University of Nevada, Reno

Paleozoic Stratigraphy and Structure at Iron Point, Humboldt County, North-Central Nevada

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ABSTRACT

Detailed mapping and reconsideration of biostratigraphic data provide new insights into how the Comus Formation at its type locality at Iron Point, Humboldt County, Nevada fits into the regional stratigraphic framework. The age designation of the Comus Formation was reevaluated by this study using the most current understanding of Ordovician graptolite biostratigraphy. Previous studies at Iron Point had determined that the graptolites found in the siltstone units in the Comus Formation were middle Ordovician. This study determined that the species of graptolites found at Iron Point had been reclassified as late Ordovician since the original biostratigraphic study had been performed.

A portion of the Vinini Formation at Iron Point was remapped in this study as the uppermost unit of the Comus Formation. Originally, this part of the Vinini Formation was mapped as a thrust “klippe” over the Comus Formation. This area originally mapped as “Vinini” is a very different rock type than the rest of the Vinini mapped at Iron Point. The lithology was much more similar to the underlying Comus Formation. Additionally, the geometry of the contact did not make sense; the thrust contact between the Vinini and the Comus was mapped as subhorizontal, however all of the strata dipped moderately to the west. The boundary between the "Vinini" and Comus Formation in this area was remapped and determined to be depositional, not structural.

The Comus Formation in its type locality at Iron Point is not correlative with the "Comus Formation" that hosts Carlin-style gold deposits to the north in the Osgood Mountains. The Comus Formation at Iron Point is a sequence of interbedded carbonate and siliciclastic rocks deposited on the continental slope during the late Ordovician. The

"Comus Formation" mapped in the Osgood Mountains is a sequence of carbonate, siliciclastic, and mafic volcanic rocks deposited on or near a carbonate seamount from the late Cambrian to late Ordovician (Hotz and Willden, 1964; Breit et al., 2005). The Comus at Iron Point and the "Comus" in the Osgood Mountains are composed of some similar types of Ordovician rocks, but their internal stratigraphy is too different to be classified as the same continuous unit.

The Comus Formation at Iron Point is here interpreted to be correlative with the late Ordovician Hanson Creek Formation. These units have similar internal stratigraphy and timing of deposition. The Hanson Creek Formation was deposited on the continental shelf during the late Ordovician, and the Comus Formation is interpreted here to be the continuation of the Hanson Creek Formation onto the continental slope.

A new unit composed of conglomerate, breccia, and a mature quartzite was identified at Iron Point underlying the Comus Formation. The quartzite portion of the unit was previously associated with the lower part of the Comus Formation, but the conglomerate and breccia were never recognized. The quartzite is composed entirely of quartz, and the conglomerate and breccia have a quartz sand matrix. The quartzite may be correlative with the middle Ordovician Eureka Quartzite. The Eureka Quartzite is the only widespread siliciclastic deposit on the continental shelf or slope during Paleozoic time. Additionally, the Eureka Quartzite underlies the Hanson Creek Formation and its correlative units in other areas in the Great Basin.

Structural analyses using the new detailed mapping yielded evidence of six different deformational events at Iron Point. Their relative ages were determined through cross-cutting relationships and comparison to deformation recorded at Edna Mountain less than a kilometer east of Iron Point. The first fold set (F_1) is west-vergent, and likely correlative to mid-Pennsylvanian folds observed at Edna Mountain (Villa, 2007; Cashman et al., 2011). F_1 folds are asymmetric, steeply inclined, and locally overturned to the west. The second fold set (F_2) records north-south contraction and is likely correlative to early Permian folds observed at Edna Mountain (Villa, 2007; Cashman et al., 2011). F_2 folds are upright, symmetrical, and trend west-southwest. The King fault is a normal fault that strikes north-south and dips east. It post-dates the first two fold sets, and has not been active since the early Permian. The Silver Coin thrust strikes east-west, places the Vinini Formation over the Comus Formation, truncates the King fault, and is not affected by the first two fold sets. The West fault strikes southeast and dips southwest. The West fault truncates the Silver Coin thrust and juxtaposes the Comus and Vinini Formations in the footwall with the Cambrian Preble Formation in the hanging wall. Finally, Iron Point is bounded on the east side by the Pumpernickel fault, a normal fault that strikes north-south and dips east. The Pumpernickel fault Eureka Quartzite and Comus Formation in the footwall and the rock unit in the hanging wall is covered by Quaternary alluvium, so is not exposed. The movement on this structure is likely related to Basin and Range faulting starting in the Miocene.

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INTRODUCTION

It is unclear how the Ordovician rocks at Iron Point, on the eastern edge of Edna Mountain, Humboldt County, Nevada (Figure 1), relate to the regional stratigraphy (Erickson and Marsh, 1974; Villa, 2007; Cashman et al., 2011). The Ordovician units are erosionally removed approximately one kilometer to the west at Edna Mountain, where Pennsylvanian rocks unconformably overlie the Cambrian Preble Formation. This puzzling relationship has impeded previous attempts at correlating the two areas (Figure 2) (Villa, 2007; Cashman et al., 2011). Furthermore, the type section of the Ordovician Comus Formation is located at Iron Point (Ferguson, 1952); however, rocks mapped as the "Comus Formation" elsewhere in the region do not match those at Iron Point (Hotz and Willden, 1964; Madden-McGuire and Marsh, 1991). The "Comus Formation" in the Osgood Mountains is the host rock for gold deposits, so its distribution and regional correlation may have economic significance.

Existing geologic maps and cross-sections at Iron Point depict some contradictory and geometrically impossible relationships (Erickson and Marsh, 1974a). It is also unclear how the mapped structures relate to the regional tectonic history. A thrust fault juxtaposes two Ordovician units, the Comus and Vinini formations, at Iron Point. The timing of this thrust is unknown, and it has not been correlated with other well-known thrust faults in the region. A subsequent high-angle fault of unknown age and extent separates Iron Point and Edna Mountain (Erickson and Marsh, 1974a; 1974b). Paleozoic stratigraphy and structures have been well-documented on the western side of Edna Mountain (Villa, 2007; Cashman et al., 2011). These are inconsistent with the mapped structure and stratigraphy at Iron Point, which requires a significant strike-slip fault or other major structure between the two areas. Additionally, the Comus Formation at Iron Point is interpreted to be in the hanging wall of the Roberts Mountains thrust (Erickson and Marsh, 1974a), but to the northwest in the Osgood Mountains, it is interpreted to be in the footwall (Hotz and Willden, 1964) (Figure 1).

The purpose of this study is two-fold. First, detailed description of the stratigraphy, including a measured section and petrographic analyses, is used to resolve confusion regarding regional correlation of the type Comus Formation at Iron Point. Second, detailed structural mapping and kinematic analyses is used to resolve the structural inconsistencies between Iron Point and Edna Mountain, and to determine how Iron Point fits into the regional tectonic history.

This thesis begins with a brief outline of the regional Paleozoic tectonic history and regional Ordovician stratigraphy as they relate to what is mapped at Iron Point. The internal stratigraphy of the Comus Formation, including new subunit names, at Iron Point is described in detail. The structures observed at Iron Point are described, followed by a detailed deformation history at Iron Point based on cross-cutting relationships and the tectonic history that is well-documented at Edna Mountain. The relationship of the stratigraphy and structures at Iron Point to those at Edna Mountain are discussed, followed by a conclusion section that outlines the contributions this study has made to the understanding of the regional stratigraphy and tectonic history.

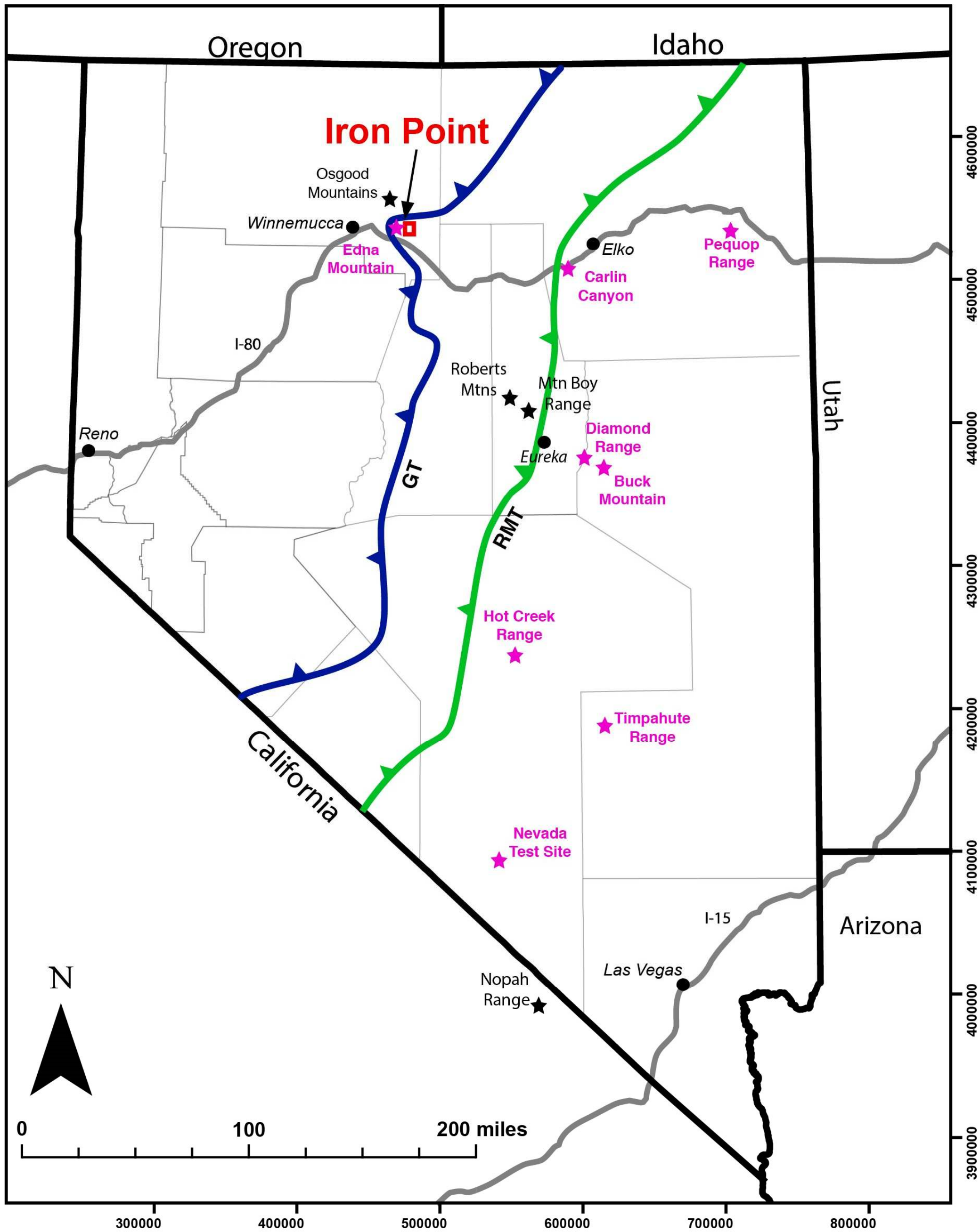


Figure 1. Location Map. This map shows Iron Point and other locations mentioned in this paper. Regional thrusts and locations of late Paleozoic deformation are also represented. RMT: Roberts Mountains thrust – this line represents the easternmost extent of deformation related to the Antler Orogeny. GT: Golconda thrust – this line represents the eastern-most extent of deformation related to the Sonoma Orogeny. Location names in pink represent places where structures of late Paleozoic age (i.e., occurring between the Antler and Sonoma orogenies) have been documented. This figure is modified from Villa (2007). Map is in UTM NAD 1983 Zone 11N.

BACKGROUND

Regional Paleozoic Tectonic Setting

The western margin of North America was a passive continental margin during Cambrian though late Devonian-early Mississippian time. Based on this understanding, Roberts et al. (1958) interpreted three distinct types of Paleozoic rock assemblages in Nevada: 1) an eastern assemblage characterized by carbonate rocks with minor amounts of shale and quartzite that represents deposition on a continental shelf, 2) a western assemblage characterized by clastic sedimentary rocks and chert with minor amounts of volcanic rocks that represents deposition in an ocean basin, and 3) a transitional assemblage characterized by carbonate, clastic, and volcanic rocks that represents deposition on a continental slope.

The Antler Orogeny disrupted this passive margin in late Devonian-early Mississippian time. The Antler Orogeny emplaced deep marine basin sediments over slope and platform shallow marine sediments along an east-vergent thrust system regionally known as the Roberts Mountain thrust (Figure 1) (Roberts et al., 1958). East-directed thrust faults and east-vergent folds developed in the allochthonous rocks indicate east-vergent motion in northern and central Nevada (e.g. Oldow, 1984). An expansive foreland basin developed east of the Antler allochthon. The timing of the Antler Orogeny is dated based on the oldest (i.e., latest Devonian-early Mississippi) foreland basin deposits (Trexler et al., 2003). The younger age bracket is poorly constrained; but rocks known as the “Antler Overlap Assemblage” range in age from early Pennsylvanian to mid-Permian (Villa, 2007; Cashman et al., 2011).

The Permo-Triassic Sonoma Orogeny also emplaced deep marine basin sediments over slope and platform shallow marine sediments, along a southeast-vergent thrust system regionally known as the Golconda thrust (Figure 1) (e.g. Silberling and Roberts, 1962). Rocks in the upper plate of the Golconda thrust are composed of Devonian through Permian strata (e.g. Gabrielse et al., 1983). Thrust faults imbricate the allochthonous units, and folds in the upper plate of the Golconda thrust are southeast-vergent and asymmetric (Miller et al., 1984). There are no known foreland basin deposits associated with the Sonoma Orogeny (Gabrielse et al., 1983).

Although the period between the Antler and Sonoma orogenies was typically regarded as tectonically quiescent, numerous investigations throughout Nevada have documented late Paleozoic deformation that cannot be attributed to either orogenic event (e.g., Trexler et al., 2003). Late Paleozoic deformation in Nevada was first recognized by Ketner (1977) and was called the “Humboldt Orogeny” to differentiate it from the other Paleozoic orogenies. Studies in the Pequop Mountains (Sweet and Snyder, 2002), the Adobe Range (Cashman et al., 2004), Edna Mountain (Villa, 2007; Cashman et al., 2011), the Diamond Mountains (Linde, 2010), and Buckskin Mountain (Whitmore, 2011), among many others (Trexler et al., 2003 and references therein), have all yielded evidence for late Paleozoic tectonism that is neither of Antler or Sonoma age. This late Paleozoic tectonic activity is recorded by numerous unconformities that truncate faults and folds within the late Paleozoic Antler foreland basin and Antler “overlap” deposits.

The rocks of Edna Mountain and Iron Point could record deformation from all of the aforementioned Paleozoic tectonic events. The Comus Formation at Iron Point was interpreted to be in the hanging wall of the Roberts Mountains thrust (Erickson and

Marsh, 1974a) (Figure 2). East-vergent deformation attributed to the Antler Orogeny has also been described in the Osgood Mountains, north of Iron Point (Hotz and Willden, 1964) (Figure 1). Several sets of structures related to late Paleozoic tectonism have been documented less than a kilometer west of Iron Point at Edna Mountain (Villa, 2007; Cashman et al., 2011) (Figure 1). East-vergent deformation and emplacement of the Golconda allochthon, related to the Sonoma Orogeny, overprint all older structures at Edna Mountain (Villa, 2007; Cashman et al., 2011).

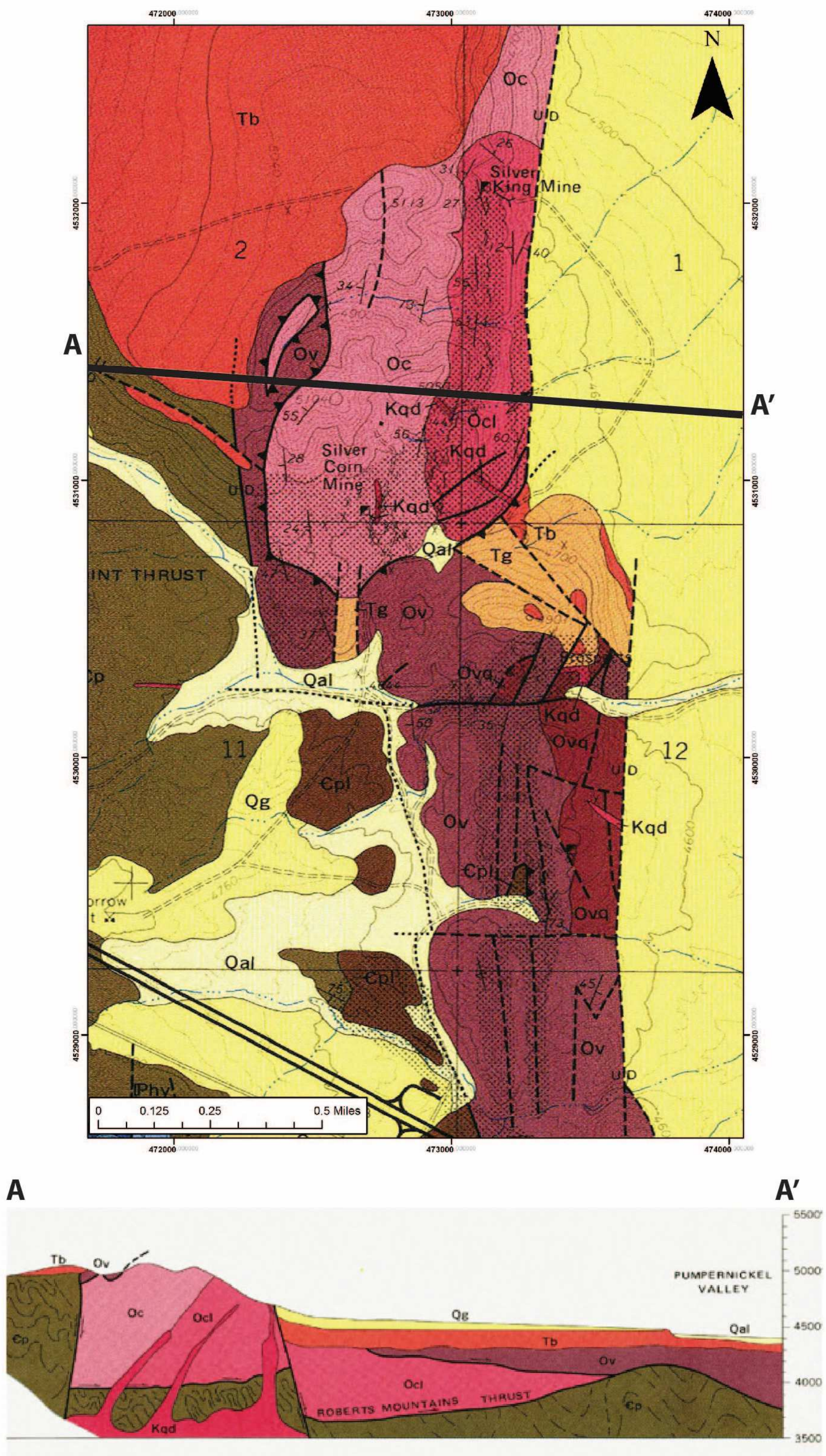


Figure 2. Original map and cross-section at Iron Point that highlights the Ordovician rocks at Iron Point depicted in the hanging wall of the Roberts Mountains thrust (Erickson and Marsh, 1974a).

Regional Ordovician Stratigraphy

The following section provides a review of regional Ordovician stratigraphy as it relates to Iron Point. The units are discussed according to where on the continental margin they were deposited, moving from the continental shelf to the basin. First, the Ordovician Eureka Quartzite and Hanson Creek Formation are described, representing continental shelf deposits. It has been suggested the Hanson Creek Formation is present at Iron Point, but it has not been mapped (Harry Cook, personal communication, 2013). The Comus Formation is described; this formation represents deposition on a continental slope. The type section of the Ordovician Comus Formation is located at Iron Point (Ferguson et al., 1952), but it is not mapped at many other locations in Nevada. The historical usage of the name “Comus Formation” and differing lithologic descriptions associated with it are discussed. The Ordovician Vinini Formation is also mapped at Iron Point, and represents deposition in a deep ocean basin. The Vinini Formation is well known and has been described by many authors throughout Nevada; a brief summary of these descriptions are provided.

Continental Shelf: Ordovician Eureka Quartzite

The Eureka Quartzite is a sequence of mature quartz arenite that was deposited across parts of southeastern California, most of Nevada, and western Utah during middle Ordovician time (Figure 3) (McBride, 2012). It is the only substantial quartzarentite interval deposited on the continental shelf from the middle Cambrian to Devonian time (Druschke, et al., 2009; McBride, 2012). It is composed of over 99 percent quartz grains, and is quartz-cemented (McBride, 2012). The quartz grains are well-rounded and well-sorted; this textural and compositional uniformity makes the Eureka Quartzite an easily identified unit in the Great Basin (Workman, 2012).

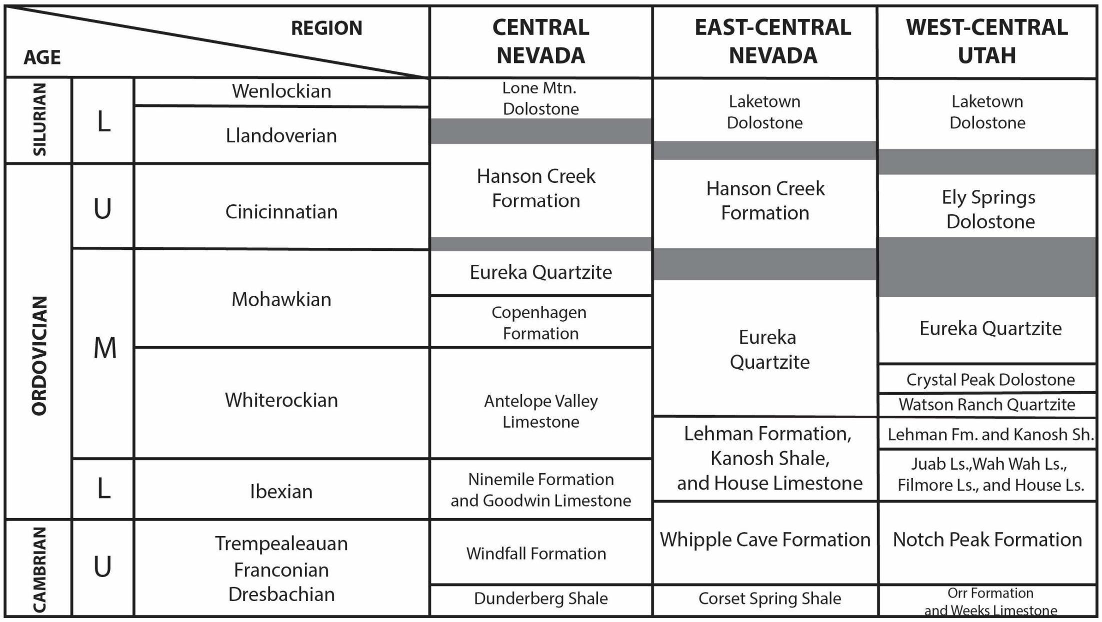


Figure 3. Regional stratigraphic correlation chart for the Eureka Quartzite and Hanson Creek Formation. Figure modified from McBride, 2012 and references therein.

The Eureka Quartzite was described in detail by Webb (1958) in the Eureka district near Eureka, NV (Figure 1). It has three distinct units starting from the lowest: 1) a red-brown cross-bedded quartz arenite with some calcareous quartz arenite and shaly beds, 2) a vitreous white quartz arenite, and 3) a dolomitic quartz arenite (Webb, 1958). In central Nevada, the Eureka Quartzite conformably overlies the middle Ordovician Copenhagen Formation, and conformably underlies the late Ordovician Hanson Creek Formation (Figure 3). There are no fossils within the Eureka Quartzite; age control is based on fossil assemblages in the underlying and overlying units (McBride, 2012).

Though the Eureka Quartzite is distinct in its compositional uniformity across the Great Basin, there is one area in southeastern California where it is particularly different. In the Nopah Range, Inyo County, California, there are breccia bodies within the Eureka Quartzite (Figure 1) (Regenfuss, et al., 1999). The breccia bodies are composed of fragments of the overlying Ely Springs Dolostone (correlative with the Hanson Creek Formation in central Nevada), and the matrix is composed of quartz sand that is compositionally identical to that which composes the Eureka Quartzite (Regenfuss et al., 1999). These breccia bodies are interpreted to be cave-fill features that formed in the Eureka Quartzite prior to silica cementation (Regenfuss et al., 1999).

Continental Shelf: Ordovician-Silurian Hanson Creek Formation

The Hanson Creek Formation is a sequence of dolomitized limestone and dolostone deposited on the continental shelf during the late Ordovician to early Silurian time (Bhatt, 1976). The Hanson Creek Formation is defined as the section of limestone and dolostone that depositionally overlies the Eureka Quartzite, and underlies the Silurian-Devonian Roberts Mountains Formation (Bhatt, 1976). It is time-equivalent to the Fish Haven dolostone and the Ely Springs Formation (Mullens and Poole, 1972; Ross et al., 1979) (Figure 3).

The type section of the Hanson Creek Formation is in Pete Hanson Creek in the Roberts Mountains (Merriam and Anderson, 1942) (Figure 1). The Hanson Creek Formation at its type section was originally split into five informal subunits (Merriam and Anderson, 1942). Starting with the lowest member, the subunits are described as follows: 1) massive, dark gray, fine-grained limestone, 2) poorly stratified limestone, 3) dark blue-gray, shaly, fossiliferous limestone that weathers light gray, 4) poorly bedded, non-crinoidal limestone that commonly contains small, black chert nodules, and 5) dark gray dolomitic limestone that contains crinoid columnals. The Hanson Creek Formation is approximately 545 feet thick at its type section (Merriam and Anderson, 1942).

Biostratigraphic dates have not been reported at the type section. However, a detailed biostratigraphic study was performed by Ross et al. (1979) in the Mountain Boy Range approximately 35 miles southeast of Pete Hanson Creek in the Roberts Mountains (Figure 1). The same lithostratigraphic and biostratigraphic sequence of the Hanson Creek Formation that is present at its type section is also present in the Mountain Boy Range (Ross et al., 1979). There, a wide array of fauna characterizes the Hanson Creek Formation, including conodonts, corals, bryzoans, brachiopods, and trilobites (Ross et al., 1979). The study by Ross et al. (1979) yielded a late Ordovician to early Silurian age for the limestone and dolostone members of the Hanson Creek Formation.

Though not present at the type section, a quartz-sand-bearing unit at the top of the Hanson Creek Formation has also been identified at various localities in Nevada (Mullens and Poole, 1972). Throughout central Nevada, the uppermost 8-30 feet (2-9 meters) of the Hanson Creek Formation contains medium to very-fine quartz grains in limestone and dolostone. The quartz content is variable; some locations have intervals of sparsely disseminated quartz sand, while in others the quartz comprises almost 75 percent of the rock (Mullens and Poole, 1972). This sandy interval has been a good stratigraphic marker for the upper part of the Hanson Creek Formation in Nevada (Mullens and Poole, 1972). This quartz-sand-bearing interval is early Silurian, based on conodont biostratigraphy from above the sandy zone in the overlying carbonate rocks (Mullens and Poole, 1972).

Continental Slope: Ordovician Comus Formation

The type section of the Comus Formation was first described at Iron Point (Ferguson et al., 1952). It has only been mapped at Iron Point and in the Osgood Mountains, north of Iron Point (Figure 4) (Ferguson et al., 1952; Hotz and Willden, 1964; Ericson and Marsh, 1974a). According to these workers, the Comus Formation is fault-bounded at Iron Point and in the Osgood Mountains, and it is unknown what formations the Comus was stratigraphically in contact with prior to faulting (Hotz and Willden, 1964; Erickson and Marsh, 1974a).

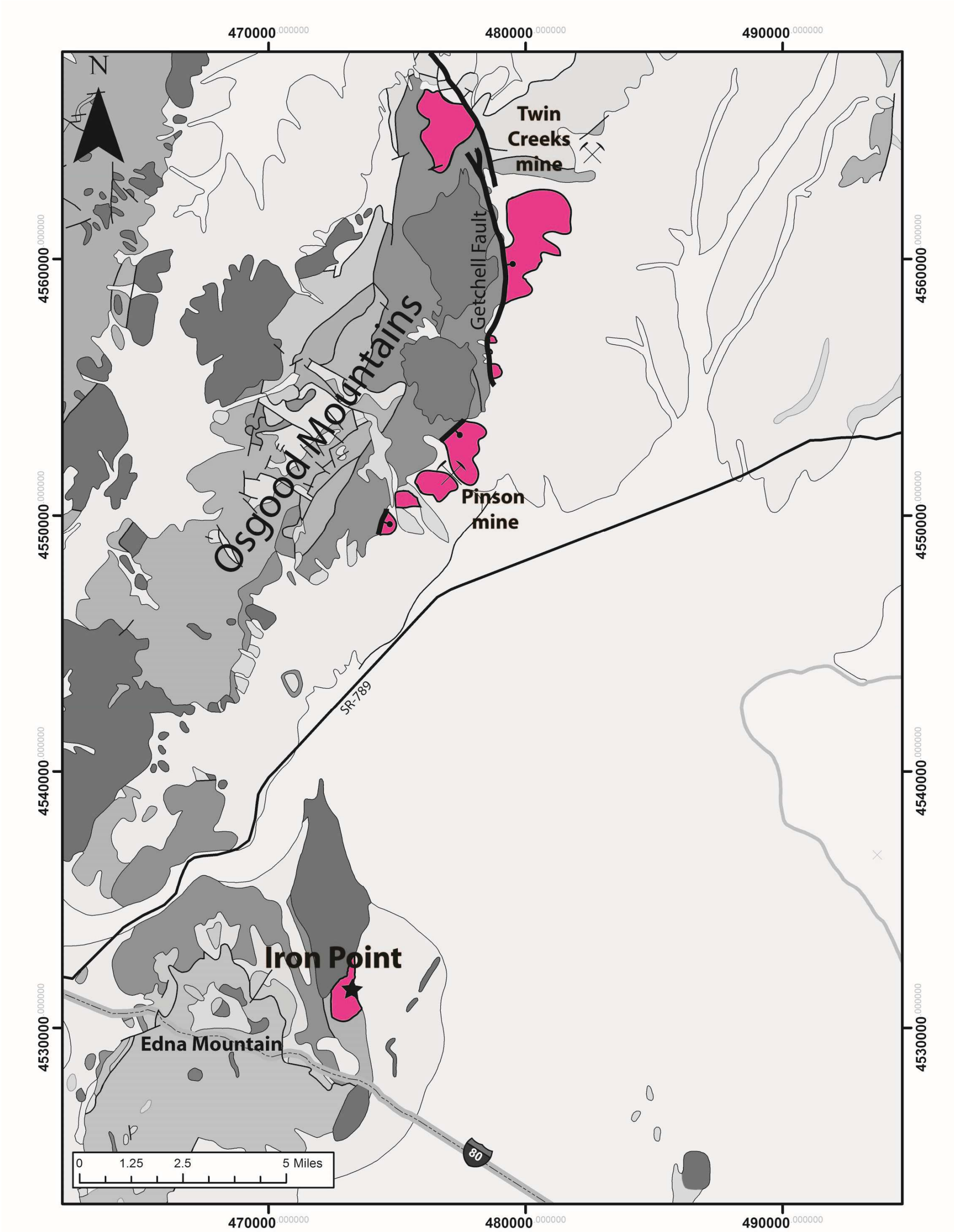


Figure 4. Simplified geologic map highlighting locations where the Comus Formation is mapped in the region. Pink polygons represent all of the locations where the Comus Formation is mapped in Nevada. Mine locations where the Comus Formation hosts gold mineralization are labeled with crossing pickax symbols. The geologic base map is a compilation of mapping completed by Hotz and Willden (1964) and Erickson and Marsh (1974a; 1974c). Map is in UTM NAD 1983 11N.

At the type section at Iron Point, the Comus Formation is composed of chert, siliceous slate, limestone, and minor quartzite (Ferguson et al., 1952). Erickson and Marsh (1974a) later split the Comus Formation at Iron Point into an upper and lower unit. The lower member contains black chert, gray dolostone, and siltstone with minor tuffaceous material. The upper member is composed of orange-weathering silty dolostone and dolomitic siltstone. The thickness of the Comus at its type section is approximately 3000 feet (900 meters) (Ferguson et al., 1952). Erickson and Marsh (1974a) used graptolites from the siltstone units at Iron Point to date the formation as middle Ordovician (unpublished USGS report, 1971). It is unclear from Erickson and Marsh's report whether the graptolites came from the "upper" or "lower" Comus.

Hotz and Willden (1964) proposed that a more complete section of the Comus Formation is exposed in the Osgood Mountains quadrangle, approximately 12 miles north. There, although the Comus Formation is mapped as one unit, the description of the Comus Formation is split into seven distinct units (Hotz and Willden, 1964). Starting with the lowermost member, the Comus Formation is composed of 1) light gray to dark greenish-gray shale and phyllite; 2) light to dark gray thin- to medium-bedded limestone with interbedded shale and shaly limestone; 3) brown-weathering, light gray to brownish-gray sandy platy dolostone with some beds of buff medium to thick-bedded dolostone; 4) medium dark gray to gray-black massive dolostone, with lenses and nodules of dark chert; 5) green to gray shale, with some fine-grained tuff and siltstone; 6) medium dark gray to gray-black massive dolostone with some chert beds, with a flat-pebble intraformational conglomerate near the top of the unit; 7) light gray to gray-brown thin to thick-bedded limestone with minor shale beds and minor lenses of brown chert (Hotz and

Willden, 1964). The exposed thickness of the Comus in the Osgood Mountains is estimated to be between 3200 and 4600 feet (975 – 1400 meters) (Hotz and Willden, 1964).

The age of the Comus Formation in the Osgood Mountains is based on early to middle Ordovician graptolites and a single upper Cambrian trilobite mold (Hotz and Willden, 1964). The graptolites were recovered from shale beds that overlie massive and thin-bedded chert, but it is unclear to which subunit these shale beds belong (Hotz and Willden, 1964). The single trilobite mold was recovered from beds that were in a different location than the graptolites, but determined to be in approximately the same stratigraphic position, or slightly lower than, the graptolites (Hotz and Willden, 1964, and references therein).

The Ordovician Comus Formation in the Osgood Mountains is thought to have been deposited on a continental slope, based on the combination of calcareous and clastic sediment (Hotz and Willden, 1964). The Comus Formation in the Osgood Mountains is considered to be a part of the "autochthonous transitional sequence" of Roberts et al. (1958) and therefore in the lower plate of the Roberts Mountain thrust (Hotz and Willden, 1964). However, Erickson and Marsh's cross-section through a portion of the Iron Point quadrangle (1974a) shows the Comus Formation there in the upper plate of the Roberts Mountain thrust.

Deep Ocean Basin: Ordovician Vinini Formation

The Ordovician Vinini Formation is one of the best-documented units within the Antler allochthon and is present at Iron Point (Erickson and Marsh, 1974a). It was first described by Merriam and Anderson (1942) at Vinini Creek, on the eastern side of the Roberts Mountains, where it crops out extensively (Figure 1). The Vinini Formation was further described in detail by Finney and Perry (1991) with a comprehensive measured section through Vinini Creek. The Vinini is approximately 9900 feet (3000 meters) thick and is split into upper and lower sections at its type locality (Finney and Perry, 1991).

The lower Vinini is characterized as black shale, quartz arenite with quartz wacke, calcareous sandstone, siltstone, shale, limestone, and rare quartzite, conglomerate, chert, and greenstone, and siltstone with quartz wacke, calcareous sandstone, and shale (Finney and Perry, 1991). Compared to the upper Vinini, the lower part of the formation contains considerable quartz arenite and calcareous sandstone. The quartz arenite and calcareous sandstone interbedded with shale and siltstone indicate turbidity flows in a submarine fan (Finney and Perry, 1991).

The upper Vinini is characterized as a thick sequence of shale, siliceous shale, and bedded chert, with a minor interval of siltstone and fine-grained sandstone at approximately the middle of the unit (Finney and Perry, 1991). The upper Vinini is composed of primarily shale and siltstone, and therefore contains the majority of the graptolite diversity within the unit (Finney and Perry, 1991). Detailed biostratigraphy of the Vinini Formation in the Roberts Mountains using graptolites indicates deposition throughout the Ordovician (Finney and Perry, 1991). The abundance of shale and siltstone suggests that the upper Vinini likely represents pelagic sedimentation in a calm, deep marine setting (Finney and Perry, 1991).

Iron Point Stratigraphy

The following section outlines the stratigraphy that has been mapped at Iron Point. The Vinini and Comus Formation are the two Ordovician units exposed at Iron Point (Erickson and Marsh, 1974a). These two units are separated by a thrust fault (Erickson and Marsh, 1974a). The Ordovician Comus Formation is overlain by Tertiary basalt at the north end of the study area. The Cambrian Preble Formation is in high-angle fault contact with both the Vinini and Comus Formations on the west boundary of the study area, however, the Preble Formation will not be of focus in this study.

The Vinini Formation at Iron Point is separated into upper and lower members based on composition (Erickson and Marsh, 1974a). Rocks that are predominantly black on their weathered surfaces characterize both the upper and lower members. The lower member is composed of fine-grained quartzite, and the upper member contains thin-bedded chert with lenses of black vanadiferous shale (Erickson and Marsh, 1974a). The upper and lower units at Iron Point are similar to the turbidite sequences that comprise the lower Vinini Formation at its type section. The stratigraphic thickness of the Vinini Formation at Iron Point cannot be measured due to intense internal deformation of the unit.

The Vinini Formation in the study area is remarkably devoid of graptolite or other fauna, as is the lower Vinini Formation at the type section. This has prevented attempts to confirm its age at Iron Point. The designation of this formation as "Vinini" was based on similar composition, color, bedding thickness, etc., to that of the type section (Erickson and Marsh, 1974b).

The Comus Formation at its type section at Iron Point is separated into an upper and lower member based on composition (Erickson and Marsh, 1974a). The lower member contains black chert, gray dolostone, and siltstone that may contain tuffaceous material or volcanic ash. The upper member is composed of silty dolostone and dolomitic siltstone, which is readily identified at Iron Point by flaggy weathering and orange to red weathered surface. The Comus Formation is approximately 2000 feet (600 meters) thick at Iron Point.

The Ordovician age designation for the Comus was based on middle Ordovician graptolites. The graptolites are rare and found in the siltstones (Madden-McGuire and Marsh, 1991). Siltstone is mapped in both the upper and lower members of the Comus Formation at Iron Point (Erickson and Marsh, 1974a).

At Iron Point, the Comus Formation contains carbonate, whereas the Vinini Formation does not, which indicates different depositional environments (Erickson and Marsh, 1974b). The relative age of the Vinini and Comus formations are unclear because there are no biostratigraphic data to confirm the age of the Vinini Formation. The thrust fault that juxtaposes these two formations was invoked to explain the similar age but different depositional environment (Erickson and Marsh, 1974b).

Iron Point Structure

The following section outlines the previously mapped structures at Iron Point. An east-west-trending thrust fault trace separates the Comus Formation to the north (footwall) and the Vinini Formation to the south (hanging wall). Iron Point is bounded on the east and west side by north-striking high-angle faults, and is covered by Tertiary basalt at the north end of the study area.

The Ordovician units at Iron Point display different styles of deformation. The Vinini Formation is isoclinally folded, whereas Comus Formation dips moderately to steeply westward, and contains a broad, north-trending fold (Erickson and Marsh, 1974b). Erickson and Marsh (1974a; 1974b) could not determine if the two distinctly different styles of deformation were related to the same event.

The Vinini and Comus formations are separated by an unnamed thrust fault at Iron Point (Erickson and Marsh, 1974a). The Vinini was thrust over the Comus, but the age and emplacement direction of this fault were unknown (Erickson and Marsh, 1974b). The deformation at Iron Point has been attributed, in part, to the late Devonian-early Mississippian Antler Orogeny (Erickson and Marsh, 1974b). The Roberts Mountain thrust has been described at Iron Point in cross section, but does not crop out at the surface (Figure 2) (Erickson and Marsh, 1974a).

Iron Point is fault-bounded on the east and west sides. The eastern side is bounded by a north-striking, steeply east-dipping normal fault that juxtaposes the Ordovician units (footwall) and Quaternary alluvium (Figure 5) (Erickson and Marsh, 1974a). The western side of Iron Point is bounded by an unusual north-striking, west-dipping high-angle fault with unknown age and displacement (Erickson and Marsh, 1974a). This fault separates Edna Mountain and Iron Point, and juxtaposes the Cambrian Preble and Comus Formations (Erickson and Marsh, 1974a).

Stratigraphy and Structure of Edna Mountain

The following section will outline the newly defined Paleozoic stratigraphy and structure of Edna Mountain, approximately 1 mile west of Iron Point (Figure 5). The age of the Paleozoic units present at Edna Mountain are Cambrian, Pennsylvanian, and Permian (Villa, 2007; Cashman et al., 2011). There is no record of deposition between early Ordovician and late Mississippian time at Edna Mountain (Villa 2007; Cashman et al., 2011). At Iron Point, only Ordovician units are mapped; it is unknown if Edna Mountain and Iron Point share any of the same stratigraphy. Late Paleozoic rocks deposited at Edna Mountain may have also been deposited at Iron Point and subsequently eroded away. Additionally, published mapping shows an unusual fault that separates the eastern side of Edna Mountain from Iron Point (Figure 5).

There are well-documented late Paleozoic structures at Edna Mountain (Villa, 2007; Cashman et al., 2011). Presence or absence of these structures at Iron Point constrains the timing of movement on the fault that separates the two areas. If Edna Mountain and Iron Point share the same structures, and therefore the same deformation history, it will mean that there has not been much movement on the fault that separates these two areas since the late Paleozoic. If Edna Mountain and Iron Point are structurally dissimilar, it will mean that the structure that separates these two areas has had significant offset since the late Paleozoic, and might be a previously unknown major regional structure.

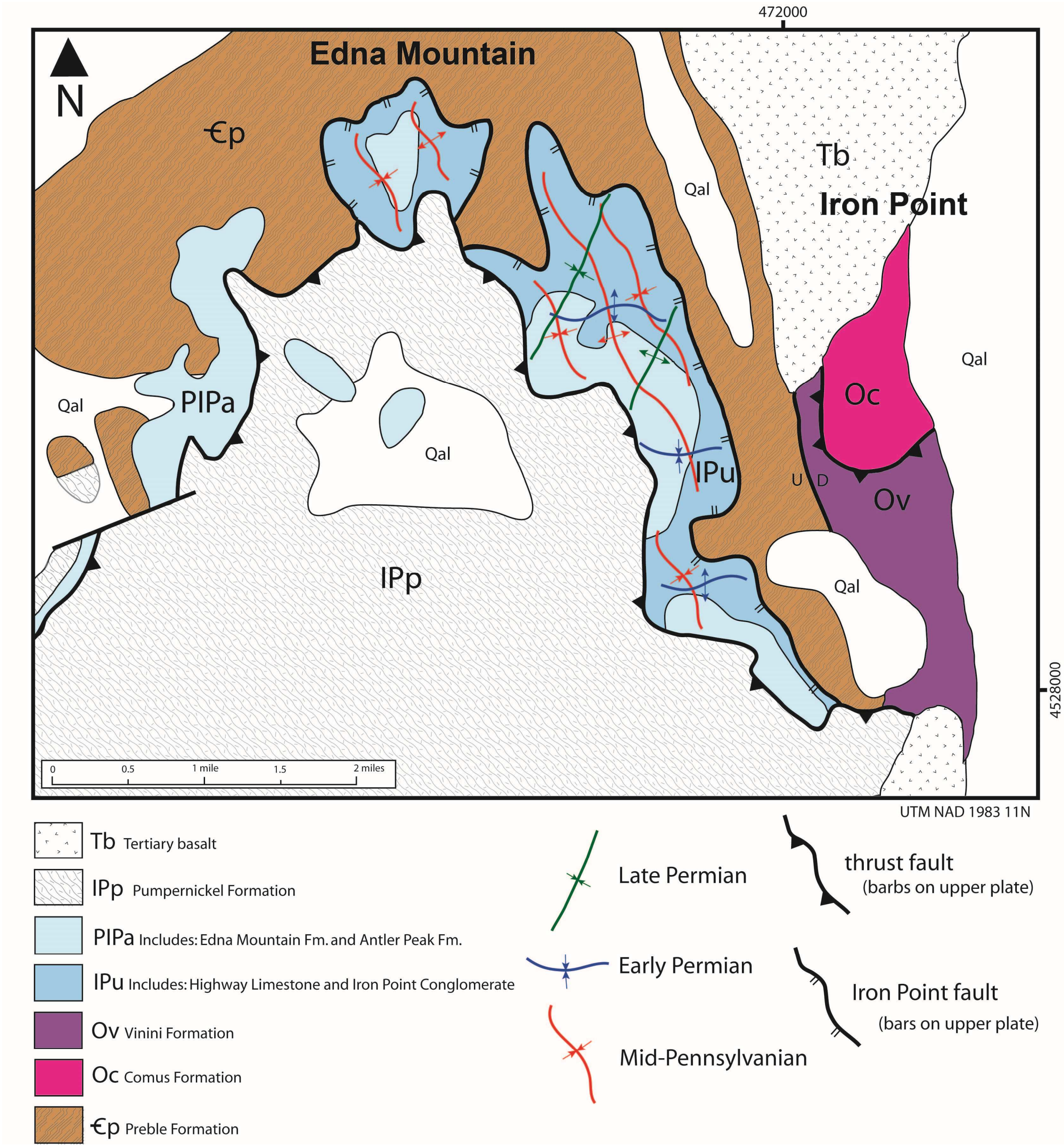


Figure 5. Simplified geologic map of Edna Mountain and Iron Point with traces of documented late Paleozoic structures (modified from Villa, 2007; Cashman et al., 2011).

Edna Mountain Stratigraphy

The Paleozoic rocks exposed at Edna Mountain range in age from Cambrian to Permian. The oldest unit exposed is the Preble Formation, which ranges in age from early Cambrian to early Ordovician (Erickson and Marsh, 1974c). It is composed primarily of thin-bedded light tan, green, and red argillite to phyllite, interbedded with limestone, fine- to medium-grained micaceous quartzite (Madden-McGuire and Marsh, 1991; Villa, 2007; Cashman et al., 2011). The Preble Formation is intensely deformed and underwent greenschist-facies metamorphism prior to deposition of the upper Paleozoic sequence at Edna Mountain (Erickson and Marsh, 1974c; Madden-McGuire, 1991).

The early Pennsylvanian Iron Point Conglomerate, part of the Antler Overlap Sequence, unconformably overlies the Preble Formation. The conglomerate is composed primarily of well-sorted, rounded, pebble- to boulder-sized clasts of quartzite, with localized sand channels (Villa, 2007; Cashman et al., 2011). The conglomerate is well stratified (Villa, 2007; Cashman et al., 2011). The Iron Point Conglomerate is ledge forming, and is distinguished from other conglomerates at Edna Mountain by its large, well-rounded clasts that are predominately quartzite (Villa, 2007; Cashman et al., 2011). The Iron Point Conglomerate was originally mapped (Erickson and Marsh, 1974c) as the mid-Pennsylvanian Battle Formation; however, a recent study at Edna Mountain recovered early Pennsylvanian fusilinids from the overlying Highway Limestone (Villa, 2007; Cashman et al., 2011). The early- to mid-Pennsylvanian Highway Limestone is in depositional contact over the Iron Point Conglomerate, thus necessitating an early Pennsylvanian or older age designation of the Iron Point Conglomerate.

The late-Pennsylvanian Highway Conglomerate overlies the Highway Limestone along a karst surface. A mid-Pennsylvanian unconformity is represented by the karst surface that separates the Highway Limestone and the Highway Conglomerate (Villa, 2007; Cashman et al., 2011). The Highway Conglomerate is composed of poorly sorted, angular clasts of silty limestone and phyllite (Villa, 2007; Cashman et al., 2011). The silty limestone clasts likely originated from the underlying Highway Limestone, based on similar color, texture, and composition (Villa, 2007; Cashman et al., 2011). The phyllite clasts are dark red and green, which indicate that the Preble Formation was the likely source (Villa, 2007; Cashman et al., 2011). The conglomerate is noticeably devoid of internal stratification (Villa, 2007; Cashman et al., 2011).

The late Pennsylvanian Antler Peak Limestone unconformably overlies the older units at Edna Mountain. Fusilinids indicate it is late Pennsylvanian (Villa, 2007; Cashman et al., 2011). The Antler Peak Limestone consists primarily of blue-gray limestone. The base of the formation is locally marked by a poorly sorted granule to pebble conglomerate. The conglomerate contains clasts of sandy limestone, limestone, chert, phyllite and quartzite (Villa, 2007; Cashman et al., 2011).

The youngest late Paleozoic unit at Edna Mountain is the mid-Permian Edna Mountain Formation. It is composed of a fine- to coarse-grained calcareous litharentite. The basal conglomerate is composed of poorly sorted, pebble- to granule-sized, angular to subangular clasts of phyllite, quartzite, chert, and limestone (Villa, 2007; Cashman et al., 2011). The Edna Mountain Formation unconformably overlies all older units (Villa, 2007; Cashman et al., 2011).

Edna Mountain Structure

Four sets of late Paleozoic structures are well documented on the western side of Edna Mountain (Villa, 2007; Cashman et al., 2011). The most pervasive structures are mid-Pennsylvanian west-southwest-vergent folds (F_1) (Villa, 2007; Cashman et al., 2011). These folds are preserved in the Iron Point Conglomerate and the Highway Limestone (Villa, 2007; Cashman et al., 2011). F_1 fold axes trend roughly 319° and are subhorizontal (Villa 2007; Cashman et al., 2011). F_1 folds are map scale and have an average wavelength of 30 feet (10 meters) (Villa 2007; Cashman et al., 2011). F_1 folds represent the first phase of Paleozoic deformation ($D_{1)}$ at Edna Mountain (Villa, 2007; Cashman et al., 2011). D_1 represents east northeast-west southwest contraction (Villa, 2007; Cashman et al., 2011).

The "Iron Point thrust" of Erickson and Marsh (1974a, 1974c) is now called the Iron Point fault. The Iron Point fault was reinterpreted to be a mid-Pennsylvanian northeast-dipping low-angle normal fault (Villa, 2007; Cashman et al., 2011). It places folded Iron Point Conglomerate and Highway Limestone in the hanging wall over the Cambrian Preble Formation in the footwall (Villa, 2007; Cashman et al., 2011). The average strike of the fault is 336° and dips ~25° NE (Villa, 2007; Cashman et al., 2011). There are no exposures of the Iron Point fault at Edna Mountain to directly measure kinematic data; the structural analyses of the Iron Point fault relied on map relationships (Villa 2007; Cashman et al., 2011). The deformational event that created the Iron Point fault is known as D_2 (Villa 2007; Cashman et al., 2011). However, there are no folds associated with the Iron Point fault, so there is no set of F_2 folds associated with D_2 (Villa 2007; Cashman et al., 2011).

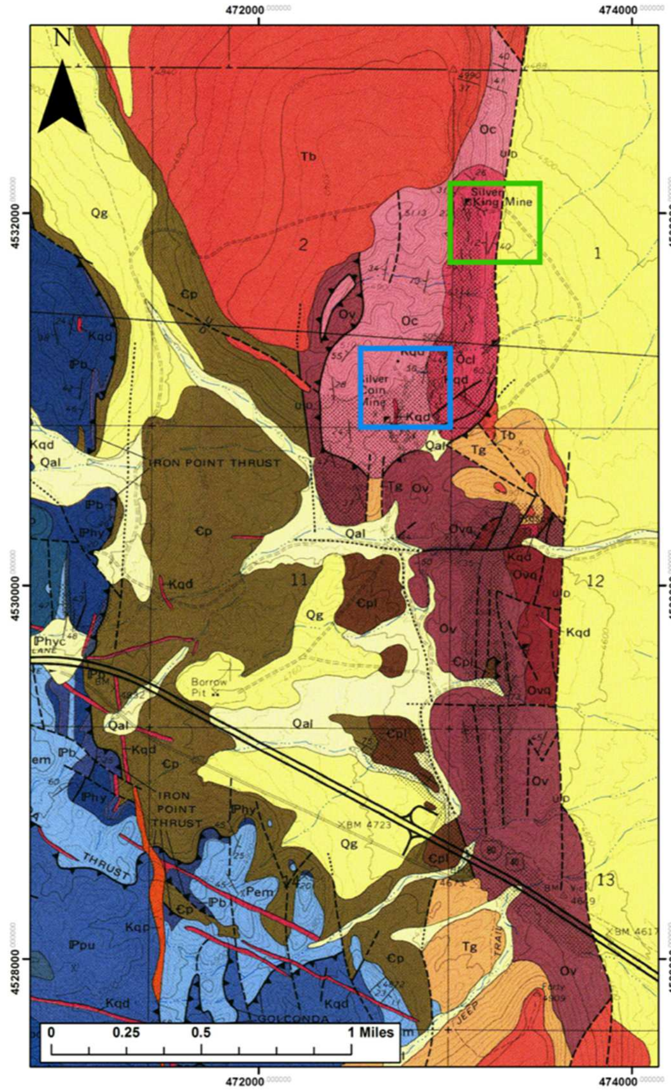
Early Permian east-trending folds, F_3, overprint all of the Pennsylvanian deformation (Villa, 2007; Cashman et al., 2011). F_3 folds are open, upright, symmetric, and have an average wavelength of 30 meters (Villa, 2007; Cashman et al., 2011). The F_3 folds overprint the F_1 folds to create a dome and basin interference pattern at Edna Mountain (Villa, 2007; Cashman et al., 2011). This third deformational event, D_3, affects the Antler Peak Limestone, the underlying Pennsylvanian units, and the Iron Point fault, but not the Permian Edna Mountain Formation (Villa, 2007; Cashman et al., 2011). The east-trending folds record north-south shortening during Permian time, a style of deformation not recognized elsewhere in Nevada (Villa, 2007; Cashman et al., 2011).

Southeast-vergent folds, F_4, affect all of the Antler overlap units at Edna Mountain and the Iron Point fault (Villa, 2007; Cashman et al., 2011). F_4 folds are asymmetric, tight to close, moderately to steeply inclined, and locally overturned to the southeast (Villa, 2007; Cashman et al., 2011). These folds trend, on average, 033° and plunge moderately, on average, 24° (Villa, 2007; Cashman et al., 2011). F_4 folds record southeast-northwest shortening (Villa, 2007; Cashman et al., 2011). This deformation, D_4, is attributed to the emplacement of the Golconda allochthon during the Sonoma Orogeny (Villa, 2007; Cashman et al., 2011).

IRON POINT STRATIGRAPHY

This section focuses on new work characterizing the two Ordovician formations that have been identified at Iron Point: the Comus and Vinini Formations. The lithologic characteristics of the Comus, Vinini, and one previously unidentified unit are described here. The descriptions and subdivisions within each formation are based on field observations, hand samples, thin section petrography, and reanalysis of published data. All data in this section are original unless otherwise noted.

The internal stratigraphy of the Comus Formation is based on one measured section (Figure 6) at the type area. The Comus Formation is well exposed in the study area. Six distinctly different subunits were identified based on differences in composition. Limited biostratigraphic data dates the Comus Formation as early to middle Ordovician (Erickson and Marsh, 1974a). A reinterpretation of the biostratigraphic dates for the Comus Formation is discussed. As mentioned earlier, the Comus Formation was previously identified as early to middle Ordovician in age (Erickson and Marsh, 1974a, 1974b; Madden-McGuire and Marsh, 1991). However, based on reinterpretation of the biostratigraphic dates, it is actually a late Ordovician unit (Figure 6).


472000
474000
4532000
4530000
4528000
N
Tb
Oc
Qg
Silver King Mine
Cp
Ov
Kqd
Ocl
Silver Coin Mine
Qal
Tg
IRON POINT THRUST
Qal
Cpl
Borrow Pit
IRON POINT THRUST
THRUST
GOLCONDA
Pem
Pb
Ppu
1
2
11
12
13
0
0.25
0.5
1 Miles

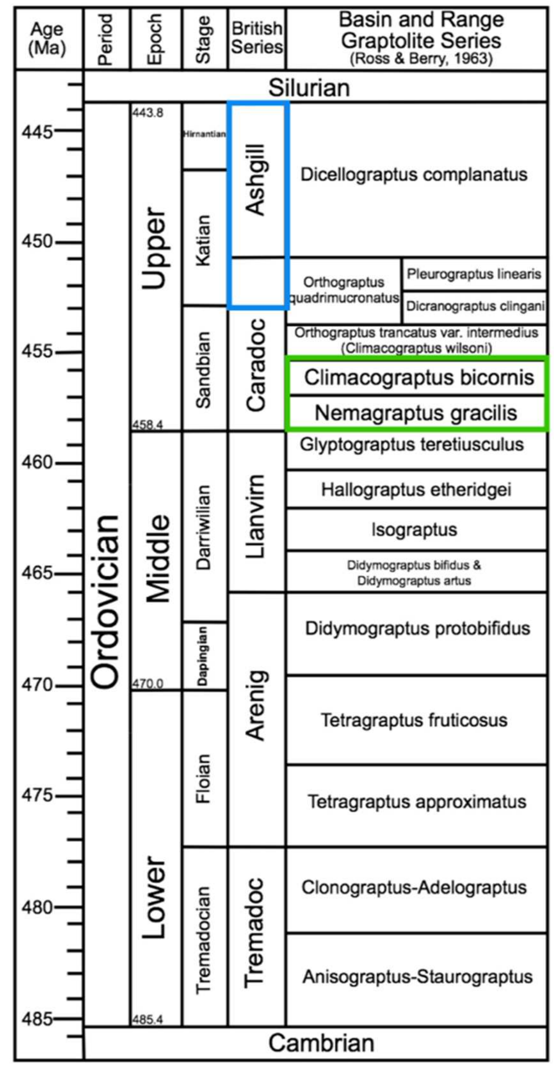

Age (Ma)
Period
Epoch
Stage
British Series
Basin and Range Graptolite Series
(Ross & Berry, 1963)
Silurian
Ordovician
Upper
Middle
Lower
443.8
458.4
470.0
485.4
445
450
455
460
465
470
475
480
485
Hirnantian
Katian
Sandbian
Darriwilian
Dapingian
Floian
Tremadocian
Ashgill
Caradoc
Llanvirn
Arenig
Tremadoc
Dicellograptus complanatus
Orthograptus quadrimucronatus
Pleurograptus linearis
Dicranograptus clingani
Orthograptus trancatus var. intermedius
(Climacograptus wilsoni)
Climacograptus bicornis
Nemagraptus gracilis
Glyptograptus teretiusculus
Hallograptus etheridgei
Isograptus
Didymograptus bifidus & Didymograptus artus
Didymograptus protobifidus
Tetragraptus fruticosus
Tetragraptus approximatus
Clonograptus-Adelograptus
Anisograptus-Staurograptus
Cambrian

Figure 6. Graptolite sample locations within the Comus Formation and the corresponding ages. The green boxes correspond to the graptolite sample collected from the subunit Ocls1 in the Comus Formation and its age. The blue boxes correspond to the graptolite sample location collected from the subunit Ocls2 in the Comus Formation and its age. Graptolite samples were analyzed by R.J. Ross and the data are contained in an unpublished USGS report (1971). Graptolite zones are modified from Ross and Berry (1963). Geologic map modified is from Erickson and Marsh (1974a). Map is in UTM NAD 1983 11N.

Location 1: Sample IP454 contains *Dicellograptus sextans* var. exilis, Dicellograptus cf. D. sextans, Glossograptus, Orthograptus? sp., possibly *O. calcaratus,* and *Diplograptus*.
Age: Best guess, zone of *Climacograptus bicornis*. May be next older zone of *Nemagraptus gracilis*.

Location 2: Sample IP427 contains *Caryocaris?*, *Orthograptus* sp. possibly *O. truncatus* var. *peretenuis*, *Climacograptus?* possibly C. *supernus*, and *Climacograptus*.
Preservation of this collection is not enough to permit conclusive identifications of any of the graptolites. Alteration of the graptolites to clay minerals has also destroyed detail. Therefore identifications and age are only a best guess.
Age: Possibly late Caradoc or Ashgill and therefore younger than IP454.

The internal stratigraphy of the Vinini Formation is then described, starting with the stratigraphically lowest units. The Vinini Formation was originally identified at Iron Point based on similar lithologic characteristics to the type section (Erickson and Marsh, 1974). Three distinctly different subunits within the Vinini Formation were identified by this study, based on changes in composition. Outcrops of the Vinini Formation at Iron Point are limited and did not permit a measured section. Where the Vinini Formation is exposed, the rocks are very altered due to abundant folding, late-stage faulting, and apparent hydrothermal alteration. No biostratigraphic data exist for the Vinini Formation at Iron Point, and none were collected in this study.

A new unit at Iron Point was also identified by this study. This unit is lithologically distinct from the Comus and Vinini formations. Field descriptions and petrographic analysis of this unit will be discussed in this section. All data in this section are original unless otherwise noted.

Ordovician Comus Formation

Ocl S_1

The stratigraphically lowest unit in the Comus Formation is Ocl S_1, dark gray siltstone (Figure 7). Ocl S_1 is a recessively weathering, dark gray siltstone. Outcrops are rare, but where it is exposed, it weathers light gray and typically displays leisegang banding. This unit has been referred to as "Nevada wonderstone" due to the pervasive leisegang banding (Madden-McGuire and Marsh, 1991). Very fine laminations are visible on weathered surfaces, however, this may be incipient cleavage development. Bedding is not readily apparent on fresh surfaces, but is visible in thin section (Figure 8). Ocl S_1 is at least 200 feet (63 meters) thick, but the unit is faulted at the lower contact and the depositional base is not exposed.

The age of graptolites in Ocl S_1 has been reclassified in this study based on work done by Finney and Perry (1991). The most notable locality for graptolite collection is at the Silver King mine, which is at the north end of the study area (Figure 6). Samples of these graptolites were collected by Erickson and Marsh (1974a), analyzed by R.J. Ross (USGS unpublished report), and determined to be of the graptolite zones *Climacograptus biocornis* or *Nemagraptus gracilis* (Caradoc age) (Figure 5). Previous age interpretation of these graptolite zones was early to middle Ordovician. The understanding of graptolite zones has been revised since the initial analysis. Based on current understanding of graptolite zones in the Basin and Range, the graptolites are actually early Late Ordovician (Carter, 1972; Finney and Perry, 1991).

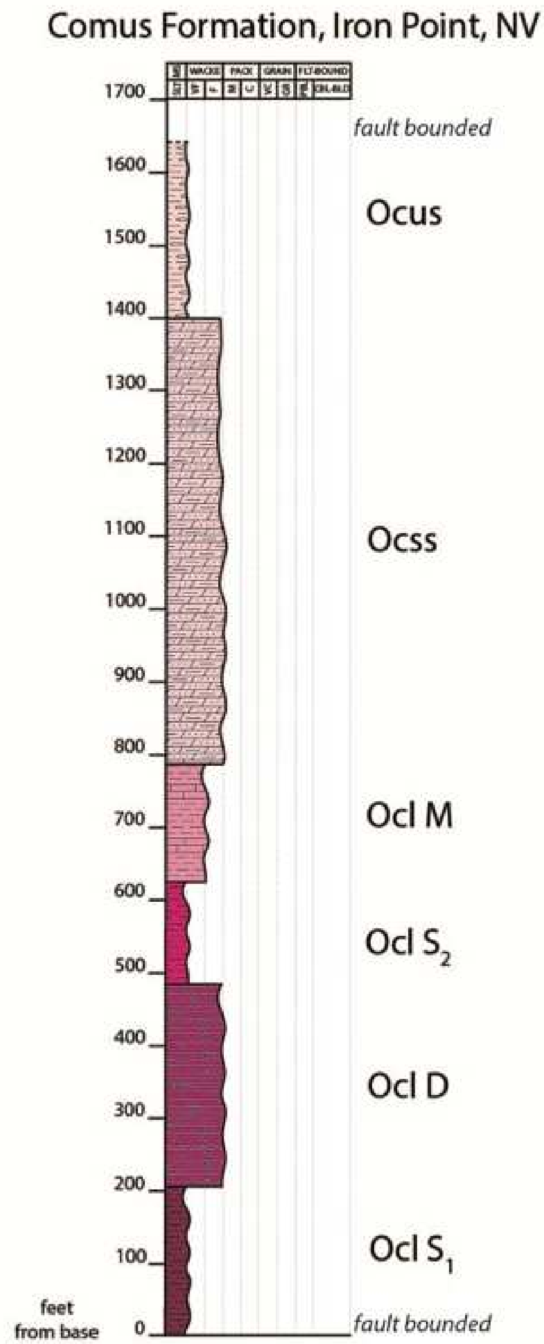


Figure 7. Stratigraphic column of the Comus Formation displaying subunits described in the text.

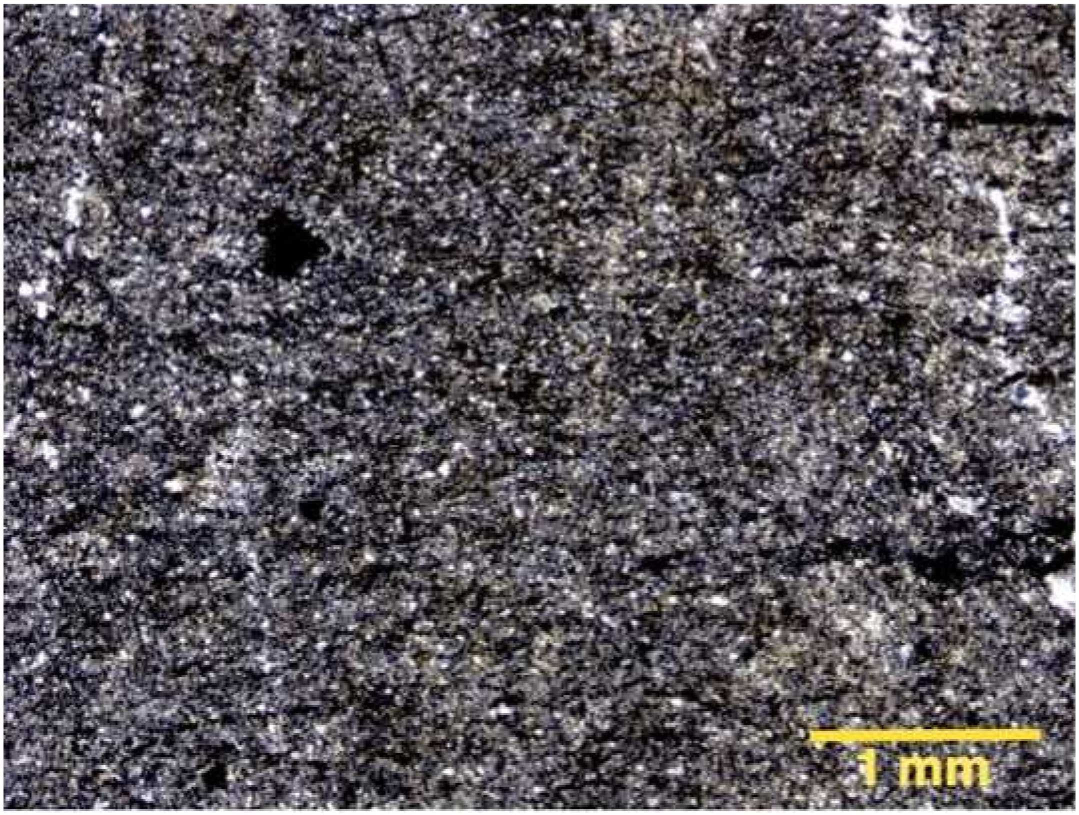


Figure 8. Cross-polarized photomicrograph of subunit Ocl S_1 within the Comus Formation. This is a siltstone composed of quartz, rare mica, and opaque minerals, with minor late-stage clay alteration. Lineations throughout this thin section are interpreted to represent bedding.

Ocl D

The next subunit, Ocl D, is ledge forming, blue-gray to dark gray recrystallized dolostone with secondary black chert (Figure 9). The dolostone often displays the elephant-skin weathering texture that typically characterizes carbonate outcrops in Nevada. This unit is silicified and also contains thin (~1 cm), white quartz veins of varying orientations. Quartz veins are more abundant near the contacts with the overlying and underlying subunits. Ocl D is approximately 270 feet (83 meters) thick and is not known to host any graptolites or other fauna.

Figure 9. Photo of massive silty dolostone with secondary black chert that is characteristic of the subunit Ocl D within the Comus Formation. Note that the quartz veins are perpendicular to bedding and are prominent, while the surrounding carbonate is more easily weathered away.

Ocl S_2

The next unit, Ocl S_2, is a recessively weathering, very fine-grained, silty quartz arenite (Figure 10). Outcrops of this unit are rare. Ocl S_2 is flaggy and thin-bedded. It is tan on fresh surfaces, and weathers pink. It also occasionally has leisegang banding. Ocl S_2 is approximately 45 meters thick.

The age of graptolites in Ocl S_2 has also been reclassified in this study. Graptolites that were collected near the Silver Coin mine by Erickson and Marsh (1974a) and analyzed by R.J. Ross (USGS unpublished report) were determined to be late Caradoc to Ashgill in age (Figure 6). The age designation for the graptolites in Ocl S_2 required reanalysis based on revisions in the understanding of graptolite zones in the Basin and Range (Carter, 1972; Finney and Perry, 1991). The species of graptolites that were found in Ocl S_2 were reclassified have since been reclassified as middle to late Upper Ordovician (Carter, 1972; Finney and Perry, 1991).

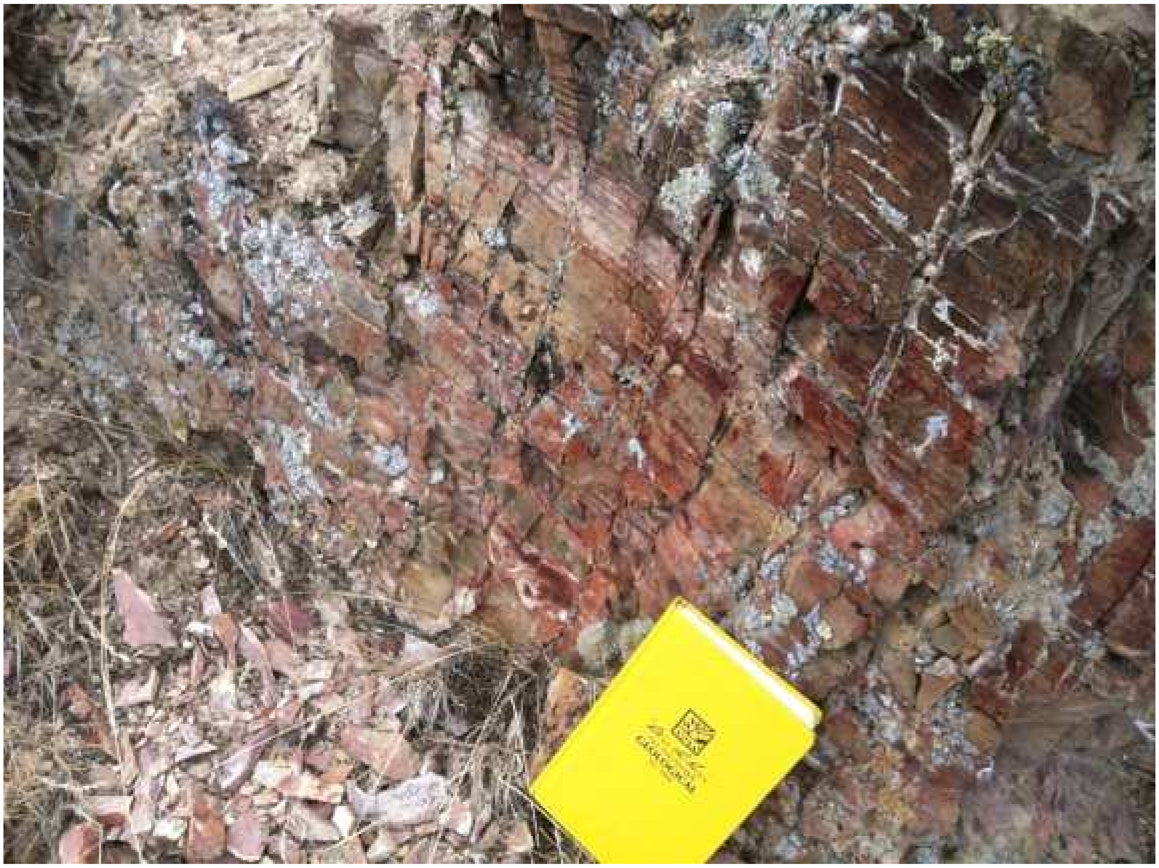


Figure 10. Photo of recessively weathering, very fine-grained, pink silty quartz arenite that is characteristic of Ocl S_2. Outcrops of this unit are rare at Iron Point.

Ocl M

The next unit, Ocl M, is a ledge-forming mudrock with interbedded siltstone and lenses of carbonate (Figure 11). The mudrock is black and medium bedded (5-10 cm). The entire unit is silicified, so it is impossible to determine whether or not the mudrock was originally carbonate mudstone. The lenses of carbonate are blue-gray on fresh surface, weather dark gray, and display a typical elephant-skinned weathering texture. Intervals of carbonate are unreactive to HCl and are variable in thickness (~5-30 cm). The siltstone intervals can range in color, but are often pink, tan, or black, and very fissile. Oclm is approximately 170 feet (51 meters) thick, and is not known to host any graptolites or other fauna.

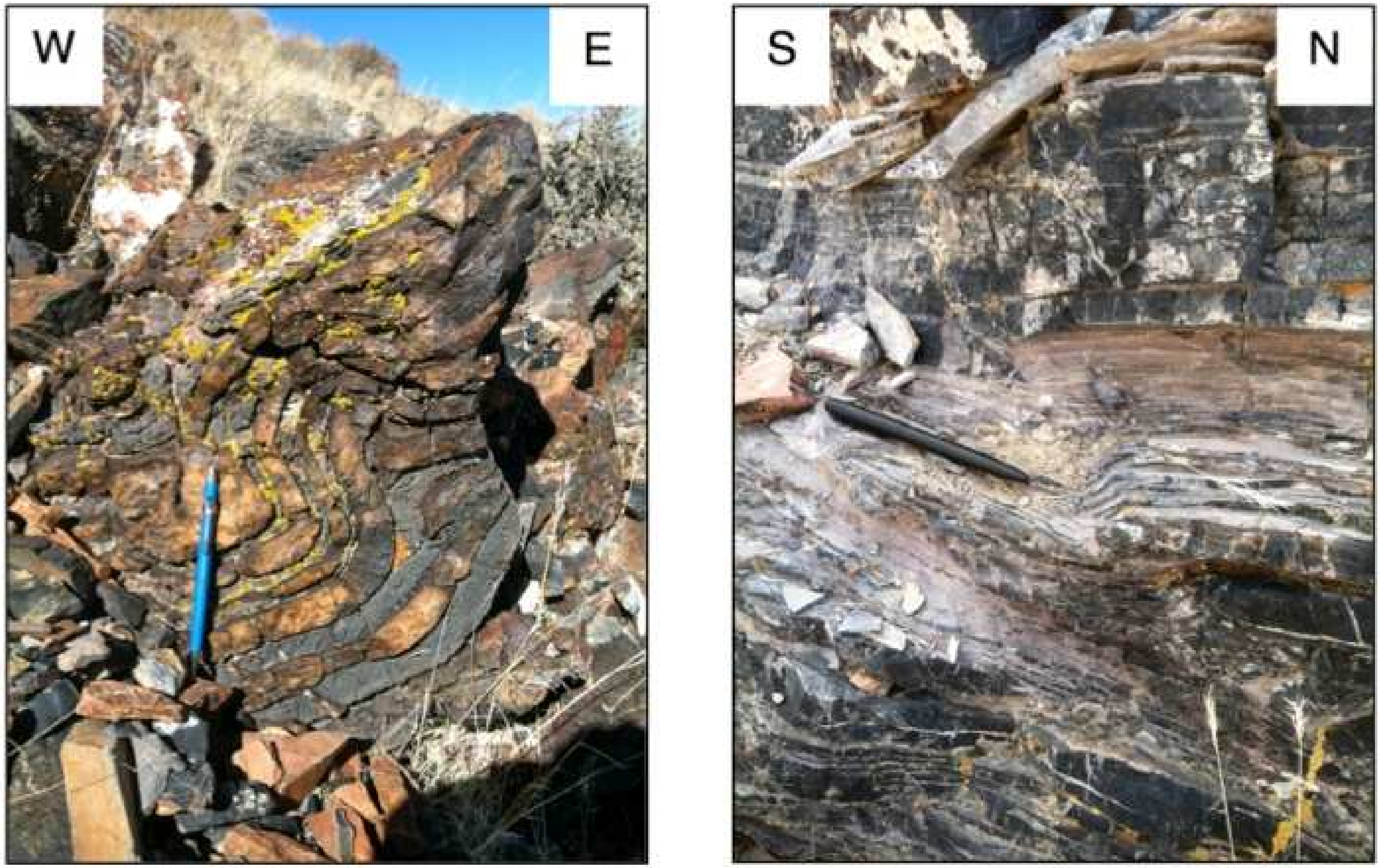


Figure 11. Photos of characteristic features of the subunit Ocl M in the Comus Formation. The photo on the left shows silicified mudrock with lenses of blue-gray carbonate of varying thickness. The photo on the right shows bedded silicified mudrock interbedded with pink, fissile shale.

Ocss

The next unit, Ocss, is a recessively weathering, very fine to fine-grained quartz arenite (Figure 12). Previous studies have referred to this unit as a “dolomitic siltstone”. (Erickson and Marsh, 1974a). Ocss is light gray to tan on fresh surface, and is characterized by orange, flaggy weathering. This unit reacts poorly to HCl when powdered, and may be cemented with dolomite; no test was made to confirm this. The bedding is wispy, which suggests bioturbation, but no graptolites or other fauna were identified (Figure 13). Ocss is approximately 470 feet (144 meters) thick.

Ocss contains a pervasive axial planar cleavage, due in part to its fortuitous position within the Comus stratigraphy (Figure 14). The widespread cleavage in Ocss could be confused with bedding. Bedding was determined based on changes in grain size. Graded bedding and rare mud lenses were observed and used to determine both stratigraphic “up” and bedding (Figure 15). The significance of this cleavage to the deformational story at Iron Point is examined in the discussion section of this thesis.

Figure 12. Photo of flaggy weathering that is characteristic of the subunit Ocss within the Comus Formation.

Figure 13. Photo of wispy beds that suggest bioturbation within Ocss.

Figure 14. Photo of cleavage and bedding within Ocss. Red line represents bedding, and blue line represents cleavage.

Figure 15. Photo of mud stringer within Ocss that indicates bedding. Red arrow is pointing to a near-vertical black mud lens.

New unit within Comus Formation - Ocus

The uppermost exposed unit of the Comus Formation at Iron Point is Ocus (Figure 7). Ocus is a recessively weathering, tan to buff siltstone and contains rare mud lenses (Figure 16). This unit is thin-bedded and poorly exposed; Ocus has only been identified in canyon walls and road cuts. Ocus is at least 246 feet (75 meters) thick, but the top of the unit is truncated by the western bounding fault, so the true thickness is unknown (Plate 1). It is not known to host any graptolites or other fauna.

This unit was previously identified as a part of the Vinini Formation (Erickson and Marsh, 1974a). The contact with the underlying Ocss unit is not exposed, but is presumed to be conformable based on continuity of bedding orientation across the section. Additionally, Ocus represents a rock type from a depositional environment that is more lithologically similar to the Comus Formation than the Vinini Formation.

Figure 16. Photo of the subunit Ocus within the Comus Formation. Black circle indicates location of small, black chert nodule. Blue lines highlight axial planar cleavage.

Ordovician Vinini Formation

Ovs

The lowermost subunit of the Vinini Formation that is exposed at Iron Point is Ovs (Plate 1A). This unit is a recessively weathering gray, thin- to medium-bedded siltstone (Figure 17). There is extensive hydrothermal clay and silica alteration throughout the unit, so the color of the fresh surface was difficult to determine. The depositional base of Ovs is covered by Quaternary alluvium at Iron Point, and may be faulted to the east by the basin-bounding fault. The contact with the overlying unit is faulted in some places and gradational in others. There are many faults and folds within this unit, so it is possible that the upper contact is conformable with the overlying unit, but has experienced slip along it during previous deformational events due to contrasting lithologic competence.

Ovs has no clear bedform markers, such as dewatering structures or bioturbation, to indicate which direction is stratigraphically “up”. If these markers do exist, they have likely been obliterated due to the intense deformation that is observed in Ovs. This unit was determined to be the stratigraphically lowest based on how the stratigraphic sequence of the Vinini Formation exposed at Iron Point compares to the well-defined stratigraphy at its type section in the Roberts Mountains (Figure 1) (Finney and Perry, 1991).

Figure 17. Photo of the subunit Ovs within the Vinini Formation.

Ovq

Overlying the altered gray siltstone is Ovq (Plate 1A). This unit is a very fine-grained quartzite with no discernable bedding, which weathers in many different colors, from tan to black. Ovq has a sugary texture and is typically white to gray on a fresh surface. In areas where the quartzite has been hydrothermally altered, the quartzite can be black on fresh surface. Due to poor and discontinuous outcrop exposure, and unclear contacts with the underlying and overlying units, the thickness of Ovq is unknown.

In thin section, Ovq is primarily composed of sub-millimeter, well-sorted detrital quartz grains (Figure 18). The quartz grains are very small, but overall uniform in size, which suggests that this unit was deposited far from a clastic source. The quartz grains have irregular grain boundaries, due to dissolution from high stress conditions. This unit also contains minerals other than quartz, including opaque oxide minerals and rare mica; however, these could be a result of later hydrothermal alteration.

As with Ovs, there are no clear bedform markers to indicate which direction is stratigraphically up in the subunit Ovq. The thin section of Ovq does not show any indications of lamination. The Vinini Formation at its type section contains abundant quartz sandstone intervals in the lower section, whereas the upper section is noticeably devoid of quartz sandstone intervals (Finney and Perry, 1991). Based on this, this study assumes that the quartzite within the Vinini Formation exposed at Iron Point represents an interval within the lower section of the formation.

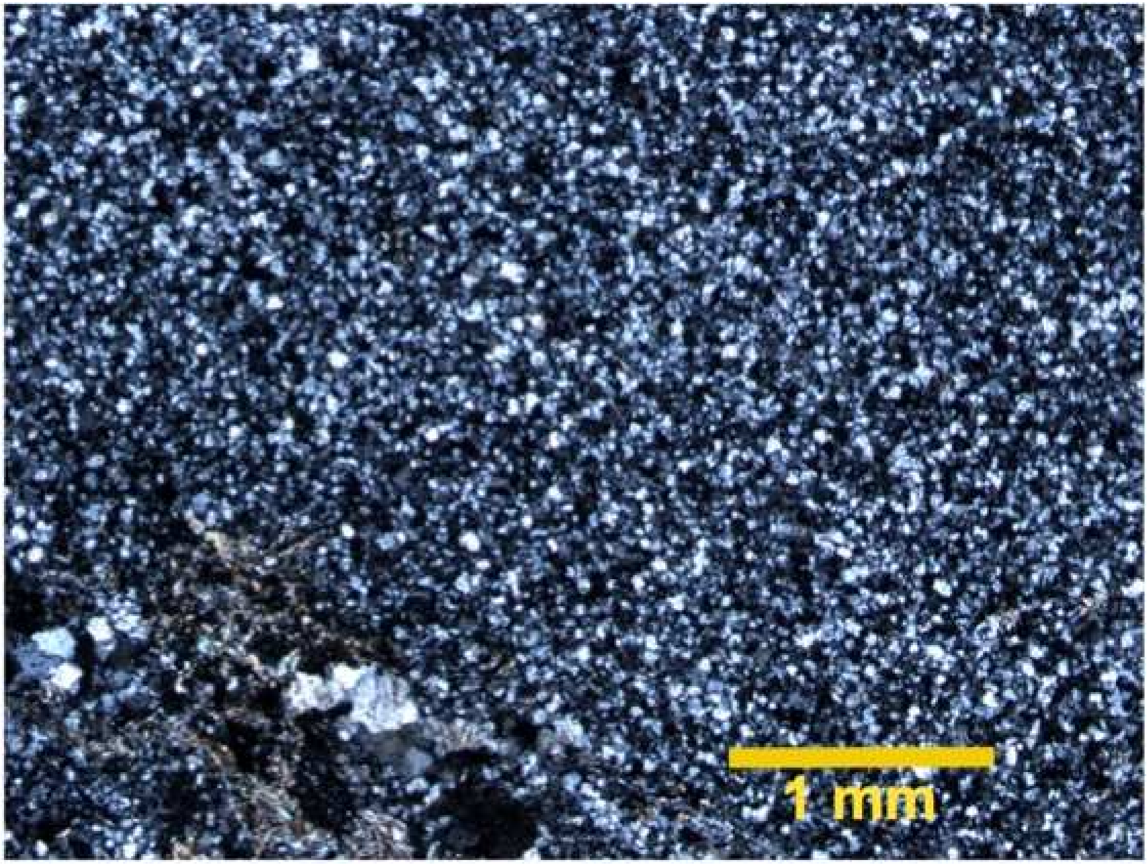


Figure 18. Cross-polarized photomicrograph of the subunit Ovq within the Vinini Formation. This is a quartzite composed primarily of very fine, well-sorted quartz grains and opaque minerals. Some late stage clay alteration is visible in the lower left corner of the figure.

Ovms

Overlying the quartzite is Ovms (Plate 1A). Ovms is composed of interbedded black chert, black mudrock, and shale (Figure 19). The mudrock sometimes contains starved ripples, and very faint ripple marks can be seen in hand sample (Figure 20). The shale is typically gray, however, in some areas has a green, phyllitic sheen.

The Vinini Formation at Iron Point is a known host of vanadium mineralization (Erickson and Marsh, 1974a). The rock is black and very altered in the mineralized areas, and historic mining disturbance is present. The entrance to Silver Coin mine is in Ovms (Figure 21). At the portal to the mine, the mudrock is black and is not silicified like other areas of hydrothermally altered Ovms. The mudrock and shale are very friable and have a matte look.

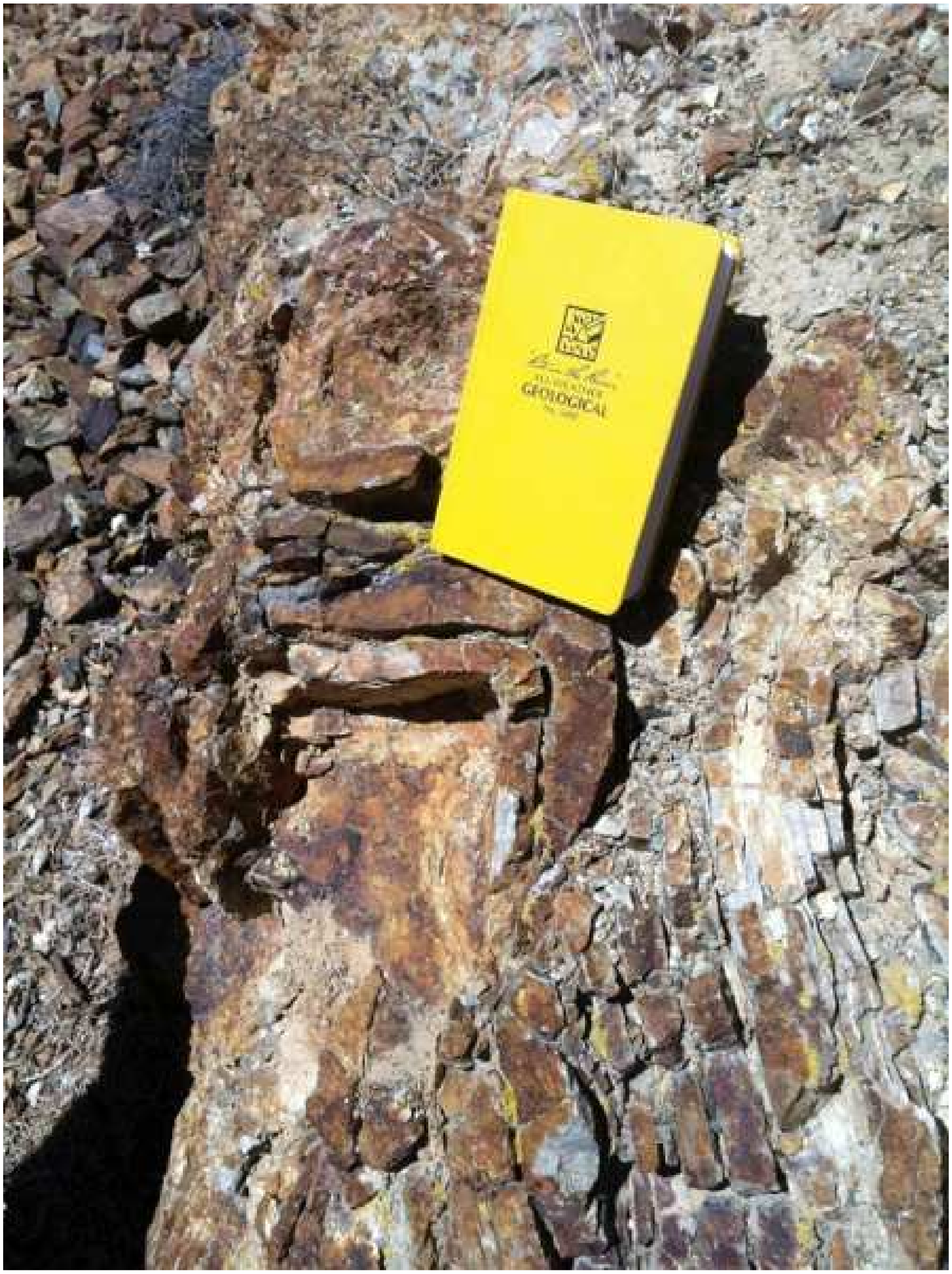


Figure 19. Ribbon chert that is characteristic of the subunit Ovms within the Vinini Formation at Iron Point.

Figure 20. Starved ripples (light tan, stained red by weathering of secondary iron oxides) within black mudrock in the subunit Ovms within the Vinini Formation.

Figure 21. Photo of bedded mudrock of Ovms at the entrance to the Silver Coin Mine at Iron Point.

Newly identified unit – Oe

A new unit at Iron Point was identified during this study, and is herein referred to as Oe. This unit contains breccia, conglomerate, and quartzite. The quartzite was mentioned in the original description of the lowest part of the Comus Formation (Erickson and Marsh, 1974a). The breccia and conglomerate in the Oe were not mentioned in the original description of the Comus Formation (Erickson and Marsh, 1974a).

The breccia within subunit Oe is composed primarily of black chert and quartzite (Figure 22). The breccia clasts are variable in size from 2 to 6 cm. The matrix is composed of fine to very fine quartz sand and is typically hydrothermally altered to orange or red iron oxide.

The conglomerate within Oe is composed of pebble- to cobble- sized clasts of black chert and quartzite (Figure 23). The matrix is composed of fine to very fine quartz grains. The clasts are sub-angular to rounded.

Weak grading was observed in some outcrops of the conglomerate (Figure 24). Outcrops of the conglomerate are discontinuous, so it is difficult to determine which direction was stratigraphically “up”.

The quartzite weathers brown, white and crystalline on fresh surface. In thin section, irregular quartz grain boundaries show dissolution, which suggests that the quartzite was under high pressure, likely due to deep burial, prior to uplift (Figure 26). Thin section shows that the quartzite is not laminated. The quartzite is composed entirely of quartz; no feldspars or other detrital grains were observed in thin section.

The nature of the contact between Oe and the Comus Formation is unclear. Oe may be bounded by faults, but no fault surfaces were found at the surface. However, the presence of breccia and hydrothermal alteration suggests faults are present within the unit.

Figure 22. Breccia within the newly identified unit Oe. Matrix is fine- to very fine-grained sand and angular white clasts are quartzite.

Figure 23. Stratified pebble to cobble heterolithic conglomerate within the newly identified unit Oe. Dark gray to black clasts are chert, tan clasts are quartzite. Clasts are angular to sub-rounded.

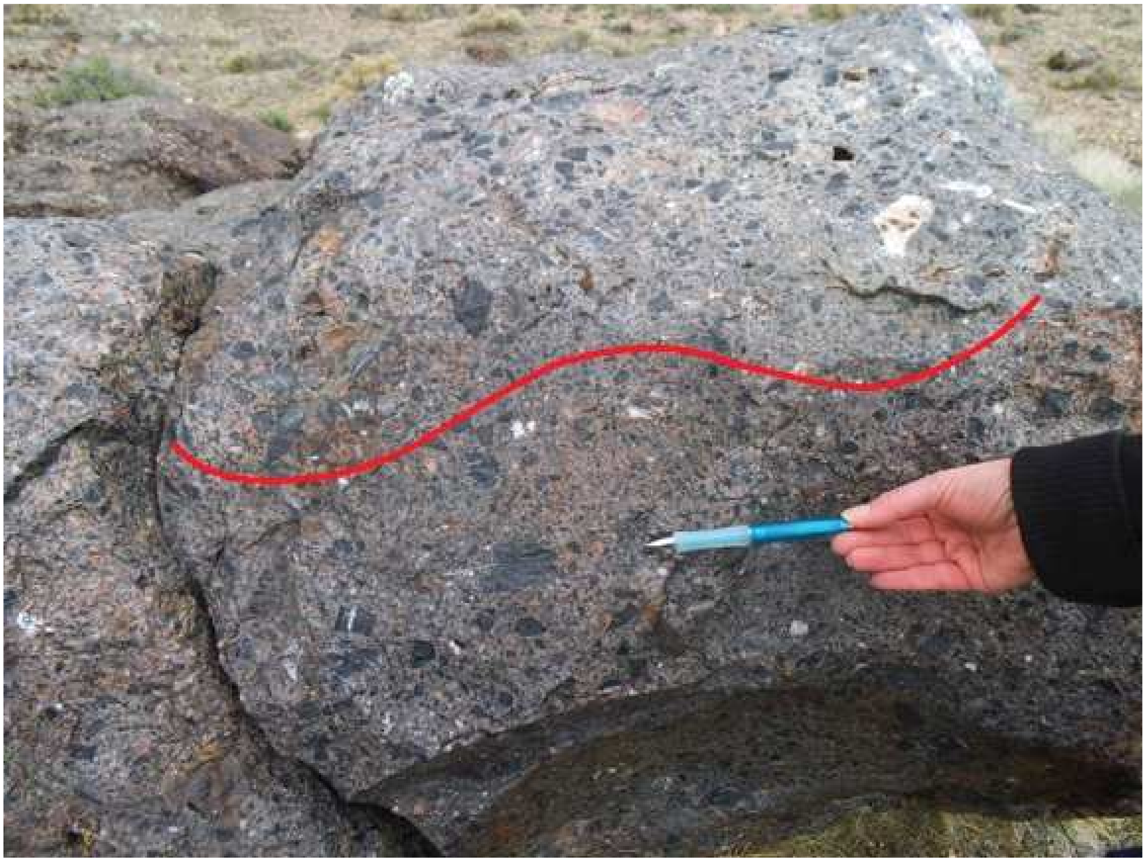


Figure 24. Weakly graded heterolithic conglomerate within the newly identified subunit Oe. Above red line, clasts are, on average, larger, below red line, clasts are, on average, smaller.

Figure 25. Possible brecciated conglomerate within subunit Oe. White, rounded clasts are quartzite; black, angular clasts are chert.

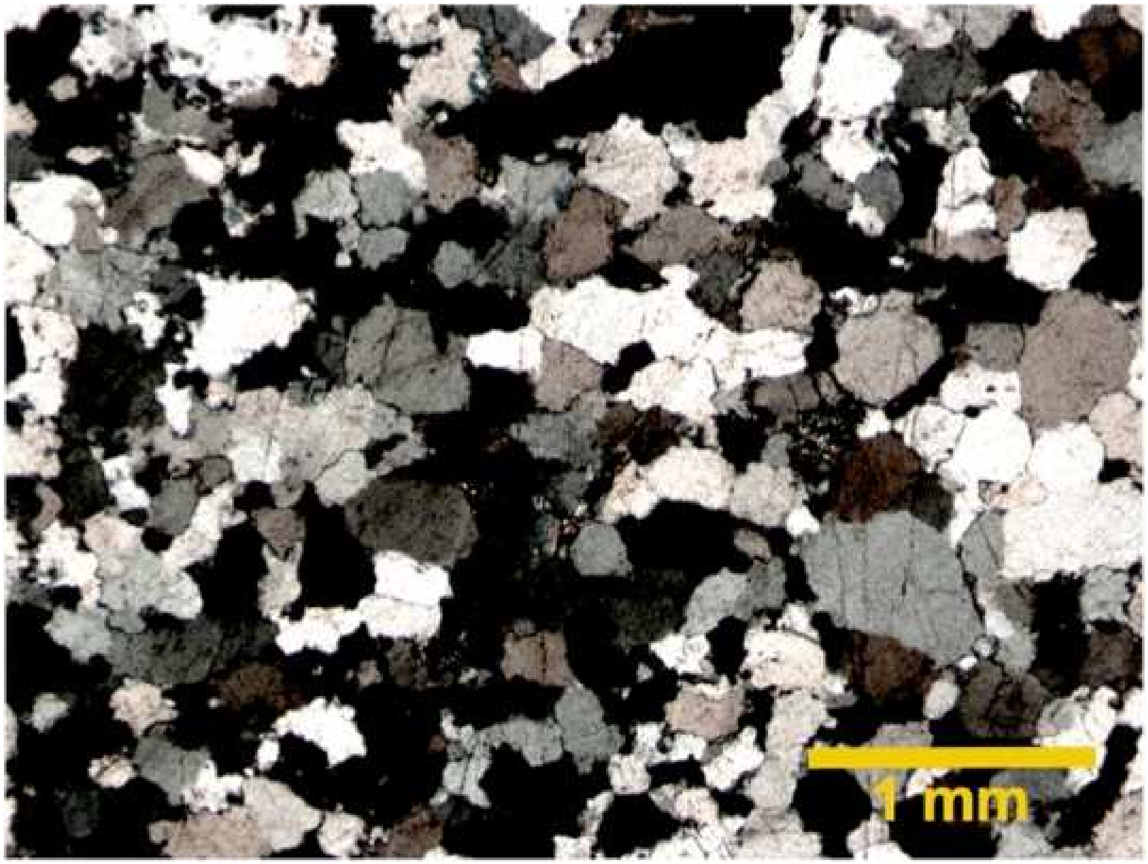


Figure 26. Photomicrograph of quartzite within the newly identified unit Oe. This quartzite is composed almost entirely of clastic quartz grains. Grain boundaries are irregular, likely due to deep burial and partial recrystallization.

STRATIGRAPHIC SUMMARY

The following section highlights the new data produced by this study on the stratigraphy at Iron Point. Prior to this study, the internal stratigraphy of the Comus Formation had not been mapped in detail. The Vinini Formation is present, but is stratigraphically distinct and structurally separate from the Comus Formation at Iron Point. A unit previous identified as the Vinini Formation was remapped as the uppermost exposed subunit, Ocus, of the Comus Formation. Finally, a new, previously unidentified unit that underlies the Comus Formation, composed of conglomerate, breccia, and quartzite was mapped.

The three Ordovician units exposed at Iron Point are compositionally distinctive from one another. The Comus Formation is composed of six distinguishable subunits, characterized by interbedded carbonate and siliciclastic material. The amount of siliciclastic material increases up-section in the Comus Formation. The Vinini Formation at Iron Point is composed of siltstone, quartzite, and bedded chert. There is no carbonate in the Vinini Formation at Iron Point.

A new subunit within the Comus Formation was identified by this study and named “Ocus”. This unit had previously been mapped as a part of the Vinini Formation (Erickson and Marsh, 1974a). Ocus is characterized by siltstone with some sand sized quartz grains and rare chert nodules. Ocus may be cemented with dolomite, however, no test was made to confirm this. Ocus is overall composed of finer grained siliciclastics compared to the underlying unit Ocss, which suggests a fining upward sequence. Ocus also preserves the axial planar cleavage that is observed in the underlying unit Ocss. While no fossils were found, based on the compositional similarities to the underlying unit, Ocss, and the dip of bedding, Ocus was determined to be the stratigraphically highest unit of the Comus Formation exposed at Iron Point.

A newly recognized occurrence of the Eureka Quartzite, stratigraphically beneath the Comus Formation, herein referred to as “Oe”, was mapped at Iron Point during this study. This Eureka Quartzite here is composed of conglomerate, breccia, and quartzite. Quartzite was recognized in original description of the Comus Formation (Ferguson, 1952) but the conglomerate and breccia were not. The quartzite weathers brown, is light gray and has a sugary texture on its fresh surface, and does not have any bedding features. In thin section it is almost completely composed of quartz grains that are homogenous in size. The conglomerate contains pebble to cobble sized clasts of dark gray rock that look to be sourced from the overlying carbonate rocks in the Comus Formation. The conglomerate also contains pebble- to cobble-sized clasts of quartzite. The matrix is composed of quartz grains. The breccia is composed of the same clasts and contains the same matrix as the conglomerate.

Very few Paleozoic quartzites are mapped in the Great Basin, and only one is widespread in the Ordovician: the middle Ordovician Eureka Quartzite. It is a mature and clean quartz arenite; the textural and compositional uniformity of the Eureka Quartzite makes it easily recognizable in the Great Basin. Conglomerate and breccia are not typical of the Eureka Quartzite. However, one anomalous area in the Nopah Range in southeastern California (Figure 1), cave deposits within the Eureka Quartzite crop out. These cave deposits are composed of clasts from the overlying late Ordovician Ely Springs Dolostone and cemented by quartz sand, similar to the “Oe” unit at Iron Point.

IRON POINT STRUCTURE

Based on the new detailed internal stratigraphy of the Comus Formation, it is possible to identify folds and faults based on repetition of the stratigraphy at Iron Point. The following section will describe and interpret these folds and faults. These structures are discussed in chronological order, from oldest to youngest. Each fold set and fault is interpreted to represent a distinct deformational event. Conventional notation is used to describe each fold set, starting with the oldest (F_1), and each deformational event, starting with the oldest (D_1). A brief explanation summarizes cross-cutting relationships that describe order of formation; however, the details of these interpretations is discussed further in the deformational history section. The data used in this section are original unless otherwise noted.

F_1 Folds

The first generation of folds at Iron Point, F_1, are developed in the Comus Formation, but were not observed in the Vinini Formation. F_1 folds are asymmetric, steeply inclined, and locally overturn to the west. These folds are large and spread out over the study area; the average wavelength of F_1 folds is ~2600 feet (~800 meters). A moderate to steeply east-dipping (60-80°) well-developed axial planar cleavage is best observed in the upper units of the Comus Formation (Figure 14; Figure 16). The average orientation of the cleavage is roughly north-south.

The most apparent F_1 fold at Iron Point is the map-scale anticline within the Comus Formation, herein referred to as the “Comus anticline” (Figure 27). The Comus anticline is locally overturned to the west (Plate 2). The Comus anticline is doubly-plunging, and plunges more steeply to the north (~31°) than to the south (~12°) (Figure 27). The west limb of the Comus anticline is truncated by the King fault, and the east limb is truncated by the Pumpernickel fault.

F_1 folds are rootless and affected by all other folds and faults observed at Iron Point. Based on these cross-cutting relationships with the other structures observed in the Comus Formation, F_1 folds represent the oldest phase of deformation at Iron Point. The Comus anticline is doubly plunging and plunges more steeply to the north than to the south; this suggests that there has been later reorientation of F_1 folds. The Comus anticline is also truncated by the King fault and Silver Coin thrust fault.

A west-verging deformational event (D_1), with an east-west shortening direction is recorded by the asymmetry of the F_1 folds and the well-developed axial planar cleavage in the upper Comus Formation. This style of deformation has been observed regionally (Trexler et al., 2004 and references therein; Villa 2007; Cashman et al., 2011) and is discussed further in the deformational history section.

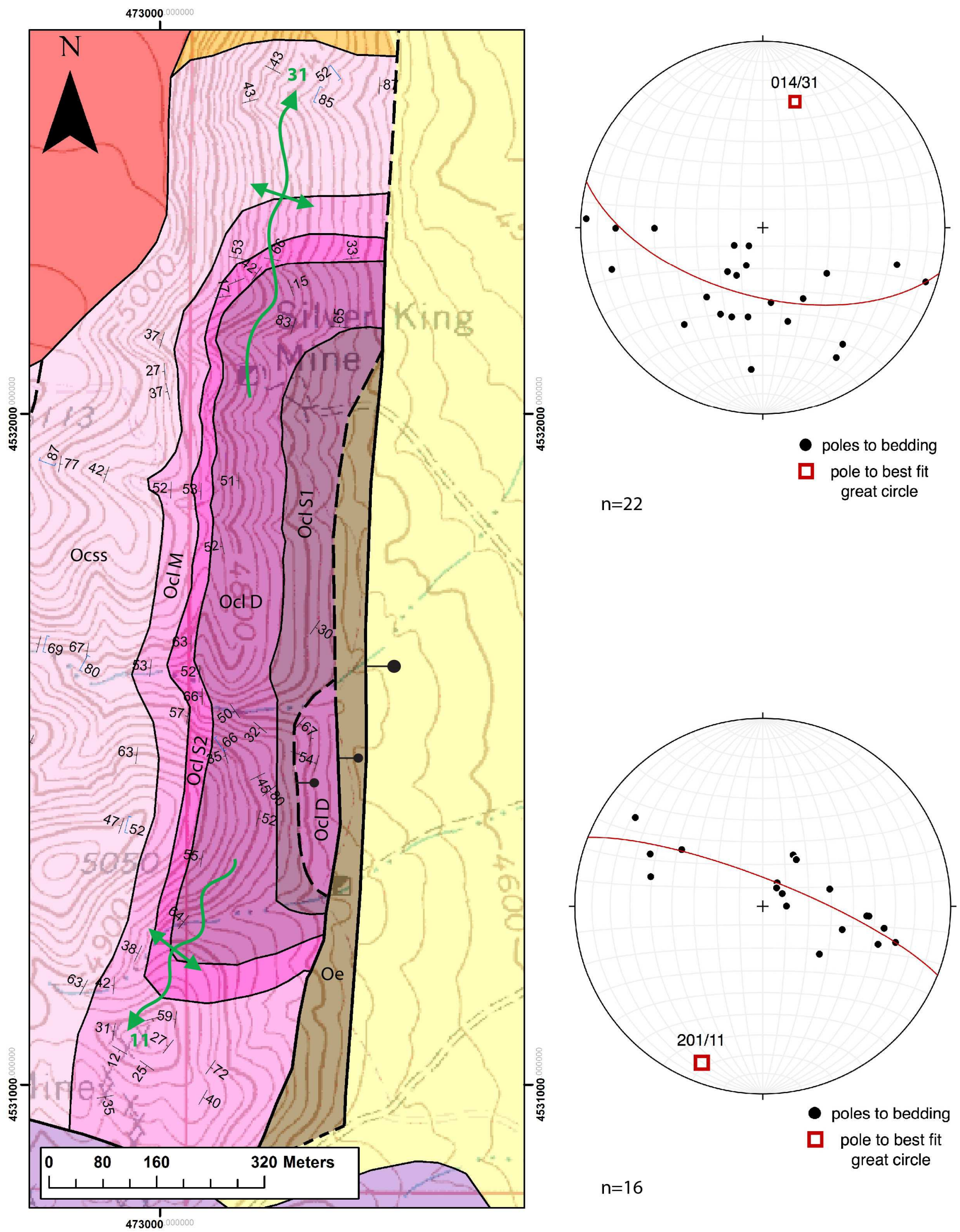


Figure 27. Map of the "Comus anticline" illustrating the location and geometry of F_1 folds in the Comus Formation. The trace of the Comus Anticline is in green, and has been overprinted by later deformation. Grid is in NAD 83 UTM Zone 11N.

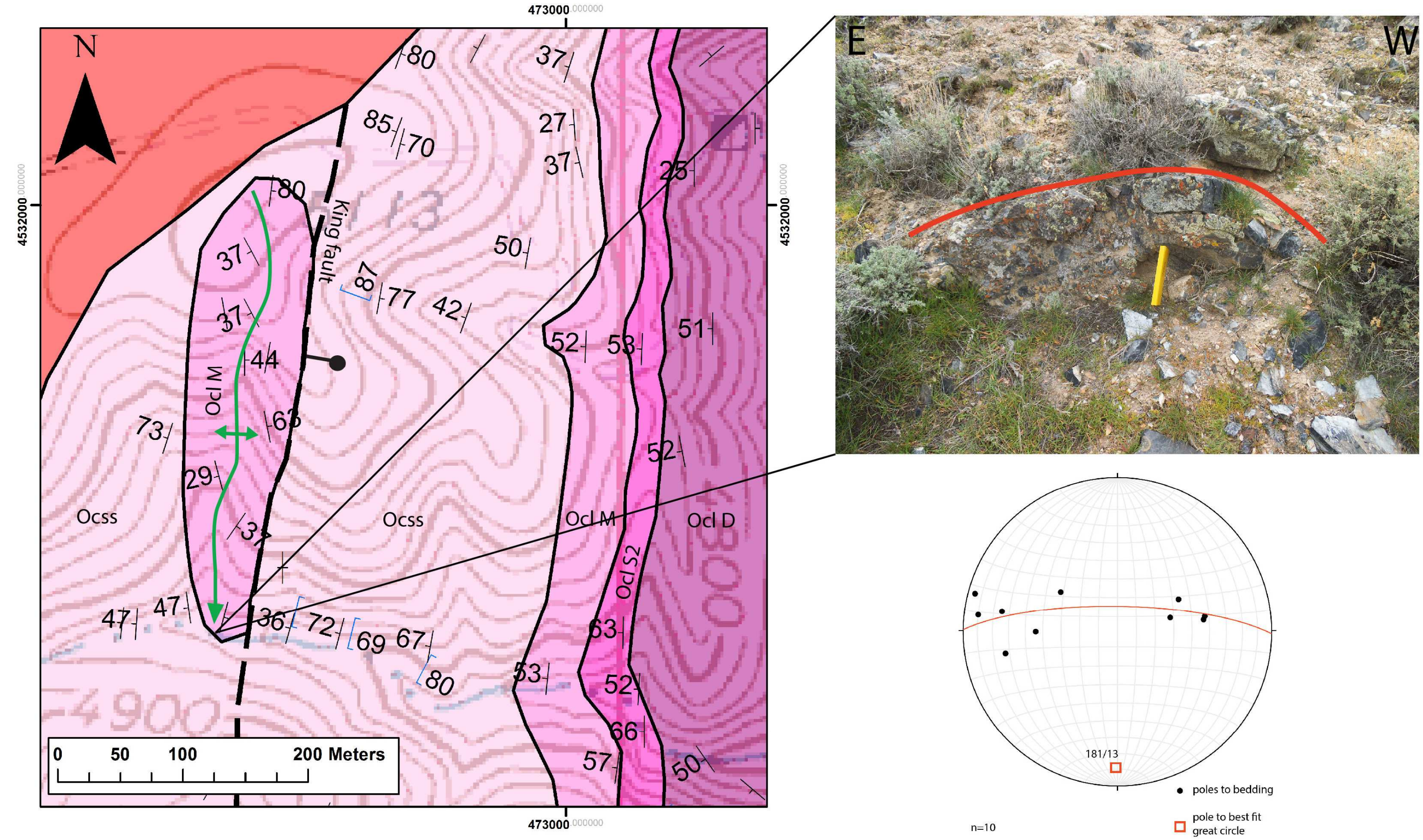


Figure 28. Trace of an F_1 fold in the subunit Ocl M in the Comus Formation (green). The photo shows the fold in Ocl M plunges to the southwest into the hillside. Stereonet shows the calculated trend and plunge of the Ocl M fold based on bedding measurements taken from Ocl M exposures in the footwall of the King fault. Grid is in NAD 83 UTM Zone 11N.

F_2 Folds

The second generation of folds at Iron Point, F_2, is developed in the Comus Formation, but has not been observed in the Vinini Formation. These folds are upright, open, and symmetric (Figure 29). F_2 folds are map-scale; average wavelength is approximately 2600 feet (800 meters). Stereographic analyses of the outcrop-scale F_2 folds show that they trend west-southwest (~303°) and plunge gently (~10) (Figure 29).

F_2 folds overprint the F_1 folds in the Comus Formation. As mentioned previously, the Comus anticline (F_1) is a doubly plunging anticline (Figure 27), which is a result of overprinting by F_2 folds. Stereographic analysis of other F_1 folds in the Comus Formation shows a gentle to moderate plunge to the north and south, as well (Figure 27; Figure 28).

The King fault offsets F_2 folds; this is most apparent in Comus Canyon, where the stratigraphic section of the Comus Formation was measured. The subunit Ocl M is exposed in the footwall of the King fault, and the subunit Ocss is exposed in the hanging wall (Figure 28). The exposure of Ocl M in the footwall of the King fault is folded about a north-south axis (F_1) and plunges gently to the south (Figure 28). The map pattern of the King fault does not show any effects of F_1 or F_2 folding (Figure 27). This indicates that the Comus Formation was folded (F_1) and refolded (F_2) prior to being offset by the King fault (Plate 3).

F_2 folds at Iron Point record a north-south shortening event (D_2). The dip of the axial plane suggests southward vergence (Figure 29), but in the absence of widespread evidence across the field area, that's a tentative interpretation only. This style of deformation has been observed less than a kilometer west at Edna Mountain (Villa, 2007;

Cashman et al., 2011). It is represented by a dome and basin map pattern that affects only the Pennsylvanian units and older deformation, but not the middle Permian or younger units (Villa, 2007; Cashman et al., 2011). Based on this, the north-south shortening occurred sometime during the early Permian (Villa, 2007; Cashman et al., 2011).

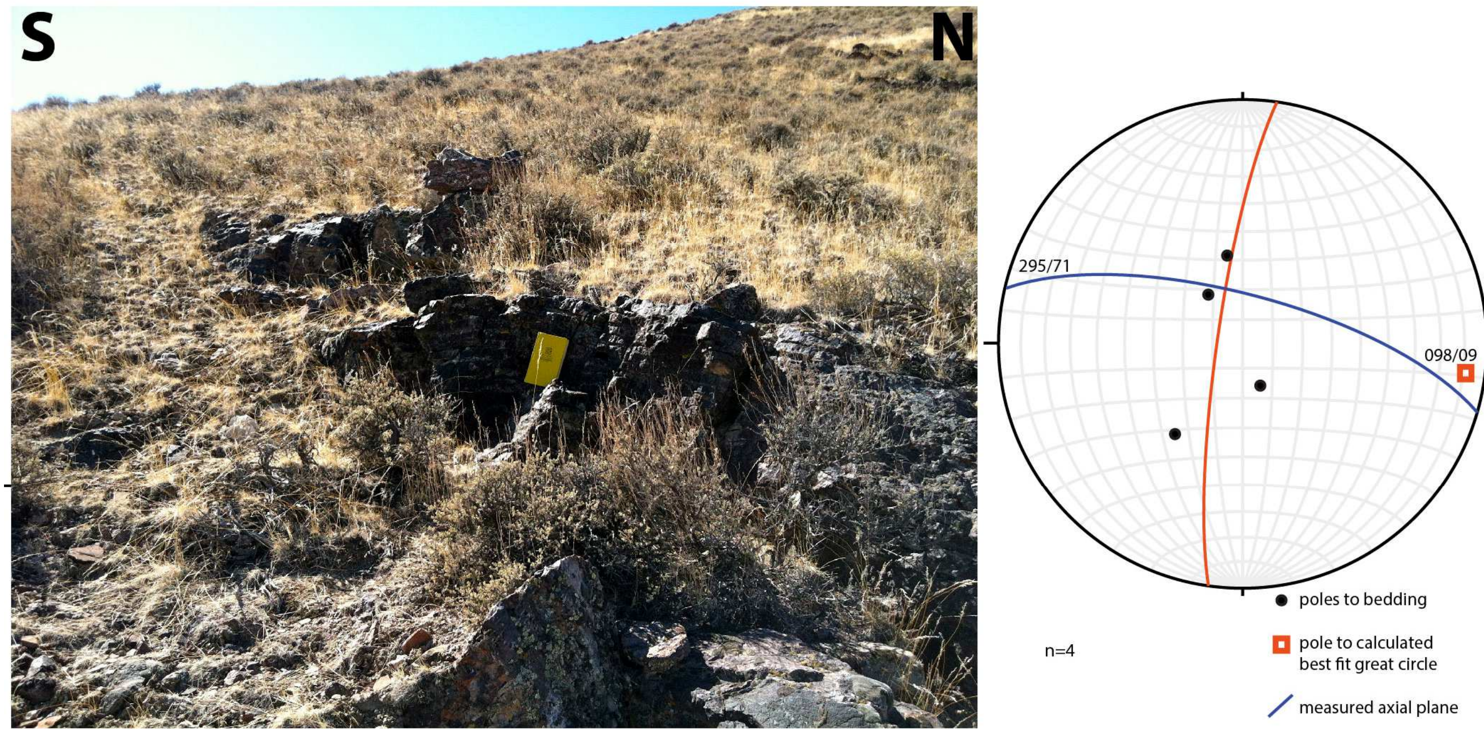


Figure 29. East-trending F_2 fold in subunit Ocl D in the Comus Formation at Iron Point. The stereonet shows the calculated trend and plunge of the fold in the photo plotted with the axial plane of the fold that was measured at the outcrop.

King fault (D_3)

The King fault is a normal fault within the Comus Formation. The King fault strikes approximately north-south (~010°) and dips moderately (~60°) east (Figure 30; Plate 2). This interpretation of fault sense is based on the map pattern of the repeated Ocl M subunit in the footwall of the fault and the duplicated thickness of the subunit Ocss north of the Silver Coin thrust (Figure 30). The King fault is truncated by the Silver Coin thrust; no evidence of faulting or repetition of units within the Vinini Formation was observed south of the Silver Coin thrust. There are no physical exposures of the King fault in the Comus Formation at Iron Point, so direct analysis of the fault surface or kinematic indicators is not possible.

The King fault is younger than the north-south shortening event that created the F_2 folds in the Comus Formation. Based on correlation with the different styles of folding at less than a kilometer to the west at Edna Mountain, F_2 folds are early Permian (see previous section on F_2 folds). Since the King fault offsets F_2 folds, the King fault is, at its oldest, post-early Permian. However, it is impossible to determine the youngest age since the King fault only affects the Comus Formation. There are not similar structures to the King fault at Edna Mountain for comparison.

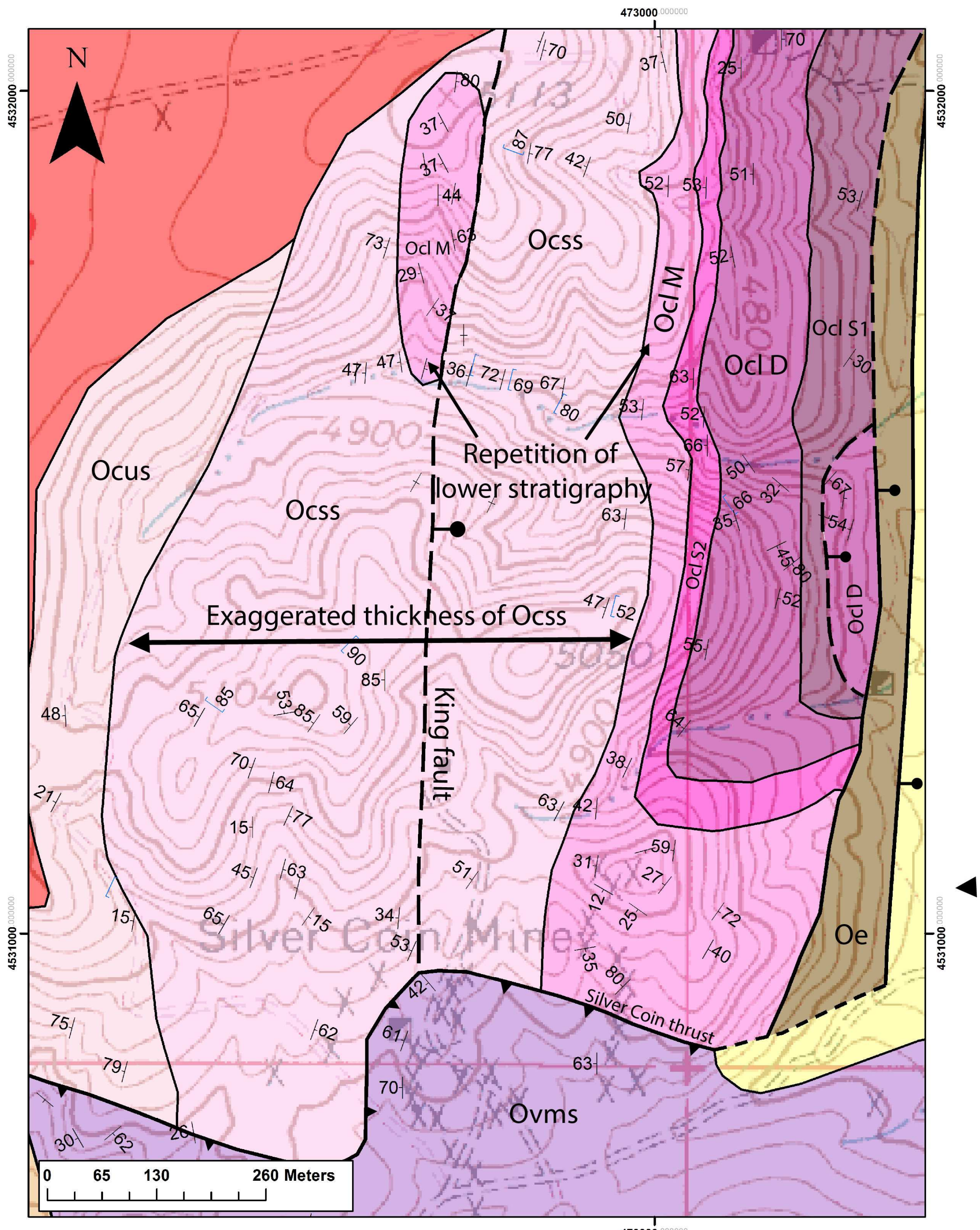


Figure 30. Map displaying the trace of the King fault. Note the repetition of the subunit Ocl M in the footwall of the King fault. The thickness of the subunit Ocss is exaggerated north of the Silver Coin thrust. Grid is in NAD 83 UTM Zone 11N.

Silver Coin thrust (D_4)

The Silver Coin thrust fault truncates the F_1 and F_2 folds and the King fault (Plate 3). There are no physical exposures of the thrust surface, so direct analysis of the fault surface or kinematic indicators is not possible. The dip angle of the thrust fault was determined using the map pattern and three-point solutions (Figure 31). The map patterns of the thrust at Iron Point suggest a gently south-dipping (~7-10°) fault with an east-northeast (074°) strike.

The Silver Coin thrust truncates F_1 and F_2 folds and the King fault. Intense internal deformation within the Vinini Formation made correlations with structures in the Comus Formation impossible (Figure 32). This intense deformation does not appear to record the same structural history as the Comus Formation. Additionally, no evidence of the King fault is present in the subunit Ovms south of the Silver Coin thrust. It is likely that folds recorded in the Vinini Formation happened prior to emplacement along the Silver Coin thrust, and therefore represent deformation that occurred elsewhere.

Internal deformation within the Vinini Formation was also analyzed in order to determine the kinematics of the thrust. However, the folds recorded in the Vinini Formation did not prove useful in determining sense of motion on the Silver Coin thrust. Many folds were measured, but there was no clear pattern to determine the style of deformation. As mentioned in the stratigraphy section, the Vinini Formation is composed of predominately interbedded chert and shale at Iron Point, which are easily deformed. The rocks of the Vinini Formation may display deformation differently that those of the Comus Formation.

The Silver Coin thrust records a shortening event, but there are no north-south shortening events documented anywhere in Nevada. It is likely that the Silver Coin thrust represents a piece of a larger thrust fault that has been truncated by the subsequent high-angle West fault and the Pumpernickel fault. Since the Silver Coin thrust truncates the King fault, it is therefore younger than post-early Permian. There is one thrust fault that is younger than the early Permian documented less than a kilometer east at Edna Mountain: the Golconda thrust. The Silver Coin thrust is interpreted to be related the emplacement of the Golconda allochthon, which will be discussed further in the deformational history section.

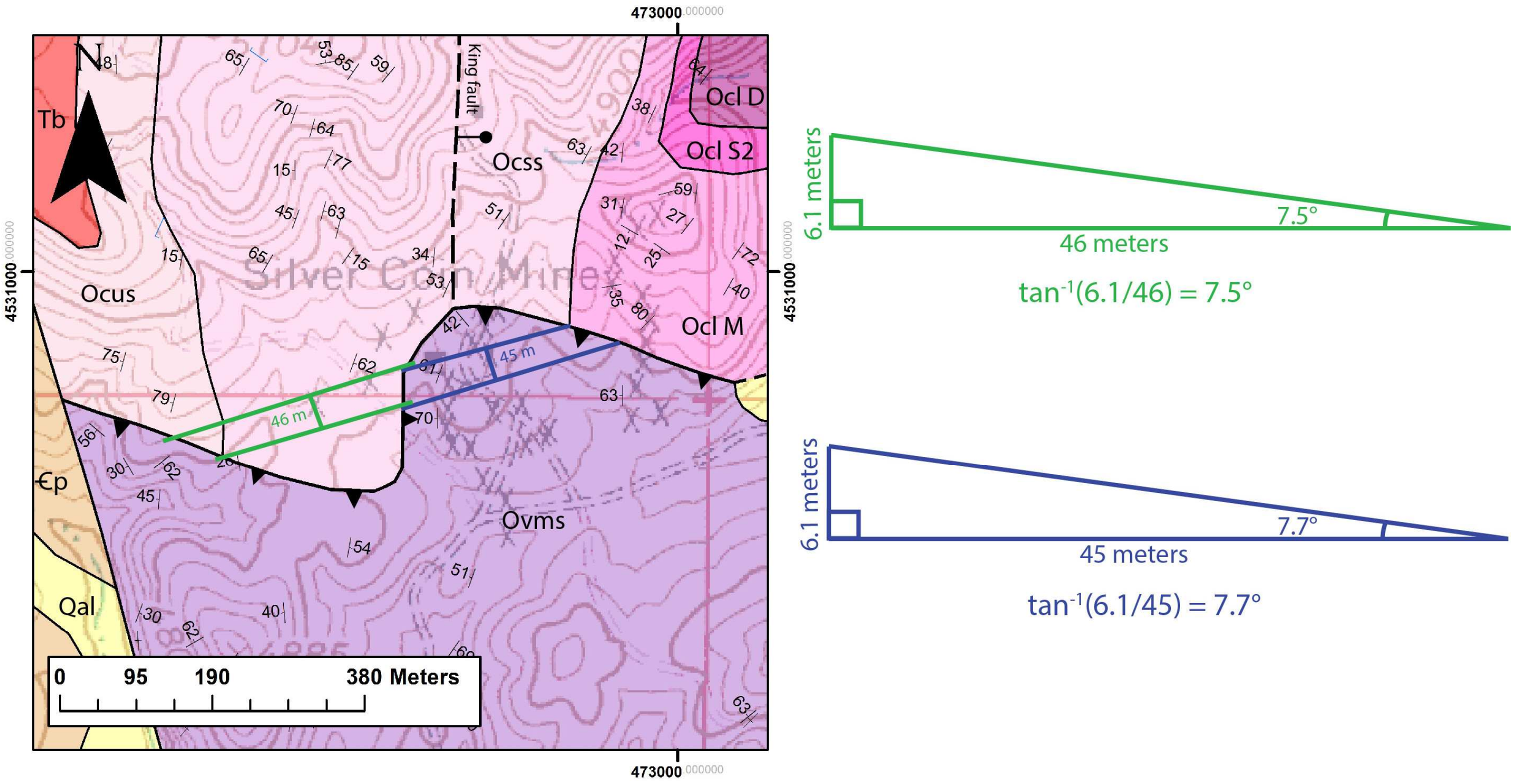


Figure 31. Two examples of three-point problems that were used to determine the approximate strike and dip of the Silver Coin thrust based on the map pattern of contact that separates the Vinini (purple) and Comus (shades of pink) Formations. Grid is in NAD 83 UTM Zone 11N.

Figure 32. Erratic folds in Ovms subunit of the Vinini Formation at Iron Point. Complex folding within the Vinini Formation made fold analysis impossible. Fold measurements and stereographic analysis of folds in the Vinini Formation did not yield any discernable pattern.

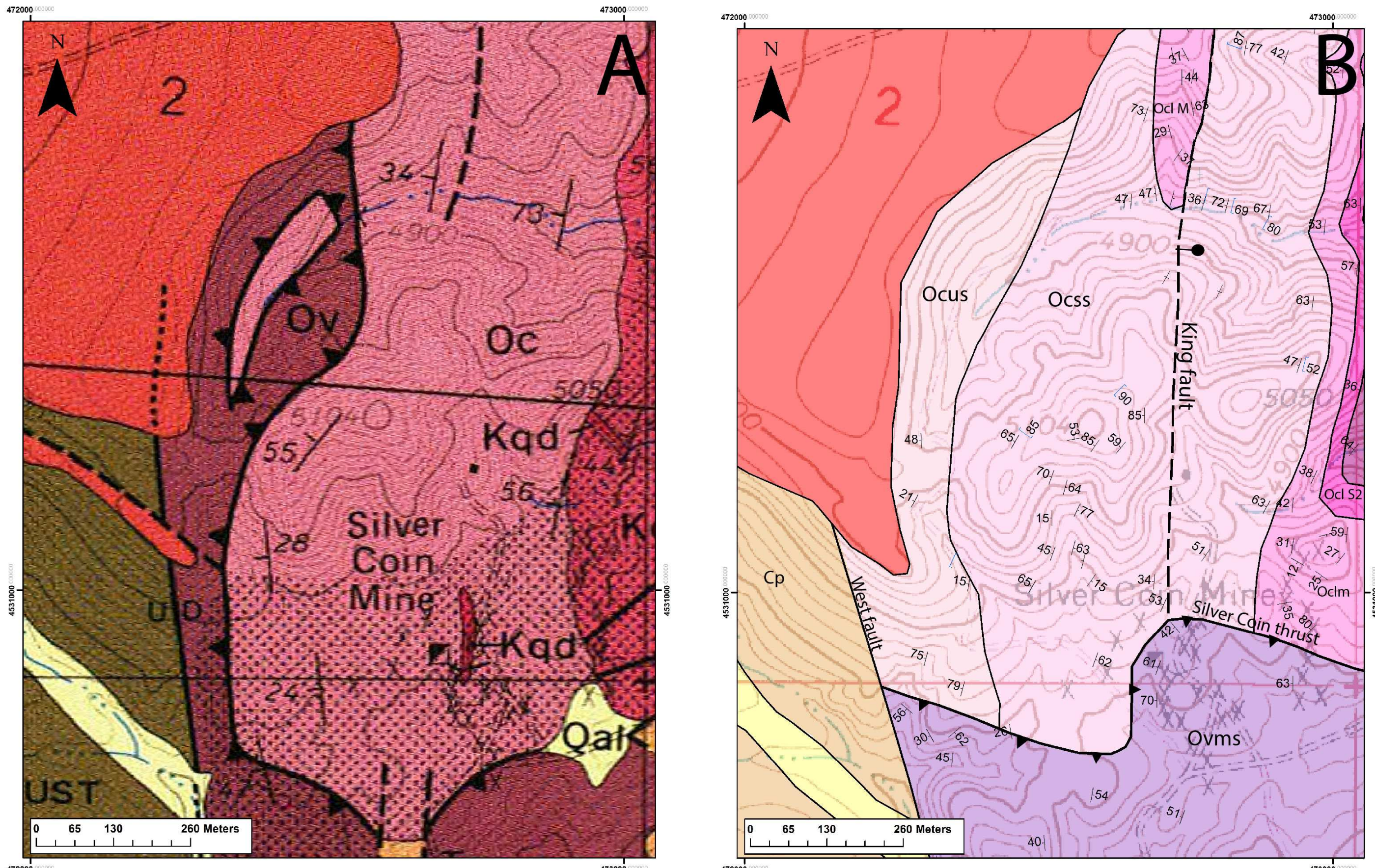


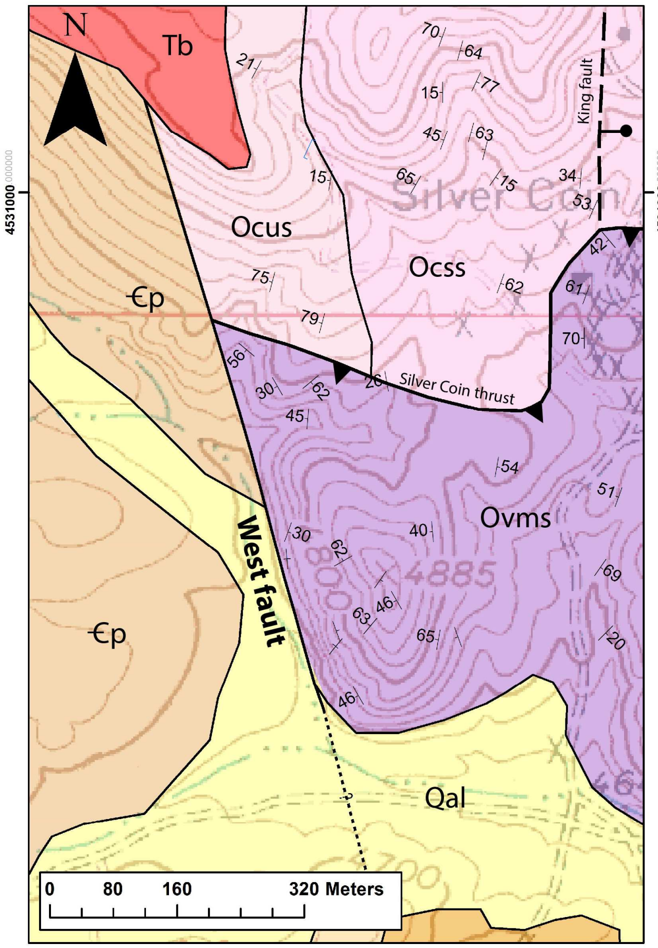
Figure 33. Original mapping (map A) displaying the Vinini and Comus thrust contact (Erickson and Marsh, 1974a), versus the new mapping (map B). Map A: Ov = Vinini Formation, Oc = Comus Formation. The original mapping at Iron Point (map A) showed the Vinini Formation wrapping around the Comus Formation. Note that on map A, the contact between the Vinini and Comus Formations is considerably south of the Silver Coin Mine, whereas on map B, the contact between the the two units has been moved to intersect the Silver Coin Mine.

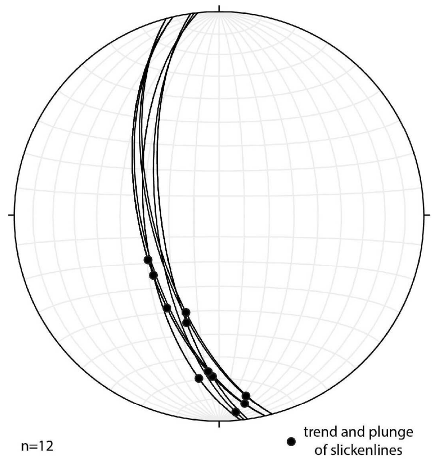
West fault (D_5)

The West fault separates Iron Point from Edna Mountain and is located on the western side of the study area (Figure 33). The Cambrian Preble Formation is in the hanging wall of the West fault, and the Comus and Vinini Formations are in the footwall. The West fault has been interpreted to be a reverse fault, based on its current orientation and juxtaposition of strata (Erickson and Marsh, 1974a). The West fault strikes, on average, southeast (~168°) and dips moderately southwest (~58°). The West fault is well exposed for a length of approximately 400 feet (120 meters) along the west-central edge of the study area, and is covered by basalt talus to the north and alluvium to the south (Figure 33).

The West fault records multiple slip events. The surface of the fault is very planar; multiple episodes of movement on the fault planed off original grooves or irregularities (Figure 34). Several sets of slickenlines were observed on the fault surface, and also support the interpretation that this fault has been reactivated. Although several sets of slickenlines were measured, however, kinematic indicators were not preserved. The styles of motion recorded on the fault surface vary from dip-slip to sinistral oblique slip (Figure 33). The dip of this fault is on average, 58°, which is a reasonable angle for a normal fault, but is unusual for a strike-slip fault. The original event that created the West fault is impossible to determine because the fault has been reactivated numerous times. However, the published interpretation of simple reverse offset does not explain the complete absence of Ordovician rocks west of the fault. It must have had significant strike-slip motion to juxtapose such dissimilar stratigraphic assemblages.

The West fault truncates the Silver Coin thrust, and all other Paleozoic deformation that has been documented at Iron Point. The Silver Coin thrust is interpreted to be post-early Permian, and likely related to the emplacement of the Golconda allochthon. The West fault truncates the Silver Coin thrust, and is therefore interpreted to post-date the Sonoma Orogeny. Map relationships along strike of the West fault south of the study area (Figure 34; Plate 1) do not show a fault along strike from the exposure of the West fault that truncates the Golconda Allochthon. However, the published map does show a southward continuation of the fault between the Preble and Vinini formations (Plate 1) (i.e., apparently the West fault); it is offset 1600 feet (500 meters) in a left lateral sense not far south of the Silver Coin thrust. There are no similar structures that correlate to the West fault at Edna Mountain, or anywhere else in the region, so the only age constraint on the West fault is that it is post-late Permian.





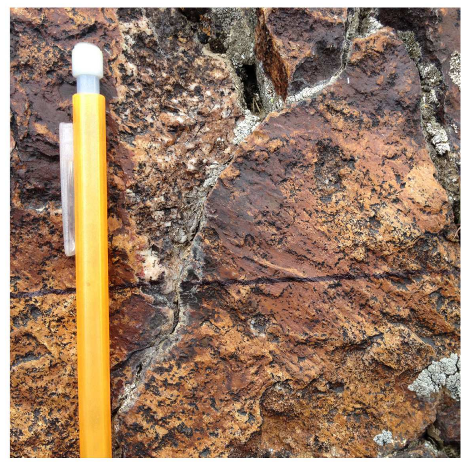


Figure 34. Map shows the trace of the West fault at Iron Point. The stereonet shows the trend and plunge of slickenlines calculated using the strike and dip of the fault surface and the rake measured at the outcrop. The photo is a close up of the fault surface, and shows how the slickenlines are preserved.

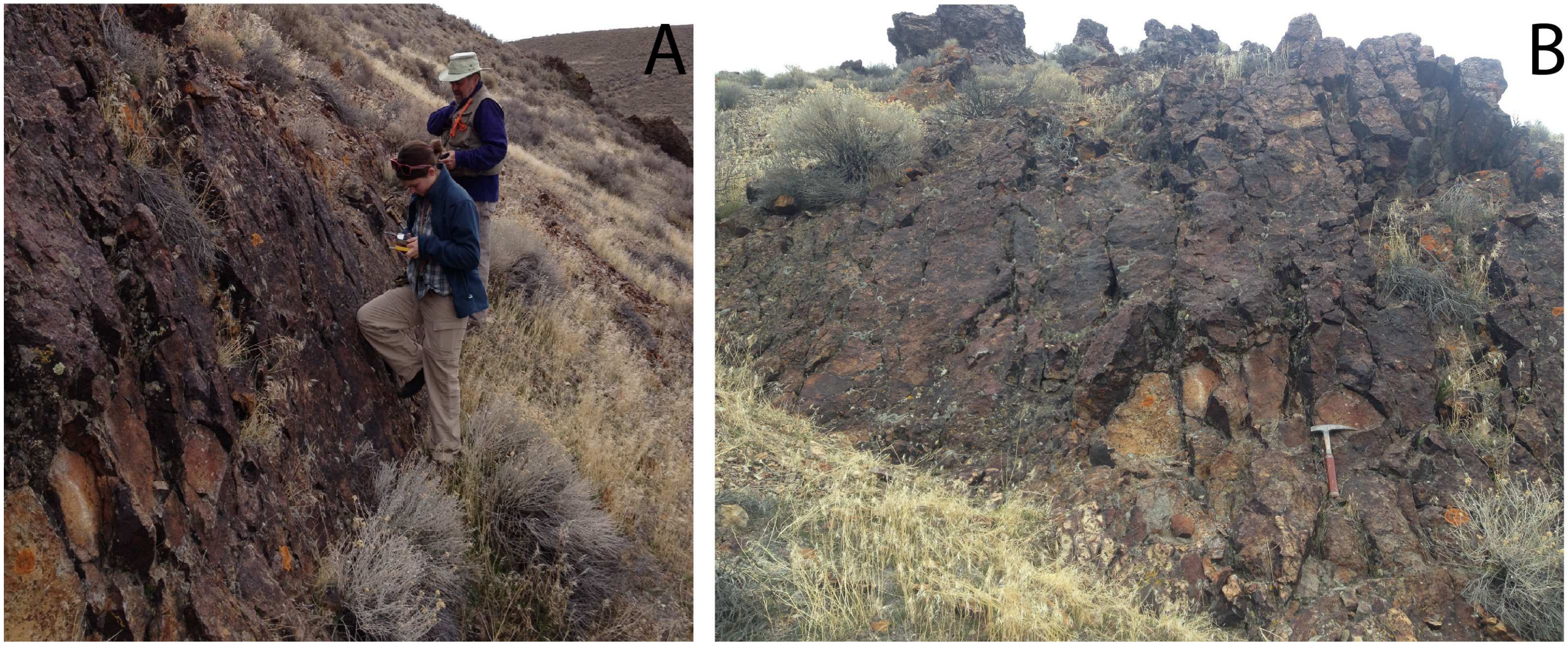


Figure 35. Photos of surface exposure of West fault at Iron Point.

Pumpernickel fault (D_6)

The Pumpernickel fault borders the study area on the east side, and separates Iron Point to the west and Pumpernickel Valley to the east (Plate 1A). Comus Formation and a fault-bounded slice of Ordovician Eureka Quartzite form the footwall of the fault, but the rock unit in the hanging wall is not exposed. The Eureka Quartzite is exposed only adjacent to the Comus Formation, and has not been observed farther south in the study area in contact with the Vinini Formation. Pliocene basalt covered by Quaternary alluvium is mapped in the hanging wall of the Pumpernickel fault, (Plate 1B) (Erickson and Marsh, 1974a). There are no physical exposures of the fault because it is covered by Quaternary alluvium, so direct analyses of the fault surface are not possible. The presence of the fault is interpreted based on the abrupt, linear change in topography on the eastern side of the study area, from hilly terrain at Iron Point on the west side of the fault, to the flat Pumpernickel Valley on the east side of the fault.

The Pumpernickel fault exposes a thin <330 foot (<100 meters) wide in map view), continuous wedge of the Eureka Quartzite (Oe) along the east edge of the study area (Plate 1A). The exposure of the Eureka Quartzite is interpreted to be a fault-bounded block, based on stratigraphic juxtaposition and on the presence of hydrothermally altered breccia. The brecciation within the Eureka quartzite increases eastward, consistent with a genetic relationship between the brecciation and the range-front fault. The west side of the wedge of Eureka Quartzite is in fault contact with the Comus Formation (Plate 1A; Plate 1B). Here, the breccia includes clasts of Comus and Eureka rocks. Further faulting within the Comus Formation produced a wedge of the subunit Ocl D in the hanging wall and Ocl S_1 in the footwall (Plate 1A). This strand of the fault was not exposed, but the relationship supports the interpretation that the Pumpernickel fault is a normal fault, based on the juxtaposition of a younger Comus subunit in the hanging wall.

The Pumpernickel fault represents east-west extension. There is no evidence to suggest that the fault has had recent movement; the fault surface is covered by alluvium, and there are no hallmarks of recent movement on a normal fault, such as triangular facets. Because the Pumpernickel fault bounds a basin, it is likely that its formation was a result of Basin and Range-style faulting.

DEFORMATIONAL HISTORY

Detailed mapping, a revised understanding of the internal stratigraphy of the Comus Formation, and structural analyses have resulted in the identification of at least six deformational events at Iron Point. Two fold sets, the King fault, the Silver Coin thrust, the West fault, and the Pumpernickel fault record these six deformational events. The following section summarizes each event in order of formation, from oldest to youngest.

D_1 event (mid-Pennsylvanian)

The oldest deformational event at Iron Point, D_1, is recorded by the north-trending F_1 folds in the Comus Formation. The F_1 folds are open to closed, asymmetric, and locally overturned to the west (Plate 1B). F_1 folds have an average wavelength of ~2600 feet (~800 meters). A steeply east-dipping (60-80°), north-south striking, well-developed axial planar cleavage to these folds is best observed in the upper units of the Comus Formation. F_1 folds record an east-west, west-directed, shortening event.

D_1 deformation at Iron Point is interpreted to have occurred during mid-Pennsylvanian time. The timing of the development of the F_1 folds at Iron Point is based on the well-constrained timing of the development of west to southwest –vergent folds at Edna Mountain (Villa, 2007; Cashman et al., 2011). F_1 folds at Edna Mountain are tight to closed, asymmetric, and locally overturned to the west or southwest; they are developed in early to mid-Pennsylvanian rocks, and are unconformably overlain by late Pennsylvanian rocks (Villa, 2007; Cashman et al., 2011). The F_1 folds identified at Edna

Mountain are of a similar geometry and style to the F_1 identified at Iron Point, so are interpreted to have formed during the same event.

D_2 event (early Permian)

The second deformational event D_2 at Iron Point was recorded by F_2 folds in the Comus Formation. F_2 folds at Iron Point are open and upright; their axes trend west-southwest (~303°) and plunge gently (~10) (Figure 29). F_2 folds have an average wavelength of ~2600 feet (~800 meters). F_2 folds record north-south shortening. Stereographic analysis of these folds suggests southward vergence, however, due to the paucity of data, this is only a tentative interpretation.

F_2 folds are interpreted to have formed during early Permian time, based on the well-constrained timing of the development of F_3 folds at Edna Mountain (Villa, 2007; Cashman et al., 2011). F_3 folds refold the F_1 folds at Edna Mountain, making a dome and basin map pattern; this affects only the Pennsylvanian and older units, but not the middle Permian or younger units (Villa, 2007; Cashman et al., 2011). The F_3 folds identified at Edna Mountain are the same geometry and style to the F_2 folds identified at Iron Point, so are interpreted to have formed at the same time.

D_3 event (post-early-Permian)

The King fault represents the third deformational event (D_3) at Iron Point. The King fault is a normal fault that strikes approximately north-south (~010°) and dips moderately (~60°) east (Figure 30; Plate 1B). There are no physical exposures of the King fault in the Comus Formation at Iron Point, so direct analysis of the fault surface or kinematic indicators is not possible. However, the map pattern shows the orientation and the stratigraphic repetition records the normal offset.

The King fault represents east-west extension. It post-dates F_1 and F_2 folds and is truncated by the Silver Coin thrust. Since the King fault offsets F_2 folds, it is, at its oldest, post-early Permian. However, it is impossible to determine the youngest age since the King fault is exposed only in the Comus Formation at Iron Point. There are not similar structures to the King fault at Edna Mountain for comparison.

D_4 event (late Permian?)

The Silver Coin thrust represents the fourth deformational event (D_4) at Iron Point. At present, the Silver Coin thrust is a gently south-dipping (~7-10°) fault with an east-northeast (074°) strike. Its sub-horizontal orientation is consistent with a thrust fault, but its slip history is unknown. There are no physical exposures of the thrust surface, so direct analysis of the fault surface or kinematic indicators is not possible.

The Silver Coin thrust represents a shortening event. The Silver Coin thrust truncates F_1 and F_2 folds and the King fault and is therefore younger than post-early Permian. The Golconda thrust is documented within 1 km to the south, southwest and west of Iron Point, and is the only known thrust in the area that is younger than the early Permian. It is likely that the Silver Coin thrust represents a piece of a larger thrust fault, possibly the Golconda thrust, that has been truncated by the subsequent high-angle West fault and the Pumpernickel fault.

D_5 event (post-late-Permian)

The West fault represents the fifth deformational event (D_5) at Iron Point. The West fault strikes, on average, southeast (~168°) and dips moderately southwest (~58°). The West fault is well exposed for a length of approximately 120 meters along the west-central edge of the study area, and is covered by basalt talus to the north and alluvium to the south (Figure 33). It has been interpreted to be a to be a reverse fault, based on its current orientation and juxtaposition of strata (Erickson and Marsh, 1974a). However, this interpretation results in relationships that are inconsistent with the results of this study (see "Discussion", below). The planar surface of the fault suggests it has been reactivated many times, and striae measured in this study (Figure 34) document several slip directions. In addition, the stratigraphic juxtaposition across the fault requires significant strike-slip movement across it (see "Discussion", below).

The West fault truncates the Silver Coin thrust, and is interpreted to post-date the Golconda thrust (late Permian Sonoma Orogeny). However, map relationships along strike of the West fault south of the study area (Figure 34; Plate 1) do not show a fault that truncates the Golconda Allochthon along strike from the exposure of the West fault. This area was outside of the scope of this study, and requires re-evaluation to determine whether or not the West fault affects rocks in the Golconda Allochthon. There are no similar structures that correlate to the West fault at Edna Mountain, or anywhere else in the region, so the only age constraint on the West fault is that it is post-late Permian.

D_6 event (Post-Pliocene)

The Pumpernickel fault represents the last deformation event (D_6) recorded at Iron Point. The Pumpernickel fault is thought to be a normal fault that strikes north-south and dips steeply east, with Ordovician Eureka Quartzite and Comus Formation in the footwall and likely Pliocene basalt in the hanging wall (Plate 1B). There are no physical exposures of the fault at Iron Point due to onlapping Quaternary alluvium, so direct analysis of the fault surface is not possible. South of the study area, the previous mapping shows that the Pumpernickel fault offsets Pliocene-age basalt (Erickson and Marsh, 1974a).

The Pumpernickel fault represents east-west extension. There has been significant offset on the fault, based on the juxtaposition of middle Ordovician Eureka Quartzite in the footwall, and possible Pliocene basalt in the hanging wall (Erickson and Marsh, 1974a). Because the Pumpernickel fault bounds a large basin (Pumpernickel Valley), it is likely that its formation was a result of Basin and Range-style faulting.

DISCUSSION

Stratigraphy

The following section discusses the contributions of this study to the understanding of the stratigraphy at Iron Point, and the resulting implications for the regional geology. The topics include: internal stratigraphy and revised age control on the type Comus Formation at Iron Point; reinterpretation of the uppermost unit in the Comus Formation at Iron Point; and the revised understanding of how the Comus Formation at Iron Point fits into the current understanding of Paleozoic stratigraphy along the western margin of North America.

This study shows that at Iron Point, the Comus Formation is composed of siliciclastic and carbonate-rich rocks. The measured section is carbonate-dominated at the base, with a gradual increase in quartz sand up-section. The uppermost units contain significant very fine-grained quartz sand. Based on this sequence, the Comus Formation likely represents deposition on a continental shelf. The increase in siliciclastic material up-section could indicate a prograding shoreline and shelf, or a change in sediment source.

This study also revised the biostratigraphic age of the Comus Formation at Iron Point. The graptolites that were originally given an age of middle Ordovician (Ross, unpublished data, 1971) have been reclassified as late Ordovician (Finney et el., 1991; Finney et al., 1993).

In the Osgood Mountains, the "Comus Formation" is also a sequence of siliciclastic and carbonate strata (Hotz and Willden, 1964). However, it is considerably thicker and contains more carbonate strata than the Comus at Iron Point, and does not exhibit an increase in quartz sand up-section (Hotz and Willden, 1964). Furthermore, the "Comus" in the Osgood Mountains also includes intervals of basalt that have been dated at about 500 Mya (Briet et al., 2005). There are no volcanic rocks in the Comus Formation at Iron Point. The "Comus Formation" in the Osgood Mountains is interpreted to record a carbonate seamount setting (Breit et al., 2005; Cook, personal communication, 2013).

The "Comus Formation" of the Osgood Mountains was dated using graptolites and one trilobite mold (Hotz and Willden, 1964). These fossils yielded an age of late Cambrian to late Ordovician. This age range overlaps with the Comus Formation at Iron Point, but represents deposition over a much larger span of time (Figure 36).

In summary, although the Comus Formation at Iron Point and the "Comus Formation" in the Osgood Mountains share some parallel lithologic characteristics, they represent two separate units. The ages and depositional environments are too different to classify these two units as the same. Since the type section of the Comus Formation is at Iron Point, it should retain the name "Comus Formation" of late Ordovician in age. The "Comus Formation" in the Osgood Mountains should therefore be called by a different name, as suggested by Madden-McGuire and Marsh, (1991) and Cook, (personal communication, 2013). This will be difficult, since the term "Comus" is used in the Osgood Mountains by various mining companies in Nevada and has come to be known as a unit that is favorable for Carlin-style gold deposition. The regional relationship of the "Comus Formation" in the Osgood Mountains requires future reevaluation, but that is outside the scope of this study.

This study supports the interpretation that the Comus Formation at Iron Point is correlative to the Ordovician Hanson Creek Formation (Cook, personal communication, 2013). Based on the new designation of the Comus Formation at Iron Point as late Ordovician, these two units overlap in age (Figure 36). The Hanson Creek Formation represents predominately carbonate deposition on the continental shelf, but the uppermost unit is a quartz sand-rich dolostone (Mullens and Poole, 1972). Carbonate content of the uppermost Comus Formation at Iron Point was not quantified in this study, but the units Ocss and Ocus do contain considerable quartz sand. Additionally, one of the upper units of the Hanson Creek Formation is known for containing black chert nodules; the Ocus unit of the Comus Formation at Iron Point also contains chert nodules (Figure 16).

The new, undated unit, “Oe”, that bounds the Comus Formation on the east side of Iron Point could be correlative to the middle Ordovician Eureka Quartzite (Plate 1A). The Eureka Quartzite underlies the Hanson Creek Formation and its correlative formations, and is the only significant Paleozoic siliciclastic deposit on the continental shelf. Oe is much coarser-grained than the quartzite within the Vinini Formation at Iron Point (Figure 18; Figure 26). The OE quartzite is composed of quartz grains that are homogenous in size. This is very similar to the Eureka Quartzite, which is also a very clean and mature quartzite.

The conglomerate and breccia within Oe are unique, but could still be a part of the Eureka Quartzite. The clasts in the conglomerate and breccia are composed of mudrock or micrite, and look like they could be fragments of the adjacent Comus Formation. There is no preferred orientation to clasts, however some areas show segregation based on clast type and size (Figure 23; Figure 24). The breccia could be a result of movement on segments of the Pumpernickel fault that bound the exposure of Oe. There is no alternative explanation for the clasts of the Comus Formation. There is one occurrence of the Eureka Quartzite that is similar to this; it is in the Nopah Range of southeastern California, described in the background section of this study. These were determined to be cave deposits in the Eureka Quartzite (Regenfuss et al., 1999). If Oe is correlative with the Eureka Quartzite, this is further evidence that the overlying Comus Formation is correlative with the Hanson Creek Formation.

If the Comus Formation is analogous to the late Ordovician-early Silurian Hanson Creek Formation, and Oe is analogous to the middle Ordovician Eureka Quartzite, the rocks at Iron Point are part of the North American Paleozoic continental margin. These rocks are in the footwall of the Roberts Mountains thrust (RMT); the RMT is not present at Iron Point in the subsurface as previously illustrated (Erickson and Marsh, 1974a).

The autochthonous Ordovician stratigraphic section at Iron Point contrasts dramatically with the Pennsylvanian-over-Cambrian unconformity at adjacent Edna Mountain. The rocks of Edna Mountain experienced pre-mid-Pennsylvanian uplift and erosion that did not occur to the rocks at Iron Point, and the two areas must have been juxtaposed along a fault or faults after that time.

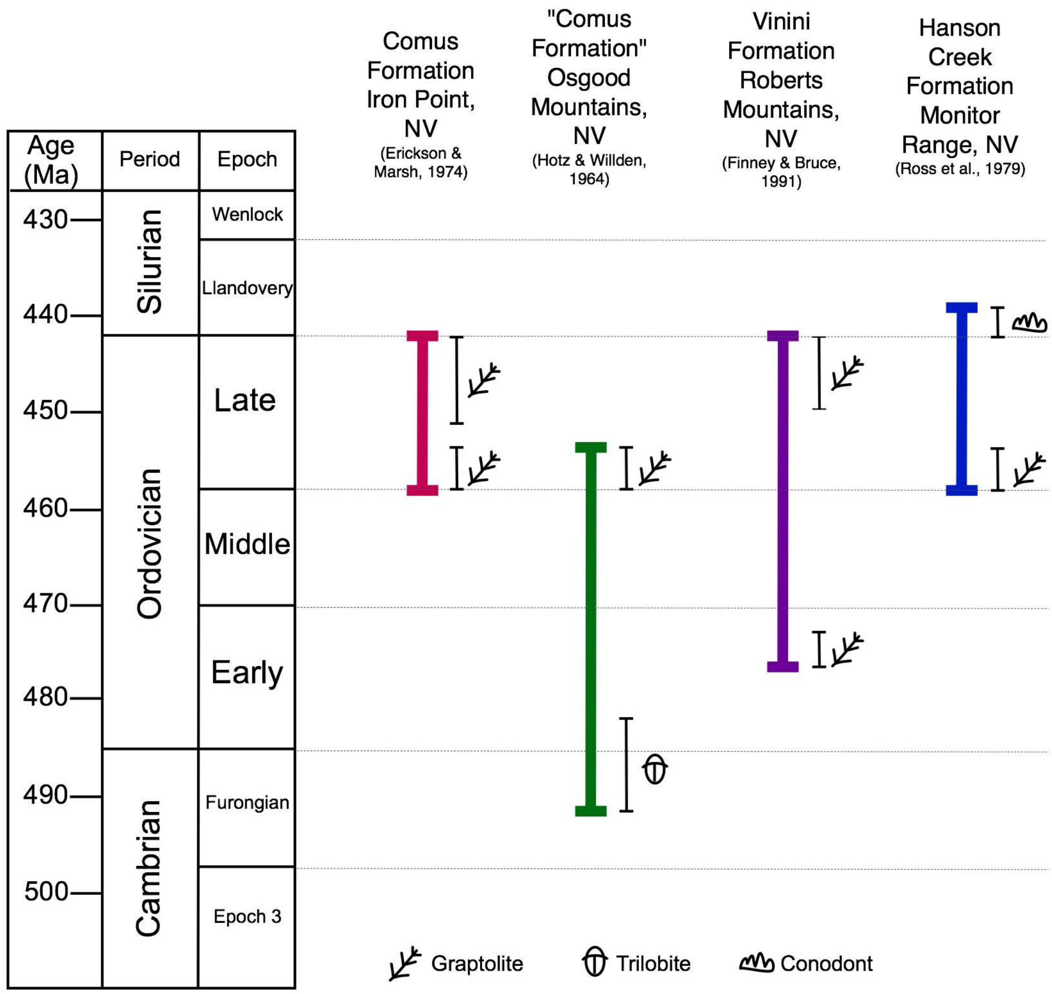


Figure 36. Stratigraphic comparison chart showing the Comus Formation at Iron Point and how it compares to the Comus Formation in Osgood Mountains, the Vinini Formation at its type section at the Roberts Mountains, and the Hanson Creek Formation where it was dated.

Structure

The following section discusses the contributions of this study to the understanding of the structural history at Iron Point. First, the newly recognized structural evolution of Iron Point is summarized. This is followed by structural correlations between Iron Point and Edna Mountain and their implications for regional structural evolution.

New data presented here document late Paleozoic structures at Iron Point. Mid-Pennsylvanian west- to west-southwest vergent folds have has been observed at Edna Mountain (Figure 37) (Villa, 2007; Cashman et al., 2011) and are now also documented at Iron Point. Folds that record early Permian north-south shortening are present at Edna Mountain (Figure 37) (Villa, 2007; Cashman et al., 2011) and have also been documented at Iron Point in this thesis. Based on these observations, Iron Point and Edna Mountain were located close enough to each other during the mid-Pennsylvanian and early Permian to develop these two folds sets.

The King fault is an extensional structure that formed sometime after the early Permian folds. Since the King fault is truncated by the Silver Coin thrust, the east-west extension that created the King fault happened sometime before the emplacement of the Silver Coin thrust. There are no other known post-early Permian extensional features in the region. This suggests that the King fault is not a large-scale regional feature with significant offset. The King fault is more likely a small feature with minor offset, possibly related to extension during or relaxation after east-west shortening.

The coeval Comus and Vinini formations overlap in age and represent two different depositional environments; at Iron Point, they are juxtaposed along the low-angle Silver Coin thrust. Throughout most of Nevada, the Vinini Formation is mapped as part of the Roberts Mountains Allochthon (RMA). Following this convention, the published cross-section of Iron Point shows both the Comus Formation and the overlying thrust fault carrying Vinini (herein called the Silver Coin thrust) to be in the RMA (Erickson and Marsh, 1974a; 1974b). However, this study has determined that the lower plate of the Silver Coin thrust is a continuous sequence of “autochthonous” continental shelf and slope rocks; the Roberts Mountains thrust is not present below Iron Point. Furthermore, The Silver Coin thrust is not Antler–age; it post-dates early Permian structures.

This study has shown that Iron Point and Edna Mountain share a similar structural history - in spite of their different Paleozoic stratigraphic records. The rocks at Iron Point record deposition on the Ordovician continental margin and basin.. In contrast, no Ordovician through Mississippian rocks are preserved at Edna Mountain, where the rocks were uplifted and eroded down to the Cambrian Preble Formation before deposition of mid-Pennsylvanian rocks (Villa, 2007; Cashman et al., 2011). The Edna Mountain and Iron Point areas were subsequently moved next to each other by a fault or faults (see above). Importantly, this study has shown that both areas contain Late Paleozoic folds and subsequent structures. They were therefore in close enough proximity in Late Paleozoic time to be subjected to the same Late Paleozoic continental margin tectonism.

At the onset of this study, one goal was to determine the significance of the West fault, between Iron Point and Edna Mountain; the West fault must have a significant component of strike-slip motion to explain the stratigraphic mismatch. Kinematic indicators on surface of the West fault record multiple events, including some with a strike-slip component. However, these appear to show reactivation of an already-planar fault surface; the slip that juxtaposed Iron Point and Edna Mountain cannot be distinguished based on field relationships or preserved kinematic indicators on the fault surface.

Importantly, there are faults at Iron Point that do not continue across the West fault to Edna Mountain (e.g., the Silver Coin thrust); similarly, there are structures at Edna Mountain that are not at Iron Point (e.g., the low-angle normal fault). Therefore, the juxtaposition of these two areas along the West fault post-dated all of these structures. If the Silver Coin thrust formed during the Sonoma orogeny, as suggested above, final motion on the West Fault must be Triassic or later. Further mapping is warranted, over a much larger area, to determine the extent and slip history of this important fault in the Late Paleozoic continental margin.

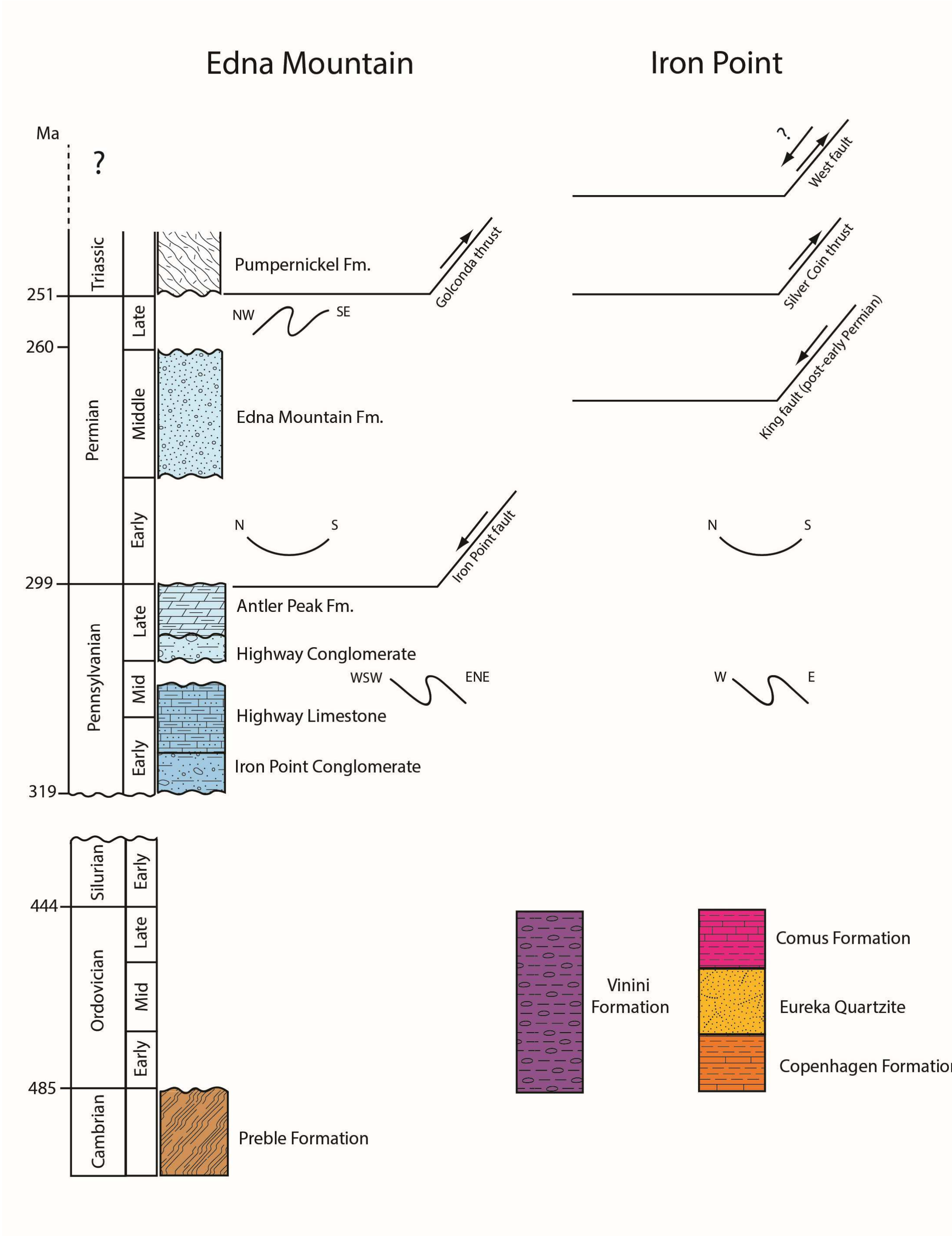


Figure 37. Diagramatic summary of the current understanding of the stratigraphy at Edna Mountain and Iron Point. Note that the timing of the King fault is only constrained to post-early-Permian; its placement on the time scale does not represent timing of formation. Figure is modified from Cashman et al., 2011.

CONCLUSION

This thesis presents new data on the stratigraphy and structure at Iron Point. This study documents evidence for the correlation of the Comus Formation at Iron Point with the late Ordovician-early Silurian Hanson Creek Formation, identifies the middle Ordovician Eureka Quartzite, which had gone previously unrecognized, and documents structures of six different ages, including late Paleozoic structures. Late Paleozoic structures had been documented in the region, but had not been recognized at Iron Point.

The internal stratigraphy of the Comus Formation at Iron Point was mapped in detail during the course of this study. The Vinini Formation is now recognized as stratigraphically distinct and structurally separate from the Comus Formation. A unit previous identified as the Vinini Formation was remapped as the uppermost exposed subunit, Ocus, of the Comus Formation. Finally, a new unit composed of conglomerate, breccia, and quartzite was mapped. This new unit is possibly correlative with the middle Ordovician Eureka Quartzite.

This study presents evidence for six deformational events at Iron Point. During the mid-Pennsylvanian, east-west shortening created north-south-trending folds. These folds are west-vergent to locally overturned, and provide further evidence for a period of tectonic unrest after the Antler Orogeny and before the Sonoma Orogeny along the western margin of North America. During the early Permian, north-south shortening created west-southwest-trending folds. These folds are upright with no clear sense of vergence. This deformation was much less intense than the west-vergent folds during the mid-Pennsylvanian. The north-south-striking King fault offset the earlier folds sometime after the early Permian. Subsequently, the Vinini Formation was thrust over the Comus

Formation along the Silver Coin thrust. All of these Paleozoic structures at Iron Point are truncated by the West fault, which separates Iron Point and Edna Mountain. The initial kinematics of the West fault are unknown, but it has been reactivated numerous times. Finally, Basin and Range-style faulting separated Iron Point from the Pumpernickel Valley to the east along the Pumpernickel fault.

Questions have arisen from this thesis that could guide future research at Iron Point and in the surrounding area. First, in what environment was the conglomerate and breccia within the Eureka Quartzite at Iron Point deposited? Second, what is the geographic extent and offset history of the West fault? Is it covered to the south beneath the Golconda Allochthon? Is there a continuation of the West fault as far north as the Osgood Mountains? What is its structural and tectonic significance of the West fault? Finally, why is the Antler autochthon exposed at the ground surface a significant distance west of the mapped “Roberts Mountain thrust” (Figure 1)? This constitutes a "window" through the thrust, and represents significant uplift in the Iron Point - Osgood Mountains area. Answering these questions will help to better refine our understanding of the Paleozoic paleogeography and tectonic evolution of western North America.

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