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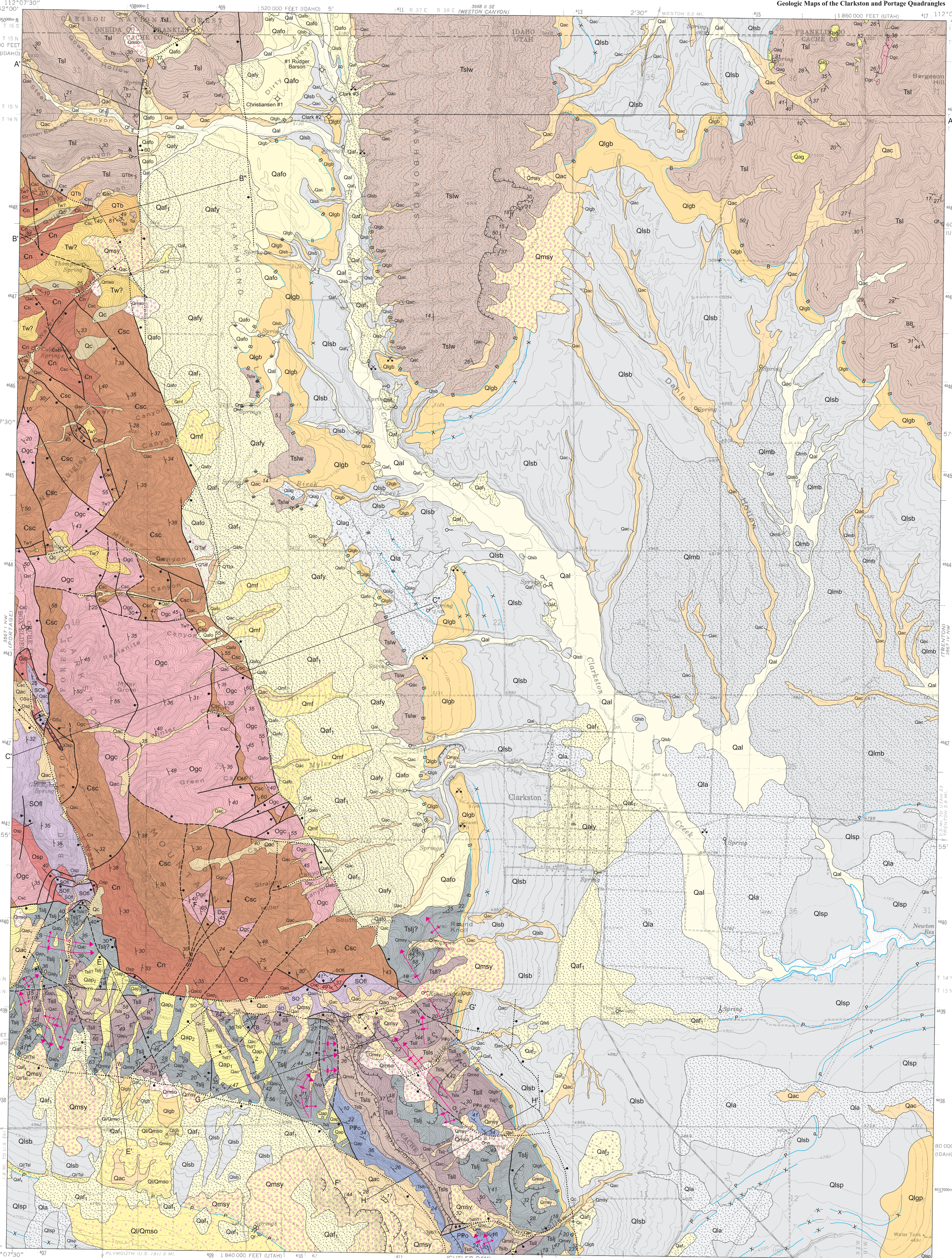

Base map from U.S. Geological Survey
Clarkston 7.5' Quadrangle, 1964

SCALE 1 : 24,000

1 1/2 0 1 MILE

1000 0 1000 2000 3000 4000 5000 6000 7000 FEET

1 .5 0 1 KILOMETER

CONTOUR INTERVAL 40 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

Field work by authors 1996-1998
Lori J. Douglas, cartographer

13°36'
242 mils

2003 MAGNETIC DECLINATION
AT CENTER OF SHEET

GEOLOGIC MAP OF THE CLARKSTON QUADRANGLE, BOX ELDER AND CACHE COUNTIES, UTAH AND FRANKLIN AND ONEIDA COUNTIES, IDAHO

by

Robert F. Biek,
Robert Q. Oaks, Jr.,
Susanne U. Janecke,
Barry J. Solomon,
and L. Marie Swenson Barry

2003

UTAH

QUADRANGLE LOCATION

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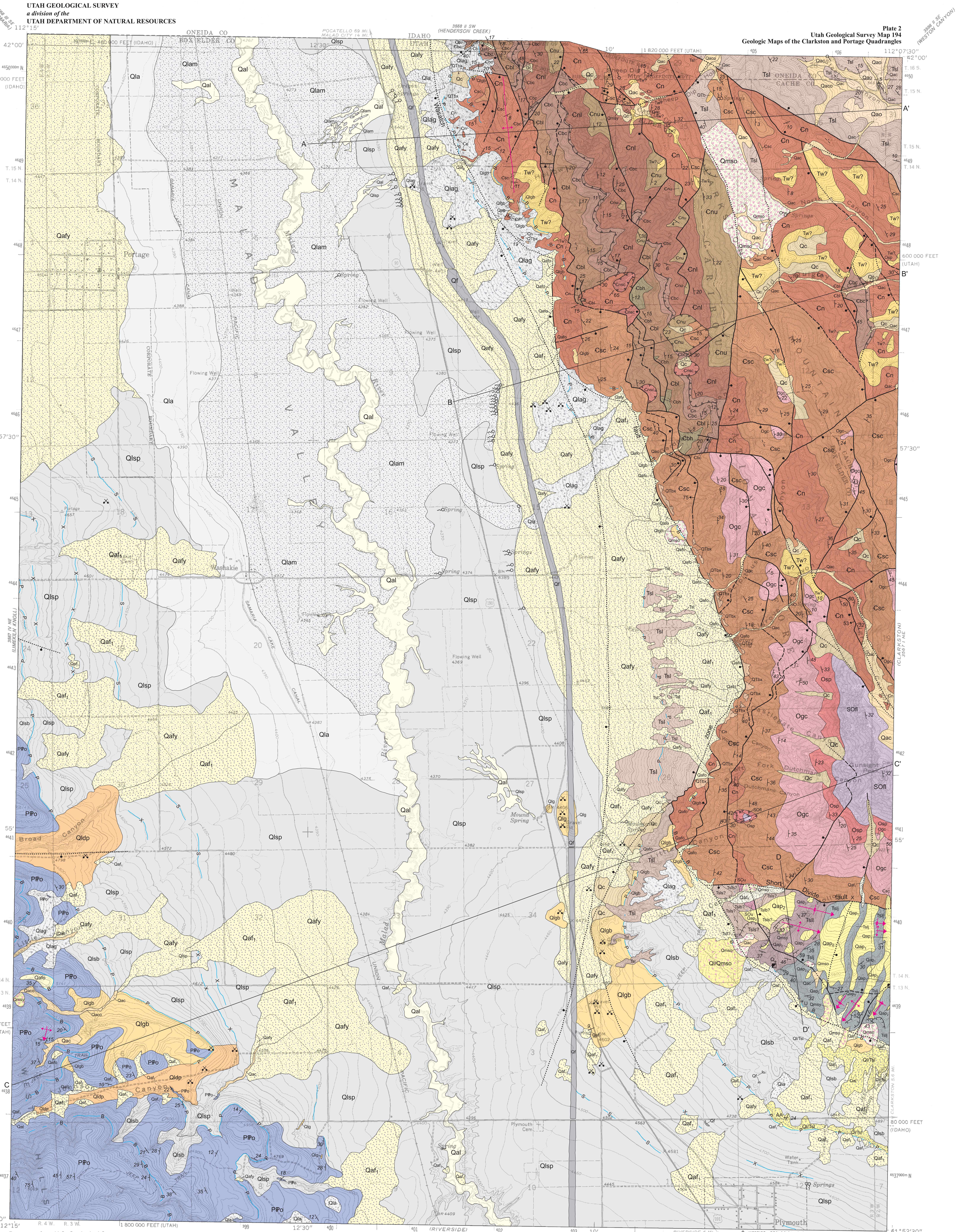

Base map from U.S. Geological Survey
Portage 7.5' Quadrangle, 1964

Field work by authors 1996-1998
Lori J. Douglas, cartographer

SCALE 1 : 24,000

1 1/2 0 1 MILE

1000 0 1000 2000 3000 4000 5000 6000 7000 FEET

1 .5 0 1 KILOMETER

CONTOUR INTERVAL 40 FEET
NATIONAL GEODETIC VERTICAL DATUM OF 1929

13°39'
243 mils

2003 MAGNETIC DECLINATION
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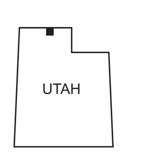
GEOLOGIC MAP OF THE PORTAGE QUADRANGLE, BOX ELDER AND CACHE COUNTIES, UTAH AND FRANKLIN AND ONEIDA COUNTIES, IDAHO



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MAP UNIT DESCRIPTIONS

QUATERNARY

Qal Alluvial deposits - *Moderately sorted sand, silt, clay, and local pebble to boulder gravel. Includes river-channel, flood-plain, local small alluvial-fan and colluvial deposits, and minor terraces up to 10 feet (3 m) above current base level; 0 to 20 feet (0-6 m) thick.*

Qat_1 Stream-terrace deposits - *Moderately sorted sand, silt, clay, and local pebble to boulder gravel that forms isolated, level to gently sloping surfaces up to 30 feet (10 m) above modern drainages. Deposited principally in river-channel and flood-plain environments; 0 to 20 feet (0-6 m) thick.*

Qao Older alluvial deposits - *Moderately sorted, locally well-consolidated, sand and pebble- to cobble-size gravel exposed in Steel Canyon and near Short Divide; white to light gray with carbonate-coated clasts. Probably less than 40 feet (12 m) thick.*

Qag Older gravel deposits - *Unconsolidated, carbonate-coated, pebble- to cobble-size gravel scattered throughout Bergeson Hill. Probably represents lag deposits of resistant lithologies eroded from the Salt Lake Formation. Less than about 10 feet (3 m) thick).*

Qaf_1 Modern alluvial-fan deposits - *Poorly to moderately sorted, clay- to boulder-size sediments deposited principally by debris flows at the mouths of active drainages. Upper portions are characterized by abundant boulders and debris-flow levies that radiate away from canyon mouths. Probably less than several tens of feet thick.*

Qaf_2 Level 2 alluvial-fan deposits - *Poorly to moderately sorted, clay- to boulder-size sediments. Exposed thickness less than 15 feet (5 m).*

Qafy Younger undifferentiated alluvial-fan deposits - *Poorly to moderately sorted, clay- to boulder-size sediments deposited principally by debris flows at the mouths of active drainages. Forms upper portion of thick, coalesced, active alluvial apron that slopes gently away from Clarkston Mountain and the West Hills.*

Qafo Older alluvial-fan deposits - *Poorly to moderately sorted, boulder- to clay-size sediments deposited along the flanks of Clarkston Mountain and in the West Hills. Forms deeply dissected alluvial apron locally overlain by Lake Bonneville deposits, and small, fault-bounded, isolated deposits at mountain front. Exposed thickness up to several tens of feet.*

Qap_{1-2}, Qap Pediment-mantle deposits - *Poorly to moderately sorted, boulder- to clay-size sediment that forms planar, deeply dissected, gently sloping surfaces along the southern flank of Clarkston Mountain, south of the Short Divide fault. Subscript denotes relative level, with Qap_1 lower and Qap_2 higher; mapped as Qap where undifferentiated; 0 to about 20 feet (0-6 m) thick.*

Qf Artificial fill - *Engineered fill used in the construction of Interstate 15, and general borrow material used to create small stock and retaining ponds. Although we mapped only the larger deposits, fill should be anticipated in all developed areas, many of which are shown on the topographic base map; thickness variable.*

Qc Colluvial deposits - *Poorly to moderately sorted, clay- to boulder-size, locally derived sediment deposited on slopes principally by slopewash and soil creep. 0 to 20 feet (0-6 m) thick.*

Qlgb, Qlgp, Qlg Lacustrine sand and gravel - *Moderately to well-sorted, moderately to well-rounded, clast-supported, pebble to cobble gravel, and lesser pebbly sand; thin to thick bedded. Contains mixed clast assemblage. Gastropods common in sandy lenses. Qlgb deposited at and immediately below highest Bonneville shoreline, and Qlgp deposited at and immediately below the Provo shoreline, in beaches, bars, and spits. Qlg mapped northwest of Portage and on the flanks of the West Hills where it was probably deposited in spits below the Provo shoreline; 0 to several tens of feet thick.*

Qldp Deltaic deposits - *Moderately to well-sorted, thin- to thick-bedded, pebble- to cobble-size gravel, sand, and silt that form a planar, gently sloping surface at the Provo shoreline in the West Hills; 0 to about 80 feet (0-24 m) thick.*

Qlsb, Qlsp Lacustrine sand and silt deposits - *Coarse- to fine-grained lacustrine sand and silt, with minor clay. Typically rhythmically bedded and well sorted, locally iron stained, with locally common gastropods. Grades downslope from sandy nearshore deposits to finer grained offshore deposits. Qlsb deposited between the Bonneville and Provo shorelines, and Qlsp deposited below the Provo shoreline. Exposed thickness less than 30 feet (10 m).*

Qlmb Lacustrine silt and clay deposits - *Calcareous silt (marl) with minor clay and fine-grained sand. Exposed thickness less than 15 feet (5 m).*

Qmsy Younger landslide deposits - *Very poorly sorted, clay- to boulder-size sediment deposited principally by rotational slump processes. Characterized by hummocky topography and numerous internal scarps. Thickness highly variable.*

Qmso Older landslide deposits - *Very poorly sorted, clay- to boulder-size sediment deposited principally by rotational slump processes, but which is characterized by subdued hummocky topography and poorly preserved main scarps. Forms large, deeply dissected landslide deposit on a near dip slope of mostly Nounan strata at the head of North Canyon, and a small landslide in Nounan and lower St. Charles strata at the head of Water Hollow; other deposits developed principally in Salt Lake strata, with local truncation by the Bonneville shoreline and subdued morphology that suggest late Pleistocene age. Thickness highly variable.*

Qmf Debris-flow deposits - *Cobble- to boulder-size gravel in a matrix of silt, sand, clay, and minor pebbles; typically with well-developed levee deposits of angular rubble. Exposed thickness less than 30 feet (10 m).*

Qac, Qaco Alluvial and colluvial deposits - *Poorly to moderately sorted, clay- to boulder-size, locally derived sediment deposited in swales and small drainages. Older deposits (Qaco) are in Steel Canyon and the West Hills where they form deeply dissected surfaces. Qac is 0 to 20 feet (0-6 m) thick; Qaco is 0 to about 40 feet (0-12 m) thick.*

Qla Lacustrine and alluvial deposits - *Moderately to well-sorted, fine-grained sand, silt, and clay found principally below the Provo shoreline; 0 to several tens of feet thick.*

Qlag Lacustrine and alluvial gravelly deposits - *Moderately sorted, moderately to well-rounded, clast-supported, pebble- to cobble-size sand and gravel mapped below the Bonneville shoreline on the flanks of Clarkston Mountain and the West Hills where it is not practical to differentiate lacustrine and alluvial-fan deposits; 0 to several tens of feet thick.*

Qlam Lacustrine and alluvial fine-grained deposits - *Well-sorted, saline and alkaline silt and clay deposited in the northern portion of the Malad Valley. Forms white, irregularly vegetated ground where not disturbed by cultivation; 0 to several tens of feet thick.*

Qst Spring deposit - *Light-gray to brownish-gray, porous, calcareous tufa mapped near the head of Elbow Canyon.*

Ql/Qmsy, Ql/Qmso Lacustrine deposits over landslide deposits - *Mapped along south flank of Clarkston Mountain below the Bonneville shoreline (Ql/Qmso) where fine-grained lacustrine deposits form a discontinuous veneer over older, pre-Bonneville-shoreline landslide deposits, and similarly below the Provo shoreline (Ql/Qmsy). Lake deposits are 0 to several feet thick.*

Ql/Tsl Lacustrine deposits over the Salt Lake Formation - *Mapped along south flank of Clarkston Mountain below the Bonneville shoreline where fine-grained lacustrine deposits form a discontinuous veneer over the Salt Lake Formation. Lake deposits are 0 to several feet thick.*

QUATERNARY-TERTIARY

QTaf Alluvial-fan deposits - *Poorly to moderately sorted, clay- to boulder-size sediment that forms steeply sloping, deeply dissected surface at the entrance to Mikes Canyon. Exposed thickness less than 100 feet (30 m).*

QTb Remobilized Wasatch(?) Formation - *Bouldery, Wasatch(?)-like strata that overlie the Salt Lake Formation on the northeast flank of Clarkston Mountain. Thickness uncertain, but probably several tens of feet.*

QTbx Fault breccia - *Brecciated and re-cemented strata of the St. Charles, Nounan, and Salt Lake Formations mapped along the flanks of Clarkston Mountain. Typically forms a planar, steeply dipping veneer on faceted spurs that is 0 to 6 feet (0-2 m) thick; considerably thicker at the entrance to Mikes Canyon.*

unconformity

Tb Basaltic sills in Salt Lake Formation - *Light-gray to light-olive-gray, very fine-grained, commonly vesicular basalt. Poorly exposed and enclosed by highly altered zones that grade from black glass with pervasive conchoidal fractures; to greenish-black, light-olive-gray, and yellowish-gray, conchoidally fractured porcellanite and mudstone; to overlying beds with a phyllonitic sheen. Basalt sills and altered beds commonly have iron-manganese stains and so contrast sharply with typically white Salt Lake strata. Mapped near the entrance to North and Steel Canyons and Gowans Hollow. Less than one foot to about 10 feet (<1-3 m) thick. Yielded a K-Ar age of 8.0 ± 0.5 Ma (Don Fiesinger, Utah State University, written communication, March 8, 1999).*

Salt Lake Formation

Tsl Salt Lake Formation, undifferentiated - *Laterally and vertically variable sequence of interbedded, generally white to light-gray, thin- to medium-bedded, poorly to well-cemented, generally poorly exposed, tuffaceous siltstone, sandstone, oolitic limestone, micritic limestone, volcanic ash, porcellanite, and pebble to cobble conglomerate. Subdivided into six subunits in the Junction Hills, south flank of Clarkston Mountain, and Bergeson Hill; subunits are queried where uncertain. Collectively, Salt Lake strata are up to about 9,250 feet (2,820 m) thick.*

Tslw Subunit of the Washboards - *Moderately sorted, thin- to medium-bedded, poorly consolidated, calcareous, very light-gray silt and clay and lesser interbedded sand, pebbly sand, and gravel. Clasts are subrounded to rounded, pebble- to cobble-size Mutual and Geertsen Canyon quartzites, Cambrian and Ordovician carbonates, green Cambrian limestone nodules, black chert, resistant greenish Salt Lake Formation mudstone clasts, and rare granitic rocks. Forms west-dipping slope veneered with lag gravels and loess at the Washboards, and isolated exposures near the Bonneville shoreline northwest of Clarkston. Estimated thickness 3,000 to 4,000 feet (915-1,220 m).*

Tslp Oolite subunit of Plymouth - *Oolitic limestone, micrite, and less common thin interbeds of pebble to cobble conglomerate, tuffaceous marl, and siltstone; 210 feet (64 m) thick.*

Tslj Tephra subunit of Junction Hills - *Volcanic ash with interbedded tuffaceous siltstone, tuffaceous limestone, porcellanite, marl, and micrite. Up to 2,550 feet (780 m) thick.*

Tsll Zeolite subunit of Long Divide - *Tuffaceous sandstone with volcanic ash and porcellanite at the base and volcanic ash at the top; common pale-greenish yellow, zeolite-bearing siltstone. About 322 to 700 feet (98-215 m) thick.*

Tsls Micrite subunit of Short Divide - *Tuffaceous pebble to cobble conglomerate, tuffaceous limestone, very fine crystalline oolitic limestone, and lesser volcanic ash beds; 400 feet (122 m) thick.*

Tslb Conglomerate subunit of Bensons Hollow - *Tuffaceous sandstone and pebble to cobble conglomerate; 203 feet (62 m) thick.*

unconformity

Tw(?) Wasatch(?) Formation - *Unconsolidated, subrounded to rounded pebbles to boulders of locally derived lithologies, including Cambrian, Ordovician, and Silurian carbonates, quartzites, and chert, and cobbles of Mississippian and Pennsylvanian strata not now present at Clarkston Mountain. Contains distinctive upper Swan Peak orthoquartzite boulders up to 6 feet (3 m) in diameter. Uncommon clasts have red, coarse sand matrix that adheres to protected recesses, and clasts locally have a thick pedogenic carbonate rind that imparts a white appearance to deposits. Forms a deeply dissected, brown to light-gray, coarse rubble sheet scattered throughout north end of Clarkston Mountain, and commonly forms a thin cover or lag that drapes over ridge tops. May be in excess of 200 feet (60 m) thick immediately north of Elgrove Canyon.*

unconformity

PALEOZOIC, undifferentiated

SOu Garden City, Swan Peak, Fish Haven, and Laketown Formations, undifferentiated - *Poorly exposed, highly brecciated upper Swan Peak, Fish Haven/Laketown, and Garden City strata preserved as a fault sliver near the west end of the Short Divide fault.*

PENNSYLVANIAN-PERMIAN, undifferentiated

PPo Oquirrh Formation - *Interbedded, typically thin- to medium-bedded, medium- to dark-gray limestone and thin- to medium-bedded, medium-gray, very fine- to fine-grained sandstone with poorly developed ripple cross-stratification; sandstone is both calcareous and non-calcareous. Weathers to colors of grayish orange, moderate yellowish brown, and pale to moderate reddish brown. Forms steep, rounded slopes of the West Hills with few ledgy outcrops; exposures in the Junction Hills are medium- to thick-bedded, medium- to dark-gray, locally cherty, finely crystalline limestone, sandy limestone, and sandstone. Incomplete thickness in excess of 6,000 feet (1,830 m).*

unconformity

ORDOVICIAN-SILURIAN, undifferentiated

SO Fish Haven Dolomite, Laketown Dolomite, and Swan Peak Formation, undifferentiated - *Brecciated, fault-bounded blocks mapped near Short Divide.*

SOfl Fish Haven Dolomite and Laketown Dolomite, undifferentiated - *Medium- to very thick-bedded, medium to coarsely crystalline dolomite with local black and light-brown chert nodules and lenses; contains solitary and colonial corals. Forms rugged cliffs. Fish Haven strata are darker gray and less than about 100 feet (30 m) thick; Laketown strata are in excess of 2,000 feet (600 m) thick.*

unconformity

ORDOVICIAN

Osp Swan Peak Formation - *Lower 108 feet (33 m) consists of slope- and ledge-forming, thin- to medium-bedded, moderate-brown orthoquartzite with abundant fucoidal markings, lesser medium-gray to olive-gray shale, and minor thin-bedded limestone; upper part forms prominent cliffs and ledges of very thick bedded, white to light-gray, white- to grayish-orange-pink-weathering, fine- to medium-grained orthoquartzite; 574 feet (175 m) thick.*

Ogc Garden City Formation - *Slope-and ledge-forming, thin- to medium-bedded, medium-gray to medium-bluish-gray limestone and limestone intraformational conglomerate that weathers to thin, gray to tan plates; upper part contains abundant black chert nodules and stringers and local dolomite; 1,764 feet (538 m) thick.*

unconformity

CAMBRIAN

Ꞓnsc Nounan and St. Charles Formations, undifferentiated - *Brecciated, fault-bounded blocks of Nounan and St. Charles strata mapped north of Gardner Canyon.*

Ꞓsc St. Charles Formation - *Divisible into three units (not mapped separately): a ledge-forming lower sandstone unit (Worm Creek Quartzite Member), a slope- and ledge-forming middle limestone unit, and an upper, unnamed, cliff-forming, very thick-bedded dolomite unit. The lower unit, about 70 feet (20 m) thick, is interbedded, thin-bedded, silty and sandy carbonate sandwiched between two beds of medium- to very thick-bedded, brown-weathering, light-gray to brownish-gray to pale-red, fine- to medium-grained silica-cemented feldspathic sandstone that has a conchoidal fracture. The middle limestone unit, about 250 feet (75 m) thick, consists of thin- to medium-bedded, medium-gray limestone with thin interbeds of intraformational limestone conglomerate and fossil hash throughout, while its lower part contains thin interbeds of brown-weathering siltstone. The upper unit, about 700 feet (210 m) thick, consists of mostly thick- to very thick-bedded, medium- to dark-gray, medium to coarsely crystalline dolomite with local light-brown chert nodules and stringers; 1,020 feet (311 m) thick.*

Nounan Formation

Ꞓn Nounan Formation, undifferentiated - *Mapped south of Gardner Canyon, along the northwest margin of Clarkston Mountain, and east of the crest of Clarkston Mountain where the distinction between the upper and lower members is poorly developed.*

Ꞓnu Upper member - *Thin- to medium-bedded, light- to dark-gray, locally oolitic, variably sandy and silty dolomite and limestone, and interbedded, locally iron-stained and micaceous, calcareous siltstone and fine-grained sandstone. Weathers to poorly exposed, brown and reddish-brown slopes. Thins from at least 500 feet (150 m) thick in the north to about 100 feet (30 m) thick in the south.*

Ꞓnl Lower member - *Cliff-forming, thick- to very thick-bedded, medium to coarsely crystalline, light- to medium-gray dolomite; locally mottled light and dark gray with relict oncolites, oolites, fossils, and wavy to planar algal stromatolites. Thickens from about 800 feet (245 m) thick in the north to about 1,200 feet (365 m) thick in the south.*

Bloomington Formation

Ꞓb Bloomington Formation, undifferentiated - *On plate 1, may include Hodges Shale and middle limestone strata exposed in fault blocks along the northwest flank of Clarkston Mountain.*

Ꞓbc Calls Fort Shale Member - *Brown-weathering, slope-forming, laminated to very thin-bedded, brown shale and micaceous siltstone with some interbedded, thin- to very thick-bedded, locally oolitic, gray limestone and limestone intraformational conglomerate. Shales contain distinctive greenish-gray-weathering, irregularly shaped limestone nodules; 429 feet (131 m) thick.*

Ꞓbl Middle limestone member - *Ledge- and cliff-forming, thin- to very thick-bedded, medium-gray limestone, oolitic limestone, and local intraformational limestone conglomerate; 444 feet (135 m) thick.*

Ꞓbh Hodges Shale Member - *Slope- and ledge-forming, thin-bedded, medium-gray limestone, oolitic limestone, and local intraformational limestone conglomerate, and lesser interbedded light-olive-gray shale. Incomplete section 439 feet (134 m) thick.*

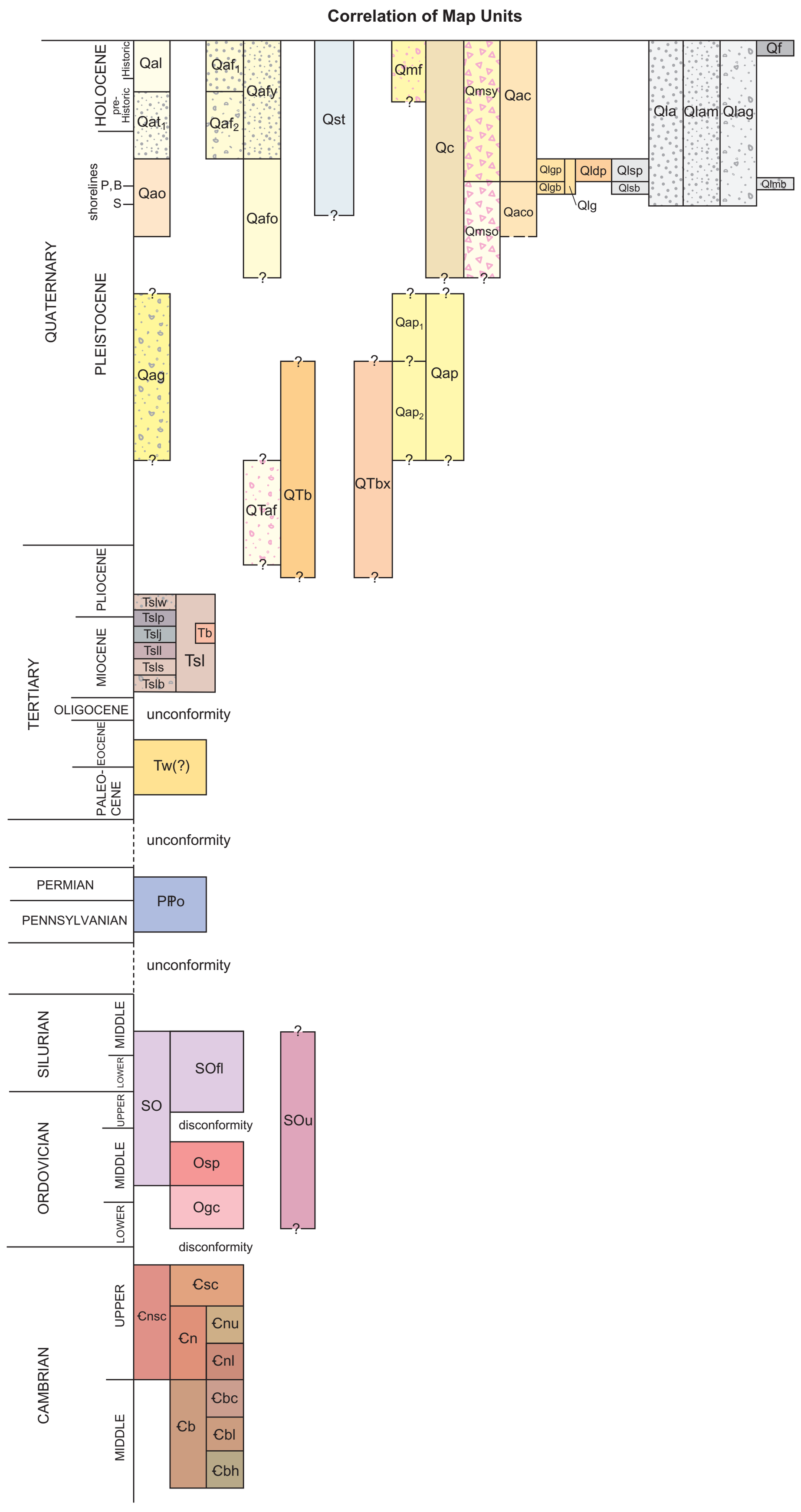
Subsurface units

Ꞓ-pꞒ Cambrian and Precambrian, undivided - *shown in cross-section only.*

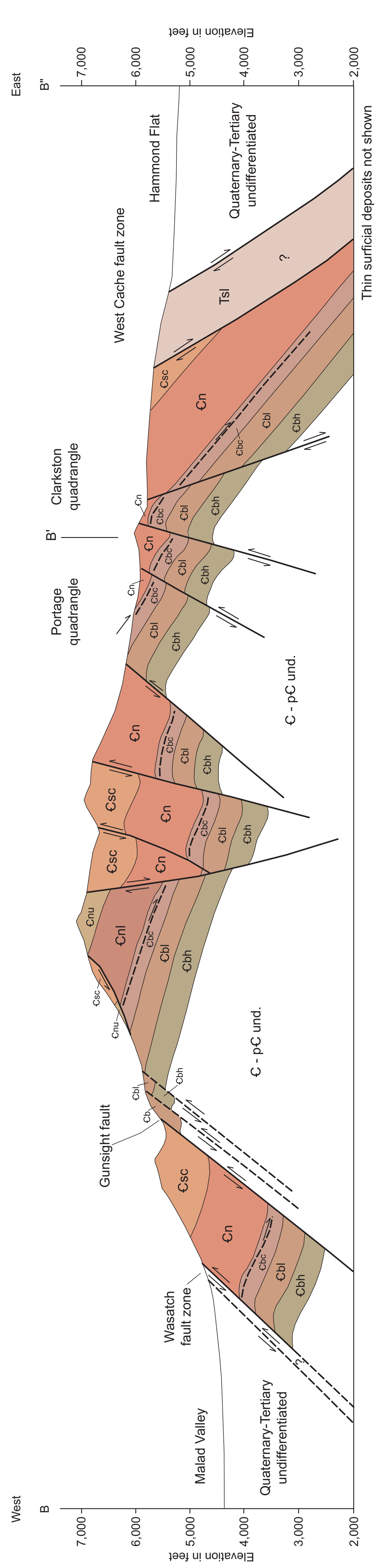
Map Symbols

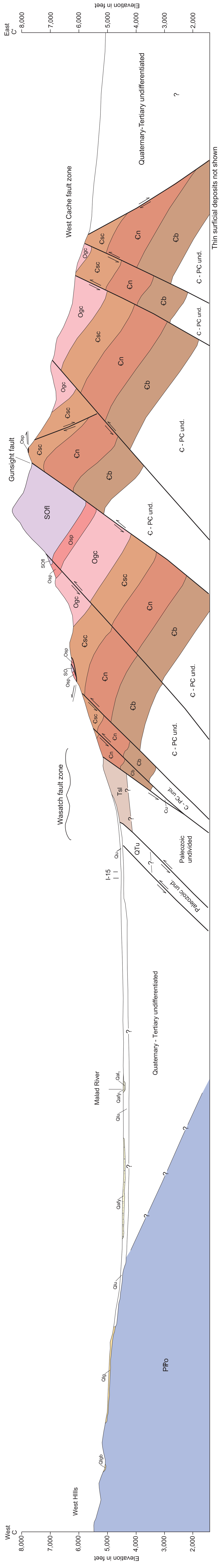
- Contact
- Normal fault, dashed where approximately located, dotted where concealed; bar and ball, or arrow with number showing dip, on down-dropped side; queried where uncertain
- Gently dipping normal fault, dotted where concealed; barbs on upper plate
- Landslide scarp, hachures on down-dropped side
- Axial trace of anticline, showing direction of plunge
- Axial trace of syncline, showing direction of plunge
- Lake Bonneville shorelines, B = Bonneville, P = Provo, S = Stansbury, X = unnamed
- 14 Strike and dip of inclined bedding
- 44 Strike and dip of overturned bedding
- Approximate directions of strike and dip of inclined bedding
- Sand and gravel pit
- Prospect
- Shaft
- Petroleum exploration well, plugged and abandoned
- Spring
- Seep
- + C Location of volcanic ash sample (see also table 1 in text)

Correlation of Map Units



SYSTEM	SERIES	FORMATION		MAP UNIT SYMBOL		THICKNESS feet (meters)		LITHOLOGY
TERTIARY	PLIOCENE	Salt Lake Fm. (Tb = basaltic sills in Tsl)	Subunit of the Washboards	Tsl Tb	Tslw	3,000-4,000 (915-1,220)	9,250 (2,820)	
			Oolite subunit of Plymouth		Tslp	210 (64)		
	MIOCENE		Tephra subunit of Junction Hills		Tslj	2,550 (780)		7.9 Ma ash
			Zeolite subunit of Long Divide		Tsll	322 - 700 (98 - 215)		
			Micrite subunit of Short Divide		Tsls	400 (122)		10.94 Ma ash
			Conglomerate subunit of Bensons Hollow		Tslb	203 (62)		unconformity
	PALEOCENE-EOCENE	Wasatch? Fm.		Tw?		200 (60)		FAULT OMITS SECTION
PENNSYLVANIAN/PERMIAN		Oquirrh Fm.		PPo		6,000+ (1,830)		top not exposed; base not exposed; FAULT OMITS SECTION
SILURIAN	MIDDLE / LOWER	Fish Haven - Laketown undiff.	Laketown Dolomite	SOu SO	SOfl	2,000+ (600+)	2,100+ (630+)	top not exposed; solitary and colonial corals
ORDOVICIAN	UPPER		Fish Haven Dolomite			100 (30)		disconformity
	MIDDLE	Swan Peak Fm.			Osp	574 (175)		fucoidal markings
	LOWER	Garden City Fm.			Ogc	1,764 (538)		cherty; Intraformational limestone conglomerate; disconformity
CAMBRIAN	UPPER	St. Charles Fm.		Ꞓnsc	Ꞓsc	1,020 (311)		Worm Creek Quartzite Member
		Nounan Fm.	upper member	Ꞓn	Ꞓnu	N S 500 - 100 (150 - 30)	1,300 (400)	
			lower member		Ꞓnl	N S 800 - 1,200 (245 - 365)		
	MIDDLE	Bloomington Fm.	Calls Fort Shale Member	Ꞓb	Ꞓbc	429 (131)	1,312+ (400)	green limestone nodules
			middle limestone member		Ꞓbl	444 (135)		oolitic limestone
			Hodges Shale Member		Ꞓbh	439+ (134+)		base not exposed





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by

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MAP 194
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UTAH GEOLOGICAL SURVEY

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11/03

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ABSTRACT

The Clarkston and Portage quadrangles lie astride the boundary between the Basin and Range and Middle Rocky Mountains physiographic provinces. The quadrangles straddle Clarkston Mountain, the southern extension of the Malad Range of southern Idaho, which consists of a complexly faulted sequence of generally east-dipping lower Paleozoic strata. Nearly 8,000 feet (2,400 m) of Middle Cambrian to Silurian carbonate and fine-grained clastic strata are exposed on Clarkston Mountain. The oldest rocks were previously assigned to the Ute, Blacksmith, and Bloomington Formations, but are herein reinterpreted, respectively, as the Hodges Shale, middle limestone, and Calls Fort Shale members of the Bloomington Formation. The Bloomington Formation is overlain by the Nounan, St. Charles, Garden City, Swan Peak, Fish Haven, and Laketown Formations. Oquirrh Formation strata are exposed in the West Hills and Junction Hills, and the Garden City Formation at Bergeson Hill. We also mapped the Paleocene-Eocene Wasatch(?) Formation, the Miocene-Pliocene Salt Lake Formation, and a variety of unconsolidated sediments, including extensive Pleistocene Lake Bonneville deposits, in the quadrangles.

Detailed stratigraphic and tephrochronologic studies have enabled subdivision of the Salt Lake Formation into six informal subunits. The formation consists of a laterally and vertically variable sequence of interbedded, generally white to light-gray, tuffaceous siltstone, sandstone, conglomerate, limestone, volcanic ash beds, and porcellanite. Based on tephrochronologic studies, the Clarkston and Portage quadrangles contain both the oldest (10.94 ± 0.03 Ma) and youngest (about 4.4 to 5.1 million years old) known Salt Lake Formation beds in the nearby area, making it late-middle Miocene to at least early Pliocene. Salt Lake strata are offset by at least 30 faults that bound at least 25 separate blocks along the south flank of Clarkston Mountain and the Junction Hills; folds with a wide variety of trends also deform these rocks. Chemical fingerprinting of volcanic ashes has proven useful in mapping the Salt Lake Formation in this structurally complex area.

The east-dipping, lower Paleozoic strata of Clarkston Mountain may be part of the east limb of a large anticline, the axis of which is faulted and now buried under Malad Valley. These strata were transported eastward and folded during the Sevier orogeny as the upper plate of the Paris-Willard thrust. Clarkston Mountain is cut by numerous, mostly down-to-the-west normal faults that may reflect the range's position in the upper plate of the Miocene-Pliocene Bannock detachment. Other structures include a minor bedding-plane fault of possible early Tertiary age, and several klippen bounded by gently dipping faults of probable late Tertiary age. In plan view, a fault map of Clarkston Mountain looks like a shattered pane of glass.

The Wasatch fault zone forms the western margin of Clarkston Mountain and Junction Hills, and bounds the eastern margin of Malad Valley, an east-tilted asymmetric graben or half graben filled with up to 5,000 feet (1,500 m) of Cenozoic sediments. Clarkston Mountain is bounded on the east by the West Cache fault zone, which has documented Holocene offset in the Clarkston quadrangle, and on the south by the Short Divide fault. We interpret the Short Divide fault, a major transverse structural feature, as a segment boundary of both the Wasatch and West Cache fault zones.

The principal economic resources in the quadrangles include sand and gravel, and ground water. Springs in the quadrangles provide an important source of culinary water for local communities. Because the Wasatch and West Cache fault zones traverse the Clarkston and Portage quadrangles, and because of their proximity to other major normal-fault zones, the Clarkston and Portage quadrangles are faced with significant seismic hazards, including surface fault rupture, ground shaking, liquefaction, and other seismically induced hazards. Other geologic hazards in the quadrangles include mass movements, flooding, shallow ground water, problem soil and rock, and radon.

INTRODUCTION

The Clarkston and Portage quadrangles are about 90 miles (145 km) north-northwest of Salt Lake City (figure 1). The two quadrangles straddle Clarkston Mountain and Junction Hills, the crest of which forms the boundary between Box Elder and Cache Counties. To the west, Malad Valley is traversed by Interstate 15 and is home to the small towns of Portage and Plymouth. Cache Valley, and the town of Clarkston, lie to the east. These and nearby towns, which depend on local springs and ground water for their water supply, experienced water shortages during the drought of the early

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1990s. Geologic hazards including earthquakes, landslides, flooding, problem soil and rock, shallow ground water, and radon are known in the Clarkston and Portage quadrangles and surrounding area.

The Clarkston and Portage quadrangles lie astride the boundary between the Basin and Range and Middle Rocky Mountains physiographic provinces (Fenneman, 1931; Stokes, 1986). Clarkston Mountain consists of a complexly faulted sequence of generally east-dipping Middle Cambrian to Silurian strata that are well exposed along the mountain's rugged western flank, and generally poorly exposed along its eastern dip slope. The Wasatch fault zone forms the western margin of the mountain range and the eastern margin of Malad Valley, an asymmetric graben or half graben filled with up to 5,000 feet (1,500 m) of Cenozoic sediments (Peterson, 1974; Zoback, 1983). The Wasatch fault zone offsets older alluvial-fan deposits (Qafo) at the entrance to Elgrove Canyon, above the Bonneville shoreline; this scarp is probably pre-Bonneville in age. Clarkston Mountain and the Junction Hills are bounded on the east by the West Cache fault zone, which shows evidence of Holocene offset in the Clarkston quadrangle (Black and others, 1999, 2000). The Short Divide fault marks the southern end of Clarkston Mountain and is also a segment boundary of the Wasatch and West Cache fault zones (Machette and others, 1992; Solomon, 1999; Black and others, 1999, 2000). Deposits and shoreline features of Pleistocene Lake Bonneville are well preserved in both Cache and Malad Valleys.

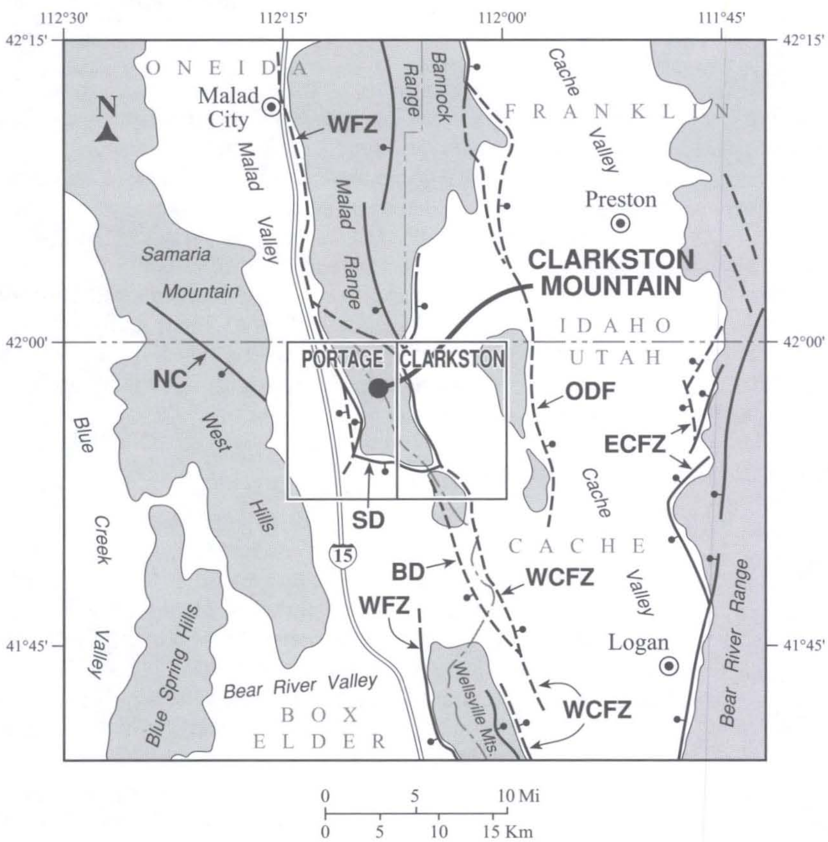


Figure 1. Location of the Clarkston (right) and Portage (left) quadrangles with simplified traces of the North Canyon fault (NC), Wasatch fault zone (WFZ), Short Divide fault (SD), West Cache fault zone (WCFZ), Beaver Dam fault (BD), Oxford-Dayton fault (ODF), and the East Cache fault zone (ECFZ). Modified from Solomon (1999).

Clarkston Mountain, Junction Hills, and adjacent Cache and Malad Valleys have been the focus of numerous topical geological investigations, many of which are cited under other sections of this report. The bedrock of Clarkston Mountain was first mapped in detail by Hanson (1949) on a 1:24,000-scale planimetric base, and later at 1:12,000 by Burton (1973), Gray (1975), and Green (1986), who mapped the northern, southern, and middle portions of Clarkston Mountain, respectively. Prammani (1957) mapped the east-central part of the Malad Range, including the area north of Steel Canyon in the Clarkston quadrangle. Cluff and others (1974) used 1:12,000-scale, low-sun-angle aerial photographs to map the traces of suspected surface-fault ruptures along the Wasatch and West Cache fault zones in the Clarkston and Portage quadrangles. Solomon (1999) produced a 1:50,000-scale Quaternary geologic map of western Cache Valley, including the West Cache fault zone and adjacent mountain fronts.

Recent 1:24,000-scale geologic maps of adjacent areas include those of Murphy and others (1985), who mapped the Limekiln Knoll quadrangle, and Oviatt (1986a), who mapped the Cutler Dam quadrangle. Beus (1963, 1968), Platt (1977), and Allmendinger and Platt (1983) also mapped parts of the West Hills and Samaria Mountains. Goessel (1999) mapped the Junction Hills and part of the northern Wellsville Mountains. Williams (1948, 1958) provided 1:125,000-scale geologic maps of the Logan 30 x 60-minute quadrangle and Cache County, respectively. Dover (1995) later provided a revised geologic map of the Logan 30 x 60-minute quadrangle. Doelling (1980) produced a 1:125,000-scale geologic map of Box Elder County. Oriel and Platt (1968) provided a 1:62,500-scale reconnaissance geologic map of the Preston 30 x 60-minute quadrangle in southern Idaho. Rember and Bennett (1979) and Oriel and Platt (1980) provided 1:250,000-scale geologic maps of the Pocatello and Preston 1° x 2° quadrangles, respectively.

The original fieldwork and aerial photo interpretation for the Clarkston (plate 1) and Portage (plate 2) quadrangles was begun by Jon King (Utah Geological Survey – UGS) in the mid-1990s. Biek mapped the geology of Clarkston Mountain in 1996. Solomon completed his map of the Cache Valley portion of the Clarkston quadrangle in 1997, as part of a larger study of the West Cache fault zone. Subsequent to King's work, Oaks, with Utah State University students L. Marie Swenson and Kathryn M. Goessel, recognized, mapped, and reported on several subunits of the Salt Lake Formation in the Junction Hills and southern flank of Clarkston Mountain. Janecke spent many days in the field and office, critically evaluating the mapping of Wasatch(?) and Salt Lake strata and structure, and her insight has enabled a more realistic assessment of these units. Biek compiled these maps in addition to mapping the remainder of the quadrangles. Biek wrote the report and prepared plate 3 materials in 1998, and in

1999 incorporated Oaks' new mapping and text of the Salt Lake Formation in the Junction Hills and Clarkston piedmont.

STRATIGRAPHY

Nearly 8,000 feet (2,400 m) of lower Paleozoic carbonate and fine-grained clastic marine strata are exposed on Clarkston Mountain. These strata were deposited in a variety of shallow-marine to offshore-marine environments during the Middle Cambrian to Silurian, and were transported eastward in the middle Cretaceous (Neocomian to Turonian) during the Sevier orogeny as part of the upper plate of the Paris-Willard thrust fault (Crittenden, 1972; Allmendinger and others, 1984; Allmendinger, 1992; DeCelles, 1994).

Cambrian

Bloomington Formation

The oldest strata at Clarkston Mountain were previously assigned to the Ute, Blacksmith, and Bloomington Formations (Hanson, 1949; Burton, 1973; and Green, 1986), but we reinterpret these rocks as the Hodges Shale, middle limestone, and Calls Fort Shale members, respectively, of the Bloomington Formation. Hanson (1949) first mapped strata exposed in several fault blocks along the northwest flank of Clarkston Mountain as the Ute, Blacksmith, and Bloomington Formations, but in doing so acknowledged the resulting discrepancies forced upon what he called the Bloomington Formation. Hanson's Bloomington strata are just 429 feet (131 m) thick, one-third the thickness of these beds elsewhere in northern Utah. Furthermore, Hanson's Bloomington stratigraphy does not match sections elsewhere in northern Utah, evidence that the upper part (what would later be known as the Calls Fort Shale and much of the middle limestone) may be missing. Hanson (1949) suggested that these differences were the result of an unconformity, but admitted he had found no physical or faunal evidence for such a hiatus. In an effort to resolve this discrepancy, Burton (1973) and Green (1986) mapped a thrust fault at the contact of the Bloomington and Nounan Formations, which they believed truncated the upper part of the Bloomington Formation.

The lithostratigraphic units in question are all Middle Cambrian (Maxey, 1958; Oviatt, 1986b; Jensen and King, 1999), but are largely unfossiliferous at Clarkston Mountain. Still, stratigraphic relationships appear to be best explained by assuming that beds previously mapped as the Ute, Blacksmith, and Bloomington Formations are, respectively, the Hodges Shale, middle limestone, and Calls Fort Shale members of the Bloomington Formation. The best evidence in support of this interpretation is the fact that shale beds mapped by Hanson (1949), Burton (1973), and Green (1986) as Bloomington, and herein mapped as the Calls Fort Shale, contain distinctive greenish-gray-weathering, irregularly shaped limestone nodules. These nodules are absent in the Hodges Shale and middle limestone members, but are characteristic of the Calls Fort Shale, the upper member of the Bloomington Formation (Maxey, 1958).

Despite the different names assigned to these units, our contacts and those used by previous workers are substantially the same. The middle limestone member, which forms bold cliffs and ledges, separates the slope- and ledge-forming shales of the Hodges Shale and Calls Fort Shale Members. Both contacts are gradational in that thin ribbon limestone beds similar to those of the middle limestone are present in both the upper and lower shale members. Hanson (1949) measured in excess of 1,300 feet (400 m) of strata that we map as Bloomington along the northwest flank of Clarkston Mountain; the base is not exposed. The Bloomington Formation is 1,085 feet (550 m) thick in the Wellsville Mountains (Maxey, 1958) and at least 1,000 feet (300 m) thick in the Bear River Range (Dover, 1995).

The middle limestone member was deposited in a variety of lagoon, shoal, and tidal-flat environments (Hay, 1982). Local thin interbeds of oolitic limestone and intraformational limestone conglomerate also support a shallow-marine depositional environment for the Hodges Shale and Calls Fort Shale Members. The Bloomington Formation is Middle Cambrian based on fossil data (Maxey, 1958). Jensen and King (1999) reported Middle Cambrian brachiopods from the Hodges Shale Member in the Brigham City quadrangle. Oviatt (1986b) found Middle Cambrian trilobites from the Calls Fort Shale in the northern Wellsville Mountains. The fossil-poor middle limestone, located between shales containing Middle Cambrian fossils, is also Middle Cambrian in age.

Hodges Shale Member (€bh): The Hodges Shale Member consists of interbedded, generally thin-bedded, fine-grained, medium-gray limestone with light-yellow-brown-weathering silty partings, and lesser interbedded light-olive-gray shale. Thin-bedded oolitic limestone and intraformational limestone conglomerate are common though not abundant. The Hodges Shale weathers to ledgy slopes that are best exposed in Gardner Canyon. Hanson (1949) measured an incomplete section of 439 feet (134 m) of strata he mapped as the Ute Formation, but that we reassign to the Hodges Shale Member of the Bloomington Formation. The contact with the overlying middle limestone is gradational and is marked by a pronounced break in slope. It corresponds to a change from thin-bedded limestone with minor silt and shale partings to cliff-forming, thick- to very thick-bedded limestone. We recovered several unidentified linguloid brachiopods from Hodges Shale strata.

Middle limestone member (€bl): The middle limestone member is a ledge- and cliff-forming, lithologically monotonous sequence of thin- to very thick-bedded, planar-bedded, locally cross-stratified, medium-gray limestone and oolitic limestone. Intraformational limestone conglomerate and cryptalgal laminae are locally present. The middle limestone member forms prominent cliffs and small box canyons, and is exposed north of Gardner Canyon along the western flank of Clarkston Mountain. Hanson (1949) measured 444 feet (135 m) of strata he assigned to the Blacksmith Formation but that we reassign to the middle limestone member of the Bloomington Formation. Hay (1982) measured 368 feet (112 m) of beds that he assigned to the Blacksmith Limestone (middle limestone of this report) on the northwest flank of Clarkston Mountain (presumably measured in Gardner Canyon, but the published location is incorrect). The contact of the middle limestone and Calls Fort Shale members is gradational and corresponds to the base of the first shale bed.

Calls Fort Shale Member (€bc): The Calls Fort Shale

Member is laminated to very thin-bedded brown shale and brown micaceous siltstone, with characteristic greenish-gray limestone nodules, and interbedded, thin- to very thick-bedded, gray limestone. The limestone is locally oolitic, and contains some beds of intraformational limestone conglomerate. The upper Calls Fort Shale contains several ledge-forming units 10 to 15 feet (3-5 m) thick of medium- to very thick-bedded, locally oolitic limestone. The Calls Fort Shale weathers to brown, splintery shale slopes littered with greenish-gray limestone nodules and lesser platy limestone. Calls Fort strata are exposed north of Gardner Canyon along the northwest flank of Clarkston Mountain; the member is also in Old Quigley Canyon along the mountain's east flank. Hanson measured 429 feet (131 m) of Bloomington strata on the northwest flank of Clarkston Mountain that we reassign to the Calls Fort Shale Member of the Bloomington Formation. Hanson (1949) found only two fossils in his Bloomington strata, one agnostid and one undetermined trilobite cranidium.

The contact of the Calls Fort Shale and overlying Nounan Formation corresponds to a prominent break in slope, but it is poorly exposed on Clarkston Mountain. Burton (1973) and Green (1986) mapped the contact as a thrust fault in order to account for irregular thickening and thinning of their Bloomington strata. The Calls Fort Shale Member does thicken and thin irregularly in some exposures on the northwest flank of Clarkston Mountain, and Calls Fort strata likely contain a minor bedding-plane fault. We mapped the fault primarily within Calls Fort strata because we observed no deformation of overlying Nounan strata.

Bloomington Formation, Undifferentiated (€b)

We mapped the Bloomington Formation as undifferentiated in section 11, T. 14 N., R. 3 E., and near the state line in section 34, T. 15 N., R. 3 E., where complexly faulted, incomplete sections of the Hodges Shale and middle limestone members are exposed. Most of these exposures are probably Hodges Shale. The middle limestone may be present in the lower reaches of the canyon immediately north of Gardner Canyon, and in several ridge crests to the north. Calls Fort Shale strata, with their distinctive limestone nodules, are not present in these fault blocks.

Nounan Formation

A complete, unfaulted section of the Nounan Formation is not exposed in the Clarkston and Portage quadrangles, but the formation crops out widely in numerous fault blocks along the northwest flank of Clarkston Mountain, and in generally poorly exposed fault blocks on the mountain's eastern and southern flanks. Nounan strata are best exposed along the northwest flank of Clarkston Mountain, where they are readily divisible into two informal members: a lower member of thick- to very thick-bedded, medium- to coarsely crystalline, light- to medium-gray dolomite, and a lithologically variable upper member of thin- to thick-bedded, brown- to reddish-brown-weathering, light- to dark-gray, sandy and silty limestone and dolomite. A distinctive, coarsely crystalline, dark-gray dolomite with white fossil relicts forms the basal 100 feet (30 m) of the lower member. In exposures to the south and east, the distinction between the upper and lower members is much less obvious and so we mapped them as Nounan undifferentiated.

On Clarkston Mountain, the contact with the overlying St. Charles Formation is marked by the first appearance of ledge-forming, light-brown to pale-red, silica-cemented feldspathic sandstone – the base of the Worm Creek Quartzite Member of the St. Charles Formation – in an otherwise similar interval of slope-forming silty carbonates and calcareous siltstones. The contact appears conformable and gradational. It is best exposed on the southern flank of Clarkston Mountain, east of Water Canyon (figure 2), where it corresponds to a pronounced lithologic and color change at the top of a 30-foot-thick (10 m), cliff-forming, thin- to medium-bedded, medium-gray limestone. Farther east toward Short Divide, Nounan and St. Charles strata are heavily fractured and stained by iron-manganese oxides, which, except for the presence of the Worm Creek Quartzite Member, blurs the distinction between the two formations. Along the northwest flank of Clarkston Mountain, in the NE1/4 section 10, T. 14 N., R. 3 E., the base of the Worm Creek Quartzite Member shows slight channeling into thin to medium, planar dolomite beds of the upper Nounan Formation.

Dolomites of the lower Nounan and upper St. Charles Formations are similar and difficult to distinguish in faulted exposures that lack intervening upper Nounan and lower St.

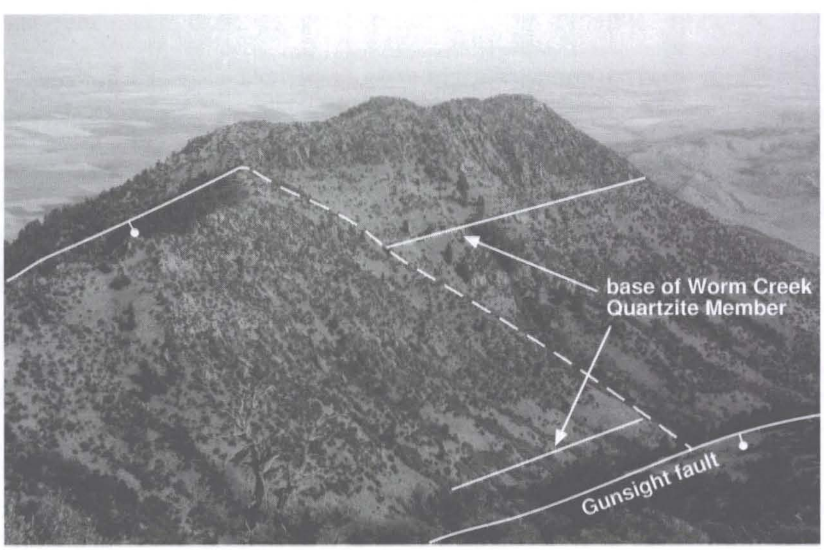


Figure 2. View southeast across Water Canyon to the southeast end of Clarkston Mountain. The St. Charles Formation consists of a lower unit of mostly ledge- and slope-forming sandstone and limestone about 320 feet (95 m) thick, the lower part of which belongs to the Worm Creek Quartzite Member, and an upper dolomite unit about 700 feet (210 m) thick. A normal fault, partially hidden from view (dashed line), displaces the Worm Creek Quartzite Member about 900 feet (275 m) down to the west. The Gunsight fault trends up Water Canyon.

Charles strata. Nounan dolomites, however, commonly contain mottled dark- and medium-gray oncolitic and pisolitic beds.

The Nounan Formation was deposited in a high-energy, subtidal to intertidal, shallow-marine environment (Gardiner, 1974). Nounan strata were probably deposited on a broad shallow shoal in which intratidal and supratidal flats became increasingly dominant in upper Nounan time. The Nounan Formation is Late Cambrian based on trilobites (Williams, 1948; Oviatt, 1986b). The Nounan Formation appears to maintain a relatively constant thickness of about 1,300 feet (400 m) throughout Clarkston Mountain.

Lower member (€nl): The lower member of the Nounan Formation forms rugged cliffs and pinnacles and consists of thick- to very thick-bedded, medium- to coarsely crystalline, light- to medium-gray dolomite. It is locally mottled light and dark gray, and relict oncolites, oolites, and both wavy and laminated algal stromatolites are common. Oncolites are typically structureless and darker gray than the enclosing dolomite. Bedding is commonly difficult to distinguish. Hanson (1949) measured 908 feet (277 m) of lower Nounan strata at the north end of Clarkston Mountain. At the south end of Clarkston Mountain, east of Water Canyon, an incomplete section of lower Nounan strata is 1,034 feet (315 m) thick; the base is not exposed. Lower Nounan strata probably vary from about 800 feet (245 m) thick in the north to about 1,200 feet (365 m) thick in the south.

Upper member (€nu): The upper member of the Nounan Formation is generally poorly exposed and weathers to brown or locally reddish-brown slopes. It consists of generally thin- to medium-bedded, light- to dark-gray, locally oolitic, sandy and silty dolomite and limestone, and interbedded, locally iron-stained and micaceous, calcareous siltstone and fine-grained sandstone. Hanson (1949) reported on trilobites from the upper Nounan Formation. Hanson (1949) estimated, and we concur, that the upper unit is at least 500 feet (150 m) thick at the north end of Clarkston Mountain, and thins to about 100 feet (30 m) thick at the mountain's south end.

Nounan Formation, Undifferentiated (€n)

The Nounan Formation is readily divisible into lower and upper members only in select exposures north of Gardner Canyon along the west flank of Clarkston Mountain. Elsewhere, we mapped it simply as Nounan undifferentiated.

St. Charles Formation (€sc)

Throughout much of northeastern Utah and southeastern Idaho, the St. Charles Formation is divisible into two members, the lower Worm Creek Quartzite Member and an upper unnamed dolomite member. On Clarkston Mountain, the distinction between these two members is well expressed only on the south flank of the mountain, where a complete section lies east of Water Canyon. Thus, the members are not mapped separately. Elsewhere, incomplete exposures of the St. Charles Formation are present in numerous fault blocks on the west flank of Clarkston Mountain, and in faulted dip-slope exposures on the mountain's east flank.

We measured 1,020 feet (311 m) of St. Charles strata at the south end of Clarkston Mountain, east of Water Canyon, similar to the thickness of 1,073 feet (327 m) reported by Hanson (1949). There, the St. Charles Formation is divisible into three units: a ledge-forming lower sandstone unit (assigned to the Worm Creek Quartzite Member), a slope- and ledge-forming middle limestone unit, and a much thicker cliff-forming, very thick-bedded dolomite unit. The lower unit, about 70 feet (20 m) thick, is interbedded, thin-bedded, silty and sandy carbonate sandwiched between two beds of medium- to very thick-bedded, brown-weathering, light-gray to brownish-gray to pale-red, fine- to medium-grained, silica-cemented feldspathic sandstone (commonly referred to as "quartzite" or "orthoquartzite")(Haynie [1957] apparently also measured the Worm Creek Quartzite Member east of Water Canyon, but neither the published location nor written description of the location match actual exposures). The middle limestone unit, about 250 feet (75 m) thick, is thin- to medium-bedded, medium-gray limestone. It contains thin interbeds of intraformational limestone conglomerate and fossil hash throughout, while the lower part contains thin interbeds of brown-weathering siltstone. Many bedding planes have abundant trace fossils, including horizontal, branching burrows similar to *Chondrites*; unburrowed areas weather to light brown. The upper unit, about 700 feet (210 m) thick, is mostly thick- to very thick-bedded, medium- to dark-gray, medium- to coarsely crystalline dolomite with local light-brown chert nodules and stringers. The dolomite is locally mottled light and dark gray, and some beds contain well-preserved planar laminae, ripple cross-stratification, and intraformational flat-pebble conglomerate. The upper St. Charles contains some thin beds of fossil hash with probable molluscan skeletal debris.

The two feldspathic sandstone beds of the Worm Creek Quartzite Member are also well developed on the north-trending ridge immediately west of Sheep Dip Mountain. Incomplete exposures there are at least 50 feet (15 m) thick and show that the sandstone thickens northward at the expense of interbedded limestones. The upper of the two beds, which is approximately 20 feet (6 m) thick and caps the ridge, is more prominent and lighter in color, unlike southern exposures wherein the lower sandstone appears more prominent; intervening tan, silty and sandy carbonates present in southern exposures are largely lacking. Both beds are pinkish-gray to light-brown to pale-red, thin- to thick-bedded, fine- to medium-grained, silica-cemented feldspathic sandstone with local planar cross-stratification. Both have a conchoidal fracture and weather medium brown with local iron staining and Liesegang banding. The lower sandstone is generally thinner bedded, silty, and burrowed.

Other instructive exposures of the Worm Creek Quartzite Member are present on the hilltop in the SW1/4 section 36, T. 15 N., R. 3 W.; on hill 6558 just east of Cold Water Springs, section 7, T. 14 N., R. 2 W. ; and in the NE1/4 section 10, T. 14 N., R. 3 W., where the base of the Worm Creek Quartzite Member is marked by a thin-bedded, burrowed, noncalcareous sandstone about 5 feet (2 m) thick, above which lies thick-bedded, planar-bedded, silica-cemented feldspathic sandstone. At this latter exposure, the sandstone shows slight channeling into thin- to medium-bedded, planar-bedded dolomite of the upper Nounan Formation.

We placed the contact of the St. Charles and Garden City Formations at the first appearance of thin-bedded, planar-bedded, medium- to dark-gray limestone; underlying St. Charles strata are thick- to very thick-bedded, light- to medium-gray, coarsely crystalline dolomite. While this lithologic break appears sharp, the contact is nowhere well exposed and, in western exposures, is generally not discernable on true-color aerial photographs. The contact was particularly difficult to place on the south side of Gunsight Peak due to thin colluvial cover and proximity to the Short Divide fault. Where beds form a dip slope on the east side of the mountain, the thin, slope-forming Garden City beds stand in stark contrast to underlying ledge- and cliff-forming St. Charles strata. In the NE1/4 section 13, T. 14 N., R. 3 W., the contact between the St. Charles and Garden City Formations, which is covered, may be a fault contact. There, breccia and calcite-healed fractures are common, and Garden City beds are locally discordant.

Trilobite and conodont faunas suggest that the St. Charles Formation is Late Cambrian to earliest Ordovician in age in the Bear River Range to the east (Landing, 1981; Taylor and others, 1981); we use the Cambrian symbol following well-established usage in the region. A sharp lithologic break, as well as faunal data, suggest that the St. Charles Formation is disconformably overlain by the Garden City Formation (Taylor and others, 1981). Oviatt (1986b) reported that St. Charles strata are 1,169 feet (356 m) thick at the northern end of the Wellsville Mountains.

Nounan and St. Charles Formations, Undifferentiated (€nsc)

Dolomites of the lower Nounan and upper St. Charles Formations are similar and difficult to identify in faulted exposures that lack intervening upper Nounan and lower St. Charles strata. Such is the case for exposures in sections 2 and 11, T. 14 N., R. 3 W. In each of these areas, the strata in question are klippe bounded by gently dipping normal faults.

Ordovician

Garden City Formation (Ogc)

The Garden City Formation is a thick sequence of thin- to medium-bedded, locally fossiliferous limestone, intraformational flat-pebble conglomerate, cherty limestone, and minor dolomite. Garden City strata are exposed on Clarkston Mountain south of Gardner Canyon where they weather to slopes and low ledges. The best and only complete exposures are on the west side of Gunsight Peak. Some of the best exposures on the east side of Clarkston Mountain are along the ridge crest just south of Elbow Canyon, but even there the Garden City Formation is much covered and cut by several small unmapped brecciated zones. Garden City strata are also exposed on the northwest flank of Bergeson Hill.

Hanson (1949) recognized but did not map four units of the Garden City Formation that totaled 1,805 feet (550 m) thick on the southwest flank of Gunsight Peak. Ross (1951) provided a more detailed measured section of the same interval that totaled 1,764 feet (538 m) thick. Ross (1951) broke this sequence into two main units. The lower unit, which makes up two-thirds of the formation, is predominantly thin-bedded, gray, intraformational limestone conglomerate and lesser finely crystalline to microcrystalline limestone and silty limestone. The upper unit is predominantly finely crystalline, irregularly laminated limestone with abundant black chert stringers and nodules that is capped by about 160 feet (49 m) of light-gray, finely crystalline, dolomitic limestone and dolomite with lesser chert. Ross (1951) noted that the comparatively thin upper dolomite unit is not uniformly present and that its unusual thickness at Clarkston Mountain may be due to faulting.

The most conspicuous Garden City rock type is thin-bedded, medium-bluish-gray, ledge-forming, intraformational, flat-pebble limestone conglomerate. The clasts are up to about 6 inches (15 cm) long, although most are 2 to 4 inches (5-10 cm) long; they are typically rounded and imbricated. Roughly equant clasts to 4 inches (10 cm) in diameter are present in some beds. These, as well as other Garden City beds, typically weather to thin, gray to tan plates, in contrast to the commonly mottled light- and dark-gray, coarsely crystalline, blocky-weathering upper St. Charles dolomite. The Garden City Formation can appear similar to limestones in the lower St. Charles Formation, but is distinguished by stratigraphic position and prominent intraformational limestone conglomerate. Garden City limestones are also characterized by slight bluish hues, in contrast to the plain grays of St. Charles and Nounan limestones. While the formation appears to be planar bedded, Ross (1951) noted elsewhere that Garden City beds typically thicken and pinch out within short distances.

The cherty, dolomitic upper portion of the Garden City Formation is best exposed immediately south of Elgrove Canyon, in the NW1/4NE1/4 section 24, T. 14 N., R. 3 W. There, black chert nodules and stringers make up more than 50 percent of some beds.

The contact of the Garden City and Swan Peak Formations represents a sharp lithologic and topographic break, although it is everywhere poorly exposed at Clarkston Mountain. We placed the contact at the base of poorly exposed, noncalcareous, gray to black shale and lesser interbedded, thin-bedded, fine-grained, light- to moderate-yellowish-brown, silica-cemented quartz siltstone. The contact is only exposed on the west and south sides of Gunsight Peak.

Morgan (1988) studied the petrology of the Garden City Formation and suggested that it was deposited in a storm-influenced, eastward-transgressive, shallow-marine environment. The Garden City Formation is Early Ordovician to earliest Middle Ordovician based on a prolific trilobite fauna (Ross, 1951), and disconformably overlies the St. Charles Formation (Landing, 1981). Hanson (1949) reported on Garden City fossils from Clarkston Mountain, including trilobites, brachiopods, and graptolites.

Swan Peak Formation (Osp)

The Swan Peak Formation forms a distinctive marker in an otherwise thick sequence of gray carbonates. The Swan Peak Formation is divisible into three informal members that total in excess of 1,500 feet (460 m) thick in southeastern Idaho (Oaks and others, 1977), but that thin markedly to the south. In the Portage quadrangle, Ross (1951) measured 570 feet (174 m), and James (1973) 574 feet (175 m), of Swan Peak strata on the southwest flank of Gunsight Peak; Oviatt (1986b) reported the Swan Peak Formation is 385 feet (117 m) thick in the northern Wellsville Mountains. Given poor exposures, gradational contacts between the lower and middle members, and comparative thinness of Swan Peak strata, we did not map individual members. Even so, portions of each member are readily identifiable in the field. The lower two members form saddles where they cross ridge crests.

The poorly exposed, slope-forming lower member appears to consist of interbedded, thin-bedded, dark-gray, noncalcareous shale, siltstone, and minor limestone. James (1973) assigned 79 feet (24 m) to the lower member.

The middle member consists of ledge- and slope-forming, interbedded, moderate-brown orthoquartzite and lesser medium-gray to olive-gray shale. The orthoquartzite is fine grained, in thin to medium generally planar beds, and is characterized by abundant fucoidal markings (horizontal feeding traces)(earlier investigators regarded similar markings first discovered in Ohio in 1838 as impressions of seaweed; the term "fucoid" was adopted from the generic name of a seaweed. The term "fucoid" is now used for indefinite trail-like or tunnel-like trace fossils). Van Dorston (1970) and Francis (1972) provided detailed descriptions of these trace fossils. Van Dorston (1970) interpreted them to be feeding burrows of orthoconic cephalopods, whereas Francis (1972) recognized that striations along the burrow walls likely resulted from worm-like organisms that could flex at the bases of the vertical shafts that lead to the horizontal burrows. Most workers note that these markings are most prominent on the undersides of beds or within beds; in the Portage quadrangle, they commonly appear on upper bedding surfaces. Except for a small exposure of fucoidal orthoquartzite along the east side of the klippe east of Gunsight Peak, in the SW1/4SE1/4SW1/4 section 19, T. 14 N., R. 2 W., the middle member is only exposed on the west and south sides of Gunsight Peak. James (1973) assigned 29 feet (9 m) of Swan Peak strata to the middle member.

The upper member forms prominent cliffs and ledges of white to very light-gray orthoquartzite. The orthoquartzite is fine to medium grained and weathers white to grayish orange pink, with brownish hues and local Liesegang banding toward the base. The upper portion is very thick bedded, whereas the lower part is generally planar to cross-bedded. In addition to exposures on the west and south sides of Gunsight Peak, the upper member is exposed in klippen north of Little Canyon, on the ridge crest east of Gunsight Peak, and at Short Divide. Upper Swan Peak strata are also exposed as fault breccia west of Short Divide and between North Canyon and Old Quigley Canyon. Orthoquartzite of the upper member is brecciated and intensely fractured in each of these displaced blocks. James (1973) measured 466 feet (142 m) of upper-member strata on the southwest side of Gunsight Peak.

Throughout northern Utah and southeastern Idaho, Middle Ordovician strata of the Swan Peak Formation and its lateral equivalents are disconformably overlain by the Upper Ordovician Fish Haven Dolomite (Oaks and others, 1977; Leatham, 1985). Although it represents a sharp lithologic change, the contact of the Swan Peak Formation and overlying Fish Haven Dolomite is not well exposed at Clarkston Mountain.

Oaks and others (1977) summarized the regional stratigraphy of the Swan Peak Formation and its equivalents in northern Utah and southern Idaho. The lower two members constitute a progradational sequence of interbedded shale and orthoquartzite that was deposited in transgressive nearshore, offshore, shoreface, and intertidal shallow-marine environments. A regional unconformity marks the base of the upper, very thick-bedded quartzite member, which was deposited on a broad shallow-marine shelf and which is correlative with the Eureka Quartzite of western Utah.

Lower Swan Peak strata are early Middle Ordovician in age based on trilobites (Ross, 1951); graptolites of the middle Swan Peak are also characteristic of the Middle Ordovician (Whiterockian). Although diagnostic fossils have not been recovered from the upper Swan Peak member in Utah, Oaks and James (1980) reported a late Middle Ordovician fossil assemblage from the Kinnikinic Quartzite of central Idaho, which is correlative with the upper member.

Ordovician and Silurian, Undivided

Fish Haven Dolomite and Laketown Dolomite, Undifferentiated (SOfl)

The Fish Haven and Laketown Dolomites are similar and commonly mapped as a single unit, as we have done here. No conspicuous lithologic change marks the Fish Haven-Laketown contact at Clarkston Mountain, only a subtle change in shades of gray, with darker Fish Haven beds below and generally lighter Laketown beds above. In his study of the Laketown Dolomite of northern Utah, Budge (1966) also noted the gradational nature of the Fish Haven-Laketown contact.

Undifferentiated strata of the Fish Haven and Laketown Dolomites are incompletely but well exposed in rugged cliffs on the west and south sides of Gunsight Peak. The best exposures of Fish Haven strata are on the northwest flank of Gunsight Peak, north of Castlegate Canyon in the NE1/4NE1/4 section 25, T. 14 N., R. 3 W. Elsewhere, exposures surrounding the Fish Haven-Laketown contact are poor and the color contrast less obvious. Hanson (1949) estimated Fish Haven strata are only about 50 feet (15 m) thick and an incomplete section of overlying Laketown beds is about 2,000 feet (600 m) thick. Somewhat less than 200 feet (60 m) of Fish Haven strata and 1,000 to 1,200 feet (300-365 m) of Laketown strata are exposed at the northern end of Wellsville Mountain (Oviatt, 1986b; Jensen and King, 1999). Budge (1966) measured 1,862 feet (568 m) of Laketown beds in the West Hills, 2 miles (3 km) west of Portage in the adjacent Limekiln Knoll quadrangle. The Fish Haven Dolomite thickens markedly to the west (to 1,020 feet [311 m] in the Promontory Mountains), but collectively Fish Haven and Laketown strata maintain a uniform thickness of about 1,440 to 1,776 feet (439-541 m) in northwestern Utah (Leatham, 1985). The upper contact of the Laketown Dolomite is not present on Clarkston Mountain. Elsewhere in north-central Utah, the Laketown Dolomite is disconformably overlain by the Early Devonian Water Canyon Formation. Undifferentiated Fish Haven and Laketown strata are also present in displaced blocks north of Little Canyon, on the ridge east of Gunsight Peak, and at Short Divide.

Both the Fish Haven and Laketown Dolomites are medium- to very thick-bedded, medium- to coarsely crystalline dolomite. Both formations contain black and light-brown chert nodules and lenses, and both contain rugose and tabulate corals, including *Favosites* sp. and *Halysites* sp. Crinoid columnals are common in upper Laketown beds. Based on conodonts, Leatham (1985) showed the base of the Fish Haven Dolomite is Late Ordovician to Early Silurian in age; brachiopods and corals also suggest a Late Ordovician age for Fish Haven strata. In northern Utah, the Laketown Dolomite is Early and Middle Silurian (Budge and Sheehan, 1980; Leatham, 1985) and may locally be latest Ordovician (Budge, 1966; Jensen and King, 1999). Fish Haven strata are probably less than 100 feet (30 m) thick at Clarkston Mountain. An incomplete section of Laketown Dolomite is in excess of 2,000 feet (600 m) thick at Clarkston Mountain.

Fish Haven Dolomite, Laketown Dolomite, and Swan Peak Formation, undifferentiated (SO)

Highly brecciated beds of the Fish Haven, Laketown, and upper Swan Peak Formations are present in fault blocks at Short Divide. Poor exposures and structural complexity preclude differentiating individual units.

Garden City, Swan Peak, Fish Haven, and Laketown Formations, Undifferentiated (SOu)

We mapped a block of undifferentiated Paleozoic bedrock at the southwest end of Clarkston Mountain, south of the Short Divide fault. Oaks recognized upper Swan Peak, Fish Haven/Laketown, and Garden City rock types, but poor exposures and structural complexity preclude differentiating individual units. The block may be a fault sliver caught up in the Short Divide fault, similar to blocks east of Water Canyon.

Pennsylvanian and Permian, Undifferentiated

Oquirrh Formation (PIPo)

Oquirrh Formation strata form steep, rounded slopes of the West Hills in the southwest corner of the Portage quadrangle, and are also exposed in the Junction Hills south of Short Divide. Much of the West Hills is shrouded in a thin mantle of colluvium, with good exposures restricted to widely scattered ledgy outcrops. We did not map the colluvium because of its thin, discontinuous nature and difficulty in distinguishing it from poorly exposed bedrock. In the West Hills, good exposures of Oquirrh Formation strata are generally restricted to areas where bedding attitudes were measured.

In the West Hills, the Oquirrh Formation consists of interbedded, blocky weathering, generally thin- to medium-bedded, medium- to dark-gray limestone and similarly bedded, medium-gray, very fine- to fine-grained sandstone. Both lithologies weather to colors of grayish orange, moderate yellowish brown, and pale to moderate reddish brown. Both calcareous and noncalcareous sandstones are present, and both are generally more poorly exposed than carbonate strata. An incomplete section of Oquirrh Formation strata estimated to be in excess of 6,000 feet (1,800 m) thick is present in the Portage quadrangle. No significant variations were noted that might allow subdivision of the Oquirrh Formation in the Portage quadrangle.

To the west, in the adjacent Limekiln Knoll quadrangle, Murphy and others (1985) divided the Oquirrh Formation into six informal units that total in excess of 14,900 feet (4,450 m) thick. In the Portage quadrangle, their unit 2 (Late Pennsylvanian to Early Permian) is present immediately north of Broad Canyon, while the overlying unit 3 (Early Permian) is exposed to the south, across a concealed west-trending, down-to-the-south normal fault. Both unit 2 and unit 3 consist of similar interbedded limestone and sandstone. Because no significant variation was noted in Oquirrh strata in the Portage quadrangle, these informal units were not used.

Oquirrh strata in the Junction Hills consist mainly of medium- to thick-bedded, medium- to dark-gray, finely crystalline limestone and sandy limestone. The latter appears to be reddish-orange to medium-brown sandstone on weathered surfaces. Chert nodules up to 4 inches (10 cm) in diameter are abundant locally, and display convex-up relics of original algal laminae. Sandstones similar to those in the West Hills are also present.

Tertiary

Wasatch(?) Formation (Tw?)

The Wasatch(?) Formation forms a poorly exposed, deeply dissected, mostly unconsolidated, brown to light-gray, coarse rubble sheet of variable thickness. These deposits are scattered throughout the north end of Clarkston Mountain, where they unconformably overlie Nounan, St. Charles, and Garden City strata; they are conspicuously absent from the southern half of Clarkston Mountain. Although the deposits at Clarkston Mountain lack the deep red colors typical of the Wasatch Formation in northern Utah and southeastern Idaho, we now believe these deposits belong to the Wasatch Formation based on tentative correlation with exposures in the Deep Creek graben to the north (Janecke and Evans, 1999). Earlier, Biek (1999) interpreted these deposits as late Tertiary to Quaternary boulder deposits. The query indicates our uncertainity.

The Wasatch(?) Formation drapes over adjacent map units so that the contact is everywhere concealed by colluvium and thicknesses are difficult to ascertain. The deposits may exceed 200 feet (60 m) thick immediately north of Elgrove Canyon, and may be up to 80 feet (24 m) thick where they cap ridges between North Canyon and Cold Water Canyon. Thick deposits may also be present along the northwest flank of the mountain in the NE1/4 section 3, T. 14 N., R. 3 W. Elsewhere, the unit probably forms a comparatively thin cover that drapes over ridge tops. In a few areas, such as along the crest of the ridge in the south-central portion of section 35, T. 15 N., R. 3 E., only scattered lag boulders remain. Locally, such as near the mouth of Old Quigley Canyon, the deposits appear to overlie the Salt Lake Formation. However, we believe that in these areas the Wasatch(?) Formation was remobilized and redeposited over Salt Lake strata, principally through a combination of mass movement and soil creep; we mapped such deposits as QTb, remobilized Wasatch(?) strata.

At Clarkston Mountain, the Wasatch(?) Formation consists principally of subrounded to rounded pebbles to boulders of locally derived lithologies, including Cambrian, Ordovician, and Silurian carbonates, orthoquartzites, and chert. The most distinctive clasts are rounded, white orthoquartzite cobbles to very large boulders from the upper Swan Peak Formation. Uncommon clasts have a red, coarse-sand matrix that adheres to protected recesses. Hanson (1949) noted that an appreciable part of these gravels was derived from strata not now present in the southern Malad Range, and that cobbles bearing Mississippian brachiopods and Pennsylvanian fusilinids are common. Rarely these deposits form poorly exposed, well-cemented, gray, pebble-conglomerate lenses with a calcareous, medium- to coarse-sand matrix.

The thick deposits on the north side of Elgrove Canyon are of interest in that they consist of two distinct, unconsolidated lithologic suites. Those in a small graben at the east half of the exposure contain abundant boulders of Swan Peak orthoquartzites and are representative of most other Wasatch(?) Formation deposits. Deposits immediately west of this graben consist almost exclusively of gray carbonate pebbles and boulders to 6 feet (3 m) in diameter; Swan Peak clasts are rare. Clasts in this western exposure are generally subrounded and coated with caliche. This sequence of carbonate-bearing boulder deposits overlain by orthoquartzite-bearing deposits is consistent with Wasatch stratigraphy in the Deep Creek graben to the north (Janecke and Evans, 1999).

Wasatch(?) strata are preserved in numerous fault blocks across the northern end of Clarkston Mountain, from the base to the crest of the range. These boulder deposits thus predate all or most of the normal faults in the range. Because they are not found on Salt Lake strata, the boulder deposits also likely predate the Salt Lake Formation. If our correlation is correct, these boulder deposits are Paleocene to Eocene synorogenic deposits of the Sevier fold and thrust belt (Oaks and Runnells, 1992; Coogan, 1992). Possible source rocks of the Wasatch(?) Formation at Clarkston Mountain are exposed in the West Hills and probably underlie Malad Valley.

Salt Lake Formation

The Salt Lake Formation is a laterally and vertically variable sequence of interbedded, generally white to light-gray, tuffaceous siltstone, sandstone, conglomerate, limestone, volcanic ash, and porcellanite. The Salt Lake Formation is widely, though generally poorly, exposed in the Junction Hills, along the south and northeast margins of Clarkston Mountain, at Bergeson Hill and the Washboards, and in the Elgrove Canyon embayment along the west flank of Clarkston Mountain. Regionally, earlier workers (see, for example, Adamson and others, 1955) subdivided Salt Lake strata into the basal Collinston Conglomerate, the widespread and generally tuffaceous deposits of the Cache Valley Formation, and the upper Mink Creek Conglomerate. Williams (1958, 1964) attempted to raise the Salt Lake to group status. However, like Smith (1997) to the south, Janecke and Evans (1999) to the north, and Goessel and others (1999) to the south, we treat Salt Lake strata as a formation. In this report, we divide Salt Lake strata into six informal subunits, ascending: conglomerate subunit of Bensons Hollow, micrite subunit of Short Divide, zeolite subunit of Long Divide, tephra subunit of Junction Hills, oolite subunit of Plymouth, and subunit of the Washboards.

Goessel (1999) recognized a basal conglomerate and three subunits of the Cache Valley Member in the three sections of the Salt Lake Formation that she and her thesis chairman R.Q. Oaks measured south of Bensons Hollow in the Clarkston and Cutler Dam quadrangles. Their measured section #3 (Goessel, 1999; Goessel and others, 1999) is tied to two ash beds with correlative age determinations, and the positions of three additional ash beds are added in their approximate stratigraphic position (figure 3). This sequence, aided by numerous tephra correlations (M.E. Perkins, unpublished data, table 1) enabled Oaks to recognize and map five subunits in the Junction Hills and in the highly faulted piedmont at the south end of Clarkston Mountain in the Clarkston and Portage quadrangles (figures 4 and 5). Where uncertain, these subunits are queried. Janecke and Evans (1999) recognized four members of the Salt Lake Formation in the Deep Creek half graben to the north in Idaho. These are the basal Skyline member (conglomerate with a tuffaceous matrix and interbedded limestone), Cache Valley Member (tuffaceous sedimentary rocks), Third Creek member (tephra with interbedded limestone and conglomerate), and the New Canyon member (conglomerate and gravel). Our preliminary correlations suggest that our conglomerate subunit of Bensons Hollow (Tslb) and micrite subunit of Short Divide (Tsls) are equivalent to the Skyline member of Janecke and Evans (1999), and that our zeolite subunit of Long Divide (Tsll) is equivalent in part to their Cache Valley Member. Lithologically similar deposits differ in age, preventing additional correlations between these two areas. In areas lacking detailed stratigraphic studies of the Salt Lake Formation elsewhere in the Clarkston and Portage quadrangles, we simply mapped Salt Lake strata as undifferentiated.

The Salt Lake Formation is a synextensional basin-fill deposit that records the onset of Miocene extension in this part of the Basin and Range Province (Bryant and others, 1989; Perkins and others, 1998; Janecke and Evans, 1999). Perkins and others (1998) used tephrochronology to establish a composite sequence of 213 tuffs ranging in age from 16 to 6 million years old in the Great Basin. Fifty-nine of these tuffs are widely distributed and help to establish a high-resolution stratigraphy of the Salt Lake Formation and other coeval basin-fill deposits. Perkins and others (1998) reported that the Cache Valley basin became active beginning about 11 to 8 million years ago and was one of many basins in which volcanic ashes derived from the Snake River Plain volcanic province accumulated. The oldest correlatable tuff in the Cache Valley basin is 10.94 ± 0.03 Ma and the youngest is about 4.4 to 5.1 million years old (Goessel and others, 1999)(for simplicity, once the full age is introduced, we drop the standard deviation following ages of correlated tuffs). The Salt Lake Formation is thus late middle Miocene to early Pliocene; as discussed below, the formation is younger still given the presence of about 6,000 feet (1,390 m) of Salt Lake strata above the youngest correlatable ash. Conglomeratic Salt Lake strata, including the conglomerate subunit of Bensons Hollow, were deposited locally along basin margins, whereas the widespread Cache Valley Member and correlative strata were deposited in streams and shallow lacustrine environments (Adamson, 1955; Goessel, 1999; Goessel and others, 1999).

Perkins and others (1998) calculated average sedimentation rates in Miocene basins of the northern Basin and Range of 920 feet (280 m) per million years, with values that ranged from 200 to nearly 2,000 feet (60-600 m) per million years. Based on ages of volcanic ashes determined by tephrochron-

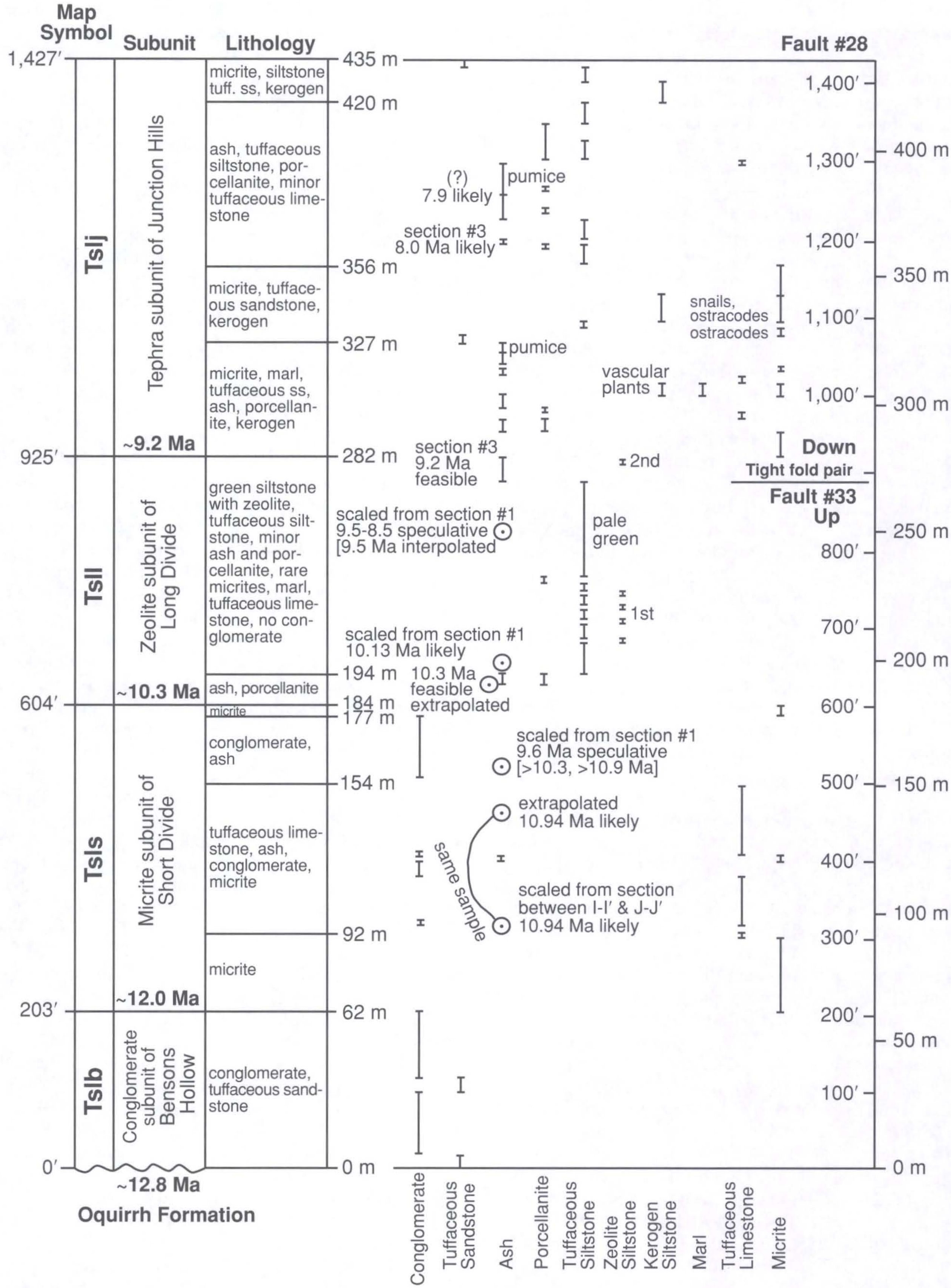


Figure 3. *Measured section #3 of the lower Salt Lake Formation on ridge south of Bensons Hollow, sections 10 and 15, T. 13 N., R. 2 W. From Oaks (2000).*

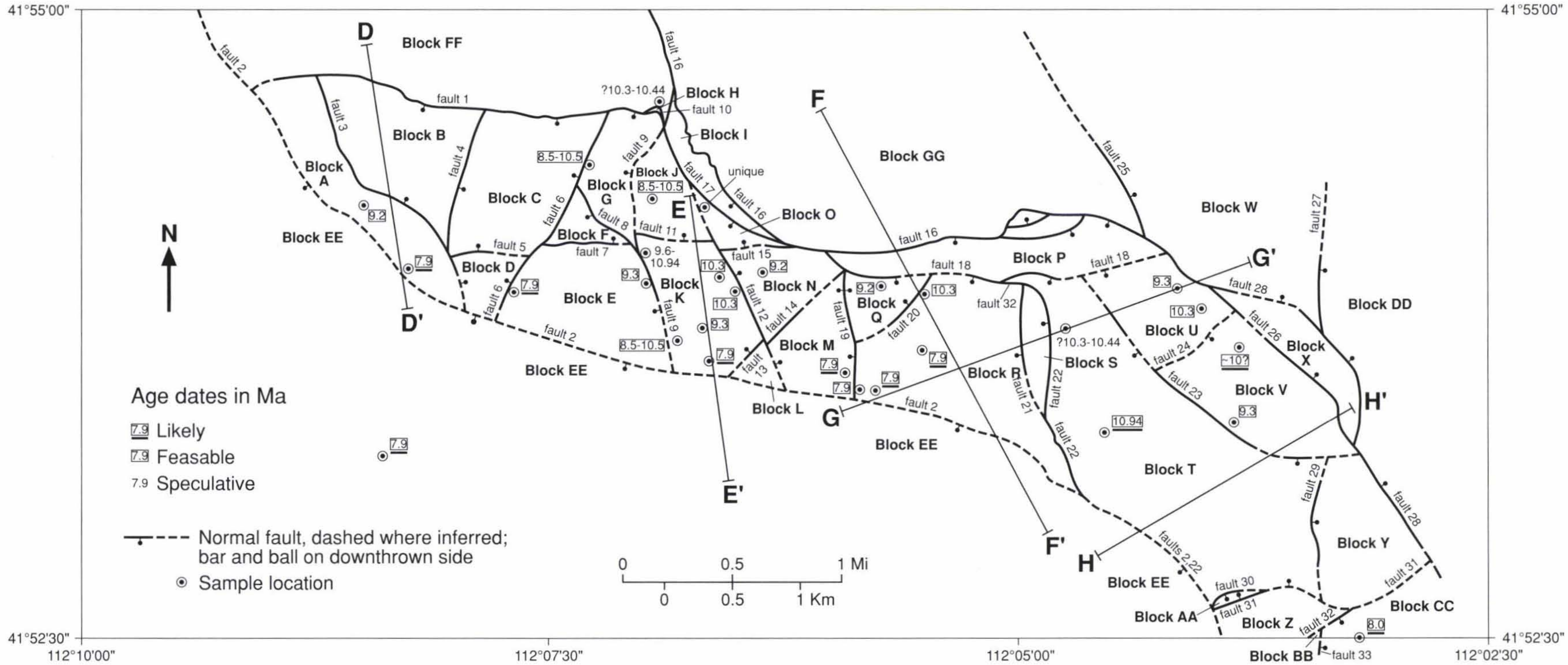


Figure 4. Fault map of the northern Junction Hills and Clarkston piedmont. See figure 5 for cross sections, plates 1 and 2 for locations. Faults are numbered and fault blocks are lettered from left to right. The locations and ages of ash samples are also shown. Ages are based on geochemical correlations with ashes of known ages, determined by Michael E. Perkins, University of Utah.

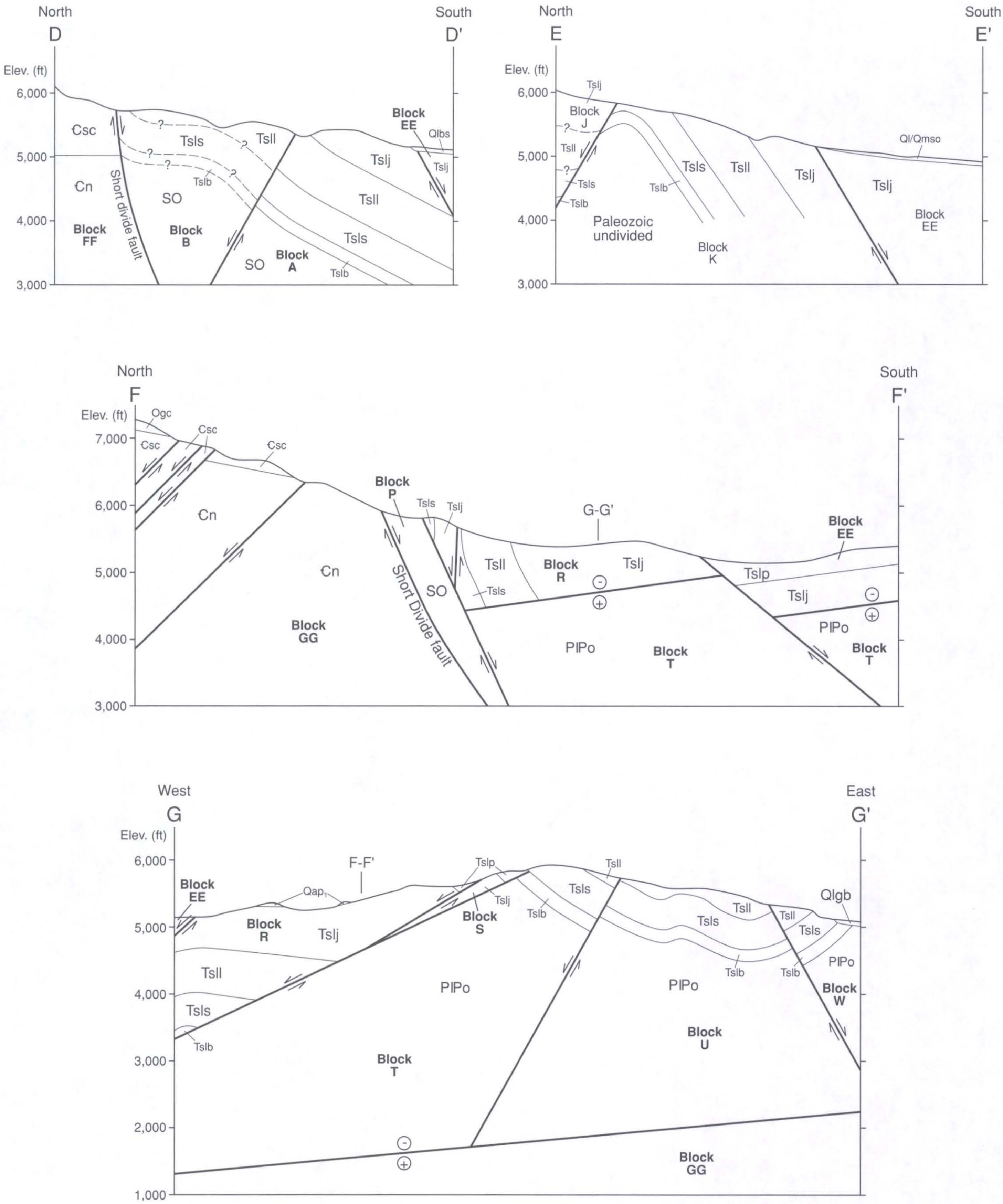
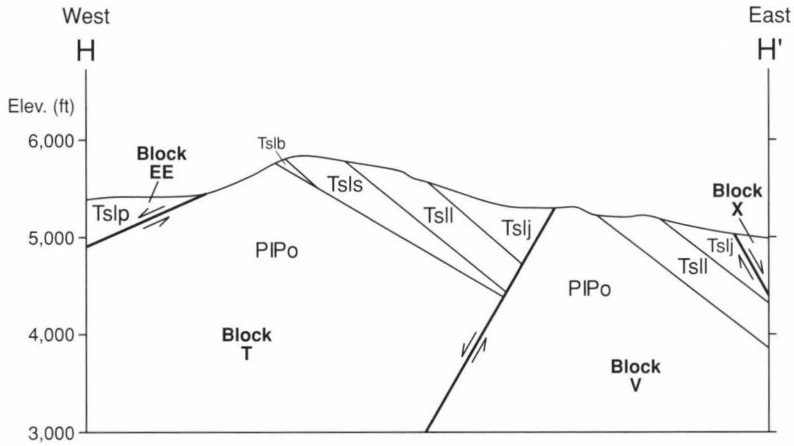


Figure 5. Cross sections of the northern Junction Hills and Clarkston piedmont (see figure 4 for locations). No vertical exaggeration. See text or plate 3 for symbol descriptions. Arrows show relative movement of faults; circle with dash shows fault with displacement into the page whereas circle with plus sign shows fault with displacement out of the page.

Figure 5. *(continued)*



ology and on stratigraphic thicknesses determined from geologic mapping, sedimentation rates in the northern Wellsville Mountains and along the south flank of Clarkston Mountain are between 425 and 690 feet (130-210 m) per million years, whereas that at measured section #3 was only about 250 feet (75 m) per million years (Goessel, 1999). A north-striking normal fault below two correlated tephras may omit strata in measured section #3. However, the fault is nearly parallel to bedding, so that stratigraphic displacement may be minor.

Mapping of Salt Lake strata in the Clarkston and Portage quadrangles, in combination with tephrochronologic dates, shows that sedimentation rates varied both over time and across the quadrangles. Rates of deposition after about 5 million years ago were much higher than those for the interval between about 11 and 6.4 million years ago based on the presence of about 6,000 feet (1,390 m) of Salt Lake strata above the 4.4 to 5.1 million-year-old ash (sample "BB" in the Bergeson Hills). Depositional rates in the past 5 million years were at least 2,180 feet (665 m) per million years, if deposition is assumed to end about 2 million years ago, or about 1,600 feet (490 m) per million years if deposition of Salt Lake strata continued to about 1 million years ago. Some post-depositional time is required to fold, fault, then erode the pediment surfaces across Salt Lake strata, then renew faulting and dissect the pediments. A rapid increase in the rate of deposition north of Clarkston is consistent with evidence, discussed in the structure section of this report, that a segment boundary exists between the Clarkston and Junction Hills segments of the West Cache fault zone, and that throw on the Clarkston segment increases northward.

Because of its deposition in a number of discrete basins, regional thicknesses of the Salt Lake Formation vary widely. We estimate the total thickness of Salt Lake strata in the Clarkston and Portage quadrangles is about 9,250 feet (2,820 m). Janecke and Evans (1999) reported that in the Deep Creek half graben to the north, their four members collectively exceed 7,265 feet (2,215 m) thick. To the south in the Junction Hills and northern Wellsville Mountains, Goessel (1999) reported at least 3,300 feet (1,000 m) of Salt Lake strata above about 900 feet (275 m) of the problematic Collinston Conglomerate, and Smith (1997) reported at least 4,710 feet (1,436 m) of Salt Lake strata in the southern Cache Valley.

Conglomerate subunit of Bensons Hollow (Tslb): The conglomerate subunit of Bensons Hollow consists of tuffaceous sandstone and tuffaceous pebble to cobble conglomerate exposed south of Short Divide, immediately west of the Box Elder-Cache County line. This subunit is 203 feet (62 m) thick just south of Bensons Hollow (Goessel, 1999; Goessel and others, 1999; Oaks, 2000), and is present at least locally along the Cache Butte Divide to the south. Smith and Oaks (see Smith, 1997) mapped correlative strata in the southern Cache Valley. The Bensons Hollow subunit is older than the 10.94 Ma ash "A," currently the oldest age in the Salt Lake Formation (table 1), which was collected from the lower part of the overlying Short Divide micrite subunit.

The conglomerate subunit of Bensons Hollow varies from very pale orange to pale grayish orange in color. Conglomerates in the Clarkston and Portage quadrangles are typically about 50 percent carbonate and chert and 50 percent siliciclastics, and clasts range from subrounded pebbles to angular boulders up to 6.5 feet (2 m) across. Although poorly exposed, rare outcrops exhibit shallow scours and medium to thick bedding. We reinterpret the bed of angular boulders at measured section #3, which was mapped as Oquirrh Formation bedrock by Oviatt (1986a), as a debris-flow deposit within the conglomerate subunit of Bensons Hollow.

Micrite subunit of Short Divide (Tsls): The micrite subunit of Short Divide consists of tuffaceous pebble to cobble conglomerate, tuffaceous limestone, very fine crystalline oolitic limestone (micrite), and lesser volcanic ash beds. Micrites near the middle of the subunit have distinctive chalky, white-weathered surfaces, but fresh surfaces are pale yellowish brown. This subunit is 400 feet (122 m) thick just south of Bensons Hollow and contains the 10.94 Ma ash "A," currently the oldest age from the Salt Lake Formation in Cache and Box Elder Counties. Ash "A" lies stratigraphically below these chalky micrites and is estimated to be about 410 feet (125 m) above Oquirrh strata. The micrite subunit of Short Divide is absent across a paleotopographic high where Oquirrh strata are exposed between Al and Bob Archibald Canyons.

Limestones of the Short Divide subunit range from pinkish gray and pale grayish orange to light brownish gray and pale olive gray. They are thin to medium bedded and exhibit wavy, planar bedding. The pebble to cobble conglomerates are dark brownish gray and typically contain about 60 percent siliciclastics and 40 percent carbonate and chert. They

Table 1. Tephrochronologic ages of volcanic ashes in Salt Lake Formation in the Clarkston and Portage quadrangles. Sample locations shown on plates 1 and 2.

LETTER DESIGNATION/ ERUPTIVE CENTER	ORIGINAL SAMPLE NUMBER	APPROXIMATE AGE (Ma) AND PROPOSED CORRELATIVE ASH [Tsl subunits]	CONFIDENCE	LATITUDE	LONGITUDE	STRIKE AND DIP	APPROXIMATE ALTITUDE	LOCATION	COMMENTS	COLLECTED BY
BB Supra-Walcott	98-03 (4-29-98)	4.75+/-0.35 Santee [Tsl]	Likely	41° 58.14'	-112° 00.23'	No data (128/31SW)	5390'	E edge of crest, in road cut	South of Ogc outcrops; likely correlation with ash CV 96-880; currently youngest date in Tsl	RQO
AA Late TF	98-01 (4-23-1)	7.9 Rush Valley [Tslj]	Likely	41° 53.22'	-112° 08.36'	23/24SE	4845'	N side of road E from Nucor Steel Plant, above Qlbp	Rare outcrop below Qlgb shoreline, same age as in scarps facing Qlgb ~1 mile to NE; likely correlations with ashes U, V, W, X, Y, Z, 9-18-14 pg, Cache A (KS)	RQO
Z Late TF	97-50 (11-6-1)	7.9 Rush Valley [Tslj]	Likely	41° 53.65'	-112° 05.50'	108/38SW	5460'	E side wide gully SW of Short Divide	Above 10.3 Ma (97-12), below Tso; likely correl's with ashes U, V, W, X, Y, AA, 9-18-14 pg, Cache A (KS)	RQO
Y Late TF	98-20 (9-24-98)	7.9 Rush Valley [Tslj]	Likely	41° 53.48'	-112° 05.78'	No data	5270'	Small gully E of gully with fault	Dirty ash, may be reworked	SUJ/ Bob Biek
X Late TF	98-10 (9-24-98)	7.9 Rush Valley [Tslj]	Likely	41° 53.55'	-112° 05.91'	No data	5430'	SW of Short Divide, N of mouth of gully	W side of gully with fault, NW of 8-15-1 (9.3 Ma), WNW of 98-20 (7.9 Ma); likely correlations with ashes U, V, W, Y, Z, AA, 9-18-14 pg, Cache A (KS)	RQO
W Late TF	10-8-1	7.9 Rush Valley [Tslj]	Likely	41° 53.61'	-112° 06.62'	85/57SE	5260'	E side of canyon E of Water Canyon, at mouth	Likely correlations with ashes U, V, X, Y, Z, AA, 9-18-4 pg, Cache A (KS)	LMS/RQO
V Late TF	97-45	7.9 Rush Valley [Tslj]	Likely	41° 53.88'	-112° 07.69'	45/04SE	5330'	E side of gully W of Bishop Canyon	Likely correlations with ashes U, W, X, Y, Z, AA, 9-18-4 pg, Cache A (KS)	RQO
U Late TF	97-07 (7-21-1)	7.9 Rush Valley [Tslj]	Likely	41° 53.96'	-112° 08.26'	43/30SE	5190'	NW side of mouth of 2nd gully SE of Mine Hollow	Likely correl's. with ashes V, W, X, Y, Z, AA, 9-18-14 pg, Cache A (KS)	RQO
T TF	8-17-5	(>7.9) 8.5-10.5 (<9.3) RV88-9B [Tslj]	Feasible	41° 53.77'	-112° 06.77'	109/40NE	5400'	E scarp of landslide, mouth of Water Canyon	Between 7.9 and 9.3 Ma dates; feasible correlation with ash S	LMS
S TF	8-31-1	8.5-10.5 (<9.3) RV88-9B [Tslj?] Loc'n not secured	Feasible	41° 54.36'(?)	-112° 07.27'(?)	30/30SE	5820'(?)	Bend in road, upper part of Bishop Canyon Road, road cut	Feasible correlation with ash T	LMS
R TF	98-02 (4-26-98)	9.2 TC90-25 [Tsll]	Feasible	41° 53.96'	-112° 06.37'	No data (185/41E)	5790'	E side 2nd canyon east of Water Canyon	Likely correlation with ashes P, Q	RQO
Q TF	97-08 (7-21-2)	9.2 TC90-25 [Tsll]	Feasible	41° 54.22'	-112° 08.48'	90/43S	5240'	E side of Mine Hollow, near mouth	Likely correlations with ashes P, R	RQO
P TF	8-13-2	9.2 TC90-25 Loc'n not secured [Tsll]	Feasible	41° 53.91'(?)	-112° 05.68'(?)	30/32SE	5790'(?)	S face of W-most knoll W of Short Divide	Location is in area of SE-facing vertical beds and NW of 10.3 Ma; anomalous unless faulted; likely correlations with ashes Q, R	LMS
O Late TF	97-43 (10-28-2)	9.3 MCMI [Tsll]	Feasible	41° 53.37'	-112° 03.86'	91/08N	5670'	Just above scarp at head of landslide	Slightly above PIPo outcrops; likely correlations with ashes M, N, 7-16-7 pg, 9-11-1 pg	RQO
N Late TF	97-56	9.3 MCMI [Tsll]	Feasible	41° 53.90'	-112° 04.16'	152/41NE	5565'	N-S head of E-W gully	Likely correlations with ashes M, O, 7-16-7 pg, 9-11-1 pg	RQO
M Late TF	9-5-1	9.3 MCMI [Tsll]	Feasible	41° 53.92'	-112° 06.96'	No data	5530'	Trail along Water Canyon N scarp of landslide	Likely correlations with ashes N, O, 7-16-7 pg, 9-11-1 pg	LMS
L Late TF	10-8-96	9.3 MCMI [Tsll]	Feasible	41° 53.75'	-112° 06.67'	95/53SW	5460'	W side of canyon E of Water Canyon, opposite side canyon	Feasible correlations with ashes U, V, W, X, Y, Z, AA, 9-18-14 pg, Cache A (KS)	RQO
K Late TF	8-15-1	9.3 or 7.9 MCMI or Rush Valley [Tslj]	Feasible	41° 53.50'	-112° 05.84'	50/60SE	5230'	Mouth of valley along fault, at slope break, SW of Short Divide	9.3 Ma incompatible with 7.9 Ma just E, W, & NW	LMS

Table 1. *(continued)*

LETTER DESIGNATION/ ERUPTIVE CENTER	ORIGINAL SAMPLE NUMBER	APPROXIMATE AGE (Ma) AND PROPOSED CORRELATIVE ASH	CONFIDENCE	LATITUDE	LONGITUDE	STRIKE AND DIP	APPROXIMATE ALTITUDE	LOCATION	COMMENTS	COLLECTED BY
J TF	9-17-2	~9.6 (older??) RV88-6 [Tsls?]	Speculative	41° 54.03'	-112° 06.98'	70/26SE	5580'	Trail along Water Canyon, start of tight meanders, S of spring boxes	Light gray, shattered, slickenlined Tsl just NW; feasible correlations with ashes H, I, 8-7-7 pg, 8-27-11 pg; probably >9.6 Ma	LMS/RQO
I TF	10-11-96	~9.6 (older??) RV88-6 [Tsls?]	Speculative	41° 54.63'	-112° 06.89'	No data	6400'	N of unvegetated ridge at E side of Water Canyon	About 20' above OSu outcrops in fault sliver in Water Canyon; feasible correlations with ashes H, J; speculative with ash 7-11-4 pg; probably >9.6 Ma	RQO
H TF	7-16-1	~9.6 (older??) RV88-6 [Tsls?]	Speculative	40° 53.74'	-112° 04.75'	133/57NE	5955'	~200' W of radio tower, S of Short Divide	Footwall of fault scarp; feasible correlations with ashes I, J, 7-11-4 pg; probably >10.3 and <10.94 Ma	LMS
G TF	97-14 (4-15-1)	~10? [Tsll?] ONN94-628 Loc'n not secured	Likely	41° 53.68'(?)	-112° 03.83'(?)	75/35NW	5290'(?)	Spur ridge S of Al Archibald Hollow		LMS
F TF	9-17-4	8.5-10.5 RV93-244 [Tsll?]	Speculative	41° 54.23'	-112° 06.94'	120/25NE	5870'	NE into spur creek of Water Canyon	Feasible correlation with ash Cache M (KS)	LMS/RQO
E TF	10-8-4	10.3 TC90-18A [Tsll]	Feasible	41° 53.89'	-112° 06.48'	70/47SE	5650'	W side of spur canyon off Canyon E of Water Canyon	Near top of thick ash; likely correlations with ashes D, 97-06 pg	LMS
D TF	10-8-97	10.3 TC90-18A [Tsll]	Feasible	41° 53.94'	-112° 06.57'	109/64SE	5680'	E side of canyon E of Water Canyon	Base of thick ash sequence; likely correlations with ashes E, 97-06 pg	RQO
C TF	97-13 (4-12-5 ½)	10.3 TC90-18 [Tsll] Loc'n not secured	Feasible	41° 53.82'(?)	-112° 04.03'(?)	No data	5440'(?)	Spur ridge N of Al Archibald Hollow	Likely correlation with ash B	LMS
B TF	97-12 (5-10-97)	10.3 TC90-18 [Tsll]	Feasible	41° 53.89'	-112° 05.51'	35/61NW(O/T)	5750'	SE face of W-most knoll W of Short Divide	Close to NE-SW fault; cf. sample P; likely correlation with ash C	LMS/RQO
A Bruneau-Jarbridge	98-04 (5-2-98)	10.94 Cougar Point Tuff XIII [Tsls]	Likely	41° 53.32'	-112° 04.55'	No data	5750'	SW of saddle in ridge S of Short Divide	Slightly above Tslb basal cgl and underlying Po; currently oldest date in Tsl	RQO
Unique	10-8-6	[unique]	-	41° 54.18'	-112° 06.34'	180/15E	6000'	E side of spur canyon of Water Canyon, near head		LMS

Eruptive centers of volcanic ashes:
Bruneau-Jarbridge = mid-late Miocene volcanic field of the Snake River Plain
TF = Twin Falls volcanic field of the Snake River Plain/late Miocene
Late TF = distinctive ashes produced toward the close of TF eruptions
Walcott = Heise volcanic field of the Snake River Plain/late Miocene
Supra-Walcott = late Miocene - early Pliocene

Processed ashes were analysed with electron microprobe at University of Utah, Department of Geology and Geophysics, and results were compared to results from ashes of known radiometric ages and evaluated by M.L. Perkins.

Explanation:
LMS = L. Marie Swenson
RQO/RO = Robert Q. Oaks, Jr.
PG = Kathryn M. (Piper) Goessel
(KS) = Kristine A. Smith
SUJ = Susanne U. Janecke

contain subequal amounts of Oquirrh carbonate, other Paleozoic carbonates, chert, and Swan Peak quartzite.

Zeolite subunit of Long Divide (Tsll): The zeolite subunit of Long Divide consists mainly of thin- to medium-bedded tuffaceous siltstone and sandstone with volcanic ash and porcellanite at the base and volcanic ash at the top. Unweathered surfaces of the tuffaceous siltstone and sandstone are white to pale greenish yellow, whereas those of the ash and porcellanite are white to very pale orange and pale olive gray. Most volcanic ashes or reworked ashes are altered to zeolite-bearing pale-yellowish-green siltstone, which characterizes this subunit. In and west of block "R" (figure 4), the zeolite subunit of Long Divide is about 700 feet (215 m) thick, over twice as thick as in section #3. This difference in thickness is believed to reflect a higher average rate of deposition north and northwest of measured section #3. No micrite, marl, or tuffaceous limestone was found in this subunit in section #3 (Goessel, 1999; Goessel and others, 1999; Oaks, 2000), although a fault is probably present in section #3 at the base of the lower (9.24 ± 0.50 Ma) ash. North of measured section #3, this subunit contains two distinctive, brown-weathering, rough ("nubbly") oncolitic micrite beds, each up to 6 feet (2 m) thick. One lies near the middle and the other near the top of the zeolite subunit, the latter just above a widespread 9.24 Ma ash.

Tephra subunit of Junctions Hills (Tslj): The tephra subunit of Junction Hills contains numerous volcanic ashes, as well as tuffaceous siltstone, porcellanite, tuffaceous limestone, marl, and micrite. Colors in the tephra subunit generally vary from white to very pale orange to light greenish gray, and except for thicker bedded volcanic ash and thinner bedded porcellanite, most bedding is characterized by thin to medium, planar beds. Kerogen is present in several beds, and casts and molds of shells of ostracodes and snails are common. Tuffaceous pebble to cobble conglomerate is common near the top of the subunit in block "R" (figure 4). The clasts consists of Paleozoic carbonate and Precambrian quartzite in highly variable proportions. Eight tephrochronologic correlations to a 7.9 ± 0.50 Ma ash come from this subunit. The tephra subunit of Junction Hills is at least 502 feet (153 m) thick in measured section #3. The thickness of the Junction Hills subunit is estimated to be 2,550 feet (780 m) in section C-C′ (plate 3). This is based on the total thickness of 3,250 feet (990 m) between the 10.3 ± 1.0 Ma ash "B" and the base of the Plymouth oolite subunit, minus an estimated 700 feet (215 m) of the Long Divide zeolite subunit above the 10.3 Ma ash in block "R."

Oolite subunit of Plymouth (Tslp): The oolite subunit of Plymouth consists of very pale-orange to yellowish-gray, porous, oolitic limestone that contains pebbles and small cobbles of Paleozoic bedrock and older Salt Lake strata, as well as casts and molds of low-spired robust snails. The middle part consists of micritic limestone and less common thin interbeds of pebble and cobble orthoconglomerate and tuffaceous marl and siltstone. One pebble count by Goessel (1999) at her section #2 yielded 72 percent limestone, 6 percent Oquirrh sandstone, 20 percent chert, and 2 percent Precambrian quartzite. Oolitic limestone is generally medium to thick bedded, and locally exhibits shallow, channel-form bedding with planar cross-laminae. The oolite subunit of Plymouth is named for exposures in measured section #2, which is just south of Bensons Hollow in the adjacent Cutler Dam quadrangle (Goessel, 1999; Goessel and others, 1999). In the Clarkston quadrangle, the oolite subunit of Plymouth is restricted to exposures immediately south-southwest of Short Divide where only one oolitic sequence was encountered in block "R" (figure 4).

Goessel (1999) and Goessel and others (1999) reported that the oolite subunit of Plymouth is probably younger than 6.4 ± 0.10 Ma on the flanks of the Wellsville Mountains. Extrapolation of sedimentation rates along the south flank of Clarkston Mountain suggests that the base of the oolite subunit of Plymouth there, in section C-C′, is between about 5.0 and 5.2 million years old. The oolite subunit of Plymouth is at least 210 feet (64 m) thick in section #2 (Goessel, 1999).

Subunit of the Washboards (Tslw): The subunit of the Washboards is informally named for the west-dipping, corrugated surface north of Clarkston known as the Washboards. The subunit of the Washboards is also present near the Bonneville shoreline northwest of Clarkston. These deposits are poorly exposed, except along the southeast margin of the Washboards, but are readily identifiable by their distinctive clast composition. They consist of mostly very light-gray, generally thin- to medium-bedded, calcareous clay and silt with lesser interbedded sand, pebbly sand, and gravel. The clasts are quartzites from the Proterozoic and Cambrian Mutual and Geertsen Canyon (Camelback Mountains) Formations, green carbonate nodules from Cambrian shales, Ordovician and Cambrian carbonates, black chert from Ordovician and Silurian carbonates, rare granitic igneous rocks, and resistant, greenish Salt Lake mudstone clasts. This diverse lithologic assemblage contrasts with the less diverse clasts in alluvial-fan deposits shed off Clarkston Mountain.

The subunit of the Washboards appears to be poorly cemented, but the overall resistance of the Washboards suggests better cementation than is shown in most exposures. Bedding at the Washboards dips more steeply west than the planated, west-sloping surface of the Washboards itself, which is covered by a veneer of lag gravels and loess. The subunit of the Washboards probably represents the youngest strata of the Salt Lake Formation based on its general conformity with the Salt Lake Formation at Bergeson Hill and similar lithology to other parts of the Salt Lake Formation. The apparent lack of significant volcanic ash is in marked contrast, however, to typical Salt Lake strata. The subunit of the Washboards is younger than the 4.4 to 5.1 million-year-old ash at Bergeson Hill and is probably Pliocene. The subunit of the Washboards may be as young as Quaternary, but lacking definite ages we restrict it to the late Tertiary. These deposits are estimated to be 3,000 to 4,000 feet (915-1,220 m) thick (Oaks, 2000).

Salt Lake Formation, undifferentiated (Tsl): Lacking detailed stratigraphic studies, we mapped the Salt Lake Formation as undifferentiated along the north and west flanks of Clarkston Mountain, and at Bergeson Hill. In the Elgrove Canyon embayment, Salt Lake strata are mostly very poorly lithified, calcareous, laminated, silty, fine-grained sandstone that contains small pebble-size dropstone clasts, common small snail shells, and local iron-manganese stains. These beds appear to be deeply weathered, gently west-dipping Salt Lake strata that may be overlain by a discontinuous veneer of lacustrine sediment of the Bonneville lake cycle. Identical beds are well lithified to the south, east of Mountain Spring.

Salt Lake strata present in the graben at the north end of Clarkston Mountain are poorly exposed, but include medium-bedded, white to light-gray limestone. North of the main range-bounding fault, the Salt Lake Formation forms a north-northeast-dipping sequence of light-gray to white, variably calcareous, thin- to medium-bedded, tuffaceous siltstone, mudstone, and fine-grained sandstone. Similarly colored calcareous tuff and finely crystalline limestone are also present. These beds are locally altered and baked by basaltic intrusions, described below. North and west of Dirty Head, Salt Lake strata are locally concealed by a discontinuous boulder lag that may be derived from Wasatch(?) or remobilized Wasatch(?) strata.

Salt Lake strata present around the west flank of Bergeson Hill consist primarily of siltstone, marl, volcanic ash, and conglomerate. This sequence contains the 4.4 to 5.1 million-year-old ash "BB," currently the youngest correlated tephra from the Salt Lake Formation in this area, and underlies the poorly consolidated, gravel-bearing deposits of the subunit of the Washboards (Tslw). This correlation implies that these beds may lie stratigraphically above the oolite subunit of Plymouth. The presence of Garden City strata at Bergeson Hill can be explained by a down-to-the-west normal fault west of those outcrops, or paleotopography. The regular westward slope of the Bouguer gravity anomalies beneath the Washboards suggests a uniform westward dip of the contact between Salt Lake and Paleozoic strata, although two down-to-the-west normal faults may be present between the Washboards and Bergeson Hill (Oaks, 2000). Conservative projection of dips westward beneath the Washboards suggest the presence of at least 6,000 feet (1,830 m) of Salt Lake strata to the west. Salt Lake strata on the west side of Bergeson Hill may be as much as 2,500 feet (760 m) thick.

Basaltic Sills in Salt Lake Formation (Tb)

Basaltic sills are exposed near the entrances to North Canyon, Steel Canyon, and Gowans Hollow in the northwest corner of the Clarkston quadrangle. Adamson (1955) described the sill exposed at the entrance to North Canyon. Prammani (1957) mapped two of these sills at Steel Canyon and Gowans Hollow, and noted additional basaltic rocks across the border in Idaho that lack alteration zones. Burton (1973) also mapped the general locations of the sills. The sills, which vary from less than 1 to 10 feet (<1-3 m) thick, are light-gray to light-olive-gray, very fine-grained basalt that weathers to mottled colors of brown, reddish brown, and gray. The basalt is commonly vesicular, and is oriented parallel or subparallel to bedding. Both upper and lower contacts reveal baked and altered Salt Lake strata. The sill immediately north of the entrance to Steel Canyon yielded a K-Ar age of 8.0 ± 0.5 Ma (Don Fiesinger, Utah State University, written communication, March 8, 1999).

Although inconspicuous and generally poorly exposed, the sills are readily located because they are associated with dark-colored, altered Salt Lake strata. These baked and altered zones are particularly well exposed on the north side of Steel Canyon. There, they grade from a thin, black glass with pervasive conchoidal fractures; to greenish-black, light-olive-gray, and yellowish-gray, conchoidally fractured porcellanite and mudstone; to overlying mudstone beds with a phyllonitic sheen. Collectively these altered beds are up to several tens of feet thick. The sills and altered beds commonly have iron-manganese stains and so contrast sharply with typically white Salt Lake strata. Salt Lake strata are similarly stained elsewhere in the Steel Canyon area although baked sediments or basalt were not found.

Quaternary and Tertiary

Remobilized Wasatch(?) Formation (QTb)

We interpret Wasatch(?)-like strata that overlie the Salt Lake Formation near the mouth of Old Quigley Canyon as remobilized from thick, upslope Wasatch(?) deposits. Because of the unconsolidated nature of most Wasatch(?) strata, contacts between Wasatch(?) and remobilized Wasatch(?) are unclear; we simply restrict remobilized Wasatch(?) (QTb) to areas underlain by Salt Lake strata. As discussed for the Wasatch(?) Formation previously, much of the Wasatch(?) itself appears to be in part remobilized. These remobilized deposits are probably several tens of feet thick.

Alluvial-Fan Deposits (QTaf)

Alluvial-fan deposits of uncertain Quaternary to late Tertiary age are present at the entrance to Mikes Canyon on the east flank of Clarkston Mountain. These deposits are in an embayment in the mountain front formed by en-echelon offset of the Clarkston fault. The alluvial-fan deposits consist of poorly sorted, clay- to boulder-size sediment that forms a steeply sloping, deeply dissected surface. The deposits were mapped as Tertiary undifferentiated by Solomon (1999), but he noted that they may be in part of Pleistocene age. Inconclusive geomorphic evidence suggests that these deposits may be cut by poorly preserved, unmapped fault scarps. The exposed thickness is less than 100 feet (30 m).

Fault Breccia (QTbx)

Along the Clarkston Mountain segment of the Wasatch fault zone, Nounan and St. Charles dolomites are highly fractured and locally form well-developed fault breccias. These well-cemented breccias vary from 0 to about 6 feet (0-2 m) thick and form planar, west-dipping surfaces on faceted spurs at the mountain front (figure 6). Unmapped, comparatively small talus deposits commonly conceal the base of these breccias. We mapped similar, though thicker and less well exposed fault breccia, just east of the entrance to Mikes Canyon along the Clarkston segment of the West Cache fault zone; this breccia, developed in St. Charles strata, is partly concealed by alluvial-fan deposits believed to be late Tertiary to Quaternary in age. We also mapped a single exposure of fault breccia developed in Salt Lake strata along the Clarkston fault north of Old Quigley Canyon. Although fault breccias are common along many well-exposed faults in the Clarkston and Portage quadrangles, we mapped these particular fault breccias because (with the exception of exposures near Mikes Canyon) they form relatively large, planar surfaces that are more resistant to erosion than adjacent strata.

Quaternary

Alluvial Deposits

Pediment-mantle deposits (Qap_1, Qap_2, Qap): We mapped south-sloping, planar surfaces – armored with unconsol-

Figure 6. View north along the Clarkston Mountain segment of the Wasatch fault zone from Castlegate Canyon. Note west-dipping fault breccia in foreground that caps faceted spur developed on east-dipping middle St. Charles strata. Alluvial-fan deposits in the middle distance overlie a bedrock bench developed on the Salt Lake Formation.

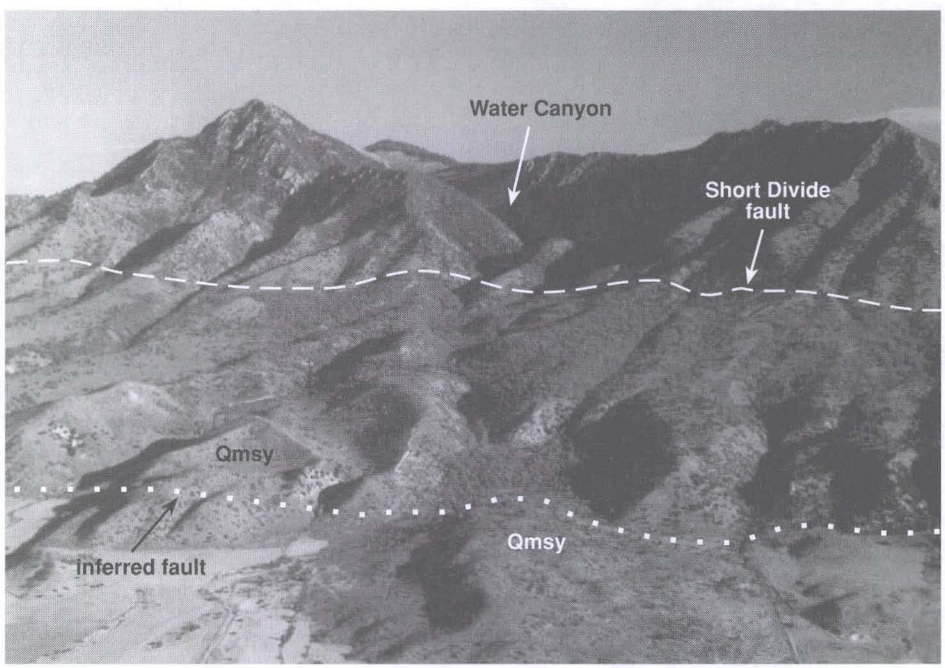


Figure 7. View north of Clarkston Mountain. Pediments south of the Short Divide fault are covered by loess and truncate complexly faulted and folded Salt Lake strata. An inferred down-to-the-south fault (dotted line), parallel and coincident with the Bonneville shoreline, in turn truncates the pediment surfaces. Large landslides (Qmsy) at the mouth of Water Canyon cover the Bonneville shoreline. The Gunsight fault trends northwest through the upper reaches of Water Canyon, and places undifferentiated Ordovician to Silurian Fish Haven-Laketown strata down on the west against the Cambrian Nounan and St. Charles Formations to the east.

idated, poorly exposed pebble- to small-boulder gravels – along the south flank of Clarkston Mountain (figure 7). These surfaces truncate bedding in the Salt Lake Formation, and in turn are truncated by steep slopes as much as 400 feet (120 m) high along their down-gradient ends, parallel and coincident with the Bonneville shoreline. Incision decreases up-gradient to a few tens of feet at the mountain front. The deposits appear to maintain a relatively constant thickness of up to 20 feet (6 m). The morphology and characteristics of the deposits suggest a pediment-mantle origin blanketed by loess. They are graded to a level well above the Bonneville shoreline and so predate Lake Bonneville.

Pediment-mantle deposits are at various levels and slightly different slope angles. Where in contact with one another, they can be readily grouped into two levels: an older, higher surface (Qap_2) into which is inset a slightly younger, lower surface (Qap_1). Elsewhere, correlation of surfaces across deep ravines becomes problematic and so the deposits are locally mapped as undifferentiated pediment-mantle deposits (Qap). In mapping these pediments, Oaks also differentiated older alluvial deposits slightly inset into the younger pediment surface. Biek chose to lump these generally small surfaces into Qap_1 to simplify an already complex portion of the map. The formation of the pediment deposits is discussed more fully under the structural geology section of this report, and they are shown on plate 1 of Goessel (1999).

The age of pediment formation is likely younger than 4.4 to 5.1 million years based on a dated ash (Smith, 1997) in truncated Salt Lake strata below the McKenzie Flat pediment surface south of Avon in southern Cache Valley. Clear evidence for only one widespread episode of tectonic quiescence and pediment formation in the region suggests that the pediment surfaces along the south flank of Clarkston Mountain are part of the McKenzie Flat pediment. From the great thickness of Salt Lake strata stratigraphically above the 4.4 to 5.1 million-year-old ash in the Clarkston quadrangle, projected westward beneath the Washboards, the pediments in the Clarkston Mountain area likely formed within about the past 2 million years and are therefore Quaternary in age.

Older gravel deposits (Qag): Unconsolidated, caliche-coated, pebble- to cobble-size clasts of Salt Lake Formation strata in a fine-grained matrix are present locally on Bergeson Hill, in the northeast corner of the Clarkston quadrangle. The deposits are not related to modern drainages, and probably represent a lag deposit of resistant lithologies eroded from the Salt Lake Formation. Older gravel deposits are less than about 10 feet (3 m) thick.

Older alluvial deposits (Qao): Gravelly alluvial deposits of uncertain age unconformably overlie Salt Lake strata near the state line in Steel Canyon. These deposits consist of moderately sorted sand and subangular to subrounded, pebble- to cobble-size gravel that locally forms well-cemented ledges several feet thick. Clasts are mostly carbonates, although green-weathering siliceous mudstones are common. The matrix of these deposits is white to light gray, similar to adjacent Salt Lake strata, and the clasts are coated with carbonate. They lie from a few feet to several tens of feet above the base level of Steel Canyon and tributary drainages, and probably represent the remains of exhumed channel fill. We also mapped older alluvial deposits near Short Divide, where they were reworked from pediment-mantle deposits and from Paleozoic bedrock. Older alluvial deposits are probably less than 40 feet (12 m) thick.

Stream-terrace deposits (Qat_1): Stream-terrace deposits are restricted to modern drainages, where they form level to gently sloping surfaces generally less than 30 feet (10 m) above the modern flood plain. These deposits consist of moderately to well-sorted sand, silt, clay, and local pebble to boulder gravel deposited principally in river-channel and flood-plain environments. They vary from 0 to 20 feet (0-6 m) thick and are incised by alluvial deposits (Qal).

Alluvial deposits (Qal): We mapped alluvial deposits along the Malad River, Clarkston Creek, and other principal drainages in the quadrangles. They consist of moderately to well-sorted sand, silt, clay, and local gravel normally less than about 20 feet (6 m) thick. Alluvial deposits include river-channel and flood-plain sediments and minor terraces up to about 10 feet (3 m) above current stream levels; small alluvial-fan and colluvial deposits too small to map separately are included in this map unit. Deposits along the two major drainages in the area, the Malad River and Clarkston Creek, are marked by numerous meander scars and oxbow lakes. Alluvial deposits are gradational with mixed alluvial and colluvial deposits.

Older alluvial-fan deposits (Qafo): Older alluvial-fan deposits form isolated, incised surfaces along the flanks of Clarkston Mountain, in the Dirty Head area between the Washboards and Clarkston Mountain, and in the West Hills. The fan deposits themselves consist of poorly to moderately sorted, boulder- to clay-size sediments derived from up-gradient drainage basins, and thus vary considerably in clast composition. Adjacent to the Clarkston Mountain front, the deposits appear to be bounded on their up-gradient side by either the Clarkston fault of the West Cache fault zone or the Clarkston Mountain segment of the Wasatch fault zone; the upper portions of these fans are normally covered by unmapped colluvial deposits that obscure the fault itself.

Based on cross-cutting relationships and relative height above modern drainages, the age of these deposits varies considerably, from pre- to post-Bonneville. In the Clarkston quadrangle, older alluvial-fan deposits at lower elevations are truncated by the Bonneville shoreline and are thus older than 14.5 ka. Isolated, fault-bounded deposits along the east flank of Clarkston Mountain are thought to be of similar age based on morphology and stratigraphic position. In the West Hills, two small deposits, one of which is cut by the Bonneville shoreline, lie in excess of 200 feet (60 m) above adjacent drainages and thus may be considerably older than 14.5 ka. In contrast, most older alluvial-fan deposits along the west flank of Clarkston Mountain lie below the Bonneville shoreline and are probably in part younger than 14.5 ka. The small deposit at the entrance to Little Canyon, north of Plymouth, is graded to the Bonneville shoreline and is therefore about 14,500 years old. These deposits were mapped simply as undivided older alluvial-fan deposits due to inadequate age constraints.

Older alluvial-fan deposits are distinguished from younger alluvial-fan deposits on the basis of morphology and elevation above modern drainages. Most older alluvial-fan deposits are simply the isolated and incised remnants of large, coalesced alluvial-fans. The exposed parts of older alluvial-fan deposits are several tens of feet thick.

Level 2 alluvial-fan deposits (Qaf_2): We mapped inactive alluvial-fan remnants on the eastern piedmont slope of the Junction Hills below the Bonneville shoreline near Al Archibald Hollow and Bensons Hollow; we also mapped small deposits above the Bonneville shoreline at Dirty Head in the Clarkston quadrangle, and at the Provo shoreline southwest of Oregon Springs in the Portage quadrangle. The inactive fan deposits consist of poorly to moderately sorted clay- to boulder-size sediments. The southern deposits bury Lake Bonneville sediments, and lacustrine shorelines are absent on surfaces formed by this unit. After fan deposition, drainage from the hollows incised underlying deposits, isolating fan remnants on slopes above modern channels. These inactive fan remnants are therefore older than modern alluvial-fan deposits (Qaf_1) but younger than Lake Bonneville deposits, ranging in age from middle Holocene to latest Pleistocene. Inactive fan deposits typically thin downslope, having an exposed thickness less than 15 feet (5 m).

Modern alluvial-fan deposits (Qaf_1): Modern alluvial-fan deposits form active, isolated alluvial fans throughout the Cache and Malad Valleys. They consist of poorly to moderately sorted clay- to boulder-size sediments deposited principally by debris flows at the mouths of active drainages. These fans are active post-Bonneville depositional surfaces, although somewhat older sediments may be present at depth. Most modern alluvial-fan deposits are probably less than several tens of feet thick.

Younger undifferentiated alluvial-fan deposits (Qafy): Large alluvial-fan complexes in the Cache and Malad Valleys form coalescing aprons of sediment shed off Clarkston Mountain, the Junction Hills, and the West Hills. We mapped these deposits as older alluvial-fan deposits (Qafo), level 2 alluvial-fan deposits (Qaf_2), modern alluvial-fan deposits (Qaf_1), and younger undifferentiated alluvial-fan deposits (Qafy). Well logs and gravity data indicate that these valley-fill deposits, including Lake Bonneville and pre-Bonneville lake deposits and other late Tertiary to Quaternary deposits, are up to 5,000 feet (1,500 m) thick in the Malad Valley portion of the Portage quadrangle (Peterson, 1974; Zoback, 1983), and about 4,600 feet (1,400 m) thick in the Clarkston trough portion of Cache Valley (Evans and Oaks, 1996).

Younger undifferentiated alluvial-fan deposits slope gently away from the mountain fronts, and show an overall down-fan decrease in clast size typical of alluvial-fan deposits. The upper portions of these fans are characterized by abundant boulders. The lower and middle portions of the fans along the West Hills and the east flank of Clarkston Mountain, and locally along the mountain's west flank, are commonly cultivated. In Malad Valley, the distal portions of these fans cover Lake Bonneville deposits, indicating that the active portions of these fans are younger than about 14 thousand years old. Similar shoreline deposits along the Clarkston embayment in Cache Valley are only locally concealed. There, younger undifferentiated alluvial- fan deposits overlie the Washboards subunit and older alluvial-fan deposits.

Younger undifferentiated alluvial-fan deposits (Qafy) are distinguished from modern alluvial-fan deposits (Qaf_1) by subtle geomorphic differences. Qaf_1 deposits form active, isolated alluvial fans at the mouths of active drainages whereas Qafy deposits are generally incised in their upper portions. Qafy deposits are correlative with undifferentiated Qaf_1 and Qaf_2 deposits.

Artificial Deposits (Qf)

Artificial fill consists of general borrow material used in the construction of small stock and retaining ponds throughout the Clarkston and Portage quadrangles. We also mapped engineered fill used in the construction of Interstate 15. Although we mapped only a few areas of artificial fill, fill should be anticipated in all developed areas, many of which are shown on the topographic base maps.

Colluvial Deposits (Qc)

Colluvial deposits consist of poorly to moderately sorted, clay- to boulder-size, locally derived sediment deposited principally by slope wash and soil creep on moderate slopes. Colluvium is common on most slopes in the quadrangles, but is only mapped where deposits are thick and extensive enough to conceal large areas of bedrock. These deposits locally include talus and mixed alluvial and colluvial deposits that are too small to be mapped separately. Colluvial deposits range from 0 to about 20 feet (0-6 m) thick.

Lacustrine Deposits

Sediments and landforms of Lake Bonneville dominate the late Pleistocene geology of Cache and Malad Valleys. Lake Bonneville is the latest in a series of large lakes that occupied portions of the Bonneville basin, an area of internal drainage for much of the past 15 million years. At least two Pleistocene pre-Bonneville lakes occupied Malad Valley, and at least one occupied Cache Valley (Scott and others, 1983; Oviatt, 1986a; Oviatt and Curry, 1987; Oviatt and others, 1992). The Bonneville lacustral cycle was coincident with the last glacial maximum of marine isotope stage 2 (Currey and Oviatt, 1987; Oviatt and others, 1992) and rose and fell through its transgressive/regressive cycle from about 30 to 12 ka (Oviatt and others, 1992). Although other pre-Bonneville lakes are known in the basin, Lake Bonneville was the deepest and most extensive, and the only one known to have overflowed (Oviatt and others, 1992). It was once the largest late Pleistocene pluvial lake in western North America, and at its maximum covered an area of about 20,000 square miles (52,000 km^2) and was about 1,000 feet (300 m) deep near the present Great Salt Lake.

A time-altitude diagram of Lake Bonneville, modified from Oviatt and others (1992), is shown in figure 8, and the following summary is also extracted from Oviatt and others (1992); ages are in radiocarbon years. Lake Bonneville began to rise from near historical levels about 30 ka, following increased precipitation at the start of the last ice age. From about 22 to 20 ka, climatic fluctuations caused at least one major lake-level fall of about 150 feet (45 m), termed the Stansbury oscillation. Formation of the Stansbury shoreline complex was followed by a rapid transgression to the lake's highest level, the Bonneville shoreline, leading to initial overflow at the Zenda outlet in southern Idaho about 15 ka. Intermittent overflow may have continued for as long as 500 years while the Bonneville shoreline formed. Headward erosion and seepage at the drainage divide, a large alluvial-fan complex, caused catastrophic failure of the outlet about 14.5 ka. This lowered the lake level about 355 feet (108 m) in a matter of months, where it stabilized at the Red Rock Pass

Table 2. *Elevations (in feet) of Lake Bonneville shorelines in the Clarkston and Portage quadrangles as determined in this study.*

Shoreline	**West Hills**	**West flank Clarkston Mtn north**	**West flank Clarkston Mtn south**	**East flank Clarkston Mtn north**	**East flank Clarkston Mtn south**	**Bergeson Hill**
Bonneville	5150-5170	5160-5170	5150-5180	5140-5160	5160-5180	5140-5170
Provo	4780-4800	4780-4800		4790-4800		na
Stansbury	4500-4520	4520		na		na

na - not applicable

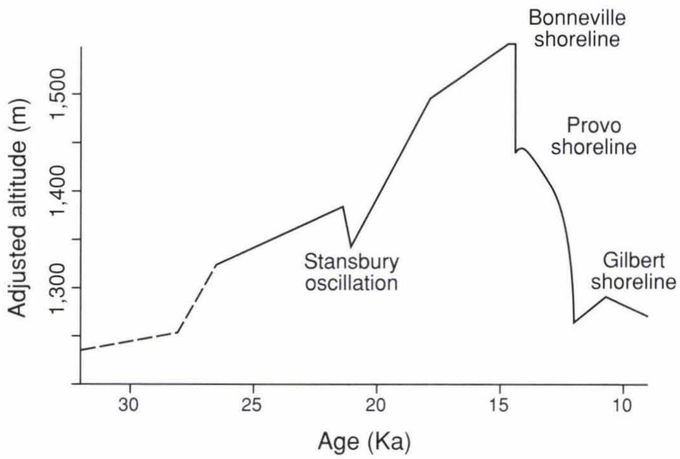


Figure 8. *Time-altitude diagram showing major shorelines of Lake Bonneville and the Gilbert shoreline of the Great Salt Lake. Modified from Oviatt and others (1992).*

bedrock threshold (Jarrett and Malde, 1987; O'Conner, 1993). The flood released about 1,128 cubic miles (4,700 km^3) of water at a peak discharge rate of about 33,000,000 ft^3/s (935,000 m^3/s) (Jarrett and Malde, 1987), lowering the lake from the Bonneville to the Provo shoreline. The lake stabilized at the Provo shoreline until about 14 ka, when the climate warmed. With rapid regression from the Provo shoreline, the lake basin again became closed, and by about 12 ka the lake had dropped to very low levels, possibly lower than average historical Great Salt Lake levels, thus marking an end to the Bonneville cycle. Between about 10.9 to 10.3 ka, the Great Salt Lake transgressed to form the Gilbert shoreline at about 4,250 feet (1,296 m) (Currey, 1990); the Gilbert phase did not reach the Clarkston or Portage quadrangles.

Three major strandlines, the Stansbury, Bonneville, and Provo shorelines of the Bonneville lake cycle are present in the Clarkston and Portage quadrangles; a number of minor strandlines are present as well. With the regression of Lake Bonneville, an enormous weight was removed from the Earth's crust, leading to isostatic rebound (Crittenden, 1963). Shoreline elevations thus vary throughout the basin, with shorelines on islands near the center of the basin showing proportionately more rebound than distal portions of the lake covered by shallow water. Solomon (1999) noted that in western Cache Valley, isostatic rebound and regional tectonics have uplifted the Bonneville shoreline as much as 89 feet (27 m), and the Provo shoreline as much as 72 feet (22 m). In the Clarkston and Portage quadrangles, the Bonneville shoreline has rebounded about 58 feet (17 m) and the Provo shoreline about 42 feet (13 m) (based on outlet elevations of 5,092 and 4,738 feet [1,552-1,444 m] and local strandline elevations of 5,150 and 4,780 feet [1,570-1,457 m]). The lake reached a maximum depth of about 840 feet (255 m) in the Portage quadrangle, but was only about 400 feet (120 m) deep in the Clarkston quadrangle.

Currey (1982) used aerial-photo and map interpretation to gauge Bonneville shoreline heights in both the Clarkston and Portage quadrangles. He identified the Bonneville shoreline at the northwest-trending spit at Little Canyon at an elevation of 5,156 ± 6.5 feet (1,572 ± 2 m), and at the north-trending spit west of Clarkston at an elevation of 5,150 ± 6.5 feet (1,570 ± 2 m). In the adjacent Limekiln Knoll quadrangle, Murphy and others (1985) identified the Bonneville shoreline at elevations of 5,160 to 5,180 feet (1,573-1,579 m), and the Provo shoreline at 4,780 feet (1,457 m). Table 2 summarizes shoreline elevations across the Clarkston and Portage quadrangles based on our mapping. The spread of values for a given shoreline is due to different morphostratigraphic components upon which elevations are measured, and documented or possible post-Bonneville faulting. Differential isostatic rebound across this relatively small area likely would have a negligible effect on shoreline elevations. In the Clarkston quadrangle north of Short Divide, the Bonneville shoreline is about 30 feet (10 m) lower than it is south of Short Divide (Solomon, 1999). This difference in shoreline elevation suggests surface rupture of the Clarkston fault independent from the Junction Hills fault to the south, with the shoreline elevation lowered on the hanging wall of the Clarkston fault (Solomon, 1999). The Short Divide fault marks a segment boundary between the Clarkston and Junction Hills segments of the West Cache fault zone (Solomon, 1999) and the Clarkston Mountain and proposed Beaver Dam segments of the Wasatch fault zone (Machette and others, 1992; Goessel, 1999; Goessel and others, 1999).

The Bonneville shoreline in the Portage quadrangle also shows elevation differences, but these are more difficult to explain. Differences may be due to complex interaction of movements along the Wasatch, West Cache, and Short Divide faults. Along the west flank of Clarkston Mountain, the Bonneville shoreline is only preserved in the footwall of the Wasatch fault zone where it is present at an elevation of about 5,160 to 5,170 feet (1,573-1,576 m). At the southwest margin of Clarkston Mountain at Little Canyon, the Bonneville shoreline is at an elevation of about 5,150 feet (1,570 m). At the south end of Clarkston Mountain, in the hanging wall of the Short Divide fault, poorly preserved Bonneville shoreline features are at an elevation of about 5,180 feet (1,579 m).

Except for prominent nearshore deposits and comparatively small areas of mostly fine-grained lacustrine deposits, most lacustrine deposits in the Clarkston and Portage quadrangles are mantled by loess and commonly cultivated. Exposures are generally poor. The deposits may be gravelly in poorly developed shoreline deposits. Typically, lacustrine deposits grade downslope from sandy nearshore deposits to fine-grained offshore deposits that are white to yellowish-gray, thick- to very thick-bedded, blocky-weathering, locally iron-stained calcareous silt.

Lacustrine sand and gravel deposits (Qlgb, Qlgp, Qlg): We mapped moderately to well-sorted, moderately to well-rounded, clast-supported, pebble- to cobble-gravel and sand, and lesser pebbly sand, along portions of the Bonneville (Qlgb) and Provo (Qlgp) shorelines. We also mapped undifferentiated lacustrine sand and gravel (Qlg) northwest of Portage and on the flank of the West Hills where it was probably deposited in spits below the Provo shoreline.

In the Clarkston quadrangle, Bonneville shoreline deposits are along the middle portion of the piedmont slope east of Clarkston Mountain, at the distal end of the alluvial-fan complex, and ringing the base of the Washboards, Bergeson Hill, and the Junction Hills. They commonly form a poorly exposed veneer on wave-cut benches. Many of these deposits are partly covered by loess and cultivated, thereby obscuring obvious contacts with finer grained offshore sediments (Qlsb). The contacts used generally coincide with a break in slope believed to mark the limit of effective wave base during the Bonneville highstand. Well-developed spits of lacustrine sand and gravel are present immediately west of Clarkston and at the south end of the Washboards. A prominent sand and gravel bar at the Bonneville shoreline connects the Washboards with Bergeson Hill; this landform enclosed a sheltered bay as the level of Lake Bonneville rose. Bonneville shoreline deposits reach a maximum local thickness of several tens of feet in a spit just west of Clarkston.

In the Portage quadrangle, evidence of the Bonneville shoreline is only locally preserved along the Clarkston Mountain front. There, it forms small gravel-capped benches carved in Nounan and St. Charles strata, or, locally, Quaternary deposits. In the West Hills, the Bonneville shoreline forms prominent, narrow benches cut into Oquirrh strata. These benches are armored with angular Oquirrh boulders and generally lack rounded clasts typical of other shoreline deposits. However, we did map Bonneville gravels along what may have been an east-trending spit on the north side of Johnson Canyon. These deposits lie in excess of 140 feet (43 m) below the Bonneville shoreline and reach downslope to the Provo shoreline. They appear to form a comparatively thin cover over an Oquirrh bedrock high. They may represent a prograding spit deposited by longshore currents as the lake rose to the Bonneville level. Along the south margin of Clarkston Mountain and the west flank of the Junction Hills, the Bonneville shoreline coincides with a prominent escarpment approximately 400 feet (120 m) in height. For reasons discussed later, we believe this escarpment to be wave-modified but primarily fault-generated. Shoreline deposits there are mostly concealed by younger alluvial and mass-movement deposits.

In the Portage quadrangle, Provo shoreline sand and gravel (Qlgp) of uncertain thickness forms prominent, wide benches at the southwest corner of Clarkston Mountain. The Provo shoreline itself is well developed north of Plymouth and in the West Hills, although generally lacks mappable sand and gravel deposits. The shoreline is only moderately developed along the west flank of Clarkston Mountain, where it forms subtle wave-cut benches on Salt Lake strata and the middle portions of piedmont slopes. In the Clarkston quadrangle, exposures of gravelly Provo shoreline sediments are restricted to the extreme southeast corner of the quadrangle.

In Cache Valley, lacustrine sand and gravel is distinguished from alluvial-fan gravels and the Washboards subunit by their greater diversity in clast composition, commonly better sorting, and geomorphic expression, in addition to proximity to the Bonneville and Provo shorelines. Bonneville shoreline gravels contain a mixed clast assemblage from both alluvial-fan and Washboards deposits. In the Portage quadrangle, such deposits are distinguished principally on the basis of morphology and proximity to the Bonneville and Provo shorelines.

Deltaic deposits (Qldp): Well-developed deltaic deposits graded to the Provo shoreline are present at the entrances to Broad and Johnson Canyons in the West Hills. These deposits consist of thin- to thick-bedded, moderately to well-sorted, pebble- to cobble-size gravel, sand, and silt that forms planar, gently east-sloping surfaces below the Provo shoreline. Topset and foreset beds are well exposed in a gravel pit at the entrance to Johnson Canyon, in the NW1/4NW1/4SW1/4 section 5, T. 13 N., R. 3 W. The exposed thickness is 0 to 80 feet (0-24 m).

Lacustrine sand and silt deposits (Qlsb, Qlsp): Coarse- to fine-grained lacustrine sand and silt, with minor clay, was deposited as nearshore beach sediments downslope of coarser grained deposits (Qlgb, Qlgp, and Qlg). Exposures are poor and obscured by cultivation, while contacts are gradational with adjacent lacustrine units. Solomon (1999) noted that where exposed elsewhere in Cache Valley, lacustrine sand and silt is typically rhythmically bedded and well sorted, with common ripple laminations. In the Clarkston and Portage quadrangles, transgressive sand and silt (Qlsb) was deposited as the level of Lake Bonneville rose, and is preserved upslope of the Provo shoreline. Transgressive deposits grade downslope into finer grained deposits (Qlmb) east of Clarkston. Regressive sand and silt (Qlsp) was deposited downslope of the Provo shoreline as the lake level fell. These deposits become increasingly fine grained along the central part of Malad Valley, although it is not practical to differentiate the predominantly finer silt and clay deposits. The exposed thickness of lacustrine sand and silt is less than 30 feet (10 m).

Lacustrine silt and clay deposits (Qlmb): Calcareous silt (marl), with minor clay and fine sand, was deposited in the quiet-water environment of the sheltered bay between headlands east of Clarkston. These sediments overlie sandy deposits (Qlsb), implying deposition in increasingly deeper or quieter water in a transgressive lake. Shorelines are not developed on this unit. Exposures are poor, but lacustrine silt and clay is commonly thick bedded to very thick bedded, with an exposed thickness of less than 15 feet (5 m).

Mass-Movement Deposits

Landslide deposits (Qmsy, Qmso): Landslides in the Clarkston and Portage quadrangles are grouped into younger and older mass movements based on degree of preservation of characteristic features, and, where possible, on cross-cutting relationships with the Bonneville shoreline. We mapped landslides that conceal shoreline features, or that have comparatively fresh morphology, as younger mass-movement deposits (Qmsy); all others are mapped as older mass-movement deposits (Qmso).

A large number of landslides, including a complex mass movement several square miles in extent, are along the Bonneville shoreline in both the Clarkston and Portage quadrangles. Most of these deposits formed in Salt Lake strata and many conceal shoreline features, evidence that they are contemporaneous with the Bonneville shoreline or are post-Bonneville in age; they probably formed as a result either of wave erosion of the Bonneville shoreline, rapid dewatering of slopes following the Bonneville flood, or earthquakes (figure 6). Others clearly are modified by shoreline erosion. Despite the differentiation into younger and older deposits, most mass movements have a subdued morphology and degraded main scarps, suggesting that most are latest Pleistocene in age; some mass movements, however, were likely active in the Holocene, and some may have evidence of historical movement.

The largest mass movement in the area is on the west flank of the Junction Hills (figure 9). This deposit formed principally in the Salt Lake Formation and coherent, back-tilted blocks of Salt Lake strata are visible. The main scarp follows the crest of the Junction Hills to a point just south of Short Divide. Numerous subsidiary scarps are present within the slide mass itself. Bonneville shoreline features are not clearly expressed on these deposits in the Clarkston quadrangle, although this strandline elevation does approximately correspond to a prominent break in slope within the deposits. Oviatt (1986a) mapped the Bonneville shoreline across the length of this deposit in the Cutler Dam quadrangle, based apparently on this break in slope and two small bench-like features. The northern bench-like feature, near the center of section 16, T. 13 N., R. 2 W., appears tilted and at a slightly lower elevation than the Bonneville shoreline, suggesting that it may be a Bonneville shoreline remnant disturbed by subsequent landsliding. Because the Bonneville shoreline is not apparent on this large landslide deposit and the Provo shoreline clearly cuts across its middle portion in the Clarkston quadrangle, evidence in the Clarkston quadrangle brackets emplacement of the slide between about 14.5 and 14.3 ka (Oviatt and others, 1992). The slide was probably initiated by Bonneville shoreline erosion that undercut slopes, rapid dewatering of slopes following the Bonneville flood, or earthquakes. Variable scarp preservation and internal morphology of the deposit suggest that it has been intermittently active since that time. The landslide deposit is modified by lacustrine processes below the Provo shoreline where it is mapped as Ql/Qmsy. The deposit extends southward into the Cutler Dam quadrangle where it covers an additional several square miles.

We mapped a large mass-movement deposit where it conceals Bonneville shoreline features along the east side of the Washboards. This deposit lacks hummocky topography characteristic of other mass movements in the quadrangles, indicating that it may have formed as a flow failure. A series of steep, poorly vegetated, arcuate slopes that bound the up-slope margin of this deposit are believed to represent an eroded headscarp.

Other landslides in the quadrangles include a large, complex landslide at the head of North Canyon in the Portage quadrangle. The slide is primarily in upper Nounan strata that form a near dip slope on the east side of Clarkston Mountain. The slide is characterized by chaotic bedding attitudes and subdued hummocky topography. It is deeply incised and in part covered by mixed alluvial and colluvial sediments. At the mountain front between Old Quigley and Cold Water Canyons, we mapped a large mass-movement deposit in remobilized Wasatch(?) Formation and Salt Lake strata. The deposit protrudes from the mountain front and is

Figure 9. *View north of the Junction Hills and large mass movement, with Clarkston Mountain in the distance and the head of Bensons Hollow in the foreground. The gently west-dipping Beaver Dam fault, partly in shadow, places Salt Lake strata down on the west (which subsequently was involved in a large mass movement) against the Oquirrh Formation and basal Salt Lake strata to the east.*

apparently unfaulted. Midway up Water Canyon just north of Gunsight Spring, we mapped a small mass-movement deposit. It is composed of angular blocks of Swan Peak, Laketown, and St. Charles strata. A broad expanse of mixed alluvium and colluvium was deposited behind this dam-like deposit.

Debris-flow deposits (Qmf): We mapped upper Holocene debris-flow deposits at the mouths of several active drainages along the eastern margin of Clarkston Mountain, from Cold Water Canyon in the north to Old Canyon in the south. The deposits, commonly on surfaces of modern alluvial-fan deposits (Qaf1), consist of cobble and boulder gravel in a matrix of silt, sand, clay, and minor pebbles. They are unsorted and unstratified except for sparse interbedded fluvial sand and gravel layers. The surfaces of the deposits are commonly covered with coarse, angular rubble and fresh-appearing levees and channels graded to modern stream level. Debris-flow deposits have an exposed thickness of less than 30 feet (10 m).

Mixed-Environment Deposits

Alluvial and colluvial deposits (Qac, Qaco): We mapped mixed alluvial and colluvial deposits along secondary drainages and larger swales. These deposits consist of poorly to moderately sorted, clay- to boulder-size, locally derived sediment modified by both alluvial and colluvial processes. Older mixed alluvial and colluvial deposits are in the Steel Canyon drainage and are differentiated by their elevation above modern drainages. Mixed alluvial and colluvial deposits (Qac) are generally less than 20 feet (6 m) thick, while older deposits (Qaco) in Steel Canyon are probably less than 40 feet (12 m) thick.

Lacustrine and alluvial deposits (Qla): Mixed lacustrine and alluvial deposits consist of moderately to well-sorted, fine-grained sand, silt, and clay deposited principally below the Provo shoreline. These deposits typically have a well-developed soil profile, and in the Clarkston quadrangle are mapped principally along Clarkston Creek, where the extent of thin, fine-grained alluvium cannot readily be differentiated from fine-grained lacustrine deposits. In the Portage quadrangle, we mapped mixed lacustrine and alluvial deposits along the central portion of Malad Valley, west of the Malad River, where it is impractical to differentiate distal alluvial-fan, flood-plain, and fine-grained lacustrine sediments. Smaller deposits are along the flanks of the West Hills. These deposits are probably 0 to several tens of feet thick.

Lacustrine and alluvial gravelly deposits (Qlag): We mapped mixed lacustrine and alluvial gravelly deposits along the alluvial apron surrounding Clarkston Mountain, at and below the Bonneville shoreline, where it is not practical to differentiate gravelly lacustrine and alluvial-fan deposits. These deposits consist principally of moderately sorted, moderately to well rounded, pebble- to cobble-size gravel and sand. A coarser alluvial-fan component predominates in deposits near the mountain front east and south of Oregon Springs in the Portage quadrangle.

In Cache Valley, mixed lacustrine and alluvial gravelly deposits are distinguished from alluvial-fan gravels and the Washboards subunit by their greater diversity in clast composition and geomorphic expression, in addition to proximity to the Bonneville shoreline. Bonneville shoreline gravels contain a mixed-clast assemblage from both alluvial-fan and Washboard deposits. In the Portage quadrangle, such deposits are distinguished principally by morphology and proximity to the Bonneville and Provo shorelines.

Lacustrine and alluvial fine-grained deposits (Qlam): Mixed lacustrine and alluvial fine-grained deposits consist of silt and clay deposited along the Malad River in the northern part of the Portage quadrangle. These deposits are commonly saline and alkaline and, where undisturbed by cultivation, are distinguished by marsh vegetation and characteristic bright, blotchy appearance on aerial photographs (Chadwick and others, 1975). The contact with undifferentiated lacustrine sediments east of the river is gradational and poorly expressed. The contact with mixed lacustrine and alluvial sediments west of the river is gradational but marked by a subtle change in slope. These deposits are probably several tens of feet thick.

Spring Deposit (Qst)

A single exposure of light-gray to brownish-gray, porous, calcareous tufa is present near the head of Elbow Canyon. The deposit is on Garden City strata along a northwest-trending, down-to-the-west normal fault.

Stacked-Unit Deposits

Undifferentiated lacustrine deposits over the Salt Lake Formation (Ql/Tsl): We mapped a discontinuous veneer of mostly fine-grained lacustrine deposits over the Salt Lake Formation along the south flank of Clarkston Mountain. These lacustrine deposits are below the Bonneville shoreline and are typically 0 to several feet thick.

Undifferentiated lacustrine deposits over landslide deposits (Ql/Qmsy, Ql/Qmso): Along the west flank of the Junction Hills, we mapped a discontinuous veneer of mostly fine-grained lacustrine deposits over a large landslide deposit. These lacustrine deposits are below the Bonneville shoreline and are typically 0 to several feet thick. The Provo shoreline clearly cuts across the mid-portion of the landslide deposit. These deposits have a subdued rolling topography and main scarps are typically not identifiable. Similar deposits over slightly older mass movements along the south flank of Clarkston Mountain are below the Bonneville shoreline.

STRUCTURE

Regional Setting

The Clarkston and Portage quadrangles lie astride the boundary between the Basin and Range and Middle Rocky Mountains physiographic provinces (Fenneman, 1931; Stokes, 1986). The Basin and Range Province is characterized by roughly east-west extensional tectonics that created mostly north-trending, fault-bounded, isolated mountain ranges separated by deep sediment-filled basins. In northern Utah, the Middle Rocky Mountains Province consists of a series of eastward-directed and generally eastward-younging thrust plates emplaced during the Late Cretaceous to early Eocene Sevier orogeny; east of the northern Wasatch Range, these thrust plates are widely concealed by synorogenic deposits. Both provinces have experienced broad epeirogenic uplift. The Wasatch fault zone, commonly interpreted to mark this provincial boundary in this area (for example, see Stokes, 1986), separates the West Hills and Malad Valley on the west from Clarkston Mountain, Junction Hills, and Cache Valley on the east. Cache Valley and several smaller back valleys to the south point to the transitional nature of the Basin and Range-Middle Rocky Mountains provincial boundary.

The lower Paleozoic rocks of Clarkston Mountain were transported eastward in the upper plate of the Paris-Willard thrust in the middle Cretaceous (Neocomian to Turonian) during the Sevier orogeny (Allmendinger and others, 1984; Allmendinger, 1992; Coogan, 1992; Yonkee, 1997). Lower Paleozoic Clarkston Mountain strata were probably folded as post-Paris-Willard (Crawford and Absaroka) thrust faults ramped upward near the present-day Wasatch Front and carried rocks farther eastward (Rodgers and Janecke, 1992; Yonkee, 1992). The part of the West Hills in the southwest corner of the Portage quadrangle is carved from Oquirrh Formation strata in the upper plate of the Samaria Mountain thrust, the frontal ramp of the Hansel allochthon. The Hansel allochthon was emplaced during or prior to the Late Cretaceous and shows an estimated minimum eastward displacement of 9 miles (15 km) (Allmendinger and others, 1984).

Janecke and Evans (1999) reported that a major detachment terrane in the Bannock Range of southeastern Idaho is bounded by low-angle detachment faults in which the hanging wall moved to the west-southwest during the late Cenozoic (Miocene to Pliocene), not Mesozoic or early Tertiary as earlier inferred (Link, 1982). Janecke and Evans suggested that this detachment terrane may extend southward to Huntsville, Utah, thus placing Clarkston Mountain in the hanging wall of the detachment fault. If this interpretation is correct, many of the internal faults that cut Clarkston Mountain may be the result of extension in the hanging-wall block of the Bannock detachment system.

Wasatch Fault Zone

The Wasatch fault zone is a major, active, steeply west-dipping normal fault that stretches 240 miles (386 km) from central Utah to southern Idaho. The Wasatch fault zone consists of 10 discrete segments that tend to rupture independently during major earthquakes; parts of two of these segments are in the Clarkston and Portage quadrangles (Machette and others, 1992). The Clarkston Mountain segment trends about 15 miles (24 km) from near Woodruff, Idaho, southward to the southern end of Clarkston Mountain. The Collinston segment trends about 17 miles (27 km) southward from the Short Divide area to Honeyville on the northwest flank of the Wellsville Mountains. The boundary between these two segments of the Wasatch fault zone is generally considered to be at the Short Divide fault (Machette and others, 1992).

The Clarkston Mountain segment of the Wasatch fault zone bounds the west flank of Clarkston Mountain and in the Portage quadrangle consists of two straight-line parts. North of Elgrove Canyon, the fault trends about N. 25° W., whereas to the south it trends about N. 30° E. Older alluvial-fan deposits (Qafo), in fault contact with Paleozoic bedrock, are locally preserved in the embayment in the mountain front thus created. Fault breccia (QTbx) is present on faceted spurs along the length of the Clarkston Mountain segment, and is especially well exposed along the Elgrove Canyon embayment (figure 6). Three-point solutions and well-developed fault breccia on faceted spurs show that the Clarkston Mountain segment of the Wasatch fault zone dips about 45 degrees west. Like Klauk and Budding (1984), but unlike Machette and others (1992), we infer a concealed splay of the Wasatch fault zone west of the main range-front fault to account for a bedrock bench of Salt Lake Formation exposed in this embayment.

Cluff and others (1974) used 1:12,000-scale low-sun-angle aerial photography to map the trace of the Wasatch fault zone in the Clarkston and Portage quadrangles, although they locally confused other lineaments such as shorelines with fault scarps. In the Portage quadrangle, they mapped numerous lineaments interpreted as faults near the common corner of sections 10, 11, 14, and 15, T. 14 N., R. 3 W. All of these lineaments are below the Bonneville shoreline, and it is unclear which if any are wave-modified fault scarps. The southern end of the Clarkston Mountain fault is marked by a prominent, wave-modified escarpment at the southwest end of Clarkston Mountain. Machette and others (1992) noted that the Provo shoreline wraps around this escarpment, and suggested that the fault scarp slightly predates the Bonneville lake cycle. They thus considered the latest movement on the Clarkston Mountain fault as late Pleistocene in age.

We mapped a short but prominent scarp in older alluvial-fan deposits (Qafo) at the entrance to Elgrove Canyon. This scarp is 15 feet (5 m) high and lies above the Bonneville shoreline. Similar scarps are not present elsewhere along the Clarkston fault in the Portage quadrangle. However, the mouths of most other drainages are below the Bonneville shoreline and are characterized by the presence of younger alluvial-fan deposits. This suggests that the most recent scarp-forming event on the Clarkston Mountain segment in the Portage quadrangle predates the formation of the Bonneville shoreline.

The Collinston segment of the Wasatch fault zone is not apparent in the Clarkston quadrangle, and is poorly developed to the south in the adjacent Cutler Dam quadrangle (Oviatt, 1986a; Goessel, 1999; Goessel and others, 1999). Oviatt (1986a) and Personius (1990) found no evidence for late Quaternary offset along the fault, and only mapped the fault for about 4 miles (6.5 km) north of Collinston. Goessel (1999; see also Goessel and others, 1999) remapped the Junction Hills and northern end of the Wellsville Mountains, including the Beaver Dam fault (figure 9), a gently west-southwest-dipping normal fault along the west side of the Junction Hills that may represent a newly recognized section of the Wasatch fault zone. Goessel (1999) and Goessel and others (1999) suggested that the Collinston segment of Machette and others (1992) dies out at the north end of the Wellsville Mountains, and that throw is transferred east to the reactivated, gently west-dipping Beaver Dam fault. Based on interpretation of additional gravity data, Oaks (2000) suggested that the Collinston segment continues north to the Fielding area, and that a newly proposed Fielding section continues to the southwest end of Clarkston Mountain. Pending further paleoseismic investigations, we use the fault segment terminology of Machette and others (1992) in this report.

The Beaver Dam fault may merge with an inferred fault at a prominent escarpment at the Bonneville shoreline (figure 10). Cluff and others (1974) interpreted this escarpment as the Wasatch fault zone, although Machette and others (1992) believed this escarpment formed by erosion at the Bonneville shoreline. The escarpment is up to 400 feet (120 m) high, an order of magnitude higher than other escarpments associated with the Bonneville shoreline in this region. Furthermore, the scarp lies in an embayment protected from prevailing northwest winds, so that wave action there would have been less than on northwest-facing shores. The projection of all pediment surfaces into the air and the size of the scarp strongly support a fault origin for this escarpment. We reinterpret this escarpment as a wave-modified fault scarp. Whether this queried fault is a northwestern continuation of the Beaver Dam fault or is simply a cross fault between the Clarkston Mountain segment and Beaver Dam fault is uncertain. If the escarpment is interpreted as a fault, the Clarkston Mountain and Collinston segments may not be separated by a 4-mile-long (7 km) left step and gap in late Pleistocene faulting as proposed by Machette and others (1992).

The thickness of Cenozoic rocks and deposits in Malad Valley may be up to about 5,000 feet (1,500 m) based on gravity data and two-dimensional gravity modeling (Peterson, 1974; Zoback, 1983). Malad Valley appears to be an asymmetric graben or half graben bounded on the east by the Wasatch fault zone. Within the valley, local gravity highs just north of Plymouth, Utah (figure 11), and north of Woodruff, Idaho are thought to be the result of transverse ridges of pre-Cenozoic rocks (Peterson, 1974; Zoback, 1983; Oaks, 2000). Because bedrock in the hanging wall is concealed by Cenozoic deposits, the total stratigraphic separation on the Clarkston Mountain segment of the Wasatch fault zone is unknown. If Clarkston Mountain is part of the east limb of a large anticline, the faulted axis of which lies below Malad Valley, then in order to reconstruct the fold, the apparent stratigraphic separation along the Clarkston Mountain fault must be about 5,500 feet (1,675 m).

Short Divide Fault

The Short Divide fault trends roughly east-west and marks the southern boundary of Clarkston Mountain (figures 10, 12). The Short Divide fault places the Salt Lake Formation, and in the Junction Hills the Oquirrh Formation, down against Cambrian through Silurian strata of Clarkston Mountain. Hanson (1949) noted that these relationships suggest a normal stratigraphic separation in excess of 10,000 feet (3,000 m), but admitted that such a number seemed excessive for a west-trending normal fault. Williams (1948) also struggled with this problem and suggested, among other interpretations, that Clarkston Mountain might be part of the upper plate of a thrust sheet with the Junction Hills to the south in the footwall.

The down-to-the-south Short Divide fault is roughly on strike with the steeply dipping, down-to-the-south North Canyon fault mapped by Murphy and others (1985) in the West Hills. Following Allmendinger and others (1984), Murphy and others (1985) suggested that rocks southwest of the North Canyon fault lie in the upper plate of a block (Hansel allochthon) that moved east. Platt (1977) interpreted the northward continuation of the North Canyon fault in Idaho as the steepened edge of a thrust sheet. The colinearity of the Short Divide and North Canyon faults, their common sense of most recent normal displacement, and the juxtaposition of similar stratigraphic packages across both sides of the faults suggest that the two structures are genetically related. Pre-Cenozoic rocks of Short Divide may be part of the upper plate of the Samaria Mountain thrust (Hansel allochthon). If this interpretation is correct, normal-fault offset on the Short Divide fault is superimposed on a Sevier-age structure. The age of most recent movement on the Short Divide fault can only be constrained as post-Salt Lake Formation.

West Cache Fault Zone

The West Cache fault zone is a down-to-the-east normal fault that extends about 58 miles (93 km) along the west side

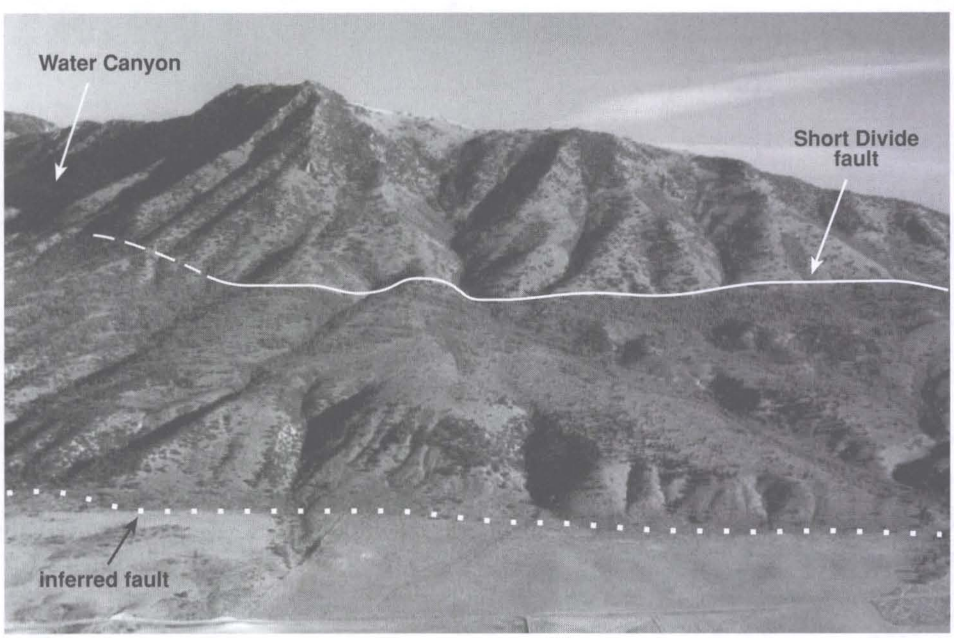


Figure 10. *View north-northeast of the Short Divide area. Water Canyon is at the left side of the photo. Pediments south of the Short Divide fault are covered by loess and truncate complexly faulted and folded Salt Lake strata. An inferred down-to-the-south fault (dotted line), parallel and coincident to the Bonneville shoreline, in turn truncates the pediment surfaces.*

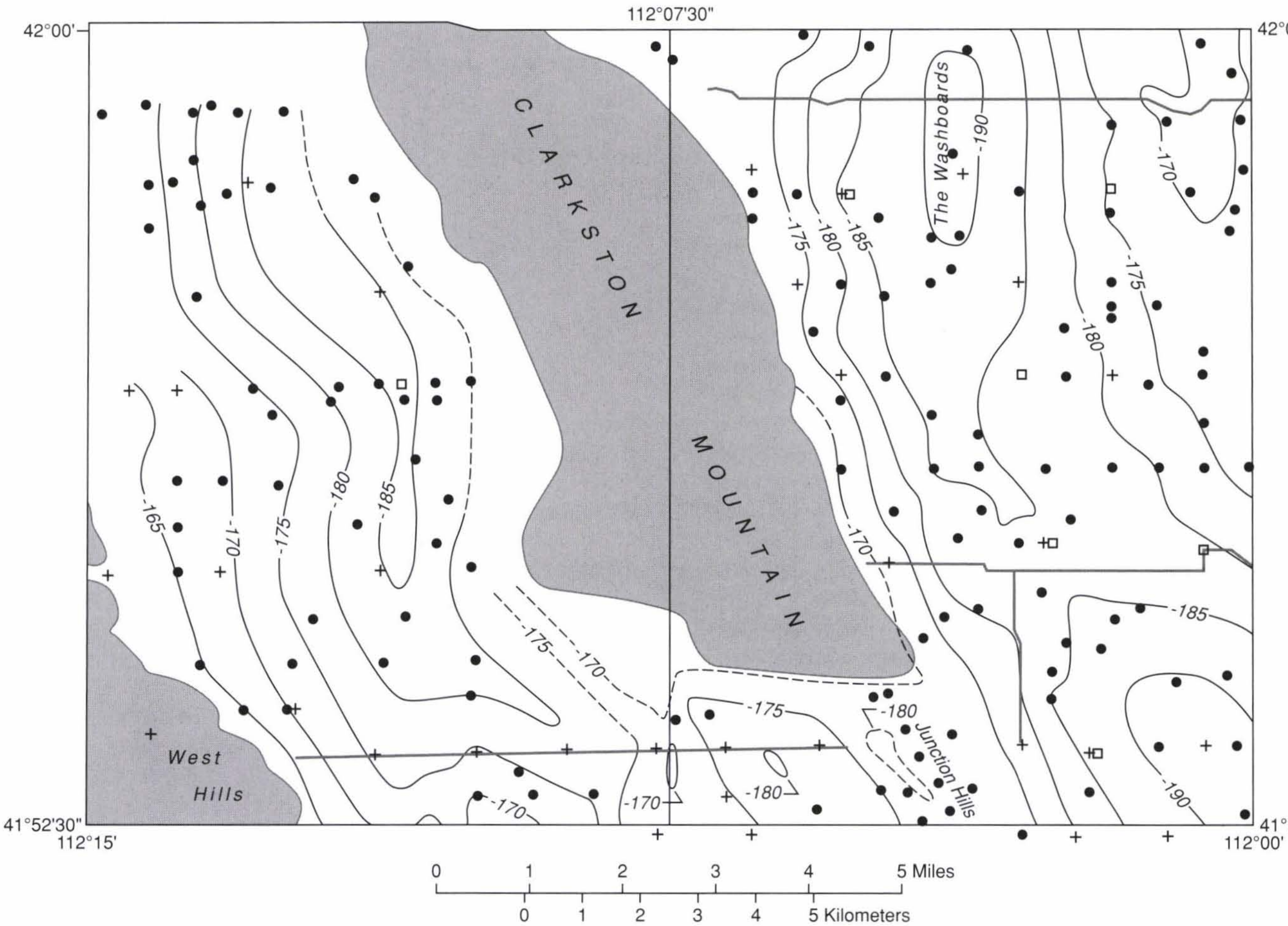


Figure 11. Bouguer gravity map of the Clarkston and Portage quadrangles, based on U.S. Geological Survey database (plus signs) and 317 new stations (heavy lines) and additional stations (dots) by R.Q. Oaks, Jr. Contour interval 5 mgals. Approximate limit of Paleozoic outcrop shown for Clarkston Mountain and the West Hills.

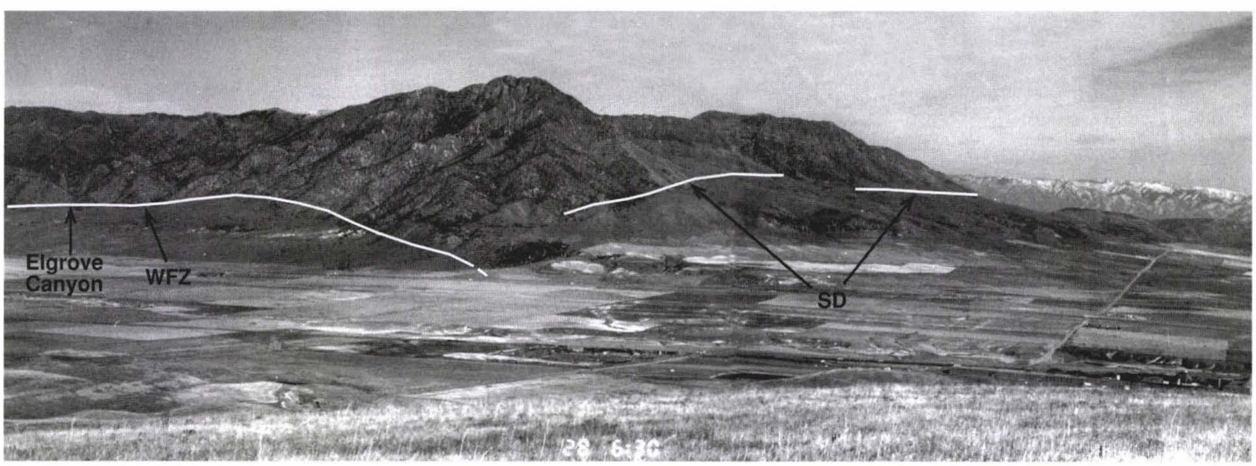


Figure 12. View east-northeast across Malad Valley to the southern end of Clarkston Mountain. The Short Divide fault (SD) is at the southern end of Clarkston Mountain and is a segment boundary of both the Wasatch fault zone (WFZ) (Machette and others, 1992) and the West Cache fault zone (Solomon, 1999; Black and others, 1999, 2000). Note light-colored strata of the Salt Lake Formation exposed in the Elgrove Canyon embayment.

of Cache Valley in north-central Utah and southeastern Idaho. Solomon (1999) divided the West Cache fault zone in Utah, from north to south, into the Clarkston, Junction Hills, and Wellsville faults, and suggested that the Clarkston fault is a seismically independent structural segment. Paleoseismic investigations of Black and others (1999, 2000) confirmed that these faults rupture separately and can therefore be considered separate segments of the West Cache fault zone. Only the Clarkston fault and the northern part of the Junction Hills fault are in the Clarkston quadrangle. Seismic-reflection data (Smith and Bruhn, 1984; Evans, 1991; Evans and Oaks, 1996) indicate that, at least south of Logan, the West Cache fault zone has significantly less displacement than the East Cache fault zone, evidence that the West Cache fault zone is probably antithetic to the East Cache fault zone (Sullivan and others, 1988).

The Clarkston fault extends at least 7 miles (11 km) along the eastern base of Clarkston Mountain from Short Divide northward to and possibly beyond the Utah-Idaho state line. South of Cold Water Canyon, the fault consists of two en echelon sections that overstep to create an embayment at Mikes Canyon, where older alluvial-fan deposits (QTaf) are preserved. These two en echelon sections trend N. 20° W. and form a series of discontinuous down-to-the-east fault scarps. The fault is generally covered by Holocene deposits at the mouths of canyons, but older alluvial-fan deposits (Qafo) are preserved between these drainages. Solomon (1999) noted two areas of potential Holocene displacement at the mouths of Winter and Raglanite Canyons. Black and others (1999, 2000) trenched a fault scarp 13 feet (4 m) high on the north side of Winter Canyon and exposed a single main fault trace and evidence for one surface-faulting earthquake having an estimated displacement of 11.5 feet (3.5 m). Radiocarbon dates on organic material exposed in the trench indicate the most recent surface-faulting event on the Clarkston fault occurred 3,600 to 4,000 years ago. Total stratigraphic separation on the Clarkston fault is unknown.

North of Cold Water Canyon, in an area covered by apparently unfaulted landslide deposits (Qmsy), remobilized Wasatch(?) strata (QTb), and Wasatch(?) strata (Tw?), the Clarkston fault branches into north- and northwest-trending splays. The north-trending splay places older alluvial-fan deposits and Salt Lake strata down on the east against the Salt Lake Formation; local well-developed fault breccias suggest that the fault dips 60 to 65 degrees to the east. The northwest-trending splay places Salt Lake strata down on the northeast against Cambrian carbonates and continues through the northeast corner of the Portage quadrangle.

Solomon (1999) noted that the elevation of the Bonneville shoreline north of the Short Divide fault is about 5,150 feet (1,570 m), distinctly lower than its elevation of about 5,180 feet (1,579 m) to the south of the Short Divide fault. The Short Divide fault marks the boundary between the Clarkston Mountain segment and the proposed Beaver Dam section of the Wasatch fault zone (Machette and others, 1992; Goessel, 1999; Goessel and others, 1999), and Solomon (1999) and Black and others (1999, 2000) showed that the Short Divide fault is probably a segment boundary of the West Cache fault zone as well. Whether this 30 feet (9 m) of post-Bonneville displacement is due to movement on the West Cache or Wasatch fault zones is unknown. If due solely to movement on the Clarkston fault, the maximum slip rate on the Clarkston fault since the late Pleistocene is 0.021 in/yr (0.54 mm/yr) (Solomon, 1999). Black and others (1999, 2000) discussed evidence relating to this difference in elevation that suggests two or three surface-faulting earthquakes occurred on the Clarkston fault since formation of the Bonneville shoreline. Gravity studies of co-author Oaks (figure 11) (see also Oaks, 2000) show a distinct transverse zone at the south end of the Clarkston segment, further evidence that the Short Divide area is a segment boundary of the West Cache fault zone.

The Junction Hills fault extends 16 miles (25 km) from Short Divide southward along the eastern flank of the Junction Hills and Cache Butte Divide. Oviatt (1986a) and Black and others (1999, 2000) described the Junction Hills fault where it is exposed in a stream cut about 4 miles (6 km) south of the Clarkston quadrangle. The fault shows about 9.5 feet (2.9 m) of displacement of transgressive Bonneville gravels. Radiocarbon analyses by Black and others (1999, 2000) indicate that the most recent surface-faulting event at this site occurred around 8,250 to 8,650 years ago. They further suggested that a minimum of 13,850 years elapsed between this most recent surface-faulting event and the penultimate event. The maximum post-Bonneville slip rate for the Junction Hills fault was thus determined to be 0.005 in/yr (0.13 mm/yr). Evans (1991) estimated a net dip slip of 2,000 to 4,000 feet (600-1,200 m) on the West Cache fault zone at the Junction Hills since Miocene extension began.

East of Short Divide, in the NW$^1/_4$ section 3 and NE$^1/_4$ section 4, T. 13 N., R. 2 W., a splay of the Junction Hills fault displaces the Salt Lake Formation. Displacement along this fault appears to be small, however, such that the principal displacement on the Junction Hills fault may be taken up on a concealed fault to the east, just below and parallel to the Bonneville shoreline. Lineaments in the E$^1/_2$ of sections 3 and 10, T. 13 N., R. 2 W., however, are believed to be fault scarps. Although these scarps are along the same trend as the fault studied by Oviatt (1986a) and Black and others (1999, 2000), it is unclear whether they cut Bonneville lacustrine deposits or if thin lacustrine deposits are draped over a pre-Bonneville structure in the Salt Lake Formation. While the lineaments are mapped as concealed by lake beds, they are clearly visible on aerial photos and in the field as discontinuous breaks in slope up to a few tens of feet high. The Junction Hills fault appears to branch in the SE$^1/_4$ section 3, T. 13 N., R. 2 W. into a series of north- and northwest-trending splays. One or more of these splays may continue northward along the east base of Round Knoll, as suggested by a steep, wave-modified escarpment in Salt Lake strata. The westernmost splay of the Junction Hills fault appears to trend northwest across Al Archibald Hollow, where it probably links up with the Clarkston fault.

Cache Valley changes symmetry from a dominantly east-tilted half graben in the south, to a true graben in the middle, to a dominantly west-tilted half graben in the north. This change in basin geometry occurs over a broad area centered on the Utah-Idaho border (Evans and Oaks, 1996; Janecke and Evans, 1999). Gravity studies of co-author Oaks (figure 11) (see also Oaks, 2000) show a strong gravity low beneath the Washboards that disappears southward at the south end of the Washboards. This gravity low may indicate increased displacement northward along the Clarkston segment of the West Cache fault zone, which is consistent with the high rate of deposition inferred for Salt Lake strata in the Bergeson Hills and Washboards. If this interpretation is correct, the Clarkston fault must have been active during deposition of strata younger than 4.4 to 5.1 million years old, if not before.

The thickness of Cenozoic strata in Cache Valley in the Clarkston quadrangle (Clarkston trough) is poorly constrained, but is probably less than that for the main portion of the valley. Peterson and Oriel (1970) interpreted about 8,000 feet (2,500 m) of Cenozoic rocks in the Clarkston trough based on gravity data and two-dimensional gravity modeling. Zoback (1983) interpreted 6,990 feet (2,130 m) of Cenozoic rocks in the main part of the Cache Valley and about 3,280 feet (1,000 m) in the Clarkston trough. Evans and Oaks (1996) estimated about 4,590 feet (1,400 m) of Cenozoic sediments in the Clarkston trough. East of the Clarkston trough, the fault-bounded bedrock ridge of Little Mountain (Newton Hill), Pete McCombs Hill, and Bergeson Hill apparently joins with the Bannock Range to the north in Idaho (Janecke and Evans, 1999). This ridge corresponds to a gravity high (Peterson and Oriel, 1970; Evans and Oaks, 1996; Oaks, 2000) (figure 11) and is bounded on the east by the Oxford-Dayton fault. The Clarkston trough is bounded on the south by a concealed, transverse ridge of pre-Cenozoic rocks about where the town of Newton is located (Peterson and Oriel, 1970; Zoback, 1983); Oaks (2000) showed the southern boundary farther north, at his Plymouth transverse zone.

Clarkston Mountain

Clarkston Mountain consists of highly faulted, moderately east-dipping Middle Cambrian to Silurian strata (Biek, 1999). The range is bounded on the west by the Clarkston Mountain segment of the Wasatch fault zone, on the north and east by the Clarkston segment of the West Cache fault zone, and on the south by the Short Divide fault. Strata of Clarkston Mountain are moderately well exposed on the west flank of the mountain, but generally poorly exposed on the near-dip slope of the mountain's east flank. As originally suggested by Hanson (1949), Clarkston Mountain may be the east limb of a large anticline, the axis of which is faulted and now buried under Malad Valley. Pre-Cenozoic Clarkston Mountain strata were probably folded as post-Paris-Willard (Crawford and Absaroka) thrust faults ramped upward near the present-day Wasatch front and carried the overlying rocks farther eastward (Rodgers and Janecke, 1992; Yonkee, 1992). Rodgers and Janecke (1992) created a pre-Miocene paleogeographic map of southeast Idaho and used the apparent structural relief to infer fold and fault geometries associated with the western part of the Idaho-Wyoming-Montana thrust belt. They identified the Malad ramp beneath Malad Valley, a major footwall ramp of the Paris-Putnam thrust system. Clarkston Mountain is cut by numerous, mostly down-to-the-west, moderately to steeply dipping normal faults, most of which probably formed during Basin and Range extension; by a minor bedding-plane fault of possible early Tertiary age; and by several gently dipping normal faults of likely early Basin and Range age.

Normal Faults

In plan view, a geologic map of Clarkston Mountain looks like a shattered pane of glass. Most faults in Clarkston Mountain trend north-northwest, subparallel to the axis of the range, and most are moderately to steeply dipping, down-to-the-west normal faults. The largest fault in the range is the Gunsight fault mapped by Hanson (1949) as a major, northwest-trending, down-to-the-west normal fault east of Gunsight Peak (figure 13). The trace of the fault is concealed by mixed alluvial and colluvial deposits (Qac) in Water Canyon and in the upper reaches of Elgrove Canyon, but map patterns suggest that there it dips about 55 degrees west. Planar breccias on the southward extension of the Gunsight fault – where it appears to offset the Short Divide fault – suggest that there it dips only 30 to 40 degrees west. Near the entrance to Water Canyon, the Gunsight fault places undivided Fish Haven-Laketown strata against the Nounan Formation and shows about 5,000 feet (1,525 m) of stratigraphic separation. Northward, the topography of Gardner Canyon is such that the Gunsight fault trends sharply westward to the lower reaches of Gardner Canyon where it places St. Charles strata against Bloomington strata. The Gunsight fault continues northward along the lower slopes of Clarkston Mountain where it splits into a number of down-to-the-west normal faults. Displacement decreases to the north so that at cross-section B-B′, total stratigraphic separation on these down-to-the-west faults is only about 3,800 feet (1,160 m).

The Gunsight fault divides Clarkston Mountain into two structural blocks. Strata of the western structural block are cut by a few faults having mostly small normal displacement. The largest of these, which places St. Charles Formation against the Nounan Formation, is at the southwest corner of Clarkston Mountain near a fault-segment boundary of the Wasatch fault zone. The eastern structural block is cut by numerous faults, many of which are related to the development of four small grabens. The largest graben trends northwest at the northern end of the range, and is the only place

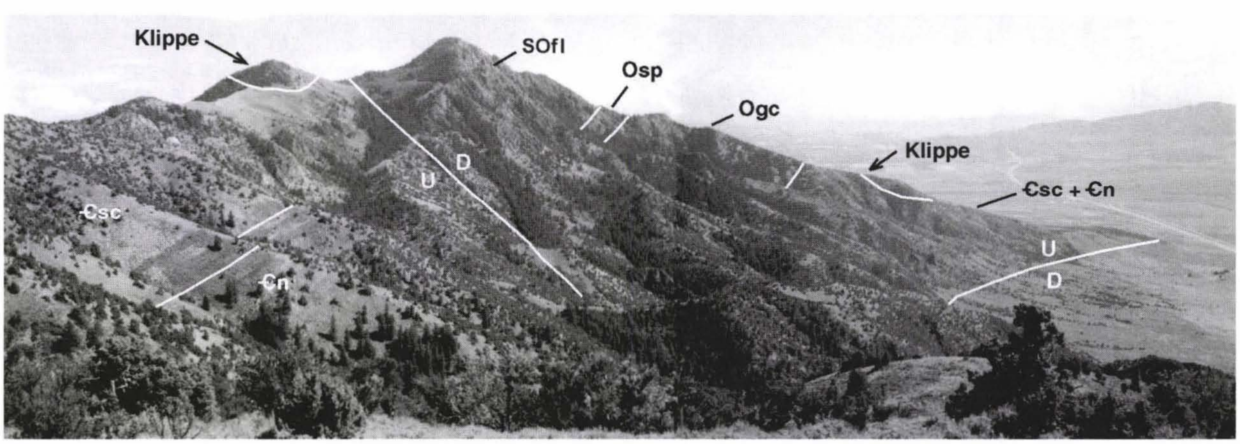


Figure 13. *View south to Gunsight Peak from the NE1/4 section 13, T. 14 N., R. 3 W. The Gunsight fault cuts through the saddle to the left (east) of Gunsight Peak, and places east-dipping Ordovician-Silurian Fish Haven and Laketown Dolomites (SOfl) down on the west against the east-dipping Upper Cambrian St. Charles Formation (€sc). The skyline to the right (west) of Gunsight Peak shows a mostly unfaulted section of Nounan (€n), St. Charles (€sc), Garden City (Ogc), and Swan Peak (Osp) strata below the Fish Haven-Laketown Dolomites (SOfl). The Clarkston Mountain segment of the Wasatch fault zone is at the base of the range. The hill east (left) of Gunsight Peak is capped by a klippe of Swan Peak and Fish Haven-Laketown strata bounded by a subhorizontal normal fault. This klippe overlies east-dipping St. Charles and Garden City strata.*

within Clarkston Mountain where the Salt Lake Formation is preserved. Based on offset of the Worm Creek Quartzite Member, stratigraphic separation on the eastern fault of this graben is about 500 feet (150 m). Displacement on the graben's western fault is less certain, but appears to be about one-half that amount. At the east-central and southeast margins of the range, small grabens are present on a near-dip slope of St. Charles and Garden City strata. These grabens are delineated by offset of the St. Charles-Garden City contact, which is a good marker on Clarkston Mountain.

Burton (1973), Gray (1975), and Green (1986) each interpreted many west-dipping normal faults at Clarkston Mountain as reactivated thrust faults, although they did not offer compelling evidence to support their interpretations. We believe their conclusions were based on misidentified stratigraphy and incorrect map relations. Furthermore, their "thrust faults" are at a high angle to bedding, unlike most thrust faults. Sprinkel (1979) summarized two such examples of "reactivated thrust faults" based on what we believe to be incorrect mapping of Burton (1973) and Gray (1975). Steeper dips of bedding near some of these faults may simply be a result of reverse drag on listric normal faults.

Janecke and Evans (1999) proposed that a major detachment terrane is present in the Bannock Range of southeastern Idaho, bounded by low-angle detachment faults in which the hanging wall moved to the west-southwest during the late Cenozoic (Miocene to Pliocene), not Mesozoic or early Tertiary as earlier inferred (Link, 1982). Janecke and Evans suggested that this detachment terrane may extend southward to Huntsville, Utah, thus placing Clarkston Mountain in the hanging wall of the detachment fault. If this interpretation is correct, many of the internal faults that cut Clarkston Mountain may be the result of extension in the hanging-wall block of the Bannock detachment system.

Bedding-Plane Fault

A nearly complete section of Bloomington strata is present on the northwest flank of Clarkston Mountain. However, the upper member of the Bloomington Formation, the Calls Fort Shale Member, thickens and thins irregularly, as is best shown in the NE1/4SW1/4 section 2 and the SE1/4 section 11, T. 14 N., R. 3 W. These map relations suggest that a small fault, substantially parallel to bedding, is in the upper Calls Fort Member. The fault is nowhere well exposed, but it is inferred in other exposures of Calls Fort strata. Burton (1973) and Green (1986) first interpreted this feature as a thrust fault separating the Bloomington and Nounan Formations. However, as described earlier, they believed it cut out most of the middle and all of the upper members of the Bloomington Formation. The fault could be an early Tertiary normal fault that formed upon relaxation of Sevier-age compression, prior to Basin and Range extension.

Klippen on the West Slope of Clarkston Mountain

Several klippen of younger-over-older strata are present on the crest and west flank of Clarkston Mountain. Where exposed, the bases of the klippen are pervasively brecciated, whereas strata within the blocks are relatively intact. The small southernmost klippe is on the west flank of Gunsight Peak and consists of the upper orthoquartzite member of the Swan Peak Formation and lower Fish Haven-Laketown strata that overlie the St. Charles Formation. This block has been displaced about 4,500 feet (1,370 m) down to the west from exposures nearer to Gunsight Peak. A large block of middle and upper Swan Peak and lower Fish Haven-Laketown strata is present on the crest of Clarkston Mountain northeast of Gunsight Peak. These strata dip east and are bounded by a sub-horizontal fault; the block itself is cut by younger steeply dipping normal faults. We mapped five klippen of Nounan and St. Charles strata over the Bloomington Formation outcrop belt north of Gardner Canyon on the west flank of Clarkston Mountain. These blocks are probably erosional remnants of a once much larger klippe, the western part of which is cut by the Gunsight fault.

These younger-over-older klippen are probably the hanging-wall remnants of gently dipping normal faults as originally suggested by Hanson (1949). The positions of most of these klippen suggest a stratigraphic separation of less than a few thousand feet, although offset is difficult to determine because source beds of some blocks are now down-faulted and lower in elevation than the klippen themselves. Because source beds of the large block east of Gunsight Peak are at an elevation below the block itself, emplacement of the block must have occurred prior to or during early movement on the Gunsight fault. Similarly, an early or pre-normal-fault emplacement is suggested for the northern set of klippen, whose source beds are now down-faulted on the east flank of Clarkston Mountain. Because they are deeply dissected, cut by moderately to steeply dipping normal faults, and lie on the deeply eroded crest and flanks of the range, the blocks were probably emplaced in the late Tertiary, prior to extensive movement of the Gunsight fault.

Junction Hills and Clarkston Piedmont

South of the Short Divide fault in the Clarkston and Portage quadrangles, the Salt Lake Formation is offset by at least 30 faults that bound at least 25 separate blocks (figure 4). Folds with a wide range of trends also deform these rocks. These strata exhibit two major domains separated by a down-to-the-west normal fault (fault #22) along the west side of the Junction Hills (the Plymouth fault of Swenson [1997] and the Beaver Dam fault of Goessel [1999; Goessel and others, 1999]). Fault #22 places the oolite subunit of Plymouth in the west down against older strata of the conglomerate subunit of Bensons Hollow and micrite subunit of Short Divide and Oquirrh Formation. The fault appears to trend uphill into the saddle just west of Short Divide, where it may curve west and merge with a splay of the Short Divide fault (fault #18) just north of a zone of steep to overturned beds in the zeolite subunit of Long Divide and tephra subunit of Junction Hills. East of the Beaver Dam fault, strata of the lower part of the Salt Lake Formation overlie Oquirrh bedrock and dip generally east-northeast. Steeply dipping normal faults #30, #31, and #32 predate the gently west-dipping Beaver Dam fault (fault #22). An inlier of Oquirrh strata between Bob and Al Archibald Hollows and local flat dips in the zeolite subunit of Long Divide there support the presence of down-to-the-west normal fault #23.

The western domain, along the southern base of Clarkston Mountain, is highly faulted. The Short Divide fault forms the north boundary of this domain, while a queried fault coincident with the Bonneville shoreline forms its southern boundary. Numerous north-striking faults of modest offset are also present. West of Water Canyon, folds in the Salt Lake Formation near the Short Divide fault plunge east parallel to the Short Divide fault. To the south, near and west of Water Canyon and south of east-striking faults #5, #7, #11, and #15 (figure 4), small folds in the Salt Lake Formation plunge mainly southward and define an overall south-plunging antiform. Faults #5, #7, #11, and #15 appear to be an older west-striking fault system that continues west from the Short Divide fault (fault #16) where the latter swings abruptly northwest toward Water Canyon. This older fault appears to be offset by the younger, north-striking faults. Faults #14 and #20 are parallel to a set of northeast-striking faults in Paleozoic bedrock of Clarkston Mountain. Inactivity on these two faults that cut Salt Lake strata but not the pediment may signify recent similar inactivity of northeast-striking faults in bedrock within the range. A similar conclusion of inactivity may be drawn for north-northwest-striking fault #12, based on the lack of offset of pediment surfaces, even though this fault appears to have offset the older east-striking fault system.

Correlation of pediment surfaces is difficult, and those along the south flank of Clarkston Mountain are no exception. Oaks, who mapped this area, suggests that the two different pediment levels at the center of section 5, T. 13 N., R. 2 W. record about 100 feet (30 m) of dip-slip throw on fault #19 between block "M" in the footwall and blocks "Q" and "R" in the hanging wall. He also notes that tephra correlations and lack of matching lithologies in the tephra subunit of Junction Hills across fault #19 suggest about 170 feet (50 m) of pre-pediment, net dip-slip throw with block "M" in the hanging wall. Inferred pre- and post-pediment displacement on this fault is thus in opposite directions. Oaks' co-authors, however, interpret the different pediment levels as having resulted from changes in base level due to some combination of movement on the Wasatch, Short Divide, or West Cache fault zones.

West Hills

Only a small part of the West Hills is in the southwest corner of the Portage quadrangle. The West Hills consist of poorly exposed, mostly west-dipping Oquirrh Formation strata. To the west in the Lime Kiln Knoll quadrangle, Murphy and others (1985) mapped the northwest-trending, down-to-the-south North Canyon normal fault, which separates generally west-dipping lower Paleozoic rocks on the northeast from folded upper Paleozoic rocks on the southwest. The North Canyon fault is concealed by Quaternary deposits at the east margin of the West Hills, but it may continue eastward into Malad Valley west of Washakie. The North Canyon fault has a small amount of apparent reverse offset at its northern end in Samaria Mountain of southern Idaho, but shows a normal displacement at its southern end in the West Hills (Allmendinger and others, 1984). Both Allmendinger and Platt (1983) and Murphy and others (1985) interpreted the down-dropped block southwest of the North Canyon fault as the upper plate of a block that was thrust eastward along the Samaria Mountain thrust. Strata in the West Hills in the Portage quadrangle belong to this upper plate, the Hansel allochthon.

Bjorklund and McGreevy (1974) and Klauk and Budding (1984) inferred a concealed fault along the west side of Malad Valley. Based on a fairly uniform slope of limited gravity data in Malad Valley, a fault with significant offset may not be present along this part of the west side of Malad Valley.

ECONOMIC GEOLOGY

Aggregate

In the Clarkston and Portage quadrangles in the late 1990s, sand and gravel were being mined only from deposits associated with the Provo shoreline at the southwest corner of Clarkston Mountain. The Utah Department of Transportation Materials Inventories of Box Elder and Cache Counties (Utah State Department of Highways, 1965, 1967) contain basic analytical information on these and other aggregate deposits in the quadrangles. Inactive or abandoned sand and gravel pits are common in both quadrangles and are shown on the geologic maps with a symbol. In the Portage quadrangle, sand and gravel pits are commonly associated with Provo shoreline deposits, and in the Clarkston quadrangle, they are commonly associated with Bonneville shoreline deposits.

In the Portage quadrangle, Bonneville shoreline sand and gravel deposits tend to be small and difficult to access, and they contain almost no gravel in the West Hills. However, we mapped a large sand and gravel deposit (Qlgb) of uncertain thickness north of Johnson Canyon. This deposit appears to form a veneer over an Oquirrh bedrock high. We mapped comparatively large and mostly unexploited sand and gravel deposits, formed as Provo shoreline deltas (Qldp), at the entrances to Broad and Johnson Canyons in the West Hills. In the Clarkston quadrangle, most sand and gravel deposits associated with the Bonneville shoreline are thin, although comparatively extensive deposits of uncertain thickness are mapped along the south end of the Washboards and between the Washboards and Bergeson Hill. Although not as well sorted as lacustrine deposits mapped as Qlgp and Qlgb, enormous quantities of poorly to moderately sorted sand and gravel are in alluvial-fan deposits shed off Clarkston Mountain and the West Hills. Gravels derived from the Salt Lake Formation are generally undesirable because of their relative softness and tendency to flake apart when they are wetted and dried.

Oil and Natural Gas

In northern Utah, most petroleum exploration and development activity has been on the eastern three major thrust plates – the Crawford, Absaroka, and Hogsback plates – with very little activity in the Paris-Willard thrust plate in which the Clarkston and Portage quadrangles are located. The potential for petroleum discoveries in the quadrangles, for example within the Crawford plate beneath the Paris-Willard thrust, is unknown and largely unexplored. However, Oaks and Kendrick (1992) believed that rocks beneath the Paris-Willard thrust west of longitude 111°30′ W. have been buried too deeply for too long to be productive.

Four relatively shallow, unsuccessful, closely spaced wildcat wells were drilled, however, in the northwest corner of the Clarkston quadrangle. According to records at the Utah Division of Oil, Gas and Mining (DOGM), the Harold and Vivian Clark #3 well (API 43-005-30011), in the SW1/4 section 33, T. 15 N., R. 2 W., was spudded in 1981 and abandoned in 1990 at a depth of 3,006 feet (916 m). The #1 Rudger Barson well (API 43-005-30006), in the NW1/4 section 33, T. 15 N., R. 2 W., was drilled to a depth of 1,730 feet (527 m) and abandoned in 1981. The Harold and Vivian Clark #2 well (API 43-005-30005), in the NW1/4 section 4, T. 14 N., R. 2 W., was drilled to a depth of 2,135 feet (651 m) in 1980; in 1984 it was turned into a water well. The Christiansen #1 well (API 43-005-30008), in the SE1/4 section 32, T. 15 N., R. 2 W., was drilled to a depth of 595 feet (181 m) and in 1984 was completed as a water well.

Natural gas was produced for local use from several wells near Brigham City (Campbell, in Doelling, 1980). The gas, locally known as swamp gas or marsh gas, was probably generated from the decay of organic material in lacustrine sediments. The potential for similar small reserves in the Clarkston and Portage areas is unknown. Similar occurrences in Cache Valley are all found where the ground surface is below the Provo shoreline.

Prospects

We mapped several small prospects near the margins of Clarkston Mountain. Most prospects are on or near faults in brecciated carbonates stained by iron-manganese oxides. Two prospects in Salt Lake Formation strata are just south of the entrance to Steel Canyon, in beds stained by iron-manganese oxides and locally baked by basaltic intrusions. An adit is located at the mouth of Mine Hollow, near the stratigraphic contact of the zeolite subunit of Long Divide and tephra subunit of Junction Hills.

Geothermal Resources

DeVries (1982) assessed low-temperature geothermal resources of Cache Valley, and Klauk and Budding (1984) provided a similar assessment geothermal resources of the eastern portion of Box Elder County. In Malad Valley, Klauk and Budding (1984) identified an area west and south of Plymouth as having low-temperature geothermal potential. This area includes Uddy hot springs (Belmont Hot Spring Resort), which are about one mile (1.6 km) southwest of Plymouth on the Malad River flood plain. These springs are developed with a swimming pool and mineral bath, and discharge an estimated 1,560 gallons per minute (6,000 L/min) at 127°F (53°C) (Blackett, 1995). Murphy and Gwynn (1979) summarized the geology and hydrogeology of Uddy hot springs, and noted that this system results from the deep circulation of meteoric water. Geothermal-gradient data for wells in Malad Valley are summarized by Klauk and Budding (1984) and Blackett (1994).

DeVries (1982) noted that the area around Trenton, about 4 miles (8.5 km) east of Clarkston in Cache Valley, has anomalously warm ground water and that it is therefore an area of low-temperature geothermal potential. Ground-water temperatures of 73 to 122°F (23-50°C) are known in the area.

Building and Ornamental Stone

No building-stone or ornamental-stone quarries are in the Clarkston or Portage quadrangles, although this area does contain stratigraphic units that were quarried elsewhere in northern Utah for local use as building stone. Oviatt (1986a) noted that oolitic limestone from the Salt Lake Formation was quarried on a small scale in the adjacent Cutler Dam quadrangle. Purple quartzite of the middle member of the Swan Peak Formation and limestone of the Garden City Formation were used locally as building stone in some early structures, including the Logan Temple, Logan Tabernacle, and the foundations of some older buildings on the Utah State University campus (Morgan, 1992). However, both units are difficult to access in the Clarkston and Portage quadrangles.

The Garden City Formation, which is exposed as a near dip slope west of Clarkston, contains thin- to medium-bedded, planar-bedded, bluish-gray limestone with black chert nodules that could serve as good flagstone or rough dimension stone. The middle limestone member of the Bloomington Formation contains planar-bedded, light-gray oolitic limestone that could serve similar uses, but it is only exposed along the rugged northwest flank of Clarkston Mountain.

Zeolites

The zeolite subunit of Long Divide, which is widely exposed in the Junction Hills and along the south flank of Clarkston Mountain, is characterized by zeolite-bearing, pale-green siltstones. Mayes and Tripp (1991) noted the occurrence of the zeolite clinoptilolite in tuffs of the Salt Lake Formation in this area. These beds may have potential for slow-release chemical fertilizers and other uses.

WATER RESOURCES

Surface Water

The average annual precipitation in the Clarkston and Portage quadrangles is about 25 to 30 inches (64-76 cm) along the crest of Clarkston Mountain, and about half that in the adjacent valleys and low hills (Bjorklund and McGreevy, 1974). The boundary between Box Elder and Cache Counties, which follows the crest of Clarkston Mountain and the Junction Hills, marks a drainage divide between Cache and Malad Valleys.

To the east, Clarkston Creek serves as the principal drainage in the Cache Valley portion of the Clarkston quadrangle. Clarkston Creek is not gauged, but Kariya and others (1994) estimated its mean annual flow at an anomalously low 144 acre-feet/year (175,634 m^3/yr) based on records at Newton Reservoir. Bjorklund and McGreevy (1971), however, estimated the mean annual flow of Clarkston Creek at 4,000 acre-feet/year (4,878,720 m^3/yr) during the period 1960 to 1968. We suspect that the mean annual flow of Clarkston Creek is closer to that reported by Bjorklund and McGreevy (1971). Newton Reservoir, part of which is in the Clarkston quadrangle, stores water from Clarkston Creek for use by the Newton Water Users Association.

To the west, the Malad River is the trunk stream in the Box Elder County portion of the Portage quadrangle. Bjorklund and McGreevy (1974) reported an average annual flow of 36,000 acre-feet/year (43,908,480 m^3/yr) for the Malad River at Woodruff, Idaho, just north of the quadrangle. At low flows, the Malad River is moderately saline, and as the Malad River enters the Portage quadrangle, most of the base flow comes from Woodruff Springs, about 4 miles (6.4 km) north of the state line in Idaho (McGreevy, 1972). The Malad River gains ground-water discharge southward through the Portage quadrangle.

Numerous intermittent streams drain the highlands surrounding Cache and Malad Valleys. Except during periods of exceptional flow, discharge from most of these streams disappears into the upper portions of alluvial fans that surround the highlands, thereby recharging ground water.

Ground Water

Unconsolidated late Tertiary to Quaternary basin-fill deposits form the principal aquifers in the Cache Valley portion of the Clarkston quadrangle. Because these deposits are thinner in the Clarkston area than to the east, Kariya and others (1994) considered the Clarkston bench to be a separate ground-water system from the main Cache Valley, upon which their study focused. They noted that near the Clarkston Mountain front, several hundred feet of unconsolidated basin fill may remain unsaturated and that ground water is generally unconfined there, although lenses of perched water may be present above impermeable clay layers. Eastward, along Clarkston Creek and adjacent lowlands, the depth to ground water is generally less than 10 feet (3 m) (Bjorklund and McGreevy, 1971; Kariya and others, 1994). Kariya and others (1994) noted that ground water from the Clarkston bench recharges the main ground-water system of Cache Valley and also discharges directly to Cutler Reservoir. On the Clarkston bench, ground-water recharge is by infiltration of precipitation and seepage of streams; most recharge probably happens as intermittent streams flow across coarse alluvial-fan deposits at the mountain front (Kariya and others, 1994). Ground-water quality in the Clarkston area is generally good with 300 to 900 mg/L total dissolved solids (TDS) (Bjorklund and McGreevy, 1971).

Numerous springs occur along the flanks of Clarkston Mountain and the Junction Hills. Springs near Clarkston provide water for the Clarkston, Trenton, and Newton public water systems. Northwest of Clarkston, many of these springs issue from the toe of alluvial-fan deposits where they overlie the subunit of the Washboards, suggesting that the silty and clayey Washboard deposits act as a barrier to downward ground-water flow.

Bjorklund and McGreevy (1974) summarized ground-water resources of the lower Bear River drainage basin, which includes the Malad Valley of the Portage quadrangle. They showed the central Malad Valley as an area of natural discharge, with ground water flowing toward the center of the valley and then southward toward Great Salt Lake. Except for local areas of perched water, the depth to ground water is in excess of 100 feet (30 m) along the margins of Malad Valley (Bjorklund and McGreevy, 1974). Because of the thickness of unconsolidated late Tertiary to Quaternary basin-fill deposits in Malad Valley, Bjorklund and McGreevy (1974) noted that the possibility for development of large-discharge wells in the Portage area is good; conversely, while development of wells with discharges to 50 gpm (180 L/min) or more is possible in the Plymouth area, the comparatively shallow depth to Salt Lake bedrock there will likely preclude the development of wells with much larger discharges.

Springs are numerous along the central portion of Malad Valley. Bjorklund and McGreevy (1974) noted that Plymouth obtains water for its public-water system from eight springs and one well on the southern flank of Clarkston Mountain, while Portage obtains water from one nearby well and one spring in Portage Canyon in the adjacent Limekiln Knoll quadrangle.

Bjorklund and McGreevy (1974) reported on ground-water quality in Malad Valley. They noted that the TDS concentration for springs and shallow wells was generally less than 600 mg/L, although a TDS concentration of 900-1,800 mg/L is common in the Washakie area. They classified ground-water sodium and salinity hazards and noted that most areas are low-sodium and low- to medium-salinity hazards, with a few areas near Portage, Plymouth, and Washakie as low-sodium and high-salinity hazard areas. Because the central Malad Valley is a discharge area, ground-water quality is generally not affected by the quality of Malad River water.

The numerous faults that offset the Salt Lake Formation along the south flank of Clarkston Mountain break this unit up into small ground-water compartments. The predominance of tuffaceous strata, typically altered to swelling clays, and the presence of folds further restrict the transmission of water from compartment to compartment. The scarcity of extensive, well-sorted sand and gravel reduces the likelihood of sustainable high yields in that area.

In the Clarkston quadrangle, the overall westward dip of Salt Lake strata from Bergeson Hill west to at least the west margin of the Washboards, and the rather smooth east-west gravity profiles, suggest a less complex domain there. Thin gravels exposed in the east face of the Washboards, and relatively unaltered tuffaceous sediments, suggest the potential for wells with modest yields near the west margin of the Washboards. The possible effect of such wells on Clarkston Creek is unknown, as is the quality of ground water in this agricultural area. Between the Washboards and Bergeson Hill, wells with modest yields may be recharged from outcrops above the Bonneville shoreline.

Oaks (2000) used field mapping, tephrochronology, oil and gas and water-well logs, and interpretation of gravity and seismic reflection data to evaluate ground-water resources in western Cache Valley and adjacent Box Elder County, including much of the Clarkston and Portage quadrangles. Among his findings are that aquifers in the Salt Lake Formation are limited in extent by folds, faults, and complex stratigraphy. Thus, Salt Lake Formation aquifers are typically of low to moderate yield and of variable quality. He further noted that locally it may be possible to drill through the Salt Lake Formation into Paleozoic bedrock along the valley margins, or move laterally to the Paleozoic bedrock of the mountain itself, in search of promising wells.

GEOLOGIC HAZARDS

The following descriptions of earthquakes hazards, mass movements, flooding, shallow ground water, problem soils, and radon identify potential problem areas and should not be used in place of site-specific hazard investigations.

Earthquakes and Seismic Hazards

The Clarkston and Portage quadrangles lie within the Intermountain seismic belt, a north-trending zone of pronounced seismicity that trends from northern Arizona to western Montana (Smith and Arabasz, 1991). Two major faults – the Wasatch fault zone and the West Cache fault zone – are known to have Quaternary offset in the Clarkston and Portage quadrangles (Machette and others, 1992; Solomon, 1999; Black and others, 1999, 2000; this report). In addition, several other normal-fault zones are present in the vicinity, including the Oxford-Dayton fault, and the East Cache fault zone, on which a large prehistoric earthquake of about magnitude 7.0 (M_S) occurred about 4,000 to 7,000 years ago (McCalpin and Forman, 1991). Earthquake hazards in these two quadrangles are therefore significant and include surface fault rupture, ground shaking, slope failures, liquefaction, flooding, rock falls, and other seismically induced hazards.

Black and others (1999, 2000) estimated paleo-earthquake magnitudes (moment magnitudes) for surface-faulting events on the Clarkston and Junction Hills segments of the West Cache fault zone of M_W 7.0-7.1. They further suggested that two or three surface-faulting events have occurred on the Clarkston segment since the Bonneville highstand about 14.5 ka. Black and others (1999, 2000) estimated a maximum slip rate for the Clarkston segment of 0.027 in/yr (0.68 mm/yr) since the late Pleistocene and a maximum slip rate of 0.008 in/yr (0.21 mm/yr) for the Junction Hills segment. Paleoseismic studies have not been undertaken on the Clarkston Mountain or Collinston segments of the Wasatch fault zone.

Other normal faults with Quaternary movement, though concealed, are present in and adjacent to the quadrangle. The Oxford-Dayton fault is concealed along the east side of the ridge that contains Bergeson Hill, Pete McCombs Hill, and Little Mountain (Newton Hill). Because scarps are not visible in Lake Bonneville deposits along this ridge, large earthquakes (M_L >6.0) probably have not occurred there in the Holocene. The inferred fault at the Bonneville shoreline at the south end of Clarkston Mountain must have been active in the Quaternary to produce such a large scarp, although there is no evidence of post-Bonneville movement.

The damage potential from surface fault rupture and associated deformation is greatest along and adjacent to both exposed and concealed normal faults that exhibit evidence of Holocene or latest Pleistocene activity (plates 1 and 2). Holocene fault scarps are present on the West Cache fault zone in the Clarkston quadrangle. Trenching and scarp profiles by Black and others (1999, 2000) showed 10.2 to 12.1 feet (3.1-3.7 m) of Holocene displacement on the West Cache fault zone at Winter and Raglanite Canyons. A scarp about 15 feet (4.5 m) high is present at the entrance to Elgrove Canyon along the Wasatch fault zone. Any development along or across an active normal fault could be damaged when the fault ruptures.

Four earthquakes of magnitude 4.0 or greater have been recorded in northern Utah from 1962 through 1990 (Evans, 1991). The most recent large earthquake in the greater Clarkston-Portage area occurred in August 1962, and is known as the Richmond or Cache Valley quake. The earthquake had a magnitude of 5.7 (M_S) and the epicenter was approximately 4.7 miles (7.5 km) east of Richmond, Utah, in the Bear River Range; the focus might have been on the Temple Peak fault zone (Westaway and Smith, 1989), although Evans (1991) noted that there is no consensus as to the form or location of the fault responsible for the earthquake. Major structural damage occurred in Richmond, Lewiston, Logan, and Smithfield, Utah during this earthquake (Lander and Cloud, 1964). Because prehistoric Holocene earthquakes along the East and West Cache fault zones were of greater magnitude, and because considerable new construction has occurred since 1962, the potential exists for much greater earthquake damage than that sustained during the 1962 earthquake. Intense historical earthquake activity has occurred in the Hansel Valley-Pocatello Valley corridor about 25 miles (40 km) west of the Clarkston and Portage quadrangles, including the 1909 M 6, 1934, M_L 6.6, and the 1975 M_L 6.0 earthquakes (McCalpin and others, 1992). The potential for damage in the Clarkston-Portage area resulting from large earthquakes in the Hansel Valley-Pocatello Valley corridor is uncertain.

Youngs and others (1987) and Frankel and others (1996) mapped ground-shaking potential along the Wasatch fault zone. However, neither of these estimates of peak ground acceleration take into account local geology, such as the fine-grained, unconsolidated sediments in the Cache and Malad Valley grabens. Local geologic characteristics can increase ground acceleration. Structures built on Quaternary units that contain fine-grained sediments and have shallow ground-water levels (Bjorklund and McGreevy, 1971; these maps) are most prone to damage from ground shaking. These units include alluvial, lacustrine, and mixed alluvial and lacustrine deposits mapped in Cache and Malad Valleys. For any given earthquake, ground shaking tends to be more severe on unconsolidated deposits like those in the valley than for bedrock, which is present in Clarkston Mountain, the Junction Hills, and north of Clarkston Creek. Buildings should be constructed in accordance with the International Building Code (2000).

During ground shaking, pore-water pressure can increase in saturated granular sediments. This increase leads to a loss of contact between grains and loss of strength. These materials can then behave as a dense fluid and flow. This transformation from a solid to a fluid state is termed liquefaction. The resulting ground failure can cause structural damage to buildings, roads, railways, utilities, and other construction. Saturated fine-grained sand and silt are most susceptible to liquefaction (Youd, 1984; Obermeier, 1996). Liquefaction features were observed in Cache and Bear River Valleys following the 1962 Cache Valley earthquake (Lander and Cloud, 1964). Bay (1987) (see also Anderson and others, 1994) used engineering methods to map liquefaction potential in Cache and Box Elder Counties, and showed very low potential on alluvial fans and Lake Bonneville deposits south of Clarkston Mountain in Box Elder County, and very low potential for the Cache County portion of the Clarkston quadrangle. Liquefaction potential along Clarkston Creek and in the low area east of Clarkston should probably be re-examined in light of the shallow water table and fine-grained sediments in these areas.

Various kinds of earthquake-induced slope failures are a potential local hazard. Liquefaction, even of buried sediments, can lead to lateral spread of material on slopes as low as 0.5 percent (0.3°) and flow landslides on slopes steeper than 5 percent (≥ 3°) (Youd, 1978; National Research Council, 1985). This can lead to the failure of stream banks and dikes. Other serious, seismically induced slope-failure hazards in the quadrangles include rock falls and landslides in the mountains and along the mountain front. See the mass movement sub-section for potential hazard areas in the quadrangle.

Earthquake-related flooding can result from dam failure, canal failure, landslides that dam drainages, diversion of streams, lowering and tilting of ground levels, and rupturing of water lines.

Mass Movements

Mass movement is a natural hillslope process in which rock and soil move downslope under the direct influence of gravity. Hazardous forms of mass movement in the quadrangle include debris flows, landslides (slides, slumps, and flows), and rock falls. Most buildings and homes in the Clarkston and Portage quadrangles are presently located away from the steep slopes and canyons where these mass movements are most likely to happen. Still, other structures, including roads and utilities, could be damaged by mass movements, as could future development in landslide-prone areas.

We mapped debris-flow deposits (Qmf) on alluvial fans along the east flank of Clarkston Mountain. Less well preserved, unmapped debris-flow deposits doubtless are present at the mouths of most drainages along the flanks of Clarkston Mountain, West Hills, and Junction Hills. The highest potential for damage due to debris flows is thus in the upper parts of alluvial-fan deposits mapped as Qaf_1, Qafy, and Qlag along these mountain fronts. Small debris-flow deposits are also present in Holocene alluvium and colluvium (Qac) in mountain drainages, making these areas of debris-flow hazard as well.

Landslides – including slides, slumps and flows mapped as Qmsy, Qmso, Ql/Qmsy, and Ql/Qmso – in the Clarkston and Portage quadrangles are mostly developed on the Salt Lake Formation, although significant landslide hazards may also be found on steeper slopes of the Lake Bonneville shorelines. The Hodges Shale and Calls Fort Shale Members of the Bloomington Formation are also susceptible to landsliding in this region. Landslide scarps are shown as hachured lines on plates 1 and 2. Most of the largest landslides in the quadrangle are in Salt Lake strata exposed near the Bonneville shoreline, and may have resulted from erosion of shoreline slopes, rapid dewatering of slopes following the Bonneville flood, or earthquakes. We also mapped smaller Holocene and historical slides and slumps in Salt Lake strata and overlying colluvium.

Evidence of rock falls is present throughout the mountainous portions of the quadrangles as accumulations of boulders at the base of steep slopes. Rock falls are a natural part of the erosion process, where resistant, fractured or jointed strata break apart and tumble downslope. They are commonly associated with heavy rainfall events, snow avalanches, or earthquakes, but many probably happen as isolated random events after prolonged freeze-thaw weathering. Slopes that are oversteepened by construction activities may present additional rock-fall hazards. The extent of the hazard can be assessed by the relative abundance of rock-fall debris at the base of a slope. The relative hazard varies locally and depends upon the distance from the base of the slope, nature and stability of slope debris, and local topography.

Flooding

Canyons in the quadrangles are susceptible to flash flooding during heavy rainfall and rapid snowmelt. Cloudburst storms produced some flooding in the Clarkston area on July 30, 1917 (Woolley, 1946); August 22, 1958, and August 25, 1961 (Butler and Marsell, 1972); and in 1980 and 1981 (Utah Division of Comprehensive Emergency Management, 1981). Large volumes of snowmelt runoff in the spring of 1983 resulted in floods, debris flows, and landslides along mountain fronts in most of northern Utah, but damage in Cache and northern Box Elder Counties was minimal (Wieczorek and others, 1983, 1989; Anderson and others, 1984). Alluvial fans (Qaf1, Qafy, Qlag) along the mountain front and mixed alluvium and colluvium (Qac) in mountain drainages in the quadrangles are especially vulnerable as potential sites of flash floods.

Federal Emergency Management Agency (FEMA) and Flood Insurance Agency (FIA) flood insurance rate maps covering the Cache County portion of the quadrangles show special flood-hazard areas along Clarkston Creek (FEMA, 1987a, 1987b), along City and Myler Creeks (FIA, 1980; FEMA, 1987b), and along unnamed drainages near Round Knoll and heading in Al Archibald Hollow (FEMA, 1987b). FEMA maps covering the Box Elder County portion of the quadrangles show special flood hazard areas along the Malad River and several tributaries (FEMA, 1987c, 1987d, 1987e, 1987f). These FEMA areas correspond in part to areas mapped as alluvial deposits (Qal). FEMA maps, however, do not address hazards associated with flash flooding that are present on younger alluvial-fan deposits. In contrast to major riverine floods, flash floods are highly localized and unpredictable. They quickly reach a maximum flow and then quickly diminish. Flash floods commonly contain high sediment or debris loads and commonly begin or end as debris floods or flows, further adding to the destructiveness of such events.

Shallow Ground Water

Ground water is within 10 feet (3 m) of the ground surface in the central part of the Clarkston quadrangle. Spring locations and soil mapping by Erickson and Mortensen (1974) suggest that shallow ground water is present in the flood plain along Clarkston Creek and in the low area east of Clarkston. The actual extent of shallow ground water is not known because water levels are only available for a couple dozen wells, most of which are close to the town of Clarkston. Seasonal variations in precipitation and runoff can alter the water levels (Bjorklund and McGreevy, 1971). Shallow ground water hampers excavation and construction of foundations, and can preclude installation of septic tanks and subsurface structures; it also increases the possibility of liquefaction during an earthquake. Placing underground storage tanks, landfills, or waste sites in areas where ground water is at or near the surface is ill-advised because of the potential for ground-water contamination.

Shallow ground water is also known in the Malad Valley portion of the Portage quadrangle. Bjorklund and McGreevy (1974) noted this to be an area of natural ground-water discharge, and spring locations and soil mapping by Chadwick and others (1975) also suggest that shallow ground water is present in the central portion of Malad Valley. Severe high-water-table problems plagued the town of Plymouth in 1983 and probably resulted in part from a shallow hardpan layer (Schick International, 1983).

Problem Soil and Rock

Expansive, collapsible, and saline and alkaline soils and rocks are known in the Clarkston and Portage quadrangles. Expansive soil and rock contain clay minerals that swell conspicuously when wet and shrink as they dry. This swelling and shrinking can cause significant foundation problems and can damage roads and underground utilities. Montmorillonite clay of the smectite group is the principal material susceptible to such changes. This type of clay is commonly formed by the alteration of volcanic glass, of which the Salt Lake Formation contains large amounts. Chadwick and others (1975) showed soils having a moderate to high shrink-swell potential on some gravelly surfaces (Qap_1 and Qap_2) on the Salt Lake Formation on the south flank of Clarkston Mountain, and areas of low to moderate shrink-swell potential elsewhere throughout the Box Elder County portion of the Portage quadrangle. Erickson and Mortensen (1974) showed soils having a moderate to high shrink-swell potential in the Cache Valley portion of the Clarkston quadrangle, especially in clayey lake-bottom sediments (Qlmb); in mixed, fine-grained, lake-bottom and flood-plain sediments (Qla) in areas near Clarkston Creek and Dahle Hollow; and on scattered alluvial fans (Qaf_1, Qafy) on the east flank of Clarkston Mountain. They also showed areas of low to moderate shrink-swell potential elsewhere throughout the Cache County portion of the Clarkston quadrangle. Generally, Quaternary units derived in large part from Salt Lake strata may contain expansive soils.

Collapsible soils may be in geologically young, loose, dry, low-density deposits such as are common in Holocene-age alluvial-fan and colluvial depositional environments. Collapsible soils have considerable strength and stiffness when dry, but can settle dramatically when wet, thereby causing significant damage to roads and other structures. This collapse, called hydrocompaction, can happen when susceptible soils are wetted below the level normally reached by rainfall, destroying the clay bonds between grains. Irrigation water, lawn watering, or water from leach fields can initiate hydrocompaction.

Chadwick and others (1975) mapped soils that are moderately to strongly alkaline or saline in the central and northern parts of the Malad Valley in the Portage quadrangle. Such soils limit agricultural use of the land, and construction activities in these soils may require special foundation materials. These soils are underlain by deposits herein mapped in part as mixed lacustrine and flood-plain deposits (Qlam). Erickson and Mortensen (1974) and Chadwick and others (1975) did not note any saline or alkaline soils in the Clarkston quadrangle.

Erosional pipes form when surface runoff infiltrates and percolates through soils at a rate sufficient to cause subsurface erosion in poorly lithified sediments. Piping can produce a system of tunnels, small caves, and pseudokarst topography that collectively serves to channel runoff underground. The principal danger associated with piping is roof collapse. Piping we observed in the Clarkston and Portage quadrangles is generally restricted to landslide deposits and fine-grained lacustrine and alluvial sediments near cut-bank exposures and is generally of small scale. Erickson and Mortensen (1974) also noted piping in some soils.

Radon

Radon is an odorless, tasteless, colorless radioactive gas that is found in small concentrations in nearly all rocks and soil. Radon can become a health hazard when it accumulates in sufficient concentrations in enclosed spaces such as buildings. A variety of geologic and non-geologic factors combine to influence concentrations of radon indoors, including soils or rocks with naturally elevated levels of uranium, soil permeability, ground-water levels, atmospheric pressure, building materials and design, and other factors. Indoor-radon concentrations can vary dramatically within short distances due to both geologic and non-geologic factors.

The radon hazard has been depicted only in a general way in the Clarkston and Portage quadrangles (Sprinkel and Solomon, 1990; Black, 1993). These generalized maps, which should not be used in place of site-specific studies, show that most of the Clarkston and Portage quadrangles have a moderate to high radon-hazard potential, with high radon-hazard potentials generally restricted to areas underlain by bedrock. It is important to note, however, that a quantative relationship between geologic factors and indoor-radon levels does not exist, and that localized areas of higher or lower radon potential are likely to be present in any given area. Actual indoor-radon levels can vary widely over short distances, even between buildings on a single lot.

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REFERENCES

Adamson, R.D., 1955, The Tertiary Salt Lake Group in Cache Valley, Utah-Idaho: Logan, Utah State Agricultural College, M.S. thesis, 59 p.

Adamson, R.D., Hardy, C.T., and Williams, J.S., 1955, Tertiary rocks of Cache Valley, *in* Eardley, A.J., editor, Tertiary and Quaternary geology in the eastern Bonneville basin: Utah Geological Society Guidebook to the Geology of Utah, no. 10, p. 1-22.

Allmendinger, R.W., 1992, Fold and thrust tectonics of the western United States exclusive of the accreted terranes, *in* Burchfiel, B.C., Lipman, P.W., and Zoback, M.L., editors, The Cordilleran orogen – conterminous U.S.: Boulder, Colorado, Geological Society of America, The Geology of North America, volume G-3, p. 583-607.

Allmendinger, R.W., Miller, D.M., and Jordan, T.E., 1984, Known and inferred Mesozoic deformation in the hinterland of the Sevier belt, northwest Utah, *in* Kerns, G.J., and Kerns, R.L., Jr., editors, Geology of northwest Utah, southern Idaho and northeast Nevada: Utah Geological Association Publication 13, p. 21-34.

Allmendinger, R.W., and Platt, L.B., 1983, Stratigraphic variation and low-angle faulting in the North Hansel Mountains and Samaria Mountain, southern Idaho, *in* Miller D.M., Todd, V.R., and Howard, K.A., editors, Tectonic and stratigraphic studies in the eastern Great Basin: Geological Society of America Memoir 157, p. 149-164.

Anderson, L.R., Keaton, J.R., and Bay, J.A., 1994, Liquefaction potential map for the northern Wasatch Front, Utah: Utah Geological Survey Contract Report 94-6, 150 p., scale 1:48,000.

Anderson, L.R., Keaton, J.R., Saarinen, T.F., Wells, W.G., II, 1984, The Utah landslides, debris flows, and floods of May and June 1983: Washington, D.C., National Academy Press, 96 p.

Bay, J.A., 1987, Liquefaction potential mapping of Weber, Box Elder and Cache Counties, Utah: Logan, Utah State University, M.S. thesis, 140 p., 6 plates, scale 1:48,000.

Beus, S.S., 1963, Geology of the central Blue Spring Hills, Utah-Idaho: Los Angeles, University of California, Ph.D. dissertation, 233 p., scale 1:48,000.

—1968, Paleozoic stratigraphy of Samaria Mountain, Idaho-Utah: American Association of Petroleum Geologists Bulletin, v. 52, p. 782-808.

Biek, R.F., 1999, Geology of Clarkston Mountain (southern Malad Range), Box Elder and Cache Counties, Utah, *in* Spangler, L.E., editor, and Allen, C.J., co-editor, Geology of northern Utah and vicinity: Utah Geological Association Publication 27, p. 27-43.

Bjorklund, L.J., and McGreevy, L.J., 1971, Ground-water resources of Cache Valley, Utah and Idaho: Utah Department of Natural Resources Technical Publication No. 36, 66 p., scale 1:200,000.

—1974, Ground-water resources of the lower Bear River drainage basin, Box Elder County, Utah: Utah Department of Natural Resources, Division of Water Rights Technical Publication no. 44, 58 p.

Black, B.D., 1993, The radon-hazard-potential map of Utah: Utah Geological Survey Map 149, 12 p., scale 1:1,000,000.

Black, B.D., Giraud, R.E., and Mayes, B.H., 2000, Paleoseismology of Utah volume 9 – Paleoseismic investigation of the Clarkston, Junction Hills, and Wellsville faults, West Cache fault zone, Cache County, Utah: Utah Geological Survey Special Study 98, 23 p.

Black, B.D., Solomon, B.J., and Giraud, R.E., 1999, Surficial geology and paleoseismicity of the West Cache fault zone, Cache County, Utah, *in* Spangler, L.E., editor, and Allen, C.J., co-editor, Geology of northern Utah and vicinity: Utah Geological Association Publication 27, p. 181-201.

Blackett, R.E., 1994, Low-temperature geothermal water in Utah – A compilation of data for thermal wells and springs through 1993: Utah Geological Survey Open-File Report 311, 74 p., scale 1:750,000.

—1995, Thermal springs in Utah: Utah Geological Survey Public Information Series No. 32, scale 1:1,000,000.

Bryant, B.C., Naeser, C.W., Marvin, R.F., and Mehnert, H.H., 1989, Ages of late Paleogene and Neogene tuffs and the beginning of rapid extension, eastern boundary of the Basin and Range Province near Salt Lake City, Utah: U.S. Geological Survey Bulletin 1787-K, 11 p.

Budge, D.R., 1966, Stratigraphy of the Laketown Dolostone, north-central Utah: Logan, Utah State University, M.S. thesis, 86 p.

Budge, D.R., and Sheehan, P.M., 1980, The Upper Ordovician through Middle Silurian of the eastern Great Basin, Part 1, Introduction – Historical perspective and stratigraphic synthesis: Milwaukee Public Museum Contributions in Biology and Geology Number 28, 26 p.

Burton, S.M., 1973, Structural geology of the northern part of Clarkston Mountain, Malad Range, Utah-Idaho: Logan, Utah State University, M.S. thesis, 54 p., scale 1:12,000.

Butler, Elmer, and Marsell, R.E., 1972, Developing a state water plan – Cloudburst floods in Utah, 1939-1969: U.S. Geological Survey and Utah Division of Water Resources Cooperative Investigations Report no. 11, 103 p.

Chadwick, R.S., Barney, M.L., Beckstrand, Delyle, Campbell, Ludene, Carley, J.A., Jensen, E.H., McKinley, C.R., Stock, S.S., and Stokes, H.A., 1975, Soil survey of Box Elder County, Utah, eastern part: U.S. Department of Agriculture, Soil Conservation Service, U.S. Department of the Interior, Fish and Wildlife Service, and Bureau of Land Management, in cooperation with Utah Agricultural Experiment Station, 223 p.

Cluff, L.S., Glass, C.E., and Brogan, G.E., 1974, Investigation and evaluation of the Wasatch fault north of Brigham City and Cache Valley faults, Utah and Idaho; a guide to land-use planning with recommendations for seismic safety: Oakland, California, Woodward-Lundgren and Associates, unpublished report for U.S. Geological Survey, contract no.

14-08-001-13665, 147 p.

Coogan, J.C., 1992, Structural evolution of piggyback basins in the Wyoming-Idaho-Utah thrust belt, *in* Link, P.K., Kuntz, M.A., and Platt, L.B., editors, Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 55-82.

Crittenden, M.D., Jr., 1963, New data on the isostatic deformation of Lake Bonneville: U.S. Geological Survey Professional Paper 454-E, p. E1-E31.

—1972, Willard thrust and the Cache allochthon, Utah: Geological Society of America Bulletin, v. 83, p. 2871-2880.

Curry, D.R., 1982, Lake Bonneville – Selected features of relevance to neotectonic analysis: U.S. Geological Survey Open-File Report 82-1070, 30 p., scale 1:1,000,000.

—1990, Quaternary paleolakes in the evolution of semidesert basins, with special emphasis on Lake Bonneville and the Great Basin, U.S.A.: Paleogeography, Paleoclimatology, Palaeoecology, v. 76, p. 189-214.

DeCelles, P.G., 1994, Late Cretaceous-Paleocene synorogenic sedimentation and kinematic history of the Sevier thrust belt, northeast Utah and southwest Wyoming: Geological Society of America Bulletin, v. 106, p. 32-56.

deVries, J.L., 1982, Evaluation of low-temperature geothermal potential in Cache Valley, Utah: Utah Geological and Mineral Survey Report of Investigation No. 174, 96 p.

Doelling, H.H., with J.A. Campbell, J.W. Gwynn, and L.I. Perry, 1980, Geology and mineral resources of Box Elder County, Utah: Utah Geological and Mineral Survey Bulletin 115, 251 p., scale 1:125,000.

Dover, J.H., 1995, Geologic map of the Logan 30′ x 60′ quadrangle, Cache and Rich Counties, Utah and Uinta County, Wyoming: U.S. Geological Survey Miscellaneous Investigations Series Map I-2210, scale 1:100,000.

Erickson, A.J., and Mortensen, V.L., 1974, Soil survey of Cache Valley area, Utah: U.S. Department of Agriculture, Soil Conservation Service and Forest Service, and Utah Agricultural Experiment Station, 192 p.

Evans, J.P., 1991, Structural setting of seismicity in northern Utah: Utah Geological Survey Contract Report 91-15, 37 p., 3 plates, scale 1:100,000.

Evans, J.P., and Oaks, R.Q., Jr., 1996, Three-dimensional variations in extensional fault shape and basin form – The Cache Valley basin, eastern Basin and Range Province, United States: Geological Society of America Bulletin, v. 108, no. 12, p. 1,580-1,593.

Federal Emergency Management Agency, 1987a, Flood insurance rate map, Cache County, Utah, unincorporated areas, panel 1: Federal Emergency Management Agency, Community Panel no. 490012 0001B, scale 1:24,000.

—1987b, Flood insurance rate map, Cache County, Utah, unincorporated areas, panel 2: Federal Emergency Management Agency, Community Panel no. 490012 0002B, scale 1:24,000.

—1987c, Flood insurance rate map, Box Elder County, Utah, unincorporated areas, panel 8: Federal Emergency Management Agency, Community Panel no. 490005 0008C, scale 1:24,000.

—1987d, Flood insurance rate map, Box Elder County, Utah, unincorporated areas, panel 9: Federal Emergency Management Agency, Community Panel no. 490005 0009C, scale 1:24,000.

—1987e, Flood insurance rate map, Box Elder County, Utah, unincorporated areas, panel 17: Federal Emergency Management Agency, Community Panel no. 490005 0017C, scale 1:24,000.

—1987f, Flood insurance rate map, Box Elder County, Utah, unincorporated areas, panel 18: Federal Emergency Management Agency, Community Panel no. 490005 0018C, scale 1:24,000.

Fenneman, N.M., 1931, Physiography of western United States: New York, McGraw-Hill Book Co., 534 p.

Flood Insurance Agency, 1980, Flood insurance rate map, Town of Clarkston, Utah, Cache County: U.S. Department of Housing and Urban Development, Flood Insurance Agency, Community Panel no. 490014 0001A.

Francis, G.G., 1972, Stratigraphy and environmental analysis of the Swan Peak Formation and Eureka Quartzite, northern Utah: Logan, Utah State University, M.S. thesis, 125 p.

Frankel, Arthur, Mueller, Charles, Barnhard, Theodore, Perkins, David, Leyendecker, E.V., Dickman, Nancy, Hanson, Stanley, and Hopper, Margaret, 1996, National seismic-hazard maps - documentation June 1996: U.S. Geological Survey Open-File Report 96-532, 110 p.

Gardiner, L.L., 1974, Environmental analysis of the upper Cambrian Nounan Formation, Bear River Range and Wellsville Mountain, north-central Utah: Logan, Utah State University, M.S. thesis, 121 p.

Goessel, K.M., 1999, Tertiary stratigraphy and structural geology, Wellsville Mountains to Junction Hills, north-central Utah: Logan, Utah State University, M.S. thesis, 220 p., scale 1:24,000.

Goessel, K.M., Oaks, R.Q., Jr., Perkins, M.E., and Janecke, S.U., 1999, Tertiary stratigraphy and structural geology, Wellsville Mountains to Junction Hills, north-central Utah, *in* Spangler, L.E., editor, and Allen, C.J., co-editor, Geology of northern Utah and vicinity: Utah Geological Association Publication 27, p. 45-69.

Gray, W.E., 1975, Structural geology of the southern part of Clarkston Mountain, Malad Range, Utah: Logan, Utah State University, M.S. thesis, 43 p., scale 1:12,000.

Green, D.A., 1986, Structural geology of the central part of Clarkston Mountain, Malad Range, Utah: Logan, Utah State University, M.S. thesis, 55 p., scale 1:12,000.

Hanson, A.M., 1949, Geology of the southern Malad Range and vicinity in northern Utah: Madison, University of Wisconsin, Ph.D. dissertation, 128 p., approximate scale 1:38,400.

Hay, H.W., Jr., 1982, Petrology of the Middle Cambrian Blacksmith Formation, north-central Utah: Logan, Utah State University, M.S. thesis, 157 p.

Haynie, A.V., Jr., 1957, The Worm Creek Quartzite Member of the St. Charles Formation, Utah-Idaho: Logan, Utah State University, M.S. thesis, 39 p.

International Building Code, 2000: Falls Church, Virginia, International Code Council, Inc. 756 p.

James, W.C., 1973, Petrology and regional relationships of the Ordovician Kinnikinic Formation and equivalents, central and southern Idaho: Logan, Utah State University, M.S. thesis, 268 p.

Janecke, S.U., and Evans, J.C., 1999, Folded and faulted Salt Lake Formation above the Miocene to Pliocene New Canyon and Clifton detachment faults, Malad and Bannock ranges, Idaho – Field trip guide to the Deep Creek half graben and environs, *in* Hughes, S.S., and Thackray, G.D., editors, Guidebook to the geology of eastern Idaho: Pocatello, Idaho Museum of Natural History, p. 71-96.

Jarrett, R.D., and Malde, H.E., 1987, Paleodischarge of the late Pleistocene Bonneville flood, Snake River, Idaho, computed from new evidence: Geological Society of America Bulletin, v. 99, p. 127-134.

Jensen, M.E., and King, J.K., 1999, Geologic map of the Brigham City 7.5-minute quadrangle, Box Elder and Cache Counties, Utah: Utah Geological Survey Map 173, 46 p., scale 1:24,000.

Kariya, K.A., Roark, D.M., and Hanson, K.M., 1994, Hydrology of Cache Valley, Utah, and adjacent part of Idaho, with emphasis on simulation of ground-water flow: Utah Department of Natural Resources Technical Publication no. 108, 120 p.

Klauk, R.H., and Budding, K.E., 1984, Geothermal assessment of the lower Bear River drainage and northern east shore ground-water areas, Box Elder County, Utah: Utah Geological and Mineral Survey Report of Investigation No. 186, 64 p., scale 1:100,000.

Lander, J.F., and Cloud, W.K., 1964, United States earthquakes, 1962: Washington, D.C., U.S. Department of Commerce, U.S. Coast and Geodetic Survey, p. 15-21.

Landing, Ed, 1981, Conodont biostratigraphy and thermal color alteration indices of the upper St. Charles and lower Garden City Formations, Bear River Range, northern Utah and southeastern Idaho: U.S. Geological Survey Open-File Report 81-740, 22 p.

Leatham, W.B., 1985, Ages of the Fish Haven and lowermost Laketown Dolomites in the Bear River Range, Utah, *in* Kerns, G.J., and Kerns, R.L., Jr., editors, Orogenic patterns and stratigraphy of north-central Utah and southeastern Idaho: Utah Geological Association Publication 14, p. 29-38.

Link, P.K., 1982, Structural geology of the Oxford Peak and Malad Summit quadrangles, Bannock Range, southeastern Idaho, *in* Powers, R.B., editor, Geologic studies of the Cordilleran thrust belt: Rocky Mountain Association of Geologists, p. 851-858.

Machette, M.N., Personius, S.F., and Nelson, A.R., 1992, Paleoseismology of the Wasatch fault zone – A summary of recent investigations, interpretations, and conclusions: U.S. Geological Survey Professional Paper 1500A, p. A1-A71.

Maxey, G.B., 1958, Lower and Middle Cambrian stratigraphy in northern Utah and southeastern Idaho: Geological Society of America Bulletin, v. 69, p. 647-688.

Mayes, B.H., and Tripp, B.T., 1991, Zeolite minerals in Utah: Utah Geological Survey Open-File Report 210, 170 p., 1 plate, scale 1:750,000.

McCalpin, James, and Forman, S.L., 1991, Late Quaternary faulting and thermoluminescence dating of the East Cache fault zone, north-central Utah: Bulletin of the Seismological Society of America, v. 81, p. 139-161.

McCalpin, James, Robison, R.M., and Garr, J.D., 1992, Neotectonics of the Hansel Valley-Pocatello Valley corridor, northern Utah and southern Idaho: U.S. Geological Survey Professional Paper 1500-G, p. G1-G18.

McGreevy, L.J., 1972, Specific-conductance survey of the Malad River, Utah and Idaho: U.S. Geological Survey Professional Paper 800-C, p. C239-C242.

Morgan, S.K., 1988, Petrology of passive margin epeiric sea sediments – The Garden City Formation, north-central Utah: Logan, Utah State University, M.S. thesis, 156 p.

—1992, Geologic tours of northern Utah: Utah Geological Survey Miscellaneous Publication 92-1, 98 p.

Murphy, B.E., Beus, S.S., and Oviatt, C.G., 1985, Provisional geologic map of the Limekiln Knoll quadrangle, Box Elder County, Utah: Utah Geological and Mineral Survey Map 79, 9 p., scale 1:24,000.

Murphy, Peter, and Gwynn, J.W., 1979, Geothermal investigations at selected thermal systems of the northern Wasatch Front, Weber and Box Elder Counties, Utah: Utah Geological and Mineral Survey Report of Investigation No. 141, 50 p.

National Research Council, 1985, Liquifaction of soils during earthquakes: Washington, D.C., National Academy Press, 240 p.

Oaks, R.Q., Jr., 2000, Geologic history of Tertiary deposits between the lower Bear River drainage basin and the Cache Valley basin, north-central Utah, based on study of the Salt Lake Formation, the Collinston Conglomerate, and the Wasatch Formation, with applications to groundwater resources and fault-related geological hazards: Logan, Utah State University, Final report for the Bear River Conservancy District, Box Elder County, and Utah Division of Water Resources, 116 p.

Oaks, R.Q., Jr., and James, W.C., 1980, The Kinnikinic Quartzite (Middle Ordovician) in the type area, central Idaho, and a new reference section near Arco, Idaho: Brigham Young University Geology Studies, v. 27, part 2, p. 1-9.

Oaks, R.Q., Jr., James, W.C., Francis, G.G., and Schulingkamp, W.J., II, 1977, Summary of Middle Ordovician stratigraphy and tectonics, northern Utah, southern and central Idaho, *in* Heisey, E.L., Lawson, D.E., Norwood, E.R., Wach, P.H., and Hale, L.A., editors, Rocky Mountain thrust belt geology and resources: Wyoming Geological Association, Twenty-ninth Annual Field Conference Guidebook, p. 101-118.

Oaks, R.Q., Jr., and Kendrick, D.R., 1992, Seismic reflection suggests poor petroleum prospects west of 111°30′W., Bear River Range, north-central Utah [abs.]: Geological Society of America Abstracts with Programs, v. 24, no. 6, p. 56.

Oaks, R.Q., Jr., and Runnels, T.R., 1992, The Wasatch Formation in the central Bear River Range, northern Utah: Utah Geological Survey Contract Report 92-8, 79 p.

Obermeier, S.F., 1996, Using liquefaction-induced features for paleoseismic analysis, *in* McCalpin, J.P., editor, Paleoseismology: San Diego, Academic Press, p. 331-396.

O'Conner, J.E., 1993, Hydrology, hydraulics, and geomorphology of the Bonneville flood: Geological Society of America Special Paper 274, 83 p.

Oriel, S.S., and Platt, L.B., 1968, Reconnaissance map of the Preston 30′ x 60′ quadrangle, southeastern Idaho: U.S. Geological Survey Open-File Report 68-205, scale 1:62,500.

—1980, Geologic map of the Preston 1° x 2° quadrangle, southeastern Idaho and western Wyoming: U.S. Geological Survey Miscellaneous Investigations Series Map I-1127, scale 1:250,000.

Oviatt, C.G., 1986a, Geologic map of the Cutler Dam quadrangle, Box Elder and Cache Counties, Utah: Utah Geological and Mineral Survey Map 91, scale 1:24,000, 7 p.

—1986b, Geologic map of the Honeyville quadrangle, Box Elder and Cache Counties, Utah: Utah Geological and Mineral Survey Map 88, scale 1:24,000, 13 p.

Oviatt, C.G., and Currey, D.R., 1987, Pre-Bonneville Quaternary lakes in the Bonneville basin, Utah, *in* Kopp, R.S., and Cohenour, R.E., editors, Cenozoic geology of western Utah--Sites for precious metal and hydrocarbon accumulations: Utah Geological Association Publication 16, p. 257-263.

Oviatt, C.G., Currey, D.R., and Sack, Dorothy, 1992, Radiocarbon chronology of Lake Bonneville, eastern Great Basin, USA: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 99, p. 225-241.

Perkins, M.E., Brown, F.H., Nash, W.P., McIntosh, William, and Williams, S.K., 1998, Sequence, age, and source of silicic fallout tuffs in middle to late Miocene basins of the northern Basin and Range Province: Geological Society of America Bulletin, v. 110, no. 3, p. 344-360.

Personius, S.F., 1990, Surficial geologic map of the Brigham City segment and adjacent parts of the Weber and Collinston segments of the Wasatch fault zone, Box Elder and Weber Counties, Utah: U.S. Geological Survey Miscellaneous Investigations Series Map I-1979, scale 1:50,000.

Peterson, D.L., 1974, Bouguer gravity map of part of the northern Lake Bonneville basin, Utah and Idaho: U.S. Geological Survey Miscellaneous Field Studies Map MF-627, scale 1:250,000.

Peterson, D.L., and Oriel, S.S., 1970, Gravity anomalies in Cache Valley, Cache and Box Elder Counties, Utah, and Bannock and Franklin Counties, Idaho: U.S. Geological Survey Professional Paper 700-C, p. C114-C118.

Platt, L.B., 1977, Geologic map of the Ireland Springs-Samaria area, southeastern Idaho and northern Utah: U.S. Geological Survey Miscellaneous Field Studies Map MF-890, scale 1:48,000.

Prammani, Prapath, 1957, Geology of the east-central part of the Malad Range, Idaho: Logan, Utah State University, M.S. thesis, 60 p.

Rember, W.C., and Bennett, E.H., 1979, Geologic map of the Pocatello 1° x 2° quadrangle, Idaho: Idaho Bureau of Mines and Geology Geologic Map Series, scale 1:250,000.

Rodgers, D.W., and Janecke, S.U., 1992, Tertiary paleogeographic maps of the western Idaho-Wyoming-Montana thrust belt, *in* Link, P.K., Kuntz, M.A., and Platt, L.B., editors, Regional geology of eastern Idaho and western Wyoming: Geological Society of America Memoir 179, p. 83-94.

Ross, R.J., Jr., 1951, Stratigraphy of the Garden City Formation in northeastern Utah and its trilobite faunas: Yale Peabody Museum Bulletin 6, 161 p.

Schick International, Inc., 1983, High water-table investigations; Plymouth, Utah, unpublished consultant's report, 9 p.

Scott, W.E., McCoy, W.D., Shroba, R.R., and Rubin, Max, 1983, Reinterpretation of the exposed record of the last two cycles of Lake Bonneville, western United States: Quaternary Research, v. 20, p. 261-285.

Smith, K.A., 1997, Stratigraphy, geochronology, and tectonics of the Salt Lake Formation (Tertiary) of southern Cache Valley, Utah: Logan, Utah State University, M.S. thesis, 245 p.

Smith, R.B., and Arabasz, W.J., 1991, Seismicity of the Intermountain Seismic Belt, *in* Slemmons, D.B., Engdahl, E.R., Zoback, M.D., and Blackwell, D.D., editors, Neotectonics of North America: Boulder, Colorado, Geological Society of America Decade Map Volume 1, p. 185-228.

Smith, R.B., and Bruhn, R.L., 1984, Intraplate extension tectonics of the eastern Basin and Range – Inferences on structural style from seismic reflection data, regional tectonics and thermal-mechanical models of brittle-ductile deformation: Journal of Geophysical Research, v. 89, p. 5733-5762.

Solomon, B.J., 1999, Surficial geologic map of the West Cache fault zone and nearby faults, Box Elder and Cache Counties, Utah: Utah Geological Survey Map 172, 20 p., scale 1:50,000.

Sprinkel, D.A., 1979, Apparent reverse movement on previous thrust faults along the eastern margin of the Basin and Range Province, north-central Utah, *in* Newman, G.W., and Goode, H.D., editors, Basin and Range symposium and Great Basin field conference: Rocky Mountain Association of Geologists, p. 135-144.

Sprinkel, D.A., and Solomon, B.J., 1990, Radon hazards in Utah: Utah Geological Survey Circular 81, 24 p.

Stokes, W.L., 1986, Geology of Utah: Utah Museum of Natural History and Utah Geological and Mineral Survey, 307 p.

Sullivan, J.T., Nelson, A.R., LaForge, R.C., Wood, C.K., and Hansen, R.A., 1988, Central Utah regional seismotectonic study for USBR dams in the Wasatch Mountains: Denver, U.S. Bureau of Reclamation Seismotectonic Report 88-5, 269 p.

Swenson, L.M., 1997, Structural geology and stratigraphy of the Tertiary Salt Lake Formation at the south end of Clarkston Mountain: Logan, Utah State University, senior honors thesis, 22 p., scale 1:24,000.

Taylor, M.E., Landing, Ed, and Gillett, S.L., 1981, The Cambrian-Ordovician transition in the Bear River Range, Utah-Idaho – A preliminary evaluation, *in* Taylor, M.E., editor, Short papers for the Second International Symposium on the Cambrian System: U.S. Geological Survey Open-File Report 81-743, p. 222-227.

Utah Division of Comprehensive Emergency Management, 1981, History of Utah floods, 1847-1981: Utah Division of Comprehensive Emergency Management, Floodplain Management Status Report, variously paginated.

Utah State Department of Highways, 1965, Materials inventory, Box Elder County: Utah State Department of Highways, 17 p., 7 maps, scale approximately 1:200,000.

—1967, Materials inventory, Cache and Rich Counties: Utah Department of Highways, Materials and Research Division, Materials Inventory Section, 19 p., scale approximately 1:160,000.

Van Dorston, P.L., 1970, Environmental analysis of Swan Peak Formation in Bear River Range, north-central Utah and southeastern Idaho: American Association of Petroleum Geologists Bulletin, v. 54, no. 7, p. 1,140-1,154.

Westaway, R., and Smith, R.B., 1989, Source parameters of the Cache Valley (Logan), Utah, earthquake of 30 August, 1962: Bulletin of the Seismological Society of America, v. 79, p. 1410-1425.

Wieczorek, G.F., Ellen, S.D., Lips, E.W., Cannon, S.H., and Short, D.N., 1983, Potential for debris flows and debris floods along the Wasatch Front between Salt Lake City and Willard, Utah, and measures for their mitigation: U.S. Geological Survey Open-File Report 83-635, 45 p.

Wieczorek, G.F., Lips, E.W., and Ellen, S.D., 1989, Debris flows and hyperconcentrated floods along the Wasatch Front, 1983 and 1984: Bulletin of the Association of Engineering Geologists, v. 26, no. 2, p. 191-208.

Williams, J.S., 1948, Geology of the Paleozoic rocks, Logan [30'] quadrangle, Utah: Geological Society of America Bulletin, v. 59, no. 11, p. 1121-1164, scale 1:125,000.

—1958, Geologic atlas of Utah, Cache County: Utah Geological and Mineralogical Survey Bulletin 64, 98 p., scale 1:125,000.

—1964, The age of the Salt Lake Group in Cache Valley, Utah-Idaho: Utah Academy of Arts, Letters, and Sciences Proceedings, v. 41, part II, p. 269-277.

Woolley, R.R., 1946, Cloudburst floods in Utah, 1850-1938: U.S. Geological Survey Water Supply Paper 994, 128 p.

Yonkee, W.A., 1992, Basement-cover relations, Sevier orogenic belt, northern Utah: Geological Society of America Bulletin, v. 104, p. 280-302.

—1997, Kinematics and mechanics of the Willard thrust sheet, central part of the Sevier orogenic wedge, north-central Utah, *in* Link, P.K., and Kowallis, B.J., editors, Proterozoic to Recent stratigraphy, tectonics, and volcanology, Utah, Nevada, southern Idaho and central Mexico: Brigham Young University Geology Studies, v. 42, Part I, p. 341-354.

Youd, T.L., 1978, Major cause of earthquake damage is ground failure: Civil Engineering, v. 48, no. 4, p. 47-51.

—1984, Liquefaction effects – Liquefaction and associated ground failure, *in* Williams, M.E., compiler, Proceedings of the geologic and hydrologic hazards training program: U.S. Geological Survey Open-File Report 84-760, p. 210-232.

Youngs, R.R., Swan, E.H., Power, M.S., Schwartz, D.P., and Green, R.K., 1987, Probabalistic analysis of earthquake ground shaking hazard along Wasatch Front, Utah, *in* Gori, P.L., and Hayes, W.H., editors, Assessment of regional earthquake hazards and risk along the Wasatch Front, Utah: U.S. Geological Survey Open-File Report 87-585, paper M, 110 p.

Zoback, M.L., 1983, Structure and Cenozoic tectonism along the Wasatch fault zone, Utah, *in* Miller, D.M., Todd, V.R., and Howard, K.A., editors, Tectonic and stratigraphic studies of the eastern Great Basin: Geological Society of America Memoir 157, p. 3-27.