2015-07-01

Geologic Mapping of Ice Cave Peak Quadrangle, Uintah and Duchesne Counties, Utah with Implications from Mapping Laramide Faults

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Duchesne Counties, Utah with Implications

from Mapping Laramide Faults

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A thesis submitted to the faculty of
Brigham Young University
in partial fulfillment of the requirements for the degree of

Master of Science

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September 2015

ABSTRACT

Geologic Mapping of Ice Cave Peak Quadrangle, Uintah and
Duchesne Counties, Utah with Implications
from Mapping Laramide Faults

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Geologic mapping (1:24,000 scale) of the Ice Cave Peak quadrangle, Uintah and
Duchesne Counties, Utah has produced a better understanding of the geologic structures present
in the quadrangle and has increased our understanding of faulting in northeastern Utah. Map
units in the quadrangle range in age from late Neoproterozoic to Quaternary and include good
exposures of Paleozoic rocks (Mississippian to Permian), limited exposures of Mesozoic rocks,
and good exposures of Tertiary strata (Duchesne River Formation and Bishop Conglomerate)
deposited during uplift of the Uinta Mountains.

Lower Mississippian strata along the south flank of the Uinta Mountains have typically
been mapped as Madison Limestone. Our preliminary mapping suggested that the Madison could
perhaps be subdivided into an upper unit equivalent to the Deseret Limestone, and a lower unit
separated by a phosphatic interval equivalent to the Delle Phosphatic Member of the Deseret
Limestone found farther west. Upon further investigation, we propose not extending the use of
Deseret Limestone, with the equivalent to the Delle Phosphatic Member at its base, into the
south-central Uinta Mountains. Microprobe analysis revealed no phosphorus in thin sections of
this unit. Instead, the unit is composed almost entirely of calcite and dolomite.

A zone of northwest-trending faults, called the Deep Creek fault zone, occurs mainly east
of the Ice Cave Peak quadrangle. However, our mapping shows that this fault zone extends into
the quadrangle. These faults are both strike-slip and normal/oblique faults as documented by
mapping and kinematic indicators and cut the folded hanging-wall sedimentary rocks above the
Uinta Basin-Mountain boundary thrust fault. These faults may be part of an en echelon fault
system that is rooted in the Neoproterozoic and reactivated during Laramide deformation above a
possible transfer zone between segments of the buried boundary thrust.

Keywords: Uinta Mountains, Laramide faulting, Ice Cave Peak

ACKNOWLEDGEMENTS

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Introduction

The southern flank of the Uinta Mountains located approximately 17 miles northeast of Roosevelt, Utah is home to beautiful rock exposures that range in age from Neoproterozoic to Quaternary. Excellent exposures of Paleozoic rocks (Mississippian to Permian), limited exposures of Mesozoic rocks, and good exposures of Tertiary strata (Duchesne River Formation and Bishop Conglomerate) produced by uplift of the Uinta Mountains come together to make the beautiful scenery that occurs within the Ice Cave Peak 7.5' quadrangle (Figure 1).

The study area is bordered on the north by the South Flank fault zone (Figure 2) and on the south by the Uinta Basin-Mountain boundary fault (Haddox et al., 2005). The Deep Creek fault zone (Untermann and Untermann, 1969) has previously been mapped to the east of the study area, however, as a result of this mapping project it has been extended farther west into the Ice Cave Peak quadrangle.

Other smaller-scale mapping projects have included the Ice Cave Peak quadrangle as part of their mapping (Kinney, 1955, Sprinkel, 2006). A 1:24,000 scale geologic map of the Ice Cave Peak 7.5' quadrangle with accompanying unit descriptions, stratigraphic column, and cross section has been produced to further understand the region. The breadth of this project will address the nature of the exposed units within the study area, the stratigraphy, the subsurface geology, the structures that are present within the field area, the economic potential, and the geologic hazards of the Ice Cave Peak quadrangle. During this process new discoveries were made that have local and regional implications, which will also be discussed in the following chapters.

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The Deep Creek fault zone displays a variety of different fault types (Untermann and Untermann, 1969; Haddox, 2005). How do these different fault types form in one fault zone? Does this represent different periods of faulting? Or can they be explained within a single phase and system of faulting? These questions are discussed and possible solutions proposed in the following chapters.

Stratigraphic questions related to the Ice Cave Peak quadrangle include: What units are exposed within the study area? Can we clarify stratigraphic nomenclature for this region? Which members of the Duchesne River Formation, are exposed in the quadrangle?

Economic questions are addressed for the region. What is the economic potential for the tar sands mine within the field area? Has the geology within Ice Cave Peak quadrangle produced natural resources that could benefit the region? How can regional implications that result from this study be used for economic benefit in other localities?

Lastly, geologic hazards are discussed for the study area and region. The moderate risk of landslide due to the Bishop Conglomerate could threaten some homes. Earthquakes, flooding, and rock-falls all pose a minor risk and are discussed further within the following chapters.

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Chapter 1: Mapping Ice Cave Peak 7.5' quadrangle

<u>Introduction</u>

Ice Cave Peak quadrangle is located on the southeastern flank of the Uinta Mountains at the northern margin of the Uinta Basin. The Uinta Basin-Mountain- boundary fault is a blind thrust fault that boarders the southern flank of the Uinta uplift. It has been mapped as either underneath the southern edge of the quadrangle or just south of the quadrangle (Sprinkel, 2006). The quadrangle is named for the beautiful and prominent Ice Cave Peak, which is found on the east side of Whiterocks Canyon high above the river. The core peak is made up of Mississippian Madison Limestone. Caves occur in the Paleozoic carbonate formations within the field area. The Ice Cave Peak quadrangle has both rugged mountains and valleys in the north and the characteristically flatter topography of the Uinta Basin to the south.

Elevation in the quadrangle ranges from above 9,800 feet in the north to 6,320 feet in the south. There are two main canyons that run north and south within the quadrangle. The larger, Whiterocks Canyon is to the east and contains the Whiterocks River. The smaller Farm Creek Canyon is located to the west and has been formed by Farm Creek. The southern portion of the Ice Cave Peak quadrangle is located on the Uinta and Ouray Indian Reservation and is populated with a few homes. About six homes exist, in Whiterocks Canyon, which are not on reservation land. The northern portion of the quadrangle is located within Ashley National Forest and is a popular location for wilderness recreation, fishing, and hunting.

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Traditional geologic field mapping methods combined with state of the art ArcGIS and Cardinal Systems VR-One and VR-Two 3-D mapping software were used for mapping. These software packages also assisted in the production of the plates, figures, and diagrams included herein.

Stratigraphy

Map units in the Ice Cave Peak quadrangle range in age from late Neoproterozoic to Quaternary. The field area includes good exposures of Paleozoic rocks (Mississippian to Permian), limited exposures of Mesozoic rocks, and good exposures of Tertiary strata (Duchesne River Formation and Bishop Conglomerate) resulting from uplift and erosion of the Uinta Mountains. Pre-Quaternary strata in the quadrangle typically dip to the southwest. Laramide faulting associated with the Uinta Basin-Mountain boundary thrust and the South Flank fault zone, which flank the quadrangle to the south and north respectively, are the main cause of the tilting of these strata (Haddox, 2005; Sprinkel, 2006).

Neoproterozoic, Red Pine Shale (Zur)

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Neoproterozoic rocks within the Ice Cave Peak quadrangle comprise the Red Pine Shale, which is the uppermost formation of the Uinta Mountain Group (Gregson and Chure, 2000). The Red Pine Shale was deposited during a period of time when this region was actively rifting away from Laurentia and was juxtaposed next to what would eventually become the Pacific Ocean (Hintze and Kowallis, 2009). Red Pine Shale deposits represent a marine deltaic system that was supplied with mature sediments from the east and immature sediments from the north (Myer, 2008).

The Red Pine Shale is a silty shale interbedded with silty sandstone. It displays a dark-green-gray color on fresh surfaces and weathers brown-gray to reddish-brown, purple to pale pink, and pale olive. Shale units within the Red Pine Shale are composed of over fifty percent clay particles. Shale units are very thin- to medium-bedded with ripple and loading structures present. The shale units are slope-formers (especially near faulted zones) and poorly cemented. Sandstone units within the Red Pine Shale are fine to coarse-grained. They are poorly to well sorted, and very thick to medium-bedded. The sandstone units are friable to well cemented and quartz-rich to arkosic. Cross-bedding is present and the sandstone units tend to be ledge formers. The Red Pine Shale is 1800 to 4000 feet thick.

The shale is organic-rich, ranging from 1-5.9% TOC (Dehler and Sprinkel, 2006; Myer, 2008) and may contain over 10% TOC in some areas (Sprinkel, personal communication).

Red Pine Shale is the oldest exposed formation within the Ice Cave Peak quadrangle with an age of 740– 750 Ma (Dehler and Sprinkel, 2006). It is exposed in the northern portion of the Ice Cave Peak quadrangle and is usually capped by Mississippian Madison Limestone (Mm) or by middle Pleistocene-aged piedmont gravels (Qap3) (Sprinkel, 2006). Good exposures of the

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formation can be found along the Whiterocks River as well as at higher elevations. The Red Pine Shale is susceptible to mass movement, with landslides, slumps, and debris flows (Qms) common. A significant angular unconformity separates it from the overlying Madison Limestone.

Mississippian System

The Madison Limestone (Middle to Lower Mississippian), Doughnut and Humbug Formations (Upper Mississippian) comprise the exposed Mississipian rocks within the Ice Cave Peak quadrangle. They represent carbonate-rich shelf deposits to marginal-marine and nearshore depositional environments (Sandberg et al., 1982; Haddox, 2005).

Madison Limestone (Mm)

The Madison Limestone is a wackestone to packstone that is dark-gray to light-gray on fresh surfaces. The Madison Limestone weathers dark-gray. Outcrops are coarse- to medium-crystalline, massive, and contain an abundance of chert. The chert nodules are dark to light gray and pebble- to cobble-sized. There are localized areas of angular to subangular brecciated limestone clasts enclosed in a crystalline matrix of dark gray limestone (near Ice Cave Peak). The Madison Limestone is approxiately 1100 feet thick.

The Madison Limestone is well exposed in the northern half of the Ice Cave Peak quadrangle. It forms the dominant topographic feature of the area, making up the skyline along much of Whiterocks Canyon. The Madison Limestone typically dips to the south-southeast at an average of 55°. The formation is overlain by the Humbug Formation, and is underlain by the Red

Pine Shale (Kinney, 1955; Gregson and Chure, 2000). The Madison Limestone is predominately a resistant, cliff-forming unit and in places develops karst topography.

An attempt was made early in the mapping process to separate the Madison Limestone Formation into two separate Mississippian units, the Deseret Limestone underlain by the Madison Limestone. The reason for contemplating this division relates to the inconsistent use of nomenclature for these Mississippian rocks in the eastern Uintas. Some of the early workers in the region mapped them as the Madison Limestone overlain by the Deseret Limestone (Huddle and McCann, 1949; Baker et al., 1949). Later in 1955 the region was mapped as Deseret Limestone, including its basal member, the Delle Phosphatic Shale, overlying the Madison Limestone (Kinney, 1955). Even recently they have been mapped as one unit, Madison Limestone (Sprinkel et al., 2000), and as two units, Deseret Limestone and Madison Limestone (Haddox et al., 2010).

Where the Delle Phosphatic Member is mapped in other localities it has several unique indicative features; that include: 1) phosphate, 2) siliceous microfossils, 3) secondary chert, 4) typically a dense lime mudstone, or phosphatic mircofossils and 5) is stratigraphically sandwiched between subtidal and peritidal Mississippian shelf carbonates (Nichols and Silberling, 1990).

A detailed section was measured and multiple thin sections were examined to determine the character of the suspected Delle Phosphate Member in the Ice Cave Peak quadrangle, including polished thin sections for electron microprobe analysis. We examined these polished thin sections for Ca, P, Si, K and other elements and found the samples to be almost entirely

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calcite and dolomite and to contain no phosphorus. We did, however, find the presence of several small flecks of gold.

The goal of this part of our study was to find a phosphate-rich zone that would have correlated to the Delle Phosphatic Member found elsewhere at the base of the Deseret Limestone. This would have allowed for the separation of the current Madison Limestone into an upper Deseret Limestone and lower Madison Limestone. The target strata that we studied displayed many of the characteristic features of the Delle Phosphatic Member. Field observations yielded secondary chert nodules, and our measured section included an argillaceous, recessive unit that is stratigraphically sandwiched between more resistant carbonate rocks. Thin section investigation revealed siliceous microfossils in the form of sponge spicules and small lamina within the lime mudstones. These characteristics might indicate that these rocks are equivalent to the Delle Phosphatic Member of the Deseret Limestone; however, no phosphate was present in any of the thin sections. Due to the lack of phosphate, the key indicator for the Delle Phosphatic Member, we are not confident that the upper part of the formation is equivalent to the Deseret Limestone and conclude that the Madison Limestone should not be separated into two units until further evidence is found.

Humbug Formation (Mh)

The Humbug Formation is a sandstone interbedded with limestone and shale. The sandstones are dark-yellowish-orange to red in color on fresh surfaces and weather to a black or orange color. The sandstones are also very fine-grained to fine-grained, locally cross-bedded,

and hematitic. The Humbug Formation contains some shallow caves and sinkholes. The limestone units within the Humbug Formation are mudstones to packstones. They are light-gray on fresh surface and weather to a dark-gray. The shale units are slope-forming and covered by vegetation making an accurate description difficult. The Humbug Formation is 150 to 250 feet thick.

The Humbug Formation is stratigraphically positioned between the Madison Limestone and the Doughnut Shale. It is a slightly weaker, more recessive formation that is partially overlain by landslide material (Qms) within the Ice Cave Peak quadrangle. The Humbug Formation, combined with the much weaker Doughnut Shale make up east-west trending, strike valleys on either side of the Whiterocks Canyon with outcrops of the Humbug Formation exposed on the northern slopes of those valleys.

Doughnut Shale (Md)

The Doughnut Shale is entirely covered by Quaternary colluvium and alluvium within the Ice Cave Peak quadrangle and thus cannot be characterized from exposures within the quadrangle. However, in nearby regions the Doughnut Shale is a dark-gray, organic-rich, shale with reddish shale near its base. There are units of coarse sandstone and limestone interbedded with the shale (Sprinkel, 2006). The Doughnut Shale is 80 to 300 feet thick in neighboring quadrangles and by measurement of the landslide material that covers it within Ice Cave Peak quadrangle.

Pennsylvanian System

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The Round Valley Limestone (Early and Middle Pennsylvanian), Morgan Formation (Middle Pennsylvanian), and the Weber Sandstone (Middle Pennsylvanian to Permian) make up the Pennsylvanian system exposed within the Ice Cave Peak quadrangle. All three of these formations are thought to have been deposited in beach or shallow subtidal depositional environments (Welsh and Bissell, 1979).

Round Valley Limestone (IPr)

The Round Valley Limestone consists of limestone beds interbedded with shale and mudstone beds. The limestones are wackestones to packstones. They are light-blue-gray in color on fresh surfaces and weather dark-bluish-gray. The Round Valley Limestone contains abundant fossils (brachiopods, crinoids, etc.) and red, blue-gray, and gray chert nodules. The shale and mudstone units are gray to green and purple to red in color. These units are thinly bedded. The Round Valley Limestone is 260 feet thick.

The Round Valley Limestone, like most exposed units within the Ice Cave Peak quadrangle, strikes predominately east-west and dips about 40° – 50° to the south. It is best exposed in the center of the quadrangle where it forms a north facing cliff above the colluvium and landslide material that overlies the Doughnut Shale Formation. The best exposures within the quadrangle can only be accessed by ATV or hiking in from roads. The limestone cliffs of the Round Valley Limestone have a minor element of folding in Farm Creek Canyon (see Structural Geology section).

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Morgan Formation (IPm)

The Morgan Formation is a sandstone interbedded with, relatively thin beds (2-15 feet thick), of limestone and silty shale. The sandstone is light-red-gray in color. The sandstone units are medium- to fine-grained, and locally cross-bedded. The sandstone units are ledge-forming in both Farm Creek and Whiterocks Canyons (Plate 1). The limestone units are wackestone, light- to medium-gray in color on fresh surfaces. They usually weather red, due to staining from the overlying reddish sandstone beds. The limestone units are fossiliferous and contain gray to red chert. The interbedded silty shale is light-red to gray and the silts are very fine-grained. The shale is thinly bedded and slope-forming. The Morgan Formation is 500 to 950 feet thick.

The Morgan Formation is a widely exposed formation within the Ice Cave Peak quadrangle with excellent outcrops occurring in both Farm Creek and Whiterocks Canyons. The formation forms cliffs, shallow caves, and spires. The Morgan Formation is faulted within the Ice Cave Peak quadrangle and contains most of the slickenlines that were measured in this thesis project (see Chapter 2). The dip of the Morgan Formation steepens from south to north in the field area, with an average dip of 49°, but with dips as high as 80°. The Morgan Formation weathers over the top of topographically lower formations, staining them with its red color.

Weber Sandstone (PIPw)

The Weber Sandstone is a sandstone (60%) with some interbedded limestone (40%) in the lower portion. The sandstone units are light-gray to yellowish-gray on fresh surface and weather light-gray to very-pale-orange. The sandstone units are medium- to coarse-grained,

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rounded to subangular, and well sorted. Clay particles and glauconite are both present; however, these are not as abundant as found farther east in this formation (Haddox, 2005). The Weber Sandstone also contains abundant burrowing that is thought to have destroyed sedimentary structures in some locations. It is very thick-bedded to thickly laminated, and forms resistant ridges. We also locally observed cross-bedding. The limestone units are a pale-bluish-purple, thinly-bedded, and cliff-forming. The Weber Sandstone is about 1000 feet thick.

The Weber Sandstone is well exposed within the Ice Cave Peak quadrangle where it lies stratigraphically above the Morgan Formation and below the Park City Formation. In Whiterocks Canyon, the Weber Sandstone is overlain by the Brennan Basin Member of the Duchesne River Formation. The Weber Sandstone dips to the south at 55° to 70°.

Permian System

The Permian strata found within the Ice Cave Peak quadrangle consist of the upper most portion of the Weber Sandstone and the Park City Formation, which were part of backshore, beach, and shelf depositional environments (Haddox, 2005).

Park City Formation (Pp)

The Park City Formation is comprised of dolomite beds (60%) and sandstone beds (40%). The dolomites are tan to whitish-pink on a fresh surface and weather gray to moderate-yellowish-orange. The dolostones may be glauconitic and quartz-rich. The glauconite grains

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range from 1 to 5 mm in diameter and the quartz is secondary. Resistant dolostone beds contribute to the Park City Formation being a cliff-former. On fresh surfaces, it smells of sulfur and it weathers to a vuggy surface texture. Sandstones are reddish-yellow and coarse- to medium-grained with grains that are subrounded and moderately sorted. Sandstones are thick-bedded and locally cross-stratified. Shale and siltstone beds are less prevalent within the Ice Cave Peak quadrangle than in other nearby areas because the whole formation with all its members is not exposed. The Park City Formation is about 150 feet thick.

The Park City Formation, probably the Grandeur Member, outcrops in Whiterocks Canyon and can also be found in some gullies in the central portion of the quadrangle near Farm Creek (Plate 1). It strikes nearly east-west and dips around 55° south near Farm Creek; however, dip increases to the east of Whiterocks River to 73° and dip shifts to a more south-southeast direction.

Triassic and Early Jurassic Strata

The Moenkopi Formation (Early Triassic), Chinle Formation (Late Triassic), formation of Bell Springs, and Nugget Sandstone (Late Triassic to Early Jurassic) make up the Triassic to Early Jurassic units exposed within the Ice Cave Peak quadrangle. The Moenkopi Formation is transitional in depositional environment from the open marine deposits of the Park City Formation (MacLachlan, 1972) to the terrestrial deposits of the Chinle, Bell Springs, and Nugget formations (Sprinkel et al., 2011a, 2011b).

Moenkopi Formation (TRm)

The Moenkopi Formation is a siltstone and mudstone interbedded with gypsum. The siltstones and mudstones are reddish-brown to reddish-orange on fresh surfaces and weather to dark-reddish-brown. The siltstones are fine-grained and thick-bedded. The formation is slope-forming. Mudstones are often ripple-laminate and thin-bedded. The gypsum units are light-red to grayish-orange-pink in color and ledge-forming. The Moenkopi Formation is overlain in most places by colluvium and vegetation providing only incomplete exposures within the Ice Cave Peak quadrangle. The Moenkopi Formation is about 820 to 1120 feet thick.

The Moenkopi Formation is part of the Mesozoic section that is only exposed in a small portion of the quadrangle along Whiterocks Canyon. It is well exposed to the east in the Lake Mountain and Dry Fork quadrangles (Kinney, 1955; Gregson and Chure, 2000; Haddox et al., 2010), where portions of the formation are repeated due to faulting within the Deep Creek fault zone (Haddox, 2005).

Some fossils have been found within the Moenkopi Formation. Hamblin and Foster (2000) discovered Triassic bones and trackways just to the east of Ice Cave Peak quadrangle. The Moenkopi Formation found on the southern flanks of the Uinta Mountains is similar in lithology to the more well-known outcrops on the Colorado Plateau. Due to limited exposure within the Ice Cave Peak quadrangle, this formation has not been subdivided into members.

The Moenkopi Formation is stratigraphically bounded by two unconformities (McKee, 1954; Pipiringos and O'Sullivan, 1978), placing it between the Park City Formation below and the Chinle Formation above. It is best exposed on the north slope near the top of an east-west

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trending ridge within the Ice Cave Peak quadrangle located in Whiterocks Canyon. In the easternmost outcrops within the quadrangle, an angular unconformity separates the Moenkopi Formation from the Tertiary Bishop Conglomerate (Plate 1) and in the west it is capped by piedmont gravels (Qap).

Chinle Formation (TRc)

The Chinle Formation is shaley mudstone interbedded with siltstone. Mudstones are moderate-yellow to grayish-red on fresh surface and weather gray to dark-grayish-red. Mudstones are fissile and poorly cemented. Mudstone units are weak and form slopes. Siltstones are pale-purple to dark-reddish-brown in color, and moderately- to well-sorted. Siltstone units are also slope-formers. The Chinle Formation is not well exposed within the Ice Cave Peak quadrangle and is mostly covered by colluvium, alluvium, and overgrown with vegetation. It is exposed just east of the main road through the Whiterocks Canyon where the beds strike to the northeast and dip southwest at 57° to 65°. The formation is about 290 feet thick.

The Chinle Formation is another of the Mesozoic formations exposed in a small portion of Whiterocks Canyon near the center of the Ice Cave Peak quadrangle. The formation lies stratigraphically above the Moenkopi Formation and below the formation of Bell Springs (Sprinkel et al., 2011b). Within the Ice Cave Peak quadrangle the Chinle Formation is in contact with its stratigraphic neighbors as well as overlain along an angular unconformity by the Brennan Basin Member of the Duchesne River Formation, which caps the eastern mid elevations. On the eastern side of Whiterocks Canyon, the Chinle Formation is partially buried

by a large alluvial fan. It is capped by piedmont gravels on the western side of the main canyon. The Chinle Formation contains multiple members (Kinney, 1955; Pipiringos and O'Sullivan, 1978; Gregson and Chure, 2000; Jensen and Kowallis, 2005) however, due to poor exposure they are not distinguished within the Ice Cave Peak quadrangle. The Gartra Member, which is a prominent cliff-former to the east of Ice Cave Peak, is either not exposed or not present within the Ice Cave Peak quadrangle. This member tends rapid thickness changes in exposures located in the Lake Mountain, Dry Fork, and Steinaker Reservoir quadrangles to the east (Haddox et al., 2010a; Haddox et al., 2010b; Bart Kowallis, personal comm., 2015).

Formation of Bell Springs (TRb)

The formation of Bell Springs is comprised of sandstone and siltstone. Outcrops are grayish-orange to light-reddish-gray on fresh surfaces and weather light-grayish-red to moderate-reddish-brown. The sandstone is fine- to medium-grained and contains ripple laminations. The formation also contains rip-up clasts. It is sparsely bioturbated, and cross-bedding was also present. The formation of Bell Springs is 90 to 130 feet thick.

The main outcrop for the formation of Bell Springs is found on the eastern side of Whiterocks Canyon on private land. Access was granted by the owner of the land for us to obtain a unit description. The outcrops face west and east within the Whiterocks Canyon. The formation of Bell Springs is thought to be the transition from fluvial deposition to the eolian deposits of the Nugget Sandstone (Sprinkel et al., 2011b; May, 2014; Kowallis et al., 2015).

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<u>Nugget "Navajo" Sandstone (JTRn)</u>

The Nugget Sandstone is a light-gray to light-brown sandstone on fresh surfaces and weathers to dark- to light-red with iron oxide staining. The sandstone is medium- to very fine-grained and well sorted with grains that are subrounded and well cemented. The formation is very thick-bedded to massive, and typically forms ridges and cliffs. We observed both high- and low-angle trough cross-stratification (Figure 3). Within the Ice Cave Peak quadrangle, the Nugget Sandstone is highly fractured and saturated with hydrocarbons. The fractures appear to control the level of hydrocarbon saturation. Dark gray tar-rich sediments are found within inches of light gray to pink hydrocarbon free sediments, separated by healed fractures. The Nugget Sandstone is about 800 feet thick.

The best exposure of the Nugget Sandstone within the quadrangle is located on private property to the east of Whiterocks Canyon (Plate 1). A private road, trending west from the main Whiterocks Canyon road, dead ends in a tar sands pit within the Nugget Sandstone (Figure 4). The formation in the pit is stained dark gray. There is a second abandoned quarry on the eastern side of Whiterocks Canyon as well. The quarrying has produced excellent exposures of the fractures within the formation (See Economic Geology section).

Middle Jurassic and Cretaceous Strata

The Middle Jurassic Carmel and Stump formations, the Late Jurassic Morrison Formation, and the Early Cretaceous Cedar Mountain Formation occur in very limited exposures within the Ice Cave Peak quadrangle. The Carmel and Stump formations are marine deposits

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within an inland sea (Hintze and Kowallis, 2009; Sprinkel et al., 2011a). The Morrison and Cedar Mountain Formation were both deposited in fluvial and lacustrine environments (Kinney, 1955; Gregson and Chure, 2000; Haddox, 2005)

Carmel Formation (Jc)

The Carmel is composed of siltstone, mudstone, and gypsum, with the lower portion consisting of a micritic and sandy limestone and gypsiferous siltstone. The upper siltstone and mudstone is moderate-red-brown to pale-green on fresh surfaces and weathers moderate-reddish-orange. The siltstone is fine- to very-fine-grained, subangular to subrounded, well sorted, and medium-bedded to thickly laminated. The laminations are massive to planar. The formation as a whole is slope-forming. Gypsum beds in the Carmel Formation are common within the Ice Cave Peak quadrangle. They are light gray, and found in thin beds 2 to 10 feet thick, and veins inches to 2 feet thick. The lower limestone beds in the formation are poorly exposed and often covered by colluvium in an east-west trending gully near the contact of the Carmel Formation and Nugget Sandstone. Due to the gully, the terrain is steep and the limestone beds are weathering ledges with the overlying Brennan Basin Member of the Duchesne River Formation weathering debris on top of the limestone beds. Where exposed they are micritic to sandy, moderate-gray in color, and thick- to thinly-bedded. The Carmel Formation is about 150 to 220 feet thick.

The best exposure of the Carmel Formation is on the eastern side of Whiterocks Canyon in the gully where it comes in contact with the Nugget Sandstone; however, this location is located on private land. At this locality two small pit mines for gypsum have been excavated

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near the prospect label on the geologic map (Plate 1). Access was given by the land owner for unit descriptions to be obtained.

Stump Formation (Js)

The part of the Stump Formation exposed in the quadrangle is sandstone and probably equivalent to the Curtis Member mapped in the quadrangles to the east (Haddox et al., 2010a, 2010b). The upper, Redwater Member, which is prominent in these other quadrangles to the east and is typically more resistant to erosion, is likely present but was not observed due to the poor exposures of this formation within the quadrangle. Sandstones of the Stump Formation are light-gray to grayish-orange on fresh surfaces and weather greenish-gray to gray. Sandstones are very-fine to fine-grained and consists of angular to subrounded, very well sorted to well sorted grains. The Stump Formation is cross-bedded and contains glauconite. It tends to be a weaker formation that is friable, and slope-forming. The Stump Formation strikes northeast and dips to the south at 71 ° on the eastern side of Whiterocks Canyon and is about 130 to 250 feet thick.

The Stump Formation is only exposed within the Whiterocks Canyon area near the Whiterocks River where it has been mapped by a number of earlier workers (Thomas and Kruegar, 1946; Eicher, 1955; Kinney, 1955; Stephenson, 1961; Patterson, 1980; Eschner, 1983; Gregson and Chure, 2000). The exposures are almost exactly in the center of the Ice Cave Peak quadrangle and just south of the Nugget Sandstone tar sands pit on the western side of Whiterocks Canyon. It is capped on the east side of the river by the Brennan Basin Member of the Duchesne River Formation and by landslides (Qms). It contributes to a colluvial deposit that

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lie directly downslope. On the west side of the canyon, the Stump Formation is capped by both piedmont gravels and by the Brennan Basin Member. It lies stratigraphically between the Carmel Formation and the Morrison Formation.

Cedar Mountain and Morrison Formations (KJcm)

The Morrison Formation is composed of mudstones interbedded with channel conglomerates and sandstone lenses (Kowallis et al., 1998). Mudstones are variegated, moderate-red to light-red on fresh surfaces and weather yellow-green to light-gray. The formation is predominately a slope-former within the Ice Cave Peak quadrangle. Pebble conglomerates are poorly sorted and thinly-bedded (Sprinkel 2006), but well cemented, which cause them to be ledge-forming (Gregson and Chure, 2000). Sandstones are more abundant elsewhere in the region (Kinney, 1955; Turner and Peterson, 1992) and are moderate reddish orange, coarse- to medium-grained, moderately sorted, and friable (Sprinkel, 2006). Exposures of this formation are very poor in the quadrangle. The Morrison Formation is about 300 feet thick.

The Cedar Mountain Formation is very similar to the Morrison Formation. It is comprised of mudstones interbedded with pebble conglomerate lenses and beds (Kinney, 1955; Gregson and Chure, 2000; Sprinkel, 2006). The conglomerates have a silty to muddy matrix. Mudstones are a moderate-red, pale-purple, light-gray, or pale-greenish-yellow (Currie, 1997). The formation contains abundant calcrete beds that weather out as concretions (Sprinkel, 2006). The Cedar Mountain formation is also a slope-former, and poorly exposed within the quadrangle. This formation is indistinguishable from the Morrison Formation within the Ice Cave Peak

quadrangle, and they are mapped as one unit. The Cedar Mountain Formation is approximately 200 feet thick (Sprinkel, 2006).

The Cedar Mountain and Morrison formations the youngest pre-Tertiary strata exposed in the Whiterocks Canyon area. Very limited outcrops of these units occur in the center of the quadrangle and only on the western side of Whiterocks Canyon where the underlie slopes south of the Nuggets tar sand quarry (Plate 1).

Brennan Basin Member of the Duchesne River Formation (Tdb)

The Brennan Basin Member of the Duchesne River Formation is composed of conglomerate interbedded with sandstone, siltstone, mudstone, and bentonitic ash beds (Figure 5). The conglomerates contain boulder-(up to 5 feet in size) to gravel-sized clasts. The clasts are in a moderate red to pale yellowish-orange sandy to muddy matrix. The matrix is very poorly sorted and contains subangular to rounded clasts. It is thick- to very thick-bedded. The clasts are predominantly Paleozoic limestone (dark gray to moderate gray) with some clasts of moderate-reddish-orange sandstone and multi-colored (Mesozoic) rocks as well. Occasional clasts of reddish quartz sandstone from the Neoproterozoic Uinta Mountain Group. The conglomerate is poorly- to well-cemented with quartz and typically forms resistant cliffs and ledges.

Sandstones within the Brennan Basin Member of the Duchesne River Formation are pale yellowish-orange to grayish-pink on fresh surfaces and moderate reddish-orange on weathered. They are coarse- to fine-grained, moderately sorted to well sorted, and very thick- to thin-bedded. The sandstone units contain some tar, particularly when in contact with the Nugget

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Sandstone. The tar is thought to originate from the Green River Formation (Blackett, 1996). The sandy beds pinch and swell and scour into underlying mudstone beds. Sandstone beds range in lateral extent from a few feet to approximately 1/4 mile in distance.

The siltstone and mudstone beds of this formation are light-brown to pale-yellowish-orange on fresh surfaces and moderate-red- to dark-yellowish-orange on weathered. They are thick- to thinly-bedded and massive to planar laminated. The muddy beds within the Brennan Basin Member of the Duchesne River Formation contain ripples and tend to be slope-forming due to their fissile nature.

Bentonitic ash beds within the Brennan Basin Member of the Duchesne River Formation are yellowish-gray to light gray on fresh surface and weather grayish-yellow-green. They can contain pebble- to sand-sized lithics and range from 2-12 feet thick. Ash beds from this member have given isotopic ages ranging from 40-41 Ma (B. Kowallis and D. Sprinkel, unpublished data). The Brennan Basin Member of the Duchesne River Formation is about 280 to 580 feet thick.

The Duchesne River Formation outcrops across much of the southern portion of the Ice Cave Peak quadrangle (Anderson and Picard, 1972; Hansen, 1984). The Duchesne River Formation has four members: the oldest, and only member exposed in the Ice Cave Peak quadrangle, is the Brennan Basin Member. The other members of the formation (Dry Gulch Member, LaPoint Member, and Starr Flat Member) are exposed south of the quadrangle (D. A. Sprinkel, personal communication).

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The distinguishing difference between the four members of the Duchesne River Formation is the relative amount of conglomerate, sandstone, siltstone, and mudstone (D.A. Sprinkel, personal communication). All four members contain some of each lithology. There is also a fining upward element to the oldest three members. The youngest, Starr Flat Member, reflects a return to conglomerate. The Brennan Basin Member has a higher proportion of sandstone and conglomerate compared to the other members. The Dry Gulch Member is predominately siltstone, and the LaPoint Member contains more mudstones and numerous layers of altered volcanic ash (D.A. Sprinkel, personal communication). The fining upward progression then terminates as the Starr Flat Member is predominately conglomerate. The break in the fining upward progression coupled with the fact that the Starr Flat Member is the only member entirely sourced from the Uinta Mountains that does not occur to the south of the Uinta Basin syncline supports speculations that perhaps the Starr Flat Member is more closely related to the Bishop Conglomerate than it is to the Duchesne River Formation (Haddox, 2005; Kowallis et al., 2005; Sprinkel, 2006).

Within the Ice Cave Peak quadrangle, the Brennan Basin Member of the Duchesne River Formation is usually a slope-former. However, the formation does form a fairly resistant cap over several areas in the quadrangle due to the cobbles and clasts of the conglomerate layers. The member is relatively easy to identify due to it reddish color and occasional gray-green beds of bentonite. The Brennan Basin Member of the Duchesne River Formation is almost always capped by Quaternary piedmont gravels in the western half of the field area. It does host a few landslide deposits in both the west and east sides of the quadrangle. On the eastern side of the quadrangle, the Brennan Basin Member of the Duchesne River Formation is being exposed as the overlying piedmont gravels are being eroded away. Visible headward erosion into the

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overlying gravels can be observed both in the field and from aerial photos exposing the formation as the piedmont armor is removed. There is also a large landslide deposit that overlies the southeastern most exposures of Brennan Basin Member of the Duchesne River Formation within the quadrangle.

Bishop Conglomerate (Tb)

The Bishop Conglomerate is composed of polymict conglomerate interbedded with sandstone, siltstone, and thin ash beds. The conglomerates contain boulder- to pebble-sized clasts supported by a sandy matrix, and are poorly sorted. Clasts are subangular to rounded and include Paleozoic limestone, sandstone, chert, as well as quartzitic sandstone clasts from the Neoproterozoic Uinta Mountain Group (Figure 6). Uinta Mountain Group clasts are rare near the base of the formation, but become more common upwards indicative of the unroofing of the Uinta Mountains. Bedding varies from thin to very thick. The conglomerates are loosely cemented and moderately resistant.

The sandstone and siltstone units are light-gray to pinkish-gray on fresh surfaces and weather to a moderate-reddish-orange to pale-orange. The sandstones are friable and contain fine- to coarse-grained sand. The sandstone units may be either ledge-formers or slope-formers. Siltstones within the Bishop Conglomerate are fine-grained, moderately sorted, and mostly slope-formers.

Ash beds present within the Bishop Conglomerate are white to very light-gray and tend to be thin (< 1 ft). Tuffaceous sandstone beds are also common in the upper part of the unit. Ash

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beds from this formation have given isotopic ages ranging from 30 to 34 Ma from outcrops farther to the east (Kowallis et al., 2005). The Bishop Conglomerate is as much as 500 feet thick and thickens from west to east across the Ice Cave Peak quadrangle.

The Bishop Conglomerate is exposed in the higher elevations of the Ice Cave Peak quadrangle. This confines the Bishop Conglomerate to the northern half of the quadrangle and mainly the northeastern section where it is deposited on the higher flanks of the Uinta Mountains. One of the more well know units within the Uintas, the Bishop Conglomerate, was given its name by the famous soldier, geologist, and American west explorer John Wesley Powell (Powell, 1876; Sears, 1924).

During a pause in the Uinta Uplift the Gilbert Peak erosional surface formed and allowed the Bishop Conglomerate to be deposited (Hansen, 1984; Hansen, 1986; Hansen, 1987 Gregson and Chure, 2000; Sprinkel et al., 2000). The Bishop Conglomerate is the source of the material that makes up the majority of the landslides within the Ice Cave Peak quadrangle. The largest concentration of landslides are found in the eastern half of the quadrangle; however, the Bishop Conglomerate on the west side of Ice Cave Peak quadrangle sources landslides as well. Landslides form mostly due to the abundance of weak Mesozoic and upper Paleozoic bedrock underlying the Bishop Conglomerate (Kowallis and Bradfield, 2005; Bradfield and Kowallis, 2007).

Quaternary Deposits (Q)

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The Ice Cave Peak quadrangle field area has been incorporated in much smaller (1:100,000) scale maps before (Sprinkel, 2006). Producing this map at a 1:24,000 we were aware that we would be mapping at a much greater detail, and thus have more Quaternary units. We also wanted our final map to match the 1:24,000 scale maps in surrounding areas. With this in mind, the Quaternary deposits have been defined based upon definitions established by the Utah Geological Survey that classify them by their type (dominate processes and lithology) and rank (the singular dominant process of deposition). We identified six different Quaternary units, of which some have sub sets within them. These are: alluvium, colluvium, mass movement, human-influenced, piedmont, and glacial deposits.

Alluvial and Colluvial Deposits (Qa and Qc)

Alluvial deposits are the most diverse of the Quaternary deposits with six map subunits in the Ice Cave Peak quadrangle. The youngest alluvial deposits (Qa, and Qa1) are closely related. Active river channel and floodplain deposits are mapped as Qa. Their character varies throughout the Ice Cave Peak quadrangle. In the large Whiterocks River drainages, the deposits are composed of gravel- to boulder-sized clasts and well sorted, well-rounded sands. In the smaller streams, found in the western half of the quadrangle and at higher elevations throughout the field area, these Qa deposits tend to be moderately sorted, sub-rounded to sub-angular with pebble to cobble size clasts and occasional boulders. Qa1 represents older river terrace deposits along active river channels and are distinguished from Qa by their higher elevations along these channels and lack of boulder-sized clasts.

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Older alluvial deposits (Qa2) are typically found on gentle slopes at even higher elevations than Qa and Qa1. Qa, Qa1, and Qa2 are all unconsolidated silt, sand, and gravel. They range in thickness from 1-100 feet.

Alluvial fan deposits (Qaf) of Holocene age in the Ice Cave Peak quadrangle are concentrated in the Whiterocks Canyon. These alluvial fan deposits are composed of unconsolidated and poorly sorted boulders, gravel, sand, and silt. They display color attributes of their source formations, which are typically nearby with the exception of some of the southern alluvial fans where the tributary feeding the fan has cut back into more distant formations, often the Duchesne River Formation. Alluvial fans range up to 130 feet thick.

Mixed alluvium and colluvium (Qac) of Holocene and upper Pleistocene age within the Ice Cave Peak quadrangle often merges with the alluvial fans along the flanks of the valley in Whiterocks Canyon. This unit is characterized by boulders, cobbles, and gravel-sized clasts in a sandy or silty matrix that are deposited as thin blankets on slopes and in drainages along the boundaries between more clearly alluvial or colluvial deposits. Mixed alluvium and colluvium is as much as 50 feet thick.

Several levels of Quaternary piedmont gravels (Qap units 1 - 4) have been mapped in the quadrangle. Qap and Qap1 are estimated to be Holocene and Pleistocene, respectively (Sprinkel, 2006). They are located in the eastern portion of Ice Cave Peak quadrangle and confined to lower elevations where they cap the Brennan Basin Member of the Duchesne River Formation and often provide sediment to other Quaternary units, such as Qa2 and landslides (Figure 7). The piedmont gravels are more resistant to erosion than the underlying Duchesne River Formation and therefore form caps on almost all of the Duchesne River Formation outcrops in the eastern

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portion of the field area (Plate 1). Qap and Qap1 are composed mainly of unconsolidated to poorly consolidated Uinta Mountain Group quartzite boulders, cobbles, and pebbles in a sandy or silty matrix. Elevation and clast size are the distinguishing characteristics between Qap and Qap1 (Figure 8). Qap is found at lower elevations and has smaller clasts (less than 2 feet diameter), while Qap1 is found at higher elevations and has clasts ranging up to 2 feet in diameter. Thickness of these deposits ranges from 1-30 feet.

Qap2 is thought to be Upper Pleistocene in age (Sprinkel, 2006). It is deposited only in the eastern portion of the field area at mid elevations. It caps the Duchesne River Formation as well and is located within the large eastern landslide (Plate 1). The Qap2 deposits continue into the Lake Mountain quadrangle. Along the eastern border they are being eroded away exposing the Brennan Basin Member of the Duchesne River Formation. Lithologically Qap2 is very similar to the previously mentioned piedmont gravels, and the depositional processes are thought to be the same as well; however, their elevations and gradients do not allow them to be correlated and suggest that these are older deposits. Qap2 is a higher elevation with slightly larger clast sizes on average (up to 3 feet in diameter). Qap2 also has a slightly better developed soil on top of it than Qap and Qap1. Thickness is relatively thin and probably does not exceed 10 feet in most locations within the quadrangle.

Qap3 and Qap4 are Middle Pleistocene in age (Sprinkel, 2005) and found only at the highest elevations within the Ice Cave Peak quadrangle. Qap3 and 4 are similar in character and lithology with the other piedmont gravel deposits being composed almost exclusively of boulders and large cobbles of reddish purple Uinta Mountain Group quartzite, but these clasts tend to be less rounded and more subangular to angular than the younger piedmont gravels. Well-developed

soils cover these deposits. Again, elevation and boulder average size distinguishes Qap3 (up to 4 feet in diameter) from Qap4 (often has clasts greater than 4 feet in diameter). Qap4 is found at the highest elevations of any of the piedmont gravel deposits within the Ice Cave Peak quadrangle. Exposed surfaces of this unit are littered with large 4-7 feet diameter boulders. These units rest on the Bishop Conglomerate and, in some places, on the Madison Limestone and Red Pine Shale near the rim of the Whiterocks Canyon. Qap3 deposits range in thickness from 100-1000 feet thick and Qap4 is typically much thinner at about 10 feet thick.

Glacial Deposits (Qg)

Smiths Fork Till (Qgs) deposit is an unconsolidated, matrix-supported, diamicton that is a result of the last glaciation (ca. 32,000 to 14,000 years ago) (Munroe and Laabs, 2009). The matrix ranges from a reddish-brown silty clay to a pink sand. The pink sand matrix is the most common within the Ice Cave Peak quadrangle where the clasts are composed of Uinta Mountain Group pebbles, cobbles, and boulders. Boulders dominate the unit, and the cobbles and pebbles are only found by digging into the unit. Clasts are angular to subrounded. Local areas that contain rounded clasts are also present and have been interpreted as the, "incorporation of proglacial outwash deposits during glacial advance" (Munroe and Laabs, 2009). The topography of the Smiths Fork Till deposits is mainly hummocky but can be smooth in localities. It ranges up to 30 feet thick.

Smiths Fork Till is located within Whiterocks Canyon along the eastern banks of the Whiterocks River. Three main exposures have persisted and have yet to be covered by the

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colluvium or alluvial fans. The Smiths Fork Till is thought to have been deposited directly by glacial ice (Munroe and Laabs, 2009). No moraines of Smiths Fork age were mapped within the Ice Cave Peak quadrangle; however, sampling of terminal moraines in nearby areas for cosmogenic [10]Be surface-exposure dating revealed that Smith Fork Till terminal moraines were abandoned by the retreating glaciers before 16,000 years ago (Munroe et al., 2006; Laabs et al., 2007; Refsnider et al., 2008).

The Smiths Fork Outwash (Qgas) is similar to the till. Smiths Fork Outwash is unconsolidated, clast-supported, with cobble- to boulder-sized clasts. Boulders average around three feet in diameter. Clasts are reddish-brown to pale-pink, rounded to subangular and composed of Uinta Mountain Group quartzite in a matrix of pebbles and sand, which is unsorted, subrounded to subangular, reddish-brown to pink in color. The Smiths Fork Outwash unit is late to middle Pleistocene in age (Munroe and Laabs, 2009) and is less than 15 feet thick within the Ice Cave Peak quadrangle.

Smiths Fork Outwash is the most extensive glacial deposit within the Ice Cave Peak quadrangle. It is found on both the east and west sides of the Whiterocks River in the southern half of the quadrangle. Much of the Smith Fork Outwash deposit is on private land or on the Ute Indian Reservation. The restricted access to the land did not pose too much of a problem as the unit is fairly easy to distinguish from surrounding geology from public roads, and aerial photos.

Blacks Fork Outwash (Qgab) is very similar to Smiths Fork Outwash in that it consists of unconsolidated cobbles to boulders that are clast supported. The clasts are rounded to subrounded, sandstone and quartzite from the Uinta Mountain Group. Blacks Fork Outwash is found at higher elevations than Smiths Fork Outwash. It is typically deposited at elevations

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above all modern streams or rivers with a thickness that is generally unknown, but thought to be between 1 and 65 feet (Munroe et al., 2006; Laabs, 2007; Refsnider, 2008) and around 30 feet in the Ice Cave Peak quadrangle.

Blacks Fork Outwash is located only in the southern most section of the Ice Cave Peak quadrangle on the eastern edge of the Smiths Fork Outwash. The distinction was originally made with the aid of aerial photos when a break in elevation was recognized. The difference in elevation was then confirmed in the field and consequently mapped as the Blacks Fork Outwash of middle to late Pleistocene age (Munroe and Laabs, 2009).

Colluvium Deposits (Qc)

Colluvium deposits (Qc) consist of boulder, cobble, and gravel deposited on slopes. They are unconsolidated and contain angular to subangular, poorly to well sorted clasts. The lithology of the material varies based on the lithology of the units that weathered to form the colluvial deposit. Colluvium deposits are often located in steep topography near piedmont gravels, the Madison Limestone, and the Red Pine Shale. They range from 1 to 30 feet thick.

Colluvium deposits are found throughout the Ice Cave Peak quadrangle at all elevations. In the lower elevations (south portion of the Ice Cave Peak quadrangle) the colluvium deposits tend to be a result of weathered piedmont gravel surfaces. Landslide deposits contribute to colluvium deposits throughout the entire quadrangle. In the higher elevations, the colluvium deposits are derived from the Madison Limestone and the Red Pine Shale.

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Human-Influenced Deposits (Qh)

Human-influenced deposits (Qh) are only mapped in one location within the Ice Cave Peak quadrangle in the southeastern part where a relatively large agricultural field is located. Farming has obscured any earlier contacts in this area. Access to the private land was not obtained, and aerial photos could not conclusively identify the contact between Qa and Qa2 that occurs within the Qh.

Mass Movement Deposits (Qms)

Mass movement deposits (Qms) within the Ice Cave Peak quadrangle occur as both slumps and landslides (Holocene to upper Pleistocene) (Sprinkel, 2006) throughout the quadrangle over a range of sizes. However, landslides are by far the most prevalent and the largest of the mass movement deposits. Slumping within the field area is most common in the southeastern corner of the quadrangle. In that area, the Brennan Basin Member of the Duchesne River Formation is capped by piedmont gravels, near the toe of a large landslide (2 miles long over 3 miles wide) that flowed out of the adjacent Lake Mountain quadrangle. Blocks capped with piedmont gravel remain coherent as slumping shifts and rotates them. Smaller slumps are found throughout the quadrangle.

Landslides are located in all parts of the quadrangle. They range in size from barely mappable at the 1:24,000 scale to over 2 miles long and over 3 miles wide. The largest, as mentioned earlier, is located in the southeastern corner of the quadrangle and continues into the Lake Mountain quadrangle to the east. The scarps of the larger landslides can exceed 200 ft.

Within the larger landslides (typically to the east of Whiterocks River) we find younger mass movement deposits mapped within mass movement deposits demonstrating that recurrent movement has occurred within the larger landslides. This was also noted to the east in the Lake Mountain and Dry Fork quadrangles (Kowallis and Bradfield, 2005; Bradfield and Kowallis, 2007).

As previously stated, the majority of the large landslides occur where the Bishop Conglomerate has been deposited on weaker Mesozoic and Paleozoic units like the Chinle Formation, Moenkopi Formation, or Doughnut Shale. These weaker formations erode away easily causing over steepening of the Bishop Conglomerate that then leads to landslides. Occasionally "islands" of bedrock remain exposed after landslide material has flowed around them. These islands are more common in the large landslide on the eastern edge of the quadrangle where the Brennan Basin Member of the Duchesne River Formation protrudes in places above the hummocky topography of the landslide. The thickness and lithology of each landslide depends on its size and the rocks from which it was derived.

Structural Geology

An index map of the Ice Cave Peak quadrangle (Figure 2) shows the field area's proximity to two large fault systems, the Uinta Basin-Mountain boundary fault, and the South Flank fault zone. These fault systems combine with nearby smaller systems (the Deep Creek fault zone and splays from the previously mentioned larger systems) to have a controlling effect on the deformation and structure of the geological units within the Ice Cave Peak quadrangle.

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The Uinta Mountains block is bounded to the south by the Uinta Basin-Mountain boundary fault, which is a blind thrust fault. This thrust, along with the North Flank and Henry's Fork thrusts on the north side of the range, has created a pair of large east-west trending anticlines that account for the steep southerly dip of most units within the Ice Cave Peak quadrangle (Sprinkel, 2006, 2007, 2014).

The northern half of the quadrangle contains significant east-west trending normal faults that appear to be splays off of the South Flank fault zone (Plate 1, Sprinkel, 2006). These normal faults can be clearly seen from the vantage point of Ice Cave Peak (Figure 9). The faults offset the Mississippian Madison Limestone and Neoproterozoic Red Pine Shale. They are covered by younger (Middle Pleistocene) piedmont gravel deposits that the faults do not cut. The down-dropped blocks are to the north with footwalls to the south. These are thought to be relaxation faults that formed as Laramide compression eased.

The Deep Creek fault zone was named by Untermann and Untermann (1969). It has been extended into the Ice Cave Peak quadrangle by mapping during this project (see Chapter 2). The Deep Creek fault zone consists of multiple steeply dipping to vertical faults that show both oblique normal and strike-slip movement. A structural study of the kinematics of the faults is found in Chapter 2.

Minor folding occurs within the Ice Cave Peak quadrangle. Near the contact of the Round Valley Limestone and Morgan formations a fold can be seen on the western side of Whiterocks Canyon (Figure 10). The fold continues west from there, just over one mile, to Farm Creek Canyon (Figure 11). The fold plunges slightly to the east. The beds of the Morgan Formation display a slight increase in dip direction and then abruptly flatten out (near horizontal). The bed

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orientation returns to a dip angle and direction that is closer to the average within a few hundred feet.

The southern portion of the quadrangle is dominated by Brennan Basin Member of the Duchesne River Formation, alluvium (fluvial, glacial, and piedmont), and mass movement deposits. This results in a consistent slight dip to the south.

Economic Geology

The geology within the Ice Cave Peak quadrangle has some economic potential. Perhaps the most significant value would come from a better understanding of the subsurface, as there is active hydrocarbon production from the nearby cities of Neola, Whiterocks, and Leeton (and from the greater Uinta Basin as well). The southwest corner of the Ice Cave Peak quadrangle is within two miles of multiple producing oil wells. Furthermore, many world-class oil fields that provide hydrocarbons at economic levels are related to thrust systems (Oliver, 1986). A better understanding of the Uinta thrust system could benefit, not only exploration on the flanks of the Uinta Mountains, but serve as a model for other geologically similar localities throughout the world.

The Ice Cave Peak quadrangle has minor tar sand deposits that have been mined from the Nugget Sandstone, as well as ground and surface water resources, along with areas for tourism, camping, and hunting. Whiterocks Canyon and the greater region is widely known as a potential location for the fabled Spanish gold that is rumored to be buried in the canyons of the southeastern Uintas and attracts treasure hunters from across the country (Conn, 2005).

Nugget Tar Sands

A small tar sands open pit mine is located at the mouth of Whiterocks Canyon within the Nugget Sandstone. This tar sand deposit has been visited and revisited by both academics and industry alike throughout the last half century (Peterson, 1985; Bukka et al., 1991; Blackett, 1995). The hydrocarbons are not native to the Nugget Sandstone and isotopic studies indicate that the source was the Green River Formation (Covington, 1963). Further sulfur isotope analysis supports the conclusion that the hydrocarbons came from the Green River Formation (Mauger et al., 1973).

The bitumen saturated zone is almost entirely within the Nugget Sandstone and is about 900 feet thick (Blackett, 1995). The saturated Nugget Sandstone units are found on both sides of Whiterocks Canyon where they are capped by Duchesne River Formation. The Nugget Sandstone in this locality is highly fractured. The fractures do not have a preferred orientation. Permeability (linked to fractures) controls bitumen saturation within the Nugget Sandstone. Fractures cause zones of saturation to be found adjacent to bitumen barren zones (Blackett, 1995).

The multiple studies have estimated variable amounts of bitumen in place. Severy (1943) mapped the saturated outcrops and concluded that around 9.5 million barrels was in place. Shirley (1961) used data from 11 cores and estimate 105 million barrels (57 million barrels as proven reserves) at the mouth of Whiterocks Canyon. Covington (1963) and Lewin and Associates (1984) concluded that there was 60 million barrels in place. Rocky Mountain

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Exploration Company along with Parley Peterson concluded that the Navajo 'Nugget' Sandstone holds more than 100 million barrels of bitumen in place (Peterson, 1985). Other studies have been conducted with results that lie within the previously mentioned range of estimated barrels in place. Interestingly, there are pebbles of oil saturated Nugget Sandstone within the Duchesne River Formation (Blackett, 1995). This may suggest that oil migration occurred prior to the deposition of the Brennan Basin Member of the Duchesne River Formation.

The tar sands of Whiterocks Canyon have a storied history of production that starts in the 1940s and continues to the present. Peterson (1985) presents the timeline of production activity as follows; 1940s consisted of smaller operations such as pit mining and adits. The late 1950s saw the completion of three exploration wells in hopes of finding liquid hydrocarbons. Peterson states that in the early 1960s two extraction plants were built, which used hot water and solvents to separate the hydrocarbons from the sands. White Rocks Oil Properties drilled 11 cores within the Nugget in the early 1960s as well. Nine of the eleven wells intersected the entire interval of saturation. The eastern side of the river was home to a pilot plant and strip mine under Western Industries Company. Major Oil Company followed suite in the early 1970s with a strip mine and pilot plant on the west side of Whiterocks River. The quarry on the west side of Whiterocks River is the only evidence of this previous activity and is currently used by Duchesne County to assist in paving the county highways (Peterson, 1985; Blackett, 1995). As with all hydrocarbon reservoirs and reserves, the price of crude oil determines its economic potential.

Ground Water

As the great Mark Twain said, "Whiskey is for drinking and water is for fighting over." With drought a common occurrence throughout the West and ground water shortages being a very real risk for Utah and other states, the seasonal snowmelt that reanimates Whiterocks River every spring is a vital resource to the region. The high elevation of the Ice Cave Peak quadrangle can be packed with tens of feet of snow during a wet winter. The snow melt makes its way to the crops, gardens, and glasses of the Uinta Basin through a web of canals and irrigation plumbing. There are also a number of springs within the southern portion of Ice Cave Peak. Many of these springs have been dammed by local residence for homestead crops and cattle.

Tourism

Tourism is one of Utah's major industries. The Ashley Nation Forest, which is home to the majority of Ice Cave Peak quadrangle, receives over 2.5 million visitors each year for outdoor recreation (Utah.com). Camping is a main tourist attraction of the quadrangle with multiple, well-maintained campsites found within the field area. Beyond the beautiful outcrops of limestone cliffs and abundant wildlife, rock hounds frequent the area. Petrified wood, iron concretions, fossils, and numerous unmapped caves are all to be found within the Ice Cave Peak quadrangle. Many people come for the thrill of treasure hunting. As mentioned previously, the Whiterocks Canyon is legendary for the myth of buried Spanish gold. Many of the aspen trees within the canyon have carvings on them. Legend has it that some of these carvings act as ancient treasure maps and mark the way to the buried treasure.

<u>Hunting</u>

If one spends any time within the high country of Ice Cave Peak quadrangle, they are sure to see both large and small game. Mapping during the fall required hunter orange clothing for safety, as hundreds of hunters stake claim to the mountain sides during deer and elk season. The Whiterocks River is a very popular location for fly fishing. It is located near two blue ribbon rivers (Duchesne River and Green River).

Geologic Hazards

Geological hazards are of relatively low risk within the Ice Cave Peak quadrangle. However, multiple fault systems dissect the quadrangle, unstable units (The Bishop Conglomerate) are found near homes, and homes are built near river systems and at the mouth of smaller canyons. The largest hazard is due to unstable units and the potential for human or infrastructure damage as the result of mass movement.

The faults that run through Ice Cave Peak have not been active since pre-Oligocene (See Chapter 2) and do not pose a known threat to the residence within the field area and region. However, there is unconventional hydrocarbon production in the area, and the usage of waste water is not made public. If the water is reinjected into the subsurface (a common practice of unconventional hydrocarbon production) the regional seismic activity could greatly increase (Keranen et al., 2013). The practice of waste water injection is understood to have caused an increase of seismic activity in Oklahoma (Keranen et al., 2013). Oklahoma seismic activity

increased from around one event per year before waste water injection to multiple per month since the practice has been made routine.

Landslides also pose a minimal threat within the quadrangle. The Bishop Conglomerate is unstable and associated many mass movement events. With the amount of traffic that the region sees during hunting season and the homes located at the base of steep topography, landslides are a geologic hazard to the quadrangle. Furthermore, a few of the homes on the east side of the river are built within the rock shadow of nearby cliffs.

Most of the homes within the Ice Cave Peak quadrangle are built on glacial till and outwash and should be safe in a 100 year flood event (floodplain maps were unavailable for this region; homes are around 40 feet above active floodplain). However, there are homes and infrastructure just south of the mouth of Whiterocks Canyon that are on the active floodplain of the Whiterocks River. Landslide dams, and subsequent failure, within off shoot canyons to the east of Whiterocks River could pose a threat to a few of the homes within Whiterocks Canyon however, a catastrophic event resulting from this is unlikely.

Chapter 2: Kinematic Structural Analysis of Ice Cave Peak Quadrangle

Abstract

Geologic mapping (1:24,000 scale) of the Ice Cave Peak quadrangle, Uintah and Duchesne Counties, Utah has produced a better understanding of the geologic structures present in the quadrangle and has increased our understanding of faulting in northeastern Utah. Faulting

appears to be a result of the Laramide orogeny with the possibility of prolonged movement along some regional faults.

Mapping and kinematic analysis of structures within the Ice Cave Peak quadrangle. This project resulted in the extension of the Deep Creek fault zone into the Farm Creek brecciated zone within the Ice Cave Peak quadrangle. Faults within the Farm Creek brecciated zone are both strike-slip and normal/oblique faults as documented by mapping and kinematic indicators and cut the folded hanging-wall sedimentary rocks above the Uinta Basin-Mountain boundary thrust fault. Regional stresses indicate northeast-southwest compression combined with near east-west stresses.

These faults may be part of an en echelon fault system that is rooted in the upper Neoproterozoic and reactivated during Laramide deformation above a possible transfer zone between segments of the buried Uinta Basin-Mountain boundary thrust.

Introduction

New faults mapped as part of a USGS-EDMAP project focused on the Ice Cave Peak quadrangle along the south flank of the Uinta Mountains in Utah, has shown that the Deep Creek fault zone, previously mapped to the east of the Ice Cave Peak quadrangle (Figure 12) extends farther west into this quadrangle. The Deep Creek fault zone is an array of northwest trending faults that display a wide variety of slip, including: normal, reverse, thrust, and strike-slip movement. This fault zone has a northern termination at the South Flank fault zone, a large east-west trending normal fault, and its associated fault splays, that run along nearly the entire southern flank of the Uinta Mountains (Figure 12). The Deep Creek fault zone had previously

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been mapped by several workers, including Untermann and Untermann (1968), Sprinkel (2002), Haddox (2005), and Haddox et al. (2010). We have now extended the western end of the Deep Creek fault zone into the Ice Cave Peak quadrangle with the southern extent of this zone trending east from the Ice Cave Peak quadrangle through the Little Water Hills of the Lake Mountain quadrangle and into the Dry Fork quadrangle where the zone cuts through Little Mountain and into Coal Mine Basin, finally terminating on the east near the city of Vernal, Utah.

The strata cut by the Deep Creek fault zone range from Paleozoic to Mesozoic in age and dip to the south 10° to 50°. Oblique slip movement is predominant on the faults within the Deep Creek fault zone, although there are faults with almost pure strike-slip or pure normal slip motion. In general, the faults show left-lateral movement combined with normal and thrust faulting. The trend for mostly left-lateral movement continues with the newly discovered faults in Ice Cave Peak quadrangle.

Furthermore, the Deep Creek fault zone is located at a hinge point of the greater Uinta anticline structure (Billingsley, 1933; Ritzma, 1969; Hansen, 1986). This area is located between the Uinta Arch and Uinta Anticline (Sprinkel, 2006). The regional setting of the Deep Creek fault zone is a significant factor in the nature of the structure and faults that are present.

In order to better understand this fault zone, we collected and analyzed fault orientation and slickenline data within the Ice Cave Peak quadrangle and combined it with data collected earlier from the Deep Creek fault zone by Haddox (2005).

Background and Previous Work

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The Uinta Mountains are a Laramide orogenic structure (Hansen and Bonilla, 1954; Williams, 1955; Hansen, 1969; Stone, 1993; Marshak et al., 2000; Johnston and Yin, 2001). The Laramide orogeny was an east-west compressional event that formed predominately north-south trending structures throughout its duration (Armstrong, 1974; Hintze and Kowallis, 2009). These structures make up the basins and orogenic belts of much of Utah, Colorado, Wyoming, and the Rocky Mountain region (Conner and Harrison, 2003; Hintze and Kowallis, 2009). The Uinta Mountains, which trend east-west, are an exception to the generally north-south trend of most other Laramide uplifts.

The uncommon orientation of the Uinta Mountains has been the focus of several studies. Different hypotheses have been given to explain this unusual orientation (Gries, 1983; Hamilton, 1988; Bradley, 1995; Gregson and Erslev, 1997). Stone (1993), Paulsen and Marshak (1999), Marshak et al. (2000), Johnston and Yin (2001), and Ashby et al. (2005) proposed that this east-west trend may be explained by the reactivation of east-west trending basin-bounding faults adjacent to a paleo-aulacogen. This paleo-aulacogen was filled with Precambrian sediment that become both coherent, and semi-metamorphosed (Marshak et al., 2000). These strata were apparently very well lithified and maintained their integrity during Laramide east-west stresses (Johnston and Yin, 2001). The pre-existing faults bounding the aulacogen were reactivated and uplifted the Precambrian sediments that are now exposed at the core of the Uinta Mountains and called the Uinta Mountain Group (Johnston and Yin, 2001). These pre-existing faults would coincide with the north and south flank thrust faults that border the Uinta Mountains uplift.

Thrust faults are exposed along the northern flanks of the Uinta Mountains, but are buried, blind thrusts on the southern flanks. Associated with the northern bounding thrust faults

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are north-northeast trending left-lateral oblique faults (Johnston and Yin, 2001). Johnston and Yin (2001) predicted that similar left-lateral oblique faults would be found on the southern flank of the Uintas. We see these faults within the Ice Cave Peak quadrangle and the Deep Creek fault zone.

The Uinta Mountains are comprised of two misaligned, large anticlines. Recently, Sprinkel (2014) proposed that the two sections are separated by a northwest trending fault or fault zone cutting across the entire mountain range (Sprinkel, 2014). As previously mentioned, there are alternate explanations for this two-part division, but this new idea merits further examination. The northwest trending fault that separated the West Dome from the East Dome is thought to have first formed during the late Neoproterozoic (Sprinkel, 2014). Supporting evidence for this inferred fault zone includes: 1) the truncations of younger structural features against the inferred fault zone, 2) the westward tilting and offset of the Precambrian age Uinta Mountain Group rocks while the Paleozoic rocks show little or no offset and are lacking any pre-Laramide deformation, and 3) large angular unconformities at the contact of the Uinta Mountain Group and Paleozoic rocks (Sprinkel, 2014). Interestingly, the Deep Creek fault zone also trends northwest and is parallel to sub-parallel with Sprinkel's (2014) inferred fault zone, but lies approximately ten miles south of the saddle between the East and West Dome where his proposed fault zone sits (Figure 13).

Two other major structural features that impact the Ice Cave Peak quadrangle are the Uinta Basin-Mountain boundary thrust and the South Flank fault. The Uinta Basin-Mountain boundary thrust is a blind fault that trends east-west along the southernmost flanks of the Uinta Mountains and northern Uinta Basin (Sprinkel, 2007). The Uinta Basin-Mountain boundary

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thrust has been inferred and mapped both within the Ice Cave Peak quadrangle (Schamel, 2013) and to the south of the quadrangle (Sprinkel, 2014). The oil and gas industry has conducted some seismic studies along the northern-flank of the Uinta Basin, however much of the data is not published. The little published seismic data that exists is ineffective in locating exactly where the Uinta Basin-Mountain boundary thrust is located with respect to the Ice Cave Peak quadrangle (Fouch et al., 1992).

The South Flank fault is a large east-west trending fault that runs from the western Uintas to the hinge zone of the two anticlines just northeast of Ice Cave Peak quadrangle (Figure 12). The termination on the east end of the fault could be due to the fault being buried by Tertiary and Quaternary rocks (see Sprinkel, 2006), or the cross-cutting relation of the inferred fault zone proposed by Sprinkel (2014). The South Flank fault is an understudied fault. Ritzma (1969), Hansen (1986), and Bradley (1995) included the South Flank fault in their studies. Within the Ice Cave Peak, there is a splay of the South Flank fault that is a large normal fault (Plate 1A). Less than a mile to the north of the Ice Cave Peak quadrangle the main strand of the South Flank fault is also mapped as a large normal fault (Sprinkel, 2006). However, the fault has been mapped as having other portions that exhibit reverse or thrust movement (Sprinkel, 2002).

Haddox (2005) described the Deep Creek fault zone as a segmented array of northwest-southeast trending faults with accompanying half-grabens and horst-and-graben pairs. Most of the faults within the Deep Creek fault zone have near vertical dips. Bradley (1995) established that the faults within the Deep Creek fault zone are younger than the Late Cretaceous-Early Paleocene rocks that they cut. He also observed that the faults were buried under the Bishop Conglomerate and thus older than Tertiary (Oligocene). Recent mapping of the Dry Fork and

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Lake Mountain quadrangles, just to the east of the Ice Cave Peak quadrangle, confirms that most of these faults are older than the Oligocene Bishop Conglomerate, but several of them do cut up through the Bishop Conglomerate (Haddox et al., 2010; personal comm., Bart Kowallis, 2014) indicating that activity on the these faults continued into the Oligocene. The Deep Creek fault zone is thought to have formed during approximately the same time period as the South Flank fault (Haddox, 2005; Haddox et al., 2005).

Data Collection

Over 50 new fault motion indicators (slickenlines, slip directions, and fault plane orientations) were measured on outcrops in the quadrangle. These were entered into the FaultKinWin computer program (Allmendinger, 2015) to produce stereonet fault and contour diagrams as well as fault plane solutions and kinematic axes. Slickensided surfaces were not overly abundant within the Ice Cave Peak quadrangle and preservation was likely limited due to a number of factors. These perhaps include: the brittle nature of many of the faulted formations, their solubility's and degree of weathering, limited surface exposures of the faults due to overlying younger units, and possibly erosion along north-northwest trending canyons and gullies parallel to faults obscuring their presence (personal comm., Ron Harris, 2014).

The fault motion indicators that we were able to collect come mostly from newly mapped faults in the southern and central part of the quadrangle (see Figure 14 for slickensided locations within the Farm Creek brecciated zone), and from the previously mapped Ice Cave Peak fault. Not all of the slickensided surfaces were conclusive of movement direction. Furthermore, some

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fault surfaces have preserved two periods of movement observed as two directions of slickenline on a single surface.

Fault Descriptions and Kinematic Data

Mapping in the Ice Cave Peak quadrangle revealed several previously unmapped faults, as well as confirming the existence of faults shown on earlier mapping by Sprinkel (2006). In this section we will describe and present data collected from each of these faults including their location, exposure, and the type or types of movement recorded on each fault.

The Farm Creek Brecciated Zone

Two of the new faults, the Farm Creek North and Farm Creek South faults form the boundaries of a broader brecciated fault zone. The extent of this brecciated zone is unknown, due to burial under younger strata; however, the zone has a topographic expression that can be observed on aerial photos and DEMs for at least one and a half miles (Figure 14). The entire zone, and the two mappable faults within the zone are northwest-southeast trending, similar to the overall trend of the Deep Creek fault zone to the east. The zone crosses and is exposed on both sides of the Farm Creek Canyon. Slickenline trends and plunges, along with slip movement directions, where possible, were measured from slickensided surfaces within this zone but not on either of the main faults. The compiled kinematic data for this zone show that the surfaces had mostly oblique movement and include normal faulting with some left- and right-lateral movement, as well as thrust faulting with some left- and right-lateral movement. This is also similar to the Deep Creek fault zone (Figure 15). Eighty-three percent of the data show left-oblique slip while 17% show right-oblique slip movement. The faults within the zone are vertical

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to near vertical, which produce linear surface scars, even when they cross canyons and uneven topography (Plate 1A).

To show the data from this zone, we began by plotting separately the two major fault types (oblique normal and strike-slip) found within the Farm Creek brecciated zone into two main fault plane solution diagrams (Figure 15) using with the FaultKinWin computer program (Allmendinger, 2015). The orientation of the maximum horizontal stress direction acting on the Farm Creek brecciated zone was also calculated. Oblique normal slip motion indicators give an orientation of maximum horizontal stress in the northeast-southwest direction (Figure 16) consistent with other studies in the area (Gregson and Erslev, 1997; Johnston and Yin, 2001). Strike-slip motion indicators showed a calculated maximum horizontal stress orientation in a more east-west to slightly northeast-southwest direction (Figure 16).

Farm Creek South Fault

The Farm Creek South fault is located approximately 0.3 miles north of the Farm Creek gauging station and is the northern boundary of the Farm Creek brecciated zone. Kinematic indicators show both right- and left-oblique movement on this fault, but the majority of indicators measured favor normal left-oblique movement (Figure 17). This fault is entirely contained within the Weber Sandstone and does not cross-cut any formation contacts. The slickenlined surfaces found for this fault were predominately on the eastern side of Farm Creek Canyon where they tend to be preserved close to the creek where erosion has freshly exposed the fault plane. The fault plane has a dip of approximately 48° mainly to the N.E. The average rake is 40° mainly to the east (Figure 17).

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Some surfaces along the Farm Creek South fault have northeast-southwest strikes perpendicular to the general trend of the mapped faults. These are smaller oblique slip planes that are often found along fault zones (Grant and Kattenhorn, 2004) and are not unexpected in a left-stepping echelon fault zone (Figure 15) (Crider, 2001; Grant and Kattenhorn, 2004).

The Farm Creek South fault is well located at the bottom of Farm Creek Canyon and becomes approximately located on the east side about 100 feet up the canyon side and then disappears beneath the Brennan Basin Member of the Duchesne River Formation. On the west side of the canyon the fault is well located to the crest of the canyon wall, then becoming approximately located as it continues further west through less well-exposed outcrops (Plate 1A).

Farm Creek North Fault

The Farm Creek North fault is located 0.2 miles north of the Farm Creek South fault (Plate 1A). This fault trends northwest-southeast and is mappable for up to one and a half miles. The amount of displacement is not known. There is no noticeable thickening or thinning of either the Weber Sandstone or Morgan Formation near the fault. Slickenlined surfaces are found on both the east and west side of Farm Creek Canyon. On the eastern side of the canyon the slickenlined surfaces are limited to the base of the canyon and often found in and around small caves that form along the fault plane. On the western side of the canyon slickenlines can be found alongside a large talus field that covers and parallels the Farm Creek North fault. Slickenlines were also found further west near Sulphur Spring. Right- and left-oblique strike-slip, as well as normal and thrust, movement directions were determined from the kinematic indicators on this fault. The average slip direction, however, was left-lateral strike-slip as shown

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by the rake of the slickenlines on these surfaces, which was shallow at approximately 7° from horizontal to the east. The averaged a dip of 63° to the southwest (Figure 18).

Fault and fracture planes that are oblique to the dominate northwest-southeast trend are also present along the Farm Creek North fault. They are similar to those found along the Farm Creek South fault.

The best exposures of the Farm Creek North fault are to the east of Farm Creek Canyon. This fault is well located along its eastern end, is buried under the Brennan Basin Member of the Duchesne River Formation in its central portion, and west of Farm Creek it is somewhat covered and approximately located until it disappears beneath the Bishop Conglomerate.

Farm Peak Fault

The northernmost new fault, the Farm Peak fault, is located about 0.4 miles southwest of Farm Peak, 0.4 miles northeast of the Farm Creek North fault, and one mile north of Sulphur Spring. It has a mappable extent of approximately a quarter mile. Like the Farm Creek North and South faults, it too trends northwest-southeast; however, it has a slightly more northerly trend than those faults. The Farm Peak fault was first discovered during examination of the area using aerial photos. It is a left-lateral strike slip fault contained entirely within the Morgan Formation. No fault kinematic indicators were found in the field for this fault. The displacement appears to be minimal and is less than 200 feet based on offset beds. The fault is well located for a few hundred feet and then is obscured by vegetation and cover.

Elevation 9821 Fault

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The northernmost fault, Elevation 9821 fault, is a normal fault located just to the southwest of an unnamed peak with an elevation of 9,821 feet, or about 0.5 miles north of Ice Cave Peak. The fault trends northwest-southeast and is approximately one mile long running for most of its length along the northern edge of a large landslide. The offset on this fault is unknown but could be upwards of 500 feet based on the offset of the Madison Limestone. From aerial photos, the fault can be approximately located for a short distance on its western end within the Red Pine Shale, but it quickly becomes obscured by talus and debris farther to the west. The Elevation 9821 fault is buried by piedmont gravels to the east. Strike measurements were taken on the fault plane; however, no slickenlines were found probably due to the dissolution of exposed surfaces of the Madison Limestone.

Ice Cave Peak Fault

The Ice Cave Peak fault is located 1,000 feet to the north of Ice Cave Peak. It is the longest and best-exposed fault within the quadrangle. Four miles of it are mapped within the field area, and it continues to be mappable outside the quadrangle to the northwest (Sprinkel, 2006). The Madison Limestone on the west side of Whiterocks Canyon is offset by approximately 1,430 feet dip-slip, with 600 feet throw, and 1300 feet heave (Plate 1A). Slickenlines and movement indicators that are preserved in a sandstone unit within the Red Pine Shale for this fault and were measured by us, contributing eleven data points, and also by Haddox and others (2005). A large talus slope is made up mostly of slickensided boulder- to cobble-sized fragments of sandstone just below the in-place slickenlines. To the west, the Ice Cave Peak fault is buried by piedmont gravels and it is buried by alluvium and colluvium at lower elevations along Whiterocks Canyon and by piedmont gravels farther east. The average

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dip of the fault is 82° to the north and the average rake of slickenlines is near 84° to the west (Figure 19). As with the Farm Creek faults, minor fault planes that are approximately conjugate to the main fault were also found and kinematic data collected from them.

Bridger Canyon Fault

The Bridger Canyon fault is located along Bridger Canyon at its eastern end and extends west to the western side of Whiterocks Canyon (Plate 1A). On the western side of Whiterocks Canyon the Bridger Canyon fault has nearly 540 feet of dip-slip, with 200 feet throw, and 500 feet heave in the Madison Limestone. No slickenlines were found for this fault. It is buried under alluvium to the west and in the lower elevations of Whiterocks Canyon (Plate 1A). It is well located in both the Red Pine Shale and the Madison Limestone. To the east, we lose the surface expression of the fault under the Bishop Conglomerate. The Ice Cave Peak, Bridger Canyon, and Elevation 9821 faults are thought to be splays off of the South Flank fault.

Interpretation and Discussion

As can be observed in Figure 4, normal and thrust faults with components of left lateral strike-slip are present throughout the Deep Creek fault zone. We see both left- and right-lateral oblique movement within the Farm Creek brecciated zone in Ice Cave Peak quadrangle. Furthermore, in at least one case, at Ice Cave Peak, both of these fault types were observed in slickenlines on the same fault surface. The normal faulting appears to be overprinted by the strike-slip (Haddox, 2005), which suggests earlier normal movement followed by strike-slip movement. So, how does this variety of different fault types form in one fault zone? Does it represent different periods of faulting? Or can it be explained within a single phase and system of

faulting? In our mapping, we have observed that most of the faults do not cut the Oligocene Bishop Conglomerate but are confined to pre-Oligocene strata. To the east of Ice Cave Peak quadrangle the Deep Creek fault zone appears to have faults that indeed do cut the Bishop Conglomerate (Douglas Sprinkel, personal comm., 2015). Observation from within the Ice Cave Peak quadrangle suggests that the fault movement is of Laramide age but there could be an extended period of movement along some of these or related faults (personal communication Doug Sprinkel, 2015). The following section will include interpretations of data and discussion of their implications.

Due to the similarities in the fault trends and kinematic indicators on faults discussed in this report, we are extending the Deep Creek fault zone west into the Ice Cave Peak quadrangle. Several springs (Sulphur Spring, Bench Spring, Upper Big Tom Hollow Spring, Corral Spring, etc.) within the Ice Cave Peak quadrangle form a linear trend that connects the Farm Creek brecciated zone with previously mapped portion of the Deep Creek fault zone to the east (Figure 15). This may indicate that the brecciated zone continues off to the east underneath younger cover.

To better understand the movement history observed within Ice Cave Peak quadrangle and the Deep Creek fault zone, we look to the bigger picture of the Uinta uplift. Johnston and Yin (2001) observed left-slip transpressional deformation with the maximum compressive stress direction oriented northeast-southwest along the north flank of the Uinta Mountains (Figure 16). They were uncertain, however, as to whether the Uinta uplift was bounded by two left-slip transpressional structures (one on the north and one on the south) or by one reverse left-slip transpressional structure on the north and one thrust structure on the south (Johnston and Yin, 2001). Our data suggest that the Uinta uplift is bounded by left-slip transpressional structures on

both the north and south flanks. The model created by Johnston and Yin (Figure 20) predicts that the left-lateral strike-slip motion in the Farm Creek brecciated zone and in most of the Deep Creek fault zone is to be expected. This model also accounts for the mixed oblique (thrust and normal) components of the Deep Creek fault zone through east-west movement causing shortening in some orientations and extension in others.

Furthermore, having the Uinta uplift bound on the north and south by thrust faults with a left-lateral strike-slip component supports the work of Brown (1988) who concluded that the Colorado Plateau to the south of the Uinta Mountains is softer than the more brittle Wyoming Province to the north. This contrast in rheology, he proposed, would result in the Laramide structures to the north and east of the Uinta Mountains providing a buttress or barrier that would produce transpressional shortening of the Uinta uplift (Brown, 1988; Johnston and Yin, 2001). Furthermore, this left-slip transpressional deformation would lead to the development of the en echelon anticlines found at the core of the Uinta uplift (Johnston and Yin, 2001).

Gregson and Erslev (1997) collected slip data at the eastern end of the Uinta uplift on both the north and south flanks. Their results showed that the maximum horizontal stress directions for the Laramide orogeny around the eastern end of the uplift were variable (Figure 16). For example, on the north flank they found a maximum horizontal stress direction that was parallel to the regional stress orientation of northeast-southwest, but on the south flank they found the orientation to be more north-south, even slightly northwest. They proposed three possible explanations for this difference, one of which is compatible with the left-lateral oblique motion we observe on the south flank as well as the similar motion observed by Johnston and Yin (2001) on the north flank. This was Gregson and Erslev's (1997) second proposed model, which suggested that the anomalous north-south orientation for the maximum horizontal stress

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on the eastern end of the south flank of the Uinta uplift might be simply a localized structural transition between the Uinta arch and the White River uplift.

Another contributing factor for the variable movement within the Deep Creek fault zone is the idea that the Uinta Basin-Mountain boundary thrust is not a continuous sinusoidal feature on the southeastern flank of the Uinta uplift. To the east of Ice Cave Peak quadrangle near Asphalt Ridge the Uinta Basin-Mountain boundary thrust is thought to be trending more north-south (Figure 12). Rather than a sinusoidal curvature of the Uinta Basin-Mountain boundary thrust, we suggest that perhaps the fault is composed of segmented blocks that occupy different north-south vertical planes (Figure 21). The segments may represent zones of Precambrian weakness, similar to the main bounding thrusts of the Uinta Mountain block. Differential movement along these boundaries have created the faults in the younger, overlying strata. Strike-slip, normal, thrust, and oblique faulting would be expected to form as a result of variable relative movement of the underlying segmented thrust blocks (Figure 21).

In addition we propose that the maximum horizontal stress was rotational over time. As Larmide compression begins the orientation is northeast-southwest. Then due to buttressing of the Uintas by the White River uplift and the more brittle rocks of the Wyoming Province to the north the local maximum horizontal stress orientation rotates to an east-west direction, as suggested by Brown (1988). Finally, as Laramide compression relaxes the maximum horizontal stress becomes northwest-southeast, perpendicular to the previous regional stress orientation (Figure 16). This type of stress rotation has been observed in approximately the same time frame to have occurred in the Basin and Range Province, although perhaps for different reasons (Ren et al, 1989; Kowallis et al. 1995).

Conclusions

The Deep Creek fault zone is made up of an array of northwest-southeast trending faults that extend as far west as the Farm Creek brecciated zone in the Ice Cave Peak quadrangle. Faults in the Deep Creek fault zone record a variety of directions of movement. These faults formed during the Laramide orogeny, and perhaps during a phase of relaxation following that orogeny, possibly in a transfer zone between blocks of the main Uinta Basin-Mountain boundary thrust. We have looked at several different models to explain the variable orientation of the maximum horizontal stresses for this region during the Laramide, but our preferred model includes the following elements. First, the northeast-southwest orientation present within Ice Cave Peak quadrangle and within the rest of the Deep Creek fault zone to the east is parallel with regional Laramide stress orientation and has been documented in other nearby related studies (Figure 16). The second orientation observed in the Deep Creek fault zone, which is approximately east-west perhaps represents a rotation of the stress field due to buttressing of the lateral motion of the Uinta Mountain block by the White River uplift to the east. Finally as Larmide compression ended the maximum horizontal stress rotated to a more northwest-southeast orientation due to relaxation following the orogeny.

Figures

Figure 1: Figure modified from Google earth; Index Map of Ice Cave Peak quadrangle located on the southeastern flanks of the Uinta Mountains. 17 miles northeast of Roosevelt, UT. Yellow star indicates field location.

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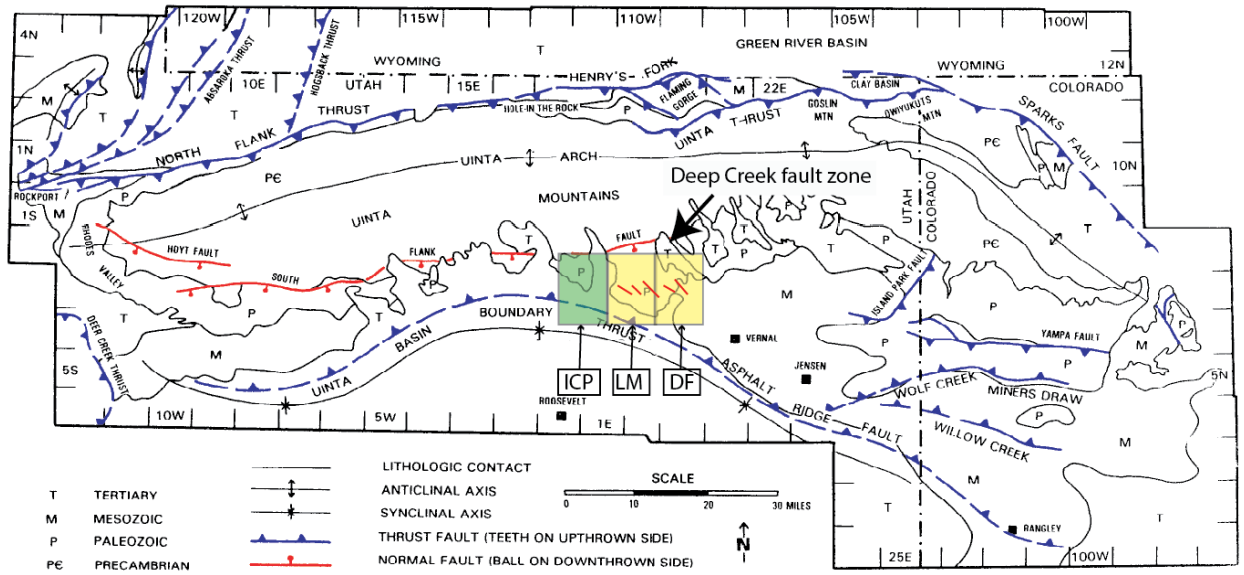
Figure 2: Structural index map of Ice Cave Peak (ICP), Lake Mountain (LM), and Dry Fork (DF) quadrangles. Green rectangle indicates location of the study area. Notice proximity to South Flank Fault Zone and Uinta Basin-Mountain boundary fault.

Figure 3: Nugget Sandstone with field book for scale. High angle cross- stratification with hydrocarbon saturated sandstone in contact with less saturated portion. This photo was taken on the eastern side of Whiterocks Canyon near the contact of Nugget Sandstone and Carmel Formation.

Figure 4: Looking northwest across Whiterocks River at the current tar sands open mine. Red rectangle highlights the tar sands pit mine within the Nugget Sandstone. Production of these hydrocarbons began in 1940s and continues today. The tar sands are currently used for paving highways within Duchesne County. Notice the red beds of the over lying Brennan Basin Member of the Duchesne River Formation. Oil is thought to have been in place within the Nugget Sandstone prior to Brennan Basin Member deposition because the Brennan Basin Member contains tar saturated clasts on Nugget Sandstone within its beds.
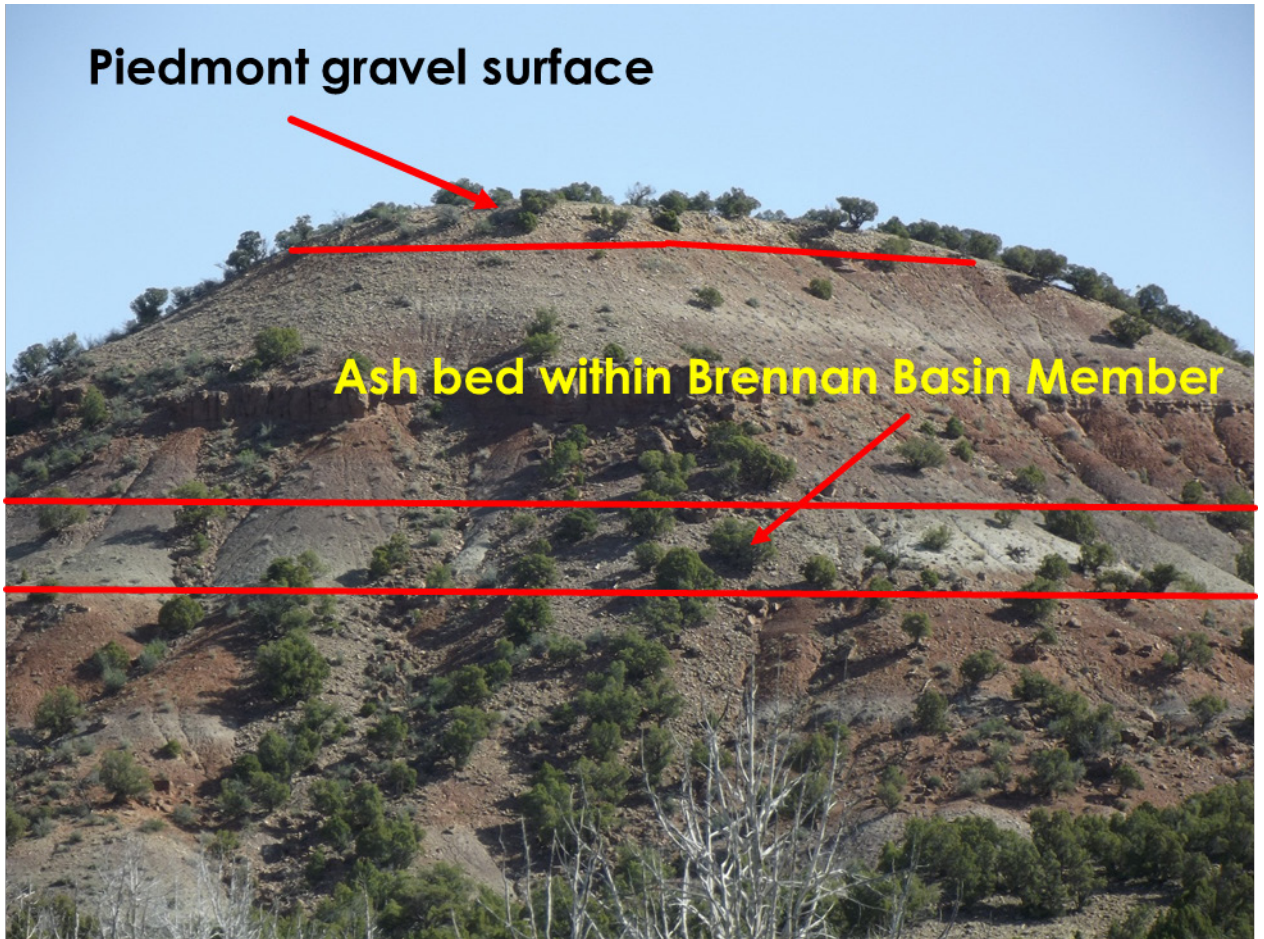
Figure 5: Looking west within the southern portion of the Ice Cave Peak quadrangle at the Brennan Basin Member of the Duchesne River Formation. Gray layers are bentonite. The formation is often capped by piedmont surfaces as seen here.

Figure 6: Bishop Conglomerate with predominately Paleozoic limestone clasts. Rock Hammer for scale. This photo was taken looking north on the ridge above Farm Creek Canyon to the east. The base of the Bishop Conglomerate (shown here) contains less Uinta Mountain Group clasts than the upper portion located in the north eastern region of Ice Cave Peak quadrangle. The Bishop Conglomerate is prone to landslides.
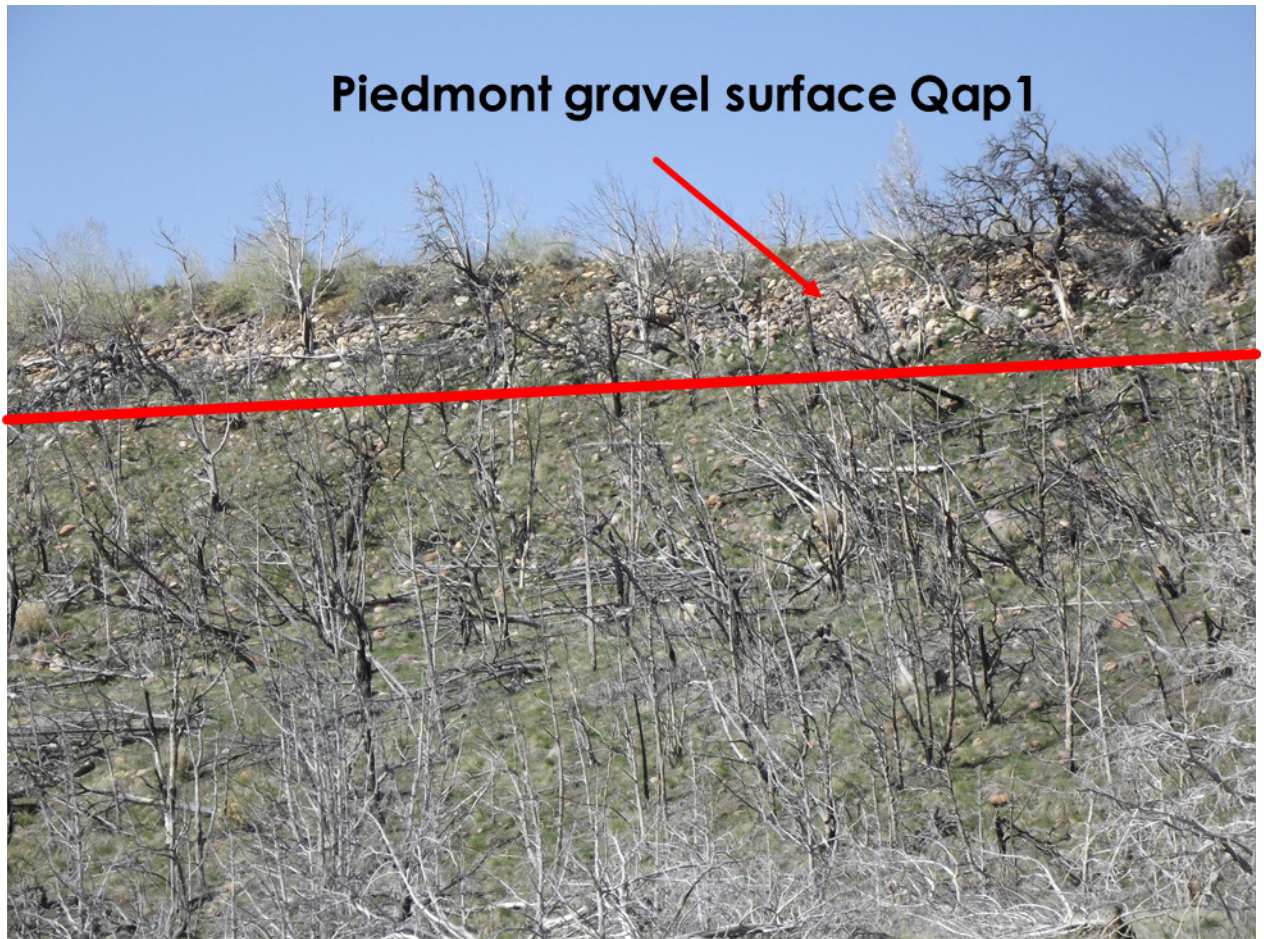
Figure 7:Qap1 in the eastern portion of the field area. Piedmont gravels capping Duchesne River Formation that is covered with burned out vegetation. This is the second lowest piedmont gravel within the Ice Cave Peak quadrangle and clast range up to two feet in diameter.
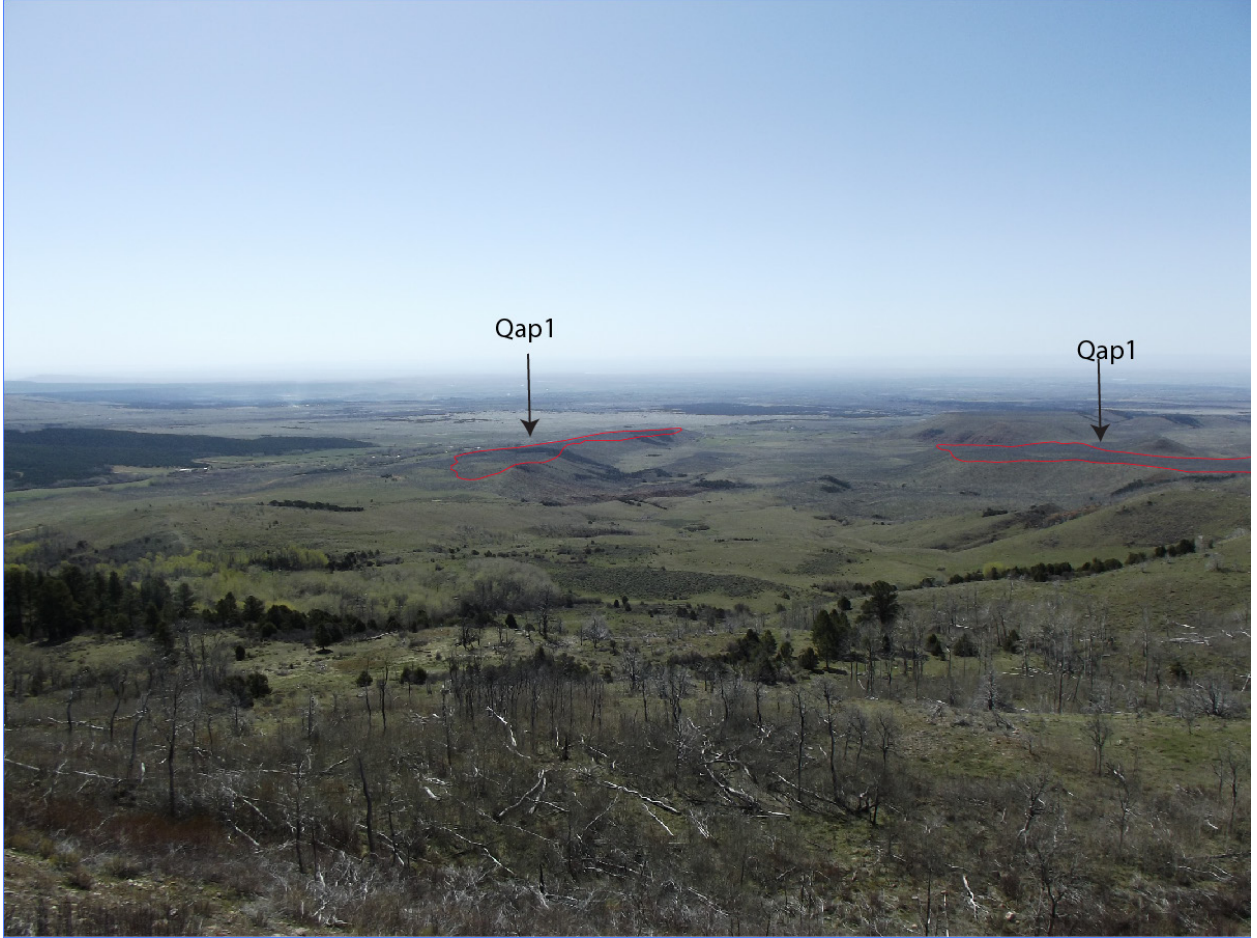
Figure 8: View looking south from within Ice Cave Peak quadrangle. Qap1 piedmont gravels can be observed in the distance. Notice the consistent elevation (7,000 feet above sea level) of the gravels.
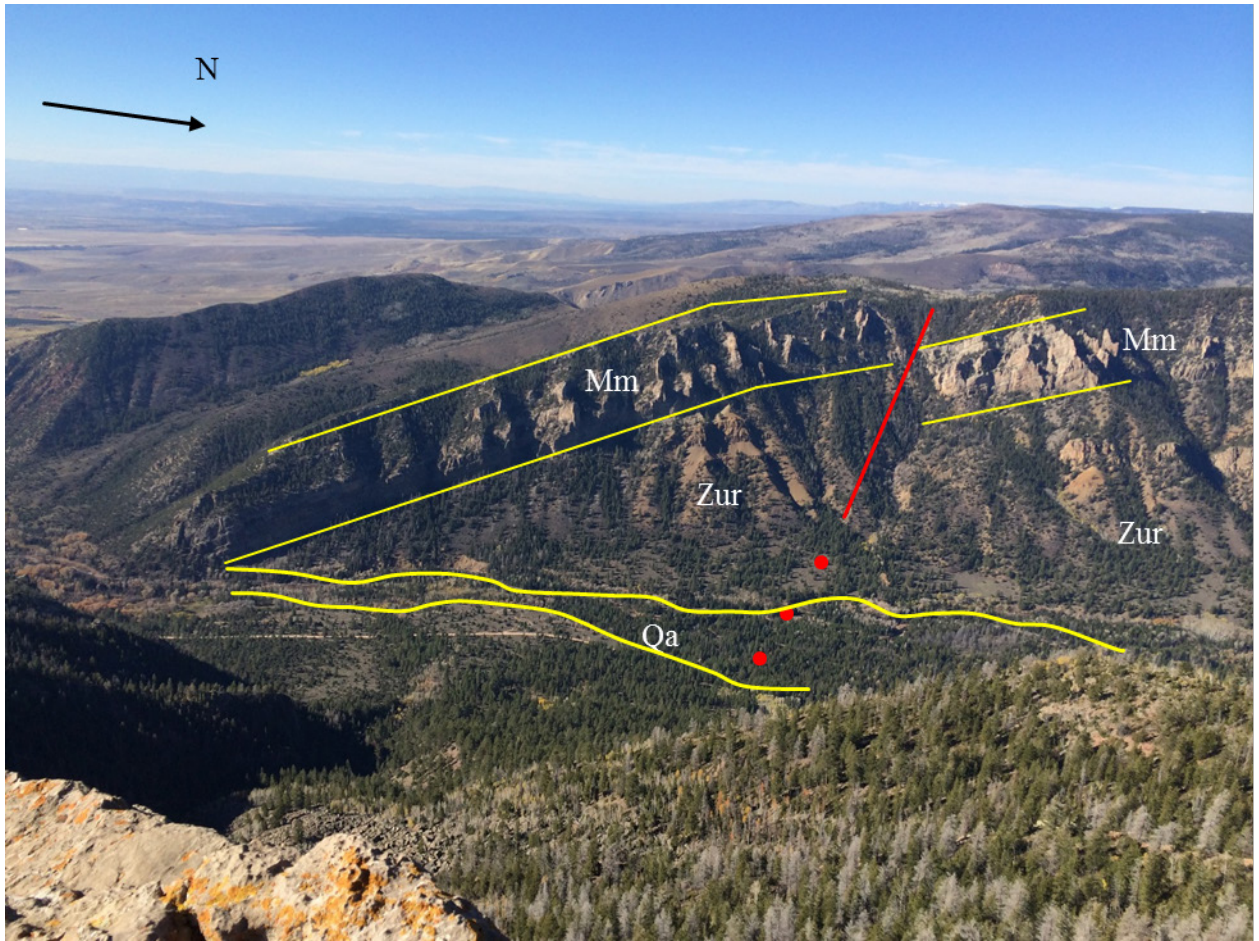
Figure 9: View from Ice Cave Peak looking west by southwest. Notice displacement in the Madison Limestone due to a splay from the South Flank fault zone. Also shown here is the angular unconformity between the Red Pine Shale and the Madison Limestone.
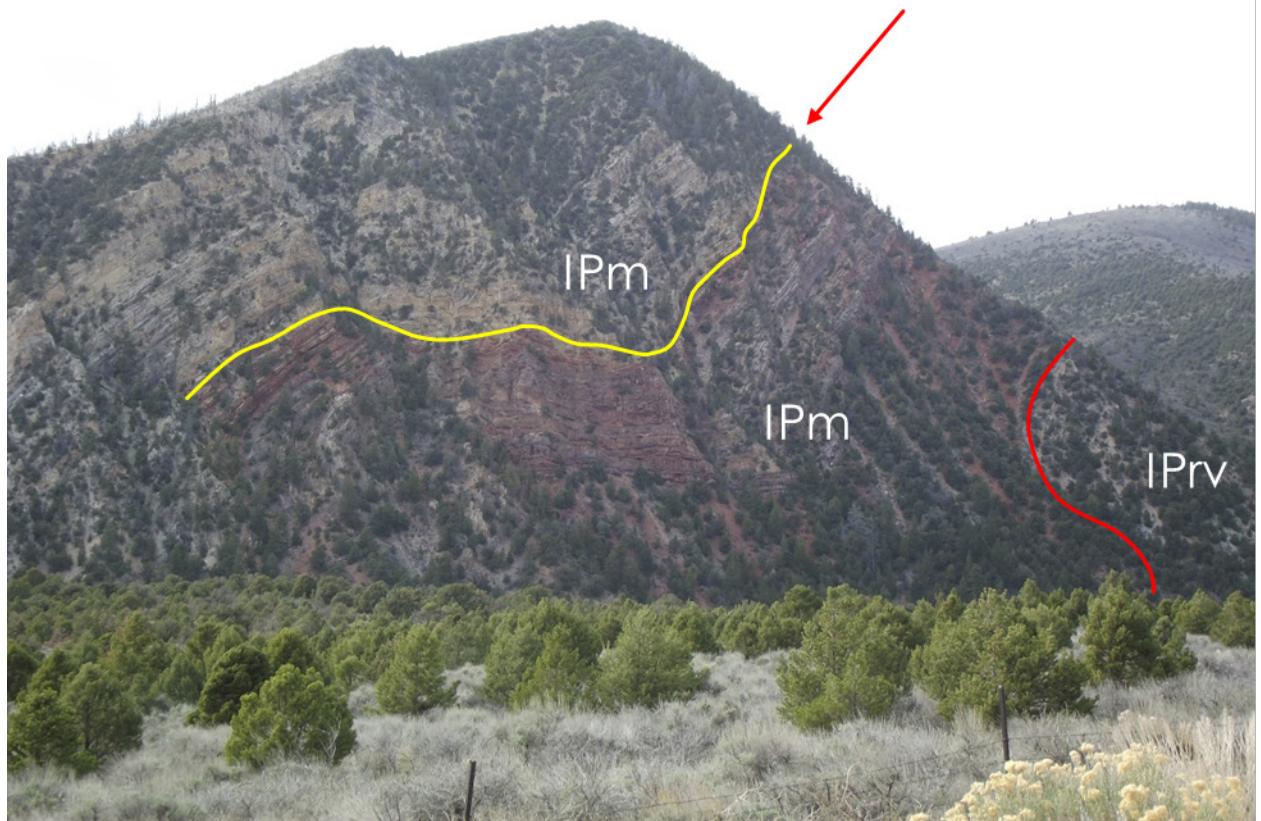
Figure 10: Looking west across Whiterocks River to the fold near the contact of the Morgan Formation and Round Valley Limestone. The yellow line indicated the fold of beds within the Morgan Formation.
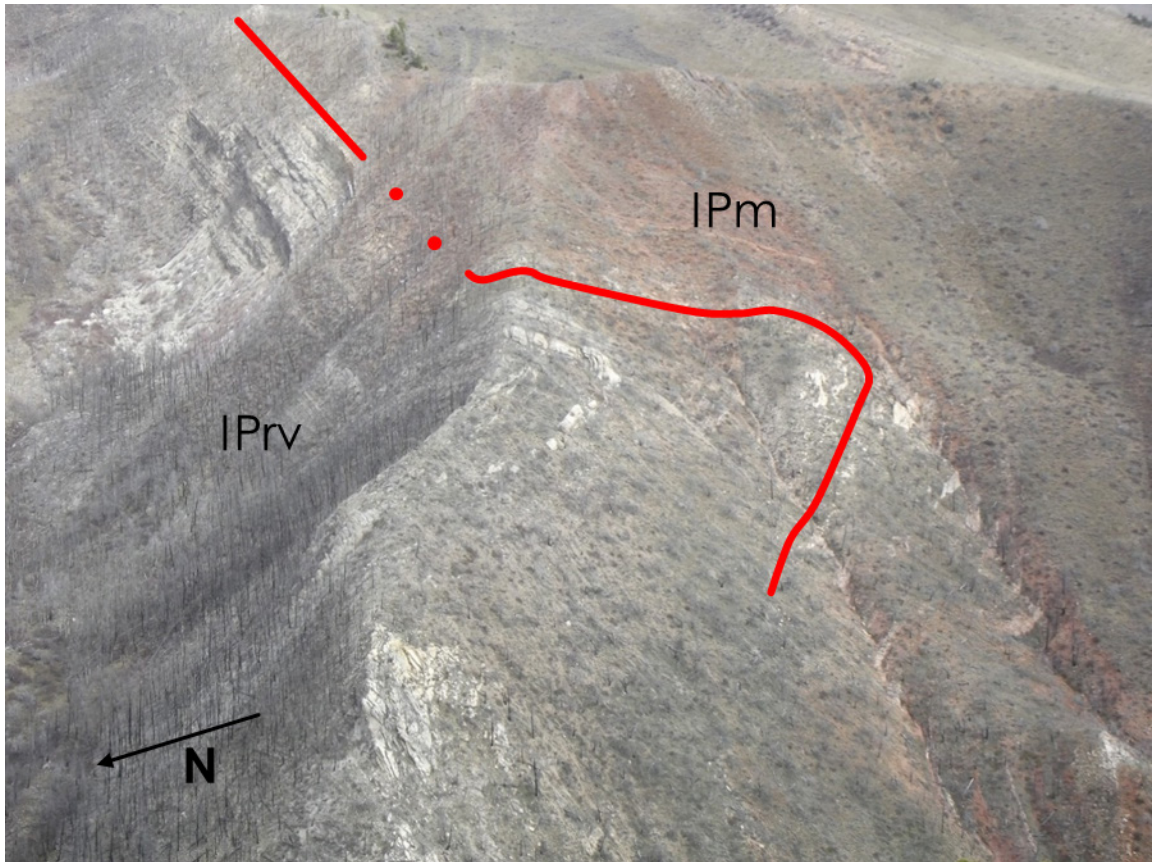
Figure 11: Round Valley Limestone cliffs (Gray ledges to the left of red line) are shown folded with red beds of the Morgan Formation. The fold is plunging slightly to the east. Red lines show folding of a single bed within the formation. The view is looking east by southeast.
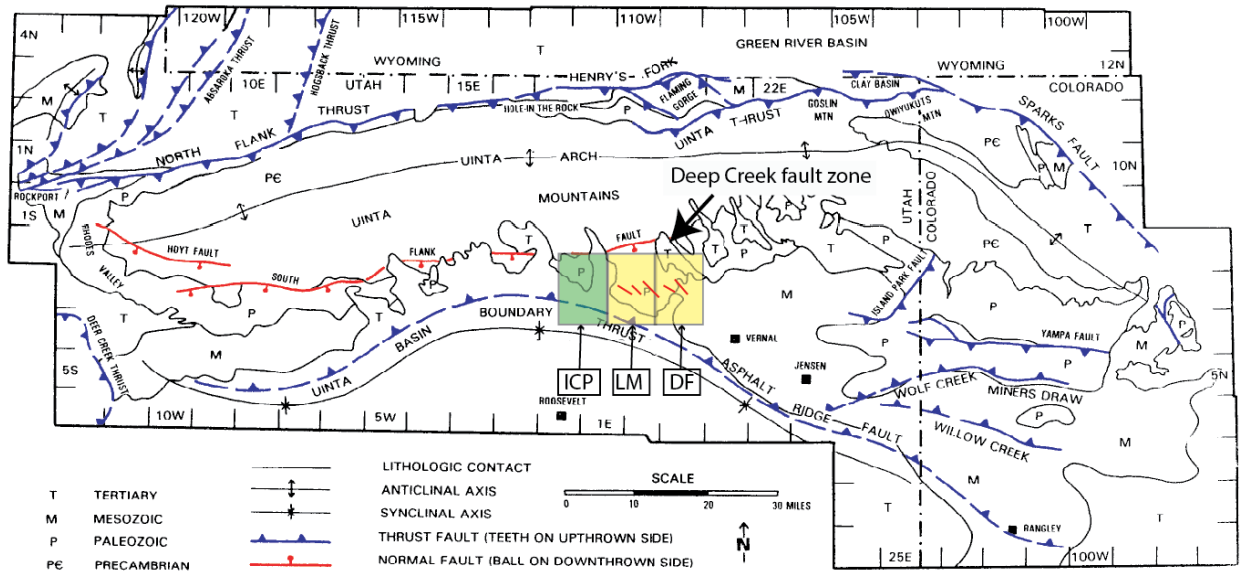
Figure 12: Structural index map of Ice Cave Peak quadrangle (ICP), Lake Mountain quadrangle (LM), and Dry Fork quadrangle (DF), showing the location of the Deep Creek fault zone within the Lake Mountain and Dry Fork quadrangles, which has been extended into the Ice Cave Peak quadrangle as a result of this project.
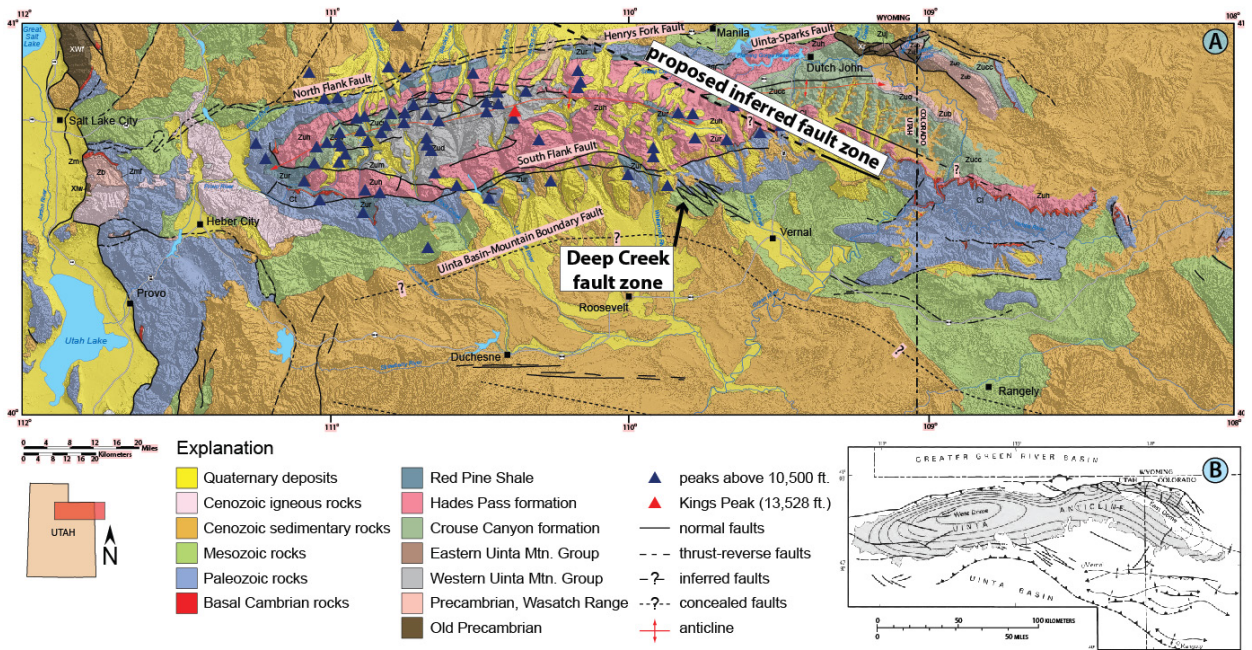
Figure 13: Map modified from Sprinkel (2014) showing location of the Deep Creek fault zone in relation to the inferred fault zone (A) separating the two anticlines of the Uinta Mountains (B).
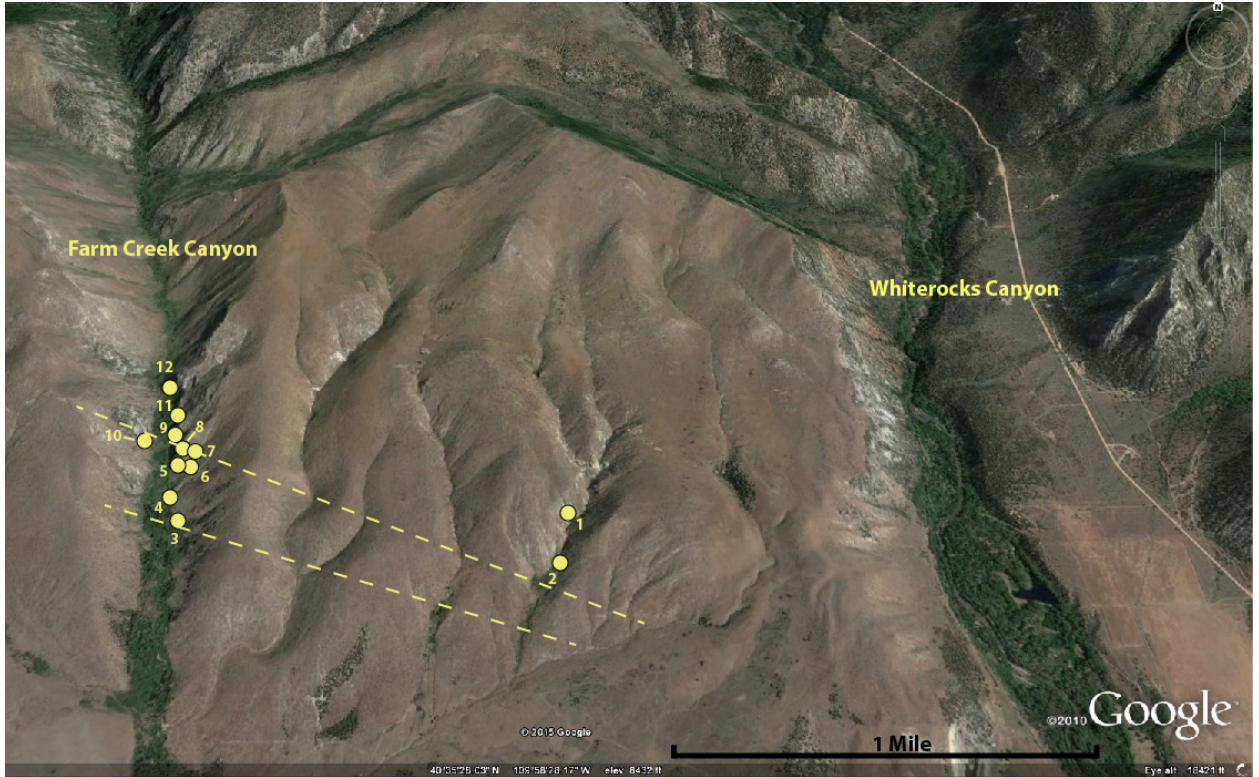
Figure 14: Figure modified from Google Earth. Locations of the slickenlines for the Farm Creek north and south faults. The dashed lines indicate visible linear features that parallel the Farm Creek brecciated zone.
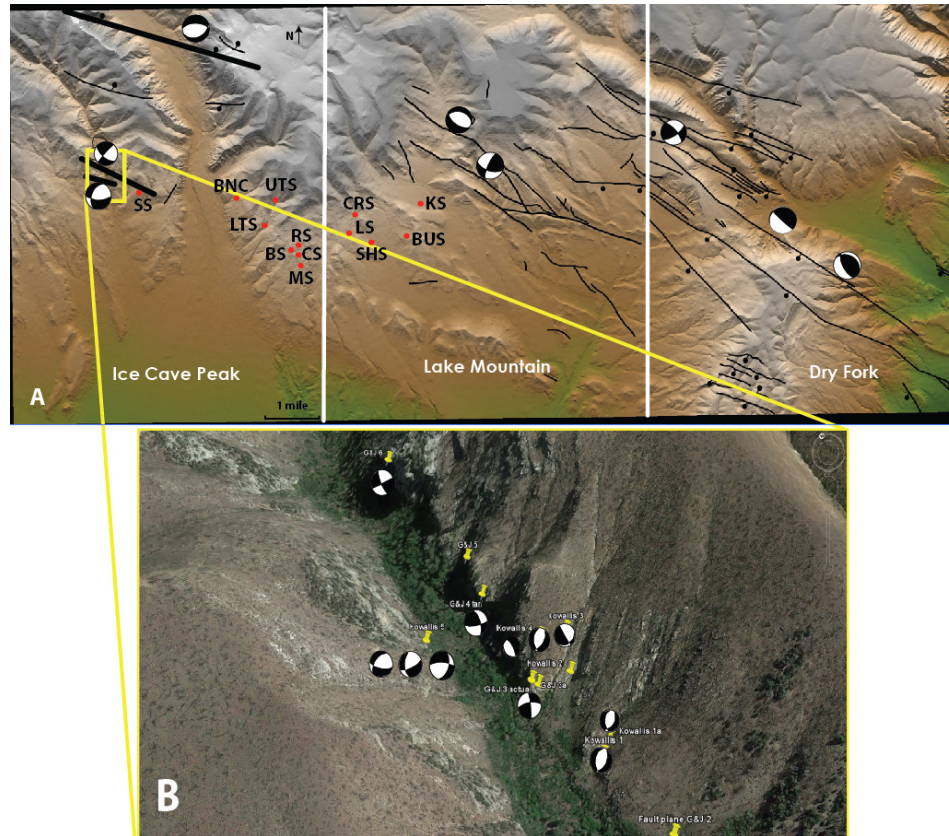
Figure 15: DEM image of Ice Cave Peak quadrangle showing an enlarged image from Google Earth that displays the location of the slickensided surfaces within Farm Creek Canyon. The variety of fault plane solutions in 5b illustrate the range of data found within the Farm Creek brecciated zone. The fault plane solutions in 5b do not represent all data collected, but all the data collected within the Farm Creek brecciated zone are represented by the two diagrams in 5a. Red dots represent springs within the Ice Cave Peak and Lake Mountain quadrangles. Sulphur Spring (SS), Bench Spring (BNC), Upper Big Tom Hollow Spring (UTS), Lower Big Tom Hollow Spring (LTS), Rat Spring (RS), Birch Spring (BS), Cuch Spring (CS), Murray Spring (MS), Corral Spring (CS), Lyman Spring (LS), Sage Hen Spring (SHS), Burton Spring (BUS), and Knoll Spring (KS) form a semi-linear trend that connects the Farm Creek brecciated zone with previously mapped portion of the Deep Creek fault zone to the east. This may indicate that the brecciated zone continues off to the east underneath younger cover.

70
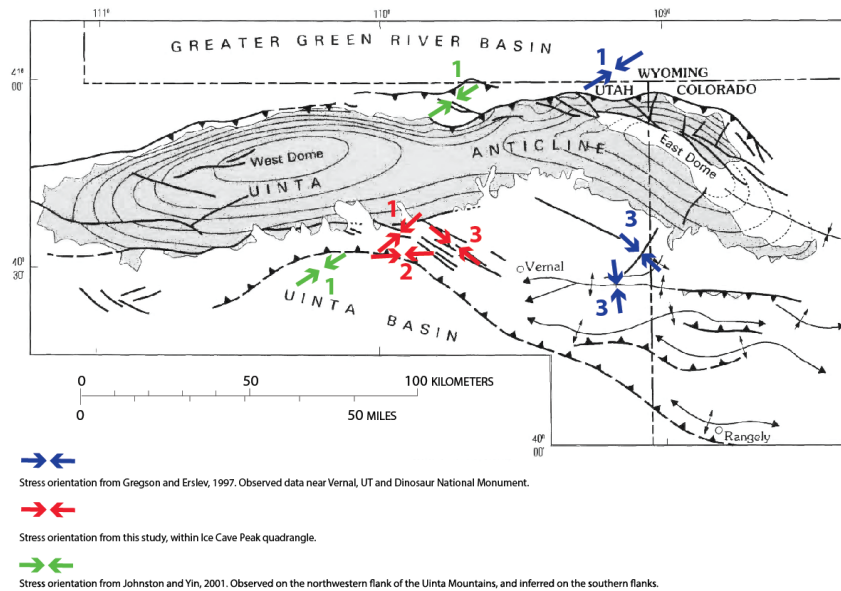
Figure 16: Map modified from Sprinkel (2014) Regional depiction of the Uinta uplift. The arrows represent Laramide regional stresses from corresponding studies. All three studies observed northeast-southwest Laramide compression. Gregson and Erslev also observed a localized transitional zone. Our observations include northeast-southwest and east-west stresses, and Johnston and Yin saw left-slip movement combined with northeast-southwest compression. The numbers represent the phases of localized horizontal stress rotation from northeast-southwest, to nearly east-west, and finally northwest-southeast.
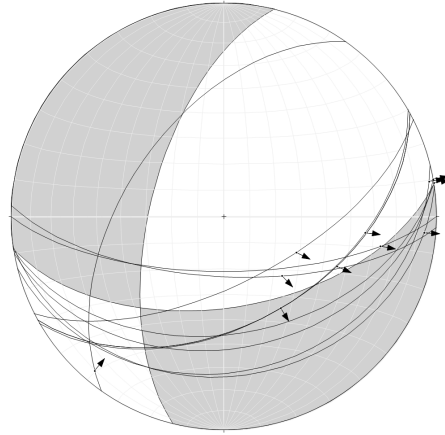
Figure 17: Fault plane solution for kinematic data collected from within the Farm Creek brecciated zone near the Farm Creek South fault. The data show near east-west trending slip surfaces with mostly normal fault movement but with a component of left-lateral strike-slip motion.
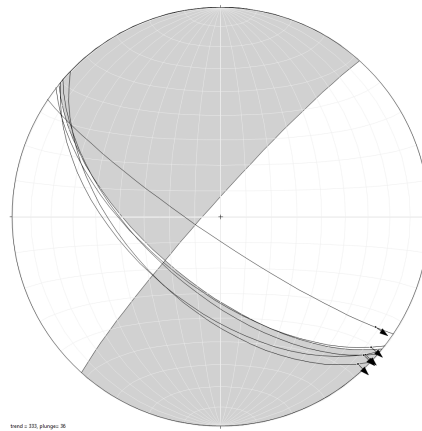
Figure 18: Fault Plane solution for kinematic data collected from within the Farm Creek brecciated zone near the Farm Creek North fault. The data show northwest trending slip surfaces with an average dip of 63° and mostly left-lateral strike-slip motion.

72

N = 11
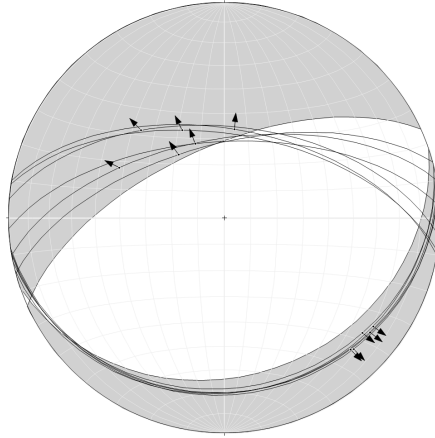
Figure 19: Fault plane solution for kinematic data collected from the Ice Cave Peak fault. Two conjugate orientations of normal slip surfaces occur at this locality. One group dips shallowly to the southeast and the other group dips more steeply at 55-60° to the north.

A. Initiation of Laramide faults along
pre-existing Proterozoic rift zone

N

regional compressional
direction

future Laramide faults
bounding Uinta Uplift

strain marker

Pz-Mz

Proterozoic Uinta
Mountain Group
in a Rift Zone
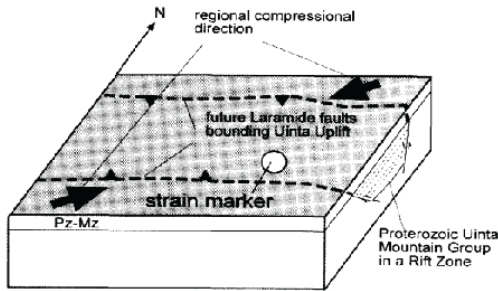
B. Development of Laramide left-slip
transpressional across the entire Uinta uplift

N

observed finite
strain ellipse

instantaneous
strain ellipse

rotation of
strain axes

Colorado Plateau

Pz-Mz

Proterozoic Uinta
Mountain Group
in a Rift Zone
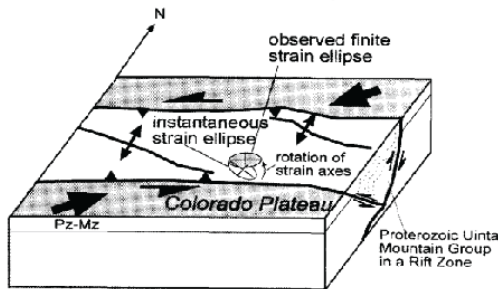
Figure 20: Figure from Johnston and Yin (2005) showing the evolution of the Uinta uplift. This illustrates the hypothesis that both the northern and southern bounding thrust faults and their associated smaller faults, like the Deep Creek fault zone, should have components of left-lateral strike-slip motion. The model also accounts for the en echelon anticlines at the core of the Uinta Mountains.
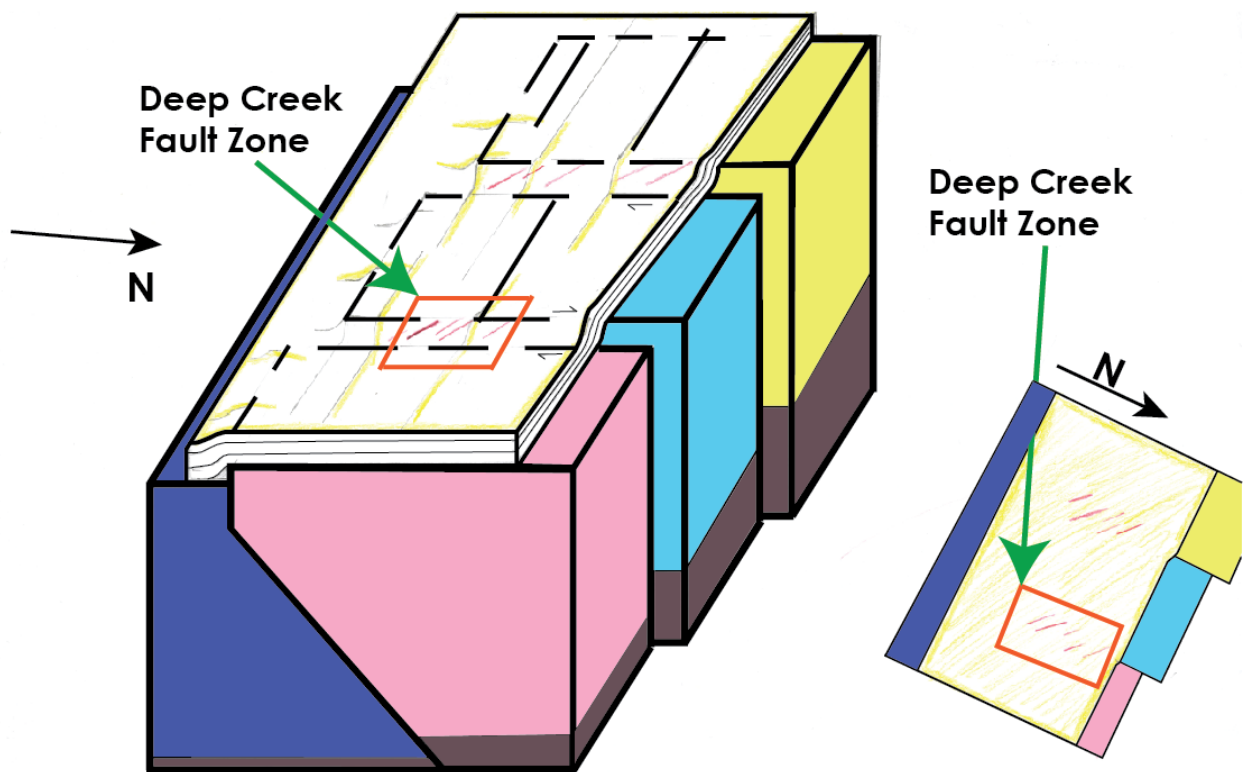
Figure 21: Model of possible segmented blocks developed by offsets of the Uinta Basin-Mountain boundary thrust along the south flank of the Uinta Mountains. The segments may be controlled by reactivation of Precambrian zones of weakness. These segment boundaries form transfer zones between thrust blocks where strike-slip, normal, and oblique-slip faults have formed depending on the relative motions of the blocks. Stress orientation from Johnston and Yin, 2001. Observed the northwestern flank of the Uinta Mountains inferred on the southern flanks.

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UNITED STATES
DEPARTMENT OF THE INTERIOR
GEOLOGICAL SURVEY

ICE CAVE PEAK QUADRANGLE
UTAH
7.5 MINUTE SERIES (TOPOGRAPHIC)
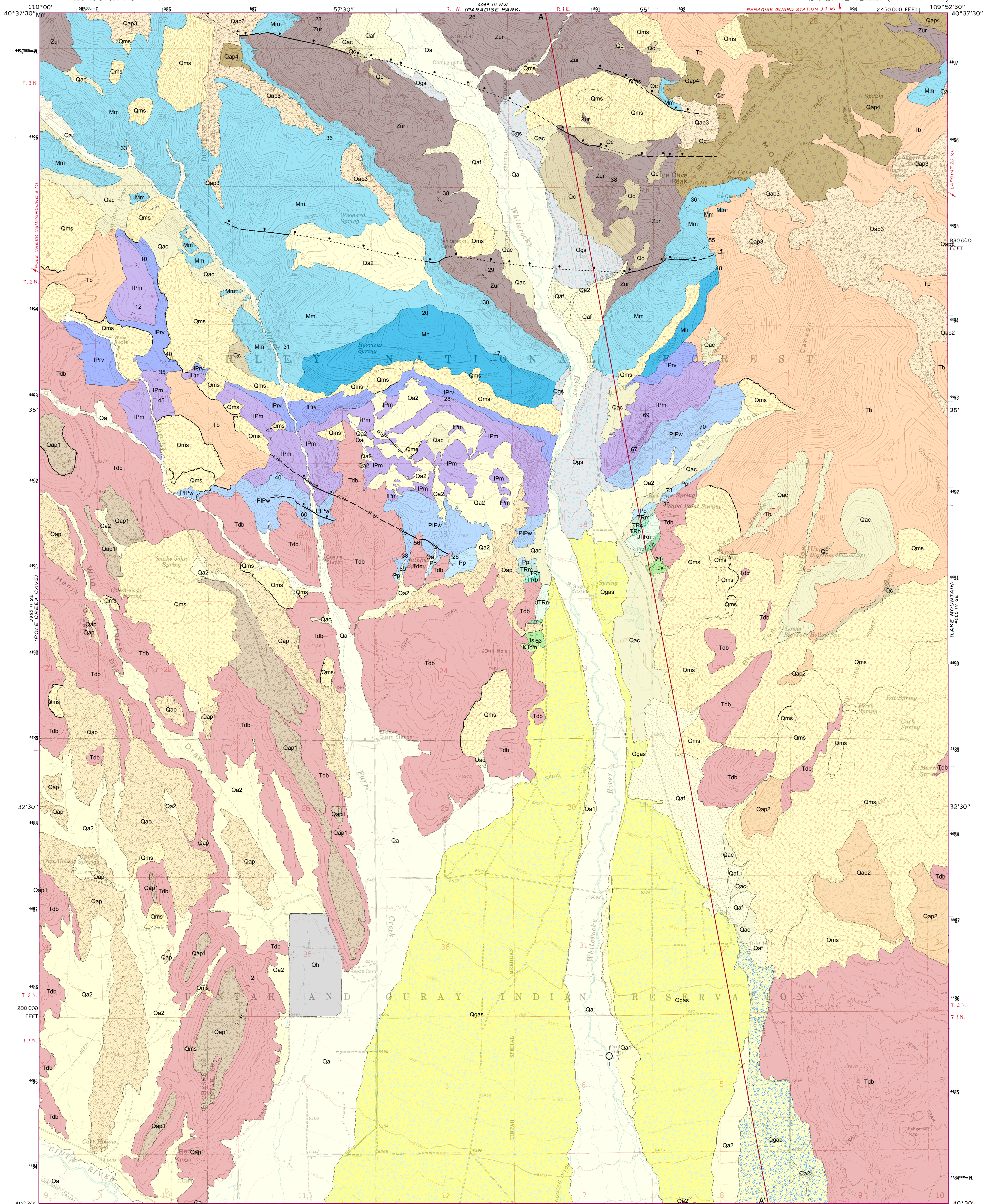
Mapped by the U. S. Forest Service
Edited and published by the Geological Survey

Control by USGS, USC&GS, and USFS

Topography by photogrammetric methods from aerial
photographs taken 1962. Field checked by USGS 1965

Polyconic projection. 1927 North American datum
10,000-foot grid based on Utah coordinate system, central zone
1000-meter Universal Transverse Mercator grid ticks,
zone 12, shown in blue

Fine red dashed lines indicate selected fence lines

Where omitted, land lines have not been established

SCALE 1:24000

CONTOUR INTERVAL 40 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

ROAD CLASSIFICATION

Light-duty _____ Unimproved dirt _____

QUADRANGLE LOCATION

ICE CAVE PEAK, UTAH
N4030—W10952.5/7.5

FOR SALE BY U. S. GEOLOGICAL SURVEY, DENVER, COLORADO 80225, OR RESTON, VIRGINIA 22092

DESCRIPTION OF MAP UNITS

FILL DEPOSITS

Qh — Human disturbances, undivided (Historical) – Unconsolidated, fine-grained sediments modified by human activities related to agriculture; not all disturbed areas were mapped; 1 to 12 feet thick.

ALLUVIAL DEPOSITS

Qaf — Alluvial-fan deposits (Holocene) – Mostly boulders, cobbles, sand, and gravel in a fine-grained (silt and clay) matrix; color of deposit reflects color of locally weathered material; unconsolidated, poorly sorted, slope forming; deposits are occasionally stratified and contain cobble and boulder lenses; fine-grained deposits are usually derived of locally weathered material; boulders and cobles are concentrated to active channels; 1 to 130 feet thick.

Qa / Qa1 / Qa2 — Alluvium (Holocene to lower Pleistocene) – Gravel, sand, and silt mixed with fine-grained matrix; color of deposit reflects color of locally weathered material; unconsolidated, well- to poorly sorted, subangular to well-rounded, occasionally stratified; active channels contain rounded cobbles and boulders; 1 to 100 feet thick.

Qap / Qap1 / Qap2 / Qap3 / Qap4 — Piedmont gravel deposits (upper to lower Pleistocene) – Boulders, cobbles, gravel, and sand in a sandy silt matrix; subangular to subrounded, poorly sorted, unconsolidated to poorly consolidated; rock fragments, boulders and cobbles are dominated by quartzite from Uinta Mountain Group; Qap and Qap1 poorly developed soil profile (Sprinkel, 2006), clasts sizes range from cobble to small boulder (up to 2 feet diameter); Qap2,3, and 4 have well developed soils (Sprinkel, 2006), larger clasts and boulders; Qap2 boulders range up to 3 feet in diameter with an increase in boulder lenses; Qap3 boulders range up to 4 feet in diameter; Qap4 boulders range to greater than 4 feet in diameter and are concentrated locally; subunits are divided by soil development (poorly to well-developed, increasing to the older units), clast size (increasing in the older units) and elevation (increasing in the older units) ; 10 to 1000 feet thick.

COLLUVIAL DEPOSITS

Qc — Colluvial deposits (Holocene) – Boulders, cobbles, and gravel slope deposits; unconsolidated, angular to subangular, poorly- to well-sorted, lithology of material varies based on the lithology of the units that weathered to form the colluvial deposit; often located in steep topography near Piedmont surfaces, Madison Limestone, and the Red Pine Shale; 1 to 30 feet thick.

GLACIAL DEPOSITS

Qgs — Smiths Fork Till (upper Pleistocene) – Coarse boulders and cobbles in a sandy matrix; boulders are quartzite and sandstone from Uinta Mountain Group, boulder dominated, angular to subrounded, moderately to well sorted; matrix is reddish-brown silty clay to pink sand; forms low ridge in Whiterocks Canyon due to armoring effect of boulders, Smiths Fork till was deposits approximately 14,000 to 32,000 ago but cosmogenic 10Be surface-exposure dating reveal that terminal moraines were abandoned by retreating glaciers before 16,000 years ago a (Laabs and Carson, 2005, Laabs and Munroe, 2009; Laabs and others, 2009) ; 1 to 130 feet thick.

Qgas — Glacial outwash of Smiths Fork age (upper Pleistocene) – Cobble to gravel-sized clasts in a sandy or silty matrix; Qgas- clasts are reddish-brown to pale pink (Uinta Mountain Group quartzite), rounded to subangular; matrix is reddish-brown to pink, pebbles to sands, unsorted, subrounded to subangular; approximately 12-32 ka (Laabs and Carson, 2005, Laabs and Munroe, 2009); 1 to 65 feet thick.

Qgab — Glacial outwash of Blacks Fork age (upper to middle Pleistocene) – Unconsolidated, clast-supported, well-rounded, mostly red sandstone and quartzite (Uinta Mountain Group) boulders to pebbles and sand deposited by meltwaters of the Blacks Fork-age glaciers; timing of deglaciation has not been directly determined in the Uinta Mountains but is thought to be roughly correlative with and to follow Bull Lake Glaciation ages, which range from 95,000 to 160,000 years as dated in other mountain ranges in the Rocky Mountains (see Phillips and others, 1997; Munroe, 2001; Laabs and Carson, 2005; Munroe and Laabs, 2009); preserved southeastern part of the quadrangle about 40 feet above Smiths Fork outwash deposits (Qgas); less than 65 feet thick.

MASS-MOVEMENT DEPOSITS

Qms — Landslide and slump deposits (Holocene to upper Pleistocene [?]) – Slump and landslide deposits which vary in scale; small slump deposits are usually caused by failure of the Brennan Basin Member of the Duchesne River Formation, as large as a few thousand square feet; large landslide deposits are usually covered with cobbles and boulders of Uinta Mountain Group quartzite, deposits are common where the Bishop Conglomerate has been deposited on less competent units; some landslides form large complexes made of several individual landslides; landslide complexes are as large as six square miles and more than three miles wide; relative ages of mass-movement deposits were not mapped because even landslides considered old and inactive actually may continue to move by slow creep, are capable of renewed movement, and pose a risk (Ashland, 2003) ; as much as 100 feet thick.

MIXED DEPOSITS

Qac — MIXED ALLUVIUM AND COLLUVIUM (Holocene upper Pleistocene) - Boulder, cobble, and gravel sized coarse-grained clasts in a sandy or silty matrix, deposits as thin blankets, deposits occur around alluvial and colluvial boundaries; cobble-sized grains to fine sediment; tend to be deposited in the Bishop Conglomerate Formation; as much as 50 feet thick.

Tb — BISHOP CONGLOMERATE (OLIGOCENE) - Polymictic conglomerate interbedded with sandstone and siltstone lenses and beds with some thin ash beds; conglomerate, boulder- to pebble-sized clasts, sandy matrix supported, poorly sorted, subangular to rounded clasts, thin- to very thick-bedded, massive; clasts of Paleozoic limestone, sandstone, and chert as well as quartz sandstone clasts from the Neoproterozoic Uinta Mountain Group; loosely cemented, moderately resistant, mostly slope forming; sandstone and siltstone, light-gray to pinkish-gray on fresh surface but weathers moderate-reddish-orange to pale-orange; sandstone, friable, fine- to coarse-grained, ledge-forming; siltstone, fine-grained, moderately sorted, ledge- to slope-forming; ash beds are white and very light-gray, thin, tuffaceous sandstone beds common in upper part of unit; ash beds from this formation have given isotopic ages ranging from 30 to 34 Ma (Kowallis and others, 2005); as much as 500 feet thick.

Tdb — BRENNAN BASIN MEMBER OF THE DUCHESNE RIVER FORMATION (EOCENE) - Conglomerate interbedded with sandstone, siltstone, mudstone, and bentonitic ash beds; conglomerate, boulder (up to 5 feet in size) to gravel sized clasts in a moderate-red to pale-yellowish-orange sandy to muddy matrix, very poorly sorted, subangular to rounded clasts, thick- to very thick-bedded, massive; dominantly clasts of Paleozoic limestone (dark gray to moderate gray) with some clast of moderate-reddish-orange sandstone and multi-colored (Mesozoic), and rare clasts of reddish quartz sandstone from the Neoproterozoic Uinta Mountain Group; well cemented, resistant, cliff- to ledge-forming; pebble conglomerate beds form scour channels, pebble- to gravel-sized clasts in a coarse sandy matrix, subrounded to rounded grains, poorly sorted, massive, thick- to very thick-bedded, ledge-forming ; sandstone, pale-yellowish-orange to grayish-pink on fresh surface but weathers moderate-reddish-orange, coarse- to fine-grained, moderately sorted to well sorted, very thick- to thin-bedded, massive to planar laminated, sandy lenses pinch and swell and scour underlying mudstone beds; sandstone lenses can be as thick as 15 feet and extend 1/4 mile distance, well-cemented, resistant, cliff- to ledge-forming, some beds contain traces of bitumen probably derived from the Green River Formation (Blackett, 1996); siltstone and mudstone, light-brown to pale-yellowish-orange on fresh surface but weathers moderate-red to dark-yellowish-orange, thick- to thin-bedded, massive to planar laminated; muddy units contain ripples and are slope-forming; bentonitic ash beds, yellowish-gray to light-gray on fresh surface but weathers grayish-yellow-green; can contain pebble to sand-sized lithics, range from 2-12 feet thick, ash beds from this member give isotopic ages ranging from 40-41 Ma (Kowallis and Sprinkel, unpublished data); 280 to 580 feet thick .

Tu — UINTA FORMATION (EOCENE) - Shown only in cross section.

Tg — GREEN RIVER FORMATION (EOCENE) - Shown only in cross section.

Tw — WASATCH FORMATION (EOCENE TO PALEOCENE [?]) - Shown only in cross section.

Kmv — MESAVERDE GROUP (UPPER CRETACEOUS) - Shown only in cross section.

Km — MANCOS SHALE (UPPER CRETACEOUS) - Shown only in cross section.

Kf — FRONTIER SANDSTONE (UPPER CRETACEOUS) - Shown only in cross section.

Kmo — MOWRY SHALE (UPPER TO LOWER CRETACEOUS) - Shown only in cross section.

Kd — DAKOTA SANDSTONE (LOWER CRETACEOUS) - Shown only in cross section.

KJcm — CEDAR MOUNTAIN AND MORRISON FORMATIONS – Mapped together in quadrangle due to poor outcrops of these units; see below for description, age, and thickness of each formation; combined thickness is 500 feet.

Kc — CEDAR MOUNTAIN FORMATION(LOWER CRETACEOUS, Albian to Barremian) - Mudstone interbedded with pebble conglomerate lenses and beds; mudstone moderate-red, pale-purple, light-gray, or pale-greenish-yellow; contains abundant calcrete beds that weather out as concretions, slope forming, and poorly exposed; pebble conglomerates, gravel- to sand-sized rock fragments and chert pebbles, silty to muddy matrix, subangular to subrounded, poorly sorted, medium- to very thin-bedded, ledge-forming; poorly exposed in the quadrangle; 200 feet thick.

Jm — MORRISON FORMATION (UPPER JURASSIC, Tithonian to Kimmeridgian) - Mudstone interbedded with channel conglomerate and sandstone lenses; mudstone is variegated, moderate-red to light-red on fresh surface but weathers yellow-green to light-gray, slope-forming; pebble conglomerate, poorly sorted, thin-bedded, well cemented, ledge-forming; sandstone, moderate reddish orange, coarse- to medium-grained, moderately sorted, friable, lenticular, ledge-forming; exposures of this formation are very poor in the quadrangle; 300 feet thick.

Js — STUMP FORMATION (UPPER JURASSIC, Oxfordian) - Sandstone; light-gray to grayish-orange on fresh surface but weathers greenish-gray to gray; very-fine to fine-grained, angular to subrounded, very well sorted to well sorted, cross-bedded, glauconitic, friable, and slope forming. 130 to 250 feet thick.

Je — ENTRADA SANDSTONE (MIDDLE JURASSIC, Callovian) - Shown only in cross section, not exposed within the quadrangle; upper reddish-brown siltstone and fine-grained sandstone and lower light-gray, pink, and light-brown sandstone; lower sandstone forms cliffs and ridges; 160 to 215 feet thick.

Jc — CARMEL FORMATION (MIDDLE JURASSIC, Callovian to Bajocian) – Upper siltstone and mudstone with gypsum (Winsor Member) and lower micritic and sandy limestone and gypsiferous siltstone (Paria River, Crystal Creek, and Co-op Creek Limestone Members); upper siltstone and mudstone, moderate-red-brown to pale green on fresh surface but weathers moderate-reddish-orange; siltstone, fine- to very fine-grained, subangular to subrounded, well sorted, medium-bedded to thickly laminated, massive to planar laminated, slope-forming; mudstone, soft, slope-forming; gypsum, light gray, thin beds to veinlets; lower limestone , micritic to sandy, moderate-gray, thick- to thin-bedded, ledge-forming, fossiliferous; contains beds of altered volcanic ash; local and regional 40Ar/39Ar dating of sanidine and biotite crystals and U-Pb dating of zircon crystals yield ages between 166 and 168 Ma (Sprinkel and others, 2011a); 150 to 220 feet thick.

JTRn — NUGGET SANDSTONE (LOWER JURASSIC TO UPPER TRIASSIC, Toarcian to Rhaetian) - Sandstone; light-gray to light-brown on a fresh surface but weathers dark- to light-red with iron oxide staining; medium- to fine-grained, well sorted, subrounded, well cemented, very thick-bedded to massive, high-angle cross-stratified, ridge and cliff forming; 800 feet thick.

TRb — FORMATION OF BELL SPRINGS (UPPER TRIASSIC, Rhaetian) - Sandstone and siltstone; grayish-orange to light-reddish-gray on fresh surface but weathers light-grayish-red to moderate-reddish-brown; sandstone, fine- to medium-grained, ripple laminated, contains rip-up clasts, sparse bioturbation, and cross-bedding; 90 to 130 feet thick.

TRcu — CHINLE FORMATION (UPPER TRIASSIC, Rhaetian to Carnian) - Shaley mudstone interbedded with siltstone, mudstone, moderate yellow to grayish-red on fresh surface but weathers gray to dark-grayish-red; fissile, poorly cemented and slope forming; siltstone, pale-purple to dark-reddish-brown, fine-grained, moderately to well sorted, and slope-forming; 290 feet thick.

TRm — MOENKOPI FORMATION (LOWER TRIASSIC, Olenekian to Induan) - Siltstone and mudstone interbedded with gypsum; siltstone and mudstone, reddish-brown to reddish-orange on fresh surface but weathers to dark-reddish-brown; siltstone, fine-grained, thick bedded, and slope forming; mudstone, ripple-laminate, thin-bedded, and slope forming; gypsum, light-red to grayish-orange-pink, massive, and ledge-forming; 820 to 1120 feet thick.

Pp — PARK CITY FORMATION (Middle to LOWER PERMIAN, Wordian to Artinskian) - Dolomite and sandstone; dolomite, tan to whitish-pink on fresh surface but weathers gray to moderate-yellowish-orange, glauconitic, quartz rich, and cliff forming; sandstone, reddish-yellow, coarse- to medium-grained, subrounded, moderately sorted, thick-bedded and locally cross-stratified; 150 feet thick.

PIPw — WEBER SANDSTONE (LOWER PERMIAN TO UPPER PENNSYLVANIAN, Sakmarian to Kasimovian) – Sandstone with interbedded with limestone in the lower part; sandstone, light-gray to yellowish-gray on fresh surface but weathers light-gray to very-pale-orange; medium- to coarse-grained, rounded to subangular, well sorted, clay particles and glauconite present; contains excessive burrowing, very thick-bedded to thickly laminated, ridge forming, and locally cross-bedded; limestone, blueish-purple, thinly-bedded, and cliff forming; 1000 feet thick.

IPm — MORGAN FORMATION (MIDDLE PENNSYLVANIAN, Moscovian) – Sandstone interbedded with limestone and silty shale; sandstone, light-red-gray, medium- to fine-grained, locally cross bedded, ledge forming; limestone (wackestone), light-gray, medium to fine-grained, fossiliferous, contains gray to red chert; interbedded silty shale, light-red to gray, very fine-grained, thinly bedded, slope forming; 500 to 950 feet thick.

IPrv — ROUND VALLEY LIMESTONE (Lower Pennsylvanian, Bashkirian) – Limestone interbedded with beds of shale and mudstone; limestone (wackestone to packstone), light-blue-gray on fresh surface but weathers dark-bluish-gray, contains abundant fossils (brachiopods, crinoids, etc.) and red, blue-gray, and gray chert nodules; mudstone, gray to green and purple to red; thinly bedded; 260 feet thick.

Md — DOUGHNUT SHALE (UPPER MISSISSIPPIAN, Serpukhovian to Visean) – Dark-gray shale with some reddish shale near base and interbeds of coarse sandstone, limestone, and organic-rich shale; mostly covered by Quaternary deposits, float, and vegetation within the quadrangle; shale is slope forming and clay rich; 80 to 300 feet thick.

Mh — HUMBUG FORMATION (Middle MISSISSIPPIAN, Visean) – Sandstone interbedded with limestone and shale; sandstone, dark-yellowish-orange to red on fresh surface but weathers black to orange; sandstone, very fine-grained to fine-grained , locally cross-bedded, hematitic, contains caves and sinkholes; limestone (mudstone to packstone), light-gray on fresh surface but weathers dark-gray; shale, slope forming and generally covered with vegetation making an accurate description difficult. 150 to 250 feet thick.

Mm — MADISON LIMESTONE (Middle to Lower Mississippian, Visean to Tournaisian) – Limestone (wackestone to packstone), dark-gray to light-gray on fresh surface but weathers dark-gray; coarse to medium crystalline, massive, and contains an abundance of chert; chert nodules are dark to light-gray, pebble to cobble sized; angular to subangular brecciated limestone clasts enclosed in a medium crystalline matrix of dark-gray limestone are present; 1100 feet thick.

Zur — RED PINE SHALE (middle Neoproterozoic, Cryogenian) - Silty shale interbedded with silty sandstone, dark green gray on fresh surface but weathers brown gray to reddish brown, purple to pale pink, and pale olive; shale units are fine to very fine grained, very thin to medium bedded, ripples and loading structures are present; slope forming and poorly cemented; sandstone units are fine to coarse grained, poor to well sorted, very thick to medium bedded, friable to well cemented, quartz rich to arkosic, cross-bedded and ledge forming; 1800 to 4000 feet thick.

Zu — UINTA MOUNTAIN GROUP (middle Neoproterozoic, Cryogenian) - Shown only in cross section. 14,000+ feet thick.

CORRELATION OF SURFICIAL UNITS

CORRELATION OF BEDROCK UNITS

MAP SYMBOLS

CONTACT – Dashed where inferred or approximately located.

FAULT – Dashed where inferred or approximately located, dotted where covered; bar and ball on downthrown side; arrows show stike-slip movement.

LANDSLIDE SCARP – Teeth on down slope side of scarp.

STRIKE AND DIP OF BEDS

DRILL HOLE LOCATION – Formation tops available in Sprinkel (2006).

STRATIGRAPHIC COLUMN

AGE		FORMATION	SYMBOL	THICKNESS Feet	LITHOLOGY
Quat.		unconsolidated deposits	Q*		
Tertiary	Oligocene	Bishop Conglomerate	Tb	0-500	
	Eocene	Brennan Basin Member of the Duchesne River Formation	Tdb	280-580	
		Uinta Formation**	Tu	600	
		Green River Formation**	Tg	3000+	
	Paleocene	Wasatch Formation**	Tw	1200	
Cretaceous		Mesaverde Group**	Kmv	1-200+	
		Mancos Shale**	Km	4000	
		Frontier Formation**	Kf	140-300	
		Mowry Shale**	Kmo	90-120	
		Dakota Sandstone**	Kd	115-260	
		Cedar Mountain Formation	Kc	200	
Jurassic		Morrison Formation	Jm	300	
		Stump Formation	Js	130-250	
		Entrada Sandstone	Je	160-215	
		Carmel Formation	Jc	150-220	
Triassic		Nugget Sandstone	JTRn	800	
		Formation of Bell Springs	TRb	90-130	
		Chinle Formation	TRcu	290	
		Moenkopi Formation	TRm	820-1120	
Permian		Park City Formation	Pp	150	
		Weber Sandstone	PIPw	1000	
Carboniferous	Pennsylvanian	Morgan Formation	IPm	500-950	
		Round Valley Limestone	IPrv	260	
	Mississippian	Doughnut Shale	Md	80-300	
		Humbug Formation	Mh	150-250	
		Madison Limestone	Mm	1100	
Neoproterozoic		Red Pine Shale	Zur	1800-4000	
		Uinta Mountain Group**	Zu	14000+	

*See Correlation of Quaternary Units for symbols
**not exposed in quadrangle - on cross section only

not to scale

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