3.0 GEOLOGIC CONDITIONS

The SNWA has studied and thoroughly described the stratigraphy, structural geology, and hydrogeology of the study area in a report titled "Geology of White Pine and Lincoln Counties and Adjacent Areas, Nevada and Utah: The Geologic Framework of Regional Groundwater Flow Systems" (SNWA, 2007a). An electronic version of the report has been attached as Volume 1 on the CD-ROM accompanying this report. This section provides an overview of the geologic information presented in that report.

3.1 Geology and Stratigraphy

The geology of the geologic study area (Plates 1 and 2, in Volume 1) is characterized by a thick stratigraphic sequence of rocks from Proterozoic to Holocene age that has been structurally deformed during several tectonic episodes. The thick sequence includes three major assemblages that are important aquifers:

- Carbonate aquifer of Paleozoic age
- Volcanic rocks of Tertiary age
- Basin-fill sediments of Tertiary to Quaternary age.

Along with the aquifers are moderate to thick confining units or low-permeability units, including:

- Early to Late-Proterozoic metamorphic and igneous rocks
- Late Proterozoic to Lower Cambrian quartzite and shale
- Shale, sandstone, and conglomerate of Mississippian age
- Triassic to Cretaceous shale, siltstone, and sandstone
- Mesozoic to Cenozoic plutons.

Three tectonic episodes, plus an intervening episode of extensive volcanism, have affected the hydrogeology of the region. The oldest tectonic episode is the Antler deformation (Late Devonian to Late Mississippian). This episode included east-verging thrust sheets. The second tectonic episode was the Sevier deformation (Jurassic through early Cenozoic) that resulted in east-verging thrust sheets in which Paleozoic carbonate rocks were placed over each other and over younger rocks.

In Eocene to middle Miocene time, volcanism resulted in the development of thick blankets of ash-flow tuff and related lava flows, including many scattered calderas that were the sources of the tuff. The caldera margins formed new groundwater flow paths and barriers.

The third tectonic episode is the middle Miocene to Holocene basin-range deformation that shaped the current topography of the Great Basin, including most of Nevada and parts of western Utah and southeastern California. Basin-range faulting produced graben and horst topography, resulting in deep basins and relatively high mountain ranges, generally oriented north-south. In general, the mountain ranges provide areas of groundwater recharge, and accumulations of alluvial fill within the basins provide areas of aquifer storage and avenues of groundwater flow. Basin-range faults may provide hydrogeologic barriers to groundwater flow. But it has also been interpreted that more commonly, basin-range faults provide conduits to groundwater flow, especially from north to south. These north-south conduits may double as barriers to east or west flow in certain flow systems such as the GSLDFS.

The age of the rocks in the geologic study area is summarized in a Geologic Time Scale chart (Figure 3-1). The oldest rocks are Early and Late Proterozoic metamorphic and igneous units. These rocks are overlain by thick sequences of quartzite and subordinate shale, which are locally metamorphosed to slate and schist, of Late Proterozoic age. The Proterozoic rocks pass conformably upward into rocks of similar type and thickness, though less metamorphosed, that are Late Proterozoic to Early Cambrian in age. During Middle Cambrian time, carbonate deposition was initiated, and thick sequences of marine limestone and dolomite were deposited from the Middle Cambrian through the Permian Periods. These rocks make up the carbonate aquifer of Nevada and adjacent parts of Utah and range in thickness between 5,000 and 30,000 ft throughout this area (Harrill and Prudic, 1998).

Locally, marine sandstone and shale are intertongued with the carbonates. These units are not interpreted to form significant impediments to regional groundwater flow, with the exception of the Chainman Shale and related shale and sandstone of Late Mississippian age. This unit locally exceeds 2,000 ft in thickness, and in all but the southern part of the geologic study area, this unit divides the carbonate aquifer into two distinct aquifers, the lower and upper carbonate aquifers. The Chainman Shale and related clastic units were derived from erosion of a structural highland, the Antler Highland, in and northwest of the geologic study area. The highland, made up in large part of the Roberts Mountain allochthon, was produced by the Antler compressive deformational event.

Mesozoic rocks in the geologic study area are largely clastic, nonmarine, and thin where deposited, but in most places they have been removed by erosion. They and older rocks were deformed during the Sevier deformational event. At this time, the geologic study area was a highland, also known as a hinterland, and an episode of erosion of the area removed most Mesozoic rocks.

Plutons of Late Jurassic to Paleocene age were intruded during pulses during Sevier deformation. These plutons probably had associated extrusive volcanic units, but all of these units have been removed by erosion. Mesozoic plutons commonly led to significant mineralization in the geologic study area.

Middle Tertiary (Eocene to middle Miocene) time marked the beginning of calc-alkaline intrusion and resulting volcanism, the terminal product of subduction beneath western North America that began in the Triassic Period (Atwater, 1970; Lipman et al., 1972; Hamilton, 1995). Above individual source plutons, vent deposits included andesitic and dacitic lava flows and volcanic mudflow breccia that locally exceeded several thousand feet of thickness. Caldera deposits consist of dacitic to rhyolitic ash-flow tuffs, which are similarly thick within individual calderas. Farther outward from the vents above the plutons, lava flows are sparse because they do not flow more than a few miles

ERA	PERIOD	EPOCH	TIME	PROCESSES AND ROCK TYPES
Cenozoic	Quaternary	Holocene Pleistocene	Present 1.8 Ma *5 Ma*	Valley-Fill Alluvium
	Tertiary	Pliocene Miocene Oligocene Eocene Paleocene	*38 Ma*	Start Basin-Range Faulting (20 Ma) Volcanics and Older Sediments Emplacement of Calderas
Mesozoic	Cretaceous Jurassic Triassic		65 Ma	Sevier Orogeny, Intrusions Continental Sediments
Paleozoic	Permian Pennsylvanian Mississippian Devonian Silurian Ordovician Cambrian		248 Ma	Antler Orogeny, Intrusions Chainman Shale, Carbonates
Precambrian			543 Ma	Quartzite and shale
			4.5 Ga	

Source: Adapted from Geological Society of America, 1999

Figure 3-1
Geologic Time Scale, Including Rock Type and Tectonic Events

from their vents, but outflow ash-flow tuffs accumulated to aggregate thicknesses exceeding 1,000 ft in most of the geologic study area.

Starting at about 20 Ma ago (middle Miocene), subduction ceased and extensional deformation increased in the map area (Christiansen and Lipman, 1972; Christiansen and Yeats, 1992; Rowley and Dixon, 2001). Basin-range deformation, characterized by vertical (normal) faulting, began to form alternating mountain ranges and valley basins. The main pulse of this basin-range faulting began about 10 Ma ago, during which time the present topography formed. As valleys formed, they were filled by debris eroded from the adjacent mountain range, creating basin-fill deposits.

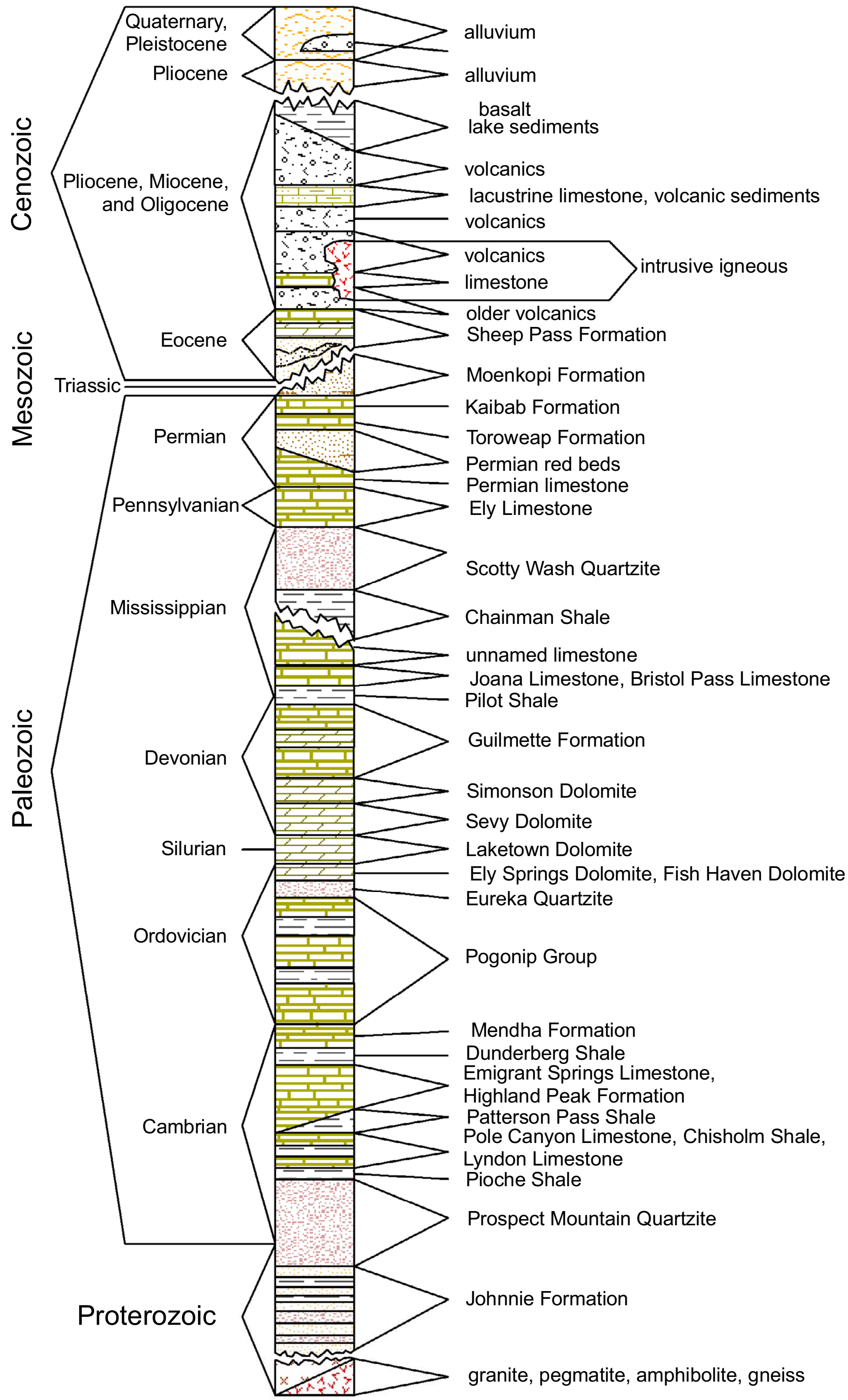
Individual rock units, structures, basins, and ranges are described in detail in Volume 1 on the CD-ROM accompanying this report. Thicknesses of most units are from the county reports of the area where the unit is exposed. The relationships between geologic units in the different areas of the map can be determined from Figures 3-2 to 3-5. These figures illustrate geologic columns for Lincoln (Figure 3-2), White Pine (Figure 3-3), and Clark counties (Figure 3-5), and western Utah (Figure 3-4). The Utah area consists of western Iron, Beaver, and Millard counties and the southwestern corner of Juab County.

3.2 Structural Geology

This section discusses the structural framework of the geologic study area. Three main structural events affected the geologic study area: (1) Late Devonian to Late Mississippian Antler compressive deformation, (2) Late Jurassic to early Tertiary Sevier compressive deformation, and (3) late Cenozoic basin-range extensional deformation. In addition to these structural events, middle Cenozoic time was characterized by mild extension (Rowley, 1998; Miller et al., 1999; Rowley and Dixon, 2001) and voluminous calc-alkaline volcanism that profoundly affected the topography and hydrology of the geologic study area.

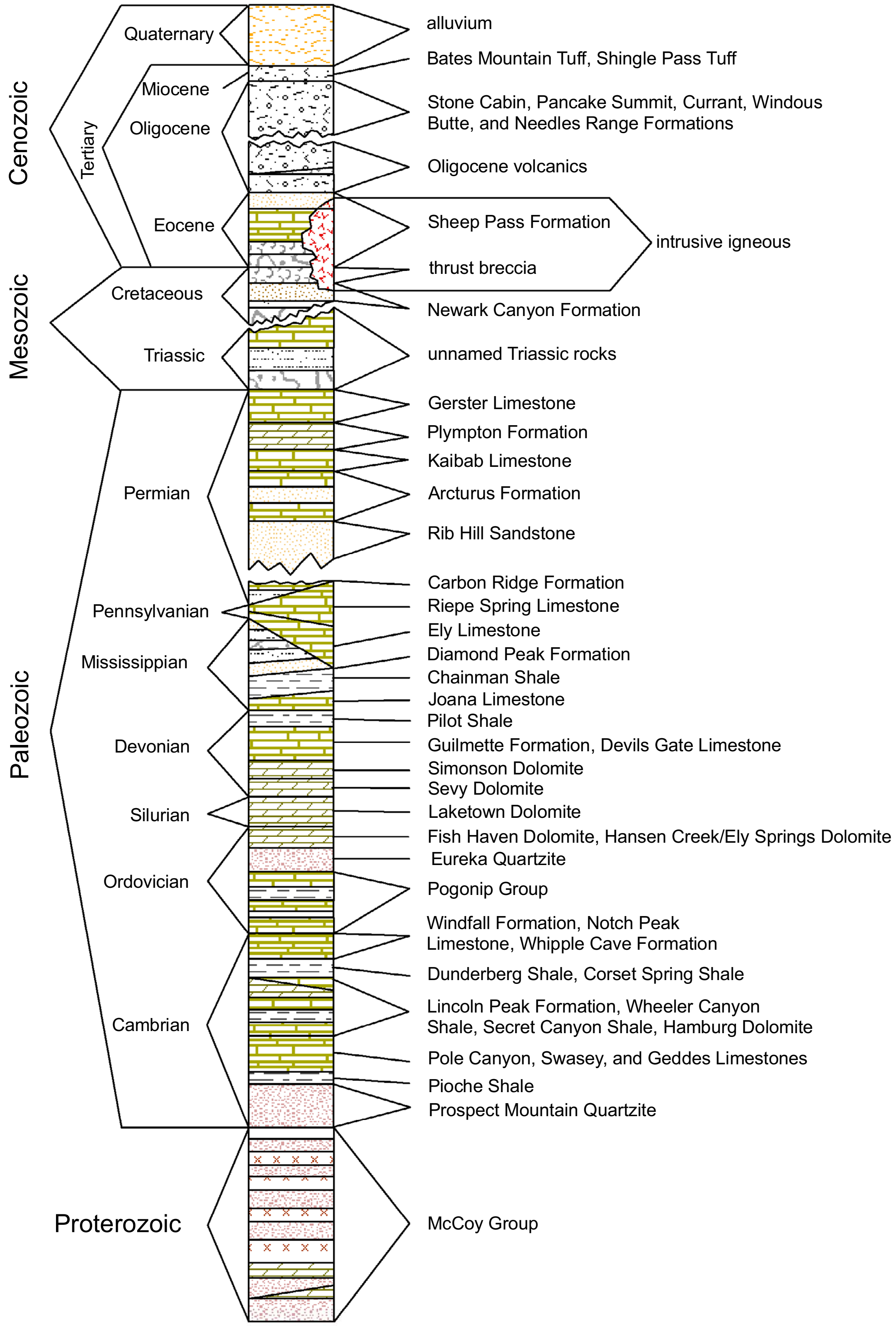
The Late Devonian to Late Mississippian Antler compressive deformation affected the northwestern part of the geologic study area, creating a north-trending highland (Larson and Langenheim, 1979; Carpenter et al., 1994; Poole and Sandberg, 1977 and 1991). This event formed folds and thrusts of the Roberts Mountain allochthon, which was at least 8,000 ft thick and passed through the western side of Eureka, Nevada (Carpenter et al., 1994; Saucier, 1997). The thrusts transported deeper-water sedimentary rocks eastward as much as 100 mi. Coarse synorogenic siliceous clastic detritus was shed from the highland into the foreland basin to the east, transitioning to shale farther east. The main synorogenic rock units that resulted were the Chainman Shale and Diamond Peak Formation, and farther south the Scotty Wash Quartzite.

The second structural event, the Middle Jurassic to early Tertiary Sevier compressive deformation, resulted in generally north- to north-northeast-striking, east-verging folds and thrust faults. Scattered Middle Jurassic to lower Tertiary plutons were emplaced in many mountain ranges of the geologic study area. Eastward-directed overthrusts emplaced Late Proterozoic to middle Paleozoic rocks over Late Proterozoic to Mesozoic rocks (Armstrong, 1968). At least a half dozen large thrusts are well exposed in the Las Vegas area, each with displacements ranging from several to 20 mi (Page et al., 2005b). Tectonic shortening caused by thrusting in southern Nevada is at least 22 to 45 mi (Stewart, 1980; Burchfiel et al., 1974). Except for the southern part of the geologic study area, most of the area



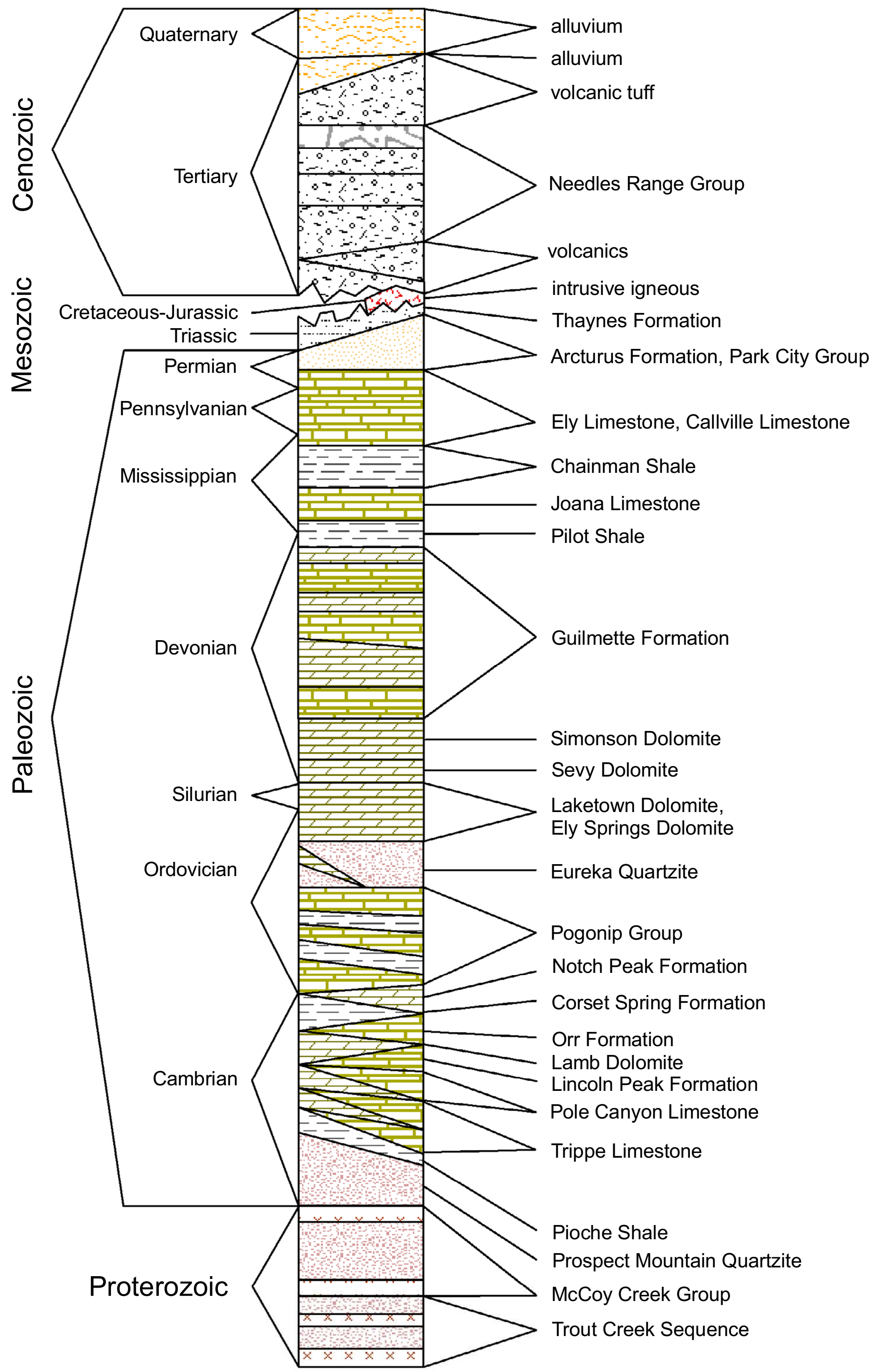
Source: Tschantz and Pampeyan, 1970

Figure 3-2
Geologic Units of Lincoln County, Nevada



Source: Hose and Blake, 1976

Figure 3-3
Geologic Units of White Pine County, Nevada



Source: Hintze, 1988

Figure 3-4
Geologic Units of Western Utah

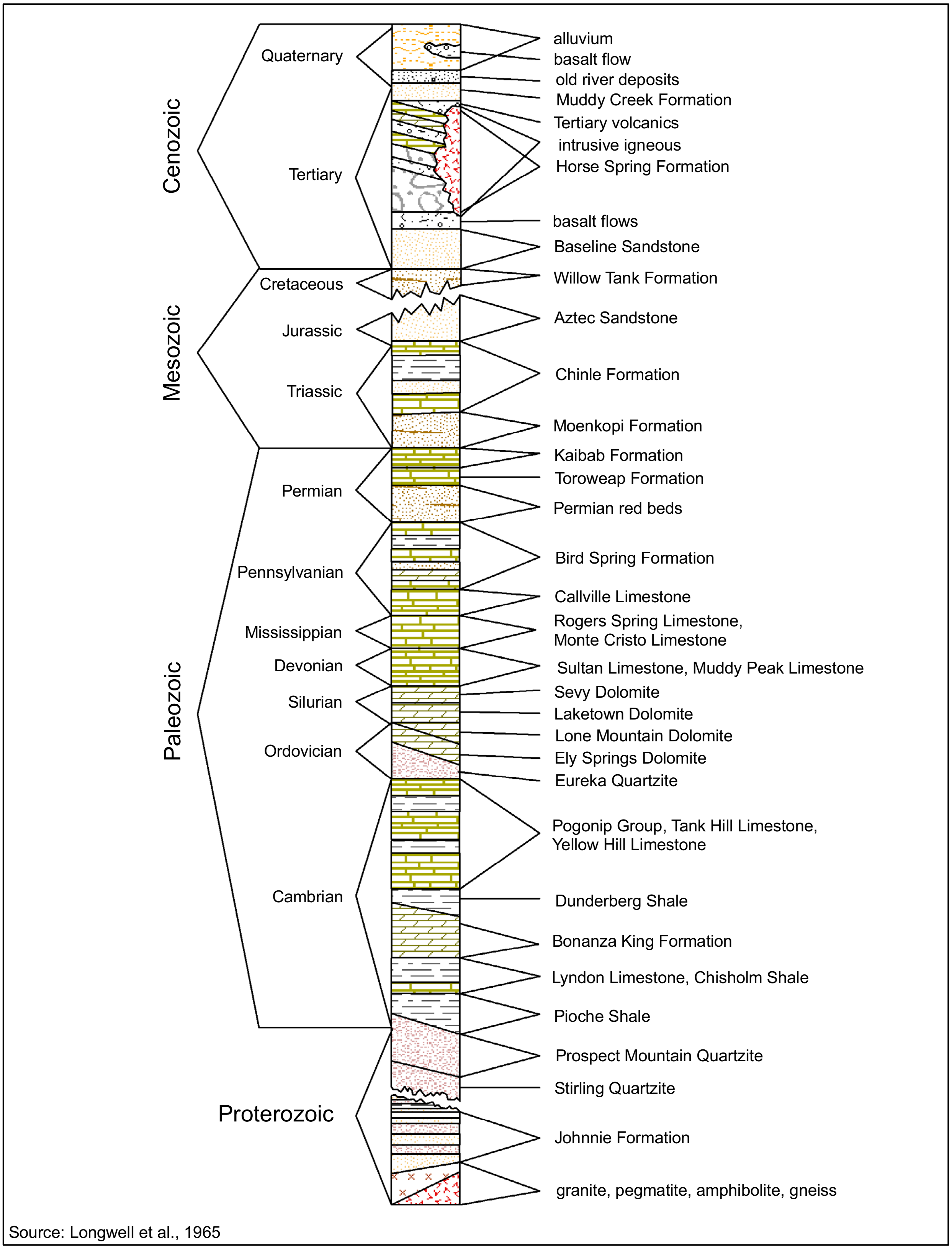


Figure 3-5
Geologic Units of Clark County, Nevada

has been considered to be the western hinterland of the deformation. In other words, the leading edges of most major thrusts are east of the map area, and the deformation created highlands within the hinterland of the map area that in turn eroded and shed clastic material primarily to the east. Some of the thrusts, including the Gass Peak, however, have been projected northward into the hinterland in the central and northern part of the geologic study area, including the Timpahute Range, Worthington Mountains, Golden Gate Range, Grant Range, Pancake Range, and Newark Valley (Vandervoort and Schmitt, 1990; Dobbs et al., 1994; Taylor et al., 2000). Sevier-type deformation is shown schematically on Figure 3-6, and the Sevier-age Glendale/Muddy Mountains thrust in the Muddy Mountains is shown on Figure 3-7.

East-striking faults and folds, alignments of plutons and volcanic vents, alignments of geophysical anomalies, local alignments of basins and ranges, hot springs, hydrothermally altered rocks, and mineral deposits have been noted in the Great Basin for years, primarily by geologists of the mining industry. Ekren et al. (1976 and 1977), Rowley et al. (1978), and Stewart et al. (1977) called these alignments "lineaments" with an origin similar to transform faults in the ocean basins. Ekren et al. (1976) also suggested that the lineaments began to form in the Cretaceous, if not earlier, and continued to be active throughout both Tertiary calc-alkaline magmatism and basin-range deformation. Like transform faults, these lineaments seem to represent boundaries between areas to the north and south that had different amounts, rates, and types of structural deformation. Rowley (1998) and Rowley and Dixon (2001) referred to them as transverse zones, and we follow their terminology here. They are poorly known and have been mapped in detail only locally, so they are projected with limited evidence between these areas where they are known. Therefore, transverse zones are delineated as speculative zones of potential disruption on Plates 1 and 2 in Volume 1.

Transverse zones bound parts of most igneous belts in the Great Basin. They also define the northern and southern sides of the Caliente caldera complex, representing structures by which this caldera spread east and west to a degree much more profound than most other calderas in the Great Basin. Transverse zones may both provide barriers to the southward flow of groundwater and act as conduits to east or westward flow of groundwater (Prudic et al., 1995; Rowley, 1998; Rowley et al., 2001).

The third structural event, the basin-range episode of extensional deformation, began at about 20 Ma and continues today. It is characterized by east-west extension and resulted primarily in north-striking normal faults. Over some parts of the Great Basin, early phases of this deformation produced north-striking basins and ranges due partly to gentle folding. Sediments were deposited in basins formed by these early faults and broad warps, but these basins were not necessarily in the same locations as they are today. The present topography was produced later, during the main pulse of basin-range deformation that began after 10 Ma for most parts of the Great Basin. The axes of basins and ranges since 10 Ma were commonly different from those created during the early phase of deformation. Some parts of the older basins were uplifted as part of the new ranges and some parts of the older ranges were downthrown as part of the new basins. An example is the presence of Miocene lacustrine limestones and associated clastics in the North Pahroc and Pahranagat ranges (Tschanz and Pampeyan, 1970) that were originally deposited in one or more basins.

The dominant fault type since major deformation began (about 10 Ma) continued to be north-striking normal faults, but locally strike-slip and oblique-slip faults accommodated the east-west extension. Examples of such accommodation zones are the east-northeast, left-lateral Pahranagat shear zone at

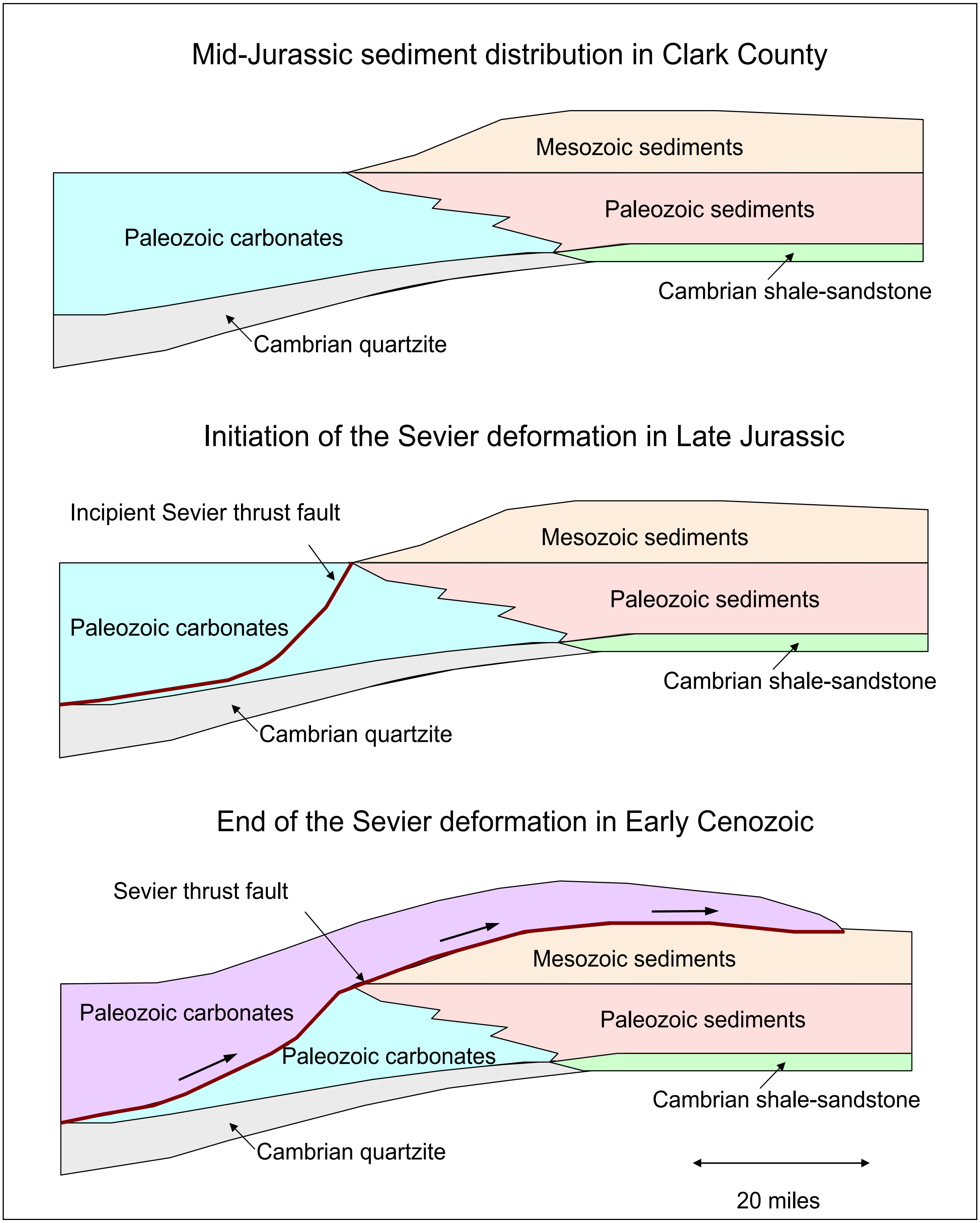
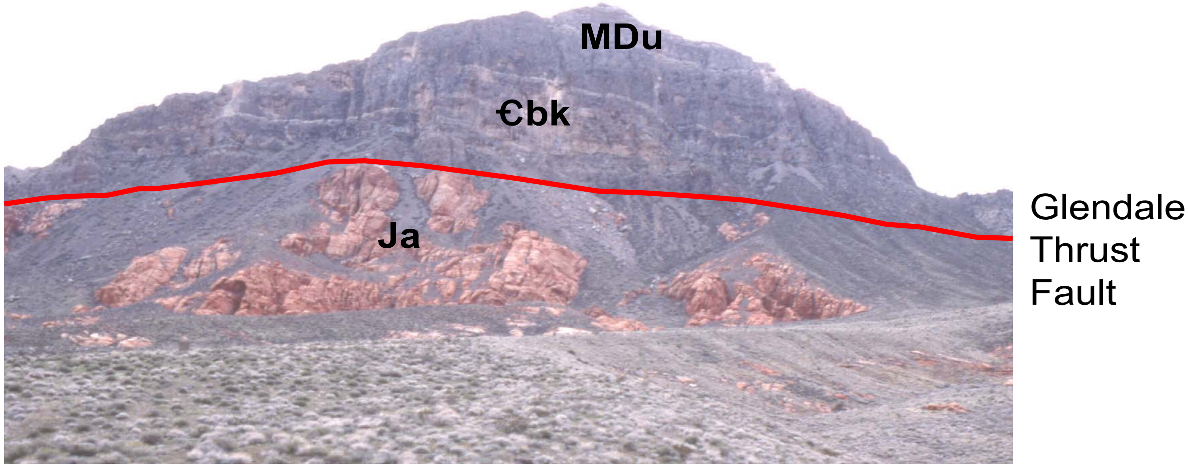


Figure 3-6
Schematic Diagram of Sevier Thrust Sheets, Illustrating the Movement of Paleozoic Carbonates over Cratonic Sediments



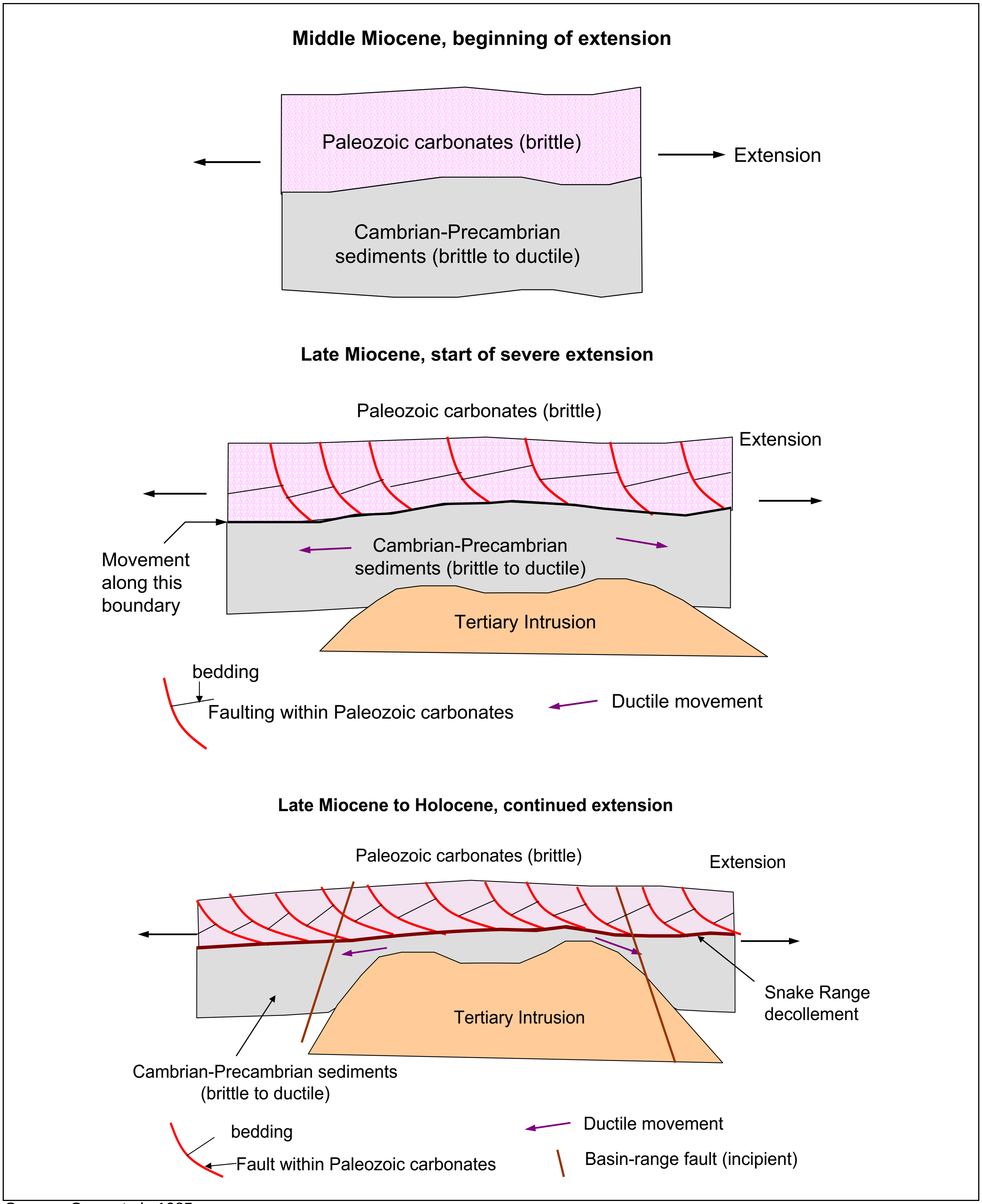
Ja – Jurassic Aztec Sandstone
MDu – Mississippian-Devonian carbonates undifferentiated
€bk – Cambrian Bonanza King Formation

Figure 3-7
Paleozoic Carbonates Thrust over Jurassic Aztec Sandstone in the Muddy Mountains near Muddy Peak

the southern end of Pahranagat Valley and the northeast trending, left-lateral Kane Spring fault zone west of the Meadow Valley Mountains (Ekren et al., 1977). East-striking transverse faults continued to be active at the same time, segmenting the Great Basin into broad east-trending corridors of different types and amounts of east-west pulling apart.

In some parts of the map area, low-angle faults were previously mapped as thrust faults. These faults, however, place younger rocks on older rocks. In some places, the direction of movement of the upper plates of these faults is westward rather than eastward. Most of these faults are considered to be much younger, Tertiary in age, expressions of structural extension and formed during the basin-range deformational event. The faults are interpreted to be detachment faults, although the general synonyms "attenuation" or "denudation faults" that were used by some early workers who first recognized them (Moores et al., 1968; Armstrong, 1972) are more appropriate in places where many subhorizontal faults are present, notably the Eureka area. In these areas, rapid uplift of ranges resulted in their tops being structurally stripped (or attenuated or denuded) by low-angle faults that verged into the adjacent low areas, much like large gravity slides.

One major fault to which the name "detachment" fault is appropriate is the well-known Snake Range decollement. Although originally considered to be a thrust fault that placed Middle Cambrian and younger rocks over Middle Cambrian and older rocks (for example, Nelson, 1966), the fault was later mapped in greater detail and reinterpreted as an Eocene to middle Miocene low-angle fault caused by stretching and thinning during uplift of a metamorphic core complex (Miller et al., 1983; Gans et al., 1985 and 1989). This detachment may represent the ductile/brittle transition zone uplifted by the core complex (see Figure 3-8 Miller et al., 1983; Gans et al., 1985; Gans, 2000). Rocks have been thinned



Source: Gans et al., 1985

Figure 3-8
One Scenario for Development of the
Snake Range Decollement during Late Cenozoic Extension

by the elimination of strata due to the faulting. Later work indicated that, while the decollement had an older (late Eocene and early Oligocene) history, most displacement on it was middle Miocene and later, coinciding with basin-range deformation (Miller et al., 1999). The core-complex uplift that formed the decollement included the Kern Mountains and southern Deep Creek Range (Miller et al., 1999). Finally, Miller et al. (1999, p. 902) suggested that the Snake Range decollement may not be a normal fault at all but instead a "highly complex structural boundary developed above a rising and extending mass of hot crystalline rocks."

3.3 Geology of the Project Basins

The geology of the Project Basins is summarized in the following sections.

3.3.1 Spring Valley

Spring Valley is an approximately 100-mi-long broad deep graben within the GSLDFS. At its northern end, Spring Valley contains about 2,000 ft of basin-fill sediments. The amount of basin-fill increases in the south to as much as 6,000 ft of material.

In the north, the Antelope Range separates Spring Valley from Tippett Valley. The Antelope Range is a relatively small, low range of faulted, mostly Tertiary volcanic rocks that unconformably overlie mostly west-dipping Silurian to Permian sedimentary rocks, dominantly carbonate rocks.

On the west, the approximately 120 mi long Schell Creek Range bounds the valley, separating it from Steptoe Valley to the west. The northern part of the Schell Creek Range is made up of a west-dipping sequence of Precambrian through Permian rocks (Lumsden et al., 2002), with overlying Tertiary volcanic rocks along the faulted western flank of the range. Small Tertiary intrusions are exposed locally along the range.

On the southwestern side of Spring Valley, the Fortification Range separates the valley from Lake Valley to the southwest. The Fortification Range is a series of faulted, upper Paleozoic carbonate rocks including, at the northern end, a narrow, low, north-northwest-trending, northeast-dipping cuesta that joins the eastern side of the Schell Creek Range. Despite the small size of the Fortification Range, it is complexly faulted and contains repeated sections of the Mississippian Chainman Shale beneath the surface, probably more than 1,000 ft thick.

To the east, the broad, high, north-trending Snake Range separates Spring Valley from Snake Valley to the east. It contains Wheeler Peak, more than 13,000 ft high and within Great Basin National Park. The range is about 65 mi long, nearly all of it in White Pine County, but with the low southern end in Lincoln County. The range is a horst, bounded on both sides by major high-angle normal fault zones. Except for the southern end, the Snake Range is cored by Late Proterozoic to Cambrian quartzite, intruded by a massive batholith of apparent Jurassic age (Whitebread, 1970; Miller et al., 1994 and 1995). The range is a metamorphic core complex, which rose rapidly and formed the Snake Range decollement, a detachment fault (Miller et al., 1999). This low-angle Tertiary detachment formed over an extended period as the range uplifted and stretched the roof rocks apart (Gans, 2000). The fault places complexly faulted Middle Cambrian carbonate and younger rocks over a lower plate of

Middle Cambrian carbonate rocks, Lower Cambrian clastic rocks, and older rocks. The decollement is exposed on the top and eastern side of the northern half of the range. The central part of the Snake Range is narrower and lower. On the eastern side of Sacramento Pass, north-striking, down-to-the-east listric normal faults drop down a thick section of Tertiary volcanic and basin-fill rocks (Gans et al., 1989; Miller et al., 1994, 1995, and 1999). The southern end of the range consists of south-plunging tilt blocks of Paleozoic rocks as young as Mississippian. These tilt blocks become lower in elevation to the south, and the eastern tilt blocks plunge beneath the valley fill. The western tilt blocks continue southward to become the Limestone Hills, which consist mostly of east-dipping Devonian carbonate rocks bounded by normal faults on the western and eastern sides.

3.3.2 Snake Valley

Snake Valley, east of the Snake Range, is a long, broad, deep graben containing about 5,000 ft of basin-fill deposits but local holes in the basin contain thicker deposits than this (Allmendinger et al., 1983; Saltus and Jachens, 1995; Kirby and Hurlow, 2005). The Deep Creek Range makes up the northwest side of Snake Valley and is a horst, bounded by north-striking normal faults on either side. The Snake Range is located on the western side of the valley, and was described in the previous section. Snake Valley is bounded on the eastern side by the Confusion Range. The Confusion Range and small ranges of similar rocks form the entire eastern (Utah) side of Snake Valley. The area includes hills (Middle Range) connected to and east of the northern end of the Confusion Range. The Confusion Range proper is 60 mi long, with a general northerly trend. The Conger Range is a 15-mi-long, southwest-diverging fork in the southern Confusion Range. Except for the southern portion of the Confusion Range, the Mississippian Chainman Shale, 1,000 to 2,000 ft thick in the area, is repeated and thus exposed on both sides of and beneath all these ranges because it is deformed into north-striking folds (Hintze and Davis, 2002a and b, and 2003).

3.3.3 Cave Valley

Cave Valley is part of the WRFS and consists of two distinct but connected portions, separated by an oblique-slip fault at Shingle Pass. One of these portions, northern Cave Valley, is a narrow graben with mostly east-dipping Cambrian rocks at shallow depth and containing relatively thin basin-fill sediments. The southern portion, south of Shingle Pass, generally contains less than 3,000 ft of basin-fill sediments and volcanic rocks but in a narrow, central, north-trending axial part, these Cenozoic rocks are 6,000 ft or more thick.

Along the western side of Cave Valley, the Egan Range is a complexly faulted horst of east-dipping Cambrian to Permian rocks, overlain by Tertiary volcanic rocks. The Egan Range separates Cave Valley from White River Valley to the west. Halfway southward down Cave Valley a northeast-striking oblique-slip fault passes through the Egan Range at Shingle Pass. This fault has effectively partitioned the valley into two sub-basins by displacing a section of the Egan Range and forming an east-dipping tilt block that extends northeast across and beneath Cave Valley where it terminates against the range-front fault of the Schell Creek Range. Based on oil test well drilling and gravity surveys, this block contains the Mississippian Chainman Shale (Hess, 2004; Mankinen et al., 2006; Scheirer, 2005).

Farther south, the Egan Range remains an east-tilted horst of Cambrian through Tertiary rocks, then bends southeast to join the southern end of the Schell Creek Range. Here Cave Valley terminates where the Egan and Schell Creek ranges join each other in a complex of north-northeast- and north-northwest-striking normal and oblique-slip faults.

To the east, the Schell Creek Range separates Cave Valley from Lake and northern Dry Lake (Muleshoe) Valleys. This section of the Schell Creek Range contains a narrow, heavily faulted sequence of Precambrian through Tertiary rocks that dips east. Here the dominant fault is on the western flank of the range. In the northeast portion of the valley, the Schell Creek Range is cored by Precambrian to Cambrian quartzite. West of the Geyser Ranch, the rocks are mostly Late Proterozoic and Cambrian quartzite (Van Loenen, 1987), but farther south the rocks are dropped down along an east-trending fault at Patterson Pass and are mostly of middle to upper Paleozoic and Tertiary age. In the south, where Cave Valley terminates at the intersection of the Schell Creek Range and the Egan Range, a Tertiary pluton has mineralized adjacent carbonate rocks at the Silver King Mine.

3.3.4 Dry Lake Valley

Dry Lake Valley is a deep graben that contains in most places 3,000 to 5,000 ft of basin-fill sediments (Mankinen et al., 2006), but locally along the axis of the graben as much as 10,000 to 25,000 ft of sediments and underlying downfaulted volcanic and carbonate rocks are present (Scheirer, 2005).

Along the western side of Dry Lake Valley at the junction with the southern Egan and Schell Creek ranges, the North Pahroc Range extends south for about 40 mi separating the valley from Pahroc Valley to the West. The North Pahroc Range consists of upper Paleozoic rocks overlain by Tertiary volcanic rocks. These rocks dip west off major faults along the eastern side of the range. The North Pahroc Range is separated from the smaller South Pahroc Range at the southern end of the valley by an east-trending belt of faulted rocks of low relief formed by the east-striking Timpahute transverse zone. The belt of faulted rocks is the boundary between Dry Lake Valley and Delamar Valley to the south. The Seaman and the North Pahroc ranges join together at their southern ends, and the Hiko Range continues south of this intersection.

On the east side and from north to south, the Fairview, Bristol, Highland, and Chief Ranges are a 60-mi-long group of north-trending, heavily-faulted ranges of mostly east-dipping rocks that separate Dry Lake Valley from Lake, Patterson and Panaca Valleys to the east. The northern portion of Dry Lake Valley (Muleshoe Valley) is bounded on the east by the Fairview Range. The Fairview Range is a horst made up of Devonian to Pennsylvanian rocks at both the northern and southern ends of the range. The central part of the range consists of the western lobe of the Indian Peak caldera complex. To the south, the valley is bounded on the east by the Bristol, Highland, and Chief ranges. A low pass between the Fairview Range and the Bristol Range is cut by numerous east-striking faults of the Blue Ribbon transverse zone, which crosses the entire Great Basin at about this latitude (Rowley, 1998; Rowley and Dixon, 2001). The small horsts due west of this area are named the West, Ely Springs, Black Canyon, and Burnt Spring Ranges.

The Bristol Range is a horst that consists mostly of an east-dipping sequence of Cambrian carbonate rocks. The range is cored by a Tertiary pluton on the northern end that is associated with silver deposits of the Jackrabbit and Bristol districts. A low angle, west-dipping detachment or gravity-slide fault that placed Devonian rocks on Cambrian rocks is exposed in the northwestern part of the range (Page and Ekren, 1995). The Highland Range, the southward continuation of the Bristol Range, consists of east-dipping Cambrian carbonate rocks, underlain by Precambrian and Cambrian quartzite. A west-dipping, west-verging, moderately dipping fault on the western side of the range, the breakaway part of the Highland detachment fault, placed the younger carbonate rocks on the older quartzite. The Chief Range, south of the Highland Range, is made up of east-dipping Precambrian and Cambrian quartzite that is unconformably overlain by Tertiary volcanic rocks and cut by a Tertiary pluton that controls the small Chief gold district. The faults that lift the range on the western side consist of an oblique-slip fault (right lateral and normal) and the west-dipping Highland detachment fault (Rowley et al., 1994).

The small West Range consists of Devonian sedimentary rocks and Tertiary volcanic rocks on which are Devonian rocks emplaced by a low-angle fault that can be interpreted as either a detachment fault or a gravity-slide plane (Page and Ekren, 1995). The Ely Springs Range consists of Cambrian through Silurian rocks, overlain by Tertiary volcanic rocks. The Black Canyon Range consists of Cambrian sedimentary rocks and Tertiary volcanic rocks. The Burnt Springs Range consists of Cambrian sedimentary rocks unconformably overlain by Tertiary volcanic rocks.

3.3.5 Delamar Valley

Delamar Valley, just south of Dry Lake Valley, is a graben that deepens to the south with a general maximum thickness of more than 5,000 ft of basin-fill sediments east of the South Pahroc Range (Mankinen et al., 2006). Locally as much as 20,000 ft of sediments and underlying downfaulted volcanic and carbonate rocks are present (Scheirer, 2005).

Along the western side of the valley, the South Pahroc Range extends southward from the North Pahroc Range, separating the valley from Pahranagat Valley to the west. The South Pahroc Range is a series of west-titled blocks of volcanic rocks; the main faults are on the eastern side of the range. The South Pahroc Range terminates against the east-northeast-trending Pahranagat shear zone, which also terminates Pahranagat and Delamar Valleys at their southern extent.

Along the eastern side of the valley, the Delamar Mountains bound the eastern side of the valley, extending southward for 40 mi from the Burnt Springs Range in southern Dry Lake Valley. The boundary between the two ranges can be placed at the northern caldera wall of the Caliente caldera complex, here controlled by the east-trending Timpahute transverse zone (Ekren et al., 1976; Rowley, 1998; Swadley and Rowley, 1994). The Delamar Mountains consists of east-dipping Late Proterozoic to Cambrian rocks and Tertiary volcanic rocks. The range, however, is dominated by Tertiary caldera complexes. The western end of the Caliente caldera complex is in the northern part of the range, and the Kane Springs Wash caldera complex is in the central part of the range (Rowley et al., 1995; Scott et al., 1995 and 1996). The main bounding fault of the Delamar Mountains is the down-to-the-west normal fault on the western side, and this is joined from the southwest by several splays of the left-lateral Pahranagat shear zone (Ekren et al., 1977).

3.3.6 Coyote Springs Valley

Coyote Spring Valley is underlain by thin basin-fill sediments, generally less than 1,000 ft deep with carbonates directly below (Burbey, 1997). The northern Sheep Range is located on the west side of Coyote Spring Valley, and is a simple, narrow, and abrupt horst of Cambrian and Ordovician sedimentary rocks. The northeast side of Coyote Springs Valley contains the Delamar Mountains as described for Delamar Valley in the previous section. The east side of Coyote Spring Valley is comprised of the Meadow Valley Mountains and the Arrow Canyon Range. The southern end of the Meadow Valley Mountains, just east of Coyote Spring Valley, is made up of mostly thrust-faulted and normally faulted Paleozoic rocks. The Arrow Canyon Range is a sharp, narrow north-trending range consisting of a syncline of Cambrian to Mississippian carbonate rocks. It is uplifted along its western side by normal faults of the Arrow Canyon Range fault zone (Page and Pampeyan, 1996; Schmidt and Dixon, 1995; Page, 1998).

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