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Geology of the Republic Area and Portions of the Okanogan Highlands, WA

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The Tobacco Root Geological Society, Inc.

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Conference Organizers, 2014 Field Conference:

Skye Cooley, Geologist, Cirque Geoscience & GIS Consultants
Heather Vick, Hydrogeologist 4, Washington State Department of Ecology

Editor: Marie M. Garsjo

Layout: Susan Barth, MBMG

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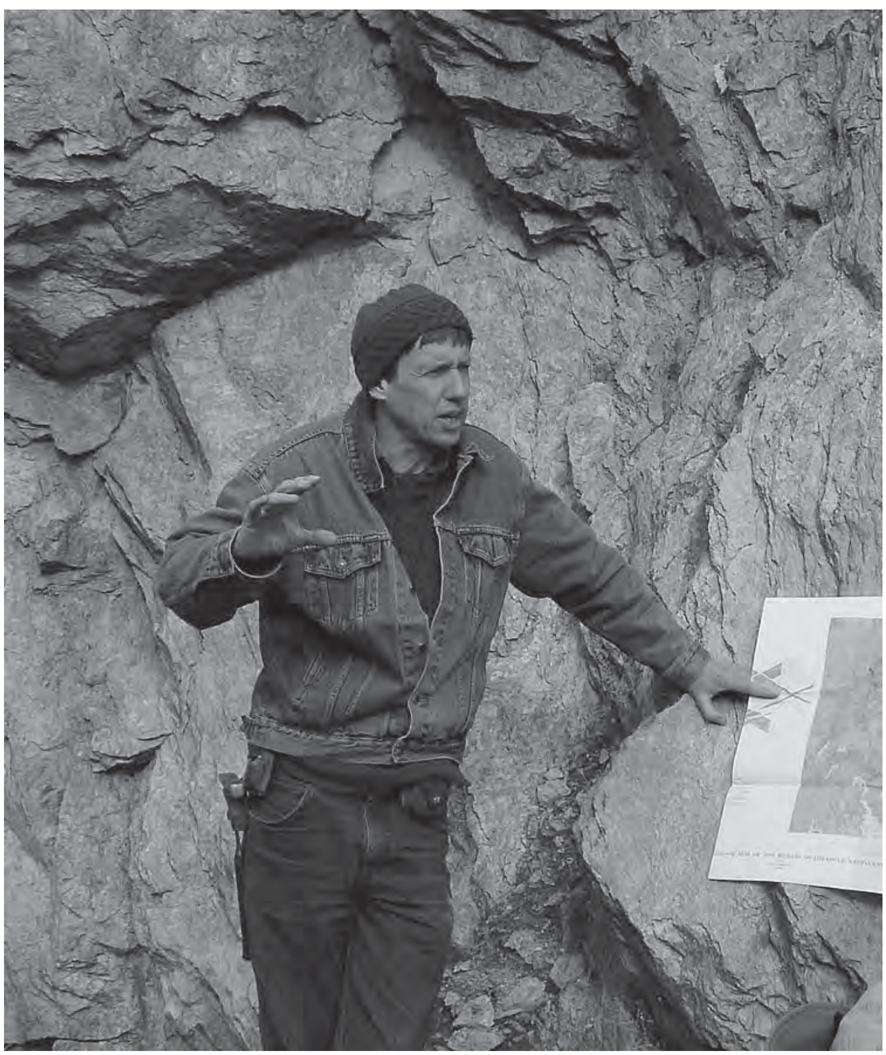
2014 HAMMER AWARD: REED LEWIS

A native of Moscow, Idaho, Reed Lewis developed an interest in rocks at a young age that set him on a career path in geology. Reed completed his undergraduate geology degree at the University of Idaho in Moscow in 1980. He continued his geology education by attending graduate school at the University of Washington, where he completed a Masters degree in 1984 and then went on to Oregon State University to complete a PhD in 1990. Following graduate school, Reed spent four years mapping the geology of north-central Idaho with the Idaho Geological Survey.

Seeking a change of scenery, culture, and geology, Reed spent three years exploring for gold deposits in northern Saudi Arabia as an employee of the U.S. Geological Survey from 1992 to 1995. Reed returned to the U.S. and worked for two years with the Montana Bureau of Mines and Geology in Butte, where he mapped in the Phillipsburg area and published the geologic map of the Butte 1° x 2° quadrangle and the Montana portion of the Missoula 1° x 2° quadrangle, both in 1998. Reed returned to his hometown in 1998 to work for the Idaho Geologic Survey. Reed has worked throughout northern and central Idaho, specializing in Precambrian rocks, the Idaho Batholith, and accreted terranes. During his tenure with the Idaho Geological Survey, Reed has been an author on over 120 publications—mainly geologic maps. A crowning achievement came in 2012 with the publication of the new geologic map of Idaho, a decade-long effort on which Reed is the lead author. Reed is currently serving as the interim director of the Idaho Geological Survey and the Idaho State Geologist.



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- Leif Anderson (TRGS Scholarship), University of Colorado, Absolute Chronology of Pre-Illinoisan glacial advances in the Rocky Mountains
- Lori Babcock (TRGS Scholarship), Montana State University, Structurally controlled diagenesis of upper Devonian (Frasnian Stage) dolomite: southwest and north central Montana and southern Alberta
- Jack Borski (TRGS Scholarship), Montana State University, Depositional Setting, Sequence Stratigraphic Analysis, and Diagenetic Evolution of the Devonian -Mississippian Transition in Western Montana
- Peter Christophersen (TRGS Scholarship), University of Montana, Deformational and thermal history of the Bitterroot mylonite constrained by electron backscatter diffraction (EBSD) and Ti-in-quartz (TitaniQ) thermometry
- Anna Downey (TRGS Scholarship), Kansas State University, Quaternary mafic volcanism in the upper Wind River Basin, WY
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- Julian Stahl (TRGS Scholarship), Montana State University, Fracture analysis of circum-Bighorn Basin anticlines, Wyoming-Montana

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Tikoff, B., and Teyssier, C., 1992, Crustal-scale, en-echelon P-shear tensional bridges: a possible solution to the batholithic room problem: Geology, v.20, p. 927-930.

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FUNDAMENTALS OF THE GEOLOGY OF NORTH-CENTRAL WASHINGTON

Eric Cheney

Professor Emeritus, University of Washington

ABSTRACT

The geology of north-central Washington includes portions of pre-Jurassic North America (Laurentia), the Quesnel terrane, and Cenozoic cover sequences. Eocene antiformal metamorphic core complexes dominate the topography and the structure of the region. The Sanpoil syncline between the Okanogan and Kettle metamorphic core complexes is too often called the "Republic graben." The volcanic and sedimentary strata of regional Eocene unconformity-bounded sequences are preserved in the Sanpoil syncline and other structural lows in the region. The largest epithermal gold producer was the Knob Hill mine (1937 to 1995), which is intimately tied to the history of Republic.

INTRODUCTION

The annual field conference of the Tobacco Root Geological Society was last held here in Republic, Washington in 1980. As a participant of that conference (Cheney et al., 1982), I have survived to review the geology of North Central Washington (NCW) for the 2014 conference. For the sake of the field trips of the 2014 conference, this review is primarily descriptive and does not dwell on plate tectonic and other orogenic processes.

NCW is the area between the Okanogan River valley on the west and the Columbia River valley on the east (fig. 1). Physiographically, this area is the western portion of the Okanogan Highlands, the highest summits of which are less than 2,500 m in elevation. NCW also is in the northeastern portion of what might be called the Ponderosa crescent of Washington, which extends along the eastern slope of the Cascade Range and most of the Okanogan Highlands.

NCW is geologically diverse, containing representatives of the three major geologic components of Washington. These are Laurentia (ancestral) pre-Jurassic North America, terranes that accreted to Laurentia from the middle Jurassic through the Paleocene, and Eocene to Holocene cover sequences that unconformably overlie both Laurentia and the terranes (fig. 2). Washington is characterized by westerly and northerly trending folds formed in the Eocene and Neogene, but the paucity of Eocene and younger rocks in NCW makes the westerly folds difficult to recognize.

Specifically, this review summarizes the characteristics of the metamorphic core complexes (which expose the local crystalline basement), Laurentia, the Quesnel terrane (which accreted to Laurentia in the middle Jurassic), and cover sequences bounded by Cenozoic unconformities. Figure 2 shows these entities, although for simplicity and clarity granitic intrusions and many mapped faults are not shown. In figure 2, Laurentia is comprised of map units B, SW, SA, and Z. The Quesnel terrane is unit Q. Map unit T represents the Cenozoic cover sequences, which in this area are Eocene and to a much lesser extent the Miocene Columbia River Basalt Group. The article also notes the mineral deposits of NCW.

The 1:250,000 map of Stoffel et al. (1991) shows the major geologic units of the northeastern quarter of Washington. Of course, refinement of some units continues, and geologic interpretations of them are still evolving. NCW was glaciated in the Pleistocene (Stoffel et al., 1991). Although they extensively cover bedrock, glacial and post-glacial sediments are ignored herein.

METAMORPHIC CORE COMPLEXES

The topography and geology of the Okanogan Highlands are dominated by northward trending antiforms cored by metamorphic and batholithic rocks. These are crustal-scale boudins bounded by low-angle normal faults, called detachment faults. Most of the rocks in the fault zones are mylonitic (lineated and well foliated), indicating that deformation was ductile. Such antiforms bounded by mylonites are called metamorphic core complexes (MCCs), and they record significant extension of the crust. The MCCs in the northwestern United States and adjacent British Co-

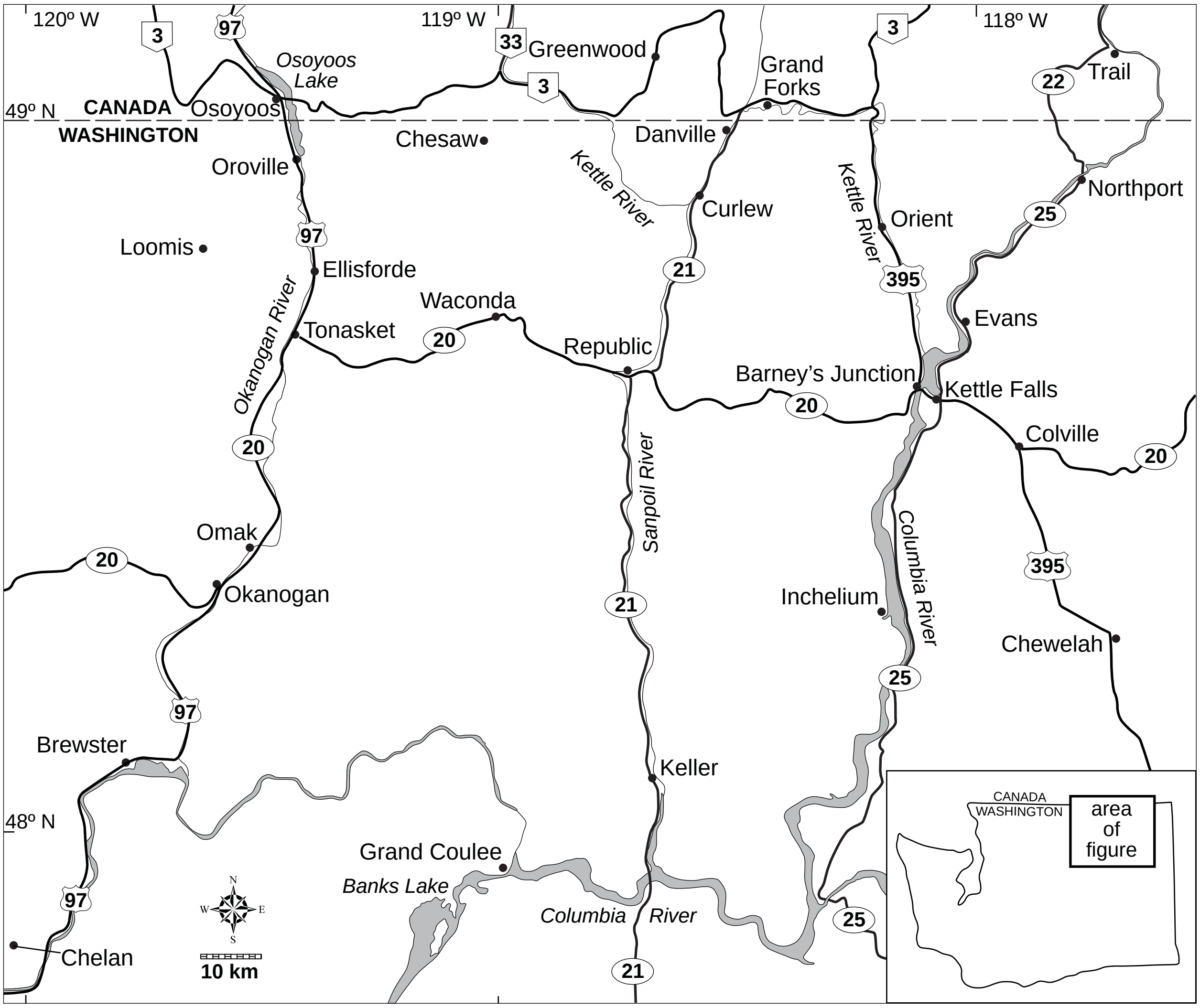

Figure 1. Index map. Figure 1 is in the southwest corner of Figure 2.

lumbia (fig. 2) formed in the middle Eocene; whereas, the MCCs in the Basin and Range province south of the Great Falls tectonic zone of Montana and Idaho formed after the middle Miocene. The antiformal MCCs in the Okanogan Highlands disrupt the continuity of Eocene and older rocks (and much geologic thinking about them).

Rocks in the MCCs of the Okanogan Highlands range from 1.7 Ga pelitic schists in the Canadian portion of the Kettle MCC to Mesozoic orthogneisses and Eocene batholiths. Some of the Mesozoic orthogneisses and Eocene batholiths have cm-scale feldspar megacrysts (phenocrysts), which in mylonitic zones have become porphyroclasts that are quite photogenic. Virtually all of the metamorphic rocks of the MCCs are in amphibolite facies. In contrast, the pre-Tertiary rocks in the hanging walls of the detachment faults are in greenschist facies, and the Tertiary rocks are unmetmorphosed.

The MCCs of NCW include a widespread succession of metamorphic rocks. These were first described by Parker and Calkins (1964) as the Teans Mary Creek sequence (TMC) in the northeastern thumb of the Okanogan MCC. The TMC occurs throughout the Kettle MCC and in the northeastern part of the Okanogan MCC (Cheney, 1980, table 2). The distinctive aspect of the TMC is that it has two thick megacrystic orthogneisses, two thick quartzites, and intercalated sillimanite-bearing biotitic paragneisses (table 1). On a regional scale, the TMC is simple layer cake geology, which is nearly horizontal along the antiformal crest of the Kettle MCC. The structurally low biotitic paragneisses in the TMC have biotitic pegmatites that are uranium-bearing, and larger bodies of biotitic alaskite to aplite that contain only sub-economic amounts of uranium (table 1); both were prospected during the uranium booms of the 1950s and 1970s.

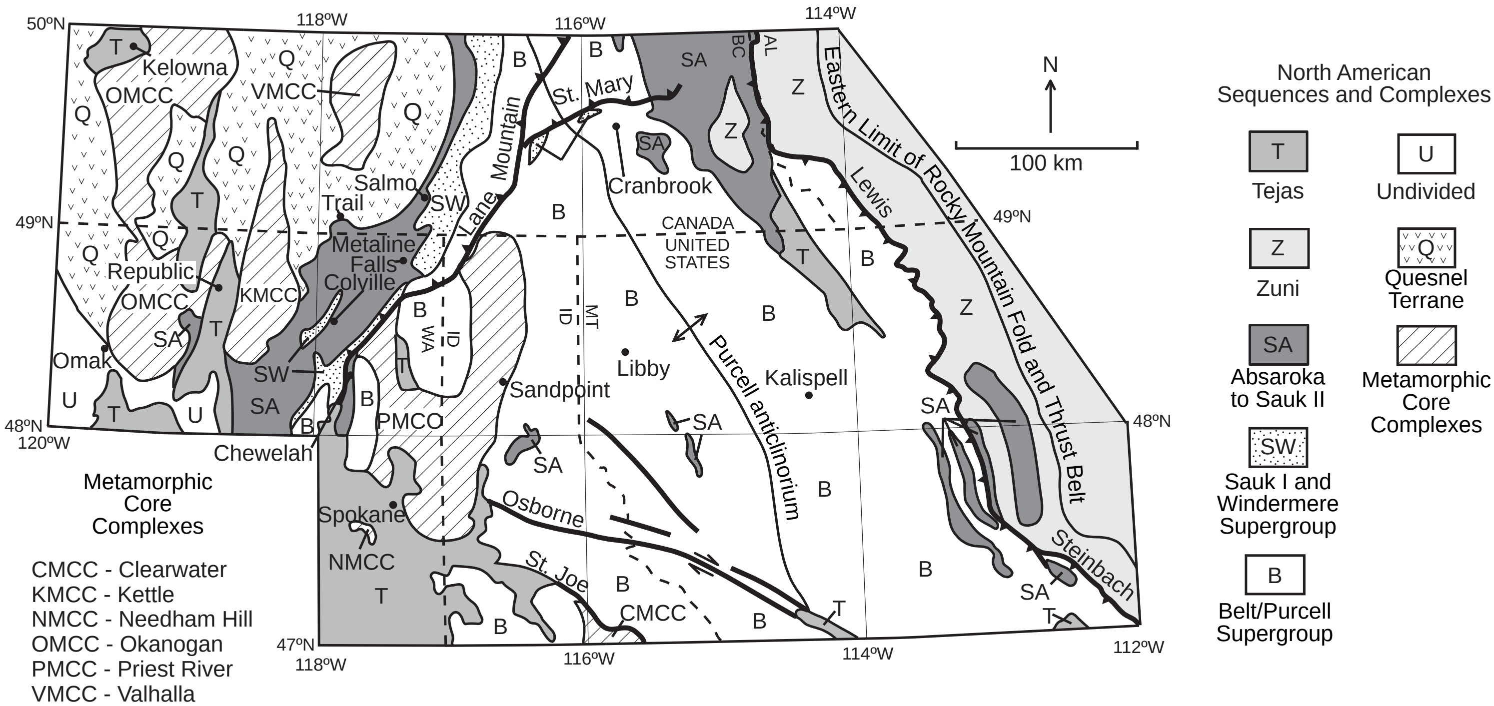

Figure 2. The regional geological setting of north-central Washington. For clarity, all granitic intrusions and many mapped faults are omitted from the figure. North American sequences indigenous to Laurentia are Mesoprotrozoic to Cretaceous (B, SW, SA, and Z). Q is the Quesnel terrane. T represents post-accretionary Eocene to Miocene sequences.

The TMC quartzites are so thick that their protoliths most likely were continental, that is, Laurentian. The structurally lower feldspathic quartzite might be the feldspathic Neoproterozoic Three Sisters Formation of the Windermere Supergroup. The upper, more pure quartzite might be either the Neoproterozoic to early Cambrian Addy Quartzite or the middle Cambrian quartzite of Chewelah (which is equivalent to the Flathead Quartzite of the craton). The upper quartzite formed the famous and historic Indian fishing site at Kettle Falls on the Columbia River near Barney's Junction; the falls, the migratory salmon runs, and Indian fishing were obliterated in 1942 when FDR Lake formed behind Grand Coulee dam.

Since the pioneering work of Snook (1965), it has become clear that most of the detachment faults in the Okanogan Highlands are kilometer-thick zones of mylonite (ductile deformation) with a thin zone of cataclastic rock (brittle deformation) that has been

Table 1. The Tenas Mary Creek succession in the Kettle and Okanogan metamorphic core complexes (after Cheney, 1980, tables 1 and 2 and text).

Major Lithologic Unit	Thickness of Major Unit	Minor Lithologies	Thickness of Marble	Uranium Pegmatite
Weakly micaceous quartzite, only on eastern margin	>300 m			none
Fine-grained biotite schist, only on eastern margin	600 m	Quartzite, marble	< 30 m	none
Amphibolite, only on eastern margin	200 m			none
Megacrystic tonalitic orthogneiss , minor feldspars megacrysts mostly < 1 cm	800 m			none
Biotite schist and gneiss	< 300 m	Quartzite, marble	< 15 m	few
Feldspathic quartzite, 5 to 10 % of 1 to 5mm feldspar	>650 m	Biotite gneiss, marble	< 60 m	none
Biotite schist and gneiss	150 m	Quartzite, marble	< 60 m	common
Megacrystic biotitic orthogneiss, cm-scale orthoclase , irregular pegmatitic patches < 1 m across	>850 m	none		none
Biotite schist and gneiss	>700 m	Quartzite, marble	< 60 m	very common

Notes: Progressively structurally higher major units are progressively higher in the table. All biotitic units are sillimanite-bearing. The last column refers to uranium-bearing biotitic pegmatites and to uranium-bearing bodies of biotitic alaskite to aplite.

chloritically altered, and superimposed along the upper contact of the mylonite. The extreme opposite to the chloritic cataclastic rock are the bands of ultramylonite derived from the tonalitic Tonasket Gneiss on the western limb of the Okanogan MCC; megascopically, ultramylonites look like chert. Chloritic zones vary from widely spaced fractures to breccia; they commonly are less than several tens of meters thick. Of course, chloritic zones are absent in compositionally inhospitable rock, such as quartzite. The chloritic zones in MCCs are so thin that they are commonly obscured by overburden, lakes, and forests.

The style of detachment faults varies. Low-angle detachment faults, such as the Okanogan Valley fault on the western limb of the Okanogan MCC and the Kettle River fault on the eastern limb of the Kettle MCC, have mylonite zones that are up to 4 km thick. On the steeper, opposite limbs of these two MCCs, such as the Bacon Creek fault on the eastern margin of the Okanogan MCC, the mylonite is absent, and only a chloritic zone exists.

Slip-on detachment faulting was not necessarily oriented directly down-dip. An excellent example is the western part of the Okanogan MCC. The lineations in the mylonitic Tonasket Gneiss strike northwestward, as do antiforms defined by foliations in the gneiss (Snook, 1965, plate 1; Kruckenberg et al., 2008, fig. 2a). The Okanogan Valley detachment fault has a sinuous trace along the western side of the MCC, with northwesterly bulges to the west (Stoffel et al., 1991; Cheney et al., 1994, fig. 2). These bulges are on strike with the antiforms defined by foliations, implying that movement on the fault was to the northwest. The antiforms indicate that overall, the detachment fault has the geometry of an obliquely inclined, old-fashioned washboard; so movement directly down dip would have been difficult.

The radiometric dating by Kruckenberg et al., (2008) and their predecessors documents the following history of the Okanogan MCC. The original age of the tonalitic Tonasket Gneiss in the western portion of the MCC is 100 to 70 Ma. At structurally deeper levels the gneiss underwent partial melting from 61 to 47 Ma to form migmatites; the resultant crustal instability may have caused mylonitization at upper levels. Granitic intrusions range from 54 to 47 Ma and locally are mylonitic. Cooling of the gneisses and the granitic plutons lasted until 47 Ma.

The Sanpoil syncline is between the antiformal Okanogan and Kettle MCCs. The synclinal pattern of the three Eocene formations is accentuated if the unconformably underlying Mesozoic and Paleozoic rocks of the Quesnel terrane are lumped into a single basal unit (Cheney and Rasmussen, 1996, fig. 2). The Eocene and Quesnellian rocks of the Sanpoil syncline are bounded by the detachment faults of the MCCs. The Sanpoil syncline is popularly known as the Republic graben. The structural low between the main body of the Okanogan MCC and its northeastern thumb is popularly known as the Toroda Creek half graben.

LAURENTIA

In the northwestern US and adjacent Canada, Laurentia consists of crystalline basement older than 1.71 Ga and several overlying sequences that are bounded by unconformities. The basement does not crop out west of Idaho, except in part of the Kettle MCC exposed in British Columbia. An intriguing question is how far to the west does the Laurentian crystalline basement extend in the subsurface. The initial strontium 0.706/0.704 line, east of which granitic plutons record their passage through Precambrian crystalline basement, appears to extend into the central Cascade Range (Armstrong et al., 1977, fig. 10). Worldwide, alkalic igneous complexes, no matter what their age, are restricted to areas of Proterozoic or Archean basement; the Jurassic Shasket Creek alkalic complex is near Danville (Parker and Calkins, 1964; Cheney, 2014, fig. 2), and the Cretaceous Kruger complex is 10 km west of Osoyoos (Buddington, 1990). An exotic body of two-mica, garnet schist in the Chesaw thrust in Quesnellian rocks between Curlew and Danville (Cheney, 2014, Stop 7) implies that crystalline crust is somewhere below. As noted above, TMC quartzites in the Kettle and Okanogan MCCs probably are Laurentian. All of these hint that in the subsurface, the Laurentian crystalline basement extends west of the Okanogan valley, and, therefore, underlies virtually all of the Quesnel terrane exposed in Washington.

Westward from the axial trace of the Purcell anticlinorium in Montana and British Columbia (fig. 2), the Laurentian cover sequences vary from cratonic, to miogeoclinal, to distal, and from Mesoproterozoic to Mississippian. The Phanerozoic sequences in Washington are the same as those recognized by

Sloss (1988) on the craton to the east, but they are thicker and more conformable. The distal Laurentian rocks in NCW (unit SA of fig. 2) are the Cambrian to Silurian Covada Group and the Devonian Bradeen Hill assemblage (Smith and Geherls, 1992); both are predominantly pelitic with minor greenstone and some limestone. Because these rocks are badly deformed, poorly exposed, and contain few fossils, they are difficult to map. They are informally known as the black shale belt, and less affectionately as the "black crap". They extend as far southwest as the Sanpoil syncline (Smith, 1991), where they are truncated by the detachment fault on the southeastern side of the Okanogan MCC (fig. 2).

The western margin of the Laurentian sequences is a northwestward to westward verging, middle Jurassic thrust belt in the black crap and other rocks; this is the Pend Oreille River fold and thrust belt (Cheney, 2010). The basal of eight thrusts in this belt is the Waneta thrust in distal rocks in British Columbia. This thrust strikes southwestward across the International Border and is truncated by the Kettle River detachment fault on the eastern side of the Kettle MCC. In British Columbia the Waneta thrust dips 30° to 35° to the southeast, thereby placing Laurentian strata over the Quesnel terrene (Fyles and Hewlett, 1959). That is, the Quesnel terrane was inserted between the Laurentian basement and its cover sequences. This insertion, or "delamination" of crustal rocks, likely generated the Pend Oreille River fold and thrust belt.

As noted before, the quartzites of the TMC succession in the Kettle and Okanogan MCCs probably are Laurentian. If so, Laurentian rocks were deformed in the Antler, the mid-Jurassic, or other orogenies and were incorporated into the basement that became MCCs in the Eocene.

QUESNEL TERRANE

The Quesnellian rocks consist of three packages. The first is the ophiolitic Knob Hill Group and the mostly pelitic Attwood Group (with minor limestone); these are Carboniferous to Permian. The Knob Hill Group consists of ultramafic rocks (mostly serpentinized), metadiorite, fine-grained metabasaltic greenstone, and quartzite (metachert). Unconformable on both the Knob Hill Group and the Attwood Group is the Triassic Brooklyn Formation (sharpstone conglomerate grading upward into clastic limestone and argillite), which, in turn, is unconformably overlain by the Jurassic Rossland Group.

Regionally, the Rossland Group consists of pelitic rocks with a middle volcanic unit, the Elise Formation, which is predominantly mafic. The distinctive lithology of the Elise Formation is augite-bearing greenstone, some of which is coarsely fragmental (Höy and Dunne, 2001). Another distinctive unit in the Danville area is the Lexington quartz-feldspar porphyry, which appears to be intrusive (Fyles, 1990). Formations similar to the Attwood, Knob Hill, Brooklyn, and Elise occur west of the Okanogan MCC but were given different lithostratigraphic names; these names are without precedence (Cheney et al., 1994, fig. 3).

Felsic metavolcanic rocks appear to be underreported in the Quesnellian rocks. Such felsic rocks can be hosts for, and guides to, volcanogenic deposits of massive sulfide. Felsic metavolcanic rocks occur at the Morning Star, Lamefoot, and Overlook deposits (Cheney, 2014, figs. 2 and 5). Similar rocks occur at several other localities (Cheney et al., 1994, table 2). However, a word of caution is necessary: the Lexington quartz-feldspar porphyry is definitely an intrusion (J. T. Fyles, written communication, 1992), but it is so intensely foliated beneath the Chesaw fault (Church, 1986), which is discussed next, that it resembles metavolcanic schist.

The Chesaw, or No. 7 Fault (Cheney et al., 1994; Cheney and Rasmussen, 1996), is the major pre-Cenozoic fault in the area; it is at the base of the Knob Hill Group. Regionally, rocks of the Knob Hill Group (including serpentinite) are in the hanging wall of the fault. The Attwood Group is below the fault. Although regionally the Triassic Brooklyn Formation and the Jurassic Elise Formation are both above and below the fault, at Danville they are below the fault and are exposed in a half window; the area of Knob Hill Group rocks east of the Kettle River at Danville is a klippe (Cheney, 2014, fig. 2).

In addition to serpentinite, the Chesaw fault is also marked by bodies of listwanite, most of which are less than several tens of meters long. Listwanite is carbonated ultramafic rock, resulting in the assemblage magnesite + dolomite + quartz, commonly with minor pyrite and the bright green chrome mica, fuchsite. The quartz commonly occurs as numerous veinlets and coarse-grained knots. Unweathered listwanite without quartz veinlets or knots is a soft, gray gneiss. Weathered listwanite is a distinctive rusty orange and also hosts orange lichens. Because of their distinctive color, fuchsite (which resembles malachite), and quartz veinlets or knots, listwanites in NCW commonly are pocked by old prospecting pits. Especially west of the Okanogan MCC, the Chesaw fault can be mapped by "connecting the dots" of listwanite.

The Chesaw fault is of regional extent. Including the younger intervening Okanogan MCC, the fault extends for 120 km from Omak, Washington, to Grand Forks, British Columbia (Cheney et al., 1994, fig. 2). Variably serpentinized ultramafic rocks near Rossland, British Columbia (Höy and Dunne, 2001) and just south of the International Border (Weaver, 1920) suggest that the fault is present 55 km east of Grand Forks (east of the intervening Kettle MCC).

The Lexington quartz porphyry and the Shasket Creek alkalic complex bracket the age of the Chesaw fault. The Lexington quartz porphyry, which is 199.4 + 1.4 Ma (Church, 1992), is deformed by the Chesaw fault along the International Border west of Danville (Church, 1986, 1992). The Shasket Creek alkalic complex, which is 163 + 0.4 Ma (Berger et al., 1991), intrudes the thrust southwest of Danville (Cheney et al., 1994; Cheney, 2014, fig. 2).

Folds in Quesnellian rocks are of at least two ages. The older ones tend to be tight to isoclinal, and they fold the Chesaw thrust (Cheney et al., 1994, fig. 4; Cheney, 2014, fig. 5). Some folds are (or were) recumbent (Cheney, 2014, fig. 6). Broad Eocene folds (the MCCs and the Sanpoil syncline) and subsequent erosion caused klippe and windows in the Chesaw thrust (Cheney et al., 1994, figs. 4 and 7; Cheney, 2014, fig. 2).

TERTIARY SEQUENCES BOUNDED BY UNCONFORMITIES

The Eocene triad of formations (Muessig, 1962) from the base up, is the arkosic O'Brien Creek Formation, the rhyodacitic Sanpoil Formation, and the tuffaceous, clastic, and rhyolitic Klondike Mountain Formation. Each of these is locally over 1 km thick and bounded by unconformities. Pearson and Obradovich (1977) realized that representatives of each of these formations occur across northeastern Washington and into southern British Columbia (the columns for Pend Oreille, Republic, and White Lake in fig. 3). Thus, these formations were not deposited in local syndepositional basins, such as the putative Republic "graben". Cheney (1994) extended these and the other Eocene unconformity-bounded sequence across Washington (fig. 3) and called them the Challis sequence.

The O'Brien Creek has some general characteristics. Although most of the formation is dark siltstone and shale, they weather easily and most outcrops are sandstone. Sandstones typically have mm- to cm-scale clasts of black argillite and felsic volcanic rocks. Some of the siltstones contain thin beds of lignite (Muessig, 1967). The sandstones are typically white weathering and chalky to somewhat friable in outcrop. Thus, the sandstones were regarded as tuffaceous (e.g., Muessig, 1967). However, these characteristics are more likely due to pervasive regional alteration of feldspars to zeolites (such as laumantite), as is the case in the Eocene arkosic Roslyn (Chumstick) Formation on the eastern flank of the Cascade Range (R. J. Stewart, 1998, personal communication).

The rhyodacitic Sanpoil Volcanics remain enigmatic. The formation is composed of flows, pyroclastic breccias, and lithic tuffs (Tschauder, 1989). Some local stratigraphy is recognizable around the mines of the Republic district. However, elsewhere the formation is generally without bedding or marker units; so its internal stratigraphy is largely unknown.

The Klondike Mountain Formation consists of three members (Muessig, 1967; Gaylord et al., 1996), which are the basal Tom Thumb Tuff Member, an arkosic middle member, and an upper member with rhyolitic flows. The Tom Thumb Tuff has very coarse volcaniclastic rocks at is base and grades up into water-lain tuff (the "lake beds" of the older literature). This tuff is renowned for its leaf, flower, fruit, seed, pollen, conifer cones, insect, and fish fossils (as shown by several authors in the June, 1996, issue of Washington Geology, v. 24, no. 2). In the subsurface the tuff is so carbon-rich and uncemented that drill core from exploration in the 1990s had to be sheathed in plastic lest it disaggregate.

Despite its lithostratigraphic assignment to the

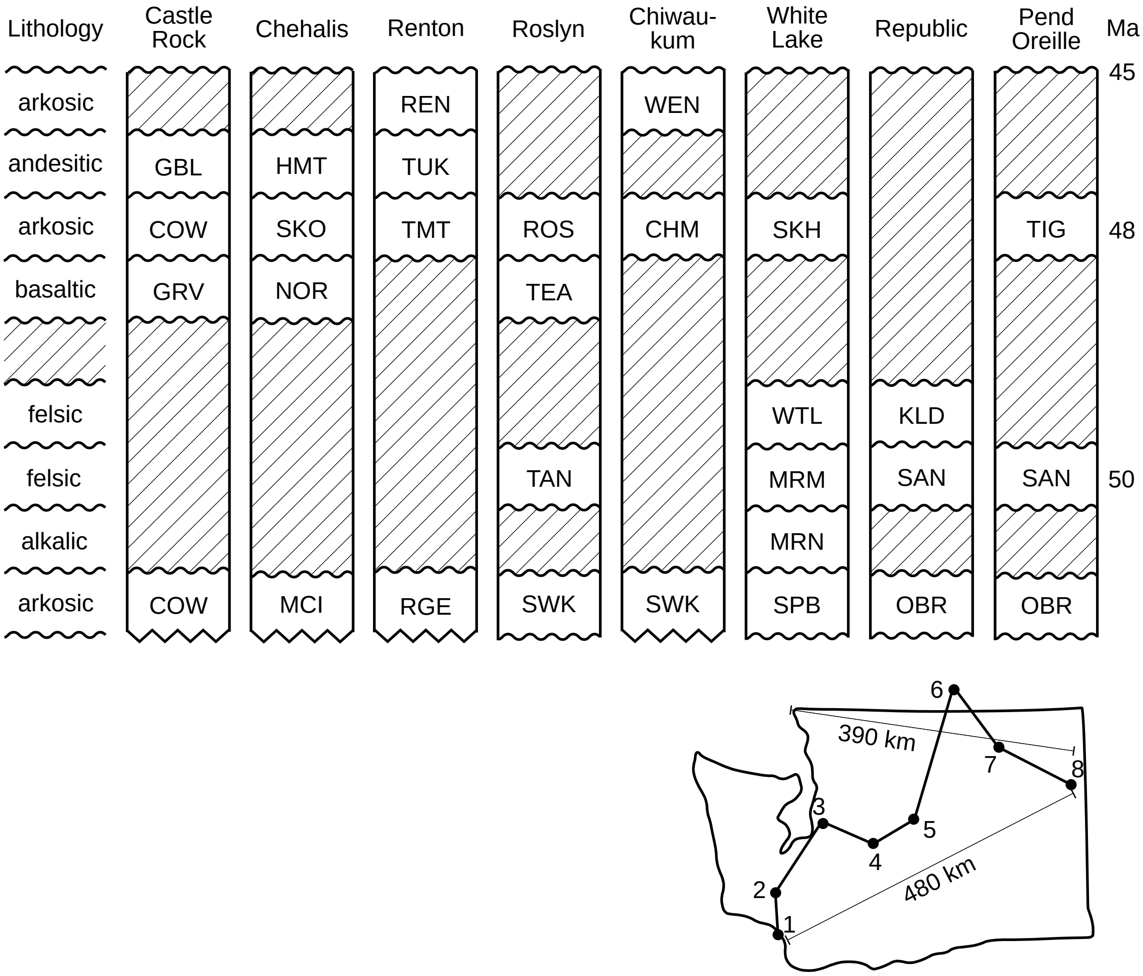

Figure 3. Inter-regional sequence stratigraphy of Eocene strata (after Cheney and Rasmusen, 1996, fig. 3). Note that the time scale is not linear. Wavy lines between formations are unconformities. Abbreviations for units: CHM, Chumstick Formation; COW, Cowlitz Formation; GBL, Goble Volcanics; GRV, Grays River volcanic rocks; HMT, Hatchet Mountain Formation; KLD, Klondike Mountain Formation; NOR, Northcraft Formation; MCI, McIntosh Formation; OBR, O'Brien Creek Formation; MRM, Marama Formation; MRN, Marron Formation; REN, Renton Formation; RGE, Raging River Formation; ROS, Roslyn Formation; SAN, Sanpoil Volcanics; SKH, Skaha Formation; SKO, Skookumchuck Formation; SLP, Silver Pass Volcanic Member of the Swauk Formation; SPB, Springbrook Formation; TAN, Taneum Formation; TEA, Teanaway Formation; TIG, Tiger Formation; TMT, Tiger Mountain Formation; TUK, Tukwila Formation; SWK, Swauk Formation; WEN, lower part of the Wenatchee Formation; WTL, White Lake Formation. The arrows indicate that some of the sequences extend beyond Washington.

Klondike Formation (Muessig, 1967), the Tom Thumb Tuff is compositionally similar to the rhyodacitic Sanpoil Volcanics, and should be assigned to the upper part of the formation. The epithermal gold mineralization at Republic occurs in the Sanpoil Volcanics and the tuff, but not in the unconformably overlying members of the Klondike Mountain Formation. Furthermore, the contact between the tuff and the Sanpoil Volcanics is concordant (Gaylord et al., 1996).

In the Toroda Creek "half graben", the middle arkosic member of the Klondike Mountain includes very large blocks of granitic rocks. These are rock-avalanche deposits from the northeastern thumb of the Okanogan MCC (Malte, 1995). For avalanches to have happened, one or more of the detachment faults bounding the northeastern thumb of the MCC must have breached the surface. Although the avalanche deposits effectively provide a stratigraphic age for the local detachment fault(s), other detachment faults in the Okanogan Highlands may have somewhat different ages.

The major Eocene structures are, of course, the antiformal MCCs and the Sanpoil syncline. The Eocene strata are also preserved elsewhere in synformal lows bounding the MCCs. Examples are First Thought Mountain near Orient east of the Kettle MCC, and areas from Omak to Ellisford along the western margin of the Okanogan MCC (Stoffel et al., 1991; Cheney et al., 1994, fig. 2). Vertical dips in the O'Brien Creek Formation near the Lamefoot mine (Cheney, 2014, fig. 5) might be due to listric faults along the western margin of the Kettle MCC.

The Columbia River Basalt Group (CRBG) is the predominant lithostratigraphic unit in the Miocene Walpapi sequence (Cheney, 1994). The main area of the CRBG is south of the Okanogan Highlands in southeastern Washington, eastern Oregon, and westernmost Idaho. However, remnants of the CRBG occur north of the Spokane and Columbia rivers along the southern margin of the Okanogan Highlands (Stoffel et al., 1991). Presumably, the CRBG once extended across much of what subsequently became the Okanogan Highlands.

ECONOMIC GEOLOGY

NCW has a rich and varied mining history, but most deposits have either been mined out or are now dormant for various reasons. Table 2 illustrates the diversity of the types of deposits in NCW, but this table certainly is not complete. Sand and gravel and common rock quarries are omitted from the table. Inclusion of deposits in the Greenwood district (Church, 1986) and the Rossland district several kilometers west of Trail in (Höy and Dunne, 2001) in adjacent British Columbia would generate more entries in the table. Providing references for each deposit in the table would unduly lengthen this paper, but Cheney (2014) described the Morning Star and Lamefoot deposits. The interested reader is encouraged to search the publications and databases of the Washington Division of Geology and Earth Resources and other online sources. Two particularly informative printed compendia are Hughes (1897), and Derkey et al., (1990).

The only active metal mine in Washington in early 2014 was the Buckhorn Mountain gold mine in Quesnellian rocks between Chesaw and Republic. The unmined Mount Tolman porphyry Cu-Mo at Keller (Lasmanis and Utterback, 1995) is the metallic deposit with the greatest potential in the entire State, but the Confederated Colville Indian Tribes, on whose reservation it occurs, continue to oppose its development.

By far the largest of the epithermal producers was the Knob Hill mine, which operated from 1937 to 1995. The following description is from Tschauder (1989). It is a low-sulfidation, adularia-sericite type of hot spring deposit. By 1989 it had produced 2.2 million tons of ore with an average grade of 0.55 oz/ton gold and 4.92 oz/ton silver. It and the adjoining Golden Promise deposit combined produced 2.5 million ounces of gold by 1995 (Lasmanis, 1996). The veins are very fine-grained to chalcedonic quartz that is vuggy, brecciated, and crustified, with mm-scale bands of very fine-grained sulfide and selenide minerals. Mineralization flared upward from discrete veins into zones of stockworks and into uppermost zones of breccia near the paleosurface; the upward increase in volume was accompanied by lower ore grades. Adjacent to mineralization the rocks are sericiticly altered. Around the sericitic zone is a much larger zone of propylitic alteration (chlorite, epidote, calcite, and pyrite), which makes the rocks green. Mineralization, including sinter, topped out at the unconformity which bounds the top of the Tom Thumb Member of the Klondike Mountain Formation. The age of mineralization, based on ^{40}Ar/^{39}Ar dating of adularia (Berger et al., 1991), is 50.1 $\pm$ 0.2 Ma, and this date also establishes the approximate age of the upper part of the Tom Thumb Tuff (fig. 3). The veins at both the Knob Hill mine and the K-2 mine near Curlew were dextral strike-slip systems (Tschauder, 1989; Chutas, 2000) and are above the Bacon Creek detachment fault.

CONCLUSIONS

The geology of NCW is diverse and complex, but rather poorly known, in part due to extensive Quaternary sediments and forest cover. Regional mapping at scales larger than 1:48,000 is still almost nonexistent. The TRGS last visited Republic in 1980. When TRGS reconvenes in Republic 1/3 of a century from now, our successors will have learned a lot if they have combined the many new tools in their kits with renewed regional geologic mapping.

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Table 2. Representative mineral deposits of north-central Washington and their nearby towns. Most of the towns are shown in figure 1.

Type of Deposit	Metamorphic Core Complexes	Laurentia	Quesnel Terrane	Eocene Sequences
Porphyry Cu			Lone Star, Danville Sherman, OmaK Silver Nail, Oroville	
Porpyhyy Mo-Cu, Mo			Aeneas, XXXX Mt.Tolman, Keller	
Skarn			Magnetic, Chesaw	
Intrusion related Au veins			Kabah Texas, Nighthawk Ruby, Nighthawk Wannacut Lake	
Low sulfidation Epithernal gold				Bodie, Toroda First Thought, Orient Graphite Ck., Toroda Knob Hill, Republic K2, Curlew Surprise, Republic
Au massive sulfide/oxide			Buckhorn Mtn., Chesaw? Lamefoot, Republic Overlook, Republic	
Au Banded Iron Formation	Lone Ranch Ck., Danville			
Volcanogenic massive sulfide			New World, Loomis Morning Star, Danville?	
Bedded Barite		Flagstaff Mtn., Northport		
Sedex Zn-Pb-Ag		Bonanza, Evans		
Podiform Cr			Haeberle, Omak Chopaka Lake, Loomis	
Mississippi Valley Sb?			Lucky Knock, Ellisford	
Rössing U	Mt.Leona, Malo			
limestone			Spectacle Lk, Tonasket Several quarries, Evans	
High-Ca marble	Wauconda			
Decorative quartzite	Kifer, Barney's Junction			
Dolomite		quarries, Northport		

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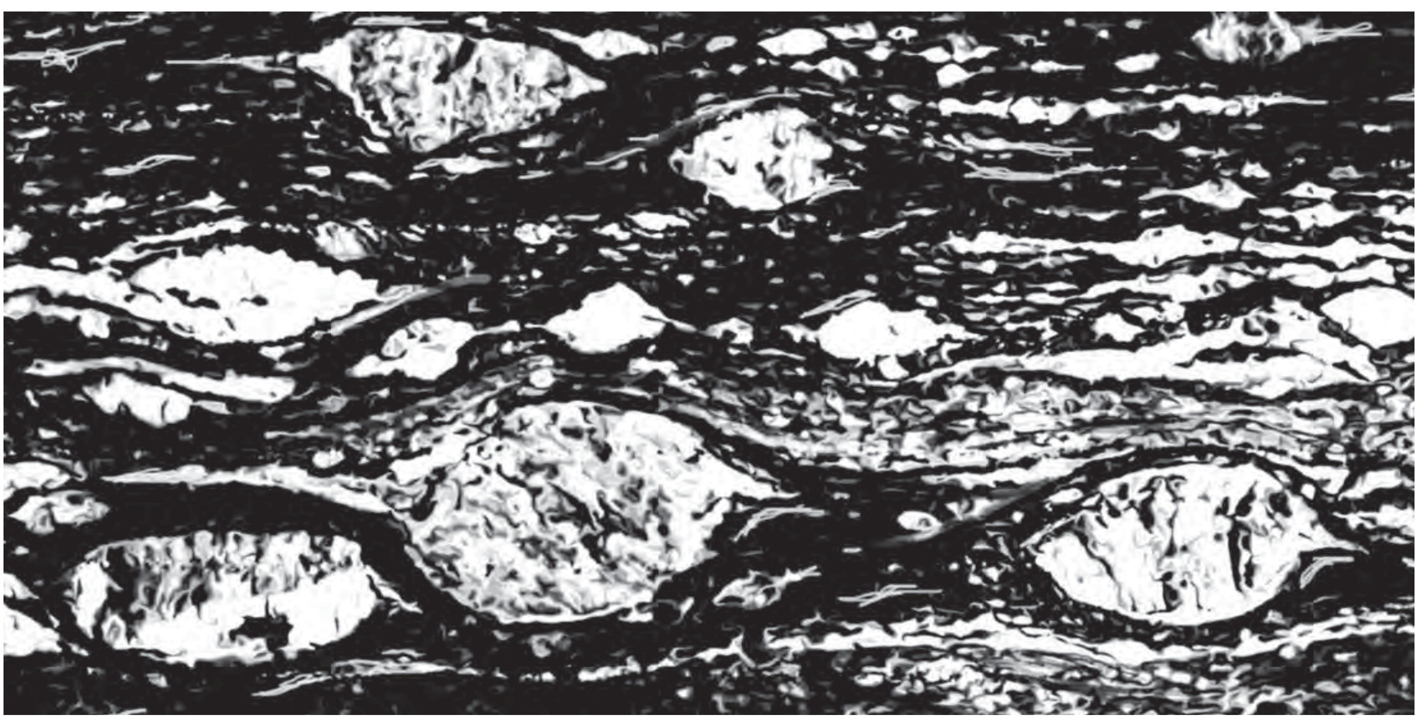


GLACIAL HYDROLOGY OF THE MONTANE CORDILLERAN ICE SHEET IN SOUTHERN BRITISH COLUMBIA: AN ANALOGUE FOR INTERPRETING LANDFORMS AND SEDIMENTS IN NORTHERN WASHINGTON STATE AND ALONG THE FRINGE OF THE CHANNELED SCABLAND?

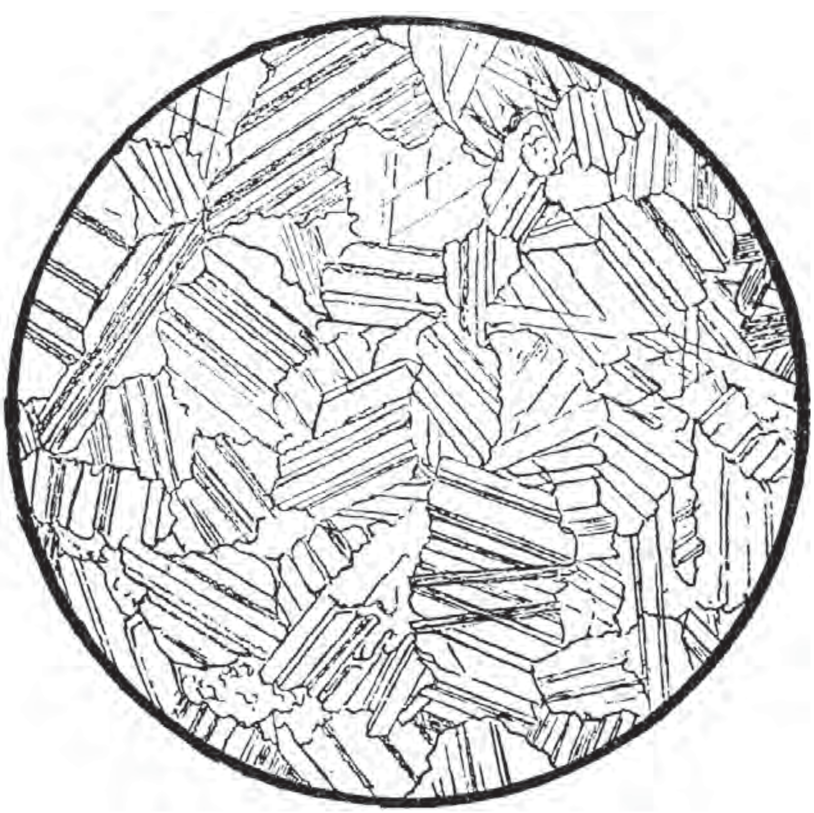
Jerome-Etienne Lesemann

Vancouver Island University

Growth and decay of the Cordilleran Ice Sheet (CIS) led to abundant meltwater production, storage and drainage. In Okanagan Valley, southern British Columbia, some of this meltwater was stored, and drained, through subaerial channel systems connecting deglacial lakes developed at the close of the last glaciation. Various lines of evidence also suggest that a portion of this meltwater may have been stored and transferred through subglacial lake systems and subglacial channels (i.e., tunnel valleys).

Six decades of intermittent work in southern British Columbia have led to a regional understanding of deglacial lake systems. The possibility of subglacial storage/drainage has more recently been added to the regional geologic context. More significantly, these glaciohydraulic reconstructions represent time/space 'snapshots' associated with approximate ice marginal positions at times when the Okanogan Lobe of the CIS had retreated from its terminus at the Withrow Moraine. Along Okanogan Valley, the area between the 49th parallel and the Withrow Moraine contains landforms and sediments similar to those used to infer subaerial and subglacial meltwater flow conditions in Okanagan Valley (north of the 49th Parallel). As well, little is known about the potential development of deglacial lakes along Okanogan Valley during retreat of the Okanogan Lobe from its terminus. Given physiographic similarities with southern British Columbia and extensive sedimentary deposits resembling those to the North, it is reasonable to infer that similar deglacial processes may have taken place along Okanogan Valley and along similar valleys fringing the southern CIS margin, though this remains speculative due to a paucity of field data.

This presentation will consist of a review of the regional context, processes, and products of various types of meltwater flow (subaerial and subglacial) associated with the CIS around Okanagan Valley in southern British Columbia. Evidence for subglacial flow in northern Washington State will also be discussed. Finally, using the geologic and geomorphic context developed in southern British Columbia, more speculative elements of CIS hydrology such as lake drainage(s), interpretations of sedimentary sequences fringing the Channeled Scabland, and development of deglacial lakes in northern Washington State will be discussed.



EOCENE MAGMATISM ACROSS THE NORTHERN CORDILLERA

Kirk Schleiffarth and Peter Larson

Washington State University

ABSTRACT

Eocene magmatism in the northwest Cordillera extends from the Clarno Formation in north central Oregon eastward to the Black Hills of South Dakota, and as far north as the Alaska-Yukon border. In the northwestern United States, Eocene complexes include the Clarno, the Absaroka and Challis volcanics, and their related intrusive phases in Wyoming, Montana, and Idaho; the Penticton and Colville complexes in south central British Columbia and north central Washington; and the Montana alkalic intrusions that extend into Wyoming and South Dakota. The Buck Creek, Kamloops, and Penticton complexes lie in central British Columbia. Many of the igneous complexes were emplaced within a few million years of 50 Ma. Magmatism in most of these occurrences is intimately associated with contemporaneous crustal extension and often with core complex emplacement. They are not arc-type subduction magmas, with the possible exception of the Clarno.

There is a tendency for the magmas to be alkaline and in several locations the complexes include carbonatites (e.g., Bear Lodge complex, northeast Wyoming). Initial Sr isotope ratios are usually continental and are commonly as high as 0.710. The εNd values are normally low and can be near -20. Lead ratios are typically radiogenic and high. These data suggest the involvement of old cratonic rocks in their petrogenesis.

Melting/assimilation likely occurred at various crustal levels. Several authors have suggested that the magmas are the result of adiabatic melting in the uppermost mantle followed by delamination of a gravitationally unstable and overthickened lithosphere. The lithosphere had probably been enriched in incompatible elements via devolatilization of a slab in an earlier subduction event. The magmas then rise into the crust aided by extensional features such as fractures.

INTRODUCTION

Understanding how continental crust is created and destroyed is critical to unraveling Earth's history and the evolution of the North American continent. Specifically, knowing whether new crust is derived from juvenile mantle material or is created by partial melting of preexisting continental crust (Morris et al., 2000), or both, is essential to interpreting the formation and evolution of the Earth. The North American Cordillera is a complex geologic province where magmatic and tectonic events extending from Archean to Holocene time have been overprinted, including the Eocene Challis-Kamloops Volcanic Belt, or CKVB (e.g. Dickinson and Synder, 1978; Lewis et al., 1987; Link and Janecke, 1999; Tysdal, 2000; DeCelles, 2004). The petrogenesis of the CKVB is complex and poorly understood, yet is critical to deciphering the continent's development.

The CKVB extends south-southeast from central British Columbia to central Idaho and is related to the paleotectonic plate boundary between the Farallon/Kula and North American plates, and the widespread extension throughout the North American Cordilleran thrust belt following the Sevier-Laramide orogeny (fig. 1; Morris et al., 2000; Breitsprecher et al. 2003; Dostal et al., 2003). The CKVB forms a narrow belt in British Columbia and becomes progressively wider towards the south in the northwestern United States (Morris et al., 2000).

The volcanic province is composed of many volcanic fields including the Buck Creek, Kamloops, Penticton, and Princeton fields of British Columbia; and the Colville, Clarno, Challis, Absaroka, and Montana alkalic fields of the northwestern United States. Most of these volcanic fields produced rocks that have a wide range of silica concentrations and high-K calc-alkaline geochemistry. They contain enriched large ion lithophile elements (LILE), light rare-earth elements (LREE), and relatively depleted high field strength

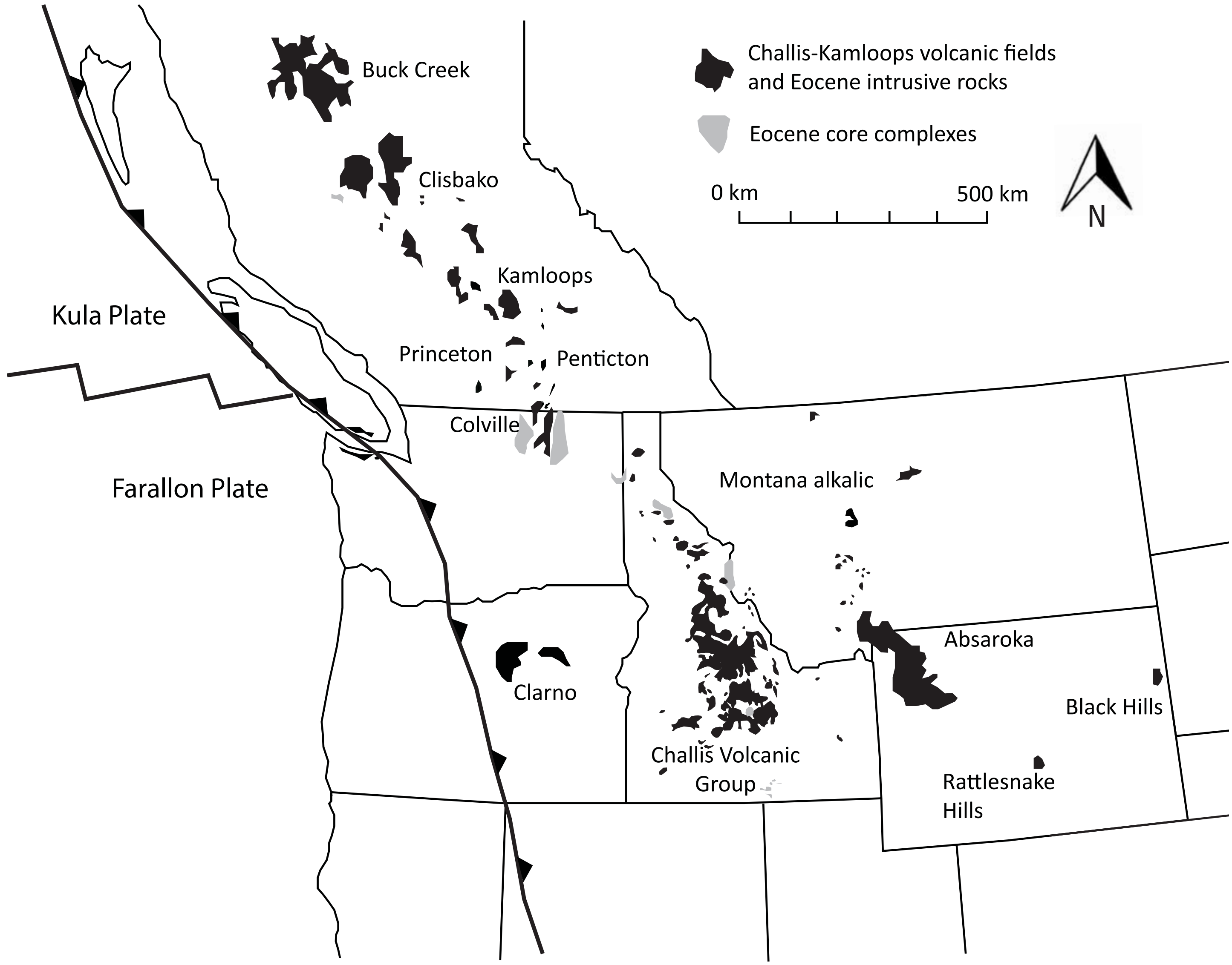


Figure 1. Map of tectonic setting and major geologic features of early Eocene time (54–40 Ma) exposed in the northwestern United States and southern British Columbia. This figure shows the location of the major CKVB volcanic fields and associated intrusions, and metamorphic core complexes within present-day geography. Modified from Bendick and Baldwin (2009), Breitsprecher et al. (2003), Ikert et al. (2009), Norman et al. (1992), Moye et al. (1988), and Lewis et al. (2012).

elements, or HFSE (Saunders et al., 1991; Ewart et al., 1992). The volcanic fields within the CKVG represent a magmatic flare-up between 54 and 40 Ma and are all related to extensional tectonics (McIntyre et al., 1982; MacDonald et al., 1992; Morris et al., 2000; Dostal et al., 2003).

Large normal fault systems, extensional basins, and metamorphic core complexes are present throughout the CKVG and are spatially and temporally associated with volcanism. Most models for the formation of the CKVB attempt to account for both the extensional stress field prevalent throughout the province and the calc-alkaline signature of the Eocene volcanic rocks.

Existing models include:

- Shallow subduction and slab rollback (Vance, 1979)
- Delamination of the Farallon plate (Humphreys, 1995; Feeley, 2003)
- Rifted arc (Dostal et al., 2001), and
- Slab window (Breitsprecher et al., 2003)

There is no broadly accepted explanation for the petrogenesis of the CKVB; however, the subduction model is generally disregarded (Moye et al., 1988). Here, we briefly describe the major Eocene igneous complexes in the Cordillera (fig. 1).

CHALLIS VOLCANIC GROUP

The Challis Volcanic Group (CVG) is located in east-central Idaho. It is the largest of the Eocene volcanic fields and much of it lacks detailed mapping. The CVG is located in the southern portion of the CKVG and is exposed over an area of 25,000 km^2 (Moye et al., 1988). It lies directly north of the Snake River Plain and east of the Idaho batholith. The CVG represents a flare-up of explosive volcanism that erupted from 51–40 Ma (McIntyre et al., 1982; Larson and Geist, 1995). Generally, voluminous and explosive eruptions originated from cauldron complexes, while effusive lava flows originated from small vents scattered throughout the volcanic field (McIntyre et al., 1982). Volcanic deposits of the CVG covered irregular terrain composed of Paleozoic and Precambrian sedimentary rocks (Moye et al., 1988). Due to the scattered vent locations, irregular topography, various eruption styles, and significant erosion between eruptive events, the CVG stratigraphy can be laterally restricted and quite complicated to unravel (Schleiffarth, 2013).

BUCK CREEK VOLCANIC COMPLEX

The Eocene Buck Creek volcanic complex of west-central British Columbia is associated with strike-slip fault-bounded basins trending northwest-southeast. The volcanic rocks were emplaced during between 51-40 Ma and have high-K, calc-alkaline compositions. The volcanic rocks occur in mostly thin, intermediate to mafic, porphyritic lava flows, with interbedded volcanic breccia, and are associated with subvolcanic plugs and dikes (Dostal et al., 2001). Most of the lava flows have large euhedral plagioclase laths that are accompanied by clinopyroxene and minor phlogopite phenocrysts (Dostal et al., 2001). The εNd and $^{87}Sr/^{86}Sr(i)$ values (fig. 2) suggest that a mantle source produced the Buck Creek volcanics. Dostal et al. (2001) proposed that the volcanic complex records a gradual transition from compression and subduction to an extensional tectonic environment.

KAMLOOPS AND PRINCETON GROUPS

The Eocene Kamloops Group of south-central British Columbia is an intermediate to mafic, calc-alkaline volcanic field (Dostal et al., 2003). The Eocene Princeton Group, located in south-central British Columbia just south of the Kamloops Group, consists of volcanic rocks that erupted as cinder cones and stratovolcanoes within extensional basins (Ickert et al., 2009). Compositions range from basaltic andesite to rhyolite. Magmatism took place between 53-47 Ma (Dostal et al., 2003). Major and trace element patterns of the Princeton group resemble those of many continental arcs (Ickert et al., 2009). The entire assemblage has an adakitic component (Ickert et al., 2009), interpreted to be the result of melting young, warm, oceanic crust that had been subducted (Garrison and Davidson, 2003). Ickert et al. (2009) attributed the subduction signature to a slab window, a gap in the slab where a spreading center is subducted, beneath southern British Columbia.

PENTICTON AND COLVILLE IGNEOUS COMPLEXES

The Penticton Group of south-central British Columbia and the Colville igneous complex of northeast Washington are correlative volcanic assemblages that have a calc-alkaline to alkaline affinity (Dostal et al., 2003). The Penticton group is 2,500 m thick and is

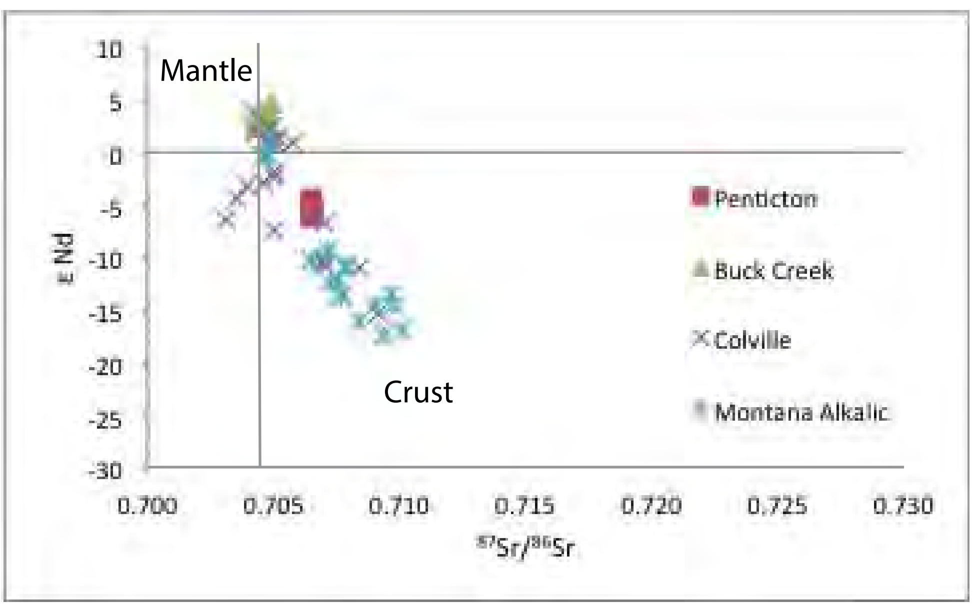


Figure 2. εNd and $^{87}Sr/^{86}Sr(i)$ variation diagram illustrating the wide range of isotopic signatures present in CKVB volcanic fields. Buck Creek (green triangles) represents the most mantle-like isotopic signature (positive εNd and low $^{87}Sr/^{86}Sr(i)$), while the Montana alkalic province shows more continental isotopic values (negative εNd and high $^{87}Sr/^{86}Sr(i)$). Data from Dostal et al. (2001), Dostal et al. (2003), Morris et al. (2000), and MacDonald et al. (1992).

well preserved in half-grabens and basins bounded by strike-slip faults (Ewing, 1981). The igneous complex consists of intrusive, hypabyssal, and extrusive rocks that are distinguished from the older Cretaceous intrusions by their association with extensional structures (Morris et al., 2000). The emplacement of the igneous complex was synchronous with the formation of the Kettle and Okanogan core complexes and four associated grabens that trend north-northeast (Morris et al., 2000). The major extrusive group, the Sanpoil Volcanic Formation, consists of andesite, dacite, and rare trachyandesite. The earliest members were deposited as ash-flow tuffs, while the majority was erupted as lava flows (Morris et al., 2000). A wide range of εNd and $^{87}Sr/^{86}Sr(i)$ values suggests that the Colville and Penticton provinces were derived from both mantle and crustal sources (fig. 2). Morris et al. (2000) interpreted the magmatism to be the result of decompressional melting of both the crust and mantle during Eocene extensional orogenic collapse.

CLARNO FORMATION

The Eocene to Lower Oligocene volcanic rocks of the Clarno Formation are exposed in the Blue Mountains in central Oregon. The Clarno is primarily composed of andesitic and basaltic lavas, pyroclastic rocks, breccias, and intrusions (Gromme et al., 1986). However, compositions range from basalt to rhyolite (Rogers and Ragland, 1980). Clarno rocks have both calc-alkaline and tholeiitic affinities and show enriched LILE and depleted HFSE signatures, typical of arc magmas (Rogers and Ragland, 1980). Feeley (2003) attributed Clarno magmatism to subduction due to slab rollback of the Farallon Plate.

ABSAROKA VOLCANIC FIELD

The Absaroka volcanic field is located in northeast Wyoming and southwest Montana and is the second largest volcanic field (23,000 km^2) within the CKVB (Feeley, 2003). The province contains calc-alkaline to alkaline rocks ranging from basalt to rhyolite. The Absaroka volcanic field was once considered to be related to subduction (Chadwick, 1970). More recently, Feeley (2003) showed that it has distinctive radiogenic isotopic and trace element compositions that suggest the source regions were within ancient sub-continental lithospheric mantle. He suggested that during the Eocene, the Farallon plate detached and sank, which triggered partial melting and magma generation in the asthenosphere (Feeley, 2003). This model would explain the across-arc variations in potassium, which is considered typical arc behavior.

MONTANA ALKALIC VOLCANIC FIELD

The Montana alkalic volcanic field is composed of several small Eocene volcanic fields scattered throughout much of Montana, northeastern Wyoming, and the western Dakotas. The province erupted contemporaneously (54-50 Ma) with the other CKVB provinces. The rocks range from calc alkaline to highly potassic and locally include carbonatites. They were emplaced as intrusions, lavas, and pyroclastic flows (MacDonald et al., 1992). Radiogenic isotope ratios show a large range with εNd and $^{87}Sr/^{86}Sr(i)$ values lying between -16 to -10 and 0.707 to 0.710, respectively (fig. 2; MacDonald et al., 1992). The thickness of lithospheric mantle beneath the Archean Wyoming craton is problematic for models invoking shallow subduction. The slab would be forced to depths greater than reasonable source regions for subalkalic basalts derived from the mantle (Dudás, 1991). Dudás (1991) suggested alternate models involving mantle upwelling and adiabatic melting. Crustal thinning occurring in response to overthickening in the Cordilleran thrust belt could have triggered the Eocene magmatic event (Dudás, 1991). MacDonald et al. (1992) suggested that the province had a more complex petrogenesis with several sources of melting within the lithospheric mantle.

SUMMARY

Volcanic fields of the CKVB have a wide range of silica concentrations, calc-alkaline to high-K alkaline affinities, and a wide range of radiogenic isotope ratios. Volcanism in the southern portion of the CKVB covers a width of over 1,000 km, while volcanism in British Columbia occurred in a narrower area within and adjacent to the North American craton (Morris et al., 2000). Eocene volcanism in western North America represents an anomalous flare-up that began and ended contemporaneously throughout the province. The flare-up followed a ten-million year gap in magmatism in Paleocene time after the long-lived subduction and compression that occurred throughout Mesozoic time. All of the volcanic fields within the province are associated with extensional faults and basins, which are likely indirectly related to previous crustal shortening. The width of the volcanic province, the abrupt initiation and halt of volcanism, and the widespread extension, including the emergence of metamorphic core complexes during Eocene time, suggests that subduction is not an appropriate model for the generation of the CKVB magmas.

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USING PRE-EXISTING THERMAL INFRARED AND SIDE-LOOKING AIRBORNE RADAR TO OUTLINE THE LIBERTY MINING DISTRICT

D.H. Vice

Pennsylvania State University

ABSTRACT

The Liberty Mining District is located in Kittatas County, Washington, about 32 km (20 miles) north of Ellensburg. It is a small, gold-producing area in the center of an unnamed regional dome on the eastern flank of the Cascade Mountains. The mineralization occurs in quartz and calcite veins associated with a series of diabase dikes. The boundaries of the Liberty Mining District have been poorly defined in the past, leading to unnecessary expenditures of exploration dollars. Pre-existing thermal infrared (TIR) and side-looking airborne radar (SLAR) imagery of the Liberty Mining District was studied to determine the areal extent of the district. The use of two separate geophysical methods made it possible to outline the mining district with greater accuracy.

A TIR survey was conducted over the Liberty Mining District to identify areas of potential geothermal activity. Diabase dikes outside of the district were found to have a brighter (i.e., warmer) thermal signature than diabase dikes inside the district. Dikes inside the district are more altered and have a thermal signature that is closer to the surrounding bedrock. This distinction helps in outlining the extent of the Liberty Mining District. A SLAR survey of the same area was studied to identify structural features, particularly any lineaments which may represent conduits for hydrothermal fluids. The lineaments inside the Liberty Mining District showed more of an east-west orientation while those outside of the district showed more of a northeast-southwest and northwest-southeast orientation. Used together, these two remote sensing systems outline the mining district, and indicate that it represents only about 13 percent of the area of the dome.

INTRODUCTION

Remote-sensing methods such as thermal infrared imagery (TIR) and side-looking airborne radar (SLAR) have often been used to study mineral deposits. In this study, existing TIR and SLAR imagery were re-examined about 2010 to outline the extent of the Liberty Mining District, and thus delineate where exploration expenditures would be more effective. Burlington Northern (BN) conducted the TIR survey over the Liberty Mining District during the fall of 1978 as part of a geothermal exploration program; they then conducted the SLAR survey in 1983 as part of a gold exploration program on company lands adjacent to the Liberty Mining District.

The problem has been that basalt dikes occur throughout the area, but only those dikes within the mining district have gold-bearing quartz and calcite veins associated with them. This association has led many people to prospect basalt dikes that contain calcite veins outside of the district. One example of this is the M3 tunnel (fig. 1) approximately four miles north of the Liberty Mining District. The diabase dike at this site has a quartz-calcite breccia zone (Mullen et al., 1942) that did not contain any gold, thus the expenditure of time and money spent driving the tunnel was wasted.

GEOLOGY OF THE LIBERTY MINING DISTRICT AND SURROUNDING AREA

The Liberty Mining District is a small, gold-producing area located in the center of a regional dome or northwest-trending arch on the east flank of the Cascade Mountains (Walsh et al., 1987). A series of broad, open, northwestern-trending folds are present within the dome. One of these structures, the Green Canyon Monocline, forms the southern edge of the structure (fig. 2) (Rector, 1963; Dahl, 1973). The district is located approximately 32 kms (20 miles) north of Ellensburg in Kittitas County, Washington, in Sections 1& 2, T. 20 N., R. 17 E., and Sections 5 & 6, T. 20 N., R. 18 E. (fig. 3). Access to the mining district and surrounding area is via U. S. Highway 97.

Bedrock in the Liberty district and the surrounding area (table 1) consists of sediments of the Swauk Formation (Tabor et al., 1982; Walsh et al., 1987). Basalt

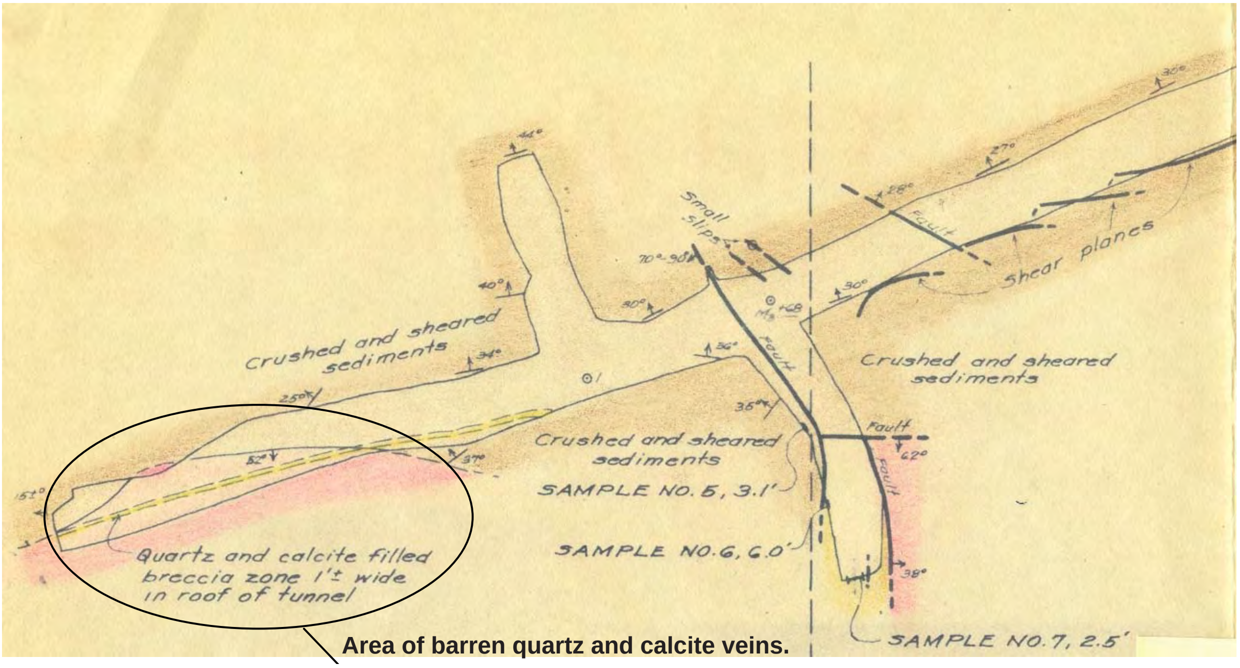


Figure 1. Principal Workings, Nairns First Claim (Mullen et al., 1942) in SW¼, NE¼, SW¼ of Section 13, T. 21 N., R. 17 E., which occurs just north of the northern boundary of figure 3.

and andesitic basalt of the Eocene Teanaway Basalt are exposed in a semi-circle around the south and east sides of the Swauk Formation. A swarm of basalt dikes related to the Teanaway Basalt cut sediments of both the Swauk and Silver Pass Formations (Dahl, 1973; Margolis, 1994). The occurrence of calcite veins associated with the basalt dikes throughout the domal structure surrounding the Liberty district has raised the question about the extent of the district. Gold-bearing veins are associated with two distinct zones of dikes within the Liberty district (Margolis, 1994).

A three hundred meter (984 foot) thick sandstone within the Swauk Formation is exposed in the hinge (or axis) of an anticline extending through the Liberty district (Tabor et al., 1982). Margolis (1994) indicates that this sandstone is pervasively altered; however, alteration in the rocks of the Swauk Formation lying above the sandstone is restricted to the vicinity of veins.

MINERALIZATION

Gold in the Liberty Mining District was first discovered in placers in Williams Creek in 1868 and in lode deposits in 1887. Intermittent mining activity has occurred to the present time (Rector, 1963). Production records are limited but total production probably has not exceeded $250,000 (Rector, 1963) which is estimated to be less than 7,100 oz. of gold. This amount is below the threshold level of 10,000 oz. that the U. S. Geological Survey uses to delineate major mining districts (Koschmann and Bergendahl, 1968, p. 3) so the Liberty district would be classified as a minor district. The lode gold deposits occur in pockets associated with quartz-calcite-dolomite veins that border some of the basalt dikes, and are related to late-stage hydrothermal activity (Rector, 1963; McMahan et al., 1973).

Table 1. Stratigraphic Column for the Liberty Mining District.

Formations	**Age**	**Description**
Tenaway Basalt	Eocene	Basalt flows surrounding dome and dikes cutting earlier sediments
Silver Pass Formation	Eocene	
Swauk Formation	Early Eocene	Continental arkose and shale

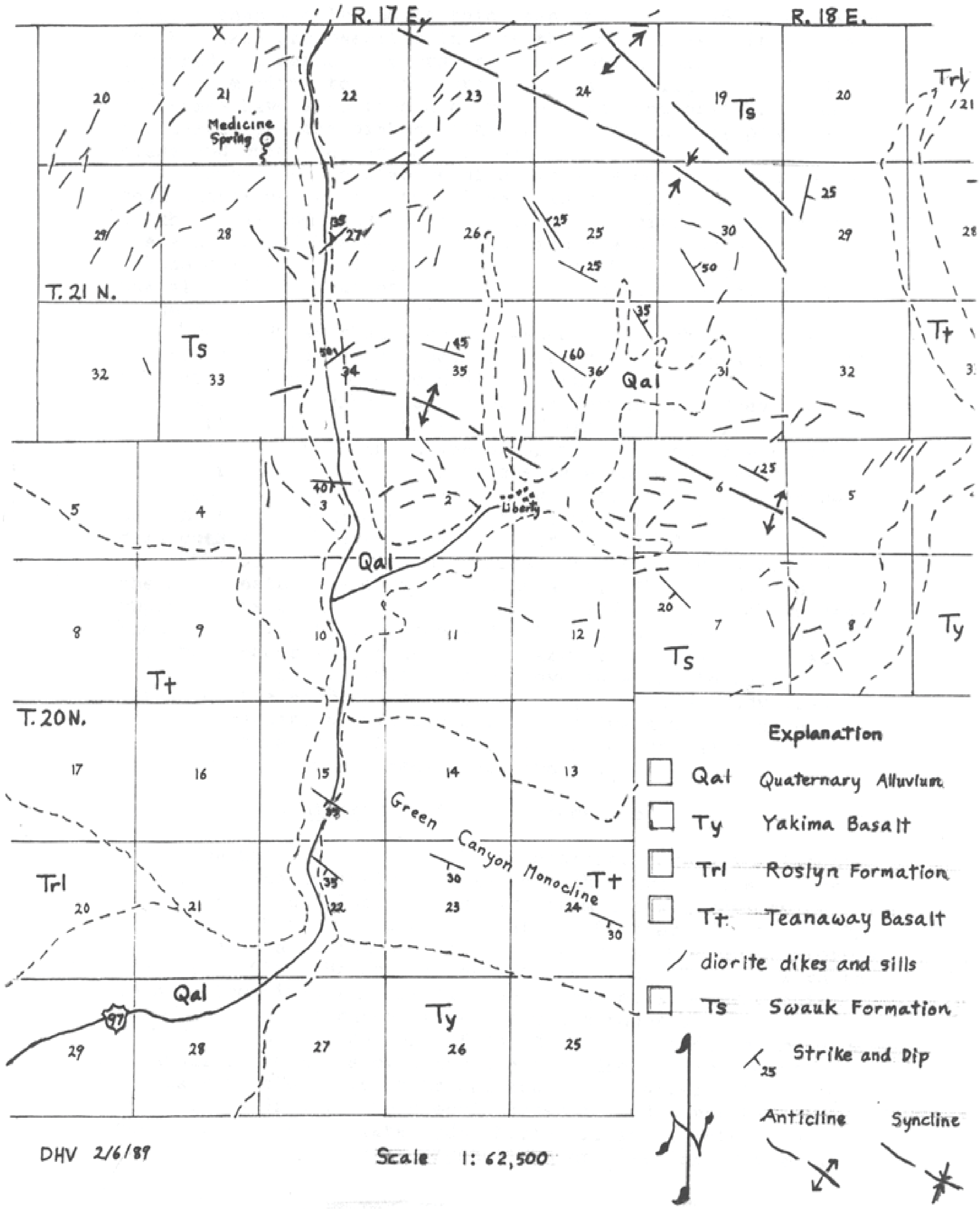
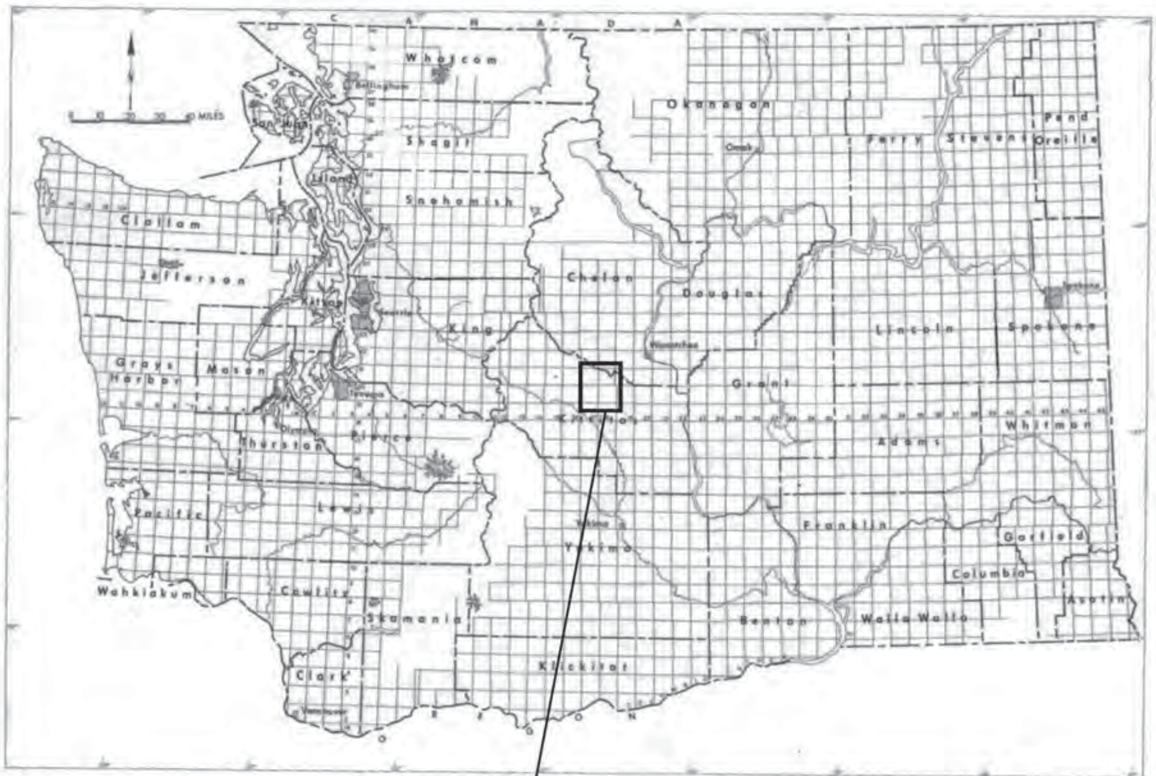


Figure 2. Geologic map of the Liberty Mining District (modified from Dahl, 1973; Walsh, 1987). Please note that although the scale of the map is given as 1 inch per mile, digitizing the map and inserting it into this paper has changed the scale to approximately 1:72,000.



Area of dome that contains the Liberty Mining District

Figure 3. Index map of Washington showing the location of the Liberty Mining District (dark rectangle). Background lines give township and range for state of Washington.

As an example of the rich character of the pockets within the district, Livingston (1963) notes that in 1956 two prospectors found 45 pounds of gold in a discontinuous vein or pocket that was approximately 2 m (7 ft.) long, 1.2 m (4 ft.) wide, and 0.6 (2 ft.) high. At the fixed selling price of $35 per ounce in 1956 (Koschmann and Bergendahl, 1968), this discovery had a value of $24,640. At the May 7, 2014 selling price of approximately $1,289.50 per ounce (www.goldprice.org), the value of this discovery would have been $928,440.00. Gold in the Liberty District often occurs in the form of wire, crystals, and dendrites, as well as matted aggregates, which increases the value because collectors are willing to pay a higher price (Rector, 1963; Dahl, 1973).

THERMAL INFRARED SURVEY (TIR)

Thermal infrared sensing is a passive remote sensing method that detects variations in the amount of energy (surface radiant flux) that is emitted from the surface of the Earth (Loughlin, 1991; Sabins, 1997). This energy is primarily in the thermal infrared range of the electromagnetic spectrum (from 8 to 14 μm) and is a good approximation of the temperature of the Earth's surface at the time the imagery was acquired.

Based on the occurrence of the Medicine Mineral Spring within the Liberty area, BN conducted a TIR survey in 1978 over the Liberty Mining District to determine if the area contained any potential for geothermal resources (Crowley Environmental and Planning Associates, 1982).

The TIR data were acquired in digital format and then processed to five-inch black and white format imagery with an approximate scale of 1:24,000 (Crowley Environmental and Planning Associates Inc., 1982). A more precise scale for the TIR imagery is not possible because the system does not use a lens; therefore, a focal length is not available with which to calculate a scale. The black and white imagery was assembled into a mosaic which was then used for interpretation. Variations in temperature on the land surface are shown in gray in the TIR imagery; the darker shades are cooler and the lighter gray shades are warmer (Sabins, 1997).

A Daedalus model 1210 thermal IR linescanner was used for the survey, which is similar to the system shown by Sabins (1997, p. 144) although no direct film recorder was available. A light aircraft was

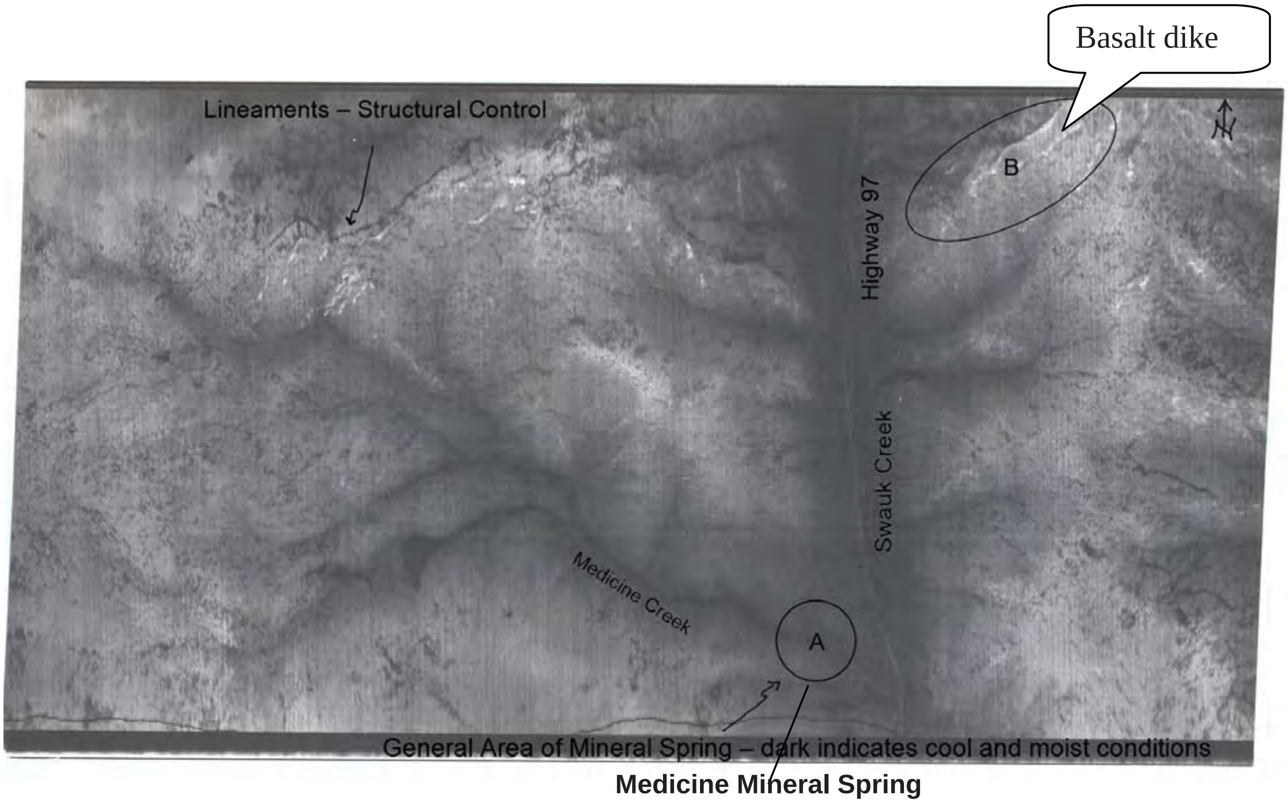


Figure 4. TIR imagery of the Medicine Springs area. (A) The Medicine Mineral Spring. (B) One of the basalt dikes outside of the Liberty mining district. The scale is nominally 1:24,000.

used as the platform for the scanner and the survey was flown at an altitude of 1,500 to 1,600 m (5,000 to 6,000 feet) above ground level. The Daedalus system had an instantaneous field of view of less than one milliradian, thus giving a spatial resolution of approximately 2 m (7 feet) (Sabins, 1997). The temperature resolution of the system was less than 1°F (Crowley Environmental and Planning Associates, 1982).

The TIR survey showed no temperature anomaly associated with Mineral Spring (fig. 4) or the immediate vicinity suggesting that there is no surface geothermal activity in the Liberty area (Vice, 1979). The darker gray areas along both Medicine and Swauk Creeks are due to greater moisture in the soil near the spring causing cooler temperatures in the sediments adjacent to and underneath the streams.

In interpreting the TIR imagery for geothermal anomalies, one interesting feature was observed. Basalt dikes outside of the Liberty Mining District showed strong anomalies in the TIR imagery while those inside the district were not as prominent (Vice, 1979). A field investigation found that the dikes outside the district had only a thin soil cover with sparse vegetation consisting mostly of grass with scattered conifers and brush (Vice, 1979) which made them appear much warmer in the imagery than the surrounding Swauk sediments which had a heavy cover of brush and conifers (fig. 5). This difference in vegetative and soil cover provided a sharp contrast in the TIR imagery, and thus the dikes appeared as warm lineaments (A in fig. 5).

The basalt dikes within the Liberty Mining District were more poorly defined in the TIR imagery because they were altered by the hydrothermal activity associated with the formation of the gold-bearing veins. Biotite was altered to sericite and the pyrite was later altered to limonite (Margolis, 1994). In a thin-section study of the basalts, Rector (1963) noted the presence of quartz and calcite, altered pyroxene minerals and plagioclase feldspars that had become more sodium-rich. The altered character of the rock has reduced the contrast within the TIR imagery between the dikes and the surrounding Swauk sediments (fig. 6). The reduced contrast is distinctive to the basalt dikes within the district compared to the response of the dikes outside of the district and helps to define the district boundaries.

SLAR LINEAMENT STUDY

Side-looking airborne radar (SLAR) is considered to be an active remote sensing system because it

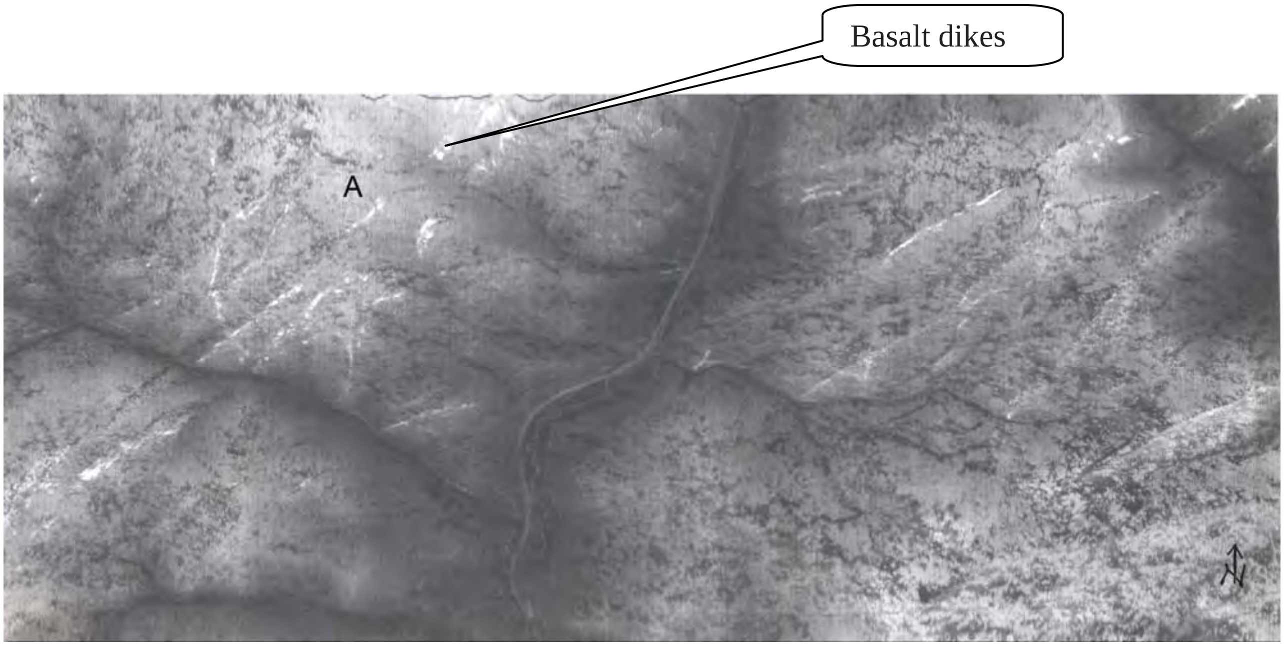


Figure 5. TIR imagery of unaltered dike outside of Liberty Mining District (A). The scale is nominally 1:24,000.

provides its own energy. The radar system sends out radio waves from the platform (normally an airplane) on which the system is mounted, and then records those waves as they are reflected back from the ground surface (Sabins, 1997).

SLAR is used to identify structural features, in particular, lineaments. Lineaments are mappable straight or curvilinear patterns and may reflect faults or dikes within the bedrock (Sabins, 1997, p. 93). Lineaments are targeted during exploration because the structures they represent could have controlled the movement of hydrothermal fluids and thus the location of mineral deposits.

In 1979, BN used SLAR imagery prepared for the U. S. Army Corps of Engineers, Seattle District, by MARS Inc., to study the Liberty Mining District. A complex set of lineaments were mapped in this study. Within the mining district, the lineaments mostly trended east-west and northwest-southeast; while mostly north-south and northeast-southeast lineaments occurred outside of the district (fig. 7). This change in attitude correlates with structural changes described by Tabor et al., (1982) who observed that the anticline in the Liberty Mining District appears to be offset and trends east-west, while an anticline and syncline north of the district are trending northwest (see fig. 3).

Figure 6. TIR imagery of altered dike within Liberty Mining District (D). The scale is nominally 1:24,000.

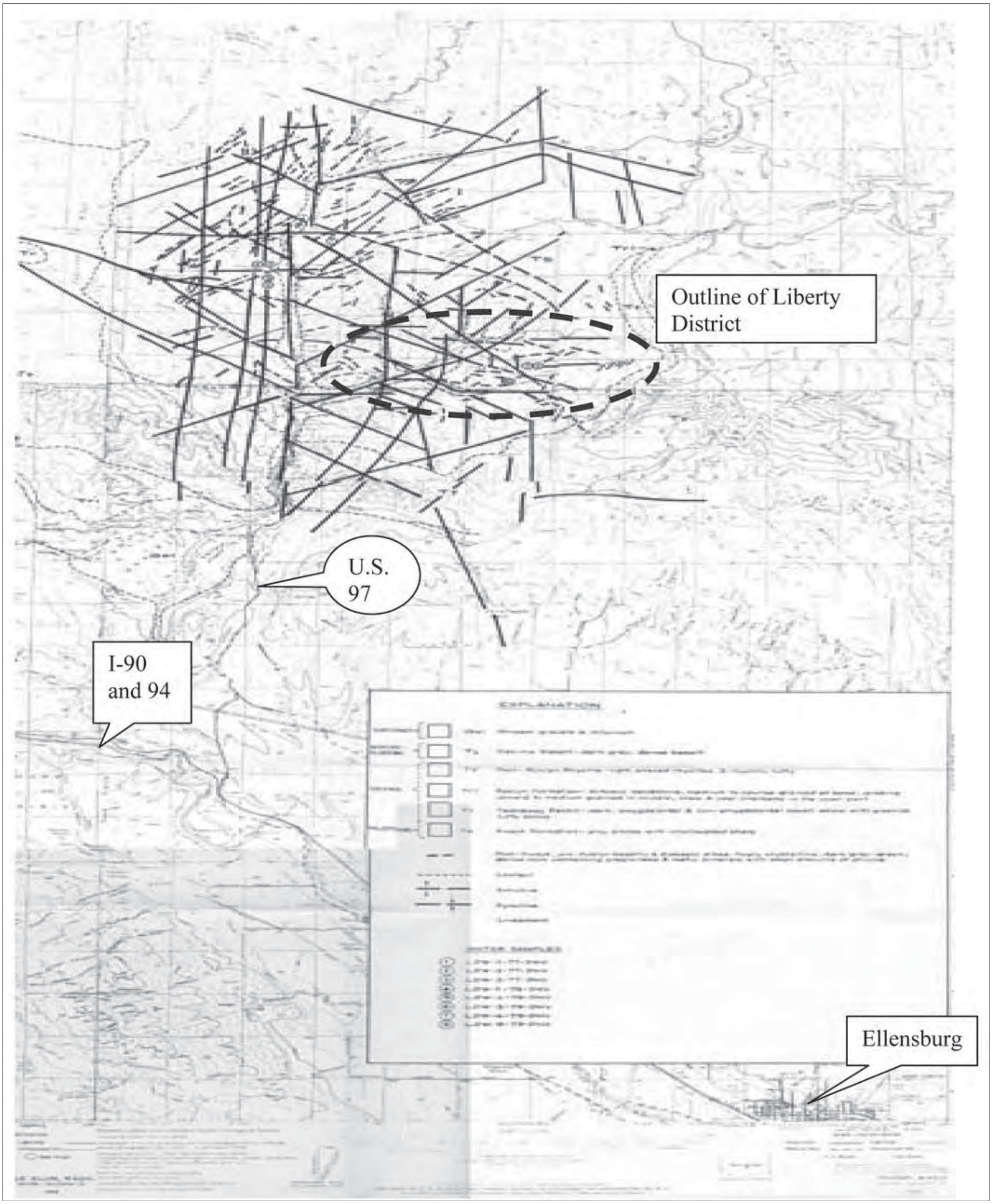


Figure 7. Lineament map of the Liberty Mining District and the surrounding area.

This change in orientation suggests that gold mineralization in the Liberty district was controlled by a localized structural pattern that is restricted to the area.

SIZE OF MINING DISTRICT

An estimate of the size of the Liberty Mining District is useful because it can provide an estimate of the total productive area. Defining the boundaries of the district allows exploration work to be concentrated where it can be the most productive.

Both the TIR and the SLAR data suggest that the Liberty Mining District comprises approximately four sections (2560 acres), which is about 13 percent of the approximately 30 sections (19,200 acres) of the surrounding dome. Mineral deposition occurred where a change in the orientation of structure provided pathways that enhanced fluid circulation (Bouchot and Genter, 2009). A modern analog might be the Bouillante geothermal field on the Island of Guadeloupe in the Caribbean, where geothermal activity is located at the intersection of a regional transcurrent fault (the Montserrat-Bouillante fault system) and a series of normal faults that are perpendicular to the main fault (Bouchot and Genter, 2009). Geothermal energy is just a small part of the overall igneous activity in the area.

SUMMARY AND CONCLUSIONS

When used together, TIR and SLAR imagery are an effective way to outline the boundaries of the Liberty Mining District. TIR imagery showed differences in hydrothermal alteration of the basalt dikes outside of the district compared to those inside it; the SLAR data showed a different lineament pattern for the Liberty district compared that of the surrounding area. In addition to determining the extent of the Liberty district, these two remote sensing surveys can be used to help answer a larger question: what is the relative area affected by hydrothermal activity compared to the total area of the surrounding dome. This study shows that to be 13 percent. This is significant in exploration because it effectively outlines the productive area of a mining district or a geothermal system.

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MODELS OF RESOURCE DEPLETION AND ORE PRICE: A REGIONAL CASE STUDY

Daniel H. Vice
Penn State–Hazleton Campus

Harold Aurand, Jr.
Penn State–Schuylkill Campus

ABSTRACT

The two main models of resource use, Hubbert's Curve and McCabe's Pyramid, have widely divergent ways of looking at Earth's total reserves of mineral resources and the approximate time of their depletion. Hubbert's Curve treats total resources as being finite and encloses them in a bell curve. Production is predicted to rise to a peak and then fall off at an inflexible rate. McCabe treats total resources (e.g., gold) as finite and represents in a pyramid with the highest grade ore at the peak and the lowest grade at the base. This model has no restrictions on the rate of production so that a stop-start history of production can occur. The history of three mining districts is used to illustrate the discontinuous history of production due to changes in technology and price.

INTRODUCTION

For the past fifty years a debate has been unfolding in the geological community over the use and depletion of our mineral resources. On one side are the Malthusists (or doomsters) who believe we are using resources at an unsustainable rate, and will leave nothing but shortages and poverty to future generations, a concept known as generational inequity. On the other side are the cornacopians (or boomsters) who argue that technology and society's ability to practice conservation and substitute one resource for another will prevent any resource shortages from threatening our future. Both groups have developed models to measure and illustrate their theories on resource use and depletion.

Although this debate is obviously of long-term importance, models of resource use also impact decisions we make today. Daniel Yergin's 2011 book *The Quest* gives several examples, three of which are discussed below.

First, according to Yergin, electric utilities build power plants with the expectation that they will be in operation for forty years. Determining which fuel to use requires developing a long-range estimate of available quantities and their long-term costs.

Secondly, the United States government recently persuaded General Motors to bring the Chevrolet Volt, a plug-in electric car, onto the market. This decision was based in part on the governments belief that a declining availability of oil would drive up the cost of gasoline and make electric cars more competitive. Had their model predicted a flatter price for gasoline they might have encouraged General Motors to invest more in battery research or other basic technologies to make the Volt more competitive.

Finally, Yergin believes speculators at the end of the last decade misread one of the models of resource depletion resulting in an artifical increase in the price of oil, contributing to the economic crisis (Yergin, 2011). All three of these examples show the importance of having accurate models and understanding their limits.

This paper will test the Malthusian and Cornacopian models of resource use as they apply to copper and gold mining by looking at the history of specific mining districts. These are the Robinson District in White Pine County, Nevada; the Round Mountain District in Nye County, Nevada; and the Golden Sunlight Mine in Jefferson County, Montana.

THEORIES ON AVAILABLE RESOURCES

The basic model of the Malthusian group is M. King Hubbert's Curve (fig. 1) which was created for the oil and gas industry (Hubbert, 1978), and similar, unnamed models such as those used by the Club of Rome and other groups in the 1970s to set depletion rates for most of the metals we use. McCabe's Pyramid (fig. 2) is a typical model for the Cornacopian group. It plots each resource on a pyramid, with the highest quality, most easily accessable deposits at the

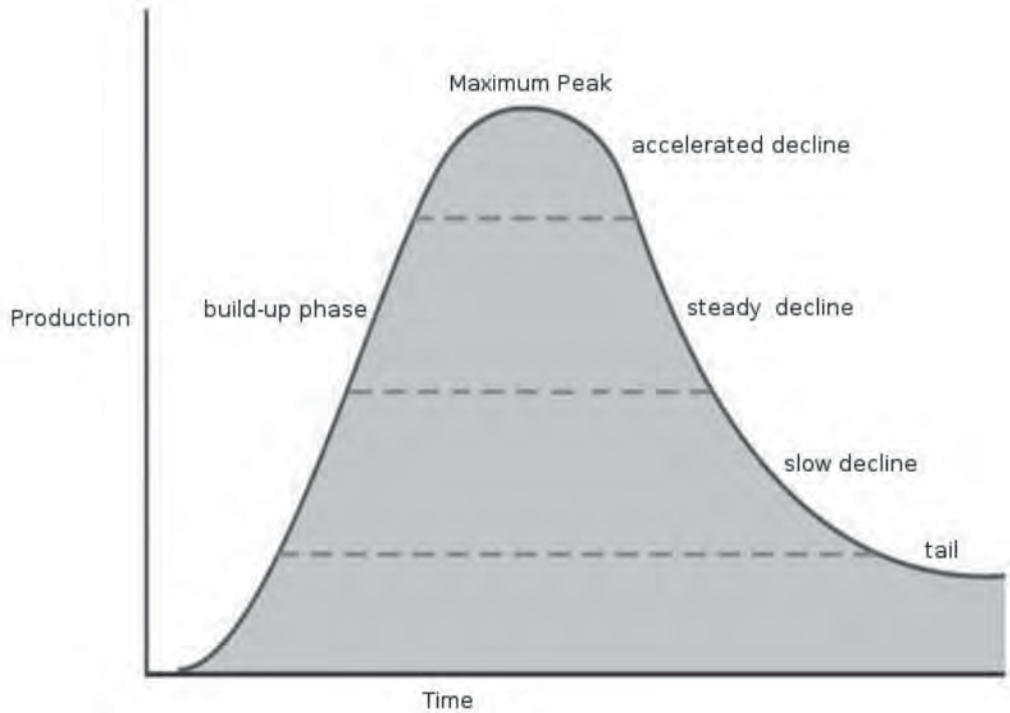

Figure 1. Hubbert's Curve.

top, and lower grade, less accessable deposits at the bottom.

The Club of Rome is a think tank founded in 1968. Their website, www.clubofrome.org, describes them as an informal association of independent leaders from politics, business, and science, who are interested in contributing to a better world in a systematic, long-term, interdisciplinary, and holistic manner. Their aims are to identify the most crucial problems which will determine the future of humanity and evaluate alternative solutions to the problems, and then communicate these solutions to decision makers in the public and private sectors and to the general public.

HUBBERT'S CURVE AND RELATED MODELS

The Hubbert model began with the estimation of the earth's total physical reserve of oil and natural gas. Using historical records, he then calculated the amount of those total resources that had already been used. He believed that resource use would follow a simple curve, so that he developed his model to demonstrate that historic production, then extrapolated into the future using the same simple curve until the resource was depleted (fig. 1). The model did not allow for changes in resource use either through new production technology affecting supply, or increased conservation, or resource substitution affecting the demand. Essentially, once a society began to use a resource, Hubbert believed they had started down a pathway with no exit. They would use that resource in a steady predictable way until the resource was depleted.

The Club of Rome followed a similar model in 1972 when they predicted that the world's supply of copper would be exhausted between 1993 and 2020, and the world's supply of gold would be exhausted between 1981 and 2001 (Arndt and Ganino, 2012). The Club of Rome arrived at these predictions by dividing the annual amount of use into the total known reserves to derive the date at which the resource would be depleted. They did not consider the effect of changes in technology, price, nor the possiblity of new additions to the total amount of resources.

MCCABE'S PYRAMID

McCabe imagined the total physical reserves of each resource as a pyramid (fig. 2) with the highest grade (i.e., the richest resources) at the peak and the lowest grade (i.e., the leanest resources) at the base. McCabe's Pyramid was not plotted against time and had an economic component in its definition. For McCabe's Pyramid the lower limit for a material to be economically produced was set by a combination of price and technology. Resources that were the most economical to extract were plotted at the top and therefore used first. As the resources that were the more accessible were depleted, production moved downward on the pyramid to more challenging resources until depletion. Changes in either technology or price would allow a lower grade of resources to be used which increased the total amount of resources available.

Production was not perceived as following a predetermined path in McCabe's pyramid, in fact, there was no reason why a resource couldn't be exploited for a while, and then stop, only to have the exploi-

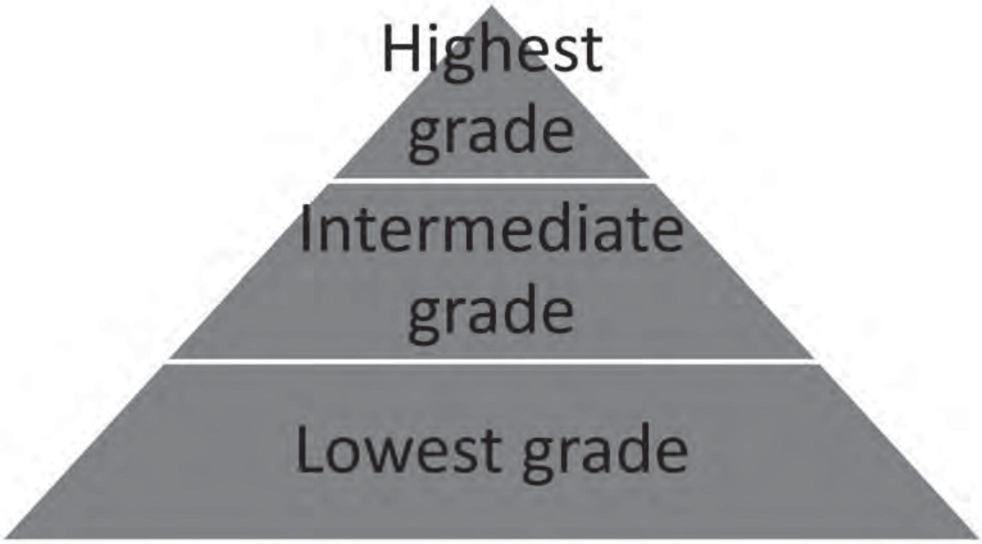

Figure 2. McCabe's Pyramid.

tation resume after changes in technology or price had occurred. The lack of a predetermined path in McCabe's pyramid allows for resource substitution, changing price and/or changing technology to affect resource availability. Resource substitution reduces demand. Changes in technology can increase supply by making a marginal ore more economical, while changes in price either increase or decrease supply of a resource by making marginal ore either more or less economical.

A COMPARISON OF DISTRICT HISTORIES WITH THE THEORIES

The Golden Sunlight Mine in Jefferson County, Montana; the Robinson District in White Pine County, Nevada; and the Round Mountain District in Nye County, Nevada provide interesting test cases for both Hubbert's Curve and McCabe's Pyramid. Figure 3 gives the general locations of the districts we discuss in this paper.

The Golden Sunlight mine in Montana started producing micron-size gold particles in late 1982 from an Eocene breccia pipe in a latite porphyry that had intruded the Precambrian Greyson and Newland Formations (Berg and Zeihen, 1985). In 1985, the indicated reserves were 25.8 million tons which was projected to be a 14-year supply for the mine. The mine has operated nearly continuously since then (Gevock, 2012) although it came close to shutting down in the late 1990s because of low gold prices. Gevock (2012) noted that the mine has had its life extended several times beyond the

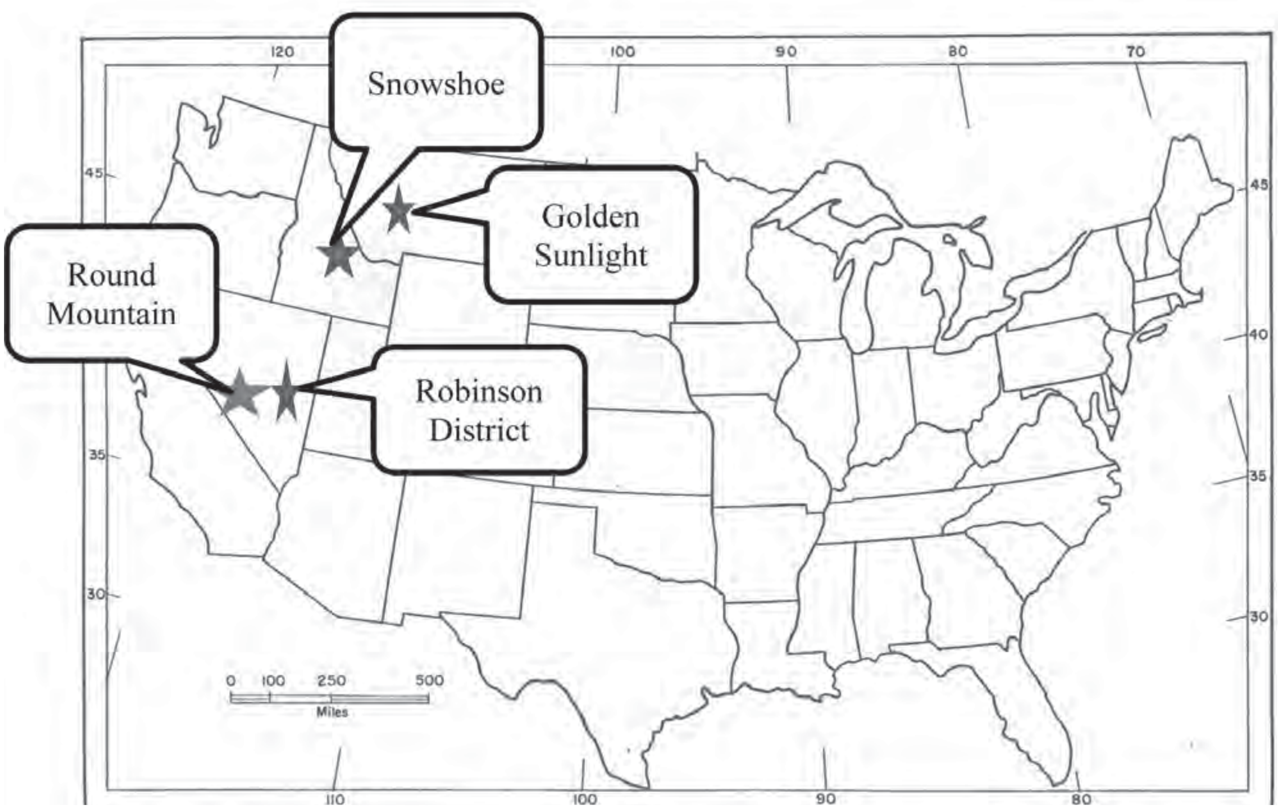

Figure 3. Locations of mines discussed in this paper.

original fourteen years after additional exploration found more ore. The most recent extension came in 2008 when Barrick Gold Corporation invested $130 million to extend surface mining operations into a new area. That expansion is expected to allow the mine to continue operations through 2015.

The Robinson District in White Pine County, Nevada, is a copper porphyry type of deposit (i.e., large, low-grade ore body) with some gold deposits around the perimeter. The copper ore in the district was discovered in 1902 and production began in 1908. Mining continued until 1978 when the operations were shut down. When copper prices rose in the late 1980s a new company came in and started mining in 1992 but shut down with low prices in 1999 (Tingley and Pizarro, 2000). However, the Polish mining company, KGHM International, has gone back into the Robinson district and is mining copper again (Prenn, 2013).

The Round Mountain deposit in Nye County, Nevada, is another example. It was mined from the time of discovery in 1906 until the start of World War II. Mining started again in 1977 and was projected to have a six-year life (Werschky, 2013). However, mining has continued to the present and the mine is still projected to have a six-year life (Werschky, 2013).

All three mining districts show a stop-start history dependent largely on price. This is what McCabe's Pyramid would have suggested, but very much against Hubbert's predetermined prediction curve. This conclusion is supported by mineral supply studies conducted by the USBM from the late 1970s until 1996 showing a wide range of reserves (Bob Weldin, 2014, email).

In many ways, this should come as no surprise. While Hubbert's Curve has remained popular and is often defended (Deffeyes, 2001 and 2010) its actual predictive value has fallen short. Hubbert himself believed that oil production would peak in the United States in the early 1970s (Hubbert, 1978). In reality, production is now at an all-time high because of new drilling technology that has been introduced.

In 1972 the Club of Rome made the predictions discussed above; that the world's supply of copper would be exhausted between 1983 and 2020 and the supply of gold between 1981 and 2001. Arndt and Ganino (2012) described predictions of reserve depletion made in 2009 in an article in the industry journal, *Mining Environmental Management*. These estimates found that times of exhaustion had barely changed or even increased since the Club of Rome had made their predictions forty years previously. For example, the predicted time of depletion for copper was then thirty-two years and that for gold was sixteen years. They noted that two factors were important for this change in the time of depletion for these metals. The first factor was that new deposits were being found and developed at a rate that has kept pace with demand. The second factor was that once a company had sufficient resources for twenty to thirty years of mining with current technology, it made no sense to spend more money for exploration and development of new orebodies. Thus the amounts of known reserves tend to remain stable for the long-term.

Technology has also played a role in undermining predictions. In the early 1970s when the Club of Rome was issuing its dire warning about gold depletion, the Homestake Mine in South Dakota was the principal American producer. The Homestake has since closed, but the United States has continued to vary between being the world's second and third largest gold producer. The United States has maintained this position because of the development of technology that can recover gold from low grade ore. Today the United States produces 8.8 percent of the world's gold, with nearly 70 percent coming from Nevada (George, 2013).

There are other examples. In the anthracite coal fields of Pennsylvania the development of a fluidized bed combustion chamber has allowed cogeneration plants like Northeastern Power in McAdoo to burn waste coal with only 5000 or 6000 btus per pound, where once 12,000 btus was considered optimal (fig. 4). The use of hydaulic fracking in the Marcellus Shale in Pennsylvania and other gas fields has increased natural gas production to the point where the United States is becoming a net exporter of that fuel (Yergin, 2011).

In all of these examples changes in technology or price created new reserves where none had existed before. The result was to shift, often dramatically, what Hubbert had seen as a predetermined curve. Mineral supply studies based solely on price by the USBM

Figure 4. The McAdoo Cogeneration power plant.

from the late 1970s until 1996 show a similar stop-start pattern (Bob Weldin, 2014, email).

One exception to these changes is when land use decisions permanently block access to certain locations. For example in the late 1960s and early 1970s a mineral survey was conducted in the Idaho Primitive Area (fig. 5). At that time the Snowshine Mine had been inactive for many years. Part of this was because of its relatively isolated location, and part because the Federal Government had frozen the price of gold at $35 per ounce. Today, due to changes in the market, gold is valued at $1291.80 per ounce (About.com, May 1, 2014), and economic mining might again be possible, but the land classification system prevents anyone from acting on the new price or technological advances.

SUMMARY AND CONCLUSIONS

This paper discusses two theoretical models on resource use, Hubbert's Curve and McCabe's Pyramid, and compares them to the history of three mining districts, the Golden Sunlight, Robinson District, and Round Mountain. The history of these districts best fits the McCabe's Pyramid model.

ACKNOWLEDGMENTS

The senior author wishes to thank Margaret Delaney for digging a lot of Golden Sunlight material out of the files and a number of people with whom he has worked on the Idaho Primitive Area or discussed this paper over the years including Dave Lockard and Bob Weldin. The authors wish to thank a number of friends who have reviewed this article and made many helpful suggestions.

Figure 5. The Snowshoe Mine.

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CONSTRUCTION OF THE CANNON GOLD MINE TAILINGS FACILITY, WENATCHEE, WASHINGTON

Jack Caldwell, P.E., MSc (Eng), LLB

It was 1983 and the mining industry was in a deep downturn. Work was hard to find, particularly consulting assignments. Our Vancouver office of Steffen, Robertson, and Kirsten (now SRK) was nearly deserted. One day, Asamera Minerals, Inc., an American mining company, phoned our office and asked if we would be interested in designing the tailings facility for a gold mine they wanted to develop near Wenatchee, Washington (fig. 1). Of course we said yes, and I travelled to the mine the next day.

Figure 1. The downstream face of the embankment of the Cannon Mine Tailings Facility.

I was met by John Toften, the underground mine manager, but our first meeting did not go well….. After a short tour around the site he had selected for the tailings facility, he asked me what I would do next. I told him that I needed to excavate test pits and develop a drilling plan to characterize overburden and bedrock---standard geotechnical approaches.

“Jack,” he said, “We are miners. We dig and see what we have got, and we respond to what we have got. You will come and live on site, be a miner, see what we have got, and respond to what we have got.”

I had no idea of how to do that. I had a family in Vancouver. But I had no other job. In short order, we packed up the house, and moved to Leavenworth. The first job was developing a conceptual design and the Environmental Impact Statement, and acquiring the necessary permits.

When I told my boss that the embankment would be over three-hundred feet high, he got nervous. He called on Syd Hillis and John Gadsby to be my peer reviewers. John was a consulting geotechnical engineer in Vancouver, still working today at age eighty. Syd had been the chief geotechnical engineer of the Revelstoke and Tarbela dams, in British Columbia and Pakistan, respectively, the highest and biggest earth and rockfill dams constructed up to that time. He was then a specialist consultant on dams for the Asian Development Bank.

Syd came to the site first. I walked him around the steep valley intended for the tailings facility. My idea was to construct a rockfill embankment to impound the tailings, which would be discharged as a fluid slurry into the resulting reservoir.

He made me show him potential material sources. The basalt the covered the high hills would be perfect as the rockfill. The loess on flatter plains could be compacted into an effective, low-permeability core. And the sandstone quarry that occupied part of the site could be used for sand for drains and filters, necessary if the fine-grained materials in the core started to fail through internal erosion.

Syd was convinced but wary. In 1976, in the Teton Dam in Idaho had failed, killing eleven, and causing serious flooding and property damage. The root cause of failure had been the friable sandstones of the foundation---they had piped when the dam was filling and failure resulted. The foundation for our embankment was composed of friable sandstone similar to that occurring at Teton Dam.

Over many a bottle of saki we deliberated and argued. The solution we implemented was this: construct a zoned embankment and place sand from the sandstone quarry on site as a filter across the entire foundation. The sand acts as a filter for finer-grained materials in the foundation, thus preventing piping of the foundation and embankment. It worked well. The mine manager was not pleased at the additional expense, but a discussion on dam failure convinced him of its necessity.

Kim D'Rubertis was our engineering geologist. He taught me a great deal about the local geology and how that geology affected my design. He helped me find appropriate soil and rock materials for the many zones of the embankment. He helped me characterize groundwater seepage into and out of the reservoir–the regulators in Seattle, from whom we needed permits, had started to ask difficult questions about the impact of groundwater on the structure, and the impact of the structure on the groundwater.

Peter Kiewit was retained as the contractor to construct the embankment. They were good people to work with, always ready to accommodate the design changes I was forever making—and charge the mine for them. For as new, or unexpected, conditions were revealed, we had to make changes to the contract. This is characteristic of major embankment construction around the world.

In the depth of the first winter we installed a million dollar grout curtain. Ken Weaver came to the site to guide us. Many years later he wrote the definitive book: *Dam Foundation Grouting*. Published in 1991, it is still available at a cost of $393.96. The update 2007 version is only $112.00, and it is still used in the industry.

Others from SRK came to the site to help. They worked on the site geochemistry, the design and construction of the surface water management facility, and they provided assistance with the regular soils testing we did in our small on-site laboratory---a few pieces of geotechnical testing equipment in a rented trailer.

And so the embankment grew ever higher (fig. 2). Its design and construction are described in papers that are readily available on the Cannon Mine website.

The mine produced about 1.25 million ounces of gold and two million ounces of silver, and was the second largest underground gold mine in the United

Figure 2. An aerial view of the Cannon Mine Tailings Facility.

States. The underground operations were closed in 1994, and were plugged, regraded, and replanted (fig. 3). The mine buildings were converted into offices and an equipment maintenance facility for the local school district.

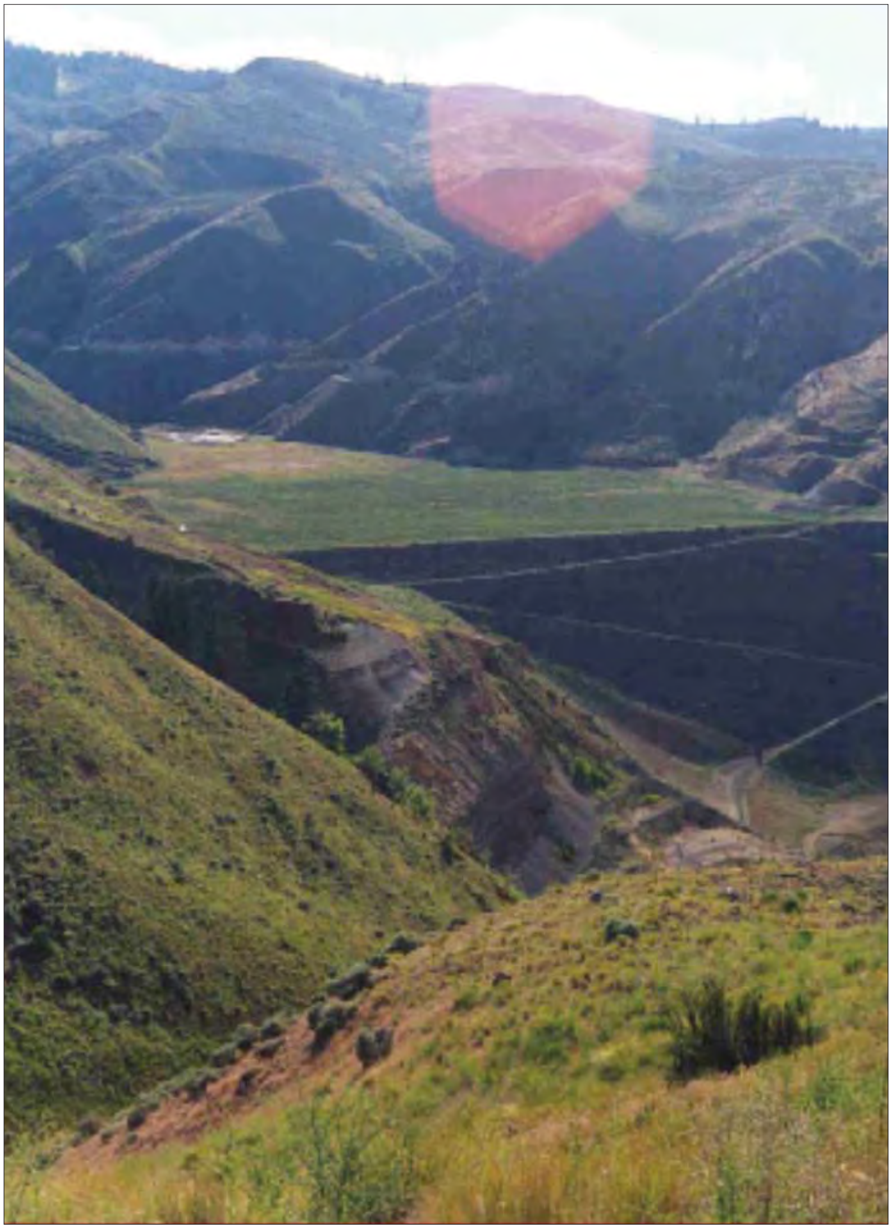
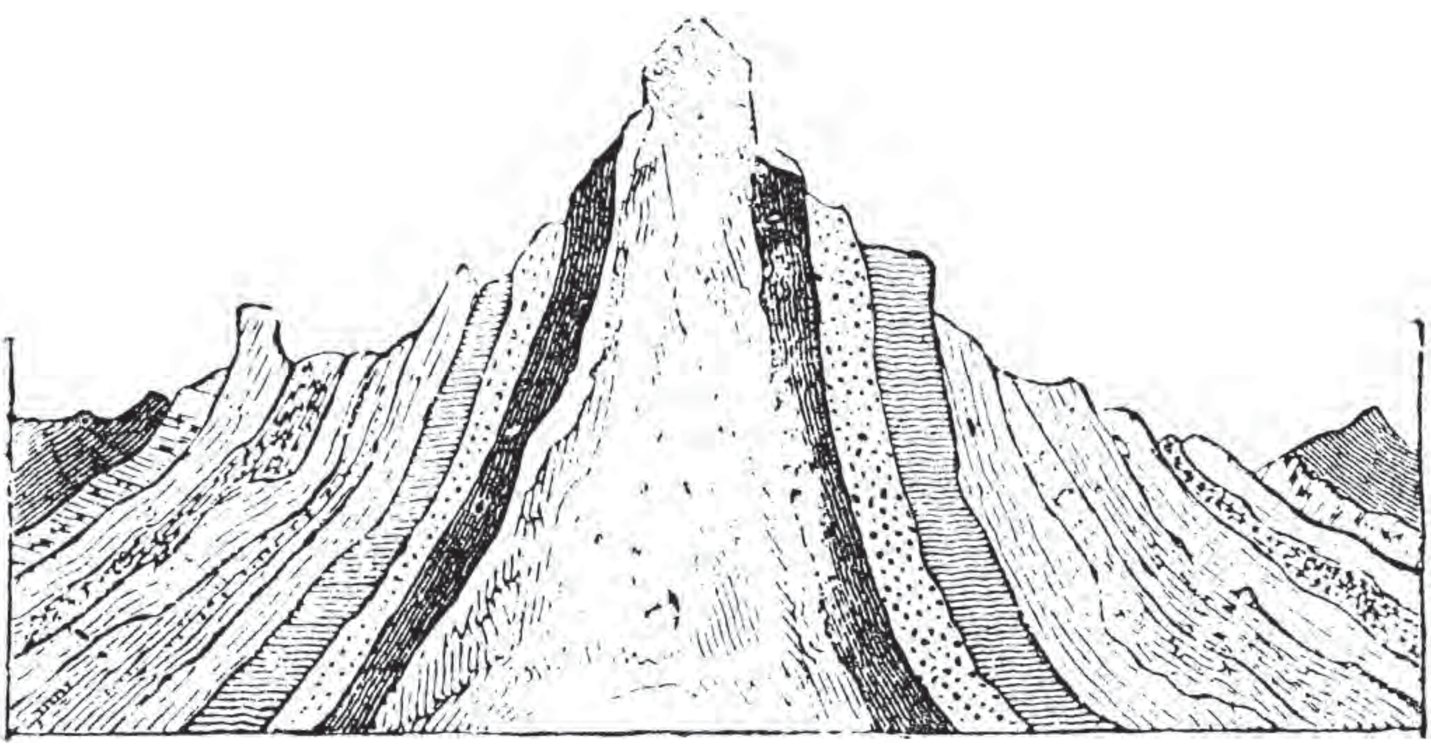


Figure 3. Cannon Mine Gold Tailings Facility after reclamation.



MOUNT TOLMAN MINING HISTORY

Skye Cooley

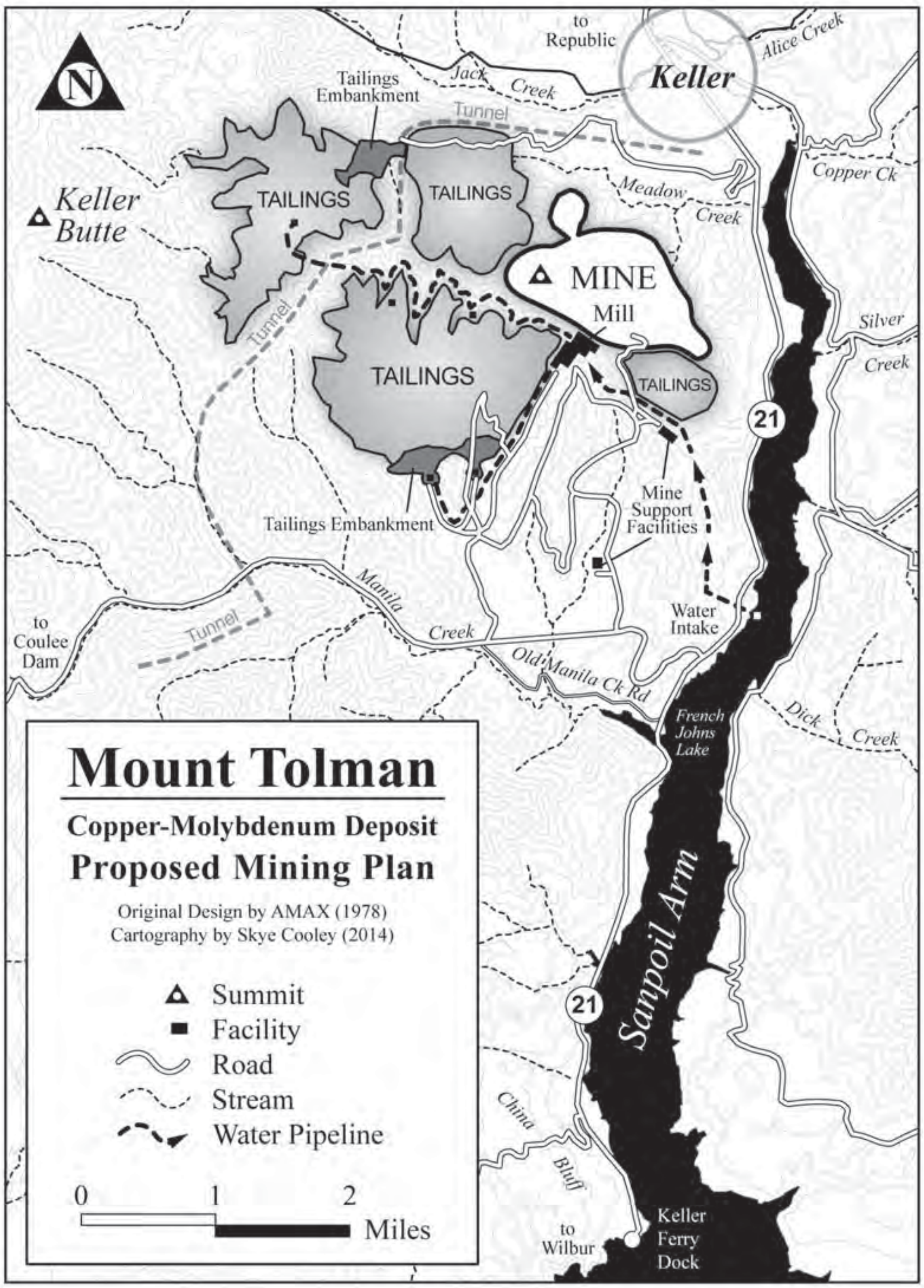
Cirque Geoscience & GIS Consultants
Adjunct, Boise State University Geosciences Department
Boise, Idaho

SUMMARY

In 2005, a worldwide shortage caused the price of molybdenum to increase from $5 to $33 per pound. Overnight Mount Tolman became worth 20 billion dollars. Mount Tolman is located on the Colville Indian Reservation 40 miles south of Republic and 95 miles northwest of Spokane. The porphyry copper-molybdenum ore body is hosted in fractured, hydrothermally-altered granodiorite of the Colville Igneous Complex (51-60 Ma). The mountain is located at the south end of the Eocene-age fault-bounded syncline known as the Republic Graben.

Mineral exploration from 1896 to 1918 on a dozen small claims around the base of the peak hinted at a rich and geologically complex deposit. Subsequent testing by the Mount Tolman Gold Company in the 1930s sparked the interest of investors, but no mine was opened. A subsequent round of exploration in the 1960-70s by the Bear Creek Mining Company better constrained the extent of the body and bolstered confidence of its potential.

By 1982, the AMAX Corporation had completed a full Environmental Impact Statement (EIS) on the proposed project, won approval on the bulk of the engineering, built roads, and was busy analyzing over 350,000 feet of core. AMAX initially estimated the 1,200-foot deep, 1½-square mile ore body to contain 900 million pounds of molybdenum and 1,100 million pounds of copper. According to the Washington Division of Geology and Earth Resources, the Mount Tolman deposit



"contains the third or fourth largest molybdenum reserve in the United States, with 2.4 billion tons of ore averaging 0.093 percent MoS2 and 0.09 percent Cu" (WDGER website, 2014). Both the company and the Colville Tribe could reasonably expect a multi-billion dollar return over the life of the mine if these projections held true.

In 1982, AMAX pulled out suddenly in response to a precipitous downturn in the moly market. The Mount Tolman Project never advanced beyond the final stages of planning and the tide of public opinion turned. In 1994, lingering doubts were made concrete with the passage of a Reservation-wide moratorium on exploration and mining. A 2006 referendum vote upheld the mining moratorium. Despite its promising geology, remote location, struggling regional economy, and increased global demand for molybdenum, the mountain remains as quiet today as it ever was.

EARLY EXPLORATION

Discovery of lode gold in north-central Washington occurred in the 1870s, replacing dwindling placer deposits discovered 20 years earlier by Chinese settlers (Collier, 1907). Molybdenum was first discovered at Crown Point near Lake Chelan in 1897 (Minobras Mining Services and Research, 1979). News of fresh mineral discoveries in the vast Okanogan Highlands region drew the attention of investors and governments alike at the turn of the century. Volume 1 of the Washington State Geological Survey Bulletin describes the geology and ore bodies of the Republic Mining District (Umpleby, 1910) as does an important follow-up paper by Bancroft (1914). A survey of the Colville Indian Reservation by J.T. Pardee culminated with the publication of USGS Bulletin 677 (Pardee, 1918) and the first geologic map of the area surrounding Mount Tolman. Pardee, recognized today for his contributions to the geology of the Glacial Lake Missoula floods, noted the considerable travel and transportation difficulties the Columbia River gorge imposed upon miners in the area.

Columbia River crossings designed to carry industrial traffic are few in north-central Washington. For more than a century, the ranchers, loggers, and miners of the region have crossed the river at the same locations by barge or ferry. Ferry service across the Columbia gorge at the mouth of the Sanpoil River (Keller

LARGE TONNAGE IN MT. TOLMAN

WILBUR, Wash., April 11.—A report of George W. Bartholomew, as a geological engineer, told officers of the Mount Tolman Gold company, that the quartz-poryhpry formation indicates great depth to the ores and he gave his written opinion, covering the company's holdings and adjacent area, that there is enough ore to operate on a large scale for a long period, according to President Snyder.

"A mining engineer representing a major railway system of the United States, whose name we have not the privilege of using now, inspected that part of our holdings where the major development has been effected," said Snyder recently. "He told me December 12, while examining the face of the Lincoln tunnel, that, in his opinion, the tunnel face was within 30 feet of our main ore body, and added that he would not be surprised if we found a low grade body of pay ore across virtually all of the next 600 feet directly ahead.

"He had examined the surface between the nephelite-syenite and the diorite dykes which are approximately 600 feet apart, the present tunnel face being up the south side of the first named dyke.

"This property of 39 claims, extending from the western slope of Mount Tolman peak nearly to the head of Meadow and Jack creeks, in Ferry county, about 14 miles northeast of the Grand Coulee dam, lies in a continuous body. It has many outcroppings. John C. Hamond, a mining engineer, is directing the development."

President-Manager Alva Snyder announced that the Hammond vein had been cross cut at a vertical depth of 310 feet at a distance of 497 feet from the portal of the Lincoln tunnel. He is taking samples to the assayer, and is confident of good values in gold, silver and copper. The vein is 20 feet wide, with the gangue chiefly quartz.

Much inconvenience because of water and not a little danger was experienced during a month of tunnel driving by Mr. Snyder and his son, Paul. The pressure was so great that water gushed from the drilled holes. He was determined to keep his promise to stockholders that Lincoln tunnel would be driven to the Hammond vein this spring, and not being able to get miners he operated the jackhammer himself. Two weeks he worked alone at night, and five of his shifts averaged 16 hours and 20 minutes.

Ferry) has been ongoing in one form or another since the late 1800s. Washington State began operating the crossing at Keller Ferry in 1930. The original cable barge was replaced by a side-wheeler in the 1930s, then a tug-and-barge in the 1940s.

The State continued crossing service during construction of Grand Coulee Dam and the filling of Lake Roosevelt. The state ferry *M/V Martha S* began service in 1948. The 80-foot, twin-diesel vessel and its two-man crew made the 1.25 mile crossing 40 times a day, 7 days a week, for 65 years. The new ferry, the *M/V Sanpoil* (116 feet long, with 20 cars), a one-of-a-kind ship built by Foss Maritime and operated by Washington State Department of Transportation, assumed the route in 2013. The ferry is free to all cars and passengers. The crossing links sections of SR 21 from landings located fifteen miles north of Wilbur, Washington, and ten miles south of Keller, Washington.

The “North Half” of the Colville Indian Reservation (Hodges, 1897) was opened to mineral exploration in 1897 when Congress rescinded Indian ownership, established by treaty decades earlier. Ore reserves were believed to be so rich and plentiful that the reduction of Reservation land by half was justified in the eyes of the Federal government. Today, the Tribes enjoy hunting, fishing, and gathering rights on the North Half, but no longer manage the lands.

OPENING OF THE SOUTH HALF

In 1898, the Reservation’s “South Half” (the modern Reservation), was thrown open to non-Indian prospectors, who quickly staked some 3,000 claims in the Keller-Sanpoil, Nespelem-Moses, Covada-Meteor, and Park City districts (Pardee, 1918). Distance to the LeRoi copper smelter at Northport, Washington was considerable. A 100-ton copper smelter was erected at Keller by the Keller and Indiana Consolidated Smelting Company, but never opened. Another at Turk,WA (now called Fruitland) failed (Bancroft, 1914). Few were able to profit from their South Half claims. USGS geologist J.T. Pardee comments:

“Little more than annual assessment work was being done in the Sanpoil in 1912. Most of the claim owners were marking time, awaiting the coming of railroads, of which one or more have been projected down the Sanpoil Valley. Under present conditions, transportation is too expensive to permit the profitable working of the lodes except to a very small extent” (Pardee, 1918, p. 105-106).

In 1932, the Sunnyside Sun published an article titled, “The Largest Deposit of Molybdenum in the World” (Sunnyside Sun, May 19, 1932). The piece did not describe Mount Tolman, rather Moses Mountain, also located on the “South Half”, some 25 miles to the northwest near the town of Omak.

MOUNT TOLMAN GOLD COMPANY

News reports indicate the first ores from Mount Tolman were assayed by C.M. Fassett in Spokane at Idaho’s Bunker Hill smelter (Spokane Chronicle, November 12, 1931). The Mount Tolman Gold Company (MTGC) was established in 1934 by Alva Snyder (President), C.A. Drinkard, and A.R. Gregory from an initial capital investment of $10,000. Between 1935 and 1937, the Spokesman Review published a flurry of short, enthusiastic articles describing the exploration successes of the company.

Test holes and tunnels into nepheline-syenite and diorite yielded economic concentrations of gold, molybdenum, silver, and copper. By all accounts, the Mount Tolman deposit appeared to be large and continuous, though water production was a problem in some of the tunnels. The Mount Tolman property had “an area of 820 acres and is penetrated by 14 tunnels having a total length of 2,150 feet.” (Spokesman Review, November 10, 1935).

MTGC at their high point aggregated 39 claims, including:

- “Jannot”, 22-feet thick, with gold, silver, copper
- “King Richard”, 27-feet thick, with gold, silver, copper
- “Clay”, 9-feet thick, with silver
- “Emerald”, 4-feet thick, with molybdenum
- “Ivanhoe”, 3-feet thick, with gold, silver
- “Bluebird”, with gold, silver, copper
- “Hammond”, 20-feet thick, with gold, silver, copper

At that time MTGC staff included Frank V. Taylor, Vice President; Carol C. Snyder, Secretary; and

William S. Kirkendall, Treasurer; John C. Hammond, mining engineer; and George W. Bartholomew, Geological Engineer.

Other noted claims from MTGC and others include the “Manila”, “Last Chance”, “Umatilla”, “Dewey”, “Rover Bonanza”, “Addison”, with lead, copper, zinc, silver, and gold, “Walla Walla”, and least seven notable, but unnamed ore bodies exposed in the Lincoln and Twin tunnels. Enthusiastic assessments by several promoters, including Colonel H.C. Demming, (former) State Geologist for Pennsylvania; F.W. Callaway, from Bunker Hill and Sullivan; R. King, mining engineer from Spokane; and J.C. Hammond, mining engineer from the town of Inchelium, WA, sought to generate additional outside investment (Spokesman Review, November 10, 1935). Pardee also observed, “The volume of mineralized rock is so great that a very small profit to the ton would render the deposits extremely valuable” (Pardee, 1918, p. 116).

CONSOLIDATED MINING AND SMELTING COMPANY, COMINCO

Cominco, now Teck Resources Ltd., began in 1906 as The Consolidated Mining and Smelting Company of Canada (CM&S), or by its nickname, “Smelters”. CM&S acquired 37 Mount Tolman claims from MTGC and others between 1931 and 1937. These included the “Glen Lord”, “Advance”, “Silver Ridge”, “Iconoclast”, and “California” claims. An economic copper body was located in 1945, but CM&S was limping along and ultimately failed to produce it.

The Bureau of Indian Affairs (BIA) and Bureau of Land Management (BLM) reassessed the company’s claims between 1954 and 1964. The agencies invalidated the claims on the basis of their uneconomic and “non-mineral” character. CM&S sued the BIA in Federal District Court in Spokane, charging “errors, mismanagement, and cover ups” by agency staff. CM&S lost the suit and their subsequent appeal to the Ninth Circuit Court in San Francisco (Spokane Daily Chronicle, Firm Blames Agencies, Feb 26, 1981).

Shamrock Lead Mines, Inc. operated from 1935 to 1937 at Mount Tolman, targeting gold, silver, lead, and nickel. Shamrock had a cabin-compound near Keller, and employed 18 people in 1935. They employed only 12 in 1936. Drilling for Shamrock was done by Kirkbride Engineering Company of Spokane. In 1937, operations were overseen by R.S. Welles and later Eldon Davidson, both of whom were mining engineers. Shamrock held 50 local claims on 1,000 acres in 1937. James E. Angle managed the mine for Shamrock, employing as many as 17 miners. Its board of directors, established in 1937, included G.B. Walker, Seattle; A.A. Bowman Yakima; J.E. Angle, Secretary-Treasurer; L.B. Walker, trustee; and J.R. Abraham, trustee.

Various small outfits worked claims at Tolman for brief periods. Pacific Mutual Silver and Lead operated a few claims at Mount Tolman in the 1930s, though little of its history was recorded. The Molybdenum Mines Company of Omak employed W. B. Hancock, a mining engineer from Seattle, to conduct exploration at Moses Mountain in 1931. Bear Creek Mining Company, a subsidiary of Kennecott, was another early entry to exploration at Mount Tolman.

MODERN EXPLORATION AND MINING ON INDIAN LANDS

Three acts of Congress cover modern mining on lands managed by Indian tribes recognized by the Federal Government. The initial legislation of 1909, permitted allotted lands (lands owned by individual Indians) to be leased for mining (35 Stat. 781-783; 25 USC 396). A similar act was passed in 1935 for unallotted lands, lands typically considered Reservation lands (52 Stat. 347; 25 USC 396 a-g). In 1982, the Indian Mineral Development Act (96 Stat. 1938; 25 USC 2101-2108) empowered Tribes to negotiate directly with mining companies and decide their own agreements on exploration, development, leases, royalties, rights, and all their related dealings. In 1977, Congress approved funding to assess the mineral potential of Indian lands in hopes of promoting “economic well being by means of mineral development”. Eleven tribes published articles on geology, permitting, and opportunities in a report by Manydeeds and Smith (1991).

No new exploration on the Colville Reservation has taken place since the early 1990’s. There are just two small mines on Mineral Ridge that are somewhat active today (Hurst and Passmore, 2010, unpublished report). Factors are remoteness, difficulty of travel, and a population leery of mining impacts, though to the last point, Gary Passmore, Director of the Colville

Tribes Environmental Trust Program, states, “Water quality and sediment analysis from areas with past mining activity have revealed little or no pollution emanating from old mine workings and tailings material” (Gary Passmore, personal communication in Office of Environmental Trust Report, 2011, p. 29). Also, the initial 1994 tribal moratorium (Resolution 1994-561), which halted all exploration and mining activities on the Reservation, did not restrict data collection under BIA 638 funding, tribe-sponsored geology projects, or walk-on permits activities initiated by tribal members.

GEOLOGY OF THE ORE BODY

Mineralization at Mount Tolman occurs within an east-west oriented zone that covers an area approximately 4 miles long, 1.5 miles wide, and 1,000 feet thick. Molybdenite, copper, and pyrite are accompanied by magnetite, chalcopyrite, rutile, sphalerite, and galena. A thick phyllitic cap overlays the stockwork deposit. Steep, north-trending normal faults, associated with the Republic Graben, offset the ore body tens to hundreds of feet. A wide band of quartz porphyry trending northeast across the southern slopes of Mount Tolman is believed to be the source of the ore fluids (BIA, 1983). Numerous small claims pursue quartz-zinc-copper-lead-silver-gold trends that surround the main ore body. Uranium was discovered on the Spokane Indian Reservation, some 30 miles east of Mount Tolman in similar intrusive rocks (NUEXCO, 1979; Midnite Mine).

Porphyry deposits are ore bodies formed by the complex interaction of intruding magma, dikes, and fluids circulating near the contact with the surrounding rock. Ores of copper, molybdenum, gold, and silver are commonly found in networks of stockwork fractures and veins within this hydrothermally-altered zone. Porphyry deposits form at relatively shallow depths in the crust (0.5 to 2.5 miles) and are associated with A-type granitic magmas of intermediate to silicic composition.

The Mount Tolman Granodiorite is part of the Colville Igneous Complex (CIC). Calc-alkaline geochemical signatures have led many to interpret the CIC as a subduction arc, in the past called the “Challis Arc” (Muessig, 1962; Lipman, 1972; Atwater and Rinehart, 1984; Moye, 1984; and Holder and Holder, 1988). More recent work on extrusive rocks in the Sanpoil Volcanics and Klondike Mountain Formation complex suggest an alternative origin for the CIC, one involving decompression melting driven by post-Laramide collapse of overthickened crust (Morris et al., 2000).

Above an elevation of 1,700 feet, slopes at Mount Tolman are blanked by deeply-weathered granitic soils and colluvium. Below this, varved lake beds of the Nespelem Silt (Pardee, 1918) fill valleys and form remnant benches along streams. The late Pleistocene Nespelem Silt was deposited in Glacial Lake Columbia. Highly impermeable and landslide-prone lake beds are 400- to 600-feet thick in the valleys of Meadow, Manila, and Last Chance Creeks. Coarser-grained flood deposits from the Glacial Lake Missoula outburst occur as interbeds in the silt and cap silt terraces locally (Atwater, 1986). Glacial meltwater gravels are also present at lower elevations.

STRATEGIC MOLYBDENUM IN DEMAND

The name molybdenum derives from the Ancient Greek word molybdos, meaning lead. The primary ore mineral is molybdenite (MoS_2), though it is found in wulfenite ($PbMoO_4$) and powellite ($CaMoO_4$). The silvery-gray ore was often mistaken for galena (PbS) and sometimes graphite (C). Molybdenum was first discovered in 1778 by the pharmaceutical chemist, Carl Wilhelm Scheele, a German-speaking Swede who was a pioneer of early chemistry. The dark metal powder was first isolated in 1781 by another Swede, the chemist Peter Jacob Hjelm. Scheele is credited today with isolating oxygen, molybdenum, tungsten, barium, hydrogen, chlorine, several common organic acids (tartaric, oxalic, uric, lactic, citric), and a few inorganic acids including hydrofluoric, hydrocyanic, and arsenic acids. He was, however, scooped in publishing many of his findings by two Englishmen, the natural philosopher Joseph Priestley and chemist Humphry Davy. Those damn Brits are always knocking off.

Molybdenum, like platinum, cobalt, and chromium, is a “strategic” metal due to its widespread use in manufacturing for the defense and electronics industries. Worldwide demand for molybdenum is increasing and while the market for moly is volatile, few discoveries of new ore bodies have been made in recent decades. The future for the graphite-gray metal is bright.

Molybdenum's special properties include a high melting point (2,623° C, 4,753° F), strength retention at high temperatures, thermal conductivity, and resistance to corrosion. It has a low coefficient of expansion and is a superior alloying element. Moly is used to harden and lighten steel, as a dye, as a catalyst in the petroleum refining, as an additive to lubricants, and in innumerable electronic components. The metal is of moderate hardness (Mohs 5.5). Froth floatation is used to isolate and recover moly.

Despite its widespread use in modern industrial manufacturing, moly production started slowly. Greeks and Romans recognized the lead-like mineral (Tooker, 1991). Master sword makers of 14th century Japan are reported to have used moly as an alloying agent, but western factories would scarcely notice it for nearly 100 years after its isolation. In 1891, the French group Schneider and Company first used molybdenum to make armor-plate steel. With World War I (1914-1918) came the first industrial-scale demand for moly for steel used in high speed tools, tougher armor plate, and well casing.

The Climax mine opened near Leadville, Colorado, in 1915 and the Questa mine near Taos, New Mexico, in the 1920's to help meet the demand. Demand increased again during World War II (1939-1945) when it was discovered that its lower molecular weight (95.94 g/mol) lightened steel considerably more than tungsten (183.84 g/mol). Forging and heat treating techniques for Mo-alloyed steels were perfected in the 1930's, opening many new markets. Today, China possesses the largest molybdenum reserves and is its top producer (Zeng, 2013).

The copper-molybdenum porphyry deposit is hosted in faulted granodiorite and cut by dikes of intermediate composition related to extensional volcanism and the opening of the Republic Graben. It is an elliptical ore body formed within a former subduction arc where low-Fluorine (<0.1%), calc-alkaline magmatic and hydrothermal mineralization probably occurred during the early stages of magmatism. Early descriptions often mention seams in the wallrock coated with molybdenite and pyrite, "Molybdenite and part of the chalcopyrite, pyrite and quartz appear to have been first introduced, followed by a second generation of the same minerals except molybdenite…" (Pardee, 1918, p. 112).

The Mount Tolman deposit is similar to deposits at the Endako Mine (Fraser Lake, BC), the Ruby Creek Mine (Atlin, BC), Buckingham and Pine Nut Mines (Lander County, NV), Quartz Hill Mine (southeast AK), and Thompson Creek Mine (Challis, ID) (Cox and Singer Model 21b, 1986; Tooker, 1991; Westra and Keith, 2008). Copper and molybdenum sulfides occur as fracture fillings and as veinlet stockworks within and adjacent to intrusive bodies. The ore is identified by pyrite, chalcopyrite, and scheelite, with quartz as the primary gangue mineral.

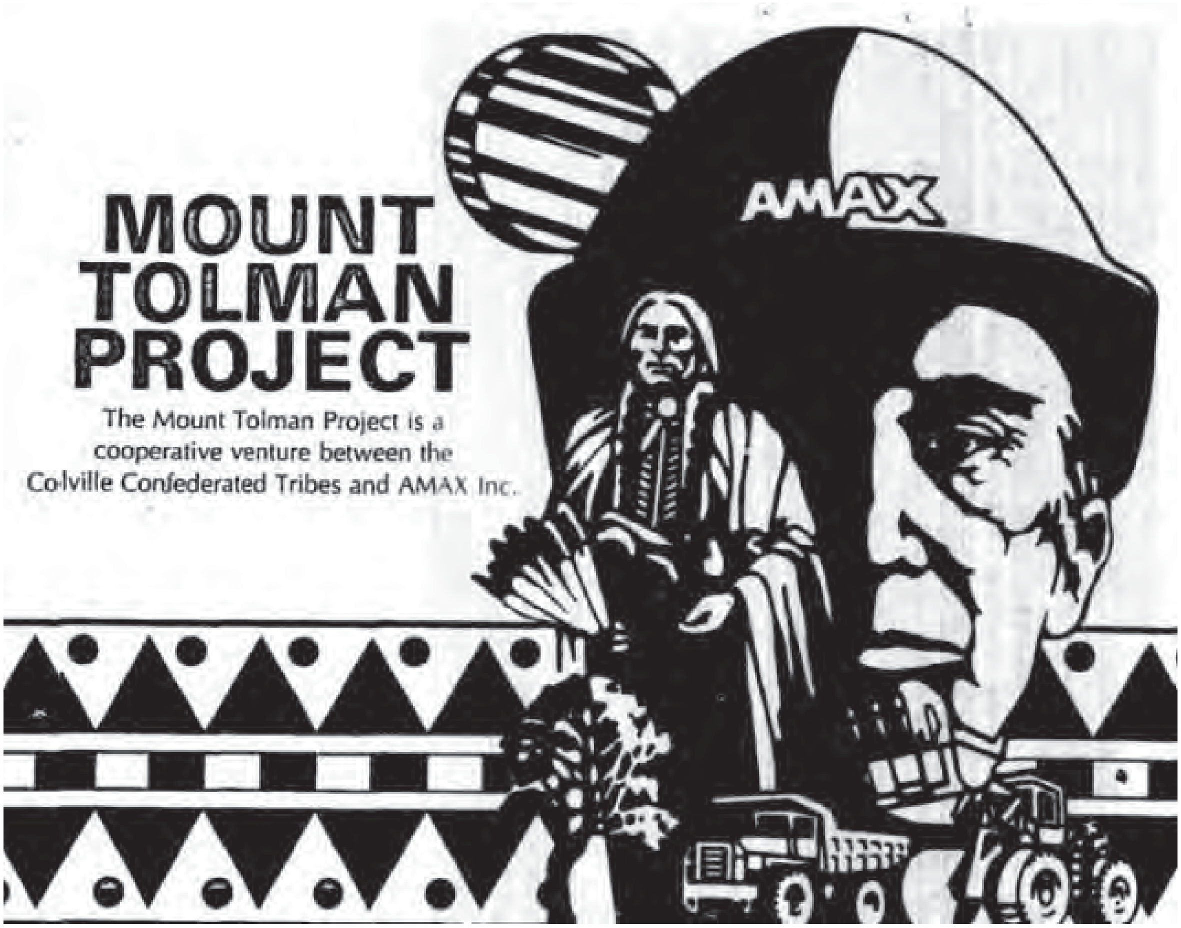
PROSPECTING BY COMPUTER

Some of the earliest predictive computer modeling of an ore deposit was performed on data from Mount Tolman (Bell, 1982; Campbell et al., 1982). PROSPECTOR was state-of-the-art software that used geologic mapping, geochemical data, core logs, and expert knowledge to predict the size of the ore body in the subsurface. The model described it as a "plate-like rind topping a large batholith" slightly larger than that estimated by a traditional geologic survey.

THE AMAX YEARS: 1978–1982

In the late 1970s, the AMAX Corporation of Greenwich, Connecticut, was expanding its molybdenum interests into Colorado, British Columbia, and Washington. In 1978, 18 companies initially submitted mining proposals to the Colville Tribe, including one from the Bear Creek Mining Company, a subsidiary of Kennecott Minerals Company, who had been assessing the deposit since 1964. AMAX won the contract and the company ramped up exploration immediately. The company moved fast, conducting both geologic exploration and engineering studies at the same time. During the three years of exploration at the site, they spent $50 million on equipment, drilling, and testing. Another $20 million went to the Tribe and its members in the form of direct payments to tribal programs.

** Note: AMAX merged with Cyprus Minerals Company in 1993 to form Cyprus-AMAX Minerals Company. Cyprus-AMAX was a leading producer of molybdenum, lithium, copper and coal worldwide. In 1999, the Phelps Dodge Corporation acquired Cyprus-AMAX. Subsequently, Freeport-McMoRan Copper and Gold, Inc. (FMCG, www.fcx.com) of Phoenix, Arizona acquired Phelps Dodge in 2007.*



Associates, 1980) with recirculating plumbing.

The tailings impoundment design called for a 600-foot high, compacted embankment from lake bed material to be built atop a 300-foot high rock embankment. It would have been one of the largest such tailings structure ever built. The initial design was reviewed and reworked by Dr. Geoff Blight, from the University of the Witwatersrand in Johannesburg, South Africa, and the preeminent expert on earthen dams, Arthur Casagrande (Harvard University). The original engineer, Jack Caldwell (ithinkmining.com), recalled "I was eviscerated by [Casagrande] for my early mistakes". The improved design was approved by Roy L. Soderberg of the Washington State Bureau of Mines and the BIA.

The terms of agreement with the Tribe stated that AMAX would spend $8.5 million on an EIS (Bureau of Indian Affairs, 1980). Upon successful completion and adoption of the EIS, AMAX would pay an additional $8.5 million to the Tribe. In return, AMAX would be granted a tribal mining permit and have first rights to a federal lease at Mount Tolman.

By 1979, a handful of small drill rigs had completed 160,000 feet of diamond core drilling. By 19781 they had completed 360,000 feet. Core drilling to a depth of 1,200 feet was typical. Some 13 miles of roads had been constructed on the mountain to facilitate the exploration.

Mining and milling was slated to begin in 1986 once the exploratory drilling, metallurgical testing, processing facilities construction, and worker training was completed. The pit-filling lake would remain at the summit after the mine closed. Waste rock and tailings would fill and level about 2,800 acres of the adjacent valleys of Manila, Meadow, Jack, and Last Chance Creeks, and Consolidated Basin. Tailings would be contained by unlined earth-rock embankments (Robertson-Pincock, 1980; Robinson and

A 194,000 square-foot concentrator mill was to be built on the mountain. The concentrate would be trucked out of the area for smelting, though no smelter was identified to receive it. Water for the mill would be drawn from the Columbia River at 8,000 gpm via a new pump station located two miles from the mine. No water from the operation would be discharged back to the Columbia or its tributary the Sanpoil River; rather it would be recovered and recycled through the mill or used for road-dust suppression (Hydro-Triad, 1980b).

In total, some 3,700 acres would be disturbed by the operation at Mount Tolman. The Final EIS, by Beck Consultants, did not address the potential extraction of the gold, silver, zinc, or lead also known to be present at the site.

The mine was expected to produce 60,000 tons per day and remain active for 43 years. Mining would bring about 400 long-term jobs to Ferry and Okanogan Counties, two of the poorest and least populated counties in the Pacific Northwest. The Tribe would not see profits immediately, but could expect sizable returns after 10 to 15 years. If moly prices remained at $15 per pound (three times its average price over recent decades), the mine would make a projected $90 million per year, or $10.3 billion over 43 years. The lease specified the Tribe would receive the larger of a) $5 million per year, b) 5 percent of the revenue, or c) 50 percent of the operating margin.

Promotional advertisements for the Mount Tolman Project were taken out in local newspapers and college scholarships were granted on behalf of the project to students from Lake Roosevelt, Pateros, and Ridgefield High Schools (Quad-City Herald, May 28, 1981). New structures were built at the Nespelem Indian Agency Headquarters and at the base of the mountain near Keller. The Mount Tolman Fire Center and Colville Tribe's Keller Forestry District offices occupy the site today.

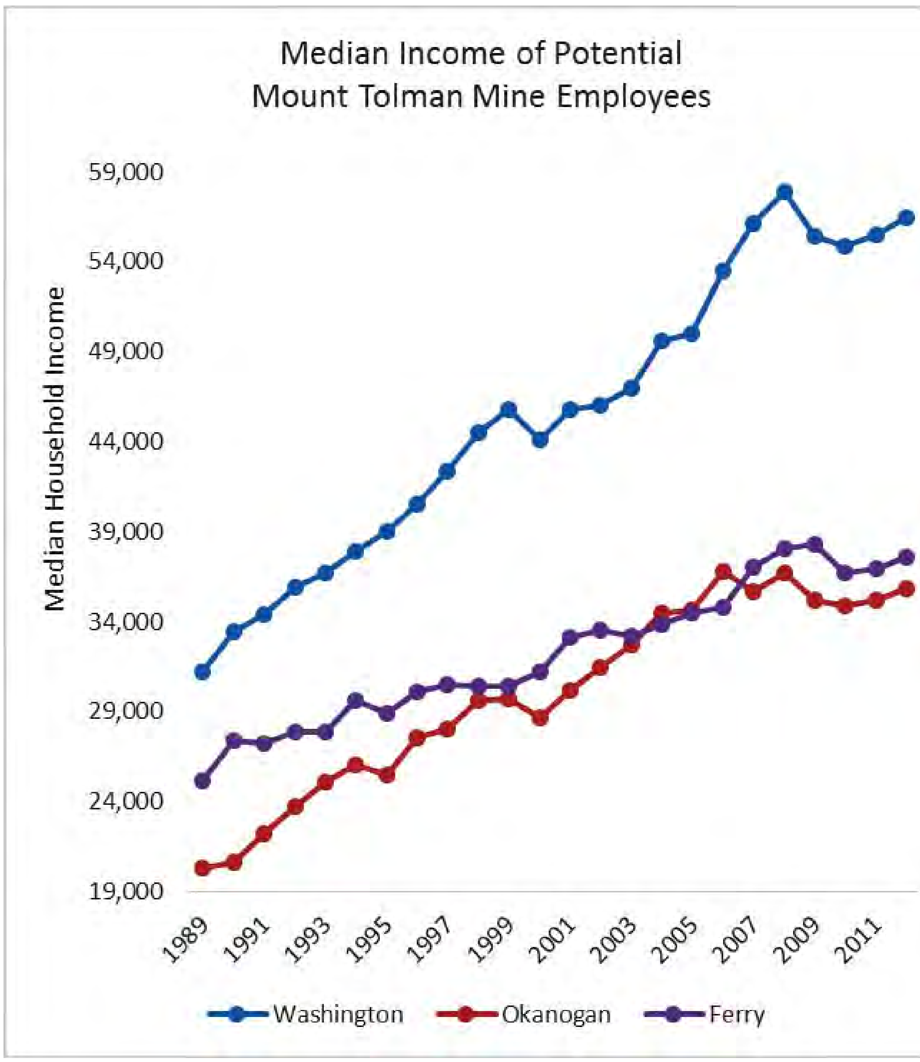
A drop in the price of copper and molybdenum abruptly halted development at the mine in 1981. Moly dropped from $8 to $3 per pound. Rumors that AMAX would abandon the project began to circulate. The company initially denied the rumors, but by late 1981 they had placed the project on "caretaker" status and pulled out.

** Note: Don Aubertin, a tribal member, was the director of mine engineering for the Colville Reservation at the time. William Utterback was AMAX's Chief Geologist. Dale Kohler was Chair of the Tribe's Land and Forestry Committee. Charles E. Stott Jr. was the AMAX General Manager for the Mount Tolman Project. Stan Dempsey was a Vice President at AMAX. Les Darling was the Project Environmental Manager for AMAX. Pierre Gousseland was Chairman of the Board for AMAX.*

RENEWED INTEREST IN MINING: 2004–2006

A 2005 spike in moly prices reinvigorated interest in the Mount Tolman deposit. Prices soared to $45/pound in 2005. The Colville Business Council hired Don Aubertin, from Lakewood, Colorado, who had been involved during the AMAX years, to prepare educational materials on a new mining plan for evaluation by tribal members. The proposed reopening of Mount Tolman to mining was politically charged and garnered considerable interest. Many tribal and non-tribal groups got involved, appealing for sunlight on the decision process in which the Colville Business Council was engaged.

The following are excerpts from letters written to The Eagle Review newspaper by the Chairman of the Business Council, D.R. Michel. They show his efforts to publicly address the concerns raised by tribal members.



"As you can see by [Colville Business Council] Resolution 2005-305 Don Aubertin did not receive $100,000 dollars. The Resolution funds a Mining Referendum Project Proposal. Don is involved because of his past experience with the project back in the 1980s. He is under contract to help put together information on the pros and cons of mining the Mt. Tolman Project. There will be a series of meetings set up in the near future. There will also be a packet of information sent out to tribal members. This packet will show the footprint of the project, the projected revenues, and jobs involved. It will also show the environmental risks and concerns that the project will entail."

"The education process will focus on the Mt. Tolman Project. The mining and milling process will be explained; it will show there is no cyanide used in the processing of the ore on site. It will also address the environmental concerns. After the education process is complete there will be a referendum vote on whether or not to go forward with the project. I feel it is the right of all eligible voting Tribal members to cast a vote. We are recommending that the polls be opened and absentee ballots are available for voting. That decision should be based on the facts, not scare tactics. "

"If the referendum vote fails then mining on the reservation will remain closed. If approved we will move forward with the project. I want to ensure the membership that Council will do everything possible to make this the most environmentally protective plan possible. We will use only the most current technology available. The Environmental Impact Statement (EIS) must be updated to comply with current law. The EIS will hold a company accountable to the Environmental Protection Agency (EPA). EPA administers the Clean Water Act and Clean Air Act. In the EIS they must show how they will ensure water and air are protected. Also in the process a reclamation fund will be established up front that will cover reclamation during the early stages of mining, plus annual additions to the fund throughout the project will cover reclamation at any given time."

MORATORIUM ON MINING UPHELD

On March 18, 2006, tribal members went to the polls to vote on mining at Mount Tolman. Thirty percent of the 6,684 eligible voters participated. The referendum posed a clear choice:

YES: Lift the moratorium on mining and allow the Colville Business Council to pursue proposals from mining companies, received 847 votes (40%).

NO: Keep the moratorium on mining in place [CBC Resolution 1994-561], received 1,254 votes (60%).

**Note: In 1977, the Confederated Tribes of the Colville Reservation held a referendum vote on mining. 875 of 3500 eligible voters cast ballots. According to the much contested count, 567 (65%) voted in favor of mine development, 308 (45%) against.*

FUTURE OF MINING AT MOUNT TOLMAN

Despite its promising geology, remote location, struggling regional economy, and increased global demand for molybdenum, the mountain remains as quiet today as it ever was. If a mine is ever to be opened at the mountain, industry representatives need to both understand the desires of people of the Colville Reservation and target substantial investment in them over the long term. Two important views are captured in the following quotes:

"From talking to people, I would say it was more the old tribe members who'd voted against the mine, and it was the young people who were more in favor of it. And that's a worry, because they'll be back. I told my own kids this, I said, 'Be ready, because every generation, they're going to come back, and offer us even more money for that mountain.' Next time it'll be $50 billion."

- Soy Redthunder (tribal member) in Brian Schofield, 2009, Selling Your Fathers Bones, p. 303

"As a Tribal member I will cast my vote in support of the project. As a Council member I support the project because it will allow us to offer our future generations a secure foundation to build on."

- D.R. Michel, Former Chairman of the Colville Business Council

Author's Note: In 2006, I was the employed as the Soil Scientist for the Colville Reservation. The consensus around the water cooler was that the vote against mining skewed older and that most pro-mine members lived east of the Sanpoil Valley. I recall a sense of relief amongst the administration when the polls closed

without incident. The opening of a mine at Mount Tolman would have fundamentally changed life on the Reservation. My thought at the time was that the Tribe and its leaders were just not ready to embrace that magnitude of change. In the future, however, I think they will be.

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THE OBLIQUE CONVERGENCE OF NORTH AMERICA

John Whitmer

Issaquah, Washington

INTRODUCTION

The following discussion contains what I have always thought of as a pretty neat mechanism for shaping the North and South American continents. For me, the mechanisms of plate tectonics are exciting, and they help make the geologic story more logical and understandable.

ACCRETED TERRANES AND THE BEHAVIOR OF THE MANTLE

Figure 1 was taken from the cover of the July, 1996, issue of *Geology* magazine. It shows the deformation caused by the movement of the Caribbean and Scotia Plates. Russo and Silver performed seismic tomographic studies of mantle flow in those areas, and in the area west of South America. They found that the mantle flows northward offshore in the northern half of South America. In the southern half of South America the mantle flows southward. At the same time, the South America Plate moves inexorably westward. The South America plate sinks rapidly, and extends to a depth of about 400 km, constituting an effective barrier to the flow of asthenospheric mantle. To manage the increased pressure, the mantle must flow past the South America Plate. The asthenospheric mantle therefore flows around the northern and southern edges of the South American Plate, (see fig. 2). The pressure of the mantle impinging on that barrier is greatest at

Figure 1. Image from Russo, R.M., and Silver, P.G., 1996, Cordillera Formation, Mantle Dynamics, and the Wilson Cycle: Geology, Volume 24, p. 511–514.

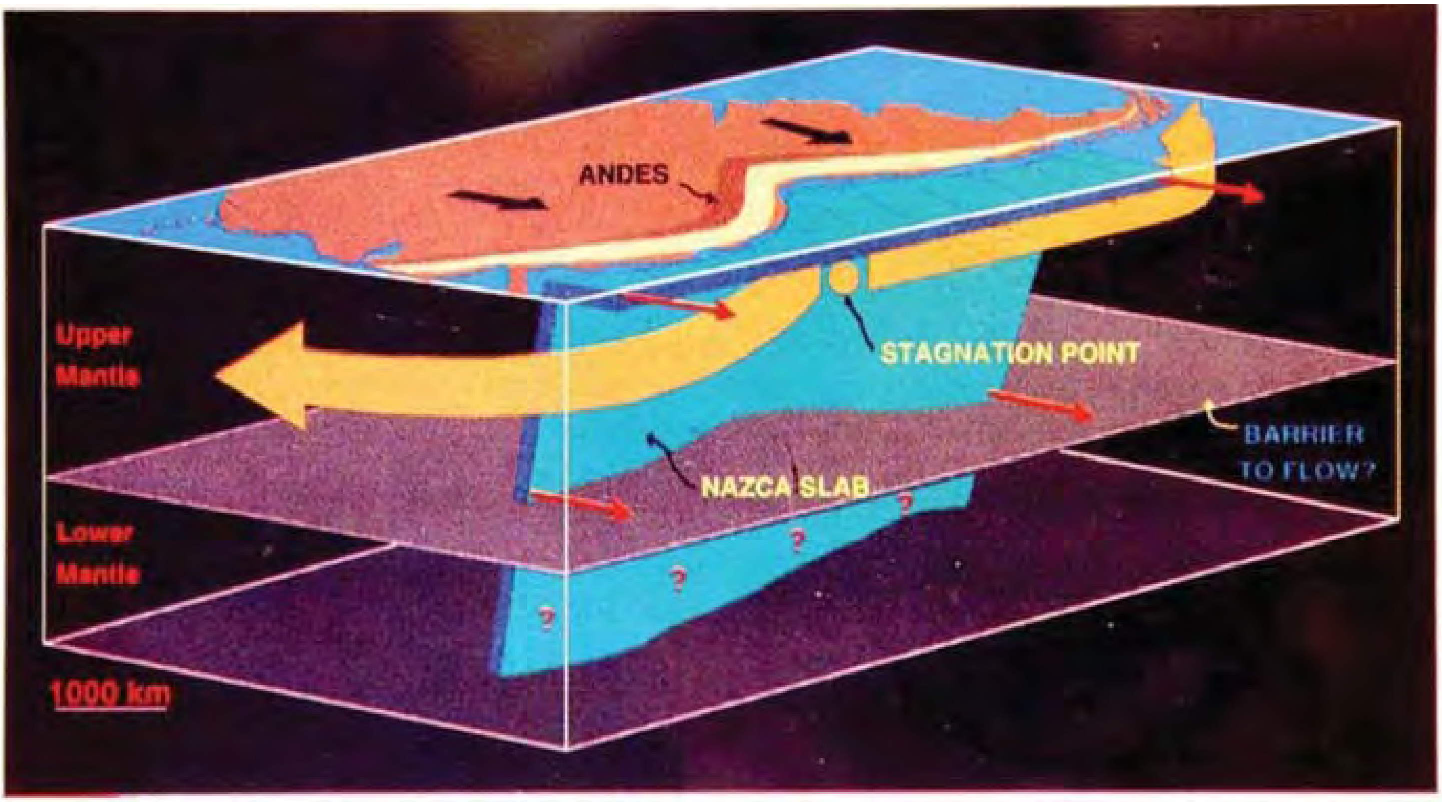

Figure 2. This figure represents a model of trench-parallel mantle flow beneath the Nazca slab (Russo and Silver, 1994). Flow is driven by westward retrograde notion of the Nazca slab (red arrows) and westward motion of South America (black arrows). Because fast shear-wave directions, and hence mantle flow directions, are parallel to the trench, we posit a barrier to mantle flow regardless of whether the slab penetrates into the lower mantle. The mantle flows from a region of high pressure coincident with a stagnation point beneath the central Andes, toward regions of lower pressure that lie beneath the Caribbean and Scotia Sea plates (orange arrows). The vertical exaggeration is 3.3-1.

the center of the converging margin. It retards roll-back of that slab and causes the indentation which is so prominent on the west side of the continent. The authors suggested that a similar condition prevailed on the west coast of North America during the Mesozoic, resulting in the indentation we now call the Columbia Embayment.

I developed this understanding from three papers by Russo, R.M., and Silver, P.G., published in 1994, 1995, and 1996. I think you would enjoy reading them.

Warren Hamilton (1969) and others (fig. 3) recognized more than thirty years ago that the Sierra Nevada Batholith, the Idaho Batholith, the Okanogan Mountains, and the Coast Plutonic Complex were parts of a volcanic arc which extended from Mexico to Alaska. They also recognized accreted terranes related to the arc, including the Klamath Mountains and mountains in eastern Oregon.

Those terranes and batholiths mark the location of a long, high, volcanic arc, which in the Mesozoic resembled the modern Andes Mountains. They inferred a coastline which was deeply indented, similar to that of the modern South American coastline. That indentation is expressed on our landscape as a lowland extending into Idaho, outboard from the Blue Mountains, the Okanogan Mountains, and the Rockies of Idaho and Northeastern Washington. It is known as the Columbia Embayment.

In figure 3, the trench of the Columbia Embayment strikes northwest-southeast. The imprint of that trench appears on the modern landscape as the Olympic-Wallowa Lineament (OWL). It is called a “lineament” because geologists have been unable to agree on what it means. Raisz (1944) was the first to recognize it. Raisz was an expert cartographer who drew pencil sketches of regions using the topographic data that was available. His maps were relief maps of the highest quality obtainable until the advent of powerful computers and the digital images they produce. His 1944 map of Washington showed a prominent trend extending in a northwest-southeast direction from the northern Olym-

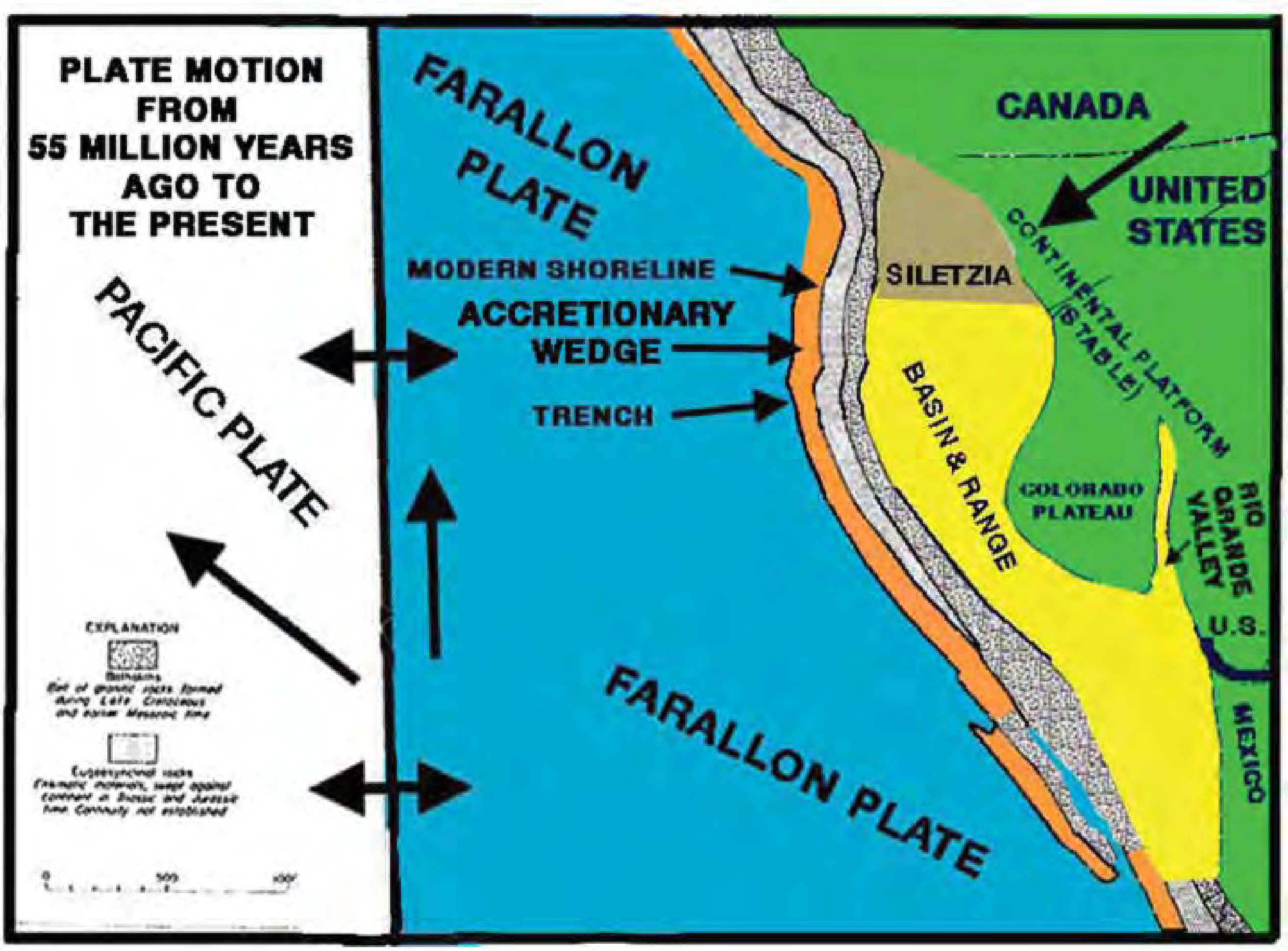

Figure 3: Plate Motion from 55 years ago to the present. The modern trench is shown at the edge of the white areas which were below sea level prior to 55 million years ago. The Baja California Terrane was east of the trench, and firmly attached to the mainland.

pic Mountains to and beyond the Wallowa Mountains in Oregon.

Geologists who looked only at the big picture ("Armwavers") saw large scale folds and strike-slip faults. Field geologists, who walked over the landscape carefully observing and documenting the nature of the rocks, saw only a series of parallel (en echelon) normal faults. Field geologists believe only what they can see, feel, and measure on the ground. They scorn "armwavers." Consequently, there are two factions having nothing to do with one another. They will never agree until enough data is accumulated to settle the argument. So, the OWL is still known as a lineament. Because of its ancestry in the old subduction zone and trench, it marks the boundary between the older continental crust to the north, which is thicker, cooler, and more rigid, and the younger, more fragile oceanic rock, which is thinner and warmer, that has been accreted to the south.

PLATE MOVEMENT

Note that in figure 4, the oceanic plate north of the Mendocino Fracture Zone (MFZ) is rotated clockwise relative to the Pacific Plate south of the MFZ. North America has overridden the southern part of the Pacific Plate to the extent that the youngest oceanic crust exposed is more than 25 million years old. North of the MFZ, clockwise rotation has moved the oceanic crust westward past the rifts (spreading centers, age zero) that separate the Gorda and Juan de Fuca (JDF) plates to the east, from the rotated Pacific Plate to the west. North of the MFZ, crust as old as 15 million years is exposed west of the trench. This rotation is the result of interaction between the colliding plates. The Pacific Plate moves continually toward the northwest, dragging the southern edge of the northern block with it. The eastern edge of the oceanic plates is locked onto the North America Plate, resulting in rotation about a pivot beneath western North America.

A similar torque is exerted upon the JDF Plate. The

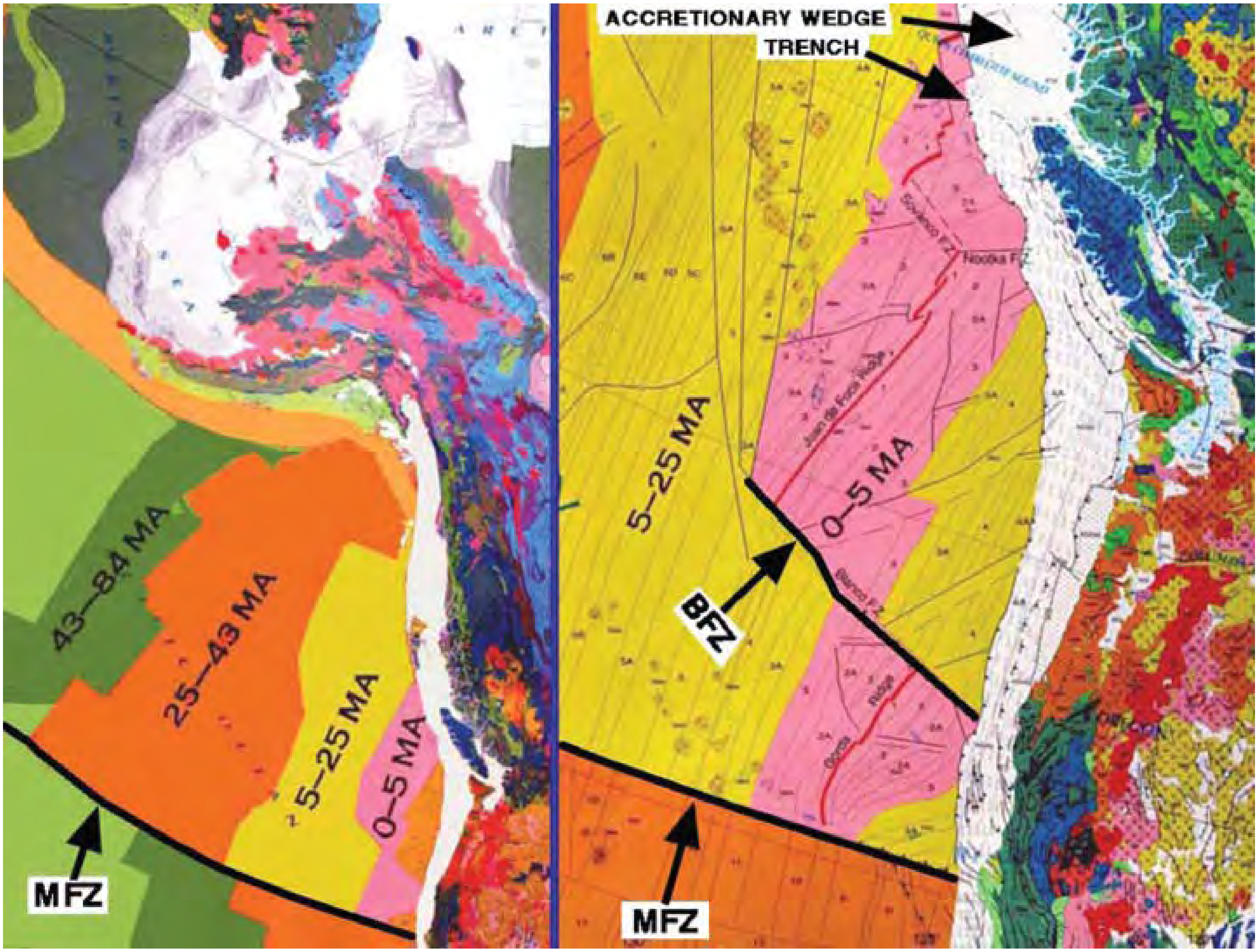

Figure 4. In the left panel, the amount of clockwise rotation can be seen by comparing the location of the western edge of the orange (25-43 Ma) stripe on the Pacific Plate with the corresponding margin on the plate north of the MFZ. In the right panel, the offset of the spreading ridge (pink, 0-5 Ma) along the Blanco Fracture Zone (BFX) is obvious. Oceanic crust younger than 10 Ma is still so warm, thin, and fragile, that it is more inclined to break into microplates than is the encroaching continental crust.

Gordo Plate is sutured onto the North America Plate and cannot rotate. About eight million years ago the accumulated stress became so large that rupture occurred on the Blanco Fracture Zone (BFZ), enabling significant clockwise rotation of the JDF Plate. Some two million years ago, rupture of the Nootka Fault Zone to the north occurred, instigating additional clockwise rotation of the JDF Plate (McNeill, L.C., Goldfinger, C., et al., 2000). The effect of the rotation at either end of North America was to produce significant extension at the southern latitudes and little or no extension at the northern end. A salient index of extension is the volume of volcanics erupted. The Mount Baker Volcanic Field is small, and is surrounded by Mesozoic rocks. Toward the south, the volume of volcanics increases progressively. South of White Pass the volcanics cover the entire region, consequently there are no Mesozoic rock outcrops between there and the Klamath Mountains, the northern end of which begins at Canyonville, Oregon, on Interstate 5.

STRUCTURAL EVOLUTION OF THE PACIFIC NORTHWEST

I have made a lot of effort to understand the interaction of North America with the oceanic plates, as shown in figure 5. The most helpful and remarkably informative document for me is the AAPG Tectonic Map of North America. That map and several interesting papers have been indispensable to my formulating the following story. The following is an attempt to summarize the structural evolution of our region:

1. North America has a 200 million year history of oblique convergence with the oceanic plates of the northeastern Pacific.
2. Deformation has almost entirely been partitioned into margin-parallel, strike-slip faulting and margin-normal thrust faulting or subduction.
3. Subduction events yielding active volcanism have been relatively brief (Tobisch, O.T., Saleeby, J.B., et al., 1995; Godfrey, N.J., Beaudon, B.C., et al., 1997).
4. Strike-slip regimes with associated continental folding and thrusting have been long-lasting and have been responsible for much of the volcanism and magmatism in North America (Brown, E.H., and Talbot, J.L., 1989; Maxon, J., and Tikoff, B., 1996).
5. The eastern edge of the Farallon Plate was subducted beneath North America between 200 and 30 Ma. This regime configured the western margin of North America with a deep concavity or embayment toward the east, and an orogen (the Rocky Mountains) similar to the modern Andes, with analogous relief and elevations. The magmatic arc was continuous and parallel to the trench. Today, the dismembered fragments of that arc are the Peninsular Ranges (which extend about 800 miles from Southern California to the tip of Baja), the Sierra Nevada Batholith, the Klamath Mountains, the Idaho Batholith, the Wallowa Mountains, the Okanogan and Kettle Mountains, and the Coast Plutonic Complex (Russo R.M., and Silver, P.G., 1994, and 1995; Hamilton, W.B., 1969, and 1978).

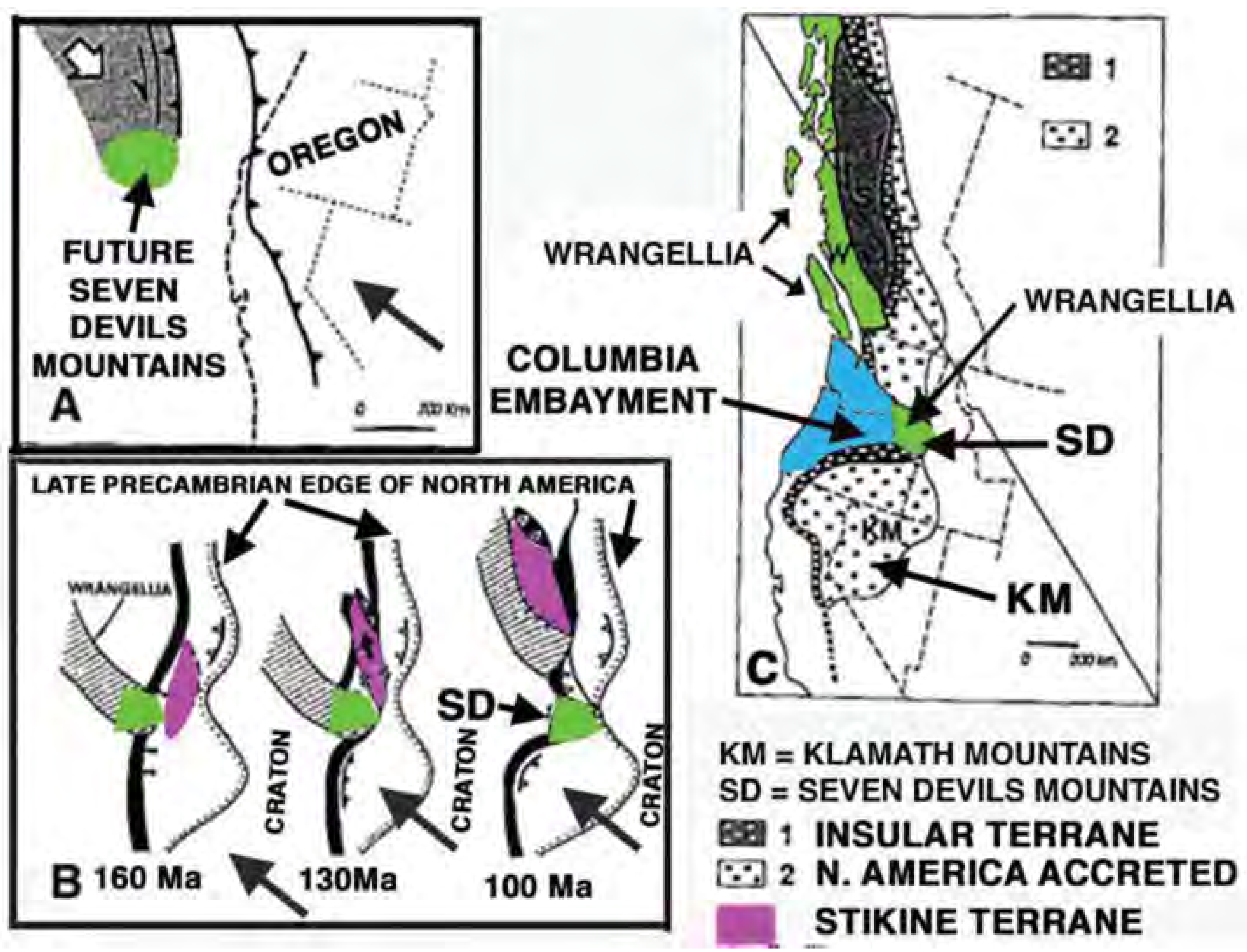

Figure 5. This image, taken from Vauchex and Nicolas (1988) and from Wernicke and Klepacki (1988), is their interpretation of the mode of collision between Wrangellia and North America.

6. Wrangellia began as a large igneous province on the Farallon Plate, most likely as a result of hot spot eruptions (Richards, M.A., and Jones, D.L., et al., 1991). It moved northward, parallel to, and in step with North America. When it collided with the Aleutian Trench, it jammed the subduction system because it was too thick and buoyant to be subducted. Its northward movement stopped; it was sutured to the Bering Plate (fig. 5C).

7. The consequence was that between about 94 and 40 Ma, there was a transpressional relationship between the Farallon and North American plates because Wrangellia and Baja, California, could not be subducted in the oblique convergence regime. Uplift at the rift yields a ridge that rises some 3,000 feet above the abyssal plain, and under those conditions the continent cannot over-ride it. It will simply push the ridge in front of it. This collisional relationship resulted in the Laramide Revolution (Orogeny) (Maxon, J., and Tickoff, B., 1996). America moved more vigorously above a nearly horizontal oceanic slab (flat slab subduction as shown in fig. 6) making an exceptionally wide volcanic arc extending to central Montana. At its eastern edge, the subducting slab sank rapidly and ruptured, producing the Central Montana Igneous Province (Nolet, G., 2009).

8. During the events summarized in paragraph 7, Wrangellia collided with North America in a left-lateral regime (Monger, J.W.H., Van der Hayden, P., et al., 1994; Monger, J., 2002). See figure 5C. This is not compatible with the long-standing oblique convergence of North America with the Farallon Plate, or with the right-lateral movement of Wrangellia and Baja California relative to North America.

9. Prior to 40 Ma, Wrangellia jammed the Aleutian subduction zone and was sutured to the Bering Plate. North America moved northwest in a left-lateral transpressive relationship to Wrangellia and Stikinia, a tectonostratigraphic terrane found in the Canadian cordillera that formed in a volcanic arc environment during Paloezoic and Mesozoic time (Currie and Parrish, 1997). Wrangellia indented North America, trapping Stikinia in a wedge with its apex to the south. Tectonic wedging enabled the escape of Stikinia to the north. Its oceanic basement was subducted beneath the Bering Plate and its supracrustal component was obducted onto the Bering Plate to form a fold and thrust belt (Vauchez, A., and Nicholas, A., 1991; Wernicke, B., and Klepacki, D.W., 1988). See figure 5A.

The indentation of North America was so extreme that the southern end of Wrangellia now crops out as the Seven Devils Mountains of Idaho (fig. 7). The indentation is manifested on the Oregon Coast just north of Bandon, where outcrops of the Franciscan Assemblage (Otter Rock Formation) end and the Tyee Basin begins.

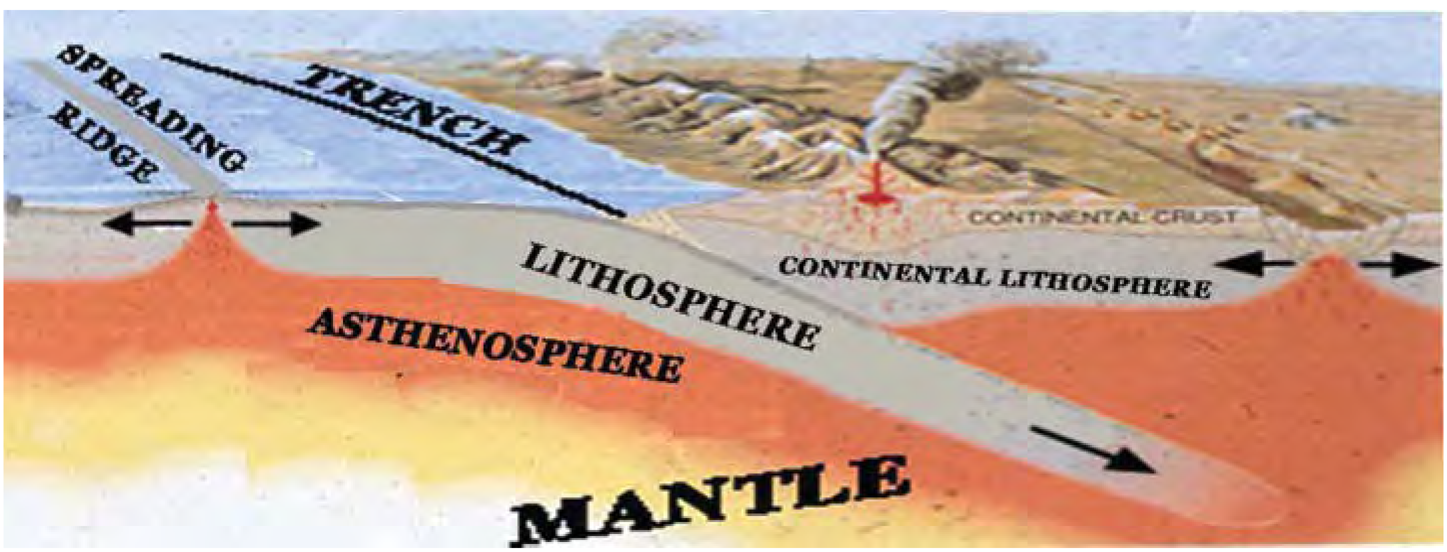

Figure 6. Flat-slab subduction. With rapid convergence, the continent impinges over the oceanic plate more rapidly than the oceanic plate can sink. In consequence, the dip of the oceanic plate is much less steep, or “flat”. The result is significant widening of both the forearc and the back arc.

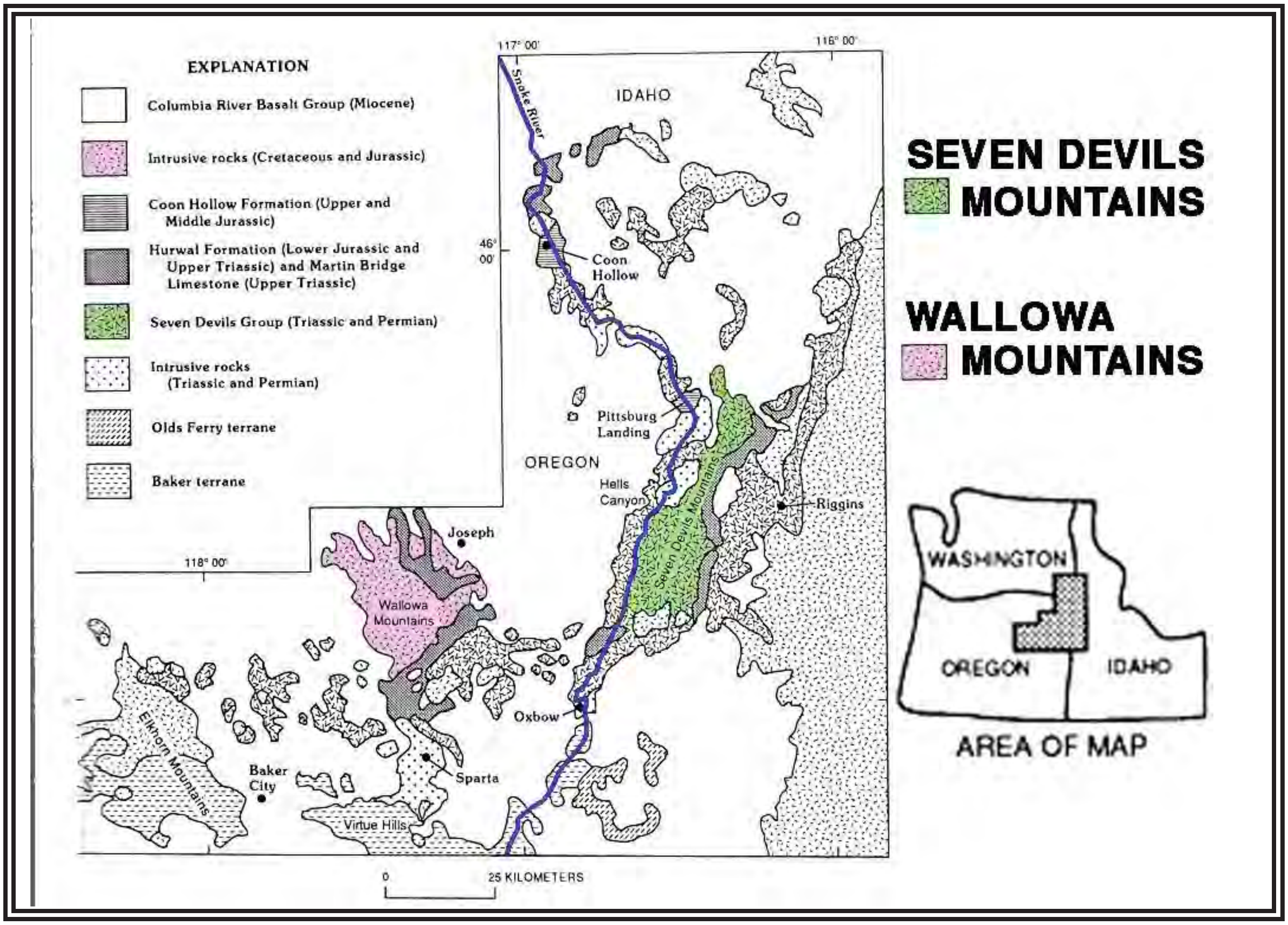

Figure 7. The Seven Devils Mountains are remnants of the southeastern end of Wrangellia. Subsequent deposition of sedimentary and volcanic rock buried the portion of Wrangellia between the Seven Devils Mountains and Vancouver Island. The Wallowa Mountains are part of a volcanic arc that docked into Wrangellia.

The Tyee Basin has a basement of oceanic basalt (Siletz Volcanics). Inland, at Roseburg, are outcrops of Siletz Volcanics, the equivalent of the Crescent Formation.

A few miles to the south, near Dillard, Oregon, are outcrops of Franciscan Formation, revealing how much that unit was offset to the east by the indentation of North America by Wrangellia. A few miles farther south, near Canyonville, the Klamath Mountains begin. They are the southern remnant of the Mesozoic Volcanic Arc. The northern remnants have been offset far to the east by indentation of the continent from collision with Wrangellia.

10. The wedge containing Stikinia was extruded to the north. The eastern margin of the wedge was the Straight Creek-Fraser Fault (SCFF). The western margin of the wedge was largely a subduction zone. See figure 6. There was a maximum rate of convergence because the oceanic plate moved northeast whilst North America moved southwest. The oceanic plate did not have time to sink, consequently flat slab subduction prevailed.

11. The SCFF was a right-lateral, strike-slip transform fault aligned north-south, driven by the northward movement of the extruding wedge (Stikinia) relative to North America (a transform fault cuts completely through the lithosphere, separating two tectonic plates). Drag on the SCFF produced a series of north-south, strike-slip faults (en echelon) in northwestern North America. The imprint of that structure can be seen on the digital image of Washington, figure 8. The valleys north of the OWL (followed by I-90) are aligned north-south.

12. About 55 million years ago the direction of motion of the Pacific Plate changed from

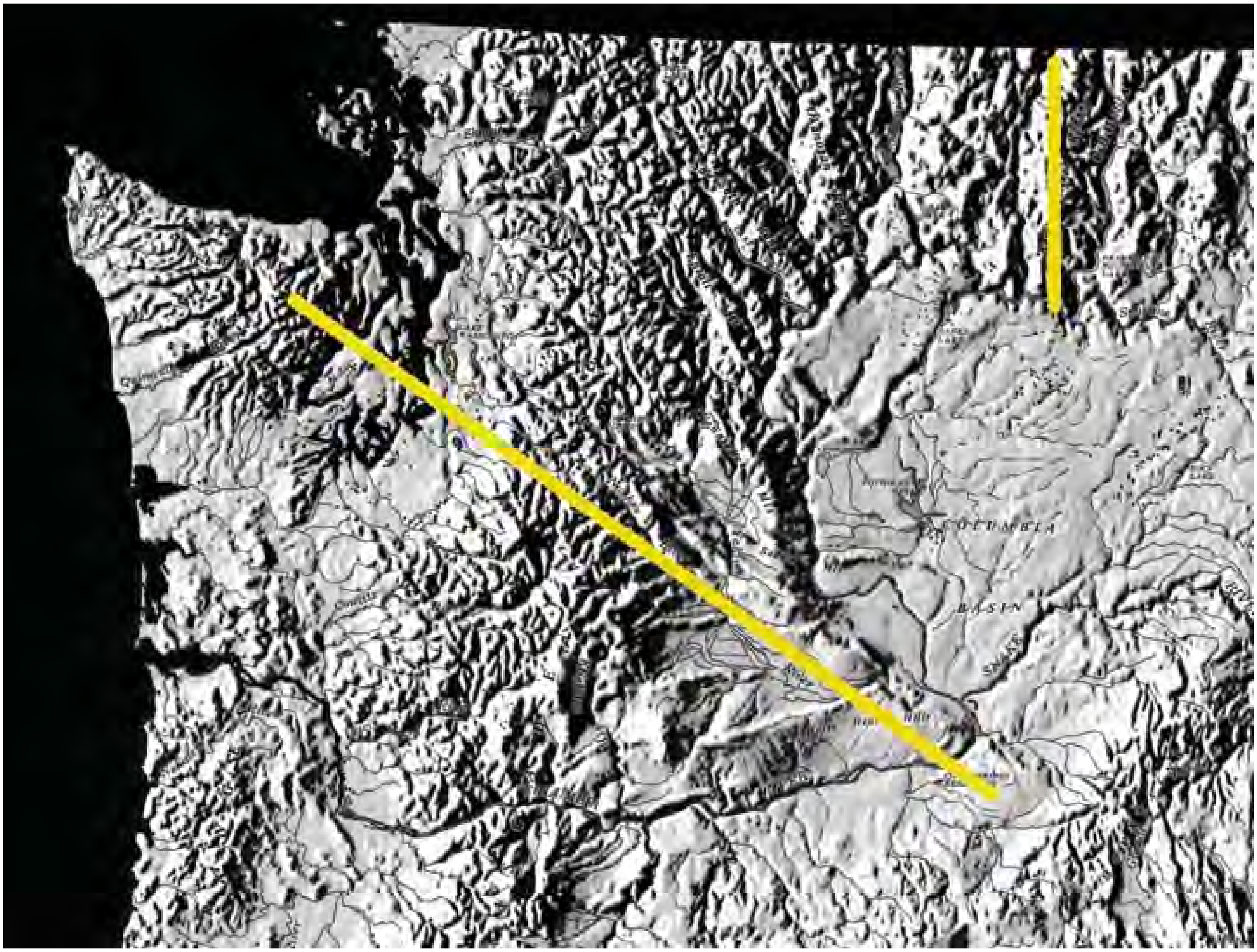

Figure 8. This computer-generated relief map of Washington shows the OWL and the Straight Creek Fault. The major structures to the northeast are predominately aligned north–south, parallel to the Straight Creek Fault.

northward to northwest. See figure 9. The likely driver of this change was the collision of India with Tibet, an event that caused a major, world-wide reorganization of tectonic plate interaction. The westward component of this motion caused the eastern margin of the Pacific Plate (rift or spreading center) to move westward, making space for North America to expand (extend) toward the west. This abruptly ended the flat-slab subduction and the fold and fault regime. The force of gravity impelled the Cordillera to become thinner (collapse) as it expanded, moving the coast and the trench westward. The lower crust, below the brittle-ductile transition, stretched like taffy whilst the upper, brittle crust expanded by means of motion on curved (listric) faults. See figure 10.

13. In figure 11, the block containing the Puget Lowland and the Olympic and Coast ranges encounters Vancouver Island as an immovable barrier. It cannot move northward. Instead it develops folds and faults which enable the necessary shortening and yield the array of synclines and anticlines along the Oregon and Northern California Coasts (i.e. Cascadia) including the Strait of Juan de Fuca, the Olympic Mountains, Grays Harbor, the Doty Hills, Willapa Bay, the Willapa Uplift, the Columbia River estuary, and on and on along the Oregon and Northern California Coast, i.e., Cascadia. By contrast, the western block in the accretionary wedge encounters the trench of Vancouver Island at a very oblique angle. Part of its northward motion is accommodated by subduction beneath Vancouver Island. Another part is enabled by transform faulting along the oblique contact.
In addition, North America, moving southwest, impinges directly on the western accretionary wedge block, opening the transtensional rift even more.

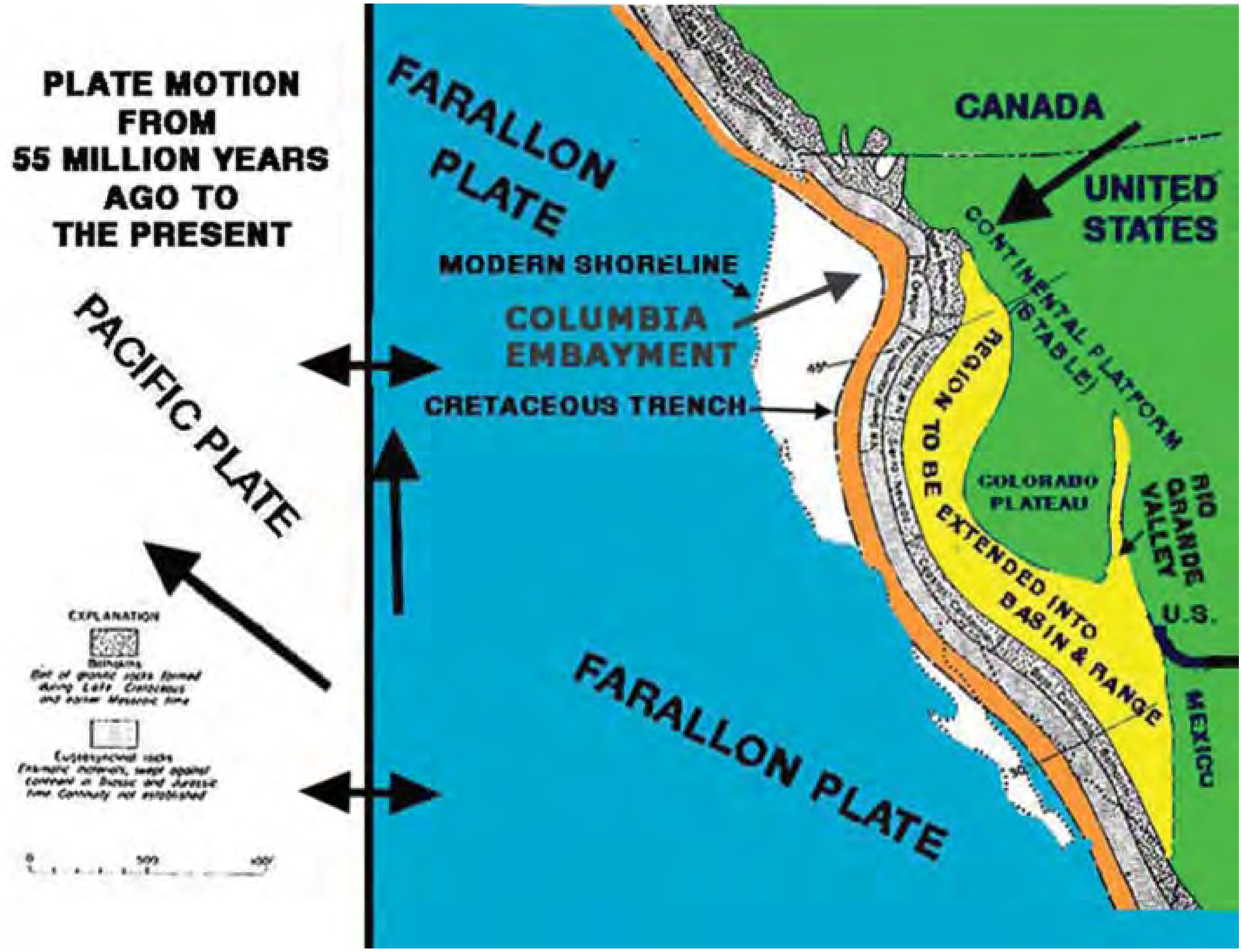

Figure 9. Plate motion from 55 million years ago to the present (Hamilton, 1969). Prior to about 30 million years ago, Baja California was inside the trench and firmly attached to North America. After the Pacific Plate began moving northwest, the continent stretched, moving the trench westward and causing the extension that widened both the forearc and volcanic arc.

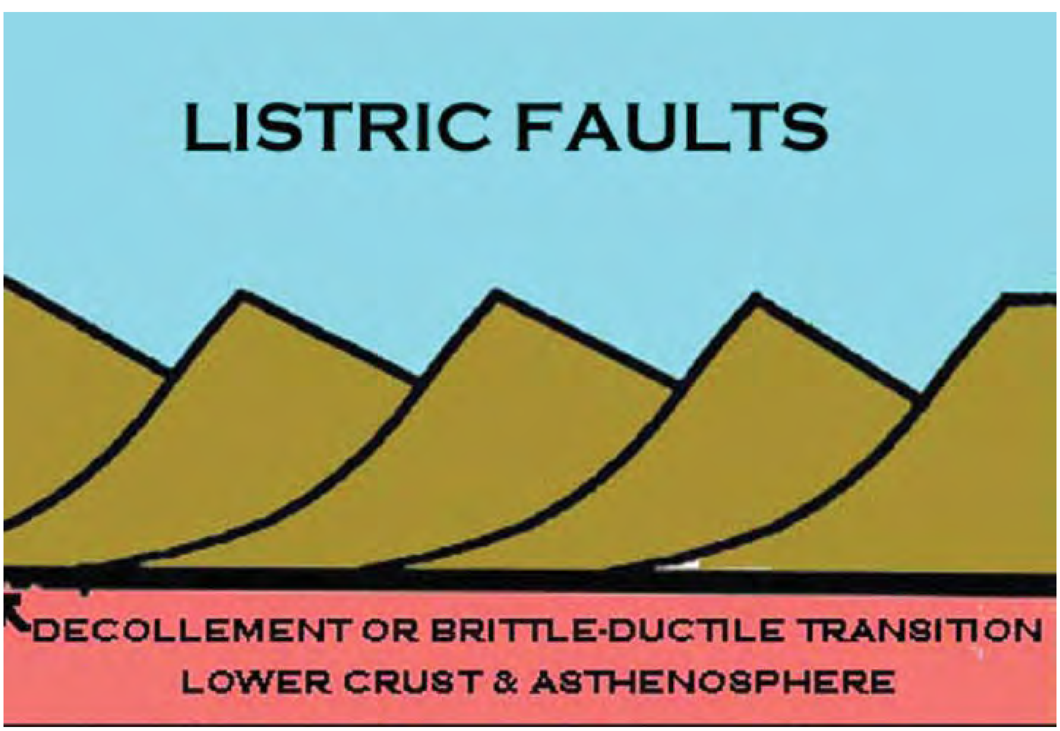

Figure 10. Listric faults. Below the brittle-ductile transition, the lower crust and the asthenosphere, the uppermost layer of the mantle are ductile, and can stretch without breaking.

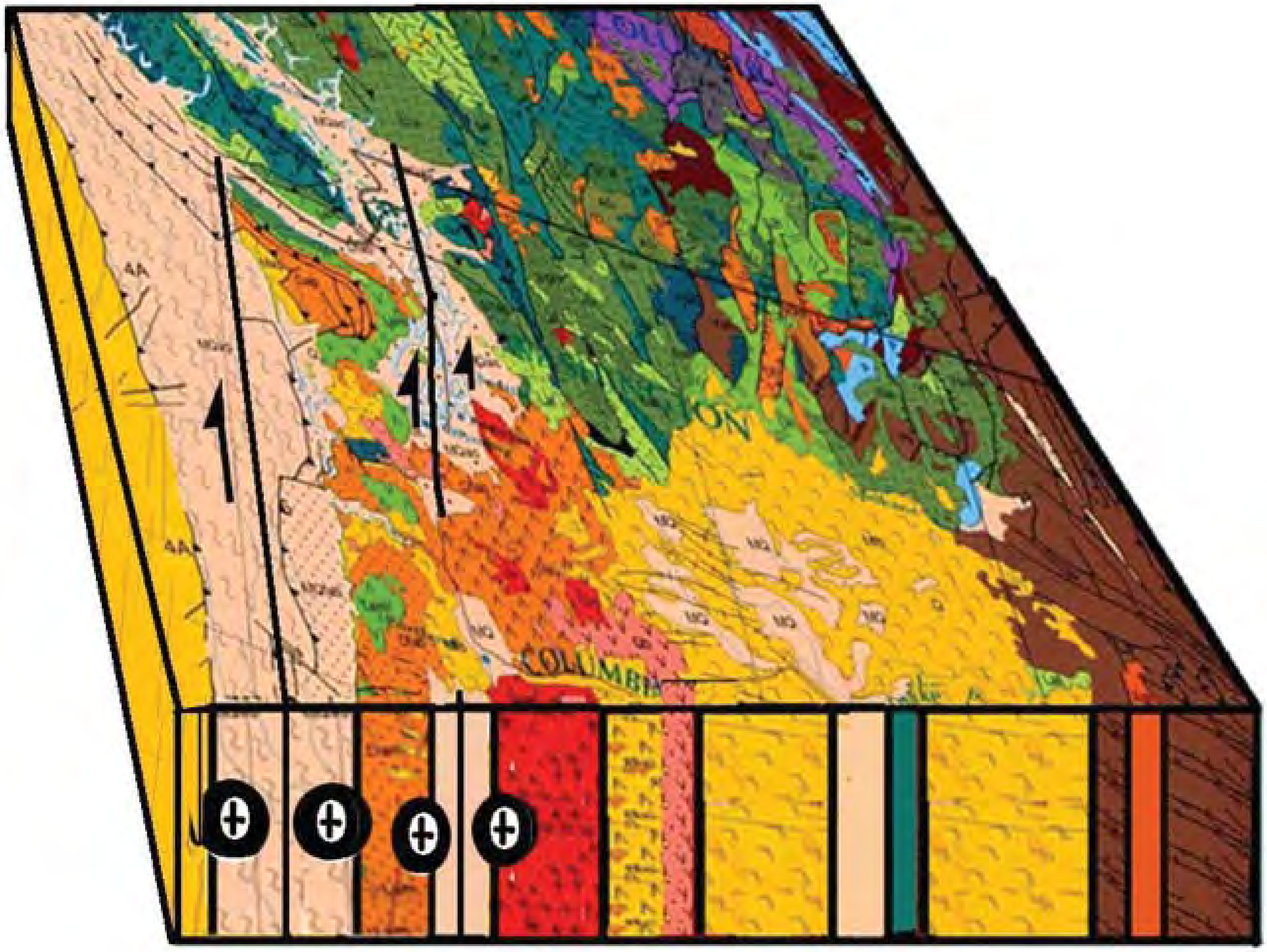

Figure 11. This block diagram depicts two north-striking, right lateral transform faults, one in the accretionary wedge and the other in the Puget Lowland. No doubt there are many similar faults in the forearc but it is not feasible to put them in a diagram. With each fault, the western block moves north relative to the eastern block.

14. In Cascadia, the forearc is moving northward nearly continuously, shown in figure 12. The Olympic Mountains are colliding with the east-west backstop of Vancouver Island. The result is compression of the Olympic Mountains and Puget Lowlands into east-west folds and faults, e.g., the Strait of Juan de Fuca, the Olympic Mountains, Grays Harbor, etc. The block west of the fault in the accretionary wedge is in oblique contact with the trench. Part of its motion is absorbed by subduction. Another part involves right-lateral offset on an oblique transform fault. That offset imparts counterclockwise rotation of the western block, and transtension, making a rift, depicted by the red zone in figure 12.

There is similar but east-west rifting to the east. I suspect that this yielded volcanism in British Columbia. My contention is that the north-south transtensional rift led to the immense outpouring of Siletzia basalt. (The northern part of Siletzia was mapped by geologists of Washington State, who named it the Crescent Basalt. Oregon-based geologists named their unit the Siletz Volcanics, after the type locality in Siletz Bay. Siletzia is the terrane underlying much of the Oregon and Washington forearc, consisting of Eocene mafic crust with high seismic velocities. It acts as a backstop for accretion of marine sedimentary rocks from the obliquely subducting Juan de Fuca slab, (Parsons, Wells, Fisher, et al., 1999)]. That unit was so thick that it made a number of subaerial volcanoes in addition to voluminous submarine deposits. The oceanic plate thus became too thick to be subducted and it jammed the subduction system. The result was the accretion of the Siletzia to North America. See figure 13. This led to the development of a new subduction zone some distance west of the Challis Trench. The Challis Arc became inactive and the new Cascade Arc was born.

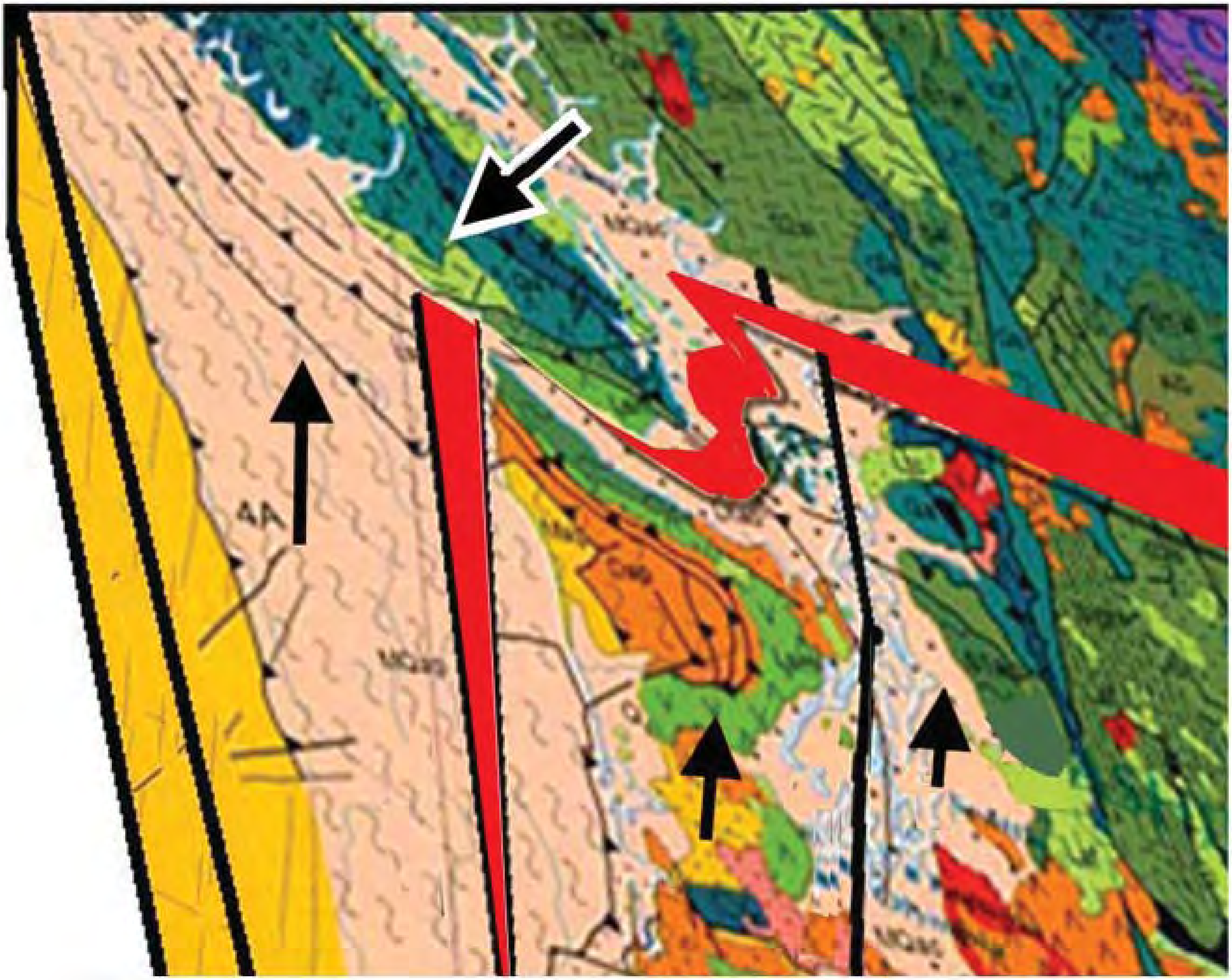

Figure 12. This image is from the AAPG Tectonic Map of North America. The intent is to show the effect of right-lateral transform faulting on the edge of the continent. The red areas depict zones of extension produced by rotation and transform faulting.

CONCLUSION

I propose that the events postulated in figures 11 and 12 brought about the outpouring of Siletz Volcanics and the accretion of that terrane to North America. Extension of western North America has prevailed ever since that accretion event. In addition, there has been significant rotation of the Juan de Fuca Plate.

This tectonic regime has led to the formation of metamorphic core complexes in the Okanogan/Kettle Mountains and instigated widespread volcanism. Those remarkable phenomena deserve a story of their own.

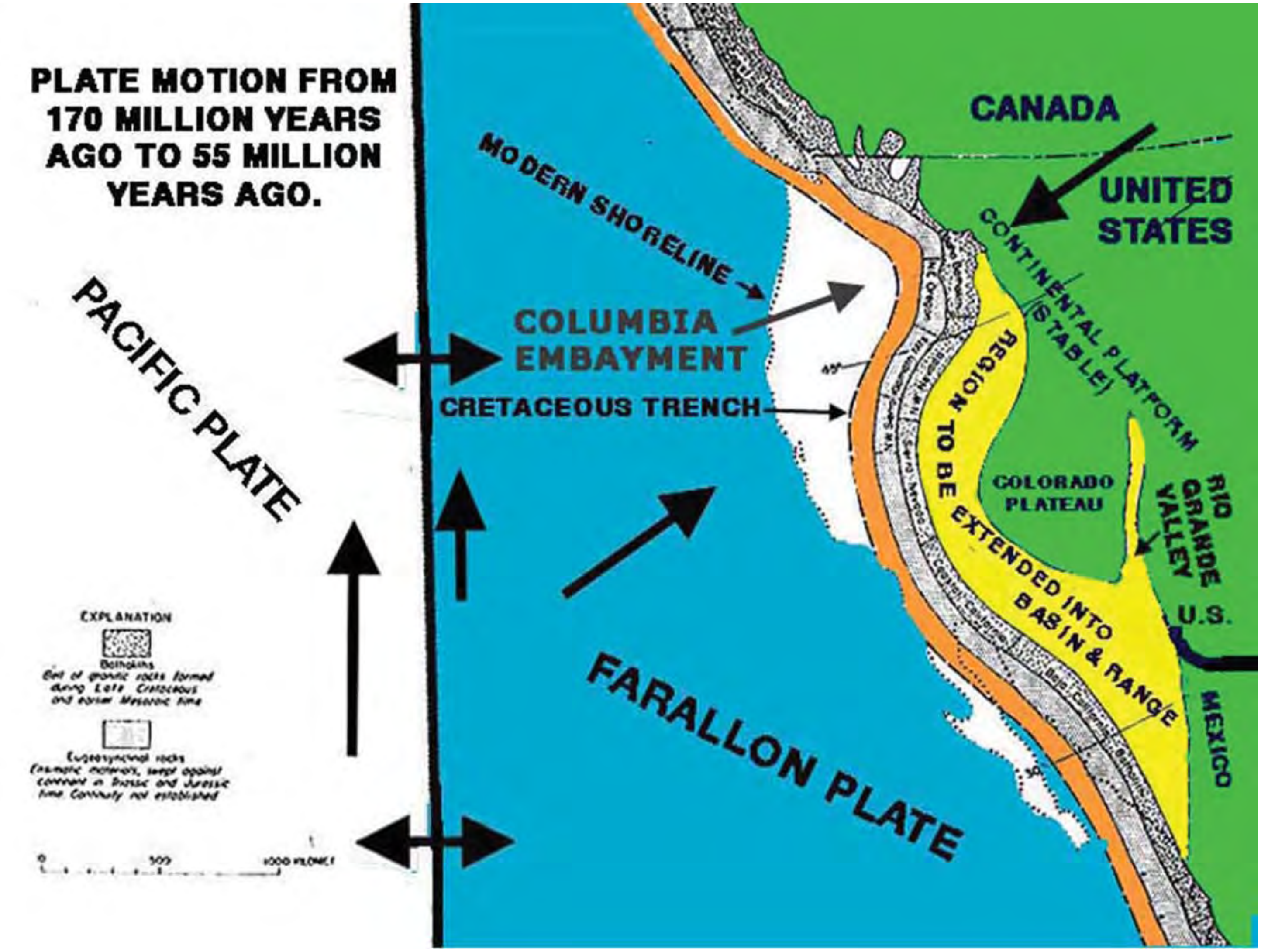

Figure 13. The Farallon Plate here is moving northward in opposition to the southwest-moving North America Plate. The Crescent Basalt-Siletzia Volcanics jammed the subduction system and the unit was accreted to North America. The map is derived from Hamilton, 1969.

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GEOLOGY—A BLESSING

John Whitmer

Issaquah, Washington

My interest in geology began in the 1930's, when I was about 13 years old. My community (Spokane, Washington) was abuzz with excitement about changes coming to the Grand Coulee and Dry Falls, which once dwarfed Niagara Falls. The excitement was fueled by the proposal to build Grand Coulee Dam and by the zealous opposition of the electric utilities, Washington Water Power Company in particular. Although I did not realize it, the rejection of the flood hypothesis of J. Harlan Bretz was reverberating in the geological community as well. My high school general science teacher augmented my interest. He told us why there were so many lakes near Spokane, and about how Hayden Lake nearly drained catastrophically when its impervious bottom layer of clay ruptured.

If I had followed this interest in choosing my career, geology would have been my choice. However, I did not see geology as a way to make a living. In response to many influences, I originally chose to study Electrical Engineering, but switched to Medicine upon realizing that I was smart enough to go to medical school.

In innumerable ways, becoming a physician was an immense blessing to me, but I was never comfortable with the business aspects of medicine. In addition, I wanted to hike and explore the nearby mountains. I was known as the "doctor who was always up in the mountains when you needed him."

I solved that problem by becoming the Student Health Director at the University of Wyoming. Among the perquisites of that job was eligibility to take three semester hours of class without paying tuition. In my years at the University I studied Geomorphology, Paleontology, and Structural Geology. Don Blackstone and Brainerd ("Nip") Mears became mentors, friends, and role models to me. I also became acquainted with Dave Love, a neighbor.

Shortly after the 1959 Madison Canyon Earthquake, Don Blackstone, Nip Mears, and Ron Parker invited me to go with them to the earthquake site, and to a Friends of the Pleistocene field trip in the Wind River Range. That was the most exciting, awesome, intellectually stimulating experience of my life.

My passion for geology has never waned since. My love for mountain climbing combined perfectly with geology. I have seen world-class outcrops in Wyoming, Montana, Idaho, Washington, Oregon, the southwestern states, California, Canada, England, New Zealand, and Iceland among others.

Few career geologists have been on as many geological field trips led by heavy-hitters as I have. TRGS has contributed mightily to my education, for I have missed only a few of its annual meetings. I have been on many GSA field trips since 1973, and I became a professional member of the GSA in 1985. Several Friends of the Pleistocene field trips have greatly enhanced my understanding.

I joined the Northwest Geological Society in 1965. In my view, it is the greatest enhancement to the quality of life in the entire Seattle region. I have missed few of its meetings and served as the editor of its newsletter for about 20 years. I was president of the NWGS for one term.

For decades I whined because I thought I should have been a geology professor. Nevertheless, I persisted in medicine. It simply would have been too burdensome for my family if I had left medicine to pursue the academic study of geology.

Since retiring, I have devoted my full attention to geology. My motto is, "Anything that takes my mind off geology is not good for me." For 20 years I have taught two non-credit geology classes for senior citizens for three quarters every year.

Nowadays, I look back on my life and realize that if I had it to do over, I would not change a thing. I feel richer than Bill Gates although I virtually lack any disposable income. I have had the best of two worlds. There is no way to overstate the blessing that medical education and practice have been to me. Geology has been the perfect avocation, totally eliminating boredom from my life and providing a wonderful *raison d' etre* in retirement. I am so busy that one of my friends remarked that I should get a job so I could have some spare time.

FACIES MODEL AND SEQUENCE STRATIGRAPHIC FRAMEWORK OF THE DEVONIAN-MISSISSIPPIAN SAPPINGTON FORMATION IN SOUTHWESTERN AND CENTRAL MONTANA

T. Nagase, M.H. Hofmann,* and M.S. Hendrix

Department of Geosciences, University of Montana, Missoula

**Corresponding author (michael.hofmann@umontana.edu)*

ABSTRACT

The Devonian/Mississippian Sappington Formation in central and southwest Montana is the surface equivalent of the prolific Bakken petroleum system in the Williston Basin. Although studied heavily in the 50s and 60s, it has not been on the radar screens of researchers for several decades. Here we provide a fresh look at the Sappington Formation using a modern sequence stratigraphic approach in order to better understand the depositional history and facies distribution.

The Sappington Formation in southwestern and central Montana can best be characterized by eight lithofacies. These eight lithofacies are organized into four facies associations, describing distinct depositional environments, ranging from restricted offshore (FA1) and normal marine offshore (FA3), to a storm dominated shoreface (FA4) and open marine carbonate build-ups (FA2).

Stacking of the facies and facies associations suggests that Sappington sedimentary rocks are part of two complete higher order stratigraphic sequences and a partial higher order stratigraphic sequence. The basal boundary of the lower sequence is marked by the erosional contact between the lower Sappington Shale and the underlying Three Forks Formation. Half way through the middle Sappington Member, there are progradationally stacked, shallow marine facies abruptly overlain by offshore-transitional facies, marking the contact between the first and the second sequence. Renewed progradational facies stacking above this surface is characteristic for the upper part of the middle Sappington member, reflecting the gradual basinward shift during the second cycle. The base of the third sequence is located at the bottom of the upper Sappington Shale. This last Sappington sequence encompasses the upper Sappington Shale and part of the Lodgepole Limestone.

1. INTRODUCTION

Over the past decade, exploration and development of the Devonian Bakken-Three Forks petroleum system in North Dakota and Montana has increased, and it is now the top producing petroleum play in the United States (Gaswirth et al., 2013). Despite the magnitude of this economic boom, surprisingly little information has been published that directly addresses the detailed stratigraphy and sedimentology of equivalent late Devonian and early Mississippian strata. These rocks are well exposed in outcrops within the Bridger and Tobacco Root Ranges of southwestern Montana, and the Big Snowy Mountains in central Montana (fig. 1).

This paper summarizes results from initial documentation of three stratigraphic sections of the Sappington Formation in central and southwestern Montana as part of a recently-completed M.S. Thesis at the University of Montana (Nagase, 2014). The main objective is to provide baseline stratigraphic information regarding the lithofacies, sedimentary and biogenic structures, mineralogy, and geochemistry of the Sappington Formation.

Previous workers (e.g., Achauer, 1959; Gutschick et al., 1962; McMannis, 1962: Sandberg, 1965; fig. 2) typically differentiated the Sappington Formation into five lithostratigraphic units:

(1) a basal black, organic-rich mudstone
(2) a lower argillaceous, calcareous siltstone or silty argillaceous limestone
(3) a middle gray, laminated, calcareous mudstone;
(4) an upper yellowish gray to brownish orange, flaggy to massive, calcareous siltstone
(5) an uppermost dark gray to black, sandy, organic-rich mudstone.

Compilation of results from regional studies and conodont analyses indicate that strata of Late Devo-

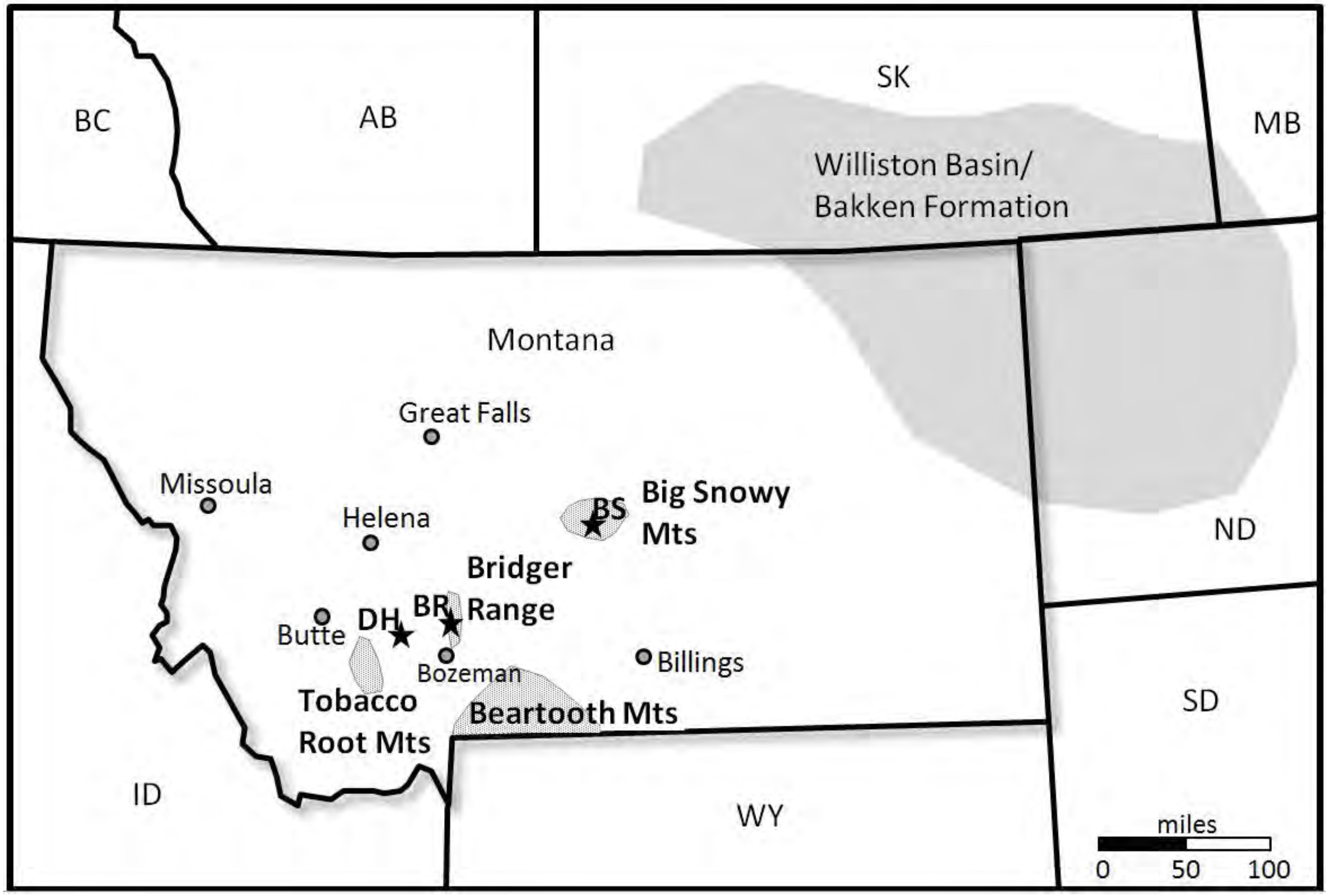


Figure 1. The study area is located in central and southwestern Montana. Sappington outcrop locations presented in this manuscript are highlighted with stars [Dry Hollow (DH), Bridger Range (BR), and Big Snowy Mountains (BS)]. Other geographic features referred to in the text are also included.

nian (Famennian) to Early Mississippian (Tournaisian) age in Montana include the Bakken Formation in the Williston Basin, the Exshaw Formation in the Western Canadian Basin, and the Sappington Formation in central and southwestern Montana (Sandberg and Klapper, 1967; Baars, 1972; Peterson, 1986; Savoy and Harris, 1993; Smith and Bustin, 2000; Kaufmann, 2006; Johnston et al., 2010; fig. 3).

The basal, organic-rich Sappington Formation is time equivalent to the lower member of the Bakken and Exshaw Formations, whereas the middle and upper members of the Sappington Formation are time equivalent to the lower part of the middle Bakken Formation and the upper part of the lower Exshaw Formation. The missing conodont zones *Siphonodella Sulcata* and lower *S. Dupulicata* above the middle and upper members of the Sappington Formation caused early workers to interpret the presence of an unconformity that spans the Devonian-Mississippian boundary (Achauer, 1959; Gutschick et al., 1962; McMannis, 1962; Sandberg and Klapper, 1967). In contrast, the Bakken and Exshaw Formations contain the two missing conodont zones.

During the Late Devonian (Famennian) to Early Mississippian (Tournaisian), the western margin of the North American craton was located between the equator and 30° N, and was characterized by a subtropical climate (Scotese and McKerrow, 1990). Smith and Bustin (1998) inferred that an east-directed equatorial undercurrent provided nutrients to the western margin of the continent through a series of interconnected basins that included the Williston Basin (fig. 4). At this time, the tectonic emplacement of an allochthon along the western margin of the North American continent initiated the Antler Orogeny (e.g. Sandberg, et al., 1983; Giles and Dickenson, 1995; Kevin, 2001; Warme et al., 2008), forming a mountainous ‘Antler Highlands’ which separated the western margin of the craton from the open Panthalassa Ocean to the west (fig. 4).

Prior geologic analysis of the region has suggested that, in addition to the Antler Highlands and foredeep

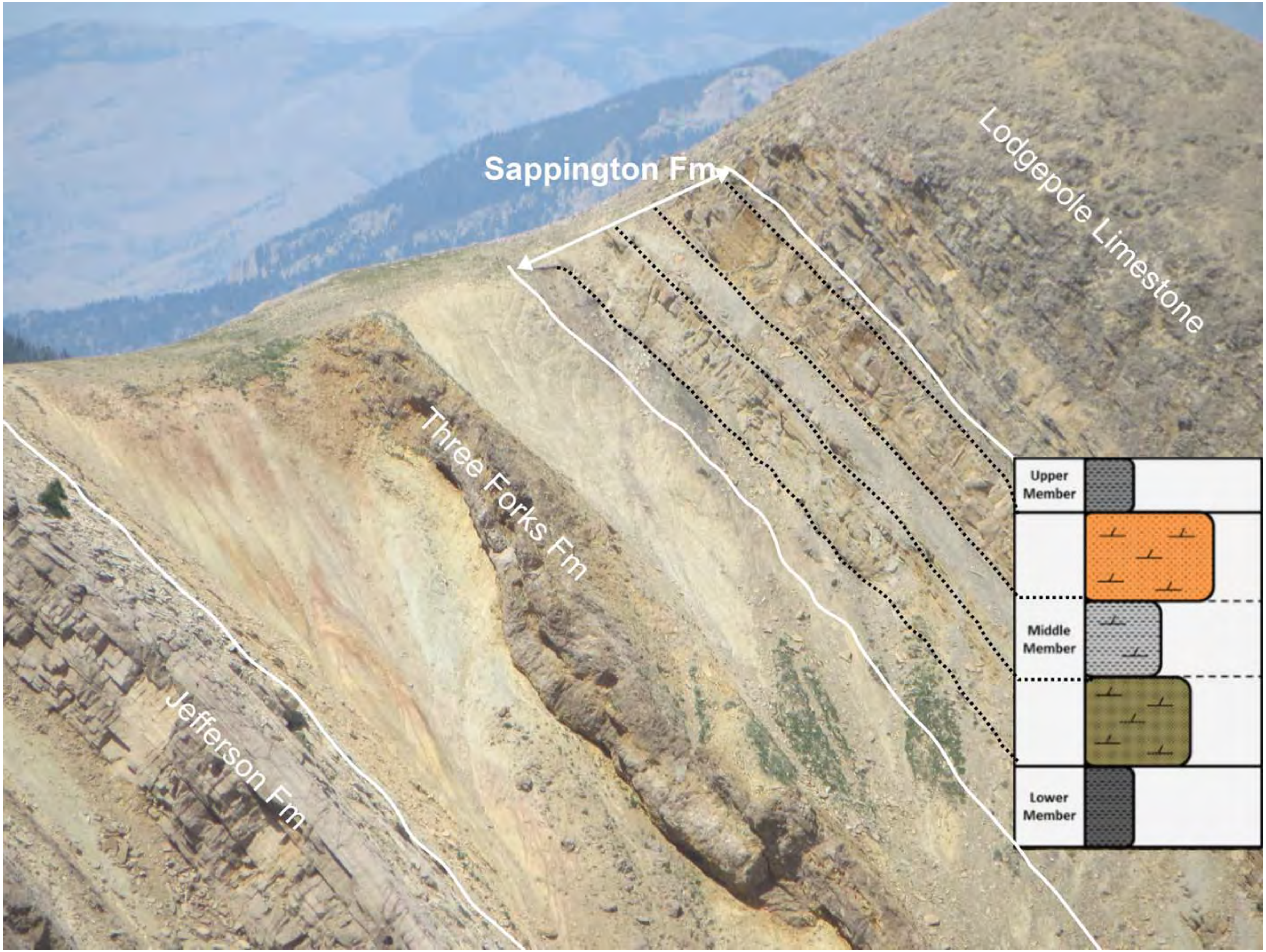


Figure 2. Example of the Sappington Formation in the Bridger Range (BR section), showing five distinct lithostratigraphic units, bounded below by the Three Forks Formation and above by the Lodgepole Formation: lower, middle, and upper mudstone members are slope formers, and lower and upper units of a middle calcareous siltstone-silty limestone member are cliff formers (e.g., Achauer, 1959; Gutschick, et al., 1962; McMannis, 1962: Sandberg, 1965). View looking toward the northwest.

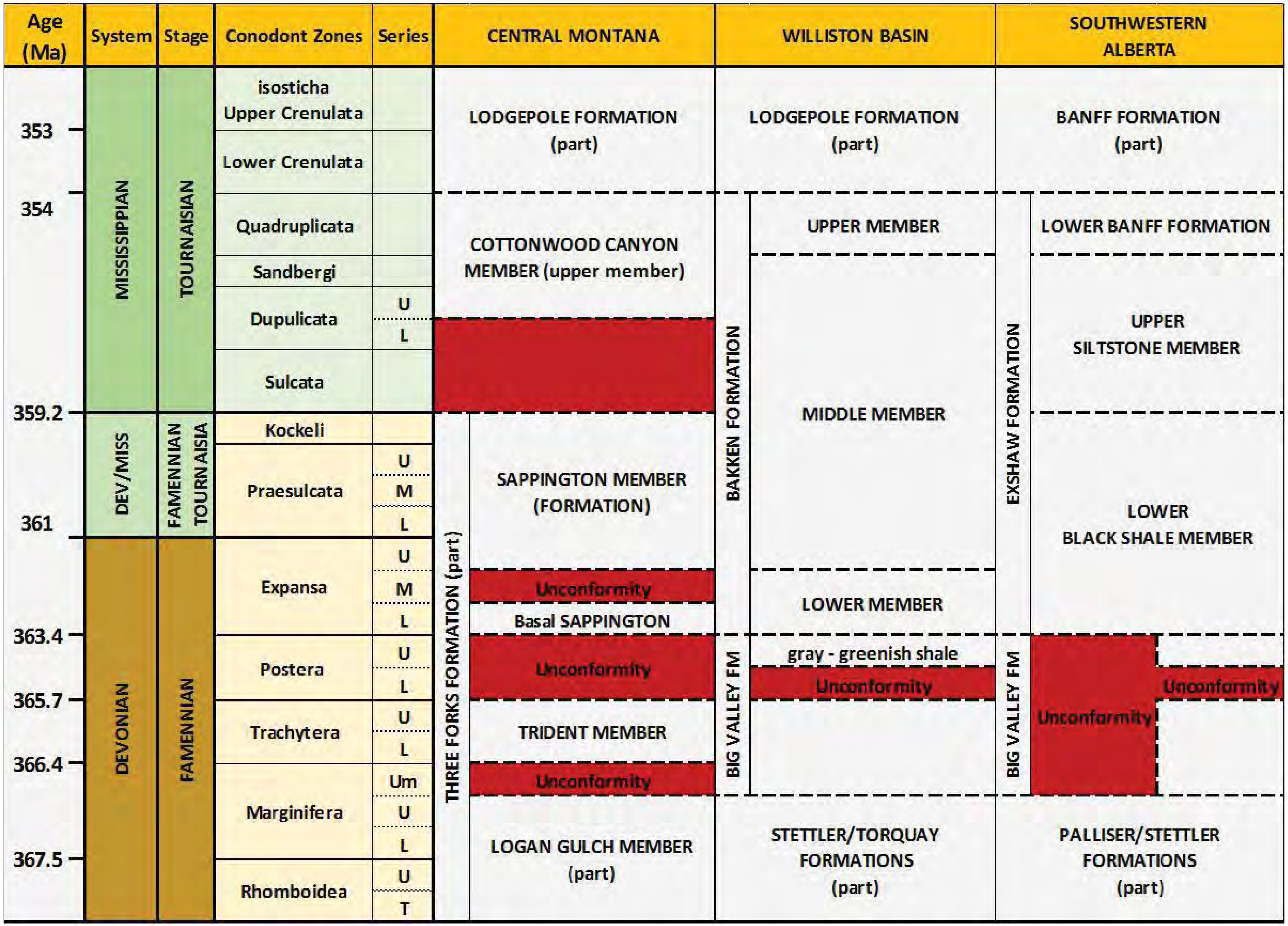


Figure 3. Biostratigraphic correlation of the Devonian-Mississippian strata in Montana based on conodont zones (e.g., Savoy and Harris, 1993; Kaufmann, 2006; Johnston, et al., 2010). The biostratigraphic correlation shows that the Sappington Formation is time equivalent with a lower member and a lower unit of a middle member of the Bakken Formation, and it is time equivalent with the lower member of the Exshaw Formation. The upper unit of the Cottonwood Canyon Member of the Lodgepole Formation unconformably overlies the Sappington Formation above two missing conodont zones (Siphonodella Sulcata and S. Dupulicata). Calibration of the time scale is based on Haq and Schutter (2008) for Devonian strata and Buggisch, et al., (2008) for the Mississippian strata. The Devonian-Mississippian boundary is placed at 359.2 Ma according to Haq and Schutter (2008), whereas Kaufmann (2006) placed the boundary at 360.7 ± 2.7 Ma, which makes the Devonian-Mississippian boundary variable.

Figure 4. Paleogeographic map of Montana during the Devonian-Mississippian, showing major sedimentary basins (dark blue) and potential exposed (source) areas. The study area is located in the Sappington Basin (map compiled from several sources, including Scotese and McKerrow, 1990; Smith and Bustin, 1998; Sonnenberg, et al., 2011; Blakey, 2011).

to the west, paleogeographic and tectonic elements in Montana included three major highlands and three major basins or troughs (Baars, 1972; Peterson, 1986; Smith and Bustin, 2000; fig. 4). These paleogeographic elements and the role they played in controlling Devonian-Mississippian depositional systems have led to differing interpretations regarding the distribution of lithostratigraphic units across the region. For instance, Sonnenberg et al., (2011) described the Bakken Formation in the Williston Basin as extending to the Elk Point Basin in the Western Canadian Basin where the Exshaw Formation is present, whereas Smith and Bustin (1998) suggested that the Bakken Formation in the Williston Basin extends through the Central Montana Trough where the Sappington Formation is present. Although lateral variations of the Sappington Formation are not well documented, early work on the formation concluded that the Sappington was deposited in a west-east trending basin in central Montana and that it thins both to the north and south (e.g. Gutschick et al., 1962; McMannis, 1962; Rau, 1962).

2. METHODS

We measured a series of stratigraphic sections from outcrops in central and southwestern Montana to document sedimentary facies of the Sappington Formation. We described bedding style and characterized bedding thickness, lithology, grain size, and sedimentary and biogenic structures. We measured natural gamma radiation (GR) at each outcrop using a handheld scintillometer (RS-125 Super-SPEC), collecting measurements every 50 cm in fine-grained portions of the section and every 100 cm in coarser-grained portions of the section. Our field observations were supplemented by petrographical, geochemical, and mineralogic information acquired during laboratory analysis at the University of Montana. Here we report quantitative X-ray diffraction (XRD) analysis for 59 samples, determined by using a PANalytical X'Pert PRO X-ray diffractometer, and total organic carbon (percent TOC) for 78 samples, determined using an Elemental Analyzer (CE Instruments EA 1110 CHNS-O).

3. SAPPINGTON FACIES

The Sappington Formation in central and southwestern Montana can be subdivided into eight facies and three sub-facies based on lithology, sedimentary structures, microfabrics, and ichnology. The mineralogical (XRD), geochemical (TOC), and petrophysical (GR) characteristics for each facies are listed in table 1.

Facies 1: Organic-rich Mudstone

The organic rich mudstone facies is commonly associated with the upper and lower informal members of the Sappington Formation. The facies can be subdivided into three different sub-facies representative of distinct depositional processes.

Facies 1A: Organic-rich Mudstone with Microfossils

Facies 1A consists of organic-rich mudstone con-

Table 1. Bulk mineralogic, geochemical (TOC) and petrophysical (GR) characteristics of the Sappington facies in the study area.

Facies	Mineralogy (average wt %, [2 *sigma* standard					
			Carbonate			
	Quartz	K-spar	Calcite	Dolomite	Ankerite	Illite
1A	30.0 [4.1]	22.8 [3.1]	0.5 [0.3]	1.9 [1.4]		32.5 [2.9]
1B	31.6 [12.1]	24.9 [10.3]	1.6 [2.4]	1.5 [2.0]		28.7 [9.1]
1C	34.6 [5.1]	15.7 [4.7]	9.9 [9.3]	21.7 [17.5]		16.7 [1.2]
2	25.0 [4.0]	4.0 [3.4]	20.5 [9.5]	26.5 [16.0]	2.3 [2.7]	19.1 [7.4]
3	31.1 [4.2]	5.0 [3.3]	16.6 [11.3]	29.0 [12.5]	2.6 [3.2]	12.8 [7.5]
4	20.9	0.3	3.5	63.2	7.3	2.5
5	41.1 [10.9]	4.7 [3.8]	27.5 [16.5]	17.5 [20.7]	1.6 [3.6]	7.5 [3.2]
6	27.2 [7.1]	0.2 [0.2]	17.2 [28.0]	43.6 [27.2]	3.2 [2.2]	8.6 [5.9]
7	36.4	4.6	42.1	15.4		1.5
8	21.1 [6.8]	1.6 [2.0]	30.8 [32.9]	35.9 [28.9]	2.4 [1.6]	7.7 [4.2]

taining medium to coarse silt, detrital quartz, and microfossils in a clay-rich matrix (plate 1). The mudstone is homogeneous and lacks structure and bioturbation (Biotrubation Index (BI) = 0).

The total GR response of Facies 1A is the highest of any facies (table 1) reflecting the high TOC (average 10.4 percent, table 1) and clay content (average 44.5 percent) of this facies.

The dominantly pelagic clay minerals and additional hemiplegic silt-sized siliciclastic grains settled out of suspension under anoxic, bottom water conditions with low energy.

Facies 1B: Organic-rich Mudstone with Pinch-and-Swell Lamination

Facies 1B consists of organic-rich mudstone containing medium to coarse silt-sized detrital quartz in a clay-rich matrix with pinch-and-swell laminations. The laminations are defined by medium silt to very-fine sand composed of skeletal debris from radiolaria, which are mostly spherical, silicified, and/or phosphatized (plate 1). The pinch-and-swell laminations range in thickness from 0.5 to 2.0 mm and are characterized either by sharply truncated or irregular surfaces. The bioturbation index of Facies 1B is 0. XRD analysis of Facies 1B reveals a mixture of quartz, feldspar, and clay (table 1).

Pelagic clay minerals with subordinate amounts of hemiplegic, silt-sized siliciclastic detritus were transported in suspension and deposited in anoxic bottom water conditions. In contrast, medium silt to very-fine sand sized radiolarian grains were probably deposited by turbidity currents, as is shown by grading upward sequences and laminae composed of silt and sand with sharp truncation surfaces at their base. Other laminae, defined by radiolarian debris, display gradational lower and upper contacts, suggesting that they settled out of suspension.

Facies 1C: Bioturbated, Organic-rich Calcareous/Dolomitic Mudstone

Facies 1C consists of bioturbated, calcareous/dolomitic mudstone with a siliciclastic and carbonate mud matrix and medium to coarse silt-sized detrital quartz grains and dolomite rhombs (plate 1). This mudstone is thinly-bedded (3- to 5 cm thick) and the bioturbation index ranges from 1 to 2. Ichnofacies include mainly *Chondrites* and subordinate *Planolites*. Burrows are typically filled with coarser sediments composed of detrital quartz in a clay-rich matrix.

High natural radioactivity (GR) and a high spectral GR response on the uranium curve in particular, are consistent with the relatively high TOC concentration in this facies (table 1). The thorium and potassium curves are consistent with relatively high clay and K-feldspar contents.

The inferred sedimentary processes of Facies 1C are similar to processes described for Facies 1B. Pelagic clay minerals and additional hemiplegic silt-sized detrital quartz settled out of suspension under

d deviation])				TOC	GR	
Clay		Other		(average wt %)	(nSv/h)	
I/S	Chlorite	Anhydrite	Gypsum		min	max
12.0 [*8.7*]				10.4	197.3	619.9
10.2 [*10.4*]	1.6 [*2.3*]			3.5	108.2	578.3
0.3 [*0.1*]				3.0	116.3	251.2
1.4 [*4.2*]				0.2	83.0	146.8
1.7 [*5.0*]	1.3 [*1.5*]			0.2	85.1	164.5
		2.3		0.2	66.7	122.0
0.2 [*0.5*]				0.2	56.6	96.0
				0.2	66.7	109.2
				0.3	70.5	83.9
0.1 [*0.1*]	0.1 [*0.3*]		0.4 [*0.8*]	0.2	83.0	121.1

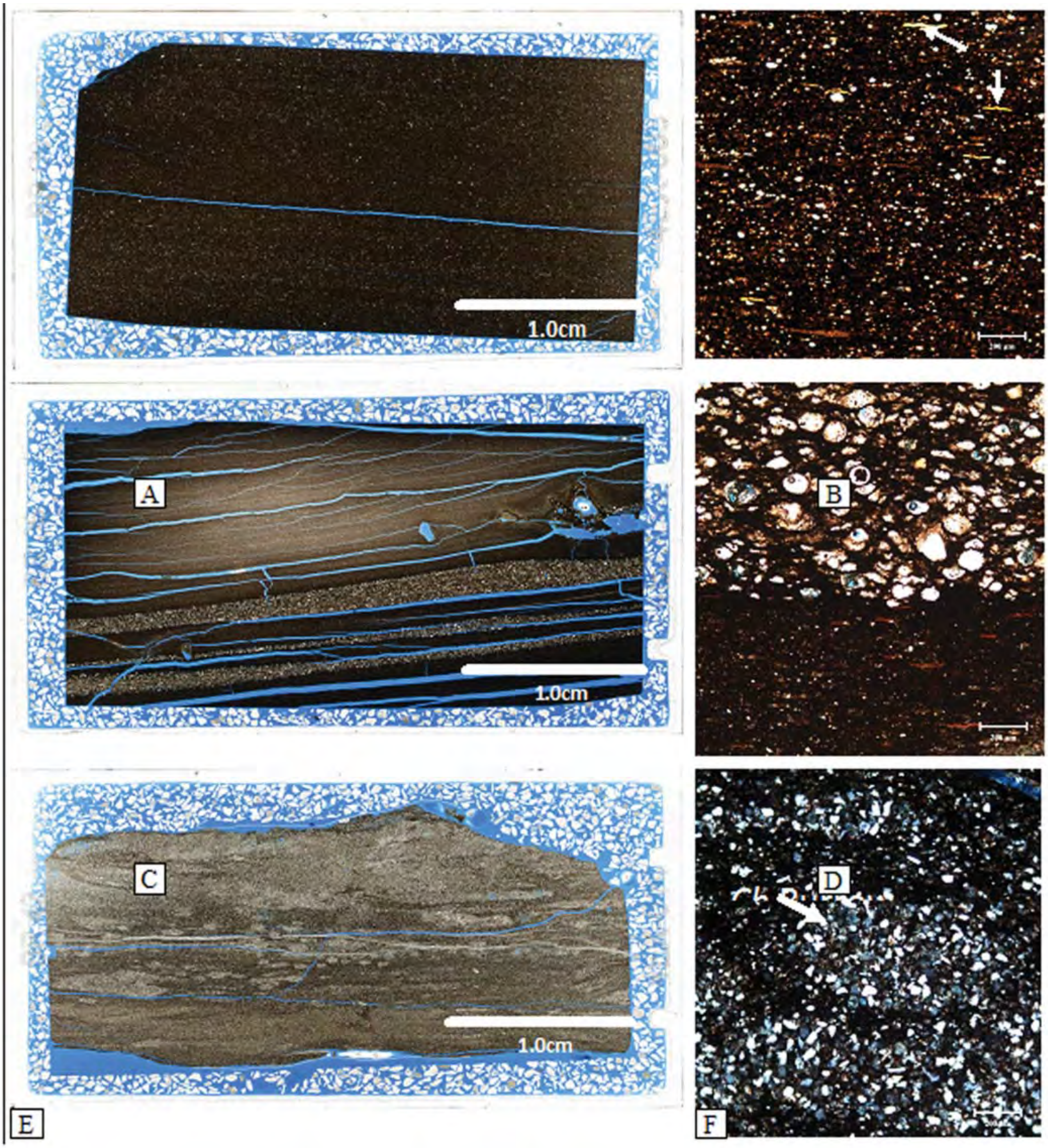


Plate 1. Examples of fine-grained organic rich facies (F1), the primary facies of the upper and lower Sappington Shales. (A) Thin section of structureless to faintly laminated organic rich mudstone (F1A). (B) Flattened Tasmanites cysts (white arrows) are common in this facies. (C, D) Sharp based, thinly bedded pinch-and-swell structures are common in facies 1B. Silt sized components are Radiolarians, Tasmanites and detrital quartz. (E,F) Bioturbated silty mudstone (F1C) with Chondrites (Ch) burrows is most common facies in upper Sappington Shale member. Burrows are commonly filled with silt-sized detrital quartz.

dysoxic conditions capable of supporting a sparse infauna. The presence of *Chondrites*, whose burrow system can occur within the sediment in the anaerobic zone (Bromley and Ekdale, 1984), is consistent with the sedimentation processes inferred for Facies 1C. *Chondrites* burrows indicate feeding and dwelling in the substrate, implying low rates of sediment accumulation (Schieber, 1999).

Facies 2: Bioturbated, Calcareous/Dolomitic Muddy Siltstone

Facies 2 consists of bioturbated, calcareous/dolomitic siltstone containing siliciclastic and carbonate mud, and medium to coarse silt-sized detrital quartz grains and dolomite rhombs (plate 2). This muddy siltstone is thinly bedded (3- to 5-cm thick) and locally

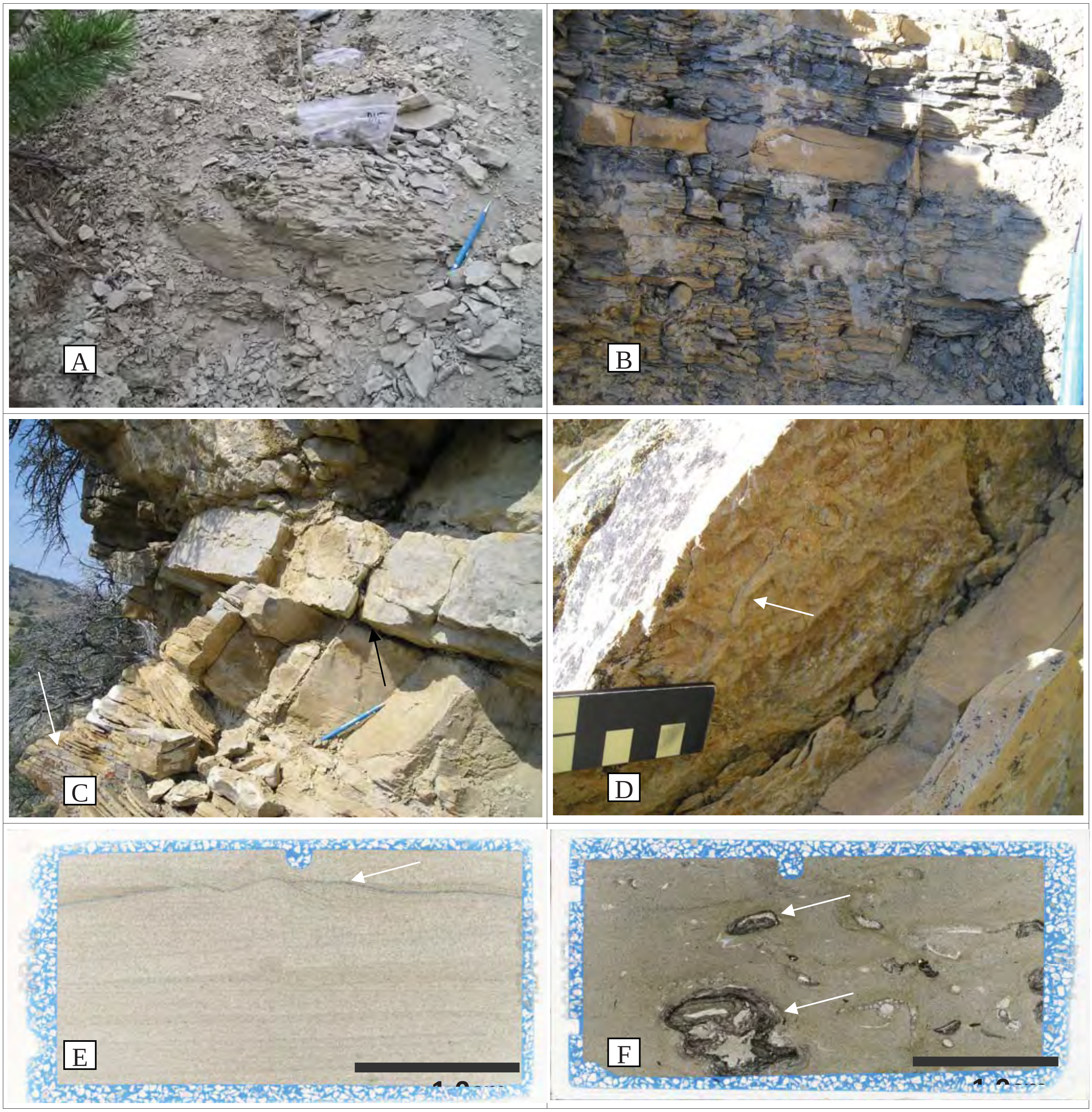


Plate 2: (A) Typical, slope forming exposure of the middle Sappington shale in the Dry Hollow section. Facies F2 is the predominant facies in this interval. (B) Interbedded siltstones and mudstones of Facies F3 in Bridger Range section. (C) Ripple-laminated silty dolostone (Facies F6, bed with pencil leaning against) and parallel laminate silty dolostone (Facies F7, white arrow) just below contact to overlying Lodgepole Limestone (black arrow) in the Dry Hollow section. (D) Bioturbated, ripple laminated siltstone (Facies F5) in the Bridger Range section. The predominant traces visible on the bedding plane are Palaeophycus traces (white arrow); scale bar in centimeter. (E) The low angle bedforms in the lower part of this thin section are HCS beds, typical for Facies F4; the ripple laminated upper part (white arrow) of the thin section is indicative for Facies F6. (F) Thin section photomicrograph of Facies F8; Algal-encrusted Oncolites (white arrows) and other bioclastic material embedded in fine-grained carbonate matrix are common constituents in this facies.

interlaminated with silty dolostone. The bioturbation index ranges from 0 to 3, and ichnofacies identified include *Palaeophycus*, *Planolites*, and *Teichichnus*. Burrows are often filled with coarser sediments composed of detrital quartz and dolomite rhombs in a clay matrix.

The total GR response in Facies 2 is largely controlled by potassium and thorium, reflecting the overall abundance of clay in this facies and the relatively low organic content (table 1).

The dominant sedimentary process inferred for Facies 2 is clay minerals and silt-sized detrital quartz settling out of suspension in a low energy environment. The presence of *Cruziana* ichnofacies, including *Planolites* and *Teichichnus* burrows suggests the sublittoral zone in a marine environment (Catuneanu, O., 2006; MacEachern et al., 2010).

Facies 3: Bioturbated, Wavy-laminated/Lenticular Silty Dolostone with Interbedded/Interlaminated Mudstone

Facies 3 consists of thinly bedded (1- to 5-cm), wavy-laminated/lenticular silty dolostone interbedded with mudstone that has been bioturbated (plate 2). The silty dolostone contains medium silt to very-fine sand-sized grains of detrital quartz and authigenic dolomite rhombs. The bioturbation index ranges from 0 to 1, and ichnofacies identified include *Planolites* and *Skolithos*.

The natural GR for Facies 3 is lower than the GR response of Facies 1, reflecting the overall low organic content in this facies. The relative high clay content, in particular in the mudstone interbeds, results in an intermediate GR response, very similar to Facies 2 (table 1).

Facies 3 is interpreted as a product of storm deposition during which medium silt to very fine-grained sand-sized particles of detrital quartz were introduced and deposited above the storm wave base but below the fair-weather wave base. The sharp basal contact characteristic of silty dolostone beds in Facies 3, the presence of combined flow ripples in the dolostone, and the presence of overlying mud drapes are consistent with deposition associated with a waning storm event. The mudstone interbeds were deposited during periods of quiescence between storm events that allowed for settling of the fine-grained suspended load in an environment below the fair-weather wave base.

Facies 4: Hummocky Cross-Stratified (HCS) Silty Dolostone

Facies 4 consists of hummocky cross-stratified (HCS) silty dolostone containing medium silt to very fine-sand-sized detrital quartz grains, and abundant authigenic dolomite rhombs and finely crystalline calcite cement (plate 2). HCS beds are typically a few decimeters thick and each bed displays stacked, low-angled cross-laminations with sharp basal contacts. Sedimentary structures are well preserved and bioturbation is absent (BI = 0). Abundant soft sediment deformation is present in some of the measured sections. Total and spectral gamma ray responses are consistent with low TOC, low clay, and low K-feldspar contents (table 1).

Facies 4 is interpreted as a product of deposition resulting from storm-generated combined flow currents and associated relaxation (geostrophic?) flow. HCS beds with sharp basal contacts represent the main storm deposits; HCS beds associated with beds containing soft sediment deformation represent rapid sedimentation through storm-related processes.

Facies 5: Bioturbated, Ripple-Laminated, Calcareous/Dolomitic Siltstone-Silty Dolostone

Facies 5 consists of calcareous/dolomitic siltstone and silty dolostone that has ripple laminations and has been bioturbated. It contains medium silt to very fine sand-sized detrital quartz grains, and abundant authigenic dolomite rhombs and finely crystalline calcite cement (plate 2). The silty dolostone beds are wave-rippled, thinly bedded (a few cm thick), and laterally continuous. Current-ripples occur locally. Some rippled beds are separated by silt partings or mud drapes. The trace fossil assemblage is the most diverse of any facies (BI ranges from 1 to 3), including *Chondrites*, *Planolites, Teichichnus, Thalassinoides, and Rosselia*. The total and spectral GR are consistent with low TOC, and the presence of clay and K-feldspar in this facies (table 1).

The dominant sedimentary process inferred for Facies 5 is episodic wave reworking of medium silt to very fine-sand-sized detrital quartz grains above the fair-weather wave base. The presence of a variable, mainly *Cruziana,* ichnofacies implies deposition under oxygenated conditions in a normal, sublittoral, marine environment (Catuneanu, O., 2006; MacEachern et al., 2010).

Facies 6: Ripple-Laminated Silty Dolostone

Facies 6 consists of ripple-laminated silty dolostone containing medium silt to very fine sand-sized detrital quartz grains, and authigenic dolomite rhombs and finely crystalline calcite cement (plate 2). Beds of silty dolostone are generally 1- to 5-cm thick and wavy, with infrequent current ripples. Locally, the rippled beds contain brachiopod and echinoderm skeletal debris, and in places the beds are separated by silt partings or mud drapes. Bioturbation is absent in Facies 6 (BI = 0).

The total GR response for this facies is consistent with its low organic content, and the thorium and potassium response curves are consistent with the measured total clay and K-spar percentages (table 1).

The dominant sedimentary process inferred for Facies 6 is wave reworking of medium silt to very fine sand-sized detrital quartz grains above the fair-weather wave base. The absence of ichnofacies in Facies 6 implies relatively high rates of sediment accumulation and/or physical reworking.

Facies 7: Planar-Bedded Silty Dolostone

Facies 7 consists of planar-bedded, silty dolostone containing coarse silt to very fine sand-sized detrital quartz grains and dolomite rhombs (plate 2). Secondary porosity due to dissolution of dolomite rhombs is common. Dolostone beds are typically 1- to 2-cm thick, parallel, and separated by sharp contacts. Internally, beds display planar horizontal lamination. The bioturbation index of Facies 7 is 0.

The spectral GR log for uranium shows little variation, consistent with the low overall organic carbon content (table 1). The response of the thorium and potassium curves is consistent with low overall clay and K-spar percentages.

Transport and deposition of silt and sand-sized particles was associated with high energy flows that included high concentrations of suspended load and bedload. The absence of bioturbation also implies relatively high rates of sediment accumulation or high energy levels, consistent with the inferred deposition in a shallow marine environment.

Facies 8: Oncolitic, Fossil-bearing Floatstone

Facies 8 consists of floatstone rich in oncoids and fossil debris. Oncoids commonly contain brachiopods, bryozoans, and other irregular-shaped carbonate clasts as nuclei, and multiple generations of algal coatings can be either rounded or irregular. Most oncoids are a few cm in size but individual oncoids as large as ~10 cm in diameter were observed. Fossil debris consists of brachiopods, bryozoans, echinoderms, and subordinate gastropods and ostracods. These fossils typically are delicately preserved, although fragmented bioclasts were identified as well. The matrix consists of siliciclastic and carbonate mud, medium to coarse silt-sized detrital quartz grains and dolomite rhombs, calcite cement, and organic debris (plate 2).

The floatstone beds are massive and thicker (~3 m) at the Bridger Range (BR) locality, and they are interbedded with calcareous/dolomitic mudstone beds (Facies 3) at the Dry Hollow (DH) locality. At the BR locality, the oncolitic, fossil-bearing floatstone becomes less fossiliferous and more fine-grained and matrix dominant upsection. The bioturbation index of Facies 8 ranges from 0 to 1.

The total carbonate content of Facies 8 is nearly 70 percent (table 1). The total GR log values for this facies are consistently low, reflecting the absence of any significant organic carbon in the facies and the high carbonate content (table 1).

Facies 8 is interpreted as being deposited along the flanks of a carbonate build-up where heterotrophic filter feeders were abundant within a biologically-productive photic zone. The abundance of delicately-preserved fossils in this facies suggests deposition of the floatstone was proximal to the source of the fossil debris, and the multiple truncation surfaces and generations of encrusting laminae in many of the irregularly shapes oncoids indicate occasional reworking (Flügel, 2010). The fine-grained matrix in the floatstone implies deposition in a low energy environment, as well as sediment baffling and binding by organisms living in the carbonate build-up.

4. FACIES ASSOCIATIONS

Based on our facies analysis of the Sappington Formation in central Montana, we define four facies associations and interpret these as representing the specific sedimentary environments and processes involved.

Facies Association 1: Partly Restricted Offshore Marine Environment

Facies Association 1 (FA1) consists exclusively of Facies 1 and is interpreted as a partly restricted offshore marine environment. FA1 forms the lower and upper members of the Sappington Formation at the DH and BR localities, and the upper member in the BS localities (fig. 5).

The absence of bioturbation and the high organic content in the lower mudstone member suggests that anoxic conditions prevailed long enough to severely limit biological activities. Overall the lower mudstone member is thicker, more organic rich, and has a higher

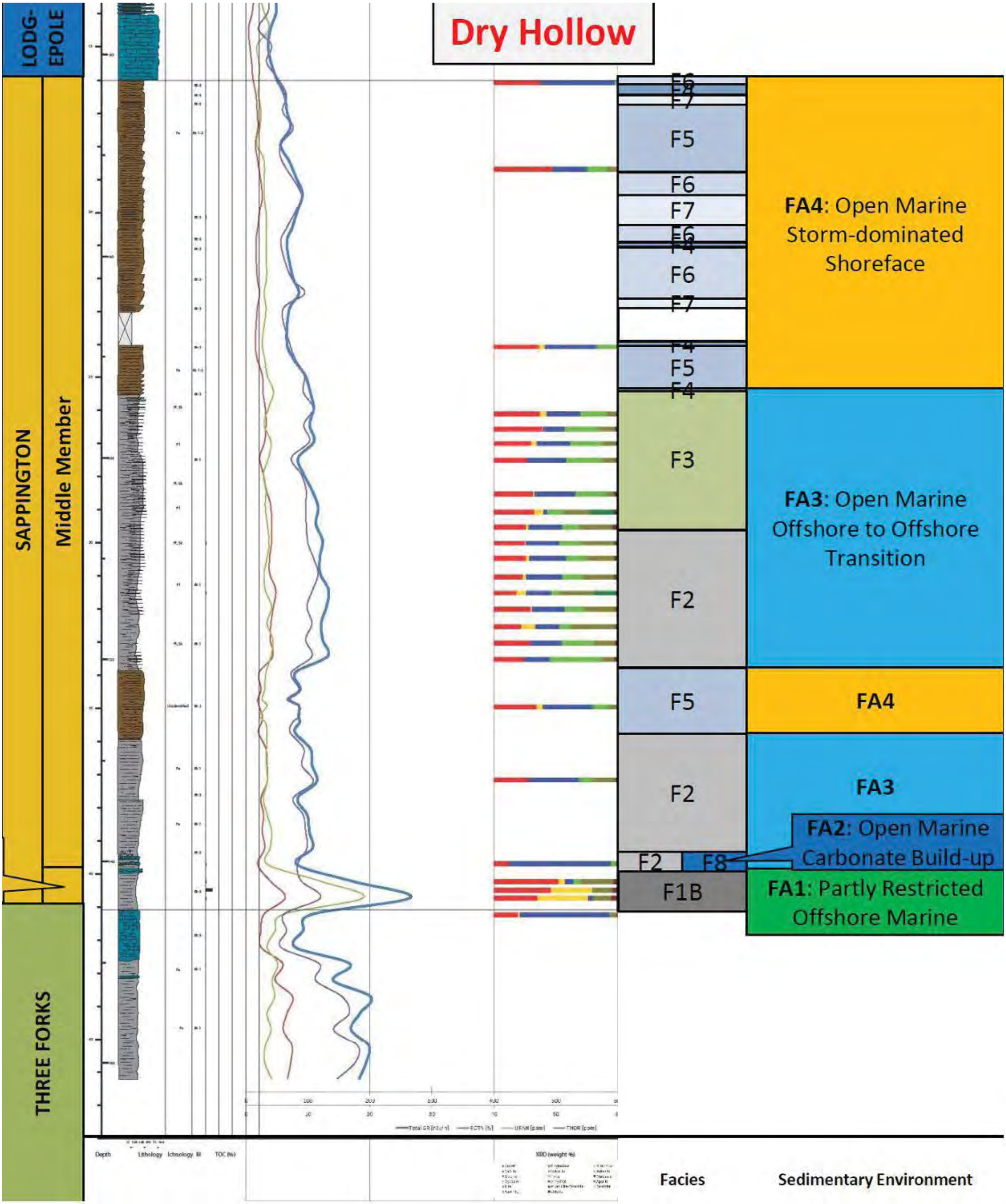
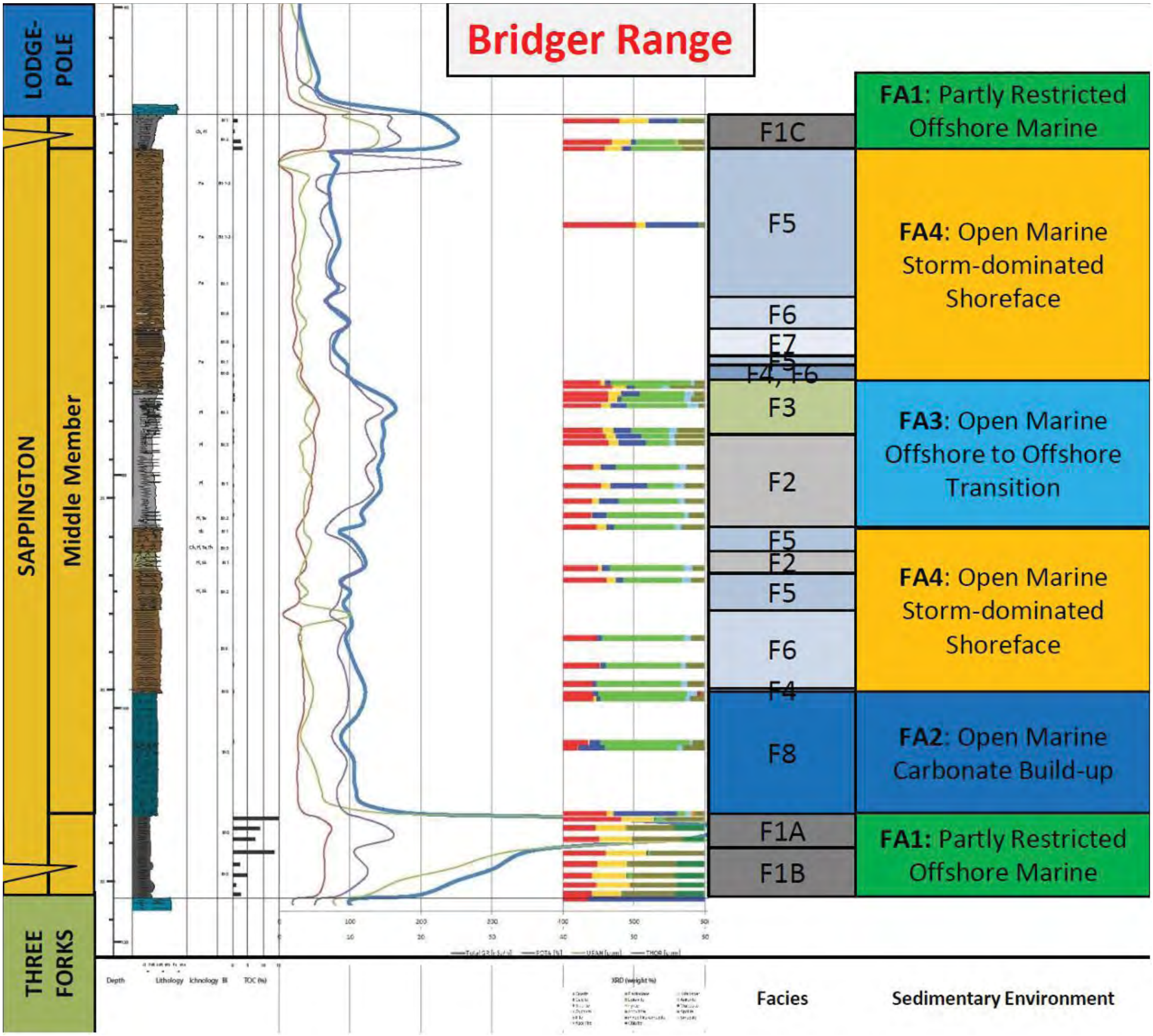

Dry Hollow
LODG-EPOLE
SAPPINGTON
Middle Member
THREE FORKS
F6
F7
F5
F6
F7
F6
F4
F6
F7
F4
F5
F4
F3
F2
F5
F2
F2
F8
F1B
FA4: Open Marine Storm-dominated Shoreface
FA3: Open Marine Offshore to Offshore Transition
FA4
FA3
FA2: Open Marine Carbonate Build-up
FA1: Partly Restricted Offshore Marine
Depth
Lithology
Ichnology
BI
TOC (%)
XRD (weight %)
Facies
Sedimentary Environment

Figure 5A



B

Figure 5. Stratigraphic sections for the Dry Hollow section (left) and the Bridger Range section (right). In each display, the column farthest to the left lists the stratigraphic formation and members, the second column from the left displays the scale in meters; the third column from the left grain size and lithology; in the next column we list ichnology and bioturbation indices (BI), followed by TOC in weight percent; the line graph column are the total and spectral GR values; the bar graph to the right of the GR lists the bulk mineralogic data in weight percent; in the second column from the right we display the facies, and in the column farthest to the right we show the inferred depositional environment and facies association. For example, the lower Sappington shale is a good marker bed and recognizable by the highest GR values and TOC in both sections. It is entirely composed of Facies F1A and F1B (facies column, second from the right) and is interpreted to have been deposited in a restricted marine shelf environment (depositional environment column to the far right).

clay content at the BR locality. Its characteristics are:

- 2.1 m thick
- average TOC of 6.97 percent
- maximum TOC of 14.78 percent, and
- average 48.3 percent clay

This is in comparison to the DH locality, which has the following characteristics:

- 1.0 m thick;
- average TOC of 1.09 percent
- maximum TOC of 2.74 percent, and
- average 23.6 percent clay

We interpret these variations as the result of increasingly anoxic conditions from a marginal basin position at the DH section to a more central basin position closer to the BR section.

The presence of bioturbation and relatively high organic content in the upper mudstone member at the BR and BS localities (combined average TOC of 3.04 percent) suggests deposition in an offshore to offshore transitional environment under dysoxic conditions capable of supporting a sparse infauna.

Facies Association 2: Open Marine Carbonate Build-up Environment

Facies Association 2 (FA2) consists of oncolitic, fossil-bearing floatstone (Facies 8). It is restricted to the lower unit of the middle Sappington Formation (fig. 5), and is interpreted as having been deposited along the flanks of a carbonate build-up.

The diverse suite of delicately-preserved heterotrophic filter feeders, including brachiopods, bryozoans, and echinoderms observed in the floatstone implies deposition in an open marine environment with relatively consistent temperature, salinity, and current activity, and relatively low levels of turbidity. The presence of photosynthesizers strongly suggests a shallow subtidal environment within the photic zone. The abundance of fine-grained sediments in the floatstone is interpreted to have been in part trapped by these organisms but also having settled out of solution in a low energy environment. The very loose packing of the allochems, which generally float in a fine-grained matrix of mixed siliciclastic and carbonate sediment, suggests that the facies association represents an allochthonous mixture that includes contributions from a nearby, active carbonate build-up as well as from continental weathering.

Facies Association 3: Open Marine Offshore to Offshore Transitional Environment

Facies Association 3 (FA3) includes bioturbated, calcareous/dolomitic muddy siltstone (Facies 2) and wavy-laminated/lenticular silty dolostone with interbedded/interlaminated mudstone (Facies 3). FA3 occurs in the lower and middle units of the middle Sappington member at the Dry Hollow (DH) and Bridger Range (BR) localities (fig. 5). FA3 is interpreted as being deposited in an offshore to offshore transitional environment. The presence of a *Cruziana* ichnofacies assemblage and relatively low TOC content (< 0.65 percent) in the middle mudstone member suggests that the basin was generally well oxygenated and normal marine conditions prevailed. The calcareous/dolomitic muddy siltstone generally represents deposition in a low energy offshore environment and the deposition of silty dolostone with interbedded/interlaminated mudstone reflects a decrease in mud content associated with shallower water and higher energy of an offshore transition zone.

Facies Association 4: Open Marine Storm-dominated Shoreface Environment

Facies Association 4 (FA4) includes HCS silty dolostone (Facies 4), bioturbated, ripple-laminated silty dolostone (Facies 5), ripple-laminated silty dolostone (Facies 6), and planar bedded silty dolostone (Facies 7). It occurs in the lower and upper units of the middle Sappington member at the DH, BR, and Big Snowy (BS) localities (fig. 5). FA4 is interpreted as reflecting deposition in the lower to middle shoreface along a storm-dominated coastline.

The high degree of bioturbation in the ripple-laminated silty dolostone within the lower unit of the middle Sappington Formation suggests a rate of deposition that was low enough to allow for the establishment of a thriving infauna.

In the upper unit of the middle Sappington member at the DH and BR localities, FA4 is represented by a vertical succession of facies interpreted to reflect deposition by individual storm events, as well as an amalgamation of depositional events from multiple storms. A sharp basal contact of HCS silty dolostone is interpreted to represent the initial energy increase associated with the storm. The overlying silty dolostone containing HCS, combined flow ripples, and local soft sediment deformation are interpreted to represent storm deposition. All three structures are consistent with rapid sedimentation during storm events. Locally observed unidirectional current ripples that cap the succession imply decreasing energy levels possibly associated with post-storm relaxation and/or geostrophic flow. Some beds of silty dolostone with HCS are overlain by current ripple laminae; others are overlain by planar stratification. Both transitions are interpreted to be storm-related. Some successions of planar-bedded silty dolostone are overlain by current ripples that transition upward into wave ripples, interpreted to reflect relaxation or geostrophic current flow associated with waning storm events, followed by wave reworking of the uppermost deposit. Centimeter-scale interbeds of bioturbated, ripple-laminated silty dolostone are interpreted to reflect wave reworking and infaunal colonization during intervening periods of relatively fair weather.

5. SEQUENCE STRATIGRAPHIC FRAMEWORK

5.1. Dry Hollow (DH) Section

The Sappington Formation at the Dry Hollow locality (DH) is approximately 25 meters thick (fig. 6). The basal Sappington-Three Forks contact is placed between a crinoidal limestone bed of the Three Forks Formation and the organic rich mudstone bed of the lower Sappington member. This sharp contact is a regional unconformity marked by the absence of several conodont zones (fig. 3), and is interpreted as a sequence boundary (SB).

The lower organic-rich mudstone of the Sappington Formation represents an abrupt landward shift of facies from the underlying shallow marine limestone of the Three Forks Formation. Within the lower organic-rich mudstone, facies are interpreted to be retrogradationally stacked and accompanied by an increase in clay content and TOC. A thin bed with the highest clay content and the highest total GR measurement was observed immediately below the contact with the overlying oncolitic, fossiliferous floatstone. This high GR interval is interpreted to contain the maximum flooding surface (mfs) that can be correlated across the region. The retrogradationally stacked facies of the lower shale, bounded by the composite SB/fs at the base and mfs at the top, is interpreted as a Transgressive Systems Tract (TST).

The lower unit of the middle Sappington member, consisting of oncolitic, fossil-bearing floatstone, calcareous/dolomitic muddy siltstone, and silty dolostone, represents a shoaling-upward succession above the mfs. The interpreted sedimentary environments shoal from carbonate build-up, to offshore transition, and finally to shoreface deposits in an open marine environment. The shoaling upward succession is topped by a regional correlative flooding surface (fs), and is interpreted as a Highstand Systems Tract (HST).

An abrupt shift of facies from the silty dolostone to the muddy siltstone of the middle unit of the middle Sappington member is interpreted as representing a composite SB and fs. The SB marks the end of the first depositional sequence reflected by the most basinward progradation of the sedimentary system. The subsequent flooding surface indicates the beginning of base-level rise and the base of the second depositional sequence. The muddy siltstone generally fines upward, and the bed with the highest clay content is interpreted as containing the mfs, forming the top of a second TST.

The muddy siltstone transitions into interbedded/interlaminated silty dolostone. This gradual coarsening of facies above the mfs is interpreted as the onset of the next shoaling-upward succession. Within this shoaling-upward succession, facies are stacked progradationally, also reflected by the decrease in total GR. An abrupt shift in facies, at the upper contact of the shoaling-upward succession is interpreted as a forced regressive surface (FRS) that can be correlated across the region. The shoaling-upward succession, bounded by the mfs at the base and FRS at the top, is interpreted as representing a HST of the second depositional sequence.

The upper unit of the middle Sappington member is interpreted as another shoaling-upward succession. Above the FRS, facies are stacked aggradationally to progradationally, consistent with the overall decrease in total GR response. The upper contact between this shoaling-upward succession and the overlying Lodgepole limestone is interpreted to be a SB. It is marked by an abrupt shift of facies from silty dolostone to limestone. This youngest shoaling-upward succession, bounded by the FRS at the base and SB at the top, is interpreted as representing a Falling Stage Systems Tract (FSST).

5.2 Bridger Range (BR) Locality

The Sappington Formation at the Bridger Range (BR) locality is approximately 20.5 meters thick (fig. 7). It is bounded at the base by crinoidal limestone of the Three Forks Formation and at the top by bioclastic limestone of the Lodgepole Formation (Madison Group). The basal Sappington mudstone in the Bridger section is the most organic-rich observed in this study. The contact between this organic-rich mudstone and the underlying carbonate beds of the Three Forks Formation is interpreted to be a flooding surface (fs) with regional significance, following a well-constrained hiatus (fig. 3). Within the lower organic-rich mudstone, facies fine upward from the base to 2 m above the flooding surface and are accompanied by an overall increase in TOC, clay content, and total GR response.

Figure 6. Cycles, Surfaces, and Systems Tracts of the Sappington Formation in the Dry Hollow Section. Inset map in upper left corner shows the equivalent contact between the Sappington Formation and the Lodgepole Limestone at an outcrop in Milligan Canyon, about one km to the east of the measured section and the outcrop photograph. We recognized two full cycles in this outcrop, bound by sequence boundaries (SB); the lower composed of a transgressive systems tract (TST) and, separated by a maximum flooding surface (mfs), a highstand systems tract (HST); the upper composed of TST, HST and Falling Stage Systems Tract (FSST). The surface marking the transition from HST to FSST is interpreted as Forced Regressive Surface (FRS).

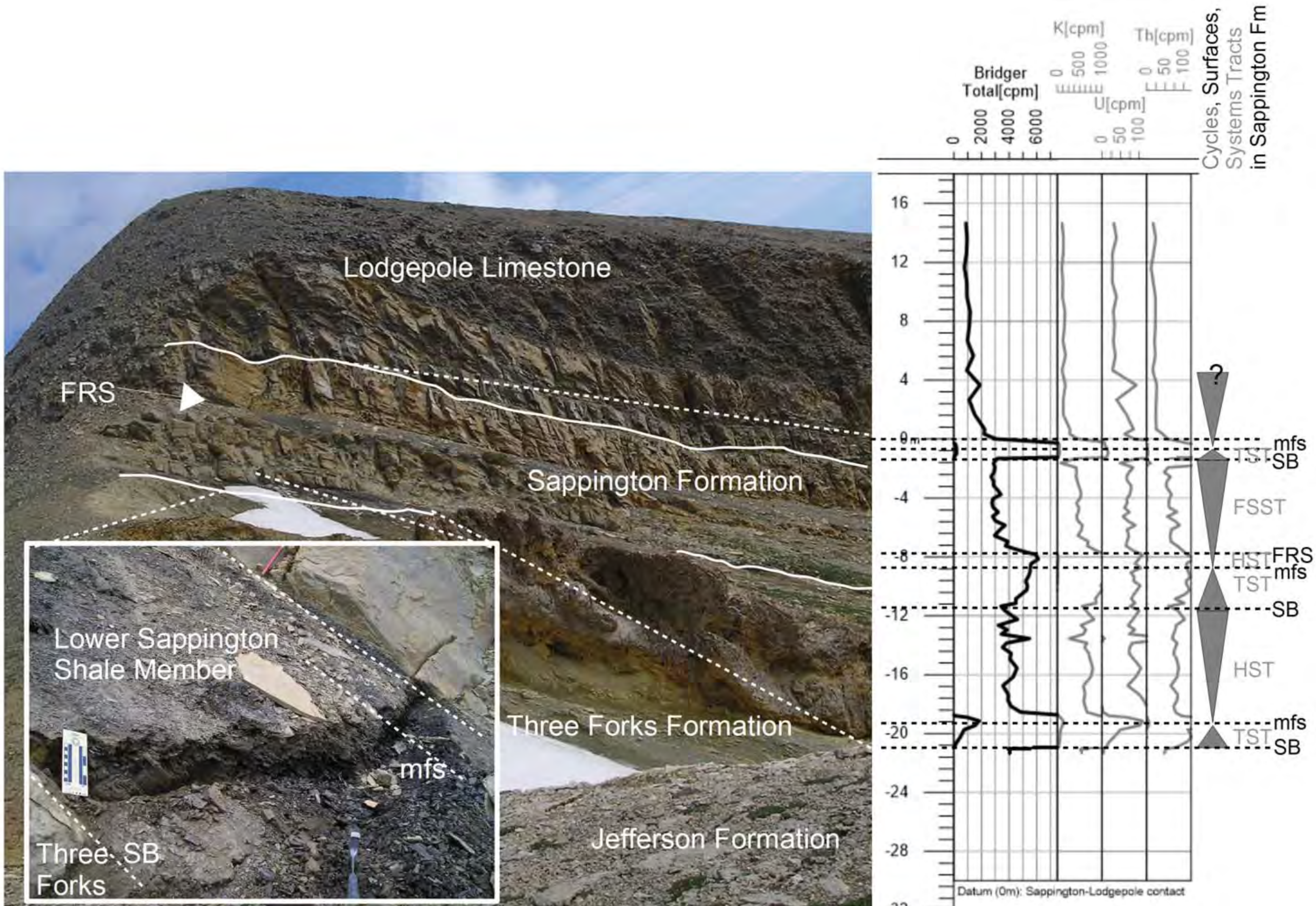


Figure 7. Cycles, Surfaces, and Systems Tracts of the Sappington Formation in the Hardscrabble Peak section in the Bridger Range. Similar to the Dry Hollow section, we recognized two full cycles in this outcrop, bound by sequence boundaries (SB); the lower composed of a transgressive systems tract (TST) and, separated by a maximum flooding surface (mfs), a highstand systems tract (HST); the upper composed of TST, HST, and Falling Stage Systems Tract (FSST). The surface marking the transition from HST to FSST is interpreted as a Forced Regressive Surface (FRS). Inset map in lower left corner shows the lower Sappington Shale member. The contact with the underlying Three Forks Formation is interpreted as SB; the mfs is located approximately 0.5 m below the contact to the middle Sappington member in an interval with notably higher clay content.

A thin interval approximately 2 m above the basal contact contains the highest clay mineral percentage and is bracketed by the highest measured TOC values. This interval is interpreted to contain a mfs that can be correlated across the region. The lower part of the lower member, bound below by the fs and above by the mfs, is interpreted to represent a TST.

An increase of silt content in the upper part of the lower member, above the mfs, is interpreted to represent the onset of shoaling. Shoaling continued upward during deposition of the lower unit of the middle Sappington member, which consists of oncolitic, fossil-bearing floatstone, calcareous/dolomitic muddy siltstone, and silty dolostone. The change in lithologies from floatstone to silty dolostone is interpreted to represent a transition from a depositional environment dominated by carbonates to one dominated by clastic deposition. Notably, extrabasinal siliciclastic silt is abundant in the floatstone, suggesting that – though subordinate to deposition by intrabasinal carbonate – extrabasinal contributions from continental weathering were still significant. Within the overall shoaling-upward succession, facies are stacked progradationally. The inferred sedimentary environments changed from restricted marine shelf, to normal marine carbonate build-up, to shoreface, consistent with the observed overall decrease in total GR measurements.

An abrupt shift of facies above the shoaling upward succession, from silty dolostone to muddy siltstone, is interpreted to represent a composite SB/fs, marking the end of base-level fall and the end of the first depositional sequence, as well as the beginning of a regional base-level rise associated with the onset of the second depositional sequence. This shoaling-upward succession, bounded by the mfs at the base and SB/fs at the top, is interpreted as a HST.

Above the SB/fs, within the fining-upward succession, facies are stacked retrogradationally, consistent with an increase in total GR measurements up section. We interpret an interval with high clay content to coincide with the mfs of the second depositional sequence. These retrogradationally stacked, open marine offshore deposits are interpreted as a TST, bounded by a SB/fs at the base and a mfs at the top.

Above the mfs, the middle Sappington member consists of calcareous/dolomitic muddy siltstone interbedded/interlaminated with silty dolostone that represents a shoaling-upward succession. The contact between the shoaling-upward succession and the overlying silty dolostone is interpreted as a FRS, as suggested by an abrupt basinward shift of facies from muddy siltstone to silty dolostone. This shoaling-upward succession, bounded by the mfs at the base and FRS at the top, is interpreted as the HST of the second depositional sequence.

The upper unit of the middle Sappington member above the FRS consists of a shoaling-upward succession. Within this shoaling-upward succession, silty-dolostone facies are stacked aggradationally to progradationally. The abrupt shift of facies from silty dolostone to the bioturbated, organic-rich calcareous/dolomitic mudstone of the upper Sappington member is interpreted as a SB that marks the end of base-level fall. This shoaling-upward succession, bounded by the FRS at the base and SB at the top, is interpreted as a FSST.

Following this most basinward progradation of the depositional system and the formation of a SB at the contact between the middle Sappington member and the upper organic-rich mudstone member, the abrupt landward shift of facies above this contact marks the next onset of flooding. Within the upper, organic-rich mudstone, facies are stacked retrogradationally and are accompanied by an increase in total GR measurements. We interpreted the surface at the inflection point from retrogradational to progradational stacked facies as the maximum flooding surface. The retrogradationally stacked, restricted offshore marine facies are interpreted as a TST bounded by the composite SB/fs at the base and mfs at the top.

A coarsening-upward succession within the upper organic-rich mudstone, above the mfs, represents the onset of a subsequent HST that continues in the overlying Lodgepole Limestone.

5.3 Big Snowy (BS) Locality (Incomplete Section)

In the Big Snowy locality a 7.5-meter thick interval of the upper Sappington Formation and the lowermost Lodgepole Limestone are exposed in a road cut (fig. 8).

The upper unit of the middle Sappington member consists of a shoaling upward succession of silty dolostone. The contact between the shoaling-upward suc-

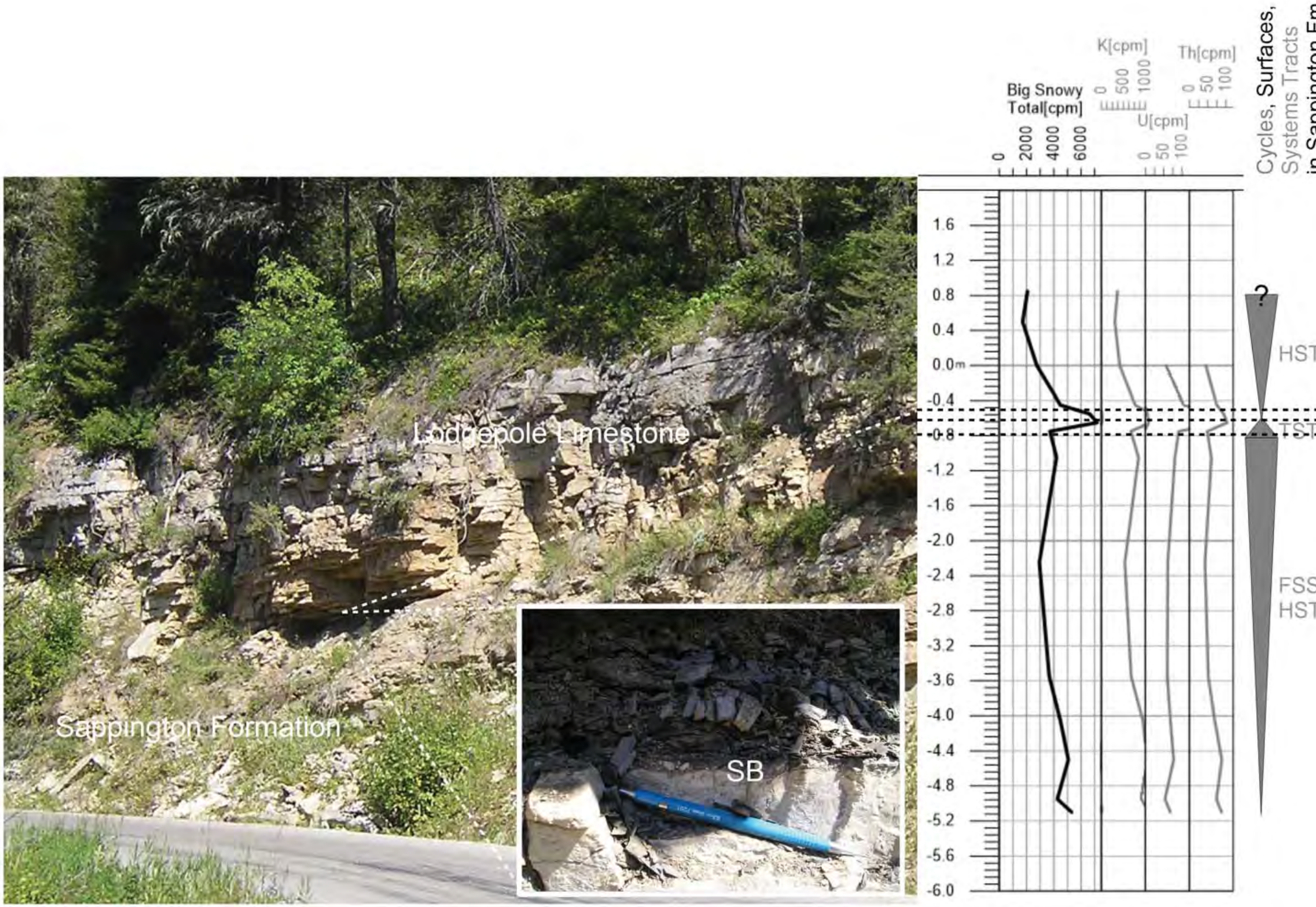


Figure 8. Cycles, Surfaces, and Systems Tracts of the Sappington Formation in the Big Snowy section. In the Big Snowy section only the uppermost part of the Sappington Froma-tion is exposed in an outcrop along Crystal Lake Road. The exposed strata of the middle Sappington member is interpreted as part of a highstand systems tract (HST) or Falling Stage Systems Tract (FSST), bound by a sequence boundary (SB) on top. The thin upper Sappington Shale Member is interpreted as transgressive systems tract (TST), bound by a SB at the base and a maximum flooding surface (mfs) on top. The mfs is located at the interval with the highest clay content. The transition to the Lodgepole formation is interpreted as part of the next HST. The inset map in the lower right shows the SB between the middle Sappington HST and the upper Sappington TST. Rip-up clasts along the wavy surface are the result of wave reworking during early transgression.

cession and the overlying organic-rich upper mudstone is an erosional, wavy surface with abundant mud rip-up clasts. This surface was interpreted as a sequence boundary (SB) and a transgressive wave ravinement surface (TRS). The shoaling-upward succession is interpreted as either the HST or FSST of the second Sappington sequence.

The upper organic-rich mudstone of the Sappington Formation represents an abrupt landward shift of facies overlying the TRS. Within the upper organic-rich mudstone, facies are stacked retrogradationally, which is also reflected by an increase in total GR. A maximum flooding surface (mfs) was placed where the most fine-grained mudstone and the highest total GR measurements were observed. The retrogradationally stacked facies is interpreted to represent a thin TST in a basin marginal position bounded by the TRS at the base and mfs at the top. The subsequent coarsening-upward succession within the upper organic-rich mudstone, above the mfs, was interpreted as a HST with an upper bounding surface in the overlying Lodgepole Limestone.

6. CONCLUSIONS

Detailed sedimentologic, petrographic, petrophysical, and geochemical analyses of the Devonian-Mississippian Sappington Formation and its equivalents in central and southwestern Montana leads to the following conclusions:

(1) The Sappington Formation can be subdivided into eight facies:

(F1) organic-rich mudstones
(F2) bioturbated, calcareous/dolomitic muddy siltstone
(F3) bioturbated, wavy-laminated silty dolostone with interbedded/interlaminated mudstone
(F4) hummocky cross-stratified silty dolostone
(F5) bioturbated, ripple-laminated, calcareous/dolomitic siltstone-silty dolostone
(F6) ripple-laminated silty dolostone
(F7) planar-bedded silty dolostone, and
(F8) oncolitic, fossiliferous floatstone

(2) The Sappington Formation is interpreted as representing four facies associations:

(FA1) a partly restricted offshore marine environment
(FA2) an open marine carbonate build-up environment
(FA3) an open marine offshore to offshore transition environment, and
(FA4) an open marine storm-dominated shoreface environment

(3) Based on a depositional sequence approach, the Sappington Formation is interpreted to represent two higher order sequences with an additional sequence continuing into the overlying Lodgepole Limestone. The oldest depositional sequence contains a Transgressive Systems Tract and Highstand Systems Tract, the second depositional sequence contains a Transgressive Systems Tract, Highstand Systems Tract, and Falling Stage Systems Tract, and the youngest depositional sequence contains a Transgressive Systems Tract and Highstand Systems Tract.

7. ACKNOWLEDGMENTS

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MESOARCHEAN PLUTONISM IN THE SOUTH SNOWY BLOCK (YELLOWSTONE NATIONAL PARK): NEW EVIDENCE FOR AN OLD ARC

Paul Mueller,[1] David Mogk,[2] Darrell Henry,[3] David Foster,[1] Tyler Berndt,[2] Timothy Grip,[3] Matt Hanson,[4] Alisa Kotash,[5] Patrick Maloney,[6] Kate Philbrick,[7] and Bryant Ware[8]

Participants with Research Experience for Undergraduates are listed alphabetically

[1]University of Florida, [2]Montana State University, [3]Louisiana State University, [3]University Massachusetts-Amherst, [4]Whitman College, [5]SUNY Genesco, [6]Franklin and Marshall, [7]Wellesley College, [8]University of Wyoming

ABSTRACT

Petrographic, geochemical, and geochronologic data for a suite of Mesoarchean plutonic rocks that intrude the Jardine metasedimentary sequence (JMS) in the South Snowy block (SSB) of the Beartooth Mountains indicate that they are likely arc-related magmas. The plutons range from distinct, bulbous bodies that crosscut and, in some cases, inject and migmatize layers in the adjacent JMS to batholith-scale aggregations of sheets. The JMS exposed in the SSB, including sections in Yellowstone National Park, are contiguous with the gold-bearing, low-grade, metasedimentary rocks of the Jardine district that have been mined intermittently since the 1880's. The limited age-range of these plutons (2.79-2.83 Ga) places a firm upper limit on the depositional age of the JMS at 2.79 Ga, and extends the geographic range and emplacement styles of the arc-related magmatism that dominates the Beartooth-Bighorn magmatic zone of the Wyoming Province.

INTRODUCTION

The Beartooth-Bighorn magmatic zone (BBMZ) of the northern Wyoming Province is volumetrically dominated by plutonic and meta-plutonic rocks of the TTG association (tonalite-trondhjemite-granodiorite) (Mueller and Frost, 2006). Recent U-Pb zircon geochronology and geochemistry have shown that this vast volume of igneous rocks was intruded in a relatively narrow time interval between 2.8 and 2.9 Ga as a result of subduction-driven-magmatic processes (e.g., Wooden and Mueller, 1988; Frost and Fanning, 2006; Mueller et al., 2010). Most exposures in the BBMZ indicate intrusion occurred in a ductile, mid-crustal environment in which magmas of a range of compositions (diorite to granite) were present simultaneously. Mutually crosscutting relationships between magmas of distinct compositions are common.

In contrast, plutons exposed in the southwestern corner of the Beartooth Mountains (South Snowy block, SSB, fig. 1) range from bulbous, epizonal forms to much larger sill- or sheet-like bodies (e.g., Berndt et al., 2012; Grip et al., 2012; Maloney et al., 2011; Philbrick et al., 2011). In all cases, however, they intrude the low- to medium-grade metasedimentary rocks of the Jardine metasedimentary sequence (JMS). The JMS underlies much of the northern part of Yellowstone National Park, and extends north of the Park to include the Jardine gold district. The JMS includes quartzites, iron formations, and schists with clear preservation of sedimentary structures (e.g., graded bedding, cross-beds, etc.) as described previously by Casella et al. (1982), and Thurston, (1986). Radiometric dating of the plutons, however, was not part of these studies. Consequently neither the age of the plutonism or the time of deposition of the JMS was constrained.

This contribution reports results from an NSF-supported Research Experience for Undergraduates project devoted to the Precambrian igneous and metamorphic rocks exposed in Yellowstone National Park and the adjacent terrain. We sampled eleven distinct plutonic bodies and determined their compositions (mineralogic as well as major and trace element) by x-ray fluorescence and inductively coupled plasma (ICP) techniques and their ages by U-Pb dating of single zircons using laser ablation (LA), multi-collector (MC), and ICP-MS (mass spectrometry) methods as described in Mueller et al. (2008).

RESULTS AND DISCUSSION

As shown in table 1, ages for all plutonic bodies fall in a relatively narrow range from 2.79 to 2.83 Ga, which is identical to the range of ages determined for the various members of the Long Lake magmatic complex (LLMC, Mueller et al., 2010). The ages of

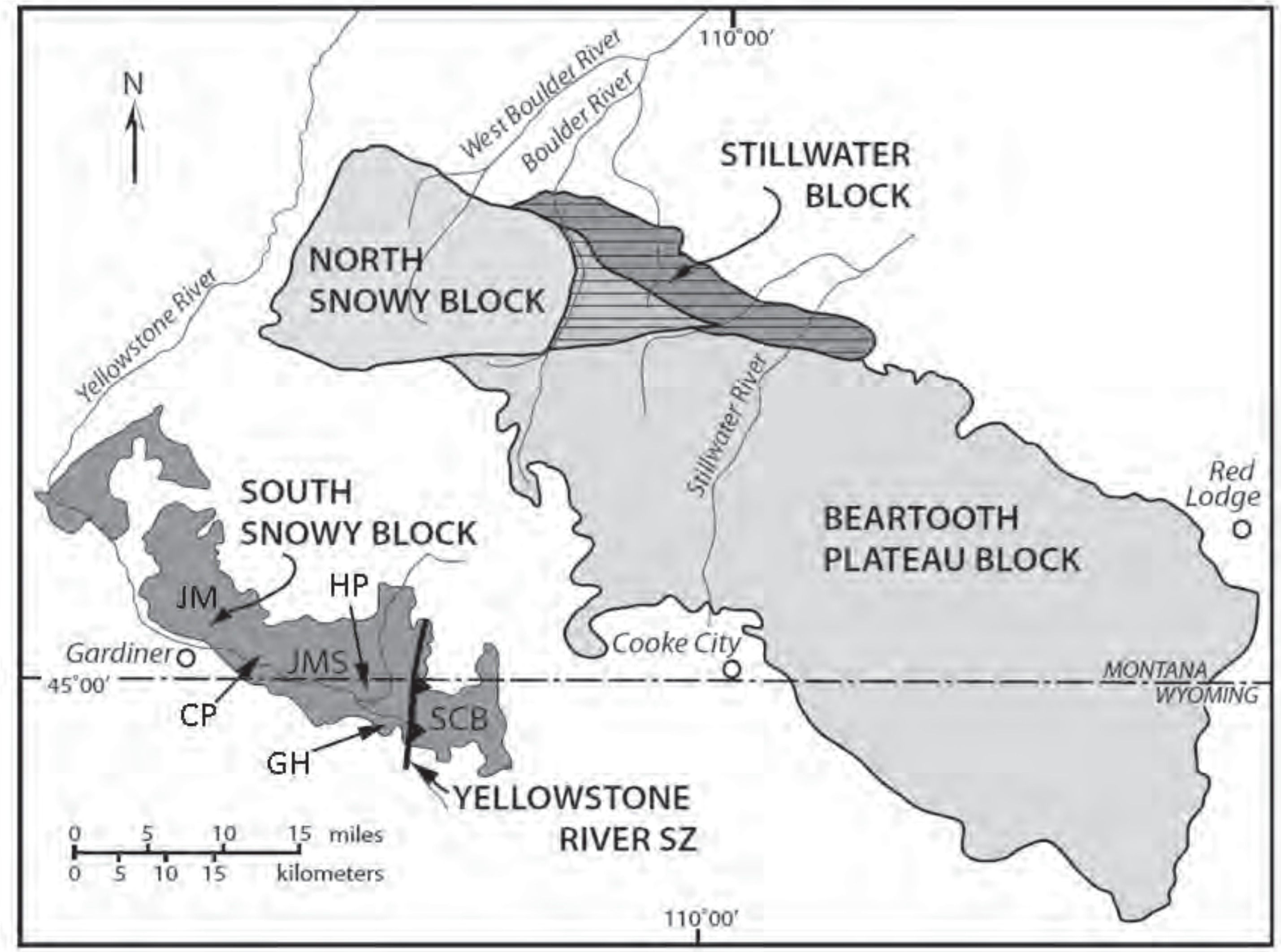

Figure 1. Schematic depiction of major Archean geologic units of the Beartooth Mountains. The JMS label and Yellowstone River shear zone are within Yellowstone National Park.

the youngest plutons (~2.80 Ga; fig. 2) and the youngest detrital zircons (~2.9 Ga) constrain the depositional age of the JMS (Mueller et al., 2014).

Although intruded over a limited time interval, the SSB plutons exhibit a range of compositions and intrusive forms. In the western SSB two 10 km-scale, peraluminous granitic plutons that are bulbous and locally muscovite-bearing (Crevice Mountain and Hellroaring Creek stocks) intruded the JMS (fig. 3). These granitic rocks are spatially associated with a suite of dioritic and granodioritic sheets that created injection migmatites in parts of the JMS. U-Pb zircon ages for these sheets (leucosomes) also range from 2.79 to 2.81 Ga, and have TTG compositions. A second, distinct suite of intrusive rocks is exposed in the eastern SSB and referred to as the Slough Creek batholith (SCB). These rocks range in composition from diorite to granite with TTG suite affinities and comprise a composite plutonic suite of sheet-like intrusions with compositional variations on the 10- to 100-m scale; magmatic epidote in hornblende-bearing plutons indicate crystallization at about eight kilobars (Berndt et al., 2012).

In terms of intrusive style, then, the pluton geometries change generally from west to east in consort with increasing metamorphic grade of the JMS. Plutons in the western part of the SSB are discrete, bulbous bodies that exhibit sharp contacts with their host JMS lithologies. As metamorphic grade and depth of burial increase to the east, the plutons become more sill-like and comprise larger assemblages (e.g., Slough Creek batholith). These easternmost magmas were emplaced at higher ambient pressures and temperatures and developed more migmatitic lithologies, which appear to represent both injection of new magma and some anatexis of the JMS. Pressure estimates from metamorphic parageneses in the JMS suggest a mini-

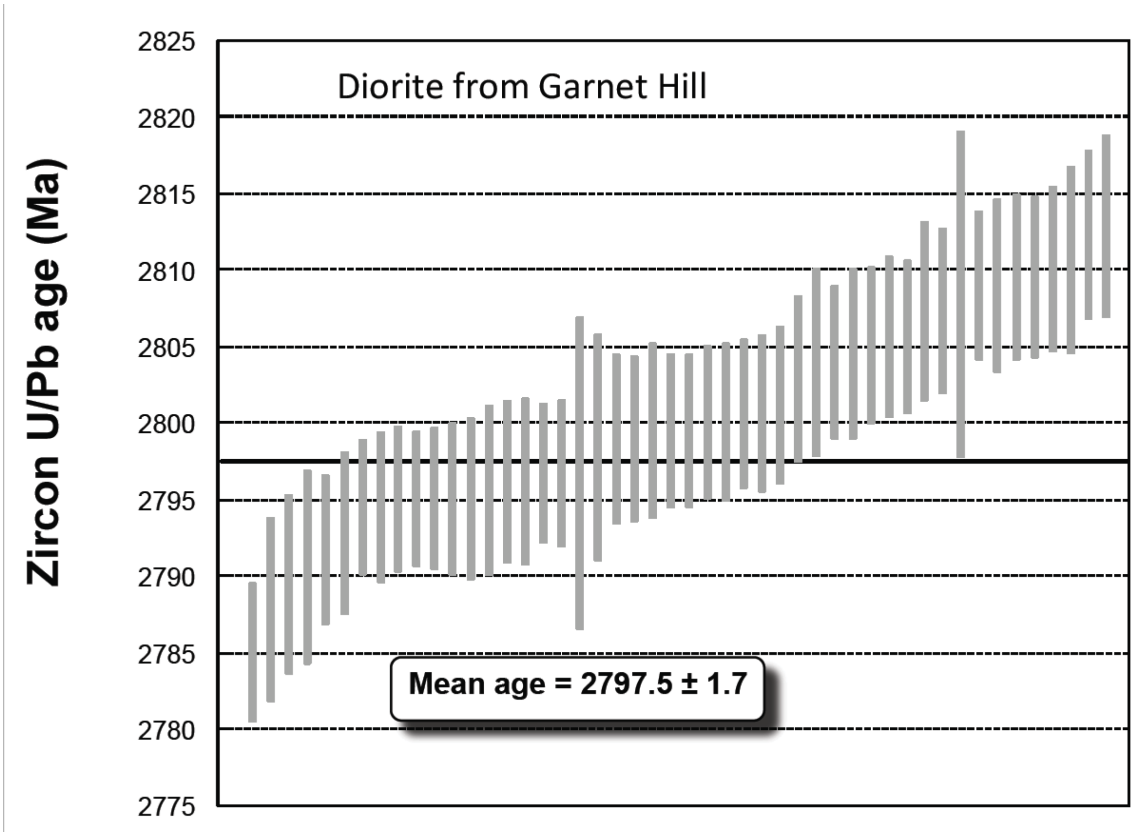

Figure 2. Plot of individual ages of zircons from the peraluminous diorite of Garnet Hill (error bars and age are given at the 2-sigma level based on 207Pb/206Pb measurements). Data acquired by laser ablation MC-ICP-MS.

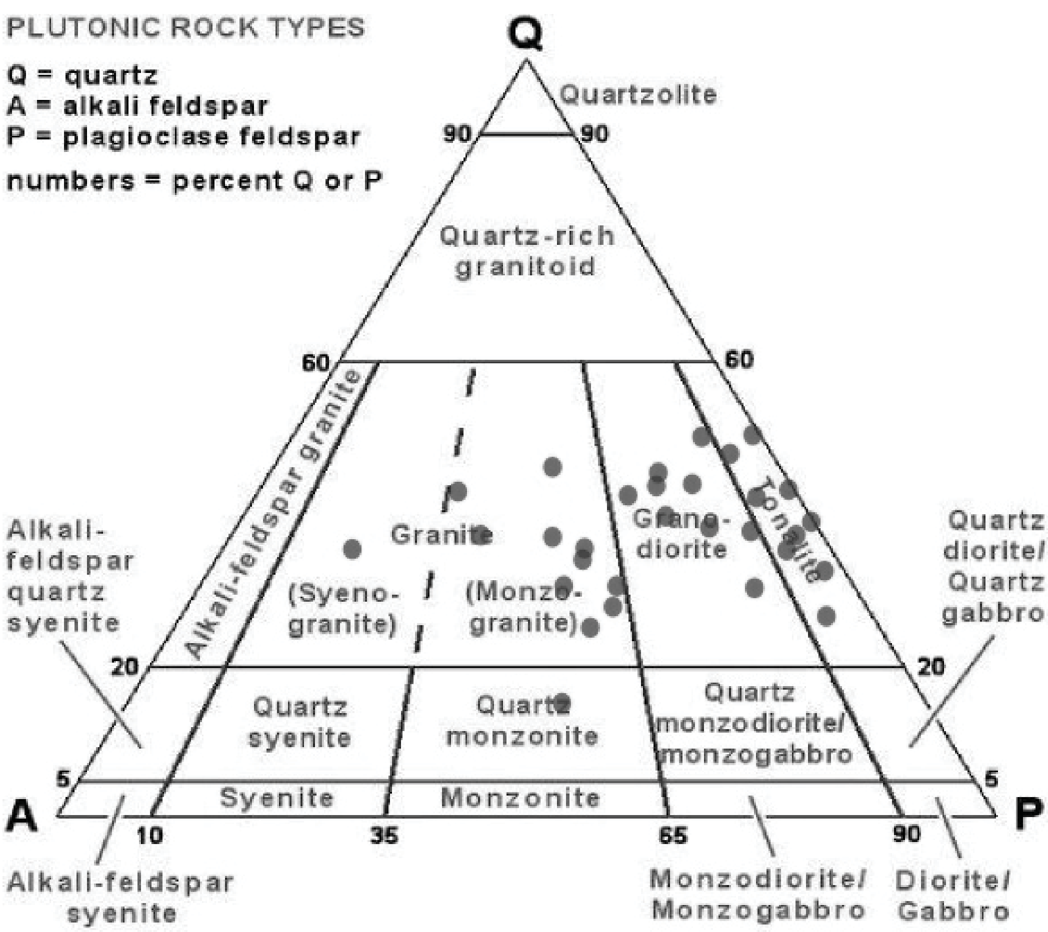

Figure 3. A normalized quartz (Q)-Plagioclase (P)-Potassium feldspar (A) diagram showing the range of mineralogic compositions represented in the SSB plutonic rocks.

mum depth of emplacement of 10 km in the west to 24 km in the easternmost plutons, which is supported by the presence of magmatic epidote in some components of the Slough Creek batholith. The metamorphic gradient is not continuous from west to east. A two-kilobar discontinuity in metamorphic grade occurs across the Yellowstone River shear zone, suggesting 5-6 km of displacement (fig. 1).

Regardless of pluton geometry, the SSB plutons all show evidence of a strong genetic relationship to magmas generated in modern subduction-related magmatic provinces, particularly young, hot subduction systems in which slab melting is thought to occur (e.g., Drummond and Defant, 1990). As such, the plutonic rocks of the SSB resemble the dioritic to granitic rocks of the LLMC (Mueller et al., 2010; fig. 4) because many plutons have similar, depleted HREE contents (average Yb(n) <10; fig. 5) indicative of residual garnet in their source(s). Overall REE patterns, however, are not strongly fractionated with La/Yb(n) from 4 to 150; only two have ratios over 100. The rocks also exhibit the relative depletion in high field strength elements (HFSE) characteristic of modern arc magmas (fig. 6). As a consequence, the rocks typically plot in the volcanic arc field (Rb vs. Y+Nb) and volcanic arc plus syn-collisional (Nb vs. Y) fields of the granitic tectonic discrimination diagrams (fig. 7) of Pearce et al., (1984). The western SSB stocks (west of the Yellowstone River shear zone) and the Slough Creek batholith in the eastern SSB are, therefore, interpreted to be upper and middle crustal equivalents of the LLMC intruded at about 12 km and 20 km depth, respectively. The similarities in elemental composition to the LLMC rocks also extend to the Pb and Nd isotopic systems (Mueller et al., 2014).

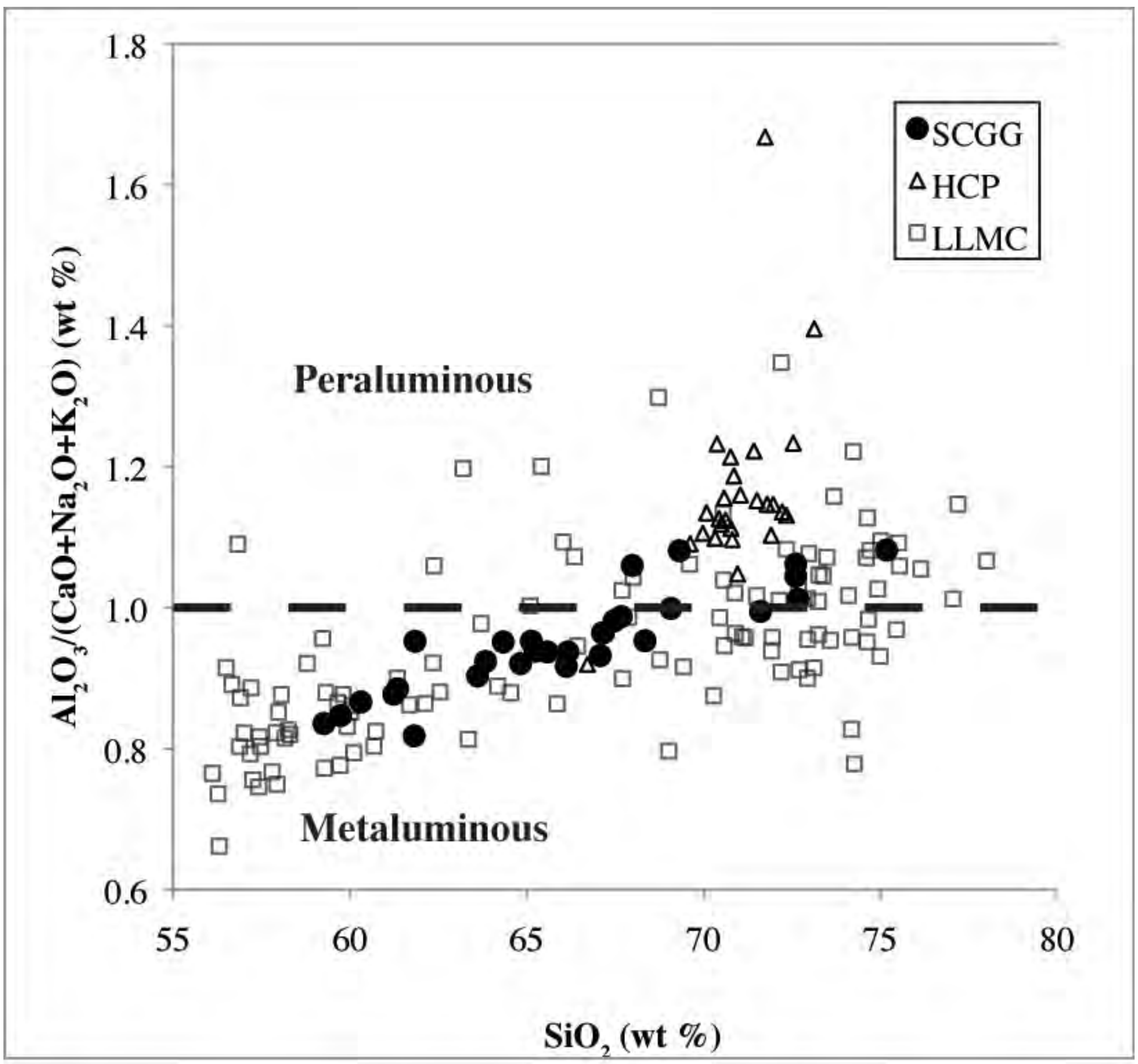

Figure 4. A comparison of the extent of overlap in aluminosity of the SSB (solid circles) and the LLMC (squares) plutonic suites. The peraluminous Hellroaring Creek stock is also plotted (triangles).

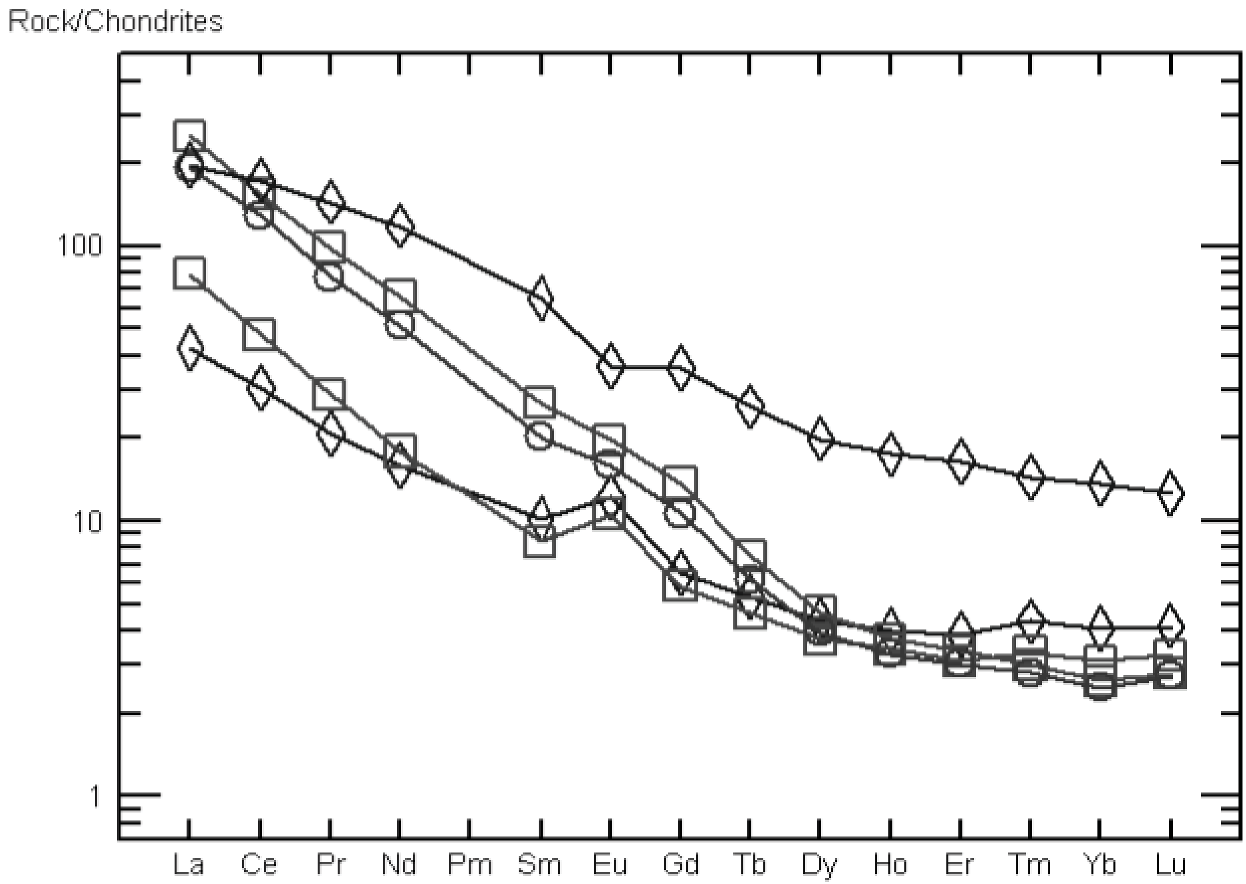

Figure 5. Chondrite-normalized REE (rare earth) diagram showing the typical depletion (<10x) in HREE (heavy rare earth) contents (McDonough and Sun, 1995).

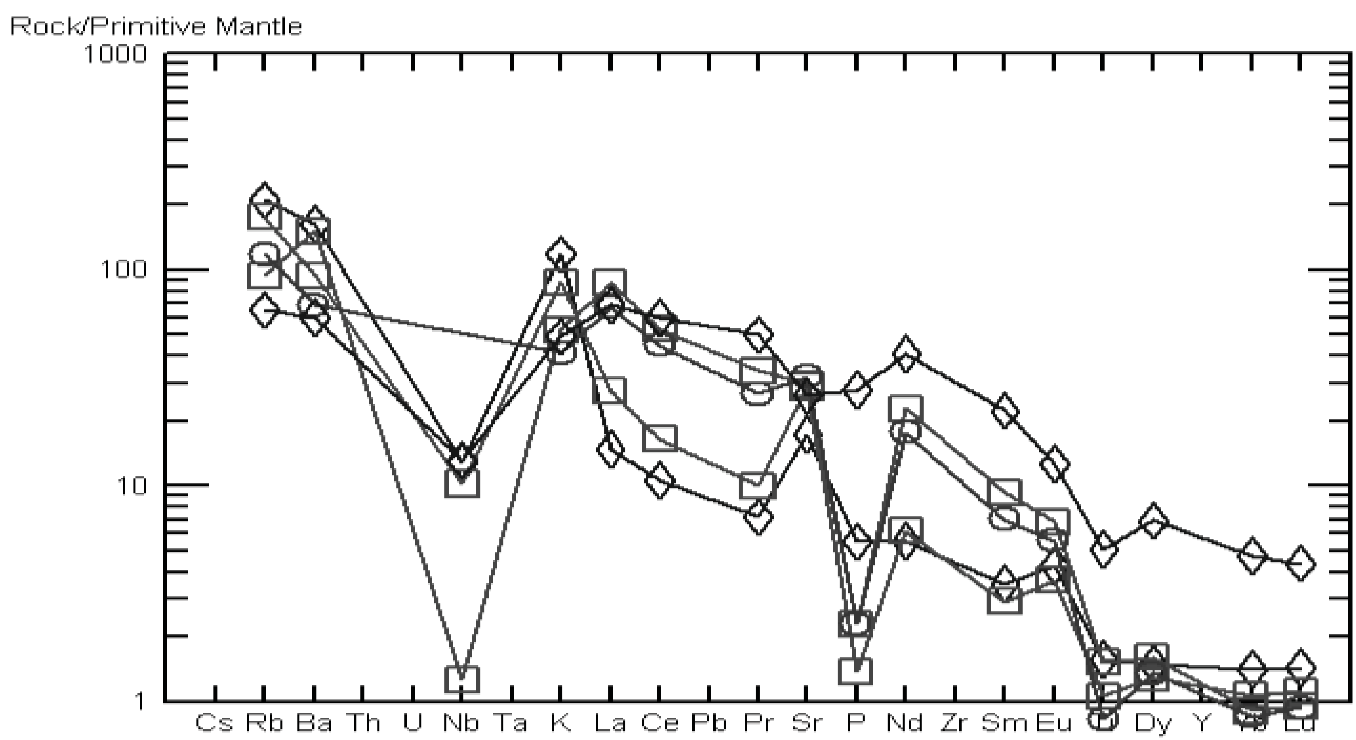

Figure 6. Primitive mantle normalized trace element contents (McDonough and Sun, 1995) of SSB plutonic rocks showing relative depletions of HFSEs characteristic of modern arc magmatism.

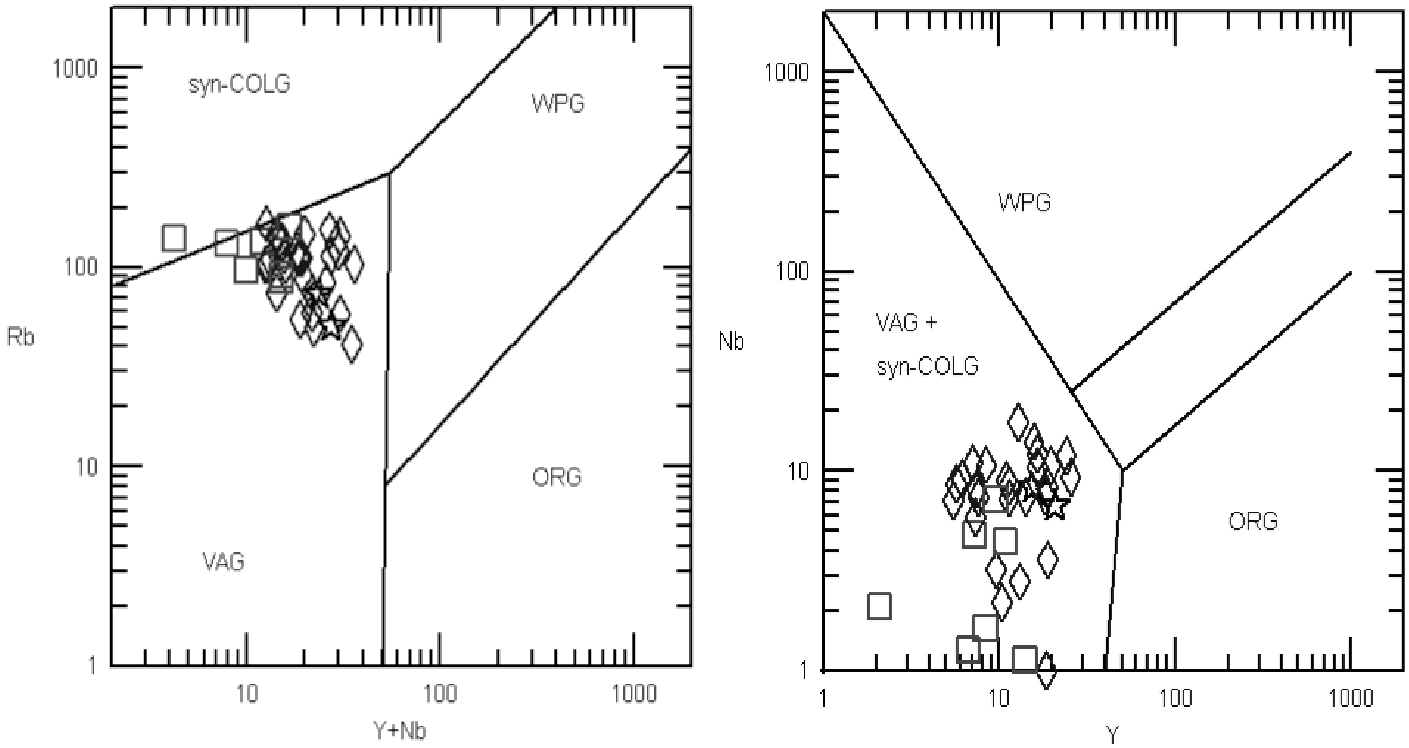

Figure 7. Trace element discrimination diagram for granitic rocks from Pearce et al., (1984).

CONCLUSIONS

The strong age and geochemical similarities between the epizonal plutons of the SSB and the mesozonal TTG rocks that dominate the BBMZ as a whole suggest that the SSB magmas formed in the same overall tectonic environment. Consequently, the combination of SSB and LLMC plutonic rocks provide a cross section of a Mesoarchean arc plutonic complex over a distance of over100 km and a paleo-depth range of 10 to 25 km. This cross section is unique within the Wyoming Province and one of the oldest arc sections preserved on earth.

ACKNOWLEDGMENTS

Gratitude is expressed to the NSF for financial support of a two-year Research Experience for Undergraduates site project devoted to the Precambrian rocks of Yellowstone National Park (individual grants to Montana State University, University of Florida, and Louisiana State University). In addition to the authors, many students engaged in the project provided useful data and geological context for this study. Personnel from the Science Office of Yellowstone National Park were also most helpful in arranging sampling and logistical permissions.

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ROAD LOGS

ITINERARY FROM SEATTLE TO REPUBLIC, WASHINGTON ROAD LOG AND SELECTED GEOLOGY

John Whitmer, Leader

REGIONAL GEOLOGY AND THE LANDSCAPE

The landscape of this region bears the imprint of no fewer than six advances of the Puget Lobe of the Cordilleran Glacier (Troost, K, recent mapping), with the greatest effect from the most recent (Fraser) advance (Owen, Caffee, et. al, 2006). The history of this event is so peculiar that it has taken many decades to achieve the current understanding, and the task is not yet fully completed.

The Last Glacial Maximum (LGM) is dated about 20,000 years ago, in the Pleistocene. Until then, the mean annual temperature was so low that the ice lacked ductility and was frozen to its bed. In consequence, the ice sheet grew by thickening and there was precious little horizontal movement. At the LGM, the ice was about 6,000 feet thick over Vancouver, B.C., but there were very few places where it advanced south of the U.S. Border. By 18,000 years ago, they had retreated greatly. The Puget Lobe reached its maximum advance about 3000 years later (Troost, K., Booth, D., et al., 2009).

Only in 1990 was it recognized that the transition from Arctic to Temperate conditions was incredibly rapid, perhaps occurring in less than a decade. Within a few years, climate change had increased the average temperature to warmer than it is now. Rapid melting of the ice ensued. Nearly all the meltwater disappeared into crevasses or glacier mills (glaciermuehl), creating a large body of subglacial water which lubricated the ice-rock interface. The ice surged rapidly into the adjacent lowlands. The minimum estimated sliding velocity was 1,500 feet per year. Unlithified sediments beneath the glacier were little affected by shearing. The rate of ice advance in the Puget Lowland is estimated to have been about 440 feet per year (Troost, K., Booth, D., et al., 2009).

Immense torrents of water flowed from north to south beneath the ice through the Black Lake Spillway and smaller channels to the east into the Chehalis Valley (Troost, K., Booth, D., et al. (2009). The effect on the regional landscape was to leave interfluves of long, drumlinoid ridges alternating with valleys or troughs, the deepest of which are Puget Sound, Lake Washington, Lake Sammamish, and the Snoqualmie Valley. Since the regional slope is from south to north, the preglacial streams flowed in that direction. The advancing Puget Lobe ice sheet forced reversal of flow from north to south via the Tacoma Narrows, the Black Lake Spillway, Hobart Canyon (to May Valley and the Cedar River) and the Snoqualmie Valley (to the Cedar River). See figure 1.

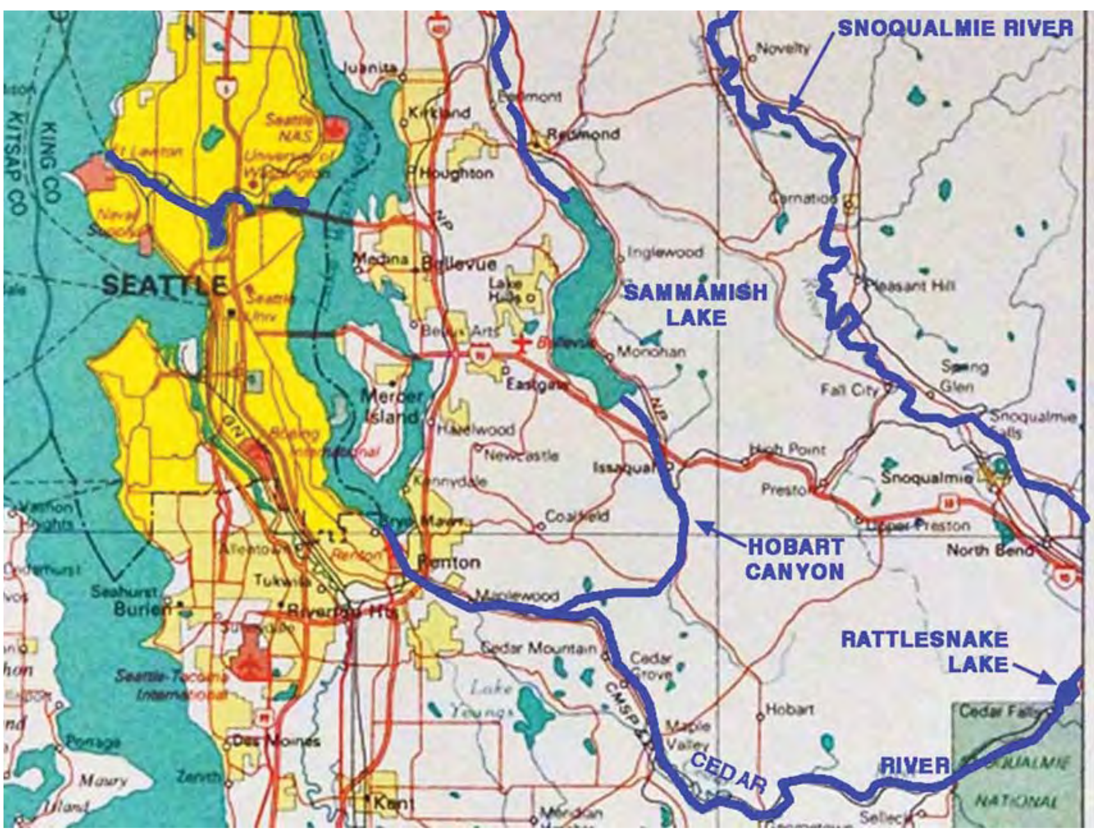

Figure 1. The Snoqualmie River Drainage.

The maps in this field guide utilize images from the USGS Atlas of the United States (1971). They are entirely schematic, with the goal of illustrating the general configuration of the landscape.

Figure 2 shows the Seattle Fault and the anticline in its upper plate. The Seattle Fault is a blind thrust fault that dips northward. It constitutes a tectonic wedge which separates (delaminates) the supracrustal rocks of the region from the crustal rock, i.e., the Crescent Basalt (Potter, C.J., 1994). Splays of the Seattle Fault reach the surface at a number of places, notably the Toejam Hill Fault on Bainbridge Island and the Vasa Park Fault west of Lake Sammamish. Current thinking is that such faults are "fault bend" faults, the nature of which is illustrated in figure 3.

The anticline in the upper plate of the Seattle Fault dominates the view from I-90 near Issaquah. It is divided into three sections by Hobart Canyon south of Issaquah and by Tibbets Creek Canyon immediately west and south of Issaquah. From west to east, these sections are named Cougar Mountain, Squak Mountain, and Tiger Mountain (Sites B, C, and D on fig. 5). Local lore explains that if you were between a cougar and a tiger, you would squawk too. Tiger Mountain is the largest of the three. Its summit is about 3,400 feet above sea level. The region is called the Issaquah Alps owing to this high relief.

Squak and Tiger Mountains are the deformed remnants of a late Eocene Volcano that spawned numerous andesitic debris flows, leaving a footprint that extended to the Snoqualmie Valley and Rattlesnake Lake. The Cascade Volcanic Arc had not been erupted at this time. Instead, a host of shield volcanoes was scattered throughout Cascadia (southern British Columbia, Washington, Oregon, and northern California). Each volcano had a large footprint, separated by lowlands through which streams from the Okanogan Mountains and the Idaho Batholith flowed to the ocean. Alternat-

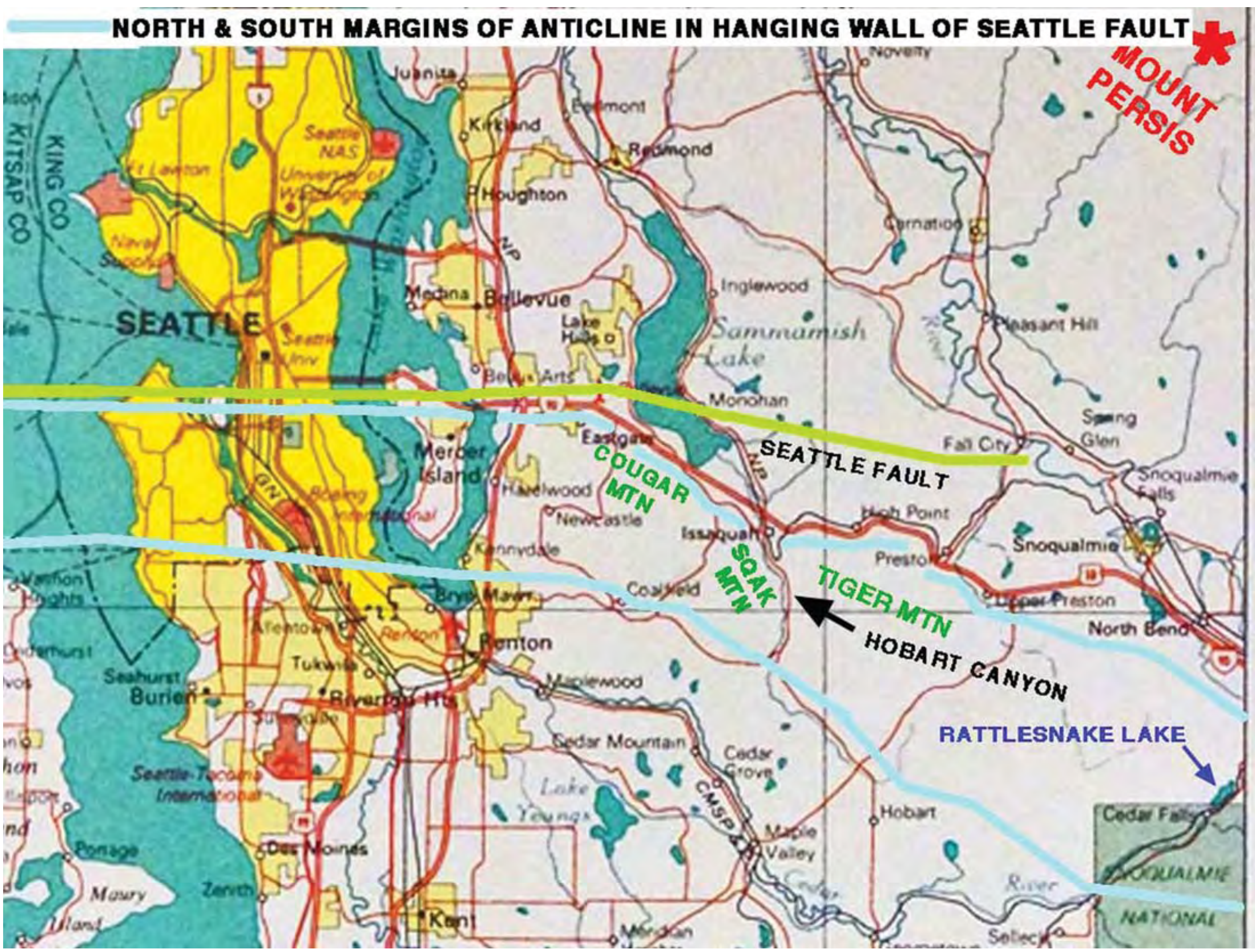

Figure 2. Northern and southern margins of anticline in the hanging wall of the Seattle thrust fault.

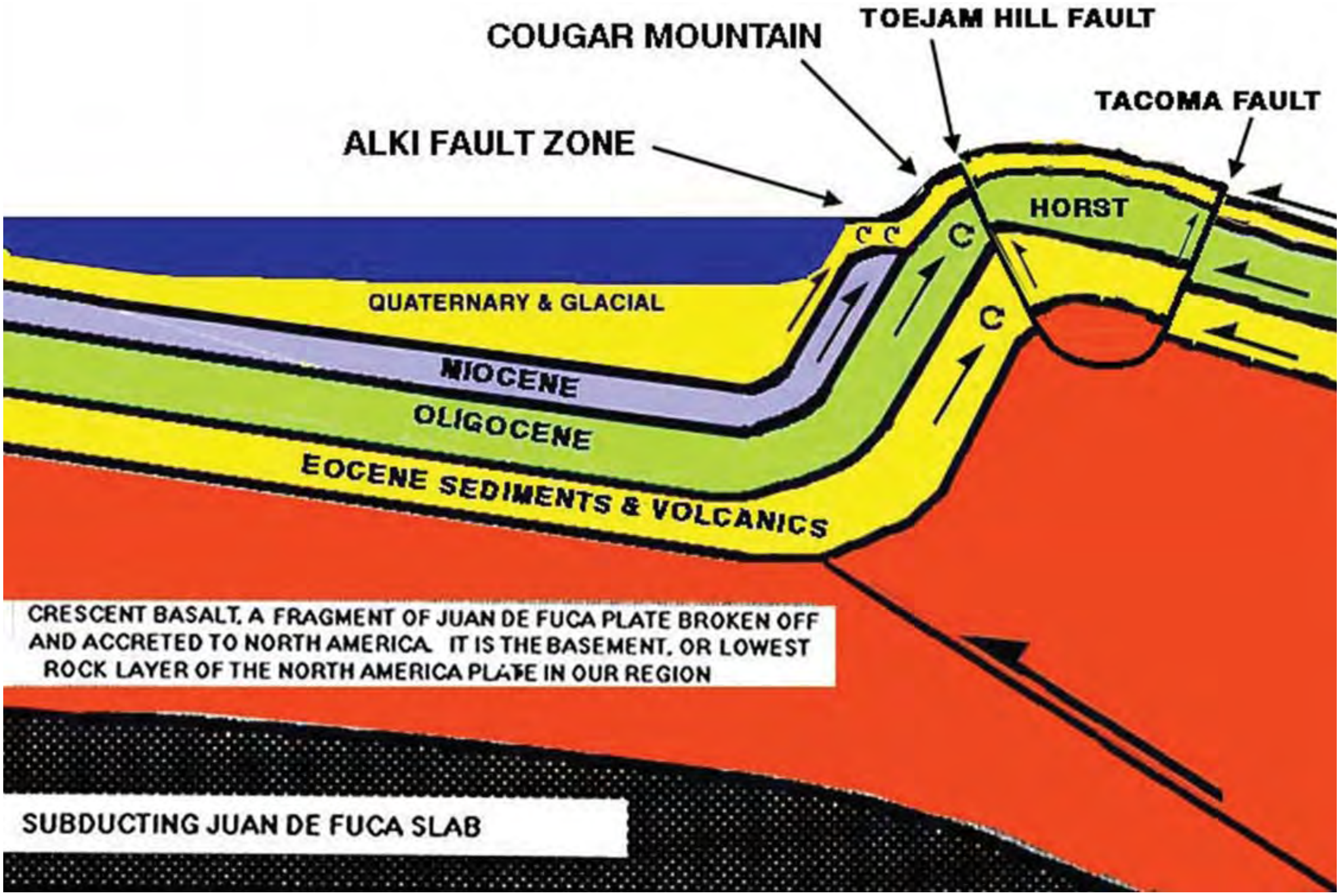

Figure 3. Movement on the Seattle fault produced north-directed compression of the Cougar Mountain Anticline. Opposite-facing faults ensued, along which a wedge (horst) moved upward. This movement occurred 1,100 years ago.

ing layers of silt, sand, and gravel were deposited in these valleys. Sand and gravel were deposited in the stream channel; and finer-grained materials were deposited on the flood plain during infrequent episodes of high run-off. Dense forests grew on the floodplain only to be buried in subsequent flood events. These sediments are lumped into the Puget Group, in which coal and peat deposits abound. In the Seattle-North Bend area, the Eocene volcanoes are Tiger Mountain and Mount Persis (marked by a red asterisk in the northwest corner of fig. 3). Their footprints meet in the Snoqualmie Valley.

As is usually the case with large shield volcanoes, Tiger Mountain had an outer shell of resistant lava flows and indurated volcanic breccia. Its core, in contrast, was shattered by frequent explosions of water-rich, partially melted, crystalline mush. Over long periods, streams penetrated the shell, gaining access to the shattered core which yielded rapidly to stream erosion. This process is augmented by the radial fractures of the volcano.

When the subglacial torrent of the advancing Puget Lobe flowed south in Hobart Canyon, it cut right through the Eocene Volcano, leaving a remnant of the volcano on the eastern face of Squak Mountain and a larger fragment on Tiger Mountain. Hobart Canyon has high walls that are nearly vertical, and a flat flood-plain that is subject to frequent flooding. This flooding has caused severe property damage. Public policy in recent decades has been directed toward restricted use of the flood plain and purchasing private land from landowners who had experienced frequent losses. North of I-90 from Seattle eastward is a series of plateaus with poorly-drained surfaces of low relief on which there are kettles, depressions, shallow channels and ravines, some with perennial streams and others with intermittent streams. The landscape is characteristically glaciated, covered with till (locally called “hardpan”) and abundant glacial erratics of all sizes. The till averages 45 feet thick in the Seattle area, and it is of similar thickness throughout the Puget Lowland (Brown, N.E., Hallet, B., and Booth, D.B., 1987).

Where not altered by development, the landscape is green with vegetation; outcrops of bedrock are rare. Few geologists are willing to work in this environment. Hoover Mackin famously said that there is nothing about the green hell of Washington that would not be improved by a good forest fire. The people who map this area frequently use machetes to gain access to outcrops.

These plateaus lie at elevations between 400 and 450 feet above sea level, marking the level to which the entire Puget Lowland was filled by outwash from torrents emanating from the Puget Lobe as it advanced southward. In Seattle, Queen Anne Hill and the Magnolia District exemplify this surface. Other examples are the plateau west of Puget Sound, the region between Eastgate and Lake Sammamish, and the Sammamish Plateau between Lake Sammamish and the Snoqualmie Valley.

As the ice sheet advanced to the south, the major streams, which had previously flowed to the north, were forced to reverse their direction and flow to the south. Their valleys became proglacial lakes. The advancing ice floated on these lakes. The water in Lake Sammamish created a sub-glacial tunnel that drained through Hobart Canyon, which became an ice tunnel. Such tunnels precisely regulate the pressure in the subglacial lakes. Hydrostatic pressure supports the ice. As the depth of the water increased, and therefore the hydrostatic pressure, the enclosing ice in the tunnels melted, responding to the increased pressure. The hydrostatic head beneath the glacier averaged 90% of the thickness of the ice overburden (Brown, N. E., Hallet, B., and Booth, D.B., 1987). The flow of water increased, enlarging of the tunnel. Conversely, when the volume of meltwater declines, so does the pressure, and the tunnel walls respond by increasing the ice thickness and decreasing tunnel size. This balance maintains sufficient pressure to support the overlying ice.

It is estimated that at its maximum advance, the ice was 3,000 feet thick above Seattle. This implies immense hydrostatic pressure, enabling water in the tunnel to act as a powerful jet – not unlike a pressure washer. By this means, very large boulders were plucked out of the bedrock.

In the 1970's I did extensive exploration of the Lake Missoula Floods region using the work of J Harlan Bretz (1959 and 1966) as my guide. The channeled scabland resembled classic glacial landforms but the sculpting had been done entirely by water, for the elevation was too low and no glacier ever reached there. I concluded that the water under the ice did the work of erosion to produce deep U-shaped valleys. The effect of erosion by sliding ice resembles sandpapering, rounding edges and corners and making striations. The volume of rock removed by this means is tiny compared to the immense volume removed by water (Iverson, N., 2012).

Between the glaciated plateaus, the landscape in this region consists of long valleys: swales alternating with long ridges that resemble drumlins in their shape and their sediment, composed of water-transported sand and gravel. They strike north-south. The interridge valleys and swales are *tunnel valleys* (Wright, H., 1973; Gerath, R. F., Fowler, B. K., et al. 1985). The swales mark the location of subglacial tunnels in which water flowed with sufficient velocity to sweep the sediment out from beneath the glacier. The ridges were shaped by water with only enough velocity to imprint a streamlined (drumlinoid) shape. Some of these ridges are tens of miles long. There are many such undulations between Seattle and Lake Sammamish (see fig. 4). Lake Union, the Ship Canal, and the Montlake Cut are believed to occupy a valley eroded by an ice-marginal stream during recession of the ice sheet.

This paragraph pertains to points of interest located at lettered sites on the following map, figure 5. Site A, Mercer Island, is one of the smaller drumlinoid ridges. Both Mercer Island and Bellevue were virtually undeveloped until the opening of the first floating bridge on July 2, 1940. On November 25, 1990, during renovation of the bridge, a severe storm struck unexpectedly. Rainwater entered one of the pontoons through hatches that were unwisely left open. The flooded pontoon sank and dragged an additional seven pontoons with it. Fortunately, a second, larger span had been opened on June 4, 1989. Rebuilding of the sunken part of the original bridge was ultimately completed and the structure remains as part of the floating bridge complex today (Dorpat, P., and McCoy, G., 1998).

Subsequent rapid, dense development, aided by the

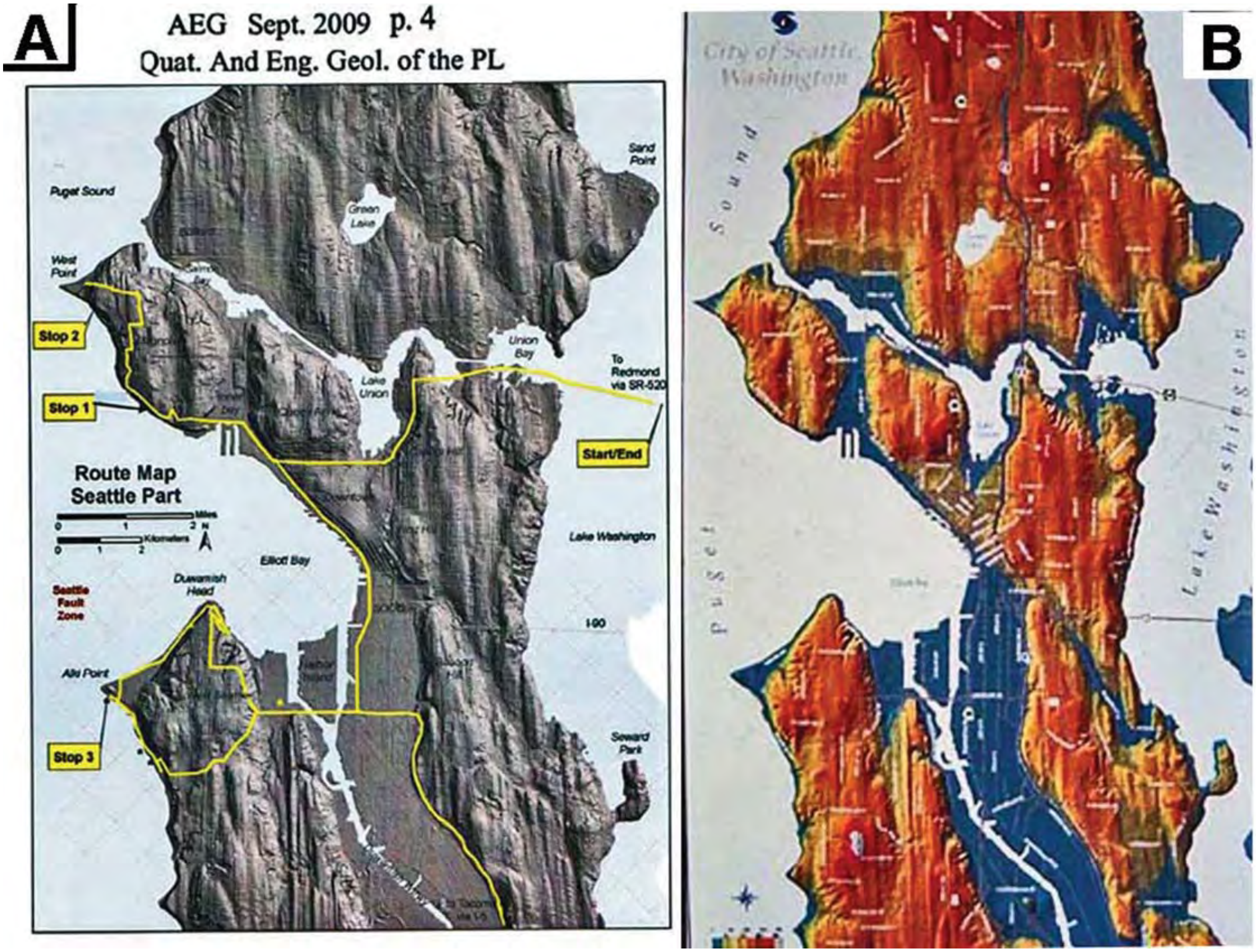

Figure 4. The dark areas in Figure 4B denote flood plains of recent alluvium, including the Duwamish Valley, tidewater marshland covered with landfill; Harbor Island; the Pioneer Square district and the extensive area south of Pioneer Square, known as SODO for "south of the Dome," now marked at its north end by the two large stadiums. Alki Point and the narrow fringe along the coastline denote uplifted wave-cut platforms, mostly turbidites. The Duwamish Valley, Harbor Island, Pioneer Square, and SODO are highly vulnerable to earthquake damage owing to amplification of seismic shaking, liquefaction, and lateral spreading. During the Earthquake of 1949, several houses in the Duwamish Valley were jacked off their foundations by sand volcanoes which ruptured the basement floors. In contrast, the drumlinoid ridges (with the exception of West Seattle) and the plateaus had been compacted by the over-riding ice sheet, rendering them nearly as damage-resistant as bedrock.

bridge, totally changed the environment. Today, Bellevue rivals Seattle as an urban center.

The I-90 Tunnel west of the floating bridge was completed in 1989. With a diameter of 63 feet, it is the world's largest tunnel through soft earth. A ring of 24 smaller, interlocking tunnels, each of which is filled with concrete, surrounds the central bore. It was constructed through the heavy blue clay of Mount Baker Ridge, one of the region's many drumlinoid ridges. Since the Ridge is in the Seattle Basin, the footwall of the Seattle Fault, it is 1,700 feet above Miocene turbidite bedrock. At downtown Seattle, bedrock is 2,700 feet below the surface. From downtown the basin floor rises gently northward. Bedrock reaches the surface at the northern end of Whidbey Island, a short distance south of Deception Pass.

At Eastgate, where I-90 and I-405 intersect, the Mormon Temple is visible to the north. Beyond it is the large, glaciated plateau of eastern Bellevue. Cougar Mountain, Site B, is south of I-90, the top of which has extensive coal and clay deposits which supported heavy industry for some time. Newcastle, on Cougar Mountain, had such a large coal mine that its population exceeded that of Seattle in the late nineteenth century. Much of the coal was used by the railroads until the 1920's when the locomotives west of the Rocky Mountains were converted to oil burners. The

Figure 5. Noteworthy sites between Seattle and Snoqualmie Falls. A, Mercer Island; B, Cougar Mountain; C, Squak Mountain; D, Tiger Mountain; E, Recently discovered splay of Seattle Fault; F, Lake Sammamish Pleistocene Delta; G, Issaquah Highlands; H, Snoqualmie Falls.

mines are abandoned now, but mine tunnels, collapse pits, and remnants of mine structures remain. There is one coal seam on the top of Cougar Mountain that is still burning. Clay at the site was mined to make brick until the 1990's.

At Site E, a splay of the Seattle Fault reaches the surface at Vasa Park. I believe that it continues eastward under Lake Sammamish and deforms the landscape north of the Sammamish Delta, Site F. That delta (F) was an immense gravel fan, emplaced there beneath the ice sheet when the subglacial lake was more than 400 feet deep. It was fed by a subglacial stream that crossed the Sammamish Plateau from the Snoqualmie Valley. All of the late Pleistocene, Fraser gravel has been mined, leaving a cirque-like landform now occupied by a large shopping center. The gravel from earlier glacial events is never mined because it is too weathered to be an acceptable building material. On the plateau just northeast of the headwall is an immense glacial erratic boulder.

Site G denotes the Issaquah Highlands, an area that is being densely developed. The developers avoided the northern part of the Highlands because it is riddled with abandoned mine tunnels. That area, together with Squak Mountain, Site C, supported Issaquah's thriving coal mines until the market collapsed in the 1920's. The mine operators solved their mine waste problem by dumping the tailings into adjacent branches of Issaquah Creek. This accelerated the building of the large delta where Issaquah Creek enters Lake Sammamish in Lake Sammamish State Park. Dewatering of the Delta during the Earthquake of 2001 produced many fissures and sand volcanoes. Coal slack (a mixture of coal fragments, coal dust, and dirt that remains after screening coal) was visible a few inches down in the fissures, and the sand volcanoes were black with coal slack.

Dense residential development hides the mine sites of Squak Mountain. Collapse of tunnel roofs has damaged much of the property in the area. During one period of heavy precipitation the groundwater pore pressure became so high that a torrent escaped from an abandoned mine ventilator shaft causing a debris flow which damaged a major street.

All of the slopes in this region are prone to landslides. The hazard increases with the steepness of the slope. Soil creep is widespread, producing drunken forests and pistol butt trees. The Earthquake of 1949 triggered sufficient soil creep on both sides of the Duwamish River that the piers of the draw bridges converged, jamming the draw spans so that they could not be opened.

STOP 1 **SITE 1.** Snoqualmie Falls (fig. 6), known for its scenic beauty, is held up by an intracanyon flow of andesite breccia from an Eocene eruption of Mount Persis. It is an example of inverted topography. During the eruption, andesite flows followed the topography and flowed down existing drainage ways. During subsequent uplift, softer rock around the flows eroded, then the landscape was buried by Pleistocene outwash sediment. In re-establishing its course after deglaciation, the River had to cross the resistant andesite breccia.

STOP 2 **SITE 2.** Mount Si. Exit 27. From here eastward, Mount Si dominates the landscape north of I-90. A member of the Western Melange Belt, Mount Si is an accreted seamount composed of Mesozoic metavolcanic rock. Its western face was remarkably oversteepened by the torrent of water beneath the floating margin of the Puget Lobe. A smaller, rounded peak, Little Si, abuts it on the west. Little Si is the metasedimentary member of the Mesozoic seamount. I suspect that Mount Si is a major landslide hazard to the city of North Bend, for it will

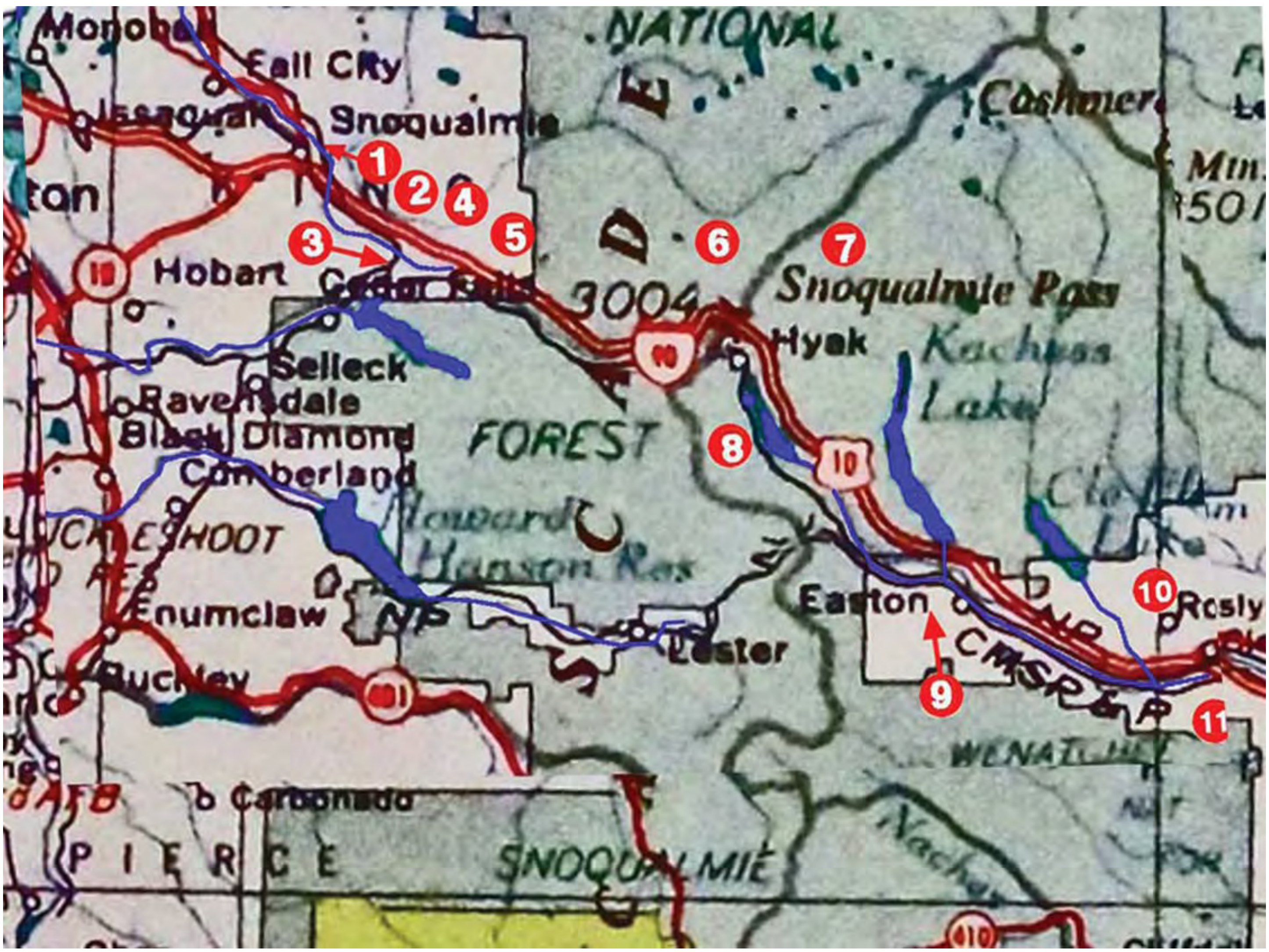

Figure 6, Sites 1-11. Noteworthy locations between Snoqualmie Falls and the Cle Elum coal field.

swing widely forth and back in the seismic wave of a BIG earthquake.

STOP 3

SITE 3. Rattlesnake Lake. Exit 32, road to Cedar Falls. Rattlesnake Lake occupies a closed depression at the pass between the Snoqualmie and Cedar Rivers. The depression was eroded by turbulent water when the Snoqualmie River was forced to flow southward into the Cedar River during the advance of the Puget Lobe.

Settlers built a small village on this prairie. About a century ago, the City of Seattle built a dam for hydroelectricity and to supply water to the growing city. Unfortunately, the reservoir was lined with porous gravel. The designers assumed that the basin was lined with glacial till - “A glacier must flow downhill!” The till was actually on the downstream side because the ice had advanced up the valley from the Puget Lobe (Ed. Again, showing the importance of a good geologic investigation). The operators were never able to fill the reservoir to capacity. On their first attempt, springs discharged water at an alarming rate until a blowout created a new canyon (Boxley Canyon). This disastrous flood destroyed three sawmills, and severed the main Milwaukee Railroad Line, and several logging railroads. Despite remedial efforts and drastic limitations of water storage, the local water table rose, creating Rattlesnake Lake. At low water stage, foundations of the abandoned village are visible above water level. See Mackin, J., 1941.

STOP 4

SITE 4. Edgwood. Exit 34, site of the Boxley Canyon Flood disaster.

STOP 5

SITE 5. Milepost 35. Downstream, western edge of the Puget Lobe glacial embankment. The ice sheet dammed the Snoqualmie River here, creating a proglacial lake. The sub-glacial stream that oversteepened Mount Si flowed into this lake, forming an immense gravel delta.

In time, the Snoqualmie River filled the lake with alluvium. Here, I-90 begins its ascent through the embankment in a canyon cut by the post-glacial river. There are gravels of the Pleistocene delta and stream deposits of the embankment on the canyon slopes. Bedrock above the sediment is mostly Oligocene and Miocene granodiorite. The Miocene rock is part of the Snoqualmie Batholith, most of which is north of I-90. Owing to uplift of that buoyant pluton, the peaks and ridge crests on each side tower 4,000 feet or more above I-90. Climbing any of these peaks, including Mount Si, involves the risk of disaster on their steep slopes. South of I-90, the rocks dip steeply southward, away from the Batholith. The dip slope of that structure is prominent from I-90 west of North Bend.

Milepost 37. Large gravel quarry in Pleistocene Delta to left.

Milepost 39. Midway up the south slope is the abandoned grade of the Milwaukee Railroad. The railway was electrified for most of its traverse of the Rocky Mountains and the Cascade Range. Completed in about 1914, it was the most recent and the best engineered railroad in Washington, but it did not survive bankruptcy in the early 1980’s.

Milepost 40. Highway is at level of proglacial lake fill. The road grade is nearly flat as it traverses the fill.

Milepost 42. I-90 passes three hanging valleys crossed by high railroad bridges. At Hall Creek, a segment of the bridge collapsed when a debris torrent undermined a support tower. That railroad grade is now a trail. Thanks to the Rails-to-Trails program, the missing span has been replaced by a footbridge. The trail is a good place to hike and snowshoe in any season, but with each increment of elevation gain toward the east, the wind becomes stronger and colder. The Cascade glaciers descending this canyon reached their maximum advance some 20,000 years ago. By the time the Puget Lobe reached here, the Cascade ice had melted away.

Exit 47. From here to Snoqualmie Pass, eastbound I-90 is on the southeast side of the canyon and the westbound lanes are on the northwest side. Denny Creek, a large tributary canyon, enters from the north. It is the site of Alpental, a major ski resort. This exit also provides access to the site of our nation’s worst railroad disaster (Flynn, L. J., 2007; Wandell, B., 1999).

STOP 6

SITE 6. Mileposts 50-54. I-90 bends around the nose of a major anticline, following a strike valley to cross Snoqualmie Pass. The area has suffered significant urban sprawl owing to the ski resorts here.

STOP 7 **SITE 7.** Gold Creek Valley is to the north. In the 1980's several families used the canyon for winter camping. The activity ended when several of the families were wiped out by a snow avalanche.

I-90 follows the Yakima River Valley to Cle Elum. "Lake" Keechelus is the reservoir for an irrigation district near Yakima.

STOP 8 **SITE 8.** Mileposts 57-60. There is a serious rockslide problem here. Although it looks like granite and appears to be solid, it is tuff that has been extensively jointed in a way that facilitates rockslides. An obsolete shed spans the eastbound lanes of I-90. Extensive remedial construction has been underway for years.

Exit 62. Stampede Pass-Lake Kachess exit. Ridges and peaks around here have rounded crests; there have been no matterhorns since the Pass. This indicates to me that none of the peaks in this area projected above the ice in major glacier advances. The road from here to the south goes to Stampede Pass and the Green River watershed. Most of that basin is closed to the public, reserved as the water supply for Tacoma. The Northern Pacific Railroad ascended from Pasco and crossed Stampede Pass in a tunnel. A few miles downstream there was a rail facility at the Village of Lester. Upon closure of the watershed to the public, the residents had no access to their property except by rail. One by one, over many years, the people reluctantly moved away.

Mileposts 67-68. Multiple basaltic dikes in road cut in median (left) side of I-90. These are thought to be in strands of the Straight Creek Fault (Eric Cheney). Lake Kachess and Lake Easton are in the trace of that fault.

STOP 9 **SITE 9.** "Lake Easton" another reservoir for irrigation, marks the trace of the Straight Creek Fault.

STOP 10 **SITE 10.** Milepost 81, shown in both figures 6 and 7. Cross the Cle Elum River. Gravel here is outwash from a large glacier in the Cle Elum River Valley. The Roslyn Coal Field is north of here.

STOP 11 **SITE 11.** Exit 84, shown in both figures 6 and 7. Exit here to Cle Elum. Stop at SAFEWAY store for rest room. Cle Elum and Roslyn were major coal mining areas. Coal was mined primarily to fire steam locomotives of the Northern Pacific Railroad. Mining declined rapidly after the railroad switched to fuel oil and is practically nonexistent now, although there is still coal here. There is a huge residential development at Roslyn. One developer accidentally started a fire in a coal seam a few years ago. It burned for a year or more and caused respiratory distress in residents of Roslyn. It may be still burning, as these underground fires are difficult to put out.

STOP 12 **SITE 12 (fig. 7).** Lookout Mountain, prominently in view to the east, is a cuesta of uplifted Columbia River Basalt. The basalt makes cliffs on mountains to the south of Cle Elum as well. The basalt attains an elevation of 5,600 feet not far from here. It was no doubt deposited at the level of the modern Columbia River – at less than 600 feet in elevation, suggesting uplift of nearly a mile. Note that the strata on the east flank of Lookout Mountain dip to the east.

Follow S.R. 970 and S.R. 10 east from Cle Elum. Follow S.R. 970 when it veers to the left about three miles east of Cle Elum.

For several miles the highway follows the Teanaway Valley, with the lofty Wenatchee Mountains to the north.

Milepost 6. The Teanaway River enters from a deep canyon to the northwest. S.R. 970 continues northeast and begins its ascent over the terminal moraine of the Pleistocene Yakima Glacier, which began at Snoqualmie Pass. Look for small to medium-sized glacial erratics.

STOP 13 **SITE 13.** Milepost 9. Crest of Yakima Moraine at an elevation of 2,394 feet. From here the road descends to Swauk Creek (elevation 2,220 feet) and ends at U.S. Highway 97 on which we will travel north.

Turn north on U.S. Highway 97.

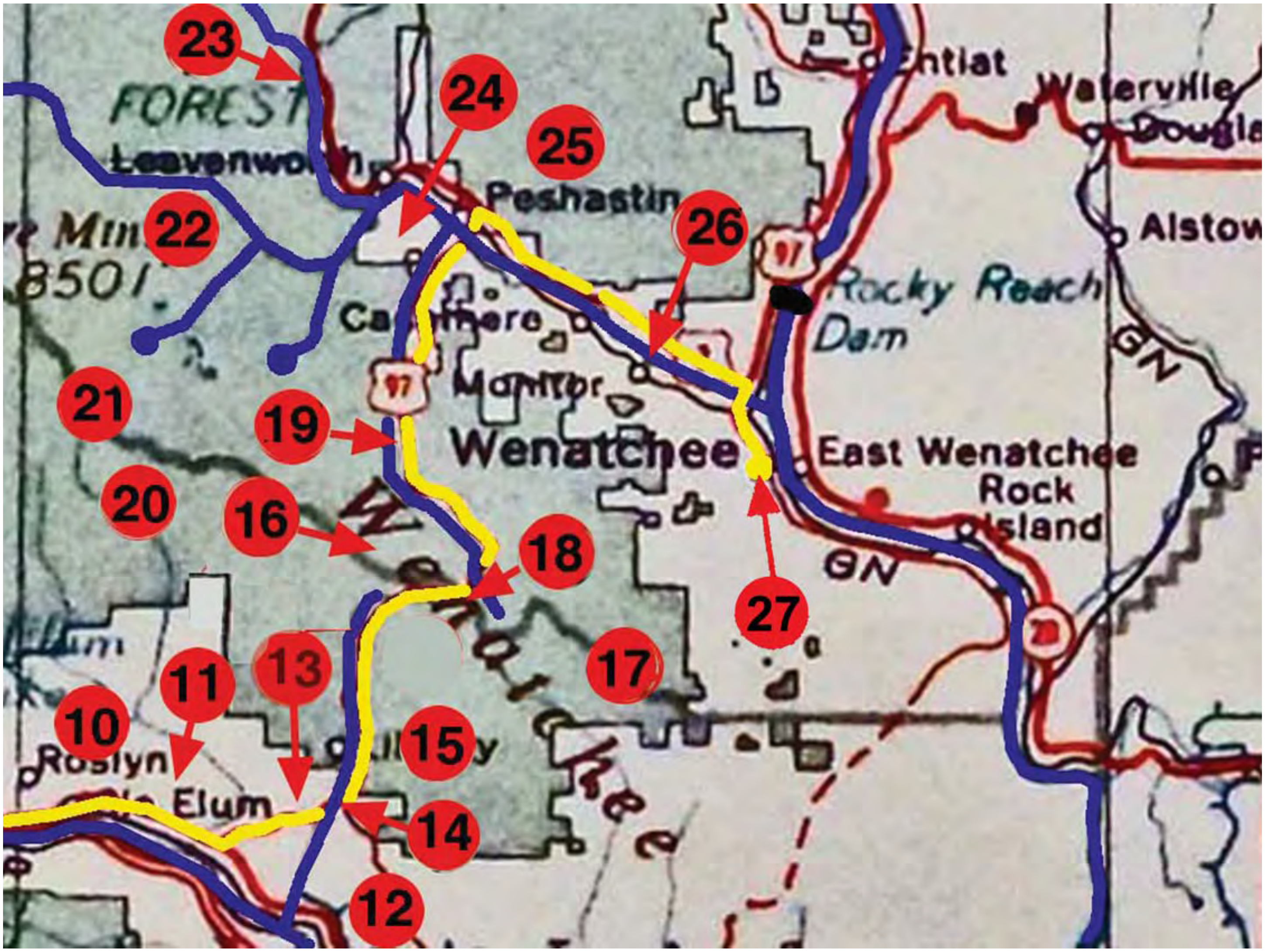

Figure 7. Sites 10-26. Noteworthy locations between Cle Elum and Wenatchee.

STOP 14

SITE 14. From one-half mile north of the U.S. Highway 97 junction to just downstream from Liberty Road, the floodplain of Swauk Creek is badly disrupted by dredge tailings.

STOP 15

SITE 15. Liberty Road leads to a fine view of a Teanaway Dike traversing a mountain side. The Liberty community was a gold mining center. It had an arrastra that operated until 1932. An arrastra is a primitive mill for grinding gold or silver ore. The simplest form of the arrastra is two or more flat-bottomed drag stones placed in a circular pit paved with flat stones, connected to a center post by a long arm.

STOP 16

SITE 16. Blewett Pass on the abandoned Blewett Pass Road. Travelers in the 1930's were terrified on this road because of the steep terrain, icy conditions and inadequate brakes characteristic of automobiles of that era.

STOP 17

SITE 17. At 5,600 feet above sea level, this is the highest outcrop of the Columbia River Basalt. It records the magnitude of uplift of the Cascade Range in the 15 million years since the Columbia River Basalt was emplaced. There is good reason to believe that the flood basalt flowed onto a flood plain at about the elevation of the modern Columbia River in the Pasco Basin, about 600 feet above sea level.

STOP 18

SITE 18. Swauk Pass, shown in figures 7 and 8. The modern highway crosses the Wenatchee Mountains here, but the public clings to the name Blewett Pass. The old Blewett townsite is about ten miles north of Swauk Pass. Blewett was a gold mining center where 260 miners worked until 1880. At a site west of the highway are grooves left by a waterwheel-powered arrastra. About

1.6 miles north of the townsite is the junction with the old Blewett Pass highway, which follows Peshastin Creek southward. The Peshastin Creek Canyon provides a cross-section through the suture zone, which is called the Ingalls Tectonic Complex. The roadcuts display outcrops of serpentenized peridotite from the mantle, metavolcanics, and meta-argillite. These units are shown as Jurassic on the map, but they may be considerably older, as heat from the Mount Stuart Batholith has recrystallized them (fig. 8).

STOP 19

SITE 19. U.S. 97 follows Tronson Creek northwest, parallel to the strike of a major suture zone. The Mount Stuart Batholith, which constitutes the highest part of the Wenatchee Mountains, represents the volcanic arc associated with the suture zone. Tronson Creek enters Peshastin Creek, which flows northward, parallel to the regional pattern of north-striking strike slip faults.

These two valleys exemplify the stress regime that prevailed from about 200 million to eight million years ago. This imprinted a northwest-southeast fold and thrust belt upon the landscape. Simultaneously, there was right-lateral, strike-slip faulting striking north-south. Additional clockwise rotation of the Juan de Fuca Plate about eight million years ago changed the strike of the least principal stress to nearly east-west, establishing the Yakima Fold and Fault Zone.

STOP 20

SITE 20. The Ingalls Tectonic Complex marks the suture zone where the Mount Stuart volcanic arc was accreted to North America (fig. 8).

STOP 21

SITE 21. Mount Stuart, the highest, most prominent peak in the region, is part of the Mount Stuart Batholith, the Cretaceous volcanic arc. Owing to the stress regime of the rapidly converging oceanic and continental plates, the Batho-

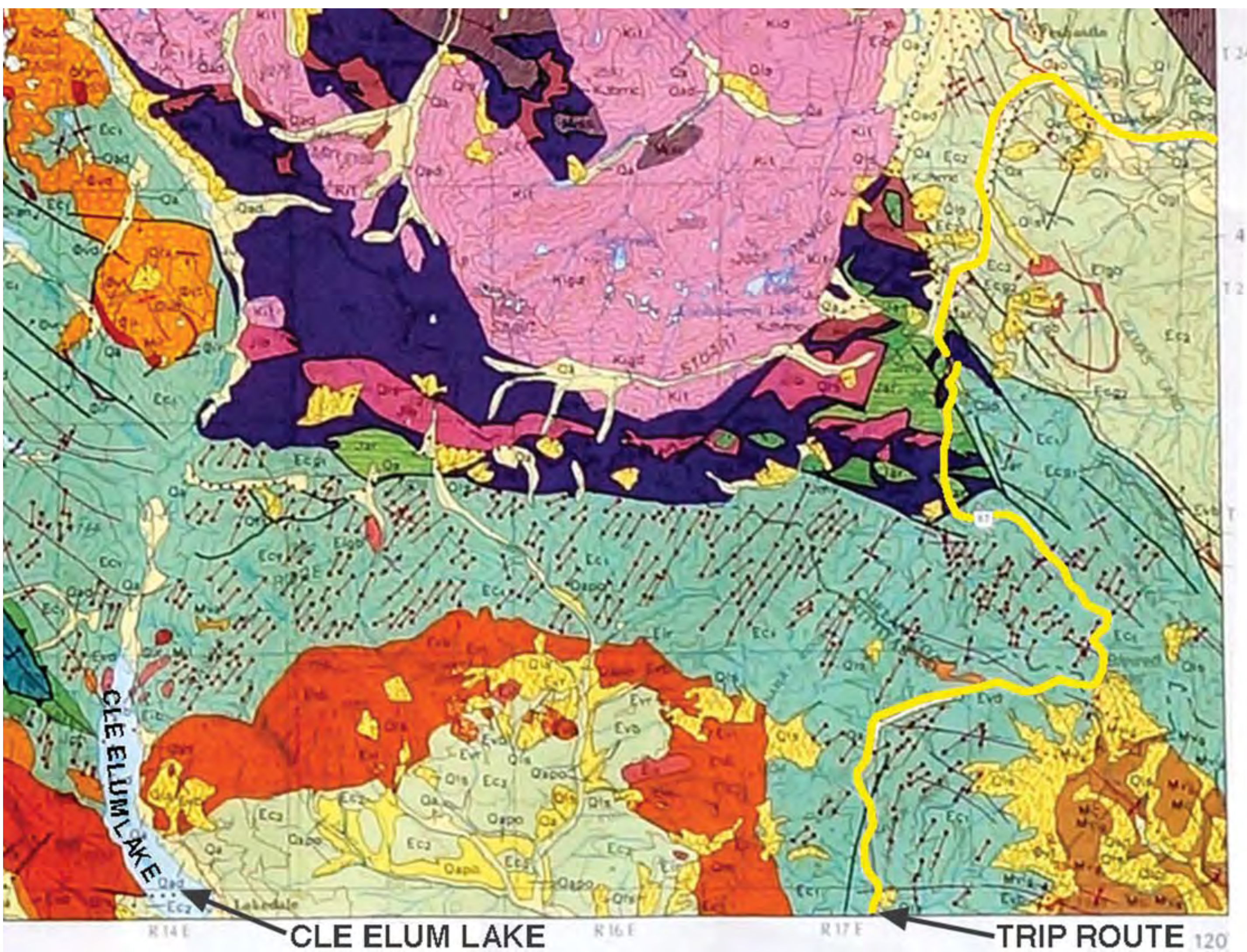

Figure 8. Swauk Pass, Ingalls Ophiolite, and Mount Stewart.

lith over-rode the suture zone, cutting out the entire forearc.

STOP 22

SITE 22. Icicle Creek. There is remarkable geology in this watershed. The road there provides access to the Enchantment Lakes Wilderness area, a magnificent alpine area of glaciated granitic rock of the Mount Stuart Batholith. Its grandeur rivals the High Sierra and Trinity Alps in central and northern California respectively, and the Wind River Range of Wyoming.

STOP 23

SITE 23. Tumwater Canyon of the Wenatchee River.

STOP 24

SITE 24. The lower reach of Peshastin Valley. There are many fruit orchards here which require irrigation.

Junction with U.S. Highway 2 in the Wenatchee River valley. Turn right onto eastbound lanes.

We are now in the Chewaukum Graben, bounded to the northeast by the Entiat Fault, a normal fault that juxtaposes the Eocene volcanics and sediments of the Chumstick Formation against the Eocene Swakane Gneiss. In the regional orogeny here, the Swakane Gneiss is at the bottom of a stack of thrust sheets (Miller and Paterson, 2001; Miller, et al., 2007). Until it was recently dated by isotopic methods, it was considered very old, possibly Precambrian because of its high-grade metamorphism.

There is debate about the nature of the Chumstick Formation and its relation to the Cascade terrane east of Leavenworth. Eric Cheney has dubbed the unit "Swaukstick" for he correlates it with the Swauk Formation.

The eastern slope of the Cascade Range is in a rain shadow. The arid climate supports a forest of Ponderosa Pine. Nearly every summer there are large, disastrous forest fires that emit dense smoke which makes

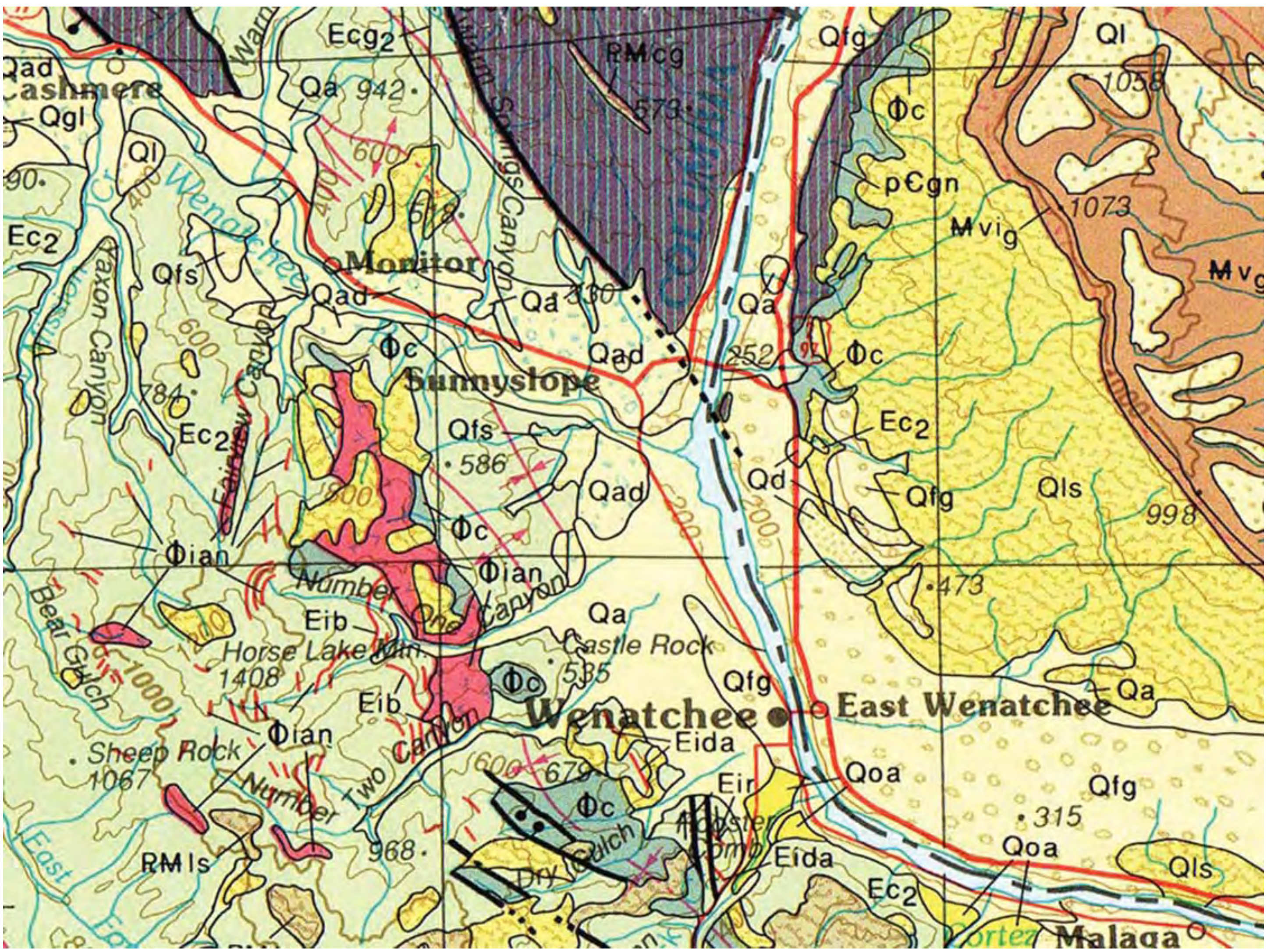

Figure 9. Andesite dike complex northwest of Wenatchee.

breathing difficult from the lower Peshastin Valley to East Wenatchee.

STOP 25

SITE 25. Peshastin Pinnacles. These crags consist of metamorphosed Chumstick Formation. They provide the best conditions for rock climbers in Washington and Oregon.

STOP 26

SITE 26. Town of Monitor. South of here is a complex of Oligocene intrusive andesite dikes which make a high, jagged ridge west of Wenatchee (fig. 9). In Stemilt Canyon at Wenatchee's southern edge, a lode of gold was developed into the Cannon Mine in the 1970's. It was one of Washington's largest sources of gold. Owing to modern mine technology, all the recoverable gold had been removed and the mine was shut down after only 20 years of operation.

STOP 27

SITE 27. Wenatchee. The city is at the confluence of the Wenatchee and Columbia Rivers. It has miles of interesting waterfront park along the Columbia River.

SITE 28 is not included in this publication because they are not shown in either figure 7 or 10**.**

STOP 29

SITE 29 (fig. 10). Cannon Mine Site. This was a major gold deposit which was mined out in about 20 years in the late 20th Century. Eric Cheney and his two partners were instrumental in recognizing the lode.

STOP 30

SITE 30. Stop at the Forest Service Headquarters for a view of andesite dike complex northwest of Wenatchee (fig. 9).

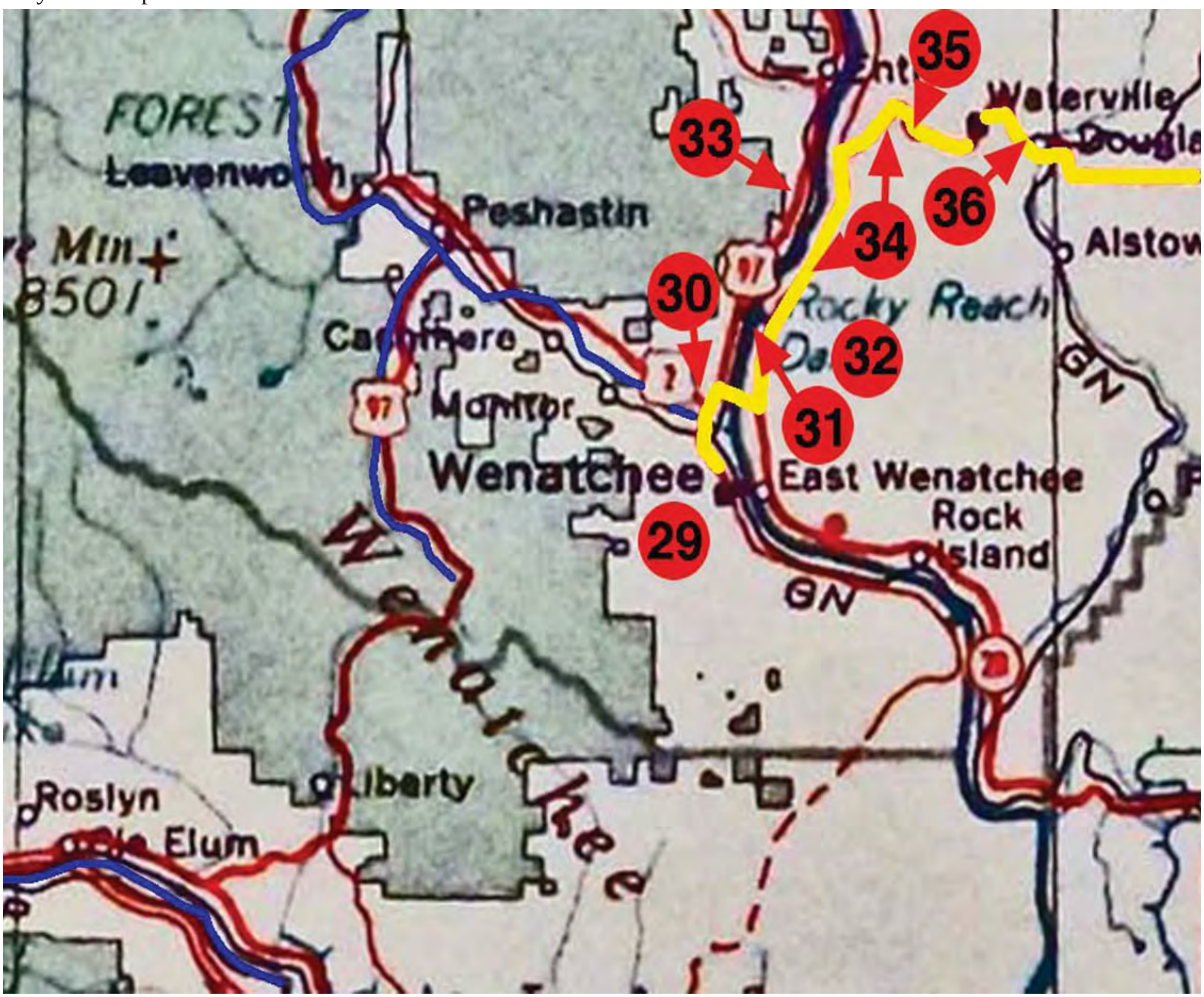

Figure 10, Sites 29-36. Noteworthy locations between Wenatchee and Waterville.

STOP 31

SITE 31. Rocky Reach Dam was built and is operated by the Chelan County Public Utility District. The county, with its small tax base, could never have financed it except through the Washington Public Power Supply System (WPPS). It has a large visitor center with many interesting displays about dam construction, local history, and archeology. Bighorn sheep are frequently seen on the steep slope to the west.

STOP 32

SITE 32. Badger Mountain. This anticline towers above the Columbia River, making a spectacular backdrop, with gravel and kame terraces left by the Lake Missoula Floods.

STOP 33

SITE 33. Earthquake Point. The Earthquake of 1872 caused a landslide here which temporarily dammed the Columbia River and pushed the channel eastward.

STOP 34

SITE 34. Swakane Gneiss

STOP 35

SITE 35. Contact between the Swakane Gneiss and the Columbia River Basalt, that contains invasive basaltic dikes.

STOP 36

SITE 36 (fig. 11). Lava Delta of the Columbia River Basalt

STOP 37

SITE 37. Withrow Moraine, the terminal moraine of the Okanogan Lobe of the Cordilleran Glacier.

STOP 38

SITE 38. Jameson Lake, which is impounded in Moses Coulee by the Withrow Moraine. Moses Coulee was eroded by glacial outburst floods, or jokulhlaups, from the Okanogan Lobe of the Cordilleran Glacier in an earlier stage of advance than Withrow and Grand Coulee.

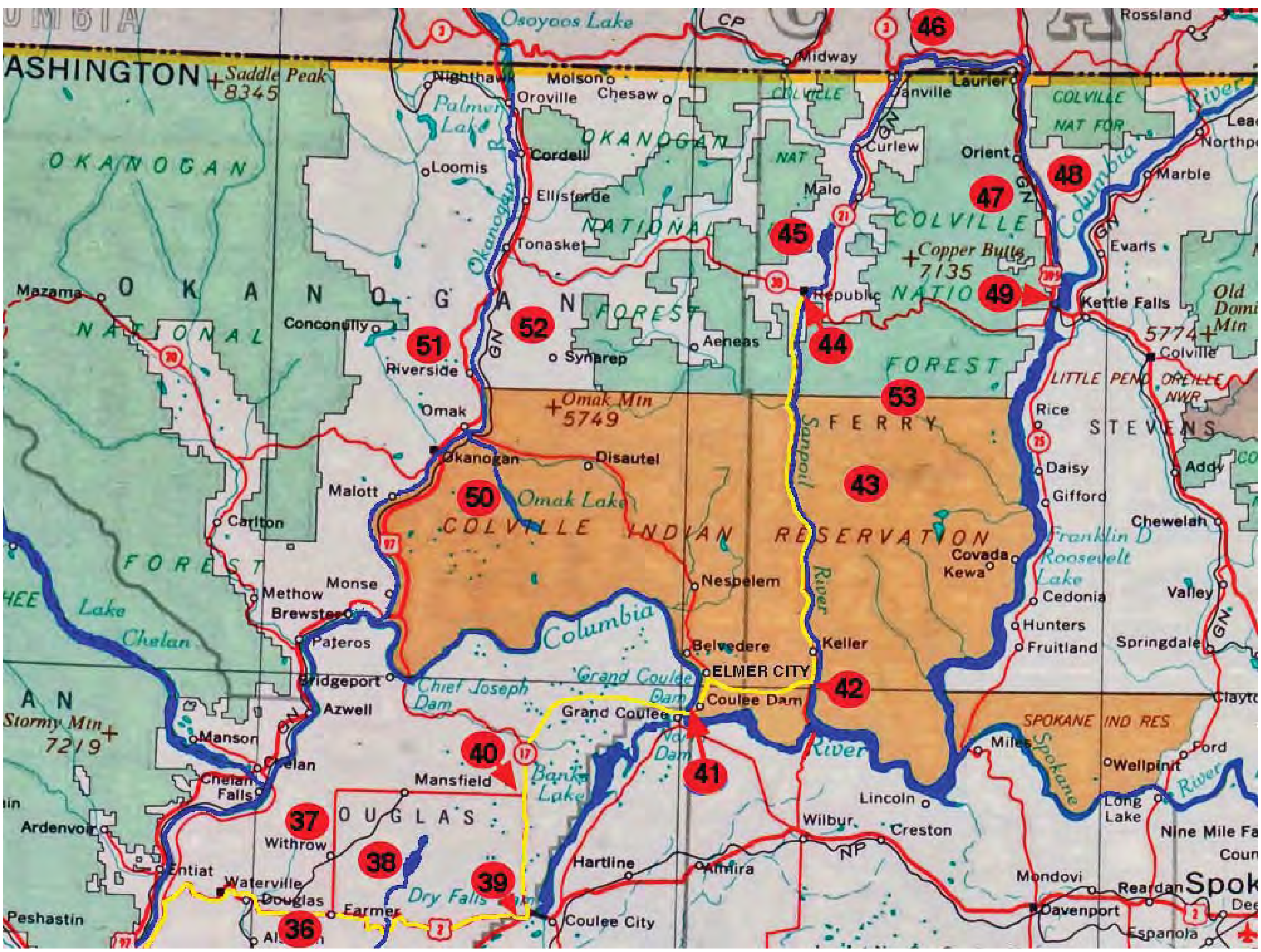

Figure 11, Sites 37-53. Noteworthy locations between the Withrow Moraine and the Kettle Metamorphic Core Complex.

STOP 39

SITE 39. East end of the Withrow Moraine. There are widely scattered haystack rocks on both sides of U.S. Highway 2.

STOP 40

SITE 40. Sims Corner, which is surrounded by immense kames, eskers and kettles. There are more kames and many haystack rocks to the West on S.R. 172.

STOP 41

SITE 41. Stop at vista point for great panorama of Grand Coulee Dam and its surroundings. The dam has a most interesting Visitor Center.

STOP 42

SITE 42. Sanpoil Bay and Manila Canyon. Brian Atwater mapped this area in great detail and identified stratigraphic evidence for at least 88 jokulhlaups in the latest (Fraser - Pinedale) glacial advance.

STOP 43

SITE 43. Sanpoil Volcaniclastics - andesite ash flows alternating with debris flows (lahars) which partially filled the Republic Graben. The Republic Graben is bounded by normal faults separating the Okanogan metamorphic core complex (MCC) on the west side, from the Kettle MCC to the east. The graben extends from the Columbia River into British Columbia.

STOP 44

SITE 44. Town of Republic. Eocene plant fossils abound within the town. Republic had an underground gold mine that shut down about 2001 after 105 years of operation.

STOP 45

SITE 45. Curlew Lake. The Sanpoil River flows south from here, and the Kettle River flows north. The lake is at the high point in the Republic Graben. When the Okanogan Lobe blocked the natural drainage, the outwash stream flowed south and its turbulence at the high point eroded the closed basin that is now occupied by Curlew Lake.

STOP 46

SITE 46. Grand Forks, British Columbia on the Kettle River. The Kettle River loops around the nose of the Kettle Dome (MCC), then flows south in the detachment fault between the Kettle MCC and Quesnellia.

STOP 47

SITE 47. Anticline composed of folded quartzite and limestone in the Kettle Detachment Zone.

STOP 48

SITE 48. Quesnellia - detached upper plate of the Kettle Metamorphic Core Complex.

STOP 49

SITE 49. Amphibolite with granitic boudins in the Kettle Detachment Fault Zone.

STOP 50

SITE 50. Omak Lake in the Okanogan Detachment Fault Zone.

STOP 51

SITE 51. Detached upper plate of Okanogan Detachment Fault Zone.

STOP 52

SITE 52. Here the Okanogan Metamorphic Core Complex verges to the west.

STOP 53

SITE 53. Here the Kettle Metamorphic Core Complex verges to the east.

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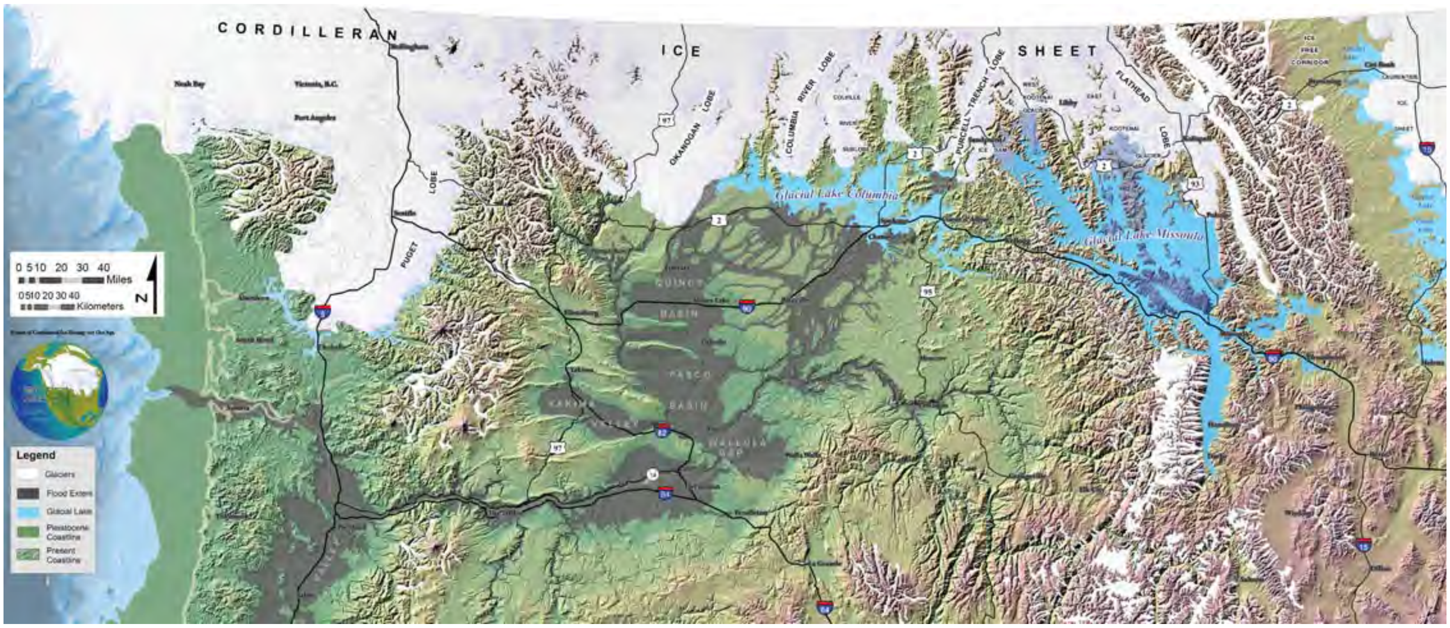
OUTBURST MISSOULA FLOODS AND GLACIAL LAKE COLUMBIA ALONG THE NORTHERN CHANNELED SCABLAND

Bruce Bjornstad

Senior Research Scientist
Pacific Northwest National Laboratory



The Pacific Northwest 20,000 years ago during the maximum extent of the Cordilleran Ice Sheet. Outburst floods from Glacial Lake Missoula repeatedly overran the pre-existing Glacial Lake Columbia and overfilled the Columbia Valley. This forced floodwaters to spill southward across the Channeled Scabland.

This field-trip guide explores the evidence for repeated outburst floods from Glacial Lake Missoula. It also explores Glacial Lake Columbia, another huge body of water that persisted along the northern margin of the Channeled Scabland (Baker and Nummedal, 1978) during the time of the last Pleistocene glaciation and the Glacial Lake Missoula floods. While outwash flood deposits from Glacial Lake Missoula are conformable and interbedded with quiet-water silts and clays from Glacial Lake Columbia at Stop 1, it is likely that Glacial Lake Columbia existed for many centuries after the last outburst from Glacial Lake Missoula.

Mileage

0.0 *Leave TRGS meeting place in Republic, WA. Head south on SR 21 to Keller Ferry.*

Features and stops along the tour route are numbered in figure 1.

51 **STOP 1** **Stop 1. Keller Ferry - Missoula Flood Rhythmites and Glacial Lake Columbia (fig. 2)**

Multiple exposures like this occur around the Sanpoil Arm of Lake Roosevelt (a modern, much-lower version of Glacial Lake Columbia). In this photo, darker sandy rhythmites, each laid down in a matter of hours during the Missoula flooding, are capped by lighter-colored, varved lake beds from Glacial Lake Columbia. Based on these deposits, up to 89 Missoula flood events, separated by only a few dozen years, are hypothesized for the late Wisconsinan glaciation (Atwater, 1984, 1987; Waitt, 1984; Waitt, Denlinger, and O'Connor, 2009.

Cross Keller Ferry and continue south on SR 21. Just before the town of Wilbur turn northwest onto SR 174.

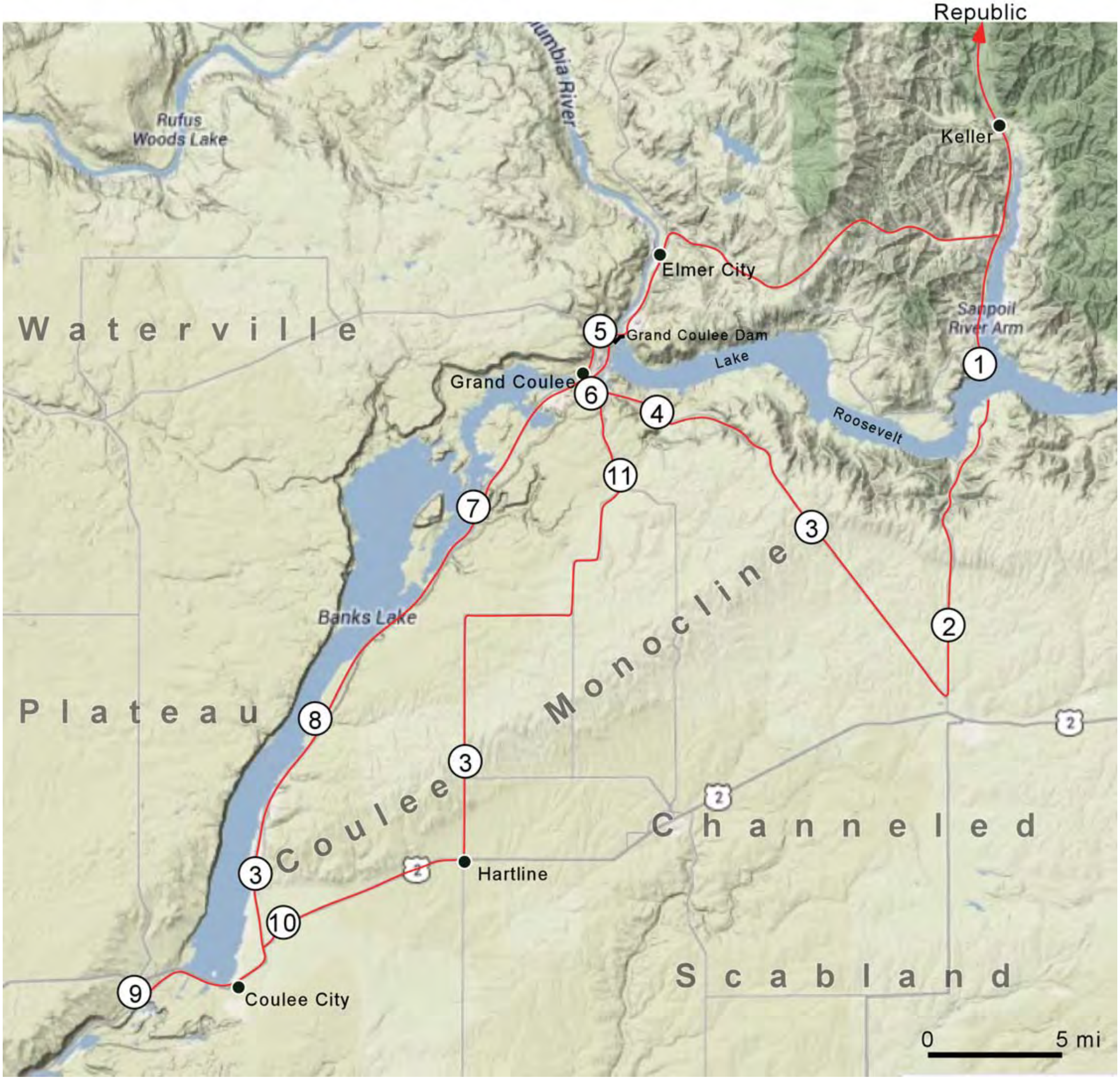


Figure 1. Location Map. Features and stops are numbered along the tour route.

61 **STOP 2** **Stop 2 (drive by). Corbett Draw Flood-Spillover Coulee (fig. 3, right)**

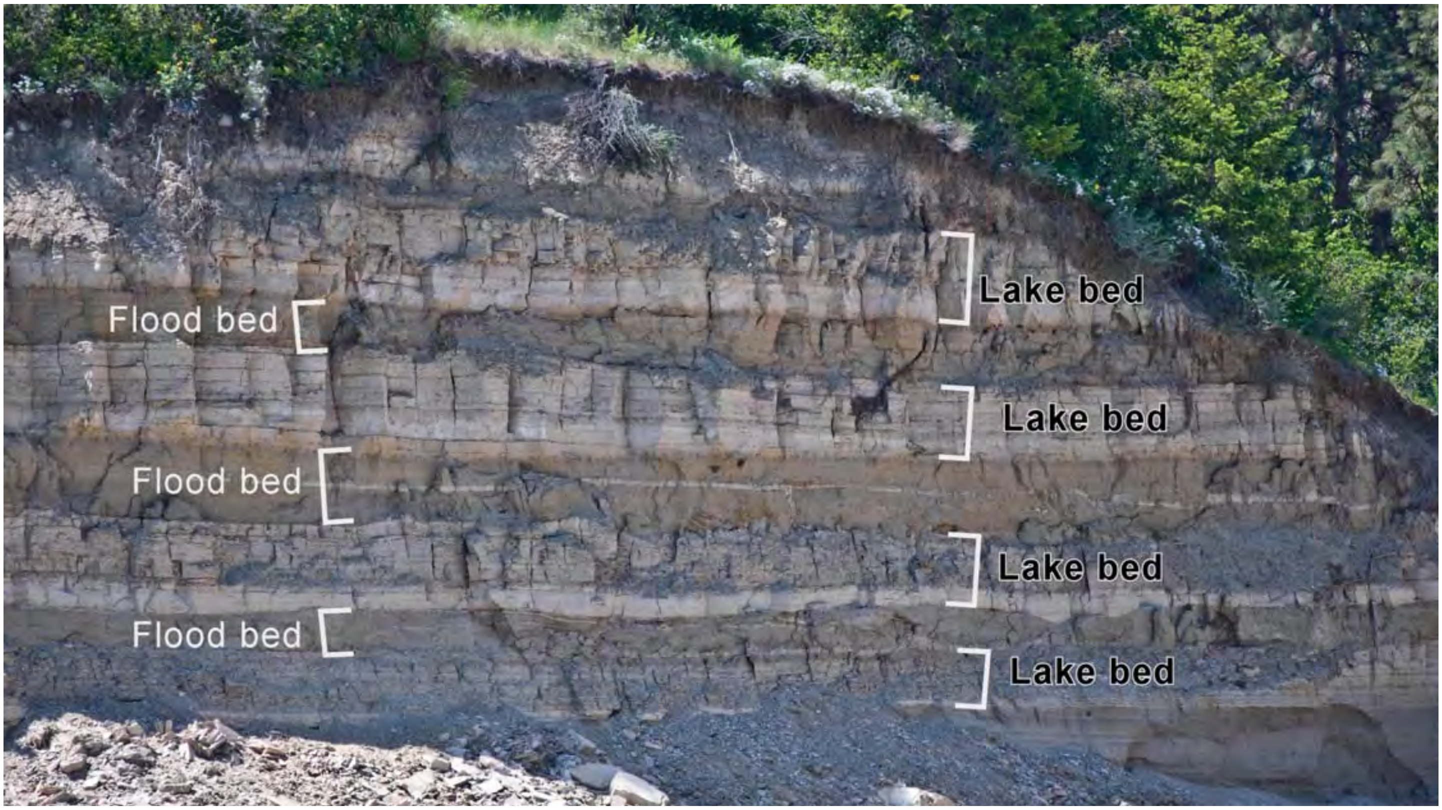


Figure 2. (Stop 1) Outburst flood deposits associated with Glacial Lake Missoula are conformable and interbedded with quiet-water lacustrine deposits from Glacial Lake Columbia.

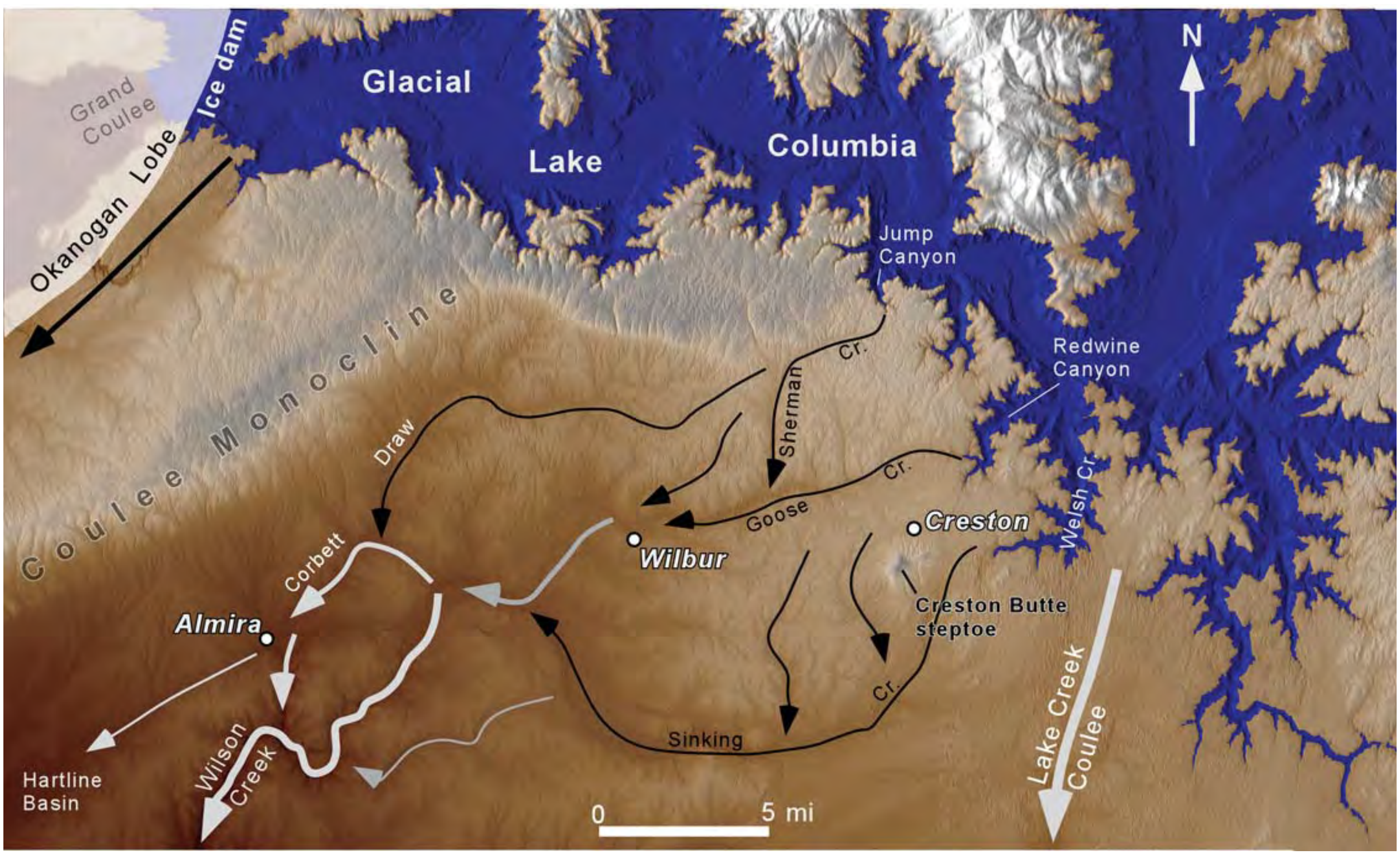


Figure 3. (Stop 2) After the Missoula floods overran pre-existing Glacial Lake Columbia and overfilled the Columbia River Valley, they spilled south onto the Channeled Scabland. Anastomosing spillover channels are indicated with arrows. Glacial Lake Columbia, dammed by the Okanogan Ice Lobe, is shown with a surface elevation of 2,300 feet. The Okanogan Lobe is shown near its maximum extent, when only a limited amount of floodwater was able to flow through a narrow opening between the ice sheet and the Coulee Monocline near Grand Coulee. With the ice in this position, most Missoula floodwaters were diverted down the Telford-Crab Creek and Cheney-Palouse scabland tracts to the east.

73 **STOP 3** **Stop 3 (drive by). Coulee Monocline**

Long before Miocene volcanic flows of the Columbia River basalt were folded, they flowed out of vents and spread out in broad, horizontal sheets across the Columbia Plateau. After the basalt cooled, tectonic forces locally compressed the flows, which caused bending and cracking of the hardened, brittle basalt into a monoclinal structure, the Coulee Monocline.

The Coulee Monocline (figs. 1 and 3) is an upland area that lies above the elevation of most of the Ice Age floods. It forms a bounding ridge that runs along the west side of the Lower Grand Coulee before crossing the mouth of Upper Grand Coulee and continuing northeast (see fig. 1). It was a key structural feature that strongly influenced the development of Grand Coulee. It diverted the floodwaters spilling out of the Columbia Valley from the north and west down Grand Coulee, while to the east it diverted floodwaters down the Telford-Crab Creek scabland tract (Bjornstad and Kiver, 2012). (Flood tracts are the major channel networks visible where floodwaters eroded away all topsoil and incised into the underlying basalt bedrock).

The recessional cataract canyon that formed the Upper Grand Coulee started here, where the coulee crosses the monocline. The youngest basalt flows consist of Wanapum Basalt (Stoffel et al., 1991), emplaced 15.6 and 14.5 million years ago. They are thicker on the downfolded side of the Coulee Monocline, indicating the monocline was actively being uplifted and folded as the volcanic flows lapped up against this rising ridge. In places the rocks at the top of the monocline are 1,000 feet higher than the same rocks at the base of the monocline.

82 **STOP 4** **Stop 4. Glacial Lake Columbia Strandlines (fig. 4)**

Strandlines go up to about 2,400 feet in elevation, marking the highest level of Glacial Lake Columbia. These strandlines formed prior to the last Missoula floods, which are known to have breached the head of Grand Coulee, significantly lowering the outlet for Lake Columbia from 2,400 to 1,500 feet in elevation.

90 **STOP 5** **Stop 5. Okanogan Ice Dam at Crown Point (fig. 5)**

These parallel grooves in polished granitic bedrock at Crown Point are glacial striations left behind as the slowly creeping Okanogan Ice Lobe moved over this rock surface (Bjornstad and Kiver, 2012). Total thickness of the ice in this area at times approached 5,000

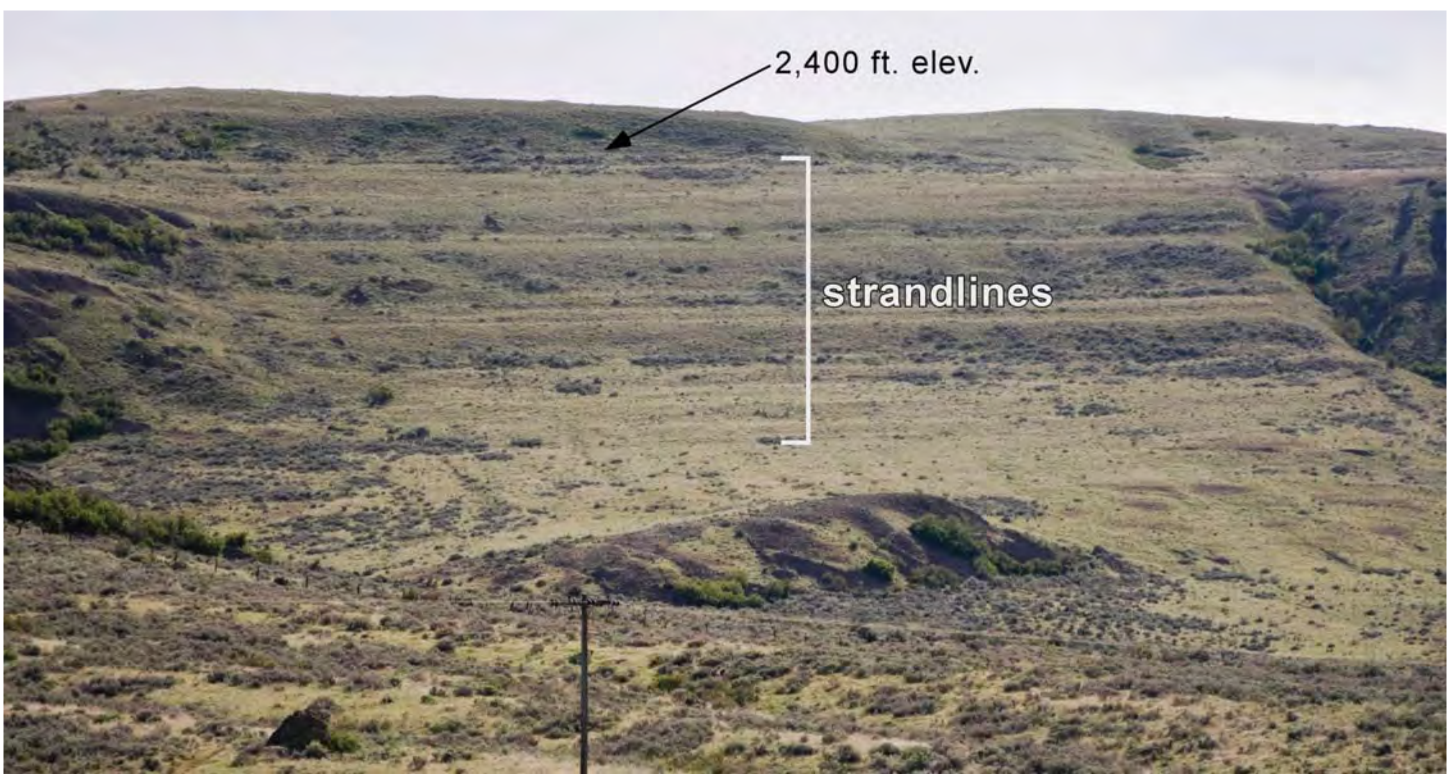


Figure 4. Horizontal strandlines (wave-cut benches) are etched into the high, southern wall of the Columbia River Valley between Spring and Neal Canyons (Bjornstad and Kiver, 2012).

Figure 5. Looking south from Crown Point Vista towards Grand Coulee Dam and Lake Roosevelt (upper left).

feet, which completely blocked the Columbia River and created Glacial Lake Columbia. It persisted for thousands of years, before breaking apart and releasing a last flood down the Columbia Valley somewhere between 14,000 to 15,000 years ago.

Retrace route to Grand Coulee and turn southwest onto SR 155.

100 **STOP 6** **Stop 6. Missing Cataract at Head of Grand Coulee**

As floodwater eroded the Upper Grand Coulee, the 900-foot-high cataract eroded headward (receded) a full 25 miles until it broke through here into the Columbia River Valley, which was much lower in elevation (fig. 6). (Cataracts formed from large waterfalls flowing over a high cliff.) Ice Age floods were diverted down the Upper Grand Coulee through most of the last glacial cycle when the Okanogan Ice Lobe blocked the Columbia River near present-day Grand Coulee Dam (fig. 7). Glacial Lake Columbia formed behind the ice dam and quickly filled with glacial meltwater.

Grand Coulee developed when Ice Age floods flowed along the edge of the ice sheet over an upland plateau at approximately 2,400 feet in elevation and across the Coulee Monocline where the Upper Grand Coulee recessional cataract originated. With each successive flood, the cataract eroded headward (north) carving out a deep, narrow canyon that lengthened with each flood outburst. Toward the end of the last glacial cycle (about 17,000 years ago) the cataract

Figure 6. Looking southwest at Grand Coulee Dam, which is located near the location of the ice dam for Glacial Lake Columbia.. Grand Coulee is in the background, with the Banks Lake reservoir occupying the floor of the coulee. This reservoir is filled with water pumped from Lake Roosevelt as part of the Columbia Basin Irrigation Project. In the distance are the Banks Lake inselbergs and the Steamboat Rock Monolith.

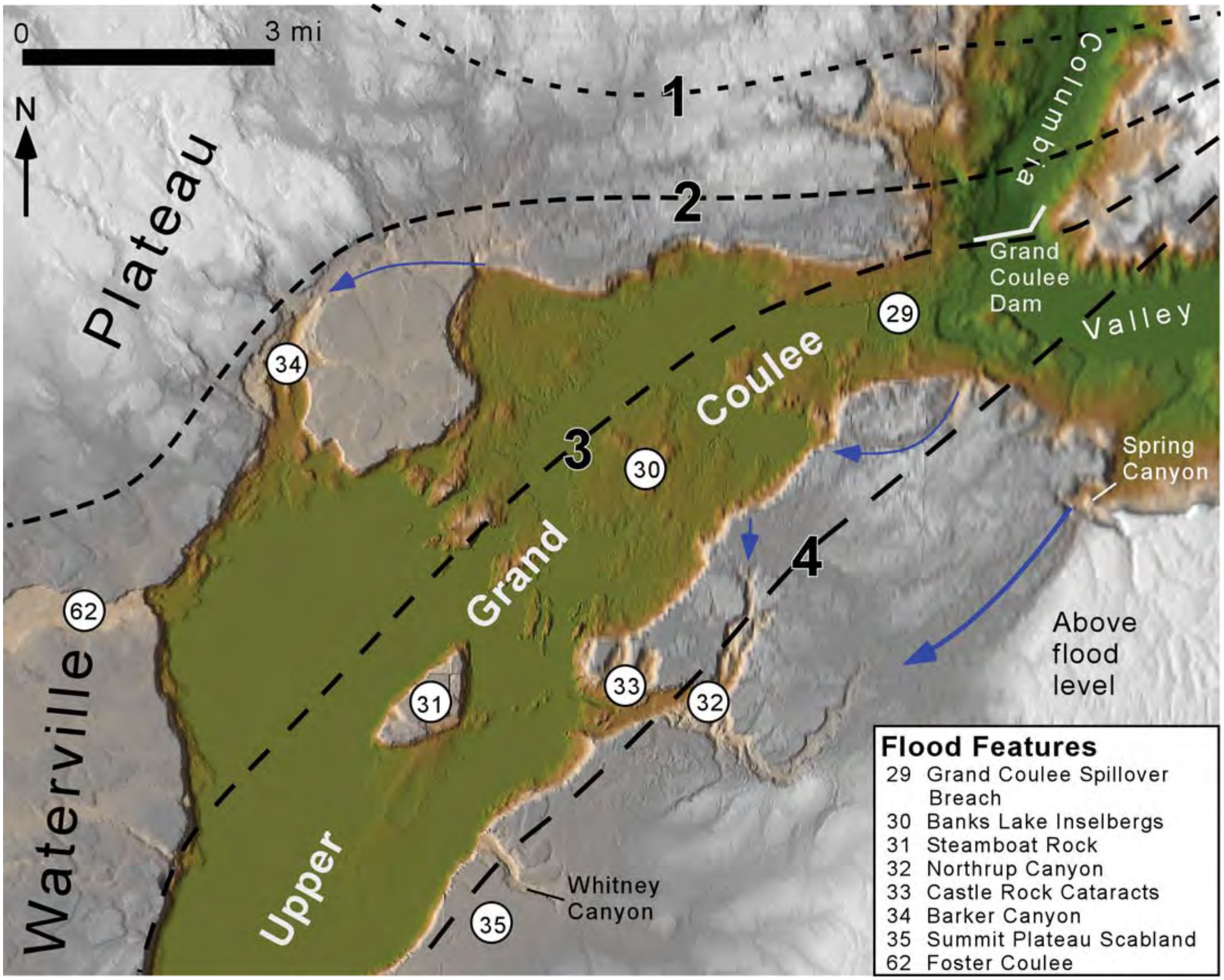


Figure 7. Flood flow into Grand Coulee funneled between the uplands to the southeast and glacial ice to the north and west (dashed lines). During the maximum ice advance (position 4), when most of the main coulee was plugged with ice, some floodwater flowed southwest across the high plateau to the east. At other, less advanced positions of glacial ice (1 through 3) floodwaters invaded and eroded wider swaths of Grand Coulee.

finally breached the head of Grand Coulee (fig. 6). In essence, when the 900-foot high cataract, reached the head of the Grand Coulee, it self destructed (Bretz, 1932). Today the coulee walls on either side indicate the height of the missing cataract.

The two-mile wide breach lowered the spillover channel a full 900 feet across this divide. Today the head of Grand Coulee still hangs more than 500 feet above the natural level of the Columbia River below. The breached spillover was 800 feet lower than any of the other spillovers into the Channeled Scabland. After the head of Grand Coulee was breached, some of the last, smaller floods may have travelled entirely down Grand Coulee, bypassing the higher spillovers into the Cheney-Palouse and Telford-Crab Creek scabland tracts to the east. The Grand Coulee, which was suddenly deeper, may have hijacked most or all the water from subsequent outburst floods. With its lower lip into the Channeled Scabland it became the first coulee to fill and the last to drain. The breached Grand Coulee likely continued to carry floodwaters from Glacial Lake Missoula for over ten days (Waitt, Denlinger, and O’Connor, 2009). When compared to highly jointed basalt, the granitic rocks that make up the floor of the upper coulee were more resistant to erosion. If not for the more-resistant granitic rocks, the floods might have carved an even deeper channel across the Upper Grand Coulee.

106.5 STOP 7 **Stop 7. Inselbergs and the Northrup Canyon Eddy Bar (figs. 8 and 9)**

Inselbergs are smooth, rounded rocky knobs of granite that protrude at the head of Grand Coulee (Keszthelyi et al., 2009). Inselbergs also exist as a group of islands (i.e., archipelago) within Banks Lake (fig. 8). The granitic rocks here originated in granitic batholiths intruded from 50 to 100 million years ago before being slowly unroofed and exposed at the surface for eons. About 17 million years ago the granite was reburied beneath lava flows of Columbia River basalt. Only in recent geologic history did the granitic rocks become re-exhumed along the recessional cataract canyon of Grand Coulee. Cosmogenic age dates (that measure the amount of certain isotopes produced by exposure of minerals to cosmic rays) of the granite indicate the inselbergs weren't exhumed until about 17,000 years ago (Keszthelyi et al. 2009). The morphology of the granitic inselbergs differs markedly from that of the basalt primarily due to the contrasting fracture patterns, layering of the basalts, and relative homogeneity of the granite rock.

116.5 **STOP 8** **Stop 8. Glacial Lake Columbia Varves along Paynes Gulch Bar**

A thick deposit of yellowish, laminated silt (fig. 10) indicates that a long-lived arm of Glacial Lake Columbia occupied the Upper Grand Coulee for several centuries after the last Missoula outburst flood (Atwater, 1984, 1987). This deposit, referred to here as the Lake Columbia Silt, is also known as the Nespelem Silt, the Nespelem Formation, and the Steamboat Rock Silt.

Lake Columbia and the silt deposits did not occupy the Upper Grand Coulee until after glacial floods breached the head of the coulee and significantly lowered its elevation. The silt deposits are composed of up to hundreds of annual varves that blanket the youngest flood deposits. The Lake Columbia silt lies on the floor of the Upper Grand Coulee and continues

Figure 8. The light-colored, irregular knobs of granitic basement rock in foreground are inselbergs. The granite was exhumed after Ice Age floods locally eroded through the overlying basalt in the Upper Grand Coulee.

Figure 9. Looking east at a close up of high-energy flood sediments exposed in an eddy bar at the mouth of Northrup Canyon. The huge flood-deposited boulders and poorly sorted debris are stark testimony to the speed and power of the floods. This pile of flood debris was probably deposited as an eddy at the mouth of Northrup Canyon by the last floods that came down Grand Coulee.

up the sides to an elevation of 1,620 feet, although it is reported as high as 1,800 feet elsewhere in the region. Thus, it appears Glacial Lake Columbia may have filled the coulee to an elevation between 1,700 to 1,800 feet as silt was accumulating at the bottom of the quiet waters.

The post-Missoula-flood Lake Columbia silt also appears to blanket giant ripples and coarser flood deposits produced by water moving with higher energy than those that underlie the Paynes Gulch flood bar. A map produced by J Harlen Bretz (1932) shows that the silt blankets the entire floor of the coulee, including the area submerged under Banks Lake.

More evidence for a long-lived glacial lake in the Upper Grand Coulee is provided by delta bars, located in several places along the west side of the coulee. Elevated deltas that advanced into the ancient Lake Columbia indicate that a lot of glacial meltwater from the Okanogan Lobe was draining off the Waterville Plateau into Upper Grand Coulee via Barker Canyon, as well as Foster and Horse Lake Coulees. The delta bars reflect deposition of sediment by these glacial-meltwater streams into Glacial Lake Columbia, where the water slowed as it entered the lake. Perfect preservation of these delta bars further suggests that Glacial Lake Columbia outlasted the last Missoula outburst floods. Had more floods moved through Grand Coulee, they would have certainly destroyed or modified the delta bars, which is not the case (Bjornstad and Kiver, 2012).

Near Coulee City head south and west along US 2. At the west end of Dry Falls Dam turn south onto SR 17.

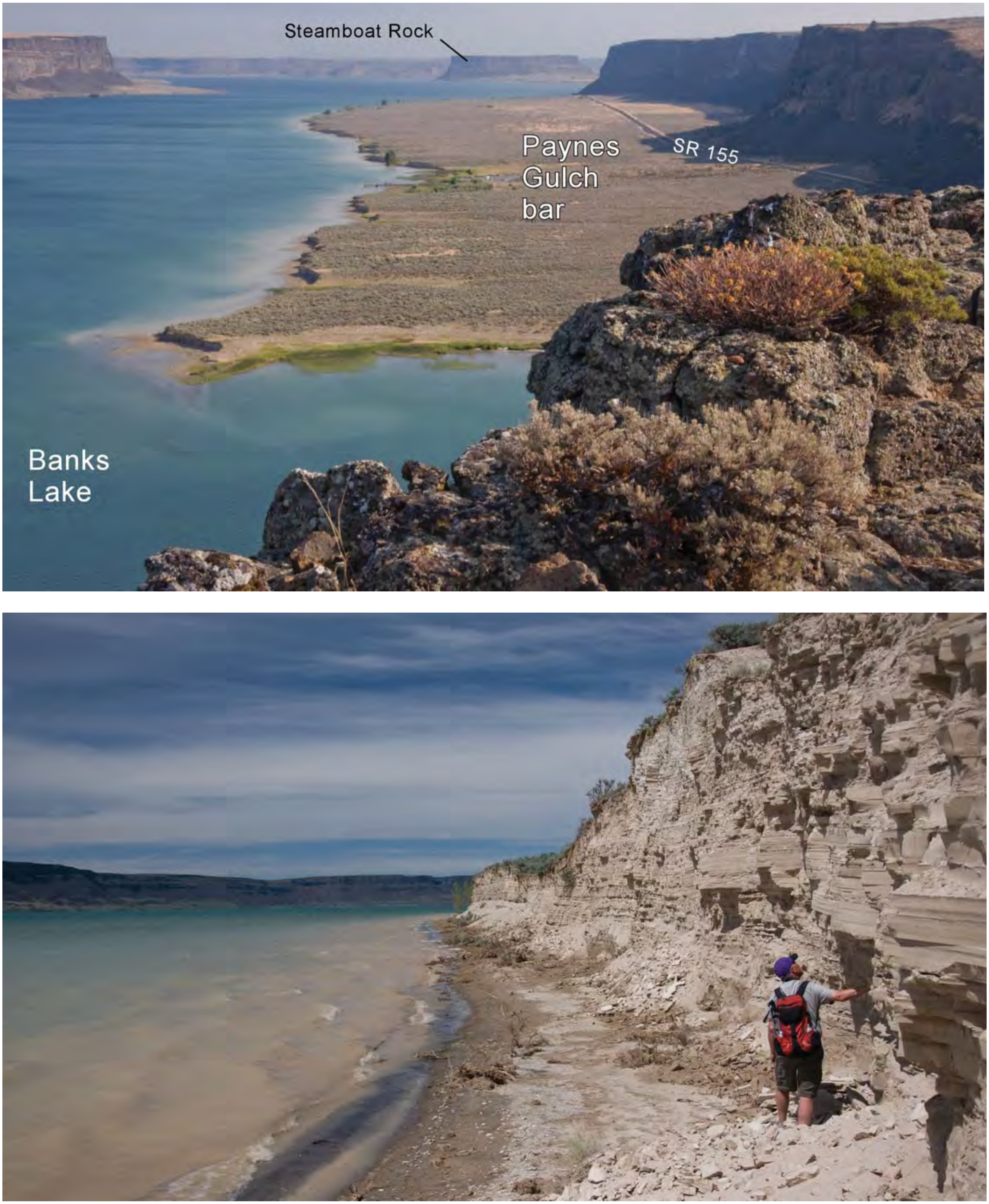


Figure 10. Top: Paynes Gulch Bar in the Upper Grand Coulee. Bottom: Glacial Lake Columbia deposits of well-laminated yellowish silt exposed along Banks Lake.

132 **STOP 9**

Stop 9. Dry Falls and the Great Cataract Group

Dry Falls is perhaps the premier flood attraction of all the Channeled Scabland, primarily due to easy access to a spectacular overview from the precipitous edge of the falls and an informative visitor's center, the J Harlen Bretz Memorial Museum (fig. 11).

Amazing as the view is from Dry Falls Visitor Center, one can really only see a portion of the multi-alcoved cataract that makes up Dry Falls (fig. 12). The eastern alcove, called Monument Coulee, is mostly obscured by Umatilla Rock, a tall blade of basalt that divides the coulee into two almost equal segments (fig. 11). Two plunge-pool lakes, Dry Falls and Green Lake, are visible from the viewpoint; a third lake (Red Alkali) is hidden against the cataract at the head of Monument Coulee. So much water flowed through here that the falls would have been almost imperceptible, with perhaps only a minor drop in the level of the floodwaters at the falls. The geology looks much like it did when cataclysmic floods last poured over the falls about 15,000 years ago. The dry, semi-arid climate with slow rates of weathering and erosion have helped to preserve flood features both here and elsewhere in the scabland.

Retrace route to Coulee City before heading east on US 2.

139 **STOP 10**

Stop 10 (drive by). Coulee City Expansion Bar

A huge expansion bar fanned out into the Hartline Basin at the mouth of Upper Grand Coulee as floods 500 feet deep and moving up to 70 mph temporarily slowed, causing sediment to accumulate (fig. 13). A high bouldery surface up to 1,850 feet in elevation gently descends from the mouth of Upper Grand Coulee into the Hartline Basin. Coarse boulders at the gaping mouth of the coulee grade into finer debris further down the fan. At one time the total width of the expansion bar must have been an incredible 16 miles, from Coulee City on the east to Dry Falls on the west (fig. 13). The expansion bar is composed completely of basalt clasts, no granitic rock is present. This indicates that the bar formed during older Ice Age floods at a time before the cataract of the Upper Grand Coulee had receded to Steamboat Falls (former cataract near Steamboat Rock and exhumed granitic rocks north of there). Had the bar developed after granitic rocks were exposed these rocks would be present in the bar.

The expansion bar acted as a debris dam that temporarily impounded Glacial Lake Columbia at the southern end of the Upper Grand Coulee. The central portion of the expansion bar was lowered and removed by later floods. The demise of the Coulee City Expansion Bar may have come during the 900-foot breach of Upper Grand Coulee. An alternate explanation for later incision of the bar could be more intense ero-

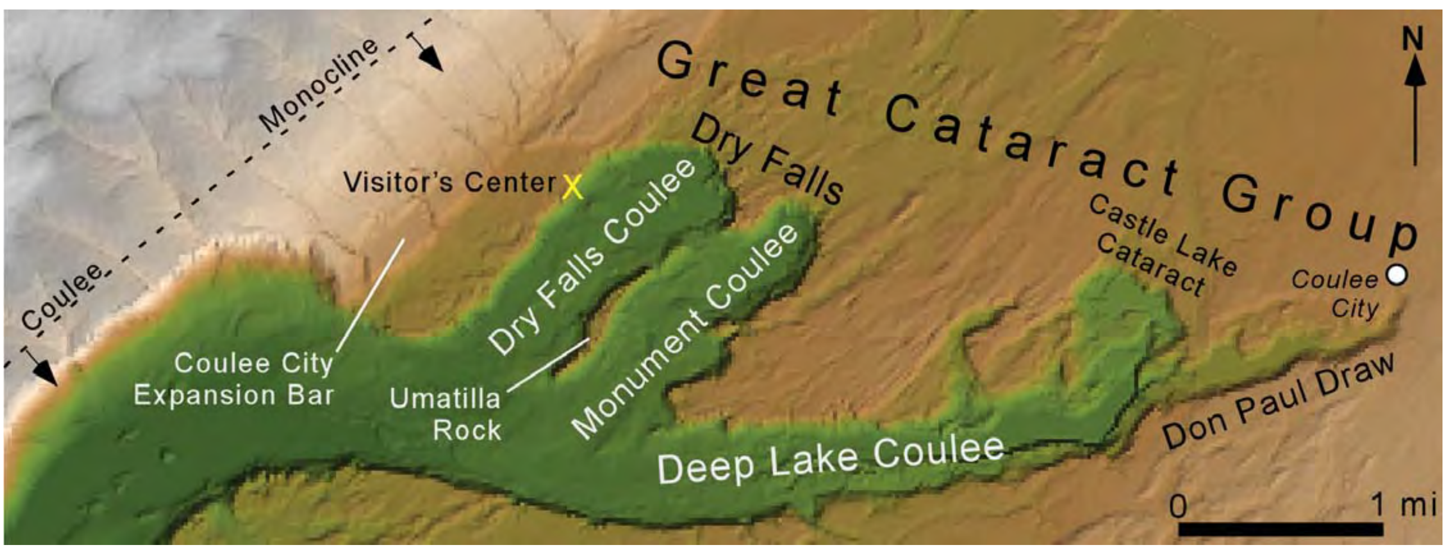


Figure 11. Shaded-relief map of the four-mile-wide Great Cataract Group. At the west end, the Umatilla Rock Blade neatly divides the Dry Falls and Monument canyons which were carved by recessional cataracts. The Castle Lake canyon, also carved by a recessional cataract, lies further to the east. The eastern end of the Great Cataract Group extends a mile further to the head of Don Paul Draw at Coulee City. The Dry Falls recessional cataract began about 15 miles to the south, where floodwaters initially dropped over a basalt ledge into the Quincy Basin near Soap Lake at the mouth of Lower Grand Coulee.

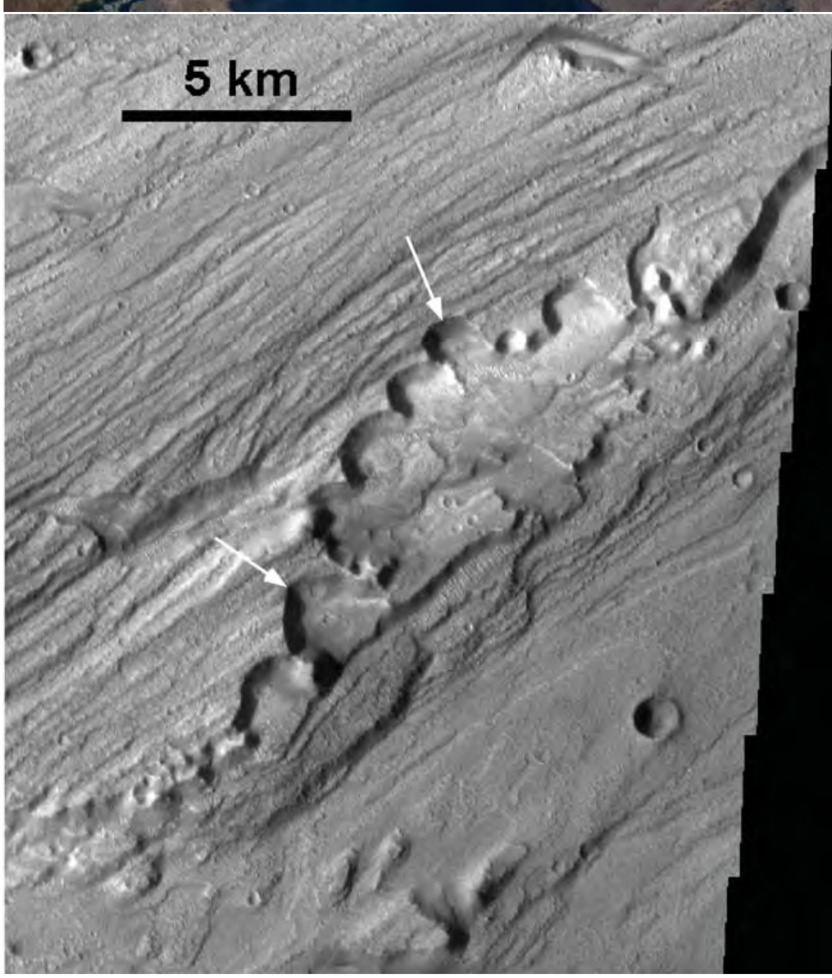


Figure 12. Looking northwest. Grooved upland plateau around the scalloped head of Dry Falls cataract. Top: Umatilla Rock divides Dry Falls from Monument coulees. The south end of Banks Lake Reservoir is visible at upper right. Bottom: Kasei Valley, Mars. Compare with similarly grooved plateau above. This comparable string of cataract alcoves on the planet Mars is about twice as long as those of the Great Cataract Group shown in figure 11.

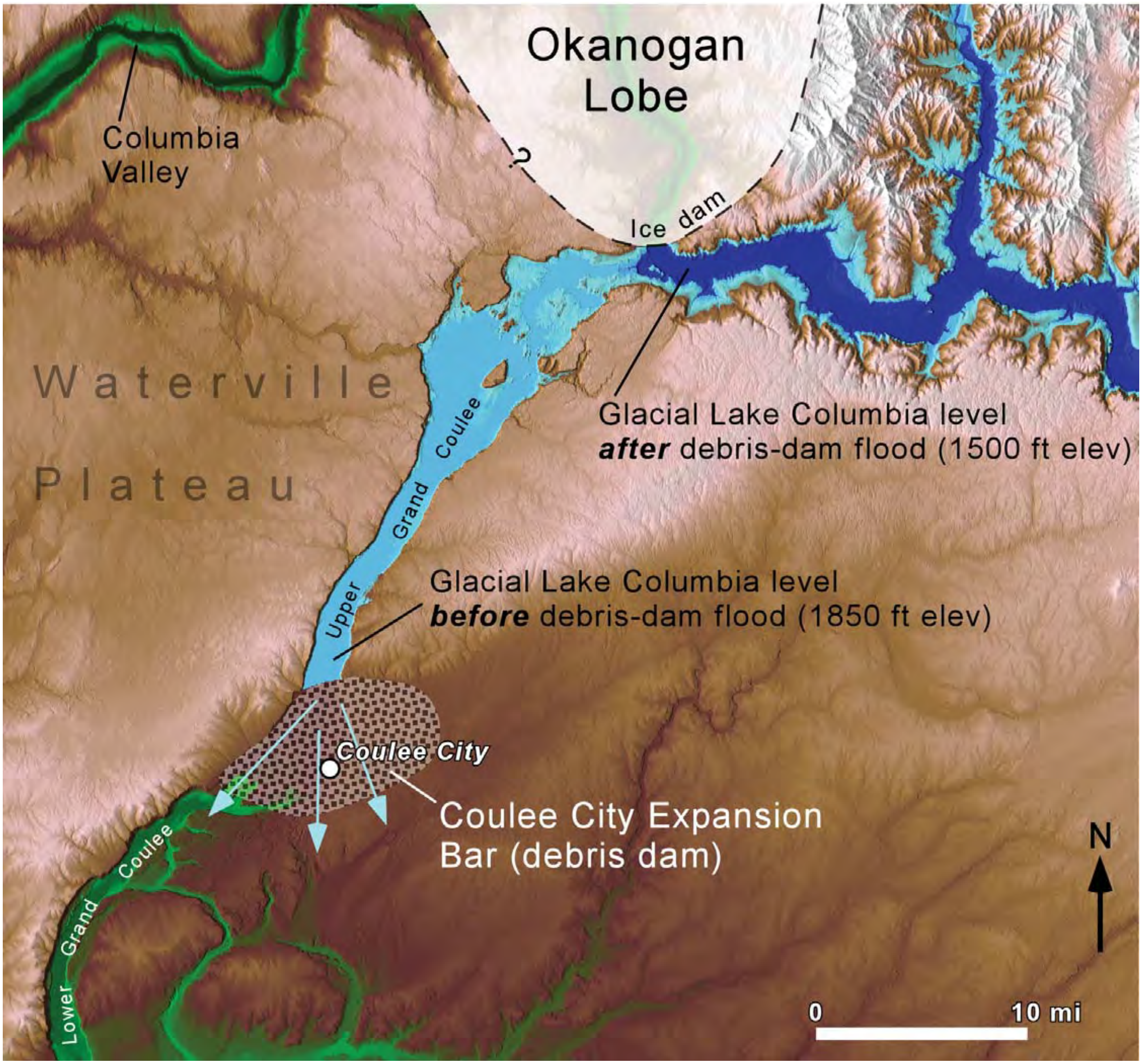


Figure 13. One of the last floods occurred when Glacial Lake Columbia overtopped a debris dam at the southern end of the Upper Grand Coulee. The ensuing flood drained south into the Lower Grand Coulee, lowering the level of Lake Columbia by more than 300 feet.

sion that occurred as Dry Falls and the Great Cataract Group receded into the area. In either case, the floods would have incised into and removed the loose bar deposits, leaving behind erosional remnants of the bar along the sides of the coulee where flow was slower and erosion was less intense. A number of terraces etched into the sides of the bar (fig. 14) indicate that the floods that came after bar incision were successively smaller.

At Hartline turn north onto a series of right-angle turns that cross the Coulee Monocline. You will eventually want to follow Old Coulee Road north back to Grand Coulee.

Figure 14. A series of subtle terraces (parallel dashed lines at right) lie at the head of the Coulee City Expansion Bar where the Upper Grand Coulee expands into the Hartline Basin. These record a series of successively smaller floods that coursed through the coulee late in the last glacial cycle. Note the steeply dipping basalt flows on the left that has been folded into hogbacks along the Coulee Monocline.

173 **STOP 11** **Stop 11. Northrup – Spring Canyon Spillover Complex (fig. 15)**

At Grand Coulee head northeast on SR 155, crossing the Columbia River at Coulee Dam. Continue past Elmer City before heading east onto Peter Dan Road and then Manila Creek Road, passing through the Colville Reservation. Turn north onto SR 21.

248 Return to Republic. End of field trip.

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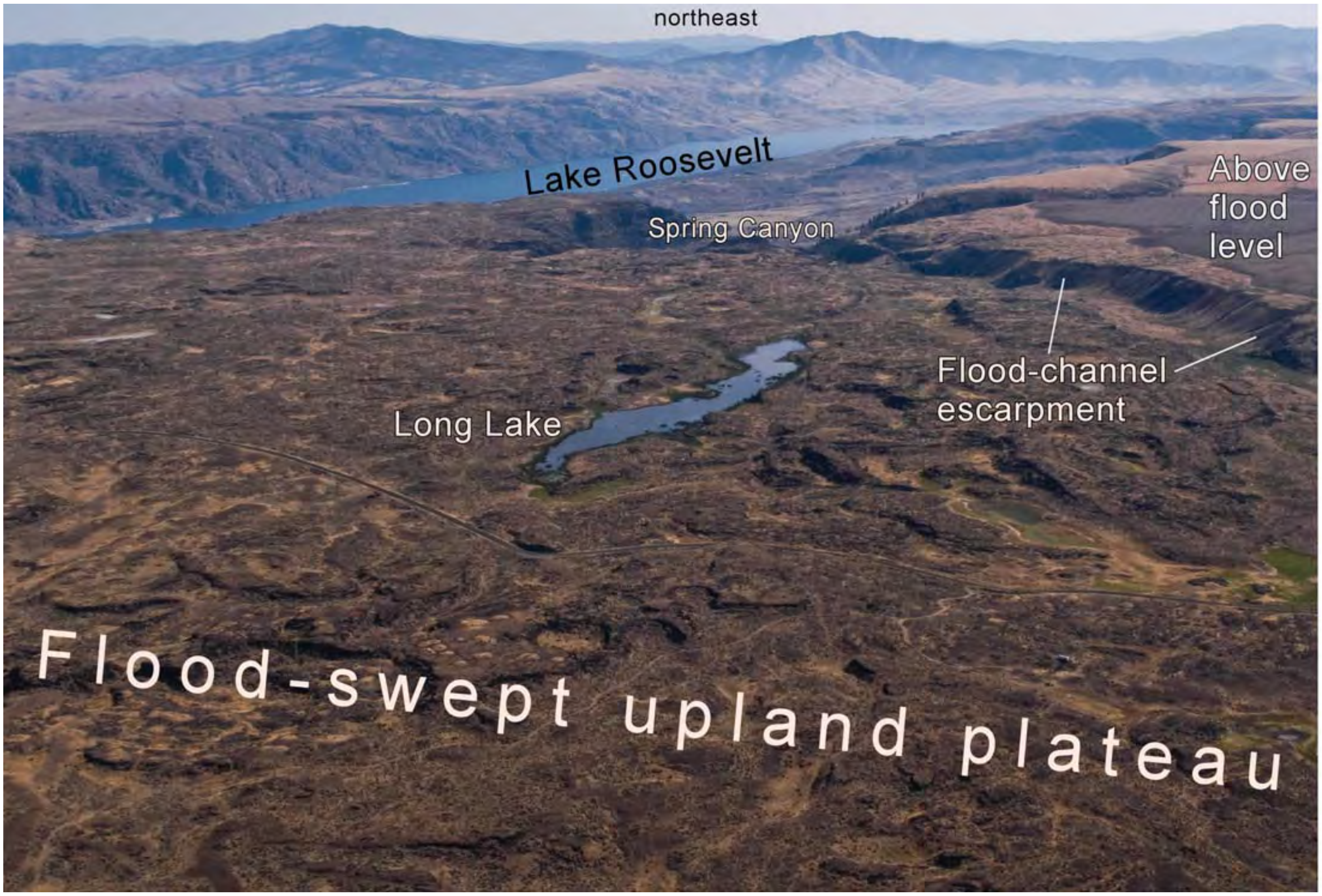

Figure 15. Looking northeast. Flood spillover area in uplands between Spring Canyon and Northrup Canyon. Long Lake fills one of the many shallow rock basins excavated by high-energy currents that swept across this upland plateau. The maximum elevation of the floods here was about 2,500 feet, based on the elevation of the flood-beveled escarpment at upper right. These spillover areas into Northrup Canyon probably developed prior to breaching at the head of Grand Coulee. Right: Looking north. Highest observed ice-rafted erratic (see lens cap) along upland plateau behind Northrup Canyon (N47.8618, W118.9966). This angular granodiorite boulder may have been moved slightly from the adjacent wheat field.

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EXPOSURES OF LARGE CLASTIC DIKES IN COLUMBIA BASIN
A GEOLOGIC TRAVERSE THROUGH WASHINGTON, OREGON, AND IDAHO

Skye Cooley

Cirque Geoscience & GIS Consultants
Adjunct, Boise State University Geosciences Department
Boise, Idaho

GEOLOGY OVERVIEW

The Glacial Lake Missoula Floods left behind slackwater and flood-channel sediments that contain vertically-sheeted clastic dikes. The origin of the dikes has been the subject of debate for decades. Several explanations have been proposed, including:

a) desiccation
b) ground ice (Black, 1979)
c) liquefaction (Jenkins, 1925)
d) hydrofracture (Pogue, 1998)
e) flood loading (Baker, 1973)
f) multi-genetic (Lupher, 1944)

Most clastic dikes described around the world are “seismites”, features of soft sediment deformation resulting from cyclic shaking, pore pressure elevation, and fluid escape. The vast majority of clastic dikes described in the literature originates at depth from a liquefied layer and terminate upwards in younger sediments.

The clastic dikes in the Columbia Basin are peculiar. The dikes originate in unconsolidated outburst flood deposits, but are nowhere present above the local maximum flood stage. The features are found from Lewiston, Idaho, to Foster Coulee, Washington, to Portland, Oregon. They were forcefully injected from the top down, taper downward, and have fluted silt walls that record the downward filling. They coalesce to form polygons in map view and measure from 1 cm to over 3.5 m in width. Some of the silt-sand dikes cut indurated post-Columbia River Basalt gravels (Neogene age basin fill units), interbeds in the basalt, and the Miocene basalt itself.

Two ages of Pleistocene dikes are recognized. The younger set, numerous and well exposed in the region, formed during Missoula flooding (~21-12 ka). The older set is partially lithified and of Early and Middle Pleistocene age (~780-35 ka), or “pre-late Wisconsin”. Anomalous downward-intruding clastic dikes of similar size have been described elsewhere, though the

Outcrops with Large Clastic Dikes: Columbia, Yakima, Walla Walla, Umatilla & Lewiston Basins

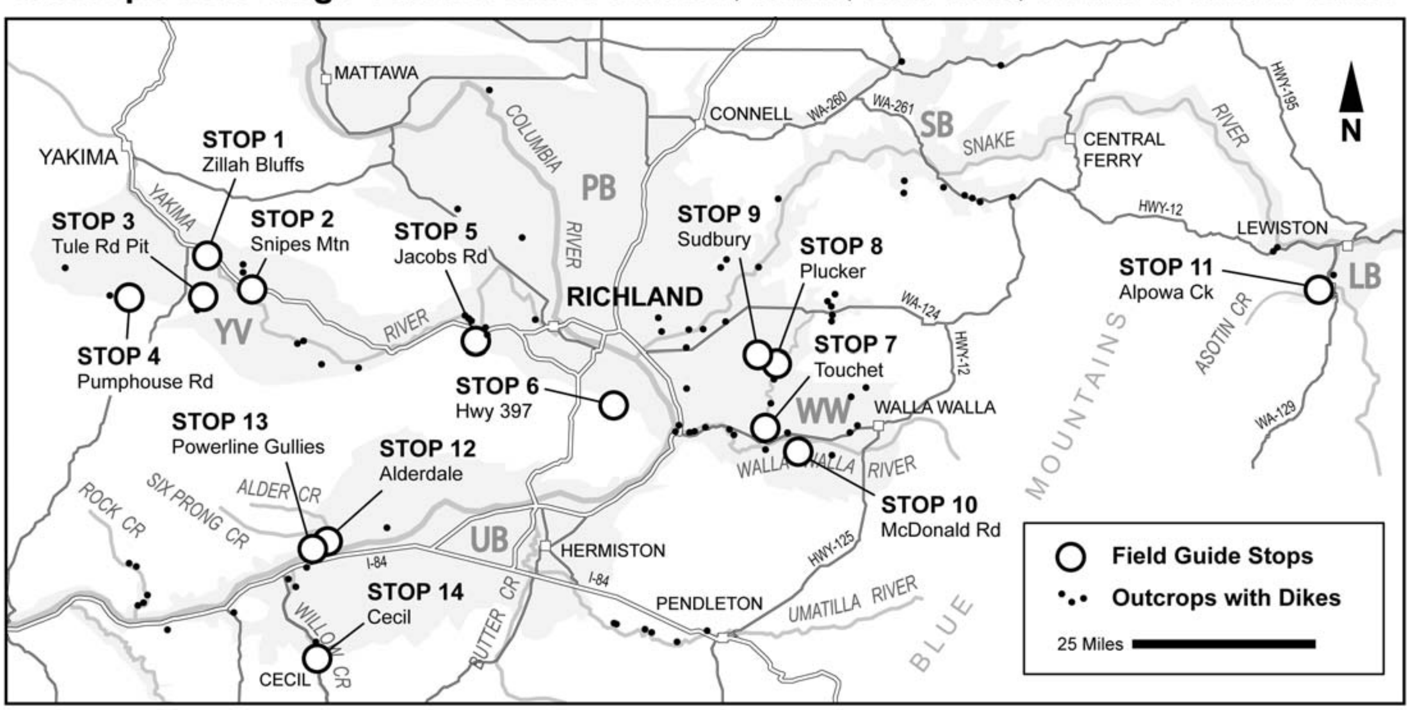


Photo 1. A typical "Touchet type" clastic dike. Multiple fill bands (about 10 in this dike) are composed of unconsolidated sand and silt derived from Missoula flood slackwater deposits.

Photo 2. Silt walls of the dikes exhibit flute casts with upward-pointing noses, a clear indication that sediment entered from the top.

record is sparse. Dikes originating in fluvial gravels penetrate the underlying Bishop Tuff (Lipshie, 1976); dikes sourced in thick debris flows penetrate the underlying sandstone in Black Dragon Canyon, Utah, and some dikes penetrate various glacial deposits beneath the toes of glaciers (Goldthwait, 1951; Larson and Mangerud, 1992; Dreimanis and Rappol, 1997; Van der Meer et al., 2009).

This field guide describes several excellent exposures of large, “Touchet-type” clastic dikes, where dike width measurements and counts of fill bands have been made by the author. To date, I have collected data on some 1,400 dikes from 200 sites and 13 different geologic units. These constitute one of the world’s largest and most complete records of paleoseismic(?) features in a single region. The data will be used to test whether seismicity or flood loading was responsible for producing the dikes. A future paper will compare maximum dike widths and number of fill bands to seismic energy attenuation curves (compiled in Galli et al., 2000) , which control the distribution of soft sediment deformation features versus local variations in Missoula flood-water depth, flood velocity from data by Denlinger and O’Connell (2010), and subbasin hypsometry from a GIS.

ROAD GUIDE OVERVIEW

Many guidebooks have been published on the Missoula flood geology with stops at the dozen or so classic outcrops and overlooks. This road guide seeks to introduce the reader to new outcrops and different areas along the flood route, but decidedly off the beaten path. Many of these stops are seldom, if ever, visited. The Columbia Basin is a vast region, thus the stops in this guide are farther apart than they may appear to be on the map. Allow three days to visit all the stops. GPS coordinates are provided in UTM NAD83 Zone 10N and Zone 11N. Coordinates will match Google Earth (WGS84).

Author's Note on Field Work in Columbia Basin

For two weeks each year over the past decade, I have traversed the Columbia Basin in search of dike-bearing outcrops. It is solitary, fundamental geology fieldwork. I have recorded the widths of several thousand dikes at hundreds of outcrops throughout the region. The dikes cut 13 different geologic units and are clearly the product of a regional-scale perturbation of one form or another. My goal is a better understanding of the features' origin. I seek to reach the goal by way of a comprehensive, regional assessment of size, spatial distribution, relative age, stratigraphic contexts, and cross-cutting relationships. My long days afoot have often been punctuated by new discoveries. The Basin is beautiful, vast, and unpretentious. The blistering heat of noonday yields to coyote-dark nights, which open to sweet, green desert mornings. Fieldwork here, amidst these northern canyons, is joyful work.

ROAD GUIDE

STOP 1 Stop 1: Zillah Bluffs Hike

GPS: 711570 mN, 5141351 mE

This is a stop for the intrepid, sure-footed geologist who does not mind getting her boots wet. A small group is appropriate here. Trenching tools, hand lenses, and helmets are recommended.

From the east on I-82 in Yakima Valley, take Exit 54 at Zillah. Turn left on the Yakima Valley Highway and cross the bridge over the freeway. Turn right on Railroad Avenue E and drive to its end. Park at the turnaround at the end of the road near the railroad bridge over I-82. Hike northwest to the twin railway bridges over the freeway. Cross the tracks and continue on towards a wire fence at left. Follow the fenceline down the hill towards the freeway. Cross the fence when a sandy outcrop of Touchet Beds becomes visible to your left. Make your way through trees and sloping sand along the base of the bluff for approximately ¼ mile. You will be just above the river level. Return to cars via the same route.

Numerous large clastic dikes (over 60 cm wide), about 15 rhythmites, numerous bedforms, the Mt. St. Helens Set S couplet, and a berg mound deposit (granite clasts), are well exposed in the bluffs.

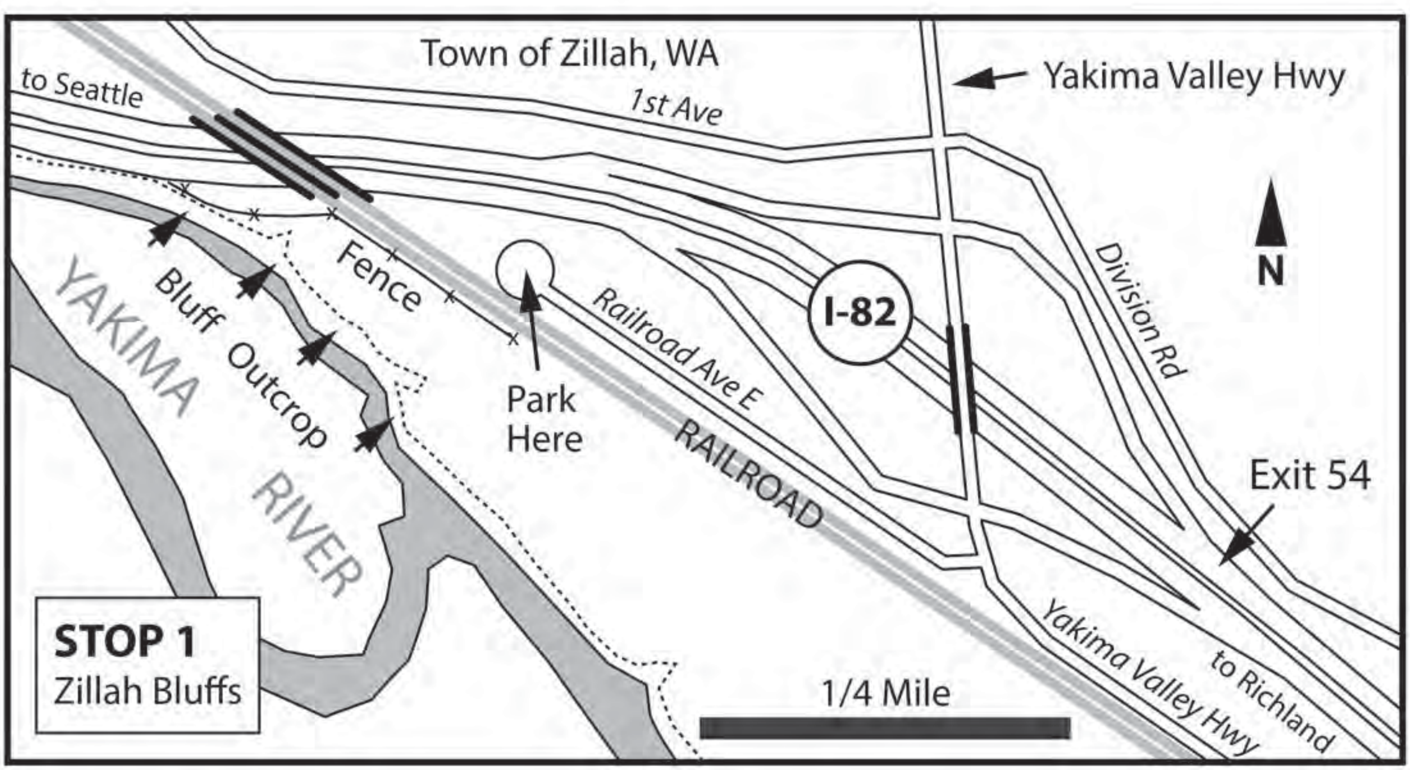


Photo 3. Stop 1. Expect difficult footing and brush along the base of Zillah Bluffs.

STOP 2 Stop 2: Snipes Mountain / Emerald Road

GPS: 718500 mE, 5134140 mN

From the east on I-82, take exit 58 to Granger via Hwy 223. Turn left under the freeway (southwest) on Hwy 223/Van Belle Road. Continue just past the town of Granger but not over the Yakima River. Look for the left turn onto Emerald Road at RV sign. Turn right on Emerald Road past a large gravel quarry. Continue atop the bluff overlooking the Yakima River floodplain to GPS location. Park in one of the informal pullouts, roadcuts, and overhead power lines.

Snipes Mountain is a fault-bounded block of basalt capped by oxidized, quartzite-dominated Snipes Mountain Conglomerate and Late Pleistocene Missoula flood slackwater beds (Touchet Beds). The conglomerates are deposits of the ancestral Columbia River deposited on the basalts.

Several large sand-, silt-, and gravel-filled clastic dikes intrude both the flood rhythmites and the underlying Miocene-Pliocene conglomerate. Large downward-intruded dikes are visible from embankments at the edge of the large, inactive gravel quarry, located one mile to the west. Also, there are excellent exposures of dikes and deformed host sediments in the roadcuts located along the private driveways between Emerald Road and the river. Large petrified logs bearing opal are present in-situ. Ask local ranchers and orchard owners for access.

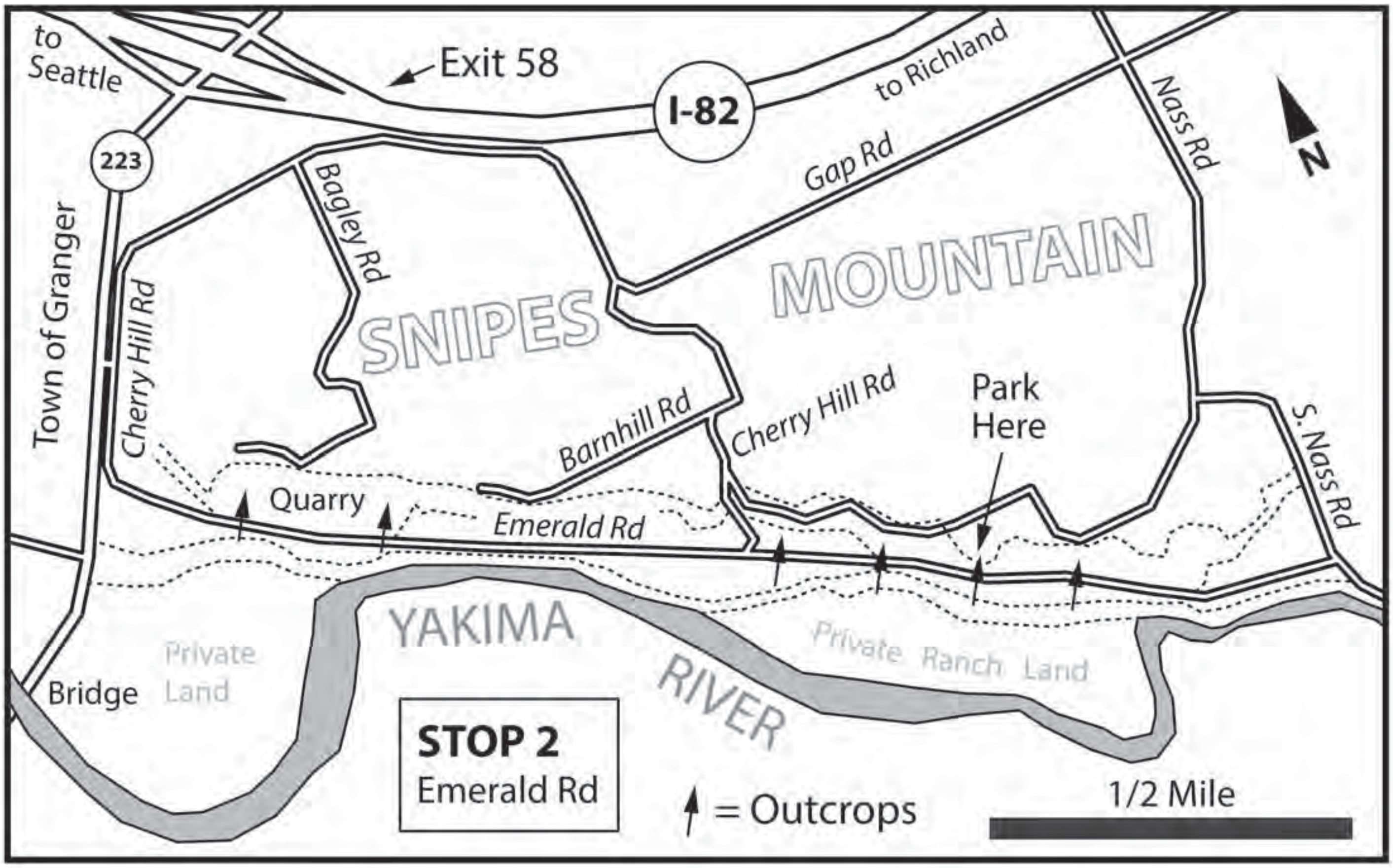
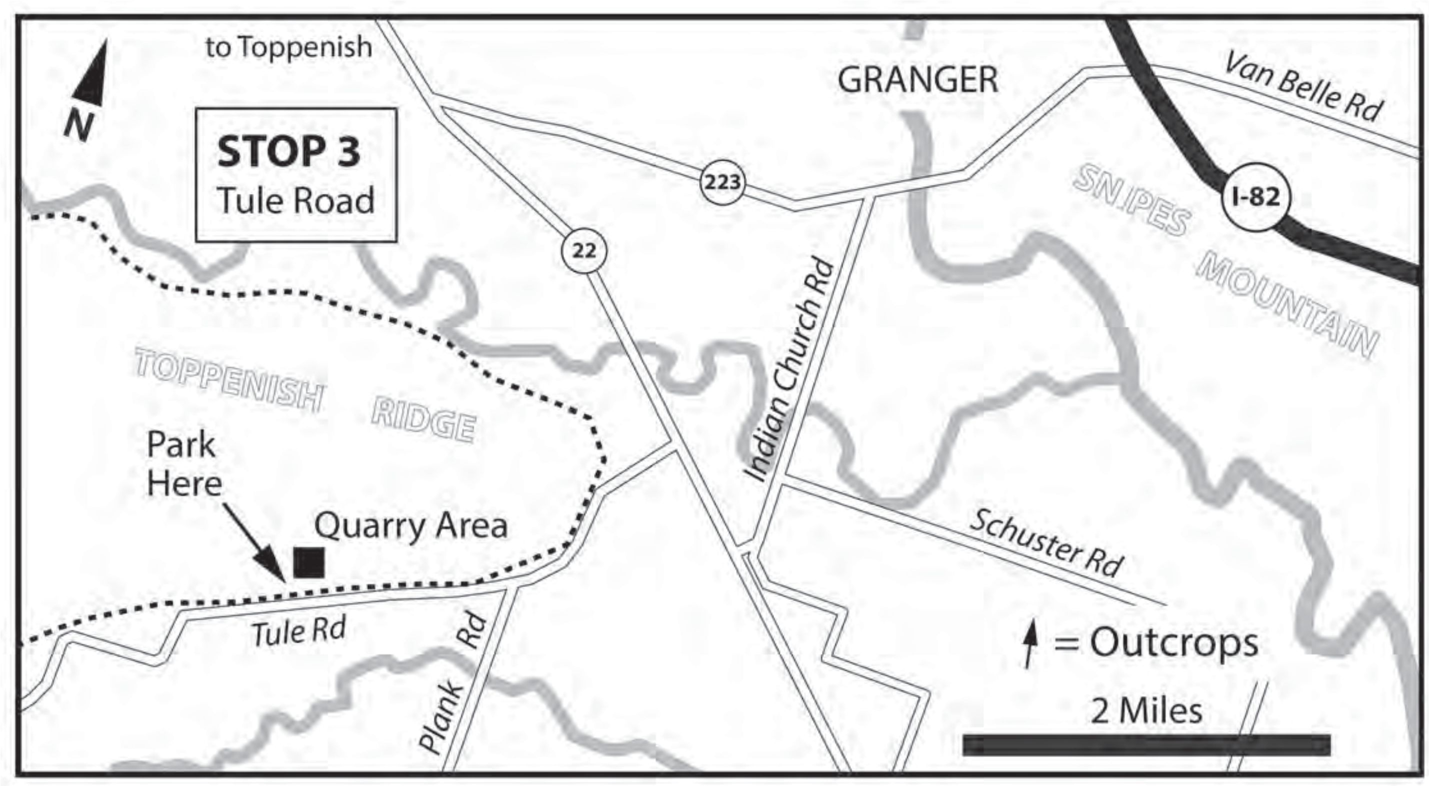


Photo 4. Stop 2. Gravel-filled dikes as well as the more typical silt-sand dikes intrude the Snipes Mountain Conglomerate. Dike shown cuts through more than 10m of section.



STOP 3 Stop 3: Tule Road Gravel Pit

GPS: 712700 mE, 5130070 mN

Drive southwest from Granger on Hwy 223 to the junction with Hwy 22/Wappenish Road. Go left (SE) on Hwy 22 to Plank Road. Turn right on Plank Road, then right on Tule Road. Continue one mile to gravel pit area. This is Yakama Indian Reservation Land, and you must ask permission for access at the Tribal Headquarters in Yakima. DO NOT COLLECT ANYTHING!

Cut faces within 200 m of Tule Road expose large, vertical clastic dikes intruding several meters into the Snipes Mountain Conglomerate. Explore the nearby cuts for a better look at the features.

Photo 5. Stop 3. Light colored Missoula flood rhythmites overlie petrified-wood bearing Snipes Mountain Conglomerate at Tule Road gravel pit. Several large silt-sand dikes cut downward through the gravels.

STOP 4 Stop 4: Pumphouse Road

GPS: 702921 mE, 5130647 mN

This is a lonely outcrop exposing an unusually large dike for this part of the Yakima Valley. Drive about five miles south on Hwy 97 from downtown Toppenish to Toppenish Ridge. At the base of the long incline up the ridge, turn right on Pumphouse Road, leading west. Cross the cattle guard and bypass the Toppenish National Wildlife Refuge entrance. Drive 1.9 miles on Pumphouse Road to a primitive road to the left (South) located a couple hundred feet BEFORE BIA Road #96. The road you want does not lead to the house. It traverses along a wash. Drive or walk 1/3 mile from the pavement to an outcrop on the left.

STOP 5 Stop 5: Jacobs Road Cul-de-Sac

GPS: 308624 mE, 5124932 mN

A very large dike cuts flood and hillslope deposits in roadcuts of the cul-de-sac. Cuts along Weber Canyon Road and cutbanks off Demoss Road also expose sizeable dikes in Touchet Beds.

STOP 6 Stop 6: HWY 397 / Locust Grove Road

GPS: 339368 mE, 5110645 mN

From the junction of I-182 and Hwy 395 at Pasco, drive south across the Columbia River on Hwy 395. Continue until Hwy 395 merges into I-82 South. Exit the freeway shortly at Exit 114 towards Finley via Hwy 397. Drive 6.5 miles east to the GPS location.

Lithified clastic dikes of pre-Late Wisconsin age are present here. The dikes stand out in relief from the cut face in places. Some have tops truncated by a Missoula flood-age unconformity. The dikes cut a white shaley interbed of the Latah Formation in the Columbia River Basalt. Younger dikes are also present here, originating in the Touchet Beds. This is a great place to explore for the sure of foot.

STOP 7 Stop 7: Touchet River Road Double Cut

GPS: 371725 mE, 5104004 mN

From the gas station in Touchet, Washington (located at the intersection of Hwy 12 and Touchet River Road), drive north on Touchet River Road for 2.7 miles to a large double-sided roadcut. Park in pullouts at either end of the cuts. Walk the roadside cut faces.

Photo 6. Stop 7. Road cuts along Touchet River Road expose a wide variety of internal sedimentologic characteristics and cross cutting relationships between the dikes and the host sediment, ash layers, and small faults.

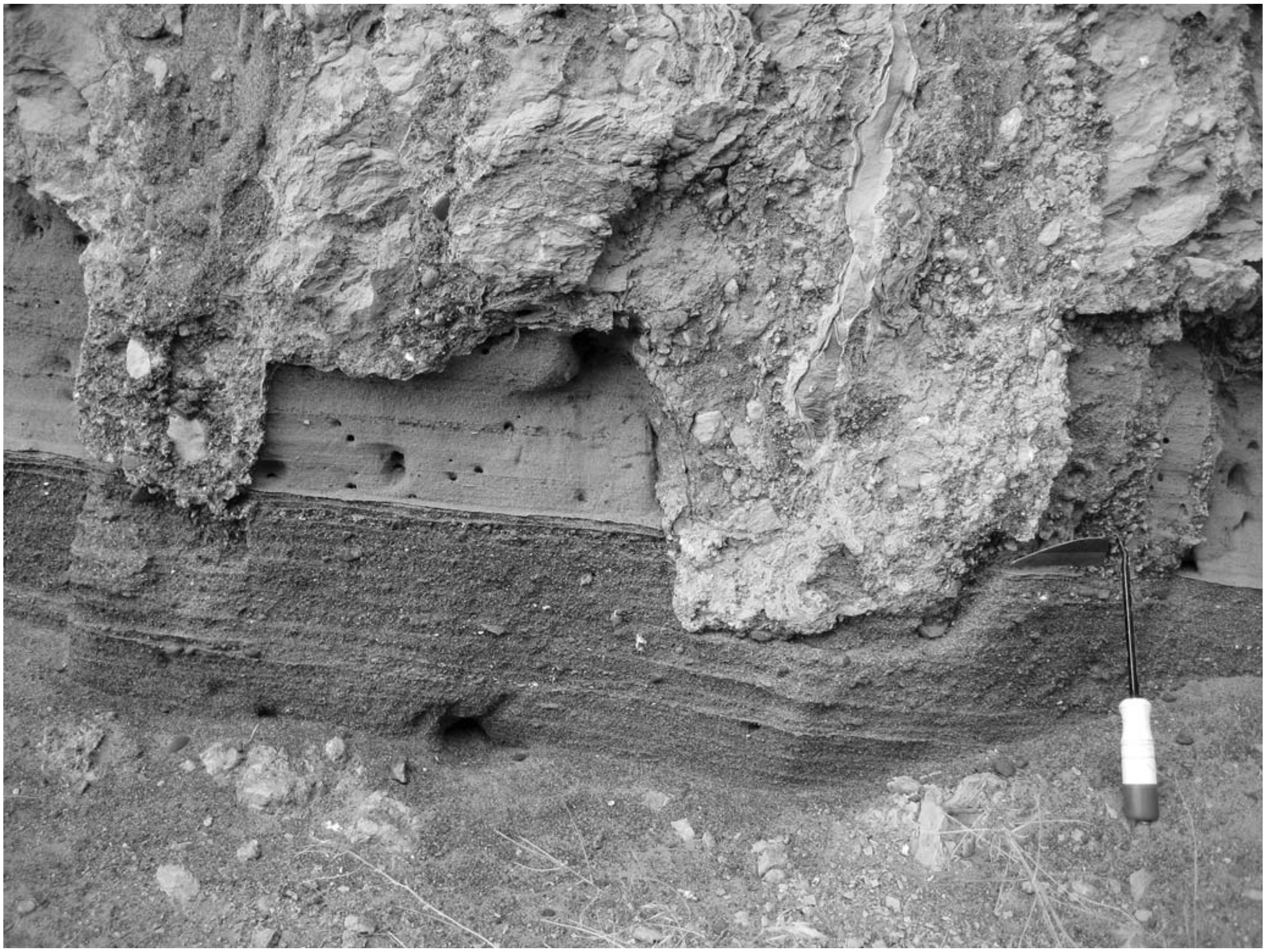


Photo 7. Stop 7. The type section for the slackwater Touchet Beds is located about 5 miles south of Stop 7, in the bluffs of the Walla Walla River. The bluffs are difficult to access. This photo, from the bluffs, captures what appears to be two dikes caught in the act of intruding downward through the rhythmite section.

If you have only the time to visit one outcrop for clastic dikes, this is it. The large, vertical road cut through Touchet Beds exposes a continuous double-sided section for about ¼ mile. This is a hands-on exposure where the full variety of internal sedimentary features and crosscutting relationships between dikes and faults can be observed. Watch for cars and wine tourists.

If you have only the time to visit one outcrop for clastic dikes, this is it. The large, vertical road cut through Touchet Beds exposes a continuous double-sided section for about ¼ mile. This is a hands-on exposure where the full variety of internal sedimentary features and crosscutting relationships between dikes and faults can be observed. Watch for cars and wine tourists.

STOP 8 Stop 8: Plucker Historical Sign

GPS: 373404mE, 5116641 mN

From Walla Walla, drive 17.5 miles west on Hwy 12 to Touchet. Turn north on Touchet River Road and continue 13.5 miles to the Lewis and Clark Historical marker/pullout. The road is just downhill. Watch for traffic.

The story here is shown by an older set of clastic dikes, features of pre-late Wisconsin age that have been modified by soil forming processes, with conspicuously truncated tops. The outcrop exposes at least a couple of generations of clastic dikes, each cut by unconformities within pebbly, fossiliferous strata from the pre-Late Wisconsin floods. The younger set is here too. Unconsolidated sand-silt dikes originate near the top of the exposure and cut both late and early-middle Pleistocene rhythmites.

Photo 8. Stop 7. Huge clastic dikes, some containing dozens of fill bands, are exposed in roadcuts for several miles along the Touchet River Road.

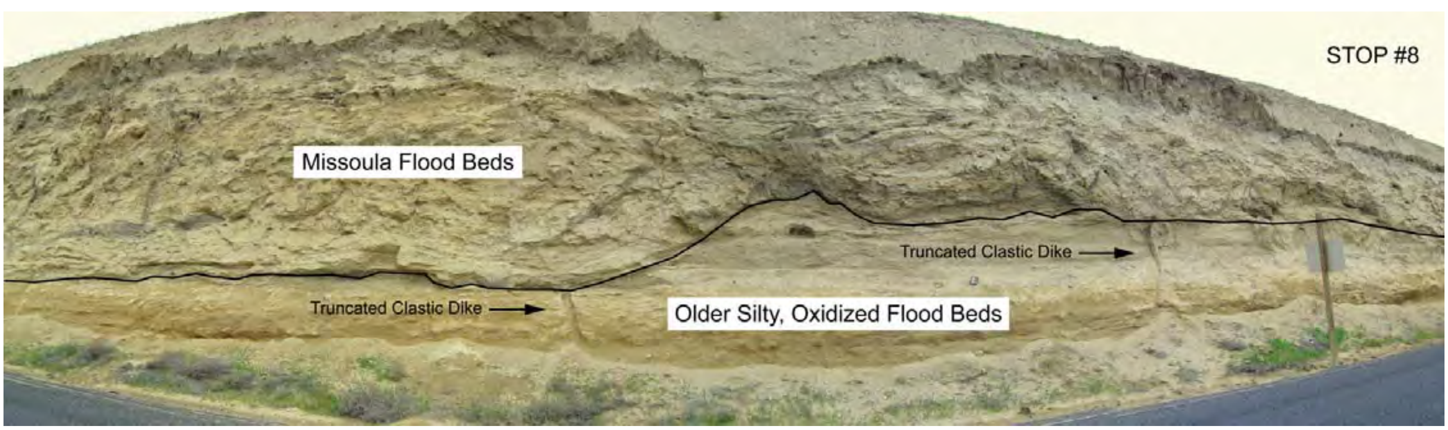


Photo 9. Stop 8. Missoula flood beds overlie silty, oxidized beds deposited in an earlier glacial flood episode. The older unit is not fully understood in large part due to poor exposure in the region. Much of these so-called “pre-late Wisconsin” deposits have been eroded by the late Pleistocene floods. Sheeted clastic dikes, nearly identical to those originating in Missoula flood deposits, are present in the older beds. They are easily recognized by higher clay content, pedogenic alteration (iron staining), and tops that are truncated by the younger flooding surface (black line). Whitman College professor, Patrick Spencer, has done much to forward understanding on the pre-late Wisconsin deposits in Walla Walla Valley and Pasco Basin.

Both dikes sets share in common a vertically sheeted character (multiple fill bands). This seems to argue for a common origin. Both sets of dikes formed during a glacial-interglacial transitional period (Alwin, 1970; McDonald and Busacca, 1988; Spencer and Jaffee, 2002) and may be related to cyclic outburst flooding.

Pre-late Wisconsin deposits ("Pre-Missoula gravels" of Richman, 1981) include at least three unconformity-bound flood sequences that contain layers of loess, tephra, and paleosols. The units disconformably overlie Tertiary units (Ringold Formation, CRBG). The unit's age is about 32 to over 790 ka (McDonald and Busacca, 1988). The sediment varies with location, but generally consists of well-rounded and sorted pebble- to cobble-size gravels, exotic-bearing diamict, quartz sandstone, Stage I-IV caliche layers, and sand-silt-clay rhythmites (Patton and Baker, 1978; Bjornstad, 1980; McDonald and Busacca, 1988; Vrooman and Spencer, 1990; Spencer and Gilk, 1999; Spencer and Jaffee, 2000; Bjornstad et al., 2001; Spencer and Jaffee, 2002).

Unconformities exposed in the Walla Walla valley (Spencer and Jaffee, 2002) appear to correlate with erosional surfaces in Illinoian-age loess (> 780-35 ka) on the Palouse Slope (McDonald and Busacca, 1988) and in gravel bars in the Pasco basin (Patton and Baker, 1978; Bjornstad et al., 2001). The ultimate origin of the deposits remains murky, though they may record outburst floods, possibly from glacial lakes in southern British Columbia (Shaw et al., 1999; Kovanen and Slaymaker, 2004; Lesemann and Brennand, 2009; Sweeney et al., 2011).

STOP 9 Stop 9: Sudbury Double Cut

GPS: 383970 mE, 5113450 mN

From Walla Walla, drive west on Hwy 12 to the Sudbury Road exit. Continue 9.8 miles on Sudbury Road to a large, double-sided road cut. Park downhill a few hundred feet in paved pullouts. The smaller roadcuts near pullouts are of interest as well.

Whitman College geology professor, Patrick Spencer, introduced me to this outcrop. His work on pre-late Wisconsin deposits is reshaping our understanding of early-middle Pleistocene sedimentation in the Columbia Basin (Spencer and Jaffee, 2002).

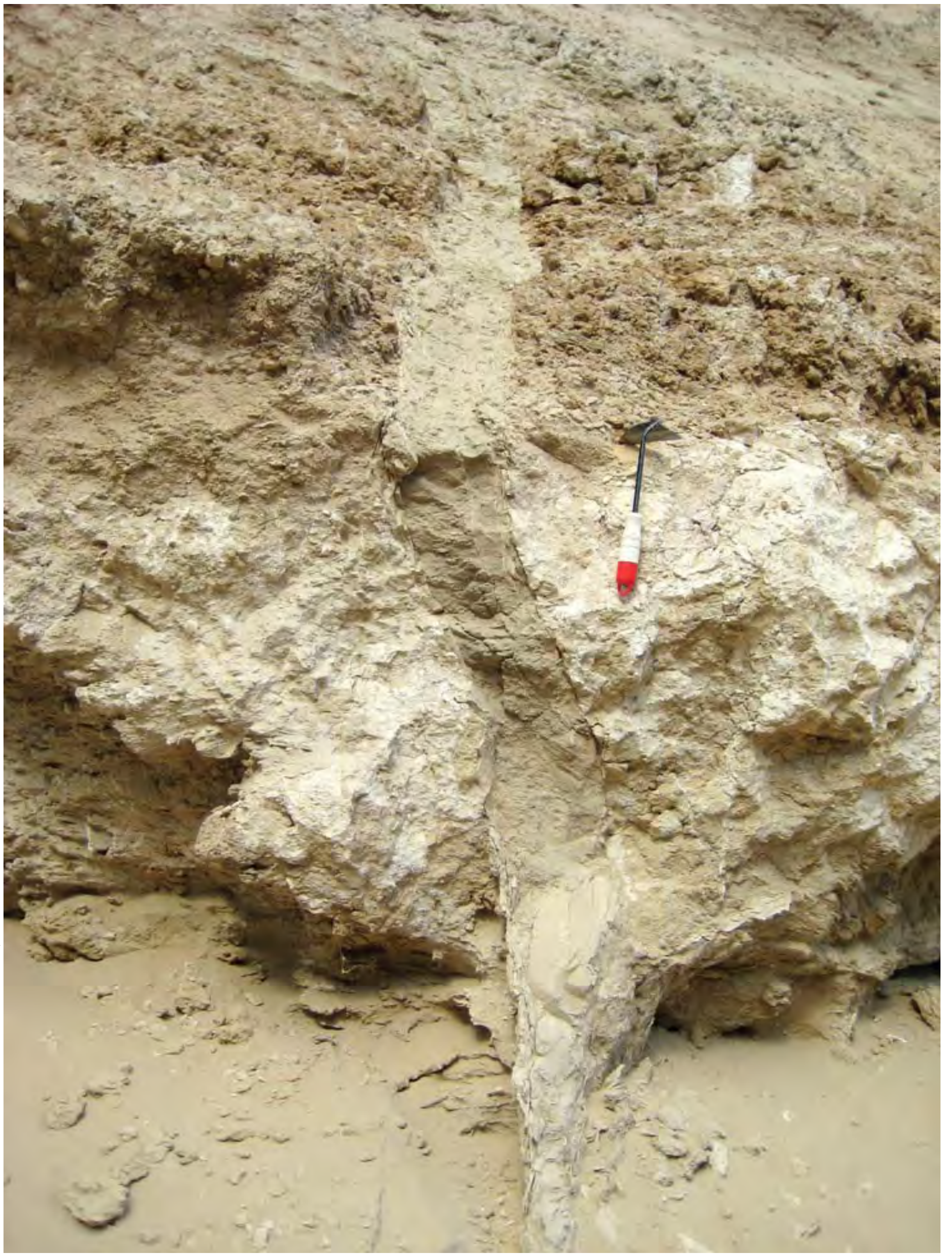


Photo 10. Stop 9. Several clastic dikes cut through layers of Pleistocene loess and paleosols developed in them. Here a dike cuts a meter-thick layer ofwell indurated nodular caliche. Dike bearing units at the site contain mammal fossils.

Despite how high this hillcrest location feels, it is still below the maximum flood stage elevation of the Missoula floods. Several oxidized loess layers are well exposed here, some with thick caliche calcrete layers and/or nodules, which attests to long periods of aridity and surface stability. Loess beds are separated by gentle angular unconformities. Mammal and rodent bones are present here, too. Offset in a small thrust fault in the basalt near the pulloff is shown particularly well in a thin, water-lain gravelly unit that contains exotic clasts overlying the basalt. Thin, white ash layers appear fuzzy in the silty cut face. Oxidized, truncated clastic dikes are present as well as the more typical gray-tan, unconsolidated silt-sand dikes ("Touchet-type" dikes). A long, vertical dike cuts the large face, its top and bottom terminations obscured. It appears to originate from below. Other dikes clearly pinch downward.

STOP 10

Stop 10: McDonald Road

GPS: 379678 mE, 5099136 mN

Leave Hwy 12 about one mile east of Lowden (between Touchet and Walla Walla), heading south on McDonald Road. Drive 1.2 miles, crossing the river and Detour Road, to a tidy road cut located at the rise in the road. Pull off onto the shoulder.

A large clastic dike is exposed here. Large dikes such as this one are numerous in the cutbanks of the Walla Walla River nearby (Touchet Bed type section), but access is easy here, not so there.

STOP 11

Stop 11: Alpowa Creek Wildlife Area

GPS: 483527 mE, 5139884 mN

From Lewiston, Idaho, drive 7.5 miles west on Hwy 12/Inland Empire Hwy to the old trestle bridge over Alpowa Creek. Turn right onto Westlake Drive and immediately right into the Wildlife Area parking lot. Hike about 1,000 feet along the base of the cliff rimming the reservoir (parallel to, but below Westlake

Photo 11. Stop 11. A very large Touchet type clastic dike is exposed in a bluff along below Westlake Drive. This location is 450 km from identical features in Willamette Valley, OR. Ruler in the the photo is in centimeters.

Road) to find outcrops of Touchet Beds and gravels over basalt. A few large clastic dikes are hidden amongst the trees. Return to car via the same route.

STOP 12

Stop 12: Alderdale Road

GPS: 273355 mE, 5081114 mN (UTM Zone 10)

Drive west on Hwy 14 from the junction with I-82 and the bridge over the Columbia to Umatilla. After 31 miles, look for the Alderdale Road leading north. Continue up the canyon for one mile. Look for a wide spot to pull off in the vicinity. Sandy bluffs rising from the road's shoulder are of interest.

One of the largest clastic dikes in the Columbia Basin is exposed here; it is approximately 120 cm wide. The bluffs nearby also contain a few dikes. Cut banks along nearby Sixprong Creek also contain a wealth of dikes. However, the creek has been closed to visitors by a local rancher who has successfully lobbied to have Sixprong Road abandoned by Klickitat County.

STOP 13

Stop 13: Powerline Gullies Hike / Alder Ridge

GPS: 270894 mE, 5079717 mN

Drive West on Hwy 14 from the junction with I-82 and the bridge over the Columbia. Near milepost 148, two miles west of Alderdale Road, locate the gully-mouth trash rack across from a little sign "160" near the railroad tracks. Pull off onto the wide shoulder.

This hike is adventurous, if of only modest mileage. The gully system is comprised of three branches. Each deepens and widens as you climb.

Skirt the trash rack structure and washed out fencing to gain good views of clastic dikes up to 40-cm wide. These dikes cut basaltic colluvium, debris flow

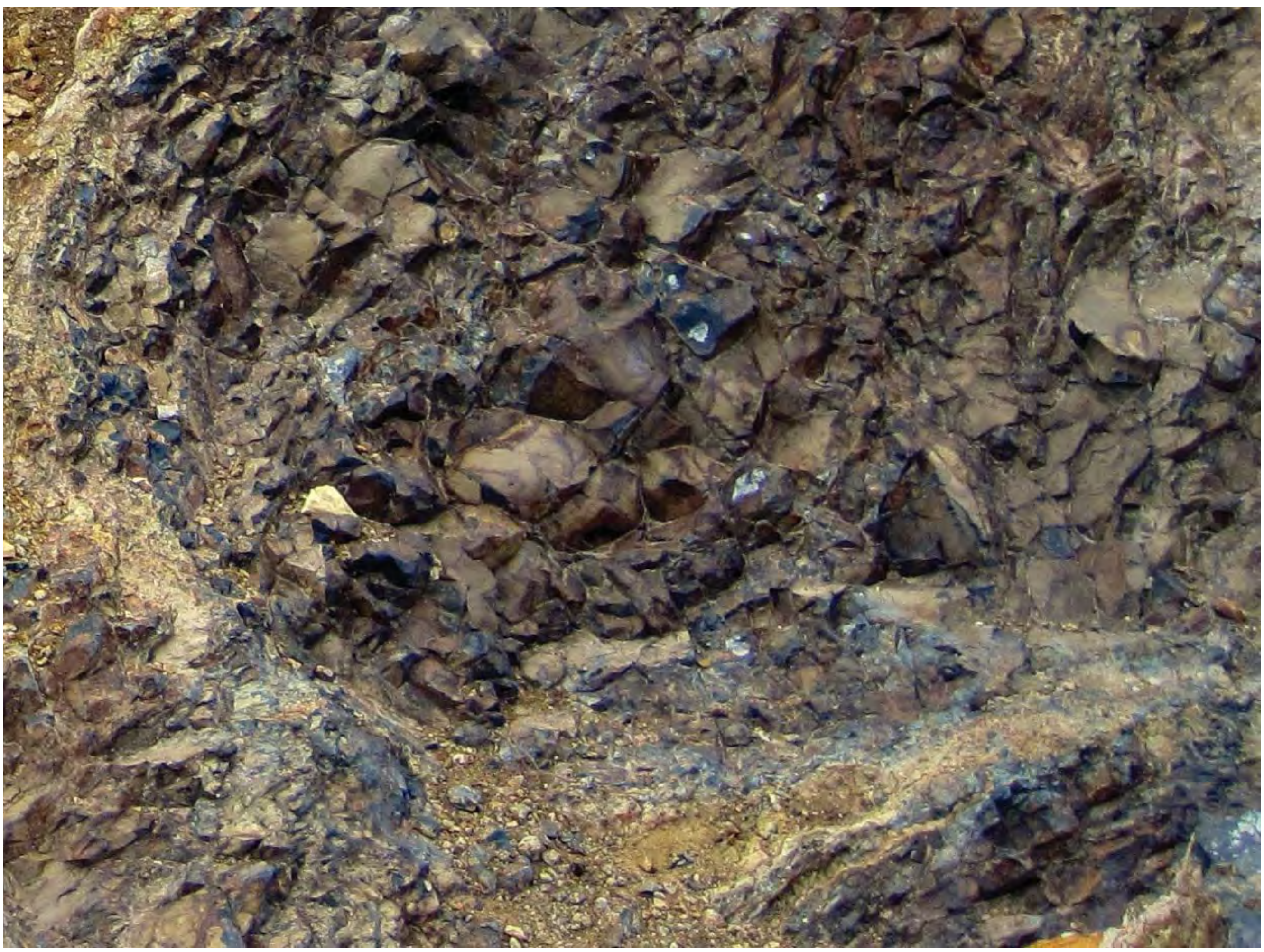


Photo 12. Stop 13. A silt-sand dike, originating in overlying Missoula flood sediment, intrudes Miocene pillow basalt in a gully along Hwy 14 at Alder Ridge. Similar dikes, exposed in the walls of the gullies, cut basaltic colluvium (Pleistocene), thick caliche paleosols developed in loess (Pleistocene), and Dalles Group sediments (Neogene).

deposits, nodular caliche developed in loess (paleosol), Missoula flood gravel beds and rhythmites, green/white Dalles Group sediments, pillow basalts, and white fault gouge. A fault cuts the basalt, forming Alder Ridge, near the gully mouth. Large chunks of opal are found both in-situ and as float in many places. Some of the dikes are offset along small faults. Others show strong parallel orientations. Lenses of Mazama ash containing backfilled rodent and insect burrows overlie all of the dike-cut units. The ash is exposed at the upper end of the gullies, near the power line towers. Towers #4 and #5 are located about 1/2 mile from the road (it feels longer). The latest major debris flow event was during the winter of 2013.

Another hikeable gully is located 0.7 miles west of milepost 148.

Photo 13. Stop 14. The slackwater rhythmites at Cecil formed during the filling of transient Lake Condon, due to a bedrock constriction downstream of Wallula Gap. These Touchet Bed-equivalent units are well exposed along the quiet road through Willow Creek Valley (Hwy 74).

STOP 14

Stop 14: Cecil / Willow Creek Valley

GPS: 269460 mE, 5056080

From Umatilla, drive 19 miles west to Boardman. Continue west on I-82 to Exit 147 to Ione and Heppner via Hwy 74. Drive 13.7 miles on Hwy 74 to Cecil, Oregon.

Missoula floodwaters backed far into the Willow Creek valley. About 15-20 slackwater Touchet Beds are exposed at Cecil. Sheeted clastic dikes identical to those in Touchet River (60 miles away), at Alpowa Creek (140 miles away), and Emerald Road (50 miles away) intrude the section. There about a dozen sizeable roadcuts in slackwater rhythmites along this stretch of Hwy 74. Dikes intrude green-yellow-gray, fractured mudstones and sandstones of the Latah Formation (possibly Dalles Group) in a road cut near red barns and new wind turbines (269680 mE, 5059711 mN).

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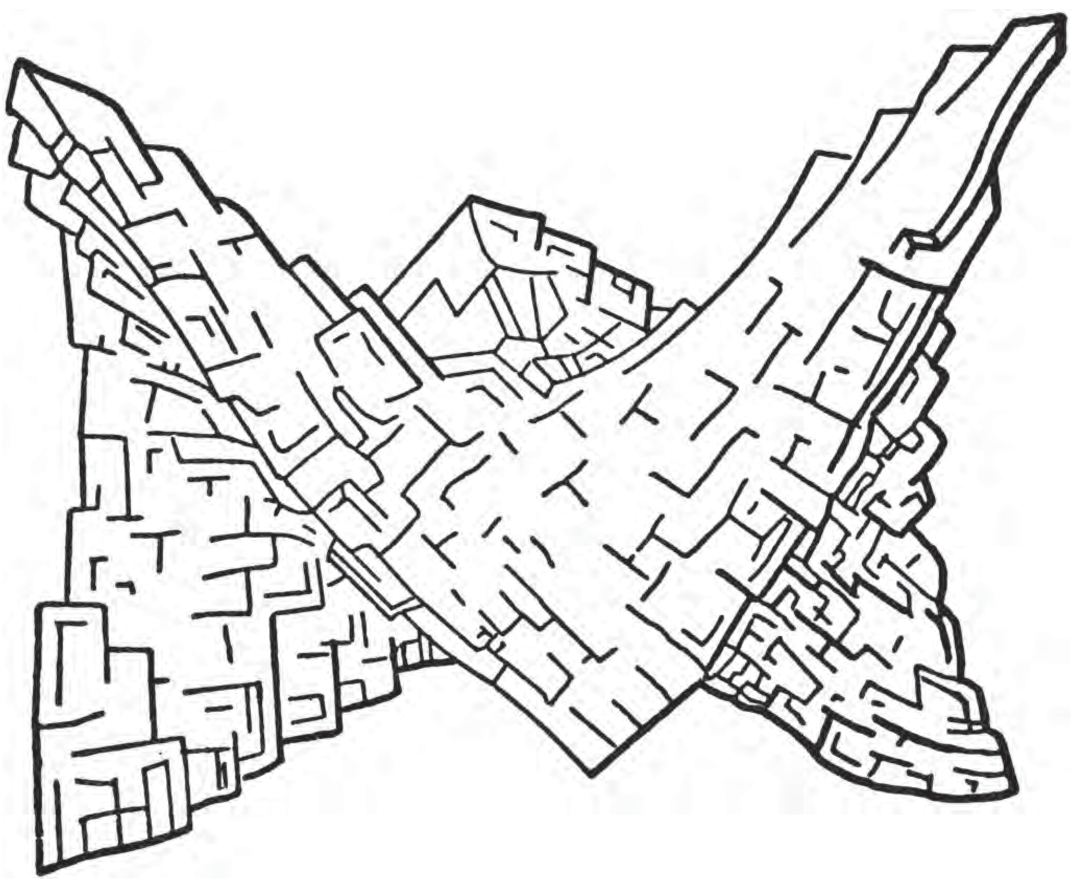


ROAD LOG: FIELD GUIDE TO SELECTED SITES IN THE OKANOGAN HIGHLANDS

Ralph L. Dawes

Wenatchee Valley College

INTRODUCTION

Following this background on the geology of the Okanogan Highlands is a road log for a field trip conducted for the Okanogan Highlands Alliance in August of 2013. We will conduct essentially the same field trip; however, the road log has been revised for the Tobacco Root Geological Society in 2014 with the goal of examining and discussing the geology in greater depth. Because the field trip will be based in Republic rather than Tonasket as the 2013 field trip was, the sequence of stops was altered. We also have a new stop planned at an active quarry near Wauconda, which has been arranged with Columbia River Carbonates, who manage the site.

Aspects of the geology of the Okanogan Highlands to consider while on this field trip:

1. The Okanogan gneiss dome or Okanogan metamorphic core complex?

Both the Okanogan and Kettle metamorphic core complexes have been referred to as gneiss domes, mainly in older references. As a gneiss dome, the rock is viewed as having risen as a diapir, a more or less mushroom-shaped, ductile mass of flowing rock (Eskola, 1948; Fox et al., 1976; Holder and Holder, 1988, Kruckenberg et al., 2008). Although the concepts and terminology of metamorphic core complexes largely displaced that of gneiss domes in the 1980s, since 2000 there has been a revival of the term gneiss dome and renewed discussion of the timing, mechanisms, and tectonics involved. The tendency lately has been to classify metamorphic core complexes as one class of gneiss domes (Teyssier and Whitney, 2002; Whitney et al., 2004).

The metamorphic core complex model became prominent in the 1980s (Crittenden et al., 1980; Rhodes and Cheney, 1981; Davis, 1983; Coney and Harms, 1984). As a metamorphic core complex, the structure is seen as having been uplifted along a low-angle normal fault, referred to as a detachment fault in this context. The detachment fault and deformation in the hanging wall above is brittle. The relatively cold and brittle roof rocks have slid down and off to the side. Grabens or half-grabens may occur in the roof rocks, formed at about the same time as the core complex and detachment fault. Beneath the detachment fault, the upper zone of the footwall is mylonitic and commonly contains synkinematic intrusive rock. In general, the metamorphic core comples is the footwall of one or more detachment faults, and non-mylonitized parts of core complexes typically contain granitic rock that was at least in part intruded at the same time as the core complex was being uplifted and exhumed.

Rocks to the west of the Okanogan detachment fault have been interpreted as being the hanging wall or roof of the Okanogan metamorphic core complex. The Okanogan fault itself runs through the Okanogan River valley and is largely covered by alluvium and glacial drift. The mylonitic shear zone in the footwall of the Okanogan fault extends several km eastward into the Okanogan metamorphic core, and is over one km thick in some places (Kruckenberg et al., 2008).

2. Structural expressions of changes in the regional stress regime during the Eocene

During the Eocene epoch, the stress regimes in the inland Cordillera changed from predominantly compressional and transpressional to tensional and transtensional. In the Okanogan Highlands, the main Eocene structures are the Okanogan and Kettle metamorphic core complexes, and the Republic and Toroda grabens. This change in stress across the region has most often been ascribed to a change in tectonic interactions along the western margin of the North American plate; normal plate convergence and subduction during the late Mesozoic changed either to transform faulting, highly oblique subduction, or both, during the Eocene (Engebretson, 1985; Haeussler et al., 2003). Parts of the North American margin moved northward and, in the hinterland, extension began in zones previously subject to compression.

In addition, plate interactions along the continental margin of the Pacific Northwest during the Eocene may have led to the formation of a slab window in the mantle beneath the region (Thorkelson and Taylor, 1989; Hooper et al., 1995; Breitsprecher et al., 2003). By providing a pathway for the asthenosphere to rise to the base of the continental lithospherein place of a colder plate that had previously subducted beneath the region, the slab window may have caused heating, thermal expansion, decompression, and partial melting of the lithosphere, contributing to the regional uplift and collapse (spreading and thinning) of previously over-thickened crust.

3. Location of the craton margin relative to the Okanogan Highlands

Most maps that reconstruct the pre-Cenozoic or pre-Mesozoic margin of the Laurentian/North American craton draw the boundary just east of the Okanogan Highlands. In northeastern Washington, miogeoclinal strata that formed on the largely passive continental margin are involved in the Jurassic-age fold and fault structures of the Kootenay arc. The Sr 0.706 line, commonly taken as a proxy for the boundary of the craton, is generally projected passing to the east side of the highlands (Armstrong et al., 1977; Ghosh, 1995).

However, there are some metamorphic and plutonic rocks in the Okanogan and Kettle core complexes with early Paleozoic and Precambrian ages (Ross and Parrish, 1991; Armstrong et al., 1991), and volcanic rocks in the Canadian Okanogan Highlands with nodules interpreted as lithospheric mantle of Precambrian age (Dostal et al., 2003). These clues suggest the possibility that part of the craton might be somewhere within the Okanogan Highlands. Telescoping of cratonic crust hasdisrupted and obscured the original margin of the continent in the vicinity of the Okanogan Highlands (Colpron and Price, 1995; Lund, 2008, Ghosh, 1995). This was caused first by Jurassic Jurassic and Cretaceous compression during the formation of the Kootenay arc,accretion of the Intermontane superterrane, and eastward thrusting of the Intremontane superterrane.Subsequently, during the Eocene, extension-driven formation of grabens and metamorphic core complexes, along Eocene igneous intrusions, volcanic accumulations, and metamorphism, has further complicated the picture. Although the boundary of the old craton may be within the Okanogan Highlands instead of along its east side, the continental margin has been buried and modified by the geologic events that formed the highlands.

4. Correlations and histories of stratified formations and terranes in the Okanogan metamorphic core complex and adjacent areas

The stratified rocks in the Okanogan metamorphic core complex and adjacent areas have been disrupted and overprinted by faulting, metamorphism, and intrusion, making it a challenge to establish correlations and agree to a common set of stratigraphic names. A compounding factor is that many of the rocks and structures continue across the Canada-U.S. border. It is not unusual for geologists mapping in different countries to have applied different names to the same rock units (Fox, 1970; Little, 1982; Orr and Cheney, 1987).

Correlating and mapping formations of stratified rock is important for mapping the geologic structure and locating zones of possible economic value. It is also important for deciphering the history of terrane accretion of the area. As it stands, the most cited model for terrane accretion in the Okanogan Highlands and adjacent areas is Jurassic accretion of the Intermontane superterrane (Monger et al., 1982), a composite terrane combining oceanic, ocean plateau, and island-arc crust, to the continental margin. Most, if not all, of the accreted rock in the Okanogan Highlands is thought to be part of Quesnellia, one of the Intermontane terranes.

5. Eocene magmatism

The large volume of igneous rock that intruded into and erupted onto the area of the Okanogan Highlands during the Eocene epoch raises a number of questions. The timing overlaps with the Eocene episode of igneous activity that broke out across much of the Pacific Northwest, from the Cascades to the area east of the Rocky Mountains, known as the Challis episode or Challis event.

In some parts of the Pacific Northwest, northern Rocky Mountains, and the High Plains, the Challis volcanics are much more potassic, or alkaline, than typical arc rocks. However, in most Eocene igneous suites in the Okanogan Highlands and northeastern Washington, the rocks are predominantly calc-alkaline

(Holder and Holder, 1988; Hooper et al., 1995; Morris et al., 2000). The slab-window hypothesis explains the derivation of calc-alkaline rocks with island arc-like geochemistry from older, arc-derived basement beneath the region.

Another feature of Eocene volcanic rocks in and around the Okanogan Highlands is their association with Eocene faults. The Republic and Toroda grabens are largely filled by Eocene volcanic rocks that are contemporaneous with extension of the grabens by normal faulting (Fox et al., 1976; Pearson and Obradovich, 1977; Atwater et al., 1984; Holder and Holder, 1988). Some of the intrusions in the core complexes are thought to have occurred during active movement on the detachment faults.

West of the Okanogan fault, on the hanging wall of the Okanogan metamorphic core complex, there are batches of Eocene volcanic rock with ages close to the ages reported for movement on the fault (Stoffel et al., 1991). It is not clear if the proximity to the detachment fault zone of volcanic rock in the hanging wall of the core complex arose because the magma was channeled along the fault zone to the surface, or if its location on the down-dropped hanging wall of a normal fault has led to the preservation of volcanic rocks which erupted from volcanic vents further from the faults.

6. Glaciation of the Okanogan Highlands during the Pleistocene epoch

The Okanogan lobe of the Cordilleran ice sheet covered all of the Okanogan Highlands except for a few isolated nunataks. The ice sheet reached its terminus on the Columbia Plateau, on the other side of the Columbia River, south of the highlands. All the bedrock peaks, knobs, ridges and hills in the highlands were glacially sculpted to some extent, as were the valleys. Nearly all the valleys and basins are partly to largely underlain by till, outwash, and, in some cases, clay and silt from glacial-lacustrine and glacial-margin environments. Erratic boulders that originated further north are common across the area. Sequences of nested glacial terraces can be seen along the sides of most large valleys in the highlands.

There is evidence in British Columbia (Fulton et al., 1992; Kovanen and Slaymaker, 2004) and on the Columbia Plateau to the south (Booth et al., 2003) that there were pre-Wisconsin glacial advances across the area, but most if not all the glacial features in the Okanogan Highlands have been interpreted to be Late Wisconsinan. Based on correlation with dated glacial deposits occurring both north and south of the area, the Okanogan lobe advanced southward across the border from Canada approximately 17,000 years ago and retreated back into Canada about 11,000 years ago.

It has been argued that subglacial melt water, under high hydraulic pressure due to confinement by the thick, overlying ice, created many of the drumlin-shaped bedrock ridges and valley-connecting cross-channels found in the highlands, including the highlands west of the Okanogan River (Lesseman and Brennand, 2009). Similarly, the Okanogan Valley, including the portion in Canada known as the "Okanagan" where there has been more detailed recent study of the glacial history, has been posited by some researchers as a site where large volumes of subglacial melt water built up into a system of subglacial lakes, which probably broke out on several occasions into jökulhlaups that flooded part of the Channeled Scabland to the south (Kovanen and Slaymaker, 2004; Lesseman and Brennand, 2009; Shaw at al., 1999). Whether subglacial melt water, rather than glacial ice, did a significant amount of eroding and sculpting of the Okanogan Highlands and whether outbreaks of the subglacial melt water contributed significantly to large-scale flooding of the Channeled Scablands to the south, are unresolved questions.,.

Most studies have argued that the glacial erosional features in the highlands are the result of direct erosion by glacial ice, and that the lakes were subaerial features along the glacier's margins (Atwater et al., 1984). The complex arrangement of the peaks and valleys of the Okanogan Highlands may have lent itself to the formation of temporary, ice-dammed lakes. The glacially-impounded lakes deposited varves and other lacustrine sediments which are present in many places in the highlands and adjacent valleys.

Along the southern border of the Okanogan Highlands, the Okanogan lobe advanced across the Columbia River west of Grand Coulee, forming glacial Lake Columbia and diverting the drainage through Moses and Grand Coulees. Glacial Lake Columbia, which lapped up against the southeast side of the Okanogan


Okanogan Highlands Alliance
Geology Tour 2013
Route Map
Marble Outcrop at Chesaw
Chesaw
8
Granodiorite at Pontiac Ridge Rd N
7
Paleozoic Spectacle Fm/Beaver Canyon
6
Beth Lake Campground
5
Sample Rock: Sanpoil Lavas/Beaver Cnyn
Sample Rock: Pontiac Ridge Rd S
4. Bodie/Klondike Volcanic
3. Moccasin Lake Overlook
Wauconda
Nealy/Havillah Rd - Vista Point
9
Havillah
10
Antoine Crk Ovrlk/Gneiss
2. Kame Terraces
Image USDA Farm Service Agency
© 2013 Google
Image © 2013 DigitalGlobe
97
7
20
Tonasket
1. Gneiss Borrow Pit
Turn-off for Stop 1
Google earth
Imagery Date: 9/25/2011 48°47'51.02" N 119°07'03.86" W elev 5959 ft eye alt 30.67 mi

Highlands, is thought to have broken through its ice dam only once, as the glacier melted and wasted away toward the end of the Pleistocene epoch (Atwater, 1987; Waitt et al., 2009).

GEOLOGY OF THE OKANOGAN HIGHLANDS TOUR 2013

(http://okanoganhighlands.org/education/geology2013)

INTRODUCTION

Two hundred million years ago, the area that is now the Okanogan Highlands was located at the western margin of the North American continent. Since that time, three major geologic processes have shaped the area: terrane accretion, Eocene extension and its related volcanism, and glaciation.

Between 200 and 180 million years ago, volcanic islands built on oceanic crust were accreted onto the continent in the Highlands region during a period of compression from the west. Quesnellia is the largest accreted terrane in the Okanogan Highlands and the one from which we will see rocks on this tour.

Between 60 and 40 million years ago, the crust under the Okanogan Highlands was stretched and broken apart at the same time as volcanoes erupted in the area. Today we will see rocks representing two different forms of eruptions – some occurred as lava flows and some as exploded bits and pieces. The extension and resulting faulting of the crust that occurred at this time created metamorphic core complexes and grabens, which are the definitive geologic structures of the Highlands. We will see a metamorphic core complex, a zone of igneous and metamorphic rocks that were domed upward beneath a detachment fault, at Stop 1, and grabens at Stop 4.

Between 2.6 million and 12,000 year ago, continental glaciers repeatedly advanced from the north, reshaping the Okanogan Highlands. The glaciers smoothed the ridges and peaks, deposited flat layers of sediment on the valley bottoms, formed flat benches along the sides of the valleys, and created a system of lakes and streams. This unique combination of bedrock, geologic structures, and glacial features create the wondrous landscape of the Okanogan Highlands.

STOP 1

Stop 1: Amphibolite Gneiss Borrow Pit

<u>Location:</u> On Clarkson Mill Road, 1.6 miles south of Tonasket (48° 40’ 56.27”N 119° 27’ 27.24”W)
<u>Parking:</u> Approach from the southern end of Clarkson Mill (48° 40’ 34.94”N 119° 28’ 02.13”W), and pull out on the right-hand side of the road heading north, just before the borrow pit.

This borrow pit is in the Okanogan fault zone, which forms the western border of the Okanogan Metamorphic Core Complex. Here in the borrow pit we are looking at a mylonite (Greek for “milled rock ”) composed of amphiblolite and granite from the Tonasket Gneiss, which is part of the Okanogan Metamorphic Core Complex. It was recrystallized from shear stress exerted by movement on the Okanogan fault. Note the parallel manner in which the minerals in the granite (quartz, feldspar, and biotite) have been recrystallized.

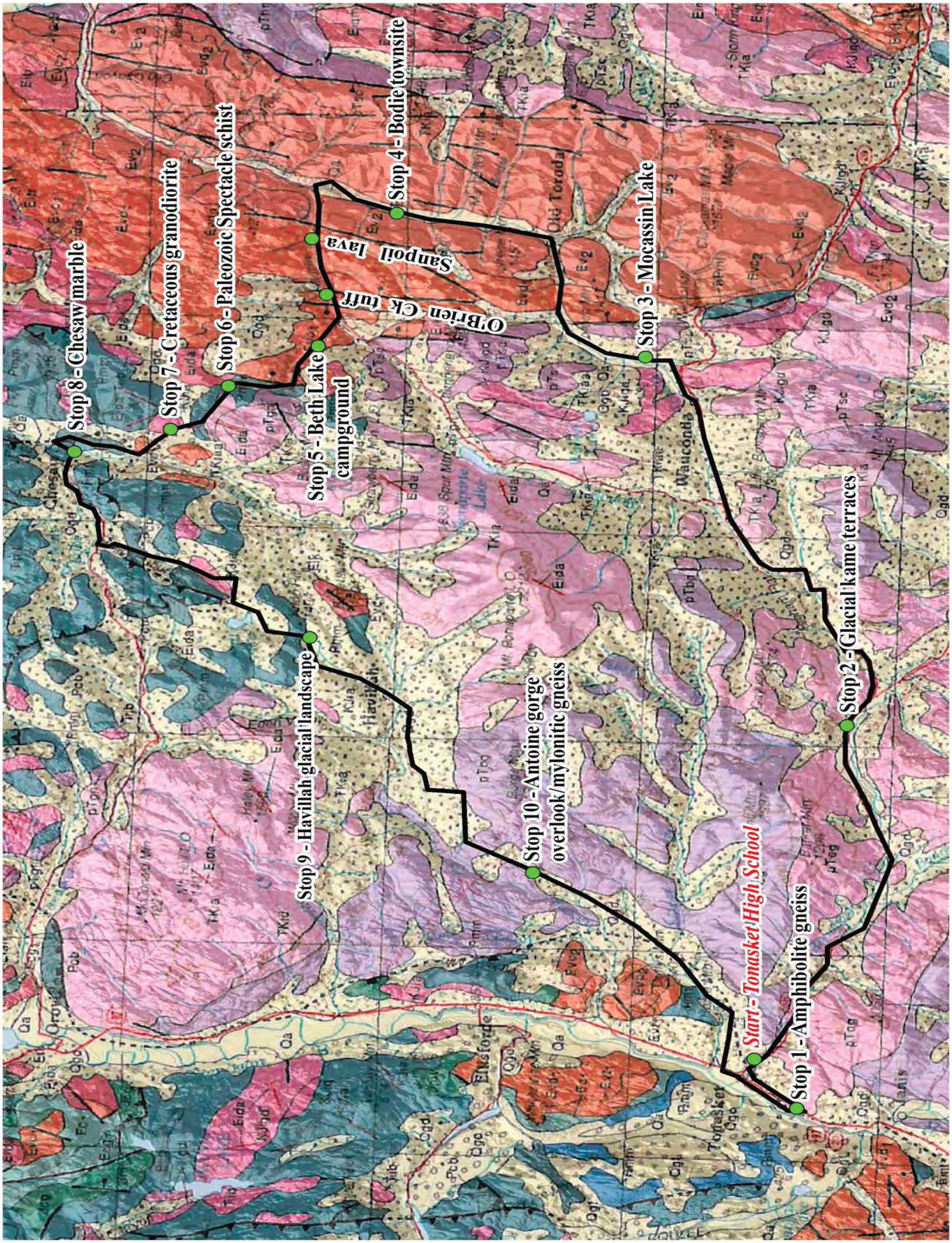
The Okanogan River valley follows the Okanogan fault zone from Canada to Omak, following the weaker materials in the fault zone. The hills on the west side of the Okanogan River consist of rocks that had been above the Okanogan fault and slid off to the west. The “flat irons” that line the east side of the Okanogan valley between Riverside and Tonasket are composed of mylonite originating below the fault zone that was sheared by sliding.

STOP 2

Stop 2: Bonaparte Creek widens and Kame Terraces are prominent

<u>Location:</u> Between Mileposts 273 and 274, just before Cayuse Mountain Road (48.66990,-119.22489)
<u>Parking:</u> Large, paved pullout

Starting as much as 2.6 million years ago, the Okanogan Highlands were glaciated several times by enormous ice sheets that flowed out of western Canada. The most recent glaciation of the Highlands ended about 12,000 years ago and left the landscape we see today. As the most recent ice sheet stagnated and melted, remnant glaciers were left occupying the major valleys. Between the sides of the remnant glaciers and the valley walls, glacial meltwater deposited sand, gravel, and boulders, forming kame terraces. Terraces at more than one elevation along the valley suggest that the glacier melted in stages.


Start - Tonasket High School
Stop 1 - Amphibolite gneiss
Stop 2 - Glacial kame terraces
Stop 3 - Mocassin Lake
Stop 4 - Bodie townsite
Sanpoil lava
O'Brien Ck. tuff
Stop 5 - Beth Lake campground
Stop 6 - Paleozoic Spectacle schist
Stop 7 - Cretaceous granodiorite
Stop 8 - Chesaw marble
Stop 9 - Havillah glacial landscape
Stop 10 - Antoine gorge overlook/mylonitic gneiss
Old Toroda
Wauconda
Tonasket
Ellisforde
Chesaw
Havillah

STOP 3

Stop 3: Moccasin Lake

Location: Wauconda (State Land, before Kiesecker's property) (48.74289, -118.98775)
Parking: Pull in facing the gate

Note that the outcrop of marble just ahead on route (at 48.75161, 118.98769) has been metamorphosed longer and more intensely than the marble at Chesaw, Stop 8.

The lake is on private property. It is located along Walker Creek, which flows north into Toroda Creek. Local folklore has several stories about this lake , which include the story that Moccasin Lake is bottomless or at least unfathomably deep, and a story of a person falling into the lake and the body being found in a stream to the north. Soundings performed with a weight and line in the summer of 2013 found the lake to be no more than 10 m deep and to have a muddy bottom. The Toroda Creek valley shows signs of being occupied by a remnant glacier while the ice sheet was melting in the Highlands. There are what appear to be kame terraces along parts of the valley. Here, the Toroda Creek valley is flat, suggesting that the whole valley may have been temporarily occupied by a lake when the remnant glacier melted. It may also be all that is left of a once larger lake in a stream drainage that had been disrupted by glaciation, or it may be a kettle lake. Just northwest of Moccasin Lake is a landform that may be a stranded delta formed when a side stream discharged into the temporary lake.

What is the origin of Moccasin Lake? It may be all that is left of a once larger lake in a glacially disrupted stream drainage, or it may be a kettle lake, a hollow formed by sediment accumulating around a chunk of ice left from a glacier.

Glaciers commonly excavate deeper into soft or weak zones in the bedrock than they do in harder, more competent materials. Bedrock that crops out through the glacial sediments on the hillsides around Moccasin Lake includes marble, which is relatively soft and soluble in water. The same bedrock presumably underlies the glacial outwash, varved lake sediments, and stream sediments that fill the bottom of the valley. This can lead to speculation about the possible role of bedrock in the formation of Moccasin Lake.

Leaving Moccasin Lake, we will drive- by an outcrop of strongly foliated and lineated white marble, that probably originated as limestone from an accreted terrane. The Quesnellia terrane includes limestone deposits formed in a shallow, tropical ocean over 200 million years ago. In some parts of Quesnellia, the limestone is fossiliferous, which help narrow its age to late Paleozoic to early Mesozoic (roughly between about 300 and 200 million years ago). Here the marble has been so strongly metamorphosed that no fossils remain. Note the sugary, shiny calcite crystals that make up this marble. From the quarry a mile or two to the north, relatively pure, high-calcium marble is excavated, taken in trucks down to a train siding south of Tonasket, and loaded onto railroad cars to ship to a processing facility near Portland. The marble is used as a filler in paper, filling in spaces between paper fibers to make paper less expensive per pound, brighter, and whiter.

STOP 4

Stop 4: Bodie townsite/Klondike Volcanic Outcrops

Location: Just before the historic townsite of Bodie
Parking: Park at second pullout and walk back to the first outcrop of andesite (48.83175, -118.89762)

There is a volcanic breccia cropping out just across from the parking spot.

Backtracking just a little south on the road, the second outcrop is also composed of andesite. It has curved flow banding suggesting that lobes and bulges formed as the flow grew (48.83058, -118.89933).

We have driven out of the Okanogan metamorphic core complex and into the Toroda Creek graben. The rocks in the Toroda Creek graben, like the Republic graben to the southeast, largely consist of Eocene volcanic rock, extruded between 48 and 57 million years ago. Down dropping of the Toroda Creek and Republic grabens, eruption of the volcanic rocks, and formation of the Okanogan Metamorphic Core Complex were contemporaneous.

The two outcrops at Stop 4 contain two different types of volcanic rock from the Klondike Mountain Formation, showing contrasting styles of eruption. The southern outcrop is an andesitic flow that displays flow banding, which is common in viscous, slow-moving flows. Such viscous lavas have more silica in them than the runnier type of lava that forms basalt, and, as seen here, solidify into a much lighter colored rock than basalt. The northern outcrop just across from the parking spot is a volcanic breccia formed during an explosive eruption and welded together by the heat of the molten material.

Though we don't see it here, the Klondike Mountain Formation also contains sedimentary units including layers of sandstone, siltstone, and shale, deposited in streams and lakes during extended lulls in the volcanism. The Tom Thumb Tuff, the basal member of the Klondike Mountain, contains many plant fossils (and a few fish and insect fossils). They are well exposed, and open to the public, in the Stone Rose fossil quarry in Republic.

Volcanic Stratigraphy of the Toroda Creek and Republic Grabens (youngest at top)

Klondike Mountain Formation – lava flows, sedimentary layers, and volcanic breccia
Sanpoil Volcanics – thick lava flows
O'Brien Creek Formation – tuff, lapilli, volcaniclastic rocks, and sedimentary rocks

The O'Brien Creek Formation is characterized by thick white layers of volcanic ash, crystals, lapilli, and other volcanic debris. Much of the volcanic debris was volcaniclastic, which was then reworked by streams. The Sanpoil Volcanics are notable for thick flows that form the steep slopes and high hills in the area, including the eastern portion of Beaver Creek.

STOP 5 — Stop 5: Beth Lake Campground and lunch

Location: Beaver Canyon
Parking: Campground sites 13 and 14, no fee, day use permission granted. The group can eat at #13. Go left, following the sign for Campsites.

We will share samples of volcanic rock from outcrops we passed on the road . They include:

Samples from a flow within the Sanpoil Volcanics : (Lat/Long: 48° 51' 33" 118° 55' 11"; GPS Lat/Long: 48.85901, 118.91974), and

Samples of lapilli tuff from the O'Brien Creek Formation on Pontiac Ridge Road (Lat/Long: 48° 51' 11" 118° 56' 58"; GPS Lat/Long: 48.85287, -118.94937).

STOP 6 — Stop 6: Outcrop of the Paleozoic Spectacle Formation

Location: In Beaver Canyon, at Milepost 26, 2.8 miles beyond Beth Lake Campground; look for dead-top alder to mark the spot just beyond pullout (48° 53' 16"N 119° 00' 41"W).
Parking: Pull out on the right-hand side of road, just before the outcrop; the group will walk along roadside to the outcrop.

At this stop we will see metamorphic schists and amphibolites of the Quesnellia terrane, which were originally oceanic crust on which volcanoes built up into an island arc. Modern day examples of island arcs include Japan and the Aleutian Islands. The variety of rock types in Quesnellia reflects the origin of the terrane. They include ocean-floor and island-arc volcanics; intrusive granodiorite that solidified beneath the volcanoes; and a variety of marine sedimentary rocks including mudstone, sandstone, and limestone. All this rock accreted to the edge of North America early in the Jurassic period (between 170 and 200 million years ago). Over the millions of years during which the accretion occurred, parts of Quesnellia were broken and thrust over other parts of Quesnellia along a thrust fault. The Chesaw thrust, which is near here, is one of these faults, but it is covered by glacial sediment where our route crosses the fault.

STOP 7 — Stop 7: Granodiorite at the Mouth of Beaver Canyon

Location: 1.8 miles from the previous stop, at Chesaw end of gorge, at the north end of Pontiac Ridge Road (48° 54' 11" 119° 02' 16")
Parking: Turn left into ample space across from the outcrop, in a wide across from Pontiac Ridge Road.

The rock at Stop 7 is Cretaceous granodiorite, as discussed at the previous site.

The Buckhorn Mountain mine is not far from here. The gold deposit at the Buckhorn mine formed as a result of contact metamorphism and metasomatism that occurred when granodiorite – similar to what we see here – intruded marble of the Quesnellia terrane.

In view across the valley, note the *roche moutonnée*, a rock feature caused by movement of the glacier across the bedrock, creating an asymmetrical landform in which rock has been smoothed on one side and plucked on the other. Literally translated from French, the term means “sheep rock” and was first applied to landforms like this in the European Alps.

STOP 8

Stop 8: Marble Outcrop at Chesaw

Location: Just south of Chesaw at dramatic bend in the road (48° 56’ 25” 119° 03’ 18”)
Parking: The landowners just before the turn in the road, the Leslies, have granted permission for us to park in their driveway, although it may be easier to park in the pullout on the right.

The rock here is marble of the Quesnellia terrane. Some outcrops of Quesnellia marble are so little metamorphosed that fossils are still discernible in them – fossils of marine plankton and seafloor dwelling organisms that lived in a warm, shallow ocean. The fossils are of Permian age (between 252 and 299 million years old). The marble here is less metamorphosed, strained, and recrystallized than the marble near Moccasin Lake at Stop 3. The steeply tilted beds in the marble here may be the original sedimentary beds, rather than layers created by metamorphism.

STOP 9

Stop 9: Hungry Hollow Hummocks

Location: Along Hungry Hollow Road 1 mile from its junction with Chesaw Road.

The landscape here is kame and kettle topography. Kames and associated kame deltas and eskers occur when meltwater atop a glacier thermally and mechanically erodes moulins (i.e., vertical shafts) that deliver sediment to the interior or base of a glacier. This sediment accumulates in thermally- and mechanically-eroded hollows and tunnels. When the ice melts, a variety of topographically positive features appear depending on the original shape of these hollows. Conical hills reflecting roughly circular ice voids are kames, while sinuous ridges known as eskers develop in meltwater channels. Kame deltas are roughly circular, but flat-topped because they formed in subglacial lakes and were confined by the overlying ice. As such, eskers are often associated with kame deltas. Finally, the depressions seen on kames and kame deltas are kettles, formed by the melting of chunks of glacial ice once buried in glacial drift. From the ground (but much better from the air via Google Earth), we can see an esker in the foreground and kames, kame deltas, and kettles in the background. While it does not appear to be true here, kames, eskers, and kame deltas are often quarried because of the well-sorted sands and gravels associated with them. Google Earth shows such a quarry north of our stop and just north of the Chesaw Road.

These features formed during ice stagnation in the area, near the end of the Pleistocene probably 15,000 to 11,000 years before the present.

STOP 10

Stop 10: Antoine Creek Gorge Overlook and Gneiss Outcrop

Location: Along Havillah Road (48° 46’ 27” 119° 19’ 04”)
Parking: In pullout just after gneiss outcrop and near gorge overlook

Since the last glacial ice melted away (or possibly starting before then), Antoine Creek incised a steep gorge within a broader glaciated valley.

Tonasket Gneiss crops out along the road: Compare the rock here with the amphibolite we saw at Stop 1. Here the Tonasket Gneiss consists of mylonite and several types of granitic rock that reflect various degrees of metamorphism and recrystallization. Like the rocks we saw in the borrow pit, the rock here was caught up in the shear stress of the Okanogan detachment fault during the Eocene epoch. The fabrics and cross-cutting relationships in some of the light-colored, granitic rock in the outcrops at Stop 10 indicate that the magma was intruding contemporaneously with the metamorphism of the mylonite.

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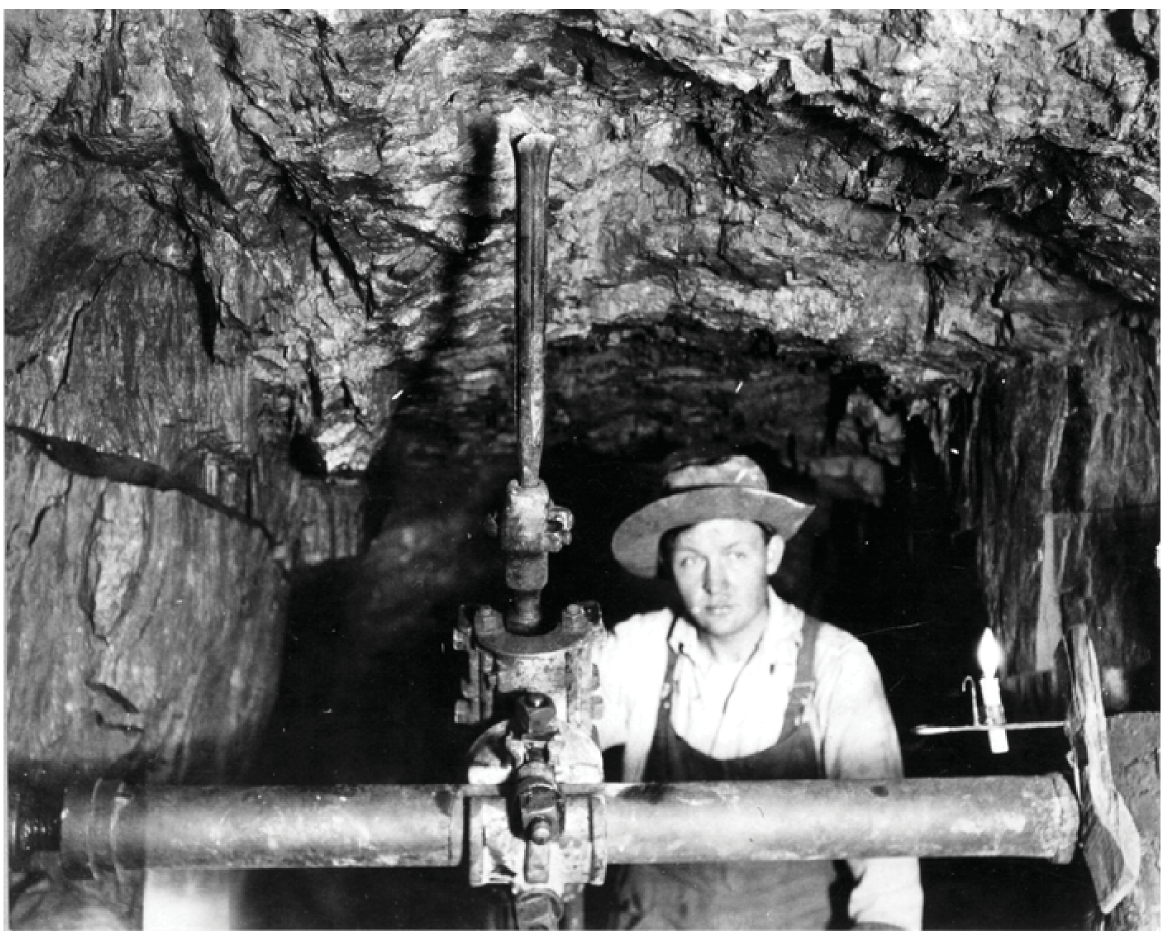

VERTICAL ZONING IN EPITHERMAL GOLD-SILVER DEPOSITS, REPUBLIC MINING DISTRICT, FERRY COUNTY, WASHINGTON

Arthur L. Smith

Hecla Silver Valley, Inc., Coeur d'Alene, Idaho

INTRODUCTION

The Republic Mining District has produced 2,973,708 ounces of gold from near surface epithermal deposits from 1986 to 1995. Hecla Mining Company entered the district in 1981 when it acquired Day Mines, Inc. Remarkably, major exploration companies paid little attention to the area until Hecla Mining Company discovered the Golden Promise ore body in late 1984. The deposit represents a "blind" discovery that was buried beneath 400 to 700 feet of post-mineralization volcanic and sedimentary rocks (Tschauder, 1985; Fifarek, et al., 1996). Hecla Mining Company is currently working on a 3D computer model of the district to aid in the understanding of deposit geometries and future exploration potential.

REGIONAL GEOLOGY

The Republic region is located at the southern end of the Omineca crystalline belt, which is distinguished by high grade metamorphic core complexes (gneiss domes) and intrusion of the Colville batholith. Near Republic, the Okanogan and Kettle gneiss domes are separated by a set of northeast-trending en echelon grabens filled with Eocene-age volcanics. The Republic graben is 50 miles long and six to ten miles wide, see figure 1. Local basins likely developed as a result of extensional faulting at a time of waning volcanism around 47 to 54 Ma. Hydrothermal activity driven by deep heat sources and channeled through structural conduits formed mineral deposits near the paleosurface which were subsequently covered by lake-bed sediments. The deposits were uncovered again by Pleistocene glacial scouring (Chapman, et al., 2009; Fifarek, et al., 1996; Braun, 1989).

DISTRICT GEOLOGY

Eocene volcanics in the graben have been divided into the older tuffs of the O' Brien Creek Formation, overlain by thick porphyritic andesite flows with interbedded flow breccias, epiclastic breccias, and sediments of the Sanpoil Formation. Hypabyssal feeder dikes of the Scatter Creek Formation intrude the Sanpoil. An angular unconformity separates the Sanpoil from overlying lake bed sediments of the Klondike Mountain Formation. The unconformity marks a time of relative volcanic quiescence, hot-spring activity, and associated gold mineralization. Post-mineral amygdaloidal basalt dikes and sills intrude both the Sanpoil and Klondike Mountain Formations (figs. 2 and 3). Pleistocene-age glacial till is common over much of the region (Braun, 1989; Chapman et al., 2009; Fifarek et al., 1996).

GEOLOGIC HISTORY

The geologic history and setting of hydrothermal activity in the Republic Mining District can be inferred from the depositional facies of the upper Sanpoil Volcanics and middle to lower Klondike Mountain Formation. During the waning stages of Sanpoil magmatism, a small, asymmetric graben trending northwest formed within the larger northeast-trending Republic graben (fig. 4a). Braun, 1989, believes this graben formed from evacuation of an underlying magma chamber. According to Fifarek, et al., 1996, the deposition of the epiclastic dominant (ED) facies; normal movement along the Eureka, Boundary, and South Basin faults; and hot springs activity including hydrothermal eruptions were synchronous.

Sinter and hydrothermal eruption breccias were deposited along a common, district-wide surface that is recognized as the approximate upper contact of the ED facies. These deposits were ultimately buried and preserved by alluvial fan deposits of the basin margin (BM) facies along the Eureka and South Basin fault scarps (fig. 4b). Hydrothermal eruption breccias, brecciated sinter, and clasts of chalcedonic quartz are locally present in the basin margin facies, particularly near its base. The basin margin facies and hot springs deposits were eventually inundated by rising lake waters and buried by mud-dominated sediments of the lacustrine facies (L) (fig. 4c; Fifarek et al., 1996).

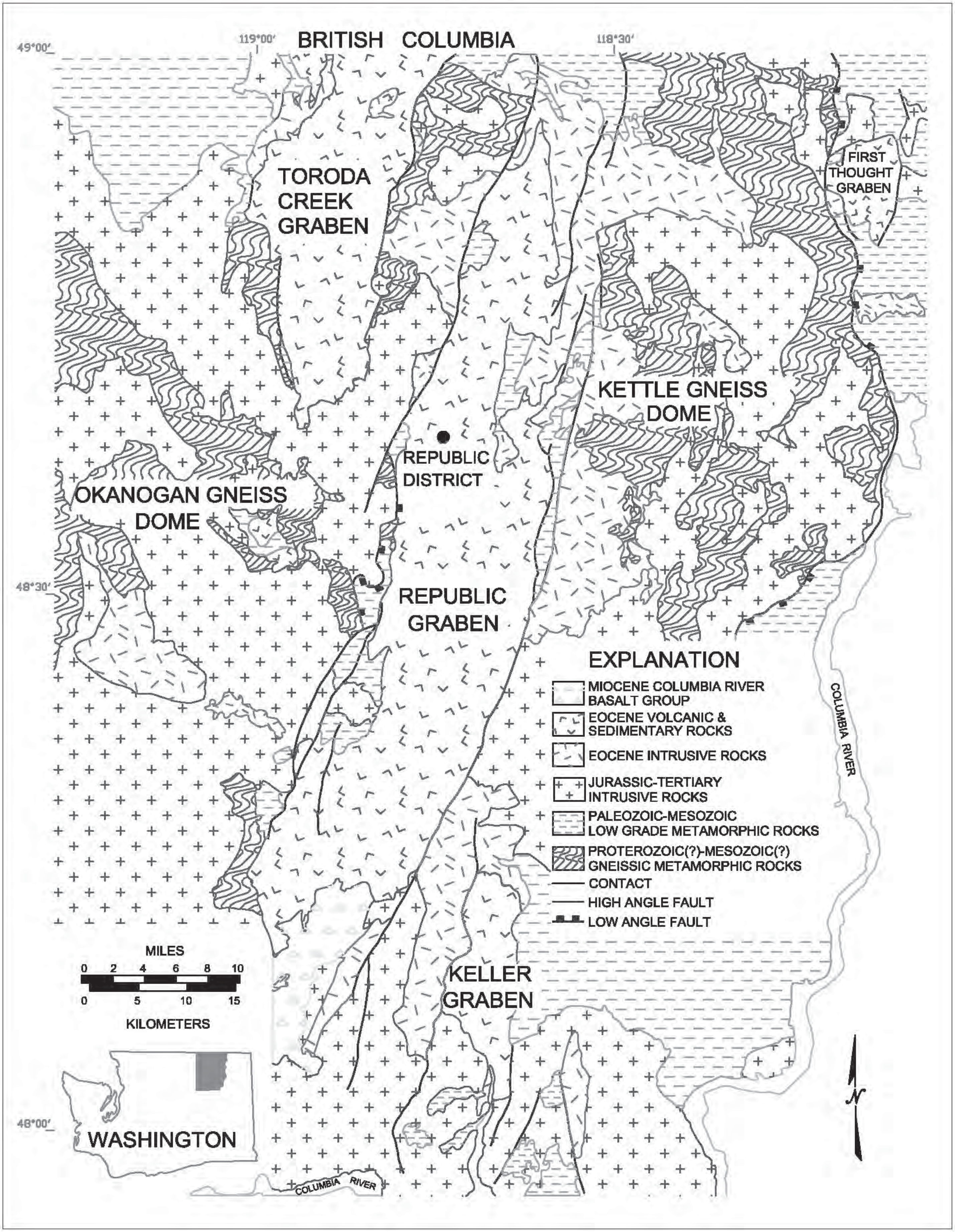

Figure 1. Principal structural features and rock types of the Okanogan Highlands of north-central Washington and the location of the Republic District (adapted from Fifarek et al., 1996).

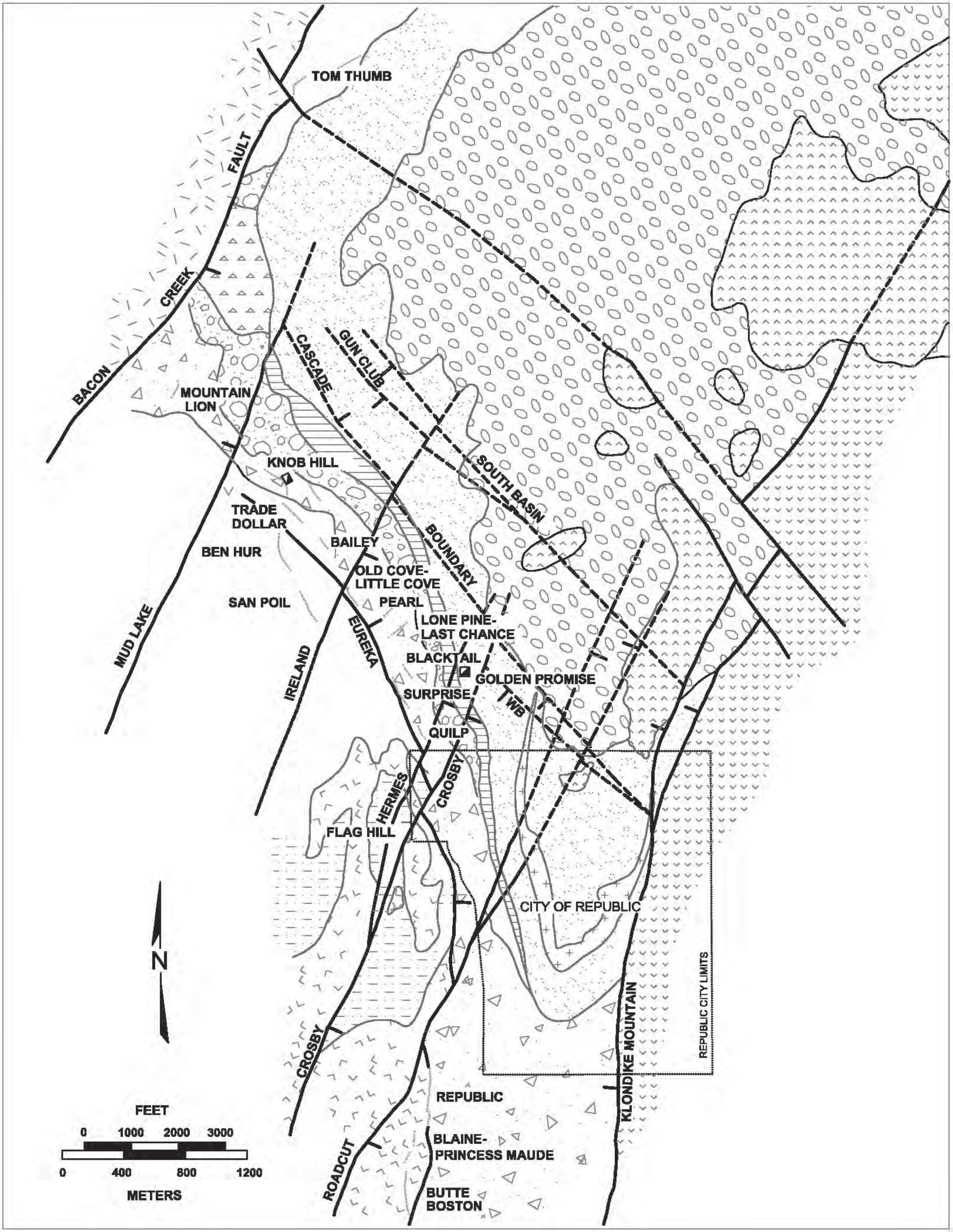

Figure 2. Geologic map for the Republic Mining District showing the distribution and stratigraphic relations of the major rock types, epithermal vein deposits, and faults (Fifarek et al., 1996).

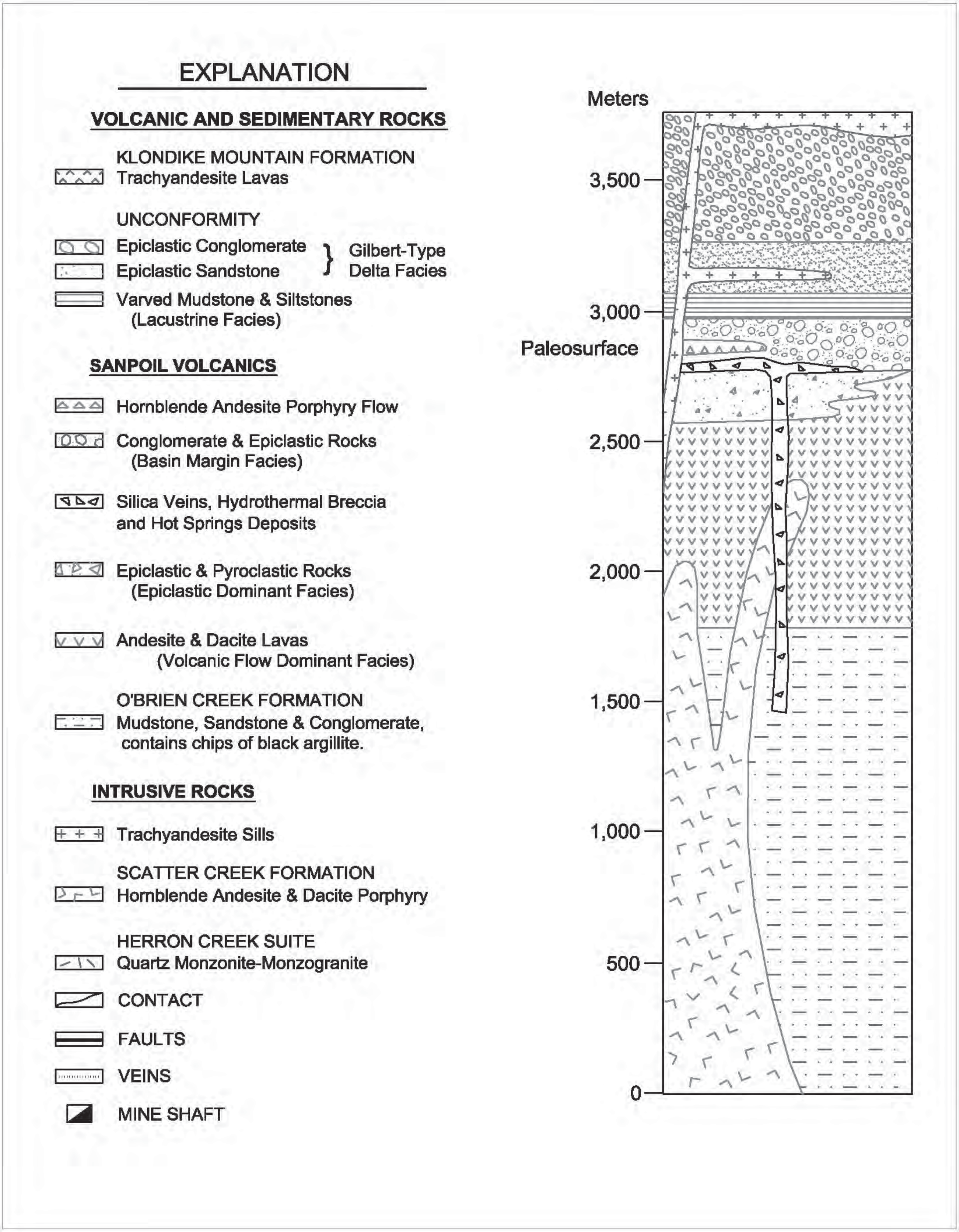

Figure 3. Stratigraphic column for the Republic Mining District (Fifarek, et al., 1996).

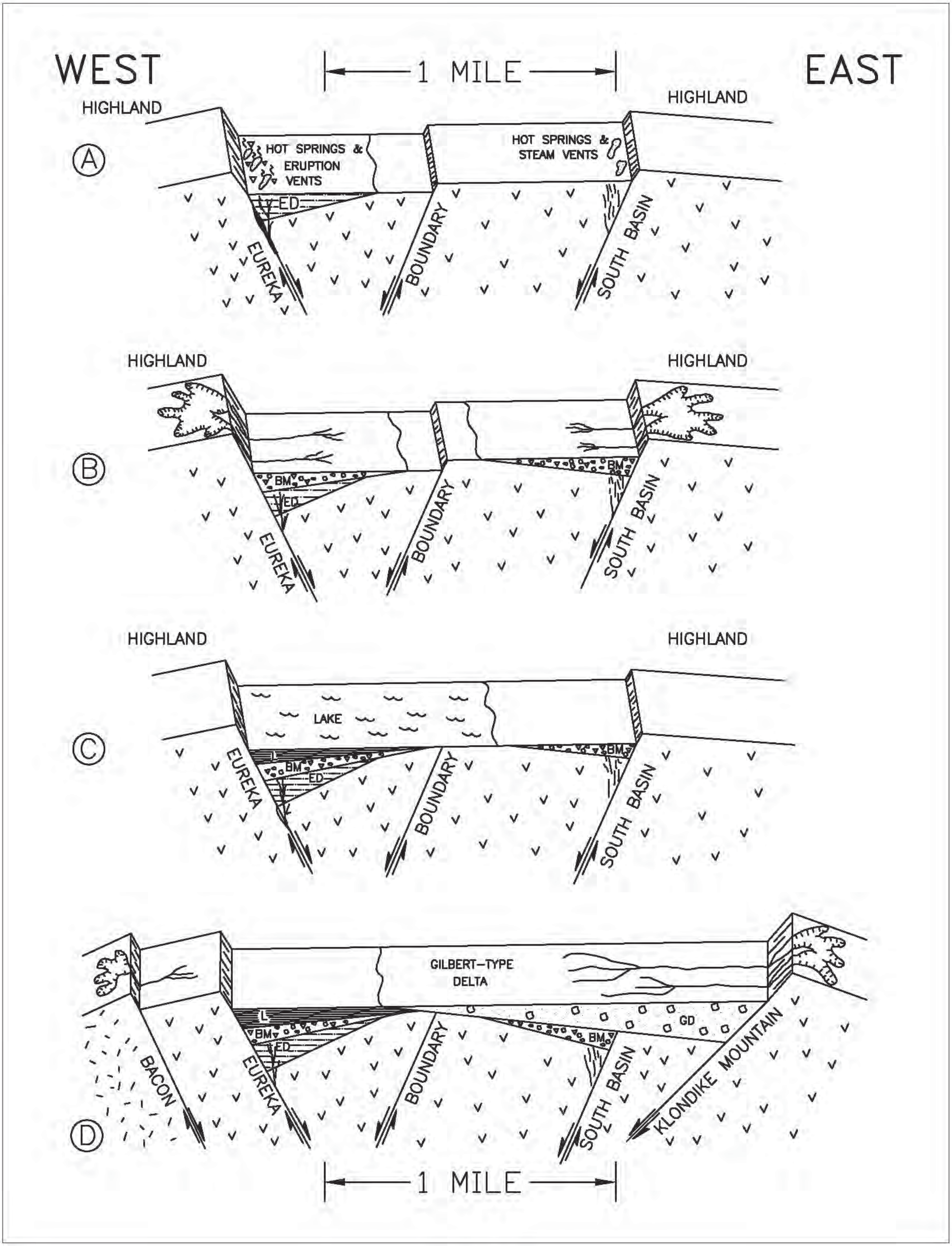

Figure 4. Schematic model depicting the development and filling of an asymmetric graben in the Republic district and the timing of epithermal mineralization relative to the basin-fill facies.

Price, 1991, indicates that the hydrothermal activity continued sometime after lacustrine deposition began. Deposition of Gilbert-type deltaic sand and gravel deposits onto the lacustrine plain signaled an increase in sediment due to a combination of better developed drainage basins, continued basin subsidence, and uplift of the highlands along the northeast-striking Bacon and Klondike Mountain faults (fig. 4d; Fifarek et al., 1996).

a) Deposition of the epiclastic dominant, ED, facies and contemporaneous hot springs activity including hydrothermal eruptions (upper Sanpoil Volcanics).

b) Formation of alluvial fan deposits by debris flow and fluvial mechanisms that buried and preserved the epithermal deposits (basin margin facies, BM, upper Sanpoil Volcanics).

c) Inundation by lake waters and deposition of organic-rich mud-and silt-dominated sediments (lacustrine facies, L, lower Klondike Mountain Formation).

d) Deposition of Gilbert-type deltaic sands (Gilbert-type delta facies, GD, middle Klondike Mountain Formation), Fifarek et al., 1996.

GEOLOGIC CHARACTERISTICS OF EPITHERMAL AU-AG DEPOSITS

A typical epithermal gold-silver deposit can be broken down into four vertical zones, each of which has certain characteristics of geometry, Au grades, Ag:Au ratios, geochemistry, and alteration. A cross-section through a typical epithermal system is shown in figure 5. The paleosurface of the system contains a siliceous vuggy sinter of chalcedony in the form of a mound or terrace. Below this is a wedge-shaped stockwork of banded chalcedonic stringers which coalesce downward into a large bonanza vein of delicate, colloform-banded chalcedony. Vein mineralogy in the bonanza zone consists of electrum, native gold, naumannite, chalcopyrite, and pyrite. This large vein gradually transitions into a glassy, low-grade quartz vein that is coarser grained at depth (Braun, 1989; Tschauder, 1989).

Gold grades within the sinter zone typically run from 0.05 to 0.30 ounce per ton (opt), while the stockwork zone contains erratic grades. As the bonanza (boiling) zone is approached, gold grades steadily increase to over 1.0 opt and locally can be in excess of 100 opt. The bonanza zone can have a vertical thickness of several hundred feet. Gold grades taper off to sub-economic values less than 0.25 opt in the deep zone (fig. 5; Braun, 1989).

Vertical zoning can be seen in the epithermal system based on patterns of Ag:Au ratios (fig. 5). In general, the sinter zone has a low Ag:Au ratio while the stockwork zone is erratic. There is an increase in the Ag:Au ratio as the bonanza zone is approached followed by a steady decrease with depth (Braun, 1989).

The distribution of trace elements in the wallrocks around the veins is typical of an epithermal gold system. In general, wallrocks adjacent to a vein are enriched in Au, Ag, and Sb, while erratically distributed Hg and As anomalies are located near the paleosurface (Tschauder, 1989). The geochemical signature of Hg, As, Sb, and Mo can determine vertical depth in the system (fig. 5). The sinter zone is strongly enriched in Hg which diminishes rapidly with depth. The consistent decrease of As and Sb makes them good depth indicators, while the consistent increase in Mo with depth makes it a good deep-level indicator (Braun, 1989).

According to Tschuader, 1989, “The wallrocks have been silicified and argillically altered. The silicified zone, in the shape of a downward tapering funnel centered on the vein, tends to grade outward into an argillic zone, which in turn grades outward and downward into a propylitic zone.”

ROAD LOG

This half day field trip will examine an Eocene epithermal hot spring system at various levels below the paleosurface in the central portion of the Republic Mining District. The trip route and stops are shown on figure 6. All mileage was taken from Google Earth.

Many of the stops will take place on public roads. Park as far off the road as possible and please exercise extreme caution when leaving your vehicles. Permission must be obtained prior to entering private land.

Cumulative mileage is given on the left: interval mileage is shown in parentheses.

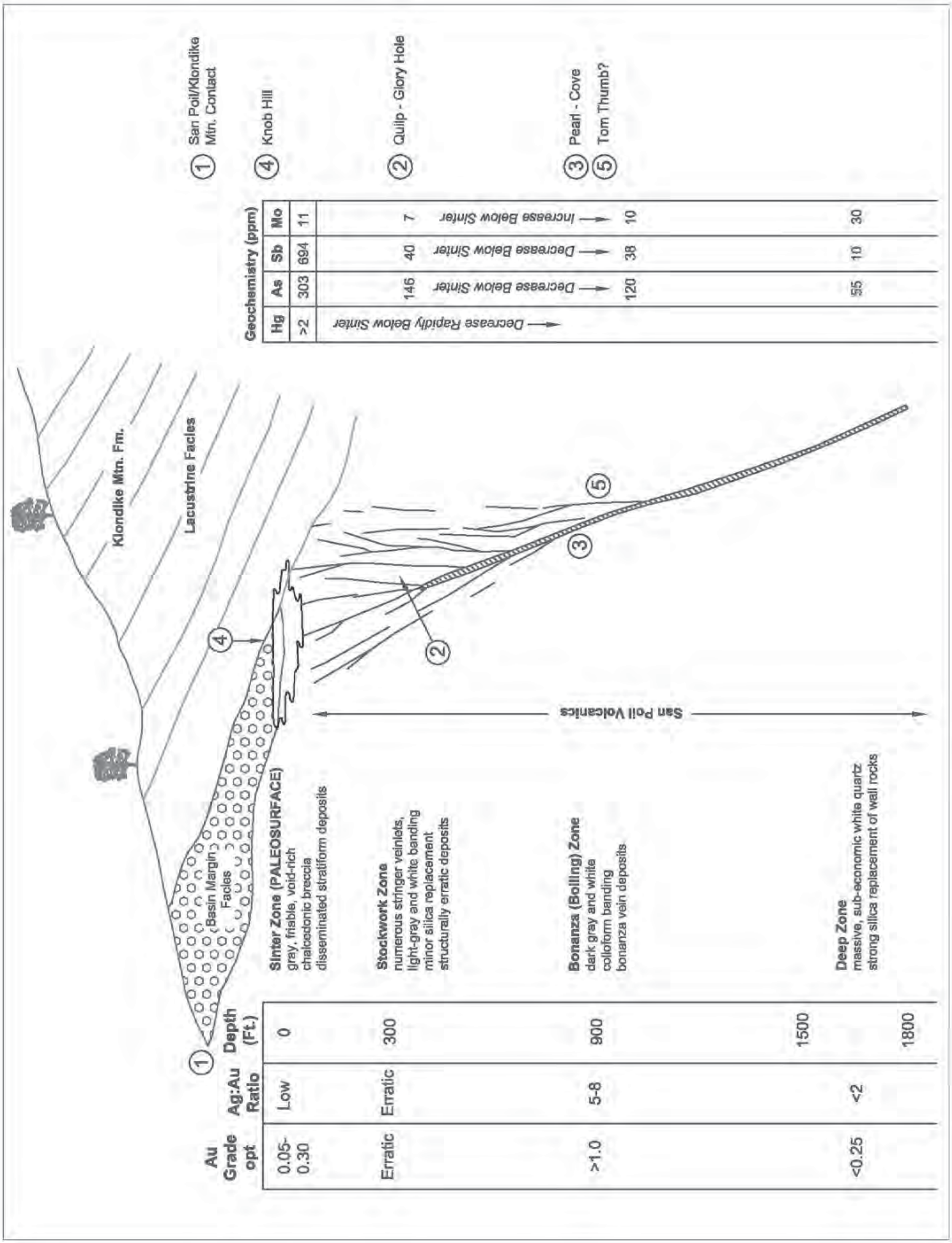

Figure 5. Physical characteristics of hot spring gold deposits along the Eureka graben (modified from Braun, 1989).

Figure 6. Map showing field trip stops along Knob Hill road.

0.0 (0.0) Start at the intersection of Creamery Road (Fairgrounds) and Washington State Route 20 (Sherman Pass Scenic Byway); turn left and follow into the town of Republic. Route 20 turns into Clark Street; go north to end of Clark Street.

3.1 (3.1) Turn left on Knob Hill Road and go past the Stonerose fossil site on the right. Pull off on left side of road. Be careful crossing the road to the outcrop.

STOP 1 Stop 1: Sanpoil Volcanics/Klondike Mountain Formation contact

3.3 (0.2) The outcrop on the right is one of the few places where the contact between the Sanpoil Volcanics and the Klondike Mountain Formation is exposed. Note the clasts of chalcedonic quartz in the light-green conglomerate (Tschauder, 1989).

4.3 (1.0) Continue northwest on Knob Hill Road and pull off on the primitive road to the right. Go through the locked gate and follow the primitive road for 0.3 miles. Park on the large mine dump of the RAD decline.

The Golden Promise veins were first intercepted by drill holes from an exploration drift in 1963. That drilling program resulted in three mineralized intercepts that contained in excess of 0.20 opt Au; the best intercept was three feet with 6.62 opt Au and 32.46 opt Ag. This area was not explored for 20 years, partly because the geologists in 1963 believed the pyroclastic rocks of this area were poor hosts for gold mineralization. In 1984, a reevaluation of geologic data led to a successful program of diamond drilling and exploratory drifting by the Hecla Mining Company. Nine of ten holes drilled intersected intervals in excess of five feet with 0.25 opt Au, including one intercept of 28 feet with 1.57 opt Au and 11.95 opt Ag. Production from the mine exceeded 723,116 tons at an average grade of 0.75 opt Au and 3.84 opt Ag (Tschauder, 1989; Fifarek, et al., 1996).

Walk 450 feet south along the dump to the Quilp glory hole.

STOP 2 Stop 2: The Quilp mine

4.7 (0.4) Rock outcrops are tuff breccias and lithic tuffs of the upper Sanpoil Volcanics. There are two important structural trends exposed in the Quilp glory hole, the northwest-striking and northeast-dipping Eureka fault, seen at the north end of the pit, and a set of northeast-striking and southeast-dipping faults and veins, seen at the south end of the pit. The Eureka fault system controls mineralization along the Republic, Quilp, Surprise, Pearl, Cove, Bailey, Gold Dollar, No. 3, and Mountain Lion veins. The northeast-striking structures parallel the Hermes fault, which lies buried in the gulch southeast of the Quilp Mine. The Hermes and Crosby faults collectively offset the older Eureka fault approximately 1,640 feet in a dextral sense (Tschauder, 1989; Fifarek, et al., 1996).

The glory hole is a good example of the style of mineralization near the top of the stockworks zone (approximately 200-300 feet below the paleosurface). The stockworks merge underground into a vein which averaged 0.410 opt Au. Production from the upper part of the Quilp ore shoot, mined from the glory hole, totaled 50,393 tons of ore at an average grade of 0.186 opt Au and 1.14 opt Ag (fig. 5; Tschauder, 1989).

5.1 (0.4) Continue on the primitive road back to Knob Hill Road, then turn right.

5.4 (0.3) Turn right onto the dirt road and drive about 500 feet, park on the right.

STOP 3 Stop 3: The Pearl-Cove mine

5.5 (0.1) Walk up the old road to the Pearl-Cove stopes. The Pearl-Cove development is on the north extension of the Quilp-Surprise vein zone. The average Ag:Au ratio is 7.1 which puts the Pearl-Cove near the bonanza (boiling) zone, approximately 900 feet below the paleosurface. See figure 5. Production from the Pearl-Cove totaled 30,325 tons of ore at an average grade of 0.22 opt Au and 1.1 opt Ag (Full, et. al., 1959; Braun, et. al., 1986).

5.6 (0.1) Return to Knob Hill Road, then turn right.

6.2 (0.6) Turn right onto the gravel road and park. ***This is a congested area. Please watch for traffic.***

STOP 4

Stop 4: Knob Hill (Stewart Pit) Sinter

6.4 (0.2) Walk up the road and go through the second gate on the left. Continue walking uphill to the ridge. Hecla Mining Company is currently building a 60-million gallon lined impoundment for groundwater management. ***For safety considerations, foot traffic may be limited in areas with active construction. Please do not access any areas without permission from tour guide.***

The contact between the Sanpoil Volcanics and the Klondike Mountain Formation lies just to the east of Knob Hill (fig. 2). Sinter-style mineralization was mined from the Mud Lake and Stewart pits from 1937-1941. The Mud Lake pit lies to the left of our position on the ridge while the Stewart pit is directly downslope.

The sinter zone consists of a combination of hydrothermal breccia, subaerial silica, and silica replacement of lahars and lake sediments deposited on the paleosurface. The Mud Lake pit lies at or just below the paleosurface and exploited a gently east-dipping breccia body in a tabular shape estimated to be 700 feet long, 120 feet wide, and 60 feet thick. Remnants of the bleached and silicified breccia can be seen in the west wall of the pit along with stratified lacustrine sediments capping the hill. There does not appear to be a physical connection between the breccia and the underlying vein, which suggest that the deposit formed from a zone of lateral hot spring discharge. Production from the pit totals 499,076 tons at an average grade of 0.097 opt Au and 0.85 opt Ag (Nielsen, 1982; Braun, 1989; Tschauder, 1989).

The Stewart pit lies at or just below the paleosurface and mined a downward-tapering breccia body about 200 feet long, 40 feet wide, and 30 feet thick that is centered over the Stewart vein. Carbon fragments are encapsulated in silica within the breccia body, and a layer of organic carbon trash crops out along the north wall of the pit. Pyritic mineralization has occurred in places, and the wall rocks are argillically altered. The geometry of the breccia body coupled with the abundant carbon trash suggests the Stewart ore body formed in a hydrothermal eruption crater. Production from the pit and underlying stope totals 44,197 tons at an average grade of 0.280 opt Au and 1.43 opt Ag (Nielsen, 1982; Tschauder, 1989).

6.6 (0.2) Return to the vehicles and continue north on Knob Hill Road.

7.2 (0.6) Pull across the road and park. The Mountain Lion mine lies just to the west of the parking area. The Golden Eagle deposit of Midway Gold is located beneath the flat field to the east. Mineralization outcrops at the Mountain Lion mine and plunges 15° to 20° northeasterly under post-mineralization cover. The bulk of the deposit consists of a strong- to moderately-silicified hydrothermal breccia which is similar to the Mud Lake breccia. Midway Gold's website indicates that the deposit contains 31,400,000 Tonnes at a grade of 1.88g/t for a total of 1,744,000 contained Au ounces.

7.9 (0.7) Continue on Knob Hill Road until it merges with Trout Creek Road (SR) 257.

8.4 (0.5) Turn right onto dirt road.

STOP 5

Stop 5: Tom Thumb outcrop

8.9 (0.5) Park and walk about 170 feet up the short hill to the northwest. This is the outcrop of the Tom Thumb breccia vein which is up to 10 feet thick and dips easterly at about 25°. The contact between the Sanpoil Volcanics and the Klondike Mountain Formation lies near the gulch bottom about 150 feet east. Tschauder, 1989, believes the veins occur along the contacts of rhyodacite flows, and that mineralization in the area was controlled by stratigraphic contacts rather than by steeply dipping faults. He also states that all the veins in this area contain brecciated and re-cemented, banded chalcedony. Isolated chalcedony clasts can be found in brecciated flows adjacent to the veins. Production from the Tom Thumb ore shoot totals 25,000 tons at an average grade of 0.334 opt Au and 1.61 opt Ag. The Ag:Au ratio is 4.8 which would put this near the bonanza (boiling) zone.

End of road log.

ACKNOWLEDGMENTS

Thanks to Hecla Mining Company for permission to publish this data; Bruce Cox for suggesting a field guide of the epithermal deposits would be interesting and for reviewing and editing this paper; Steve Petroni for suggesting I write the paper and for reviewing and editing the guide; and to Brandi Rollins for her help in preparation of the illustrations.

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KINROSS' KETTLE RIVER OPERATIONS AND THE BUCKHORN MINE

Peter Cooper

Chief Geologist

Kinross' Kettle River Operations (KRO) are centered near the town of Republic, in north central Washington State. The KRO is anchored by a conventional grinding-flotation-CIL gold concentrator that processes 2,000 ton/day (tpd). It has treated gold ores from eight separate sources in four production centers since 1990.

The deposits that have been mined over this period represent at least three different deposit types in rocks of significantly different ages. The majority of historical production tonnage has come from three deposits of volcanogenic/replacement (skarnoid?) bodies of magnetite-hematite-sulfide in Permo-Triassic sediments; the balance was developed from four deposits of Tertiary epithermal quartz lodes, and one deposit of pyroxene-garnet-magnetite skarn in Permo-Triassic sediments and volcanics. The Buckhorn skarn deposit, currently in production, has contributed proportionally more gold (37 percent through the end of 2013). Since 1995, all mine production has been transported to the mill by highway haul trucks.

The KRO processing facility only operated near its nominal capacity of 2,000 tpd in six of its first ten years, and has been significantly under-utilized since the waning of production from the Lamefoot and K2 ore bodies in 2000. Custom toll milling of ores delivered from other locations has become an increasingly important component of KRO's business activities.

All of the various deposits that have been mined and processed by the KRO were discovered and/or delineated by Echo Bay Minerals Company, initially in joint venture with Crown Resources Corporation (CRC). The Buckhorn deposit is also a CRC discovery, although full deposit delineation was accomplished by Battle Mountain Gold Company.

The Buckhorn Mountain gold deposit is hosted by skarn occurring adjacent to the Mesozoic Buckhorn Mountain granodiorite-diorite pluton, approximately 47 road miles northwest of the KRO mill site. Within the Buckhorn skarn system, exoskarn is widespread within Paleozoic metasediments, and to a much lesser extent within overlying Mesozoic basaltic volcanics. Endoskarn occurs sporadically within dikes, sills, and irregular bodies of diorite and granodiorite that have intruded the sequence. There are also intrusive phases that did not develop associated skarns.

Post-mineral faulting has produced locally significant offset of the main ore zone. The deposit occurs within a mile of the western limit of the Tertiary Toroda Creek Graben, and graben-parallel faults are an important component of the local structural fabric. The volcanic and volcaniclastic rocks that occur within the graben structure can be subdivided and correlated with Eocene formations occurring in the Republic Graben to the southeast.

Ore-grade gold mineralization within the Buckhorn skarn system occurs in multiple settings. In order of decreasing importance, these include: 'stratabound' skarn in gently dipping calcareous sedimentary rocks (main ore zone), massive skarn in undifferentiated protolith, skarned intrusive bodies, and within hornfels and calc-silicate altered volcanics overlying the sedimentary skarn.

Gold is most highly concentrated in the distal portions of the skarn. Accessory sulfides are dominated by pyrrhotite (average 5-10 percent in ore) with ubiquitous traces of chalcopyrite (locally to 1 percent). Arsenopyrite, pyrite, and bismuthinite are localized, and while they are minor components overall, they locally exceed several percent. The only metal that correlates well with gold is bismuth, and the visible occurrence of any of a suite of bismuth minerals is a good indicator of high gold concentrations. The overall gold to silver ratio is about 2:1. There are no economic concentrations of metals other than gold.

Gold mineralization within the main ore zone is more or less continuous along approximately 800 feet of strike and 1,200 feet down-dip. Vertical thickness ranges up to 60 feet. Gold is concentrated at or toward the footwall contact of skarn with marble for much of the zone's extent.

The Buckhorn mine achieved commercial ore production in June of 2008, and full-scale production of 1,000 tpd a year later. Cumulative production through the end of 2013 has been 2,165,000 tons with an average mined grade of 0.43 ounces gold per ton (opt). Reserves and resources contained in the YE2013 mine plan ($1,200-Au) total 592,400 tons at an average grade of 0.28 opt-Au.

The Buckhorn deposit has become the single largest source of gold for KRO. As presently outlined, it will be second only to the Lamefoot deposit in terms of ore tonnage to the mill. Depletion of the current reserve and resource is expected by mid to late 2015, at which point it is anticipated that cumulative gold production from the deposit will have been close to 1.1 million ounces.

STRATIGRAPHY, STRUCTURE, AND MINERALIZATION IN THE QUESNEL TERRANE NORTH OF REPUBLIC, WASHINGTON

Eric S. Cheney

Department of Earth and Space Sciences

ABSTRACT

In the Republic area, Paleozoic and Mesozoic marine rocks of the Quesnel terrane are restricted to the Sanpoil syncline between the Eocene Okanogan and Kettle metamorphic core complexes. The Quesnellian rocks are cut by the Jurassic regional Chesaw thrust fault, which is discontinuously marked by serpentinite. The Quesnellian rocks and the thrust fault are at least locally isoclinally folded. A klippe of Quesnellian rocks near Danville, Washington is caused by the Sanpoil syncline, and erosion of the Kettle River valley. Mineralization at the Morning Star mine near Danville and the Lamefoot mine near Curlew Lake is associated with felsic metavolcaniclastic rocks.

INTRODUCTION

The purpose of this field trip is to examine the stratigraphy, structure, and some of the mineralization in the pre-Cenozoic rocks of the Quesnel terrane north of Republic, Washington (fig. 1). Although ore specimens are not available, we will contemplate mineralization in the Jurassic Elise Formation at the Morning Star mine near Danville and in the Triassic Brooklyn Formation at the Lamefoot mine near Curlew Lake.

The trip begins at the junction of SR 21 and the Kettle River Road on the south side of the Kettle River at Curlew, Washington. Kettle River Road leads northwestward to the International Border crossing at Ferry, Washington, and Midway, British Columbia. Most of the field trip will be in the Danville area adjacent to the International Border (fig. 2), and we will not be crossing the border on this trip. The trip ends in Republic, Washington.

Most of the stops are along highways and roads. Participants are urged to be cognizant of traffic. One stop involves a 0.8 km (round trip) walk along an abandoned railroad grade. The final stop is a 1 km (round trip) moderate bushwhack with an ascent of 150 meters. This stop inspects a classic detachment fault in non-Quesnellian rocks, and because it is the final stop, it is optional. It is, however, recommended.

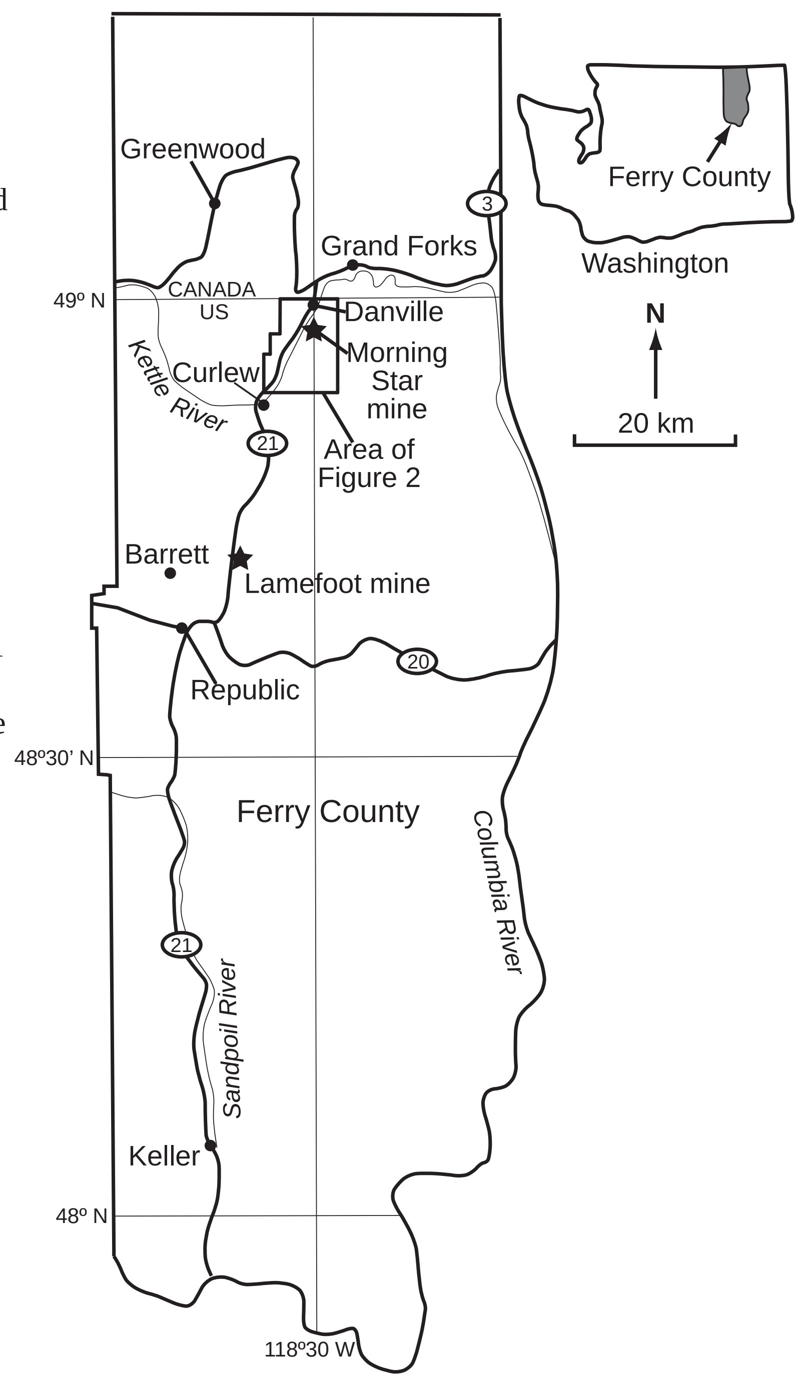

Figure 1. Location map.

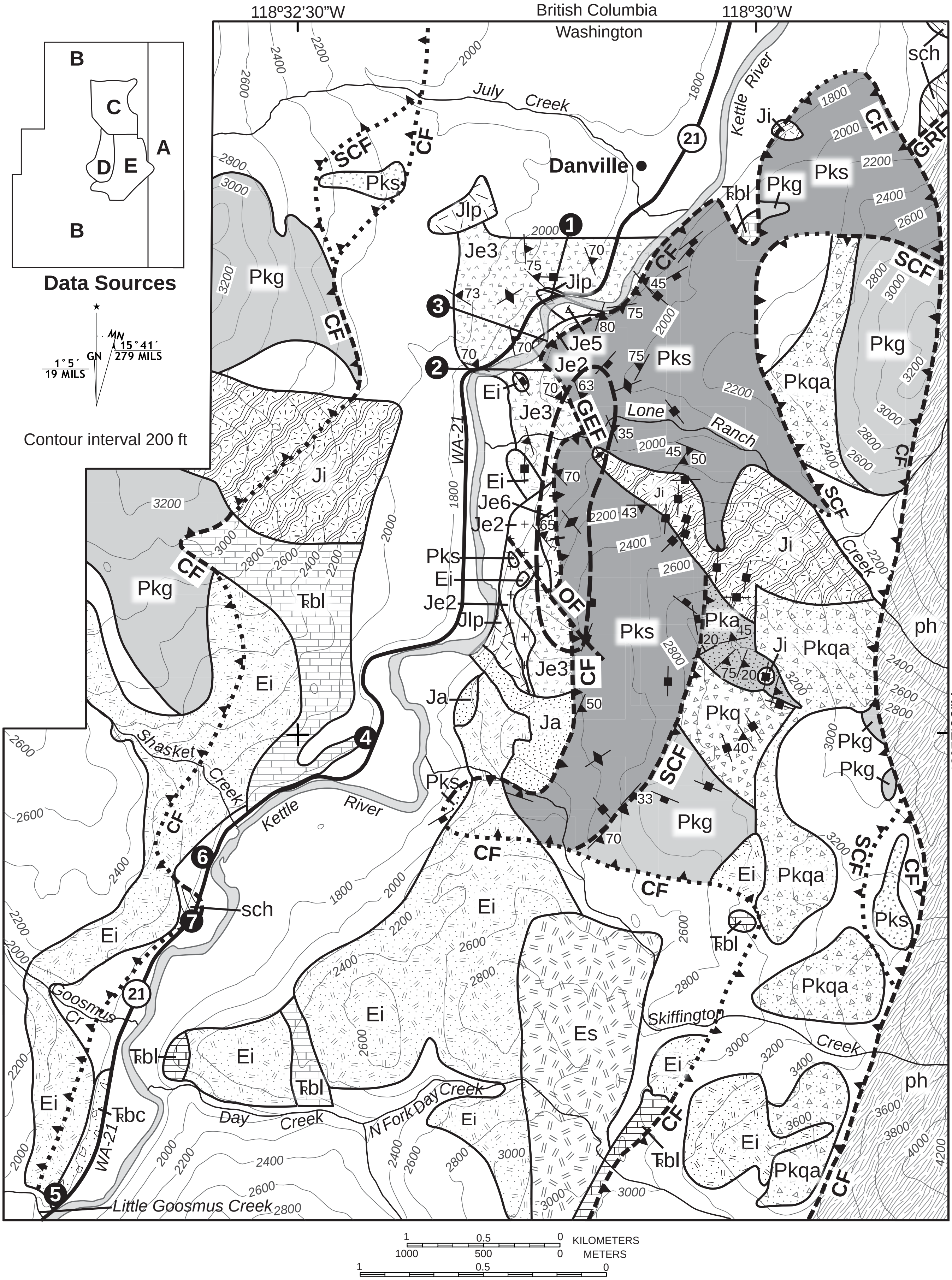

Figure 2. Geologic map of the Danville area. Sources of data are: A, Pearson, 1977; B, Parker and Calkins, 1964; E, S. Cheney 1:2400 mapping in 2010 and 2011; D, Caron, 2005; E, E.S. Cheney 1:62,500 mapping in 2011.

REGIONAL GEOLOGY

Introduction

The regional geology along the International Border consists of three major components (Cheney et al., 1994; Cheney and Rasmussen, 1996). One component is composed of Proterozoic to metamorphic and igneous rocks in Eocene metamorphic core complexes (the Kettle complex east of SR 21 and the Okanogan complex west of SR 21). The other two components are composed of Carboniferous to Jurassic rocks of the Quesnel terrane (a.k.a. Quesnellia) and regionally extensive mid-Eocene sedimentary and volcanic rocks. Quesnellia was an oceanic terrane that accreted to ancestral North America (Laurentia) in the middle Jurassic. Rocks of the metamorphic core complexes are mostly in amphibolite facies; Quesnellian rocks are in greenschist facies, and the Eocene rocks are unmetamorphosed.

Quesnellian Stratigraphy and Structure

The Quesnellian rocks, which are the main focus of this trip, consist of three packages. The ophiolitic Knob Hill Group and the mostly pelitic Attwood Group (with minor limestone) are Carboniferous to Permian. The Knob Hill Group consists of ultramafic rocks that are mostly serpentinized, metadiorite, fine-grained metabasaltic greenstone, and quartzite (metachert). Unconformable on both the Knob Hill Group and the Attwood Group is the Triassic Brooklyn Formation, in which sharpstone conglomerate grades upward into clastic limestone. The Brooklyn Formation is, in turn, unconformably overlain by the Jurassic Rossland Group. Regionally, the Rossland Group consists of pelitic rocks with a middle, predominantly mafic volcanic unit, the Elise Formation. The distinctive lithology of the Elise Formation is augite-bearing greenstone, some of which is coarsely fragmental (Höy and Dunne, 2001). In the Danville area the Lexington quartz-feldspar porphyry intrudes the Elise Formation.

Our knowledge of the Quesnellian rocks is largely imported from Canada. The Knob Hill, Attwood, and Brooklyn formations were well described by Fyles (1990) in the Greenwood mining district northwest of Danville. Höy and Dunne (2001) described the Rossland Group.

The major pre-Cenozoic fault in the area is at the base of the Knob Hill serpentinite; this is the Chesaw thrust, or No. 7 Fault (Cheney et al., 1994; Cheney and Rasmussen, 1996). The fault extends discontinuously 120 km from Omak, Washington, to Grand Forks, British Columbia (Cheney et al, 1994, fig. 2). Regionally, rocks of the Knob Hill Formation (including the serpentinite) are in the hanging wall of the fault (fig. 2). The Attwood Formation is below the fault. Although regionally the Triassic Brooklyn Formation and Jurassic Elise Formation occur both above and below the fault, at Danville they are below the fault (that is, structurally below the serpentinite). Incision of the Kettle River between Curlew and the International Border has exposed the Brooklyn and Elise formations below the fault in the Danville half-window, which is open to the north.

Drilling at the Morning Star mine area southeast of Danville indicates that the base of the serpentinite (the Chesaw thrust) dips 14 to 20 degrees eastward and that the maximum thickness of the serpentinite above it is about 200 m (Cheney, 2011). The zone of foliated rock below the fault is about 160 m thick. The 14 to 20 degree eastward dip and the map pattern of figure 2 indicate that the area of Knob Hill Group rocks east of the Kettle River is a klippe above the Chesaw thrust.

The Lexington quartz porphyry and the Shasket Creek alkalic complex bracket the age of the Chesaw fault. The Lexington quartz porphyry, which is 199.4 ± 1.4 Ma (Church, 1992), is very well foliated below the Chesaw fault along the International Border west of Danville. The Shasket Creek alkalic complex, which is 163 ± 0.4 Ma (Berger et al., 1991), intrudes the thrust southwest of Danville (Cheney et al., 1994; fig. 2 of this report).

Cenozoic Stratigraphy and Structure

The major Cenozoic structures in the area are the antiformal Okanogan and Kettle metamorphic core complexes, and the intervening Sanpoil syncline. The syncline, which is all too commonly called the "Republic graben", is primarily bounded by the low angle normal faults (detachment faults) of the metamorphic core complexes on the east and west (Cheney et al., 1994). The Quesnellian and the Eocene sedimentary and volcanic rocks occur in the hanging walls of the detachment faults. That is, the Quesnellian rocks, Chesaw thrust, and Eocene rocks were originally continuous, but are now segmented between the metamorphic core complexes.

The three Eocene sedimentary and volcanic formations in the Republic areas have a regional synclinal pattern (Cheney and Rasmussen, 1996, fig. 2). The synclinal pattern is accentuated if all of the unconformably underlying Quesnellian rocks are combined into a single (basal) unit. Because the Sanpoil syncline plunges southward, Tertiary volcanic and sedimentary rocks are not preserved to the north near the International Border. The klippe near Danville is on or near the axial trace of the syncline. Down plunge to the south, the Eocene strata are preserved in the core of this structural low (Cheney and Rasmussen. 1996, fig. 2).

The area was extensively glaciated; so, till and the forests on which it grow are common, even at higher altitudes. At altitudes less than 2,000 feet, glacial outwash and younger alluvium predominate.

LOCAL GEOLOGY

The field stops elucidate the local geology. Table 1 describes the lithologies of the Quesnellian rocks near the Morning Star mine (fig. 2). The Brooklyn Formation is omitted from this table because it does not crop out at the mine. The Knob Hill and Brooklyn formations also occur at the Lamefoot mine (Stops 8 and 9).

ECONOMIC GEOLOGY

Historical production at Morning Star (Parker and Calkins, 1964, table 5) from 1903 to 1932 was five thousand tons from tectonic stringers and lenses of massive pyrite and magnetite along or near the basal contact of the serpentinite. These production statistics suggest that the Chesaw fault disrupted a volcanogenic massive sulfide deposit that is at least 2 m thick and averages about 0.5 percent copper and 0.5 ounces of gold per ton. Its location is presently unknown.

Another two thousand tons of ore were produced at the Morning Star from 1935 to 1943, primarily from mesothermal quartz-pyrite-gold veins. These veins varied from hairline fractures to < 0.8 m wide in northwesterly fault zones that cut the serpentinite.

The Lamefoot mine was developed in the Brooklyn Formation about 30 km south of the Morning Star mine. It was discovered in 1990 and operated at about 2,000 tons per day from 1994 to 2000. Most of the gold came from quartz veinlets crosscutting massive magnetite and pyrrhotite, or interbanded magnetite, jasper, and quartz plus carbonate. Additionally, significant amounts of ore came from 1- to 20-cm wide veinlets of quartz ± pyrite ± chalcopyrite ± pyrrhotite ± arsenopyrite ± gold in the adjacent felsic metavolcanic rock.

ROAD LOG

The trip begins at the junction of SR 21 and Kettle River Road at MP 181 about 0.3 miles south of Curlew, Washington. Proceed 8.2 miles north on SR 21 past Lone Ranch Creek Road on the right (east) at MP 189.2. We are now in the northern part of figure 2. Continue another 0.7 miles to a long road cut (MP 189.9) and pull off on the shoulder of the road adjacent to the Kettle River.

STOP 1

Stop 1. Elise Formation, Lexington porphyry, and Scatter Creek intrusion

The southern end of the roadcut is a photogenic example of the Elise Formation (unit Je5 of table 1). The formation is andesitic to basaltic. Here it is blocky weathering and unfoliated. Note the large clasts, the augite (and feldspar) phenocrysts, and the conspicuous epidote.

Northward in the middle part of the roadcut are much lighter colored bodies of quartz and feldspar porphyry. This is the Lexington porphyry (Jlp of table 1). It is 199.4 ± 1.4 Ma (Church, 1992). The contacts here are enigmatic, but the different orientations of the northern and southern contacts suggest that the bodies are intrusive.

At the northern end of roadcut (at MP 190), unmetamorphosed Scatter Creek monzonite is distinctly different than the andesitic to basaltic Elise Formation and the Lexington porphyry. Although the Scatter Creek is commonly regarded as the hypabyssal equivalent of the Sanpoil rhyodacitic volcanic rocks, here its grain size is > 1 mm. The Scatter Creek is resistant to weathering, and so widespread that it commonly appears to be more abundant than the older rocks.

Turn around and take the Lone Ranch Creek Road at MP 189.2. Continue 0.5 miles to the first outcrop on the left (west) side of the road.

Table 1. LITHILOGIC UNITS IN THE MORNING STAR AREA

Note: Asterisks (*) denote units defined by Caron (2005)

NAME	SYMBOL	DESCRIPTION	FIELD STOP	REFER- ENCES
Scatter Creek	Ei*	Dikes and sills of unaltered and unmineralized (black and white) feldspar and hornblende phyric, fine- to medium-grained monzonite. Rare, cm-scale feldspar and microdiorite inclusions.	1	Parker and Calkins (1964)
Shaske t Creek	Jsp	Syenite porphyry, with 5 to 10 % pink feldspar < 1 cm	4	Parker and Calkins (1964)
	Jsi	Trachytic, fine-to medium-grained monzonite, with 5 to 20 % amphibole, no biotite, accessory sphene, and minor epidote		
Lexing -ton porphy -ry	Jlp*	Quartz porphyry. Rock is white, slightly rusty weathering, megascopically unfoliated, and has 1 to 10 % each of quartz and feldspar phenocrysts ranging from 2 to 10 mm. The quartz phenocrysts are rounded to ovoid.	1	Church (1992)
Elise Forma- tion	Je6*	Fine- to medium-grained equigranular metadiotite (green)		Caron (2005) Cheney (2009)
	Je5c* and Je5f*	Black augite phyric greenstone. Greenstone is blocky weathering, unfoliated, contains epidote, and commonly has 5 to 30 % augite up to 1 cm. Je5c has clasts (commonly augite-bearing) up to 1 meter. Je5f has no meascopically recognizable clasts.	1	Caron (2005)
	Je3*	Black augite-bearing greenstone. Rocks are commonly fine-grained and feldspathic, have a trace to < 5% of 1 to 5 mm augite, and are variably foliated. Intercalated gray units without augite may be more siliceous.	2	Caron (2005)
	Je2c* and Je2f*	Bradley unit of felsic metavolcanic rock. Rock is white to variably chloritic (green) and pyritic (rusty weathering) and has less than a few percent rounded quartz grains < 2mm. Clasts commonly are siliceous and angular (flattened or deformed). Je2c has polymict clasts > 5 cm. Je2f has clasts $\leq$ 2 cm and appears to be more feldspathic.	3	Glover (1994) Caron (2005)
Archi- bald Forma- tion	Ja*	Black, sulfidic, banded and massive argillite. Banded argillite has 1 to 10 mm light gray silty laminae, 1 to 4 mm black laminae, and some weak graded bedding. Banded units are 5 to 15 m thick. Massive argillite is aphanitic and in units 6 to 35 m thick.		Caron (2005) units Ja and Je1 Cheney, (2010)
Knob Hill Forma- tion	Pkq	Quartzite (metachert) ranging from aphanitic (tan, gray) to coarse-grained (white). Commonly intensely fractured and sulfidic. Minor intercalated argillite similar to PKa.		Caron (2005) unit M2
	Pka	Black, weakly foliated argillite with minor intercalated quartzite similar to Pkq. Commonly rusty weathering (sulfidic).		Caron (2005) unit M1
	Pkg	Fine-grained, gray-green, sub-phyllitic greenstone. Probably massive where not adjacent to Pks.	8	
	Pks and Pkl	Serpentinite (Pks) is aphanitic, gray to black, and very magnetic; with some relict pyroxene texture (bastite). Incipient listwanite causes apple-green to lemon-yellow flakes and surfaces. Pkl is pervasive listwantite, which commonly is gray, gneissic, and non-magnetic. Pkl is common along shear zones and adjacent to quartz veins. Pks and Pkl are commonly foliated near contacts.	2	

STOP 2 **Stop 2. Serpentinite and greenstone**

Admire this fine example of serpentinite (Jks of table 1). Variably slickensided, yellow-green surfaces of listwanite are diagnostic of serpentinite, but elsewhere in the area serpentinite is dense, very fine-grained, and black. All varieties of serpentinite are quite magnetic. Note our position and altitude with respect to Stop 1. The Chesaw fault must occur beneath the valley fill been Stops 1 and 2.

Walk back down the road to an outcrop of variably foliated greenstone with minor malachite (Je3 in table 1) on the east side of the road. The foliation is caused by the northwest-trending Grey Eagle fault (fig. 2).

Return downhill to the abandoned railroad grade. It may be possible to drive a short distance along the west side of the grade. In any case, walk north along the grade to an outcrop of felsic metavolcaniclastic rock.

STOP 3 **Stop 3. Felsic metavolcaniclastic rocks**

These felsic metavolcaniclastic rocks are Je2 in table 1. We are now below the Chesaw fault; so the generally southeastward dipping foliation may be tectonic rather than approximately parallel to the original bedding. Note that the rocks are pyritic and have flattened metavolcanic clasts. These are the types of felsic volcaniclastic rocks that might host the source of the lenses of massive sulfide and magnetite that occur near the base of the serpentinite in the Morning Star mine. The strike length of the felsic volcaniclastic rocks in the railroad cut is only 160 m. At 140 ppm Cu and 0.02 ppm Au, the rocks here are not geochemically anomalous. Walk another 150 m along the railroad grade where these rocks appear to be structurally underlain by the augite-bearing, clastic Elise Formation.

Return to SR 21 and turn left (south). This is MP 189.2. At a curve at MP 187.4, beware of on-coming traffic and pull out on the left shoulder.

STOP 4 **Stop 4. Limestone of the Brooklyn Formation, syenite dike**

Limestones occur in the middle part of the Brooklyn Formation and typically are clastic. On weathered surfaces some sand-sized clasts of quartz or chert weather "up", giving the rock a sandpaper-like feel. Other, generally darker clasts of carbonate weather like the matrix. Look for cm-sized clasts of red chert and > 5 cm clasts of limestone. The Brooklyn Formation is the most fossiliferous unit in this area of Quesnellia. Cheney et al. (1994) reported that elsewhere poorly preserved megafossil are Late Triassic (Norian); whereas, conodonts are Middle Triassic (Ladinian). Evidently, conodonts, which many geologists love and trust for paleontological dating, can be recycled (like heavy minerals).

The syenite dike here contains phenocrysts of two feldspars. It may be part of the Shasket Creek alkalic complex.

Continue southbound on SR 21 to MP 185.2 and park at the wide entrance to Little Goosmus Creek Road. Walk 0.2 miles north on SR 21.

STOP 5 **Stop 5. Sharpstone Conglomerate of the Brooklyn Formation**

This roadcut is in the lower part of the Brooklyn Formation. It consists of green siltstones and sandstones with interbeds of conglomerate. Clasts in the conglomerate are predominantly chert, and can be up to 12 cm long. The conglomerates have crudely graded bedding, indicating that these beds are overturned.

We can now contemplate the structure of the Quesnellian rocks. Figure 2 shows that SR 21 trends southwestward across a westerly dipping homocline. In this homocline, the Triassic lower Brooklyn Formation (Stop 5) overlies the middle Brooklyn Formation (Stop 4), which, in turn, overlies the Jurassic rocks of the Elise and Archibald formations in the area of the Morning Star mine. In other words, the entire homoclinal section is overturned.

In figure 2 note that the Archibald Formation appears to occupy the nose of a northerly plunging syncline. Drilling also found phyllitic argillites of the Archibald Formation beneath the serpentinite. Figure 3 is a schematic cross section though the Morning Star mine that illustrates the relationships recognized by geologic mapping and drilling. Specifically, the overturned homoclinal section is inferred to be the western limb of an overturned syncline.

Return to the vehicles and proceed back (northbound) on SR 2. Park in a very small turnout at MP 186.3 adjacent to the bend in the Kettle River. Walk 0.2 miles northward on SR 21.

STOP 6

Stop 6. Chesaw Thrust

Note the abundance of fragments of serpentinite along the northwest side of the road. This serpentinite marks the trace of the Chesaw fault. Serpentinite crops out over a vertical interval of 200 m on the hillside above the road but is bounded by Scatter Creek rocks to the north, west, and south. The base of the hill on the skyline to the northeast (across the river) is the serpentinite of the Morning Star mine area; whereas, the intervening topographic low is underlain by Brooklyn, Elise, and Scatter Creek rocks of the Danville window. Figure 2 shows that the window is open to the north, so that actually it is a half window. To the south, rocks of the window are truncated by the Granby River detachment fault on the western limb of the Kettle metamorphic core complex (Parker and Calkins, 1964); they also are unconformably overlain by Eocene rocks (Cheney and Rasmussen, 1996, fig. 2).

In summary, Stops 5 and 6 indicate that the westward dipping homocline is overturned below the Chesaw thrust (fig. 3).

Walk back south along SR 21 past where the vehicles are parked to the first outcrop on the west side of the road.

STOP 7

Stop 7. Rosetta Stone

Determine the lithology, metamorphic facies, probable mode of emplacement, ultimate origin, and possible age of this Rosetta Stone. Unfortunately, the next outcrop to the south is not much help. It is Scatter Creek, but it is extremely well jointed (fractured).

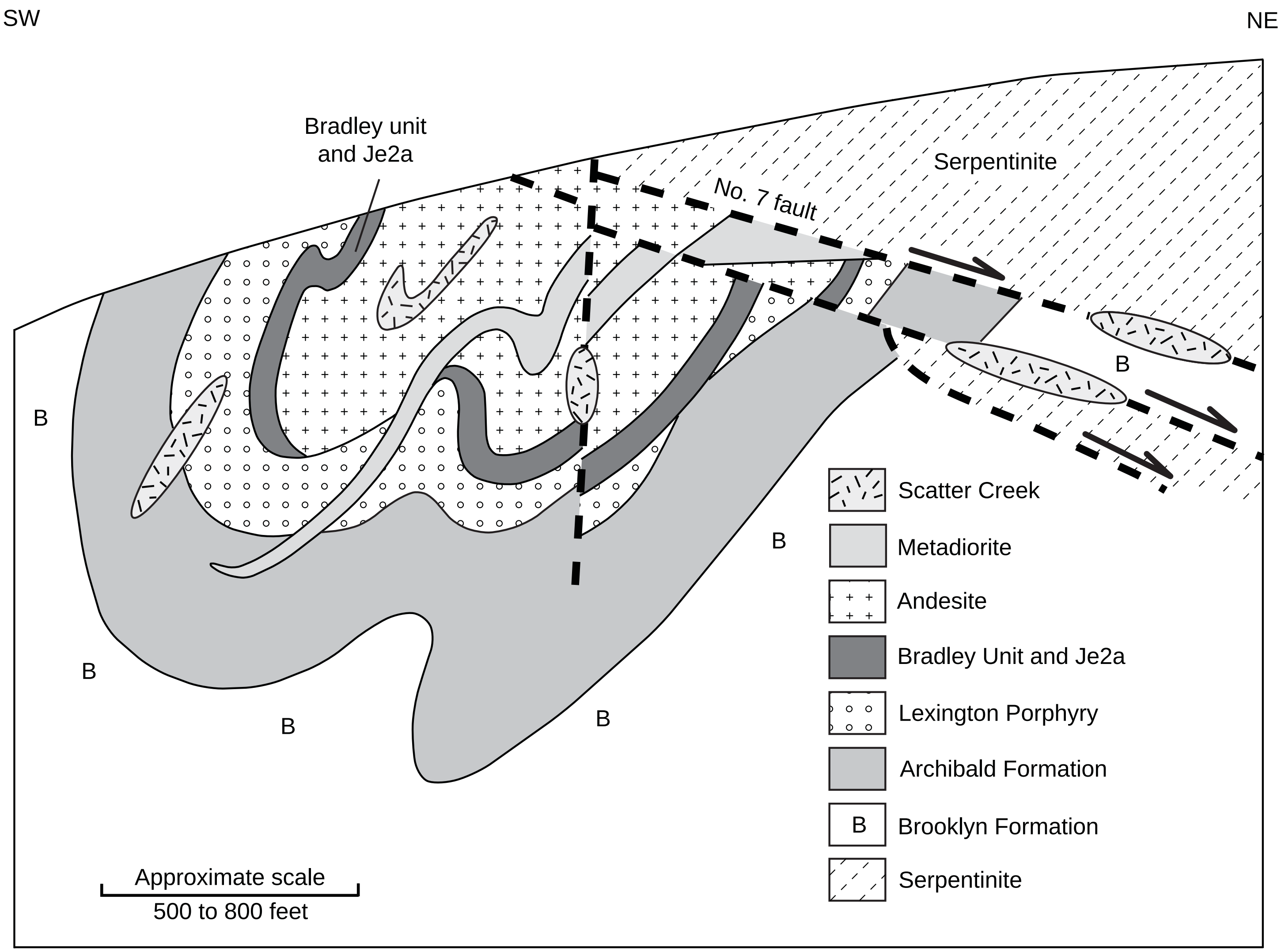

Figure 3. Schematic cross section of the area of the Morning Star mine.

Turn around and drive south past Curlew (MP 181.3). On the right (west) at MP 168.7 is the entrance to Curlew State Park. Continue 0.1 mile and turn left on Wolfcamp Road (MP 168.6). The surface plant and portal of the Lamefoot mine were located at this junction: admire the reclamation.

STOP 8 Stop 8. Lamefoot mine site

The Lamefoot mine is in a N-S belt of Quesnellian rocks about 8 km long and 1.5 to 3 km wide (Muessig, 1962). Figures 4, 5, and 6 show the geology of the Lamefoot deposit. Because the enclosing limestones have Triassic conodonts, they are part of the Brooklyn Formation (M.G. Rasmussen, 1998, personal communication). Figure 4 shows the geology of the main sulfide/oxide body. As already noted, most of the gold came from quartz veinlets crosscutting massive magnetite and pyrrhotite, or interbanded magnetite, jasper, and quartz plus carbonate. Additionally, significant amount of ore came from 1- to 20-cm wide veinlets of quartz ± pyrite ± chalcopyrite ± pyrrhotite ± arsenopyrite ± gold in the adjacent felsic (sericitic) volcaniclastic rocks to the east.

Because of the presence of Tertiary dikes, limestone, and magnetite, a similar deposit nearby was originally interpreted as a skarn (Lowe and Larson, 1996), as were the deposits at the Overlook and Lamefoot mines. However, figures 4 and 5 show that although the massive and banded parts of the mineralization at Lamefoot do have deformational shapes, they are strongly controlled by the stratigraphy. Additionally, the similar but overturned nearby Overlook deposit has sulfidic clasts in stratigraphically overlying strata (Rasmussen, et al, 1998), indicating that the sulfides formed early in the history of the deposit. So the massive parts of these deposits were chemically precipitated sediments in the Brooklyn Formation. At Overlook, D. E. Archibald (in Rasmussen et al., 1998) dated the gold mineralization as Jurassic (197 ± 3 Ma) by using $^{40}Ar/^{39}Ar$ to date sericite in envelopes around gold-bearing quartz veinlets. Evidently, the massive Triassic sulfide/magnetite and banded units acted as some sort of chemical (and or physical) trap that precipitated the gold during the Jurassic.

The different belts of limestone at Lamefoot were originally given different informal names, as indicated in figure 5, but all are now recognized as the Brooklyn limestone. The black area of the “Wardlaw” limestone

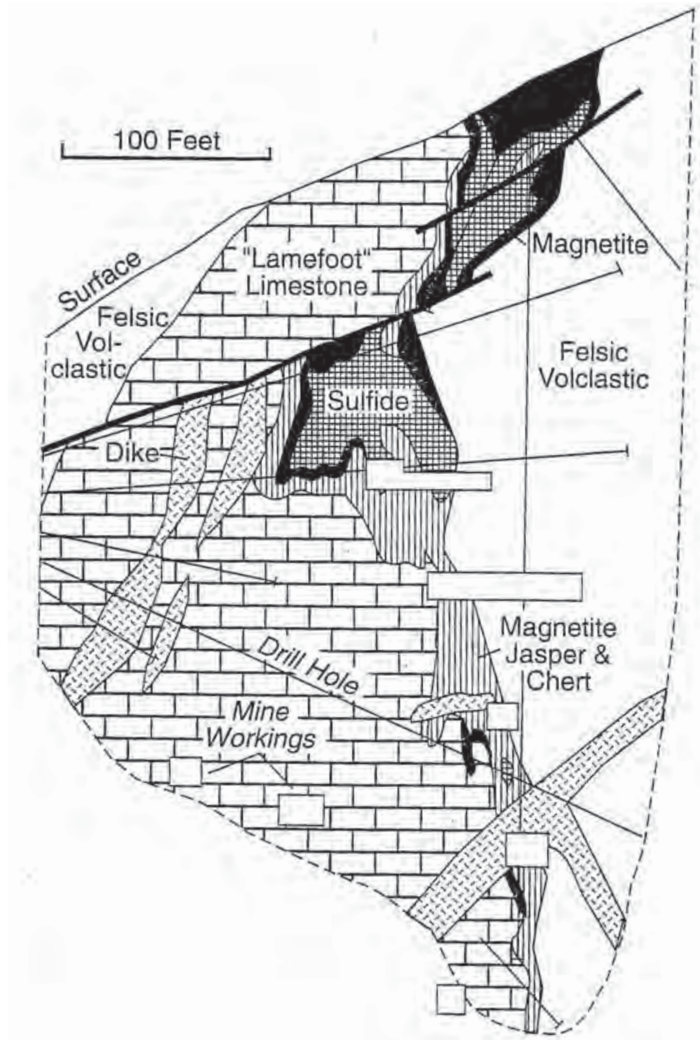

Figure 4. Interpretation of the Lamefoot deposit as a deformed volcanogenic massive sulfide/oxide deposit. The section was compiled by E.S. Cheney in 1996 along section 26700 N of the mine grid using data collected up to 11/19/1994 by geologists of Echo Bay Mines, Ltd.

in figure 5 is olistostromal, with fragments at least 0.3 m in diameter.

Note that in figure 5, the main sulfide/oxide body is thickened in the nose of an antiform that is an overturned, isoclinal syncline. M. G. Rasmussen (1998, personal communication) collected a specimen from overturned graded beds of pyrite (a “sulfide wacke”) in the nose of this fold.

Also note that in figure 5, the sandstones of the Eocene O’Brien Creek Formation dip vertically. Rotation of these beds clockwise back to horizontal shows that the isoclinal syncline was originally recumbent below the Chesaw thrust (fig. 6).

Continue 0.9 miles up Wolfcamp Road. The retaining walls of concrete blocks 0.2 to 0.6 miles up the

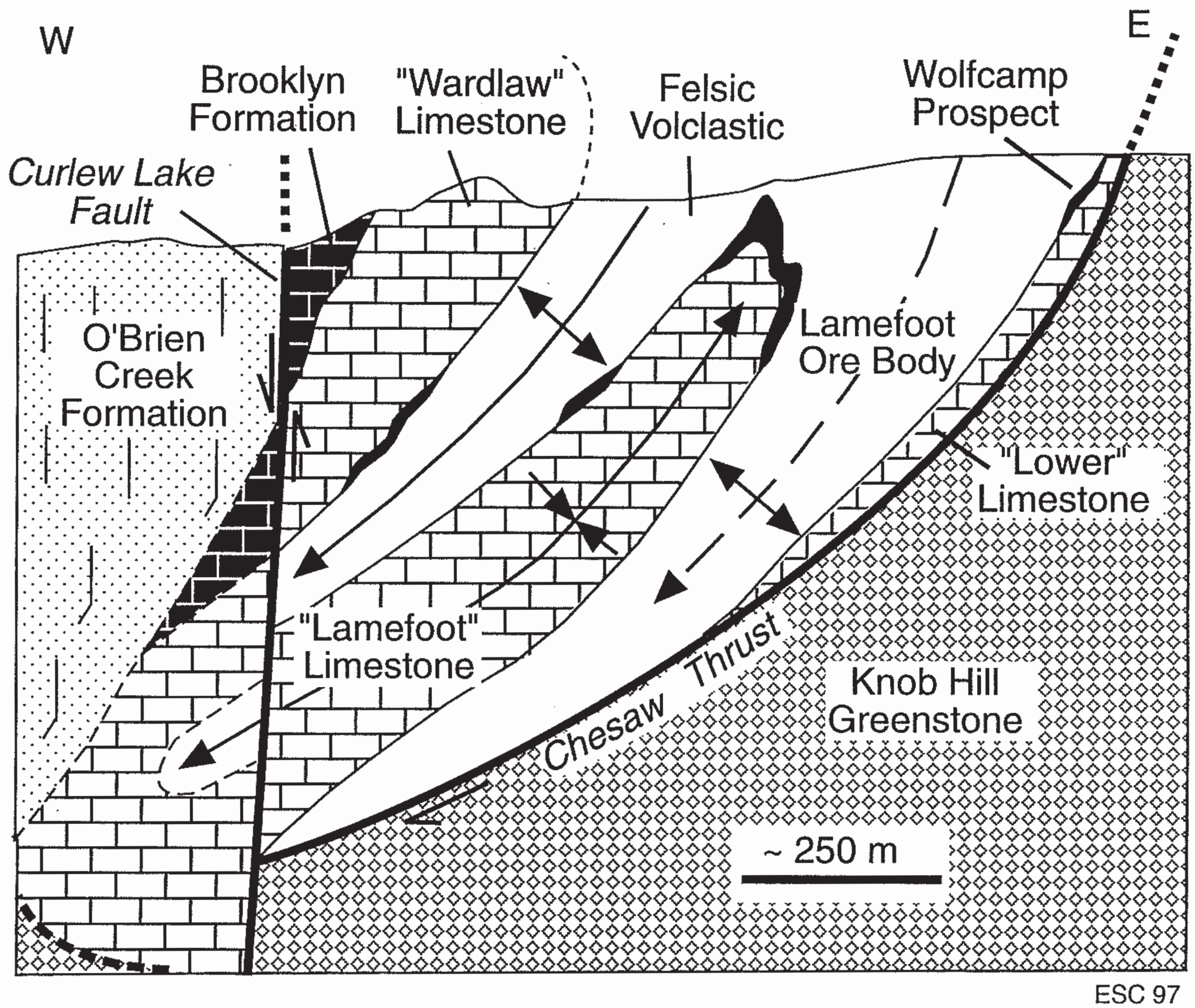

Figure 5. Schematic cross section of the Lamefoot deposit at about 26700 N. The section was compiled by E.S. Cheney in 1996 using data collected up to 11/19/94 by geologists and consultants of Echo Bay Mines, Ltd.

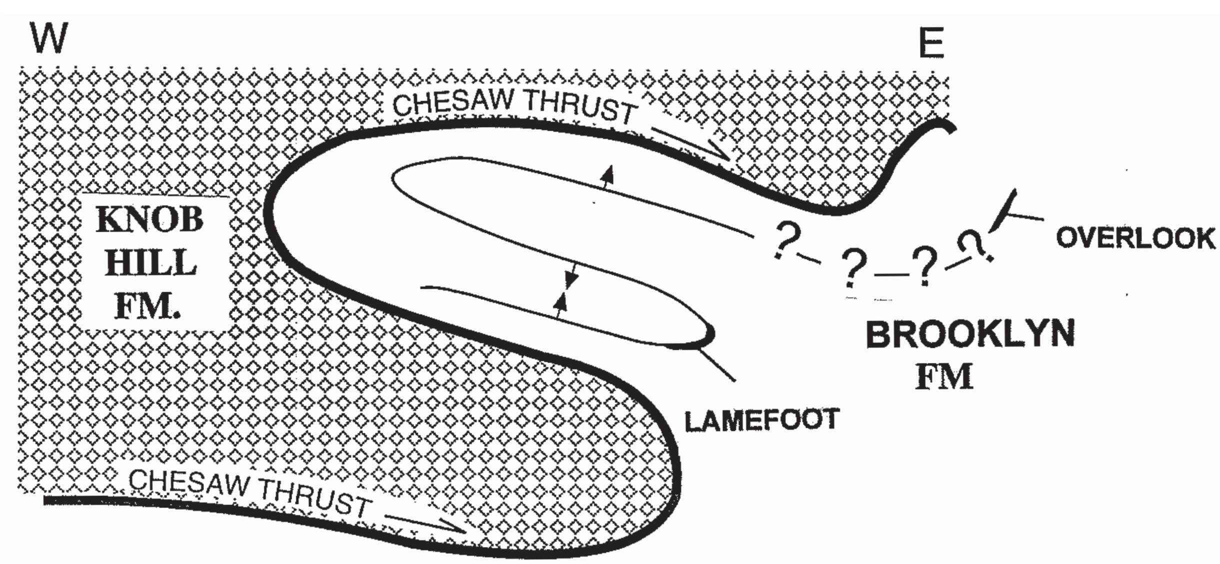

Figure 6. Schematic pre-Tertiary cross section of the Lamefoot-Overlook district as inferred by E.S. Cheney in 1996.

road were used in 1990 to 1991 to collar drill holes into the Lamefoot deposit. Some of these drill holes are shown in figure 4. The distinctly rusty-weathering rock on both sides of the road is the pyritic felsic metavolcanic rock shown in figures 4 and 5. Stop (at 0.9 miles) at a long roadcut where the road curves and starts uphill.

STOP 9

Stop 9. Knob Hill Greenstone and the Chesaw thrust

Start at the uphill end of the roadcut in typically nondescript Knob Hill greenstone. Walk southward back down the road noting (or even sketching) that the greenstone is progressively scalier. Minor interleaved pelitic phyllite subtly marks the Chesaw thrust. South of the greenstone is metalimestone, magnetite-bearing sulfidic rock, and scruffy outcrops of rusty weathering, sericitic volcaniclastic rock. Compare this traverse with the eastern portion of figure 5.

Turn the vehicles around and return to SR 21 (MP 168.6). Turn left (southbound) and at MP 166.5 turn right (west) on Herron Creek Road. At the T junction at 1.0 mile turn left (south) toward Republic. Bold outcrops in the next two miles are dikes of Scatter Creek and sandstones of the O'Brien Creek Formation (Muessig, 1962). Very gently dipping tan strata at 3.0 miles are the basal Tom Thumb Tuff Member of the Klondike Mountain Formation (Muessig, 1962). Continue another 0.6 miles to the intersection of Clark Avenue, the main Street of Republic.

At the intersection with Clark Avenue, go straight on to Knob Hill Road. In 0.1 mile on the right is the Stonerose fossil quarry (to dig purchase a ticket at the museum opposite the city park). At 0.7 miles the road enters Eureka Gulch, historically the main gold producing area of the Republic district. At 1.5 miles on the hillside to the right is the former site of the Knob Hill mine, the largest producer in the district. The mine produced about 2.5 million ounces of epithermal gold and three times as much silver from 1937 to 1995. Turn right on Swamp Creek Road 1.8 miles from here. In another 1.0 mile park at the junction of Barrett Creek and Swamp Creek roads. We are at the site marked Barrett on figure l.

STOP 10

Stop 10. Bacon Creek fault

The Bacon Creek fault is the eastern bounding detachment fault of the Okanogan metamorphic core complex. We will bushwhack up the hill to examine the fault, the view across the Sanpoil syncline, and, possibly, ticks on each other.

Limonitically weathering granitic rocks are on the northwest side of Swamp Creek Road. The limonite documents the former presence of pyrite. The presence of pyrite and the lack of mafic minerals indicate that the rock was hydrothermally altered, presumably by fluids moving along the fault.

Scramble up the hill to the base of the first big slab of granitic bedrock. Note that chlorite fills anastomosing fractures and the feldspar is pink. The rock has closely spaced joints. Locally this rock deserves to be called breccia. Identify the former mafic mineral(s).

Continue uphill along the left (south) side of the slabby bedrock. Traverse onto the slabs as curiosity and safety permit.

Continue over the crest of the hill until in unaltered (non-chloritic) rock. Note that neither the chloritic or non-chloritic rocks are mylonitic (foliated or lineated). This is common on the steeper limbs of metamorphic core complexes in this region.

Return to the crest of the hill and find the scenic viewpoint about 20 m northeast of the uphill traverse. Look below to the Sanpoil Volcanics on the southeastern side of the Barrett Creek Road Muessig, 1962). Estimate the minimum and maximum possible dip of the Bacon Creek fault at this locality. Using the minimum possible dip, estimate the minimum possible thickness of the zone of chloritic rock (perpendicular to the dip of the fault). The chloritic portions of detachment faults in Washington are so thin that they rarely crop out.

Remember the evidence for hydrothermal pyrite at the base of the hill, and note that the Knob Hill gold mine 3 km to the south is in the hanging wall of the fault.

From this viewpoint look northeastward across the Sanpoil syncline (mostly below 1,100 m) to the Kettle

Crest range (locally higher than 2,100 m) 15 km in the distance. The Kettle Crest is in the Kettle metamorphic core complex.

Descend the hill and examine the Sanpoil Volcanics (Muessig, 1962) on the far side of Barrett Creek Road. Determine whether the rhombic shapes in the outcrop are depositional (clasts) or fractures in a damaged zone adjacent to the fault.

This is the end of the field trip. Remove any ticks on your fellow participants.

CONCLUSIONS

Due to extensive cover and the lack of detailed geologic mapping, the lithologic, stratigraphic, and structural complexities of the Quesnellian rocks probably are greater than commonly realized. For example, felsic metavolcaniclastic rocks may be more numerous than currently appreciated.

The regional Chesaw fault displaces rocks as young as the Lexington porphyry (199.4 Ma). Perhaps, folding of this fault occurred soon afterwards in the mid Jurassic when Quesnellia accreted to Laurentia. The largest structural features in the area are Eocene antiformal metamorphic core complexes. They dominate the topography and restrict the Quesnellian rocks to the intervening Sanpoil syncline. Locally steep dips in Eocene strata may have been caused by movement on listric detachment and other faults.

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GEOLOGY OF THE PEND OREILLE MINE, METALINE FALLS, WA

Geology Department, Pend Oreille Mine, Teck Washington, Inc.

INTRODUCTION

The Pend Oreille Mine (POM) is a Mississippi Valley Type (MVT) deposit located in the Metaline District in northeastern Washington. The Metaline district is part of the Kootenay Arc, which extends from Washington into British Columbia. It consists of at least two main deposits, the Josephine and the Yellowhead. The Josephine is located along the contact between the Ordovician Ledbetter Slate and the Cambro-Ordovician Metaline Carbonates. The Josephine is characterized by brown to red sphalerite with low iron, due to little or no pyrite; a 3:1 zinc to lead ratio, and heterolithic, silicified (late) solution-collapse breccia, and it is often lower grade and less continuous than the Yellowhead. The Yellowhead horizon lies between the Upper and Middle Metaline. The Yellowhead is characterized by yellow to tan sphalerite, high pyrite content, a 5:1 zinc to lead ratio, and is rarely silicified. The host rock is a stratabound solution breccia, or unbrecciated, bedded dolomite with coarse, sparry dolomitic alteration. The breccia is almost always homolithic and zebra texture is common. Zig-zag vugs are commonly found in the hanging wall of the ore. A less explored deposit that is deeper in the mine is known as the Yellowhead 2.

General descriptions of the district geology, stratigraphy and structure are most recently presented by St. Marie and Kesler, 2000; Zieg et al. 2000; and Otto, 2001. The summary below is taken from these papers.

STRATIGRAPHY AND LITHOLOGY

The Metaline MVT zinc-lead district comprises several carbonate-hosted sulfide deposits within the Cambrian-Ordovician Metaline Formation. The Metaline Formation is part of the Kootenay arc, a structural northeast-trending belt extending from northern Washington into south central British Columbia (fig. 1). The Kootenay arc contains a sequence of platform sedimentary rocks extending from the Upper Proterozoic to Upper Paleozoic and is interpreted as the westernmost extent of the autochthonous North American craton.

The Metaline Formation is 3,000 to 5,000 feet thick and comprises massive to bedded dolomite with lesser amounts of limestone. Its internal stratigraphy is poorly understood (figs. 2 and 3). The Metaline Formation is, arguably, conformably underlain by the Cambrian Maitlen Phyllite and early Cambrian Gypsy Quartzite. The Metaline Formation is conformably overlain by Ordovician Ledbetter Slate. The Ledbetter Slate comprises distinctly black, carbonaceous argillite and slate.

There is no clear agreement on the vertical or lateral spatial variations of many of the carbonate lithofacies within the Metaline Formation, due to the subtle nature of many primary sedimentary features and the variable effects of early-late diagenetic and weak to intense hydrothermal dolomite alteration and solution collapse. The Metaline Formation was originally subdivided into three members by Dings and Whitebread (1965):

- Upper Metaline Formation (~200-1,550 feet thick) comprising massive, grey limestone including subtidal bioturbated lime mudstone and intraclastic peloidal packstone lithofacies, the Josephine horizon (breccia) and the Fish Creek breccia;
- Middle Metaline Formation (~3,800 feet thick) comprising predominantly bedded, fine-grained subtidal dolostone and lesser peritidal dolomitic shale lithofacies; and
- Lower Metaline Formation (~950 feet thick) comprising thin bedded, dark, carbonaceous, subtidal lime mudstone (limestone), and shale lithofacies.

Later workers (Bush et al., 1992) have subdivided the Metaline Formation into a number of laterally equivalent lithofacies which are complicated by a number of alteration facies described as “varied dolostone” (fig. 4).

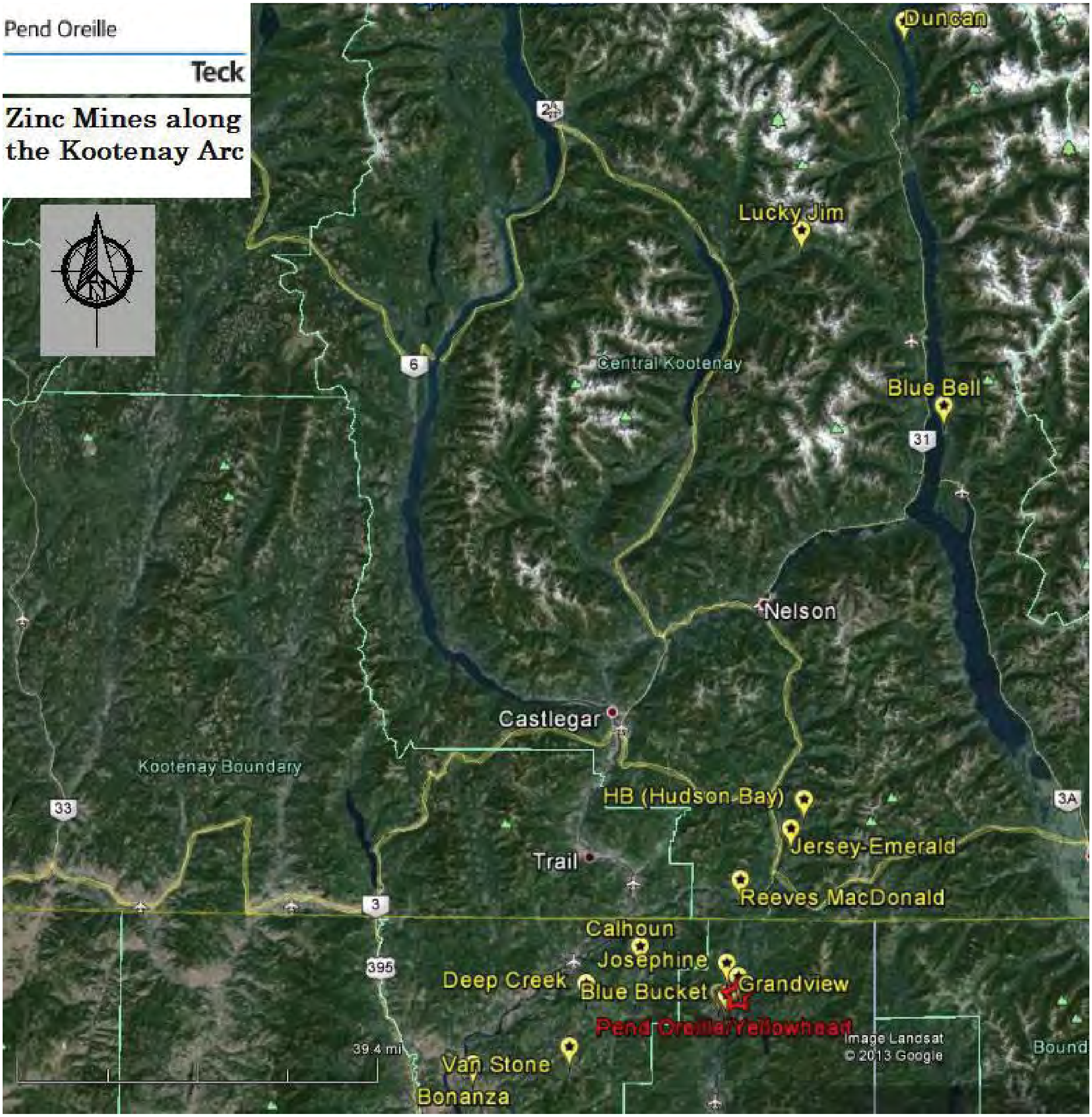

Figure 1. Air Photo of the Kootenay Arc showing zinc mines with >100,000 tons production (Google Earth).

ALTERATION AND STRUCTURE

Dolomitization, primarily in the Middle and Upper Metaline Formation, is by far the dominant alteration type in the district and has been subdivided into three to five stages by different workers (Fisher, 1981, 1988; Bending, 1983). Early diagenetic dolomitization is typically patchy to pervasive, which produces fine-grained crystalline dolomite. This is overprinted by pervasive medium and late coarse-grained (up to 0.5 cm), white, sparry and saddle dolomite (a variety of dolomite that has a warped crystal lattice characterized by curved crystal faces and cleavage) that is locally zebra textured (showing conspicuous banding, generally parallel to bedding, consisting of light-gray, coarsely textured layers alternating with darker finely textured layers.)

Several phases of late dolomitization are evident as are multiple episodes of solution collapse breccia-

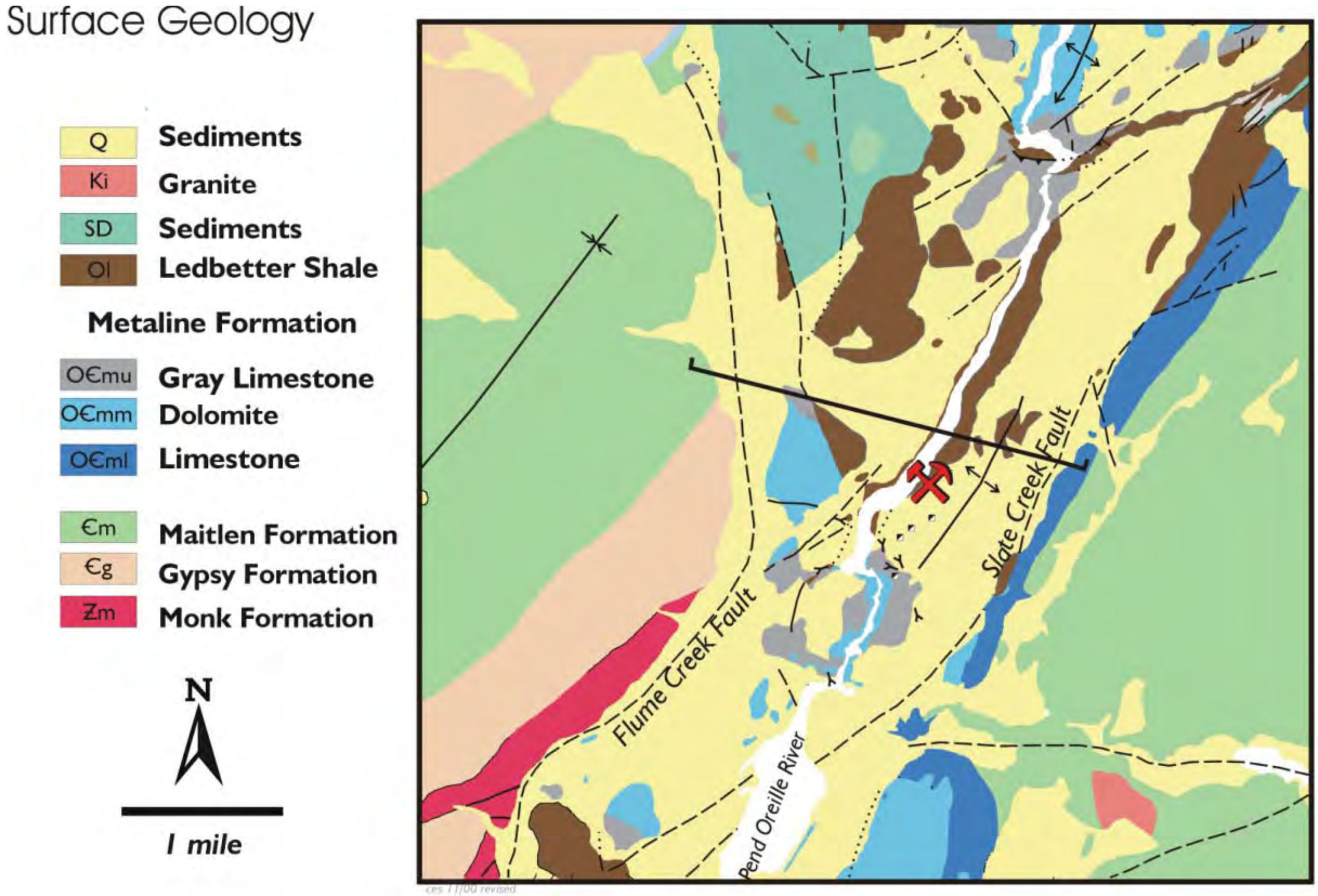

Figure 2. Detailed geologic map of the immediate Metaline District centered on the Pend Oreille Mine (Teck American, 2001).

tion, extending from pre-diagenesis collapse of limestone to late stage re-brecciation of late stage dolomite and accompanying mineralization. Late silicification (jasperoid) in the Josephine breccia and the precipitation in open spaces of fine-grained, drusy quartz are thought to have occurred late in the sequence of alteration. The latest alteration event formed a matrix of coarsely-crystalline calcite. Re-brecciation of the Josephine breccia is common along late faults (tectonic breccias?).

Deformation within the Kootenay arc initiated with compression during the Devonian Antler orogeny and culminated in the late Jurassic to Cretaceous Nevadan orogeny, producing the prominent northeast-trending folds present today. Granitic stocks and batholiths, including the Jurassic Nelson batholith and the Cretaceous Kaniksu batholith, were intruded contemporaneously with the deformation. Rock in the Metaline district is metamorphosed to greenschist grade characterized by muscovite-chlorite. Tertiary extension is reflected by steeply dipping, NE and WNW-striking, normal faults, which locally cut Eocene dikes. These faults, coupled with pre-existing, shallow-dipping planes of weakness, such as bedding, clearly demonstrate the most recent period of deformation. This latest stress regime (Eocene) resulted in reactivation of many earlier-formed structures oriented north and northeastward, and is discernible as the dominant structural feature in the district.

MINERALIZATION

The Metaline District hosts two distinct styles of carbonate-hosted mineralization: 1) breccia-hosted (Josephine-type) mineralization and 2) stratabound replacement (Yellowhead-type) mineralization (Zieg et al., 2000).

The Josephine breccia (host to Josephine-type mineralization) ranges up to 250 feet in thickness and occurs at the top of the Upper Metaline Formation. The breccia is heterolithic, varying depending on the nature of the enclosing stratigraphy and degree of alteration. Framework- to matrix-supported breccia clasts consist of angular to subrounded dolomite, chert, quartzite, and black argillite, in a matrix of

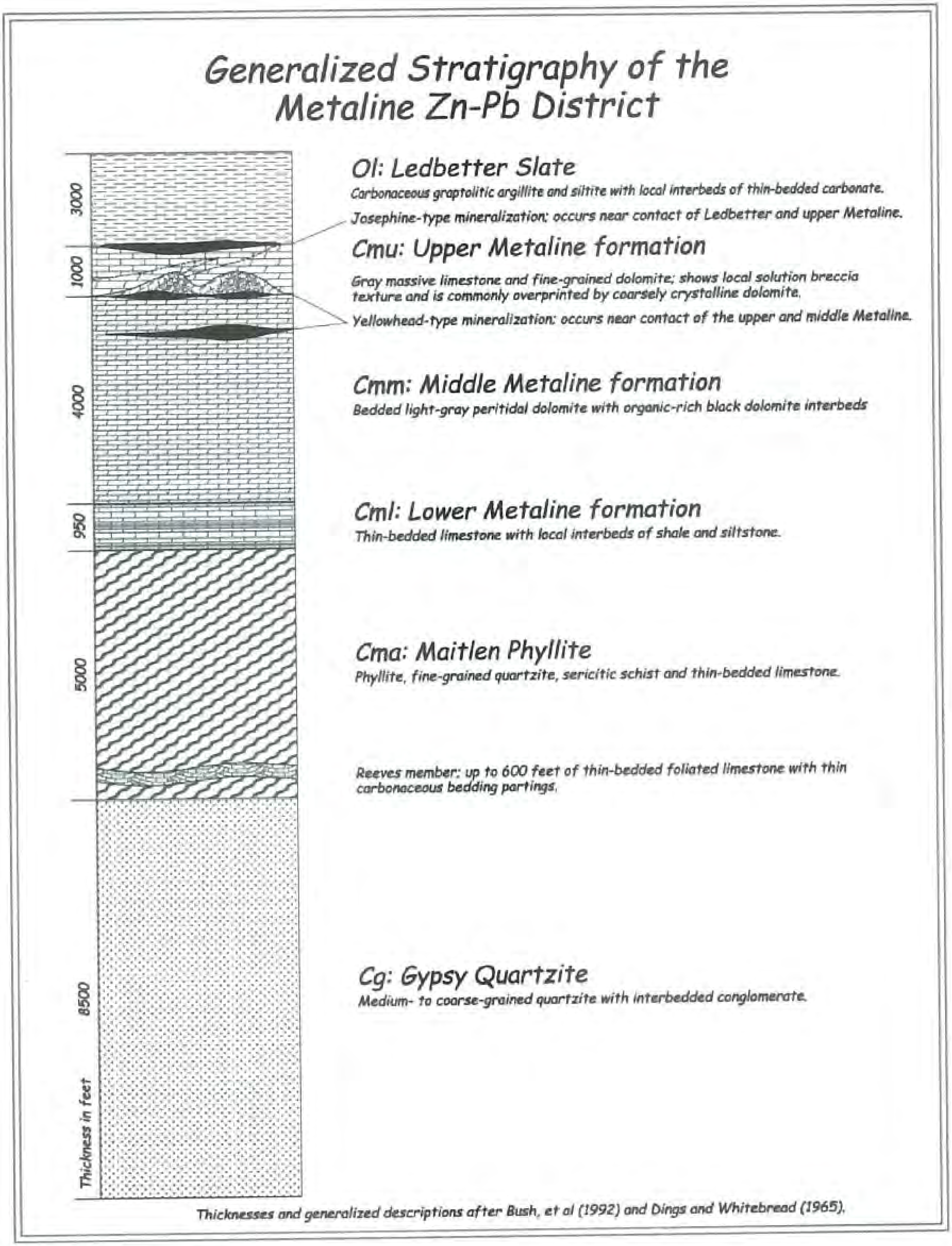

Figure 3. Schematic stratigraphic column (Otto, 2001).

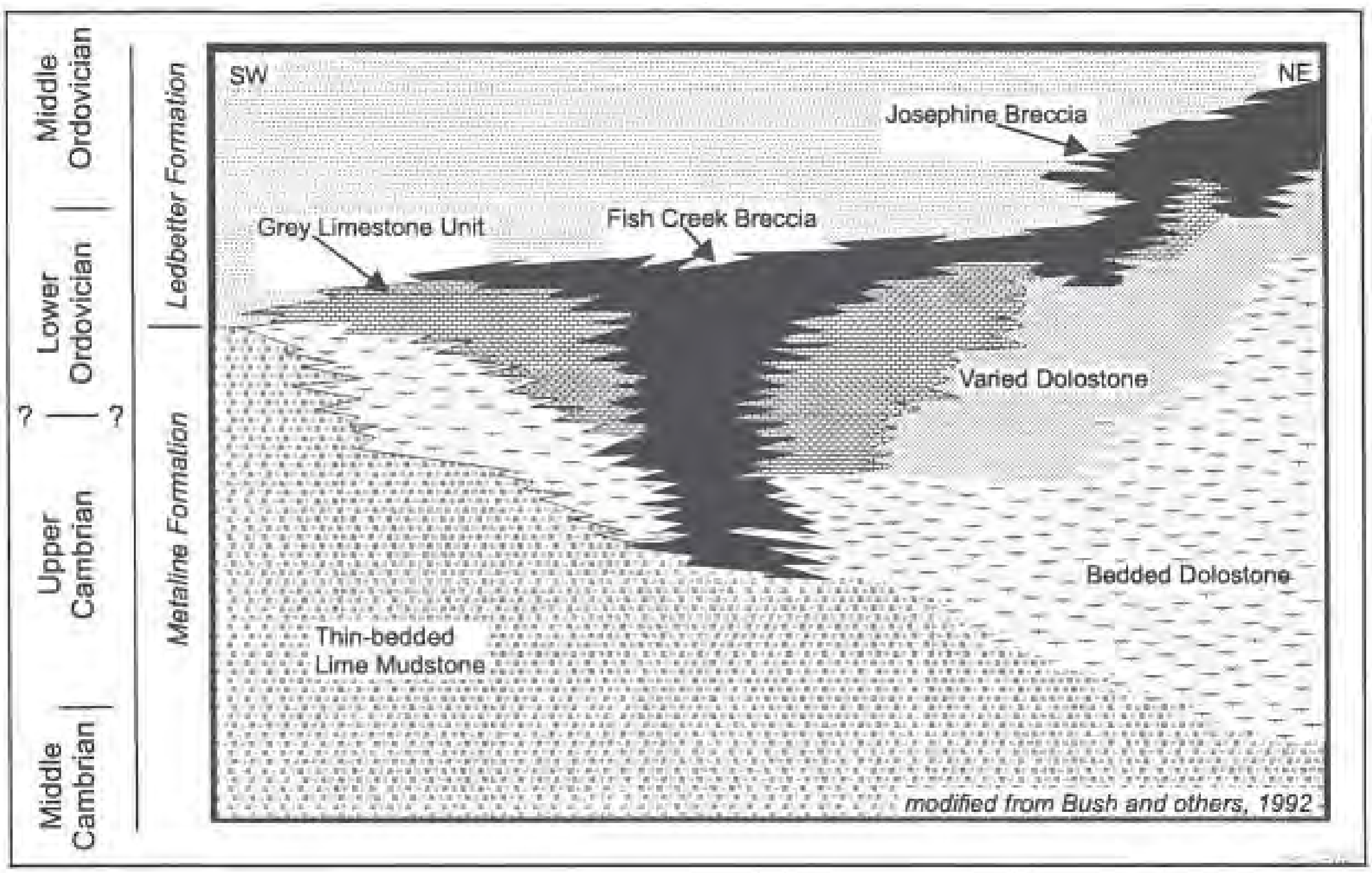

Figure 4. Schematic cross section showing relationships among lithofacies in the Metaline Formation (Zieg et al., 2000).

black, fine-grained rock fragments and insoluble residue. Internal sediments are common as pockets or layers within the breccia bodies. Very similar breccias occur in the Middle Metaline above and below the Yellowhead horizons. Josephine breccias are now interpreted to have formed by multiple solution collapse events (Zieg et al., 2000).

Josephine-type mineralization occurs as irregular, fine- to coarse-grained, reddish brown sphalerite and galena (~3:1 ratio) disseminations and banded masses or pods occurring as clast rims, matrix replacement, and late breccia infill. At least three paragenetic stages of sphalerite mineralization are recognized, with later sphalerite typically being yellow, banded to colloform in texture and associated with late quartz-calcite-barite alteration. Mineralized bodies can appear pipe-like in form, up to 3,000 feet long, 300 feet wide, and up to 100 feet thick, and show some control by minor, mine scale, NE and NNE-trending structures.

Yellowhead-type mineralization occurs in at least two significant "horizons" (Yellowhead 1 and 2) in the district at or near the transition from the Middle Metaline to Upper Metaline Formation. Mineralization is interpreted to be hosted by stratabound zones of dissolution and solution collapse breccias within the bedded dolomite. Within the Yellowhead, rocks are overprinted by intense dolomitization and sulfide mineralization to the point where original textures are obscured or destroyed. Distinctive features common in the hanging wall of the Yellowhead horizon called "concentrically banded dolomite" (or, zig-zag vugs) are thought to reflect evidence of "cavern voids" or fluid pathways now in-filled with successive generations of dolomite. Other common textures include internal sediments and several varieties of zebra dolomite.

Unlike the Josephine, Yellowhead-type mineralization is iron-rich, occurring as tabular replacement "horizons" with brecciated textures or wispy laminated to colloform banded textures, and comprising fine- to medium-grained pyrite-sphalerite. Sphalerite is typically pale tan to yellow and pink. Pyrite (after early marcasite) is more widespread than sphalerite, which is generally confined to central and lower portions of the sulfide zones as well as barren zones less than five feet thick that are adjacent to, but along the same horizon, as the Zn-Pb-rich zones. Galena developed later and is generally coarser-grained. Multiple phases of mineralization suggest multiple events with significant amounts of carbonate dissolution and replacement.

At the Pend Oreille Mine, the Main Yellowhead 1 (YH1) horizon is found 600-900 feet below the Josephine breccia and occurs as a linear, northeast-trending, 300-400 foot wide zone comprising at least five en echelon zones (Southwest, Central, Northeast, Warren, and Kinney zones), extending over 8,500 feet along strike. Thickness for the resource averages about 17 feet, but can be up to 60 feet. A strong, "local" structural control on mineralization is apparent and has been defined through detailed underground geologic mapping and grade distribution plots.

A second, lower Yellowhead-type horizon (YH2) is well defined in some areas, but has only been partially drill tested in areas below the POM Main YH1 horizon. In all occurrences, the lower YH2 horizon appears to be almost directly below YH1 mineralization suggesting similar structural controls.

Ore textures and mineral chemistry in both types of mineralization indicate strong non-equilibrium among adjacent minerals. Although fluid inclusion homogenization and freezing temperatures are similar for both types of ore; fluid inclusion leachate and gas analyses differ for each mineral (St. Marie and Kesler, 2000). These data all suggest that ore minerals were deposited from two parent brines that changed composition slightly to deposit each mineral. Isotopic analyses indicate that both types of mineralization had the same sulfur source (reduced seawater) while lead (and by inference zinc) also came from two distinct sources. These observations indicate that ore formed when two parent metal-rich brines mixed with the same source of sulfur.

The typical colloform to botryoidal textures of sphalerite and pyrite in Yellowhead-type mineralization suggest rapid deposition from supersaturated solutions. In contrast, the generally coarse-grained nature of Josephine-type mineralization is more suggestive of growth from less saturated solutions. Levels of saturation are attributed to the H_2S content, requiring higher levels in the Yellowhead necessary to produce saturated fluids.

Therefore, Yellowhead-type mineralization required an acidic, low S, metal (Fe-Zn-Pb)-rich, hot (~160-180˚C) brine which mixed with a H_2S-rich brine (or gas). In contrast, data suggests that Josephine-type mineralization formed from an acidic, moderate S, metal (Zn-Pb)-rich, cooler (~100-160˚C) fluid which mixed with H_2S-bearing modified seawaters from the overlying Ledbetter shale (?). Temperature and S content are thought to control Fe content.

SUMMARY

The deposit at the Pend Oreille Mine is hosted bya the carbonate sequence of the Metaline Formation. Several other zinc-lead deposits exist along the Kootenay Arc. The trends indicate local structural controls of MVT deposits. Several horizons exist along the northeasterly trend at the Pend Oreille. The upper Josephine horizon, which has low Fe content, is found along the Upper Metaline-Ledbetter contact with different characteristics than the lower horizon, the Yellowhead, which is Fe rich and is found along the transition between the Upper and Middle Metaline. The different characteristics and mineral chemical studies indicate different brine compositions and temperatures, but the same sulfur source.

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