Mapping and Kinematic Structural Analysis of the Deep Creek Fault Zone, South Flank of the Uinta Mountains, Near Vernal, Utah

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MAPPING AND KINEMATIC STRUCTURAL ANALYSIS OF THE DEEP
CREEK FAULT ZONE, SOUTH FLANK OF THE UINTA
MOUNTAINS, NEAR VERNAL, UTAH

by

David A. Haddox

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

Master of Science

Department of Geology
Brigham Young University
August 2005
This thesis has been read by each member of the following graduate committee and by majority vote has been found to be satisfactory.

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ABSTRACT

MAPPING AND KINEMATIC STRUCTURAL ANALYSIS OF THE DEEP CREEK FAULT ZONE, SOUTH FLANK OF THE UINTA MOUNTAINS, NEAR VERNAL, UTAH

David A. Haddox
Department of Geology
Master of Science

The geology along the southern flank of the Uinta Mountains, located north of Vernal, Utah, has been mapped at the 7.5’ scale within two quadrangles: the Dry Fork and Steinaker Reservoir Quadrangles. Ambiguities dealing with stratigraphy, structural geology, and geohazards are currently being addressed as a result of this and other mapping projects in the vicinity.

The geologic units in the area range in age from Mississippian to Late Cretaceous and include Uinta-sourced Tertiary units. Brief unit descriptions are provided for each of the units exposed in the map area.

The main structural influence on the rocks within the area is that of the Uinta Uplift and its southern bounding fault, the Uinta Basin Boundary thrust. Locally, the Deep Creek fault zone overprints and dissects the southernmost flank of the broad
Uinta Anticline. Other smaller structurally complex areas and folds exist east of the Deep Creek fault zone.

The Deep Creek fault zone is made up of a series of NW-SE trending faults, likely related to the South Flank fault zone. Many authors have inferred dip-slip movement along the South Flank fault zone, but have not supported these claims using kinematic data. Detailed mapping and kinematic data collected within the study area has produced a better understanding of the deformation history along the fault zones in question.

The faults within the Deep Creek fault zone have steep, linear traces upon which both vertical dip-slip and very nearly strike-slip (left-lateral oblique-slip, mainly) movement has occurred. The faults of the Deep Creek fault zone are likely Paleocene in age. The data suggest a bimodal history of deformation which the principal stress field does not seem to be influenced by typical east-northeast-west-southwest Laramide orogenic far-field stresses. The creation and early history of these faults may have been due to localized stress fields related to activity of the underlying Uinta Basin Boundary thrust, or a later period of uplift, a possible accommodation zone between the western and eastern domes of the Uinta Mountain Range, a transfer zone between the Uinta Basin Boundary thrust and the Asphalt Ridge fault, or a combination of these.
ACKNOWLEDGEMENTS

Deepest thanks to those who supported me throughout this endeavor.

Special thanks to my wife, Karen, for her sacrifice, support, and patience as I gave this project the attention it needed in order for it to be a success.

Thanks go to my committee: Doctors McBride, Harris, and especially Doctor Kowallis for his guidance and patient correspondence with me when I was out of town, especially when I placed him in any difficult situations. A great deal of appreciation and respect goes to all the professors who instructed me throughout the time I spent here.

Great thanks to the students that most aided me: Jeremy Shakespeare and Paul Jensen. I thank Jeremy for the knowledge he attributed to this project while in the field; I thank Paul for his support during the later stages of my project.

Thanks also go to the BYU Field Camp students of the years 2003 and 2004 and their individual contributions to the mapping process.

Special thanks go to Doug Sprinkel of the Utah Geological Survey for his expertise and advice concerning the geology of the area.

The funding provided for the mapping portion of the thesis project and subsequent expenses was provided by an EDMAP grant. Additional funding was provided by Brigham Young University.

Thanks also go to Parks Rangers Kurt and Mike from Steinaker State Park.
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Introduction

The area along the southern flank of the Uinta Mountains north of Vernal, Utah contains one of the best exposures of Mesozoic-aged strata along the northern rim of the Uinta Basin. Many unique and beautiful rock formations exist here and the area has long been enjoyed as a center of recreation for Basin residents and even as a refuge of solitary living in the canyons and lowlands at the base of the Uinta Mountains.

The study area rests at the base of the Uinta Mountains and along their southernmost flank (Figure 1). It is located above the Uinta Basin Boundary thrust fault (whose southern tip is thought to exist just a few miles south of the area) and within the eastern edge of the Deep Creek fault zone which dissects and displaces the rocks often concealed under extensive landslide deposits sourced from the thick, boulder-rich Bishop Conglomerate.

Two geologic maps including accompanying unit descriptions, cross sections, and a stratigraphic column have been created for the Dry Fork and Steinaker Reservoir 7.5’ (1:24,000 scale) quadrangles in order to better understand the nature of the Deep Creek fault zone and additional structures in the area (see Plates 1 & 2). Detailed mapping has uncovered a measure of geologic issues that did not correlate well with the current beliefs about the area. These issues rage in matters from structural, stratigraphic, economic, and geohazards.

The South Flank and Deep Creek fault zones consist of an array of apparent normal faults, but currently no published kinematic data exists for these fault zones. The structural issues discussed in this document address and interpret the kinematic
data collected within the Deep Creek fault zone (Untermann and Untermann, 1969) and in a few locations along the South Flank fault. Descriptions of folds and small structurally complex areas are also addressed.

Stratigraphic questions include as to whether or not the Buckhorn Conglomerate exists within the map area and marks the contact between the Upper Jurassic Morrison Formation and the Early Cretaceous Cedar Mountain Formation (Currie, 1997). Another stratigraphic question is the formation of the Lower Glen Canyon Sandstone, its age, depositional environment, and correlation to time-equivalent geologic units in surrounding regions (Jensen and Kowallis, 2005). The remaining stratigraphic issues deal with the following geologic units: the Meade Peak Member of the Park City Formation (Schell and Dyni, 1973), the Upper Glen Canyon Sandstone oasis deposit (Anderson et al., 2000), and the Frontier Formation “cannonballs” (Doug Sprinkel, personal communication). These will be described and addressed in the body of this document.

The economic issues in the area deal with current questions related to, among describing less significant issues, additional strip-mining the Meade Peak Member of the Park City Formation west of the current phosphate mine.

Finally, the geohazards related to faulting of the Deep Creek fault zone and the threat of earthquakes (of which there is virtually none) are briefly discussed (see University of Utah Seismic Station website for historical earthquakes: http://www.quake.utah.edu/). More importantly, the age, extensiveness, and frequency of landslide hazards is also described (Kowallis and Bradfield, 2005).
Chapter 1: Mapping the Dry Fork and Steinaker Reservoir 7.5’ Quadrangles

Introduction

The Dry Fork and Steinaker Reservoir 7.5’ Quadrangles are adjacent east-west quadrangles which lay in the Uinta Mountain physiographic province just north of the Colorado Plateau (see Figure 1). Topographically, the quadrangles straddle the transition from the Uinta Basin to that of the southern flank of the Uinta Mountains. The southwest corner of Dry Fork Quadrangle intersects 40° 30’ N latitude 109° 45’ W longitude. The shared northern corner of both quadrangles is at the 40° 37’30” N latitude 109° 37’30” W longitude intersection and the southeastern corner of Steinaker Reservoir Quadrangle is at 40° 30’ N latitude 109° 30’ W longitude.

The Dry Fork Quadrangle includes prominent local geographic features such as Little Mountain, which dominates the southern portion of the quadrangle and, north of there, Dry Fork Canyon. Both quadrangles share Ashley Gorge, while the Steinaker Reservoir Quadrangle features a small portion of Big Brush Gorge at its northeast corner. The Steinaker Reservoir Quadrangle also includes Red Mountain, the dominant feature, and Steinaker State Park, which is concentrated around Steinaker Reservoir (see Plates 1 & 2).

From west to east across the quadrangles, the landslide-covered, horst-and-graben characteristics of the Deep Creek fault zone in Dry Fork Quadrangle transitions into broad, barren folds in the Steinaker Reservoir Quadrangle.
Residences in the two quadrangles are rural, but exist only 15 minutes or so from Vernal, the largest city in the Uinta Basin. The higher elevations around Little Mountain and the northern slopes of both quadrangles are used mostly for range cattle. Recreational opportunities include hiking slot canyons, ATV/ORV use in the hills and dune deposits, mountain biking, enjoying Steinaker State Park, and admiring the petroglyphs found in the area. Dry Fork Canyon is the western entrance to the Red Cloud Loop Scenic Byway.

**Stratigraphy**

Many of the first geologists to map the Mesozoic-aged rocks in the Vernal area had experience in the Colorado Plateau. For that reason many of the geologic units have similar names and—rightly so—most of the units have equivalent counterparts in the Colorado Plateau (Kinney, 1955; MacLachlan, 1957; Anderson et al, 2000). In a few cases the Mesozoic units in Vernal may have been misnamed or the names may be misleading (see Plates 1 & 2).

The geologic units exposed in the map area range in age from the Mississippian to the Oligocene with several major unconformities throughout (Figure 2; Kinney, 1955, Gregson and Chure, 2000). Other than several Quaternary deposits, there may also exist an unnamed unit capping the Bishop Conglomerate most likely of Pleistocene age (Doug Sprinkel, personal communication). The units in the map area consist of the Mississippian-aged Madison Limestone up to the Cretaceous-aged Mancos Shale; Uinta Basin-fill units represented are the Duchesne River Formation and Bishop Conglomerate (Kinney, 1955).
The total average thickness of these units is 9396 feet. These strata are exposed within the Uinta Uplift above the Uinta Basin Boundary thrust fault. The major cause of deformation is uplift along the thrust. The average dip of the units within the Steinaker Reservoir and Dry Fork Quadrangles is approximately 20° to the southeast—although there are numerous small tilted fault blocks and folded beds, this average denotes the major structural influence of the Uinta Basin Boundary thrust. The average thickness for each geologic period represented in the area is as follows: Mississippian = 467 feet, Pennsylvanian = 1785 feet, Permian = 138 feet, Triassic = 1315 feet, Jurassic = 2178 feet, Cretaceous = 3003 feet, and Tertiary = 510 feet (Figure 2). Of the Quaternary deposits, there are landslide deposits estimated to be as thick as 50-100 feet, several well-developed alluvial pediment deposits ranging in thickness between 3 feet and 50 feet, and the unnamed Pleistocene-aged unit deposited above the Bishop Conglomerate, which is approximately 40-50 feet thick (although this estimate could be disputed because the shared contact is difficult to locate—there really is no easy way to define it because the Bishop Conglomerate and the unnamed Pleistocene unit have a similar composition, although the Pleistocene unit may have much larger clasts) (see Plates 1 & 2).

The following section will address the geologic units found within the map area in detail. Each unit will be described briefly including where it can be found within each quadrangle, its composition, and its thickness. Sprinkel (2000) and Gregson and Chure (2000) provide a good source summarizing the same units in nearby areas. For those instances where controversy surrounds nomenclature or
where the given name of a unit near Vernal just doesn’t fit its namesake in the Colorado Plateau, additional information will be given.

**Mississippian Rocks**

The Mississippian System consists of the Madison Limestone (Lower Mississippian), the Humbug Formation (Upper Mississippian), and the Doughnut Shale (Upper Mississippian) (Sandberg et al, 1982; Welsh and Bissell, 1979). These were largely deposited in shelf, marginal-marine, and near-shore environments.

**Madison Limestone (Mm)**

The Madison Limestone (Kinney, 1955; Gregson and Chure, 2000) is only exposed in Dry Fork on the north side of the canyon and only in small ledges (Figure 3). It usually forms cliffs in the region. Only 250 feet is exposed at a maximum in the Dry Fork Quadrangle, but it is fully exposed in nearby canyons and can be as thick as 1220 feet.

Because the Madison Limestone is poorly exposed in the map area, it is only possible to characterize the uppermost 100 feet or so. It is made-up of a light grey to medium-grey colored, massively bedded, wackestone and packstone carbonate that contains abundant dark grey chert nodules and is interbedded at the top by tan-colored, dense, fine-grained sandstone beds. Few invertebrate fossils were found in the section.
Humbug Formation (Mh)

The Humbug Formation (Kinney, 1955; Gregson and Chure, 2000) is poorly exposed in the map area (Figure 3). It, less than the Madison Limestone, only occasionally crops out through colluvium in Dry Fork.

The Humbug Formation consists of alternating beds of mudstone, shale, sandstone, and carbonate. The carbonate beds are dark or light grey in color and range in texture from mudstone to packstone; they are fossiliferous and contain chert.

The Humbug Formation has an apparent thickness of 125-160 feet.

Doughnut Shale (Md)

The Doughnut Shale (Kinney, 1955; Gregson and Chure, 2000) is also poorly exposed in the map area (Figure 3). It is exposed in Dry Fork, but may exist at the very bottom of Ashley Gorge under colluvium.

Because it is almost wholly covered by colluvium, it cannot be characterized. In the region, it is a dark grey, clayey, marine shale.

The Doughnut Shale has an apparent thickness of 50-80 feet.

Pennsylvanian Rocks

The Pennsylvanian System consists of the Round Valley Limestone (Early and Middle Pennsylvanian), the Morgan Formation (Middle Pennsylvanian), and the Weber Sandstone (Middle Pennsylvanian to Permian) (Mallory, 1979; Welsh and Bissell, 1979). These rocks were deposited in beach and near-shore environments.
The Round Valley Limestone (Kinney, 1955; Gregson and Chure, 2000) is only exposed in Dry Fork and Ashley Gorge in the Dry Fork Quadrangle (Figure 4). The best exposure is at a road-cut just over three miles from the mouth of Dry Fork Canyon where 120 feet is exposed. Otherwise, it usually outcrops in the form of one single ledge, where, at best, there is a maximum of 25 feet exposed in a wash—the rest is covered by colluvium. The best exposures are found deep in Ashley Gorge, where a significant section can be seen.

The Round Valley Limestone is made up of light to dark, bluish-gray to lavender-colored carbonate beds; it is a wackestone to packstone in classification and contains numerous fossils of the age (mainly crinoids, brachiopods, etc). It is occasionally interbedded with thin, dark shaly beds that are commonly purple, red, green, and gray. The Round Valley Limestone is usually colored red on steeper cliffs because the overlying red-colored Morgan Formation washes over it—much like the Redwall Limestone of the Grand Canyon.

Where the Round Valley Limestone is mostly covered (under colluvium draping the canyon walls), there aren’t any large-scale observations that could be made, except to say that one particular bed, as mentioned, is more durable and can usually be seen through the colluvium. At a maximum, the best exposure (in Ashley Gorge) is approximately 200 feet thick; this exposure is estimated to be near the covered contact with the underlying Doughnut Shale and may represent the total thickness of the unit. In Dry Fork, it has an apparent thickness of 260 feet.
Morgan Formation (IPm)

The Morgan Formation (Dreise, 1983; Kinney, 1955; Gregson and Chure, 2000) is exposed in two areas: in Dry Fork Canyon in the Dry Fork Quadrangle and in Ashley Gorge in both quadrangles. It is best exposed in Ashley Gorge in the northeast corner of the Dry Fork Quadrangle, but an excellent exposure occurs in a side canyon off of Dry Fork Canyon just south of Charley’s Park (Figure 5). Where it is exposed in Dry Fork Canyon it forms ledgy cliffs; in Ashley Gorge it is often covered with colluvium, but outcrops in its entirety as you move further up the gorge.

The Morgan Formation contrasts sharply with the other geologic units surrounding it. It is made up of medium-gained, dark red to red-orange, massive sandstone beds occasionally interbedded with lavender-colored carbonate beds and thin shale beds. The carbonate beds are dense wackestones to packstones—in fact, they are difficult to even chip with a hand-sized rock hammer. The shale beds are much thinner, cause the breaks in the sandstone and carbonate bedding, and range in color from purple to gray. The sandstone and carbonate beds are about the same size and produce ledges on average about 3 feet to 4 feet thick.

The complete thickness of the Morgan Formation is approximately 360-400 feet.
**Weber Sandstone (PlPw)**

The Weber Sandstone (Whitaker, 1975; Bissell and Childs, 1958; Kinney, 1955; Gregson and Chure, 2000) is one of the more spectacular units in the map area. The high, steep cliff walls in the majority of canyons or gorges in the region are made up of the Weber Sandstone (Figure 6, see also Figure 5). (Nearby, the Split Mountain Anticline north of Dinosaur National Monument features this unit.) It is also exposed at the bottom of the smaller canyons cutting the foothills in the northern part of each quadrangle.

The Weber Sandstone is made up of mostly multi-colored, medium-sized, eolian or backshore sands; it contains a fair amount of clay and glauconite as interstitial grains. Near the bottom of the unit, the Weber Sandstone is interbedded with the same lavender-colored carbonates found in the Morgan Formation (though they are rattier and thinner). Large-scale (6 feet to 10 feet in size) fluid-escape structures and soft-sediment deformation structures are present throughout the unit. Occasionally, one can find bivalve- and gastropod-mixed sandy fossil beds.

The Weber Sandstone is completely exposed in both quadrangles, but rarely in a complete vertical section. Where the top is exposed, the base is covered under colluvium or is buried; where the base is exposed, the top has usually been truncated by erosion and covered by deposition of the Bishop Conglomerate. The total thickness of the Weber sandstone is as much as 1275 feet.
Permian Rocks

The Permian System is represented by the uppermost third of the Weber Sandstone (Middle Pennsylvanian to Permian) and the Park City Formation (Early Permian) (Whitaker, 1975). The upper Weber Sandstone consists of rocks deposited in back-shore eolian and beach depositional facies. The Park City Formation consists mainly of marine rocks deposited on a broad shelf, but also contains terrestrial-derived rocks.

Park City Formation (Pp)

The Park City Formation (Jado, 1980; Kinney, 1955; Gregson and Chure, 2000) forms the main southward dipping slope off of the Uinta Mountains in both quadrangles (Figure 6, caps the dip-slope). It is also occasionally exposed in fault blocks dissected by the Deep Creek fault zone in the Dry Fork Quadrangle.

The Park City Formation is made up of four distinct members: two dolostone members—the Lower and Upper Franson Members—and two mudstone/shale members—the MackEntire and Meade Peak Members, respectively. The bottommost member, the Grandeur Member, is locally absent or present as a sandstone bed only as thick as one foot. The dolostone members are either dark brown (Lower Franson) or a white (Upper Franson) in color; they are sandy, glauconitic, and they contain chert nodules; both members are occasionally interbedded with thin shaly beds that break up the individual dolostone beds. The Lower Franson is much sandier and contains more chert than the Upper Franson.
The shale/mudstone members vary significantly. The Meade Peak Member (

*Ppm*) (Smith et al, 1952; Schell and Dyni, 1973) is known elsewhere as the Phosphoria Formation and is made up of dark, olive- to black–colored shallow marine shale bedded with a few thin sandstone beds near its base. It also contains red and green chert nodules characteristic only to the Meade Peak Member of all the geologic units in the area. Because of its economic importance, the Meade Peak Member has been mapped separately from the other Members of the Park City Formation within both quadrangles. The MackEntire Member, on the other hand, is made up of reddish-orange terrestrial mudstones that rarely outcrop in either quadrangle. These mudstones have been diagenetically altered significantly in the western part of the Dry Fork Quadrangle—extensive cementation likely do to hydrothermal fluid flow within the Deep Creek fault zone.

Where exposed at the surface, which is a significant portion of the area of both quadrangles, the Park City Formation weathers to a thin blanket of colluvium, with little soil, but only occasionally crops out at the surface. Because of the thinness of this cover, it has been mapped as exposed bedrock. It is 90-145 feet thick. Each member is of nearly equal thickness: the Meade Peak Member is 20 feet thick, the Lower Franson Member is 30 feet thick, the MackEntire Member is 30 feet thick, and the Upper Franson Member is 35 feet thick. The contact between each member is relatively easy to pick out as they each differ noticeably from one another in lithology and color.


Triassic Rocks

The Triassic System consists of the Dinwoody Formation (Early Triassic), the Moenkopi Formation (Early Triassic), the Chinle Formation (Late Triassic), and the Glen Canyon Sandstone (Late Triassic to Early Jurassic) (MacLachlan, 1972; Dubiel, 1994). The Triassic-Jurassic boundary is somewhere in the bottommost part of the Upper Glen Canyon Sandstone. These rocks are largely made-up of continental deposits in origin, except the lower units (Dinwoody and Moenkopi Formations) represent a transition from a marine environment to terrestrial that of a terrestrial environment.

Dinwoody Formation (Rd)

The Dinwoody Formation has only recently been mapped as a separate unit in this area (Sprinkel, 2002). It is exposed in both quadrangles at the base of the Park City Formation dip-slope, but begins to pinch out from east to west across the quadrangles. It also outcrops under isolated gravel mantles on the dip-slope in the northern section of each quadrangle or under the Bishop Conglomerate higher on the flanks of the Uintas. The white-colored shale and gypsum beds of the Dinwoody Formation interfinger with the overlying red-beds characteristic of the Moenkopi Formation east of Red Mountain in the Steinaker Reservoir Quadrangle (Figure 7). This red- and white-striped interfingering diminishes to the west as the white-colored beds pinch-out below a dark brown-colored, dense, micaceous sandstone bed of which the top represents the contact between the two formations. This bed seems to become enveloped entirely by Moenkopi Formation red-beds. Though the Dinwoody
Formation pinches-out to the west, it does exist in the Dry Fork Quadrangle under Pine Ridge; inversely, in the Donkey Flat Quadrangle, it thickens with little or no interfingering to the east.

The Dinwoody formation is made up of thin, tan- to brown-colored coarse-grained sandstone beds, which at times include a significant amount of mica. The sandstones are interlayered with massive gypsum beds and very thin, medium- to fine-grained gypsiferous sandstone and shale beds. The base of the Dinwoody Formation contains pink and white chert nodules and one bed of thick, massive, white-colored gypsum. Because of the abundance of gypsum and thinness of its sandstone beds, the Dinwoody is readily eroded and, when exposed, weathers at the surface in the form of punky colluvium.

In the Steinaker Reservoir Quadrangle, the Dinwoody Formation was not mapped along gross lithologic criteria, but was mapped along what is likely a local chronostratigraphic surface. This particular dense, micaceous, sandstone bed can be readily traced across the base of Red Mountain to the western border of the quadrangle. In the Dry Fork Quadrangle, the upper contact (the sandstone marker bed) is usually covered, so other lithostratigraphic surfaces were used to map the contact instead, such as the basal gypsum beds that are easily recognized. In light of this, the thickness of the Dinwoody Formation is likely underestimated in places.

The Dinwoody Formation is approximately 120 feet thick in the Steinaker Reservoir Quadrangle, but is only 40 feet thick in the Dry Fork Quadrangle. In the Donkey Flat Quadrangle, where its thickness reaches a maximum, it has an estimated thickness of 170 feet.
Moenkopi Formation (Rm)

The Moenkopi Formation (McKee, 1954; Pipiringos and O’Sullivan, 1978; Kinney, 1955; Gregson and Chure, 2000) is one of the more scenic formations found within the region because of its thickness and contrasting color compared to the lighter, tan-colored sandstones around it like the Weber Sandstone and Glen Canyon Sandstone (Figure 7). Lithologically, it is much like the Moenkopi Formation found on the Colorado Plateau. It is made up of thinly-bedded, orange- and red-colored fine-grained sandstones, siltstones, and mudstones that are often ripple-laminated and contain sparse mica. In places it is interbedded with thin, easily-eroded gypsiferous beds. Near its base, a few thin, chippy, green- or white-colored ripple-laminated beds can occasionally be found. Approximately in the middle of the unit, there is a large, massive, gypsum bed that may correlate (author’s speculation) with the Sinbad Limestone Member of the Moenkopi Formation found in the Colorado Plateau as near as the San Rafael Swell (Stewart et al, 1972a; Blakey, 1974). Amphibian bones and trackways have been found in the Moenkopi Formation in the map area (Hamblin and Foster, 2000).

The Moenkopi Formation is not split into members in the map area, but could possibly be split into three: a ‘Lower Red’, an ‘Upper Red’, and a capping ‘Chocolate Brown’. The ‘Lower’ and ‘Upper Red members’ are separated by the thick gypsum bed previously mentioned otherwise they are similar ad fit the general description given above. The ‘Chocolate Brown member’ contains thick, massively-bedded,
fine- to medium-grained, dark brown sandstone beds; the beds in this member are, on average, much thicker than those found within the rest of the Moenkopi Formation.

The Moenkopi Formation outcrops in its entirety in both quadrangles and is 820-1120 feet in thickness. It is exposed along the northern portion of both quadrangles in stratigraphic position above the Dinwoody Formation at the base of the Park City Formation dip-slope. It makes up many of the higher elevation ridges and spurs in the map area (e.g. the ominous Red Mountain in the Steinaker Reservoir Quadrangle is named after the red color of the Moenkopi Formation). In the Dry Fork Quadrangle and adjacent Lake Mountain Quadrangle, the Moenkopi Formation outcrops are repeated by Deep Creek fault zone fault blocks in several places.

**Chinle Formation (TRc)**

The Chinle Formation (Figure 8) (Poole and Stewart, 1964; Stewart et al, 1972b; Jensen and Kowallis, 2005; Kinney, 1955; Gregson and Chure, 2000; Pipiringos and O’Sullivan, 1978) is exposed in both quadrangles in stratigraphic position above the Moenkopi Formation. Because of the contrasting durability between the differing members of the Chinle Formation, their spatial distribution is disproportionate to their thicknesses—the thin, resistant Gartra Member produces a wide outcrop band, while the bulk of the formation outcrops as only a thin strip across both quadrangles.

The Chinle Formation is made up of four distinct members: the Gartra Member (TRcg), the Mottled member, the Ocher member, and the Upper Red member (‘Siltstone member’ in some accounts) (High et al, 1969; Poole and Stewart, 1964).
The most spatially prevalent, as indicated above, is the Gartra Member. Lithologically, it is a light-colored pebble conglomerate that has been interpreted to be deposited in a braided fluvial environment. It occasionally contains two conglomerate beds; the upper bed is thinner and interbedded with the overlying Mottled member. The Gartra Member is mostly one, thick, pinching and swelling bed made up of a conglomeration of accreted fluvial channels (McCormick and Picard, 1969; Salter, 1966). Locally, these fluvial channels cut into the Moenkopi Formation. This contact has been shown to be a regional unconformity on a larger scale (High et al, 1969). The Gartra Member contains abundant petrified logs and wood fragments and its thickness ranges from 20 feet to 80 feet.

The Mottled member of the Chinle Formation is easily recognized because of its distinctive deep-purple color. When exposed at the surface, it weathers into badlands-type topography. A fresh surface cannot be found cropping out in either quadrangle, but road-cuts and self-dug trenches show it is made up of highly-deformed, variegated paleosols with a significant component of altered volcanic ash. The Mottled member occasionally contains lenses of isolated fluvial sandstone beds and freshwater carbonates. Reptile bones, tracks, and fossilized fern-like plant impressions (Hamblin et al, 2000) have been found by the author in this lower member (Figure 9). This unit is approximately 30-50 feet thick.

The next unit, the Ocher member, has been interpreted to be a saline lacustrine deposit, in origin, and is easily recognized by its (again) very distinctive mustard-coloring (High et al, 1969). It consists of shale and very thin sandstone beds that occasionally contain calcite-aragonite fragments and thin, isolated, very coarse-
grained to pebbly conglomerate lenses. This member also contains a bed of large, cobble-sized calcite-aragonite nodules. This unit has been shown to correlate with part of the Popo Agie Formation in Southeastern Wyoming (High et al, 1969; Picard, 1977). It is about 40-50 feet thick and, like the Mottled member, weathers into badlands topography, in which a fresh surface is never exposed in either quadrangle.

The Upper Red member locally has not been described in the literature in any detail (High et al, 1969; Picard, 1977). It has been termed the Upper Carbonate unit in other areas nearby, but this name does not describe what is found in the map area. For the most part, it is a homogenous, earthy-colored (reddish), easily deformed, chippy siltstone unit that contains an isolated bed of very-coarse grained, pebbly sandstone near the base. This unit is approximately 40-50 feet thick, but, unlike the other mudstone members of the Chinle Formation, the Upper Red member will often outcrop in a relatively fresh surface under the protective ledges of the overlying Lower Glen Canyon Sandstone. In at least one location, the jumbled, deformed stratification of the siltstones is truncated unconformably by the Lower Glen Canyon Sandstone (Figure 10). It is unclear as to whether or not this represents a regional angular unconformity or simply truncation of locally deformed beds (soft sediment deformation). The nature of other exposures of this contact does not clear this matter up.

Member contacts, although often covered, are rather easy to pick out because of the color contrasts between them. However, there are also recognizable beds that separate the contacts. Between the Mottled and Ocher members, there is a very coarse-grained to pebbly sandstone that contains rip-up clasts of the Mottled member.
Between the Red Siltstone Member and the Ocher Members is a bed of cobble-sized carbonate concretions (see Figure 2).

The total thickness of the Chinle Formation is 170-230 feet. It should be noted that the Chinle Formation slumps readily below jointed Lower Glen Canyon Sandstone, blocks, especially on Red Mountain.

Glen Canyon Sandstone

The Glen Canyon Sandstone (Poole and Stewart, 1964; High and Picard, 1975; Imlay, 1980; Picard, 1977; Jensen and Kowallis, 2005; Kinney, 1955; Gregson and Chure, 2000) has been named thus because the package of rocks of which it is derived was thought to be equivalent to the Glen Canyon Group (Wingate Sandstone, Kayenta Formation, and Navajo Sandstone) of the Colorado Plateau (Anderson et al, 2000). It is the unit that is the most extensively exposed in the map area, spatially. It is also one of the thickest units (a total of 810-1140 feet), forming steep-sided canyons and slot canyons and occasionally an arch. The Glen Canyon Sandstone is jointed in conjugate sets across the Dry Fork and Steinaker Reservoir Quadrangles, and, even though the orientations change slightly across the map area, the conjugate orientations remain mainly north-northwest and north-northeast directions. The formation is faulted in many areas in Dry Fork Quadrangle (and thus very difficult to approximate what stratigraphic level one might be in when on the unit), but, in contrast, is only broadly folded in the majority of the Steinaker Reservoir Quadrangle.

The Glen Canyon Sandstone can be divided into two members: the Upper member (J.TRgu) and the Lower member (J.TRgl); the upper cross-stratified facies and
lower thinly-bedded facies of Picard (1977), respectively. Currently, there is a question as to whether or not these members should be named according to Utah or Wyoming nomenclature. In Wyoming, the Nugget Sandstone is divided into a lower member called the Bell Springs Member equivalent to the Lower Glen Canyon Sandstone and an upper unnamed member equivalent to the Upper Glen Canyon Sandstone. (For more information regarding this topic see Paul Jensen’s Masters Thesis expected Summer, 2005, Brigham Young University.)

The Lower Glen Canyon Sandstone is largely made up of medium-grained quartz sand (Figure 8). It outcrops in dark red, green, and mottled white- and orange-colored beds and is interbedded with shale beds of light green and purple coloring. The base of the Lower Glen Canyon Sandstone is planar ripple-laminated. It is at the base that the red sandstones are present. The bulk of the unit has less bedding throughout, but is made up of ripple-laminated sandstones similar to the sandstone found at the base, except with orange and mottled color. Small, lensoidal channels cut into the top of this more massive, thick, mottled-orange bed. Shales of the Lower Glen Canyon Sandstone exist in the uppermost portion and are often separated by thin to medium-sized beds of orange- and white-colored sandstone. These sandstone beds are either: (1) thin, 1-2 feet thick, ripple-laminated, green sandstone to marlstone beds, or (2) thicker, orange-colored, bioturbated beds, which occasionally have a laterally thicker isolated channel, which cuts 5-10 feet into the underlying shale beds.

In the Steinaker Reservoir Quadrangle, the Lower Glen Canyon Sandstone is exposed along the base of Red Mountain and continues westward in correct stratigraphic position above the Chinle Formation across Ashley Creek into the Dry
Fork Quadrangle. There, it is exposed very near the base of the alluvium-filled valleys under the Upper Glen Canyon Sandstone cliffs. It is also exposed under the Upper member at the confluence of Dry Fork and Ashley Creek and also along the north end of Little Mountain buried under colluvium, but crops out near Castle Cove. The Lower Glen Canyon Sandstone changes in thickness across the quadrangles. In the Steinaker Reservoir Quadrangle, it is 110 feet thick; in the Dry Fork Quadrangle, near Castle Cove, it is 80-90 feet thick.

The Upper Glen Canyon Sandstone (equivalent to the famous Navajo Sandstone in Southern and Central Utah and the Nugget Sandstone along the Wasatch Range) is exposed over a large area in both quadrangles—usually taking up much of the southern half of each quadrangle (Figure 6) (Anderson et al, 2000). It is 720-1030 feet thick, but this is only an estimate since a complete vertical section is only exposed at the base of Red Mountain, where the horizontal distance covered while measuring across this unit may have introduced some error into thickness.

The base of the Upper Glen Canyon Sandstone is in contact with a six-foot thick, purple-colored shale of the Lower Glen Canyon Sandstone. Sedimentary features at this contact—either hexagonal mud-cracks or some type of ordered burrowing features—are filled with the well-sorted sands of the Upper Glen Canyon Sandstone. Features indicative of a long-standing exposure surface (hard, dense lumps of mud separated by clay) are also present at this contact. Dinosaur tracks in the basal sandstone at this contact near Dinosaur National Monument have been determined to be of Triassic age (MacLachlan, 1957; Hamblin et al, 2000). Though this contact is very abrupt, the approximately 30 feet of sandstone above this contact
contains a mix of sedimentological structures that seem to record the climatic and depositional changes expected when transitioning from a wetter lacustrine to an arid erg environment. Starting at the base, immediately above the Lower-Upper Glen Canyon Sandstone contact, there is 3 feet of high angle trough cross stratification, which is truncated by planar, ripple-laminated stratification which then transitions into reddish mudstone breaks (paleosol or tidal flat mud) in the sandstone. This then transitions back into planar, ripple-laminated sandstone and then high angle trough cross stratification indicative of eolian dunes. Three to four of these cycles exist above this contact. These are also semi-regional—they can be seen across the map area. After this 30 to 50 foot transitional interval, the unmistakable, “Navajo-famous”, large amplitude trough cross stratification begins in the Upper Glen Canyon Sandstone. The rock type within this interval is of the same color and size as that of its southern and western counterparts (buff-colored, well-sorted, medium-grained wind-blown sand) (Picard, 1977).

**Glen Canyon Sandstone Oasis Deposit (JṖgo)**

One of the more unexpected and exciting discoveries found within the map area was that of a freshwater carbonate, or, specifically, a light brown to tan to cream-colored dolostone that demonstrates classic fenestral fabric and contains risoliths on it’s upper surface; it is located about 2/3 up from the base of the Upper Glen Canyon Sandstone (Figures 2 & 11). This bed is thin (approximately 10 feet thick), but extensive in out crop. It is a good stratigraphic marker within the Upper Glen Canyon Sandstone and provided a means for correlation within this otherwise massively bedded
formation. For this reason, it was mapped in each quadrangle as a single unit. Carbonate and marlstone beds much like this found within the Navajo Sandstone in southern and central Utah have been interpreted to be interdune “oasis” deposits (Gilland, 1979; Anderson et al, 2000). These oasis deposits have also been recently described by Erik Waiss in the Canyonlands National Park area, near Moab, Utah (Master’s Degree, University of Nebraska, expected 2004/2005). Additional information on interdune oasis deposits in the region can be found in Driese (1983).

Perhaps keying into this lone bed embedded within a massive erg deposit—void of other chronological markers would reveal additional important, time-relevant discoveries that might help give an age for this formation. There exists a semi-regional, silty, reddish surface immediately under this carbonate oasis bed that may to be equivalent to the red-beds near Red Fleet Reservoir that contain dinosaur tracks (Hamblin et al, 2000). More detailed stratigraphic studies are needed to verify this possible correlation.

**Jurassic Rocks**

The Jurassic System consists of the Upper Glen Canyon Sandstone (Late Triassic-Early Jurassic), the Carmel Formation (Middle Jurassic), the Entrada Sandstone (Middle Jurassic), the Stump Formation (Middle and Upper Jurassic), and the Morrison Formation (Upper Jurassic) (Peterson, 1972; Imlay, 1980; DeCelles et al, 1992; Uygar and Picard, 1985). Except for the Carmel Formation and Stump Formation, which were deposited as separate inland seas transgressed across this region, the Jurassic formations are all terrestrial eolian and fluvial overbank deposits.
**Carmel Formation (Jc)**

The Carmel Formation (Imlay, 1967; Fritz, 1977; Jordan, 1987; Kinney, 1955; Gregson and Chure, 2000; Pipiringos and O'Sullivan, 1978) is exposed throughout both quadrangles. Within the Steinaker Reservoir Quadrangle, it is exposed in the southern part most often as a low-lying, curved ridge that eventually snakes its way northward, then disappears into alluvium (Figure 12). The Carmel Formation is exposed within the Dry Fork Quadrangle in fault blocks within the ravines of Coal Mine Basin, exposures near the top of Castle Cove, and as a solitary ridge west of Little Mountain under landslide debris.

The Carmel Formation is mostly made up of thick, deep red-colored siltstone- and mudstone-mixed beds capped by thinner, greenish-colored beds of the same lithology; these are in turn capped by a thin, durable, sandy, fossiliferous, wackestone carbonate bed. This red-green carbonate sequence is repeated three to four times within the formation. The final cycle is capped by a thick, massive gypsum bed (as thick as four feet in some areas) instead of a carbonate bed. In the Steinaker Reservoir Quadrangle, west of Steinaker State Park, there are two gypsum beds each of about three feet thick (maximum) separated by a thin mudstone bed. After this gypsum bed, a 10 foot thick bed of several thin, variegated siltstones and mudstones exist at the top of the unit. The Carmel Formation also contains completely jasperized fossil bivalves and a few thin, biotite-bearing ash beds.
The Carmel Formation is not split into different members. It is approximately 150-220 feet thick. The package of sequences (parasequences) represents several transgression-regression fluctuations of the Jurassic-aged Sundance Seaway.

**Entrada Sandstone (Je)**

The Entrada Sandstone (Stephenson, 1961; Otto, 1973; Kocurek, 1981; Otto and Picard, 1975a & 1975b; Imlay, 1967; Kinney, 1955; Gregson and Chure, 2000; Pipiringos and O’Sullivan, 1978) is rarely exposed in the map area—in fact, with a few exceptions it almost always forms a strike valley. The upper portion is well exposed in Steinaker State Park north of the ranger station in the Steinaker Reservoir Quadrangle (Figure 12); the lower and upper portions are exposed at differing localities in Coal Mine Basin, in the Dry Fork Quadrangle. Where the Entrada Sandstone is exposed, it weathers to rounded outcrops. Many adjacent eolian and alluvium deposits are derived from the Entrada Sandstone because it weathers easily. In this region, the Entrada Sandstone pinches-out to the north, though this is not apparent in the map area (Gregson and Chure, 2000).

The Entrada Sandstone is made up of three marine- and terrestrial-mixed, earthy, orange-colored, sandy- and silty-mudstone layers of about equal thickness (3-15 feet thick), separated by two, thick (70-90 feet), yellow–stained, white- to grey-colored, medium-grained, friable, planar and trough cross-bedded sandstones. The sandstone beds are generally so weakly cemented that one can gouge a hole in them with little effort (a finger even).
The Entrada Sandstone is approximately 160-215 feet thick. Because of the colluvial nature of this unit, this total is based on best estimates and regional thicknesses.

**Stump Formation**

The Stump Formation (Eschner, 1983; Eicher, 1955; Stephenson, 1961; Thomas and Kruegar, 1946; Patterson, 1980; Kinney, 1955; Gregson and Chure, 2000) is exposed throughout both of the map quadrangles (Figure 12). In the Steinaker Reservoir Quadrangle, the Stump Formation forms an east-west oriented ridge in the south, which is folded and bends northward in an S-shape, before reorienting itself back to an east-west direction south of Red Mountain. In the Dry Fork Quadrangle, the formation is exposed in several places: in Coal Mine Basin, on the northeastern rim of Little Mountain, at the western foot of Little Mountain under the landslides, and in several other places on the western edge of Little Mountain where it is cut by faults into isolated faulted blocks. The Stump Formation is often completely exposed, except where covered by Quaternary deposits such as landslides. The formation is split into two members.

The basal member is a white, very coarse-grained, pebbly, glauconitic, cross-bedded sandstone called the Curtis Member ($J_{sc}$). The upper stacked sequences of brown- to olive-colored, sandy, oolitic, sometimes cross-bedded carbonate beds interbedded with distally-deposited, dark olive-colored, gypsiferous, shale beds is called the Redwater Member ($J_{sr}$). The Redwater Member is made up of
progressively shallowing-up parasequences. In places, the shale beds also host abundant cigar-shaped belemnite fossils.

The Curtis Member-Redwater Member contact is easy enough to locate as its white sands terminate below a contrastingly dark colored shale bed. One of the questions posed of the Stump Formation was whether or not its members could be mapped separately, perhaps indicating a possibility that they should be formally separated into individual formations (Doug Sprinkel, personal communication). True, the Curtis Member is relatively thin, but thick enough that it can be mapped at the 1:24,000 scale. It is the author’s opinion, though, that they be kept within the Stump Formation because the Curtis Member, though mappable, might indeed be too thin.

The whole unit is 220-270 feet thick (the Curtis Member is 40-90 feet thick and the Redwater Member is 180 feet thick).

**Morrison Formation (Jm)**

The Morrison Formation (Peterson, 1986; Curie, 1997; Dawson, 1970; Turner and Peterson, 1992; Kowallis et al, 1998; Kinney, 1955; Gregson and Chure, 2000; Pipiringos and O’Sullivan, 1978), like the Entrada Sandstone, is rarely exposed in the quadrangles, but instead forms strike-valleys. The Morrison Formation is present west near Little Mountain in two places in the Dry Fork Quadrangle. Northwest of Little Mountain, it is mostly covered by colluvium, but west of Little Mountain, it is exposed where it has not been covered by landslides. Bedding at this location is difficult to recognize because it is adjacent to a fault. Occasionally, in the Steinaker Reservoir Quadrangle, the Morrison Formation is exposed as spurs on the
backs of ridges held up by the Redwater Member of the Stump Formation. Also, in the alluvium-filled strike valleys formed in the Morrison Formation, fluvial channel sandstones may outcrop occasionally near the surface. The best exposure of the Morrison Formation, however, is at the eastern edge of the Steinaker Reservoir Quadrangle, south of highway 191. There, the Morrison formation is stratigraphically intact and nearly fully exposed.

The Morrison Formation is made up of variegated terrestrial mudstones (much like the Mottled member of the Chinle Formation) occasionally interbedded by somewhat well-cemented, pebble-conglomerate fluvial channels. Colors in the unit change across the map area and range from a mixture of the following colors in any given place: purple, red, green, and white, mainly, with pink, gray, and yellow as accessory colors. The Morrison Formation is famous for the dinosaur quarry not 30 miles from here at Dinosaur National Monument (Gregson and Chure, 2000). Petrified wood can also be found in the Morrison Formation.

The contact between the Morrison Formation and the Cretaceous Cedar Mountain Formation is not easily recognized in the map area. Currently, there is work being done to address this issue, but this will be addressed in the following section detailing the Cedar Mountain Formation.

Complete exposures of the Morrison Formation can be found just east of the Steinaker Reservoir Quadrangle in the Donkey Flat Quadrangle. The best exposures there are along the strike-valley following Highway 191 on the east and west side of Red Fleet Reservoir—a complete section is exposed at each of those locations. The Morrison formation is approximately 620-650 feet thick.
Cretaceous Rocks

The Cretaceous System is represented in the area by the Cedar Mountain Formation (Early Cretaceous), the Dakota Sandstone (Early Cretaceous), the Mowry Shale (Early Cretaceous), the Frontier Formation (Late Cretaceous), and the Mancos Shale (Late Cretaceous) (McGookey, 1972; Weimer, 1986; DeCelles et al., 1992; Molenaar and Cobban, 1991). The Cedar Mountain Formation was deposited in a continental environment with rivers, lakes, and overbank deposits. The other units represent fluctuations of the Cretaceous Western Interior Seaway as it first transgressed and then dominated this part of the continent during the Cretaceous Period.

Cedar Mountain Formation (Kc)

The Cedar Mountain Formation (Young, 1975; Currie, 1997; Kinney, 1955; Gregson and Chure, 2000; Molenaar and Cobban, 1991) is exposed along the base of the ridge found south and east of Highway 191, in the Steinaker Reservoir Quadrangle (Figure 13). From there, it remains exposed (except for the presence of the Steinaker Dam) at the base of the same ridge on the western side of Steinaker Reservoir, south of the Steinaker Feeder Canal. In the Dry Fork Quadrangle, the Cedar Mountain Formation is exposed partly at the ridge south of the Coal Mine Basin and is exposed under since-eroded landslide debris on the east and west sides of Little Mountain.
The Cedar Mountain Formation is made up almost wholly of mudstones of two main colors: light red or pink, and gray. Freshwater carbonate beds pinch and swell at differing levels within the formation, but are mainly located in the upper half of the formation at the base of and within the gray-colored portion. The Cedar Mountain Formation could be separated in the map areas into a lower pink member and an upper gray member.

The lower pink member is mainly made up of the reddish or pink mudstones previously mentioned, but has some banded purple- and white-colored calcareous mudstone beds at its base. It contains small flint and jasper chert pebbles throughout, and contains an abundance of fist-sized, dense, carbonate soil concretions that create (and make it difficult to walk on) the semi-steep slopes characteristic of the lower pink member. It also contains an abundance of gastroliths—smooth stones believed to have been swallowed by dinosaurs and kept in their stomachs to help them digest their food.

The upper gray member is made up of dark-colored siltstone and calcareous mudstone beds. It contains several sandstone beds and the freshwater carbonates. Some of the carbonate beds break up into large, boulder-sized concretions that contain light blue chalcedony and crystallized-quartz veins.

One of the more obscure stratigraphic contacts in the map area is the one between the Jurassic Morrison Formation and the Cretaceous Cedar Mountain Formation, since both units are lithologically similar and seem to grade into each other. Ash beds near the contacts show, however, that the unconformity between these formations is approximately 50 million years. This large gap in time is not very
evident because of the nature of the contact. Currently, this contact is tentatively
designated by a yellowish “alteration layer” near the base of the lower pink member
(Scott Madsen, personal communication). Although this layer has been confirmed in
other nearby areas and seems to at least have a semi-regional occurrence, the author
has discovered two of these layers about 2-3 feet apart south of the Steinaker Feeder
Canal. This discovery complicates the existence of a single unconformity.

Elsewhere in the Colorado Plateau and as nearby as Dinosaur National
Monument, the Buckhorn Conglomerate (Kcb), located at the base of the Cedar
Mountain Formation, marks this contact (Currie, 1997; Gregson and Chure, 2000).
Until now, the Buckhorn Conglomerate was thought not to exist near Vernal. In fact,
it may exist in discreet, isolated channels, which pinch-and-swell intermittently along
the Morrison Formation-Cedar Mountain Formation contact, but even though this
system of channels appears to correlate within the map area, they have not been
confirmed as the true Buckhorn Conglomerate (Figure 13). Because this system of
channels exists so close to the Morrison Formation-Cedar Mountain Formation
contact, it was used as the contact between the two for mapping purposes within both
quadrangles. Though nearby ash beds or traces of pollen would be ideal in
determining whether or not this channel is the Buckhorn Conglomerate, these have
yet to be found. This issue is being worked on by Brent Greenhalgh (BYU Master’s
Thesis expected Fall, 2006).

The Cedar Mountain Formation is 210 feet thick (the lower pink member and
the Upper Gray Member have nearly equal thicknesses.
Dakota Sandstone (Kd)

The Dakota Sandstone (Kinney, 1955; Gregson and Chure, 2000; Molenaar and Cobban, 1991) represents the first transgressive sequence of the Cretaceous Western Interior Seaway. It is fully exposed everywhere above the Cedar Mountain Formation about halfway up the same ridges in the map area (Figure 13). In the Dry Fork Quadrangle, it can be seen in additional locations through the landslide debris on the edges of some ravines.

The Dakota Sandstone is made up of two 20-45 feet thick, pebbly and coarse-grained nearshore sandstones with an occasional occurrence of conglomerate lenses; the two thick sandstone beds are, in turn, separated by a 40-50 foot thick layer of black, offshore shale, which is occasionally locally bedded with thin sandstone and coal facies. The sandstones are often well cemented, but may be friable in other areas. The sandstones are often stained with a particular yellow coloring (limonite staining); this staining is very indicative of the Dakota Sandstone and is helpful in identifying it in poorly exposed areas. Also of note, the top of the lowest sandstone in contact with the base of the shaly layer above contains petrified wood.

The Dakota Sandstone is 115-140 feet thick.

Mowry Shale (Kmo)

The Mowry Shale (Molenaar and Wilson, 1990; Kinney, 1955; Gregson and Chure, 2000; Molenaar and Cobban, 1991) is exposed above the Dakota Sandstone in the map area (Figure 13). Its silvery appearance and distinct lithology makes it one of the easiest units to spot along the ridges in the south and east parts of the Steinaker
Reservoir Quadrangle and even under colluvium and landslide debris in the Dry Fork Quadrangle. The Mowry Shale is exposed nearly everywhere the other Cretaceous units are exposed.

The Mowry Shale consists of an indurated, chippy, siliceous, marine shale that is dark bluish-gray in color on a fresh surface, and appears dusty-grey to light blue on a weathered surface. It contains abundant fish scales and bones, shark’s teeth, and even a species of crustacean has been found (Stewart et al, 1994; Kowallis and Anderson, 2005). The Mowry Shale has preserved several tens of ash beds now altered to bentonite, much of which is mined in Wyoming.

The Mowry Shale has been determined to be a relatively unstable unit; as many slumps and abundant intra-formational deformation has been observed. It has also acted as a structural detachment surface—even when the shale within the Dakota Sandstone and mudstones of the Cedar Mountain Formation seem relatively undeformed by the same processes directly around and below the highly deformed Mowry Shale.

The Mowry Shale is 90-120 feet thick. An excellent exposure is located along Highway 191 at the road cut north of the Steinaker Reservoir Dam, but other nearly complete exposures exist along the ridges in the Steinaker Reservoir Quadrangle.

**Frontier Formation (Kf)**

The Frontier Formation (Maione, 1971; Ryer and Lovekin, 1986; Trexler, 1957; Molenaar and Wilson, 1990; Kinney, 1955; Gregson and Chure, 2000; Molenaar and Cobban, 1991) is the capping unit of the southern most and eastern
most ridges in the Steinaker Reservoir Quadrangle (Figure 13). It caps the ridge south of Coal Mine Basin and can be found elsewhere in the Dry Fork Quadrangle in the ravines west of Little Mountain, south of Pine Ridge, and even at high elevations, very near the contact with the Bishop Conglomerate on the western edge of Little Mountain.

The Frontier Formation consists of cream-colored to almost rusty, brown-colored sandstones interbedded by shale and mudstone beds and coaly facies. It can be separated it into three, Upper, Middle, and Lower members.

The Lower member consists of a shallowing-up set of brown-colored, gypsiferous shale beds capped by durable, carbonate-rich, brown-colored sandstone beds. These sandstones are the beds which cap the high ridge on the eastern and southern portion of Steinaker Reservoir Quadrangle. Trace fossils and oyster beds are found within this Lower member.

The Middle member consists of the cream-colored, wave-dominated, nearshore sandstones interbedded by coaly or very carbonaceous beds. Two of these thick sandstone beds exist. The coaly facies pinches and swells as does the quality of the coal. The coal here, though no longer economic, was an important source of fuel for locals before better and cheaper sources of energy, like the Flaming Gorge Dam and hydro-electric plant, became available. This Middle member is very often stained by hydrothermal fluids and can be confused with the Dakota Sandstone, but the Frontier Formation has a darker, often brown stain.

Finally, the Upper member is made up of a mantle of 30-40 feet thick, black marine shale capped by a thin sandstone layer which is embedded with a peculiar, but
locally common phenomenon, known as the “cannonballs” (Figure 14) (Doug Sprinkel, personal communication). The top of this layer marks the contact between the Frontier Formation and the Mancos Shale. The “cannonballs” are made up of a sandy carbonate that can be as dark in color as the darker Mississippian- and Pennsylvanian-aged shelf carbonates found in the region. They also weather in a horizontally crinkly fabric. It is believed that the “cannonballs” in the Frontier Formation (which, in the Donkey Flat Quadrangle, are as large as a Volkswagen Beetle) are carbonate-cemented, sandy concretions produced by groundwater processes (Dutton, et al, 2002). These “cannonballs” are often protruding from the topmost sandstone bed of the Frontier Formation and seem to have subsided into the underlying, thin sandstone bed. That observation, the fact that they weather in a crinkly, horizontal fabric (not indicative of concentric formation from a central nucleus), and that the composition of the “cannonballs” contains such a high percentage of calcite-to-sand within the map area might cast doubt toward the fact that these are nothing more than large, groundwater-influenced carbonate concretions. It is the author’s opinion that these may represent stromatalitic patch-reef buildups, but that is yet to be worked out. Additional information on this topic is necessary to clear up this matter.

The Frontier Sandstone is approximately 140-270 feet thick in the area.

**Mancos Shale (Kms)**

The Mancos Shale (Johnson, 1990; Cole and Young, 1991; Kinney, 1955; Gregson and Chure, 2000; Molenaar and Cobban, 1991) is the thickest geologic unit
in the map area, and is also one of the least variable units, lithologically. It is exposed in the eastern and southeastern portion of the Steinaker Reservoir Quadrangle. In the Dry Fork Quadrangle, it is exposed on both sides of Little Mountain in the southern portion below much of the landslide debris, but can also be found in contact with the Bishop Conglomerate high up on Little Mountain.

The Mancos Shale is made up almost entirely of distally deposited, dark gray marine shale (Cole and Young, 1991). It is very homogenous and there is little color change throughout, though it lightens in color where of coarser silt and sand grains are present. The Mancos Shale weathers into badlands topography, in gray and yellowish tints, and is often capped by thick alluvial pediment surfaces. Only the lowest section is known to be exposed in the two quadrangles (there is no simple field technique of telling the intraformational position of outcrops in Dry Fork Quadrangle).

The thickness of the Mancos Shale is 4,700 feet thick. A complete stratigraphic section is very difficult to find, but one exists south of Coal Mine Basin, south of the Dry Fork Quadrangle—barring any offset from hidden faults. The difficulty finding a complete section is due to a few factors: (1) it is such a thick unit, (2) there are only subtle marker beds or changes of lithology within the unit, and (3) there is a fair amount of folding between Coal Mine Basin and Split Mountain, and south where the Mesa Verde Formation caps the Mancos Shale—this folding is difficult to see in the field, but can be interpreted from lithology changes (represented by color patterns) on aerial photos. The differences alone in the dip and strike, and
the inferred thickness between the Frontier Formation and the Mesa Verde Formation vary greatly, which suggests some folding between them.

**Duchesne River formation (Td)**

The Duchesne River Formation (Hansen, 1984; Anderson and Picard, 1972) is only exposed on Little Mountain below the Bishop Conglomerate and in the southwestern corner of the Dry Fork Quadrangle (Figure 15). It shows evidence of at least one major intraformational unconformity in this area as the steeper dipping beds (30° to 40°) at the base of the formation seem truncated by more horizontally-deposited beds at the top of the formation.

The Duchesne River Formation is made up of thick beds of well-cemented conglomerates occasionally interbedded with light-colored, coarse-grained sandstones and orange-colored mudstone and siltstone beds. Clasts are commonly 1-2 feet in diameter and of Paleozoic origin, which inversely grade into a majority of Precambrian Uinta Mountain Group quartzite clasts—commonly purple in color (Doug Sprinkel, personal communication).

The Duchesne River Formation is a well known unit in the Uinta Basin as it is the youngest of the major basin-fill formations and outcrops locally over a wide area. It is broken up into four members: the oldest is the Brennan Basin Member, next is the Dry Gulch Member, then the LaPoint Member, and finally the Starr Flat Member. It is unclear if all four members exist in the Dry Fork Quadrangle. The Brennan Basin Member may be the unit making up the high angle bedded slopes (as it represents the first deposition of the Duchesne River Formation). It may be truncated.
by the younger part of the Duchesne River Formation (Starr Flat Member), creating the angular unconformity. The Starr Flat Member is deposited high on Little Mountain and may be gradational, rather than disconformable, with the Bishop Conglomerate. (It is interesting to note that the Starr Flat Member is the only member that is sourced from the Uinta Mountains and does not appear on the south side of the axis of the Uinta Basin syncline. Could it be that this member is closer related to the Bishop Conglomerate?)

It was difficult to determine the thickness of the Duchesne River Formation under the many gravel deposits, but its estimated thickness is 520 feet.

**Bishop Conglomerate (Tb)**

The Bishop Conglomerate (Hansen, 1984; Hansen, 1986; Gregson and Chure, 2000; Sprinkel et al, 2000) is exposed in the northwest corner of the Steinaker Reservoir Quadrangle high on the flanks of the Uintas (Figure 15). Similarly, it is exposed at about the same elevations in the northern portion of the Dry Fork Quadrangle. It is also the bulk of the rock capping Little Mountain in the same quadrangle.

The Bishop Conglomerate was recognized and named by John Wesley Powell in one of his expeditions in the eastern Uinta Mountains (Powell, 1876; Sears, 1924). It is deposited on a regional peneplane surface called the Gilbert Peak erosion surface (Marsell, 1969). This surface was created during a hiatus of uplift during the mountain-building period of the Uinta Uplift. The Bishop Conglomerate, however, rarely outcrops except in small windows here and there where, in the map area, it is
exposed as several thick, massive, well-cemented conglomerate beds made up mostly of Uinta Mountain Group quartzite clasts and several Paleozoic clasts. These conglomerate beds are interbedded with thin coarse-grained sandstone lenses, and, near the contact with the Duchesne River Formation, fine-grained, orange-colored mudstones and siltstones. In one location, high on Little Mountain, a three foot thick, well-sorted, medium-grained sandstone bed has been found and, in other areas in the region the Bishop “Conglomerate” is not a conglomerate at all—it is entirely made up of a massive, pink-colored, tuffaceous sandstone (Figure 16). One of the unknowns about the Bishop Conglomerate is that its internal facies relationships are unknown. Whether or not this sandstone is older or younger than the conglomeritic portion of the Bishop Conglomerate or whether it represents a lateral change in facies has yet to be determined.

The Bishop Conglomerate is often responsible for the many large-scale landslides in the region, provided certain conditions are right for failure. More on this topic will be discussed in the mass movement deposits portion of the Quaternary Deposits section.

As mentioned, there is an issue, supported by the detailed mapping of the two quadrangles, that may show that the Bishop Conglomerate, originally thought to be younger than, and not equivalent to, the Uinta Basin’s Duchesne River Formation, may in part be contemporaneous with it (Kowallis et al, 2004). They showed that the upper part of the Duchesne River Formation is equivalent in age to the Bishop Conglomerate, but facies relationships in the field had not, until now, demonstrated a relationship between the two. Because of this relationship, picking a contact between
the two units while mapping was a problem, and more detailed study of this issue may change the current location of the contact (Figure 15). The contact was chosen low on Little Mountain to honor the older conclusions that they are separate units—until further evidence can support a stronger correlation between the two units.

The thickness of the Bishop Conglomerate seems to change across in the map area. In fact, the Bishop Conglomerate thickens from the western highlands of the Dry Fork Quadrangle toward the east, and then thins again into the Steinaker Reservoir Quadrangle. Little Mountain may exist along the axis of a thickened wedge of the unit, perhaps along the axis of a master drainage system or channel.

On average, when present, it is 200-500 feet thick.

**Quaternary Deposits (Q)**

In order to be as precise as possible while mapping, the level of detail and spectrum of Quaternary-aged units or deposits defined in the Steinaker Reservoir and Dry Fork Quadrangles was initially open ended; Quaternary Deposits were defined according to type (depositional process and lithology) and rank (hierarchy of dominant processes). For example, if a partly eolian and partly alluvial and partly colluvial deposit of sand were blanketing the Upper Glen Canyon Sandstone, then the character representing the dominant process (e.g. ‘a’ for alluvium) would be indicated first after the ‘Q’ in the symbol, ending in the least dominant (e.g. ‘c’ for colluvium and ‘e’ for eolian), then that would be followed by the character representing the sediment type (e.g. ‘s’ for sand). In this case, the symbol, ‘Qaces,’ may be the resulting combination of characters for that particular deposit. This type of objective
logic was used for all of the Quaternary-aged deposits. The final version of the maps, however, does not represent the original names resulting from this hierarchy of naming symbols, but for the reader’s sake, they have been simplified to show only the major processes that best describe each deposit.

With that in mind, there are at least six different types of Quaternary deposits that occur in the Dry Fork and Steinaker Reservoir Quadrangles (with their accompanying sub-types within the majority of the groups). These groups are named as follows and will be addressed in detail hereafter: alluvium, colluvium, eolian, mass movement, and human-influenced deposits, and what is called the unnamed Pleistocene unit.

**Unnamed Pleistocene Unit (Qp)**

The unnamed Pleistocene unit is regional in extent sitting usually on top of the Bishop Conglomerate. Mapping of the Dry Fork and Steinaker Reservoir Quadrangles by the author combined with the previous mapping of the 30° x 60° Dutch John Quadrangle (of which the Steinaker Reservoir and Dry Fork Quadrangles are included) by Doug Sprinkel, of the Utah Geological Survey, have led to the belief that the loosely consolidated material laying directly above the Bishop Conglomerate is not simply weathered Bishop Conglomerate. Because of the colluvial nature of the contact, only a few solid observations collected within the region show this unit cutting into the Bishop Conglomerate. For that reason, the contact between the two units is often inferred.
Many very large boulders (upwards of 6-10 feet in diameter, Figure 17) are part of this deposit sitting above the Bishop Conglomerate and within the landslide debris sourced from it. Boulders of this size do not occur within the Bishop Conglomerate itself, but are common in this younger deposit. The wetter climate during the Pleistocene may have contributed to moving such large-sized boulders so far from their source (approximately seven miles).

The unnamed Pleistocene unit is approximately 50 feet thick in the Dry Fork Quadrangle but only 30-40 feet thick in the Steinaker Reservoir Quadrangle. It is believed that this unit, like the Bishop Conglomerate, is thickened in the Dry Fork area.

**Alluvial Deposits**

Alluvium within the quadrangles differs significantly from one location to another, as one would imagine, according to depositional processes and nearby geologic source units. In many cases, especially in the Steinaker Reservoir Quadrangle, the deposits are dominantly fine-grained or sandy and contain isolated lenses of coarser material (pebble- to boulder-sized). This is largely because there is much less gravel originating from the Bishop Conglomerate and because units like the Moenkopi Formation, the Glen Canyon Sandstone, Entrada Sandstone, and Morrison Formation—the sandy and fine-grained units—are the bulk of the units being eroded in the area. The Dry Fork Quadrangle alluvium is coarser because of the prevalence of conglomeritic units and gravel deposits within the quadrangle (maybe 50-60% of the area).
Present ephemeral and permanent stream beds carry cobble and boulder-sized clasts, except in much of the Steinaker Reservoir Quadrangle, where only fine-grained sediment and sand are found.

Alluvial pediments are a common deposit type—especially in the Dry Fork Quadrangle (Figure 18). These deposits are commonly made-up of Precambrian Uinta Mountain Group boulders mixed in a matrix made up of the nearby, softer-unit-derived clasts and sediment. These are located in various places and at various elevations throughout the quadrangles. In the Steinaker Reservoir Quadrangle, there is a semi-consolidated alluvial pediment near Steinaker State Park that is derived mostly of Glen Canyon Sandstone and Entrada Sandstone sands, which includes root casts and risoliths.

Spring alluvium deposits are rare, but exist in small amounts in both quadrangles. These deposits are usually very fine-grained and their extent is defined on the maps.

Several other alluvial deposits are mixed with colluvium and eolian deposits; these are quite common and are spread across the map area. These deposits range in thickness from a couple feet to up to over ten feet in thickness. The Spring Creek cut-banks may be as tall as 30 feet, showing the thickness of the deposits in some valleys.

**Colluvium Deposits**

Colluvial deposits are common throughout both quadrangles. These may be gravel deposits, sand deposits, silt and mud deposits, or mixed gravel-, sand-, and finer-grained deposits. The lithology of the deposit is largely due to the rock type
from which it is derived and, occasionally, from older alluvial and colluvial deposits, such as the gravels and boulders sourced from higher-elevation landslide or alluvial pediment deposits. Because colluvium is usually situated on a slope, these deposits are not very thick. Like the alluvium deposits, these deposits can also be found as mixed-process deposits.

**Eolian Deposits**

Virtually all eolian deposits are found within the environs of the Upper Glen Canyon Sandstone. These deposits may be accumulated as trapped, windblown sediment under baffling vegetation, as low-lying, basin- or canyon-trapped windblown sediment, or as dune deposits, which are commonly found along the wind-ward side of high ridges. Eolian deposits can also be found as mixed-process deposits.

**Human-Influenced Deposits**

These deposits are few, but can be very large. They range in diversity from mine till, highway fill, dam fill, or even where an old airstrip had been created. These “deposits” are defined in the few places where the geology or surface has been altered by man-made operations.
**Mass Movement Deposits**

Two types of mass movement deposits exist in the map area. These are: slumps and landslides. These vary in age between each other and within the differing types. While mass movement deposits exist in both quadrangles, the type of deposit dominant each quadrangle is different. Only a few slump deposits exist in the Dry Fork Quadrangle, while in the Steinaker Reservoir Quadrangle, only a few landslide deposits exist.

Slumping in the map area is generally restricted to two incompetent, culprit formations: The Chinle Formation and the Mowry Shale (Figure 19). The overlying weight of the lower Glen Canyon Sandstone and Frontier Formation, respectively, is what usually causes either the Chinle Formation or Mowry Shale to weaken and fail.

As indicated in the previous section, there is evidence that landslides have been an active process in this area (Osmond, 1969). In fact, landslide deposits prevail in the Dry Fork Quadrangle, and many of the surrounding quadrangles in the area (Kowallis and Bradfield, 2005). The landslide deposits are much larger deposits than the slump deposits; they are up to 1 mile long and 2-3 miles wide (Figure 20). These deposits occasionally weather into alluvial and colluvial deposits within closed drainages within their coverage. “Windows” within the coarse surficial landslide debris show underlying geologic units sometimes intact—it is unclear whether or not these are in correct stratigraphic position or whether they re included as part of the landslide debris (e.g. the Cedar Mountain Formation, Dakota Sandstone, Mowry Shale, Frontier Formation can be seen in succession on the eastern slopes of Little Mountain within the heart of the landslide debris).
As indicated in the previous section, the cause of the landsliding is due to over-steepening of the Bishop Conglomerate on the flanks of the Uinta Mountains. It has been determined, through this detailed mapping project, that the likelihood of failure of the Bishop Conglomerate is much higher when it rests on top of older, incompetent materials that, in turn, were deposited above a stronger, durable layer—specifically, the combination of the Morrison Formation-Redwater Member of the Stump Formation, the Dinwoody Formation-Park City Formation, and, in some cases, the Mancos Shale-Frontier Formation have contributed significantly to the possibility of failure.

The landslides in the immediate area are quite spectacular. Mapping around the landslides has been difficult, but rewarding. Most errors mapping the Deep Creek fault zone, for example, have been due to misinterpolation of faults through or under the landslide debris. Careful attention to the rocks under the landslides has cleared up many of these types of problems. Todd Bradfield is currently dating the landslide deposits within the region; his findings should be presented in Fall of 2006 (BYU, Master’s Thesis).

**Structural Geology**

The geologic structures within the study area are deformed by both the regional-scale Uinta Uplift and the more localized Deep Creek fault zone. The Dry Fork and Steinaker Reservoir Quadrangles are located on the southern fringe of the Uinta Mountains physiographic province. Approximately the northern halves of the quadrangles are characterized by gently south-dipping strata off of the southern flank.
of the Uinta Uplift. The southern halves, on the other hand, are influenced mainly by
the Deep Creek fault zone and broad folding probably associated with it (Untermann
and Untermann, 1969). The folding may be due to an unknown tectonic cause, but
both the folds and faults appear to be late Laramide age structures.

**Deep Creek Fault Zone**

The Deep Creek fault zone is a major structural element within the two
quadrangles (Untermann and Untermann, 1969). It is made up of a combination of
horst, graben, and half-graben fault blocks produced by steep faults with evidence for
both normal- and oblique-slip movement. An in-depth study of the kinematics of the
Deep Creek fault zone is contained in Chapter 2.

**Folding**

Aside from the faulting within the Deep Creek fault zone, folding is the
dominant form of deformation within the map area, especially in the Steinaker
Reservoir Quadrangle (Untermann and Untermann, 1969). The folding is usually
broad and sometimes nearly imperceptible, but a few significant folds in the central
and southeastern part of the Steinaker Reservoir Quadrangle will be described here.

**Ashley Creek Syncline**

The Ashley Creek syncline (named here) is located just southeast of the mouth
of Ashley Gorge (Figure 21). It is an asymmetrical syncline, with dips as steep as 80°
toward the south on the northern hinge, and as low as 5° to the north on the southern side. The Ashley Creek syncline was probably formed by a subsurface graben related to the Deep Creek fault zone as the axis trends the same direction as the bulk of the faults in the Deep Creek fault zone. The Ashley Creek syncline is not exposed west across the river bed along its axis; it dies-out, or, flattens in the opposite direction across Spring Creek.

**Kink-and-Racetracks Folds**

The Kink-and-Racetracks folds (named here) may partly be an eastward extension of the Ashley Creek syncline. They are made up of three plunging folds: a syncline, an anticline, and a syncline, beginning at the base of Red Mountain, on the north, and progressing south toward the very southeastern edge of the Steinaker Reservoir Quadrangle. These folds are east-plunging folds with bedding that largely keeps a constant dip of about 20°. The Kink anticline (named here), however, has bedding that dips upwards of 50° or so on its northern hinge, and 20° south-dipping, to 80° overturned, north-dipping beds on its southern hinge. A few small faults also exist within the Kink anticline of which, one in particular, has nearly zero displacement at the top of the ridge (within the Frontier Formation), but nearly 15-20 feet of displacement occurs just 50 feet or so along the fault trace from the tip (juxtaposing the Mowry Shale and Dakota Sandstone, Figure 22). West of the Kink anticline, where the Carmel Formation outcrops as a ridge, there exists a small reverse fault with about 2 feet offset; this fault appears to have a left-lateral strike-slip
component along its trace (no slickenlines were found at the fault surface, grooves were used instead).

Confluence Dome

The Confluence dome (named here) is located in the south-central section of the Steinaker Reservoir Quadrangle, but continues west, into the Dry Fork Quadrangle. This is the most subtle of the structural features in the map area. It has shallow dips of about 7° maximum and trends east-northeast and west-southwest. The apex of the Confluence dome is somewhere near where Ashley Creek and Dry Fork converge. It is best viewed along Taylor Mountain Road, north of Maeser City, where the durable Glen Canyon Sandstone oasis deposit holds up a broad ridge along which a shallow, but narrow, doubly-plunging syncline perpendicular to the axis of the Confluence dome is super imposed (northwest-southeast).

The Racetracks Structural Area

One of the more exciting combinations of geologic structures in the area is also one of the most enigmatic in origin. The Racetracks structural area, (named here) located at the eastern edge of the Steinaker Reservoir Quadrangle along the ridge south of Highway 191 (on the back side of the ridge), has several tightly packed anticlines and synclines, which are occasionally overturned (Figure 23 & 24). Yet the northern side shows only subtle traces of deformation (Figure 25). Careful mapping has shown that at least three of these types of highly-deformed areas exist along the
same ridge shared by the Steinaker Reservoir Quadrangle and the Donkey Flat Quadrangle.

It took several days mapping to understand the structures here because of the complexity and nature of the Frontier Formation, but a few facts can be stated about the Racetracks structural area: (1) the Frontier Formation has been slightly thrust up over the Mancos Shale, (2) there is significant shortening within the Mowry Shale and Frontier Formation on the back side of the ridge, (3) the structural detachment is within the uppermost Mowry Shale, and (4) there is no obvious trigger that would cause this type of folding (regional forces would not likely concentrate folds like these in such a small area without interfering with underlying units).

Tureva-like, large-scale gravity sliding is the leading theory for the creation of the Racetracks structural area. Although there is virtually no evidence that this kind of event happened (there is a lack of debris from up dip geologic units), there is strong evidence that landsliding had once been an active process up-dip, on Red Mountain. There, gravel-capped piles of rock consisting of intact Chinle Formation and Lower Glen Canyon Sandstone are superimposed over the Upper Glen Canyon Sandstone. In at least five instances, a portion of the Upper Red member of the Chinle Formation is in direct contact with the ripple-laminated sandstones of the Lower Glen Canyon Sandstone—this contact is occasionally preserved and in good condition (Figure 26). These deposits are sometimes found very low off of Red Mountain, near where the Glen Canyon Sandstone is in contact with the Carmel Formation. This juxtaposition of older units above a younger unit is very likely the result of a landslide event. This same type of unit-preservation within a landslide can
be found in the Dry Fork Quadrangle, where a 6-foot by 15-foot “block” of the Morrison Formation is found at a very low elevation, just over one mile away from its source with a 1,400 foot change in elevation. This, coupled with the observation that mass movement processes have historically occurred throughout the region, supports the theory that there were landslides on Red Mountain in the past. But, could they have been large enough to deform and fold the Frontier Formation to the degree it has been today where such significant shortening and minor thrusting has occurred on the ridge?

**Joints**

Joints in the map area can best be seen in the Glen Canyon Sandstone and the Gartra Member of the Chinle Formation. Occasionally, they can be seen in the Upper Franson Member of the Park City Formation and along the ledges and cliffs of the other remaining sandstone units. The joints trend in northwest-southeast direction (see Figure 34). This fits with the data provided by Untermann and Untermann (1969) as they studied joints within the Uinta Mountains and concluded that the two prevailing trends are in a north-south and northwest-southeast direction. Near the Deep Creek fault zone, the joints parallel the dominant northwest-southeast fault directions. Jointing can best be seen in aerial photos within the Upper Glen Canyon Sandstone on the southwest flank of Red Mountain and on the Confluence dome and also within the Gartra Member of the Chinle Formation on both sides of Ashley Creek and on Red Mountain.
The most significant economic resource found within the map area is strip mining the highly-concentrated phosphorus beds of the Meade Peak Member of the Park City Formation (Smith et al, 1952). Other useful, but less significant economic resources that have been identified in the map area are: groundwater, building and landscaping stone, coal, sand and gravel, and tourist-curio materials. The Steinaker Reservoir Quadrangle also contributes economically from camping, tourism, and water-related recreation at Steinaker State Park. It is interesting to note that, besides Steinaker and nearby Red Fleet Reservoirs, another dam is proposed on Spring Creek just before it empties into Ashley Creek. The potential for petroleum reserves has also been explored in the area in the past (evident from seismic acquisition clear-cuts); there are currently no producing petroleum wells in the area. Economically-speaking, it might be worthy to note that the subject of ‘Spanish Gold’ and abandoned ‘Spanish Gold Mines’ are a hot-topic in the annals of urban legends among locals, but these, unfortunately, do not contribute to the current economy and likely will not in the future.

Strip mining of the Meade Peak Member of the Park City Formation (Phosphoria Formation equivalent) has been going on in the hills above Vernal, for nearly 46 years (Figure 27) (mine administrator, personal communication). After the sand and gravel are winnowed out, the phosphorus-rich slurry (30% in solution) is then piped to a refinery in Rock Springs, Wyoming (pamphlet, SF Phosphates
Limited Company). From there, the phosphorus is sold and then used in many
different products and applications. On the maps, the Meade Peak Member of the
Park City Formation has been mapped separately from the Park City Formation
because of its economic implications. In fact, there is a debate as to whether or not
another mine (proposed by the Ashley Creek Phosphate Company in 2002) should be
created along the north slopes of the Steinaker Reservoir Quadrangle above Spring
Creek (Doug Sprinkel, personal communication).

Ground Water

Aside from good aquifers like the Glen Canyon Sandstone and Weber
Sandstone, there are many springs within both quadrangles. These springs issue from
joints or fractures, faults, lithology changes under alluvium, complex changes within
landslide deposits (likely due to lithology), and under the Bishop Conglomerate along
the Gilbert Peak erosion surface (Figure 28). The higher elevation springs are most
usually used to water range cattle. The accessible, low elevation springs are used for
culinary and irrigation water. The water is good quality (largely free from
contaminants); the discharge fluctuates seasonally and, over the past years, many
springs have been slowing down or drying up due to Utah’s current 5 year drought.

Coal

Coal mining was economic during the late 1800’s and early 1900’s before the
railroad was completed connecting Vernal with other cities in the region (Doelling
The coal beds occur within the Frontier Formation and were mined in Coal Mine Basin in the Dry Fork Quadrangle. The coal is of poor quality and low volume (18 tons in 3 seams “greater than 42 inches,” reported by the National Coal Resource Data System, 1970) because the coal seams are small they are no longer economic even with a potential need only a few miles away in Vernal.

**Sand and Gravel**

A few sand and gravel pits are located in the Steinaker Reservoir Quadrangle, but no operations of a similar type seem to be present within the Dry Fork Quadrangle. Near the banks of Spring Creek, where it crosses the Taylor Mountain Road, a private operation is targeting mixed red sand from the Moenkopi Formation and Glen Canyon Sandstone for landscaping and building purposes. The bulk of the gravel pits are located in the southeast corner of the Steinaker Reservoir Quadrangle, where alluvial pediment gravels are easily extracted from on top of the Mancos Shale.

**Building /Landscaping Stone**

One of the common types of decorative stone used in the region is from the Frontier Formation—the “cannonballs.” Because of their sphericity, the “cannonballs” are popular and can range from melon-size to over 6 feet in diameter. The sub-rounded, purple Uinta Mountain Group quartzite cobbles and boulders have also been used as decorative stone on outside walls of buildings and structures.
Tourist-curio materials

These types of materials include petrified wood, fossils, dinosaur bones, and ironstone sheets and concretions. Petrified wood is found within the Chinle Formation, the Morrison Formation, and the Dakota Sandstone. The largest specimens are found within the Morrison Formation (as in-place ‘stumps’) while the smallest (chips) are found within the Chinle Formation, except the Gartra Member, where logs of up to 4 feet in length can be found. Iron concretions (small versions of the somewhat famous “Moqui Marbles” stones of Southern Utah and Northern Arizona, which are believed by Native Americans to possess supernatural powers) are found in many places within the Upper Glen Canyon Sandstone. These are created by hydrothermal fluids and they litter the ground in some areas (Doelling, 1968). (Some ant hills are even protected by a mantle of BB-sized iron concretions.) Sheets of this material can be cut and sold to tourists as “moonstone” which has been said to have a fluorescent glow when viewed under a black light (Suggawa Sunnavi, person communication). The popular fossils found in the area are belemnites, dinosaur bone and tracks, and many more. State and Federal Laws protect many of these.

Geologic Hazards

Mass movement is the principal geologic hazard in the map area—especially the Dry Fork Quadrangle (Kowallis and Bradfield, 2005). Other geologic hazards would include swelling clays and soils in many of the muddy and shaly geologic units (notably the Chinle Formation, Morrison Formation, and even the Mancos Shale), ground shaking, and flash flooding in the narrow canyons at the base of Red
Ground shaking within the area would be less likely attributed to fault activity than it would be to the blasting that goes on at the Phosphate Mine. Although many faults exist within the quadrangles, these have not been active since the pre-Oligocene, 37 million years ago (see discussion chapter 2), and are not thought to be a hazard under present tectonic conditions (see the University of Utah Seismic Stations website for historical earthquakes: http://www.quake.utah.edu/). The hazards related to ground shaking are minimal, but rockfall is one potential geologic hazard due to its effects.

The landslide deposits within the Steinaker Reservoir Quadrangle are small and few and, though there is still potential that other events may occur, the hazard is deemed minimal compared to that of the landslide potential within the Dry Fork Quadrangle. There, several mass movement deposits ranging from slumps to mudslides to large debris flows like those that originate from the Bishop Conglomerate exist. Most of the landslides are thought to be old and occurred during wetter climate cycles, but the Bishop Conglomerate is still perched over soft, shale and clay deposits like the Mancos Shale, Morrison Formation, and Dinwoody Formation on both Little Mountain and on the northern slopes. If conditions were again favorable, the potential for the occurrence of a massive landslide would increase as would the inherent dangers, especially since the landslides have reached the valley floors in the past, near the settlement of Dry Fork, for example.
Geologic Scenic Attractions

The Dry Fork and Steinaker Reservoir 7.5’ Quadrangles could be considered a ‘jewel’ in the sagebrush-covered Uinta Basin. Geologic units within and adjacent to these quadrangles, are very similar, the same in some cases, to the geologic units that give Southern and Central Utah such acclaim. The scenery is no less colorful and could be considered a window into, or, a condensed version of the Colorado Plateau landscape. There exist a handful of arches in the area, known less by the general public than by locals. Dry Fork Canyon and Ashley Gorge are deeply incised (up to 1,600 feet) canyons within the Weber Sandstone. Similarly, the slot canyons in the Upper Glen Canyon Sandstone (cut as much as 200 feet) at the base of Red Mountain are no less spectacular. Petroglyphs found throughout the area give an even more satisfying flavor to the scenery in which they are chiseled.

Chapter 2: Kinematic Structural Analysis of the Deep Creek Fault Zone, Near Vernal, Utah

Introduction

No previous detailed studies exist for two Uinta Mountain area fault zones—the Deep Creek and the South Flank fault zones. The Deep Creek fault zone, named by Untermann and Untermann (1969), is located on the south-central flank of the Uinta Mountains, northwest of Vernal, Utah (Figure 1). It is oriented northwest-southeast and cuts Paleozoic and Mesozoic rocks that dip south and southwest...
approximately 20º into the Uinta Basin. The northwestern extent of the fault array is approximately limited by a major Uinta Mountain fault, called the South Flank fault (Figure 1).

In an attempt to better understand the formation of the South Flank fault and the Deep Creek fault zone, several quadrangles (and their associated faults) were mapped along the south flank of the Uinta Mountains in more detail (1:24,000) than previously accomplished (Untermann and Untermann, 1968; Sprinkel, 2002). Kinematic data from fault surfaces were collected where they could be found to assist in understanding the movement history in this fault zone.

An initial hypothesis was that this array of faults was associated with the eastward termination of the South Flank fault, even though the South Flank fault continues eastward past the Deep Creek fault zone for about another ten to twenty kilometers (Figure 1). Possibly the Deep Creek fault zone and the South Flank fault zone were generated by the partitioning of strain on the upper plate of the Uinta Basin Boundary thrust. A second hypothesis is that the Deep Creek fault zone is one of diffuse deformation formed in response to a local perturbation of the regional stress field around a change in trend in the greater Uinta Mountain structure. The fault zone is situated within a subtle structural hinge, which divides the Uinta Mountains into two anticlines or domes (Billingsley, 1933; Ritzma, 1969; Hansen, 1986). Ritzma (1969) called the two domes the “west dome” and the “east dome” (Figure 1). The west dome of the Uinta Mountains runs nearly east-west and projects into the more northwest-southeast trending east dome north of the Deep Creek fault zone. South of the fault zone, there exists a similar hinge where the subsurface Uinta Basin
Boundary thrust moves from a proximal position with respect to the west dome anticline to an increasingly distal position to the south of the east dome. The Deep Creek fault zone exists where the east dome and the basal thrust (name changed to the Asphalt Ridge fault) diverge (Figure 1). In contrast to the faulting found in the Deep Creek fault zone, the region between the Asphalt Ridge thrust fault and the east dome of the Uinta Mountains is relatively free of faults (at least at the surface) and deformation in this area is accommodated by folding (discussed briefly in Chapter 1).

**Background and Previous Work**

The Uinta Mountains have received considerable attention because of their unusual orientation. They are a Laramide orogenic structure, but trend nearly perpendicular to the more typical north-south trending Laramide uplifts and basins of Colorado and Wyoming. Although there has long been debate surrounding the question of how the Uinta Mountains were uplifted (Sales, 1969; Gries, 1983; Bradley, 1995), recent publications suggest that basin inversion, which appeals to structural inheritance, may be the reason for their unusual orientation (Paulsen and Marshak, 1999; Marshak et al., 2000). These authors suggest that the thick “pod” of coherent, dense, semi-metamorphosed, Precambrian sediment (interpreted to have been deposited in an aulacogen) was brought upward along pre-existing Precambrian basin-bounding faults when placed under Laramide stresses.

The thrust faults found on the north flank of the Uinta Mountains (Figure 1) have received more attention when it comes to structural analysis (e.g., Johnston and Yin, 2001). This is likely because these faults are exposed at the surface, while their
southern counterpart, the Uinta Basin Boundary thrust fault, is buried. Deformation along the North Flank fault has been described as north-northeast left-lateral oblique thrusting (Johnston and Yin, 2001) with deformation first propagating along this northern fault and then later propagating along the southern Uinta Basin Boundary thrust (Bradley, 1995).

**South Flank Fault**

The South Flank fault trends east-west nearly the entire length of the western Uinta Mountains (Figure 1). At its western end, it disappears under valley-fill near Woodland, Utah, but it has been interpreted to extend into American Fork Canyon on the Wasatch Front, where it is truncated by the Wasatch fault (Paulson and Marshak, 1999). At its eastern end, the South Flank fault abruptly disappears north of the study area (Sprinkel, 2002). This termination is obscured by a thick covering of Tertiary and Quaternary rocks, including the Bishop Conglomerate. Though their names may suggest a relation, it is important to remember that the North Flank thrust fault is the northern bounding thrust fault of the Uinta uplift, while the South Flank fault is not its southern equivalent (the Uinta Basin Boundary thrust is the southern bounding fault).

As mentioned, little data—if any—exists on the fault zones in question. No slip data has been published on the South Flank fault and its kinematic history is essentially unknown. A number of authors have included the South Flank fault on their index maps, cross-sections, or other illustrations of the Uinta Mountains, but then only mention the fault briefly and do not explain its origin in the accompanying text (e.g., Ritzma, 1969; Hansen, 1986; Bradley, 1995). The type of faulting and
direction of offset along the South Flank fault has been illustrated differently by
different authors. Most have described it as a normal fault because, from geologic
maps of the region, the rock units south of the fault zone are younger than those on
the north (Ritzma, 1969; Bradley, 1995). Still others have portions of it mapped as a
reverse or thrust fault (Sprinkel, 2002).

Results

The Deep Creek Fault Zone

The Deep Creek fault zone is made up of a segmented array of northwest-
southeast trending faults that have created a series of half-grabens and horst-and-
graben pairs (Figure 29). The faults have semi to near vertical dips and subsequently,
in most areas produce linear map trends. These faults cut and are, therefore, younger
than the Latest Cretaceous-Early Paleocene rock units that were deformed during the
early phases of the Uinta Uplift (Bradley, 1995). They are older than the Tertiary
(Oligocene) Bishop Conglomerate, which was deposited on the Gilbert Peak erosion
surface across the faults (Figure 30). Both Deep Creek fault zone and the South
Flank fault fall within this age range and likely formed at the same time.

In order to understand the faults within the Deep Creek fault zone, each of the
faults and/or fault blocks (all first named in this document) will be described here in
detail.
**Dry Fork Fault Array**

The faults in this complex (Figure 29) generally have smaller displacements when compared to the rest of the Deep Creek fault zone. This group of faults is a set of apparently normal faults with variable apparent vertical displacement (10-400 feet) ranging from 0.3 to 4 miles in length (Figure 29). They are dissected by canyons and covered by landslide material, which hinders mapping and data collection. Dry Fork Canyon is semi-parallel to the orientation of the faults, and has exposed them in such a way that correlation from the north side of the canyon across to the south (where slopes are covered by talus and vegetation) is difficult. The faults of this group strike in two common orientations—to the west-northwest and northwest. At the surface, these faults are displacing the Weber Sandstone, Dinwoody Formation, some Moenkopi Formation, and the Park City Formation, which forms the major dip-slope in this area (see Figure 2). Within the Park City Formation, two distinctive units (the Upper and Lower Franson Member) aided greatly in mapping of the faults. Within the Dry Fork fault array, one of the main structures is the Dry Fork graben, which has an apparent vertical displacement of approximately 400 feet and is about 2.5 miles long.

**Sawtooth Ridge Graben**

The Sawtooth Ridge graben is also formed by a set of apparent normal faults. The southeast tips of the graben exist about a hundred meters or so southwest of the mouth of Dry Fork Canyon (Figure 29). Here, the Weber Sandstone is juxtaposed next to the Park City Formation. Vertical displacement increases to the northwest.
along the graben where the Upper Glen Canyon Sandstone and perhaps the Carmel Formation are faulted down against the Weber Sandstone (approximately 2,000 feet of apparent vertical offset). The faults bounding the Sawtooth Ridge graben are about 5 miles in length; the bedding between them dips consistently to the northwest and increases with increasing offset along the faults.

**Castle Cove Fault**

The best exposed fault in the Deep Creek fault zone is the Castle Cove fault (Figure 31). It is also one of the longest faults in the fault zone at over 6 miles. Previous studies have extended the Castle Cove fault to the northwest in a wash below Pine Ridge (Figure 29). In the wash, the Dinwoody or Moenkopi Formations seem to be displaced stratigraphically below the Park City Formation. Detailed mapping has shown that these Dinwoody Formation and Moenkopi Formation “rocks” are deposits of Quaternary alluvium derived from those formations that have not been transported far from their source. Instead, the fault plane parallels the edge of Pine Ridge where the Gartra Member of the Chinle Formation forms cliffs to the northeast.

Evidence for this conclusion comes from intra-formation relationships within the Moenkopi Formation. Tracing a prominent gypsum marker-bed within the Moenkopi Formation shows that it is shortened (cut-out) with respect to the Gartra Member of the Chinle Formation (Figure 32). Near Castle Cove, on the southernmost edge of Pine Ridge, the thickness of the Moenkopi Formation between the Dinwoody Formation gypsums at the base of the slope and the Gartra Member of the Chinle
Formation is approximately 300 feet compared to the average thickness of about 970 feet. Along the fault to the northwest, the true thickness of the Moenkopi Formation can be seen where it is faulted against the Dinwoody Formation and the Gartra Member of the Chinle Formation. The maximum apparent vertical displacement for the Castle Cove fault is 1,200 feet. The largest offset seems to be in Castle Cove and lessens along the fault in both directions. The northwest tip of this fault starts to taper out when it is truncated by the Sawtooth Ridge graben, while the southern tip is covered by pediment gravels, but may extend across the Ashley Creek where a small set of strike-slip or oblique-slip faults exist in the southwest corner of the Steinaker Reservoir quadrangle (Figure 29).

Deep Creek Graben

Another graben, much like the Sawtooth Ridge graben, lies further to the southwest. But unlike the Sawtooth Ridge graben where the faulted units have a uniform dip direction parallel to the strike of the bounding faults, the Deep Creek graben (Figure 29) has faulted beds with multiple orientations; on the northwest, the direction of strike parallels the trend of the bounding faults, whereas, on the southeast portion of the graben at the base of Little Mountain, the beds gradually change their strike orientation perpendicular to the bounding faults (perhaps suggesting a period of folding previous to the faulting). In the northwest portion of the graben, on the north-bounding fault, the Upper Glen Canyon Sandstone is juxtaposed next to the Gartra Member of the Chinle Formation (apparent vertical offset of about 900 feet). This north-bounding fault is the longest fault in the Deep Creek fault zone at over 11 miles
long. The Deep Creek graben’s maximum displacement is located directly across the Deep Creek graben from there, on the south-bounding fault, where the Park City Formation is juxtaposed against the Redwater Member of the Stump Formation (vertical offset of about 2,750 feet). This fault is approximately 5 miles long before it merges with the Mosby Creek fault on the northwest. The displacement of this fault lessens to the southeast, eventually placing the Cedar Mountain Formation or Dakota Sandstone next to the Lower Glen Canyon Sandstone—an apparent vertical offset of 2,300 feet.

The north-bounding fault reappears on the east side of Little Mountain, but because of extensive landslide cover, the south-bounding fault cannot be traced out the east side. The Bishop Conglomerate caps Little Mountain further to the southeast, covering the graben. It is under this cover where the displacement on the north-bounding fault likely increases and reaches a maximum. It may be that it is because of this graben that the Bishop Conglomerate has been thickened over Little Mountain—capturing the coarse material in weaker, less robust units like the Mancos Shale.

**West Little Mountain Fault**

The West Little Mountain fault (Figure 29) is buried along much of its trace. Nonetheless, mapping this fault has not been difficult because quite a bit of offset exists along its trace (approximately 4 miles long). Offset to the northwest disappears near Crow Creek in the Lake Mountain Quadrangle; the southeastern tip is covered by landslide debris from Little Mountain. On the west side of Little Mountain, the
fault places Cedar Mountain Formation on the north next to the Redwater Member of the Stump Formation on the south. The apparent vertical displacement there (approximately 780 feet) is not necessarily the maximum since most of the fault trace is concealed.

**Chalk Cliffs Fault**

The Chalk Cliffs fault (Figure 29) places the Upper Glen Canyon Sandstone next to the Gartra Member of the Chinle Formation on the northeast (apparent vertical displacement of approximately 900 feet). The extent of this fault is not known because, like many other faults in the Deep Creek fault zone, it is concealed by the Bishop Conglomerate on the northwest. The mappable extent of the fault is 2 miles. Quite a few exposures with slickenlines were found along this fault, but only few appeared to be in place. The few available data points have left-lateral slip directions at approximately 15° from horizontal.

**Mosby Creek Fault**

The Mosby Creek fault (Figure 29) is likely an extension of the Deep Creek graben’s south-bounding fault. It is approximately 4.5 miles long and has an apparent vertical displacement of 950 feet that can be estimated at near Sprouse Spring. There, the Upper Glen Canyon Sandstone is displaced on the south of the fault plane next to the Moenkopi Formation on the north. The Mosby Creek fault is a north-bounding fault of an unnamed graben. This graben was not named in lieu of naming its
bounding faults (the Mosby Creek fault and the Smelter Spring fault) because, to the southeast, the graben is dissected by several other faults.

**Smelter Creek Fault**

The Smelter Creek fault (Figure 29) is 5 miles long and its trace includes a complex set of fault junctions. It may be related to the West Little Mountain fault to the southeast because it seems to project northwestward from where the West Little Mountain fault has been terminated at its northern extent. This fault and the other faults connected to it are quite similar to the Dry Fork Canyon fault array. Both have secondary fault trends that connect but rarely cut through the primary trend. The Smelter Creek fault dissects and rotates fault blocks made up of the Moenkopi Formation and capped by the Gartra Member of the Chinle Formation. The Upper Glen Canyon Sandstone has an apparent displacement of approximately 2,000 feet on the north of the fault plane next to the Weber Sandstone on the south, along the road paralleling Mosby Creek in the northwestern part of the Lake Mountain Quadrangle. There, exists one of the best slickensided surfaces found anywhere in the Deep Creek fault zone (referenced in Figure 35). It is interesting to note that almost equal numbers of both near vertical and sub-horizontal slickenline rakes were recorded at this locality. These data suggest the possibility that this fault surface has recorded two periods of offset.
Kinematic Data

Preservation of kinematic data is generally poor along most of the faults in the Deep Creek fault zone. The poor preservation of slickenlines is probably related to several factors: (1) exposure to weathering, (2) fluid alteration along faults (Figure 33), and (3) extensive Quaternary cover. In many places slickensided rocks exist as float, but ‘in place’ slickensided surfaces are rare. Nonetheless, we were able to collect data from a few sites and believe that they do give a better picture of the sense of movement across these faults.

The majority of slip indicators, where present, exist as slickensided surfaces, where Riedel shear fractures indicate the slip direction. Other indicators were occasionally present in a few localities. At some sites along the fault zones, slip indicators were rare and perhaps unreliable. These were not published here. Although, when compared with the more robust sites, they do seem to correlate in a general sense.

Although only a few faults have slip indicators, joints are common across the whole study area and strike consistently northwest-southeast (Figure 34)—essentially in the same direction as the main trend of the faults making up the Deep Creek fault zone.

Three sites within the Deep Creek fault zone and one along the South Flank fault produced usable slickenline data (Figure 35). Two of the localities were located along or near the north-bounding fault of the Sawtooth Ridge graben (see Figure 29 for localities). One locality was recorded within a small set of surfaces antithetic to the main fault plane existing where the Gartra Member of the Chinle Formation is
juxtaposed next to the Weber Sandstone. The second locality is completely within
the Weber Sandstone on a fault which is antithetic to and near the southern tip of the
graben. The data show that the movement along this fault is largely strike-slip, with
rakes on average up to 30 degrees from horizontal (Figure 35). In the Lake Mountain
quadrangle immediately west of the Mosby Mountain Road, a surface of Weber
Sandstone along Mosby Creek, as stated earlier in the Smelter Creek fault description,
contained a good collection of slickenlines (Figure 35). Interestingly, along this fault
the slickenlines are somewhat bimodal with some showing essentially normal dip-slip
motion and others near strike-slip motion. Although a single slickenlined surface that
showed both directions was not found at this locality, the data suggest that perhaps
this fault experienced more than one period of movement under quite different stress
fields.

Another good site exists along the South Flank fault near Ice Cave Peak (see
Figure 1 for the sample location). There, the kinematic data were similar to those
found along the Smelter Creek fault in the sense that two directions of motion are
indicated; one mainly normal dip-slip and the other strike-slip (Figure 36). In
addition, at this locality, one slickenlined surface showed both directions together and
it appeared that the dip-slip lines were older and had the strike-slip or oblique-slip
lines superimposed over them.

Discussion

Several factors indicate that the Deep Creek fault zone is likely a fold
accommodation fault zone situated above a neutral surface on the back limb of the
folded rocks associated with the Uinta Basin Boundary thrust. It may be that the Deep Creek fault zone is a localized fault zone along the southern flank of the Uinta Mountains. However, little or no published evidence exists as to the shape and origination of the main basin bounding thrust.

The joints suggest a maximum direction of stress parallel to and a minimum direction of stress perpendicular to the faults. This orientation does not seem to correlate with the overall southwest-northeast to east-west principal stress direction of the Laramide Orogeny (Gries, 1983, Johnston and Yin, 2001, among others) and may represent a local rotation of stress along the south flank of the Uinta Mountains. We are unsure whether or not the joints or the faults formed first in this area.

Because the faults trend in the general orientation of the joints, there is a possibility that they may be related. It is possible that the joints formed first, and then the faults formed using the existing weaknesses caused by the joints. On the other hand, the faults (and folds) could have formed first and the joints formed later during uplift in a different stress field. Xiaofen et al (1989) and Kowallis et al (1995) (see also Zoback et al (1981) and Best (1988)) show that the regional stress orientations determined from quartz microcracks in several plutons in western and central Utah had migrated principally from an east-west orientation to that of a northwest-southeast orientation during an interval between approximately 30 Ma and 14 Ma, respectively. This data suggests that a possible change in stress orientations may have been likely sometime after fault deformation.

Kinematic evidence suggests that oblique slip (nearly strike-slip) displacement occurred along both the South Flank fault and within the Deep Creek fault zone.
At our locality along the South Flank fault and at a few locations within the Deep Creek fault zone, it appears that at least two phases of deformation took place. At the Ice Cave Peak locality, at least two different surfaces showed a set of oblique- or strike-slip indicators overprinting a set of dip-slip indicators. We propose that an initial period of normal faulting was followed by some strike-slip motion.

First, during the initial stages of deformation, the maximum principal strain orientation was approximately NW-SE near and within the study area. During this time, extension was significant enough to create offset producing normal faults. On a regional scale, these extension may have been caused by the stretching of the Mesozoic and Paleozoic cover (perhaps even including some of the Precambrian Uinta Mountain Group) over propagating tip of the subsurface Uinta Basin Boundary thrust (localized fold accommodation faulting).

An analogous example of extension over propagation of the tip of a subsurface thrust fault can be found in the Owl Creek Uplift, a Laramide uplift located in Northern Wyoming. Wise (1963) described a zone of extension near the crest of the flanks of the Owl Creek Uplift, which he indicated likely represents a keystone graben feature (see Figure 37). Wise also showed that part of the Owl Creek Uplift had subsided over less competent lithological units (Mesozoic-aged) underlying the detachment. In the Uinta Mountains, the South Flank fault may or may not have been created as a result of similar subsidence of dense basement rocks superposed over weaker Paleozoic, Mesozoic, and Tertiary units (analogous to the Boysen fault labeled in Figure 37). Even though the current thrust models for
Laramide structures—specifically the nature of the thrust fault in Figure 37—have changed, this does not affect the fact that extension can and does occur above any particular Laramide-style uplift (providing the right local stresses).

This type of deformation has been described in laboratory experiments (Friedman et al., 1988). These experiments did not require collapse of a propagating fault tip into less competent underlying units for there to develop a zone of extension, but show that, for specific cases, a zone of extension may be created on the flanks of any particular “uplift” superposed over a major propagating thrust fault. In fact, Friedman and others (1988) stated that “if layer-parallel extension in the upthrown block is large enough, normal faults occur to form a graben.” They also stated, “The normal faults provide the major compensatory extension required by the low angle faulting.”

It could be that the ramp angle of the Uinta Basin Boundary thrust shallows from west to east across the Deep Creek fault zone (compare cross sections, Figure 38) and that this created the stresses necessary to form the Deep Creek fault zone. This hypothesis cannot currently be tested until additional data is acquired (e.g. seismic along indicated traverse paths in Figure 1). There is little doubt that the distance between the Uinta Basin Boundary thrust and the axis of the Uinta Uplift anticline changes significantly (diverge) from west to east across the study area (see Figure 1).

Another hypothesis suggests that the Deep Creek fault zone may also represent a system of faults that served to transfer strain from the South Flank fault, mostly north and west of the Deep Creek fault zone (horse-tail splay termination), to
the folded, but unfaulted rocks to the east across a regional structural swale that divides the east dome of the Uinta Mountains from the western one (zone of accommodation). Although this hypothesis could in fact be combined with the latter hypothesis, it may also be that, instead of a shallowing of the Uinta Basin Boundary thrust from west to east across the area, there exists a transfer zone between the Uinta Basin Boundary thrust and the Asphalt Ridge fault. This transfer zone might be located on the western edge of the Deep Creek fault zone and could also explain the sudden southward translation of the Asphalt Ridge fault versus that of the Uinta Basin Boundary thrust. More study of this area and these faults is warranted to test these hypotheses.

Conclusions

In the words of Donald Wise (1963):

A fault pattern produced by range uplift in the middle Rocky Mountains can be extremely complex and consist of nearly simultaneous movement of seemingly mutually exclusive fault types. In detail, however, these faults can be quite rational stress reorientations….Comparatively simple structural environments can have their systems of stresses locally reoriented to produce a chaos of seemingly unrelated but nearly synchronous fault and fold types.
The Deep Creek fault zone consists of an array of northwest-southeast trending faults in which it is likely that an episode of dip-slip deformation was followed by at least one other episode of oblique- or strike-slip deformation. The fault zone is likely a zone of fold accommodation faults related to the fold produced by the underlying Uinta Basin Boundary thrust fault. It seems that there is little doubt that the stress field in which the Deep Creek fault zone and perhaps parts of the South Flank fault zone was a localized stress field nearly perpendicular to that of the Laramide far-field stress orientation. Several working hypotheses exist as to how the Deep Creek fault zone was produced, but more work is necessary to test each of these hypotheses. It could be that a combination of some or all of the hypotheses suggested above had come into play, in some part, to create the complexity seen along the fault zones. The data collected in this study along the South Flank fault are preliminary possibilities, but they are too few to come to a concrete conclusion.

Summary

Two specific tasks were tackled in this study: completing the geologic maps and the kinematic study of the faults along the south flank of the Uinta Mountains. Though, by themselves, the only thing these projects seem to have in common is spatial (shared location), the fact is, neither could correctly and accurately be accomplished without unpaid focus on the other project.

Thoroughly detailed mapping can lead to interesting, if not exciting discoveries. When compiling a geologic map it is important to ask the right questions and make many objective observations. When one understands the processes or
reasons behind any particular observation—or to any abnormality or questionable observation, then that knowledge can be used to correctly interpret the surroundings and justly represent the “truth”, as it is believed, on paper.

Mapping the geology of the Dry Fork and Steinaker Reservoir 7.5’ Quadrangles in such a short time was very difficult, but rewarding. Completing this project any other way is hard to imagine. Many additional stratigraphic, structural, and otherwise issues had been discovered during the mapping project. Though some of them have, in the author’s opinion, been answered sufficiently, the majority have yet to be worked on conclusively.

The kinematic structural analysis of the Deep Creek and South Flank fault zones project was also quite difficult because of the veritable lack of data, on the whole. Additional data along the full length of the South Flank fault zone would likely contribute to what is understood between it and the Deep Creek fault zone as a result of this study.

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the state of stress and style of tectonism of the Basin and Range province of the
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Figure 1. Index Map of the Uinta Mountains, including the study area. (Pz - Paleozoic rocks, Mz - Mesozoic rocks.)
| FORMATION                        | SYMBOL | THICKNESS
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<td>Duchesne River Formation</td>
<td>Td</td>
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<td>Kms</td>
<td>4700 (1424)</td>
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<td>Kf</td>
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<td>Dinwoody Formation</td>
<td>Td</td>
<td>30-170 (9-52)</td>
</tr>
<tr>
<td>Park City Formation</td>
<td>Undivided</td>
<td>Pp</td>
</tr>
<tr>
<td>Meade Peak Mbr</td>
<td>Ppm</td>
<td>20 (6)</td>
</tr>
<tr>
<td>Weber Sandstone</td>
<td>Pipw</td>
<td>1015-1275 (308-386)</td>
</tr>
<tr>
<td>Morgan Formation</td>
<td>Ip</td>
<td>360-400 (109-121)</td>
</tr>
<tr>
<td>Round Valley Limestone</td>
<td>Ir</td>
<td>260 (79)</td>
</tr>
<tr>
<td>Doughnut Shale</td>
<td>Md</td>
<td>50-80 (15-24)</td>
</tr>
<tr>
<td>Humbug Formation</td>
<td>Mh</td>
<td>125-160 (38-48)</td>
</tr>
<tr>
<td>Madison Limestone</td>
<td>Mm</td>
<td>0-520 (0-158)</td>
</tr>
</tbody>
</table>

Figure 2. Stratigraphic column of geologic units exposed north of Vernal, Utah.
Figure 3. Image showing the Madison Limestone (Mm), the Humbug Formation (Mh), and the Doughnut Shale (Md); this was taken looking at exposures northwest up Dry Fork into the adjacent Lake Mountain Quadrangle. These are better exposures than crop out in the map area.
Figure 3. Image showing the Madison Limestone (Mm), the Humbug Formation (Mh), and the Doughnut Shale (Md); this was taken looking at exposures northwest up Dry Fork into the adjacent Lake Mountain Quadrangle. These are better exposures than crop out in the map area.
Figure 4. Outcrop of the Round Valley Limestone showing the durable, blocky bedding and the orange-coloring colored by the Morgan Formation deposited above.
Figure 5. Image shows the Morgan Formation, which is capped by the Weber Sandstone in the background; the lowest bed in the foreground is one of the very dense carbonate beds, the beds above consist of the typical red-colored sandstones. (P|Pw - Weber Sandstone, and I|Pm - Morgan Formation.)
Figure 6. Photo of the thick sandstones found in the area, looking NE across Dry Fork toward Ashley Gorge. (Tb - Bishop Conglomerate, PIPw - Weber Sandstone, Pp - Park City Formation, JTgu - Upper Glen Canyon Sandstone, JTrgl - Lower Glen Canyon Sandstone.)
Figure 7. Red Mountain gets its name from its characteristic red color derived from the Moenkopi Formation. Note that the Gartra Member of the Chinle Formation pinches and swells in thickness and cuts into the Moenkopi Formation. A ledge of the Lower Franson Member of the Park City Formation can also be seen. (Rcg - Gartra Member of the Chinle Formation, Rm - Moenkopi Formation, Rd - Dinwoody Formation, and Pp - Park City Formation.)
Figure 8. The Lower Glen Canyon Sandstone becomes thinner on the western side of the map area. This image shows the Lower Glen Canyon sandstone (J Rgl) deposited above the three upper-most members of the Chinle Formation: The Upper Red member (Rcu), the Ocher member (Rcm), and the Mottled member (Rcl).
Figure 9. Various reptile tracks--type A and type B (brachycheirotherium?) and plant frond (type C) impressions have been made in a carbonate deposited in the lower-most Mottled member of the Chinle Formation.
Figure 10. Suspect local angular unconformity between the Lower Glen Canyon Sandstone (J\textsuperscript{B}r\textsuperscript{gl}) and the Upper Red member of the Chinle Formation (T\textsuperscript{c}).
Figure 11. A. This image shows how resistant to weathering the Glen Canyon Sandstone oasis deposit is as it “holds-up” the Upper Glen Canyon Sandstone under the eastern portion of the Confluence dome. B. The Glen Canyon Sandstone oasis deposit may be as thick as 2 meters; here, one can see hints of the characteristic fenestral fabric in the bedding.
Figure 12. This image was taken near the Ranger Station at Steinaker State Park. It shows the Jurassic-aged Upper Glen Canyon Sandstone (JrGu), the Carmel Formation (Jc), the upper part of the Entrada Sandstone (Je), and the two members of the Stump Formation: the Curtis Member (Jsc) and the Redwater Member (Jsr).
Figure 12. This image was taken near the Ranger Station at Steinauer State Park. It shows the Jurassic-aged Upper Glen Canyon Sandstone (JRgu), the Carmel Formation (Jc), the upper part of the Entrada Sandstone (Je), and the two members of the Stump Formation: the Curtis Member (Jsc) and the Redwater Member (Jsrt).
Figure 12. This image was taken near the Ranger Station at Steinaker State Park. It shows the Jurassic-aged Upper Glen Canyon Sandstone (J†ggu), the Carmel Formation (Jc), the upper part of the Entrada Sandstone (Je), and the two members of the Stump Formation: the Curtis Member (Jsc) and the Redwater Member (Js†).
Figure 13. This image shows the bulk of Cretaceous-aged units in the map area; this was taken on the north side of the east 'racetrack' across Highway 191. (Kf - Frontier Formation, Kmo - Mowry Shale, Kd - Dakota Sandstone, Kc - Cedar Mountain Formation, and Kcb - unconfirmed Buckhorn Conglomerate bed.)
Figure 14. A. Cannonballs embedded into the top-most sandstone bed of the Frontier Formation are shown numbered; they can become much larger than shown. B. Cannonball enveloped at the base by topmost sandstone bed. C. These “concretions” are made up of a high concentration of carbonate-based lithologies.
Figure 15. This photo was taken on the western ridge of Little Mountain. A tentative contact has been picked between the Duchesne River Formation (Td) underlying the Bishop Conglomerate (Tb). The contact between the formations is gradational and was picked where sandier beds were overlain by courser-grained beds. Included on the right are two images showing more detail within the two units where the two squares are located.
Figure 15. This photo was taken on the western ridge of Little Mountain. A tentative contact has been picked between the Duchesne River Formation (Td) underlying the Bishop Conglomerate (Tb). The contact between the formations is gradational and was picked where sandier beds were overlain by courser-grained beds. Included on the right are two images showing more detail within the two units where the two squares are located.
Figure 16. A. The Bishop Conglomerate tuffaceous sandstone facies is very thick, but is not correlated well to the conglomerate facies of the same unit. B. A portion of the unnamed Pleistocene unit, which is deposited just over the top of the tuffaceous sandstone. It is peculiar that both of these units at this location are relatively fine-grained compared to their type-lithologies.
Figure 17. Very large sub-rounded boulders made up of Precambrian Uinta Mountain Group purple quartzite are removed far from their origin of deposition across relatively flat land. Rock hammer, field notebook, and Paul Jensen for scale.
Figure 18. This image shows a thick alluvial pediment, which is located northeast of Coal Mine Basin in the Dry Fork Quadrangle. This particular pediment consists of a conglomerate with a sandy matrix and is up to 15 feet thick. (*Qap* - alluvial pediment, *Je* - Entrada Sandstone, *Jc* - Carmel Formation, and *JRgu* - Upper Glen Canyon Sandstone.)
Figure 19. Slump deposit (outlined in yellow) near Highway 191 in the Steinaker Reservoir Quadrangle north of Steinaker State Park. The Mowry Shale is the unit that failed under the Frontier Formation and can be seen at the bottom of the deposit (Kmo). (Qa - alluvium deposits, Qms - slump deposit, Kf - Frontier Formation, Kmo - Mowry Shale, Kd - Dakota Sandstone, Kc - Cedar Mountain Formation, and Js - Redwater Member of the Stump Formation.)
Figure 20. Landslides originating from the higher elevations may be as large as 1.3 miles across at the head and over 3 miles long, like this one covering the western borders of Coal Mine Basin. (Qms - landslide, Tb - Bishop Conglomerate, Ku - Cretaceous units undifferentiated, Js - Stump Formation, and Je - upper part of the Entrada Sandstone, JRégu - Upper Glen Canyon.)
Figure 21. This image shows a near-axis view of the Ashley Creek syncline. The beds on the northern hinge dip near 80°, while those on the south dip only a few. (Jc - Carmel Formation, JRGu - Upper Glen Canyon Sandstone, JRLg - Lower Glen Canyon Sandstone, RCh - Chinle Formation, RCG-Gartra Member of the Chinle Formation, and RM-Moenkopi Formation).
Figure 22. Within the Kink anticline there is deformation and faulting of geologic units. This image shows the local deformation near a fault juxtaposing the Mowry Shale (Kmo) next to the shale bed within the Dakota Sandstone (Kd); there is no displacement along the fault up the hill in the Frontier Formation (Kf).
Figure 23. Simple reconstruction of the structures found within the Racetracks structural area.
Figure 24. A. Image looking southwest shows the back side of the Racetracks structural area. Usually the structures here are in anticline-syncline pairs, shown here. The steeply dipping, brown-colored beds on the south are the lower Frontier Formation, these are usually thrust over the upper portion of the Frontier Formation. (Kms - Mancos Shale, Kf - Frontier Formation, Kmo - Mowry Shale.) B. Another anticline-syncline pair; this one exposes the Mowry Shale-cored anticline.
Figure 25. Front of the ridge along Highway 191 opposite the deformation found within the Racetracks structural area. The uppermost Mowry Shale is the detachment for the structures in this area. (*Kf* - Frontier Formation, *Kmo* - Mowry Shale.)
Figure 26. A. Lower Glen Canyon Sandstone (J†gil) and the uppermost units of the Chinle Formation (Upper Red - Rcr, and Ocher - Rco) are preserved in succession and exposed on top of the younger Upper Glen Canyon Sandstone (J†gu). B. Blocks of the Lower Glen Canyon Sandstone and carbonate concretions distinctive of the Ocher unit of the Chinle Formation are found in other localities near the base of Red Mountain.
Figure 27. View of the Phosphate Mine (in background on high slope) north of Vernal. The ridge in the foreground is capped by the Redwater Member of the Stump Formation, with the Morrison Formation at the foot of the ridge.
Figure 28. Regional map showing the relationship between springs, landslides, and the Bishop Conglomerate (modified from Kowallis and Bradfield, 2005).
Figure 29. Structure map showing names and locations of many faults within the Deep Creek fault zone. Data sample localities are indicated by polygons [ ]: a, Smelter Creek fault, b, Willow Springs fault, c, Fell-off fault.
Figure 30. The relative ages of the faults in the Deep Creek fault zone are shown in this photo looking east from the western edge of the Lake Mountain Quadrangle. The Oligocene-aged Bishop Conglomerate truncates the tops of the faults on Lake Mountain and Little Mountain; the faults cut Late Cretaceous rocks. (Tb - Bishop Conglomerate, Ku - Cretaceous units undivided, Ju - Jurassic units undivided, JRGu - Upper Glen Canyon Sandstone, Rm - Moenkopi Formation, Pp - Park City Formation.)
Figure 31. Photo looking to the southeast at the Castle Cove fault trace. Here, one can see that the top of the Upper Glen Canyon Sandstone (J\textit{R}gu) is juxtaposed very near (past, even) the base of the same unit—a vertical displacement of approximately 1,200'. (\textit{Jsr} - Redwater Member of the Stump Formation, \textit{Jsc} - Curtis Member of the Stump Formation, \textit{Je} - Entrada Sandstone, \textit{Jc} - Carmel Formation, \textit{J\textit{R}gu} - Upper Glen Canyon Sandstone, \textit{J\textit{R}gl} - Lower Glen Canyon Sandstone, \textit{Rc} - Chinle Formation, \textit{Rcg} - Gartra Member of the Chinle Formation, and \textit{Rm} - Moenkoppi Formation.)
Figure 32. New mapping of the Castle Cove fault places the fault trace along the top of Pine Ridge. Truncation of a distinct bed of gypsum within the Moenkopi Formation (‘marker bed’) provides evidence supporting this new interpretation. 
(Qms - landslide debris, Rcg - Gartra Member of the Chinle Formation, Rd - Dinwoody Formation, Rm - Moenkopi Formation, Pp - Park City Formation.) Photo looking west at Pine Ridge from the eastern edge of the Dry Fork Quadrangle.
Figure 33. Typical representation of the core of a fault in the Deep Creek fault zone (map-board is approximately 2.5 feet in length). Kinematic data was difficult to obtain and record in the Deep Creek fault zone because many faults, like this one, have been altered by hydrothermal fluids—most likely by basin brines (Keith, 2002).
Figure 34. Rose Diagram showing the joint orientations taken from within the Dry Fork and Steinaker Reservoir 7.5' Quadrangles. Mean orientation is 314° +/- 8°. The red set represent the primary orientations; the blue, the secondary. The green set represent the curvy cross joints. This Rose Diagram was created using GeoOrient 9.2 (R. J. Holcombe, Dept. of Earth Sciences, University of Queensland, Queensland, Australia).
Selected Data: Deep Creek Faults

Figure 35. Fault orientations and slickenline data for select faults from the Deep Creek fault zone. Compressional (P) and extensional (T) axes are centered in the P and T dihedra solutions for several of the diagrams; they are placed at the bisector of the conjugate faults for the Smelter Creek fault diagram. (See Angelier and Mechler (1977) and Angelier (1984) for details of the P and T dihedra method.) Slickensided surfaces recorded within the field suggest that two different periods of motion occurred. The slip data seem to confirm this observation (See also Figure 36). Data for the Willow Springs and Fell-Off faults are fairly simple and show both faults to be strike-slip faults with P-axes oriented approximately E-W. The data are more complex for the Smelter Creek fault with most of the surfaces showing normal slip.
Ice Cave Peak Fault Data

Figure 36. Fault orientations and slickenline data for the Ice Cave Peak fault. Compressional (P) and extensional (T) axes are centered in the P and T dihedra solutions for several of the diagrams. (See Angelier and Mechler (1977) and Angelier (1984) for details of the P and T dihedra method.) The Ice Cave Peak data are complex, but can be divided into two subsets. The first set includes all of the strike slip surfaces which suggest right-lateral movement along surfaces oriented NE-SW. The second set includes the normal faults, which suggest near N-S extension. Slickensided surfaces recorded within the field suggest that two different periods of motion occurred. The slip data seem to confirm this observation (See also Figure 35).
Figure 37. Cross section of the Owl Creek Uplift (after Wise, 1963). This diagram may be analogous to the Uinta Uplift along the southern flank where a “zone” of extension accommodates space created by thrusting of basement rock over younger, less-competent units. The Boysen fault would be analogous to the South Flank fault. The main Owl Creek bounding thrust has been more recently interpreted as a low angle thrust (see p. 17, Fig. 14, Gries (1983).
Figure 38. Schematic cross sections comparing the ramp angle of the Uinta Basin Boundary thrust (UBB) with the occurrence of the Deep Creek fault zone (DCFZ). The South Flank fault is labeled as ‘SFF’. It may be important to remember that these cross sections were not based on actual data, but may represent a possible solution to the presence of the Deep Creek fault zone with respect to regional structures. (See Figure 1 for approximate locations of the traverses.)
DESCRIPTION OF MAP UNITS

Alluvium -- mostly coarse sand and gravel mixed with fine silt deposits are of alluvial-hyporelief type. These deposits are usually coarser than gravel with gravelly color. They are usually formed in alluvial valleys and are moderately well sorted. These deposits are usually formed in alluvial valleys and are moderately well sorted.

Morrison Formation -- consists of silt and sand. The beds are light grey, medium-grained, and form thin beds. The unit is 110-170 feet thick.

Mancos Shale -- consists of marine shale and gypsum interbedded with sandstone; shale is white and light grey, forms slopes; sandstone beds are light grey, medium-grained, micaceous, and form thin beds; unit is 180-410 feet thick.

Bipper Conglomerate -- consists of sand and silt. The beds are red, fine-grained, slope-forming; sandstone beds are red, fine-grained, slope-forming; mudstones are white and yellowish, medium-grained, friable, commonly slope-forming; mudstones are white and yellowish, medium-grained, friable, commonly slope-forming.

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