GEOLOGIC MAP OF THE PANGUITCH 30' X 60' QUADRANGLE, GARFIELD, IRON, AND KANE COUNTIES, UTAH

Robert F. Biek, Peter D. Rowley, John J. Anderson, Florian Maldonado, David W. Moore, David B. Hacker, Jeffrey G. Eaton, Richard Hereford, Edward G. Sable, Harry F. Filkorn, and Basia Matyjasik

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UTAH GEOLOGICAL SURVEY
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UTAH DEPARTMENT OF NATURAL RESOURCES
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2015

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SCALE: 1:62,500

A companion virtual geologic field trip and transparent 3D geologic map overlay of this quadrangle will soon be available on the Utah Geological Survey website under our geologic guides web page.

Cover photo: *Pinnacles eroded from Tertiary-age lava flow breccia along State Route 20, Garfield County.*

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GEOLOGIC MAP OF THE PANGUITCH 30' X 60' QUADRANGLE, GARFIELD, IRON, AND KANE COUNTIES, UTAH

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ABSTRACT

The Panguitch 30′ x 60′ quadrangle spans the southernmost High Plateaus in southwest Utah, a structural and stratigraphic transition zone between the highly extended Basin and Range Province on the west and the colorful, mostly flat-lying strata of the Colorado Plateau on the east. The structural grain of the map area is dominated by north- to northeast-trending basin-range normal faults that bound relatively unfaulted blocks represented by the Markagunt and Paunsaugunt-Sevier Plateaus. That grain is imprinted on the leading edge of the Sevier orogenic belt, whose thrust faults and folds are dramatically displayed in the Red Hills and in Cedar Canyon at the west edge of the map area, and whose subtle effects are documented eastward to the Paunsaugunt fault zone.

The map area also includes the southern margin of the Marysvale volcanic field. Gravitational spreading of the southern sector of the volcanic field created two unusual, east-trending, south-vergent structures, one deep-seated and the other shallower. The deeper structure consists of thrust faults and folds of the Rubys Inn thrust fault zone, which is much younger than and trends nearly at a right angle to Sevier-age structures. Our mapping suggests that the Rubys Inn thrust fault continues westward into the Markagunt Plateau as a blind thrust fault roughly coincident with the southern margin of the shallower structure, the Markagunt gravity slide. The Markagunt gravity slide, which apparently covers at least 1600 square miles (4160 km^2) of the central and northern Markagunt Plateau and adjacent area, resulted from catastrophic collapse of the southwestern sector of the Marysvale volcanic field in the early Miocene, about 21 to 22 million years ago. It is thus somewhat larger than the famous Heart Mountain detachment of northwestern Wyoming, long known as the largest subaerial gravity slide in the world. We remain uncertain of what caused the collapse, but suggest that it may be related to pre-caldera inflation of the 20–18 Ma Mount Belknap caldera.

By mapping coarse alluvial strata associated with major sequence boundaries, we correlate Upper Cretaceous strata of the Markagunt Plateau with better exposed and more thoroughly studied strata of the Kaiparowits basin east of the map area. Specifically, we restrict the Grand Castle Formation to its upper conglomerate member and show that the lower conglomerate member of the Grand Castle Formation is in fact the Drip Tank Member of the Straight Cliffs Formation, and that the middle sandstone member of the Grand Castle Formation is in fact the capping sandstone member of the Wahweap Formation. Our mapping also constrains the evolution and demise of the Claron depositional basin, whose colorful strata are famously known at Cedar Breaks National Monument and Bryce Canyon National Park. We report new late Eocene ages for the inception of volcanism in this part of southwest Utah, we map the distribution of regional ash-flow tuffs that erupted from Oligocene and Miocene calderas near the Utah-Nevada border (documenting paleotopography of the region and constraining the inception of basin-range faulting), and we map 38 relatively young basaltic lava flows and cinder cones, the most recent phase in the volcanic legacy of the southernmost High Plateaus. Several of these lava flows cross and are offset by faults associated with basin-range extension and so serve as control points for long-term slip rates on major normal faults in the area.

The Panguitch 30′ x 60′ quadrangle consists of 32 individual 7.5′ quadrangles, about half of which had previously published mapping useful for compilation. Much of that geologic mapping resulted from the U.S. Geological Survey's (USGS) BARCO project during the mid-1980s to mid-1990s. The BARCO project (a multidisciplinary federal effort to better understand the geology of the Basin and Range–Colorado Plateau transition in southwest Utah and adjacent Nevada and Arizona) was disbanded following reorganization of the USGS in the mid-1990s. Nevertheless, the combined work laid a solid foundation for future mapping in the region, which we completed from 2007 to 2013. This resulting new mapping and compilation shows the regional geology of Utah's southern High Plateaus in unprecedented detail.

INTRODUCTION

The Panguitch 30′ x 60′ quadrangle spans the southernmost High Plateaus in southwest Utah, a structural and stratigraphic transition zone between the highly extended Basin and Range Province on the west and the colorful, mostly flat-lying strata of the Colorado Plateau on the east (figure 1). Parts of four major, north- to northeast-trending, down-to-the-west, basin-range normal faults cross the map area, cutting and bounding these physiographic regions. From west to east, these are (1) the Hurricane fault zone at the west edge of the Red Hills, (2) the Parowan-Paragonah fault zone that forms the west edge of the Markagunt Plateau (and thus the east margin of the Basin and Range Province at this latitude), (3) the Sevier fault zone, which bisects the High Plateaus, and (4) the Paunsaugunt fault zone, which forms the eastern structural margin of the Paunsaugunt and Sevier Plateaus. A simple west-to-east cross section of the map area (figure 2) shows the horst block of the Red Hills bounded by grabens of northern Cedar Valley and Parowan Valley, the gently east-tilted block of the Markagunt Plateau, the Paunsaugunt and Sevier Plateaus, and finally the westernmost edge of the Colorado Plateau. The structural grain of the map area is thus dominated by north- to northeast-trending normal faults that bound relatively unfaulted blocks represented by the Markagunt and Paunsaugunt-Sevier Plateaus. That grain is imprinted on the leading edge of the Sevier orogenic belt (Cretaceous to early Tertiary), whose thrust faults and folds are dramatically displayed in the Red Hills and in Cedar Canyon, and whose subtle effects are noticed eastward to the Paunsaugunt fault zone.

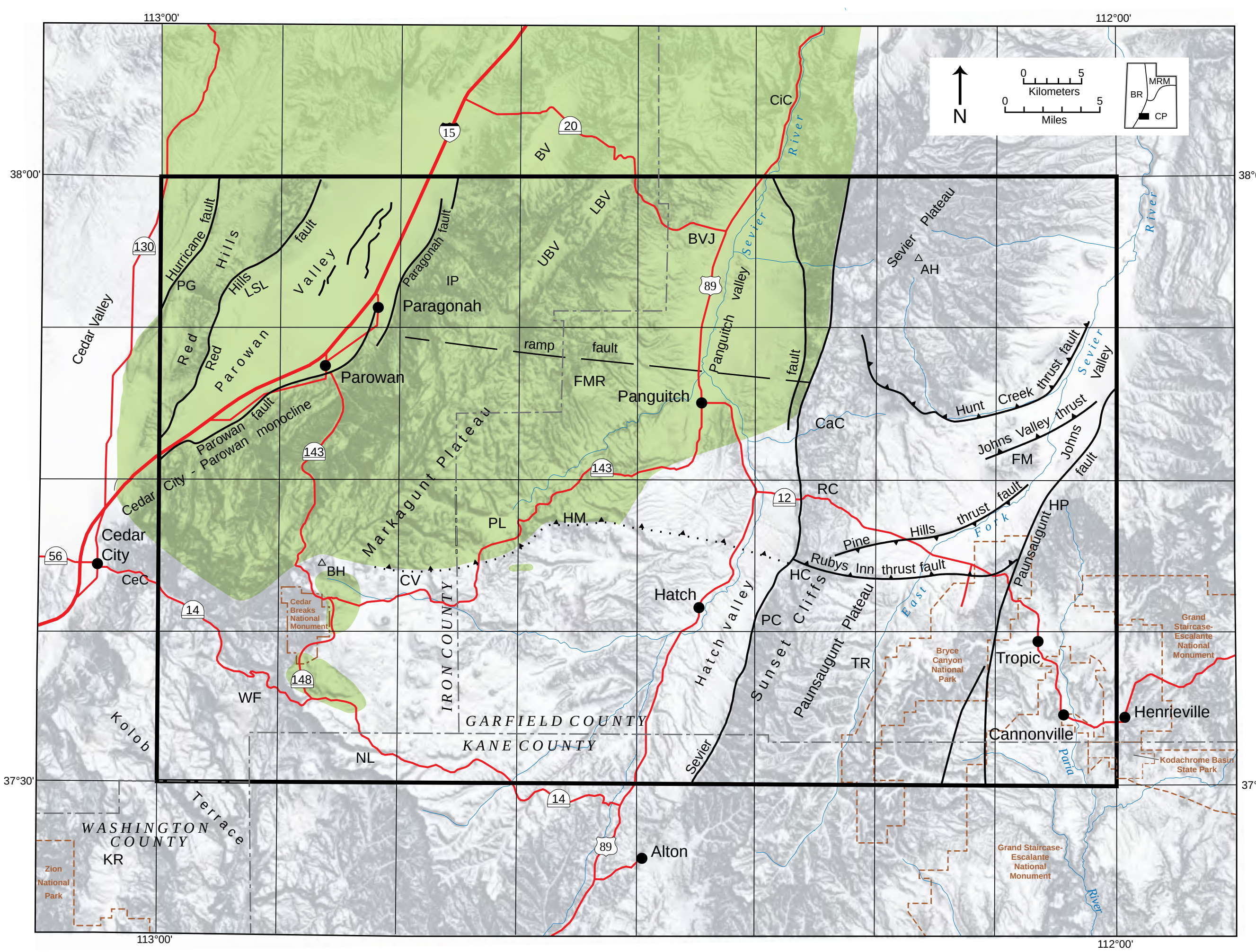


Figure 1. *Shaded-relief image of the Panguitch 30' x 60' quadrangle, showing major fault zones that form the boundaries of the Basin and Range Province and Colorado Plateau; the Markagunt, Paunsaugunt, and Sevier Plateaus, part of Utah's High Plateaus, together form a structural and stratigraphic transition zone between the two provinces. Heavy line marks boundary of the map area; grid shows boundaries of 7.5' quadrangle maps (see figure 6 for quadrangle names). Green area shows extent of Markagunt gravity slide.*

Inset shows major physiographic provinces of Utah: BR, Basin and Range; CP, Colorado Plateau; MRM, Middle Rocky Mountains. Key sites discussed in text: AH, Adams Head; BH, Brian Head peak; BV, Buckskin Valley; BVJ, Bear Valley Junction; CeC, Cedar Canyon; CiC, Circleville Canyon; CaC, Casto Canyon; CV, Castle Valley; FM, Flake Mountain; FMR, Fivemile Ridge; HC, Hillsdale Canyon; HM, Haycock Mountain; HP, Henderson Point; IP, Iron Peak; KR, Kolob Reservoir; LBV, Lower Bear Valley; LSL, Little Salt Lake; NL, Navajo Lake; PC, Proctor Canyon; PG, Parowan Gap; PL, Panguitch Lake; RC, Red Canyon; TR, Tropic Reservoir; UBV, Upper Bear Valley; WF, Webster Flat.

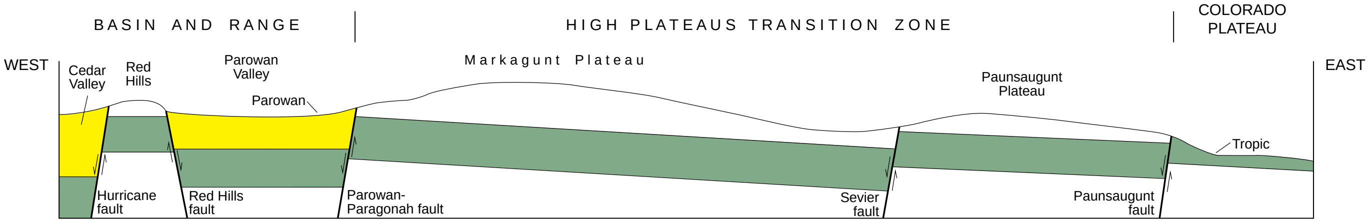


Figure 2. *Schematic west-to-east cross section through the northern Panguitch 30′ x 60′ quadrangle. Note gentle east dip of strata on the Markagunt and Paunsaugunt Plateaus, which are dropped down to the west along the Sevier and Paunsaugunt faults. The Red Hills horst block is bounded by the Hurricane and Red Hills faults to create the easternmost range of the Basin and Range Province at this latitude. Yellow denotes basin fill, green denotes Cretaceous strata.*

The map area also includes the southern margin of the Marysvale volcanic field, one of the largest eruptive centers in the western United States, characterized by Oligocene-Miocene stratovolcanoes and subordinate yet important calderas. Gravitational spreading of the southern sector of the volcanic field created two unusual, east-trending, south-vergent structures, one deep-seated and the other shallower. The deeper structure consists of thrust faults and folds of the Paunsaugunt thrust fault system, which is much younger than and trends nearly at a right angle to Sevier-age structures. We present evidence that of these thrust faults, the Rubys Inn thrust fault continues westward into the Markagunt Plateau as a blind thrust fault roughly coincident with the southern margin of the shallower structure, the Markagunt gravity slide. The Markagunt gravity slide spans at least 1600 square miles (4160 km^2) and resulted from catastrophic collapse of the southwestern sector of the Marysvale volcanic field in the early Miocene, 21 to 22 million years ago. It is thus somewhat larger than the famous Heart Mountain detachment, long known as the largest subaerial gravity slide in the world, which resulted from catastrophic collapse, in the Eocene, of the Absaroka volcanic field in northwest Wyoming (see, for example, Malone and Craddock, 2008; Beutner and Hauge, 2009; Craddock and others, 2009).

The oldest rocks exposed in the map area, at Parowan Gap, belong to the Lower Jurassic Navajo Sandstone. Middle Jurassic strata are also exposed there, as well as in Cedar Canyon and south of Tropic. However, it is the Upper Cretaceous, Paleocene, and Eocene strata for which the quadrangle is most noteworthy (figure 3). Our mapping correlates Upper Cretaceous strata at the west edge of the Markagunt Plateau with better exposed and more thoroughly studied strata of the Kaiparowits basin east of the map area. The mapping also constrains the evolution and demise of the Claron depositional basin, whose colorful strata are famously known at Cedar Breaks National Monument and Bryce Canyon National Park. We report new late Eocene ages for the inception of volcanism in this part of southwest Utah, we map the distribution of regional ash-flow tuffs that erupted from Oligocene and Miocene calderas near the Utah-Nevada border (documenting paleotopography of the region and constraining the inception of basin-range faulting), and we map 38 relatively young basaltic lava flows and cinder cones, the most recent phase in the volcanic legacy of the southernmost High Plateaus (figure 4). Because the map area contains such a wide range of young and old lava flows and ash-flow tuffs, it is adorned with inverted valleys representing each stage in the evolution of these enigmatic features. Several of these lava flows crossed and are offset by faults associated with basin-range extension and so serve as control points for estimating long-term slip rates on major normal faults in the area.

The Panguitch 30′ x 60′ quadrangle straddles parts of three ecoregions (large areas having geographically distinct plant and animal communities): Central Basin and Range, Wasatch and Uinta Mountains, and Colorado Plateaus (figure 5), each corresponding roughly to the three physiographic regions described above. The western escarpment of the Markagunt Plateau represents an especially abrupt transition from the dry, high-elevation basins below to the spruce- and fir-clad plateau above. The Markagunt, along with the Sevier and Paunsaugunt Plateaus, serve as critical watersheds that sustain life in the dry basins.

The map area is relatively sparsely populated, with most residents living in the small towns of Parowan, Paragonah, Summit, Panguitch, Hatch, Cannonville, and Tropic. Yet the region is one of great natural beauty and year-round recreational interest. It includes the Pink Cliffs, the uppermost riser and tread of the Grand Staircase, showcased at Cedar Breaks National Monument and Bryce Canyon National Park. The northwest corner of Grand Staircase–Escalante National Monument and part of Kodachrome Basin State Park are in the southeast corner of the map area. The area's diverse geology, accessibility, and spectacular scenery make it an ideal outdoor geological classroom. We are working on a virtual geologic fieldtrip that highlights many of our favorite geologic sites and features in the map area, similar to the virtual geologic guide of the St. George 30′ x 60′ quadrangle (see http://geology.utah.gov/geo_guides/st_george/index.htm).

In the text that follows we discuss the most widely exposed sedimentary rocks in the Panguitch 30′ x 60′ quadrangle, especially as they relate to changing depositional environments and the structural evolution of this part of southwest Utah during the Late Cretaceous and early Tertiary. We describe middle Tertiary volcanic rocks and latest Tertiary and Quaternary basaltic lava flows and show how they can be used to un-

Figure 3. *Among its sedimentary strata, the Panguitch 30' x 60' quadrangle is most noteworthy for its record of Late Cretaceous, Paleocene, and Eocene time. This map correlates Upper Cretaceous strata at the west edge of the Markagunt Plateau with better exposed and more thoroughly studied strata of the Kaiparowits basin east of the map area, and it also offers new insights and age control of the Claron depositional basin.*

Here, in the lower reaches of Left Hand Canyon south of Parowan, the lower two members of the former Grand Castle Formation are now correlated with the Drip Tank Member of the Straight Cliffs Formation (Ksd) and the capping sandstone of the Wahweap Formation (Kwcs); the Grand Castle Formation, now restricted to its upper member (Kgc), underlies colorful strata of the pink member of the Claron Formation (Tcp). John Henry Member of the Straight Cliffs Formation (Ksj); exhumed basaltic vent area (Qbc)

Figure 4. *Panguitch Lake lava flows immediately south of Panguitch Lake, the youngest of several dozen basaltic lava flows on and near the Markagunt Plateau. A radiometric age is pending on the youngest of these lava flows, but we suspect that it may be as young as mid-Holocene. This and other lava flows ultimately resulted from ongoing crustal extension in the region and document the erosional history of the plateau. They are part of a younger bimodal (basalt and high-silica rhyolite) phase of volcanism that accompanied modern basin-range extension.*

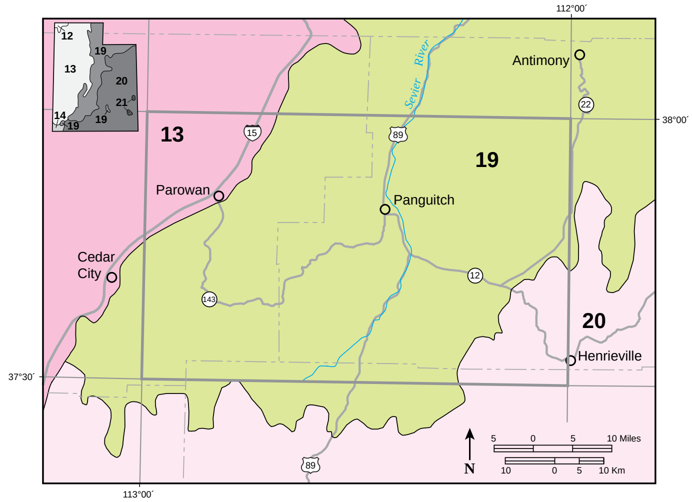


Figure 5. *Large areas having similar physiographic and climatic characteristics, and thus geographically distinct plant and animal communities, are known as ecoregions or bioregions. In southwest Utah, ecoregions (numbered on the map and inset) include the Central Basin and Range (13), Mojave Basin and Range (14), Wasatch and Uinta Mountains (19), and Colorado Plateaus (20), reflecting the great diversity of plant and animal communities found here. Modified from Ecoregions Map of Utah, available at the U.S. Geological Survey website www.usgs.gov.*

derstand the tectonic and geomorphic evolution of the present landscape. We also provide a summary of the major geologic structures in the map area, including Sevier-age thrust faults, major basin-range normal faults, and the enigmatic Paunsaugunt thrust faults. 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 in part compiled (figure 6), and in a number of popular accounts of the geology of southwest Utah and its parks, including those of DeCourten (1994), Hamblin (2004), Hintze (2005), Orndorff and others (2006), Hintze and Kowallis (2009), Baer and Steed (2010, 2012), Biek and others (2010b), Davis and Pollock (2010), Doelling and others (2010, 2012), Hatfield and others (2010, 2012), and Pollock and Davis (2012). 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.

REGIONAL GEOLOGIC SETTING

Parts of four north-trending, regional-scale tectonic features dominate the geology of southwest Utah (figure 7): (1) the leading edge of the Sevier fold-thrust belt, (2) the transition zone between the 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. This boundary resulted from Neoproterozoic rifting of the supercontinent Rodinia (see, for example, Karlstrom and others, 1999) and is manifest today as the boundary between the Basin and Range and Colorado Plateau–Middle Rocky Mountains physiographic provinces. Collectively, these four regional-scale features help us to understand the distribution and structure of the rocks that make this part of the state so interesting.

The Sevier orogeny was one of the most important events in Utah's geologic history. It was part of the larger Cordilleran orogeny, a mountain-building event that produced a zone of deformation that extends from southern Mexico to Alaska along the western margin of North America and that is widely considered the archetypal example of an ancient mountain belt formed between converging oceanic and continental plates (see, for example, DeCelles, 2004; Dickinson, 2009). The Sevier orogeny was a direct result of the subduction of the Farallon tectonic plate (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, DeCelles, 2004; Willis, 1999, 2000). In the Panguitch 30′ x 60′ quadrangle, such deformation is dramatically displayed in the Red Hills with the Iron Springs thrust faults (the large, frontal thrust in this part of the orogenic belt) and overturned Middle Jurassic and Upper Cretaceous strata, and in Cedar Canyon where the Middle Jurassic Carmel Formation is disharmonically folded below overlying Upper Cretaceous strata. Subtle latest Cretaceous compressional deformation, however, extends eastward to the Paunsaugunt fault, which exhibits an early history of pre-basin-range, down-to-the-east displacement. Late Cretaceous sediments shed from the southern Utah part of the Sevier thrust belt display dramatic west-to-east changes across the map area, in part recording the encroachment, maximum extent, and retreat of the Western Interior Seaway.

The area of what is now the Colorado Plateau and Basin and Range physiographic provinces was uplifted several thousand feet during early Tertiary time (see, for example, references in Davis and Bump, 2009, p. 118). 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; Dickinson, 2006). Today, the Earth's crust is still being pulled apart west-to-east across the Basin and Range. Compared to the Colorado Plateau—a broad area underlain by mostly gently warped sedimentary rocks only locally disrupted by broad uplifts and igneous intrusions—the Basin and Range crust is

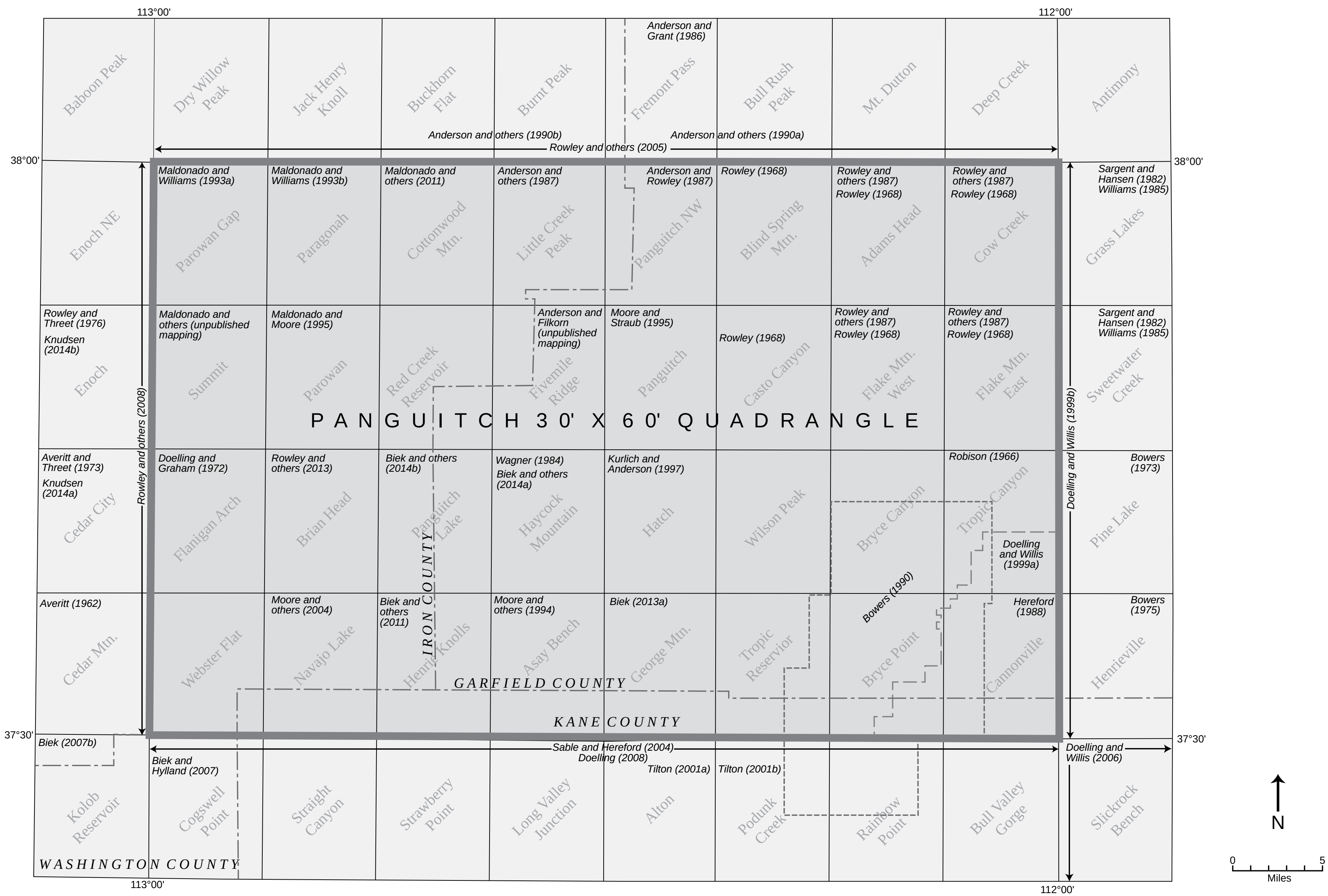


Figure 6. *Index to U.S. Geological Survey 7.5' quadrangles and principal sources of geologic mapping used in this compilation.*

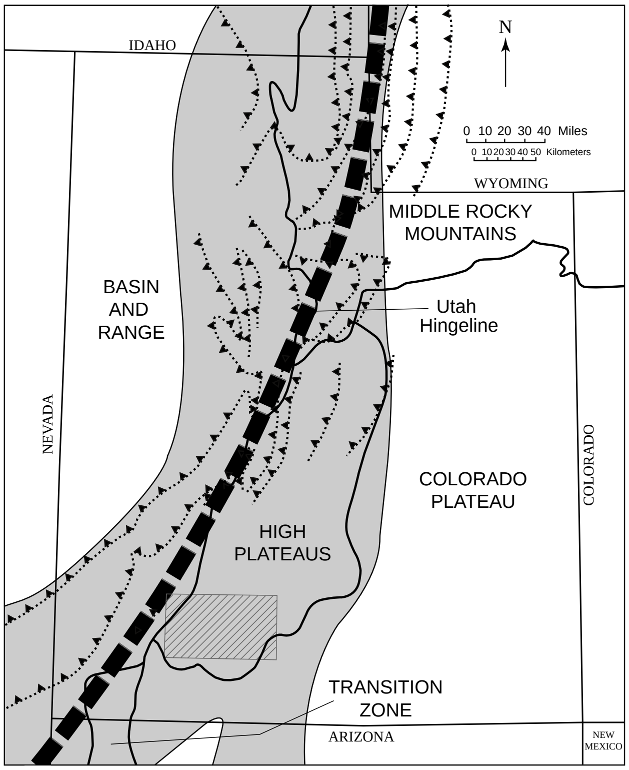


Figure 7. *Regional-scale tectonic features of Utah. The Utah hingeline (also known as the Wasatch hingeline), which originated as a long-lived boundary between a stable continental shelf and a subsiding marine basin, now controls the location of three other regional-scale features: the Sevier orogenic 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. Box shows location of Panguitch 30′ x 60′ quadrangle.*

thinner, is broken into north-trending fault blocks, and has experienced widespread igneous activity. Utah's High Plateaus are a structural and stratigraphic transition zone between these two great physiographic provinces that is broken by several north-trending faults that step down from the Colorado Plateau to the Basin and Range. The boundaries of this transition zone are well defined in the Panguitch 30′ x 60′ quadrangle (bounded on the west by the Parowan-Paragonah fault zone and on the east by the Paunsaugunt fault), but southward the transition zone loses its character and in southwesternmost Utah most geologists consider it to encompass the area between the Gunlock-Reef Reservoir-Grand Wash fault system on the west and the Hurricane fault zone on the east (Biek and others, 2009). Regardless, all these faults comprise the southern part of the Intermountain seismic belt in Utah, a north-trending zone of pronounced seismicity that extends from northern Arizona to western Montana (Smith and Arabasz, 1991). These major faults not only form the boundaries of the transition zone, but also greatly influenced the late Tertiary-Quaternary landscape evolution of southwest Utah.

STRATIGRAPHY

The Panguitch 30′ x 60′ quadrangle contains exposed rocks as old as the Early Jurassic Navajo Sandstone near Parowan Gap at the west edge of the map area, and Middle Jurassic Carmel Formation near Parowan Gap, in Cedar Canyon, and south of Tropic in the southeast corner of the map area, but the most widespread exposures of sedimentary strata belong to several Late Cretaceous formations and the overlying early Tertiary Claron Formation. Late Eocene to Miocene volcanic rocks form much of the northern Red Hills, central and northern Markagunt Plateau, and the Sevier Plateau, whereas latest Tertiary and Quaternary basaltic lava flows are scattered across the western two-thirds of the map area. What follows is an overview of these rocks, emphasizing Cretaceous and Tertiary strata. Additional information on these and other rocks and sediments exposed in the quadrangle are in the map unit descriptions (see appendix).

The stage on which the geology of this part of southwest Utah is set began in earnest in latest Neoproterozoic time, about 575 million years ago, when the southern part of the supercontinent Rodinia began to break up (rifting apparently began about 750 million years ago in what is now the northern Cordillera) (Dickinson, 2006, 2009) (figure 8). Laurentia, the Precambrian core of North America, split away from other continental blocks, creating the proto-Pacific Ocean and leaving the western edge of North America, then facing north in central Nevada, as a passive margin (passive margins form on the trailing edge of Earth's tectonic plates) (figure 9). During the early Paleozoic Era, what is now Utah was located near the equator in this passive margin setting—along a coastline probably geologically similar to that of the southeastern United States today. The rifting created a broad north-trending zone of attenuated continental crust that reached inland as far as central Utah. Today, this north-trending boundary between weaker, thinned crust to the west and thicker, more stable crust to the east is known as the Utah hingeline. Subsidence along the passive western margin of North America created accommodation space in which to preserve a great thickness of shallow-marine strata, which are wonderfully exposed in the Hurricane Cliffs and Beaver Dam Mountains of southwest Utah.

During latest Devonian to earliest Mississippian time, intraoceanic volcanic arc complexes accreted to the southwestern margin of North America as part of the Antler orogeny (DeCelles, 2004; Dickinson, 2006, 2009). Although still in an equatorial shallow-marine setting, Utah now felt the distant effects of this mountain-building episode. In southwest Utah, widespread erosion associated with the Antler orogeny removed hundreds of feet of marine rocks of Ordovician and Silurian age before deposition resumed in the Late Devonian (see, for example, Crafford, 2008). Several thousand feet of mostly shallow-marine carbonates of Mississippian, Pennsylvanian, and Permian age in southwest Utah record resumption of sedimentation following the Antler orogeny.

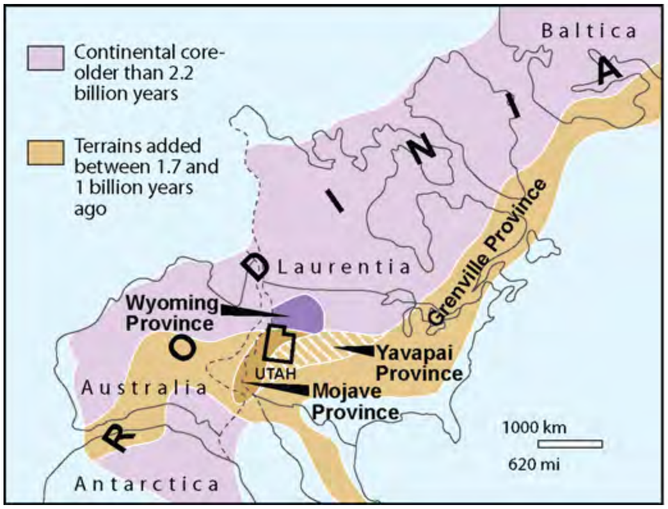


Figure 8. *The Rodinian supercontinent formed between 1.7 and 1.0 billion years ago as island-arc terranes accreted to the southern margin of older, Archean cratonic rocks. Rodinia included Laurentia (North America), Australia, Antarctica, Greenland, and Baltica (northern Europe). About 750 million years 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).*

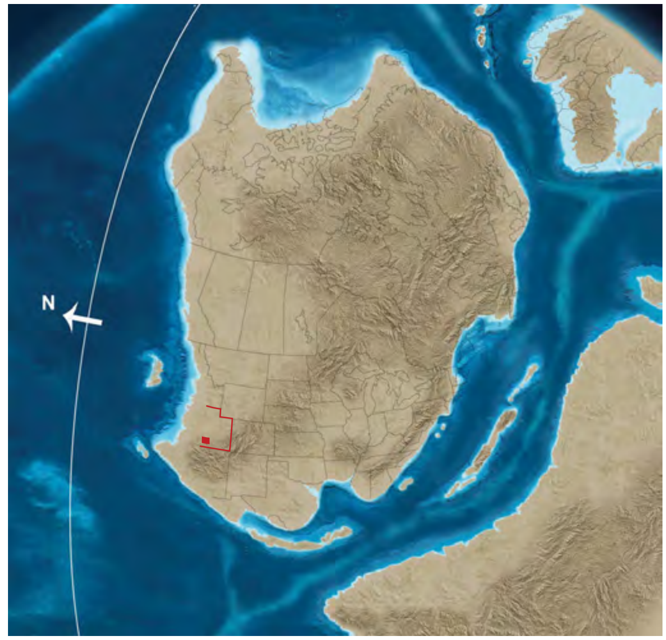


Figure 9. *Schematic reconstruction of late Neoproterozoic paleogeography of the southwest U.S. (about 550 million years ago). Note the active rifts (light blue) that separate North America from adjacent continental tectonic plates (Baltica at upper right and Gondwana at lower right), marking the breakup of the Rodinian supercontinent. Over the following 100 million years, the western margin of North America (which was actually the northern margin at the end of the Neoproterozoic) was eroded down to a low-relief coastal plain, setting the stage for accumulation of a great thickness of shallow-marine Paleozoic strata. The Utah hingeline (also known as the Cordilleran or Wasatch line; compare to figure 7) separates thicker strata that accumulated in a subsiding basin from equivalent but thinner strata that accumulated on a more stable shelf platform. Box shows location of Panguitch 30′ x 60′ quadrangle. Modified from color image available on Northern Arizona University Emeritus Geology Professor Ronald Blakey's website http://www.cpgeosystems.com/paleomaps.html, accessed June 18, 2013. Used by permission.*

The long period of mostly shallow-marine sedimentation came to an end in Late Permian time as additional oceanic island arcs continued to collide and accrete to the western Pangean margin (Pangea was the latest of Earth's several supercontinents) in what is now California and southwest Arizona during the Late Permian to earliest Triassic Sonoma orogeny (see, for example, Speed, 1977). No rocks of Late Permian to earliest Triassic age are present in southwest Utah due to erosion and non-deposition associated with this orogeny. Dramatic effects of erosion are present southwest of St. George where deep, conglomerate-filled channels are cut into Lower Permian shallow-marine limestone of the Kaibab Formation (Hayden, 2011). The Permian-Triassic 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). Accretion of island arcs and subduction of oceanic crust deformed the western continental margin, forming a large chain of mountains and volcanoes, transforming the passive margin to a volcanic arc-bounded interior basin drained by north- and northwest-flowing rivers. Volcanic ash from the arc drifted east to the interior basin where it was deposited as the colorful layers of the Chinle Formation, a unit that is well known as the host for spectacular petrified wood at Petrified Forest National Park in Arizona. Chinle strata differ dramatically from the underlying red siltstone, gray limestone, and white gypsum of the Moenkopi Formation, which records a series of incursions and retreats of a shallow ocean across a gently sloping continental shelf, where small changes in sea level translated into shoreline changes of many miles (Blakey and others, 1993; Dubiel, 1994). These two formations are separated by a major unconformity, the TR-3 regional unconformity of Pipiringos and O'Sullivan (1978). Moenkopi strata below the unconformity record the last passive-margin shallow-marine sedimentation at the western margin of Pangea, whereas Chinle strata above the unconformity document a remarkably diverse ecosystem of an inland basin. Chinle and Moenkopi strata are exposed immediately west of the map area in Cedar Canyon, on the east limb of the Kanarra anticline (Rowley and others, 2008; Knudsen, 2014a).

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). With the discovery of exceptionally well-preserved dinosaur tracks and associated fossils at the St. George Dinosaur Discovery Site at Johnson Farm in 2000, 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). Moenave strata are unconformably overlain by the prominent pinkish-brown ledge of the Springdale Sandstone Member of the Kayenta Formation, left behind as streams partly eroded Moenave strata; the Springdale itself is overlain by siltstone and sandstone that weathers to steep, deep reddish-brown slopes below the Navajo Sandstone. In Cedar Canyon, immediately west of the map area, an eolian (wind-blown) sandstone known as the Shurtz Tongue of the Navajo Sandstone forms a ledge near the middle of the Kayenta Formation (Knudsen, 2014a), other parts of which were deposited in distal river and playa environments. Biek and others (2009) and Hintze and Kowallis (2009) provide more information on Jurassic and older strata in the St. George 30′ x 60′ quadrangle, immediately southwest of this map area.

Jurassic
(201 to 145 million years ago)

As movement of the Earth's tectonic plates continued to carry North America northward, and as mountains formed in what is now Nevada and California created a rain shadow, Utah entered the arid, low-latitude, trade-winds climatic belt. Vast dune fields similar to those of 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; Milligan, 2012) (figure 10). The Shurtz Tongue of the Navajo Sandstone, exposed immediately west of the map area in Cedar Canyon, represents a brief incursion of the dune field onto the Kayenta playas, and in the St. George area, the gradual increase in wind-blown sand is clearly recorded in strata that enclose the gradational contact between the Kayenta Formation and Navajo Sandstone (Biek and others, 2009). During this 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. 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 interdunal playas and lakes ("oases").

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 fractures or joints, weathers into bold, rounded cliffs. However, only part of the Navajo is exposed in the map area at Parowan Gap, where it is fault-bounded, thoroughly fractured, and tilted steeply to the east in the upper plate of the Iron Springs thrust faults. 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 then blown southward and incorporated into the Navajo-Nugget-Aztec dune field (Dickinson and Gehrels, 2003, 2009a, 2009b; 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, described below.

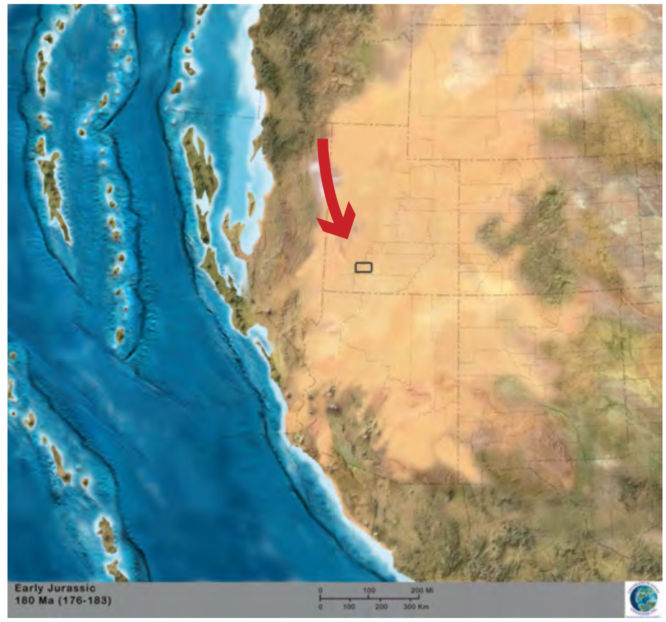


Figure 10. *Schematic reconstruction of Early Jurassic paleogeography of the southwest U.S. (about 180 million years ago). By this time, what is now southwest Utah entered the dry, low-latitude, trade-winds climatic belt, where the Navajo Sandstone, part of the world's largest coastal and inland dune field (Milligan, 2012), was deposited by winds from the north (red arrow). Much of the sand in the Navajo was originally derived from the ancestral Appalachian Mountains (Dickinson and Gehrels, 2003, 2009a, 2009b; Rahl and others, 2003). Box shows location of Panguitch 30′ x 60′ quadrangle. From Northern Arizona University Emeritus Geology Professor Ronald Blakey's website http://www.cpgeosystems.com/paleomaps.html, accessed June 18, 2013. Used by permission.*

The west end of Parowan Gap, listed on the National Register of Historic Places in 1974, has long been significant to the Paiute and Hopi Tribes and is widely recognized for the number and quality of its petroglyphs carved into Navajo Sandstone. More than 1500 geometric designs and images of snakes, mountain sheep, lizards, and human figures adorn over 90 separate rock panels at the Gap (Bureau of Land Management, undated). Most of the petrogylphs are thought to have been created by the Fremont culture about 3000 years ago, but some may be almost 5000 years old. The interpretation of petroglyphs is always controversial and the meaning of many petroglyphs is lost to time, doubtless the reason there are so many competing interpretations of what is seen at the Gap. Interpretations of petroglyphs as sign language recall actual events of the region's earliest inhabitants and as maps showing routes of travel, as well as images related to solar and lunar calendars suggested by those interested in the possible intersection of archaeology and astronomy. Of all the conflicting interpretations of geometric petroglyphs at the Gap, the archaeoastronomy connection seems the most compelling; even today, the Gap draws scores of people to witness important astronomical events. However, that a single petroglyph can be interpreted in multiple ways, including the Zipper Gylph (figure 11), one of the most recognizable images at the Gap, illustrates the difficulty in accurately deciphering images for which we have no "dictionary" and for which we have precious little understanding. As most archaeologists admit, the interpretation of many petroglyphs is ambiguous; possibly there is no one right interpretation.

But one thing we do know is that there is a short tridactyl trackway (Milner and Spears, 2007), made about 190 million years ago, when a small three-toed dinosaur walked across the lower part of a sand dune in the largest sand desert in Earth history (Milligan, 2012). The intersection of that long-gone creature and its long-gone sand desert environment, overlain with a multitude of petroglyphs, makes the west end of Parowan Gap a profoundly special place. The area is unusual too in that today a wind gap, now occupied by an underfit stream that only in exceptional years drains Little Salt Lake playa, is present where Parowan Gap pierces the Navajo Sandstone horst (figure 12). How could such a tiny stream that seldom carries water carve such an impressively narrow, deep gap in the fairly resistant Navajo Sandstone? And why wouldn't the stream have been deflected to the north to erode through less resistant basin-fill sediments and volcanic rocks? The short answer is that this stream did not carve the gap. This is in essence a fossil valley, a classic example of an antecedent stream course. The stream that carved Parowan Gap must have existed prior to uplift of the horst block and the development of Parowan Valley; it may be the ancestral, western part of the stream that now drains Parowan Canyon.

Figure 11. *The Zipper Glyph panel, one of many panels of petrogylphs carved into Lower Jurassic Navajo Sandstone at The Narrows at the west end of Parowan Gap. Most of the petroglyphs here are thought to be about 3000 years old (BLM, undated). The Narrows remains a significant historical and spiritual site for the Paiute and Hopi Tribes, and even today draws scores of people to witness important astronomical events.*

By Middle Jurassic time, deformation associated with subduction of the Farallon oceanic plate along the western edge of North America spread eastward into western Utah (DeCelles, 2004). These eastward-directed compressional forces created the Sevier orogenic belt and associated broad deformation zone, which consists of, from west to east, a thrust belt, foredeep basin, forebulge, and back-bulge basin (DeCelles, 2004; Willis, 1999, 2000) (figure 13). 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. Recent research in western Utah and eastern Nevada suggests the presence of an older, Middle to Late Jurassic episode of deformation that Thorman (2011) called the Elko orogeny, and it too is envisioned to have consisted of the classic four-part zone of deformation (figure 13). We now recognize the first effects of the Cordilleran orogeny (related to the Elko pulse of deformation) in southwestern Utah as recorded in the gently undulating but remarkably flat surface at the top of the Navajo Sandstone, the J-1 unconformity (Pipiringos and O'Sullivan, 1978). The unconformity resulted from a poorly understood combination of continued development of the Uinta Arch and early growth of the Twin Creek basin, which apparently cut off the erg's sand supply. Ultimately, rising sea levels resulted in the southward incursion of a warm, shallow inland sea in the Utah-Idaho trough (a foredeep basin of the Elko orogeny), resulting in the deposition of reddish-brown siltstone, mudstone, and gypsum of the Manganese Wash and Sinawava Members of the Temple Cap Formation in coastal-sabkha and tidal-flat environments that occupied paleolow areas eroded into the top of the Navajo Sandstone; the intervening White Throne Member heralds a brief return to sand desert conditions, but is not present west of the transition zone (Blakey, 1994; Peterson, 1994; Sprinkel and others, 2011a; Doelling and others, 2013).

Figure 12. *The Narrows at the west end of Parowan Gap, which pierces a resistant fault-bounded block of Navajo Sandstone, thus creating a puzzling wind gap now occupied by a normally dry, underfit stream. Parowan Gap is in essence a fossil valley, a classic example of an antecedent stream course. The stream that carved Parowan Gap must have existed prior to development of Parowan Valley; it may be the ancestral, western part of the stream that now drains Parowan Canyon.*

The overlying Carmel Formation was deposited in a shallow inland sea of this foredeep basin (figure 14) (Sprinkel and others, 2011a). 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 this map area, parts of the formation are exposed in structurally complicated outcrops at Parowan Gap and in Cedar Canyon, and the upper two members are present south of Cannonville. Most geologists used to think that a significant erosional episode, the J-2 unconformity of Pipiringos and O'Sullivan (1978), separated Carmel from the underlying Temple Cap strata. Although incursion of the Utah-Idaho seaway marks a significant change in depositional environments between the two formations, new radiometric and fossil age control of this interval across the Colorado Plateau and adjacent areas shows that it does not mask a significant gap in the rock record (Kowallis and others, 2011; Sprinkel and others, 2011a). The Thousand Pockets Tongue of the Carmel Formation (Doelling and others, 2013) is present along the Paria River immedi-

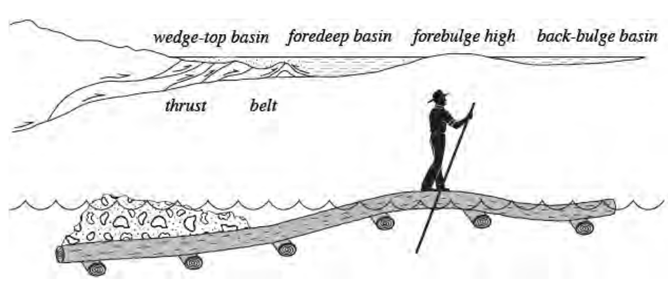


Figure 13. *Typical parts of a thrust system. The thickened, eastward-moving, leading-edge thrust wedge on the left overloads the Earth's crust, which flexes in response, similar to loading rock on a wooden raft floating on water. In Utah, the entire thrust system migrated eastward over time during the 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).*

ately south of the southeast corner of the map area, pinching out westward near Skutumpah Creek (Sable and Hereford, 2004; Doelling, 2008), but it is doubtless present in the subsurface in the Cannonville area. It overlies Crystal Creek strata and forms a prominent, massively cross-bedded sandstone ledge as much as 100 feet (30 m) thick.

Carmel strata record changing conditions within the foredeep basin, beginning with maximum incursion of the seaway (gray muddy limestone of the Co-op Creek Limestone Member), seaway retreat to coastal sabkha and tidal-flat environments (reddish-brown mudstone, siltstone, and sandstone of the gypsiferous Crystal Creek Member), a period of marine incursions and retreats (thick gypsum beds, reddish-brown mudstone, and gray limestone of the Paria River Member), and final retreat of the seaway to expose broad sandy mud flats (reddish-brown siltstone and sandstone of the Winsor Member). Thin beds of altered volcanic ash, derived from the Cordilleran magmatic arc in what is now California and western Nevada, are present throughout the formation and allow remarkable age control and correlation throughout the basin (Sprinkel and others, 2011a). Most carbonate beds in the formation are muddy and properly classified as micritic limestone or even calcareous shale, reflecting significant clay input into the shallow, epicontinental seaway. Locally, however, thin, relatively pure limestone beds are present, and these commonly contain *Isocrinus* sp. crinoid columnals, pelecypods, and gastropods (figure 15). Carmel gypsum, particularly mottled pink or light-gray alabaster gypsum of the Paria River Member, is locally used for carving.

The broad sandy mud flat in which Winsor strata accumulated eventually was overwhelmed by reddish-brown silty sand now assigned to the Gunsight Butte Member of the Entrada Sandstone. The three-member Entrada Sandstone records deposition in tidal-flat, sabkha, and coastal dune environments (Peterson, 1988, 1994). It is cut out by a major unconformity just west of the Paunsaugunt fault zone (Sable and Hereford, 2004; Doelling, 2008), but is widely exposed across the Colorado Plateau where it is a prominent unit at several national and state parks, including Kodachrome Basin State Park in and adjacent to the southeast corner of the map area. At and near Kodachrome Basin, the Gunsight Butte Member (and underlying Winsor Member of the Carmel Formation) are renowned for their sedimentary breccia pipes, which are thought to have formed as fluid escape structures from overpressured, underlying Carmel strata in Middle Jurassic time (Baer and Steed, 2010). Several such breccia pipes are prominently displayed

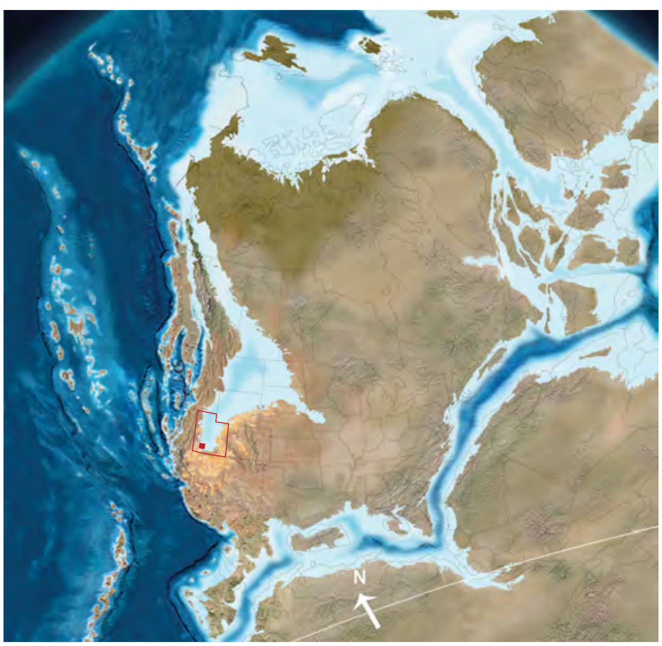


Figure 14. *Schematic reconstruction of North American Middle Jurassic paleogeography (about 170 million years ago). What is now southwest Utah was near the south end of a shallow inland sea, part of the foredeep basin of the Elko orogeny. Continued encroachment of the orogenic belt (the forebulge high) into southwest Utah led to erosion or non-deposition of Late Jurassic and Early Cretaceous strata. Box shows location of Panguitch 30′ x 60′ quadrangle. From Northern Arizona University Emeritus Geology Professor Ronald Blakey's website http://www.cpgeosystems.com/paleomaps.html, accessed June 18, 2013. Used by permission.*

Figure 15. *Star-shaped fossils, the disarticulated stems of shallow-marine animals commonly known as sea lilies (*Isocrinus *sp., formerly and more descriptively known as* Pentacrinus *sp.), are common at some horizons of the Co-op Creek Limestone.*

at Shepard Point (figure 16). A conspicuous banded slope of reddish-brown and gray silty sandstone (Cannonville Member) overlies Gunsight Butte strata, and is in turn overlain by light-yellow-brown, commonly cross-bedded sandstone (Escalante Member) (figure 17). Thompson and Stokes (1970) defined the Henrieville Sandstone, the type section of which is immediately east of the map area near Henrieville, which they interpreted as unconformably overlying Entrada strata. Peterson (1988), however, suggested that the Henrieville is simply a bleached upper part of the Escalante Member of the Entrada Sandstone, with which we concur.

Cretaceous
(145 to 66 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 latest Early Cretaceous time, relatively thin but widespread Lower Cretaceous gravels formed a broad alluvial plain over most of the Western Interior. As noted in the classic paper by Heller and Paola (1989), the distribution of these gravels is believed to reflect regional thermal uplift associated with Jurassic-Cretaceous magmatism in the hinterland, immediately prior to onset of thrusting in the Sevier orogenic belt and development of a sediment-trapping foredeep basin. Detrital zircon studies of Hunt and others (2011b) showed that on the Markagunt Plateau the clasts were largely derived

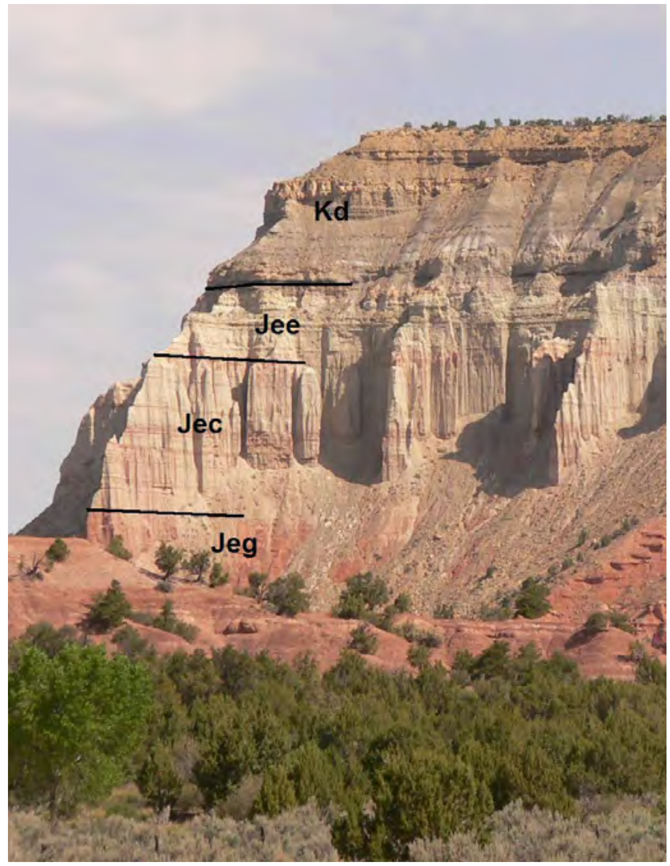


Figure 17. *The south end of Bulldog Bench immediately southwest of Cannonville, which exposes a complete section of the Entrada Formation (Jee, Escalante Member; Jec, Cannonville Member; Jeg, Gunsight Butte Member). Uppermost Winsor strata form the low hill at left; a thin coal bed marks the base of the Dakota Formation (Kd).*

Figure 16. *Sedimentary breccia pipe in Winsor Member strata at Shepard Point in the southeast corner of the Panguitch 30' x 60' quadrangle. This and other pipes in the Gunsight Butte Member of the Entrada Formation and Winsor Member of the Carmel Formation are thought to have formed as fluid escape structures from loading of overpressured, underlying Carmel strata in Middle Jurassic time (Baer and Steed, 2010). Inset shows chaotic, unbedded nature of pipe and disturbed bedding adjacent to pipe; note overlying gravel-capped pediment deposits. View is towards the north; road to Kodachrome Basin State Park is in the foreground.*

from Ordovician to Devonian strata in the Sevier thrust belt, and they suggested correlation with the Short Canyon Conglomerate of central Utah (Doelling and Kuehne, 2013). Interestingly, the Markagunt Plateau conglomerate exhibits a detrital zircon signature different from that of the conglomerate interval near Gunlock (which is identical to the older Buckhorn Conglomerate). This is puzzling given the fact that, except for a gap where these strata are eroded over the crest of the Kanarra anticline, they can be traced continuously from the eastern Kolob Plateau, across the southern flank of the Pine Valley Mountains, and westward beyond Gunlock (Biek and others, 2009). The base of these two conglomerate intervals marks the Cretaceous unconformity, which progressively cuts out much of the Carmel Formation as it is traced westward. Apparently the deeper incision near Gunlock reflects preservation of the oldest part of the Cedar Mountain Formation (the Buckhorn Conglomerate) whereas the conglomerate of the Markagunt Plateau reflects the younger Short Canyon interval; doubtless they reflect the variability of sources that must exist in a complicated thrust belt. In this map area, pebble conglomerate of the Cedar Mountain Formation overlies badland slopes of reddish-brown siltstone and sandstone of the Winsor Member whose uppermost beds are bleached white under the Cretaceous unconformity.

The pebble conglomerate (and overlying, pastel-colored smectitic mudstone) were previously known as the lower Upper Cretaceous (Cenomanian) Dakota Formation. However, on the basis of new radiometric ages and distinctive lithology, we assign these beds to the 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). Still, assignment of this interval, which is well developed in Cedar Canyon (figures 18 and 19), remains controversial among some stratigraphers. We do not recognize the interval in the southeast part of the map area; there, a relatively thin Dakota Formation unconformably overlies white and pale-yellowish-brown silty sandstone of the Entrada Formation.

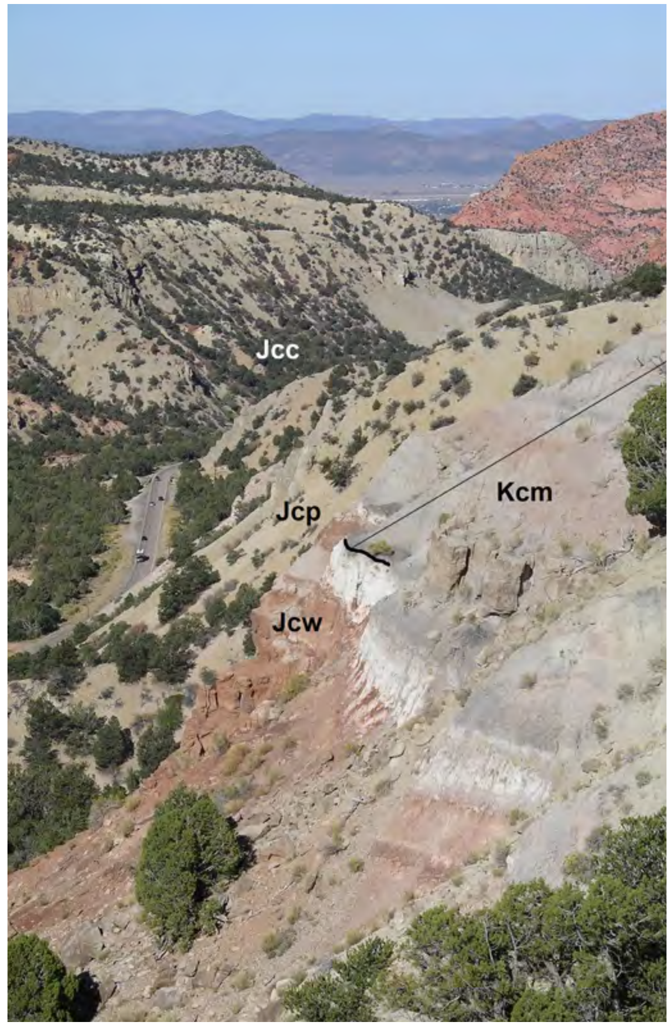


Figure 18. *Cedar Mountain Formation (Kcm) and Carmel Formation in Cedar Canyon near the west edge of the map area. Base of Cedar Mountain Formation is marked by a thin pebble conglomerate and overlying dark-gray bentonitic ash; note thin, lenticular channel sandstone near base of Cedar Mountain strata and bleached white upper part of Winsor Member of the Carmel Formation (Jcw). Co-op Creek Limestone Member of the Carmel Formation (Jcc) is exposed at road level along State Highway 14; Crystal Creek strata are hidden from view; Paria River Member (Jcp) on next ridge underlies Winsor Member. Outcrop is in the SW1/4NE1/4NW1/4 section 21, T. 36 S., R. 10 W.; view is towards the west.*

Encroachment of the Sevier orogenic belt in early Late Cretaceous time created a foredeep basin in which thousands of feet of sediment shed from the thrust belt accumulated. These Upper Cretaceous strata undergo significant west-to-east and north-to-south facies changes on the Markagunt and Paunsaugunt Plateaus, thus presenting significant challenges to correlation and mapping as described by Tilton (1991), Eaton and others (2001), Moore and Straub (2001), Moore and others (2004), and Rowley and others (2013). The lower part of this Upper Cretaceous section is divided into the alluvial-plain and brackish estuarine and lagoonal Dakota Formation; the westward-thinning and intertonguing wedge of marine Tropic Shale; and the nearshore, brackish-water, and coastal-plain Straight Cliffs Formation (see, for example, am Ende, 1991; Eaton and Nations, 1991; Eaton and others, 2001; Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003). Collectively, this sedimentary package was deposited during the Greenhorn Marine Cycle, a large-scale sea-level rise and fall recognized worldwide and that here corresponds to the maximum transgression of the Western Interior Seaway (figure 20) (see, for example, McGookey, 1972; Kauffman, 1984). This package of rock is overlain by Upper Cretaceous fluvial and floodplain strata of the Wahweap Formation, and, locally on the Paunsaugunt Plateau, by newly discovered remnants of the Kaiparowits Formation.

Figure 19. *Cedar Mountain Formation (Kcm) and enclosing units in Cedar Canyon, in the SW1/4SE1/4SW1/4 section 16, T. 36 S., R. 10 W. Swelling mudstone of light-gray, reddish-brown, and purplish hues contrasts sharply with yellowish-brown and olive-gray mudstone of overlying Dakota Formation (Kd). About 40 feet (12 m) above the base of the Dakota Formation, a ledge-forming 20-foot-thick (6 m) pebbly sandstone and conglomerate with rounded quartzite and black chert clasts is present. Jcw = Winsor Member of Carmel Formation.*

Complicating this picture is the now-superceded three-member Grand Castle Formation of the western Markagunt Plateau, originally inferred to be Paleocene in age. However, the lower two members are now known to be Late Cretaceous; the upper member is undated but is likely Late Cretaceous as described below. Our mapping shows that the lower conglomerate member of the Grand Castle Formation is in fact the Drip Tank Member of the Straight Cliffs Formation (as suggested by Eaton and others, 2001; Moore and Straub, 2001; Lawton and others, 2003; and Eaton, 2006), and that the middle sandstone member of the Grand Castle Formation is in fact the capping sandstone member of the Wahweap Formation (as suggested by Pollock, 1999; and Lawton and others, 2003). Mapping coarse alluvial strata associated with major sequence boundaries has been the key to working out these lithostratigraphic correlations.

Dakota Formation

Although the Dakota Formation is not correlative with the type Dakota in Nebraska, the term is used loosely in Utah for deposits of an overall transgressive sequence below the Tropic Shale, the lower part of which was deposited in floodplain and river environments, whereas the upper part represents estuarine, lagoonal, and swamp environments of a coastal plain (Gustason, 1989; Eaton and others, 2001; Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003; Titus and others, 2013). The Dakota thus records changing depositional environments landward of an encroaching sea, the western time-equivalent facies of the marine Tropic Shale. Dakota strata include ledge-forming, yellowish-brown, fine- to medium-grained sandstone and siltstone and less resistant, slope-forming, gray smectitic mudstone, thin coal beds, and rare marly beds. Two significant coal zones on the Paunsaugunt Plateau, the lower Bald Knoll zone and the upper Smirl zone (Doelling and Graham, 1972; Quick, 2010), are now being mined near Alton on the Plateau's southern flank; thinner coal beds were also mined on the west flank of the Markagunt Plateau in Cedar Canyon and on the Kolob Terrace (Doelling and Graham, 1972). Abundant invertebrate and palynomorph fossil assemblages in the Dakota indicate shallow-marine, brackish, and freshwater deposits of Cenomanian age (Nichols, 1997; Eaton and others, 2001; Eaton, 2009; Titus and others, 2013). A smectitic bed from the middle Dakota south of Tropic yielded an early Cenomanian $^{40}Ar/^{39}Ar$ age of 96.06 ± 0.30 Ma (Dyman and others, 2002).

Dakota strata thicken dramatically westward from about 80 feet (24 m) thick near Tropic to as much as 1400 feet (425 m) thick in Cedar Canyon, a result of greater foredeep basin subsidence adjacent to the thrust belt compared to farther offshore, a pattern repeated with the overlying Straight Cliffs Formation. The uncommonly good exposures of the Dakota in Cedar Canyon have yielded a wealth of stratigraphic information, allowing correlation of Dakota fluvial strata with orbital cycles of marine sedimentation in the offshore parts of the Western Interior Seaway (Gustason, 1989). Laurin and Sageman (2001a, 2001b) expanded on that work, constructing a high-resolution temporal and stratigraphic framework of middle Cretaceous marginal-marine deposits—they documented changes in shoreline position and also linked these changes to rhythmic, Milankovitch-driven deposition of marine limestone of the Western Interior Seaway.

Wet mudstones of the Dakota Formation are infamous for their slipperiness; until they dry out, dirt roads constructed on Dakota mudstones are typically impassible even by 4-wheel drive vehicles. The culprit is smectitic mudstone, which swells when wet and shrinks when dry. The mudstones are also why much of the Dakota outcrop belt on the Markagunt Plateau is covered by large landslides. The

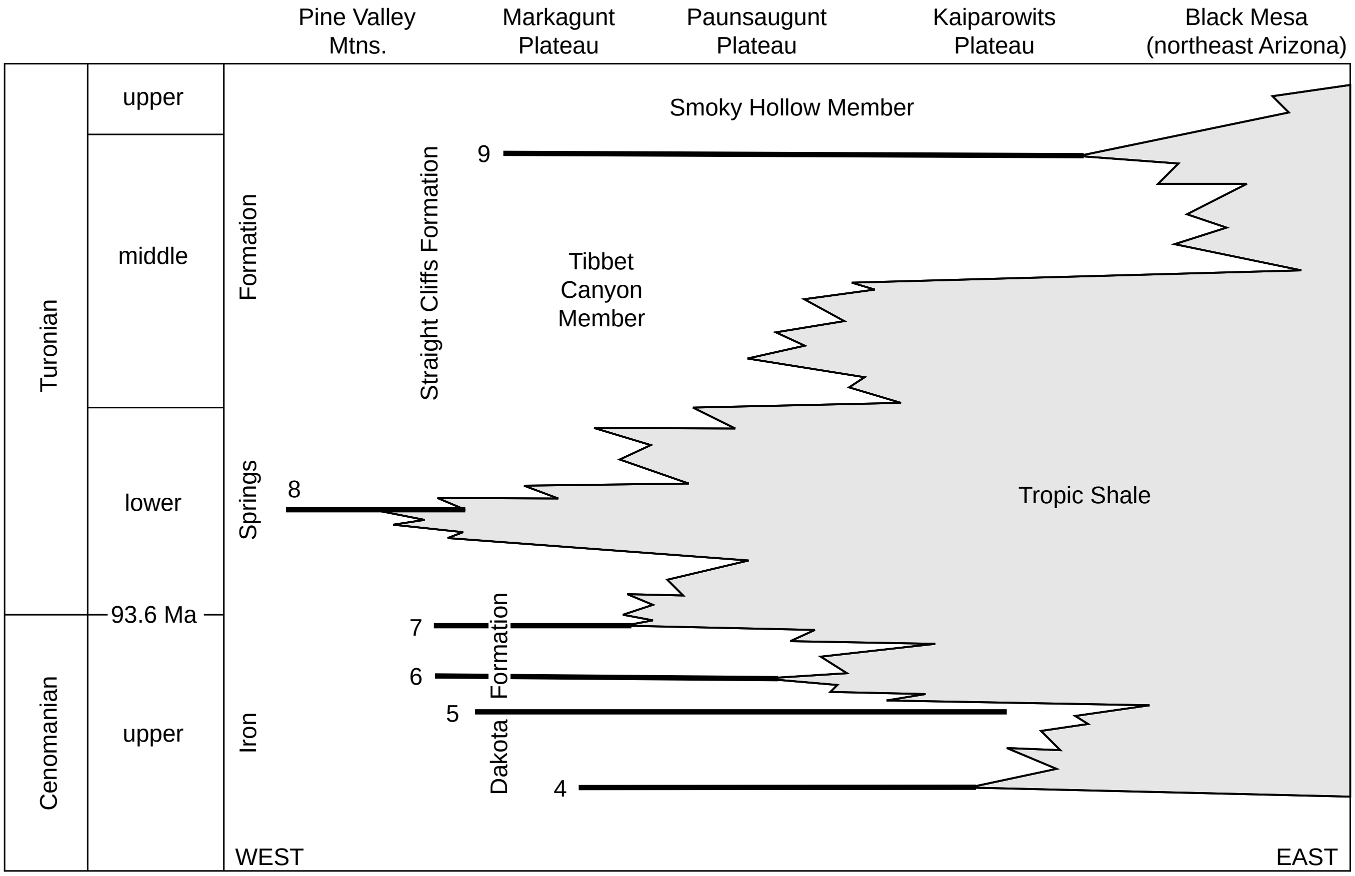


Figure 20. *Strata of the Greenhorn Cycle (in gray), showing maximum flooding surface represented by the open-marine strata of the Tropic Shale and intermediate flooding surfaces represented by coal zones (4 to 9) that accumulated in brackish, estuarine environments near the western margin of the Western Interior Seaway. Note numerous smaller cycles superimposed on the larger Greenhorn Cycle, which are due to changes in subsidence, compaction, and climate. Note also the diachronous nature of the strata, meaning that the same facies differ in age from place to place. The upper Dakota Formation is equivalent in age to the lower part of the Tropic Shale exposed farther east—that is, they are the time-correlative coastal-plain and estuarine facies of the deeper water, offshore mud deposits of the Tropic Shale. Similarly, the Tibbet Canyon Member of the Straight Cliffs Formation is older in western exposures; it represents eastward-prograding shoreline deposits that also are time-correlative with offshore Tropic muds. The Iron Springs Formation was deposited principally in braided-stream and floodplain environments of a coastal plain and is considered correlative with the Straight Cliffs Formation, Tropic Shale, and Dakota Formation. Simplified from Tibert and others (2003).*

Dakota Formation is the culprit in recurring landslide and rockfalls that periodically block Utah Highway 14 in Cedar Canyon (figure 21) (Lund and others, 2012).

Tropic Shale

The Tropic Shale, named for exposures near Tropic at the east edge of the map area, was deposited in a shallow-marine environment dominated by fine-grained clastic sediment, marking the maximum incursion of the Western Interior Seaway in Late Cretaceous (Turonian) time (Tibert and others, 2003; Titus and others, 2005). That epicontinental seaway, which divided North America into western and eastern parts, formed in response to continental-scale crustal downwarping associated with subduction at the continent's western margin (DeCelles, 2004) (figure 22). The Tropic Shale is the lateral equivalent of the Tununk Member of the Mancos Shale, the Allen Valley Shale of the Indianola Group, and the Mowry Shale, all of central and eastern Utah (see, for example, Hintze and Kowallis, 2009, and references therein). Tropic strata form a westward-thinning wedge of gray shale and silty shale that, in western exposures of the Markagunt Plateau, is a thin interval of yellowish-brown sandy mudstone, silty fine-grained sandstone, and minor shale that overlie brackish, estuarine facies of the Dakota Formation (figure 23); this westernmost wedge of the Tropic Shale is entirely Turonian in age (Eaton and others, 2001). A thin tongue of marine strata is present at Parowan Gap (in overturned Iron Springs Formation now thought to be equivalent to the Tibbet Canyon or Smoky Hollow Members of the Straight Cliffs Formation), marking the westernmost occurrence of the Western Interior Seaway.

The late Cenomanian and Turonian age of the Tropic Shale, and thus the unfolding incursion of the Western Interior Seaway, is well known from its many dated volcanic ash beds

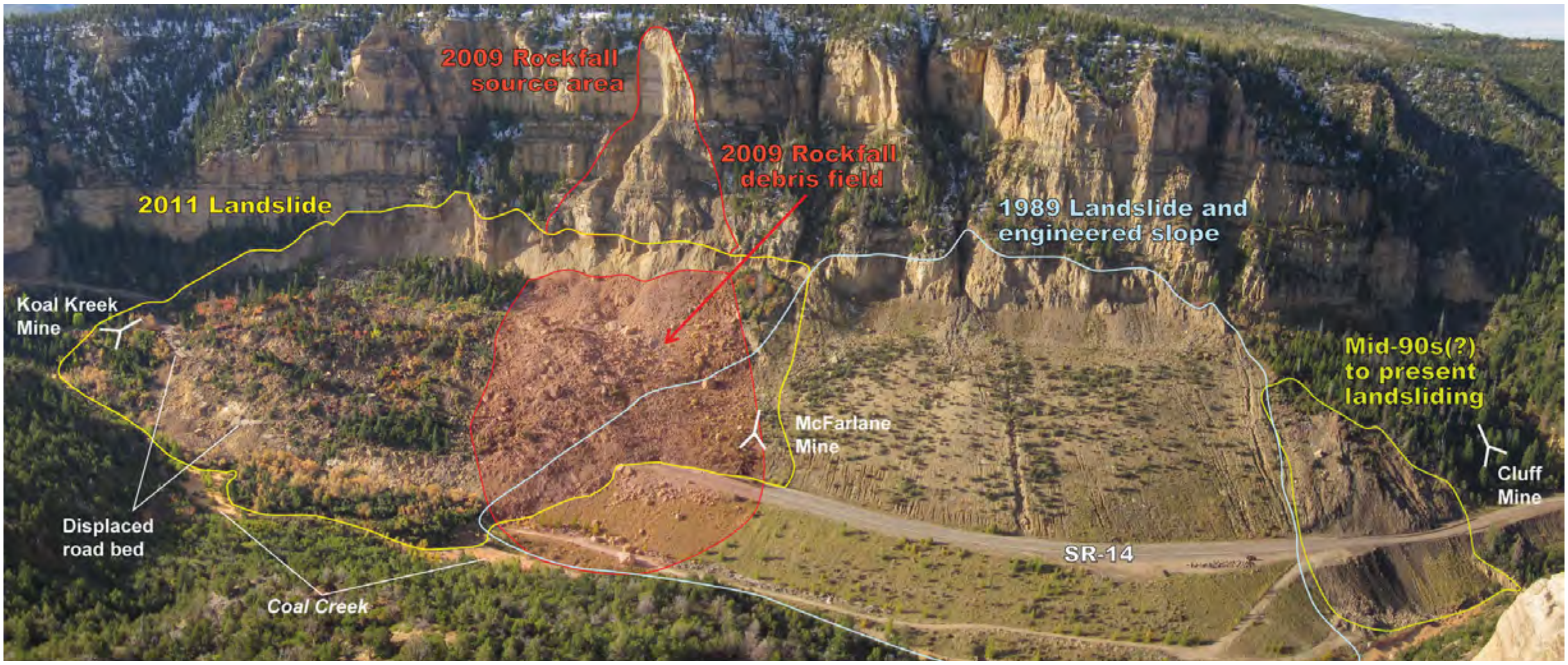


Figure 21. *Landslides and rockfalls that have affected State Highway 14, about 8 miles (13 km) east of Cedar City. The 2011 landslide occurred early in the morning of October 8th, burying and displacing a 1200-foot-long (365 m) section of the highway. The landslide occurred in the Dakota Formation, which here forms a steep, colluvium-covered slope below cliffs of the Tibbet Canyon Member of the Straight Cliffs Formation. A thin interval of Tropic Shale is at the base of the Tibbet cliffs. The approximate locations of three abandoned coal mines in the Dakota Formation are also shown. From Lund and others (2012).*

Figure 22. *Schematic reconstruction of middle Cretaceous paleogeography of North America (about 85 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. Note the pronounced foredeep basin (darker blue shading) just in front of the thrust belt. Box shows location of Panguitch 30' x 60' quadrangle. From Northern Arizona University Emeritus Geology Professor Ronald Blakey's website http://www.cpgeosystems.com/paleomaps.html, accessed June 18, 2013. Used by permission.*

and marine fossils. Ammonites, inoceramid and mytiloidid bivalves, gastropods, and oysters, as well as vertebrate fossils of sharks, fish, marine turtles, and plesiosaurs, all indicative of an open shallow-marine environment, have been recovered from Tropic strata (see, for example, Eaton and others, 2001; Titus and others, 2005; Albright and others, 2013). The ability to tie these fossils to radiometrically dated volcanic ash beds—which erupted from volcanoes at the western margin of North America, including those whose granitic roots are now exposed in the Sierra Nevada Mountains—gives geologists unprecedented precision and confidence in correlating these marine strata throughout the midcontinent region (see, for example, Titus and others, 2013).

Straight Cliffs Formation

The Straight Cliffs Formation forms an overall regressive sequence that followed the retreat of the Western Interior Seaway (see, for example, Eaton and others, 2001; Moore and Straub, 2001; Tibert and others, 2003). Peterson (1969) divided the Straight Cliffs Formation into four members in the Kaiparowits basin: in ascending order, the Tibbet Canyon, Smoky Hollow, John Henry, and Drip Tank Members. Significant research into the stratigraphy of Upper Cretaceous strata on the Markagunt and Paunsaugunt Plateaus, spurred in part by the U.S. Geological Survey's BARCO project of the 1980s and 1990s and continuing to the present, coupled with our new mapping, has enabled us to carry Kaiparowits basin nomenclature westward into the Markagunt Plateau.

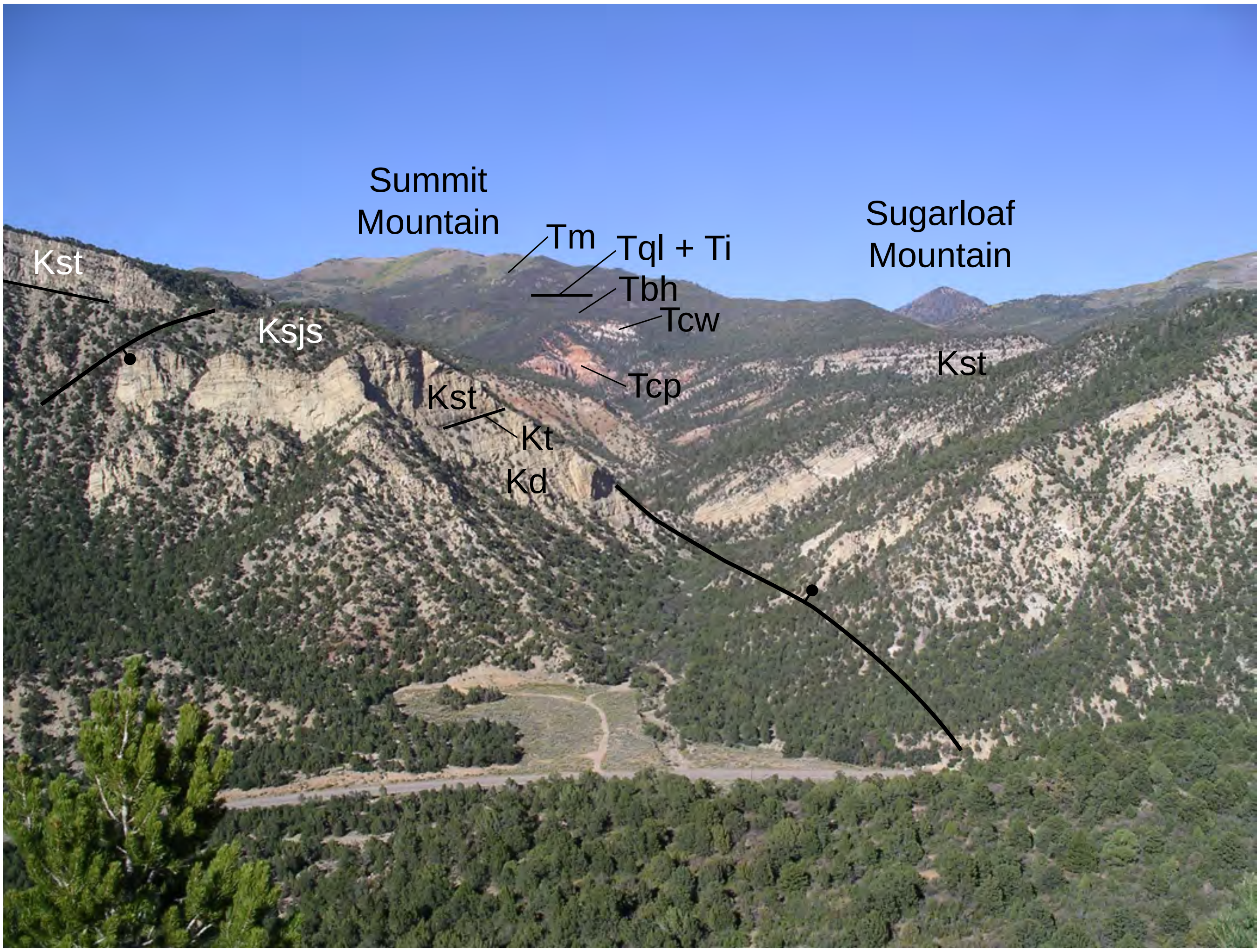


Figure 23. *View north into Maple Canyon, tributary to Cedar Canyon at the west edge of the Markagunt Plateau; State Highway 14 is in the foreground. The Tropic Shale (Kt) is represented by a thin, dark-gray mudstone and siltstone that forms a slope between ledge- and cliff-forming sandstone of the Dakota Formation (Kd) and the Tibbet Canyon Member of the Straight Cliffs Formation (Kst). The thin slope of Tropic represents the maximum incursion of the Western Interior Seaway in Late Cretaceous (early Turonian) time. Underlying Dakota strata—deposited as an overall transgressive unit of floodplain, estuarine, lagoonal, and swamp environments of a coastal plain—record the encroachment of that seaway, whereas overlying Tibbet Canyon strata were deposited in an overall progradational sequence of marginal-marine and beach environments following retreat of the Western Interior Sea.*

Several normal faults cut strata of Maple Canyon, which partly follows the south end of the Summit Mountain graben. Smoky Hollow and John Henry Members of the Straight Cliffs Formation (Ksjs); Tcp (pink member) and Tcw (white member) of the Claron Formation; regional ash-flow tuffs of the Leach Canyon Formation (Tql) and Isom Formation (Ti), which overlie the vegetated Brian Head Formation (Tbh), are unconformably overlain by the Markagunt gravity slide (Tm).

That the Western Interior Seaway withdrew slowly is reflected in the fact that its sedimentary deposits are diachronous, meaning that the same facies differ in age from place to place. In this area, marine deposition ended as the sea slowly withdrew, leaving behind shoreline deposits of the Tibbet Canyon Member of the Straight Cliffs Formation. The yellowish-brown quartzose sand of the Tibbet Canyon Member constitutes eastward prograding shoreface, beach, lagoonal, and estuarine deposits that are time-correlative with offshore Tropic muds (Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003). Tibbet Canyon strata typically form a prominent cliff and are an order-of-magnitude thicker on the west margin of the Markagunt Plateau than they are farther east, once again the result of greater subsidence along the west margin of the foredeep basin (figure 24) (Eaton, 1991; Lawton and others, 2003).

The prominent Tibbet Canyon cliff is overlain by slope-forming carbonaceous shale and few thin coal beds, several thin oyster coquina beds, and brown and gray mudstone, shale, and interbedded yellowish-brown fine-grained sandstone, all assigned to the Smoky Hollow Member of the Straight Cliffs Formation. A diverse assemblage of middle to late Turonian mollusks, benthic foraminifera, and ostracods from exposures in Cedar Canyon shows that the lower part of the member was deposited in estua-

Figure 24. *Bold cliff of Tibbet Canyon sandstone immediately south of the junction of Coal Creek and Crow Creek in Cedar Canyon; Utah Highway 14 is at the base of the cliffs.*

rine environments and that its upper part represents fluvial and floodplain deposits of a coastal plain (Eaton and others, 2001; Tibert and others, 2003). Like underlying Tibbet Canyon strata, the Smoky Hollow Member is far thicker in western exposures. A series of fluvial channel deposits of white, fine- to medium-grained sandstone and conglomeratic sandstone constitute the Calico Bed, which caps the member; this coarse sediment was deposited by braided and meandering streams that tapped both the Sevier orogenic belt and the Mogollon Highlands (Bobb, 1991). Because the Calico Bed is poorly exposed and only locally well developed on the heavily forested Markagunt Plateau, there we map the lithologically similar Smoky Hollow and overlying John Henry strata as a single unit.

John Henry strata are variegated, gray, brown, and reddish-brown mudstone and yellowish-brown, fine-grained subarkosic sandstone. Overall, the member weathers to ledgy slopes and coarsens upsection such that stacked or amalgamated sandstone beds make up most of the upper part of the unit. A volcanic ash bed 700 feet (213 m) below the top of the member (likely in the middle part of the member) in Parowan Canyon yielded a radiometric age of 83 million years, whereas an ash bed in the lower part of the member about 400 feet (120 m) above the base in Cedar Canyon yielded an age of 87 million years (Eaton and others, 1999c). The striking difference in facies and outcrop habit of Upper Cretaceous strata between Cedar and Parowan Canyons has long been noted (see, for example, Eaton and others, 2001). Parowan Canyon exposures are characterized by repetitive ledge-forming tabular sandstone beds and interbedded, slope-forming mudstone, whereas in Cedar Canyon, laterally equivalent strata are poorly exposed, stacked or amalgamated sandstone and relatively little mudstone. We are uncertain how to interpret this apparent stratigraphic variation, but note that collectively, exposures in Summit and Parowan Canyons are remarkably similar to gross characteristics of the Straight Cliffs Formation in Cedar Canyon. During the 2001 Utah Geological Association–American Association of Petroleum Geologists conference in Cedar City, much was made of the apparent difference in stratigraphic packages between Parowan Canyon and Cedar Canyon, but we now know that Parowan Canyon only exposes the upper part of the full section that is present in Cedar Canyon.

John Henry strata are unconformably overlain by the Drip Tank Member, which was deposited by east- and northeast-flowing braided rivers of a coastal plain (Tilton, 1991, 2001a, 2001b; Goldstrand and Mullett, 1997; Lawton and others, 2003); the Drip Tank's early Campanian age is constrained by the ages of enclosing well-dated strata. This coarse unit likely records erosion associated with a pulse of deformation in the still-active Sevier orogenic belt to the west where rivers tapped mountains formed on the Wah Wah and Blue Mountain thrust sheets. On the Markagunt Plateau, the Drip Tank (formerly lower conglomerate member of the Grand Castle Formation) is a massive, cliff-forming, light-gray conglomerate with well-rounded, pebble- to boulder-sized clasts of quartzite, limestone, and minor sandstone and chert. It is 135 feet (41 m) thick at the type section in First Left Hand Canyon southeast of Parowan (Goldstrand and Mullet, 1997) and of similar thickness southwest to Sugarloaf Mountain (about 3 miles [5 km] west of Brian Head). South of this area, however, the Drip Tank thins irregularly southward and locally appears as two conglomerate intervals separated by a few feet to a few tens of feet of yellowish-brown, fine-grained sandstone or variegated mudstone. At the Utah Highway 14 "S curve" in Cedar Canyon, near mile marker 13, a thin pebbly conglomerate bed occurs at the top of the roadcut exposure. Those who struggled to correlate

Upper Cretaceous strata at the west edge of the Markagunt Plateau (including Eaton and others 2001; Moore and Straub, 2001; Lawton and others, 2003; and Eaton, 2006) remarkably and presciently suggested that this thin bed is the Drip Tank. Just to the south, this bed thickens to nearly 100 feet (30 m) where it is well exposed in the scar left behind by the 2005 Black Mountain debris flow (even so, we map it as a marker bed on the southwest Markagunt Plateau due to typically poor exposure and heavy forest cover) (figure 25). The Drip Tank Member typically overlies stacked or amalgamated sandstone beds, but locally, as along Ashdown Creek, overlies variegated mudstone. On the northern Markagunt Plateau, the member locally weathers to form conically shaped hoodoos that resemble old-fashioned beehives known as bee skeps (figure 26), but south of Summit it forms a resistant ledge in the upper reaches of Summit Creek canyon and the upper reaches of Pickering Creek canyon. We traced the Drip Tank around the south edge of the Markagunt Plateau, south of the map area, to exposures just west of Highway 89, thus physically linking the lower conglomerate of Goldstrand's Grand Castle Formation with Drip Tank strata at the south end of the Paunsaugunt Plateau. The rules of stratigraphic nomenclature give precedence to the first name applied to a distinctive package of strata, which is why the name lower conglomerate member of the Grand Castle Formation is now superceded by the Drip Tank Member of the Straight Cliffs Formation.

Because the Drip Tank commonly caps a series of ledge- and cliff-forming sandstone beds, its lower contact with John Henry strata is commonly difficult to discern, particularly on the Paunsaugunt Plateau (figure 27). Thus we restrict Drip Tank strata to a white quartz arenite and pebbly conglomerate facies. Pebbly conglomerate everywhere forms the top of the member, commonly forming ledges or a small cliff, whereas the sandstone forms distinctive, manzanita-covered slopes and saddles. In western exposures, the Drip Tank appears to be conformably overlain by yellowish-brown, fine-grained sandstone and lesser interbedded, varicolored and mottled mudstone of brown, gray, reddish-brown, and pinkish hues of the Wahweap Formation. However, new research in the Kaiparowits basin to the east suggests that this interval is unconformable (Tim Lawton, Centro de Geosciencias, Universidad Nacional Autonoma de Mexico, written communication, February 25, 2014). Tilton (1991) described the Drip Tank Member as the most prominent and important marker horizon in the Upper Cretaceous section on the southern Paunsaugunt Plateau, but we find that it is remark-

Figure 25. *Scar left by the 2005 Black Mountain debris flow on the south side of Coal Creek canyon, which crossed Utah Highway 14 and flowed down Crow Creek. Drip Tank (Ksd) strata form a prominent ledge near the middle of the scar. Mostly concealed, reddish-brown strata are part of the Wahweap Formation (Kw). Ksj (John Henry Member of the Straight Cliffs Formation). Black Mountain is capped by the 850,000-year-old Black Mountain lava flow and is mantled in a thick apron of basalt talus.*

ably similar in lithology and outcrop habit to the capping sandstone member of the Wahweap Formation.

Geologists continue to refine the Upper Cretaceous stratigraphy of southwest Utah, and as mentioned above, there is some uncertainty about the nature of the contact between the Drip Tank and overlying Wahweap strata; it appears conformable in western exposures yet unconformable in eastern exposures. Had these units been defined based on their western exposures, doubtless geologists would have placed the Drip Tank as the basal member of the Wahweap Formation, not the upper member of the Straight Cliffs Formation. The distinction seems a minor one, but in reality the sequence stratigraphic framework presages a more wholistic approach to understanding past sedimentary environments; in this view, coaser grained strata signal the start of a new depositional cycle, not the end (the end of such a cycle is marked by a significant unconformity and the loss or non-deposition of strata from the end of the previous cycle). Jinnah and others (2009), for example, showed that the contact between the upper sandstone member and the capping sandstone member, both of the Wahweap Formation, represents an approximately 2-million-year-long gap in time, whereas Roberts and others (2013) showed that sedimentation was relatively continuous across the capping sandstone-lower Kaiparowits interval. But following historical precedent, we treat the Drip Tank as the upper member of the Straight Cliffs Formation; regardless, it is a key unit that enabled geologists to finally correlate Upper Cretaceous strata between the Markagunt Plateau and Kaiparowits basin.

Wahweap Formation

The Wahweap Formation is mostly fine-grained sandstone, siltstone, and mudstone deposited in braided and meandering river and floodplain environments of a coastal plain (Tilton, 1991; Pollock, 1999; Lawton and others, 2003; Jinnah and Roberts, 2011). Wahweap strata were deposited between about 81 and 77 million years ago on the basis of radiometric ages of

Figure 26. *Conically shaped outcrops of the Drip Tank Member of the Straight Cliffs Formation (Ksd, formerly the lower member of the Grand Castle Formation) in First Left Hand Canyon southeast of Parowan. From this vantage point, the capping sandstone member (Kwcs) of the Wahweap Formation (formerly middle sandstone member of the Grand Castle Formation) is mostly hidden from view. Kgc (Grand Castle Formation, redefined); Tcp (pink member of the Claron Formation).*

Figure 27. *Upper Cretaceous strata in Johnson Canyon, western Paunsaugunt Plateau, showing the pink member (Tcp) of the Claron Formation in the distant cliffs. The prominent ledge of white pebbly sandstone in the middle of the photograph is the Drip Tank Member (Ksd) of the Straight Cliffs Formation; the canyon bottom is eroded into the John Henry Member (Ksj). Kw = Wahweap Formation.*

volcanic ash, numerous invertebrate and vertebrate fossils, and palynomorphs (Eaton and others, 1999a; Lawton and others, 2003; Eaton, 2006; Jinnah and others, 2009; Jinnah, 2013). Based principally on recovery of microvertebrate fossils using wet screen washing techniques, the Wahweap Formation contains the most diverse middle Campanian terrestrial fauna in North America. In their summary of Wahweap paleontology, DeBlieux and others (2013) noted that Wahweap strata have yielded at least 5 freshwater shark species, 3 freshwater ray species, 8 bony fish species, 11 amphibian species, 10 turtle species, 4 lizard taxa, 5 crocodilian taxa, 15 dinsoaur taxa, and 13 mammal species. In its type area in the Kaiparowits basin it contains four informal members, each with distinctive sandstone-to-mudstone ratios and fluvial architecture (Eaton and others, 2001). We map Eaton's lower three members as a single unit due to inadequate exposure across most of the map area, but we map his distinctive capping sandstone member separately. Detrital zircon and provenance studies of the lower three members show that these rivers flowed longitudinally to the foreland basin and tapped sources in the Cordilleran magmatic arc in southern California or western Nevada and the Mogollon Highlands of southern Arizona, but that the capping sandstone member was deposited by transverse streams that tapped Mesozoic quartzose sandstones in the Sevier orogenic belt (Pollock, 1999; Lawton and others, 2003; Eaton, 2006; Jinnah and others, 2009). Thus the basal contact of the capping sandstone member represents an abrupt change in color, petrology, grain size, and fluvial style, documenting a major shift in source areas from arc to orogenic belt and in depositional environments from meandering to braided rivers.

The lower three members of the Wahweap Formation thicken dramatically westward across the Paunsaugunt Plateau from about 300 feet (90 m) thick north-northeast of Tropic to more than 700 feet (210 m) thick on the west flank of the plateau. These three members are about 800 feet (245 m) thick in Cedar Canyon on the west flank of the Markagunt Plateau, as initially suggested by Eaton and others (1999a) and Moore and Straub (2001). However, something unusual happens to this interval between Cedar and Parowan Canyons. In Parowan Canyon, only a few tens of feet of Wahweap strata are present at the base of the capping sandstone member (formerly middle sandstone member of the Grand Castle Formation). This dramatic southward thickening of the unit south of Parowan Canyon is puzzling and we do not know if the Parowan Canyon lower Wahweap represents a condensed section or just part of one or more of the lower three members.

We do know, however, that the white, mostly medium-grained, trough cross-bedded quartz arenite that overlies the thin Wahweap beds in Parowan Canyon—originally defined as the middle member of the Grand Castle Formation—is the same interval that is exposed in a State Highway 14 road cut at the Webster Flat turnoff just west of Blowhard Mountain, as suggested by Goldstrand and Mullet (1997) and Lawton and others (2003). For mapping purposes, we restrict the capping sandstone member to its distinctive white quartz arenite facies and note that its upper part locally contains abundant pebble stringers with rounded quartzite, dolomite, chert, and limestone clasts. It typically weathers to a distinctive white, manzanita-covered slope-and-bench topography. The capping sandstone is present east of the Sand Pass fault on the west flank of the Paunsaugunt Plateau, and it is exceptionally thick at Hillsdale Canyon. There, at Hillsdale Canyon, the classic white, friable, capping sandstone is overlain by as much as 200 feet (60 m) of ledge- and cliff-forming, yellowish-brown, fine- to medium-grained sandstone with numerous thin, pebbly conglomerate stringers (mapped as Kwcg, a pebbly sandstone unit of the capping sandstone member; figure 28).

Quartz grains of the capping sandstone are well rounded and commonly frosted, recycled from the Navajo Sandstone and other Mesozoic eolian sand bodies (Goldstrand and Mullett, 1997; Pollock, 1999; Lawton and others, 2003). Detrital zircon populations of three samples from the Markagunt and Paunsaugunt Plateaus are similar, lending independent support for our mapping and correlation of this interval (Johnson and others, 2011). Goldstrand and Mullett (1997) and Lawton and others (2003) also showed that the member was deposited in a braided fluvial environment with a paleoflow direction principally to the east to south-southeast, and concluded that it represents sediments of a Late Cretaceous braided stream system that drained thrust sheets of the Sevier orogenic belt to the west.

Interestingly, this distinctive, white quartzose sandstone was originally inferred to be Paleocene in age (as was the entire Grand Castle Formation) (Goldstrand and Mullett, 1997). However, our early mapping suggested that it must be Late Cretaceous because it underlies strata in Cedar Canyon that yielded Late Cretaceous palynomorphs from multiple locations. Still, this distinctive white sandstone lacked internal age control, so in 2010 we gathered together several geologists and paleontologists interested in this problem and set out to examine the outcrops searching for something to date. We visited a steeply dipping section at

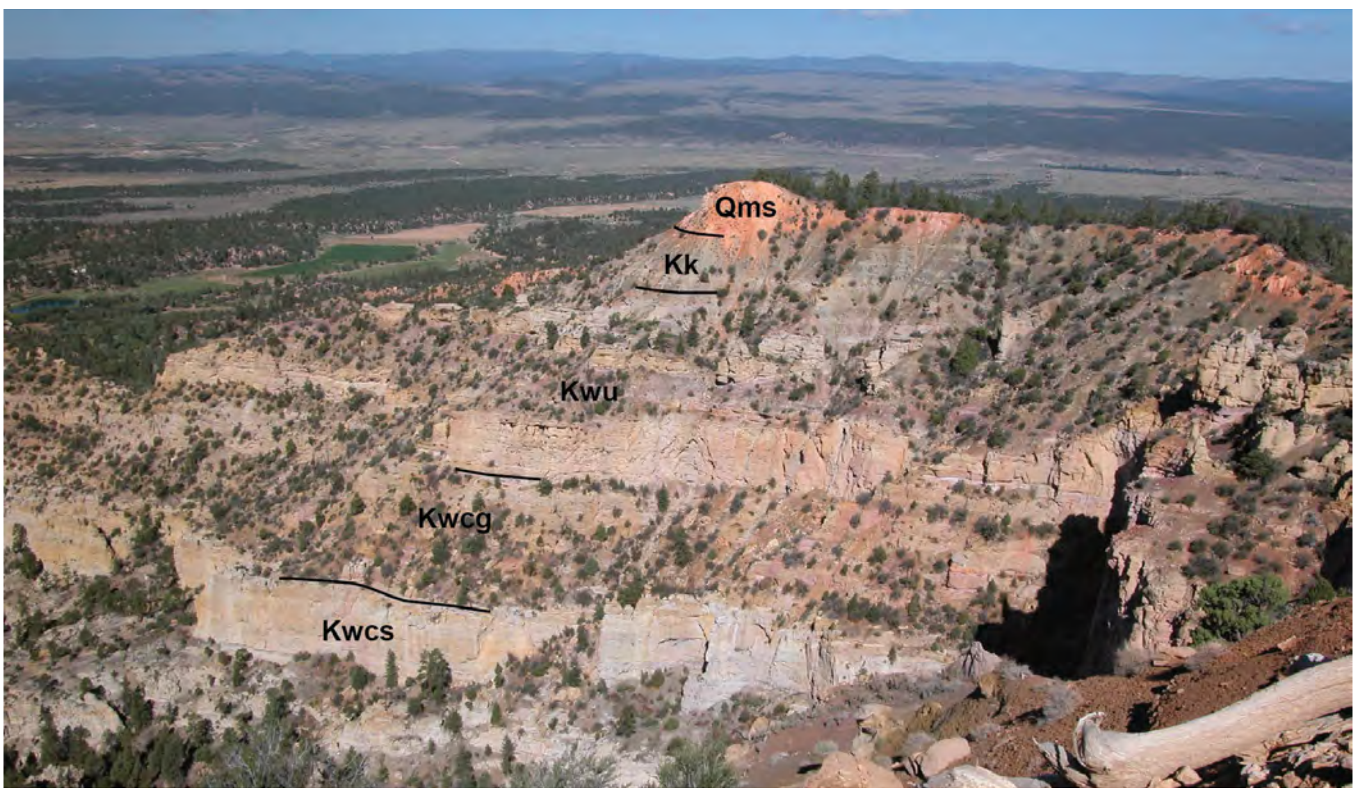


Figure 28. *Upper Cretaceous strata at the entrance to Hillsdale Canyon, west side of Paunsaugunt Plateau. Kwcg denotes pebbly sandstone and mudstone that may be an unusual facies of the capping sandstone member (Kwcs) of the Wahweap Formation. The lower part of the Kaiparowits Formation (Kkl) is mudstone and sandstone that lacks pebbles, above which is distinctive bluish-gray feldspathic lithic sandstone and mudstone of the main body of the Kaiparowits Formation (Kk). Qms denotes landslide deposits derived from the Claron Formation. The Markagunt Plateau is in the distance.*

the range front between Summit and Parowan and began what seemed like a fruitless search for datable materials in the white quartz arenite (figure 29). Luckily, we found and sampled a thin gray mudstone for palynomorphs, and while doing so, Eric Roberts, then at Southern Utah University and now at James Cook University, Australia, looked up at an overhanging ledge and saw the cast of a theropod track, proving a Cretaceous age (Hunt and others, 2011a) (figures 30 and 31). Still, we had the palynomorph sample analyzed and it yielded palynomorphs of indeterminate Campanian to Santonian age. It was one of those precious moments where a key piece of the puzzle drops into place—the quartz arenite was indeed Upper Cretaceous, giving us confidence in our correlation of Wahweap and Straight Cliffs strata between the Paunsaugunt and Markagunt Plateaus.

Kaiparowits Formation

The Kaiparowits Formation is abundantly fossiliferous, with one of the richest and most diverse terrestrial vertebrate faunas of the Cretaceous Western Interior Basin, much of which was discovered in the past decade as part of a larger effort to better understand the fossil resources and geology of Grand Staircase–Escalante National Monument (Roberts, 2007; Roberts and others, 2013). The Kaiparowits Formation was deposited near the western margin of the Late Cretaceous Western Interior Seaway as an eastward-prograding clastic wedge in a relatively wet, subhumid alluvial plain with periodic to seasonal aridity (Roberts, 2007; Roberts and others, 2013). Kaiparowits strata are mostly distinctive gray and bluish-gray, fine-grained, feldspathic, lithic sandstone, mudstone, and siltstone that are poorly cemented and so weather to badland slopes whose bluish-gray hues contrast sharply with overlying pinkish Claron paleosols (fossil soil horizons) and underlying yellowish-brown Wahweap strata (figure 32). In the Kaiparowits basin just east of the map area, the formation is as much as 2820 feet (860 m) thick (Eaton, 1991; Doelling and Willis, 1999b; Roberts and others, 2005), yet it spans just 2 million years of upper Campanian time, 76.6 to 74.5 million years ago (Roberts and others, 2005, 2013). It thus records extremely rapid sediment accumulation rates of 16 inches per thousand years (41 cm/kyr)—a rapidly subsiding basin perfect for preserving fossils. This interval of time is well represented by other equally fossiliferous strata along the length of the Western Interior Basin, thus allowing unprecedented opportunities to study evolution and ecology in ancient continental ecosystems (Miller and others, 2013; Roberts and others, 2013; Sampson and others, 2013).

Figure 29. *Northwest-dipping capping sandstone member (Kwcs) of the Wahweap Formation, Grand Castle Formation (redefined, Kgc), and pink member of the Claron Formation (Tcp) along the west flank of the Markagunt Plateau; Parowan is in the distance at left. Here, the capping sandstone yielded Late Cretaceous (Campanian to Santonian) palynomorphs and theropod dinosaur tracks, confirming our suspicion that the former middle sandstone member of the Grand Castle Formation is the same interval as the capping sandstone. Note that the Grand Castle Formation (redefined) consists of two parts, a lower gray cliff-forming conglomerate and an upper ledge- and slope-forming yellowish-brown conglomerate.*

Figure 30. *Eric Roberts and Gary Hunt below sandstone ledge of the capping sandstone member of the Wahweap Formation; note cast of theropod track (at arrow). Photo by Don DeBlieux (Utah Geological Survey).*

The westward extent of the Kaiparowits Formation has long been problematic, largely due to significant sub-Claron erosion along the Table Cliff monocline and on the Paunsaugunt Plateau and also because of the difficulty of defining its lower contact. Prior to 1990, geologic maps of the region lumped Wahweap and Kaiparowits strata, inferring the presence of the latter unit but not defining its distribution. In the area southwest of Tropic Reservoir, Bowers (1990) assigned light-brown, very fine grained sandstone and gray sandy mudstone (above the capping sandstone member of the Wahweap Formation) to the Kaiparowits Formation; we recovered Campanian to Maastrichtian palynomorphs from this interval. Tilton (1991, 2001a, 2001b) mapped the southernmost Paunsaugunt Plateau, correctly noting the absence there of Kaiparowits Formation.

Resolution of the Kaiparowits problem is found on the west flank of the Paunsaugunt Plateau in Hillsdale Canyon. There, an apparently unusually thick capping sandstone member of the Wahweap Formation is overlain by as much as about 250 feet (75 m) of yellowish-brown, fine-grained sandstone and varicolored and mottled, reddish-brown, purplish-gray, and gray mudstone. The sandstone forms two prominent ledges at the base and near the middle of the unit (see figure 28). This interval is similar to the basal Kaiparowits in the Henrieville Creek area (Eaton, 1991), which exhibits a different detrital zircon age distribution, with more thrust-belt-derived grains than classic blue-gray Kaiparowits strata (Welle, 2008; Lawton and Bradford, 2011). We report a new U-Pb age on zircon of 75.62 +3.08/-1.66 Ma for a bluish-gray smectitic mudstone at the base of the classic bluish-gray Kaiparowits in Johnson Canyon, just south of Hillsdale Canyon (table 5; UGS and A2Z, Inc., 2013b; Gary Hunt, written communication, September 26, 2011); we also recovered late Campanian to Maastrichtian palynomorphs from this location (see table 1). Thus we concur with the intuition of Bowers that these beds represent the basal Kaiparowits. Our mapping documents the distribution of Kaiparowits Formation on the Paunsaugunt Plateau, offering tantalizing hints into the early history of the Paunsaugunt fault. In this map area, we find no evidence of down-to-the-east movement on the Paunsaugunt or other faults that might have bounded the western margin of the Kaiparowits basin; rather we suggest that the formation's apparent abrupt thinning west of the Paunsaugunt fault is simply a result of erosion associated with the sub-Claron unconformity.

Figure 31. *Close-up of theropod track shown in figure 30. Photo by Don DeBlieux (Utah Geological Survey).*

Figure 32. *The northeast end of the Paunsaugunt Plateau preserves about 450 feet (135 m) of lower Kaiparowits strata (Kk) in the footwall of the Paunsaugunt fault, which appear to dip slightly more steeply west than overlying Claron strata (Tcp). Kwcs = capping sandstone member of the Wahweap Formation. View is north towards tree-covered hill 8400 in the NW1/4 section 30, T. 35 S., R. 2 W., south of Johns Valley.*

In the headwaters of Cedar Canyon, immediately underlying the Claron Formation, are about 200 feet (60 m) of fluvial and floodplain strata that have long puzzled geologists. Like Nichols (1997) we recovered Late Cretaceous pollen from this interval, and like Moore and Straub (2001) we find no evidence of a significant unconformity at the base of the Claron Formation in this area. In fact, where well exposed in the Cedar Breaks amphitheater, the interval is commonly mistaken for Claron strata because of the pink iron oxide staining from overlying beds (figure 33). We remain uncertain as how to correlate these beds; they may represent deposits of a small piggyback basin near the leading edge of the thrust belt, possibly the westward, temporal equivalent of the Kaiparowits Formation, or possibly the Pine Hollow Formation (a poorly dated interval below the Claron Formation on the Table Cliff Plateau immediately east of the map area [Larsen and others, 2010]).

Iron Springs Formation

West of the Parowan-Paragonah fault zone, in the Red Hills, and west of the Hurricane fault in the Pine Valley Mountains (southwest of the map area), the Upper Cretaceous section remains undivided as sandstone, mudstone, and rare conglomerate of the Iron Springs Formation, whose sediments were eroded from the Blue Springs and Wah Wah thrust sheets and deposited principally in an alluvial-plain environment (Fillmore, 1991). However, following Maldonado and Williams (1993a), we subdivide the formation into informal lower and upper members, both of which are characterized by thick, repetitive, yellowish-brown, tabular sandstone beds and thinner, non-resistant, brown and gray mudstone. The lower member—likely equivalent to the Tibbet Canyon Member of the Straight Cliffs Formation (Anderson and Dinter, 2010) or possibly the Smoky Hollow Member, and exposed in overturned beds of the upper plate of the easternmost Iron Springs thrust fault—contains numerous ledge-forming oyster beds and thin coal beds indicative of brackish, estuarine environments. The marine bivalve *Mytiloides kossmati* was collected from these beds as well (Eaton and others, 2001), indicative of the maximum westward transgression of the Western Interior Seaway in early Turonian time (Eaton, 1999). The upper member records deposition in fine-grained braided river and floodplain environments of a coastal plain. Sparse vertebrate fossils, gastropods and ostracods, and locally abundant fossil leaves are clues to the diverse flora and fauna that once inhabited floodplains of the upper Iron Springs Formation.

In Parowan Gap, in gently east-dipping beds east of the Iron Springs thrust, Milner and others (2006) and Milner and Spears (2007) reported dinosaur tracks and other trace fossils, as well as a variety of fossil mollusks, turtles, fish, and a diverse and abundant flora. The dinosaur tracks were first reported by Vice and others (2003) and are the first tracks known in the Iron Springs Formation. The tracks, natural casts found at the base of yellow-brown sandstone beds and originally formed in underlying gray mudstone, are found on talus blocks and in at least five beds in the adjacent canyon walls. Most tracks at Parowan Gap are of ornithopods (most likely hadrosaurs), but theropod tracks and a single set of ceratopsian tracks are also

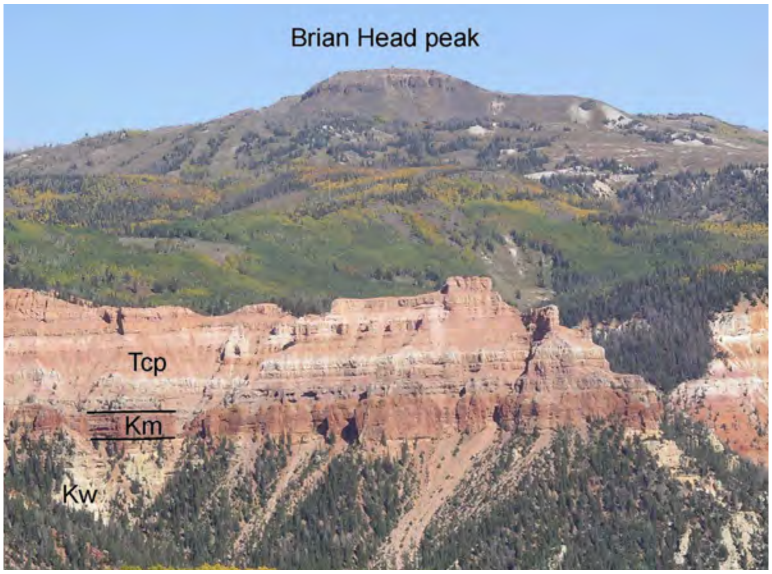


Figure 33. *Brian Head peak as seen from High Mountain. Note sandstone cliff (Cretaceous strata of the Markagunt Plateau, Km), stained dark-reddish-brown from runoff from overlying pink member of the Claron Formation (Tcp). In most areas south of Parowan Canyon, the base of Km corresponds to the top of a thin pebble to small cobble conglomerate containing rounded quartzite and limestone clasts, although in some areas, as here, the conglomerate appears to be missing. Underlying yellowish-brown mudstone, siltstone, and sandstone are assigned to the Wahweap Formation (Kw). The base of the Claron Formation corresponds to the base of the first limestone bed, likely a calcic paleosol.*

present; the latter are the oldest known ceratopsian tracks in North America (Milner and others, 2006). Iron Springs strata here are interpreted to represent river channel, overbank, and floodplain deposits of a fine-grained braided river system (Milner and others, 2006).

Grand Castle Formation

We now redefine the Grand Castle Formation of Goldstrand and Mullett (1997) to consist only of its original upper conglomerate member; the lower conglomerate is now known to be the Drip Tank Member of the Straight Cliffs Formation and the middle sandstone is now known to be the capping sandstone member of the Wahweap Formation as described above (see figure 26). Like the Drip Tank, the Grand Castle Formation (redefined) contains well-rounded, pebble- to boulder-size clasts of quartzite, limestone, sandstone, and chert and was deposited in a braided fluvial environment with paleoflow principally to the east to south-southeast, suggesting source areas in the Wah Wah, Blue Mountain, and Iron Springs thrust sheets of southwest Utah (Goldstrand and Mullett, 1997). Its age is not well constrained, but we suggest that it is indeed Late Cretaceous. A debris-flow deposit within the upper conglomerate member at its type section yielded Late Cretaceous (Santonian?) pollen that Goldstrand and Mullett (1997) interpreted as recycled from older strata. However, Nichols (1997) reported Late Cretaceous *Proteacidites* sp. pollen, which he interpreted as Coniacian and Santonian, from overlying beds here mapped as Cretaceous strata on the Markagunt Plateau (Km) west and south of Blowhard Mountain, and we recovered late Campanian to Maastrichtian pollen from this same interval (table 1).

Paleocene to Middle Eocene
(66 to 38 million years ago)

Whereas the thin-skinned deformation of the Sevier orogenic belt and accompanying foredeep basin controlled Late Cretaceous sedimentation in southwestern Utah, it was the broad basin developed during the latest Creteaceous to Eocene (75 to 45 Ma) Laramide orogeny that gave us the colorful fluvial and lacustrine sedimentary rocks of the Claron Formation. The Laramide orogeny partly overlapped in time and space with the Sevier orogeny, and both were the result of the collision of the Farallon and North American plates; they are regional names for two parts of the Cordilleran orogeny (which affected the entire western part of North America) and are distinguished by differing styles of deformation (DeCelles, 2004). The Sevier orogeny as we have seen was characterized by long folds and thrust faults that took advantage of weak layers in Paleozoic and Mesozoic strata in the western part of the orogenic belt. But east of the Utah hingeline in the central part of western North America, where those strata are much thinner and did not easily decouple from basement rocks, plate tectonic compression produced a different structural style characterized by basement-cored uplifts—the Laramide orogeny.

The San Rafael Swell, Circle Cliffs Uplift, Monument Upwarp, and Kaibab Plateau Uplift are examples of classic Laramide structures in the western Colorado Plateau (figure 34). Each is an asymmetric, doubling plunging anticline (with a conspicuous monocline along its steep eastern flank) formed above generally blind reverse fault zones of low angle (20° to 40°). These faults are commonly thought to have formed above reactivated Neoproterozoic extensional faults that initially formed during intracratonic rifting of Rodinia (Davis and Bump, 2009); the Laramide structures are thus fault-propogation folds formed at the tip of faults rooted in Precambrian basement. Laramide uplifts thus reflect "inversion tectonics"—fault-bounded former Neoproterozoic basins that now stand as structurally high features (Marshak and others, 2000). Although timing is typically poorly constrained, Laramide deformation is widely thought to have begun in the latest Campanian across the Colorado Plateau (Davis and Bump, 2009). In this map area, erosion associated with the pre-Claron unconformity shows that the Paunsaugunt fault was active as a reverse fault following middle Campanian deposition of the Kaiparowits Formation.

Laramide deformation and accompanying quiescence of magmatic arc volcanism is thought to have resulted from subduction at an unusually shallow angle along part of

Table 1. *Palynomorph samples from the Panguitch 30' x 60' quadrangle.*

Sample Number	Map #	Longitude (W)	Latitude (N)	7.5' Quadrangle	Formation or Member	Map Unit Symbol	Identification*	Notes
P062110-1	4	-112.86567	37.81505	Parowan	capping sandstone	Kwcs	Santonian to Campanian	about 75 feet above base
P062110-2	19	-112.86567	37.81505	Parowan	capping sandstone	Kwcs	Santonian to Campanian	1 foot below P062110-1
WP080410-1	18	-112.32274	37.68169	Wilson Peak	capping sandstone	Kwcs	Campanian	
PG110210-1	5	-112.94879	37.89765	Parowan Gap	Iron Springs	Ki	late Cenomanian to Turonian	about 20 feet below Kgc
WP080310-1	15	-112.36435	37.63129	Wilson Peak	John Henry	Ksj	Late Cretaceous, Santonian or older	UMNH Loc. 122
WP080310-3	17	-112.32817	37.6848	Wilson Peak	Kaiparowits, basal	Kk	late Campanian to Maastrichtian	about 30 feet above base of Kk
FA062210-1	1	-112.87973	37.65083	Flanigan Arch	Wahweap	Kw	Santonian to Campanian	about 150 feet below Kwcs
FA062210-2	2	-112.88001	37.64953	Flanigan Arch	Wahweap	Kw	Santonian to Campanian	about 250 feet below Kwcs
FA062210-3	21	-112.88001	37.64953	Flanigan Arch	Wahweap	Kw	Santonian to Campanian	about 30 feet below FA062210-2
TR070611-1	6	-112.2929	37.54985	Tropic Reservoir	Wahweap	Kw	Campanian to Maastrichtian	
WF070611-1	12	-112.88442	37.58065	Webster Flat	Wahweap	Kw	late Campanian to Maastrichtian	UMNH locality 11
WP070411-1	13	-112.30067	37.68045	Wilson Peak	Wahweap	Kw	Campanian to Maastrichtian	
WF062210-1	10	-112.89625	37.59367	Webster Flat	Wahweap, basal	Kw	Turonian to Coniacian	UMNH locality 10
TR070611-3	8	-112.34292	37.6084	Tropic Reservoir	Wahweap, upper	Kw	Late Cretaceous	just below Kwcs
WF062210-2	11	-112.88436	37.58061	Webster Flat	Wahweap, upper	Kw	Santonian to Campanian	UMNH locality 11
NL070611-1	3	-112.8675	37.5785	Navajo Lake	Wahweap, upper unit	Kwu	late Campanian to Maastrichtian	
TR070611-2	7	-112.3177	37.54273	Tropic Reservoir	Wahweap, upper unit	Kwu	late Campanian to Maastrichtian	
TR070611-4	9	-112.34495	37.60167	Tropic Reservoir	Wahweap, upper unit	Kwu	late Campanian to Maastrichtian	just above Kwcs
WP070611-1	14	-112.25888	37.69898	Wilson Peak	Wahweap, upper unit	Kwu	late Campanian to Maastrichtian	
WP080310-5	16	-112.32824	37.68433	Wilson Peak	Wahweap, upper unit	Kwu	Late Cretaceous	about 60 feet below base of Kk
WP080410-2	20	-112.32041	37.68291	Wilson Peak	Wahweap, upper unit	Kwu	Cretaceous	about 20 feet below Tcp cliff

* Identifications by Gerald Waanders.

western North America from southern Montana to northern New Mexico (see, for example, Humphreys, 2009). The cause of this northeast-directed "flat slab" subduction is unknown, but it may have resulted from some combination of subduction of young, buoyant oceanic crust that was comparatively thick by association with oceanic plateaus, in concert with physical dynamics of a subduction zone process known as mantle-wedge suction (itself ultimately a result of gravitational instabilities in the crust and upper mantle) (Humphreys, 2009; Manea and others, 2011). Additionally, Davis and Bump (2009) suggested that the topographically high Sevier orogenic belt imparted southeast-directed compressive stress to the Colorado Plateau. The Colorado Plateau was thus squeezed in two nearly orthogonal directions simultaneously, resulting in mostly northwest- and northeast-trending Laramide uplifts flanked by deep basins that collected sediment eroded from the adjacent highlands.

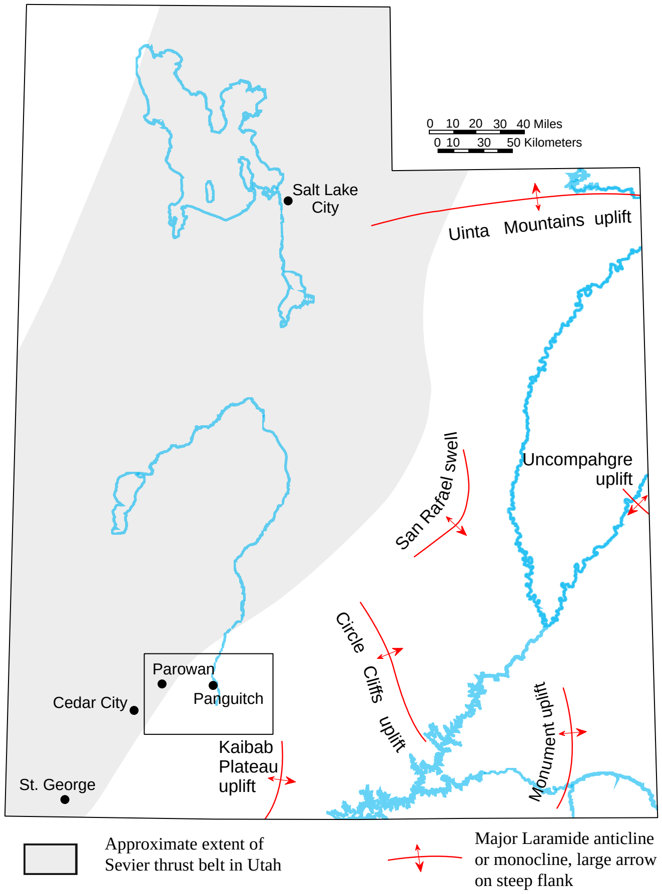


Figure 34. *Major Laramide anticlines and monoclines in Utah. Box shows location of map area. Simplified from Willis (1999).*

Claron Formation

Multi-hued pink, orange, and white strata of the Claron Formation, famously exposed at Cedar Breaks National Monument and Bryce Canyon National Park, were deposited in the southern part of one such Laramide basin (figures 35, 36, and 37). In the central and northern parts of the basin or adjacent basins (the Flagstaff and Uinta Basins, respectively), conglomerate and mudstone of the North Horn Formation, lacustrine limestone of the Flagstaff Limestone, fluvial deposits of the Colton Formation, muddy lacustrine deposits of the Green River Formation, and the mostly fluvial deposits of the Crazy Hollow and Aurora Formations were deposited during Paleocene and Eocene time (Hintze and Kowallis, 2009). The remnants of this Paleocene-Eocene Claron-Flagstaff-Uinta Basin or basins are roughly outlined by today's High Plateaus and Uinta Basin, but its margins are mostly poorly known because of subsequent erosion over today's Colorado Plateau and at the edges of what at the time were relict highlands of the Sevier orogenic belt. The interconnectedness of the basins is also poorly understood, but Davis and others (2009) used isotopic and elemental records preserved in authigenic calcite from samples in each basin to better understand Paleogene landscape evolution of Utah, showing an along-strike migration of a high-elevation landscape from north to south over time and, in southwest Utah, a transition from a closed to open basin beginning with deposition of the white member. When reading older maps and reports of the region, it is important to remember that the definition of the Claron has changed over time, generally becoming more restrictive and purging overlying volcaniclastic strata, and we add additional modifications to its upper and lower contacts as described in the appendix.

The Claron Formation signals the end—almost—of thin-skinned, east-vergent compressional deformation in southwest Utah, and this is dramatically displayed in exposures at Parowan Gap and south and west of the map area along the crest of the Kanarra anticline. Because the Claron basin developed on eroded Mesozoic strata, Claron strata unconformably overlie a variety of Upper Cretaceous formations in this map area. At Parowan Gap, the Claron Formation (and underlying redefined Grand Castle Formation) rests in profound angular unconformity above overturned lower Iron Springs strata; both formations are cut by late-stage movement of the Iron Springs thrust fault, as described in the structural geology section of this report. (Immediately to the west of the map area near Cedar City, and to the southwest near Pintura, Claron strata unconformably overlie strata as old as the Early Jurassic Navajo Sandstone now exposed in the core of the Kanarra anticline [see, for example, Biek and others, 2009; Biek and Hayden, 2013; Knudsen, 2014a].) On the Paunsaugunt Plateau, the Claron unconformably overlies erosionally beveled Kaiparowits, Wahweap, and Straight Cliffs strata, documenting latest Cretaceous to early Tertiary east-vergent displacement on the ancestral Paunsaugunt fault, then a reverse fault but now reactivated as the easternmost large normal fault bounding the west edge of the Colorado Plateau. Between these two areas, on the Markagunt Plateau, Claron strata concordantly overlie Upper Cretaceous conglomerate, sandstone, and mudstone. In the northwestern Markagunt Plateau north of Parowan, the contact between Claron and underlying strata appears gradational and is difficult to pick in a largely conglomeratic interval; we place the contact at the first appearance of reddish-brown siltstone, below which is typically ledge- and cliff-forming conglomerate of the Grand Castle Formation (redefined). Basal Claron strata are similarly conglomeratic in the Red Hills, suggesting that the northwest margin of the Claron basin lay not far northwest of the map area.

On the Markagunt Plateau, conglomerate in the basal Claron decreases markedly southward, so that east of Cedar City the base of the Claron corresponds to the first classic Claron paleosol as also defined by previous workers (for example, Moore and Straub, 2001; Moore and others, 2004). This paleosol overlies about 200 feet (60 m) of yellowish-brown fine-grained sandstone and minor mudstone and siltstone that we assign simply to an informal map unit (Km), Cretaceous strata on the Markagunt Plateau; it may represent local deposition in a small piggyback basin developed near the leading edge of the thrust belt. It is possible that these beds are time-equivalent

Figure 35. Cedar Breaks National Monument, eroded into the thick pink member and comparatively thin white member of the Claron Formation.

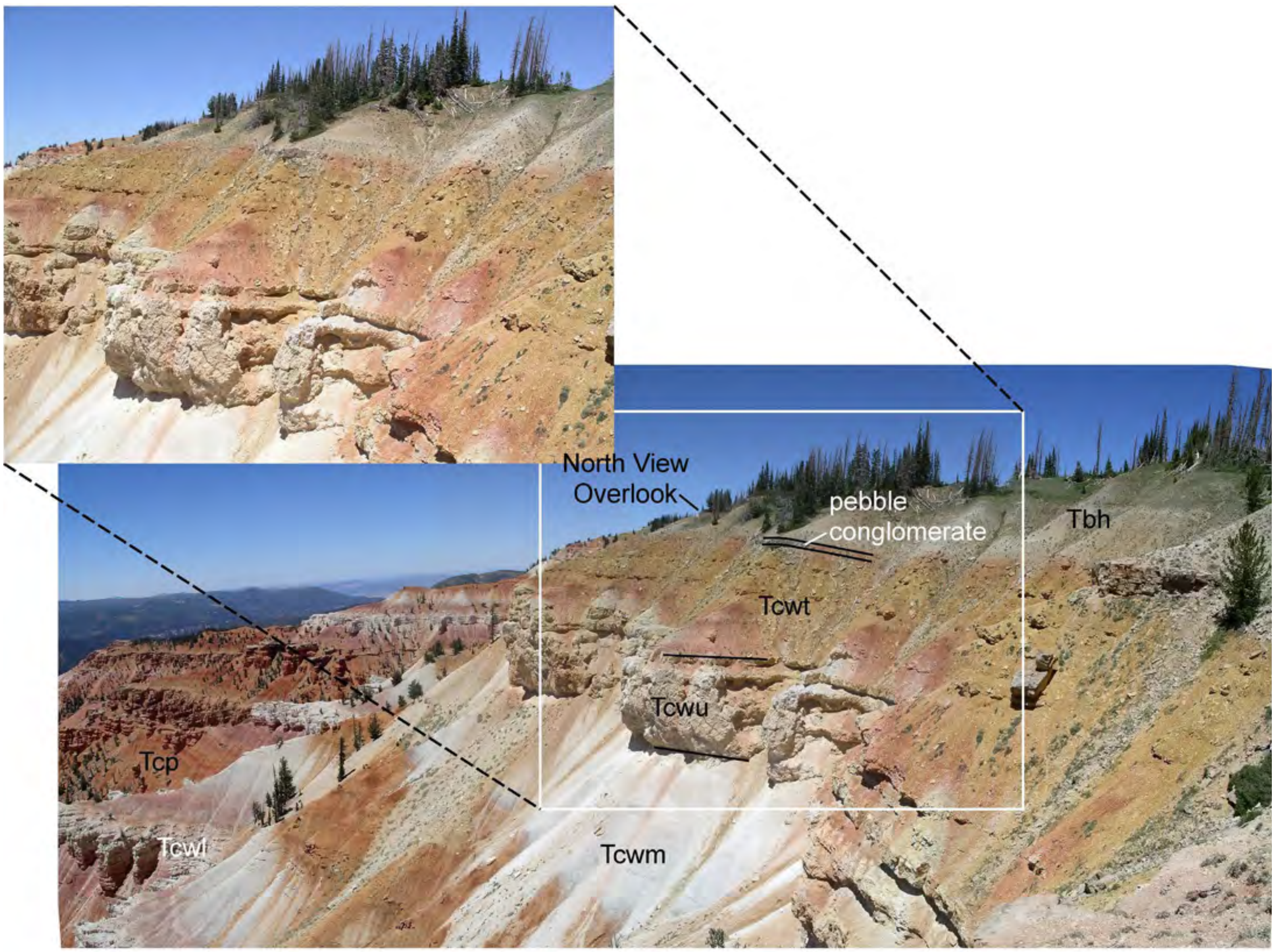


Figure 36. North View Overlook at Cedar Breaks National Monument showing contact between Claron and Brian Head strata (the North View Overlook is on basal strata of the gray volcaniclastic unit of the Brian Head Formation, Tbh). Sable and Maldonado (1997b) assigned variegated, nontuffaceous mudstone, siltstone, and minor sandstone and pebble conglomerate (here labeled Tcwt, 109 feet [33 m] thick) to their lower Brian Head Formation. However, these strata appear identical to strata of the middle white unit (Tcwm); they are nontuffaceous and appear simply to be an uppermost facies of the white member of the Claron Formation, to which we assign them. The top of the Claron, as defined here, is marked by a thin, calcareous, pebbly conglomerate that has rounded clasts of chert, quartzite, and Claron limestone but no volcanic clasts; this conglomerate may be equivalent to the informally named conglomerate at Boat Mesa on the Paunsaugunt Plateau, which marks a significant unconformity in southwest Utah. Tcp = pink member, Tcwl = lower limestone unit of the white member, of the Claron Formation.

Figure 37. *Pink member of the Claron Formation as seen from Bryce Point Overlook. Where exposed along steep escarpments, Claron strata are famous for weathering into a fantastic variety of hoodoos and fins.*

to the Kaiparowits Formation or to the undated Pine Hollow Formation present on the Table Cliff Plateau east of the map area.

Much of the Claron Formation was extensively modified by soil-forming processes such that it now represents a stacked sequence of paleosols (ancient soils) interlayered with fluvial and minor lacustrine deposits (figure 38) (Mullett and others, 1988a, 1988b; Mullett, 1989; Davis and Pollock, 2010). Locally abundant trace fossils of ants, wasps, and bees in the upper part of the pink member and lower part of the white member record insect nest activity during paleosol formation (Bown and others, 1995a, 1995b, 1997). Crayfish burrows in Claron strata of the Markagunt Plateau record relatively deep and highly fluctuating water tables in the pink member, and relatively shallow water tables in alluvial parts of the white member (Hasiotis and Bown, 1997). But as well known as the Claron is for its spectacular scenery—it does after all form the Pink Cliffs, the uppermost riser and tread of the Grand Staircase—it is equally known among geologists for being frustratingly unfossiliferous and devoid of datable volcanic ash or age-constraining detrital zircons. The lack of volcanic ash stems directly from the near cessation of volcanism that accompanied relatively rapid, "flat-slab" subduction at the western margin of North America during the Paleocene to early Eocene (see, for example, Ward, 1991).

The lack of fossils is intriguing in that the formation was apparently deposited during a period of abrupt mammalian diversification that accompanied a gradual 6-million-year-long global warming trend culminating in the Eocene Climatic Optimum (see, for example, Röhl and others, 2000). It was during this time, when the globe was free of continental glaciers, that many major mammalian orders, including horses and primates, first appeared. The beginning of the climatic optimum corresponds with the Paleocene-Eocene Thermal Maximum (PETM), the most abrupt and significant climatic disturbance of the Cenozoic era. The PETM lasted about 20,000 years when global temperatures abruptly rose by about 9°F (5°C) (Wright and Schaller, 2013). The cause of the PETM is unclear, but it led to degassing of clathrates (methane ice on the sea floor), further accentuating climatic warming. Atmospheric carbon levels during the period are known to be much higher than today's concentration of about 400 ppm; interestingly, the rate of CO_2 release to the atmosphere was similar to what is happening today (Cui and others, 2011).

Sinkholes are common in the pink member in the central Markagunt Plateau (figure 39) (Moore and others, 2004; Biek and others, 2011; Hatfield and others, 2010; Rowley and others, 2013). Large sinkholes visible on 1:20,000-scale aerial photographs are plotted on the geologic map, and doubtless many smaller sinkholes are present. These sinkholes capture local runoff and serve to shunt shallow groundwater rapidly down dip where it emerges as springs, including the large Mammoth, Asay, and Cascade Springs (Wilson and Thomas, 1964; Spangler, 2010;

Weaver, 2010) (figure 40). Karst terrain developed in the pink member of the Claron Formation makes the groundwater of the Markagunt Plateau particularly susceptible to contamination.

The white member of the Claron Formation is locally missing on the west flank of the Markagunt Plateau, for example, southeast of Parowan and northeast of Paragonah, and the conglomerate at Boat Mesa is thin or missing there as well. We are uncertain if this represents nondeposition or erosion of white member strata, but suspect the former, implying that this area was the former northwestern margin of lacustrine deposition. Clastic strata of the uppermost pink member in western exposures are possibly coeval basin-margin facies of lacustrine strata of the white member in eastern exposures.

The maximum age of the pink member is poorly constrained as Eocene to Paleocene(?) on the basis of unspecified late Paleocene palynomorphs from lower Claron strata on the east side of the Pine Valley Mountains, and the Paleocene to Eocene gastropods *Viviparus trochiformis, Physa* sp., and *Goniobasis* sp. (table 2) (Goldstrand, 1990, 1994). Goldstrand (1992, 1994) suggested that the pink member may be time transgressive, being older in western exposures and possibly no older than middle Eocene on the Table Cliff Plateau. This idea, however, was based on fission-track analysis of a single sample from the underlying Pine Hollow Formation, which we consider suspect. For one, such a young age seems at odds with the time required to accummulate such a thick stack of mature paleosols. The inferred young age also appears at odds with the nature of the contact between Upper Cretaceous strata on the Markagunt Plateau (Km) and Claron strata described above. Geologists have had no better luck constraining the age of the underlying Pine Hollow Formation on the Table Cliff Plateau east of the map area—it presents a complete lack of datable materials and although a Paleocene(?) (Bowers, 1972) or even Paleocene to early Eocene age (Larsen and others, 2010) is preferred, Bowers (1972) correctly noted that a latest Cretaceous age cannot be ruled out, as did Anderson and Rowley (1975) and Rowley and others (1979). Given our current understanding of the lower Claron For-

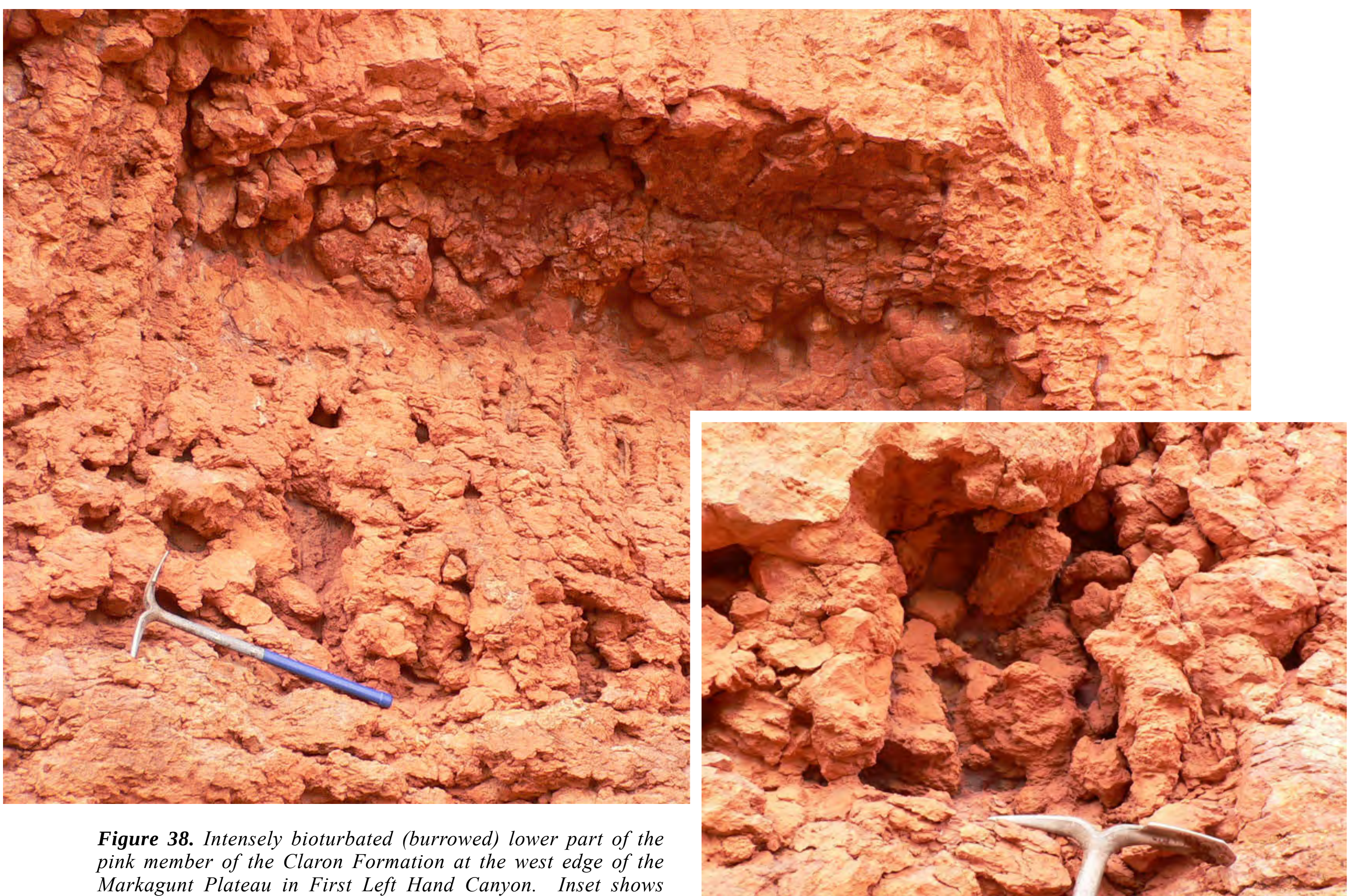


Figure 38. *Intensely bioturbated (burrowed) lower part of the pink member of the Claron Formation at the west edge of the Markagunt Plateau in First Left Hand Canyon. Inset shows close-up of burrows, which may be crayfish burrows.*

mation and its paucity of datable materials, we consider it possible that basal beds of the pink member are latest Cretaceous in age. This possibly older age is intriguing given the fact that Anderson and Dinter (2010) showed that lower Claron strata are cut by and gently folded above the Iron Springs thrust fault at Parowan Gap, showing that the last stages of thrust faulting in southwest Utah continued into lower Claron time; our new mapping further shows that minor extensional relaxation occurred on this thrust fault. Previously, Claron and underlying Grand Castle strata were thought to postdate thrust faulting in southwest Utah. Additionally, Laramide depositional basins become progressively younger from central Utah to central Colorado (Decelles, 2004), further implying that the Claron basin may be older than previously thought.

The age of the white member is well constrained as late middle Eocene (Duchesnean Land Mammal Age) on the basis of sparse vertebrate fossils from this unit on the eastern Markagunt Plateau (Eaton and others, 2011); by limiting U-Pb zircon ages of 35.77 ± 0.28 Ma and 36.51 ± 1.69 Ma for overlying basal Brian Head Formation on the Markagunt and Sevier Plateaus, respectively (Rowley and others, 2013); and by a U-Pb detrital zircon age of 37.97 +1.78/-2.70 Ma from the conglomerate at Boat Mesa on the southwestern Sevier Plateau (this mapping). Middle Eocene vertebrate fossils and charophytes are also known in basal Brian Head strata on the southwestern Sevier Plateau (Feist and others, 1997; Eaton and others, 1999b).

Figure 39. *Typical sinkhole on the southeast flank of Blowhard Mountain, immediately west of Utah Highway 148. The sinkhole formed in the pink member of the Claron Formation and here propagated upward through Markagunt gravity slide residuum.*

The Claron Formation has long been considered correlative with the Flagstaff Limestone of central Utah (see, for example, Hintze and Kowallis, 2009), but our new ages show that the white member postdates or is age-equivalent to the Aurora Formation (Willis, 1988). Given the poor age control on the pink member, we thus see two possibilities for correlation to central Utah strata: (1) the Claron Formation is age-equivalent to at least part of the North Horn Formation and the Flagstaff, Colton, Green River, Crazy Hollow, and Aurora Formations in central Utah, or (2) the Claron is mostly Eocene and thus overlies an unconformity of at least 10 million years duration. Additional age control, especially of the pink member, is needed to assess such correlations.

Conglomerate at Boat Mesa

We know that Claron deposition ended by the latest middle Eocene, about 38 million years ago, with deposition of the conglomerate at Boat Mesa above an erosionally beveled surface. That surface cuts gently down from north to south on the Paunsaugunt Plateau where the conglomerate is best developed (figure 41). Westward, on the Markagunt Plateau, a similar conglomerate, typically less than a foot (0.3 m) thick and so not mapped separately, is locally present at the top of the Claron. The conglomerate is distinctive in that it lacks volcanic clasts, but it contains a robust suite of late Eocene detrital zircons that yielded a U-Pb detrital age of 37.97 +1.78/-2.70 Ma (Gary Hunt, UGS, written communication, March 7, 2012; UGS and A-to-Z, 2013b). The source of the zircon is unknown, but the lack of volcanic clasts suggests that the formation predates the inception of volcanism in this part of southwest Utah. Our detrital zircon sample also yielded a Middle Jurassic peak of about 168 Ma from a coherent group of 34 grains, indicating that Middle Jurassic volcanic or intrusive rocks provided a significant source of sediment to the formation even though it lacks such clasts.

The conglomerate at Boat Mesa represents deposits of braided stream channels and minor floodplains incised into the Claron Formation. On the Paunsaugunt Plateau, this conglomerate is typically 50 to 100 feet (15–30 m) thick and consists of two parts differentiated by color and lithology (figure 42). The lower part is less widespread,

Figure 40. *Cascade Falls, which feeds the North Fork of the Virgin River, derives its water from Navajo Lake via a cave system developed in the pink member of the Claron Formation; inset shows water emerging from the cave. This is an unusual example of interbasin water transfer, albeit of natural origin. Here, water that would formerly have drained east to the upper Sevier River and from there north and west to the Great Basin now flows into the Colorado River basin. The lake first formed when the Henrie Knolls lava flows dammed the Navajo Lake drainage, likely in late Pleistocene time, and the cave system itself developed parallel to a small-displacement normal fault. A small earthen dam prevents the entire lake from draining through a sinkhole in the east part of the small basin.*

Table 2. *Goldstrand (1991) Claron samples.*

Sample Number	UTM Northing	UTM Easting	7.5′ Quadrangle	Formation	Identification
23-88	4162620	397890	Bryce Point	Claron	*Physa* sp.?
31-88	4177720	425000	Upper Valley	Claron	*Viviparus trochiformis*
3 89	4137550	296410	Pintura	Claron	palynomorph
1 90	4137910	295110	Pintura	Claron	*Goniobasis* sp.?

Note: Sample 3 89 easting originally stated incorrectly as 496410.

likely restricted to stream channels incised into the Claron surface; it is a ledge-forming interval that is yellowish- to reddish-brown, very thick bedded pebbly conglomerate with rounded pebbles of black chert, brown and gray quartzite, and lesser Paleozoic limestone overlain by slope-forming, fine- to medium-grained sandstone, siltstone, and mudstone. The upper part is mostly light-gray conglomerate, lesser light-gray to light-brown calcareous sandstone and conglomeratic sandstone, and minor white to light-gray limestone and conglomeratic limestone. This upper part contains distinctive green quartzite pebbles, and in the limestone intervals, clasts commonly appear to float in a carbonate mud matrix. The upper part forms a prominent white ledge; where it overlies the white lacustrine beds of the white member the two units form a single white ledge, making it difficult to differentiate the units from a distance.

Figure 41. *Bryce Point is capped by a 30-foot-thick (6 m) ledge of the conglomerate at Boat Mesa. Volcanic mudflows of the Mount Dutton Formation cap the Sevier Plateau in the distance. Bryce Canyon National Park preserves and celebrates the fantastically eroded landscape eroded into the Claron Formation at the east edge of the Paunsaugunt Plateau.*

Late Eocene to Oligocene
(38 to 23 million years ago)

Brian Head Formation

Although the conglomerate at Boat Mesa completely lacks volcanic clasts, it ushered in a new stage in the geologic history of southwestern Utah, one of widespread volcanism that swept southward across western North America from about 55 to 20 million years ago. In southwestern Utah, the oldest volcanic rocks belong to the Brian Head Formation, named for poor exposures of white volcaniclastic mudstone, siltstone, silty sandstone, sandstone, volcanic ash, micritic limestone, and minor conglomerate and multi-hued chalcedony at its type area of Brian Head peak (figure 43). These strata, rich in volcanic ash, were deposited in low-relief fluvial, floodplain, and lacustrine environments (Sable and Maldonado, 1997b) (figure 44). The base of the section is well exposed near the North View Overlook in Cedar Breaks National Monument; there, a thin rhyolitic ash bed overlies a thin pebbly conglomerate likley equivalent to the conglomerate at Boat Mesa (figure 45). This ash bed yielded a U-Pb age on zircon of 35.77 ± 0.28 Ma, and several additional radiometric ages from Brian Head strata from across the map area show that it was deposited from about 37 to 33 million years ago (table 3). It is mostly late Eocene in age, now barely reaching into the Oligocene (Sable and Maldonado [1997b] assigned Brain Head strata to the Oligocene; at that time, the Eocene/Oligocene boundary was about 38 Ma).

A more complete section of the Brian Head Formation is well exposed on the southwest flank of the Sevier Plateau (figure 46). There, a transitional interval as much as 160 feet (50 m) thick, not present elsewhere, of fine-grained, slope-forming sandstone, siltstone, and mudstone of red, pink, yellowish-

Figure 42. *At Sand Wash, the conglomerate at Boat Mesa consists of two parts: (1) a lower, ledge-forming, reddish-brown pebbly conglomerate and overlying slope-forming sandstone, siltstone, and mudstone (Tbml), and (2) an upper white pebbly conglomerate and conglomeratic limestone (Tbm). Volcaniclastic sandstone and mudstone of the Brian Head Formation form the overlying pinyon-juniper-covered slopes, whereas volcanic mudflow deposits of the Mount Dutton Formation form the bold cliffs of Blind Spring Mountain.*

brown, and purplish-gray hues forms the base of the formation. This variegated interval is mostly non-volcaniclastic, but some mudstone intervals exhibit swelling soils that suggest a volcanic ash component. Fossil turtles, charophytes, and fish—dominated by aquatic taxa suggestive of lacustrine paleoenvironments—are known from this interval and assigned to the Duchesnean North American Land Mammal Age (end of middle Eocene) (Feist and others, 1997; Eaton and others, 1999b; Korth and Eaton, 2004).

Like the underlying Claron Formation, the Brian Head Formation has a long and convoluted nomenclatural history reflecting the evolving state of our knowledge as different workers mapped and studied different parts of the formation. We further restrict its definition by reassigning the Markagunt Plateau "variegated beds" of Sable and Maldonado (1997b) to the uppermost Claron Formation so that at its type section, the Brian Head Formation consists of just its middle volcaniclastic unit. An upper volcanic unit of mudflow breccia, lava flows, autoclastic lava-flow breccia, volcaniclastic sandstone and conglomerate, and lesser ash-flow tuff is present on the northern Markagunt Plateau west-southwest of Bear Valley Junction (figure 47), although this interval, mapped as local volcanic rocks by Anderson and Rowley (1987), may better be assigned to a distal, early phase of the Mount Dutton Formation; similar strata were reported by Maldonado and Williams (1993a, 1993b) in the northern Red Hills (northern end of Jackrabbit Mountain). In contrast, the excellent exposures on the southwestern Sevier Plateau enabled us to map the Brian Head Formation there as four informal units, the up-

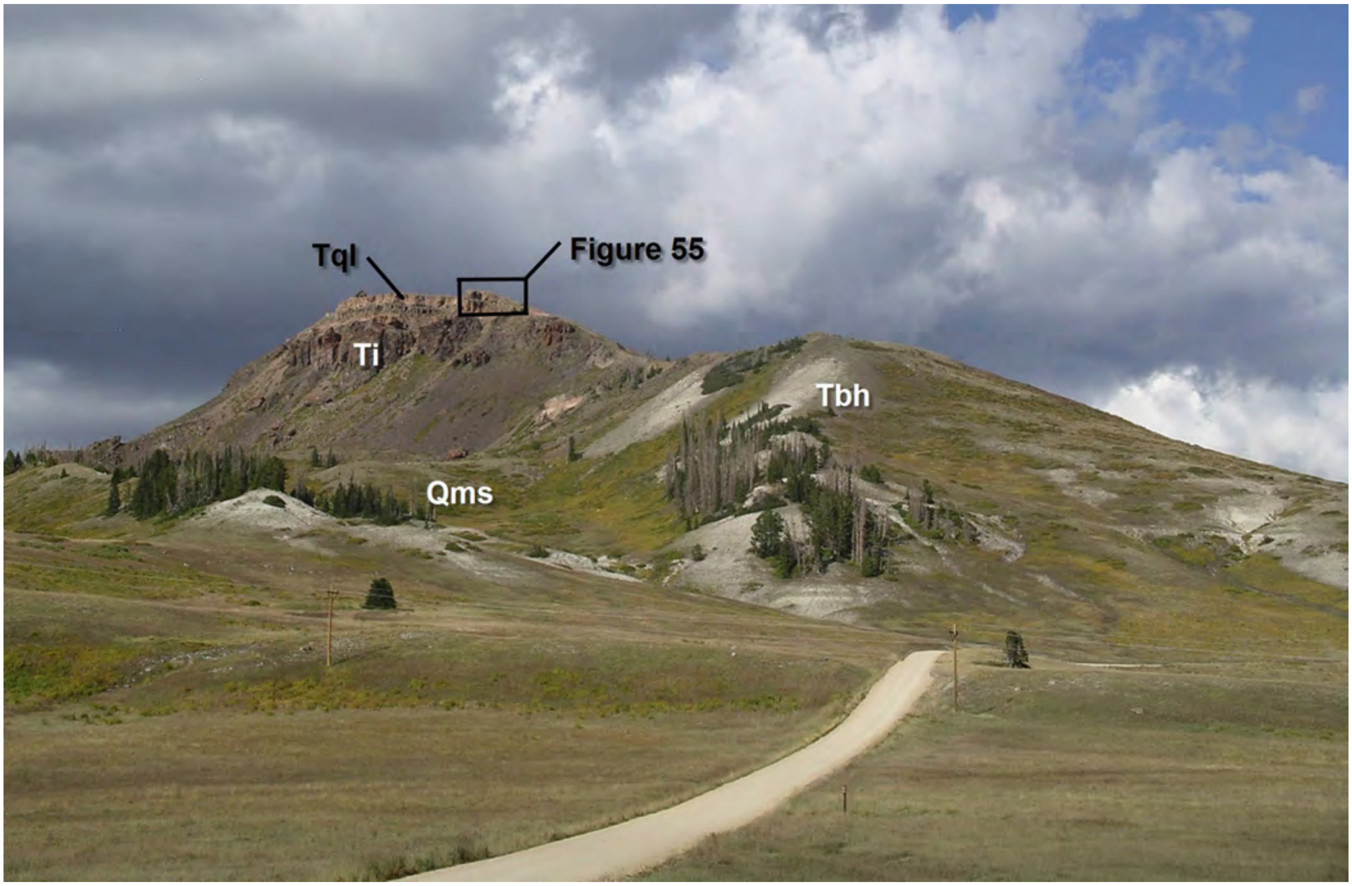


Figure 43. Brian Head peak. The type section of the Brian Head Formation (Tbh) is on the ridge just right of center (by the Tbh label). Brian Head peak is capped by the Leach Canyon Formation (Tql; see also figure 55), which overlies the Isom Formation (Ti). A large landslide complex (Qms) is west and south of the peak.

per three of which correspond to the middle volcaniclastic unit of the formation on the Markagunt Plateau.

The Brian Head Formation is known for abundant trace fossils, including possible crayfish burrows and root traces (Golder and Wizevich, 2009; Golder and others, 2009), but aside from its basal variegated interval it is surprisingly unfossiliferous. It is also known for its colorful beds of chalcedony in various shades of white, gray, yellow, red, black, and brown, all typically with a white weathering rind. The chalcedony, probably derived from remobilization of silica in glass shards from the tuff beds, forms resistant beds commonly 1 to 3 feet (0.3–1 m) thick but locally as much as 8 feet (2.5 m) thick, and is suggested to have resulted from silicification of limestone beds (Maldonado, 1995; Sable and Mal-

Figure 44. Air-fall ash bed in the Brian Head Formation just south of Haycock Mountain. Sample HM071809-4 from this bed yielded a U-Pb age on zircon of 34.95 ± 0.83 Ma.

Figure 45. *Basal Brian Head strata just south of the North View Overlook at Cedar Breaks National Monument. Utah Geological Survey geologists Don DeBlieux (standing) and Gary Hunt are at contact with underlying Claron Formation (here partly covered by Brian Head slopewash debris). Inset shows thin rhyolitic ash (the upper part of which is a deep reddish brown) at base of the Brian Head Formation that yielded a U-Pb age on zircon of 35.77 ± 0.28 Ma.*

donado, 1997b; Schinkel, 2012) or possibly volcanic ash beds (Bakewell, 2001). Being resistant, the chalcedony commonly litters slopes developed on Brian Head strata, and it was commonly used for tools and arrowheads by Native Americans. The chalcedony, however, is highly fractured and without heat treatment is of little use for lapidary purposes (Strong, 1984).

Above all, however, because of its abundant smectitic clay derived from weathered volcanic ash, the Brian Head Formation is known for its swelling soils and for its susceptibility to landsliding. Nearly everywhere it is exposed on steep hillsides it forms large landslide complexes, the largest of which fills the 15-mile (24 km) length of Yankee Meadows graben. The non-resistant, clay-rich Brian Head Formation was also the principal detachment surface for the Markagunt gravity slide.

Regional Ash-flow Tuffs

Utah's middle Tertiary landscape looked unimaginably different from that of today. Geologists refer to that former landscape as the Great Basin altiplano or Nevadaplano, a high-elevation region that stretched from the Sierra Nevada batholith in eastern California eastward to what is now the Colorado Plateau (DeCelles, 2004; Best and others, 2009, 2013). The altiplano was studded with volcanic mountains and intervening basins, analogous perhaps to the modern Andean Altiplano of South America. It was onto this landscape that dozens of widespread ash-flow tuffs accumulated—part of the so-called middle Cenozoic "ignimbrite flare-up" of southwestern North America, which represents one of the largest episodes of subduction-related, silicic volcanism known on Earth (figure 48). Eocene to Miocene volcanism in Utah is part of a broad pattern of volcanism that migrated southward through time, in large igneous belts that contained the eruptive centers (Mackin, 1960; Cook, 1965; Armstrong and others, 1969; Noble, 1972; Stewart and Carlson, 1976; Stewart and others, 1977; Rowley, 1998; Rowley and Dixon, 2001) due to progressive detachment and foundering of the "flat slab" portion of the subducted Farallon oceanic plate (figure 49 and 50) (Dickinson, 2006; Humphreys, 2009). According to Humphreys (2009), foundering of the flat slab is thought to have been caused by accretion of a large oceanic block

Table 3. *$^{40}Ar/^{39}Ar$, K-Ar, and U-Pb formation ages of volcanic rocks, Panguitch 30' x 60' quadrangle and adjacent area.*

Lava Flow or Formation	Map Symbol	Sample number	Map #	K-Ar age (Ma)	K-Ar age (Ma) (corrected)	$^{40}Ar/^{39}Ar$ age (Ma)	U-Pb age (Ma)	Mineral	7.5' Quadrangle	Longitude	Latitude	Lab used	Reference	Comments
Asay Knoll	Qbak	pan-5		0.52 ± 0.05				whole rock	Asay Bench	-112.5481	37.5525	USGS	Best and others (1980)	also used labs at BYU and UofA - Tucson
Bear Valley Fm.	Tbv	R-10	2	24.0 ± 0.4	24.6 ± 0.4			biotite	Fivemile Ridge	-112.5333	37.8700	Ohio State University	Fleck and others (1975)	tuff
Bear Valley Fm.	Tbv	R-9	3	23.9 ± 0.5	24.5 ± 0.5			plagioclase	Fivemile Ridge	-112.5600	37.8617	Ohio State University	Fleck and others (1975)	vitric ignimbrite
Bear Valley Formation tuff	Tbv	CM081612-3	45				24.58 ± 1.92	zircon	Cottonwood Mountain	-112.7051	37.9029	AtoZ	this report	
Bear Valley Formation tuff	Tbvt	89USa2a	25	15.8				glass	Haycock Mountain	-112.5790	37.7170	USGS (Mehnert)	Sable unpublished data (1992)	unsuitable; basal vitrophyre of Bear Valley Tuff
Bear Valley Formation tuff	Tbvt	89USa2a	34	22.3 ± 1.1				plagioclase	Haycock Mountain	-112.5790	37.7170	USGS (Mehnert)	Sable and Maldonado (1997a)	basal vitrophyre of Bear Valley Tuff
Bear Valley Formation tuff	Tbvt	89USa2a	35			24.23 ± 0.17		plagioclase	Haycock Mountain	-112.5790	37.7170	USGS (Snee)	Sable and Maldonado (1997a)	basal vitrophyre of Bear Valley Tuff
Black Mountain	Qbbm	C-350-5	4	0.80 ± 0.24				whole rock	Webster Flat	-112.9592	37.5889	USGS	Anderson and Mehnert (1979)	
Black Mountain	Qbbm	zion-63	5	0.87 ± 0.04				whole rock	Webster Flat	-112.9372	37.5675	USGS	Best and others (1980)	also used labs at BYU and UofA - Tucson
Blue Spring Mountain	Tbbm	626BS1				2.78 ± 0.16		whole rock	Panguitch Lake	-112.6700	37.6700	NIGL	Stowell (2006)	location imprecise
Brian Head Fm.	Tbh	BH062310-1	6				35.77 ± 0.28	zircon	Brian Head	-112.8302	37.6548	AtoZ	this report	base of formation
Brian Head Fm.	Tbh						34.7 ± 0.6	zircon	Brian Head	-112.8311	37.6811	unknown	Davis and others (2009)	Brian Head peak
Brian Head Fm.	Tbh						35.2 ± 0.8	zircon	Brian Head	-112.8311	37.6811	unknown	Davis and others (2009)	Brian Head peak
Brian Head Fm.	Tbh	CC101310-1	7				36.51 ± 1.69	zircon	Casto Canyon	-112.2570	37.8196	AtoZ	this report	about 80' above base of formation
Brian Head Fm.	Tbh	93USa10a				34.99 ± 0.22***		biotite	George Mountain	-112.4700	37.5872	NMGRL	Sable unpublished data (1996)	excess Ar; unusual K/Ca ratio
Brian Head Fm.	Tbh	HM111810-1	9				34.95 ± 0.83	zircon	Haycock Mountain	-112.5522	37.7059	AtoZ	this report	same as HM071809-2

Table 3. continued

Lava Flow or Formation	Map Symbol	Sample number	Map #	K-Ar age (Ma)	K-Ar age (Ma) (corrected)	$^{40}Ar/^{39}Ar$ age (Ma)	U-Pb age (Ma)	Mineral	7.5' Quadrangle	Longitude	Latitude	Lab used	Reference	Comments
Brian Head Fm.	Tbh	HM111810-2	8				33.55 ± 0.80	zircon	Haycock Mountain	-112.5502	37.7066	AtoZ	this report	same as HM071809-4
Brian Head Fm.	Tbh	HM071809-2	40			35.04 ± 0.05**		sanidine	Haycock Mountain	-112.5522	37.7059	NIGL	this report	
Brian Head Fm.	Tbh	HM071809-4	39			33.80 ± 0.05**		sanidine	Haycock Mountain	-112.5502	37.7066	NIGL	this report	
Brian Head Fm., tuff	Tbht	Rh-573		26.3 ± 1.3				biotite	Parowan Gap	-112.8750	37.9772	USGS (Mehnert)	Sable unpublished data (1992)	see also Maldonado and Moore (1995)
Brian Head Fm., tuff	Tbht	Rh-573		34.2 ± 2.1				plagioclase	Parowan Gap	-112.8750	37.9772	USGS (Mehnert)	Sable unpublished data (1992)	see also Maldonado and Moore (1995)
Brian Head Fm., tuff	Tbht	Rh-573				33.00 ± 0.13		plagioclase	Parowan Gap	-112.8750	37.9772	USGS (Snee)	Sable unpublished data (1994)	see also Maldonado and Moore (1995)
Brian Head Fm., tuff	Tbht	Rh-573				33.70 ± 0.14		biotite	Parowan Gap	-112.8750	37.9772	USGS (Snee)	Sable unpublished data (1994)	see also Maldonado and Moore (1995)
Coopers Knoll	Qbck	HM101408-1	39			< 0.92		whole rock	Haycock Mountain	-112.6071	37.6879	NIGL	this report	
Dickinson Hill	Tbdh	AC-PANG	12	5.3 ± 0.5				whole rock	Panguitch	-112.4292	37.8000	USGS	Rowley and others (1994a)	approximate coordinates; NE wall DD Hollow
Haycock Mountain Tuff	Thm	R-11	1	22.1 ± 0.6	22.7 ± 0.6			plagioclase	Fivemile Ridge	-112.5167	37.8333	Ohio State University	Fleck and others (1975)	misidentified as Bauers Tuff Member
Haycock Mountain Tuff (type section)	Thm	94UPh-Thm2	13			22.75 ± 0.12		sanidine	Haycock Mountain	-112.6083	37.7300	NMGRL	Sable unpublished data (1996)	approximate location
Henrie Knolls	Qbhk	HK092106-1	14			0.058 ± 0.036*		whole rock	Henrie Knolls	-112.6595	37.5903	NMGRL	Biek and others (2011)	
Houston Mountain	Tbhm	HK092006-3	15			5.27 ± 0.14		whole rock	Henrie Knolls	-112.7232	37.6175	NMGRL	Biek and others (2011)	
Iron Peak laccolith	Tip	R-28		19.7 ± 0.5	20.2 ± 0.5			whole rock	Cottonwood Mountain	-112.7017	37.8483	Ohio State University	Fleck and others (1975)	latitude uncertain; possibly 37.9111
Iron Peak lava flow?	Tipl?	R-27		20.7 ± 0.5	21.2 ± 0.5			whole rock	Red Creek Reservoir?	-112.6767	37.7550	Ohio State University	Fleck and others (1975)	location uncertain

Table 3. *continued*

Lava Flow or Formation	Map Symbol	Sample number	Map #	K-Ar age (Ma)	K-Ar age (Ma) (corrected)	$^{40}Ar/^{39}Ar$ age (Ma)	U-Pb age (Ma)	Mineral	7.5' Quadrangle	Longitude	Latitude	Lab used	Reference	Comments
Isom Fm. (Bald Hills Tuff Mbr.)	Ti	R-8	16	25.0 ± 0.4	25.7 ± 0.4			plagioclase	Fivemile Ridge	-112.5667	37.8550	Ohio State University	Fleck and others (1975)	
Isom Fm. (Blue Meadows Tuff Mbr.)	Ti	R-7		25.2 ± 0.4	25.9 ± 0.4			plagioclase	Panguitch NW	-112.4333	37.9700	Ohio State University	Fleck and others (1975)	
Leach Canyon Fm.	Tql	89USa1a	17	22.8 ± 1.1				biotite	Panguitch Lake	-112.6944	37.7003	USGS (Mehnert)	Rowley and others (1994a)	
Leach Canyon Fm.	Tql	89USa1a	41	24.3 ± 1.0				sanidine (rerun)	Panguitch Lake	-112.6944	37.7003	USGS (Mehnert)	Sable and Maldonado (1997a)	
Leach Canyon Fm.	Tql	89USa1a	42	24.8 ± 1.0				sanidine	Panguitch Lake	-112.6944	37.7003	USGS (Mehnert)	Rowley and others (1994a)	
Leach Canyon Fm.	Tql	89USa1a				23.86 ± 0.26		biotite	Panguitch Lake	-112.6833	37.7083	USGS (Snee)	Sable and Maldonado (1997a)	
Limerock Canyon Fm.	Tl	H111810-1	18				20.52 ± 0.49	zircon	Hatch	-112.4721	37.6588	AtoZ	this report	
Limerock Canyon Fm.	Tl	LT-1B-89	19	19.8 ± 0.8				sanidine	Hatch	-112.4681	37.6569	USGS (Mehnert)	Sable and Maldonado (1995b)	also see Kurlich and Anderson (1997)
Limerock Canyon Fm.	Tl	LT-1B-89	32	20.2 ± 1.4				biotite	Hatch	-112.4681	37.6569	USGS (Mehnert)	Sable and Maldonado (1995b)	also see Kurlich and Anderson (1997)
Limerock Canyon Fm.	Tl	LT-2-89	33			20.48 ± 0.8		biotite	Hatch	-112.4681	37.6569	USGS (Snee)	Sable and Maldonado (1995b)	
Limerock Canyon Fm.	Tl	LT-4-89	20	21.0 ± 1.0				sanidine	Hatch	-112.4778	37.6444	USGS (Mehnert)	Sable and Maldonado (1995b)	also see Kurlich and Anderson (1997)
Limerock Canyon Fm.	Tl	LT-4-89	31	21.5 ± 0.6				biotite	Hatch	-112.4778	37.6444	USGS (Mehnert)	Sable and Maldonado (1995b)	also see Kurlich and Anderson (1997)
Long Flat	Qblf	LEA71SS2				0.60 ± 0.25		whole rock	Brian Head	-112.7500	37.6600	NIGL	Stowell (2006)	location imprecise

Table 3. *continued*

Lava Flow or Formation	Map Symbol	Sample number	Map #	K-Ar age (Ma)	K-Ar age (Ma) (corrected)	$^{40}Ar/^{39}Ar$ age (Ma)	U-Pb age (Ma)	Mineral	7.5' Quadrangle	Longitude	Latitude	Lab used	Reference	Comments
Mount Dutton Fm.	Td	R-5	21	25.1 ± 0.7	25.8 ± 0.7			whole rock	Little Creek Peak	-112.5783	37.9517	Ohio State University	Fleck and others (1975)	
Pine Spring Knoll		C-311-34		1.06 ± 0.28				whole rock	Cedar Mountain	-113.1028	37.5625	USGS	Anderson and Mehnert (1979)	vent (source) uncertain; overlies Ta
post-Claron tuff		R-4		31.1 ± 0.5	31.9 ± 0.5			biotite	Burnt Peak	-112.5917	38.0067	Ohio State University	Fleck and others (1975)	
Red Canyon	Qbrc	pan-28	22	0.56 ± 0.07				whole rock	Wilson Peak	-112.3333	37.7447	USGS	Best and others (1980)	also used labs at BYU and UofA - Tucson
Red Canyon	Qbrc	SF-2	23			0.49 ± 0.04**		whole rock	Casto Canyon	-112.3292	37.7544	NMGRL	Lund and others (2008)	footwall; maximum age
Red Canyon	Qbrc	SF-6	50			0.51 ± 0.01**		whole rock	Wilson Peak	-112.3284	37.7525	NMGRL	Lund and others (2008)	hanging wall
Red Hills	Qbrh	C342-5A	24	1.28 ± 0.4				whole rock	Parowan Gap	-112.9875	37.8833	USGS	Anderson and Mehnert (1979)	reported as 1.3 ± 0.3 (Best and others, 1980)
Rock Canyon	Tbrc	SF-4	26			4.94 ± 0.03**		whole rock	Wilson Peak	-112.3089	37.7241	NMGRL	Lund and others (2008)	footwall
Rock Canyon	Tbrc	SF-7	27			4.98 ± 0.03**		whole rock	Wilson Peak	-112.3304	37.7402	NMGRL	Lund and others (2008)	hanging wall
Rock Canyon	Tbrc	H101508-4	168			5.25 ± 0.03		whole rock	Hatch	-112.4030	37.6751	NIGL	this report	
Spry intrusion	Tis	BRP110410-1					26.24 ± 0.62	zircon	Bull Rush Peak	-112.3597	37.0244	AtoZ	this report	Hwy 89 road cut exposure
Spry intrusion	Tis	84-257		26.1 ± 1.8				plagioclase	Bull Rush Peak	-112.3556	38.0361	USGS	Rowley and others (1994a)	Hwy 89 road cut exposure
Summit	Qbs	C-1730-8A	10	0.94 ± 0.14				whole rock	Summit	-112.9592	37.7700	USGS	Anderson and Mehnert (1979)	
Summit	Qbs	C-1730-9	11	1.00 ± 0.16				whole rock	Summit	-112.9597	37.7733	USGS	Anderson and Mehnert (1979)	incorrectly listed as 112º 58' 35' '

Table 3. *continued*

Lava Flow or Formation	Map Symbol	Sample number	Map #	K-Ar age (Ma)	K-Ar age (Ma) (corrected)	$^{40}Ar/^{39}Ar$ age (Ma)	U-Pb age (Ma)	Mineral	7.5' Quadrangle	Longitude	Latitude	Lab used	Reference	Comments
volcanic rocks of Bull Rush Creek - dike		84-259		26.4 ± 3.2				plagioclase	Bull Rush Peak	-112.3236	38.0319	USGS	Rowley and others (1994a)	derived from Spry intrusion
volcanic rocks of Bull Rush Creek - dike		84-259		29.4 ± 2.2				hornblende	Bull Rush Peak	-112.3236	38.0319	USGS	Rowley and others (1994a)	derived from Spry intrusion
Wah Wah Springs Fm.	Tnw	USA60A	28	30.4 ± 1.1				hornblende	Brian Head	-112.7967	37.6811	USGS (Mehnert)	Rowley and others (1994a)	
Wah Wah Springs Fm.	Tnw	USA60A	36	30.4 ± 3.1				biotite	Brian Head	-112.7967	37.6811	USGS (Mehnert)	Rowley and others (1994a)	
Wah Wah Springs Fm.	Tnw	R-25		28.7 ± 0.5				biotite	Brian Head	-112.7667	37.7067	Ohio State University	Fleck and others (1975)	
Wah Wah Springs Fm.	Tnw	USA60B	37	29.1 ± 1.0				hornblende	Brian Head	-112.7967	37.6811	USGS (Mehnert)	Rowley and others (1994a)	
Wah Wah Springs Fm.	Tnw	USA60B	38	32.4 ± 3.4				biotite	Brian Head	-112.7967	37.6811	USGS (Mehnert)	Rowley and others (1994a)	
Water Canyon	Qbw	R-29	29	0.44 ± 0.04	0.45 ± 0.04			whole rock	Parowan	-112.7733	37.8633	Ohio State University	Fleck and others (1975)	
Wood Knoll	Qbwk	CCB	30	0.63 ± 0.10				whole rock	Flanigan Arch	-112.9339	37.6350	NMGRL	Lund and others (2007)	approximate location

NOTES:

Map Symbol is the symbol used in this report

$^{40}Ar/^{39}Ar$ age is plateau age unless otherwise noted

age uncertainity = 2 standard deviations

Pre-1976 K-Ar ages corrected according to Dalrymple (1979)

* = low confidence

** = weighted mean age

*** = isochron age

NMGRL = New Mexico Geochronology Research Laboratory

NIGL = Nevada Isotope Geochronology Laboratory

AtoZ = Apatite to Zircon, Inc., Viola, Idaho

Whole rock means groundmass concentrate

Longitude and latitude of non-UGS samples may be slightly off due to projection uncertainties.

Figure 46. *Exceptional exposures of the Brian Head Formation just north of Hancock Canyon on the southwest flank of the Sevier Plateau, the locus of studies by Mike Wizevich (Central Connecticut State University) and his students. Here, Brian Head strata are divisible into four parts: (1) a basal variegated unit (below the hikers and so not visible), (2) a lower light-gray, fine-grained volcaniclastic unit, which includes a thick, bluish-gray bentonitic mudstone at its base (Tbh_1), (3) a distinctive red-green-gray banded, fine-grained volcaniclastic unit (Tbh_2), and (4) a thick, upper volcaniclastic unit (Tbh_3). Here, Brian Head strata are capped by Mount Dutton Formation volcanic mudflow deposits (Td). Inset shows bentonitic ash bed at base of Tbh_1, which yielded a U-Pb age on zircon of 36.51 ± 1.69 Ma.*

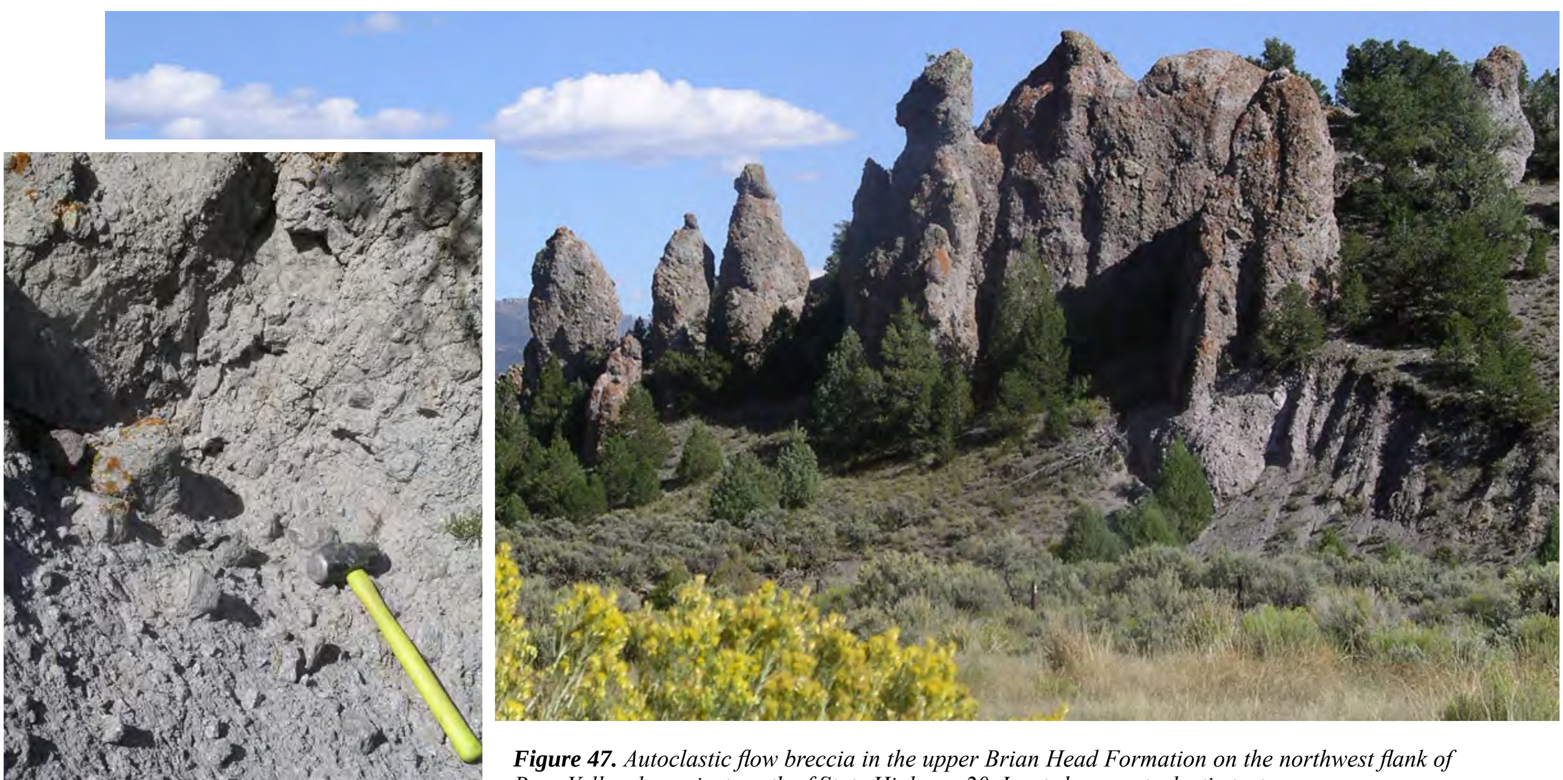


Figure 47. *Autoclastic flow breccia in the upper Brian Head Formation on the northwest flank of Bear Valley dome, just south of State Highway 20. Inset shows autoclastic texture.*

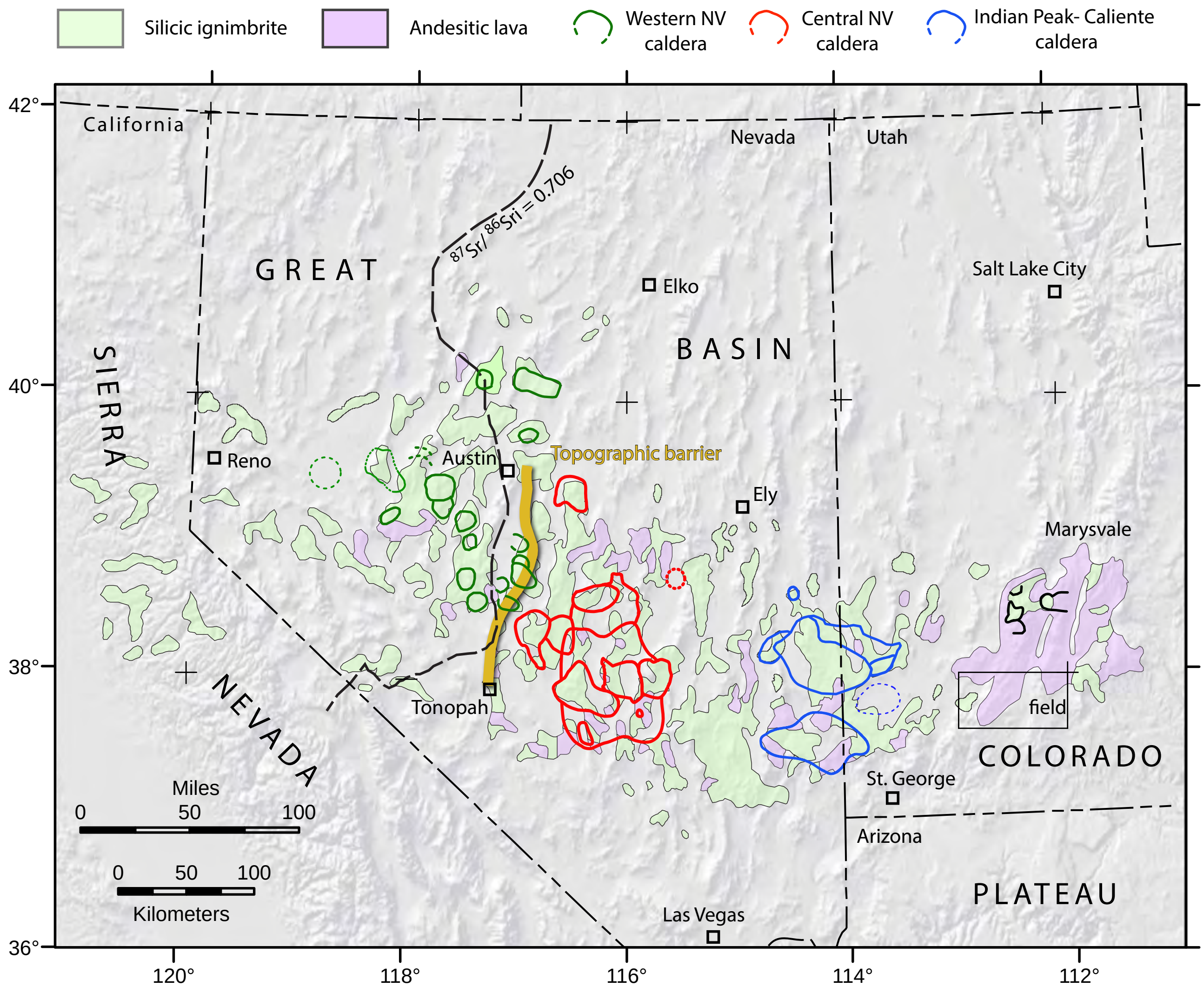


Figure 48. *Southern Great Basin ash-flow tuff province that resulted from the middle Cenozoic (36 to 18 Ma) "ignimbrite flare-up" of Best and others (2009, 2013). This figure does not show pre-36 Ma ash-flow tuffs and calderas from more northern igneous belts, nor does it show post-18 Ma tuffs and calderas from more southern igneous belts. Several ash-flow tuffs (ignimbrites) from the Indian Peak and Caliente caldera complexes (blue line) on the Utah-Nevada border spread eastward into southwest Utah and as far east as this map area. The coeval Marysvale volcanic field (eastern purple area) and calderas (black) are shown to emphasize the contrasting dominance of andesitic lavas over silicic ignimbrites, but Best and others (2009, 2013) did not consider the Marysvale field to be part of the ignimbrite province. The western edge of Precambrian continental basement is indicated by the dashed $^{87}Sr/^{86}Sr = 0.706$ line. The thick yellow line marks the western edge of the Great Basin altiplano; tuffs erupted from western Nevada calderas flowed mostly west down the west flank of the altiplano and are now preserved in exhumed paleovalleys across today's Sierra Nevada Mountains. Box shows location of Panguitch 30' x 60' quadrangle. From Best and others (2013).*

(Siletzia) at the northwestern margin of North America about 48 million years ago, jamming the subduction zone and leading to the formation of the modern Cascadia subduction zone and volcanic arc west of the accreted block. The westward jump in the subduction zone caused a tear in the subducted Farallon plate at the southern margin of the accreted terrain, opening a window in the relatively cold and dense oceanic crust through which relatively hot asthenosphere ascended, feeding the magmatic flare-up. Hydration of the lithosphere from subduction of wet oceanic crust is an important component of the magmatic flare-up, acting as a flux that lowered melting points, creating less dense minerals and thus less dense, more buoyant crust.

Best and others (2013) provided a comprehensive summary of the Great Basin ash-flow tuff province of Nevada and western Utah, where, from about 36 to 18 million years ago, more than 200 large eruptions from 42 calderas resulted in more than 16,500 cubic miles (70,000 km^3) of tuff deposited over the landscape (see figure 48). In the Indian Peak and Caliente caldera complexes in the eastern part of the tuff province, more than 50 large eruptions produced an estimated 7600 cubic miles (32,000 km^3) of ash-flow tuffs now spread over an area of 15,000 square miles (63,000 km^2) in east-central Nevada and southwestern Utah (this volume is roughly enough to fill the Grand Canyon nearly eight times over). Nine of those eruptions are popularly know as "super eruptions," each hav-

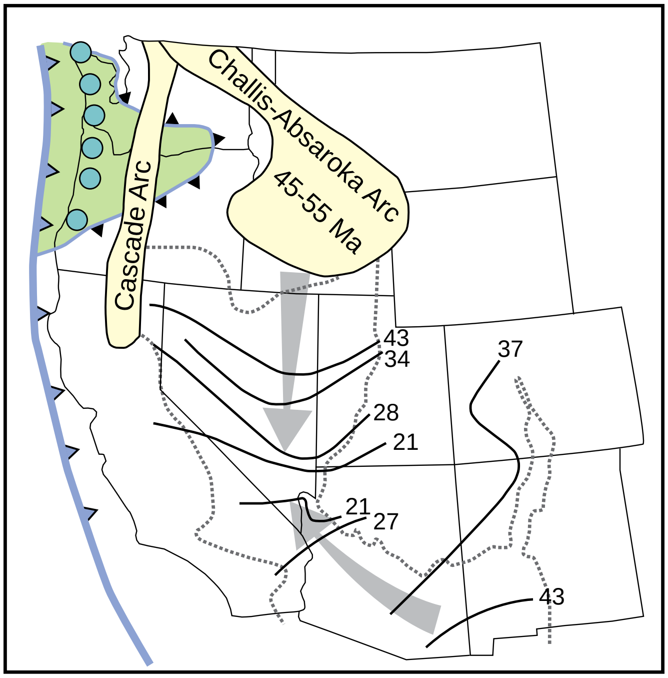


Figure 49. Progression of major volcanic activity in the western U.S. from 55 to 20 Ma. Green area shows oceanic Siletzia terrane with seamounts (blue dots), which accreted to western North America about 48 Ma. Following accretion, the subduction zone (blue lines) and arc-related volcanism (yellow areas) jumped west to the Cascadia subduction zone and the Cascade arc. This initiated "ignimbrite flare-up," which propagated southward through the Basin and Range, as shown by lines with ages of initial volcanism. From Humphreys (2009).

ing ejected more than 240 cubic miles (1000 km^3) of rock, including the Wah Wah Springs, Lund, Isom, and Leach Canyon Formations in this map area. In addition to ash-flow tuffs, these eruptions produced voluminous ash falls—fine-grained ash fall from the Wah Wah Springs eruption, for example, is recognized in western Nebraska (Best and others, 2013). Over time, repeated eruptions of the Indian Peak and Caliente caldera complexes produced ash-flow tuffs that filled topographic low areas in the Great Basin altiplano.

Explosive volcanic eruptions in the Great Basin ash-flow tuff province lasted for about 18 million years. The Indian Peak caldera complex is known for having first erupted enormous volumes of phenocryst-rich dacites over a span of a few million years from about 31 to 29 million years ago, followed by phenocryst-poor trachydacites from 28 to 24 million years ago. The phenocryst assemblage of each phase shows that the dacites equilibrated at depths of about 4 to 5.5 miles (7–9 km) in the upper crust, whereas the trachydacites equilibrated at mid-crustal levels of about 18 miles (30 km).

The tuff province is so called because of its preponderance of ash-flow tuffs over lava flows and volcanic mudflow breccia. Ash-flow tuffs are the deposits of pyroclastic flows, density currents of hot volcanic rock, ash, and gases derived from explosive volcanic eruptions. Pyroclastic flows can travel more than 100 miles (160 km) across the landscape, filling valleys that radiate away from volcanic highlands. One interesting and very useful characteristic of ash-flow tuffs, noted by Mackin (1960), is that ash-flow tuffs are emplaced in a geological instant over broad areas, and thus serve as important time horizons for correlating rock formations and understanding structural development of the region (showing among other things that little extension accompanied the "ignimbrite flare-up" and that significant basin-range extension did not begin in southwest Utah until about 20 million years ago). Best and others (2013) summarized how our understanding of these ash-flow tuffs has evolved, beginning in the 1950s with J. Hoover Mackin who first realized that they were indeed the products of enormous catastrophic eruptions—ash-flow tuffs—not simply lava flows. Today,

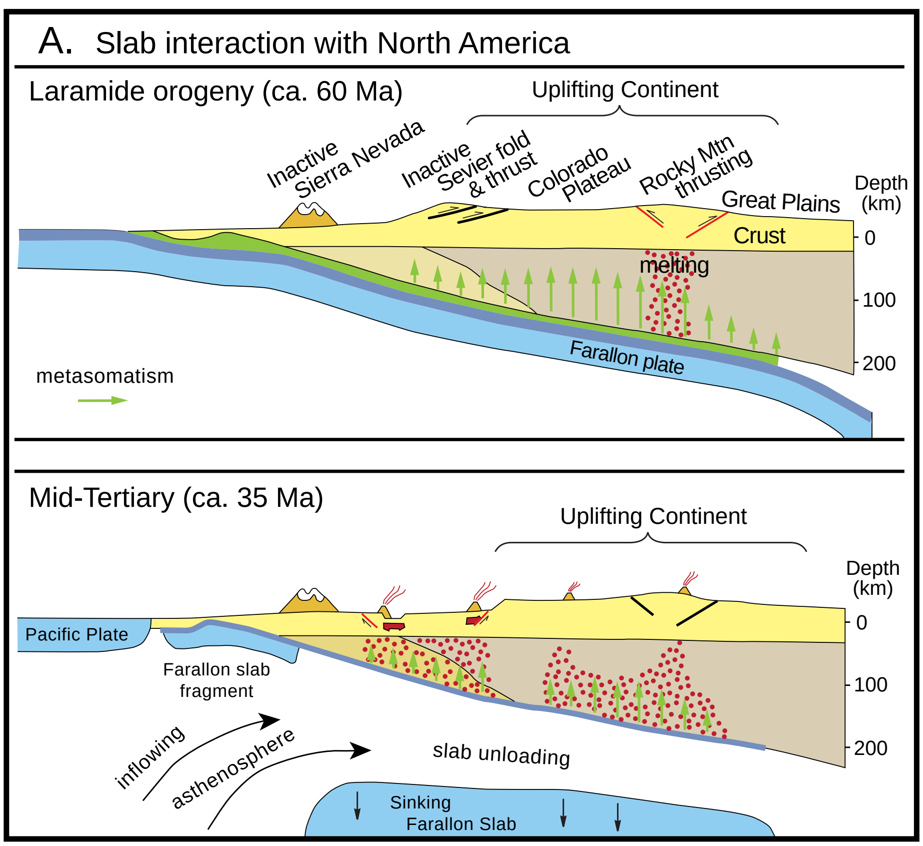


Figure 50. Model for flat-slab subduction and initiation of Laramide orogeny. The infusion of water (green arrows) from the subducting slab caused melting of the basal lithosphere. As the slab rolled back or detached, hot asthenosphere came in contact with the lower crust, initiating the "ignimbrite flare-up." From Humphreys (2009).

the calderas themselves are recognizable only through mapping of stratigraphic and structural relations between caldera in-fill and outflow deposits—20 million years of subsequent basin-range extension, erosion, and burial under intervening basins makes the calderas all but invisible in the modern landscape.

Outflow tuffs may consist of a simple cooling unit or multiple cooling units and range from a few inches to more than 1000 feet (300 m) thick depending on the size of the eruption, topography of the pre-eruption landscape, and distance from the caldera source. The Panguitch 30′ x 60′ quadrangle lies more than 60 miles (100 km) east of the Indian Peak and Caliente caldera complexes, so here we only have relatively distal (outflow) parts of several ash-flow tuffs. These include the crystal-rich dacites of the 30.5 Ma Wah Wah Springs Formation and 27.9 Ma Lund Formation, the densely welded 26 to 27 Ma trachydacite of the Isom Formation, and the distinctive 23.8 Ma rhyolite of the Leach Canyon Formation, all found today across parts of the northern and central Markagunt Plateau; Brian Head peak provides spectacular exposures of the latter two tuffs. Outcrops of two younger regional ash-flow tuffs, the 22.8 Ma Bauers Tuff Member of the Condor Canyon Formation and the 22 Ma Harmony Hills Tuff, are also locally present in the map area, as are several ash-flow tuffs erupted from calderas in the southern Marysvale volcanic field.

Because ash-flow tuffs tend to be confined to paleovalleys far from their source area, the distal end of many ash-flow tuffs behave geomorphically like basaltic lava flows—a resistant volcanic rock that buried former stream valleys eroded into commonly less resistant rocks of the valley walls. Topographic inversion of these ash-flow tuffs naturally ensues as it does for the younger basaltic lava flows. The distribution of regional ash-flow tuffs on the Markagunt and Sevier Plateaus suggests that they may have been controlled by broad, east-trending paleovalleys. It may be that the Wah Wah Springs Formation blocked an east-trending broad paleovalley in the northern part of the map area, and subsequent drainage was re-established in non-resistant Brian Head strata along its southern outcrop belt. Subsequent ash-flow tuffs were thus channeled into this new drainage. The southern depositional limits of both Isom and Leach Canyon ash-flow tuffs were apparently controlled by a paleohigh of Brian Head strata in the vicinity of what is now Cedar Breaks National Monument.

The Wah Wah Springs Formation is only present across the northern part of the map area. However, displaced Wah Wah Springs strata are present as far south as the Brian Head area as part of the upper plate of the Markagunt gravity slide, described below. It is thickest in the northern Red Hills at Jackrabbit Mountain (about 200 feet [60 m] thick), but is typically about 40 to 50 feet (12–15 m) thick in the northern Markagunt and southern Sevier Plateaus. The Wah Wah Springs Formation erupted from the Indian Peak caldera, the first of several super eruptions that rocked the Indian Peak caldera complex in the early Oligocene. Today, the Wah Wah Springs covers at least 8500 square miles (22,000 km^2) with an estimated volume of as much as about 720 cubic miles (3000 km^3) (Best and others, 1989a) (for comparison, the volume of erupted products from the 1980 Mount St. Helens eruptions was about 1 cubic mile [4.2 km^3]). The Wah Wah Springs Formation is a crystal-rich dacite that contains about 40% phenocrysts of plagioclase, hornblende, and biotite (plus minor quartz, Fe-Ti oxides and sanidine); it exhibits reversed magnetic polarity (Best and Grant, 1987). At Jackrabbit Mountain and in the northern Red Hills, the Wah Wah Springs is overlain by the 27.9 Ma Lund Formation, a petrographically similar but normally magnetized crystal-rich dacitic ash-flow tuff of similar volume; together, the two units form the Needles Range Group in this part of Utah. Both the Wah Wah and Lund are chemically and petrographically similar to the Three Creeks Tuff Member of the Bullion Canyon Volcanics, which erupted about 27 million years ago from the Three Creeks caldera in the northwest part of the Marysvale volcanic field (Steven and others, 1979; Steven, 1981). In the southeastern Sevier Plateau, the Three Creeks Tuff locally rests on the Wah Wah Springs, making it difficult to distinguish the two formations. Three Creeks Tuff, however, is typically autobrecciated in the Panguitch 30′ x 60′ quadrangle. The Wah Wah Springs Formation is also petrographically similar to the 20 Ma Harmony Hills Tuff, but their major- and trace-element chemistry is sufficiently distinct to differentiate the two dacitic ash-flow tuffs; additionally, the Wah Wah Springs is unique among Great Basin ash-flow tuffs for its abundance of hornblende over biotite. Excellent exposures of the Wah Wah Springs are present at Bear Valley Dome just north of State Highway 20, and good exposures of both Wah Wah and Three Creeks are present along Deer Creek in the northeast corner of the map area (figure 51).

The 26 to 27 million year old, crystal-poor, densely welded, trachydacitic ash-flow tuff of the Isom Formation overlies Wah Wah Springs Formation at Jackrabbit Mountain and northern Markagunt Plateau, but it is absent from the Sevier Plateau. On the Markagunt Plateau, Isom is present as far south as Brian Head peak and is spectacularly exposed for many miles along Black Ledge where it is about 350 feet (110 m) thick (figure 52). The distribution of the thickest parts of the Isom suggests that it too may have filled a broad, east-trending paleovalley, one that was cut into relatively non-resistant Brian Head strata south of the Wah Wah Springs outcrop belt. That it did not reach as far east as today's Sevier Plateau is likely due to a topographic barrier developed in response to early activity in the Marysvale volcanic field, including the Spry intrusion and its eruptive products.

The Isom Formation is unusual in that it was so hot when erupted that it flowed like a lava flow during its final stages of emplacement. Many Isom outcrops reveal secondary flow characteristics, including flow breccias, contorted flow layering (figure 53), and linear vesicles, such that the unit was considered a lava flow until Mackin (1960) mapped its widespread distribution (300 cubic miles [1300 km^3] today spread over an area of 9500 square miles [25,000 km^2] [Best and others, 1989a]) and found evidence of glass shards, thus showing

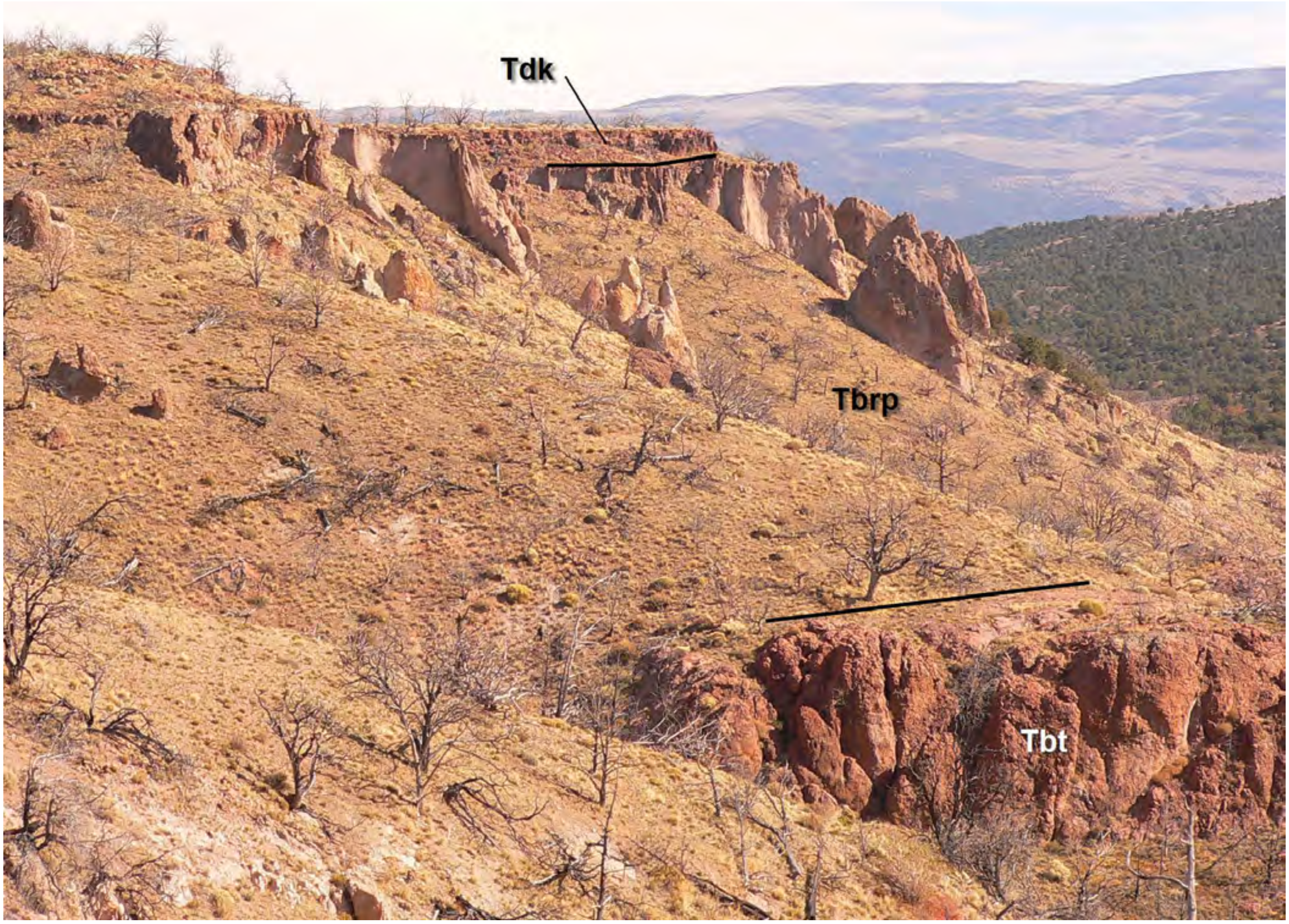


Figure 51. *Volcanic strata of the Deer Creek drainage on the east flank of the Sevier Plateau. Here, the 26 Ma Three Creeks Tuff Member (Tbt) of the Bullion Canyon Volcanics overlies the 30 Ma Wah Wah Spings Formation at the bottom of the drainage (just out of view); locally, a few tens of feet of Mount Dutton mudflow breccia separates the two formations. The volcanic rocks of Bull Rush Peak (Tbrp) consist of autoclastic flow breccia, locally with a basal vitrophyre, whose clasts and matrix are identical to those of the Spry intrusion; they locally overlie thin lenses of Mount Dutton mudflow breccia. The Kingston Canyon Tuff Member (Tdk) of the Mount Dutton Formation forms a prominent ledge along the north side of the canyon and is in turn overlain by more mudflow breccias of the Mount Dutton alluvial facies (just out of view). View is towards the east; the Table Cliff Plateau is in the distance.*

its true ash-flow tuff nature. For that reason it is commonly referred to as a tufflava or a rheomorphic ignimbrite, an ash-flow tuff that was sufficiently hot to move with laminar flow as a coherent ductile mass—see, for example, Anderson and Rowley (1975, 2002), Andrews and Branney (2005), and Geissman and others (2010). Isom is spectacularly exposed at Brian Head peak where the lower part of the formation is classic tufflava about 80 feet (24 m) thick and the upper part is a flow breccia 60 to 90 feet (18–27 m) thick. East of Yankee Meadows Reservoir, the lower part of the Isom is unusual in that welded Isom ash-flow tuff overlies as much as 30 feet (9 m) of coarse sandstone with angular Isom pebbles and planar and low-angle cross-stratification that in turn overlies calcite-cemented Isom breccia (figure 54). The breccia and sandstone appear to represent an early phase of the Isom eruption.

The eruption of the Isom Formation was followed by the eruption of three regionally distinctive ash-flow tuffs: the 23.8 Ma Leach Canyon Formation, the 22.8 Ma Bauers Tuff Member of the Condor Canyon Formation, and the 22.0 Ma Harmony Hills Tuff, all part of the Quichapa Group. Only the Leach Canyon is preserved as far east as Panguitch Lake; the Bauers and Harmony Hills tuffs are restricted to the west part of the Markagunt Plateau. The distribution of these ash-flow tuffs suggests that a west-facing topographic escarpment associated with early basin-range extension may have impeded eastward distribution of the latter two tuffs. Both the Leach and Bauers erupted from the Caliente caldera, whereas the Harmony Hills Tuff likely erupted from the eastern Bull Valley Mountains (Williams, 1967; Rowley and others, 1995).

The Leach Canyon Formation is a distinctive, light-pink to white, unwelded to poorly welded, crystal-rich rhyolite tuff that contains abundant white or light-pink collapsed pumice fragments and several percent lithic clasts, many of which are reddish brown; phenocrysts of plagioclase, slightly less but

***Figure 52.** Black Ledge near Yankee Datum (elevation 10,172 feet) just east of Yankee Meadows Reservoir. Here, Black Ledge appears to consist of a single cooling unit of the Isom Formation, in addition to an unusual underlying pebbly sandstone and breccia phase (see figure 54). View is towards the southwest.*

***Figure 53.** Contorted flow layering in the Isom Formation at Brian Head peak, an example of the rheomorphic nature of the Isom Formation.*

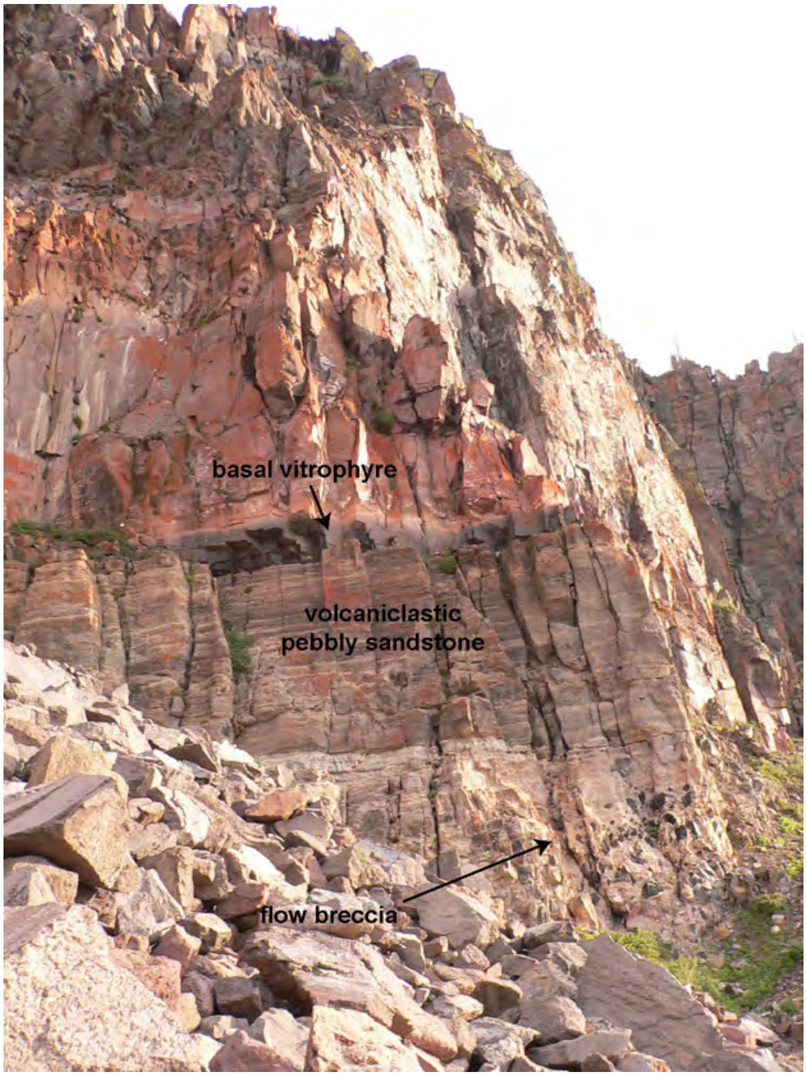


Figure 54. *The lower part of the Isom Formation, just east of Yankee Meadows Reservoir, contains two unusual components that we have not observed elsewhere. First, the base of the Isom is several tens of feet of flow breccia whose angular Isom clasts are completely cemented by calcite, and second, this breccia is overlain by as much as 30 feet (9 m) of volcaniclastic pebbly sandstone with planar and low-angle cross-stratification and whose clasts are angular to subangular fragments of Isom Formation. This volcaniclastic unit is overlain by a 3-foot-thick (1 m) vitrophyre and densely welded Isom that is reddish-brown in its lower part and that grades upward into typical dark-gray Isom.*

subequal amounts of quartz and sanidine, and minor biotite, hornblende, Fe-Ti oxides, and a trace of pyroxene make up 25 to 35% of the rock. It is typically about 100 feet (30 m) thick and forms the resistant caprock of Brian Head peak and the southern part of Black Ledge, is exposed eastward nearly to the Panguitch Lake area, and is also exposed in fault blocks at the west edge of the Markagunt Plateau and in the northern Red Hills. Some of the best exposures are at Brian Head peak, where it consists of four parts, including a basal ash-fall tuff and vitrophyre that are rarely exposed elsewhere (figure 55). The Leach Canyon Formation unconformably overlies the Isom Formation at Brian Head peak and the southern part of Black Ledge. North of Castle Valley and at Prince Mountain, however, the Leach Canyon unconformably overlies Brian Head strata. This distribution suggests that the Prince Mountain-Castle Valley area was a paleohigh of Brian Head strata during Isom time, and that, once the resistant Isom was in place, this paleohigh was preferentially eroded to form a broad, east-trending stream valley in which the Leach Canyon Formation accumulated. The Leach Canyon is not present in the Panguitch 30′ x 60′ quadrangle on the Markagunt Plateau north of Clear Creek, a west-northwest-trending tributary to Panguitch Lake.

The Leach Canyon Formation is petrographically and chemically similar to the Haycock Mountain Tuff, which led Sable and Maldonado (1997a) to suggest that the latter is a distal facies of Leach Canyon. While it is true that the two formations are not reliably distinguishable on the basis of their major- and trace-element chemistry, the Haycock Mountain Tuff is typically even less welded than the Leach Canyon and contains conspicuous black lithic fragments, unlike the reddish-brown lithic fragments of the Leach Canyon, facts previously noted by Anderson (1993), Rowley and others (1994a), and Hatfield and others (2010). Detailed mapping of the Panguitch Lake and Haycock Mountain 7.5′ quadrangles (Biek and others, 2014a, 2014b) reconfirms that these are indeed two different units. The Leach Canyon Formation can be traced in continuous outcrop from Brian Head peak northward to the head of Bunker Creek and then east to the east end of Prince Mountain just west of Panguitch Lake. It is structurally overlain by the Markagunt gravity slide, which here consists mostly of the Isom Formation. Samples from the south side of Prince Mountain yielded K-Ar ages of 22.8 ± 1.1 Ma (biotite) and 24.8 ± 1.0 Ma (sanidine) (Rowley and others, 1994a, sample 89USa-1a, which they mistakenly called Haycock Mountain Tuff) and a duplicate K-Ar age of 24.3 ± 1.0 Ma (sanidine) as well as an $^{40}Ar/^{39}Ar$ age of 23.86 ± 0.26 Ma (biotite) (Sable and Maldonado, 1997a, on the same sample 89USa-1a). The Leach Canyon Formation is widely agreed to be about 23.8 Ma (Best and others, 1993; Rowley and others, 1995). As noted by Hatfield and others (2010), both Rowley and others (1994a) and Sable and Maldonado (1997a) misinterpreted this tuff to be the Haycock Mountain Tuff, which yielded a slightly younger $^{40}Ar/^{39}Ar$ age of 22.75 ± 0.12 Ma (sanidine) at its type section one mile (1.6 km) northeast of Panguitch Lake (Sable, unpublished data, 1996; see also Hatfield and others, 2010; our U-Pb age on zircon for the Haycock Mountain Tuff is slightly younger yet at 21.63 ± 0.73 Ma). The facts that the tuff atop Prince Mountain yielded an age analytically indistinguishable from the Leach Canyon Formation, and that it can be traced contin-

Figure 55. *Leach Canyon Formation on the south side of Brian Head peak (see figure 43), the most complete section of this formation exposed on the Markagunt Plateau. Here, the classic three-part section of an ash-flow tuff is exposed, including an unwelded basal air-fall tuff, a thick vitrophyre, and moderately welded ash-flow tuff that caps Brian Head peak. A flow breccia of the Isom Formation (Ti) is present in the lower left corner of the photograph; its contact with the Leach Canyon is concealed by talus.*

uously to classic Leach Canyon outcrops at Brian Head peak, are strong evidence that it is the Leach Canyon Formation and not the Haycock Mountain Tuff.

In the map area, the Bauers Tuff Member is restricted to the west flank of the Markagunt Plateau. It is a densely welded, crystal poor, rhyolitic ash-flow tuff that superficially resembles the Isom Formation, but it contains minor bitotite that is conspicuously bronze-colored in the upper, vapor-phase part of the tuff. The Bauers underlies the Markagunt gravity slide in Summit Canyon and Parowan Canyon. In the nearby northwest corner of the Brian Head quadrangle, Rowley and others (2013) interpreted poorly exposed Bauers Tuff Member to be in the upper plate of the gravity slide, although it is possible that this Bauers is part of a fault block associated with the Sugarloaf Mountain fault and thus not part of the gravity slide. The crystal-rich, moderately welded, dacitic Harmony Hills Tuff is superficially similar to the Wah Wah Springs Formation, but it is restricted to a narrow band at the west edge of the Markagunt Plateau, south of autochthonous Wah Wah Springs Formation. We discovered that the Harmony Hills Tuff (and boulder gravels eroded from Harmony Hills Tuff) underlies the Markagunt gravity slide and is a key unit in constraining its maximum age, described below.

Marysvale Volcanic Field

The Marysvale volcanic field is one of the major volcanic fields of the southwestern United States. It encompasses the central part of the High Plateaus and adjacent Basin and Range Province, roughly centered on the small town of Circleville (figure 56). The volcanic field lies at the east end of the east-northeast-trending Pioche-Marysvale igneous belt, a zone of extensive volcanism and mineralization above a major batholithic complex (Rowley and others, 1994a; Cunningham and others, 2007). Most igneous activity in the Marysvale volcanic field occurred in the Oligocene and early Miocene, from about 32 to 21 million years ago, as a series of mostly andesitic, calk-alkaline volcanic rocks and related intrusions, followed by bimodal (basalt and high-silica rhyolite) volcanism that accounts for only about 5% of the 2860 cubic miles (12,000 km^3) of igneous rocks in the field (Rowley and others, 1994a; Cunningham and others, 2007). In its heyday, the Marysvale volcanic field was a cluster of towering stratovolcanoes and subordinate calderas, in contrast to the calderas that dominated the west end of the igneous belt; more than 90% of the volume of Marysvale volcanic rocks are lava flows, volcanic mudflow breccia (lahar deposits), or flow breccias, whereas less than 10% consists of ash-flow tuff (Steven and others, 1984; Rowley and others, 1994a; Cunningham and others, 2007).

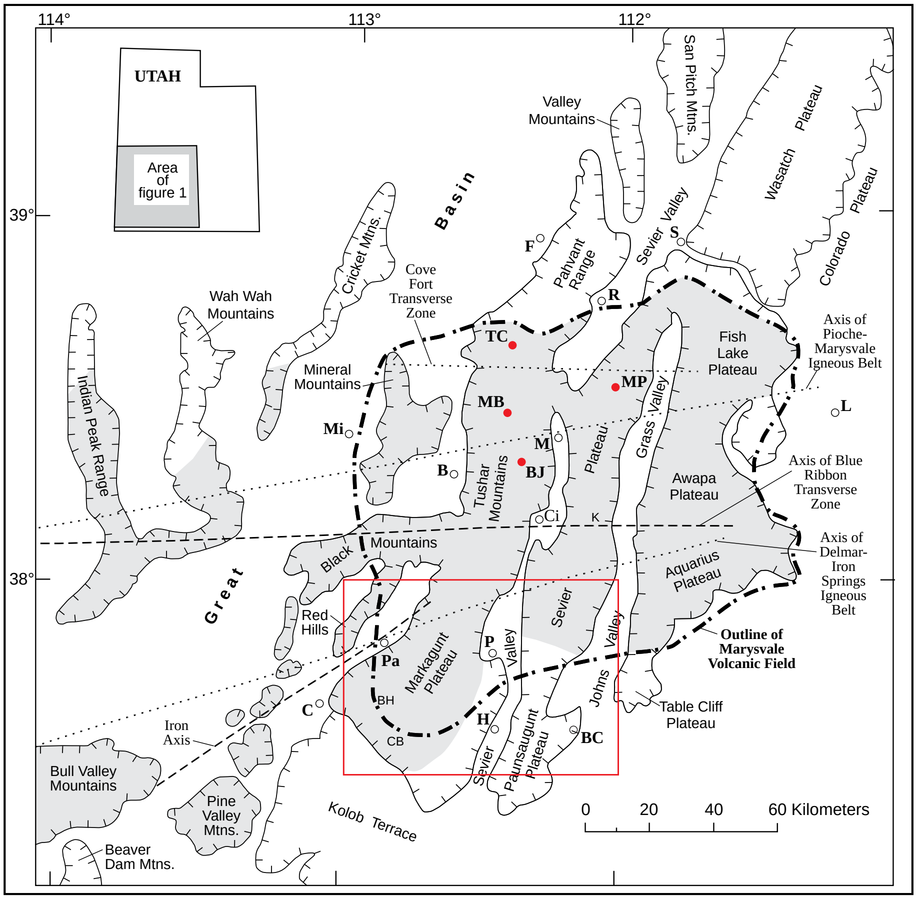


Figure 56. *The Marysvale volcanic field outlined by heavy dashed-dotted line. Red dots show central part of calderas (BJ = Big John, MB = Mount Belknap, TC = Three Creeks, and MP = Monroe Peak). B, Beaver; BC, Bryce Canyon; BH, Brian Head; C, Cedar City; CB, Cedar Breaks; Ci, Circleville; F, Fillmore; H, Hatch; K, Kingston Canyon; L, Loa; M, Marysvale; Mi, Milford; P, Panguitch; Pa, Parowan; R, Richfield; S, Salina). Gray areas are underlain by volcanic rocks. Panguitch 30′ x 60′ quadrangle shown by red rectangle. Modified from Rowley and others (1994a) and Cunningham and others (2007).*

Calc-alkaline igneous sequence: Most Marysvale-derived volcanic rocks in the map area, which encompasses the southwest margin of the volcanic field, belong to the alluvial facies of the Mount Dutton Formation. These are drab gray and brown, monotonous, andesitic to dacitic volcanic mudflow breccia (figure 57) and lesser interbedded volcaniclastic conglomerate and tuffaceous sandstone that are at least 1000 feet (300 m) thick in the map area in the northern Markagunt Plateau (Anderson and Rowley, 1987) and at least 6000 feet (2000 m) thick farther north (Anderson and others, 1990a, 1990b; Rowley and others, 2005). In the northern part of the map area and especially farther north in the Beaver 30′ x 60′ quadrangle, the map unit also contains subordinate lava flows, flow breccia, and minor felsic tuff. Anderson and Rowley (1975) defined the Mount Dutton Formation as consisting of most of the rocks—a volume of at least 1200 cubic miles (5000 km^3)—exposed on the south flank of the Marysvale volcanic field, and divided it into complexly interfingering and cross-cutting vent and alluvial facies derived from clustered stratovolcanoes and dikes. Except for the thin, densely welded trachytic ash-flow tuff of the 25 Ma Kingston Canyon Tuff Member (see figure 51) and unmapped felsic tuff on the west flank of the Sevier Plateau, only those rocks associated with the alluvial facies are present in the map area, where they typically overlie the Bear Valley Formation on the Markagunt Plateau and the Brian Head Formation on the Sevier Plateau.

On the Markagunt Plateau, the Mount Dutton Formation overlies volcaniclastic sandstone, lesser interbedded lava flows, volcanic mudflow breccia, conglomerate, mudstone, and ash-fall and ash-flow tuff beds of the Bear Valley Formation. Anderson and others (1987) described the difficulty of differentiating similar lava flows and mudflow breccias of the Bear Valley and overlying Mount Dutton Formations, and we are uncertain if these two formations are consistently mapped across the northwestern Markagunt Plateau. At its type section northwest of Bear Valley Junction, the Bear Valley Formation is volcaniclastic sandstone with high-angle cross stratification; Anderson (1971) and Anderson and others (1990a, 1990b) suggested that the sand was derived from the south and accumulated in a low-relief basin bounded on the north by an east-trending fault scarp possibly associated with the 26 Ma Spry intrusion, which intruded as a laccolith into the Claron Formation immediately north of the map area (Grant, 1979; Anderson and others, 1987, 1990a).

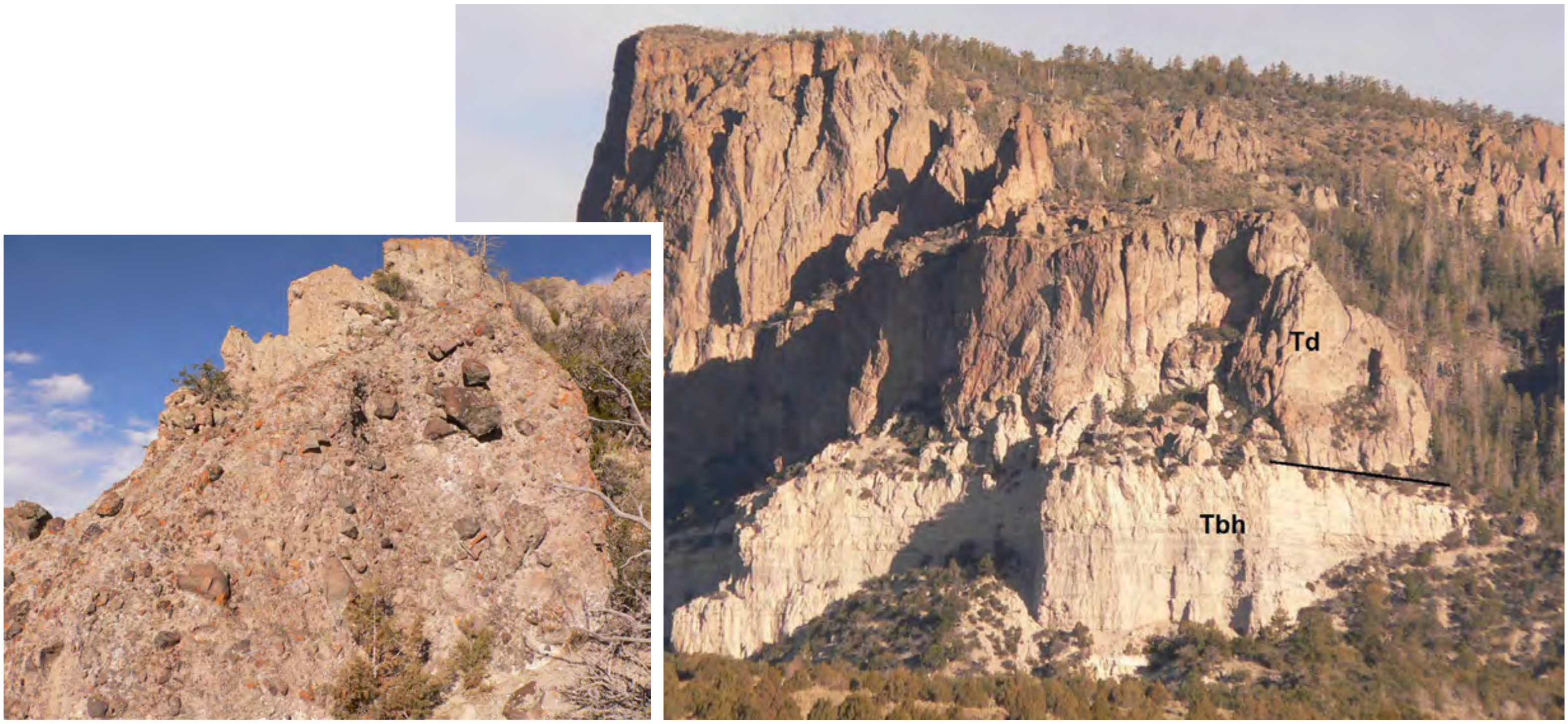


Figure 57. *Casto Bluff at the south end of the Sevier Plateau. On the Sevier Plateau, Mount Dutton volcanic mudflow deposits (Td) overlie light-gray, volcaniclastic sandstone of the Brian Head Formation (Tbh). Inset shows typical, matrix-supported nature of Mount Dutton volcanic mudflows. Immediately to the north, along the southwest flank of the plateau, uppermost Brian Head and basal Mount Dutton strata exhibit shearing and small south-vergent folds, evidence of gravitational collapse of the southern part of the Marysvale volcanic field.*

Like other laccoliths in southwest Utah, the Spry intrusion was emplaced as molten rock from deep within the Earth moved upward into the shallow crust where it spread out to form a mushroom-shaped intrusion (figure 58). Many of these laccoliths produced gravity slides and erupted lava flows and (or) ash-flow tuffs. The Spry intrusion also erupted to produce ash-flow tuff (primarily the Buckskin Breccia), as well as volcanic mudflow breccia and autoclastic flow breccia (volcanic rocks of Bull Rush Peak).

Iron Peak forms the easternmost exposed laccolith of the Iron Axis, a northeast-trending belt of early Miocene calc-alkaline laccoliths and concordant stocks that rose above the roof of an inferred large batholith at about 22 to 20 Ma (Blank and Mackin, 1967; Cook and Hardman, 1967; Rowley, 1998; Rowley and others, 1998). Iron Peak is the second youngest and most mafic of the Iron Axis intrusions and its floor is well exposed in the north canyon wall of Little Creek, northeast of Paragonah, which has incised through the laccolith to reveal numerous feeder dikes that are resistant and so stand as tall fins (Anderson, 1965; Spurney, 1984; Hacker and others, 2007) (figure 59). The laccolith is a 20 Ma gabbro-diorite porphyry, with diorite the dominant phase, composed almost entirely of augite and plagioclase (calcic labradorite) and about 8% opaque oxide minerals, mostly magnetite (Anderson, 1965; Spurney, 1984). Most of the other plutons of the Iron Axis are quartz monzonite porphyry and appear to be partly controlled by northeast-striking, southeast-verging

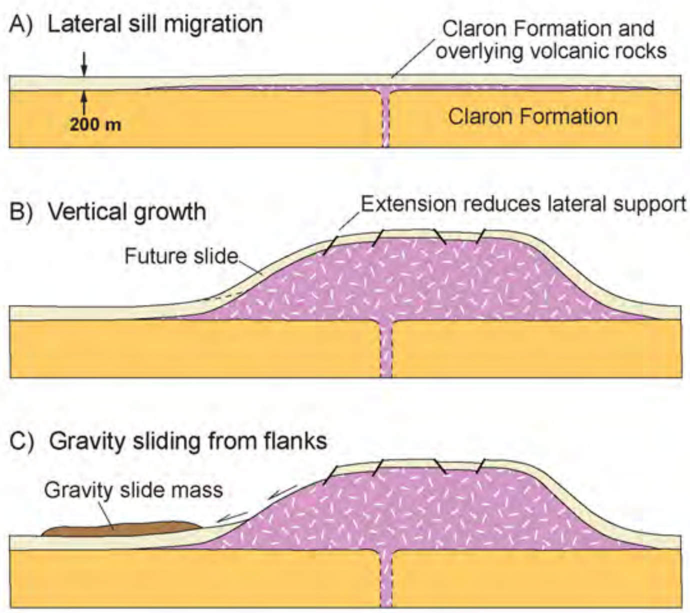


Figure 58. *Schematic diagram illustrating growth of a typical laccolith. (A) Initial lateral migration of a sill within the Claron Formation to its fullest extent at a relatively shallow depth, (B) vertical growth of the laccolith by continued injection of magma, and (C) gravity sliding of oversteepened flanks. Like most laccoliths of the nearby Iron Axis, the Spry intrusion likely produced gravity slides, but if so, they are eroded away or buried by younger Mount Dutton volcanic mudflow deposits. The Spry laccolith erupted to produce the Buckskin Breccia and the volcanic rocks of Bull Rush Creek, the latter as a stratovolcano that once towered over the intrusion. Modified from Willis (2002) and Hacker and others (2002).*

***Figure 59.** Erosion has cut through the floor of the Iron Peak laccolith (Tip), a gabbro-diorite porphyry and the easternmost of the Iron Axis laccolithic intrusions, to reveal a swarm of dikes (inset) trending northeast that are as thick as 25 feet (8 m). The dikes are inferred to be feeder dikes of the laccolith. The Iron Peak laccolith intruded both Brian Head volcaniclastic strata (Tbh) and Bear Valley ash-flow tuff (Tbvt) that are juxtaposed across a pre-Iron Peak-intrusion graben. Tcp, pink member of Claron Formation.*

Sevier-age thrust faults; the plutons were emplaced at shallow depths, mostly within about 1.2 miles (2 km) of the surface (Mackin and others, 1976; Van Kooten, 1988; Hacker and others, 2002, 2007; Rowley and others, 2008). The Iron Peak laccolith was previously mapped as intruding the Brian Head Formation (Spurney, 1984; Maldonado and others, 2011), but our mapping shows that it intruded both Brian Head volcaniclastic strata and Bear Valley ash-flow tuff that are juxtaposed across a pre-Iron Peak-intrusion graben now preserved at the west edge of the Markagunt Plateau. Magnetite veins are common in the laccolith (Spurney, 1984), but are insufficient to have been of economic importance, unlike the nearby Iron Springs mining district west of Cedar City, 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, 2008).

Like the other laccoliths of the Iron Axis, the Iron Peak laccolith probably formed rapidly following a two-stage emplacement process—injection of a sill immediately followed by inflation—at shallow crustal depth of less than 4000 feet (1.2 km) on the basis of stratigraphic reconstructions (Spurney, 1984; see also Hacker and others, 2002, 2007; Willis, 2002). Rapid inflation and doming of most laccoliths of the Iron Axis led to their partial unroofing by gravity sliding, immediately followed by volcanic eruptions (Mackin, 1960; Blank and Mackin, 1967; Hacker and others, 1996, 2002, 2007; Hacker, 1998; Willis, 2002; Rowley and others, 2008). Maldonado and others (1997) and Sable and Maldonado (1997a) suggested that emplacement of the laccolith or nearby intrusive bodies may have been one trigger for formation of the Markagunt Megabreccia, but we now know that the laccolith is too small, is located too far south, and may be too young to possibly have triggered the Markagunt gravity slide as discussed later.

Considerable confusion remains about the extent of possible Iron Peak eruptive products and the Bear Valley Formation northeast of Paragonah. Coauthor Anderson's student John Spurney interpreted exposures immediately east of the Iron Peak laccolith as a peripheral breccia complex (Spurney, 1984), and this area was reinterpreted as older Bear Valley breccia (Maldonado and others, 2011). We tentatively follow Maldonado's lead but note that additional mapping, geochemistry, and age control are needed to sort out the eruptive history of the Iron Peak laccolith. Spurney (1984) also described volcanic rocks of similar composition to the south on the divide between Red Creek and Little Creek canyons, northeast of Paragonah, that we interpret as gravity-slide deposits of lava flows associated with the Iron Peak laccolith.

Immediately east of Upper Bear Valley, strata are domed above an inferred laccolith, and, although not exposed, the intrusive body is likely a diorite or gabbro of early Miocene age (Anderson and others, 1987). Numerous small gravity slide blocks south-southeast of the dome may be related to this intrusion.

Miocene
23 to 5.3 million years ago

In the early Miocene—about 23 million years ago in the Marysvale volcanic field, and about 17 million years ago in extreme southwestern Utah—volcanism began to change from calc-alkaline to bimodal. 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. Rifting is a result of torsion of the North American continent as a result of right-lateral transform shear along the San Andreas transform fault system between the North American and Pacific plates (figure 60) (Dickinson, 2006). The transform plate boundary developed following subduction of the Farallon plate spreading ridge and continues to evolve into a broad zone of distributed deformation eastward to the Walker Lane structural belt near the California-Nevada border. Unlike the intermediate-composition calc-alkaline rocks, 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). Bimodal volcanism was dominant during the Miocene and early Pliocene but continues to the present time. The most visible evidence of bimodal volcanism in this map area are the many basaltic lava flows that dot the landscape; local rhyolitic volcanism in the Panguitch 30′ x 60′ quadrangle is preserved in the largely volcaniclastic Limerock Canyon Formation.

The first products of bimodal volcanism in the Marysvale volcanic field are 23 to 21 Ma potassium-rich mafic volcanic rocks exposed immediately north and northeast of the map area (Rowley and others, 1994b, 2005). These rocks erupted from local vents throughout the volcanic field, but are thickest and most widespread in its southeastern part, particularly on the Aquarius and Awapa Plateaus. The main mass of bimodal rocks, however, is high-silica rhyolite of the 22 to 14 Ma Mount Belknap Volcanics in the west part of the volcanic field. A smaller but highly informative rhyolitic eruptive center is in Kingston Canyon, a deep antecedent canyon that bisects the Sevier Plateau about 15 miles (24 km) north of the map area. There, rhyolitic flows on the plateau, and a lava dome on the canyon's floor, show that the canyon was carved by the East Fork of the Sevier River between about 8 and 5 million years ago, thus signaling the age of the main basin-range faulting in that area (Rowley, 1968; Rowley and others, 1981).

The only rocks of bimodal, rhyolitic affinity in the map area belong to the Limerock Canyon Formation, several hundred feet of mostly white, tuffaceous, volcaniclastic sandstone, pebbly sandstone, gritstone, pebbly conglomerate, mudstone, and minor limestone and ashfall tuff beds. The Limerock Canyon Formation was deposited in fluvial, floodplain, and minor lacustrine environments (Kurlich and Anderson, 1997). It is present only on the east part of the Markagunt Plateau near Hatch, south of the Markagunt gravity slide, where it may be preserved in a subtle basin in front of an inferred blind west-trending thrust fault (the inferred westward continuation of the Rubys Inn thrust fault). Limerock Canyon strata are puzzling for a mapper in that they are similar to but much younger than the Brian Head Formation, and both formations overlie the white member of the Claron For-

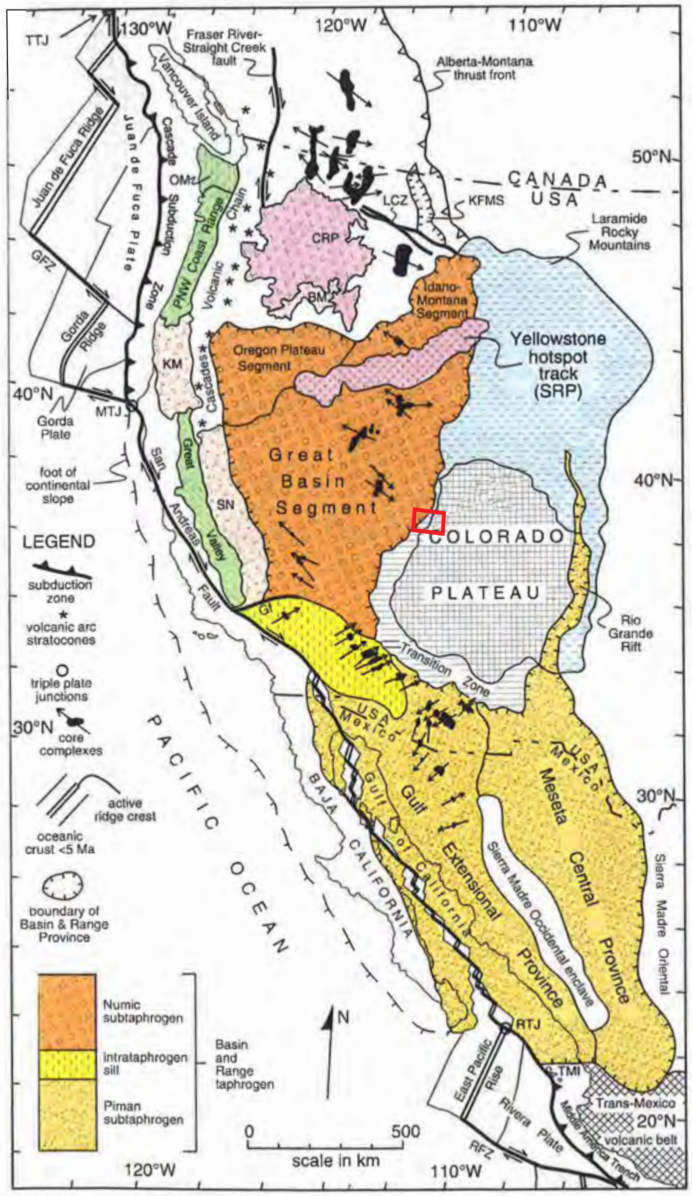


Figure 60. *Major tectonic features of the western Cordillera. Panguitch 30' x 60' quadrangle shown by red box. The term "taphrogen" is a general term for rift phenomena considered to be the first stage in continental rifting and separation of tectonic plates. From Dickinson (2011).*

mation. Without additional radiometric age control, we are unable to consistently distinguish the two formations in the field and so locally map them as a combined unit.

Pliocene and Quaternary
5.3 million years ago to Present

Basaltic Lava Flows

More than 40 basaltic lava flows erupted onto the Markagunt Plateau in the past 5 million years. Most, in fact, erupted within the past 1 million years, and several are so young—thought to be of latest Pleistocene or even Holocene age—that they look like they erupted almost "yesterday." These relatively small-volume, widely scattered, mostly basaltic lava flows and cinder cones lie at the north edge of the Western Grand Canyon basaltic field, which extends across the southwest part of the Colorado Plateau and transition zone with the Basin and Range Province in southwest Utah and northeast Arizona (Hamblin, 1963, 1970, 1987; Best and Brimhall, 1970, 1974; Best and others, 1980; Smith and others, 1999; Johnson and others, 2010). This volcanic field contains hundreds of such flows (figure 61; see also illustration on plate 2) that range in age from Miocene to Holocene; they are synchronous with modern basin-range extension (and thus with our modern topography), which began in southwestern Utah between 23 and 17 million years ago.

The oldest-known basaltic lava flows in the map area are the 5.3 Ma Houston Mountain, Dickinson Hill, and Rock Canyon lava flows, but the undated Sidney Peaks lava flow may be older still. The youngest lava flows are the Dry Valley and Panguitch Lake lava flows south of Panguitch Lake, and though undated, they are likely younger than the 32,000-year-old Santa Clara basaltic lava flow, the youngest dated lava flow in southwest Utah (Willis and others, 2006; Biek and others, 2009). Red-hot lava flows, an integral part of the Southern Paiute legend "How the whistler [bird] and badger got their homes," were suggested to relate to the Panguitch Lake-area lava flows (Palmer, 1957; Southern Paiutes lived in southwest Utah beginning about A.D. 1100 [Canaday, 2001]), but it is unlikely the flows are this young. Schulman (1956) briefly reported on 850- to 950-year-old juniper (*Juniperus scopulorum*) trees growing on young lava flows, thus showing that the lava flows are at least that old but still could be many thousands of years older (Schulman's lava flows are apparently near Panguitch Lake although definitive sample locations are unavailable; see his samples BRY 2104 and BRY 2110, table "Overage drought conifers"). We are currently attempting to date one of the Panguitch Lake lava flows.

Lava flows in the map area 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. Several lava flows, including the Duck Creek and Bowers Knoll lava flows, contain open lava tubes; the best known is Mammoth Cave (6 miles [10 km] northeast of Duck Creek Village) (figure 62). Most of the lava flows are dark gray and fine grained, and contain small olivine phenocrysts and common crystal clusters of olivine, plagioclase, and clinopyroxene. With few exceptions, these lava flows are difficult to distinguish by hand sample alone. They are distinguished for this geologic map by detailed geologic mapping, major- and trace-element geochemistry, and isotopic ages.

Basaltic lava flows on the Markagunt Plateau provide stunning yet little-known examples of inverted topography—former stream valleys that now stand as sinuous ridges that meander down the plateau's gently east-tilted surface. Because the plateau contains such a wide range of young and old lava flows, it is adorned with inverted valleys representing each stage in the evolution of these enigmatic features (figures 63 and 64). The Markagunt Plateau exhibits (1) deeply dissected, isolated lava flow remnants that bear little relation to modern drainages; (2) long, sinuous, classic inverted valleys that rival those of the St. George area (with its basalt-capped ridges that stand in stark contrast to underlying Jurassic and Triassic red beds, as first described in detail by Hamblin [1963, 1970, 1987] and Hamblin and others [1981]); and (3) modern drainages plugged with basaltic lava, whose streams are only now beginning to erode the margins of the flows. The distal ends of the Asay Knoll (Qbak), Bowers Knoll (Qbbk), and Coopers Knoll (Qbck) lava flows form classic inverted valleys. Not only is the Markagunt Plateau a showcase for scenic volcanic geology, the volcanic rocks themselves allow geologists to reconstruct the plateau's structural and geomorphic evolution, discussed in the structural geology section of this report.

The Panguitch Lake and Dry Valley lava flows are the youngest lava flows in southwest Utah (figures 65 and 66). The Panguitch Lake and Miller Knoll flows blocked Bunker Creek drainage, a former tributary of the unnamed main-stem creek of Rock Canyon. Rock Canyon is now occupied by a conspicuously underfit stream whose former headwaters are now diverted into the Panguitch Creek drainage (figure 67). The expansive meadow of Blue Spring Valley formed as sediment accumulated upstream of the Panguitch Lake and Miller Knoll lava flows. Bunker Creek, having carved a new channel through non-resistant volcaniclastic sandstone and mudstone of the Brian Head Formation, now feeds Panguitch Lake. Gregory (1949) suggested that Rock Canyon was also fed by the Ipson and Clear Creek drainages, but if so, evidence for that drainage configuration is lost to erosion and must predate formation of Panguitch Lake.

The outlet of Panguitch Lake was first dammed in 1872 and modified repeatedly until completion of the current structure in the mid 1970s. This dam is 28 feet (8.5 m) high and impounds 23,550 acre-feet (29,000,000 m^3) of water. Like many

Figure 61. *Hancock Peak, a tree-covered cinder cone near the west edge of the Markagunt Plateau. This vent area produced the middle to early Pleistocene Hancock Peak lava flows, the lower reaches of which form classic inverted valleys near the community of Mammoth. View is towards the southwest.*

Figure 62. *Mammoth Cave is one of the largest lava tubes on the Markagunt Plateau. Basaltic lava can flow long distances in tunnels such as this because the ceiling and walls help retain the lava's heat. A tunnel or tube remains when the lava drains away.*

small dams throughout Utah, it merely enlarged a pre-existing natural lake—which was about 40 feet (12 m) deep and just over one square mile (2.6 km^2) in size prior to pioneer settlement of the region. That lake was an important fishery for the Southern Paiutes; the name "Panguitch" means "big fish" in their native language. We are uncertain when the lake first formed; it appears to postdate the eruption of the 900,000-year-old Cooper Knoll lava flow, present near the southeast margin of the lake, and predate the latest Pleistocene to Holocene Black Rock Valley and Panguitch Lake lava flows. The lake lies at the southwest end of the Panguitch syncline, a structure we interpret to be related to a fault segment boundary of the Sevier fault zone.

Glacial Deposits

The Crystal and Lowder Creek basins east of Brian Head on the Markagunt

Plateau hold the southernmost glacial deposits in Utah (figure 68). Terminal, recessional, and lateral moraines, and hummocky, stagnant-ice topography, are locally well developed in these drainages, but sculpted bedrock is absent or inconspicuous, probably owing to the relatively small size and suspected short duration of the glaciers. These deposits are of the Pinedale alpine glacial advance—coincident with the global Last Glacial Maximum (LGM, 21 ± 2 ka; see, for example, Marchetti, 2007)—and an older glaciation of uncertain Quaternary age, and were first briefly described by Gregory (1950) (figure 69). We show the full extent of these glacial deposits, particularly those associated with the older advance, and note that the Lowder Creek bog has long been the site of significant alpine research into post-glacial plant recolonization of the landscape (Mulvey and others, 1984; Currey and others, 1986). The shallow lake and marsh deposits of Lowder Creek bog, which formed in a small moraine-dammed basin, span about 17,000 years. Interestingly, Lowder Creek bog contains a thin volcanic ash layer that Madsen and others (2002) identified as the 14,300 ^{14}C yr B.P. Wilson Creek #3 ash. This ash erupted from Mono Craters in California, 340 miles (550 km) due west of Brian Head, and represents the farthest east occurrence of ash from this eruption.

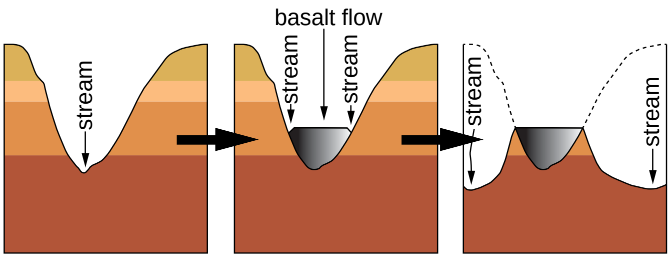


Figure 63. *Schematic diagram showing the evolution of an inverted valley. Many classic inverted valleys adorn the landscape of the Markagunt Plateau. The inverted valleys formed as lava flowed down former canyons and washes, causing the stream to shift laterally where it preferentially eroded softer sedimentary bedrock. Over time, the former valley walls are eroded away, leaving the old lava-filled valley floor, protected by the resistant lava flow, as a sinuous elevated ridge. Modified from Milligan (2002).*

Alpine glacial deposits are remarkably hard to date (Marchetti, 2007; see also Biek and others, 2010a). And because younger glacial advances typically scour away or cover deposits of older advances, the alpine glacial record is remarkably incomplete. Of the dozens of glacial advances that have occurred worldwide over the past 2 to 3 million years, only the last two are widely recognized in the Rocky Mountain region. Traditionally, researchers have used radiocarbon dating of plant debris associated with glacial deposits to determine the age of younger (less than about 60,000 years) glacial deposits. Recently, a new dating technique that uses cosmogenic exposure ages has proven useful (the technique measures the amount of time a glacier-deposited boulder has been exposed to the sun since last being moved), enabling more robust age control and correlation to other glacial deposits throughout the region. Western State College of Colorado geologist David Marchetti, who combines geologic mapping and cosmogenic dating techniques to tease apart the history and past climates recorded by these deposits, has successfully dated Pinedale glacial deposits on nearby Boulder Mountain and on the Fishlake Plateau (local last glacial maximum about 21.1 ka; Marchetti and others, 2005, 2007, 2011; Weaver and others, 2006), but the older deposits of Castle Valley lack boulders amenable to this method.

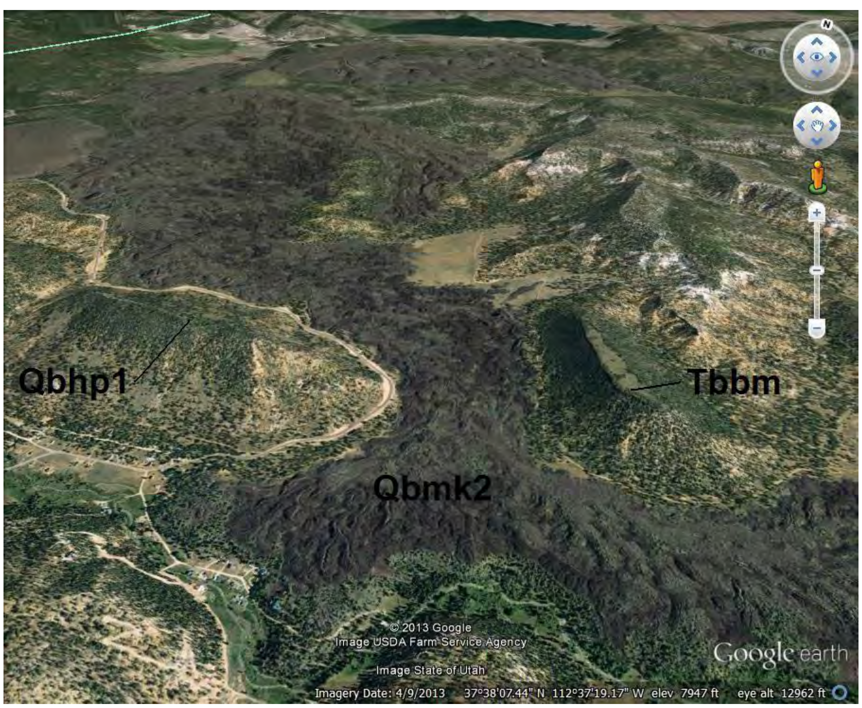


Figure 64. *Oblique Google Earth image of the young Miller Knoll lava flow (Qbmk$_2$) in Black Rock Valley. A remnant of the Hancock Peak (Qbhp$_1$) lava flow now forms a classic inverted valley above the community of Mammoth at lower left. The Blue Spring Mountain (Tbbm) lava flow remnant forms an older inverted valley to the east. Panguitch Lake is in the distance.*

We sampled pre-Pinedale sandy glacial till in an erosional scarp high above the confluence of Mammoth and Crystal Creeks for optically stimulated luminescence dating. The sample yielded a puzzling age of 48.95 ± 19.24 ka (UGS and USULL, 2013), suggesting that the deposits may correspond to the MIS 3-4 advance (generally considered a smaller advance not typically recognized in the western U.S.). Given the widespread extent of this older till, its poorly preserved or complete lack of glacial landforms, and its degree of incision along Mammoth and Crystal Creeks,

Figure 65. *Panguitch Lake lava flow immediately south of Panguitch Lake. Note the steep, blocky flow fronts that characterize this latite lava flow, unlike the more common, fluid basaltic lava flows that dot the plateau.*

Figure 66. *Panguitch Lake lava flow ($Qbpl_3$) in the distance; note steep, blocky flow front. The chemically similar lava flow that erupted from Miller Knoll ($Qbmk_1$) is in the foreground.*

we interpret the age to be unreliable and these glacial deposits to be older, more likely of Bull Lake age. This older till forms broad, open, boulder-strewn and sage-brush-covered, eastward-sloping surfaces of the Castle Creek and Mammoth Creek areas.

STRUCTURE

The Panguitch 30′ x 60′ quadrangle spans the southernmost High Plateaus in southwest Utah, covering an area of structural overlap between Late Cretaceous compressional deformation associated with the Sevier orogeny, middle Tertiary voluminous calc-alkaline volcanism, and late Tertiary to modern bimodal volcanism and basin-range extensional deformation. The bulk of the map area contains parts of three gently east-tilted plateaus bounded by high-angle normal faults: from west to east, the Parowan-Paragonah fault zone forms the west edge of the Markagunt Plateau (and thus the east margin of the Basin and Range Province at this latitude); the Sevier fault zone bisects the High Plateaus, forming the west edge of the Paunsaugunt and Sevier Plateaus; and the Paunsaugunt fault zone forms the west structural margin of the Colorado Plateau (see figures 1 and 2). The northwest part of the map area contains the Red Hills, the easternmost range in the Basin and Range at this latitude, bounded on both sides by the deep basins of Cedar and Parowan Valleys. The structural grain of the map area is thus dominated by north- to northeast-trending normal faults that bound relatively unfaulted blocks represented by the Markagunt and Paunsaugunt-Sevier Plateaus. That grain is overprinted on the leading edge of the Sevier orogenic belt, whose thrust faults and folds are dramatically displayed in the Red Hills and in Cedar Canyon, and whose subtle effects are noticed eastward to the Paunsaugunt fault zone.

Figure 67. *Blue Spring Valley and the Miller Knoll cinder cone, which is the vent area of the Miller Knoll lava flows of Black Rock Valley. Blocky, Panguitch Lake latite lava flows are to the north (left). These flows blocked the Rock Canyon drainage in latest Pleistocene to Holocene time, creating the meadow of Blue Spring Valley and ultimately diverting Bunker Creek into the Panguitch Lake drainage. The Sunset Cliffs of the Paunsaugunt Plateau are in the background.*

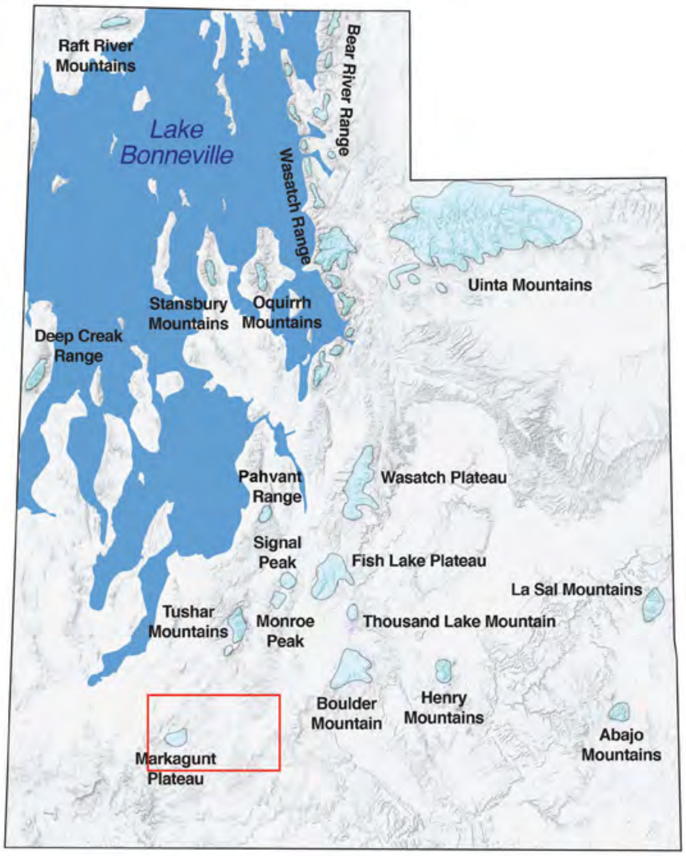


Figure 68. *Utah map showing mountain and plateau crests high enough to have been glaciated during the Last Glacial Maximum (LGM). In northern Utah, glacial ice accumulated above elevations of about 8200 feet (2500 m), whereas in southern Utah mountains needed to be above 10,000 feet (3050 m) to collect ice. Glaciologists use the term "equilibrium-line altitude" (ELA) to refer to the elevation above which snowfall will accumulate faster than it will melt (averaged over multiple years), thus thickening, compacting, and crystallizing into glacial ice. The position of an ELA is controlled by climate and thus varies over time, but when it is below mountain crests for long periods of time, glaciers develop. ELAs can also be used as a proxy to estimate temperature and precipitation during glacial advances. At the height of the LGM, central Utah was on average 7°F to 25°F (4–14°C) colder than today (Marchetti, 2007), as shown on figure 69. Box shows location of Panguitch 30′ x 60′ quadrangle.*

Late Cretaceous Compressional Structures

Iron Springs Thrust Faults

Three large-displacement thrust faults present in Parowan Gap are collectively known as the Iron Springs thrust faults. The easternmost thrust fault places overturned lower Iron Springs Formation over upright upper Iron Springs strata, the middle thrust places isoclinally folded Carmel strata over overturned lower Iron Springs Formation, and an inferred thrust at the west edge of Parowan Gap places steeply east-dipping but upright Navajo Sandstone on top of overturned Carmel Formation. The thrust sheets are erosionally beveled by the pre-Grand Castle unconformity, and the whole is marvelously exposed in the antecedent valley that is Parowan Gap.

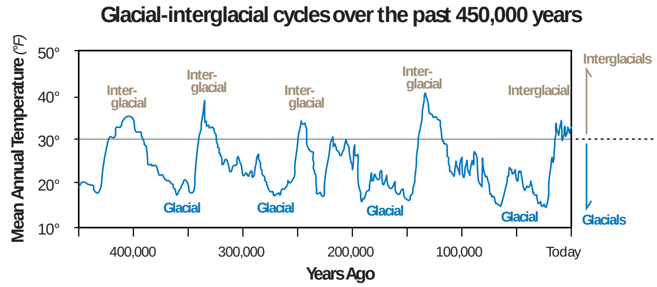


Figure 69. *Four fairly regular glacial-interglacial cycles occurred during the past 450,000 years. The shorter interglacial cycles (10,000 to 30,000 years) were about as warm as present and alternated with much longer (70,000 to 90,000 years) glacial cycles substantially colder than present. Notice the longer time span with cooling events dropping into the colder glacials followed by faster, abrupt temperature swings to the warmer interglacials. This graph, modified from Eldredge and Biek (2010), combines several ice-core records from Antarctica and is modified from several sources, including Sime and others (2009).*

Many people use the informal name "Ice Age" as a synonym for the Pleistocene Epoch (the period of time from 2.6 Ma to 11,700 years ago), and in particular for the last major ice advance about 21,000 years ago. While descriptive, such terminology is confusing in that the Pleistocene comprises multiple glacial and interglacial cycles, and, further, that the earth is still gripped within an ice age cycle (fortunately in a warm interglacial period!). Scientists thus prefer to use the more precise terms of Last Glacial Maximum (LGM) or Marine Isotope Stage 2 (MIS 2) to refer to the last glacial episode that peaked about 21,000 years ago; these terms are roughly synonymous with Late Wisconsinan, Pinedale, Smiths Fork, and a host of other local names bestowed by researchers on the youngest widespread glacial features prior to our ability to obtain numerical ages (and thus confident correlations) of these features. Similarly, the penultimate glaciation is known as pre-LGM or MIS 6, roughly coeval with the Illinoian glaciation, Bull Lake, Blacks Fork, and other local names, which exhibit glacial maxima about 140,000 to 160,000 years ago in the Rocky Mountain region

More geologists have mapped and studied the Parowan Gap area than perhaps any other place in this map area. Parowan Gap was first mapped in some detail by University of Washington student Richard Threet in the early 1950s (Threet, 1952, 1963a), and was remapped using modern techniques and terminology by co-author Florian Maldonado (Maldonado and Williams, 1993a) as part of the U.S. Geological Survey BARCO project. It was mapped yet again by University of Utah Professor David Dinter and one of his students in 2007–08 (Anderson and Dinter, 2010). While it remains a classic field camp locality for budding geologists, there are still new discoveries to be made, as our mapping attests.

The easternmost thrust fault places overturned lower Iron Springs strata (marginal marine beds that may correlate with the maximum transgression of the Western Interior Seaway of early Turonian age and assigned to the Straight Cliffs Formation by Anderson and Dinter [2010]) on top of upright, subhorizontal upper Iron Springs fluvial and floodplain strata (correlative with the Straight Cliffs or Wahweap

Formation). Anderson and Dinter (2010) first recognized that thrust faulting here continued later than previously thought because late-stage movement on the Iron Springs thrust fault cuts overlying, subhorizontal Grand Castle and Claron Formations (figure 70). Furthermore, we discovered that the thrust fault exhibits down-to-the-west extensional relaxation on the order of 200 feet (60 m). This relaxation event may be part of a larger pattern of middle Eocene to early Miocene (about 40 to 20 Ma) gravitational collapse of the Sevier orogenic belt (see, for example, Constenius, 1996; Constenius and others, 2003), or it may reflect early basin-range extension.

At Parowan Gap, the lower few feet of Grand Castle strata are locally iron stained. Anderson and Dinter (2010) called this interval their informal conglomerate of Parowan Gap, which they interpreted as distinct from and unconformable under the Grand Castle Formation; furthermore, they found this unit only west of the Iron Springs thrust. They interpreted their conglomerate at Parowan Gap as debris-flow deposits on overturned and beveled Iron Springs and Carmel strata juxtaposed by the central Iron Springs thrust, which was later transported eastward above the Iron Springs thrust. We re-examined these exposures and concluded that they are a basal clast-supported channel conglomerate of the Grand Castle Formation. We interpret local iron staining as a result of the permeability contrast between the coarse conglomerate and underlying fine-grained sandstone and mudstone of the Iron Springs Formation and micritic limestone of the Carmel Formation. Reported differences in clast composition, size, sorting, and color between the two units were also not recognized. Perhaps local, basal beds of the Grand Castle Conglomerate represent debris-flow deposits that are more poorly sorted than typical Grand Castle strata, but our re-examination failed to find widespread evidence of a separate stratigraphic unit.

The western splay of the Iron Springs thrust is inferred from exposures at the west end of Parowan Gap. There, steeply east-dipping but upright Navajo Sandstone is preserved in a horst at the west margin of the Red Hills; poorly exposed, but likely overturned Carmel and Temple Cap beds abut the Navajo Sandstone on the horst's east flank. We interpret this relationship as resulting from a down-to-the-east normal fault, part of the western bounding fault of the Enoch graben, but recognize that a west-dipping thrust fault must place Navajo over overturned and isoclinally folded Carmel strata in the shallow subsurface as shown on cross-section C–C′ (plate 3).

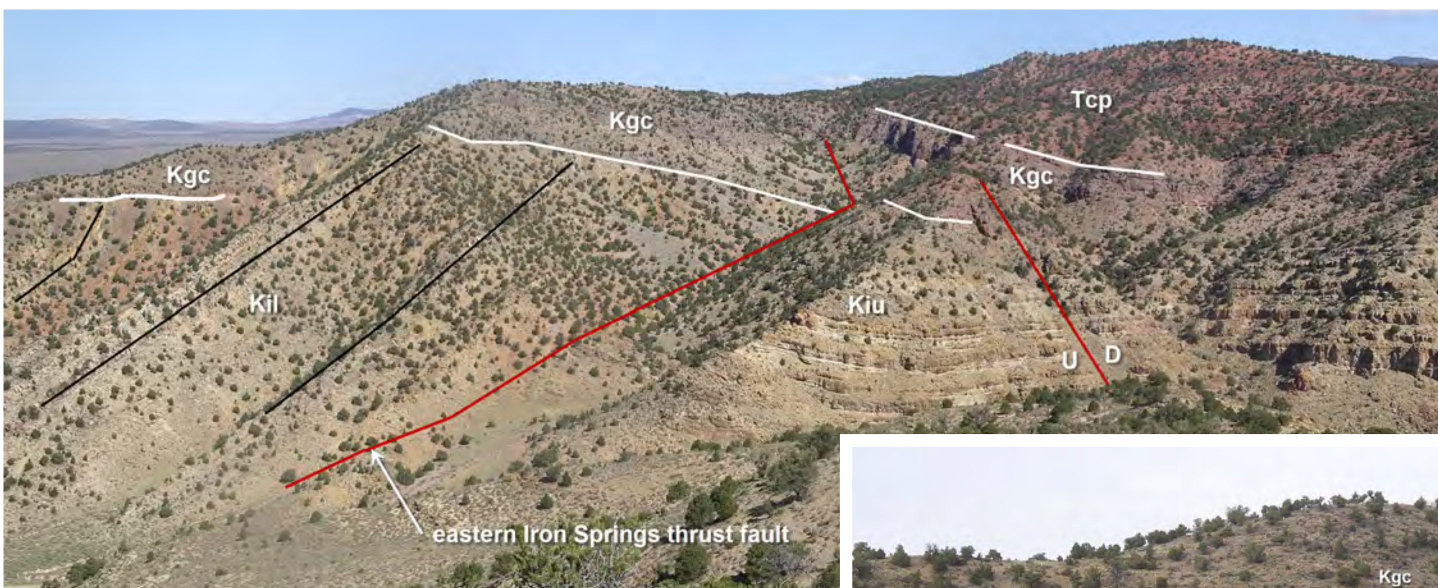


Figure 70. *The eastern Iron Springs thrust fault in Parowan Gap, which here places overturned and steeply dipping lower Iron Springs strata (Kil, black lines indicate bedding) on upright and gently east-tilted upper Iron Springs strata (Kiu). Following pronounced beveling and erosion, the Grand Castle Formation (Kgc) was deposited; it is conformably and gradationally overlain by the pink member of the Claron Formation (Tcp). Iron Springs, Grand Castle, and Claron strata in the footwall dip gently and concordantly east, showing that thrust faulting continued into early Claron time. Map relationships show about 200 feet (60 m) of extensional, down-to-the-west relaxation on the thrust fault, which likely accompanied early basin-range extension. Inset shows close-up of unconformity between overturned Iron Springs strata (Kil) and overlying, subhorizontal conglomerate of the Grand Castle Formation (redefined) (Kgc) located at the west (left) edge of the photograph.*

Interpretation of the Iron Springs thrust faults is difficult given limited exposure of upper-plate strata. Much of the thrust plate lies deeply buried under northern Cedar Valley, west of the map area, and it is locally intruded by shallow quartz monzonite bodies that are likely early Miocene laccoliths associated with the Iron Axis intrusions. We envision two plausible interpretations: (1) the overturned beds represent a large ramp anticline whose footwall cutoff must be deeply buried under northern Cedar Valley, or (2) overturned strata represent subthrust beds of the footwall. Our cross section C–C′ (plate 3) shows this latter interpretation, but we remain uncertain of the magnitude of displacement reflected in the easternmost Iron Springs thrust.

The Iron Springs thrust faults are the easternmost large thrust faults of the Sevier fold and thrust belt at this latitude. However, significant Sevier-age compressional deformation is present eastward across the map area, as evidenced by two faults at the west edge of the Markagunt Plateau that reveal an earlier history of thrust displacement (see cross section C–C′; plate 3), and by the Paunsaugunt fault, which exhibits an earlier history of pre-Claron, east-directed movement (see Paunsaugunt fault zone discussion).

Cedar City-Parowan Monocline

The northwest margin of the Markagunt Plateau is cut by paired high-angle normal faults that create a series of horsts and grabens that step down from the plateau to the adjacent Basin and Range (figure 71) (Thomas and Taylor, 1946; Threet, 1963b; Maldonado and others, 1994, 1997). Horst blocks expose increasingly older Upper Cretaceous strata westward across the margin of the plateau, reflecting the east flank of the Kanarra anticline. Grabens preserve younger Claron, Brian Head, and Oligocene and Miocene ash-flow tuffs as well as the early Miocene Markagunt gravity slide.

Close to the margin of the plateau at the southern end of Parowan Valley, the dip of Upper Cretaceous and Claron strata within horst blocks is warped down moderately to steeply northwest, parallel to the Parowan fault, defining a monocline. Immediately west of the map area, these strata are warped steeply west parallel to the Hurricane fault zone (Averitt and Threet, 1973; Knudsen, 2014a). In this map area, the monocline is especially well expressed south of Summit and midway between Summit and Parowan (see figure 29). This folding creates what Threet (1963b) called the Parowan monocline, and what Anderson and Christenson (1989) later called the Cedar City–Parowan monocline (Maldonado and Sable also produced a prelimnary, unpublished map of this area, parts of which are incorporated into our compilation of the Panguitch 30′ x 60′ quadrangle). This fold has long been considered a result of deformation associated with basin-range extension (see, for example, Anderson and Mehnert, 1979; Maldonado and others, 1997). Threet (1952, 1963b) suggested that the monocline formed as a structural link between the Hurricane and Paragonah fault zones and was thus of Neogene age. In particular, Threet (1963b) reasoned that the monocline represented oblique unfolding of the east flank and, immediately west of the map area, accentuation of the west flank of the Kanarra anticline. Maldonado and others (1994, p. 111; see also Maldonado and others, 1997) also interpreted the monocline as Miocene or younger, and described it as "… a flexure along the east edge of Parowan Valley," implying that it may be related to their Parowan thrust fault zone (described later) and thus a result of southeast-directed compression possibly due to intrusion of a hypabyssal igneous body.

Importantly, however, apart from possible tilting of blocks within the Parowan fault zone at the mountain front, regional ash-flow tuffs preserved in the intervening grabens are not affected by monoclinal folding. Isom, Leach, Bauers, and Harmony Hills ash-flow tuffs are subhorizontal at the north end of the Summit Mountain graben immediately southeast of Summit, and Leach and Bauers are subhorizontal at the north end of the Fiddlers Canyon graben near the west edge of the map area. Monoclinal folding must thus predate Oligocene volcanism—and thus basin-range extension—in the region. We interpret the monocline as a thrust propagation fold, a fold formed above a blind thrust fault now reactivated as a basin-range normal fault (see cross section C–C′; plate 3). Alternately, it is possible that the monocline is instead related to a northwest-vergent backthrust. Compressional, Sevier-age deformation extended east of the Iron Springs thrust, as well documented by stratigraphic and structural relations on the Paunsaugunt Plateau (see discussion of Paunsaugunt fault), and it extended into Claron time as evidenced by late-stage movement of the Iron Springs thrust fault (discussed above).

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, 1963b; Kurie, 1966; Averitt, 1967; Averitt and Threet, 1973; Hurlow and Biek, 2003; Biek and others, 2009; Biek and Hayden, 2013; Knudsen, 2014a); part of its east limb is preserved at the west edge of the map area in Cedar Canyon. 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 downdropped to the west where they are concealed by younger deposits. It appears that the blind thrust over which the Kanarra anticline formed was reactivated as part of the Hurricane fault zone. The east limb of the fold is overturned near Toquerville (Hurlow and Biek, 2003) and near Kanarraville (Averitt, 1967; Biek and Hayden, 2013), but is elsewhere upright, with beds on the east limb generally dipping between 20° and 35° east, shallowing abruptly as they pass under the great cliffs of Navajo Sandstone in the vicinity of Kolob Canyons.

Southwest of the map area, 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,

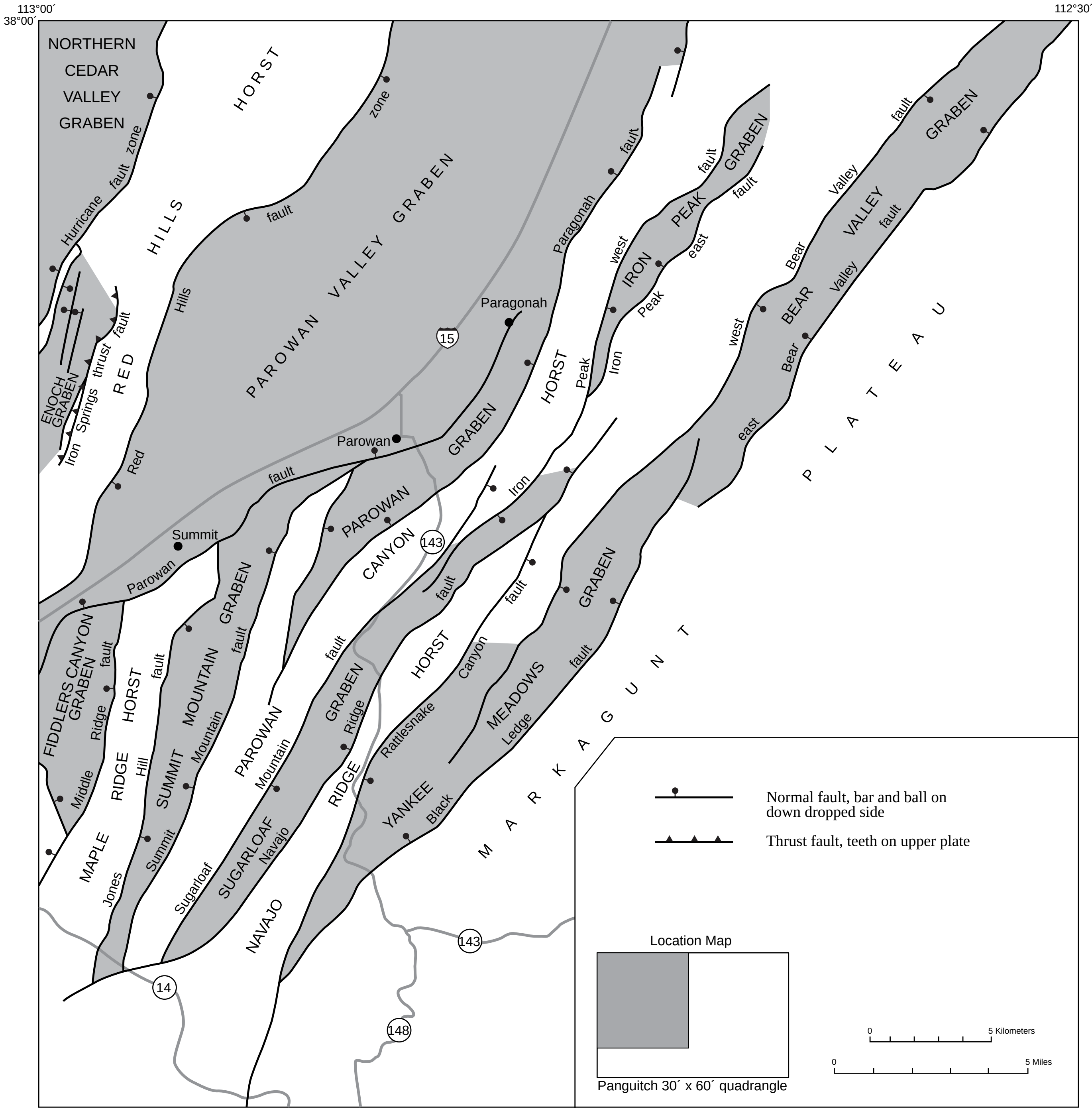


Figure 71. *Major faults of the western Markagunt Plateau and Red Hills, and named grabens (shaded) and horsts. Modified after Maldonado and others (1997).*

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 (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). However, depending on the inferred geometry of the thrust faults, horizontal shortening along these faults may be two to three times what Kurie (1966) depicted. The Taylor Creek thrust faults extend from Toquerville on the south to Kanarraville on the north, but similar west- and east-directed thrust faults are present along the fold's eastern limb northward to Cedar City (Knudsen, 2014a). In Cedar Canyon at the west edge of this map area, the Carmel Formation is disharmonically folded on the east limb of the anticline.

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, which also marks the eastern limit of significant Sevier-age compressional deformation in southwest Utah.

The relationship of the Kanarra anticline and Iron Springs thrust faults is unknown because the structures are dismembered by basin-range faulting and their area of possible overlap is covered by basin-fill deposits. We agree with the reasoning of Threet (1963b) who argued that the two structures are related, but we are uncertain if the Iron Springs thrust is the same blind thrust that formed the Kanarra anticline—it is possible that they are two closely spaced, en echelon anticlines separated by a zone of complex folding, similar to that present between the Kanarra and Virgin anticlines to the south (Biek and others, 2009).

"Parowan Thrust Fault Zone" of Maldonado and others

On the basis of unpublished mapping by Maldonado and Sable in the Summit quadrangle between Summit and Parowan, Maldonado and others (1992a, 1992b, 1994, 1997) named the Parowan thrust fault zone for exposures in three separate structural blocks. The senior author re-examined these areas and reinterprets their northern thrust fault segment (southwest of Parowan) as two high-angle normal faults whose dip locally appears to be gently northwest due to the presence of modern landslide deposits. Biek reinterpreted their middle segment as a high-angle normal fault that drops Claron down against the John Henry Member of the Straight Cliffs Formation, again locally concealed by modern landslide deposits, and their southern segment as a high-angle normal fault that drops Isom Formation down against John Henry strata. The senior author thus suggests that structural relations in these areas are best explained by a network of Neogene high-angle normal faults—locally concealed by previously unmapped modern landslide deposits—that cut the relay ramp between the Hurricane and Paragonah fault zones. Co-author Maldonado disagrees and prefers his original conclusion that the faults are low-angle, south- and east-vergent thrust faults that place Cretaceous strata over Claron and Isom Formations, further suggesting that the faults resulted from intrusion of a hypabyssal body under what is now Parowan Valley.

Middle Tertiary

Paunsaugunt Thrust Fault System

On the Paunsaugunt Plateau, south-vergent thrust faults place Upper Cretaceous strata on the pink member of the Claron Formation. The faults were rediscovered in the mid-1980s, when University of Arizona geologists George Davis and Robert Krantz noticed small-displacement thrust faults at Bryce Canyon National Park; their "discovery" fault is none other than the small south-dipping thrust at Bryce Point (figure 72) (Davis and Krantz, 1986). Though small, this fault is so prominent that it is hard to believe it could go undetected for so long, especially considering the scores of geologists who have cast their eyes over that landscape (Davis, 1990, 1991, 1999). But Davis was primed for the discovery, as just the day before he examined a low-angle fault zone where State Highway 12 cuts through the Claron Formation at the north end of the park, brought to his attention by U.S. Geological Survey geologist Richard Hereford, which he reasoned may exhibit thrust movement. Davis (1997a, 1997b, 1999) provided an engaging account of the rediscovery of what is now known as the Paunsaugunt thrust fault system, the main parts of which were originally interpreted by Gregory (1951) as a west-trending horst bounded by high-angle normal faults. Davis's student Eric Lundin first mapped the fault zone in detail, outlining many of its salient features (Lundin, 1987, 1989; Lundin and Davis, 1987). Davis (1997b, p. 72) related that "During the course of Lundin's mapping, I took him to the site where Bob Krantz and I had reinterpreted the Rubys Inn fault as a thrust. There, in an out-of-the-way place on the poor outcrop I was amazed to find a Chevron business card neatly tucked underneath a fist-sized rock placed on the trace of the Rubys Inn thrust, at a location approximately midway between the Paunsaugunt and Sevier faults. It was Frank Royse's business card. He had left it there (on a later trip with his student Jim Coogan), and had written on the back side of the card, "Hi George! How 'bout this crazy fault?" Though their work was unpublished, Chevron geologists discovered this thrust fault in the mid-1950s!

The Paunsaugunt thrust fault system consists of several thrust faults of which the frontal Rubys Inn thrust fault and the Pine Hills back thrust fault are the best known. Because the thrust faults are at right angles to the expected orientation of Sevier and Laramide structures in the region, and because they involve strata that were thought to post-date those orogenies, the cause of the thrust faulting remained poorly understood until 1993. Lundin (1987, 1989) showed that the thrust faults form an arcuate belt around the southeast margin of the Marysvale volcanic field and that they sole into Middle Jurassic evaporites (figures 73 and 74), but he was uncertain of their tectonic significance. Nickelsen and Merle (1991) and Nickelsen and others (1992) showed that southward-directed compressional deformation extended as much as 12 miles (20 km) south of the thrust fault zone on both the Paunsaugunt and Markagunt Plateaus. Maldonado and others (1997) also showed evidence of conjugate shears in Claron strata near the west edge of the Markagunt Plateau, and May and others (2011) and Leavitt and others (2011) documented compressional deformation as much as 18 miles (29 km) south of the thrust on the Paunsaugunt Plateau and noted similar structures in Cedar Canyon. Furthermore, as described below, our new mapping shows that a related blind thrust fault must be present south of Panguitch Lake on the Markagunt Plateau, possibly the westward extension of the Rubys Inn thrust fault.

The pervasive small-scale deformation noted above includes small-displacement (millimeters to centimeters) conjugate thrust and strike-slip faults in Upper Cretaceous and Claron strata, which Davis (1999) called deformation band shear zones, as well as stylolites and spaced cleavage in impure Claron carbonate beds. Classic deformation bands are well developed in the capping sandstone of the Wahweap Formation in Hillsdale Canyon (figure 75) and are discussed in detail by Davis (1999, p. 73). Such shears are also common in hoodoos of Bryce Canyon, which led Davis (1999) to show that these features are as important, and locally more important, than vertical joints in controlling the formation of hoodoos.

Nickelsen and others (1992) and Davis and Rowley (1993) first suggested that this unusual thrust deformation may be due to laccolith emplacement in the southern part of the Marysvale volcanic field, an idea expanded upon by Merle and others (1993), who suggested that the thrust fault zone formed about 20 to 30 million years ago in response to gravitational collapse of, and/or emplacement of batholithic intrusions into, the Marysvale volcanic field. Interestingly, Merle and others (1993) found no evidence of compressional deformation in the conglomerate at Boat Mesa, which they thus thought postdated thrust deformation but which we now know to be about 38 Ma. On the basis of exposures northwest of Johns Valley, in the northwest part of the Flake Mountain East quadrangle, the conglomerate at Boat Mesa is clearly folded concordantly with underlying Claron strata in the upper plate of the Johns Valley thrust fault, as is overlying Brian Head Formation, and, we suspect, Mount Dutton Formation. Recently, Anderson and others (2013) suggested that the Paunsaugunt thrust fault system is related not to gravitational spreading of the Marysvale volcanic field, but rather to a left-lateral component of strike-slip movement on the Paunsaugunt fault. This interpretation, however, does not take into account the inferred continuation of the Rubys Inn fault westward into the Markagunt Plateau as shown by our new mapping, nor the fact that we find no kinematic indicators of anything other than normal dip-slip

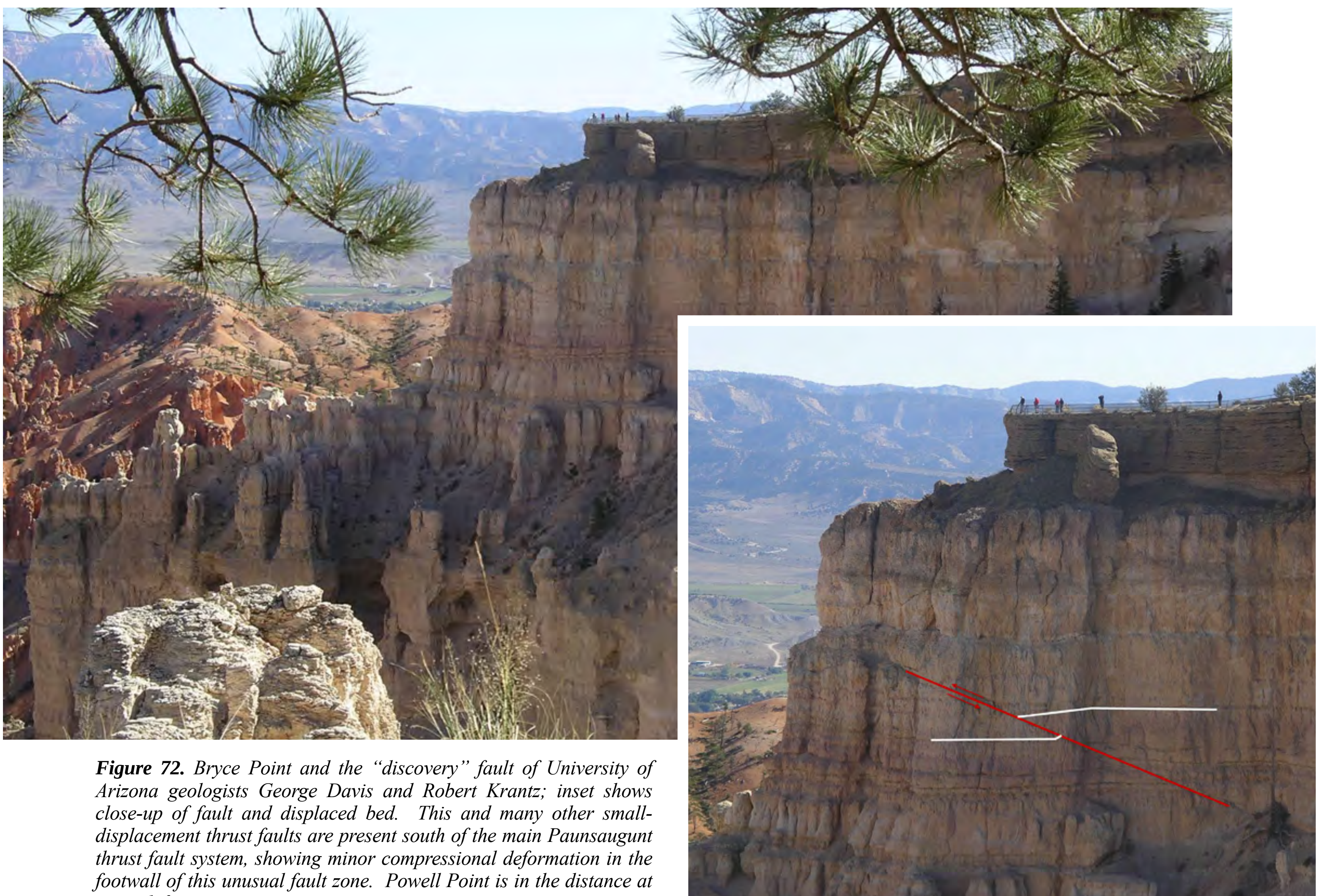


Figure 72. *Bryce Point and the "discovery" fault of University of Arizona geologists George Davis and Robert Krantz; inset shows close-up of fault and displaced bed. This and many other small-displacement thrust faults are present south of the main Paunsaugunt thrust fault system, showing minor compressional deformation in the footwall of this unusual fault zone. Powell Point is in the distance at upper left.*

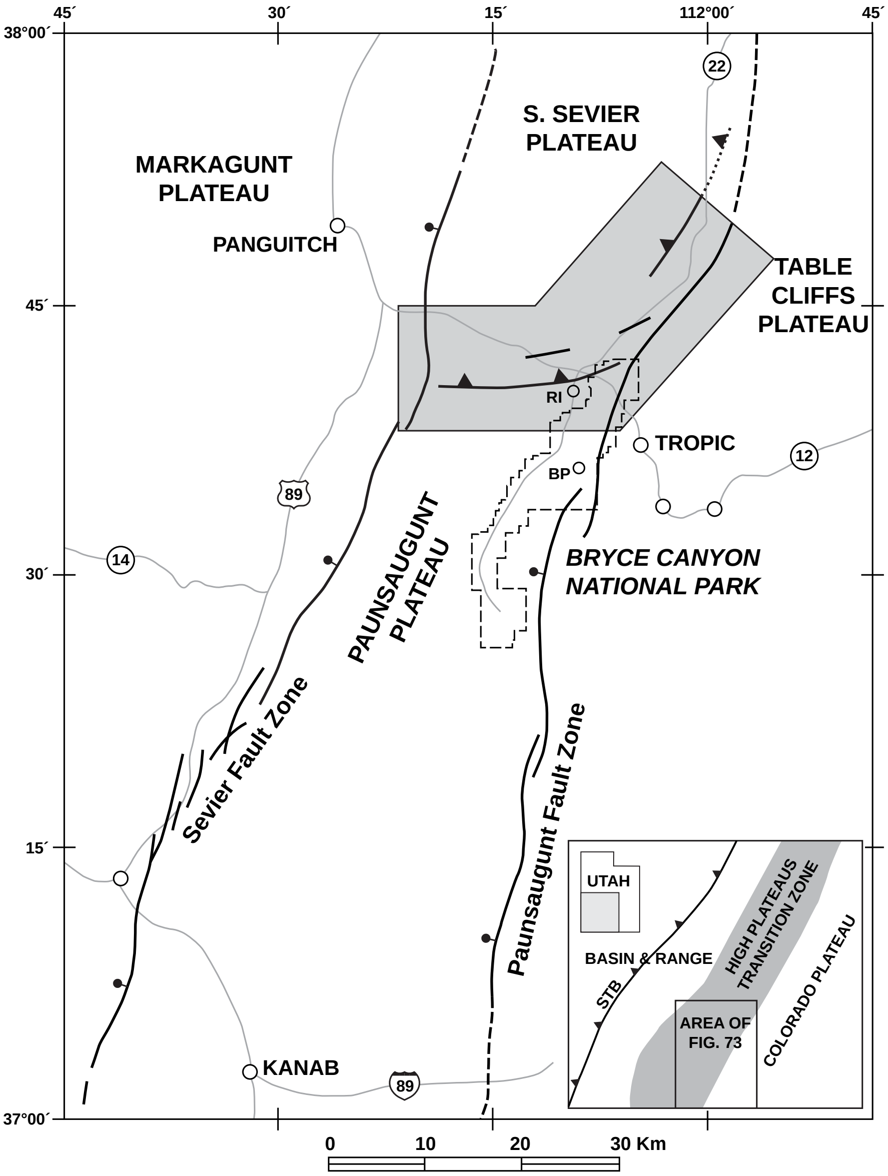


Figure 73. *Map of the southern High Plateaus showing location of the Paunsaugunt thrust fault system. BP, Bryce Point; RI, Ruby's Inn; STB (in inset box), leading edge of Sevier thrust belt. Modified from Lundin (1989).*

movement on the Paunsaugunt and Sevier faults. Like Merle and others (1993), we agree that gravitational spreading of the Marysvale volcanic field produced the unusual Paunsaugunt thrust fault system.

On the basis of earlier work and our new mapping, the age of thrusting on the Paunsaugunt Plateau can only be definitively constrained as postdating the 37 to 33 Ma Brian Head Formation and predating basin-fill deposits (Taf, Tvg) of poorly constrained Miocene age. If gravitational loading by the Marysvale volcanic pile caused the thrusting as we suspect, then it is reasonable to conclude the faults were coincident with late-phase development of the approximately 20 to 30 Ma volcanic field. Why the conglomerate at Boat Mesa fails to show significant evidence of widespread, small-scale compressional deformation, as is so well developed in the underlying Claron Formation, we do not know. Still, following the "two-tiered" model of Davis (1997a, 1997b, 1999), we envision the thrusts as directed outward from the Marysvale volcanic field, which spread and collapsed under its own weight, resulting in southward-directed thrust faults rooted in evaporite strata of the Middle Jurassic Carmel Formation. The magnitude of offset on the major faults is probably about 1500 to 3000 feet (Lundin, 1989; Davis, 1999).

The Rubys Inn thrust fault is spectacularly exposed in Hillsdale Canyon and at the east edge of the Paunsaugunt Plateau where Utah Highway 12 drops down into the Tropic amphitheater. At Hillsdale Canyon, the fault places Wahweap strata over the pink member of the Claron Formation (figure 76). Two splays of the thrust fault are present at the east edge of the plateau, where they dip about 30° north and place pink Claron on top of itself (figure 77). Spectacular exposures of the grooved and striated fault plane are present above Highway 12 (figure 78). Part of the thrust is also preserved in a hoodoo not far north of the highway (see figure 21 of Davis and Pollock [2010]).

On the Markagunt Plateau, we find no evidence of large-scale older-on-younger relationships indicative of thrust faulting. However, along trend with the Rubys Inn thrust fault, tilted Claron and Brian Head strata near Panguitch Lake and tilted Isom Formation at Haycock Mountain suggest that these rocks were folded above a blind thrust that, at Haycock Mountain, ramped up and soled into the Brian Head Formation; a deeper thrust would be required to fold Claron strata near Panguitch Lake. On the south side of Haycock Mountain, the upper few tens of feet of Brian Head strata are typically deformed by shears and small recumbent folds, but it is unclear if this deformation is due to a Rubys Inn-related thrust fault or emplacement of the Markagunt gravity slide. The 20 to 21 Ma Limerock Canyon Formation, present only at Hatch Mountain and nearby areas to the south, may be preserved in a subtle basin south of this inferred blind thrust, implying that thrust faulting took place at the transition between major calc-alkaline to bimodal volcanic activity in the southern part of the Marysvale volcanic field. Alternatively, tilting of the Claron, Brian Head, and Isom may be much younger, reflecting the location of these units on the southeast limb of a poorly expressed syncline that clearly folds the 5 Ma

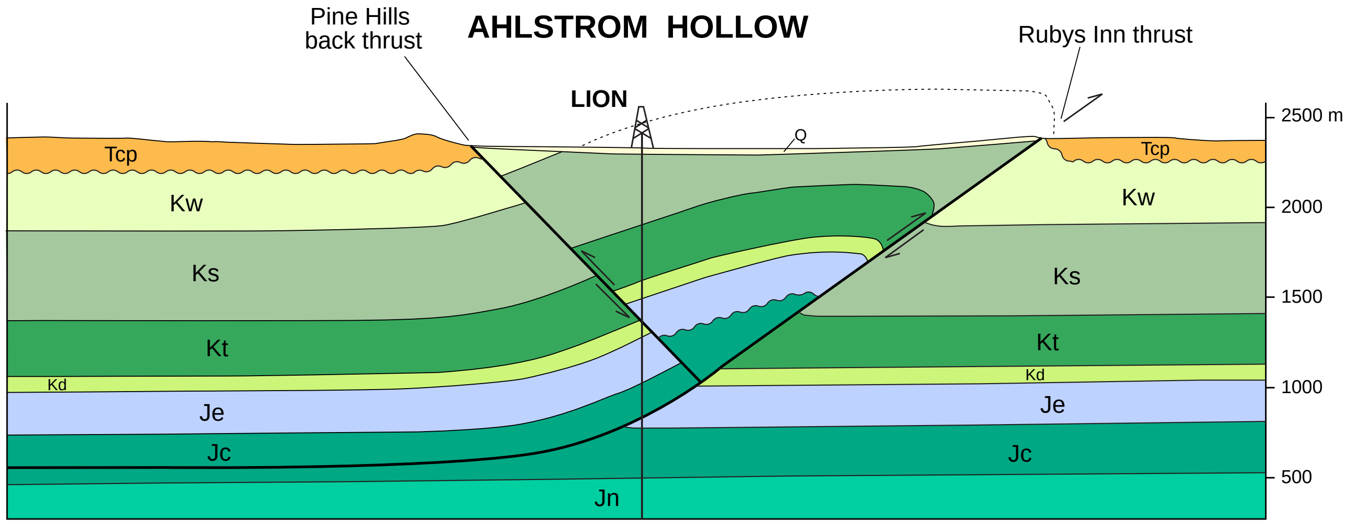


***Figure 74.** North-south cross section through the central portion of the Paunsaugunt thrust fault system near Ahlstrom Hollow, showing the Rubys Inn thrust fault and Pine Hills back thrust. Note that Rubys Inn thrust fault soles into Middle Jurassic evaporite-bearing strata (Jc, Carmel and Arapien Formations). Modified from Lundin (1989).*

***Figure 75.** Small-displacement conjugate shears (deformation bands) in the capping sandstone member of the Wahweap Formation at Hillsdale Canyon.*

Fivemile Ridge lava flow northeast of Panguitch Lake, a fold likely related to an inferred segment boundary on the Sevier fault zone.

Middle Tertiary Markagunt Gravity Slide

Introduction

With an apparent areal extent of at least 1600 square miles (4160 km^2), larger than the state of Rhode Island, the Markagunt gravity slide is among the largest subaerial gravity slides on Earth. It is larger than the famous 1300-square-mile (3400 km^2) Heart Mountain detachment in northwestern Wyoming, long considered to be the largest landslide on land (see, for example, Malone and Craddock, 2008; Beutner and Hauge, 2009; Craddock and others, 2009). Grooves, striations, Riedel shears, pseudotachylyte, crushed and rehealed clasts, basal cataclastic breccia, and clastic dikes provide strong evidence of catastrophic emplacement from the north by gravity sliding. The uniformity of directional indicators, the stratigraphic sequence of volcanic rocks that make up upper-plate strata, and the overall geometry of the gravity slide show that it represents a single emplacement event and is not composed of multiple, smaller gravity slides derived from various sources (Biek and others, 2014; Hacker and others, 2014). The gravity slide was emplaced on rocks as young as the 22 Ma Harmony Hills Tuff and is overlain by and thereby apparently predates undeformed Haycock Mountain Tuff, which yielded an $^{40}Ar/^{39}Ar$ age of 22.75 ± 0.12 Ma (Sable, unpublished data, 1996), but for which we report a new U-Pb zircon age of 21.63 ± 0.73 Ma (UGS and AtoZ, 2013a).

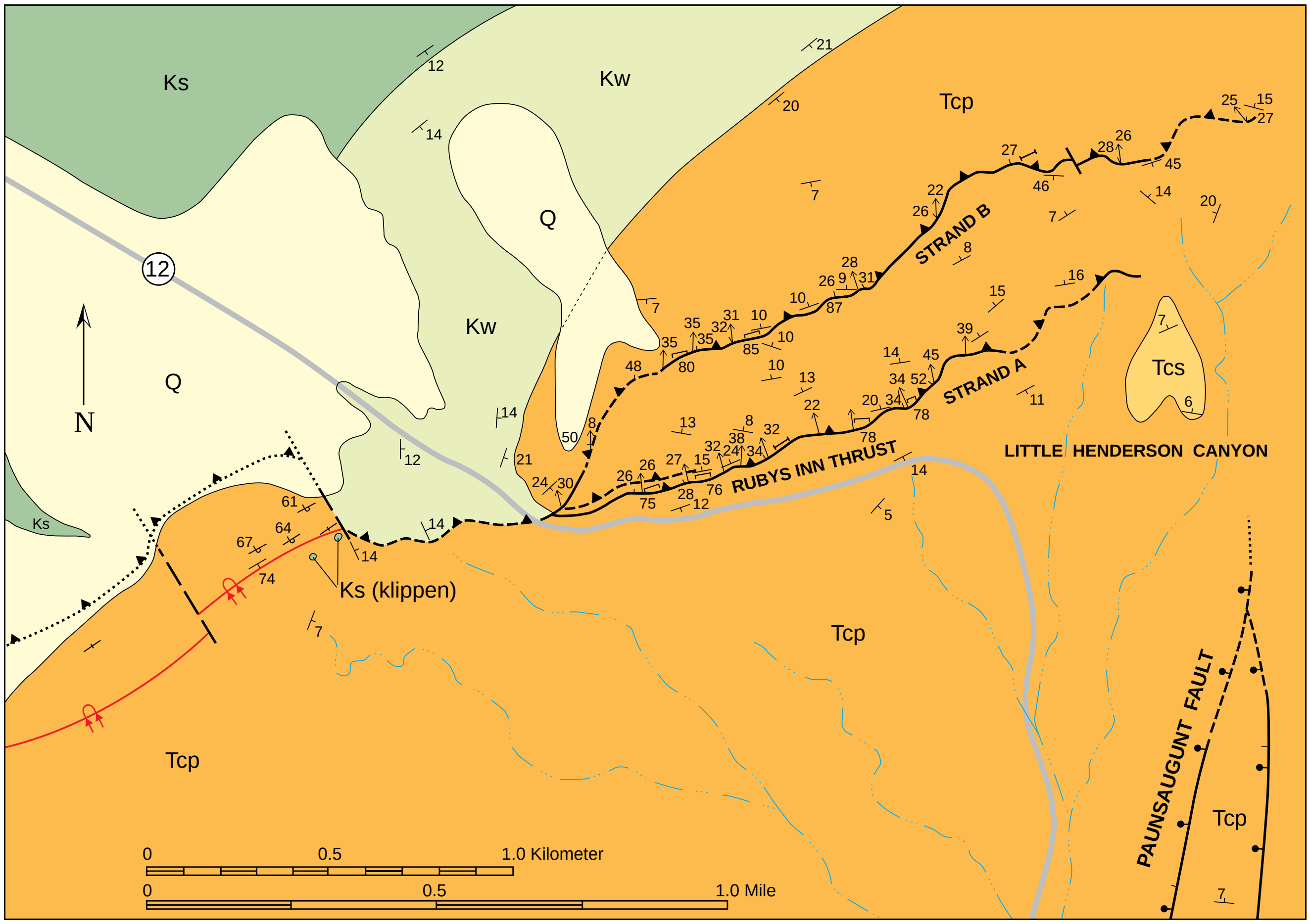
The Markagunt gravity slide represents catastrophic gravitationally induced collapse, about 22 to 21 million years ago, of the southwestern sector of the Oligocene to early Miocene Marysvale volcanic field, one of the largest fields in the United States. From its inferred breakaway zone in the Tushar Mountains to the southern limit of its debris avalanche deposits, the gravity slide is about 60 miles (95 km) long; it is at least 40 miles (65 km) wide at the latitude of the ramp. We can now document catastrophic southward transport of at least 20 miles (32 km) over the former early Miocene land surface. The southwestern part of the Marysvale volcanic field, which consists of clustered stratovolcanoes and subordinate yet important calderas, is built on a weak substrate of mostly fine-grained volcaniclastic strata of the Brian Head Formation, which to this day are famously prone to modern landsliding. We remain uncertain what triggered the gravity slide, but sug-

Figure 76. *Steeply south-dipping Wahweap Formation (Kw) thrust over subhorizontal pink member of the Claron Formation (Tcp) at hill 8170 in Hillsdale Canyon; view is towards the east.*

gest that it was triggered by pre-caldera inflation of the 20–18 Ma Mount Belknap caldera.

The Markagunt gravity slide was discovered in the mid-1980s and its eroded remains were formally named the Markagunt Megabreccia in 1993, nearly two decades after geologists first began to realize that the gently east-tilted, high-elevation Markagunt Plateau was capped by something other than the normal sequence of volcanic rock commonly found in southwest Utah (Anderson, 1993). We use the term Markagunt Megabreccia to refer to the eroded remains of the Markagunt gravity slide as to better tie its deposits to its structural origin. Also, the deposits were so named early after their discovery, when it was realized that they represented the product of a gravity slide but when knowledge of the extent and source of the slide was based on incomplete and preliminary mapping of areas for which we now have a much fuller understanding. The Markagunt Megabreccia (map unit Tm) is used for those parts of the gravity slide where individual rock units are not mapped separately; map symbols with a unit name, for example Tm(Tnw) for Wah Wah Springs Formation, are used where the map unit is mapped separately as part of the upper plate of the gravity slide. Typically, we map only the basal shear as a gravity slide fault; boundaries between the many units that compose its upper plate are mapped as contacts for simplicity and to avoid map clutter, though we recognize that many such contacts are indeed slip surfaces that accommodated variable amounts of deformation including extension and thinning of upper-plate strata and, especially near the ramp area, compression and juxtaposition of units by thrusting. One exception is that we retain gravity slide faults that bound relatively small allochthonous blocks mapped by previous workers in the northern Markagunt Plateau; some of these blocks apparently represent post-Markagunt Megabreccia gravity slides shed off known and inferred laccoliths, whereas others appear to represent blocks shed off the pre-Markagunt Megabreccia Spry intrusion.

At its simplest, the gravity slide is a great sheet of volcanic rock that slid many miles southward and at its distal southern end placed older rock on younger rock above a subhorizontal surface. It blankets the entire central and northern Markagunt Plateau and adjacent areas (figure 79) and consists of large blocks many square miles in size of Miocene and Oligocene regional ash-flow tuffs (most originally erupted from calderas along the Utah-Nevada border) and local volcanic and volcaniclastic rocks (derived from the Marysvale volcanic field). One way to think of the Markagunt gravity slide is like a stack of thick cards or blocks that are sheared between one's hands—strata are intensely deformed along the shears themselves but remain relatively undisturbed in the interior of the blocks. The fact that the gravity slide consists mostly of undeformed large blocks, bounded below by an inconspicuous shear plane, is one of the reasons it remained undiscovered for so long. Locally however, the gravity slide is a structurally chaotic assemblage or consists of large tilted blocks of these volcanic rocks. Nearly everywhere, rocks immediately below the gravity slide are undisturbed. Mapping and stratigraphic studies during the 1970s to 1990s show how our understanding of this complex unit has evolved, as summarized by Maldonado and others



SYMBOLS

Contact, dotted where concealed

Normal fault showing dip: ball on the downthrown side

Thrust fault showing dip and trend and plunge of striae

Tear fault

Overturned nonplunging syncline

Strike and dip of bedding: overturned: vertical

Strike and dip of cleavage: vertical

Figure 77. *Geologic map of the Rubys Inn thrust fault where it crosses Utah Highway 12 in Bryce Canyon National Park. Ks, Straight Cliffs Formation; Kw, Wahweap Formation; Q, surficial deposits; Tcp, pink member of the Claron Formation; Tcs, unusual facies of both the white member of the Claron Formation and conglomerate at Boat Mesa. Modified from Lundin (1989).*

(1992a, 1997), Anderson (1993), Moore and Nealey (1993), Sable and Maldonado (1997a), Hatfield and others (2004, 2010), Moore and others (2004), Biek (2013b), Rowley and others (2013), and Biek and others (2014).

There is no single place where we can go to see the entire story encapsulated by the Markagunt gravity slide. Different geologists have seen different parts of the beast, and thus understandably came to different conclusions about this complex unit. It reminds us of the allegory of the “blind men and the elephant.” Part of it feels like one thing, other parts like something else; in the beginning it was simply too big and too strange for any one person to understand. This is especially true given that until recently the entire Markagunt Plateau had not yet been mapped in sufficient detail. Early reconnaissance-scale geologic maps of the plateau did not even recognize the gravity slide. Those pioneering geologists mapped just the back of the elephant and understandably did not recognize the gravity slide for what it was at the time. Most later detailed mapping was north of the ramp fault—an area of translation but not significant deformation of the gravity slide—and understandably these maps, apart from relatively small gravity slides shed off small laccoliths that apparently both pre- and post-date emplacement of the Markagunt Megabrecia, also did not recognize the allochthonous nature of this much larger feature. Mapping of the entire Panguitch 30′ x 60′ quadrangle allowed us to see most of the “elephant.” Especially important was the recognition of

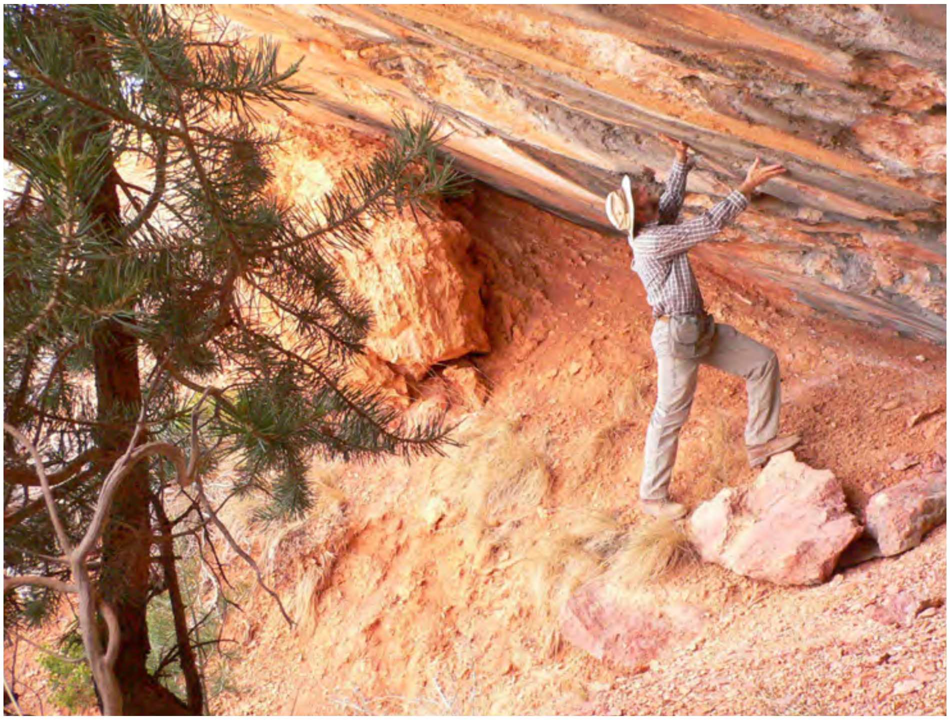


Figure 78. *Lower splay of Rubys Inn thrust fault just north of the "Dark Skies" turnout on Utah Highway 12 in Bryce Canyon National Park.*

a ramp, the location where the gravity-slide plane comes to the surface and allows the toe of the slide to move across the former land surface. Importantly, however, early mapping by Maldonado in the northern Red Hills in the westernmost Markagunt Plateau led to the discovery of what he called the Red Hills shear zone, part of which we now reinterpret as the basal slip surface of the Markagunt gravity slide, as discussed below.

The discovery and our still-unfolding understanding of the Markagunt gravity slide began in the early 1970s when James Judy, during thesis mapping supervised by John Anderson of Kent State University (Judy, 1974), identified older-on-younger relationships (Wah Wah Springs Formation resting on Isom Formation) near Sidney Peaks northeast of Brian Head as well as farther north along Black Ledge, a result, he speculated, of gravity sliding off a horst block west of Black Ledge and possibly from intrusion of the Iron Peak laccolith. But it was not until the late 1980s to early 1990s, when Anderson and half a dozen U.S. Geological Survey (USGS)-supported Master's students continued mapping in this area that they finally had enough information to grasp what they were dealing with. Unfortunately, their work was cut short following reorganization of the USGS in the mid 1990s, when they had only just begun to appreciate the scale of this beast. Nevertheless, their combined work laid a solid foundation for future mapping in the region. In our new mapping, we have stumbled across the beast's trunk—its basal shear planes complete with clastic dikes, pseudotachylyte-filled fractures, and pervasive cataclasis of the lower few feet of the slide mass—which let us revise much of what we now know about the gravity slide.

Sable and Maldonado (1997a) noted that four separate rock units, listed below, were termed megabreccia, the first two of which we now ascribe to other origins. The units are (1) mudflow and lava-flow breccias and fluvial volcaniclastic sedimentary rocks containing scattered clasts of volcanic rocks (Anderson, 1985), which are long since known to be primary volcano-sedimentary breccia, the alluvial facies of the 23–32 Ma Mount Dutton Formation derived from the southern margin of the Marysvale volcanic field, (2) megabreccia interpreted to have resulted from collapse of high-angle fault scarps (Anderson, 1993, 2001), now known to be modern landslide deposits (Maldonado and others, 1997; Hatfield and others, 2010; Rowley and others, 2013; see also this map), (3) strata associated with the Red Hills shear zone of Maldonado and others (1989, 1992a), Maldonado (1995), and Sable and Maldonado (1997a), parts of which we now assign to the Markagunt gravity slide, and (4) the principal mass of the Markagunt gravity slide that covers much of the central and northern Markagunt Plateau, on which all workers agree. Sable and Maldonado (1997a) restricted the term to unit 4, but were uncertain of the relationship of megabreccia associated with their Red Hills shear zone to that of the main mass of the Markagunt Megabreccia on the Markagunt Plateau.

One additional unit is referred to by some workers as Markagunt Megabreccia, which senior-author Biek maps as Markagunt gravity slide residuum (QTbx); it is well developed at the west rim of the Markagunt Plateau in and near Cedar Breaks National Monument. Hatfield and others (2004, 2010), Moore and others (2004), and Rowley and others (2013) interpreted this rubble as Markagunt Megabreccia let down by dissolution of underlying Claron limestone, although they noted its unconsolidated nature. Moore (1992) interpreted this unit as landslide debris derived from the Markagunt Megabreccia. Senior author Biek interprets this unit to be the weathering product of what may be a distal, debris avalanche phase of the gravity slide that is locally present along its distal, eroded southern margin.

Red Hills Shear Zone

Maldonado (1995; see also Maldonado and others, 1989, 1990, 1992a, 1994, 1997, 2011; Maldonado and Sable, 1993) described an Oligocene and Miocene volcanic section about

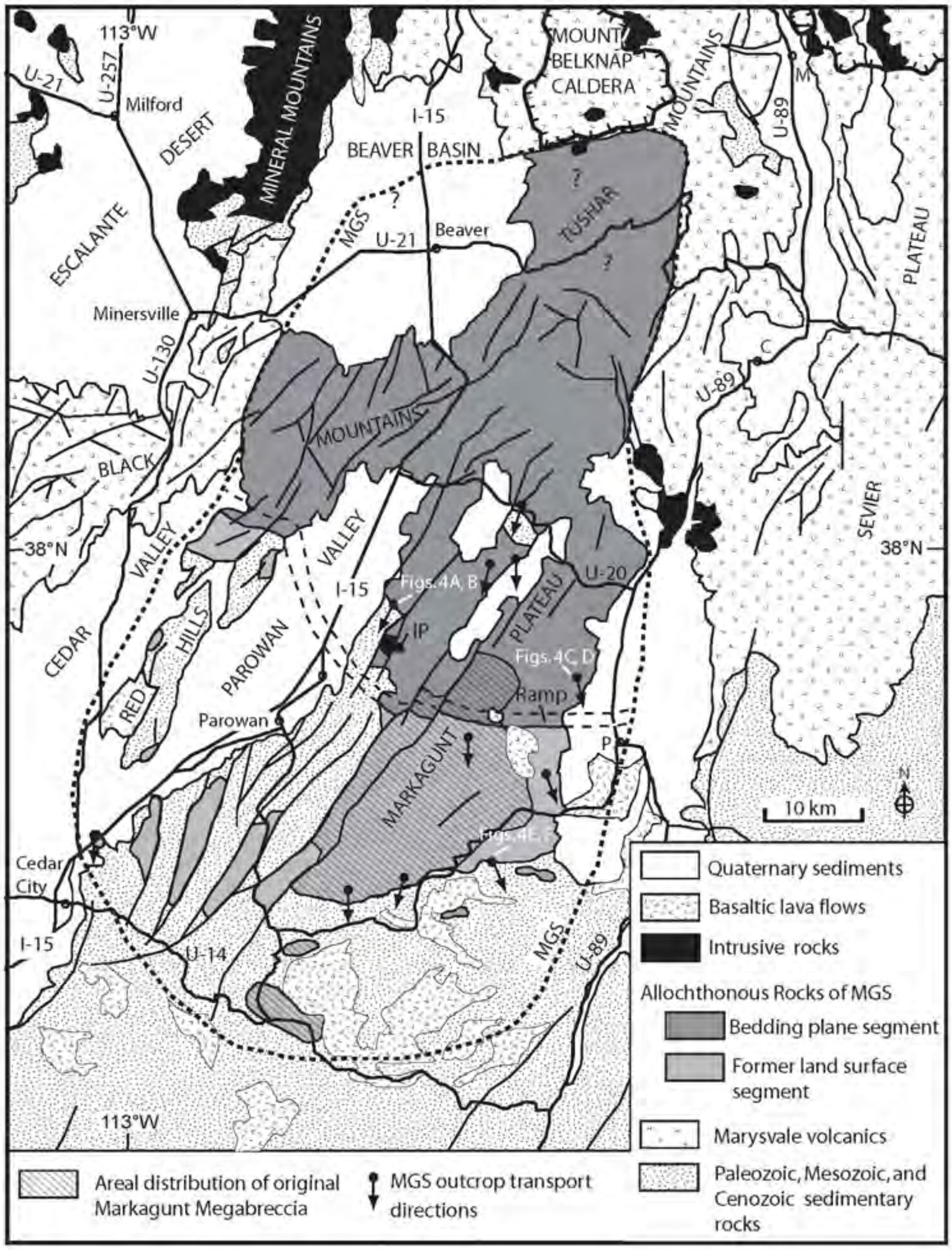


Figure 79. *Simplified geologic map showing extent (heavy dashed line) and features of the Markagunt gravity slide (MGS). C, Circleville; IP, Iron Peak intrusion; M, Marysvale; P, Panguitch. Note that the eastern flanking fault is now known to be near the west edge of the Sevier Plateau (see plate 1) and that ongoing geologic mapping is pushing the extent of the MGS farther west and north. See Hacker and others (2014) for figures 4A–F.*

2000 feet (600 m) thick in the northern Red Hills and western Markagunt Plateau that was tectonically detached from a lower plate sedimentary section along what he called the Red Hills shear zone. He recognized this detached plate on the basis of (1) a low-angle pulverized zone, variously present within or at the top of the Brian Head Formation, (2) upper-plate strata that dip more steeply than lower plate strata, (3) intense faulting and fragmentation of upper-plate rocks, (4) faults that appear to be restricted to the upper plate, and (5) omission of strata along the shear zone. Based on the age of the youngest rocks in the upper plate and of dikes that apparently cut the shear zone, he suggested that the shear zone was active 22.5 to 20 million years ago. He interpreted the Red Hills shear zone as a shallow structure that formed at a depth of less than about 6500 feet (2000 m) because of the absence of structures associated with detachments that form at mid-crustal depth. Further, he mapped numerous blocks of megabreccia derived from upper-plate rocks and noted similar blocks on the Markagunt Plateau, including those described by Sable and Anderson (1985) and Anderson (1993), and speculated on emplacement directions or mechanisms.

The Red Hills shear zone is locally the same slip surface that separates the Markagunt Megabreccia from undisturbed lower plate strata. Importantly, however, this statement holds true only north of what we now recognize as the ramp of the Markagunt gravity slide. We interpret the main horst of Jackrabbit Mountain, and the Markagunt Plateau south of the ramp (including the prominent Isom cliff that forms Black Ledge escarpment) to be undisturbed lower plate strata, not part of the upper plate of the Red Hills shear zone as Maldonado originally envisioned. Given our new discoveries of the extent of the Markagunt Megabreccia and our conclusions about its catastrophic mode of emplacement as outlined in this map and report, and given the requirements of nomenclatural precedence, we abandon the name Red Hills shear zone in favor of geomorphic (versus structural) terms to describe the Markagunt Megabreccia. We now refer to the Red Hills shear zone as the basal slip surface of the Markagunt gravity slide. Co-author Maldonado strongly disagrees with our new mapping and interprets that volcanic strata everywhere above the Brian Head Formation are detached along his Red Hills shear zone.

Geometry of the Markagunt Gravity Slide

The Markagunt gravity slide exhibits the full range of structural features commonly seen in modern landslides, including compression and resultant folding and thrust faulting in the landslide's toe area, simple translational movement across the main

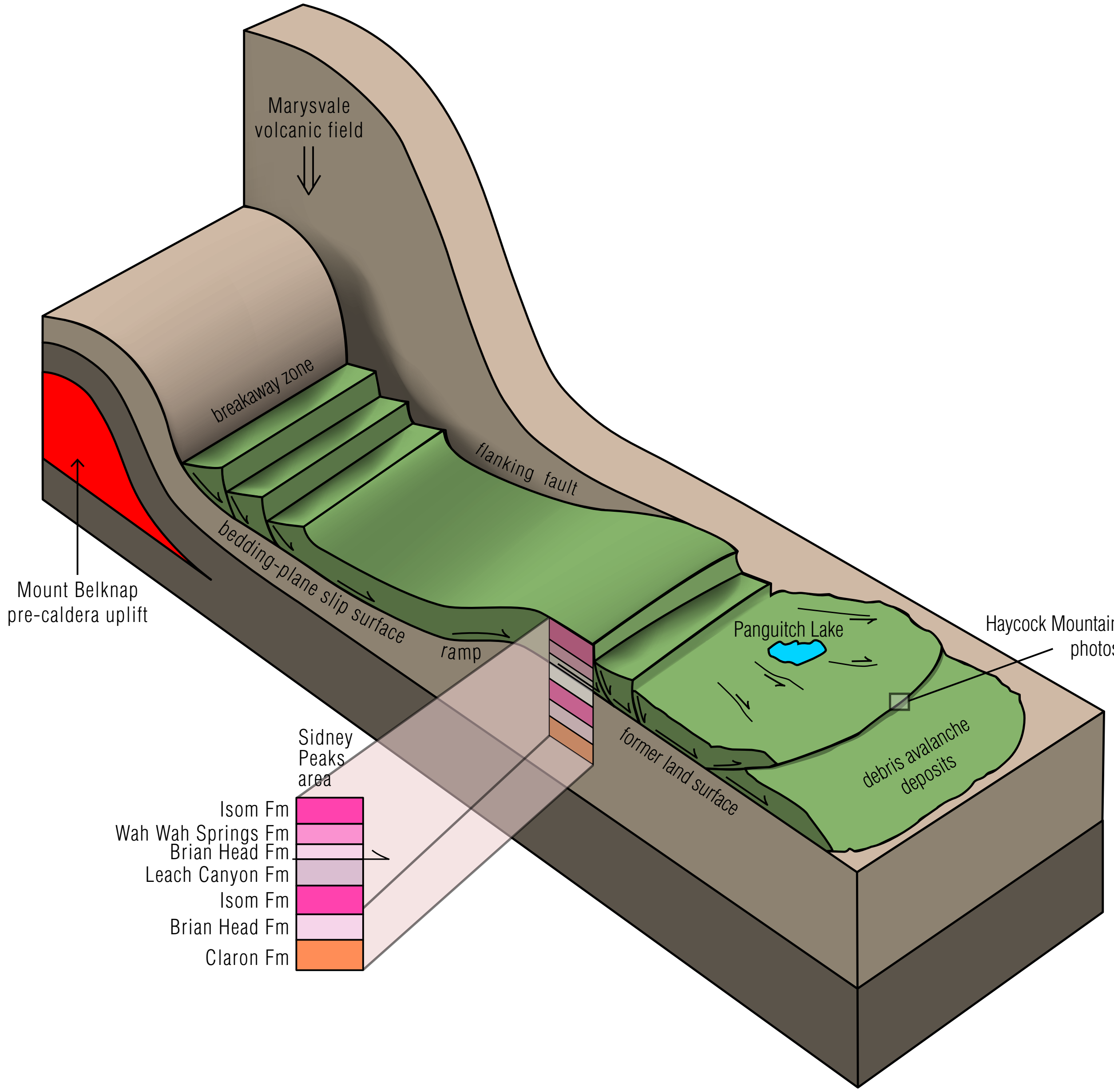


Figure 80. *Vertically exaggerated block diagram of an idealized gravity slide. Here, the trigger is suggested to be pre-caldera inflation of the Mount Belknap area, causing arching of overlying strata and consequent failure on over-steepened slopes. Note the four main bounding surfaces: the bedding-plane slip surface in mechanically weak, clay-rich rocks of the Brian Head Formation; the ramp, where the slide mass breaks upward to the surface; the former land surface, now covered by the slide mass; and the flanking failure, in essence a tear fault that bounds the lateral margin of the slide. The basal slip surface resembles shallow low-angle faults, complete with slickensided and striated surfaces, cataclastic zones, local pseudotachylyte, and brittle microfabrics. Extensional deformation characterizes the upper part of the slide, whereas compressional deformation characterizes the toe area. The main part of the gravity slide remains mostly intact, with individual blocks as much as several square miles in size, preserving a stratigraphy inherited from the source area. Distal portions of the slide mass disaggregate into debris avalanche deposits. Because gravity is the ultimate driver of such large landslides, the dip of the slip surface must be sufficient to overcome the shear or frictional strength of the detachment layer. Once moving, however, the slides can travel many miles over former land surfaces.*

body of the landslide, and extensional faulting in the upper parts of the landslide. Figure 80 shows a block diagram illustrating major components of such gravity-slide deposits, which are a special class of extremely large landslides.

Our new mapping, with that of Maldonado and Williams (1993a) and Knudsen (2014a, 2014b), suggests that the western extent of the Markagunt Megabreccia is west of the Red Hills and Jackrabbit Mountain and likely is buried under Long Hollow in the northwest corner of the map area. We map the east flanking fault on the western part of the Sevier Plateau. The location of the west and east flanking faults in the Beaver 30′ x 60′ quadrangle to the north is the subject of ongoing mapping. The ramp fault (figure 81) is best exposed about 2 miles (3 km) east-southeast of Red Creek Reservoir. The breakaway zone is on the south flank of the Mount Belknap caldera, about 25 miles (40

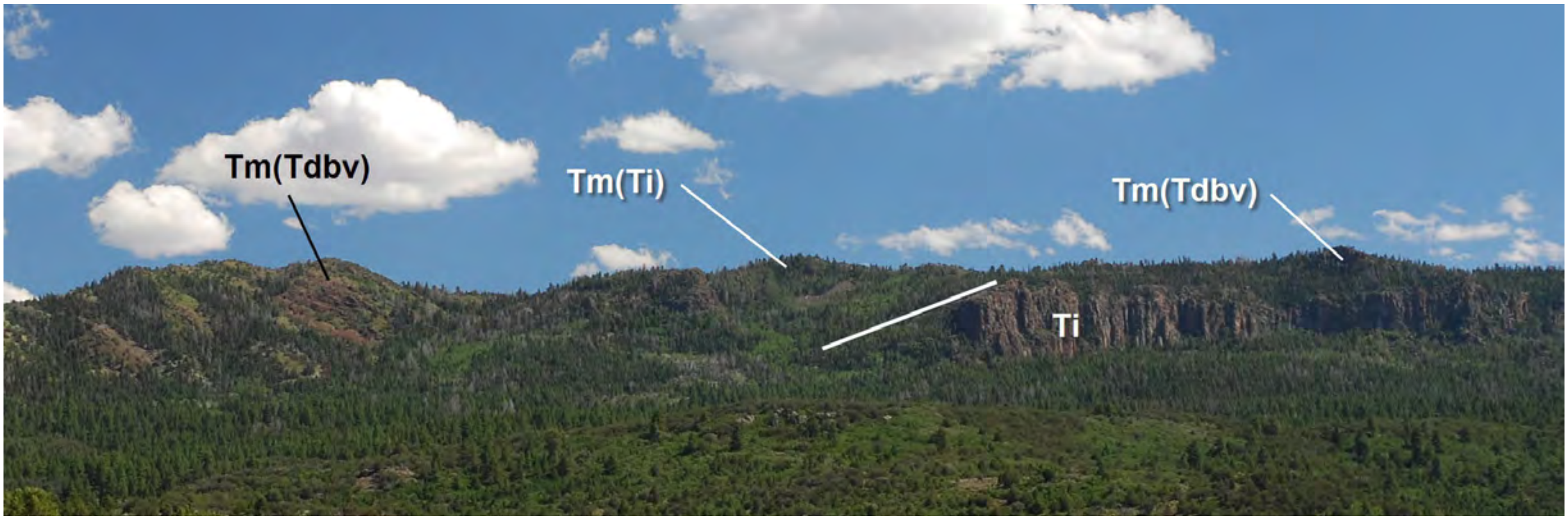


***Figure 81.** Black Ledge, showing possible ramp fault (white line) that places the Mount Dutton and Bear Valley Formations (Tm[Tdbv]), and Isom Formation (Tm[Ti]) on top of the Isom Formation (Ti) as part of the Markagunt gravity slide. South of this structure, which is about 2 miles (3 km) east-southeast of Red Creek Reservoir, Isom Formation forms an unbroken cliff that extends southward to Brian Head peak. View is towards the southeast.*

km) north of the map area, now being mapped in more detail by co-author Rowley. Given these constraints, we now estimate that the Markagunt gravity slide is at least 1600 square miles (4160 km^2) in size. Thus, the Markagunt gravity slide appears to be larger than the famous Heart Mountain detachment (HMD), long considered to be the largest subaerial gravity slide on earth at about 1300 square miles (3400 km^2) in size (the HMD resulted from catastrophic collapse, in the Eocene, of the Absaroka volcanic field in northwest Wyoming) (Malone and Craddock, 2008; Beutner and Hauge, 2009; Craddock and others, 2009).

The Markagunt Megabreccia was previously described as consisting of house-size to city-block-size blocks, or even blocks that are as much as one square mile (2.5 km^2) in size (Anderson, 1993; Sable and Maldonado, 1997a). We now know that the deposits of the Markagunt gravity slide exhibit considerable variation in deformation of upper-plate strata depending on where in the gravity slide one is looking. Between the breakaway zone and the ramp, strata above the bedding plane fault form intact blocks many tens of square miles in size; some blocks are gently tilted, but many are subhorizontal and, apart from deformation associated with the basal shear plane, reveal little or no evidence of being out of place. At and near the ramp, upper-plate rocks exhibit intense compressional deformation, with numerous thrust splays creating older-on-younger structural relations. Southward, on the former Miocene land surface, upper-plate rocks are commonly pervasively fractured and locally back-tilted into the slide plane, but commonly they too form large, mostly intact blocks many square miles in size. The Markagunt gravity slide is relatively easy to identify across the southern margin of its outcrop belt due to these older-on-younger relations. North of the ramp, however, identification of the gravity slide relies on rare glimpses of its basal shear plane, one of the reasons its flanking faults and breakaway zone are still poorly defined.

One remarkable feature of the gravity slide is the presence of sheared or crushed and rehealed clasts found in upper-plate strata near the basal shear or near smaller-displacement internal shears (figure 82). The clasts are rounded cobbles and boulders of quartzite—from the Claron Formation or from Miocene alluvial gravels, both ultimately recycled from Upper Cretaceous strata—and subangular volcanic clasts present in volcanic mudflow deposits of the Mount Dutton Formation. They account for much less than one percent of clasts in any given outcrop, but nonetheless some clasts exhibit severe in-place brittle deformation, some with mini-faults in a strain ellipsoid configuration, but are now rehealed; commonly, the clasts are present in colluvium overlying weathered outcrops, a sign that a significant-displacement shear must be nearby.

Markagunt gravity slide on the Miocene land surface: South of the ramp fault, on the former Miocene land surface, most of the Markagunt gravity slide was emplaced on resistant Isom Formation. Near its southern extent, the gravity slide was also emplaced on Leach Canyon Formation, Bauers Tuff Member, Harmony Hills Tuff, and Miocene alluvial gravels eroded into the Brian Head Formation.

The basal slip surface of the Markagunt gravity slide generally dips gently east (mimicking the regional dip of the plateau that was subsequently tilted following emplacement of the gravity slide) and south (because the source of the gravity slide is to the north; Sable and Maldonado, 1997a; Anderson, 2001), but near the toe at Haycock Mountain the basal slip surface dips north. The northward-dipping Isom Formation (caprock of Haycock Mountain) was reasonably interpreted by Anderson (1993) and Sable and Maldonado (1997a) as autochthonous, and they also

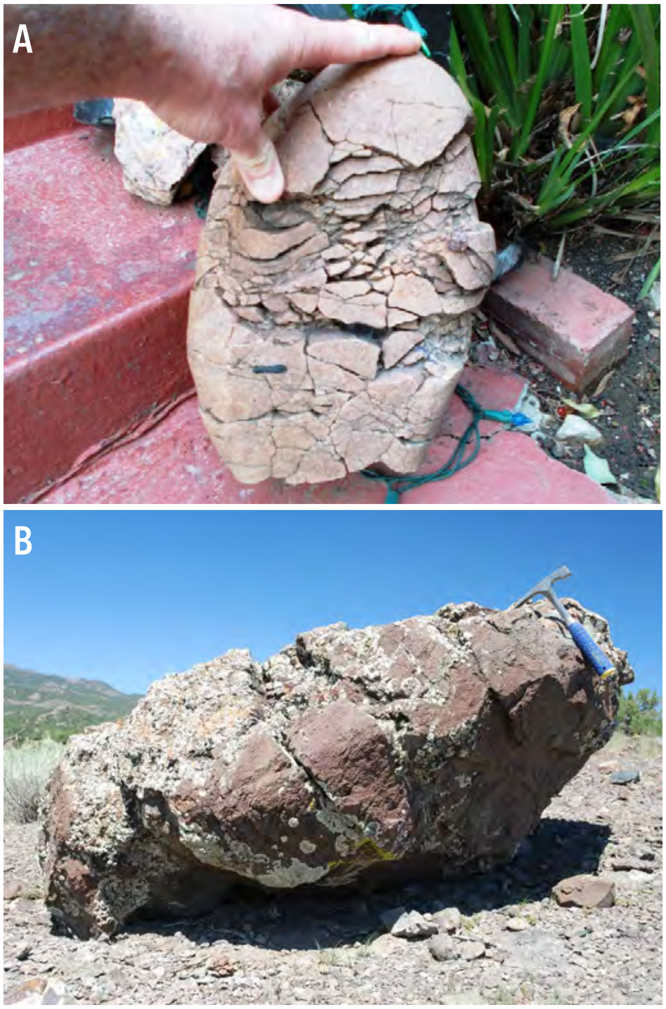


Figure 82. *Crushed and rehealed orthoquartzite cobble (A) collected from map unit (Tm[Ta]) northeast of Little Valleys, and a large andesitic boulder (B) from unit Tm(Td) about 5 miles (8 km) north-northeast of Panguitch. Clasts such as these are common near major shear planes of the Markagunt gravity slide.*

interpreted autochthonous Isom Formation at the type area of the gravity slide along Highway 143 east of Panguitch Lake.

However, we identified previously unreported basal breccias, shear zones, and associated clastic dikes exposed at the base of the gravity slide on the south side of Haycock Mountain (figures 83, 84, 85, 86, and 87). These exposures show that the entire Isom in this area is allochthonous and collectively demonstrate catastrophic emplacement by gravity sliding. Furthermore, slickenlines at the base of the gravity slide, as well as clastic dikes and Riedel shears (see, for example, Angelier and others, 1985; Petit, 1987), demonstrate north-to-south transport. Because moderately northwest-dipping Claron and Brian Head strata are present just south of Panguitch Lake, the northward dipping Isom likely reflects tilting due to a blind thrust fault in underlying strata, the westward equivalent of the Miocene Rubys Inn thrust fault of Lundin and Davis (1987), Lundin (1989), Nickelsen and others (1992), Merle and others (1993), and Davis (1999).

Sable and Maldonado (1997a) reported slickenlines on the basal slip surface of the gravity slide, as well as roche-moutonnée-like features and tilted beds that collectively suggest southward transport. Sable (1982) reported striations oriented about N40E on the top of the Leach Canyon on the northern side of Sidney Peaks, but could not rule out a glacial origin. We found identical striations at the top of the Leach Canyon along the western side of Sidney Peaks, where there are no glacial deposits, and thus interpret the striations as having resulted from emplacement of the Markagunt gravity slide.

Ramp of the Markagunt gravity slide: Part of the ramp fault is exposed at the northern end of Black Ledge, about 2 miles (3 km) east-southeast of Red Creek Reservoir (see figure 81). There, the ramp is roughly coincident with the west-northwest-striking northern Black Ledge fault of Anderson (2001). The area is of difficult access and has not been mapped in detail, but it corresponds to the northern terminus of the several-hundred-foot-high cliff of autochthonous Isom Formation that stretches southward to Brian Head peak. At the ramp, undivided Bear Valley and Mount Dutton Formations, as well as Isom Formation, are thrust southward over the Isom Formation. The ramp has not yet been adequately mapped eastward toward Panguitch and may be mostly concealed beneath post-gravity slide lava flows and basin-fill deposits. Westward, the ramp is concealed by modern landslide deposits and farther west is buried beneath Parowan Valley basin-fill deposits; erosion has stripped off upper-plate strata of intervening horst blocks. Still farther west, we infer the ramp is north of Jackrabbit Mountain in the adjacent Beaver 30′ x 60′ quadrangle, its exact location concealed by monotonous volcanic mudflow deposits of the alluvial facies of the Mount Dutton Formation. We infer a north-trending tear fault, buried under northern Parowan Valley, separates these two parts of the ramp.

Markagunt gravity slide north of the ramp: More than half of the Markagunt gravity slide lies above its bedding-plane slip surface (north of the ramp fault) and, there, the gravity slide is noted for large, subhorizontal or gently tilted blocks of mostly Bear Valley and Mount Dutton volcanic and volcaniclastic strata. Within these blocks there is little or no indication that they have moved. Indeed, to say that 5-mile-long (8 km) Cottonwood Mountain or Sandy Peak, for example, slid catastrophically many miles southward would leave most observers incredulous.

However, erosion of upper-plate rocks provides several windows down into lower plate strata, thus exposing the Markagunt gravity slide's basal shear plane. Two of the most revealing exposures found so far are those a few miles northwest of Panguitch and at the west end of Cottonwood Mountain. Collectively, these two as well as other sites compel us to sug-

Figure 83. *Markagunt gravity slide exposure (area of box; see figure 84) just south of Haycock Mountain on the southwest side of hill 8652, NW1/4SE1/4 section 5, T. 36 S., R. 6 W.; NAD83 coordinates: 37° 42' 39.66" N., 112° 33' 03.08" W. View is towards the southeast.*

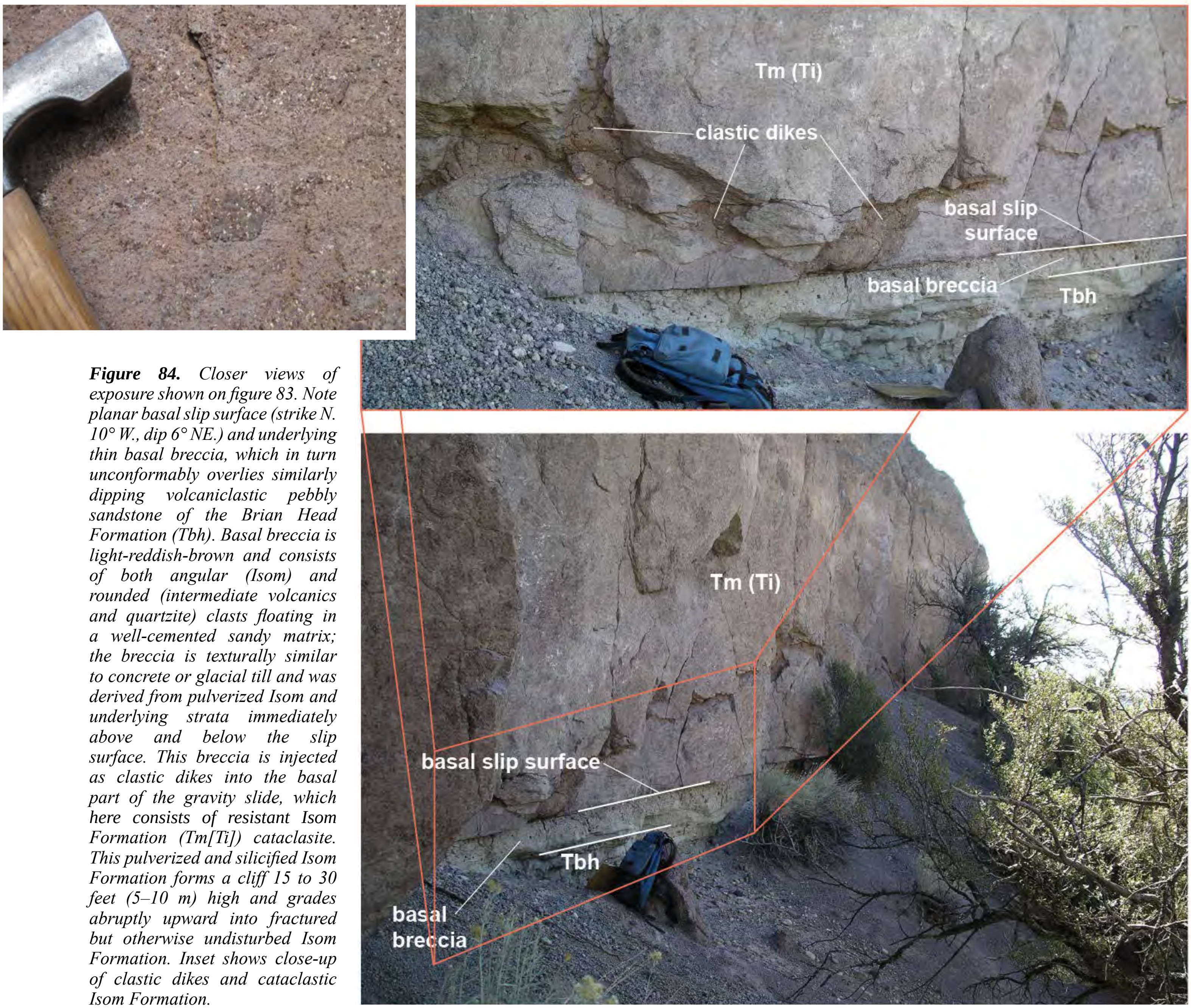


Figure 84. *Closer views of exposure shown on figure 83. Note planar basal slip surface (strike N. 10° W., dip 6° NE.) and underlying thin basal breccia, which in turn unconformably overlies similarly dipping volcaniclastic pebbly sandstone of the Brian Head Formation (Tbh). Basal breccia is light-reddish-brown and consists of both angular (Isom) and rounded (intermediate volcanics and quartzite) clasts floating in a well-cemented sandy matrix; the breccia is texturally similar to concrete or glacial till and was derived from pulverized Isom and underlying strata immediately above and below the slip surface. This breccia is injected as clastic dikes into the basal part of the gravity slide, which here consists of resistant Isom Formation (Tm[Ti]) cataclasite. This pulverized and silicified Isom Formation forms a cliff 15 to 30 feet (5–10 m) high and grades abruptly upward into fractured but otherwise undisturbed Isom Formation. Inset shows close-up of clastic dikes and cataclastic Isom Formation.*

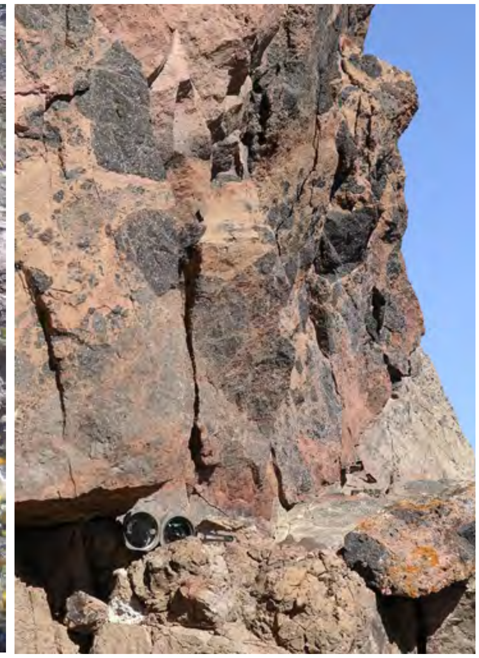



__Figure 85.__ Base of Markagunt gravity slide exposed just south of Haycock Mountain on the southeast side of an unnamed hill at the head of Little Coal Pit Wash, about one mile (1.6 km) west of the locale shown on figure 84, NE1/4 section 6, T. 36 S., R. 6 W.; NAD83 coordinates: 37° 42' 43.86" N., 112° 34' 14.98" W. The basal part of the gravity slide is a cataclasite of about 30 feet (10 m) of brecciated, pulverized, and resilicified Isom Formation (Tm[Ti]), which grades abruptly upward into fractured but otherwise undisturbed Isom Formation. Basal breccia, in shadow, unconformably overlies Miocene stream gravels (Ta) eroded into the Brian Head Formation. These gravels contain rounded clasts of the 26–27 Ma Isom Formation and so are younger than the Isom Formation, thus creating an older-on-younger relationship and showing that the entire Isom caprock of Haycock Mountain is part of the upper plate of the Markagunt gravity slide. Right photo shows cataclastic nature of Isom, which here locally appears similar to that of an autoclastic flow breccia.

__Figure 86.__ Close-up of slickenlines exposed at the west (left) side of figure 85. Slickenlines trend 20° NW and plunge about 15°. The base of the Markagunt gravity slide forms a planar surface that strikes N. 50° W. and dips 15° NE. Riedel shears at the base of the gravity slide demonstrate transport from north to south. Note basal breccia at base of gravity slide.

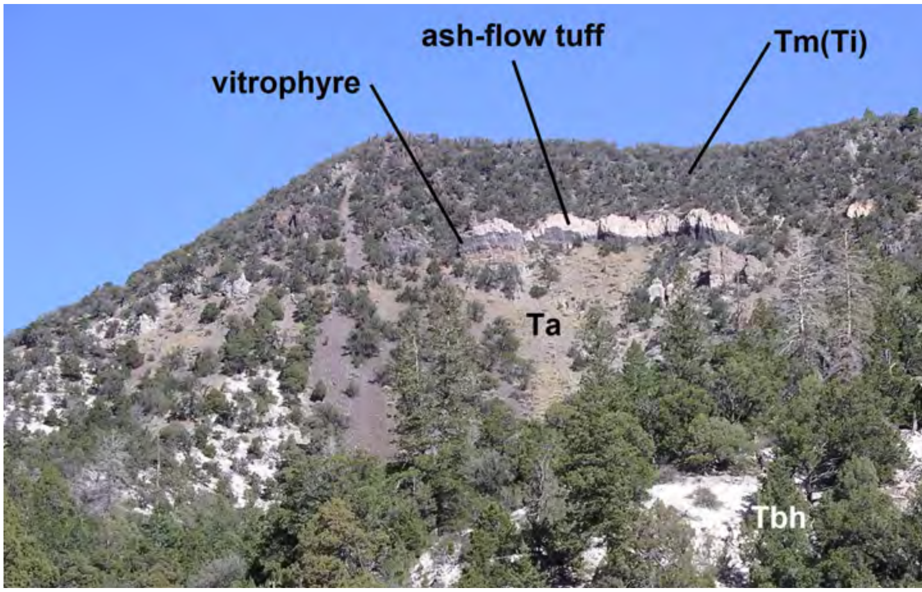


Figure 87. *Southwest flank of Haycock Mountain, just south of hill 9047. Note light-colored ash-flow tuff and its basal vitrophyre near the top of the mountain, shown in more detail in inset (map scale prevents showing the tuff on the map, but Biek and others (2014a) show it at a scale of 1:24,000). Volcaniclastic strata of the Brian Head Formation (Tbh) are in the foreground, overlain by several tens of feet of Miocene alluvial gravel (Ta), at least some of which is deformed and incorporated within the Markagunt gravity slide; "caprock" of Haycock Mountain, above the light-colored ash-flow tuff, is the Isom Formation as part of the Markagunt gravity slide (Tm[Ti]).*

Inset shows basal vitrophyre and associated ash-flow tuff; note underlying reddish-brown conglomerate (Tm[Ta]), which consists of rounded, pebble- to boulder-size clasts of intermediate volcanic rocks and quartzite—importantly, including clasts of Isom Formation—that here is about 100 feet (30 m) thick. The vitrophyre is about 20 feet (6 m) thick and is marked by a thin, brick-red devitrified zone at its top. It is overlain by a similar thickness of unwelded light-brown ash-flow tuff; chemically, both the vitrophyre and unwelded tuff are trachydacite. This vitrophyre yielded a K-Ar age on plagioclase of 22.3 ± 1.1 Ma (Rowley and others, 1994a; their sample 89USA2A) and an $^{40}Ar/^{39}Ar$ plagioclase age of 24.23 ± 0.17 Ma on the same sample (Sable and Maldonado, 1997a). We are uncertain how to correlate this ash-flow tuff —probably a local ash-flow tuff (shown separately on the more detailed map of Biek and others, 2014)—but it appears certain that it is not the Haycock Mountain Tuff; Rowley (in Hatfield and others, 2010) re-interpreted it to be an older (pre-Markagunt gravity slide) Bear Valley rhyolitic tuff. Regardless, its presence again demonstrates that the entire caprock of Haycock Mountain is allochthonous.

gest that the entire northern Markagunt Plateau is capped by allochthonous strata of the Markagunt gravity slide.

Northwest of Panguitch near Wide Hollow and Sandy Creek, co-author Hacker discovered shear planes and fractures lined and filled with obsidian-like glass, known as pseudotachylyte (see, for example, Spray, 2013) (figures 88 and 89). The glass is locally devitrified and is bounded by thin baked zones. Slickenlines, Riedel shears, and flow features on the glass-lined shears all demonstrate north-to-south transport. Glass-filled fractures penetrate upward and downward from the slip surface into adjacent strata where they extend for a few feet to a few tens of feet in length; they are commonly about 0.5 to 3 inches (1.3–7.5 cm) in width. Southward, we find Isom Formation thrust over Bear Valley strata, thus showing that these glass-lined shears are simply secondary shears above a more deeply buried basal shear zone. We are currently attempting to date this pseudotachylyte.

At the west end of Cottonwood Mountain, we find basal Wah Wah Springs Formation with a cataclastic and sheared texture above a striated slip surface that again demonstrates north-to-south transport. Maldonado and others (1989, 1992a, 1994, 1997, 2011) and Maldonado (1995) first reported on this locality. Clastic dikes of pulverized Wah Wah and Brian Head beds are found in the basal Wah Wah Springs Formation (figure 90), similar to clastic dikes we found at Haycock Mountain. To help constrain the age of the gravity slide, we sampled a basaltic dike that cuts the gravity slide (the same dike shown on figure 17 of Maldonado, 1995; see also Maldonado and others, 2011); an age is pending. This area is part of the Red Hills shear zone of Maldonado and others (1989, 1992a, 1994, 1997, 2011) and Maldonado (1995), the northern part of which we now reinterpret as the main failure plane of the Markagunt gravity slide as described previously. Basal Wah Wah Springs strata are similarly sheared and cataclastically deformed along the margins of Upper Bear Valley.

Timing of Emplacement of the Markagunt Gravity Slide

With our new mapping, we are able to further constrain the timing of emplacement of the Markagunt gravity slide. At the premature close of the USGS BARCO project in the mid-1990s, disagreement remained as to the age and extent of the

Figure 88. *Sandy Wash pseudotachylyte discovery site, about 5 miles (8 km) north-northwest of Panguitch (UTM NAD83 coordinates: 37° 53′ 47.69″ N., 112° 29′ 8.66″ W.). Note sharp contact between highly fractured sandstone of the Bear Valley Formation below and volcanic mudflow deposits of the Mount Dutton Formation above. Their contact is a shear plane lined with obsidian-like pseudotachylyte. Inset shows close-up of pseudotachylyte-filled dike at GPS receiver in center of photo. View is towards the north.*

Figure 89. *Pseudotachylyte-filled fracture at the Sandy Wash discovery site.*

Markagunt gravity slide, as described by Anderson (2001). The resolution of the age and extent of the gravity slide involves, among other issues, the Haycock Mountain Tuff in the type area of the Markagunt gravity slide, first described in detail by Anderson (1993) (figure 91). Because it is undeformed, he reasoned that the Haycock Mountain Tuff (^{40}Ar/^{39}Ar sanidine age of 22.75 ± 0.12 Ma; Sable, unpublished data, 1996) and underlying alluvial gravels are unconformable on and thus postdate the Markagunt gravity slide, as did Rowley and others (1994a) and Hatfield and others (2010). However, Sable and Maldonado (1997a) interpreted the Haycock Mountain Tuff to be a distal facies of the Leach Canyon Formation and part of the upper plate of the Markagunt gravity slide. Mapping in the Panguitch Lake 7.5′ quadrangle, described below (see description of the Leach Canyon Formation), however, now reconfirms that the 23.8 Ma Leach Canyon Formation and 22.8 Ma Haycock Mountain Tuff are different units of slightly different age as first suggested by Anderson (1993). Thus, the interpretation of Anderson (1993, 2001), and of Rowley and others (1994a) and Hatfield and others (2010), that the Haycock Mountain Tuff represents a post-Markagunt

Figure 90. *Basal shear plane of the Markagunt gravity slide at the west end of Cottonwood Mountain (near east-central edge of section 1, T. 33 S., R. 8 W.; NAD83 coordinates: 37° 57' 35.68" N., 112° 42' 0.72" W.). Here, basal Wah Wah Springs is a cataclasite that rides above a 1- to 2-foot-thick (0.3–0.6 m) basal breccia of finely comminuted Brian Head and Wah Wah debris; this breccia is also injected as dikes (right photo) that reach upward as much as 30 feet (9 m) into the 40-foot-thick (12 m), highly fractured Wah Wah Springs Formation. Grooves, striations, and Riedel shears on the basal shear plane suggest northeast-to-southwest movement.*

Figure 91. *Haycock Mountain Tuff (Thm) exposed in the north wall of Panguitch Creek, SE1/4 section 4, T. 35 S., R. 6 W. Here, the tuff is preserved in an early Miocene stream channel, underlain by several tens of feet of moderately cemented gravel (Thma, interpreted to postdate emplacement of the Markagunt gravity slide) that is in turn underlain by volcanic mudflow deposits of the Mount Dutton Formation (concealed by trees but interpreted to be part of the gravity slide).*

gravity slide tuff that partly filled a stream channel eroded into the gravity slide, appeared eminently reasonable.

However, the 22.75 Ma age of the undeformed Haycock Mountain Tuff is at odds with the fact that the Markagunt gravity slide overlies not only 22.03 Ma Harmony Hills Tuff but, at Haycock Mountain, Miocene alluvial gravels that contain rounded cobbles of the Harmony Hills Tuff. Exposures in Parowan Canyon (sections 14 and 15, T. 35 S., R. 9 W.) were interpreted by Maldonado and Moore (1995) as Harmony Hills Tuff in normal fault contact against autochthonous Isom Formation, but we reinterpret this Isom to be part of the Markagunt gravity slide that slid over the Harmony Hills Tuff; this fault is a gently southeast-dipping gravity-slide plane, not a west-dipping normal fault. This interpretation is consistent with similar exposures of the Harmony Hills Tuff in Summit Canyon to the west. The Markagunt gravity slide must thus be younger than about 22 million years old. We resampled the Haycock Mountain Tuff and report a new U-Pb age on zircon of 21.63 ± 0.73 Ma (UGS and AtoZ, 2013a). The Haycock Mountain Tuff does indeed represent a post-Markagunt gravity slide ash-flow tuff that partly filled a stream channel eroded into the gravity slide.

To summarize, catastrophic emplacement of the Markagunt gravity slide postdates the 22.03 Ma Harmony Hills Tuff. The gravity slide is overlain by undeformed Haycock Mountain Tuff now dated at about 21.6 Ma. We are currently attempting to date pseudotachylyte, which may yield a tighter constraint on the timing of emplacement. We are also dating a basaltic dike that intrudes the gravity slide near Cottonwood Mountain. Currently, we interpret the gravity slide to have been emplaced between about 21 and 22 million years ago, near the end of peak calk-alkaline volcanic activity in the Marysvale volcanic field.

Emplacement Mechanism of the Markagunt Gravity Slide

While we can confidently demonstrate catastrophic emplacement of the Markagunt gravity slide and interpret its emplacement as a single event, we remain uncertain as to what initiated gravity sliding. Previously, the gravity slide was interpreted as possibly having formed by either gravity sliding off the 20 Ma Iron Peak laccolith or other large unexposed intrusive bodies, or by low-angle, thin-skinned thrusting away from the intrusions, or failure off the backslope of inferred west-northwest-striking Miocene fault blocks. Co-author Rowley suggests that it may be related to a south-facing topographic scarp at the south end of the Tushar Mountains. But our current preferred hypothesis is that the Markagunt gravity slide is related to pre-caldera inflation of the Mount Belknap caldera. We describe each of these proposed scenarios below.

Relationship to Iron Peak laccolith: The Markagunt gravity slide was interpreted by Maldonado (1995) and Sable and Maldonado (1997a) to possibly have formed either by southward gravity sliding off the 20 Ma Iron Peak laccolith or other large unexposed intrusive bodies, or by low-angle, thin-skinned thrusting away from the intrusions. Anderson (2001) noted that the Iron Peak laccolith may be too small to have produced a dome large enough to trigger the Markagunt gravity slide. Even if the Iron Peak laccolith was once much larger—part of a much larger intrusive complex that underlies the Red Hills, northern Parowan Valley, northern Markagunt Plateau, and most of Panguitch valley as suggested by aeromagnetic anomaly and well data (Blank and Crowley, 1990; Blank and others, 1992, 1998; Rowley and others, 1994a)—we now know that Iron Peak is located too far south and west to possibly have been a trigger of the gravity slide. We also suspect that the laccolith is too young. We are attempting to redate the Iron Peak laccolith, but now interpret the laccolith as postdating emplacement of the Markagunt gravity slide; we interpret the excellent exposures at Cottonwood Mountain to be a result of subsequent doming of upper-plate strata above the intrusion.

In the past, the Iron Peak laccolith was originally thought to be a prime candidate as the trigger of the Markagunt gravity slide because it is about the correct age and because in this larger intrusive complex, most, if not all, intrusions are laccoliths (Anderson, 1965; Anderson and Rowley, 1975; Anderson and others, 1990a, 1990b). It was thought that inflation of this larger complex, or several individual laccoliths within it, may have triggered catastrophic emplacement of the gravity slide. Mapping in the Iron Springs mining district and along the Iron Axis west and southwest of Cedar City, where a complex of laccoliths produced many gravity slides similar to but much smaller than the Markagunt gravity slide (Hacker, 1998; Hacker and others, 2002, 2007; Rowley and others, 2008; Knudsen and Biek, 2014), demonstrates laccolith emplacement as a trigger for catastrophic gravity slides.

Because of the apparently inadequate size of the Iron Peak laccolith, Anderson (2001) suggested that gravity slides originated from southward failure off the backslope of inferred west-northwest-striking Miocene fault blocks. Our new mapping shows that Anderson's northern Black Ledge fault is the ramp fault of the Markagunt gravity slide, and his central Black Ledge fault is a relay ramp between en echelon parts of the Black Ledge and East Bear Valley faults. We find no evidence for south-tilted fault blocks as originally inferred during early mapping of the northern Markagunt Plateau.

Relationship to southern Tushar Mountains topographic scarp: The trigger of the Markagunt gravity slide may be related to the current south-facing topographic scarp between the Tushar Mountains and the Markagunt Plateau. The top of the present scarp, about 11 miles (18 km) north of the map area, is held up by a volcanic field of resistant, 22 to 23 Ma basalt flows ("older basalts" of Anderson and Rowley, 1975) covering an area of about 50 square miles (130 km^2) to a thickness of about 1500 feet (500 m), not including 1000 feet (300 m) of the mafic gravels of Gunsight Flat derived from the flows. In their compilation of the Beaver 30′ x 60′ quadrangle, Rowley and others (2005) mapped the basalts as potassium-rich mafic lava flows; in detailed mapping they were called the mafic lava flows of Birch Creek Mountain and mafic lava flows of Circleville Mountain, from base to top. These rocks are the oldest extension-related bimodal volcanic rocks in the area (Anderson and others, 1990a, 1990b). A generally east-trending fault is mapped on the Tushar Mountains–Markagunt Plateau scarp, but it is concave to the north and downthrown to the north, opposite to the throw one would expect if it had any causal relationship to the scarp. Yet this fault may have an important relationship to the gravity slide if it represents subsidence of a trap-door caldera-like feature into the vent area or magma chamber that erupted the basalts, presumably later in the eruptive cycle. The northern side of this fault, as with a caldera, was a high north-facing scarp that shed the mafic gravels of Gunsight Flat northward into the basaltic volcanic field (Anderson and others, 1990a, 1990b). This northward transport of sediment also is the opposite of what one would expect from looking at the 2000-foot-high (600 m) south-facing scarp. So clearly the south-facing scarp did not exist at 22 Ma, yet one may interpret that extension had just begun in the area, resulting in the start of uplift of the Tushar Mountains and Markagunt Plateau.

As mapped by Anderson and others (1990a, 1990b), the 22 to 23 Ma basaltic field appears to be domed, as indicated by radial (outward, to the south) dips of the flows and mafic gravels in the southern part of the field. Furthermore, faults in the field are crudely radial and concentric, and result in a distinctive mosaic pattern of intersecting faults, an indicator of roofs of underlying intrusions (Steven, 1989). The crest of the dome is at Anderson Meadow Reservoir, in a canyon near the center of the basaltic field, about 10 miles (16 km) east-southeast of Beaver; here the rocks are hydrothermally altered and silicified. The map patterns suggest that a mafic pluton underlies the field, perhaps a laccolith similar to the 20 Ma gabbroic Iron Peak laccolith (Anderson and Rowley, 1975; Fleck and others, 1975; Spurney, 1984; Maldonado and others, 2011). Gravity data reveal a gravity high at or beneath the basaltic field (Campbell and others, 1999), but these data are on the southern edge of their gravity map and could be equally well explained by the basalts themselves.

Relationship to Mount Belknap caldera: Based on our new mapping that extends the Markagunt gravity slide north of the Iron Peak laccolith, our preferred trigger is thus farther north and is the subject of ongoing mapping. Answers to all geologic conundrums in a map area cannot necessarily be answered before deadlines require publication of the subject map, especially when the answers may lie outside the map area. Therefore, lacking specific data on the possible trigger requires speculation based on the geology to the north, as mapped by Cunningham and others (1983) and Rowley and others (2002, 2005). Based on this mapping, it seems possible that the gravity slide formed due to failure on oversteepened slopes that resulted from pre-caldera inflation in the Marysvale volcanic field, an idea first suggested to senior-author Biek by volcanologist Gene Smith (University of Nevada, Las Vegas). A caldera that is of permissible age is the bimodal, 18 to 20 Ma Mount Belknap caldera 25 miles (40 km) north of the map boundary (Cunningham and others, 1998; Rowley and others, 2002, 2005). Structural overprinting by the Mount Belknap caldera may have destroyed the breakaway zone. The caldera erupted the 19 Ma Joe Lott Tuff Member of the Mount Belknap Volcanics, which may partly bury the breakaway zone.

We suggest that the continued late-stage growth of the southern Marysvale volcanic field, stacking ever more volcanic rocks on a structurally weak substrate of Brian Head strata, created the conditions necessary for gravity sliding. It seems possible that a nudge by an intruding pluton beneath the basaltic field or beneath what was to become the Mount Belknap caldera, or one of many earthquakes during magmatism or the start of regional extension in the area, could have triggered the catastrophic gravity slide, even if the structural slope on the Brian Head Formation were only a few degrees to the south. Further, we suggest that the critical point of failure would have been near the distal, southern toe of the Marysvale volcanic field. At some point, the southward-thinning wedge of volcanic rock was no longer an adequate buttress for the ever-growing pile of volcanic rock to the north. The entire volcanic section on the south flank of what is now the Mount Belknap caldera may have slumped southward to initiate the gravity slide, but the thick wedge of volcanics in the source area need not have moved far. That initial slump would have provided the “push” needed to get the slide moving over the characteristically weak Brian Head substrate. Structural overprinting by subsequent basin-range extension complicates estimates of the initial dip of the slip surface; it may have been less than a few degrees near the breakaway zone and nearly subhorizontal south of the ramp.

Relationship to Paunsaugunt thrust fault system: Interestingly, as noted by Sable and Maldonado (1997a) and Davis (1997a, 1997b), the south margin of the Markagunt gravity slide is on trend with well documented, east-trending, south-vergent thrust faults involving Upper Cretaceous and Paleocene-Eocene Claron Formation on the Paunsaugunt Plateau. These thrust faults, including the Rubys Inn thrust fault, are interpreted to represent gravitational loading and collapse of the southern part of the Marysvale volcanic field (or possibly coeval batholithic emplacement) (Davis and Krantz, 1986; Lundin, 1989; Davis and Rowley, 1993; Merle and others, 1993; Davis, 1999). In the “two-tiered” model of Davis (1997a, 1997b, 1999), the Markagunt gravity slide is but one structure—an upper-level part—of a second, deeper series of Tertiary thrusts directed outward from the southern Marysvale volcanic field, which spread and collapsed under its own weight, resulting in southward-directed thrust faults rooted in evaporite strata of the Middle Jurassic Carmel and Arapien Formations (see discussion of the Paunsaugunt thrust fault system). On the Markagunt Plateau, gently northwest-dipping Claron and Brian Head strata south of Panguitch Lake, and gently north-dipping Isom Formation at Haycock Mountain, may reflect folding in the upper plate of an east-trending, south-vergent, blind thrust fault (see plate 1), which may be the westward continuation of the Rubys Inn thrust fault.

Late Tertiary and Quaternary Extensional Structures of the Transition Zone

Hurricane Fault Zone

The Hurricane fault zone is a major, active, west-dipping normal fault that stretches at least 155 miles (250 km) from south of the Grand Canyon northward to Cedar City and beyond (figure 92). 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 about 7500 feet (2300 m) along the north part of the fault (Hurlow, 2002), but estimating such displacement is complicated, as described by Biek and others (2009). Like most long normal faults, the Hurricane fault zone is composed of discrete segments that tend to rupture independently. The Cedar City segment lies immediately west of the map area where it forms the east margin of the Basin and Range Province at that latitude; impressive fault scarps near the south end of the segment near Shurtz Creek, just southeast of Cedar City, attest to the cumulative effect of several late Pleistocene to Holocene surface faulting earthquakes on this segment of the fault (Lund and others, 2007).

Some workers end the Hurricane fault just north of Cedar City (see, for example, Hurlow, 2002; Lund and others, 2007); however, we suggest, as did Maldonado and others (1994), that the Hurricane fault extends farther north where it bounds the west side of the Red Hills (figure 1). In his evaluation of groundwater resources of Cedar Valley, Hurlow (2002) called the fault zone on the west side of the Red Hills the eastern basin-bounding fault system (EBBFS), the master fault bounding the east margin of Cedar Valley. Whether this is a continuation of the Cedar City segment of the Hurricane fault zone or perhaps a new fault segment is uncertain, and some workers may still prefer to end the Hurricane fault at the north side of Cedar City. The naming of faults is problematic in part because of the way faults tend to evolve over time through linkage of once-discrete sections. A corollary of this is the development of sub-basins within a larger extensional basin, as revealed in the model relating faulting and basin development by Schlishce and Anders (1996) (figure 93) and described in detail for Cedar Valley by Hurlow (2002).

The paleoseismicity of the Hurricane fault zone was investigated by Lund and others (2001, 2002, 2007), who noted that the most recent surface faulting earthquake on the fault in Utah occurred in the latest Pleistocene or early Holocene, on what they deemed the northern segment of the fault southwest of 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. The fault is considered capable of generating damaging earthquakes of about magnitude 7.0; the 1992 magnitude (ML) 5.8 St. George earthquake may have occurred on the west-dipping subsurface projection of the Hurricane fault (Arabasz and others, 1992; Pechmann and others, 1995).

Lund and others (2007) used a remnant of the Wood Knoll lava flow, high above the confluence of Ashdown and Crow Creeks in Cedar Canyon (figure 94), as a surrogate for a long-term slip-rate estimate of 21 inches per 1000 years (0.53 mm/yr) for the Cedar City segment. This estimate is about twice what they determined for displaced late Pleistocene fan alluvium at the south end of the segment near Shurtz Creek. Southward, near Ash Creek Reservoir on the Ash Creek segment of the Hurricane fault, they used the displaced Pintura lava flow to document an average slip rate of about 22 inches per 1000 years (0.56 mm/yr) during the last 850,000 years for that segment of the fault.

The Hurricane fault may vary in structural style along its length. Anderson and Barnhard (1993a, 1993b) and Anderson and others (2013) summarized evidence showing that the northern part of the Hurricane fault zone—the Cedar City segment and possibly the Ash Creek segment—may have 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 is a reactivated Sevier-age thrust fault. Near Toquerville, however, the Hurricane fault trends due south, away from the southwest trend of Sevier-age folding. Pliocene to Quaternary displacement on the southern four segments of the fault is normal dip slip, and Biek and others (2009) and Billingsley and Workman (2000) reported no evidence suggesting this part of the Hurricane fault is a reactivated Sevier-age structure.

The inception of faulting along the Hurricane fault zone is poorly constrained but appears to have begun earlier on its northern two segments. Faulting probably started by about 10 to 12 million years ago, the age of the first basaltic volcanism in northern Cedar Valley (Anderson and Mehnert, 1979; Rowley and others, 1979, 1981, 2008; Hurlow, 2002). Rowley and others (1978) used the then-known distribution of regional ash-flow tuffs to suggest that basin-range extension may have begun as early as about 20 million years ago, creating a west-facing topographic highland, whose west edge was generally coincident with the ancestral Hurricane fault and that blocked or partly blocked 20 to 26 Ma ash-flow tuffs from flowing east of the Hurricane fault more than about 10 miles (15 km). However, since then, additional mapping in the Cedar City-Marka-

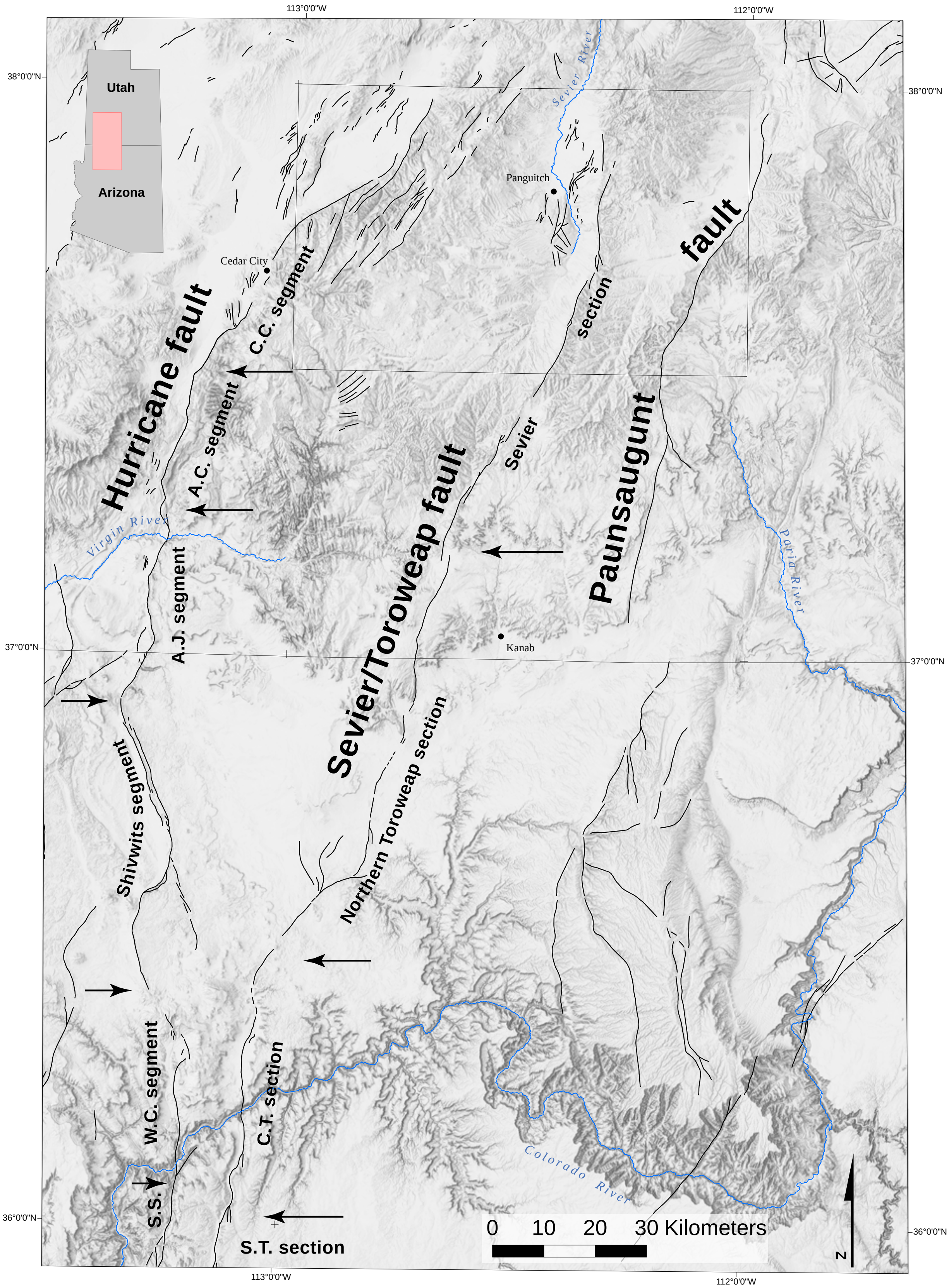


***Figure 92.** The Hurricane, Sevier-Toroweap, and Paunsaugunt fault zones in the Basin and Range–Colorado Plateau transition zone. Arrows indicate fault section boundaries, each segment or section of which has a different rupture history and rate of long-term slip. Our mapping suggests that the Hurricane fault zone continues north of Cedar City, but this portion of the fault has not been studied. Hurricane fault abbreviations: S.S., southern segment; W.C., Whitmore Canyon segment; A.J., Anderson Junction segment; A.C., Ash Creek segment; C.C., Cedar City segment. Sevier/Toroweap fault abbreviations: S.T., Southern Toroweap section; C.T., Central Toroweap section. Box shows location of Panguitch 30' x 60' quadrangle. Modified from Lund and others (2007, 2008).*

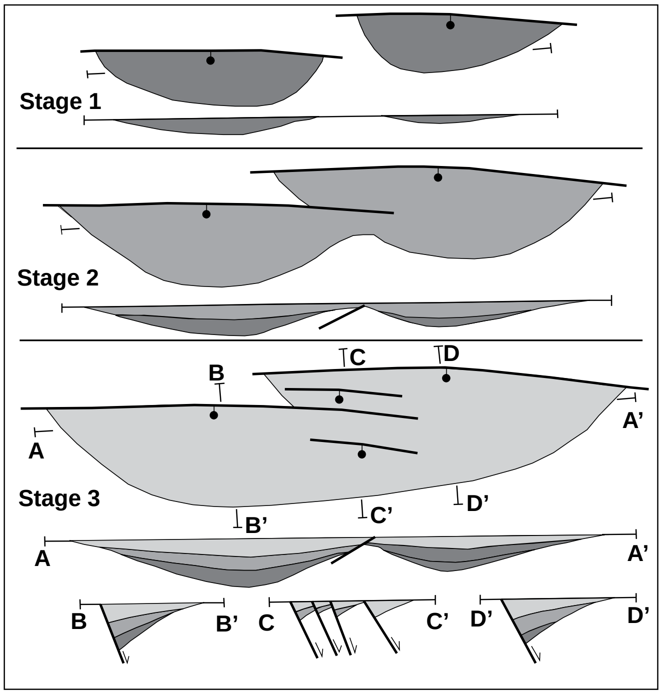


Figure 93. Model relating faulting and basin development; each of the three stages shows a plan view and cross section. Note how once-discrete fault segments and associated basins become linked over time. From Schlische and Anders (1996).

gunt Plateau area has failed to find conclusive control by faulting on the distribution of these ash-flow tuffs, a conclusion also reached by Anderson and Mehnert (1979). As described earlier, we now interpret the distribution of regional ash-flow tuffs to reflect the filling and incision of paleovalleys that radiated away from calderas at the Utah-Nevada border. Finally, Hurlow (2002) noted that the oldest (10–12 Ma) basaltic lava flows noted above are interbedded in his seismically defined youngest basin-fill unit, and that older basin-fill deposits are exposed in the North Hills and Cross Hollow Hills west and south of Cedar City. Although these older basin-fill deposits are undated, Hurlow (2002) suggested that basin development and thus the initiation of faulting on the Hurricane fault zone began about 12 million years ago. We do know that the southern segments of the fault are clearly younger. Billingsley and Workman (2000) described offset relations of late Tertiary and Quaternary basaltic lava flows in Arizona and showed that, on the basis of 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.

Parowan Valley Faults

Parowan Valley is an asymmetric graben bounded on the east by the Parowan-Paragonah fault zones and on the west by the antithetic Red Hills fault zone. Scarps are present on old fan alluvium east-southeast of Summit and at the entrance to Parowan Canyon, but we did not find scarps on unconsolidated deposits along the Paragonah portion of the fault zone. The central part of the valley is cut by numerous scarps in old fan alluvium (Anderson and Christenson, 1989; Maldonado and Williams, 1993a, 1993b). Total displacement on the faults is constrained by interpretation of seismic-reflection profiles. Hurlow (2002) interpreted more than 6600 feet (2000 m) of displacement on the Parowan fault and about 3300 to 6600 feet (1000–2000 m) on the Red Hills fault.

Figure 94. Remnant of the Wood Knoll lava flow, high above the confluence of Ashdown and Crow Creeks in Cedar Canyon; note thin, reddish-brown alluvial gravels below flow. Lund and others (2007) used this flow as a surrogate for a long-term slip-rate estimate of 21 inches per 1000 years (0.53 mm/yr) for the Cedar City segment of the Hurricane fault zone.

Paleoseismic investigations have not been done on these faults other than several scarp profiles (Maldonado and Williams, 1993b). Scarp morphology led Black and Hecker (1999) to classify the most recent movement on the faults as less than 15,000 years ago; latest movement on the Paragonah and Red Hills faults was estimated to be less than 130,000 years ago. However, Williams and Maldonado (1995) reported that most faults in the area have small (<1 m) offsets in possible Holocene fan alluvium. Just southeast of Paragonah, the 0.45 Ma Water Canyon lava flow crosses a relay ramp between two en echelon sections of the Parowan and Paragonah faults (Maldonado and Moore, 1995). At the mouth of Water Canyon, the flow reveals about 250 feet (75 m) of displacement, yielding a long-term slip rate of about 0.17 mm/yr (about 550 feet/mya) for the Paragonah fault, although to what extent if any the lava flow cascaded over a possible pre-existing fault scarp is unknown. The 1.0 Ma Summit lava flow is truncated by the Parowan and Red Hills faults and thus must underlie the southernmost end of Parowan Valley.

Interpretation of gravity data and seismic reflection profiles shows that Parowan Valley basin-fill deposits are in excess of 2000 feet (600 m) thick just a mile (1.6 km) north of Summit, and that they are at least 3000 feet (900 m) thick in the central part of the basin (Hurlow, 2002). This basin fill is the principal source of drinking water for most area residents. Recent land subsidence and earth fissure formation in Parowan Valley, and in areas immediately southwest of the map area in Cedar Valley (Katzenstein, 2013), are tied to excessive groundwater pumping (DuRoss and Kirby, 2004), the subject of ongoing Utah Geological Survey investigations.

Markagunt Plateau Horsts and Grabens

The northwest margin of the Markagunt Plateau is cut by paired high-angle normal faults that create a series of horsts and grabens that step down from the plateau to the adjacent Basin and Range (figure 71) (Maldonado and others, 1997). Horst blocks expose increasingly older Upper Cretaceous strata westward across the margin of the plateau, reflecting its position on the east flank of the Kanarra anticline. Grabens preserve younger Claron, Brian Head, and Oligocene and Miocene ash-flow tuffs as well as the early Miocene Markagunt gravity slide. Displacement on the faults decreases southward and many of the faults terminate at an east-northeast-trending, down-to-the-north, high-angle normal fault (figure 95). Overall, the collection of horsts and grabens form a highly faulted relay ramp between the Paragonah fault zone and the Hur-

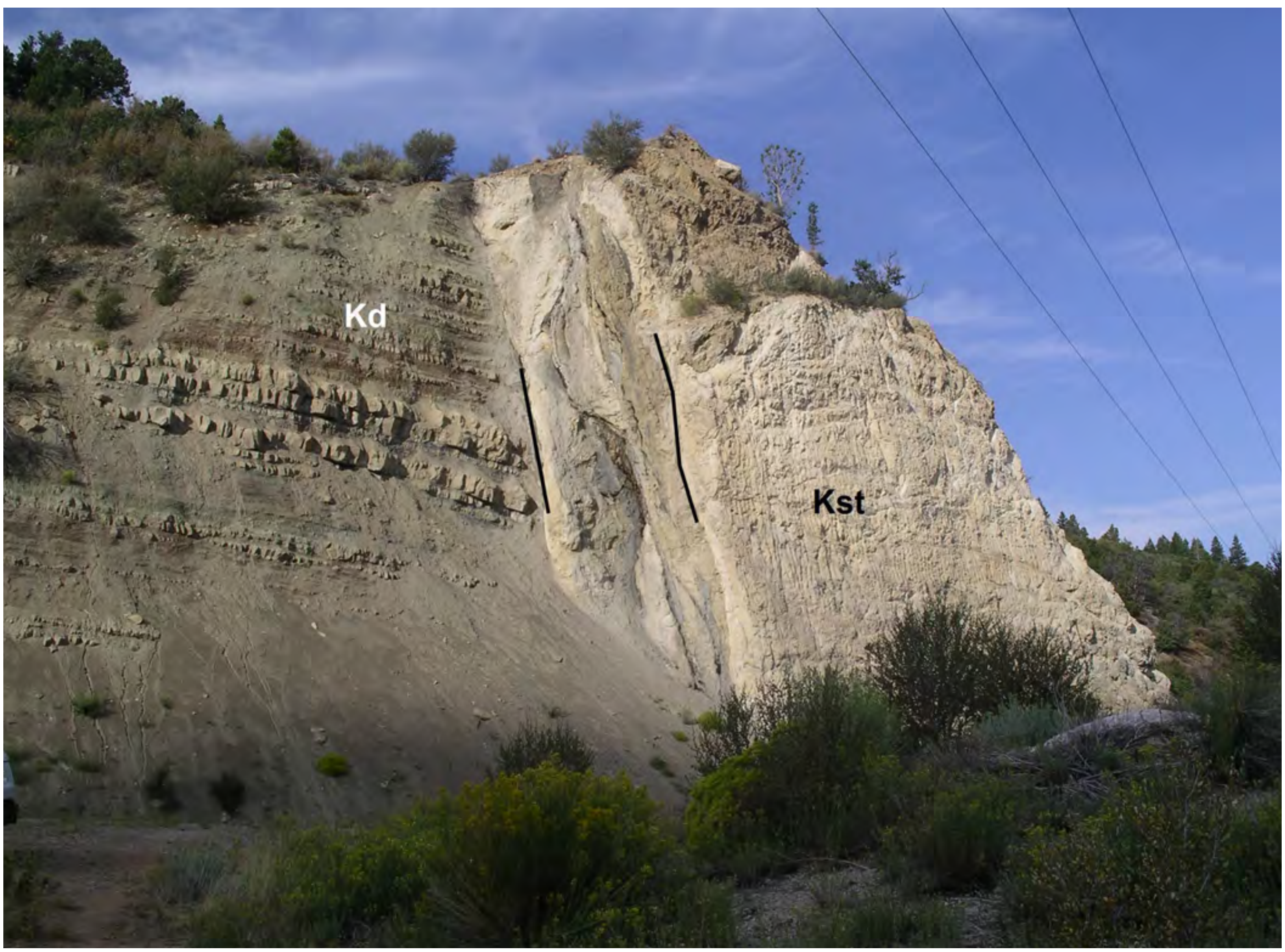


Figure 95. *Road cut on the south side of Utah Highway 14 near the junction of Coal Creek and Last Chance Canyon. Here, the southern part of the Navajo Ridge fault places Tibbet Canyon Member of the Straight Cliffs Formation (Kst) against gently tilted Dakota Formation (Kd). The fault zone itself (between the black lines) includes slivers of steeply north-northwest dipping Dakota coal, mudstone, and sandstone beds. Compare to figure 13 of Maldonado and others (1997).*

ricane fault zone. Preliminary mapping by Anderson (1965) suggested that the faults were en echelon, and Maldonado and others (1997) described them as anastomosing, but our more complete map of the west flank of the plateau shows that, while the interiors of the graben and horst blocks are cut by numerous normal faults, the major bounding faults themselves are mostly large, through-going, high-angle normal faults.

Because displacement on the horst and graben faults dies out to the south, we envision that the faults formed due to simple torsion of the relay ramp. We find no evidence of left-lateral strike-slip or oblique-slip displacement on these faults. Anderson and Bucknam (1979) reported probable Holocene displacement on the Middle Ridge fault, but misidentified landslide structures for the fault itself; there, modern landslide deposits derived principally from remobilization of the Markagunt gravity slide clearly fill the Fiddlers Canyon graben. Maldonado and others (1997) speculated that the grabens and horsts were the result of minor accommodation above the Black Ledge fault, which they inferred exhibited a magnitude of displacement similar to that of the Hurricane and Paragonah fault zones. However, displacement on the Black Ledge fault clearly dies out southward in Coal Canyon and we now view it simply as the easternmost large down-to-the-west normal fault at the west margin of the Markagunt Plateau.

Rowley and others (1981) and Hurlow (2002) reasoned that the inception of modern basin-range faulting likely began about 10 to 12 million years ago in the Cedar City area. However, our mapping shows that the Iron Peak graben juxtaposes Bear Valley and Brian Head strata that were later intruded by the 20 Ma Iron Peak laccolith. This suggests that early basin-range extension in the region began prior to about 20 million years ago.

Sevier Fault Zone

The Sevier fault zone stretches from south of the Grand Canyon northward to at least the north edge of the map area, a distance of about 150 miles (250 km); by convention, it is known as the Toroweap fault in Arizona (Lund and others, 2008) (figure 92). Northward, fault-bounded basins in the heart of the Marysvale volcanic field, including the Circleville, Marysvale, and Monroe areas, are on structural trend with the Sevier fault zone, indicating that the fault bisects the volcanic field and links up with down-to-the-west normal faults in southern Sevier Valley south of Richfield (Hintze and others, 2003; Rowley and others, 2005).

Paleoseismic reconnaissance of the Sevier fault zone suggests that it may be divided into several seismogenic segments that have unique rupture histories, but Lund and others (2008) conservatively described the inferred segments as fault sections until distinct rupture histories can be demonstrated. The southernmost part of the fault in Utah is part of the northern Toroweap section, which terminates near Clay Flat south of Mt. Carmel. Following Black and others (2003), the part of the fault from Clay Flat northward to Panguitch is termed the Sevier section. Although Lund and others (2008) suggested that Hillsdale Canyon may mark a segment boundary on the fault zone, our mapping strongly supports the presence of such a boundary farther north, east of Panguitch. There, a complexly faulted bedrock high and associated graben in the hanging wall separate the Hatch part of the valley on the south from the Panguitch part of the valley on the north. In this zone, numerous scarps are present on late Pleistocene fan alluvium where they collectively form a northeast-trending graben, and a broad, northeast-trending syncline folds late Tertiary fan alluvium and the Fivemile Ridge basaltic lava flow on the west slope of the Markagunt Plateau (figure 96). The course of Panguitch Creek

Figure 96. *View southwest past Panguitch, subparallel to the axis of the Panguitch syncline. The syncline, which may be related to a possible fault segment boundary of the Sevier fault zone, corresponds to the juniper-covered slope underlain by the Fivemile Ridge lava flow. Hancock Peak is the prominent cinder cone on the skyline beyond Panguitch.*

occupies the trough of this syncline. This area also corresponds to the south end of the Sevier Plateau, which is of higher elevation and relief than the Paunsaugunt Plateau. This northern, "Panguitch" section of the fault zone appears to terminate several miles north of the map area, where it devolves into an anastomosing series of down-to-the-west and down-to-the-east normal faults before linking up with the eastern bounding fault of Circleville Valley (Rowley and others, 2005).

Tectonic displacement on the Sevier fault zone generally increases south to north along the length of the Sevier fault (Lund and others, 2008), but in this map area, displacement is hard to constrain because basin-fill deposits conceal bedrock strata in the hanging wall. At Red Canyon, seismic reflection data indicate about 3000 feet (900 m) of displacement (Lundin, 1987; Davis, 1999). Near the southern edge of the map area, Biek (2013a) estimated about 4500 feet (1370 m) of tectonic offset.

Lund and others (2008) noted that, apart from scarps in the hanging wall near Panguitch, no scarps on unconsolidated surficial deposits are present along the main trace of the Sevier fault in Utah, thus precluding trenching to evaluate the paleoseismic history of the Utah portion of the fault zone. However, they used dated basaltic lava flows displaced by movement on the fault to determine vertical slip rates and recurrence intervals at three locations on the fault—two locations near Red Canyon, and, south of the map area, at Black Mountain.

The 0.5 Ma Red Canyon lava flow erupted on the downthrown side (hanging wall) of the Sevier fault immediately north of Red Canyon; at that time, the fault juxtaposed tilted, non-resistant late Tertiary basin-fill deposits in the hanging wall against non-resistant Brian Head strata in the upthrown footwall. Thus, the fault did not form a significant topographic barrier as it does today, and the Red Canyon flow crossed onto the footwall. It has since been displaced about 650 feet (200 m), yielding a vertical slip rate of 0.4 mm/yr and a recurrence interval between surface faulting earthquakes of about 5000 years (figure 97). Similarly, just south of Red Canyon, the 5 Ma Rock Canyon lava flow, which erupted on the eastern

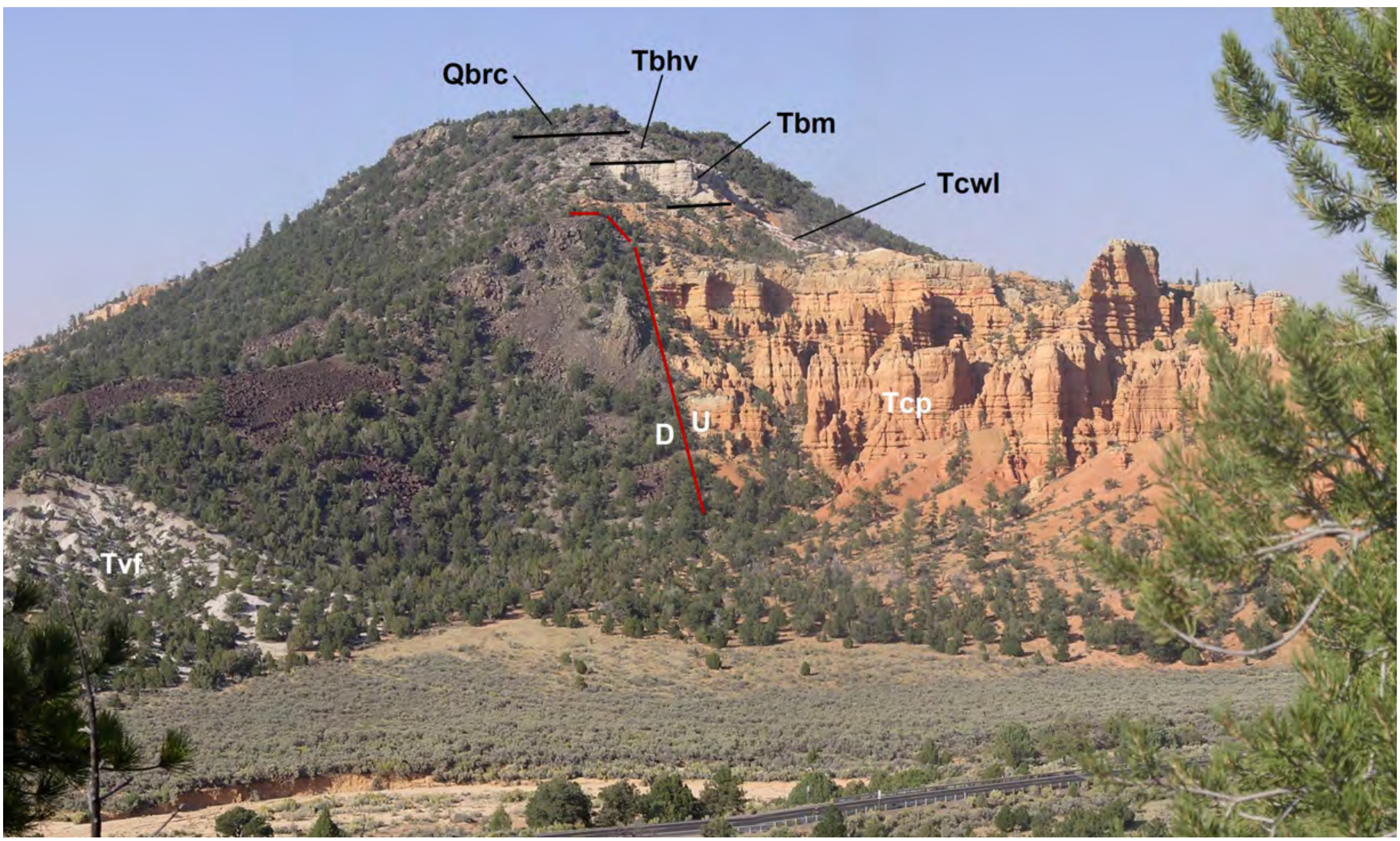


Figure 97. *Utah Highway 12 at the entrance to Red Canyon. The 500,000-year-old Red Canyon lava flow (Qbrc), on the north side of the canyon, erupted from vents on the hanging wall and flowed eastward across the Sevier fault (red line), the trace of which is largely concealed by talus and landslides; it has since been displaced about 650 feet (200 m), yielding a vertical slip rate of 0.4 mm/yr (Lund and others, 2008). At the time the lava flow erupted, the fault juxtaposed tilted, non-resistant late Tertiary basin-fill deposits (Tvf) in the hanging wall against non-resistant, basal Brian Head strata (Tbhv) in the footwall; thus, the fault did not form a significant topographic barrier as it does today. Tbm, conglomerate at Boat Mesa; Tcp, pink member of the Claron Formation; Tcwl, lower unit of the white member of the Claron Formation.*

slope of the Markagunt Plateau, flowed eastward and crossed the fault (which at the time juxtaposed non-resistant fan alluvium against coarse-grained volcaniclastic deposits). Lund and others (2008) estimated displacement on the Rock Canyon flow to be 775 to 1130 feet (237–344 m), which yields a much lower vertical slip rate of about 0.05 mm/yr.

However, the main branch of the Sevier fault here accounts for only part of the total displacement on the fault zone at this latitude (figure 98). A concealed, down-to-the-west fault must be present west of coarse-grained volcaniclastic strata exposed at the base of the Claron cliffs; displacement on that fault is unknown, but total displacement on the fault zone here is about 2700 feet (900 m) (Lundin, 1987; Davis, 1999). Thus, the 0.05 mm/yr displacement rate must be a minimum rate. The actual rate can be no more than 0.18 mm/yr assuming all displacement postdates the 5 Ma Rock Canyon flow, an unlikely scenario. We remain uncertain why the fault exhibits lower long-term slip rates over the past 5 million years compared to post-0.5 Ma slip rates. The inception of faulting on the Sevier fault zone is poorly known but widely cited as middle Miocene, 12 to 15 million years ago (see, for example, Davis, 1999); however, most displacement on the fault zone at the north end of Circleville Valley occurred 8 to 5 million years ago (Rowley, 1968; Rowley and others, 1981).

Lund and others (2008) discussed the uncertainty of determining a reliable vertical slip rate for the Black Mountain part of the fault, located about 10 miles (16 km) south of the map area, due to the difficulty of distinguishing fault scarps from landsliding or from possible cascading of the Black Mountain lava flow over pre-existing topography. The total elevation difference between the 0.57 Ma lava flow at the top of Black Mountain and that at its base is about 750 feet (230 m). Interestingly, if this elevation difference is due entirely to displacement along the Sevier fault zone, it yields

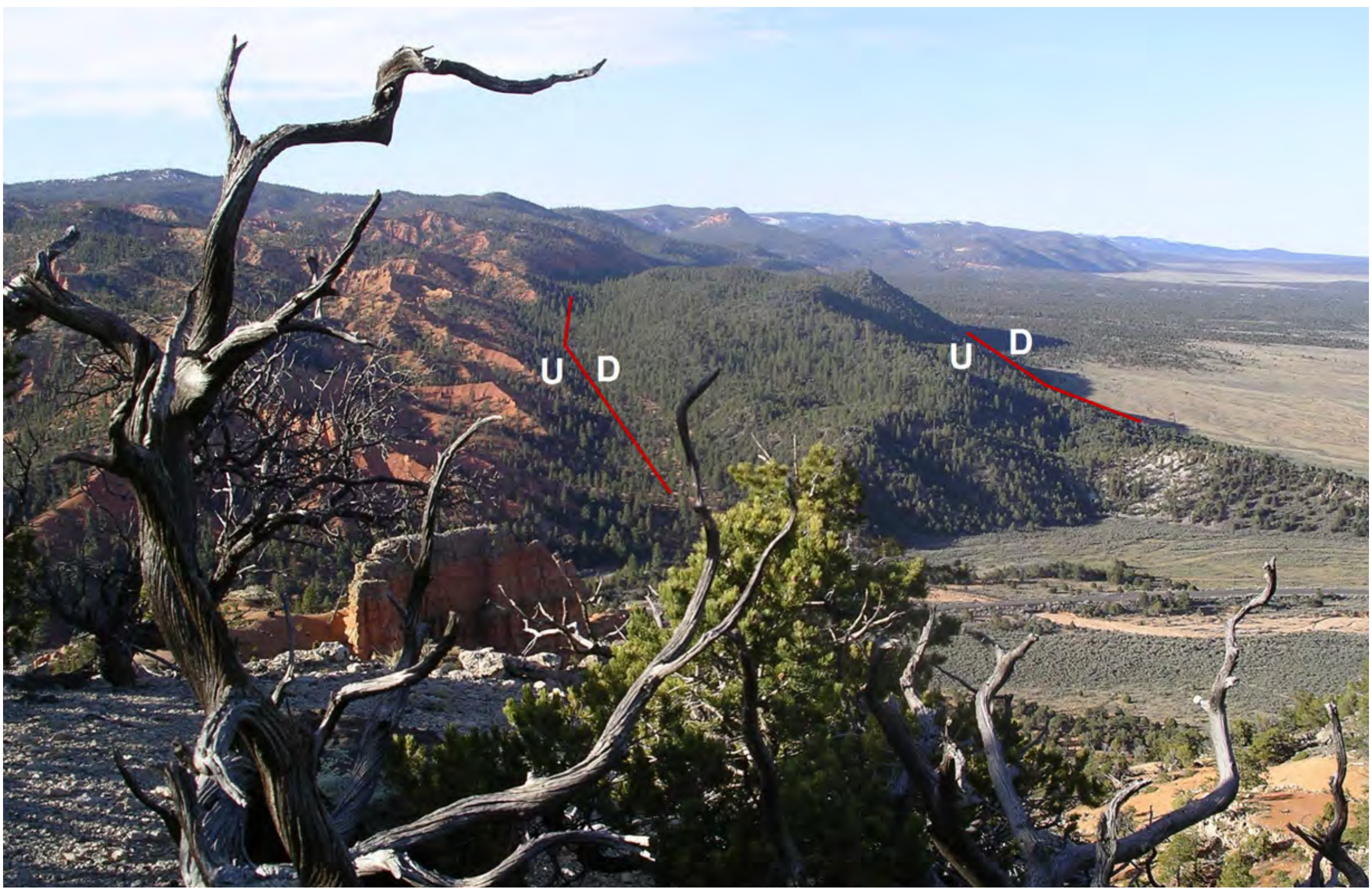


Figure 98. *View south along the Sevier fault zone just south of Red Canyon. Pinyon-juniper-covered hill is capped by the 5 Ma Rock Canyon lava flow, a remnant of which is preserved on the hanging wall (just out of view to left in this photo), and which erupted from a vent several miles to the west on the east flank of the Markagunt Plateau (and thus underlies Sevier Valley). The Rock Canyon flow is thus displaced by two main strands of the Sevier fault at this location. Wilson Peak is at upper left.*

a vertical slip rate of 0.4 mm/yr, an order of magnitude greater than their preferred vertical slip rate of 0.04 mm/yr, but comparable to the slip rate determined for displacement of the Red Canyon lava flow. However, given a vertical slip rate of 0.4 mm/yr, the absence of scarps on unconsolidated deposits along the main trace of the Sevier fault zone is puzzling; scarps a few feet to tens of feet high should be expected (Lund and others, 2008), although it is possible that the lack of scarps may reflect decreased late Quaternary activity on this part of the fault. Lund and others' (2008) preferred slip rate of 0.04 mm/yr results from difficulties assessing displacement due to landsliding at Black Mountain and assumptions about the middle Miocene timing of the inception of faulting.

Paunsaugunt Fault Zone

The Paunsaugunt fault is the easternmost large, down-to-the-west normal fault in southwest Utah, forming the eastern structural margin of the Paunsaugunt and Sevier Plateaus and thus the boundary between the Transition Zone and the Colorado Plateau. Unlike the Hurricane and Sevier fault zones, the Paunsaugunt fault is little studied and its northern and southern ends are not defined. It is, however, the largest of an en echelon set of down-to-the-west normal faults that stretch from the Grand Canyon area northward to the northern part of the Marysvale volcanic field, a distance of about 170 miles (270 km) (see figure 92). Davis (1999) suggested that displacement on the Paunsaugunt fault may die out as a listric normal fault in evaporite strata of the Middle Jurassic Carmel Formation about 13 miles (21 km) south of the map area. This idea was likely based in part on the difficulty in tracing the fault through massive Navajo Sandstone—Sable and Hereford (2004), for example, did not map the Paunsaugunt fault through Navajo, whereas Doelling (2008) did map the fault, linking up well-expressed parts of the fault north and south of the Navajo outcrop belt. Grass Valley and Johns Valley occupy the hanging wall along the northern parts of the fault, whereas the southern part, in the Colorado River drainage, lacks sediment-filled basins on the down-dropped block.

Within the Panguitch 30′ x 60′ quadrangle, the Paunsaugunt fault consists of several splays. Near the south edge of the map area, two splays of the fault place lower Straight Cliffs Formation against Entrada Sandstone and Entrada against Entrada, and just northwest of Tropic, the fault places the pink member of the Claron Formation against the lower John Henry Member of the Straight Cliffs Formation; both areas reveal about 1500 feet (460 m) of tectonic displacement, somewhat less than the 2000 feet (600 m) estimated by Davis (1999). Farther north, collective tectonic displacement among the many strands of the Paunsaugunt fault may decrease, with deformation taken up in part by folding associated with the Paunsaugunt monocline (Davis, 1999).

Bowers (1990) mapped one small fault scarp in surficial deposits near Bulldog Hollow, due west of Tropic, and we mapped a possible fault scarp on late Pleistocene fan alluvium just north of the head of Little Henderson Canyon, about 5 miles (8 km) north-northwest of Tropic (Bowers also described a second small scarp at Squaw Bench, south of the map area). The Bulldog Hollow scarp cuts pediment-mantle deposits that now lie about 240 feet (70 m) above nearby Bryce Creek; adjacent slopes are forested and covered by colluvium, rendering interpretation of the scarp ambiguous, but a scarp profile suggested the surface was displaced about 15 feet (4.5 m) (RJH Consultants, 2012). The scarp near Henderson Canyon is smaller and poorly expressed in partly eroded fan alluvium. A preliminary paleoseismic investigation of the Paunsaugunt fault in Grass Valley, more than 12 miles (19 km) north of the map area, was undertaken in 2010 for a proposed pumped-storage hydroelectric generating facility. That study found antithetic scarps in fan alluvium of likely late Pleistocene age (RJH Consultants, 2012).

At Bryce Canyon National Park, the Paunsaugunt fault zone juxtaposes relatively resistant Claron Formation in the hanging wall against relatively non-resistant Upper Cretaceous strata in the footwall (figure 99). The Paria River and its tributaries have preferentially eroded the softer Cretaceous beds, thus leaving Claron strata topographically high, forming the resistant cap of the Paunsaugunt Plateau. (The Paria River flows into the Colorado River below Lake Powell, which drains to the Gulf of Mexico, whereas the Sevier River flows into Sevier Lake at an elevation of 4500 [1370 m]), thus there is a huge difference in potential energy between the two drainages.)

The effect is puzzling at first as one gazes east from the rim at Bryce Canyon—one naturally assumes that the fault dropped non-resistant strata down to the east, thus forming the lowlands of the Tropic amphitheater, yet all that is needed is to look farther east to the Table Cliff Plateau to see Claron strata soaring at higher elevations (figure 100)—but Claron is indeed dropped down to the west about 1500 feet (460 m) at this latutude. The unusual topographic inversion forms what geologists call an obsequent fault-line scarp, and this must be one of the world's premier examples of such a structure. The seldom-used word "obsequent" is fitting for it is applied to landforms whose orientation is opposite to what may have been expected; an obsequent fault-line scarp faces the opposite way of the original fault scarp.

Stratigraphic relations of the sub-Claron unconformity demonstrate that the Paunsaugunt fault exhibits an early history of west-side-up reverse movement. East of the map area at the Table Cliff Plateau, Claron strata overlie the Pine Hollow Formation, which overlies the Canaan Peak Formation, and which in turn unconformably overlies the Kaiparowits Formation (Doelling and Willis, 1999b). Nei-

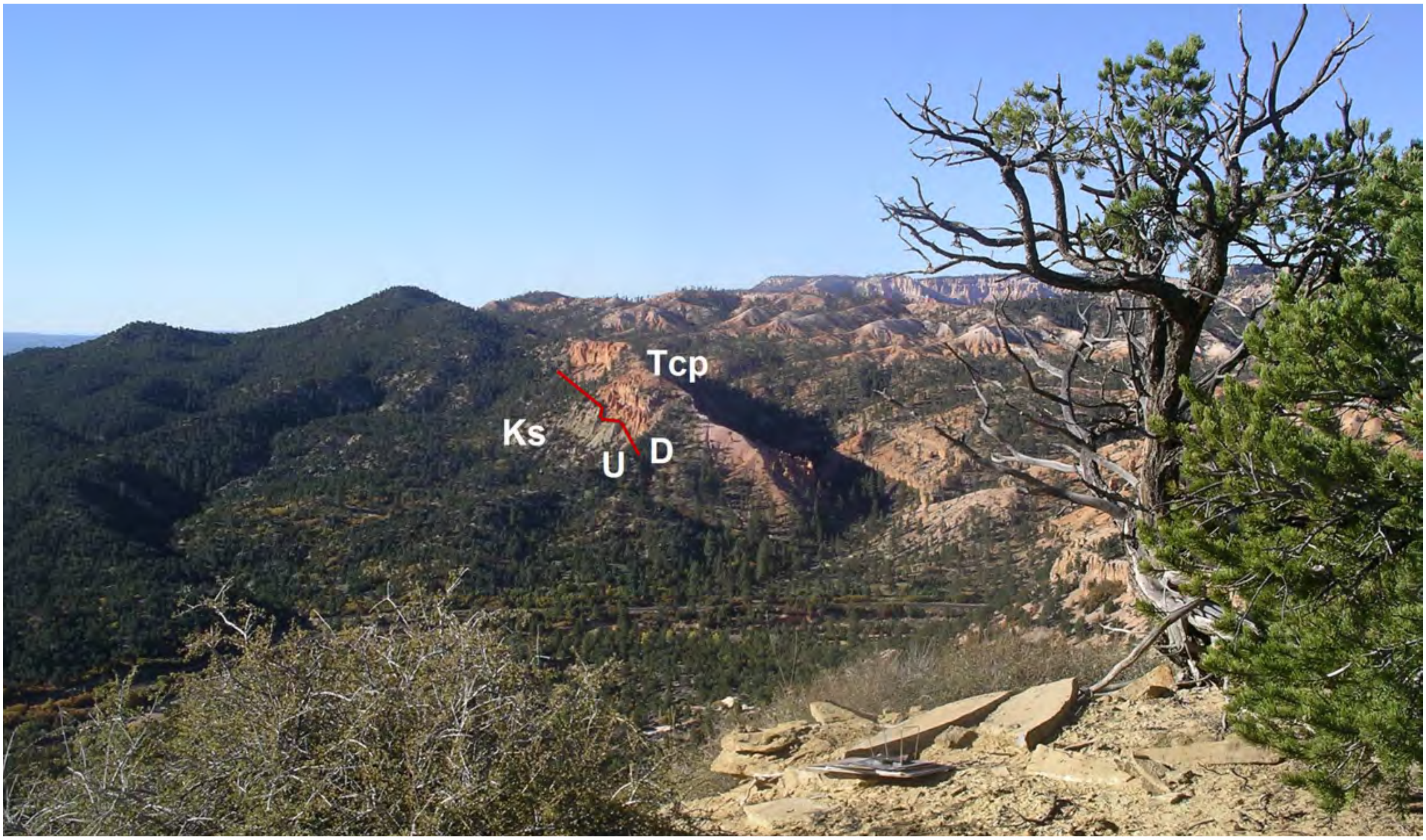


Figure 99. *View south across Utah Highway 12 in Tropic Canyon, showing the Paunsaugunt fault where it places the pink member of the Claron Formation (Tcp) down on the west against the mostly forested Straight Cliffs Formation (Ks).*

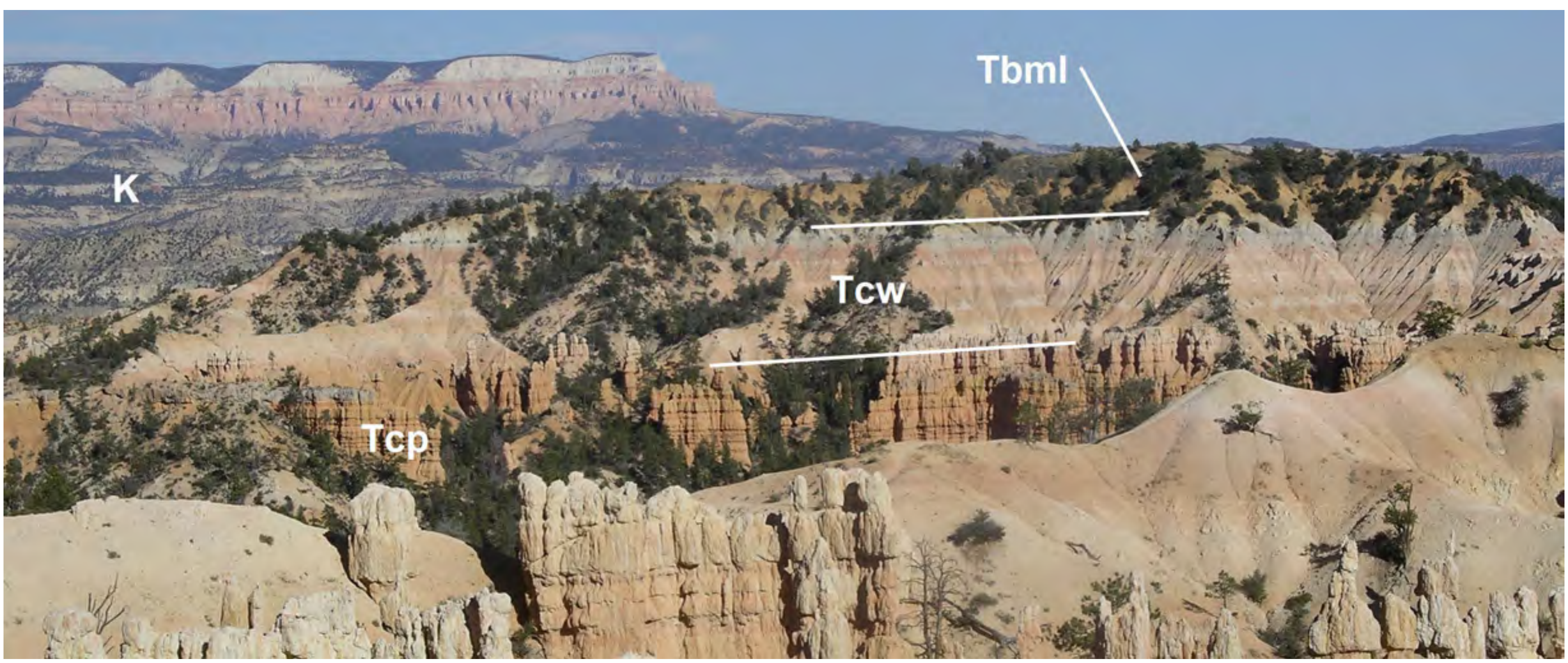


Figure 100. *Powell Point, southernmost point of the Table Cliff Plateau, is capped by the resistant white member of the Claron Formation. The unnamed hill in the middle is capped by the conglomerate of Boat Mesa (Tbml), which overlies an unusual facies of the Claron's white member (Tcw). The Paunsaugunt fault lies beyond this hill and reveals about 1500 feet (460 m) of down-to-the-west displacement. Kaiparowits, Wahweap, and Straight Cliffs Formations (K) form the broad basin of the Tropic Amphitheater. Tcp, pink member of the Claron Formation.*

Table 5. *Optically stimulated luminescence and uranium-lead detrital zircon ages, Panguitch 30' x 60' quadrangle.*

Formation	Map Symbol	Sample number	Map #	U-Pb age (Ma)	OSL age (ka)	7.5' Quadrangle	UTM easting	UTM northing	Lab used	Reference	Comments
Pre-Pinedale glacial till	Qgtou	PL090109-1	44	N/A	48.95 ± 19.24	Panguitch Lake	348527	4169280	USULL	this report	
pink member of Claron Formation	Tcp	WP110211-1	47	age unavilable	N/A	Wilson Peak	388836	4173199	AtoZ	this report	siltstone from base of basal Tcp conglomer-ate; youngest grain is 243 Ma
conglomerate at Boat Mesa	Tbm	CC110311-1		37.97 +1.78 -2.70	N/A	Casto Canyon	387944	4183097	AtoZ	this report	
middle sand-stone unit of white mbr. Claron Fm.	Tcwm	CC110311-2		39.6 ± 2.93	N/A	Casto Canyon	387834	4183113	AtoZ	this report	1 grain only; about 80 feet below base of Tbm
Kaiparowits Formation, lower unit	Kkl	WP050111-1		75.62 +3.08 -1.66	N/A	Wilson Peak	382016	4168801	AtoZ	this report	

Notes:
Map Symbol is the symbol used in this report
AtoZ, Apatite to Zircon, Inc., Viola, Idaho
USU LL, Utah State University Luminescence Laboratory

ther Pine Hollow nor Canaan Peak is present in this map area. Instead, the Claron rests with slight angular discordance on Kaiparowits and older strata, showing that the sub-Claron unconformity cuts down sharply to the west towards the Paunsaugunt fault zone. The unconformity cuts progressively down-section to the west and south so that Claron rests on Wahweap along most of the east rim of the Paunsaugunt Plateau at Bryce Canyon National Park. In the vicinity of Paria View, the unconformity cuts farther down into underlying Drip Tank strata, reflecting that area's position on the crest of the pre-Claron Bryce Canyon anticline. A subtle pre-Claron syncline is present along the axis of the plateau, preserving a complete section of the Wahweap Formation and locally the basal Kaiparowits Formation. Basal Kaiparowits strata are also preserved on the west flank of the plateau and are well exposed at Hillsdale Canyon. The overall geometry of the pre-Claron unconformity thus reveals a broad anticlinal welt along the east flank of the plateau over which Kaiparowits and locally the entire Wahweap Formations were removed by erosion prior to Claron deposition. We envision that this anticline formed above a blind, east-directed, pre-Claron thrust fault, which we interpret to be the easternmost expression of thin-skinned, Sevier-age deformation in this part of the fold-thrust belt. The thrust fault was later reactivated as the Paunsaugunt normal fault during basin-range extension.

FUTURE GEOLOGIC RESEARCH

Geologic maps graphically communicate vast amounts of geologic information—they display three-dimensional features on a flat piece of paper, with the added benefit of depicting the relative age, composition, and relations among rocks and sediments at and near the earth's surface. A detailed geologic map shows what you are standing on; where similar rocks or sediments may be found; how old they are; what they are composed of; how they formed; how they have been affected by faulting, folding, or other geologic processes; and what existing or potential mineral resources and geologic hazards are nearby (Biek, 1999). It should come as no surprise then that geologic maps are the foundation for virtually all other geologic studies.

A little-appreciated fact about geologic maps is that they also reveal both practical and purely scientific questions that remain unanswered. The very act of making a geologic map inevitably raises more geologic questions than the map alone can answer or that were beyond the scope of the mapping effort. Astute map readers will eventually recognize some of these outstanding geologic problems, but we offer this highly subjective top-ten list of some of the most intriguing geologic issues that remain to be adequately addressed in the Panguitch 30′ x 60′ quadrangle.

1. Map high-priority 7.5′ quadrangles at 1:24,000-scale detail. Chief among these are the Parowan, Parowan Gap, and Summit quadrangles. LiDAR imagery is now available for the greater Parowan Valley area, which will enable much more detailed mapping of surficial deposits, fault scarps, and earth fissures.

2. Investigate the age, extent, and stability of the many significant landslides in the map area, including those

of Yankee Meadows graben, near Panguitch Lake, and along the Utah Highway 14 corridor.

3. Determine the age of basal Claron strata across its basin. It is conceivable, though unlikely, that basal Claron strata are at least locally of Late Cretaceous age.

4. Refine the correlation and age of "Cretaceous strata on the Markagunt Plateau" (map unit Km).

5. Determine the age of young basaltic lava flows of the central Markagunt Plateau and reconstruct the drainage history of the plateau. The greater Panguitch Lake area in particular appears to be a fruitful area for research.

6. Continue more detailed mapping of the Markagunt gravity slide, especially its ramp area, and work to elucidate the mechanics of its emplacement and trigger.

7. Investigate the extent and tectonic setting of the Limerock Canyon Formation and develop criteria to distinguish it from the Brian Head Formation.

8. Map the Iron Peak area in detail and determine if the laccolith erupted lava flows.

9. Create a more comprehensive map of sinkholes on the Markagunt Plateau and further evaluate groundwater movement through this karst terrain. Ultimately, such a map could be used to help protect water resources of the area.

10. Reevaluate the age, extent, and relationship between the Bear Valley and Mount Dutton Formations and determine why Bear Valley strata are apparently not present on the Sevier Plateau.

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APPENDIX

MAP UNIT DESCRIPTIONS

APPENDIX

MAP UNIT DESCRIPTIONS

QUATERNARY

Human-derived deposits

Qh **Artificial fill** (Historical) − Engineered fill used to construct the dam at Navajo Lake; fill of variable thickness and composition should be anticipated in all developed or disturbed areas; typically less than 20 feet (6 m) thick.

Qhd **Disturbed land** (Historical) – Disturbed area in Castle Valley (about 5 miles [9 km] southwest of Panguitch Lake) mapped because it obscures extent of glacial deposits and landforms; also includes large sand and gravel pits southeast of Panguitch and northwest of Bryce Canyon.

Alluvial deposits

Qal **Stream alluvium** (Holocene) − Moderately sorted sand, silt, clay, and pebble to boulder gravel deposited in active, main-stem stream channels and floodplains of the Sevier River and its few well-graded major tributaries; locally includes minor stream-terrace alluvium as much as about 10 feet (3 m) above current stream level; inferred to overlie older alluvial and fan deposits in the valley of the Sevier River; probably less than 30 feet (9 m) thick.

Qat **Stream-terrace alluvium** (Holocene to middle? Pleistocene) − Moderately sorted sand, silt, and pebble to boulder gravel that forms incised gently sloping terraces above the Sevier River and its few well-graded major tributaries; deposited in a stream-channel environment, but locally includes colluvium and small alluvial fans; terraces are at elevations of about 10 to 120 feet (3–35 m) above adjacent streams but are not subdivided here due to limitations of map scale; typically less than 20 feet (6 m) thick.

Qaly **Young stream alluvium** (Holocene) − Combined stream alluvium (Qal) and the youngest (lowest elevation) part of stream-terrace alluvium (Qat), but undivided here due to limitations of map scale; this lumped unit is commonly mapped in upland drainages where it may include small alluvial-fan deposits from tributary drainages and colluvium from adjacent slopes; commonly grades downslope into alluvial fans; locally includes historical debris-flow and debris-flood deposits derived from tributary drainages, as, for example, the deposits of the 2005 Black Mountain debris flow that entered the upper reaches of Coal Creek in Cedar Canyon (Giraud and Lund, 2010); typically less than 20 feet (<6 m) thick, but deposits in major stream valleys may locally exceed 30 feet (9 m) thick.

Qalo **Old stream alluvium** (Holocene and upper Pleistocene) – Similar to lower- to middle-elevation parts of stream-terrace alluvium (Qat), but these dissected deposits are largely restricted to upland drainages not well graded to the Sevier River; typically less than 20 feet (6 m) thick.

Qao **Oldest stream alluvium** (Pleistocene) – Moderately sorted sand, silt, and pebble to boulder gravel that forms topographically inverted channel deposits at the mouth of Clear Creek, which drains eastward into Panguitch Lake, and on the south side of the lake itself; the deposits of the latter area were well exposed in excavations associated with a new housing development that revealed interbedded sand and pebbly to cobbly, locally iron-stained gravel containing clasts mostly of the Isom Formation (Ti, which is commonly grussy weathering) and subordinate chalcedony and quartzite, but apparently lacking basalt; map unit includes deposits that underlie the nearby Cooper Knoll lava flow (Qbck) just southeast of Panguitch Lake and that consist of subrounded to rounded pebbles to boulders of the Isom Formation, mafic volcanic rocks, chalcedony, and, especially near the base of the deposits, quartzite pebbles and cobbles; the source of the quartzite pebbles and cobbles is unknown, but they are doubtless ultimately recycled from the Grand Castle Formation (redefined) or Drip Tank Member of the Straight Cliffs Formation now exposed in grabens below the western topographic rim of the Markagunt Plateau; Qao deposits record an early phase in the development of the Clear Creek drainage and the related Rock Canyon drainage, which also drains eastward across the central part of the Haycock Mountain quadrangle and the lower part of which is now drained by its markedly underfit northern tributary, Pass Creek; in other words, ancestral Clear Creek apparently used to drain southeast into Rock Canyon, thus giving the canyon a much larger drainage basin before Clear Creek was diverted by the Panguitch Lake lava flows; deposits near Panguitch Lake are 40 to 60 feet (12–18 m) thick and those underlying the Cooper Knoll lava flow are as much as 120 feet (35 m) thick.

Qam **Marsh alluvium** (Holocene and upper Pleistocene) − Dark-yellowish-brown clay, silt, sand, and minor gravel lenses deposited in closed depressions on landslides and glacial moraines in the Lowder Creek area east of Brian Head peak; forms small marshy areas characterized by cattails and other hydrophilic vegetation; typically less than 10 feet (3 m) thick, but marsh alluvium of Lowder Creek bog, described in more detail under the unit description for "glacial till of Pinedale age (Qgtp)," is at least 21 feet (7 m) thick (Mulvey and others, 1984).

Qap **Pediment alluvium** (Holocene and Pleistocene) – Poorly sorted sand and gravel containing subangular to rounded clasts that forms a locally resistant cap on eroded bedrock surfaces or, locally, on old fan alluvium (Taf); represents multiple surfaces from several tens of feet to more than 100 feet (30 m) above modern drainages; deposited principally as debris flows, debris floods, and in ephemeral stream channels; typically less than 20 feet (6 m) thick.

Qaf_1 **Young fan alluvium** (Holocene) − Poorly to moderately sorted, non-stratified, clay- to boulder-size sediment containing subangular to subrounded clasts deposited principally by debris flows and debris floods at the mouths of active drainages; equivalent to the upper part of young and middle fan alluvium (Qafy), but differentiated because Qaf_1 typically forms smaller, isolated fans; probably less than 30 feet (9 m) thick.

Qaf_2 **Middle fan alluvium** (Holocene and upper Pleistocene) − Similar in composition and morphology to young fan alluvium (Qaf_1), but forms inactive surfaces incised by younger stream and fan deposits; equivalent to the older, lower part of young and middle fan alluvium (Qafy) and coalesced fan alluvium of Parowan Valley (Qafc); these deposits preserve previously unreported fault scarps east and north of Summit in southern Parowan Valley that appear to be the southwest continuation of the Parowan Valley faults that Williams and Maldonado (1995) and Black and Hecker (1999) inferred to be latest Pleistocene to early Holocene in age; probably less than 30 feet (9 m) thick.

Qafy **Young and middle fan alluvium, undivided** (Holocene and upper Pleistocene) − Poorly to moderately sorted, non-stratified, boulder- to clay-size sediment containing subangular to subrounded clasts deposited at the mouths of streams and washes; forms both active depositional surfaces (Qaf_1 equivalent) and low-level inactive surfaces incised by small streams (Qaf_2 equivalent) that are undivided here; deposited principally as debris flows and debris floods, but colluvium locally constitutes a significant part; small, isolated deposits are typically less than a few tens of feet thick, but large, coalesced deposits in Sevier and Parowan Valleys are much thicker and form the upper part of basin-fill deposits.

Qafc **Coalesced fan alluvium of Parowan Valley** (Holocene and upper Pleistocene) – Similar to young and middle fan alluvium (Qafy) but forms large, coalesced fans in Parowan Valley; typically exhibits a lower overall slope than young and middle fan alluvium, which we mapped as smaller fans close to the range front; forms unfaulted, active surfaces and was deposited principally as debris flows and debris floods; thin planar beds with small snails exposed in arroyo walls immediately east of Winn Gap (at the south end of the Red Hills) may represent deposits of a shallow lake or playa, possibly blocked at Winn Gap by the Red Hills fault zone, which also blocked the east end of Parowan Gap to form Little Salt Lake (Threet, 1952); thickness uncertain, but Hurlow (2002) showed that Quaternary and Neogene basin fill is in excess of 2000 feet (600 m) thick in southern Parowan Valley west of Parowan and that this basin fill thickens to the northeast; only the uppermost part of this basin-fill is included in map unit Qafc, which we assume to be in excess of several tens of feet thick.

Qafo **Older fan alluvium** (upper and middle Pleistocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, boulder- to clay-size sediment with moderately developed calcic soils (caliche); forms broad, gently sloping, incised surfaces in Parowan, Panguitch, and Hatch Valleys; fault scarps locally prominent on these deposits; deposited principally as debris flows and debris floods; exposed thickness as much as several tens of feet.

$Qafo_2$ **Oldest fan alluvium** (middle and lower? Pleistocene) – Similar in composition to older fan alluvium, but forms deeply dissected surfaces with little or no remaining fan morphology; preserved in the footwall of inferred faults in Hatch valley; maximum exposed thickness is about 150 feet (45 m).

Colluvial deposits

Qc **Colluvium** (Holocene and upper Pleistocene) − Poorly to moderately sorted, angular to subrounded, clay- to boulder-size, locally derived sediment deposited by slope wash and soil creep on moderate slopes and in shallow depressions; locally grades downslope into deposits of mixed alluvial and colluvial origin; mapped only where it conceals contacts or fills broad depressions; the Claron and Brian Head Formations and Upper Cretaceous strata shed enormous amounts of colluvium, such that an apron of heavily vegetated colluvium (unmapped because it forms a veneer having poor geomorphic expression) typically envelops at least the lower part of steep slopes along their outcrop belt; typically less than 20 feet (6 m) thick.

Qco **Older colluvium** (upper? Pleistocene) – Similar to colluvium but deeply incised by modern drainages on the flanks of the Sevier and Paunsaugunt Plateaus; typically less than 40 feet (12 m) thick.

Eolian deposits

Qed **Eolian dune sand** (Holocene) − Grayish-pink to pale-red, well-sorted silt and fine-grained sand largely stabilized by vegetation on the Cedar Breaks escarpment; most of the sand consists of tiny clay pellets eroded from the Claron Formation and carried eastward up the scarp by strong winds; typically less than 15 feet (5 m) thick.

Qes **Eolian sand** (Holocene) − Yellowish-brown to reddish-brown, moderately well sorted sand and silt derived from deflation of Little Salt Lake playa deposits (Qlp), which are located to the south and west; forms thin sheets and poorly developed dunes partly covered by sparse vegetation; generally more saline than underlying alluvium and so allows *Sarcobatus* sp. (greasewood) to flourish at the expense of *Artemisia* sp. (sagebrush) (Maldonado and Williams, 1993b); typically less than 6 feet (2 m) thick.

Glacial deposits

Glacial till and outwash are present only east of Brian Head peak in the Castle Creek and Lowder Creek drainages and in the greater Castle Valley area. These deposits are of the Pinedale alpine glacial advance and an older glaciation of uncertain Quaternary age (possibly Bull Lake). Pinedale deposits in their type area in the Wind River Range of Wyoming are about 12 to 24 ka (Imbrie and others, 1984) (with glacial maxima about 16 to 23 ka on the basis of cosmogenic ^{26}Al and ^{10}Be dating; Gosse and others, 1995), and they are roughly coeval with the late Wisconsin glaciation, Last Glacial Maximum (LGM), and Marine Oxygen Isotope Stage 2 (MIS 2). In contrast, deposits of the Bull Lake alpine glacial advance in their type area in the Wind River Range are about 128 to 186 ka (Imbrie and others, 1984) (with glacial maxima about 140 to 160 ka; Gosse and Phillips, 2001; Sharp and others, 2003), and they are roughly coeval with the Illinoian glaciation or MIS 6.

Qgtp **Glacial till of Pinedale age** (upper Pleistocene) – Non-stratified, poorly sorted, sandy pebble to boulder gravel in a matrix of sand, silt, and minor clay; clasts are matrix supported, subangular to subrounded, and derived from the Leach Canyon, Isom, and Brian Head Formations and the Markagunt gravity slide exposed in the headwaters of the Castle Creek and Lowder Creek drainage basins; terminal moraine at the west end of Castle Valley is at an elevation of about 9750 feet (2973 m), whereas the terminal moraine of the smaller Lowder Creek basin is at Long Flat at an elevation of about 10,100 feet (3080 m); recessional and lateral moraines and hummocky, stagnant-ice topography are locally well developed, but sculpted bedrock is absent or inconspicuous, probably owing to the relatively small size and suspected short duration of the glaciers (Mulvey and others, 1984); well-developed terminal and recessional moraines are as much as 120 feet (37 m) thick, but till is much thinner elsewhere and locally consists only of scattered boulders or a veneer of meltout till on bedrock.

The Brian Head–Sidney Peaks area marks the southernmost occurrence of late Pleistocene glaciation in Utah (Mulvey and others, 1984), as first briefly described by Gregory (1950). Agenbroad and others (1996) interpreted glacial deposits and features that they attributed to their “Mammoth Summit glacier” at the southwest side of Brian Head peak and north edge of Cedar Breaks National Monument, but which we interpret as landslide deposits and in-place Brian Head Formation, the latter partly covered by a lag of large blocks of the Isom Formation.

Till is Pinedale age on the basis of distinct, well-preserved morainal morphology and relatively unweathered clasts, and a minimum limiting age of 14,400 ± 850 ^{14}C yr B.P. from marsh deposits of the Lowder Creek bog that overlie the till (Mulvey and others, 1984; Currey and others, 1986; see also Anderson and others, 1999). Madsen and others (2002) identified the 14,300 ^{14}C yr B.P. Wilson Creek #3 ash (erupted from Mono Craters in California) in the Lowder Creek bog. Marchetti and others (2005, 2007, 2011) and Weaver and others (2006) reported boulder exposure ages from four different moraines that indicate a local last glacial maximum of about 21.1 ka for the main Pinedale advance on Boulder Mountain approximately 80 miles (130 km) to the northeast. Their ages coincide with the global LGM (21 ± 2 ka) and thus likely are the age of the main Pinedale moraines on the Markagunt Plateau. Marchetti and others (2005, 2011) also reported a smaller advance at about 16 ka on Boulder Mountain.

Qgop **Glacial outwash of Pinedale age** (upper Pleistocene) – Moderately to well-sorted, generally subrounded, clast-supported, pebble to boulder sand and gravel; clasts are typically little weathered and of the same provenance as glacial till (Qgtp); mapped on the east side of Castle Valley where the deposits likely represent the waning stages of Pinedale glaciation; probably about 20 to 30 feet (6–9 m) thick.

Qgtu **Older glacial till of uncertain pre-Pinedale age** (upper? and middle? Pleistocene) – Similar to glacial till of Pinedale age, but glacial landforms are poorly preserved or absent; forms a low-relief, rubble-covered, locally hummocky surface both northeast and southwest of the southern Long Flat cinder cone (peak 10,392, north of lower Lowder Creek; map unit Qblfc); the northeast flank of this cinder cone is conspicuously truncated, perhaps by this glacial advance; also forms low hills south of Castle Valley, in the southwest part of the Panguitch Lake 7.5′ quadrangle, that are composed almost entirely of large blocks of Leach Canyon Formation, with minor blocks of Isom Formation and chalcedony, that we infer to be the deeply eroded remains of a medial or recessional moraine; thickness uncertain, but probably about 10 to 30 feet (3–10 m) thick.

Mulvey and others (1984) and Currey and others (1986) first suggested that glacial till older than Pinedale age may be present in the Brian Head quadrangle, west of Castle Valley. We sampled a sandy till exposed in a bluff northwest of the confluence of Mammoth and Castle Creeks (map unit Qgtou) that yielded an optically stimulated luminescence age of 48.95 ± 19.24 ka (table 5; see also UGS and USULL, 2013), but we remain uncertain how to interpret this age. The deposits may correspond to the Early Wisconsin MIS 3-4 advance (about 59 to 71 ka; Imbrie and others, 1984), and although Laabs and Carson (2005) reported that MIS 3-4 glacial moraines are not known in Utah, such deposits may be more widespread in the west than originally thought (Tammy Rittenour, Utah State University, written communication, August 3, 2010). However, given the widespread extent and degree of incision of Qgtou deposits, we interpret these glacial deposits to be older, more likely of Bull Lake age.

Qgtou **Older glacial till and outwash, undivided** (upper and middle? Pleistocene) – Similar to older glacial till of uncertain pre-Pinedale age, but forms broad, open, boulder-strewn and sage-brush-covered, eastward-sloping surfaces of the Castle Creek and Mammoth Creek areas; exposures just north of the junction of Castle Creek and Mammoth Creek suggest that most of this surface is underlain by till now deeply incised at its eastern end; glacial outwash deposits, especially those graded to the Pinedale terminal moraines, are presumed to be present locally on this till plain, but are not readily differentiated at this map scale; Mulvey and others (1984) and Currey and others (1986) briefly reported on possible ice wedge polygons as evidence for periglacial features on the southwest side of Castle Valley; glacial till is as much as 60 feet (18 m) thick where exposed near the confluence of Castle and Mammoth Creeks.

Lacustrine and playa deposits

Qlp **Little Salt Lake playa deposits** (Holocene) – Calcareous, saline, and gypsiferous gray clay, silt, and fine-grained sand deposited on the flat playa floor of Little Salt Lake in the southwest part of Parowan Valley; locally includes small dunes of eolian silt; at least 25 feet (8 m) thick.

The Little Salt Lake playa formed in response to relative uplift of the Red Hills structural block along the Red Hills fault (Threet, 1952; Maldonado and Williams, 1993a, 1993b). The playa reflects ponded drainage and represents the latest stage in the history of antecedent drainage through Parowan Gap (Maldonado and Williams, 1993b). We infer that a playa has occupied this area intermittently throughout the Pleistocene, but deposits at and near the surface are doubtless Holocene in age.

Qlm **Little Salt Lake playa-margin deposits** (Holocene and upper Pleistocene) – Calcareous, saline, and gypsiferous gray clay, silt, sand, and local volcanic and quartzite pebbles, deposited on gentle slopes around the margin of Little Salt Lake playa; periodically flooded during high lake levels; includes small alluvial fans, eolian sand and silt, and alluvium; less than 12 feet (4 m) thick.

Qlg **Coarse-grained lacustrine sediment** (Holocene and upper Pleistocene) – Sand and gravel deposited at the east end of Navajo Lake, which formed behind a lava dam created by the Henrie Knolls lava flows; probably 10 to 15 feet (3–5 m) thick.

Mass-movement deposits

Qms, Qmsh, Qms(Kd), Qms(Ti), Qms(Tql), Qms(Tm)
Landslides (Historical to upper? Pleistocene) – Very poorly sorted, locally derived material deposited by rotational and translational movement; composed of clay- to boulder-size debris as well as large, partly intact, bedrock blocks; characterized by hummocky topography, numerous internal scarps, chaotic bedding attitudes, and common small ponds, marshy depressions, and meadows; the largest landslide complexes are several square miles in size and involve tuffaceous strata of the Brian Head (Tbh) and Dakota (Kd and Ktd) Formations, and to a lesser extent the combined Limerock Canyon and Brian Head Formations (Tlbh); Qmsh denotes landslides known to be active in historical time, but

any landslide deposit may have been historically active even if not so identified; large rotational slump blocks of Isom Formation (Qms[Ti]) and Leach Canyon Formation (Qms[Tql]) are mapped in the Yankee Meadows graben and in the lower part of the Lowder Creek basin, and slump blocks of Dakota Formation (Qms[Ktd]) are mapped in Cedar Canyon; Qms(Tm) represents Quaternary landslide debris derived from the Markagunt gravity slide present northwest of Cedar Breaks National Monument; query indicates areas of unusual morphology that may be due to landsliding; thickness highly variable, but typically several tens of feet or more thick and the largest landslides, at Yankee Meadows graben, may be as much as 600 feet (200 m) thick (Maldonado and others, 1997); undivided as to inferred age because 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).

The most active and problematic landslide in the map area is in Cedar Canyon where State Highway 14 crosses the upper part of the Dakota Formation. Lund and others (2009) reported on a large rockfall associated with this landslide that closed the highway for six days in January 2009. The landslide moved again on October 8, 2011, destroying the road and closing the highway for nearly a year (Lund and others, 2012).

Dense forests and widespread colluvium may conceal unmapped landslides, and more detailed imaging techniques such as LiDAR may show that many slopes, particularly those developed on the Brian Head (Tbhv) and Bear Valley (Tbv) Formations and on Upper Cretaceous strata, host surficial deposits that reveal evidence of creep or shallow landsliding. Understanding the location, age, and stability of landslides, and of slopes that may host as-yet unrecognized landslides, requires detailed geotechnical investigations.

Qmt **Talus** (Holocene and upper Pleistocene) – Poorly sorted, angular cobbles and boulders and finer-grained interstitial sediment deposited principally by rockfall on or at the base of steep slopes; talus that is part of large landslide complexes is not mapped separately; talus is common at the base of steep slopes across the map area, but is only mapped where it conceals contacts or forms broad aprons below cliffs of resistant bedrock units; commonly grades downslope into colluvium; typically less than 30 feet (9 m) thick.

Mixed-environment deposits

Qac **Alluvium and colluvium** (Holocene and upper Pleistocene) – Poorly to moderately sorted, generally poorly stratified, clay- to boulder-size, locally derived sediment deposited in swales and small drainages by fluvial, slope-wash, and creep processes; generally less than 20 feet (6 m) thick.

Qaco **Older alluvium and colluvium** (upper? Pleistocene) – Similar to mixed alluvium and colluvium (Qac), but forms incised, isolated remnants, typically along the upper reaches of streams; probably about 20 to 30 feet (6–9 m) thick.

Qacf **Alluvium, colluvium and fan alluvium** (Holocene and upper? Pleistocene) – Poorly to moderately sorted, non-stratified, clay- to boulder-size sediment deposited principally by debris flows, debris floods, and slope wash at the mouths of active drainages and the base of steep slopes; locally reworked by small, ephemeral streams; forms coalesced apron of fan alluvium and colluvium impractical to map separately at this scale; typically 10 to 40 feet (3–12 m) thick.

This map unit is also mapped in sediment-choked drainages that drain the Paunsaugunt and Sevier Plateaus, where it consists of modern alluvium, low-level terrace alluvium, colluvium, and fan alluvium impractical to differentiate at the map scale. The upper reaches of these deposits are typically characterized by debris-flow deposits partly reworked by braided channels, probably a result of flash-flood runoff on the Claron Formation. The morphology of many of these drainages suggests that they are partly backfilled following an earlier, deeper level of canyon erosion, and in the larger drainages these deposits are likely several tens of feet thick.

Qacfo **Older colluvium and fan alluvium** (Pleistocene) – Mapped in the western Markagunt Plateau, where it consists of poorly sorted, boulder- to clay-size sediment derived mostly from the Claron, Brian Head, and Isom Formations, and locally below the rim of the Paunsaugunt Plateau where it consists principally of reworked Claron debris; deposited principally by debris flows, debris floods, and slope wash; typically forms a resistant cap on isolated hill tops and ridges underlain by Upper Cretaceous strata, remnants of a once larger apron of sediment shed off the plateaus and now preserved as deeply dissected inverted valleys; also forms a broad bench, preserved in the Iron Peak graben west of the town of Brian Head, where it is locally exposed in the main scarp of a large landslide complex southeast of Sugarloaf Mountain (SE1/4SW1/4 section 8, T. 36 S., R. 9 W.); also forms incised, isolated remnants south of Haycock Mountain, in the upper reaches of the Clear Creek drainage, and a single deposit southeast of Brian Head peak; typically about 20 to 30 feet (6–9 m) thick, but larger deposits may locally exceed 50 feet (15 m) thick.

Qae **Alluvium and eolian sand** (Holocene and upper Pleistocene) − Moderately to well-sorted, mostly light-reddish-brown silt and sand deposited by sheetwash and ephemeral streams in small drainages and swales on the Henrie Knolls lava flow in the west-central part of the Henrie Knolls quadrangle; probably less than 10 feet (3 m) thick.

Qea **Eolian sand and alluvium** (Holocene and upper Pleistocene) − Moderately to well-sorted, yellowish-brown sand deposited by wind and locally reworked by ephemeral streams; includes sand, silt, clay, and pebble to boulder gravel of stream channels; mapped in the southern Red Hills; probably less than 20 feet (6 m) thick.

Qaec **Alluvium, eolian sand, and colluvium** (Holocene and upper Pleistocene) − Moderately sorted, light-reddish-brown and moderate- to dark-yellowish-brown silt and sand and locally gravelly lenses deposited in swales and small drainages on and adjacent to the Henrie Knolls lava flow (Qbhk); the margins of the deposits include significant colluvium derived from adjacent hillslopes developed on the Claron Formation and basaltic lava flows; soils developed on this unit have an argillic horizon 1 to 1.5 feet (0.3–0.5 m) thick of moderate-reddish-brown sandy clay and clayey fine-grained sand; typically less than 10 feet (3 m) thick, although deposits in the Cow Lake area, south of the Henrie Knolls flows, are likely as much as 20 feet (6 m) thick.

Qca **Colluvium and alluvium** (Holocene to middle Pleistocene) − Poorly to moderately sorted, clay- to pebble-size, locally derived sediment deposited principally by slope wash and locally reworked by alluvial processes; typically mapped where lava flows dammed local washes, causing ponding of mixed colluvial and alluvial sediment; distal, finer-grained parts form broad, open meadows; thickness uncertain, but likely less than about 20 feet (6 m) thick.

Qce **Colluvium and eolian sand** (Holocene and upper Pleistocene) − Poorly to moderately sorted, clay- to boulder-size, locally derived sediment—partly covered by a veneer of eolian sand—deposited by slope wash on moderate slopes and in shallow depressions in the Red Hills graben south of Parowan Gap and on the Red Canyon lava flow at the east margin of the valley of the Sevier River; colluvial debris is derived from adjacent lava flows and, in the Red Hills area, from the Navajo Sandstone; deposits in the Red Hills are probably less than 20 feet (6 m) thick; those near Red Canyon are probably less than about 6 feet (2 m) thick.

Qmtc **Talus and colluvium** (Holocene and upper Pleistocene) – Poorly sorted, angular to subangular, cobble- to boulder-size and finer-grained interstitial sediment deposited principally by rockfall and slope wash on steep slopes throughout the quadrangle; includes minor alluvial sediment at the bottom of washes; generally less than 30 feet (9 m) thick.

Qmtco **Older talus and colluvium** (upper? Pleistocene) – Similar to talus and colluvium but deeply incised by modern drainages where it is mapped at the south end of the Sevier Plateau; generally less than 30 feet (9 m) thick.

Qmsc **Landslides and colluvium** (Holocene and upper Pleistocene) – Landslides and colluvium impractical to differentiate at this scale; as much as several tens of feet thick.

Qmsco **Older landslides and colluvium** (upper? Pleistocene) – Older landslides and colluvium deeply incised by modern drainages on the west flank of the Sevier Plateau, about 1.5 miles (3 km) southeast of Blind Spring Mountain; as much as several tens of feet thick.

Qla **Lacustrine sediment and alluvium** (Holocene) − Forms the meadow of Blue Spring Valley about 2 miles (3 km) southwest of Panguitch Lake, which we interpret to be a lake deposit made up of moderately to well-sorted, thinly bedded, light-gray and light-brown, fine-grained sand, silt, and clay derived principally from Brian Head strata in the Bunker and Deer Creek drainages; upper surface is marked by numerous small stream channels and meander cutoffs; also mapped near the east end of Navajo Lake where it consists of fine-grained sediment eroded from the pink member of the Claron Formation, and at Co-op Valley Sinks east of Parowan; likely several tens of feet thick in Blue Spring Valley; recent remedial work on the Navajo Lake dam shows that sediments there, as much as 37 feet (11 m) thick along the central portion of the dam, overlie basalt presumably of the Henrie Knolls lava flow (Qbhk) (Carl Ege, Utah Division of Water Rights, written communication, May 8, 2013).

Blue Spring Valley was flooded to form a shallow reservoir following completion of the Blue Spring Valley dam in the late 1800s or early 1900s; the small dam was breached by 1917 (Ipson and Ipson, 2008). The valley is now drained at its north end by Blue Spring Creek, which may have formed in response to the Miller Knoll lava flows that blocked the original Bunker Creek outlet at the southeast end of the valley, possibly as late as middle Holocene time.

Qlao **Older lacustrine sediment and alluvium** (Holocene and upper Pleistocene) − Similar to lacustrine sediment and alluvium (Qla), but forms incised, planar surfaces 5 to 10 feet (2–3 m) above the main meadow of Blue Spring Valley; likely several tens of feet thick.

Residual deposits

Qr **Relict Houston Mountain and Blue Spring Mountain lava flows** (Holocene and Pleistocene) − Mapped about 4 miles (7 km) north-northeast of Navajo Lake, where blocky remnants of the 5.3 Ma Houston Mountain lava flow (Tbhm) have been let down by sapping and dissolution of underlying beds; angular to subangular blocks of the lava flow, typically 3 feet (1 m) or less in diameter but locally as large as about 12 feet (4 m), locally form a basaltic pavement on the white member of the Claron Formation, but typically they are widely scattered; other than uncommon small fragments of chalcedony (itself likely the remains of the Brian Head Formation once buried by the Houston Mountain lava flow), no other exotic rock types are present; map unit probably formed as former basalt-capped hilltops succumbed to chemical weathering of carbonate beds in the underlying Claron Formation and concomitant hillslope erosion, undermining the lava flow and scattering resistant basalt blocks over the underlying bedrock, processes that began following development of inverted topography now capped by the Houston Mountain lava flow; unmapped colluvium derived from this unit blankets much of the nearby Claron Formation; also mapped on the south side of Blue Spring Mountain (about 5 miles [8 km] southwest of Panguitch Lake) where blocks of the Blue Spring Mountain lava flow (Tbbm) conceal the upper part of the white member of the Claron Formation and possibly the lower part of the Brian Head Formation; typically less than a few feet thick.

Stacked unit deposits

Stacked unit deposits comprise a discontinuous veneer of Quaternary deposits that mostly conceal underlying bedrock units. Although most bedrock in the quadrangle is partly covered by colluvium or other surficial deposits, we use stacked units to indicate those areas where bedrock is almost wholly obscured by surficial deposits that are derived from more than just residual weathering of underlying bedrock.

Qlao/Qbmk$_3$

Older lacustrine sediment and alluvium over the Miller Knoll lava flow (Holocene and upper Pleistocene/Holocene to upper Pleistocene) − Mapped along the southeast edge of Blue Spring Valley (about 2 miles [3 km] southwest of Panguitch Lake) where the oldest Miller Knoll lava flow (Qbmk$_3$) is partly concealed by a veneer of sediment interpreted as a mixture of lacustrine and alluvial, and possibly eolian, sand and silt; Blue Spring Valley likely drained through Black Rock Valley prior to the drainage being blocked by the Miller Knoll lava flows, with lacustrine and alluvial sediment accumulating in the basin upstream of the flows; surficial cover is likely less than 6 feet (2 m) thick.

Qc/Tbh

Colluvium over the Brian Head Formation (Holocene to Pleistocene/Oligocene to Eocene) – Mapped on the west flank of Houston Mountain (6 miles [10 km] east of Cedar Breaks National Monument) and south of the town of Brian Head, where colluvium, residual deposits, and possibly landslide deposits conceal the underlying Brian Head Formation; at Houston Mountain, colluvium includes large blocks of the Houston Mountain lava flow enclosed in a matrix of colluvium derived from weathered, tuffaceous Brian Head strata; also mapped on the flanks of the Sevier Plateau where colluvium derived mostly from the overlying Mount Dutton Formation obscures underlying Brian Head strata; surficial cover may exceed 20 feet (6 m) thick.

Qc/Tcwu

Colluvium over the upper limestone unit of the white member of the Claron Formation (Holocene and upper Pleistocene/Eocene) – Mapped on the southwest side of Houston Mountain (6 miles [10 km] east of Cedar Breaks National Monument) where colluvium conceals the underlying upper limestone unit of the white member of the Claron Formation; colluvium includes large blocks of the Houston Mountain lava flow enclosed in a matrix of colluvium derived from weathered, tuffaceous Brian Head strata and the upper limestone unit of the white member of the Claron Formation; surficial cover may exceed 10 feet (3 m) thick.

Qc/Tcw

Colluvium over the white member of the Claron Formation (Holocene and upper Pleistocene/Eocene) – Mapped southeast of the Town of Bryce, where the contact between the pink and white members of the Claron Formation is poorly exposed; strata mapped there as the white member may be an unusual facies of the uppermost pink member; surficial cover may exceed 10 feet (3 m) thick.

Qc/Tcp

Colluvium over the pink member of the Claron Formation (Holocene and upper Pleistocene/Eocene) – Mapped on the Paunsaugunt Plateau where a veneer of colluvium mostly conceals underlying bedrock; surficial cover mostly less than about 10 feet (3 m) thick.

Qc/Kws

Colluvium over the Wahweap and Straight Cliffs Formations (Holocene and upper Pleistocene/Upper Cretaceous) – Mapped west of Johns Valley where a veneer of colluvium mostly conceals underlying bedrock; surficial cover mostly less than about 10 feet (3 m) thick.

QUATERNARY-TERTIARY

Holocene(?) to upper Tertiary lava flows

Basaltic lava flows in the Panguitch 30′ x 60′ quadrangle are at the northern edge of the Western Grand Canyon basaltic field, which extends across the southwest part of the Colorado Plateau and the adjacent High Plateaus transition zone into the Basin and Range Province in southwest Utah, northwest Arizona, and adjacent Nevada (Hamblin, 1963, 1970, 1987; Best and Brimhall, 1970, 1974; Best and others, 1980; Smith and others, 1999; Johnson and others, 2010). This volcanic field contains hundreds of relatively small-volume, widely scattered, mostly basaltic lava flows and cinder cones that range in age from Miocene to Holocene. In southwestern Utah basalts are synchronous with basin-range deformation and are part of mostly small, bimodal (basalt and high-silica rhyolite) eruptive centers (Christiansen and Lipman, 1972; Rowley and Dixon, 2001).

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; Nealey and others, 1995, 1997; Nelson and Tingey, 1997; Nusbaum and others, 1997; Smith and others, 1999; Downing, 2000; Johnson and others, 2010). Major- and trace-element data for volcanic rocks in the map area are available at http://geology.utah.gov/online/analytical_data.htm. Nb/La ratios for virtually all samples of basaltic and andesitic lava flows from the map area are less than 1.0, thus suggesting a lithospheric mantle source (Fritton and others, 1991). Rock names are from LeBas and others (1986). Table 3 summarizes ages of basaltic rocks in this map area.

QTb **Basaltic lava flow, undivided** (Pleistocene? to Miocene?) – Medium- to dark-gray basalt lava flow that caps a ridge north of Wilson Creek, a southern tributary of Mammoth Creek, about 3 miles (5 km) northwest of Asay Bench; undated and correlation is uncertain, but major- and trace-element geochemistry shows affinities to the 5.3 Ma Houston Mountain lava flow, although its degree of topographic inversion suggests that it is not that old; about 20 to 30 feet (6–9 m) thick.

$Qbpl_1$, $Qbpl_2$, $Qbpl_3$

Panguitch Lake lava flows (upper Holocene? to upper Pleistocene?) – Mapped as three separate lava flows, with $Qbpl_1$ being the youngest, all three of which are mostly unvegetated and blocky, and exhibit steep flow fronts 100 to 200 feet (30–60 m) high: $Qbpl_1$ is dark-gray to black latite (potassium-rich trachyandesite) containing small (1 mm), stubby plagioclase phenocrysts in a glassy to aphanitic groundmass; $Qbpl_2$ and $Qbpl_3$ are dark-gray latite containing small stubby plagioclase and abundant acicular hornblende phenocrysts in a fine-grained groundmass; age uncertain, but may be as young as late Holocene; individual lava flows are typically about 200 feet (60 m) thick.

The $Qbpl_1$ lava flow lacks collapsed lava tubes and exhibits blocky flow lines similar to those of the Dry Valley lava flow (Qbdv). The smaller $Qbpl_2$ lava flow has collapsed lava tubes and partly buries the $Qbpl_3$ lava flow. The $Qbpl_3$ flow, which has abundant collapsed lava tubes and branching distributary lobes, erupted from a vent apparently now concealed by the younger vents of the $Qbpl_1$ and $Qbpl_2$ lava flows (immediately northeast of Miller Knoll, which is the large cinder cone [Qbmkc] about 3 miles [5 km] south of Panguitch Lake) and flowed northward about 3 miles (5 km) nearly to Panguitch Lake; this is the “northern Panguitch flow” of Stowell (2006).

Qbdv **Dry Valley lava flow** (upper Holocene? to upper Pleistocene?) – Dark-gray latite (potassium-rich trachyandesite) that contains olivine and abundant hornblende phenocrysts in an aphanitic to fine-grained groundmass; forms a thick, blocky, laterally restricted flow west of Black Rock Valley and north of Mammoth Creek that exhibits high, steep flow fronts (except at Dry Valley, immediately west of the vent, where a slightly older more fluid phase erupted); upper surface shows prominent arcuate ridges that reveal flow directions; vent area lacks scoria or cinders and there is no "tuff ring" as stated by Stowell (2006); northern flank of flow is partly vegetated, but upper surface and south-facing slopes are not vegetated; age uncertain, but overlies and is younger than the Miller Knoll lava flow ($Qbmk_2$—the "arcuate andesite flow" of Stowell, 2006); typically 100 to 120 feet (30–35 m) thick.

$Qbmk_1$, $Qbmk_2$, $Qbmk_3$, Qbmkc

Miller Knoll lava flows and cinder cone (middle Holocene? to upper Pleistocene) – Mapped as three separate lava flows in the Black Rock Valley area south of Panguitch Lake, with $Qbmk_1$ being the youngest flow: $Qbmk_1$ is dark-gray to black andesite that contains small (1 mm), stubby plagioclase phenocrysts in a glassy to aphanitic groundmass; $Qbmk_2$ and $Qbmk_3$ are dark- to medium-gray basaltic trachyandesite containing clusters of olivine, plagioclase, and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass and include both sodium-rich (mugearite) and potassium-rich (shoshonite) rock types, locally containing small, thin plagioclase phenocrysts; the $Qbmk_2$ lava flow, the "southern Panguitch flow" of Stowell (2006), yielded cosmogenic exposure ages of about 37,000 years (David Marchetti, Western State Colorado University, written communication, August 4, 2009); the $Qbmk_2$ and $Qbmk_3$ flows are thus likely late Pleistocene in age; the $Qbmk_1$ flow unit may be as young as middle Holocene; lava flows are typically 30 to 100 feet (10–30 m) thick, but may be thicker where they fill paleotopography.

The $Qbmk_1$ lava flow erupted from a vent near the top of the Miller Knoll cinder cone (Qbmkc, at the northwest end of Black Rock Valley) and forms a blocky, mostly unvegetated flow that looks morphologically similar to, and may be chemically transitional with, latite of the Panguitch Lake lava flows (Qbpl). The much larger $Qbmk_2$ lava flow erupted from vents on the south side of the Miller Knoll cinder cone and flowed about 4 miles (6 km) southeast through Black Rock Valley to Mammoth Creek, forming a young-looking, blocky, poorly vegetated flow that has abundant collapsed lava tubes and branching distributary lobes. The $Qbmk_3$ lava flow erupted from a vent now concealed by the Miller Knoll cinder cone; the lava flow is mostly well vegetated and was the first flow to block Blue Spring Valley—the western part of this flow is partly covered by old mixed lacustrine and alluvial deposits (Qlao) that we interpret as having accumulated upstream of the lava-flow dam.

The southern extent of the $Qbmk_2$ lava flow along Mammoth Creek was clearly limited by pre-existing topography of the pink member of the Claron Formation, but the flow now lies at the modern base level of Mammoth Creek. The lava flow once blocked Mammoth Creek, which has since eroded the adjacent, less-resistant Claron strata. Lacustrine sediments are absent upstream of the lava flow along Mammoth Creek, but stream terraces there may record partial infilling and subsequent exhumation of the valley.

Qbnl, Qbnlc

Navajo Lake lava flows and cinder cone (upper Pleistocene?) – Medium- to dark-gray mugearite (sodium-rich basaltic trachyandesite) containing clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; some lava flows contain common small plagioclase phenocrysts; lava flows (Qbnl) erupted from vents at a cinder cone (Qbnlc) about 3 miles (5 km) north of Navajo Lake (Moore and others, 2004) and coalesced into flow complexes not mapped separately; margins of flows typically form steep, blocky flow fronts 10 to 30 feet (3–9 m) high; cinder cone is well vegetated; lava flows are locally well vegetated, but more commonly barren and characterized by a rough, blocky surface; vegetated areas collect wind-blown sediment that forms a sparse soil cover on parts of the flows; age uncertain, but likely late Pleistocene on the basis of degree of incision and weathering, although Moore and others (2004) considered the lava flow as probably Holocene; lava flows are typically several tens of feet thick, but thicker where they fill paleotopographic lows.

Qbrd, Qbrdc

Red Desert lava flows and cinder cone (upper Pleistocene?) – Medium- to dark-gray basalt and basaltic andesite that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; some lava flows contain common small plagioclase phenocrysts; lava flows (Qbrd) erupted from vents at a cinder cone (Qbrdc) nearly 6 miles (10 km) north of Navajo Lake (Moore and others, 2004) and from a small vent about 3 miles (5 km) to the southeast, and they coalesced into flow complexes not mapped separately; margins of flows typically form steep, blocky flow fronts 10 to 30 feet (3–9 m) high; cinder cone is well vegetated; lava flows are locally well vegetated, but more commonly are barren and have a rough, blocky surface; vegetated areas collect wind-blown sediment that forms a sparse soil on parts of the flows; age uncertain, but lava flows are likely late Pleistocene on the basis of degree of incision and weathering, although Moore and others (2004) considered the lava flows as probably Holocene; lava flows are typically several tens of feet thick, but thicker where they fill paleotopographic lows.

Qbhk, Qbhkc

Henrie Knolls lava flows and cinder cones (upper Pleistocene) – Medium- to dark-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; some lava flows, particularly those between Duck Creek Sinks and Dry Camp Valley Spring (about 4 to 5 miles [6–8 km] east and northeast of Navajo Lake), also contain common plagioclase phenocrysts and have a slightly coarser groundmass; lava flows that erupted from the northeasternmost group of cinder cones, including Henrie Knolls, tend to be of basaltic andesite composition; margins of flows typically form steep, blocky flow fronts 10 to 30 feet (3–9 m) high; cinder cones are well vegetated; lava flows are locally well vegetated but more commonly barren, exhibiting a rough, blocky surface; age unknown, but probably late Pleistocene because the north end of flow complex is incised by Tommy Creek and capped by level 4 stream-terrace deposits (Biek and others, 2011, here mapped as Qat) inferred to be of late Pleistocene age; sample HK092106-1 near Henrie Knolls yielded a low-precision $^{40}Ar/^{39}Ar$ age of 0.058 ± 0.035 Ma (UGS and NMGRL, 2009); lava flows are typically several tens of feet thick, but likely exceed 200 feet (60 m) thick where they fill paleotopographic lows.

The wide chemical variation of the Henrie Knolls lava flows (Qbhk) reflects the fact that these flows erupted from at least 20 separate vents marked by cinder cones (Qbhkc), including the largest two cones at Henrie Knolls, in the northeast part of the flow complex. The cinder cones are strikingly aligned along a northeast trend, subparallel to mapped normal faults in the quadrangle. Although no fault that postdates eruption of the Henrie Knolls lava flows has been identified along this trend, a concealed, unmapped fault likely controls the alignment of vents. The southernmost of the Henrie Knolls lava flows blocked the Navajo Lake and nearby Dry Valley drainages, forming Navajo Lake and intermittent Cow Lake.

Qbmc, Qbmcc

Midway Creek lava flow and cinder cones (Pleistocene) – Medium- to dark-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; lava flow (Qbmc) erupted from a vent at a cinder cone (Qbmcc) and is partly covered by the Navajo Lake lava flow (Qbnl) (Moore and others, 2004); this cinder cone may also be the source of the Duck Creek lava flow (Qbdc); lava flow is typically several tens of feet thick, but thicker where it fills paleotopographic lows.

Qbde, Qbdec

Deer Valley lava flow and cinder cone (Pleistocene) – Medium- to dark-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; small lava flow (Qbde) erupted from a vent at a cinder cone (Qbdec) 1.5 miles (2.5 km) north of Navajo Lake; lava flow is typically several tens of feet thick, but thicker where it fills paleotopographic lows.

Qbho, Qbhoc

Horse Pasture lava flow and cinder cone (Pleistocene) – Medium- to dark-gray basalt and hawaiite (sodium-rich trachybasalt) containing clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; lava flow (Qbho) erupted from a vent at a cinder cone (Qbhoc) 4 miles (6 km) north of Navajo Lake; this cinder cone may also be the source of the Duck Creek lava flow (Qbdc); lava flow is typically several tens of feet thick, but thicker where it fills paleotopographic lows.

Qbdc **Duck Creek lava flow** (Pleistocene) – Medium-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts and abundant small plagioclase phenocrysts in a fine-grained groundmass; lava flow is typically partly concealed by a veneer of unmapped surficial deposits of alluvial, colluvial, and eolian origin; maximum exposed thickness is about 15 feet (5 m) near Aspen Mirror Lake, but likely several tens of feet thick where it fills paleotopography in the Duck Creek drainage.

The location of the vent of the Duck Creek lava flow is unknown, but it may be concealed by the Henrie Knolls (Qbhk) or Navajo Lake (Qbnl) lava flows. Alternatively, geochemical data suggest that the Duck Creek lava flow may be the distal part of either the Midway Creek or Horse Pasture lava flows. Regardless, the lava flowed east down the ancestral Duck Creek drainage and continued northeastward to at least the Bowers Flat area at the west edge of the Asay Bench quadrangle. The lava flow contains a long, open lava tube near Aspen Mirror Lake, just west of Duck Creek village (U.S. Forest Service restricts access). The age of the Duck Creek flow is uncertain, but it locally covers the Bowers Knoll lava flow (Qbbk) and in turn is locally covered by the Henrie Knolls lava flow (Qbhk), thus is probably late to late-middle Pleistocene. However, Johnson and others (2010) suggested that the distal end of the Bowers Knoll flow as mapped here, including the part that contains Mammoth Cave, may be the Duck Creek flow—if so, incision there suggests that the Duck Creek flow is about 500,000 years old, far older than indicated by the degree of incision along the upstream part of the flow.

Qbsk, Qbskc

Strawberry Knolls lava flows and cinder cones (Pleistocene) – Medium- to dark-gray potassic trachybasalt that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; lava flows (Qbsk) erupted from Strawberry Knolls (Qbskc), two cinder cones located about 2 miles (3 km) east of Duck Creek village, and flowed mostly northeast along Strawberry Creek to Uinta Flat; age uncertain, but cinder cones are well vegetated and flow is incised by Strawberry Creek as much as 40 feet (12 m) at its downstream end and so is probably late to middle Pleistocene; lava flows are typically 20 to 30 feet (6–9 m) thick, but doubtless many tens of feet thick near vent areas.

Qblhc **Lake Hollow cinder cone** (Pleistocene) – Forms a small, partly eroded cinder cone about 1.5 miles (3 km) north of Mammoth Creek and east of Black Rock Valley, with a small lava flow (not differentiated on this map) at the base of the cone of medium- to dark-gray hawaiite (sodium-rich trachybasalt) that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; vent is on-trend with the Henrie Knolls lava flows, to which it may be related; age unknown, but likely late to middle Pleistocene on the basis of its degree of erosion; lava flow is less than about 20 feet (6 m) thick.

Qbef, Qbefc

East Fork Deep Creek lava flow and cinder cone (Pleistocene) – Medium- to dark-gray, fine-grained olivine basalt lava flow (Qbef) west of Navajo Lake; cinder cone (Qbefc) is deeply eroded due to its location just below the western escarpment of the Markagunt Plateau, just west of Navajo Lake; the distal southern end of this flow was called the Three Creeks lava flow by Biek and Hylland (2007), which they estimated to be less than 300,000 years old on the basis of degree of incision and comparison with nearby dated lava flows; lava flow is probably 20 to 40 feet (6–12 m) thick.

Qbw, Qbwc

Water Canyon lava flow and cinder cone (middle Pleistocene) – Dark-gray potassic trachybasalt and shoshonite (potassium-rich basaltic trachyandesite) that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; quartz xenocrysts common; lava flow (Qbw) erupted from cinder cone (Qbwc) in Water Canyon about 3 miles (5 km) southeast of Paragonah (Maldonado and Moore, 1995); Fleck and others (1975) reported a K-Ar age (corrected according to Dalrymple, 1979) of 0.44 ± 0.04 Ma for this flow; lava flow is as much as 200 feet (60 m) thick where it partly fills Water Canyon.

Qbbk, Qbbkc

Bowers Knoll lava flow and cinder cones (middle Pleistocene) – Medium-gray mugearite (sodium-rich basaltic trachyandesite) containing abundant clusters of olivine, plagioclase, and clinopyroxene phenocrysts in a fine-grained groundmass; lava flow erupted from Bowers Knoll, a cinder cone (Qbbkc) about 3 miles (5 km) northeast of Duck Creek village, about 10 miles (16 km) south of Panguitch; forms rugged, heavily vegetated, blocky surface having steep flow fronts 40 feet (12 m) or more high; as mapped, contains Mammoth and Bower caves, large open lava tubes, but that part of the flow may belong to the Duck Creek flow (Johnson and others, 2010); age uncertain, but locally underlies the Duck Creek lava flow (Qbdc), so is probably middle Pleistocene; Best and others (1980) reported a K-Ar age of 0.52 ± 0.05 Ma for the nearby Asay Knoll lava flow (Qbak), which exhibits a similar degree of incision and weathering; typically 40 feet (12 m) or more thick near flow margins, but may exceed 100 feet (30 m) thick near the central part of the flow.

Qbrc **Red Canyon lava flow and cinders** (middle Pleistocene) – Medium- to dark-gray andesite and mugearite (sodium-rich basaltic trachyandesite); lava flow apparently erupted from several vent areas (which have abundant cinders but no cone morphology) on the hanging wall of the Sevier fault and flowed eastward across the fault where it caps Black Mountain near the entrance to Red Canyon; Lund and others (2008) reported $^{40}Ar/^{39}Ar$ ages of 0.49 ± 0.04 Ma (footwall) and 0.51 ± 0.01 Ma (hanging wall), and Best and others (1980) reported a K-Ar age of 0.56 ± 0.07 Ma for this flow on the hanging wall; Lund and others (2008) documented 630 to 740 feet (192–225 m) of displacement on the flow, yielding a vertical-slip-rate estimate of 0.38-0.44 mm/yr; lava flow is typically about 80 feet (25 m) thick, but thicker near vent areas.

Qbak, Qbakc

Asay Knoll lava flow and cinder cone (middle Pleistocene) – Medium- to dark-gray potassic trachybasalt and shoshonite (potassium-rich basaltic trachyandesite) that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; lava flow (Qbak) erupted from Asay Knoll cinder cone (Qbakc) and covers Asay Bench; Best and others (1980) reported a K-Ar age of 0.52 ± 0.05 Ma for this flow; lava flow is typically 20 to 30 feet (6–9 m) thick, but is doubtless many tens of feet thick near vent area.

Qbck, Qbckc

Cooper Knoll lava flow and cinder cone (middle to lower Pleistocene) – Medium-gray basalt that contains clusters of olivine, plagioclase, and clinopyroxene phenocrysts in a fine-grained groundmass; lava flow (Qbck) erupted from a vent at a cinder cone (Qbckc) on the south flank of Cooper Knoll, about 1 mile (1.6 km) southeast of Panguitch Lake; overlies stream gravels containing rounded pebbles and cobbles of the Isom Formation, mafic and intermediate volcanic rocks of the Mount Dutton Formation, chalcedony, and minor quartzite; sample HM101408-1 yielded a discordant age spectrum showing that the flow is less than or equal to 0.92 Ma (UGS and NIGL, 2013); more likely, it may be about 500,000 years old on the basis of comparison with the similarly incised Asay Bench lava flow (Qbak) for which Best and others (1980) reported a K-Ar age of 0.52 ± 0.05 Ma; lava flow is about 20 to 40 feet (6–12 m) thick.

Qbpk **Pryor Knoll lava flow** (Pleistocene) – Basaltic lava flow present at the southwest edge of the map area; may have erupted from Pryor Knoll in the adjacent Cedar Mountain quadrangle; age uncertain, but may be about 1 Ma on the basis of comparison with nearby dated lava flows and its degree of dissection; lava flow is typically several tens of feet thick.

Qbwf, Qbwfc

Webster Flat lava flow and cinder cone (middle Pleistocene) – Medium-gray, fine-grained olivine basalt with small plagioclase phenocrysts; lava flow (Qbwf) erupted from vent at cinder cone (Qbwfc) about 1 mile (1.6 km) east of Black Mountain in the southwest corner of the map area and flowed mostly south down the Kolob Terrace; age uncertain, but probably about 500,000 years old on the basis of comparison with nearby dated lava flows and its degree of dissection; lava flow is typically several tens of feet thick.

Qbal, Qbalc

Aspen Lake lava flow and cinder cone (middle Pleistocene) – Medium-gray, fine-grained olivine basalt with small plagioclase phenocrysts; lava flow (Qbal) erupted from vent at cinder cone (Qbalc) about 1 mile (1.6 km) south of Black Mountain in the southwest corner of the map area and flowed mostly south down the Kolob Terrace; age uncertain, but probably about 500,000 years old on the basis of comparison with nearby dated flows and its degree of dissection; lava flow is typically several tens of feet thick.

Qblf, Qblfc

Long Flat lava flow and cinder cones (middle Pleistocene) – Medium-gray basalt to hawaiite (sodium-rich trachybasalt) that contains clusters of olivine and clinopyroxene phenocrysts; lava flow (Qblf) erupted mostly from hills 10,392 and 10,352 (Brian Head 7.5′ topographic quadrangle map; Rowley and others, 2013), which are two cinder cones (Qblfc) near Long Flat about 3 miles (5 km) east of Brian Head peak; a third, smaller cinder cone is near the southeast margin of the flow near State Highway 143; parts of the lava flow are covered by Pinedale-age glacial till and glacial outwash, and the cinder cones appear to be more heavily eroded than the nearby Hancock Peak cinder cone (Qbhpc); the northeast flank of hill 10,392 is conspicuously truncated and it may have been eroded by an earlier glacial advance (if so, likely the Bull Lake [Illinoian or MIS 6] advance); Stowell (2006) reported an $^{40}Ar/^{39}Ar$ maximum isochron age of 0.60 ± 0.25 Ma for sample LEA71SS2, which is likely from the Long Flat lava flow, but minor- and trace-element signatures of the Long Flat and nearby Hancock Peak flows are similar and Stowell's sample location lacks precision to be properly located, thus the age of the Long Flat flow is uncertain; lava flow is several tens of feet thick.

Qbwk, Qbwkc

Wood Knoll lava flow and cinder cone (middle Pleistocene) – Medium- to dark-gray, fine-grained olivine basalt; lava flow (Qbwk) erupted from a vent at Wood Knoll, a cinder cone (Qbwkc) about 2 miles (3 km) southwest of Cedar Breaks National Monument, and flowed northwest into Long Hollow; a remnant of the flow, perched 1100 feet (335 m) above the junction of Ashdown Creek and Coal Creek, yielded an $^{40}Ar/^{39}Ar$ age of 0.63 ± 0.10 Ma and a long-term down-cutting rate of 0.53 mm/yr (about 21 inches per thousand years or 1700 ft/myr), inferred to be a minimum rate of relative uplift on the Hurricane fault to the west (Lund and others, 2007); lava flow is typically several tens of feet thick, but is as much as about 300 feet (90 m) thick where it fills the ancestral Coal Creek channel.

Qbub, Qbubc

Upper Bear Springs lava flows and cinder cones (middle to lower Pleistocene) – Medium- to dark-gray, fine-grained olivine basalt; lava flows (Qbub) erupted from vents at cinder cones (Qbubc) about 2 miles (3 km) southwest of Navajo Lake and flowed mostly south onto the Kolob Terrace; probably about 750,000 years old because they appear to be the same lava flows as those at Horse Knoll (Sable and Hereford, 2004; Doelling, 2008), which yielded a K-Ar age of 0.81 ± 0.05 Ma and an $^{40}Ar/^{39}Ar$ age of 0.73 ± 0.02 Ma (Biek and Hylland, 2007; UGS and NMGRL, 2008); lava flows are several tens of feet thick.

Qbbm **Black Mountain lava flow** (middle to lower Pleistocene) – Medium-gray, fine-grained olivine basalt with small plagioclase and pyroxene phenocrysts; lava flow caps the northwest-sloping surface of Black Mountain in the southwest corner of the map area; vent unknown but may be concealed by nearby younger lava flows or surficial deposits to the southeast; yielded K-Ar ages of 0.80 ± 0.24 (Anderson and Mehnert, 1979) and 0.87 ± 0.24 Ma (Best and others, 1980); typically several tens of feet thick.

$Qbhp_1$, $Qbhp_2$, Qbhpc

Hancock Peak lava flows and cinder cone (middle to lower Pleistocene) – Medium-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts in a fine-grained groundmass; on the basis of chemistry and morphology, the map unit is divided into two flows, both of which are well vegetated; erupted from Hancock Peak, a large, well-preserved cinder cone (Qbhpc) 5 miles (8 km) southeast of Brian Head peak; $Qbhp_1$ appears to overlie $Qbhp_2$ and extends farther downstream where it caps an inverted valley about 600 feet (180 m) above Mammoth Creek just north of the community of Mammoth Creek; age unknown, but estimated to be middle to early Pleistocene on the basis of comparison with the apparently 60,000-year-old Long Flat lava flow (Qblf) and the 2.8 Ma Blue Spring Mountain lava flow (Tbbm); lava flows are typically several tens of feet thick, but likely exceed 100 feet (30 m) thick where they fill paleotopographic lows.

Qbtp **The Pass lava flow** (middle to lower Pleistocene?) – Medium- to dark-gray basalt that contains clusters of olivine and clinopyroxene phenocrysts in a fine-grained groundmass; caps a small knob just south of The Pass, about 1 mile (1.6 km) east of Panguitch Lake; Wagner (1984) interpreted the unit to be a small gabbroic intrusion, but more likely it is a flow remnant partly involved in a landslide; chemically similar to the 5.3 Ma Houston Mountain lava flow (Tbhm), but source is uncertain; probably about 50 feet (15 m) thick.

Qbc **Second Left Hand Canyon vent** (middle to lower Pleistocene?) – Forms deeply eroded basaltic vent area (Maldonado and others, 1997). Lower part contains abundant angular blocks of Claron Formation and basaltic volcanic rocks and minor rounded quartzite pebbles and cobbles; the whole mass is cut by several basaltic dikes; some blocks are as large as 12 feet (4 m) in size, but most are pebble to small cobble size; unbedded and appears to be a volcanic mudflow deposit. Upper part is mostly basaltic blocks and fewer Claron blocks, welded into scoriaceous matrix. Unconformably overlies the capping sandstone members of the Wahweap Formation on the northwest side of Henderson Hill in Second Left Hand Canyon, about 4 miles (6 km) south of Parowan; lies about 600 feet (180 m) above the modern drainage; may be associated with adjacent basaltic dikes; about 400 feet (120 m) thick.

Qbs, Qbsc

Summit lava flow and cinder cone (lower Pleistocene) – Medium- to dark-gray, fine-grained olivine basalt that Maldonado and others (1997) referred to as the Cinder Hill cone and flow; lava flow (Qbs) erupted from a vent at a cinder cone (Qbsc) at the base of the Hurricane Cliffs, about 2 miles (3 km) southwest of Summit; lava flow also crops out at the southeast margin of the Red Hills and is presumed to underlie the southern part of the Parowan Valley graben, where it is displaced by graben-bounding faults; yielded K-Ar ages of 1.00 ± 0.16 Ma and 0.94 ± 0.14 Ma (Anderson and Mehnert, 1979); lava flow is typically several tens of feet thick.

Qbe **Elliker Basin lava flow** (lower Pleistocene) – Medium- to dark-gray, fine-grained olivine basaltic trachyandesite; vent area unknown, but minor scoria and blocky flow breccia is present on the north rim of Elliker Basin, suggesting that the vent could underlie the basin, which is southwest of Summit; yielded K-Ar ages of 1.00 ± 0.16 Ma and 1.11 ± 0.11 Ma (Anderson and Mehnert, 1979); typically several tens of feet thick.

Qbrh, Qbrhc

Red Hills lava flows and cinder cones (lower Pleistocene) – Medium- to dark-gray, fine-grained basaltic andesite with small olivine and plagioclase phenocrysts; lava flows (Qbrh) erupted from vents at three cinder cones (Qbrhc) in the southern Red Hills

(Rowley and Threet, 1976); lava flows are mostly covered by eolian sand and silt, and locally by small areas of fan alluvium, but due to map scale only the larger of such areas are mapped; lava flow is cut by faults associated with the Red Hills horst and graben; yielded K-Ar ages of 1.28 ± 0.4 Ma (Anderson and Mehnert, 1979) and 1.30 ± 0.4 Ma (Best and others, 1980); lava flow is typically several tens of feet thick.

Tbbm, Tbbmc

Blue Spring Mountain lava flow and cinder cone (Pliocene) – Medium-gray hawaiite and mugearite (sodium-rich trachybasalt and basaltic trachyandesite, respectively) lava flow (Tbbm) that contains clusters of olivine and clinopyroxene phenocrysts in an aphanitic to fine-grained groundmass; erupted from vents at a cinder cone (Tbbmc) on Blue Spring Mountain and flowed east and south, mostly toward the ancestral Mammoth Creek drainage; an erosional outlier caps Mahogany Hill, about 500 feet (150 m) above Mammoth Creek east of its intersection with Black Rock Valley; the cinder cone is deeply eroded and the lava flow is well vegetated; between Blue Spring Mountain and Blue Spring Valley the flow is involved in a large landslide complex, which slid on the underlying Brian Head Formation; lava flow is typically several tens of feet thick, but is doubtless thicker near the vent area and where it fills paleotopographic lows.

Stowell (2006) reported an $^{40}Ar/^{39}Ar$ plateau age of 2.78 ± 0.16 Ma for what is likely the Blue Spring Mountain lava flow, but Stowell's sample location lacks precision to be properly located, thus age is uncertain. On the basis of their similar chemistry, the map unit includes a northeast-trending dike at the north end of Blue Spring Valley and a small remnant a few tens of feet above Blue Spring Creek (Biek and others, 2014). If the Blue Spring Creek remnant is indeed part of the 2.8-million-year-old Blue Spring Mountain lava flow, it means that the Blue Spring Valley area has been a topographic low for nearly the past 3 million years, an unlikely scenario. Alternately, this remnant may be a small basaltic sill.

Tbfr **Fivemile Ridge lava flow** (Pliocene?) – Medium-gray basalt containing clusters of olivine and clinopyroxene phenocrysts in a fine-grained groundmass; erupted from deeply eroded vent area on Fivemile Ridge about 8 miles (13 km) northeast of Panguitch Lake; appears to be gently folded into a northeast-trending syncline; age uncertain, but degree of incision suggests that it is Pliocene in age; typically several tens of feet thick.

Tbhb **Horse Bench lava flow** (Pliocene?) – Medium-gray andesite that caps Horse Bench, about 8 miles (13 km) north-northeast of Panguitch Lake; on the basis of major- and trace-element chemistry of a single sample, appears to be a different flow than the Fivemile Ridge flow, but location of vent is unknown; age uncertain, but degree of incision suggests that it is Pliocene in age; typically several tens of feet thick.

Tbhm **Houston Mountain lava flow** (lower Pliocene to upper Miocene) – Medium-gray basalt containing clusters of olivine and clinopyroxene phenocrysts in a fine-grained groundmass; unconformably overlies the Brian Head Formation (Tbhv) and Leach Canyon Formation (Tql) along the west edge of Blue Spring Mountain; an erosional outlier on the south side of Clear Creek, about 3 miles (5 km) west-northwest of Panguitch Lake, contains abundant 1- to 2-mm-long plagioclase phenocrysts, but is otherwise chemically similar to and included in the Houston Mountain flow map unit; unit also caps Houston Mountain (about 6 miles [10 km] east of Cedar Breaks National Monument) and other hills of lower elevation to the south (about 3 miles [5 km] northeast of Navajo Lake), where it is typically platy weathering; source vent unknown and margins of lava flow are entirely eroded away, but elevation of remnants suggests flow was derived from the west of its current exposures, probably in the Brian Head quadrangle, likely at a vent now eroded and concealed by younger deposits; sample HK092006-3 yielded an $^{40}Ar/^{39}Ar$ whole-rock age of 5.27 ± 0.14 Ma (UGS and NMGRL, 2009); maximum thickness is about 140 feet (43 m) at Houston Mountain.

Tbdh, Tbdhc

Dickinson Hill lava flows and cinder cones (lower Pliocene to upper Miocene) – Medium-gray basalt containing clusters of olivine and clinopyroxene phenocrysts in a fine-grained groundmass; interbedded with upper Tertiary fan alluvium (Taf); lava flows (Tbdh) erupted from vents at deeply eroded cinder cones (Tbdhc) southwest of Panguitch; Anderson and Christenson (1989) reported a K-Ar age of 5.3 ± 0.5 Ma for one of these lava flows; major- and trace-element chemistry and age are similar to those of the nearby Rock Canyon lava flow, making differentiation of the two lava flows uncertain in the area southeast of Panguitch; exposed thickness of lava flows is as much as 65 feet (20 m), and cinder deposits are 3 to 10 feet (1–3 m) thick.

Tbrc, Tbrcc

Rock Canyon lava flow and cinder cone (lower Pliocene to upper Miocene) – Medium-gray potassic trachybasalt and basalt that contains clusters of olivine and clinopyroxene and small plagioclase phenocrysts in a fine-grained groundmass; lava flow (Tbrc) is interbedded with upper Tertiary fan alluvium (Taf); erupted from a cinder cone (Tbrcc) about 4 miles (6 km) north-northwest of Hatch; apparent age and major- and trace-element chemistry are similar to those of the nearby 5.3 Ma Dickinson Hill lava flow (Tbdh); query indicates our uncertainty in correlating these two flows in their area of possible overlap; maximum exposed thickness is about 100 feet (30 m).

On the basis of limited geochemistry, lava flow remnants exposed in the footwall and hanging wall of the Sevier fault just south of State Route 12 and Red Canyon may be the Rock Canyon lava flow. Lund and others (2008) reported $^{40}Ar/^{39}Ar$ ages on the flow near Red Canyon of 4.94 ± 0.03 Ma (footwall outcrop) and 4.98 ± 0.03 Ma (hanging-wall outcrop), and our sample H101508-4 yielded an $^{40}Ar/^{39}Ar$ plateau age of 5.25 ± 0.03 Ma (UGS and NIGL, 2013). If this correlation is correct, it implies that the valley of the Sevier River and the footwall of the Sevier fault zone did not exist as topographic barriers to eastward movement of the Rock Canyon lava flow in early Pliocene time, about 5 million years ago. It further implies that the Rock Canyon lava flow underlies parts of Hatch valley. Regardless, Lund and others (2008) documented 775 to 1130 feet (237–344 m) of displacement on the flow near Red Canyon, yielding a vertical-slip-rate estimate of 0.05–0.07 mm/yr for this part of the Sevier fault, but this estimate failed to take into account displacement on a concealed western splay of the fault and so is significantly lower than the estimate determined from the nearby but younger 0.5 Ma Red Canyon lava flow (see Sevier fault discussion).

Tbsp **Sidney Peaks lava flow** (lower Pliocene? to upper Miocene?) – Medium-gray basalt containing clusters of olivine and clinopyroxene phenocrysts as much as 0.25 inch (5 mm) in diameter in a fine-grained groundmass; forms deeply dissected flow and flow breccia that unconformably overlie the Markagunt gravity slide; deposit just northeast of Sidney Peaks, which unconformably overlies the Leach Canyon Formation, consists of lava blocks in a cinder matrix, is locally cut by basaltic dikes, and may be a deeply eroded vent area; age uncertain, but probably early Pliocene or late Miocene on the basis of comparison with dated flows elsewhere on the Markagunt Plateau; as much as 80 feet (25 m) thick.

QTbx **Markagunt gravity slide residuum** (Holocene to Miocene?) – Mapped south of the Markagunt gravity slide, principally at and near the west rim of the Markagunt Plateau, but also mapped where it caps several small hills south of Panguitch Lake and in the upper reaches of Rock Canyon, south of Haycock Mountain. Consists of two units not mapped separately: (1) large blocks and rubble of the Isom Formation present in the northeast part of Cedar Breaks National Monument, and (2) an unconsolidated mix of Isom boulders and minor Brian Head debris present in the southern part of the monument and in the deposits near Panguitch Lake. In both units, Isom blocks are subangular and locally as much as 16 feet (5 m) long, but in the northeast part of the monument, rafted Isom blocks are as much as several hundred feet in extent; many of the Isom blocks are internally brecciated on a fine scale, then rehealed by silicification to become a resistant rock (Hatfield and others, 2010), presumably by redeposition of silica from ash shards and fine breccia fragments of the Isom ash-flow tuff itself (the brecciation is a direct result of deformation within the Markagunt gravity slide). Maximum thickness is about 150 feet (45 m) at Blowhard Mountain immediately south of Cedar Breaks National Monument, but most deposits are 5 to 30 feet (2–9 m) thick.

Exposures on the west side of Blowhard Mountain reveal large fractured blocks of Brian Head tuffaceous mudstone, sandstone, micritic limestone, and chalcedony, with some blocks as large as 100 feet (30 m) long, in addition to the ubiquitous, angular Isom boulders. There, the map unit overlies alluvial boulder gravel (undifferentiated from the residuum) as much as about 50 feet (15 m) thick that consists of subrounded Isom Formation clasts, minor Brian Head limestone and chalcedony clasts, and, near the base of the deposit, rare Claron clasts. Isom clasts are commonly 2 to 3 feet (0.5–1 m) in diameter and locally as much as 9 feet (3 m) long and include at least one large boulder of Isom flow breccias similar to that at Brian Head peak. No other clast types are present. Much of the west side of this outcrop belt at Blowhard Mountain is involved in landslides, many with historical movement, but the gravel is clearly part of a channel eroded into the Claron Formation and overlain by mass-wasting deposits composed of Isom and Brian Head debris.

Between Blowhard Mountain and Long Valley Creek, these residuum deposits form an extensive, hummocky surface draped over the Claron Formation. Elsewhere on Blowhard Mountain and areas to the east and south, the map unit is characterized by abundant or scattered, large, locally internally brecciated, angular Isom boulders that litter the surface. The hummocky surface is mostly due to dissolution and collapse of underlying Claron limestone, which created numerous sinkholes in this area (Moore and others, 2004; Hatfield and others, 2010; Spangler, 2010), but the surface is also likely due in part to ongoing slumping and slope creep that results from admixed tuffaceous Brian Head strata (Moore, 1992).

The large blocks of Isom in this map unit in and near the northeast part of Cedar Breaks National Monument rest unconformably on Brian Head and Claron strata, whereas not far to the north, at a higher elevation, the main mass of the Markagunt gravity slide rests on Leach Canyon Formation. Because the southern margin of the gravity slide (and underlying regional ash-flow tuffs) is erosional in nature, we do not know its southern depositional limit. However, because debris from the Leach Canyon Formation is missing in areas mapped as QTbx in the northeast part of Cedar Breaks National Monument and at and near Blowhard Mountain, it seems likely that the Leach Canyon (and underlying in-place Isom) pinched out southward against paleotopography and did not extend much farther south than its present-day outcrop. Because there is no evidence for a post-Leach Canyon (but pre-Markagunt gravity slide) unconformity that cuts out strata southward across the west edge of the Markagunt Plateau, senior-author Biek interprets the large Isom blocks to be old landslide remnants that are at a lower topographic and structural level than the main mass of the Markagunt gravity slide. In this interpretation, the large Isom blocks and Isom and Brian Head debris are inferred to be derived from northward retreat of the erosional escarpment that stretches from Brian Head peak eastward to Haycock Mountain. Thus in this view, similar to that suggested by Moore (1992) and Sable and Maldonado (1997a), the map unit is a remobilized part of the Markagunt gravity slide, the southern extent of which was emplaced on a paleohigh of Brian Head strata that served to constrain the southern limit of the Isom and Leach regional ash-flow tuffs. The gravity slide has since been let down to its present position principally by landsliding in late Tertiary and Quaternary time, with smectitic clays of the Brian Head Formation initially providing the weak shear surface for downslope movement of the blocks, which were then further dispersed by colluvial and slope-creep processes. An alternative interpretation, however, suggests that all but the southernmost blocks are a bona fide part of the Miocene Markagunt gravity slide (Moore and others, 2004; Hatfield and others, 2010; Rowley and others, 2013), which thus must have been emplaced as far south as Blowhard Mountain; subsequent weathering and sapping of the gravity slide and underlying Claron Formation then spread debris southward to the area beyond State Route 14. Yet another interpretation is that the map unit may be the distal remains of debris avalanche deposits at the south margin of the Markagunt gravity slide. Our differences in interpretation reflect our incomplete understanding of these deposits.

QTap **High-level pediment alluvium** (Pleistocene and Pliocene?) – Moderately sorted, subrounded to rounded pebble to boulder gravel and sand that form a gently east-dipping, locally resistant cap on upper Tertiary fan alluvium (Taf) near the east margin of the Markagunt Plateau; surface of deposit typically covered by veneer of pebble and cobble residuum; divisible into two different levels (Moore and Straub, 1995), but undivided here due to map scale; deposited principally as debris flows and debris floods, and in ephemeral stream channels; probably less than 20 feet (6 m) thick.

QTaf **Quaternary and late Tertiary fan alluvium** (Pleistocene to Pliocene) – Poorly to moderately sorted, non-stratified, subangular to subrounded, clay- to boulder-size sediment with poorly to moderately developed calcic soils (caliche); forms deeply dissected surfaces with no remaining fan morphology in the Red Hills; prominent clasts include Tertiary volcanic rocks, pale-reddish-orange and light-pinkish-gray limestone and calcareous mudstone of the Claron Formation, and yellowish-brown Cretaceous siltstone, sandstone, and coquina; includes lesser amounts of chalcedony and recycled quartzite, and rare quartz monzonite; overlain by the 1.0 Ma Summit basaltic lava flow (Qbs); deposited principally as debris flows; unconformably overlies Claron Formation and Isom Formation; several tens of feet thick in this map area, but at least 200 feet (60 m) thick in the adjacent Enoch quadrangle to the west (Knudsen, 2014b).

QTh **Basin-fill deposits of Long Hill** (Pleistocene and Pliocene?) – Poorly sorted, poorly stratified, boulder- to clay-size sediment containing subangular to subrounded clasts preserved in down-dropped blocks on the west side of the Red Hills; northern exposures consist predominantly of volcanic clasts, some as much as 3 feet (1 m) in diameter, whereas southern exposures contain abundant quartzite cobbles in a reddish-brown calcareous matrix; original depositional form is not preserved; interpreted to represent deeply eroded basin fill deposited principally as debris flows and debris floods on large alluvial fans; Maldonado and Williams (1993a) mapped kilometer-scale blocks of Oligocene and Miocene ash-flow tuffs within this basin-fill unit that they interpreted to be gravity-slide blocks of Pliocene or Pleistocene age; senior-author Biek reinterprets these blocks simply as autochthonous normal-fault-bounded blocks partly covered by basin-fill deposits; exposed thickness as much as about 300 feet (90 m).

QTlf **Fine-grained lacustrine sediment** (lower? Pleistocene to Pliocene) – Fine-grained sand and silt with abundant, small, high-spired gastropods present just west of Wildcat Hollow and north of Panguitch Creek about 5 miles (8 km) southwest of Panguitch; may represent lacustrine deposits that accumulated in an upstream basin created by the Fivemile Ridge lava flow where it dammed ancestral Panguitch Creek; exposed thickness about 20 feet (6 m).

TERTIARY

Taf **Upper Tertiary fan alluvium** (Pliocene to Miocene) – Moderately to poorly consolidated, brown and grayish-brown sandstone, siltstone, pebbly sandstone, and conglomerate that form an incised, east-tilted alluvial-fan surface of low, rounded hills along the west side of Panguitch and Hatch valleys; clasts are of various volcanic rocks (95%) and about 5% quartzite and sandstone (Kurlich and Anderson, 1997); clasts were derived from the west and north from erosion of the Mount Dutton Formation and regional ash-flow tuffs and deposited as aggrading alluvial fans, possibly in a structurally closed basin later incised by through-going drainage of the Sevier River (Moore and Straub, 1995; Kurlich and Anderson, 1997); includes uncommon, thin, ash-fall tuff beds; interbedded with upper Tertiary basaltic lava flows (including the 5.0 Ma Rock Canyon lava flows [Tbrc] and the 5.3 Ma Dickinson Hill lava flows [Tbdh]) and uncommon, thin, lenticular beds of lacustrine limestone; east part of the outcrop belt locally includes upper Tertiary stream alluvium representing an axial valley stream; unconformably overlies Claron, Brian Head, Isom, and Limekiln Knoll strata and locally capped by pediment alluvium (QTap); as much as 760 feet (230 m) thick in the Hatch quadrangle (Kurlich and Anderson, 1997) and at least 1000 feet (300 m) thick in the Panguitch 7.5′ quadrangle (Moore and Straub, 1995).

The deposits were previously referred to as the Sevier River Formation, which was named by Callaghan (1938) for partly consolidated basin-fill deposits near Sevier, Utah, on the north side of the Marysvale volcanic field (see, for example, Anderson and Rowley, 1975; Anderson and others, 1990a; Moore and others, 1994; Rowley and others, 1994a). The name Sevier River Formation formerly had value in reconnaissance-scale studies in the High Plateaus; in and near its type area near the town of Sevier, it contains ash-fall tuffs that have fission-track and K-Ar ages of 14 and 7 Ma and basaltic lava flows that have K-Ar ages of 9 and 7 Ma (Steven and others, 1979; Best and others, 1980; Rowley and others, 2002), which are older than this map unit. In later, more detailed mapping in the High Plateaus, the name Sevier River Formation was restricted to its type area for older basin-fill sediments deposited in post-20 Ma basins, but that preceded development of the present topography (Rowley and others, 2002); later basin-fill deposits of the main phase of basin-range deformation in the northern Marysvale area were referred to as “sedimentary basin-fill deposits (QTs)” (Rowley and others, 2002). Co-author John Anderson (verbal communication, November 16, 2004) referred to these deposits in the Panguitch area as the Panguitch gravels.

The Sevier River Formation and other late Tertiary basin-fill deposits provide control on structural development of the High Plateaus area. Rowley and others (1981) used K-Ar ages of mapped volcanic rocks in the Sevier Plateau to the north to constrain the main phase of basin-range faulting to between 8 and 5 Ma, during which time the Sevier Plateau was uplifted along the Sevier fault zone at least 6000 feet (2000 m). This timing is supported by $^{40}Ar/^{39}Ar$ ages of about 7 Ma for the formation of alunite and natrojarosite at Big Rock Candy Mountain, in Marysvale Canyon 35 miles (55 km) north of the map area, due to oxidation (supergene alteration) as a result of downcutting by the Sevier River to expose previously altered rocks in the mountain (Cunningham and others, 2005).

Pediment deposits preserved atop the Spry intrusion, about 400 feet (120 m) above Circleville Canyon immediately north of the map area (Anderson and others, 1990a), led Anderson (1987) to suggest that basin-fill deposits once filled the ancestral valley of the Sevier River to a similar depth above the modern river. Some support for this interpretation is a similar pediment capped by Sevier River Formation over 1000 feet (300 m) above the Sevier River in northern Marysvale Canyon about 40 miles (64 km) north of the map area (Eardley and Beutner, 1934; Rowley and others 1988a, 1988b). This interpretation may be correct, although senior-author Biek sees no evidence for such exhumation of late Tertiary fan alluvium in the map area. Rather, he suggests that the structural high of the Spry intrusion and its capping pediment deposits may be due to an inferred fault segment boundary of the Sevier fault zone; that is, the long-term displacement rate there may be lower than that in the basins to the south and to the north. Thus, Biek interprets the capping pediment deposits simply to be remnants stranded by continued downcutting of the Sevier River as a result of differential slip on the Sevier fault, not due to exhumation of

Panguitch valley and Circleville Valley basin-fill deposits.

Map unit Taf is also preserved west and south of Henderson Point immediately north of Bryce Canyon National Park. There, the deposits consist of moderately consolidated, gently west-dipping volcaniclastic sand and gravel with minor clasts of quartzite and Claron Formation limestone.

Tvf, Tvg

Upper Tertiary basin fill (Miocene) – **Fine-grained strata** (Tvf) is exposed along the valley of the Sevier River southeast of Panguitch and is light-brown, pinkish-gray, and white tuffaceous mudstone, siltstone, fine-grained sandstone, and local diatomite (Crawford, 1951); moderately to poorly consolidated; laminated to thick beds, locally with small gastropods; contains few thin beds of peloidal micritic limestone; likely deposited in small lake basins and floodplains (Moore and Straub, 1995); exposed thickness about 100 feet (30 m). **Coarse-grained strata** (Tvg) is preserved in fault blocks along the Sevier fault zone between Red and Hillsdale Canyons, and also present in nearby footwall exposures between Wilson Peak and Black Mountain; it consists of brownish-gray volcaniclastic conglomerate with pebble- to boulder-size clasts mostly of intermediate volcanic rocks but also minor basaltic clasts and rounded quartzite clasts; locally includes medium- to coarse-grained sandstone and minor mudstone; typically poorly exposed, but outcrop habit suggests that exposed parts are as much as 400 feet (120 m) thick. These basin-fill deposits are undated but predate upper Tertiary fan alluvium (Taf) and so are likely Miocene in age; they are in excess of 1000 feet (300 m) thick under Hatch valley.

Tlbh **Limerock Canyon and Brian Head Formations, undivided** (lower Miocene and lower Oligocene to middle Eocene) – While detailed geochronologic dating should be able to differentiate the formations, we remain uncertain how to distinguish in the field apparently similar strata of the Brian Head and Limerock Canyon Formations at Mammoth Ridge and areas immediately to the south. Sable and Maldonado (1997b) also described the difficulty of differentiating similar volcaniclastic strata of the Limerock Canyon, Bear Valley, and Brian Head Formations.

Tl **Limerock Canyon Formation** (lower Miocene) – White, light-gray, and pale- to olive-green, tuffaceous, volcaniclastic sandstone, pebbly sandstone, gritstone, pebbly conglomerate, mudstone, and minor tuffaceous limestone. It is commonly bioturbated and includes at least 10 thin beds of ash-fall tuff. The clasts are about 90% volcanic but include as much as 10% quartzite and sandstone; Kurlich and Anderson (1997) stated that the formation lacks Needles Range, Isom, Bear Valley, and Mount Dutton clasts, but 26- to 27-Ma Isom clasts are abundant and many of the mafic volcanic clasts that they reported could be derived from the Mount Dutton Formation (where Isom clasts are recognized, Limerock and Brian Head strata are readily differentiated); the type section of the Limerock Canyon Formation (west of Hatch) contains a few tens of feet of strata that we reassign to the Brian Head Formation, and we suggest that the limestone that Kurlich and Anderson (1997) assigned to the Brian Head Formation at the base of this type section is in fact the upper white member of the Claron Formation as originally described by Kurlich (1990); query indicates uncertain identification on the south side of Hatch Mountain; unconformably overlain by unconsolidated upper Tertiary fan alluvium (Taf); as much as 290 feet (88 m) thick in a composite type section west of Hatch (Kurlich, 1990; Kurlich and Anderson, 1997).

The Limerock Canyon Formation was deposited in fluvial, floodplain, and minor lacustrine environments (Kurlich and Anderson, 1997). It is present only on the east part of the Markagunt Plateau near Hatch, south of the Markagunt gravity slide, where it appears to be eroded into the Brian Head Formation. We suggest that it may be preserved in a subtle basin in front of an inferred blind west-trending thrust fault, the inferred westward continuation of the Rubys Inn thrust fault. Two ash-fall tuff beds, about 100 feet (30 m) and 200 feet (60 m) above the base of the formation at the type section west of Hatch, respectively, yielded K-Ar ages of 21.5 ± 0.6 Ma (biotite) and 21.0 ± 1.0 Ma (sanidine), and of 20.2 ± 1.4 Ma (biotite) and 19.8 ± 0.8 Ma (sanidine) (Sable and Maldonado, 1997b); Sable and Maldonado (1997b) also reported $^{40}Ar/^{39}Ar$ ages of 20.48 ± 0.8 Ma (biotite) and 21.0 ± 1.0 Ma (sanidine), and we obtained a U-Pb age on zircon from an ash-fall tuff near the middle of the formation of 20.52 ± 0.49 Ma (UGS and AtoZ, 2013a).

Tmf **Flake Mountain megabreccia, undivided** (Miocene?) – Comprises a basal breccia and associated clastic dikes on the southwest side of Flake Mountain; interpreted to be a cataclastic breccia made up of Mount Dutton Formation and minor Brian Head Formation and derived from a gravity slide; age unknown, but Rowley and others (1987) suggested that the megabreccia was Pliocene or Miocene or possibly early Pleistocene; as much as several tens of feet thick.

Tmf(Td) **Flake Mountain megabreccia**, Mount Dutton Formation, alluvial facies, component – Largely intact block nearly 3 square miles (8 km^2) in size of brown volcanic mudflow breccia of mostly andesitic composition; interpreted to have slid southward from the Sevier Plateau; about 300 feet (90 m) thick.

Tmf(Tbh) **Flake Mountain megabreccia, Brian Head Formation component** – White and light-gray, tuffaceous, volcaniclastic sandstone and mudstone that rest out of sequence on the white member of the Claron Formation and so are interpreted to be the basal part of the Flake Mountain megabreccia; about 40 feet (12 m) thick.

Thm **Haycock Mountain Tuff** (lower Miocene) – Consists of two similar cooling units at its type section along Panguitch Creek about one mile (1.6 km) east of Panguitch Lake: lower unit is white to very light pink, unwelded, crystal-poor rhyolite tuff that is overlain by light-pink, unwelded, moderately resistant, crystal-poor rhyolite tuff; both contain common pumice fragments and lithic fragments of black, aphanitic, mafic volcanic rock; typically forms moderately resistant ledge over undeformed volcaniclastic conglomerate (Thma) and elsewhere overlies locally deformed Bear Valley Formation [Tm(Tbv)] and Mount Dutton Formation [Tm(Td)]; mapped north of Haycock Mountain where it is about 35 feet (11 m) thick.

The Haycock Mountain Tuff is petrographically and chemically similar to the Leach Canyon Formation, but lacks red lithic fragments that characterize the Leach Canyon Formation (see the Leach Canyon Formation unit description for details). As noted by Hatfield and others (2010), the Haycock Mountain Tuff yielded an $^{40}Ar/^{39}Ar$ age of 22.75 ± 0.12 Ma (Sable, unpublished data, 1996). Additionally, Fleck and others (1975) reported a K-Ar age (corrected according to Dalrymple, 1979) of 22.7 ± 0.6 Ma (plagioclase) for what he interpreted as Bauers Tuff Member on the Markagunt Plateau, east of Fivemile Ridge, but that we re-interpret as Haycock Mountain Tuff. On the basis of the undeformed nature of the Haycock Mountain Tuff and its underlying conglomerate (Thma), Anderson (1993), Rowley and others (1994a), and Hatfield and others (2010) interpreted the Haycock Mountain Tuff to postdate emplacement of the Markagunt gravity slide. However, the 22.75 Ma age of the undeformed Haycock Mountain Tuff is at odds with the fact that the Markagunt gravity slide overlies 22.03 Ma Harmony Hills Tuff. The Markagunt gravity slide must thus be younger than about 22 million years old. We resampled the Haycock Mountain Tuff and report a new U-Pb age on zircon of 21.63 ± 0.73 Ma (UGS and AtoZ, 2013a). The Haycock Mountain Tuff does indeed represent a post-Markagunt gravity slide ash-flow tuff that partly filled a stream channel and adjacent lowlands eroded into the gravity slide.

Thma **Alluvium underlying Haycock Mountain Tuff** (lower Miocene) – Moderately sorted, moderately consolidated, pebble to boulder volcaniclastic gravel underlying undisturbed Haycock Mountain Tuff at its type section about one mile (1.6 km) northeast of Panguitch Lake and also about 6 miles (9.5 km) farther northeast where it is exposed on the north side of Panguitch Creek; maximum exposed thickness about 80 feet (24 m).

Markagunt Megabreccia (lower Miocene) – With an apparent aerial extent of at least 1600 square miles (4160 km^2), larger than the state of Rhode Island, the Markagunt gravity slide is the largest known subaerial gravity slide on Earth; it surpasses in size the famous 1300-square-mile (3400 km^2) Heart Mountain detachment in northwestern Wyoming, long considered to be the largest known landslide on land (see, for example, Malone and Craddock, 2008; Beutner and Hauge, 2009; Craddock and others, 2009). At its simplest, the Markagunt gravity slide is a great sheet of volcanic rock that slid many miles southward and at its distal southern end overrode the Earth's surface to place older rock on younger rock. The ramp is where the gravity slide plane came to the surface. The slide mass, known as the Markagunt Megabreccia, consists of large blocks—commonly many square miles in extent—of Miocene and Oligocene regional ash-flow tuffs (most originally erupted from calderas along the Utah-Nevada border) and local volcanic and volcaniclastic rocks (derived from the Marysvale volcanic field), and blankets the entire central and northern Markagunt Plateau and adjacent areas. See the accompanying text and Biek and others (2014) and Hacker and others (2014) for a discussion of the Markagunt gravity slide, including the history of its discovery and still unresolved questions about its origin.

Tm **Markagunt Megabreccia, undivided** – The Markagunt gravity slide is undivided where exposures are insufficient to delineate bedrock units at the map scale and in more remote areas due to time constraints. Most areas mapped as Tm consist predominantly of the Isom Formation, but locally includes Wah Wah Springs and Brian Head strata, and, north of Panguitch Lake, large amounts of volcanic mudflow breccia, volcaniclastic sandstone, and minor pebbly conglomerate of the Bear Valley and Mount Dutton Formations; it locally includes conglomeratic strata of the pink member of the Claron Formation on the east flank of Jackrabbit Mountain. On the Markagunt Plateau, north of the latitude of Panguitch Lake, Tm was emplaced on the resistant, planar surface of the Isom Formation, but south of the lake it was emplaced on the Leach Canyon Formation. At the west edge of the plateau, south of Iron Peak, Tm was locally emplaced on Brian Head strata. The Markagunt gravity slide locally exceeds 2000 feet (600 m) thick.

Tm(Ta) **Markagunt Megabreccia, middle Tertiary alluvium component** – Volcaniclastic conglomerate and pebbly sandstone on Haycock Mountain and areas to the north, mostly in the Fivemile Ridge quadrangle; contains quartzite cobbles and small boulders in the basal part of the deposits; typically forms cobble-covered hillsides, but is locally well-consolidated in exposures on the southwest side of Haycock Mountain; mostly overlies the Isom Formation component [Tm(Ti)], but also underlies upper-plate Isom on the south side of Haycock Mountain; maximum thickness in this area is probably about 100 feet (30 m).

Anderson (1993) suggested that this alluvium postdated emplacement and tilting of the gravity slide and that remnant alluvium at Haycock Mountain, some as much as 800 feet (250 m) above Panguitch Creek, is simply what remained after exhumation of the ancestral Panguitch Creek drainage. However, our new mapping suggests that there are multiple intervals of Miocene gravels that both predate (Ta) and postdate (Thma) the gravity slide. Importantly, the gravels atop Haycock Mountain are tilted northward, concordantly with underlying Isom Formation, and we thus interpret them to have ridden largely undisturbed on the Isom during emplacement of the gravity slide. Thus we now envision these gravels as a relatively thin sheet that blanketed the landscape prior to emplacement of the Megabreccia, not as the exhumed remains of post-gravity slide basin fill.

Tm(Tdbv)

Markagunt Megabreccia, Mount Dutton and Bear Valley components, undivided – Mapped north of Panguitch Lake where exposures are inadequate to readily differentiate these formations; exceeds 500 feet (150 m) thick.

Tm(Tds)

Markagunt Megabreccia, Mount Dutton Formation, local sandstone facies component – Pale- to dark-gray and light-yellowish-brown, cross-bedded, slope-forming, zeolite-cemented tuffaceous sandstone; present at the northern edge of the map area, about 4 miles (6 km) northwest of Bear Valley Junction; as much as about 100 feet (30 m) thick.

Tm(Td)

Markagunt Megabreccia, Mount Dutton Formation, alluvial facies component – Brown or locally reddish-brown volcanic mudflow breccia of mostly andesitic composition, volcaniclastic pebble to boulder conglomerate, and minor tuffaceous sandstone; query indicates uncertain identification; at least 1000 feet (300 m) thick on the northern Markagunt Plateau.

Tm(Tdm)

Markagunt Megabreccia, Mount Dutton Formation, mafic alluvial facies component – Dark-gray, vesicular, basaltic andesite and basalt present as angular cobble- to boulder-size blocks floating in a light-gray sandy and muddy matrix of the same composition; monolithic; interpreted to be volcanic mudflow deposits; a small area of basaltic scoria is present north of Bunker Creek in the NE1/4 NE1/4 section 12, T. 36 S., R. 8 W., about 3 miles (5 km) west of Panguitch Lake; maximum thickness is about 80 feet (24 m).

Mapped between Panguitch Lake and Sidney Peaks where it overlies 23.8 Ma Leach Canyon Formation (Tql), and north of Panguitch Lake, where it overlies the Isom Formation (Tm[Ti]). In the Bunker Creek drainage west of Panguitch Lake, the map unit is overlain by Isom Formation or locally by Wah Wah Springs Formation, each as part of the Markagunt gravity slide.

Tm(Tbv)

Markagunt Megabreccia, Bear Valley Formation, undivided, component – Volcaniclastic sandstone and, especially in the upper part of the formation, interbedded lava flows, volcanic mudflow breccia, conglomerate, mudstone, and ash-fall and ash-flow tuff beds; sandstone is white to light gray, greenish gray, yellowish gray, olive gray, and locally vivid green, typically poorly indurated, moderately to well sorted, and fine to medium grained; sand is about 60% quartz, and the remainder is feldspar, biotite, hornblende, augite, and relict pumice replaced by zeolite; cement is mostly zeolite (clinoptilolite, commonly altered to chlorite) (Anderson, 1971); most sandstone is characterized by high-angle cross-beds indicative of eolian deposition; query indicates uncertain identification; Fleck and others (1975) reported two K-Ar ages (corrected according to Dalrymple, 1979) of 24.6 ± 0.4 Ma and 24.5 ± 0.5 Ma from the upper part of the formation in the Fivemile Pass quadrangle, west of Panguitch; zircons from our sample CM081612-3 yielded a mean formation age of 24.58 ± 1.92 Ma for rhyolitic tuff intruded by the Iron Peak laccolith; query indicates uncertain presence on the northwest flank of Iron Peak; the formation is in excess of 1000 feet (300 m) thick at its type section on the northern Markagunt Plateau (north of Highway 20 and about 1.5 miles [2.5 km] west of its junction with U.S. Highway 89; Anderson, 1971), but is typically 500 to 800 feet (150–200 m) thick (Anderson and Rowley, 1987).

Anderson and others (1987) and Sable and Maldonado (1997b) described the difficulty of differentiating similar lava flows and mudflow breccias of the Bear Valley and overlying Mount Dutton Formations. Kaufman and Anderson (1981) reported a possible vent for the volcanic rocks in the formation near Twin Peaks, about 7 miles (11 km) north of Panguitch Lake, but this area appears to be part of the Markagunt gravity slide and it is uncertain if some or all of the deformation observed there resulted from gravity-slide emplacement or is associated with a small vent area. Green sandstone was locally quarried from Bear Valley strata about 4 miles (6 km) southwest of Bear Valley Junction (Anderson and Rowley, 1987).

Tm(Tbvv)

Markagunt Megabreccia, Bear Valley Formation, lava flows and volcanic mudflow breccia component – Dark-gray andesitic(?) lava and volcanic mudflow breccia mapped east of Upper Bear Valley in the Little Creek Peak quadrangle; interpreted by Anderson and others (1987) to be the remnants of a small stratovolcano, a local vent complex of the Bear Valley Formation; about 400 feet (120 m) thick.

Tm(Tbvb)

Markagunt Megabreccia, Bear Valley Formation, volcanic mudflow breccia component – Pale-yellowish-brown breccia composed of pebble- to cobble-size clasts, mostly of intermediate-composition volcanic rocks and lesser amounts of tuff and tuffaceous sandstone; as much as about 800 feet (245 m) thick south of Cottonwood Mountain, about 7 miles (11 km) northeast of Paragonah.

Tm(Tbvf)

Markagunt Megabreccia, Bear Valley Formation, lava flow component – Dark-gray mafic lava flow present about one mile (1.6 km) southeast of Red Creek Reservoir; about 60 feet (18 m) thick.

Tm(Tdli)

Markagunt Megabreccia, Mount Dutton Formation, Leach Canyon Formation, and Isom Formation components, undivided – Poorly exposed, faulted outcrops of these units at Long Hill, in the northwest Red Hills, interpreted by Maldonado and Williams (1993a) as Miocene to Oligocene(?) gravity-slide blocks shed into basin-fill deposits of Long Hill, but here reinterpreted as part of the Markagunt Megabreccia mostly buried by basin fill; exposed thickness about 100 feet (30 m).

Tm(Tql)

Markagunt Megabreccia, Leach Canyon Formation component – Mapped near Winn Gap at the south end of the Red Hills where a fews tens of feet of Leach Canyon ash-flow tuff rests on allochthonous Isom Formation, and west of Black Knoll on the east flank of Jackrabbit Mountain, where it is about 100 feet (30 m) thick.

Tm(Tqcb)

Markagunt Megabreccia, Bauers Tuff Member of Condor Canyon Formation component – Mapped northwest of Navajo Ridge in the northwestern corner of the

Brian Head quadrangle (Rowley and others, 2013) and south of Jackrabbit Wash in the northwestern Red Hills (Maldonado and Williams, 1993a); exposed thickness about 100 feet (30 m).

Tm(Tbrp)

Markagunt Megabreccia, Volcanic Rocks of Bull Rush Peak component – Mapped northeast of Bear Valley Junction, where it is interbedded with volcanic mudflow deposits of the Mount Dutton Formation; collective thickness is at least 2500 feet (750 m).

Tm(Tbb)

Markagunt Megabreccia, Buckskin Breccia component – Lithic ash-flow tuff comprising at least four undifferentiated, moderately resistant cooling units, locally separated by thin, tuffaceous sandstone, each bed of which contains angular clasts of rock petrologically identical to volcanic rocks of Bull Rush Creek and similar to those of the Spry intrusion (Anderson and Rowley, 1975; Anderson and others, 1987, 1990a, 1990b); the Buckskin Breccia was first noted by Anderson (1965) to contain clasts similar to the Spry intrusion, which increase in size toward the intrusion and which Yannacci (1986) called clasts of juvenile magma (i.e., clasts of identical composition to the matrix, broken up during eruption of pyroclastic flows), all of which indicate that the Buckskin Breccia originated as an eruptive phase of the nearby Spry intrusion; type section is just east of Lower Bear Valley, north of Highway 20 in the extreme northeast corner of the Little Creek Peak quadrangle (Anderson and others, 1987); Buckskin Breccia exposed northwest of Bear Valley Junction is here interpreted to have first slid off the 26.2 Ma Spry intrusion and later transported again as part of the Markagunt gravity slide; more than 700 feet (210 m) thick in the Little Creek Peak quadrangle (Anderson and others, 1987), but only a few tens of feet thick west of Bear Valley Junction.

Tm(Ti) **Markagunt Megabreccia, Isom Formation component** – Medium-gray, crystal-poor, densely welded, trachydacitic ash-flow tuff; small (1–3 mm) euhedral crystals constitute 15 to 20% of the rock and are mostly plagioclase (90%) and minor pyroxene, magnetite, and rare quartz set in a devitrified glassy groundmass; unlike autochthonous Isom Formation, most outcrops and blocks weather to grussy soils and rounded hills; except locally along the south edge of Haycock Mountain, rarely forms cliffs as is typical of the autochthonous Isom; maximum thickness about 400 feet (120 m).

Although generally poorly exposed, the Isom Formation constitutes the great bulk of the gravity slide along its southern margin, roughly between Haycock Mountain, Brian Head, and Horse Valley, and in down-faulted blocks at the plateau's west margin. The map unit locally includes components of the Brian Head Formation, Wah Wah Springs Formation, and volcanic mudflow breccia that are difficult to delineate given extensive forest cover and inconspicuous outcrop habit. It generally slid over the Leach Canyon Formation west of Panguitch Lake, over Isom Formation north of Panguitch Lake, and over thin lower Miocene conglomerate that unconformably overlies the Brian Head Formation east of Panguitch Lake.

Tm(Tnw)

Markagunt Megabreccia, Wah Wah Springs Formation component – Pale-red, grayish-orange-pink, and pale-red-purple, crystal-rich, moderately welded, dacitic ash-flow tuff; phenocrysts of plagioclase, hornblende, biotite, and quartz (with minor Fe-Ti oxides and sanidine) comprise about 40% of the rock; locally exhibits intense cataclasis and shearing; about 40 feet (12 m) thick on the Markagunt Plateau and as much as about 100 feet (30 m) thick on Jackrabbit Mountain and northern Red Hills.

Tm(Tn) **Markagunt Megabreccia, Needles Range Group, undivided component** – Wah Wah Springs and Lund Formations mapped on the flanks of Jackrabbit Mountain where combined thickness is about 150 feet (45 m).

Tm($Tbhu_2$)

Markagunt Megabreccia, Brian Head Formation, upper volcanic unit, upper part component – Light-gray quartz latite porphyry with phenocrysts of plagioclase, hornblende, biotite, and uncommon quartz, sanidine, and pyroxene in a holocrystalline groundmass of quartz, plagioclase, and sanidine; all exposures in this map area exhibit autobrecciation; well exposed along Highway 20 about 1.5 miles (2.5 km) west of Bear Valley Junction; previously mapped as local volcanic rocks by Anderson and Rowley (1987), but reinterpreted here as autoclastic flow breccia of the Brian Head Formation; maximum exposed thickness in the map area is about 300 feet (90 m).

Tm(Tbhu$_1$)

Markagunt Megabreccia, Brian Head Formation, upper volcanic unit, lower part component – Volcanic mudflow breccia, lava flows, and lesser ash-flow tuff west of Bear Valley Junction at Bear Valley Junction dome (also known as Orton dome), where strata were called local volcanic and sedimentary strata by Anderson and Rowley (1987); may better be assigned as a distal, early phase of the Mount Dutton Formation, but is here lumped with the Brian Head Formation; about 700 feet (200 m) thick.

Tm(Tbh$_3$)

Markagunt Megabreccia, Brian Head Formation, upper part of middle volcaniclastic unit component – Similar to Tm(Tbh) described below, but contains more sandstone and lesser mudstone and few beds of chalcedony; mapped along the Sevier fault zone east of Panguitch; incomplete thickness is at least 300 feet (90 m).

Tm(Tbh$_2$)

Markagunt Megabreccia, Brian Head Formation, middle part of middle volcaniclastic unit component – Volcaniclastic fine-grained sandstone, siltstone, and mudstone; forms distinctive green, red, and gray band in lower part of the formation along the Sevier fault zone east of Panguitch; about 80 feet (25 m) thick.

Tm(Tbh$_1$)

Markagunt Megabreccia, Brian Head Formation, lower part of middle volcaniclastic unit component – Light-gray and white volcaniclastic mudstone, siltstone, and fine-grained sandstone mapped along the Sevier fault zone east of Panguitch; incomplete thickness is about 100 feet (30 m) thick.

Tm(Tbh)

Markagunt Megabreccia, Brian Head Formation component – Poorly exposed but distinctive white to light-gray volcaniclastic mudstone, pebbly sandstone, micritic limestone, and chalcedony clasts commonly present in colluvium, thus betraying the formation's presence northwest of Castle Valley (about 5 miles [8 km] west-southwest of Panguitch Lake), where it rests out-of-sequence on autochthonous Leach Canyon Formation; on the ridge at the common border of sections 9 and 16, T. 36 S., R. 8 W., pebbly volcaniclastic sandstone of the Brian Head Formation is well exposed at the head of a small landslide, dips 27° northeast, and is overlain by similarly dipping Wah Wah Springs Formation; on the hill to the south, however, Brian Head strata appear to be subhorizontal; thickness uncertain but outcrop patterns suggest that displaced Brian Head strata likely exceed 100 feet (30 m) thick.

Unit on the southeast side of Haycock Mountain is characterized by white and light-gray, locally tuffaceous, volcaniclastic sandstone, pebbly sandstone, mudstone, minor tuffaceous limestone, and local multi-hued chalcedony that are faulted and folded, indicative of deformation as part of the Markagunt gravity slide (alternately, deformation may be due to south-vergent thrust faulting associated with the westward extension of the Rubys Inn thrust); exposed thickness as much as 150 feet (45 m).

Tm(Tcp)

Markagunt Megabreccia, Claron Formation, pink member component – Brightly colored and mottled but mostly reddish-orange, pebble and cobble conglomerate, calcareous sandstone, and mudstone mapped on the northeast flank of Jackrabbit Mountain in the northwest part of the map area; clasts are rounded orthoquartzite and limestone, some of which exhibit intense shear deformation; probably at least 100 feet (30 m) thick.

Ta **Miocene alluvium (lower Miocene)** – Moderately sorted, moderately consolidated, pebble to boulder gravel mapped in three areas across the quadrangle: (1) on the south side of Haycock Mountain, (2) at the southeast end of Flake Mountain, and (3) at Miners Peak in the southwest corner of the map area. Deposits at Haycock Mountain contain rounded volcanic clasts and lesser quartzite and Claron limestone clasts; importantly, they also contain rounded clasts of Isom and possibly of Harmony Hills Tuff (though the latter may be derived from the Needles Range Group) and so are at least as young as 26–27 Ma and may be younger than 22 Ma; regardless, the Markagunt gravity slide overlies the map unit at Haycock Mountain; maximum exposed thickness about 80 feet (25 m).

Deposits at the southeast end of Flake Mountain are poorly exposed but consist of pebble to boulder gravel; contains well-rounded volcanic clasts, Claron Formation limestone clasts, and common recycled quartzite clasts; interpreted to be stream channel deposits eroded into the white member of the Claron Formation; structurally overlain by the Flake Mountain megabreccia; maximum exposed thickness about 40 feet (12 m).

Deposits at Miners Peak on Cedar Mountain are unconsolidated, poorly sorted, clay- to very large boulder-size sediment characterized by large quartz monzonite boulders. Quartz monzonite boulders as much as about 30 feet (10 m) in diameter constitute about 90% of the deposits. These deposits also contain large boulders of Claron Formation limestone as much as 18 feet (6 m) long, as well as recycled, rounded pebbles and small cobbles of Precambrian and Cambrian quartzite, lesser Cretaceous sandstone boulders, and rare cobbles and boulders of pebbly sandstone of uncertain origin. Except for the quartzite, most clasts at Miners Peak are subangular to subrounded. These deposits are about 90 feet (27 m) thick and were probably deposited by debris flows originating in the ancestral Pine Valley Mountains from erosion of the 20.5 Ma Pine Valley Latite or Pine Valley laccolith (see Biek and others [2009] for a discussion of the provenance and age of these unusual deposits).

Tip **Iron Peak laccolith** (lower Miocene) – Medium-gray gabbro-diorite porphyry composed almost entirely of augite and plagioclase (calcic labradorite) and about 8% opaque oxide minerals, mostly magnetite, with diorite the dominant phase (Anderson, 1965; Spurney, 1984); magnetite veins are present throughout the intrusion and are as much as 10 feet (3 m) in width, but most are less than one inch (2.5 cm) wide (Spurney, 1984); base of laccolith is well exposed in the north canyon wall of Little Creek, northeast of Paragonah, which has incised through the laccolith to reveal numerous feeder dikes; originally referred to as the Iron Point laccolith by Anderson (1965) and Anderson and Rowley (1975), as the namesake peak was then known (the peak was renamed and is now referred to as Iron Peak); yielded K-Ar whole-rock age (corrected according to Dalrymple, 1979) of 20.2 ± 0.5 Ma (Fleck and others, 1975); exposed thickness is as much as about 800 feet (240 m).

Iron Peak forms the easternmost laccolith of the Iron Axis, a northeast-trending belt of early Miocene calc-alkaline 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). Iron Peak is the second youngest and most mafic of the Iron Axis intrusions. Most of the other plutons of the Iron Axis are quartz monzonite porphyry and appear to be partly controlled by northeast-striking, southeast-verging Sevier-age thrust faults; they were emplaced at shallow depths, mostly within about 1.2 miles (2 km) of the surface (Mackin and others, 1976; Van Kooten, 1988; Hacker and others, 2002, 2007; Rowley and others, 2008). Like the other laccoliths in the belt, the Iron Peak laccolith probably formed rapidly following a two-stage emplacement process—injection of a sill immediately followed by inflation—at shallow crustal depth of less than 4000 feet (1.2 km) on the basis of stratigraphic reconstructions (Spurney, 1984; see also Hacker and others, 2002, 2007; Willis, 2002). Rapid inflation and doming of most laccoliths of the Iron Axis led to their partial unroofing by gravity sliding, immediately followed by volcanic eruptions (Mackin, 1960; Blank and Mackin, 1967; Hacker and others, 1996, 2002, 2007; Hacker, 1998; Willis, 2002; Rowley and others, 2008).

Spurney (1984) interpreted exposures immediately east of the Iron Peak laccolith as a peripheral breccia complex; Maldonado and others (2011) reinterpreted this area as older Bear Valley breccia. The correlation of these rocks remains uncertain; they may be lava flows derived from the Iron Peak intrusion. Spurney (1984) also described volcanic rocks of similar composition to the south, on the divide between Red Creek and Little Creek canyons, northeast of Paragonah, that we map as Markagunt gravity slide (Tm), which may include lava flows associated with the Iron Peak laccolith. Fleck and others (1975; sample R-27, corrected according to Dalrymple, 1979) reported a K-Ar age of 21.2 ± 0.5 Ma on a lava flow within volcanic mudflow breccias that was interpreted by Anderson and Rowley (1975, p. 28) and Rowley and others (1994a, p. 12) to be derived from eruption of the Iron Peak laccolith; however, coordinates reported by Fleck and others (1975) for the latter sample are incorrect—that sample location falls on Isom Formation on the west side of Ipson Creek about 3 miles (5 km) northwest of Panguitch Lake.

The Iron Peak laccolith was previously mapped as intruding the Brian Head Formation (Spurney, 1984; Maldonado and others, 2011), but our mapping suggests that it intruded both Brian Head volcaniclastic strata and Bear Valley ash-flow tuff that are juxtaposed across a pre-Iron Peak-intrusion graben now preserved at the west edge of the Markagunt Plateau, about 5 miles (8 km) northeast of Paragonah. More detailed mapping is needed to delineate the extent of the laccolith and its host strata. We are attempting to redate the laccolith, but suggest that it is younger than the Markagunt gravity slide.

Emplacement and doming of the Iron Peak laccolith was suggested as one possible cause of sliding of the Markagunt gravity slide (Maldonado, 1995; Maldonado and others, 1997; Sable and Maldonado, 1997a), although Anderson (1993, 2001) suggested that the intrusion seemed too small to have produced such a large gravity slide. However, because the laccolith is only exposed in a graben, we do not know its original extent, particularly how far west it may have once reached. On the basis of the presence of apparent feeder dikes (Tipd) in the Claron horst west of Iron Peak, we infer that the laccolith extends west into the subsurface of Parowan Valley and thus may be considerably larger than first envisioned. However, Iron Peak laccolith doming is not our preferred trigger for the Markagunt gravity slide as discussed earlier.

Spurney (1982, 1984) suggested that magnetite veins formed late in the laccolith's emplacement, a result of alteration of augite phenocrysts. While magnetite veins are common, they are apparently of insufficient number to have been of economic importance, unlike the nearby Iron Springs mining district west of Cedar City, 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, 2008).

Tipd **Feeder dikes of Iron Peak laccolith** (lower Miocene) – Mafic dikes exposed in the north canyon wall of Little Creek, immediately south of the Iron Peak laccolith; of the same composition as the adjacent laccolith, and so are interpreted to be its feeder dikes (Anderson, 1965; Spurney, 1984; Hacker and others, 2007); dikes intruded into ash-flow tuff of the Bear Valley Formation, and they are resistant and so stand as tall fins; most dikes trend northeast, dip moderately to steeply west, and most are about 6 feet (2 m) wide but range from about 0.8 to 25 feet (0.25–8 m) wide.

Timd **Mafic dikes at the west edge of the Markagunt Plateau** (lower Miocene) – Highly altered, greenish-gray to brownish-gray, aphanitic to fine-grained mafic dikes that trend mostly north in the Cottonwood Mountain quadrangle northeast of Paragonah; some dikes contain small plagioclase phenocrysts; typically deeply weathered and so poorly exposed, but most dikes fill joints and small-displacement faults, which are especially well developed in a horst of gently northwest-tilted Claron strata at the west edge of the plateau, west of Iron Peak (the bar and ball symbol on these dikes indicates relative sense of displacement of the fault); a sample from one of the northwest-trending dikes west of the Iron Peak laccolith yielded a K-Ar age of about 20 Ma (H.H. Mehnert and R.E. Anderson, written communication to F. Maldonado, 1988), consistent with an interpretation that the dikes are related to the laccolith; dikes range from about 1 to 20 feet (0.3–6 m) wide.

Map unit symbol Timd also used to denote basaltic and andesitic dikes in the adjacent Little Creek Peak quadrangle to the east, which Anderson and others (1987) interpreted as feeder dikes for the Mount Dutton and Bear Valley Formations and which Maldonado and others (1997, 2011) interpreted as related to the Iron Peak laccolith.

Tdbrp **Mount Dutton Formation and Volcanic Rocks of Bull Rush Peak, undivided** (upper Oligocene) – Mapped along Deer Creek northwest of Johns Valley.

Td **Mount Dutton Formation, alluvial facies** (upper Oligocene) – Light- to dark-gray and brown, andesitic to dacitic volcanic mudflow breccia and lesser interbedded volcaniclastic conglomerate and tuffaceous sandstone; also contains subordinate lava flows, flow breccia, and minor felsic tuff; Anderson and Rowley (1975) defined the Mount Dutton Formation as consisting of most of the rocks exposed on the south flank of the Marysvale volcanic field, and divided it into complexly interfingering and cross-cutting vent and alluvial facies derived from clustered stratovolcanoes and dikes; except for the Kingston Canyon Tuff Member, only those rocks associated with the alluvial facies are present in the map area, where they typically overlie the Bear Valley Formation on the Markagunt Plateau and the Brian Head Formation on the Sevier Plateau.

The Marysvale volcanic field is one of several voluminous calc-alkaline, subduction-related volcanic centers and underlying source batholiths that characterized the West from Oligocene to Miocene time at this latitude (Lipman and others, 1972; Rowley and Dixon, 2001). Fleck and others (1975) and Rowley and others (1994a) reported several K-Ar ages of 23 to 30 Ma on rocks of the coeval vent facies, not including an anomalously young age (corrected according to Dalrymple, 1979) of 21.2 ± 0.5 Ma (sample R-27 of Fleck and others, 1975) ascribed by Anderson and Rowley (1975, p. 28) and Rowley and others (1994a) to a local volcano, perhaps from the Iron Peak intrusion east of Paragonah. The alluvial facies is at least 1000 feet (300 m) thick in the map area in the northern Markagunt Plateau (Anderson and Rowley, 1987) where we re-interpret it to be part of the Markagunt gravity slide, about 2000 feet (600 m) thick on the southern end of the Sevier Plateau (Rowley and others, 1987), and is at least 6000 feet (2000 m) thick in the Sevier Plateau north of the Panguitch 30′ x 60′ quadrangle (Anderson and others, 1990a, 1990b; Rowley and others, 2005). Inidvidual mudflows and other rock units pinch out radially from an east-trending string of stratovolcanoes along the southern part of the Marysvale volcanic field.

Tdk **Kingston Canyon Tuff Member** (upper Oligocene) – Reddish-brown and pink densely welded trachytic ash-flow tuff characterized by a thin basal black vitrophyre, local steeply dipping flow foliation, and light-gray “lenticules,” which are interpreted to be gas bubbles drawn out as much as 3 feet (1 m) in the plane of bedding yet only an inch (2.5 cm) or less thick; forms a single cooling unit with pumice lenticules and sparse to moderately abundant volcanic rock fragments; phenocrysts are plagioclase and minor pyroxene, Fe-Ti oxides, and biotite; basal vitrophyre as much as 6 feet (2 m) thick is commonly exposed; present in the southern Sevier Plateau where it pinches out against paleotopography near the base of the Mount Dutton Formation (Anderson and Rowley, 1975; Rowley and others, 1987, 1994a); queried on the west flank of the Sevier Plateau at the head of Limekiln Creek where it is about 80 feet (25 m) thick; Fleck and others (1975) determined a K-Ar age on biotite from the unit of 25.1 ± 0.4 Ma, as discussed by Anderson and Rowley (1975); typically about 30 feet (10 m) thick.

Tbv **Bear Valley Formation, undivided** (upper Oligocene) – Poorly indurated, moderately to well-sorted, fine- to medium-grained volcaniclastic sandstone; sandstone is white to light gray, greenish gray, yellowish gray, olive gray, and locally vivid green; sand is about 60% quartz, and the remainder is feldspar, biotite, hornblende, augite, and relict pumice replaced by zeolite; cement is mostly zeolite (clinoptilolite, commonly altered to chlorite) (Anderson, 1971); most sandstone is characterized by high-angle cross-beds indicative of eolian deposition; Fleck and others (1975) reported two K-Ar ages (corrected according to Dalrymple, 1979) of 24.6 ± 0.4 Ma and 24.5 ± 0.5 Ma from the upper part of the formation in the Fivemile Pass quadrangle, west of Panguitch; zircons from our sample CM081612-3 yielded a mean formation age of 24.58 ± 1.92 Ma for rhyolitic tuff intruded by the Iron Peak laccolith; the formation is in excess of 1000 feet (300 m) thick at its type section on the northern Markagunt Plateau (north of Highway 20 and about 1.5 miles [2.5 km] west of its junction with U.S. Highway 89; Anderson, 1971), but is typically 500 to 800 feet (150–200 m) thick (Anderson and Rowley, 1987); where not part of the Markagunt gravity slide, only present at Jackrabbit Mountain in the northwest corner of the map area where it forms a southward-thinning wedge as much as about 60 feet (18 m) thick below the Leach Canyon Formation.

Quichapa Group (lower Miocene to upper Oligocene) – Consists of three regionally distinctive ash-flow tuffs: in ascending order, the Leach Canyon Formation, Condor Canyon Formation, and Harmony Hills Tuff (Mackin, 1960; Williams, 1967; Anderson and Rowley, 1975; Rowley and others, 1995; Best and others, 2013). The Leach Canyon Formation likely erupted from the Caliente caldera complex (Williams, 1967), the two-member Condor Canyon Formation clearly erupted at least in part from the western (Clover Creek caldera) part of the Caliente caldera complex (Rowley and others, 1995), and the Harmony Hills Tuff likely erupted from the eastern Bull Valley Mountains (Rowley and others, 1995).

Tqh **Harmony Hills Tuff** (lower Miocene) – Resistant, pale-pink to grayish-orange-pink, crystal-rich, moderately welded, dacitic ash-flow tuff; contains about 50% phenocrysts of plagioclase (63%), biotite (16%), hornblende (9%), quartz (7%), pyroxene (5%), and sanidine (trace) (Williams, 1967); yielded an $^{40}Ar/^{39}Ar$ plateau age of 22.03 ± 0.15 Ma (Cornell and others, 2001); as much as 50 feet (15 m) thick.

Exposed in Parowan Canyon where it overlies the Bauers Tuff Member of the Condor Canyon Formation and is reinterpreted by senior-author Biek to structurally underlie the Isom Formation that is part of the Markagunt gravity slide; also preserved at the base of the gravity slide in Summit Creek canyon. The source of the Harmony Hills Tuff is unknown, but isopachs are centered on Bull Valley (Williams, 1967), suggesting that it was derived from the eastern 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, 2008). Consistent with this interpretation is the fact that its age is nearly identical to that of those intrusions.

Tqcb **Bauers Tuff Member of Condor Canyon Formation** (lower Miocene) – Resistant, light-brownish-gray to pinkish-gray, densely welded, rhyolitic ash-flow tuff; contains about 10 to 20% phenocrysts of plagioclase (40–70%), sanidine (25–50%), biotite (2–10%), Fe-Ti oxides (1–8%), and pyroxene (<3%), but lacks quartz phenocrysts (Rowley and others, 1995); bronze-colored biotite and light-gray flattened lenticules are conspicuous in the upper, vapor-phase part of the tuff, and a basal vitrophyre 10 to 20 feet (3–6 m) thick is normally present; typically about 50 to 100 feet (15–30 m) thick.

The Bauers Tuff Member is exposed in Parowan Canyon where it is as much as 50 feet (15 m) thick and overlies volcaniclastic sandstone and mudflow breccia of the Mount Dutton Formation. It overlies the Leach Canyon Formation in Summit Creek canyon. It is also exposed in a fault block in the northwestern Red Hills, where it is as much as about 100 feet (30 m) thick (Maldonado and Williams, 1993a).

The Bauers Tuff Member erupted from the northwest part (Clover Creek caldera) of the Caliente caldera complex and covered an area of at least 8900 square miles (23,000 km^2) (Best and others, 1989b; Rowley and others, 1995) with an estimated volume of 740 cubic miles (3200 km^3) (Best and others, 2013). The prefered $^{40}Ar/^{39}Ar$ age of the Bauers Tuff Member is 22.7 Ma (Best and others, 1989a) or 22.8 Ma (Rowley and others, 1995), which is also the $^{40}Ar/^{39}Ar$ age of its intracaldera intrusion exposed just north of Caliente, Nevada (Rowley and others, 1994b). Fleck and others (1975) reported a K-Ar age (corrected according to Dalrymple, 1979) of 22.7 ± 0.6 Ma (plagioclase) for what they interpreted as Bauers Tuff Member on the Markagunt Plateau, east of Fivemile Ridge, but that we re-interpret as Haycock Mountain Tuff.

Tql **Leach Canyon Formation** (upper Oligocene) – Grayish-orange-pink to pinkish-gray, poorly to moderately welded, crystal-rich to crystal-poor rhyolite tuff that contains abundant white or light-pink collapsed pumice fragments and several percent lithic clasts, many of which are reddish brown; contains as much as 25 to 35% total phenocrysts of plagioclase, slightly less but subequal amounts of quartz and sanidine, and minor biotite, hornblende, Fe-Ti oxides, and a trace of pyroxene; forms the resistant cap rock of Brian Head peak and the southern part of Black Ledge, is exposed eastward nearly to the Panguitch Lake area, and is also exposed in fault blocks at the west edge of the Markagunt Plateau and in the northern Red Hills, as described below; source is unknown, but it is probably the Caliente caldera complex because isopachs show that it thickens toward the complex (Williams, 1967; Rowley and others, 1995); the total volume of the Leach Canyon is estimated to be 830 cubic miles (3600 km^3), representing the largest and apparently the initial eruption of the Caliente caldera complex (Best and others, 2013); typically about 100 feet (30 m) thick in the map area.

At Brian Head peak, the Leach Canyon Formation, which unconformably overlies the Isom Formation, consists of four parts, the lower three of which are rarely exposed elsewhere. At the base is non-resistant, 6- to 10-foot-thick (2–3 m), unwelded, white, rhyolite tuff that is overlain by a 10-foot-thick (3 m) moderate-orange-pink rhyolite tuff that has sparse reddish-brown lithic clasts, which becomes slightly more indurated in the upper part of the unit. This is overlain by a massive, 12-foot-thick (4 m) black vitrophyre, which is in turn overlain by a 25-foot-thick (8 m) resistant, pale-red, moderately welded rhyolite tuff that contains pale-lavender flattened pumice lenticules and as much as 1% distinctive, small, reddish-brown lithic clasts of flow rock torn from the vent walls. This resistant upper unit forms the cap rock of Black Ledge northward to beyond the Sidney Peaks area.

To the east, west of Panguitch Lake, the Leach Canyon Formation unconformably overlies the Brian Head Formation or, locally, stream gravel too thin to map that contains clasts of Isom Formation (for example, on the southeast side of Prince Mountain at sample location PL061708-3). A non-resistant, moderate-orange-pink ash-fall tuff identical to that at Brian Head peak is present at the base of the unit on the southeast side of Prince Mountain. The main part of the cooling unit contains only rare, small, reddish-brown lithic fragments.

The Leach Canyon Formation and the Haycock Mountain Tuff are petrographically and chemically similar, which led Sable and Maldonado (1997a) to suggest that the latter is a distal facies of Leach Canyon. While it is true that the two formations are not reliably distinguishable on the basis of their major- and trace-element chemistry, the Haycock Mountain Tuff is typically less welded than the Leach Canyon and contains conspicuous black lithic fragments, unlike the reddish-brown lithic fragments of the Leach Canyon, facts previously noted by Anderson (1993), Rowley and others (1994a), and Hatfield and others (2010). Detailed mapping of the Panguitch Lake and Haycock Mountain quadrangles (Biek and others, 2014a, 2014b) reconfirms that these are indeed two different units. The Leach Canyon Formation can be traced in continuous outcrop from Brian Head peak northward to the head of Bunker Creek and then east to the east end of Prince Mountain just west of Panguitch Lake. It is structurally overlain by the Markagunt gravity slide, which here consists mostly of the Isom Formation. Samples from the south side of Prince Mountain yielded K-Ar ages of 22.8 ± 1.1 Ma (biotite) and 24.8 ± 1.0 Ma (sanidine) (Rowley and others, 1994a, sample 89USa-1a, which they mistakenly called Haycock Mountain Tuff) and a duplicate K-Ar age of 24.3 ± 1.0 Ma (sanidine) as well as an $^{40}Ar/^{39}Ar$ age of 23.86 ± 0.26 Ma (biotite) (Sable and Maldonado, 1997a, on the same sample 89USa-1a). The Leach Canyon Formation is widely agreed to be about 23.8 Ma (Best and others, 1993; Rowley and others, 1995). As noted by Hatfield and others (2010), both Rowley and others (1994a) and Sable and Maldonado (1997a) misinterpreted this tuff to be the Haycock Mountain Tuff, which yielded a slightly younger $^{40}Ar/^{39}Ar$ age of 22.75 ± 0.12 Ma (sanidine) at its type section about one mile (1.6 km) northeast of Panguitch Lake (Sable, unpublished data, 1996; see also Hatfield and others, 2010) and our new U-Pb age on zircon of 21.63 ± 0.73 Ma. The facts that the tuff at Prince Mountain yielded an age analytically indistinguishable from the Leach Canyon Formation, and that it can be traced continuously to classic Leach Canyon outcrops at Brian Head peak, are irrefutable evidence that it is the Leach Canyon Formation and not the Haycock Mountain Tuff.

The Leach Canyon Formation unconformably overlies the Isom Formation at Brian Head peak and the southern part of Black Ledge. North of Castle Valley and at Prince Mountain, however, the Leach Canyon unconformably overlies Brian Head strata. This distribution suggests that the Prince Mountain–Castle Valley area was a paleohigh of Brian Head strata during Isom time, and that, once the resistant Isom was in place in flanking areas, this paleohigh was preferentially eroded to form a broad, east-trending stream valley in which the Leach Canyon accumulated. Except for outcrops in Bear Valley north of Utah Highway 20, the Leach Canyon is not present on the Markagunt Plateau north of Clear Creek, a northwest-trending tributary to Panguitch Lake.

Tbrp **Volcanic rocks of Bull Rush Creek** (upper Oligocene) – Light-brown, light- to dark-gray, pink, and light-greenish-yellow, complexly interbedded, volcanic mudflow breccia, conglomerate, sandstone, lava flows, poorly welded ash-flow tuff with abundant lithic clasts, and autoclastic flow breccias; mineralogically and chemically similar to the Spry intrusion and Buckskin Breccia, and so probably represents a volcano complex on the roof of the intrusion, which is a laccolith (Anderson and Rowley, 1975; Rowley and others, 2005); mapped along Deer Creek northwest of Johns Valley where it consists of autoclastic flow breccia, locally with a basal vitrophyre, whose clasts and matrix are identical to those of the Spry intrusion; also mapped northeast of Bear Valley Junction, where it is interbedded with volcanic mudflow deposits of the Mount Dutton Formation and interpreted to be part of the Markagunt Megabreccia; yielded a K-Ar age of 26.4 ± 3.2 Ma (Anderson and others, 1990a), consistent with that of the Spry intrusion (Rowley and others, 2005); Anderson and others (1990a) reported an incomplete section is about 800 feet (250 m) thick in the adjacent Circleville Canyon area.

Ti **Isom Formation** (upper Oligocene) – Medium-gray, crystal-poor, densely welded, trachydacitic ash-flow tuff, typically having distinctive rheomorphic features including flow folds, elongated vesicles, and flow breccias and thus commonly known as a tufflava (Mackin, 1960; Cook, 1965, 1966; Anderson and Rowley, 1975, 2002); small (1–3 mm) euhedral crystals constitute 10 to 15% or less of the rock and are mostly plagioclase (90%) and minor pyroxene and Fe-Ti oxides set in a devitrified-glass groundmass; exhibits pronounced platy outcrop habit and is thus accompanied by extensive talus deposits; rarely, a black basal vitrophyre is exposed, and locally fracture surfaces and elongated vesicles (lenticules, described below) are dark reddish brown to dusky red; query indicates uncertain correlation in the upper reaches of the Clear Creek drainage northwest of Panguitch Lake; the Isom Formation is 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), and because it is locally interbedded with the 26 Ma Buckskin Breccia (Anderson and others, 1987); maximum exposed thickness is about 350 feet (110 m) at Black Ledge and about 250 feet (75 m) along Ipson Creek.

The best and most extensive exposures of the Isom Formation in the map area are at Brian Head peak and to the northeast along Black Ledge where at least three cooling units are locally present. At Brian Head peak, the lower part of the formation is classic tufflava about 80 feet (24 m) thick, whereas the upper part is a flow breccia 60 to 90 feet (18–27 m) thick. Along Black Ledge, about 7 miles (11 km) northeast of Brian Head peak, the flow breccia is absent and the Isom there appears to consist of a single cooling unit about 350 feet (100 m) thick. The Isom also forms prominent cliffs north of Clear Creek and Panguitch Lake. East of Yankee Meadows Reservoir, the base of the Isom is unusual in that Isom ash-flow tuff overlies as much as 30 feet (9 m) of coarse sandstone with angular Isom pebbles and planar and low-angle cross-stratification that in turn overlies calcite-cemented Isom breccia.

Regionally, many outcrops of all cooling units in the Isom Formation reveal secondary flow characteristics, including flow breccias, contorted flow layering, and linear vesicles such that the unit was considered a lava flow until Mackin (1960) mapped its widespread distribution (300 cubic miles [1300 km^3] today spread over an area of 9500 square miles [25,000 km^2] [Best and others, 1989a]) and found evidence of glass shards in its basal vitrophyre, thus showing its true ash-flow tuff nature. Such an ash-flow tuff is also called a tufflava or rheomorphic ignimbrite, one sufficiently hot to move with laminar flow as a coherent ductile mass—see, for example, Cook (1966), Anderson and Rowley (1975, 2002), Ekren and others (1984), Andrews and Branney (2005), and Geissman and others (2010). Exhibits pronounced subhorizontal lamination or platiness, which Mackin (1960) called “lenticules.” Fryman (1986, 1987), Anderson and others (1990c), and Anderson and Rowley (2002) also described the light-gray, pancake-shaped lenticules, which are typically spaced 4 to 8 inches (10–20 cm) apart and may extend for 30 feet (10 m) or more, and which are locally contorted, suggesting turbulence in the flow as it moved over uneven topography. Fryman (1986, 1987) also described fumaroles in the Isom of the northern Markagunt Plateau, a result of degassing of the flow as it came to rest.

The source of the Isom is unknown, but isopach maps and pumice distribution suggest that it was derived from late-stage eruptions of the 27–32 Ma Indian Peak caldera complex that straddles the Utah-Nevada border, possibly in an area now concealed by the western Escalante Desert (Rowley and others, 1979; Best and others, 1989a, 1989b, 2013). Estimated crystallization temperature and pressure of Isom phenocrysts are 950°C and <2 kbar (Best and others, 1993), and this relatively high temperature is supported by its degree of welding and secondary flow features. At its type area in the Iron Springs district west of the map area, Mackin (1960) defined three members, a lower unnamed member, the Baldhills Tuff Member, and the upper Hole-in-the-Wall Tuff Member; Rowley and others (1975) redefined the Baldhills Tuff Member to include Mackin’s lower unnamed member, and noted that the Baldhills consists of at least six cooling units. Maldonado and Williams (1993a, 1993b) described nine apparent cooling units in the northern Red Hills at the west edge of the map area. In the northern Markagunt Plateau, Anderson and Rowley (1975) defined the Blue Meadows Tuff Member, which underlies the Baldhills Tuff Member, but it is possible that the Blue Meadows Tuff is part of the Mount Dutton Formation, and thus a local tuff derived from the Marysvale volcanic field (Rowley and others, 1994a). Maldonado and others (2011) mapped the Blue Meadows Tuff Member of the Isom Formation at the east edge of the Cottonwood Mountain quadrangle. For this map, we simply call this ash-flow tuff the Isom Formation, noting our uncertainty in dividing the Isom on the Markagunt Plateau.

Tbt **Three Creeks Tuff Member of the Bullion Canyon Volcanics** (upper Oligocene) – Moderately welded, light-gray, light-brown, and pink, crystal-rich dacitic ash-flow tuff; contains nearly 50% phenocrysts of plagioclase, subordinate hornblende and biotite, and minor quartz, Fe-Ti oxides, and sanidine, and moderately abundant pumice and sparse volcanic rock fragments; lithologically similar to the Wah Wah Springs Formation, but contains more and significantly larger plagioclase (as long as 5 mm) and other phenocrysts; typically autobrecciated in the map area, likely due to late-stage flowage near the distal southern margin of the tuff (Rowley, 1968); yielded K-Ar age of about 27 Ma (Steven and others, 1979); maximum thickness is about 90 feet (30 m) in the map area.

Erupted from the Three Creeks caldera in the southern Pahvant Range (Steven, 1981), forming the largest ash-flow tuff in the Marysvale volcanic field. The estimated volume of the tuff is about 50 cubic miles (200 km^3), but this is a minimum because it is commonly covered near the base of the Bullion Canyon and Mount Dutton volcanic sequences (Cunningham and others, 2007).

Tn **Needles Range Group, undivided** (upper to lower Oligocene) – Lund Formation (27.9 Ma) unconformably overlying Wah Wah Springs Formation (30 Ma), locally undivided in the Red Hills due to map scale and structural complexity.

Tnl **Lund Formation** (upper Oligocene) – Grayish-orange-pink, moderately welded, crystal-rich, dacitic ash-flow tuff exposed in the Red Hills; similar to underlying Wah Wah Springs Formation, but with generally larger phenocrysts and a lighter-colored matrix; locally contains spheroidal masses of pumice as large as 1 foot (0.3 m) in diameter near the top of the unit; base of the formation includes about 12 feet (4 m) of pale-greenish-yellow tuffaceous sandstone and lesser pebbly volcaniclastic conglomerate; exhibits normal magnetic polarity (Best and Grant, 1987); derived from the White Rock caldera, the southwest part of the older Indian Peak caldera, and is of similar volume to the underlying Wah Wah Springs Formation (Best and Grant, 1987; Best and others, 1989a, 1989b, 2013); preferred age is 27.9 Ma (Best and others, 1989a); as much as about 200 feet (60 m) thick (Maldonado and Williams, 1993a).

Tnw **Wah Wah Springs Formation** (lower Oligocene) – Pale-red to grayish-orange-pink, moderately welded, crystal-rich, dacitic ash-flow tuff that rests on Brian Head strata and, on the Markagunt Plateau, is overlain by the Isom Formation; phenocrysts of plagioclase, hornblende, and biotite (plus minor quartz, Fe-Ti oxides, and sanidine) constitute about 40% of the rock; the abundance of hornblende over biotite is unique among Great Basin ash-flow tuffs; elongate collapsed pumice is common; exposed west of Cottonwood Mountain and west of Bear Valley in the northern Markagunt Plateau, at the head of Bunker Creek west of Panguitch Lake, and in the Red Hills; exhibits reversed magnetic polarity (Best and Grant, 1987); derived from the Indian Peak caldera of the 27 to 32 Ma Indian Peak caldera complex that straddles the Utah-Nevada border (Best and others, 1989a, 1989b, 2013); today, the Wah Wah Springs covers at least 8500 square miles (22,000 km^2) with an estimated volume of as much as about 720 cubic miles (3000 km^3) (Best and others, 1989a); about 30 Ma on the basis of many K-Ar and $^{40}Ar/^{39}Ar$ age determinations (Best and Grant, 1987; Best and others, 1989a, 1989b; Rowley and others, 1994a); about 40 feet (12 m) thick in the northern Markagunt Plateau and as much as 130 feet (40 m) thick in the Sevier Plateau; map patterns show that it is about 200 feet (60 m) thick in the Red Hills, about half the thickness reported by Maldonado and Williams (1993a, 1993b).

A small exposure on Lowder Creek (east of Brian Head peak) is deeply weathered, non-resistant, white, crystal-rich ash-flow tuff about 6 feet (2 m) thick. Phenocrysts of plagioclase, hornblende, biotite, and quartz (plus minor Fe-Ti oxides and sanidine) make up about 30 to 40% of the rock. The color and degree of welding contrast sharply with typical Wah Wah Springs, which led Rowley and others (2013) to suggest that the tuff at Lowder Creek was deposited in a lake. Biotite-hornblende pairs from two samples from the same bed at Lowder Creek yielded K-Ar ages of 29.1 to 32.4 Ma (Rowley and others, 1994a). The Lowder Creek exposure is overlain by 3 to 6 feet (1–2 m) of volcanic mudflow breccia, which is in turn overlain by 10 to 15 feet (3–5 m) of deeply weathered, non-resistant, crystal-poor ash-flow tuff(?) of uncertain provenance, which is itself overlain by autochthonous Isom Formation.

Brian Head Formation (lower Oligocene to middle Eocene) – The Brian Head Formation is the oldest widespread Tertiary volcaniclastic unit in the region, and like the underlying Claron Formation, it too suffers from a long and complicated nomenclatural history (Sable and Maldonado, 1997b). On the Markagunt Plateau, it disconformably overlies the uppermost mudstone, siltstone, and sandstone interval (Tcwt) of the white member of the Claron Formation (in the northwestern Markagunt Plateau, where the white member appears to be missing, Brian Head strata overlie the pink member of the Claron Formation). On the Sevier Plateau, Brian Head strata, which locally include a basal variegated unit not present to the west, disconformably overlie the conglomerate at Boat Mesa. Sable and Maldonado (1997b) designated a type section at Brian Head peak and divided the Brian Head Formation into three informal units, ascending: (1) nontuffaceous sandstone and conglomerate, (2) a volcaniclastic unit that has minor but conspicuous limestone and chalcedony, and (3) a volcanic unit, locally present in the northern Markagunt Plateau but not at the type section, characterized by volcanic mudflow breccia, mafic lava flows, volcaniclastic sandstone and conglomerate, and ash-flow tuff. We include the basal nontuffaceous sandstone and conglomerate as a new uppermost part of the Claron Formation (Tcwt), thus further restricting the Brian Head Formation to a widespread volcaniclastic unit (Tbh) and a local volcanic unit ($Tbhu_1$, $Tbhu_2$) present only in the northern Markagunt Plateau. On the Sevier Plateau, we divide the Brian Head Formation into four informal units (in ascending order, Tbhv, Tbh_1, Tbh_2, and Tbh_3).

On the Markagunt Plateau, Brian Head strata are unconformably overlain by the 30 Ma Wah Wah Springs Formation (Tnw), or locally by the 26 to 27 Ma Isom Formation (Ti) or the 23.8 Ma Leach Canyon Formation, whereas on the Sevier Plateau, Brian Head strata are typically overlain by the Mount Dutton Formation. Maldonado and Moore (1995) reported $^{40}Ar/^{39}Ar$ ages of 33.00 ± 0.13 Ma (plagioclase) and 33.70 ± 0.14 Ma (biotite) on an ash-flow tuff in the northern Red Hills that lies near the top of the formation. Davis and others (2009) reported U-Pb (SHRIMP-RG) ages of 35.2 ± 0.8 Ma and 34.7 ± 0.6 Ma from the Brian Head Formation at Brian Head peak. We obtained U-Pb ages on zircon from ash-fall tuffs at the base of the formation at Cedar Breaks National Monument of 35.77 ± 0.28 Ma, from about 80 feet (25 m) above the base of the formation on the southwest flank of the Sevier Plateau of 36.51 ± 1.69 Ma, and from the upper part of the formation near Haycock Mountain of 34.95 ± 0.83 Ma and 33.55 ± 0.80 Ma (Table 5); the two Haycock Mountain samples also yielded $^{40}Ar/^{39}Ar$ ages on sanidine of 35.04 ± 0.23 Ma and 33.80 ± 0.05 Ma (UGS and NIGL, 2012). Fleck and others (1975) reported a K-Ar age of 31.0 ± 0.5 Ma on a crystal-poor ash-flow tuff at the base of the volcanic section from just north of Showalter Mountain immediately north of the map area in the Burnt Peak 7.5′ quadrangle (Anderson and Rowley, 1975, p. 12–13). Anderson and others (1990b) mapped this section as local volcanic and tuffaceous sedimentary rock that lies below the Wah Wah Springs Formation, but it may better be lumped as an early, distal phase of the Mount Dutton Formation. Eaton and others (1999b) and Korth and Eaton (2004) reported on Duchesnean (middle Eocene) vertebrate fossils in the variegated unit, here assigned to the basal Brian Head Formation. Feist and others (1997) described sparse charophytes from the formation in Casto Canyon on the southwestern Sevier Plateau that indicate a maximum age of middle Eocene. The Brian Head Formation is thus early Oligocene to latest middle Eocene. Golder and Wizevich (2009) and Golder and others (2009) described trace fossils, including possible crayfish burrows and root traces, in the Brian Head Formation on the Sevier Plateau.

$Tbhu_1$ **Brian Head Formation, upper volcanic unit, lower part** – Volcanic mudflow breccias, lava flows, and lesser ash-flow tuff northwest of Adams Head on the southwest Sevier Plateau; may better be assigned as a distal, early phase of the Mount Dutton Formation, but is here lumped with the Brian Head Formation; about 250 feet (75 m) thick.

Tbht **Rhyolitic tuff of middle volcaniclastic unit** – Pinkish-brown, unwelded rhyolite ash-flow tuff in the upper part of the formation in the northern Red Hills, on the west flank of Jackrabbit Mountain; yielded K-Ar ages of 34.2 ± 2.1 Ma (plagioclase) and 36.3 ± 1.3 Ma (biotite), and $^{40}Ar/^{39}Ar$ ages of 33.00 ± 0.13 Ma (plagioclase) and 33.70 ± 0.14 Ma (biotite) (Maldonado, 1995); as much as about 200 feet (60 m) thick.

Tbh **Middle volcaniclastic unit, undivided** – White to light-gray volcaniclastic mudstone, siltstone, silty sandstone, sandstone, conglomerate, volcanic ash, micritic limestone, and multi-hued chalcedony; near Mineral Canyon and northwest of Little Salt Lake (hill 7292), conglomerate consists of pebble- to boulder-size, rounded clasts of intermediate volcanic rocks of unknown affinity and quartzite pebbles and cobbles; sandstone is commonly bioturbated with pencil-size root or burrow casts that weather out in relief; soft-sediment slump features are locally common; chalcedony is various shades of white, gray, yellow, red, black, and brown, typically has a white weathering rind, is commonly highly brecciated and resilicified, typically occurs in beds 1 to 3 feet (0.3–1 m) thick but locally as much as 8 feet (2.5 m) thick, and is locally stained by manganese oxides; chalcedony beds probably reflect silicification of limestone beds (Maldonado, 1995; Sable and Maldonado, 1997b; Schinkel, 2012) as silica is leached from glass shards in the volcanic ash beds; Bakewell (2001) suggested that the chalcedony may have resulted from silicification of the tuff beds themselves; Maldonado (1995) suggested a hydrothermal source for the generally low metal values in the chalcedony; chalcedony is almost always highly fractured, but some is useful for lapidary purposes (Strong, 1984); Brian Head strata are about 500 feet (150 m) thick at Brian Head peak; on the southwest flank of the Sevier Plateau, they are readily divisible into four informal parts that collectively total about 1000 feet (300 m) thick, described below.

The formation is typically non-resistant, poorly exposed, and extensively covered by colluvium, but locally well exposed near Panguitch Lake and on the southwest side of Brian Head peak. Because of abundant bentonitic clay derived from weathered volcanic ash, this unit weathers to strongly swelling soils (unlike the underlying Claron Formation) and forms large landslide complexes; it was the principal detachment surface for the Markagunt gravity slide and for the Red Hills shear zone of Maldonado (1992, 1995) and Maldonado and others (1994, 1997). The Brian Head Formation was deposited in low-relief fluvial, floodplain, and lacustrine environments in which large amounts of volcanic ash accumulated (Sable and Maldonado, 1997b).

Tbh_3 **Upper part of middle volcaniclastic unit** – Similar to the middle volcaniclastic unit described above, but contains more sandstone and lesser mudstone and few beds of chalcedony; contains several tens of feet of apparently non-volcaniclastic Claron-like red beds near the middle of the unit on the southeast flank of the Sevier Plateau; about 700 feet (210 m) thick.

Tbh_2 **Middle part of middle volcaniclastic unit** – Volcaniclastic fine-grained sandstone, siltstone, and mudstone; forms distinctive green, red, and gray band in lower part of the formation on the Paunsaugunt Plateau; about 80 feet (25 m) thick.

Tbh_1 **Lower part of middle volcaniclastic unit** – Light-gray and white volcaniclastic mudstone, siltstone, and fine-grained sandstone; basal part includes a 25-foot-thick (8 m) ash-fall tuff that weathers to bluish-gray swelling soils and which yielded a U-Pb age on zircon of 36.51 ± 1.69 Ma; about 160 feet (50 m) thick.

Tbhv **Variegated part of middle volcaniclastic unit** – Very fine to fine-grained, slope-forming sandstone, siltstone, and mudstone of red, pink, yellowish-brown, and purplish-gray hues; non-volcaniclastic, but some mudstone intervals exhibit swelling soils that suggest a volcanic ash component; overlies the conglomerate at Boat Mesa; present only at the south end of the Sevier Plateau, in Casto Canyon and areas to the south where it is 0 to about 160 feet (50 m) thick; Sable and Maldonado (1997b) reported that the unit is 50 to 125 feet (16–38 m) thick in the Casto Canyon area.

Bowers (1972) first noted the similarity of this interval to the Colton Formation of central Utah, but we now know that the Colton is older, of early Eocene age (Fouch and others, 1983). Eaton and others (1999b) and Korth and Eaton (2004) reported on vertebrate fossils from this "variegated" interval on the southwest Sevier Plateau—a fauna dominated by aquatic taxa suggestive of lacustrine paleoenvironments—which they assigned to the Duchesnean North American Land Mammal Age (end of middle Eocene). Feist and others (1997) also reported on sparse late middle Eocene vertebrate fossils and charophytes from these beds.

unconformity

Tbm, Tbml

Conglomerate at Boat Mesa (middle Eocene) – Contains three distinct, non-volcaniclastic intervals, the lower two of which are combined as Tbml. Uppermost ledge-forming interval (Tbm) is mostly light-gray conglomerate, lesser light-gray to light-brown calcareous sandstone and conglomeratic sandstone, and minor white to light-gray limestone and conglomeratic limestone; clasts are rounded pebbles of black chert, brown, gray, and distinctive greenish quartzite, and lesser Paleozoic limestone; no volcanic or intrusive clasts are present; in the limestone intervals, clasts commonly appear to float in a carbonate mud matrix, but otherwise the conglomerates are clast supported. The lower two intervals (Tbml) include an upper part of slope-forming, reddish-brown and light-gray, thin- to medium-bedded, fine- to medium-grained sandstone, siltstone, and mudstone a few feet to about 30 feet (9 m) thick, and a lower, ledge-forming interval that is yellowish- to reddish-brown, very thick bedded pebbly conglomerate containing clasts similar to the upper interval but lacking green quartzite pebbles; this combined interval pinches out at the south end of the Sevier Plateau, south of South Fork Limekiln Creek, but reappears farther south in Bryce Canyon National Park (it is lumped with Tbm between Bryce Point and Inspiration Point due to map scale); about 100 feet (30 m) thick at Boat Mesa in Bryce Canyon National Park (Bowers, 1990) and typically about 50 to 100 feet (15–30 m) thick on the southwest flank of the Sevier Plateau; on the Markagunt Plateau, a thin pebble conglomerate that unconformably overlies the Claron Formation, and that we suggest is a westward facies of the conglomerate at Boat Mesa, is a few feet to perhaps 10 feet (3 m) thick but is not mapped separately.

The conglomerate at Boat Mesa unconformably overlies the white member of the Claron Formation. It is especially well developed at its namesake Boat Mesa in Bryce Canyon National Park, and on the southwest flank of the Sevier Plateau where it unconformably overlies—and ultimately cuts out—the upper limestone interval of the white member (Tcwu) of the Claron Formation and is in turn overlain by the lower unit (Tbh_1) of the Brian Head Formation or by the variegated part of the middle volcaniclastic unit (Tbhv). The upper part of the conglomerate at Boat Mesa (Tbm) weathers to a white ledge, making it difficult to distinguish from the upper white limestone unit of the Claron Formation on aerial photos or from a distance.

The conglomerate at Boat Mesa represents deposits of braided stream channels and minor floodplains incised into deposits of the Claron Formation. Bowers (1990) suggested the conglomerate at Boat Mesa was Oligocene and Davis and Pollock (2010) thought it was Oligocene or Miocene, but lack of volcanic clasts suggests latest middle Eocene age, confirmed by an overlying 36.51 ± 1.69 Ma volcanic ash in the Brian Head Formation (unit Tbh_1) (UGS and AtoZ, 2013a). We also obtained a U-Pb detrital age on zircon of 37.97 +1.78/-2.70 Ma (table 5) (Gary Hunt, UGS, written communication, March 7, 2012; UGS and A2Z, Inc., 2013b) for the unit on the southwest flank of the Sevier Plateau. This sample also yielded a Middle Jurassic peak of about 168 Ma from a coherent group of 34 grains, indicating that Middle Jurassic volcanic or intrusive rocks provided a significant source of sediment to the formation even though it lacks such clasts.

unconformity

Claron Formation (Eocene to Paleocene) – Claron Formation strata are among the most visually arresting rocks in southwestern Utah, but because the formation lacks a type section and was named for incomplete, fault-bounded exposures in the Iron Springs mining district, the nomenclatural history of these rocks is complicated. Mackin (1947) first applied the name Claron Formation to strata of the Markagunt Plateau, noting the similarity with rocks in the Iron Springs mining district to which Leith and Harder (1908) first applied the name. Bowers (1972) subdivided the Claron Formation into three informal members on the Table Cliff Plateau immediately east of the map area: ascending, the lower pink, white limestone, and variegated sandstone members. Bowers' variegated member is what we map as the conglomerate at Boat Mesa (Tbm and Tbml) and overlying non-volcaniclastic Brian Head strata (Tbhv) on the south flank of the Sevier Plateau. Anderson and Rowley (1975) provide the best review of this nomenclatural history, although since then, additional mapping and stratigraphic work has enabled the upper part of their Claron Formation to be split off as the conglomerate at Boat Mesa and the Brian Head Formation as described above. Anderson and Rowley (1975) considered the formation to have two informal members, a lower red member and an upper white member, but, on the basis of precedence and a long informal usage of the term "pink" when referring to the uppermost part of the Grand Staircase and its strata, we retain Bowers' informal name pink member for the lower Claron. We further subdivide the white member and restrict it to non-volcaniclastic strata unconformably overlain by the conglomerate at Boat Mesa, and note that the pink member has long been known informally as the red member on the Markagunt Plateau.

We thus map the Claron Formation as two members made up of five informal lithostratigraphic units described below in descending order: the upper white member (which is divided into an uppermost mudstone unit, an upper limestone unit, a middle mudstone and sandstone unit, and a lower limestone unit) and the lower pink member. The several lithologic facies of the white member of the Claron are here mapped separately for the first time, which has proven useful to better understand faulting at the west edge of the Markagunt Plateau, the distribution of overlying volcaniclastic strata of the Brian Head and Limerock Canyon Formations, and the southern depositional limit of the Markagunt gravity slide and location of inferred toe thrusts.

The Claron Formation consists of mudstone, siltstone, sandstone, limestone, and minor conglomerate deposited in fluvial, floodplain, and lacustrine environments of an intermontaine basin bounded by Laramide uplifts; the pink member is almost wholly fluviatile and the white member is both lacustrine and fluviatile (Goldstrand, 1990, 1991, 1992, 1994; Bown and others, 1997). Ott (1999) recognized a 130-foot-thick (40 m) interval of mostly medium-bedded bioclastic limestone and thin-bedded micritic limestone with gastropods, ostracods, charophytes, and algal filaments in the lower part of the pink member in Bryce Canyon National Park; this lacustrine interval does not appear to be present in western exposures, suggesting that lacustrine strata are better developed in the central part of the basin. Much of the pink member, and clastic parts of the white member, were greatly modified by bioturbation and pedogenic processes, creating a stacked series of paleosols (Mullett and others, 1988a, 1988b; Mullett, 1989; Mullett and Wells, 1990; see also Bown and others, 1997). Ott (1999) reported depositional cyclicity within the Claron Formation at Bryce Canyon National Park, with multiple regressive cycles, each with increasing pedogenesis toward its top, stacked one upon the other. Bown and others (1995a, 1995b, 1997) reported on trace fossils of ants, wasps, and bees in the upper part of the pink member and lower part of the white member, recording nest activity during paleosol formation. Hasiotis and Bown (1997) reported on crayfish burrows in Claron strata of the Markagunt Plateau that indicate relatively deep and highly fluctuating water tables in the pink member and relatively shallow water tables in alluvial parts of the white member. Davis and others (2009) used isotopic and elemental records preserved in authigenic calcite from samples in the Claron, Flagstaff, and Uinta lake basins to better understand Paleogene landscape evolution of Utah, showing an along-strike migration of a high-elevation landscape from north to south over time and, in southwest Utah, a transition from a closed to open basin beginning with deposition of the white member; Ott (1999) also showed that the pink member was deposited in a hydrologically closed basin. Detrital zircon studies of the Claron Formation from the Escalante Mountains east of the map area show that the formation there was largely derived from erosion of lower Paleozoic sandstones exposed in surrounding Laramide uplifts (Link and others, 2007; Larsen and others, 2010). The Claron Formation is typically forested and covered by colluvium, but it forms the Pink Cliffs, the uppermost riser of the Grand Staircase, and is spectacularly exposed at Cedar Breaks National Monument and Bryce Canyon National Park.

The Claron Formation is unconformably overlain by the conglomerate at Boat Mesa, or, where the conglomerate is apparently locally absent on the west flank of the Markagunt Plateau, by the Brian Head Formation. It appears that the white member of the Claron Formation is locally missing on the west flank of the Markagunt Plateau, for example, southeast of Parowan and northeast of Paragonah, and the conglomerate at Boat Mesa is thin or missing there as well. We are uncertain if this represents nondeposition or erosion of white member strata, but suspect the former, implying that this area was the former northwestern margin of lacustrine deposition. It is possible that clastic strata of the uppermost pink member in western exposures are coeval basin-margin facies of lacustrine strata of the white member in eastern exposures.

The age of the white member is well constrained as late middle Eocene (Duchesnean Land Mammal Age), but the maximum age of the mostly nonfossilferous pink member is poorly constrained as Eocene to Paleocene(?) as described in the accompanying text. The lower part of the pink member is likely Paleocene in age, but given its paucity of datable materials, we cannot yet rule out the possibility that it is latest Cretaceous.

Tc **Claron Formation, undivided** (Eocene and Paleocene) – Mapped west of Brian Head where we are unable to differentiate pink and possible white member strata; exposed thickness is about 400 feet (120 m).

Tcw **White member, undivided** (Eocene) – Lithologies are described below for individual units. On the Markagunt Plateau, used for areas south of Blue Spring Mountain (southwest of Panguitch Lake) and west of Brian Head peak where incomplete and isolated exposures preclude subdivision; also used on the southeastern Sevier Plateau where intervening clastic intervals lose character and most of the member appears to be white micritic limestone, and at Flake Mountain, where what we map as the white member may in fact be the pink member.

We also map the white member at Boat Mesa and northward along the escarpment in the northeast part of Bryce Canyon National Park; Bowers (1990), however, mapped these strata as the pink member. Ledge-forming white limestone typical of the white member is present on hill 8155 west of Boat Mesa, but appears to grade east and north into interbedded varicolored mudstone, siltstone, and limestone that weathers to distinctly lighter colored slopes than the underlying pink member.

In aggregate, the white member maintains a relatively uniform thickness of about 350 to 450 feet (105–135 m) on the Markagunt and Paunsaugunt Plateaus. However, the member thickens to the east where it is dominated by white micritic limestone and is about 550 feet (170 m) thick on the Table Cliff Plateau (Bowers, 1973). The entire white member is about 340 feet (100 m) thick in Rock Canyon southeast of Panguitch Lake. Hatfield and others (2010) reported that it is 360 feet (110 m) thick at Cedar Breaks National Monument, but including the lower sandstone and conglomerate unit of Sable and Maldonado (1997b), as suggested here, the thickness is 440 feet (135 m) (regardless, the white member is truncated south of Cedar Breaks National Monument by late Tertiary and Quaternary erosion associated with development of the Markagunt Plateau). Moore and others (1994) reported significant facies changes in the white member near Asay Bench, but there, in aggregate, it is 448 feet (137 m) thick. The white member is about 350 feet (105 m) thick on the southern Sevier Plateau. Ott (1999) measured an incomplete section of 233 feet (71 m) of the white member near Bryce Point, but it is unclear if her lower contact is the same as that used on this map.

Tcwt **Uppermost mudstone, siltstone, and sandstone unit of white member** (upper and middle Eocene) – Varicolored and commonly mottled, pale-reddish-orange, reddish-brown, moderate-orange-pink, dark-yellowish-orange, and grayish-pink calcareous mudstone and siltstone, locally with minor fine-grained silty sandstone and micritic limestone; indistinguishable in lithology and color from the middle white (Tcwm) and pink members (Tcp) of the Claron Formation.

Forms a brightly colored slope at the top of the upper white member of the Claron Formation in the northern part of Cedar Breaks National Monument, where it is best exposed near the North View Overlook. There, it is 109 feet (33 m) (Schneider, 1967) of mudstone and siltstone capped by a thin calcareous sandstone and pebbly conglomerate, described below. Also exposed immediately south of Panguitch Lake, where it is about 50 feet (15 m) thick, and in the upper reaches of Rock Canyon. This unit is queried near Winn Gap at the south end of the Red Hills, where, on the basis of a similar four-part limestone-clastic-limestone-clastic section above the pink member, we infer that the white member may be present. The unit is thin or absent on the southeastern Markagunt Plateau and on the Sevier Plateau.

Schneider (1967) reported biotite in some of these beds, and while some beds exhibit slightly expansive soils, we found no biotite—even so, it was the apparent presence of biotite-bearing strata, and possible correlation to variegated strata on the southern Sevier Plateau (see Feist and others, 1997), that led Sable and Maldonado (1997b) to provisionally include these strata as a basal part of their Brian Head Formation. However, in light of the absence of tuffaceous material in these beds, these same exposures strongly suggest to us that the non-tuffaceous sandstone and conglomerate as defined by Sable and Maldonado (1997b) is simply an uppermost unit of the Claron Formation (Tcwt) and the overlying conglomerate at Boat Mesa (Tbm, Tbml). We place an unconformity at the base of the thin sandstone and conglomerate (not at the top of the limestone ledge of the white member), thereby including the Claron-like red beds as a new upper unit of the white member. Similar strata are present above the conglomerate at Boat Mesa on the southwest flank of the Sevier Plateau, but because they lie above, not below, the conglomerate, we assign them to the basal unit (variegated unit, Tbhv) of the Brian Head Formation.

Tcwu **Upper limestone unit of white member** (Eocene) – White, pale-yellowish-gray, pinkish-gray, and very pale orange micritic limestone and uncommon pelmicritic limestone, locally containing intraformational rip-up clasts; locally contains sparse charophytes and planispiral snails; typically poorly bedded and knobby weathering; locally vuggy with calcite spar and commonly cut by calcite veinlets; resistant and so forms prominent ledges and flat ridge tops; upper conformable contact with Tcwt corresponds to a pronounced color change from white to very pale orange micritic limestone below to brightly colored reddish-orange mudstone and siltstone above; queried at the south end of the Red Hills.

The upper limestone unit of the white member thickens irregularly to the east and ranges from about 30 to 180 feet (10–55 m) thick, but some of this variation may be due to difficulties in placing the locally gradational contacts. The unit is 45 to 60 feet (14–18 m) thick at Cedar Breaks (Schneider, 1967; Moore and others, 2004; Rowley and others, 2013) and about 30 feet (10 m) thick in the southern Red Hills (Threet, 1952). It is about 80 to 100 feet (24–30 m) thick in the Black Rock Valley area south-southeast of Panguitch Lake, 80 to 165 feet (24–50 m) thick southwest of Hatch in the Asay Bench quadrangle (Moore and others, 1994), and about 150 to 180 feet (45–55 m) thick near Houston Mountain north-northeast of Navajo Lake (Biek and others, 2011). It is typically about 80 to 100 feet (25–30 m) thick on the southern flank of the Sevier Plateau, but north of the latitude of Limekiln Creek it grades into a mudstone-dominated interval similar to that of the underlying middle unit.

Tcwml **Middle mudstone, siltstone, and sandstone unit and lower limestone unit of white member, undivided** (Eocene) – Locally undivided at Cedar Breaks National Monument due to map scale, and about 5 miles (8 km) to the east at Houston Mountain due to poor exposure. As mapped, less than about 250 feet (75 m) thick.

Tcwm **Middle mudstone, siltstone, and sandstone unit of white member** (upper middle Eocene) – Varicolored and commonly mottled, pale-reddish-orange, reddish-brown, moderate-orange-pink, yellowish-gray, dark-yellowish-orange, and grayish-pink calcareous mudstone and siltstone, and minor fine-grained calcareous sandstone and chert-pebble conglomerate that weathers to a poorly exposed slope; upper conformable contact corresponds to a pronounced color change from brightly colored reddish-orange mudstone and siltstone below to white to very pale orange micritic limestone above; queried at the south end of the Red Hills.

As mapped on the southwest side of the Sevier Plateau, consists of a deep reddish-orange siltstone and mudstone interval about 60 feet (18 m) thick that becomes difficult to identify north of Limekiln Creek. About 120 feet (36 m) thick near Cameron Troughs south of Panguitch Lake, but appears to thin abruptly to about 50 feet (15 m) thick about one mile (1.6 km) to the east. At Cedar Breaks National Monument, Schneider (1967) measured 227 feet (69 m) of strata we assign to Tcwm, but Rowley and others (2013) reported that this interval is 310 feet (94 m) thick in this same area. Moore and others (1994) reported that their middle sandy unit is 175 to at least 220 feet (54–67 m) thick in the Asay Bench quadrangle southwest of Hatch.

Eaton and others (2011) reported the first sparse late middle Eocene (Duchesnean Land Mammal Age) vertebrate fossils and ostracods of *Cypris* sp. from this unit on the eastern Markagunt Plateau.

Tcwl **Lower limestone unit of white member** (Eocene) – Micritic limestone similar to the upper white limestone interval (Tcwu); forms cliff or steep, ledgy, white slope above more colorful but typically subdued slopes of the pink member (Tcp); upper conformable contact corresponds to a pronounced color change from white to very pale orange micritic limestone below to brightly colored reddish-orange mudstone and siltstone above; query indicates uncertain identification on Navajo Ridge northwest of Brian Head peak, at the south end of the Red Hills, and near Willow Creek northeast of Paragonah.

As mapped on the southwest flank of the Sevier Plateau, the lower limestone interval is multi-hued mudstone, siltstone, and limestone that weathers to white and pink, subdued slopes. As such, its lower and upper contacts are difficult to pick except from a distance or on aerial photographs, and we are uncertain of correlation to exposures on the Markagunt Plateau.

The lower limestone unit attains its maximum thickness of about 300 feet (90 m) at Bryce Point in Bryce Canyon National Park (Bowers, 1990), but is only about 160 feet (50 m) thick to the north on the southwest flank of the Sevier Plateau. To the west, on the Markagunt Plateau, it is about 100 to 120 feet (30–35 m) thick in the upper reaches of Rock Canyon. Moore and others (1994) reported that their lower white limestone is generally 85 to 120 feet (26–36 m) thick, but as much as 180 feet (55 m) thick, in the Asay Bench quadrangle. This unit is only about 47 feet (14 m) thick at Cedar Breaks National Monument, where it is informally called the “lower white limestone” (Schneider, 1967; Rowley and others, 2013), and about 30 feet (10 m) thick in the southern Red Hills (Threet, 1952).

Tcp **Pink member** (Eocene and Paleocene) − Alternating beds of varicolored and commonly mottled, pale-reddish-orange, reddish-brown, moderate-orange-pink, dark-yellowish-orange, and grayish-pink sandy and micritic limestone, calcite-cemented sandstone, calcareous mudstone, and minor pebbly conglomerate that weather to colluvium-covered ledgy slopes. Limestone is poorly bedded, microcrystalline, generally sandy with 2 to 20% fine-grained quartz sand, and is locally argillaceous; contains common calcite veinlets, calcite spar-filled vugs, calcite spar- and micrite-filled burrows, and stylolites; also contains sparse small bivalves and planispiral gastropods; many of these limestone beds are calcic paleosols (Mullett and others, 1988a, 1988b; Mullett, 1989; Mullett and Wells, 1990). Sandstone is thick-bedded, fine- to coarse-grained, calcareous, locally cross-bedded quartz arenite that typically weathers to sculpted or fluted ledges that pinch out laterally and that locally contain pebble stringers. Mudstone is generally moderate reddish orange, silty, calcareous, contains calcareous nodules, and weathers to earthy, steep slopes between ledges of sandstone and limestone. Pebbly conglomerate forms lenticular beds typically 5 to 15 feet (2–5 m) thick containing rounded quartzite, limestone, and chert pebbles, cobbles, and, locally, small boulders; conglomerate is uncommon on the Markagunt Plateau south of Parowan Canyon, but lower pink member strata are abundantly conglomeratic in the Red Hills and at the northwest edge of the Markagunt Plateau north of Parowan; at Sugarloaf Mountain west of Brian Head, several tens of feet of conglomerate (or several thinner beds within this interval) overlie the basal Claron limestone; Gregory (1951) and Bowers (1990) reported conglomeratic basal Claron strata on the Paunsaugunt Plateau. Upper, conformable contact corresponds to a pronounced color and lithologic change from brightly colored reddish-orange mudstone and siltstone below to white to very pale orange micritic limestone above.

Sinkholes are common in the pink member in the central Markagunt Plateau (Moore and others, 2004; Biek and others, 2011; Hatfield and others, 2010; Rowley and others, 2013). Large sinkholes visible on 1:20,000-scale aerial photographs are plotted on the geologic map, and doubtless many smaller sinkholes are present. These sinkholes capture local runoff and serve to shunt shallow groundwater rapidly down dip where it emerges as springs, including the large Mammoth and Asay Springs (Wilson and Thomas, 1964; Spangler, 2010).

The pink member is mostly nonfossiliferous and its age is poorly constrained as Eocene to Paleocene(?) (Goldstrand, 1994) as described above. Measurements from the map suggest that the pink member is about 1000 feet (300 m) thick at Cedar Breaks National Monument, similar to the measured thickness of Schneider (1967), who reported that the pink member there was 993 feet (303 m) thick (however, the lower 56 feet [17 m] of his section includes beds we assign to Km, thus the pink member there is 937 feet [286 m] thick), considerably less than the 1300 feet (400 m) reported by Sable and Maldonado (1997b). Strata that we include in the pink member are likely of similar thickness in more structurally complicated outcrops of the Red Hills (Threet, 1952, 1963a). The pink member is about 600 feet (180 m) thick at Bryce Canyon National Park.

Tcpl **Pink member, limestone marker bed** (Eocene?) – White to very pale orange micritic limestone presumed to be near the top of the pink member, but that may represent the northwest feather edge of either the lower or upper limestone units of the white member. Forms conspicuous ledge about 60 feet (18 m) thick near Mineral Canyon northeast of Paragonah, where it underlies a few hundred feet of Claron red beds.

CRETACEOUS-TERTIARY

The Pine Hollow and Canaan Peak Formations are present on the Table Cliff Plateau immediately east of the Panguitch 30′ x 60′ quadrangle (Doelling and Willis, 1999b), but are not present in this map area. Both are poorly constrained as Late Cretaceous to early Tertiary based on their stratigraphic position (Bowers, 1972; Larsen and others, 2010).

CRETACEOUS

Km **Cretaceous strata on the Markagunt Plateau** (Upper Cretaceous, Maastrichtian[?] to upper Campanian) – Consists of yellowish-brown, commonly stained dark-reddish-brown, fine-grained sandstone and lesser interbedded, similarly colored mudstone and siltstone; bedding is thin to very thick and appears tabular from a distance; weathers to ledgy slope or cliff; outcrop habit and surficial color make it look like the pink member of the Claron Formation from a distance; not mapped in Parowan Canyon and areas to the north, where basal Claron strata are conglomeratic and identification of this interval, if present, is uncertain; upper contact placed at the base of the lowest sandy limestone bed (calcic paleosol) of the pink member of the Claron Formation, following Moore and Straub (2001); about 200 feet (60 m) thick near State Highway 14 at the west edge of the Markagunt Plateau, but apparently thins to the north where it may be about 60 feet (20 m) thick in Parowan Canyon.

These Markagunt Plateau strata represent fluvial and floodplain environments apparently gradationally overlain by the Claron Formation. Like Moore and Straub (2001), we recognize no significant erosion beneath the Claron Formation at the west edge of the Markagunt Plateau, leading to uncertainty as to the age of this interval and the age of basal Claron strata. Nichols (1997) reported Late Cretaceous (Santonian?) pollen from strata we map as Km south and west of Blowhard Mountain, and we recovered late Campanian to Maastrichtian pollen from this same interval (table 1). The statigraphic position of Km strata precludes it being Santonian in age. The apparently gradationally overlying basal Claron Formation is widely believed to be late Paleocene(?) (Goldstrand, 1994) on the basis of one report of late Paleocene palynomorphs from basal Claron strata on the east flank of the Pine Valley Mountains and the gastropods *Viviparus trochiformi*s, *Goniobasis*, and *Physa* from the pink Claron on the Table Cliff Plateau, eastern Pine Valley Mountains, and Bryce Point, respectively (Goldstrand, 1991). We treat Km strata as a separate informal unit, but it is possible that they are better assigned to the base of the Kaiparowits Formation.

Ku **Grand Castle Formation (redefined), capping sandstone member of the Wahweap Formation, and Drip Tank Member of the Straight Cliffs Formation, undivided** (Upper Cretaceous) – Undivided along the northwest flank of the Markagunt Plateau between Red Creek and Little Creek where the three units are too thin to map separately at this scale; about 200 feet (60 m) thick.

Kgc **Grand Castle Formation (redefined)** (Upper Cretaceous?, Maastrichtian? to upper Campanian?) – Light-gray and light-red massive conglomerate; clasts are well-rounded, pebble- to boulder-size quartzite, limestone, sandstone, and chert; redefined and restricted to only the upper member of three informal members (resistant upper conglomerate, non-resistant middle sandstone, and resistant lower conglomerate) of the Grand Castle Formation of Goldstrand and Mullett (1997); cliff-forming in the Parowan Canyon area, where it locally weathers to form hoodoos similar to those of the Drip Tank Member (formerly lower conglomerate member of the Grand Castle Formation); thins dramatically south of Parowan Canyon where it is mapped as a poorly exposed marker bed; we also map Grand Castle Formation (redefined) locally on the Paunsaugunt Plateau, where it is typically thin and poorly exposed at the base of the Claron Formation.

Grand Castle strata are as much as about 200 feet (60 m) (Threet, 1952, 1963a) to 300 feet (90 m) (Maldonado and Williams, 1993a) thick near Parowan Gap;

Anderson and Dinter (2010) reported that there it is about 230 feet (70 m) thick. On the Markagunt Plateau, thins abruptly to the south from 183 feet (56 m) thick at the type area in First Left Hand Canyon southeast of Parowan (Goldstrand and Mullet, 1997) and the conglomerate may locally be absent south of Navajo Ridge (where it was not recognized in the measured sections of Goldstrand, 1991), but this interval is typically mantled in talus and colluvium that may obscure its presence. However, our mapping of Upper Cretaceous strata between Parowan and Cedar Canyons shows this conglomerate to be present at Sugarloaf Mountain (about 8 miles [13 km] south of Summit), in Last Chance Canyon (a tributary to Cedar Canyon) where it is about 25 feet (8 m) thick, in the upper reaches of Spring Creek Canyon below Cedar Breaks National Monument, west of Blowhard Mountain, and west of Navajo Lake where it is no more than a few feet thick. We thus conclude that though dramatically thinner than at its type section, the upper conglomerate member is, apart from a few local areas, present along the entire west margin of the Markagunt Plateau.

The upper contact with strata here mapped as Km on the Markagunt Plateau, and mapped as basal pink member of the Claron Formation in the Red Hills and northwestern Markagunt Plateau, appears gradational. On the Markagunt Plateau south of Parowan, the upper contact corresponds to the base of ledge- and cliff-forming, tabular-bedded sandstone stained dark-reddish-brown from overlying Claron strata. Elsewhere, the upper contact corresponds to the top of the cliff-forming conglomerate, above which is interbedded reddish-brown siltstone, sandstone, mudstone, sandy limestone, and pebbly conglomerate of the Claron Formation.

The Grand Castle Formation (redefined) was deposited in a braided fluvial environment with paleoflow principally to the east to south-southeast, suggesting source areas in the Wah Wah, Blue Mountain, and Iron Springs thrust sheets of southwest Utah (Goldstrand and Mullett, 1997). Goldstrand and Mullett (1997) inferred a Paleocene age for their entire Grand Castle Formation on the basis of distant correlations with Canaan Peak and Pine Hollow Formations on the Table Cliff Plateau, but we found evidence that the lower two members are Late Cretaceous and are thus reassigned as the Drip Tank Member of the Straight Cliffs Formation and the capping sandstone member of the Wahweap Formation as described below. The age of their upper conglomerate member of the Grand Castle Formation (here, the Grand Castle Formation [redefined]) is not well constrained, but we suggest that it is indeed Late Cretaceous. A debris-flow deposit within the upper conglomerate member at its type section yielded Late Cretaceous (Santonian?) pollen that Goldstrand and Mullett (1997) interpreted as recycled from older strata. However, Nichols (1997) reported Late Cretaceous *Proteacidites* sp. pollen, which he interpreted as Coniacian and Santonian, from overlying beds here mapped as Km west and south of Blowhard Mountain, and we recovered late Campanian to Maastrichtian pollen from this same interval (table 1).

In the northern Red Hills, the Grand Castle Formation (redefined) was earlier referred to informally as the conglomerate of Parowan Gap by Maldonado and Williams (1993a). However, like Goldstrand and Mullet (1997), we infer this interval to be their upper conglomerate member of the Grand Castle Formation because it is gradationally overlain by the pink member of the Claron Formation and their underlying middle Grand Castle sandstone is absent. Anderson and Dinter (2010) reported a 10- to 15-foot-thick (3–5 m), poorly sorted, matrix-supported conglomerate at the base of the Grand Castle that they informally called their conglomerate of Parowan Gap. They described this unit, which is apparently restricted to the hanging wall of the Iron Springs thrust, as distinct from overlying Grand Castle conglomerate, but we found mostly clast-supported conglomerate identical to the Grand Castle conglomerate at this horizon. The basal few feet of Grand Castle Formation are locally iron stained throughout the Parowan Gap area, likely a result of a strong permeability contrast between underlying Upper Cretaceous strata and the overlying Grand Castle conglomerate.

unconformity

Kk **Kaiparowits Formation** (Upper Cretaceous, upper Campanian) – Bluish-gray, gray, and locally greenish- or brownish-gray, fine-grained, feldspathic, lithic sandstone, mudstone, and siltstone; contains locally abundant gastropods; typically poorly cemented, weathering to badland slopes whose bluish-gray hues contrast sharply with overlying pinkish Claron paleosols and underlying yellowish-brown Wahweap strata; on the west side of the Paunsaugunt Plateau, however, commonly heavily forested and involved in large landslides, where it is recognized as far south as Proctor Canyon but presence of landslide deposits along its outcrop belt suggests that it may extend even farther south to Big Hollow; on the west flank of the plateau, base of map unit is a bluish-gray smectitic mudstone as much as several meters thick, the slip surface for the landslides.

Incomplete sections of the Kaiparowits Formation are preserved about 3 miles (5 km) southeast of Flake

Mountain where it is as much as about 450 feet (135 m) thick, and on the west flank of the Paunsaugunt Plateau where it is as much as about 200 feet (60 m) thick. The entire formation is about 2820 feet (860 m) thick in the Kaiparowits basin east of the map area (Eaton, 1991; Doelling and Willis, 1999b; Roberts and others, 2005).

The Kaiparowits Formation was deposited as an eastward-prograding clastic wedge in a relatively wet, subhumid alluvial plain with periodic to seasonal aridity near the western margin of the Late Cretaceous Western Interior Seaway (Roberts, 2007; Roberts and others, 2013). It is abundantly fossiliferous, with one of the richest and most diverse terrestrial vertebrate faunas of the Cretaceous Western Interior Basin (Roberts, 2007; Roberts and others, 2013). Roberts and others (2005) reported four $^{40}Ar/^{39}Ar$ ages on sanidine from altered volcanic ashes that bracket the age of the formation in the Kaparowits Basin between 76.1 and 74.0 Ma, and that demonstrate extremely rapid sediment accumulation rates of 16 inches/kyr (41 cm/kyr); using the new Fish Canyon Tuff reference age of 28.201 ± 0.046 Ma (Kupier and others, 2008), Roberts and others (2013) recalibrated these ages to be about 76.6 to 74.5 Ma. We report a new U-Pb age on zircon of 75.62 +3.08/-1.66 Ma for the bluish-gray smectitic mudstone at the base of the formation in Johnson Canyon on the west flank of the Paunsaugunt Plateau (table 5) (UGS and A2Z, Inc., 2013b; Gary Hunt, written communication, September 26, 2011); we also recovered late Campanian to Maastrichtian palynomorphs from this location (table 1).

Tilton (1991, 2001a, 2001b) first subdivided Upper Cretaceous strata of the southern Paunsaugunt Plateau, correctly noting the absence there of Kaiparowits Formation. In the area southwest of Tropic Reservoir, Bowers (1990) assigned light-brown, very fine grained sandstone and gray sandy mudstone (above the capping sandstone member of the Wahweap Formation) to the Kaiparowits Formation, and although these beds are unlike typical bluish-gray, feldspathic, lithic Kaiparowits strata, they are correlative to strata we map as the lower unit of the Kaiparowits Formation (Kkl). Intermediate-scale maps of the southernmost Paunsaugunt and Markagunt Plateaus (Sable and Hereford, 2004; Doelling, 2008) incorrectly show Kaiparowits Formation on the Markagunt Plateau, and although unmapped, incorrectly suggest that it is intermittently present on the southern Paunsaugunt Plateau, which Tilton (2001a, 2001b) clearly showed is not the case.

Kkl **Lower unit** (Upper Cretaceous, upper Campanian) – Yellowish-brown, fine-grained sandstone and varicolored and mottled, reddish-brown, purplish-gray, and gray mudstone; sandstone forms two prominent ledges at the base and near the middle of the unit in Hillsdale Canyon on the west flank of the Paunsaugunt Plateau; upper contact placed at the base of the first sandy limestone bed (calcic paleosol) or pebble to cobble conglomerate of the pink member of the Claron Formation, or at the base of bluish-gray, fine-grained, feldspathic, lithic sandstone, mudstone, and siltstone of the Kaiparowits Formation; as much as about 250 feet (75 m) thick in the Hillsdale Canyon area, but thins southward under the Claron unconformity; not differentiated from the main body of the Kaiparowits in exposures north of Tropic; as much as about 400 feet (120 m) thick south of Hillsdale Canyon, between Big Hollow and Proctor Canyon, but there it may include strata elsewhere mapped as the gravelly unit (Kwcg) of the Wahweap Formation.

Although undated and lacking bluish-gray feldspathic sandstone and mudstone that characterize the bulk of the Kaiparowits Formation in the Kaiparowits Basin, we assign this interval to an informal lower unit of the Kaiparowits Formation because it appears lithologically similar to basal Kaiparowits strata along Henrieville Creek on the west flank of the Kaiparowits Plateau (Eaton, 1991) and because it is lithologically unlike the underlying capping sandstone member of the Wahweap Formation. Welle (2008) and Lawton and Bradford (2011) showed that the lower unit of the Kaiparowits Formation in the Kaiparowits basin has a different detrital zircon signature than that of its middle and upper units, with more thrust-belt-derived grains, recording a transition in sediment source areas from the thrust-belt-sourced capping sandstone to arc-derived Kaiparowits.

Kws **Wahweap and Straight Cliffs Formations, undivided** (Upper Cretaceous) – Mapped in the Johns Valley area where it includes uppermost John Henry, Drip Tank, and basal Wahweap strata; at least 1000 feet (300 m) thick.

Kwcsd **Capping sandstone member of the Wahweap Formation and Drip Tank Member of the Straight Cliffs Formation, undivided** (Upper Cretaceous) – Mapped north of Red Creek (north of Paragonah) where too thin to map separately at this scale.

Wahweap Formation (Upper Cretaceous, middle Campanian) – Eaton (1991) divided the formation into four informal members in the Kaiparowits basin based principally on sandstone-to-mudstone ratios and fluvial architecture. In ascending order, these include his lower, middle, upper, and capping sandstone members. However, because of extensive vegetative cover and poor geomorphic expression in this map area, we map his lower three members simply as Wahweap Formation, undivided (Kw). The distinctive capping sandstone (Kwcs) is mapped separately, except on the south side of Johnson and Hillsdale Canyons due to poor exposure. We map one additional unit that overlies the capping sandstone member on the west flank of the Paunsaugunt Plateau (Kwcg). It is possible that Kwcg is simply an unusual facies of the capping sandstone member.

The Wahweap Formation is mostly fine-grained sandstone, siltstone, and mudstone deposited in braided and meandering river and floodplain environments of a coastal plain (Tilton, 1991; Pollock, 1999; Lawton and others, 2003; Jinnah and Roberts, 2011). Detrital zircon and provenance studies of Eaton's lower three members show that these rivers flowed longitudinally to the foreland basin and tapped sources in the Cordilleran magmatic arc in southern California or western Nevada and the Mogollon Highlands of southern Arizona, but that the capping sandstone member was deposited by transverse streams that tapped Mesozoic quartzose sandstones in the Sevier orogenic belt (Pollock, 1999; Lawton and others, 2003; Eaton, 2006; Jinnah and others, 2009). Thus the basal contact of the capping sandstone member represents an abrupt change in color, petrology, grain size, and fluvial style, documenting a major shift in depositional environments from meandering to braided rivers, and in source areas from arc to orogenic belt.

Jinnah (2013; see also Jinnah and others, 2009) reported an $^{40}Ar/^{39}Ar$ age of 80.6 ± 0.3 Ma (Campanian) on a devitrified volcanic ash located about 130 feet (40 m) above the base of the Wahweap Formation on the Kaiparowits Plateau, and further noted that the formation was deposited between about 81 and 77 Ma. Eaton and others (1999a) and Eaton (2006) reported enigmatic fossil mammals from near the base and top of the formation in Cedar Canyon that may be Campanian, and Lawton and others (2003) reported middle Campanian pollen from the upper part of the formation near Webster Flat, just southwest of Blowhard Mountain on the west rim of the Markagunt Plateau. We also recovered Santonian to Campanian pollen immediately below Kwcs near Webster Flat (table 1).

Kw **Wahweap Formation, upper, middle, and lower members, undivided** (Upper Cretaceous, middle Campanian) – Varicolored and mottled mudstone of brown, gray, reddish-brown, and pinkish hues, and yellowish-brown fine-grained sandstone and silty sandstone.

Exposures on Markagunt Plateau: Typically heavily vegetated and poorly exposed on the Markagunt Plateau, but likely equivalent to the upper, middle, and lower members as defined by Eaton (1991). The upper part, below the capping sandstone member, contains more sandstone than mudstone, also noted by Moore and Straub (2001) and Moore and others (2004). Measurements from the map show that the Wahweap Formation, excluding the capping sandstone member, is about 800 feet (245 m) thick below Cedar Breaks National Monument and south of Blowhard Mountain, and Moore and Straub (2001) measured 760 feet (230 m) of strata in Cedar Canyon that we assign to the Wahweap Formation (not including the capping sandstone member). Only a few tens of feet of Wahweap Formation, undivided are present at the base of the capping sandstone member (formerly middle sandstone member of the Grand Castle Formation) in Parowan Canyon, illustrating a dramatic southward thickening of the unit south of Parowan Canyon. Eaton and others (1999a) and Moore and Straub (2001) initially proposed correlation of Cedar Canyon strata in the map area with early to middle Campanian Wahweap strata on the Kaiparowits Plateau.

Exposures on Paunsaugunt Plateau: Moderately well exposed and equivalent to the upper, middle, and lower members as defined by Eaton (1991). Collectively, these members thicken from east to west, from about 300 feet (90 m) thick about 6 miles (10 km) north-northeast of Tropic to in excess of 700 feet (210 m) thick on the west flank of the plateau. Thickness estimates of the Wahweap Formation on the west flank of the plateau are complicated by the Sand Pass fault, which cuts out the middle part of the formation. In Hillsdale Canyon, an exceptionally thick section of the uppermost Wahweap Formation (here mapped as Kwcg and Kwcs, described below), is about 350 feet (105 m) thick, so we estimate that our entire Wahweap Formation is about 950 to 1050 feet (290–320 m) thick on the west flank of the plateau.

Kwcg **Pebbly sandstone unit** (Upper Cretaceous, middle Campanian) – Yellowish-brown, fine- to medium-grained sandstone with numerous, thin pebbly conglomerate stringers and lenticular beds as much as about 3 feet (1 m) thick; clasts are rounded quartzite, chert, and minor limestone; forms prominent cliff in the lower reaches of Hillsdale Canyon, but elsewhere weathers to ledgy slopes; mapped between Wilson Canyon (immediately north of Hillsdale Canyon) and Johnson Canyon, but was not definitely identified south of Johnson Canyon; appears to thin and become less pebbly eastward from Hillsdale Canyon and may pinch out north of Johnson Bench; as much as about 200 feet (60 m) thick in Hillsdale Canyon.

Kwcs **Capping sandstone member** (Upper Cretaceous, middle Campanian) – White to very pale orange, locally iron stained, very fine to coarse-grained, mostly medium-grained, trough cross-bedded quartz arenite that "caps" the Wahweap Formation in its type area; upper part contains abundant pebble stringers and conglomeratic beds with rounded quartzite, dolomite, chert, and limestone clasts; clasts are typically about 1 inch (2.5 cm) in diameter but as large as 2 to 3 inches (5–7.5 cm), and include common reddish-brown and purple quartzite clasts, unlike underlying Drip Tank strata; quartz grains are typically well rounded and commonly frosted, recycled from Mesozoic eolianites (Pollock, 1999; Lawton and others, 2003; see also UGS and AtoZ, 2013b); locally contains carbonized or petrified plant debris, small mudstone rip-up clasts, iron concretions, and soft-sediment deformation features; typically poorly cemented, forming distinctive white, manzanita-covered slope-and-bench topography; attains its maximum thickness of 277 feet (85 m) in First Left Hand Canyon southeast of Parowan (this site was the type section of the equivalent and now abandoned middle member of the Grand Castle Formation), and elsewhere, map patterns show that the member is about 200 feet (60 m) thick.

Among the best exposures of the capping sandstone on the Markagunt Plateau are those about 2 miles (3.5 km) southwest of Parowan, at the mouth of Summit Creek canyon, at the type area in First Left Hand Canyon (southeast of Paraowan), and in a State Highway 14 road cut west of Blowhard Mountain. We discovered Campanian to Santonian palynomorphs and a theropod dinosaur track (the latter found by Eric Roberts, formerly with Southern Utah University and now at James Cook University, Australia) in the lower part of the interval in an unnamed canyon about 2 miles (3 km) southwest of Parowan, confirming our suspicion of a Late Cretaceous age for this member on the Markagunt Plateau (Hunt and others, 2011a).

As defined in the Kaiparowits basin, includes yellowish-brown, fine-grained sandstone locally with thin lenses of mudstone (similar to parts of Eaton's lower Wahweap strata), but as mapped here restricted to distinctive white quartz arenite facies. Equivalent to the middle member of the Grand Castle Formation as suggested by Lawton and others (2003), confirmed by this mapping, and supported by preliminary detrital zircon analyses of Johnson and others (2011) who examined detrital zircon populations of three samples from the Markagunt and Paunsaugunt Plateaus and concluded that they represent sediments of a Late Cretaceous braided stream system that drained thrust sheets of the Sevier orogenic belt to the west. Goldstrand and Mullett (1997) and Lawton and others (2003) also showed that the member was deposited in a braided fluvial environment with a paleoflow direction principally to the east to south-southeast, suggesting source areas in Navajo Sandstone exposed in the upper plate of the Iron Springs thrust, now exposed in the Red Hills.

Along much of the west flank of the Markagunt Plateau south of Parowan Canyon, the capping sandstone member is commonly covered by talus and colluvium derived from overlying Claron Formation. There, we used a dashed lower contact to indicate our uncertainty as to its true thickness. Our mapping confirms the finding of Goldstrand and Mullet (1997), who first correlated the sandstone at the Websters Flat turnoff with their middle sandstone member of the Grand Castle Formation.

Ki **Iron Springs Formation** (Upper Cretaceous, Santonian or lower Campanian to Cenomanian) – Shown as undivided on cross sections and mapped separately as upper and lower parts in the Red Hills at the west edge of the map area; incomplete section is about 2500 feet (750 m) thick in the Red Hills (Maldonado and Williams, 1993a), but the entire formation is about 3500 to 4000 feet (1070–1220 m) thick in the Pine Valley Mountains (Cook, 1960).

The Iron Springs Formation was deposited principally in braided-stream and floodplain environments of a coastal plain (Johnson, 1984; Fillmore, 1991; Eaton and others, 2001; Milner and others, 2006) and is typically correlated to the Dakota Formation, Tropic Shale, and Straight Cliffs Formation (Eaton, 1999; Eaton and others, 2001). Late Cretaceous age is from Goldstrand (1994) and an ash that is 712 feet (217 m) below the top of the formation in Parowan Canyon (here reassigned to the middle part of the Straight Cliffs Formation), which yielded an $^{40}Ar/^{39}Ar$ age of 83.0 ± 1.1 Ma (Eaton and others, 1999c). Lower Iron Springs strata (Kil) in the upper plate of the Iron Springs thrust may be associated with the maximum transgression of the Greenhorn Sea of late Cenomanian or early Turonian age (Eaton and others, 1997; Eaton, 1999).

Kiu **Upper unit** – Interbedded, ledge-forming, calcareous, cross-bedded, fine- to medium-grained sandstone and less-resistant, poorly exposed sandstone, siltstone, and mudstone present in the footwall of the Iron Springs thrust, the easternmost thrust fault in the Red Hills; variously colored grayish orange, pale yellowish orange, dark yellowish orange, white, pale reddish brown, and greenish gray; locally stained by iron-manganese oxides, and Liesegang banding is common in the sandstone beds; sandstone beds range from quartz arenite to litharenite (Fillmore, 1991; Goldstrand, 1992); weathers to repetitive, thick, tabular sandstone beds and thinner interbedded mudstone; upper contact with the Grand Castle Formation (redefined) is difficult to map on the east side of the Red Hills because of abundant Grand Castle-derived colluvium and faults; Milner and others (2006) reported on dinosaur tracks in upper Iron Springs strata near Parowan Gap, and also noted a diverse assemblage of plant fossils, bivalves, gastropods, turtles, fish, and trace fossils suggestive of late Santonian to early Campanian age (Milner and Spears [2007] mistakenly reported an early Turonian age for these same beds); incomplete section is about 260 feet (80 m) thick in Parowan Gap (Anderson and Dinter, 2010).

Kil **Lower unit** – Interbedded sandstone, siltstone, and mudstone similar to that of the upper unit (Kiu) but restricted to the upper plate of the Iron Springs thrust; contains numerous oyster coquina beds commonly 1 to 3 feet (0.3–1 m) thick; incomplete, thrust-fault bounded section is about 2215 feet (675 m) thick at Parowan Gap (Anderson and Dinter, 2010).

Straight Cliffs Formation (Upper Cretaceous, lower Campanian to Turonian) – Peterson (1969) divided the Straight Cliffs Formation into four members in the Kaiparowits basin: in ascending order, the Tibbet Canyon, Smoky Hollow, John Henry, and Drip Tank Members. Several geologists mapped these members (separately or as lumped upper and lower Straight Cliffs strata) on the Paunsaugunt Plateau, including Tilton (1991, 2001a, 2001b), Doelling and Willis (1999a), Sable and Hereford (2004), and Doelling (2008), and in the text above we described the difficulty encountered in early attempts to carry this nomenclature westward into the Markagunt Plateau. On the Markagunt Plateau, we map the Tibbet Canyon Member where it forms bold cliffs at the west and south margins of the plateau; we lump the Smoky Hollow and John Henry Members since the intervening Calico bed is typically poorly developed; and we map the Drip Tank Member, which is the same interval as the now-abandoned lower conglomerate member of the Grand Castle Formation, as originally suggested by Eaton and others (2001), Moore and Straub (2001), Lawton and others (2003), and Eaton (2006). On the Paunsaugunt Plateau, we map the Tibbet Canyon and Smoky Hollow Members together following Bowers (1990) and Doelling and Willis (1999a), and map the John Henry and Drip Tank Members separately.

The Straight Cliffs Formation is an overall regressive sequence that formed during the last marine incursion of the Western Interior Seaway (see, for example, Eaton and others, 2001; Moore and Straub, 2001; Tibert and others, 2003). The Tibbet Canyon Member represents initial progradational (overall regressive) strata of the Greenhorn Cycle deposited in shoreface, beach, lagoonal, and estuarine environments adjacent to a coastal plain (Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003). The overlying Smoky Hollow, John Henry, and Drip Tank Members were deposited in fluvial and floodplain environments of a coastal plain (Peterson, 1969; Eaton and others, 2001).

Ksd **Drip Tank Member** (Upper Cretaceous, lower Campanian) – On the Paunsaugunt Plateau, Drip Tank strata are white to light-gray, fine- to medium-grained quartzose sandstone, and, in the upper part of the unit, pebbly sandstone and pebbly conglomerate; very thick bedded with prominent cross-stratification; clasts are subrounded to rounded, white and gray quartzite, gray Paleozoic limestone, and black chert, but lack the abundant reddish-brown and purple quartzite clasts found in capping sandstone strata; locally iron stained and locally contains casts of tree limbs; lower sandstone forms distinctive, manzanita-covered slopes and saddles, but upper part of unit tends to form cliffs and ledges; upper contact with the Wahweap Formation appears to be conformable and corresponds to the top of a white sandstone and pebbly sandstone, above which is yellowish-brown, fine-grained sandstone and lesser interbedded, varicolored and mottled mudstone of brown, gray, reddish-brown, and pinkish hues; as mapped, restricted to the white quartz arenite and pebbly conglomerate facies and thus ranges from about 100 to 200 feet (30–60 m) thick on the Paunsaugunt Plateau; Tilton (2001a) reported that the member is 185 to 215 feet (56–66 m) thick on the southern Paunsaugunt Plateau, in the Alton quadrangle immediately south of the map area.

On the Markagunt Plateau, the Drip Tank Member (formerly lower conglomerate member of the Grand Castle Formation) is a massive, cliff-forming, light-gray conglomerate with well-rounded, pebble- to boulder-sized clasts of quartzite, limestone, and minor sandstone and chert. It is 135 feet (41 m) thick at the type section in First Left Hand Canyon southeast of Parowan (Goldstrand and Mullet, 1997) and of similar thickness southwest to Sugarloaf Mountain (about 3 miles [5 km] west of Brian Head). South of this area, however, the Drip Tank thins irregularly southward, ranging from a few feet thick to nearly

100 feet (30 m) thick, and locally appears as two conglomerate intervals separated by a few feet to a few tens of feet of yellowish-brown, fine-grained sandstone or variegated mudstone. The Drip Tank Member typically overlies stacked or amalgamated sandstone beds, but locally, as along Ashdown Creek in the southwest corner of the map area, overlies variegated mudstone. On the northern Markagunt Plateau, the member locally weathers to form conically shaped hoodoos that resemble old-fashioned beehives known as bee skeps, but south of Summit it forms a resistant ledge in the upper reaches of Summit Creek canyon and the upper reaches of Pickering Creek canyon.

Tilton (1991) described the Drip Tank Member as the most prominent and important marker horizon in the Upper Cretaceous section on the southern Paunsaugunt Plateau, but we find that it is remarkably similar in lithology and outcrop habit to the capping sandstone member of the Wahweap Formation. The Drip Tank was deposited by east- and northeast-flowing braided streams (Tilton, 1991, 2001a, 2001b; Lawton and others, 2003); age from Jinnah and others (2009, 2011).

unconformity

Ksjs **John Henry and Smoky Hollow Members, undivided** (Upper Cretaceous, Santonian to Turonian) – Undivided on the Markagunt Plateau, where the Calico bed is poorly exposed and apparently only locally well developed; Smoky Hollow strata are described below and John Henry strata are described separately; combined unit is about 1250 to 1350 feet (380–410 m) thick on the Markagunt Plateau (Eaton and others, 2001; Moore and Straub, 2001).

Smoky Hollow strata are slope-forming, brown and gray mudstone, shale, and interbedded yellowish-brown fine-grained sandstone; lower part contains a few thin coal beds, common carbonaceous shale, and several thin oyster coquina beds; upper contact corresponds to the top of the Calico bed, a stacked series of fluvial channel deposits of white to light-gray, fine- to medium-grained sandstone and conglomeratic sandstone; Smoky Hollow strata are middle to upper Turonian on the basis of a diverse assemblage of mollusks, benthic foraminifera, and ostracods from exposures in Cedar Canyon (Eaton and others, 2001; Tibert and others, 2003); Moore and Straub (2001) assigned 313 feet (95 m) of strata in Cedar Canyon as likely equivalent to the Smoky Hollow Member; Eaton and others (2001) measured 364 feet (110 m) of strata that likely belong to the Smoky Hollow Member in Cedar Canyon, the lower 167 feet (54 m) of which were deposited in brackish-water environments and that are an order of magnitude thicker than equivalent brackish-water strata on the Kaiparowits Plateau, reflecting greater subsidence rates in the western part of the foredeep basin.

Moore and Straub (2001, their subunit 4 of interval A) suggested that the Calico bed may be present in Cedar Canyon about 285 feet (87 m) above the base of the formation. Subunit 4 is sandstone about 30 feet (9 m) thick but it is not distinctive in as much as it lacks pebbles and we were unable to use this bed as a marker horizon. However, we did locally map a thin pebble conglomerate about 330 feet (100 m) above the base of the formation on the Kolob Terrace in the southwest corner of the map area, which may be the Calico bed. Bobb (1991) noted that sediments of the Calico bed were deposited by braided and meandering streams that tapped both the Sevier orogenic belt and the Mogollon Highlands.

Ksj **John Henry Member** (Upper Cretaceous, Santonian to upper Turonian) – Slope-forming, variegated, gray, brown, and reddish-brown mudstone and thin- to thick-bedded, grayish-orange to yellowish-brown, fine-grained subarkosic sandstone; forms ledgy slopes; sandstone is commonly bioturbated and locally stained by iron-manganese oxides; stacked or amalgamated sandstone beds make up most of the upper part of the unit; upper contact corresponds to a break in slope at the base of the Drip Tank Member (on the Markagunt Plateau, formerly the lower conglomerate of the Grand Castle Formation); biotite from an ash bed about 800 feet (245 m) above the base of the member in Cedar Canyon yielded an $^{40}Ar/^{39}Ar$ age of 86.72 ± 0.58 Ma (late Coniacian), and biotite from an ash bed 700 feet (213 m) below the top of the member in Parowan Canyon yielded an $^{40}Ar/^{39}Ar$ age of 83.0 ± 1.1 Ma (early Campanian to late Santonian) (Eaton and others, 1999c); Eaton (2006) reported on mammal fossils in Cedar Canyon that suggest the lower part of the member there is late Turonian; about 900 to 1000 feet (275–300 m) thick in Cedar Canyon (Moore and Straub, 2001), 800 to 1100 feet (240–335 m) thick at Bryce Canyon National Park (Bowers, 1990), and about 670 feet (200 m) thick at the south end of the Paunsaugunt Plateau (Tilton, 2001a).

Ksjc **John Henry Member, lower part** (Upper Cretaceous, Santonian to upper Coniacian) – White to light-gray, fine- to medium-grained, cross-bedded sandstone and pebbly conglomerate likely equivalent to the Calico bed; mapped east of the Paunsaugunt fault north of Tropic Canyon; 80 to 100 feet (25–30 m) thick.

unconformity

Ksst **Smoky Hollow and Tibbet Canyon Members, undivided** (Upper Cretaceous, Turonian) – Undivided on the east flank of the Paunsaugunt Plateau following Bowers (1990), where the combined members are about 240 to 300 feet (70–90 m) thick. Smoky Hollow strata are described below and Tibbet Canyon strata are described separately.

Smoky Hollow strata are slope-forming, brown and gray mudstone, shale, and interbedded yellowish-brown fine-grained sandstone; lower part contains a few thin coal beds, common carbonaceous shale, and several thin oyster coquina beds; upper contact corresponds to the top of the Calico bed, a stacked series of fluvial channel deposits of white to light-gray, fine- to medium-grained sandstone and conglomeratic sandstone; Smoky Hollow strata are middle to upper Turonian on the basis of a diverse assemblage of mollusks, benthic foraminifera, and ostracods from exposures in Cedar Canyon (Eaton and others, 2001; Tibert and others, 2003); Titus and others (2013) reported nearly identical U-Pb and $^{40}Ar/^{39}Ar$ ages of 91.9 Ma for zircon and sanidine, respectively, recovered from a bentonitic ash bed near the top of the Smoky Hollow just southeast of the map area; not including the Calico bed, Bowers (1990) assigned 200 to 250 feet (60–75 m) of strata to the Smoky Hollow Member at Bryce Canyon National Park, and 80 to 100 feet (25–30 m) to the Calico bed itself; Tilton (2001a), however, reported that the entire Smoky Hollow Member is just 125 to 140 feet (40–45 m) thick at the south end of the Paunsaugunt Plateau.

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 minor interbedded, grayish-orange to gray mudstone and siltstone; locally contains pelecypods, gastropods, and thin to thick beds of oyster coquina; forms bold cliffs in Cedar Canyon and in the West and East Forks of Braffits Creek south of Summit, and a prominent though thinner cliff on the east flank of the Paunsaugunt Plateau; upper contact corresponds to a pronounced break in slope and is placed at the top of a coquinoid oyster bed and base of overlying thin coal and carbonaceous shale interval that caps the member; forms the riser of the Gray Cliffs part of the Grand Staircase; represents initial progradational (overall regressive) strata of the Greenhorn Cycle deposited in shoreface, beach, lagoonal, and estuarine environments adjacent to a coastal plain (Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003); thins eastward from about 650 to 800 feet (200–245 m) thick on the west flank of the Markagunt Plateau, 120 to 160 feet (37–50 m) thick at the south end of the Paunsaugunt Plateau (Tilton, 2001a, 2001b), and only 40 to 50 feet (12–15 m) thick on the east flank of the Paunsaugunt Plateau.

Ktd **Tropic Shale and Dakota Formation, undivided** (Upper Cretaceous, Turonian to Cenomanian) − Undivided in Cedar Canyon where the Tropic Shale is a few feet to at most a few tens of feet thick.

Kt **Tropic Shale** (Upper Cretaceous, Turonian to Cenomanian) − Gray to olive-gray, very thin bedded shale and silty shale in eastern exposures, but thins dramatically westward where it is dark-gray and yellowish-brown sandy mudstone, silty fine-grained sandstone, and minor shale; weathers to form badlands in the type area near Tropic; on the Markagunt Plateau, the base of the formation is locally characterized by a lag of septarian nodules, and Titus and others (2005) noted that the nodules are characteristic of the lower and middle parts of the formation east of the Paunsaugunt Plateau; numerous smectitic volcanic ash beds occur throughout the formation (Titus and others, 2005); contains locally abundant, well-preserved fossils, including ammonites, inoceramid and mytiloidid bivalves, gastropods, and oysters, as well as vertebrate fossils of sharks, fish, marine turtles, and plesiosaurs, all indicative of an open shallow-marine environment (see, for example, Eaton and others, 2001; Titus and others, 2005); on the Markagunt Plateau, typically poorly exposed, but forms subtle, vegetated slope at the base of the Straight Cliffs Formation and above the prominent “sugarledge sandstone” (Cashion, 1961) at the top of the Dakota Formation; upper, conformable contact placed at the base of the cliff-forming, planar beds of the Straight Cliffs Formation; deposited in shallow-marine environment dominated by fine-grained clastic sediment, marking the maximum incursion of the Western Interior Seaway (Tibert and others, 2003); age well constrained from ammonite and inoceramid biostratigraphy and dated interbedded volcanic ash beds (see, for example, Titus and others, 2013); thins westward across the map area, from about 700 feet (215 m) thick in the east near Tropic, to 40 feet (12 m) thick in the southwest part of the map area, to just a few feet thick in Cedar Canyon; Doelling and Willis (1999a) reported that it is 600 to 900 feet (180–275 m) thick in the southeast corner of the Panguitch 30′ x 60′ quadrangle.

Ktu **Tropic Shale, upper unit** (Upper Cretaceous, Turonian) − Gray and brown silty shale, mudstone, and fine-grained sandstone that forms an interval transitional to the overlying Tibbet Canyon Member of the Straight Cliffs Formation; mapped north of Tropic where it forms a light-colored slope nearly 200 feet (60 m) thick below the prominent Tibbet ledge.

Kd **Dakota Formation** (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 very thick bedded, fine to medium grained; includes several prominent cliff-forming sandstone beds each several tens of feet thick in the upper part of the formation, the upper one of which may correspond to the "sugarledge sandstone" of Cashion (1961); mudstone and claystone are gray to yellowish brown and commonly smectitic; the Dakota Formation contains two significant coal zones on the Paunsaugunt Plateau, the lower Bald Knoll zone and the upper Smirl zone (Doelling and Graham, 1972; Quick, 2010); oyster coquina beds, clams, and gastropods, including large *Craginia* sp., are common, especially in the upper part of the section; thin marl beds above the "sugarledge sandstone" locally contain small, distinctive gastropods with a beaded edge (*Admetopsis* n. sp., indicative of a latest Cenomanian brackish water environment [Eaton and others, 2001]); thins dramatically eastward, from about 1300 to 1400 feet (400–425 m) thick on the Markagunt Plateau at the south end of Jones Hill west of Maple Canyon, to about 80 to 275 feet (24–85 m) thick on the Paunsaugunt Plateau (Doelling and Willis, 1999a; Tilton, 2001a).

Dakota strata are typically poorly exposed and involved in large landslides on the Markagunt Plateau; they are the culprit in recent landslides that damaged Utah Highway 14 in Cedar Canyon (Lund and others, 2009, 2012). Most workers divide the Dakota Formation into three members, the lower one of which we re-assign to the Cedar Mountain Formation and the upper two of which we combine given the difficulty of mapping their mutual contact. The upper contact with the Tropic Shale is conformable. On the Markagunt Plateau, the contact corresponds to the top of the thin marl beds overlying the "sugarledge sandstone," whereas on the Paunsaugunt Plateau, it is placed at the top of the highest coal bed, which has locally burned, creating baked mudstone (clinker) at this horizon. The Dakota Formation is not correlative with the type Dakota in Nebraska, but the term is used loosely in Utah for deposits of an overall transgressive sequence below the Tropic Shale, the lower part of which was deposited in floodplain and river environments, whereas the upper part represents estuarine, lagoonal, and swamp environments of a coastal plain (Gustason, 1989; Eaton and others, 2001; Laurin and Sageman, 2001a, 2001b; Tibert and others, 2003). Gustason (1989), based in part on study of exposures in Cedar Canyon, correlated fluvial packages of the Dakota with orbital cycles of marine sedimentation of the deeper parts of the Western Interior Sea. Laurin and Sageman (2001a, 2001b) expanded on that work, constructing a high-resolution temporal and stratigraphic framework of middle Cretaceous marginal-marine deposits—they documented changes in shoreline position and also linked these changes to rhythmic, Milankovitch-driven deposition of marine limestone in the Western Interior Seaway. Invertebrate and palynomorph fossil assemblages in the Dakota indicate shallow-marine, brackish, and freshwater deposits of Cenomanian age (Nichols, 1997; Eaton, 2009); Dyman and others (2002) reported a Cenomanian $^{40}Ar/^{39}Ar$ age of 96.06 ± 0.30 Ma from middle Dakota strata about 2 miles (4 km) south of Tropic.

Kcm **Cedar Mountain Formation** (Cretaceous, Cenomanian to Albian) − Consists of a basal pebble conglomerate overlain by brightly colored variegated mudstone in Cedar Canyon. Mudstone is variegated gray, purplish-red, and reddish-brown, distinctly different from the gray and yellowish-brown hues of overlying Dakota strata; clay is smectitic and weathers to "popcorn-like" soils; includes minor light-gray to dark-yellowish-brown, fine- to medium-grained channel sandstone. Basal conglomerate is grayish brown and typically poorly cemented and non-resistant; clasts are subrounded to rounded, pebble- to small-cobble-size quartzite, chert, and limestone; red quartzite clasts are common; entire formation is about 60 feet (18 m) thick in Cedar Canyon, and the conglomerate ranges from less than one foot (0.3 m) to about 10 feet (3 m) thick.

Except for the thin conglomerate ledge at its base, weathers to generally poorly exposed slopes covered with debris from the overlying Dakota Formation. 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, but regionally, the Cedar Mountain Formation is unconformably overlain by the Dakota Formation (see, for example, Kirkland and others, 1997). Volcanic ash from correlative strata on the Kolob Plateau yielded a single-crystal $^{40}Ar/^{39}Ar$ age of 97.9 + 0.5 Ma on sanidine (Biek and Hylland, 2007), pollen analyses indicate an Albian or older age (Doelling and Davis, 1989; Hylland, 2010), and Dyman and others (2002) obtained an $^{40}Ar/^{39}Ar$ age of 101.7 + 0.42 Ma (latest Albian) on similar strata near Gunlock, Utah. Additionally, palynomorphs including rare occurrences of *Trilobosporites humilis* and possibly *Pseudoceratium regium*, collected from a thin mudstone interbed within basal conglomerate beds in Cedar Canyon immediately to the west of the map area (NW1/4 NW1/4 SE1/4 section 17, T. 36 S., R. 10 W., Cedar City 7.5′ quadrangle) suggest a late Albian age for this horizon (M.D. Hylland, Utah Geological Survey, unpublished data, November 9, 2001). The Cedar Mountain Formation was deposited in a floodplain environment of a broad coastal plain (Tschudy and others, 1984; Kirkland and others, 1997; Cifelli and others 1997; Kirkland and Madsen, 2007). This interval was previously mapped as the lower part of the Dakota Formation, but the lithology, age, and stratigraphic position of these beds suggest correlation to the Cedar Mountain Formation (Biek and Hylland, 2007; Biek and others, 2009). Specifically, the mudstone interval appears to be time-correlative with the Mussentuchit Member of the Cedar Mountain Formation of central and eastern Utah (Hylland, 2010).

Unconformity (K). No rocks of late Middle Jurassic to middle Early Cretaceous age are preserved in southwest Utah. This is because during this time, the back-bulge basin that developed in front of the Sevier orogenic belt had migrated eastward, and much of Utah was a forebulge high, a broad, gentle uplift that was high enough to undergo a prolonged period of modest erosion (see, for example, Willis, 1999). In this area, this 60-million-year-long gap in the rock record is marked by a bleached zone at the top of the Winsor Member of the Carmel Formation. The Cretaceous unconformity cuts down section to the west, where, on the south flank of the Pine Valley Mountains, first Winsor, then Paria River, and finally Crystal Creek strata are completely eroded away, so that at Gunlock the Cedar Mountain Formation rests upon the Co-op Creek Limestone, the lower member of the Carmel Formation (Biek and others, 2009).

JURASSIC

Je **Entrada Sandstone** (Middle Jurassic) – Consists of upper, middle, and lower members, which here are described following Doelling and Willis (1999a) but not mapped separately; the upper, Escalante Member is white, light-gray, pale-orange, and yellow-brown, fine- to coarse-grained, massive, high-angle cross-bedded, cliff-forming sandstone; the middle, Cannonville Member is mostly reddish-brown and gray silty sandstone and sandy siltstone that forms a banded slope; the lower, Gunsight Butte Member is mostly reddish-brown, fine-grained, cross-bedded, silty sandstone that commonly forms cliffs. As mapped here, Baer and Steed (2010) reported that the Escalante Member is 175 feet (54 m) thick, the Cannonville Member is 138 feet (42 m) thick, and the Gunsight Butte Member is 280 feet (85 m) thick in nearby Kodachrome Basin State Park; collectively, the formation there is about 530 feet (160 m) thick.

Thompson and Stokes (1970) noted that the Henrieville Sandstone, the type section of which is immediately east of the map area near Henrieville, is truncated westward under the Cretaceous unconformity; they envisioned that the Henrieville unconformably overlies the Escalante Member of the Entrada Sandstone and that it may correlate with the Salt Wash Member of the Morrison Formation. Peterson (1988) suggested and we concur that the Henrieville Sandstone is simply a bleached upper part of the Escalante Member of the Entrada Sandstone.

The Entrada Sandstone records deposition in tidal-flat, sabkha, and coastal-dune environments (Peterson, 1988, 1994). At and near Kodachrome Basin, the Gunsight Butte Member is renowned for its sedimentary breccia pipes, which are thought to have formed as fluid escape structures from overpressured, underlying Carmel strata in Middle Jurassic time, probably prior to deposition of the "Henrieville Sandstone" of Baer and Steed (2010).

Jct **Carmel and Temple Cap Formations, undivided** (Middle Jurassic, Bajocian to Aalenian) – Poorly exposed in fault blocks near Parowan Gap where it consists of light-gray micritic limestone and calcareous shale (Co-op Creek Limestone Member of the Carmel Formation) and reddish-brown mudstone and siltstone (Temple Cap Formation); also used to denote the entire Carmel and Temple Cap Formations on cross sections; Sprinkel and others (2011a) reported that the Temple Cap Formation ranges from about 173 to 171 Ma on the basis of several $^{40}Ar/^{39}Ar$ and U-Pb zircon ages; deposited in coastal-sabkha and tidal-flat environments (Blakey, 1994; Peterson, 1994); incomplete section about 30 feet (10 m) thick near Parowan Gap (Maldonado and Williams, 1993a).

Jcn **Carmel Formation** (Co-op Creek Limestone Member), **Temple Cap Formation, and Navajo Sandstone, undivided** (Middle to Lower Jurassic) – Poorly exposed in fault blocks near Parowan Gap where it consists of light-gray micritic limestone and calcareous shale (Co-op Creek Limestone Member of the Carmel Formation), reddish-brown mudstone and siltstone (Temple Cap Formation), and massively cross-bedded, light-gray or white sandstone that consists of well-rounded, fine- to medium-grained, frosted quartz (Navajo Sandstone); incomplete section several tens of feet thick.

Jc **Carmel Formation, undivided** (Middle Jurassic) – Undivided in the Parowan Gap area following Maldonado and Williams (1993a), where it is mostly light-gray micritic limestone and lesser yellowish-gray and reddish-brown fine- to medium-grained sandstone likely of the Co-op Creek Limestone and Crystal Creek Members; about 800 feet (250 m) thick.

Nomenclature of the Carmel Formation follows that of Doelling and Davis (1989), Sprinkel and others (2011a), and Doelling and others (2013). The Carmel Formation was deposited in a shallow inland sea of a back-bulge basin, and together with the underlying Temple Cap Formation, provides the first clear record of the effects of the Sevier orogeny in southwestern Utah (Sprinkel and others, 2011a). Middle Jurassic age is from Imlay (1980) and Sprinkel and others (2011a). Measured thicknesses in Cedar Canyon are from Doug Sprinkel (Utah Geological Survey, written communication, June 22, 2010).

Sprinkel and others (2011a) noted that several oil and gas exploration wells in the eastern part of the map area encountered Middle Jurassic salt at depth and so assigned those beds to the Twelvemile Canyon Member of the Arapien Formation. Carmel and Arapien strata thus interfinger at depth under the eastern Markagunt Plateau and Panguitch valley. For simplicity, we refer to this entire interval as Carmel Formation on the cross sections.

Pipiringos and O'Sullivan (1978) interpreted that Temple Cap and Carmel strata were separated by their J-2 unconformity, but new radiometric ages and palynomorph data suggest that the J-2 does not exist or is a very short hiatus in southern Utah (Sprinkel and others, 2011a; Doelling and others, 2013).

Jcw **Winsor Member** (Middle Jurassic, Callovian to Bathonian) – Light-reddish-brown, fine- to medium-grained sandstone and siltstone; uppermost beds typically bleached white under the Cretaceous unconformity; poorly cemented and so weathers to vegetated slopes, or, locally, badland topography; Baer and Steed (2010) separated the member into two informal units at Kodachrome Basin State Park, a lower sandstone and shale unit and an upper gypsiferous unit; upper contact on the Markagunt Plateau is at the base of a pebble conglomerate, which marks the Cretaceous unconformity, whereas it is conformably overlain by the Entrada Sandstone in the southeast part of the map area; deposited on a broad, sandy mud flat (Imlay, 1980; Blakey and others, 1983); 250 feet (75 m) thick in Cedar Canyon and 50 to 150 feet (15–45 m) thick in the southeast part of the Panguitch 30′ x 60′ quadrangle (Doelling and Willis, 1999a).

Jcp **Paria River Member** (Middle Jurassic, Bathonian) – Consists of three parts not mapped separately: (1) upper part is about 50 feet (15 m) of cliff-forming, olive-gray, micritic and argillaceous limestone and calcareous mudstone; laminated in very thick beds; locally contains small pelecypod fossils; (2) middle part is about 20 feet (6 m) of reddish-brown and greenish-gray shale that forms slope; and (3) lower part is gypsum and minor interbedded shale as much as 80 feet (25 m) thick in nodular, highly fractured and contorted beds and as thin, laminated beds. Upper contact is sharp and planar; deposited in shallow-marine and coastal-sabkha environments (Imlay, 1980; Blakey and others, 1983); Sprinkel and others (2011a) reported an $^{40}Ar/^{39}Ar$ age on zircon from a volcanic ash of 165.9 ± 0.51 Ma on lower Paria River strata in south-central Utah; 173 feet (53 m) thick in Cedar Canyon and 150 to 400 feet (45–120 m) thick in the southeast part of the Panguitch 30´ x 60´ quadrangle (Doelling and Willis, 1999a).

Jcx **Crystal Creek Member** (Middle Jurassic, Bathonian) – Thin- to medium-bedded, reddish-brown siltstone, mudstone, and fine- to medium-grained sandstone; commonly gypsiferous and contains local contorted pods of gypsum; forms vegetated, poorly exposed slopes; upper contact is sharp and broadly wavy and corresponds to the base of the thick Paria River gypsum bed; Kowallis and others (2001) reported two $^{40}Ar/^{39}Ar$ ages of 166 to 167 Ma 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); 294 feet (90 m) thick in Cedar Canyon.

Jcc **Co-op Creek Limestone Member** (Middle Jurassic, Bajocian) − Thin- to medium-bedded, light-gray micritic limestone and calcareous shale; locally contains *Isocrinus* sp. crinoid columnals, pelecypods, and gastropods; forms sparsely vegetated, ledgy slopes and cliffs; Kowallis and others (2001) reported several $^{40}Ar/^{39}Ar$ ages of 167 to 168 Ma for altered volcanic ash beds within the lower part of the member in southwest Utah that were likely derived from a magmatic arc in what is now southern California and western Nevada; Sprinkel and others (2011a) reported $^{40}Ar/^{39}Ar$ ages of 169.2 ± 0.51 Ma and 169.9 ± 0.49 Ma on two ash beds in the lower part of the member in southwestern Utah; deposited in a shallow-marine environment (Imlay, 1980; Blakey and others, 1983); probably about 400 feet (120 m) thick; the member is as much as about 350 feet (105 m) thick on the Kolob Terrace north of Zion National Park (Biek and Hylland, 2007).

Temple Cap Formation (Middle to Lower Jurassic) – Not mapped separately; Maldonado and Williams (1993a) mapped reddish-brown, highly sheared, fine- to medium-grained sandstone east of The Narrows at Parowan Gap as an incomplete section of Temple Cap strata; see Biek and others (2009) and Sprinkel and others (2011a) for a description of this formation in southwesternmost Utah.

Unconformity (J-1). The J-1 unconformity of Pipiringos and O'Sullivan (1978), formed prior to 173 million years ago in southwest Utah (Sprinkel and others, 2011a).

Jn **Navajo Sandstone** (Lower Jurassic) − Massively cross-bedded, poorly to moderately well-cemented, light-gray or white sandstone that consists of well-rounded, fine- to medium-grained, frosted quartz sand; upper, unconformable contact is sharp and planar and regionally corresponds to a prominent break in slope, with cliff-forming, cross-bedded sandstone below and reddish-brown mudstone 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), part of the world's largest coastal and inland paleodune field (Milligan, 2012); correlative in part 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; Sprinkel, 2009; Sprinkel and others, 2011b); 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 Ga) crust involved in Appalachian orogenesis of eastern North America (Dickinson and Gehrels, 2003, 2009a, 2009b; Rahl and others, 2003; Reiners and others, 2005); Mohn (1986) measured an incomplete section at Parowan Gap of 1090 feet (332 m); the entire formation is about 1800 to 2300 feet (550–700 m) thick in southwest Utah (Biek and others, 2009).

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Plate 1
Utah Geological Survey Map 270DM
Geologic Map of the Panguitch 30' x 60' Quadrangle

GEOLOGIC MAP OF THE PANGUITCH 30' X 60' QUADRANGLE, GARFIELD, IRON, AND KANE COUNTIES, UTAH

2015

by

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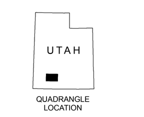
SCALE 1:62,500

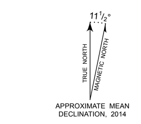
1 0.5 0 1 2 3 4 MILES

1 0.5 0 1 2 3 4 5 6 KILOMETERS

CONTOUR INTERVAL 50 METERS



QUADRANGLE LOCATION



APPROXIMATE MEAN DECLINATION, 2014

Disclaimer

This geologic map was funded by the Utah Geological Survey and the U.S. Geological Survey, National Cooperative Geologic Mapping Program through USGS STATEMAP award numbers 08HQAG0096, G09AC00152, G10AC00386, and G11AC20249. The views and conclusions contained in this document are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the U.S. Government.

Base from USGS Panguitch 30' x 60' Quadrangle (1980)
Projection: UTM Zone 12
Datum: NAD 1927
Spheroid: Clarke 1866

Project Manager: Douglas A. Sprinkel

GIS and Cartography: Jay C. Hill, and Basia Matyjasik

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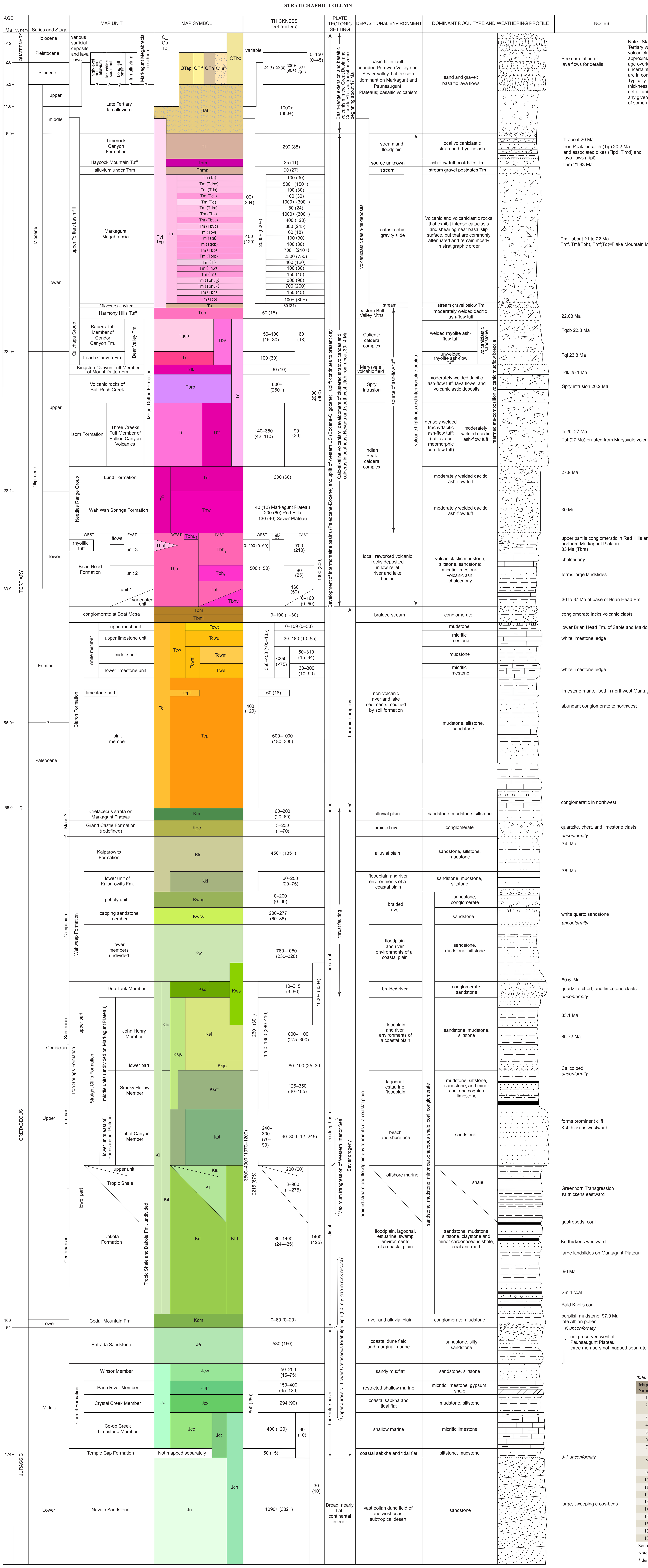
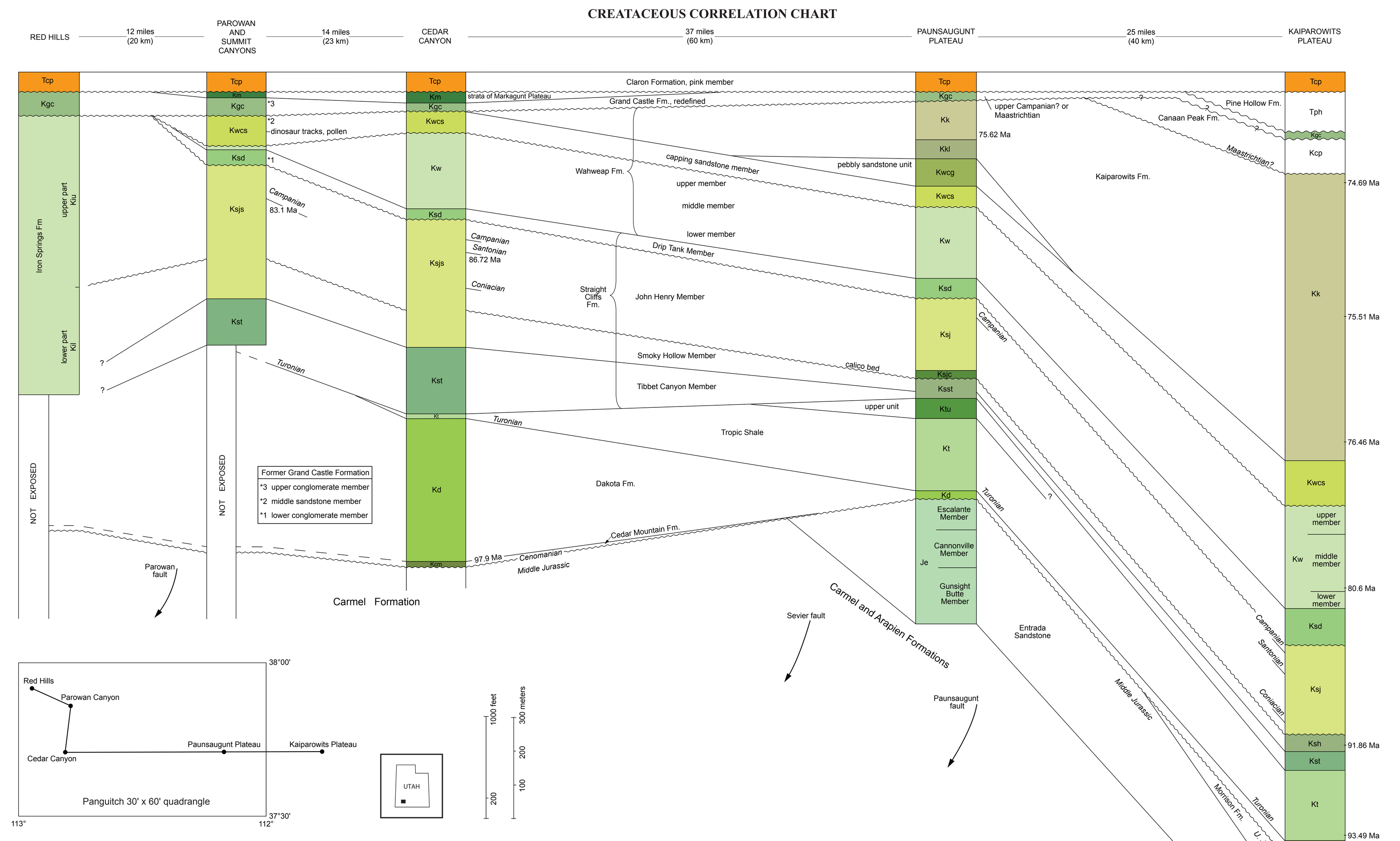
P.O. Box 146100, Salt Lake City, UT 84114-6100
(801) 537-3300
geology.utah.gov

STRATIGRAPHIC COLUMN



CREATACEOUS CORRELATION CHART



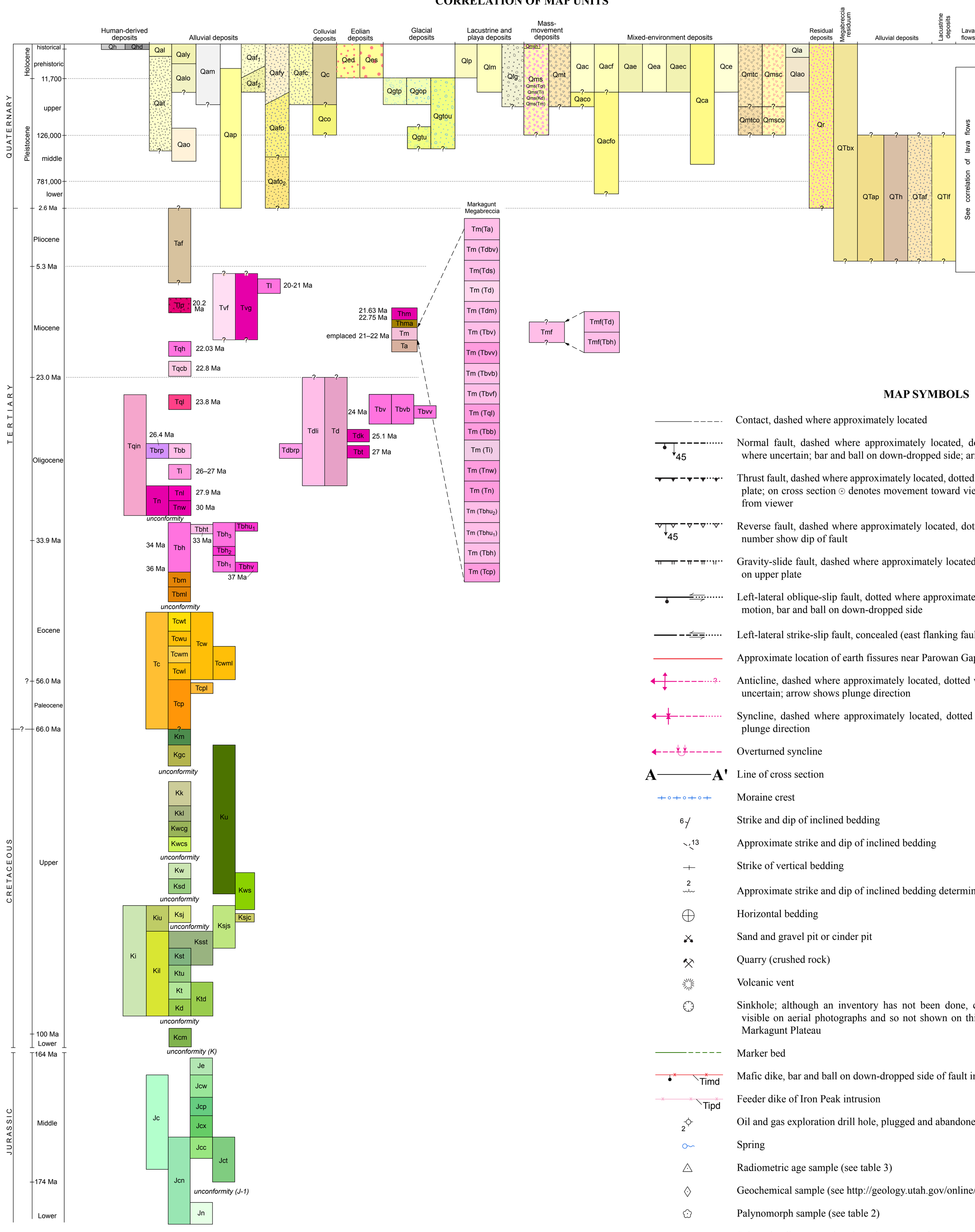
Correlation of Upper Cretaceous strata across the Panguitch 30' x 60' quadrangle and eastward into the Kaiparowits Plateau. The Grand Castle Formation of Goldstrand and Mullett (1997) is here redefined and restricted to their upper conglomerate member. Our mapping shows that their lower conglomerate member of the Grand Castle Formation is in fact the Drip Tank Member of the Straight Cliffs Formation (as originally suggested by Eaton and others, 2001; Moore and Straub, 2001; Lawton and others, 2003; and Eaton, 2006), and that their middle sandstone member of the Grand Castle Formation is in fact the capping sandstone member of the Wahweap Formation (as originally suggested by Pollock, 1999; and Lawton and others, 2003).

As much as about 200 feet (60 m) of strata are present above the Grand Castle Formation (redefined) and below the Claron Formation on the Markagunt Plateau, here assigned to an informal unit, Cretaceous strata of the Markagunt Plateau (Km), of poorly constrained Upper Cretaceous age. Lithologically similar strata are as much as about 400 feet (120 m) thick on the Paunsaugunt Plateau, but there they underlie the Kaiparowits Formation and are mapped as an informal lower unit of the Kaiparowits Formation (Kkl).

Note that newly recognized Kaiparowits Formation is preserved below the Sunset Cliffs on the west flank of the Paunsaugunt Plateau, and immediately east of the Paunsaugunt fault near the plateau's northeast margin.

Radiometric ages of volcanic ash beds are shown on right side of columns; a short line marks the base of Upper Cretaceous stage names.

CORRELATION OF MAP UNITS



MAP SYMBOLS

- Contact, dashed where approximately located
- Normal fault, dashed where approximately located, dotted where concealed; queried where uncertain; bar and ball on down-dropped side; arrow and number show dip of fault
- Thrust fault, dashed where approximately located, dotted where concealed; teeth on upper plate; on cross section ⊙ denotes movement toward viewer, ⊕denotes movement away from viewer
- Reverse fault, dashed where approximately located, dotted where concealed; arrow and number show dip of fault
- Gravity-slide fault, dashed where approximately located, dotted where concealed; barbs on upper plate
- Left-lateral oblique-slip fault, dotted where approximately located, arrows show relative motion, bar and ball on down-dropped side
- Left-lateral strike-slip fault, concealed (east flanking fault of Markagunt gravity slide)
- Approximate location of earth fissures near Parowan Gap (see DuRoss and Kirby, 2004)
- Anticline, dashed where approximately located, dotted where concealed, queried where uncertain; arrow shows plunge direction
- Syncline, dashed where approximately located, dotted where concealed; arrow shows plunge direction
- Overturned syncline
- A——A' Line of cross section
- Moraine crest
- 6 Strike and dip of inclined bedding
- 13 Approximate strike and dip of inclined bedding
- Strike of vertical bedding
- 2 Approximate strike and dip of inclined bedding determined photogrammetrically
- Horizontal bedding
- Sand and gravel pit or cinder pit
- Quarry (crushed rock)
- Volcanic vent
- Sinkhole; although an inventory has not been done, countless sinkholes, not readily visible on aerial photographs and so not shown on this map, are known to be on the Markagunt Plateau
- Marker bed
- Timd Mafic dike, bar and ball on down-dropped side of fault intruded by dike
- Tipd Feeder dike of Iron Peak intrusion
- 2 Oil and gas exploration drill hole, plugged and abandoned; number refers to table 4
- Spring
- Radiometric age sample (see table 3)
- Geochemical sample (see http://geology.utah.gov/online/analytical_data.htm)
- Palynomorph sample (see table 2)
- Selected sedimentary pipes in and near Kodachrome Basin State Park (see Baer and Steed, 2010)

CORRELATION OF LAVA FLOWS

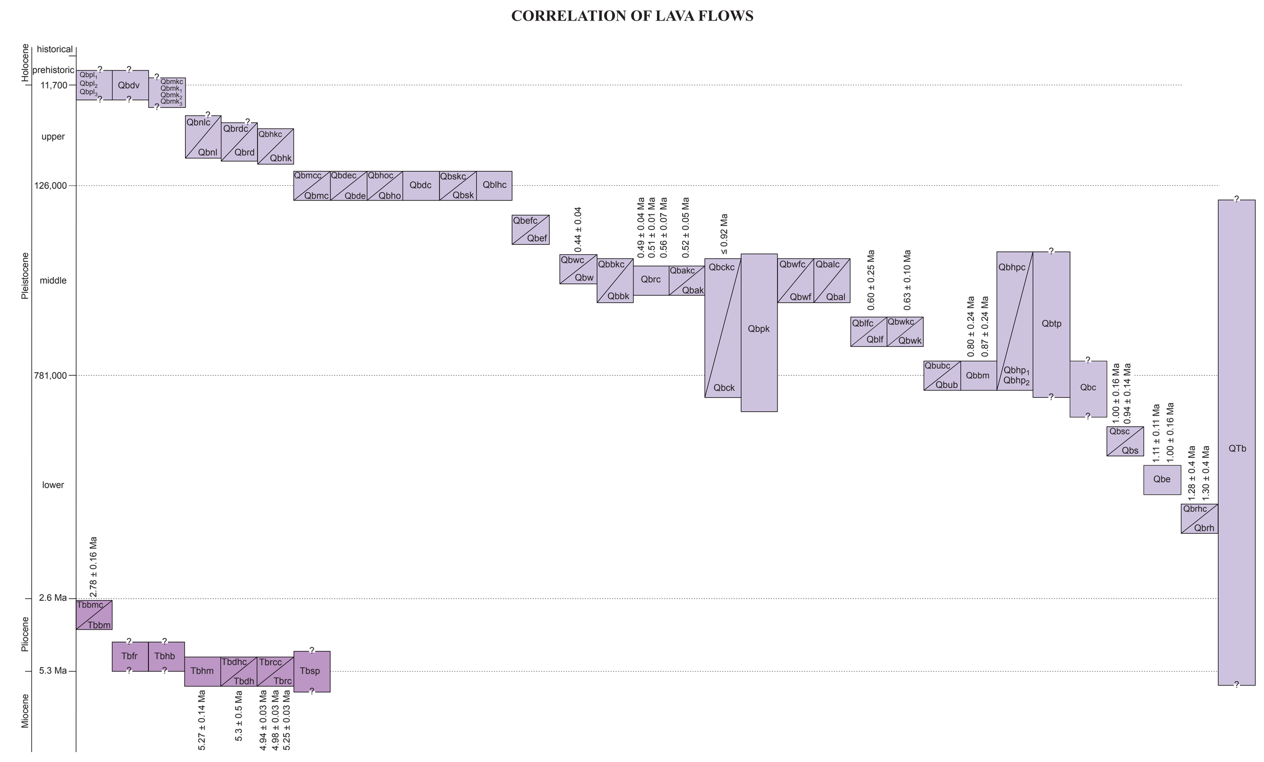


Table 4. *Wildcat exploration drill holes, all plugged and abandoned, in the Panguitch 30' x 60' quadrangle.*

Map ID Number	API Well Number	Operator	Well Name	Year Abandoned	County	7.5' Quadrangle	Qtr/Qtr	Section	Township-Range	Ft. NS*	NS	Ft. EW*	EW	UTM Eastings	UTM Northings	Latitude	Longitude	TD (feet)
1	43-021-30009	DELTA PETROLEUM CORP	FEDERAL 23-44	2007	IRON	Parowan Gap	SESE	23	33S-10W	1062	S	1073	E	329132	4198227	37.91555	-112.94384	12094
2	43-021-10758	MOUNTAIN FUEL SUPPLY CO	LITTLE SALT LAKE GOVT 1 (WSW)	1963	IRON	Summit	NESE	9	34S-10W	2310	S	330	E	325975	4192037	37.8592	-112.97825	4400
3	43-021-30011	CEDAR MOUNTAIN GAS LLC	CLARK 1-28	2009	IRON	Webster Flat	SWNW	28	37S-10W	665	S	655	W	323920	4157241	37.54721	-112.99318	2017
4	43-017-30068	MAY PETROLEUM INC	PANGUITCH 1	1978	GARFIELD	Panguitch Lake	NESW	33	36S-7W	2546	S	2171	W	355193	4166654	37.63746	-112.64119	11610
5	43-017-10902	PHILLIPS PETROLEUM CO	HATCH FEE B-1	1963	GARFIELD	Asay Bench	SESE	2	37S-7W	820	S	549	E	359150	4164461	37.61832	-112.59593	4971
6	43-017-10901	PHILLIPS PETROLEUM CO	HATCH FEE A-1	1964	GARFIELD	Asay Bench	NESE	16	37S-6W	1980	S	660	E	365454	4161419	37.59185	-112.52397	4917
7	43-017-30106	NORTHWEST EXPLORATION CO	PANGUITCH 1	1981	GARFIELD	Panguitch	SWNW	7	35S-5W	2287	N	670	W	370747	4182717	37.78267	-112.46781	6752
8	43-017-30104	TEXAS INTERNATIONAL PETRO	DIXIE UNIT 1-19	1981	GARFIELD	Casto Canyon	NESW	19	34S-4.5W	1391	S	1910	W	380978	4188520	37.83635	-112.35261	14075
9	43-017-30115	ARCO OIL & GAS COMPANY	DIXIE UNIT 2	1982	GARFIELD	Blind Spring Mtn.	SWSW	2	33S-4.5W	960	S	1060	W	386527	4202827	37.96598	-112.29182	15251
10	43-017-30146	FELLOWS ENERGY LTD	10-33C2	2004	GARFIELD	Cow Creek	NWSE	33	33S-2W	1986	S	1782	E	411273	4194102	37.89011	-112.00907	1716
11	43-017-30112	SOUTH LOUISIANA PROD CO	CLAY CREEK FEDERAL 13-29	1982	GARFIELD	Flake Mtn. East	SWSW	29	34S-2W	660	S	660	W	408711	4185659	37.81377	-112.03714	12894
12	43-017-30084	CHAMPLIN PETROLEUM CO	CLAY CREEK FED 11-32	1982	GARFIELD	Flake Mtn. East	NWNW	32	34S-2W	660	N	661	W	408693	4185257	37.81014	-112.03729	8785
13	43-017-10375	FOREST OIL CORP	GOVT 1 (RED CANYON)	1959	GARFIELD	Wilson Peak	SENE	1	36S-4.5W	2970	S	990	E	387227	4173692	37.70353	-112.27931	6242
14	43-017-10662	LION OIL COMPANY	BRYCE-MONSANTO 1	1957	GARFIELD	Bryce Canyon	NWNW	10	36S-4W	990	N	990	W	392688	4172411	37.69264	-112.21718	11221
15	43-017-20350	BRYCE CANYON OIL CO	1	1928	GARFIELD	Tropic Canyon	NENE	36	36S-3W	2640	N	2640	E	406076	4165090	37.62998	-112.06441	400
16	43-017-30117	U R C CORPORATION	TROPIC STATE 1-29	1983	GARFIELD	Cannonville	NWNE	29	37S-2W	660	N	1980	E	409288	4157825	37.56299	-112.02712	6260
17	43-025-30009	TEXACO INC	GOVT SCHNEIDER A NCT-1	1971	KANE	Cannonville	SESW	5	38S-2W	1320	S	1320	W	408655	4153600	37.52485	-112.03376	5958
18	43-017-30126	ARCO OIL & GAS COMPANY	BUCK KNOLL U 1	1985	GARFIELD	Tropic Reservoir	SESW	36	37S-4.5W	880	S	1885	W	386488	4155281	37.53938	-112.28483	10119

Source: Utah Division of Oil, Gas, and Mining; http://oilgas.ogm.utah.gov/Data_Center/LiveData_Search/well_data_lookup.cfm; accessed March 13, 2012.

Note: Location of wells 9, 15, and 17 as reported above are slightly off true location, which is shown on plate 1.

* denotes feet north, south, east, or west from section boundary

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Plate 3
Utah Geological Survey Map 270DM
Geologic Map of the Panguitch 30' x 60' Quadrangle