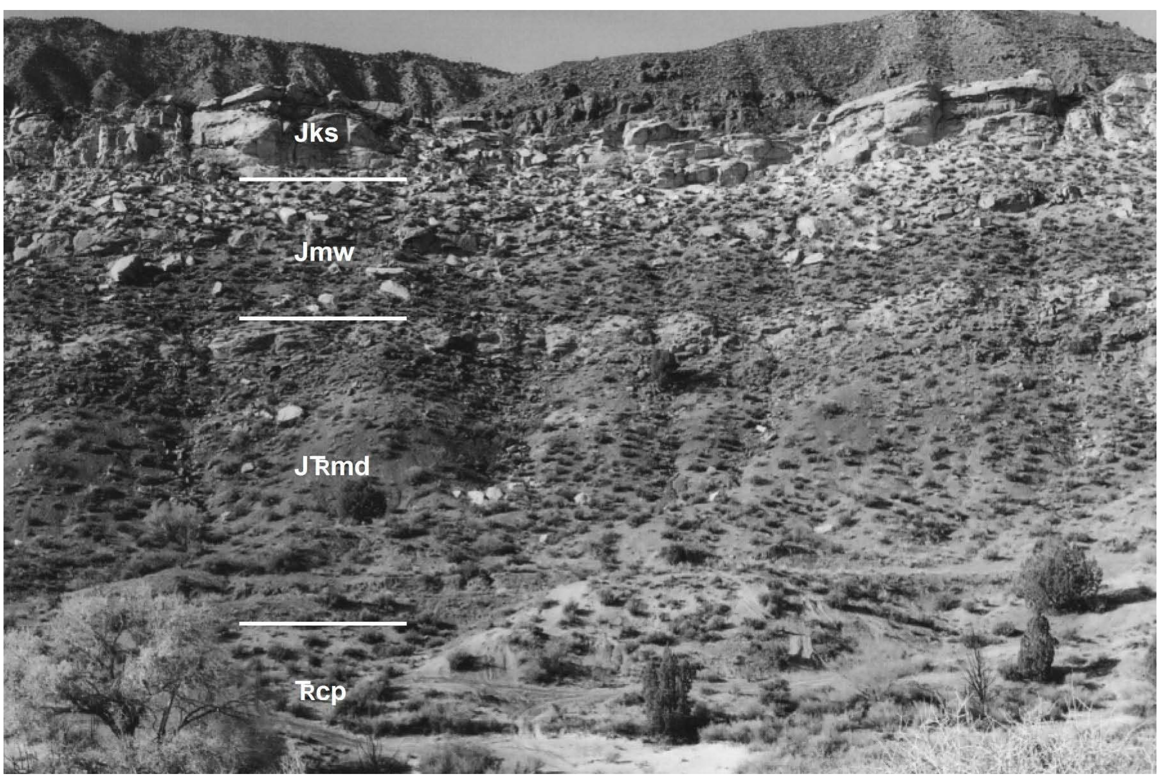

Figure 26. *View west to the north end of White Reef, northwest of Leeds, in the Silver Reef mining district. The Moenave Formation consists of the Dinosaur Canyon Member (JTRmd) and Whitmore Point (Jmw) Member. The Springdale Sandstone Member of the Kayenta Formation (Jks) in the district is famous for the unusual occurrence of ore-grade silver chloride in sandstone. The Petrified Forest Member of the Chinle Formation (TRcp) is exposed at the bottom of the wash.*

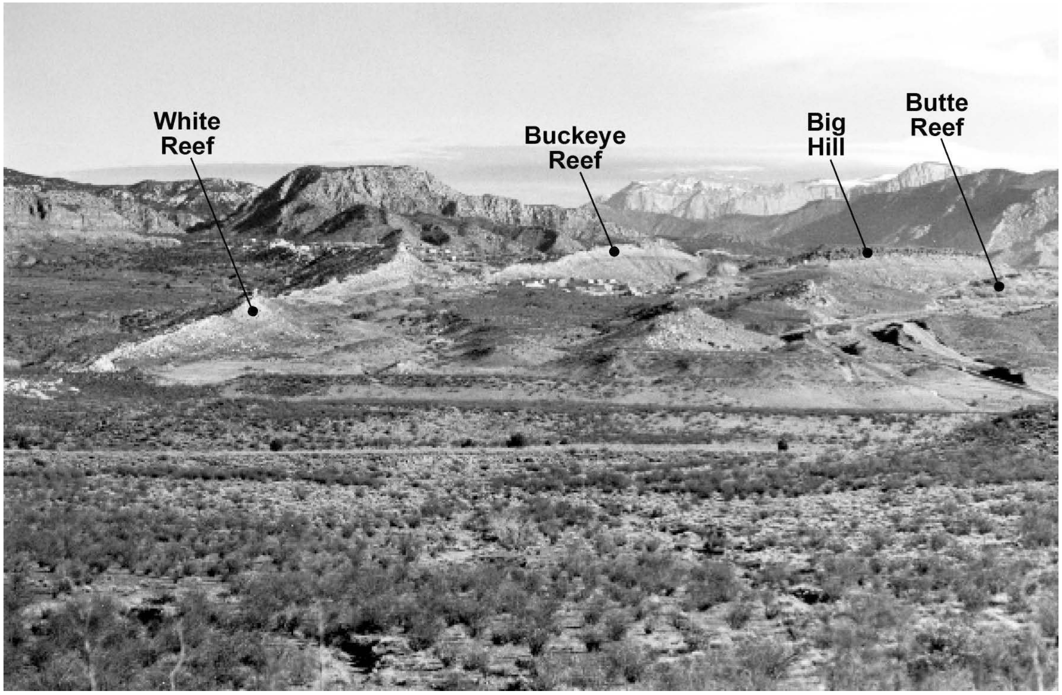

Figure 27. *View northeast of the Silver Reef mining district. Cottonwood Creek is in the foreground, the snow-capped peaks in the distance are in the Kolob Canyons section of Zion National Park, and Interstate 15 and Leeds are on the right. Prior to 1910, the district produced more than 7 million ounces (200,000 kg) of silver; sporadic production through the mid-1900s yielded additional silver, copper, gold, and uranium. White Reef, Buckeye Reef, Big Hill, and Butte Reef are each capped by the resistant, ledge-forming Springdale Sandstone Member of the Kayenta Formation.*

Figure 28. *View north to Mt. Kinesava near the south end of Zion Canyon from near the radio towers southeast of Rockville. Note the deep red slope of the upper part of the Kayenta Formation (Jk) between the cliffs of Navajo Sandstone (Jn) and the Springdale Sandstone Member of the Kayenta Formation (Jks). The Shinarump Conglomerate Member of the Chinle Formation (TRcs) caps the prominent mesa above Rockville (above the upper red member of the Moenkopi Formation, TRmu), above which are the Petrified Forest Member of the Chinle Formation (TRcp), landslide deposits (Qms), and the slopes of the Dinosaur Canyon Member (JTRmd) and Whitmore Point Member (Jmw) of the Moenave Formation.*

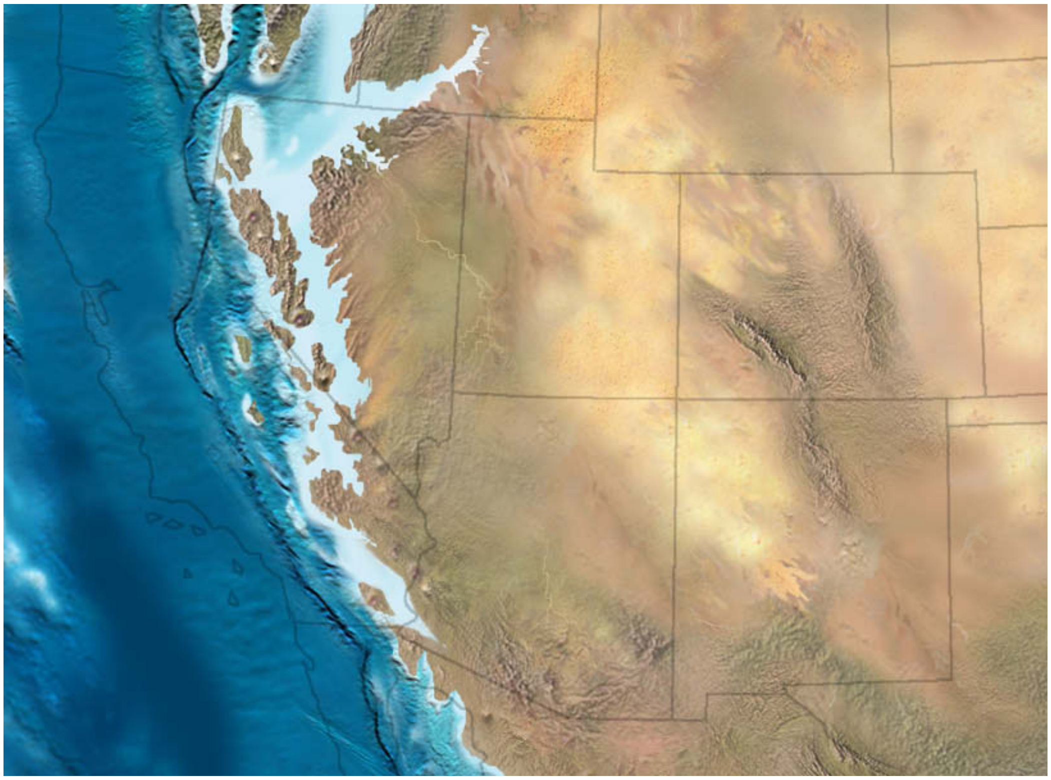

Figure 29. *Schematic reconstruction of Early Jurassic paleogeography of the southwest U.S. (about 200 million years ago). By this time, what is now southwest Utah entered the dry, low-latitude, trade-winds climatic belt, where the Navajo Sandstone, part of the world's largest coastal and inland dune field, was deposited. Much of the sand in the Navajo was ultimately derived from the ancestral Appalachian Mountains. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

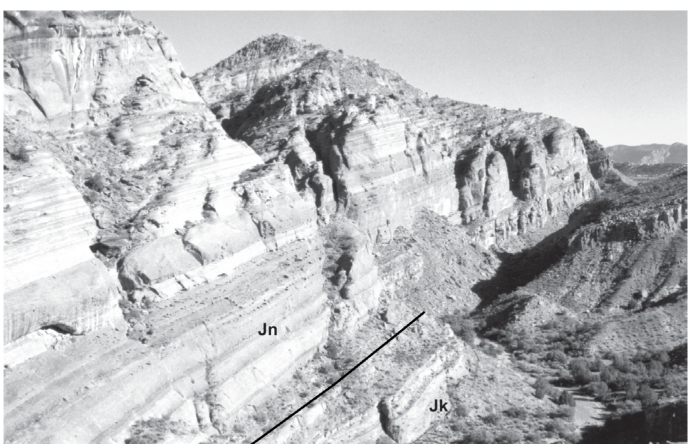

Figure 30. *View north to the transition zone between the upper part of the Kayenta Formation (Jk) and the Navajo Sandstone (Jn), as seen from the Red Cliffs Recreation Area overlook. Note the planar-bedded sandstone in the lower part of the Navajo Sandstone. The transition zone records inundation of a coastal sabkha by wind-blown sand.*

The Lower Jurassic Navajo Sandstone is renowned for its uniformity and great thickness, locally exceeding 2000 feet (600 m). It consists of moderately well-cemented, well-rounded, frosted, fine- to medium-grained quartz grains and, where little deformed by joints, weathers into bold, rounded cliffs. Above its basal transition zone, the Navajo Sandstone is characterized by large-scale cross-beds. Recent research on geochronology of detrital zircon grains in the Navajo suggests that most of the sand was eroded from the ancestral Appalachian Mountains, transported westward by a continental-scale river system to the western shore of Jurassic North America, and was then blown southward and incorporated into the Navajo-Nugget-Aztec dune field (Dickinson and Gehrels, 2003; Rahl and others, 2003). Preservation of this tremendous thickness of wind-blown sand was made possible because of basin subsidence associated with Early Jurassic compressional deformation near the west margin of North America (Allen and others, 2000). This deformation caused the continental interior to flex downward, creating accommodation space for sediment accumulation, much as it did in Late Cretaceous time as the Sevier orogenic belt migrated eastward into Utah as described later.

The top of the Navajo Sandstone forms a remarkably flat surface, the J-1 unconformity (figure 31) (Pipiringos and O'Sullivan, 1978). Over short distances, little relief is noticeable along the contact, but over several miles the relief may be as much as several hundred feet. The lower part of the Temple Cap Formation (the formation gets its name from strata that cap the West Temple in Zion National Park) marks a brief respite in desert conditions. The red mudstone of the Sinawava Member was deposited unconformably over the Navajo Sandstone in warm shallow seas in the Utah-Idaho trough; wind-blown sand dunes of the White Throne Member herald a similarly brief return to desert conditions of a coastal dune field (Blakey, 1994; Peterson, 1994). The White Throne Member thins westward and pinches out east of the Hurricane fault; only the Sinawava Member, which thickens westward, is present in western exposures. New research suggests that the Temple Cap Formation, lower part of Page Sandstone (Harris Wash Tongue of south-central Utah), and Gypsum Springs Member of the Twin Creek Limestone (of central and northern Utah) are time equivalent (D.A. Sprinkel, Utah Geological Survey, verbal communication, January 12, 2009; Dickinson and Gehrels, 2009). Thus, whereas the J-2 unconformity marks a significant change in depositional environments, recording encroachment of a shallow inland sea described below, it may not represent a significant gap in the rock record as envisioned by Pipiringos and O'Sullivan (1978).

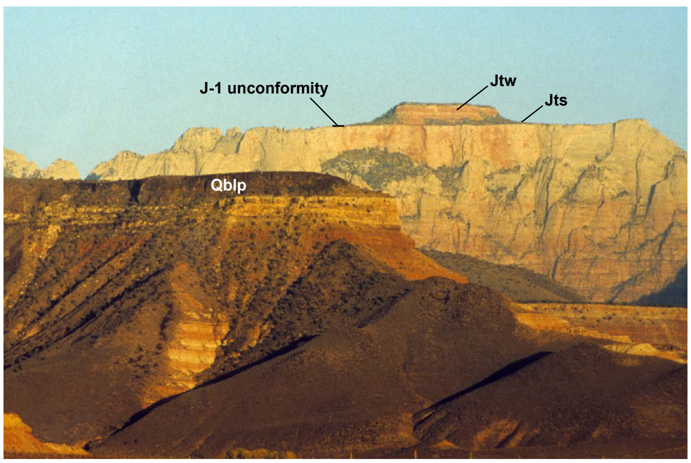

Figure 31. *View east, from State Highway 9 just west of Virgin, to the West Temple of Zion National Park, which is capped by the thin, tree-covered, slope-forming Sinawava Member (Jts) and cliff-forming White Throne Member (Jtw) of the Temple Cap Formation. The J-1 unconformity separates Temple Cap strata from the cliff-forming Navajo Sandstone (Jn) below. The mesa in the foreground is capped by the 1.0-million-year-old Lava Point basaltic flow (Qblp), which overlies the ledge-forming Shinarump Conglomerate Member of the Chinle Formation. This flow documents 1300 feet (400 m) of incision since it was emplaced and shows that most of Zion Canyon was carved within the past 2 million years.*

By Middle Jurassic time, deformation associated with subduction of the Farallon oceanic plate spread eastward into western Utah. These eastward-directed compressional forces created the Sevier orogenic belt, which consists of, from west to east, a thrust belt, foredeep basin, forebulge, and back-bulge basin (Willis, 1999, 2000) (figure 32). Each of these four parts of the thrust system migrated eastward through the area over time, and each created unique environments of deposition or erosion. The Carmel Formation, for example, was deposited in a shallow inland sea of the back-bulge basin, the first clear record of the effects of the Sevier orogeny in southwestern Utah (figure 33). The Carmel Formation consists of four members in southwest Utah: in ascending order, the Co-op Creek Limestone, Crystal Creek, Paria River, and Winsor Members (Imlay, 1980). In southwest Utah, these members form an eastward-thickening wedge preserved beneath the Cretaceous (K) regional unconformity of Pipiringos and O'Sullivan (1978). Because of increasingly deeper erosion, upper members are missing in western exposures (figure 34).

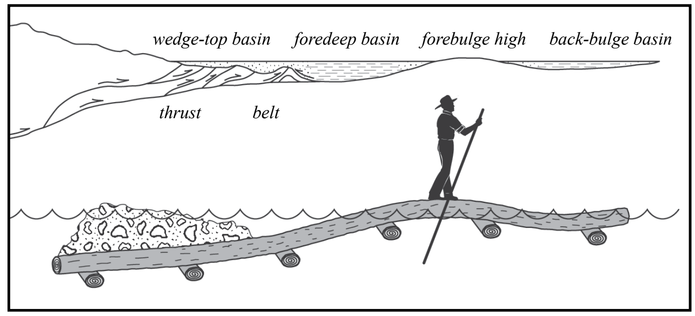

Figure 32. *Typical parts of a thrust system. The thickened, eastward-moving, leading-edge thrust wedge on the left overloads the Earth's crust, which flexes in response, similar to loading rock on a wooden raft floating on water. In Utah, the entire thrust system migrated eastward over time during the middle Mesozoic to early Tertiary, but this simple pattern is commonly complicated due to variations in crustal strength and pre-existing faults (from Willis, 1999).*

Cretaceous
(145 to 65 million years ago)

By late Middle Jurassic time, the back-bulge basin had migrated eastward, and most of Utah was a forebulge high, a broad, gentle uplift that was high enough to undergo a prolonged period of modest erosion; thus, no rocks of late Middle Jurassic to middle Early Cretaceous age are preserved in southwest Utah. By late Early Cretaceous time, the first coarse synorogenic sediments, represented by pebbly conglomerate, sandstone, and mudstone, were deposited in the foredeep basin from sources to the west in the orogenic belt. Today, this distinctive interval can be traced from the Gunlock area, across the southern base of the Pine Valley Mountains, and across the Upper Kolob Plateau north of Zion National Park (figure 35). Identification of this interval, which consists of a thin conglomerate and overlying bentonitic mudstone, is controversial. These strata were previously known as the lower Upper Cretaceous (Cenomanian) Dakota Formation; however, we now believe this conglomerate and mudstone represents previously unrecognized Cedar Mountain Formation, a lithologically similar unit considered to be mostly of late Early Cretaceous age (Tschudy and others, 1984; Kirkland and others, 1997, 1999). Additional stratigraphic study is needed to resolve middle Cretaceous nomenclature in southwest Utah.

Younger foredeep basin strata—including alluvial-fan and alluvial-plain sediments that grade eastward into finer grained coastal-plain, marginal-marine, and marine deposits—record continued encroachment of the orogenic belt (figure 36). They total several thousand feet thick in the Pine Valley Mountains and on the Upper Kolob Plateau. Much work has been done on these units on the Markagunt Plateau and areas to the east, where they are divided into the alluvial-plain and brackish estuarine and lagoonal Dakota Formation; the westward-thinning and intertonguing wedge of marine Tropic Shale; the nearshore, shallow-marine, and brackish-water Straight Cliffs Formation;

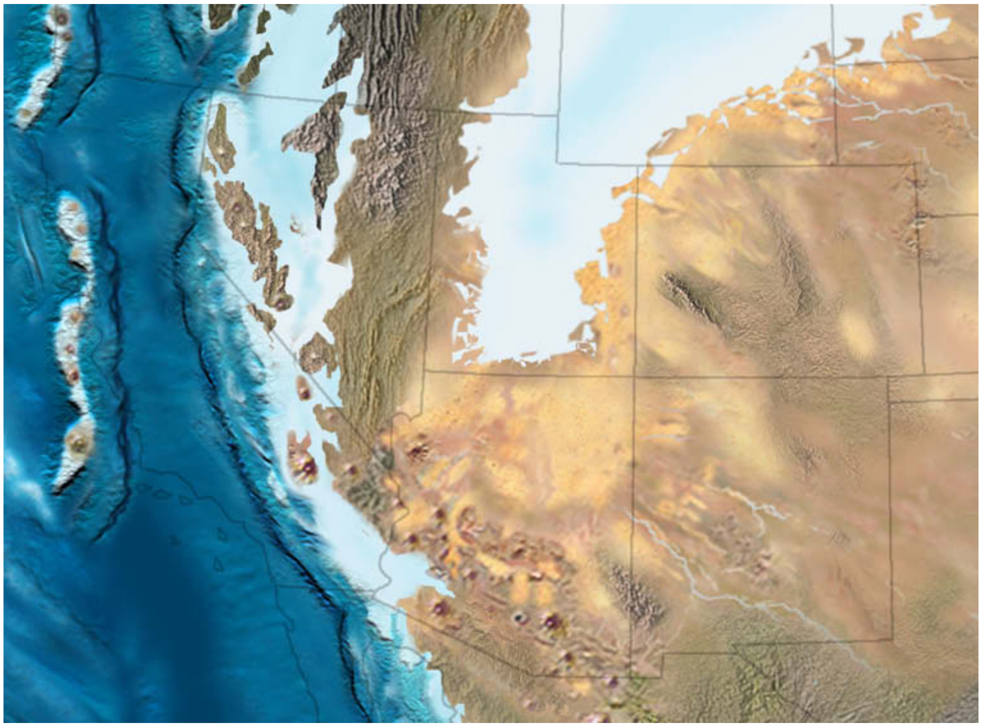

Figure 33. Schematic reconstruction of Middle Jurassic paleogeography of the southwest U.S. (about 170 million years ago). What is now southwest Utah was near the south end of a shallow inland sea, part of the back-bulge basin that developed in front of the Sevier orogenic highlands. Continued encroachment of the orogenic belt (the forebulge high) into southwest Utah led to erosion or non-deposition of Late Jurassic and Early Cretaceous strata. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

Figure 34. View northwest across Gunlock Reservoir to the northeast-dipping calcareous mudstone and limestone slopes of the Co-op Creek Limestone Member of the Carmel Formation (Jcc); the underlying red mudstone and siltstone of the Sinawava Member of the Temple Cap Formation (Jts) is mostly concealed by high-level alluvial-fan deposits. The upper members of the Carmel Formation, present on the east side of the Pine Valley Mountains and on the Kolob Plateau, are truncated beneath the K unconformity. Pebble conglomerate that we assign to the Cedar Mountain Formation (Kcm) overlies the unconformity and can be traced eastward beyond the east edge of the map area. Square Top Mountain and Jackson Peak, both in the upper plate of the Square Top Mountain thrust, are on the skyline. The 1.0-million-year-old Magotsu Creek lava flow is in the foreground. Jn, Navajo Sandstone; Ki, Iron Springs Formation.

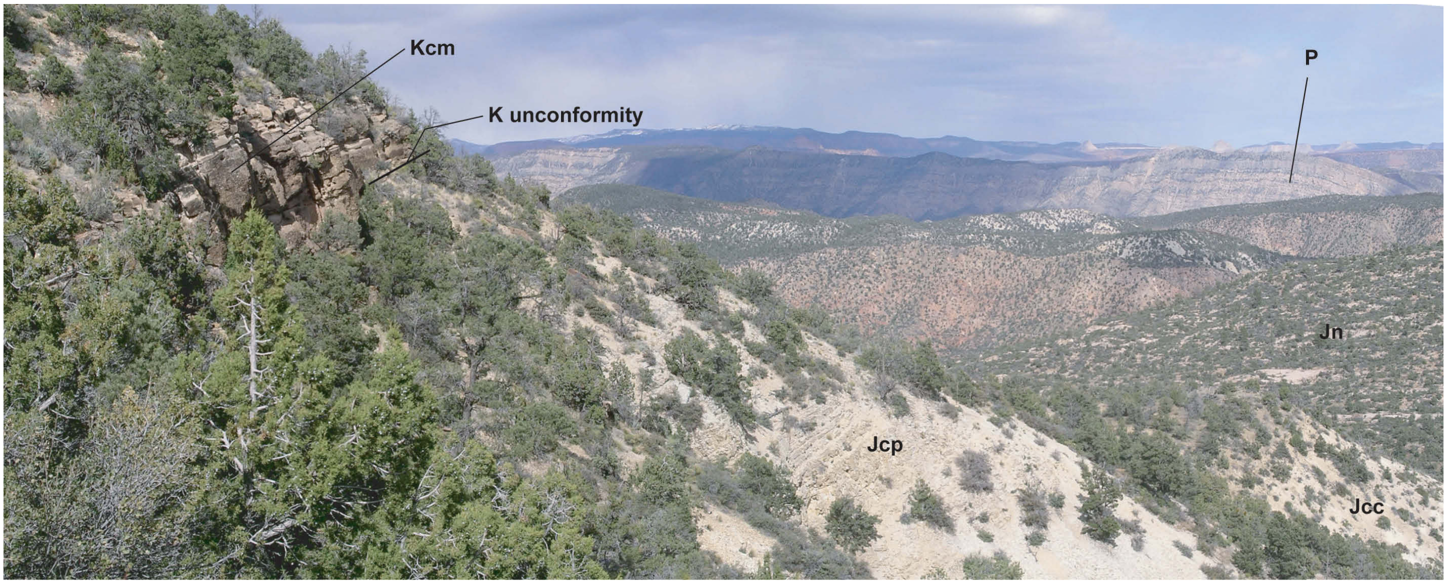

Figure 35. *Ledge-forming pebble conglomerate (Kcm) overlies the Cretaceous unconformity that here is cut into the Paria River Member of the Carmel Formation (Jcp) on Cedar Ridge on the southeast flank of the Pine Valley Mountains. This distinctive interval, as much as 100 feet (30 m) thick, can be traced from the Square Top Mountain area, through Gunlock and across the southern base of the Pine Valley Mountains, and continuing eastward across the Upper Kolob Plateau north of Zion National Park. Many geologists interpret this conglomerate and overlying bentonitic mudstone as belonging to the Upper Cretaceous Dakota Formation. However, the lithology, age, and stratigraphic position of these beds lead us to suggest correlation to the Cedar Mountain Formation (Kcm) of central Utah. Jn, Navajo Sandstone; Jcc, Co-op Creek Limestone Member of the Carmel Formation; the Crystal Creek Member is concealed in this view. Permian strata (P) exposed on the east limb of the Kanarra anticline are visible in the middle distance.*

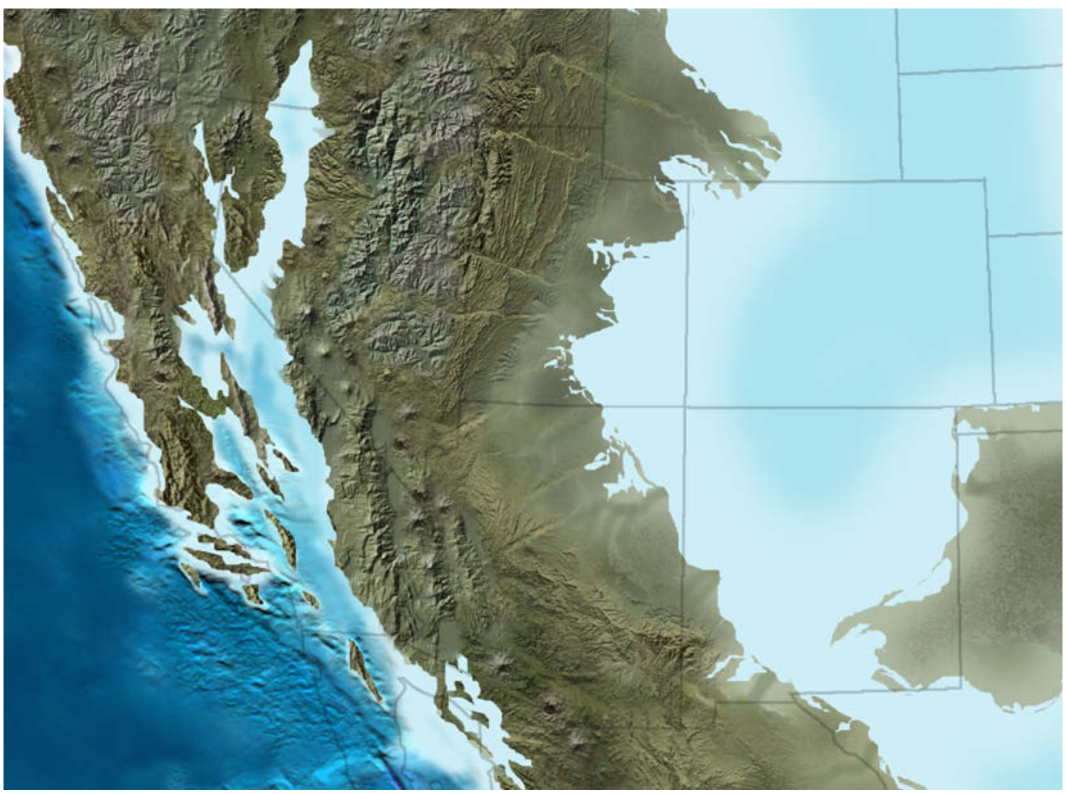

Figure 36*. Schematic reconstruction of Middle Cretaceous paleogeography of the southwest U.S. (about 90 million years ago). Middle Cretaceous time marks continued encroachment of the Sevier orogenic belt into what is now southwest Utah. At its maximum transgression, the Cretaceous Seaway reached westward into southwest Utah as far as the eastern Pine Valley Mountains. Clipped from color image created by Northern Arizona University Geology Professor Ronald Blakey; see his website http://jan.ucc.nau.edu/~rcb7/RCB.html, or* Ancient Landscapes of the Colorado Plateau *(Blakey and Ranney, 2009), for further information on these paleogeographic images; used by permission.*

and still younger Late Cretaceous strata (see, for example, am Ende, 1991; Eaton and Nations, 1991; Eaton and others, 2001; Laurin and Sageman, 2001; Tibert and others, 2003). However, west of the Hurricane fault, in the Pine Valley Mountains, most of the equivalent-age section remains undivided as sandstone, mudstone, and conglomerate of the Iron Springs Formation, which was eroded from the Blue Springs and Wah Wah thrust sheets and deposited principally in an alluvial-plain environment (Fillmore, 1991); brackish-water (marginal-marine) molluscan faunas in the lower part of the formation in the eastern Pine Valley Mountains probably indicate maximum westward transgression of the Western Interior Seaway (Eaton, 1999).

Late Cretaceous to Early Tertiary
(100 to 34 million years ago)

Late Cretaceous and early Tertiary sedimentary rocks doubtless once covered much of southwestern Utah. They are still widespread in the Pine Valley and Bull Valley Mountains, but were removed by erosion from the southern and eastern parts of the map area. The conglomerate of the upper Campanian(?) to Paleocene Canaan Peak Formation documents continued encroachment of thrust sheets from the west and deposition in river and alluvial-fan environments (Goldstrand, 1992). By Paleocene time a broad basin developed throughout southwest Utah in which the Paleocene-Eocene fluvial and lacustrine sedimentary rocks of the Claron Formation were deposited; these multi-hued orange, white, and red strata, famously exposed at Cedar Breaks National Monument and Bryce Canyon National Park, were extensively modified by soil-forming processes and now represent a stacked sequence of paleosols (ancient soils) interlayered with lacustrine and fluvial deposits (Mullett and others, 1988a, 1988b; Mullett, 1989; Davis and Pollock, 2003). The Claron Formation typically unconformably overlies older strata and signals the end of thin-skinned compressional deformation in southwest Utah, and this is dramatically displayed in exposures of the Pintura anticline.

Middle Tertiary to Present

During Oligocene and early Miocene time, explosive calc-alkaline (andesitic to rhyolitic) volcanism dominated in areas to the west and north, where huge calderas and other volcanoes erupted (Mackin, 1960; Rowley and Dixon, 2001). Thousands of feet thick to the north and west, distal parts of these volcanic rocks doubtless covered parts of the St. George basin with ash-flow and airfall tuff that has since eroded away.

The Pine Valley Mountains laccolith and other smaller laccoliths farther north and west are calc-alkaline intrusions rising above a huge batholith in the eastern part of the Delamar-Iron Springs igneous belt. The Pine Valley laccolith, which today forms the massif of the Pine Valley Mountains (figure 3), was emplaced in the early Miocene, about 20.5 million years ago, as molten rock from deep within the Earth moved upward into the Claron Formation, where it spread out into a shallow, mushroom-shaped intrusion (figures 37 and 38) (Hacker, 1998; Hacker and others, 2002, 2007; Willis, 2002). The laccolith was emplaced rapidly, which resulted in catastrophic slope failure on oversteepened topography. The

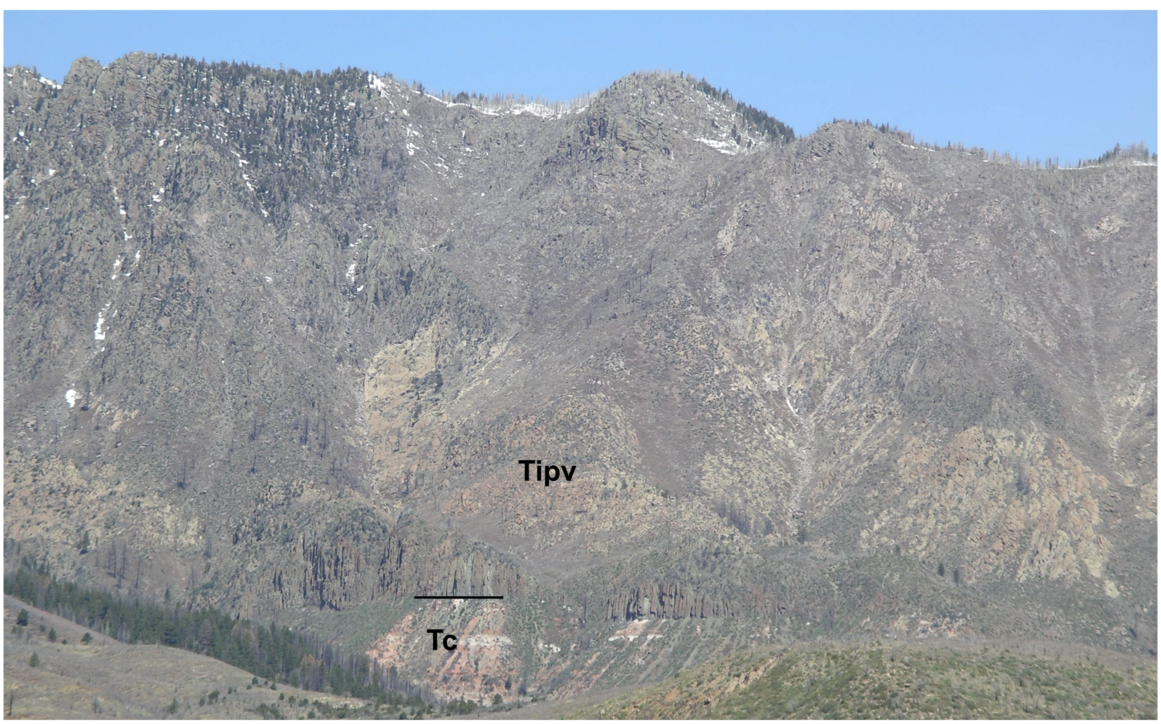

Figure 37. *View northwest into the Mill Creek drainage on the rugged east side of the Pine Valley Mountains. Here, the intrusive contact between the Pine Valley laccolith (Tipv) and the Claron Formation (Tc) is well exposed. Note the vertical columnar joints at the base of the laccolith, which formed as the intrusion cooled. Hacker and others (2007) provide a good summary of the laccolith and the mechanism of its emplacement.*

result was enormous gravity slides—highly sheared, allochthonous rock masses underlain by low-angle faults and laterally bounded by tear faults. This unroofing of the laccolith led to concomitant eruption of lava flows that blanketed the gravity-slide masses; other laccoliths in the igneous belt erupted ash-flow tuffs.

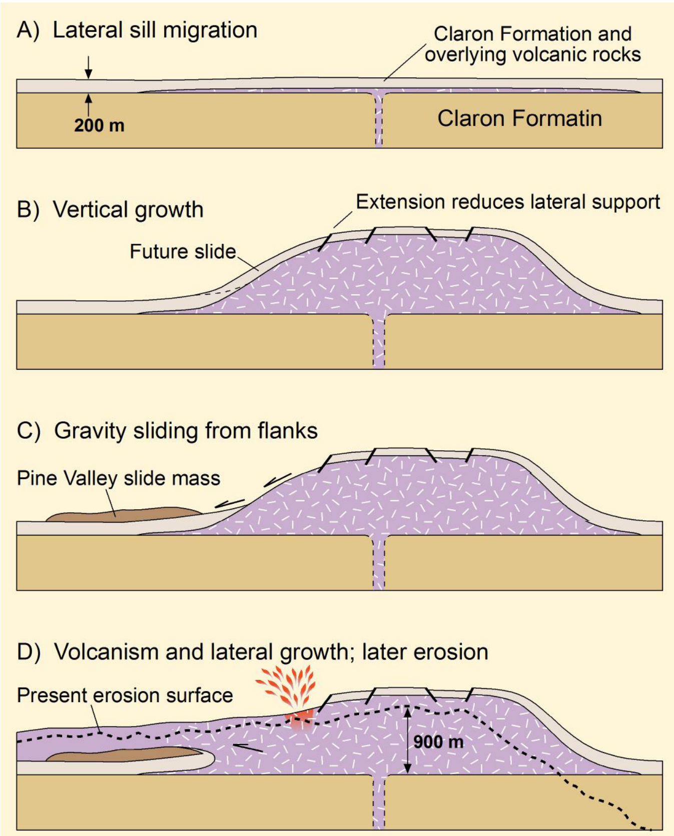

Figure 38. Schematic diagram illustrating growth of the Pine Valley laccolith. (A) Initial lateral migration of a sill to its fullest extent, likely at a depth of only about 650 feet (200 m); (B) vertical growth of the laccolith by continued injection of magma, (C) gravity sliding of oversteepened flanks, and (D) lateral growth by overturning flanks and extrusion of lava flows. Note that unlike many of the laccoliths of the nearby Iron Axis to the north, the Pine Valley laccolith did not induce explosive eruptions of ash-flow tuff, but it did produce viscous lava flows. From Willis (2002) after Hacker and others (2002).

About 17 million years ago, in the early Miocene in southwesternmost Utah, volcanism began to change from calc-alkaline to bimodal (it began changing somewhat earlier, about 23 million years ago, just to the north of the map area). Bimodal volcanic rocks are dominated by two different compositions: rhyolitic (silica- and aluminum-rich and iron- and magnesium-poor) and basaltic (silica- and aluminum-poor and iron- and magnesium-rich). This change, which is present in volcanic terranes throughout the western U.S., marks the beginning of extension or basin-range tectonics, the time when the western U.S. began to rift apart. Bimodal volcanism was dominant in the Miocene and early Pliocene, but continues to the present time. The most visible evidence is the many basalt flows prevalent throughout Washington County and surrounding areas, the oldest of which is the 17.39-million-year-old Harrison Peak basalt (part of map unit Tb). Rhyolitic volcanic rocks are less common in southwest Utah.

The youngest rocks and deposits in the county are basaltic lava flows and local thin alluvial, mass-movement, and eolian deposits. These are short-lived ephemeral deposits that are being actively weathered and transported out of the area by the Virgin River and its tributaries. The scattered deposits we see today have only found temporary respite in some protected location—they will soon (geologically speaking) be eroded and transported down the Virgin River to the Colorado River and ultimately to the Gulf of California. The basalt flows are more resistant and act as a temporary armor over whatever they cover, protecting areas for a short time. Many of the lava flows now form inverted valleys, for which the greater St. George area is justly famous. As the lava flows blocked drainages, displaced streams moved off to the side of the lava flow where they preferentially eroded softer sedimentary bedrock, ultimately leaving the resistant lava flows stranded as elevated, sinuous ridges—inverted valleys—that mark the location of former channels (figure 39). But even they cannot withstand the relentless onslaught of erosion. Within a few million years, most basalt flows are reduced to scattered remnants capping isolated mesas.

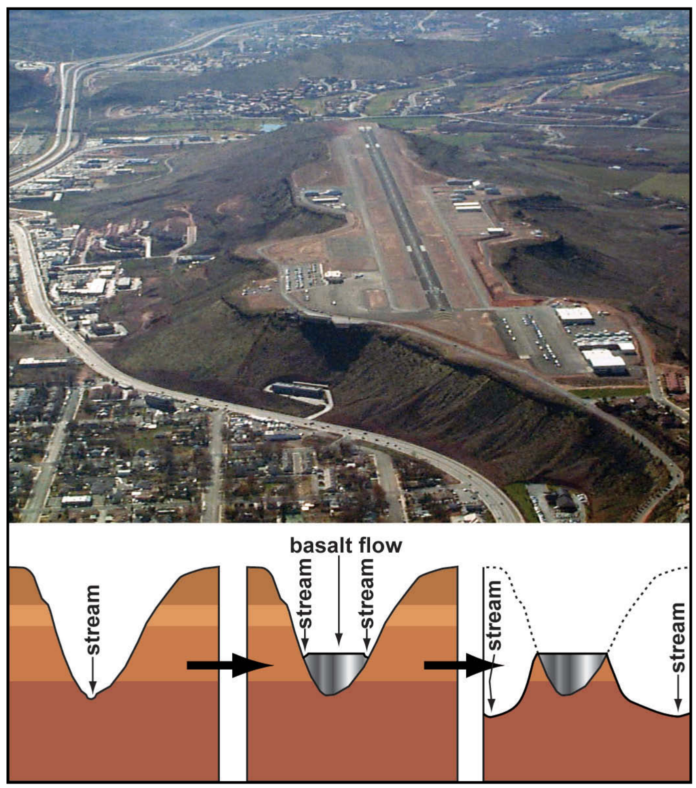

Figure 39. Many classic inverted valleys adorn the landscape of the greater St. George area. The inverted valleys formed as lava flows moved down former canyons and washes, causing the stream to shift laterally where it preferentially eroded softer sedimentary bedrock. Over time, the former valley walls are eroded, leaving the old valley floor, protected by a resistant lava flow, as a sinuous elevated ridge. This photo shows a pilot's view of the St. George Municipal Airport, which is capped by the resistant 1.2-million-year-old Airport part of the Cedar Bench lava flow. The schematic diagram shows evolution of an inverted valley. Modified from Milligan (2002). Photo courtesy of the Civil Air Patrol.

STRUCTURE

The St. George 30' x 60' quadrangle covers an area of overlap between Late Cretaceous compressional deformation associated with the Sevier orogeny, middle Tertiary voluminous calcalkaline volcanism, and late Tertiary to modern bimodal volcanism and complex basin-range extensional deformation that includes normal- and oblique-slip faulting, basin formation, and rotation of rocks along steep axes. This overprinting of older structures by younger folds and faults created extraordinarily complex geology in the west and northwest parts of the mapped area, leading to intense and long-standing controversy over interpretations of the region's structure and geologic history. Here, we attempt to explain the main controversies and offer our interpretation of the major geologic structures in the mapped area. We divide these structures into compressional and extensional features.

Late Cretaceous Compressional Structures

Square Top Mountain Thrust Fault

The Square Top Mountain thrust fault, in the northwest part of the map area, is the easternmost large thrust fault of the Sevier orogenic belt exposed in southwest Utah (figure 40) (smaller thrust faults, described below, are present to the east in the map area). It places late Paleozoic strata over Mesozoic strata and can be traced for nearly 4 miles (6 km) between the Red Hollow fault and Jackson Wash fault (Hintze, 1986; Hintze and others, 1994). In western exposures, the Pennsylvanian Callville Limestone is thrust over Late Triassic to Early Jurassic strata (Chinle, Moenave, and Kayenta Formations). The thrust fault ramps upsection to the east, so that at Jackson Peak it places Early Permian Queantoweap Sandstone over Late Cretaceous Iron Springs Formation. The upper plate is concealed by highly faulted early Tertiary strata north and east of Jackson Peak that obscure the leading edge of the thrust sheet. A steep, northwest-trending, right-lateral tear fault in structurally incompetent Middle Jurassic Temple Cap and Carmel strata links the Red Hollow and Square Top Mountain parts of the thrust fault.

Hintze and others (1994) reported that the attitude of the Square Top Mountain thrust is generally not known due to a combination of colluvial cover, partial dismemberment by later oblique-slip faults, and concealment by early Tertiary strata that are faulted down against the allochthon. Lower-plate strata generally dip moderately northeast as the west limb of the Shivwits syncline, and change to a more northerly strike with steep to overturned dips near the Red Hollow fault. The tilting of beds near Red Hollow is due to some poorly understood combination of fault drag associated with thrust faulting and later fault drag associated with left-lateral oblique slip on the Red Hollow fault (Anderson and Barnhard, 1993a; Hintze and others, 1994). Directly northeast of the easternmost exposure of upper-plate Permian sandstone are steep to overturned, northeast-striking Jurassic strata interpreted by Hintze and others (1994) to be part of a lower-plate fold of the thrust fault, thus suggesting a southeast direction of thrusting in this area.

Hintze (1986) and Carpenter and Carpenter (1994) suggested that the Shivwits syncline must predate emplacement of the thrust sheet because it appears that the Square Top Mountain thrust fault cuts across the west limb of the fold. In map view, except near Red Hollow, it does appear that the trace of the thrust is subhorizontal, but Hintze and others (1994) cautioned against this interpretation for the reasons noted above. In their cross section A-A', Hintze and others (1994) showed the thrust dipping gently northeast, subparallel to lower-plate strata. As described below, following Anderson and Barnhard (1993a, 1993b), we interpret the Shivwits syncline as a synextensional feature related to left-lateral oblique slip on the Gunlock–Reef Reservoir–Grand Wash fault zone and structural crowding against the Beaver Dam Mountains culmination.

__Figure 40.__ Oblique aerial view to the north of Square Top Mountain and Jackson Peak. Both peaks are part of the upper plate of the Square Top Mountain thrust, the easternmost large thrust fault of the Sevier orogenic belt exposed in southwest Utah. The thrust fault ramps upsection from west to east and places late Paleozoic strata over Mesozoic strata. In the foreground, northeast-dipping Kayenta Formation and Navajo Sandstone form the west limb of the Shivwits syncline; note the dissected, planar, gently south-dipping surface overlain by old alluvial-fan deposits.

The orientation of the Square Top Mountain thrust fault, and associated thrust faults and folds in upper-plate strata, suggest southeast-directed maximum compression during thrusting, which is also reflected in the orientation of the Virgin anticline and associated structures. The Square Top Mountain thrust may link up with the Keystone–Muddy Mountain–Tule Spring thrust system, which comprises the frontal thrust of the Sevier orogenic belt in southeast Nevada (Axen and others, 1990; Page and others, 2005). Carpenter and Carpenter (1994) estimated about 30 miles (50 km) of crustal shortening on this Late Cretaceous thrust system. At Square Top Mountain, emplacement of the thrust plate is bracketed between the Iron Springs Formation (Late Cretaceous—Santonian or early Campanian to Cenomanian) and the Grapevine Wash Formation (early Tertiary to Late Cretaceous), whose sediments Hintze and others (1994) interpreted as recording the approach and arrival, respectively, of the allochthon. The synorogenic Grapevine Wash Formation represents debris eroded from the upper plate of the Square Top Mountain thrust and deposited as alluvial fans and monolithic scarp breccia in front of a southeast-advancing thrust sheet (Hintze, 1986; Hintze and others, 1994).

Virgin Anticline

The Virgin anticline is a 30-mile-long (50 km), open, upright, northeast-trending, symmetrical fold that, along with the Kanarra anticline, effectively marks the eastern limit of significant Sevier-age compressional deformation in southwest Utah. The limbs of the fold generally dip 25 to 35 degrees, providing spectacular exposures of Triassic and Jurassic strata, and the fold itself is made all the more visible by hogbacks underlain by the resistant Shinarump Conglomerate Member of the Chinle Formation (figure 6). The anticline has three structural domes along its length. From south to north these are Bloomington Dome, Washington Dome, and Harrisburg Dome, each of which exposes the gypsum-bearing Lower Permian Harrisburg Member of the Kaibab Formation. East-dipping axial reverse faults at Harrisburg Dome (figure 41) and at Washington Dome place Lower Permian strata on top of Lower Triassic beds (Biek, 2003a; Hayden, 2005). The Virgin anticline appears to die out just southwest of Bloomington Dome, yet oddly is on strike with a structural saddle (Anderson and Barnhard, 1993a) between the North Virgin Mountains (of northwest Arizona) and Beaver Dam Mountains.

The Virgin anticline plunges to the northeast at about 10 to 15 degrees near Leeds. Several subsidiary folds—including the Leeds anticline, Leeds syncline, and other smaller folds and their associated thrust and high-angle faults—complicate the structure of the anticline's northeast-plunging nose (figures 42 and 43) (Biek, 2003a, 2003b; Biek and Rohrer, 2006). Proctor (1948, 1953) was the first to recognize the largest and westernmost of these thrust faults amid considerable controversy over structural interpretations of the Silver Reef mining district. Proctor's recognition of this and other thrust and high-angle faults in the district was the key to understanding that just one sandstone bed—the Springdale Sandstone Member of the Kayenta Formation—hosts the silver ore deposits of the mining district. The presence of these subsidiary folds on the plunging nose of the Virgin anticline suggests that, although the Virgin and Kanarra anticlines are co-linear and doubtless genetically related, they are separate structures—not simply the same fold cut by the Hurricane fault zone. Dramatic differences in fold geometry between the Virgin and Kanarra anticlines also suggests that they are separate structures. Small subsidiary folds are also present on the southwest-plunging southern end of the Virgin anticline, on the southern flanks of Bloomington Dome.

The Virgin anticline may have formed above a blind thrust fault in Cambrian strata (Davis, 1999), possibly in Late Cretaceous time (Hurlow and Biek, 2003). The age of formation of the Virgin anticline and subsidiary folds is difficult to determine because of inadequate cross-cutting relationships. The early Late Cretaceous Iron Springs Formation is the youngest bedrock unit involved in folding of the

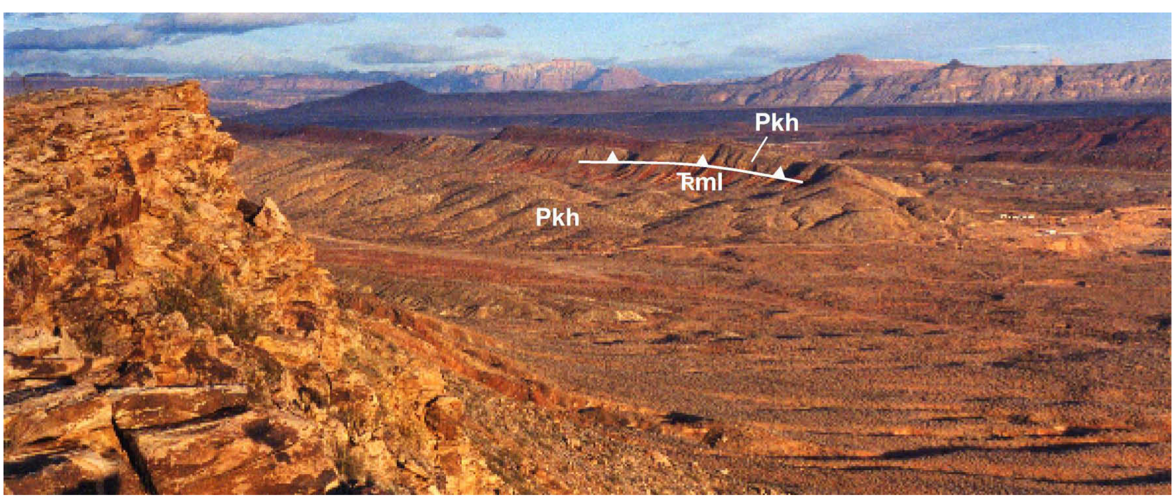

Figure 41. *Harrisburg Dome, one of three structural domes along the Virgin anticline, exposes the gypsum-bearing Harrisburg Member of the Kaibab Formation (Pkh). A west-directed back thrust is present on the east side of the dome, placing Harrisburg strata over the lower red member of the Moenkopi Formation (TRml). The Virgin anticline is the easternmost Sevier thrust belt structure in southern Utah, and it records much less shortening than the Square Top Mountain thrust shown on figure 40.*

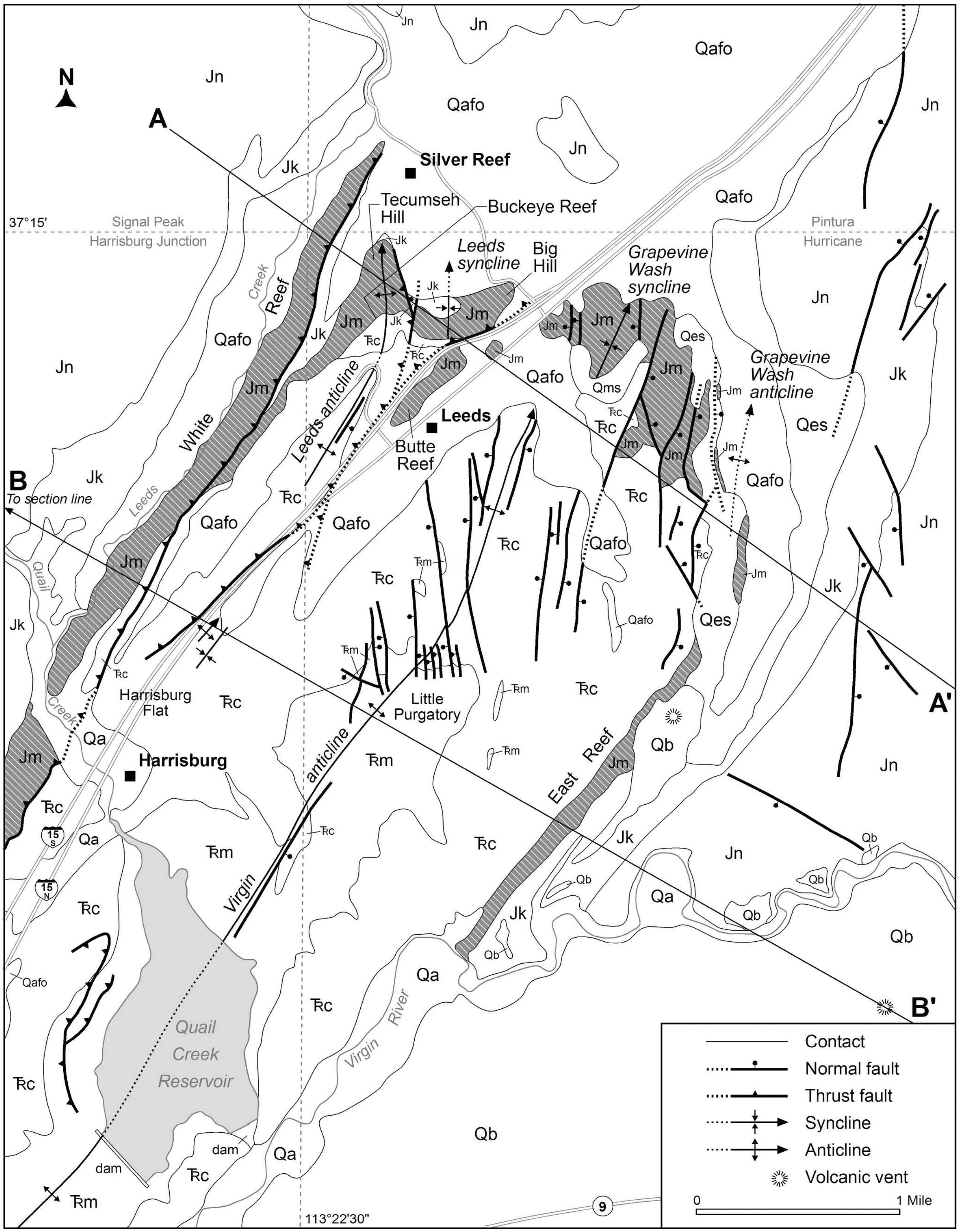

Figure 42. *Simplified geologic map of the Silver Reef mining district, showing Moenave strata repeated by thrust faults. Note that the Moenave Formation on this map (Jm) includes the Springdale Sandstone (now part of the Kayenta Formation). Compare to figure 27. TRm, Moenkopi Formation; TRc, Chinle Formation; Jk, Kayenta Formation; Jn, Navajo Sandstone. Quaternary units include Qa, undifferentiated alluvium; Qafo, old alluvial-fan deposits; Qes, eolian sand; and Qb, undifferentiated basaltic lava flows. Harrisburg Junction, Hurricane, Pintura, and Signal Peak quadrangle boundaries also shown. Cross sections A-A' and B-B' shown on figure 43. From Biek and Rohrer (2006).*

Virgin anticline. However, we suggest that the formation of the Virgin anticline is related to the Pintura anticline, which is exposed to the north in the Pintura quadrangle. The Pintura anticline is unconformably overlain by (and thus predates) the Canaan Peak Formation, the oldest beds of which may be Late Cretaceous (late Campanian) in age. The Virgin anticline thus may have formed between early and late Campanian time (about 84 to 71 million years ago), the youngest known age of the Iron Springs Formation and the oldest known age of the Canaan Peak Formation, respectively (Goldstrand, 1992, 1994), but both ages are poorly constrained. This uncertainty is compounded because similar thrust structures developed well into the Tertiary in central Utah (see, for example, Lawton and others, 1997; Willis, 1999; DeCelles and Coogan, 2006).

Kanarra Anticline

The Kanarra anticline extends nearly 40 miles (65 km) from near Toquerville on the south to near Cedar City on the north (Gregory and Williams, 1947; Threet, 1963; Kurie, 1966; Averitt, 1967; Averitt and Threet, 1973; Hurlow and Biek, 2003). Only the east limb of the fold is exposed along most of its length; the fold's crest and its west limb are sheared off by the Hurricane fault and down-dropped to the west where they are concealed by younger deposits. The east limb of the fold is overturned near Toquerville (Hurlow and Biek, 2003) and near Kanarraville (Averitt, 1967), but is elsewhere upright with beds on the east limb dipping between 20º and 35º east, shallowing abruptly as they pass under the great cliffs of Navajo Sandstone in the vicinity of Kolob Canyons, and under Smith Mesa farther south. Part of the west limb of the fold is exposed near Pintura.

The east limb of the Kanarra anticline is complicated by the Taylor Creek thrust fault zone, a west-directed back thrust first understood and mapped in some detail by Kurie (1966). Along much of its length, the Taylor Creek thrust fault consists of two or three main thrust splays, best illustrated by repetition of the resistant, cliff-forming Springdale Sandstone Member of the Kayenta Formation in the Kolob Canyons part of Zion National Park (figure 44) (Biek, 2007a). At the south end of Black Ridge, the Taylor Creek back thrust places Permian Fossil Mountain (Kaibab) strata on the Triassic lower

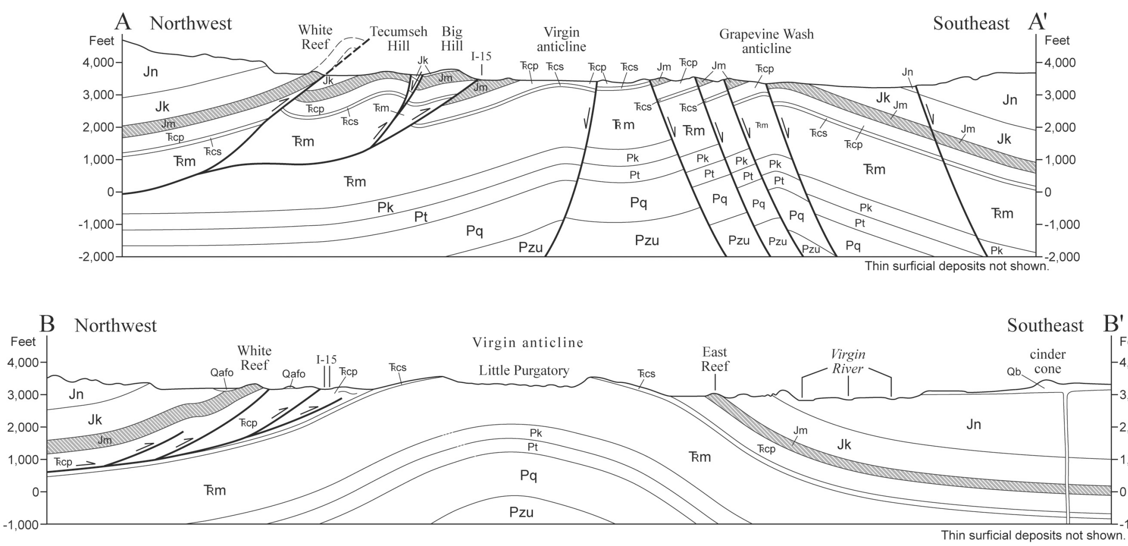

Figure 43. *Simplified cross sections of the Silver Reef mining district. Location of cross sections shown on figure 42. See figure 42 for identification of unit labels. From Biek and Rohrer (2006).*

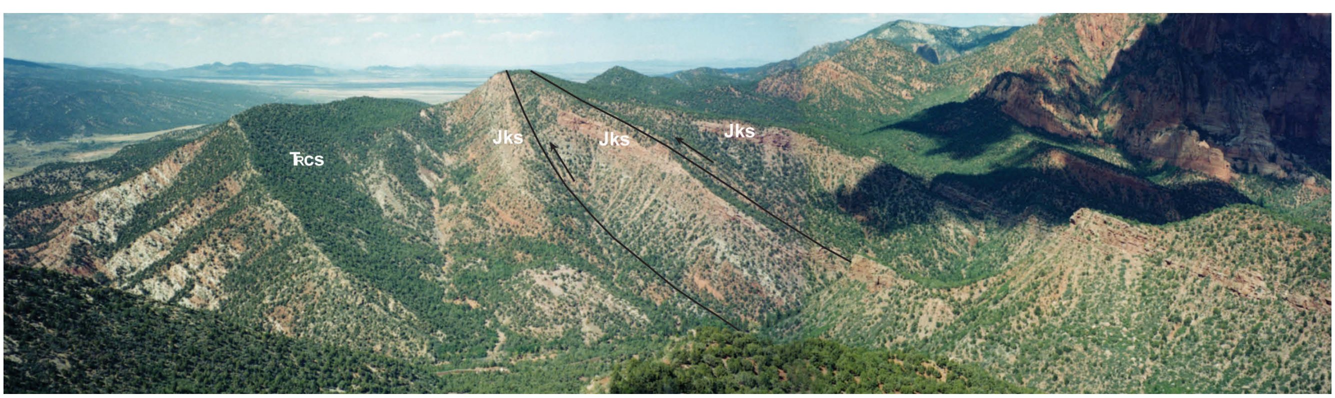

Figure 44. *View north across Taylor Creek canyon of the Taylor Creek thrust fault on the east limb of the Kanarra anticline. Two main splays of the west-directed back thrust result in three repeated east-dipping sections of the Moenave and basal Kayenta Formations. Hogbacks of Springdale Sandstone Member of the Kayenta Formation (Jks) and Shinarump Conglomerate Member of the Chinle Formation (TRcs) are labeled.*

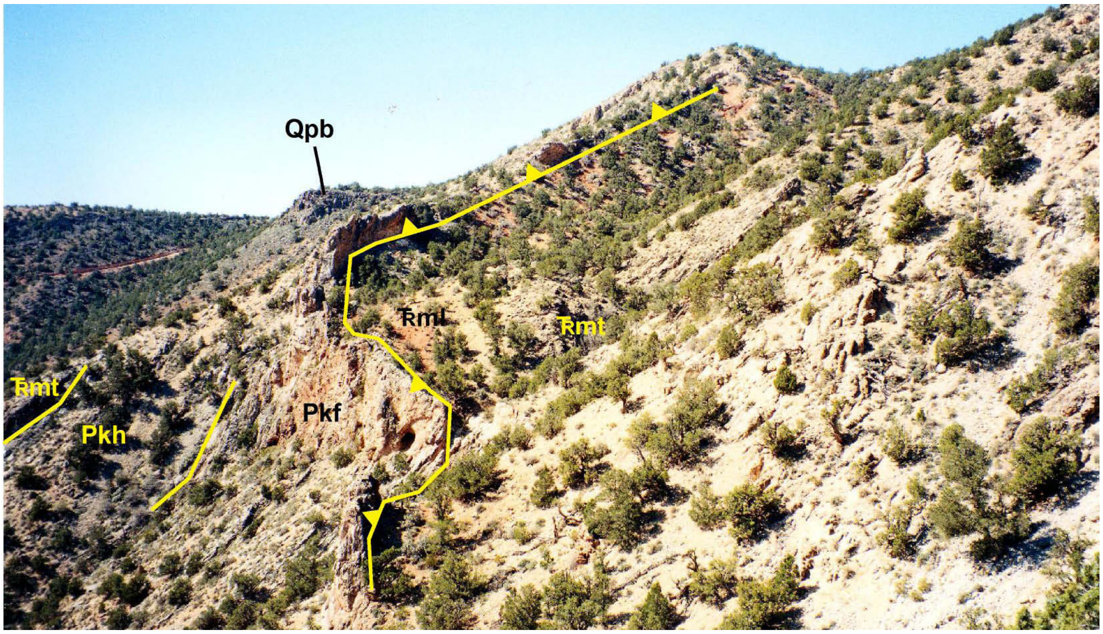

Figure 45. *View southwest of the Fossil Mountain Member of the Kaibab Formation (Pkf) at the south end of Black Ridge, just northeast of Toquerville. Here, near-vertical Fossil Mountain strata are thrust westward over the lower red member of the Moenkopi Formation (TRml); this is the southward continuation of the Taylor Creek thrust fault. Pkh, Harrisburg Member of the Kaibab Formation; TRmt, Timpoweap Member of the Moenkopi Formation. A small remnant of the Pintura basaltic lava flow (Qbp) is also visible.*

red member of the Moenkopi Formation on the vertical east limb of the fold (figure 45), and smaller thrust faults are also present (figure 46) (Hurlow and Biek, 2003). Between these two areas, the Taylor Creek thrust fault zone is mostly concealed by the Pintura lava flow and large landslide complexes, thus complicating interpretations as described by Biek (2007a). Kurie (1966) estimated at least 2000 feet (600 m) of vertical and about 2500 feet (760 m) of horizontal displacement on the Taylor Creek thrust fault (likely determined at Taylor Creek). Depending on the inferred geometry of the thrust faults, horizontal shortening along these faults may be two to three times what Kurie (1966) depicted.

As mentioned above, the Hurricane fault zone and axial plane of the Kanarra anticline are nearly coincident. Here, it appears that the blind thrust over which the Kanarra anticline formed was reactivated as part of the Hurricane fault zone.

Figure 46. *View north to small west-directed back thrust in the Virgin Limestone Member of the Moenkopi Formation on the east limb of the Kanarra anticline, near the south end of Black Ridge.*

The Kanarra anticline exposes Lower Permian to Upper Cretaceous strata, and the age of this fold is probably linked to the development of the nearby Pintura structural culmination, possibly during the early to late Campanian, about 84 to 71 million years ago (Hurlow and Biek, 2003). Grant and others (1994) depicted the Kanarra anticline as a fault-propagation fold related to an east-directed thrust fault in the subsurface; we envision that this is the thrust in Cambrian Bright Angel Shale inferred by Davis (1999). The Kanarra anticline is parallel and en echelon (stepped east) to the Virgin anticline to the south, and it too marks the eastern limit of significant Sevier-age compressional deformation in southwest Utah.

Pintura Anticline

The Pintura anticline is mapped west of Interstate 15 along the east flank of the Pine Valley Mountains. Hurlow and Biek (2003) summarized evidence for and previous interpretations of the fold, whose age provides important constraints on the timing of Sevier-age folding in southwestern Utah. In the north-central Pintura quadrangle, the Canaan Peak and Claron Formations depositionally overlie the Navajo Sandstone, and the Claron Formation overlies the Carmel Formation just to the northeast and west. The Cannan Peak and Claron Formations were, therefore, deposited above an angular unconformity developed on an erosionally beveled structural culmination. Where the Canaan Peak and Claron Formations unconformably overlie the Navajo Sandstone, erosion had previously removed more than 4000 feet (1200 m) of stratigraphic section, including all of the Middle Jurassic Temple Cap and Carmel Formations and the Upper Cretaceous Iron Springs Formation. This unconformity establishes the age of the structural culmination as between early and late Campanian time (about 84 to 71 Ma), the youngest known age of the Iron Springs Formation and the oldest known age of the Canaan Peak Formation, respectively (Goldstrand, 1992, 1994), but both dates are poorly constrained and are derived from outside the quadrangle. Why deformation associated with the Sevier orogeny apparently ended earlier in southwest Utah than it did in central Utah (where thrusting is documented into the early Eocene) is unknown; although we lack radiometric or fossil evidence, it may be that the Canaan Peak Formation is younger than previously thought.

The existence of the Pintura anticline as a Late Cretaceous compressional structure, however, has long been debated. We agree with the assessment of Cook (1960b) that direct evidence of the Pintura anticline as a Sevier-age structure lies hidden in the great, sweeping cross-beds of Navajo Sandstone, and that when the Claron Formation is rotated back to horizontal the Carmel Formation everywhere dips to the northwest. If a pre-Hurricane-fault Pintura anticline existed, its hinge is concealed beneath younger deposits on the downthrown side of the Hurricane fault. Hurlow and Biek (2003) reasoned that the Pintura anticline was a broad anticline whose east limb formed the west limb of a syncline that separates the Pintura and Kanarra anticlines. An alternate interpretation of the Pintura anticline is that it is simply the dismembered west limb of a structurally higher level of the Kanarra anticline modified by reverse drag on the Hurricane fault zone. In this latter interpretation, the two anticlines would not be separated by a buried syncline and the Navajo Sandstone would be at structurally deeper levels adjacent to the Hurricane fault zone than is shown on cross section B-B' (plate 2). On the map we place the axial trace of the Pintura anticline along the *present* dip reversal in the Carmel Formation in the north-central part of the Pintura quadrangle, and continue it south along the outcrop belt of Canaan Peak Formation, recognizing that these features occupy a broad, poorly defined, composite hinge zone. Either way, erosional beveling and deposition of the Canaan Peak Formation over the crest of the anticline occurred in the Late Cretaceous or possibly Early Tertiary. We interpret this hinge zone to have formed as a contractional fold, later modified in the axis of a large-scale reverse-drag fold in the hanging wall of the Hurricane fault. This large-scale reverse drag is to us the most likely explanation for the eastward tilting of the Claron Formation.

St. George Syncline

The St. George syncline is a broad, open fold with a poorly defined axial trace. We show the fold trending northeast, subparallel to the trace of the Virgin anticline, from southwest of St. George through the city of Washington, and into the Pine Valley Mountains. The fold axis is cut by the St. George and Washington fault zones, which, coupled with the gentle dip of beds, makes the fold axis difficult to locate. The fold involves rocks as young as the Iron Springs Formation, and we assume that it too formed during the Late Cretaceous or possibly Early Tertiary as a subtle fold behind the prominent Virgin anticline.

Small Folds

The many small, northeast-trending synclinal folds southwest of St. George, commonly associated with circular collapse structures, are likely due to gypsum dissolution (figure 47) (Higgins, 1997). The age of these folds is not well constrained, but sinkholes in Virgin River terrace gravel deposits suggest that dissolution is an ongoing process where gypsiferous strata are in proximity to ground water.

Figure 47. *In this part of the White Hills, the gypsum-bearing Harrisburg Member of the Kaibab Formation, exposed in this hillside, contains many small circular collapse structures that are commonly associated with small synclines. Most of these unusual features are probably due to gypsum dissolution, although some may be breccia pipes that resulted from deeper-seated dissolution and collapse in the underlying Mississippian Redwall Limestone. The resistant medial limestone of the Harrisburg Member forms the exposed top of this collapse feature.*

Late Tertiary and Quaternary Extensional Structures of the Transition Zone

Hurricane Fault Zone

The Hurricane fault is a major, active, steeply west-dipping normal fault that stretches at least 155 miles (250 km) from south of the Grand Canyon northward to Cedar City (figure 48). The fault lies at or near the base of the Hurricane Cliffs, which in the map area form an impressive fault scarp several hundred feet high (figure 14). Conspicuous, west-tilted, faulted slivers of mostly Triassic and Jurassic red beds are at the base of the cliffs, and contrast strongly with gray Permian carbonates. Several Pleistocene basaltic lava flows flowed across and are now offset by the fault zone, dramatically recording long-term slip rates as described below.

Like most long normal faults, the Hurricane fault zone is composed of discrete segments that tend to rupture independently. Parts of two of these segments, the Anderson Junction and Ash Creek segments, are in the map area (figure 48) and are linked by a structurally complicated segment boundary near Toquerville (Stewart and Taylor, 1996; Biek, 2003b; Hurlow and Biek, 2003; Lund and others, 2007a). The segment boundary coincides with the south end of the Kanarra anticline, a Late Cretaceous frontal fold of the Sevier orogenic belt, which is spectacularly exposed at the south end of Black Ridge (figure 49). An equally dramatic relay ramp, which exposes a much-faulted yet complete panel of northwest-dipping Moenkopi Formation, is at the north end of the Anderson Junction segment (figure 50). Total stratigraphic separation generally increases northward along the fault, from 800 to 1300 feet (250–400 m) at the Colorado River in the Grand Canyon (Karlstrom and others, 2007) to more than 8000 feet (2450 m) along the north part of the fault, but estimating such displacement is complicated as described below.

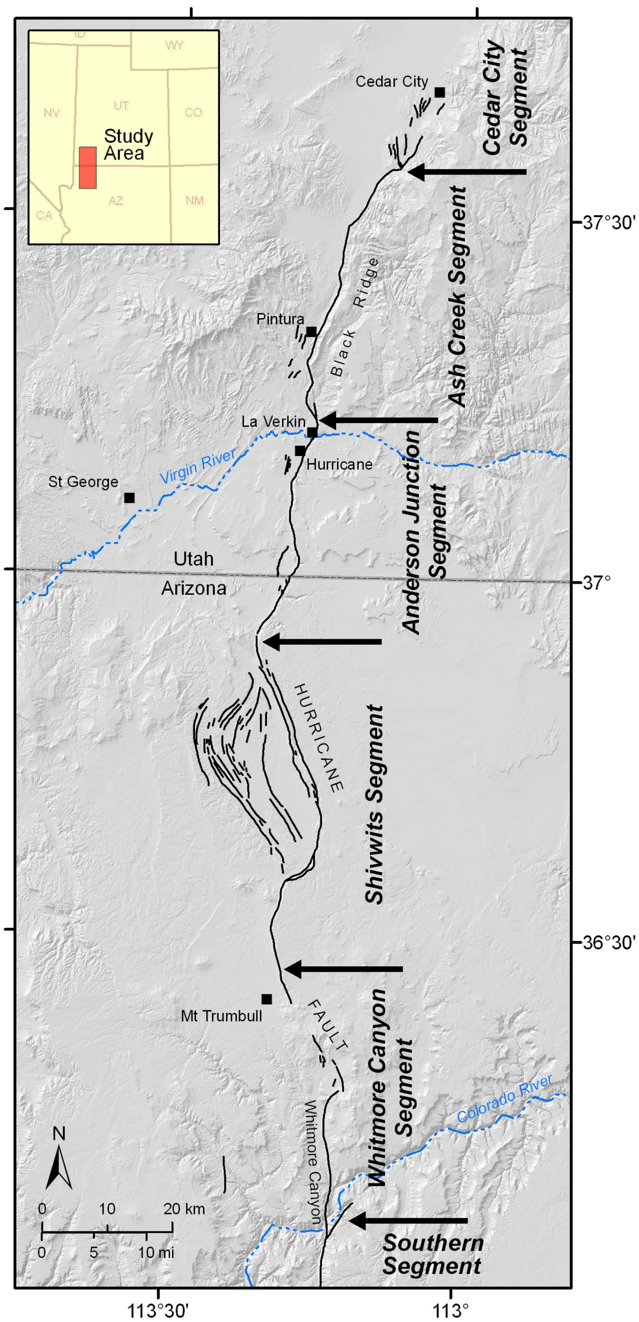

Figure 48. *The six segments of the Hurricane fault zone, each of which has a different rupture history and rate of long-term slip; arrows indicate segment boundaries. Part of the Anderson Junction segment and most of the Ash Creek segment are within the St. George 30'x60' quadrangle. From Lund and others (2007a).*

The paleoseismicity of the Hurricane fault zone was investigated by Lund and others (2001, 2002, 2007a), who noted that the most recent surface faulting event on the fault in Utah occurred in the latest Pleistocene or early Holocene, at the north end of the fault near Cedar City. They further noted that multiple surface faulting earthquakes have occurred in the late Quaternary along most, if not all, of the Utah portion of the fault. Based on dated basaltic lava flows that crossed and are now offset by the fault, Lund and others (2007a) calculated an average slip rate of about 8 inches per 1000 years (0.21 mm/yr) for the Anderson Junction segment of the Hurricane fault since about 350,000 years ago (figure 51). Near Ash Creek Reservoir, on the Ash Creek segment of the Hurricane fault, they similarly documented an average slip rate of about 22 inches per 1000 years (0.57 mm/yr) since about 850,000 years ago for that segment of the fault. The fault is considered capable of generating damaging earthquakes of about magnitude 7.0; the 1992 magnitude (M_L) 5.8 St. George earthquake is thought to have occurred on the west-dipping subsurface projection of the Hurricane fault (Arabasz and others, 1992; Pechmann and others, 1995).

Published estimates of normal separation on the Hurricane fault in Utah vary by an order of magnitude, from 1410 to 13,120 feet (430–4000 m) (Anderson, 1980). This discrepancy arises in part from measurements taken along different segments of the fault, but is also due to failure to subtract from the apparent, or stratigraphic, throw: (1) Late Cretaceous folding associated with the Sevier orogeny, (2) reverse-drag flexure of the hanging wall, and (3) rise-to-the-fault flexure in the footwall (Anderson and Christenson, 1989); it also reflects the fact that hanging-wall strata are largely covered by basaltic lava flows and surficial deposits, thus leading to uncertainty in hanging-wall relationships along lines of cross section. To avoid these complications, Anderson and Mehnert (1979) measured the top of the Navajo Sandstone on either side of the Hurricane fault at a sufficient distance from the fault to be representative of block interiors, and found a tectonic displacement (throw) of just 2000 to 2800 feet (600–850 m) at about the latitude of Pintura (see, for example, Swan and others [1980] for a discussion of apparent displacement versus tectonic displacement). Anderson and Christenson (1989) revised these estimates and found tectonic displacements of about 3600 feet (1100 m) and 4900 feet (1500 m) at the latitudes of St. George and Toquerville, respectively, with which we concur.

The Hurricane fault has been called a normal dip-slip fault (Huntington and Goldthwaite, 1904; Gardner, 1941; Cook, 1960b; Averitt, 1962; Hamblin, 1965, 1970; Kurie, 1966; Stewart and Taylor, 1996; Billingsley and Workman, 2000; Hurlow and Biek, 2003; Biek, 2003b, 2007a), a reverse fault (Lovejoy, 1964), a fault having a significant component of left-lateral slip (Moody and Hill, 1956; Anderson and Barnhard, 1993a, 1993b), and some have suggested that the fault is in part a reactivated Sevier-age thrust fault (Hintze, 1986; Huntoon, 1990). The different ideas stem in part from workers studying different parts of the fault and in part from our increased understanding of the fault zone through subsequent geologic mapping. It may also be that the fault zone has a more complicated history than previously envisioned.

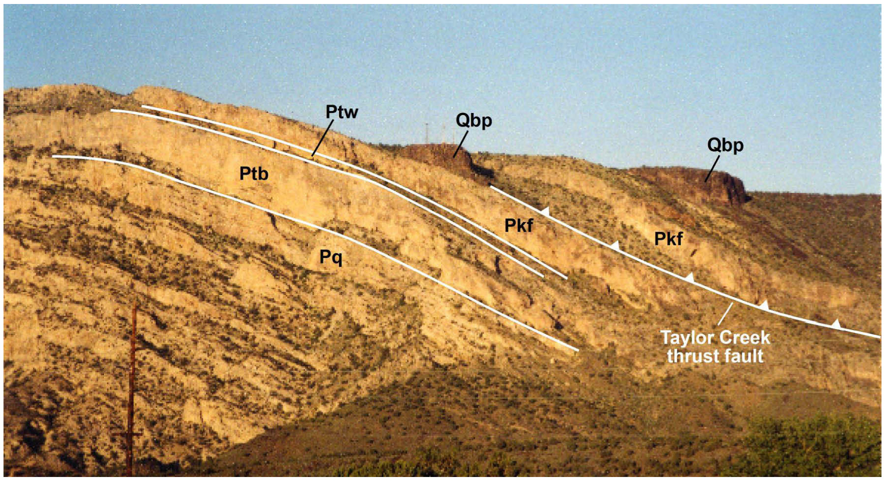

Figure 49. *The south end of Black Ridge, showing the ledgy slopes of the Queantoweap Sandstone (Pq), the cliff-forming Brady Canyon Member (with a narrow slope of Seligman Member below, Ptbs) and slope-forming Woods Ranch Member (Ptw) of the Toroweap Formation, and the cliff-forming Fossil Mountain Member of the Kaibab Formation (Pkf). Note how Fossil Mountain strata are repeated by a west-directed back thrust, the Taylor Creek thrust fault. Two isolated remnants of the 850,000-year-old Pintura lava flow (Qbp) are also visible.*

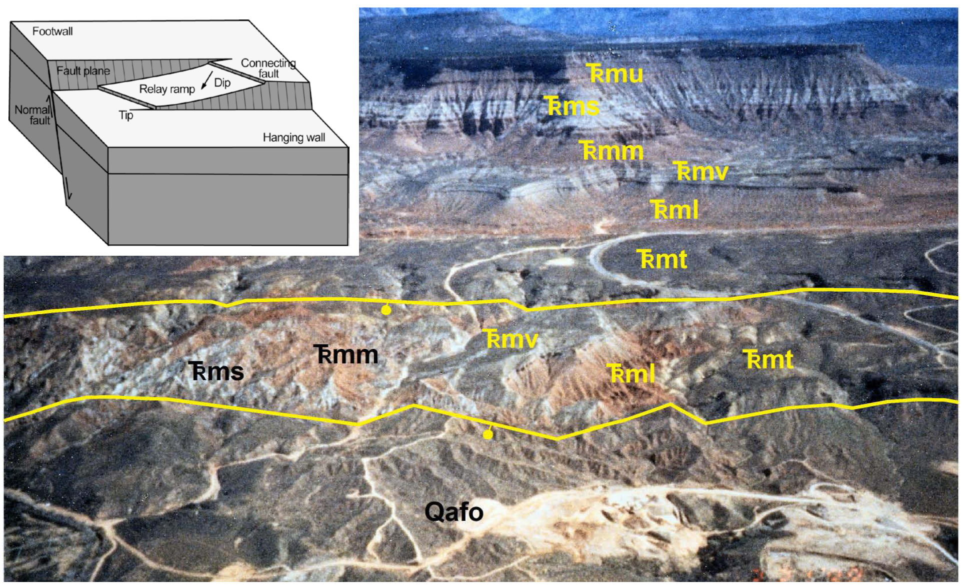

Figure 50. *Oblique aerial photograph (view east) of the relay ramp between two parts of the Hurricane fault zone. Inset block diagram shows the main features of a relay ramp (modified from Peacock and Sanderson, 1994). Highway 9 loops in and out of the photo on the right, and a gravel pit in old alluvial-fan deposits (Qafo) is in the foreground. Qafo conceals complexly faulted and folded Triassic, Jurassic, and Cretaceous strata exposed to the south in the Virgin River canyon and to the north near LaVerkin Creek. Hurricane Mesa, with its flat-lying Moenkopi strata capped by the Shinarump Conglomerate Member of the Chinle Formation (TRcs), is in the distance. Members of the Moenkopi Formation are: TRmu, upper red; TRms, Shnabkaib; TRmm, middle red; TRmv, Virgin Limestone; TRml, lower red; and TRmt, Timpoweap.*

Anderson and Barnhard (1993a, 1993b) summarized evidence showing that the northern part of the Hurricane fault zone—the Cedar City segment and possibly the Ash Creek segment—has a component of left-lateral displacement, and that near or just north of the latitude of the Pine Valley Mountains, this left-lateral displacement is transferred westward to the Gunlock–Reef Reservoir–Grand Wash fault zone. This northern part of the Hurricane fault zone closely follows the axial trace of the Kanarra anticline, suggesting that this part of the fault may also be a reactivated Sevier-age thrust fault. Also, the northern part of the fault zone forms the east structural margin of the New Harmony and Cedar Valley basins, whose strata show basin-bounding faulting began in the Miocene, about 20 million years ago (Hurlow, 2002; Rowley and others, 2006). In the Cedar City area, Rowley and others (1978) suggested that a west-facing topographic highland, whose west edge is generally coincident with the Hurricane scarp, blocked or partly blocked Great Basin-derived ash-flow sheets of 26 to 20 million years old from flowing east of the Hurricane fault more than about 10 miles (15 km).

However, Pliocene to Quaternary displacement on the Anderson Junction segment of the Hurricane fault is normal dip-slip, and we see no evidence suggesting this part of the Hurricane fault is a

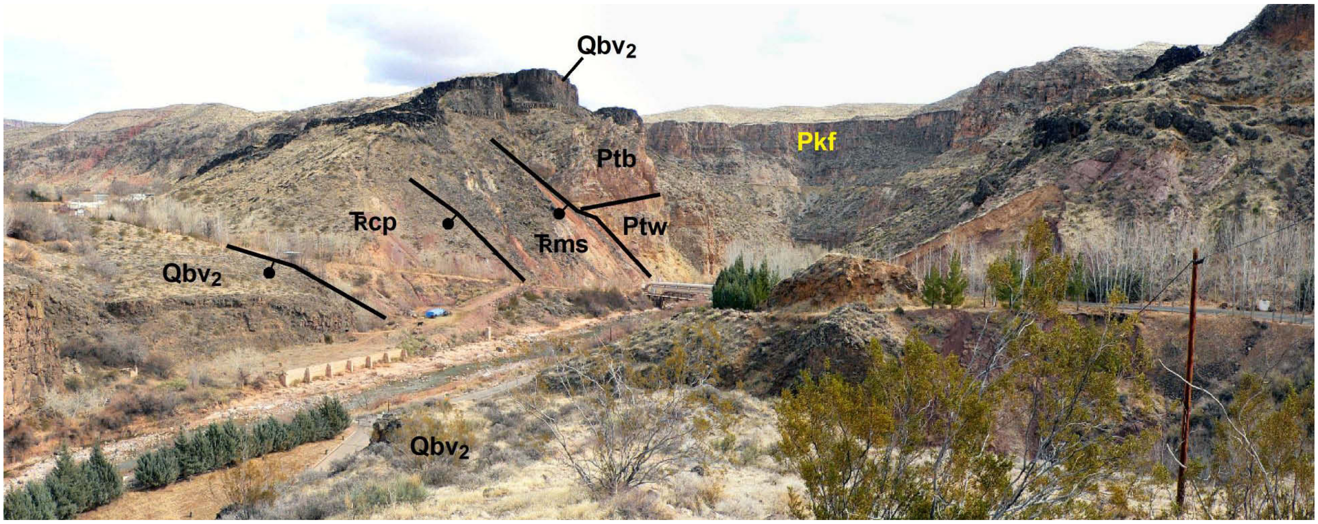

Figure 51. *View northeast to the entrance of Timpoweap Canyon, where the Virgin River exits the Hurricane Cliffs. Three main strands of the Hurricane fault zone are shown. Here, the 350,000-year-old Volcano Mountain lava flow (Qbv_2) is offset about 240 feet (73 m), yielding an average slip rate of about 8 inches per 1000 years (0.21 mm/yr) for this strand of the fault. The fault zone is characterized by a west-dipping panel of Triassic strata caught between splays of the fault; these red beds act as a barrier to ground-water flow, causing warm water to rise through permeable Permian strata in the footwall and emerge as Pah Tempe Hot Springs. TRcp, Petrified Forest Member of the Chinle Formation; TRms, Shnabkaib Member of the Moenkopi Formation; Ptw and Ptb, Woods Ranch and Brady Canyon Members of the Toroweap Formation, respectively; Pkf, Fossil Mountain Member of the Kaibab Formation. The Harrisburg Member of the Kaibab Formation and the Timpoweap Member of the Moenkopi Formation form the rolling hills on the skyline.*

reactivated Sevier-age structure. Indeed, the Hurricane fault trends due south, away from the southwest trend of the Virgin anticline. The inception of faulting on the Anderson Junction segment can only be constrained between Late Cretaceous and middle Quaternary time. Stewart and Taylor (1996) considered it to have begun as early as late Miocene or early Pliocene based on a displacement rate determined from Quaternary basalt offset by the Hurricane fault. This estimate, however, is based on total stratigraphic separation and not tectonic displacement as described above and may thus yield an inappropriately old age for initiation of faulting.

Anderson and Christenson (1989) considered the Hurricane fault to be a Pliocene to Quaternary feature. We concur with this age assignment for most offset along the fault, but consider it likely that initiation of faulting was earlier, at least on the Cedar City and possibly Ash Creek segments. It probably started by about 10 million years ago, the age of basaltic volcanism in northern Cedar Valley (Rowley and others, 2006), and possibly by about 20 million years ago based on distribution of regional ash-flow tuff sheets (Rowley and others, 1978). However, the southern segments of the fault are clearly younger. Billingsley and Workman (2000) described offset relationships of late Tertiary and Quaternary basaltic lava flows in Arizona and showed that, based on equal offset of lava flows and underlying Mesozoic strata, most normal faults on the Shivwits and Uinkaret Plateaus became active after 3.6 and possibly after 2.6 million years ago. Thus, it appears that the Ash Creek and Cedar City segments of the Hurricane fault zone have a longer, more complicated history than the southern segments.

In the Kolob Arch quadrangle, highly fractured, deeply weathered, fault-bounded blocks of the Claron Formation(?) and quartz monzonite porphyry are exposed as fault-bounded blocks in the Hurricane fault zone at the base of the Hurricane Cliffs. These exposures suggest that rocks of the Pine Valley latite or laccolith and the Claron Formation at least locally underlie the New Harmony basin west of the Hurricane fault, and once extended east of the fault. If the Pine Valley laccolith once extended east of what is now the Hurricane fault, it could have served as a nearby source of Pine Valley quartz monzonite megaboulders now scattered on the Kolob Plateau (figure 52) (Averitt, 1962, 1967; Anderson and Mehnert, 1979; Biek and others, 2003; Biek, 2007b; Biek and Hylland, 2007). Alternatively, these megaboulders may have been deposited by a northeastward-moving gravity slide or debris flow along high-gradient streams from the upper slopes of the Pine Valley laccolith, which still, despite being in the down-dropped block of the Hurricane fault zone, towers over the Hurricane scarp.

Figure 52. *Large quartz monzonite boulder, probably derived from the ancestral Pine Valley Mountains, near the east shore of Kolob Reservoir. The presence of these boulders documents a complete reversal of drainage in southwest Utah, from east-flowing to west-flowing streams following the inception of Basin and Range extension, as described in the "Evolution of the Modern Landscape" section of this report.*

Washington Fault Zone

The Washington fault is at least 40 miles (65 km) long, stretching from northern Arizona to Washington City. It bifurcates into several splays north of Washington City and may link up with the Washington Hollow fault to the north, although it is difficult to trace the fault through massively cross-bedded and jointed Navajo Sandstone. Stratigraphic separation decreases northward along the fault from 2200 feet (600 m) about 6 miles (10 km) south of the border, to about 1650 feet (500 m) at the state line, to about 700 feet (200 m) near Washington City (Hayden, 2005).

One fault scarp, about 5 miles (8 km) south of the Warner Valley turnoff, cuts older colluvial and alluvial sediments that, based on scarp profiling and comparison to dated scarps elsewhere, Anderson and Christenson (1989) considered to be latest Pleistocene in age. Just north of the Virgin River, several scarps on splays of the fault cut the 900,000-year-old Washington lava flow. Other prominent scarps in the quadrangle seem to be the result of differential erosion, a conclusion reached by Peterson (1983) and Anderson and Christenson (1989). The Utah Geological Survey began a paleoseismic study of the fault in 2007; initial trenches across the fault zone near Washington Dome uncovered preliminary evidence for three surface-faulting events about 70,000, 30,000, and 18,000 years ago.

The Washington fault truncates the southwest part of Washington Dome (figure 53). Fault drag created narrow, anticlinal footwall folds in the Moenkopi Formation and Shinarump Conglomerate Member of the Chinle Formation from Washington Dome south into Arizona. One of the largest and best defined of these drag folds is Beehive Dome, just north of Fort Pearce Wash.

Late Tertiary and Quaternary Extensional Structures of the Basin and Range

Geologic structures in the west part of the map area, including the Gunlock–Reef Reservoir–Grand Wash fault zone, Beaver Dam Mountains culmination, Shivwits syncline, Shebit anticline, Jackson Wash–Pahcoon Flat fault, and other structures are wonderfully exposed but terribly puzzling. Interpretations of their origin continue to be the subject of considerable disagreement as to the relative influence of Late Cretaceous Sevier-age compressional deformation and late Tertiary and Quaternary extension associated with development of the Basin and Range. For example, many researchers interpreted the principal folds of the Beaver Dam Mountains culmination and the Shivwits syncline as a result of Late Cretaceous compressional deformation (Dobbin, 1939; Reber, 1951; Moore, 1972; Steed, 1980; Hintze, 1986; Carpenter and others, 1989; Carpenter and Carpenter, 1994; Christie-Blick and others, 2007). Wernicke and Axen (1988) and Axen and Wernicke (1989) concluded that some structures were related to Late Cretaceous compression and that others, including the main Beaver Dam culmination, were due to basin-range extension. Through detailed mapping of key structural elements in the Beaver Dam Mountains and adjacent area, Anderson and Barnhard (1993a, 1993b) concluded that none of the structures in question resulted from Late Cretaceous compressional deformation.

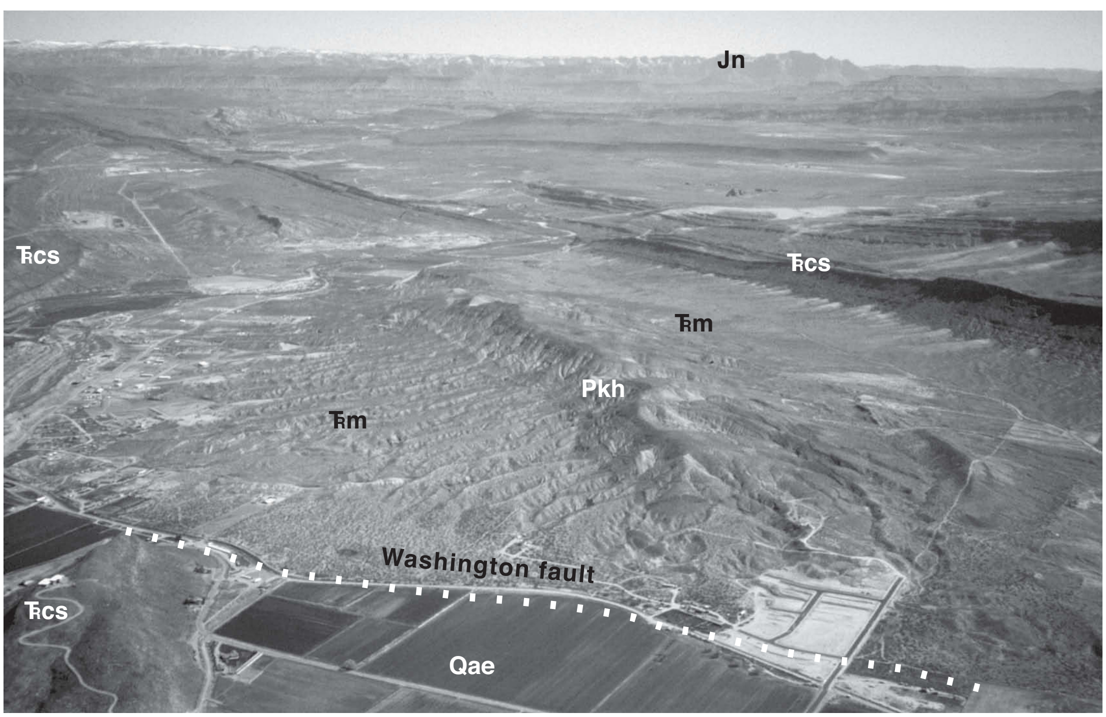

Figure 53. *Aerial view northeast to Washington Dome and the concealed trace of the Washington fault. Note how the fault truncates Washington Dome. TRcs, Shinarump Conglomerate Member of the Chinle Formation; TRm, Moenkopi Formation; Pkh, Harrisburg Member of the Kaibab Formation; Qae, mixed alluvial and eolian deposits. The towering cliffs of Navajo Sandstone (Jn) on the horizon are in Zion National Park.*

These structures—unusual folds and faults that exhibit puzzling offset relationships along strike, including the Beaver Dam mountains culmination, Shivwits syncline, and Jackson Wash-Pahcoon Flat fault zone—are best explained as a result of late Tertiary and Quaternary displacement on left-lateral oblique-slip faults that are consistent with a regional pattern of significant left-lateral strike- and oblique-slip faulting at the east margin of the Basin and Range, in combination with footwall uplift associated with the Piedmont–Red Hollow range-front fault zone that created the main Beaver Dam culmination (figure 54) (Anderson and Barnhard, 1993a, 1993b).

Anderson and Barnhard (1993b, p. 37) suggested an elegant model that describes the formation of the Beaver Dam culmination and associated structures, which we quote here. “We suggest that the formation of this culmination can be likened to resting one’s right arm (Colorado Plateau) and fist (palm side down, the future Beaver Dam Mountains) on a table and flexing the fist upward. In the process, the knuckles and fingers (extension-parallel corrugations beneath the Castle Cliff detachment) move toward the arm as the proximal slope of the fingers decreases. The wrist produces a synclinal or homoclinal fold in which the skin is locally crumpled (Shivwits syncline). At what stage the corrugations form is not known, but they could represent early ductile footwall shortening normal to the extension direction of the same deformation field represented by late strike-slip faults that displace the flexure axis. Rocks in the ‘hanging wall’ above the knuckles are severely extended and attenuated, and record extreme extension across the culmination. The magnitude of attenuation and extension appears to have been about proportional to the magnitude of uplift and shortening. The deformation was highly localized and involved critical elements of in-place rotations around subhorizontal axes.”

Gunlock–Reef Reservoir–Grand Wash Fault Zone

The Gunlock–Reef Reservoir–Grand Wash fault zone is widely regarded as forming the west boundary of the transition zone

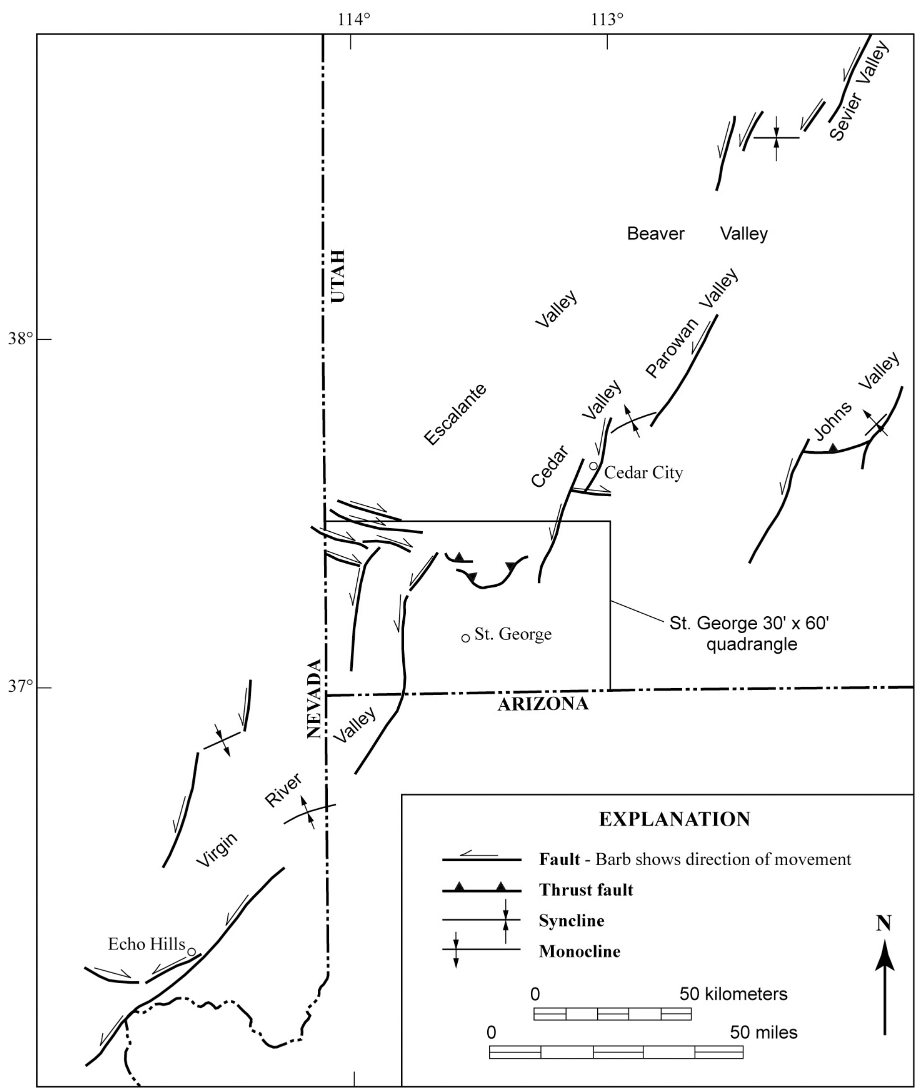

Figure 54. Map of the southeastern Great Basin showing traces of selected faults having known or suspected oblique-slip displacement, associated thrust faults, and folds. Modified from Anderson and Barnhard (1993b).

between the Basin and Range and Colorado Plateau provinces in southwest Utah (see, for example, Anderson and Barnhard, 1993a, 1993b; Karlstrom and others, 2007). Both the Gunlock and Grand Wash faults exhibit down-to-the-west offset of well-documented late Tertiary age (Lucchitta, 1979; Anderson and Barnhard, 1993a, 1993b; Bohannon and others, 1993; Billingsley and Workman, 2000; Faulds and others, 2001). The Reef Reservoir fault, however, shows down-to-the-east offset; it is inferred to be late Tertiary in age based on its association with the Gunlock fault and uplifted Beaver Dam Mountains (Anderson and Barnhard, 1993a, 1993b). All three faults are steeply dipping, but exactly how they are connected is not well understood. The Gunlock fault appears to be also linked to the Antelope Range fault near Mountain Meadows at the north edge of the map area (Rowley and others, 2006).

Gunlock fault: Hintze and others (1994) and Anderson and Barnhard (1993a, 1993b) showed that the Gunlock fault has a significant component of left-lateral displacement. The fault is associated with Tobin Wash basin, a northeast-trending synextensional synclinal basin north of Gunlock whose basin-margin strata are coarser and more steeply dipping than central-basin strata (Anderson and Barnhard, 1993a). The basin is coeval with left-lateral slip on the Gunlock fault, and records shortening normal to the regional extension direction. Displacement on the Gunlock fault is mostly late Tertiary in age based on the fault's association with strata of Tobin Wash basin. Most displacement on the Gunlock fault appears to predate the 1.6 million-year-old Gunlock–Dameron Valley North lava flow, which crosses the fault and shows little displacement (Hintze and others, 1994; Higgins, 2002). Offset relationships of Mesozoic strata suggest a maximum stratigraphic separation near Gunlock, but determining the magnitude of this offset is complicated by the fact that critical exposures are concealed by the Veyo and other basaltic lava flows. (Outcrop patterns of the Middle Jurassic Carmel Formation suggest that these lava flows may conceal the hinge zone of an anticline whose axis trends north and northeast parallel to the Gunlock fault but which cannot be traced through the massively cross-bedded Navajo Sandstone.) The apparent left-lateral, down-to-the-west stratigraphic offset may exceed 15,000 feet (5000 m) near Gunlock, but displacement dies out rapidly to the south where it is partly accommodated by folding of the Shivwits syncline (Anderson and Barnhard, 1993b). The Gunlock fault dies out directly south of the Santa Clara River, in what Reber (1951) called the Shebit anticline.

Reef Reservoir fault: The Reef Reservoir fault is on strike and overlaps slightly with the Gunlock fault, and is a steeply dipping reverse fault that bounds the southeast-plunging nose of the Shivwits syncline (figure 55). The Reef Reservoir fault trends from just south of the Santa Clara River southward into Mine Valley (Hintze, 1986; Hammond, 1991; Hintze and Hammond, 1994). All previous workers interpreted the Shivwits syncline as a Sevier-age compressional structure, but Anderson and Barnhard (1993a, 1993b) set forth an interpretation of the Shivwits syncline involving left-lateral displacement on the Gunlock–Reef Reservoir–Grand Wash fault zone in combination with structural crowding associated with coeval eastward tilting of the Beaver Dam Mountains culmination. In effect, left-lateral displacement on the Gunlock fault dies out southward into the Shivwits syncline, and reverse displacement on the Reef Reservoir fault accommodates tight folding of the syncline's east limb.

Grand Wash fault: The west-side-down Grand Wash fault increases in displacement southward and forms the abrupt boundary between the Colorado Plateau and the Basin and Range at the Grand Canyon. Structural relief across the fault in the vicinity of the Grand Canyon exceeds 10,000 feet (3000 m) (Lucchitta, 1979); stratigraphic separation decreases to the north to about 1500 feet (450 m) at the state line, and essentially dies out just 5 miles (8 km) farther north near the Reef Reservoir fault (Hintze, 1986; Hammond, 1991; Hintze and Hammond, 1994). In the vicinity of the Grand Canyon, movement on the fault is restricted to middle Miocene, about 11 to 16 million years ago (Lucchitta, 1979; Bohannon and others, 1993; Faulds and others, 2001). The timing of movement on the Utah part of the fault is unrestricted due to lack of late Tertiary deposits.

Shivwits Syncline

The northward-plunging axial trace of the Shivwits syncline extends for 9 miles (15 km) directly west of and parallel to the Reef Reservoir fault and the southern part of the Gunlock fault (figure 55). These faults bound the steeper east limb of the syncline. Hintze (1986) and Hintze and Hammond (1994) interpreted the fold as a Late Cretaceous compressional feature, suggesting that its coincidence with the Gunlock–Reef Reservoir faults was due to superposition of later extensional structures over earlier compressional features. However, the association of the faults and syncline led Anderson and Barnhard (1993a, 1993b) to suggest that the fold is a result of left-lateral oblique slip on the Gunlock and Reef Reservoir faults, but they also noted that the fold is a larger feature than relative fault displacement alone would suggest. Anderson and Barnhard (1993a, 1993b) thus proposed a model wherein the fold is a combined result of uplift of the Beaver Dam Mountains, uplift on an inferred convex-upward Pahcoon Flat fault, and left-lateral displacement on the Gunlock–Reef Reservoir faults (figures 55 and 56). In effect, the Shivwits syncline developed as a result of left-lateral displacement on the Gunlock–Reef Reservoir fault zone and structural crowding against the flank of the Beaver Dam Mountains culmination, with deformation dying out in the reverse sense of displacement observed on the Reef Reservoir fault. The steep west dip and structural thinning of beds on the east limb of the fold are due in part to fault-tip folding that preceded rupture of the Reef Reservoir fault.

Jackson Wash–Pahcoon Flat Fault Zone

The Jackson Wash–Pahcoon Flat fault zone, on the northeast flank of the Beaver Dam Mountains culmination, is the most enigmatic structure in the map area (figure 55). The following discussion is summarized from Anderson and Barnhard (1993a, 1993b). The Pahcoon Flat fault is composed of two parts, a northwest-striking part that is mostly buried by surficial deposits of Pahcoon Flat and a southern part that cuts down-section to the

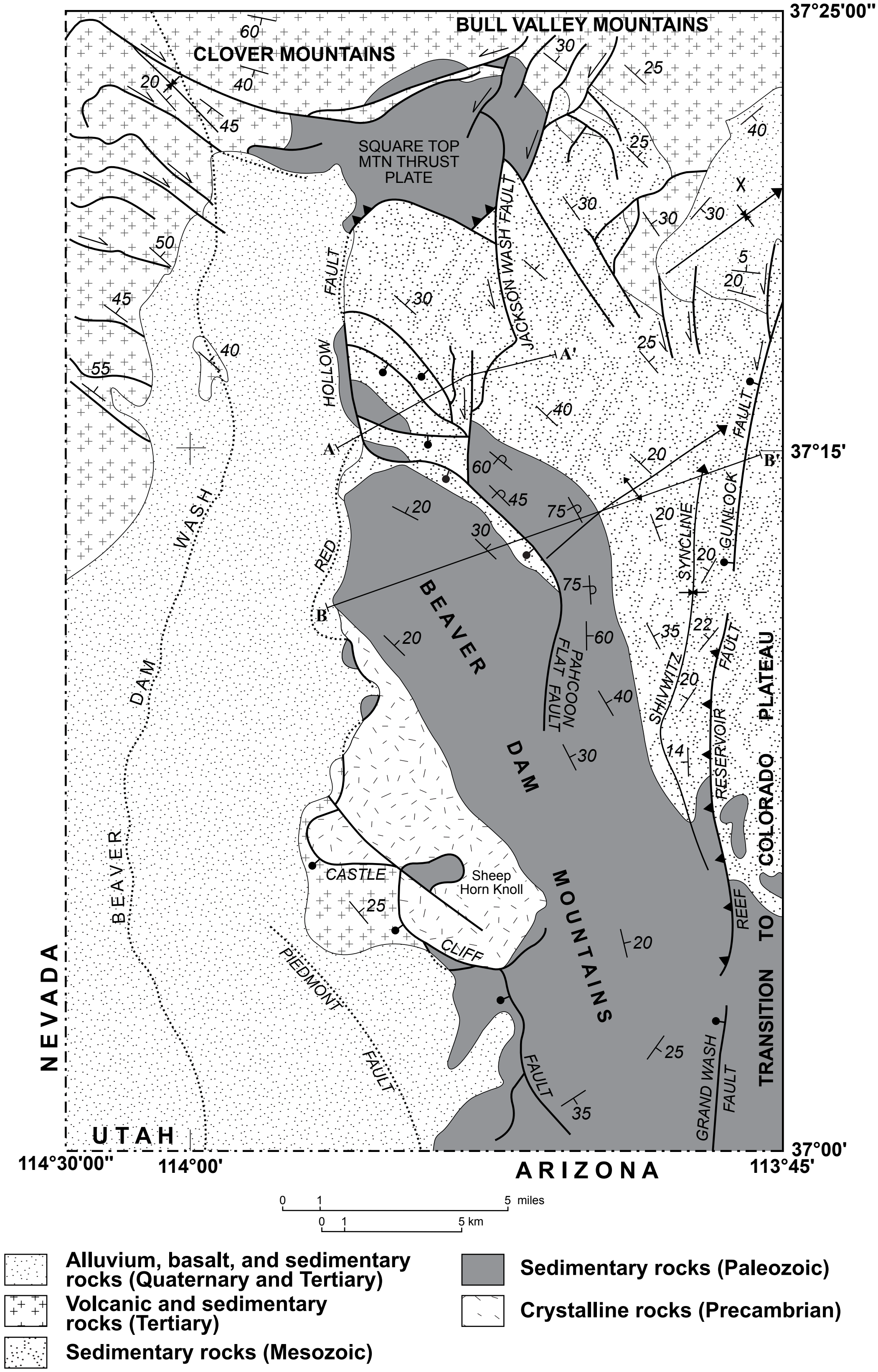

__Figure 55.__ Geologic map of the northern Beaver Dam Mountains area; 'X' in the upper right corner of the map marks the Tobin Wash basin. Cross sections A-A' and B-B' are shown on figure 56. The Shivwits syncline, long interpreted by many as a Late Cretaceous structure that formed during the Sevier orogeny, is here interpreted to be the result of left-lateral displacement on the Gunlock–Reef Reservoir fault zone and structural crowding against the flank of the Beaver Dam Mountains culmination, with deformation dying out in the reverse sense of displacement observed on the Reef Reservoir fault. From Anderson and Barnhard (1993a).

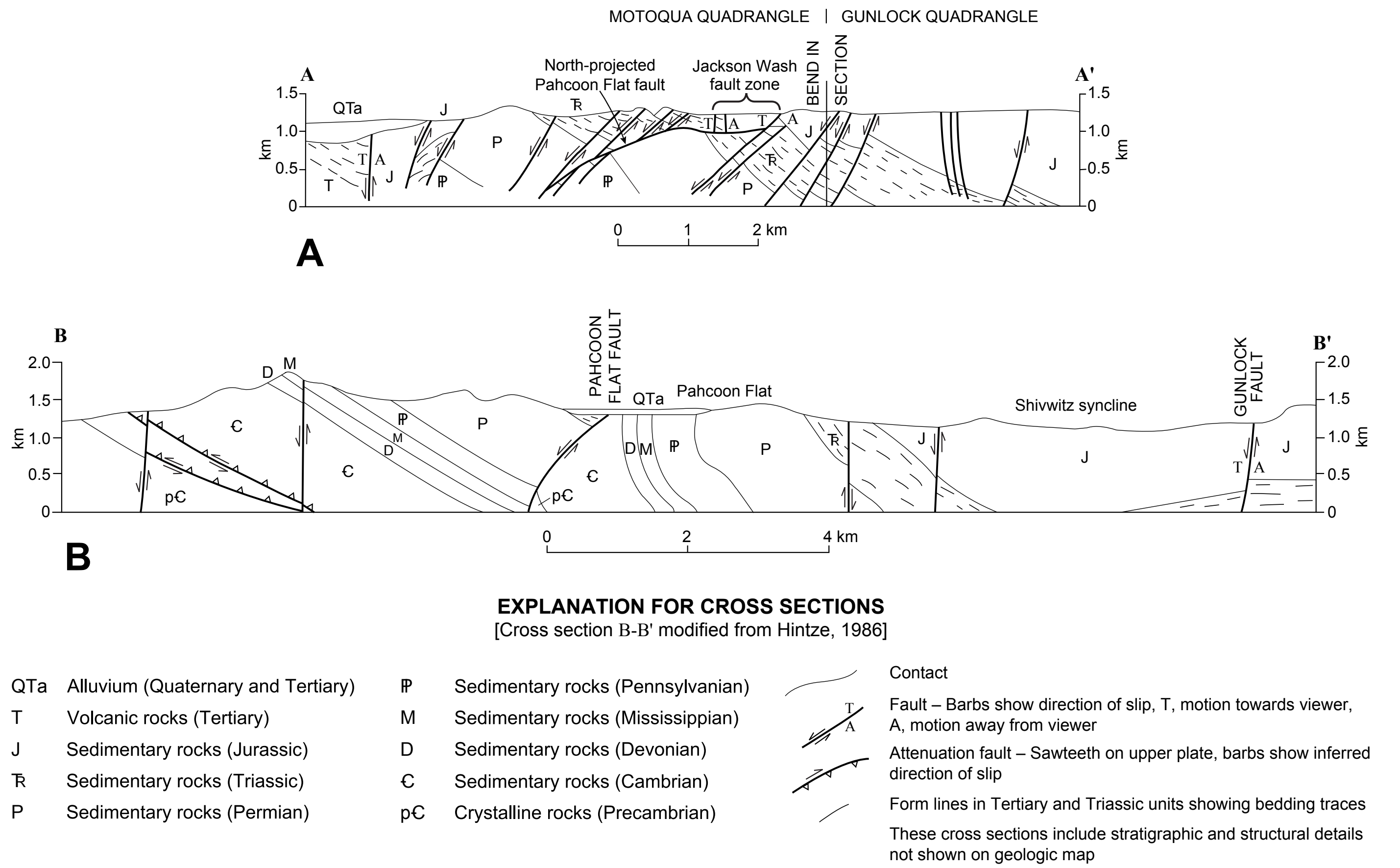

Figure 56. *Cross sections showing the inferred convex-upward shape of the Jackson Wash and Pahcoon Flat faults associated with footwall dips that are steeper than hanging-wall dips. Hanging-wall strata that have been eroded from above the vertically dipping footwall in section B-B' are assumed to have been extended and thinned similar to, but more intensely than, those preserved above the Jackson Wash fault as shown in section A-A'. Location of cross section shown on figure 55. From Anderson and Barnhard (1993a).*

south. The southern part of the fault is steeply dipping based on its relationship to topography, but its dip decreases northward and directly south of Pahcoon Flat is just 35 degrees west (Hintze, 1986). The sense of apparent displacement along the fault changes from down-to-the-east along the southern part of the fault to down-to-the-west along the northern part of the fault. Beds in the footwall of the fault are overturned along much of its length, and it is this aspect of the fault that is most puzzling.

The Jackson Wash fault trends north from directly west of Pahcoon Flat to the southern Bull Valley Mountains, where it appears to be linked to northeast-trending faults with left-lateral displacement. Several northwest-trending oblique-slip faults intersect but do not cut the hanging wall of the Jackson Wash fault along its southern trace, suggesting that they are mechanically coupled with the Pahcoon Flat fault. Outcrop patterns of Navajo, Temple Cap, and Carmel strata along the northern trace of the Jackson Wash fault reveal significant stratigraphic separation, but southward, displacement is distributed across a broad zone of structurally incompetent Chinle and Moenkopi strata in the hanging wall of the fault.

Anderson and Barnhard (1993a, 1993b) interpreted the Jackson Wash–Pahcoon Flat fault zone and associated footwall culmination as having an origin similar to that of the Castle Cliff fault and Beaver Dam Mountains culmination. That is, they suggested that the fault zone has a convex-upward shape, which may explain footwall rocks that dip more steeply than hanging-wall rocks (figures 55 and 56). An alternate interpretation, suggested here, is that the Jackson Wash–Pahcoon Flat fault zone developed as a result of broadly distributed right-lateral shear between the left-lateral oblique-slip faults of the Gunlock–Reef Reservoir–Grand Wash fault zone and the Piedmont–Red Hollow fault zone (figure 57). In this model, the northwest-trending part of the Pahcoon Flat fault serves as a gently southwest-dipping relay ramp between the more steeply dipping Jackson Wash fault and the north-trending part of the Pahcoon Flat fault. The overturned beds in the Pahcoon Flat area are interpreted to be a result of structural crowding on the northeast flank of the Beaver Dam culmination in combination with continued movement on the Gunlock–Reef Reservoir fault. Neither interpretation is completely satisfactory, and the Jackson Wash–Pahcoon Flat fault zone remains one of the most enigmatic features in the map area.

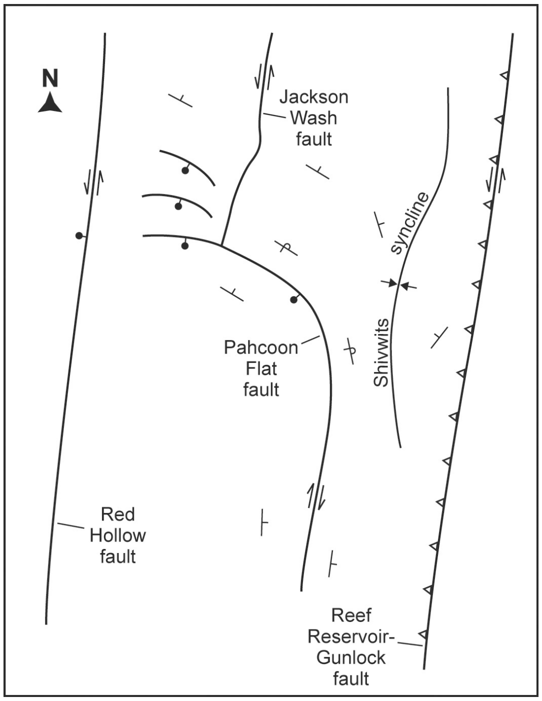

Figure 57. *In this model, the Jackson Wash–Pahcoon Flat fault zone is interpreted to have developed as a result of broadly distributed right-lateral shear between the left-lateral oblique-slip faults of the Gunlock–Reef Reservoir fault zone and the Red Hollow fault zone. Compare to figure 55 for area covered by this sketch.*

Beaver Dam Mountains Culmination

The Beaver Dam Mountains are a structural culmination cored by Paleoproterozoic crystalline rocks overlain by a well-exposed section of Paleozoic and Mesozoic sedimentary rocks. The overall structure of the range is an internally faulted and folded, northeast- and east-dipping homocline bounded on the east by the Shivwits syncline and the Gunlock–Reef Reservoir–Grand Wash fault zone (figure 58). The range is bounded on the west by the Piedmont–Red Hollow fault zone and by the Castle Cliff fault. The Beaver Dam Mountains culmination is on structural trend with similar structural culminations in the Virgin Mountains to the south in northwest Arizona and adjacent Nevada that Anderson and Barnhard (1993a) referred to as the Black Ridge, north Virgin, and Mount Bangs culminations, each cored by crystalline basement rocks. A structural saddle with 11,500 to 13,000 feet (3500–4000 m) of relief separates the Beaver Dam and Mount Bangs culminations (Anderson and Barnhard, 1993a).

Considerable disagreement remains as to the age and interpretation of the Beaver Dam Mountains culmination. Dobbin (1939), Reber (1951), Moore (1972), Steed (1980), Hintze (1986), Carpenter and others (1989), and Carpenter and Carpenter (1994) interpreted the Beaver Dam Mountains culmination as a doubly plunging anticline, with significant structural complications on each nose, which formed during the Late Cretaceous Sevier orogeny. However, no compelling evidence was offered to support this interpretation, and the inferred east- to northeast-directed compression required to create such a fold is at odds with the southeast-directed compression evidenced by the Square Top Mountain thrust and frontal fold of the Virgin anticline. If the culmination formed in the Late Cretaceous as suggested by others, it implies that the Square Top Mountain thrust is a backward-breaking thrust and that the Virgin anticline, the frontal fold in this part of the fold-and-thrust belt, formed above a blind thrust in basement rocks, not in structurally incompetent Cambrian strata.

Wernicke and Axen (1988) suggested that the culmination resulted from late Tertiary isostatic uplift of the footwall of a low-angle normal fault, which they called the Castle Cliff detachment fault. Carpenter and others (1989) and Carpenter and Carpenter (1994), however, showed that the Castle Cliff fault is a rootless feature and that the Mesquite Basin west of the Beaver Dam Mountains is not a shallow basin floored by Proterozoic rocks as required by a rolling-hinge model of isostatic footwall uplift.

Christie-Blick and others (2007) summarized conflicting interpretations of these models, principally as they relate to low-angle normal faults and associated structures in the nearby and much better studied Mormon Mountains of southeast Nevada, and cited two major objections to the Beaver Dam Mountains culmination having resulted from late Tertiary isostatic uplift of the footwall of a low-angle normal fault: (1) Pennsylvanian strata in the footwall of the Pahcoon Flat fault are overturned (figure 55), correctly noted by many as at odds with a model of simple isostatic footwall uplift, and (2) the axial trend of the culmination is apparently oblique to and truncated by the range-bounding fault zone. Regarding point (1): strongly folded strata associated with the Pahcoon Flat fault are thought by some to demonstrate profound east-west compression, but as described previously, we suggest that they are best explained by structural crowding on the northeast limb of the culmination due to left-lateral oblique slip on the Gunlock-Reef Reservoir fault zone. Regarding point (2): the trend of the Beaver Dam Mountains and Virgin Mountains culminations is indeed parallel to the range-bounding Piedmont fault, as correctly noted by Axen and Wernicke (1989). Gravity data indicate a northwest-trending 18 mgal/km drop northwest of Castle Cliff that may reflect unconsolidated, low-density basin-fill sediment in the hanging wall of the northwest continuation of the Piedmont fault, which is parallel to the Beaver Dam Mountains culmination (Hintze, 1986). It is the Red Hollow fault that appears to truncate the axial trace of the culmination north of Welcome Spring, and this is accentuated by drag associated with left-lateral displacement on that fault.

As described below, the Castle Cliff fault and gravity-slide blocks document rapid and major uplift of the Beaver Dam Mountains during the late Miocene along the range-front Piedmont fault. Further uplift and exhumation of the range followed in the Pliocene and Quaternary—this later uplift may account for the apparent truncation of the axial trace of the culmination. In the interpretation of Anderson and Barnhard (1993a, 1993b), followed here, the Beaver Dam Mountains culmination is viewed as a synextensional structure of late Tertiary age.

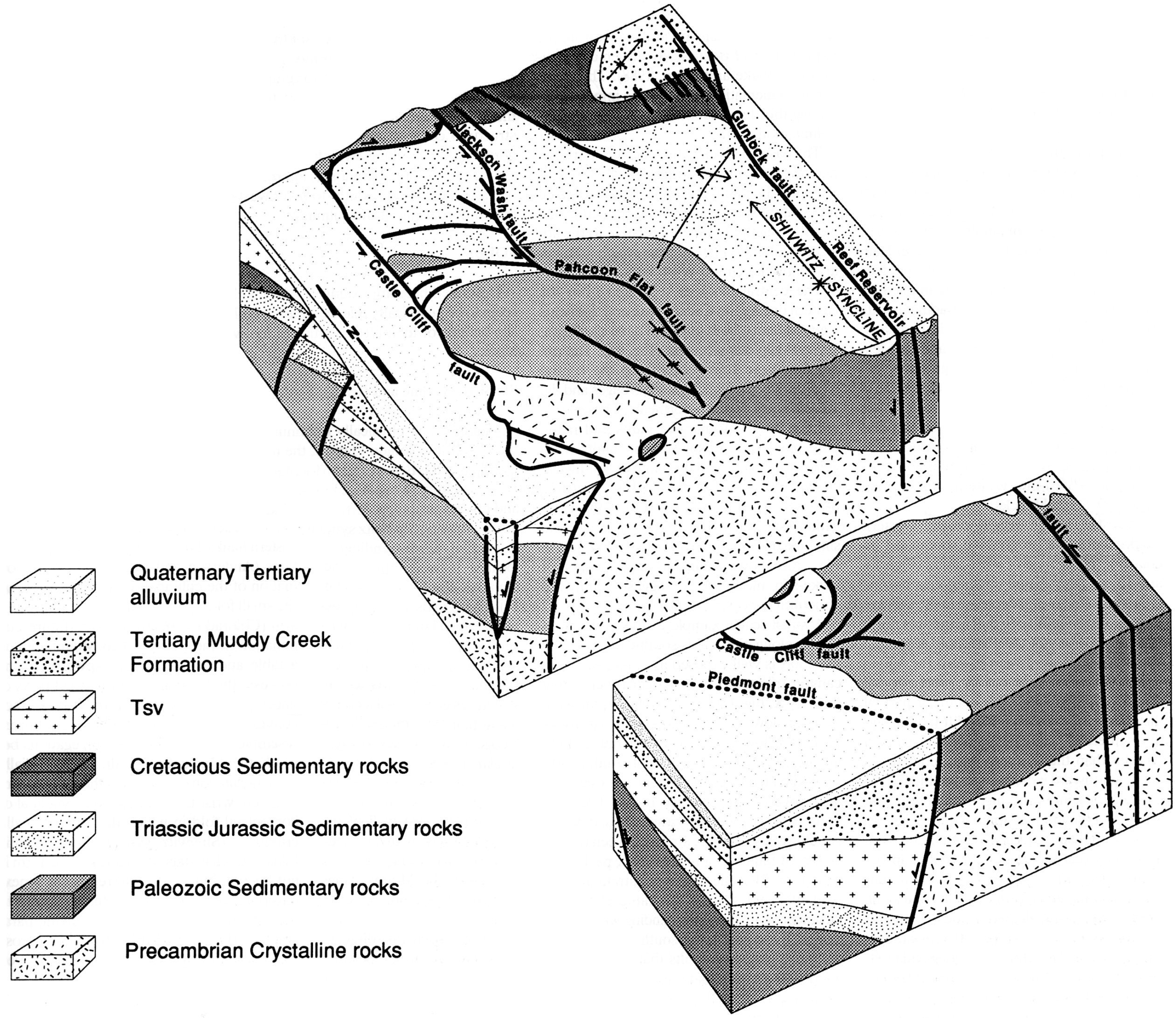

Figure 58. *Block diagram of the Beaver Dam Mountains showing interpreted convex-upward shape of the Castle Cliff fault, concave upward shape of the Piedmont fault zone, and planar shape of the predominately left-lateral Gunlock–Reef Reservoir fault zone. The north-trending folds are interpreted as resulting from structural crowding caused by uplift and tilting of the footwall of the Castle Cliff fault. The northeast-trending folds are interpreted as structures that formed in response to synextensional contraction normal to the extension direction coeval with oblique-slip faulting. Red Hollow fault not shown on this illustration from Anderson and Barnhard (1993a).*

O'Sullivan and others (1994) used two apatite fission-track ages from the Tapeats Sandstone as a proxy for evaluating the uplift history of the Beaver Dam Mountains. Modeling of those ages suggests rapid uplift of the range at about 17 million years ago at a rate of 1300 to 3300 feet per million years (400–1000 m/Myr) during the early Miocene, followed by a slower rate of 230 to 360 feet per million years (70–110 m/Myr) from the late Miocene to present, with a total uplift of at least 13,000 to 20,000 feet (4000–6000 m). Additional apatite fission-track ages determined from Lower Cambrian and Paleoproterozoic rocks in the Beaver Dam Mountains also suggest rapid exhumation beginning about 16 Ma (Stockli, 1999).

Piedmont Fault Zone

The Piedmont fault zone bounds the west side of the southern Beaver Dam Mountains and the Virgin Mountains (figure 55), trending south and southwest into adjacent Arizona and Nevada. It also forms the east side of the Virgin River depression, which consists of two deep basins separated by a buried ridge; part of the Mesquite Basin underlies the southwest corner of the map area, and seismic and gravity data show it to be one of the deepest basins in the Basin and Range Province, containing as much as 26,000 feet (8000 m) of Oligocene to Quaternary basin-fill deposits (Bohannon and others, 1993; Carpenter and Carpenter, 1994; Langenheim and others, 2001). The Piedmont fault is a major high-angle, down-to-the-west normal fault, is listric at depth (Bohannon and others, 1993; Carpenter and Carpenter, 1994), and may have significant left-lateral oblique slip (Anderson and Barnhard, 1993a, 1993b). The fault is also known as the Mesquite fault, which in adjacent Arizona and Nevada has fault scarps as much as 60 feet (18 m) high on early to middle Quaternary alluvial-fan deposits and as much as 13 feet (4 m) high on late Pleistocene fans, although Holocene and some latest Pleistocene fans are not faulted (Mayer, 1982).

Red Hollow Fault Zone

The Red Hollow fault zone trends north about 16 miles (25 km) from the heart of the Beaver Dam Mountains culmination to just north of the Square Top Mountain thrust, thus forming the east margin of a relatively shallow basin beneath northern Beaver Dam Wash (figure 55). We extend the fault farther south than did Anderson and Barnhard (1993a, 1993b) or Hintze and others (1994) based on the presence of Muddy Creek strata in the footwall of the fault north of Reber Wash. The northern part of the fault consists of two main splays that bound Permian strata. Hintze (1986) described the intriguing "fish hook" curve of Mesozoic strata in the footwall of the fault that he interpreted to be the result of normal fault drag on previously tilted strata (see plate 1). We interpret the "fish hook" and associated down-to-the-southwest faults to be the result of oblique, left-lateral slip on the steeply west-dipping Red Hollow fault, first suggested by Anderson and Barnhard (1993a, 1993b). The down-to-the-southwest faults on the east limb of the "fish hook" are localized in structurally incompetent Moenkopi strata and are believed to accommodate extension in the hanging wall of the Jackson Wash–Pahcoon Flat fault zone. Anderson and Barnhard (1993a, 1993b) reasoned that the buried hanging wall of the Red Hollow fault zone consists of a thick section of northeast-tilted and faulted Miocene volcanic strata, and suggested that, although the magnitude of left-lateral slip is unknown, Miocene and younger throw on the fault could exceed 2 miles (3 km).

Castle Cliff Fault

Paleozoic strata adjacent to the south and west sides of the crystalline core of the Beaver Dam Mountains are sheared, brecciated, and tectonically thinned compared with equivalent strata directly east of the range. Examples of the structurally thinned strata are well exposed at and near Castle Cliff, where most of the Cambrian section, which is over 5000 feet (1500 m) thick a few miles to the east, is missing, with only a few hundred feet of the Tapeats Sandstone and Bonanza King Formation preserved (figure 59). The massive, cliff-forming Redwall Limestone, normally as much as 850 feet (260 m) thick, is reduced to only about 50 feet (15 m) thick, and the overlying Callville Limestone is unusually thin as well. Other well-exposed, highly thinned sections of Paleozoic rock are found at Sheep Horn Knoll (Bonanza King Formation and Redwall Limestone) and northwest of Castle Cliff at the base of the Beaver Dam Mountains in fault contact with Proterozoic rocks (mostly Redwall Limestone, but also Bonanza King Formation, Muddy Peak Dolomite, and Tapeats Sandstone). Thinning occurred in the upper plate of a low-angle fault, the Castle Cliff fault. The thinned block at Sheep Horn Knoll has been dropped down-to-the-east by a northwest-trending normal fault first recognized by Cook (1960b).

The Castle Cliff fault dips gently west and south from 10° to 30°, and in plan view, as noted by Anderson and Barnhard (1993a, 1993b), exhibits a wavy or corrugated surface on a wavelength of 1 to 2 miles (2–4 km). It hugs the mountain front as a zone of intense structural thinning atop the main crystalline culmination of the Beaver Dam Mountains. Diehl and others (1996, 1997; see also Anderson and Diehl, 1999) provided evidence that much of the stratal attenuation in upper-plate rocks resulted from protracted dissolution during uplift and tilting of the Beaver Dam Mountains culmination. The fact that the Castle Cliff fault is preserved at the base of the range at the core of the culmination suggests that it is no longer active but is being exhumed due to footwall uplift associated with the Red Hollow and Piedmont faults.

Hintze (1986) summarized early interpretations of the enigmatic Castle Cliff fault, which continues to inspire significant debate as to its origin. Dobbin (1939) and Reber (1951) first interpreted the thin upper-plate strata as part of an east-directed thrust sheet, as was then the standard interpretation of low-angle faults, even though the faults place younger strata over older strata, which is unusual for thrust faults. Cook (1960a) also interpreted the larger brecciated blocks at Sheep Horn Knoll and Castle Cliff as thrust blocks following Dobbin and Reber. Jones (1963), however, found kinematic indicators that showed the Sheep Horn Knoll and Castle Cliff blocks are west-directed features that he interpreted as gravity-slide blocks. Carpenter and others (1989) and Carpenter and Carpenter (1994) also interpreted a gravity-slide origin for these blocks, and discussed evidence against their origin as part of the upper plate of a large, rooted, low-angle detachment fault suggested by Wernicke and Axen (1988), Wernicke and others (1989), and Axen and others (1990). Anderson and Barnhard (1993a, 1993b) interpreted the fault to be a convex-upward fault zone and estimated a maximum of about 5 miles (8 km) of stratigraphic separation over the crest of the Beaver Dam culmination, suggesting that the fault accommodated extreme extension over the core of the Beaver Dam Mountains culmination. Christie-Blick and others (2007) summarized critical aspects of these conflicting interpretations and concluded that the Castle Cliff fault is a feature developed during landsliding.

Anders and others (2006) discussed distinguishing features of catastrophic gravity-slide blocks versus low-angle detachment-fault blocks in their study and reinterpretation of the structure of the nearby Mormon Mountains in southeast Nevada. They noted that the Mormon Mountain detachment lacks many characteristic features of seismically cycled faults, but instead displays many features of catastrophically emplaced gravity-slide blocks, including a basal conglomerate, clastic dikes, and lack of lower-plate deformation features. Echoing the work of others cited above and of Anders and others (1998), they also interpreted the Castle Cliff fault and its upper-plate rocks as formed by catastrophic landsliding, not by rooted, low-angle detachment faulting.

However, the Castle Cliff detachment fault and its thinned upper-plate strata have not been studied in detail as have similar structures in the Mormon Mountains. More detailed investigation of the fault surfaces and of the upper-plate rocks themselves is needed before we can confidently assess conflicting interpretations of catastrophically emplaced landslide blocks versus seismically cycled faults and tectonically thinned upper-plate strata. Thus, we treat the thinned upper-plate strata of the Castle Cliff fault

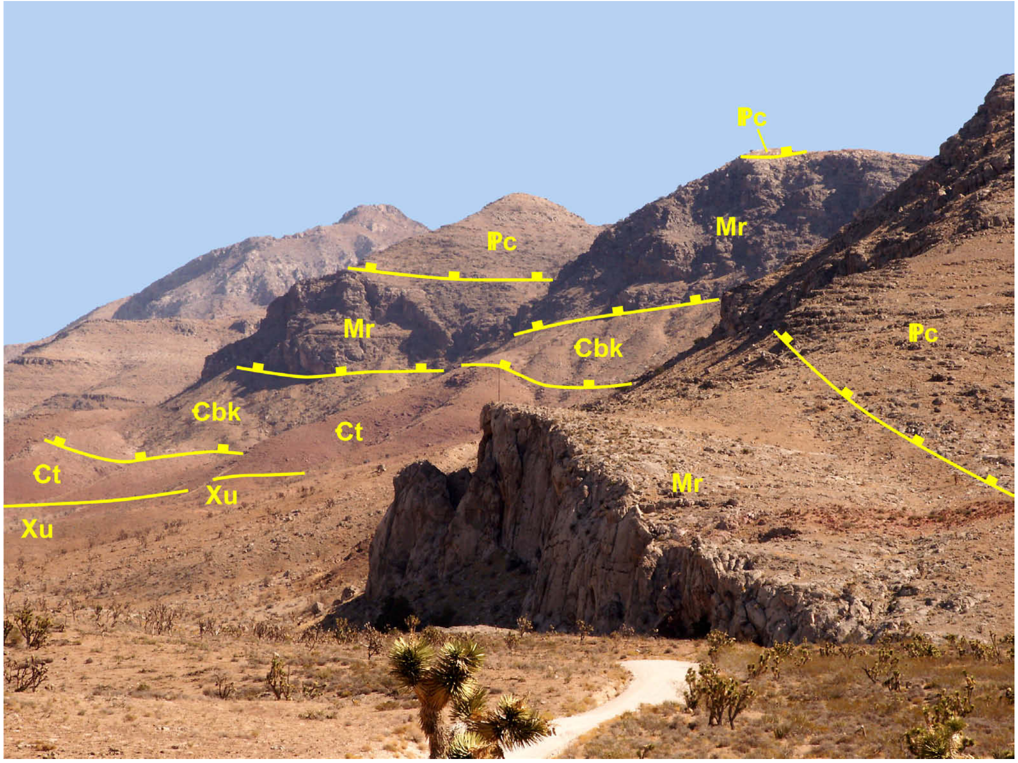

Figure 59a. *View east-southeast to Castle Cliff, showing highly attenuated lower Paleozoic section over metamorphic Paleoproterozoic rocks (Xu). Most of the Cambrian section is missing, and the normally massive, cliff-forming Redwall Limestone (Mr) is reduced to only about 50 feet (15 m) thick. This lower Paleozoic section has been tectonically thinned in the upper plate of the Castle Cliff fault. Ct, Tapeats Sandstone; Cbk, Bonanza King Formation; IPc, Callville Limestone.*

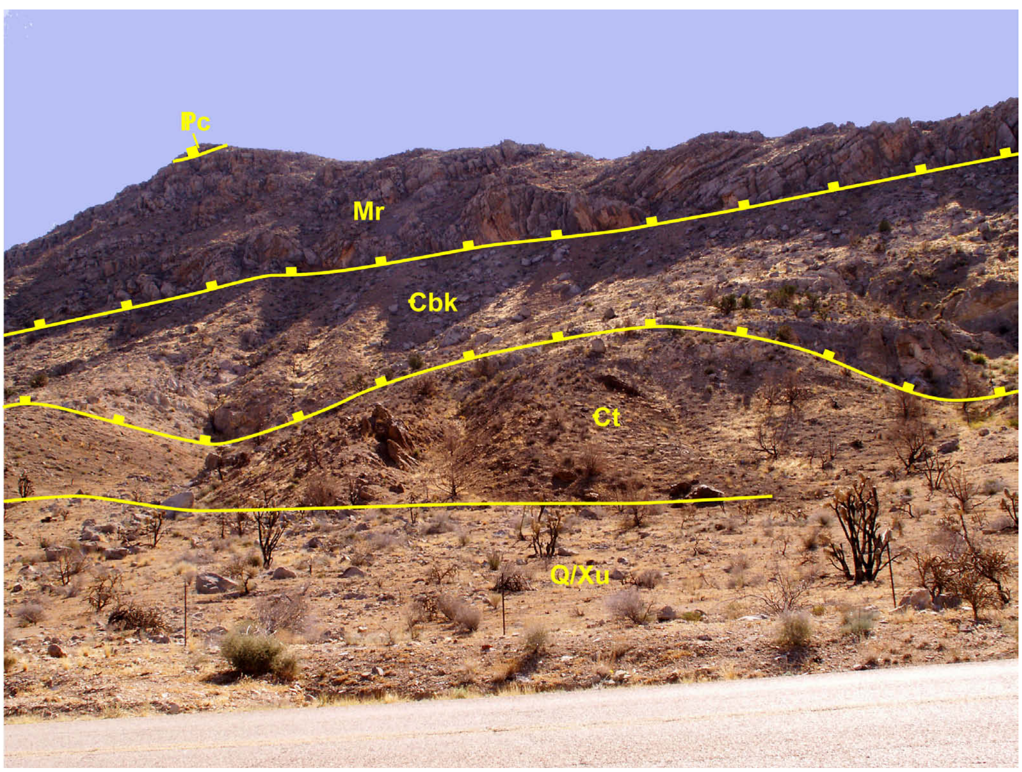

Figure 59b. *View southeast to upper plate of Castle Cliff fault just east of Castle Cliff. Note east-tilted block of Redwall Limestone (Mr) bounded by planar low-angle normal faults. Paleoproterozoic rocks in the foreground are mostly concealed by colluvium and alluvial-fan deposits (Q/Xu). Ct, Tapeats Sandstone; Cbk, Bonanza King Formation; IPc, Callville Limestone.*

and gravity-slide blocks as two separate entities: (1) the Castle Cliff fault and associated upper-plate strata, which we interpret as having accommodated extreme extension over the core of the Beaver Dam Mountains culmination, possibly in the upper plate of a convex-upward fault, and (2) bona fide gravity-slide blocks derived from the upper plate of the Castle Cliff fault that are incorporated in the Muddy Creek Formation, described below.

Gravity-Slide Blocks

Large blocks of severely brecciated and attenuated Mississippian Redwall Limestone derived from the upper plate of the Castle Cliff fault are present within basin-fill deposits west of the Beaver Dam Mountains (figure 60). These blocks and enclosing basin-fill deposits typically dip east, back towards the Beaver Dam Mountains. One of the best exposures of these gravity-slide blocks and underlying strata is in the SE1/4 section 36, T. 42 S., R. 19 W. where a large gravity-slide block of Redwall Limestone both overlies and is overlain by alluvial-fan deposits interpreted to be a coarse alluvial facies of the Muddy Creek Formation (figures 60 and 61). The gravels are subangular to subrounded, clast-supported, pebble- to small cobble-size Paleozoic carbonate clasts (although some clasts are as much as 1 to 3 feet (0.3–1 m) in diameter), and minor red silty sandstone clasts likely from the Kayenta or Moenkopi Formations, all set in a moderately well-cemented, reddish-brown, sandy matrix. Interestingly, no clasts of the Proterozoic crystalline basement are present in these deposits even though today Proterozoic rocks form the exposed bulk of the western Beaver Dam Mountains. A shear zone in Muddy Creek deposits directly underlies the brecciated block of Redwall Limestone. This shear cuts through bedding in the gravel deposits but otherwise does not greatly disturb bedding. Where undeformed to the east in the Beaver Dam Mountains, parts of the Redwall Limestone are conspicuously bedded with alternating chert and limestone, but in this and other gravity-slide blocks bedding is destroyed and chert is highly concentrated in breccia that contains little limestone, suggesting significant dissolution of the limestone layers prior to emplacement of the slide blocks (figure 62) (Diehl and others, 1996, 1997; Anderson and Diehl, 1999).

Another instructive exposure of a gravity-slide block and enclosing deposits is on the northwest side of the hill near the center of section 35, T. 42 S., R. 19 W., where Muddy Creek exposures include interbedded fine- and coarse-grained facies with a more diverse clast composition (roughly 50% subrounded volcanic clasts of intermediate compostion, 20% to 30% subrounded clasts of likely Moenkopi or Kayenta strata, and 20% to 30% subrounded Paleozoic carbonate clasts), but lacking clasts of Proterozoic rocks. Anderson and Barnhard (1993a) reported sparse Proterozoic basement clasts in these basin-fill deposits, but lead-author Biek failed to find any Proterozoic clasts in numerous exposures of coarse-grained Muddy Creek strata at the west edge of the Beaver Dam Mountains; however, such clasts do make up the greater part of alluvial-fan deposits shed off the range in post-Muddy Creek time.

Figure 60. *View north to Castle Cliff fault and large gravity-slide block of Redwall Limestone; road to Welcome Spring traverses wash in middle of photograph (visible near the right edge of the photo). Here, the Castle Cliff fault dips gently west and places highly sheared Mississippian Redwall Limestone (Mr) against Paleoproterozoic crystalline basement rocks (Xu). Redwall Limestone is in turn overlain on the west by coarse alluvial-fan deposits interpreted to be a coarse facies of the Muddy Creek Formation (Tmc), but we are uncertain if their contact (white line) is a depositional contact or a low-angle normal fault. Figure 61 shows close-up of large gravity-slide block (Tmc[Mr]).*

Figure 61. *Close-up of large, east-dipping gravity-slide block of Redwall Limestone (shown on figure 60) that overlies similarly dipping alluvial-fan deposits interpreted to be a coarse facies of the Muddy Creek Formation; view north just southwest of Welcome Spring. Inset shows brecciated Mississippian bedrock. These coarse alluvial-fan deposits lack clasts of the Paleoproterozoic crystalline basement, now widely exposed in the core of the Beaver Dam Mountains culmination, showing that this and other gravity-slide blocks were emplaced prior to unroofing of the Beaver Dam Mountains culmination, likely in late Miocene time.*

Figure 62. *Another view of the brecciated block of Redwall Limestone shown on figures 60 and 61. Undeformed Redwall Limestone typically contains alternating beds of chert and limestone, but in the gravity-slide blocks, chert is typically concentrated in breccias that contain little limestone. Limestone dissolution and breccia formation are interpreted to have formed in the upper plate of the Castle Cliff fault, prior to emplacement of the gravity-slide blocks.*

The lack of Proterozoic clasts in the sedimentary strata suggests that the gravity-slide blocks were emplaced prior to unroofing of the crystalline basement and were derived from the highly attenuated sheath of Paleozoic strata inferred by Anderson and Barnhard (1993a) to have once been continuous over the crest of the range. Because the blocks are clearly incorporated in what we interpret as the Muddy Creek Formation, their age of emplacement is late Miocene to earliest Pliocene, as first suggested by Anderson and Barnhard (1993a, 1993b), not Quaternary-latest Tertiary as reported by others. Anders and others (1998; see also Anders and others, 2000) suggested that the blocks incorporated in basin-fill deposits west of the Beaver Dam Mountains were emplaced by a single catastrophic event in late Tertiary time. However, the gravity-slide blocks in the embayment west of Welcome Spring appear to be distributed through several hundred feet of Muddy Creek strata, suggesting multiple episodes of emplacement in late Miocene time. Also, strong block-to-block variations in non-catastrophic, pre-emplacement dissolution and collapse structure suggests derivation from contrasting locations in the highlands.

The location of the Muddy Creek Formation and associated gravity-slide blocks in an embayment in the mountain front west of Welcome Spring suggests preservation on a relay ramp between en echelon parts of the Red Hollow and Piedmont faults, as well as in the footwall of the Red Hollow fault, which are now being exhumed due to footwall uplift associated with the Red Hollow and Piedmont faults. It may be that the slide blocks and enclosing Muddy Creek strata represent growth strata deposited at and near the edge of an evolving Miocene basin. If this is true, the older Muddy Creek strata, now exposed farther from the mountain front, should dip more steeply than younger Muddy Creek strata. Extensive cover by alluvial fans and colluvium make this hard to determine, and may also mask normal faults that almost surely cut the Muddy Creek Formation.

Caliente-Enterprise Zone

In map view, the Bull Valley Mountains, in the northwest part of the map area, look like a shattered pane of glass, with long, sinuous, mostly east-trending, oblique-slip and right-lateral faults that contrast strongly with the widely spaced, mostly north-trending faults elsewhere on the map. This structurally complex, east-trending zone of faulting is part of what Axen (1998) and Hudson and others (1998) called the Caliente-Enterprise zone, which reaches westward into Nevada. The Caliente-Enterprise zone can be considered one of many east-trending transverse zones—which are like transform faults in the ocean basins in that they separate areas north and south of them that extended or deformed by different amounts or rates or by different types of structures (Rowley, 1998; Rowley and Dixon, 2001). Unlike modern basin-range extension that produced north-south-oriented basins and ranges, transverse zones formed parallel to extension. Transverse zones are defined by an alignment of east-west faults, igneous belts, anomalous geothermal activity, and aeromagnetic and gravity anomalies that have long been recognized in the Basin and Range Province (Rowley, 1998). Transverse zones are partly dismembered and overprinted by modern basin-range extension and so are difficult to recognize, yet they are important in that they may influence regional ground-water flow, and because they commonly coincide with zones of hydrothermally altered rock, mining districts, and segment boundaries of major normal faults. They certainly were active during the middle Cenozoic episode of voluminous calc-alkaline volcanism and throughout the time of basin-range extension, and some zones probably began to form during the Jurassic or Cretaceous regional compression of Sevier deformation (Ekren and others, 1976, 1977; Rowley and others, 1978; Rowley, 1998; Rowley and Dixon, 2001).

Paleomagnetic data revealed significant counterclockwise vertical-axis rotation of large, fault-bounded blocks that make up the Caliente-Enterprise zone. Rotations differ systematically westward from 45° to more than 90° in the eastern part (Bull Valley Mountains), to 30° to 40° in the central part, to 10° to 15° in the western part of the zone; rotation probably began between 18 and 14 million years ago (Hudson and others, 1998). Hudson and others (1998) interpreted the Caliente-Enterprise zone as a broad, left-lateral transfer zone that transferred 25 miles (40 km) or more of extension from areas to the north to areas to the south (figure 63). Significant north-south shortening accompanied the block rotations.

EVOLUTION OF THE MODERN LANDSCAPE

To reconstruct the events leading to the present landscape, it is convenient to establish a starting point—some well understood "pause" in the continuous drama of geologic events. For our purposes, one such horizon is the top of the youngest marine deposits, which represent the last time that the land surface was at sea level. In southwest Utah, the Late Cretaceous Tropic Shale marks the last incursion of the Western Interior Seaway; it is present on the Markagunt Plateau above Zion National Park and grades westward

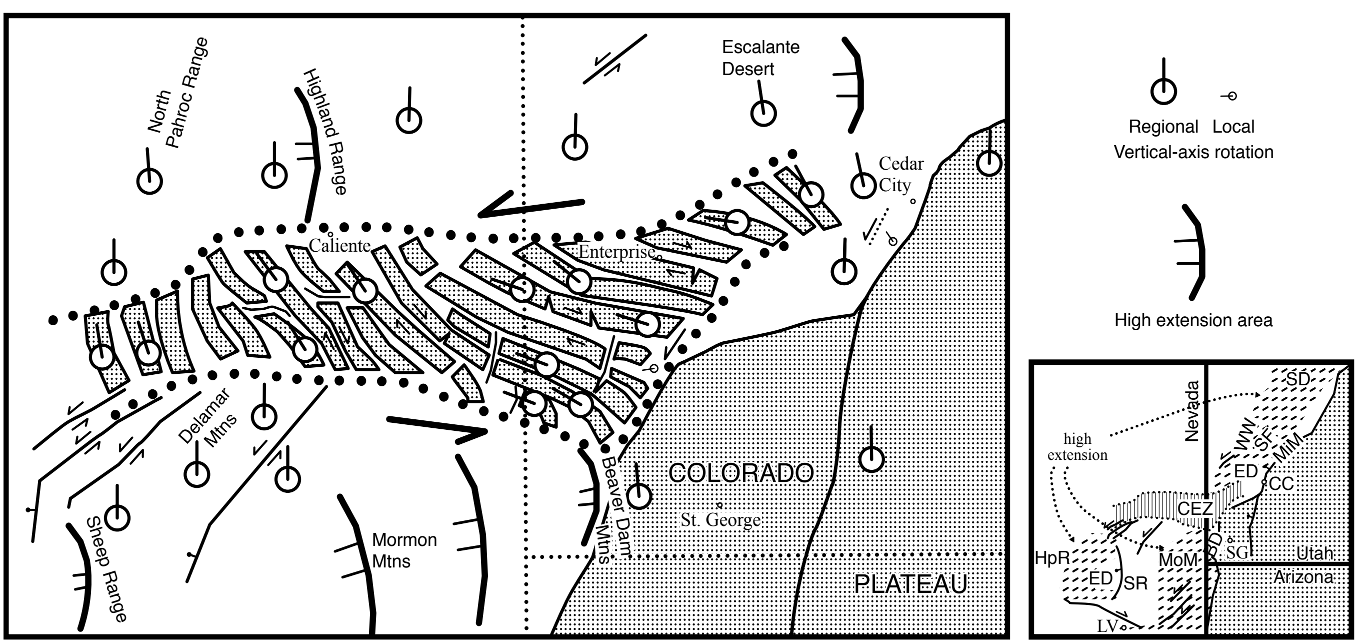

Figure 63. Tectonic interpretation of the Caliente-Enterprise zone as a transfer zone. Counterclockwise rotation and discontinuous en echelon faulting accommodated distributed left-lateral shear in a broad transfer zone that formed in response to heterogeneous Miocene extension. Zones of high extension depicted on small figure of lower right are modified from Wernicke (1992). Compare with figure 1 for location of St. George quadrangle. BD, Beaver Dam Mountains; CC, Cedar City; CEZ, Caliente-Enterprise zone; DR, Desert Range; ED, Escalante Desert; HpR, Halfpint Range; LV, Las Vegas; MiM, Mineral Mountains; MoM, Mormon Mountains; SD, Sevier Desert detachment; SF, San Francisco Mountains; SG, St. George; SR, Sheep Range; WW, Wah Wah Mountains. From Hudson and others (1998).

into the coastal-plain strata of the lower Iron Springs Formation of the Pine Valley Mountains. The west part of the map area was probably a coastal plain just above sea level at that time. By projecting this now-uplifted, deformed, and eroded surface over the map area and estimating the total thickness of the rock column before erosion, we get a picture of the amount of uplift that has occurred in each area. The numbers for southwesternmost Utah are truly impressive. In the northeast part of the map area, this "sea-level" surface is now at an elevation of about 7500 feet (2300 m) above sea level. It is projected to about 12,000 to 14,000 feet (3700–4300 m) above sea level near the Hurricane Cliffs (due partly to upwarp associated with folding of the Virgin and Kanarra anticlines), and to about 7000 feet (2100 m) above sea level around St. George. The greatest amount of uplift is in the Beaver Dam Mountains where this surface would be at elevations over 20,000 feet (6000 m) above sea level. In contrast, in parts of the Bull Valley Mountains this surface may be near or below sea level, and in parts of the Mesquite Basin west of the Beaver Dam Mountains it lies far below sea level.

When did this great amount of vertical structural relief occur, what caused it, and what caused the large differences between various areas? A decade or two ago, geologic convention held that the Colorado Plateau was uplifted from low elevations within the past 10 to 15 million years, although no one could provide a good mechanism (Lucchitta, 1979; Gable and Hatton, 1983; Hintze, 1993). During the past several years, however, ideas have changed. Geologists now generally agree that regional uplift began in the early Tertiary due to compression during the late stages of the Sevier and Laramide orogenies, and has continued at varying rates to the present (Kelley and others, 2001; Young, 2001; Sahagian and others, 2002; Davis and Pollock, 2003; Morgan, 2003; Flowers and others, 2008). As described by Willis (2000), the Laramide orogeny was synchronous with the late stages of the Sevier orogeny and both resulted from collision of the Farallon and North American tectonic plates, but they are distinguished by style of deformation. The Sevier orogeny defines a more western event that took advantage of weak bedding planes in thick Paleozoic and Mesozoic sedimentary rock. Shortening in also-weak basement metamorphic and igneous rock was transferred tens of miles eastward along weak shale and evaporite layers, producing "thin-skinned" thrust faulting that, in the eastern part, only involved sedimentary strata. In contrast, the Laramide orogeny produced "basement-cored" uplifts because thin sedimentary rock in those areas did not easily "decouple" from the basement rock, and because basement rock was more rigid than in the Basin and Range to the west.

Large-scale, regional uplift of western North America, including the Colorado Plateau, is thus ascribed to the Laramide orogeny. But this uplift was uneven, with some local areas rising several thousand feet more than others, forming large domal anticlines (such as the San Rafael Swell and Kaibab Uplift). Uplift may have slowed or paused during the middle Tertiary, then accelerated during the late Tertiary. Sahagian and others (2002) determined that the regional uplift rate from 25 to 5 million years ago was about 130 feet (40 m) per million years, and that this rate increased abruptly to about 720 feet (220 m) per million years about 5 million years ago. Morgan (2003) presented evidence that much of this latter uplift can be attributed to a major change deep beneath the Earth's surface near the mantle-crust boundary in which minerals change phases from dense ecologite to less-dense garnet-granulite, resulting in an increase in volume. Part of this uplift can also be attributed to crustal rebound as regional erosion removed several thousand feet of overlying rock, beginning in earnest about 6 million years ago with integration of the Colorado River system and exhumation of the Colorado Plateau (Pederson and others, 2002; Pederson, 2008).

About 20 million years ago, the topography of what is now southwest Utah must have looked very different. The overall drainage pattern was probably to the east and northeast, directly opposite the southwest-flowing drainage we see today (see, for example, Young, 2001). The entire late Mesozoic to Tertiary sedimentary package preserved today around the Pine Valley Mountains would have extended southward across the Utah-Arizona border. These strata were doubtless blanketed by the distal part of calc-alkaline volcanic rocks derived from the Indian Peak and Caliente caldera complexes and the Iron Axis suite of intrusions. There were no Hurricane Cliffs, no red-rock canyons.

Beginning about 17 million years ago, the Basin and Range Province, which was previously possibly higher than the Colorado Plateau, began to founder and collapse away from the Colorado Plateau (see, for example, Rowley and others, 1979; Duebendorfer and others, 1998; Hudson and others, 1998; Faulds and others, 2001; Rowley and Dixon, 2001). The present topography of the transition zone began to form. (Most of this topography, however, dates to less than 10 million years ago, when the main part of basin-range deformation began in force.) The inception of basin-range tectonics was a direct result of the interaction of the East Pacific Rise (mid-ocean ridge or spreading center) with the North American plate (Atwater, 1970; Severinghaus and Atwater, 1990; see also Stoffer, 2006) (figure 64). The ridge approached the North American plate at an oblique angle, such that it progressively contacted the continent south-to-north up the continental margin (the Juan de Fuca plate off the modern coast of northern California, Oregon, and Washington is a remnant of the Farallon plate that was subducting during the Cretaceous and early Tertiary to produce the Sevier orogeny; the Juan de Fuca plate is still actively subducting beneath the continent, giving rise to the Cascade volcanoes). Consumption of the spreading center caused the cessation of earlier subduction and ultimately the establishment of the San Andreas transform boundary (fault). This effectively ended regional compression and caused the southwestern U.S. to rift or pull away from the rest of North America (the Pacific plate west of the East Pacific Rise is moving northwest toward Alaska). It also led to the opening of the Gulf of California about 6 million years ago (Oskin and Stock, 2003; see also Karlstrom and others, 2008) and thus integration of the westward-flowing Colorado River system, which renewed regional exhumation of the Colorado Plateau, including carving the Grand Canyon and canyons of the Virgin River and its tributaries. Fiero (1986) provided a good summary of hypotheses that may explain the inception of basin-range extension.

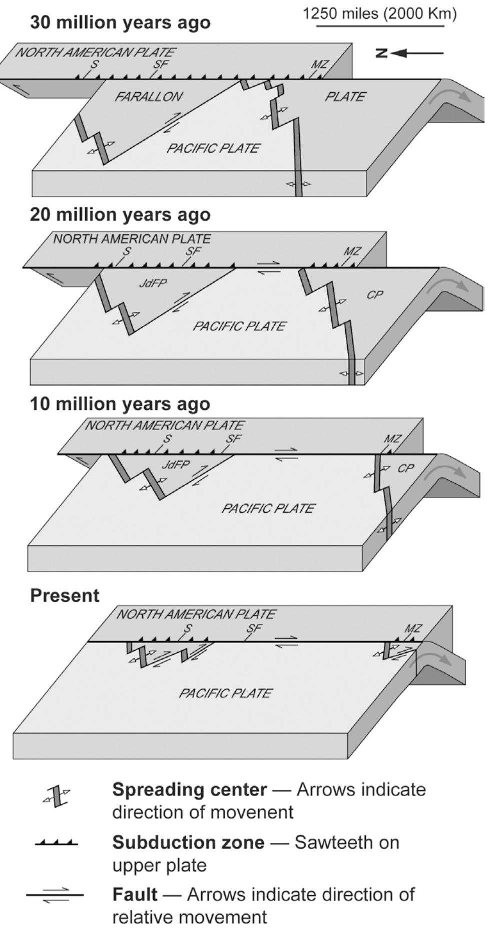

Figure 64. *Beginning about 30 million years ago, the North American plate progressively overrode the East Pacific Rise spreading center, leading to the cessation of subduction and development of the San Andreas transform plate boundary. As the spreading center was overridden, the Farallon plate was divided into two smaller plates, the northern Juan de Fuca plate (JdFP) and the southern Cocos plate (CP). Development of the transform boundary ended compressional deformation that had affected Utah during the late Mesozoic and early Tertiary and led to major extension of the Basin and Range. S=Seattle; SG=St. George, SF=San Francisco; MZ, Manzanillo, Mexico. Modified from Stoffer (2006).*

Beginning in Miocene time in southwest Utah and continuing nearly to the present day, dozens of widely scattered basaltic lava flows poured out onto the landscape (figure 65). Today, they constitute part of the Western Grand Canyon basaltic field, which extends across the southwest part of the Colorado Plateau and adjacent transition zone in southwest Utah, northern Arizona, and easternmost Nevada (Hamblin, 1963, 1970; Best and Brimhall, 1970, 1974; Best and others, 1980; Smith and others, 1999). The many Pliocene to Pleistocene basaltic lava flows of the Virgin River corridor provide important constraints on long-term erosion rates across the region. Incision rates of the Virgin River increase from west to east across southwest Utah from about 0.2 feet per thousand years (0.06 m/kyr) on the St. George block, to 0.35 feet per thousand years (0.11 m/kyr) on the Hurricane block, and to about 1.15 feet per thousand years (0.35 m/kyr) on the Zion block (Willis and Biek, 2001) (figure 66 and table 1). These rates reconfirm and expand on many of the findings of Hamblin and others (1981), who similarly documented incision rates in the St. George basin. Using these long-term incision rates, we can show, for example, that most of Zion Canyon was carved within the past 2 million years (Biek and others, 2003). Variation in the amount and rate of long-term incision along parts of the Virgin River is largely a function of base-level lowering along faults at and near the boundary of the Basin and Range; the rates are fastest in the east part of the map area because the Hurricane fault is the most active fault in the region. Long-term incision rates of Beaver Dam Wash may be comparable to those east of the Hurricane fault, but this remains to be studied.

LIVING IN UTAH'S DIXIE—GEOLOGIC RESOURCES AND GEOLOGIC HAZARDS

The spectacular geology of Utah's Dixie, as the area around St. George is affectionately known (for its warm climate and pioneer's attempts to grow cotton), has always played a central role in the lives of its inhabitants. First, tectonism—notably the north-trending basin-range faults—created the mountains and valleys they inhabit. Second, erosion of this faulted topography—primarily a function of the Virgin River and its tributaries—beveled the landscape down to the rocks we now walk on, exposing the red rock that so many of us love. Third, geologic and water resources partly led to early settlement and an economy that brought more people to the area. Significant population growth and ensuing development of recent decades has brought an awareness of the many geologic hazards present in the region.

The Virgin River and its major tributaries and fertile floodplains provided the original incentive for communities of Native Americans, and later pioneers, to locate in this hot desert. Iron deposits in the Iron Axis of Washington and Iron Counties led to the founding of Cedar City, which during and after World War II, was the largest iron district in the West (Mackin, 1968), and the district is once again gearing up for iron production. Oil from the small, now shut-in Virgin oil field, one of the earliest oil discoveries in Utah, once provided jobs and income to southwest Utah (table 3 shows selected oil and gas exploration wells for the map area). The turn-of-the-century Silver Reef mining district near Leeds extracted silver and uranium from the Springdale Sandstone Member of the Kayenta Formation (Proctor and Shirts, 1991; Biek and Rohrer, 2006). The Goldstrike gold district at the west edge of Washington County (Willden and Adair, 1986; Willden, 2006) only recently ceased production in the mid-1990s. The Apex mine southwest of Santa Clara, recently a major source for rare-earth metals for high-technology electrical industries (Bernstein, 1986a, 1986b), closed in 1989.

Southwest Utah's population growth rate is among the highest in the country. This growth boom could not be sustained without ground water, which provides nearly a third of the culinary water for Washington County. Most ground water comes from wells in the Navajo Sandstone, which has good primary porosity and permeability because it is composed of poorly cemented, wind-

Figure 65. *View east (from the east-dipping limb of the Virgin anticline) up the Virgin River to cinder cones and lava flows of the Hurricane area. These and other basaltic lava flows in the map area are part of the Western Grand Canyon basaltic field, which contains hundreds of relatively small late Tertiary and Quaternary lava flows and cinder cones. In the distance, Hurricane Mesa is on the left; Gooseberry Mesa and the mesa of Little Creek Mountain are on the right, and the West Temple and cliffs of Zion National Park are in the middle skyline.*

blown (and thus well-sorted) sand. Recharge for surface and ground water comes from precipitation in the mountains in the northern part of the county, primarily the Pine Valley and Bull Valley Mountains and the Kolob and Markagunt Plateaus. Most well fields tapping the Navajo Sandstone, however, are located on basin-range faults, which provide secondary (fracture) permeability. In fact, fracture flow allows ground-water recharge not only to rapidly reach the well fields but enables the wells to be of high yield (Rowley and Dixon, 2004). The newest and most innovative well field in Washington County is alongside Sand Hollow Reservoir, which was created by the Washington County Water Conservancy District to recharge the Navajo aquifer directly beneath the reservoir. Its highest producing wells are along fault zones that act as conduits to tap ground water leaking from the reservoir. Water for the Sand Hollow Reservoir, and for the Quail Creek Reservoir, is piped from the Virgin River diversion dam, which lies upstream from Pah Tempe hot springs. The springs have the highest recorded spring-water temperature in the St. George basin—they average about 104°F (40°C), have a pH of 6.3, and discharge sodium-chloride-type water with a very high average of 9600 to 9900 mg/L total dissolved solids (Blackett and Wakefield, 2002; Nelson and others, 2009). The high TDS of the spring water contributes to poor downstream water quality and is the reason Virgin River water is collected above the springs and piped to Quail Creek and Sand Hollow Reservoirs.

Living today, as in the past, requires some understanding of and compromises with nature, because geologic hazards can lead to economic loss and physical injury. Most notorious was the 1992 magnitude (M_L) 5.8 St. George earthquake, the largest in Utah since 1975 (Christenson, 1995). It caused at least $1 million in economic losses, including triggering the Springdale landslide near the west entrance to Zion National Park, 27 miles (43 km) east of the epicenter near Washington City. The landslide, which destroyed three homes, resulted when the Petrified Forest Member of the Chinle Formation slid due to ground shaking. However, the Petrified Forest Member does not need an earthquake to cause problems. For example, it underlies a low mesa on the north side of Santa Clara where road construction and lawn irrigation have led to repeated slumping and sliding of the clay-rich rock, destroying several homes and causing damage to many back yards (Harty, 1992; Lund and others, 2007b). In 2001 at Rockville, aptly named because huge rock-fall boulders from the Shinarump Conglomerate Member of the Chinle Formation litter the area, a 200- to 300-ton boulder (figure 67) crashed into a home, narrowly missing its owner (Lund, 2002; Rowley and others, 2002; Lund and others, 2007b). In the first two weeks of January 2005, heavy rainfall led to flooding of the Santa Clara and Virgin Rivers from Hurricane, Utah, to Mesquite, Nevada, resulting in the destruction of as many as 50 homes, mostly along the Santa Clara River (figure 68) (Lund and others, 2007b).

The Utah Geological Survey recently completed a series of geologic-hazard and adverse-construction-condition maps of the St. George-Hurricane area (Lund and others, 2008). These geographic information system- (GIS-) based maps provide an important tool for government officials, developers, and consultants to identify potential hazards and thus determine where detailed site-specific geologic hazard investigations are needed prior to development. The map folio contains fourteen 1:24,000-scale maps of flood hazards, expansive soil and rock, landslides, surface faulting, liquefaction, rock fall, collapsible soil, gypsiferous soil and rock, caliche, shallow bedrock, shallow ground water, wind-blown sand, piping and erosion-susceptible soils, and breccia pipes and paleokarst. Each map has an accompanying document that provides information on the nature of the hazard or adverse condition in the study area. A GIS search application permits the maps to be queried by geologic hazard or adverse condition type and location, and produces a map and report on the hazards of adverse conditions of interest.

Table 1. *$^{40}Ar/^{39}Ar$ and K-Ar ages of basaltic lava flows, St. George 30' x 60' quadrangle and adjacent area.*

Flow	Map Symbol	UGS sample number	$^{40}Ar/^{39}Ar$ plateau age (Ma)	$^{40}Ar/^{39}Ar$ isochron age (Ma)	Mineral	Lab used	7.5' Quadrangle for UGS Samples
Aqueduct Hill	Tbah	VY8301-4	1.98 ± 0.04	1.97 ± 0.05	Whole rock	NMGRL	Veyo
Aqueduct Hill	Tbah	VY8301-4	2.05 ± 0.01	2.02 ± 0.01	Whole rock	USGS	Veyo
Aqueduct Hill	Tbah	VY9303-1	1.33 ± 0.45*	0.08 ± 0.06*	Whole rock	NMGRL	Veyo
Baker Dam	Qbbd	VR8301-3	0.670 ± 0.040		Whole rock	USGS	Veyo
Baker Dam	Qbbd	VR8301-1	0.690 ± 0.140	0.686 ± 0.022	Whole rock	USGS	Veyo
Big Sand	Qbb	VR42-09	1.13 ± 0.05	1.111 ± 0.007	Whole rock	USGS	Santa Clara
Cedar Bench	Qbcb	VR40-11	1.23 ± 0.01		Whole rock	USGS	Washington
Cedar Bench (Airport)	Qbcb	VR40-05	1.23 ± 0.01		Whole rock	USGS	St. George
Cedar Bench (Snow Canyon Overlook)	Qbso	VR42-08	1.16 ± 0.03		Whole rock	USGS	Santa Clara
Central West	Qvcw	VY11802-7	1.77 ± 0.09	1.74 + 0.27	Hornblende	NMGRL	Veyo
Central West	Qvcw	VY8301-6	0.920 ± 0.070	0.930 + 0.022?	Feldspar	USGS	Veyo
Cinder cone NE of Veyo	Qbc	VY8301-5	0.650 ± 0.08	0.614 ± 0.04	Whole rock	USGS	Veyo
Cinder Pits	Qbcp	VR123-5	0.24 ± 0.02	0.27 ± 0.04	Whole rock	NMGRL	Hurricane
Crater Hill	Qbc	VR41-02	0.310 ± 0.070	0.298 ± 0.032	Whole rock	USGS	Springdale West
Crater Hill	Qbc	VR41-03	0.320 ± 0.130	0.294 ± 0.018?	Whole rock	USGS	Springdale West
Crater Hill	Qbc	ZP1501	0.280 ± 0.080	0.228 ± 0.040?	Whole rock	USGS	Springdale West
Crater Hill	Qbc	ZP1501	0.10 ± 0.08*	0.10 ± 0.29*	Whole rock	NMGRL	Springdale West
Dameron Valley east	Qbde	VY122001-4	0.59 ± 0.02	0.59 ± 0.03	Whole rock	NMGRL	Veyo
Diamond Valley	Qbdv	VY122001-1	0.05 ± 0.07*	0.009 ± 0.006*	Whole rock	NMGRL	Veyo
Divide	Qbd	TD12999-1	0.41 ± 0.08	0.43 ± 0.19	Whole rock	NMGRL	The Divide
East Reef	Qber	VR122-2	0.20 ± 0.16*	0.07 ± 0.08*	Whole rock	NMGRL	Hurricane
Eight Mile dacite	Trdy	EMD046	2.09 ± 0.22	2.03 ± 0.76	Whole rock	NMGRL	Central East
Gould Wash	Qbgw	VR41-08	0.420 ± 0.210	0.420 ± 0.05	Whole rock	USGS	Little Creek Mountain
Granite Wash	Tbgw	VY11802-1	2.03 ± 0.20	1.97 ± 0.02	Whole rock	NMGRL	Veyo
Grapevine Wash	Qbg	ZP-0502	0.22 ± 0.03	0.30 ± 0.05	Whole rock	NMGRL	The Guardian Angels
Grapevine Wash	Qbg	ZP-0503	0.29 ± 0.02	0.26 ± 0.04	Whole rock	NMGRL	The Guardian Angels
Grapevine Wash	Qbg	ZP-0605	0.31 ± 0.04	0.34 ± 0.04	Whole rock	NMGRL	The Guardian Angels
Grapevine Wash	Qbg	ZP-0606	0.26 ± 0.01	0.26 ± 0.02	Whole rock	NMGRL	The Guardian Angels
Grapevine Wash	Qbg	ZP-0607	0.26 ± 0.03	0.22 ± 0.04	Whole rock	NMGRL	The Guardian Angels
Grass Knoll	Qbgk	SMQ-1	1.02 ± 0.36**	1.20 ± 0.17	Whole rock	NMGRL	Saddle Mountain
Grass Valley	Qbgv	VR42-03	1.09 ± 0.09	0.966 ± 0.030	Whole rock	USGS	The Divide
Gunlock - Dameron Valley north	Qbgd	VY8301-9	1.62 ± 0.02	1.63 + 0.03	Whole rock	NMGRL	Veyo
Gunlock - Dameron Valley north	Qbgd	VY8301-10	1.65 ± 0.02	1.67 + 0.04?	Whole rock	USGS	Veyo
Gunlock - Dameron Valley north	Qbgd	VR40-01	1.61 ± 0.07		Whole rock	USGS	Shivwits
Hornet Point	Qbhp	CP83100-3	0.76 ± 0.02	0.74 ± 0.05	Whole rock	NMGRL	Cogswell Point
Horse Knoll	Qbhk	CP62001-3	0.73 ± 0.02	0.73 ± 0.02	Whole rock	NMGRL	Cogswell Point
Horse Ranch Mountain	Qbhr	KA92600-1	1.03 ± 0.06	1.05 ± 0.17	Whole rock	NMGRL	Kolob Arch
Horse Ranch Mountain	Qbhr	KA92600-1	1.03 ± 0.10	0.99 ± 0.46	Whole rock	NMGRL	Kolob Arch
Horse Ranch Mountain	boulder	KR72000-6	0.97 ± 0.18*	1.9 ± 4.4*	Whole rock	NMGRL	Kolob Reservoir
Ivans Knoll	Qbi	H11299-2	0.97 ± 0.07	1.04 + 0.06	Whole rock	NMGRL	Hurricane
Ivans Knoll	Qbi	VR123-11	1.03 ± 0.02	1.05 + 0.02	Whole rock	NMGRL	Hurricane
Ivans Knoll	Qbi	VR41-06	1.06 ± 0.16	0.937 + 0.006?	Whole rock	USGS	Hurricane
Kolob Peak	Qbkp	KR81200-1	1.05 ± 0.05	1.09 ± 0.19	Whole rock	NMGRL	Kolob Reservoir
Lark Canyon	Qbla	CEQ-18	0.61 ± 0.04	0.64 ± 0.04	Whole rock	NMGRL	Central East
Lava Point	Qblp	ZP-0601	1.02 ± 0.03	1.04 ± 0.02	Whole rock	NMGRL	Kolob Reservoir
Lava Point	Qblp	ZP-0602	1.08 ± 0.02	1.09 ± 0.02	Whole rock	NMGRL	Kolob Reservoir
Lava Point	Qblp	VR41-01c	1.06 ± 0.01		Whole rock	USGS	Virgin
Lava Point	Qblp	ZP-0405	1.14 ± 0.16	0.96 ± 0.2	Whole rock	NMGRL	The Guardian Angels
Lava Point					Whole rock		The Guardian Angels
Lava Ridge (Middleton Black Ridge)	Qbl	VR40-06	1.41 ± 0.01		Whole rock	USGS	St. George
Little Creek	Qblc				Whole rock	UNLV	Little Creek Mountain
Little Creek Peak	Qbli	VR43-01	1.44 ± 0.04		Whole rock	USGS	The Guardian Angels
Magotsu Creek	Qbmc	VY11702-1	0.98 ± 0.03	0.99 ± 0.03	Whole rock	NMGRL	Gunlock
Magotsu Creek	Qbmc	VY8301-7	1.00 ± 0.09	0.994 ± 0.023	Whole rock	USGS	Veyo
Magotsu Creek	Qbmc	VY9503-1	1.69 ± 0.33*	0.68 ± 0.33*	Whole rock	NMGRL	Veyo
Mahogany Knoll					Whole rock		Saddle Mountain
Pine Valley and Grass Valley Reservoir	Qbpg	CEQ-14	0.67 ± 0.07	0.67 ± 0.08	Whole rock	NMGRL	Central East
Pine Valley and Grass Valley Reservoir	Qbpg	CEQ-8	1.08 ± 0.13	1.12 ± 0.27	Whole rock	NMGRL	Central East
Pintura	Qbp(c)	VR113-4	0.89 ± 0.02	0.89 ± 0.02	Whole rock	NMGRL	Hurricane
Pintura	Qbp(b)	AC-1	0.88 ± 0.05		Whole rock	NMGRL	Kolob Arch
Pintura	Qbp(b)	BR-1	0.84 ± 0.03		Whole rock	NMGRL	Kolob Arch
Pintura	Qbp(c)	ACG-1	0.81 ± 0.10	0.83 ± 0.10	Whole rock	NMGRL	Pintura
Pintura	Qbp	MH-1	0.87 ± 0.04		Whole rock	NMGRL	Smith Mesa
Radio Tower	Qbrt	H11299-4	0.14 ± 0.06	0.13 ± 0.07	Whole rock	NMGRL	Hurricane
Red Knoll	Qbrk	SMQ-4	0.45 ± 0.86**	0.75 ± 0.55	Whole rock	NMGRL	Saddle Mountain
Remnants	Qbr	TD11699-3	1.06 ± 0.03	1.07 ± 0.08	Whole rock	NMGRL	The Divide
Remnants	Qbr	TD50699-1	0.94 + 0.04	0.94 ± 0.05	Whole rock	NMGRL	The Divide
Remnants	Qbr	GVN-2	1.47 ± 0.34	1.12 ± 0.50	Whole rock	NMGRL	The Divide
Saddle Mountain	Qbsm	VY122001-3	0.74 ± 0.07*	0.47 ± 0.12	Whole rock	NMGRL	Veyo
Santa Clara	Qbs	VR42-02	0.120 ± 0.06*	0.089 ± 0.025?	Whole rock	USGS	White Hills
Santa Clara	Qbs	SC100605-1	27,270 ± 250		charcoal, C-14	Beta Analytic	Santa Clara
Twin Peaks	Tbt	VR40-10	2.43 ± 0.02		Whole rock	USGS	Washington
Twin Peaks (T-Bone Hill)	Tbt	VR40-12	2.37 ± 0.02		Whole rock	USGS	Washington
Twin Peaks (West Black Ridge)	Tbt	VR40-04	2.34 ± 0.02		Whole rock	USGS	St. George
Twin Peaks (West Black Ridge)	Tbt				Whole rock		St. George
Veyo	Qbve3	VY111902-7	0.69 ± 0.04	0.71 ± 0.08	Whole rock	NMGRL	Veyo
Veyo	Qbve4	VY122001-9	disturbed age spectrum		Whole rock	USGS	Veyo
Virgin Flats	Qbvf	CP71900-1	0.37 ± 0.02	0.38 ± 0.06	Whole rock	NMGRL	Cogswell Point
Volcano Knoll	Qbvk	CP71900-6	0.34 ± 0.03	0.359 ± 0.09	Whole rock	NMGRL	Cogswell Point
Volcano Mountain	Qbv1				Whole rock	USGS	Hurricane
Volcano Mountain	Qbv1				Whole rock	USGS	Hurricane
Volcano Mountain	Qbv2				Whole rock	USGS	Hurricane
Volcano Mountain	Qbv2				Whole rock		Hurricane
Volcano Mountain	Qbv2				Whole rock		Hurricane
Volcano Mountain	Qbv2				Whole rock		Hurricane
Washington	Qbw	HJ11299-2	0.87 ± 0.04	0.89 ± 0.08	Whole rock	NMGRL	Harrisburg Junction
Washington	Qbw	VR40-07	0.98 ± 0.02		Whole rock	USGS	Harrisburg Junction

NOTES:
Map Symbol is the symbol used on this map; numbered rows indicate further subdivision as shown on maps used in compilation of this map
Latitude and longitude based on NAD27 for UGS samples; coordinate system unknown for other samples
Latitude and longitude for samples reported in Lund and Everitt (1998) and Lund and others (2001) are approximate
UNLV = University of Nevada - Las Vegas
NMGRL = New Mexico Geochronology Research Lab

Age uncertainty = 2 standard deviations
Isochron ages not available for some samples
Whole rock means groundmass concentrate
** = integrated age
* = low confidence
? = preliminary number

***Table 1** continued.*

Flow	Reference	Longitude (UGS Sample)	Latitude (UGS Sample)	Other K-Ar and $^{40}Ar/^{39}Ar$ Ages	Sample Number	References
Aqueduct Hill	this report	-113.6719	37.3504			
Aqueduct Hill	this report	-113.6719	37.3504			
Aqueduct Hill	this report	-113.6678	37.3579			
Baker Dam	this report	-113.6701	37.3484			
Baker Dam	this report	-113.6933	37.3330			
Big Sand	this report	-113.6103	37.1631			
Cedar Bench	this report	-113.5697	37.1592			
Cedar Bench (Airport)	this report	-113.5940	37.1047	1.07 ± 0.04 K-Ar		Hamblin and others (1981)
Cedar Bench (Snow Canyon Overlook)	Higgins (2003)	-113.6314	37.2195			
Central West	this report	-113.6719	37.3711			
Central West	this report	-113.6617	37.3721			
Cinder cone NE of Veyo	this report	-113.6765	37.3473			
Cinder Pits	Biek (2003b)	-113.3186	37.1819			
Crater Hill	this report	-113.0834	37.1746	0.53 ± 0.215		source???
Crater Hill	this report	-113.0848	37.1802	disturbed age spectrum	CD98-5	Downing (2000)
Crater Hill	this report	-113.1057	37.2116			
Crater Hill	this report	-113.1057	37.2116			
Dameron Valley east	this report	-113.5641	37.3095			
Diamond Valley	this report	-113.6300	37.2510			
Divide	Hayden (2004a)	-113.2981	37.0719			
East Reef	Biek (2003b)	-113.3431	37.2081			
Eight Mile dacite	this report	-113.5897	37.4399			
Gould Wash	this report	-113.2491	37.1258	0.278 ± 0.0183 Ar/Ar	LD98-5	Downing (2000)
Granite Wash	this report	-113.6306	37.3667	1.5 ± 0.2 K-Ar	SG-67	Best and others (1980)
Grapevine Wash	Willis and Hylland (2002)	-113.1169	37.3381			
Grapevine Wash	Willis and Hylland (2002)	-113.1069	37.3500			
Grapevine Wash	Willis and Hylland (2002)	-113.0989	37.2981	0.26 ± 0.025		source???
Grapevine Wash	Willis and Hylland (2002)	-113.0961	37.2789	0.26 ± 0.09 K-Ar	Zion-20	Best and others (1980)
Grapevine Wash	Willis and Hylland (2002)	-113.0961	37.2800			
Grass Knoll	this report	-113.6059	37.2742			
Grass Valley	this report	-113.3244	37.0747			
Dameron Valley north	this report	-113.6675	37.3231			
Dameron Valley north	this report	-113.6664	37.3245			
Dameron Valley north (Gunlock)	this report	-113.7744	37.2381	1.6 ± 0.11 K-Ar		Hintze and others (1994)
Hornet Point	Biek and Hylland (2007)	-112.9994	37.4514	1.2 ± 0.6 K-Ar	Zion-58	Best and others (1980)
Horse Knoll	Biek and Hylland (2007)	-112.8811	37.4369	0.81 ± 0.5 K-Ar	Zion-51	Best and others (1980)
Horse Ranch Mountain	Biek (2007a)	-113.1589	37.4786			
Horse Ranch Mountain	Biek (2007a)	-113.1589	37.4786			
Horse Ranch Mountain	Biek (2007b)	-113.0444	37.4408			
Ivans Knoll	Biek (2003b)	-113.3520	37.1720			
Ivans Knoll	Biek (2003b)	-113.3639	37.1261			
Ivans Knoll	this report	-113.2971	37.1284			
Kolob Peak	Biek (2007b)	-113.0672	37.4225	1.13 ± 0.05 K-Ar	Zion-2	Best and others (1980)
Lark Canyon	this report	-113.6132	37.4036	0.56 ± 0.06 K-Ar	Zion-21	Best and others (1980)
Lava Point	Biek (2007b)	-113.0411	37.3881	1.0 ± 0.1 K-Ar	Zion-65	Best and others (1980)
Lava Point	Biek (2007b)	-113.0400	37.3861	1.6 ± 0.4 K-Ar	Zion-1A	Best and others (1980)
Lava Point	this report	-113.1468	37.2113	0.7 ± 0.24		source???
Lava Point	Willis and Hylland (2002)	-113.0600	37.3739	1.10 ± 0.08 K-Ar		Hamblin and others (1981)
Lava Point				1.40 ± 0.08 K-Ar	Zion-4	Best and others (1980)
Lava Ridge (Middleton Black Ridge)	Higgins (2003)	-113.5510	37.1121	1.5 ± 0.1 K-Ar	SG-79	Best and others (1980)
Little Creek				0.345 ± 0.015 Ar/Ar	LD98-1	Downing (2000)
Little Creek Peak		-113.0727	37.3556	1.40 ± 0.064		source???
Magotsu Creek	this report	-113.7613	37.2808	1.1 + 0.1 K-Ar	SG-70	Best and others (1980)
Magotsu Creek	this report	-113.6856	37.3671			
Magotsu Creek	this report	-113.6869	37.3619			
Mahogany Knoll				1.2 ± 0.1 K-Ar	SG-61	Best and others (1980)
Pine Valley and Grass Valley Reservoir	this report	-113.5292	37.4080			
Pine Valley and Grass Valley Reservoir	this report	-113.5178	37.4032			
Pintura	Biek (2003b)	-113.2969	37.2431	1.0 ± 0.1 K-Ar	SG-74	Best and others (1980)
Pintura	Lund and Everitt (1998)	-113.2364	37.4031			
Pintura	Lund and Everitt (1998)	-113.2144	37.4107			
Pintura	Lund and others (2001)	-113.2833	37.2833			
Pintura	Lund and others (2001)	-113.2361	37.3653			
Radio Tower	Biek (2003b)	-113.2889	37.1981			
Red Knoll	this report	-113.5544	37.3198			
Remnants	Hayden (2004a)	-113.3261	37.1050			
Remnants	Hayden (2004a)	-113.2939	37.1039			
Remnants	Lund and others (2001)	-113.3250	37.1069			
Saddle Mountain	this report	-113.6399	37.3123			
Santa Clara	this report	-113.6361	37.1167			
Santa Clara	Willis and others (2006)	-113.6564	37.1503			
Twin Peaks	this report	-113.5664	37.2224			
Twin Peaks (T-Bone Hill)	this report	-113.5687	37.1425			
Twin Peaks (West Black Ridge)	this report	-113.5991	37.1129	2.24 ± 0.11 K-Ar		Hamblin and others (1981)
Twin Peaks (West Black Ridge)				2.3 ± 0.1 K-Ar	SG-78	Best and others (1980)
Veyo	this report	-113.7049	37.2748			
Veyo	this report	-113.6857	37.3080			
Virgin Flats	Biek and Hylland (2007)	-112.9756	37.4131			
Volcano Knoll	Biek and Hylland (2007)	-112.9292	37.4217	0.36 ± 0.8 K-Ar	Zion-57	Best and others (1980)
Volcano Mountain				0.258 ± 0.024 Ar/Ar	7 21	Sanchez (1995)
Volcano Mountain				0.270 ± 0.050 Ar/Ar*	7 22	Sanchez (1995)
Volcano Mountain				0.353 ± 0.045 Ar/Ar	6 15	Sanchez (1995)
Volcano Mountain				0.293 ± 0.087 K-Ar		Hamblin and others (1981)
Volcano Mountain				0.289 ± 0.086 K-Ar	PED-53-66	Best and others (1980)
Volcano Mountain				0.303 ± 0.30 K-Ar	PED-53-66	Best and others (1980)
Washington	Biek (2003a)	-113.4719	37.1589	1.7 ± 0.1 K-Ar	SG-80	Best and others (1980)
Washington	Biek (2003a)	-113.4718	37.1378			

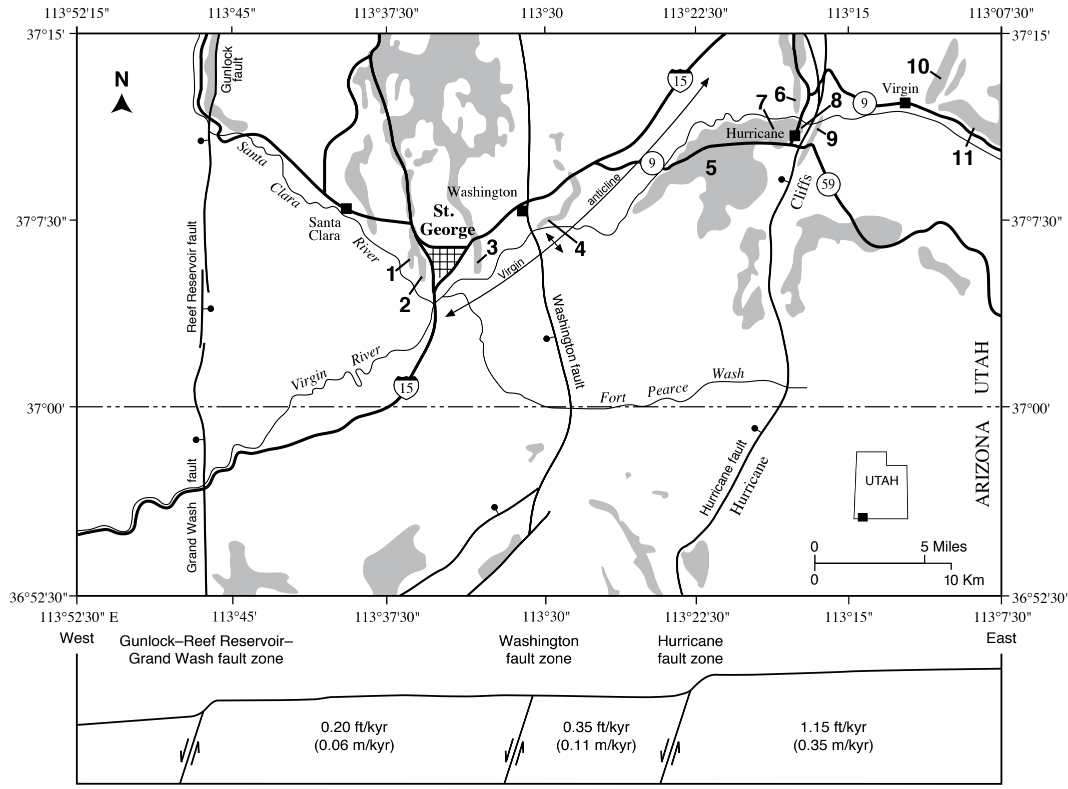

Figure 66. *Long-term incision rates for the Zion–St. George area can be determined by using radiometric ages for basaltic lava flows and the height of these flows above major drainages. This figure shows lava flows (gray) and their relationship to the Virgin River and major fault zones; numbers correspond to lava flows listed in table 2, which shows that long-term incision rates along the Virgin River differ systematically across the region from 0.2 to 1.15 feet per thousand years (0.06–0.35 m/kyr). Long-term incision rates are highest on the upthrown east side of the Hurricane fault, the most active fault in the region.*

Table 2. *Incision rates along the Virgin River, southwest Utah.*

FLOW	^{40}Ar/^{39}Ar AGE (Ma)[1]	HEIGHT[2] above Virgin River feet (m)	INCISION RATE feet/ka (m/kyr)	INCISION RATE feet/ka (m/kyr) Average near middle of block	
1 Twin Peaks	2.34 ± 0.02	550 (168)*	0.23 (0.07)	0.20 (0.06)	St. George block
2 Big Sand (Airport)	1.23 ± 0.01	250 (76)*	0.20 (0.06)		
3 Lava Ridge (Middleton)	1.41 ± 0.01	170 (52)	0.13 (0.04)		
4 Washington	0.87 ± 0.04 0.98 ± 0.02	350 (107)	0.39 (0.12) 0.36 (0.11)	0.35 (0.11)	Hurricane block
5 Ivans knoll	0.97 ± 0.07 1.03 ± 0.02	280 (85)*	0.30 (0.09) 0.26 (0.08)		
6 Pintura	0.89 ± 0.02	190 (58)	0.20 (0.06)		
7 Radio Tower	0.14 ± 0.06	70 (21)	0.49 (0.15)		
8 Volcano Mtn. (hanging wall)	0.353 ± 0.045[3]	170 (52)	0.49 (0.15)		
9 Volcano Mtn. (footwall)		410 (125)	1.15 (0.35)	1.15 (0.35)	Zion block
10 Lava Point	1.06 ± 0.01	1300 (400)*	1.31 (0.40)		
11 Crater Hill	0.12 ± 0.02[4]	125 (38)*	0.98 (0.30)		

These rates are based on dated basalt flows that either entered the channel of the ancestral Virgin River or that were projected to the river. The Washington fault separates the St. George and Hurricane blocks, and the Hurricane fault separates the Hurricane and Zion blocks.
[1]Utah Geological Survey data.
[2]Height above Virgin River based on geologic mapping by Higgins and Willis (1995), Willis and Higgins (1995), Willis and Hylland (2002), and Biek (2003a, 2003b). Measurements were made on U.S. Geological Survey 7.5' topographic maps and by photogrammetric methods. * indicates that the lava flow was projected along its paleolongitudinal profile to the Virgin River. Twin Peaks flow projected 2.5 miles (4 km), Big Sand (Airport) flow projected 1.0 mile (1.5 km), Ivans Knoll flow projected 1.2 miles (2 km), Lava Point flow projected 1.0 mile (1.5 km), and Crater Hill flow projected 0.5 mile (0.8 km). In each case, the gradient of the downstream portion of the lava flow was used to project the flow to the Virgin River.
[3]Sanchez (1995).
[4]Optically stimulated luminescence (OSL) age.

Figure 67. *Block of Shinarump Conglomerate Member of the Chinle Formation that broke free from the cliff above Rockville, destroying a house and nearly crushing its occupant. Note gouge mark left by boulder as it rolled across the yard. Rockville is known for its homes nestled among large rock-fall boulders; here we have dramatic evidence that the process of cliff retreat continues today. Photo by Bill Lund, Utah Geological Survey.*

Figure 68. *View northeast across the Santa Clara River to a subdivision off Indian Hills Drive, January 2005. Here, the Santa Clara River eroded laterally into terrace deposits, undermining and destroying several homes. While uncommon, large floods are a natural part of southwest Utah's geologic landscape, and building on floodplains and low terraces is not without risk.*

Landslides, earthquake faults, swelling soil and rock, floods, sinkholes, and rock falls are just some of the geologic hazards faced by residents of Utah's Dixie. Sand and gravel, decorative stone, gypsum, water, and, at one time, sliver, gold, rare-earth metals, and even oil are some of the main geologic resources known in the area. Coupled with a well-exposed section of rock that exceeds 8 miles (12 km) in thickness, which records a variety of ancient depositional environments, and with a bewildering array of geologic structures, southwest Utah is indeed an ideal outdoor classroom for those curious about the environment in which they live.

ACKNOWLEDGMENTS

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Table 3. *Selected Oil and Gas Exploration Wells in the St. George 30' x 60' quadrangle, Washington County, Utah.*

OBJECTID	APINUMBER	WELLNAME	TOTAL_DPTH	TOWNSHIP	TOWNSHIP_D	RANGE	RANGE_D	SECTION	MERIDIAN	FEET_N_S	DIR_N_S	FEET_E_W	DIR_E_W	QTR_QTR
1	4305320530	1 GOVT	382	43	S	15	W	17	SALT LAKE	1095	FSL	2415	FEL	SWSE
2	4305320528	1 ESCALANTE	2532	43	S	15	W	17	SALT LAKE	1708	FNL	2090	FWL	SENW
3	4305320526	1-A GOVT	2512	43	S	15	W	1	SALT LAKE	2513	FNL	1620	FEL	SWNE
4	4305320516	1	4600	42	S	14	W	16	SALT LAKE	330	FNL	330	FWL	NWNW
5	4305310214	1 ST GEORGE UNIT	6347	43	S	15	W	19	SALT LAKE	232	FNL	2461	FEL	NWNE
6	4305310302	1 FEDERAL	1357	43	S	12	W	31	SALT LAKE	660	FSL	660	FWL	SWSW
7	4305310379	1 FOUR BOYS FED	1000	43	S	15	W	24	SALT LAKE	1550	FSL	230	FWL	NWSW
8	4305310405	1 SHIVWITS	400	41	S	17	W	28	SALT LAKE	660	FSL	2080	FEL	SWSE
9	4305310602	1 PENN-US-KNOWLES-SK	3006	43	S	13	W	28	SALT LAKE	727	FNL	1906	FEL	NWNE
10	4305310635	1 UNIT	6260	42	S	13	W	25	SALT LAKE	60	FNL	200	FEL	NENE
11	4305310704	1 GOVT-WOLF	7315	40	S	13	W	23	SALT LAKE	90	FNL	3515	FWL	NWNE
12	4305310879	1 PINTURA UNIT	9501	39	S	13	W	33	SALT LAKE	675	FSL	505	FWL	SWSW
13	4305311164	1 UNIT	5496	39	S	13	W	33	SALT LAKE	398	FSL	872	FWL	SWSW
14	4305320044	1 FEDERAL	6000	43	S	13	W	14	SALT LAKE	660	FSL	660	FWL	SWSW
15	4305320542	1-X STATE HUNT	720	40	S	16	W	7	SALT LAKE	520	FNL	530	FEL	NENE
16	4305330001	30-B3X FEDERAL	5606	40	S	12	W	30	SALT LAKE	2183	FSL	1915	FWL	NESW
17	4305330002	1 WILSON-FEE	1825	41	S	13	W	23	SALT LAKE	3086	FSL	500	FWL	SWNW
18	4305330003	36-C1 STATE	4124	40	S	13	W	36	SALT LAKE	660	FNL	1980	FEL	NWNE
19	4305330005	1 TOLEDO-HIKO BELL	7060	42	S	13	W	11	SALT LAKE	700	FSL	2070	FWL	SESW
20	4305330007	1-13 FEDERAL	3171	40	S	13	W	13	SALT LAKE	1625	FNL	1140	FWL	SWNW
21	4305330014	1 STATE HUNT	584	40	S	16	W	7	SALT LAKE	500	FNL	500	FEL	NENE
22	4305330015	2 STATE HUNT	843	40	S	16	W	8	SALT LAKE	500	FNL	500	FWL	NWNW
23	4305330024	1-25 FEDERAL	2610	39	S	13	W	25	SALT LAKE	2060	FNL	721	FEL	SENE
24	4305330027	1 LOVELL-STATE	2540	43	S	14	W	32	SALT LAKE	500	FSL	506	FWL	SWSW
25	4305330031	1 DIXIE STATE	2880	43	S	14	W	32	SALT LAKE	1053	FSL	1712	FEL	SWSE
26	4305320540	1 ESCALANTE	969	43	S	15	W	32	SALT LAKE	1078	FNL	300	FEL	NENE
27	4305320522	1 ST GEORGE	970	43	S	14	W	30	SALT LAKE	1258	FNL	2131	FWL	NENW
28	4305320534	2 ARROWHEAD	4114	43	S	15	W	19	SALT LAKE	251	FNL	2524	FEL	NWNE
29	4305320524	1	1185	43	S	15	W	1	SALT LAKE	1078	FNL	300	FEL	NENE

OBJECTID	SPUD_DATE	COMP_DATE	DRILL_TYPE	OPERATOR	LATITUDE	LONGITUDE	UTM_EAST	UTM_NORTH	UTM_ZONE
1	12/26/1926	03/15/1927	CABLE	USONA OIL COMPANY	37.04044	-113.55644	272633	4102211.86	12
2	01/30/1927	05/07/1931	CABLE	MID-AMERICAN OIL	37.04771	-113.55873	272451.01	4103024.04	12
3	11/07/1932	09/19/1939	CABLE	ARROWHEAD OIL CO	37.07472	-113.48157	279392.4	4105839.09	12
4	/ /1919	/ /1919	ROTARY	CYPHER, HARRY T	37.14035	-113.43757	283491.63	4113020.1	12
5	01/31/1951	09/22/1951	ROTARY	CHEVRON USA INC	37.03712	-113.575	270972	4101888.03	12
6	05/03/1961	11/20/1961	CABLE	DAWN PETROLEUM, INC.	37.00222	-113.25221	299594.69	4097286.69	12
7	07/15/1963	09/19/1963	COMB-TL	FOUR BOYS COMPANY	37.02759	-113.49316	278224.68	4100636.5	12
8	06/22/1961	08/28/1961	ROTARY	GREAT WESTERN PETROLEUM	37.18448	-113.75888	255089.18	4118699.24	12
9	10/15/1959	03/24/1960	CABLE	INTEX OIL COMPANY	37.02325	-113.31576	293995.84	4099755.86	12
10	06/20/1909	12/30/1909	CABLE	K-T PETROLEUM CO	37.1122	-113.25545	299595.97	4109496.4	12
11	12/30/1963	02/29/1964	ROTARY	MCCULLOCH OIL CORP OF CA	37.30195	-113.27941	297973.5	4130601.63	12
12	10/02/1961	02/13/1962	ROTARY	PAN AMERICAN PETROLEUM COR	37.34993	-113.32418	294135.91	4136022.22	12
13	03/18/1951	09/19/1951	ROTARY	SUN OIL COMPANY	37.34988	-113.32364	294183.61	4136015.49	12
14	03/12/1966	06/12/1966	ROTARY	DEVEREUX CORP., THE	37.04163	-113.2889	296434.7	4101737.37	12
15	02/20/1975	05/16/1975	ROTARY	WARNER VALLEY OIL	37.32922	-113.68076	262480.73	4134561.63	12
16	02/28/1968	04/13/1968	ROTARY	BUTTES GAS & OIL CO	37.27905	-113.23971	301432.44	4127976.52	12
17	07/29/1968	04/02/1969	COMB-TL	LA VERKIN OIL COMPANY	37.20932	-113.28979	296804.58	4120345.7	12
18	01/14/1968	06/01/1969	COMB-TL	TITAN OIL INC	37.27125	-113.2621	299426.41	4127158.3	12
19	04/04/1971	05/05/1971	ROTARY	TOLEDO MINING CO	37.14356	-113.28379	297161.26	4113036.25	12
20	04/23/1972	05/11/1972	ROTARY	PEASE OIL & GAS COMPANY	37.31213	-113.26939	298888.84	4131709.82	12
21	11/04/1974	12/30/1974	ROTARY	WARNER VALLEY OIL	37.32928	-113.68066	262489.78	4134568.04	12
22	11/20/1974	11/28/1978	ROTARY	WARNER VALLEY OIL	37.32929	-113.67722	262794.67	4134560.5	12
23	04/09/1976	06/09/1976	ROTARY	NEPTUNE EXPLORATION LTD	37.37195	-113.25589	300243.97	4138318.79	12
24	03/15/1977	05/12/1977	ROTARY	LOVELL, DWAYNE B	37.00062	-113.45536	281510.46	4097556.42	12
25	11/15/1977	03/10/1978	ROTARY	ENERGY ASSOCIATES	37.00243	-113.44482	282453.68	4097733.1	12
26	11/25/1934	03/06/1935	ROTARY	ESCALANTE EXPLORATION	0	0	272433.56	4098212.3	0
27	12/01/1927	10/31/1929	CABLE	ST GEORGE OIL COMPANY	0	0	280370.94	4100094	0
28	05/13/1931	06/30/1935	ROTARY	ESCALANTE EXPLORATION	0	0	271054.75	4101750	0
29	03/25/1929	12/02/1932	ROTARY	ARROWHEAD OIL CO	0	0	279273.06	4105995.8	0

Source: Division of Oil, Gas, and Mining database accessed January 2009; all wells plugged and abandoned.

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APPENDIX

DESCRIPTION OF MAP UNITS

QUATERNARY

Artificial deposits

Qf **Artificial fill** (Historical) – Engineered fill used to create Quail Creek, Ash Creek, Sand Hollow, and Gunlock dams, and other smaller flood-control and water-supply structures, as well as the Washington County landfill; fill placed for highways, building pads, and other uses not mapped.

Alluvial deposits

Qa River and stream alluvium (Holocene) – Stratified, moderately to well-sorted gravel, sand, silt, and minor clay deposited in river and stream channels and floodplains; includes local small alluvial fans, colluvium, and stream-terrace alluvium less than about 10 feet (<3 m) above modern base level, and higher-level stream-terrace alluvium too small to map separately; typically 10 to 25 feet (3–8 m) thick.

Qat **Old river and stream alluvium** (Holocene to middle Pleistocene) – Stratified, moderately to well-sorted alluvial gravel, sand, silt, and minor clay that forms level to gently sloping terraces above modern drainages; locally divisible into six or more distinct terrace levels based on elevation above modern drainages, but undivided here due to map scale; deposited in stream-channel and floodplain environments and may include colluvium and alluvial fans too small to map separately; commonly forms a sand-and-gravel veneer 10 to 30 feet (3–9 m) thick over an eroded bedrock surface.

Qap **Pediment alluvium** (Holocene to middle Pleistocene) – Poorly sorted, subangular to rounded, silt and sand to small-boulder gravel that forms a locally resistant cap over eroded bedrock surfaces; locally divisible into several distinct levels, as south of St. George, but undivided here; deposited principally as debris flows, debris floods, and in ephemeral stream channels; 0 to about 80 feet (0–24 m) thick.

Qaf_1 **Level-1 fan alluvium** (Holocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment deposited at the mouths of active streams and washes; clast composition ranges widely and reflects rock types exposed in upstream drainage basins; deposited principally as debris flows and debris floods on active depositional surfaces; typically 10 to 30 feet (3–9 m) thick.

Qaf_2 **Level-2 fan alluvium** (Holocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment deposited at the mouths of streams and washes; clast composition ranges widely and reflects rock types exposed in upstream drainage basins; deposited principally as debris flows and debris floods and typically forms inactive surfaces incised by active drainages; typically 10 to 50 feet (3–15 m) thick.

Qafy **Younger fan alluvium** (Holocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment deposited at the mouths of streams and washes; clast composition ranges widely and reflects rock types exposed in upstream drainage basins; forms both active depositional surfaces (Qaf_1 equivalent) and low-level inactive surfaces incised by small streams (Qaf_2 equivalent) undivided here; deposited principally as debris flows and debris floods, but colluvium locally constitutes a significant part of the deposits; small, isolated alluvial fans are typically less than a few tens of feet thick, but large, coalesced fans, as in the New Harmony basin, are probably as much as 200 feet (60 m) thick.

Qafo **Older fan alluvium** (Pleistocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment with moderately developed calcic soils (hardpan or caliche); forms broad, gently sloping, deeply dissected surfaces; deposited principally as debris flows and debris floods. Deposits southeast of the Pine Valley Mountains are typically a few tens of feet thick, but deposits in the New Harmony basin likely exceed 200 feet (60 m) thick. Also interpreted as forming the upper part of basin-fill deposits of the Mesquite Basin in the southwest corner of the map area, where it forms coalesced alluvial fans shed from the Beaver Dam and Bull Valley Mountains that are incised by Beaver Dam Wash as much as 300 feet (90 m).

Colluvial deposits

Qc **Colluvium** (Holocene to upper Pleistocene) – Poorly sorted, angular, clay- to boulder-size, locally derived sediment deposited principally by slope wash and soil creep; locally includes talus, alluvium, and eolian sand too small to map separately; gradational with talus; includes older colluvium now incised by adjacent drainages; generally less than 20 feet (6 m) thick, although older colluvium on the Kolob Plateau may be as much as 100 feet (30 m) thick.

Eolian deposits

Qes **Eolian sand** (Holocene to upper Pleistocene) – Well sorted, fine- to medium-grained, well-rounded, frosted quartz sand; sand is recycled principally from the Navajo Sandstone and Kayenta Formation; locally forms small dunes partly stabilized by vegetation; locally capped by thick calcic soils (hardpan or caliche); typically less than 20 feet (6 m) thick.

Qeo **Older eolian sand** (upper Pleistocene) – Forms resistant, planar surfaces capped by well-developed calcic soils (hardpan or caliche) and lesser eolian sand; typically less than 20 feet (6 m) thick.

Lacustrine and basin-fill deposits

Ql **Lacustrine and basin-fill sediment** (Holocene to middle Pleistocene) – Well-stratified sand, silt, and clay deposited in small lakes and basins created by landslides, rock falls, and basaltic lava flows; typically grades into colluvium, alluvium, and fan alluvium at basin margins and in stratigraphically upper part of the map unit; variable thickness from a few tens of feet to as much as about 250 feet (75 m) thick; sediments from 14 such lakes and ephemeral ponds are known in the Zion National Park area; see Hamilton (1978, 1979, undated, 1995), Doelling (2002), Willis and Hylland (2002), Willis and others (2002), Biek and others (2003), and Biek (2007a, 2007b) for details on the lake deposits in this area.

Mass-movement deposits

Qmt **Talus (Holocene to upper Pleistocene)** – Poorly sorted, angular boulders and finer-grained interstitial sediment deposited principally by rock fall on and at the base of steep slopes; typically grades downslope into colluvium and may include colluvium where impractical to differentiate the two; also includes alluvium in the bottom of washes, and, in the Smith Mesa quadrangle, locally includes small landslides along the outcrop belt of the Petrified Forest Member of the Chinle Formation; generally less than 30 feet (9 m) thick.

Qms **Landslides** (Holocene to middle[?] Pleistocene) – Poorly sorted, clay- to boulder-size, locally derived material deposited by rotational and translational landslide movement; characterized by hummocky topography and small ponds, numerous internal scarps, and chaotic bedding attitudes; basal slip surfaces most commonly form in the Shnabkaib Member of the Moenkopi Formation, the Petrified Forest Member of the Chinle Formation, the Co-op Creek Limestone Member of the Carmel Formation, the Dakota Formation, and the upper unit of the Straight Cliffs Formation, and the slides incorporate these and overlying map units; the Petrified Forest Member and Dakota Formation especially form large, complex mass movements; undivided as to inferred age because new research shows that even landslides having subdued morphology (suggesting that they are older, weathered, and have not experienced recent large-scale movement) may continue to exhibit slow creep or are capable of renewed movement if stability thresholds are exceeded (Ashland, 2003); generally several tens of feet thick, but some deposits may exceed 200 feet (60 m) thick.

Qmp **Older landslides, colluvium, and pediment alluvium** (Holocene to Pleistocene) – Remnants of poorly sorted rock-fall, landslide, colluvial, and generally minor alluvial-fan debris that mantle and armor gently sloping, pediment-like benches cut across bedrock; consists of angular and subangular, house-sized boulders to fine-grained sand and lesser amounts of silt and clay derived from nearby cliffs and ledges; preserved as remnants that form inclined benches near steep bedrock slopes, high above modern drainages, in and near the southwest part of Zion National Park; as mapped, includes talus above the benches in the Smith Mesa quadrangle; these benches may be either remnants of much larger surfaces that were graded to the ancestral Virgin River, which at the time of deposition must have been at least several hundred feet above its present position, or, alternatively, these benches are the remnants of sloping erosional surfaces mantled and protected from erosion by the coarse deposits and were not graded to the river; as much as 30 feet (9 m) thick; graded to several levels that project as much as 700 feet (210 m) above the modern river channel.

Mixed-environment deposits

Qac, Qaco

Alluvium and colluvium (Holocene to upper Pleistocene) – Poorly to moderately sorted, generally poorly stratified, clay- to boulder-size, locally derived sediment deposited principally in swales, small drainages, and the upper reaches of large streams by fluvial, slope-wash, and creep processes; gradational with both alluvium and colluvium; older deposits (Qaco) include gypcrete (silt to small boulders cemented by gypsum) along the Moenkopi outcrop belt just west of St. George that is as much as 10 feet (3 m) thick; older deposits form inactive surfaces commonly incised by younger (Qac) deposits; generally less than 30 feet (9 m) thick.

Qea, Qeao

Eolian sand and alluvium (Holocene to upper Pleistocene) – Well-sorted, fine- to medium-grained, reddish-brown eolian sand reworked by alluvial processes, and poorly to moderately sorted gravel, sand, and silt deposited in small channels; younger deposits (Qea) form active depositional surfaces, whereas older deposits (Qeao) typically form incised, inactive surfaces; as much as about 20 feet (6 m) thick.

Qae, Qaeo

Alluvium and eolian sand (Holocene to upper Pleistocene) – Moderately sorted gravel, sand, and silt deposited in small channels and on alluvial flats, and well-sorted, fine- to medium-grained, reddish-brown eolian sand locally reworked by alluvial processes; younger deposits (Qae) form active depositional surfaces, whereas older deposits (Qaeo) typically form incised, inactive surfaces; as much as about 20 feet (6 m) thick.

Qca **Colluvium and alluvium** (Holocene to middle[?] Pleistocene) – Poorly sorted, angular to subrounded, fine-grained to boulder-sized sediment deposited on moderate slopes by slope-wash, debris-flow, and alluvial processes; along the base of the Pine Valley laccolith, unit includes undifferentiated alluvial fans and talus with both active and deeply dissected surfaces; near St. George, forms broad, low- to moderate-relief slopes that lack a well-defined drainage pattern, and near the Washington fault it is moderately incised; west of Beaver Dam Wash, consists of gypsiferous, clay- to cobble-size sediment that forms deeply incised erosional remnants overlying the Muddy Creek Formation (Tmc); typically 20 to 30 feet (6–9 m) thick, but deposits on the flanks of the Pine Valley Mountains may locally exceed 80 feet (25 m) thick.

Qeac **Mixed eolian sand and alluvium with well-developed calcic soils** (Holocene to middle Pleistocene) – White, laminated calcic soil (hardpan or caliche) with crinkle bedding and pisoliths (stage V carbonate development of Birkeland and others, 1991); forms resistant cap on Quaternary/Tertiary alluvial-fan deposits (QTaf) and the Muddy Creek Formation (Tmc) west of the Beaver Dam Mountains; typically 5 to 30 feet (2–9 m) thick.

Qer **Eolian sand and residuum** (Holocene to upper Pleistocene) – Reddish-orange, fine- to medium-grained sand containing residual Navajo Sandstone pebbles, cobbles, and boulders; deposited in shallow topographic depressions and on gently sloping surfaces mostly on the Navajo Sandstone, from which it is derived; generally less than 20 feet (6 m) thick.

Qmr **Residuum of basaltic lava flows** (Holocene to lower Pleistocene) – Residual lag of angular to subangular basaltic blocks derived from the Lava Point lava flow (northeast of Virgin) and the Little Creek Peak lava flow (south of Kolob Reservoir); probably as much as several tens of feet thick.

Basaltic lava flows

Basaltic lava flows in the St. George 30' x 60' quadrangle are part of the Western Grand Canyon basaltic field, which extends across the southwest part of the Colorado Plateau and adjacent transition zone in southwest Utah, northeast Arizona, and adjacent Nevada (Hamblin, 1963, 1970, 1987; Best and Brimhall, 1970, 1974; Best and others, 1980; Smith and others, 1999). This basaltic field contains hundreds of relatively small-volume, widely scattered basaltic lava flows and cinder cones that range in age from late Miocene to Holocene (in addition, a couple of andesitic lava flows are present in the northwest part of the map area). The oldest basaltic lava flows in the St. George 30' x 60' quadrangle are a 17-million-year-old (early Miocene) basaltic flow widely exposed north of Pine Valley, and a 9-million-year-old (late Miocene) basaltic flow near the base of the Muddy Creek Formation along Beaver Dam Wash. All other basaltic lava flows in the quadrangle are younger than 2.5 Ma, and the youngest lava flow is the newly ^{14}C-dated 27,000-year-old (about 32,000 calendar years) Santa Clara lava flow at and near Snow Canyon State Park (Willis and others, 2006). Most lava flows in the quadrangle are less than about one million years old; three temporal clusters occur between about 250,000 and 350,000; 550,000 and 700,000; and 950,000 and 1,050,000 years ago. Table 1 (p. 50) summarizes the age of the lava flows. The illustration on plate 2 shows apparent clustering of age populations and that many of the vents are associated with major faults. Statistically, basaltic flows in this area become younger eastward, like many other basaltic lava fields near the western margin of the Colorado Plateau, although no simple age progression exists.

The basaltic magmas are partial melts derived from the compositionally heterogeneous lithospheric mantle, and this, coupled with fractional crystallization, may account for most of the geochemical variability between individual lava flows (Lowder, 1973; Best and Brimhall, 1974; Leeman, 1974; Nelson and Tingey, 1997; Nusbaum and others, 1997; Smith and others, 1999; Downing, 2000). Whole-rock groundmass concentrate ^{40}Ar/^{39}Ar raw data and major- and trace-element geochemistry for these lava flows are available in analytical reports by NMGRL and UGS (2006a, 2006b, 2006c), UGS and NMGRL (2007a, 2007b, 2007c, 2008), Biek and Ehler (2007), and on the Utah Geological Survey Web site http://geology.utah.gov/online/analytical_data.htm. Rock names are derived from the total alkali vs. silica diagram of LeBas and others (1986).

Lava flows typically have a rubbly base, a dense, jointed middle part, and, if not eroded away, a vesicular upper part that has a rough aa (a Hawaiian term for a blocky, jagged flow) or, rarely, a poorly developed pahoehoe (a Hawaiian term for a smooth or ropy flow) surface. The flows commonly overlie stream-gravel and other surficial deposits and are partly covered by eolian sand and calic soil (caliche) not shown on this map. Most lava flows are dark gray, fine grained, and contain small olivine phenocrysts, and with few exceptions are difficult to differentiate by hand sample alone. They are differentiated only through detailed geologic mapping in concert with trace-element geochemistry and isotopic ages. Lava tubes are locally present, especially near vent areas, and in one locality, near the State Highway 9 bridge linking Hurricane and LaVerkin, pillow basalts (indicative of deposition under water) are present at the base of a flow where it blocked the ancestral Virgin River (Biek, 2003b; Biek and Hayden, 2007).

Most lava flows are 10 to 40 feet (3–12 m) thick, but they are as much as several hundred feet thick where they fill ancient stream channels. Most also consist of multiple flow or cooling units that range from a few feet to a few tens of feet thick, each unit representing a pulse of magma separated by enough time for cooling but not significant weathering to occur. For example, the south end of the Washington lava flow in the Harrisburg Junction quadrangle consists of three cooling units that together constitute the flow itself. Some lava flows, such as the Volcano Mountain and Veyo flows, have a longer, more complicated eruptive history and are themselves composed of several distinct flows.

Because each flow was emplaced in a "geological instant" (most small basaltic volcanoes are monocyclic, meaning that each vent produces only one eruptive cycle that may last less than a year or as much as a few tens of years in duration), flowed several miles or more across the landscape, and is resistant to erosion, these lava flows provide a "snapshot" of the local landscape as it existed when the flow erupted. As the lava flows blocked drainages, the streams commonly briefly re-established themselves on top of the flows as evidenced by thin gravel deposits. However, the streams soon moved off to the side where they preferentially eroded softer sedimentary bedrock, ultimately leaving the resistant lava flows stranded as elevated, sinuous ridges (called inverted valleys) that mark the location of former channels. The greater St. George area is justly famous for these classic examples of inverted topography, such as Washington and Middleton Black Ridges, as first described in detail by Hamblin (1963, 1970, 1987) and Hamblin and others (1981).

Long-term incision rates for the Zion-St. George area—calculated using isotopic ages for these lava flows and their height above major drainages—differ systematically across the region from 0.2 to 1.25 feet per thousand years (0.06 to 0.38 m/kyr) (figure 66 and table 2) (Willis and Biek, 2001). Our calculations reconfirm and expand on many of the findings of Hamblin and others (1981), who similarly documented incision rates in the St. George basin. Several of the lava flows crossed and are now offset by the Hurricane fault, providing important constraints on fault slip rates over the past one million years (Lund and others, 2001, 2002, 2007a). Thus, in this area, we know that the amount and rate of long-term incision along the Virgin River is largely a function of base-level lowering along faults at and near the boundary of the Basin and Range Province. Long-term incision rates are highest on the upthrown (east) side of the Hurricane fault, the most active fault in the region.

Thus, position on structural blocks is important when estimating relative ages of lava flows based on the amount of "topographic inversion" ("stage" designations of Hamblin, 1963, 1970, 1987). For example, the lower reaches of Middleton and Washington Black Ridges, capped by the Lava Ridge and Washington lava flows, respectively, are classic inverted valleys. The two flows are about 5 miles (8 km) apart and both flowed into a well-graded stretch of the ancestral Virgin River. The Washington flow stands about twice as high above the present Virgin River as the Lava Ridge flow, yet $^{40}Ar/^{39}Ar$ ages now show that it is about 500,000 years younger than the Lava Ridge flow (see figure 66). The greater topographic inversion on the younger flow is directly attributable to its position on the footwall (upthrown part) of a separate, relatively more elevated structural block. The old axiom that "the older the lava flow, the higher it now stands above adjacent drainages" is generally true only when comparing flows on the same structural block.

Qb, Qbc

Basaltic lava flows and cinder deposits, undivided (Pleistocene) – Isolated basaltic lava flows (Qb) and cinder deposits (Qbc) of uncertain correlation along the Virgin River near Hurricane, and along the Santa Clara River near Veyo and Central.

Qbs, Qbsc

Santa Clara lava flow and cinder cone (upper Pleistocene) – Dark-brownish-black to black subalkaline basalt (Qbs) that contains abundant small olivine phenocrysts in an aphanitic groundmass; has a jagged aa surface and, with the nearby and chemically similar Diamond Valley lava flow, is the youngest flow in the St. George basin; an iridescent sheen is locally present on protected surfaces; erupted from a vent at cinder cone (Qbsc) at south end of Diamond Valley north of St. George; our attempts at $^{40}Ar/^{39}Ar$ dating yielded unreliable ages, likely due to insufficient argon due to the young age of the flow (UGS unpublished data); we obtained a new radiocarbon age of 27,270 ± 250 ^{14}C yr B.P. (about 32,000 calendar years) from charcoal discovered in loose sand just underneath the flow (Willis and others, 2006), and the sand itself yielded an optically stimulated luminescence (OSL) age of about 40,000 calendar years (Tammy Rittenour, Utah State University Luminescence Laboratory, written communication, February 2008), suggesting that the sand is part of older mixed alluvial and eolian deposits adjacent to the ancestral Snow Canyon wash that existed before emplacement of the flow; lava flow is 10 to 30 feet (3–9 m) thick, but locally thicker where it fills paleotopography.

Qbdv, Qbdvc

Diamond Valley lava flow and cinder cone (upper Pleistocene) – Chemically and petrographically similar to the Santa Clara lava flow, but contains rare plagioclase phenocrysts (Faust, 2005; see also Faust and Smith, 2005); not dated, but morphology of lava flow (Qbdv) and cinder cone (Qbdvc) suggests that it is slightly younger than the much larger Santa Clara flow; lava flow is 10 to 30 feet (3–9 m) thick, but locally thicker where it fills paleotopography.

Qbc, Qbcc

Crater Hill lava flow and cinder cone (middle? Pleistocene) – Medium-gray basalt to trachybasalt (Qbc) with small olivine phenocrysts; erupted from a vent at Crater Hill, which is a large cinder cone (Qbcc) east of Virgin; upper surface of flow has large arcuate flow ridges and one large rafted and tilted block once considered to be a separate cinder cone and vent (Nielson, 1977); this flow blocked the ancestral Virgin River, creating Lake Grafton, once the largest Pleistocene lake in the map area, which reached upstream into the lower part of Zion Canyon; three samples yielded $^{40}Ar/^{39}Ar$ plateau ages of

0.310 ± 0.070 Ma (0.298 ± 0.032 Ma isochron), 0.320 ± 0.130 Ma (0.294 ± 0.018 Ma preliminary isochron), and 0.280 ± 0.080 Ma (0.228 ± 0.040 Ma preliminary isochron) (UGS unpublished data); however, this age of about 300,000 years is older than expected given long-term incision rates of about 1.15 feet per thousand years (0.35 m/kyr) (Willis and Biek, 2001), and the lab had low confidence in the analyses; an optically stimulated luminescence (OSL) age of 122.30 ± 15.73 ka was determined for sediments of Lake Grafton (Tammy Rittenour, Utah State University Luminescence Laboratory, written communication, August 2008), which better fits long-term incision rates; lava flow is typically 40 to 80 feet (12–24 m) thick, but locally as much as 400 feet (120 m) thick where it ponded in ancestral Virgin River and Coal Pits Wash channels.

Qbrt, Qbrtc

Radio Tower lava flow and cinder cones (middle Pleistocene) – Medium- to dark-gray, fine-grained basalt to basanite (Qbrt) with small olivine phenocrysts; erupted from vents at four overlapping cinder cones (Qbrtc) west of Hurricane; yielded an $^{40}Ar/^{39}Ar$ plateau age of 0.14 ± 0.06 Ma (0.13 ± 0.07 Ma isochron) (Biek, 2003b); lava flow is as much as 180 feet (55 m) thick where it fills the ancestral Virgin River channel.

Qbcp, Qbcpc

Cinder Pits lava flow and cinder cones (middle Pleistocene) – Medium- to dark-gray, fine-grained basalt (Qbcp) with small olivine phenocrysts; yielded an $^{40}Ar/^{39}Ar$ age of 0.24 ± 0.02 Ma (Biek, 2003b); erupted from vents at two overlapping cinder cones (Qbcpc) northwest of Hurricane; lava flow is as much as several tens of feet thick.

Qbg, Qbgc

Grapevine Wash lava flows and cinder cones (middle Pleistocene) – Medium-gray, fine-grained basaltic trachyandesite lava flows (Qbg) with small olivine phenocrysts; erupted from a number of vents on the Lower Kolob Plateau, including the Firepit Knoll and Spendlove Knoll cinder cones (Qbgc); five $^{40}Ar/^{39}Ar$ plateau ages on these flows range from 0.22 ± 0.03 Ma to 0.31 ± 0.02 Ma (Willis and Hylland, 2002), consistent with a K-Ar age of 0.26 ± 0.09 Ma (Best and others, 1980); typically 20 to 40 feet (6–12 m) thick, but locally as much as 400 feet (120 m) thick where ponded in paleodrainages.

Qber, Qberc

East Reef lava flow and cinder cones (middle Pleistocene) – Medium-dark-gray, fine-grained basanite (Qber) with small olivine phenocrysts; erupted from vents at two overlapping cinder cones (Qberc) about 2 miles (3 km) southeast of Leeds; probably about 250,000 to 350,000 years old based on comparison with neaby lava flows; lava flow is about 20 to 30 feet (6–9 m) thick.

Qbv, Qbvc

Volcano Mountain lava flow and cinder cone (middle Pleistocene) – Medium- to dark-gray to grayish-black, fine- to medium-grained alkali basalt (Qbv) with sparse olivine phenocrysts; erupted from a vent at Volcano Mountain (Qbvc) southwest of Hurricane; divisible into three separate flows (Biek, 2003a, 2003b); youngest and middle-level flows yielded $^{40}Ar/^{39}Ar$ ages of 0.258 ± 0.024 Ma and 0.353 ± 0.045 Ma, respectively (Sanchez, 1995), in accord with K-Ar ages of 0.289 ± 0.085 and 0.303 ± 0.30 (Best and others, 1980) on the middle-level flow; middle-level flow is displaced about 240 feet (73 m) by the Hurricane fault at Timpoweap Canyon and locally has pillow basalt at its base; the oldest lava flow flowed about 8 miles (13 km) down the Virgin River; lava flows are generally 35 to 45 feet (11–14 m) thick and form rough, blocky surfaces, but the middle-level flow is as much as 170 feet (50 m) thick where it fills the ancestral Virgin River channel.

Qblc, Qblcc

Little Creek lava flow and cinder cone (middle Pleistocene) – Medium-gray, fine-grained basalt to trachybasalt (Qblc) with sparse olivine phenocrysts; erupted from a vent at Gray Knoll cinder cone (Qblcc) and from a series of northwest-trending vents marked by spatter cones on top of Little Creek Mountain; yielded an $^{40}Ar/^{39}Ar$ plateau age of 0.345 ± 0.015 Ma (Downing, 2000); lava flow is generally 15 to 40 feet (5–12 m) thick.

Qbgw, Qbgwc

Gould Wash lava flow and cinder cones (middle Pleistocene) – Dark-gray, fine-grained basalt (Qbgw) with abundant olivine phenocrysts; erupted from vents at two cinder cones (Qbgwc) about 6 miles (10 km) southeast of Hurricane; yielded an $^{40}Ar/^{39}Ar$ isochron age of 0.278 ± 0.018 Ma (Downing, 2000), whereas our sample VR41-08 yielded $^{40}Ar/^{39}Ar$ ages of 0.420 ± 0.210 Ma (plateau) and 0.420 ± 0.05 Ma (isochron) (UGS unpublished data); the flow exhibits a level of incision comparable to the adjacent 400,000-year-old Divide lava flow, and so is likely about 400,000 years old; lava flow is generally 20 to 30 feet (6–9 m) thick.

Qbd, Qbdc

Divide lava flow and cinder cones (middle Pleistocene) – Dark-gray, fine-grained basalt to basanite (Qbd) with small olivine phenocrysts; forms lava cascade over Hurricane Cliffs south of Hurricane (section 3, T. 43 S., R. 13 W.); erupted from vents at two cinder cones (Qbdc) and two north-trending dikes; yielded an $^{40}Ar/^{39}Ar$ age of 0.41 ± 0.08 Ma (Hayden, 2004a); lava flow is generally 15 to 40 feet (5–12 m) thick.

Qvb, Qvbc

Andesite of Black Hills and cinder cone (middle Pleistocene) – Resistant, black andesite lava flows containing various percentages of phenocrysts as large as 0.2 inch (5 mm) of plagioclase and smaller pyroxene; erupted in part from a cinder cone (Qvbc) just south of the lava flow; mapped by Blank (1959, 1993) southwest of Enterprise, at the north edge of the map area; yielded a whole-rock $^{40}Ar/^{39}Ar$ age of 0.42 $\pm$ 0.13 Ma (Rowley and others, 2006); thickness about 650 feet (200 m).

Qbsm, Qbsmc

Saddle Mountain lava flow and cinder cone (middle Pleistocene) – Dark-gray, fine-grained lava flow (Qbsm) that ranges in composition from basalt to trachybasalt to basaltic trachyandesite; contains small olivine and plagioclase phenocrysts; erupted from a vent at a cinder cone (Qbsmc) about 6 miles (10 km) east of Veyo; yielded an $^{40}Ar/^{39}Ar$ preferred isochron age of 0.47 $\pm$ 0.12 Ma (0.74 $\pm$ 0.07 Ma low-confidence plateau) (UGS and NMGRL, 2008); lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbce, Qbcec

Central-area lava flow and cinder cone (middle to lower Pleistocene) – Dark-gray basaltic trachyandesite (Qbce) with small olivine phenocrysts; erupted from four vents at cinder cones (Qbcec) just south of Central; overlies the 600,000-year-old Lark Canyon lava flow and so is probably 500,000 to 600,000 years old; lava flow is generally 20 to 40 feet (6–12 m) thick

Qbde, Qbdec

Dameron Valley East lava flow and cinder cone (middle Pleistocene) – Dark-gray trachybasalt (Qbde) with small olivine phenocrysts in an aphanitic groundmass; erupted from a vent at a cinder cone (Qbdec) about 2 miles (3 km) southeast of Veyo; yielded an $^{40}Ar/^{39}Ar$ plateau age of 0.59 $\pm$ 0.02 Ma (0.59 $\pm$ 0.03 Ma isochron) (UGS and NMGRL, 2008); lava flow is generally 10 to 20 feet (3–6 m) thick.

Qbla, Qblac

Lark Canyon lava flow and cinder cone (middle Pleistocene) – Dark-gray basalt (Qbla) with small olivine phenocrysts; erupted from a vent at a cinder cone (Qblac) about 2 miles (3 km) southwest of Pine Valley; yielded an $^{40}Ar/^{39}Ar$ plateau age of 0.61 $\pm$ 0.04 Ma (0.64 $\pm$ 0.04 Ma isochron) (UGS and NMGRL, 2007a) and a K-Ar age of 0.56 $\pm$ 0.06 Ma (Best and others, 1980); lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbmk, Qbmkc

Mahogany Knoll lava flow and cinder cone (middle[?] Pleistocene) – Dark-gray basalt (Qbmk) with small olivine phenocrysts; erupted from vents at cinder cones (Qbmkc) on the southwest flank of the Pine Valley Mountains; yielded a K-Ar age of 1.2 $\pm$ 0.1 Ma (Best and others, 1980), but based on geomorphic expression is believed to be younger and of comparable age to nearby lava flows that are about 600,000 years old; lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbrk, Qbrkc

Red Knoll lava flow and cinder cone (middle[?] Pleistocene) – Gray andesite to trachyandesite (Qbrk) that erupted from a vent at a cinder cone (Qbrkc) on the southwest flank of the Pine Valley Mountains; yielded a low-confidence $^{40}Ar/^{39}Ar$ integrated age of 0.45 $\pm$ 0.86 Ma (UGS and NMGRL, 2007a), but based on geomorphic expression is probably about 450,000 to 700,000 years old; lava flow is generally 30 to 60 feet (9–18 m) thick.

Qbtb, Qbtbc

Truman Bench lava flow and cinder cone (middle[?] Pleistocene) – Dark-gray basalt to trachybasalt (Qbtb) with small olivine phenocrysts; erupted from a vent at a cinder cone (Qbtbc) on the southwest flank of the Pine Valley Mountains; probably about 450,000 to 700,000 years old based on comparison with nearby flows; lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbbd **Baker Dam lava flow** (middle Pleistocene) – Dark-gray basaltic trachyandesite that contains abundant small olivine and some plagioclase phenocrysts in an aphanitic groundmass; the age and chemistry of this flow suggest that it may be the distal part of the Lark Canyon lava flow; yielded $^{40}Ar/^{39}Ar$ plateau ages of 0.670 $\pm$ 0.040 Ma and 0.690 $\pm$ 0.08 Ma (0.686 $\pm$ 0.022 Ma isochron) (UGS unpublished data); as much as about 50 feet (15 m) thick.

Qbpv, Qbpvc

Pine Valley lava flow and cinder cones (middle Pleistocene) – Dark-gray basaltic lava flows (Qbpv) with small olivine phenocrysts; erupted from a number of vents at cinder cones (Qbpvc) that show a striking northeast alignment just west of Pine Valley; lava flows once blocked ancestral Santa Clara River, allowing sediment to accumulate behind the lava dam and create the wide, level floor of Pine Valley; yielded an $^{40}Ar/^{39}Ar$ plateau age of 0.67 $\pm$ 0.07 (0.67 $\pm$ 0.08 Ma isochron) (UGS and NMGRL, 2007a); lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbve, Qbvec

Veyo lava flow and cinder cone (middle Pleistocene) – Dark-gray, fine-grained basaltic trachyandesite (Qbve) with abundant olivine and sparse plagioclase phenocrysts; erupted from a vent at the Veyo Volcano cinder cone (Qbvec) just south of Veyo; divisible into five separate flows (Embree, 1970; Higgins, 2002); yielded $^{40}Ar/^{39}Ar$

plateau age of 0.69 ± 0.04 Ma (0.71 ± 0.08 Ma isochron) (UGS and NMGRL, 2007c); lava flow is generally 20 to 50 feet (6–15 m) thick.

Qbhp, Qbhpc

Hornet Point lava flow and cinder cone (middle Pleistocene) – Medium- to dark-gray, medium- to coarse-grained basalt to trachybasalt (Qbhp) with small olivine and abundant pyroxene phenocrysts; locally deeply weathered to grus-like soils, and isolated boulders typically have concentric weathering rinds; erupted from a vent at a deeply eroded cinder cone (Qbhpc) at Hornet Point; yielded $^{40}Ar/^{39}Ar$ isochron age of 0.74 ± 0.05 Ma (Biek and Hylland, 2007); lava flow is as much as 240 feet (73 m) thick.

Qbca, Qbcac

Canal lava flow and cinder cone (middle to lower Pleistocene) – Brownish-gray basalt (Qbca) with sparse olivine and some plagioclase phenocrysts; erupted from a vent at a small cinder cone (Qbcac) about 2 miles (3 km) northeast of Veyo; appears to partially conceal the Aqueduct Hill, Magotsu Creek, and Central East lava flows; probably about 600,000 to 900,000 years old based on geomorphic comparison with nearby flows; lava flow is typically about 20 feet (6 m) thick.

Qbck, Qbckc

Cedar Knoll lava flow and cinder cone (middle to lower Pleistocene) – Dark-gray basalt (Qbck) with small olivine phenocrysts; erupted from a vent at a cinder cone (Qbckc) about 2 miles (3 km) northwest of Pine Valley; probably about 600,000 to 900,000 years old based on geomorphic comparison with nearby flows; lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbgf, Qbgfc

Grassy Flat lava flow and cinder cone (middle to lower Pleistocene) – Dark-gray basaltic trachyandesite (Qbgf) with small olivine phenocrysts; erupted from a vent at a cinder cone (Qbgfc) about 3 miles (5 km) north of Pine Valley; probably about 600,000 to 900,000 years old based on geomorphic comparison with nearby flows; lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbf, Qbfc

Fourmile Bench lava flow and cinder cone (middle to lower Pleistocene) – Dark-gray basaltic trachyandesite (Qbf) with small olivine phenocrysts; erupted from a vent at a cinder cone (Qbfc) about 2 miles (3 km) north of Pine Valley; probably about 600,000 to 900,000 years old based on geomorphic comparison with nearby flows; lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbp, Qbpc

Pintura lava flow and cinder cones (lower Pleistocene) – Medium- to dark-gray, fine- to medium-grained basaltic lava flows (Qbp) with small olivine phenocrysts; geochemically divisible into three groups: basalt, trachybasalt, and basaltic trachyandesite to basaltic andesite (Hatfield, 2001); locally contains abundant small plagioclase phenocrysts; erupted principally from the Pintura volcanic center (Qbpc) about 2.5 miles (4 km) north of Pintura, and from a smaller vent atop Black Ridge; displaced in excess of 1200 feet (365 m) by the Hurricane fault near Pintura; yielded five $^{40}Ar/^{39}Ar$ ages of 0.89 ± 0.02 Ma, 0.81 ± 0.10 Ma, 0.87 ± 0.04 Ma, 0.88 ± 0.05 Ma, and 0.84 ± 0.03 Ma (Lund and others, 2001, 2002, 2007a; Biek, 2003b); lava flows are typically 20 to 40 feet (6–12 m) thick, but as much as 1140 feet (350 m) thick where they fill paleotopography on Black Ridge.

Qbgk, Qbgkc

Grass Knoll lava flow and cinder cone (middle to lower Pleistocene) – Dark-gray basalt to trachybasalt (Qbgk) with small olivine phenocrysts; erupted from a vent at the Grass Knoll cinder cone (Qbgkc) on the southwest flank of the Pine Valley Mountains; yielded an $^{40}Ar/^{39}Ar$ integrated age of 1.02 ± 0.36 Ma (1.20 ± 0.17 Ma isochron) (UGS and NMGRL, 2007a); lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbw, Qbwc

Washington lava flow and cinder cone (lower Pleistocene) – Medium- to dark-gray to dark-greenish-gray, fine-grained basanite to picrobasalt (Qbw) with abundant clinopyroxene and olivine phenocrysts; erupted from a vent at a cinder cone (Qbwc) about 3 miles (5 km) northeast of Washington; yielded $^{40}Ar/^{39}Ar$ ages of 0.87 ± 0.04 and 0.98 ± 0.02 Ma (Biek, 2003a), which fits well with regional incision rates (Willis and Biek, 2001), but Best and others (1980) reported an anomalously old K-Ar age of 1.7 ± 0.1 Ma for this flow; lava flow is 25 to 35 feet (8–11 m) thick except near its source, where it is as much as about 100 feet (30 m) thick.

Qvcw **Central West andesitic porphyry lava flow** (lower Pleistocene) – Medium-gray to light-brownish-gray, coarse-grained lava flow with large plagioclase and slightly smaller quartz phenocrysts; flow composition varies and includes trachyandesite, andesite, dacite, and trachydacite; erupted from a vent or vents just west of Baker Dam Reservoir; yielded an $^{40}Ar/^{39}Ar$ plateau age on hornblende of 1.77 ± 0.09 Ma (1.74 ± 0.27 Ma isochron) (UGS and NMGRL, 2008) and an $^{40}Ar/^{39}Ar$ plateau age on plagioclase of 0.920 ± 0.07 (0.930 ± 0.022 Ma preliminary isochron) (UGS unpublished data), but the reason for the age discrepancy is unknown; the flow is probably close to one million years old based on geomorphic comparison to nearby lava flows; forms hilly, rugged topography similar to that of the andesite of Black Hills (Qvb); locally in excess of 400 feet (120 m) thick.

Qbi, Qbic

Ivans Knoll lava flow and cinder cone (lower Pleistocene) – Medium-gray, fine- to medium-grained basalt (Qbi) with small olivine phenocrysts; erupted from a vent marked by a deeply eroded cinder cone (Qbic) immediately south of Volcano Mountain, southwest of Hurricane; remnant of flow caps Mollies Nipple at the top of the Hurricane Cliffs, showing about 1300 feet (400 m) of displacement along the Hurricane fault; yielded two $^{40}Ar/^{39}Ar$ plateau ages of 0.97 ± 0.07 and 1.03 ± 0.02 Ma (1.04 ± 0.06 and 1.05 ± 0.02 Ma isochron, respectively) (Biek, 2003b), and sample VR41-06 yielded an $^{40}Ar/^{39}Ar$ age of 0.937 ± 0.006 Ma (preliminary isochron) and 1.06 ± 0.16 Ma (plateau) (UGS unpublished data); this lava flow has a normal magnetic polarity signature (Michael J. Hozik, The Richard Stockton College of New Jersey, verbal communication, 1998) and thus appears to have erupted during the Jamarillo normal event (see, for example, Glen, 1982), a short-duration flip in the Earth's magnetic field that occurred about 0.99 to 1.07 million years ago during the Matuyama reversed polarity epoch; lava flow is generally 20 to 30 feet (6–9 m) thick.

Qbmc, Qbmcc

Magotsu Creek lava flow and cinder cone (lower Pleistocene) – Medium- to dark-gray, medium-grained trachybasalt to basaltic trachyandesite (Qbmc) with sparse plagioclase, sparse olivine, and rare quartz phenocrysts; erupted from a vent at a cinder cone (Qbmcc) about 2 miles (3 km) northeast of Veyo; yielded $^{40}Ar/^{39}Ar$ plateau ages of 0.98 ± 0.03 Ma (0.99 ± 0.03 Ma isochron) (UGS and NMGRL, 2008) and 1.00 ± 0.09 Ma (0.994 ± 0.023 Ma isochron) (UGS unpublished data) and a K-Ar age of 1.1 ± 0.1 Ma (Best and others, 1980); lava flow is generally 10 to 50 feet (3–15 m) thick.

Qbgvr, Qbgvrc

Grass Valley Reservoir lava flow and cinder cone (lower Pleistocene) – Dark-gray basaltic lava flows (Qbgvr) with small olivine phenocrysts; erupted from vent at cinder cone (Qbgvrc) southwest of Grass Valley; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.08 ± 0.13 Ma (1.12 ± 0.27 Ma isochron) (UGS and NMGRL, 2007a); lava flow is generally 20 to 40 feet (6–12 m) thick.

Qbgv, Qbgvc

Grass Valley lava flow and cinder cone (lower Pleistocene) – Dark-gray, fine- to medium-grained trachybasalt to basalt (Qbgv) with small olivine phenocrysts; erupted from a vent at a deeply eroded cinder cone (Qbgvc) about 7 miles (11 km) south of Hurricane; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.09 ± 0.09 Ma (0.966 ± 0.030 Ma preliminary isochron) (UGS unpublished data); lava flow is several tens of feet thick.

Qbr **Remnants lava flow** (lower Pleistocene) – Dark-brownish-black to dark-gray, medium-grained basanite with small olivine phenocrysts; erupted from a deeply eroded vent atop the Hurricane Cliffs at the "Three Brothers," about 5 miles (8 km) south of Hurricane; displaced by the Hurricane fault about 1450 feet (440 m); yielded preferred $^{40}Ar/^{39}Ar$ plateau ages of 1.06 ± 0.03 Ma (1.07 + 0.08 Ma isochron) and 0.94 ± 0.04 Ma (0.94 ± 0.05 Ma isochron) (Hayden, 2004a) and an anomalous $^{40}Ar/^{39}Ar$ plateau age of 1.47 ± 0.34 Ma (1.12 ± 0.50 Ma isochron) (Lund and others, 2001, 2007a); typically about 40 feet (12 m) thick.

Qblp, Qblpc

Lava Point lava flow and cinder cones (lower Pleistocene) – Light- to medium-gray, fine- to medium-grained basaltic trachyandesite to basaltic andesite and trachybasalt (Qblp) with small olivine phenocrysts; erupted from vents at Home Valley Knoll, a group of three overlapping cinder cones (Qblpc) on the Kolob Plateau; the distal edge of this flow caps a mesa 1300 feet (400 m) above the modern Virgin River, near the town of Virgin, and shows that the lower half of Zion Canyon was eroded in the past one million years; yielded $^{40}Ar/^{39}Ar$ plateau ages of 1.02 ± 0.03 Ma (1.04 ± 0.02 Ma isochron), 1.06 ± 0.01 Ma, 1.08 ± 0.02 Ma (1.09 ± 0.02 Ma isochron), and 1.14 ± 0.16 Ma (0.96 ± 0.02 Ma isochron) (Willis and Hylland, 2002; Biek, 2007b), consistent with two of four published K-Ar ages (Best and others, 1980; Hamblin and others, 1981); lava flow is typically 20 to 40 feet (6–12 m) thick, but as much as 200 feet (60 m) thick where it fills paleodrainages.

Qbhr **Horse Ranch Mountain lava flow** (lower Pleistocene) – Medium-gray, fine- to medium-grained basalt with small olivine phenocrysts; caps Horse Ranch Mountain in the Kolob Canyons part of Zion National Park; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.03 ± 0.06 Ma (1.05 ± 0.17 Ma isochron) (Biek, 2007b); source of the lava flow is unknown, but likely to the northeast of the map area in the Cedar Mountain quadrangle; 30 to 125 feet (10–38 m) thick.

Qbkp, Qbkpc

Kolob Peak lava flow and cinder cone (lower Pleistocene) – Medium- to light-gray, fine-grained basaltic trachyandesite (Qbkp) with small olivine phenocrysts; erupted from a vent at Kolob Peak, a cinder cone (Qbkpc) now eroded in half; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.05 ± 0.05 Ma (Biek, 2007b) and a K-Ar age of 1.13 ± 0.05 Ma (Best and others, 1980); thickness of lava flow uncertain, but likely in excess of 100 feet (30 m) thick where it fills paleodrainages.

Qbb, Qbbc

Big Sand lava flow and cinder cone (lower Pleistocene) – Dark-reddish-gray to dark-brownish-gray basaltic trachyandesite (Qbb), locally with large quartz and plagioclase and small olivine phenocrysts; erupted from a vent at a cinder cone (Qbbc) about 5 miles (8 km) north of St. George; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.13 $\pm$ 0.05 Ma (1.111 $\pm$ 0.007 Ma isochron) (UGS unpublished data); lava flow is generally 10 to 50 feet (3-15 m) thick.

Qbcb, Qbcbc

Cedar Bench lava flow and cinder cones (lower Pleistocene) – Dark-greenish-gray to brownish-black trachybasalt (Qbcb) with small phenocrysts of clinopyroxene and olivine; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.23 $\pm$ 0.01 Ma (UGS unpublished data); based on their similar geochemistry, includes the Snow Canyon Overlook lava flow of Willis and Higgins (1996), which yielded a K-Ar age of 1.2 $\pm$ 0.1 Ma (Best and others, 1980) and an $^{40}Ar/^{39}Ar$ plateau age of 1.16 $\pm$ 0.03 Ma (UGS unpublished data), and the Airport lava flow of Higgins and Willis (1995), which yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.23 $\pm$ 0.01 Ma (UGS unpublished data), that is somewhat older than the K-Ar age of 1.07 $\pm$ 0.04 Ma reported by Hamblin and others (1981); erupted from vents at two overlapping cinder cones (Qbcbc) about 12 miles (20 km) north of St. George; lava flow is typically 10 to 30 feet (3-9 m) thick, but as much as about 100 feet (30 m) thick where it fills paleotopography.

Qbl, Qblc

Lava Ridge lava flow and cinder cone (lower Pleistocene) – Moderate- to dark-gray to dark-brownish-gray basaltic trachyandesite (Qbl) with prominent euhedral plagioclase phenocrysts, common quartz and pyroxene phenocrysts, and small olivine phenocrysts; petrographic and limited geochemical data suggest that the Middleton lava flow of Higgins and Willis (1995), which forms the spectacular inverted valley of Middleton Black Ridge, is part of the Lava Ridge flow; erupted from a group of heavily weathered cinder cones (Qblc) on Lava Ridge, about 8 miles (13 km) north of St. George; the “Middleton” lava flow yielded a K-Ar age of 1.5 $\pm$ 0.1 Ma (Best and others, 1980) and an $^{40}Ar/^{39}Ar$ age of 1.41 $\pm$ 0.01 Ma (UGS unpublished data); lava flow is typically 20 to 30 feet (6–9 m) thick.

Qbli **Little Creek Peak lava flow** (lower Pleistocene) – Medium-gray, fine- to medium-grained basalt with small plagioclase and olivine phenocrysts; source unknown; yielded an $^{40}Ar/^{39}Ar$ plateau age of 1.44 $\pm$ 0.04 Ma (Willis and Hylland, 2002); typically 20 to 60 feet (6–18 m) thick.

Qbgd **Gunlock-Dameron Valley North lava flow** (lower Pleistocene) – Dark-gray basaltic trachyandesite to trachybasalt with common, small olivine phenocrysts; source unknown, but is at least 2 miles (3 km) east of Veyo, where it may be concealed by the younger Saddle Mountain lava flow; the Dameron Valley North flow yielded $^{40}Ar/^{39}Ar$ plateau ages of 1.62 $\pm$ 0.02 Ma (1.63 $\pm$ 0.03 Ma isochron) (UGS and NMGRL, 2008) and 1.65 $\pm$ 0.02 Ma (1.67 $\pm$ 0.04 Ma preliminary isochron) (UGS unpublished data); Embree (1970) mapped and described vertical and lateral chemical variation of the Gunlock lava flow, the age, geochemistry, and petrology of which suggest it is simply the distal part of the Dameron Valley North flow of Higgins (2002); the Gunlock part of this lava flow yielded a K-Ar age of 1.6 $\pm$ 0.11 Ma (Hintze and others, 1994) and an $^{40}Ar/^{39}Ar$ plateau age of 1.61 $\pm$ 0.07 Ma (UGS unpublished data); typically 20 to 50 feet (6–15 m) thick.

QUATERNARY/TERTIARY

Alluvial deposits

QTaf **Oldest alluvial-fan deposits** (Pleistocene to Pliocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment locally capped by resistant, well-developed calcic soils (hardpan or caliche); clast composition ranges widely and reflects rock types exposed in upstream drainage basins; forms broad, gently sloping, deeply dissected surfaces east and northeast of the Beaver Dam Mountains and west of the Gunlock-Reef Reservoir faults; may be correlative with old alluvial-fan deposits (Qafo) mapped west of the Beaver Dam Mountains, but differentiated here due to uncertain long-term incision rates on these different structural blocks; deposited principally as debris flows and debris floods; variable thickness is as much as about 150 feet (45 m).

QTac **Oldest alluvial and colluvial deposits** (Pleistocene to Pliocene) – Poorly to moderately sorted, generally poorly stratified, clay- to boulder-size, locally derived sediment deposited in the Pakoon Flat area by fluvial, debris-flow, and slope-wash processes; may locally exceed 100 feet (30 m) thick.

Mass-movement deposits

QTms(Ki), QTms(Tcl), QTms(Tipv), QTms

Oldest landslide deposits (Pleistocene to Pliocene) – Includes large blocks of the Iron Springs Formation (QTms[Ki]), lower Claron Formation (QTms[Tcl]), and Pine Valley laccolith (QTms[Tipv]) on the southeast flank of the Pine Valley Mountains, here reinterpreted from the original mapping of Hurlow and Biek (2003). QTms mapped in the northwest part of the White Hills is unsorted, clay- to boulder-size sediment that caps ridges in excess of 400 feet (120 m) above adjacent drainages; it includes large blocks of the Shinarump Conglomerate Member of the Chinle Formation that are more than 30 feet (9 m) long; these QTms deposits are about 20 to 80 feet (6–24 m) thick.

TERTIARY

Tbgw **Granite Wash lava flow** (Pliocene) – Dark-gray, fine-grained basalt to trachybasalt with small olivine phenocrysts; yielded a preferred $^{40}Ar/^{39}Ar$ isochron age of 1.97 ± 0.02 Ma (2.03 ± 0.20 Ma integrated age) (UGS and NMGRL, 2008) and a somewhat younger K-Ar age of 1.5 ± 0.2 Ma (Best and others, 1980); generally about 20 to 100 feet (6–30 m) thick.

Tbah **Aqueduct Hill lava flow** (Pliocene) – Dark-gray basaltic andesite to trachyandesite with small olivine phenocrysts and distinctive aggregates of plagioclase; yielded an $^{40}Ar/^{39}Ar$ integrated age of 1.98 ± 0.04 Ma (1.97 ± 0.05 Ma isochron) (UGS and NMGRL, 2008) and an $^{40}Ar/^{39}Ar$ plateau age of 2.05 + 0.01 Ma (2.02 ± 0.01 Ma isochron) (UGS unpublished data); generally about 20 to 50 feet (6–15 m) thick.

Tbt, Tbtc

Twin Peaks lava flow and cinder cones (Pliocene) – Dark-gray to dark-brownish-gray basaltic trachyandesite (Tbt) with large plagioclase and quartz, and small olivine and clinopyroxene phenocrysts; erupted from vents at extensively eroded cinder cones (Tbtc) at Twin Peaks, about 8 miles (13 km) north of St. George; geochemistry suggests that the West Black Ridge lava flow of Higgins and Willis (1995) and the older remnants (Tbo) of Willis and Higgins (1995) erupted from the Twin Peaks vents; yielded an $^{40}Ar/^{39}Ar$ plateau age of 2.43 ± 0.02 Ma (UGS unpublished data); the West Black Ridge part of the flow yielded K-Ar ages of 2.3 ± 0.1 Ma (Best and others, 1980) and 2.24 ± 0.11 Ma (Hamblin and others, 1981) and an $^{40}Ar/^{39}Ar$ age of 2.34 ± 0.02 Ma (UGS unpublished data), whereas the T-Bone Hill remnant (Tbo of Willis and Higgins, 1995) yielded an $^{40}Ar/^{39}Ar$ plateau age of 2.37 ± 0.02 Ma (UGS unpublished data); lava flow is generally about 20 to 80 feet (6–24 m) thick.

Tb **Basaltic lava flows, undivided** (Pliocene to Miocene) – Resistant, dark-gray and black, aphanitic to crystal-poor lava flows and local cinder cones of mostly olivine basalt that erupted in the north Bull Valley Mountains and Pine Valley area from about 17 Ma to about 2 Ma and that, along with Quaternary basaltic lava flows, are synchronous with basin-range extension (Christiansen and Lipman, 1972; Rowley and Dixon, 2001); Tertiary basalts (including trachybasalt, basaltic andesite, trachyandesite, and andesite) capping mesas and peaks north of the towns of Central and Pine Valley and extending as far as the north edge of the map area, have a wide range of ages, as suggested by the following $^{40}Ar/^{39}Ar$ ages: (1) 2.22 ± 0.05 Ma for a cap rock about 1.5 miles (2.5 km) north of Central (UGS and NMGRL, 2008); (2) about 5.7 to 5.89 ± 0.11 Ma for the basalt of Gum Hill east-southeast of Enterprise and just north of the map area (Cornell and others, 2001; UGS and NMGRL, 2007b); and (3) 17.39 ± 0.12 Ma for the basalt of Harrison Peak at the north edge of the map area (UGS and NMGRL, 2007a); includes basalts along Moody Wash about 6 miles (10 km) west of Central that both overlie and underlie Ox Valley Tuff (To) and that have K-Ar ages by Best and others (1980) of, respectively, 7.9 and 12.3 Ma; maximum thickness of lava flows about 300 feet (100 m).

Ts **Basin-fill sedimentary rocks** (Pliocene to Miocene) – Poorly to moderately consolidated, light-brown, gray, and light-reddish-brown, tuffaceous, alluvial sandstone and subordinate mudstone, siltstone, and conglomerate deposited in closed basins of various ages and origins; most basins are related to the basin-range episode of extension in the area; previously called the Muddy Creek Formation in some areas (Cook, 1957, 1960b; Hintze and others, 1994), but we prefer this generic label given the uncertain correlation to the principal Muddy Creek basin of southeast Nevada; includes the volcaniclastic rocks of Enterprise Reservoir (Blank, 1993) in the northwest corner of the map area, Miocene sedimentary rocks in the Dodge Spring quadrangle (Anderson and Hintze, 1993), and sedimentary breccia of Miocene volcanic units near the Gunlock fault in the Gunlock quadrangle (Hintze and others, 1994); maximum thickness at least 1500 feet (450 m).

Trdy **Young rhyolite and dacite lava flows** (Pliocene to Miocene) – Small centers of resistant, crystal-poor to crystal-rich, high-silica rhyolite and dacite volcanic domes and lava flows with isotopic ages of about 8.5 to 2 Ma; includes the Eight-Mile Dacite (Cook, 1957; Hacker, 1998) just northeast of the town of Central, which has an $^{40}Ar/^{39}Ar$ plateau age of 2.03 ± 0.76 Ma (2.09 ± 0.22 Ma isochron) (UGS and NMGRL, 2007a), the rhyolite of Shinbone Creek (Blank, 1959, 1993) at the north edge of the map area and northwest of Central (K-Ar ages of 5.0 ± 0.2 and 4.7 ± 0.02 Ma; McKee and others, 1997), and the rhyolite of Pilot Peak (Blank, 1959) at the north edge of the map area and northwest of Central ($^{40}Ar/^{39}Ar$ age of 6.05 ± 0.05 Ma; UGS and NMGRL, 2008); maximum thickness about 500 feet (150 m).

Tag **Old boulder gravel deposits of the Kolob Plateau** (Miocene?) – Poorly sorted, clay- to large-boulder-size sediment characterized by enormous quartz monzonite boulders; clasts also include large boulders of Cretaceous fossiliferous sandstone, cobbles and small boulders derived from the Carmel Formation, recycled rounded pebbles and small cobbles of Precambrian and Cambrian quartzite, and uncommon cobbles and boulders of Claron limestone; most clasts are subangular to subrounded, but the quartz monzonite clasts are well rounded (likely due to exfoliation); quartz monzonite boulders as much as 24 feet (7.3 m) long, 22 feet (6.7 m) wide, and at least 8 feet (2.4 m) high are present near Kolob Reservoir, and subspherical quartz monzonite boulders 10 to 15 feet (3–5 m) long are common; most quartz monzonite clasts, however, are 1.5 to 3 feet (0.5–1 m) in diameter; forms a deeply eroded surface that drapes over pre-existing topography; probably deposited by debris flows or possibly a gravity slide originating in the ancestral Pine Valley Mountains; this hypothesis, however, requires a complete eastward-to-westward reversal of drainage across the Hurricane fault; Averitt (1962, 1964) first described similar deposits farther north, east and southeast of Cedar City, and Anderson and Mehnert (1979) interpreted those exposures as debris-flow deposits shed off the ancestral Pine Valley Mountains; Hacker (1998) and Hacker and others (2002) provided evidence that the Pine Valley Mountains stood very high and shed gravity slides, volcanic rocks, and debris flows about 20.5 million years ago, long before initiation of the Hurricane fault; blocks of the Pine Valley laccolith caught in the Hurricane fault zone (Biek, 2007a) show that the laccolith once likely reached east of the Hurricane fault, thus providing a somewhat closer source for the large Pine Valley boulders; thickness uncertain, but probably less than 30 feet (9 m) thick.

Two basalt boulders clearly incorporated into the deposits at Kolob Reservoir have a chemical signature similar to the Horse Ranch Mountain lava flow, and one boulder yielded an $^{40}Ar/^{39}Ar$ disturbed age spectrum with a maximum plateau age of 0.97 $\pm$ 0.18 Ma (1.9 $\pm$ 4.4 Ma low-confidence isochron) (Biek, 2007b), nevertheless analytically indistinguishable from the 1.03 $\pm$ 0.06 Ma Horse Ranch Mountain lava flow (Biek, 2007a); how the basalt boulders came to be incorporated into these assumed Miocene-age deposits is not known, although the boulders may indicate some degree of reworking of these deposits.

Tmc, Tmc?, Tmcb, Tmc(Mr)

Muddy Creek Formation (upper Miocene) – Mostly fine-grained but locally coarse-grained, grayish-orange, pinkish-orange, and medium-reddish-brown calcareous sandstone, siltstone, mudstone, and moderately sorted, pebble- to boulder-conglomerate; includes large gravity-slide blocks Tmc(Mr) west of the Beaver Dam Mountains (as described below and in accompanying text); commonly tuffaceous, with minor light-gray to white airfall and water-lain tuff beds less than 3 feet (1 m) thick; deposited as closed-basin-fill sediment in the Virgin River depression and other connected basins to the south and west; when deposited, the formation graded from coarse conglomeratic deposits near basin margins to finer grained, laminated to thick-bedded central-basin strata, but later faulting disrupted this simple pattern; fine-grained facies commonly have thin gypsum stringers and veinlets; poorly to well cemented, generally forming slopes except where protected by younger, resistant calcic soils (caliche); here restricted to the Virgin River depression (Mesquite basin) due to uncertain correlation with other basins (Ts) in the northwest part of the map area; query (Tmc?) indicates uncertain identification west of concealed trace of Red Hollow fault; Hintze and others (1994) reported an 8.8 $\pm$ 0.3 Ma K-Ar age on basalt (Tmcb) at or near the base of the formation along Beaver Dam Wash in the Motoqua quadrangle; Bohannon (1984) noted that the Muddy Creek Formation is overlain by a 5.9 million-year-old basalt and overlies a 10.6 million-year-old sandstone near Lake Mead; ages for the entire formation range between about 11 and 5 million years old (Reynolds and Lindsay, 1999); represents one of the youngest basin-fill sedimentary units deposited prior to establishment of the through-going Colorado River system (Bohannon, 1984; Bohannon and others, 1993; Pederson, 1999, 2008; Langenheim and others, 2001); through study of sand grain provenance, Farrell and Pederson (2001) and Pederson (2008) showed that the siliciclastic component of the Muddy Creek has a predominant volcanic source in the Caliente and Kane Creek Wash caldera complexes west of the map area and a secondary source from sedimentary rocks surrounding the Virgin River basin; the map unit postdates most of the basin-range faulting that formed the basin, although significant faulting such as along the Piedmont fault zone has continued during and after Muddy Creek deposition; where exposed in this quadrangle, unconformably overlies a variety of older, mostly Miocene and Oligocene volcanic units; locally overlain by a 3- to 5-foot-thick (1–2 m), well-developed calcic soil (Qaec, stage V carbonate development of Birkeland and others, 1991); Metcalf (1982) reported a general thickness of at least 3000 feet (900 m) for the Muddy Creek Formation in southern Nevada; Bohannon and others (1993) reported that it overlies older basin-fill rocks of the Horse Spring Formation and that it fills the upper 3000 to 6500 feet (1000–2000 m) of the Virgin River depression (Mesquite basin), which is one of the deepest (more than 21,000 feet [6400 m]) basins in the Basin and Range Province, on the basis of seismic and gravity data (Bohannon and others, 1993; Carpenter and Carpenter, 1994; Langenheim and others, 2001; Dixon and Katzer, 2002).

Large blocks of tectonically thinned, brecciated Redwall Limestone, mapped as Tmc(Mr), are incorporated in both coarse- and fine-grained facies of the Muddy Creek Formation at the west edge of the Beaver Dam Mountains; the blocks are commonly 1000 feet (300 m) or more long and exposed parts are 10 to 200 feet (3–60 m) thick; the blocks are preserved in structurally high areas in the footwall of the Red Hollow fault and in a relay ramp between the Red Hollow and Piedmont faults; enclosing Muddy Creek clastic strata contain clasts of Mesozoic red beds but lack Proterozoic clasts, thus showing that the blocks were emplaced prior to unroofing of the Beaver Dam Mountains; the gravity-slide blocks in the embayment west of Welcome Spring appear to be distributed through several hundred feet of Muddy Creek strata, suggesting multiple episodes of emplacement in late Miocene time.

Trdm **Middle rhyolite and dacite lava flows** (Miocene) – Mostly resistant, generally light-gray and tan, crystal-poor, high-silica rhyolite and dacite volcanic domes and lava flows, locally with ash-flow tuff, tuff breccia, and airfall tuff that erupted from several centers in the west part of the map area; includes the rhyolite of Little Pine Creek (12.0 ± 0.4 Ma; Blank, 1959; McKee and others, 1997) and the rhyolite of Cow Hollow (15.3 ± 0.4 to 14.3 ± 0.4 Ma; Blank, 1959; McKee and others, 1997) at the north edge of the map northeast of Ox Valley; maximum thickness of each unit 400 to 1400 feet (120 to 400 m).

Tim **Mineral Mountain intrusion** (middle Miocene) – Resistant, gray and pink, high-silica granite porphyry laccolith made up of mostly fine-grained orthoclase but with distinctive abundant large "eyes" of beta quartz and with minor ferromagnesian minerals; located in the southwest Bull Valley Mountains about 4 miles (6 km) northwest of the ghost town of Goldstrike (Cook, 1960b; Bullock, 1970; Eliopulos, 1974; Morris, 1980; Adair, 1986); considered to be the southern intrusion of the Iron Axis, a northeast-trending belt of intrusions partly structurally controlled by thrust faults and characterized by iron occurrences (Mackin, 1960; Tobey, 1976; Blank and others, 1992; Hacker, 1998; Hacker and others, 2002), including the large commercial deposits of the Iron Springs mining district north of the map area; however, this stock is compositionally much more silicic and much younger than the intrusions in other parts of the Iron Axis; it, and perhaps small hypabyssal andesitic intrusions (Adair, 1986), may have been the heat source for hydrothermal (ground-water) solutions (Willden and Adair, 1986; Limbach and Pansze, 1987) that led to gold deposits at and northwest of Goldstrike (Willden and Adair, 1986; Willden, 2006; Rowley and others, 2007); the map unit is the source of eruptions of the Ox Valley Tuff (To), either (1) as an intracaldera (resurgent) intrusion that was emplaced into a caldera that has been removed by erosion, or (2) by rapid vesiculation and eruption of the magma following landsliding of oversteepened laccolith flanks during rapid emplacement, as with most other intrusions of the Iron Axis, which are also laccoliths (Rowley and others, 2007); $^{40}Ar/^{39}Ar$ integrated age from a disturbed age spectrum is about 12.1 ± 1.9 Ma (UGS and NMGRL, 2007a).

To **Ox Valley Tuff** (middle Miocene) – Moderately resistant, gray and red, poorly to densely welded, crystal-poor (including distinctive "eyes" of beta quartz), high-silica rhyolite ash-flow tuff exposed as outflow tuff in the Bull Valley Mountains and Clover Mountains; derived from the Mineral Mountain intrusion (Tim); one area of outflow tuff about 7 miles (11 km) south-southwest of Mineral Mountain was suggested to be a caldera source of the Ox Valley Tuff by Anderson and Hintze (1993) and Hintze and others (1994) because the thickness of the Ox Valley there is as much as 4000 feet (1200 m) and the rock is commonly densely welded, but Rowley and others (2007) interpreted it to represent deposition of near-source outflow tuff upon thin precursor tuff that in turn rests on andesite lava flows and mudflows on the southwest flank of a mountain formed by rapid uplift resulting from its inflation by emplacement of the laccolithic Mineral Mountain intrusion, whose oversteepened flanks failed by landsliding, exposing the magma, which erupted the Ox Valley Tuff; the age of the Ox Valley Tuff was formerly unclear and considered to be 12.6 to 12.3 Ma (Rowley and others, 1995), but the following new $^{40}Ar/^{39}Ar$ ages suggest that the age is 14.0 to 13.5 Ma (Snee and Rowley, 2000): (1) an age of 13.46 ± 0.05 Ma from a sample collected from the lowest of four cooling units exposed in the type area, at Ox Valley, 8 miles (13 km) northwest of Central (Rowley and others, 2006); (2) an age of 14.10 ± 0.03 Ma from a sample collected by R.E. Anderson just west of Beaver Dam State Park, Nevada, just northwest of the map area (Rowley and others, 2006); (3) an age of 12.19 ± 0.08 Ma from a rhyolite flow collected by R.E. Anderson that rests on Ox Valley Tuff at Docs Pass just west of the map area (Rowley and others, 2006); and (4) an age of 13.93 ± 0.08 Ma from a sample of Ox Valley Tuff collected about 3 miles (5 km) southwest of Enterprise, just north of the map area (UGS and NMGRL, 2007b); this reinterpretation of the age of the Ox Valley Tuff suggests, furthermore, that the Ox Valley Tuff may be correlative with the tuff of Etna, widely exposed as an outflow ash-flow sheet in the Caliente caldera complex, including south of Caliente, Nevada, that has similar composition and mineralogy as the Ox Valley Tuff, and with an age interpreted to be 14.0 Ma based on ages of overlying and underlying rocks (Rowley and others, 1995); maximum outflow thickness about 4000 feet (1200 m).

Tkw **Kane Wash Tuff** (middle Miocene) – Resistant, tan and reddish-gray, densely welded, crystal-poor, high-silica rhyolitic and peralkaline ash-flow tuff, locally containing a brown and black basal vitrophyre; mapped by Blank (1959) north of Gunlock as the "lower Moody tuff" of the "Cedar Spring Member" of his "Cove Mountain Formation;" is the eastern finger edge of an outflow sheet of the Kane Wash Tuff derived from the Kane Springs Wash caldera complex of the Delamar and Meadow Valley Mountains of Nevada (Harding and others, 1995; Scott and others, 1995a, 1995b); $^{40}Ar/^{39}Ar$ age of a sample collected from just west of lower Moody Wash northwest of Veyo is 14.45 ± 0.06 Ma (UGS and NMGRL, 2008), suggesting that the unit is correlative with the upper and more voluminous (Gregerson Basin Member) of two members of the Kane Wash Tuff, which in turn has $^{40}Ar/^{39}Ar$ ages of 14.39 Ma from its upper cooling unit and 14.55 Ma from its lower cooling unit (Rowley and others, 1995; Scott and others, 1995a, 1995b); maximum thickness about 30 feet (10 m).

Ta **Andesitic lava flows, flow breccia, and mudflow breccia** (lower Miocene) – Soft to moderately resistant, red, brown, and green, mostly crystal-poor andesite to dacite lava flows, flow breccia, and mudflow breccia erupted from widely scattered stratovolcanoes or other local vents over a poorly known time span of 23(?) to 18(?) Ma; occurs at several stratigraphic positions in this age range, including (1) aphanitic lava flows and mudflow breccia that overlie Racer Canyon Tuff (Tr) in the northwest part of the map area; (2) flows and flow breccia of the andesite of Maple Ridge (Blank, 1959, 1993), which underlies Racer Canyon Tuff (Tr) in the northwest part of the map area and contains abundant large phenocrysts of plagioclase, pyroxene, and biotite; (3) thin lava flows that locally underlie gravity slides (Tgb) that were shed during rapid emplacement of the intrusions of the Iron Axis, not including similar andesitic flows that erupted during emplacement of these intrusions but are mapped with those volcanic rocks (as Tcv and Tpa; see Hacker, 1998); and (4) andesitic flows and volcanic mudflow breccia at several stratigraphic positions within the Quichapa Group (Tq), both between the Harmony Hills Tuff (Tqh) and the Bauers Tuff Member of the Condor Canyon Formation (Tqc) (the andesite flows are called the andesite of Little Creek by Blank, 1993) and between the Bauers Member and the Leach Canyon Formation (Tql), that thickens southward from the north edge of the map area; the maximum thickness of individual sequences is about 600 feet (200 m), but west of the Greek Peak area, hydrothermally altered, poorly exposed andesite masses are probably well over 1000 feet (300 m) thick.

Tr **Racer Canyon Tuff** (lower Miocene) – Resistant, tan, gray, and pink, poorly to moderately welded, low-silica rhyolite ash-flow tuff; where exposed in the map area, is outflow tuff derived from the Telegraph Draw caldera of the Caliente caldera complex (Rowley and others, 2008), located north of the west part of the map area; exact age of the Racer Canyon Tuff is unclear (Rowley and others, 1995), but our best estimate is that it is about 18.7 Ma based on two ages for sample 89-314e (Rowley and others, 2006); in the Dodge Spring area, west of Motoqua, a unit correlated with Racer Canyon Tuff was mapped by Anderson and Hintze (1993) as Hiko Tuff, which is almost identical to Racer Canyon but slightly younger and clearly was derived from the west end of the Caliente caldera complex; about 12 outflow cooling units well exposed south of Upper Enterprise Reservoir (Rowley and others, 1995), along the north edge of the map area, collectively total at least 1500 feet (450 m), but the unit thins abruptly southward.

Tpr **Sedimentary rocks of Page Ranch** (lower Miocene) – Mostly soft, tan conglomerate, sandstone, and mudflow breccia derived from erosion of rapidly intruded quartz monzonite plutons of the Iron Axis; unit mapped locally but may be correlative with lower parts of basin-fill sedimentary rocks (Ts), as just north of the northeast part of the map area; maximum thickness about 200 feet (60 m).

Tpv **Pine Valley Latite** (lower Miocene) – Resistant, gray, pink, and black, flow-foliated, crystal-rich, dacitic and trachydacitic lava flows erupted primarily from vent areas at Rencher Peak and Timber Mountain and derived from eruption of the Pine Valley laccolith (Tipv) (Cook, 1957; Hacker, 1998); Tpv has three $^{40}Ar/^{39}Ar$ ages of 20.44 to 20.42 Ma on two samples (Rowley and others, 2006), similar to the reported preferred age of 20.5 Ma for the Pine Valley laccolith (Hacker and others, 1996; Rowley and others, 2006); maximum thickness in the mapped area about 1100 feet (335 m).

Tgb **Gravity-slide breccia** (lower Miocene) – Moderately resistant, mostly pink and gray, tectonic breccia resulting from gravity slides (huge landslides) of various ages, made up of sedimentary (Tc) and volcanic (Tpv, Tcv, Tre, Tpa, Tqh, Ta, Tqc, Tql, Ti, Tw) rocks shed off the roofs of rapidly rising quartz monzonite porphyry laccoliths and stocks (Tipv, Tib, Tih, Tibm, and several others outside the map area) of the Iron Axis (Cook, 1957; Blank, 1959, 1993; Mackin, 1960; Blank and Mackin, 1967; Blank and others, 1992; Hacker and others, 1996, 2002; Hacker, 1998); maximum thickness in the mapped area about 250 feet (75 m).

Tcv **Volcanic rocks of Comanche Canyon** (lower Miocene) – Moderately resistant, red and pink, crystal-rich dacitic lava flows and poorly to moderately welded ash-flow tuff derived from eruption of the Stoddard Mountain intrusion, located just north of the map area, and resting on gravity slides and fanglomerate also derived from the intrusion; the tuff issued from a small caldera about 1 mile (1.6 km) in diameter, along the east edge of the intrusion and north of the map area (Rowley and others, 2006); the tuff contains distinctive clasts of red Iron Springs Formation sandstone; unit has $^{40}Ar/^{39}Ar$ ages of 22.72 $\pm$ 0.07 and 21.78 $\pm$ 0.08 Ma on one sample (Rowley and others, 2006), the younger of which is consistent with ages of other Iron Axis plutons and that of the Stoddard Mountain intrusion; maximum thickness about 160 feet (50 m).

Tre **Rencher Formation** (lower Miocene) – Moderately resistant, white, tan, red, and purple, poorly to moderately welded, crystal-rich, dacite ash-flow tuff and tuff breccia; derived from eruption of the Bull Valley intrusion (Tib) (Blank, 1959; Hacker, 1998; Hacker and others, 2002); Tre locally rests on gravity-slide masses that partially unroofed the Bull Valley intrusion; Tre has K-Ar ages of 21.8 and 21.5 Ma (McKee and others, 1997) and $^{40}Ar/^{39}Ar$ ages of 21.65 $\pm$ 0.09 and 22.01 $\pm$ 0.08 Ma (UGS and NMGRL, 2008, and UGS and NMGRL, 2007b, respectively) from two samples collected 10 and 6 miles (16 and 10 km), respectively, west of Central, and 21.83 ± 0.17 and 21.46 ± 0.40 Ma from two samples collected from near the town of Pinto north of the map area (Cornell and others, 2001); thickness about 1600 feet (500 m).

Tpa **Rocks of Paradise** (lower Miocene) – Resistant to moderately resistant, gray, pink, and red, crystal-rich, dacitic lava flows and poorly to moderately welded ash-flow tuff derived from eruption of the Pinto Peak intrusion, located just north of the map area, and tilted westward by emplacement of the Stoddard Mountain intrusion, also just north of the map area (Hacker, 1998; Hacker and others, 2002); $^{40}Ar/^{39}Ar$ ages are 21.97 ± 0.09, 21.62 ± 0.08, and 21.75 ± 0.30 Ma on two samples (Rowley and others, 2006), in good agreement with the age of the Pinto Peak intrusion (21.96 ± 0.11 Ma; UGS and NMGRL, 2007a); maximum thickness about 600 feet (180 m).

Intrusions of quartz monzonite porphyry

Mostly resistant, gray and pink, crystal-rich, shallow (most emplaced within about 1.2 miles [2 km] of the surface), quartz monzonite laccoliths and concordant stocks that rose at about 22 to 20 Ma above the roof of an inferred large batholith (Blank and Mackin, 1967; Cook and Hardman, 1967; Rowley, 1998; Rowley and others, 1998) of the same material. The quartz monzonite plutons define the central part of a northeast-trending belt known as the Iron Axis, extending from the youngest Mineral Mountain intrusion (Tim, 13.5 to 14 Ma) in the southwest Bull Valley Mountains to at least the Iron Peak laccolith (20.2 ± 0.05 Ma [second youngest in the Iron Axis], Fleck and others, 1975) in the Markagunt Plateau east of the town of Paragonah, northeast of the map area. Most of the central quartz monzonite plutons appear to be partly controlled by northeast-striking, southeast-verging Sevier-age (Late Cretaceous to early Tertiary) thrust faults that emplaced the Temple Cap Formation over the Iron Springs Formation (Mackin and others, 1976; Van Kooten, 1988); the thrust faults provided access for the magma to move upward through the thick Navajo Sandstone. Some of the plutons of the Iron Axis produced hematite ore deposits from replacement of the Homestake Limestone Member of the Carmel Formation and from magnetite veins; most of these deposits surround three plutons, which make up the Iron Springs mining district just north of the map area, the largest iron-producing district in the western U.S. (Mackin, 1947, 1954, 1960, 1968; Blank and Mackin, 1967; Bullock, 1970; Mackin and others, 1976; Mackin and Rowley, 1976; Rowley and Barker, 1978; Barker, 1995; Rowley and others, 2006); the much smaller Bull Valley mining district, adjacent to the Bull Valley intrusion (Tib), shipped some ore (Wells, 1938; Bullock, 1970; Tobey, 1976). The quartz monzonite plutons are mapped individually because each differed slightly in age and evolved differently, although most are characterized by partial unroofing by gravity sliding, immediately followed by volcanic eruptions (Mackin, 1960; Blank and Mackin, 1967; Hacker and others, 1996, 2002, 2007; Hacker, 1998); relative ages of most plutons can be determined by mapping deformation of eruptive or gravity-slide products by younger plutons; most plutons in the central Iron Axis have several lithologic phases (thin outer peripheral-shell phase, selvage-joint phase, and interior phase) whose mapping elucidated that the iron solutions were derived from deuteric breakdown of ferromagnesian minerals in the selvage-joint phase (Mackin, 1947, 1954, 1960, 1968; Mackin and Ingerson, 1960; Mackin and others, 1976; Mackin and Rowley, 1976; Rowley and Barker, 1978; Barker, 1995).

Tipv **Pine Valley laccolith** (lower Miocene) – Perhaps the largest exposed laccolith in the world (Cook, 1957, 1960b); 3000 feet (900 m) thick and exposed over an area of 90 square miles (240 km^2), with well-exposed concordant basal and flank intrusive contacts in the upper part of the Claron Formation (Tc), a much higher structural level than that of most of the other quartz monzonite intrusions with which it is chemically similar; consists of locally flow layered, medium-gray quartz monzonite porphyry with medium- to coarse-grained phenocrysts of plagioclase, pyroxene, biotite, and sanidine; groundmass is fine-grained to microscopic plagioclase, quartz, and pyroxene; the northern flanks of the Pine Valley Mountains are covered by gravity slides, the largest having an area of at least 60 square miles (150 km^2) and more than 1800 feet (550 m) thick, made up of brecciated roof rocks, indicating that the laccolith was emplaced rapidly, then shed its roof and flanks by gravity sliding along relatively weak shale beds in Tc; removing its roof led to eruption of thick lava flows of the Pine Valley Latite (Tpv); because it was emplaced at such a shallow level (estimated at less than 600 feet [200 m]) below the former land surface, the rocks of both Tipv and Tpv are very similar in outcrop and may be distinguished from each other only by careful geologic mapping, and therefore early workers (for example, Mattison, 1972) interpreted all the rocks as volcanic; we instead follow the interpretations of Cook (1957, 1960b), Hacker (1998), and Hacker and others (2002); K-Ar age on biotite is 20.9 ± 0.6 Ma (McKee and others, 1997), but we prefer $^{40}Ar/^{39}Ar$ ages of 20.47 ± 0.04 and 20.63 ± 0.12 Ma from a sample from the base of the laccolith, and 20.32 ± 0.08 and 20.46 ± 0.05 Ma from a sample collected 500 feet above the base (Rowley and others, 2006), in excellent agreement with the $^{40}Ar/^{39}Ar$ ages of Tpv.

Tid **The Dairy intrusion** (lower Miocene) – A small, apparent laccolith located about 3 miles (5 km) southeast of the small community of Pinto (Cook, 1957); interpreted by Hacker (1998) to be connected at depth with the much larger Stoddard Mountain intrusion, located just north of the map area; the Stoddard Mountain intrusion was emplaced into the Iron Springs Formation at a lower structural level than the Pine Valley laccolith, but at a higher structural level than most of the other quartz monzonite plutons; the Stoddard Mountain intrusion produced gravity slides, then erupted to form the volcanic rocks of Comanche Canyon (Tcv); it has an $^{40}Ar/^{39}Ar$ age of 21.86 ± 0.09 Ma (Rowley and others, 2006), similar to the $^{40}Ar/^{39}Ar$ age of Tcv.

Tib **Bull Valley intrusion** (lower Miocene) – Exposed 6 miles (10 km) west-northwest of Central, the intrusion forms the southwest end of an intrusive arch that appears to connect with the Hardscrabble Hollow intrusion (Tih) farther northeast and with the Big Mountain intrusion (Tibm) at the northeast end (Blank, 1959, 1993; Blank and others, 1992; Hacker, 1998; Hacker and others, 2002); formed replacement hematite ore bodies within Temple Cap and Carmel strata, into which it intruded (Wells, 1938; Blank, 1959; Tobey, 1976); during rapid intrusion to form a mountain, the intrusive arch shed large gravity slides, followed by eruption of the Rencher Formation (Tre); it or other parts of the arch may have had an earlier episode of eruption, resulting in the significantly larger ash-flow tuff – Harmony Hills Tuff (Tqh) (Blank, 1959; Blank and others, 1992); unit has an $^{40}Ar/^{39}Ar$ age of 21.98 ± 0.10 Ma (UGS and NMGRL, 2007b).

Tih **Hardscrabble Hollow intrusion** (lower Miocene) – Exposed 5 miles (8 km) northwest of Central, the intrusion is a small body that appears to be the central part of the same intrusive arch containing the Bull Valley and Big Mountain intrusions; has an $^{40}Ar/^{39}Ar$ age of 22.02 ± 0.11 Ma (UGS and NMGRL, 2007b).

Tibm **Big Mountain intrusion** (lower Miocene) – Exposed as a mountain along the north edge of the map area, just west of Mountain Meadow; this is the north end of the intrusive arch that includes the Bull Valley and Hardscrabble Hollow intrusions and extends north of the map area (Blank, 1993); has an $^{40}Ar/^{39}Ar$ age of 22.00 ± 0.07 Ma (UGS and NMGRL, 2007b), almost identical to the ages of the other intrusions of the arch.

Quichapa Group

Tq **Quichapa Group, undivided** (lower Miocene and upper Oligocene) – Regional ash-flow sheets that were described in detail by Mackin (1960), Williams (1967), Anderson and Rowley (1975), and Rowley and others (1995); consist of the petrographically and chemically distinctive Harmony Hills Tuff, Condor Canyon Formation, and Leach Canyon Formation, here locally lumped where too thin to show separately.

Tqh **Harmony Hills Tuff** (lower Miocene) – Resistant, gray and tan, crystal-rich, moderately welded, dacitic ash-flow tuff; lithologically resembles the Rencher Formation (Tre) except that the Rencher has no quartz or sanidine; source of Harmony Hills Tuff unknown but isopachs are centered on Bull Valley (Williams, 1967), suggesting that it was derived from the east Bull Valley Mountains, probably from an early, much more voluminous eruptive phase of the Bull Valley/Hardscrabble Hollow/Big Mountain intrusive arch, as suggested by Blank (1959), Williams (1967), and Rowley and others (1995); consistent with this interpretation is the fact that the $^{40}Ar/^{39}Ar$ plateau age of the Harmony Hills is 22.03 ± 0.15 Ma (Cornell and others, 2001), nearly identical to that of Tib, Tih, and Tibm; maximum thickness about 500 feet (150 m).

Tqcl **Bauers Tuff Member of the Condor Canyon Formation and Leach Canyon Formation, undivided.**

Tqci **Bauers Tuff Member of the Condor Canyon Formation, Leach Canyon Formation, and Isom Formation, undivided**.

Tqcw **Condor Canyon Formation, Leach Canyon Formation, Isom Formation, and Wah Wah Springs Formation, undivided**.

Tqc **Condor Canyon Formation** (lower Miocene) – Consists of two members, here not mapped separately; the upper one, the **Bauers Tuff Member**, is a resistant, brown, gray, and purple, crystal-poor, densely welded, dacitic to trachydacitic ash-flow tuff, commonly with a black basal vitrophyre, derived from the northwest part (Clover Creek caldera) of the Caliente caldera complex (Rowley and others, 1995); $^{40}Ar/^{39}Ar$ age of the Bauers Tuff Member is about 22.8 Ma (Best and others, 1989a), which is also the $^{40}Ar/^{39}Ar$ age of its intracaldera intrusion exposed just north of Caliente (Rowley and others, 1994b); the underlying member, the **Swett Tuff Member**, is a resistant, brown, crystal-poor, densely welded, dacitic ash-flow tuff, probably derived from the Caliente caldera complex; $^{40}Ar/^{39}Ar$ age is 23.87 ± 0.04 Ma (Rowley and others, 2006), but this age may be too old because it is slightly older than the $^{40}Ar/^{39}Ar$ age of the underlying Leach Canyon Formation (Tql); maximum thickness of the Bauers is about 300 feet (100 m), whereas the Swett is only exposed in the north part of the map area, where its maximum thickness is about 30 feet (10 m).

Tql **Leach Canyon Formation** (upper Oligocene) – Moderately resistant, tan and gray, crystal-poor, poorly welded, low-silica rhyolite ash-flow tuff containing abundant cognate pumice and red lithic fragments; source is unknown but probably is the Caliente caldera complex because isopachs show that it thickens toward the complex (Williams, 1967); the $^{40}Ar/^{39}Ar$ age of the formation is about 23.8 Ma (Best and others, 1993; Rowley and others, 1995); maximum thickness about 500 feet (150 m).

Tiw **Isom Formation and Wah Wah Springs Formation of the Needles Range Group (Mackin, 1960; Anderson and Rowley, 1975), undivided**.

Ti **Isom Formation and enclosing sedimentary rocks, undivided** (upper Oligocene) – Consists of two members, here not mapped separately – the upper **Hole-in-the-Wall Tuff Member** and the lower **Baldhills Tuff Member,** both resistant, brown and reddish-brown, crystal-poor, densely welded, trachydacitic ash-flow tuffs derived perhaps from the Indian Peak caldera complex (Best and others, 1989a, 1989b) north of the map area; the Baldhills Tuff Member consists of several cooling units, some with black basal vitrophyres; the Hole-in-the-Wall Tuff Member is a single cooling unit, commonly with a black basal vitrophyre, exposed only in the north part of the map area; tuffs of the Isom Formation are interbedded with lacustrine limestone and fluvial tuffaceous sandstone that thicken southwestward into the Bull Valley Mountains (Blank, 1959; Hintze and others, 1994; Hacker, 1998); age of the Isom appears to be about 26 to 27 Ma, on the basis of many $^{40}Ar/^{39}Ar$ and K-Ar ages (Best and others, 1989b; Rowley and others, 1994a). Maximum thicknesses: sedimentary rocks above the tuffs—250 feet (80 m); Hole-in-the-Wall Tuff Member—30 feet (10 m); Baldhills Tuff Member—120 feet (40 m); sedimentary rocks below the tuffs—200 feet (60 m).

Tw **Wah Wah Springs Formation of the Needles Range Group** (lower Oligocene) – Soft to resistant, light-greenish-gray, tan, grayish-pink, and light-purple, crystal-rich, moderately welded, dacite ash-flow tuff; derived from the Indian Peak caldera complex (Best and others, 1989a, 1989b); identified as Wah Wah Springs Formation in the Gunlock-Motoqua area (Hintze and others, 1994), and probably is Wah Wah Springs elsewhere in the map area (the Wah Wah Springs Formation is the upper formation of the Needles Range Group in most areas); preferred age, based on many K-Ar and $^{40}Ar/^{39}Ar$ age determinations, is about 30 Ma (Best and others, 1989a, 1989b); maximum thickness about 50 feet (15 m).

unconformity

Claron Formation (lower Oligocene to upper Paleocene)

Tc **Claron Formation, undivided** – Upper and lower members combined in the Bull Valley Mountains and along the flanks of the Pine Valley Mountains; may locally include pre-Isom lacustrine strata of Hintze and others (1994) in the Bull Valley Mountains (i.e., strata younger than "typical" Claron); Goldstrand (1994) considered the formation to be Eocene and Paleocene over most of southwest Utah, but the age is poorly constrained and as mapped here in the Bull Valley Mountains includes beds in the upper part that are clearly early Oligocene; deposited in fluvial, lacustrine, and alluvial-plain environments, and extensively modified by post-depositional bioturbation and pedogenic processes (Mullett and others, 1988a, 1988b; Mullett, 1989; Taylor, 1993); the entire formation in the St. George quadrangle ranges from about 300 to 1600 feet (90–490 m) thick; the maximum thickness in the northern Bull Valley Mountains is about 1000 feet (300 m), but it thins southward to about 300 to 500 feet (90–150 m) thick on the south margin of the range.

Tcu **Upper member** – White to light-gray, resistant limestone interbedded with mottled pale-gray, pinkish-gray, and maroon mudstone; limestone is medium to thick bedded and includes nonfossiliferous micrite, fine-grained bioclastic calcarenite, pelletal calcarenite, and local laminated algal structures; unconformably overlain by a variety of Miocene and Oligocene volcanic and sedimentary rocks; about 520 feet (160 m) thick in the Pintura area (Hurlow and Biek, 2003).

Tcl **Lower member** – Interbedded mudstone, siltstone, sandstone, conglomerate, and limestone; mudstone is orangish red to reddish brown; sandstone is light-brown, medium- to coarse-grained, cross-bedded to structureless litharenite; sandstone grades into, or is locally incised by thick-bedded pebble conglomerate with limestone, quartzite, and chert clasts; lower to middle part of the lower member contains resistant, thick-bedded micrite that is mottled orange, pink, tan, lavender, and yellow and that contains sparse bivalve and gastropod shell fragments; upper part of the lower member contains medium-gray, thin- to medium-bedded, resistant micrite to calcarenite with locally common bivalve fragments, oncolites, and calcite-filled veins and vugs; in the Pintura area, upper contact corresponds to a pronounced color change from brightly colored reddish-orange mudstone below to gray to pinkish-gray mudstone above; about 1090 feet (330 m) thick in the Pintura area (Hurlow and Biek, 2003).

unconformity

TERTIARY-CRETACEOUS

TKg **Grapevine Wash Formation** (Paleocene to Upper Cretaceous) – Present only in the Gunlock – Square Top Mountain area where it is divisible into three units (Wiley, 1963; Hintze and others, 1994); in descending order, but undivided here (unit descriptions summarized from Hintze [1986] and Hintze and others [1994]): **Upper conglomerate unit** – Poorly sorted, subangular to subrounded pebble to cobble conglomerate, but with some boulders as much as 6 feet (2 m) in diameter; most clasts derived from the Queantoweap Sandstone and Callville Limestone; forms white to pale-yellowish-gray ledges and cliffs; about 330 to 430 feet (100–130 m) thick. **Middle sandstone and conglomerate unit**

– Contains more sandstone and siltstone and forms less-resistant, less-vegetated slopes than enclosing units; about 60% of the clasts are limestone derived from the Callville, Toroweap, and Kaibab Formations, with the remainder derived from the Queantoweap Sandstone; about 1300 feet (400 m) thick. **Lower conglomerate unit** – Resistant, pebble to cobble conglomerate and minor cross-bedded gritstone, sandstone, and siltstone; clast composition similar to that of the middle unit, but includes well-rounded pebbles and cobbles of gray quartzite; forms reddish-brown cliffs; about 360 to 500 feet (110-150 m) thick. The Grapevine Wash Formation represents stream and fan alluvium derived principally from the Square Top Mountain thrust plate (Wiley, 1963; Hintze, 1986; Goldstrand, 1992); unconformably overlain by the Claron Formation or Oligocene and Miocene volcanic and sedimentary rocks generally with an angular discordance of 5° to 10°; as mapped, may locally include Claron Formation; the entire formation is as much as about 2000 feet (600 m) thick.

unconformity

TKcp **Canaan Peak Formation** (Paleocene to Upper Cretaceous) – Conglomerate with minor interbedded sandstone and mudstone; clast-supported conglomerate is reddish brown, contains pebble- to cobble-sized, rounded clasts of brown, purple, gray, reddish-brown, and black, structureless to cross-bedded quartzite, welded dacitic to andesitic ash-flow tuff, chert, and pale-gray to brownish-gray, fine- to medium-grained sandstone; welded tuff clasts contain altered feldspar and quartz phenocrysts in a gray, pale-green, orange-tan, or purple-brown devitrified groundmass; Goldstrand (1992) correlated the tuff clasts with Jurassic volcanic rocks in southeast California; sandstone is reddish-brown, medium- to coarse-grained, structureless to weakly laminated litharenite in 0.5- to 1-foot-thick (0.2–0.3 m) beds; mudstone is present near the base and top of the formation, and is reddish brown and finely laminated; preserved in paleochannels on the Navajo Sandstone (Goldstrand, 1992) along a structural culmination interpreted to be the hinge zone of the Pintura anticline (Hurlow and Biek, 2003); palynomorphs collected from the Canaan Peak Formation on the Table Cliff Plateau in south-central Utah indicate a Maastrichtian (or possibly late Campanian) to early Paleocene age (Goldstrand, 1994); 0 to 120 feet (0–35 m) thick (Hurlow and Biek, 2003).

unconformity

CRETACEOUS AND JURASSIC

KJu **Iron Springs, Carmel, and Temple Cap Formations, undivided** – Steeply dipping, faulted strata of the uppermost Iron Springs Formation, Co-op Creek Limestone Member of the Carmel Formation, and the Sinawava Member of the Temple Cap Formation exposed in the Virgin River canyon just west of Utah Highway 9.

CRETACEOUS

Two different nomenclatural schemes are used to describe Upper Cretaceous strata on either side of the Hurricane fault zone. East of the Hurricane fault, where they have been extensively studied, Upper Cretaceous strata in the St. George 30' x 60' quadrangle are divided into, in ascending order, the Dakota Formation, Tropic Shale, and Straight Cliffs Formation. Offshore-marine deposits of the Tropic Shale pinch out westward on the Upper Kolob Plateau (Biek and Hylland, 2007; Biek, 2007b) and in Cedar Canyon east of Cedar City (Eaton and others, 2001). West of the Hurricane fault, Cook (1960b) mapped what he called the Dakota-Tropic, Straight Cliffs–Wahweap, and Kaiparowits Formations on the east side of the Pine Valley Mountains, but noted that these units became increasingly sandy and indistinguishable westward, where he mapped them as undivided Cretaceous strata. Additional biostratigraphic work in the Pine Valley Mountains may allow nomenclature of the Markagunt Plateau to be carried westward across the Hurricane fault, but for now, these strata remain undivided as the Iron Springs Formation.

Ktm **Tectonic breccia** (Upper Cretaceous) – Scattered angular blocks of red sandstone, light-gray limestone, and conglomerate as much as 2 feet (0.5 m) in diameter in an unconsolidated, unbedded, reddish-brown silty matrix; clasts derived from the Moenkopi Formation, Shinarump Conglomerate Member of the Chinle Formation, and Kayenta Formation; description from Hintze and others (1994), who interpreted it to be a chaotic mix of material caught under the Square Top Mountain thrust as it moved eastward over Triassic and Jurassic strata; 0 to 330 feet (1–100 m) thick.

Ki **Iron Springs Formation** (Upper Cretaceous, Santonian or lower Campanian to Cenomanian) – Interbedded, ledge-forming, calcareous, cross-bedded, fine- to medium-grained sandstone and less-resistant, poorly exposed sandstone, siltstone, and mudstone; contains a few thin sandy coquina beds, minor carbonaceous shale, and uncommon pebbly sandstone; lower part contains numerous light-gray to reddish-brown smectitic mudstone intervals; the formation is variously colored grayish orange, pale yellowish orange, dark yellowish orange, white, pale reddish brown, and greenish gray and is locally stained by iron-manganese oxides; Liesegang banding is locally common in the sandstone beds; sandstone beds range from quartz arenite to litharenite (Fillmore, 1991; Goldstrand, 1992); deposited principally in braided-stream and floodplain environments (Johnson, 1984; Fillmore, 1991), but Eaton and others (1997) noted that brackish-water molluscan faunas collected about 1000 feet (300

m) above the base of the formation in the Pine Valley Mountains probably indicate the maximum transgression of the Cretaceous seaway; map unit tentatively correlated to the Dakota Formation (as used here), Tropic Shale, and Straight Cliffs Formation of the Markagunt Plateau (Eaton, 1999); age from Goldstrand (1994); about 3500 to 4000 feet (1070–1220 m) thick in the Pine Valley Mountains (Cook, 1960b) and about 3000 feet (900 m) thick near Gunlock (Hintze, 1986).

Straight Cliffs Formation

Ksu **Upper unit** (Upper Cretaceous, Santonian[?] to Turonian) – Slope-forming, grayish-orange to yellowish-brown, thin- to thick-bedded, fine-grained subarkosic sandstone and gray mudstone and shale; contains a few thin coal beds, common carbonaceous shale, and several thin oyster coquina beds; forms broad, rounded hills typically mantled with colluvium; believed to be equivalent to the Smoky Hollow Member and possibly John Henry Member of the Straight Cliffs Formation of the Kaiparowits Plateau (see, for example, Eaton and others, 2001); deposited in fluvial, floodplain, and lagoonal environments of a coastal plain (Eaton and others, 2001); incomplete thickness as much as 320 feet (100 m) in the Kolob Reservoir quadrangle (Biek, 2007b), but upper part not preserved.

Kst **Tibbet Canyon Member** (Upper Cretaceous, Turonian) – Grayish-orange to yellowish-brown, generally medium- to thick-bedded, planar-bedded, fine- to medium-grained quartzose sandstone and lesser interbedded, grayish-orange to gray mudstone and siltstone; locally contains pelecypods, gastropods, and thin to thick beds of oyster coquina; typically forms cliffs, but here more commonly weathers to steep, vegetated slopes; upper contact corresponds to a break in slope and is placed at the top of a coquinoid oyster bed that caps the member; deposited in shoreface, lagoonal, estuarine, and floodplain environments of a coastal plain (Laurin and Sageman, 2001; Tibert and others, 2003); thickens eastward across the Kolob Reservoir quadrangle from about 240 to 450 feet (75–135 m) (Biek, 2007b).

Kdt **Dakota Formation and Tropic Shale, undivided** (Upper Cretaceous, Cenomanian) – Interbedded, slope- and ledge-forming sandstone, siltstone, mudstone, claystone, carbonaceous shale, coal, and marl; sandstone is yellowish brown or locally white, thin to thick bedded, fine to medium grained; includes two prominent cliff-forming sandstone beds, each several tens of feet thick, in the upper part of the formation; mudstone and claystone are gray to yellowish brown and commonly smectitic; oyster coquina beds, clams, and gastropods, including large *Craginia* sp., are common, especially in the upper part of the section; uppermost marl beds above the uppermost sandstone cliff contain distinctive gastropods with a beaded edge (*Admetopsis* n. sp., indicative of a latest Cenomanian brackish environment [Eaton and others, 2001; Hoffman, 2005]); Dakota strata are typically poorly exposed and involved in large landslides; includes the overlying Tropic Shale, which is restricted to the east part of the Kolob Reservoir quadrangle where it is silty and sandy and no more than a few feet thick (Biek, 2007b), representing the farthest-west encroachment of offshore-marine environments of the Cretaceous seaway in this part of Utah (see, for example, Eaton and others, 2001); upper contact placed at the top of a slope-forming, coaly and marly mudstone sequence and at the base of the typically cliff-forming sandstone of the Tibbet Canyon Member of the Straight Cliffs Formation; deposited in floodplain, estuarine, lagoonal, and swamp environments (Gustason, 1989; Laurin and Sageman, 2001; Tibert and others, 2003); invertebrate and palynomorph fossil assemblages indicate shallow-marine, brackish, and freshwater deposits of Cenomanian age (Nichols, 1995); about 850 feet (260 m) thick (Biek, 2007b).

unconformity

Kcm **Cedar Mountain Formation, undivided** (Cretaceous, Cenomanian to Albian) – Consists of two units in southwest Utah (Hylland, 2000; Biek and Hylland, 2007) in ascending order: **Conglomerate unit** – Thick-bedded, yellowish-brown, channel-form conglomerate, pebbly sandstone, and pebbly gritstone; clasts are subrounded to rounded, pebble- to small-cobble-size quartzite, chert, and limestone; locally stained reddish brown to dark yellowish brown; best developed on the Upper Kolob Plateau east of Kolob Reservoir, on the east flank of the Pine Valley Mountains, and in the Gunlock-Veyo area, where it is 30 to 100 feet (10–30 m) thick, but nearly everywhere present as a thin, commonly less than 1-foot-thick (0.3 m), pebbly conglomerate deposited on paleotopography developed on top of the various members of the Carmel Formation. **Mudstone unit** – Gray to variegated smectitic mudstone with minor light-gray to yellowish-gray fine-grained sandstone that weathers to poorly exposed slopes; near the base of this predominantly mudstone interval on the Upper Kolob Plateau there is a distinctive, 1- to 4-foot-thick (0.3–1.2 m), locally ledge-forming, pale-olive to greenish-gray, thin- to medium-bedded, fine- to medium-grained sandstone containing subangular, reddish-brown chert granules; upper contact is poorly exposed and corresponds to a color and lithologic change, from comparatively brightly colored smectitic mudstone below to gray and light-yellowish-brown mudstone and fine-grained sandstone above; less than a few tens of feet thick.

Unconformably overlain by the Dakota Formation east of the Hurricane fault (see, for example, Kirkland and others, 1997) and by the Iron Springs Formation west of the fault; deposited in river-channel and floodplain environments on a broad, coastal plain (Tschudy and others, 1984; Kirkland and others, 1997); Biek and Hylland (2007) reported a single-crystal $^{40}Ar/^{39}Ar$ age of 97.9 ± 0.5 Ma (earliest Cenomanian) on sanidine from a volcanic ash in Cedar Mountain mudstone immediately above the conglomerate bed in the Straight Canyon quadrangle east of Zion National Park; Dyman and others (2002) obtained an $^{40}Ar/^{39}Ar$ age of 101.7 ± 0.42 Ma (latest Albian) on sanidine from a stratigraphically similar mudstone near Gunlock, the same 20-foot-thick (6 m), moderate-red bentonitic bed near the Gunlock Reservoir for which Hintze and others (1994) reported an anomalously young fission-track age of 80 ± 10 Ma; pollen analyses indicate an Albian age for the mudstone and underlying conglomerate (Doelling and Davis, 1989; Hylland, 2000).

Correlation of these strata to the Cedar Mountain Formation of central Utah is controversial (see, for example, Titus and others, 2005). The conglomeratic member was previously mapped as the lower part of the Dakota Formation, and the mudstone unit (in the Gunlock area) as an Upper Cretaceous bentonite, but the lithology, age, and stratigraphic position of these beds suggest correlation to the Cedar Mountain Formation of central Utah; specifically, the mudstone unit appears to be time-correlative with the Mussentuchit Member of the Cedar Mountain Formation of central and eastern Utah (Cifelli and others 1997; Kirkland and Madsen, 2007); ongoing detrital zircon studies may help resolve the provenance and correlation of the underlying conglomeratic unit; where conglomerate is thickest, it likely represents deposits of a major Cedar Mountain river channel, whereas thin conglomerate deposits may represent terrace remnants of this river system or possibly younger gravels that were reworked in the early Late Cretaceous; 0 to about 100 feet (0–30 m) thick.

K unconformity (Pipiringos and O'Sullivan, 1978)

JURASSIC

Carmel Formation

Nomenclature follows that of Doelling and Davis (1989); deposited in a shallow inland sea of a back-bulge basin, the first clear record of the effects of the Sevier orogeny in southwestern Utah; forms an eastward-thickening wedge preserved beneath the basal Cretaceous unconformity; age from Imlay (1980).

Jc **Carmel Formation** (Middle Jurassic, Callovian to Bajocian) – Mapped on the nose of the Pintura anticline.

Jcw **Winsor Member** (Middle Jurassic, Callovian to Bathonian) – Light-reddish-brown, fine- to medium-grained sandstone and siltstone; poorly cemented and weathers to densely vegetated slopes; preserved only northeast of Leeds Creek, under the basal Cretaceous unconformity; typically overlain by the conglomerate unit of the Cedar Mountain Formation, but locally overlain by the Dakota Formation where the conglomerate appears to be missing; deposited on a broad, sandy mudflat (Imlay, 1980; Blakey and others, 1983); thins westward due to truncation under the K unconformity; 0 to 320 feet (0–98 m) thick.

Jcp **Paria River Member** (Middle Jurassic, Bathonian) – Laminated to thin-bedded, yellowish-gray to light-olive-gray argillaceous limestone and micritic limestone that locally overlies a thick, white, alabaster gypsum bed in the basal part of the Paria River Member; limestone weathers to small chips and plates and contains rare, small pelecypod fossils; forms steep, ledgy slopes; upper contact is sharp and planar on the Kolob Plateau, but is unconformably overlain by the Cedar Mountain Formation in western exposures; deposited in shallow-marine and coastal-sabkha environments (Imlay, 1980; Blakey and others, 1983); thins westward due to truncation under the K unconformity; 0 to 160 feet (0–50 m) thick.

Jcx **Crystal Creek Member** (Middle Jurassic, Bathonian) – Thin- to medium-bedded, pale- to moderate-reddish-brown gypsiferous siltstone, mudstone, fine- to medium-grained sandstone, and gypsum; typically friable and weakly cemented with gypsum; contains local thin interbeds of yellowish-gray to light-olive-gray mudstone with abundant fine- to coarse-sand-size biotite flakes; moderate-reddish-brown blebs of jasper are common in the lower part of the member; uppermost part on the southeast flank of the Pine Valley Mountains consists of white, grayish-orange-pink, light-brown, and dark-yellowish-orange, fine- to medium-grained sandstone and minor mudstone and siltstone; forms vegetated, poorly exposed slopes; in the Zion National Park area, upper contact is sharp and broadly wavy and corresponds to the base of a thick Paria River gypsum bed or argillaceous limestone, but in western exposures the upper part is truncated under the basal Cretaceous unconformity; Kowallis and others (2001) reported two $^{40}Ar/^{39}Ar$ ages of 167 to 166 million years old for altered volcanic ash beds within the member near Gunlock that were likely derived from a magmatic arc in what is now southern California and western Nevada; deposited in coastal-sabkha and tidal-flat environments (Imlay, 1980; Blakey and others, 1983); thins westward due to truncation under the K unconformity; 0 to 250 feet (0–75 m) thick.

Jcc **Co-op Creek Limestone Member** (Middle Jurassic, Bajocian) – Divisible into two informal unmapped units: **Upper unit** – Thin- to medium-bedded, light-gray, yellowish-gray, or yellowish-brown micritic limestone and, especially in the upper part, oolitic and sandy limestone; forms sparsely vegetated, ledgy slopes and cliffs; locally contains abundant *Isocrinus* sp. columnals, pelecypods, and gastropods. **Lower unit** – Mostly thinly laminated to thin-bedded, light-gray micritic limestone, calcareous shale, platy limestone, and, near the base, lesser gypsum and fine-grained sandstone; forms steep, vegetated slopes; contact with upper unit is gradational and corresponds to a subtle break in slope and vegetation patterns. Contact with Crystal Creek Member is sharp and planar and corresponds to a prominent color change from yellowish- and brownish-gray to reddish-brown, with fossiliferous sandy limestone and micritic limestone giving way to gypsiferous mudstone and siltstone; locally, moderate-reddish-brown blebs of jasper and an oyster coquina are present in uppermost mudstones of the Co-op Creek Limestone; in the west half of the map area, locally unconformably overlain by the conglomerate unit of the Cedar Mountain Formation; Kowallis and others (2001) reported several $^{40}Ar/^{39}Ar$ ages of 168 to 167 million years old for altered volcanic ash beds within the lower unit in southwest Utah that were likely derived from a magmatic arc in what is now southern California and western Nevada; deposited in a shallow-marine environment (Imlay, 1980; Blakey and others, 1983); generally thins to the west, but ranges from about 285 to 580 feet (87–177 m) thick.

J-2 unconformity? (Pipiringos and O'Sullivan, 1978), formed about 169 to 168 million years ago in southwest Utah (Kowallis and others, 2001). New research suggests that the the J-2 unconformity may not represent a significant gap in the rock record as envisioned by Pipiringos and O'Sullivan (1978) (D.A. Sprinkel, Utah Geological Survey, verbal communication, January 12, 2009).

Temple Cap Formation

Jtw **White Throne Member** (Middle Jurassic, Bajocian) – Very thick bedded, yellowish-gray to pale-orange, well-sorted, fine-grained quartz sandstone with high-angle, tabular-planar, and wedge-planar cross-beds in sets as much as 20 feet (6 m) thick, similar to the Navajo Sandstone; forms prominent cliff in the south part of Zion National Park, but pinches out westward under the Upper Kolob Plateau; deposited in a coastal dune field (Blakey, 1994; Peterson, 1994); 0 to 130 feet (0–40 m) thick.

Jts **Sinawava Member** (Middle Jurassic, Bajocian) – Slope-forming, moderate-reddish-brown mudstone, siltstone, and fine-grained silty sandstone; west of the Hurricane Cliffs, contains several white, gray, and pink alabaster gypsum beds as much as 10 feet (3 m) thick and several thin intervals with white and pinkish-gray chert nodules that may be silicified bentonite; western exposures also contain numerous thin, greenish-gray mudstone (altered volcanic ash) beds with common biotite; a ledge-forming, pinkish-gray to light-greenish-gray, calcareous, medium- to coarse-grained, locally pebbly sandstone is present about one-third up the section in the Harrisburg Junction quadrangle (Biek, 2003a); forms narrow, but prominent, deep-reddish-brown, vegetated slope at the top of the Navajo Sandstone in the Zion National Park area; elsewhere forms conspicuous bright-red and gray slopes that weather to soft, gypsiferous soils; upper contact is gradational and interfingering with the White Throne Member in the Zion National Park area, but unconformable with light-gray calcareous shale and micritic limestone of the Co-op Creek Limestone Member to the west; deposited in coastal-sabkha and tidal-flat environments (Blakey, 1994; Peterson, 1994); member mapped as a colored line where too thin to show separately; Kowallis and others (2001) reported several $^{40}Ar/^{39}Ar$ ages of 170 to 169 million years old for altered volcanic ash beds within the formation that were likely derived from a magmatic arc in what is now southern California and western Nevada; thins abruptly over the crest of the Pintura anticline; thickens west and south from less than 10 feet (< 3 m) thick in the Kolob Canyons part of Zion National Park to about 400 feet (120 m) thick near Gunlock.

J-1 unconformity (Pipiringos and O'Sullivan, 1978), formed prior to about 170.5 million years ago in southwest Utah (Kowallis and others, 2001).

Jn **Navajo Sandstone** (Lower Jurassic) – Pale-reddish-orange, reddish-brown, or white, cliff-forming, cross-bedded, quartz sandstone; forms spectacular sheer cliffs, deep canyons, and impressive spires, promontories, and monoliths; consists of well-sorted, well-rounded, fine- to medium-grained, frosted quartz sand; bedding consists of high-angle, large-scale cross-bedding in tabular-planar, wedge-planar, and trough-shaped sets 10 to 45 feet or more (3–14+ m) thick; contains sparse planar interdune deposits and locally common ironstone bands and concretions; locally prominently jointed (see, for example, Rogers and Engelder, 2004; Rogers and others, 2004); lower part forms transition zone characterized by planar-bedded, fine-grained sandstone and silty sandstone with thin siltstone interbeds with wavy bedding, flaser-like laminae, and soft-sediment deformation and bioturbation features, and less common but resistant cross-stratified sandstone; divided into three informal non-stratigraphic units of variable but roughly equal thickness based on color and weathering habit in the south part of Zion National Park (Doelling and others, 2002): (1) "white Navajo," which forms the upper part of the Navajo Sandstone and is pale-gray, yellowish-gray, orangish-gray, and white because of alteration, remobilization, and bleaching of limonitic and hematitic (iron-bearing) cement, probably due to hydrocarbon migration (see, for example, Beitler and others, 2003); (2) "pink Navajo," which forms the middle part of the Navajo Sandstone, is generally less resistant than the "white Navajo" above and "brown Navajo" below, and is pale reddish orange due to more uniformly dispersed hematitic cement; and (3) "brown Navajo," which is streaked medium to dark reddish brown due to iron oxide remobilization caused by ground-water or hydrocarbon migration, and which forms the lower massive cliff of the Navajo Sandstone and is correlative in part with the lower transitional beds of the Navajo; Nielsen and Chan (in preparation) described diagenetic facies and fluid-related alteration of the Navajo Sandstone in southwest Utah; the upper, unconformable contact of the Navajo is sharp and planar and corresponds to a prominent break in slope, with cliff-forming, cross-bedded sandstone below and reddish-brown mudstone of the Sinawava Member of the Temple Cap Formation above; deposited in a vast coastal and inland dune field with prevailing winds principally from the north (Blakey, 1994, Peterson, 1994); correlative with the Nugget Sandstone of northern Utah and Wyoming and the Aztec Sandstone of southern Nevada and adjacent areas (see, for example, Kocurek and Dott, 1983; Riggs and others, 1993); the lower transition interval, which reaches its maximum thickness of about 300 feet (100 m) in the Red Cliffs area northeast of St. George, represents deposition in a sand-dominated sabkha environment (Tuesink, 1989; Sansom, 1992); much of the sand may originally have been transported to areas north and northwest of Utah via a transcontinental river system that tapped Grenvillian-age (about 1.0 to 1.3 billion-year-old) crust involved in Appalachian orogenesis of eastern North America (Dickinson and Gehrels, 2003; Rahl and others, 2003; Reiners and others, 2005); forms the principal aquifer throughout much of the quadrangle (Clyde, 1987; Hurlow, 1998; Heilweil and others, 2000; Heilweil and others, 2002; Rowley and Dixon, 2004; Rowley and others, 2004); about 1800 to 2300 feet (550–700 m) thick.

Kayenta Formation

Marzolf (1994) and Blakey (1994) presented evidence to restrict the Moenave Formation to the Dinosaur Canyon and Whitmore Point Members, with a major regional unconformity at the base of the Springdale Sandstone. Further work supports this evidence, indicating that the Springdale Sandstone is more closely related to the Kayenta Formation and should be made its basal member (see, for example, Lucas and Heckert, 2001; Molina-Garza and others, 2003; Lucas and Tanner, 2006).

Jk **Kayenta Formation (entire upper part above the Springdale Sandstone Member west of Zion National Park; middle part [main body] in and near Zion National Park where Springdale Sandstone and Lamb Point and Tenney Canyon Tongues are mapped separately; and entire formation on cross section)** (Lower Jurassic) – Moderate- to dark-reddish-brown, thin- to thick-bedded siltstone, fine-grained sandstone, and mudstone with planar, low-angle, and ripple cross-stratification; contains several thin, light-olive-gray-weathering, light-gray dolomite beds in the lower third of the formation above the Springdale Sandstone; forms steep ledgy slope grading up to ledgy cliffs at top; upper contact is conformable and gradational and generally corresponds to the top of the highest thin siltstone and mudstone beds, above which are the towering cliffs of Navajo Sandstone; deposited in distal river, playa, and minor lacustrine environments (Tuesink, 1989; Sansom, 1992; Blakey, 1994; Peterson, 1994); entire upper part thickens westward from about 550 to 700 feet (165–210 m) thick in the south part of Zion National Park, to 900 to 1200 feet (275–365 m) thick in the greater St. George area, to about 1500 feet (460 m) thick in the northern Beaver Dam Mountains; middle part below the Lamb Point Tongue is about 290 to 400 feet (88–120 m) thick.

Jkt **Tenney Canyon Tongue of Kayenta Formation** (Lower Jurassic) – Forms the upper part of the Kayenta Formation in the Zion National Park area, above the Lamb Point Tongue; consists of lenticular beds of pale-reddish-brown to moderate-reddish-orange siltstone and fine-grained sandstone, and minor claystone and limestone; forms a steep slope grading up to ledgy cliffs at top; pinches out to the east near the Paunsaugunt fault, about 50 miles (80 km) east of the quadrangle; thickens westward from about 140 to 315 feet (43–95 m).

Jnl **Lamb Point Tongue of Navajo Sandstone** (Lower Jurassic) – Mostly reddish-brown, fine-grained, well-sorted, strongly cross-bedded, quartzose sandstone; contains scattered thin lenses of flat-bedded, pale-reddish-brown siltstone and claystone similar to Kayenta Formation beds; forms a vertical ledge in the upper one-third of the Kayenta Formation in the Zion National Park area; upper contact placed at top of a thick, laterally continuous, ledge-forming sandstone; locally contains a 1-foot-thick (30 cm) bed of limestone near the top; deposited in a coastal dune field and sabkha environment; thins westward and pinches out near Right Fork North Creek; mapped as a colored line where thin; 0 to 60 feet (0–18 m) thick.

Jks **Springdale Sandstone Member of Kayenta Formation** (Lower Jurassic) – Medium- to thick-bedded, fine- to medium-grained sandstone with planar and low-angle cross-stratification, and minor, thin, discontinuous lenses of intraformational conglomerate and thin interbeds of moderate-reddish-brown or greenish-gray mudstone and siltstone; contains locally abundant petrified and carbonized fossil plants; forms conspicuous cliffs and mesas and is host to the ore deposits of the Silver Reef mining district near Leeds (Proctor and Shirts, 1991; Biek and Rohrer, 2006); upper contact is conformable and gradational and corresponds to the first appearance of laterally continuous, thin-bedded, reddish-brown, fine-grained silty sandstone that overlies the Springdale cliff; the top of the Springdale Sandstone is known as the Springdale megatracksite, reflecting the common occurrence of theropod tracks found at this stratigraphic horizon (Lucas and others, 2005; Hamblin and others, 2006); deposited in braided-stream and minor floodplain environments of a northward-flowing river system (Clemmensen and others, 1989; Blakey, 1994; Peterson, 1994; DeCourten, 1998); about 90 to 150 feet (27–45 m) thick.

J-sub Kayenta unconformity (Blakey, 1994; Marzolf, 1994)

JURASSIC AND TRIASSIC

JTRkm **Kayenta and Moenave Formations, undivided** (Lower Jurassic to Upper Triassic) – Mapped in the Kolob Canyons area and just east of Motoqua where it is undivided due to structural complications and map scale.

JTRmc **Moenave and Chinle Formations, undivided** (Lower Jurassic to Upper Triassic) – May include parts of all members of these formations where mapped along the Hurricane fault.

JTRm **Moenave Formation** (Lower Jurassic to Upper Triassic) – Consists of two undivided members: **Whitmore Point Member** (Lower Jurassic): pale-red-purple, greenish-gray, and blackish-red mudstone and claystone, lesser moderate-reddish-brown, fine-grained sandstone and siltstone, and uncommon dark-yellowish-orange micaceous siltstone; contains several 3- to 18-inch-thick (8-120 cm), bioturbated, cherty, dolomitic limestone beds with algal structures, small reddish-brown chert blebs, and locally common fossil fish scales of semionotid fish (Hesse, 1935; Schaeffer and Dunkle, 1950; Milner and Kirkland, 2006); the dolomitic limestones are light greenish gray, light gray, and yellowish gray, and they weather to mottled colors of pale yellowish orange, white, yellowish gray, and pinkish gray, commonly with green iron or copper-carbonate stains; upper part locally includes fine-grained, channel-like sandstone beds, and lower part locally contains brown sandstone beds similar to those of the underlying Dinosaur Canyon Member; weathers to poorly exposed, locally brightly colored slopes; upper contact corresponds to a major regional unconformity (Blakey, 1994; Marzolf, 1994); deposited in floodplain and lacustrine environments (Clemmensen and others, 1989; Blakey, 1994; Peterson, 1994) of the terminal floodplain of the Moenave alluvial system (Tanner and Lucas, 2007); lacustrine deposits are best developed in the St. George-Washington area (Kirkland and Milner, 2006), becoming thinner and less well developed to the northeast and northwest; the St. George Dinosaur Discovery Site at Johnson Farm is near the base of the member and contains exceptionally well-preserved theropod tracks (*Eubrontes* and *Grallator*) and theropod swim tracks (Kirkland and Milner, 2006; Milner and others, 2006), and a variety of invertebrate (Lucas and Milner, 2006), trace (Lucas and others, 2006), and plant (Tidwell and Ash, 2006) fossils; age from Molina-Garza and others (2003) and Lucas and Tanner (2006, 2007); ranges from about 50 to 140 feet (15-40 m) thick. **Dinosaur Canyon Member** (Lower Jurassic to Upper Triassic): generally thin-bedded, moderate-reddish-brown and moderate-reddish-orange, fine-grained sandstone, fine-grained silty sandstone, and lesser siltstone and mudstone with planar, low-angle, and ripple cross-stratification; basal part typically includes a single bed of reddish-brown, fine-grained silty sandstone several tens of feet thick; a thin chert-pebble conglomerate marks the base of the member; forms ledgy slopes; upper contact is conformable and gradational and corresponds to the base of a light-gray, bioturbated, dolomitic limestone with algal structures and reddish-brown chert blebs (see Kirkland and Milner [2006] for discussion of contact); lower part may be equivalent to the Rock Point (= Church Rock) Member of the Chinle Formation (Kirkland and Milner, 2006); deposited in stream-channel and floodplain environments of a north-northwest-flowing stream system (Clemmensen and others, 1989; Blakey, 1994; Peterson, 1994; Tanner and Lucas, 2007); age from Molina-Garza and others (2003) and Lucas and Tanner (2006, 2007); ranges from about 150 to 250 feet (45–75 m) thick.

"J-0" unconformity. Evidence now indicates that the Jurassic-Triassic boundary is within the Dinosaur Canyon Member of the Moenave Formation and that the J-0 unconformity of Pipiringos and O'Sullivan (1978) (previously assumed to be at the base of the Jurassic) is within the Upper Triassic (Molina-Garza and others, 2003; Lucas and others, 2005; Kirkland and Milner, 2006).

TRIASSIC

TRu **Chinle and Moenkopi Formations, undivided** (Upper to Lower Triassic) – May include Petrified Forest and lower, middle, or upper red strata caught in the Hurricane fault near Toquerville Springs; may include parts of all members of these formations where mapped along the Reef Reservoir fault and in the northern Beaver Dam Mountains.

Chinle Formation

TRc **Chinle Formation, undivided** (Upper Triassic) – Shown on cross sections only.

TRcp **Petrified Forest Member** (Upper Triassic) – Varicolored, typically gray to purple mudstone, claystone, and siltstone, lesser white to yellow-brown sandstone and pebbly sandstone, and minor chert and nodular limestone; regionally divisible into three parts, in ascending order: (1) the bentonitic Blue Mesa, (2) the pebbly sandstone of the Moss Back or Sonsela (depending on clast provenance), and (3) the bentonitic Painted Desert (Lucas, 1993; Woody [2006], however, suggested restricting the Petrified Forest to strata above the Sonsela, that is, to Lucas' Painted Desert unit); Lucas (1993) proposed elevating the Chinle to group status, but that change has not been formally completed; the

Owl Rock Member of the Chinle Formation may be present at the top of the unit as mapped here (Spencer Lucas, New Mexico Museum of Natural History and Science, verbal communication, August 2005), and the lowermost part of the Petrified Forest Member as mapped here locally contains light-gray, fine- to medium-grained sandstone with abundant petrified wood and cobble- to small boulder-size "cannonball" sandstone concretions that may represent a distal, unmapped facies of the Monitor Butte Formation; the middle part of the member is marked by a distinctive, thick-bedded, white to yellow-brown sandstone and pebbly sandstone with well-rounded chert and quartzite clasts likely correlative with the Sonsela; swelling, bentonitic mudstone and claystone are common throughout the member and although typically poorly exposed, their bright purple, grayish-red, dark-reddish-brown, light-greenish-gray, brownish-gray, olive-gray, and similar hues locally show through to the surface – they weather to a "popcorn" surface and are responsible for numerous building foundation problems in the region; typically poorly exposed and commonly forms landslides; contains petrified wood, especially in sandstone beds; upper, unconformable contact is at the top of brightly colored swelling mudstones with nodular limestone, above which is a thin chert-pebble conglomerate and overlying reddish-brown, non-bentonitic siltstone and fine-grained sandstone; deposited in a variety of fluvial, floodplain, palustrine, and lacustrine environments that formed inland of a magmatic arc associated with a subduction zone along the west coast of North America (Stewart and others, 1972a; Dickinson and others, 1983; Lucas, 1993; Dubiel, 1994; DeCourten, 1998; Lucas and Tanner, 2007); about 400 to 650 feet (120–200 m) thick.

TRcs **Shinarump Conglomerate Member** (Upper Triassic) – Medium- to coarse-grained sandstone, pebbly sandstone, and lesser pebbly conglomerate, locally with silty sandstone, claystone, and smectitic claystone interbeds, that forms prominent cliffs, hogbacks, and mesas; clasts are subrounded chert and quartzite; mostly thick bedded with both planar and low-angle cross-stratification, although thin, platy beds with ripple cross-stratification occur locally; predominantly pale- to dark-yellowish-orange, but pale-red, grayish-red, pale-orange, and pale-yellow-brown hues are common; commonly stained by iron-manganese oxides, locally forming "picture stone"; contains poorly preserved petrified wood and plant debris, commonly replaced in part by iron-manganese oxides; upper contact marks a sharp topographic break and typically corresponds to the first appearance of varicolored, swelling mudstone; deposited in braided streams that flowed north and northwest (Stewart and others, 1972a; Dubiel, 1994; DeCourten, 1998); typically about 100 feet (30 m) thick, but ranges from 5 to 250 feet (2–75 m) thick due to deposition over paleotopography developed on the TR-3 unconformity.

TR-3 unconformity (Pipiringos and O'Sullivan, 1978)

Moenkopi Formation

TRm **Moenkopi Formation** (Lower Triassic) – West-dipping, fault-bounded blocks of mostly lower, middle, or upper red strata along the Hurricane fault; at Washington Dome, includes strata of the lower red and Virgin Limestone members; along the north side of Bloomington Dome, includes strata of the lower red and Timpoweap members; along the Reef Reservoir and Gunlock faults, it may include parts of all the members of the formation; the Moenkopi Formation probably does not include strata of Middle Triassic age in southwest Utah (Spencer Lucas, New Mexico Museum of Natural History and Science, verbal communication, August 2005). Lucas and others (2007) suggested elevating the Moenkopi Formation to Group status in southwest Utah, restricting it to the clastic facies of the Rock Canyon Conglomerate and the lower, middle, and upper red members; they suggested removing intervening carbonate- and evaporite-dominated members of the Timpoweap Limestone (Sinbad equivalent), Virgin Limestone, and Shnabkaib from the Moenkopi and assigning them to their Thaynes Group. We retain established member names for this compilation pending further work on regional nomenclature.

TRmu **Upper red member** (Lower Triassic) – Moderate-reddish-orange to moderate-reddish-brown, mostly thin- to medium-bedded siltstone, mudstone, and fine-grained sandstone with planar, low-angle, and ripple cross-stratification; locally contains thin gypsum beds and abundant discordant gypsum stringers and typically forms ledgy slopes; locally includes a prominent, cliff-forming, medium- to thick-bedded, 100-foot-thick (30 m), fine-grained sandstone near the base of the member (locally known as the pale-yellowish-orange Purgatory sandstone where it is bleached in the core of the Virgin anticline); upper, unconformable contact is at the base of the first coarse-grained, thick-bedded, pale-yellowish-brown pebbly sandstone of the Shinarump Conglomerate; deposited in tidal-flat and coastal-plain environments (Stewart and others, 1972b; Dubiel, 1994); generally thins to the northeast from about 400 to 600 feet (120–180 m) thick west of St. George to 200 to 280 feet (60–85 m) thick in the Zion National Park area.

TRms **Shnabkaib Member** (Lower Triassic) – Forms "bacon-striped," ledgy slopes of laminated to thin-bedded, gypsiferous, pale-red to moderate-reddish-brown mudstone and siltstone, resistant, white to greenish-gray gypsum, and minor thin, laminated, light-gray dolomite beds; gypsum is present as laterally continuous structureless beds, finely laminated commonly silty or muddy beds, nodular intervals as much as about 10 feet (3 m) thick, and as secondary cavity fillings and cross-cutting veins; in the Beaver Dam Mountains, the member contains numerous limestone beds as much as about 10 feet (3 m) thick, and far fewer red beds; typically weathers to soft, gypsiferous soils; upper conformable and gradational contact corresponds to the top of the highest thick gypsum bed; deposited in supratidal, intertidal, and subtidal environments on a broad, coastal shelf of low relief (Stewart and others, 1972b; Lambert, 1984); generally thickens westward from about 350 to 500 feet (105–150 m) east of the Hurricane Cliffs to about 800 to 1000 feet (245–300 m) west of St. George and in the Beaver Dam Mountains.

TRmm **Middle red member** (Lower Triassic) – Interbedded, slope-forming, laminated to thin-bedded, moderate-reddish-brown to moderate-reddish-orange siltstone, mudstone, and fine-grained sandstone with thin interbeds and veinlets of greenish-gray to white gypsum; lower part includes several thick gypsum beds; upper, conformable and gradational contact corresponds to the base of the first thick gypsum bed; deposited in a tidal-flat environment (Stewart and others, 1972b; Dubiel, 1994); generally thins westward from about 450 to 550 feet (140–170 m) thick east of the Hurricane Cliffs to about 300 to 400 feet (90–120 m) thick west of St. George.

TRmv **Virgin Limestone Member** (Lower Triassic) – Light-gray, light-olive-gray, and yellowish-brown limestone and silty limestone that typically forms three thin, resistant ledges at the base, middle, and top of the member, but in the Virgin River Gorge and White Hills contains five such limestone ledges; locally fossiliferous with circular and five-sided crinoid columnals, gastropods, and brachiopods; intervening slopes account for about 80% of the member and are mudstone and siltstone that have variable gray, yellowish-gray, and grayish-purple hues; upper, conformable contact corresponds to the top of the uppermost Virgin limestone bed; deposited in a variety of shallow-marine environments (Stewart and others, 1972b; Dubiel, 1994); generally thickens westward, but ranges from 0 to about 250 feet (0–75 m) thick due to deposition over paleotopography (best exposed southwest of St. George) that developed on the Permian-Triassic unconformity.

TRml **Lower red member** (Lower Triassic) – Interbedded, slope-forming, laminated to thin-bedded, moderate-reddish-brown mudstone, siltstone, and fine-grained sandstone with local, thin, laminated, light-olive-gray gypsum beds and veinlets; upper contact corresponds to the base of the first Virgin limestone bed; deposited in a tidal-flat environment (Stewart and others, 1972b; Dubiel, 1994); the upper part of the member at Kolob Canyons yielded the oldest known Early Triassic fossil vertebrate tracks in North America (Mickelson and others, 2006); generally thickens eastward, but ranges from 0 to about 300 feet (0–90 m) thick due to deposition over paleotopography (best exposed southwest of St. George) that developed on the Permian-Triassic unconformity.

TRmtr **Timpoweap and Rock Canyon Conglomerate Members, undivided** (Lower Triassic) – Locally undivided atop the Hurricane Cliffs south of the Hurricane.

TRmt **Timpoweap Member** (Lower Triassic) – Lower part consists of light-brown-weathering, light-gray to grayish-orange, thin- to thick-bedded limestone and cherty limestone; chert occurs as small disseminated blebs, thus giving weathered surfaces a rough texture; upper part consists of grayish-orange, thin- to thick-bedded, slightly calcareous, fine-grained sandstone, siltstone, and mudstone; both parts form low cliffs and ledges and form a gently undulating surface on top of the Permian-Triassic unconformity; upper contact with the lower red member is conformable and gradational and corresponds to a change from yellowish-brown, fine-grained sandstone, siltstone, and minor limestone below to reddish-brown siltstone and mudstone above (Biek, 2003b, 2007a); deposited in a shallow-marine environment (Nielson and Johnson, 1979; Dubiel, 1994; Lucas and others, 2007); Lucas and others (2007) reported a late Smithian age for the unit and reiterated a correlation with the Sinbad Limestone of central and eastern Utah; ranges from 50 to 180 feet (15–55 m) thick (Nielson, 1981).

TRmr **Rock Canyon Conglomerate Member** (Lower Triassic) – Consists of two main rock types: (1) a pebble to cobble, clast-supported conglomerate, with rounded chert and minor limestone clasts derived from the Harrisburg Member of the Kaibab Formation, which was deposited in paleovalleys and is typically 0 to several tens of feet thick, but locally as much as about 200 feet (60 m) thick; and (2) a widespread but thin breccia about 3 to 10 feet (1–3 m) thick, which probably formed as a regolith deposit on Harrisburg strata (Nielson, 1991); conformably and gradationally overlain by limestone of the Timpoweap Member and may locally include Timpoweap strata in the White Hills and Jarvis Peak quadrangles;

TR-1 unconformity (Pipiringos and O'Sullivan, 1978)

TRIASSIC AND PERMIAN

TRcPk **Chinle, Moenkopi, and Kaibab Formations, undivided** (Upper Triassic to Lower Permian) – Fault-bounded, highly fractured blocks containing various members of the Chinle, Moenkopi, and Kaibab Formations along the Hurricane fault.

TRmPk **Moenkopi and Kaibab Formations, undivided** (Lower Triassic to Lower Permian) – Fault-bounded, highly fractured blocks containing various members of the Moenkopi and Kaibab Formations along the Hurricane fault.

TRmPt **Moenkopi, Kaibab, and Toroweap Formations, undivided** (Lower Triassic to Lower Permian) – Fault-bounded, highly fractured blocks containing various members of the Moenkopi, Kaibab, and Toroweap Formations along the Grand Wash fault.

PERMIAN

Pkq **Kaibab Formation, Toroweap Formation, and Queantoweap Sandstone, undivided** (Lower Permian, Leonardian to Wolfcampian) – Fault-bounded, highly fractured blocks of Kaibab, Toroweap, and Queantoweap strata along the Hurricane fault.

Pkt **Kaibab and Toroweap Formations, undivided** (Lower Permian, Leonardian) – Fault-bounded, highly fractured blocks of Kaibab and Toroweap strata along the Hurricane fault.

Kaibab Formation

Pk **Kaibab Formation** (Lower Permian, Leonardian) – Fault-bounded, highly fractured blocks of the Harrisburg and Fossil Mountain Members along the Hurricane fault.

Pkh **Harrisburg Member** (Lower Permian, Leonardian) – Thin- to thick-bedded limestone and cherty limestone, and medium- to thick-bedded, laminated gypsum and gypsiferous mudstone; gypsum abundance increases markedly westward from the Hurricane Cliffs, where it is uncommon, to the Beaver Dam Mountains, where it makes up about one-half of the member; characterized by wavy, irregular bedding likely due to gypsum dissolution; forms ledgy slopes that enclose a resistant, cliff- and ledge-forming medial white chert and limestone interval; Late Permian to Early Triassic subaerial erosion created significant paleotopography on the Kaibab Formation, with channels commonly several tens of feet deep but locally several hundred feet deep; unconformably overlain by river-channel and breccia deposits of the Rock Canyon Conglomerate at the deepest levels of incision, and locally by the Timpoweap, lower red, or Virgin Limestone Members in former paleotopographic high areas mostly southwest of St. George; deposited in shallow-marine and sabkha environments (McKee, 1938; Nielson, 1981, 1986; Sorauf and Billingsley, 1991); the contact of the Harrisburg Member and basal Moenkopi Formation is locally difficult to identify (Biek, 2003b) and the uppermost Harrisburg Member as mapped by Billingsley and Workman (2000) apparently contains strata we assign to the Rock Canyon and Timpoweap Members of the Moenkopi Formation; ranges from 0 to 350 feet (0–100 m) thick due to erosion associated with the Permian-Triassic unconformity (Nielson, 1981).

Pkf **Fossil Mountain Member** (Lower Permian, Leonardian) – Lithologically uniform, light-gray, thick-bedded, fossiliferous limestone and cherty limestone; fossils are whole brachiopods, broken bryozoans, and disarticulated crinoids; "black-banded" due to abundant reddish-brown, brown, and black ribbon chert and irregular chert nodules that locally make up 20 to 40% of the rock; forms a prominent cliff in the Hurricane Cliffs and Beaver Dam Mountains; upper conformable contact with Harrisburg strata corresponds to a sharp break in slope with gypsiferous mudstone above and limestone cliff below; deposited in a shallow-marine environment (McKee, 1938; Nielson, 1981, 1986; Sorauf and Billingsley, 1991); maintains a fairly uniform thickness of about 200 to 300 feet (60–90 m) (Nielson, 1981).

P-k unconformity (Blakey, 1996)

Toroweap Formation

Pt **Toroweap Formation** (Lower Permian, Leonardian) – Fault-bounded, highly fractured blocks of the Woods Ranch, Brady Canyon, and Seligman Members in the Beaver Dam Mountains and along the Hurricane fault in the Kolob Canyons area.

Ptw **Woods Ranch Member** (Lower Permian, Leonardian) – Laterally variable, interbedded, yellowish-gray to light-gray, laminated to thin-bedded dolomite and similarly bedded black chert, massive gypsum, yellowish-orange gypsiferous mudstone and siltstone, and limestone; forms slopes and strike valleys and so is typically poorly exposed, but contains three principal gypsum beds in the Virgin River Gorge, each 20 to 50 feet (6–15 m) thick (Hintze, 1986); contains local intraformational conglomerate and collapse breccia in Hurricane Cliff exposures; eastern exposures in the Hurricane Cliffs contain significantly more carbonate and lesser gypsum compared to exposures in the Beaver Dam Mountains and Virgin River Gorge; Nielson (1986) noted that much of the gypsum in the Beaver Dam Mountains is replaced by aragonite and calcite; upper contact with Fossil Mountain strata corresponds to a sharp break in slope with slope-forming, gypsiferous siltstone and thin limestone beds below and cliff-forming cherty limestone above—Billingsley and Workman (2000) described an unconformity with about 15 feet (4.5 m) of relief on this contact to the south; deposited in a complex sequence of shallow-marine and sabkha environments (McKee, 1938; Rawson and Turner-

Peterson, 1979, 1980; Nielson, 1981, 1986); gypsum dissolution and structural complications account for a variable thickness of 65 to 870 feet (20–265 m) in the Beaver Dam Mountains (Hintze and Hammond, 1994; Hayden and others, 2005a) and about 120 to 320 feet (35–98 m) in the Hurricane Cliffs (Nielson, 1981).

Ptbs **Brady Canyon and Seligman Members, undivided** (Lower Permian, Leonardian) – Undivided where too thin to show separately.

Ptb **Brady Canyon Member** (Lower Permian, Leonardian) – Lithologically uniform, light- to medium-gray, medium- to coarse-grained, thick-bedded, fossiliferous limestone and cherty limestone; fossils are broken brachiopods, bryozoans, and crinoids; ribbon chert and irregular chert nodules locally make up 30 to 40% of the rock; forms a prominent gray cliff, similar to that developed in Fossil Mountain strata, in the lower part of the Hurricane Cliffs and in the Beaver Dam Mountains; the base of the member is dolomitic in the Beaver Dam Mountains; upper contact with Woods Ranch strata corresponds to a sharp break in slope, with cliff-forming cherty limestone below and slope-forming, gypsiferous siltstone and thin limestone beds above; deposited in a shallow-marine environment (McKee, 1938; Rawson and Turner-Peterson, 1979, 1980; Nielson, 1981, 1986); typically about 250 feet (75 m) thick in the Beaver Dam Mountains and about 160 to 250 feet (48–75 m) thick in the Hurricane Cliffs (Nielson, 1981).

Pts **Seligman Member** (Lower Permian, Leonardian) – Forms slopes of yellowish-brown to grayish-orange, thin-bedded, planar-bedded, fine- to medium-grained sandstone and minor siltstone with brown-weathering nodular chert; contains gypsiferous siltstone and minor gypsum in the Beaver Dam Mountains and Virgin River Gorge, but Nielson (1986) noted that much of the gypsum is replaced by aragonite and calcite; upper contact with the Brady Canyon Member corresponds to a sharp break in slope, with slope-forming, thin-bedded sandstone below and cliff-forming cherty limestone above; deposited in shallow-marine and beach environments (McKee, 1938; Rawson and Turner-Peterson, 1979, 1980; Nielson, 1981, 1986); thickens westward from about 30 feet (9 m) in the Hurricane Cliffs to about 100 to 200 feet (30–60 m) in the Beaver Dam Mountains (Nielson, 1981).

P-tw unconformity (Blakey, 1996)

Pq **Queantoweap Sandstone** (Lower Permian, Leonardian to Wolfcampian) – Yellowish-brown, pale-orange, and grayish-orange, thick-bedded, cross-bedded, fine- to medium-grained sandstone that weathers to a conspicuous stair-step topography; forms the bulk of the Hurricane Cliffs near Pintura, where it is typically noncalcareous quartz arenite, likely bleached by reducing fluids in the core of the Pintura anticline; also exposed in the Beaver Dam Mountains, Virgin River Gorge, and at Square Top Mountain where it has a calcareous cement; in the Hurricane Cliffs near Pintura, lower 120 feet (37 m) is slope-forming, brown-weathering, brownish-gray, thin- to medium-bedded, fine-grained sandy dolomite, calcareous sandstone, and silty sandstone that is locally vuggy-weathering; upper half stratigraphically equivalent to the Hermit Formation of northwest Arizona and southeast Nevada and the lower half is stratigraphically equivalent to the Esplanade Sandstone of northwest Arizona (Blakey, 1996; Billingsley, 1997); upper contact with the Seligman Member corresponds to a subtle break in slope, with ledge- and cliff-forming cross-bedded sandstone below and slope- and ledge-forming, planar-bedded sandstone and gypsiferous, sandy siltstone above; deposited in shallow-marine, beach, and dune environments (Nielson, 1981; Johansen, 1988); age from Blakey (1996); about 1400 to 2000 feet (425–600 m) thick in the Beaver Dam Mountains (Hintze and Hammond, 1994; Hintze and others, 1994), and about 1400 to 1700 feet (425–520 m) thick in the Hurricane Cliffs just south of Pintura (Hurlow and Biek, 2003).

Pp **Pakoon Dolomite** (Lower Permian, Wolfcampian) – Light- to medium-gray, thin- to thick-bedded dolomite, dolomitic limestone, and minor limestone, all with yellowish-brown-weathering, light-gray to white, irregularly shaped chert nodules; in the Beaver Dam Mountains, the upper 60 feet (18 m) is mostly gypsum with minor limestone and sandstone; contains sparse macrofossils, including bryozoans, fusilinids, and crinoid columnals, especially in thin limestone beds in the upper part; weathers to low ledges and slopes; stratigraphically equivalent to the lower Esplanade Sandstone of northwest Arizona (Blakey, 1996; Billingsley, 1997) and to the upper part of the Bird Spring Formation of southeast Nevada (Hintze and Axen, 1995); in the Beaver Dam Mountains, upper contact is gradational and corresponds to the top of the gypsiferous interval above which is yellowish-brown, cross-bedded sandstone, whereas at Black Ridge, this same contact appears sharp and corresponds to the first appearance of thin-bedded, yellowish-brown-weathering sandy dolomite; deposited in a shallow-marine environment (McNair, 1951); 600 to 850 feet (180–260 m) thick in the Beaver Dam and southern Bull Valley Mountains (Hintze and Hammond, 1994; Hintze and others, 1994).

P-0 unconformity (Blakey, 1996)

PERMIAN AND PENNSYLVANIAN

PIPpc **Pakoon Dolomite and Callville Limestone, undivided** (Lower Permian to Lower Pennsylvanian) – Locally undivided near the Mineral Mountain intrusion in the southern Bull Valley Mountains.

PIPbs **Bird Spring Formation** (Lower Permian to Pennsylvanian) – Mapped only west of Beaver Dam Wash, where strata equivalent to the Pakoon Dolomite and Callville Limestone are collectively known as the Bird Spring Formation; unit description after Hintze and Axen (1995); cyclically interbedded limestone and cherty limestone, dolomite, and sandstone; fossils include brachiopods, corals, and bryozoan and crinoid debris; formation is mostly limestone in lower part and becomes more dolomitic in upper part; forms distinctive ledge-slope topography; upper contact is gradational and placed at the top of the uppermost carbonate bed; deposited in a shallow-marine environment (Rich, 1963); estimated thickness of 2000 feet (600 m) may be high due to unmapped faults that repeat parts of the formation.

PENNSYLVANIAN

IPc **Callville Limestone** (Upper to Lower Pennsylvanian, Virgilian to Morrowan) – Medium-gray, fine- to medium-grained, medium- to thick-bedded limestone with cyclic interbeds of moderate-orange-pink calcareous siltstone, sandstone, and light-gray dolomite increasing in the upper third; commonly cherty and fossiliferous; *Chaetetes* coral is common in the upper part, whereas brachiopods and bryozoans are common in limestone beds throughout (Hintze, 1986); forms ledge-slope topography in the Beaver Dam Mountains and at Square Top Mountain, similar to the overlying Pakoon Dolomite; stratigraphically equivalent to the lower part of the Bird Spring Formation of southeast Nevada (Hintze and Axen, 1995); upper contact is placed at the base of the lighter-colored, non-fossiliferous dolomite beds of the Pakoon Dolomite; represents shallow-marine deposits (Welsh and Bissell, 1979); age from Welsh and Bissell (1979); about 1500 to 2000 feet (460–600 m) thick in the Beaver Dam and southern Bull Valley Mountains (Hammond, 1991; Hintze and Hammond, 1994; Hintze and others, 1994).

unconformity

MISSISSIPPIAN AND CAMBRIAN

MCu **Redwall Limestone and Bonanza King Formation, undivided** (Lower Mississippian and Upper and Middle Cambrian) – Undivided at Castle Cliff due to map scale.

MISSISSIPPIAN

Msc **Scotty Wash Quartzite and Chainman Shale, undivided** (Upper Mississippian) – As first recognized by Adair (1986) and Willden and Adair (1986), the overlying **Scotty Wash Quartzite** is a resistant, medium- to dark-gray and brown, well-bedded, shallow-marine sandstone and quartzite, with minor thin sandy shale beds, that has a maximum thickness of about 80 feet (25 m); the underlying **Chainman Shale** is a non-resistant, black, dark-gray, and greenish-gray marine shale that has a maximum thickness of about 80 feet (25 m). The presence of these strata in the upper plate of the Square Top Mountain thrust, but not in the lower plate of the Beaver Dam Mountains, suggests substantial displacement on the Square Top Mountain thrust fault, placing strata of deeper basin affinity over correlative but thinner platform strata.

Mr **Redwall Limestone** (Lower Mississippian) – Divided into four members in the Grand Canyon by McKee (1963; see also McGee and Gutschick, 1969), in descending order the Horseshoe Mesa, Mooney Falls, Thunder Springs, and Whitmore Wash Members, which Steed (1980) described but did not map in the Virgin River Gorge, and that remain undivided in the Beaver Dam Mountains (Hintze, 1986); known as the Monte Cristo Limestone in the Basin and Range (Hewitt, 1931), with its Yellowpine and Arrowhead Members (Horseshoe Mesa lithologic equivalent), Bullion Member (Mooney Falls lithologic equivalent), Anchor Member (Thunder Springs lithologic equivalent), and Dawn Member (Whitmore Wash lithologic equivalent); typically forms a single uniform cliff of medium- to dark-gray, thick-bedded, locally cherty and fossiliferous limestone; in the Beaver Dam Mountains, the basal 60 feet (18 m) is coarse grained and dolomitic, above which is an 80-foot-thick (25 m), dark-yellowish-brown weathering, thin-bedded, cherty, bioclastic limestone—likely the Thunder Springs Member—that forms a prominent marker horizon; upper 460 feet (140 m) is bioclastic and fossiliferous, containing horn corals, colonial corals, and brachiopods (Hintze, 1985a; Hayden and others, 2005a); upper contact corresponds to a prominent break in slope, with massive Redwall Limestone cliffs below and cyclic, ledgy slopes above; also present as gravity-slide blocks west of the Beaver Dam Mountains; deposited on a gently sloping shallow-marine shelf as two transgressive (Whitmore Wash and Mooney Falls Members) and two regressive (Thunder Springs and Horseshoe Mesa Members) sequences (McKee and Gutschick, 1969); ranges from 615 feet (188 m) thick in the Virgin River Gorge (Steed, 1980) to about 850 feet (260 m) thick at Horse Canyon in the West Mountain Peak quadrangle (Hintze, 1986).

Mm **Monte Cristo Limestone** (Lower Mississippian) – Mapped only west of Beaver Dam Wash, where strata equivalent to the Redwall Limestone are known as the Monte Cristo Limestone (Hintze and Axen, 1995); thick-bedded, light-gray limestone and cherty limestone; about 1000 feet (300 m) thick.

DEVONIAN

Dm, Dm?

Muddy Peak Dolomite (Upper Devonian) – Consists of two informal units (Hintze, 1985b, 1986) not mapped separately: (1) upper "pinnacle" unit is light- to medium-gray, medium- to coarse-grained, thick-bedded dolomite with sandy laminae and scattered chert nodules, which weathers to form light-gray pinnacles below the massive Redwall Limestone cliffs; (2) lower "slope" unit is light-olive-gray to pale-yellowish-gray, fine-grained, thin- to medium-bedded, silty dolomite with ledge-forming sandstone beds near the top, which weathers to a ledgy slope between more-resistant Nopah and pinnacle dolomites; "slope" unit includes small, silicified, stromatoporoid-like structures as well as rare biohermal mounds that include crinoid debris and scattered coral, gastropod, and brachiopod fragments (Hintze, 1985b); query indicates uncertain identification in upper plate of low-angle normal fault along west margin of Beaver Dam Mountains; collectively equivalent to the Temple Butte Formation as exposed in the Virgin Mountains (Beus, 1980; Billingsley and Workman, 2000); deposited in a shallow-marine environment (Beus, 1980); "pinnacle" unit is 140 to 190 feet (45–55 m) thick, whereas "slope" unit is 300 to 450 feet (90–135 m) thick (Hayden and others, 2005a); Hammond (1991) reported a thickness of 680 feet (207 m) for the entire formation in the Shivwits quadrangle.

unconformity Hintze and Axen (1995) mapped about 1600 feet (500 m) of Ordovician and Silurian strata (Pogonip Group, Eureka Quartzite, and unnamed dolomite) in the upper plate of the Tule Springs thrust in the adjacent Scarecrow Peak quadrangle, just 2 to 4 miles (3–6 km) west of the Utah-Nevada state line; Ordovician and Silurian strata are absent in the Beaver Dam Mountains (and in all of central and eastern Utah) due to widespread erosion in the early Devonian, likely associated with the Antler orogeny in central Nevada.

CAMBRIAN

Cn **Nopah Dolomite** (Upper Cambrian) – Light-gray to brownish-gray, fine- to medium-grained, generally thick-bedded dolomite; contains algal stromatolites in upper part, small tubular trace fossils (twiggy bodies), and mottled zones that suggest bioturbation during deposition in a warm shallow-marine environment (Hintze, 1986); thin, slope-forming, orangish- to greenish-gray, glauconitic silty dolomite or limestone unit at the base of the Nopah may correspond to the Dunderberg Shale of southern Nevada (Hintze, 1986); similar to Bonanza King Formation, but, except for lowest part, is more massive and forms steeper ledges and cliffs; upper contact is a major regional unconformity and corresponds to the base of the more gentle slope of the "slope" unit of the Devonian Muddy Peak Dolomite; about 1300 feet (400 m) thick (Hintze, 1985b, 1986; Hayden and others, 2005a).

Cbk **Bonanza King Formation** (Upper and Middle Cambrian) – Medium- to light-brownish-gray, fine- to medium-grained, medium- to thick-bedded dolomite with some bluish-gray silty limestone beds in the lower part (Hintze, 1986); upper half consists of numerous thin beds of light-gray, fine-grained boundstone (Hintze, 1985b; Hammond, 1991); commonly mottled light and medium gray, and typically forms banded light- and dark-gray ledges and cliffs; commonly pervasively brecciated; olive-gray, slope-forming, argillaceous limestone near base may be equivalent to the Chisholm Shale of the Pioche, Nevada, area (Hintze, 1985b, 1986); upper contact with the more massive, ledge- and cliff-forming Nopah Dolomite is marked by a thin slope-forming interval; equivalent to the Mauv Limestone of the Grand Canyon region and so mapped on the adjacent Littlefield 30' x 60' quadrangle (Billingsley and Workman, 2000), but similarity to equivalent strata in southern Nevada led Hintze (1986) to use the miogeoclinal Bonanza King nomenclature; 2600 feet (800 m) thick along the north side of Horse Canyon in the West Mountain Peak quadrangle, but the base may not be exposed due to attenuation faulting (Hintze and Hammond, 1994; Hayden and others, 2005a).

Cba **Bright Angel Shale** (Middle to Lower Cambrian) – Olive-green, slope-forming, micaceous shale, siltstone, and fine-grained quartzose sandstone; contains a ledge-forming, pale-brown-weathering, 10-foot-thick (3 m) dolomite about 125 feet (38 m) above the base and a 6-inch-thick (15 cm) limestone bed about 50 feet (15 m) below the top; weathers to strike valleys, and is commonly thinned or absent due to attenuation faulting; upper contact is gradational and drawn at the base of the limestone ledges of the Bonanza King Formation; stratigraphically equivalent to the Pioche Shale, Lyndon Limestone, and Chisolm Shale of the Pioche, Nevada, area (Page and others, 2005); deposited in shallow-marine environment (Stewart, 1970); 250 to 350 feet (75–100 m) thick (Hayden and others, 2005a).

Ct **Tapeats Sandstone** (Lower Cambrian) – Dark-reddish-orange to pale-reddish-brown, thin- to thick-bedded, medium- to coarse-grained orthoquartzite with a few thin beds of gritstone and quartz-pebble conglomerate; forms ledges; upper contact is gradational where the quartzite gives way to the shale and siltstone of the Bright Angel Shale; deposited in sandy intertidal and nearshore environments as the coarse-grained facies of the transgressing Cambrian sea (McKee, 1945; Hereford, 1977); 1300 feet (400 m) thick in the northern and eastern Beaver Dam Mountains (Hammond, 1991; Hintze and Hammond, 1994), but, due to attenuation faulting, only about 600 feet (180 m) thick south and west of West Mountain Peak.

Great Unconformity

PRECAMBRIAN

Xu **Gneiss, schist, and pegmatite** (Paleoproterozoic) – Paragneiss, orthogneiss, para-amphibolite and felsic and intermediate intrusive rocks cut by pegmatite sills and dikes; these rocks are also exposed in the Virgin Mountains of northwest Arizona where Moore (1972) reported a complex assemblage of garnet-biotite gneiss, garnet-sillimanite-biotite schist, granite gneiss, granodiorite gneiss, and amphibolite; forms the exposed core of the Beaver Dam Mountains of Utah, with dioritic gneiss the most resistant and most extensively exposed rock type (Hintze, 1985a, 1985b, 1986; Hayden and others, 2005b); granitic pegmatite is common and intrudes both gneiss and schist, principally parallel to foliation; less common, white pegmatite composed principally of orthoclase and quartz with minor muscovite, plagioclase, and garnet commonly exhibits graphic granite texture (Hintze, 1986); Colberg (2007) and Fitzgerald and Colberg (2008) reported pressure-temperature conditions for the amphibolite and metapelitic units in the range of 700 to 800°C and 0.7 to 0.8 GPa, and evidence of an earlier, higher-pressure phase of metamorphism, suggesting that the rocks once occupied a lower crustal position; upper contact is an unconformity of approximately 1.2 billion years, referred to in the Grand Canyon region as the "Great Unconformity"; Nelson and others (2007) reported laser-ablation ICP-MS U-Pb mean ages of 1739 $\pm$ 8.4 Ma and 1728 $\pm$ 4.8 Ma for paragneiss and orthogneiss, respectively, with a few grains having apparent older $^{207}Pb/^{206}Pb$ ages (between 1800 and 1900 Ma for paragneiss and between 1760 and 1790 Ma for orthogneiss); King (1976) compared these rocks to the Paleoproterozoic Vishnu and Brahma schists of the Grand Canyon (now the Grand Canyon Metamorphic Suite); a pegmatite in similar Precambrian rocks in the nearby East Mormon Mountains, Nevada, yielded a 1.7 billion-year K-Ar age (Olmore, 1971); Karlstrom and Bowring (1993) and Karlstrom and others (2005) considered these rocks Paleoproterozoic in age and part of the Mojave Province; Quigley and others (2002) considered similar rocks in the North Virgin Mountains as Paleoproterozoic.